Journal of the VIRTUAL EXPLORER

A dynamic review electronic Earth Science journal publishing material from all continents

REVIEW ARTICLE

Keywords
Greece
Mesohellenic Basin
Cenozoic
Hellenides
Syntectonic sedimentation
Piggyback Basin
Forearc Basin

Table of Contents

Correspondence to:
Jacky.Ferriere@univ-lille1.fr

Citation
Ferrière, J., Chanier, F., Reynaud, J., Pavlopoulos, A., Ditbanjong, P. and Coutand, I. 2013. Evolution of the Mesohellenic Basin (Greece) : a synthesis. In: (Ed.) Emmanuel Skourtsos, The Geology of Greece - Part II, Journal of the Virtual Explorer, Electronic Edition, ISSN 1441-8142, volume 45, paper 1

Evolution of the Mesohellenic Basin (Greece) : a synthesis

Jacky Ferrière[1], Frank Chanier[1], Jean-Yves Reynaud[1], Andreas Pavlopoulos[2], Pitaksit Ditbanjong[1], Isabelle Coutand[3]

[1]UFR Sciences Terre, UMR 8217, Lille University, 59655 Villeneuve d'Ascq, Cedex, France

[2]Mineralogy-Geology, Agricultural University of Athens, Iera odos 75, 11855, Athens, Greece

[3]Department of Earth Sciences, 3015, 1355 Oxford street, Dalhousie University, Halifax, N.S., Canada

Abstract This article is an attempt to synthesize the knowledge about the Mesohellenic Basin (MHB), based upon available literature and also unpublished data. We focus on our interpretation but also mention alternative ones. The MHB is an orogenic basin of general importance, because of (i) its large size (300km along strike, and 150km in Greece) ; (ii) its location in the middle part of the Hellenic chain (between the Pindos accretionary prism and the Pelagonian upper unit) ; (iii) its large-scale piggyback setting. It has also a regional interest because of its Late Eocene-Middle Miocene age, a period of the Hellenic orogen which remains poorly understood.

The MHB fill is dominated by siliciclastic submarine deposits emplaced by gravity processes. After two main successive tectonic events, the deposits show a continued deepening during late Eocene (Krania basin) and Oligocene (MHB basin s.s.) times. Then, the Miocene series are characterized by shallower coarser-grained (early Miocene) or more calcareous (middle Miocene) sediments.

Water depth overall increases towards the north. In the Miocene, the southern MHB limit extends beyond the Meteora area, beneath the present-day Trikala plain.We consider that the piggyback setting is a key to the evolution of the MHB. Tectonics primarily controlled subsidence and the regime of sedimentation, therefore overprinting the effect of eustatic changes. The MHB infill reflects the timing and nature of understhrusted tectonic units : (i) in the Late Eocene, during the easy subduction of the thin Pindos basin crust and the development of the Pindos accretionnary prism in the external zones, subsidence in the MHB is localized in contrasted and likely small areas inherited from heterogeneities of the internal zones, namely the boundaries of the Pelagonian Indentor (PI); (ii) in the Oligo-Miocene, subsidence is generalized in the strike of the chain, due to collision of the thicker crusted Gavrovo-Tripolitsa block of the external zones.

In the Oligo-Miocene MHB, subsidence is first strong (Eptachorion marls), and then migrates to the east, progressively or stepping over structural highs as the Theopetra-Theotokos Structure (TTS) in front of the PI. In the Miocene, sediment supply is abuptly transferred from the Pindos accretionnary prism to the Pelagonian hinterland, in response to a severe uplift of the Pelagonian domain notably the Pelagonian Indentor (Meteora conglomerates).

While the collision is recorded as a major compressional phase at the Eocene-Oligocene boundary, most of the following tectono-sedimentary evolution reflects processes at the subduction plane, which remain hypothetical (tectonic erosion, underplating…). Also, the importance of strike-slip motion of some faults on the basin evolution remain matter of debate.

Our ongoing research on the MHB is focused on the chronostratigraphic assessment of sediment supply, based on thermochronology and basin modeling.

INTRODUCTION

To the heart of the Hellenic chain in Continental Greece, the Mesohellenic basin (MHB) is a major sedimentary basin of the Tethyan orogenic belt (Fig. 1A and 1B). This basin extends from Albania to northern Greece, at the boundary between the two main structural zones of the Hellenides : to the east, the internal zones, that were submitted to obduction in the Jurassic, and, to the west, the external zones which were only tectonized during the Cenozoic. Today, the MHB is a hilly landscape including sandstones, siltstones and conglomerates, at a present elevation of 700 m (in the south) to more than 1000 m (in the north), surrounded by the Pindos mountains to the W and the Pelagonian domain to the E.

The MHB is of importance because of its: (i) large size and a thick sedimentary pile (about 4,5 km of vertical thickness for c.a. 20 Ma) pointing to major orogenic processes ; ii) detailed sedimentary record, dominated by various submarine gravity siliciclastic deposits; (iii) lower Cenozoic age, which is a poorly known period of the Internal Hellenic chain ; iv) original geodynamic location and evolution (in the middle part of the orogenic belt and on the upper tectonic unit).

There are various interpretations as regarding to the processes at origin of the MHB, which would be either a retroarc foreland basin [Doutsos 1994], a strike-slip half graben ([Zelilidis et al., 2002], a large piggy-back basin [Ferriere et al., 2004], or mostly a pull-apart basin [Vamvaka et al., 2006]. In this article, we will present and discuss the successive interpretations of the MHB, focusing on its evolution (nature and changes of depositional setting) and on the related possible controls (eustacy, tectonics) in the perspective of the large scale geodynamics of this part of the Hellenides (subduction, collision).

Figure 1A
Location of the Mesohellenic basin (MHB) as the southern part of The Albano-Thessalian basin. The Tertiary basins in the eastern internal domain are represented in yellow color (geologic map superimposed on the MNT GTOPO_30, modified after Qirjaku Kaleshi [2000].

THE MHB: GEOLOGICAL SETTING

The Mesohellenic Basin: definition and overview

The Mesohellenic Basin (MHB), located in Northern Greece and Albania (Fig.1A and 1B), was formerly called "Albano-thessalian" by Bourcart [1925], before being named Mesohellenic Basin (MHB) by Brunn [1956] and Aubouin [1959]. It is called "Mesohellenic" as it develops in the middle part of the Hellenides. Compared to most other intermontane basins, the MHB is remarkable by its large dimensions (more than 300 km long with its Albanian part, half in Greece, 30 km wide and with a thick pile of sediments 4,5 km of vertical thickness). It has been said to be "molassic" as it is filled with

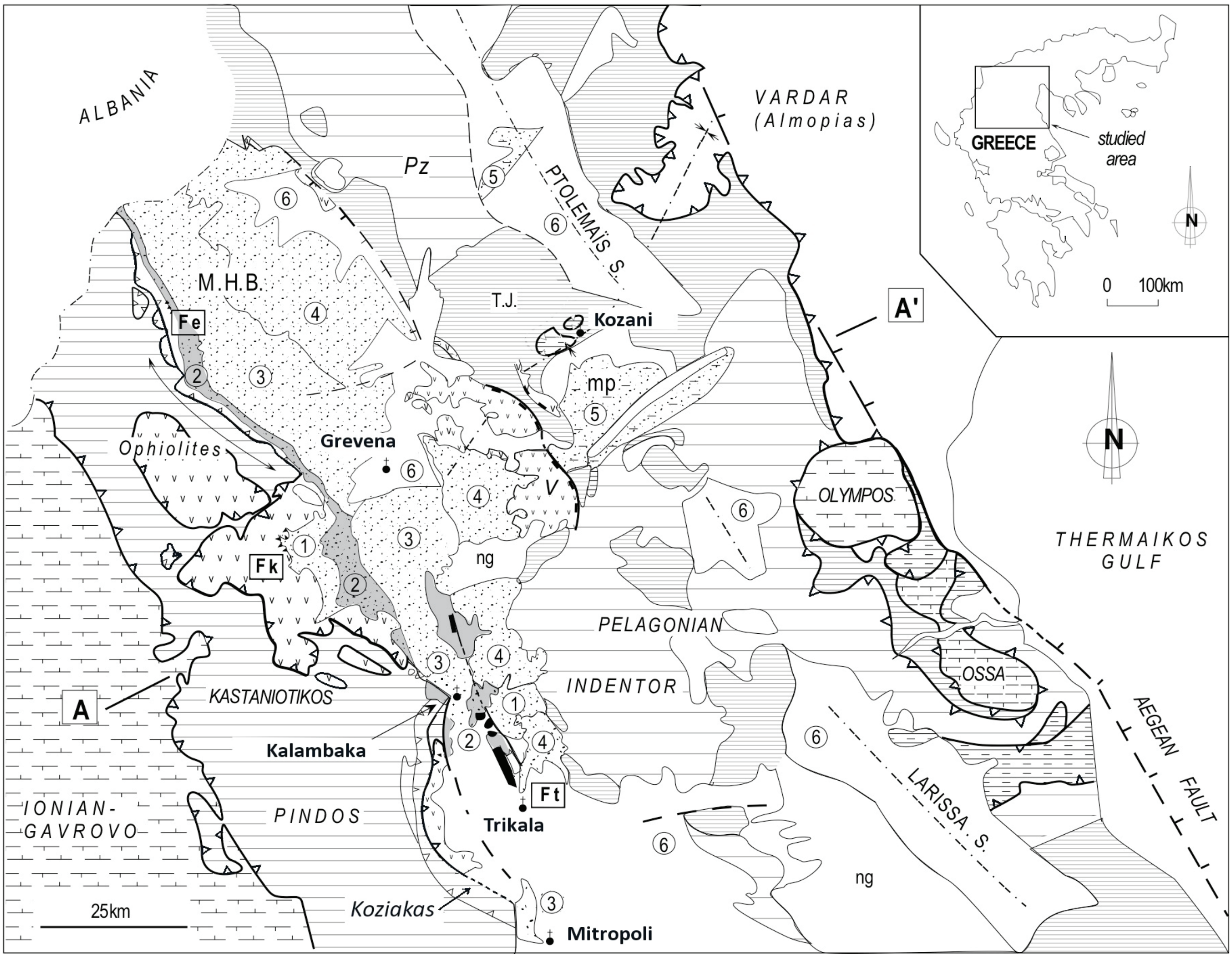


Figure 1B

Simplified geological map of the Mesohellenic Basin (MHB) in northern continental Greece (modified after Ferriere et al., 2004). 1 to 4 : main Formations of the MHB, 1: Krania and Rizoma (Late Eocene), 2: Eptachorion (Latest Eocene ?- Oligocene p.p), 3: Taliaros-Pentalofon (Late Oligocene-Early Miocene), 4: Tsotyli-Ondrias-Orlias (Early to Middle Miocene) ; 5: Ptolemais basin (Late Miocene-Pliocene, mp), 6: recent deposits. Abr. Pz: Paleozoic, TJ: Triassic and Jurassic, ng : Neogen., S: Synclines ; Fe, Fk, and Ft: faulted-flexures of Eptachorion (Fe) Krania (Fk), and Theopetra-Theotokos (Ft). AA': cross-section (Fig. 2). Bold lines: major tectonic contacts,with rectangular boxes: late Jurassic thrusts, with white triangles: main Tertiary thrusts. Lines with black triangles: tertiary back-thrusts or main reverse series. Dashed lines: normal faults.

detrital sediments (marls, shales, turbidites, conglomerates) unconformably overlying the deformed Mesozoic-Paleocene basement and some early Tertiary thrusts. However, its sedimentary fill is mostly syntectonic.

The basin mostly developed east of the main Tertiary tectonic boundary between external and internal zones of the Hellenides, known as the "Internal Zones Thrust" part of a very large thrust system located beneath the MHB (Fig. 1B and Fig. 2).

In this area, the internal zones are made up of the Pelagonian continental crust (Triassic to Jurassic metamorphic limestones and Paleozoic gneisses) partly overlapped by upper Jurassic ophiolites thrust again towards the west onto the Pindos units during the Tertiary events [Brunn, 1956 ; Aubouin, 1959] (Pl. I-A). Most of the MHB fill rests above these thick ophiolitic units obducted during the Jurassic.

The external zones consist of Pindos series, mainly of Pindos flysch nappes, just west of the Pelagonian zone below which the thin, continental or oceanic (?) Pindos crust was underthrusted to the east.

This basin is Cenozoic in age: it was infilled between the Upper Lutetian (ca 45 Ma) and the middle Miocene (ca 15 Ma), spanning over 25-30 Ma (Fig. 3). Deposition follows a major deformation episode of the internal zones in the lower-mid Eocene. Upper Lutetian-upper Eocene marine deposits unconformably rest above the basement of internal zones, while Oligocene sediments overlap unconformably both the external and internal zones and seal the "Internal zone thrust".

The MHB forms an elongated asymmetrical syncline, with steeper strata on its western flank (Fig. 4) . Another asymmetry raises as Miocene strata (Tsotyli Formation) are absent in the west and rest onto the basement in the east. (Figs. 3 and 4). Seismic profiles (Fig. 5) show a pinch out of deposits at depth [Kontopoulos et al., 1999; Zelilidis et al., 2002]. These data show that deposition is controlled by an overall eastward migration of depocentres and thus of subsidence (Figs. 3 and 4).

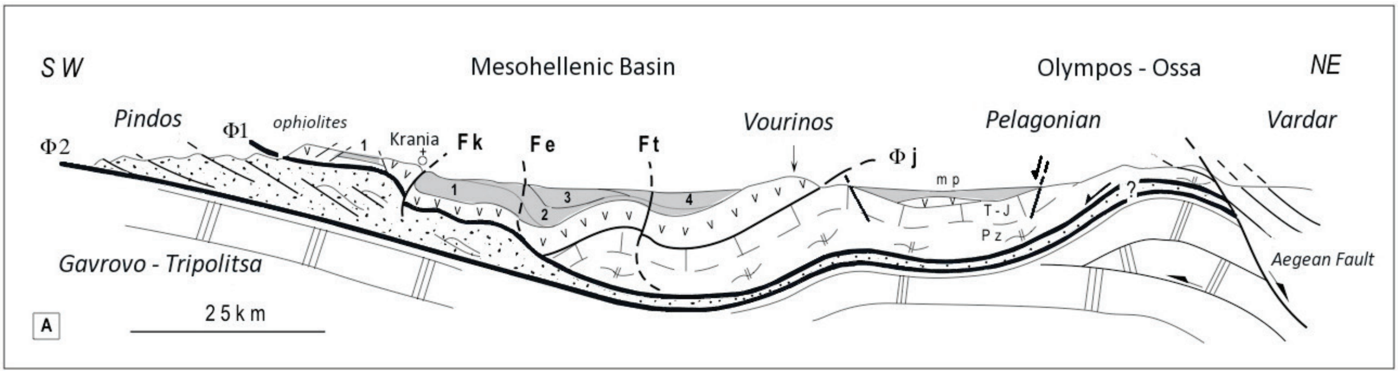


Figure 2
Cross-section showing the MHB as a piggyback basin above the main Tertiary thrusts responsible for the Olympos window (modified after Ferriere et al. [1998]). See Fig. 1 for location. 1 to 4: MHB Formations, same captions as in Fig. 1. Fe, Fk, and Ft: faulted-flexures of Eptachorion (Fe), Krania (Fk) and Theopetra-Theotokos (Ft). Φ1 and Φ2: main Tertiary thrusts j: Jurassic obduction. Vertical scale: maximum thickness of the MHB sediments on the cross-section: 4 km.

The width of the MHB decreases southward along strike. This is related to the basin squeezing to the south against the "Pelagonian Indentor", which forms a spur of the basement of the internal zones to the SE of the basin (Fig.1A and Fig.3) [Ferriere et al., 2004].

The southern part of the MHB is separated alongstrike by a horst, or faulted anticline called Theopetra-Theotokos Structure ("TTS": Fig.3 and Fig.4), which splits the basin into two parallel parts, one to the west which is occupied by the Pentalofon Formation (lowermost Miocene), and the other, to the east, by the Tsotyli-Ondria Formation (lower Miocene p.p.) (Fig.1B and Fig.3).

To the NW, the MHB extends into Albania, where it is called "Albano-Thessalian basin". It might step over the major, transverse Scutari-Pec feature, which limits the Albanic and Dinaric chains (Fig.1A). In this article we only present the available data regarding to the greek part of the basin, and we mostly focus to its southern half.

Evolution of ideas about the MHB

The older attempt concerning the litho-chronostratigraphic framework of the MHB was published by Brunn [1956] and little changed by many workers since. The first detailed studies of the MHB were focused on mapping these lithological Formations [Brunn, 1956; 1969; Savoyat et al., 1969; 1971a; 1971b; 1972a; 1972b]. New maps concerning the MHB were published later [Mavridis et al., 1979; 1993; Koumantakis et al.,, 1980; Vidakis et al., 1998].

Other publications provided some refinements: i) biostratigraphic refinements based on Foraminifera [Soliman and Zygojiannis, 1980] and nanofossils [Zygojiannis and Muller, 1982]. New data on nanofossils were published by Kontopoulos et al., [1999] and Ferriere et al., [2004]; ii) source rock studies from heavy minerals [Zygojiannis and Sidiropoulos, 1981] or olistoliths [Papanikolaou et al., 1988; Wilson 1993]; iii) dynamics of depositional systems [Faugères 1977a; 1977b; Desprairies 1979 for the different Formations; Ori and Roveri, 1987 for the Meteora conglomerates].

New modern studies concerning the MHB were initiated since the ninety's, essentially applied to i) sedimentological analyses and large-scale industrial seismic data [Zelilidis et al.,1996; 1997; Zelilidis and Kontopoulos, 1996; Kontopoulos et al., 1999; Zelilidis et al., 2002] and ii) tectonic and geodynamic data used to assess the first basin models [Doutsos et al., 1994; Ferrière et al., 1998; 2004; 2011; Vamvaka et al., 2006].

Following these various studies, the main stages of evolution of the MHB were established but divergent interpretations still exist about the geodynamic setting and also the interplay of eustacy regarding to the basin stratigraphy. Here below we first briefly present the lithological formations and tectonic deformations of the MHB, and then propose an attempt of reconstructing the basin evolution and discussing the main related mechanisms.

THE SEDIMENTARY FILL OF THE MHB

Overview

Knowledge limits

The large thickness of the MHB strata (Fig.3), the repetitive stacking of gravity deposits and uncertainties about their lateral and vertical correlation across the basin bring

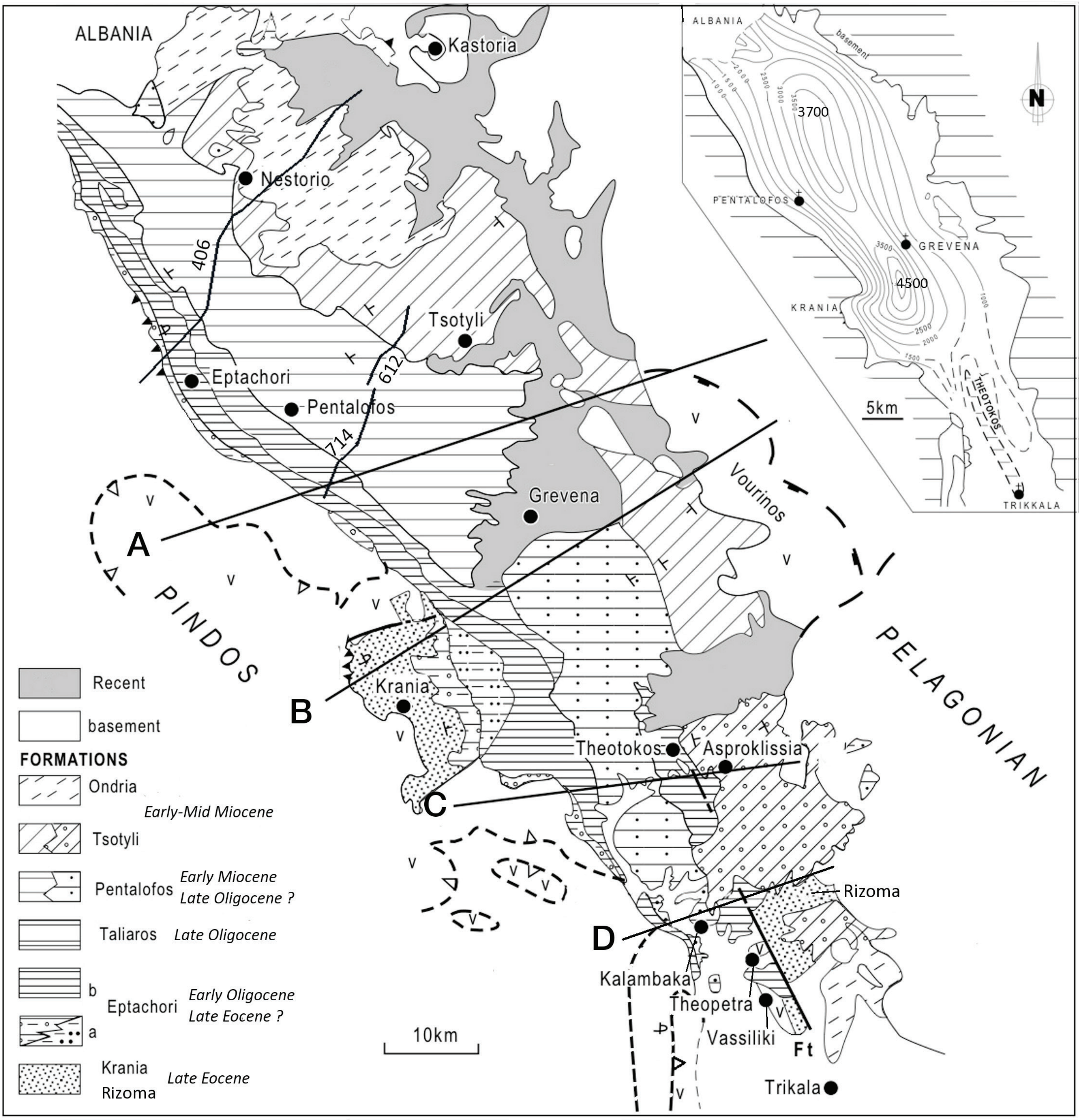


Figure 3

Map of the lithological formations of the MHB. Modified after the Geological Maps of Greece at 1:500,000 [Bornovas and Rondogianni-Tsiambaou, 1983] and at 1:50,000 [cf. References], the synthetic map of Doutsos et al. [1994] and from our field study in the southern half of the MHB. Cross sections A to D see Fig. 4. Up right: depth contours of the basement below the basin from seismic data from Kontopoulos et al. [1999] completed in the south from field data (maximum thickness : 4500m near Grevena).

about uncertainties on the definition of large-scale lithological units of stratigraphic significance (Fig. 6). Moreover, the chronostratigraphy of MHB Formations is still not very precise, mostly because of the scarcity of fossils or due to their reworking in gravity dominated facies.

