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Aspects of the tectonic significance of syn-tectonic Tertiary sediments in the Hellenide orogen: a literature study

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Cover image: traditional Greek arch stone bridge with the Pindos Mountains in the back.
Photo: Uwe Ring.

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Abstract

The complex geological history of Greece, involving subduction, collision and extension, has resulted in formation of several major basins that have since been filled with various sediments. By studying the heavy minerals and fission track ages of detrital apatites in these sediments, it is possible to reconstruct the source terrains and tectonic processes during the Paleogene. This thesis aims to summarize heavy mineral analyses from several published papers, as well as fission track analysis of detrital apatite. The focus is on several major flysch (Pelagonian, Pindos, Gavrovo-Tripolitza and Ionian) and foreland (Thrace and Mesohellenic) basins in Greece. These studies reveal several different sources, including granitic, metamorphic and ophiolitic ones, which reflect processes such as oceanic closure, nappe stacking, basin progradation and extensional faulting.

Introduction

The Hellenides in Greece formed by sustained subduction of Neotethys (Vardar Ocean in Greek transect) and intervening underthrusting of the Apulian continental fragments (e.g., Adria) underneath Eurasian since the Jurassic (Stampfli *et al*., 1998). During the Tertiary, a number of flysch and foreland basins formed (e.g., Mountrakis, 1986; Doutsos *et al*., 1993).

This thesis examines the tectonic evolution of the Hellenides through the analysis of syntectonic sedimentary basins that have developed within several tectonic zones. The primary objective is to investigate the sediment provenance, tectonic contexts and geologic history of these basins by summarizing heavy minerals and detrital apatite age analysis from existing literature.

The Hellenides are subdivided into tectonic zones, which are characterized by rock type, stratigraphy, tectonometamorphic history, and preorogenic paleogeography (Robertson *et al*., 1991). In the Hellenides of the Aegean transect, these are, from northeast (top) to southwest (bottom), (a) the Rhodope zone, (b) the Vardar oceanic zone, (c) the Pelagonian zone, (d) the Pindos zone (including the Cycladic Blueschist Unit), (e) the Gavrovo-Tripolitza zone, (f) the Ionian zone (Jacobshagen, 1986; van Hinsbergen *et al*., 2005). The tectonic zones with relevant flysch basins for this study are:

Rhodope, Pelagonian, Pindos, Gavrovo-Tripolitza and Ionian zones (Fig. 1).

The foreland/extensional basins do not belong to a distinct tectonic zone, instead they were formed after the tectonic zones were juxtaposed with each other. In this study, we focus on the Thrace and Mesohellenic basins (Fig. 1). The Mesohellenic basin evolved as a foreland basin above Neotethyan ophiolites and the Pelagonian zone. The Thrace basin evolved as a Paleogene supra-detachment basin during the Eocene-Oligocene, above the extended Rhodope zone (Vamvaka, 2015).

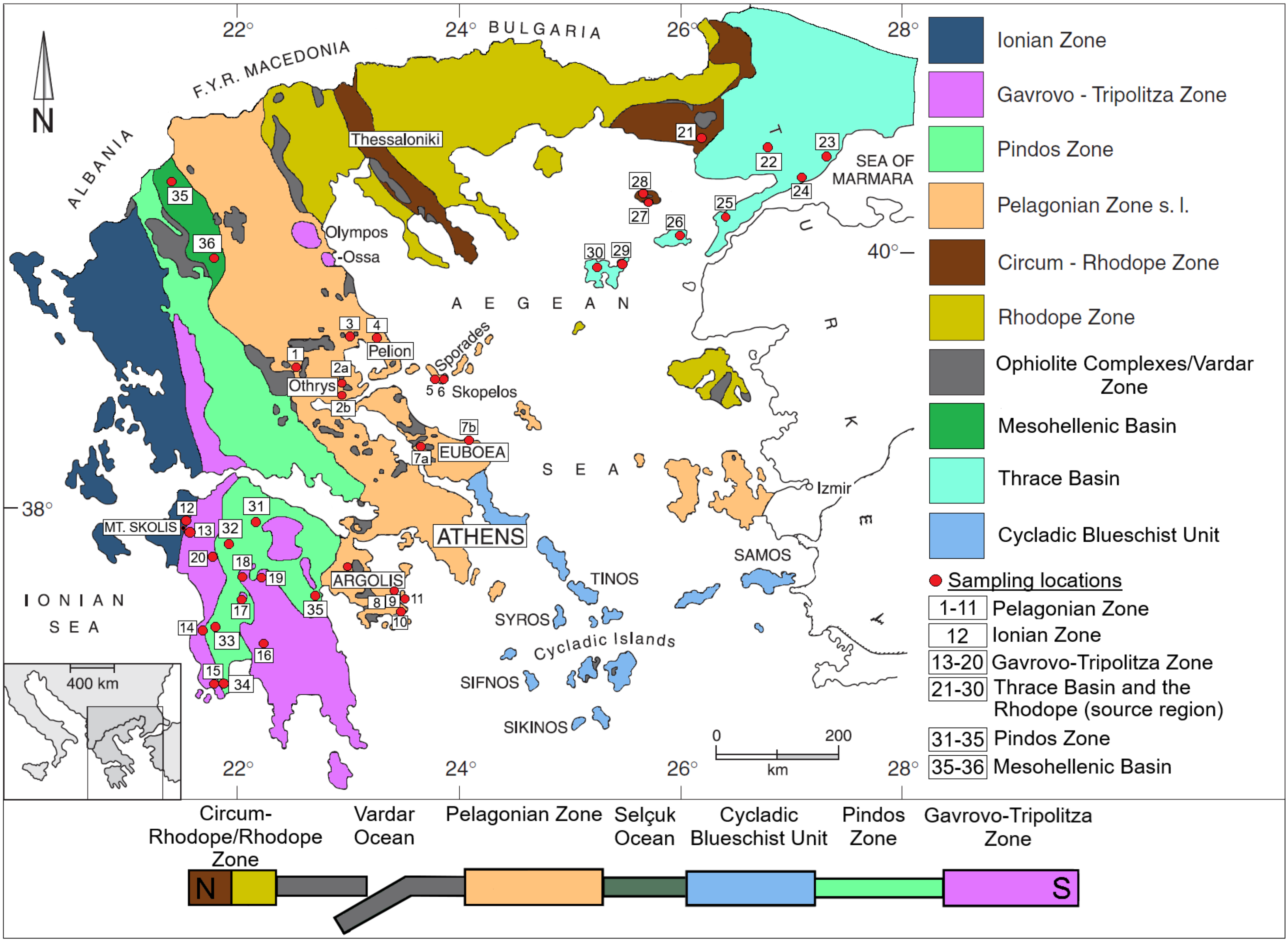

Figure 1. *Geographical location of the tectonic zones and sampling locations based on studies from Faupl et al. (1999), Faupl et al. (2002), Caracciolo et al. (2014), and Vamvaka et al. (2009), with a paleosketch showing the zones from north to south (Ring et al., 2022). The Circum-Rhodope and Rhodope zones, as well as Ophiolite Complexes are also shown. The ophiolite complexes belong to the Vardar Zone. The tectonic zones are divided into External and Internal Hellenides. The Pindos, Gavrovo-Tripolitza and Ionian zones are considered as External Hellenides (Brunn, 1956). The Internal Hellenides include the Pelagonian, Circum-Rhodope and Rhodope zones (Jacobshagen, 1986; Katsikatsos, 1992).*

Geological setting of the tectonic zones and foreland/extensional basins

We first summarize the depositional history of the tectonic zones important for this study followed by a summary of the Thrace and Mesohellenic basins.

Rhodope zone

The Rhodope zone is of interest because it is a potential source region for the sediments in the Thrace basin (Fig. 1). It is characterized by high to medium grade metamorphic units that form a complex metamorphic nappe pile. This includes metamorphic basement which consists of gneisses and schists, Mesozoic ophiolites and volcano-sedimentary rocks and Tertiary volcano-sedimentary basins (Nesbitt *et al*., 1988).

The process of Alpine nappe stacking and crustal thickening was primarily driven by the accretion of several continental units during the Early to Late Cretaceous, which were located between Eurasia to the north and Pelagonia to the southwest (Bonev *et al*., 2006; Burg *et al*., 1996; Georgiev *et al*., 2010; Jahn-Awe *et al*., 2010; Zagorchev, 1998). In the Greek part of the Thrace basin and extending into Bulgaria, Tertiary sediments, approximately 3–5 km thick overlie the Rhodope basement complex (Kilias et al., 2013).

Pelagonian zone

The NW-SE striking Pelagonian zone (Fig. 1), located in the Republic of North Macedonia and Greece, is characterized by thrust sheets that record Alpine orogenic events (Aubouin *et al*., 1976). The Pelagonian zone experienced Late Paleozoic-Early Triassic rifting during the breakup of Pangea, leading to the formation of a passive margin facing the Neotethys/Vardar ocean (e.g., Bortolotti and Principi, 2005; Stampfli and Borel, 2002), with thick sedimentary deposits consisting of Permian and Early Triassic synrift and postrift sequences on top of Variscan (370-290 Ma) basement, as well as extensive carbonate platform sediments from the Middle Triassic to the Middle Jurassic (e.g., De Bono, 1998; De Bono *et al*., 2001; Scherreiks *et al*., 2010).

