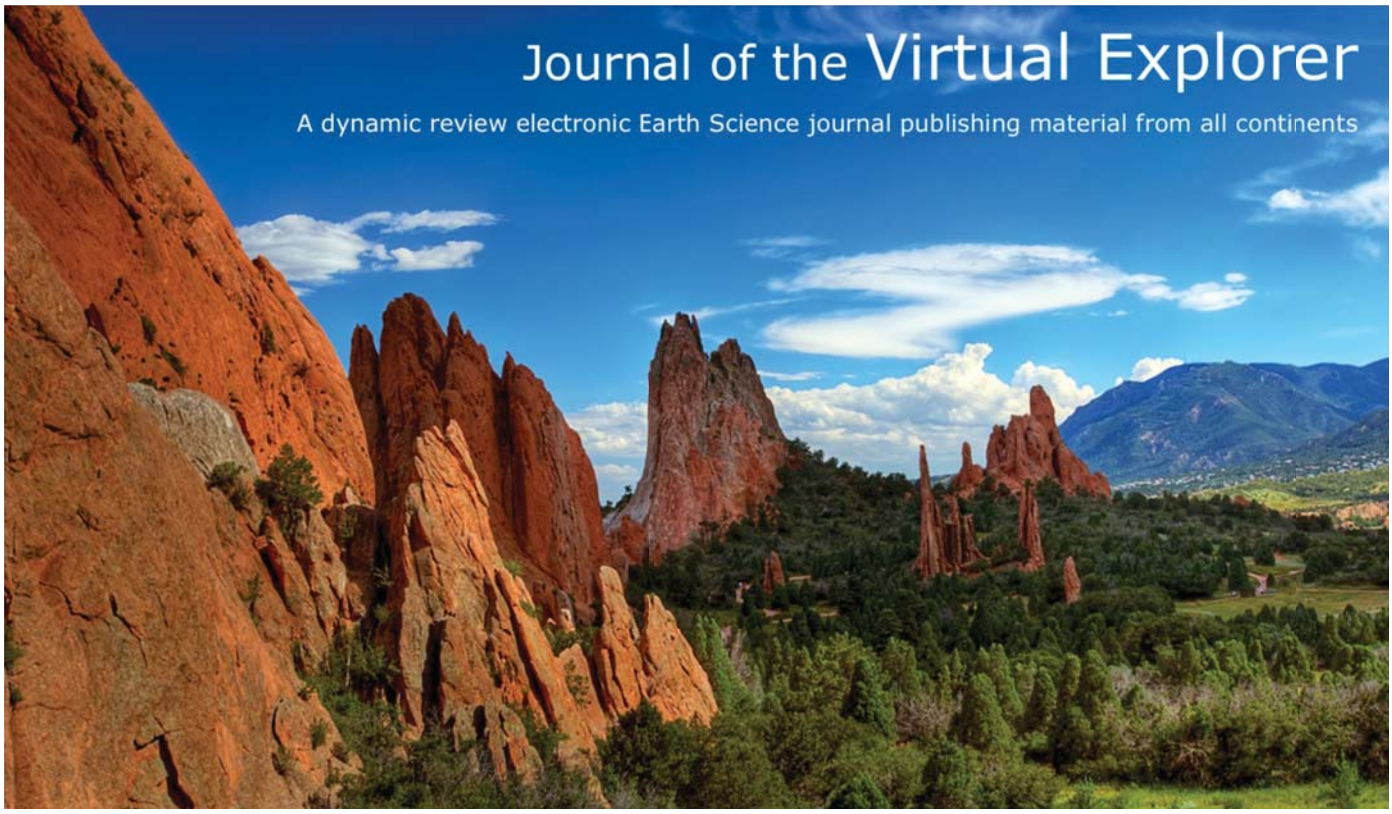

A geological cross-section through northern Greece from Pindos to Rhodope Mountain Ranges: a field guide across the External and Internal Hellenides

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IN THIS GUIDE

The aim of this work is to introduce the architecture and structural evolution of the Hellenic orogen in northern Greece, during the Alpine orogeny, through observations of selected outcrops, samples and view-points, along an east-west directed traverse in northern Greece, from the Pindos Mountain range (External Hellenides) to the Rhodope Mountainous province (Internal Hellenides). For this purpose, we describe and discuss in detail, during an 8-days field trip, the most interesting tectonic and sedimentary structures in macro- and micro-scale, of the geological units, occupied along this cross-section.
From the west to the east we cross: I. The Ionian, Gavrovo and Pindos nappes, II. The Tertiary sediments of the Meso-Hellenic trough, III. The Pindos and Vourinos ophiolite nappes with their underlain and overlain Mesozoic sedimentary units and ophiolite mélanges. IV. The Pelagonian nappe, IV. The Ampelakia unit and its underlain Olympos-Ossa unit, V. The Axios zone and VI. The Serbo-Macedonian/Rhodope metamorphic province of the Internal Hellenides.
The External Hellenides zones are characterized by a continuous sedimentation from the Triassic to the Paleogene-Neogene with a flysch deposition, as the last synorogenic sediment, being progressively younger from the east to the west due to the orogeny migration. A Tertiary thin-skinned tectonics marks the structural evolution of the External Hellenides.
The geometry and kinematics of the ophiolite bodies' emplacement on top of the Pelagonian nappe and the External Hellenides remain until today an open question. Are they obducted from one or more ocean basins?
The Pelagonian nappe constitutes a complicated, polymetamorphic, internal imbricated nappe, composed of a Paleozoic or older basement (gneisses, schists and Carboniferous granitoids) overlain by a Triassic-Jurassic clastic and carbonate cover. The Pelagonian nappe was emplaced on top of the External Hellenides in the Early Oligocene together with the Ampelakia unit, metamorphosed during Tertiary under high pressure - low temprature conditions. The tectonic contact between External Hellenides and the overlain Ampelakia-Pelagonian nappe pile has been reworked by an Oligocene-Miocene normal detachment fault related to the uplift and exhumation of the External Hellenides in the form of tectonic windows in the Olympos, Pieria, Kamvounia and Ossa Mountains.
The Axios zone is traditionally divided into the Almopias, Paikon and Peonias subzones, each one characterized by its own structure and deposition features.
The Serbo-Macedonian/Rhodope metamorphic province occupies the eastern most part of our section. It forms an Alpine heterogeneous nappe stack, consisting of both continental and oceanic crustal rocks. Synmetamorphic compression versus syn- to post- orogenic extension and progressively exhumation of deep crustal levels characterize the Alpine structural evolution of the Serbo-Macedonian/Rhodope province. Tertiary to Early Miocene granitoids intruded the whole massif during the extensional event. Ultra-high- and high-pressure metamorphic rocks occurring between the nappe stack and in the volca-

no-sedimentary Circum-Rhodope belt around the Serbo-Macedonian/Rhordope massif, indicate subduction processes during the nappe stacking.
The most important structures of the above described Hellenides geological sequences will be presented, in order to show the main structure of the Hellenic orogen and its structural evolution during the Alpine orogeny, taking also into account our views and interpretations concerning the tectonic setting and evolution history of the Hellenides..

INTRODUCTION

In this work we construct a perpendicular to the Hellenides cross-section in northern Greece, from the Rhodope province in the east to the Pindos Mountain range in the west (fig. 1, 2). It is considered as a highly representative cross-section of the Hellenides that involves the most important geological units of both continental and oceanic origin. Through this, we'd like to present the main geological structure and architecture of the complicated Hellenic orogen, formed during the Alpine orogeny. In this frame, we discuss the Alpine structural evolution of the Hellenides, taking into account our former knowledge and experience, as well as all related published international literature.

The proposed cross-section is described in an 8-days' field-trip across the main geological units of the Hellenides in northern Greece (fig. 2). Towards this, we've selected particular outcrops along the planned traverse, for which we describe, both in micro- and macro-scale, the composition and the characteristic tectonic-metamorphic and stratigraphic features of all occurring geological units. Finally, we reconstruct the possible structure of the Hellenic orogen and we depict its evolutionary stages during the Alpine orogeny, indicating the different tectonic events affected the Hellenides.

The guide was mainly planned for geologists, but it could also be useful for trekkers interested in the geology and natural aspects of the Hellenic mountainous region. In the first part of our work, we talk about the geology of Greece, and introduce the main composition and structures of the geological units that occur from Rhodope to Pindos Mountain. In the second part, we describe in

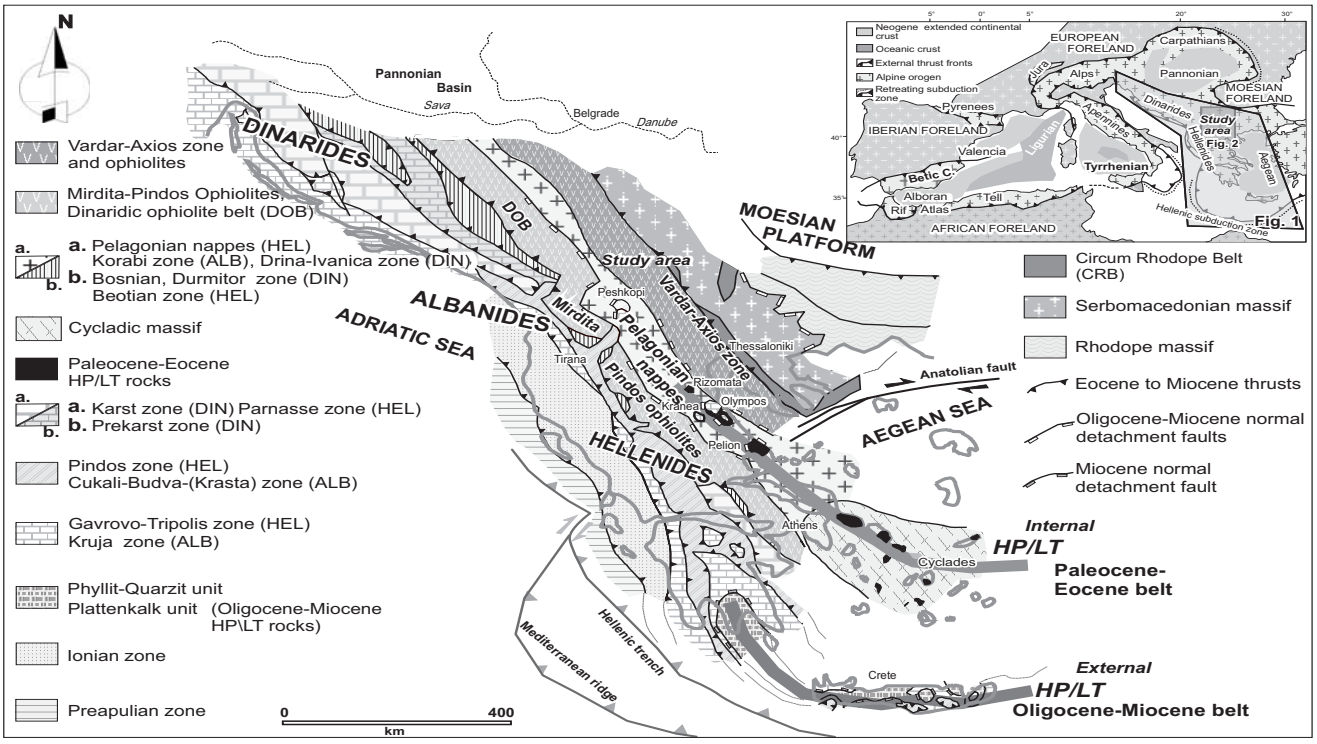


Figure 1
The main structural domains of the Dinarides and Hellenides. Insert: The extension of the Alpine orogenic belt in the Europe (Kilias et al. 2002).

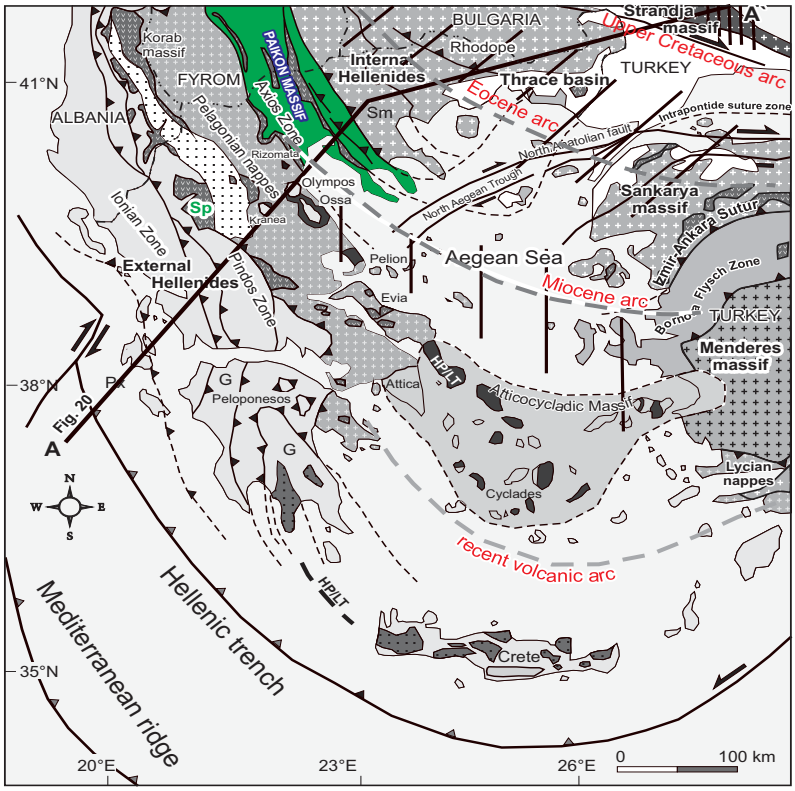


Figure 2
Geological map of the Hellenides with the main structural domains and their continuation to the adjacent orogenic belts (Kilias et al. 2013). The proposed cross-section is also shown.

detail the most important geological features and tectonic relationships of those geological units, through observations of outcrops at selected locations, samples and view-points along our E-W cross-section.

FIRST PART

Geological setting

As a branch of the broader Alpine orogenic system in Eurasia, the Hellenic orogen (fig. 1, 2) resulted from the continental collision of Laurasia with Apulia, where the latter corresponds to the eastern continental margin of Gondwana. Others again, rather envision a final continental collision of several Laurasia's and Gondwana's segments. Therefore, the Hellenic orogen is respectively associated with the closure of either one or more ocean basins between two continents or several continental segments (Frisch 1981, Stampfli and Borel 2002, Jolivet et al. 2004, Frisch and Meschede 2007, Gawlick et al. 2008). The question around the existence of one or more ocean basins between Laurasia and Gondwana continents remains under debate until today.

More specifically, some authors suggest that only one main ocean-basin existed, situated eastern and western of the Apulia and Laurasia continental margins respectively, and it corresponds to the Neotethyan Meliata-Axios/Vardar ocean basin. In this case, the Meliata-Axios/Vardar ocean basin is regarded as the main origin of all ophiolites belts met in the Hellenides, which were obducted on the continental margins with a mainly west-ward movement direction (fig. 2,3; Frisch 1981, Gawlick et al. 2008, Schmid et al. 2008, Kilias et al. 2010).

On the contrary, other researchers propose that the Axios/Vardar ocean-basin was a narrow basin between the Pelagonian microcontinent in the west and the Serbo-Macedonian/Rhodope province in the east, operating about simultaneously with a second, Pindos, ocean that was situated western of the Pelagonian microcontinent (Mountrakis 1986, Sharp & Robertson 2006, Robertson 2012). Alternatively, Axios/Vardar basin is also viewed as just a small ocean, part of a series of narrow ocean domains extending in the middle of small continental blocks between Gondwana and Laurasia continents (Papanikolaou 2009, Robertson 2012). In any case, according to any of those latter scenarios, the Hellenides' ophiolite belts derive from more than one ocean basins with different emplacement directions, either top-to-the-east or top-to-the-west (Robertson et al. 1996, Papanikolaou 2009, Sharp and Robertson 2006, Robertson 2012).

Recent works (e.g. Jahn-Awe et al. 2010, Burg 2012, Froitzheim et al. 2014) suggest a new theory, so-called the "maximum allochthony hypothesis". According to that, all the ophiolite belts of the Hellenides are allochthonous, originated from one main ocean rooted along the northeastern border of the Rhodope and the south western margin of the Laurasia. The Pindos ocean is regarded as a narrow ocean basin operating western of the Pelagonian block during the Late Cretaceous and closed in the Tertiary, when subducted completely under the Pelagonian.

Today, the Hellenides form an arcuate type orogenic belt extending from Dinarides (i.e. in Albania, FYROM, Bulgaria) to Taurides (i.e. in Turkey), which is traditionally subdivided into the Internal and External Hellenides zones (fig. 1, 2). The Internal Hellenides are characterized

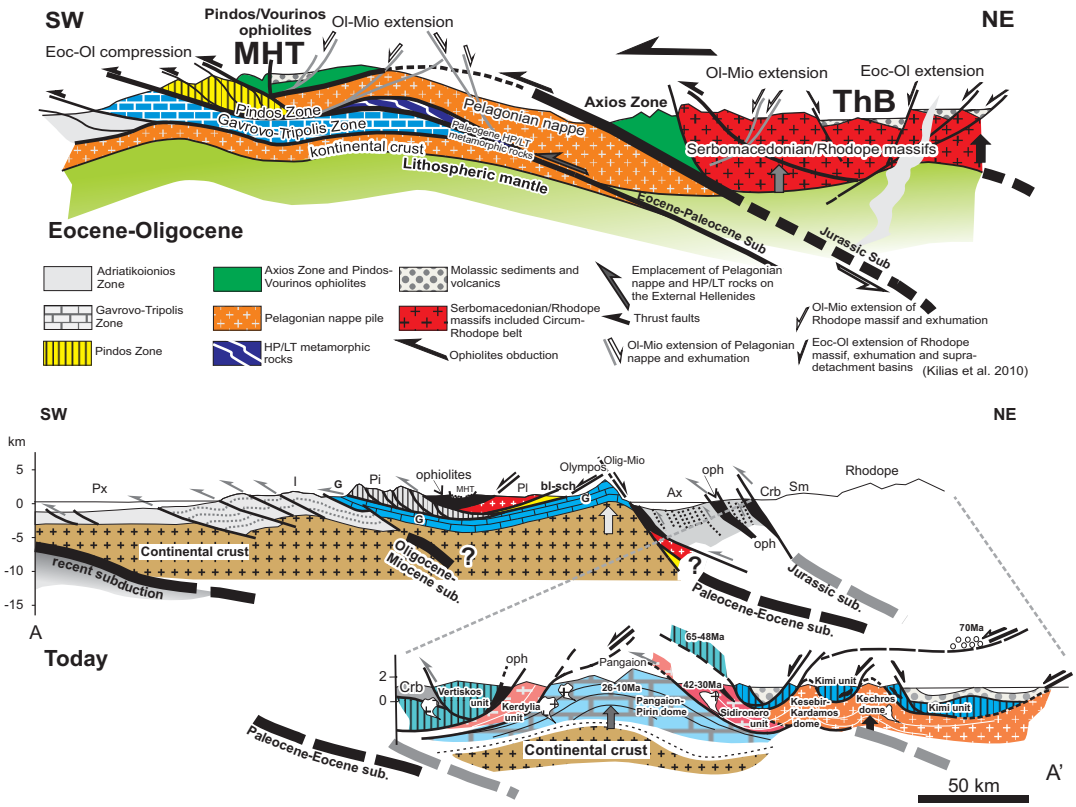


Figure 3
Geological cross-sections through the Hellenides during the Eocene-Oligocene and today. The geotectonic position of the MHT and THB molassic basins is shown (Kilias et al. 2013, 2015).

by Mesozoic, Paleozoic and older metamorphic rocks, initially affected by the Jurassic-Cretaceous tectonic event during the Alpine orogeny. On the other hand, the External Hellenides are mainly built up by Mesozoic and Cenozoic carbonate rocks, characterized by a continuous sedimentation that was terminated with a Tertiary to Neogene flysch deposition. They form a complicated SW- to SSW-verging thin thrust and fold belt of Paleogene to Neogene age, without any important metamorphism.

Maintaining the old subdivision of the Hellenides in isopic zones (i.e. Brunn 1956, Aubouin 1959), the External and Internal Hellenides are composed, from the west to the east, by the following main tectonostratigraphic domains (fig.1, 2):

A. External Hellenides

I. Paxos zone (PaZ)
II. Ionian zone (IoZ)
III. Gavrovo zone (GaZ), Parnassos zone (ParnZ)
IV. Pindos zone (PiZ), including Koziakas unit (KozU)

B. Internal Hellenides

I. Pelagonian nappe (PEL)
II. Sub-Pelagonian zone (SubPEL)
III. Axios/Vardar zone (AxZ)
IV. Circum-Rhodope belt (CRB)
V. Serbo-Macedonian massif (SRB)
VI. Rhodope massif (RHD)

A Paleocene-Eocene high-pressure/low-temperature (HP/LT) metamorphic belt is developed along the tectonic contact between the Internal Hellenides' Pelagonian nappe and the External Hellenides, where the first one is thrusted over the External (fig. 1, 2, 3; Scherner et al. 1990, Kilias et al. 1991). HP/LT metamorphic rocks of Oligocene-Miocene age are also recorded in between the External Hellendies further to the outer (western) parts of the Hellenic orogen in the southern Peloponnese and Crete island (fig. 1, 2, 3; Seidel et al. 1982, Kilias et al. 1994, 2010, 2013).

On the top of these so called tectono-stratigraphic domains of the Hellenides, Eocene to Early Miocene molassic sedimentary basins were developed, such as the Mesohellenic trough (MHT) (Ferrière et al. 1998, Vamvaka et al. 2006, Zelilidis et al. 2002) and the Thrace basin (ThB) (Maravelis et al. 2007, Kilias et al. 2013, 2015), as well as

Neogene-Quaternary intramontagne and other terrestrial sedimentary basins (fig. 3, 4).

Below, we present the main geological-structural features and composition of the Hellenides' tectono-stratigraphic domains that are cut along our traverse directed from the west to the east.

A. External Hellenides

I. Ionian zone (fig. 2,5)

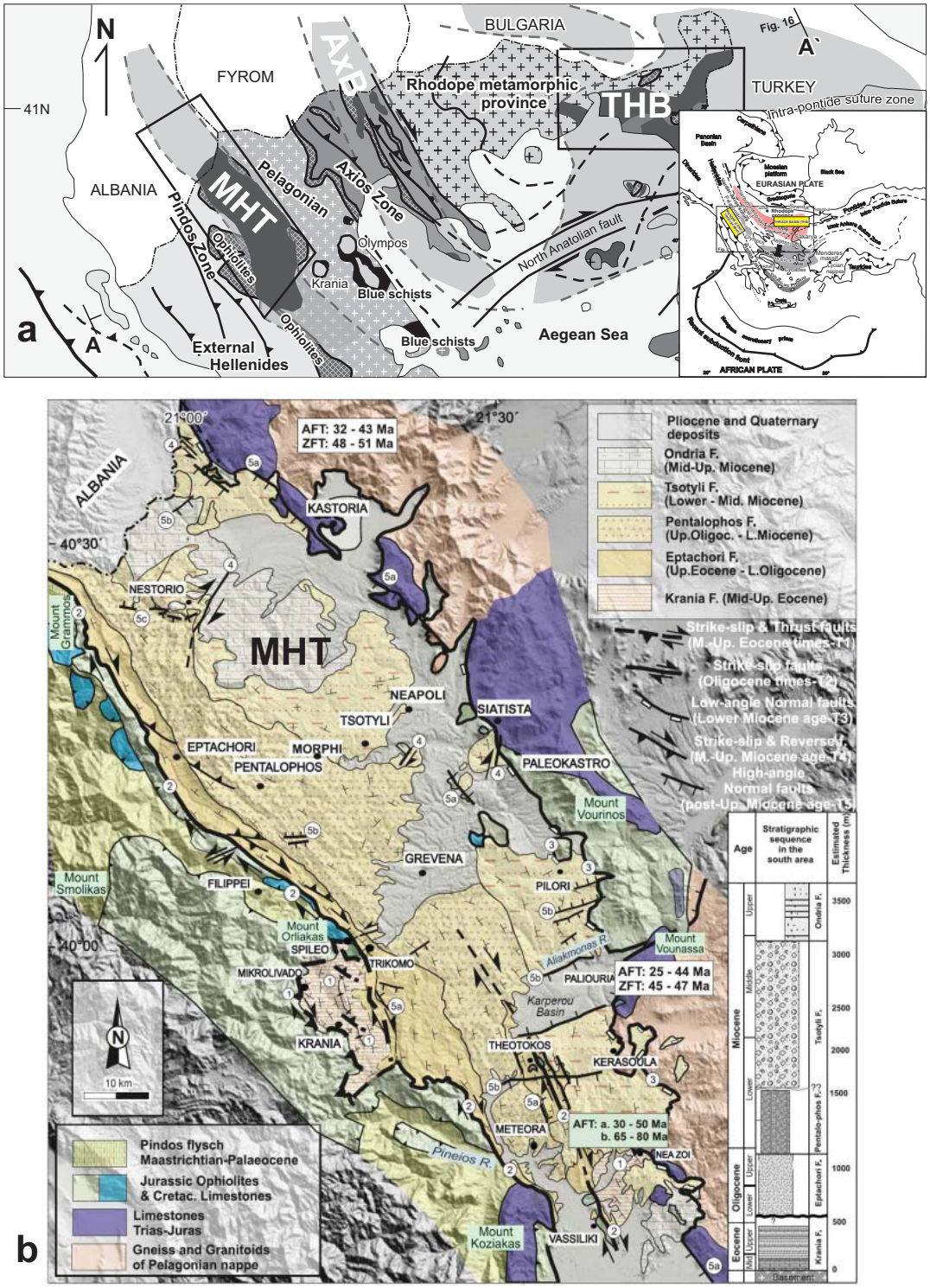


Figure 4
a. Location of the MHT and THB on the Hellenides is shown (Kilias et al. 2015). Insert: The Hellenides as part of the Alpine orogen in the Eastern Mediterranean region. The SSW-migration of the Tertiary magmatic activity in the Hellenides is also indicated (from Eocene until today). A-A´ the proposed cross-section. It is illustrated in fig. 3 and described in detail in the text. b. The geological map of the Mesohellenic Trough (Vamvaka et al. 2006).

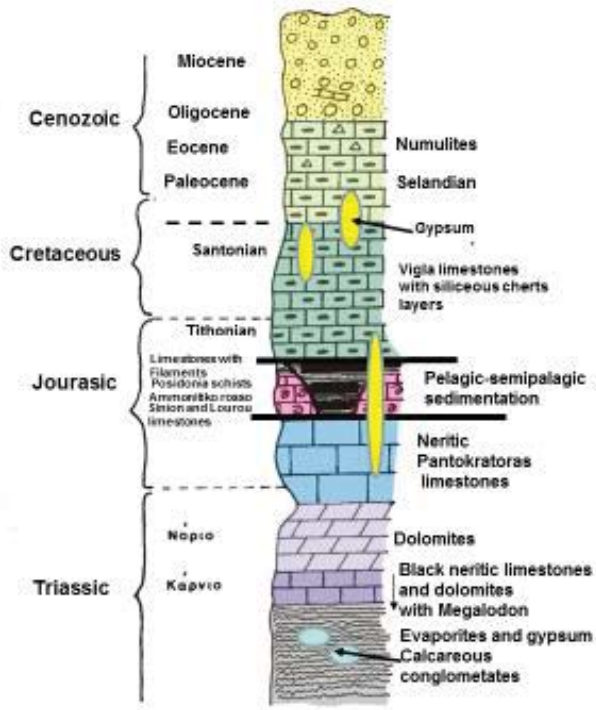


Figure 5
Schematic lithostratigraphic column of the Ionian zone (External Hellenides; Mountrakis 2010).

The Ionian zone is described by a continuous sedimentation from the Triassic to the Miocene. Until the Middle Liassic, the sedimentation was characterized by neritic calcareous deposits, while from then until the Eocene it changed to pelagic and finally ended with an Oligocene-Miocene flysch deposition (Brunn 1956, Aubouin 1959).

Equivalent to the Ionian sediments is the Plattenkalk series in Peloponnese and Crete, cropping out tectonically from below (/from beneath) the Phyllite-Quarzite unit and the Gavrovo zone. Both the Plattenkalk series and the Phyllite-Quarzite unit have been affected by the Oligocene-Miocene HP/LT metamorphic event of the External Hellenides (Seidel et al. 1982, Theye et al. 1992, Kilias et al 1994).

The Ionian zone is thrusted over the Paxos zone towards west, which forms the most external zone of the Hellenides, characterized by a continuous carbonate platform sedimentation from the Triassic to the Miocene-Pliocene.

II. Gavrovo zone (fig. 2,6)

The Gavrovo zone is described by a continuous Triassic to Late Eocene neritic sedimentation, terminated by the deposition of an Eocene-Oligocene flysch. The external (western) parts of the Gavrovo zone appear unmetamorphosed. However, its internal parts that crop out at the tectonic windows of Olympus, Ossa, Almyropotamos and Cyclades (fig. 2, 3) from beneath the Pelagonian nappe and the Paleocene-Eocene blue-schists belt, show evidence of a low- to high-grade metamorphism, or/and possibly of a HP/LT metamorphism (Schermer et al. 1990).

Recent works by Dinter (1998), Jahn-Ave et al. (2010), and Froitzheim et al. (2014) support the continuing of the Gavrovo carbonate platform until the Rhodope metamorphic province, and its outcropping as a tectonic window from below the metamorphic Rhodope units, named as the Pangaion metamorphic core complex.

The Gavrovo zone together with the Ionian and Paxos zones belong to the Mesozoic to Cenozoic passive continental margin of the Apulia continent.

III. Pindos zone (fig. 2,7,8)

The geotectonic position of the Pindos zone is widely controversial. It is believed to be either a part of an ocean basin opened progressively due to the Permo-Triassic continental rifting of Pangaea (Brunn 1956, Mountrakis 1986, Robertson et al.1996, Rassios and Dilek 2009, Rassios and Moores 2006, Robertson 2012), or a part of a narrow ocean basin initially opened during the

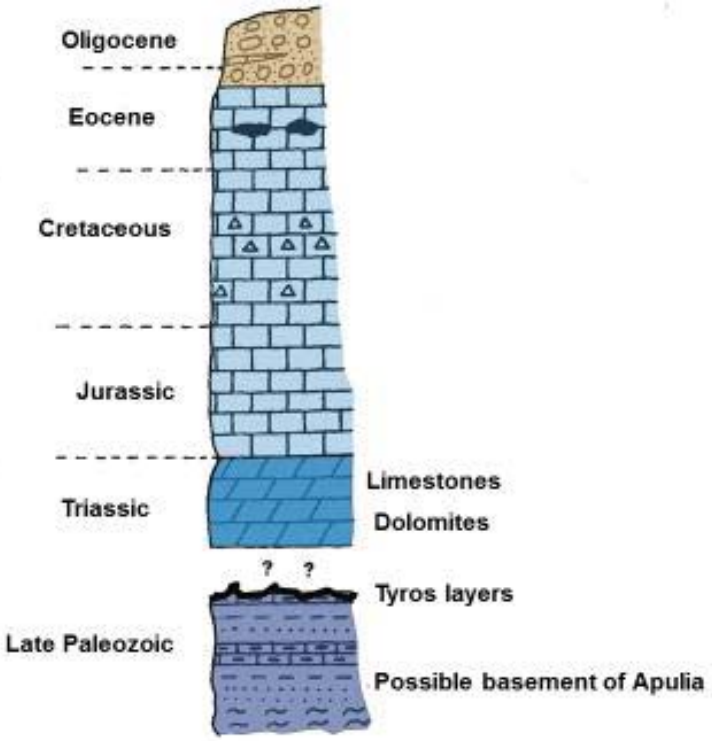


Figure 6
Schematic lithostratigraphic column of the Gavrovo zone (External Hellenides; Mountrakis 2010).

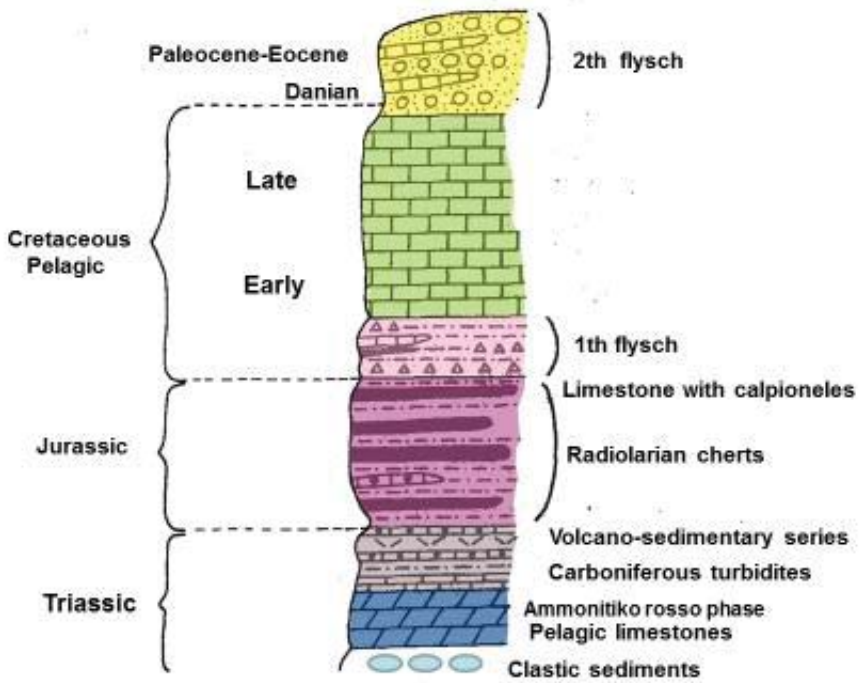


Figure 7
Schematic lithostratigraphic column of the Pindos zone (External Hellenides; Mountrakis 2010).

Late Cretaceous (Froitzheim et al. 2014), or even to be a basin formed on thin continental crust (Robertson et al. 1996). In any case, the Pindos zone is characterized by deep-sea sediments (siliceous and other pelagic deposits) of Triassic to late Late Cretaceous age, overlain in places by a Paleocene-Eocene flysch, which shares the typical features of a wild flysch (Brunn 1956, Aubouin 1959). Nevertheless, the Pindos first depositional materials are clastic neritic sediments of Ear-

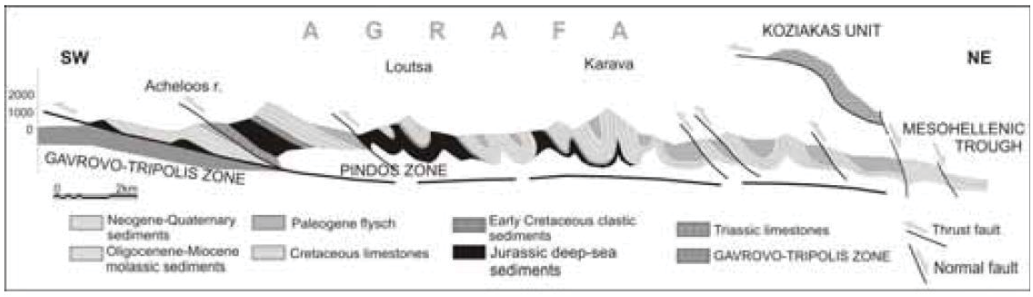


Figure 8
The characteristic W-ward vergent thin-skinned tectonic of the Pindos nappe overthrusted during the Eocene – Oligocene the Gavrovo zone. The Koziakas unit overthrusts again the Pindos nappe during the Tertiary.

ly-Middle Triassic age, with volcanoclastic products occurring in between them.

The Pindos zone is in a similar structural position as the Paleocene-Eocene blue-schists belt in Olympos and Cyclades provinces, directly above the Gavrovo zone. It possibly represents the non-metamorphic equivalent of the Paleocene-Eocene high pressure belt that has escaped the subduction under the Pelagonian continental crust (fig. 3; Bonneau 1984, Kilias et al. 2010). Stampfli et al. (2003) have challenged this interpretation by regarding the blue-schists belt of Pelagonian origin.

B. Internal Hellenides

I. Pelagonian nappe (fig. 2,9,10,11)

The Pelagonian nappe (Pelagonian zone) is composed of a Triassic-Jurassic carbonate platform sequence resting on top of a volcanosedimentary Permo-Triassic series, characterized by a bimodal volcanic activity and Paleozoic or older basement rocks (gneisses and schists), intruded by Carboniferous calc-alkaline (~300 Ma) and Triassic A-Type (~240 Ma) granitoids (fig. 10,11,12, 12a; Yarwood & Aftalion 1976, Mountrakis 1986, Poli et al. 2009, Koroneos et al. 2013). Obducted phiolites, imbricated with Mid-Late Jurassic ophiolite-mélanges, occur in many places on the Triassic-Jurassic Pelagonian carbonate cover. The ophiolites'obduction took place during the Mid-Late Jurassic, following the intra-oceanic subduction/-s (i.e. one or more) in the Neotethyan ocean basin, and was accompanied by the contemporaneous amphibolite-sole formation dated between 170 and 180 Ma (Spray and Roddick 1980, Spray et al. 1984, Kilias et al. 2001, Robertson 2012, Robertson et al. 2013). The Pelagonian nappe together with the obducted ophiolite belt crop out mainly in the continental part of Greece, but also on top of the Cyclades metamorphic complexes as small far-travelled relics, and in Crete known as the Asterousia nappe (Bonneau 1984, Kilias et al. 1994, Fassoulas et al. 1994).

Based on the two main theories about the Hellenic orogen formation, the Pelagonian basement is regarded either as I. the Mesozoic eastern passive margin of the Apulia continent, equivalent in this case to the External Hellenides, with a main wide ocean basin in the east, so-called the Neothethyan Axios/Vardar ocean basin (Frisch 1981, Gawlick et al. 2008, Schenker et al. 2014, 2015, Kilias et al. 2010, Tremblay et al. 2015) or as II. a microcontinent emerging in the middle of two separate ocean basins that were operating more or less contemporaneously during the Alpine orogeny; these were the Pindos ocean in the west and the Axios/Vardar ocean in the east (Mountrakis 1986, Robertson et al. 1996, Brown & Robertson 2004, Rassios and Moores 2006, Sharp & Robertson 2006). As accruing in these interpretations, the ophiolite rocks on the two, western and eastern, margins of the Pelagonian nappe should be originated either from one or two ocean sources, respectively. This leads to an ongoing discussion about the direction of the ophiolites' obduction onto the Pelagonian continent during the Middle to Late Jurassic times; either W- to SW-ward (e.g., Jacobshagen et al. 1978, Vergely 1984, Xoxha 2001, Gawlick et al. 2008, Schmid et al. 2008, Kilias et al. 2010, Bortolotti et al. 2013, Tremblay et al. 2015, Michail et al. 2016), or W- to SW-ward and E- to NE-ward (e.g. Mountrakis 1986, Robertson et al. 1996), or even only E- to NE-ward (e.g. Robertson & Shallo 2000, Sharp and Robertson 2006, Rassios & Dilek 2009).

A detail path of the tectono-metamoprhic history of the Pelagonian nappe, which is discussed in numerous works (e.g., Kilias et al. 2010, Most et al. 2001, Mposkos et al. 2001, Katrivanos et al 2013), is displayed in fig. 13.

The Sub-Pelagonian zone is composed of a sequence of Triassic to Jurassic pelagic carbonate and siliceous sediments, which lie tectonically

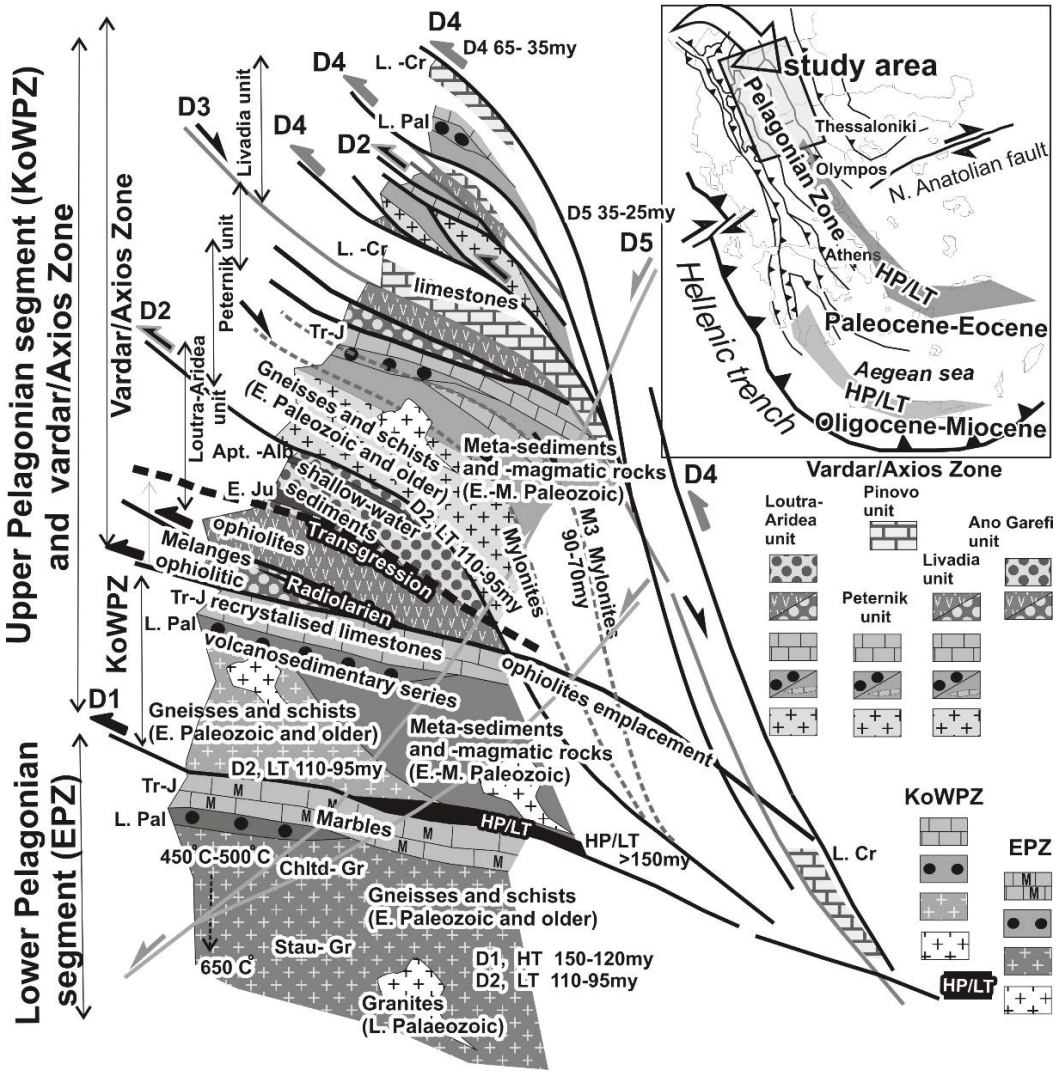


Figure 9
Alpine architecture and structural evolution of the Pelagonian nappe pile and western Axios/Vardar zone (Almopia subzone; Kilias et al. 2010).

over the Mesozoic platform carbonate sediments of the Pelagonian nappe. Today it is traced along the western Pelagonian margin. Characteristic phase of the Sub-Pelagonian zone built-up is the red, ammonites' bearing, pelagic sediments of the Hallstatt phase of Late Triassic age (fig. 14). Ophiolites and ophiolite mélanges overthrust the Sub-Pelagonian deep-water sediments. The latter emplacement took place during the Mid-Late Jurassic, following the general geotectonic history of the Alpine orogeny in the Hellenides.

However, the Sub-Pelagonian zone remains an under debate zone concerning its paleogeographic existence at the western flank of the Pelagonian nappe. One scenario wants it to be the Mesozoic continuation of the western Pelagonian continental margin to the continental slope and the deeper basin area towards the Pindos ocean basin (e.g., Ferrière 1974, Mountrakis 1986, Robertson et al. 1996, Robertson 2012). The opposing theory of only one, wide, ocean basin in the east of the Pelagonian nappe (i.e. Axios/Vardar ocean basin) explains the position of the Sub-Pelagonian zone along the western Pelagonian side as such of a tectonic nature, having been thrusted, together with the ophiolite belt, from the east to the west, on the Pelagonian nappe (Gawlick et al. 2008, Kilias et al. 2010, Bortolotti et al. 2013, Froitzheim et al. 2014, Schenker et al. 2014, 2015).

II. Axios/Vardar zone (fig. 2,15a,15b,16,17)

The Axios/Vardar zone is structurally another very complicated Hellenides' zone composed of units of both continental and oceanic origin. Traditionally, it is subdivided in three (3) subzones, which, from the west to the east, are: the Almopia

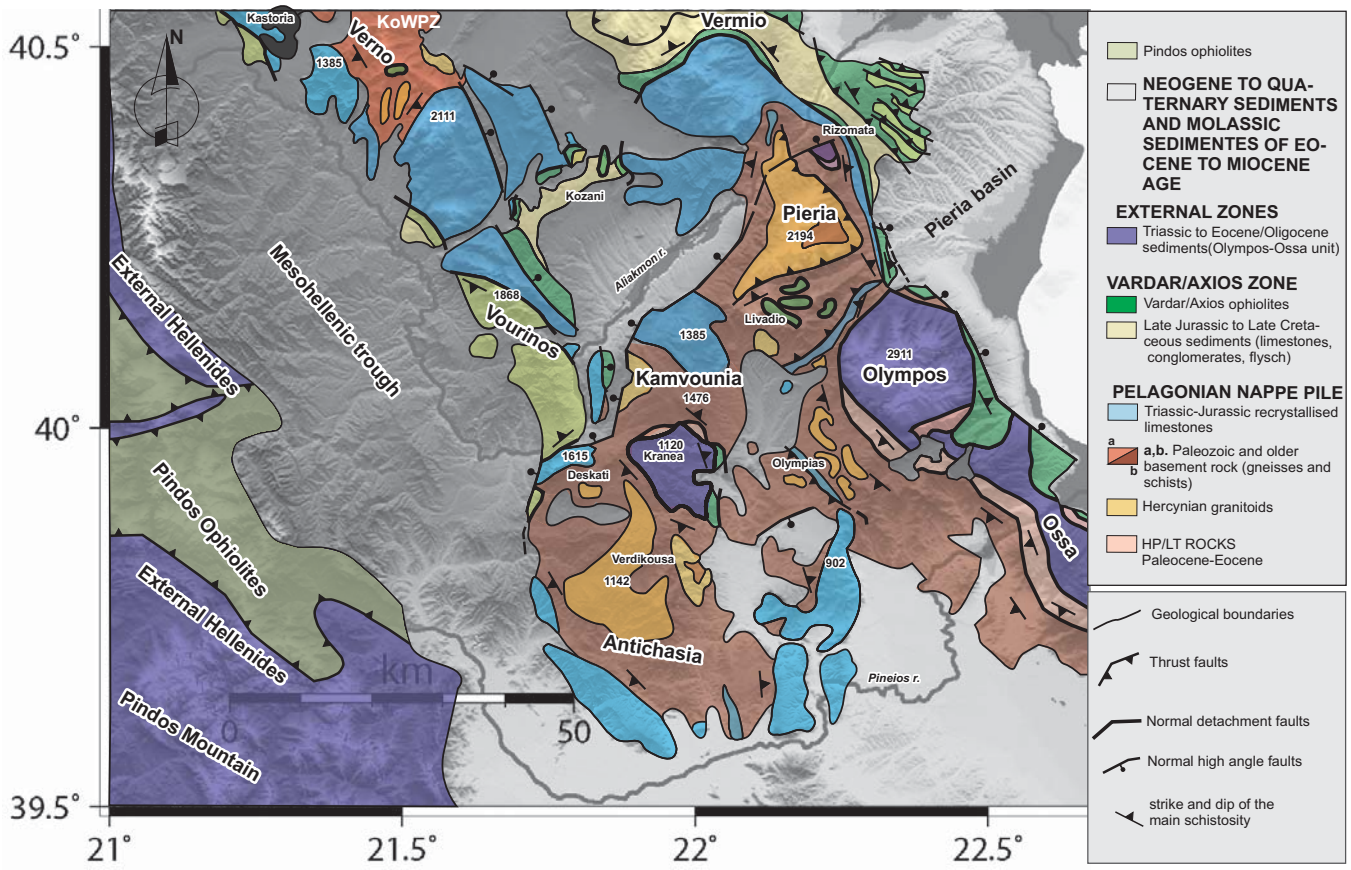


Figure 10
Geological map of the Pelagonian nappe pile in northern-central Greece (modified after Kilias & Mountrakis 1989, Koroneos et al. 2013).

(AlmS), the Paikon (PaikS) and the Paeonia (PaeoS) subzones (fig. 15, 16, 17, 18; Mercier 1968, Vergely 1984).

Several different interpretations have been proposed about the geotectonic position and structural evolution of the Axios/Vardar zone (e.g. Mercier 1968, Mountrakis 1986, Bonneau et al. 1994, Godfriaux & Ricou 1991, Galeos et al. 1994, Brown & Robertson 2004, Saccani et al. 2008, Katrivanos et al. 2013, Bortolotti et al. 2013). In a general view, the Axios/Vardar zone forms the suture zone between Gondwana and Laurasia, although its exact initial geotectonic position remains still under debate. The most popular theory is that Axios/Vardar zone's present position, between the Pelagonian and Serbo-Macedonian massifs, corresponds to the Neotethyan ocean, where the Middle-Late Jurassic subduction took place followed by the Late Jurassic obduction of the ophiolites on the Pelagonian continental margin (e.g. Robertson et al. 1996, 2013; Bortolotti et al. 2013). A more recent scenario considers the zone as allochthonous, deriving further from the East (e.g., Jahn-Ave et al. 2010, Froitzheim et al. 2014). In any case, parts of the Axios/Vardar zone were also accreted to the Laurasia margin located northeast of the Neotethyan ocean. These parts comprise ophiolites, accompanying Triassic-Jurassic pelagic and neritic sediments, as well as volcanosedimentary series of the Circum-Rhodope belt (Mercier 1968, Ricou et al. 1998, Bonev & Stampfli 2003, Mainhold et al. 2009, Jahn-Ave et al. 2010, Froitzheim et al. 2014, Schmid et al. 2008, Schenker et al. 2014). The Paikon subzone is regarded, either I. as a volcanic arc with the Paeonia subzone at the East as an ocean back-arc basin related to the subduction of the Almopias ocean (Brown & Robertson 2003, Sharp & Robertson 2006), or II. a tectonic window of the Pelagonian nappe, where the ophiolites of the Almopia's and Paeonia's subzones were obducted with a SW-ward sense of movement from a single ocean basin east of the Pelagonian continent (fig. 18,19; Kilias et al. 2010, Katrivanos et al. 2013, Michail et al. 2016). In this context the Paeonia subzone has been interpreted as a Jurassic supra-subduc-

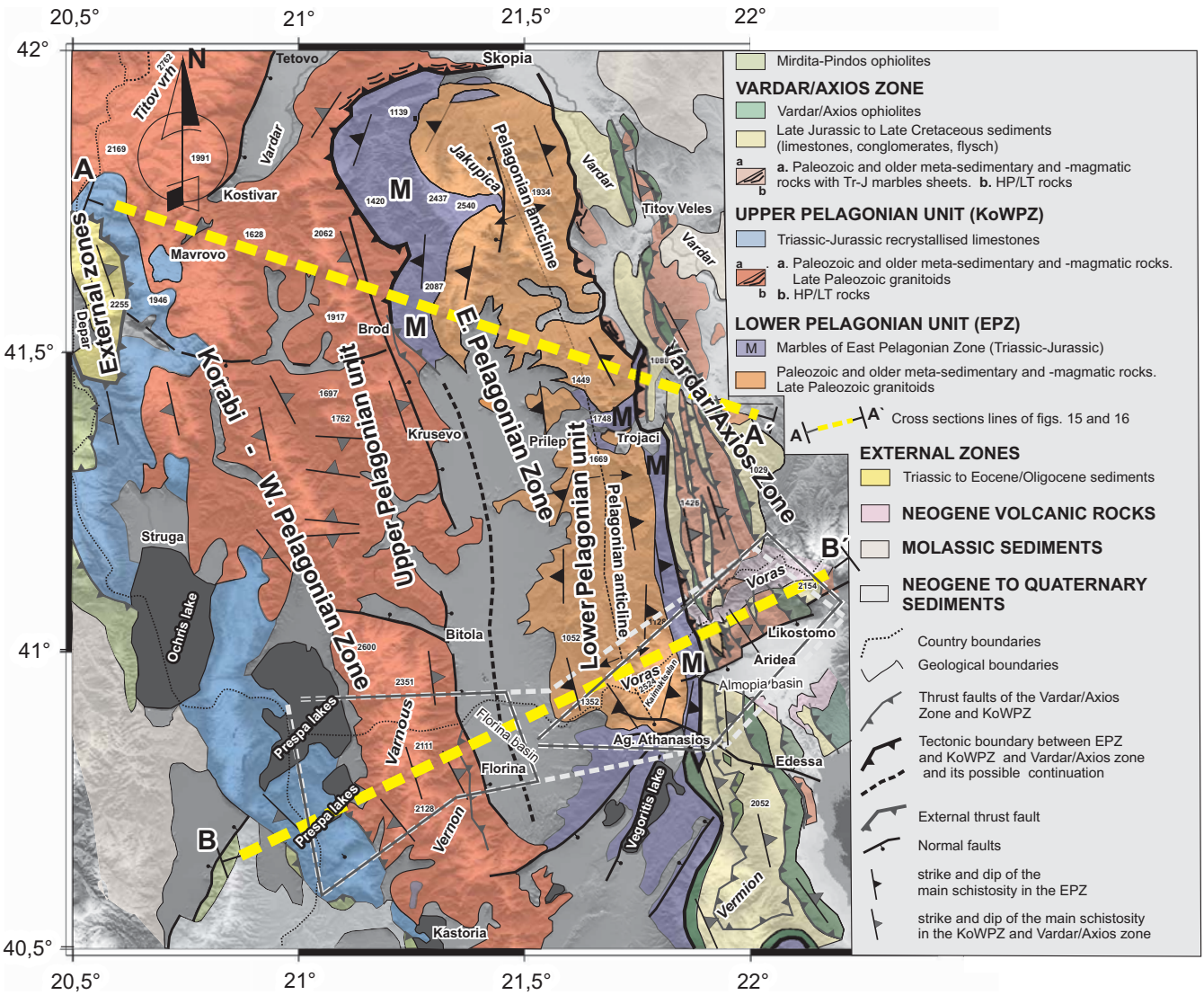


Figure 11
Geological map of the northern Pelagonian nappe pile and Axios/Vardar zone in FYROM and northern Greece. A-A′, B-B′ cross-section lines of fig. 29 b,c. The box area is shown in detail in fig. 15a & 29a (Kilias et al. 2010).

tion back-arc ocean basin behind of an ensimatic island arc during the Mid-Late Jurassic intraoceanic subduction (Kilias et al. 2010, Jahn-Ave et al. 2010, Froitzheim et al. 2014, Michail et al.2016).

III. Serbo-Macedonian/Rhodope metamorphic province (SRB/RHD; fig. 2,20,21)

In this work, we regard the Serbo-Macedonian massif as one with the Rhodope metamorphic province (Ricou et al. 1998, Jahn-Awe et al. 2010 Froitzheim et al. 2014). In this case, the Serbo-Macedonian/Rhodope metamorphic province is bordered to the west by the Axios/Vardar suture zone that is regarded as the main ocean basin and ophiolites' source of the Hellenides.

The SRB/RHD occupies the easternmost part of our cross-section. It constitutes the centre of the Hellenic arc-type orogen (fig. 1, 2), consisted of an Alpine nappe stack of several metamorphic units (nappes) of both continental and oceanic origin. The SRB/RHD is traced with the same structure further in the north, in the country of Bulgaria. The upper units of the Serbo-Macedonian/Rhodope nappes' stack represent the northeastern active margin of the Neotethyan ocean, which belongs to the Laurasia continental margin.

The several SRB/RHD nappes were progressively emplaced one over another by intense Cretaceous-Tertiary compressional tectonics and thrusting, in the peculiar position between the Apulia plate in the west and the Moesia platform in the east (i.e. part of Laurasia continent), in which the two plates were initially separated by the Axios/Vardar ocean basin (fig. 22; e.g.