The only available ages are from marls (pelagic foraminifera, nannofossils) or from a few carbonate shelf intervals (benthic foraminifera, invertebrates). Moreover, published ages are significantly divergent, even for the same faunal associations as nanofossils [Zygojiannis and Müller, 1982; Kontopoulos et al., 1999; Zelilidis et al., 2002;

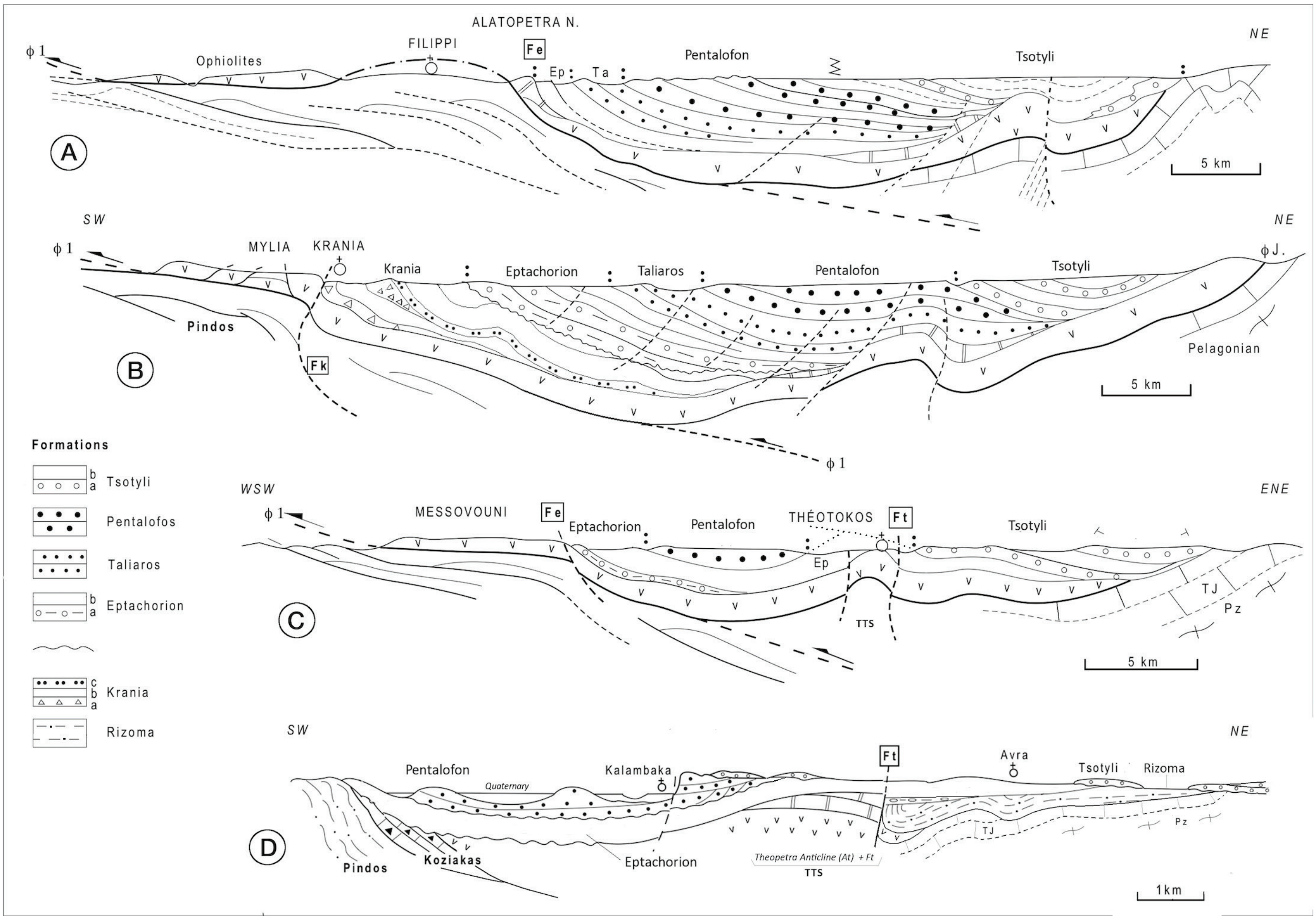


Figure 4
Cross-sections of the southern MHB with no significant vertical exaggeration (modified after Ferriere et al. [2004], compiled from our field work, Geological Maps of Greece at 1: 50 000 [cf. References] and seismic profiles published by Kontopoulos et al. [1999] and Zelilidis et al, [2002]).

See Fig. 3 for location. Abbreviations: TTS: Theopetra-Theotokos Structure; Fk, Fe, and Ft, faulted flexures of Krania, Eptachorion and Theopetra, respectively.

Lithologies: see text. Krania Formation: a: conglomerates and olistoliths of the lower Krania sequence; b) turbiditic beds; c) thick unconformable turbiditic base of the upper Krania sequence. Eptachorion Fm: a) basal conglomeratic beds, (b) turbidites and marls. Tsotyli Fm: a) mainly conglomerates and sandstones; b) mainly sandstones and siltstones.

Ferriere et al., 2004] (Fig. 7).

The thickest, deepest and more extended Oligo-Miocene marine formations crop out in the northern part of the MHB basin. Middle-upper Eocene deposits are exposed in restricted area of the center of the basin (near Krania village; Fig. 1B and Fig. 3). These formations have their chronostratigraphical equivalents to the south of the basin, but there the facies point to shallower water depths (ex. Pentalofon Formation in the Meteora area) and therefore hiatuses and lacunas are frequent..

The main lithological formations and their boundaries

The Oligocene to Miocene siliciclastic deposits were first described as six main lithostratigraphic units by Brunn [1956, 1960] from the northern part of the MHB with the addition of a late Eocene Formation (Krania Fm). These Formations are (Fig.3, 4 and 6):

- i) Eptachorion Formation (about 1000m, mainly Oligocene) dominated by silty marls in the upper part of the Formation with decimetric thick very fine sandstone beds often resting on thick conglomerates in the lower part.

- ii) Taliaros (or Tsarnos) and (iii) Pentalofos Formations (2500m, latest Oligocene and early Miocene): sandstone beds coarsening upwards to conglomeratic beds; mainly conglomeratic beds in the south [Ori and Roveri, 1987; Ferriere et al., 2011].

- iv) Tsotyli Formation (600m, early-mid ? Miocene): marls interbedded with sandstones in the northern MHB; gneissic pebbles rich conglomeratic beds in the south [Savoyat et al., 1971a; 1971b; 1972a; 1972b; Zygojiannis and Muller, 1982; Ferriere et al., 2004].

- v) Ondria and (vi) Orlias Formations (350m or more, early-mid Miocene): sandstones and marls with fossiliferous limestone beds.

Most of these formations have been adopted by later workers, although minor modifications have been made. These consist namely in additional lithologic members: i) within Pentalofon Fm: Tsarnos and Kalloni members were distinguished, pass-

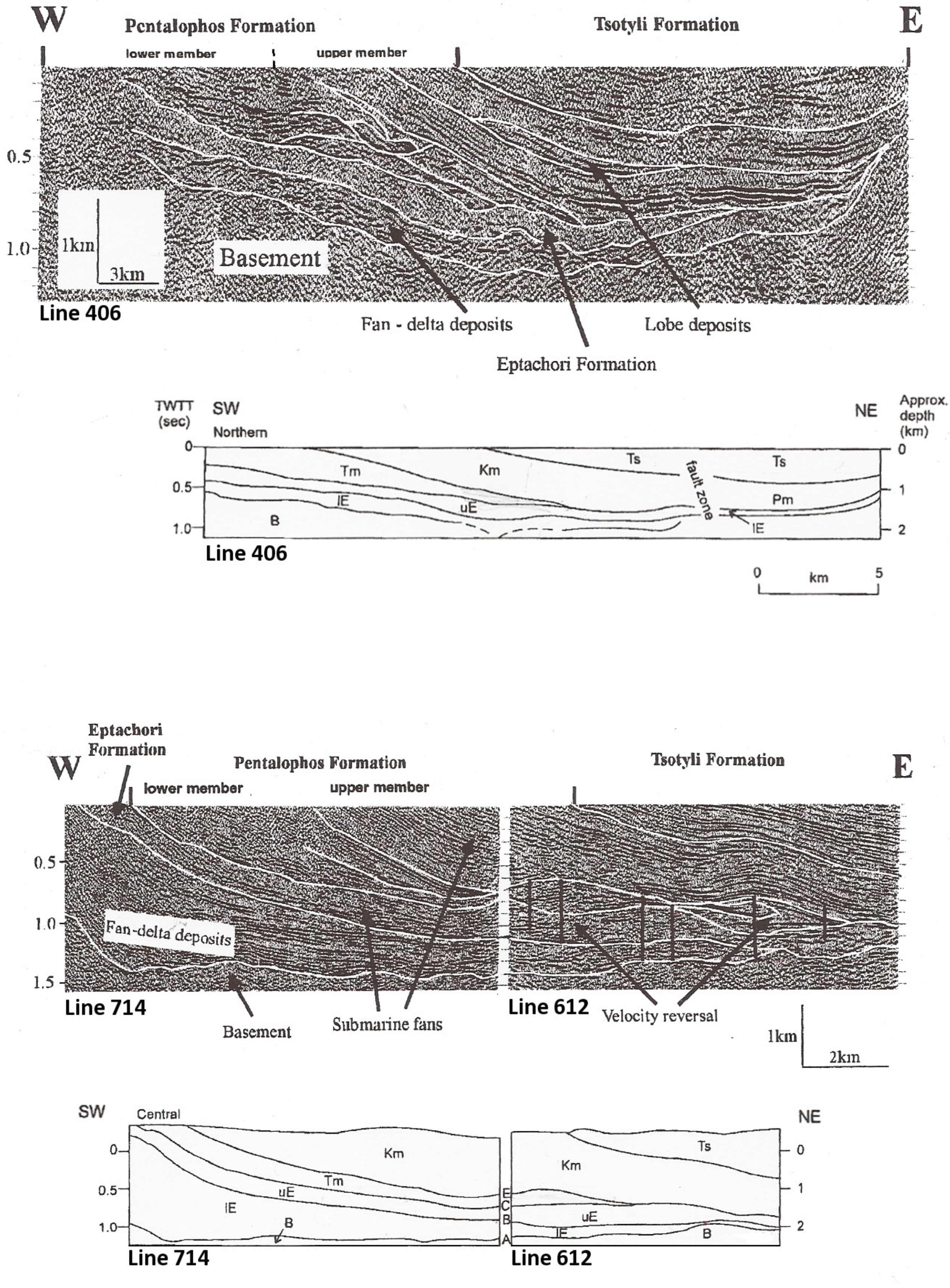
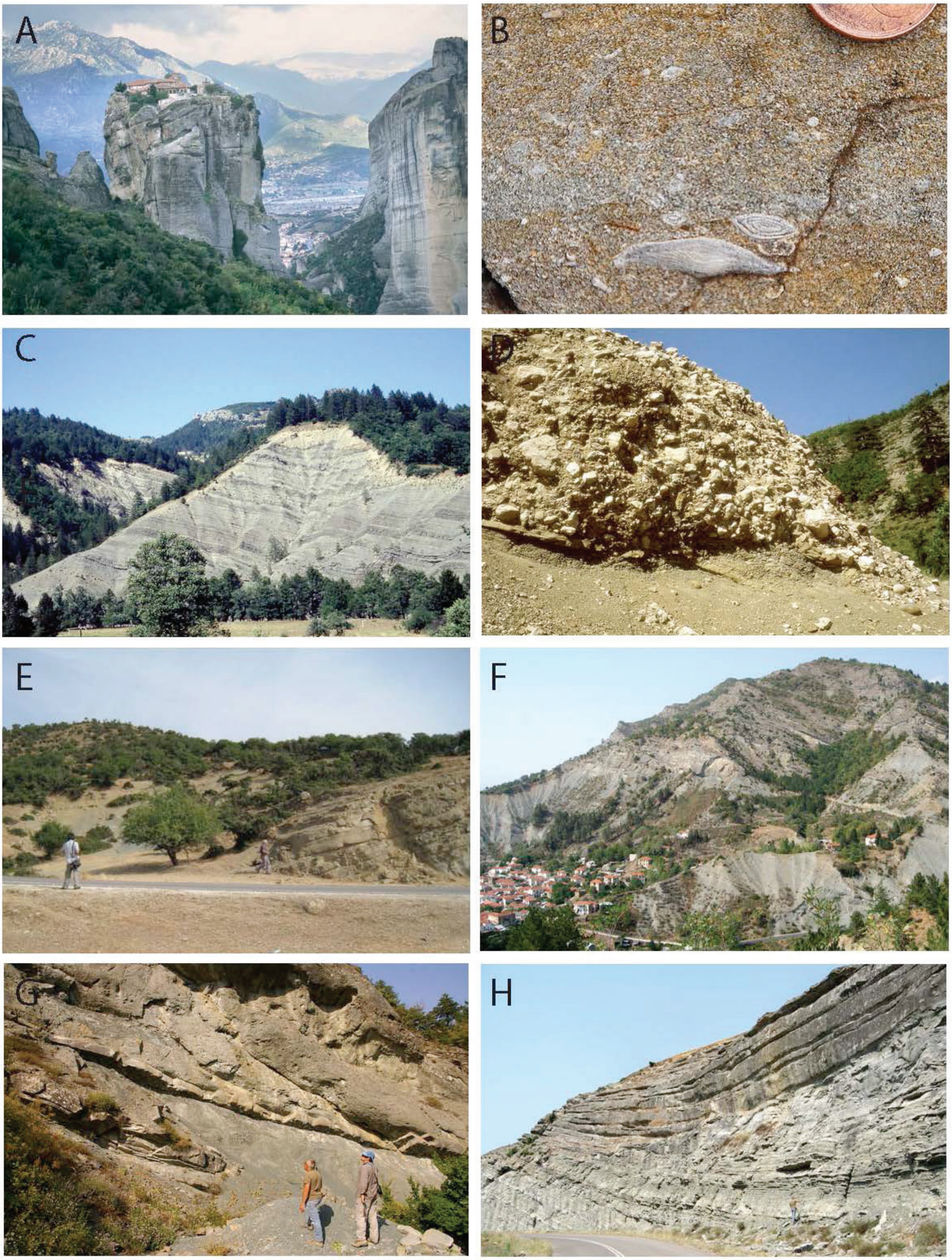


Figure 5

Seismic profiles in the northern MHB and their interpretation. After Kontopoulos et al. [1999], and Zelilidis et al. [2002]. Lines 406, 714 and 612: location see Fig. 3. Abbr: B: Basement; E: Eptachorion Fm (l: lower, u: upper); Km: Kalloni member; Tm: Tsarnos Fm; Ts: Tsotyli Fm.


A
B
C
D
E
F
G
H

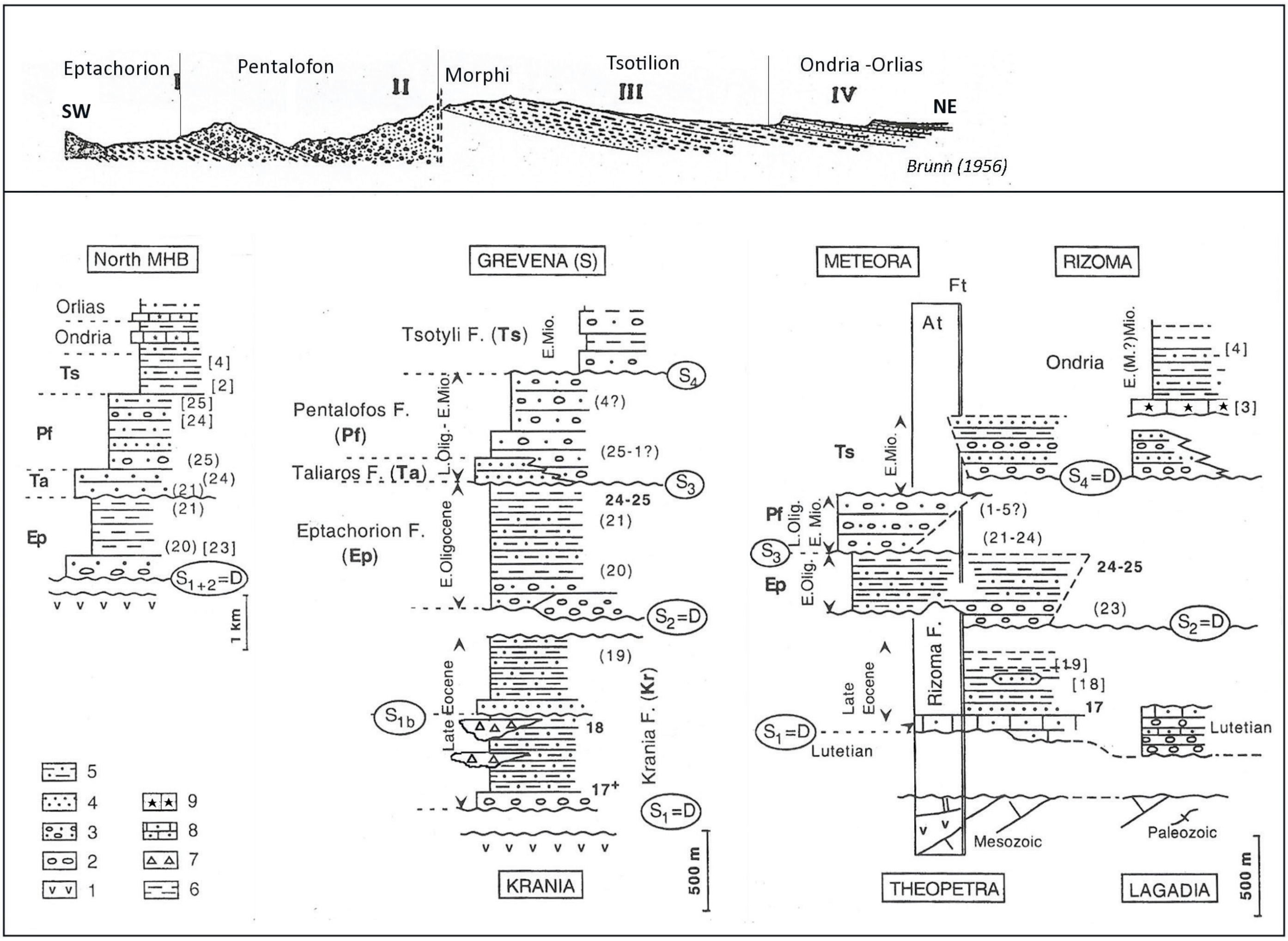


Figure 6

Schematic stratigraphic logs in the central (Grevena) and southern (Meteora-Rizoma) MHB, compared to the northern one (North MHB) (modified after Ferriere et al., 2004). 1: Pelagonian basement: ophiolites (v) with upper Cretaceous (limestones) on Mesozoic marbles and Paleozoic gneisses; 2: Basal conglomerates (unconformity); 3: conglomerates and sandstones; 4: Sandstones (mainly turbiditic); 5: Sandstones and shales (mainly turbiditic); 6: Shales and marls (partly hemipelagic); 7: major olistoliths (Krania); 8: Eocene detrital limestones with Nummulites; 9: Miocene Echinid-rich limestones.

Nannofossils Biozones after Zygojiannis and Müller [1982], Kontopoulos et al. [1994] and Zelilidis et al. [2002]; thick numbers: our own results. Abbr: D: major angular unconformities, S: other significant Surfaces (main lithological changes), Pz: Paleozoic, TJ: Triassic-Jurassic, UK: upper Cretaceous, Eo: Eocene, Olig: Oligocene, Mio:Miocene, E: Early, M: Middle, L: Late.

Upper Cross-section: Stratotypes of the MHB Formations as defined by Brunn [1956].

Plate I (previous page)

A) General view of the basin, looking westward. From the front to the background: lower Miocene (Meteora conglomerates), Oligocene sandstones (west of the Pinios river valley) resting unconformably over Jurassic limestones (Koziakas mountains) and, to horizon line, the Pindos flysch. B) Rizoma Formation: shelf limestones with large benthic foraminifera of Middle to Upper Eocene. Near Vassiliki, these limestones locally rest above the oldest conglomeratic deposits of the MHB. C) Lower Krania series (Late Eocene): ophiolite-rich, deep-sea sandy turbidites. Cretaceous limestone olistolites, up to tens of meters large, are present in the background. D) Krania series (Late Eocene): submarine channel fill with Cretacous limestone blocks incised within the upper Eocene turbidites of Lower Krania series. The sandstone bar to the right corresponds to the base of the Upper Krania series. E) Eptachorion Formation (Oligocene) : typical facies of Oligocene in vicinity (SE) of the Meteora : well-bedded sandstone (to the right) overlain by uppermost Eptachorion marls (top left: Tsotyli Fm). F) Transition from the thick marl succession of the upper part of Eptachorion Fm to the turbiditic sandstones of Tsarnos Fm close to Eptachorion village. G) Detail of the Eptachorion-Tsarnos transition of picture F. The marls are overlain by slumped turbidites and blocky slurry flows. H) Pentalofon Fm (Early Miocene): example of silty-sandy turbidites to the north of the MHB between Grevena and Pentalofon.

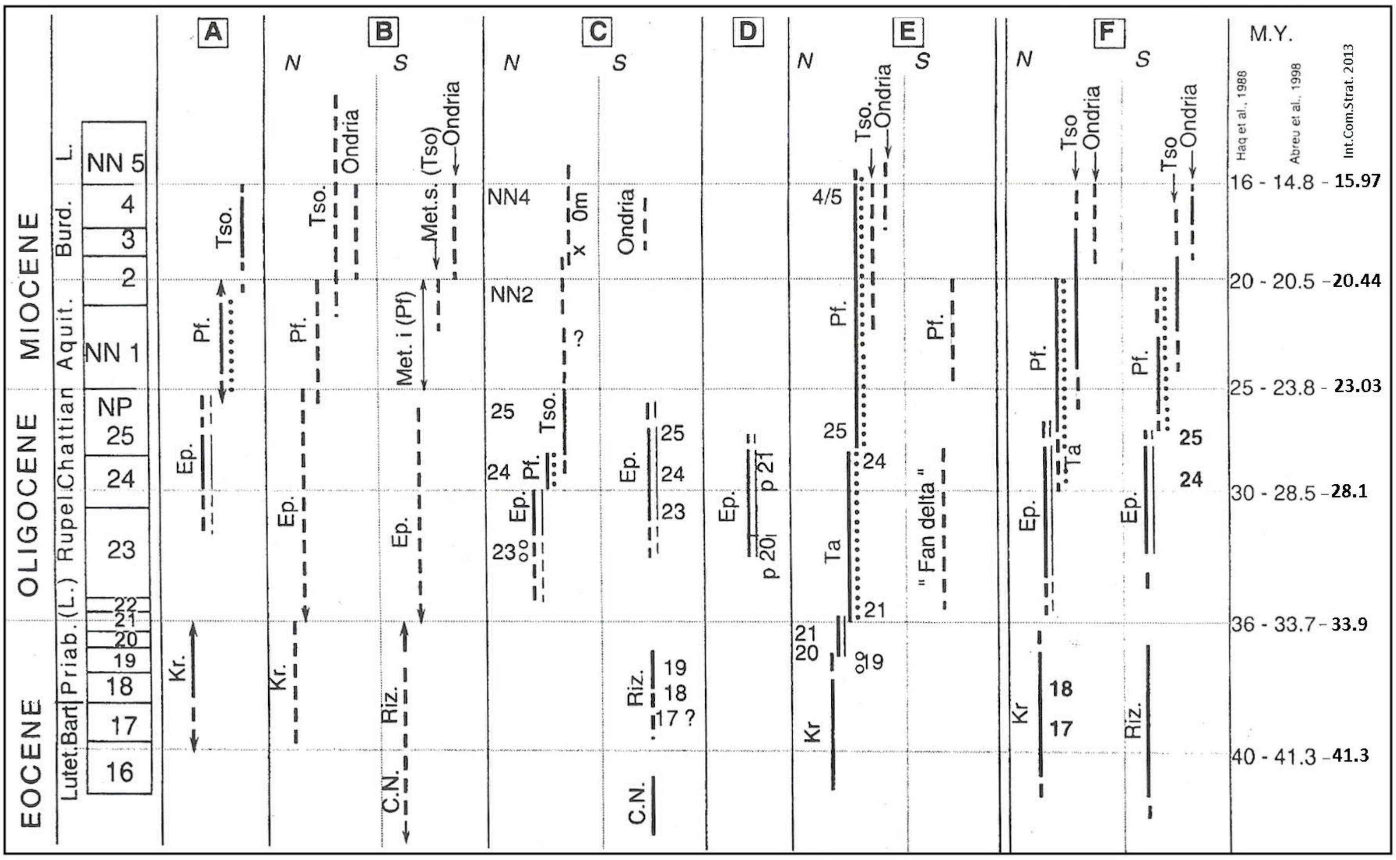


Figure 7

Compared ages of MHB deposits from Foraminifera (D) or nannoflora biozones (modified after Ferriere et al. 2004). A: Brunn [1956], B: Geological Maps of MHB areas, IGME, Greece, at 1:50,000 scale (cf References) and Bizon et al. 1968, C: Zygojiannis and Müller [1982], D: Barbieri [1992], E: Doutsos et al. [1994], Zelilidis et al. [1997, 2002] and Kontopoulos et al. [1999], F: this study.