The Pelagonian carbonate platforms drowned in the Middle-Late Jurassic, indicated by reef debris carbonates mixed with nodular cherty carbonates. Late Jurassic radiolarites and greywackes are present in the deeper parts of the Middle-Late Jurassic sequence (Danelian & Robertson, 2001; Robertson, 1991; Scherreiks, 2000). During the Paleocene and Eocene, deep-water sediments spread over the entire Pelagonian unit, forming the Pelagonian flysch which marked the end of the sedimentary evolution in this basin (Faupl *et al*., 1999).

The metamorphic grade of the rocks of the Pelagonian zone is variable and includes Variscan (380-280 Ma) and Cretaceous/Tertiary metamorphic rocks. The latter includes rare blueschist-facies rocks and more widespread greenschist to amphibolite-facies rocks (Diamantopoulos *et al*., 2008).

Pindos zone

The sedimentary sequences of the Pindos zone (Fig. 1) consist of deep-water carbonates, radiolarites, and siliciclastic rocks that date from the Late Triassic to the Eocene. These formations are tectonically arranged into SW-vergent thrust sheets, covering a significant area of mainland Greece and the Peloponnese peninsula (Faupl *et al*., 2002).
Flysch deposits on the Greek mainland are as young as Late Oligocene (Richter *et al*., 1993). Indications of the beginning of terrigenous flysch sedimentation in the southwest Peloponnese are evident from the Late Maastrichtian period (72 to 66 Ma) (Neumann *et al*., 1996), with the succession ending already in the Middle Eocene and thus earlier than on the mainland (Neumann *et al*., 2000).

The rocks of the Pindos zone are mostly unmetamorphosed, some rocks have very low-grade metamorphism. The Cycladic Blueschist Unit (Fig. 1) in the central Aegean Sea region is commonly correlated with the Pindos zone and contains numerous eclogites, blueschist and greenschist facies rocks (Jolivet and Brun, 2010; Ring *et al*., 2010).

Gavrovo-Tripolitza and Ionian zones

The Gavrovo-Tripolitza platform (Fig. 1) consists of two subzones with different geological evolution and stratigraphy (Zambetakis-Lekkas, 2007).

The Gavrovo subzone is in the southwestern (external) part of the platform. It is found in the Gavrovo area in western Greece and includes the Klokova and Varassova mountains in southern continental Greece, as well as the Skolis and Pylos mountains on the western Peloponnese. Its stratigraphic column features a thick neritic carbonate succession that dates from the Late Jurassic to the Late Eocene, which then transitions into flysch (Zambetakis-Lekkas, 2007).

The Tripolitza subzone, located in the northeastern part of the platform and outcropping in the eastern Peloponnese, Cythere, Crete, and Rhodes, features a stratigraphic column that is less thick than the Gavrovo strata, but more complete. At its base is a detrital volcanosedimentary series from the Late Paleozoic to Late Triassic period, named "Tyros beds." These beds were formed in a very shallow intertidal to shallow subtidal environment, exhibiting lagunal-marine characteristics. The sedimentation from the Late Triassic through the Late Eocene is mostly characterized by monotonous shallow water carbonate deposition. However, the carbonate deposits of the Tripolitza subunit are less rich in microfossils and exhibit a greater degree of dolomitization than the Gavrovo carbonates (Zambetakis-Lekkas, 2007). The metamorphic grade is similar to that of the Pindos zone.

The Ionian zone spans from Albania in the northeast, and extending southwestwards into Central Greece, the Peloponnese, Crete, and the Dodecanese Islands. It is comprised of sedimentary

rocks that range from Triassic evaporites to Jurassic-Late Eocene carbonates, with minor cherts and shales. It is overlain by Oligocene flysch deposits (Zoumpouli *et al*., 2022).
During the Early Triassic to the Middle Early Jurassic period, the Ionian basin was a part of an extensive shallow marine platform in the pre-rift stage. This platform later developed into a vast basin, which was surrounded by shallow platforms; the Apulia platform to the west and the Gavrovo platform to the east. The oldest rocks in this region are Triassic evaporites and carbonates, with Early to Middle Triassic evaporite (Zoumpouli *et al*., 2022).
The Ionian zone includes low-grade high-pressure rocks. However, these high-pressure rocks commonly do not contain blue amphibole.

In the northwestern Peloponnese, flysch deposits west of Mount Skolis (location shown in Fig. 1) are classified as Ionian zone, while those to the east are Gavrovo subzone. The steep summit of Mount Skolis represents the Gavrovo thrust front which consists of Cretaceous-Eocene shallow-marine carbonates (Faupl *et al*., 2002).

We discuss the Gavrovo-Tripolitza and Ionian basins together in this thesis. Richter (1974) reported sharp facies changes across flysch successions. This can be useful in identifying source terrains for both basins, as discussed in the data analysis section.

Thrace foreland basin

The Thrace basin (Fig. 1) is a major Tertiary basin in the North Aegean region and it formed on the metamorphic rocks of the Rhodope zone (Fig.1) in northern Greece (Kilias *et al*., 2006; Kopp, 1965; Lescuyer *et al*., 2003; Tranos, 2009) and the Strandja and Sakarya massifs in northwest Turkey, where the depocenter (part of the basin with maximum sediment thickness) of the basin evolved. It is debated whether the Thrace basin formed as a typical foreland basin later affected by extensional deformation or whether it formed primarily as an extensional basin. Brun and Sokoutis (2007); Burchfiel *et al*. (2003); Burchfiel *et al*. (2008); Kilias *et al*. (1999) regard the Thrace basin as an extensional basin and suggest the start of extension as early as the Eocene or possibly during the Late Cretaceous-Paleocene period (Bonev *et al*., 2006). In contrast, Dinter and Royden (1993), Dinter *et al*. (1995) attribute the onset of extension to the Miocene (Wawrzenitz and Krohe, 1998). Pre-Miocene extension coincided with compression in the outer regions of the Hellenides mountain range (Burchfiel *et al*., 2008; Jolivet and Brun, 2010; Kilias *et al*., 1991; Kilias *et al*., 1999; Schermer *et al*., 1990).

The basin deposits include Paleogene molassic sedimentary rocks, which is typical for foreland basins, along with Neogene and Quaternary sediments (Burchfiel *et al*., 2003; Christodoulou, 1958; Görür and Okay, 1996; Kopp, 1965; Meinhold and BonDagher-Fadel, 2010; Turgut *et al*., 1991; Zagorchev, 1998). The Paleogene molassic sediments are intercalated with large volumes of calc-alkaline and shoshonitic volcanic rocks. K/Ar dating indicates ages from 33.4 to 20 million years (Early Oligocene to Early Miocene) (Christofides *et al*., 2004). Intercalations of the volcanics with earliest Middle–Late Eocene clastic sediments (Burchfiel *et al*., 2003;

Christodoulou, 1958; Dragomanov *et al*., 1986; Kopp, 1965; Meinhold and BonDagher-Fadel, 2010; Zagorchev, 1998) suggest that volcanic activity already began earlier during the Middle–Late Eocene period.

The Paleogene sediments in the western Thrace basin date from the Middle–Late Eocene to Oligocene (Burchfiel *et al*., 2003; Christodoulou, 1958; Dragomanov *et al*., 1986; Kopp, 1965; Meinhold and BonDagher-Fadel, 2010; Zagorchev, 1998) and include a complex stratigraphic sequence of conglomerates, sandstones, nummulitic limestones, and shales. Sedimentation began in the Lutetian (48-41 Ma)–Bartonian (41-38 Ma) stage with continental deposits, which later transitioned to marine turbiditic deposits and limestones interbedded with volcanogenic products during the Late Eocene–Oligocene (Burchfiel *et al*., 2003; Christodoulou, 1958; Dragomanov *et al*., 1986; Kopp, 1965; Zagorchev, 1998). Overlying the Paleogene strata is a thick Neogene sequence of conglomerates, sandstones, and siltstones, with Quaternary sediments of gravel, clay, and sandstones on top (Kilias *et al*., 2013).

This thesis analyzes the Eocene-Oligocene sedimentary successions exposed at the southern Thrace basin margin to identify the dispersal pathways of eroded crustal elements from oceanic and continental sources and their contributions over time. The sampling locations are on the mainland and several nearby islands (Figure 1).

Mesohellenic foreland basin

The Mesohellenic basin (Fig. 1), also referred to as Mesohellenic Trough, MHT, spans over 200 km in length and 30–40 km in width, and it is located in northern Greece and Albania. It formed between the middle Eocene and Late Miocene along the tectonic contact of the external (non-metamorphic) and internal (metamorphic) zones of the Hellenides, following a general NW–SE orientation that aligns with the orogen's long axis. The sedimentary sequence of the MHT lies on top of the tectonic contact between these two zones (Vamvaka *et al*., 2009).
The infill of the basin mostly comprises marine turbidites and siliciclastic deposits, reaching a maximum thickness of 4 kilometers in certain sections. Variations in thickness and facies are observed across and along the basin axis among the different formations (Zelilidis *et al*., 2002).

