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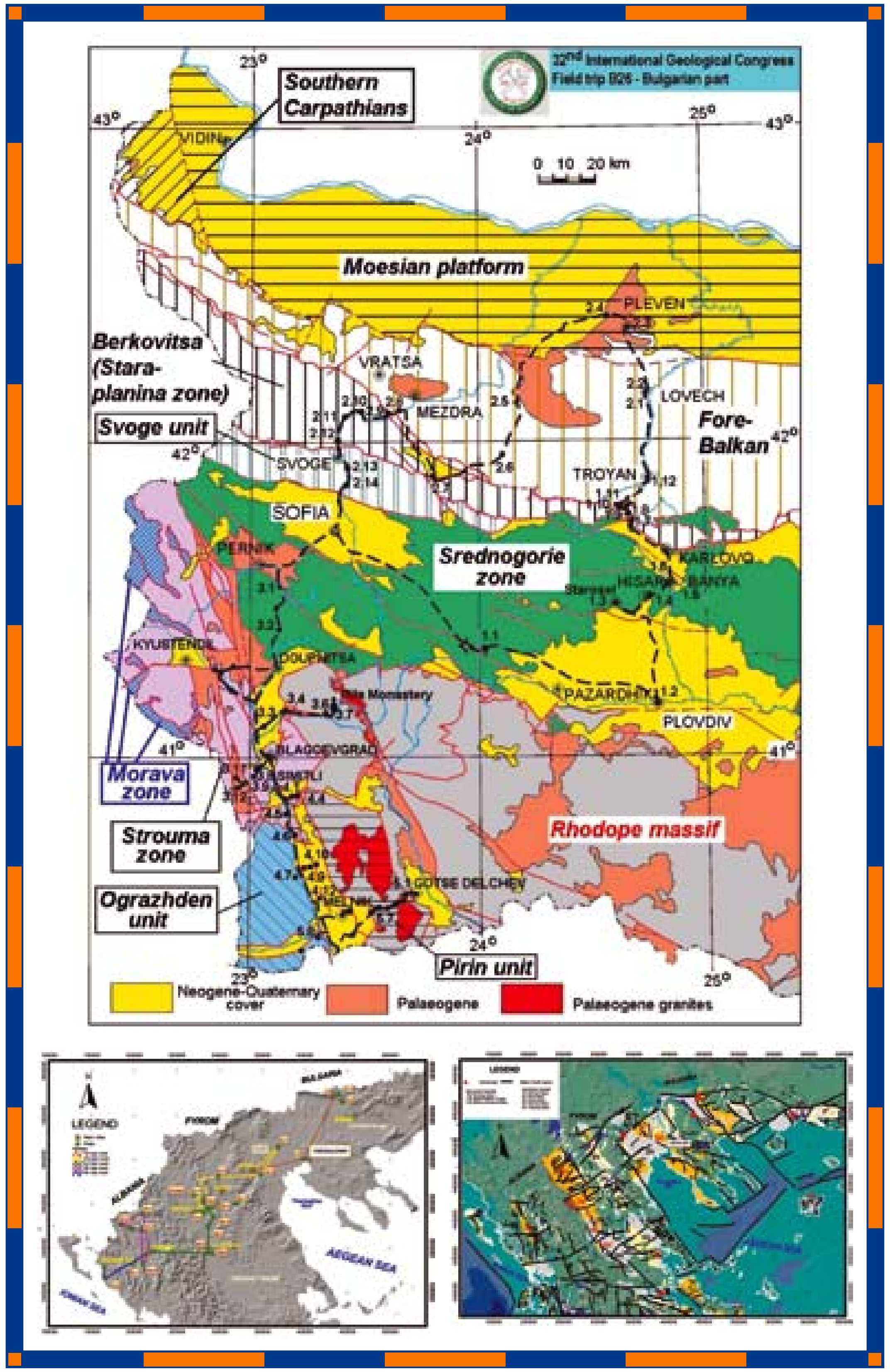
FIELD TRIP MAP



32nd INTERNATIONAL GEOLOGICAL CONGRESS

Edited by APAT

Field Trip Guide Book - B26

Florence - Italy
August 20-28, 2004

Volume n° 2 - from B16 to B33

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NEOTECTONIC TRANSEPT MOESIA-ABULIA

Leader:
I. Mariolakos

Associate Leaders:
I. Zagorchev, I. Fountoulis, M. Ivanov

Pre-Congress **B26**

The scientific content of this guide is under the total responsibility of the Authors

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field trip itinerary on Bulgarian and Greek territory

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Contents

Foreword

The field trip B26 was conceived by I. Mariolakos and I. Zagorchev as a joint initiative of Bulgarian and Greek geologists with the aim of demonstrating to the international geological community the main features of the Alpine geology of the Balkan Peninsula from a standpoint of the extensional tectonics in Neogene and Quaternary times. The preparatory work had reached its final phase when it was cruelly interrupted by the sudden death of Dimitris Mariolakos, - a bright young geologist full of vitality and enthusiasm. We, the colleagues and friends of Professor Mariolakos, wish to express again to him and his family our deep sympathy. This guidebook is dedicated to the memory of Dimitris.

Introduction

The field trip aims to present a comprehensive view over the principal neotectonic features of the Balkan Peninsula. This general idea is placed within the larger setting of the Alpine structures: the Alpine fold belts (Balkanides, Hellenides) and platforms (Moesian and Apulian) in their plate-tectonic evolution. Stress is laid also on the problems of the geological and cultural heritage.The field trip starts in Sofia, following the Maritsa fault belt east to Plovdiv, and then it crosses the principal tectonic units of the Balkanides travelling northwards to the Moesian platform. Afterwards, the itinerary crosses the Balkanides again to Sofia, and from there continues south through the main units at the margin between the Serbo-Macedonian and the Rhodope (*s.l.*) massif along the Strouma/Strymon fault belt. Then it crosses the Serbo-Macedonian massif, the Vardar/Axios zone, and the principal zones of the Hellenides to the Apulian platform. Important neotectonic structures are observed and broadly discussed, including active faults and related seismic and geotechnical hazards.

Principal field references
Geological map of Bulgaria, M 1:500,000
Geological map of Greece, M 1:500,000
Geological map of Greece, M 1:1,000,000
Geological map of Bulgaria, M 1:100,000
Geological map of Greece, M 1:50,000
Road maps of Bulgaria and Greece, M 1:500,000
Zagorchev, I. 1995. *Pirin. Geological Guidebook*. Academic Publishing House, Sofia.
The geological maps are available at the geological surveys of Bulgaria (Direction of Geology and Mineral Resources, Ministry of Environment and Waters, Sofia) and Greece (I.G.M.E. – Institute for Geology and Mineral Exploration, Athens).

Regional geologic setting

Alpine structure

The Balkan Peninsula has a complex geological structure made up of Precambrian, Cadomian, Hercynian (Variscan), Early Alpine (Cimmerian),and Alpine *s.s.* tectonic movements. Although the field trip mostly demonstrates the structures formed during the extensional collapse of the Late Alpine orogen in Palaeogene and Neogene times, it will demonstrate also some of the older features.

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The Alpine structure of the peninsula (Figure 1), is a result of the collisions between the Eurasian and the African continental plates and the closure of the Tethyan Ocean. It is dominated by two principal elements: the metamorphic crustal fragments, and the ophiolites issued from closed oceans. The metamorphic crustal fragments (from the North to the South) are as follows: Sredna-gora crystalline block; Rhodope massif, Pirin-Pangaion unit included; Osogovo-Lisets and Lisiya fragments in the core of the Strouma unit; pre-Ordovician fragments in the Morava unit; Serbo-Macedonian massif; Pelagonian massif. These crustal fragments have mostly a Precambrian age, and were amalgamated and reworked in Cadomian times. In some of the units they are covered unconformably by Palaeozoic (Ordovician and younger) sedimentary formations, and are intruded by Hercynian and Alpine granitoids.

The Alpine structure is divided into two parts by the Axios (Vardar) suture zone.

To the ENE of this zone, the Alpine structure is characterized by pre-Alpine and Alpine continental crust of the Moesian platform and its southern margin, with the development of arc basins and rifts. Along the Axios/Vardar zone, and to the WSW towards the Apulian platform, the Alpine evolution bears a pronounced Tethyan signature, with well-preserved remnants of the Tethyan Ocean.

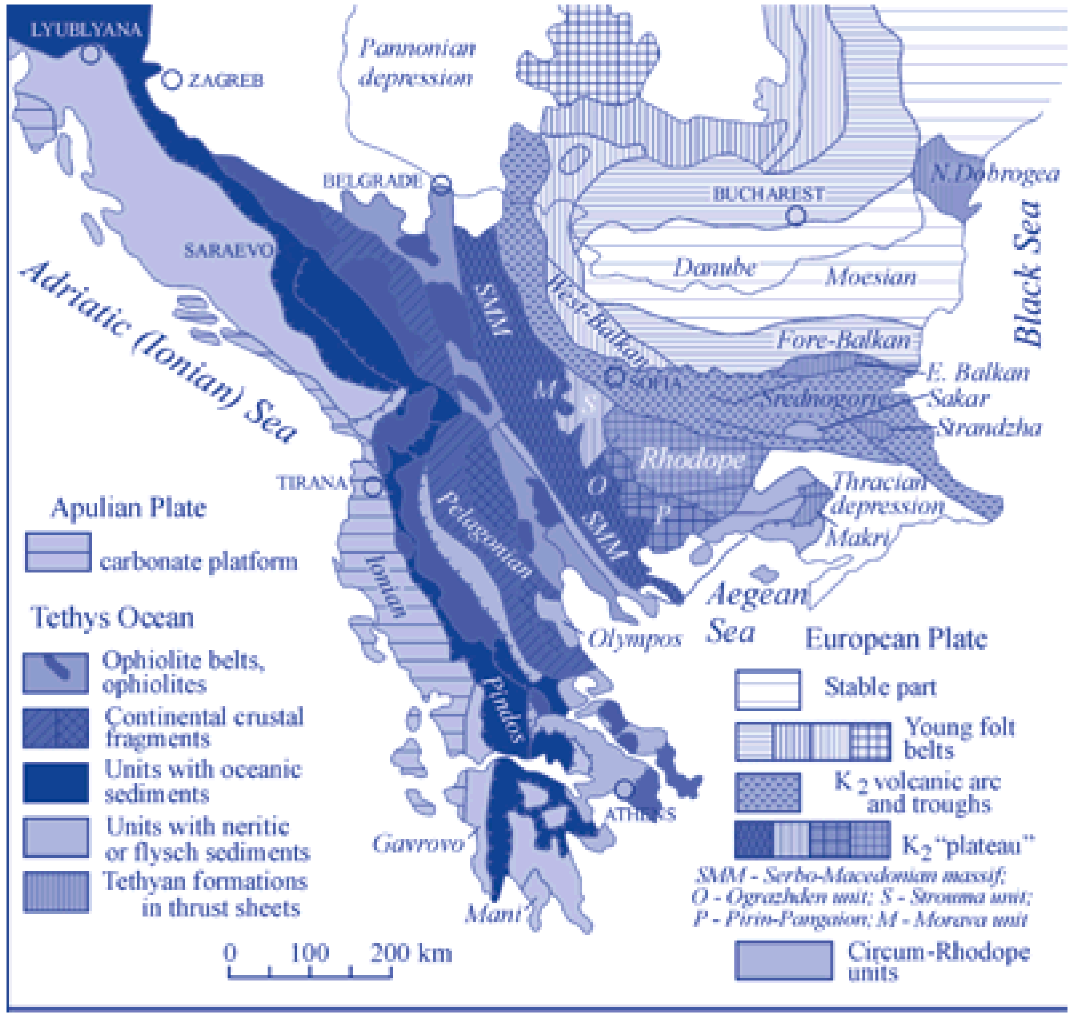


Figure 1 - Tectonic sketch map of the Balkan Peninsula (after Zagorchev, 1996, 1998)

Alpine structure of the eastern parts of the Balkan Peninsula

(Moesia – Serbo-Macedonian massif)

I. Zagorchev

The Moesian platform is geomorphically a low plain (Danube plain) with Quaternary loess cover. The oldest sedimentary complexes (Precambrian, on Romanian territory, and Devonian, in Bulgaria), are crossed by boreholes. Uplift episodes and partial erosion are reflected in several regional unconformities which occur between the Upper Permian and Lower Triassic; in the Lower Jurassic; between the Lower and Middle Cretaceous; and several unconformities in the Palaeogene and Neogene. The cover consists of continental Lower Triassic red beds, followed by Middle Triassic to Norian platform carbonates, and Carnian to Norian marine red beds; terrigenous and carbonate Jurassic, the marine transgression over the deeply eroded Triassic, beginning often directly with Middle Jurassic; Middle and Upper Jurassic and Lower Cretaceous carbonate platform sequences; Upper Cretaceous shallow-marine carbonate and terrigenous deposits, and continental or marine Palaeogene; and local Badenian to Pontian Neogene marine sediments belonging to a transitional zone between the Central and Eastern Paratethys.

The Fore-Balkan (transitional zone between the Balkan fold belt and the Moesian platform) is characterized by a low degree of deformation, and by transitional Alpine facies. In the eastern parts of the zone, the Triassic and Jurassic rocks are deeply buried beneath the Cretaceous and Palaeogene (partial) cover.

The Balkan (Stara-planina) zone is characterized by north-verging thrusting and folding in pre-Late Eocene and latest Oligocene times. The separate units (thrust sheets and anticlinoria) include fragments of units of Mid-Cretaceous and Late Cretaceous folding. The following complexes are distinguished: pre-Ordovician volcano-sedimentary complexes of oceanic (partially) and island-arc signature; Palaeozoic (Ordovician to Carboniferous) sedimentary formations intruded by Carboniferous granitoids; Permian red beds; a Triassic Peri-Tethyan sequence, ending with folding and covered unconformably by a Lower Jurassic to Kimmeridgian Peri-Tethyan sequence; an Upper Jurassic - Lower Cretaceous carbonate sequence (carbonate platform); an Upper Cretaceous shallow-water carbonate sequence; Danian to Middle Eocene continental and shallow-marine sediments; unconformable Upper Eocene molasse.

The Srednogorie zone has been set in by crustal necking and rifting in the southern edge of the Moesian platform, after the Mid Cretaceous compressive event, and developed as a volcanic island arc of Late Cretaceous age. It was transformed into an orogen by north-verging folding and thrusting in latest Cretaceous times, and later thrusted northwards over the Stara-planina zone. The Upper Cretaceous covers with unconformable depositional contact the following different complexes: slivers and rootless bodies of ultramafic (ophiolitic?) character and mantle (or oceanic crust) origin (parts of the Precambrian Prerhodopian Supergroup); the principal part of the Precambrian Prerhodopian Supergroup - a typical continental crust of granitic composition (migmatites, gneisses, metagranites); granitoid plutons of Carboniferous age (determined with Rb-Sr whole-rock isochrons), and mixed (I-type) and continental crustal signatures; Permian, Triassic and Jurassic sedimentary complexes, with Peri-Tethyan (Balkanide) signature. The Upper Cretaceous sedimentary formations are represented by coal-bearing continental to shallow-marine Cenomanian and Turonian, terrigenous to carbonate marine Coniacian, pelagic carbonates and shales, locally with radiolarites (Santonian – Campanian), and Lower-Middle Maastrichtian flysch. The Cenomanian? to Lower Maastrichtian volcanics (from pikrites and basanites to andesites and latites) form a volcano-plutonic association, with gabbro to monzonites and granodiorites of mantle signature. The Late Cretaceous folding, uplift and erosion was followed by a Danian conglomeratic formation of fluviolacustrine origin, that contains well-rounded pebbles from all older rocks. Upper Eocene and younger formations fill in syn- and post-sedimentation grabens.

The Morava-Rhodope zone embraces the tectonic units between the Srednogorie and the Axios zone. They represented in Late Cretaceous times a frontal arc ("plateau") built up of Mid-Cretaceous tectonic units. The Morava-Rhodope zone consists of several units: Rhodope *s.l.* (Rila-West Rhodope and Pirin-Pangaion subunits included), Strouma, Morava and the Serbo-Macedonian massif.

The Rhodope massif is a lenticular body of continental crust. The metamorphic complexes have been and still are the subject of discussions and controversial interpretations (Zagorchev, 1998). Two metamorphic complexes (Prerhodopian and Rhodopian Supergroup)

Volume n° 2 - from B16 to B33

are distinguished. The Prerhodopian Supergroup represents a continental crust of granitic composition (migmatites, gneisses, metagranites), that contains amalgamated rootless bodies of ultrabasic and basic rocks. The Rhodopian Supergroup originated from products (flysch-like sedimentary sequences, volcanogenic sequences, platform carbonates) of a Precambrian island arc. Both supergroups suffered Cadomian reworking, and were intruded by igneous bodies of granitic composition and Palaeozoic and Late Cretaceous age, in most cases bearing the signature of recycled and molten older continental crust. The metamorphic complexes are covered by non-metamorphic sedimentary and volcanic cover of Palaeogene age. Other hypotheses consider the Rhodope metamorphic complexes as having been formed during accretion of crustal and mantle fragments with Palaeozoic and Mesozoic ages that suffered Alpine deformations and metamorphism.

The basement of the *Strouma unit* consists of various gneisses, schists, and amphibolites (including orthoamphibolites) of Precambrian age and Cadomian reworking. The Neoproterozoic - Cambrian Frolosh Formation (metadiabases, schists, metapsammites), contain rootless bodies of mafics and ultramafics (lherzolite, troctolite, gabbro, and norite) and is intruded by the Strouma diorite formation (gabbrodiorite to granite). They are regarded as a volcanic arc association that contains some slices of oceanic crust or upper mantle. These Precambrian to Lower Palaeozoic rocks are directly covered by Permian red beds formed in the intramountain depressions of the Variscan orogen. The Triassic section begins over Permian or its basement with Lower Triassic mature conglomerate and sandstone, followed by Middle-Upper Triassic carbonates (limestones, dolomites), and Carnian – Norian marine red beds (red to purple shales, marls, and conglomerates, interbedded with oligomictic quartz sandstones and limestones). The whole unit was folded and uplifted in latest Triassic time. The gradual Early Jurassic to Bajocian transgression occurred only in the northern, in the Trun subunit. The Jurassic sedimentation ended with the Upper Jurassic – Berriasian flysch of the Nish-Troyan flysch trough (called also Perimoesian marginal flysch basin), that continued to the East, also within the future Srednogorie and Balkan zones. After the Mid-Cretaceous folding and thrusting, the unit was united with the Morava and Rhodope units into a single Morava-Rhodope superunit that represented a frontal arc ("plateau") of the complex Late Cretaceous volcanic island arc. The Late Cretaceous marine transgression of the Srednogorie basin penetrated only locally in the northeastern parts of the Strouma unit, where a trough with flysch-like sedimentation existed in Campanian and Maastrichtian times.

The Morava unit was formed as a result of Mid Cretaceous thrusting of a ridge structure (Serbo-Macedonian massif) over the Triassic to Lower Cretaceous sediments of the Nish-Troyan flysch trough. The unit consists of several thrust sheets with different lithologies. Only pre-Alpine rock units are present. The basement rocks are very similar to that of the Strouma unit, and belong to two types: pre-Cadomian high-grade gneisses, migmatites, and amphibolites intruded by Cadomian granites, and Neoproterozoic - Cambrian greenschist-facies rocks similar to the Frolosh Formation, with gabbro-diorites to diorites. Some of the thrust sheets exhibit Ordovician metasandstones and schists, Silurian – Devonian basinal schists, lydites, and limestones, and Upper Devonian flysch-like sediments.

The Ograzhden unit is a part of the Serbo-Macedonian massif. It consists of gneisses, migmatites and amphibolites, amalgamated with mantle ultrabasic and basic rocks and intruded by Cadomian granites, aplites, and pegmatites. The whole Ograzhdenian polymetamorphic and polydeformational complex was reworked in Late Cadomian times.

Alpine structure of the SW parts of the Balkan Peninsula

(Macedonia – Epirus)

I. Mariolakos, I. Fountoulis

On Greek territory *(Fig. 2)* the tectonic units largely inherited former Tethyan sedimentation zones with a continuous Alpine evolution. Greece forms a very characteristic part of the Alpine System, known as the Hellenic Arc. It is one of the major mountain chains of the Alpine-Himalayan System, which resulted from the convergence/collision between the Eurasian and the African continental plates.

The morphotectonic direction of the Hellenic Arc in Continental Greece is NNW-SSE (Figure 2), bending gradually to E-W between Kythera and Crete. Eventually, the direction becomes NE-SW east of the Dodekanissa ("twelve islands") island complex up to near Turkey.

The Hellenides comprise a large number of geotectonic units (Figure 2), corresponding to individual nappes. The overall kinematics shows a west-vergent movement directed from the core of the arc in the Aegean Sea towards the periphery, in the Ionian and Libyan Seas (Figures 3, 4).

Two main orogenic cycles have been distinguished in the Hellenides, namely:

- The Palaeo-Alpine orogeny in Late Jurassic – Early Cretaceous times;
- The Alpine orogeny, which started in Late Eocene, and culminated during Oligocene and Miocene times.

However, plate movements with resulting orogenic processes are still active along the present Hellenic Arc and Trench System.

The geotectonic units of the Hellenides can be subdivided into two groups (Internal and External Hellenides). The Internal Hellenides have undergone deformations in both orogenic cycles, whereas the External Hellenides have been deformed only during the Alpine *s.s.* orogeny. The Internal Hellenides consists of the Rhodope, Serbo-Macedonian, Circum-Rhodope, Vardar/Axios, and Pelagonian zones, and the External Hellenides, of the Pindos, Gavrovo-Tripolitza, Ionian, and Pre-Apulian (Paxos) zones.

Figure 2 - The Hellenic Arc and Trench System. KF: Kefalonia Faul; NAF: North Anatolian Fault; PT: Plini Trench; ST: Strabo Trench.

Rhodope zone. This consists of two tectonic zones, and namely, the *Pangaion (Paggaio) unit* (tectonically in a lower position), and *Sidironero unit* (in an upper position). They are separated by the

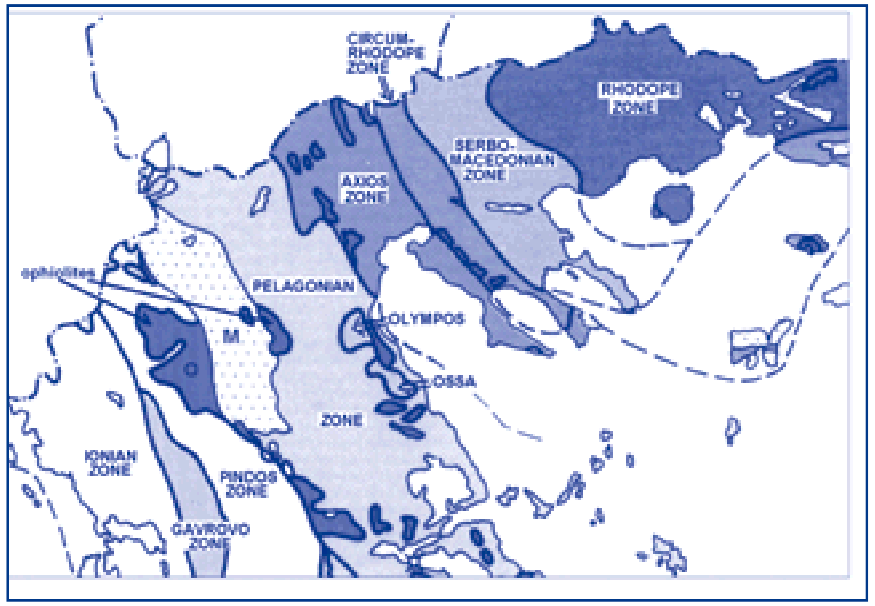


Figure 3 - Simplified map of the geotectonic units of Greece (based on Papanikolaou, 1986)

Volume n° 2 - from B16 to B33

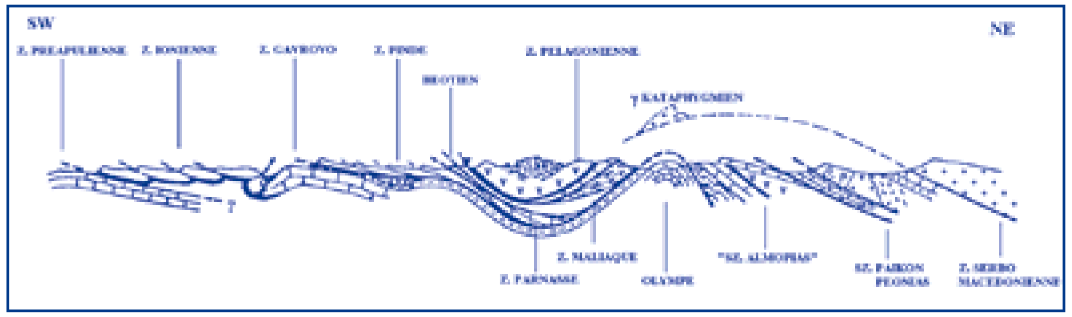


Figure 4 - Schematic cross section of the Hellenides, depicting the allochthony of the Pindos unit in relation to the Olympus window (after Aubouin et al., 1979).

SW-vergent Middle-Mesta thrust. The Paggaio unit is characterized by the extensive presence of marbles, along with amphibolites, gneisses, and mica schists. The Sidironero unit comprises gneisses (muscovitic, biotitic, augen), amphibolites, migmatites, and anatectic granites.

The *Serbo-Macedonian zone* consists, on Greek territory, of two geotectonic units: *Kerdylia* and *Vertiskos* (Kockel, 1977). Their boundary is unclear, and could be considered either as a tectonic or disharmonic contact. Both units are composed of high-grade (amphibolitic facies) metamorphic rocks (mica schists, amphibolites, and migmatites) and are crossed by Hercynian and Alpine granites (schistosed or not). The boundary between the Serbo-Macedonian and Rhodope is located within the Strymon graben (Kockel and Wallther, 1965), with the Serbo-Macedonian zone tectonically overthrusting (ENE-verging Strymon thrust) the Rhodope in the area NE of Sidirokastro.

The Circum-Rhodope zone consists of greenschist facies metamorphic formations, mainly marbles and dolomites (Triassic), clastics (Jurassic), and ophiolites. They are overthrusted with a SW vergence by the Serbo-Macedonian zone. The relationship between the Circum-Rhodope zone, and the more internal parts of Axios zone (Paionia, according to Mercier, 1968), is not clear.

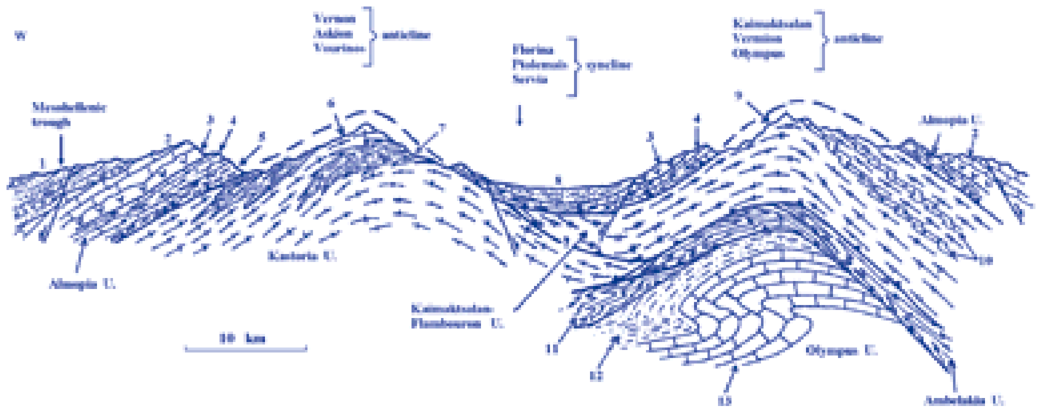


Figure 5 - Synthetic cross section in the northern part of the Middle tectonometamorphic zone (ex-Pelagonian zone) (after Papanikolaou, 1983). 1:Molasse of the Meso-Hellenic Trough; 2: sediments of the Upper Cretaceous transgression; 3: ophiolites; 4: M. Triassic - Jurassic marbles of Almopia unit; 5: phyllites, marbles and meta-volcanics of Almopia unit; 6: granites and gneisses of Kastoria unit; 8: post-Alpine deposits of the Ptolemais basin; 9: gneisses, granites, amphibolites, mica schists of Flabouro unit; 10: marbles of Flabouro unit; 11: blueschists of Ambelakia unit; 12: Eocene flysch of Mt. Olympus.; 13: Triassic – Eocene crystalline limestones of Mt. Olympus .

Axios/Vardar zone. It was Mercier (1966) whodivided the Axios zone into three sub-units: *Almopia* (later defined as a geotectonic unit by Papanikolaou, 1984), *Paikon,* and *Paionia*. The Paionia unit consists of marbles, crystalline limestones, clay schists. and phyllites of Triassic – Jurassic age. They are tectonically overlaid by basic magmatic rocks (but not a typical ophiolithic complex). of Jurassic age. The Paikon unit comprises neritic marbles, clastic sediments, both basic and acid volcanic rocks. and the “Phanos granite” of Jurassic age. It is a low-grade metamorphic unit in which some fossils are found. The metamorphism “disappears” in the Upper Jurassic – Cenomanian formations. The Almopia geotectonic unit was formerly considered as part of the Axios zone. It crops out as a tectonic nappe, “rising” from the Axios area, and reaches the Meso-Hellenic Trough, covering all the metamorphic units of the Northern Pelagonian tectono-metamorphic unit. The column of the unit begins with basic rocks (ophiolites), unconformably covered by Aptian conglomerate. It is followed by Upper Cretaceous carbonates and Upper Maastrichtian flysch.

The Pelagonian zone consists mostly of metamorphic rocks of Palaeozoic and Mesozoic age, and is subdivided into two major tectonic units: Flabouro and Kastoria.

The Flabouro unit is characterized by the presence of high-grade metamorphic rocks, gneisses, amphibolites, mica schists, and marbles. They have a polymetamorphic and polydeformational character, and are intruded by Hercynian (Carboniferous – Permian) granites.

Kastoria unit. The stratigraphic column of the Kastoria geotectonic unit is divided into two parts. The lower part consists of granites, mica schists etc., and the upper one, of low-grade metamorphic sedimentary formations (low-grade greenschist facies), such as phyllites, quartzites, sericite – chlorite schists and metatuffs. Older (schistose) and younger granites are distinguished.

Eastern Greece. It was deformed twice, during the Palaeo-Alpine, and the typical Alpine orogeny. A neritic Triassic – Jurassic sequence (*Sub-Pelagonian unit*) is overlain by the obducted ophiolites, and the transgressive Upper Cretaceous limestones. The section ends with the Eocene flysch.

Western Thessalia – Beotia zone. This zone is characterized by a continuous sequence from Triassic to Eocene. It is the most internal stratigraphically continuous unit of the Hellenides. The section begins with Triassic limestone and dolomite, followed by Jurassic pelagic limestones, shales, and radiolarites. Lower Cretaceous rhythmic sandstones and limestones follow. They are covered by the Upper Cretaceous flysch that is rich in olistolites and pebbles from ophiolites and limestones of Mesozoic age. Upwards follow pelagic limestones and marls (Palaeocene), and Eocene flysch.

Parnassos zone. A thick Triassic – Lower Cretaceous neritic carbonate sequence is interrupted by 3-4 bauxite horizons. Red pelagic limestones and shales are followed by a Palaeocene – Eocene flysch.

Pindos. The Pindos zone is characterized mostly by pelagic sediments, presumably from the Tethys slope. The oldest rocks are Middle Triassic sandstones and shales, followed by Triassic - Cretaceous pelagic limestones, pelites and radiolarites (locally related to pillow-lavas), Cretaceous cherty limestones, pelites and flysch, and fine-bedded limestones (Coniacian-Maastrichtian). The section ends with the Palaeocene – Eocene Pindos flysch.

The *Gavrovo-Tripolis zone* corresponded to a neritic carbonate platform. The base of the stratigraphic column consists of Permian – Upper Triassic clastites, carbonates, evaporites, and volcanites (Tyros beds), slightly metamorphosed to slates-phyllites. The *Tripolis subzone* is characterized by neritic limestones and dolomites (Upper Triassic – Upper Eocene). The section of the *Gavrovo subzone* begins with Upper Jurassic and Lower Cretaceous shelf (oolithic, reefal, etc.) carbonates. The Upper Cretaceous is represented by limestones with rudists, and dolomites and carbonate breccia with orbitoids, and fragments of corals and rudists. Bauxite levels and sedimentation breaks are recorded. The most important breaks are referred to the Cretaceous/Palaeogene boundary and in the Palaeocene and Eocene section. The latter consists of shallow-water carbonates replaced in the Upper Eocene by shales that pass upwards into flysch.

Ionian zone. The zone is built up of Triassic breccia and evaporites, Upper Triassic neritic carbonates, the Liassic Pantokrator Limestone, red nodular limestones and black shales (Upper Lias – Lower Malm), and platy limestones and cherts (Tithonian – Turonian). Pelagic limestones (Campanian – Eocene) pass upwards into the West-Hellenic flysch.

The non-metamorphic units of Hellenides are characterized by longitudinal NNW-SSE trending

Volume n° 2 - from B16 to B33
Volume n° 2 - from B16 to B33

b-folds without schistosity, mostly representing flexural-slip folding. The interlayering of competent and incompetent rock units control the geometry of folding. The rocks probably have not been subjected to a load of more than 3-5 km.

Extensional collapse of the Alpine orogens in the eastern parts of the Balkan Peninsula

(I. Zagorchev)

A special emphasis is laid on the extensional collapse processes in Palaeogene and Neogene times. The Late Cretaceous orogen, issued from the Srednogorie island arc, was the subject of fast uplift and erosion, already in latest Maastrichtian times, and of gradual peneplainization in the Danian (Figure 6). Exhumation of deeper levels occurred both on the former frontal

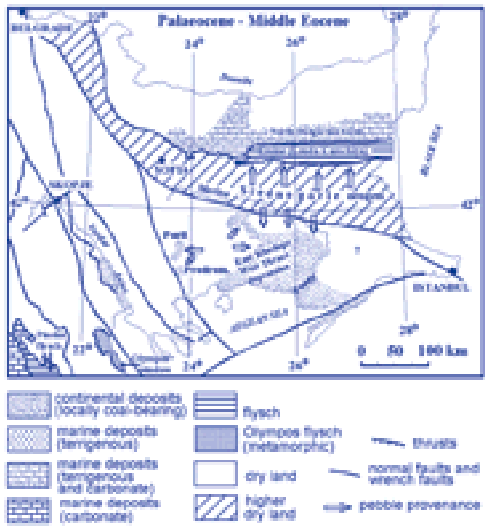


Figure 6 - Sketch map for the Palaeocene palaeogeography (without palinspastics). Modified after Zagorchev (1996)

arc ("plateau") of the Morava-Rhodope superunit, and in the Srednogorie itself.