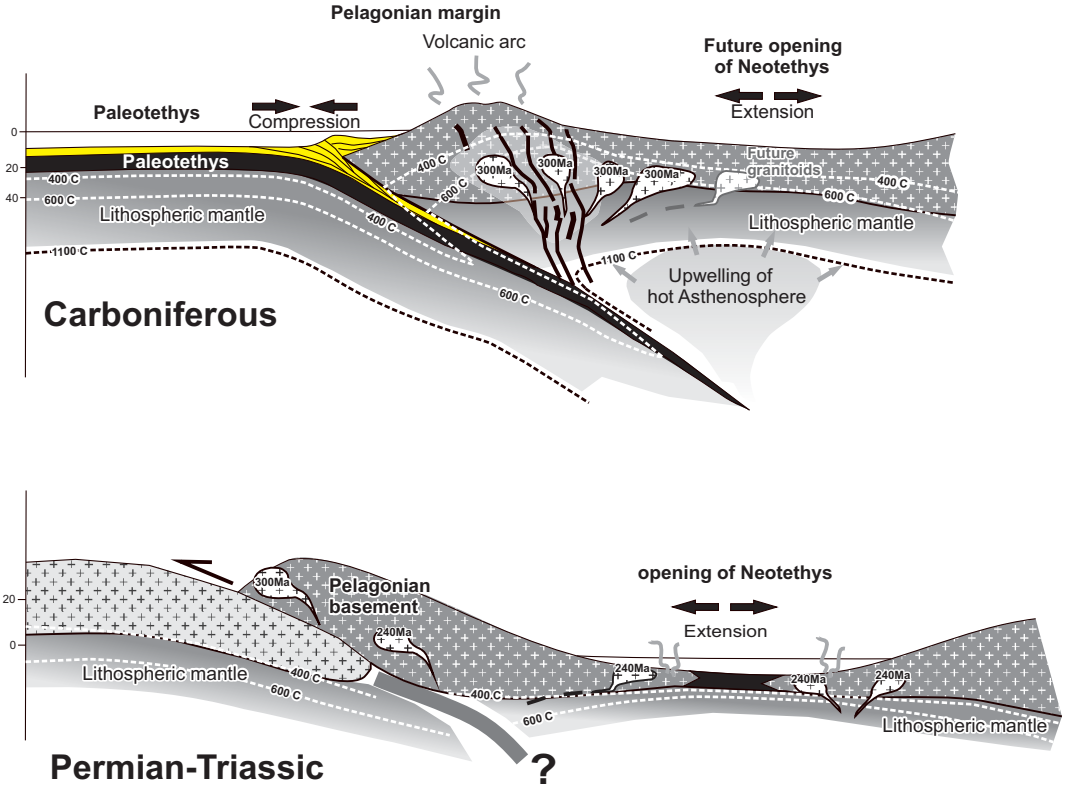


Figure 12
Schematic crustal scale transects showing the geotectonic setting of the Hercynian granitoid intrusions (~300Ma) into the Pelagonian continental margin and the Permo-Triassic granitoid intrusions (240 Ma) during the initial stages of the Neotethys opening (Koronaios et al. 2013).

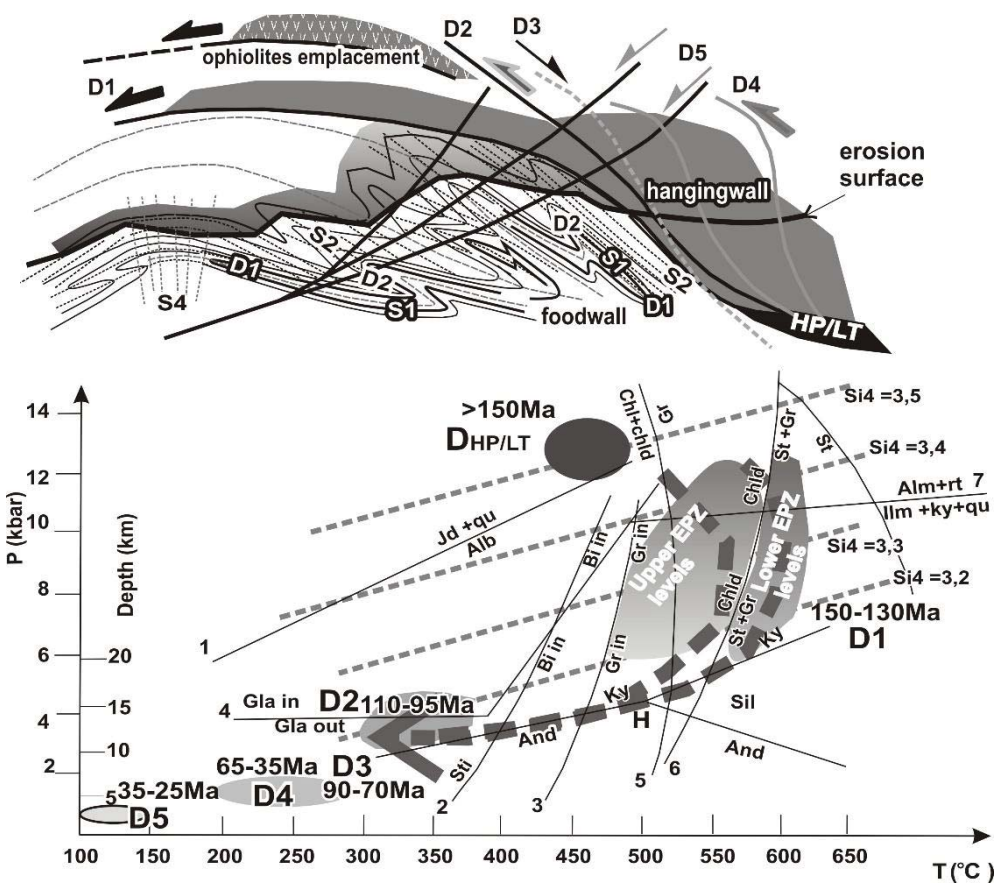


Figure 13
Schematically the geometry and kinematics of deformation of the Pelagonian nappe during the Alpine orogeny and the associated P-T metamorphic path are shown (Kilias et al. 2010).

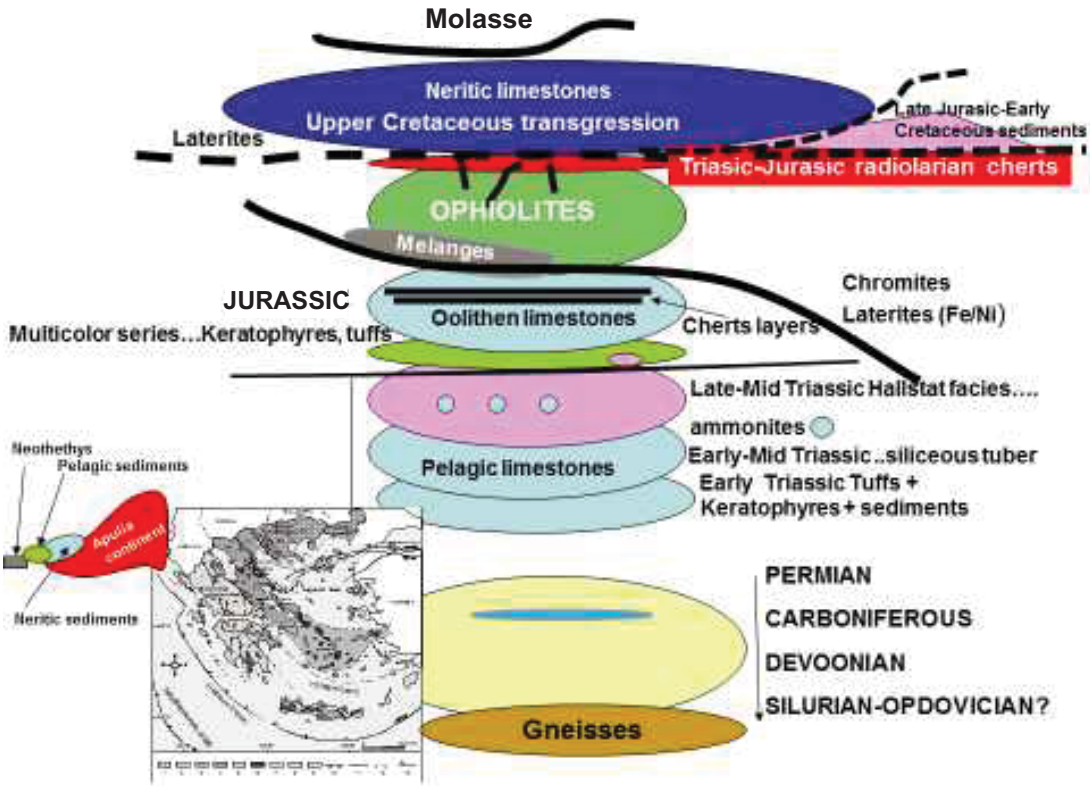


Figure 14
Schematic tectonostratigraphic column of the ?Sub-Pelagonian zone.

Burg et al. 1996, Kilias & Mountrakis 1990, Ricou et al. 1998, Bonev et al. 2006, Jahn-Awe et al. 2010, Burg 2012, Kilias et al. 2013, Froitzheim et al. 2014, Schenker et al. 2014, 2015). Nappes' stacking took place during alternating periods of continental crust segments collision and oceanic lithosphere subduction. Subsequently, extensional tectonics associated with detachment faults and ductile to brittle tectonics from Paleocene to Early Miocene times, led progressively to orogen collapse and exhumation of deep crustal levels. This extensional period was accompanied with syn- to post-tectonic granitoids' intrusions (fig. 21, 23; Dinter 1998, Kilias & Mountrakis 1998, Krohe and Mposkos 2002, Bonev et al. 2006, Krohe and Mposkos 2002).

The orogen collapse was related to a main top-to-the-SW sense of movement, at least for the Greek part of the SRB/RHD. On the other hand, the dominant thrusting direction is also detected as W- to SW-ward, although the direction of compression (thrusting) related shearing is fairly difficult to be distinguished from the direction of the corresponding extension (collapse), and the issue still remains under debate (Kilias & Mountrakis 1990, Dinter 1998, Bonev et al. 2006, Burg 2012, Kilias et al. 2013). In the Greek part, the Serbo-Macedonian/Rhodope nappe pile is composed of a series of units, which, from the tectonically lowermost to the tectonically higher one, are: the Rhodope Pangaion, Sidironero and Kimi units, and the Serbo-Macedonian Kerdylia and Vertiskos units.

The Pangaion unit consists of mica-schists' and marbles' intercalations, which in the higher levels develop into a thick marble sequence. The age of this, mainly carbonate, series is under debate. In recent works, it is appointed as Alpine (Brunn and Sokoutis 2007, Papanikolaou et al. 2004, Papanikolaou 2009, Jahn-Awe et al. 2010, Froitzheim et al. 2014,), although some older views suggest a Paleozoic age (Kronberg et al. 1970, Kronberg and Raith 1977). The overlying Sidironero and Kimi units are very heterogeneous, composed of mafic and ultramafic metaophiolites, amphibolites, ortho- and paragneisses and marbles.

The ages of the various magmatic products within the Rhodope units range from Paleozoic to Oligocene-Miocene (Kronberg et al. 1970, Dinter 1998, Marchev et al. 2005, 2004, 2013). Metamorphic paragenesis remnants of the ultrahigh- and high-pressure eclogite facies were described for

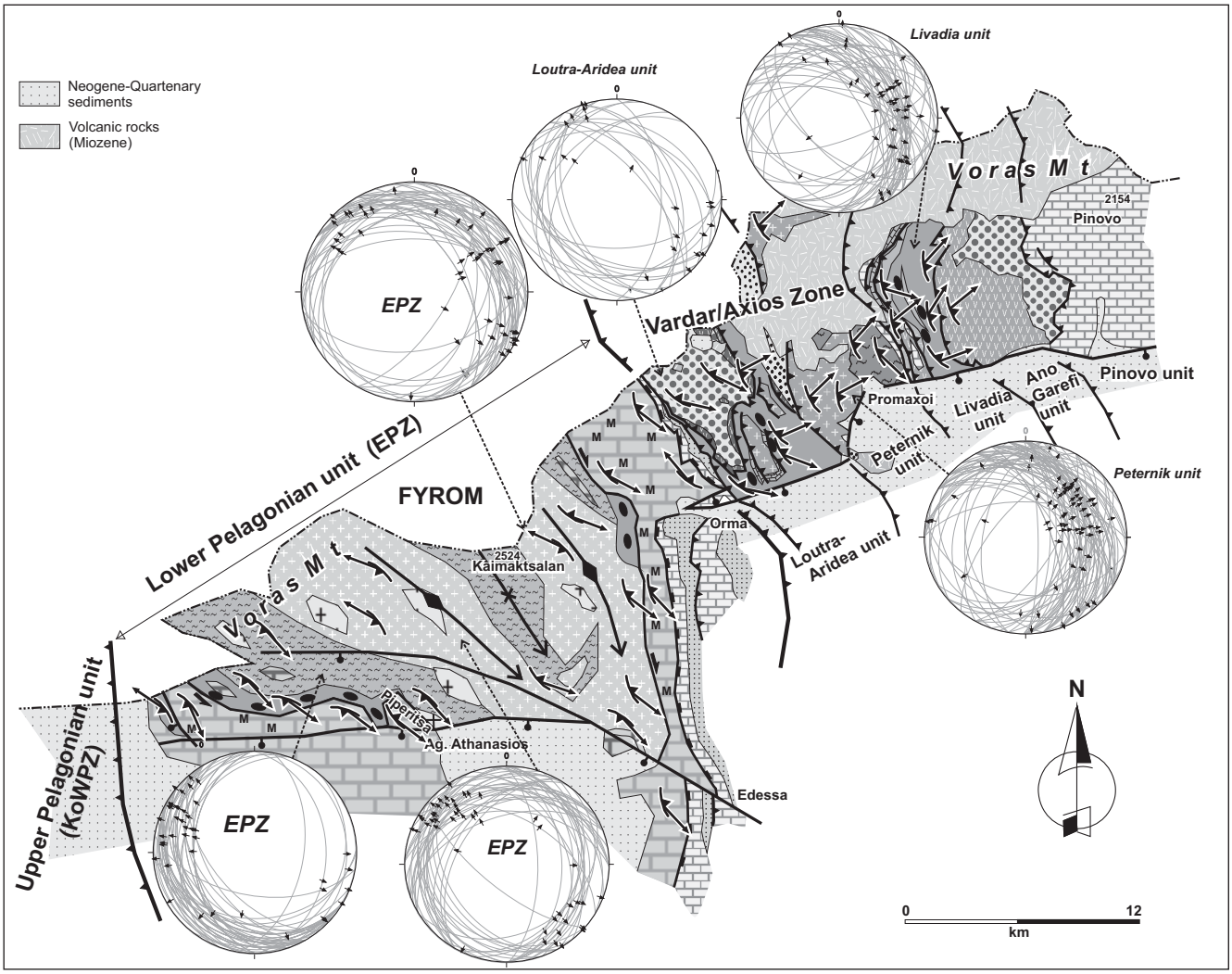


Figure 15a
Geological and structural map of the lower Pelagonian segment and the adjacent Almopia subzone (Axios/Vardar zone) in Voras Mountain. Schmidt diagrams (lower hemisphere) illustrate the orientation of the main foliation (S) and the associated stretching lineation (L) in each tectonic unit. Legend as in fig. 15b (Kilias et al. 2010).

the Sidironero and Kimi units, mainly preserved within basic bodies in high-temperature gneisses and amphibolites (Liati & Seidel 1996, Liati & Gebauer 1999, Mposkos & Kostopoulos 2001, Schmid et al. 2008). High-pressure conditions were related to subduction and compression, as well as nappe stacking. However, the ages of the ultrahigh- and high-pressure parageneses are not yet clear. Diamond-bearing ultrahigh-pressure rocks, with a possible Early Jurassic age suggested for the ultrahigh-pressure metamorphism, have been recognized in the Kimi unit (Mposkos & Kostopoulos 2001, Krenn et al. 2010, Schmidt et al. 2010, Nagel et al. 2011). On the other hand, Early Cretaceous (Wawrenitz & Mposkos 1997) and Late Cretaceous ages (Liati et al. 2002) have been also appointed for the ultrahigh- to high-pressure metamorphic event of the Kimi unit. In contrast to the latter dating though, the ultrahigh- to high-pressure metamorphism in the Kimi unit has been evolutionary related with the closure of the Paleotethys around 200 Ma ago (i.e. in the Early Jurassic) and thus represents the palaeogeographic location of the actual suture of the Paleotethyan ocean (Bauer et al. 2007, Nagel et al. 2011, Froiztheim et al. 2014).

Furthermore, high-pressure metamorphism and eclogites' formation, recorded within both Kimi and Sidironero units, have been dated between the Paleocene and the Eocene by using the U-Pb SHRIMP method on zircon (Liati & Gebauer 1999, Liati 2005) or the Lu-Hf method on whole rock and garnet fractions (Kirchenbauer et al. 2012). The latter high-pressure event was followed by a Paleocene to Eocene retrogressive high-temperature in the amphibolite facies metamorphism, partly reaching migmatization conditions, and it was identified again in both

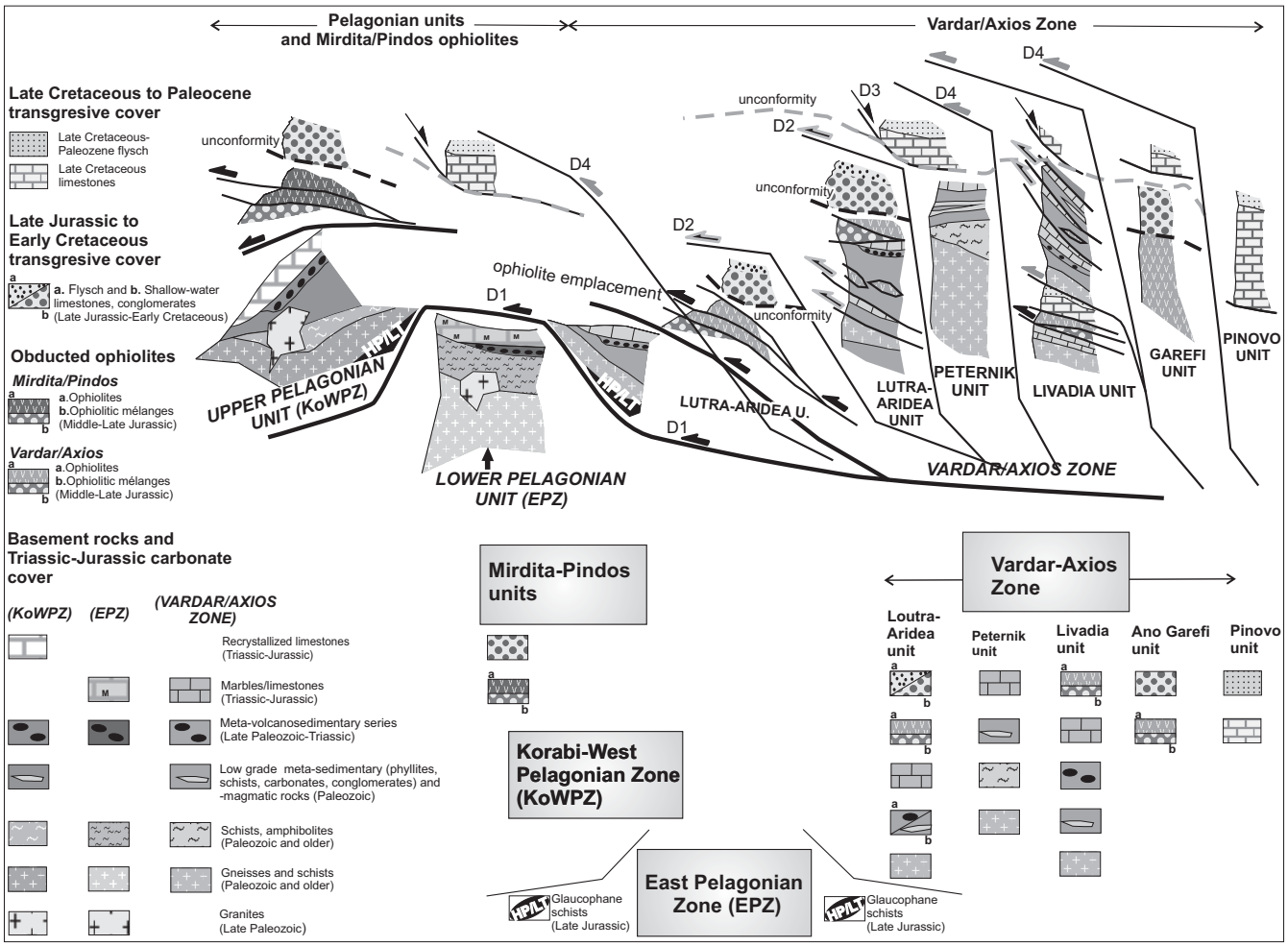


Figure 15b
Tectonostratigraphic columns (not in scale) of the imbricated Axios/Vardar units and Pelagonian nappe showing the sequence of deformation D1 to D4 (Kilias et al. 2010)

the Sidironero and Kimi units. Another Oligocene-Miocene low grade metamorphism has affected the lowermost Pangaion unit without any clear evidences for a previous high-pressure metamorphism. High- and low-grade metamorphism is related to the extension and collapse of the Serbo-Macedonian/Rhodope nappe stack, resulting in the gradual exhumation of the nappes from the top to the bottom. Initially, the Kimi unit was exhumed in the Paleocene-Eocene, followed by the Sidironero unit's exhumation in the Eocene-Oligocene, and finally that of the Pangaion unit in the Oligocene-Miocene (Krohe & Mposkos 2002, Bonev et al. 2006, Jahn-Awe et al. 2010, Kilias et al. 2013, 2014, Froitzheim et al. 2014).

According to our approach, the Serbo-Macedonian massif is an additional part of the nappe stack in the overall Serbo-Macedonian/Rhodope metamorphic province (fig. 20, 21). The Serbo-Macedonian massif is divided into the lower Kerdylia unit and the higher Vertiskos unit. At the boundary between the Vertiskos and Kimi units, metamorphic mafic and ultramafic rocks of oceanic lithosphere origin crop out tectonically, named as the Volvi ophiolite complex, (Dixon & Dimitriadis 1984, Papadopoulos & Kilias 1985).

The Kerdylia unit is consisted of gneisses and migmatites, as well as a thin marble layer of unknown age, which constitutes the highest lithostratigraphic horizon of the unit (Kockel et al. 1971, Jacobshagen et al. 1978, Himmerkus et al. 2007). The Kerdylia unit, placed on top of the western side of the Rhodope Pangaion unit, is probably equivalent to the Sidironero unit at the eastern side of the Pangaion unit, characterized by the same cooling age of Eocene-Oligocene times (fig. 20, 21; Wuethrich 2009). Moreover, dating of the granitoids and migmatites of the Kerdylia unit reveal ages of Jurassic and also of Permo-Carboniferous (Himmerkus et al. 2007), again the same as it is recorded for the Sidironero unit (Bauer et al. 2007).

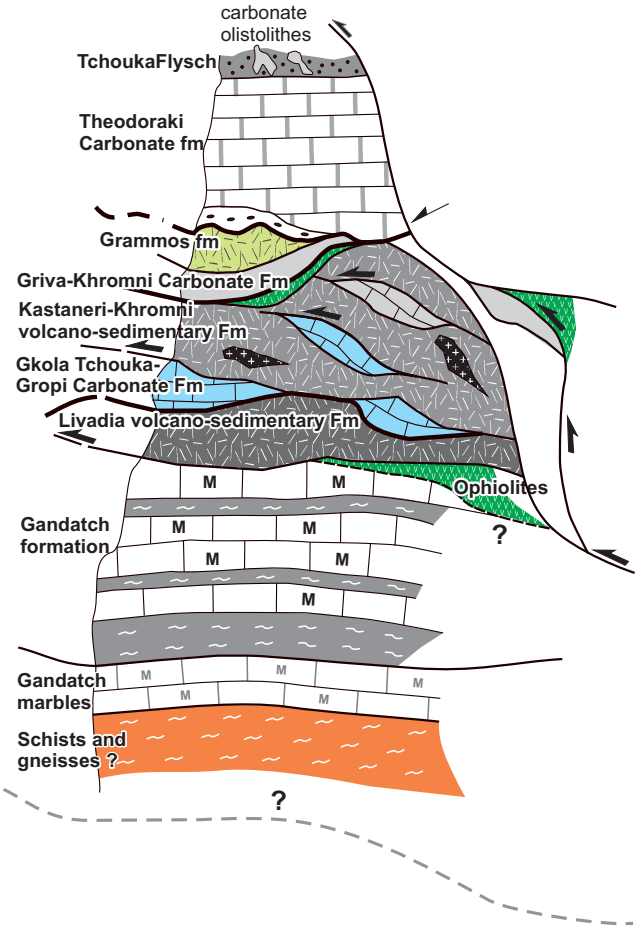


Figure 16
Tectonostratigraphy of the Paikon massif, showing also the progressive evolution of the several tectonic events affected the Paikon massif during the Alpine orogeny, from the Late Jurassic to the Cretaceous – Paleocene (Katrivanos et al. 2013).

The Vertiskos unit mainly comprises gneisses and schists of a Paleozoic age protolith, amphibolites, and in places thin Paleozoic marble layers. Migmatite rocks also occur in a great ratio in the lithostratigrapic composition of the Vertiskos unit (Chatzidimitriadis et al. 1985, Himmerkus et al. 2009a, Macheva et al. 2006,). A-type leucocratic granitoides (e.g. Arnea granite) of Early Triassic age (240 Ma) intrusions in the Paleozoic rock sequence as well as a bimodal volcanic activity of Permo-Triassic age have been formed during the continental rifting of the Pangaea supercontinent (fig. 12; Himmerkus et al. 2009b, Poli et al. 2009, Koroneos et al. 2013,). Furthermore, ophiolite bodies are also met, tectonically intercalated in between the gneissic and schist rocks.

In the Vertiskos unit, the same as in the Kimi unit, high-pressure to ultra-high-pressure diamond-bearing parageneses have been recorded (Dimitriadis & Condelitsas 1991, Kostopoulos et al. 2000). Their age once more remains under debate. It ranges from Paleozoic to Mesozoic according to the several different works on the high-pressure metamorphism of the Vertiskos unit (e.g. Dimitriadis & Gondelitsas 1991, Kostopoulos et al. 2000). Moreover, amphibolite to green schist facies metamorphic conditions of Jurassic-Cretaceous age have been also recognized for the metamorphic rocks of the Vertiskos unit, and indeed as the main metamorphic event of the Vertiskos rock sequence (Papadopoulos & Kilias 1985). The cooling of the Vertiskos unit has been dated as of Paleocene age (Wuethrich 2009).

Structurally, the Vertiskos unit occupies the same level with the Rhodope Kimi unit, and shares the same lithostratigraphic composition. Additionally, as analytically described above, it records a similar cooling age and tectono-metamorphic evolution to the Kimi unit (Krohe & Mposkos 2002, Wuethrich 2009). Therefore, both units could be regarded as parts of the same tectonic nappe in the Serbo-Macedonian/Rhodope nappe pile system (fig. 20, 21; Brunn & Sokoutis 2007, Wuethrich 2009, Kilias et al 2013, Froitzheim et al. 2014). Nevertheless, the somewhat older Paleocene cooling age of the Vertiskos unit show that

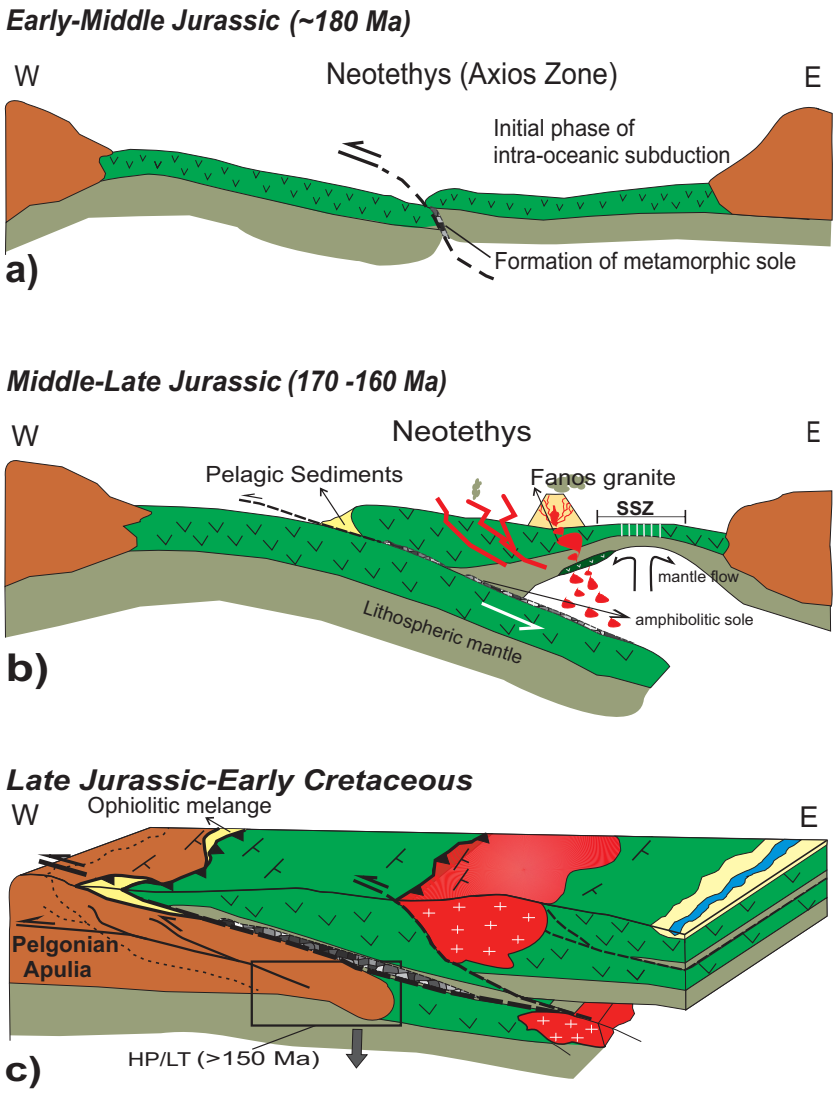


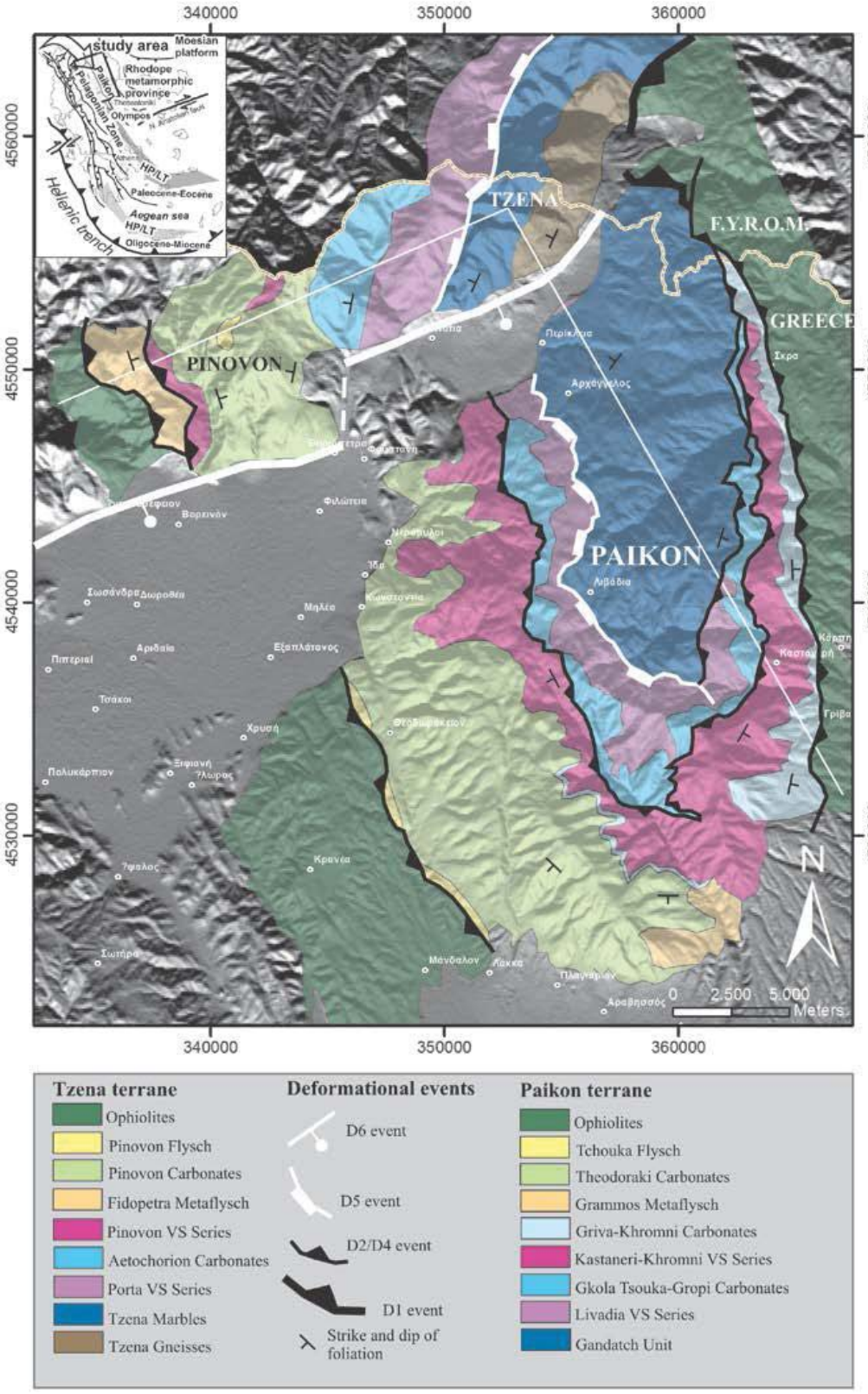
Figure 17
The Mid-Late Jurassic intra-oceanic subduction in the Neotethyan Axios/Vardar ocean related to an island-arc magmatism and the genesis of the back-arc Paeonia ocean basin. The island-arc magmatic products were emplaced together with the obducted ophiolites on the Pelagonian continental margin toward west (Michail et al. 2010).

it was at least one level higher than the Kimi unit (Burg 2012).

On top of the Vertiskos unit, at its western boundary, rests the Circum-Rhodope belt (CRB). It is an Alpine volcanosedimentary sequence, composed of deep-sea as well as of neritic Triassic-Jurassic sediments, intercalated with basic and acid-to-intermediate volcanic rocks (fig. 20, 21; Kockel et al. 1971, Kaufmann et al. 1976, Meinhold et al. 2009). Equivalent units to the Circum-Rhodope belt are the Alexandroupolis-Maronia unit in northeastern Greece, as well as the allochthonous Strandja and the Mandrica units in Bulgaria (Froitzheim et al. 2014). The CRB shows a blueschist facies metamorphism of Jurassic age, overprinted by greenschist facies metamorphic event during the Cretaceous (Michard et al. 1994).

Kockel et al. (1971) and Kaufmann et al. (1971) interpreted the CRB as the Alpine sedimentary cover of the continental margin of the Vertiskos unit towards the Axios/Vardar ocean basin. According to more recent works (Bonev & Stampfli 2003, 2008, Jahn-Awe et al 2010, Froitzheim et al 2014), the CRB and its equivalent units were tectonically placed over the Serbo-Macedonian/Rhodope nappe-stack with a main NE-ward emplacement direction during the Late Jurassic, forming the uppermost tectonic nappe of the Serbo-Macedonian/Rhodope nappes' pile. The volcanic rocks intercalated in the rock sequence of the CRB are related to an island-arc magmatic activity evolved in a Middle Jurassic intraoceanic subduction setting in the Neotethyan Axios/Vardar ocean. They emplaced on the Vertiskos continental margin together with the ophiolite rocks in the course of an arc-continent collision (e.g., Michard et al.

Figure 18
Geological map of the Paikon and Tzena units clearly showing their relationship. They are the same massif cut of the Aridea fault (D6 event; Katrivanos et al. 2016).



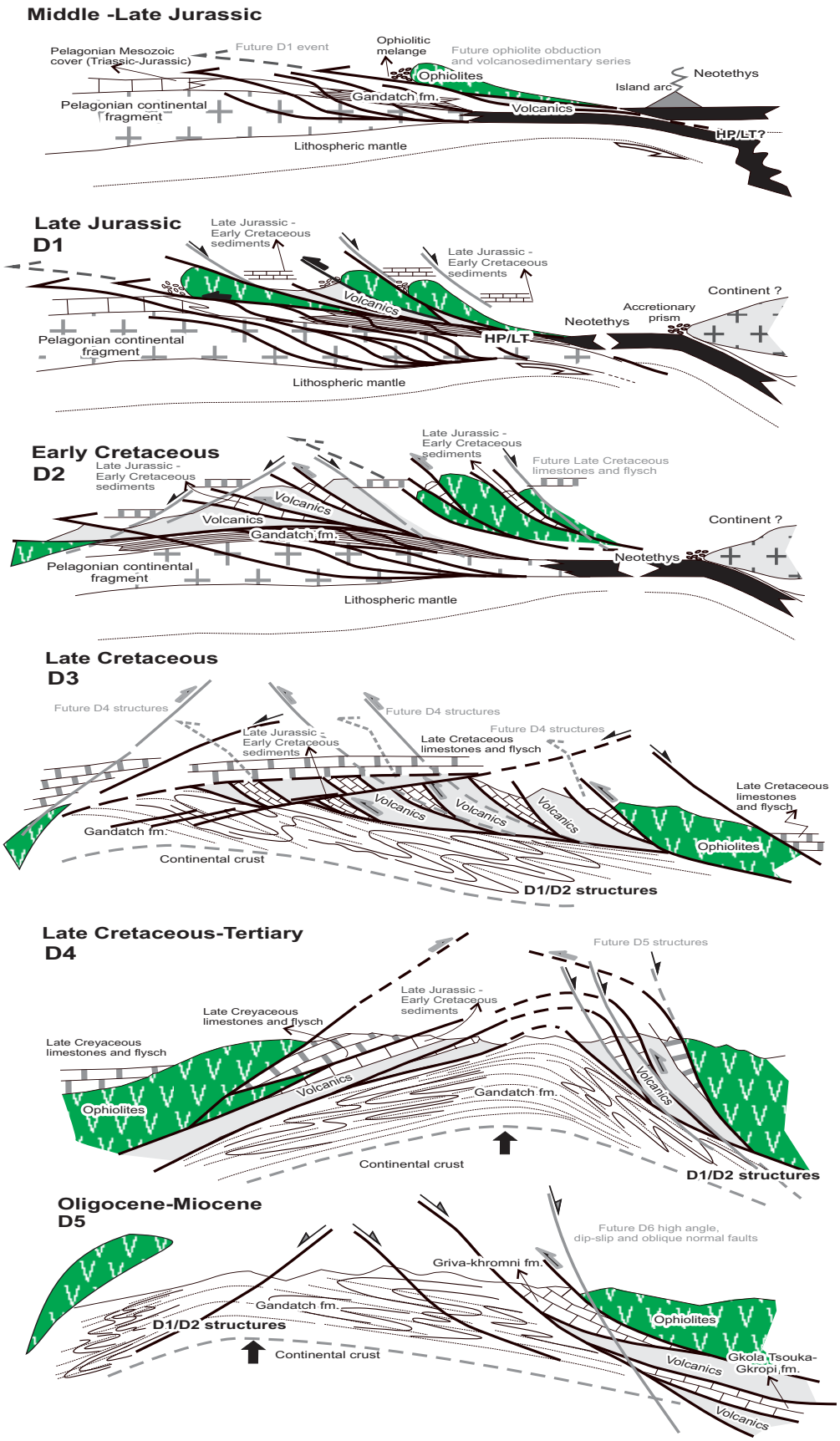


Figure 19
Schematic crustal scale transects indicating the geotectonic setting and structural evolution of the Paikon massif from the Mid-Late Jurassic to the Tertiary (D1 to D6) (Kilias et al. 2010, Katrivanos et al. 2013).

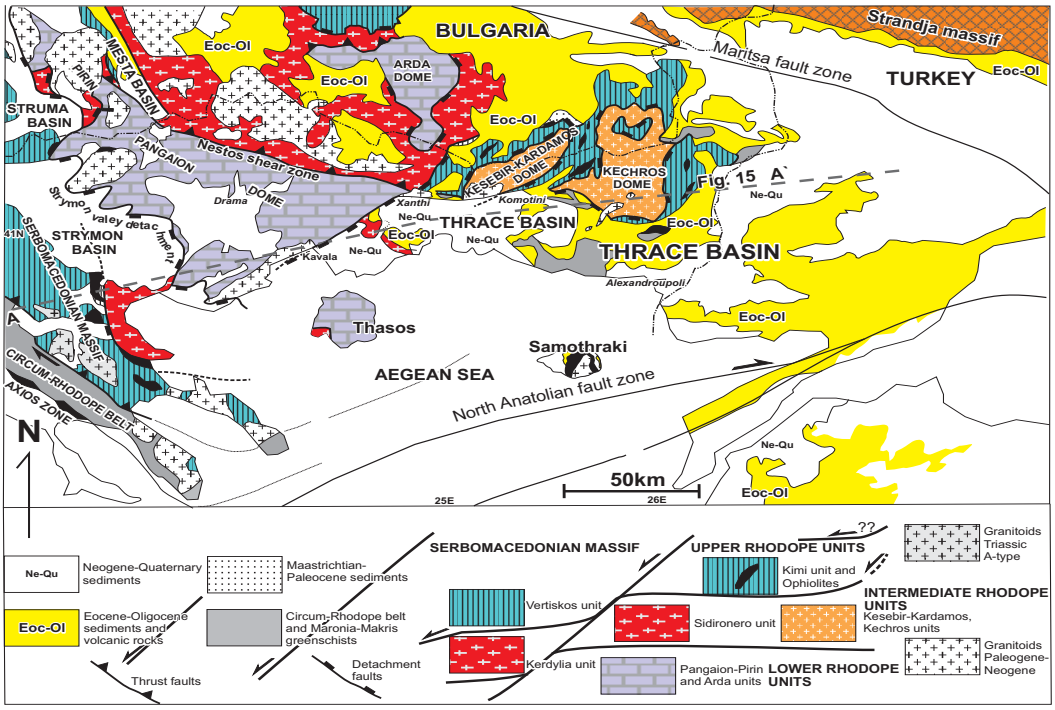


Figure 20
Simplified geological map of the SRB/RHD metamorphic province. The Thrace basin and the main structural units of the SRB and RHD massifs with their tectonic relationships, as well as the Sakarya and Strandja metamorphic rocks are shown (Kilias et al. 2013).

1998, Bonev & Stampfli 2003, 2008, Mainhold et al 2009, Michail et al 2016). Therefore, the evolution of the CRB and its equivalent units should be related to tectonic events that took place in the Neotethyan Axios/Vardar ocean realm.

Structural evolution

Based on our detailed structural studies on the Hellenides zones (e.g. Kilias et al. 2010, 2013, Katrivanos et al. 2013, Michail et al. 2016, and references therein), as well as all on recent works concerning the geological aspects of the Hellenides (i.e. Jolivet et al. 2004, Bonev et al. 2006, Jahn-Awe et al. 2010, Robertson 2012, Robertson et al. 2013, Bortolotti et al. 2013, Froitzheim et al. 2014, Schenker et al. 2014, 2015), we try in this chapter to give the main deformation architecture and geodynamic evolution of the Hellenides in the studied area during the Alpine orogeny. For the better understanding of the geodynamic evolution of the Hellenides, we consider important to initially provide here a brief summary of the recognized magmatic activity from the Paleozoic to the Neogene-Quaternary.

An Upper Paleozoic magmatism (~300 Ma) with calc-alkaline granitoids' intrusions within the Paleozoic basement units of the Internal Hellenides, related with subduction processes in the Paleotethys, has been widely described by several authors (fig. 12; Mountrakis 1986, Pe-Piper et al. 1993a, b, Koroneos et al. 1993, Anders et al. 2005, Saric et al. 2009, Koroneos et al. 2013). A-type granitoids and bimodal-type volcanic rocks of

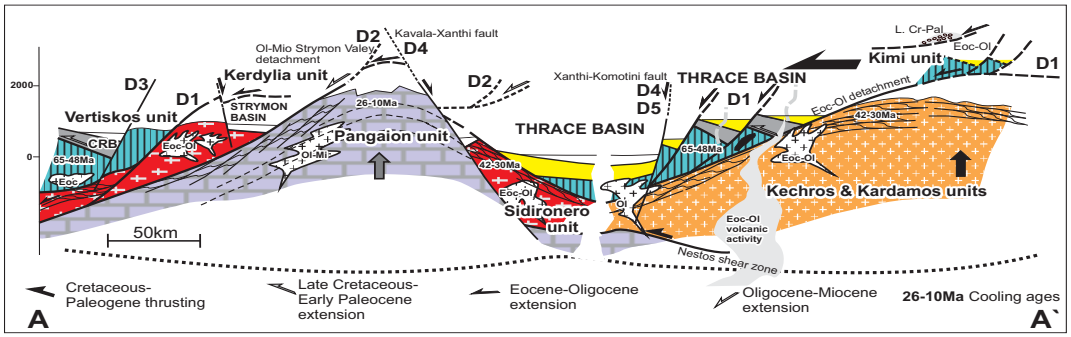


Figure 21
Geometry and kinematics of deformation during the Paleogene – Neogene extension of the Rhodope massif. Basin subsidence and extensional detachment fault system took place simultaneously with uplift and exhumation of deep crustal metamorphic units of the SRD/RHD. Extension migrates SW-SSW. Legend as in fig.20 (Kilias et al. 2013).

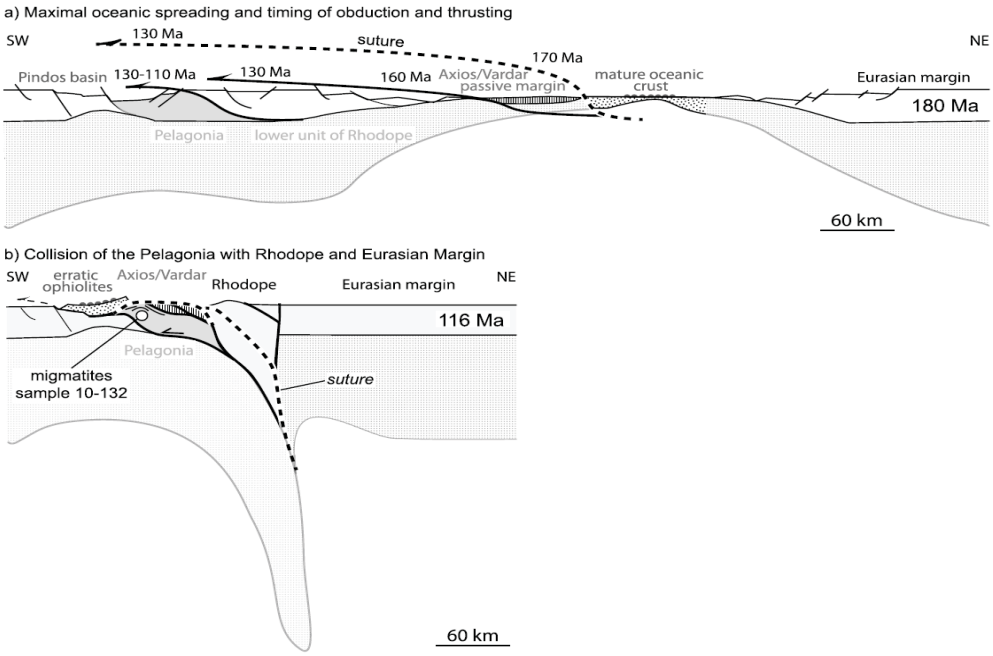


Figure 22
The structural evolution of the Neotethyan ocean from the Jurassic (intraoceanic subduction was followed by oceanic obduction on the Pelagonian continent) until the late Early Cretaceous collision of the Pelagonia with Rhodope and Eurasia margin (Schenker et al. 2014).

Early Triassic age (240 Ma) that have also intruded the Paleozoic basement units of the Internal Hellenides, have been interpreted as a magmatism related to the initial Permo-Triassic continental rifting of the Pangaea and the Neotethys ocean opening (fig. 12; Poli et al.2009, Koroneos et al. 2013). Late Jurassic arc-related magmatic activity with granitoids' intrusions, as well as volcanic products, have been recorded in the Axios/Vardar zone and the Serbo-Macedonian/Rhodope metamorphic province, cross cutting both ophiolites and continental basement rocks (fig. 17; Anders et al. 2005, Saric et al. 2009, Koroneos 2010, Katrivanos et al 2013, Froitzheim et al 2014, Michail et

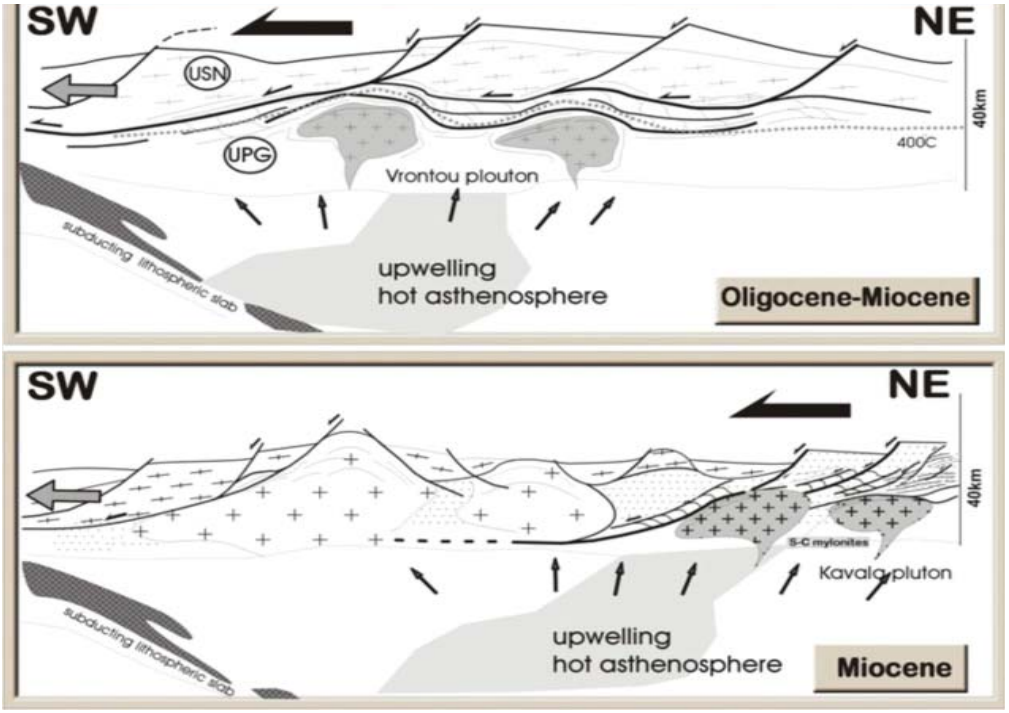


Figure 23
Schematic cross-sections through the Rhodope massif showing the relationships between syntectonicaly granitoid emplacement and continental crust subhorizontal extension and subvertical thinning (Kilias & Mountrakis 1998).

al 2016). Paleocene to Miocene magmatic activity has been mainly recognized in the Serbo-Macedonian/Rhodope province, related to subduction or crustal delaminate processes and normal detachment faulting and extension (fig. 20, 21, 23; Kilias & Mountrakis 1998, Marchev et al. 2005, 2013, Kilias et al. 2013). Finally, Neogene and Quaternary volcanic products are recorded in the broader Axios/Vardar zone and the active Hellenic volcanic arc (Soldatos 1955, Eleytheriades 1977, Fytikas et al. 1984, Vougioukalakis 2003).

I. Deformation and Kinematics

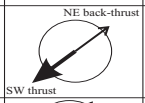
During the Alpine orogeny, the Hellenides were affected by a complicated multiphase deformation history and metamorphism, recorded in six main (D1-D6) deformational stages from the Middle Jurassic to present. In some places in the Paleozoic basement rocks of the Internal Hellenides, an older deformational event of Ercynian age is recognized, however strongly overprinted by the younger Alpine deformation (fig. 13, tables 1, 2; Kilias et al. 1999, 2010, 2013, 2014, Katrivanos et al. 2013).

In a general view, during the Alpine orogeny in the Hellenides, compression and nappe stacking alternated progressively through time with extension and orogenic collapse that was leading to exhumation of deep crustal levels. The conditions of the deformation were accordingly evolving from ductile to brittle (fig. 24, 25, 26; Kilias 2001, Kilias et al. 1994, 1999, 2001 Fassoulas et al. 1994). The kinematic pattern of extension and compression tectonics appears to be very complicated, but nevertheless for both stages, compressional and extensional, the recognized stretching lineation is roughly perpendicular to the Hellenic arc; that is NE-SW trending in the west and N-S trending in the center, with a main movement direction, at least for the compressional tectonics, SW- and S-ward, respectively (Kilias et al. 1991, 1999, 2002, Kilias 2001, Jolivet et al. 2004, Papanikolaou 2009, 2013, Jolivet and Brunn 2010). The sense of shear during the extensional stages of deformation and the nappes' collapse appears in many places bivergent, indicating an important component of bulk coaxial deformation during extension (fig. 24, 25; Kilias 1991, KIlias et al. 1994, Fassoulas et al. 1994, Hetzel et al. 1995, Kilias et al. 2002, Ring et al. 2010).

Table 1
Summarized the structural evolution of the Internal Hellenides during the Alpine orogeny

1. **Permo-Triassic**, continental rifting, bimodal magmatism and A-type granite intrusion.
2. **Triassic-Jurassic**, passive margins extension and sedimentation.
3. **Middle Jurassic**, intraoceanic subduction, amphibolite sole, ophiolite mélanges, island arc magmatism.
4. **Mid-Late Jurassic**, ophiolite obduction, high-pressure metamorphism, retrogression from greenschist to amphibolite facies conditions metamorphism, W-ward sense of movement and imbrication. E-ward sense of movement ??, **D1.**
   Deposition during extension? of the **Upper Jurassic-Lower Cretaceous** sediments above the obducted ophiolite belt or at the front of the obducted ophioites.
5. W-ward imbrication during the **Albian-Aptian (Early Cretaceous)**, syn-tectonic metamorphism, **D2.**
6. **Upper Cretaceous** extension, carbonate transgression terminated in the Paleocene internal Hellenides flysch, **D3.** Subduction of the Axios/Vardar ocean remnants under the Europa margin.
7. **Tertiary** compression **(Paleocene-Eocene)**, W-ward sense of movement, HP/LT metamorphism and building of the Internal Hellenides high pressure belt of Paleocene-Eocene age, progressively, emplacement of the Pelagonian nappe together with the HP/LT internal metamorphic belt on the External Hellenides, **D4**. Syn-orogenic extension in the Serbo-Macedonian/Rhodope metamorphic province.
8. HP/LT metamorphism and building of the External Hellenides high pressure belt of **Oligocene-Miocene** age associated with compression and nappe stacking in the External Hellenides. Syn-orogenic extension in the Internal Hellenides (**D5;** Olympos-Ossa, Kyklades and Serbo-Macedonian/Rhodope metamorphic province).
9. **Neogene-Quaternary**, active Hellenic subduction, extension and intramontagne basin formation, **D6**. Recent neo-tectonic activity.

Table 2
The main features of the tectonic events affected the Paikon massif and Pelagonian nappe from the Jurassic to Neogene (Kilias et al. 2010, Katrivanos et al. 2013).