The MHT is bound to the southwest by ophiolites and overlying Cretaceous limestones that overthrust the Maastrichtian (72-66 Ma) to Palaeocene Pindos flysch (Brunn 1956; Jones and Robertson 1991; Mountrakis *et al*., 1993). Its northeastern boundary is defined by the Pelagonian zone and remnants of obducted Vardar ophiolites (Brunn 1956; Mountrakis 1986; Rassios and Smith 2000; Robertson *et al*., 1996; Dilek *et al*., 2005).

The MHT consists of five mainly siliciclastic formations. The Krania Formation at the base, from the Middle to Late Eocene, has a maximum thickness of about 1500m (Wilson, 1993; Zelilidis *et al*., 2002). It is located at the southwestern basin margin and consists of conglomerates, Lutetian (48-41 Ma) limestones, and marly turbiditic sequences (Vamvaka *et al*., 2009).

The Oligocene Eptachori Formation is less than 1000m thick and includes conglomerates, sandstones, and marine turbiditic shales (e.g. Savoyat and Monopolis 1972; Zygogiannis and Müller 1982; Barbieri 1992; Zelilidis *et a*l., 2002). The Pentalophos Formation is of Early Oligocene to Early Miocene age and has a thickness of around 2500m (Zelilidis *et al*., 1997, 2002). It is characterized by conglomerates and turbiditic sandstones. The Early to Middle Miocene Tsotyli Formation exceeds 2000m in thickness (Mavridis and Matarangas, 1979; Mavridis and Kelepertzis, 1985), and contains conglomerates, sandstones, and shales.
The Ondria Formation, from the middle to Late Miocene, is 600m thick (Savoyat *et al*., 1971) and consists of shallow-water sandstones and limestones. All formations, except the Krania Formation, become coarser southwestward, indicating a more proximal position to the source area in the southwestern part characterized by delta fan deposits (Zelilidis *et al*., 2002).

Methods

For the flysch basins and for the Thrace basin our analysis is based on detrital heavy mineral work. For the Mesohellenic basin, such a dataset does not exist, and we use fission-track (FT) dating of detrital apatite. The circular diagrams representing heavy mineral compositions from Pindos, Ionian and Gavrovo-Tripolitza zones were constructed from the data table provided in Faupl *et al*. (2002). The same data was not publicly available for the Pelagonian zone, but circular diagrams without exact mineral percentages were provided in Faupl *et al*. (1999). The approximate mineral percentages were calculated from these diagrams using a protractor (Fig. 2).

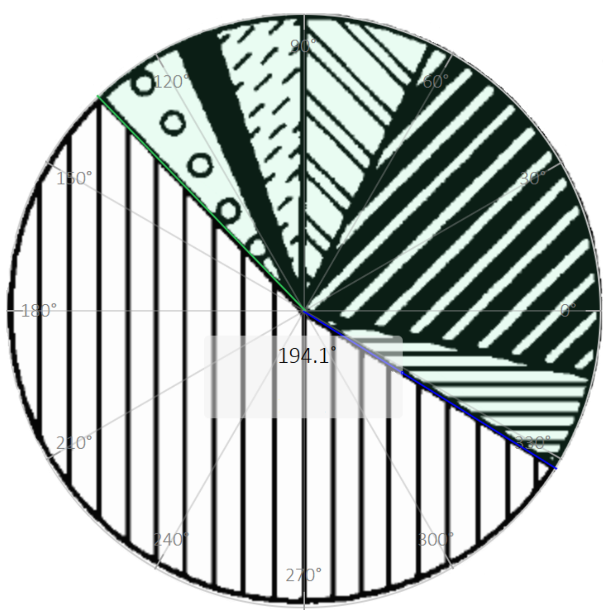

Figure 2. *Protractor method for calculating heavy mineral percentages.*

In the example in Fig. 2, apatite is represented by vertical lines, so its area is equal to 194.1° on a circle. Its percentage can then be obtained with a simple formula: 194.1/360x100=53.9%. For the Thrace basin, no data table nor circular diagrams were available in the paper, and it was stated in the paper that a data table is available as a supplementary material, but was only for authorized users, so it was not publicly available. Due to this, general descriptions of heavy mineral compositions by Caracciolo *et al*. (2014) were used. Below are more detailed methods which were used by authors when obtaining the data for each zone and basin.

Pelagonian, Pindos, Gavrovo-Tripolitza, Ionian zones

For the Pelagonian flysch basin Faupl *et al*. (1999) separated heavy minerals from crushed sandstone samples within a specific size fraction (0.4–0.063 mm). The ribbon counting method was applied to more than 200 heavy minerals per sample. Many of these minerals represented broader groups, which include tourmaline, garnet, and chrome spinel. For larger sample sets, cluster analysis facilitated the objective differentiation of heavy mineral assemblages, utilizing the software WinSTAT® version 3.0 and employing Ward's method for the incremental formation of clusters.

For the Pindos, Gavrovo-Tripolitza and Ionian flysch basins, data was collected and compiled from Faupl *et al*. (2002). The detailed methodologies employed in this study can be found in their earlier work, Faupl *et al*. (1998). For each sample, between 200 and 300 translucent heavy mineral grains were counted. To evaluate paragenetic relationships within the rock complexes of the source area, Spearman's rank correlation coefficients were used, which is a non-parametric statistical method as described by Rock (1988).

Thrace basin

Data for this basin was obtained from Caracciolo *et al*. (2014). Forty-seven sandstone samples from the southern Thrace margin were analyzed for heavy minerals. The analysis used the 63–125 µm grain size fraction, which was separated with sodium polytungstate. Additional data from several locations were referenced from previous studies.

Chemical composition was analyzed with a JEOL electron microprobe to determine concentrations of elements such as Mg, Al, Cr, Fe, V, Si, Ti, Mn, Ni, and Zn.

Table 1 shows the main detrital minerals used in this study and what their significance for potential source rocks is in general.

Table 1. *Common minerals mentioned in the study and their meaning for potential source rocks.*

Mineral	Source
Zircon, tourmaline, rutile (ZTR) group	Mainly associated with granitic source, but it can also be igneous or metamorphic
Chrome spinel	Obducted ophiolites
Staurolite	Amphibolite-facies pelitic rocks
Epidote	Low to medium grade metamorphic rocks
Hornblende	Medium to high grade metamorphic rocks, mainly of mafic composition
Pyroxene	Igneous and metamorphic rocks
Blue amphiboles	High pressure metamorphic rocks (blueschist facies)
Garnet	Most commonly found in metamorphic rocks with T >450°

Mesohellenic basin

A different method was used for this basin, since no detrital heavy mineral data exists. The summary is based on the detrital apatite FT analysis of Vamvaka *et al*. (2009). This analysis uses low-temperature thermochronology data on detrital minerals and compares it to bedrock low-temperature thermochronology data in the potential source region.

Fission tracks are damage trails in the crystal lattice formed by the spontaneous fission of 238U. They are continuously generated and are preserved below a closure temperature of 60°C (Reiners and Brandon, 2006). Fission tracks anneal at elevated temperatures (above 120°C for apatite). The temperature range in between is known as the partial annealing zone (PAZ), between 60-120°C for apatite. The length of the tracks decreases based on the temperature and time spent in the PAZ. This is used for the modeling of the rock's thermal history using track-length distributions and FT ages (Reiners and Brandon, 2006).

Sampling in the Pelagonian basement northeast of the MHT was done in two areas close to the southern and the northern parts of the MHT (Fig. 1) to gather data for fission track (FT) age analysis, as no previous FT ages were reported there. Sediments from the Eocene to Miocene were analyzed to understand provenance and thermal history. FT ages were determined via the external detector method (Gleadow and Duddy, 1981) and calibrated using established programs.

Data from sediment samples was analyzed with PopShare and BinomFit programs for age distribution, which allows to create time–temperature paths for thermal history inversion with HeFTy program by Ketcham (2005), with specific statistical evaluations to assess goodness of fit. Input parameters for model predictions included central FT ages, track length distributions, and kinetic data. First and second AFT ages were calculated from all sediment samples. When dating detrital mineral grains, the individual results cannot be averaged, so it's helpful to separate them into age gaps. This approach is useful in isolation and statistical description of these components.

Determining the sediment source from heavy mineral data

The goal was to discuss changes in heavy mineral content through time, i.e., from the bottom to the top of the flysch deposits. However, biostratigraphic information on the age of the sediments is sparse; therefore, heavy mineral data are usually presented and discussed in a geographic reference frame (Faupl *et al*., 1999).

Pelagonian flysch

During the Paleocene and Eocene, the sedimentation of terrigenous flysch was widespread across the Pelagonian Zone, due to compressional tectonics. The terrigenous contributions can be categorized into two main mineral assemblages; a garnet-rich heavy mineral assemblage found in the flysch sediments of western Othrys (Fig. 3, locality 1), the central Argolis on the eastern Peleponnese (Fig. 3, locality 9), and other assemblages containing various amounts of the stable ZTR group minerals and apatite from all sampling locations (Fig. 3).

The garnet assemblage, exclusive to western regions (Fig 1. locality 7a), is not easy to explain as garnet could reflect high-pressure or Barrovian-type metamorphic rocks in the source area. Faupl *et al*. (1999) link the garnet to Early Cenozoic subduction of an oceanic basin (Selcuk Ocean, Fig. 1) at the southwestern edge of the Pelagonian microcontinent (Fig. 4).