A new extension occurred in Palaeocene and Early to Middle Eocene times, with local flysch basins in the Balkan area, and with continental grabens with marine ingressions in the Rhodope region. Intracontinental collision processes formed the first fold-and-thrust belt of the Balkanides. Limited thrusting probably occurred in the Rhodope area, being more important in its southern part on Greek territory.

The extensional collapse of the Middle Eocene orogen occurred in Late Eocene and Early Oligocene times, and was most prominent in the area of the "plateau" – in the heterogeneous crustal areas of Srednogorie and Morava-Rhodope. Numerous small grabens were filled in by coarse terrigenous, often coal-bearing, sediments, beginning with the Bartonian, and mostly in Priabonian times. At the end of the Priabonian, marine ingression along the Strouma/Strymon fault belt and in the East Rhodope Mountains was almost coeval with outbursts of intense volcanic activity. The volcanic activity had a different composition in the different parts of the arc. It was of a bimodal composition, probably deriving from mixed mantle and crustal sources,in the East Rhodope; of a trachyandesitic composition, with a later more acidic tendency, along the Strouma belt and in the Mesta graben complex; and with an acidic (rhyolites) character in continental conditions (on dry land or in lacustrine environments), in the Central and Western Rhodope, where the continental crust reached its maximum thickness. Thus, a complex fluviolacustrine system was drained towards the two marine gulfs. The field trip demonstrates the relationship between the Palaeogene formations and their basement in the Padesh graben (a part of the gulf along the Strouma belt), and in the continental Padala graben (a part of the fluviolacustrine system near the source area).

The marine regression in Mid- to Late Oligocene times led to a total reorganization of the

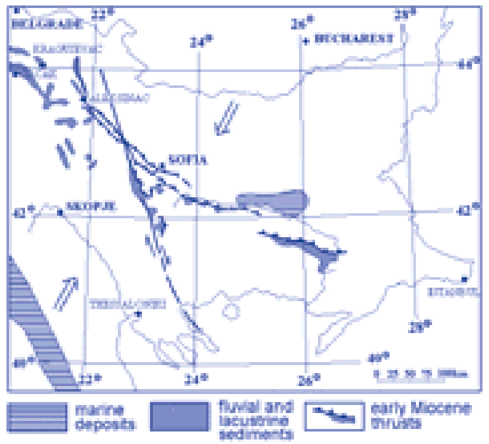


Figure 7 - Earliest Miocene transpression, thrusting and right-lateral strike-slip along the Strouma fault belt (after Zagorchev, 1992, 1996).

fluviolacustrine systems on the Balkan Peninsula. Almost all previous basins were dried out. A marine gulf became active along the eastern part of the Maritsa fault belt East of Plovdiv. In transtension conditions, a huge lake extended along the Strouma fault belt from Brezhani in SW Bulgaria (to be briefly visited), through Bobovdol, Pernik, and East Serbia, where it drained into the Pannonian sea basin, with a probable link to the East Carpathian basin. This Paratethys system gave birth to the most important coal deposits in Romania, Serbia, and Bulgaria. In earliest Miocene times, the system became reorganized, due to dramatic transpression events (Figure 7).

Molasse basins in the Hellenic area

I. Mariolakos, I. Fountoulis

The evolution of Northern Greece, especially during the neotectonic period, has been the subject of different and often controversial opinions. This evolution is greatly defined by the activity of the North Anatolian Fault (NAF), and its prolongation towards the Aegean Sea.

Three great molassic basins are distinguished within the Hellenic area. Their activity and location cover different time spans during the Alpine evolution of the Hellenic arc. These basins are (Figure 8):

Figure 8 - The main molassic basins of the Hellenic Arc system. (after Papanikolaou, 1986).

The *molassic basin* of *Rhodope* existed as a basin in the northeastern part of Greece during Eocene-Oligocene times. The molassic sediments were deposited over the metamorphic basement of Rhodope, the area of Axios, and the present-day marine area of the northern Aegean. The oldest sediments are of Early to Middle Eocene age, and span to the Oligocene/Miocene boundary (Kopp, 1966). An Oligocene volcanic activity is most significant.

The *molassic basin* of the *Meso-Hellenic Trough* existed in the midst of continental Greece during Late Eocene – Middle Miocene times. The trough sediments were unconformably deposited over the non-metamorphic formations to the west and the metamorphic formations to the east. Along different sections perpendicular to the longitudinal axis of the trough, the consistency and the origin of the deposits vary. So, every area has its own stratigraphic column. Brunn (1956) was the first to study the Meso-Hellenic Trough sediments, and divided them into different "formations", namely (from bottom to top): the Krania Formation (Upper Eocene); Eptahori Formation (Oligocene), which consists mainly of marls; the Pentalofo Formation (Upper Oligocene – Aquitanian), consisting mainly of conglomerates (Meteora area) and sandstones; the Tsotyli Formation (the thickest of all, Lower Miocene), consisting of marls and sandstones; and lastly, the Odria Formation (Middle to Upper Miocene) that consists of small thickness (20-50 m) sandy – marly limestones. New research proved that the marine sedimentation continued at least up to the Early Pliocene times (Fountoulis et al., 2001). The total thickness of the Meso-Hellenic Trough deposits reaches c. 5 km.

Volume n° 2 - from B16 to B33

The *molassic basin of Epirus – Acarnania* existed as a sedimentary basin in western continental Greece during Late Oligocene – Miocene times. In this particular basin it is difficult to distinguish between molasses and the underlain flysch (common formation to both Ionian and Gavrovo zones). This distinction can only be made with certainty in areas where the Lower – Middle Miocene molassic formations are in direct contact with the folded sediments belonging to the Ionian zone.

Neogene basins in the northern and eastern parts of the Balkan Peninsula

I. Zagorchev

Most of the Balkan Peninsula was dry land in Early and Middle Miocene times. The evidence about the evolution in these times is scarce. Marine basins still existed in the Hellenides, and in the Pannonian basin of the Paratethys. The central and eastern parts of the Peninsula were affected by slow motions and denudation that resulted in the formation of the principal planation surface (initial peneplain, orthoplain). Due to the balance between uplift, accumulation, and erosion, no traces of sediments formed in these times have been preserved. Only recently, palaeobotanical evidence has been published about the Ottnangian to Karpatian age of some formations which have very limited occurrence in the West Rhodopes.

Considerable changes in this regime began in the second half of Middle Miocene time. Marine and brackish basins flooded the surrounding areas: Ionian/Adriatic, Pannonian, Precarpathian, Euxinian (Black Sea) and later, the Aegean basin. Normal faulting (mostly along the older fault belts) led to relief contrasts and triggered faster erosion and denudation in the horsts and sediment deposition in the grabens, thus initiating a new pattern of fluviolacustrine systems centered on big palaeorivers along the Maritsa and Strouma/Strymon fault belts. The areas with the thickest continental crust (West and Central Rhodopes), became again involved in differential uplift and extension. Times of relative quiescence are documented by mountain steps (pediments, oroplains). The fault amplitudes changed over time, with a climax in the time span Pontian – Pliocene, when the thickest terrigenous sediments were deposited, parallel with the fast climate changes, and marked by the disappearance of savannahs and their big mammals (Pikermi faunas), and the coming of the Ice Age in the highest mountains (Rila, Olympus, Pirin, Belasitsa, Vitosha, etc.).

The existing evidence favors geodynamic interpretations of this area, centered upon the interference of mechanisms such as isostasy, rifting, horizontal movements along large strike-slip faults, with the formation of pull-apart basins, etc. (s. Zagorchev, 1992, 2002). These interpretations run contrary to current popular hypotheses (e.g., Dinter and Royden, 1993), which suggest the existence of a very large extension, caused by detachment systems in the Aegean and peri-Aegean regions.

Neogene basins of the Hellenic arc

I. Mariolakos, I. Fountoulis

The post-Alpine formations of Greece comprise beds unconformably deposited over the Alpine basement. They belong to the back-arc basins of the present-day Hellenic arc, and their deposition took place during Neogene and Quaternary times. These are mainly continental deposits (plus fluvial and lacustrine), and sometimes marine (coastal), deposited in neotectonic grabens.

The most important marine basins are of Late Miocene – Pliocene age, and are located mainly in the Peloponessus (Plio-Pleistocene only) and on the island of Crete, while in the rest of continental Greece, continental basins prevail. As expected, many basins show a multi-phase evolution, since continental facies alternate with lacustrine or coastal marine, and vice-versa. Especially interesting are the cases of post-Alpine (Neogene) basins formed and evolved over pre-existing molassic basins, such as in Thessaly and western Macedonia (over the Meso-Hellenic Trough) and in the Serres – Drama basin (over the Rhodope molasse).

The limited extent of marine post-Alpine sediments in Greece is due to the fact that many marine basins of Neogene age still exist today (Thassos basin).

A general observation concerning the distribution of Neogene basins is that although some characteristic trends exist, a normal spatial distribution is not present. Below, the presence of other common features, such as the sedimentary facies or their thickness, is examined.

Although the systematic study of Greece has not yet been completed, it is evident that there is a general geotectonic regime, within which some areas differ because of special tectonic conditions.

The post-Alpine (Neogene-Quaternary) basins, selected for the stops on this field trip, are the

Volume n° 2 - from B16 to B33

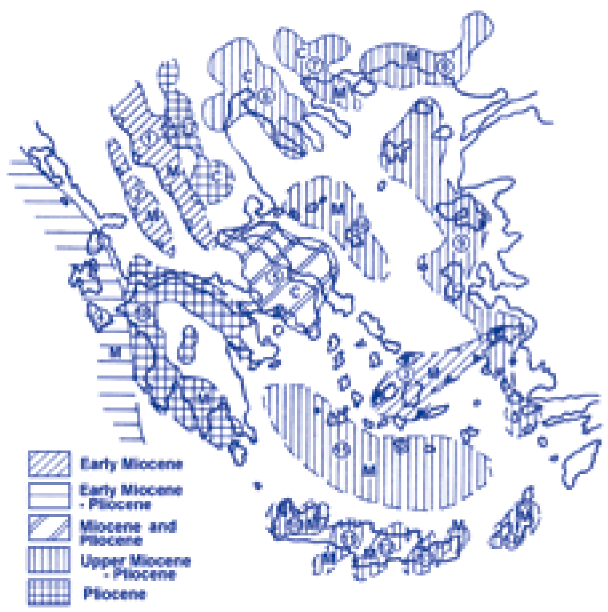


Figure 9. Main Neogene basins of the Aegean region (after Dermitzakis and Papanikolaou, 1981): a) Lower Miocene, b) Lower Miocene – Pliocene, c) Miocene and Pliocene, d) Upper Miocene – Pliocene, e) Pliocene (M: marine, C: continental). 1:Meso-Hellenic Trough, 2:Epirus, 3:Cyclades, 4:Ionian islands, 5:Attica – Boeotia, 6:Thessaloniki, 7:Strymon, 8: Thrace, 9:Eastern Aegean islands, 10:Chania, 11:Rethymno, 12: Irakleio, 13:Ierapetra, 14:Karpathos, 15:Cretan basin, 16:N. Sporades, 17:Servia – Ptolemais, 18:Peloponnesus).

following (Figure 9): Strymon basin; Axios basin; Servia – Ptolemais basin; Meso-Hellenic Through; and the Thessaly basin.

The current geodynamic regime in the Hellenic area

I. Mariolakos, I. Fountoulis

The present Hellenic Orogenic Arc is restricted to the southern part of the Hellenic territory, in contrast to all the previous arcs, which extended throughout the whole length of the Hellenides.

During the Middle Miocene, a part of the Hellenic arc, still active today, was cut off from the Tethyan chain, and since then has followed its own evolution. To the north, this part is bounded by the prolongation of the right-lateral Anatolian fault (Figure 10). In the northern Aegean region, this fault coincides with the northern limit of the active part of the Hellenic arc, bounding an area termed “Aegean microplate” (McKenzie, 1970; 1972; 1978; Galanopoulos, 1972). The present geometry of the Hellenic Arc has been developing since the Late Miocene. The back-arc basin and the volcanic arc are restricted in the Aegean plate region.

According to Le Pichon et al. (1981), the present geodynamic regime of the Hellenic arc is characterized by asymmetrical movement; along the Ionian trench, the subduction direction is NE-SW, and the regime is pure compression, in accordance with the fault plane solutions while, in the Pliny and Strabo trenches, the direction of movement is composite, featuring a substantial sinistral NNE-SSW horizontal component. In the back-arc area there are extensional structures, also with a significant horizontal component of movement.

According to Mariolakos et al. (1981, 2001), the Greek territory can be divided into three Morpho-Neotectonic regions (MNR) (Figure 10), and these are described below:

The Northeast MNR (NE-MNR) is bounded by the large North Aegean Fault Zone (prolongation of the North Anatolian Fault into the north Aegean Sea) to the south, and the Axios fault zone to the west. Compared to the NW-MNR, there are no intensive vertical movements affecting the whole region, resulting generally in a smooth relief, but they are localized in areas where fault reactivation occurred, mainly in Plio-Quaternary times, resulting in local incision. The majority of the high order tributaries of the drainage network come from the north (the Axios, Strymon, Nestos, and Evros Rivers). The 1st order basins strike NW-SE. In the northern margins of the present grabens, Plio-Pleistocene volcanic activity occurs (Aridaia, Sidirokastro, and Strymoniko).

The Northwest MNR (NW-MNR) is bounded by the Axios fault zone to the east, and the prolongation of the Malliakos - by the Amvrakikos gulf fault zone

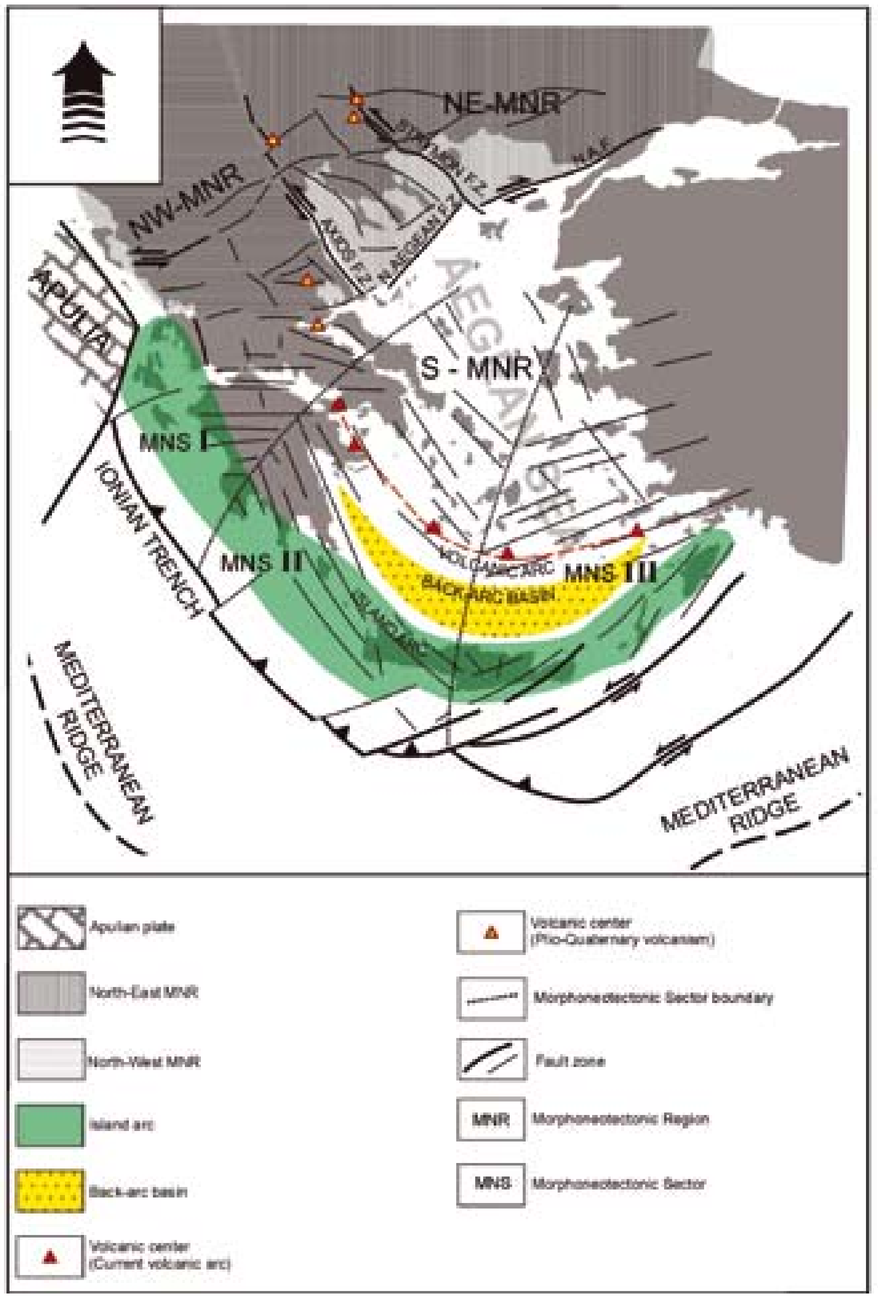


Figure 10 - Current morpho-neotectonic regime of the Greek territory.

to the west. Intensive vertical movements (mainly uplift), affecting most of the region, resulted in intensive fragmentation (partition), high-energy relief, incision, and drainage developments, mainly within the Greek territory, and characterize this morpho-neotectonic region. The mountain ranges as

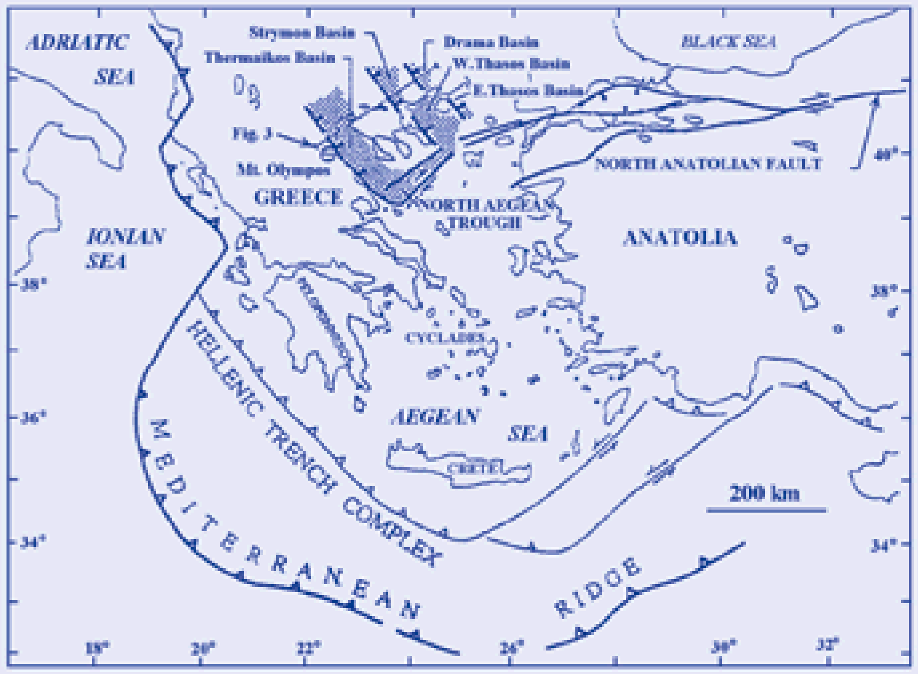


Figure 11 - Tectonic elements of NE Greek region. Subduction-related thrust contacts have barbs on hanging wall. High-angle faults bounding the modern basins have balls on downthrown side. Offshore geometry of modern depocenters (dot pattern,) north of the North Aegean trough is from Lalechos and Savoyat (1977) (after Dinter and Royden, 1993).

well as the drainage network strike mainly NW-SE. Some other characteristics, especially for the western sector, are the absence of typical horst – graben structures, and the occurrence of E-W strike-slip fault zones. For the eastern sector, the initial strike of the horst – graben structures is NW-SE (Meso-Hellenic Trough, Ptolemais – Servia basin), and later were fractured into smaller basins by faults striking NE-SW. Plio-Pleistocene volcanics occur in the SE part of the sector.

The South MNR (S-MNR) is bounded by the large North Anatolian Fault Zone and the prolongation of the Malliakos - Amvrakikos gulf Fault Zone to the north, by the Hellenic trough to the west and south, and by the Pliny and Strabo troughs to the south-southeast. The Strabo and Plinytroughs constitute the eastward prolongation of the southern part of the Hellenic Trough. Of course the Pliny and Strabo troughs present different kinematics to the Hellenic one, resulting in differences in (fault) brittle tectonics, karstification, and hydrogeological characteristics of the controlled areas.

Within these MNR a large number of neotectonic basins have been created, which have been mainly filled with deposits of Pliocene – Quaternary age.

Most geodynamic models proposed for the evolution of the Aegean domain during Neogene times have considered that the initiation of extension in this area is a direct consequence of the westward extrusion of the Anatolia (Minor Asia) plate away from the Arabia – Eurasia collision front.

According to Dinter and Royden (1993), and Dinter (1994), the Strymon basin detachment system accommodated at least 25 km of extension at the northern margin of the Aegean extensional province, from Middle Miocene through Early Pliocene time. Displacement ceased on the Strymon basin detachment with the Late Pliocene development of the Strymon and Drama basins. They propose that the Strymon and Drama basins, together with the other principal basins in the northern Aegean, are subsiding above an active northeast–dipping extensional detachment zone, that forms a unified kinematic system with the western offshore continuation of

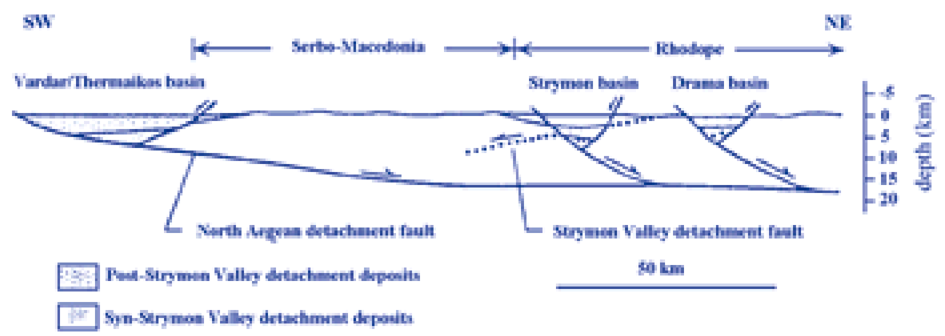


Figure 12 - Schematic cross section, showing major Upper Cenozoic extensional structures of NE Greece. Plane of section is parallel to slip on Strymon Valley detachment fault, and sub-parallel to slip on North Aegean detachment system, as proposed by Dinter and Royden(1993). No vertical exaggeration. See Figure 11 for locationand.

the dextral North Anatolian Fault, and the Northern Aegean Trough (Figures 11 and 12).

Gautier et al. (1999) suggested that regional-scale

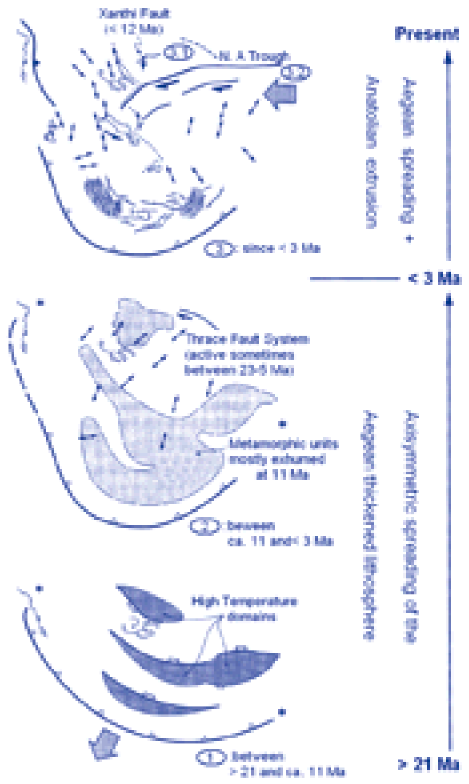


Figure 13 - Three-step scenario depicting the Neogene evolution of the Aegean (after Gautier et al., 1999).

extension, with a pattern of stretching orientations similar to that of Pliocene – Pleistocene, was already strongly active in this domain before the onset of the Arabian indentation into Eurasia (Early Miocene vs. Middle or Late Miocene). This implies that the initiation of the Aegean extension did not result from the lateral extrusion of Anatolia, and they propose that the extension started due to the gravitational spreading of the continental lithosphere that had previously been thickened during the Alpine collision (Figure 13).

Moreover, many geodynamic models have been proposed for the Hellenic arc. These models accept that the latter was the result of an extensional stress field, accompanied by grabens created by normal faulting in the back arc basin (according to publications by Ritsema, 1974; McKenzie, 1978; Mercier, 1979; Le Pichon and Angelier; 1979; Dewey and Seng?r, 1979 and others: s. Mariolakos and Papanikolaou, 1981a).

Mariolakos and Papanikolaou (1981a) suggested that marginal fault zones control the configuration of Neogene basins. These fault zones create an asymmetry with the basin morphology and sedimentation. According to the above-mentioned authors, the Hellenic arc is separated into three large parts. In part I, the major fault zones have an E – W direction. In part II, the direction is NW – SE, and in part III, the direction changes to NE ≠ SW. This arrangement shows that only parts II and III have an apparent dynamic relation to the Hellenic arc and trench system, while part I has its own peculiarity.

Data on the current deformation pattern of the Hellenic Arc have been provided by numerous researchers, and

these take the form of: (i) *in situ* measurements of the stress field in shallow (<10 m) drillings (Paquin et al., 1982); (ii) Palaeomagnetic investigations of the Neogene and Quaternary sediments (Laj et al., 1982); (iii) fault plane solutions (McKenzie 1972; 1978, Ritsema, 1974; Drakopoulos and Delibasis, 1982; Papazachos et al., 1984.) Mariolakos and Papanikolaou (1987) combined the results of various geological, seismological, and geophysical studies, and proposed a present (active) deformation model. Further data that contribute to the interpretation of the current deformation regime of the Hellenic territory, have been supplied by geodetic measurements (Billiris et al., 1991) and by the distribution of earthquake foci.

More recent investigations in the SW Peloponnessos (Messinia) showed (Mariolakos et al., 1991) that the stress field responsible for the neotectonic active deformation of the area is that of a rotational couple, which caused not only brittle but also ductile deformation structures, resulting from local transpression and transtension sectors.

Field itinerary

DAY 1

Sofia - Plovdiv - Starosel - Hisar - Karnare - Troyan (about 310 km)

The itinerary of the first day (see inside back cover) to start with follows the famous route from Rome, via Serdica, to Constantinople, along the Maritsa fault belt in the southern edge of the Srednogorie zone. From Plovdiv, it crosses towards the north the Upper Thracian plain (Neogene – Quaternary graben), the central parts of the Sredna-gora crystalline fragment, the Sub-Balkan neotectonic grabens, the Sub-Balkan normal fault, and the Central-Balkan thrust, the Balkan (Stara-planina) zone, and enters the Fore-Balkan zone.

East-Southeast of Sofia, the itinerary leaves the Sofia neotectonic graben, and follows the Maritsa fault belt in the southern edge of the Ihtiman unit, in the Srednogorie zone. The "Trayanovi vrata" (Trayan's Gates) Pass had a strategic importance in antiquity as being the only way from Europe to Constantinople.

Stop 1.1:

Quarry and road cutting: altitude 491 m; at about 70 km from Sofia

Granodiorites of the Upper Cretaceous Vurshilo pluton. Formerly believed to be of Palaeozoic age because of the superimposed schistosity, they are now referred (N. Georgiev) to the Upper Cretaceous (c. 82 Ma; A. v. Quadt). Numerous xenoliths from folded metamorphic rocks are observed. Other features exhibited at the outcrop are: vertical wrench fault, striking 118º, with striae plunging East; almost vertical dyke of granodioritic porphyrite, striking 130º, about 1.2 m thick. A superimposed, almost vertical schistosity strikes 130-135º, with striae in the schistosity planes plunging 55ºNW.

Dykes of granite-porphyries, and granodioritic, syenodioritic, and monzodioritic porphyrites are typical of the Srednogorie. Most of them have a Late Cretaceous age, and are intruded both in the Upper Cretaceous intrusive (hypabyssal to subvolcanic) bodies, and in the host Palaeozoic granites and Precambrian metamorphics. The schistosity is related to an Ihtiman shear zone, that was first a magma conductor under transtensional conditions (with mixing of magmas of mantle and crustal origin), and then developed, under transpressional conditions and right-lateral strike-slip.

Stop 1.2:

Plovdiv, Nebet-tepe Hill: 209 m

The hills of the old town are built up of syenites ("plovdivites"), with numerous xenoliths and enclaves from basic and acid rocks, hornfelses included. Moebus studied in detail the composition of the enclaves, and reported in 1959 the presence of reworked Upper Cretaceous sedimentary rocks and hornfelses. According to him, the "plovdivites" were formed through intense metasomatosis, and should not be called "syenites", this name being reserved only for rocks of purely igneous origin.

The town is more than 6000 years old, thus being older than either Rome or Constantinople. A Thracian fortified town (Eumolpia, Poulpoudeva), it was conquered in 432 BC by Phillip II of Macedonia and named "Phillipopolis" after him. The Romans called it Trimontium. In 815 the Bulgarian Kan Kroum seized the fortress. Afterwards, it was for different periods in the hands of Byzantines, Crusaders, and Turks (Filibe), and was later proclaimed by the Berlin Congress (1878) capital of the autonomous province Eastern Roumelia, and re-united with Bulgaria in 1885. Our brief visit will allow us to observe archaeological traces of this turbulent past.

>From Nebet-tepe we observe the panorama of the Plovdiv polje of the Upper Thracian depression. It

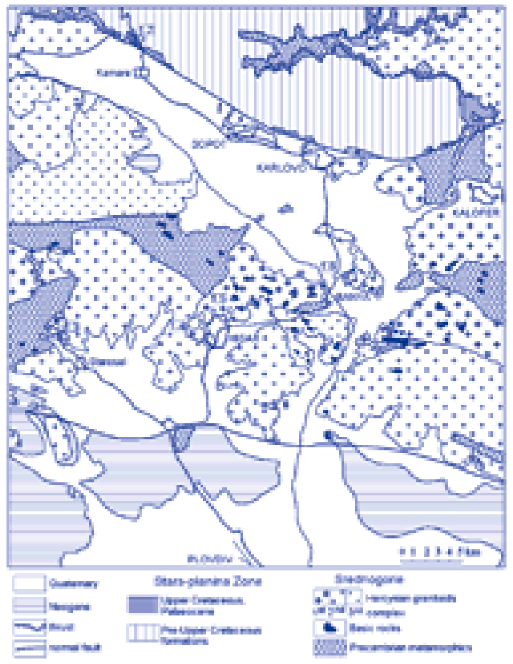


Figure 14 - Geological map of the central part of the Srednogorie zone

is a wide graben structure, superimposed over the Srednogorie zone during Late Oligocene, Neogene and Quaternary times, along the Maritsa fault belt. to the south, the graben is framed by the complex neotectonic horst of the Rhodopes, and to the north, by the complex horst of the Sredna-gora crystalline block. The normal faults and strike-slip faults of the Maritsa fault belt strike 90-120º, and have steep to moderate dip angles. With transversal and oblique faults, they form the complex block structure of the pre-Oligocene basement. Some of the faults are still active, as witnessed by the earthquakes, the Plovdiv earthquake of 18.04.1928 (M=7) included.

Stop 1.3:

Temple of Sitalkes, Starosel: 484 m

Excavations during the last 3 years have exposed a Thracian temple complex, referred to the time of King Sitalkes (5nd Century B.C.). The principal temple is built over Precambrian biotite gneisses (often garnet-bearing) and migmatites, intruded by the biotite porphyric granites to leucogranodiorites of a Palaeozoic (c. 300–320 Ma) granite pluton. Rarely, dykes of pegmatites and aplites are observed.

The temple of Sitalkes is built with local granite blocks but the well-polished elements (columns, capitels, etc.) have been made from imported zeolitized tuffs.

Stop 1.4:

Hisar

From Starosel we continue to Hisar – first a Neolithic site and a Thracian town transformed into a fortified spa (Diocletianopolis) during the Roman rule. The well-preserved walls (2.5 km long, up to 10 m high, with 44 towers, and 15 m high main gates) are a typical Roman construction of stone, brick and mortar. The stone blocks come mostly from the local biotite porphyric granite of the Hisar pluton (c. 337 Ma, Sr_i 0.706). For some important places (key stones and thresholds at the northern gates, corner blocks of towers, etc.), huge blocks of equigranular syenites have been transported from Plovdiv. Presently, 22 mineral springs have a total debit of 4000 l/min, the temperatures ranging between 27 and 52ºC. The waters are alkaline (pH 7.6 – 9.02), sodium hydrocarbonic, with sulphate, fluorine, and more than 20 microelements.

The Hisar polje represents a peculiar irregular post-sedimentation (in respect to the Neogene) graben formed by ENE- and NNE-striking steep (60-90º) normal faults. To the south, the Starosel and Hisar grabens are separated from the Upper Thracian depression by the Krasnovo normal fault that strikes ESE. Thus, a second-order step-like graben is formed between the depression and the Sredna-gora horst. The normal faults are the conductors of the thermal mineral waters.

Stop 1.4a:

Quarry "Sveti Georgi" (St George), Verigovo: 432 m

The young (III complex, c. 270 Ma) Sveti-Georgi granitoids are mostly biotite equigranular fine- to medium-grained granites, leucocratic, often with schlieren, due to the contamination with, and resorption of, basic enclaves and older granitoids. Xenoliths and enclaves originating from the Hisar granites (c. 337 Ma) are abundant.