EVENT (D)	AGE	STRESS REGIME & DEFORMATION CONDITIONS	KINEMATICS (main direction of movement)	STRUCTURES	GEODYNAMIC RESULTS	METAMORPHISM
$D_{HP/LT}$	MID-LATE JURASSIC	COMPRESSION DUCTILE	Unknown	Relicts or total destroyed structures	Subduction processes	High pressure / Low temperature
D_1	LATE JURASSIC	COMPRESSION DUCTILE		-Penetrative and synmetamorphic foliation (S_1) -Mineral streching lineation (L_1) -Isoclinal folds (F_1)	-Ophiolite obduction -Progressive nappe stacking -Crustal thickening	M_1: Greenschist facies -Bt, Chl, Act_1 Lws, high siliceous phengite, Na-ampibole
D_2	EARLY CRETACEOUS	COMPRESSION DUCTILE		-Foliation (S_2) parallel to S_1 and to axial plane of F_2 -Asymmetric tight to subisoclinal folds (F_2) -Mineral streching lineation (L_2)	-Imbrication of calcareous rocks with volcanosedimentary rocks and ophiolites -Series duplication -Continued and progressive nappe stacking	M2: Retrograde to M_1 -Bt to Chl -Act_1 to Act_2 -Wm to Ser
D_3	LATE CRETACEOUS	EXTENSION DISCRETE MYLONITIC SHEAR ZONES		-Mylonitic shear zones -Mylonitic foliation (S_3) -Mineral streching lineation (L_3)	-Basin formation -Sedimentation of Upper Cretaceous transgressive limestones and flysch -First orogen exhumation	-Qtz dynamic recrystallization & -Wm sericitization
D_4	Paleocene - Eocene	COMPRESSION BRITTLE		-Fold & Thrust belt with Reverse Faults and Folds	-Intense imbrication of all units -Ophiolite tectonic emplacement upon the west margin of Paikon massif as a back-thrusting	——
D_5	Oligocene - Miocene	EXTENSION BRITTLE		-Low angle normal faults	-Orogenic collapse and exhumation -Crustal thinning	——
D_6	Miocene - today	EXTENSION - TRANSTENSION BRITTLE	scattering of movement direction	-High angle, dip-slip to oblique normal faults and strike-slip faults	-Neogene to Quaternary basin evolution -Arideas fault zone reworking	——

D1 event (fig. 13, 27, 28, 29, 30)

The D1 is clearly imprinted on the Internal Hellenides (i.e. the Pelagonian nappe and the Axios/Vardar zone), as well as on the CRB and SRB (Mountrakis 1986, Vergely 1984, Kilias et al. 1999, 2010, Most et al. 2001, Katrivanos et al. 2013). No clear evidences for the D1 event have been recorded in the Rhodope metamorphic province, possibly due to the strong overprinting of the younger deformational event of the Internal Hellenides in this area.

The D1 event is of Late Jurassic age, characterized by a penetrative syn-metamorphic foliation (S1) associated with isoclinals folds (F1), which deforms a pre-existing foliation (S0). The F1-fold axes develop parallel to mineral stretching lineation (L1), and although they usually trend NE-SW, in some places they strike NW-SE (i.e. in the lower Pelagonian segment in Voras Mountain, fig. 15a; Kilias et al. 2010, Most et al. 2001). There are also places where the F1-fold axes show a scattering of their trend from NE-SW to NW-SE, possibly due to a progressive rotation of these folds during the deformation (e.g. Paikon basement rocks, fig. 18, 31, B11; Katrivanos et al. 2013). No clear kinematic indicators of the D1 event have been preserved, because the D1-structures have been strongly overprinted by the subsequent tectonics. However, in few places where the D1 kinematics is recognized, a main sense of movement top-to-the-W may be reconstructed. Furthermore, wherever D1 deformation is still preserved, an important coaxial component of deformation is also recorded (Kilias et al. 1999, 2010, Katrivanos et al 2013). The syn-D1 mineral parageneses, also identified in the L1-stretching lineation, show metamorphic conditions from the greenschist to the amphibolite facies, from the structurally higher to the structurally deeper crustal levels respectively (Kilias et al. 2010).

High-pressure metamorphism in the Internal Hellenides, predating the D1, has been suspected in several works (Baroz et al. 1987, Michard et al. 1994, Mposkos and Kostopoulos 2001, Most et al. 2001, Kilias et al 2010, Krenn at al. 2010, Schmidt et al. 2010,), but nonetheless without any clear evidence due to the strong overprinting by the younger tectonics. The pre-D1 high-pressure metamorphic assemblages occur only as relics and they are clearly related to compression and subduction processes or tectonic overpressure and continental crustal sinking due to overloading and nappe stacking in an arc-continent collision setting. Pre-D1 high-pressure metamorphic assemblages have been dated of Mid-Late Jurassic age (Baroz et al. 1987, Michard et al. 1994, Most et al. 2001, Kilias et al. 2010).

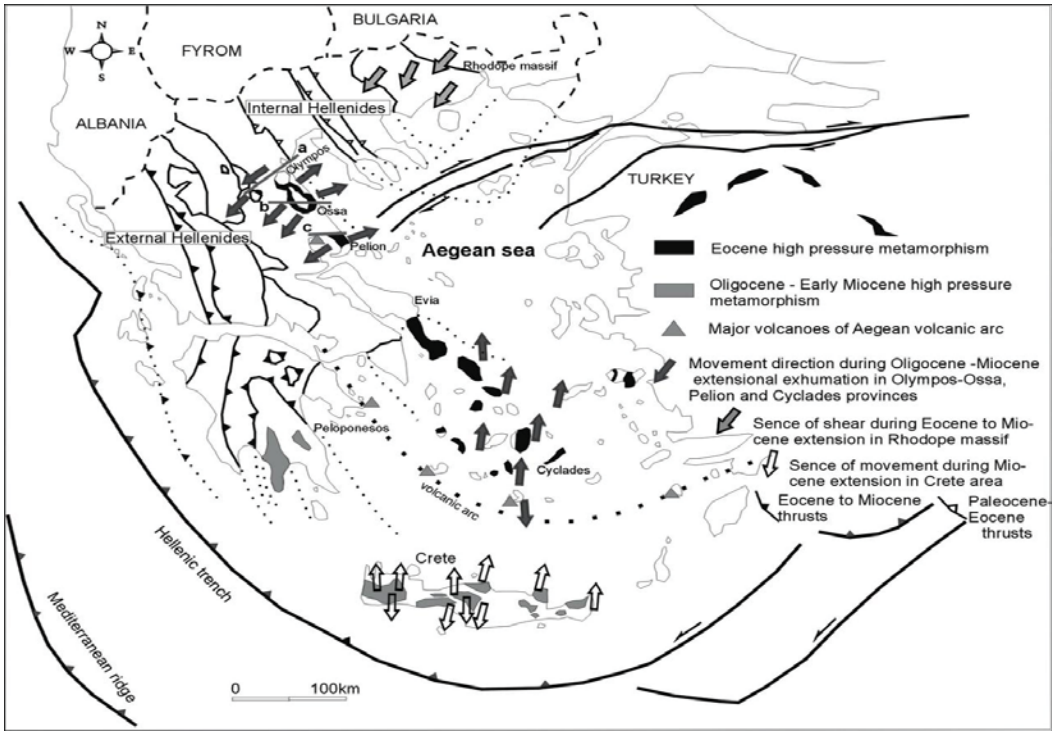


Figure 24
The two HP suture belts and the geometry of kinematics of Tertiary late-orogenic extension of the Hellenides (Kilias et al. 2002).

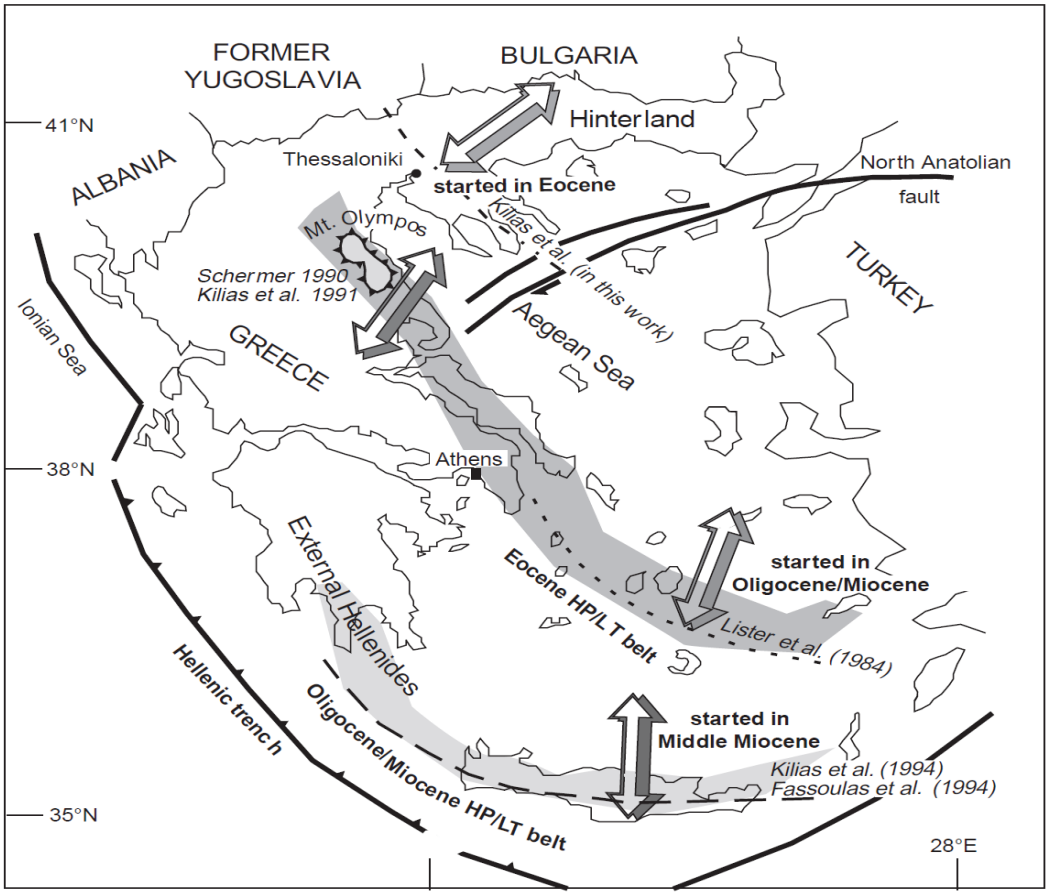


Figure 25
The SW-ward progressive migration of the extension during Tertiary time, from Internal Hellenides towards the External Hellenides. Extension and crustal exhumation take place simultaneously with compression and nappe stacking (Kilias et al. 1999).

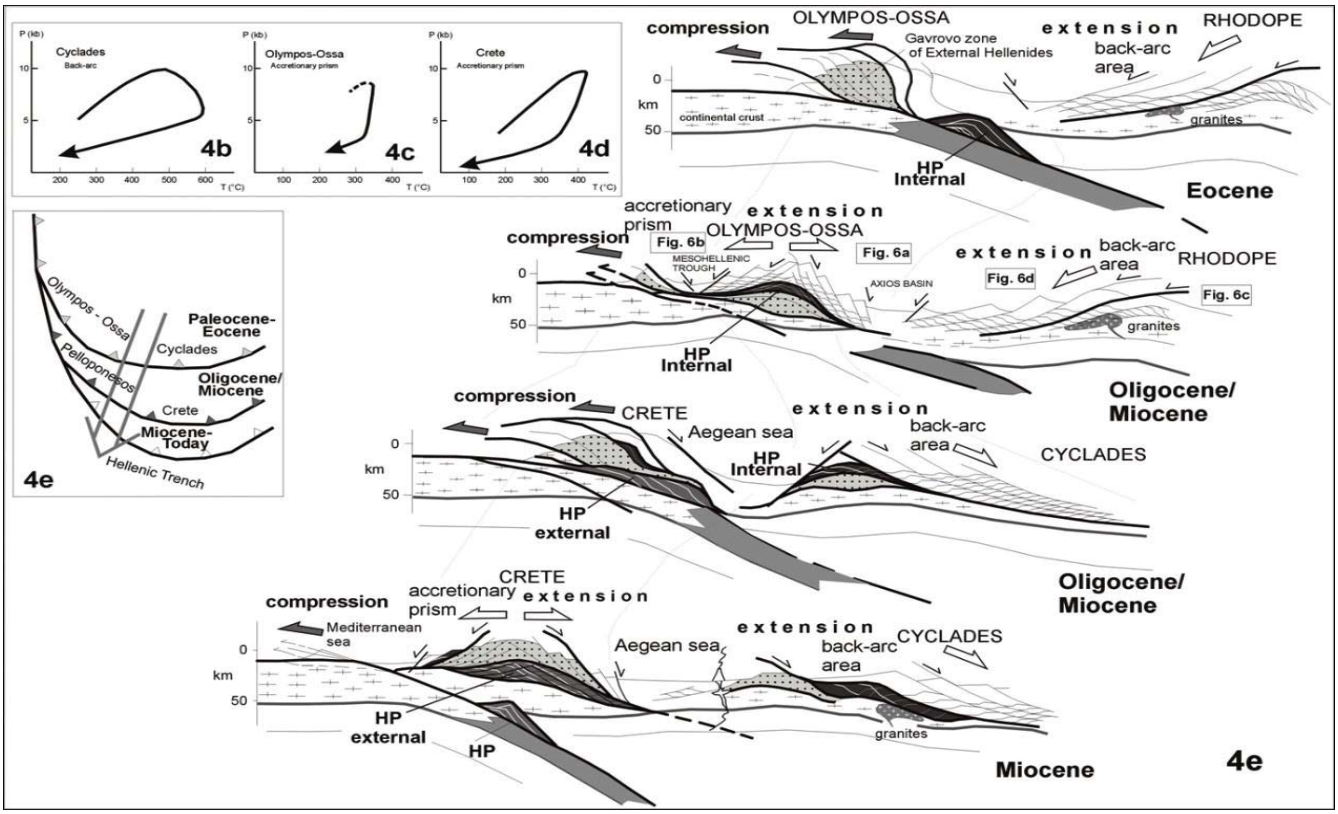


Figure 26
Schematic cross-sections illustrating the geometry and kinematic of deformation of the Tertiary syn-orogenic extension and associated compression in the Hellenides during different time periods. A SSW-ward propagation of the subduction processes and the associated compression is clearly remarkable. The representative P-T paths of the Tertiary metamorphism for the Cyclades, Olympos-Ossa, and Crete are also shown (Kilias et al. 2002).

D2 event (fig. 13, 27, 28, 29, 30)

The D1 and older structures are strongly overprinted by a younger, contractional, also ductile D2 deformation, recognized everywhere in the Internal Hellenides zones. The D2 took place until the Late Jurassic-Early Cretaceous, as can be inferred by same age sedimentary deposits that transgressively covered the obducted Neothethyan ophiolites. It is characterized by well-developed, in micro- and macro-scale, asymmetric, recumbent to overturned, tight to isoclinal folds (F2), refolding both the S1-folation and the isoclinal D1 folds. A new S2 foliation was developed parallel to the F2 folds' planes, forming in places a well-recognized crenulation cleavage with the S1 foliation. Usually though, the S1 is rotated till parallelized to S2, and as a result the two foliations coincide so that only one foliation can be observed.

The F1 and F2 folds axes are developed mostly parallel to each other, while in some places the F2 folds show even the same scattering of their trend with the F1 folds, from NW-SE to NE-SW. We attribute this scattering of the trend of the F2 folds also to rotation towards the stretching direction that took place during the D2 deformation. The latter stretching direction is E-W-ward to NE-SW-ward, as it is concluded by the development of the L2 mineral stretching lineation (i.e. defined by syn-D2 metamorphic parageneses) on the S2 planes. The D2 event has strongly affected all pre-Late Cretaceous units, also including the Paleozoic basement rocks (Kilias et al. 2010, Schenker et al. 2014, 2015). The sense of shear during the D2 is mainly top-to-the-W to SW. Opposite sense of shear is recognized in some cases, indicating the existence of a coaxial component of deformation during D2, as also mentioned during the D1. M2-metamorphic mineral parageneses show metamorphic conditions in the greenschist facies. The D2 event has been dated as late Early Cretaceous (i.e. ~110-100Ma; Most et al. 2001, Kilias et al. 2010, Katrivanos et al. 2013, Schenker et al. 2014, 2015).

D3 event (fig. 13, 19, 27, 29)

The D3 is recognized in the Pelagonian nappe pile and the metamorphic units of Axios/Vardar zone (e.g. Peternik and Paikon basement rocks). It is related to extension and has been dated of early Late Cretaceous age (100-90 Ma; Most et al.

2001, Kilias et al. 2010). The D3 structures form discrete, mylonitic shear bands with a well-developed S3-mylonitic foliation dipping mainly towards NE. The associated L3-stretching lineation also plunges down dip to the NE. The D3 shear bands are usually characterized by dynamic recrystallization of quartz and also by growth of chlorite and sericite. The sense of shear during the D3 was mainly identified down-dip towards NE, although in few cases an opposite, again down-dip, towards SW sense of shear is also observed.

In contrast to the described structural setting of the D3 event in the Pelagonian and Axios/Vardar basement rocks, ongoing subduction processes of the Axios-Vardar oceanic lithosphere remnants under the European continental margin is revealed by Late Cretaceous volcanic-arc magmatism along the southwestern European continental margin and the upper Serbo-Macedonian/Rhodope units. Therefore, a Late Cretaceous contraction at the eastern-most area of the Internal Hellenides, simultaneously with the D3 extension is inferred.

D4 event (fig. 13, 19, 27, 29, 30)

Unlike to the D3 extensional structures, the D4 structures are mainly contractional. The D4 is related to semi-ductile up to brittle deformation, and intense folding and imbrication of the affected units of the External Hellenides (i.e. Pindos zone), as well as of the upper structural levels of the Pelagonian nappe pile and the Axios/Vardar zone. On the same time, ductile deformation takes place in the lower structural levels of the Pelagonian nappe, in the Internal Hellenides high-pressure belt, and in the Serbo-Macedonian/ Rhodope metamorphic province. The related ductile D4 structures are contractional for the lower tectonic nappes (i.e. Sidironero unit, Lower Pelagonian segment and internal high-pressure belt) and extensional for the upper-most structural nappes of the Serbo-Macedonian/Rhodope metamorphic province (i.e. Kimi and Vertiskos units).

The D4 deformation, both ductile and brittle, took place during the Paleocene-Eocene, as concluded by numerous isotopic dating ages and lithostratigraphic-structural relationships (Mercier 1968, Mountrakis 1986, Vergely 1984, Most et al. 2001, Brown & Robertson 2004, 2006, Kilias et al. 2010, 2013, 2014, Froitzheim et al. 2014,). The ductile D4 structures at the lowermost Pelagonian part, the Internal high-pressure belt in Olympos-Ossa area (including the Cyclades high-pressure belt) and the Rhodope Sidironero unit, are all associated to high-pressure metamorphic conditions of Paleocene-Eocene age. Furthermore, same age ductile extension and final exhumation of the upper Serbo-Macedonian/ Rhodope (i.e. Kimi unit) nappes took place under greenschist to amphibolites facies metamorphic conditions (Wijbrans and McDougall 1986, Schermer 1993, Schermer et al. 1990, Kilias et al. 1999, 2013, Krohe and Mposkos 2002, Bonev et al. 2006, Wuethrich 2009).

A main SW-ward sense of movement is recognized for all D4 structures, wherever they were observed, related either to contraction or extension. In some cases, back-thrusting towards the NE is also identified in the broader area of the Axios/Vardar zone (Mercier 1968, Vergely 1984, Brown & Robertson 2003, 2004, Kilias et al. 2010, 2013, Katrivanos et al. 2013).

D5 event (fig. 13, 19, 27, 30)

In the Internal Hellenides, the D5 structures are generally related to extensional tectonics and exhumation of deep structural units either in form of tectonic windows or metamorphic core complexes (e.g. Olympos window, Paikon window, Pangaion window; Kilias & Mountrakis 1990, Dinter and Royden 1993, Schermer 1993, Kilias 1997, Krohe & Mposkos 2002, Brunn and Sokoutis 2007, Burg 2012, Kilias et al. 2013, 2016, Katrivanos et al. 2013).

The D5 structures have been dated as of Oligocene-Miocene age and in the Internal Hellenides are characterized by normal detachment faults and mylonitic rocks formation. Brittle deformation is recorded at the structurally higher crustal levels, and ductile at the lower ones. Along our traverse, the sense of shear during the D5 event in the internal Hellenides is down-dip mainly towards SW. Sense of movements towards NE has been also observed in a few places (e.g. Olympos-Ossa window), indicating a bivergent orogenic crustal deformation during the D5 (Dinter and Royden 1993, Kilias 1997, Kilias et al. 1999, 2002, 2010, Sfeikos et al. 1991, Katrivanos et al. 2013).

On the contrary to the Internal Hellenides, a belt of intense folding and thrusting was developed in the External Hellenides at the same time (i.e. Oligocene-Miocene times), clearly related to compression and crustal stacking. An also SW-ward sense of movement is recognized by the thrust faults' development and the asymmetric verging of the folds. Thrust and fold structures trend mainly NW-SE. During the Oligocene-Miocene, the Gavrovo zone together with the Pindos

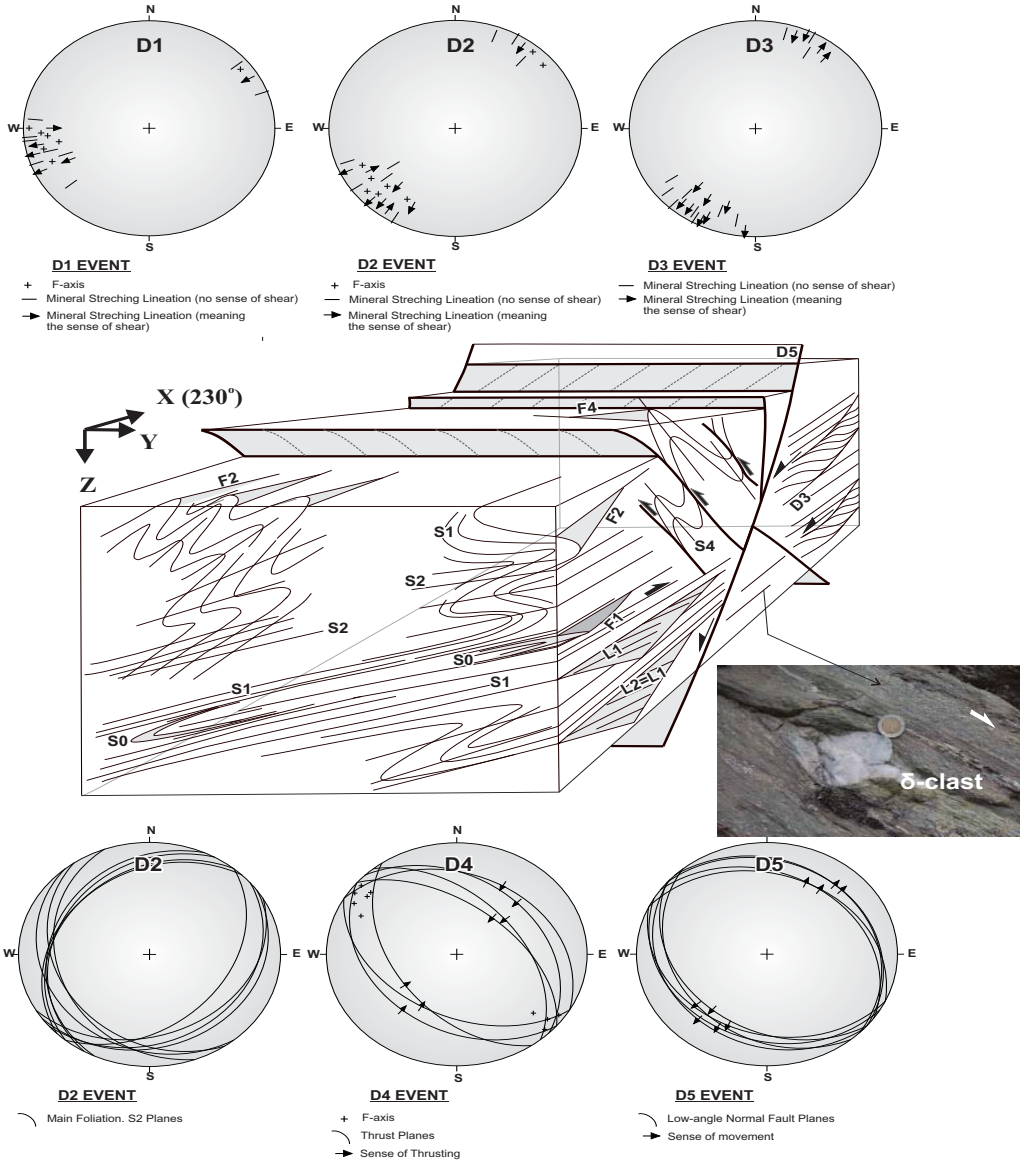


Figure 27
Schematic 3D diagram illustrating the architecture of deformation of the Paikon massif (subzone). The Schmidt diagrams show the orientation of the tectonic structures and the associated kinematics (Katrivanos et al. 2013).

nappe, which had already been thrusted over the Gavrovo zone during the Eocene (D4-event), and the Pelagonian nappe with the obducted Neotethyan ophiolites, were all thrusted towards SW over the Ionian zone, following the general SW-ern orogenic migration. The Pindos zone, possibly due to an out-of-sequence thrust fault, was again thrusted in places further to the west over the Ionian zone during the Miocene, covering completely the Gavrovo zone and its contact with the Ionian zone (Brunn 1956, Zouros 1993, Zouros et al. 1993).

D6 event (fig. 13, 18, 19, 31)

The D6 event is mainly related to extension in all the Hellenides' area, and took place under brittle conditions. The D6 structures overprint all the previously described structures and represent the final deformational stage of the orogen. At this stage, the compression and the orogenic evolution migrated into the active Hellenic subduction zone (fig.1, 2, 3, 24, 25; Papazachos & Delibasis 1969, 1997, Ring et al. 2010, Papanikolaou 2013).

The D6 structures are characterized by high-angle, both dip-slip and oblique-slip normal faults, as well as strike-slip faults, while some of them are related to the development of the Neogene-Quaternary basins in the Hellenides (Pavlides and Kilias 1987, Mountrakis 1987, Pavlides et al.1990, Pavlides & Mountrakis et al. 2006).

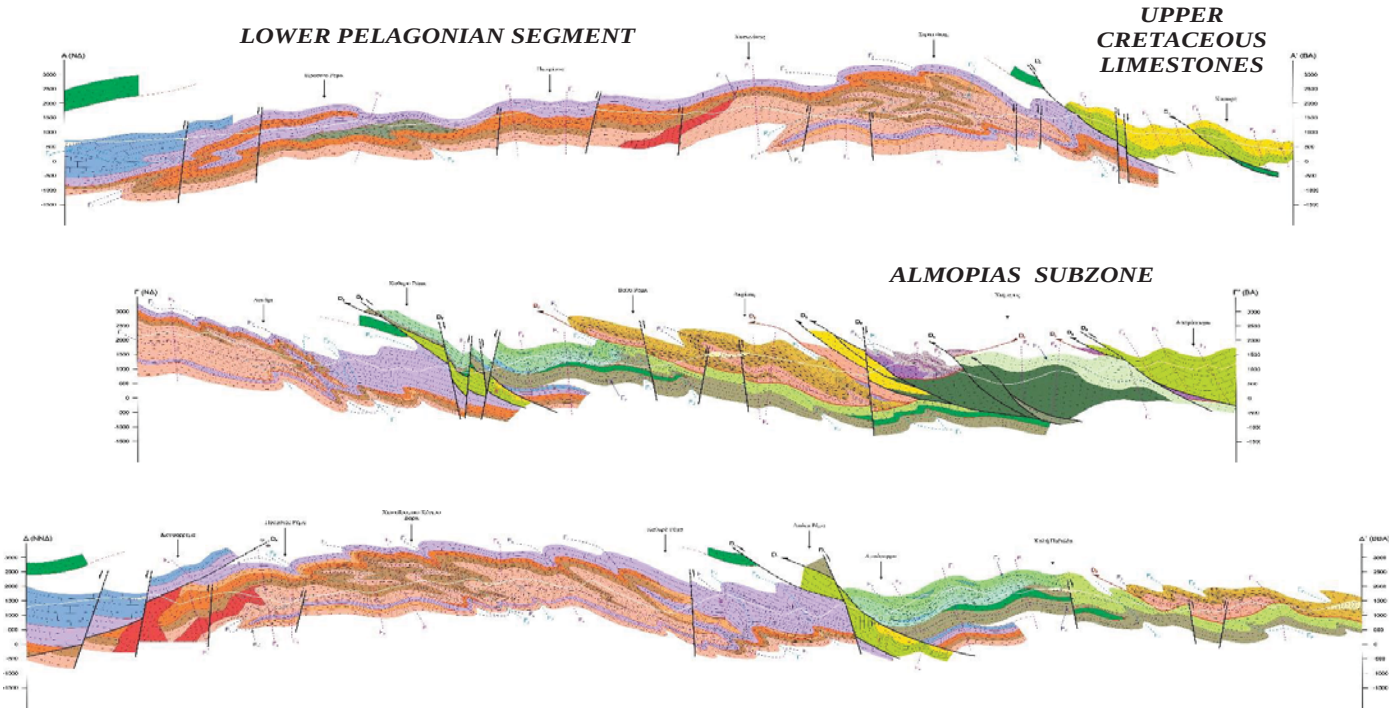


Figure 28
Representative geological cross-sections showing the geometry and kinematic of deformation of the Pelagonian nappe and the adjacent Axios/Vardar zone from the Jurassic to the recent (Avgerinas 2014).

Many of the D6-faults produced significant tectonostratigraphic gaps juxtaposing higher tectonic nappes against lower ones (fig. 31; Kilias et al. 2010, Katrivanos et al 2013). Furthermore, some of the D6 faults are still active, often associated with strong earthquakes (Pavlides et al. 1990, Tranos et al. 2003).

II. Geotectonic reconstruction of the Hellenides (fig. 3, 19, 21, 32a, 32b)

In this chapter, we attempt to summarize the several views concerning the Alpine structural evolution of the Hellenides, regarding at least the part along the proposed cross-section, included our perspective that derived by several works and accumulated experience on the deformational history of the broader area. We have to emphasize that, although a lot of scientific researches have been published about the structural evolution of the specific Hellenides' area, many questions still remain open, and hence the proposed reconstruction of the cross-section is in some respects hypothetical.

Continental rifting and opening of the Neotethys ocean basin start during the Permo-Triassic, possibly related to the closure of the Paleotethys further to the North. Continental rifting is associated with bimodal volcanic and neritic-clastic sediments deposition along the arisen continental margins, as well as A-type leucocratic granitoids intrusions in the Paleozoic basement rocks. The northeastern continental margin of the Apulia, including the Pelagonian continental domain, and the southwestern continental margin of the European continent were progressively formed, with the Neotethyan ocean basin opening in between the two margins. Carbonate platform to pelagic sediments were deposited from the Triassic to the Miocene along the continental margins of the European and Apulia plates (fig. 12; Mountrakis et al. 1983, Mountrakis 1986, Kilias & Mountrakis 1989, Pe-Piper et al. 1993 a, b, Schmid et al. 2008, Gawlick et al. 2008, 2009, Koroneos et al. 2013).

During the Mid Jurassic, part of the Neotethys ocean, the Maliac/Meliata ocean, subducted towards SE in an intraoceanic subduction setting (fig. 17, 19, 32; Schmid et al. 2008, Kilias et al. 2010, Katrivanos et al. 2013, Froitzheim et al. 2014, Schenker et al. 2014, 2015, Michail et al. 2016), which progressed in an arcuate northwestward-convex subduction zone (Schmid et al. 2008, Gawlick et al. 2008, Froitzheim et al. 2014). This geometry could be caused by a northwest-directed retreat of the subduction zone due to roll-back of the subducting slab (Bonev & Stampfli 2008,

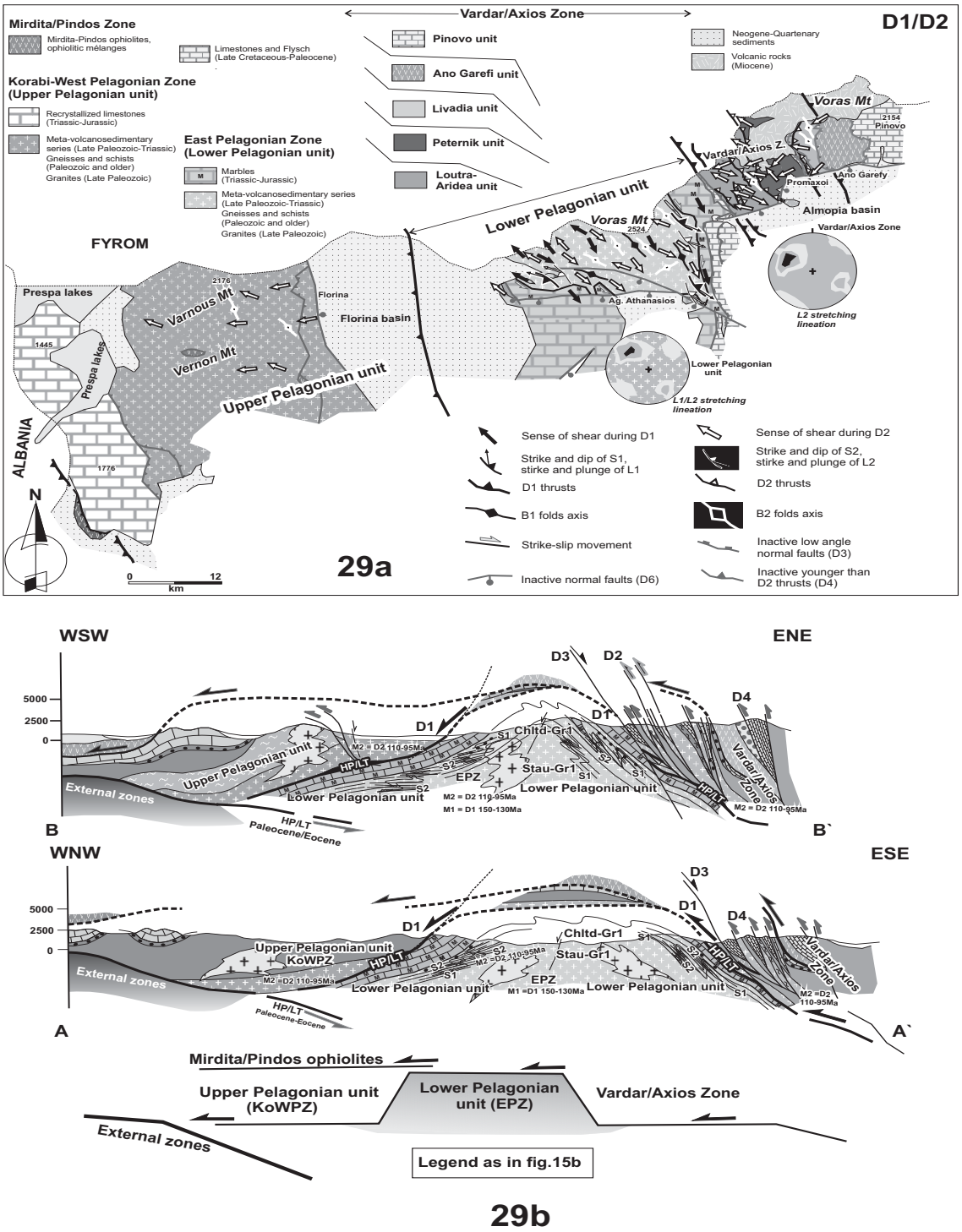


Figure 29

a. Geological – structural map of the Pelagonian nappe (Lower and Upper Pelagonian units) and the overthrusted on the Pelagonian Axios/Vardar zone units. The structural elements and the kinematics of deformation of the synmetamorphic D1 and D2 events are also shown. Schmidt diagrams (lower hemisphere) indicate the plunge direction of the associated L1 and L2 stretching lineations (Kilias et al. 2010). b. Geological cross-section (B-B') through the Pelagonian nappe and the Axios/Vardar zone in Northern Greece, showing structural relationships and stages of deformation during the Alpine orogeny in Internal Hellenides. The same structural architecture also dominates at the continuation of the Pelagonian nappe and the overthrusted Axios/Vardar zone into the FYROM area: A-A′ cross-section (Kilias et al. 2010). Legend as in fig.15b. For location of section see fig.11.

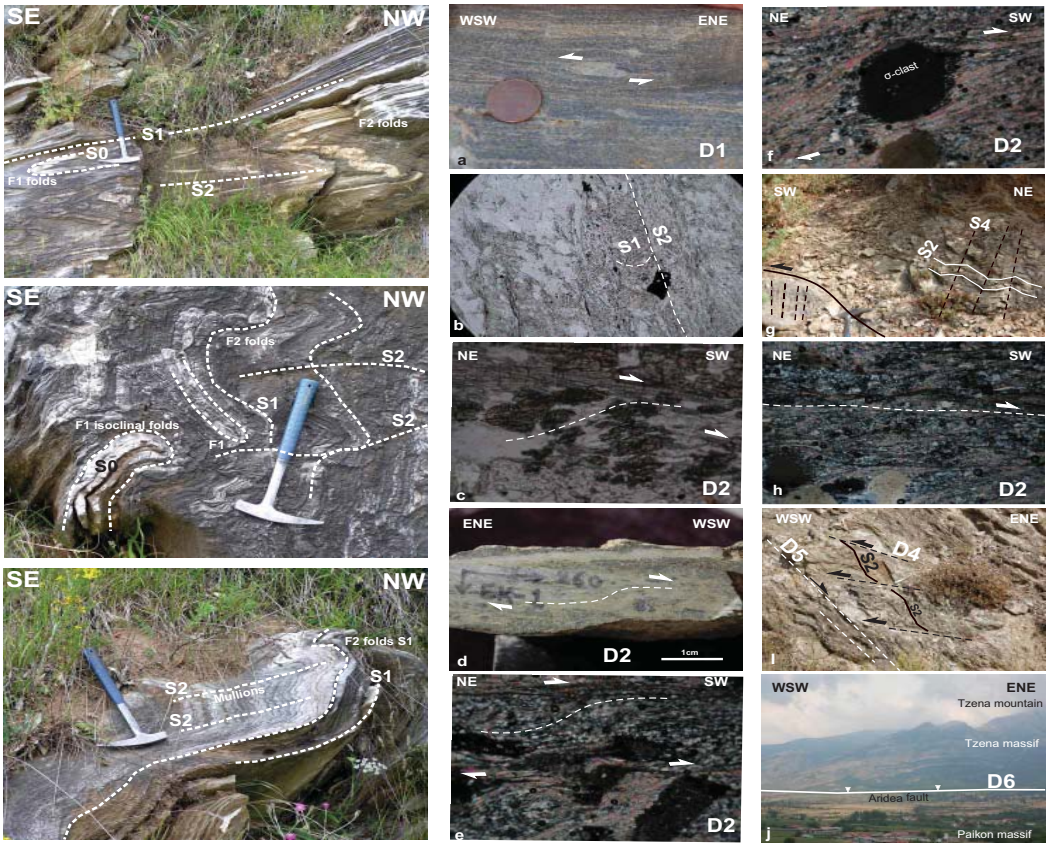


Figure 30
Micro-structures of deformation of the Paikon Massif and their relationships (Katrivanos et al. 2013).

Schmid et al. 2008). Formation of amphibolite sole of a 160-170Ma age (Spray & Roddick 1980, Trembley et al. 2015) and island-arc magmatism were related to this geotectonic setting (e.g. the Fanos granite, volcanics of the CRB, Paikon volcanoclastic series; Michard et al. 1998, Jahn-Awe et al. 2010, Katrivanos et al. 2013, Froitzcheim et al. 2014, Michail et al. 2016). At the front of the upper oceanic lithosphere, ophiolite mélanges associated with deep-water sediments were deposited, such as the Avdella mélange. Afterwards, they were strongly imbricated and folded, while the overriding upper oceanic lithosphere plate followed a main top-to-the-west sense of movement (John & Robertson 1991, Gawlick et al. 2008, Bortolloti et al. 2013).

During the intraoceanic subduction, parts of the Axios/Vardar ocean developed as a back-arc basin in a supra-subduction position (fig. 17; Schmid et al. 2008, Saccani et al. 2008, Botrtolloti et al.2013, Froitzheim et al 2014, Michail et al 2016). In the next stage, during the Late Jurassic, the island arc setting together with the Neotethyan ocean lithosphere were obducted towards West onto the Pelagonian continental margin (Gawlick et al. 2008, Kilias et al. 2010, Bortolloti et al. 2013, Froitzheim et al. 2014, Schenker et al. 2014, 2015) and also towards East, together with the CRB, onto the Vertiskos unit that consisted part of the European plate (Bonev & Stampfli 2006, Jahn-Awe et al. 2010, Froitzheim et al. 2014), due to the strong curvature of the subduction zone. Ophiolite mélanges deposits also accumulated in the basins developed along the front of the obducted ophiolite bodies.

The high-pressure metamorphism predating the D1 deformation is ascribed to this deformational stage. Parts of the continental margin and the arc formations were detached and buried beneath the advancing part of the oceanic lithosphere in response to the continental margin-trench collision, and were metamorphosed in high-pressure conditions (fig.19,32; Kilias et al. 2010, Robertson 2012, Katrivanos et al. 2013, Froitzheim et al. 2014).

During the D1 the high-pressure metamorphic rocks came up again to shallower crustal levels and were strongly deformed simultaneously with a greenschist– to amphibolite facies metamorphism, which is well exposed in the Pelagonian

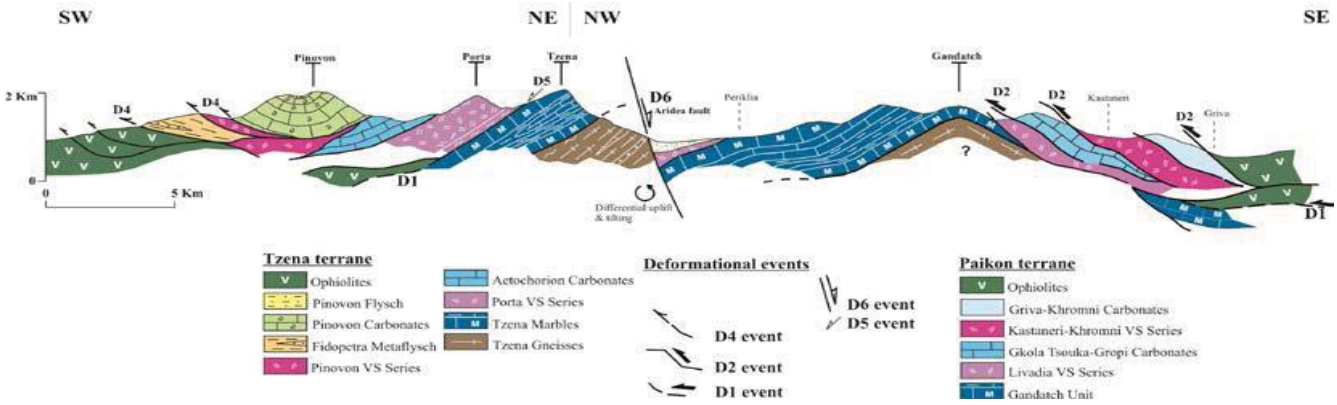


Figure 31
Schematic cross-section showing the geometry of deformation and the geological structure of the Paikon and Tzena massifs that are cut of the Aridea fault (D6 event; Katrivanos et al. 2013).

nappe, as well as in the Serbo-Macedonian massif. Furthermore, during the D1 and/or immediately after that, the accumulation of the Late Jurassic-Early Cretaceous sedimentary carbonate series took place, mainly on the top of the obducted ophiolite nappe, clearly determining the upper limit of the ophiolite emplacement (fig.19, 32). Basins' evolution where these sedimentary series were deposited is possibly related to extension and unroofing of the metamorphic rocks, followed by deposition of reworked carbonate material during the Late Jurassic – Early Cretaceous (Gawlick et al. 2008, Kostaki et al. 2013, 2014).

The depositional age of all these sediments, lying trangressively over the ophiolite rocks, is exactly the same in all ophiolite occurrences, both on the eastern and western margins of the Pelagonian nappe, as well as on the Vertiskos unit, indicating a simultaneous emplacement of the ophiolites on all continental margins (Galeos et al. 1994, Bortolotti et al. 2002, Carras et al. 2004, Gawlick et al. 2008, Mainhold et al. 2009, 2013, Kostaki et al. 2013, 2014). According to this evidence and additionally taking in account our structural works (Kilias et al. 2010, Katrivanos et al. 2013, Michail et al. 2016), as well as a lot of recent studies concerning the geodynamic evolution of Hellenides (Saccani et al. 2008, Jahn-Awe et al. 2010, Bortolotti et al. 2013, Schenker et al. 2014, 2015, Froitzheim et al. 2014), it is inferred that all ophiolites originated from a single source and this was the Neotethyan Axios/Vardar ocean basin; this in turn means that no other ocean basin than this existed, or at least there is no evidence proving so. Subsequently, the ophiolite nappes should be considered as far-travelled on the different continental parts, and this is where they were initially emplaced.

During the D2 in the late Early Cretaceous (~120-100Ma), ongoing plate convergence led further to outward W- to SW-vergent imbrication and folding of the whole Pelagonian nappe pile, including the obducted ophiolites, the Late Jurassic-Early Cretaceous sedimentary series, and the Paleozoic basement rocks, and it was generally related to a greenschist facies metamorphism (fig.19, 32; Kilias et al. 2010, Katrivanos et al. 2013, Kostaki et al. 2013, 2014, Schenker et al. 2014, 2015). On the contrary, an E- to NE-ward sense of movement has been identified for the eastern side of the Axios/Vardar ocean (Bonev & Stampfli 2008, Jahn-Awe et al. 2010, Froitzhein et al. 2014). Migmatites development in the Pelagonian basement during the late Early Cretaceous has been also recorded by Schenker et al. (2014, 2015), indicating a high grade temperature flow in places during the D2. The Beotian flysch, as well as some ophiolite mélanges occurrences, which were deposited at that time, should represent the sedimentary infill of the foreland basins formed by the D2 thrust sheets.

The crustal thickening of the late Early Cretaceous times (D2 event) was followed by an early Late Cretaceous extension (~90Ma; D3 event), which was associated with exhumation of deep crustal rocks and basins' subsidence wherein the Late Cretaceous-Paleocene carbonate series and the internal Hellenides' flysch were deposited. A stagnation of sedimentation was caused by the D2 structural activity, creating a stratigraphic gap between the Late Jurassic – Early Cretaceous sediments and the Late Cretaceous neritic sediments, with the latter lying discordantly over all the previous structures (fig.19, 32; Kilias et al, 2010, Katrivanos et al. 2013).

A continental break-up of Apulia took place locally during the D3 extension, leading to the formation of a small ocean basin in the late Cretaceous in the Cyclades area with a possible north-

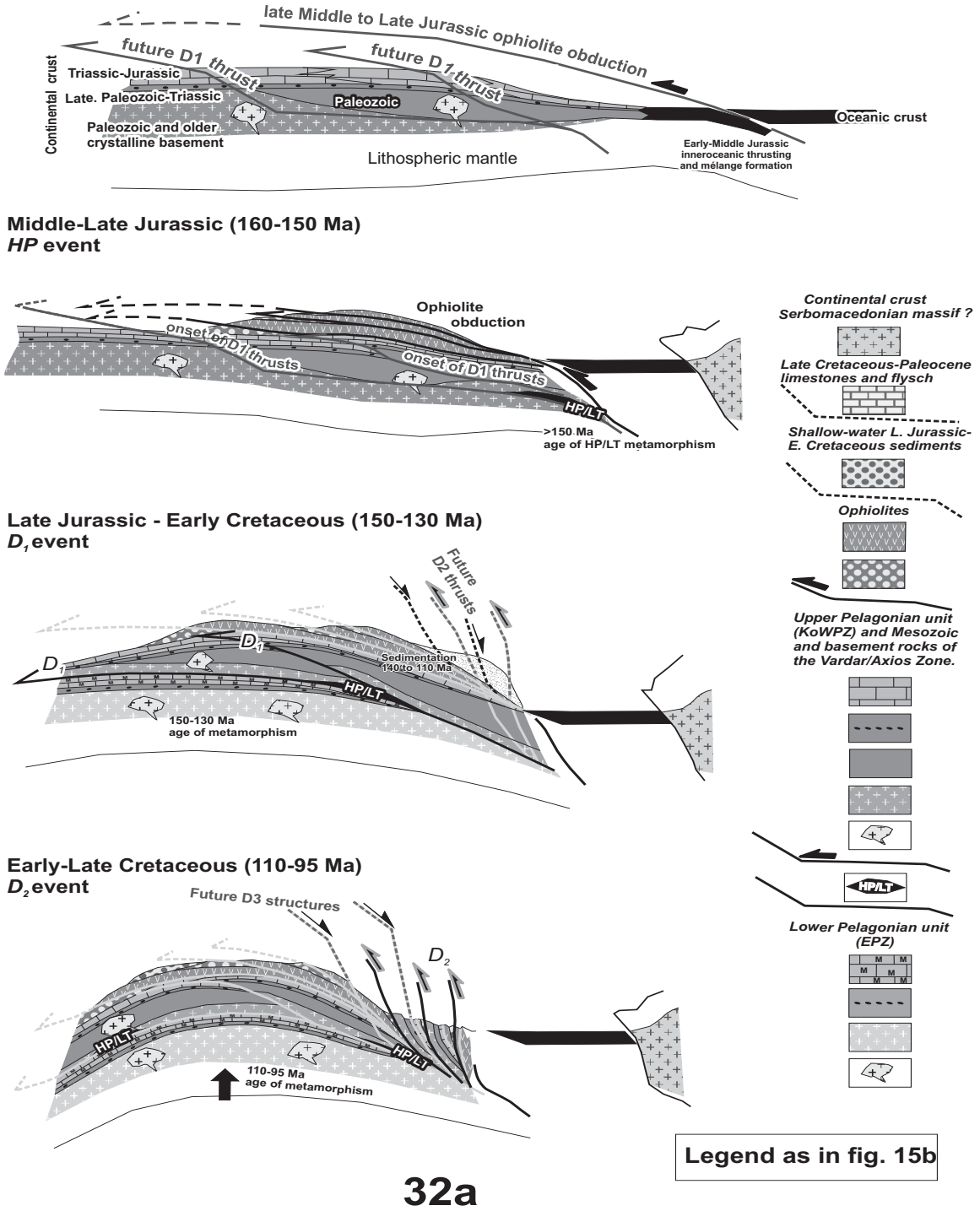
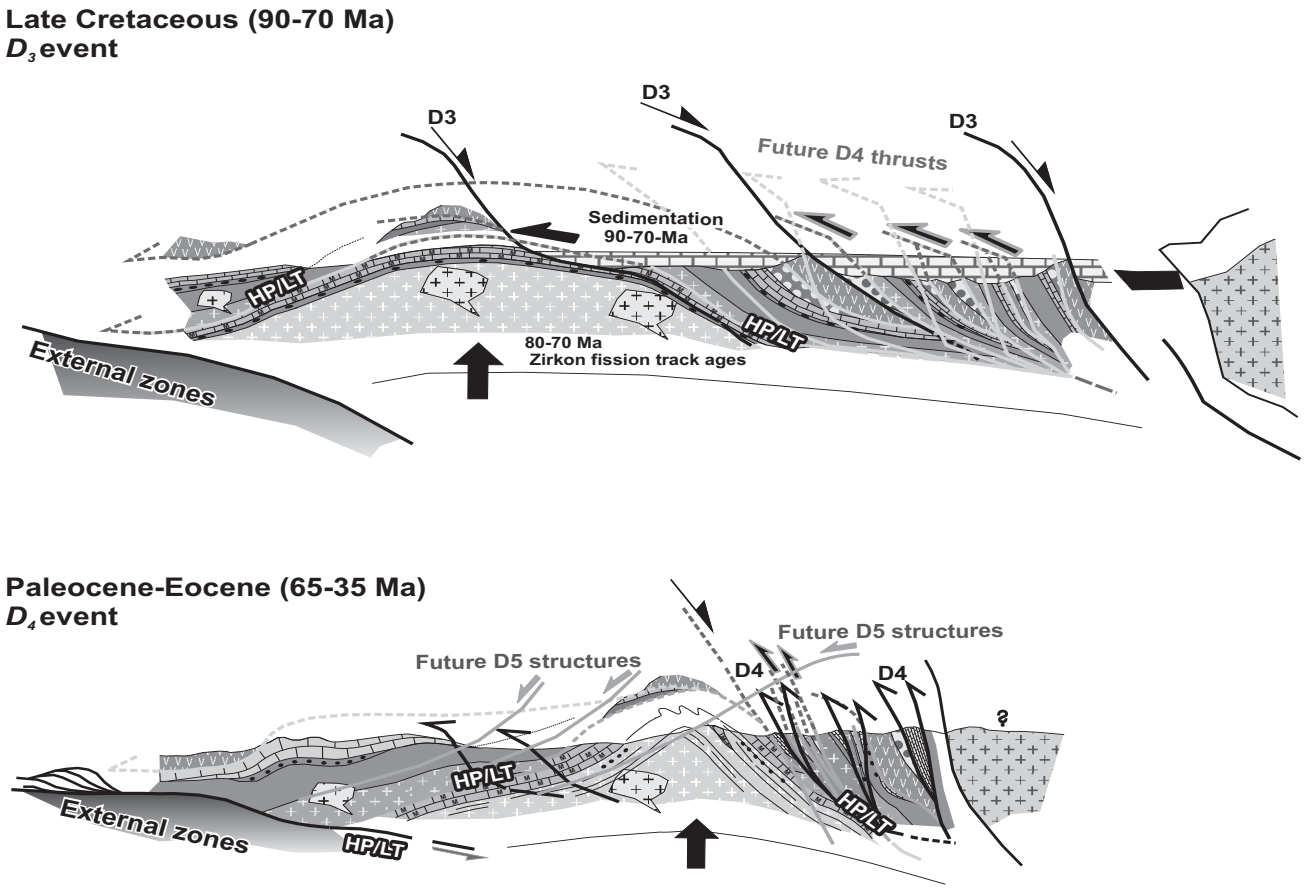


Figure 32
Schematic crustal-scale transects showing paleogeographic relation and structural evolution of the Pelagonian nappe and Axios/Vardar zone from the Mid-Late Jurassic until the Tertiary. a. Mid-Late Jurassic contraction and nappe stacking was associated with Late Jurassic – Early Cretaceous clastic and shallow-water carbonate sedimentation followed by late Early Cretaceous contraction and again nappe stacking. b. Early Late Cretaceous extension associated with basin formation and crustal unroofing was replaced by Paleocene – Eocene contraction and imbrication simultaneously with the A-type subduction of the External Hellenides under the Internal Hellenides in northern Greece or the subduction of a small Cretaceous ocean under the Pelagonian in the Cyclades area. Future Oligocene – Miocene extensional structures are also shown, leading to the unroofing and final exhumation of the External Hellenides in form of tectonic windows. Legend as in fig. 15b (Kilias et al. 2010).



Legend as in fig. 15b

32b

wards continuation to Pindos, which separated the Pelagonia from the Apulia (Papanikolaou 2009, 2013, Froitzheim et al. 2014). The extension and the ocean basin formation were associated with an about simultaneous new subduction of the Axios/Vardar ocean lithosphere remnants under the European continental margin (i.e. including the Serbo-Macedonian Vertiskos unit). This is clearly indicated by the Late Cretaceous magmatic arc that was developed along the European margin (i.e. Strandja massif and Sredna Gora zone in Bulgaria) above the subduction zone (fig. 2; Tomaschek 2003, Papanikolaou 2013, Kilias et al. 2013, Froitzheim et al. 2014).

After that, the D4 compressional event follows, which is particularly important since related to the final stages of the subduction processes in the Late Cretaceous, and the subsequent development, during the Paleocene-Eocene, of the SW-ward thrust and fold belt in the Internal Hellenides, mainly in the Axios/Vardar zone (fig.19, 32; Mercier 1968, Vergely 1984, Kilias et al. 2010, Katrivanos et al. 2013). During this structural stage, oceanic lithosphere and continental origin's segments, including the Late Jurassic calc-alkaline arc related granitoids, broke-off and by following the subducting ocean slab they finally accreted to the European active continental margin forming the Rhodope Sidironero unit and the related nappe stack (fig.22).

During the Paleocene, the entire Axios/Vardar ocean lithosphere was buried, the Axios/Vardar ocean basin closed, and the European active margin was collided with the Pelagonian (Kilias et al. 2010, 2013, Jahn-Awe et al. 2010, Robertson 2012, 2013, Froitzheim et al. 2014). During the D4 event, as compression and nappe stacking was advancing towards West through time, it was pro-

gressively followed by extension, nappes' collapse mainly towards SW, and gradual exhumation of the deeper crustal nappes associated with high-grade temperature metamorphism and migmatites creation, as well as new magmatic activity dated as of Paleocene-Eocene age (fig.3, 20, 21, 23; Kilias & Mountrakis 1998, Liati and Gebauer 1999, Christofides et al. 2001, Marchev et al. 2005, 2013, Mposkos & Krohe 2006). The magmatic activity was possibly triggered by the break-off of a deep-seated slab due to delamination or convective thinning of the previously overthickened lithosphere (Marchev et al. 2005, 2013). An upwelling of asthenoshpheric material could have possibly caused the Eocene high grade metamorphism and the high-K, calc-alkaline magmatism, as well as the migmatites formation (Marsev et al. 2005, 2013, Kirchenbauer et al. 2012).