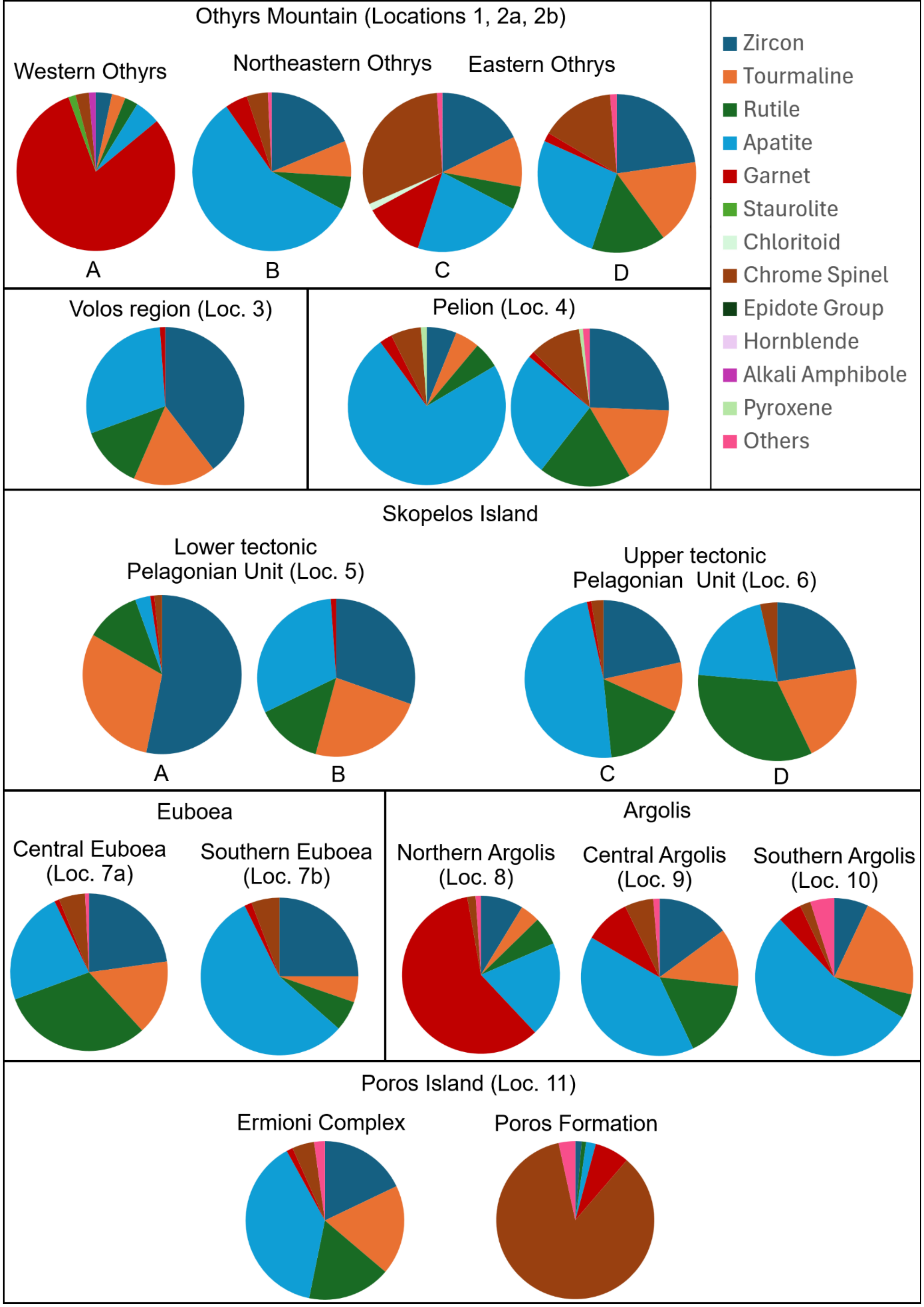

Figure 3. *Average heavy mineral compositions and clusters for the Pelagonian flysch. The letters A-D show major clusters. The location numbers correspond to sampling locations in Fig. 1. Based on the data from Faupl et al. (1999).*

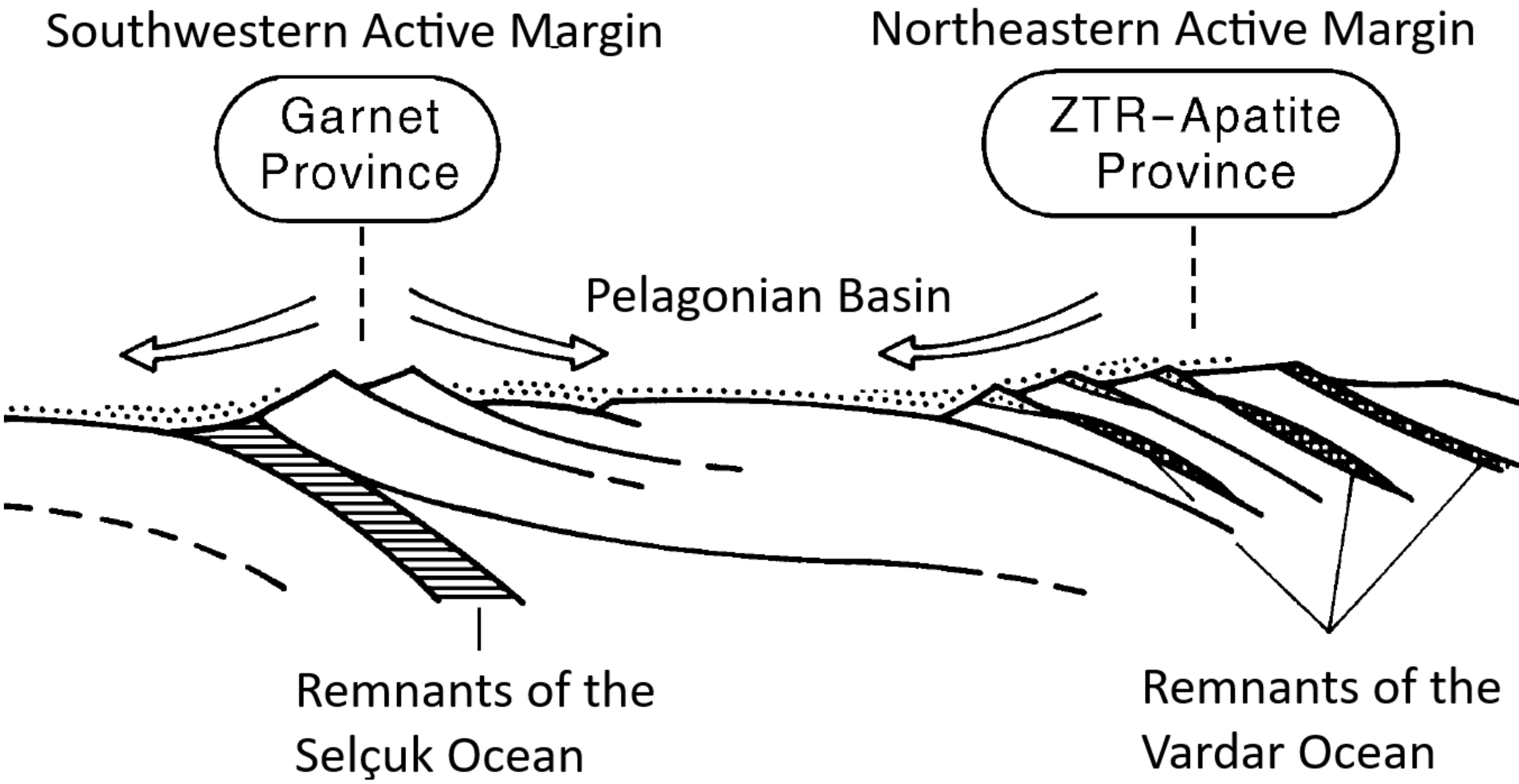

Figure 4. *Sketch of the paleotectonic position of the two major source terrains of the Pelagonian flysch during the Paleocene-Eocene. Modified from Faupl et al. (1999).*

The ZTR group heavy mineral assemblages mostly reflect the composition of a granitoid source located northeast of the Pelagonian flysch basin (Fig. 4). This granitoid source could be Pelagonian basement or the Rhodope zone (Faupl *et al*., 1999).

However, it is possible that all the material originated from the north; garnet from the previously accreted Pelagonian sediments or the Rhodope, and ZTR from the Rhodope (ZTR from the Vardar Zone is unlikely due to presence of ophiolitic materials there). The source of chrome spinel is the Vardar Zone, as this mineral typically represents ophiolites.

Ophiolite complexes contributed only marginally to the Pelagonian flysch sediments. Although an alteration of chrome spinel (which serves as an indicator mineral for ophiolite rocks) through low-grade progressive metamorphism is anticipated in several regions of the Pelagonian Zone, this is likely to occur only to a limited extent. The exposure of blueschist complexes, as indicated by blue amphibole in the detrital record, was confined to the southwestern source province (Faupl *et al*., 1999).

Pindos flysch

Based on the heavy mineral data (Fig. 5), the Paleogene Pindos flysch, source area primarily consists of metamorphic complexes with major lithologies being garnet and staurolite-bearing mica schists and gneisses. Rocks containing blue amphiboles were tectonically integrated within these complexes.