Stop 1.5:

Quarry "Momina banya": 445 m

Coarse-grained biotite porphyric granite to granodiorite of the Hisar pluton, with numerous melanocratic enclaves with dioritic composition.

Table 1 - Altitudes of the Neogene planation surfaces in Central Bulgaria

It possesses a clear platy parallelism of enclaves, schlieren, coarse (up to 5 cm) K-feldspar endoblasts, and biotite. The planar structure strikes 120-135º, and has a steep (about 80º) dip to NNE that indicates a steep northern contact of the pluton with the host Precambrian metamorphics. The granitoids of the Hisar pluton are intersected by dykes of fine- to medium-grained Sveti-Georgi granites. The thickest (1.5 m) dyke strikes 165º, and dips to the ESE at about 55º. Another 60-cm dyke strikes 140º and dips 55-70º NE. Whole-rock samples from the quarry defined a Rb-Sr isochrone, corresponding to 271 +/- 26 Ma and an initial Sr ratio of 0.709.

The road continues through biotite gneisses and migmatites, and the strongly weathered leucocratic granites of the Mihiltsi pluton (III complex) similar to the Sveti-Georgi granites.

Stop 1.6:

Road junction after the town of Banya: 316 m

The stop exhibits a neotectonic panorama of the Karlovo graben, the Sub-Balkan fault and the Stara-planina linear morphostructure (s. Figure 14). The neotectonic features of these parts of the Srednogorie zone are related to the formation and subsidence of a Sredna-gora swell, trending approximately west-east. The subsidence is linked to the collapse of the back parts of the Stara-planina Palaeogene orogen. As a result, two rows of Neogene – Quaternary grabens have been formed: the grabens of the Sub-Balkan poljes to the North (at the boundary between the Srednogorie and Stara-planina zones, along the Sub-Balkan fault), and the grabens along the Krasnovo fault, to the South. The internal structure of the Karlovo graben is complicated by second-order normal faults. It is filled with Neogene (Maeotian? – Romanian) and Quaternary sediments, with a total thickness of about 500 m. The sediments have been deposited in a fluvial system, and a lake (with diatomites and coal) has existed in the northernmost parts in latest Miocene – earliest Pliocene times.

The Upper Cretaceous Srednogorie orogen was uplifted and deeply denudated already at the end of the Maastrichtian. In some parts (southern edge of the Sredna-gora crystalline block) a fluvial system was formed in Danian time, being drained towards the marine gulf along the Stara-planina range (then below sea level). At the end of Middle Eocene times, the new folding within the Balkans led to the north-vergent thrusting of the Srednogorie over the Balkan (Stara-planina zone), followed by the uplift of the latter in Late Eocene and Oligocene times. During the most important Neogene – Quaternary uplift, a tectonic relief inversion favoured the formation of the numerous normal faults of the Sub-Balkan fault belt, and some of the old thrust surfaces rotated and were transformed into steep normal faults. Further on, some of the steep normal faults rotated in the extension process to low-angle (listric) normal faults.

Volume n° 2 - from B16 to B33
Volume n° 2 - from B16 to B33

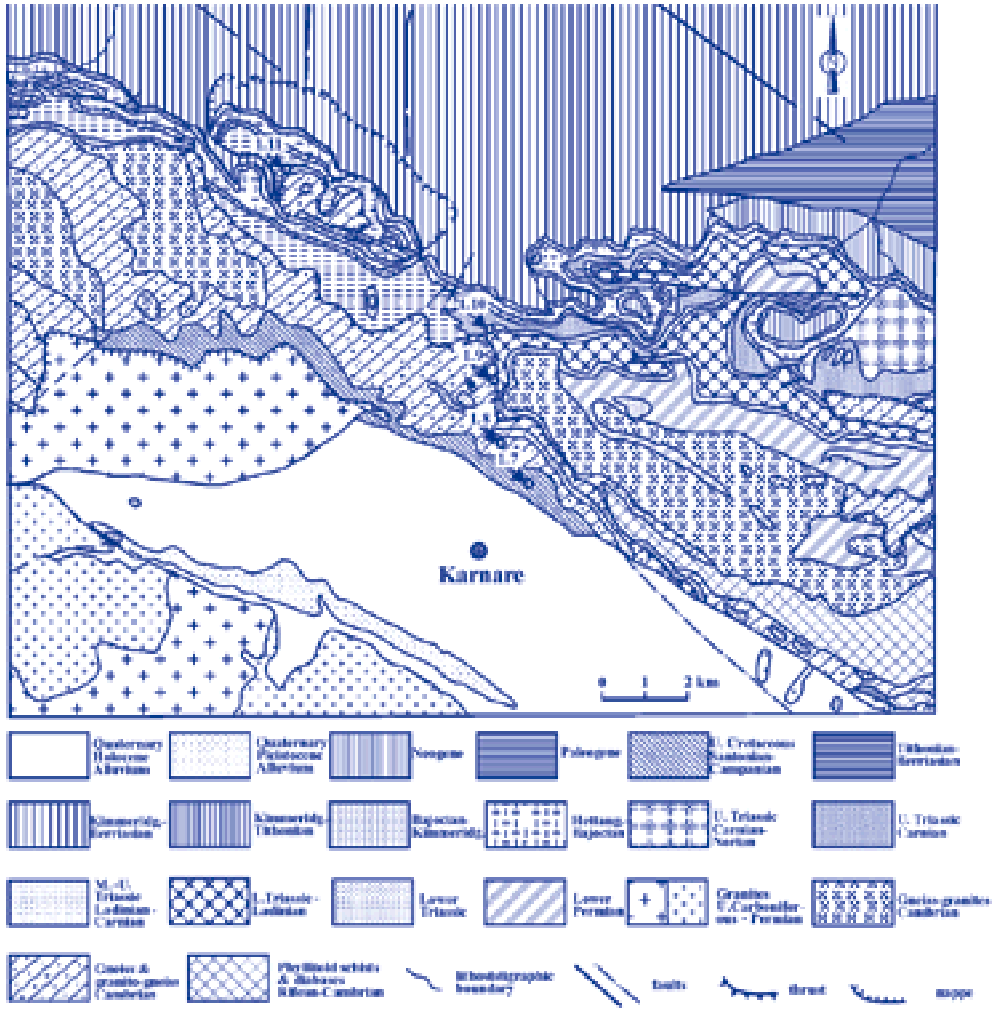


Figure 15 - Geological map of a part of the Central Srednogorie, Central Balkan and Central Fore-Balkan (after Geological map of Bulgaria 1:100000)

The itinerary continues through the Karlovo graben, and at the village of Karnare, reaches the normal faults of the Sub-Balkan fault belt.

After the village of Karnare (altitude 643 m) the itinerary leaves a high mountain step built over the Neogene at the foot of the mountain, and enters the Stara-planina zone.

Stop 1.7:
Along the road North of Karnare: 666 m

The road cutting exposes tectonized Upper Cretaceous deposits. They represent an alternation of clayey limestones (locally with flint nodules), and fine- to medium-bedded marls. The limestones are mostly micritic, and the marls, clayey or slightly silty, and fine-bedded. The two lithotypes are combined in doublets and represent periodites. They are probably climate-controlled, and represent low-order Milankovic cycles. This clayey-limestone formation has a Santonian – Campanian age. The bedding varies between 75º/35ºSSE and 120º/50ºSSW. South-

vergent (opposite to the prevalent North-vergent thrusting) upthrusts are observed. The next clayey limestones exhibit also the typical North-vergent thrust surfaces.

Stop 1.8:
Parking place to the right side: 993 m

A neotectonic panorama exhibits (Figure 15) relics of the granite thrust sheet (towards the village of Christo-Danovo), the Karlovo graben, and the Sredna-gora crystalline block (a neotectonic horst). The Precambrian (Cadomian?) metamorphics (biotite migmatized gneisses) contain numerous granite veins. Prevalent foliation 80º/60ºS. Strong mylonitization and diaphthoresis.

Stop 1.9:
1143 m

Internal thrusting along a surface with 125º/35º: strongly schistified Cadomian? granitoids, intersected by basic dykes (dolerites, lamprophyres), are thrust over Triassic dolomites, and dolomitic limestones with crinoid ossicles (Mogila and Bosnek Formation – Anisian).

Stop 1.10:

The outcrop exhibits well-bedded dark-gray to blackish limestones, with crinoid ossicles and biodetritus, locally oolithic (Vasilyov Formation, Ladinian), with bedding striking 100º/50ºN. They are covered by the Upper Triassic of the Troyan Formation: limestones, brecciated calcareous dolomites, sandstones, locally interbedded with variegated (red and greenish) clays. Rarely, badly-preserved moulds (probably, *Monotis* sp.) are visible. In the upper parts of the section, dolomites with typical weathering are observed, with bedding striking 110º/60º, changing fast to 90-110º/45-55ºNNE. The next Jurassic sequence consists (from bottom to top) of mixed continental and marine reddish-brown sandstones with coal (Bachiishte Formation - Hettangian? - Pliensbachian); quartz sandstones and quartzites (Kostina Formation, Pliensbachian); dark-gray fine-bedded limestones (Ozirovo Formation; Pliensbachian - Aalenian); and oolithic, biodetritic, and sandy limestones (Polaten Formation; Bajocian-Bathonian). Along the road, rock fragments of micritic limestones, some of them with cherty nodules, and red "ammonitico rosso" limestones, come from nearby outcrops of Callovian (Yavorets Formation) and Oxfordian-Kimmeridgian (Gintsi Formation).

Stop 1.11:
Beklemeto Pass - quarry near the mountain ridge: altitude 1399 m

The quarry exposes strongly tectonized Jurassic limestones. The klippe exhibits in an overturned position (from bottom to top): Anisian dolomite (Bosnek Formation); limestones and intraclastic limestones (Anisian Mogila Formation); sandy limestones, marls, sandstones, and fine dolomite beds (Spathian-Anisian Svidol Formation); flakey and variegated sandstones with separate quartz, quartzite, and granite pebbles (Lower Triassic Petrohan Terrigenous Group); strongly tectonized Hercynian? granite.

The neotectonic panoramic view exhibits the Neogene (Maeotian?) planation surface at 1500 - 1600 m over the Stara-planina morphostructure. To the South, the Karlovo graben, and (in the background) the Sredna-gora crystalline block (neotectonic horst). Several planation surfaces are visible, and they are clearly superimposed one upon the other due to the low-velocity uplift of the horst, in contrast with the clearly-defined planation surfaces within the Stara-planina range.

Stop 1.12:
Bridge at the entrance of the town of Troyan.

Upper Jurassic - Lower Cretaceous (in their full range - from Middle Kimmeridgian to Lower Vallanginian) flysch sediments (Cherni-Osam Formation) are exposed in the road cuttings and the picturesque cliffs. They belong to the Central-Balkan Flysch Group, deposited in the so-called Nish-Troyan flysch trough. The sedimentary sequence consists of flysch rhythms, built up of fine- to medium-bedded sandstones, siltstones and marls. Rarely, the flysch sequence is disturbed by thick turbidite beds. The flysch rhythms possess a typical texture sequence illustrative for Bouma rhythms: gradational bedding, parallel bedding, a convolution interval, and they end with marls of the background sedimentation. Some of the rhythms are incomplete. The flysch complex has been formed in the axial part of the basin (Figure 16), and shows transitions to the hemipelagic slope sediments (clayey limestones and marls) of the Salash Formation (to the west), and the coarse terrigenous sediments (conglomerates, sandstones, and marls) of the Kostel Formation (to the east). The flysch complex has a most outstanding development in the Central Fore-Balkan zone. The Nish-Troyan flysch trough had a roughly East-West trend. It is

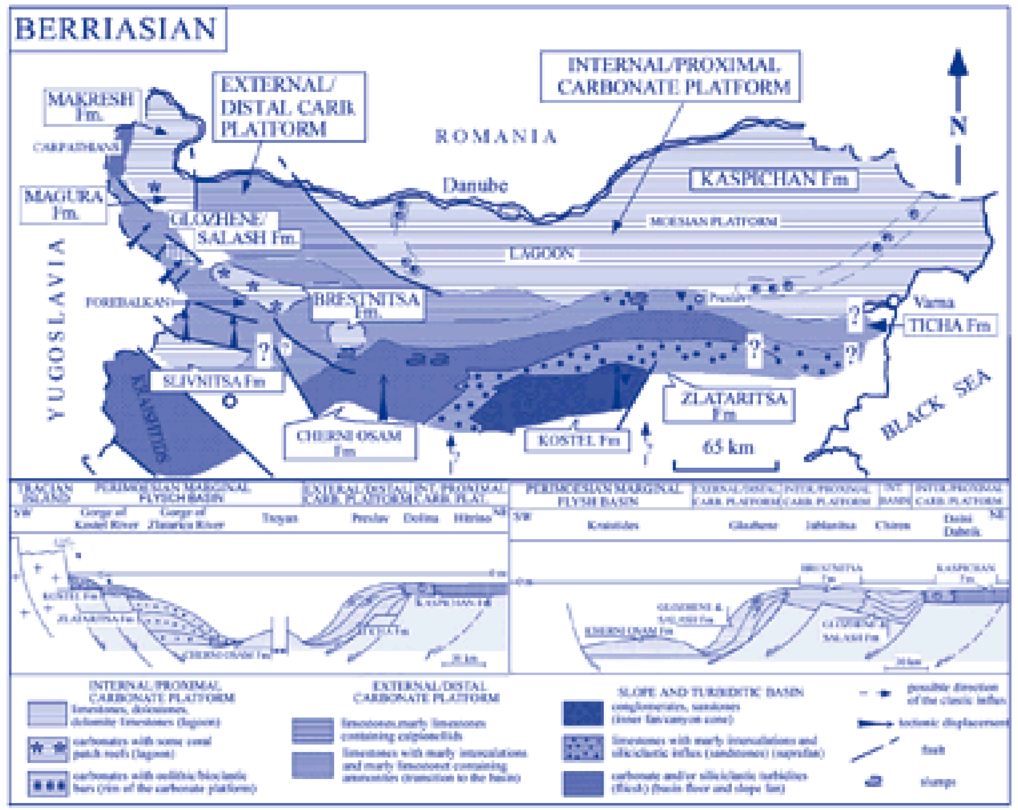


Figure 16 - Palaeogeographic and basin reconstruction of Northern Bulgaria in the Berriasian (after Minkovska et al., 2001)

regarded as a back-arc trough, a foredeep, or else, as a wrench-type basin, with a clear asymmetry.
The journey from the ridge of the Stara-planina (Balkan) mountain range towards the Danube plain (Moesian platform), allows for almost continuous observation of the Neogene surfaces (Table 1). In contrast to the steep southern slope of the Balkan (along the Sub-Balkan fault belt), the northern slope exhibits a gradual lowering of the planation surfaces that corresponds to the flexural transition from the range towards the plain, and to a steady uplift. In Neogene times, the evolution of the relief was greatly controlled by this steady uplift, and by the draining of the fluvial systems towards the Paratethyan marine basin in the Moesian platform.

DAY 2

Lovech - Pleven - Mezdra - Sofia (about 230 km)

Stop 2.1:

Lovech, Stratesh Hill: 224 m
Carbonate rocks of the Stratesh Limestone Formation (Upper Barremian) are exposed. They are a part of the Lovech Urgonian Group (Figure 17), that consists of four terrigenous and four carbonate formations. The limestones of the Stratesh Formation are biomorphic, bioconstructed, biodetritic, oolithic, etc. The principal reef-building organisms are corals, algae, and rudists (genera *Requienia, Toucasia, Monopleura* ? ??.).
The Urgonian complex is a complex biosedimentary system, formed along the northern Tethyan margin. This facies type is traced at thousands of kilometers on the territories of Europe and Asia. Carbonate sequences, typical both for inner and for outer carbonate platforms, are found at the Stratesh Hill. In their alternation, they form parasequences within a transgressive system tract.
The panoramic view of the town looking north exhibits the typical relief of the Fore-Balkan and the Danube hilly plain (Moesian platform) already

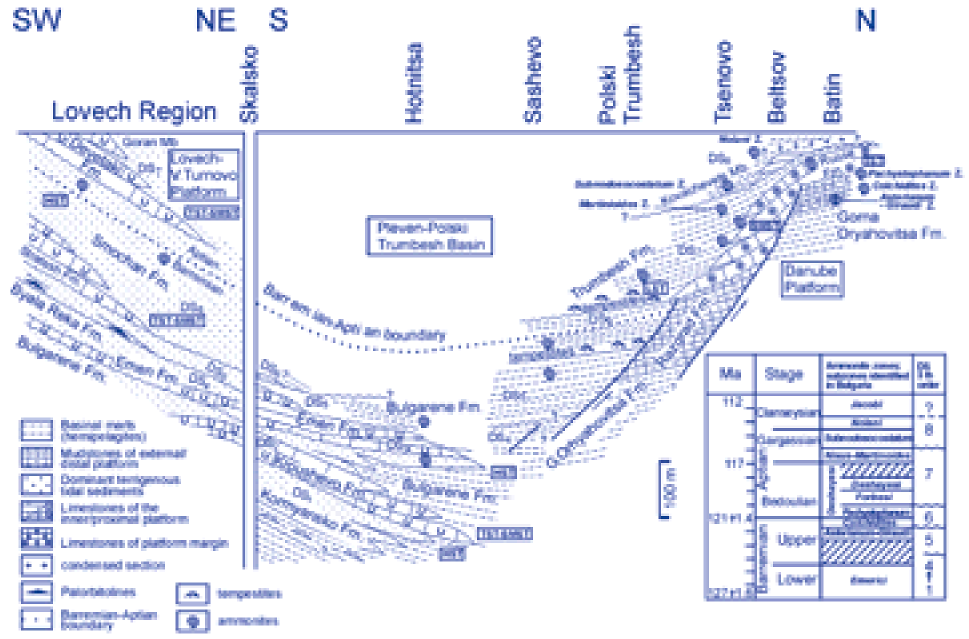


Figure 17 - Basin reconstruction of Northern Bulgaria in the interval Barremian-Aptian (after Ivanov et al., 1997)

mentioned: planation surfaces gradually dipping north, conformable to the drainage pattern of the Neogene – Quaternary fluvial system.

Stop 2.2. NW part of Lovech – road junction towards Goznitsa Suburb: 214 m

The outcrop exhibits the Smochan Terrigenous Formation (Upper Barremian). In the basal parts, it consists mostly of marls and thin layers of sandstones, and clayey and *Orbitolina* limestones. Patch reefs and biostrome bodies have been formed here. They are built of colonial corals, faceloid and dendroid, adapted to a soft substratum (clayey bottom). They are preserved in the rock in situ, the dimensions of the colonies varying from several centimeters to several tens of centimeters. Growth breaks are often visible in the big colonies, due to intense inflow of clayey material. The big colonies form the cores of isolated reefs. Massive colonies dominate the central parts, and dendroid corals, the peripheries. A high taxonomic diversity is found here, both on generic and on species level. The composition of the organisms changes in a vertical sequence. When the growth of the patch reefs is broken, the upper levels are dominated by single corals (*Montlivaultia*), in association with small bivalves and gastropods. The patch reefs and biostromes are formed at a high stand in lagoonal conditions. The rock sequence in this interval is a part of a system tract of high sea level in a third-order sequence.

Stop 2.3:

"Kayluka" Quarry near Pleven: 175 m
The quarry is situated (Figure 18) in the central part of the Moesian platform. The lower parts of the quarry exhibit the limestones of the Kayluka Formation (Upper Maastrichtian). They are biomorphic and biodetritic, and abundant in fossils: bivalves, echinids, gastropods, brachiopods, and rarely, cephalopods (ammonites and nautiloids). Its uppermost levels are dominated by biomorphic limestones. The upper boundary of the formation is marked by an wash-out, and unconformably covered by the Komarevo Formation (Upper Palaeocene): clayey limestones interbedded with marls. A glauconite-bearing layer (20-30 cm to 1-1.2 m thick), is situated at the base of the formation. The limestone - marl interbedding in the middle and upper parts of the formation is in doublets, and probably indicates a climatic Milankovic cyclicity. The Komarevo Formation is

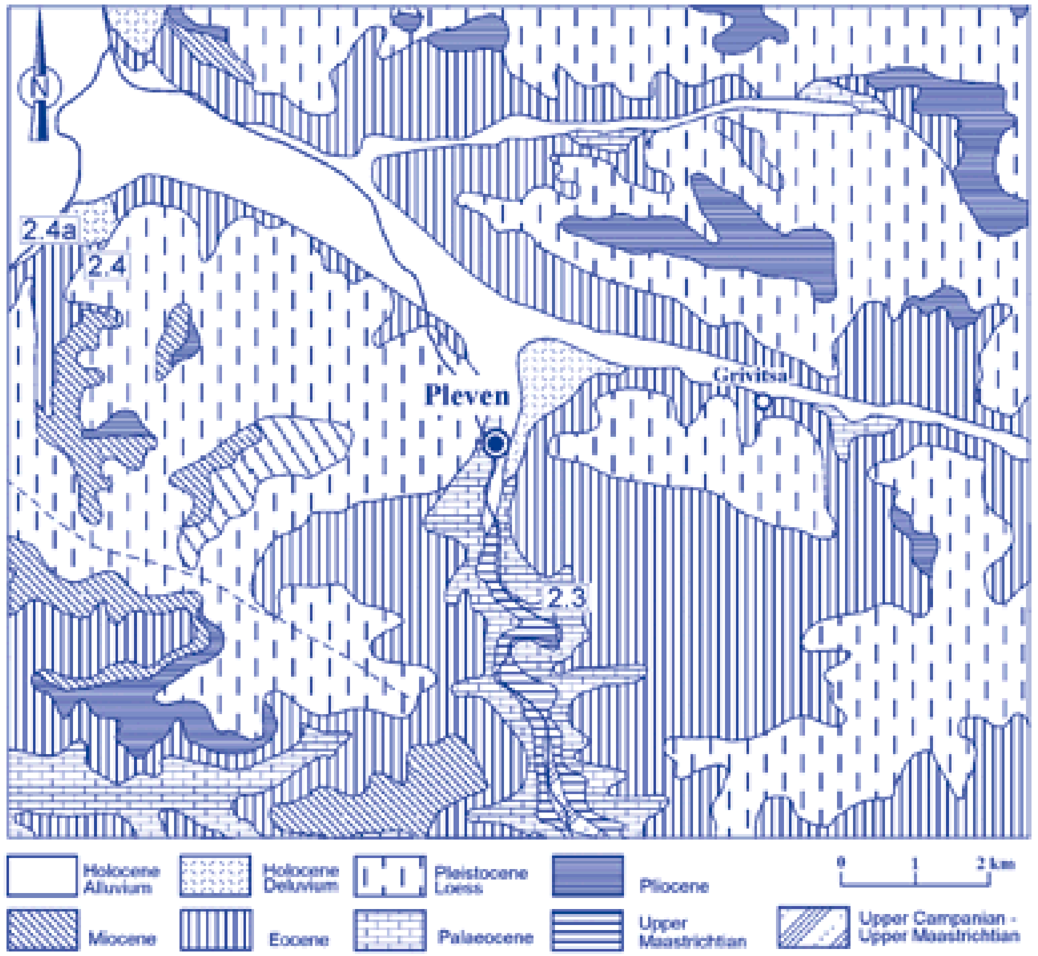


Figure 18 - Geological map of the Pleven Region (after the Geological map of Bulgaria 1:100000)

covered with a sharp boundary and wash-out by the marls of the Avren Formation (Lower Eocene). At the base of this formation, a glauconitic marl is observed, too, as well as nummulitoids.
The section illustrates a rock sequence typical of the geodynamic state of the Moesian platform, with low dips and without important tectonic disturbances.

Stop 2.4:

"Yasen" Quarry: 107 m

The quarry exhibits the marls of the Avren Formation, dated here in the boundary interval Lower-Middle Eocene. They represent a monotonous sequence, mostly of clayey marls in irregular interbedding with silty marls. The following gray-blueish to beige and yellowish calcareous clays of the Opanets Formation (Middle Miocene - Badenian) cover them with a sharp transgressive boundary. They contain abundant fossils of different groups (mostly gastropods and bivalves).
The uppermost parts of the quarry expose Quaternary sediments: loess and palaeosoils with brown color. They cover the Neogene sediments with a sharp erosional boundary. The whole section is typical for the interrelations of the Palaeogene, Neogene, and Quaternary cover sediments, and their interrelations, in the central parts of the Moesian platform.

Stop 2.4a:

Road junction at the destroyed bridge on the River Vit, towards the "Yasen" quarry: 75 m

Photo 2.4: Quarry "Yasen". Eocene transgressively covered by Miocene.

This additional stop exhibits the base of the Neogene (Badenian) section that covers with wash-out and unconformity the Middle Eocene (better exposed at the principal stop).

Stop 2.5:

Quarry next to the road, exit of the village Roumyantsevo: 175 m

This quarry exposes an Upper Cretaceous (Santonian - Maastrichtian) sequence, that transgressively covers terrigenous Lower Cretaceous sediments (Roman Formation, Upper Barremian - Lower Aptian). At the base of the Upper Cretaceous, the Kalen Formation (Upper Santonian) consists of light-brownish to greenish medium-bedded to unclearly-bedded limestones. Small amounts of glauconite are irregularly distributed among them. The next Novachene Formation (Lower Campanian) follows with a sharp boundary, and often after a wash-out. It is represented by chalk-like and clayey limestones interbedded with marls. Light-gray limestones with unclear bedding, and small amounts of glauconite (Roumyantsevo Formation, Upper Campanian; holostratotype section) follow with a sharp boundary. This formation is covered with a sharp boundary by glauconitic sandstones and glauconitic sandy limestones (Durmantsi Formation, uppermost Campanian - lowermost Maastrichtian). This formation is a typical and well-traceable marker in the Upper Cretaceous section in NW and central north Bulgaria, due to the sharp boundaries and the comparatively small thickness. With a sharp boundary again, the Durmantsi Formation is covered by clayey and micritic limestones with a nodular appearance, and abundant flint nodules (maximum range Lower Maastrichtian - Danian). The limestones build layers of moderate thickness, often interbedded by very thin (less than 2 - 3 mm) clayey-silty beds. The Upper Cretaceous section ends with a thick (>200 m) limestone sequence (Kayluka Formation, Upper Maastrichtian).
The section is typical of the platform (northern) facies of the Upper Cretaceous that is widespread over the Moesian platform and the Fore-Balkan. Here it is exposed within the flexural boundary (Bresnitsa-Preslav flexure) between these two units.

Stop 2.6:

Along the motorway, after the road junction to Dzhourovo (at the sign "Sofia 70 km").

At the stop, a nice panoramic view of the Fore-Balkan and Balkan is afforded, with the principal planation surfaces. Farther along the road, an anoxic blackish

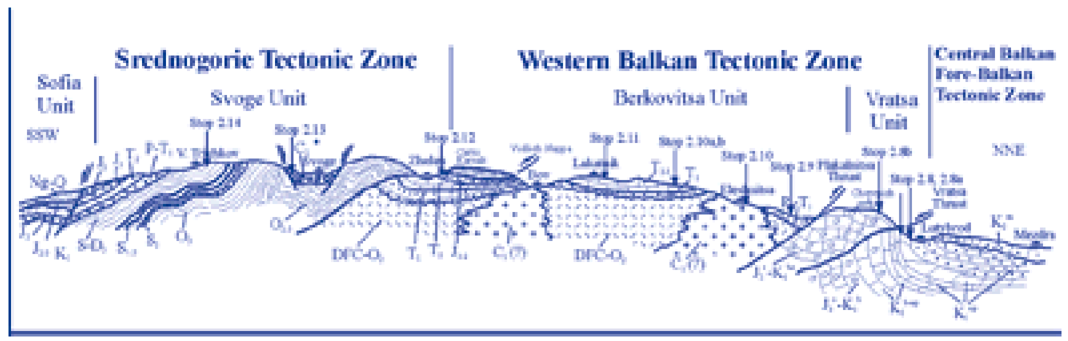


Figure 19 - Geological section along the Iskur valley (modified after the geological maps of Bulgaria 1:100000)

Volume n° 2 - from B16 to B33

flysch is exposed in the road cutting. It belongs to the Upper Jurassic - Lower Cretaceous Cherni-Osam Formation, deposited in the Nish-Troyan flysch trough.
The motorway crosses one of the ridges of the Balkan by a tunnel, and closely afterwards, the itinerary deviates along the road towards Botevgrad.

Stop 2.7:

To Botevgrad, before the right turn: 400 m

The stop exhibits a panoramic view of the Botevgrad graben. This is an internal depression within the Stara-planina positive morphostructure filled in by Neogene (partly) and Quaternary sediments. The graben is situated between the Rzhana horst (to the north; mostly built up of Mesozoic sediments of the Fore-Balkan), and the Mourgash horst (Palaeozoic, mostly Ordovician sediments), to the south. The Botevgrad graben itself has a basement typical of the Berkovitsa unit: Neoproterozoic-Cambrian low-grade metamorphics of the Berkovitsa Group intruded by Variscan granites. The graben is asymmetric: the steep high-amplitude Dragoy-Balkan normal fault to the north is well-expressed in the relief, and covered by steeper and narrower alluvial fans, whereas the southern edge is step-like, with large alluvial fans originating at small-amplitude normal faults.

The itinerary continues (Figure 19) to the valley of the River Iskur (at the Iskur Gorge). This is the only river that has its origin in south Bulgaria (Rila Mountain), and crosses the Stara-planina (Balkan) Range to the north to enter the Danube. The valley has an antecedent character: it existed in Neogene times, before the uplift of the Stara-planina Range, and had been gradually incised in the uplift process.

Stop 2.8:

Lyutibrod: 289 m
(after "Ritlite" - a geomorphic geosite).

Sediments from the upper parts of the Urgonian complex (Vratsa Urgonian Group, Lyutibrod Formation: Lower Aptian) are exposed. They are detritic, oolithic, orbitolina-bearing, and other limestones, in alternation with terrigenous-carbonatic packets (interbedding sandy limestones, calcareous sandstones, and marls). They are typical of the external carbonate platform or marginal environments. Tempestites are often present. Detailed observations prove the presence of retrogradational and progradational tendencies and sequences. In combination they form 3rd order sequences with eustatic genesis. Both limestones and marls contain abundant and various fossils (bivalves, gastropods, corals, brachiopods, and orbitolinas). The Urgonian complex along the Iskur valley, between the Cherepish Monastery and Lyutibrod, is near the boundary between two tectonic units: the West-Balkan and the Central-Balkan – Fore-Balkan ones. The boundary

Photo 2.8 - "Ritlite", Iskur Gorge. Boundary between the West-Balkan and Fore-Balkan units.

between them has a thrust or upthrust character. Consequently, the packets of the Lyutibrod Formation are in an upright position, the soft rock intervals are eroded, and the vertical limestone packets form the natural phenomenon (geomorphic geosite) "Ritlite".

Stop 2.8a:

After the tunnel: 289 m

The sediments of the Cherepish Formation (Lower – Upper Barremian) are a part of the Vratsa Urgonian Group. They are mostly limestones in thick packets, divided by terrigenous-carbonate packets. Pachiodont and bioclastic limestones are dominant. Oolithic and micritic limestones are frequent in the carbonatic packets, whereas the terrigenous-carbonatic packets are dominated by the sandy limestones and calcareous sandstones. They often contain structures typical for shallow environments: cross and oblique bedding by oolithites and the detritus, and HCS in tempestites, situated at several levels in the formation. The rhythmic sediments, with high fossil content and participation of clayey rocks, are the result of storm phenomena. Ripple-marks in the hanging wall of sandstone layers, and strong bioturbation in marls and sandstones, are also typical. Carbonatic and terrigenous-carbonatic sediments are genetically related to transgressive–regressive cycles. Several 3rd-order sequences are distinguished in this stratigraphic interval of the Vratsa Urgonian Group (Cherepish and Lyutibrod Formation). Palaeogeographically, the Urgonian complex was formed in a basin already closing in Late Barremian and Aptian times, as a typical carbonate platform probably of a ramp type, and without a well-expressed shelf margin.

Stop 2.8b (photostop):

291 m.

The stop exhibits the Cherepish Monastery and the meanders of the Iskur River around cliffs built up of light-gray to white massive limestones (Brestnitsa Formation, Upper Jurassic - Lower Barremian). Micritic and biomorphic varieties are dominant. Less frequent are biomorphic (rudist) limestones (*Diceras* sp.). The limestones of the Brestnitsa Formation are typical of the internal carbonate platform. It is an element of the chain of carbonate platforms situated along the northern margin of the Tethys in the Callovian – Vallanginian time span.

Stop 2.9:

281 m

After a fault structure, the itinerary leaves the Fore-Balkan, and enters the West-Balkan (Berkovitsa) unit of the Balkanides. Permian red conglomerates and sandstones, with pebbles of granite and lydite, are exposed. The bedding is 125-135º/50-60ºNE.

Stop 2.10:

Eliseyna, facing the railway station: 316 m

The outcrops in the road cuttings exhibit Permian andesites (porphyrites) from the core of the Berkovitsa unit.

Stop 2.10a:

349 m

In this outcrop, the basal parts of the Lower Triassic Petrohan Terrigenous Group are exposed. They cover the Carboniferous – Permian volcanics. Two formations are distinguished: lower (sandstone) and upper (sandstone-marly) formation.
In its basal parts, the sandstone formation consists of fine- to medium-grained oligomictic to polymictic sandstones, interbedding with siltstones and marls. These three rock types form cycles with different thickness. Fine and moderate rhythmicity prevail. The sandstones contain also pebbles, intraclasts (from sandstones and claystones), and extraclasts (from red Permian sedimentary rocks). The red color is dominant, beige to whitish beds are also present. Cross and oblique bedding are typical. Internal erosional surfaces, and discordant layering of some packets are frequent. The sediments of the group were formed in a fluvial environment and arid climate. They build the base of a transgressive cycle with progressive retrogradation.