Around the same time, during the Paleocene-Eocene, the Cyclades/Pindos? small ocean subducted under the Pelagonian continent, simultaneously with the A-Type subduction in Olympos-Ossa area further to the North (fig. 3,32; Schermer et al. 1990, Schermer 1993, Kilias 1997, Froitzheim et al. 2014). Further in the west, the External Pindos nappe was strongly imbricated, while some parts of it were detached and possibly driven also down under the Pelagonian nappes pile, together with the subducting lithospheric slab of the Cyclades ocean basin. The Internal Hellenides Paleocene-Eocene high-pressure belt was formed during these subduction processes (Schermer et al. 1990, Kilias et al. 1991, Schermer 1993, Kilias 1995). The same subduction zone may have been possibly extended up to the Rhodope province in the Northeast, as it is indicated by the Paleocene-Eocene high-pressure parageneses in the Sidironero unit rocks, at their tectonic contact with the underlying Pangaion unit (Liati & Gebauer 1999, Liati 2005, Kirchenbauer et al. 2012).

Summarizing, Paleocene-Eocene high-grade temperature metamorphism and migmatites formation in the Rhodope units, as well as magmatic activity (i.e. Sidironero unit), which were associated with extension and deep crustal exhumation, should have been related to the contemporaneously occurring subduction processes of the Cyclades ocean under the Pelagonian and the internal Hellenides nappes pile (fig.3, 25, 26; Kilias & Mountrakis 1998, Kilias et al. 1999, Krohe & Mposkos 2002, Kilias et al 2013). In this case, while Paleocene-Eocene compression and nappe stacking was advancing from the Axios/Vardar zone and the Olympos-Ossa-Cyclades area to the External Hellenides in the west, at the same time extension and tectonic denudation was taking place in the more eastern regions of the Serbo-Macedonia/Rhodope metamorphic province and the Internal Hellenides (fig.3, 25, 26, 32; Schermer 1993, Kilias et al. 1999, 2010, 2013, Bonev et al. 2006). As a continuation of the same process, during the Eocene-Oligocene, the Pelagonian and the entire (existing) Hellenides' nappe stack collided to Apulia, and together with the buried Cyclades/Pindos? basin formations and the internal Hellenides high-pressure belt were thrusted onto the platform sediments of the External Hellenides (i.e. Gavrovo unit), which constituted the passive continental margin of the Apulia (fig. 3, 32).

Regarding the above described architecture and structural evolution, the tectonic lower-most Rhodope unit should be the marginal part of the Apulia plate, detached from Apulia and subducted under the Internal Hellenides, and thus belonging to the External Hellenides. Therefore, it should be equivalent to the Olympos-Ossa carbonate unit. The higher, compared to the latter, metamorphic conditions recognized in the Rhodope Pangaion unit can be explained by the fact that the Pangaion unit corresponded to the deeper buried, towards East, parts of the same Apulia carbonate platform under the Internal Hellenides nappe stack.

Eocene-Oligocene nappe stacking and crustal thickening in the west were again associated with simultaneous extension at the structurally higher Hellenides' nappes and exhumation of parts of the Sidironero and Kerdylia units in the east (fig. 20, 21, 24, 25, 26). At the same time, new magmatic activity took place in the whole Serbo-Macedonia/Rhodope province (e.g. Vrondou, Elatia granites, volcanic rocks in between the molassic sediments of the Thrace basin etc).

During the Oligocene-Miocene (D5 event), when all the Internal Hellenides had already been collided with the Apulia, extension under brittle conditions took place in the structurally higher crustal levels, while ductile conditions dominated in the deeper levels, associated with low grade metamorphism and new magmatic activity (i.e. brittle conditions in the upper parts of the Pelagonian nappes pile, ductile conditions in the lower Rhodope Pangaion unit and the earlier formed during the Paleocene-Eocene high pressure metamorphic belt or else the lower-most parts of the Pelagonian nappe; fig. 3, 21, 32; Sfeikos et al. 1991, Kilias 1991, 1995, Kilias et al. 2002, 2013, 2010, Katrivanos et al. 2013). The former overthickened

lithospheric crust collapsed along normal detachment faults, which show a main top-to-the-SW sense of movement (e.g. Olympos detachment, Strymon-valley detachment). This collapse caused the formation of several tectonic windows and/or metamorphic core complexes, where the upper sedimentary series of the Apulia plate were exhumed from under the Internal Hellenides nappes (e.g. Olympos window, Paikon window, Pangaion metamorphic core complex; fig. 3, 19, 21, 32; Sokoutis et al. 1993, Brunn & Sokoutis 2007, Kilias et al. 2010, Kilias et al. 2013, 2014, Katrivanos et al. 2013). In the Olympos-Ossa area, the exhumation was very rapid and took place under an isothermal decompression path, while in the Rhodope province the exhumation followed a decompression path with an initially increasing temperature gradient (fig. 26; Schermer et al. 1990, Schermer 1990, Kilias et al. 1991, 2010, Dinter and Royden 1993, Dinter et al. 1995, Liati and Gebauer 1999).

A retreating subduction zone and roll-back of the subducted lithospheric slab under the Pelagonian and the other Internal Hellenides nappes stack could explain well the Oligocene-Miocene extension in the Internal Hellenides, which occurred contemporaneously with compression and outward-verging (i.e. westward-verging) nappes' stacking in the External Hellenides (Royden 1993, Ricou et al. 1998, Jolivet et al. 2004, Kilias et al. 2010, Ring et al. 2010, Frointzheim et al. 2014). For the Bulgarian part of the Rhodope province, a main down-dip NE-ward sense of movement has been recorded for the Oligocene-Miocene, indicating a bulk divergent extension in the Internal Hellenides and their continuation in Bulgaria. According to our recent works, a constrictional type deformation (i.e. compression along the Y-axis of the strain ellipsoid, Y<1) characterizes the described Tertiary tectonic activity in the Internal Hellenides' nappes (fig. 33; Kilias & Mountrakis 1990, 1998, Sfeikos et al. 1991, Kilias et al. 1999, 2013, 2014).

During the Neogene, an extensional regime dominates in the entire Hellenides belt, associated with high-angle normal and oblique strike-slip faults (D6 event), in some cases expressed by transpressional tectonics (Pavlidis et al. 1990, Tranos et al. 1999, Kilias et al. 2013), dismembering all pre-Neogene tectonic units and structures. In the present time, the convergence still migrates towards the SW, being directed in the active Hellenic subduction zone where the African plate is

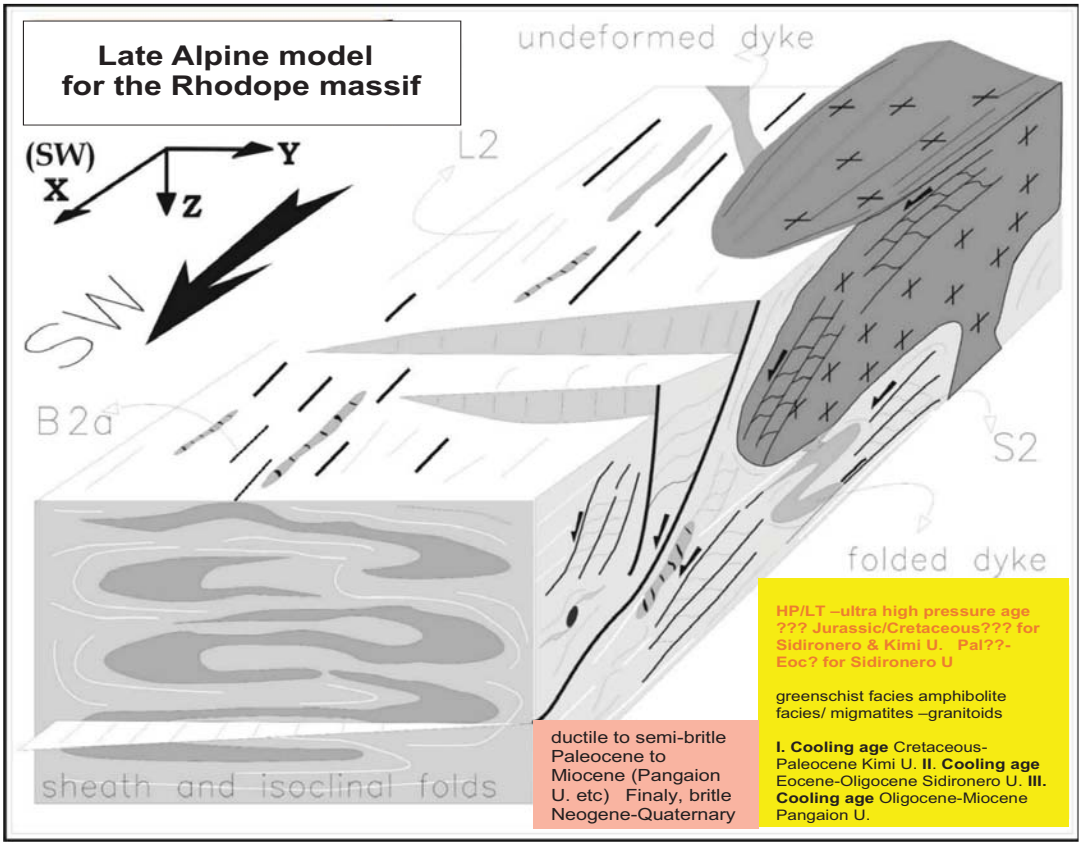


Figure 33
Schematic block diagram showing the Tertiary structure of the Rhodope massif, intruded by syntectonic granitoid bodies and characterized by a constructional type of deformation. Sense of shear top to the SW (Kilias & Mountrakis 1998).

Figure 34
The selected treck of each day during the entire field trip are shown with different colors on the geological map of Greece of a scale 1:500.000.

subducting under the Hellenides towards the NE, forming as a result the active Hellenic volcanic arc of the Aegean sea.

SECOND PART

FIELD TRIP ALONG THE CROSS-SECTION FROM PINDOS TO RHODOPE MOUNTAIN RANGES (EXTERNAL AND INTERNAL HELLENIDES).

Figure 34 shows the selected treks of each day on the geological map of Greece on a scale of 1:500.000. The locations of the selected outcrops, stops and samples discussed in the text are shown in separated figures for each day. The geological cross-sections for each day are also illustrated in detail in corresponding figures in the following.

A. 1st Day: Internal Hellenides (fig. 35)

I. Paeonia subzone
II. Circum-Rhodope belt
III. Serbo-Macedonian massif
IV. Rhodope massif

Trek: From the city of Thessaloniki through the city of Drama to the Sidironero village (fig. 36).

Overnight staying in Thessaloniki.

The first day's cross-section starts from the city of Thessaloniki and finishes at the village of Sidironero in the mountainous area of the Rhodope province. We initially follow the Thessaloniki ring-road till the exit with direction towards the village of Oraiokastro, so that we reach the village of Neochorouda on our way. After the village of Neochorouda, we turn back and take the Egnatia highway till the city of Lagadas, but then we continue via the old national road to the city of Drama (i.e. the Lagadas-Rentina-Drama road),

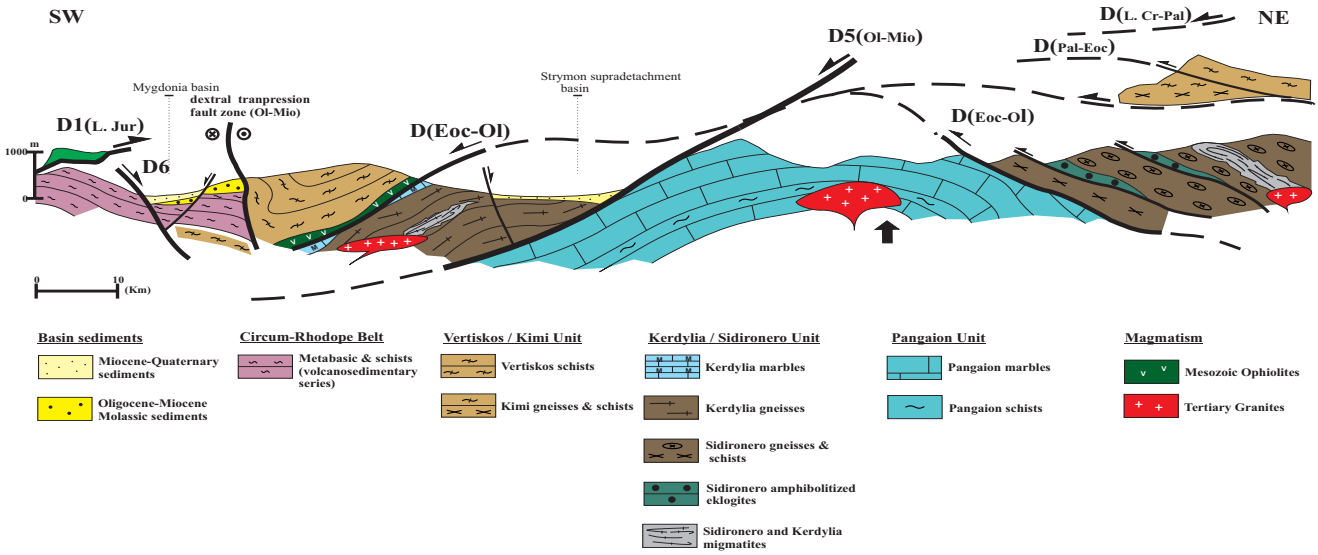


Figure 35
Schematic geological cross-section through the selected area/trek of the 1st day in the Serbo-Macedonian and Rhodope massifs, as well as the Circum-Rhodope belt and the eastern more part of the Paeonia subzone.

and onwards till our final destination, the village of Sidironero. Along this traverse, from the southwest to the northeast, we cut: the eastern-most part of the Paeonia subzone, the Circum Rhodope Belt, and the Serbo-Macedonian/Rhodope metamorphic province.

Along the ring-road of Thessaloniki, we firstly see a representative cross-section of the Chortiatis magmatic suite (fig. A1a), which is regarded as related to an island-arc magmatism of Mid-Late Jurassic age (Michard et al. 1998). It is composed of intercalations of intermediate and acid magmatic products with mainly shale deposits (fig. A1a). The Chortiatis magmatic suite is thrusted towards west over the Triassic-Jurassic Melissochori-Cholomontas turbidite unit of the CRB.

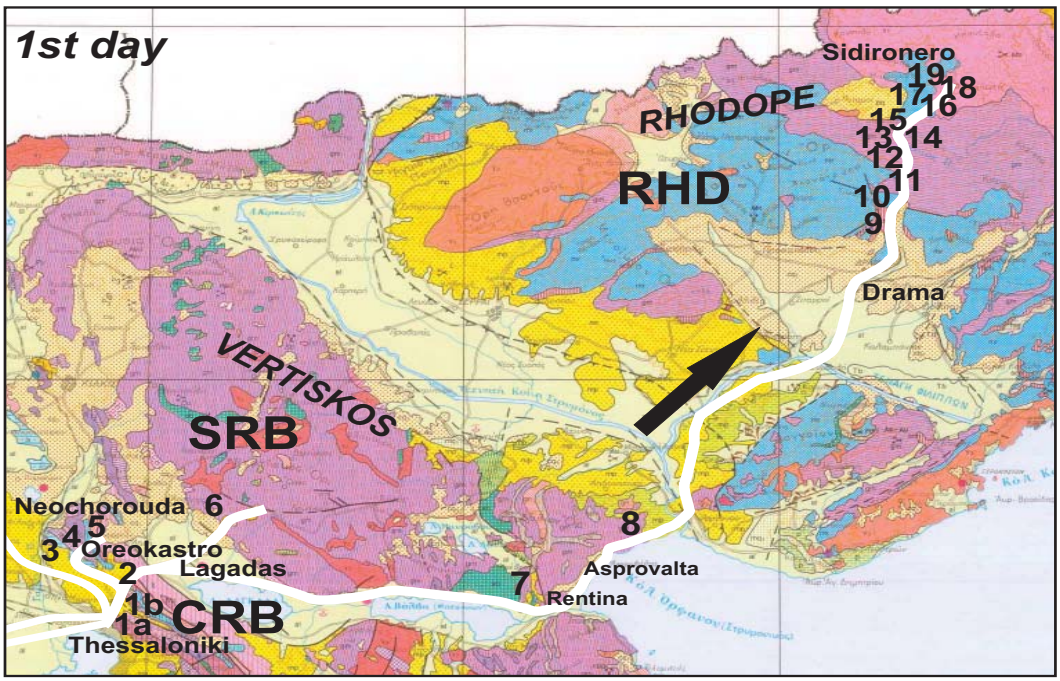


Figure 36
Location of the selected outcrops and stops are shown on the corresponding trek of the 1st day

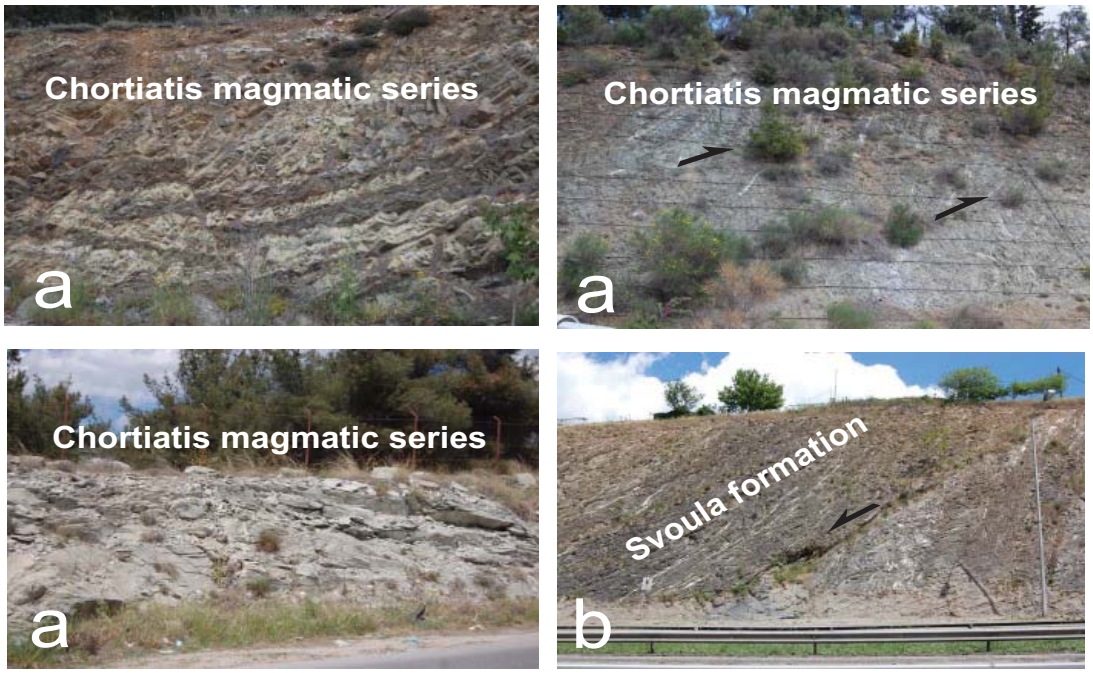


Figure A1
a. The Chortiatis Jurassic magmatic series, possibly related to an island arc magmatic activity. b. The Jurassic Svoula flyschoide formation deposited at the slope of the continental margin of the Eurasia plate. 40°37'36"N 22°58'47"E, 40°38'06"N 22°58'17"E, 40°38'42"N 22°58'08"E

The Melissohori-Cholomontas unit is composed of intercalations of sericit-chloritoid clay- and sandstone- phyllites with thin layers or boudins of limestones (fig. A1b). Usually, serpentinite layers also occur, tectonically emplaced, in between the meta-sediments of the Melissochori-Cholomontas unit.

Just before the exit to the Egnatia highway the CRB is thrusted to the West over an ophiolite mélanges zone, which is intensively overprinted by brittle, towards NE and SW, thrust faults, and younger normal faults (fig. A2). The ophiolite mélanges together with ophiolite rocks are again thrusted over a Triassic-Jurassic neritic carbonate sequence belonging to the Circum Rhodope belt. The ophiolite mélanges correspond to the remnants of the Neotythean ophiolites belt, which were emplaced during the Late Jurassic towards NE on the Trassic-Jurassic carbonate series of the Serbo-Macedonian continental margin (fig. A2; Bonev & Stampfli 2003, 2008, Jahn-Awe et al. 2010, Froitzheim et al. 2014).

In the continuing, we make a short detour from our main course, in order to realize our next stop

Figure A2
Jurassic ophiolite mélanges at the eastern most part of the Paeonia subzone. Olistoliths of Triassic carbonate rocks are tectonically incorporated in the Neotethyan ophiolite rocks during their obduction. 40°41'32"N 22°57'54"E

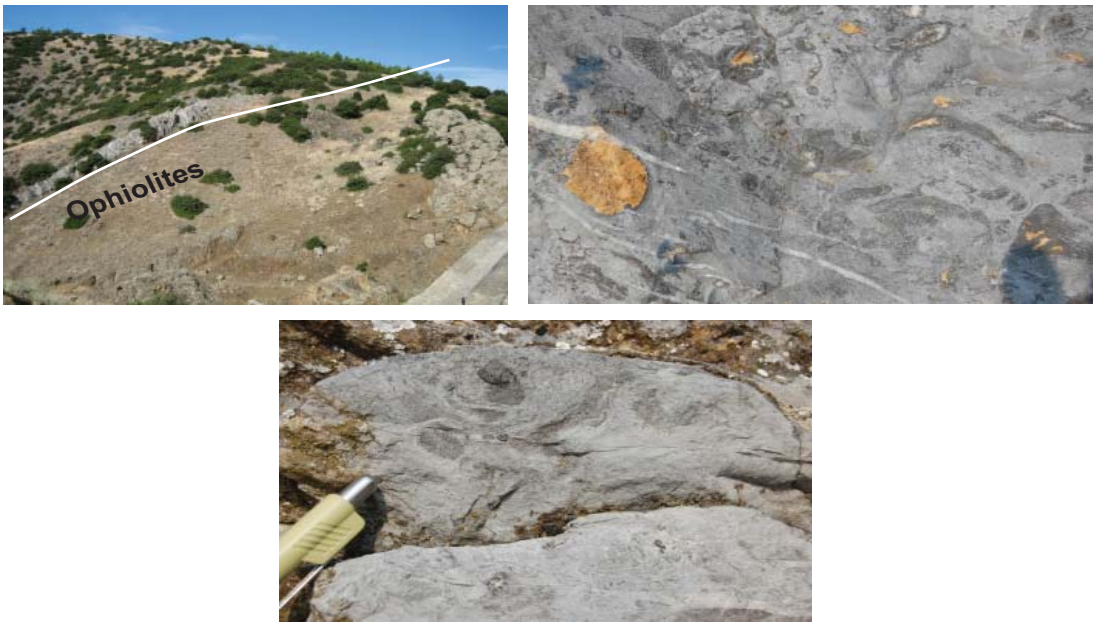


Figure A3
The transgression of the Upper Jurassic-Lower Cretaceous carbonate neritic and clastic sediments over the Paeonia ophiolite belt. The carbonate clasts and the thin carbonate layers are rich in fossils representative to an Upper Jurassic fauna collection. Axios/Vardar zone, Paeonia Subzone. 40°44'52"N 22°52'38"E

on the outskirts of the village of Neochorouda. From there, we take a gravel road along which, from the south to the north, we cross: the Late Jurassic-Early Cretaceous clastic sediments of the Paeonia subzone, the Paeonia's ophiolite belt, and finally the Circum Rhodope Belt units (fig. A3, A4, A5).

After that we turn buck on the Egnatia highway, crossing again the Chortiatis magmatic series and the Melissochori-Cholomontas turbidite unit of the CRB. As it is described earlier in the text, (geological setting), the several CRB units have been intensively imbricated towards SW and also NE under brittle conditions during the Oligocene-Miocene (Tranos et al. 1999, Mainhold et al. 2009, Mainhold and Kostopoulos 2013). The several CRB units have been metamorphosed under greenschist facies conditions during the Late Jurassic-Early Cretaceous, and are characterized by isoclinal-to-asymmetric recumbent folds and a

Figure A4
Isoclinal folding of the Upper-Jurassic-Lower Cretaceous sediments. A S1/S2 fabric is clearly distinguished. The isoclinal folds and the achsen-parallel S2-foliation are dated as of late Early Cretaceous age (D2-event). Axios/Vardar zone, Paeonia Subzone. 40°44'34"N 22°52'31"E

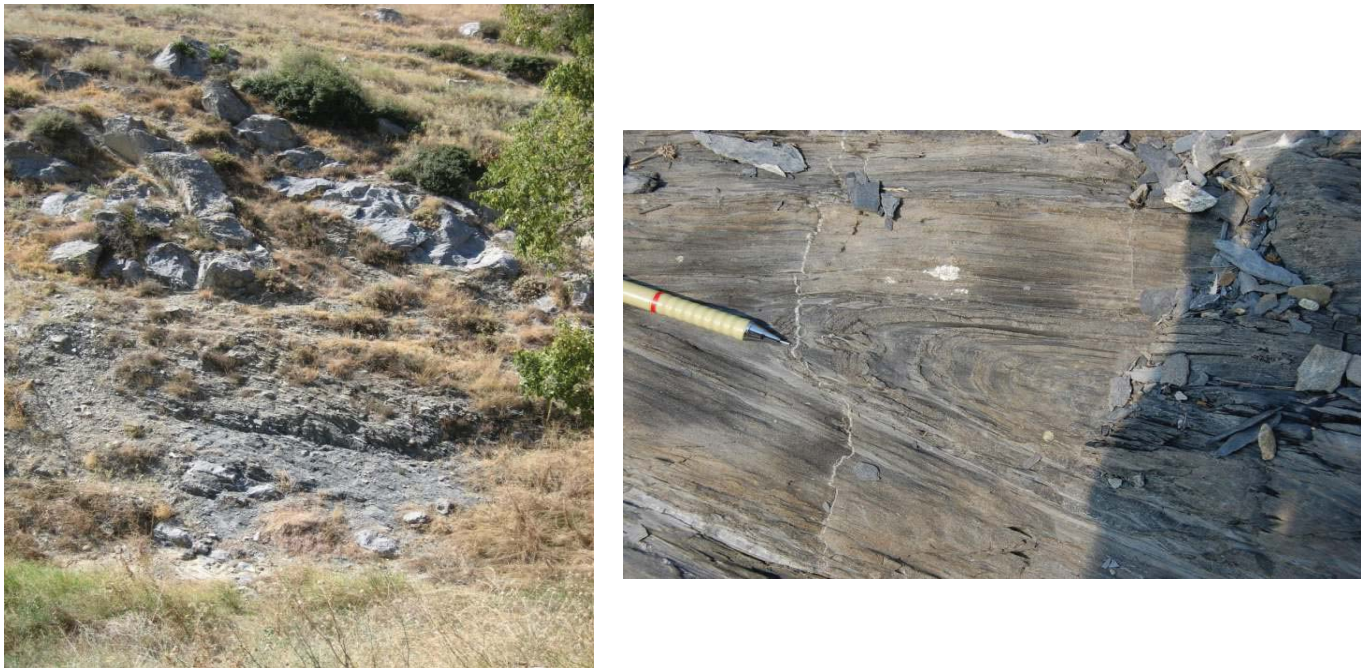


Figure A5
Fossiliferous Triassic carbonate sediments of the Serbo-Macedonian continental margin (Laurassia). Ammonites` and Gastropods` fossils are distinguished. These deposits are strongly imbricated with the Middle-Jurassic flysch-type Svoula sedimentary series, which belongs to the Circum-Rhodope belt. 40°45'46"N 22°52'38"E, 40°45'41"N 22°53'21"E

pervasive S1/S2 foliation fabric (Kaufmann et al. 1976, Tranos et al. 1999).

As we now exit the Egnatia highway at the town of Langada, we first take the provincial road to the village of Pentes Vryses, and then the national road at the northern side of the lakes Langada and Volvi, which towards the east leads us to the city of Drama.

On this trek, we initially cross the western-most boundary of the upper Serbo-Macedonian Vertiskos unit, which is defined by a dextral strike-slip fault related to a transpressional tectonic regime. Along this fault, the SRB is thrusted over the Oligocene mollasic sediments of the Langada (or else Mygdonia) basin and the Circum Rhodope belt (fig. A6). In the continuing, as we move parallel to the northern tectonic boundary of the Langada basin with the Vertiskos unit, we find he Volvi-ophiolite complex, composed mainly of metagabbros and mantle rocks, sandwiched between the lower Serbo-Macedonian Kerdylia unit and the upper Serbo-Macedonian Vertiskos unit (fig. A7). Further on, after we pass the town of Asprovalta, the migmatites of the Kerdylia unit occur. Their age is unknown, but most probably they were formed in the Tertiary (fig. A8).

Before reaching the city of Drama, and while being on the supradetachment Strymon basin, we travel almost parallel to the normal detachment fault along which the Serbo-Macedonian massif has broke-off and moved to the west. This tectonic movement was associated with the exhumation

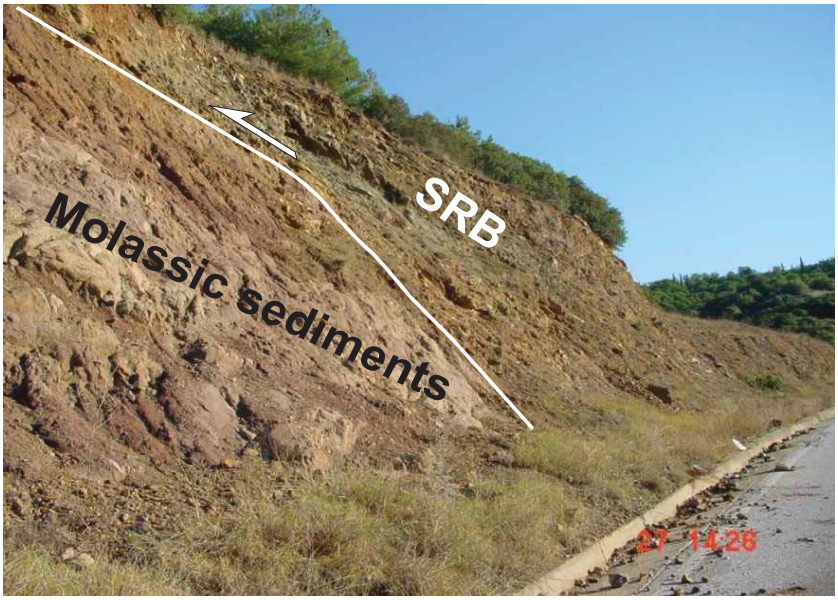


Figure A6
The Oligocene-Miocene SW-ward thrusting of the Vertiskos unit on the Eocene-Oligocene molassic sediments of the Mygdonia (Lagadas) basin. The thrusting was developed under a transpressional tectonic regime and was possibly related to the Strymon normal detachment fault, along which the SRB was detached down-ward to the SW, leading to the final exhumation of the Rhodope Pangaion unit. 40°47'22"N 23°07'22"E

Figure A7
The meta-ophiolites pointing the tectonic contact between the Vertiskos and Kerdylia units of the Serbo-Macedonian massif. Asymmetric folds verging downwards to the West is also shown. This contact is interpreted as an Eocene-Oligocene top-to-the-SW normal detachment fault, reactivating an older thrust fault zone responsible for the initial emplacement of the Vertiskos unit on top of the Kerdylia unit. Serbo-Macedonian massif. 40°41'14"N 23°32'07"E

Figure A8
Migmatites of the Kerdylia unit overprinted by a younger, Tertiary-Quaternary conjugate normal fault system. The age of the migmatites remains under debate. They are possibly of Tertiary age. Serbo-Macedonian massif. 40°46'05"N 23°46'43"E

of the lower Rhodope Pangaion unit as a metamorphic core complex, which rises up along our course (fig. A9; Pangaion Mountain).

Along the provincial asphalt road from Drama to Sidironero, we cross, from the south to the north, first the lower Rhodope Pangaion unit, composed of a thick carbonate series and intercalations of marbles and mica schists (fig. A10, A11), and then the complicate Middle Rhodope Sidironero unit that is thrusted over the Pangaion unit (fig. A12). The Sidironero unit is consisted of both continental and oceanic origin metamorphic rock sequences. It comprises gneisses, schists, amphibolites, migmatites and amphibolitized eclogites (fig. A13, A14, A15, A16, A17). Last, from there as we move along the Sidironero unit, we have an excellent view on the Nestos river (fig. A18) and the Pangaion window (fig. A19) and finally, we cut the contact of the Paleocene-Eocene Skaloti-Elatia granitoide with the Sidironero gneisses and marbles series (fig. A 20).

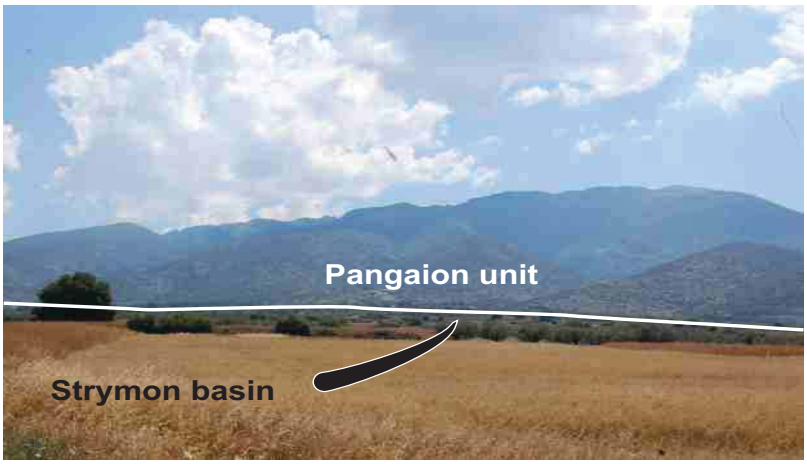


Figure A9
The Oligocene – Miocene Strymon normal detachment fault leaded to the exhumation of the Rhodope Pangaion unit. 40°54'08"N 23°56'50"E

Figure A10
Mica-schist layers of the Rhodope Pangaion unit intercalated with thin marble layers. They are intensively folded by isoclinal, recumbent folds. A mega sheath fold is also distinguished. Sense of shear to the left (SW-ward), indicated by the quartz σ-clasts. The overall structural setting is related to the SW-ward detachment of the Serbo-Macedonian massif during the Oligocene-Miocene and the progressive exhumation of the Pangaion unit as a metamorphic core complex. Rhodope massif. 41°12'09"N 24°10'57"E, 41°12'15"N 24°10'56"E

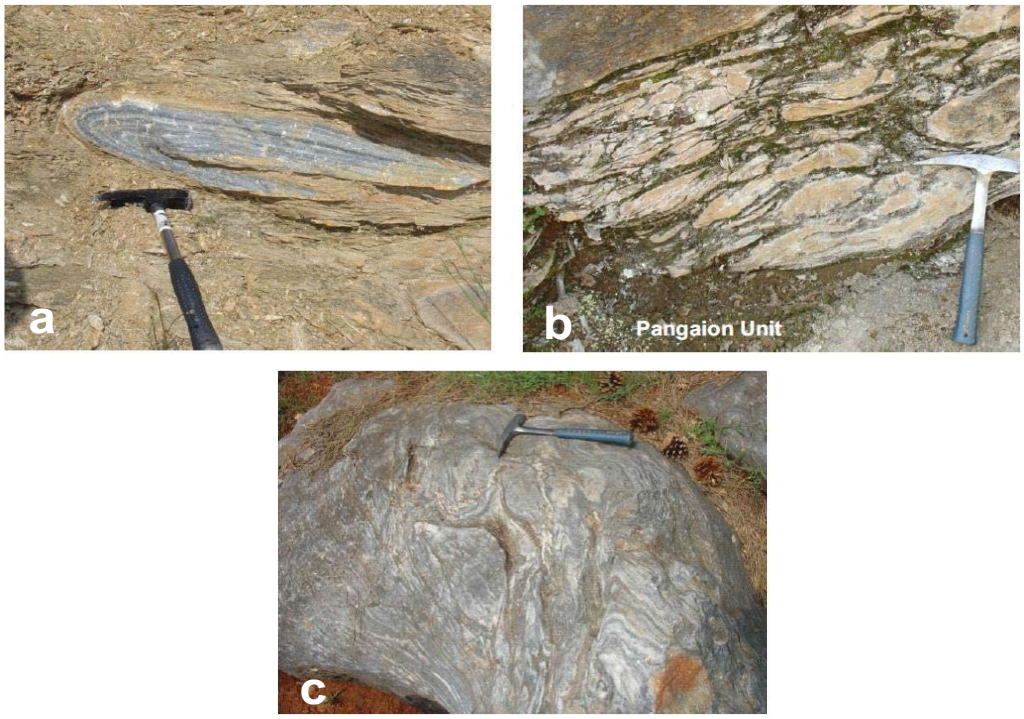


Figure A11
Fold structures in the Pangaion unit: a. isoclinal folds, b. sheath-folds, c. interference-fabric related to refolding events. All folds are parallel re-oriented to the dominant, NE-SW trending, stretching lineation (X-axis) characterizing the entire Rhodope massif. a. 41°12'17"N 24°10'56"E, b. 41°13'04"N 24°10'58"E, c. 41°12'35"N 24°10'54"E

Figure A12
The Sidironero overthrusting on the Pangaion unit. In detail: the gneisses rocks of the Sidironero unit overthrust the carbonate rocks of the Pangaion unit. Rhodope massif. 41°15'47"N 24°12'08"E

Figure A13
S-C-C` fabric of Eocene age on the Upper Paleozoic (Carboniferous age) granite of the Sidironero unit. Sense of shear top-to-the-SW. 41°19'11"N 24°12'37"E

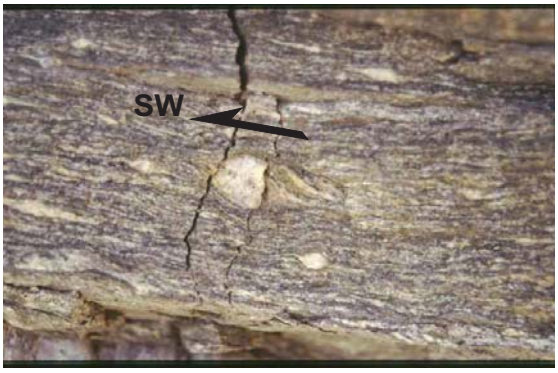


Figure A14
Sidironero augen gneisses. Feldspars σ- and δ-clasts indicating top-to-the-SW sense of shear, related to the Rhodope collapse and progressive exhumation of the Rhodope units. Rhodope massif. 41°19'58"N 24°12'34"E

Figure A15
Amphibolitized eclogites of Morb-origin, tectonically intercalated in between the Sidironero gneisses and thin marble layers. About vertical tension gashes in the amphibolitized eclogites are also distinguished, indicating a subhorizontal extension regime (X-axis). In these amphibolite rocks residual patters of diamond crystals, are also described (Schmidt et al. 2010) showing the ultra-high pressure metamorphic conditions, under wich these Morb-rocks were metamorphosed, revealing subduction processes inbetween the continental Rhodope units. The ultra-high to high-pressure metamorphic conditions are dated of Paleocene-Eocene age. 41°20'19"N 24°12'29"E

Figure A16
Folded migamtites of the Sidironero unit, related to the Tertiary extension regime and the, post high-pressure, Eocene high temperature metamorphic event affected the Rhodope Sidironero unit. 41°20'18"N 24°12'29"E

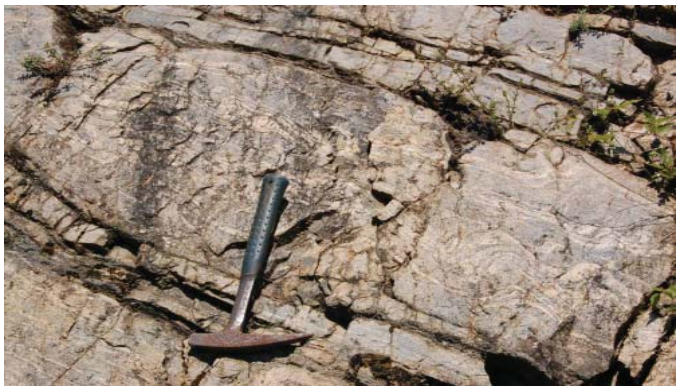


Figure A17
a. & b. ptygmatic folds associated with a high angle dipping of the migmatite foliation. A constrictional-type deformation is supposed to be related to this high angle, almost vertical, dipping of the migmatite fabric. c. asymmetric boudins of the migmatite leucosoms, indicating a top-to-the-SW sense of shear. Rhodope massif. a, b. 41°20'43"N 24°12'00"E, c. 41°20'21"N 24°12'31"E

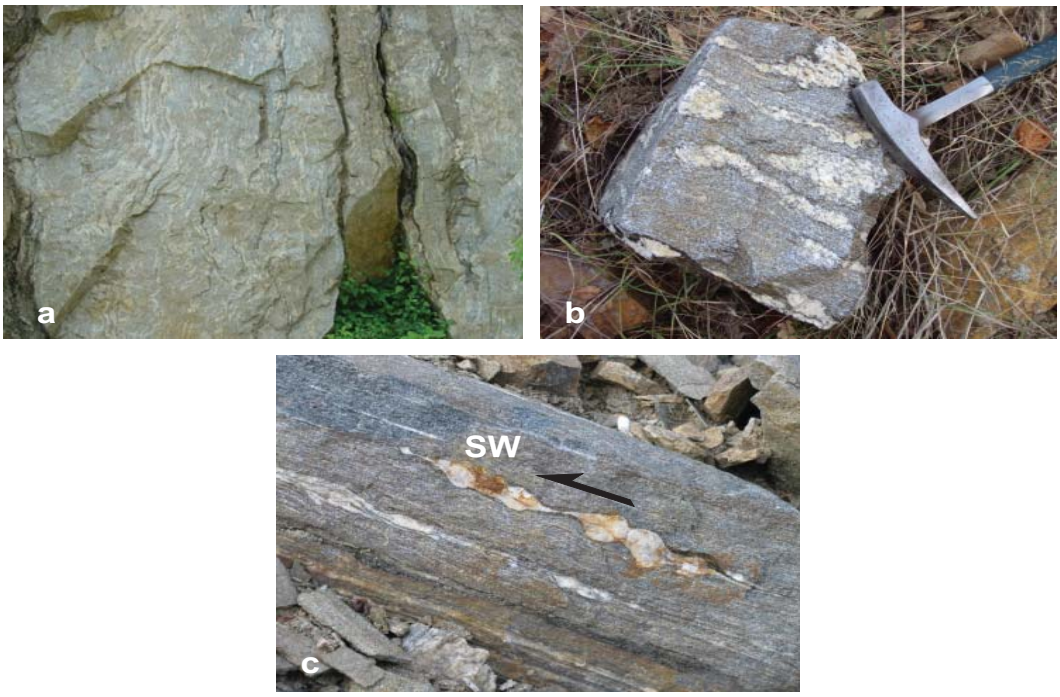
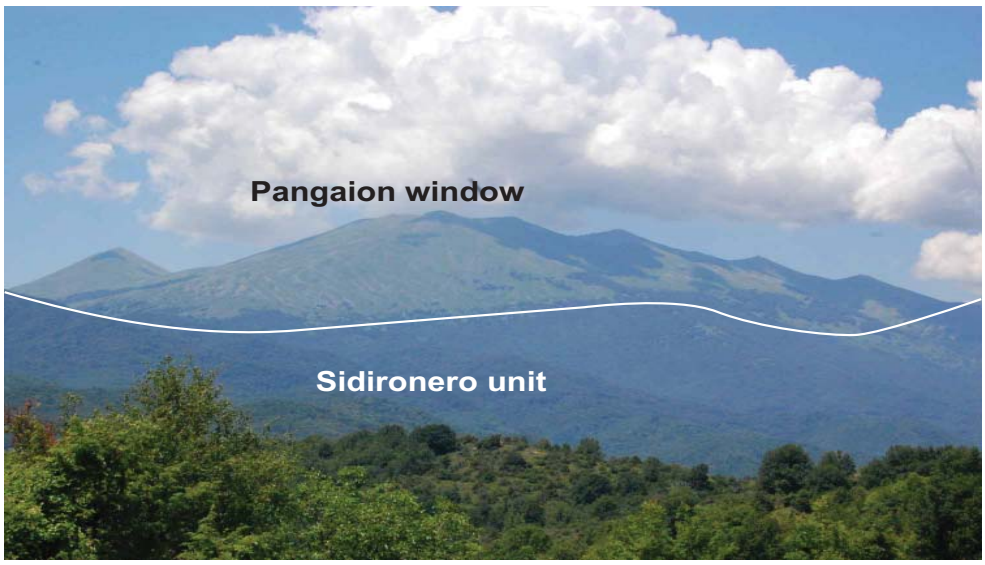


Figure A18
Panoramic view of Nestos river cutting the Rhodope units. 41°21'04"N 24°11'51"E

Figure A19
The Paleocene – Eocene Elatia granite intruded the Sidironero unit during the Tertiary extensional setting dominated at the Rhodope province. 41°22'02"N 24°14'21"E

Figure A20
Panoramic view of the Pangaion window that is possibly equivalent to the Olympos-Ossa unit of the External Hellenides. View to the South. 41°22'10"N 24°13'50"E



B. 2nd Day: Internal Hellenides (fig. 37)

I. Axios/Vardar zone:

I.1. Paeonia subzone
I.2. Paikon subzone
I.3. Almopia subzone

Trek: From the city of Thessaloniki to the villages of Livadi and Griva through the villages of Eyzonoi, Skra and Archagelos (fig. 38).

Overnight staying in Thessaloniki.

The second day's cross-section starts from Thessaloniki and finishes at the village of Griva at the foot of Mount Paikon. Starting from the ring-road of Thessaloniki with general direction to Athens, we initially turn onto the national road

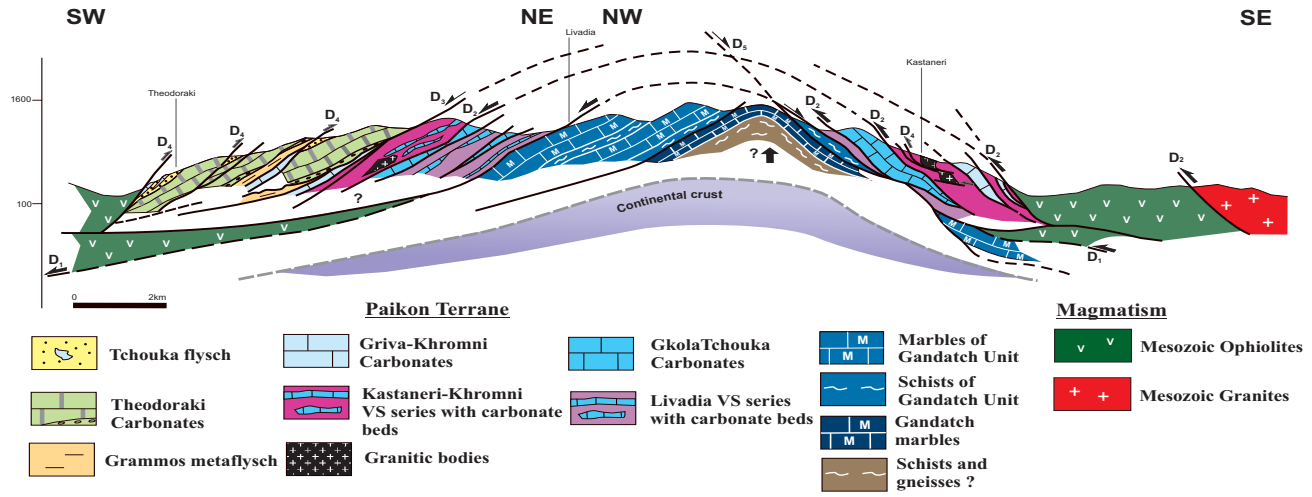


Figure 37
Schematic geological cross-section through the selected area/trek of the 2nd day in the Axios/Vardar zone (Paeonia and Paikon subzones).

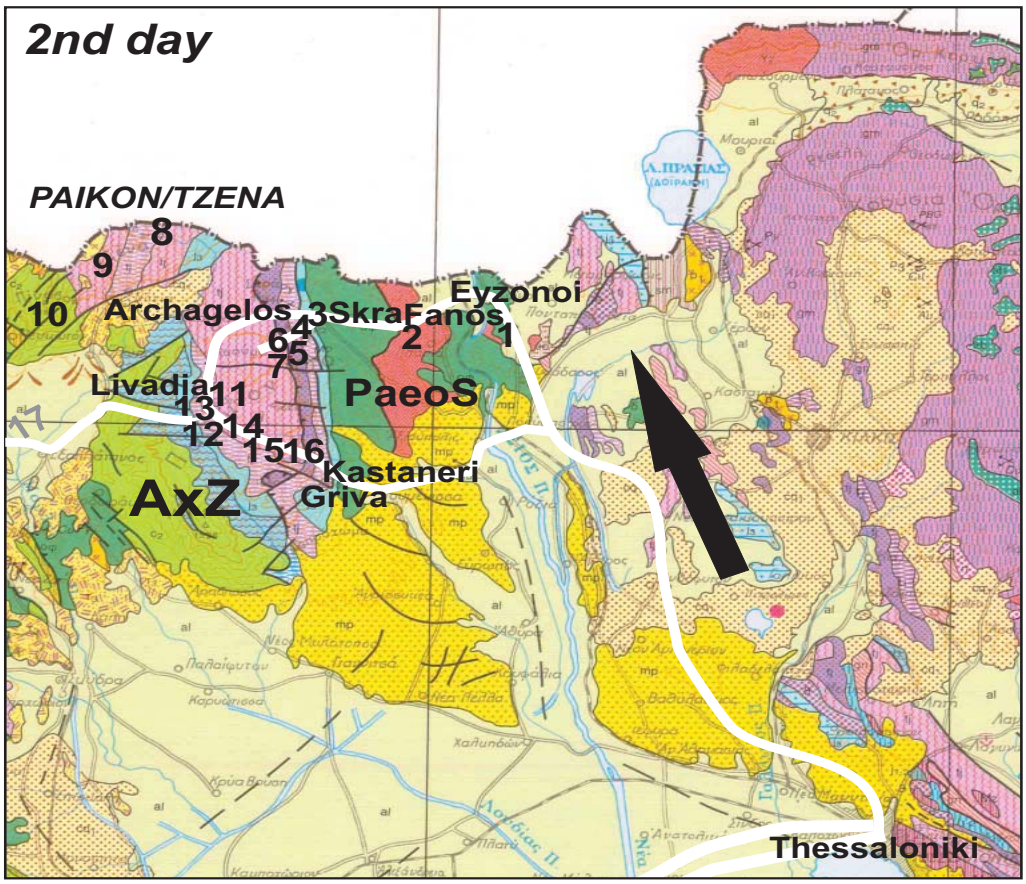


Figure 38
Location of the selected outcrops and stops are shown on the corresponding trek of the 2nd day

Figure B1
Sheeted-dykes complex and magmatic conglomerates of the Paeonia back-arc ophiolites. The ophiolite complex is mainly built by dykes of plagiogranite and gabroide rocks. Paeonia subzone, Axios zone. 41°06'28"N 22°33'34"E

leading to Edessa, and then turn on to the national highway towards FYROM. Before we reach the Greek-FYROM borders, we take the sealed road to the villages of Fanos and Livadi, and from there we continue until the day's final destination, village Griva.

Along this traverse, which actually starts with a first stop on the national road to FYROM, near the village Evzoni and the Greek customs (fig. B1), we cross the entire Axios/Vardar zone, and more specifically, first the Paeonia, then the Paikon and last the Almopia subzone.

The Paeonia subzone is consisted of rocks belonging to the ophiolitic sequence, mainly gabbroid, as well as those from the sheeted dykes' complex (fig. B1). Those ophiolites were intruded by the calc-alkaline, island-arc related, Fanos granite, during the Late Jurasssic (fig. B2; Anders et al. 2005, Saric et al. 2009, Michail et al. 2016). They are interpreted as back-arc or supra-subduction zone ophiolites with a Mid-Late Jurassic age (Zachariadis et al. 2007a, b, Saccani et al. 2008, Bortolloti et al. 2013, Michail et al. 2016). The Paeonia subzone is traced all along until its contact with the Paikon subzone near the village of Skra, on which it is tectonically emplaced during the Late Jurassic-Early Cretaceous (fig. B3, B4, B5, B6, B7).