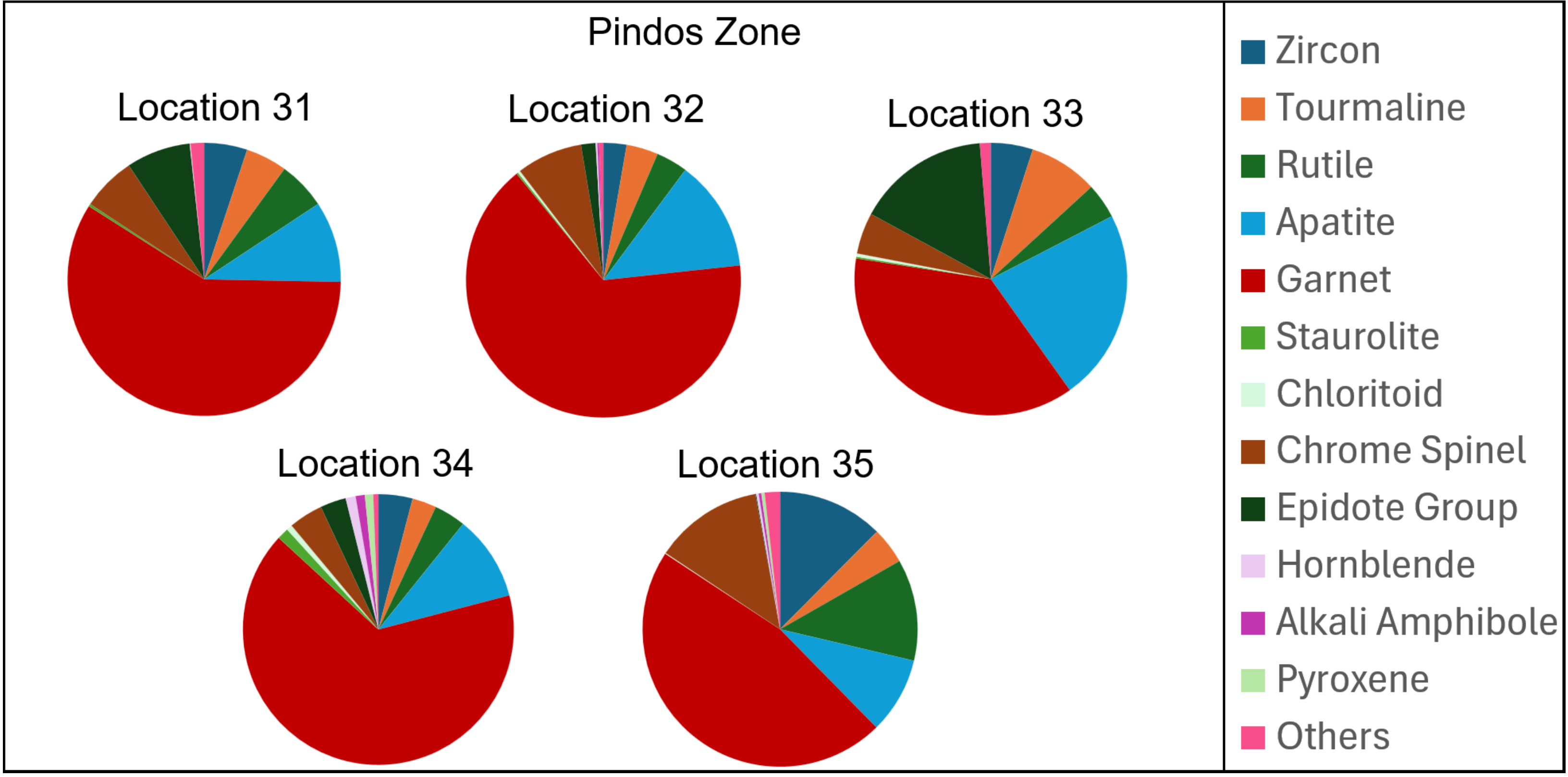

Figure 5. *Heavy mineral composition in the Pindos zone. The location numbers correspond to sampling locations in Fig. 1. Based on the data from Faupl et al. (2002).*

The occurrence of blue amphiboles with minerals indicative of Barrovian metamorphism, e.g., staurolite, chloritoid, and green calcic amphiboles, suggests tectonic stacking in the source area (Faupl *et al*., 2002).

The analysis of heavy mineral assemblages from Faupl *et al.* (2002) indicates an internal source region for these materials. The presence of detrital chrome spinel is more likely derived from obducted ophiolite complexes of the Vardar zone (Faupl *et al*., 2002). The detrital high-pressure minerals can be sourced from an exhumed tectonic wedge in front of (below) the Pelagonian zone (Faupl *et al*., 1999; Petrakakis *et al*., 2001), which would then correlate to the Cycladic Blueschist Unit.

Gavrovo-Tripolitza and Ionian flysch

The separation of Ionian and Gavrovo–Tripolitza flysch around Mount Skolis aligns with the heavy mineral compositions found in mainland Greece (Fig. 7a, b). The Ionian samples show a higher average chrome spinel content compared to the Gavrovo-Tripolitza flysch (Fig. 6). These differences in heavy mineral composition between the western region near Mount Skolis (Ionian flysch) and the central Gavrovo–Tripolitza flysch are supported by Richter (1974), who observed significant facies changes from external southwestern to internal central/northeastern eastern flysch successions on the Peleponnese (Faupl *et al*., 2002).

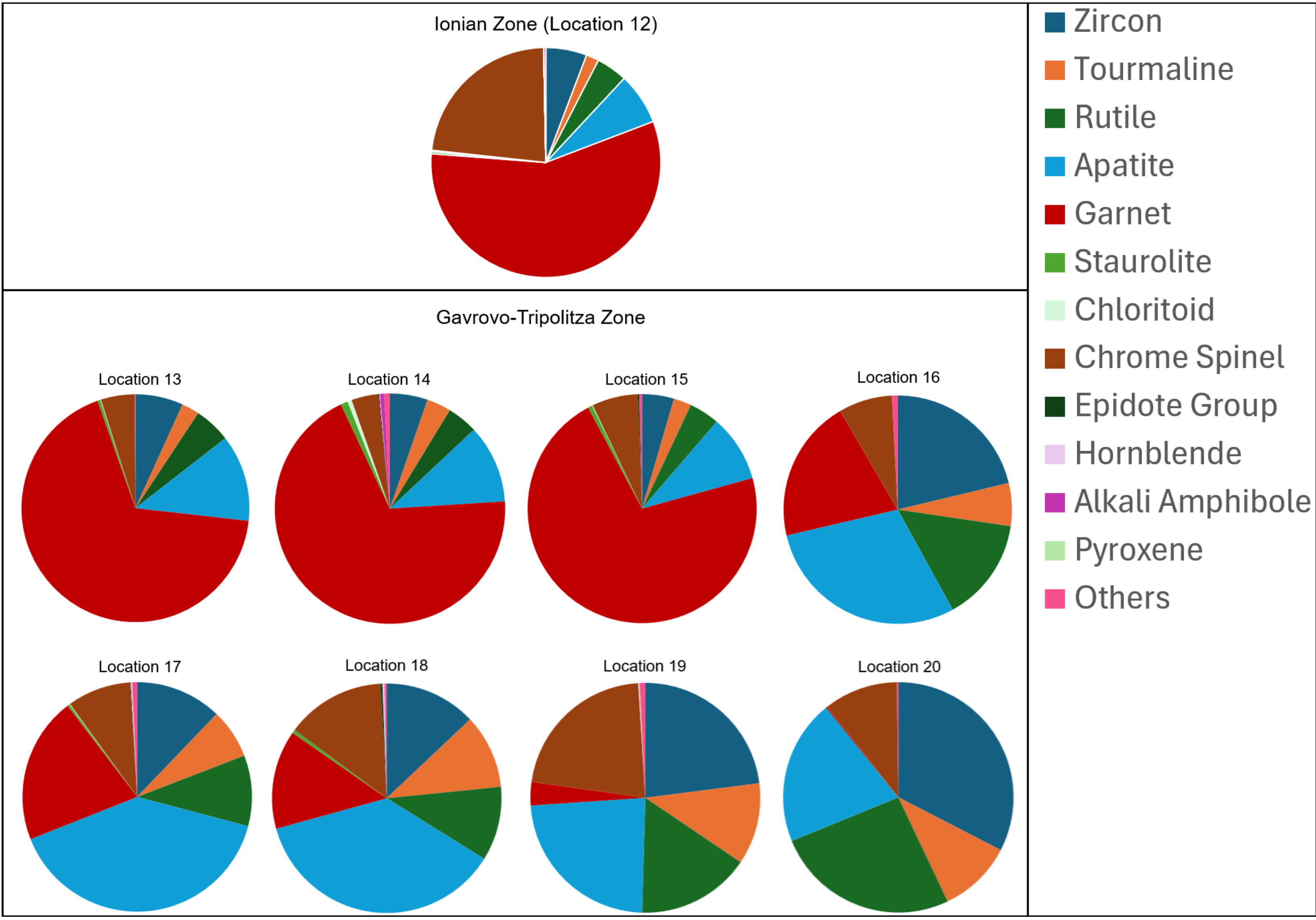

Figure 6. *Average heavy mineral composition Ionian and Gavrovo-Tripolitza Zones. The location numbers correspond to sampling locations in Fig. 1. Based on the data from Faupl et al. (2002).*

In the external successions, thick olistostromes containing material from the Pindos zone are found in younger stratigraphic horizons. This coarse material is absent in the internal successions. Garnet-rich samples from the external region resemble those from the Greek mainland (Fig. 7b) (Faupl *et al*., 1998). The heavy mineral associations in the eastern region (Gavrovo-Tripolitza), particularly the ZTR group minerals and apatite, are in general not documented on the mainland, although some apatite-rich samples were observed there (Faupl *et al*., 2002).