Abbr. CN: nummulitic limestones, Kr.: Krania, Riz.: Rizoma, Ep.: Eptachorion, Ta.: Taliaros, Pf.: Pentalofon s.l. (Pf and Ta) or s.s. (Pf), Tso.: Tsotyli, l. and u. Met.: lower (Pf) and upper (Ts) Meteora. Single and double lines : different formations ; broken thick lines : uncertainties; nannoflora biozones 16 to 25 = NP 16 to NP 25, 17+ minimum age (biozone 17 or younger), 1 to 5 = NN1 to NN5 ; Foraminifera biozones: P20 and P21 (D). Ages of the stratigraphic stages from Haq et al [1987], Abreu et al [1998] and the International Commission on Stratigraphy [2013].

ing southward to Kalabaka and Anthrakia members [Zelilidis et al., 2002] and ii) in the upper part of the MHB fill, a few more formations have been added within the Ondria Fm: X Fm, Omorphoklissia Fm and Zevgostasi Fm (see geologic maps to 1:50 000, ie: Nestorion and Argos Orestikon sheets, Savoyat, 1971a, b).

The formation boundaries are either major angular unconformities or/and abrupt changes in lithology. Angular unconformity-bounded formations record major tectonic deformations. Amongst those, one may cite, in chronological order, the lower boundaries of: Krania and Rizoma Fms (Upper Eocene) ; Eptachorion Fm (Oligocene) ; Tsotyli Fm (Miocene). In the southern part of Fms the MHB, the lower Tstotyli bounding unconformity reflects a major change in depocenter location, shifting to the Rizoma area where Miocene deposits rest above Upper Eocene, Mesozoic or Paleozoic strata (Fig. 3 and 4).

Some formation boundaries are associated with major lithologic changes without basal angular unconformity (ex. Eptachorion-Taliaros-Pentalofon Fms and Ondria-Orlias Fms). In this case, the control could be an eustatic change only. This is argued by Zelilidis et al., [2002] to explain, for instance, the Eptachorion-Pentalofon Fm boundary. However, Ferrière et al., [2011] shows that even in that case the tectonic component has to be taken into account.

Facies map of the MHB

After general maps of lithological formations [Brunn, 1956; Desprairies, 1977] and detailed geological maps to 1:50 000 (maps of IGME, from 1969 to1998), lithofacies maps have been published by Kontopoulos et al., [1999] and Zelilidis et al., [2002]. These authors distinguish the marls, conglomerates, and various facies of sandstones which they relate to submarine fans and associated turbidites (Fig. 8).

Zelilidis et al. ([2002] related these facies to subsurface depositional seismic facies and geometries. However, the likely complexity of the submarine basin topography and the lack of preservation of a complete depositional profile at the outcrop (especially for the Oligocene) make these interpretations somewhat speculative.

Besides, Ferrière et al., [2004; 2011] privileged the detailed facies mapping of selected areas in the southern half of the MHB, where unconformities are better expressed, thus providing specific arguments of the tectonic control of the depositional

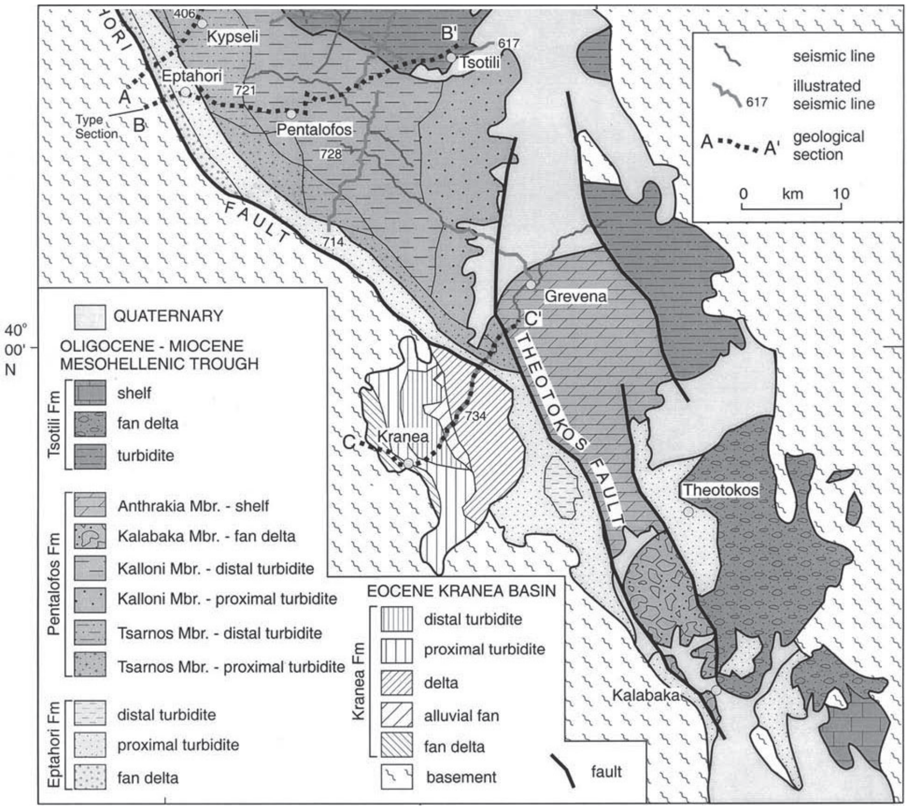


Figure 8
Map of the central part of the MHB showing different sedimentary facies, especially distal and proximal turbidites, as described by Zelidis et al. [2002].

systems (Fig.3) [Ferriere et al., 2004, 2011]. The formation descriptions presented herein mostly rest above these observations (including the Krania area), but we also have added information compiled from other available sources as regarding to the area more to the north.

The upper Lutetian-upper Eocene Formations

Rizoma Fm

This formation, which rests unconformably above the Pelagonian basement, crops out to the SE of the MHB only (Fig.3 and Fig.6). It is dominated by shales with massive sandstone interbeds, locally overlying conglomerates and limestones hosting a benthic macrofossil fauna (Pl. I-B) [Savoyat et al., 1969; Zygojiannis and Muller, 1982]. The shales and sandstones have been interpreted as a fluvial-dominated shelf delta system [Ferriere et al., 1998; 2004]. This formation is covered by Oligocene basal conglomerates or Miocene conglomeratic Tsotyli beds (Fig.3).

The Rizoma Fm comprises three lithological units, from base to top (Fig.6):

- i) Well-rounded basal conglomerates. These conglomerates are exposed in two areas, to the south of Vassiliki and more to the east near Lagadia (Fig.3 and Fig.6). They deliver numerous clasts of Cretaceous limestone, as well as radiolarites, ophiolites and triassic-jurassic marbles, all derived from the internal zones which crop out in vicinity to the east.

- ii) Nummulitic-rich limestones, pointing to a carbonate shelf setting (Pl.I-B). They also contain algae and Echinids, and are attributed to the Upper Lutetian [Savoyat et al., 1969; 1972a; Ardaens, 1978; Ferriere 1982].

- iii) The “Rizoma marls”, a thick shale succession (more than 200m) made up of distal turbiditic sequences, locally with metric sandstone beds interpreted as fluvial dominated deltaic mouth bar systems (wood fragments, floating mud pebbles, water escape features, current ripples, Skolithos traces and various burrows). This deposit is supplied by the pre-ophiolitic basement comprising schists and gneisses. Globigerinids [Bizon et al., 1968; Savoyat et al., 1969; 1972a] and calcareous nannofossils [Zygojiannis and Muller, 1982; Ferriere et al., 2004] from the Rizoma calcareous marls yielded an Upper Eocene age (biozones 17 to 19) (Fig.6 and Fig.7).

Krania Fm

The Krania Formation is well developed and preserved only inside a syn-sedimentary syncline. This formation is exposed in the western part of the MHB, to the SW of Grevena (Fig.3). It forms a flysch-like unit 1500 m thick, bounded at the base by ophiolitic conglomerates and overlying mostly the ophiolitic basement of the basin. It is bounded to the top by the major intrabasinal unconformity of the MHB. A minor unconformity has been described inside the formation [Koumantakis and Matarangas, 1980; Wilson, 1993; Ferriere et al., 2004]. The Krania Fm provided a foraminifera and nannofossil assemblage of the upper Lutetian-Upper Eocene (Fig.6 and Fig.7).

The Krania Formation exhibits a set of two sequences of deposits (Fig.6):

Lower Krania sequence

West of Krania, the deposits rest onto roughly bedded, polygenic clast-supported conglomerate beds, interpreted as alluvial fan deposits, onlapping a sole of ophiolitic epiclastites. Above these basal beds in the Krania-Microlivadon areas, the lower sequence is composed of fine-grained fining upwards and homogeneous sandstone beds interpreted as deep water and ophiolitic rich turbidites (Pl. I-C).

On the northern side of the syncline, this basal succession passes laterally to highly bioturbated, fine grained marly sandstones with thin channel bodies of sandstones showing mud pebbles at the base, interpreted as bay-fill deposition and to roughly bedded conglomerates interpreted as alluvial fans (Trikomo-Parorio area, cf Fig.14). Nannofossils in these facies yielded an Upper Eocene age (Biozones 17 and 18, Fig.6 and Fig.7) [Ferriere et al., 2004].

These deposits may be highly disrupted by slumps and locally by a large scale olistostrome (Trikomo-Monachiti area, Pl. I-D) especially at the end of deposition of the lower sequence. This olistostrome, mainly composed of Cretaceous limestones olistolites collapsed from the northern basin margin, feeds channels and gullies down to the southern basin floor (Krania and Microlivadon area) [Wilson, 1993; Papanikolaou et al., 1988; Ferriere et al., 2004]. It is coeval to a paroxysm of slumping in the deep turbiditic basin. Middle to Upper Eocene Nummulitids are found in the matrix of the olistostromes ([Ferriere et al., 1998; 2004].

Near Mylia (6 km SW of Krania, Fig.4B), flysch-like deposits deliver clasts and blocks of lavas or Lutetian and upper Eocene limestones. For some authors, these deposits belong to the top of the Pindos flysch units, which would be exposed as a tectonic window beneath the ophiolitic basement of the MHB [Desprairies, 1979]. We rather suggest they form the lowermost deposits of the MHB, preserved in small grabens at the top of the ophiolites (Fig.4B) [Ferriere et al., 2004]. This is of importance for determining the age of the main stage of deformation in the Pindos zone.

Upper Krania sequence (shales and sandy turbidites)

This unit unconformably rests above the lower sequence. The unconformity is well exposed on the northern side of the Krania syncline (Trikomo-Monachiti area), while it is less prominent southward, in the center of the syncline (cf Fig. 14). The upper Krania sequence is made up of the same turbiditic sandstones as the lower one, but exhibits at the base a sharp-based hectometric succession of thicker beds with locally abundant burrows (Skolithos), plant fragments, water escape structures and intraclastic breccias, interpreted as part of a basin floor fan (south of Monachiti) (Pl. I-D).

The Krania basin deposits are topped by the Oligocene major unconformity locally made up of reddish conglomerates and paleosoils truncating the turbidites in the middle of this basin (Fig. 3 and cf Fig.14).

Interpretation: transgression above a post-tectonic, irregular paleotopography

The late-Eocene formations of the MHB (Rizoma and Krania) record the transgression which takes place after the main Cenozoic tectonic phase of the internal zones of the Hellenides. The paleotopography was still scarped and locally active, which explains the development of olistolites (submarine slope mass wasting) and conglomerates. The first transgressive deposits are rapidly buried beneath sandy, deep-water, flysch-like turbidites to the west (Krania), while a thin calcareous shelf persists to the east (Rizoma), finally drowned by a siliciclastic shelf delta. This reflects the irregular topography on which transgression takes place, perhaps the stronger subsidence to the west, closer to the foredeep of the Hellenic chain, and likely the influence in the Rizoma area of the uplifted Pelagonian Indentor (cf infra, PI, Fig.1 and cf Fig.13) [Ferriere et al., 2004].

The Oligocene Eptachorion Formation

Overview

By contrast to Eocene formations, the Eptachorion Fm is ubiquitous throughout the MHB. It forms the lower part of the main Oligo-Miocene, NW-SE trending "Albano-Thessalian" basin. It is exposed mostly to the western border of the basin, while it is buried beneath Miocene formations to the east (Fig.3). To the south, the Theotokos-Theopetra anticline (TTS, Fig.3 and Fig.4) allows part of this formation to be exposed in the center of the MHB.

The facies are mostly marine siliciclastics but limestones are locally present at the base of the formation. The age of the formation is based on datation of benthic foraminifera (at the base), and calcareous nannofossils and planktic foraminifera (at the top). An Oligocene age is largely admitted but the time range varies from one author to another (Fig.7 and infra).

Facies of the base

<u>Limestones</u>

Where the Oligocene transgression comes onto hard substrates, namely Mesozoic limestones, shallow-water carbonate reefs develop, and they might there replace the basal conglomerates of the Eptachorion formation. This is what can be observed to the north of Alatopetra (Fig.3), where the reefs were formed by scleractinian corals associated to benthic foraminifera (Lepidocyclins).

These facies are in turn rapidly buried by fine-grained turbidites. This is also the case to the southernmost part of the MHB: (i) near Mitropoli (western border), where Oligocene limestones overlie the Mesozoic limestones of the Koziakas range [Savoyat et al., 1969b; Lekkas, 1988; Ferriere et al., 2011]; (ii) near Farkadon (eastern border) where the limestones are not followed by a siliciclastic series. At least for the first instance, the limestones are dated from the Uppermost Oligocene [Lekkas, 1988; Ditbanjong, 2013].

<u>Conglomerates</u>

Elsewhere (and in most places), the lower part of the Oligocene series is commonly composed of conglomerates, which record the major tectonic phase of the Eocene/Oligocene boundary. These conglomerates may reach 1,5 km in thickness, as interpreted from seismic profiles [Kontopoulos et al., 1999; Zelilidis et al., 2002]. However, they have not been studied in detail at the outcrop.

The clast lithology of these conglomerates reflect the bedrock lithology of the basement highs exposed in the vicinity of the outcrops. Desprairies [1979] shows that they consist of mostly ophiolitic clasts derived from the Pindos area (to the west of the basin), but this may be biased by the fact that most of his studied samples come from the western border of the MHB. In Avra (to the east of Kalambaka, Fig.4D), the conglomerate clasts are mostly composed of gneisses derived from the Pelagonian zone (to the east of the basin).

From turbidites to shelf marls

The facies of the base are overlain by sandstones and shales interpreted as turbidites (areas of Zouzouli, Alatopetra or SE Krania). These deposits largely overlap the facies of the base and also overlie the Eocene formations (near Parorio-Trikomo) or even the basin basement (Cretaceous limestones to the north of Theopetra) [Ferriere et al., 1998; 2004].

These dominantly silicic clastic deposits show an upward increase in carbonates, finally passing to marls, which is the most typical facies of the Eptachorion formation throughout the MHB. These marls contain locally pelagic foraminifera, small bivalves and floated wood fragments (Pl. I-F and I-G).

Near the Meteora, the Eptachorion deposits beneath the Lower Meteora Conglomerates are a little bit different. The marls are underlain by decametre-thick successions of alternating silty marls and cross-stratified sandstone beds, with locally preserved HCS, many trace fossils (planolites) and mud clasts (Pl. I-E). The marls contain large benthic foraminifera [Savoyat et al., 1972a] molluscs (gastropods, pectinids, oysters) and plant remnants. The marls and sandstone deposits point to a shallow water coastal or bay setting.

The Oligocene age of the Eptachorion marls was early determined owing to mollusks, large benthic (Lepidocyclins) or pelagic (Globigerinids) foraminifera, or calcareous nannofossils [Brunn, 1956; Savoyat et al., 1972b; Zygojiannis and Muller, 1982; Barbieri, 1992; Kontopoulos et al., 1999; Ferriere et al., 2004]. However, while ages based on Globigerinids (biozones 20-21) might be well established [Barbieri, 1992], those based on nannofossils vary from one author to another [Zygojiannis and Muller, 1982; Kontopoulos et al., 1999; Zelilidis et al., 2002; Ferriere et al., 2004] (see Fig.7).

Interpretation : rapid and widespread subsidence of the MHB

The Oligocene Eptachorion Formation corresponds to a transgressive sequence. The Eptachorion marls point to an offshore setting, at distance from terrigenous supply, by contrast to the underlying conglomerates and turbidites. But this facies shift might also be (partly) due to a progressive levelling of the post-tectonic landscapes surrounding the basin.

There are also significant lateral variations of the water depth associated to the upper, finer- grained part of the formation. The Eptachorion marls are interpreted as outer shelf to bathyal deposits based on their microfaunal content, mainly nanofossils and Globigerinids, which might indicate a water depth up to 600 m [Barbieri, 1992]. To the south, near the Meteora area, the interbedded marls and sandstones indicate a more proximal, lower shoreface to upper offshore/bay setting sporadically supplied by sediments derived from the coast.

Whatever, considering the shallow water of the shelf limestones of the base of the Oligocene series, eustatic changes cannot be the only cause for the recorded transgression (a rise up to 600m). It implies a strong subsidence, the causes of which are not determined. The nature of the crust beneath the MHB, which is locally loaded by thick ophiolitic nappes (Vourinos to North- Pindos section), may control subsidence variations based on density and rheology contrasts. Also, processes related to the underthrusting of external zones during the Oligocene could enhance these inherited heterogeneities (i.e. tectonic erosion, for instance; cf. infra, geodynamical setting).

The Upper Oligocene (?) - Lowermost Miocene Tsarnos - Pentalofon Formation

Overview

The Pentalofon Formation is attributed to the lowermost Miocene and possibly to the uppermost Oligocene (cf. infra). It is composed of marine conglomerates and sandstones and records an increase in energy of the sedimentary processes as compared to the Oligocene. This is the response to the uplift of the eastern border of the MHB.

One major characteristics of this formation is a sediment source fully localized in the internal zones (Pelagonian basement), to the east of the basin, which supply mostly Paleozoic gneisses and Triassic marbles to the deposit. This eastern feeder was already active in the Oligocene (cf. Avra, Fig. 4 and Fig. 16B), but subdued by comparison to the Pindos zone (to the west), which was then probably higher and supplied a more weatherable and erodible material (ophiolites and flysch).

The lower boundary of the Pentalofon Formation is a sharp lithologic contact well expressed to the south of the MHB, where the Lower Meteora conglomerates overlie the Eptachorion marls (Fig. 6). To the north, the Pentalofon Fm is finer grained. Its lower part consists of the Taliaros Fm or Tsarnos Fm of Brunn [1960], Kontopoulos et al., [1999] and Zelilidis et al.. [2002]. Based on seismic profiles, Zelilidis et al., [2002] evidence within the Pentalofon Fm a lower member (Tasrnos) and an upper member (Kalloni). These members would correspond, to the south of the MHB, to the Kalabaka and Anthrakia members respectively (Fig.8).

The lower part: Taliaros or Tsarnos formation

This formation does not exist to the south of the MHB (Fig. 3). It is composed of sandstones, marls and scaphopod - and ahermatypical coral-rich, gravelly limestones [Brunn, Pentalofon map, 1960]. The sandstones are well bedded (to the north of Alatopetra for instance). They are interpreted as submarine inner and outer fan deposits [Kontopoulos et al., 1999] or distal to proximal turbidites based on their sandstone-shale ratio [Zelilidis et al., 2002]. Near Eptachorion village, mass-wasting

Plate II (next page)
A) Early Miocene (Pentalofon Fm) to the north of Mitropoli (Southern MHB). Alternating upward-fining sandstone and siltstones. B) Pentalofon Fm: view of the Meteora Gilbert deltas (southern MHB), showing the typical westward dipping progradational wedges. C) Pentalofon Fm, Meteora Gilbert deltas (south MHB). Cliff view showing grain-size variations. The channel is of decametric scale. D) Pentalofon Fm. Conglomerate facies. The pebbles are composed of Triassic and Jurassic white marbles and Pelagonian Paleozoic gneisses, with locally minor ophiolitic clasts. E) Pentalofon Fm: imbricated pebbles showing a paleocurrent to the SW in the conglomeratic deposits. This direction is about the same for all the conglomerates of the Pentalofon Fm in the southern MHB. F) Tsotyli Fm. A few km to the north of Meteora, near Asproklissia, sandstones and silstones with interbedded conglomerate lenses. These facies occur between thick conglomerate successions not seen on the photography. G) Tsotyli Fm (early Miocene): sandstones and conglomerates with numerous schist and gneiss clasts (east of Asproklissia). H) Ondria Fm (early-mid Miocene): Clypeaster-rich limestones


A
B
C
D
E
F
G
H

features are preserved in these facies (slumps) (Pl. I-G).

The main part: Pentalofon Formation

Overview

The main Pentalofon Fm is coarser-grained and consists of materials derived from the Pelagonian basement mainly [Brunn, 1956]. It records facies variations from north to south [Desprairies, 1979; Zelilidis et al., 2002; Ferriere et al., 2004 and 2011].

To the north, it forms the main reliefs of the MHB landscapes. There, it is 2-4 km thick (Kalloni Fm, as defined from seismic profiles by Zelilidis et al. [2002]).

The mainly terrigenous deposits of the Pentalofon Fm, range from well bedded conglomerates to marly shales. These facies commonly form thickening and coarsening upward successions, first interpreted as shelf deltas [Desprairies, 1979], and then associated to deep-sea fan turbidites [Kontopoulos et al., 1999; Zelilidis et al., 2002] (Pl. I-H). Such regular bedded alternating sandstones and siltstones also exist on the southernmost western border of the MHB, near Mitropoli (Fig.1B and Pl. IIA) where they cover thick conglomerates.

In its southeastern part the Pentalofon Fm is coarser grained and has been interpreted as fan- and shelf deltas [Kontopoulos et al., 1999; Zelilidis et al., 2002], namely the famous Gilbert deltas of the Meteora area [Ori and Roveri, 1987].

The Lower Meteora Conglomerates (LMC)

The LMC rest by a locally channelized erosional contact above dominantly sandy-silty marine deposits of the Oligocene Eptachorion Formation (Fig.6). The LMC, emplaced at the narrowest part of the MHB, have been described in detail by Ori and Roveri [1987]. It is composed of homogeneous, matrix to clast-supported well rounded conglomerates (Pl. II-B and II-C). The clast lithology is dominated by Paleozoic gneisses and Triassic marbles that point to a source located NE of the Meteora area, in the Pelagonian domain (Pl. II-D and II-E). The ophiolitic pebbles are generally not abundant. It is worth noting the absence within the LMC of clasts of the lithologies originated from the Koziakas ranges, which form the basin basement on the western side of the LMC (Fig.3).