The Paikon subzone is composed, from bottom to top, of pre-Alpine basement rocks outcropping at the foothills of Tzena Mountain (fig. B8), a Triassic-Jurassic carbonate cover (fig. B9), Mid-Late Jurassic volcanoclastic rocks (fig. B5, B6; Kastaneri and Livadi formations) and Late Jurassic-Early Cretaceous neritic carbonate rocks (Gropi, Kromni, Gkola Tchouka and Griva formations; fig. 18, fig. B5, B7). All these rock sequences are intensively imbricated, forming a complicated nappe pile architecture (fig. 16, 19; Brown & Robertson 2003, Katrivanos et al. 2013). Furthermore, neritic Late Cretaceous carbonate rocks, terminating in the Late Cretaceous-Paleocene flysch of the Internal Hellenides, rest transgressively on top of all the previously described formations. While being on the road connecting the villages Archangelos and Livadi, we have a panoramic view of the Late Cretaceous limestones occurring in Pinovo Mountain (fig. B10), and the Triassic-Jurassic carbonate cover of the pre-Alpine basement rocks in Tzena Mountain (fig. B8, B9). On the same road,

Figure B2
Late Jurassic Fanos granite. This granite body intruded the Paeonia (Guevgueli) ophiolite complex at an intraoceanic subduction setting evolved in the Neotethys ocean during the Middle-Late Jurassic. Paeonia Subzone, Axios zone. 41°04'50"N 22°29'12"E

Figure B3
The late Early Cretaceous W-ward thrusting (D2) of the Paeonia, supra-subduction ophiolites on the Griva formation of the Paikon subzone. 41°04'40"N 22°23'17"E

Figure B4
Kink folds of the Griva formation related to the Paleocene – Eocene compressional event (D4). 41°04'10"N 22°23'28"E

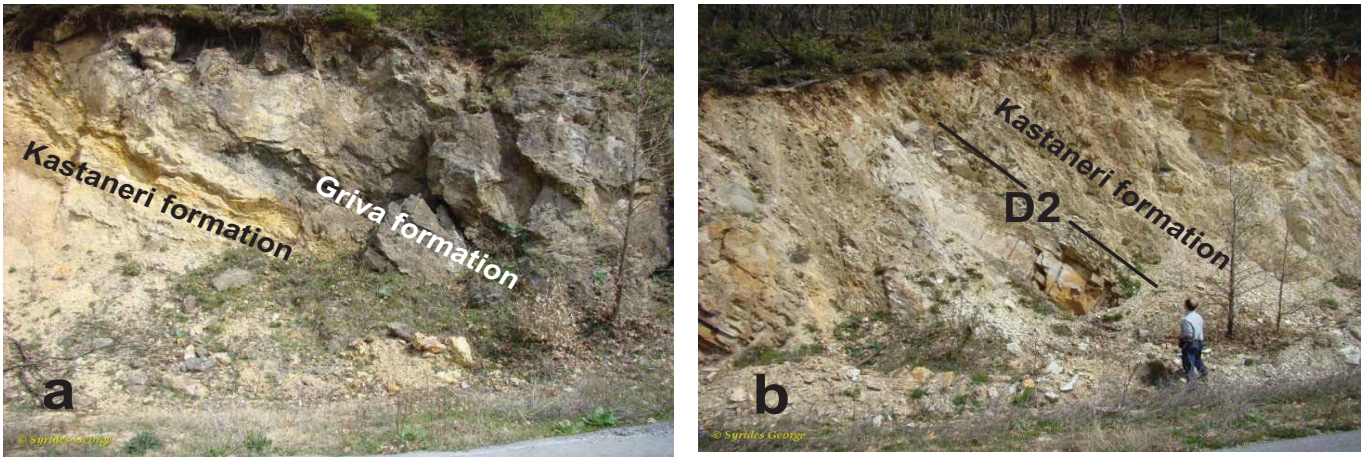


Figure B5
a. The tectonic contact (D2-compressional event) between the Upper Jurassic-Lower Cretaceous Griva formation and the Middle Jurassic volcanosedimentary Kastaneri formation, both belonging to the Paikon subzone of the Axios zone. b. The acid volcanic products of the Kastaneri formation, strongly deformed by the late Early Cretaceous D2-event, produced the illustrated dominant S2-foliation. Paeonia subzone, Axios zone. 41°04'14"N 22°23'15"E

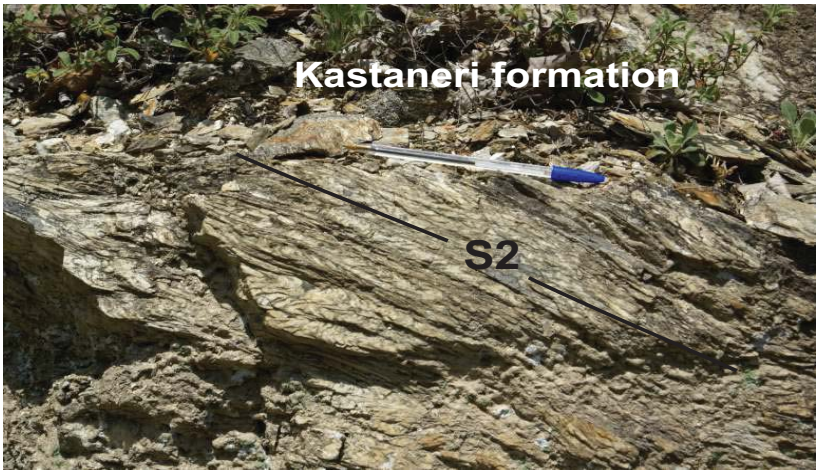


Figure B6
S1/S2 fabric in the Kastaneri formation. The old Upper Jurassic S1-foliation is strongly re-oriented along the S2-foliation, forming a granulation cleavage fabric. Due to the strong translation along the S2-planes, the two foliation are usually developed parallel to one another, so that only one fabric element seems to be recognized, on the formation and this is the S2-foliation. Paeonia subzone, Axios zone.
41°04'18"N 22°22'58"E

Figure B7
The tectonic contact between the Gola Tsouka carbonate formation and the Triassic – Jurassic Gandatch formation reworked by an E-ward normal detachment fault. The Livadi formation between both formations is completely omitted.
41°04'18"N 22°21'55"E

Figure B8
On the top of Tzena mountain subisoclinal NNE trending D2 mega-folds deform the continuation of the Triassic-Jurassic Gandatch formation of the Paikon massif to the Tzena mountain. The active, ENE-WSW striking, Aridea high angle fault divides Tzena and Paikon massifs, composed of the same tectonostratigraphic column. Due to the activity of this fault its northern footwall segment has risen, forming the Tzena Mountain top and the exhumation of the Paleozoic basement rocks of the Triassic Gandatch carbonate formation. View to the North.
41°04'46"N 22°16'54"E

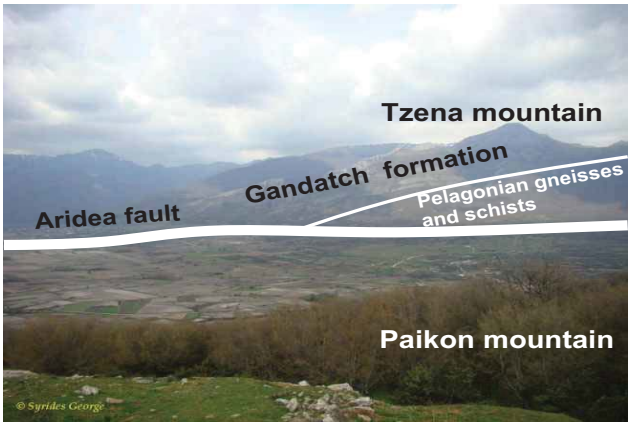
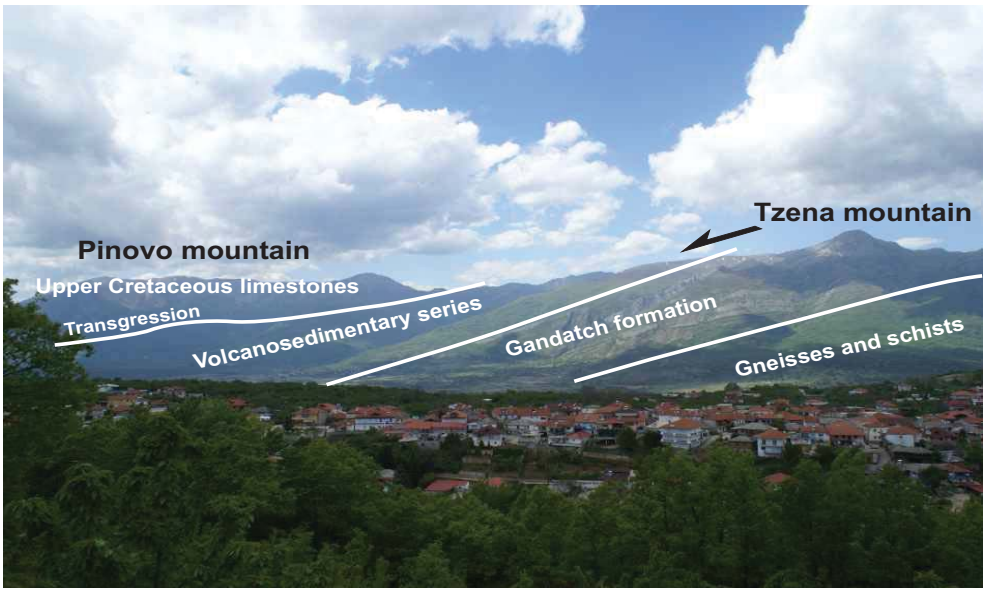


Figure B9
The Triassic-Jurassic succession in the Tzena Mountain composed by intercalation of marbles and mica-schists. It is equivalent to the Gandatch formation further to the South in Paikon Mountain but divided from it due to the Aridea fault zone cuasing the uplift of its northern footwall segment, which is the Tzena Mountain itself. This explains why (and how) the Paleozoic basement rocks under the Gandatch formation are found exhumed alongside the Tzena mountain region. 41°06'09"N 22°12'59"E

Figure B10
Panoramic view of the transgressive Upper Cretaceous Pinovo neritic carbonate rocks (Pinovo Mountain). View to the North. 41°04'46"N 22°16'54"E



all along the way to the village of Livadi and imediatelly after that, we see the Triassic-Jurassic carbonate rocks of the Paikon massif, with marbles and mica-schists intercalations (Gandatch formation), all intensively deformed, with isoclinal and recumbent asymmetric folds (fig. B11).

Just after the village of Livadi, on the road to the villages of Kastaneri and Griva villages, we turn to a gravel road in order to reach the tectonic contact between the Gandatch formation and the Mid-Late Jurassic volcanic rocks of the Livadi formation that is also intercalated with few marbles layers, maybe olistoliths (fig. B12). This tectonic contact between the two Paikon formations is a normal detachment fault along which the Livadi formation was detached from the Gandatch formation, causing the updoming and finally the exhumation of the latter during the Oligocene –Miocene (fig. B13; Katrivanos et al. 2013).

Afterwards, we return to the sealed road and continue our trek to the village Griva, where we cross again the intensively folded Gandatch formation (fig.B14, B15), the tectonic contact between the Livadi volcanic formation and the Late Jurassic-Early Cretaceous Gkola Tchouka carbonate formation (fig. B16), the volcanoclastic

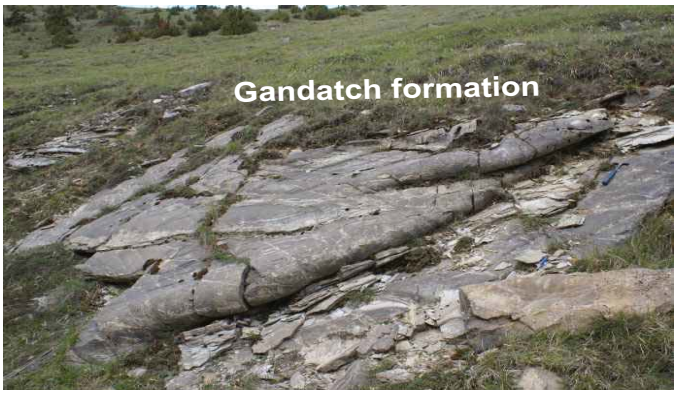
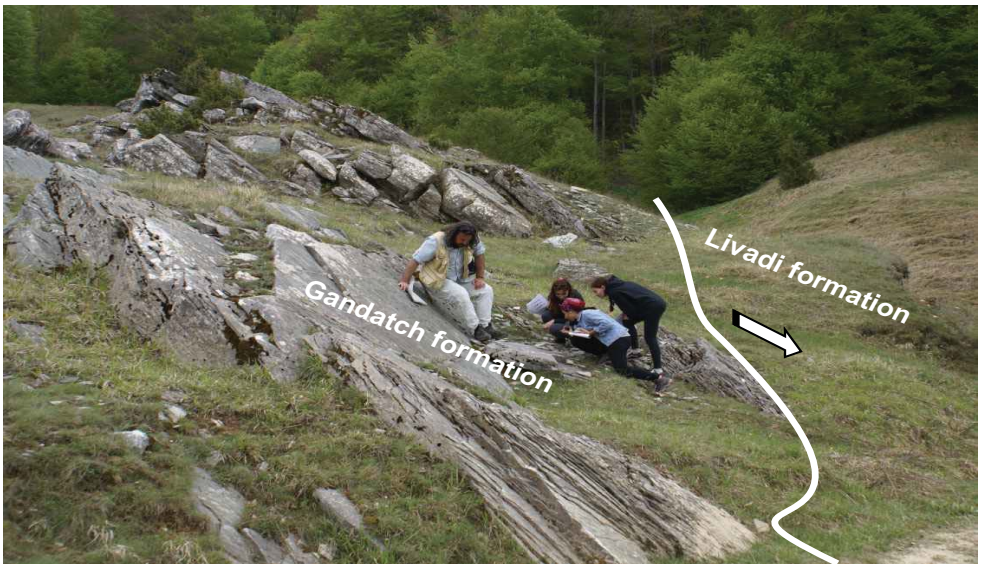


Figure B11
B-axis scattering of D2- isoclinal folds due to their re-orientation subparallel to the X-axis of the strain ellipsoide (Gandatch formation). 40°59'46"N 22°17'41"E, 40°59'42"N 22°17'44"E

Figure B12
Intercalation of carbonate layers and volcanics of the Paikon Livadi formation. 40°59'24"N 22°17'23"E

Figure B13
The Oligocene – Miocene W-ward normal detachment fault (D5) between the Livadi formation and Gandatch formation resulted to the final exhumation of the Gandatch formation. 40°59'28"N 22°17'35"E



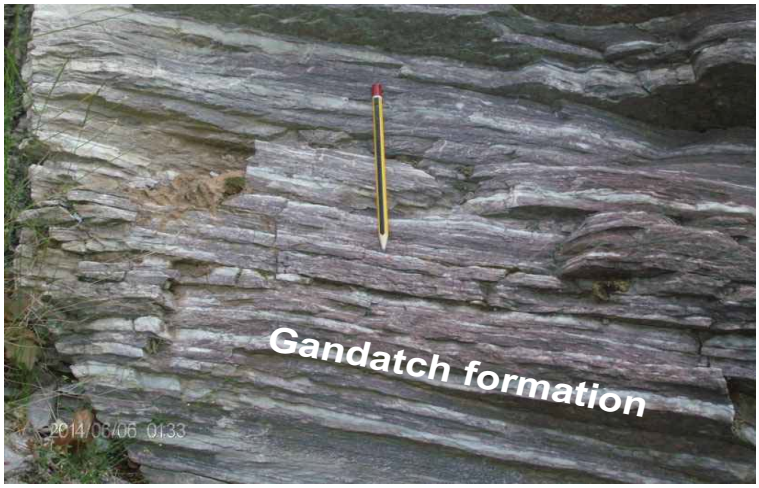


Figure B14
Red color, turbiditiic Triassic-Jurassic marbles of the Gandatch formation in Paikon Mountain. They are possible related to a continental marginal slope sedimentation at the Pelagonian continental margin (Apulia plate). They are strongly folded by the D1 and D2 tectonic events (fig. B15) that affected the entire Internal Hellenides zones during the Late Jurassic-Early Cretaceous. 40°59'34"N 22°18'00"E

Figure B15
D1 and D2 fold realm on the Triassic-Jurassic Gandatch marbles. Isoclinal and sheath folds related to the D1 event are overprinted by the D2 event. The S2 foliation dominates in these pictures. In fig. c. mullion structures are also visible, related to the D2 event. 40°59'34"N 22°18'00"E

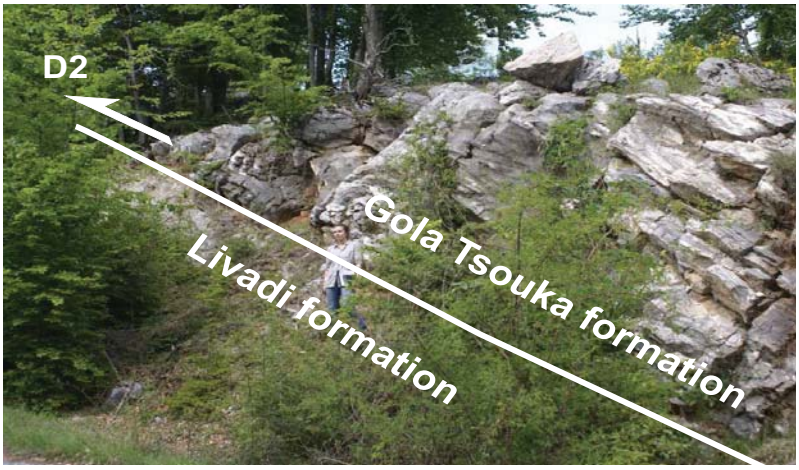


Figure B16
The tectonic contact (D2) between the Gola Tsouka and Livadi formations of the Paikon subzone (massif). 40°58'52"N 22°21'56"E

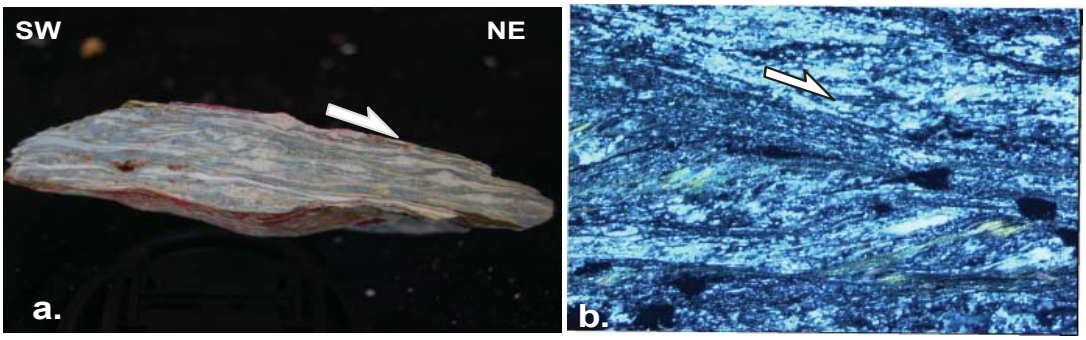


Figure B17
Mylonitic fabric in macro- and microscopic view of augen-gneisses from the Almopia subzone (Livadi unit; fig. 15) indicating a top-to-the-NE sense of shear (D3 event; XZ-section)

Kastaneri formation, and the Late Jurassic-Early Cretaceous Griva carbonate formation. Finally we reach at the outskirts of village Griva the Paeonia ophiolites again on the top of the Late Jurassic – Early Cretaceous carbonate rocks of the Griva formation.

A representative mylonitic fabric, related to the D3 extension, is also shown on the augengneisses formation from the structurally complicated Almopia subzone (fig. B17; sense of shear top-to-the-NE).

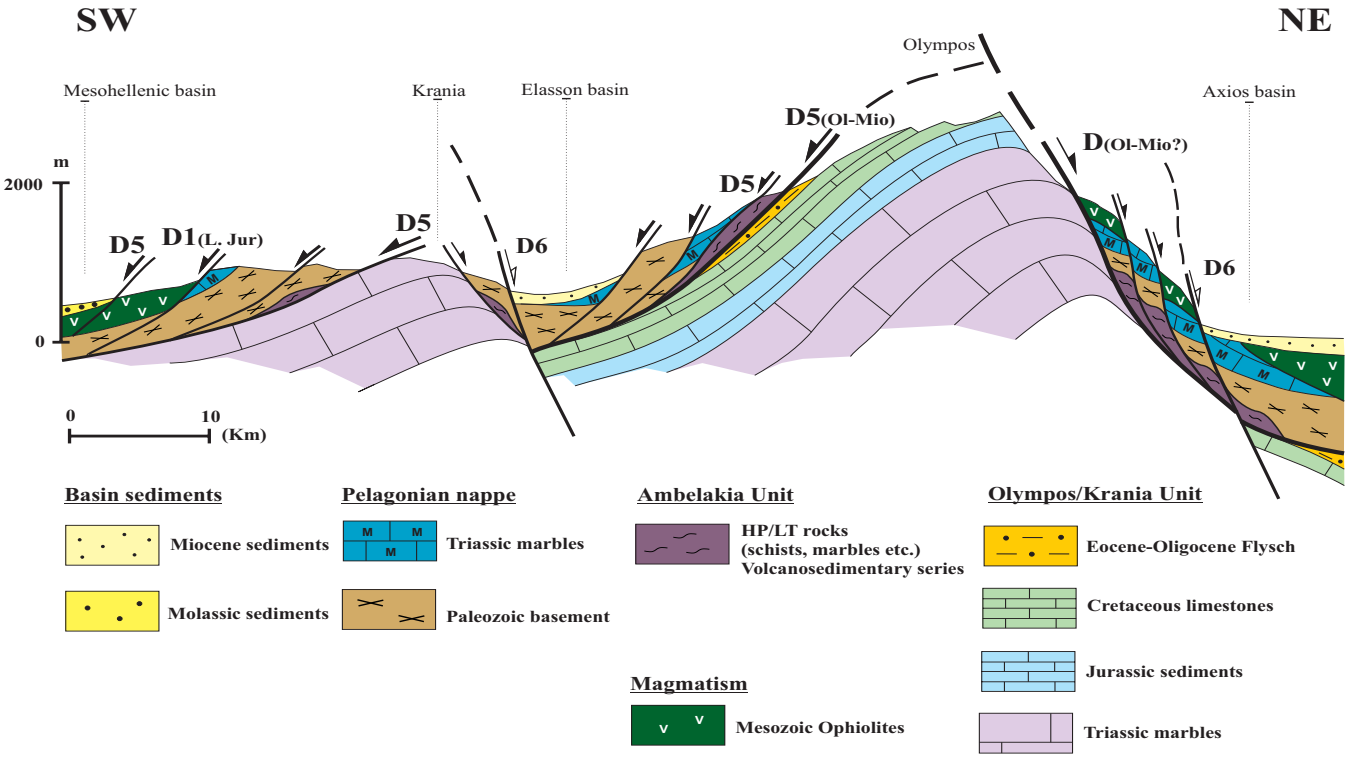
C. 3rd Day: Internal/External Hellenides (fig. 39)

I. Axios/Vardar zone (Almopia subzone)
II. Pelagonian nappe
III. Ampelakia blue schist unit
IV. Olympos-Ossa unit (External Hellenides, Gavrovo zone)

Trek: From Thessaloniki to the cities of Katerini and Elassona, and finally to the village of Ampelakia (fig. 40).

Overnight staying in Thessaloniki.

Figure 39
Schematic geological cross-section through the selected area/trek of the 3rd day in the Axios/Vardar zone, Pelagonian nappe and Olympos-Ossa unit.



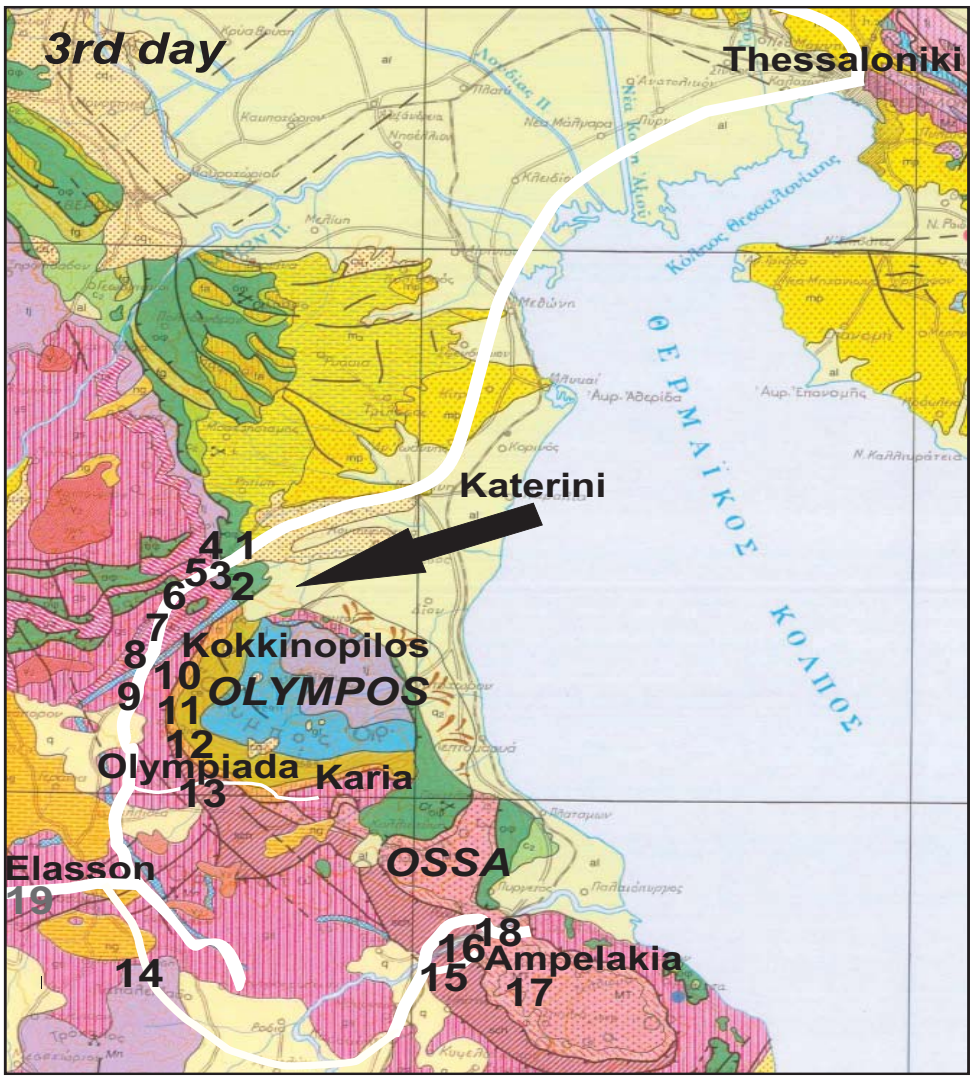


Figure 40
Location of the selected outcrops and stops are shown on the corresponding trek of the 3rd day.

The third day, we travel along the national highway from Thessaloniki to Katerini, and from there we take the old national road to the town of Elassona and onwards to the village of Ampelakia. The main purpose of the particular traverse is to study the Olympos tectonic window and its surrounding geological units and/ or nappes (fig. C1).

Our cross-section starts immediately after we exit the city of Katerini, with outcrops of the ophiolites-mélanges and ophiolites of the Almopia subzone (fig. C2), being thrusted over the Triassic-Jurassic carbonate cover of the Pelagonian nappe. The latter has undergone significant thinning at this site due to the Oligocene-Miocene post-emplacement extensional collapse (/detachment) of the Almopia's ophiolites towards NE (fig. C3, C4). A bit further we cross the Pelagonian basement (fig. C5) and we continue till its tectonic contact with the carbonate Olympos-Ossa unit that belongs to the External Hellenides (Gavrovo zone; i.e. near the small village of Kokkinopilos). The Pelagonian basement is consisted of gneises and schists, as well as mylonites (fig. C5, C6, C7, C8) and Ercynian granitoids (fig. C9), and appears intensively imbricated and multiply folded (fig. 13). On the other hand, the Olympos-Ossa unit is characterized by a continuous platform sedimentation from the Triassic to the Late Eocene, ended with an Eocene – Early Oligocene flysch deposition. The tectonic contact between those two geological formations (i.e. the Pelagonian basement and the Olympos-Ossa unit) is described by a normal detachment fault, which overprints the old thrust fault zone related to the emplacement of the Pelagonian nappe and the Internal Hellenides' blue schist belt (Ampelakia unit) on the External Hellenides during the Early Oligocene

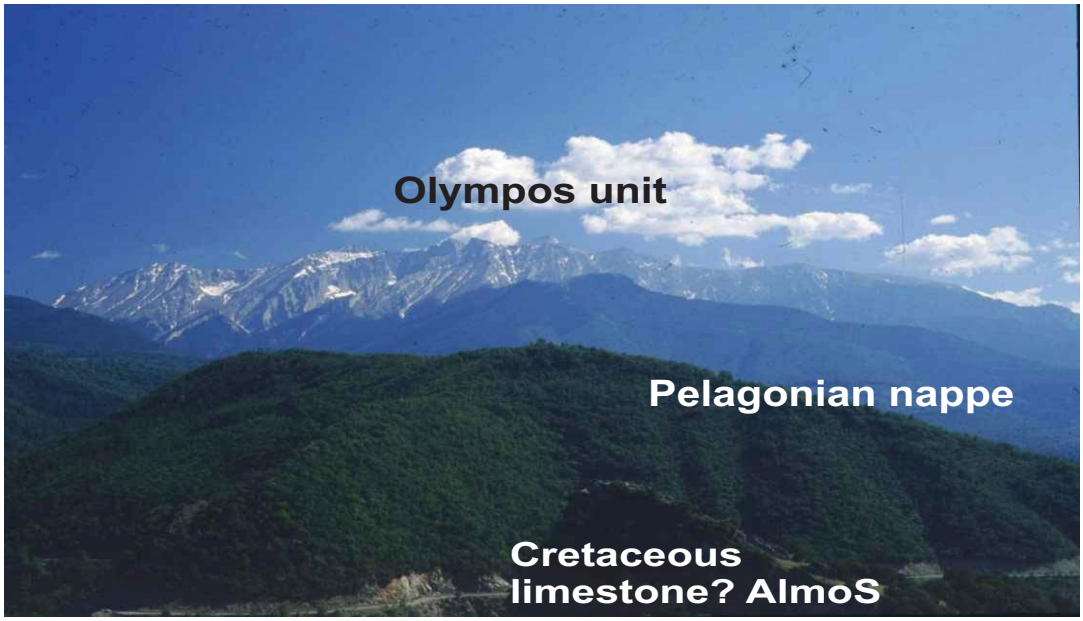


Figure C1
Panoramic view on the Olympos tectonic window, where the top Mitikas and the God Dias` Throne are distinguished. View to the South. 40°12'46"N 22°18'22"E

(fig. C10, C11). Along this younger normal detachment fault zone, the Internal Hellenides collapsed towards the SW. This normal detachment zone, in combination to the tectonic denudation of the Internal Hellenides' nappes pile to the NE at the eastern side of Mount Olympus, caused the exhumation of the External Hellenides' Gavrovo carbonates and flysch, and their exposure from below the Internal Hellenides as a tectonic window, the Olympos-Ossa tectonic window (fig. C12; Schermer et al. 1990, Kilias et al. 1991, Sfeikos et al. 1991, Schermer 1993, Kilias 1995).

At this particular location, the entire high-pressure Ampelakia unit is omitted from the tectonic

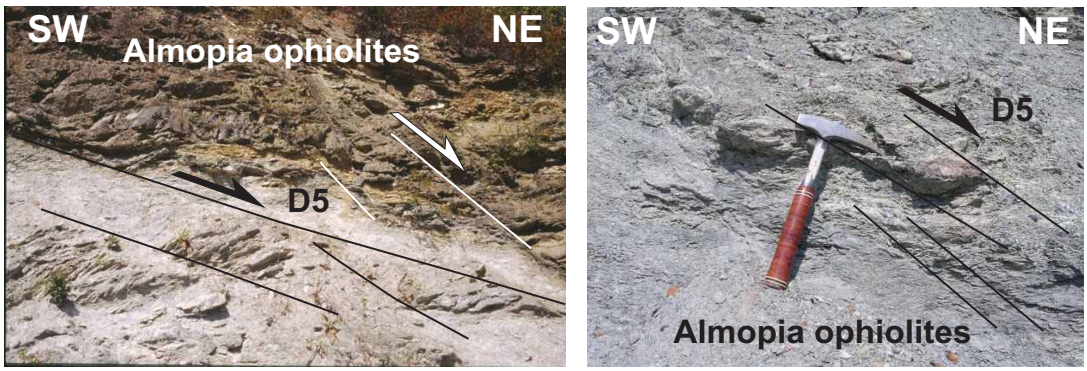


Figure C2
Extensional shear zones on the Almopia ophiolites and ophiolite mélanges that were emplaced westwards on the eastern Pelagonian continental margin during the Late Jurassic (? Kimmeridgian). The kinematics of the shear zones is top-to-the-East related to the D5 Oligocene-Miocene extension in the Olympos area and the downward detachment of the overthrusted tectonic nappes causing the finale exhumation of the Olympos carbonate unit. Tectonic nappes pile took place during the Tertiary with the finale emplacement of the Pelagonian nappe pile on the Olympos unit (External Hellenides) in the Eocene-Oligocene. At the East of the Olympos tectonic window dominates an E-ward downdip sense of movement and at its western side a W-ward sense of movement respectively. The E-ward sense of movement took place in semiductile to brittle conditions, while the W-ward sense of movement is characterized mainly by ductile conditions and the formation of the Olympos mylonites. The E-ward detachment seems to follow the westward one, causing the finale uplift of the Olympos unit and its high elevation position. Pelagonian nappe/Almopia subzone, Axios zone. 40°12'46"N 22°18'22"E 40°12'31"N 22°18'58"E (Petra village)

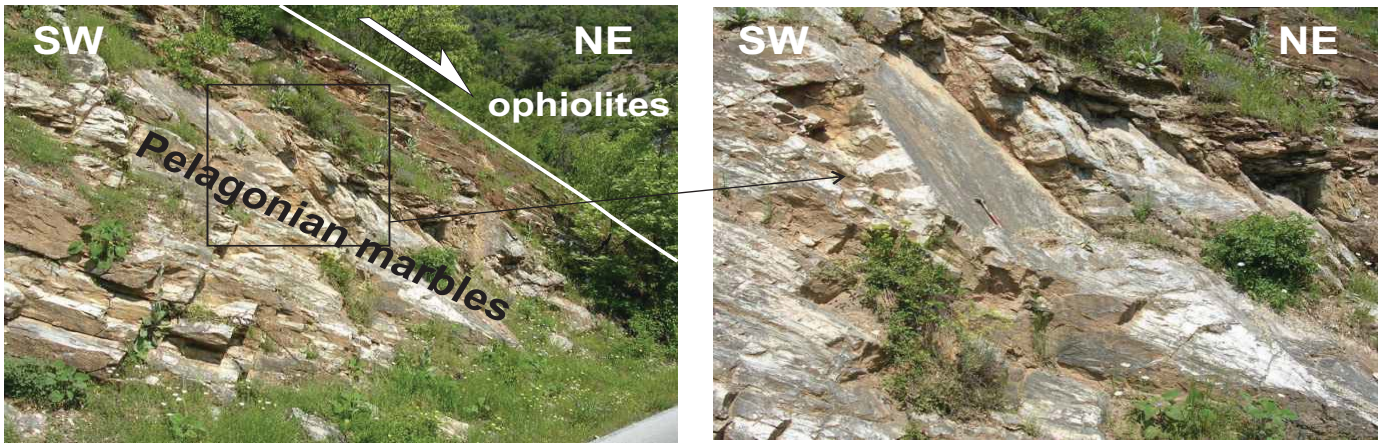


Figure C3
The tectonic contact between the Pelagonian Triassic-Jurassic platform carbonate sediments and the Almopia ophiolites. The latter were obducted W-ward on the eastern margin of the Pelagonian continent during the Late Jurassic from the Neotethyan ocean. This contact has been reworked by the Oligocene-Miocene extension causing the down-dip detachment of the ophiolites toward East. The extensional E-ward shear zone is well imprinted on the Triassic-Jurassic Pelagonian marbles, as well as on the ophiolites belt. Pelagonian nappe/Almopia subzone, Axios zone. 40°12'39"N 22°18'16"E

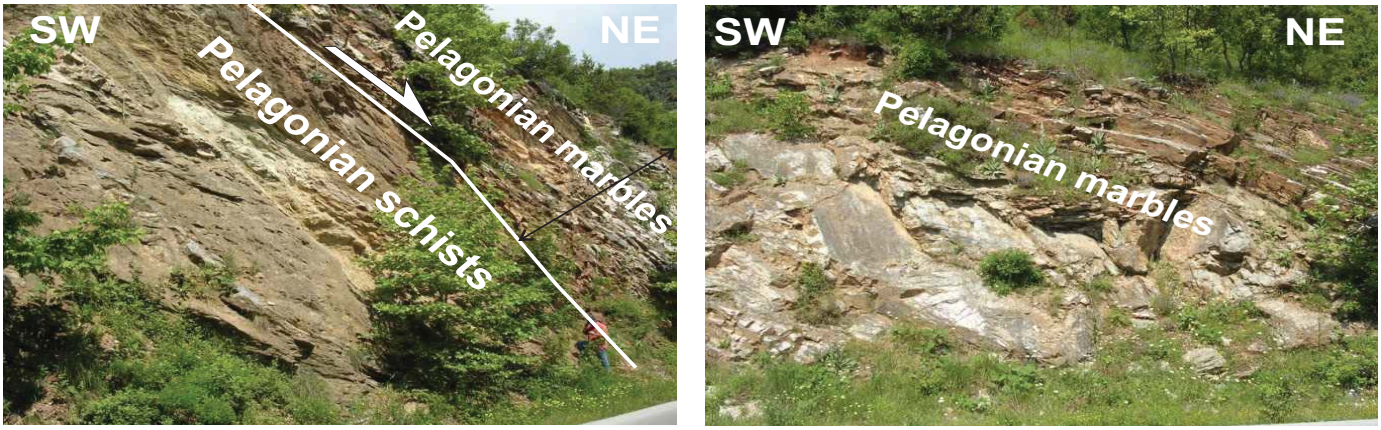


Figure C4
The tectonic contact between the Pelagonian Paleozoic basement rocks (mica schists) and its Triassic-Jurassic platform carbonate cover, which appears very thin, here (~50m), due to the strong extensional regime and the related Almopia ophiolites detachment. The tectonic contact itself is again an extensional shear zone towards the East, reworked an older thrust tectonics and crustal imbrication Pelagonian nappe. 40°12'46"N 22°18'22"E

contact between Internal and External Hellenides, confirming the extensional character of this detachment fault zone by juxtaposing the Pelagonian mylonitic basement rocks against the low grade metamorphosed to unmetamorphosed Olympos-Ossa unit (fig. C10, C11). Along the provincial road from Olympiada to Karia, we cross again the Pelagonian basement rocks and the blue schists with pervasive extensional shear bands that show a top-to-the-SW sense of movement, related to the Oligocene-Miocene SW-ward detachment of the Pelagonian and the blue schist's nappes (fig. C13).

Continuing on the new national road to the village Ampelakia, we cut the Pelagonian Triassic-Jurassic cover again, which is here characterized by a thickness of about 200m (fig. C14), much greater than that at its tectonic contact with the detached Almopia's ophiolites-mélanges and ophiolites. Further on, after we cross the national Thessaloniki-Athens highway and as we go up to the village of Ampelakia, we cross the typical

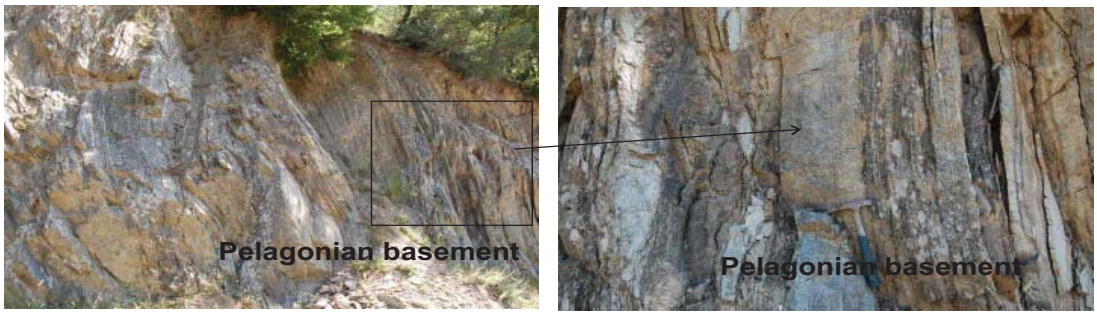


Figure C5
The Paleozoic Pelagonian basement composed by intercalations of gneisses, schists and amphibolites. A steep angle dipping of the Jurassic – Cretaceous symetamorphic foliation is recognized. 40°11'27"N 22°17'19"E

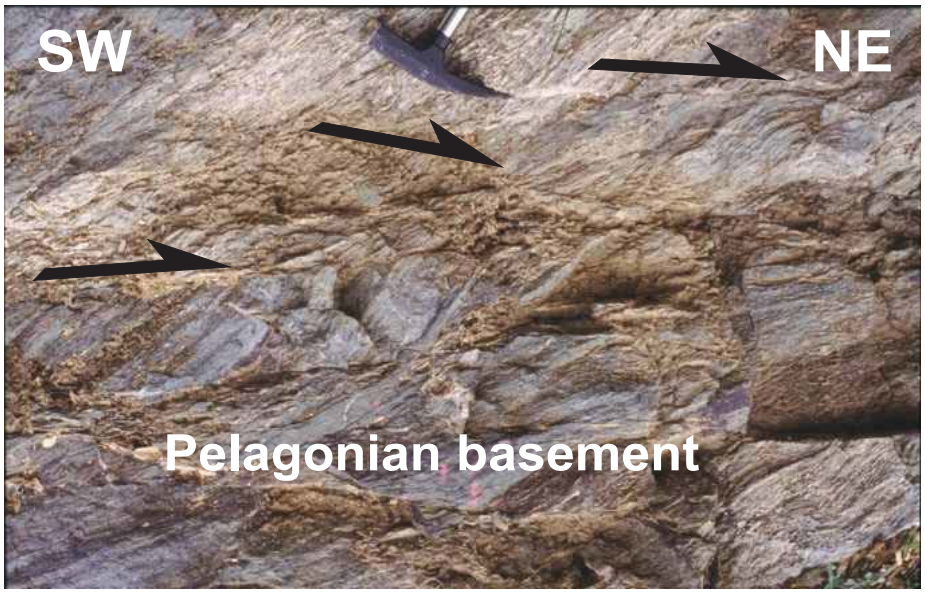


Figure C6
Extensional shear bands with a top-to-the-NE sense of shear on the Pelagonian Paleozoic basement rocks composed of gneisses and schists. Pelagonian nappe. 40°11'31"N 22°17'09."E

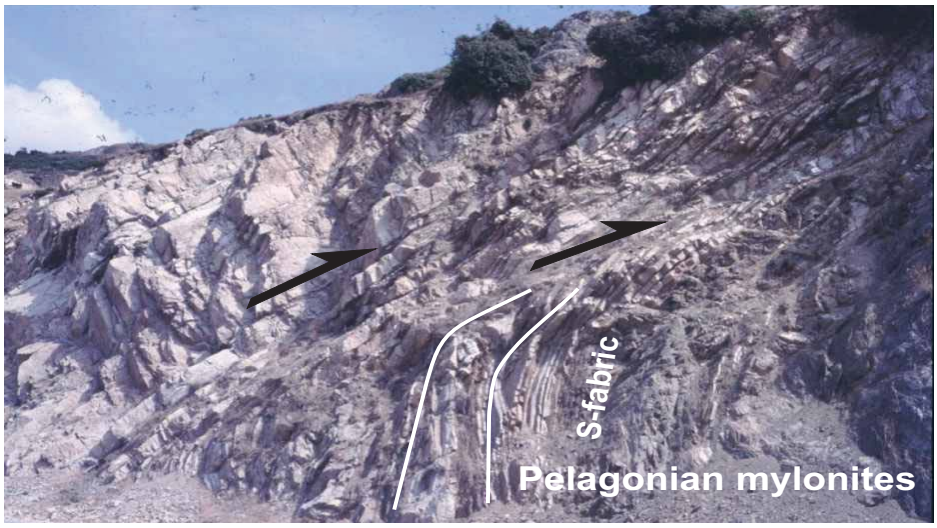


Figure C7
Olympos mylonitic gneisses affected by a thrust fault verging to the SSE. The mylonitic fabric has been dated as Oligocene-Miocene and is related to the detachment of the Pelagonian nappe towards West, dominating at the western side of the Olympos window. It is clearly a ductile deformation that took place in deeper crustal levels just after the tectonic emplacement of the Pelagonian nappe and the high-pressure blue schists Ampelakia unit on the passiv neritic margin of the Olympos carbonate unit. It is equivalent to the Gavrovo zone of the External Hellenides. During the exhumation stages the structures become progressively more brittle, as it is revealed by semi-ductile to brittle shear bands that are imprinted on mylonitic fabric with the same kinematic symmetry top-to-the-West. 40°06'24"N 22°12'18"E

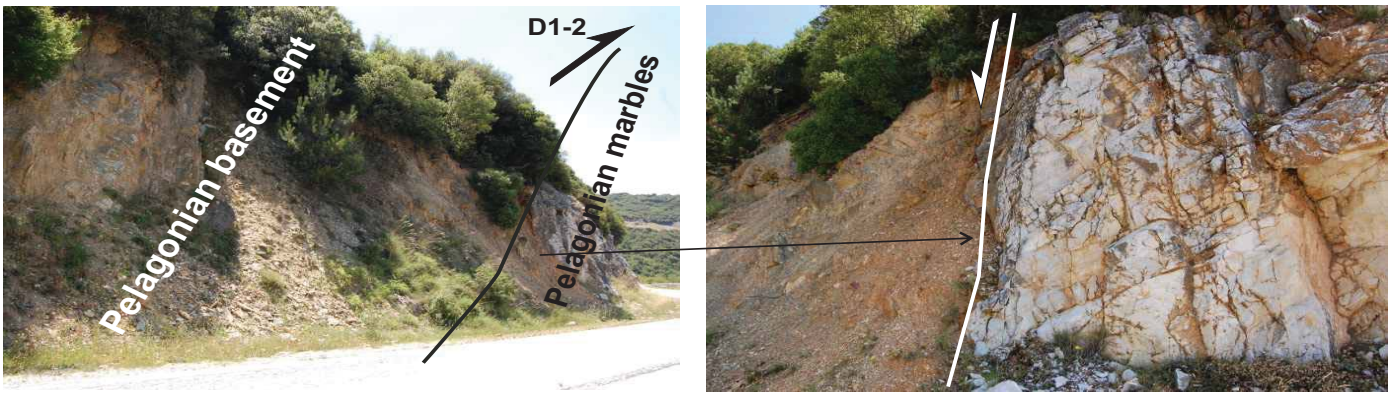


Figure C8
Thrusting of the Pelagonian basement on the Triassic – Jurassic Pelagonian carbonate cover resulting the dublication of the Pelagonian nappe. It is possibly taken place during the Jurassic-Cretaceous time and before the finale emplacement of the Pelagonian nappe on the External Hellenides during the Eocene – Oligocene. The thrust fault itself is reworked of a jounger, high angle dipping, normal fault. 40°06'19"N 22°12'35"E

Figure C9
The Carboniferous granite (~ 300Ma) intruded the Pelagonian basement during the Ercynian orogeny and the Paleo-tethys subduction. The thermal influence on the Pelagonian basement of the granite intrusion, as well as leucocratic granite veins directly at the granite-basement contact are recognized. 40°05'34"N 22°12'38"E

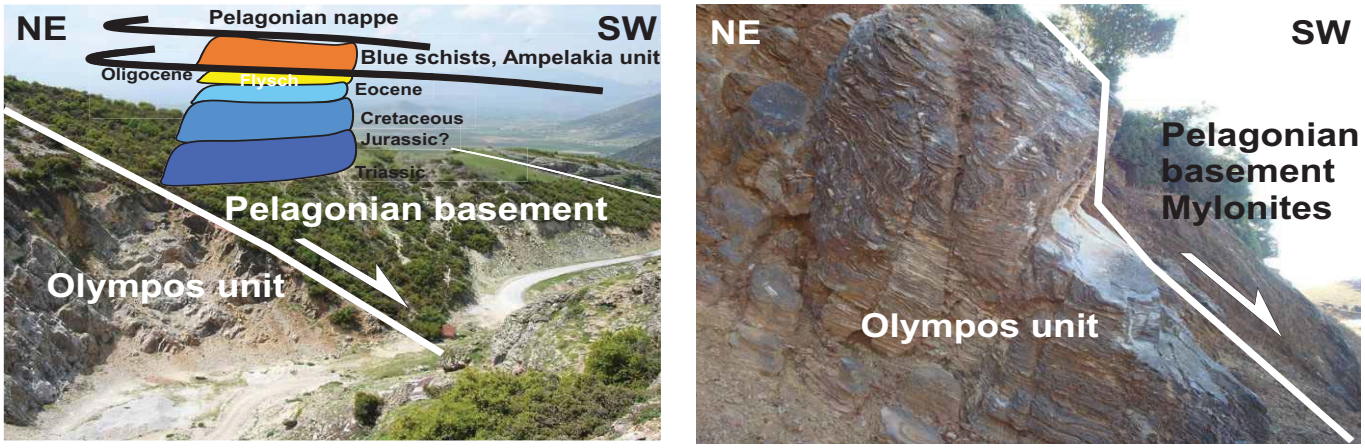


Figure C10
The Olympos normal detachment zone (D5) at the western flank of the Olympos window, between Pelagonian nappe and the Olympos Cretaceous mylonitic marbles, in some places folded during the detachment history by down-dip verging assymetric or sheath folds. Directly at the tectonic contact are formed typical ultra-mylonitic rocks of the Pelagonian basement with the total omission of the entire high-pressure blue schist unit, as well as the Eocene Olympos carbonate layer and the Eocene-Oligocene flysch. Olympos unit/Pelagonian nappe. 40°05'10"N 22°14'59"E, 40°05'14"N 22°14'54"E (sheath)

Figure C11
The Cretaceous carbonate sequence of the Olympos unit directly at the vicinity of the normal detachment fault, along which the previously overthrust the Olympos unit tectonic nappes finally collapsed. The Olympos carbonate rocks are strongly mylonitized forming carbonate mylonitic rocks. A very clear SW-trending stretching lineation is developed along the mylonitic foliation, related to a down-dip SW-ward sense of shear and the tectonic denudation of the nappe pile. 40°05'16"N 22°14'57"E

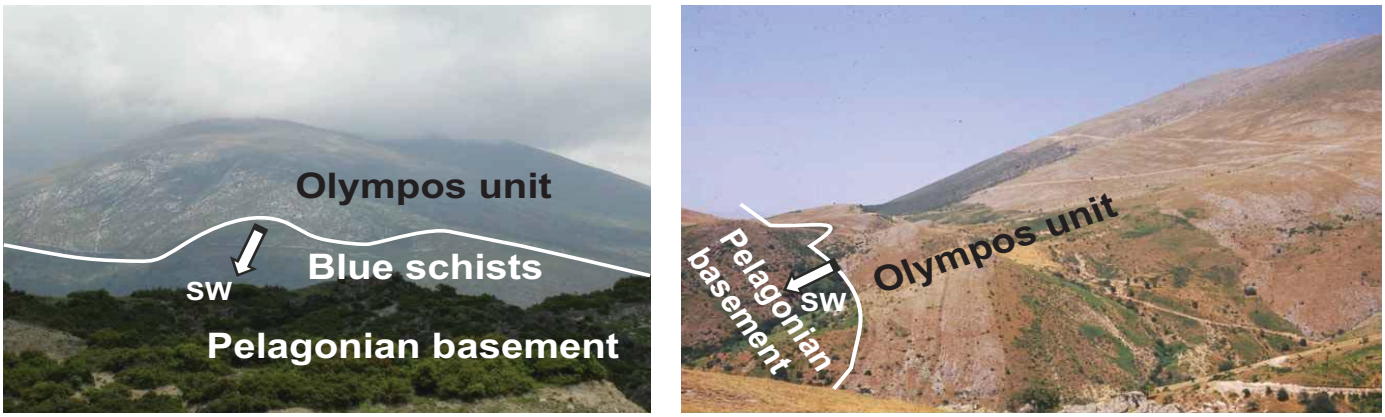


Figure C12
Panoramic view of the tectonic contact between Olympos unit and the Pelagonian nappe together with the blue schists unit. This, initially, thrust zone has been reworked by the Oligocene-Miocene Olympos normal detachment fault (D5) that caused the exhumation of the Olympos tectonic window and the exposure of the External Hellenides lithostratigraphy (Gavrovo zone) under the Internal Hellenides nappe pile architecture. View to the East. 40°01'46"N 22°10'28"E

high-pressure/low-temperature Ampelakia unit, composed of intercalations of basic to intermediate metabasites and metasediments that form a volcanosedimentary sequence (fig. C15). Studies show (Schermer et al. 1990, Kilias et al. 1991, Schermer 1993) that this sequence was initially subducted under the Pelagonian basement during the Paleocene-Eocene, and then thrusted together with the Pelagonian nappe pile on top of the External Hellenides' Olympos-Ossa unit. More specifically, at the particular location before the village Ampelakia, the Ampelakia unit overthrusts the Early Oligocene Olympos-Ossa flysch (fig. C16). The thrust zone here is once more intensively overprinted by a normal detachment fault zone with a downwards top-to-the-SW sense of shear. In addition, from the village Ampelakia, we also have a great panoramic view of the underly-

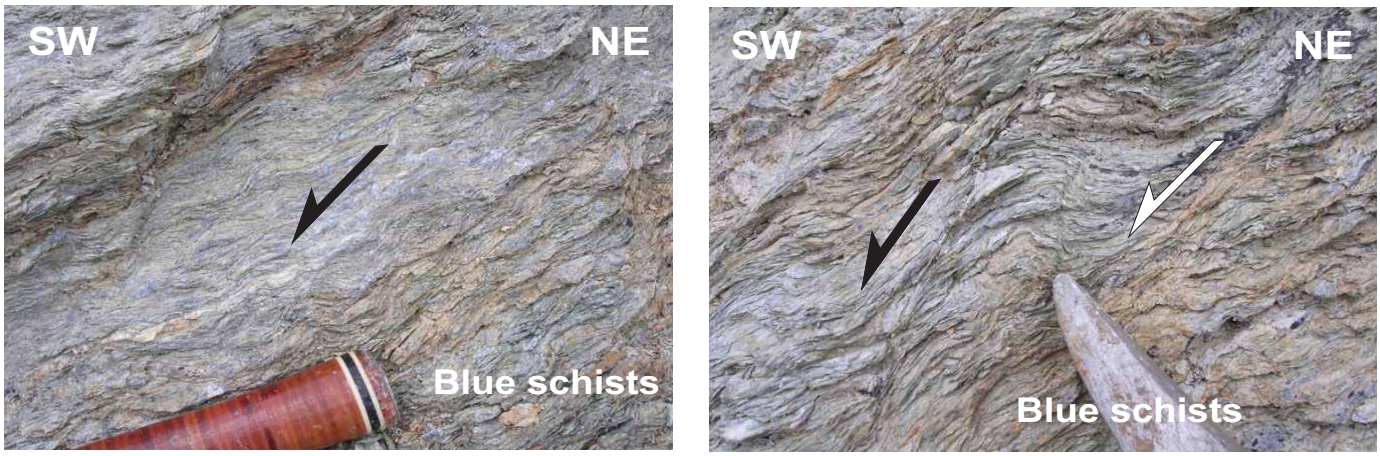


Figure C13
Extensional shear bands related to a top-to-the-SW sense of movement on blue schists, belonging possibly to the deepest parts of the Pelagonian basement affected by the Paleocene-Eocene high pressure metamorphism. The high pressure foliation planes with the syn-tectonic glaucophane growth are well preserved between the shear bands planes, indicating a very rapid exhumation of the blue schists from deep crustal levels to the top. Pelagonian nappe. 39°59'30"N 22°18'27"E

Figure C14
The Triassic-Jurassic Pelagonian platform carbonate cover, with a thickness of about 300m, in contradiction to the same but graetly thinner layer at the eastern Pelagonian margin and its tectonic contact with the NE-ward detached Almopia ophiolites. Two fold events, perpendicular to each another, are also illustrated. 39°50'55"N 22°14'49"E

ing carbonate series of the Olympos-ossa unit till the distant Tembi gorge where Pineos river flows (fig. C17). At this location, the entire high-pressure Ampelakia unit, between the Olympos-Ossa unit and the Pelagonian nappe, is well preserved.

In this section we include two more representative fotos, the first of the Ossa window, equivalent to the Olympos window (fig. C18) and the second of highly sheared schists of the Pelagonian basement western of the Kranea window (fig. C19; kilias et al. 1991, Sfeikos et al.1991).