Two distinct heavy mineral associations (Fig. 6 and 7b) may initially suggest external and internal source terrains. However, the presence of coarse Pindos material in the upper sequences of external (southwestern) flysch deposits (Richter, 1974; Neumann *et al*., 2000) contradicts this. Schneider *et al*. (1998) propose a sedimentation model favoring turbidites with a palaeocurrent pattern from an internal (northeastern) source, an interpretation supported by Gonzales-Bonorino (1996).

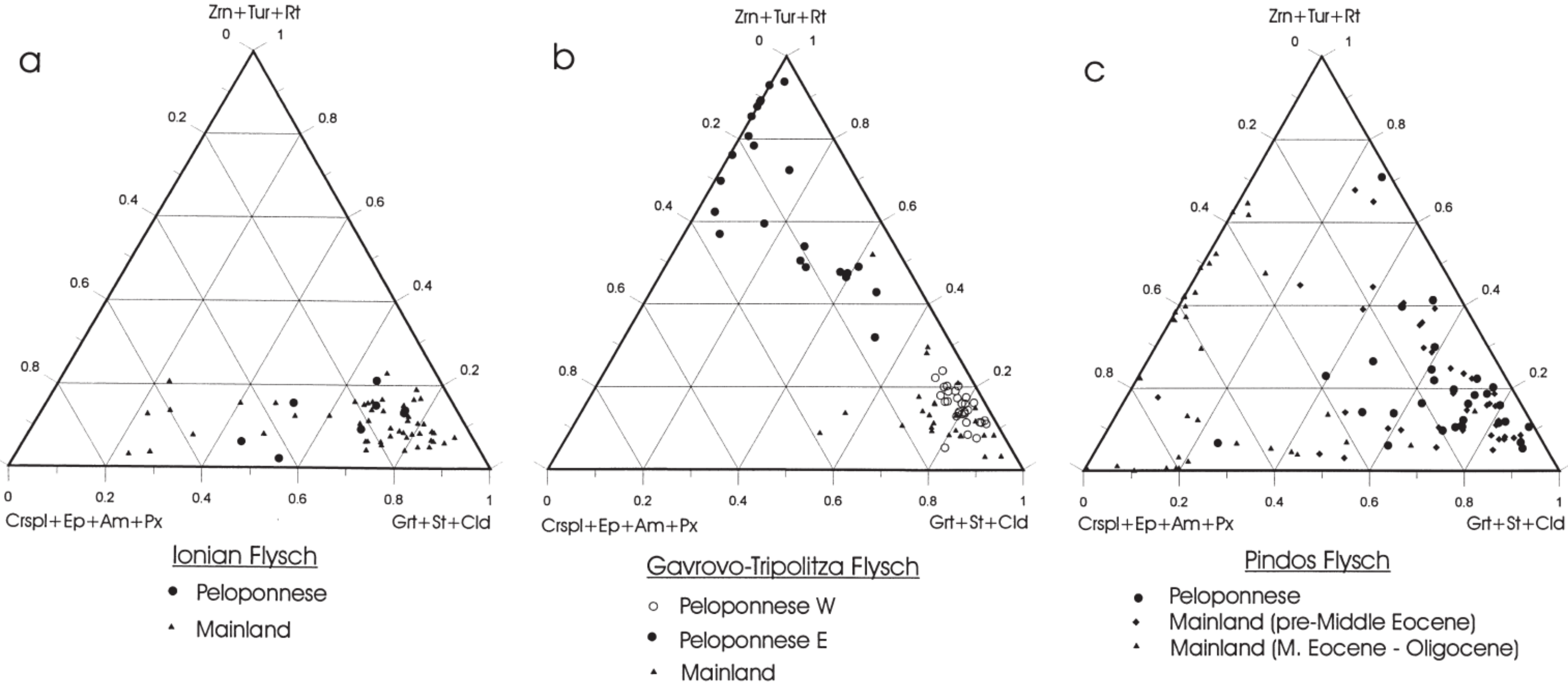

Figure 7. *Heavy mineral composition of the Ionian, Gavrovo-Tripolitza and Pindos flysch on the Peloponnese compared to the heavy mineral composition of these flysch sequences on the Greek mainland. Zrn – zircon, Tur – tourmaline, Rt – rutile, Grt – garnet, St – Staurolite, Cld – chloritoid, Crspl - chrome spinel, Ep – epidote group, Am – green amphibole, Px – pyroxene. Source: Faupl et al. (2002).*

A model addressing issues with the two heavy mineral provinces suggests significant cannibalism of Pindos flysch during the sedimentation of the Gavrovo–Tripolitza flysch, which aligns with documented pebble compositions (Faupl *et al*., 2002). The northeastern heavy mineral assemblages, which are low in garnet, likely result from the reworking of Mid-Eocene–Oligocene Pindos flysch deposits, which have low garnet content and predominance of ZTR group minerals (Faupl *et al*., 2002).

During the southwestward progradation of the Pindos nappe, the Gavrovo–Tripolitza flysch basin axis also shifted southwestward. Heavy mineral assemblages suggest erosion of lower garnet-rich Pindos flysch successions (Fig. 7b, c). The presence of detrital blue sodic amphiboles in some parts of the Gavrovo–Tripolitza flysch supports a model of cannibalism, as these high-pressure minerals are found at the same latitude as in the Pindos flysch (Faupl *et al*., 2002). This progradation indicates a younging of the Gavrovo–Tripolitza flysch succession towards the southwest, supported by biostratigraphic data from Kamberis *et al*. (2000) and Neumann *et al*. (2000).

Thrace foreland basin

The identified heavy minerals in decreasing order of abundance include zircon, Cr-spinel, garnet, rutile, epidote, titanite, tourmaline, allanite, staurolite, kyanite, and glaucophane. The abundant Cr-spinel is linked to the ophiolitic rocks surrounding the southern margin of the Thrace basin. High zircon content is explained by widespread phyllite and fine-grained schists in the nearby units, such as the Makri unit consisting of metasedimentary rocks, located in the Circum-Rhodope Zone (Fig. 1) (Caracciolo *et al*., 2014). However, it would be reasonable to assume that some zircon could have originated from the igneous rocks (granitoids), which can also be found in the Rhodope Zone (Bonev *et al*., 2014). Additionally, there was a volcanic episode which delivered detrital material to the southern part of the basin during the Eocene and Oligocene.

Ypresian (56-48 Ma) sediments in the Thrace basin show a transition from high zircon content at its base, evolving upsection into compositions rich in rutile, garnet, and Cr-spinel. This indicates initial sediment input from metamorphic/igneous rocks which evolved over time from the Ypresian to the Lutetian (48-41 Ma) to include ophiolitic and medium-grade metamorphic material (Caracciolo *et al*., 2014).

Lutetian sediments in the southern parts of the basin are primarily derived from ophiolitic terranes and reflect increasing contributions from low to medium-grade metamorphic sources, probably the Vardar zone (Fig. 1). The sedimentation in the north was influenced by the erosion of low-grade metamorphic and ophiolitic rocks (Caracciolo *et al*., 2014).

The heavy mineral composition from Priabonian (latest Eocene, 38-34 Ma) and Oligocene sediments in the southern Thrace basin indicates significant contributions from metapelitic and ultramafic sources, with garnet and Cr-spinel (Caracciolo *et al*., 2014).

Determining sediment source from apatite fission track analysis

Mesohellenic basin

The apatite FT (AFT) data is provided in Table 2, which reports the range of single-grain AFT ages from the detrital samples, as well as two AFT age populations.

There are two main apatite fission track (AFT) age populations in the clastic sediments of the southern MHT. One around 40 Ma (Eocene) and the other around 65 Ma (Paleocene) (Table 2). The younger population is consistent across all samples. This is evident in samples AV-52 and AV-107 (Table 2), which show similar AFT single-grain age distributions. Some samples date from Late Cretaceous to Early Tertiary, and those ages are also found in the Pelagonian zone. Those samples exhibit diffuse distribution rather than clusters and extend to higher ages, so the authors did not separate them into older age groups (Vamvaka *et al*., 2009).

Table 2. *AFT data for samples from Mesohellenic Trough. Source: Vamvaka et al. (2009).*

Sample Apatite data MHT	Single-grain AFT ages [Ma]	1st AFT age-population ± 1s.d. [Ma]	2nd AFT age-population ± 1s.d. [Ma]
AV-80	27.4 – 104.7	39.9 ± 8.6 30.8 ± 2.7	66.8 ± 10.1
AV-93	29.0 – 119.8	40.2 ± 4.5	64.7 ± 9.5
AV-94	21.1 – 114.8	42.7 ± 8.7	74.0 ± 15.0
AV-52	21.1 – 94.6	44.5 ± 2.4 37.9 ± 9.3	68.6 ± 9.4
AV-101	27.3 – 108.8	38.1 ± 4.3	58.7 ± 4.3
AV-107	22.2 – 96.6	44.8 ± 2.2 40.1 ± 9.8	63.2 ± 5.8

The Late Cretaceous to Paleocene ages indicate a source terrain in the Pelagonian zone or adjacent Vardar zone. The internal imbrication of the Pelagonian zone started in the the Early Cretaceous associated with the obduction of Vardar ophiolites (Most *et al*., 2001). In the Eocene parts of the leading edge of the Pelagonian was subducted with the Cycladic Blueschist Unit, which resulted in partially reset AFT ages in the Olympos area (Fig. 1). The 40 Ma AFT age population reflects exhumation and cooling of the subducted rocks (Vamvaka *et al*., 2009).