Stop 2.10b:

North of Bov (panorama of the Triassic): 373 m

The section exhibits the base and the sequences of the Triassic System of the western Balkanides. The Triassic covers with angular and regional unconformity different Palaeozoic complexes. It is subdivided into three groups.
The Lower Triassic Petrohan Terrigenous Group consists of clastic rocks: conglomerates and coarse- to fine-grained sandstones, interbedded with siltstones, and dark- to light-red shales. It is formed in fluvial systems: in an alluvial plain, in alluvial fans and under aeolic conditions, in an arid to semi-arid climate.
The Iskur Carbonate Group (uppermost Lower Triassic - Upper Triassic) contains carbonatic sequences formed in shallow marine environments. Its formal lithostratigraphy, as delineated by D. Tronkov, consists of 6 formations and several members. The lower boundary represents a fast transition from the Petrohan Terrigenous Group, and marks the marine transgression. The transition interval has a terrigenous-carbonate character, with the presence of fine-grained clastic or clayey rocks, and gray limestones and yellowish dolomites. Upwards in the sequence follow: rhythmic rocks (alternation of bioclastic micritic limestones and dolomites) formed in sublittoral and littoral environments; massive micritic and bioclastic limestones (with a carbonate bank geometry); and nodular and bioclastic limestones (formed on a carbonate ramp – submarine slumps and storm products, tempestites and tsunamites are observed). The group ends with massive dolomites.
The Triassic section ends at most places with the Moesian Group: marine red beds that correspond to a regressive sequence. It consists of various varicolored rocks: from boulder breccia-conglomerate and conglomerate, polymictic to oligomictic sandstones, siltstones, to shales, marls, limestones, and evaporites. The beginning of the Moesian Group is dated at different places as lowermost Carnian, Upper Carnian, Lower or Middle Norian. It ends also diachronously, and is followed by a significant hiatus

Volume n° 2 - from B16 to B33

(due to the Late Triassic orogeny), and erosion.

Stop 2.11:

Karst spring "Zhitolyub": 377 m

The cliffs on the left bank of the River Iskur near the karst spring (near the railway station of Lakatnik), exhibit an almost full section of the Iskur Carbonate Group, which was thoroughly studied by D. Tronkov. The lowermost parts of the section close to the spring are built up of cyclic sediments (Opletnya Member of the Mogila Formation; Lower Anisian). Each cycle consists of 3 components, in a regular alternation, and an almost constant bed thickness. The first component is a gray allochemic limestone, the second, micritic limestone, and the third one, a dolomite (that becomes yellowish to beige with weathering). Allochems are represented by ooids, bioclasts of different marine fossils, and lithoclasts. Boundaries between the adjacent cycles have often the character of a wash-out (inundation) surface. The member was formed in littoral to sublittoral environments with a varying hydrodynamic regime: dynamic medium in the lower parts of each cycle, and quiet in the upper ones. The tendency is retrogradational, and typical of a transgressive regime. Low-order cyclicity shows all signs of a climatic control, and probably represents a Milankovic cyclicity.

Stop 2.12:

Tserovo, junction to Zimevitsa: 476 m

The hill Cherni Kamak east of the river exhibits a panoramic view of the contact (Vidlich tectonic zone) between the Berkovitsa and Svoge units. West of the Iskur, it is a steep upthrust, whereas to the east, it passes into an overthrust. The allochthon is built up of pre-Mesozoic complexes of the Svoge unit, and the autochthon, of Triassic and Jurassic formations of the Berkovitsa unit.

The basement of the Berkovitsa unit consists of a Neoproterozoic - Cambrian diabase-phyllitoid complex, intruded by Lower Carboniferous granitoids. The Mesozoic cover contains Triassic (Petrohan Terrigenous Group, and lower parts of the Iskur Carbonate Group – not higher than the Anisian Mogila and Babino Formation), unconformably covered by Lower Jurassic (Kostina and Ozirovo) and Middle Jurassic formations. The allochthon (Svoge unit) is strongly tectonized, and consists of sandstones and siltstones of the Ordovician (Grohoten Formation). Jurassic sediments from the autochthon have been also involved in thrusting as parautochthonous lamellae (thrust horses) in the frontal parts of the thrust (*Figure 20*).

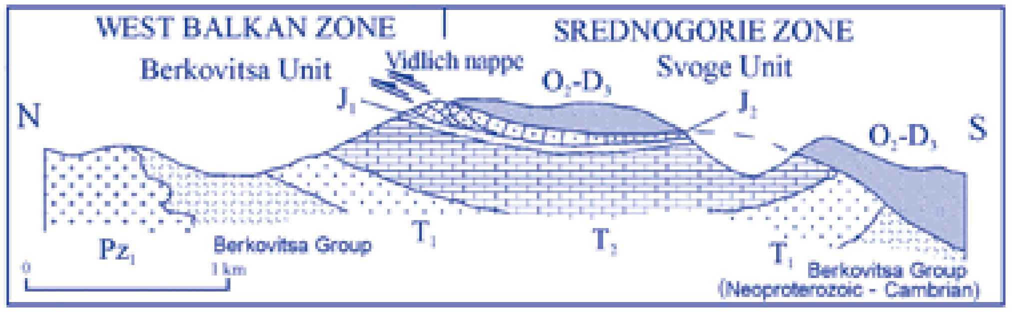


Figure 20 - Simplified sketch of the contact zone between the Svoge and Berkovitsa tectonic units near Tserovo (the Cherni Kamak hill)

Stop 2.13:

South of Svoge: 548 m

The road cuttings exhibit a part of the Carboniferous sequence of the Svoge coal basin. The continental sediments have been formed in different (fluvial, lacustrine, and paludal) environments. They cover discordantly a varied basement: Ordovician, Silurian, and Devonian formations. The lowermost Carboniferous sediments are coarse terrigenous sequences (Namurian Tsarichina and Svidnya Formation). Comparatively finer-grained (sandstones, siltstones, and shales) coal-bearing sediments follow (coal measures between 0.2 and 8 m thick) (Westphalian A: Drumsha and Svoge Formation).

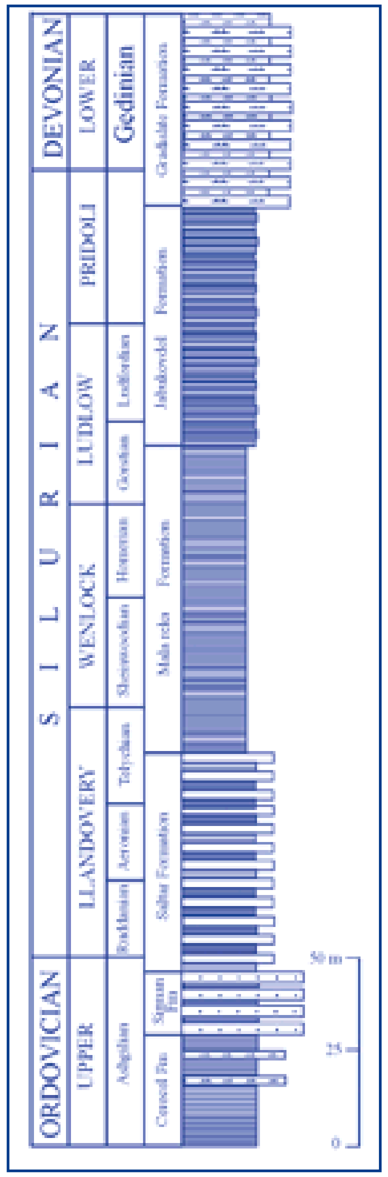


Figure 21 - Simplified section of a part of the Palaeozoic in the Svoge tectonic unit

The uppermost parts of the section are built up again of coarser terrigenous sequences of Westphalian B (Berovdol Formation), and Westphalian B – C, and (partly) Westphalian D (Chibaovtsi Formation) age. The section exhibits several wash-out and erosional boundaries. The internal architecture often exhibits a cyclic character, forming cyclothems with parasequences. The abundant fossil megaflora has been thoroughly taxonomically studied and published by Y. Tenchov.

The road cuttings expose rocks of the Svoge Formation: sandstones and conglomerates that cover an alternation of medium-bedded sandstones and shales. They are polymictic to oligomictic, with parallel or oblique bedding, typical of fluvial sediments. Huge (decimetric to metric) fragments of branches and trunks of megaflora (*Lepidodendron* and *Calamites*) are abundant. They are in allochthonous position, and possess a preferred orientation, indicating a fluvial transport. A part of the fragments are carbonized, and others changed into coal. It has been proposed to add this outcrop to the geosite list under State protection.

Stop 2.14:

Along the road NE of Vlado Trichkov: 585 m

Ordovician and Silurian sediments from the Svoge tectonic unit (Figure 21) are exposed in the road cuttings. Several formal lithostratigraphic units have been introduced and thoroughly studied by Y. Tenchov and V. Sachanski. The Tseretsel Formation (Ordovician) consists of gray-green shales with non-distinct bedding. The Upper Ordovician (Sirman Formation) is represented by interbedding of dark-gray to black sandstones and shales with graptolites. Further upwards, fine-bedded black and gray shales, gray chertified shales, and banded shales are interbedded with black or gray lydites (Saltar Formation; uppermost Ordovician - Lower Silurian). They contain rich graptolite fauna that allows the exact positioning of the Ordovician/Silurian boundary. The next Mala-Rechka Formation (Silurian) consists of monotonous (but rich in graptolites) fine-bedded black shales.

The uppermost Ordovician and the Silurian of the Svoge unit contain event markers that have been well-correlated with the event stratigraphy schemes of Europe. Such a marker is represented by the sandstones of the Sirman Formation, probably formed at a low sea stand. The interbedding of light- and dark-colored sedimentary packets is explained with climatic cyclicity.

At the town of Novi Iskur, the route enters the Sofia neotectonic graben. The northern boundary of the graben is a steep Neogene-Quaternary fault that is now also the surface expression of the boundary between the Svoge unit, and the central part of the Srednogorie zone.

DAY 3

Sofia - Doupnitsa - Rila - Rila Monastery - Padesh - Blagoevgrad (about 190 km)

Sofia graben and Vitosha unit south of Sofia

The route South of Sofia crosses the southern part of the Upper Cretaceous Srednogorie superunit. The Vitosha Mountain itself constitutes the principal part of the Vitosha unit. It consists mostly of Coniacian – Santonian volcanic rocks (mainly andesites), and their tuffs intruded by the Vitosha pluton. The Vitosha pluton is built up of three consecutively intruded magmatic phases, that illustrate the development of a

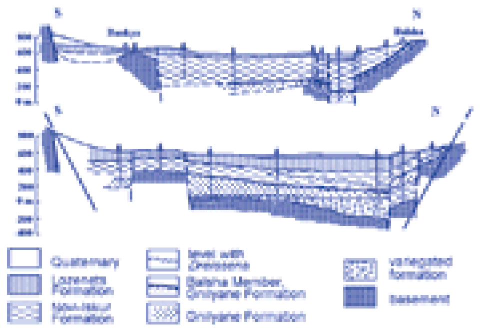


Figure 22 - Cross sections through the Sofia graben (after Kamenov and Kojumdgieva, 1982)

deeper source in the upper mantle or depleted lower crust (low initial ^{87}Sr/^{86}Sr ratio – about 0.704!) of the Srednogorie zone. The first phase consists of gabbro and gabbro-monzonite, the second, - of different varieties of monzonites, and the third one, - of granosyenite in dykes or small bodies.

The mountain represents a neotectonic horst uplifted at more than 1.5 km above the Sofia neotectonic graben (Figure 22). The graben has a heterogeneous basement: the Neogene deposits (with a total thickness of up to 1000 m) cover with depositional, unconformable contacts different rock units in small pre-Middle Miocene internal blocks: Upper Cretaceous volcanics or sediments, Jurassic and Triassic carbonate or terrigenous rocks, and even (in a very few localities), the pre-Alpine basement.

The Neogene section of the Sofia graben begins with a variegated terrigenous formation, referred to the Maeotian Stage. The second local cycle (Pontian) is related to a lacustrine environment, and is represented by the coal-bearing Gnilyane and Novi-Iskur formations. The cycle begins with a slight local unconformity over an eroded surface. A similar unconformity is even better expressed at the bottom of the last, Lozenets Formation (Dacian – Lower Romanian). It consists of well-sorted yellowish to whitish conglomerates and sands. The whole sedimentary filling of the graben is intersected and displaced by neotectonic (Quaternary) faults, some of them even displacing the Quaternary river terrace sediments.

The itinerary follows the road E-79 and crosses the westernmost periphery of the Pernik Palaeogene basin (Upper Oligocene – lowermost Miocene), which is made up of lacustrine coal-bearing sediments. The basin is a graben (Figure 23), situated over the southernmost parts of the Srednogorie superunit. In the new road cuttings, varicoloured (gray, green, and red) conglomerates, sandstones, and shales may be observed. At Stoudena, the road enters the Verila unit of the Srednogorie: mostly Triassic carbonate beds (Iskur Carbonate Group) that dip monoclinally south. The road cuttings exhibit dolomites of the Anisian Bosnek Formation, well-bedded nodular dark-gray limestones with shaly interbeds (Ladinian; Radomir Formation), and dolomites and dolomitic limestones (Carnian; Rousinovdel Formation). The latter is covered with unconformable depositional contact by the Middle Jurassic (not exposed).

Stop 3.1:

Staro selo: altitude 784 m

The road cutting exhibits Tithonian flysch: rhythmic interbedding of shales, siltstones, sandstones, and (rarely) conglomerates. The beds are striking about 140º, and dip SW at an angle of 35-40º. They are referred to as the Upper Jurassic – Lower Cretaceous Kostel Formation, and have been deposited in the Nish-Troyan flysch trough, strongly folded and displaced in Mid-Cretaceous, Late Cretaceous and Palaeogene times. Sediments of the same trough have already been observed in the Troyan area.

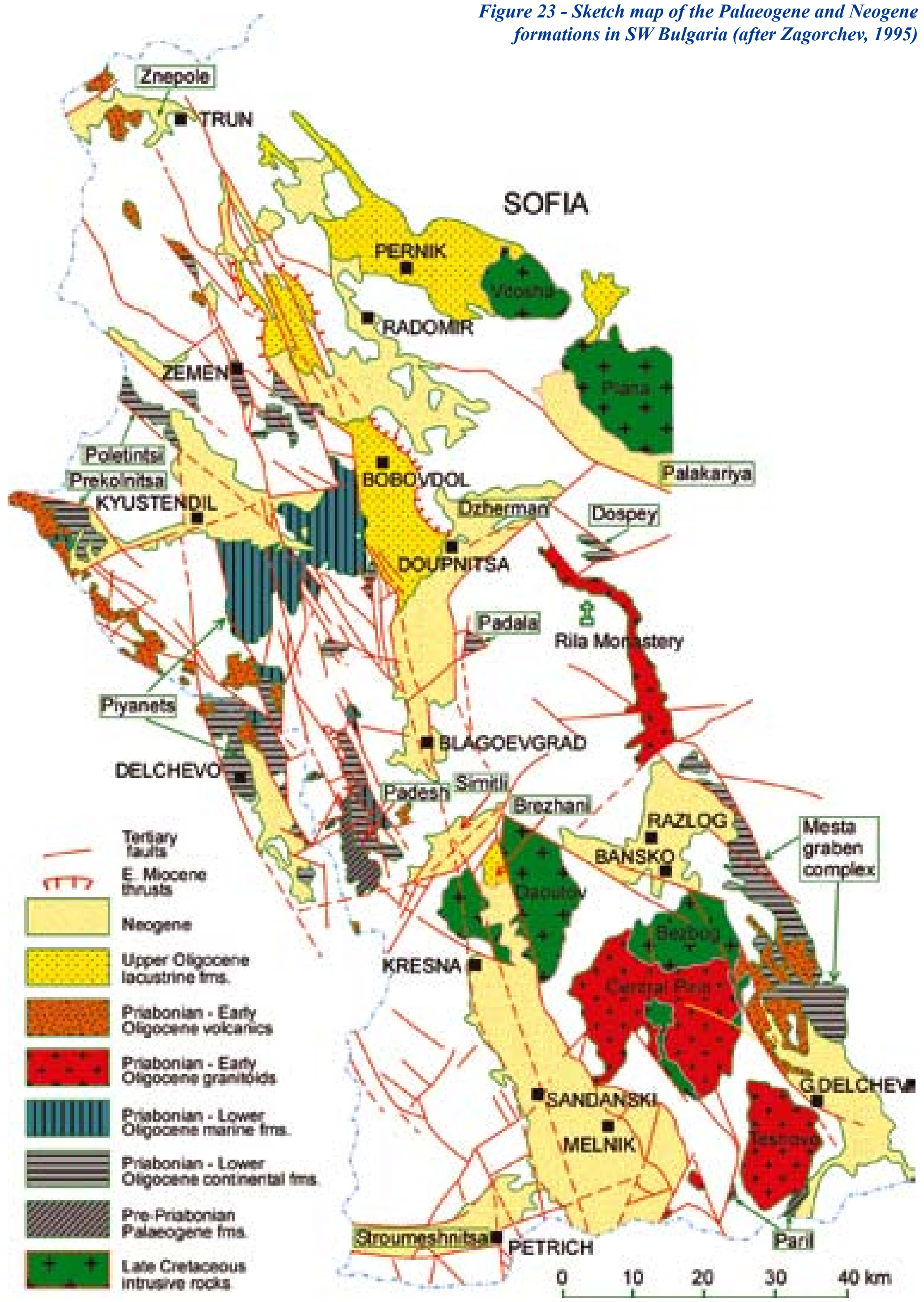
Stop 3.2:

Near Doupnitsa, 557 m: panorama of the Rila horst

The Rila horst has been uplifted relative to the Blagoevgrad and Dzherman grabens, with an upthrow of more than 3 km since Late Miocene times. The moderately inclined, to steep bounding Klisoura normal fault (Figure 24), strikes NE, and has an angle of dip between 30º and 70º to NW. It has an outstanding geomorphic expression.

Volume n° 2 - from B16 to B33

Figure 23 - Sketch map of the Palaeogene and Neogene formations in SW Bulgaria (after Zagorchev, 1995)



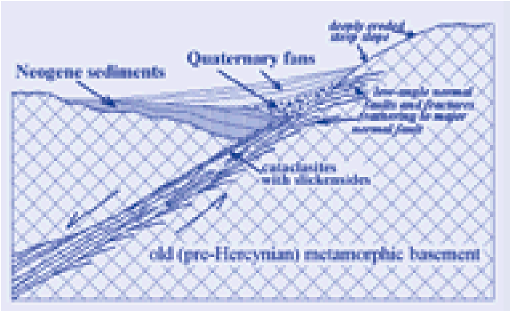


Figure 24 - Schematic section across the Klisoura fault

The stop is situated over a planation surface that represents a part of the Lower – Middle Miocene peneplain. It gradually lowers to the South, and is subsided in the Dzherman graben to altitudes about and below sea level, being uplifted to more than 2500 m in the Rila horst.

Blagoevgrad basin

The stratigraphy of the Neogene and Eopleistocene filling of the Blagoevgrad graben has been studied in detail. Here it is schematically outlined.

The section begins locally with red polymictic conglomerate that covers unconformably Palaeogene and older rock units and structures. It is followed by proluvial polymictic conglomerate, interbedded with sand and sandy clay (Pokrovnik Formation). The next Dzherman Formation (Maeotian to Pontian? after mammal fauna) consists of alluvial greenish, whitish, or yellowish sand and clay interbedded with pebble gravel lenses. The Barakovo Formation (Pontian? - Pliocene) is composed of well-sorted alluvial to proluvial whitish or yellowish pebble gravels or sands, with cross or gradational bedding. It covers the washed-out erosional surface on the older Neogene or pre-Neogene rocks. In the western slope of the Rila Mountain, an erosional surface built over the Dzherman and Barakovo Formation is covered by thick unsorted or chaotically deposited Eopleistocene to Pleistocene reddish, brownish to yellowish pebble gravel, sand, and sandy clay, all containing pebble fragments. The road cutting near the small town of Kocherinovo exposes well-sorted conglomerate and sandstone of the Barakovo Formation (Pontian? – Pliocene).

Table 2. Altitudes of the planation surfaces in SW Bulgaria

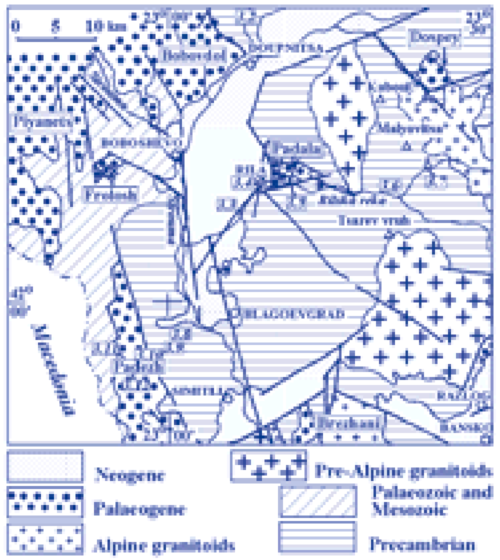


Figure 25 - Simplified map for the itinerary of the third day

Stop 3.3:
At the village of Porominovo: 394 m

Panorama (to the SE) of the very young (Quaternary) NE-striking fault along the river Rilska and the lowermost terraces of the river; in the background – erosion forms in the Neogene and Pleistocene, with the high terrace over the two formations. The West-Rila fault zone between the Rila horst and the Blagoevgrad graben is visible to the NE.

In the town of Rila, we cross a narrow horst, built up of rocks of the Cadomian Strouma diorite formation (diorites and quartz-diorites cut by dolerite dykes). At the eastern exit of the small town, south of the "Krusta" hill (the Cross), we enter the post-sedimentary Padala graben. Several stops (Figure 25) are briefly described.

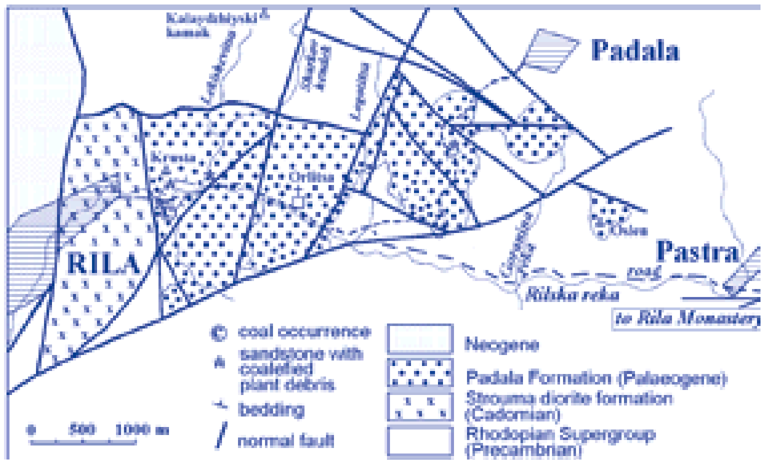


Figure 26 - Geological map of the Padala graben (after Zagorchev et al. 1999)

Stop 3.4:
Exit of the town of Rila (sign) on the road to Rila Monastery: 531 m

Panorama of beautiful erosion forms: the "Krusta" hill and adjacent hills formed in the Palaeogene conglomerate and sandstone of the Padala Formation are the majestic natural gates towards Rila Monastery.

The Padala graben (Figure 26) is filled in by the 350 -400 m thick Padala Formation. It consists of coarse breccia and conglomerate interbedded by sandstone with abundant coalefied plant debris (parts of leaves and stems). The pebbles in the conglomerate come from the typical basement rocks that occur in the Rila Mountain: biotite gneisses and migmatites, amphibolites, marble, granite, pegmatite, quartz, and diorite. These basement rocks have been attributed to the Tertiary age as early as in 1844 by Ami Boue, and in 1912, G. Bonchev reported for the first time the presence of a coal lens. An Early Oligocene age has been proven by S. Chernyavska with typical palynomorphs. Thus, the Padala Formation was formed in a part of the alluvial to lacustrine system feeding the Early Oligocene gulf. The graben has a post-sedimentary character. However, some of the normal faults at its northern periphery have pre-dated the sedimentation, and others were active during the sediment deposition. Some syn-sedimentary faults are filled in by neptunian clastic dykes. Post-sedimentary steep, moderately-inclined and low-angle normal faults, are observed to the east and within the sediments of the graben.

Photo 3.4 - Town of Rila. Palaeogene Padala Formation.

Stop 3.5:
At 150 m after the Metoh (monastery branch) "Ortlitsa", near the road junction for Padala, altitude 567 m

The metoh was first mentioned in 1378 in the Chrysovoul (Charter) of the Tsar Ivan Shishman.

The conglomerate is built up mostly of gneiss, schist and amphibolite pebbles, coming from the adjacent

Volume n° 2 - from B16 to B33

basement rocks. The pebbles are chaotically situated in the sandy matrix. Several steep normal faults are visible, with slickensides striking east (90º) (with a dip of 80ºS and subhorizontal striae) and north (360º). Two-mica gneisses, affected by diaphthoresis, follow.

The road enters deeply into the Rila Mountain (West-Rila unit of the Central-Rhodope superunit). It follows the River Rilska-reka, crossing biotite gneisses and amphibolites of the Precambrian Rhodopian Supergroup (Rupchos Group), intensely folded and intersected by the granitoids (granodiorite, granite, quartz-monzonite) of the Kalin pluton. Traditionally, the pluton has been referred to the Upper Palaeozoic, but it is highly probable to be of Alpine age.

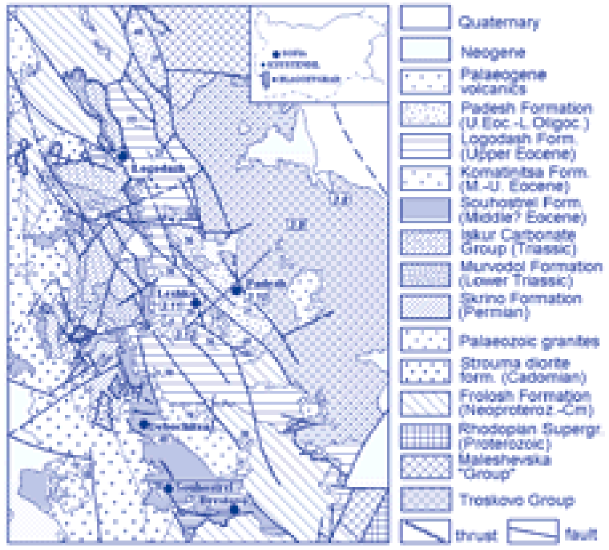


Figure 27 - Geological map of the Vlahina and Lisiya horsts and the Padesh graben (Zagorchev, 2001)

Stop 3.6:

Near the confluence of the Rivers Manastirska and Iliyna: 897 m

This stop exposes in the road cuttings the Chepelare Formation of the Rhodopian Supergroup (Proterozoic). Migmatized amphibolites (probably at least partially of ortho-origin), with a clear foliation, are intersected by pegmatite dykes (coarse-grained muscovite pegmatite). The amphibolites are interbedded with biotite gneisses (often graphite- and garnet-bearing). The foliation strikes 155-170º, and dips at 60º to the west. Late quartz veins. A significant post-pegmatite deformation is observed, with partial diaphthoresis in gneisses and amphibolites, gliding along the contacts of the pegmatites, with the formation of thin schistified zones, and fracturing inside the pegmatites.

Photo 3.7 - Rila Monastery

Stop 3.7:

Rila Monastery: 1103 m altitude

Rila Monastery is the biggest monastery in Bulgaria, and the most important national spiritual center in the Middle Ages and the National Revival in the 18th and 19th centuries. It was founded in the 10th Century by Saint Ivan Rilski (St. John of Rila), and has been ravaged and set on fire several times. The oldest building preserved is the Hrelyu Tower (1335), built by a local feudal lord. The other buildings are dated from the first half of the 19th century. Important manuscripts, treasures, and frescoes can be seen in the church and the museum.

From Rila Monastery the itinerary follows back to the main road, and from there, to Blagoevgrad and Padesh (Figure 27).

Stop 3.8:

After the village of Pokrovnik: 409 m

Cross-bedded fluvial sediments (loose yellowish and whitish sandstone and conglomerate) of the Pontian – Pliocene Barakovo Formation. They belong to the western periphery of the Blagoevgrad graben, partially covering the eastern part of the Lisiya horst.

Volume n° 2 - from B16 to B33

Stop 3.9:

529 m

The road cutting exhibits amphibolites, and hornblende-biotite and biotite gneisses, of the Troskovo Group, Ograzhdenian Supergroup. The foliation strikes NE (45º), and dips at 25º to NW. Interbeds of leptitoid (quartzo-feldspathic) gneisses are also observed. Numerous coarse-grained dykes of muscovite pegmatite have been prospected for feldspar in the past.

Stop 3.10:

At the entrance of Padesh: 614 m

Panorama of the Padesh graben, with the post-sedimentary Vlahina horst at the background.

The Lisiya horst is built up of gneisses and amphibolites of the Ograzhdenian Supergroup. The Padesh graben is separated from this horst by the Lisiya normal fault zone, that was first activated in Late Eocene time, during the sedimentation if the Logodash Formation. Tongues of coarse conglomerate and breccia, built up exclusively of rocks of Lisiya-horst provenance, are interfingering with sandstones and shales of the formation. The Logodash Formation covers coal-bearing and flysch-like sediments of Middle to Late Eocene age, and is covered with a sharp depositional contact by Upper Eocene – Lower Oligocene pyroclastic and sedimentary rocks of the Padesh Formation. The initial peneplain, built over the Padesh Formation, is covered unconformably by Neogene deposits.

The Vlahina horst is built up of Cadomian diorites, covered by Permian and Triassic sedimentary formations. Outcrops of Palaeogene deposits are preserved over the horst, thus pointing at its post-sedimentary (after deposition of the Padesh Formation) character.

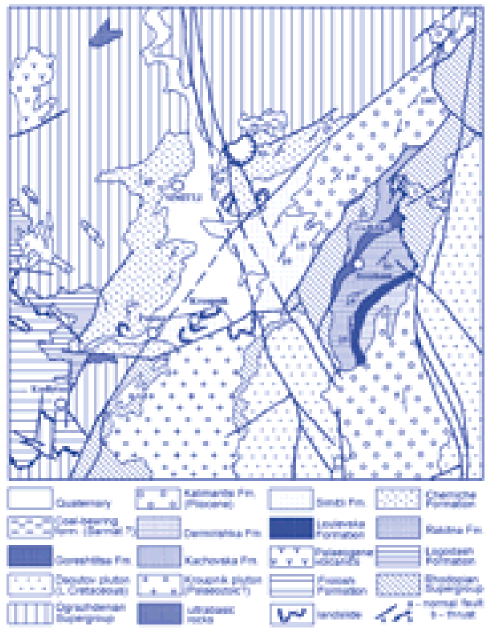


Figure 28 - Geological map of the Simitli graben (modified after I. Zagorchev and N. Dobrev)

Stop 3.11:

On the road Leshko - Gorno Leshko: 627 m

Layers (up to 100 – 120 m thick) of coarse conglomerate and breccia, built up exclusively of gneiss, amphibolite, and pegmatite pebbles and boulders, are interbedded with sandstones, siltstones, and clays (Logodash Formation). Some of the thin (5 cm) interbeds are rich in carbonate matter, with interwoven ichnofossils. Coalefied plant debris is present, too.

Stop 3.12:

On the road Padesh - Gabrovo: 588 m

Along an interval of about 200 m, the road cuttings exhibit the boundary between the Logodash and Padesh Formation. The Logodash Formation is represented by the typical fine-pebble conglomerates, sandstones, siltstones, and shales, interlayered with coarse to boulder breccia and conglomerate. It is covered with sharp depositional contact by lithoclastic, crystaloclastic, and vitrocrystalloclastic phenodacitic tuffs of the Padesh Formation. They are followed by sandstones and conglomerates, and by the Ovnarska-chouka marker bed (Eocene-Oligocene boundary), built here of carbonatic fossil-rich sandstone. Further on, sandstones and conglomerates, rich in weathered, rounded pebbles of volcanics and tuffs follow. The bedding is almost monoclinal, with beds striking 135-155º, and dips between 30 and 40º to NE. Numerous steep (70-80º) neotectonic normal faults strike 145-165º.

Volume n° 2 - from B16 to B33

DAY 4

Blagoevgrad - Brezhani - Kresna - Mikrevo - Ilindentsi - Melnik (about 160 km)

Stop 4.1:

Bridge above the main road: 308 m

A panorama of the Kroupnik active normal fault is exhibited. It bounds at the South the Neogene asymmetric Simitli graben (Figure 28). The total vertical offset is estimated at about 3 km. The fault strikes NNE-SSW, and crosses the NNW-SSE Strouma fault in the northernmost part of the Kresna Gorge of the river Strouma. The two fault zones form a peculiar seismotectonic fault knot that was the site of several devastating earthquakes in historic time, the earthquake (4.04.1904) with the highest magnitude (7.83) on the Balkan Peninsula included. To the South, the narrow Kachovska horst divides the Simitli graben from the Palaeogene of the Brezhani graben. In the background, the faults of the West-Pirin fault zone form the western boundary of the Pirin horst. The uppermost planation surface in Pirin (at an altitude of about 2600 m) and some of the lower surfaces (at c. 2000-2200, and 1500-1600 m) are visible in the far background, too.

Stop 4.2:

Road to Brezhani: 537 m

The road cutting displays the active neotectonic Kroupnik fault as a slickenside with thin mylonites. It strikes about 25º (NNE) and dips 65-70ºWNW, and is traced well in the morphology of the terrain, with the faceted slopes of the footwall. Further northeast, the fault strike changes gradually to nearly 50º, and the dip angle, to 40ºNW. The striae on the slickensides point to a normal fault character, with a slight (about 10% of the vertical component), left-lateral strike-slip.

The footwall is built of amphibolites of the Vucha Formation (Rupchos Group, Rhodopian Supergroup). They are intensely fractured and mylonitised, and the older structures (foliation, lineation, folds) are hardly visible. The hanging wall is built up of conglomerates of the Kalimantsi Formation (Pontian–Romanian). The conglomerates consist of well-rounded pebbles of equigranular granites of the Upper Cretaceous North-Pirin (Dautov) pluton. Some pebbles situated near the slickenside of the fault display shear fractures with slight (one to several millimetre) displacements, with the same sense as the master fault.