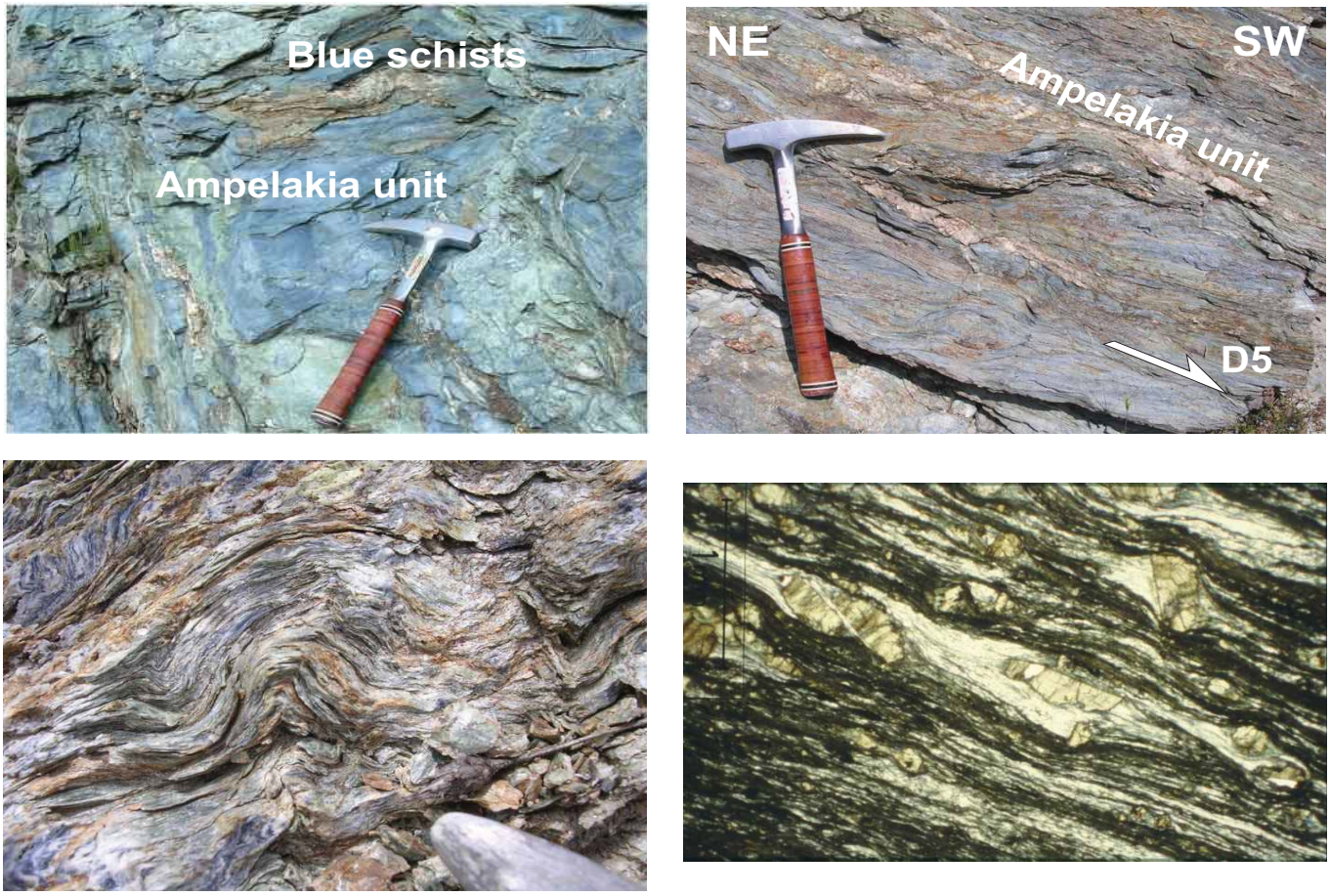


Figure C15
The high-pressure Ampelakia blue schists unit composed of basic to intermediate magmatic products in intercalation with pelitic sediments, as well as carbonate sediments in some cases, forming a typical volcanosedimentary series, however of disputable age until today. Possibly, they belong to the eastward subducted deeper parts of the Pindos zone under the Pelagonian continent. This subduction in the Olympos area has been defined as an A-type subduction of Paleocene-Eocene age. Sense of shear is top-to-the-SW during the Oligocene-Miocene detachment (D5) of the blue schists unit. The high pressure paragenesis is well preserved between the shear zones, related to retrogressive in comparison to the high-pressure event parageneses, containing mainly, new sericite and chlorite crystallization. A rapid exhumation could explain well the preservation of the high-pressure event parageneses. 39°51'29"N 22°32'06"E

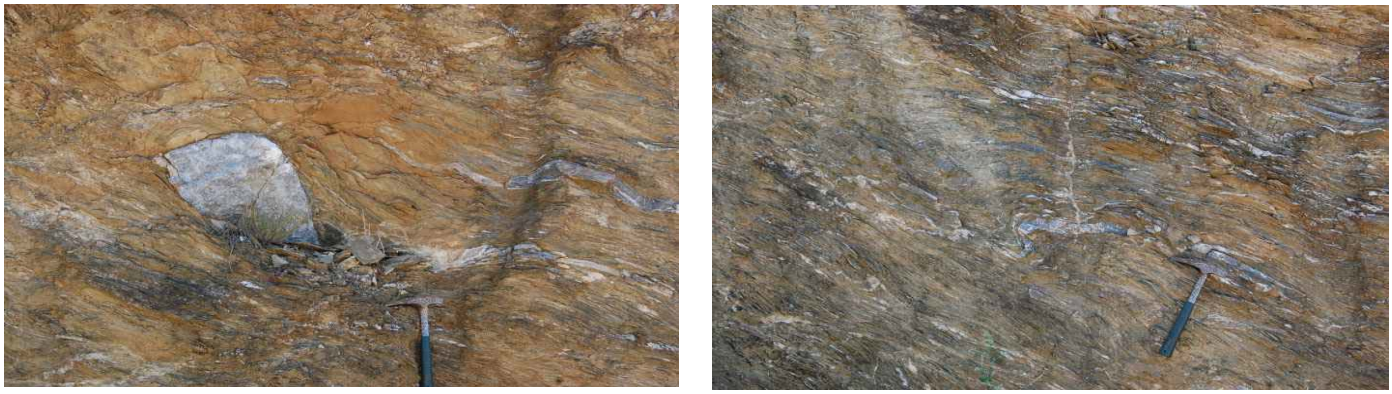


Figure C16
The Early Oligocene flysch of the External Hellenides Olympos – Ossa carbonate formation, underlain the metamorphosed during the Paleocene – Eocene under HP/LT conditions Ampelakia unit. 39°51'13"N 22°32'51"E

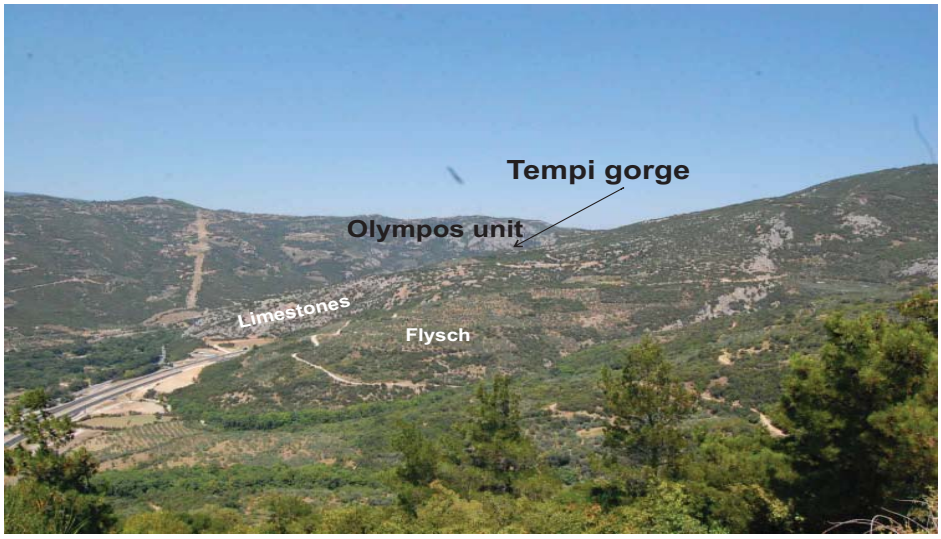


Figure C17
Panoramic view of the Tempi gorge composed by the carbonate Olympos – Ossa unit. View to the East. 39°51'24"N 22°32'29"E

Figure C18
The Ossa window further to the South, equivalent to the Olympos window. The Ossa carbonate unit shows the same structural evolution with the Olympos unit but without such a detailed stratigraphical column as this of the Olympos unit, due to its higher metamorphic path history. View to the East. 39°45'48"N 22°22'00"E

Figure C19
Highly sheared top-to-the-SW Pelagonian basement mylonites further to the west of the Olympos window and more specifically at the western flank of the Kranea window (fig. 10) that is also equivalent to the Olympos one. 39°56'55"N 21°54'25"E

D. 4th Day: Internal Hellenides (fig. 41)

I. Axios/Vardar zone (Almopia subzone)
II. Eastern margin of the Pelagonian nappe and Pelagonian nappe
III. Late Cretaceous-Paleocene clastic-carbonate sequences

Trek: From Thessaloniki to the city of Veria, and then to the villages of Rizomata, Daskio and Akrini (fig. 42).

Overnight staying in Siatista.

On the forth day, we initially reach the city of Veria from Thessaloniki via the Egnatia highway. From there, we take the mountain provincial asphalt road to the villages of Rizomata and Daskio in the Pieria Mountain until we reach Aliakmon River and the lake Polyphytos. After crossing Aliakmon River, we follow the local asphalt road leading to the village of Akrini at the foot of Vermion Mountain.

Our cross-section today starts after the city of

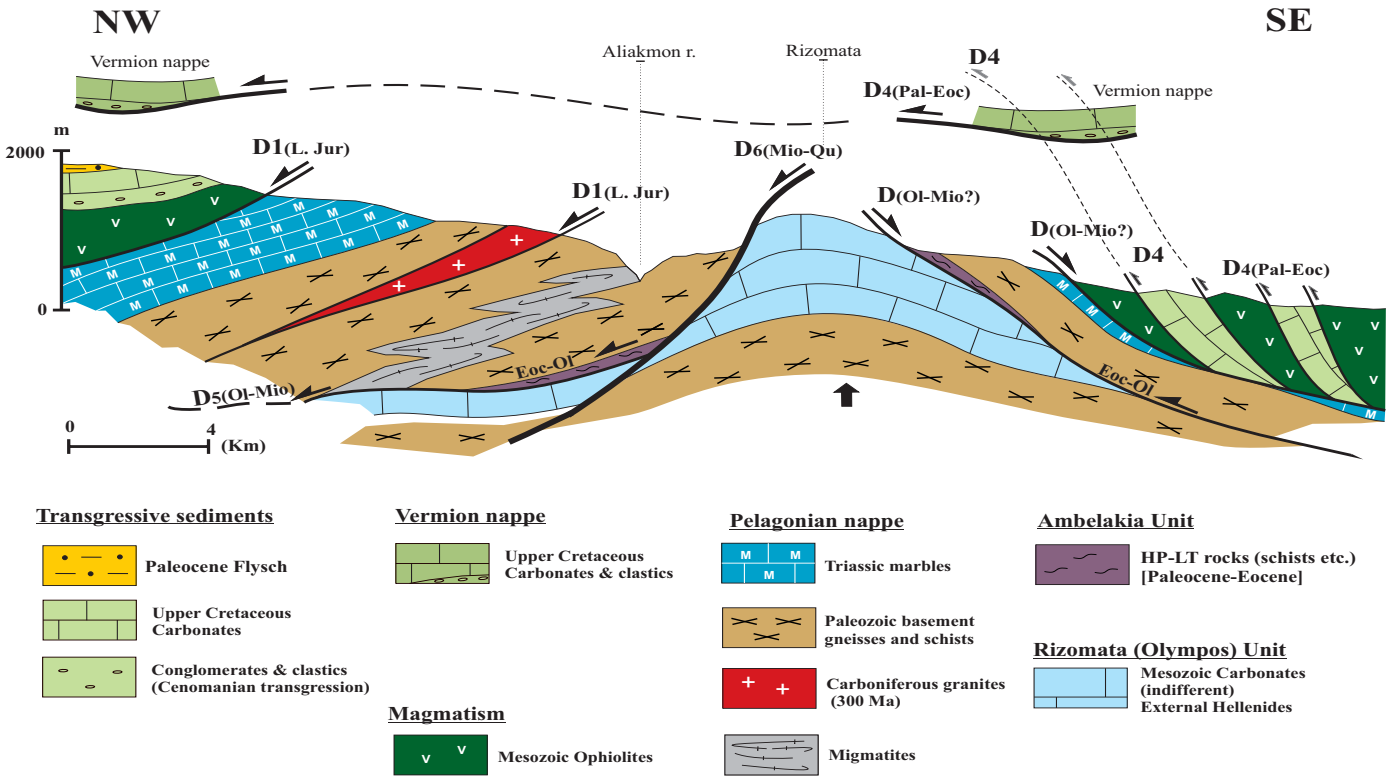


Figure 41
Schematic geological cross-section through the selected area/trek of the 4th day in the Axios/Vardar zone and the eastern part of the Pelagonian nappe, as well as the Olympos-Ossa unit.

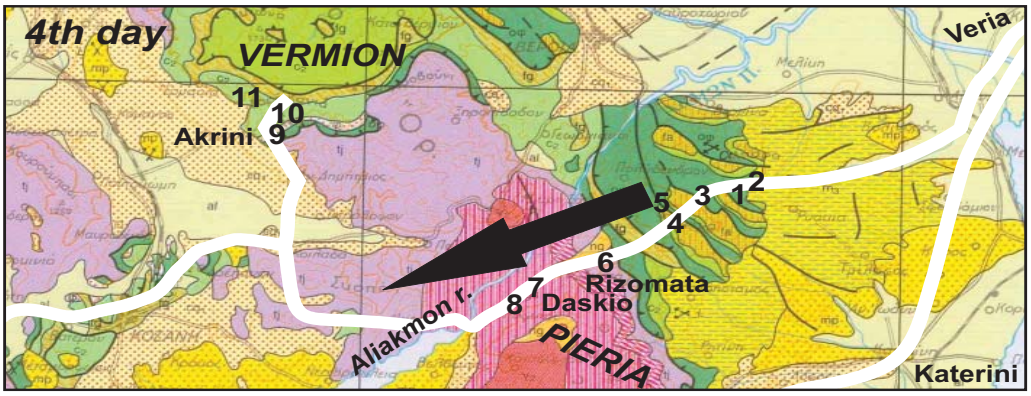


Figure 42
Location of the selected outcrops and stops are shown on the corresponding trek of the 4th day.

Veria, with the first stop located at the beginning of the mountain road towards the village Rizomata. From the east to the west, we cross the geological units of the western Almopia subzone, consisted of ophiolite mélanges and ophiolites (fig. D1, D2, D3). They, together with the Late Cretaceous (i.e. Cenomanian to Maistrichtian) neritic limestones and the Paleocene flysch of the Internal Hellenides, appear intensively imbricated (fig. D4, D5). The Almopia's units are thrusted over the Triassic-Jurassic carbonate cover of the Pelagonian basement towards the SW. However, this thrust zone has also been intensively overprinted by the earlier described Oligocene-Mio-

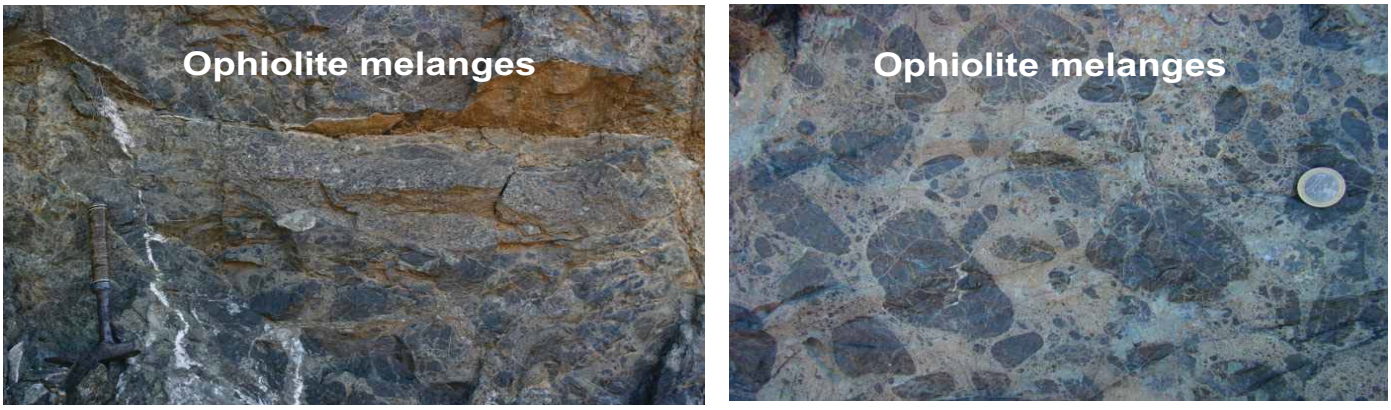


Figure D1
Ophiolite mélange, composed by ophiolite pebbles and ophiolite psamitic material. It has been deposited in a basin formed at the front of the obducted Almopia ophiolites on the eastern Pelagonian continental margin. 40°29'09"N 22°15'38"E

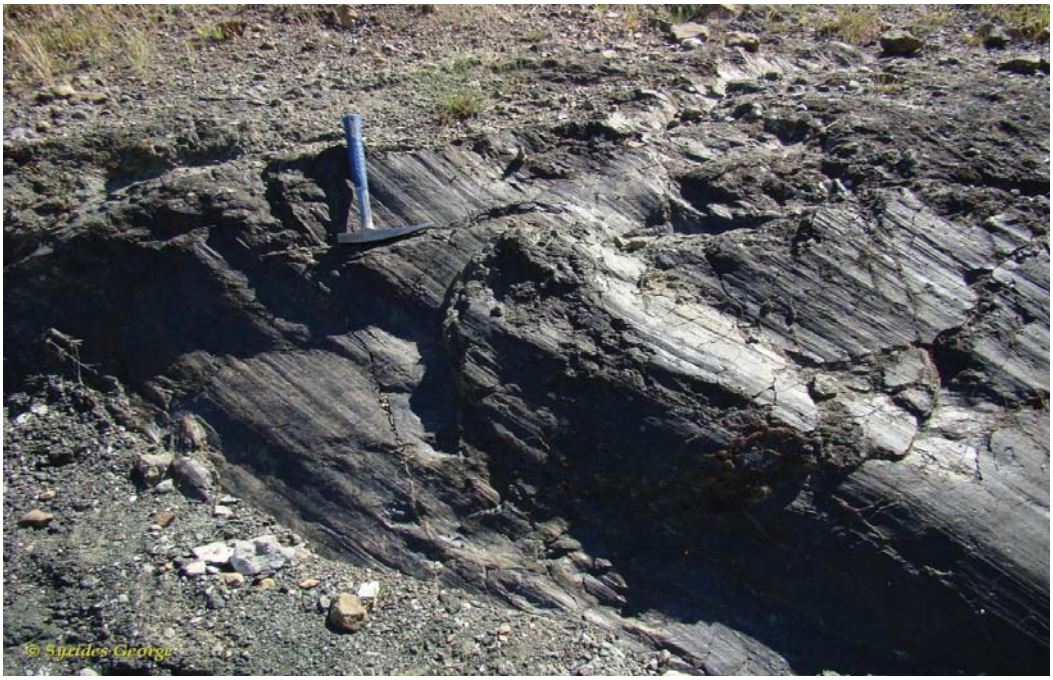


Figure D2
Low angle normal, ca. dip slip, fault (D5 event), affecting the Almopia ophiolite mélanges. It is related to the Oligocene-Miocene tectonic detachment of the Pelagonian nappe pile, before the development of the high angle normal Neogene faults and the formation of the intracontinental Neogene basins (e.g. the Axios Neogene basin). The striation lines and Riedl structures are clearly illustrated, showing the normal down-dip movement along the fault plane. A curvature of the fault plane parallel to the fault striation is also visible. Axios zone, Almopia subzone. 40°29'09"N 22°15'38"E

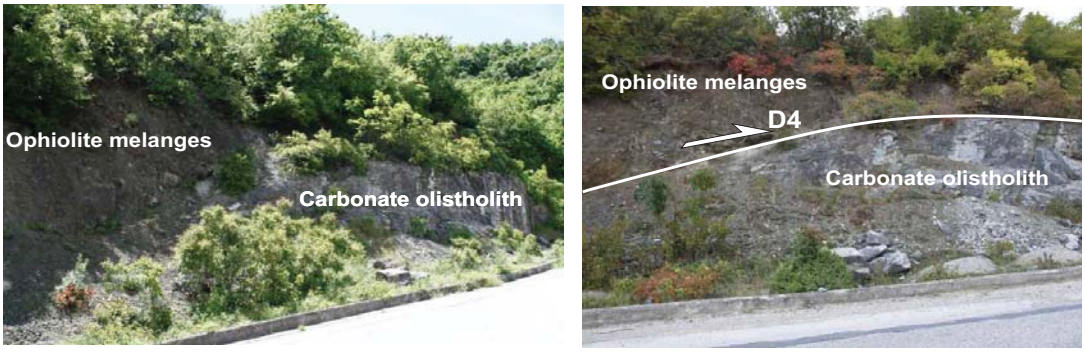


Figure D3
Ophiolite mélange of the Almopia ophiolite belt with a marble olistolith incorporated in the ophiolite mélange during the ophiolite obduction. It corresponds to an olistolith of a deep-sea carbonate origin. 40°27'43"N 22°15'35"E

Figure D4
The upper Cretaceous-Paleocene flysch of the Internal Hellenides on top of the Upper Cretaceous (Cenomanian to Maistrichtian) transgressive, neritic, carbonate sediments above the obducted ophiolites. This flysch phase appears to be rich in carbonate material. 40°28'55"N 22°15'20"E

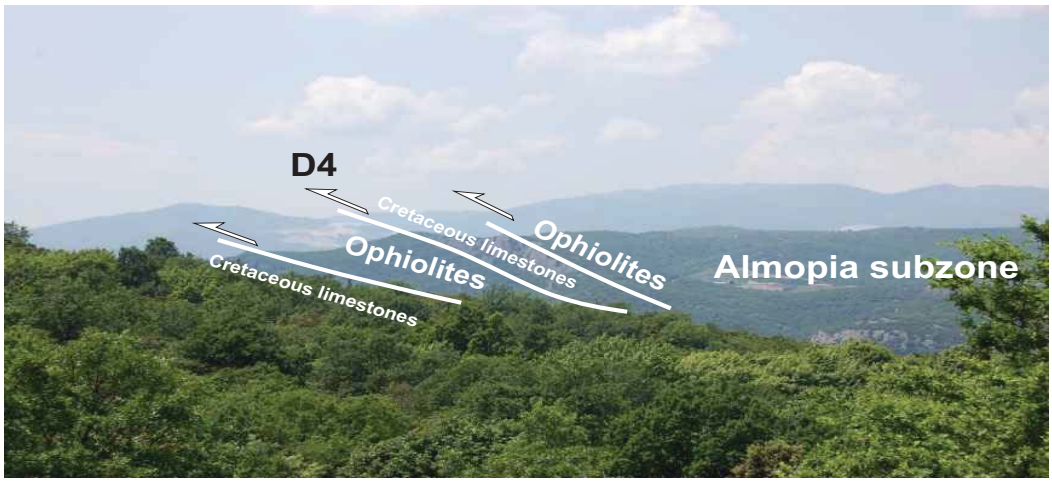


Figure D5
Panoramic view of typical Almopian slices of Paleocene-Eocene age (D4 event) between the Upper Cretaceous limestones and the Almopia ophiolites. Axios zone, Almopia subzone. View to the Nprth. 40°28'56"N 22°15'21"E

cene normal detachment fault (see 3rd day, tectonic window of Mount Olympos), causing the tectonic denudation of all the Almopia's rock sequences towards NE. At the village of Rizomata, equivalent to the Olympos-Ossa unit series of the External Hellenides are exhumed from below the internal high-pressure blue-schist belt (Ampelakia unit) and the Pelagonian nappe pile, in a small tectonic window (i.e. the Rizomata window; Kilas & Mountrakis 1985; fig. D6).

All the way from the village Rizomata to the village Daskion, and until the Aliakmon River and the Polyphyto lake, we ceaselessly see the Pelagonian basement, composed of Palaeozoic or older gneisses and schists, Ercynian granites (fig. D7), and in places migmatites of Cretaceous age (fig. D8; Schenker et al. 2014, 2015). The Pelagonian basement has been intensively affected by folding and syn-metamorphic thrusting of Late Jurassic – Early Cretaceous age (Yarvood & Aftalion 1976, Kilias et al. 2010). From the Aliakmon river until the village of Akrini, the Pelagonian recrystallized Triassic-Jurassic carbonate cover dominates, with a thickness of up to 300 m. From the village of

Figure D6
The small carbonate Rizomata tectonic window equivalent to the Olympos window. The Rizomata thin-bedded carbonate rocks are placed below the Pelagonian basement rocks, wich show in this area a high-pressure metamorphism of Paleocene-Eocene age overprinted an older amphibolites facies Upper Jurassic-Lower Cretaceous metamorphism. Green amphibole crystals are recrystallized to blue Na-amphibole of glaucophane composition. A high angle normal fault cross-cut the Rizomata carbonate rocks causing their uplift and progressive exhumation of the footwall fault segment. 40°20'54"N 22°12'26"E, 40°21'21"N 22°13'40"E

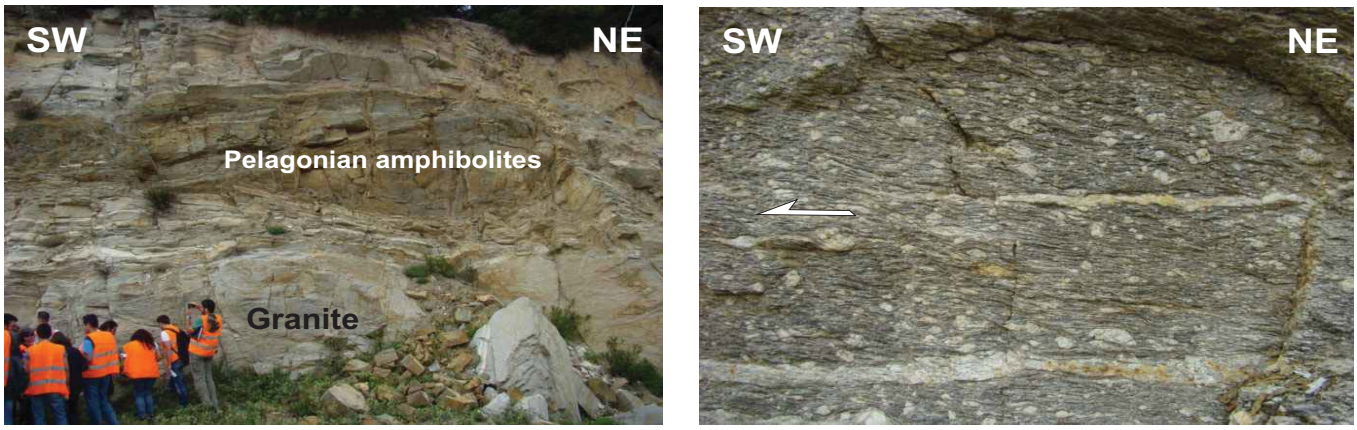


Figure D7
Carboniferous granitoid intrusions (~ 300Ma) into the Pelagonian Paleozoic or older basement. They are calk-alkaline granites related to the Ercysian orogeny and the subduction of the Paleotethys under the single Pelagonian/Serbo-Macedonian/Rhodope continental plate. Sense of shear top-to-the-SW, related to the Alpine orogeny, as clearly indicated of the σ- and δ- feldspars clasts of the augen granite- gneisses. 40°19'43"N 22°08'34"E

Akrini and all along the mountain asphalt road to the small village of Kato Vermio, we see outcrops of the Pelagonian Triassic-Jurassic carbonate cover overthrusted by the Almopia's ophiolites. The latter are here mainly composed of very thin, up to few meters thick (ca 50 m), serpentinites, possibly due to the post-emplacement extension (fig. D9). On top of the ophiolites, rest transgressively, with a clear angular unconformity, the Late Cretaceous clastic and carbonate rocks that finalize in a Paleocene flysch (fig. D10, D11). A younger, low-angle normal detachment fault related to a SW-ward sense of movement is recognized in some places between the Almopia's ophiolites and the overlying Late Cretaceous sequence. Last, on the Paleocene flysch, rests tectonically the Vermion nappe also composed of clastic and neritic carbonate rocks of Late Cretaceous age.

Figure D8
Pelagonian migmatites of late Early Cretaceous age (~120 -110Ma). They are related to the D2 compressional event affected the Pelagonian nappe and the pre-Upper Cretaceous units of the Axios zone during the collision of the Pelagonian with the Rhodope and Eurasian margin. 40°19'35"N 22°06'25"E

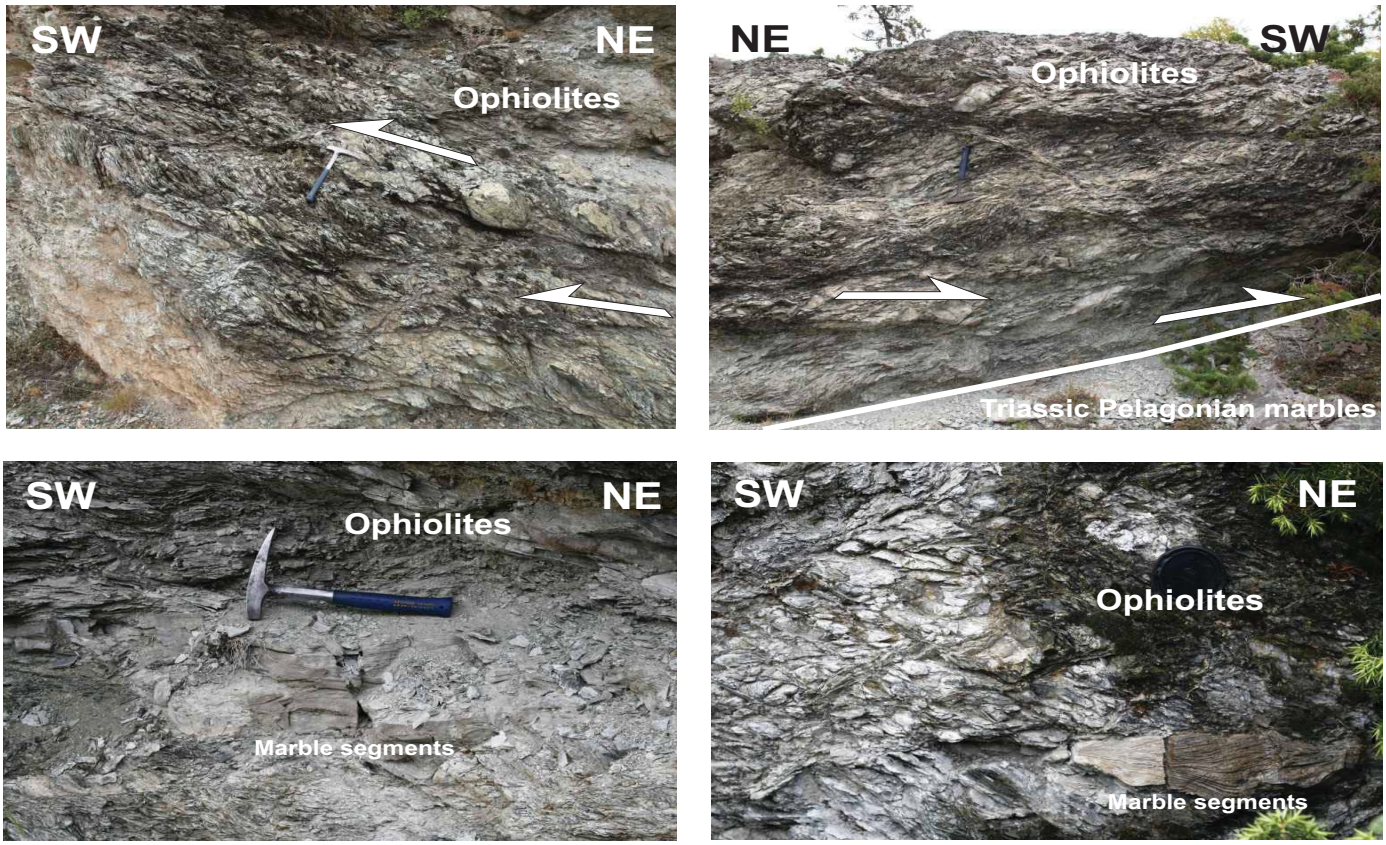


Figure D9
The Almopia ophiolites, composed here mainly of serpentinites, emplaced on the Triassic marbles of the eastern Pelagonian margin. Direclty at this tectonic contact small folded fragments of the underlying Triassic Pelagonian marbles are tectonically intercalated within the ophiolites, due to the Late Jurassic obduction. Here the thickness of the whole ophiolite sequence has been intensively reduced to only about 50m, possibly due to a post-emplacement extension, related to a very well exposed top-to-the-SW sense of movement. The latter is clearly indicated by S-C structures and shear bands, as well as σ-clasts. Axios zone, Almopia subzone. 40°26'47"N 21°55'08"E

Figure D10
Upper Cretaceous clastic and neritic carbonate sediments transgressively overlying the Almopia ophiolites. The alternation of clastic and neritic carbonate layers at the base of the stratigraphic column indicate the paleo-environmental conditions during the initial stages of the sedimentation processes. The whole carbonate sequence terminated in a late Upper Cretaceous-Paleocene flysch. Eroded ophiolite material forms usually the binding matrix of the basal conglomerate material. Axios zone, Almopia subzone. 40°27'11"N 21°55'14"E

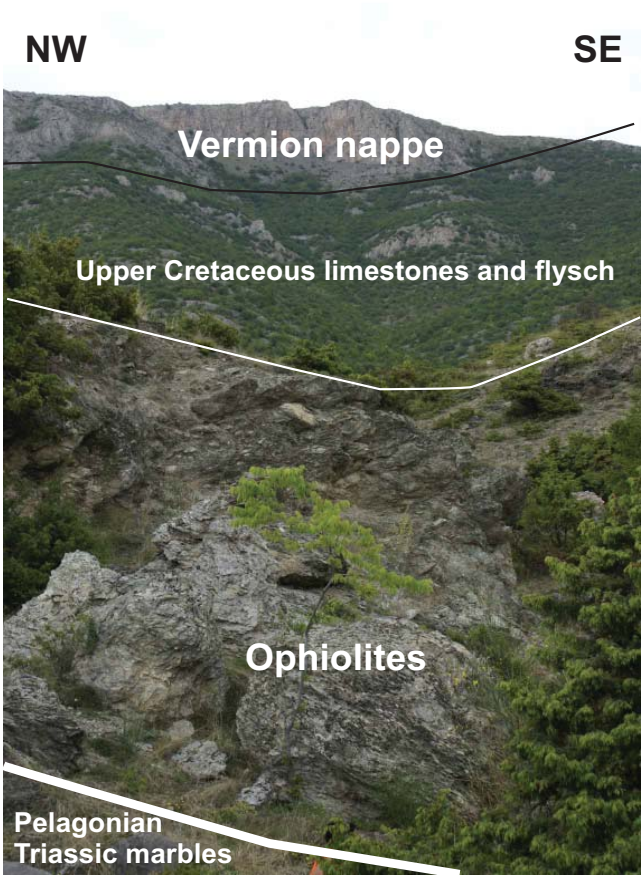


Figure D11
Panoramic view of the structural architecture of the Eastern Pelagonian margin and the Western Axios Almopian units. View to the Northeast. 40°26'47"N 21°55'08"E

E. 5th Day: Internal Hellenides (fig. 43)

I. Western margin of the Pelagonian nappe and Pelagonian nappe

II. Ophiolites and associated sediments on top of the western Pelagonian margin (Sub-Pelagonian zone)

III. Late Jurassic – Early Cretaceous sediments on top of the ophiolites

Trek: Starting from the town of Siatista, our course passes from the city of Grevena, the villages of Mikrokleisoura and Poros, and the town of Deskati, and finally ends at the peak of Vounasa Mountain (fig. 44).

Overnight staying in the town of Siatista.

Leaving the town of Siatista, we initially reach the city of Grevena via the Egnatia highway, and from there we take the provincial asphalt road "Grevena-Knidi" until the small villages of Poros and Mikrokleisoura. From Mikrokleisoura we turn back to Grevena, and via the provincial road "Grevena-Kalampaka" we reach the villages of Paliouria and Panagia near the Zavordas Monastery, and finally the town of Deskati from where we go up to the top of Vounasa Mountain (i.e. 1600m, above sea level). Along this traverse we study the western Pelagonian margin, which appears to be equivalent to the eastern one.

Our cross-section begins right at the exit as we leave the town of Siatista, where we see a typical ophiolite mélanges occurrence imbricated with the Triassic-Jurassic platform carbonate of the western Pelagonian margin. The ophiolite mélanges comprise carbonate olistoliths, clasts of serpentinites and lavas, as well as radiolarian cherts and shales (fig. E1). A top-to-the-west semi-ductile sense of movement dominates in the whole ophiolite mélanges sequence (fig. E1), despite it is developed at the western Pelagonian margin, clearly showing an ophiolite origin from the east. These ophiolite mélanges are part of the main ophiolite realm of Vourinos Mountain occurring further to the south along the western Pelagonian region.

Next stop is near the small village of Mikrokleisoura, where an ophiolite body is exhumed from beneath the Oligocene-Miocene molassic sediments of the Meso-Hellenic Trough (MHT). It

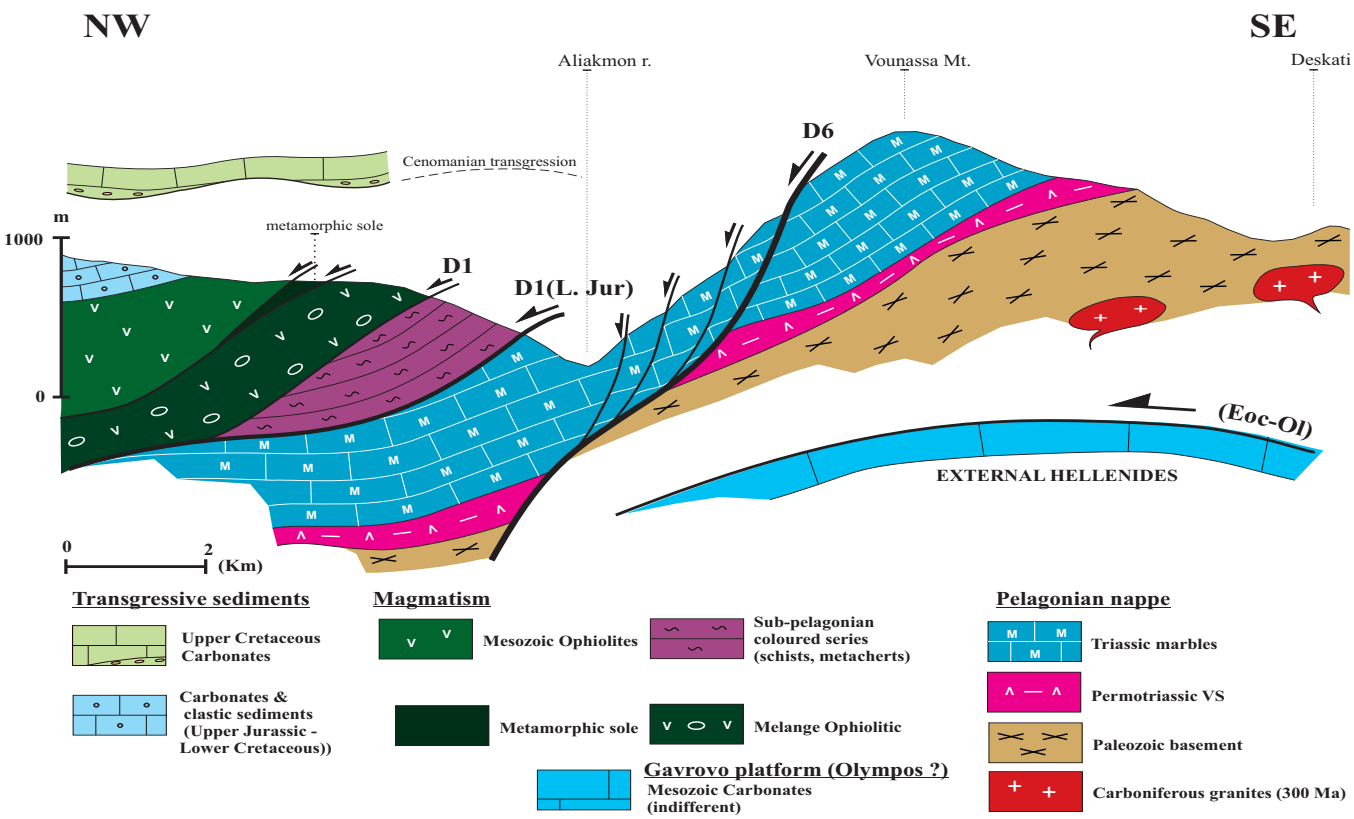
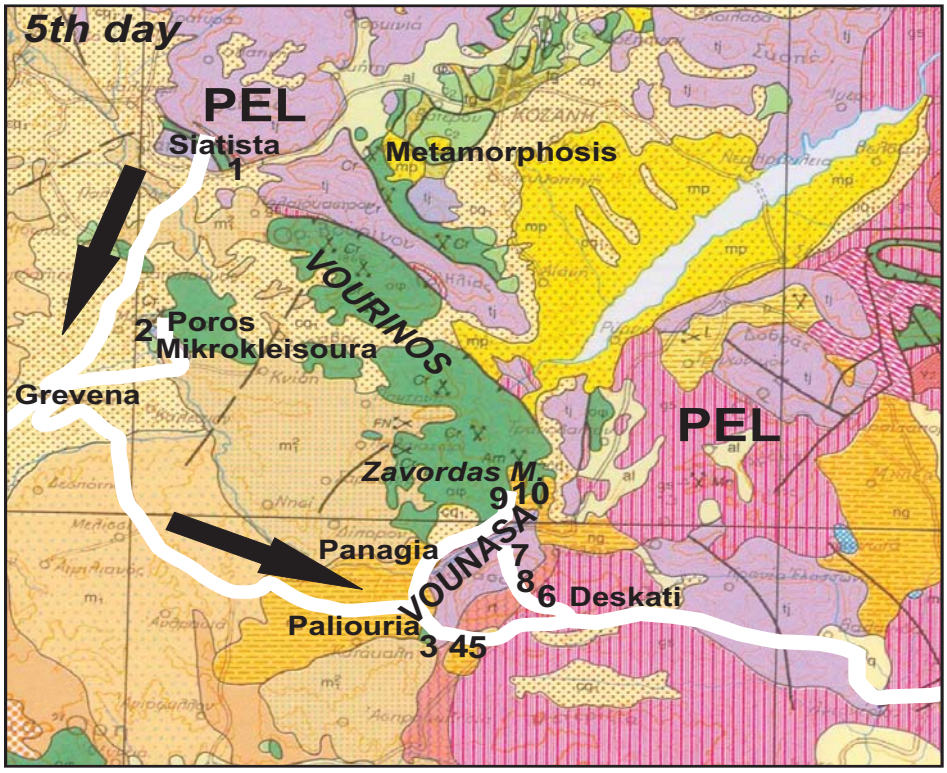


Figure 43
Schematic geological cross-section through the selected area/trek of the 5th day in the western part of the Pelagonian nappe and the ?Sub-Pelagonian zone.

Figure 44
Location of the selected outcrops and stops are shown on the corresponding trek of the 5th day.



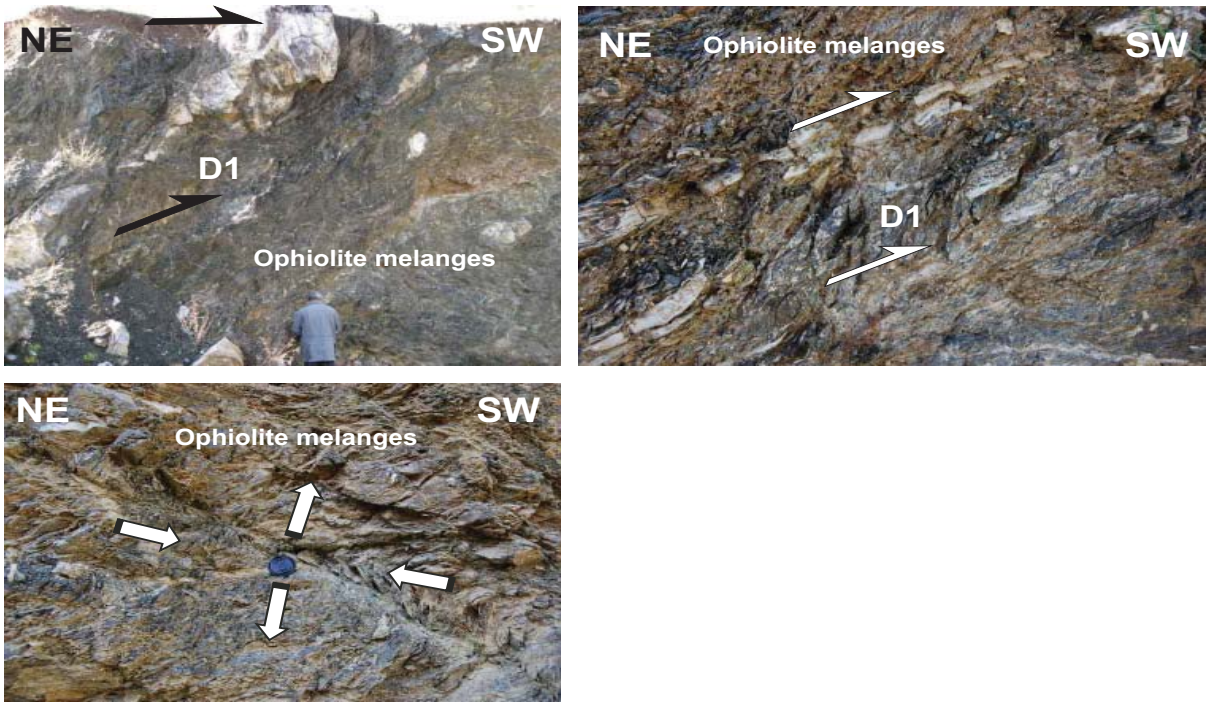


Figure E1
Mid-Late Jurassic Siatista mélanges. A sense of movement top-to-the-W is clearly indicated by S-C fabric and carbonate σ-clasts asymmetry. This ductile-semiductile movement of the ophiolites on the western edge of the Pelagonian nappe, possibly related to the ophiolites emplacement on the Pelagonian nappe, indicates the ophiolites` origin from a single ocean realm east of the Pelagonian continental margin. (?Sub-Pelagonian zone). The Siatista mélanges are mainly composed of neritic carbonate olistoliths, radiolarian cherts and serpentinites. They continue to the south at the Vourinos ophiolite complex that rests on the Triassic-Jurassic marbles` cover of the western Pelagonian flank. 40°15'28"N 21°33'19"E

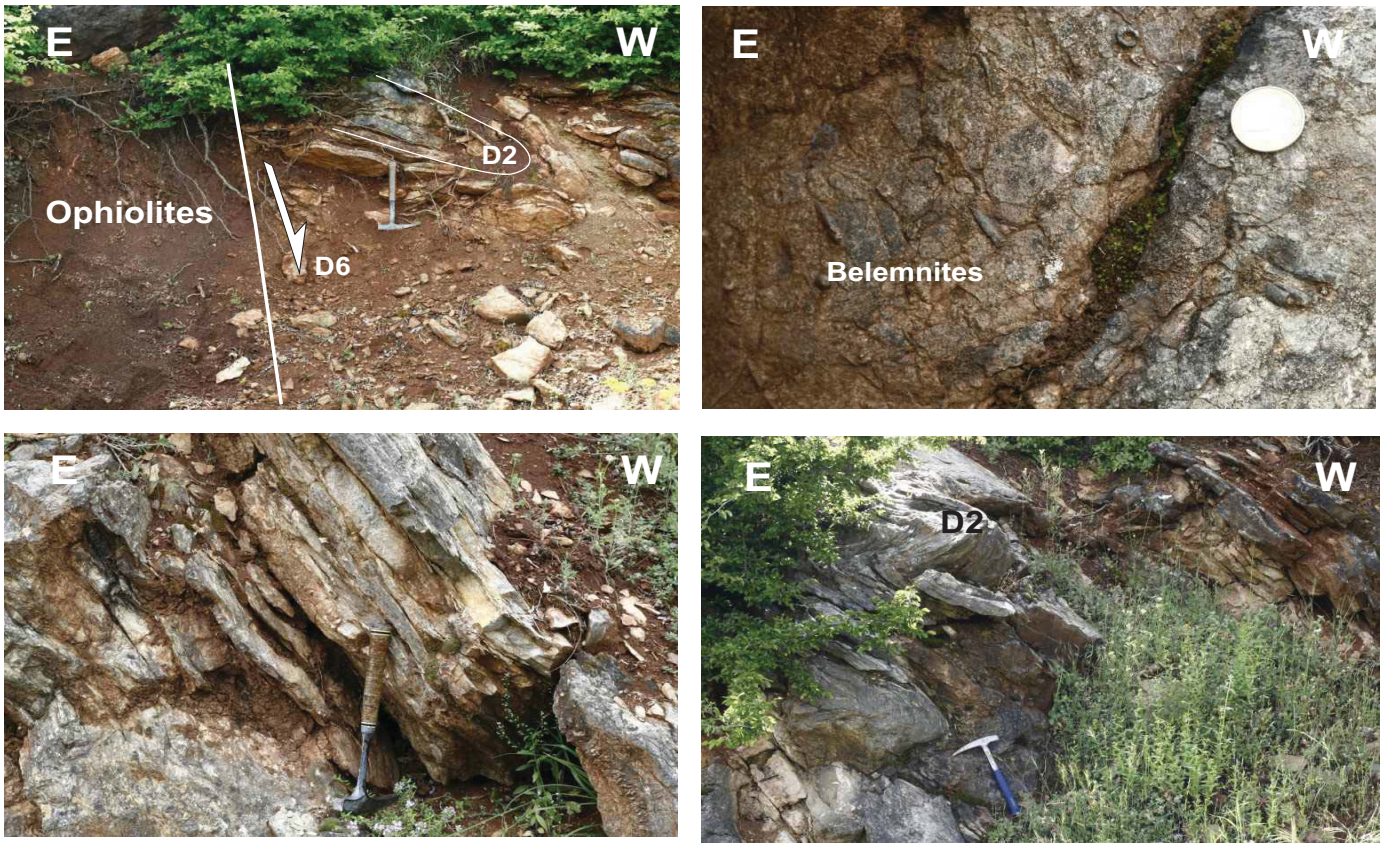


Figure E2
The transgressive clastic and carbonate semi-pelagic sediments of Upper Jurassic-Late Cretaceous age. They are at the top of the Vourinos ophiolites complex, overlying the western Pelagonian flank. Calpionelle bearing turbidite layers and clastic sediments with small belemnites are distinguished. Furthermore, two folding events are visible; the older is associated with subisoclinal, asymmetrical folds and is related to the late Early Cretaceous compressional D2-event, while the younger is characterized by open to tight about symmetrical folds, possibly of Paleocene-Eocene age (D4 event) and clearly overprints the former D2 structures. These Upper Jurassic-Late Cretaceous sediments on the top of the ophiolite belt have the same sedimentary features and age both in the Axios/Vardar ophiolite belt and in the ?Sub-Pelagonian (Vourinos) ophiolite belt, indicating at least about the same time of the ophiolites obduction on the eastern and the western Pelagonian parts. In this case the ophiolites belts are possibly, with a main W-ward emplacement direction, originated from a single Neotethyan ocean, eastern of the Pelagonian continent, regarded as the eastern part of the Apulian plate. 40°07'23"N 21°31'39"E

belongs to the ophiolite belt that was emplaced on the recrystallized Triassic-Jurassic limestones of the western Pelagonian margin. On top of this ophiolite body rest the Late Jurassic – Early Cretaceous Belemnite-bearing clastic sediments, alternating with thin layers of Calpionella-bearing limestones, identical to those recorded on top of the ophiolites emplaced at the eastern Pelagonian margin and in the Axios/Vardar zone (fig. E2). Moreover, Late Cretaceous shallow-water limestones occur higher in the column, lying transgressively over the ophiolites and the Late Jurassic – Early Cretaceous clastic sediments.

While continuing our planned route on the provincial road from Grevena to Deskati, we cross the molassic sediments of the MHT all along the way until the village of Paliouria, where their tectonic contact with the Triassic-Jurassic carbonate sediments of the western Pelagonian margin occurs. After the Palouria village, on a left bend of the road to Deskati, we see a multicolored rock series. They consist of various colors radiolarian cherts, tuffs, shales, red marbles, sipoline marbles and quartzites, intensively folded. They are clearly deep-sea sediments, associated with the obducted ophiolites on the Triassic-Jurassic platform carbonate sequence at the western Pelagonian margin, and they are possibly parts of the deep-sea deposits of the Sub-Pelagonian zone (fig. E3, E4, E5).

Further along on the road to Deskati, we cross the Late Paleozoic part (Carboniferous-?Devonian) of the Pelagonian basement rocks, composed

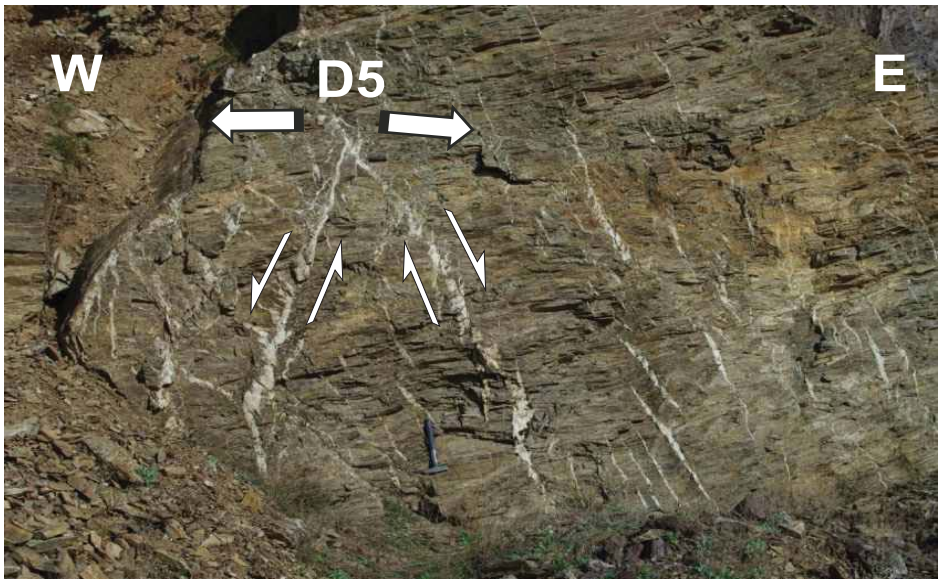


Figure E3
Greenschists and deep-water sediments of the Sub-Pelagonian zone with conjugate set of tension gashes related to a vertical maximum σ1-axis and a subhorizontal σ3-axis. A dynamic that coincides with the Oligocene-Miocene extension (D5 event) and the carbonate Olympos and Kranea units exhumation under the detached blue schists unit and the Pelagonian nappe pile. Sub-Pelagonian zone. 39°55'08"N 21°44'11"E

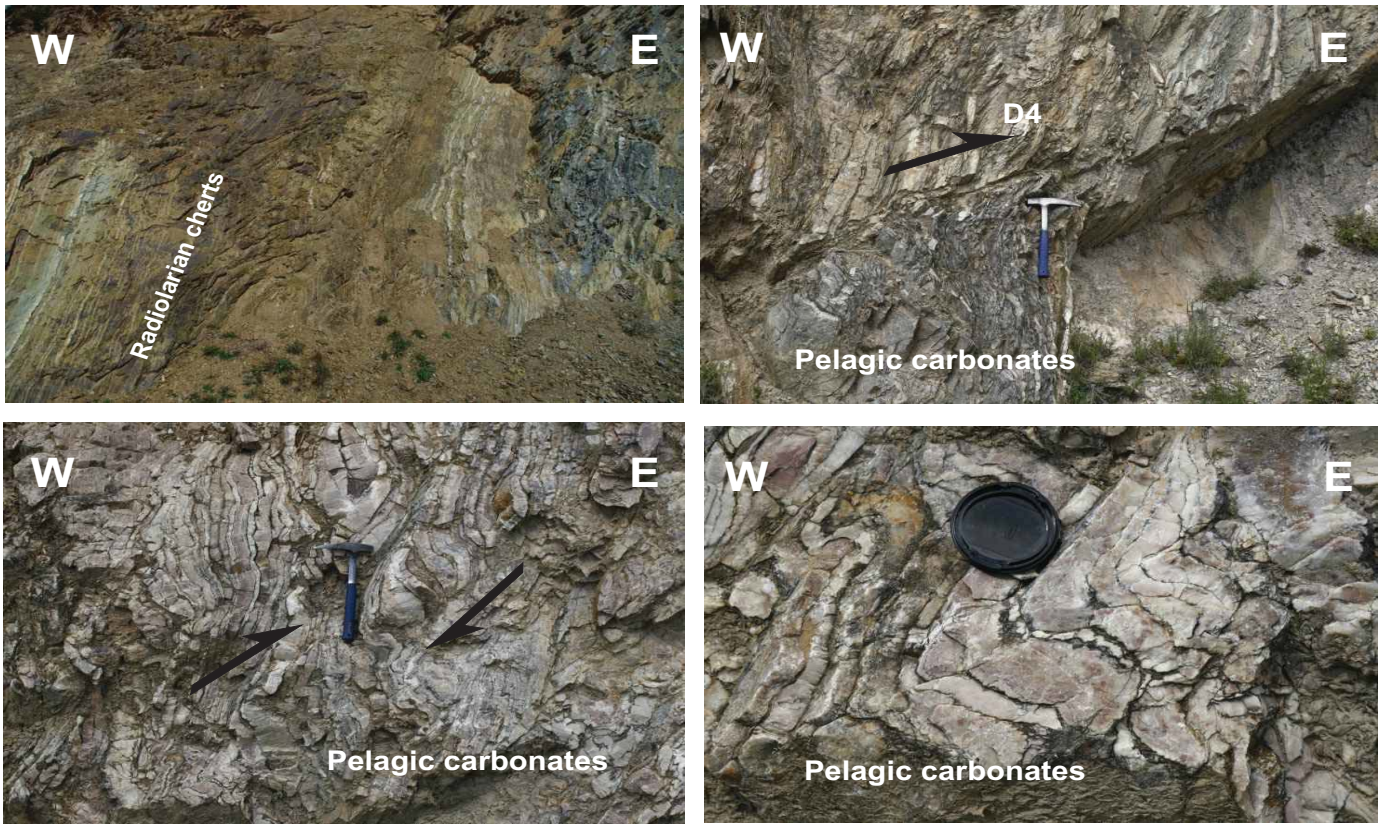


Figure E4
Multicolored deep-water sediments composed by greenschists, red and green colored radiolarian cherts, pelagic red carbonates and metapelite rocks, possibly of Jurassic age. They show low grade metamorphism but they are intensively deformed by compressional structures (e.g. isoclinal folds, reverse shear bands, and brittle thrust faults), as well as extensional structures (e.g. down-dip shear bands, S-C fabric and semiductile normal fault zones). These deep-sea sediments were possibly together with the ophiolite belt overthrust the neritic Triassic-Jurassic carbonate rocks of the western Pelagonian flank during the Late Jurassic. Today they are separated from the Pelagonian Paleozoic basement by a great Neogene-Quaternary normal fault with the total omission of the Triassic-Jurassic Pelagonian marbles. The sense of movement during their initial emplacement is not clear here due to the intensive multi-phase deformation, affected progressively these rocks. 39°55'08"N 21°44'11"E

of mica and amphibole schists, as well as schist gneisses intruded by Ercynian calc-alkaline granitoids. The same Pelagonian basement rocks occur along the sealed road from Deskati to the Vounasa Mountain peak (fig. E6). The top of Vounasa Mountain is dominated by the Triassic-Jurassic platform carbonate sequence of the western Pelagonian margin that lies on the Paleozoic Pelagonian basement (fig. E7). Between the Triassic-Jurassic carbonate cover and the Paleozoic basement rocks, the Permo-Triassic volcanoclastic products crop out (fig. E8), which are related to the Pangaea rifting and the further development of the Neotethyan ocean/-s. A bimodal-type magmatism characterizes this initial stage of the Pangaea rifting, which can be well observed at this Permo-triassic contact in Vounasa Mountain, but also in general at the contact between the Triassic-Jurassic Pelagonian carbonate cover and its Paleozoic crystalline substratum.