The oldest sediment sample (AV-80, Krania Fm., middle to Late Eocene, Table 2) shows a depositional age closely aligned with the Eocene AFT ages, suggesting rapid exhumation following the Eocene orogenic event. It also reveals a distinct cluster of single-grain ages around 30 Ma, indicating a thermal overprint likely due to Early Oligocene reheating related to compressional deformation (Vamvaka *et al*., 2009). This compressional event is dated to between 35 and 30 Ma and associated with thrusting of the Cycladic Blueschist Unit onto the Gavrovo-Tripolitza zone in the Olympos area (Schermer *et al*., 1990).

Samples from the northern study area indicate a lag time (the time difference between when a mineral grain cools through its closure temperature in the source area and when it is deposited in a sedimentary basin) of approximately 10 Myr, suggesting a slow exhumation rate of 0.3 to 0.2 km/Myr with no thermal overprint shown by the single-grain data (Table 2). Similar lag times around the Oligocene–Miocene boundary suggest that erosion may have dominated exhumation and sediment feeding to the adjacent MHT (Vamvaka *et al*., 2009). Younger sediments, post-20 Ma, show even slower exhumation (Vamvaka *et al*., 2009).

Summary and conclusions

The heavy mineral analysis revealed multiple source terrains and tectonic settings, which reflects the complex geologic history of the Hellenides.

Based on this data, I have constructed several figures depicting my own interpretation of sediment transport during the Paleogene. This was done in three stages for Paleocene, Eocene and Oligocene.
The general directions of sediment transport are also accompanied by cross sections showing tectonic evolution of relevant zones and formation of foreland basins.

During the Paleocene, the sedimentation was active in the Pelagonian and Pindos zones (Fig. 8).

In the Pelagonian flysch basin, two primary heavy mineral assemblages are a garnet-rich, and a stable mineral assemblage from granitoid sources. There were limited contributions from ophiolite complexes from the Vardar Zone. Some blue amphiboles are present, and their source corresponds to Cycladic Blueschist Unit (CBU). Garnet source is hard to pinpoint as it could have formed by high-pressure processes from subduction of an oceanic basin (Selçuk Ocean) or its source could be from the metamorphic material of the Rhodope zone. I believe it is more likely that majority of both stable ZTR minerals and garnet originated from the Rhodope zone.

The heavy mineral composition of the Pindos flysch basin points to an internal (northeastern) metamorphic complex source. Major minerals include staurolite-bearing mica and garnet, which can reflect amphibolite facies metamorphism in the Rhodope and Pelagonian zones, as well as chrome spinel which may have originated from the obducted ophiolites in the Vardar zone (Fig. 1). They are also accompanied by gneisses, schists and high-pressure minerals such as blue amphiboles originating from the CBU. A turbidite facies sedimentary model with an internal source is proposed.

Regional compression during this period caused uplift and folding (shown in cross section in Fig. 8) which likely caused the Eohellenic nappe (originally southern Vardar Ocean) to thrust towards S. This would explain its material containing chrome spinel reaching as far south as the Pindos zone.

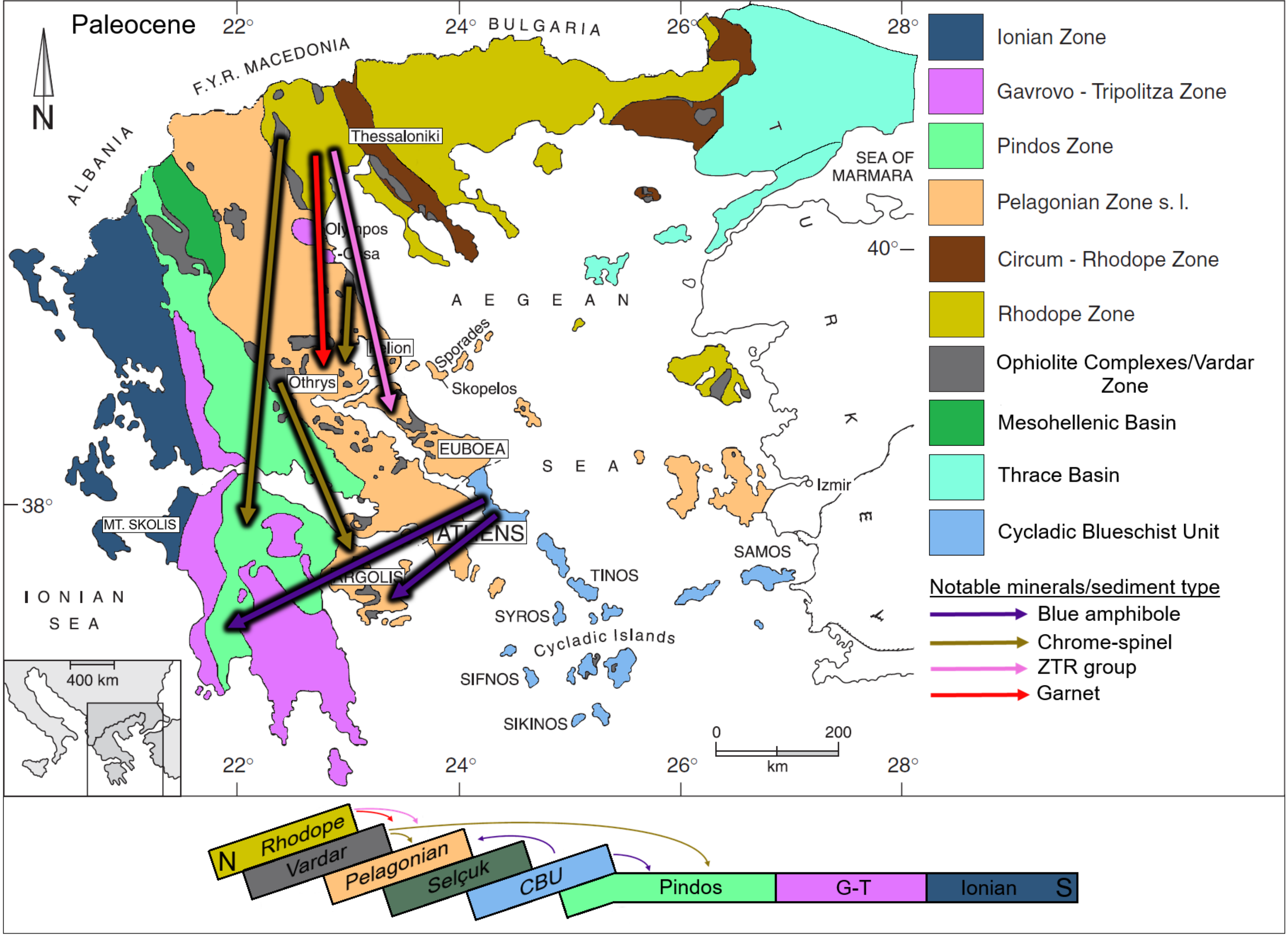

Figure 8. *Sediment transport during the Paleocene. The exact time of the Vardar Ocean closure varies from source to source, but generally it is agreed that it occurred during the Late Cretaceous or during the Paleocene, and the Cycladic Blueschist Unit (CBU) was located significantly further north than it is today before the subduction zone rollback in the Aegean Sea during the Miocene (e.g. Stampfli & Borel, 2004). As a consequence of CBU's previous position, the sediment transport would have been distinctly more N-S. G-T – Gavrovo-Tripolitza, CBU – Cycladic Blueschist Unit.*

During the Eocene, sedimentation continued in the Pelagonian and Pindos zones, and it commenced in the Gavrovo-Tripolitza and Ionian zones, as well as in the Thrace and Mesohellenic basins (Fig. 9).