The LMC are basically formed by the stacking of wedge-like units composed of clinoforms over 10° dip, which were interpreted as Gilbert-type, fan deltas prograding toward the basin axis in about 30 m water depth (Fig. 9). The feet of the deltas are overscoured by up to 30 m deep channels indicating strong currents constricted in the axis of the basin. The basin might have been overfilled by these deltas, as the same conglomerates are preserved on its other side near Mitropoli and Koziakas mountains. The wedges would correspond to highstand deposits, while the channels would be formed and backfilled during subsequent lowstand and rise of sea level. The maximum thickness of the wedge stack is about 300 m at the outcrop. The stacking pattern shows an upward decrease of the slope dip of the topsets. This points to a tectonic tilt and uplift of the hinterland, the western border of the Pelagonian Indentor (cf. infra) (Ferriere et al., 2011).

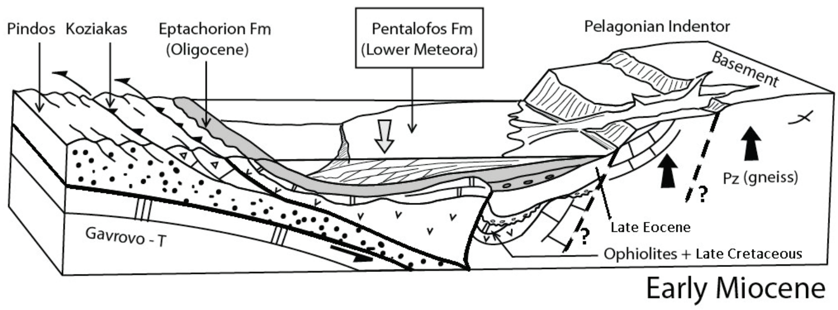


Figure 9
Meteora Gilbert delta (early Miocene): Schematic environment showing uplift in the eastern area (modified after Ferriere et al., 2011). Normal faults are hypothetical faults as they have not been proven in the field.

Age of the Pentalofon Formation

Where it is dominated by conglomerates, the Pentalofon Fm could not be dated by biostratigraphy. Elsewhere, there is a range in the published ages, depending on the area and sub-formation considered (Fig.7).

Brunn [1956] observed scleractinian corals within the Tsarnos-Taliaros Fm which he considers Upper Oligocene in age, as well as Miogypsina complanata SCHLUMB in the Pentalofon s.s. Fm which he attributes to the Aquitanian. Later datations by nannofossils confirmed this age [Zygojiannis and Muller, 1982 ; Zelilidis et al., 2002 ; Ferriere et al., 2004]. However, Late Rupelian - Chattian biozones were locally identified: NP 24 in Tsarnos member, and NP 25 in Kalloni member [Zelilidis et al., 2002]. This somewhat older age of the Pentalofon Fm was also adopted by Kontopoulos et al. [1999]. Nevertheless, the large time span of this formation raises issues as regarding to its correlation 3rd order cycles of eustatic charts.

Interpretation: raise of Pelagonian supply and rapid infilling

The deposits of the Pentalofon Fm are mainly coarse-grained detrital sediments (sandstones and conglomerates) deeper in the northern part of the MHB than in its southern one. While the main source of the Oligocene Eptachorion Fm came from the western side of the MHB, the detrital sediments of the Pentalofon Fm are feeded from the uplifted eastern side of the MHB (cf Pl. II-E). The different fan-deltas, notably the Gilbert delta outcropping in the Meteora area, fill the MHB and give rise to a regressive event.

The Lower Miocene (p.p.) Tsotyli Formation

Overview

This formation is heterogeneous, ranging from continental, coarse-grained deposits to the south, to fine-grained marine facies to the north, where it was first defined as the 'Tsotyli marls' by Brunn [1956], who aged it, based on nannofossils, to the late Aquitanian-Burdigalian (Fig.7).

To the south, in the Meteora area, it consists of mostly very coarse conglomerates alternating with poorly bedded sandstones and was there mapped as the "Upper Meteora Conglomerates" (UMC) [Savoyat et al., Kalabaka sheet, 1972a]. There, the relative amount of clasts derived from the older Pelagonian basement (mostly pre-Triassic gneisses) increases as compared to the underlying Pentalofon Fm.

These various facies are gathered into a unique formation defined as: (i) a separate, distinct depocenter on the eastern side of the MHB ; (ii) a set of deposits bounded at the base by the same, major unconformity which is well expressed in the Meteora area (angular unconformity of about 20° between the LMC and the UMC).

This depocenter has the shape of a syncline parallel to the strike of the MHB (Fig.3). The Tsotyli strata inside this syncline show an eastward offlaping, which indicates the progressive eastward displacement of accommodation and, therefore, subsidence (Fig.4). To the south of the MHB, the western side of this syncline is an old structural high, the "Theopetra-Theotokos Structure" (TTS, cf. infra and Fig.4). On the eastern flank of the syncline, the upper units of the UMC are overlying Rizoma Eocene deposits or the basement of the MHB (Fig.3 and Fig.4).

The Tsotyli deposits

The Tsotyli marls, the thickness of which is up to 1 km to the north, thin southward where they are interfingered with more sandstones and conglomerates.

To the north, the Tsotyli deposits have been first interpreted as shelf marls with some interfingering sandstones interpreted as fluvial-dominated shelf deltas [Brunn ,1956; Desprairies,1979]. More recently, they have been reinterpreted as turbidites [Kontopoulos et al., 1999] passing upward to fan-deltas [Zelilidis et al., 2002].

To the south, from Aliakmon river to the Meteora area (Asproklissia area, Fig.3), the Tsotyli deposits form a transgressive-regressive succession comprising from base to top [Ferrière et al., 2004]: (i) conglomerates and channelized fluvial sandstones with paleocurrents to the WSW (Upper Meteora Conglomerates, UMC); (ii) fine-grained and locally carbonate-rich shelf sandstones and siltstones (Pl. II-F); (iii) conglomeratic fans-deltas similar to the LMC; (iv) grey or reddish conglomerates (mostly preserved on the eastern basin border) (Pl. II-G).

Further southward, to the SE of the Meteora area, pebbles of pre-Triassic Pelagonian gneisses constitute the whole deposit.

The clast lithology within the Tsotyli Fm reflects the space and time evolution of the feeders which supplied sediments to the eastern border of the MHB:

(i) in space, one notes the predominance of ophiolitic clasts to the north, which is related to the vicinity of the Vourinos mounts, while marbles and gneisses are more abundant toward the south where exists the Pelagonian Indentor made up of these dominant lithologies ;

(ii) in time, there is an overall increase in the amount of gneisses and marbles, as the sediment source was shifted eastward into the Pelagonian hinterland, where the rejuvenating relief supplied progressively deeper rocks from the basement to the basin.

Interpretation: shift of subsidence to the east

The deposits of the Tsotyli Formation show that the MHB was deeper in its northern part as for the Pentalofon ones. The main difference between these two Formations is the major eastward migration of the depocenter and of the uplifts at the origin of the detrital material. In the southern part of the MHB (Meteora area) the Tsotyli basin is separated from the Pentalofon one by a high structural element (the TTS structure, Fig.4).

The Lower to Middle (p.p) Miocene Ondria and Orlias formations

Overview

In the Greek part of the MHB, the youngest deposits are preserved as two distinct formations at the two ends of the basin. Their lithology is not homogeneous but both have a significantly higher amount of carbonates compared to the previous ones. These two formations were initially defined in the northern part of the MHB by Brunn [1956]: The Ondria Fm is composed of alternating sandstones, limestones and marls, while the uppermost Orlias Fm, which cover a more restricted area, is composed of sandstones and bioclastic carbonates. To the south of the MHB, only the time equivalent of the Ondria Fm is exposed.

The Ondria Formation

To the north of the MHB, the Ondria Fm is well exposed. The fauna delivered by the marls provided a Burdigalian age (Fig. 7).

On the 1 :50,000 geological maps (e.g. Nestorion sheet [Savoyat et al., 1971a]), the Ondria Fm comprises several formations of higher order based on lithology: the X Fm (mostly marls and limestones, among which carbonate buildups); the Omorfoklissia Fm (sands and sandstones with interbedded Globigerinid-rich marls); and Zevgostation Fm (also alternating sands and sandstones with marly intervals).

It typically starts with marls or limestones resting above an unconformity which is more obvious to the south of the basin. Thus, near Ellinokastro-Lagadia (ca 11km east of Vassiliki, Fig.3), Echinid (Clypeaster)-rich limestones are onlapping toward the east above the basin basement. In this area, the limestones are covered by sandy turbidites which pass upward to shelf marls (cf. Fig.16B ; Pl. II-H and Pl. III-A).

The Orlias Formation

This formation occupies a restricted area. It is attributed to the Helvetian (Langhian-Serravalian p. p. ; cf. Fig.7), and comprises sandy marls, sandstones and bioclastic carbonates, mostly composed of green algae, echinids and mollusks (Ostrea cf Crassissima notamment).

Owing to the fact that they were emplaced in very shallow water depths (green algae), these deposits might correspond to the final infilling of the last marine area within the MHB.

Interpretation: final marine stage

The Ondria Fm is interpreted as a transgressive-regressive sequence. The transgressive, lower part is typically dominated by shallow-water carbonates, implying a minimum input of siliciclastics (maximum retreat of littoral sources), while the regressive, upper part of the formation records the influence of the return of these sources.

The two small areas where these formations are preserved are thought to correspond to the last marine areas of the MHB at the end of its infilling. This interpretation is supported by the fact that the strata in these formations are not much deformed (nearly horizontal) and that the facies indicate a very shallow water depth (final infilling of accommodation space).

The absence of either an erosional unconformity or conformable continental regressive strata above these formations indicate that the surrounding reliefs were eroded at the final marine stage. This is indicated by the progressive replacement of siliciclastic sediments by carbonates.

Synthesis of stratigraphic data

Overview

The sedimentary fill of the MHB, up to 4.5 km thick (Fig.3), is almost composed of siliciclastic deposits. The gravity processes dominate, as indicated by the abundance of turbidites, olistostromes, and Gilbert deltas. This depositional regime indicates an important submarine relief at the basin margins and a significant erosion in the catchments areas. It ceases for a long time only during the emplacement of the Eptachorion marls (upper Oligocene) and at the final stage of basin infilling when it is replaced by shallow-water carbonate systems (middle Miocene Ondria and Orlias Fms). The only other places where carbonates are preserved are the basin borders during major transgressions (lower Oligocene at Alatopetra, upper Oligocene at Mitropoli; Figs. 1 and 3).

The formations are organized in 5 major transgressive-regressive stratigraphic sequences, which are bounded by angular unconformities or major and abrupt facies changes (Fig.6). In most instances, even where the angular unconformity cannot be easily evidenced, the large amplitude of relative sea-level change across the boundary exceeds that of sea-level cycles as reported on eustatic charts. Therefore, the major stratigraphic sequences are controlled by tectonic deformation.

Overall evolution of water depth

The submarine, gravity flow-related facies are finer-grained moving toward the north, which is interpreted as the result of a deepening of the basin. This applies to all periods of the infilling. The deepest facies of each major stratigraphic sequence correspond to the fine-grained deposits. The Krania "flysch"sub-basin might be the deepest one, at the onset of the MHB (over 1500 m?). Water depths about 600 m are proposed by Barbieri [1982] for the Eptachorion marls. No specific water depth was determined for the Miocene sub-basins, but the submarine fan terminology used by Zelilidis et al. [2002] relates to water depths in excess of those prevailing for the shelf-to-slope range (i.e. in excess of 200 m). Thus, water depth overall decreases in the course of the basin infilling, independently from the tectonic changes which drive the major stratigraphic sequences.

This probably reflects the larger time- and space scale of formation of the Hellenic accretionary prism.

Evolution of depositional slopes

Comparison between the depositional settings of the most distal facies recorded in each major stratigraphic sequence helps to reconstruct the overall evolution of the submarine topography and infilling of the basin.

The maximum slope of the depositional system is recorded in Eocene sandstones of Krania, which also correspond to the deepest setting (common occurrence of Zoophycos traces).

The Oligocene Eptachorion marls progressively point to a lesser slope or even the distal part of a shallower basin.

The return of a steep slope, with high energy gravity deposits, is marked in the Late Oligocene Pentalofon slope fans and deltas, which recedes again within the Lower Miocene Tsotyli marls in the northern MHB. The deepest parts of the ancient formations are not at the same place than that of the youngest ones as the depocenters migrate to the east and the origin of the main part of the detrital sediments change from the W to the E of the MHB.

The basin profile achieved during the Ondria and Orlias Fms is almost flat, allowing the development of carbonate shelves even covered by terrigenous beds as in the south.

Throughout the entire basin evolution (at the exception of the latest carbonates), nevertheless, even the most distal and finest grained deposits still exhibit hints of turbidites, which suggests a likely connection at the coast with canyons or deltas. Also, the basin borders might host mostly conglomerates (alluvial fans or fan-deltas), which testifies the occurrence of neighboring steep reliefs (above sea level). A modern setting analogue could be the Gulf of Corinth, although the tectonic context is not the same.

Eastward migration of sediment sources

The clast petrography and mineralogy of the MHB deposits designate the proximal borders of the basin as the sources for sediment supply.

The Eocene Krania deposits have a high ophiolitic content. Ophiolites constitute the bulk of the basin basement and also most of the mountains that bordered the former Krania basin to the west. To the east, by contrast, Late Eocene Rizoma quartz and mica-rich deposits are not resting on ophiolites but on Mesozoic marbles and Paleozoic gneisses (due to the interplay of the basement "Pelagonian Indentor", see below).

As this lithological contrast is reflected in the clast composition of the deposits above, authors conclude that most of the basin fill was sourced in the west before the Late Oligocene (Krania and Eptachorion F.) and in the east after that time (i.e. Pentalofos and Tsotyli [Desprairies, 1979]).

That is not necessarily true for Oligocene deposits, most of which are covered by Miocene ones in the eastern part of the MHB. For instance, the Oligocene conglomerates outcropping in the eastern Kalambaka area (Avra, Fig.4D and cf Fig.16A) are not rich in ophiolitic elements but yet they are rich in gneiss pebbles coming from the pelagonian domain located on the eastern side of the MHB.

Miocene transport paths may locally be more complex, especially because of the northward deepening of the depositional setting but paleocurrent analysis based on many clasts imbrication, show that most of the basin fill was sourced in the east since the Early Miocene (i.e. Pentalofon and Tsotyli; [Desprairies, 1979 ; Ferriere et al., 2011]), even on the westernmost part of the MHB, as in the Mitropoli area (Pl. II-E).

Eastward migration of subsidence

Starting in the Oligocene, a generalized subsidence of the MHB is recorded by the basin depocenter of each major stratigraphic sequence. Subsidence progressively shift towards the east (Figs. 10 and 11).

Concerning the amount of subsidence and its behaviour in 3D, the lack of available boreholes, drillholes, and seismic profiles, as well as local erosion within the series and the fact that turbiditic slope depositional systems are poor bathymetric indicators, bring about uncertainties.

However, some curves based on extrapolated vertical sedimentary records, from outcrops, map- derived cross-sections and published seismic profiles have been proposed [Kontopoulos et al., 1999 ; Ferriere et al., 2004] (Fig.11).

The general pattern of the eastward migration of MHB depocentres and associated subsidence is synthesized by the map in Fig. 10 (see also schematic curves Fig.11). The forcing mechanisms are discussed below (see Fig. 19).

The two subsidence curves proposed by Kontopoulos et al. [1999] evidence an uplift stage during Pentalofon sedimentation, from 21 to 16 Ma, but they are only representative of the axis of the present MHB (areas of maximum residual thicknesses).

Eustatic versus tectonic controls

The paleogeographical sketch (Fig.10) is based on subsidence but also lateral and vertical facies variations. The frequent absence of shoreline deposits and possible erosion bring(s) about uncertainties concerning the true extension of the basin limits through time. Therefore, the proposed limits minimise the marine depositional area extension: i.e., the Rizoma and Krania sub-basins probably respectively extended to the north and to the east along the tectonic structures of the Pelagonian Indentor, but they were certainly partly eroded at the Eocene - Oligocene boundary.

Locally, the limits of drastic facies changes follow tectonic structures (as at the Krania sub-basin northern limit, Fig.3 and cf. fig.14). However, even if the major geographic gap between Pentalofon and Tsotyli Formations follows the Theopetra-Theotokos structure (Fig.4), no major facies change occurs at the boundary between these two conglomeratic Formations in this area.

By contrast, the main lithologic changes from Eptachorion marls to Pentalofon

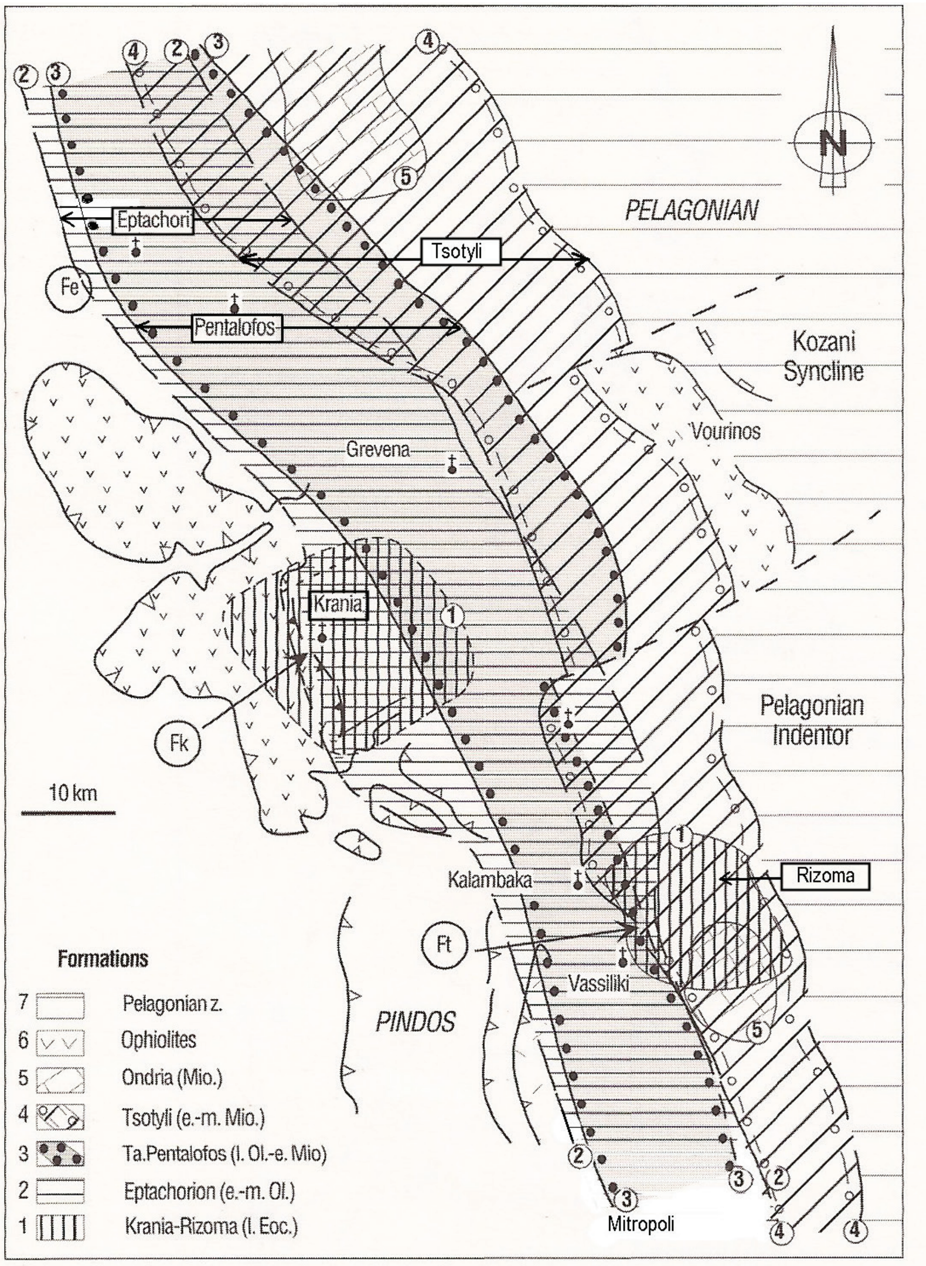


Figure 10
Paleogeographic synthesis of possible basin and sub-basins extension at different stages of MHB evolution (modified after Ferriere et al., 2004). The limits here minimise the depositional areas (e.g. we could not exclude a possible connection of the sea between Krania and Rizoma in the upper Eocene, especially along the tectonic structures bounding the Pelagonian Indentor). Abbreviations: Eoc: Eocene, Ol: Oligocene, Mio: Miocene, e: early, m: middle, l: late. 1 to 5: different MHB Formations, 6:Ophiolites and 7: infra-ophioltic Pelagonian basement.

conglomerates are not associated to major changes in basin limits. This would mean that, if the lithologic change is triggered by tectonics, the tectonic hinge line was located near the paleocoast, as was the case in the Meteora area [Ori and Roveri, 1987; Ferriere et al., 2004].

Alternatively, this suggests other kinds of controls as climate or eustatic changes. Zelilidis et al. [2002] argue that all the stratigraphic occurrences of lowstand facies compare closely with published eustatic sea-level curves (Fig. 12) . However, this apparent correlation remains more than questionnable, because (i) there is no accurate biostratigraphic control and (ii) the required tectonic calendar is not accurately established.

Zelilidis et al. [2002] suggest that the major Oligocene sea-level drop would be responsible for the abrupt change from Eptachorion marls to the Lower Meteora Conglomerate (Fig.12). However, Ori and Roveri [1987] suggested a tectonic control at origin and development of the Meteora deposits, which was evidenced by Ferriere et al. [2011].

If eustatism changes have, of course, some control on the marine deposits of the MHB, in the next section, we summary the data showing that tectonic is the most important control on the paleogeographic evolution of the MHB (cf. infra).