Last, near the Panagia village and the Zavorda Monastery on the provincial asphalt road from village Panagia to the town of Siatista, we observe a part of the Mid-Jurassic amphibolite-sole between serpentinized mantle material and ophiolite mélanges (fig. E9). This particular occurrence of

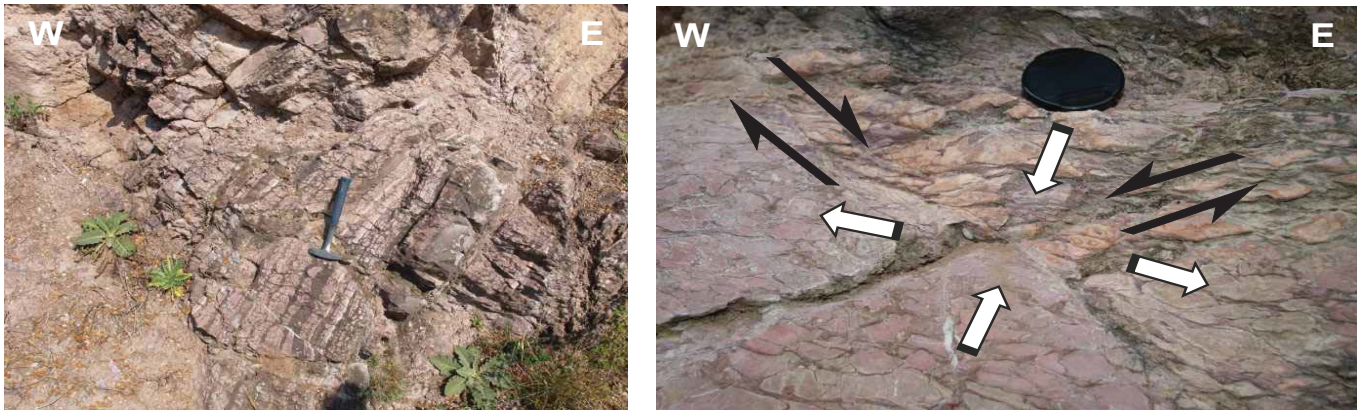


Figure E5
Red carbonates sediments, lying under the previous multicolored metasediments. They represent a deep-sea basin environment and show a possibly sedimentary brecciation, as well as a turbidite layering. Their contact with the underlying Paleozoic Pelagonian schists is a Neogene-Quaternary normal fault zone. The age of this sequence remains under debate. They should most likely be of Jurassic age equivalent to the previously ones described multicolored deep-water sediments of the ?Sub-Pelagonian zone, having been overthrusted the neritic Triassic-Jurassic Pelagonian platform carbonate cover. 39°54'52"N 21°44'34"E

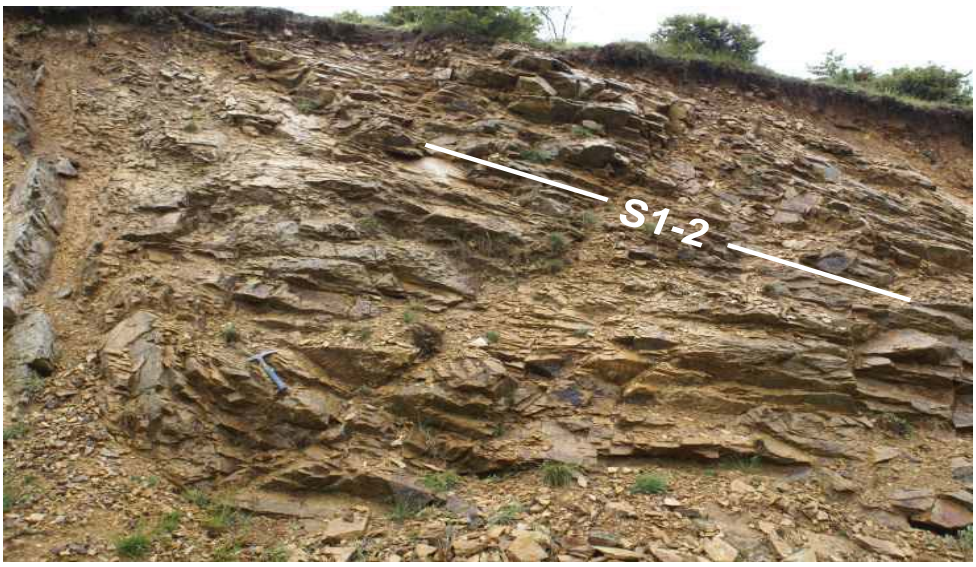


Figure E6
Paleozoic schists (Silurian-Devonian?) of the Pelagonian basement composed by intercalations of metasediments and metamagmatic products. They forme the upper part of the Pelagonian Paleozoic basement. The Carboniferous Pelagonian granites (~300Ma) intrude also this schist's sequence. 39°57'14"N 21°48'05"E

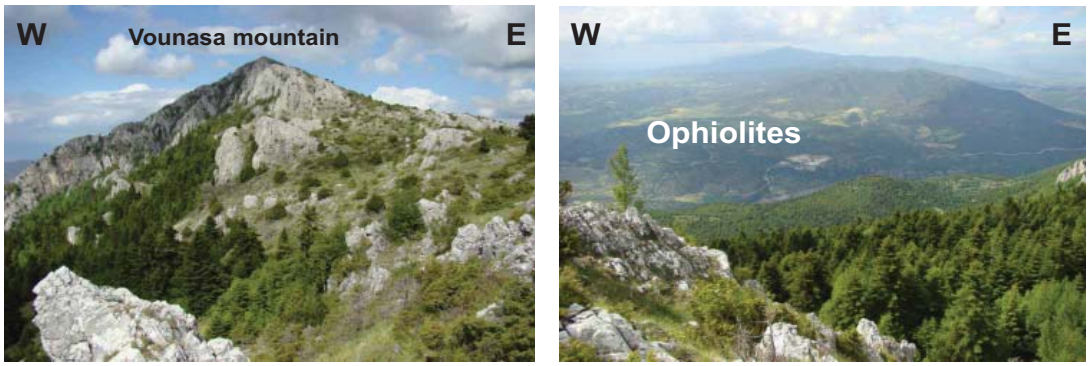


Figure E7
The Triassic-Jurassic platform carbonate cover of the Pelagonian nappe at the top of the Vounasa Mountain. Panoramic view of the Vourinos ophiolite nappe emplaced W-ward on the Pelagonian carbonate rocks, here bordered with the Pelagonian carbonate sediments with a NE-SW striking normal, possibly active fault zone, that in its hanging-wall segment with a throw of about 500m down-dips the ophiolites to the NW. View to the North. 39°57'03"N 21°46'08"E

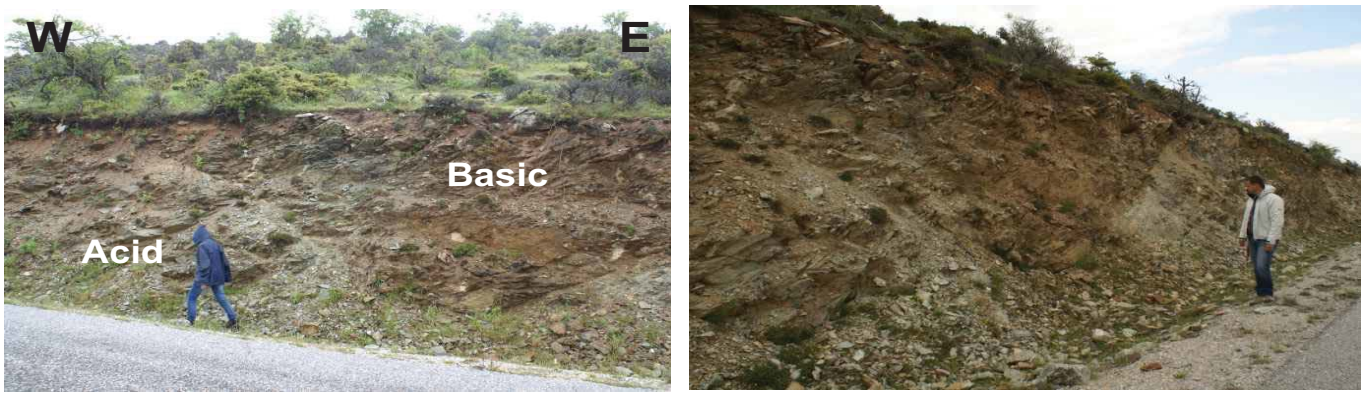


Figure E8
The Permo-Triassic bimodal magmatic series related to the Pangaia continental break and the future oppening of the Neotethyan ocean realm. 39°57'03"N 21°46'36"E

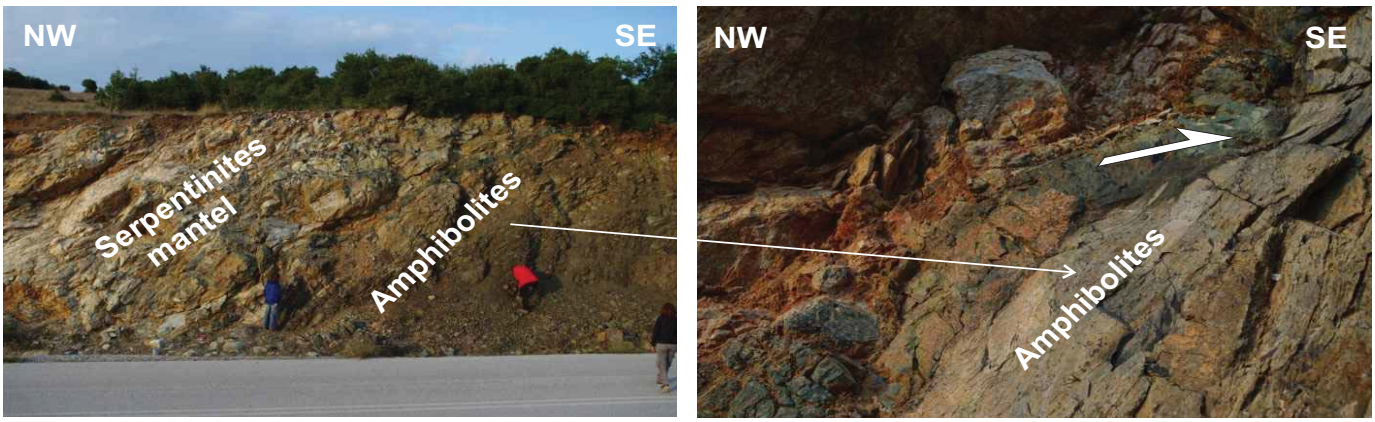


Figure E9
Ophiolite sole development between lithospheric mantel material on top (upper plate) and ophiolite mélanges at the bottom (under plate). The ophiolite sole age is 160-170Ma, and this is approximately the same as that of the Axios/Vardar zone ophiolite belt, to the east of the Pelagonian nappe and also of the Sup-Pelagonian ophiolite belt, west of the Pelagonian nappe. At this place the sense of shear appears to be towards the SE. We consider, that this sense of movement along the amphibolites sole occurrence is not representative of the real ophiolite emplacement direction due to intensive overprinting or rotation by the younger tectonic events, affected the ophiolite complex and its surrounding units after their emplacement history. 39°59'56"N 21°47'17"E

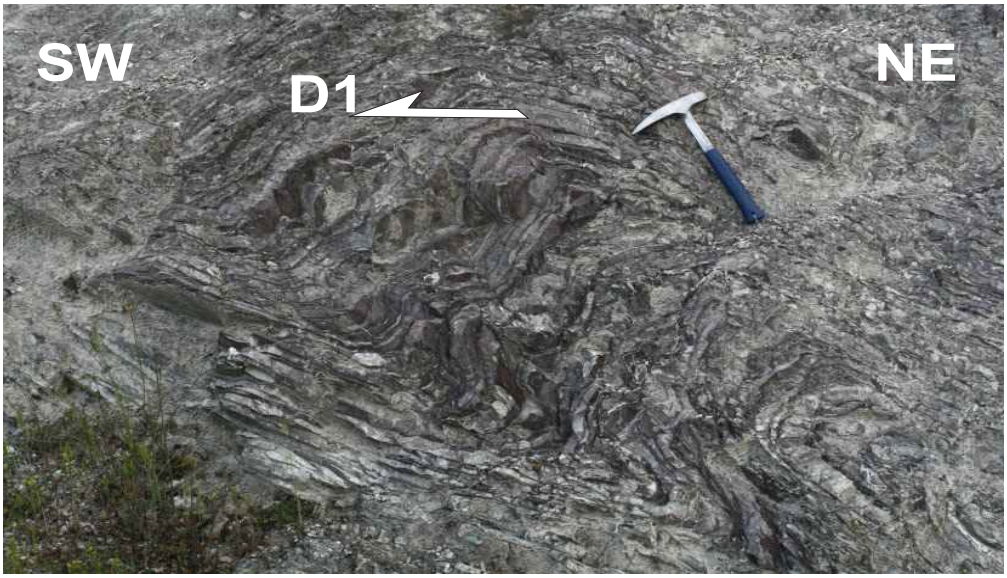


Figure E10
Deep-water sediments, intensively folded, with asymmetric sub-isoclinal folds W-wards vergent. The sedimentary sequence is composed of pelagic red carbonate sediments, metapelites and radiolarian cherts. The whole sequence has been sandwitched between the Triassic-Jurassic marbles of the western Pelagonian flank and the ?Sup-Pelagonian (Vourinos) ophiolite complex. 39°00'05"N 21°47'23"E

amphibolite-sole has a unique significance, since it clearly reveals its association with an intraoceanic subduction in a setting of one, single, ocean (i.e. the Neotythean ocean), which later led to the obduction of the ophiolites on the Pelagonian continent during the Late Jurassic. Deep-water sediments between the ophiolites and the Triassic Pelagonian carbonate cover are characterized by tight asymmetric folds verging towards the W (fig. E10).

F. 6th Day: Internal/External Hellenides (fig. 45)

I. Ophiolite belt at the western Pelagonian margin and Pindos zone
II. Mesohellenic trough
III. Pindos zone

Trek: From the town of Siatista to the village of Metamorphosis and Zygosti stream, and then to the city of Grevena and villages of Spilaeo, Avdella and Perivoli (Vourinos and Pindos Mountain; fig. 46).

Overnight staying in Siatista.

Starting from the town of Siatista, we reach the small village of Metamorphosis via the old provincial road Siatista-Kozani, and then we turn to a gravel road that leads to the village of Rodiani in order to visit the Zygosti stream on the way. From there, we turn back and take the Egnatia highway towards Grevena, until the exit to the village of Spilaeo located high up on the Orliakas Mountain. After we have reached Spilaeo, we take the provincial road connecting the mountain villages Spilaeo, Avdella and Perivoli, along which we cross the Pindos Mountain range up until the peak Avgo after the Perivoli village.

At the Zygosti stream, which we reach following the gravel road connecting the villages of Metamorphosis and Rodiani, we see once again ophiolite mélanges (fig. F1), with about the same composition as the ones described in the beginning of the previous day (5th day) at our exit from Siatista. We also see basic to ultrabasic ophiolite bodies and basic to intermediate lavas occurring in places over the ophiolite mélanges. The contact between the ophiolites/ophiolite mélanges and the Triassic-Jurassic Pelagonian carbonate cover (fig. F2) is today described by a high-angle, NE-ward dipping, normal fault, being developed parallel to the Zygosti stream.

The ophiolite lavas are overlain with a reworked tectonic contact by rich fossiliferous limestones of Late Jurassic – Early Cretaceous age (fig. F3, F4). The latter are transgressively covered by rudist-bearing Late Cretaceous neritic limestones terminated at the Paleocene flysch of the Internal Hellenides (fig. F5). In some places on the Zygosti ophiolite belt, a W-ward directed thrust is recognized, possibly related to the initial emplacement of the ophiolites on the Pelagonian continental margin (fig. F6).

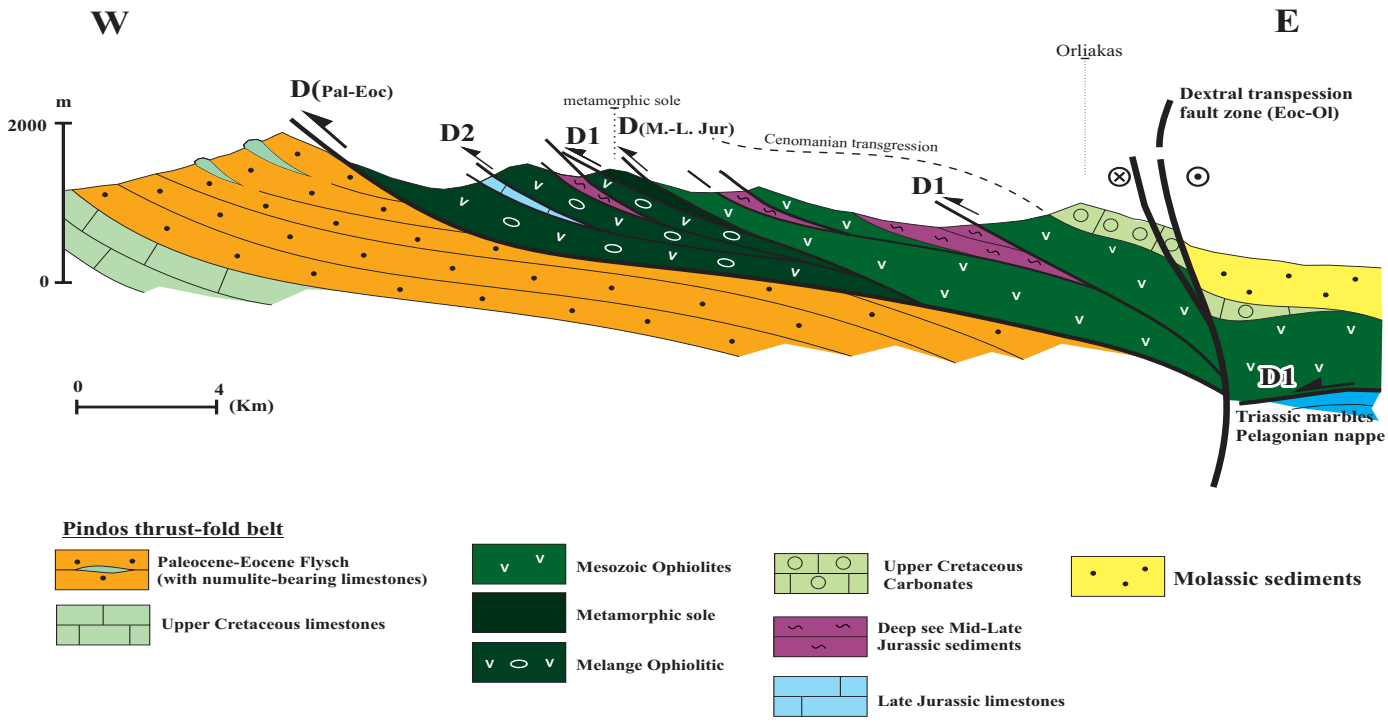


Figure 45
Schematic geological cross-section through the selected area/trek of the 6th day in the ?Sub-Pelagonian zone and Pindos ophiolite belt, as well as the Mesohellenic Trough and Pindos zone.

Figure 46
Location of the selected outcrops and stops are shown on the corresponding trek of the 6th day.

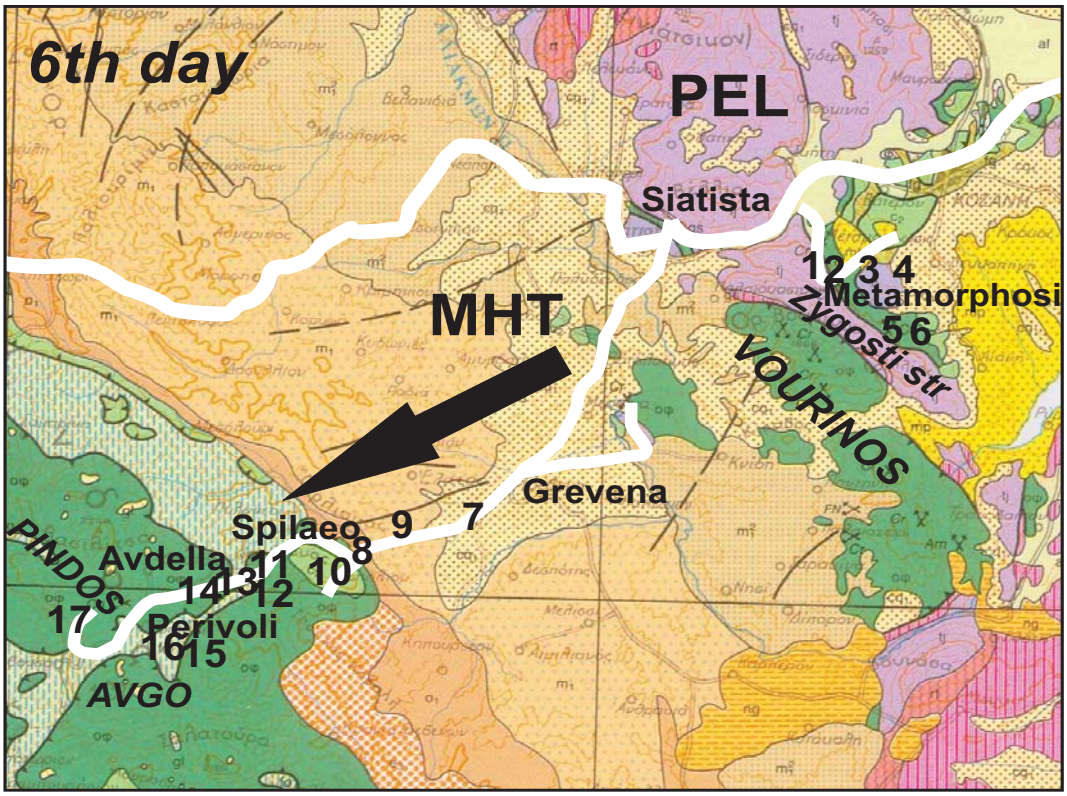


Figure F1
Zygosti ophiolite mélanges. They are of Jurassic age, related to the ophiolite emplacement on the Pelagonian continental margin, possibly of the Axios/Vardar ocean orogin. 40°14'43"N 21°40'53"E

Figure F2
Sedimentation layering of the western Triassic-Jurassic Pelagonian carbonate cover (rythmites). 40°12'38"N 21°41'44"E

Continuing our course, all along the Egnatia road to the city of Grevena and then along the provincial asphalt road to the village Spilaeo, we cross most of the molassic formations of the MHT, perpendicular to their depositional age, from the younger to the older one (i.e. the Tsotili formation of Lower-Mid Miocene age, fig. F7; the Pentalophos formation of Oligocene-Lower Miocene age, fig. F8; and last the Eptachori formation of Eocene-Oligocene age, fig. F9e). Finally, just before the village of Spilaeo, we find the tectonic contact of the last MHT formation (i.e Eptachori formation) with the rudist-bearing Late Cretaceous limestones or the ophiolite rocks and the Pindos flysch, which is characterized by an almost vertical dextral strike-slip fault (fig. F9, F10).

From the village of Spilaeo, located on the Late Cretaceous limestones close to the top of Orliakas Mountain, we have a panoramic view of the Pindos ophiolite belt and the oldest, Eocene, Kranea formation of the MHT. The Kranea formation is characterized by its restricted occurrence in two relatively small areas at the west and at the east of the MHT, and also by its compressional tectonic structures such as folds and thrust faults verging mainly towards the east but also towards the west (fig. F9e). Furthermore, by following a gravel road from the village of Spilaeo towards the old Portitsa Bridge, we can visit the amazing Spilaeo gorge developed within the Late Cretaceous limestones.

Along the sealed road from the village of Spilaeo to the mountain village of Perivoli, we cross the Mid-Late Jurassic Avdella ophiolite mélange, overlain by the Pindos ophiolite belt that is composed of mantle and oceanic crustal material (fig. F11), as well as a sheeted dykes complex and pillow lavas. The Avdella mélange is consisted of a chaotic setting of igneous and sedimentary successions in larger or smaller blocks, all highly imbricated (Johns & Robertson 1991). There is a great variation in the type and the age of those deposits. They comprise basic to intermediate extrusive and intrusive deposits of Triassic age and plateau carbonate or slope margin sediments of Triassic to Jurassic age, as well as reworked sediments. Furthermore, highly silica turbidites, radiolarites and shales are also included in the Avdella mélange composition. Those radiolarites are intercalated with Mid-Late Jurassic sediments with ophiolites-derived clastic and volcanic prod-

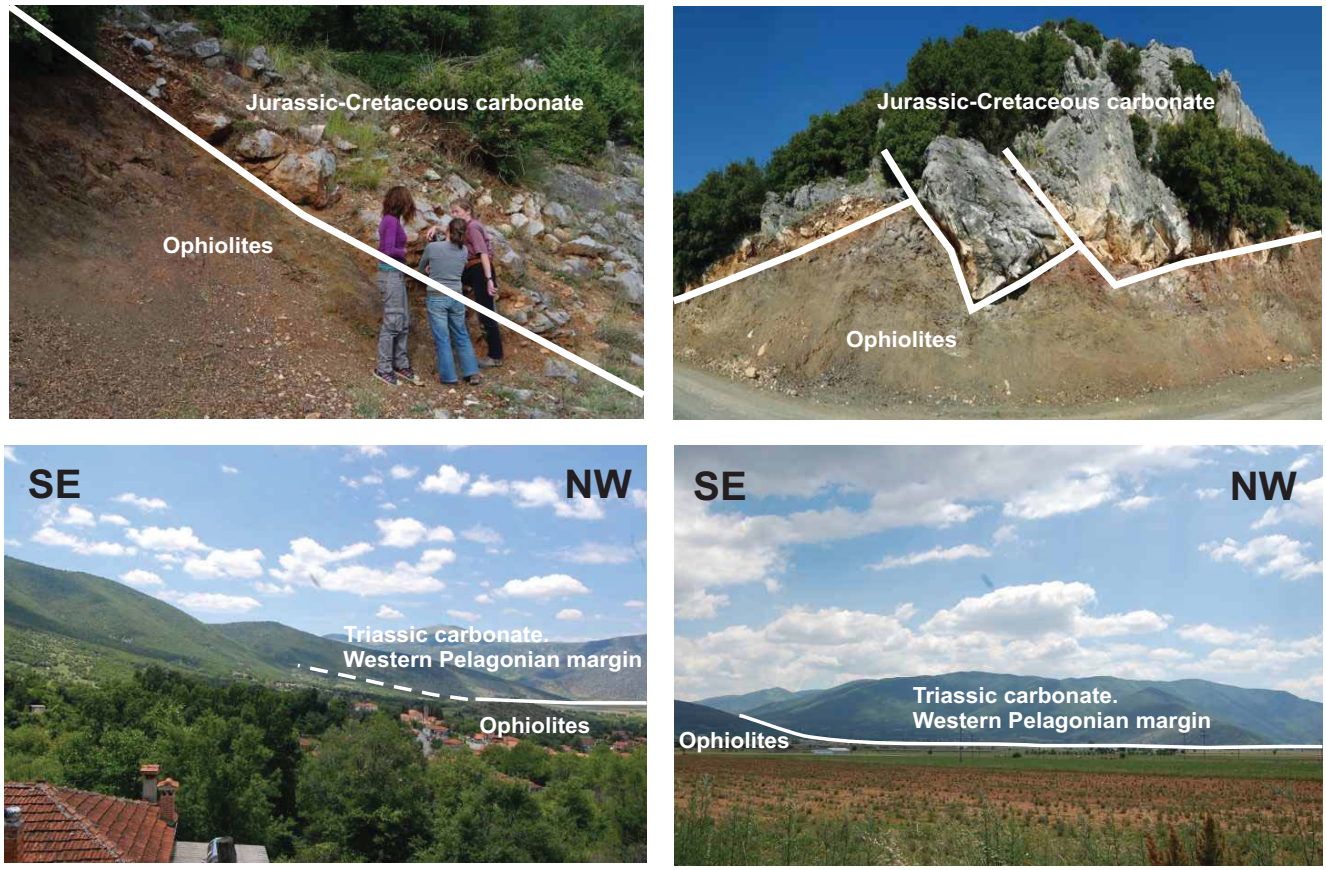


Figure F3
The contact between the Upper Jurassic-Lower Cretaceous neritic limestones and the Zygosti ophiolites, 40°12'35"N 21°42'40"E. This contact, initially of transgressive character, has been clearly overprinted by a younger extensional event equivalent to the D5 Oligocene-Miocene event, related to the Pelagonian collapse, 40°13'12"N 21°43'40"E. The Zygosti ophiolites developed along the Zygosti stream are composed of ophiolite mélanges, serpentinites and volcanic products. They form the W-ward continuation of the Axios-Almopia ophiolites, which reach as small rests the Zygosti stream area. Furthermore, the Zygosti ophiolites are bounded with the western Pelagonian carbonate cover with a younger Neogene-Quaternary high angle, NE-ward dipping, normal fault (fig. F3c, 3d)., 40°13'18"N 21°41'20"E, 40°15'50"N 21°39'15"E

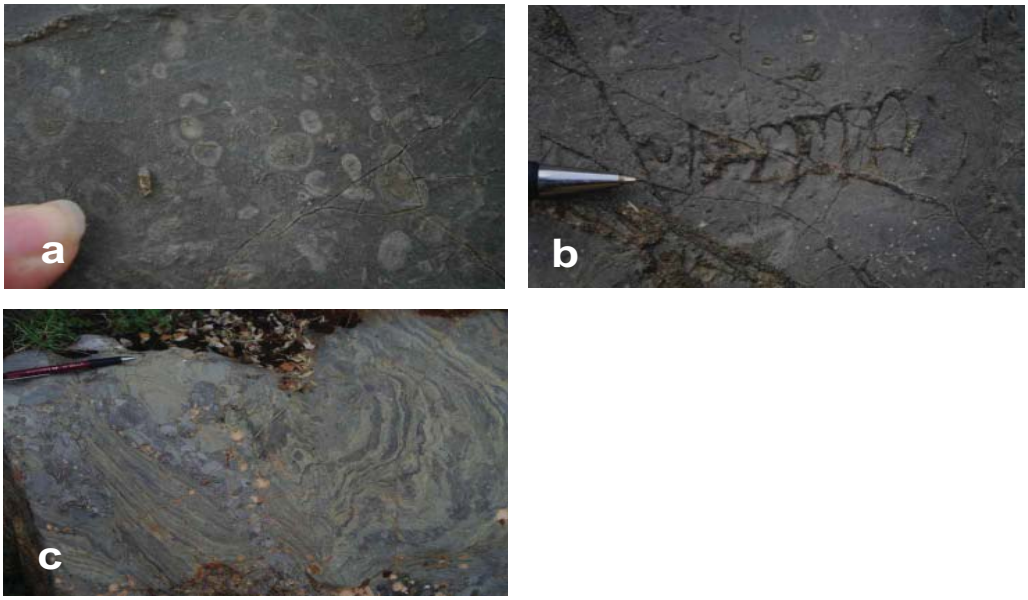


Figure F4
Fauna from the rich fossiliferous neritic Upper Jurassic (?Kimmeridgian)-Lower Cretaceous limestones, covering the obducted ophiolites belt at the Pelagonian continent. Sedimentary structures are also shown (fig. F4c). 40°12'49"N 21°42'52"E

Figure F5
W-ward vergent D4 thrust fault of Paleocene-Eocene age, affected the Upper Cretaceous neritic limestones at the Zygosti stream. The latter are related to the Upper Cretaceous transgression that started in the Cenomanian. 40°12'59"N 21°43'02"E

Figure F6
W-ward directed low angle thrust zone at the Zygosti ophiolite belt related to the ophiolite emplacement history onto the Pelagonian continental margin. The observed W-ward sense of movement supports the origin of the obducted ophiolites belt from a source east of the Pelagonian margin. 40°13'13"N 21°43'28"E

Figure F7
The Lower – Middle Miocene Tsotyli formation of the molassic Meso-Hellenic trough with characteristic crossed layers. 40°10'34"N 21°30'51"E

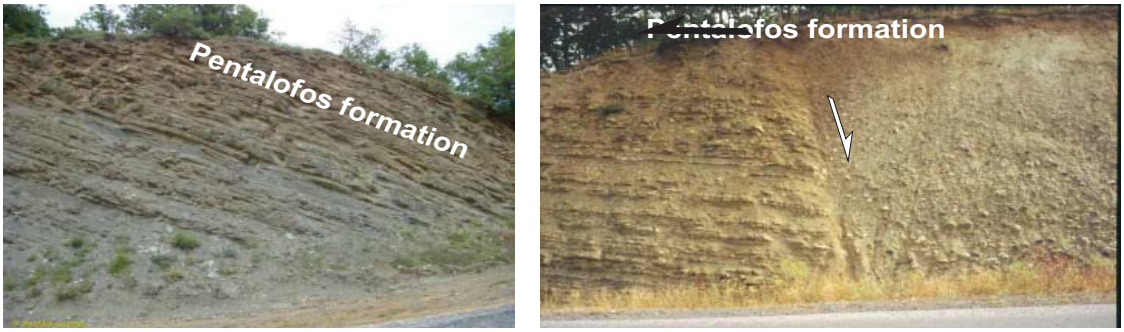


Figure F8
Turbidites of the Pentalofos formation dipping with a low angle to the East. A high angle normal fault separates the two different sedimentary phases of the Pentalofos formation captured in the picture. The conglomerates correspond to fan-delta deposits and are very common in Pentalofos formation. 40°02'22"N 21°18'57"E

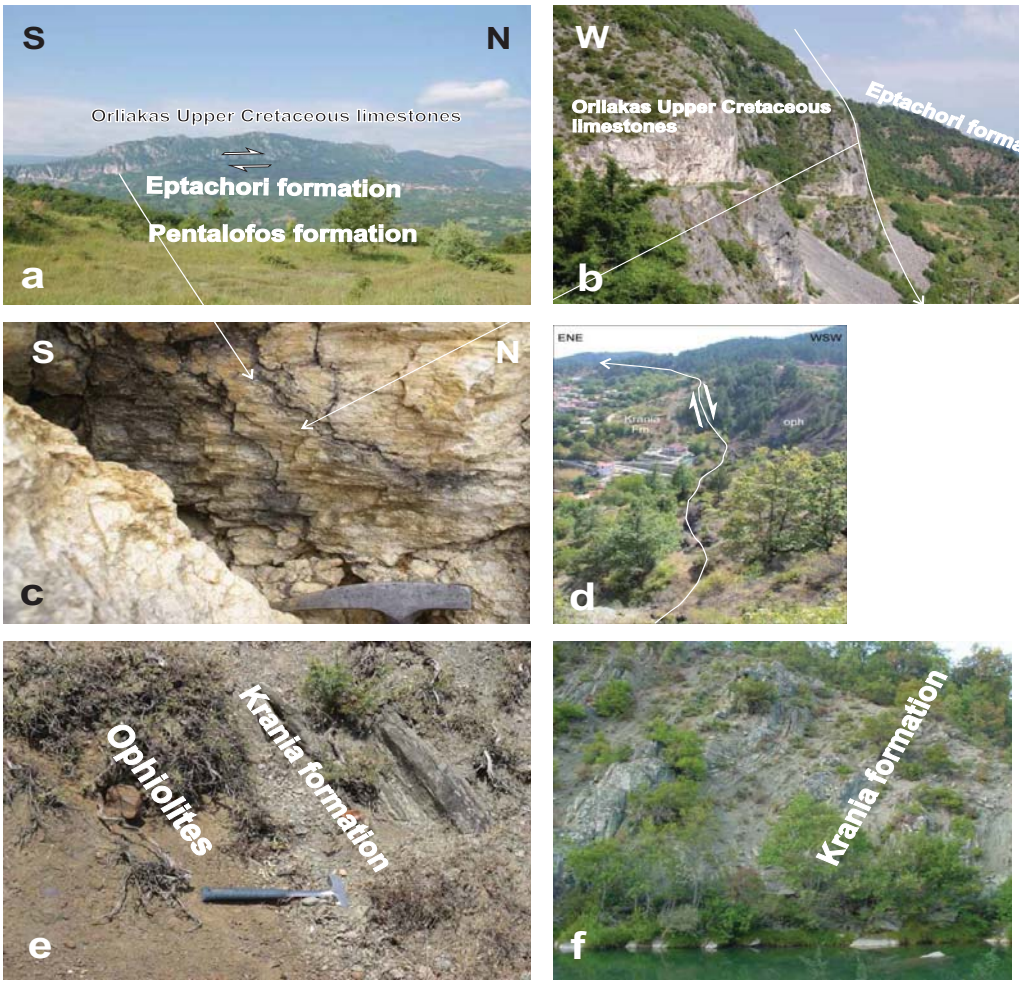


Figure F9
The N-S striking, dextral strike-slip fault zone bounding the molassic sediments of the MHT with the Orliakas Upper Cretaceous limestones (panoramic, view to the West, and close view of the fault zone). The Orliakas Upper Cretaceous limestones itself form a pop-pup-structure resulted by that strike-slip fault zone action. Along the fault zone occur the Oligocene sediments of the Eptachori formation dipping with a high angle to the East. The same fault zone continues to the North and South forming the western boundary of the whole MHT against the ophiolite rocks and the Pindos flysch. An analogous dextral strike-slip fault zone separates the Pindos ophiolite belt with the Eocene-Oligocene Kranea formation which is intensively affected by compessional tectonics (fig. F9f). The Eptachori formation discordantly overlies the Kranea formation. 40°02'22"N 21°18'53"E, 40°00'38"N 21°17'00"E

Figure F10
The Orliakas Upper Cretaceous transgressive limestones, in places with a rich fauna in rudistes and belemnites, representative fossils of the Upper Cretaceous stratigraphic period. 40°00'24"N 21°17'05"E

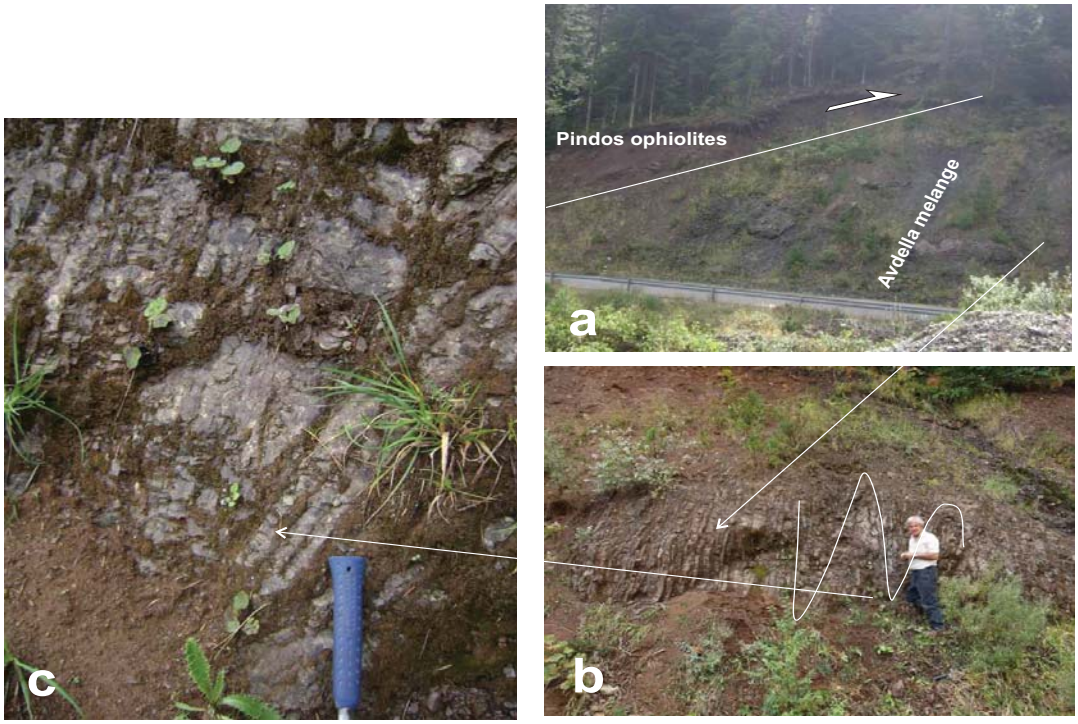


Figure F11
The tectonic contact between the Pindos ophiolite belt and the deep-water Mid-Upper Jurassic sediments of the Avdella mélange. The Pindos ophiolites overthrust towards west the deep-water sedimentary succession, which is folded by assymetrical sub-isoclinal W-ward vergent folds. We consider that this tectonic fabric is related to the initial towards West Late Jurassic ophiolite emplacement on the Pelagonian continental margin (fig. F11b, c). 40°01'31"N 21°14'31"E

ucts. Last, we also see the Avdella mélange locally imbricated with Late Jurassic – Early Cretaceous deep-sea sediments (fig. F12, F13, G14, G15). The Avdella mélange and the overlying Pindos ophiolites are thrusted over the Paleocene-Eocene Pindos flysch of the External Hellenides Pindos zone. At this location, the Pindos flysch comprises, among others, olistoliths of neritic nummulite-bearing limestones (fig. F16).

Finally, above the Perivoli village, turning right from the provincial road connecting Perivoli and Vovoussa villages to a gravel forestry road we observe the development of the ophiolite sole between the Pindos ophiolite belt and the Avdella ophiolite mélanges (fig. F17). In the broader area, the latter is developed as an extensive thrust sheet sandwiched between the underlying Pindos flysch and the overlying Pindos ophiolite belt. Further to the south, in the Koziakas Mountain, the continuation of the Avdella mélange outcrops with a similar composition, but there it is thrust over the Triassic-Jurassic deep-water pelagic carbonate formation of the Koziakas unit, which is sandwitched between the Paleocene-Eocene Pindos

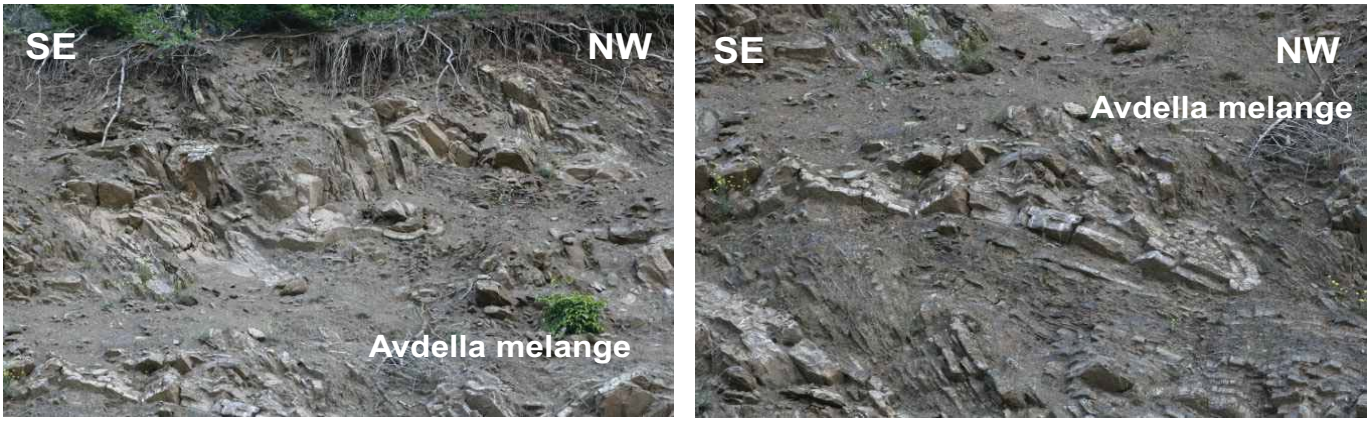


Figure F12
Sedimentary fold structures in the deep-water Middle-Late Jurassic sediments of the Avdella mélange. They are composed of redeposited radiolarian cherts and pelagic carbonate layers, as well as volcanic products. The deposition took place in deep-basins at the frontal parts of the overriding upper ophiolite plate during the Mid-Late Jurassic intraoceanic subduction in the Neotethyan ocean realm. In those basins, the other sedimentary phases of the Avdella mélange were also deposited, simultaneously, containing carbonate -mainly pelagic but also neritic- olistoliths, radiolarian cherts and ophiolites pebbles, as well as radiolarian cherts layers. 40°01'31"N 21°14'31"E

Figure F13
Deep-water Middle-Upper Jurassic sediments. They are strongly imbricated with the Avdella mélange during the ophiolite emplacement. 40°01'31"N 21°14'31"E

flysch or the Early Cretaceous Boeotian flysch and the ophiolite mélanges/ophiolite realm (fig. H11). Furthermore, equivalent mélanges formations and ophiolites also occur at the western part of the Vourinos ophiolite complexes right on the Pelagonian margin, as well as all along the Pelagonian margin, north and south of this location.

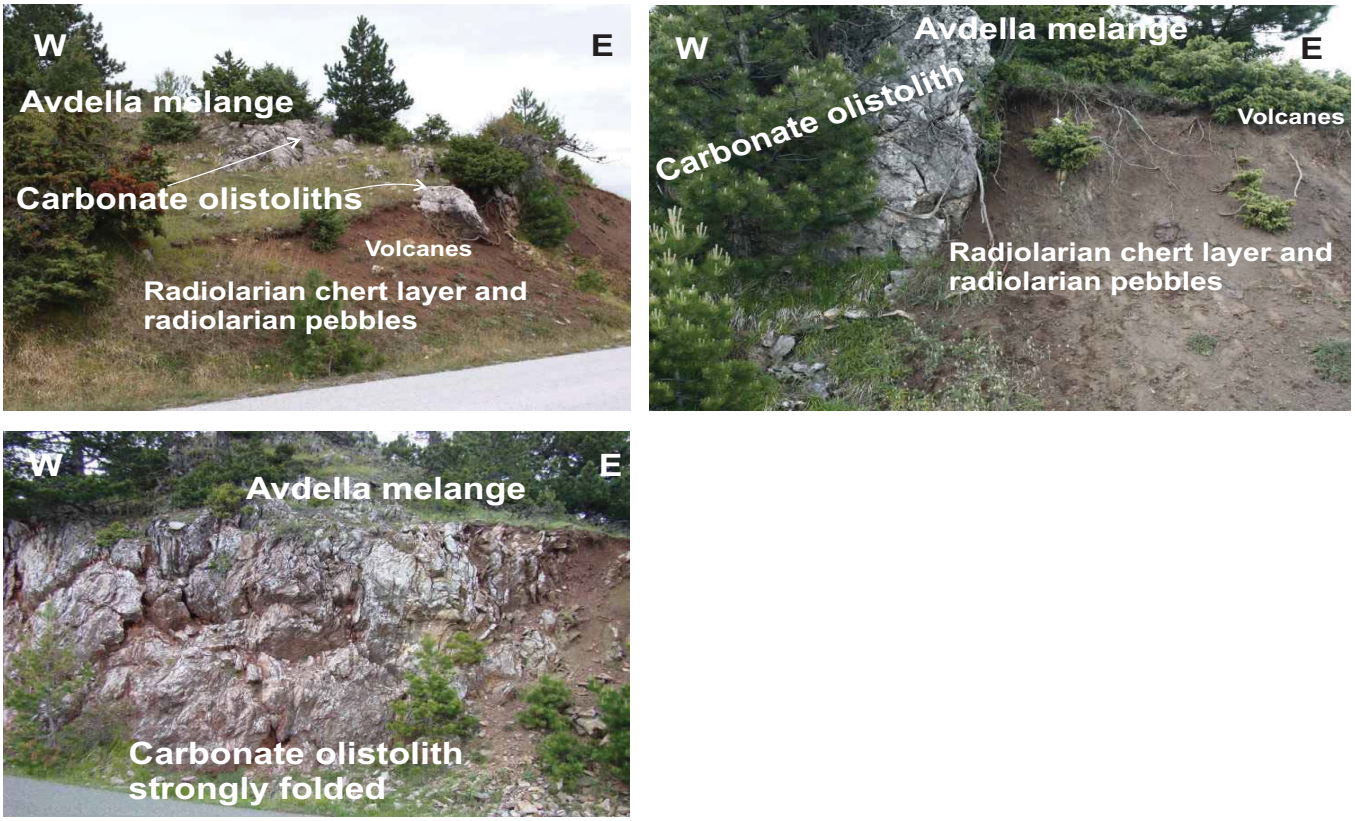


Figure F14
Pelagic red-colored carbonate olistoliths and radiolarian pebbles in between radiolarian cherts layers, and volcanic products indicating the paleo-geotectonic position of these basins in the vicinity of an ensimatic island arc. Some of those carbonate olistoliths are strongly folded with isoclinal folds, possibly related to the ductile syn-emplacement tectonic history of the ophiolites prior to their propagation over the Pindos flysch further to the West, where they were thrusted as a rigid ophiolite body without a syn-tectonic involvement with the underlying Pindos flysch. For this reason the pre-emplacement-on- the-Pindos-flysch ophiolite fabric is presented, indicative for the ductile deformation history of the ophiolites and the syn-imbricated with the ophiolites Avdella mélange. 40°00'59"N 21°13'11"E

Figure F15
A typical picture of the Avdella mélange with the chaotic mixing of heterogeneous material, composed of olistoliths and pebbles of pelagic red carbonates, radiolarian cherts, basic and ultra-basic products, as well as pelites. A reverse thrust zone is distinguished with a top-to-the-W sense of movement. 39°58'53"N 21°07'17"E

Figure F16
Nummulites bearing neritic carbonate ?olistoliths of unknown origin within the Paleocene-Eocene Pindos flysch. They usually occur near the front part of the thrust zone of the overthrusted ophiolite belt on the Pindos flysch. 39°58'12"N 21°07'04"E

Figure F17
The Middle Jurassic amphibolite sole inbetween ultrabasic manltle rocks (serpentinites) and the Avdella ophiolite mélange. 39°57'31"N 21°05'49"E

G. 7th Day: Mesohellenic trough, External Hellenides (fig. 47)

I. Mesohellenic trough
II. Pindos ophiolite belt
III. Pindos zone
IV. Ionian zone

Trek: From the town of Siatista to the town of Konitsa, through the villages of Tsotili, Pentalofos and Eptachori. Then, from Konitsa to the villages of Vitsa and Monodendri at Tymfri Mountain, above the Vikos gorge (fig. 48).

Overnight staying in the village of Metsovo.

Leaving Siatista, we take the provincial asphalt road connecting the cities of Kozani and Ioannina, which passes through a series of mountain villages of exceptional beauty, such as these of Pentalofos and Eptachori, as well as the town of Konitsa with its famous old stone-bridge. After Konitsa, we take the provincial road that starts at the exit of the Aoos gorge and goes alongside Aoos River, with direction towards the city of Ioannina. On a big right bend of this road, we cross Voidomatis River, one of the clearest rivers in Europe, also characterized by its very cold spring waters. From there, we have an excellent overview of Tymfri Mountain. Continuing the same way, we turn left after the historical village of Kalpaki towards the

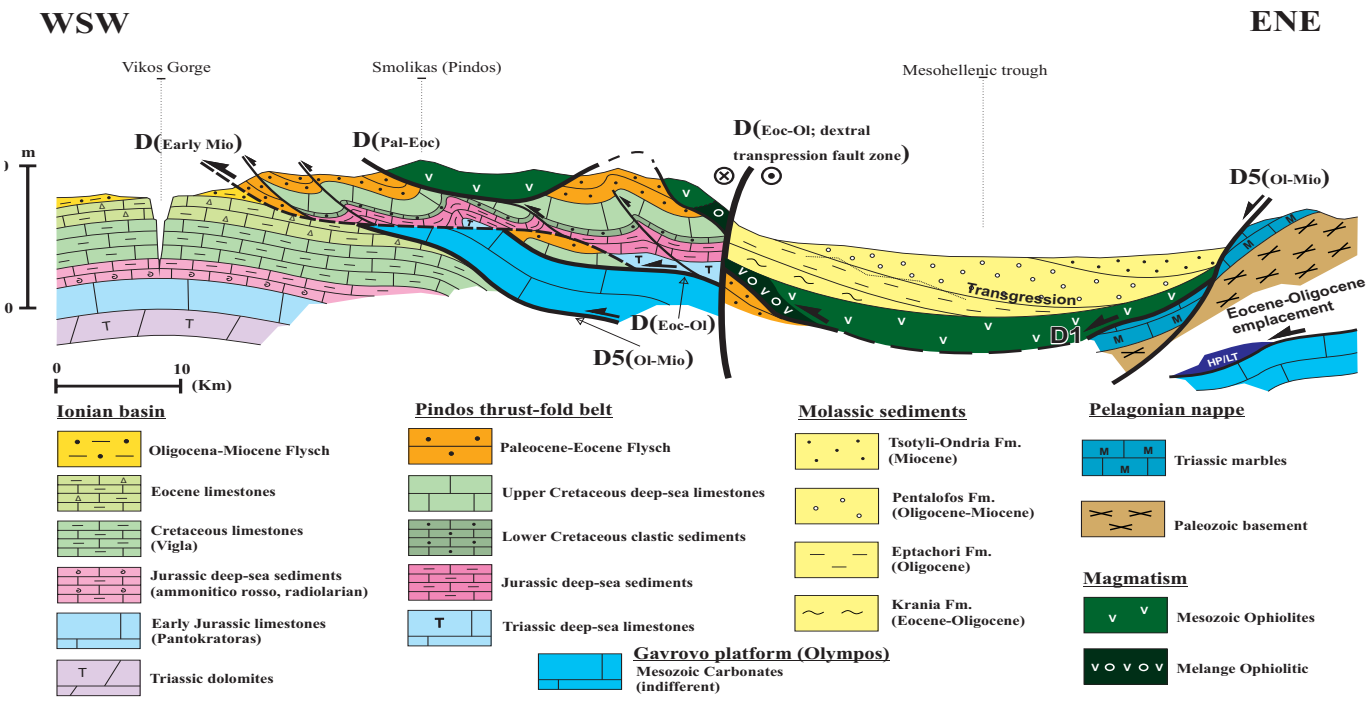


Figure 47
Schematic geological cross-section through the selected area/trek of the 7th day in the Mesohellenic Trough, Pindos ophiolite belt, Pindos zone and Ionian zone.