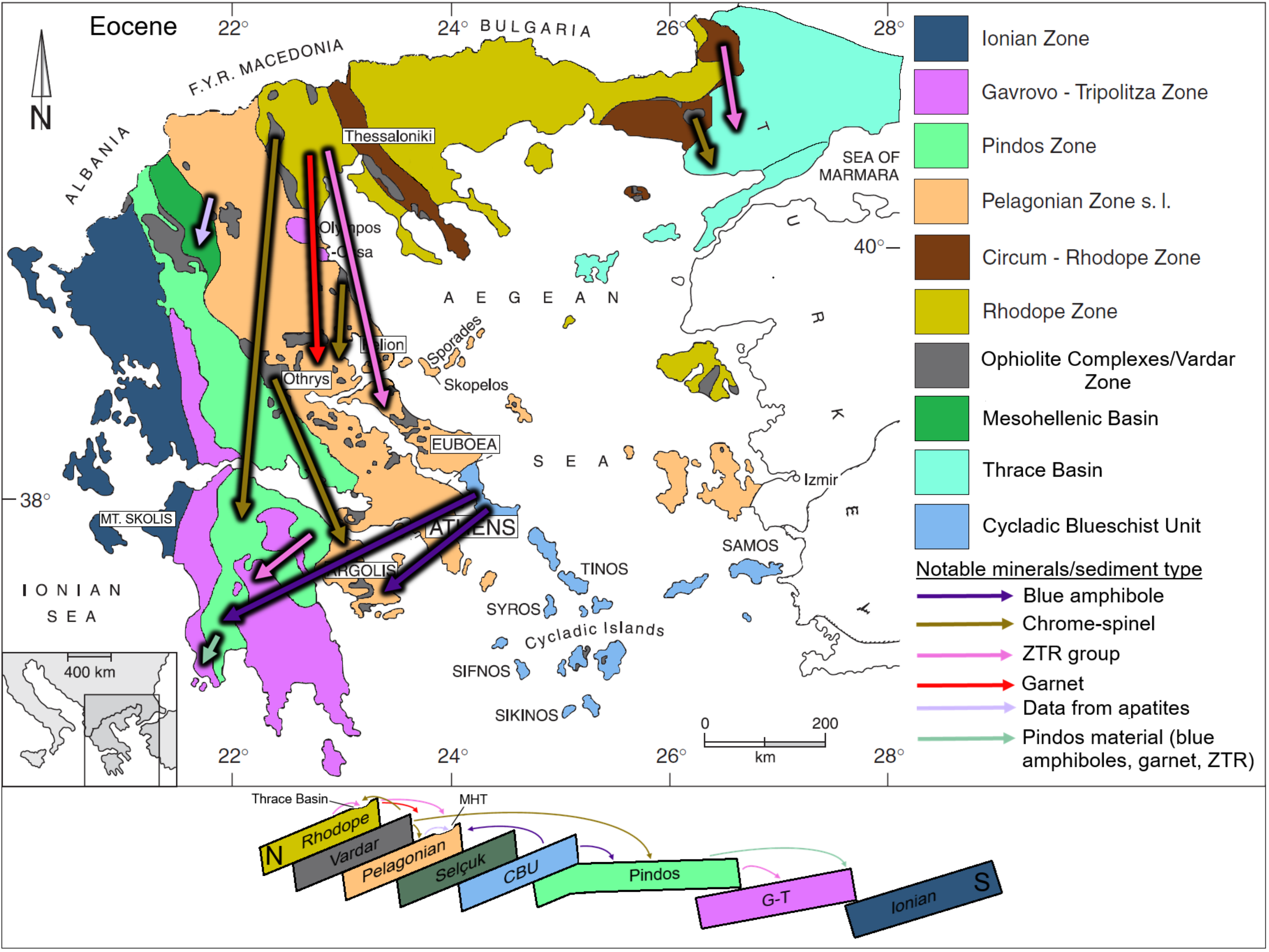

Figure 9. *Sediment transport during the Eocene. During this stage, Thrace Basin and Mesohellenic Trough (MHT) began forming as foreland basins.*

In the Gavrovo-Tripolitza and Ionian basins, heavy mineral data shows regional facies changes and provenance from mixed sources. The westward progradation of the Pindos nappe caused shifts in sediment supply and changes in the Gavrovo-Tripolitza basin axis. Biostratigraphic data shows a younging of the Gavrovo-Tripolitza southwestwards. This resulted in recycled Pindos material (ZTR group, blue amphiboles and garnet) being transferred into these basins.

The composition of sediments in the Thrace basin is characterized by high zircon content which reflects metamorphic and igneous sources, and abundant chrome spinel which points to increased ophiolitic and ultramafic input. The main source of the igneous and metamorphic sediments is the nearby Rhodope zone. The source of chrome spinel are obducted ophiolites from the Vardar zone, some of which were incorporated within the Rhodope zone during regional compression.

Due to absence of heavy mineral data for the Mesohellenic Trough, making correlations is difficult. The AFT ages suggest a source terrain in the Pelagonian microcontinent with early Cretaceous and Eocene phases of cooling. The oldest samples from the two age groups point to rapid exhumation in the Eocene, followed by Oligocene reheating. Ages of the northern samples record exhumation rates that indicate sediment transport by normal faulting. The overall AFT ages point to major sedimentation phase in the Eocene.

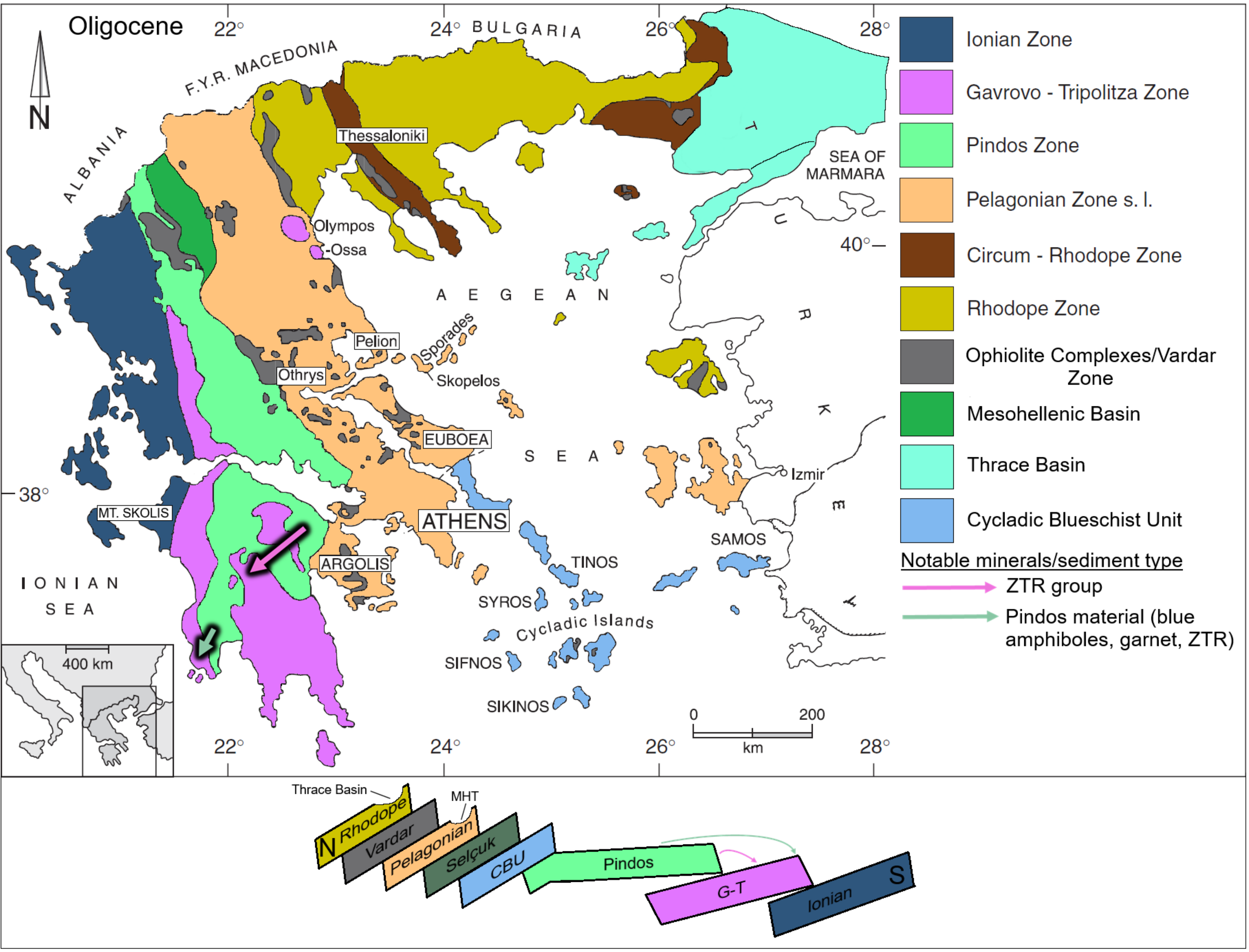

Figure 10. *Sediment transport during the Oligocene. At this stage, sedimentation is only active in the Ionian and Gavrovo Tripolitza Zones.*

Sedimentation in the Oligocene remained active in the Gavrovo-Tripolitza zone and at the southeastern margins of the adjacent Ionian zone (Fig. 10). Similarly to the Eocene, this period is characterized by continued recycling of the Pindos material as Gavrovo-Tripolitza and Ionian zones continued subducting beneath Pindos.

Flysch sedimentation ceased in other zones as regional compression caused uplift and deformation, thus ending the marine regime in the most uplifted zones. During this stage, continued filling, uplift and deformation also caused the sedimentation to cease in the Thrace basin and MHT.

This is a general interpretation of the geologic history of the Hellenides and the accompanying sedimentary processes, based on the reading material I have gathered. There are still quite a few uncertainties and disagreements among the scientific community about the history of these basins and their sediments, as I have come to realize while reading their published papers. Overall, the conclusion that can be drawn from this study is that tectonic processes such as oceanic closure, nappe stacking, basin progradation and extensional faulting have heavily influenced sediment sources and sedimentation in these basins, as well as basin development.

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