The Kroupnik fault has been active since Sarmatian time, with a total neotectonic (Upper Neogene and Quaternary) vertical offset of the order of 3500 m, and a mean velocity of 0.25 mm a^{-1}. Most of this offset (more than 3000 m), refers to the Neogene times, with two major epochs of activity (Sarmatian - Maeotian and Pontian - Romanian). The recent activity is gauged with a station that measures the relatively slow recent movements on either side of the fault. The fault was particularly active in the years 1901 – 1910. Fourteen earthquakes with epicenters along the fault trace had a magnitude exceeding 4.5,

Photo 4.1 - Panorama of the Kroupnik fault

and the total number of tremors was more than 1700. The seismic foci migrated gradually from the NNE towards SSW. The strongest earthquake (M estimated at 7.83!) occurred on 4.04.1904 with an epicenter near Kroupnik.

Stop 4.3.:

Road towards Brezhani: 540 m
Panorama of the Brezhani graben

The Brezhani graben is a post-sedimentary structure, formed mostly in earliest Miocene times. It is limited by the West-Brezhani fault to the west (from the Kachovska horst), and the faults and a west-vergent thrust belonging to the West-Pirin fault zone, to the east. The Palaeogene filling of the Brezhani graben has been studied by B. Kamenov et al. in 1956, and S. Cernjavskadetermined the Late Oligocene age by using palynomorphs. A formal lithostratigraphy was introduced by M. Vatsev in 1984 based on all previous published and unpublished research. The section is subdivided into 5 formations, with a total thickness of 1000 - 1200 m. It consists (from bottom to top) of polymictic gray conglomerate, with sandstone and siltstone interbeds (Kachovska Formation); interbedding of bituminous shales, siltstones, sandstones, and some coal seams (Goreshtichka Formation); sandstones with conglomerates and siltstones (Rakitnishka Formation); bituminous shales, sandy shales, siltstones, and coal seams (Loulevska Formation); and the topmost conglomerate and sandstone (Dermirishka Formation). The sediments are correlatable with the other coal-bearing Upper Oligocene – lowermost Miocene (Egerian) formations in western Bulgaria (Krasava, Pernik, Stanyovtsi, Bobovdol) and eastern Serbia, and are relics of a large lake. The sediments of the graben and the border thrust are sealed by Pliocene sediments of the Kalimantsi Formation.

Stop 4.4:

Entrance (road sign) of Brezhani: 546 m

The West-Brezhani normal fault strikes NNE-SSW, and dips at an angle of about 55ºW. It displaces with small amplitude the amphibolites (Vucha Formation) of the Kachovska horst from the basal conglomerates of the graben.

Stop 4.5:

Kresna Gorge, after the first tunnel and bridge: 232 m

Entering the picturesque Kresna Gorge, the itinerary crosses the Upper Cretaceous North-Pirin (Dautov) pluton. The biotite granites to quartz-monzonites exposed are equigranular, medium- to fine-grained, locally with numerous small (from several to 10 - 20 mm) segregation biotite inclusions. Thin aplite veins are also observed, as well as (rarely) pegmatoid nests, with cavities filled with adularia feldspar, quartz, and epidote crystals. Whole-rock samples gathered near the tunnel yielded a Rb-Sr isochron, corresponding to 92 +/- 22 Ma. The initial Sr isotopic ratios of the Upper Cretaceous and Palaeogene igneous rocks of SW Bulgaria, show that magmas within the thickened continental crust of the Rhodope massif (Pirin horst included) had a crustal source (anatexis), whereas magmas in the southern parts of the massif exhibit a mixed character, and the Upper Cretaceous magmas of the Srednogorie (e.g., Vitosha pluton), an upper mantle origin. The pluton is intruded both into the Ograzhdenian metamorphics of the Ograzhden unit, the northern branches of the NE-vergent Mid-Cretaceous Strymon thrust, and the Rhodopian Supergroup and Palaeozoic Kroupnik pluton of the Pirin unit.

Intense faulting and fracturing are related to the Palaeogene and Neogene development of the Strouma Lineament. Typical associations of several fault sets and the fault surface morphology and sculpture, may be observed in many outcrops along the gorge. In the outcrop, intense fracturing defines two sets of faults, striking 115-145º, and dipping 35-40º and 55-75º, with striae plunging c. 70º at angles of 30-40º. A late strike-slip fault of the Strouma fault belt strikes 150º and dips East at about 80º.

Stop 4.6:

Tunnel S of Yavorov RW station: 209 m

This stop enables observation of the rocks of the gneiss-migmatitic complex of the Ograzhdenian Supergroup (from the Ograzhden unit) West of the Strouma fault zone. The serpentinite (serpentinized harzburgite with talc, tremolite and asbestos in the peripheral zone) is preserved as a rootless lenticular body parallel to the foliation in the two-mica and biotite gneisses and amphibolites. Some desillicified pegmatitic veins cross-cut the serpentinite. A dyke of granodioritic porphyrite crops out in the vicinity. The road cutting exhibits also a case of boudinage, when biotite gneisses are boudinaged within the rich in leucosome biotite and two-mica migmatites. Obviously, the relative viscosity of the gneisses was higher than this of the "migma", and they reacted in a semi-brittle manner to the stress during the synmigmatic folding.

Volume n° 2 - from B16 to B33

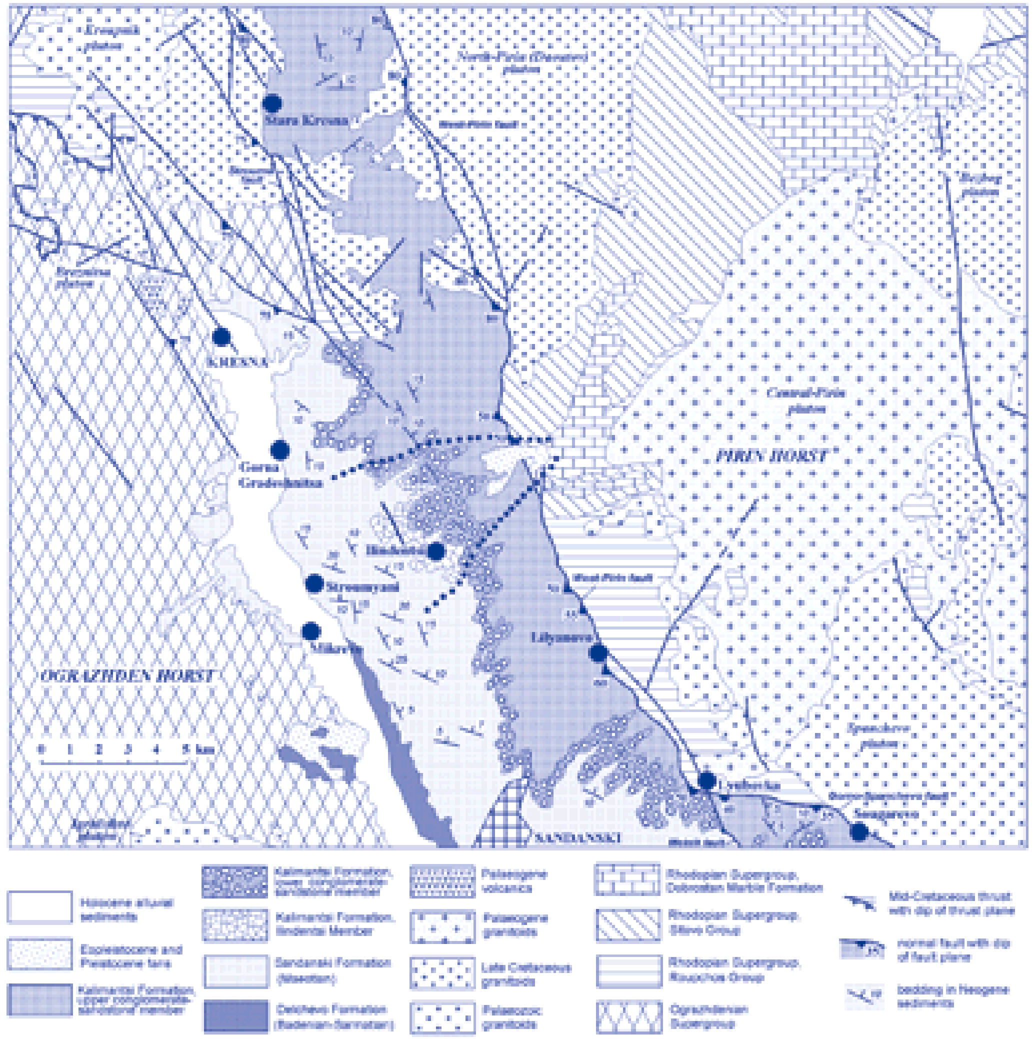


Figure 29 - Geological map of parts of the Sandanski graben and Pirin horst

Ultrabasic rocks are widespread in the Ograzhdenian Supergroup. They occur usually as rootless bodies throughout the section, their emplacement pre-dating the amphibolite-facies regional metamorphism. Their origin is considered to be related to pre-metamorphic Precambrian processes of obduction of the oceanic crust, or to penetration of mantle material in the continental crust, and their further amalgamation.

Stop 4.7:
Mikrevo: 222 m

This observation point is situated on the sinuous road from Mikrevo towards Tsaparevo, and exposes also the rocks of the upper parts of the Ograzhdenian Supergroup (Maleshevska "group") from the eastern margin of the Ograzhden unit. These are biotite and two-mica gneisses and migmatites interlayered with amphibolites and biotite schists. In the locality the biotite schists contain tiny tourmaline crystals. After folding, the schists were intruded by pegmatites (with tourmaline) and aplites, that underwent afterwards a new deformation event (again in amphibolite-facies conditions), together with the host rocks, with intense

schistification (aplites and pegmatites transformed into quartzo-feldspathic gneisses) and folding in two episodes.

The Sandanski graben is an asymmetric structure, with maximum (3000 - 3500 m) vertical displacement along the eastern margin (West-Pirin fault zone), and a small (150 - 500 m) offset along the western margin (Ograzhden fault zone). The stop displays a panoramic view (Figure 29) of the Pirin horst with the highest peak Vihren; the initial (Lower - Middle Miocene) peneplain (orthoplain) at ca. 2600 m, and the planation surfaces (oroplains) at 1950 - 2200 m and 1500 - 1600 m; the tilted section of the Sandanski graben, with the marble breccia of the Ilindentsi Member; and the post-Neogene surfaces (pediments) built over the tilted Neogene section (and partially, over the West-Pirin fault zone itself) at altitudes of 650 - 900, 500 - 600, 320 - 360, and 220 - 270 m.

Stop 4.8:
Road to Ilindentsi: 244 m

Some outstanding features of the neotectonics of the Sandanski graben (for details, Zagorcev, 1992, 1995) are visible along the road from Stroumyani towards Ilindentsi.

The Sandanski graben is filled in with Neogene sedimentary formations with a total thickness exceeding 1600 m. They are clearly divided into three parts.

The basal parts (Delchevo Formation, with Badenian to Sarmatian - Maeotian age, and a thickness exceeding 200 m) are represented by silty and shaly, reddish or greenish sandstones, interbedded with fine-pebble conglomerates, siltstones and shales. Rarely, they contain tuffite (or tephra) material, and single limestone beds. In the upper parts of the Formation, internal wash-outs are clearly visible, with thin (from centimeters to a few meters) rhythmic repetition of conglomerate or coarse sandstone over the wash-out surface of siltstone or clay (locally with coalefied plant debris), and becoming finer-grained, with graded bedding, upwards. These features are observed on the main road near Sandanski (Stop 4.7).

The next formation (Sandanski Formation, of Maeotian age, or more precisely, from uppermost Sarmatian to lowest Pontian; 180 - 1000 m thick); is built up of medium- to coarse-grained sands and sandstones, locally with cross or graded bedding, with silty or clayey interbeds (at the village of Hotovo, with the Hotovo coal seam), and some conglomerate.

At the village of Ilindentsi, the uppermost, Kalimantsi Formation (Pontian - Romanian; 200 - 600 m thick) begins over the washed-out surface of the Sandanski Formation with the 50 - 80 m thick Ilindentsi Member. It is built up of huge fragments and blocks of Dobrostan marbles coming from the nearby outcrops of the Pirin horst, in the area of the peaks Sinanitsa and Sharaliya. The breccia is well-cemented by carbonate cement, and has a sharp and sinuous lower surface.

Photo 4.8 - Pontian coarse breccia (fragments from Precambrian marbles, Pirin horst).

Locally, the lower surface is sharp and planar, and has the character of a slickenside with striae. The upper surface is also uneven, and numerous fractures into the breccia are filled in (neptunian clastic dykes) by sandstones of the next member of the formation. The

Volume n° 2 - from B16 to B33

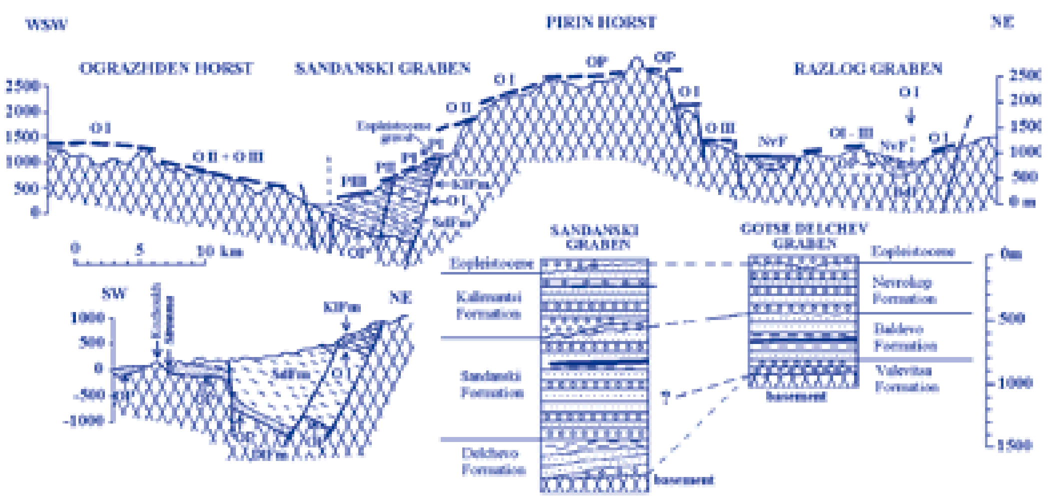


Figure 30 - Schematic neotectonic section through the Pirin horst and the adjacent grabens

breccia is displaced by normal faults that belong to the Strouma fault zone (mostly sealed by the sediments of the graben). Laterally, the Ilindentsi Member passes from the monomictic marble breccia into polymictic (with amphibolite fragments) breccia-conglomerate, and thins out at a distance of a few kilometers. The next two members of the Kalimantsi Formation are built mainly of whitish to yellowish conglomerate (well-rounded granite pebbles coming from the Palaeogene Central-Pirin pluton) and sandstone. They are separated by a thin reddish layer that may represent a lateritic palaeosoil.

The whole Neogene section described was tilted against the West-Pirin fault zone, and dips east at an angle of 5 - 20º. The post-Neogene planation surfaces (pediments), beginning with the Eopleistocene one, truncate this structure, and are inclined west, towards the Strouma valley at an angle of 2 - 5º. Locally, Eopleistocene pebble gravels with the same dip cover the Eopleistocene pediment.

The marble breccia of the Ilindentsi Member (now at an altitude of about 500 m above sea level) is correlated with the destruction of the planation surface (oroplain I) situated now at an altitude of 1950 - 2250 m in the Pirin horst (Figure 30). Therefore, two main epochs of relative uplift of the horst may be distinguished in Neogene times, the second one embracing most of the Pontian, and the Dacian and Romanian. Correspondingly, the vertical amplitudes along the West-Pirin fault zone may be calculated at 1500 - 2000 m for the Badenian to Maeotian, 1300 - 1500 m for the Pontian - Dacian (and Romanian), and 100 - 200 m, for the post-Pliocene times. Another important feature is the first mass occurrence of pebbles from the Palaeogene granitoids in the Kalimantsi Formation, that determined the time of exhumation of these granitoids not earlier than the Pontian.

Stop 4.9:

Pediment over Kalimantsi Fm: 863 m

Crossing the whole section of the Kalimantsi Formation, the itinerary arrives at this Pleistocene pediment, which is situated at an altitude of 850 – 1050 m, and truncates both the tilted section of the Neogene, and the border normal faults of the West-Pirin fault zone. The zone itself is covered by Pleistocene pebble gravels – the Pleistocene part of the Ilindentsi fan. Along the road on the way back, sandstones and conglomerates of the Kalimantsi Formation are visible. Some of the conglomerate layers are richer in marble fragments, whereas in others, abundant granite pebbles and boulders from the Palaeogene Central-Pirin pluton are related to the fast exhumation (unroofing) of the pluton in Pontian and Pliocene times.

Stop 4.10:

Above Ilindentsi: 411 m

Marble breccia over interbedding sandstones and conglomerates, with marble and granite pebbles. The panorama exhibits at least two thick layers of marble breccia of the Ilindentsi Member, partially displaced by faults of the Strouma fault belt (the most important faults of the belt are sealed by the Neogene section!). Many of the big cliffs are in fact due to landslides.

Stop 4.11:

Ilindentsi: 396 m

At this stop, the lowermost parts of the lower breccia layer of the Ilindentsi Member cover the sandstones of the Sandanski Formation. The panorama of the Ograzhden horst, west of the Sandanski graben, exhibits polyfacial and polychronous planation surfaces of Neogene age, the older truncated by the younger, and gradually lowering towards the small-amplitude Ograzhden fault. The latter is marked by a facetted slope.

Stop 4.12:

On the main road after the Ploski junction and before the railway bridge:134 m

The road cutting exhibits the Delchevo Formation (Badenian-Sarmatian), which is covered by the basal parts of the Sandanski Formation (Maeotian). The Delchevo Formation consists of reddish and greenish siltstones and sandstones, often with graded bedding and small calcareous concretions. In this outcrop, a fine cyclicity can be observed, consisting of an occurrence of wash-outs over the finest sediments, covered with sharp depositional contacts by well-sorted whitish small-pebble conglomerate or coarse sandstone, and continuing into greenish and reddish siltstone. This cyclicity is typical of the transition towards the Sandanski Formation.

Stop 4.13:

Melnik: Neogene sediments of the Sandanski graben

The “Melnishki piramidi” (Melnik Pyramids) are erosion forms that are included in the list of natural sites protected by the Bulgarian state. They developed within the loosely cemented rocks of the Kalimantsi Formation (Pontian - Romanian). The whole section is gently dipping (5 - 20º) ENE, towards the West-

Photo 4.14 - Neotectonic fault in Maeotian (Sandanski Formation)

Volume n° 2 - from B16 to B33

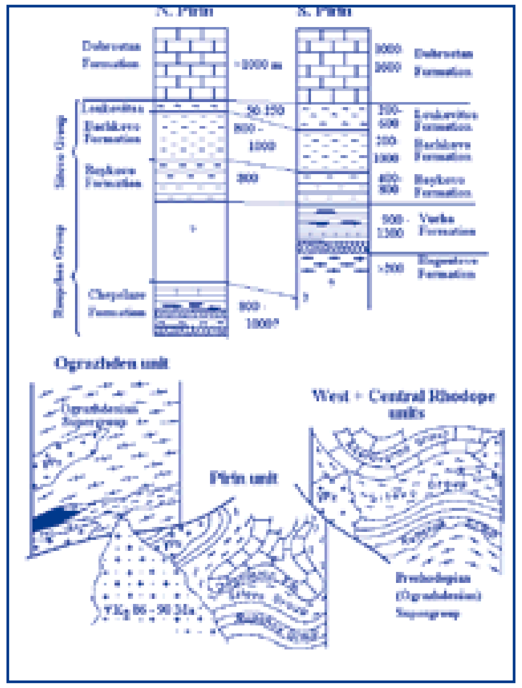


Figure 31 - Schematic columnar sections and interrelations between the Ograzhden, Pirin, and Rhodope units

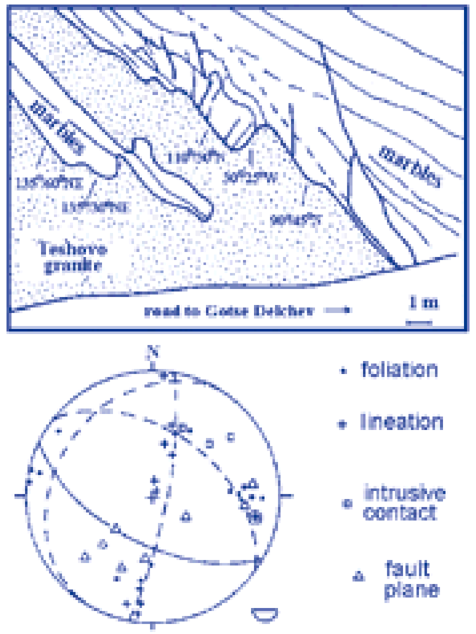


Figure 32 - Contact of the Teshovo pluton with the Dobrostan marbles (after Zagorchev, 1995)

Pirin fault zone.

The Sandanski Formation consists mainly of whitish or yellowish sandstones, interbedded with some clays and coal clays (also a thin coal seam at the village of Hotovo), and fine-pebble conglomerate. It is covered by similar (whitish to yellowish) conglomerates, with sandstone interbeds, that belong to the Kalimantsi Formation. The principal diagnostic feature of the Kalimantsi conglomerates, is the abundance of well-rounded pebbles coming from the Palaeogene granitoids that were unroofed in Pontian time. Graded bedding and cross bedding are often observed.

The rocks of the Sandanski Formation are well exposed at the crossroad to Vinogradi, 2 km West of Melnik. Typical sandstones, interbedded with clay and coal clay crop out, and the gentle dip towards east is clearly visible.

Stop 4.14:

Rozhen Monastery: erosion forms in the Kalimantsi Formation

The Rozhen Monastery was built in the beginning of the 13th century. Ruined several times during the Turkish occupation, it is now one of the most beautiful Bulgarian monasteries. Pyramids, capped pillars, and other picturesque erosion forms are typical for the badlands between the monastery and Melnik. They are observed at their best from a point situated at about a 15-minute walk above the monastery.

DAY 5

Melnik - Popovi livadi - checkpoint Koulata/ Promahon

From Melnik the itinerary crosses the Neogene of the Sandanski graben, and enters, in the Pirin horst, the Precambrian metamorphics of the Rhodopian Supergroup. At the highest point of the pass (Popovi Livadi), the itinerary enters the Palaeogene Teshovo pluton. A full section of the Rhodopian Supergroup (Figure 31) is exposed, and is briefly described hereafter, although the concentrated programme makes thorough observations impossible. This section is in an inverted position (upside down), and from the Popovi livadi pass, the itinerary follows the succession downwards (although upwards in the geometric section).

Stop 5.1:

Contact of the Teshovo pluton with the Dobrostan Marble Formation: altitude 990 m

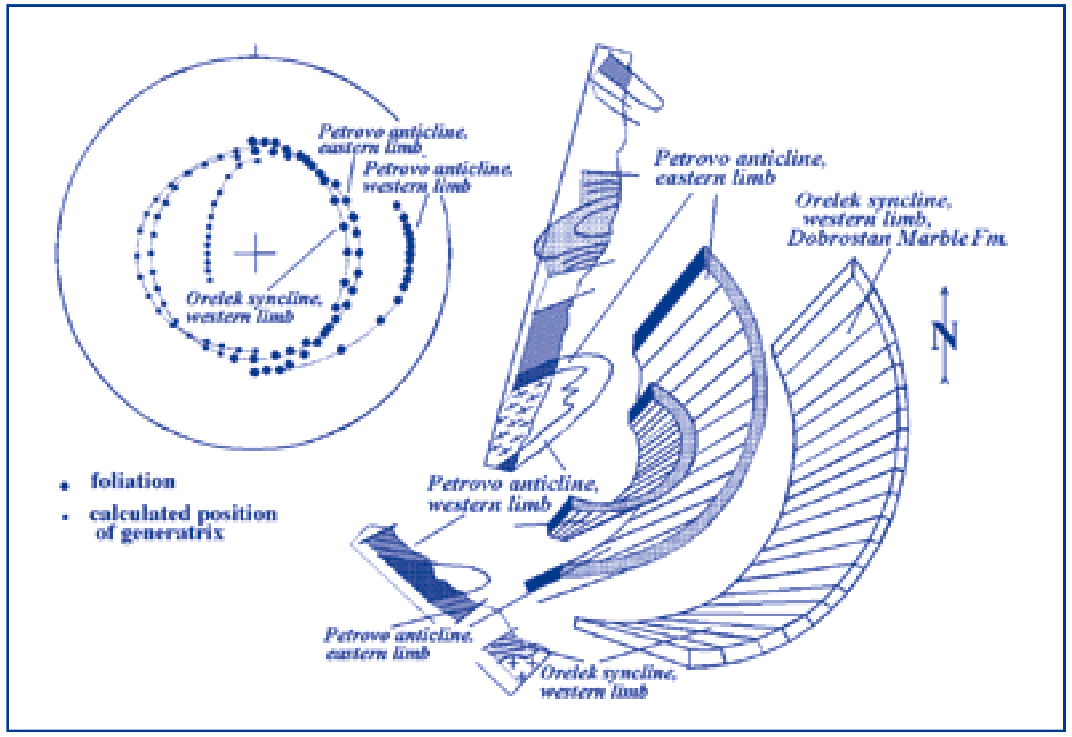
The Dobrostan Marble Formation consists of

Photo 5.1 - Extensional intrusive contact of the Palaeogene Teshovo granite (s. Figure 32)

gray or white, massive, or banded calcite marbles, locally interlayered with gray fine-grained dolomite marbles, hornblende schists, biotite schists, and quartzo-feldspathic gneisses. In the schists and impure marbles, a prominent mineral lineation parallel to synmetamorphic fold hinges is observed. In the road cutting, in very rare cases, the observations reveal that there are two generations of synmetamorphic mineral lineations and fold hinges. The earlier lineation and fold hinges are deformed (folded) and/ or obliterated by the second fold generation. The second generation lineation and hinges are usually situated at a very small angle (5 - 10º) to the first generation, and very rarely, at high (50 - 80º) angles.

The Palaeogene Teshovo granitoid pluton is built up of biotite or hornblende-biotite equigranular (locally with porphyric feldspar) fine- to medium-grained granites to granodiorites. The road cutting displays (Figure 32) complex interrelations between the pluton and the

Figure 33 - Bloc-diagram for the structure of the Belyovo conical fold. After Zagorchev (1981, 1995)



Volume n° 2 - from B16 to B33

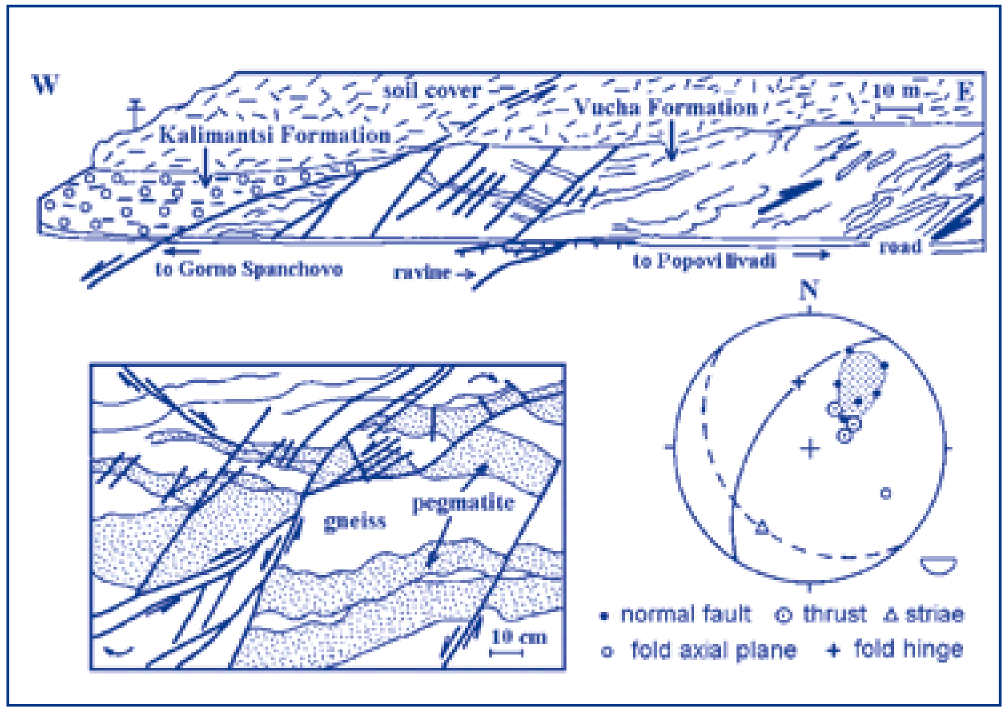


Figure 34 - Road cutting near Gorno Spanchevo. After Zagorchev (1995, Figure 25)

marbles. The intrusive contacts often follow normal faults, and locally suggest a subsidence of the pluton host rocks into the magma chamber. Subparallel apophyses intrude the marbles along the foliation planes. The overall impression is that of extension conditions during the pluton intrusion.

Stop 5.2:

Near the Popovi Livadi pass: 1399 m
Granites of the Teshovo pluton

Stop 5.3:

Boundary between the Dobrostan Marble Formation and the Loukovitsa Formation 1272 m

The contact between the two formations is marked by the disappearance of the gneiss and amphibolite layers. The foliation strikes 170º and dips steeply (70º) west, and has a polydeformational character (composite foliation). The Loukovitsa Formation consists of biotite gneisses and schists, amphibolites, calcareous schists, and marbles. Tight folds with axial planes striking about 175º and dipping 70ºW or E are abundant.

Stop 5.4:

Parking lot at the monument to Yane Sandanski: 1190 m

Banded gneisses with porphyroblastic feldspars of the Boykovo Formation, and quartzo-feldspathic gneisses (also with occasional feldspar porphyroblasts) of the Bachkovo Formation (both of the Sitovo Group, Rhodopian Supergroup) are exposed by the road cutting. They are characterised by a strong S/L structure, with prominent foliation and lineation. The sense of shear corresponds to top-to-southwest.

Stop 5.5:

Vucha Formation: 1079 m

The Vucha Formation consists of biotite gneisses (mostly garnet- and graphite-bearing), amphibolites (both ortho- and para-), schists, calcareous schists, and marbles. They are all present in the outcrops in the road cuttings. The foliation strikes 130-140º and dips at 35º to SW; a lineation plunges 10º to NW (320º). Tight folds are visible in the marble layers. The rocks are intersected by pegmatite dykes. Mylonitization along foliation surfaces and related diaphthoresis are

due to late deformations.

Stop 5.6:

Granitoids of the Spanchevo pluton: 712 m

The Spanchevo granite pluton is intruded into the rocks of the Bogoutevo and Vucha Formation from the core of the Petrovo anticline. The southern part of the anticline has now an eastern to SE vergency. The whole structure, the Spanchevo granites and their planar structures included, has been deformed with the formation of the Belyovo conical fold (Figure 33).

The Spanchevo granitoid pluton is built up of biotite and two-mica coarse-grained or porphyric granites to granodiorites. The heterogeneous internal structure is demonstrated by the frequent occurrence of schlieren and melanocratic enclaves. In the road cutting, the granites exhibit also a clear superimposed schistosity (strike 160º/35ºSW; lineation 200º/22º), that intersects also the numerous pegmatite and aplite dykes within the granitoids. The pegmatites and aplites are often folded. The superimposed schistosity intersects the dykes, and is roughly parallel to the axial planes. Late normal faults strike 115-120º/60-65ºNNE.

The pre-Palaeogene structure of the Pirin unit has a complex and polydeformational character. The following sequence of events has been described (Zagorchev, 1994): (1) Neoproterozoic(?) deposition of a thick (8 – 10 km) sequence of flyschoid island-arc deposits with basic volcanism (Chepelare, Bogoutevo, and Vucha Formation), a mostly psammitic sequence (Boykovo and Bachkovo Formation), another varied sequence of psammites, pelites, marls, and limestones (Loukovitsa Formation) and thick limestones and dolomites (Dobrostan Formation); (2) intense folding in Cadomian(?) times, with the formation of two synmetamorphic fold generations; the first generation F_1 consists of isoclinal folds, and the second one (F_2), of tight to isoclinal folds; a single planar structure is observed, and it is a foliation of a probable multiphase development (composite foliation). The hinges of the isoclinal F_1 folds and the mineral lineation L_1 trend nearly N-S, and plunge at 5 – 20º South. Mostly in the marbles of the Dobrostan Formation of the Orelek syncline, the foliation and the F_1 hinges are deformed by F_2 folds. The principal folds (Petrovo anticline and Orelek syncline) were formed during the second phase. (3) Intrusion of the Spanchevo granitoid pluton (Hercynian?) approximately in the core of the Petrovo anticline; (4) Compression and formation of a superimposed schistosity in the Spanchevo granitoids, roughly parallel to the older composite foliation – this event probably coincided with the Mid-Cretaceous thrusting (Strymon thrust) of the Ograzhden over the Pirin unit; (5) Deformation of the older structures (composite foliation included) into the Belyovo conical fold; (6) Intrusion of the Upper Cretaceous Dautov and Bezbog plutons during an extensional event coinciding with the extension within the Srednogorie zone North of the Morava-Rhodope superunit; (7) Extension and deposition of thick Palaeocene? – Middle Eocene? terrigenous sediments in the southern parts and around the Pirin unit; (8) Folding; (9) New Palaeogene extension, deposition of thick volcanic and sedimentary deposits in depressions around the unit, and Upper Eocene – Lower Oligocene intrusion of granitoids in the

Figure 36 - Stratigraphic column of the well STR-2 (after Lalechos, 1986).