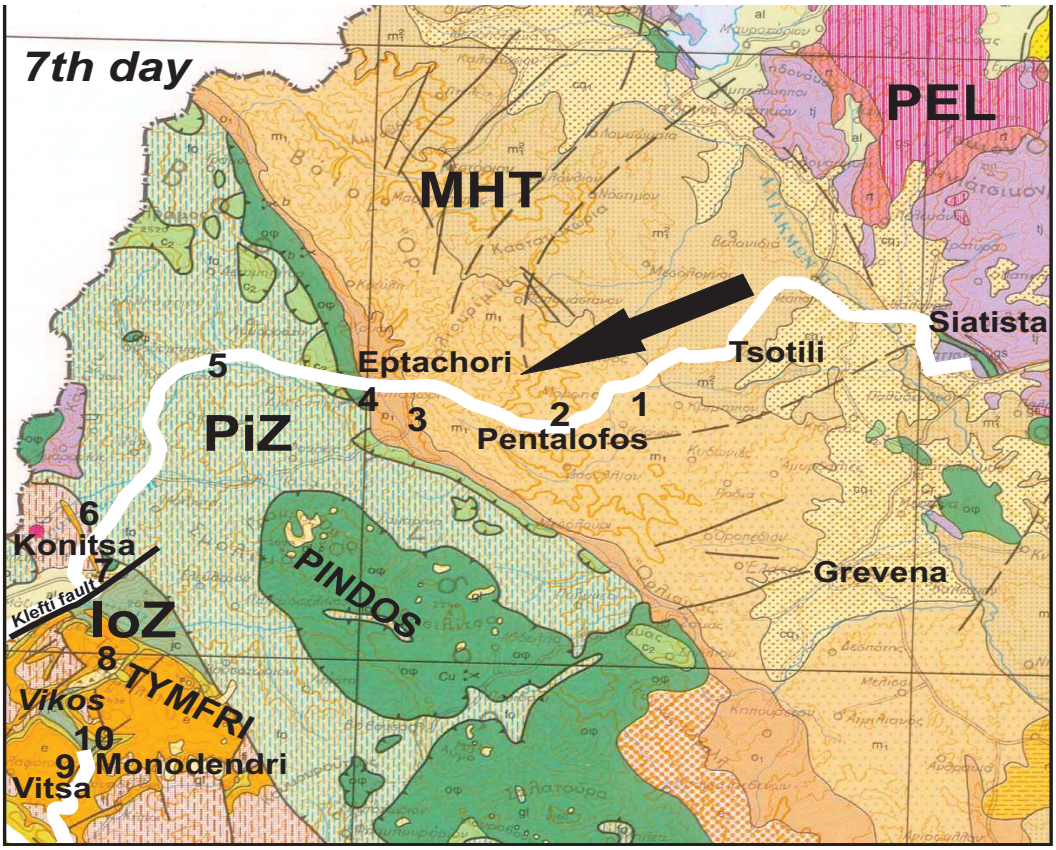


Figure 48
Location of the selected outcrops and stops are shown on the corresponding trek of the 7th day.

mountain sealed road to the traditional stone-villages of Vitsa and Monadendri, where our route for the day ends.There, just outside the village of Monodendri, is the famous Vikos gorge, the deepest Greek gorge that has been awarded a Guiness world-record for its big depth in comparison to its small width.

Along our traverse, first we cross the Tsotyli, Pentalofos and Eptachori molassic formations of the MHT in turn, characterized by a mostly low-angle NE-ward dipping of their bedding. The Tsotyli formation comprise from fine to coarse-grain deposits, while the Pentalophos formation is described mainly by coarse-grain to conglomeratic deposits, in some places with a particularly great thickness (e.g. Zelilidis et al. 2002, Vamvaka et al. 2010; fig. G1, G2). On the contrary, the Eptachori formation has an estimated restricted thickness and is characterized more or less by a fine-grain turbidite layering with more clay material in its composition (e.g. Savoyat and Monopolis 1972, Vamvaka et al. 2010; fig. G3). The Eptachori formation also differs in the amount of dipping, towards the NE, appearing steeper in most cases than the Pentalophos and Tsotyli formations, especially close to its contact with the basement rocks, like the one we reach few hundred meters after we pass the village of Eptachori. That is a tectonic contact, where the basement rocks are mainly represented by the Pindos ophiolites and the overlying neritic, rudist-bearing Late Cretaceous limestones; the latter are also found in some places in the MHT, mostly close to the basins' margin, as great olistoliths (e.g. in the Kranea formation).

However, at this contact between the Eptachori formation and the Pindos ophiolites, ophiolites

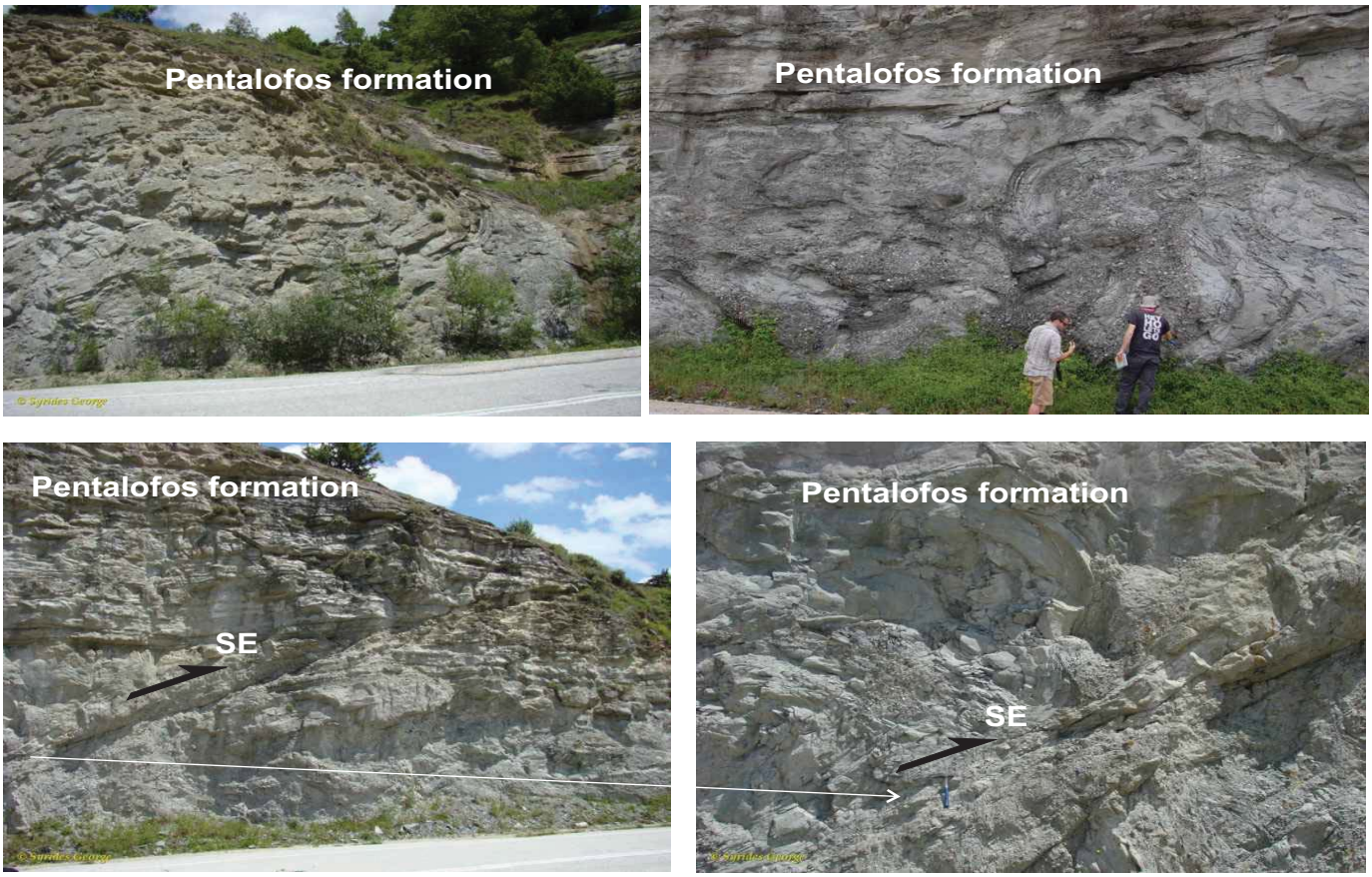


Figure G1
Pentalofos formation of the MHT. There are distinguished typical mass-flows structures and sedimentary origin's recumbent, isoclinal folds of the molassic Pentalofos deposits. The Pentalofos formation continues to the Lower-Middle Miocene Tsotyli and Ondria formations concordantly, without any important stratigraphic gap or sedimentation break. A thrust fault with a top-to-the-SE sense of movement is distinguished, as a remnant of the synorogenetic compression affected the External Hellenides during the Tertiary, forming their characteristic tectonic nappe-pile architecture. 40°11'36"N 21°11'06"E

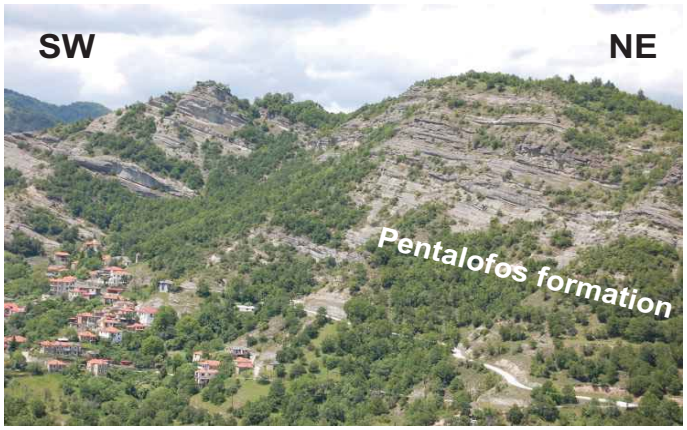


Figure G2
Panoramic view of the Oligocene Pentalofos formation with a NE-ward low angle dipping of its layering. Conglomerate fan-delta sedimentary accumulations are the characteristic feature for the entire Pentalofos formation. View to the Northeast. 40°11'50"N 21°08'25"E

Figure G3
View of the Eocene-Oligocene Eptachori formation. It is mostly characterized by a fine-grained clay material, dipping also towards the NE, as the majority of the Mesohellenic deposits. 40°13'07"N 21°01'42"E

mélanges also occur, equivalent to the Avdella mélange. The ophiolite mélanges are overthrust towards west by great serpentinites masses of mantle-origin rocks. Both, ophiolite mélanges and serpentinites, are in turn thrust over the Pindos flysch further to the west (fig. G4). Along the same road, after the ophiolites' overthrusting on the Pindos flysch, we have the chance to study the thick Pindos flysch formation itself, which is intensively deformed by compressing thrust and fold structures. In places it occurs as wild flysch with impressive structures, characteristic for the Pindos flysch formation (fig. G5). Further on, at the descent towards Konitsa, the Pindos flysch is thrust directly on the Ionian flysch towards the west, whereas the entire Gavrovo zone is omitted. At the same location, under the Ionian flysch, the Eocene thin-bedded limestones of the Ionian zone are also exposed, forming a NW-trending mega-anticline (fig. G6).

Afterwards, from Konitsa and about parallel to the provincial road we take towards the villages of Vitsa and Monodendri, at the foot of Tymfri Mountain, the active Klefti fault is developed with a main WNW dip-direction (fig. G7, G8). Our cross-section continues through the Ionian zone until the Vikos gorge cutting the whole litho-stratigraphic column of the zone (fig. G9, G10). An impressive view of the Ionian stratigraphy is available along the Vikos-gorge, from the Jurassic deep-sea sediments to the Cretaceous and Tertiary thin-bedded limestones intercalated with chert layers, up to the Oligocene-Miocene Ionian flysch (fig. G10).

Figure G4
Ophiolite mélange, tectonically lying under an ultrabasic, mantlel origin`s ophiolite complex, with dominatig the serpentinites in this region. The mélange itself forms a volcanoclastic sequence, intensively sheared and composed of serpentines, basalt and dolerite intrusions, radiolarian cherts and siliceous shale, fine-bedded marble olistoliths and debris flow deposits. 40°13'39"N 20°58'43"E

Figure G5
The characteristic "wild flysch" of the Paleocene-Eocene Pindos flysch intensively imbricated and folded. The wild flysch is composed of fine and rough clastic material and olistoliths of the same composition and origin as those of the flysch deposits, transported however in depth due to turbidite streams or mass flows in an intensively deformational regime and a highly tectonically active environment. A thrust fault is distinguished, following the general SW-ward Paleocene-Eocene imbrications` and folding`s direction of the Pindos zone. 40°10'46"N 20°48'11"E

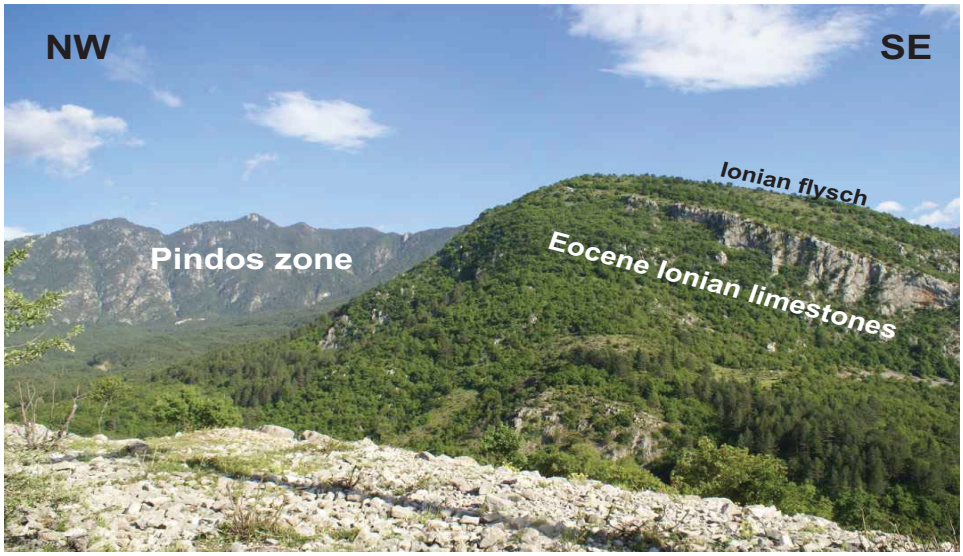


Figure G6
Panoramic view of the thrusting of the Pindos flysch on the Ionian flysch. The nummulites-bearing Ionian thin-bedded Eocene limestones are exhumed under the Ionian flysch. View to the East. 40°04'15"N 20°45'03"E

Figure G7
Panoramic view of the active normal Klefti fault. View to the Southeast. 40°03'17"N 20°44'08"E

Figure G8
Panoramic view of the Tymphri mountain with the Ionian, Cretaceous – Tertiary thin-bedded limestones. View to the Northeast. 39°57'41"N 20°39'05"E

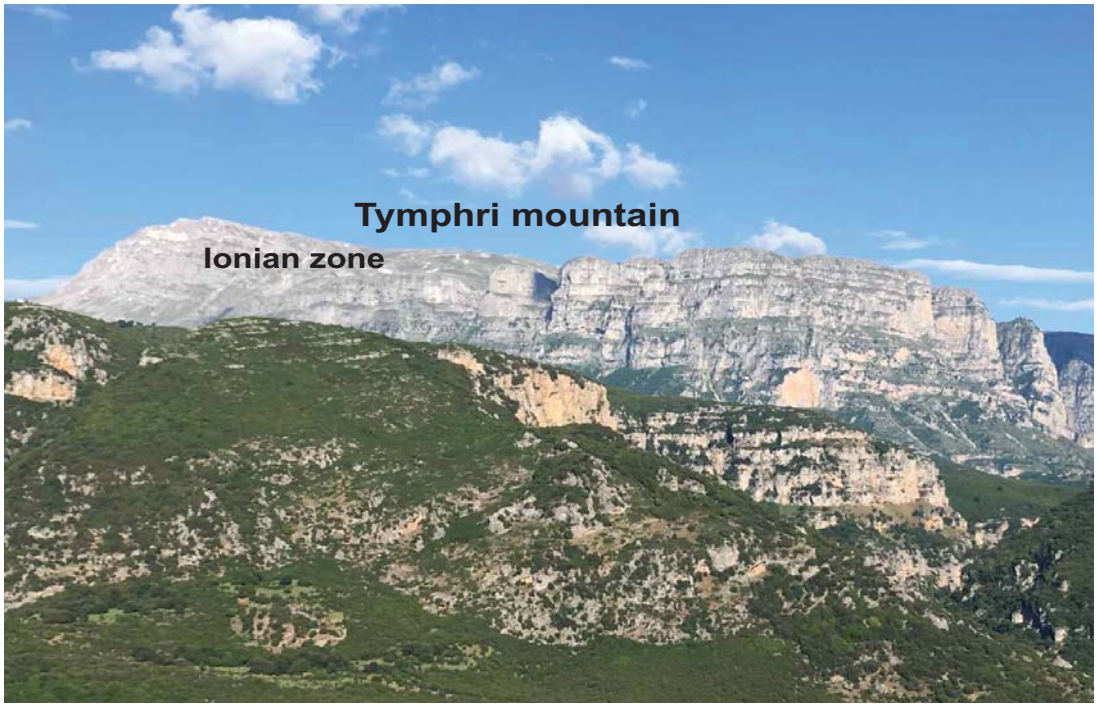


Figure G9
Nummulites-bearing thin-bedded Eocene deep-water limestones, intercalated with black siliceous cherts belonging to the Ionian zone stratigraphy. The latter is characterized by a continuous sedimentation from the Middle Triassic to the Eocene-Oligocene terminating with the Oligocene-Miocene Ionian flysch. 39°53'02"N 20°44'59"E

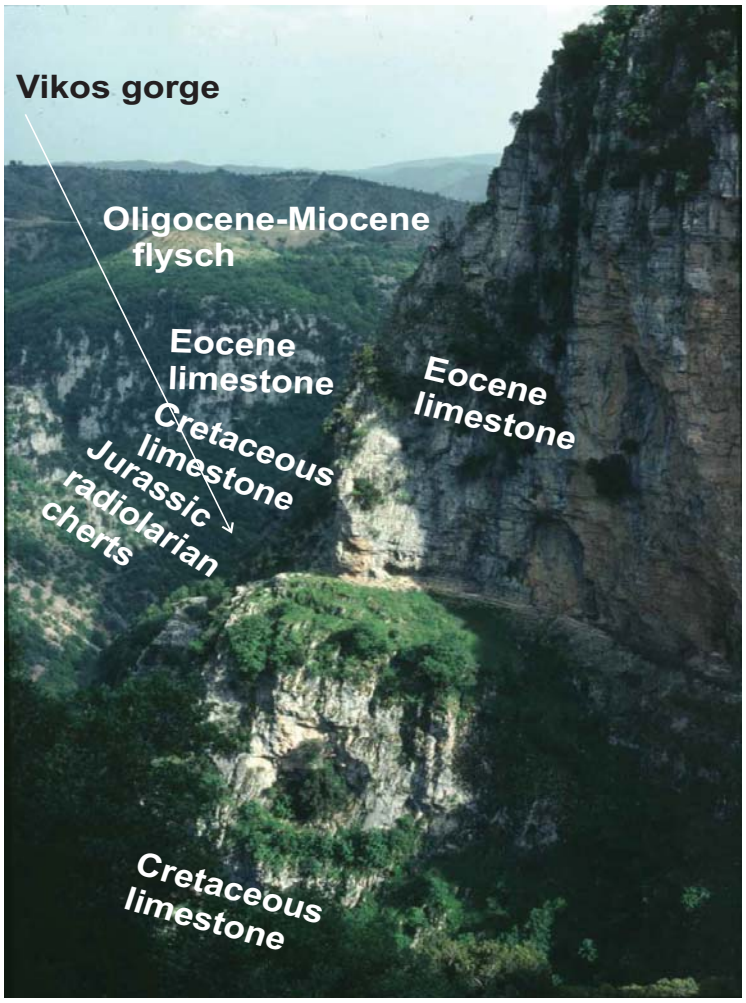


Figure G10
The Vikos gorge. The Ionian zone stratigraphy is shown, from the Jurassic radiolarian cherts layer and pelagic limestones at the bottom of the Vikos gorge to the Oligocene-Miocene Ionian flysch at the top. In the middle occur the deep-water cherts-bearing thin-bedded Cretaceous and Tertiary limestones. View to the South. 39°53'13"N 20°45'14"E

H. 8th Day: External Hellenides (fig. 49)

I. Ionian zone
II. Gavrovo zone
III. Pindos zone
IV. Koziakas unit.

Trek: From the village of Metsovo to the cities of Ioannina and Arta, then to the villages of Peta, Ano Kalentini, Piges, Argithea and Mouzaki, and finally back to Thessaloniki (fig. 50).

End of field-trip.

Leaving the village of Metsovo, we move southwards till the city of Arta, first taking the Egnatia highway till the city of Ioannina and then the national road connecting Ioannina and Arta. From there on, we move towards the ENE all the way back to the city of Thessaloniki where our field-trip comes to an end. For this, we first take the provincial asphalt road to the villages of Peta, Ano Kalentini, Piges and Mouzaki, continue till the city of Trikala, and from there we finally take the national highway that leads to Thessaloniki.

At the beginning of our section from the village of Metsovo, we move within the Pindos flysch which is cut by the Egnatia highway, until we see it overthrust directly onto the Ionian flysch. We now continue the cross-section within the Ionian flysch approximately till the city of Ioannina, while the Pindos overthrust extends alongside the mountain hang of the Egnatia road, but later, instead of the Pindos flysch it's the Pindos Triassic limestones on the Ionian flysch. Afterwards, along the national road from Ioannina to Arta, we travel across the lithostratigraphic units of the Ionian zone from the Triassic limestones to the Eocene thin-bedded limestones and the Oligocene-Miocene Ionian flysch. They form impressive NW-trending mega-anticlines and –synclines, usually with the Triassic (Ionian) limestones in the core of the anticlines and the Ionian flysch in the centre of the synclines (fig. H1). The typical Early Jurassic Pantokratoras white limestones also occur in places along the road (fig.H2).

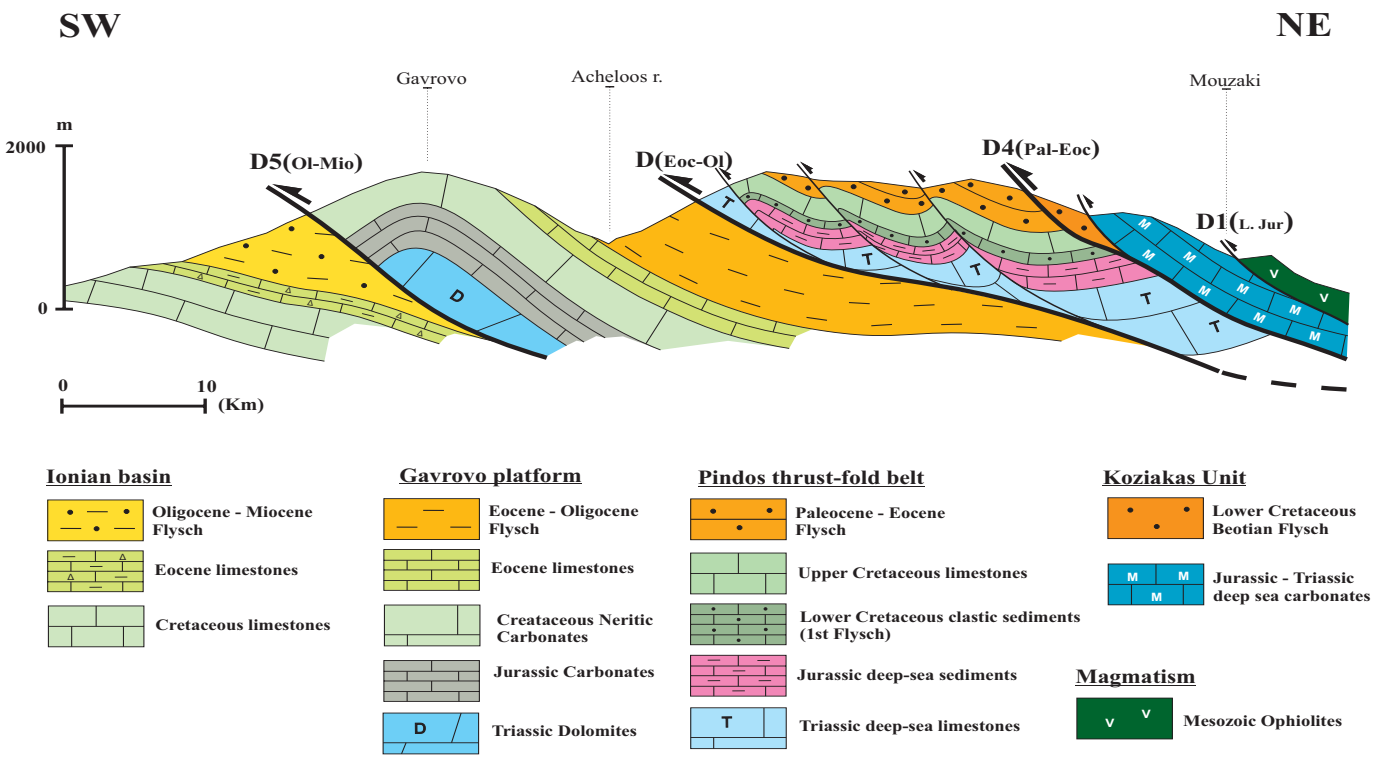


Figure 49
Schematic geological cross-section through the selected area/trek of the 8th day in the Ionian, Gavrovo and Pindos zones, as well as in the Koziakas unit.

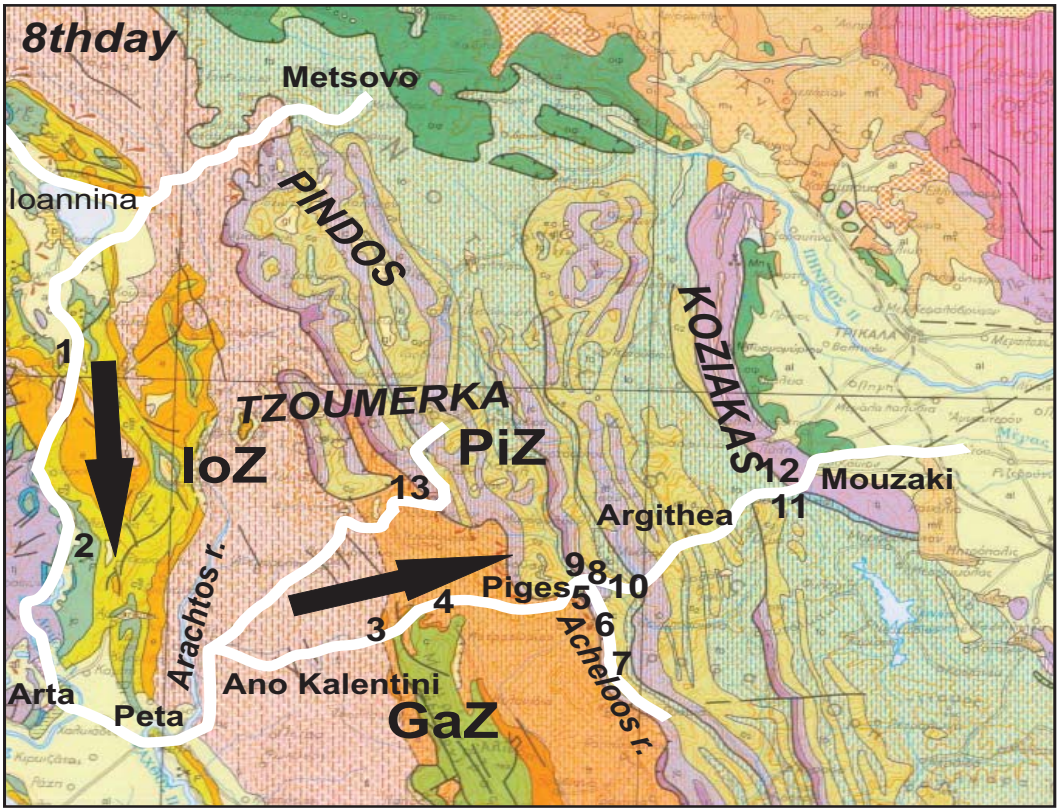


Figure 50
Location of the selected outcrops and stops are shown on the corresponding trek of the 8th day.

From Arta, known for its historical bridge, we change direction by turning left to the provincial road Arta-Mouzaki, at the beginning of which we cross again the Ionian flysch with its monotonous turbidite layering until the tectonic contact of the Ionian zone with the Gavrovo zone at the foot of Gavrovo Mountain. Here, the Ionian flysch is overthrust by the Cretaceous Gavrovo neritic limestones (fig. H3). Further to the north, in Tzoumerka Mountain, we distinguish the tectonic klippe of the Pindos units on the Ionian flysch, possibly due to a younger out-of-sequence thrust-fault within the Pindos zone (fig. H3). Our cross-section continues in the Gavrovo zone with outcrops of Cretaceous and nummulite-bearing Eocene limestones, and once again flysch, but this time the Eocene-Oligocene Gavrovo flysch, intensively folded and internally imbricated (fig. H4).

The Gavrovo flysch formation continues until Acheloos River, where the Pindos zone overthrusts the Gavrovo zone (fig. H5). Here, the Triassic Pindos clastic sediments and deep-wated limestones form an impressive thrust-sheet over the Eocene-Oligocene Gavrovo flysch (fig. H5, H6). The Pindos zone itself, as described earlier in the first part (fig. 7), is characterized by deep-sea sediments: radiolarites, cherts, shalles, pelagic carbonates usually of red colors, intercalated with cherts, and the under discussion Early Cretaceous Pindos first flysch deposits (fig. H6, H7, H8). This occurrence of Pindos zone as an overthrust on the Gavrovo zone constitues a far-travelled tectonic nappe with a very complicated internal Tertiary structure, in which numerous thrust-sheets and folds within its formations are multiply repeated (fig. 8). This complicated thrust- and fold- structure, as well as the composition of the Pindos nappe, is studied all the way across Acheloos River and the village of Piges until the Koziakas unit before the village of Mouzaki (fig. H9, H10).

At the source of the Koziakas stream, occurs the overthrust of Koziakas unit onto the Cretaceous Beotian flysch that is composed of radiolarian layers, cherts and shales, as well as serpentinite-pebbles (fig. H11). Then, the Beotian flysch and the overlying Koziakas unit, together, overthrust the Pindos flysch. The Koziakas unit is consisted of Triassic deep-water limestones imbricated with cherts and Jurassic-Early Cretaceous redeposited clastic sediments (fig. H12), as well as oolithic limestones tectonically overlain by ophiolite mélanges, equivalent to the Avdella mélange

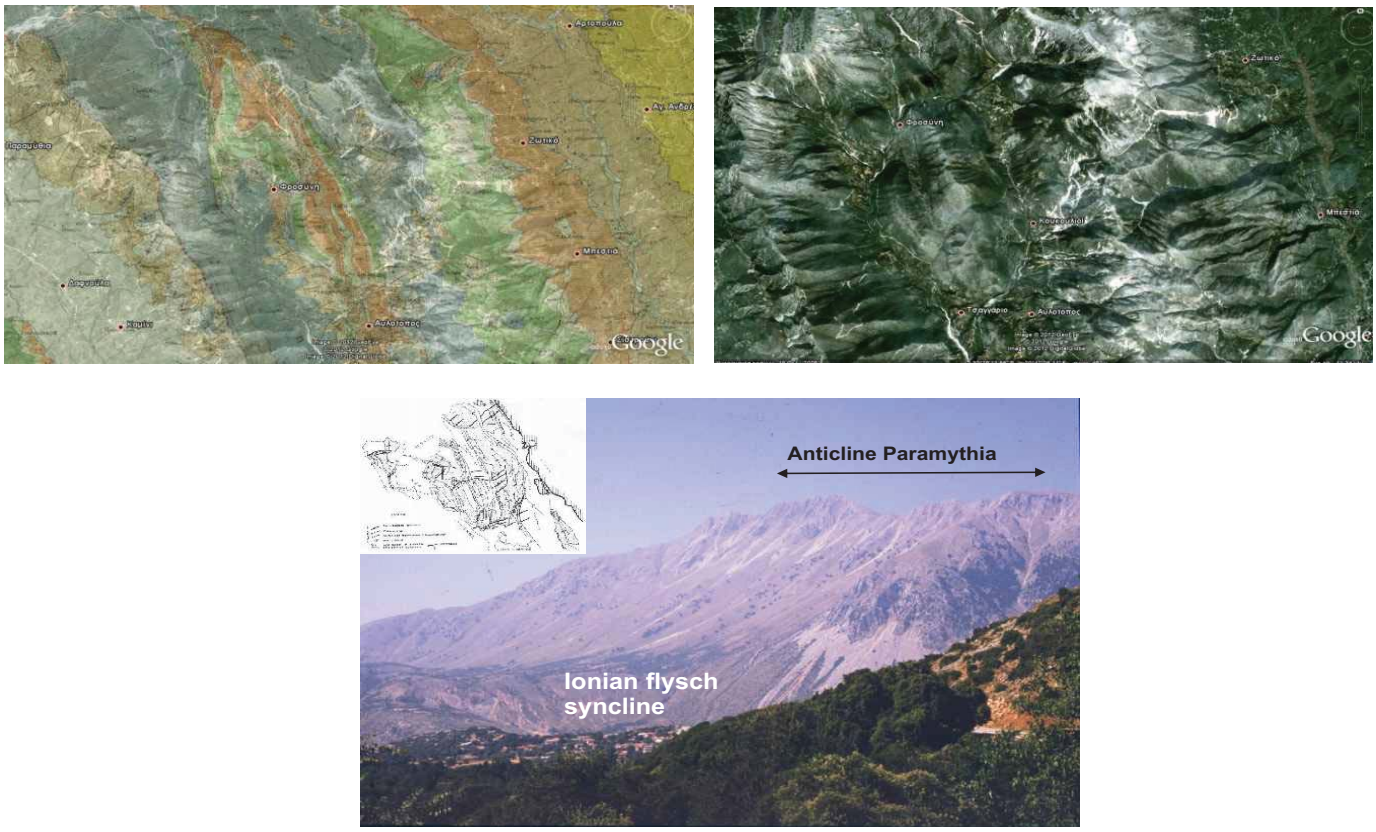


Figure H1
The Ionian zone is characterized by mega-anticline and -syncline structures trending to the NNW and usually W-wards vergent. Nevertheless, some of these such as the anticline east of the Ioannina city (i.e. the Mitsikeli anticline) show opposite kinematics with back-thrust verging to the NE.

Figure H2
The Lower Jurassic, Pantokratoras neritic limestones with their characteristic white color. 39°15'37"N 20°50'49"E

and ophiolites (of basic and ultrabasic composition; see also "Day 6"). The ophiolite mélanges and ophiolites were firstly emplaced on the Triassic Koziakas limestones in the Late Jurassic. However, the geotectonic position of the Koziakas unit remains in debate until today. Finally, an out-of-sequnece low angle thrust fault in the road from the village of Mesochora to Trikala city is included

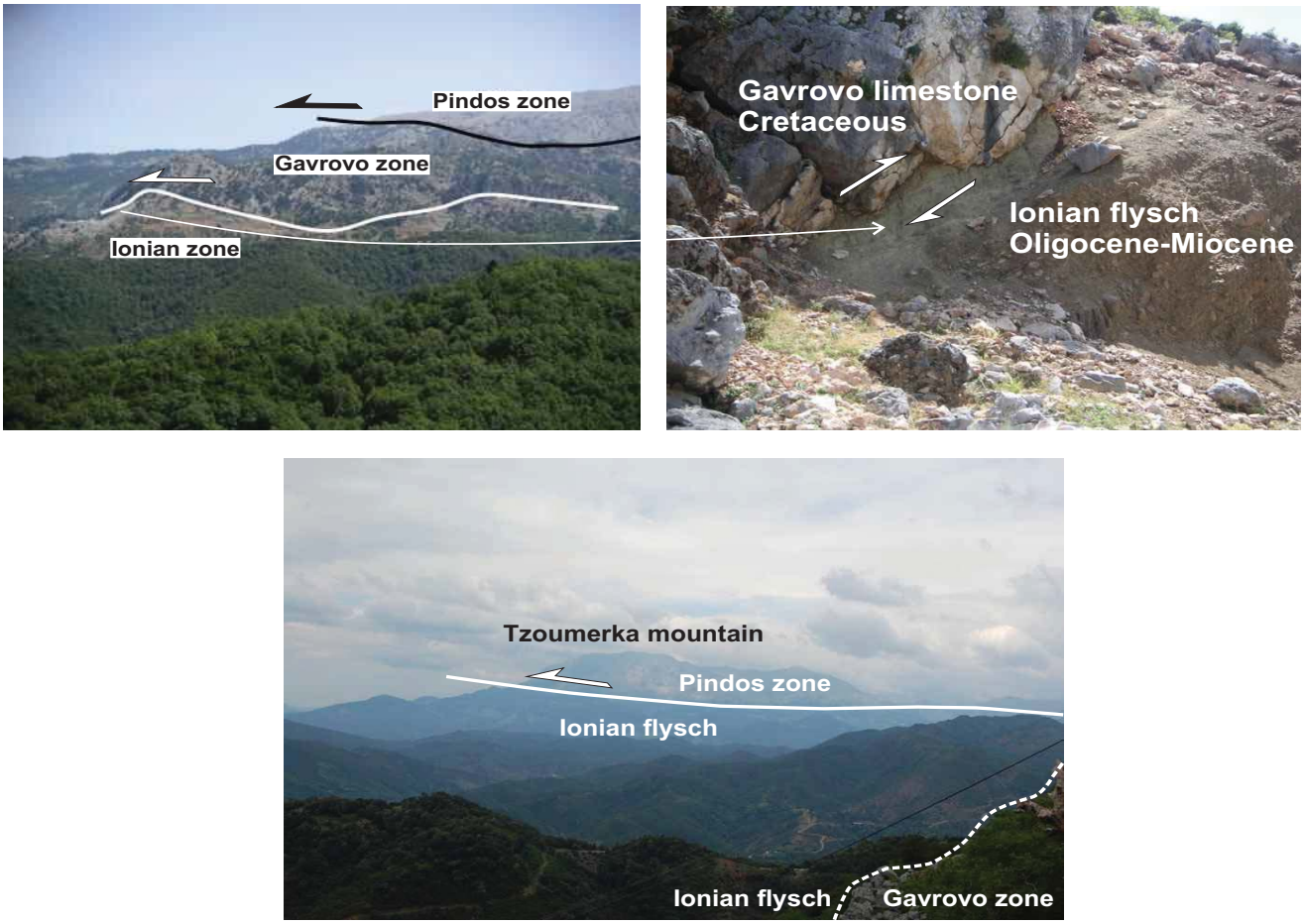


Figure H3
The nappe-pile structure of External Hellenides. The Pindos zone overthrusted the Gavrovo zone during the Eocene-Oligocene and in the next stage the latter progressively overthrusted the Ionian zone during the Oligocene-Miocene. In some places the Pindos zone, possibly due to an out-of-sequence thrust fault, has been thrusted during the Miocene over the Ionian zone, covering the Oligocene-Miocene tectonic contact between Gavrovo and Ionian zones (View to the North). The Cretaceous, Gavrovo neritic carbonatee overthrusting the Oligocene-Miocene Ionian flysch is shown. Younger, extensional low angle en-enchelon faults affect the thrust fault between the two zones. 39°14'55"N 21°13'58"E

Figure H4
The Oligocene-Miocene Gavrovo flysch, intensively deformed by thrust and fold structures, that constitute the characteristic synorogenetic compressional fabric of the flysch sedimentary phase. 39°17'14"N 21°18'50"E

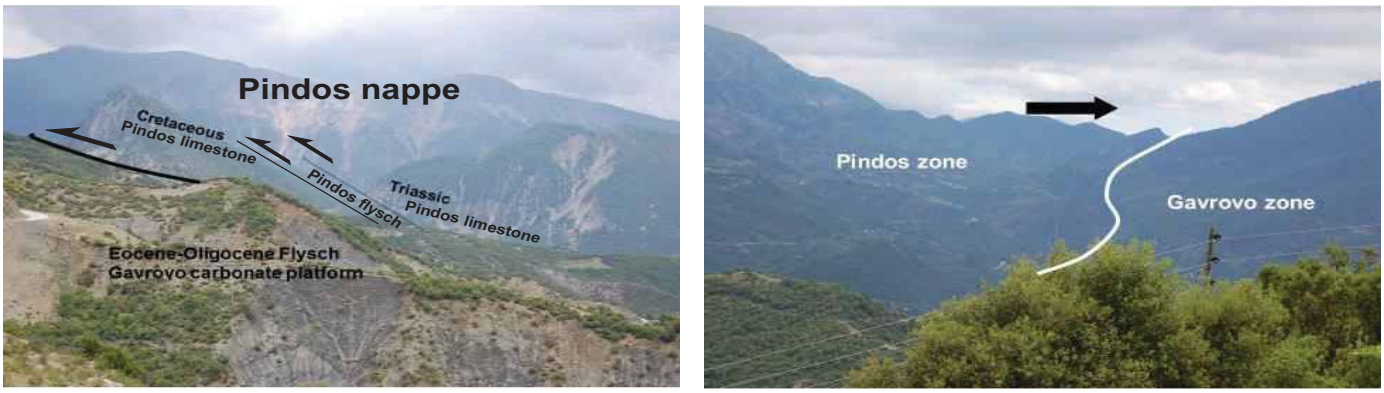


Figure H5
The Pindos zone overthrusted the Gavrovo flysch, simultaneously with the intensive, internal imbrication and folding of the Pindos series. Within the Pindos series the Triassic limestones overthrust the Paleocene-Eocene flysch lying over the Upper Cretaceous deep-water Pindos carbonate sequence. View to the East. 39°17'16"N 21°25'28"E

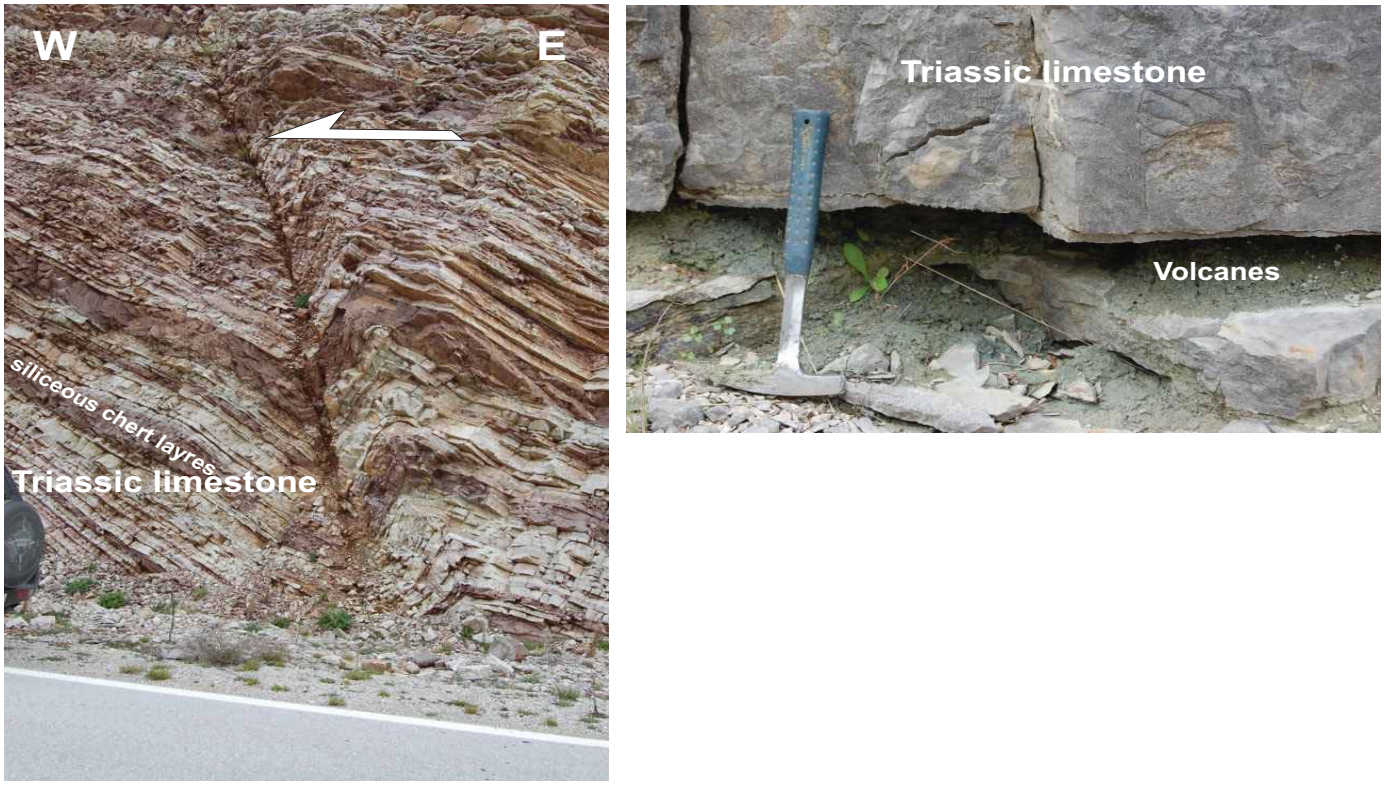
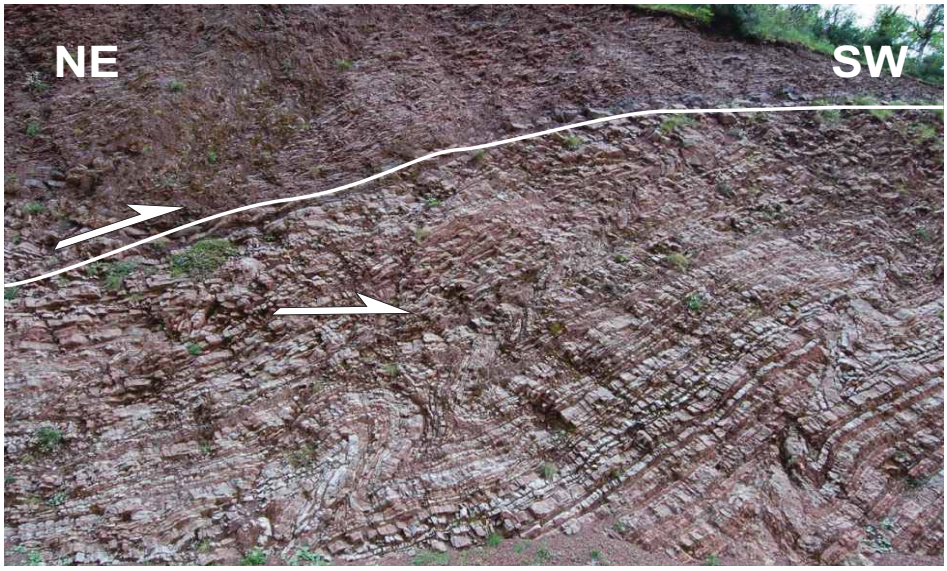


Figure H6
Triassic pelagic limestones of the Pindos zone. They are intercalated with siliceous cherts layers, as well as acid and intermediate volcanic products, possibly related to the Permo-Triassic continental break of the Pangaea supercontinent. In juxtaposition, the older lowermost sediments of the Pindos zone of Early-Mid Triassic age form a clastic neritic formation wich consists of alternations of fine to coarse grained sandstones, conglomerates, siltstones, marly- and locally platy limestones. Their deposition took place during the initial stages of the Permotriassic continental break of the Pangaea and the Neotethys opening. 39°17'52"N 21°25'54"E, 39°17'43"N 21°25'57"E

Figure H7
Intensively folded multi colored Jurassic radiolarian cherts of the Pindos zone. A top-to-the-SW vergence of the asymmetric folds is clearly visible. An out-of-sequence thrust fault cuts the formation with a SW-ward sense of movement. 39°17'32"N 21°25'55"E



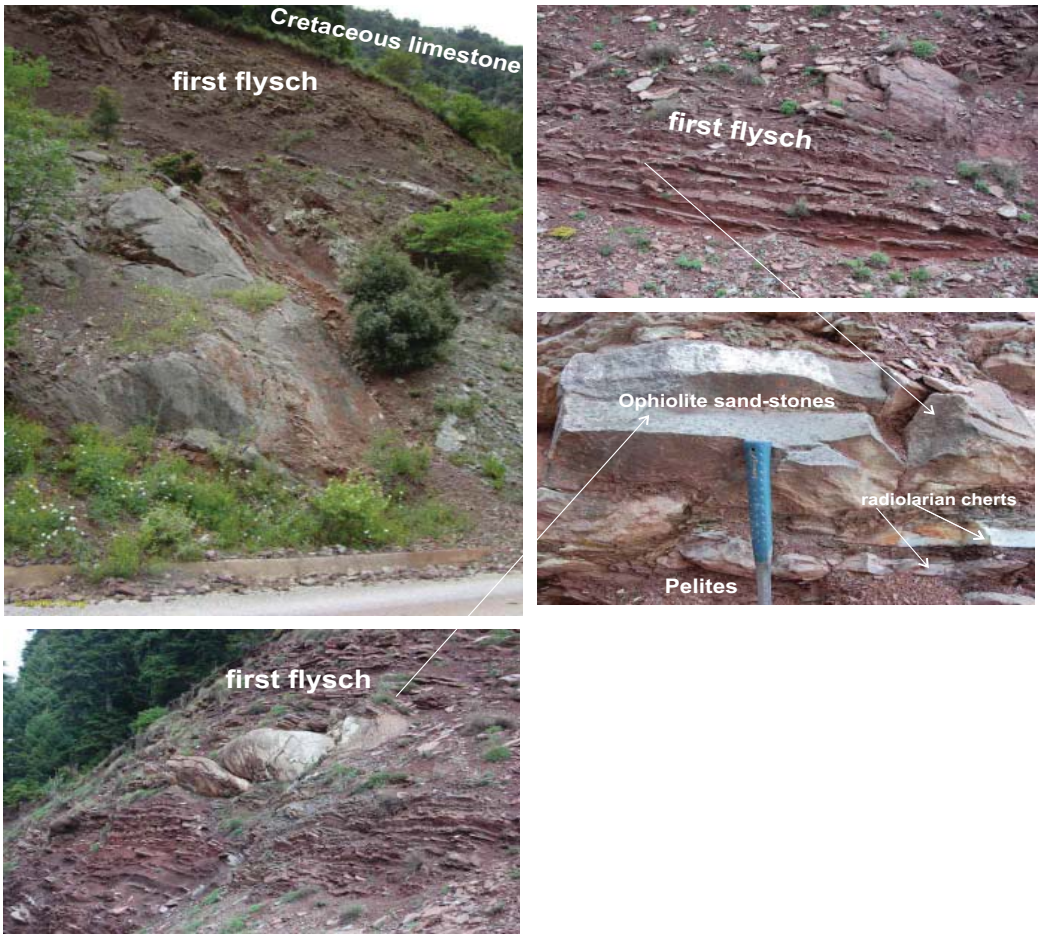


Figure H8
The Lower Cretaceous (Valanginian-Turonian) ?first flysch of the Pindos zone overlain by the Upper Cretaceous pelagic Pindos limestones and underlying by calpioneles-bearing limestones and radiolarian cherts of Tithonian-Valanginian age. All formations are strongly multifolded and imbricated, a common structure of the Pindos zone or Pindos nappe. The Pindos first flysch is composed of alternations of thin layers of red marls and radiolarian cherts, marly limestones, pelites and green fine to coarse-grained sandstones with ophiolite material, as well as ophiolite pebbles. Its geotectonic position and tectonic setting remain under debate until today. 39°18'12"N 21°24'50"E

Figure H9
The calpioneles-bearing Pindos formation of Tithonian-Valanginian age and the Upper Cretaceous Pindos limestones, intensively folded with zick-zack tight folds. The calpioneles-bearing formation consists of red and yellow-green pelagic and in places brecciated limestones intercalated with red-green radiolarian cherts, thin layers of marly-carbonate material and calcareous-sandstones. View to the West. 39°18'40"N 21°24'58"E

Figure H10
The Middle-Upper Jurassic multicolored Pindos deep-water formation, strongly folded. It is composed of alternations of green, blue, brown, red and black cherts and sometime thin layers of pelagic limestones. The formation is underlain the calpioneles-bearing formation. 39°19'40"N 21°28'39"E

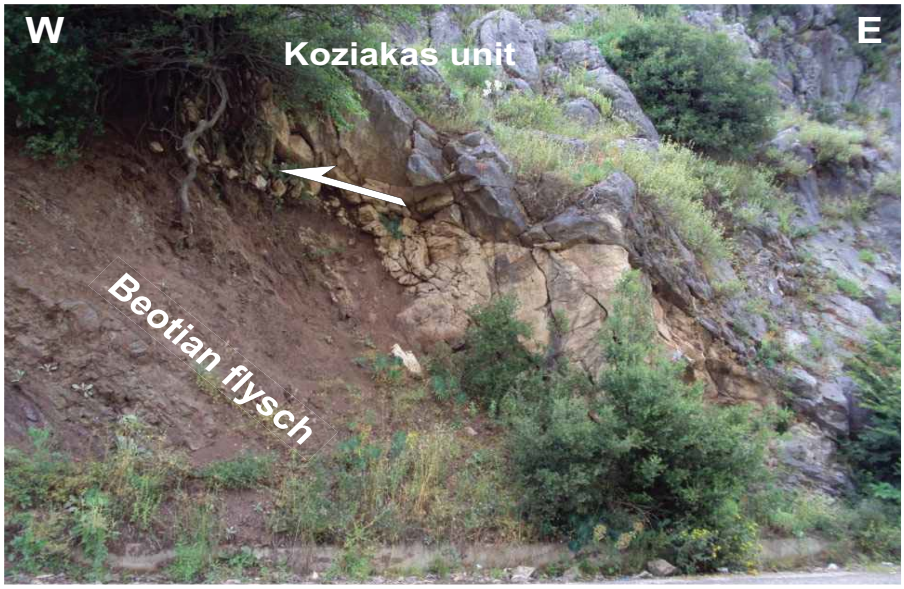


Figure H11
The Koziakas unit thrusts over the Lower Cretaceous Beotian flysch, possibly equivalent to the Pindos first flysch. The two formations together are thrusted W-wards over the Paleocene-Eocene Pindos flysch. 39°24'34"N 21°39'53"E

Figure H12
Strongly folded Triassic deep-water Koziakas limestones intercalated with thin layers of black and yellow-green radiolarian cherts. They form the deeper parts of the Koziakas unit overthrust by the Pindos/Sub-Pelagonian ophiolite belt (Middle-Late Jurassic ophiolite mélanges and ophiolites). 39°27'24"N 21°35'48"E

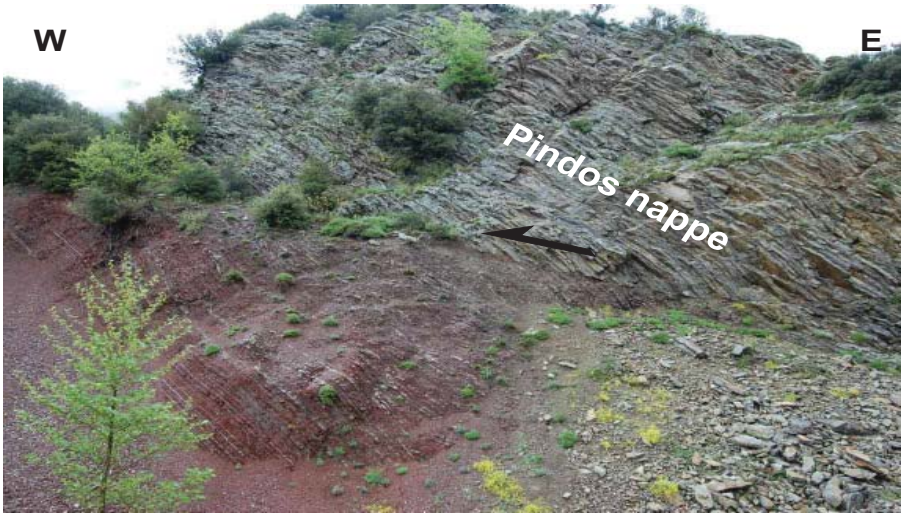


Figure H13
An out-of-sequence thrust fault in the Pindos nappe. Calpioneles- bearing red limestones thrust with a low angle towards west over red radiolarian cherts near the Pindos overthrusting on the Oligocene-Miocene Ionian flysch. The illustrated thrust plane has a lower dip- angle than the layering of the Pindos units. 39°23'28"N 21°16'07"E

in our outcrops list.

As we leave behind us the village of Mouzaki and Koziakas Mountain, we cross the Oligocene and Miocene molassic sediments of the MHT, whose contact with the Koziakas unit is defined by a high-angle dextral strike-slip fault. Last we can see the Neogene-Quaternary sediments of the Thessalian basin which covered the majority of the molassic sediments in this area.

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