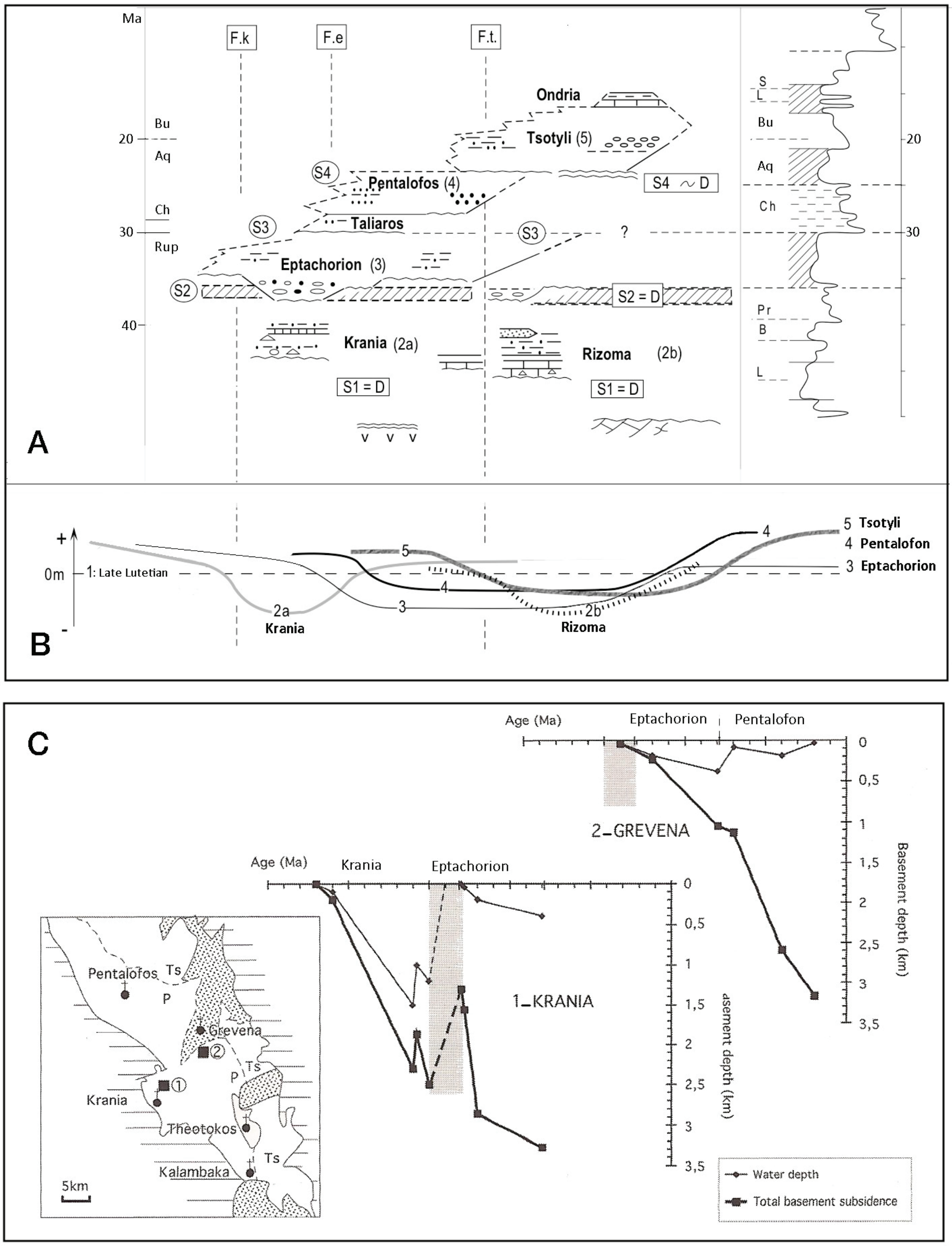
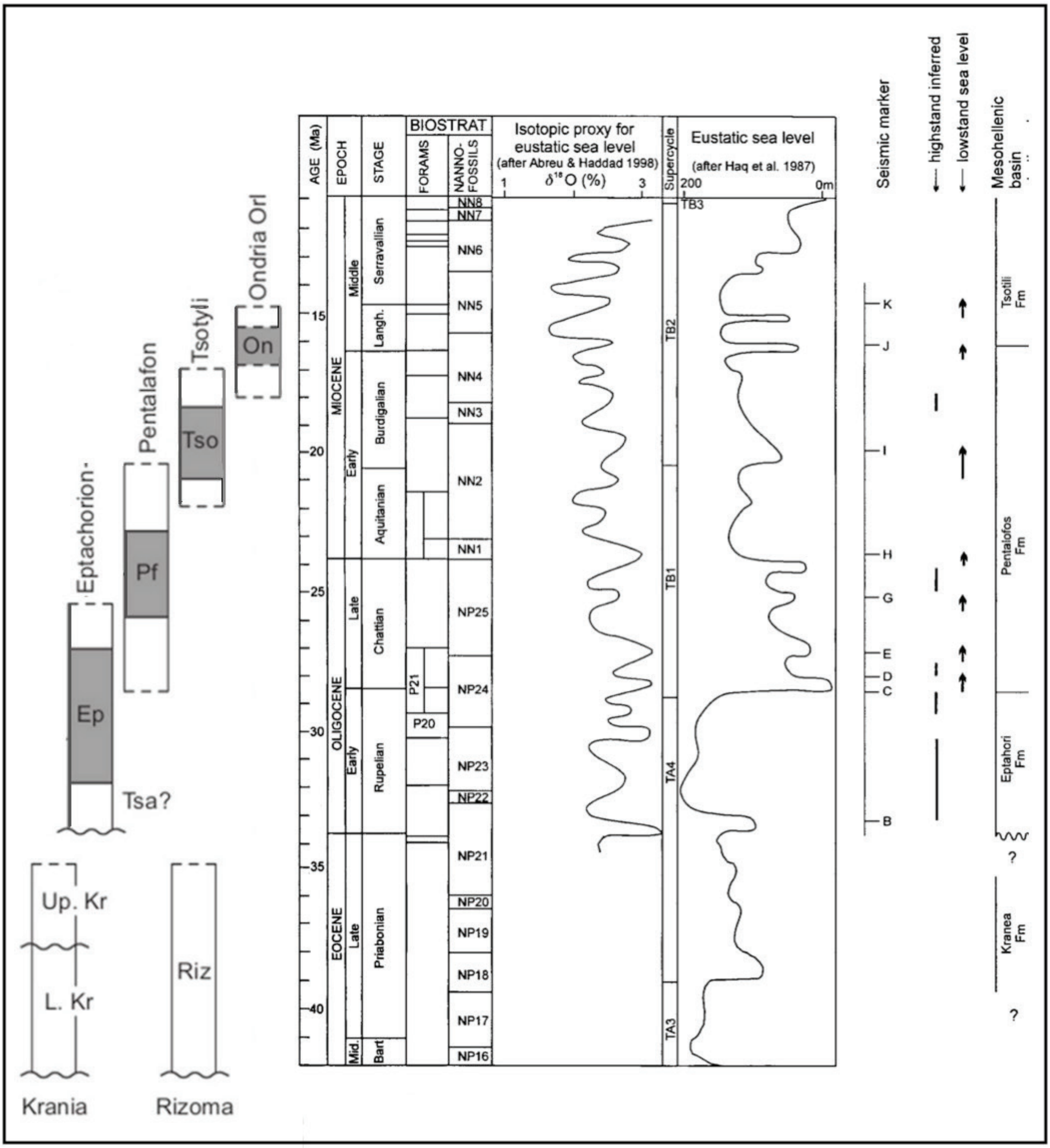


Figure 11
Subsidence characteristics.

A and B: Schematic drawings showing subsidence migration toward the east (A: lithologic Formations ; B: schematic curves). Abbreviations: L (Lutetian) to S (Serravalian) : Eocene to Miocene stratigraphic stages ; Fe, Fk, and Ft: faulted-flexures of Eptachorion (Fe), Krania (Fk) and Theopetra-Theotokos (Ft) ; S: significative surfaces ; D : Unconformities.

C: Subsidence curves concerning the central part of the MHB (1:Krania and 2: Grevena series) from our own field data and from seismic published data [Kontopoulos et al., 1999 and Zelilidis et al. 2002]. Approximations on the subsidence calculations are related to some age (see Fig.7) or Formation thicknesses uncertainties, and mainly to paleobathymetric data, particularly for deep water facies (i.e. turbidites). Backstripping has been computed with SUBSILOG (Dubois et al., 2000), using the standard parameters defined by Sclater & Christie (1980). Grey area (ca 35-33 Ma) corresponds to the main compressive tectonic event at the Eocene-Oligocene boundary. Modified after Ferriere et al. [2004].



TECTONIC DEVELOPMENT OF THE MHB

General overview

The Mesohellenic Basin corresponds to an assymetric syncline. On its western border, strata are generally steeply dipping eastward and are locally vertical to slighly overturned (Eocene and western Oligocene deposits). On the eastern border, the Paleogene formations are generally absent from outcrop, except in the South (i.e. Rizoma formation), and Miocene strata are gently dipping westward (Fig. 4).

Eocene strata are much more deformed than Oligocene and Miocene ones. The Oligocene Eptachorion Formation is uncomformably overlying the two folded small basins of Krania and Rizoma (Fig. 4).

Figure 12

Eustatic sea level variations compared to the MHB lithologic formations ages showing the uncertainties concerning the eustatic control of the MHB evolution. Right part of Fig.11 from Zelilidis et al. [2002]; left part from our own data.

Oligocene formations appear mostly on the western side of the MHB while the Miocene strata rest onto the basement on the eastern side because of eastward migration of depocenters. The southern part of the MHB is more complex as the main syncline splits into two narrow synclines separated by a major structural high : the Theopetra-Theotokos Structure (TTS) (Pl. III-B). This complex structural high corresponds broadly to a main faulted anticline (Theotokos Anticline : At), with a major fault (Theopetra Fault: Ft) on its eastern border (Figs. 4 and 13). Moreover, in this southern part of the MHB, the total width of the basin domain is much smaller than northward. This narrowing is related to the development of a particular structure raised within the basin basement : the Pelagonian Indentor (Fig. 13A).

Major faults and structures, mostly parallel to the basin strike, have been recognized by previous authors but with various interpretations [Doutsos et al., 1994 ; Ferrière et al., 1998, 2004; Vamvaka et al., 2006 ; Ferrière et al., 2011]. Small scale deformations are still poorly constrained in kinematics and age (Fig. 13B).

Some results on detailed brittle deformation analysis have been firstly proposed by Doutsos et al. [1993, 1994] but they were poorly constrained in age. The chronology of tectonic deformation of the MHB has been proposed from analysis of major and minor deformation patterns [Ferrière et al., 1998, 2004, 2011; Vamvaka et al., 2006]. These results on tectonic deformation show that the present MHB is the result of successive tectonic episodes among which the main one is late Eocene in age.

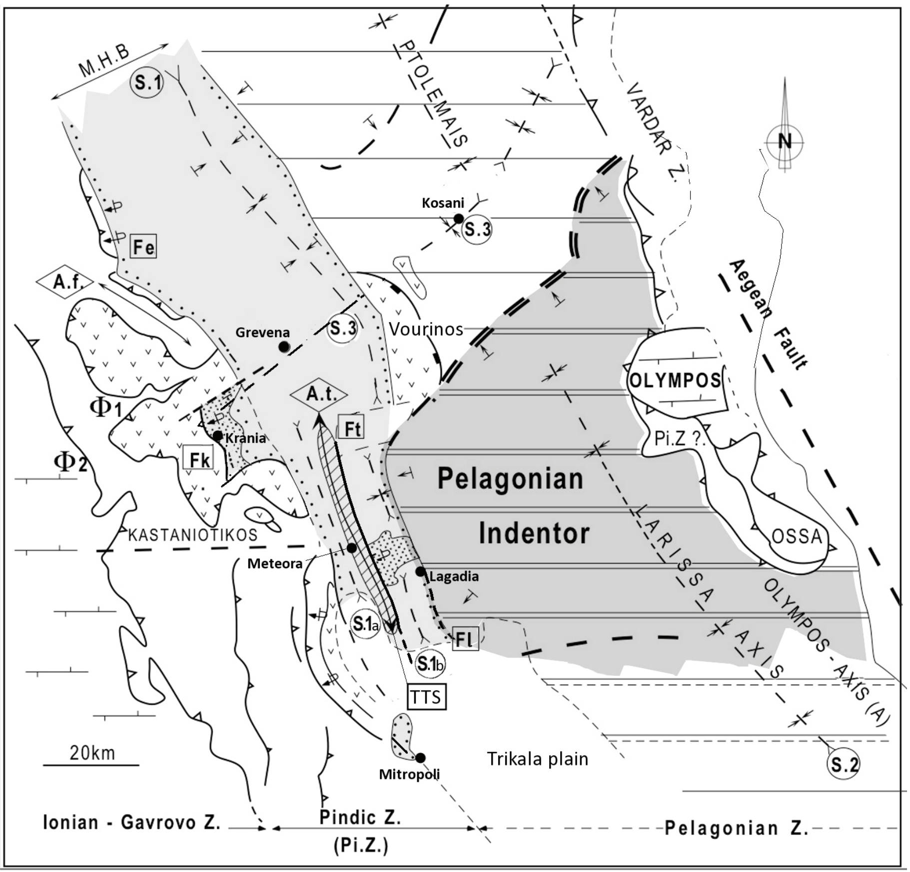


Figure 13A

Synthetic structural map of the MHB and internal zones of Hellenides (modified after Ferriere et al., 2004).

Note the relationships between the Pelagonian Indentor (double thin lines bounded by double dashed lines for the northern flexure) and: i) the Theopetra-Theotokos Structure (TTS=At+Ft) , ii) the Rizoma elongated subbasin (S.1B) on the west, and iii) the structural saddle of Kozani and Krania sub-basin on the north-west (S.3) . We can also observe that this transverse crustal structure(S.3) does not present any significant strike-slip offset that could be the result of a movement along the TTS. S.(1 to 3): Synclines ; A: Anticlines, Af: Filippi anticline, At: Theopetra-Theotokos anticline or structural high ; Fk, Fe and Ft: faulted-flexures of Krania, Eptachorion and Theopetra.

The southwestern border of the MHB

The deposits from this southwestern border of the basin are essentially Oligocene in age, except in the Krania area (late Eocene) and in the Mitropoli area (early Miocene) (Fig. 3). Various tectonic structures were described from this border and they can be complex and polyphased [Ferrière et al., 1998, 2004, 2011, Vamvaka et al., 2006].

Deformation within Oligocene series

According to some authors, the contact between the Oligocene deposits and the basement is outlined by a large fault. However the interpretation differs from a large west-

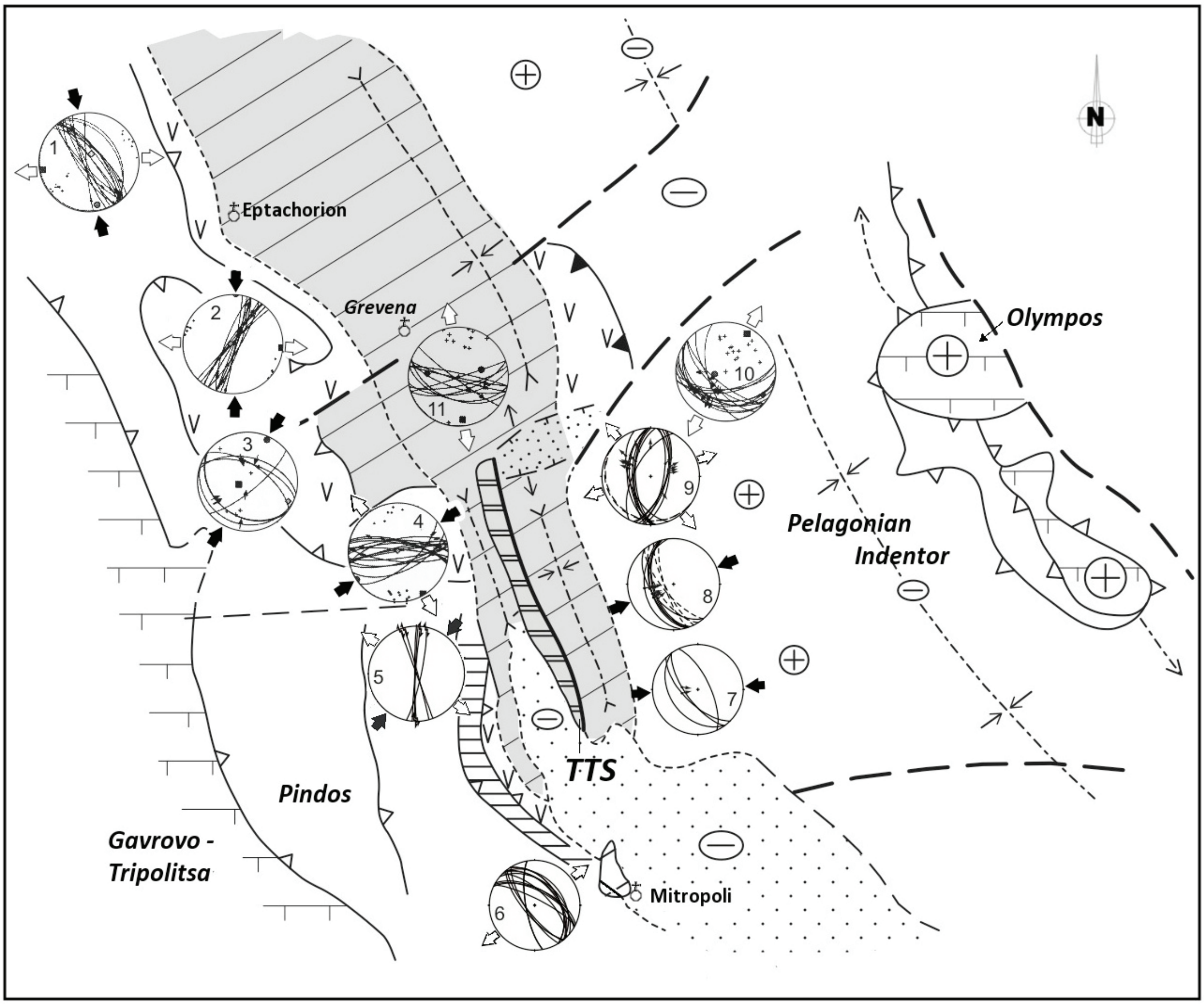


Figure 13B

Some stress stereoplots relative to the MHB (lower hemisphere). Stereoplots 1 to 4 and 10-11 after Vamvaka et al., [2006]; 5 to 9 after Ferriere et al., [2011].

On stereoplots, blacks arrows represent compressional directions from reverse faults; white arrows extensional directions from normal faults.

Sites 3, 7 and 8 in late Eocene series: reverse faults developed during the compressive late Eocene event. Sites 1, 2 and 4: Oligocene-early Miocene strike-slip event, after Vamvaka et al., [2006]. Site 6: synsedimentary early Miocene normal faults with 2 to 10 m normal offsets (no reliable striation could be observed on these fault planes) (Pl. III-E).

Site 5 in Oligocene series: strike-slip faults. Sites 9 (in Oligocene series), 10 and 11: normal faults developed during Oligocene to Quaternary times.

ward dipping thrust fault [Doutsos et al., 1994] or a steeply dipping strike-slip fault [Zelilidis et al., 2002; Vamvaka et al., 2006] (cf. Fig. 18).

We consider that three segments have to be distinguished on this MHB-SW border, from North to South:

- i) North of Krania (northern MHB, Figs.1, 3 and 13A) Oligocene series are stratigraphically resting unconformably over the basement with high dips to vertical ones (North of Alatopetra) and even locally overturned (North of Eptachorion). There is no continuous fault along this deformed border but mainly some collapse structures. This part of the western border corresponds basically to a faulted flexure (Fe, Figs. 4 and 13A) [Ferriere et al., 1998, 2004].

North of Alatopetra (Figs. 3 and 13A), the development of the Filippi anticline seems to control the progressive tilting of Oligocene strata, indicating that the development of this faulted flexure is at least partly Oligocene in age (Pl. III-D).

- ii) East of Koziakas range (southern MHB, Fig.1B), a major steeply dipping fault separates Oligocene marls from Mesozoic limestones of Koziakas Range (Fig.13A and cf. Fig 16A-B). This fault evokes a normal fault with an overall large offset toward the basin, but small scale deformations next to the fault plane reveals some reverse motion. The age of deformation, and possible inversion along this fault, could not be established in that area.

- iii) East and South of Krania (central MHB, Fig.4C), Oligocene strata rest unconformably over the ophiolitic basement with moderate tilting toward the basin (dipping NE, 10 to 40°). These areas confirm the absence of a continuous large fault bordering to the west the Oligocene formations of the MHB.


A
B
T
P
E
TTS
C
D
E
F
G
H

Structures in other formations on the SW border of the MHB

Deformations of Eocene deposits from the SW border: the Krania area

The Eocene series of the Krania area (Fig. 14) are the oldest (Lutetian to late Eocene) and the most deformed strata within the whole MHB showing faults, folds and also a lot of olistolites and slumps (Pl. III-C).

These series were deposited in a restricted area corresponding to the Krania subbasin that is bounded to the west by a large flexure striking parallel to the MHB axis. This flexure shows vertical to overturned eocene strata and faults dipping roughly vertically (Fk, Figs. 13 and 14). The faults offsets seem apparently relatively moderate because the main fault separate the ophiolitic basement from the basal conglomerates of Eocene subbasin constituted of ophiolitic detritus and blocks.

Eocene strata from Krania subbasin, where close to the faulted flexure (e.g west of Microlivadon), are locally affected by decametric reverse faults with an eastward vergence (Fig. 14). The deformation can be attributed to the late Eocene main compressional episode as the Oligocene series are sharply unconformable on top of vertical eocene beds.

Deformations in Early Miocene series from Mitropoli area, Southern MHB

The Mitropoli area (Fig 15) is characterized by the presence of thick siliciclastic series, early Miocene in age, resting locally on top of Late Oligocene reefal limestones or, more generally resting directly on top of the Mesozoic basement of the Koziakas and Pindos Ranges (Figs.1B and 13A). These strata are gently dipping, 20 to 30° NE, toward the basin axis (Fig. 15) [Ditbanjong, 2013].

Some steeply dipping large faults, NW-SE directed, are crosscutting the whole early Miocene series. They are thus younger than the early Miocene but they could not be attributed to a precise episode of deformation (Fig. 15). However, some other faults observed in the earliest Miocene strata are undoubtedly synsedimentary (Pl. III-E). The direction of extension could not be constrained precisely because of the lack of reliable striations on fault surfaces. From the spatial distribution of the fault planes, this extensional deformation has to be roughly NE-SW directed (Fig.13B, stereonet no.6). These synsedimentary faults represent the creation of a depocenter in this southern part of the basin during the early Miocene, later than the other parts of the MHB.

These listric synsedimentary faults are decametric in scale with offsets up to several meters. They are therefore not regarded as crustal faults but they reflect the deepening of the depocenter and the apparition of a significant slope toward the basin (toward the NE). The progressive onlap of marine coarse- grained followed by fine-grained sediments over the listric faults illustrates the progressive filling of that depocenter on an eastward dipping slope.

The eastern border of MHB

The eastern border of the Mesohellenic Basin is characterized by the onlap of the most recent deposits of the basin, early to middle Miocene in age, over the Pelagonian basement (Figs. 1 and 4). The basin strata are generally gently dipping westward, toward the axis of the MHB. In the southern MHB, these recent deposits are covering the small Eocene Rizoma sub-basin. Regarding the tectonic characteristics of this basin border, two segments have to be distinguished depending on the presence (southern segment) or the absence (northern segment) of the Pelagonian Indentor (Fig.13A).

Eastern border structures in the northern Segment

This border of the MHB shows, at least locally, some normal faults but the outcrop conditions and the widespread recent alluvial deposits do not allow to demonstrate a clear continuity along these faults (e.g. Orestikon sheet, Savoyat et al., 1971b). However, in many areas the Miocene series of Tsotyli Formation are unconformably covering the Pelagonian basement with a moderate dip whereas the Pelagonian basement is much higher laterally eastward. It implies a significant offset of the Pelagonian basement to account for this shift in elevation of the basement.

Eastern border structures in the southern Segment

In the southern MHB, the eastern boundary of the basin shows a westward shift because of the presence of the Pelagonian Indentor (PI, Fig. 13A) responsible for higher

Plate III (previous page)

A) Ondria Fm near Ellinokastron (SE MHB) : view of the upper, clastic part of the formation showing sandstone and siltstone alternations. This deposit is a part of the youngest formation of the MHB fill in the south of the basin. B) View of the TTS high (TTS as Cretaceous limestone quarry) with overlying Miocene Meteora conglomerates (P: Pentalofon) on its western flank covered by the Tsotyli conglomerates (T) in the background. The hills in between are composed of Oligocene Eptachorion deposits (E). C) Krania Fm (late Eocene): metre-thick slumps in the lower part of Krania Fm. The scale in the sky is the wellknown Pr. Piper. D) Steep dips (60-70°) of Lepidocyclin-bearing marly beds forming the base of Eptachorion Fm at the SW-MHB border (north of Alatopetra). These beds rest above Cretaceous limestones (to the left). E) Synsedimentary normal faults within early Miocene deposits (Pentalofon s.l.) near Mitropoli, composed of siltstones, sandstones and conglomerates. These SE dipping faults are sealed above an unconformity by a westward onlapping deposit of similar facies as those below. F) Reverse fault across sandstones of the Pentalofon Fm. Note the slump at their base. G) Strike-slip features in the Oligocene deposits near Theotokos : low dipping slickensides (8 to 10° N), subparallel to the stratal dip (lower surface) are found on subvertical fault planes directed NS. The slickensides indicate a dextral motion along the fault. H) Normal faults on the border of the post-middle Miocene basin of Karpero. This recent basin is striking ENE-WSW.