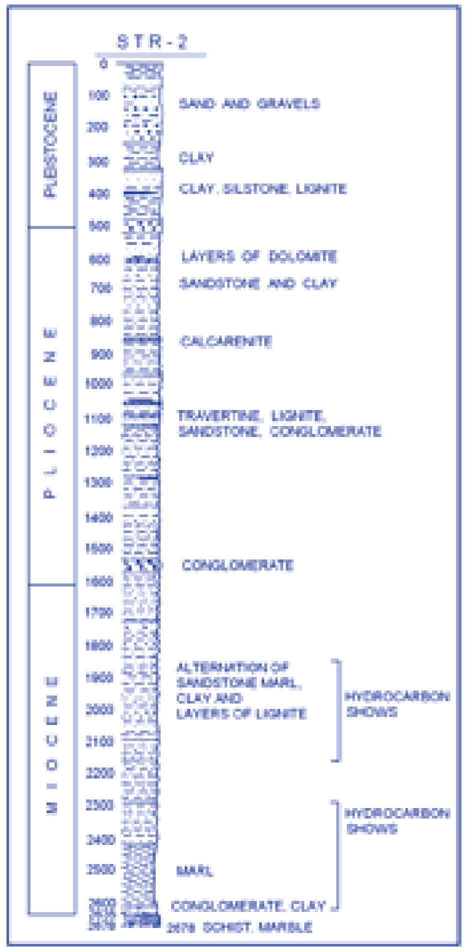


Photo 5.7 - Gorno-Spanchevo low-angle normal fault (s Figure 34)

Pirin unit; (10) Late Oligocene deposition of coal-bearing sediments (Egerian fluviolacustrine system Pernik – Bobovdol – Brezhani); (11) Thrusting of the Pirin horst over the Brezhani graben, related to the transpression along the Strouma fault belt; (12) Extension (after Lower - to Upper Miocene planation) and rifting with the formation of the neotectonic horsts and grabens, and of the new Strouma/Strymon fluviolacustrine system.

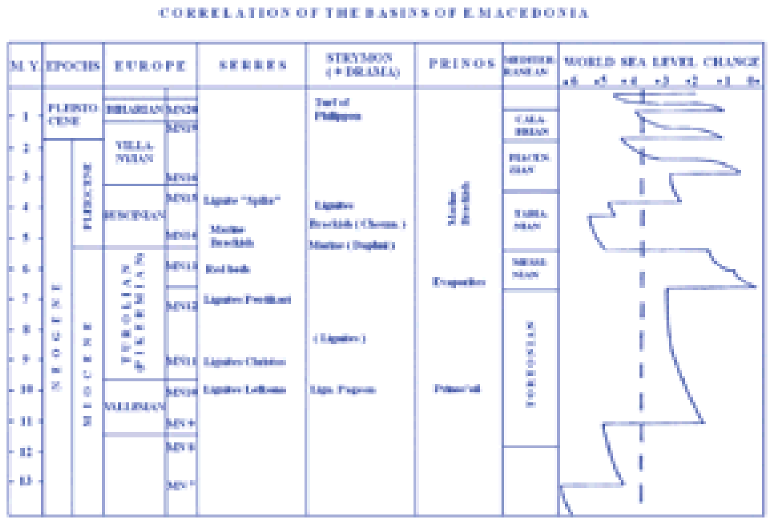
The folded rocks of the Vucha Formation exhibit several fold generations (recumbent folds included), and a last compressive event with mylonites and east-vergent movement. This structure is intersected and displaced by numerous minor normal faults dipping South-West at angles from 25 - 30º (parallel to the master fault) to 60 - 70º. The master (Gorno-Spanchevo) fault is a normal fault with a very narrow gauge along the fault surface.

Stop 5.7:

Gorno Spanchevo neotectonic fault: 599 m

East of the village of Gorno-Spanchevo, the West-Pirin fault zone is represented by a low-angle (25 - 30º) normal fault (Figure 34), that displaces the folded and mylonitized gneisses and amphibolites (with pegmatite veins) of the Vucha Formation (footwall) from the conglomerate of the Kalimantsi Formation (hanging wall). The displacement occurred in latest Pliocene or Eopleistocene time, since the Kalimantsi Formation belongs to the time interval from Pontian (or latest Maeotian) to Romanian.

Figure 37 - Correlation of the Strymon, Drama, Serres and Prinos basins, and the global sea level changes (after Karistineos and Georgiades-Dikeoulia, 1986).



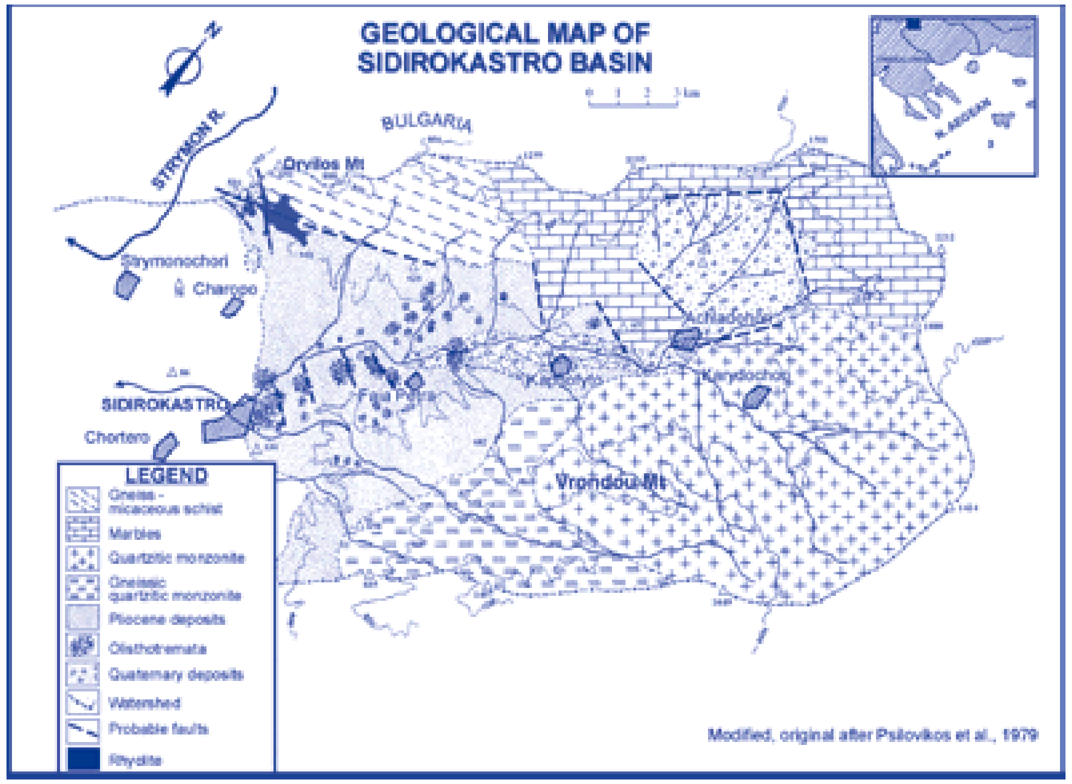


Figure 38 - Geological sketch map of the Sidirokastro basin (modified after Psilovikos et al., 1979; data for the volcanics after Maratos, 1966).

The Kalimantsi Formation is represented by the typical conglomerate, built up predominantly of granite pebbles coming from the Palaeogene Central-Pirin and Teshovo plutons.

Towards the south and southeast, a large Eopleistocene pediment (mountain step), gently-dipping towards the west, is beautifully exposed. It truncates the metamorphic rocks of the Pirin horst, the tilted conglomerate beds of the Kalimantsi Formation, and the bounding Gorno-Spanchovo low-angle fault, and is partly covered by proluvial and alluvial Eopleistocene sediments.

Stop 5.8:

Roupite – Kozhoukh: 103 m

Neogene volcanics of Kozhoukh, and the Neogene Delchevo Formation; Roupite church

The hills of Kozhouh and the Roupite locality are famous as being the site of the home and church of the sybil Vanga, who died a few years ago, and was the most popular prophetess and woman saint in the area. Something magnetic exists in that place: the pathetic hills in the flat and wide valley near the confluence of the Stroumeshnitsa River into the Strouma River; the steep northern slope of Belasitsa to the south; the ghosts of Samouil and his warriors defeated by the Byzantines in 1014, and the ghosts of thousands of dead men of different tribes that lived in the ancient and now destroyed town of Petra, and worshipped the forgotten gods of the Thracians in the nearby sanctuary. The migratory birds fly along the *Via Aristotelis* twice each year, and find a temporary refuge at that place, too.

The hills are built up of volcanic rocks of alkaline trachydacitic composition. They form an oval body about 1800 m long, and 350 - 500 m wide, with a height of 200 m above the plain. The volcanics are strongly hydrothermally altered, and the host rocks of the Ograzhdenian Supergroup and the nearby Neogene (Delchevo and Sandanski Formation) bear also traces of the hydrothermal activity. The hydrothermally-altered rocks contain opal, and are limonitized and kaolinized. The palaeohydrothermal

Volume n° 2 - from B16 to B33

activity is marked also by the irregular masses of hydrothermal carbonates (calcite and aragonite) that have been prospected as ornamental stone, due to their beautiful layered structure. They are situated at the northern side of the body, and dip steeply south. The thickness of these masses reaches up to 150 m.
The thermal mineral waters possess a sodium hydrocarbonate character, and a temperature reaching up to 67∞°C at the surface. Several other natural springs are known at the nearby villages of Levounovo (temperature from 44 to 82∞°C), and Marikostinovo (with more than 50 springs with temperatures between 42 and 61.5∞°C). The springs have been prospected with boreholes, and are only partially used for cures and greenhouses.
Kozhouh has a probable latest Miocene or Early Pliocene age. As far as no other volcanics of this age have been recorded in Southwest Bulgaria, it has an exceptional position at the crossing of the Strouma and Stroumeshnitsa fault zones. Similar volcanics of Neogene and Quaternary age, and even more alkaline composition, are known at the southern side of the Belasitsa (Kerkini) Mountain, on Greek and Macedonian territories. The age problem is still under discussion, as far as no direct contacts with Neogene rocks are observed, and the “tuffites”, known from the Neogene sedimentary succession, are determined by other scientists as hydrothermally-altered sediments (due to the mineral waters), or else their source is being sought in volcanoes outside Bulgarian territory.
At the checkpoint Koulata/Promahon the itinerary leaves Bulgarian territory and enters Greece.

Itinerary in Northern Greece (Macedonia and Epirus)

I. Mariolakos, I. Fountoulis

DAY 5

Checkpoint Koulata/Promahon - Verria

Strymon Basin

The Basin of Strymon lies between the Bulgarian/Greek border to the NNW, and the Strymon Gulf, to the SSE. The basin covers an area of 2700 km^2 of eastern Macedonia.

The onset of the Strymon basin was in Middle Miocene times. The basin is a typical postorogenic graben, trending NW-SE (Lalechos, 1986), which is still active. The basin has been formed between the Serbo-Macedonian massif to the SW, and the Rhodope massif to the NE. Based on drilling data (Figure 12), the total thickness of the post-Alpine sediments (Table 3) is estimated at about 3500 meters. The discrimination of facies is based on palynological analyses of the cores. Miocene sediments originated from N-NW, through delta type deposition. Later on, lacustrine sedimentation takes effect, periodically interrupted by sea ingressions or continental depositions.

The travertine – lignite zones show the deposition of sediments in coastal areas near the carbon-rich basement, under favorable climatic conditions.

Table 3. Stratigraphy of the Strymon basin.

Pleistocene formations:	**Clays, coarse grain sands, quartz gravels, sandstones, siltstones**
Pliocene formations:	**Alternations of clastic clays**
Continental facies:	Alternations of red clays, red sandstones, and dolomite sands. Intercalation of siltstones, sandstones, and micro-conglomerates. Travertine limestone with lignites and chalk limestone
Brackish facies:	Sandstones, siltstones, travertines and lignite layers
Miocene formations:	**Alternations of lacustrine and marine facies of sediments from bottom to top**
Marine facies:	Alternations of clastic clays, siltstone, sandstone, with intercalation of microbreccia and lignite layers
Lacustrine facies:	Alternations of sandstones, siltstone, dark brown marl, which eventually change to petroliferous limestone
	Breccia with gravels from fine grain sandstone or hard conglomerates

Many faults cross the basin sediments. Faults parallel to the initial marginal fault zones of the basin, and younger than the Alpine deformation, have created a gradual antithetic tectonic graben within the basin sediments, at the center of the basin. Within this graben, low angle normal faults are observed, creating strata bending and anticline structures. This case is similar to the marine area of the Thrace Sea, but is limited to a smaller scale.
The geothermal gradient for the Strymon basin has been estimated to fluctuate from 25 to 36^{o}C/km, to depths of over 2000 m. These values are above the normal geothermal gradient.
In the Strymon basin, several geothermal manifestations occur (at the thermal springs of Therma Nigrita, Loutra Sidirokastro, Loutra Angistro), and geothermal fields (the known fields of Therma-Nigrita, Sidirokastro, Irakleia-Lithotopos, etc.).
At the eastern margin of the basin, two minor basins, trending E-W (Sidirokastro basin and Serres basin), are located (Figure 37).

Volume n° 2 - from B16 to B33

Stop 5.9:

Strymon Straits

It is the narrowest site between the Kerkini (to the west) and Orvilos (to the east) mountains, where the Strymon River passes through. The contact between the Kerkini gneisses and amphibolites (equivalent to the Ograzhdenian Supergroup, Serbo-Macedonian massif), and marbles and schists of the Pirin-Pangaion unit (Rhodope massif) is here a steep fault. Thick travertines are locally observed.

Stop 5.10:

Sidirokastro - Kirikos and Ioulitis Monastery

Sidirokastro Basin

The basin of Sidirokastro is a part of the greater Strymon basin, and lies between the mountains of Orvilos (to the north) and Vrondou (to the south), with an E-W trend, which means that it is transversal with respect to the Strymon basin that trends NNW-SSE (Figure 38).

The Quaternary deposits of the Sidirokastro basin consist of olisthothremata (huge blocks) and breccias, originating exclusively from marbles of the Pirin-Pangaion unit. The breccias are well cemented with carbonate cement, and overlay uncomformably the Pliocene deposits. They have a very impressive relief. The main outcrops are situated along the valley of the Achladitis River. Similar rocks occur in Bulgaria, but they have been dated as Neogene (Pontian – Romanian) (at the base of the Kalimantsi formation).
Some travertine deposits also occur in the western part of the basin, under the alluvial Strymon deposits, with a thickness that varies between a few and 40 meters (Maratos, 1966).
Volcanic rocks outcrop in the northern part of the basin (towards Mt. Orvilos), at the margin between the metamorphic formations and the post-Alpine Pliocene deposits. Rhyolites cross the Pliocene deposits, and appear on top of them (Maratos, 1966). Within these deposits, no rhyolithic pebbles occur. This fact, combined with the intense topography of the volcanic cone, the presence of hydrothermally-altered metamorphosed blocks of the Alpine basement, the volcanic bombs originating from the sedimentary formations, and the obvious thermal alteration of the surrounding deposits, points to the conclusion that the volcanic activity took place during Quaternary times (Maratos, 1966). The Neogene formations consist of conglomerates, with pebbles originating from the pre-Alpine basement (granites, gneisses), followed by lacustrine and fluvial deposits which consist of alternations of marls, sandstones, micro-conglomerates with lignite intercalations. Their thickness reaches 300 m. (Kavouridis and Karydakis, 1989).
The paleogeographic evolution of this basin follows the general characteristics of the Strymon basin evolution during the Neogene, with the important difference that there is no marine influence (Sotiriadis, 1966). This means that the deposition of fluvial-lacustrine sediments in the lowermost parts of the basin must not have been interrupted from the Early - Middle Miocene to the end of Pliocene. The exact time of the end of the Neogene deposition is not known, but it is very likely that at the Pliocene – Pleistocene boundary, a phase of intense tectonic movements was activated.

The Kirikos and Ioulitis Monastery

In this area, coarse breccias occur, with fragments and olisthothremata (hill-size blocks) originating exclusively from marble. These breccias overlay unconformably a packet of generally fine-grained sediments (marl, sandstone, and occasional intercalations of conglomerate, whose pebbles originate exclusively from the Vrondou granite). The age and facies of the underlying formations is not known. However, in the eastern-northeastern adjacent areas, the olisthothremata and breccias are reported to be of Quaternary age.
A strike–slip fault occurs in both the underlying formation, as well as in the marble olisthothremata. The fault strikes N246º, dip 75º, with striation 20º to342º). More striations (284º/60º, 174º/30 º) of this fault surface were also observed in different outcrops, and so their relative dating is not possible. Near the entrance of the Monastery, another fault (50º/60º), also with striations (92º/60º, 110º/35º), can be observed. As in the previous locations, marl and sandstone, with polymictic conglomerates (pebbles exclusively from the Vrondou granitoids), underlie the monomictic breccias.

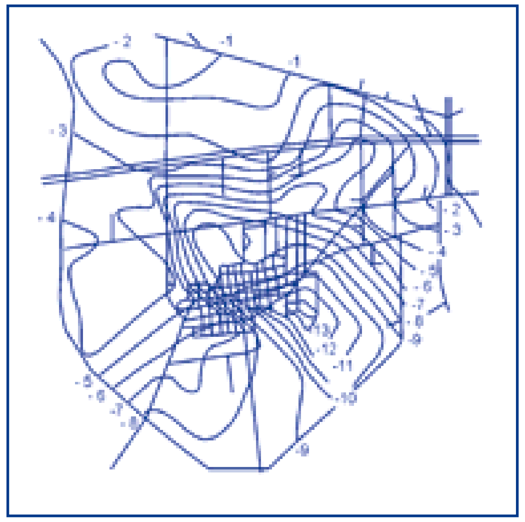


Figure 39 - The rate of subsidence in the Kalohori area, as measured by the Laboratory of Geodesy of Aristotle University of Thessaloniki for the period 1996-1998.

Stop 5.11:

Village of Vyronia: Kerkini boundary fault

Aggistro western margin

A major fault zone separates Kerkini Mt. and the reddish-brown clay and conglomerate of the plains.

The marginal fault zone of Kerkini, trending E-W, outcrops at the Aetovouni (abandoned village) location (past Nea Petritsi village, the entrance of Vyronia village). At the local amphibolite abandoned quarry, ankerites outcrop along the fault zone (amphibolite schistosity dipping 40º to 132º, ankerite dip 45º to150º, fault surface dipping 70º to150º).
Impressive talus cones of varying size occur along the front of the Kerkini fault zone. The larger of these cones occur at the foothill of the highest areas of the Kerkini Mountain.
Pliocene deposits dip towards the NW in the area between Sidirokastro and the Kerkini marginal fault zone. The watershed between the Kerkini and Doirani lake basins develops over lacustrine sandy deposits, while no fans occur in the western part of the basin margin. At this location, the Quaternary asymmetry of the Strymon River is very obvious.

Kerkini lake through time

After many geological and climatic changes, at the end of the last European glacial episode, that marked the beginning of Holocene times about 20,000 years ago, a vast lake, fed by the melting glaciers that had covered the higher elevations of the Rhodope Mountains, occupied the lower parts of the Serres/Strymon basin. During Classical and Hellenistic times, the huge interglacial lake had been reduced to two smaller, separate ones. Historians of that era first mention the name Kerkini, calling the large lake to the south "Kerkinitis" (although this name was known in mythology, Bartzoudis, 1993), and the much smaller one in the northwestern corner of the basin, "Prassias". Several Thracian tribes inhabited the wild lands of the Serres/Strymon basin, until they were subjugated or displaced by the expanding Macedonians in the 5th century BC. Among them, the Visalds lived in the valleys of Mount Dysoron, the Edones to the east of Strymon, and several clans of the Peones in the interior, around Siris (latterday Serres), Lake Prassias and the upper Strymon basin (Petrou, 1995). Since then, many historians and writers have made references to this area and to its rich flora and fauna. The lakes and marshes of Strymon were fabled for their wildfowl. A coin of the Visaltes, depicting a crane, has been found, and the wild goose was a common emblem of Edonian coinage. Fish was so plentiful that the locals "...gave

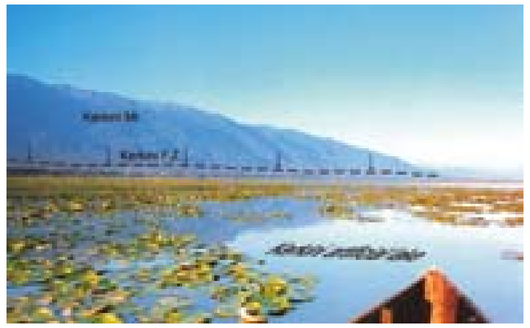


Photo 5.11 - Panoramic view of the E-W striking Kerkini fault zone. View from SW.

them for fodder to their horses and beasts of burden". Herodotus mentions two kinds of fish in Lake Prassias "paprakes" and "tilones", both so abundant that "... if a man lets an empty basket down by a line into the lake, it is no long time before he draws it up full of fish...".
For long periods of time, the people lived in harmony with the wetlands. The lake and the river provided fish and wildfowl, various raw materials, and water for crop irrigation. Human intervention was always small-scale, and the lack of technical knowledge meant that their influence on the landscape and habitats was negligible. Continuing active deposition

Volume n° 2 - from B16 to B33

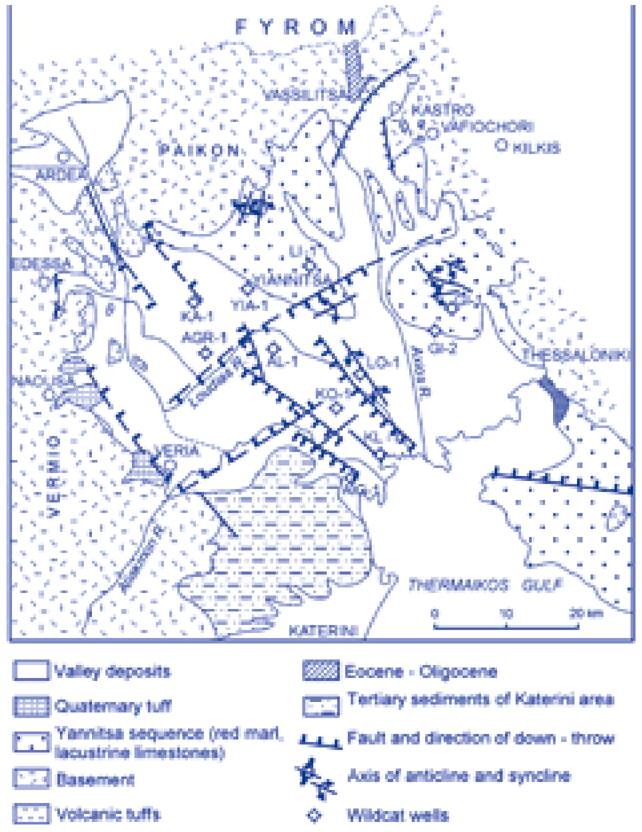


Figure 40 - Geological sketch map of the Axios basin (after Lalechos, 1986).

by the Strymon River further reduced the size of the lakes, creating large expanses of marshes and occasionally flooded areas. The northern lake was quite small, occasionally drying out completely. During the years of Ottoman occupation, it became known as "Lake Butkovo" and was later renamed "Kerkini"; the southern lake was named "Achinos".
By the turn of the 20th century, a complex unstable system of small lakes and extensive freshwater marshes covered most of the lower parts of the Serres basin. The River Strymon entered the basin from the north, through the narrow ravine of Klidi (Ruppel). Upon exiting the defile, it formed a large alluvial fan, covering an area of 180 km^2. There it spread out, with several channels meandering towards the southeast, supplying water to lake Kerkini. Downstream from Kerkini, the river turned to the east, forming a permanent bed, until finally, it emptied into the shallow lake of Achinos.

Both lakes, but mainly the larger Achinos, trapped the floodwaters and almost all of the suspended materials carried by the river. This explains the inability of the Strymon to form a delta at its mouth in the bay of Orphanos, in marked contrast to all the other rivers in eastern Greece.
The river served as a virile, undisciplined, sometimes violent, artery of life to the ecosystem, causing constant habitat modification in accordance to changes in its water level. During flood periods, the lakes increased in size, and the marshes often extended to the foot of the hills. Surrounding meadows and forests were flooded, and the water found or cut new channels, and aquatic vegetation spread and grew in profusion. During periods of drought, the lakes diminished in size, water flow was restricted to a few beds and the marshes dried out, leaving behind fertile soils.
The boundary between the Circum-Rhodope belt (situated SW of the Serbo-Macedonian and the Rhodope massif) and the Axios zone, is situated in the eastern parts of Thessaloniki, and will not be visited due to limited time and poor exposure. The Alpine Axios (Vardar) zone grossly coincides with a Neogene – Quaternary fault belt and the Thermaikos gulf.

Stop 5.12:

Kalohori: the problem of Kalohori subsidence

Kalohori is a town within the industrial area west of Thessaloniki. It is situated in the delta of Gallikos River, close to the sea. In ancient times the area was part of the sea, but the rivers accumulated enormous quantities of soil, so that they formed this part of land. According to the ex-mayor of Kalohori, "after the second world war, the local Thessaloniki authorities started to pump out water from the area for the needs of the population of the city". This pumping was more intensive year after year, because the population and the need for water were constantly increasing. The first signs of subsidence appeared in 1955, and the municipal authorities protested and tried to slow down the water pumping. The Water Company of Thessaloniki, which was under the authority of the Ministry of Macedonia-Thrace, continued the

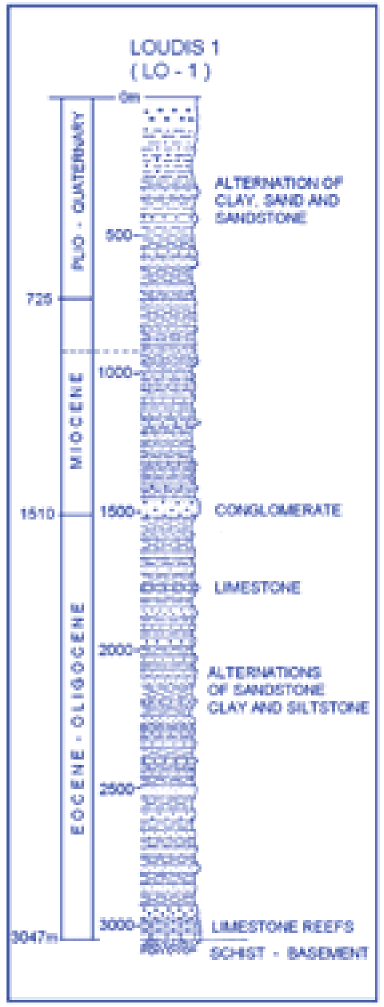


Figure 41 - Stratigraphic column of the well Loudias 1 in the Axios basin. Well location on Figure 40 (after Lalechos, 1986).

drillings in the area, resulting in further subsidence. The ground water level reached a depth of 40-45 meters from the 20 meters that it was at the beginning. Two textile factories in the area contributed to this evolution, each of them pumping 8,000 m^3 per day. As a result of this subsidence, there were several floods from the sea. In 1969 the sea reached the center of the town. By that time the department responsible for these matters from the Ministry of Public Works built a small dam to protect the town. However water pumping was still going on, resulting in further subsidence, and a bigger dam was built in 1976-77. Due to the subsidence and this construction work, the lagoon of Kalohori was formed. The subsidence went on, despite the protests of the people, their appeals to the Court, and the political interventions of local representatives. Finally, in 1995, the Water Company of Thessaloniki decided to quit its activity under the engagement of the local authorities, to allow for supplementary pumping only during July and September, and whenever it was necessary. The water company also took the responsibility to carry out research every year, to record the sinking which is still going on today at a rate of 3-4 cm every year.

Research on the ground settlements in the area of Kalohori west of Thessaloniki

Ground settlements have been observed for the last 35 years in Kalohori, a village and industrial area west of Thessaloniki. These settlements have been continuously monitored by the Laboratory of Geodesy (AUTH) since 1992, with the help of a leveling network. As it was found, the settlements have a 5 cm/year rate in the south-east part of the area (Figure 39), and can be attributed to the intense water pumping.

DAY 6

Verria – Kozani – Grevena

Verria is situated close to the western boundary of the Axios with the Pelagonian zone. Along the itinerary, ophiolites and Upper Cretaceous flysch sediments are exhibited in the road cuttings, as well as some low-grade metamorphic rocks (mostly of carbonate composition) of Triassic and Jurassic age, that belong to the Subpelagonian mantle. These rocks continue west of Kozani, where they are in tectonic contact with the Vourinos ophiolite. Towards Grevena, the itinerary enters the Palaeogene and Neogene sediments of the Meso-Hellenic trough.

Axios basin and Thermaikos Gulf

The Axios basin is extended onto the onshore to the Thessaloniki-Katerini plain, to FYROM in the North, and is connected to the Thermaikos Gulf in the South. It is bounded by the mountains of Pieria-Vermio in the west, and in the east by Chortiatis. East of Thessaloniki, it is covered with molassic Tertiary sediments. Older formations occur in the borders of the basin (Figure 40).

The Tertiary sediments

are unconformably deposited over the metamorphic basement. In Early Eocene times, a basin was formed by faults, and filled in with sediments up to the present time (Figure 41). The basin has not been continuously active, but during Middle Miocene and Oligocene times is indicated by the change from marine to lacustrine facies. The Miocene formations are lacustrine and brackish, and are transgressively situated over the older formations. The Pontian-Pliocene deposits cover unconformably lacustrine facies. It is emphasized that the thickness distribution of the sediments lacks symmetry along the E-W and N-S axes of the basin.

Table 4. Stratigraphy of the Axios basin

Plio-Quaternary	**Sandstone – marls, coarse sandstone, and gravels on the upper section.**
Miocene formations:	**Lacustrine and continental marls and sandstones of variable thickness**
M.–Upper Miocene	Sandstones, blue marls, yellow sand, red clay.
Lower Miocene:	Conglomerate and intercalation of siltstone, followed by thin layers of lignite (age: L. Miocene, palynology analysis).
Unconformity	
Eocene-Oligocene formations:	**Marine sediments, alternations of marine and continental facies**
U. Eocene – L. Oligocene:	Marls, sandstones, hard clays.
Upper Lutetian:	Marine and continental facies, marls and conglomerates (Cassandra area, thickness approx. 400 m.)
Middle Eocene:	Limestone reefs, horizon of lacustrine marls, horizon of marine facies, sandstone and conglomerates. Thickness: 50-100 m. (Vassilitsa, Kastro areas)

Figure 42 - Extension of the Axios and Strymon basins towards the Aegean Sea since Miocene times. Contours show Miocene isopachs (after Lalechos, 1986).

The modern structure of the basin is controlled by the combination of the previous tectonic deformation stages, the resulting continuous sedimentation or unconformities, or changes in the facies of the sediments. Middle – Upper Eocene marine sediments were deposited on the paleorelief of the subsiding basement. From the existing onshore and offshore wells, an increase of the thickness of the Miocene formation (422 m) can be observed, indicating that the basin in the southern part has subsided in depth. On the contrary, the thickness of the Eocene-Oligocene formations is smaller in the Thermaikos, and larger in the Thessaloniki basin. This indicates that the center of the sedimentary basin was on land, with then a rise of the south section, which then dropped during Miocene times, resulting in the considerable thickness of Miocene formations.

Stop 6.1:

Pella

Shoreline displacements often occur without the contribution of a seismic event. These cases are mainly present in large river deltas, being thus connected to the suspended material transported by rivers and deposited at their estuary. Nevertheless, even in these cases, shoreline displacements are connected to continuous tectonic uplift. There are many examples in Greece, such as the broader area of Pella. In Figure 43, the stages of evolution of the Thermaikos gulf and

Volume n° 2 - from B16 to B33

Salonica plain are shown. It is evident that, in this case too, an important role has been played by the large rivers such as Aliakmon, Loudias, Axios, and Gallikos, depositing their sediments in the northern and western parts of the Thermaikos Gulf. At the end of the 5th, and beginning of the 4th century B.C., Pella became the capital of the Macedonian kingdom. Excavations have revealed parts of the earlier city, including the cemetery and scanty architectural remains in the area of the modern drainage canal. The city was organized and expanded during the reign of Philip II and Cassander, and flourished in the middle of the 4th and during the 3rd and 2nd centuries B.C. It was captured by the Romans in 168/167 B.C., and was finally destroyed by an earthquake, possibly in the first decade of the 1st century B.C. The first excavations (between 1957 and 1963-64) brought to light the houses with the mosaic floors and part of the Palace. A second campaign was undertaken in 1976, and is still in progress. So far it has revealed the Agora, part of the Palace,

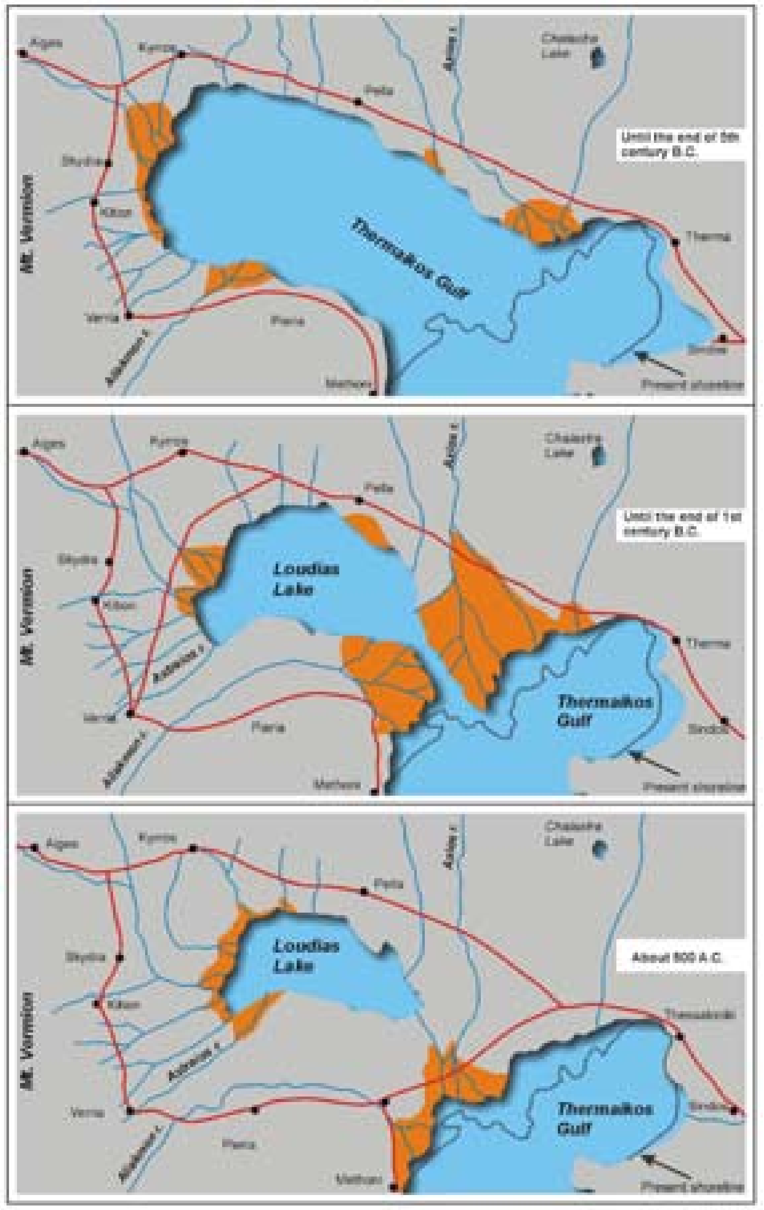


Figure 43 - Stages of evolution of the Thermaikos Gulf since 5th Century BC (modified after Struck, 1908, 1912).