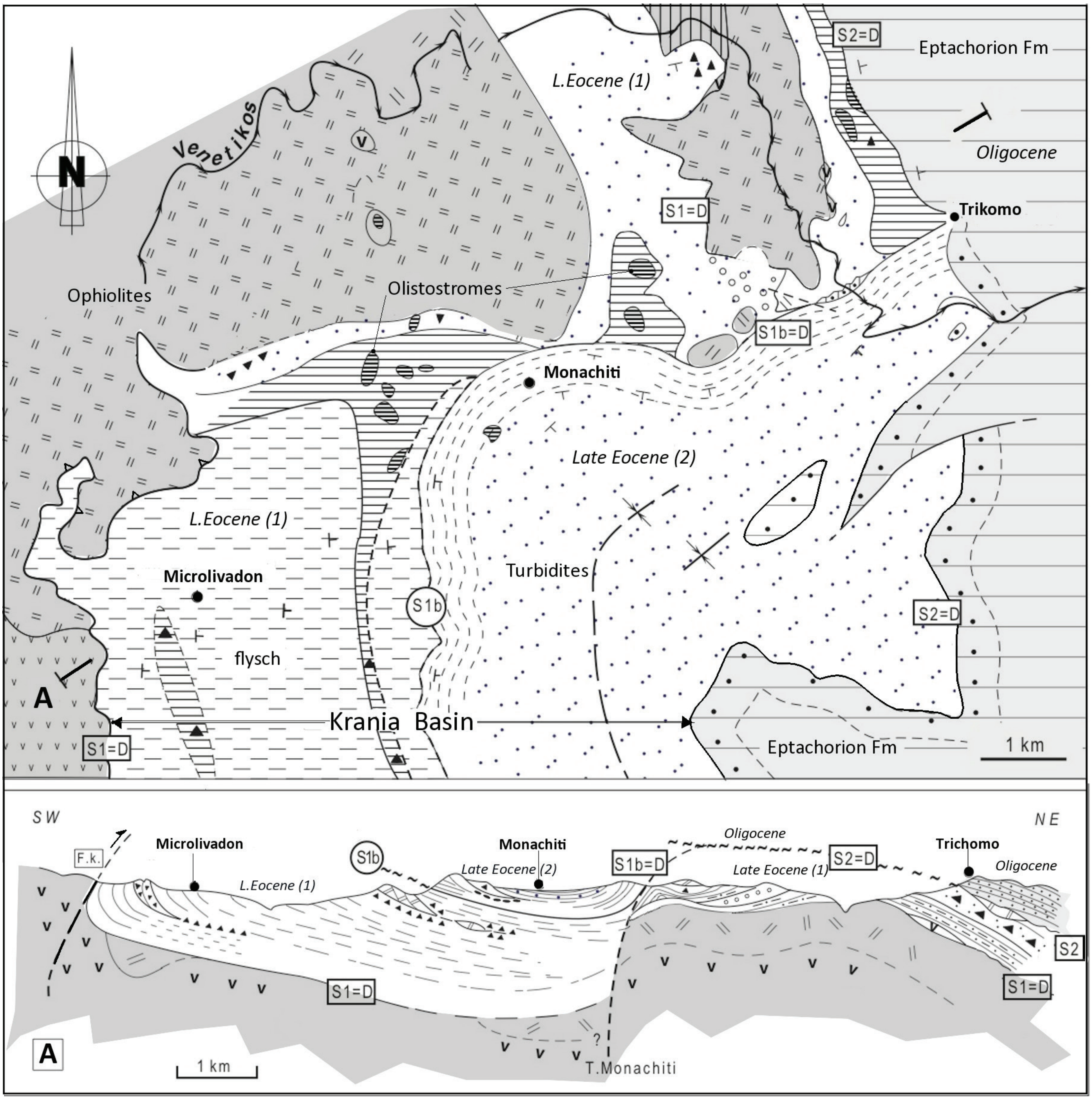


Figure 14

Geologic map and cross-section of the northern border of the Late Eocene Krania subbasin; location: see Fig.3 (modified after Ferriere et al., 2004). L. Eocene (1) and (2): Late Eocene Krania lower series (1) and Late Eocene Krania upper series (2). T. Monachiti: Monachiti Transverse structure. **Lower Krania series:** Points: fine grained sandy bay deposits; small circles: well rounded conglomerates; olistostromes: large olistostromes with a breccia matrix (horirontal lines on the map; black triangles on the cross-section) and pluridecametric, massive or little brecciated olistoliths (patches with thick horizontal lines on the map, limestones blocks on the cross-section). **Upper Krania series:** turbidites (flysch) above thick sandstones (dashed lines). Eptachorion Oligocene Formation: conglomerates at the base (points) and sandstones above. S (S1, S1b, S2) Surfaces indicating major tectonic events; D: angular unconformities (see also Fig. 6). Vertical and horizontal scales are similar (no vertical exaggeration, v.e.=1).

elevation and westward indentation of the basin. In this particular area, various Tertiary formations rest unconformably directly on top of the Mesozoic basement: late Eocene Rizoma Formation, early Miocene Tsotyli Formation, and early to middle Miocene Ondria Formation. The complex onlap relationships over the basement reveal a succession of tectonic movements within that area during the whole tertiary evolution of the MHB.

Some subvertical faults, directed NW-SE, can be observed on the border between the Pelagonian Indentor and the MHB (Figs. 16A and 16B). At least one of these faults shows an important normal motion outlined by the downward motion of Late Cretaceous limestones relative to the Triassic marbles. This particular fault is unconformably covered by Clypeaster-bearing limestones of Ondria Formation, early to middle Miocene in age (see Fig. 16B, cross-section C). This fault is likely to have been active during Oligocene to early Miocene times.

Some recent deformations can also be outlined from this area because the youngest Ondria Formation is affected by flexural folding particularly developed on the Pelagonian Indentor border (Fig. 16B). We consider therefore that this folding can be controled by some westward and/or vertical motion of the Pelagonian Indentor after the Middle Miocene.

The Theopetra – Theotokos Structure: synsedimentary tectonic activity in the southern MHB

Description of the Theopetra – Theotokos Structure (TTS)

The TTS is a 40 km long complex structural high separating the southern MHB in two synclines, the western one occupied by the Pentalofon Formation, the eastern one by the Tsotyli Formation (Fig. 3). This structure includes two distinct areas with pre-MHB basement outcroping in the Kalambaka area [Savoyat et al., 1972a] and further North on the Agiofillion geological map [Mavridis and Matarangas, 1979] also referred in earlier work as the Theotokos thrust [Doutsos et al., 1994]. We demonstrated that both of these separate areas represent outcrops of a single complex structure, the TTS, that controlled the basin geometry in this zone since the late Eocene [Ferrière et al., 1998; 2004; 2011].

The TTS structure is a complex faulted anticline in the axis of the MHB (Figs. 13A and 16). To the north, it is mainly a faulted anticline within Oligocene strata (Theotokos anticline, A.t, Fig. 13A). In the South, this anticline is dissected by faults and represents therefore a complex fault zone. The main fault appears on the eastern side of the TTS and corresponds to the Theopetra Fault (Ft, Figs. 13A and 16). It is a major boundary within the basement as ophiolites appear mainly west of the fault and are generally absent from the other side. Some faults must also separate the southern TTS from the Plio-Quaternary Trikala plain to the West. These suspected faults could not be observed from outcrop because of the recent cover of Quaternary gravels.

The TTS has a polyphased structural development as evidenced by successive unconformities (Fig. 16B). It controlled the depositional areas in the southern MHB from Eocene to Miocene times (see Fig. 20). It corresponds to the western boundary of the Rizoma outcrops, acting as a compressional structure with significant reverse motion along the Theopetra Fault. This reverse motion is clearly late Eocene in age as it affects the middle to late Eocene series and the fault is unconformably covered by Oligocene strata [Ferrière et al., 1998 ; 2004]. These Oligocene sediments are coarse conglomerates made of Pelagonien gneissic pebbles on the eastern flank of the TTS whereas they are mainly fine grained marls and fine sandstones on the western side in the Basin axis. The massive trapping of coarse material on only one side of the TTS supports also the high topographic position of the TTS during Oligocene times.

Moreover the TTS was also active during Miocene times. This is supported by the syn-tectonic fan of the early Miocene Lower Meteora conglomerates (LMC) that illustrates the differential uplift of the TTS at that time. This motion is responsible for the partition of Lower Meteora Conglomerates (LMC, Pentalofon Formation) on the West side from the Upper Meteora Conglomerates (UMC, Tsotyli Formation) on the East side of the TTS [Ferrière et al., 2011].

Origin of the TTS

The TTS developed only in the southern MHB in the front of a major indentation of the Pelagonian basement, the Pelagonian Indentor (PI, Fig. 13A). Both of these major Hellenic structures, the PI and the TTS, has therefore to be genetically linked [Ferrière et al. 2004, 2011]. The Pelagonian Indentor corresponds to a raised Pelagonian block transverse to the MHB that is bounded to the east by the Aegean fault (between

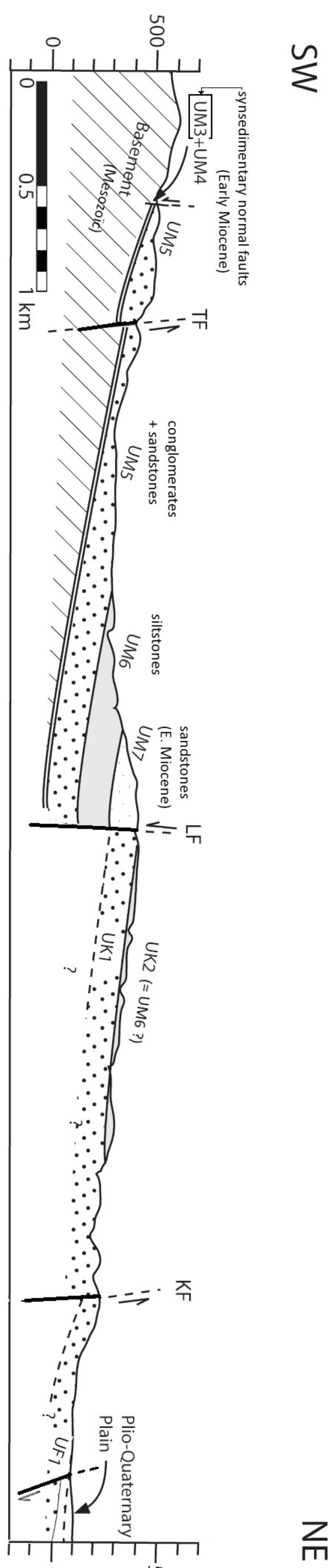


Figure 15
Cross-section in the Mitropoli area (southwestern MHB). The detritical series is only of Early Miocene age. Synsedimentary normal faults are present at the base of the series. They are responsible for the development of the MHB in this area.

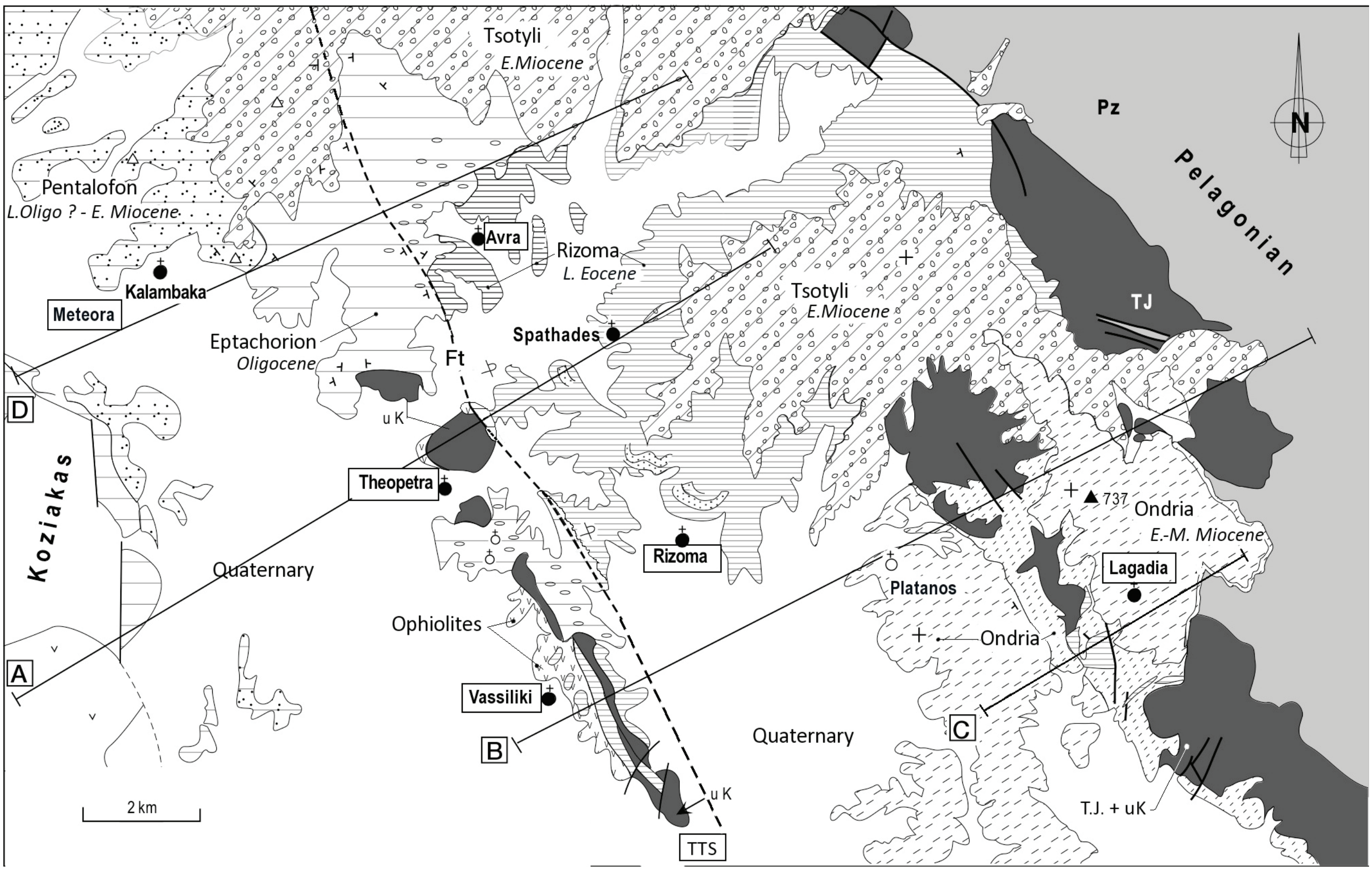


Figure 16A

Geologic map of the southeastern area of the MHB (Meteora) (modified after Ferriere et al., 2004). Abbr. Pz:Paleozoic gneisses and schists (Pz); Tj: Triassic-Jurassic marbles; uK: upper Cretaceous limestones of the Theopetra anticline (At) and South of Lagadia; L. Eocene: Late Eocene ; E.-M. Miocene: Early or/and Middle Miocene.

See Fig. 4 for cross-section D and Fig.16B for cross-sections A, B and C. Rizoma Formation: Late Lutetian-late Eocene clastic limestones overlaid by late Eocene unifites or deltaic sandstones (Rizoma sub-basin). Eptachorion Formation: Oligocene conglomerates (ellipses), sandstones and marls. Pentalofos Formation: Lower Meteora Conglomerates. Tsotyli Formation: gneiss-rich Upper Meteora Conglomerates. Ondria Formation: Early-Mid Miocene Echinid-rich limestones (small oblique dots), sandstones (turbidites) and Globigerinidae marls (large oblique dots). TTS: Theopetra-Theotokos Structure.

Olympos and Thermaikos Gulf) and to the west by the TTS. The northern boundary of the Indentor is the NE-SW Kozani-Krania saddle (Kozani straight from Brunn {1956}; Kozani saddle from Aubouin [1959]), where some remnants of the Vermion nappe and Vourinos ophiolites are preserved (Fig. 1). The southern boundary of the PI is masked by the Larissa and Trikala plio-quaternary plains. As a consequence, the narrowest section of the MHB occurs in front of the PI.

The basal beds of late Eocene series of Rizoma subbasin are transgressive on old levels of the Pelagonian basement on the western boundary of the Indentor. This suggests that the PI was already elevated and highly eroded as early as the Late Eocene. The deformation induced by the development of the PI extends far outside the MHB. In the external zones, series preserved in Pindos zone, north of the PI, are younger (Eocene flysch) than those preserved to the south of the Kastaniotikos transverse structure, more or less in front of the PI (mainly Mesozoic sediments) (Fig. 13A).

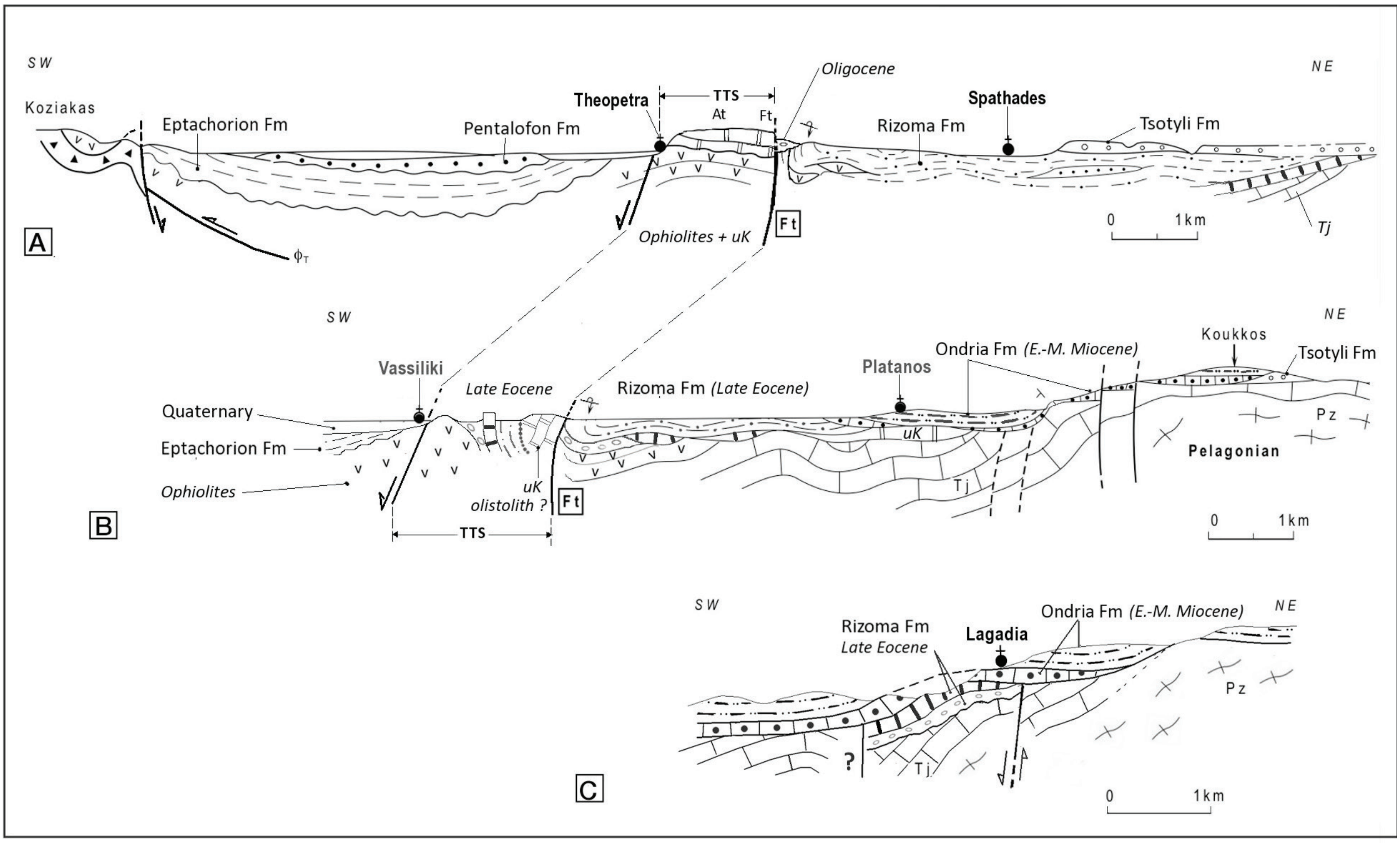
In the internal zones, the PI coincides with the location of the Olympus and Ossa tectonic windows (Fig.13A). However, at least an important part of the uplift of these Olympus-Ossa ranges occurred more recently than late Eocene - Oligocene times [Godfriaux, 1968; Fleury and Godfriaux, 1975; Schermer et al., 1990; Schermer, 1993; Lacassin et al., 2007; Migiros et al., 2011].

The pre-Late Eocene elevation of the Pelagonian Indentor could have been linked to specific paleogeography and/or specific tectonic characters in that area, notably in relation with earlier structural development of the internal zones (Early to Middle Eocene).

Putting these results together, one may hypothesize that the transverse high extending from the PI to the Olympos was created by the eastward underthrusting beneath the Pelagonian continental crust at the end of Eocene times of a thick crustal body subducted from the west of the Pindos basin. This crustal body most likely corresponds to an heterogeneity of the Gavrovo- Tripolitsa Zone [Ferrière et al., 1998; 2004].

Tranverse structures across the MHB

The main transverse structures that affect the MHB are WSW-ENE directed and they



are apparently associated with the observed longitudinal segmentation of the basin.

Krania sub-basin

The Late Eocene sediments of Krania sub-basin are the most deformed within the whole MHB. Important folding and reverse faulting are responsible for frequent sub-vertical bedding and the northern and southern boundaries of the sub-basin are transverse structures (Figs. 13 and 14).

To the north, it corresponds to the Monachiti-Trikomo Structure (MTS, Fig. 14). This structure was active at least during the late Eocene because it controlled the formation of a major unconformity between nearly horizontal Oligocene strata and the uppermost, locally subvertical, upper Eocene turbiditic sandstone strata of the Krania sub-basin [Ferrière et al., 2004]. The associated deformation caused the formation of a submarine slope, E-W trending and southward dipping, as shown by (i) olistostromes and olistolithic channels in the uppermost Eocene beds, (Pl. I-4) and (ii) the vergency of the conglomeratic alluvial fans in the lowermost Oligocene.

The orientation of this northern boundary of the Krania sub-basin and the SW-NE directed compressive deformation axis observed in the Krania sub-basin (stereonet no 3, Fig. 13B) suggest that the MTS could have developed as a WSW-ENE strike-slip fault zone as also suggested earlier [Papanikolaou et al., 1988]. The direction of a main decametric fold (near Trikomo), subparallel to the MTS, argue in favor of a large WSW-ENE directed flexure (Fig. 14) responsible for the NS steep depositional profile recorded in the upper series of the Eocene Krania sub-basin (from bay-fill to more southern basin floor-fan, via canyon-fills and large slumps (Pl. I-D and III-C).