Photo 6.1 - Mosaic depicting a hunting scene from the archeological site of Pella.

Volume n° 2 - from B16 to B33

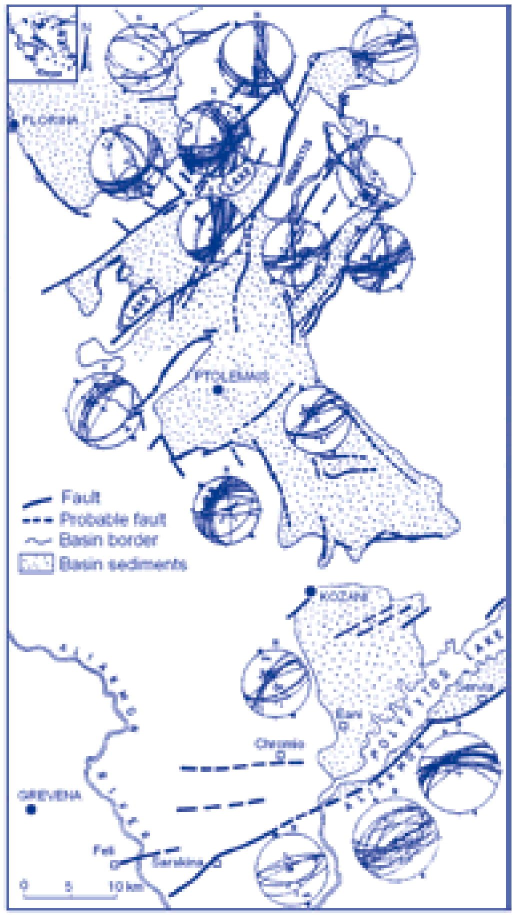


Figure 44 - Map of the Ptolemais - Servia - Florina Neogene-Quaternary basin showing the main faults. Equal-area projection of measured striated faults (after Mountrakis et al., 1998).

other houses, sanctuaries, and cemeteries. The most important monuments on the site are: (i) the private houses; (ii) house with mosaic floors of the Abduction of Helen, the Stag Hunt and Amazonomachy; (iii) the Palace; (iv) the Agora; and (v) the Sanctuaries, which consist of Thesmophorion, Sanctuary of Aphrodite, the Mother of the Gods, and Sanctuary of Darron, the curing god (identified by an inscription).

Stop 6.2:
Vergina

The ancient city situated on the northern slopes of the Pierian Mountains is securely identified as Aigai, the capital of the kingdom of Lower Macedonia. Archaeological evidence proves that the site was continuously inhabited from the Early Bronze Age (3rd millennium BC), while in the Early Iron Age (11th-8th centuries BC), it became an important center, rich and densely inhabited. The city reached its highest point of prosperity in the Archaic (7th-6th centuries BC), and Classical periods (5th-4th centuries), when it was the most important urban center of the area, the seat of the Macedonian kings, and the place where all the traditional sanctuaries were established. Moreover, it was already famous in antiquity for the wealth of its royal tombs, which were gathered in its extensive necropolis. The finds from the excavations are exhibited in the protective shelter over the royal tombs of Vergina, and in the Archaeological Museum of Thessaloniki.

The first excavations on the site were carried out in the 19th century by the French archaeologist L. Heuzey, and were resumed in the 1930s, after the liberation of Macedonia, by K. Rhomaios. After the Second World War, in the 1950s and 1960s, M. Andronicos, who investigated the cemetery of the tumuli, directed the excavations. At the same time, the Palace was excavated by the University of Thessaloniki, and a part of the necropolis, by the Archaeological Service of the Ministry of Culture.

In 1977, M. Andronicos brought to light the royal tombs in the Great Tumulus of Vergina (Megale Toumba). The most remarkable of these was the tomb of Philip II (359-336 B.C.), and its discovery is considered to be one of the most important archaeological events of the century.

The most important monuments on the site are the following: (i) the royal tombs in the Great Tumulus; (ii) the royal tombs to the NW of the city; (iii) the cemetery of the tumuli; (iv) the Palace and the Theatre; (v) the temple of Eukleia; (vi) the acropolis and the city walls.

A major marginal fault zone, trending NNW-SSE, occurs along the southern bank of the Aliakmon River and exits at Mt. Vermion (bridge).

Stop 6.3:
Mavropigi (Ptolemais)

Ptolemais - Servia basin

The Kozani – Servia basin is part of the larger Florina – Vegoritis - Kozani Neogene basin, which extends northwards to FYROM in the Bitola plain, and presents a total length, from Servia to Bitola, of about 100 km. Its average width is about 15 km. It trends in a NW-SE direction, and is divided by ridges and hills trending NE-SW (Figure 44). The basin was filled in with sediments from the Late Miocene to the

Photo 6.3 - View from NE of the northern part of the Mavrodendri - Pontokomi - Mavropigi

Quaternary. The Neogene sediments are exclusively of lacustrine origin and contain lignites. The Quaternary ones are usually fluviolacustrine and terrestrial. It is worth mentioning that there is no evidence of marine influence in the Neogene and Quaternary deposits of the basin.

The marginal fault zones trend NW-SE. Within the Plio-Quaternary deposits, a lot of NE-SW normal faults occur. Apart from these normal faults, there are some reverse faults striking NE-SW (Pavlides, 1985). To the south, the basin is bounded by the impressive Servia fault zone, which is an active one, as it affects Holocene deposits (Mountrakis et al., 1998). This kind of deformation is responsible for the Basin-and-Range type of morphology that has developed across these faults, and appears to be still active.

Mavrodendri - Pontokomi - Mavropigi fault zone

The Mavrodendri - Pontokomi - Mavropigi fault zone is the west marginal fault zone of the Ptolemais – Servia basin. It consists of almost parallel NW-SE striking faults, which are dipping 60°- 80° towards ENE (Figure 45). This fault zone bounds the Triassic-Jurassic carbonates of the Pelagonian unit with the Plio-Pleistocene deposits of the Ptolemais – Servia basin. The normal faults, occurring within the basement of the Pelagonian unit at Mt. Askion, present the same geometry (NW-SE) with the faults that constitute the fault zone.

The Mavrodendri - Pontokomi - Mavropigi fault zone is mostly covered by the young Pleistocene and Holocene deposits (scree), and only in some locations, occasional fault surfaces (slickensides) can be observed. The faults can be distinguished into two sets, according to the striations observed on their surfaces. The first set (A), presents a pitch 50-80° south, whereas the pitch for the second set (B) is 50-80° north. The second one is observed on almost all fault surfaces (Mountrakis, 1996). Secondary antithetic faults exhibit analogous kinematics to the main ones. The length of the fault zone is estimated to be more than 12 km. Mountrakis et al. (1996), characterized this fault zone as inactive because they assumed that there is no compatibility with the current stress field trending NNW-SSE, which is recognizable in W. Macedonia.

It is worth to mention that during the May 13, 1995 earthquake, the Mavrodendri village suffered damages, although the epicentral area was located far from the village.

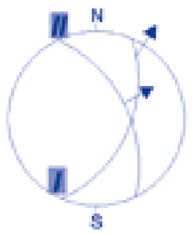


Figure 45 - Lower hemisphere projection of the average fault surfaces occurring at Mavropigi village of the Mavrodendri - Pontokomi - Mavropigi fault zone.

Stop 6.4:

Aliakmon terraces

Below the structural surface of the Plio-Quaternary deposits, which are located at a mean altitude of 624-630 m, the valleys that have been created due to the incision of the Aliakmon River and its tributaries, show usually four terraces (Brunn, 1956). A typical example is the area of Asprokampos (NE of Grevena). Terraces are found at 8, 35, 55, and 100 m above the present-day river bed, the high surface being at 150 m.

Neotectonic movements determine the drainage network of the Aliakmon River. From its source, it probably flowed northwards into Albania, but was barred by the uplift of the transverse Koritsa fault. To the south, it was barred from flowing south by the uplift of the Vounassa, Khassia Krastovon transverse accident, and was forced eastwards through the marginal Pelagonian range, where it has incised deep gorges (e.g. Zavorda gorge).

Stops 6.5 and 6.6:

Pramoritsas River

Marine Pliocene sediments

The stratigraphic structure of the Meso-Hellenic trough, presented by Brunn (1956), is the hitherto most complete analysis of the mollassic deposits. Brunn (1956) accepts that marine sediments filled the NW-SE trending Meso-Hellenic trough, from Upper Eocene (Krania Formation) till the Middle-Upper Miocene boundary (Ontria Formation). More specifically, for this particular location, Brunn suggests that: a) the youngest marine sediments date back to the Upper Miocene (Tortonian); b) the transition from the Miocene to the Pliocene consists of fluvial deposits, something observed only in Albania; c) north from Grevena town towards Neapoli, the Pliocene and Lower Pleistocene deposits are sandy and contain numerous mammal fossils of Villafranchian age; d) the Upper Quaternary deposits are lacustrine, containing; west of Servia village; numerous *Planorbis*, *Lymnaea,* etc.

Eltgen (1986) doubted the fluvio-lacustrine origin of the Pliocene deposits in the area south of Neapoli, proposing that they are marine, while sometimes there is evidence of brackish facies. His views were based on petrographic observations and paleontological

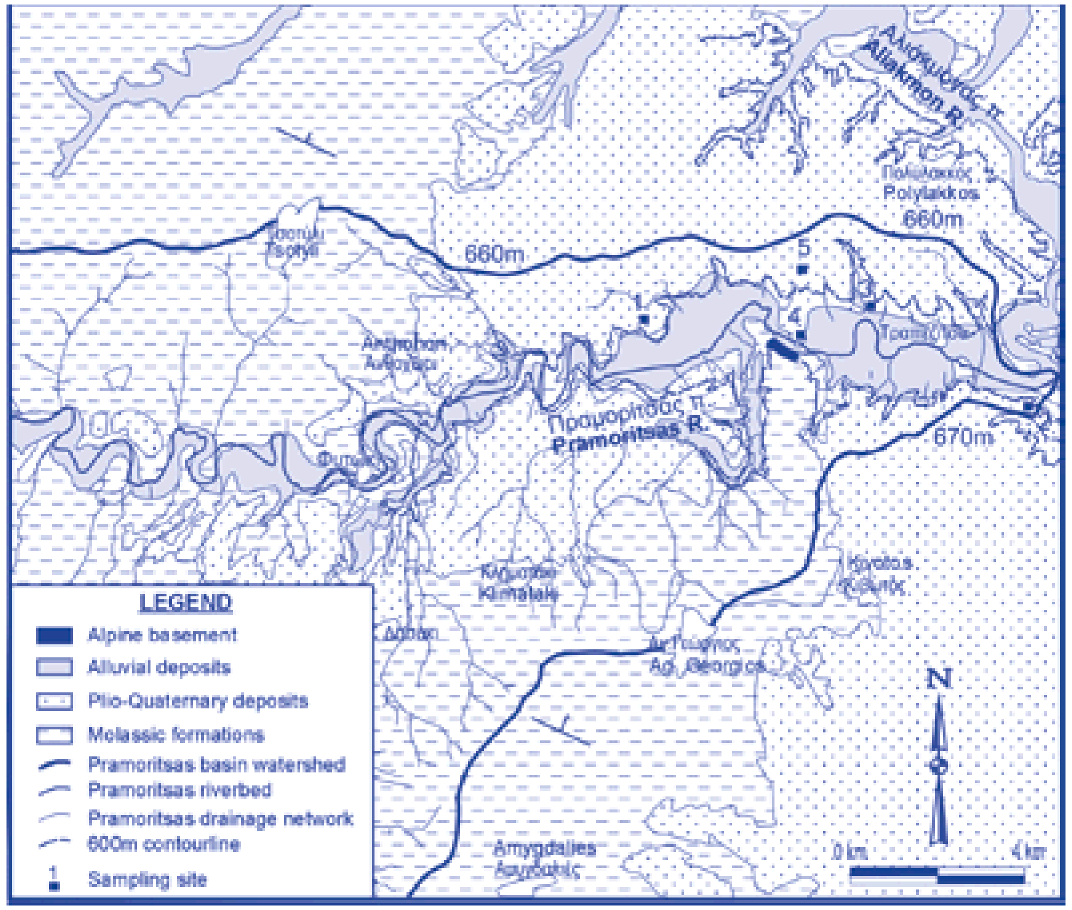


Figure 46 - Geological sketch map of the broader area of occurrence of Pliocene marine deposits in the Meso-Hellenic Trough (after Fountoulis et al., 2001).

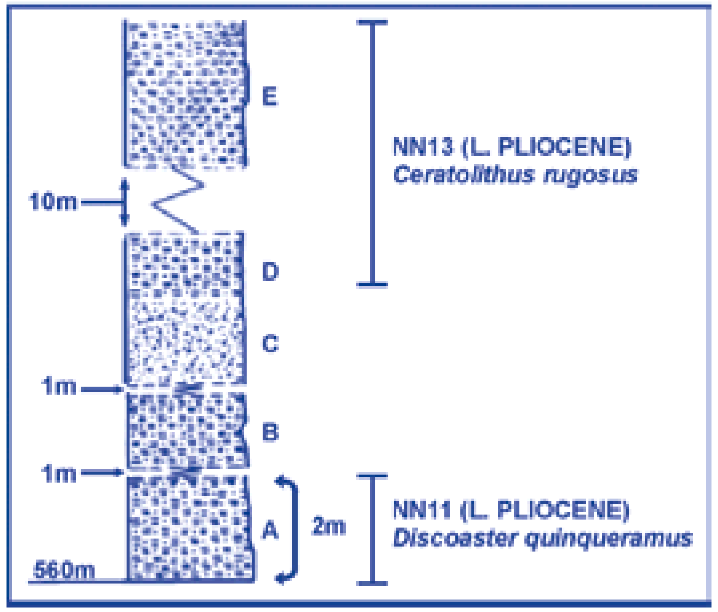


Figure 47 - Representative stratigraphic column of the Pliocene marine deposits on the banks of the Pramoritsa river in the Meso-Hellenic Trough (after Fountoulis et al., 2001).

findings; including benthic and planktonic foraminifera genera, but he did not determine any species. More specifically, he suggests that marine sedimentation in this part of the Meso-Hellenic trough continues throughout the Pliocene, without reference to specific species or locations.
Fountoulis et al. (2001), after a detailed lithostratigraphic study of the deposits on the banks of the Pramoritsa River (a tributary of Aliakmon River) north of the town of Grevena, and discovering numerous pelagic and benthic foraminifera and nannoplankton species determinations, have proven that marine sedimentation did not stop in the Late Miocene, but continued until at least the Early Pliocene (Biozone NN13) times (Figures 46 and 47).
The discovery and determination of these marine sediments of Pliocene age is very important, because it means that the sea north of Grevena communicated with the Mediterranean sea of that time. Thus, the question arises, through which routes did this communication occur?
East of the marine Pliocene basin, lie the Askion and Vourinos Mts and further to the east, the Kozani-Servia basin, filled with lacustrine Pliocene deposits. Moreover, no marine Pliocene deposits have been found to the south. To the west, the basin is bounded by the Pindic Cordilliera, and further to the west, the Ioannina and Konitsa basins develop, in which no trace of Pliocene marine deposition has been documented. This leaves a single possible exit through the northwest to the Ionian Sea, which has to be thoroughly investigated. All these data lead to the conclusion that most of the uplift movements in the broader area have taken place in the Quaternary (the last 1.6 Ma). Moreover, Bourcart (1922), as well as Brunn (1956), reported that during Villafranchian times, the Aliakmon river had to be flowing inversely to its present direction, that is, to the northwest.

The Pramoritsa strike-slip fault zone

The neotectonic structure of the broader Grevena area is characterized by the presence of two large-size tectonic structures, the common boundary of which is the Pramoritsa fault zone. The two macrostructures are: (i) the *Tsotili neotectonic structure* bounded by the Aliakmon river bed to the north, northeast, and east, and by the Pramoritsa fault zone to the south; (ii) the *Grevena neotectonic structure,* bounded by the Pramoritsa fault zone to the north, the Aliakmon river to the east and west, and by the Venetikos River to the south.

The Tsotili neotectonic structure presents the following characteristics (Figure 47): (i) In this block, occur the younger formations of the Meso-Hellenic Trough [Tsotili (Lower Miocene) and Ondria (Middle to Upper Miocene) formations]. (ii) The beds of the sediments possess constant dip direction towards the NE. (iii) The main tributaries of the drainage networks strike SW-NE and flow from SW to NE. (iv) The majority of the planation surfaces created on the molassic sediments dip towards NE. (v) The faults and fault zones occurring in the area strike mainly SW-NE and SE-NW, that is the same strike as the strike of the tributaries of the drainage networks. (vi) At the southern part of the block, and in the Pramoritsa riverbed, folded beds of molassic sediments of Tsotili Formation occur (scale of some meters), the axes of

Volume n° 2 - from B16 to B33

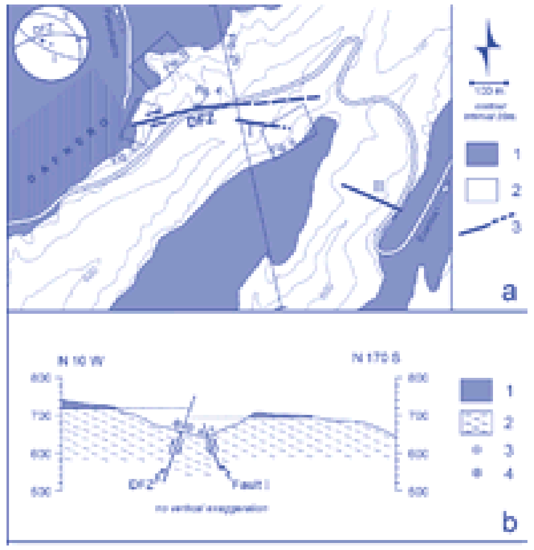


Figure 48 - (a) Geological map of Dafnero: 1: upper member, 2: lower member (see text for description), 3: faults; DFZ: Dafnero fault zone. Inset: stereographic projection (Schmidt projection, lower hemisphere) of the three mapped faults; the displacement vectors of DFZ and fault I also shown. Approximate locations of figs 3-6 - rectangles. (b) Cross section- relationship of DFZ with the outcropping Plio-Quaternary deposits: 1: upper member, 2: lower member, 3: fault displacement towards reader, 4: fault displacement away from reader. Note that fault I does not displace the upper member (polymictic conglomeratic breccias) (after Fountoulis et al., 2000).

which trend NW-SE.
The *Grevena neotectonic structure* presents the following characteristics: (i) In this neotectonic structure, occurs mainly the Pentalofon (Upper Oligocene – Aquitanian) Formation and there is no evidence of Tsotili (Lower Miocen) and Ondria (Middle to Upper Miocene) formations; (ii) The beds of the sediments do not have a constant dip direction≠ the beds dip either towards NE, or towards NW and SW; (iii) There is no sufficient geometry for the distribution of the planation surfaces, nor sufficient dip direction; (iv) The drainage network has no sufficient geometry ≠– locally (e.g. the Amygdalies area), it presents a radial arrangement; (v) The faults and fault zones occurring in the area strike mainly E-W and N-S, and occasionally occur in pairs of conjugate faults striking NNW-SSE and NNE-SSW; (vi) In the northern part of the block, and close to the Pramoritsa river-bed folded beds of the molassic sediments of the Pentalofo Formation occur in various scales, the axes of which strike NW-SE and NE-SW. Locally, the fold axes plunge in a radial way.
As already has been described, the Pramoritsa fault zone is the boundary between the Grevena and Tsotili neotectonic structures. It occurs at the southern banks of the Pramoritsa River, and has the same (E-W) mean strike. The fault surface dips at steep angles (70-80º) towards north-northeast (25º) or towards south-southwest (210º). Almost horizontal striations (126º/20º) can be observed on the fault surface. In other words, the Pramoritsa fault is a typical left-lateral strike-slip fault. The fault surfaces occur at the Pentalofo conglomerates, but it is difficult to observe the fault continuation in the marls.

Stop 6.7:

Paleokastro - Mt. Vourinos

Basal conglomerates of the molasse deposits are exhibited at the eastern margin of the Meso-Hellenic trough. The conglomerates consist of pebbles and cobbles, that come exclusively from the metamorphic rocks mostly located to the east, northeast, and southeast of the occurrence. There are no pebbles

Photo 6.7 - Fault surface with striations of the Pramoritsa fault zone. View from the North.

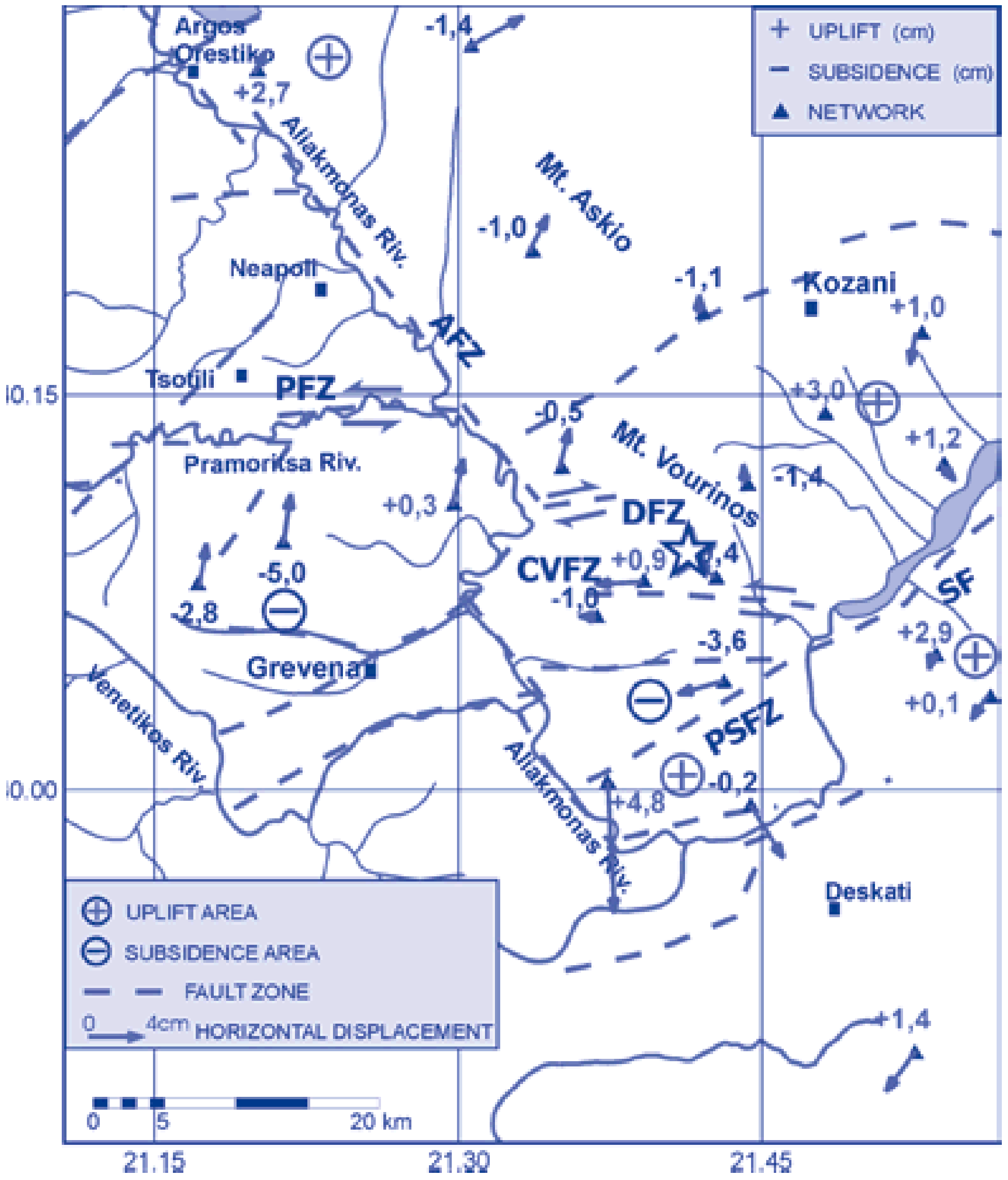


Figure 49 Horizontal and vertical displacements of the distinguished blocks, based on GPS measurements for the period May 1995 - September 2000 (after Fountoulis et al., 2002).

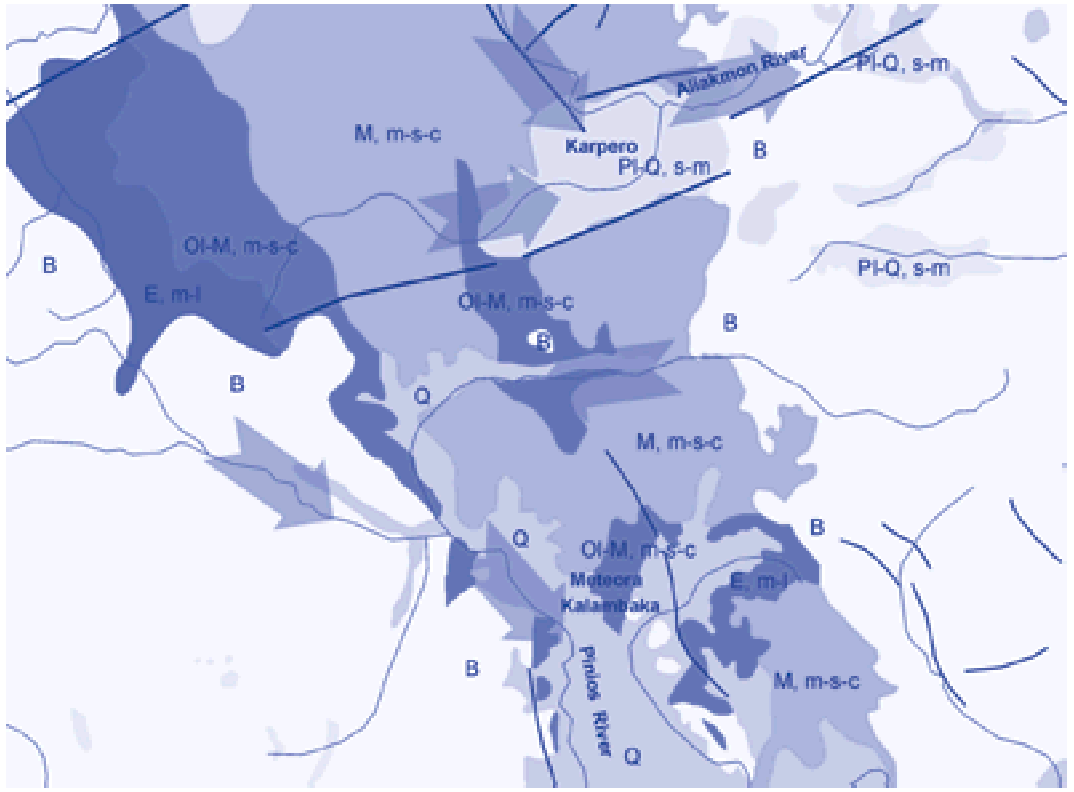


Figure 50 - Schematic geological map showing the area of the drainage divide between the Aliakmon and Pinios drainage basins. B: Alpine basement, E,m-l: Upper Eocene marls and limestones, Ol-M, m-s-c: Oligocene - Miocene marls, sandstones, conglomerates, M, m-s-c: Miocene marls, sandstones, conglomerates, Pl-Q, s-m: Pliocene-Quaternary sands, marls, Q: Quaternary deposits

coming from the underlying marbles of Mt. Askion.

The broader area has been evidently inhabited since the Paleolithic times, continuously until today. Human presence is verified by the oldest (for the Hellenic area) Paleolithic findings in Paleokastro area, while it shows dense population in Neolithic times.

Stop 6.8:
Dafnero

Dafnero fault zone

At the area of Dafnero, located at the NW edge of Mt. Vourinos, there is an impressive occurrence of an E-W trending fault zone (DFZ) (Figure 48). The outcrops there consist of Plio-Quaternary deposits that form well-expressed sub-horizontal planation surfaces throughout the eastern part of the Grevena basin and comprise: (i) a lower member (visible thickness more than 50 m.) of cohesive monomictic (ophiolitic) conglomerates (well-rounded pebbles) with sandy matrix, bearing intercalations of soil horizons, and (ii) the upper member of fairly sorted loose polymictic conglomeratic breccias, with a thickness of 10-30 m. All studies so far have left the two members undifferentiated and the age assigned to them is Plio-Pleistocene, according to Faugeres and Vergely (1974), and Mavridis and Kelepertzis (1994), or Villafranchian, (2.0 Ma BP) according to Koufos and Kostopoulos (1993) and Rassios *et al.* (1996). The Dafnero fault zone (DFZ) has a steep 70°° northern dip, and striations measured on some of its constituent surfaces were very slightly inclined (plunge 10-13°°), yielding a dextral character with a slight reverse component. It cuts the monomictic conglomerates, as well as the lower part of the overlying upper member. The reverse component of DFZ is clearly shown: (i) at the slopes of the stream running to the east of the village, where the polymictic horizon is displaced by approximately 0.50 m (northern-side up); and (ii) at a natural cut normal to DFZ, where between two

Volume n° 2 - from B16 to B33
Volume n° 2 - from B16 to B33

shear surfaces the two-generation gouge developed to follow the reverse slip component. The latter is also evidenced by the deep incision at the hanging-wall, while in the footwall incision is almost absent, if existing at all. Also, a set of veins, formed parallel to DFZ has developed at the hanging-wall for a width of about 30 m. Similar observations can be made elsewhere in Greece, where rocks of ophiolitic origin are faulted.

Another fault occurs about 200 m to the southeast; its has a NE (40°°) strike and dips SW at 70°°. The fault displaces the monomictic conglomerates, but not the overlying polymictic ones. Striation measured on it (220°°/36°°) showed a reverse character with a significant right lateral component.

No earthquake fractures could be found in the immediate area of DFZ.

DAY 7

Grevena - Kalambaka - Ioannina

The itinerary is concerned mostly with the Palaeogene, Neogene, and Quaternary sediments and tectonics of the Meso-Hellenic trough. After that, it crosses the Upper Cretaceous wild flysch of the West-Thessalia zone. The oceanic sediments of the Pindos zone are not exhumed along the road, and the itinerary enters directly the Palaeogene flysch of the Ionian zone. Then we continue to the picturesque town of Ioannina on the shores of the Ioannina Lake.

Stop 7.1:

Servia

The Ptolemais – Servia basin is bounded to the south with the impressive active Servia fault zone, that affects Holocene deposits (Mountrakis et al., 1998). This kind of deformation is responsible for the Basin-and-Range type of morphology that has developed across these faults.

The Servia area has been inhabited since Prehistoric times, and tombs, pottery and Neolithic tools were recovered in settlements, a major part of which is most of the time covered by the lake surface.

Stop 7.2:

Samara Rahi

During the recent seismic activity (Ms=6.6; 13.05.1995), few faults and fractures bearing signs of reactivation were found. Phenomena such as landslides, settlement damages as well as local distribution of damage that developed in a linear pattern, may mark the reactivation of some faults. One of these cases was the Chromio – Varis fault zone. Few *en-echelon* (right lateral) arranged fractures (E-W strike) were found at the location of Samara Rahi. These were the longest soil fractures created during the earthquake. Their total length was approximately 1 km. Vertical displacement of 20-40 cm was measured, as well as left-lateral displacement of 10-20 cm.

Photo 7.1 - Panoramic view from NW of the Servia fault zone.

Volume n° 2 - from B16 to B33

Photo 7.2 - Seismic-fracture created during the Grevena earthquakes (13-5-1995, Ms=6.6). View from the south.

Stop 7.3:

Paleohori

Recent Movements of the Upper Crust due to Creep Deformation based on GPS Measurements in W. Macedonia (NW Greece)

This is the location where the greater GPS measured subsidence was recorded for the years 1995 - 2000. In the broader area north of Paleohori, no drainage pattern has developed on the mollassic sediments, but a soft morphological lowering occurs instead. The area has been previously characterized as inactive by the seismologists, and thus one of the safest areas in Greece. The unexpected destructive seismic activity in western Macedonia (NW Greece) in May 1995, afforded the opportunity to study (Fountoulis et al., 2002) the current deformation processes using GPS measurements.

Between the years 1995 and 2000, a number of measurements were carried out at a network of observation pillars, most of them belonging to the Hellenic Triangulation Network. The coordinates were determined by terrestrial geodetic methods. The use of these pillars permits us to compare the results of satellite geodesy methods to their known coordinates on the Hellenic Geodetic Reference System 1987 (HGRS 87). The epoch of the HGRS 87 coordinates is assumed to be in the early 1980s, since the 1st order measurements were done in the period 1975 – 1979, and the 2nd order measurements, in the period 1982 – 1985.

In May 1995, a week after the main shock, 91 pillars were measured. These triangulation pillars were occupied without prior knowledge of the location of the earthquake epicenter. Further measurements took place on September 1995, May 1996, May 1998, and September 2000, where 37, 39, 56, and 59 sites were measured respectively, in order to check the aftershock behavior of the affected area. 53 of the 59 last measured (September 2000) sites belong to the Hellenic Triangulation Network. The whole network consists of 120 sites observed within 5 years. This high number of sites allowed us to draw some conclusions concerning the movements that took place due to the seismic activity, as well as the movements that took place in the period 1998 – 2000 without seismic activity.

Photo 7.3 - Panoramic view from northwest of the western part of the Paleohori-Sarakina fault zone. In this area is very characteristic the incision on the uplifted block.

The accuracy of the GPS results is estimated to be better than 1 cm for the horizontal components, and of the order of 2 cm for the vertical component.
For the time period May (?) 1980 – May 1995, there were 91 common sites, and the mean displacement was calculated to be 11 cm, and the mean velocity, 0.7 cm/y. This period includes 9 earthquakes before the main shock of the May 13, 1995, and 79 aftershocks with M>4 R.
For the time period May 1995 – May 1996, there were 19 common sites, the mean displacement being calculated to be 1.5 cm. This value is very high in comparison with the previous ones. This period includes 56 aftershocks with 4<M<5 R and only one (1) shock with M>5 R.
For the time period May 1996 – May 1998 there were 29 common sites, and the mean displacement was calculated to be 0.6 cm, and the mean velocity 0.3-cm/y. This period includes only 5 aftershocks with 4<M<5 R.
For the time period May 1998 – September 2000, there were 31 common sites and the mean displacement was calculated to be 0.8 cm, and the mean velocity 0.4 cm/y. This period includes only 3 earthquakes with 4<M<5 R. All epicenters located outside the mezoseismal area.
For the time period May 1995 – September 2000, there were 21 common sites and the mean displacement was calculated to be 2.8 cm and the mean velocity 0.5 cm/y.
It is known that a comparison of heights can be done between GPS coordinates only, since the heights in HGRS 87 are orthometric, and heights derived from GPS measurements are geometrical. So we present a comparison of heights between May 1995 and September 2000. The value of subsidence was calculated at 5 cm, and the mean subsidence velocity 1 cm/y, whereas the highest value of uplift was calculated at 4.8 cm, and the mean uplift velocity at 0.96 cm/y.
In conclusion (Figure 49), (i) The kinematics of the blocks resulting from the GPS measurements, are in good relation with the kinematics of the fault zones. (ii) Areas that are under a subsidence regime according to GPS measurements coincide with the areas that have been filled in with Plio-Quaternary sediments. The only exception is the area located NW of Grevena, which has suffered a complicated brittle-ductile deformation. (iii) For the 1996-1998, as much as for the 1998-2000 periods, there is a significant displacement of the blocks, without any important seismic activity. In W. Macedonia, the deformation processes in the moment is not only due to seismic activity, but even aseismic activity (“creep”).

Paleohori – Pontini and the Paleohori-Sarakina Fault Zone (PSFZ)

On the hills SW of Paleohori, a morphological scarp has been created by the activity of the Paleohori – Sarakina fault, trending E-W. The fault throw increases from east to west, while in the western area (at Kentro village), incision is very high on the uplifting block. This fault creates an intermediate tectonic block between the latter and the Aliakmon River, which is uplifting slower than the south margin of the Kamvounia Mts. This ENE-WSW striking fault zone is discernible through the topographic escarpment it has created in the Miocene deposits of the molassic Tsotyli Formation, while no actual fault break can be observed. The kinematics of this fault zone still remain to be clarified, as no surface kinematic indicator could be found. Therefore, it is probable that the Paleohori – Sarakina fault zone (PSFZ) is of analogous character to DFZ, so that the scarp created by it is less prominent than it should be for a normal fault.
A set of approximately E-W trending fractures was found about 500 m. before the northern entrance to Sarakina. The main fracture had a N 110°° trend, cut the asphalt road, and continued into the adjacent slope. It had no vertical displacement; however, most of the fractures had a 1 mm - 1 cm “heave”. The fracture set was *en echelon* arranged, and the fractures had a slight dextral offset.
Along the road that passes through the hills south of Paleohori, we have found a series of small-scale landslides. Two hundred meters after the road junction to Deskati, we found fractures cutting the tarmac and continuing into the soil. They are *en echelon* arranged, along an overall E-W trend, and a small ‘graben’ has been created between two sets.

Stop 7.4:
Karpero
Aliakmon drainage divide
Brunn (1958) suggests that the halting of the Aliakmon River flow towards Thessaly plains (to the south) may be explained by the transverse accident built by the Servia fault zone (which affects the Quaternary), and the Vounassa flexure. This transverse accident is prolongated westwards across the Meso-Hellenic Trough by the Kratsovon and the southern end of Krania paleogulf (Figure 50). A drainage divide between Aliakmon and Pinios drainage basins has been created on the Plio-Quaternary deposits. It has to be mentioned the opposite direction of flow of the two rivers: Aliakmon flows towards east-northeast, while Pinios flows towards west and then south.

Stop 7.5:
Meteora
Meteora is located at the northwestern boundary of the Thessalian basin, and at the southeastern limits of the Meso-Hellenic deposits. Meteora is among the biggest and most important groups of monasteries in Greece, after those in Mount Athos. The first evidence of the monasteries’ history can be traced back to the 11th century, when the first hermits settled there. The rock monasteries are characterized by UNESCO as a unique phenomenon of cultural heritage, and they form one of the most important sites on the cultural map of Greece.
The most important monasteries of Meteora are as described below:
The Holy Monastery of Great Meteoron is the biggest of the Meteora monasteries. The church “Katholikon”, dedicated to the “Transfiguration”, was erected in the middle of the 14th century and from 1387 to88, and decorated in 1483 and 1552. The old monastery is used as a museum nowadays.
The Holy Monastery of Varlaam is the second biggest monastery. The church, dedicated to three bishops, is of the Athonite type (cross-in-square, with dome and choirs), with a spacious “esonarthex” (litesurrounded by a dome. It was built in 1541/42, and decorated in 1548, while the esonarthex was decorated in 1566. The old refectory is used as a museum, while North of the Church, we can see the “parekklesion” (chapel) of the Three Bishops built in 1627, and decorated in 1637.
The Holy Monastery of Rousanou is dedicated to “The Transfiguration”, but in honour of Saint Barbara. The “Katholikon” (church) in the Athonite type, was founded in the middle of the 16th century, and decorated in 1560. Both the Katholikon and the reception halls are in the ground floor, while the “archontariki” (?), cells, and subsidiary rooms are scattered in the basement and the first floor.
The Holy Monastery of St. Nicholas Anapausas is the first we meet on our way from Kastraki to Meteora. The “Katholikon”, dedicated to St. Nicholas, is a single-nave church with a small dome, built in the

Photo 7.5 - Meteora

beginning of the 16th century. It was decorated by the Cretan painter, Theophanis Strelitzas or Bathas, in 1527.
The Holy Monastery of St. Stephen is one of the most visited, and can be reached without climbing innumerable stairs. The small single-nave church of St. Stephen was built in the middle of the 16th century, and decorated in 1545 or a little later. The "Katholikon", in honour of St. Charalambos, was built in the Athonite type, in 1798. The old refectory of the convent is used as a museum nowadays.
The Monastery of the Holy Trinity is very difficult to reach. The visitor has to cross the valley and continue high up through the rock before arriving at the entrance. The church is in the cross-in-square type, with the dome based on two columns, built in 1475-76, and decorated in 1741. The spacious barrel - vaulted esonarthex was founded in 1689, and decorated in 1692. A small skeuophylakeion (explain what it is) was added next to the church in 1684.
A great part of the monasteries (Katholika, cells, other buildings) have been restored and the rest of them are in restoration. Conservation of the frescoes has been fulfilled in most of the monasteries, too.

Stop 7.6:

Theopetra

The Theopetra Cave is located on the NE slope of the rocky limestone bulk of a hill, at whose foot lies the community of the same name. Its entrance is apsidal and large (17 x 3 m. approximately). The interior of the cave measures about 500 sq m. and small conches are formed in its periphery.
The Theopetra limestone has been dated to the Upper Cretaceous (100 – 65 Ma before present). The creation of the cave is consequently later than the above-mentioned date. The cave started being inhabited from the Middle Paleolithic period (which started at about 100,000 years before present).
The excavation started in 1987, under the direction of the prehistorian/archaeologist Catherine Kyparissi-Apostolika. The cave has been excavated for nine years now, and the excavations are still in process. Stone tools of the Paleolithic, Mesolithic, and Neolithic periods have been found, as well as Neolithic pottery, bone, and shell objects.
It is the first excavated cave in Thessaly, and the only one for the moment in Greece with deposits that start at the Middle Paleolithic, and last until the end of the Neolithic (3000 B.C.), without gaps.

DAY 8

Ioannina - Konitsa - Dodona - Ioannina
Fault zones in Epirus (NW Greece)

Epirus consists of a series of anticlines, formed and dominated by a sequence of Tertiary thrusts. These thrusts separate the land between the Ionian coast in the west, and the Pindos mountains in the east, into regions of different relief and rock types, and are dominated by Mesozoic carbonates, ranging in elevation between 1200 m to 1700 m, with some peaks at 2000 m. The highest ridges, striking NW (335º), are Mitsikeli in the east, and Paramythia, Kourenton and Kassidiaris in the west. The carbonate Ioannina plateau (P) lies in between. The eastern and western boundaries of the area have Oligocene flysch at outcrop, and are known as the Zagoria (Z) basin, and Botzara (B) synclines, respectively (Figure 51). The NW-SE fold-and-thrust trend is laterally shifted by East-West transverse faults (arrows); the Petousi-Souli (or Agia Kyriaki) to the South, the Soulopoulou fault in the plateau, and the Doliana fault zone northwest, in the Delvinaki (D) well area. Several large NE-SW trending normal faults truncate the fold and- thrust belt of N. Epirus. These include the Konitsa, Pogonianni and Doliana faults.

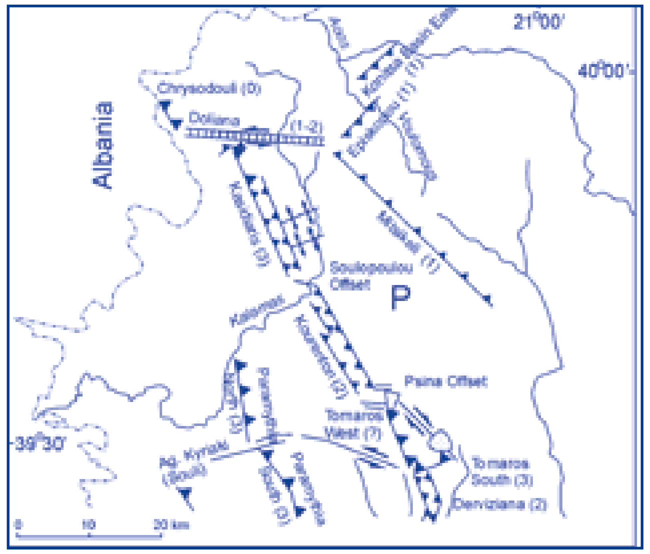


Figure 51 - Major tectonic features of northern Epirus (after Karageorgi et al., 2002).

Stop 8.1:

Konitsa marginal fault zone

Normal faulting with a NE-SW strike extends most of the way from Aridea to the fold-and-thrust belt of Epirus and coastal Albania. In particular, the Pogonianni and Doliana faults cut across the lines of anticlines that are associated with the continental collision between the Adriatic (Apulia) platform and Albania-Greece, though there is insufficient earthquake data to see how these structures are related at depth. The NE-SW normal faults cannot be traced continuously between Konitsa and the Ptolemais or the Grevena regions, but the intervening area consists mostly of flysch, in high, steep ground that is prone to landsliding. The faulting may be more continuous than it appears, but there is no historical seismicity to guide us.
The zone of NE-SW normal faults seems to cut across and mark the southern limit of the N-S normal faulting in eastern Albania and western FYR Macedonia. However, the well-constrained mechanism of the earthquake of 1 May 1967 at 39.5ºN 21.2ºE, showing normal faulting with a NNW-SSE strike, is an apparent anomaly that casts some doubt on this conclusion (Figure 52).

Both along the Konitsa and Pogonianni faults the footwall blocks (built up of Mesozoic limestones) have generated a relief of 500–1000 m. They are characterized by steep, relatively undissected ridges and exposed fault surfaces along the tops of a nearly continuous apron of scree, similar to faults in limestone elsewhere in Greece. Corrugated and striated slip surfaces along the 10 km long Konitsa fault, indicate slip vector azimuths of 295∞ in the central section, and 330∞ at the SW end. An earthquake of Mw 5.3 in the Konitsa region in 1996 had a focal mechanism with a slip vector of 290∞and may have ruptured part of the Konitsa fault. The Konitsa fault controls the northern limit of the Gamila limestone plateau, which rises up to 2400 m above sea level, and is drained by two rivers (the Aoos and Voidomatis), that cut deep gorges into the rising footwall block. The Doliana fault is much less pronounced, forming a minor ridge to the west of the Konitsa fault.

Palaeolithic human presence

The complex relief of the mountainous area of Konitsa, which is dominated by the drainage basins of the Aoos and Voidomatis Rivers, formed a pole of attraction for hunter-gatherer groups during the Late Palaeolithic period. In an environment rich in natural resources - raw materials, flora and fauna

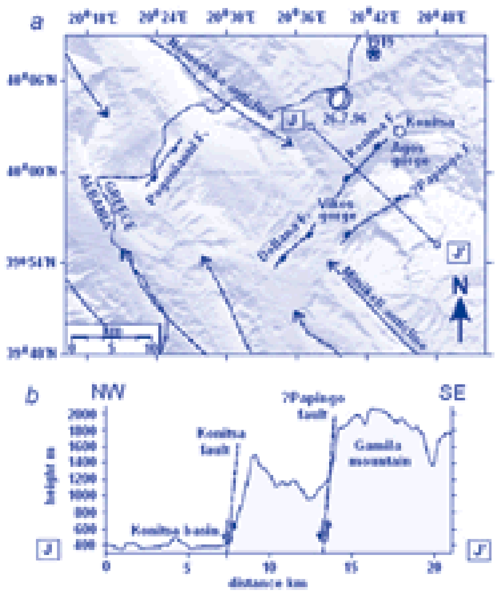


Figure 52 - (a) Major faults and earthquakes in the Konitsa region of northern Epirus. (b) Topographic profile across the Konitsa and possible Papingo faults (after Goldsworthy et al., 2002).

- the earliest evidence of human activity goes back to 17,000 A (years before the present, uncalibrated dates), and continues uninterrupted down to 10,000 A, when the local glaciers of western Pindos began to retreat and climatic conditions gradually improved. Based on natural refuges such as Klidi, Megalakkos, and Boila, groups of hunters systematically exploited the mountain biotopes. In these rock shelters, used as occasional/seasonal campsites, are preserved the remains of a way of life based on the consumption of natural available resources, rather than the growing of foodstuffs.
Flint, a hard stone procured from river beds, was used by people of the Palaeolithic period to manufacture their weapons and tools: arrows and spearheads, and a variety of cutting, scraping, and piercing tools with which to work other materials such as wood, hide and bone. They engaged in organized forays to hunt wild animals (caprines and deer), set traps for hares and birds, fished the rivers and gathered plants and fire-wood.
Although humble at first sight, this evidence is unequivocal testimony to a complex, organized society that adapted to palaeoclimatic and ecological

Volume n° 2 - from B16 to B33

fluctuations by exploiting the collective experience of a long cultural and biological history.

Although human presence may logically be inferred in the surrounding area of Konitsa from the Upper Palaeolithic to the 2nd millennium BC, with settlement centers being in accordance with the productive needs of Neolithic hunters-stockbreeders, there is no direct archaeological evidence for this. An indication of human presence in the river valley of the Aoos, at the end of the 2nd millennium BC, is furnished by the two bronze swords from Mesoyephyra. About the 13th-12th century BC, the mountainous area of the central and west Pindos was settled by the Molossoi, one of the most important tribes in the area of Epiros in terms of its political strength. Evidence for their arrival has recently been discovered on the Liatovouni hill in the Aoos valley, near its confluence with the Voidomatis.

From 1200/1100 to 730/700 BC, Epiros was isolated; a circumstance associated with the general upheavals and invasions in the Greek world. About the middle of the 8th century BC, peace was established, leading to a demographic growth that can also be seen in the cemetery of Liatovouni.

Photo 8.3 - Vikos gorge

Stop 8.2:

Kalpaki - Delvinaki - Doliana

The outcrop of Triassic anhydrite formations and evaporites in relation to the thrust structure and the Doliana fault zone will be demonstrated.

Stop 8.3:

Vikos

The Vikos Gorge, 900 metres deep and 12 kilometres long, regarded as the grand canyon of Greece, is a unique geological marvel. It is covered with lush vegetation, composed of various trees and flowers. The healing herbs, used by the renowned "quack" herbalists who worked in former centuries, were mostly collected from the Vikos. Many wild animals including the rare wild goat, still live in the gorge. Moreover, in the whole region of Zagori, th many kinds of wild animals still survive, such as wild boar, bears, wolves, ferrets, roe, deer, etc.

Vikos, and a part of Aoos, have been regarded since 1973 as the biggest national forest of Greece, covering an area of 126,000 square miles. Another well-known national forest, the one of Valia Kalnta, lies in Eastern Zagori, north of Vovousa. The national forest is considered to be one of particular natural beauty.

It takes seven hours to explore the Vikos, which extends from the bridge of Voidomatis, to the village of Koukouli. Beginning the exploration from Vitsa village (from the site called "Skala"), it takes about 20 minutes to get to the river Voidomatis, where the explorer can take a look at the Misios bridge, and the fantastic view. A path, passing between the two riverbanks, leads to the ravine, which is formed of perpendicular rock faces reaching a peak of 950 m. The width of the gorge, at riverbed level, fluctuates between 30 and 100 m. Because of its great length, some visitors do not intend to explore the Gorge. For them, it is possible to enjoy it from one of several viewpoints. We suggest three sites, which offer the best possible views into the Gorge, and namely: a) The Monastery of St. Paraskevi in Monodendri; b) Oxya above Monodendri on the mountain; c) Beloi, a view point about half an hour walk beyond Vradeto.

Volume n° 2 - from B16 to B33

Stop 8.4:

Dodona

The ancient site of Dodona is located 22 km south of Ioannina, in the narrow valley between mounts Tomaros and Manoliassa. The first remains on the site date from the prehistoric period, and the first deity worshipped here was the Earth goddess. The cult of Zeus and the sacred oak tree was brought to Dodona by the Selloi, a branch of the Thesprotian tribe, between the 19th and 14th centuries B.C., and soon became the prevalent cult of the sanctuary. The first offerings from southern Greece date from the end of the 8th century B.C., and building activity began in the 4th century B.C. The sanctuary reached the highest point of its prosperity in the 3rd century B.C., but was destroyed by the Aetolians in 219 B.C.. It was rebuilt shortly thereafter, and continued to be in use, until its destruction by the Roman invaders in 167 B.C.. In the Roman period, it had a different function, and its end came in the 4th century A.D., during the reign of Theodosius the Great. The area of the sanctuary was then covered with Christian basilicas. Today the theatre is used for performances.

The first excavations on the site, carried out by N. Karapanos in 1873-75, confirmed the location of the sanctuary and revealed a great quantity of finds. The following excavation campaign was undertaken shortly after 1913, by the Archaeological Society, under the direction of G. Soteriades but was stopped by the events of 1921 (which events?). The site was investigated again by D. Evangelides, in the period from 1929 until 1932, and systematic excavations started in the 1950s, under the direction of D. Evangelides and S. Dakaris (after Evangelides' death, they were continued by S. Dakaris). Since 1981, the excavations have been carried out under the auspices of the Archaeological Society, with the financial support of the University of Ioannina.

Systematic restoration work in the theatre, the stadium, and other monuments of the site started in 1961, and was based on the study by the architect B. Charissis. The whole project was financed by the Archaeological Society and the Program for Public Investments. Until 1975, the greatest part of the theatre had been restored, except for the third diazoma.

Photo 8.4 - Dodona, view of the ancient theatre and the Olitsikas Mt. from southeast.

The most important monuments of the site are: (i) the Sacred House (early 4th century B.C.); (ii) the Theatre, which is one of the largest in Greece, seating 18,000 people (3rd century B.C.); (iii) the Bouleuterion (end of the 4th century B.C.); (iv) the Stadium (late 3rd century B.C.); (v) the Acropolis (4th century B.C.); (vi) the Prytaneum (4th century B.C.).

DAY 9

Ioannina - Igoumenitsa

The itinerary of the last day crosses through the Ionian zone towards the Apulian platform. Mesozoic and some Cenozoic sediments are exhibited in tight folds and imbrications along the road.

Stop 9.1:

Petousi

A left lateral strike-slip fault zone forms the southern boundary of the Ioannina basin, Mt, Tomaros.

Stop 9.2:

Arillas

This stop displays the thrusting of the Ionian unit over Pliocene marine deposits.

Acknowledgments

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References cited

Armour-Brown, A., De Bruin, H., Maniati, C., Siatos, G. and Niesen, P. (1977). The geology of the Neogene sediments north of Serrai and the use of Rodent faunas for biostratigraphic control. Proc. VI Coll. Geology Aegean Region, Athens 1977, II, 615-622.

Aubouin, J., Le Pichon, X., Winterer, E. and Bonneau, M. (1979). Les Hellenides dans l’optique de la tectonique des plaques. Proc. VI Coll. Geol. Aegean Region, Athens 1977, Proc., III, 1333-1354.

Brunn, J.H. (1956). Contribution a l’ etude geologique de Pinde septentrionale et de la Macedoine occidentale. Ann. Geol. Pays Hellen., VII, 1-358.

Burchfiel, C. B., Nakov, R., Tzankov, Tz. andRoyden, L. H. (2000). Cenozoic extension in Bulgaria and northern Greece: the northern part of the Aegean extensional regime. in: Bozkurt, E., Winchester, J.A., Piper, J. D. (eds.) Tectonics and Magmatism in Turkey and the Surrounding Area. Geol. Soc., London, Spec. Publ. 173, 325-352.

Dabovski, C., Boyanov, I., Khrischev, K., Nikolov, T., Sapounov, I., Yanev, Y. andZagorchev, I. (2002). Structure and Alpine evolution of Bulgaria. – Geologica Balcanica, 32, 2-4; 9-15.

Dinter, D. and Royden, L. (1993). Late Cenozoic extension in northeastern Greece: Strymon valley detachment system and Rhodope metamorphic core complex. Geology, 21, 45-48.

Dinter, D. (1994). Tertiary structural evolution of the southern Rhodope metamorphic province: A fundamental revision. Proc. 7th Congress, Geol. Soc. Greece, Bull. Geol Soc. Greece, XXX/1, 79-89.

Eltgen, H., (1986). Feinstratigraphisch - Fazielle Untersuchungen an Pliozän - Sedimenten im Tertiarbecken Südlich Neapolis/Kozani, Nordgriechenland. ….G.M.E. Geol. Geoph. Res., Special Issue, 107-115.

Faugeres, L. and Vergely, P., (1974). Éxistence de deformations en compression d’âge quaternaire ancien (Villafranchien superieur) dans le Massif du Vourinos (Macedoine occidentale, Grèce). C.R. Acad. Sc. Paris, 278, Ser. D, 1313-1316.

Fountoulis, I. and Bakopoulo A. (1999). Morphotectonic observations in Pramoritsa River basin (Grevena, Greece). Proc. 5th Geographical Congress Geographical Soc. Greece, 94-100 (in Greek).

Fountoulis, I., Kranis, H., Lekkas, E., Lozios, S. and Skourtsos, E. (2000). Quaternary deformation in Grevena (W. Macedonia, Greece): Importance of shear and compressional strain. Ann. Geol. Pays Hellen. 38, Fasc C, p. 123-132.

Fountoulis, I., Marcopoulo-Diacantoni, A., Bakopoulo, A., Motaiti, E., Mirkou, M.R., and Saroglou, H. (2001). The presence of Pliocene marine deposits in the Meso-Hellenic Trough (Pramoritsa river banks, Grevena, Greece). Proc. 9th Congress, Geol. Soc. Greece, Bull. Geol Soc. Greece, XXXIV/2, 603-612 (in Greek).

Fountoulis, I., Paradisis, D., Veis, N. and Tsagaroulias, V. (2002). Recent Movements of the Upper Crust due to Creep Deformation based on GPS measurements

in W. Macedonia (NW Greece). In WEGENER 2002 Proceedings.

Gautier, P., Brun, J.P., Moriceau, R., Sokoutis, D., Martinod, J. and Jolivet, L. (1999). Timing, kinematics and cause of Aegean extension: a scenario based on a comparison with simple analogue experiments. Tectonophysics, 315, 31-72.

Goldsworthy, M., Jackson, J. and Haines, J. (2002). The continuity of active fault systems in Greece. Geophys. J. Int., 148, 596-618.

Gramann, von F. and Kockel, F. (1969). Das Neogen im Strimonbecken (Griechicsh-Ostmazedonien). Geol. Jb., 87, 445-484, Hannover.

Kahle, H.G., Straub, C., ReilingerR., McClusky, S., King, R., Hurst, K., Veis, G., Kastens, K., and Cross, P. (1998). The strain rate field in the eastern Mediterranean region, estimated by repeated GPS measurements. Tectonophysics, 294, 237-252.

Karageorgi, E., Watts, M. and Savvaidis, A. (2002). Integration and correlation of Geophysical data in NW Greece. EAGE 64th Conference and Exhibition – Florence May 2002, E-39.

Karistineos, N. and Gergiades-Dikeoulia, E. (1986). The marine transgression in the Serres Basin. Ann. Geol. Pays Hellen. 33/1, p. 221-232.

Karydakis, G. and Kavouridis, TH. (1988). Geothermal research of low enthalpy in Agia Varvara – Thermopigi Sidirokastrou (Serrai prefecture). Report Institute of Geology and Mineral Exploration.

Kavouridis, TH. and Karydakis, G. (1989). Geothermal research of low enthalpy in Sidirokastron area (Serres prefecture). Report E 5968 Institute of Geology and Mineral Exploration.

Kockel, F. and Walther, H.W. (1965). Die Strimon-linie als Grenze zwischen Serbo-Mazedonischen und Rila-Rhodope Massiv in Ost-Mazedonien. Geol. Jb., 83, 576-602.

Kockel, F. and Walther, H.W. (1967). Der Rhyolith von Strimonikon, sein tektonischer Rahmen und die junge Laggerstattenbildung in seiner Umgebung (Zentral – Mazedonien, Griechenland). Bull. Geol. Soc. Greece, VII/1, 1-16.

Kockel, F., Mollat, H. and Walther, H.W. (1977). Erlauterungen zur geologischen Karte der Chalkidiki und angrenzender Gebiete 1/100.000 (nord Griechenland). Bund. Fur Geowiss. V. Rohstoffe, 110 p. Hannover.

Kopp, K.O. (1966). Geologie Thrakiens III: Das Tertiar zwischen Rhodope und Evros. Ann. Geol. Pays Hellen., 16, 315-362.

Koufos, G.D. and Kostopoulos, D.S. (1993). A stenonoid horse (Equidae, Mammalia) from the Villafranchian of Western Macedonia, Greece. Bull. Geol. Soc. Greece, XXVIII/3, 131-143, Athens.

Koukouzas, K. (1972). Le chevauchement de Strymon dans la region de la frontiere greco-bulgare. Z. Deutsch. Geol. Ges., 123, 343-347, Hannover.

Lalechos, N. and Savoyat, ED. (1977). La sedimentation Neogene dans le fosse Nord Egeen. VI Coll. Geology Aegean Region, II, 591-603.

LE Pichon, X. and Angelier, J. (1981). The Aegean sea. Phil. R. Soc. London, A 300, 357-372.

Lekkas, E., Kranis, C., Fountoulis, I., Lozios, S. and Adamopoulo, E. (1996). Spatial distribution of damage caused by the Grevena -Kozani earthquake (May 13, 1995, W. Macedonia, Greece). In proceedings of Intern. Meeting on results of the May 13, 1995 earthquake of W. Macedonia: One Year after, Abstract p. 89-95, Kozani.

Lekkas, E., Lozios, S.G., Fountoulis, I.G., Kranis, H.D. and Adamopoulo, E.I., (1995). Investigation -Correlation of the geodynamic hazards at the earthquake stricken areas of Kozani - Grevena: Proposals for safe reconstruction. Unpublished Applied Research Project, Athens.

Makropoulos, K., Kassaras, I., Tzanis, A., Ziazia, M., Louis, J. andDiagourtas, D. (1996). The 13 May 1995 M-6.6 Kozani-Grevena aftershock sequence: towards understanding its dynamics and rupture processes, Proc., International Meeting: On results of the May, 13, 1995 earthquake of West Macedonia: One year after, p. 99-103, Kozani, Greece.

Maratos, G. (1967). The Sitsi-Kamen volcano on the Angistron mountain. Age and relation with the metallogenesis and the hot springs. Bull. Geol. Soc. Greece, VII/1, 93-106.

Mariolakos, I. and Papanikolaou, D. (1981). The Neogene basins of the Aegean Arc from the Paleogeographic and the Geodynamic point of view. Proc. Int. Sym Hell. Arc and Trench HEAT, I, 383-399, Athens.

Mariolakos, I. and Papanikolaou, D. (1987). Deformation pattern and relation between deformation and seismicity in the Hellenic arc. Bull. Geol. Soc. Greece, □…□, 59-76 (in Greek).

Mariolakos, I., (1975). Thoughts and viepoints on certain problems of the Geology and tectonics of Peloponnesus Greece. Ann. Geol. Pays Hellen.. 27, 215-313 (in Greek).

Mariolakos, I., Fountoulis, I., Logos, E. andLozios, S. (1991). Methods to study the torsional neotectonic deformation: the case of Kalamata area (SW

Peloponnesus, Greeece), in C. Qingxuan (Ed.) Proceedings of IGCP Project 250" Regional Crustal Stability and Geological Hazards, 3, 15-21, UNESCO-IUGS/ "Seismological Press" publications.

Mariolakos, I., Papanikolaou, D., andLagios, E. (1985). A neotectonic geodynamic model of Peloponnesus based on morphotectonics, repeated gravity measurements and seismicity. Geol. Jb., B-50, 3-17.

Mavrides, A., and Kelepertsis, A. (1993). Geological map of Greece 1: 50,000, "Knidi" quadrangle, I.G.M.E., Athens.

Mc Kenzie, D. (1970). Plate tectonics in the Mediterranean region. Nature, 226, 239-243.

Mc Kenzie, D. (1972). Active tectonics of the Mediterranean region. Geoph. J. R. Astron. Soc., 30, 109-185.

Mc Kenzie, D. (1978). Active tectonics of the AlpineHimalyan Belt. the Aegean Sea and surrounding regions. Geophys. J. R. Astron. Soc., 55, 217-254.

Mercier, J. (1966). Paleogeographie, orogenese, metamorphisme et magmatisme des zones internes des Hellenides en Macedoine (Grece): vue d'ensemble. Bull. Geol. Soc. France, 8, 1020-1049.

Mercier, J. (1968). Etude geologique des zones internes des Hellenides en Macedoine centrale (Grece). Ann. Geol. Pays Hellen., 20, 1-792 (1973).

Mountrakis, D. (1994). Introduction to the Geology of Macedonia and Thrace. Aspects of the geotectonic evolution of the Hellenic Hinterland and Internal Hellenides. Proc. 7th Congress, Geol. Soc. Greece, Bull. Geol Soc. Greece, XXX/1, 31-46.

Mountrakis, D., Pavlides, S., Zouros, N., Astaras, T., and Chatzipetros, A. (1998). Seismic fault geometry and kinematics of the May 1995 western Macedonia (Greece) earthquake. J. Geodynamics, 26/2-4, 175-196.

Papanikolaou, D. (1984). Introduction to the Geology of Greece: The pre-Alpine Units. In: I.G.C.P. No 5, 1984 Field meeting in Greece, Field guide, Part, I, 3-35.

Papanikolaou, D. (1984). The three metamorphic belts of the Hellenides: a review and a kinematic interpetation. Geol. Soc. London, Spec. Publ. 17, 551-561.

Papanikolaou, D. (1986). The geology of Greece 240 p. (in Greek).

Papanikolaou, D., Sassi, F.P., and Skarpelis, N. (1982). Outlines of the Pre-Alpine Metamorphisms in Greece. In Sassi and Varga (editors), I.G.C.P. No 5, Newsletter, 4, 56-62 and Ann. Geol. Pays Hellen., 31/1, 16-31.

Papazachos, B.C., et al. (1995). Focal properties of the 13 May 1995 large (Ms=6.6) earthquake in the Kozani area (N. Greece), Publ. Geophys. Lab., Aristotle Univ., 4, Thessaloniki.

Pavlides, S. (1985). Neotectonic evolution of the Florina-Vegoritis- Ptolemais basin (W. Macedonia, Greece). Ph.D. Thesis, University of Thessaloniki, 265 p. (in Greek).

Pavlides, S., Zouros, N.C., Chatzipetros, A.A., Kostopoulos, D.S. and Mountrakis, D.M. (1995). The 13 May 1995 western Macedonia, Greece (Kozani - Grevena) earthquake; preliminary results, Terra Nova, 7, 544-549.

Psilovikos, A., Sotiriadis, E., and Vavliakis, E. (1979). Quaternary tectonics and morphological differentiation of Sidirokastron basin. Ann. Geol. Des Pays Hellen., XXX/1, 588-601.

Struck, A. (1912). Zur Landeskunde von Griechenland, Frankfurt.

Tzankov, Tz., Angelova, D., Nakov, R., Burchfiel, B. C. and Royden, L. H. (1996). The Sub-Balkan graben system of Central Bulgaria. Basin Research 8, 125-142.

Yordanova, M., and Donchev, D. (eds.) (1997). Geography of Bulgaria. Academic Publishing House "Marin Drinov". Sofia, 730 pp (in Bulgarian).

Zagorchev, I. (1992). Neotectonics of the central parts of Balkan Peninsula: basic features and concepts. Geologische Rundschau 81, 3, 635-654.

Zagorchev, I. (1992). Neotectonic development of the Struma (Kraistid) Lineament, south-west Bulgaria and northern Greece. - Geological Magazine 129, 2, 197-222.

Zagorchev, I. (1996). Late Alpine (Palaeogene - Early Miocene) tectonics and neotectonics in the central parts of the Balkan Peninsula. Z. geol. Wiss., 24, 1/2, 91-112.

Zagorchev, I. (1998). Pre-Priabonian Palaeogene formations in Southwestern Bulgaria and Northern Greece: stratigraphy and tectonic implications. Geological Magazine 135, 1, 101-119.

Zagorchev, I. (2001). Geology of SW Bulgaria: an overview. Geologica Balcanica 21, 1-2, 3-52.

Zagorchev, I. (2002). Neogene fluviolacustrine systems in the northern Peri-Aegean Region. – Geologica Carpathica, Special issue (CD), 53.