To the South, the transverse border of the Krania sub-basin is less well documented, due to poorer outcrop quality. This southern boundary evokes a strike-slip fault zone that could have some reverse component [e.g. Panaya map; Koumantakis and Mataraga, 1980].

The Krania transverse borders are probably controlled by larger-scale transverse inherited structures, which define EW trending highs and lows of the Pelagonian substratum: the Pelagonian Indentor and the Kozani Saddle [Ferrière et al. , 2004] (Fig. 13A).

Figure 16B

Cross-sections (A to C) of the MHB southern area (see Fig.16A for location). Abbr.: Tj, uK, E.-M. Miocene, Fm, see Fig 16A. At and Ft: Theopetra Anticline and Fault

Late Eocene Rizoma Formation: i) Conglomerates (circles) and limestones (squares) at the base of Rizoma sub-basin and ii) Upper Eocene unifites and deltaic sandstones of the Rizoma sub-basin (dotted lines and points) ; uK: upper Cretaceous limestones basement or olistolites in late Eocene deposits. Oligocene Eptachorion Fm: conglomerates, sandstones and marls. Pentalofon Fm: Lower Meteora Conglomerates. Tsotyli Fm: Upper Meteora conglomerates. Miocene Ondria Fm: with Echinid-rich limestones (lower beds) sandstones and Globigerinidae marls (upper beds). TTS: Theopetra-Theotokos Structure. Vertical and horizontal scales are similar (no vertical exaggeration: v.e.=1).

Other transverse structures

Some minor transverse faults appear within the large structural blocks (more than 10 km long) defined by the major transverse structures. Some of these second order transverse faults can be observed clearly across the TTS, notably south of Vassiliki (Fig. 16A). These steeply dipping faults have some vertical motion but also some significant horizontal displacement as attested by bedding offsets and subhorizontal slickenslides. These strike-slip faults are compatible in orientation with the NE-SW direction of compressional deformation and of the NE-SW direction of extensional deformation as deduced from analysis of other faults (see Fig. 13B).

Discussion on along strike lateral displacements

Longitudinal basin-scale strike-slip faults (NNW-SSE) have been proposed by Zelilidis et al. [2002], based on some microtectonic data [Doutsos et al., 1994] and interpretations of a few seismic lines. Still, these authors do not provide any precision about the importance of the displacements implied nor about their chronology.

South of Grevena (Fig. 8), some large strike-slip faults at the western basin border have been proposed [Zelilidis et al., 2002], but they could not be evidenced by field analysis. The contact between Eptachorion and Pentalofon Formations has been considered as a major strike-slip fault (Theotokos Fault in Zelilidis et al., 2002, Fig. 8), but it is clearly stratigraphic at localities it could be observed.

More recently, the MHB has been also considered as a pull-apart basin that developed along a dextral shear zone during Oligocene and early Miocene times [Vamvaka et al., 2006]. Some large dextral strike-slip faults, NW-SE to NNW-SSE directed, would therefore correspond to the borders of the MHB at that time. One of the possible large faults that would control the development of such a large pull-apart basin corresponds to the NNW-SSE directed fault bounding the TTS to the East. Kinematic indicators from that fault zone indicate mostly some eastward verging reverse movement [Ferrière et al., 2004] and no significant strike-slip evidence could be observed on this Theopetra Fault zone.

Moreover, there is no major right-lateral offset of the main crustal structures (e.g. Kozani saddle- Krania syncline,Fig. 13A and Fig.20-A to F) on both sides of the TTS. This suggests that strike-slip displacements have been necessarily minor during the MHB development. In addition, the lack of en-echelon folds and of vertical axis folding in Oligocene to Miocene series supports the idea that if there was eventually some strike-slip motion parallel to the basin, it has to be necessarily very moderate and cannot be considered as a major process controlling the development of a large pull-apart basin.

Along the TTS, there are some dextral strike-slip faults trending about N340°to N010°E, near Theotokos and in the Meteora area, near Kalambaka (cf. Fig. 3 and Pl. III-G). These faults are compatible with the dextral motions proposed by Vamvaka et al. [2006]. However, these faults affect all strata from the Cretaceous limestones up to the Burdigalian (UMC, Tsotyli Formation) and are therefore to be considered as Burdigalian or younger in age, at least for their youngest activities. Such a post-MHB tectonic phase has been described and well-documented by Tranos et al. [2010]. This deformation could also be associated with the Neogene opening of the Karpero depression (Figs. 3 and 13B).

Tectonic development of the MHB

Tectonic structures

The main structures of the MHB are NNW-SSE to NW-SE directed, parallel to the main syncline axis of the basin. Transverse structures, including the Krania sub-basin, are mainly restricted to the northern border of the Pelagonian Indentor (Fig. 13A).

Most results from brittle deformation analysis are generally poorly constrained in age [Doutsos et al., 1994; Ferrière et al., 2004; Vamvaka et al., 2006]. The most significant results are reported on Figure 13B. Despite this difficulty to obtain precise ages for the different stages of deformation, some authors proposed successive tectonic episodes to account for the basin development [Ferrière et al., 2004; Vamvaka et al., 2006].

Brittle deformation is largely dominated by normal faulting, including some syn-sedimentary normal faults observed in the southern MHB (Mitropoli area, Pl. III-G). Reverse faults are less common and are mainly developed within the Eocene series of Krania sub-basin, but major reverse faults also exist in Early Miocene formations (e.g. near Mitropoli, Kanalia fault, Ditbanjong [2013]) or in the northern MHB (Pl.III-F). Strike-slip faulting occurs but these faults do not seem to play a major role in the Basin evolution.

The major compressional structures are faulted flexures, parallel to the basin axis, and were developed mainly on the western side of the MHB (Fk, Fe, and Ft, Fig. 13A). Longitudinal segmentation of the basin has to be related to the Pelagonian Indentor that started its development at least as early as the Late Eocene.

Chronology of deformation

First episode of compressional deformation (45-34 Ma)

Important compressional deformation occurs in the Krania area during late Eocene sedimentation. This episode is responsible for a clear angular unconformity within these series, with turbidites on top of olistostromes bearing up to hectometric-scale olistoliths of Late Cretaceous limestones. The compression continues up to the end of Eocene times and the Oligocene stata are unconformably overlying these deformations. In the Rizoma area, reverse motion of this age are observed on the main faults as the Ft fault (Fig. 13A).

Second tectonic period (34-15 Ma)

It corresponds to the main development of the MHB. During this period, the basin evolves progressively as a NW-SE directed asymmetric syncline with a steeply dipping western border. Tectonic deformation attributed to this period result essentially in vertical motions with borders uplift and basin subsidence gradually migrating eastward. This evolution is accompanied by normal faulting, minor reverse and strike-slip faulting, and km-scale folding. The main shift of subsidence is eastward, but there is also a southward jump in subsidence during the Miocene, identified by the presence of early Miocene series at the base of the western side of the basin in Mitropoli area (Fig. 1) and deposition of middle Miocene series of Ondria Formation unconformably over the Pelagonian basement (Fig. 3).

The post-MHB period (15 Ma - present-day)

This last stage of the MHB evolution is characterized by major uplift processes. This is evidenced by the high elevation of the most recent turbiditic formations of the MHB, up to 700m in elevation for Ondria Formation in the southern MHB and more than 1000m elevation for Ondria- Orlias Formations in the northern MHB.

The observed structures attributed to this period are steeply dipping faults and open folds that affected the Ondria Formation, notably close to Lagadia and Trikala (Fig. 16B).

North of Theotokos, the small Karpero sub-basin (Figs. 4 and 13B) contains Pliocene to Pleistocene continental deposits. The basin has a pull-apart shape and some of its boundaries are formed by normal faults (1:50000 Ayofillon map [Mavridis et al., 1979; Vamvaka et al., 2006]). This sub-basin is perpendicular to the main MHB direction and the brittle deformation indicates a NNW-SSE direction of extension [Vamvaka et al., 2006]. This event, with such a direction of extension, could also be responsible for eventual strike-slip reactivation of some of the former faults parallel to the MHB.

Two main tectonic events are generally considered after deposition in the MHB:

- (i) A compressional event of relatively local importance in the mid or late Miocene [Vamvaka et al., 2006, Tranos et al., 2010]. Folding of Tsotyli Formation, near Thetokos, and of Ondria Formation, near Lagadia, can be attributed to this age. According to the orientation of these southern structures, the axis of compressional deformation have to be directed NE-SW to ENE-WSW. In the north, Tranos et al. [2010] describe a NNE-SSW contraction with a transpressional-strike- slip regime.

- (ii) A widespread extensional deformation period with the development of numerous normal faults after the middle-Miocene, especially during Pliocene and Quaternary times (Pl. III-H). This extensional period can be correlate to the general evolution of the Aegean Plate that experienced major Plio-Quaternary extension [Mercier et al., 1989]. The extensional direction changes through time [Mercier et al., 1989;

Vamvaka et al., 2006; Tranos et al., 2010] with an overall evolution from NE-SW to approximately N-S.

DISCUSSION ON THE MHB EVOLUTION

Controls on sedimentation

Because sedimentation in the MHB is mainly marine and because it is contemporaneous with the Tertiary structural development of the Hellenides, tectonic deformation and eustatic sea-level variations have to be considered both as major forcing parameters on sedimentation.

Tectonic control on sedimentation

The existence of syntectonic deposits within this large piggyback basin has been described in previous sections. The tectonic control have an input in the basin sedimentation at various scales, including:

- (i) Flysch and olistolites deposition in small active sub-basins (e.g. late Eocene series in Krania area) at the onset of MHB development;

- (ii) Conglomeratic deposits in association with localized active faults (e.g. synsedimentary listric faults in Mitropoli area);

- (iii) Conglomeratic beds that developed on top of some significant angular unconformities in the basin, such as at the base of Krania, Rizoma, and Eptachorion Formations;

- (iv) The syntectonic fannig system of the Lower Meteora Conglomerates, representative of a major uplift of the source area (Pelagonian basement);

- (v) The eastward migration of the subsidence areas during the early Miocene (Pentalofon to Tsotyli Formations), associated with the tectonic uplift of a structural high (TTS) in the southern MHB.

Eustatism control on sedimentation

The MHB is filled by mainly marine sediments that were deposited at relatively shallow water- depth. Therefore the eustatic sea-level variations have to be considered as a significant driving parameter on sedimentation. This has been outlined by most sedimentological studies [i.e. Ori and Roveri, 1987; Zelilidis et al., 2002].

The high frequency rythmic sequences could be linked to some eustatic Milankovich cycles, notably within the Meteora conglomerates. However, it remains very uncertain to relate the lower frequency cycles to large eustatic events on the basis of correlations with the published eustatic charts [e.g. Haq et al. 1987; Abreu et al., 1998] because of the lack of precise datations in some formations such as in Meteora Conglomerates. For instance, the transition from Eptachorion marls to Pentalofon conglomerates (Lower Meteora Conglomerates) is clearly at least controlled by some tectonic activity [Ferrière et al., 2011] but it could be also enhanced by the large drop of sea-level proposed for the Late Oligocene [Haq et al., 1987]. However the amplitude of this sea-level drop is uncertain and some authors proposed a much smaller drop at that time [Abreu et al., 1998]. Moreover the age determination of this boundary is still questionable, being within the nannofossils biozone NP25 or Aquitanian, according to authors (Fig. 17).

Discussion on geodynamic interpretations

The first published interpretations on the geodynamic evolution of the MHB are relatively recent (Fig. 18). Firstly, the MHB has been considered as a retro-arc basin [Doutsos et al., 1994] related to the existence of eastward verging structures in the basin (Fig. 18A). However, these tectonic movements are mainly restricted to the late Eocene whereas the main elongated MHB is essentially Oligocene to Early Miocene in age. Moreover, the eastward verging reverse faults are not very developed and they cannot account for the observed eastward shift of the depocenter.

Alternative models proposed a significant strike-slip component in the development of the MHB [Zelilidis et al., 2002; Vamvaka et al., 2006]. On the basis of the presence of numerous normal faults, few angular unconformities in Oligocene and Miocene series and of interpretation of flower structures from some of the seismic

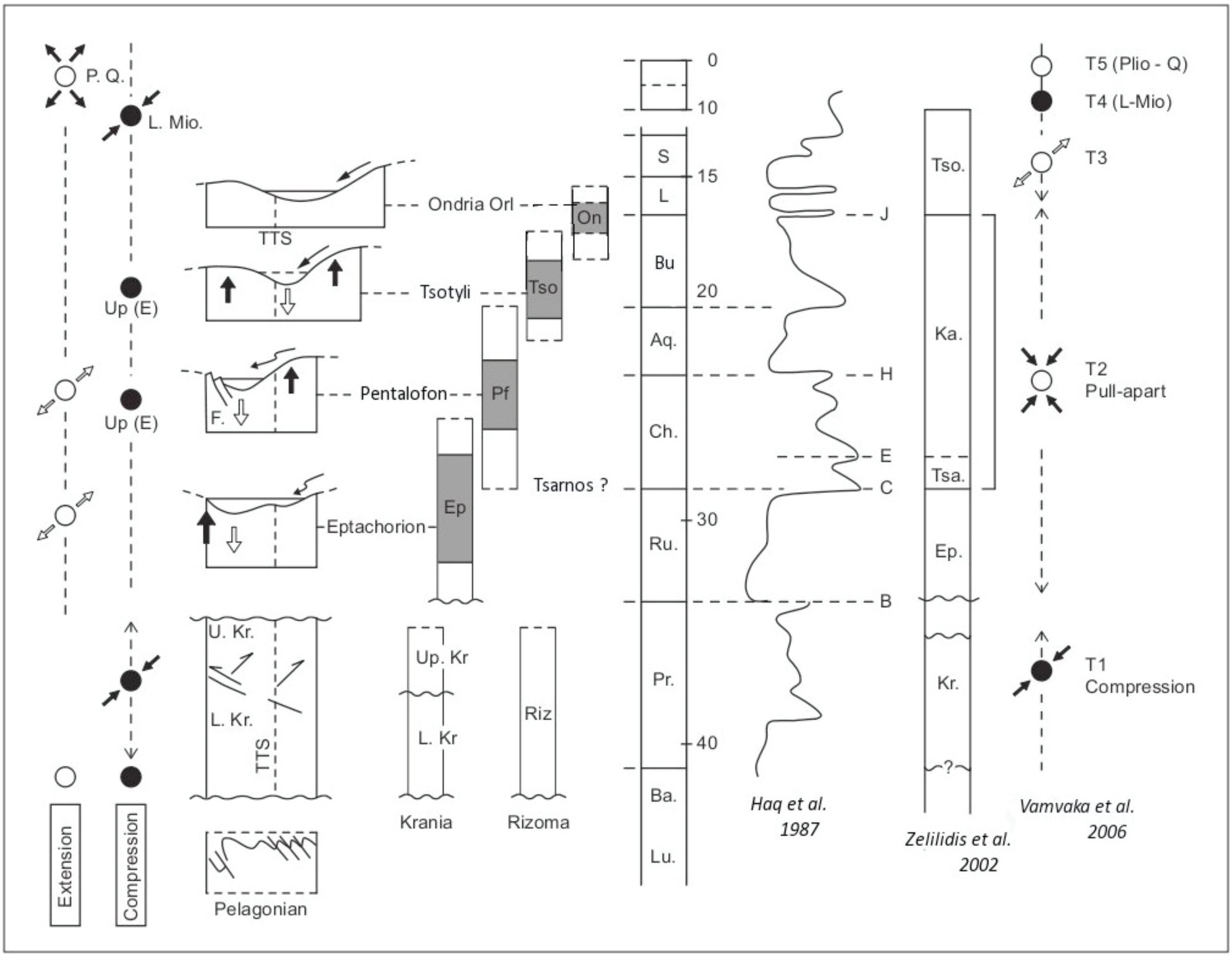


Figure 17

Eustasy vs tectonic controls on MHB evoluion.

Tectonic control. On the right, tectonic evolution after Vamvaka et al. (2006) assuming an important strike-slip event during Oligocene-Early Miocene. On the left, tectonic evolution after Ferriere et al. (1998, 2004 and 2011) considering a control by the underthrusted units under the MHB: Subduction of the Pindos basin then of the Gavrovo-Tripolitsa thick crust responsible for compressive structures followed by the migration of the main subsidence depocenters and uplifts with development of normal faults and some compressive tectonic structures.

Eustasy. Eustatic curves (from Haq et al. 1987).The eustatic control is difficult to proove as the ages of the lithologic Formations are not accurate enough to be compared with the eustatic charts (see also Fig.12). For us, the eustatic control on the MHB shallow marine sediments is necessarily efficient but with less control than the tectonic one. See text for more discussion.

lines, Zelilidis and co-authors [2002] considered the MHB as a strike-slip hemi-graben (Fig. 18B). For other authors [Vamvaka et al., 2006], the MHB is controlled by the subducting slab as described by Ferrière et al. [2004] but also behave as a pull-apart basin during the Oligocene to Early Miocene times (Fig. 18D).

We proposed earlier [Ferrière et al. 1998 ; 2004] an evolution controlled directly by the subduction of the external zones below the internal zones (Fig. 18C): firstly a forearc type of basin (Eocene), then a large crustal-scale piggy-back basin (Fig. 18C). This interpretation is based on: (i) the modifications of the basin geometry and depocenter locations that are driven by some paroxysmal tectonic episodes (e.g. near the Eocene-Oligocene boundary), and (ii) the position of the MHB domain on top of the Pelagonian basement during the downgoing motion of external zones below these internal zones. The displacement of the external zones below the internal zones is evidenced by the existence of the large tectonic windows (Ossa and Olympus windows) exposing the external zones in the inner Pelagonian zone [Godfriaux, 1968].

During the Lutetian to Late Eocene times, the Krania and Rizoma sub-basins are contemporaneous with the westward subduction of the thin (oceanic?) crust of the Pindos Basin. Possibly because of thin crust subduction, it is the upper unit and its irregularities (i.e. Pelagonian Indentor) that control the geometry of Eocene depocenters. Later on, since the basal Oligocene, the subducting crust is thicker and lighter: the continental crust bearing the Gavrovo-Tripolitza Platform. The arrival in subduction of this crust is interpreted as the driving mechanism of the brutal change in basin geometry, of major subsidence in the MHB axis and migration of the depocenter (cf. next section).

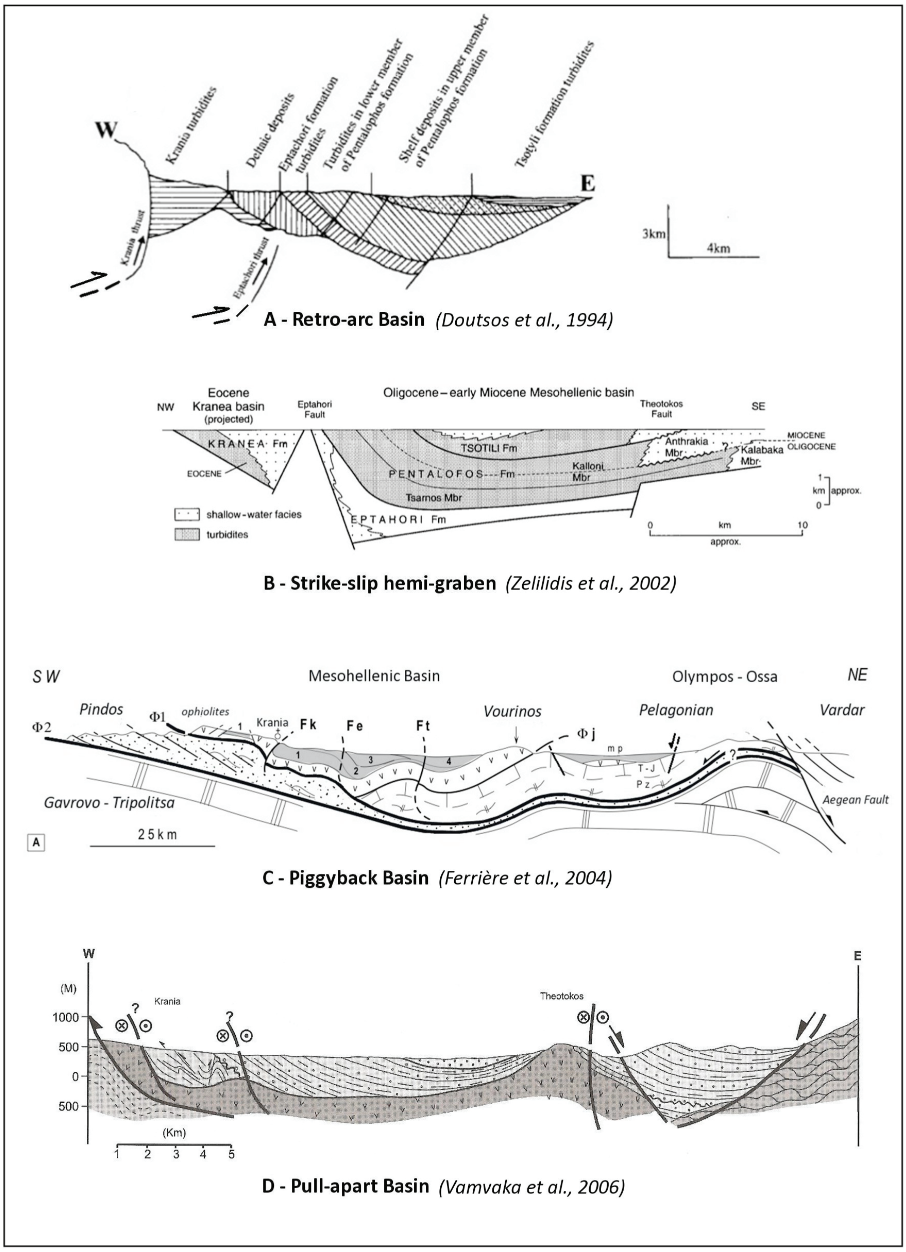


Figure 18
Different geodynamic interpretations. A: Doutsos et al., 1994: retro-arc basin in front of thrusts (Krania, Eptachori thrusts) verging to the East, linked to the main thrust of internal zones on external ones verging to the West (thrusts modified at their base, taking into account the Doutsos et al., interpretative figures). B: Zelilidis et al., 2002: the MHB is supposed to be a graben bounded by normal faults with a moderate strike-slip movement. C: Ferriere et al., 2004: the evolution of the MHB is linked to its piggyback character with different successive tectonic structures. D: Vamvaka et al., 2006: the MHB is mainly a pull-apart basin bounded by major strike-slip faults developed during Oligocene-Early Miocene times.

Discussion on mechanism of basin development

The driving mechanisms we consider for the tectonic development of the MHB are related to the subduction processes of the Hellenides external zones below the Pelagonian basement (Fig. 19).

The whole MHB is developing on the eastern side of the Pindos Fold and Thrust Belt [e.g. Skourlis and Doutsos, 2003] that can be regarded as the equivalent to an accretionary prism, notably in the early stages of its development. Comparatively with other accretionary complex, the MHB domain has therefore to be regarded as a forearc

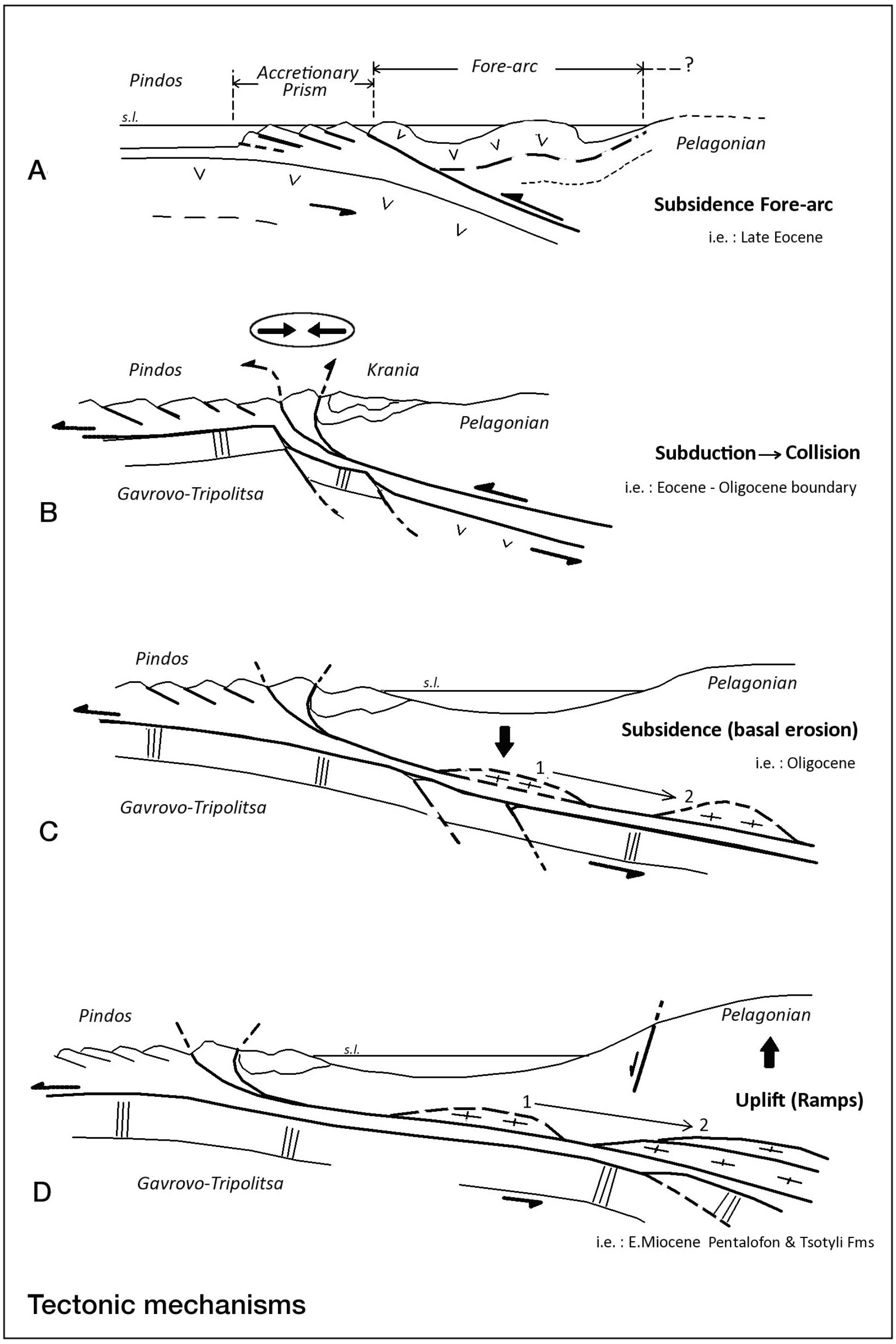


Figure 19
Main tectonic mechanisms supposed to have controlled the piggyback MHB evolution.

A: Subsidence linked to subduction: the early subsidence areas are linked to the subduction of the thin crust of the Pindos basin. The Krania and Rizoma small basins are located behind the progressively uplifted Pindos accretionary prism, in the fore-arc domain. B: Compressional tectonics: the main compressive tectonic structures (especially in the late Eocene) are linked to the arrival of a thick crust under the basins (transition between Subduction and Collision events). C: Subsidence during the piggyback stage: despite the underthrusting of a thick crust (i.e. Gavrovo crust), subsidence is active and gives rise to the main piggy-back basin. The subsidence areas could be linked to basal tectonic erosion: a tectonic slice of Pelagonian basement is pushed to the East along the basal thrust, so that the Pelagonian basement is thinned under the basin. D: Uplift during the piggyback stage: the uplifts (e.g.:uplift of the MHB eastern area) are linked to the stacking of the tectonic sheets (Pelagonian slices and possibly Gavrovo ones) linked to tectonic events along the basal thrust.

domain, possibly comparable with for instance the Hikurangi subduction margin in New Zealand, also characterized by the subduction of relatively thick crust [e.g. Nicol et al., 2007]. Such forearc domains are classically located between the highest ridge of the accretionary prism and the volcanic arc. In the present setting of the MHB, the Eocene volcanic arc is not clearly identified. Some late Eocene and Early Oligocene calco-alcaline rocks are exposed in Turkey [Dilek et al., 2009] and in northeastern Greece, but they could be linked to another subducted oceanic area [Intra-Pontide Ocean, Pe-Piper and Piper, 2006; 2007].

Accordingly, the initial subsidence of the MHB domain can be attributed to forearc setting, on top of pre-existing dense obducted ophiolites, during the subduction of Pindos thick oceanic basin (Fig. 19A).

The location and geometry of the main MHB basin (Oligocene-Early Miocene) is controlled by its piggy-back behavior. Low angle subduction of the lighter and thicker crust of the Gavrovo- Tripolitza Zone led to the uplift of the Pindos accretionary wedge. This major uplift corresponds to the transition from oceanic subduction to collision (or continental subduction) of the external zones (Fig. 19B). This process is coeval with the development of an elongated sudsiding area in the inner domain, on the back of the highest ridge of the uplifting accretionary wedge: the Oligocene to early Miocene MHB. Because of the eastward motion of the underlying units (Gavrovo-Tripolitza Zone) coeval with sedimentation, this area of subsidence corresponds at that time to a large piggy-back basin (Fig. 19C).

In addition with this process of individualization of a large elongated basin parallel to the Pindos wedge, the subsidence can be driven by basal tectonic erosion (Fig. 19C). Such tectonic erosion is not only an important process for creating subsidence areas in convergent settings, but it is also adapted to account for dominant normal faulting during basin sedimentation (e.g. Chanier et al., 1991; 1999).

The crustal Pelagonian slivers pulled out by tectonic erosion can accumulate farther east below the Pelagonian crust and therefore account for the uplift of the internal zones coeval with the MHB development (Fig.19D). A succession of superimposed crustal duplexes, from Pelagonian material, and then of Gravrovo-Tripolitza material, could be responsible for the initial uplift of the inner Pelagonian zone, leading to the first steps of the progressive exhumation of Olympus and Ossa tectonic windows. Even if the last stages of recent exhumation of these windows could be linked to the Plio-Quaternary general extension within the Hellenic region, this mechanism easily explains the observed structural geometry inside the Olympus window.

As exposed on previous sections, some horizontal striations occur, notably on some of the main faults. We do not believe that the total strike-slip displacement along the main faults was of major importance because no large lateral offsets could be evidenced. Such strike-slip displacements during convergence could be associated with some strain partitioning such as described on most oblique convergent settings [e.g. Cashman et al., 1992; Nicol et al., 2007].

EVOLUTION OF THE MHB AND SUCCESSIVE GEODYNAMIC SETTINGS

This synthesis (Figs. 20 and 21) is mainly based on our work about the MHB but also considers the results obtained from other studies in the basin [e.g. Brunn, 1956; Desprairies, 1979; Ori and Roveri, 1987; Barbieri, 1992; Doutsos et al., 1994; Kontopoulos et al., 1999; Zelilidis et al., 2002; Vamvaka et al., 2006].

Pre-MHB setting

Before the first MHB transgressive deposits, late Lutetian in age, the pre-MHB domain belongs to the internal zones that have been intensively deformed after the deposition of Paleocene flysch [e.g. Celet and Ferrière, 1976]. This major tectonic episode of the Hellenides is characterized by mainly southwest-verging folding and thrusting. The structural initiation of the Pelagonian Indentor, that is necessarly prior to MHB development, can be attributed to this event.

In the inner domains of the internal zones, this Tertiary episode of deformation is also characterized by intense metamorphism. In the outer domain, there is no significant tertiary metamorphism as the Late Cretaceous limestones are not metamorphosed. The important metamorphism of the Pelagonian zone in this area is late Jurassic in age, associated with the emplacement of ophiolitic nappes during obduction [e.g. Ferrière et al. 2012].

Lutetian to Late Eocene small confined basins

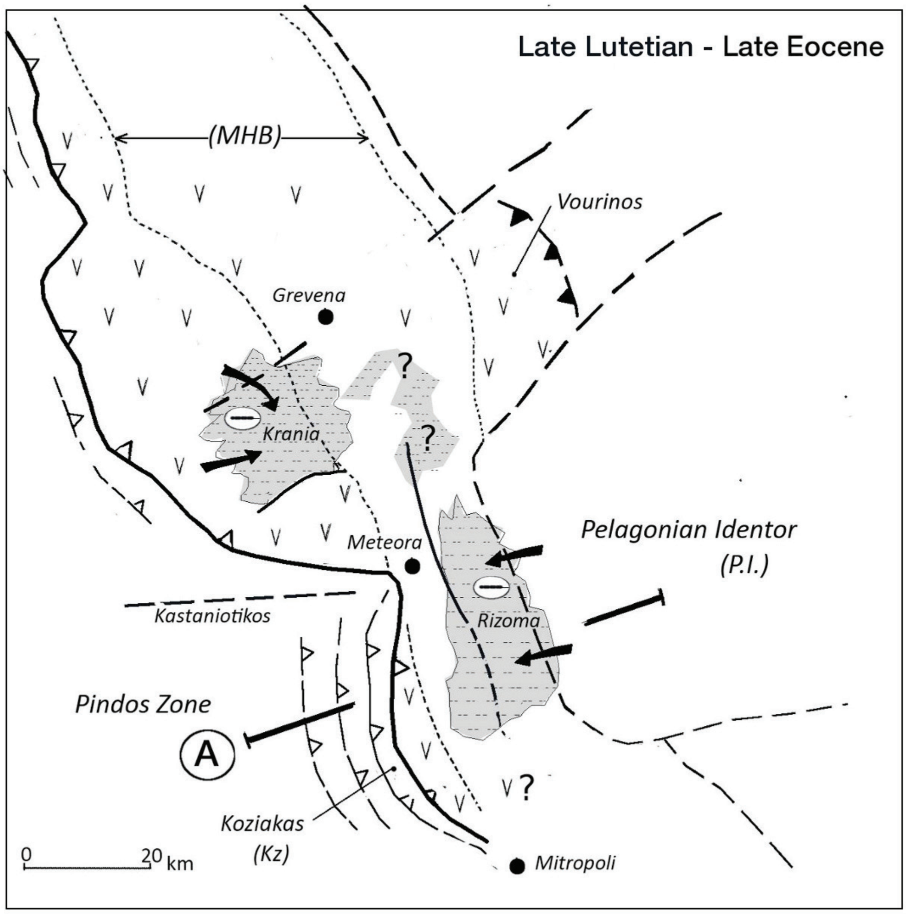
The first sedimentation areas in the MHB domain are restricted to two small confined basins: The Rizoma sub-basin resting on Pelagonian basement and Krania sub-basin resting on ophiolitic units (Fig. 20A).

The Krania sub-basin presents evidences for synsedimentary tectonic activity, such as internal angular unconformities, flysch facies with many olistostromes and large olistolites, and numerous slump events indicating instabilities on slopes. The northern boundary of the Krania sub-basin is clearly an active flexure that controls the sedimentation within the area. There is no evidence for an eventual connection between the two sedimentation areas of Krania and Rizoma, although it cannot be totally excluded.

These small basins are contemporary with the late Eocene development of the large thrust driving the internal zones over the external zones, that is to say the Pindos

Zone. This is confirmed by the younger flysch deposits in the Pindos Zone that are Lutetian in age, even up to Late Eocene [Desprairies, 1979].

Therefore, the sedimentation in Krania and Rizoma sub-basins takes place during the onset of convergence and related Pindos subduction, responsible for the development of an accretionary prism below the Pelagonian upper unit that has been covered by ophiolitic nappes during Jurassic obduction [e.g. Ferrière et al., 2012]. Because it originated at the back of a rising accretionary prism, the deep sub-basin of Krania and the sub-basin of Rizoma may be considered as forearc basins on an active margin.



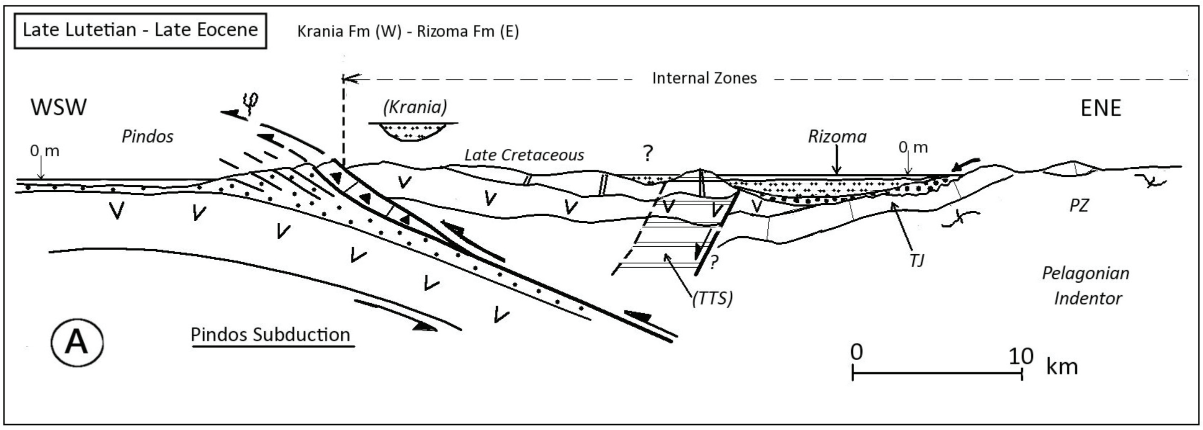


Figure 20

Detail of the MHB evolution from Pindos Subduction (late Eocene) to Collision (Oligocene – middle Miocene).

Six geologic stages are represented with, for each of them, a map of the MHB and a cross-section in the southern part of this basin. Vertical and horizontal scales are similar. See text for more explanations. Schematic maps and cross-sections illustrate the successive stages of evolution of the MHB in the framework of the Hellenic Tertiary subduction-collision of external zones under the internal ones.

Maps. Oblique black arrows: direction of sedimentary flows; + = uplift areas; - = subsiding areas. Cross-sections. Vertical black arrows: areas of major uplift and of main subsidence ; white arrows: direction of sedimentary flows.

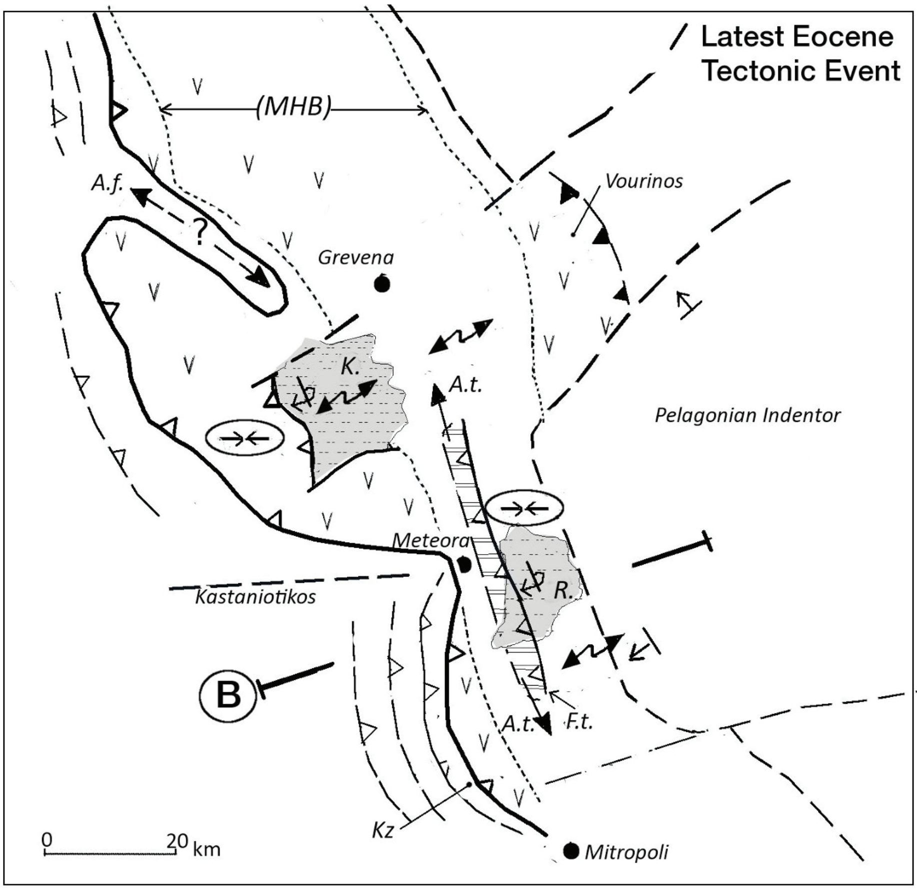
Figure 20A

Late Lutetian-late Eocene. The Krania-Rizoma subsiding areas are linked to the Pindos subduction. They developed in the fore-arc domain just on the east of the Pindos accretionary prism. The location of these two basins is linked to the Pelagonian Indentor structure. Sediments are rich in elements coming from the close areas (ophiolites for the Krania basin, Paleozoic basement for the Rizoma basin).

The pattern of subsidence, at that time, should be driven by mechanical parameters of the subducting crust (thickness, temperature, angle of the subducting slab). The boundaries of Krania and Rizoma sub-basins are controlled by heterogeneities of the upper crustal unit such as flexures at the borders of the Pelagonian Indentor (Rizoma basin on the western side and Kozani- Krania saddle on the NW side of this Indentor) (Fig. 20A).

Latest Eocene compressional episode

The Eocene-Oligocene transition is outlined by a major angular unconformity, particularly important nearby the large tectonic structures (e.g. Monachiti area, Fig. 14; Ft in Theopetra area, Fig. 16A).



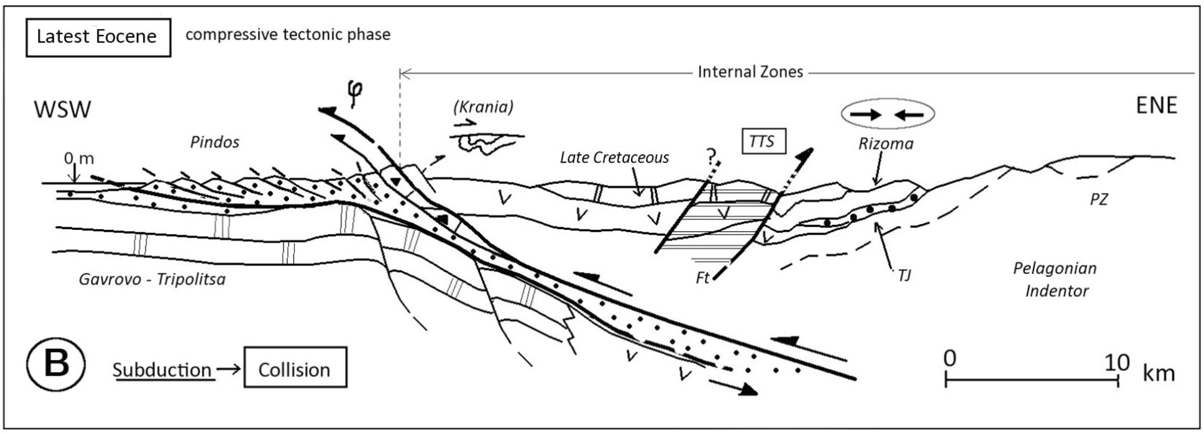
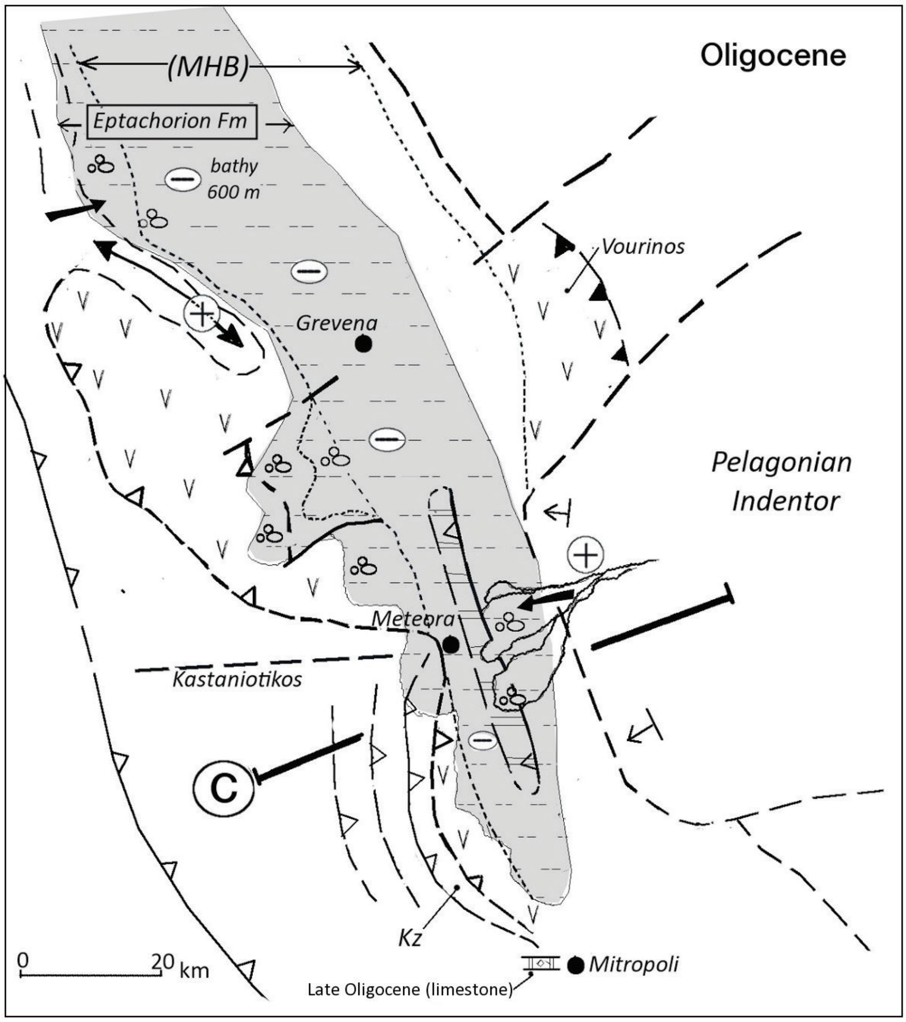


Figure 20B
Latest Eocene (till boundary with Oligocene). Development of the main compressive tectonic structures linked, for the authors, to the arrival of the thick Gavrovo-Tripolitsa crust in the subduction zone. The western boundaries of Krania and Rizoma outcrops are active (faulted flexure and reverse faults). The northern boundary of the Krania basin is also active.

This is a period of significant compressional deformation attested by the development of eastward verging reverse faults, and of folds associated locally with slight cleavage (e.g. in Krania-Mylia and Theopetra-Avra areas). It corresponds to the end of the late Eocene tectonic phase.

We consider that this major compressional event is the consequence of the arrival in the subduction zone of the Gavrovo-Tripolitsa platform unit (Fig. 20B). This is consistent with the continuing rise of the Pindos accretionary prism, the exhumation of the main flysch tectonic windows below the ophiolitic nappes (Filippi Anticline, Af, Fig. 20B and Fig. 20C), and the emersion of the Krania and Rizoma sub-basins as evidenced by the occurrence of reddish continental conglomerates at the base of Oligocene series on top of the unconformity.



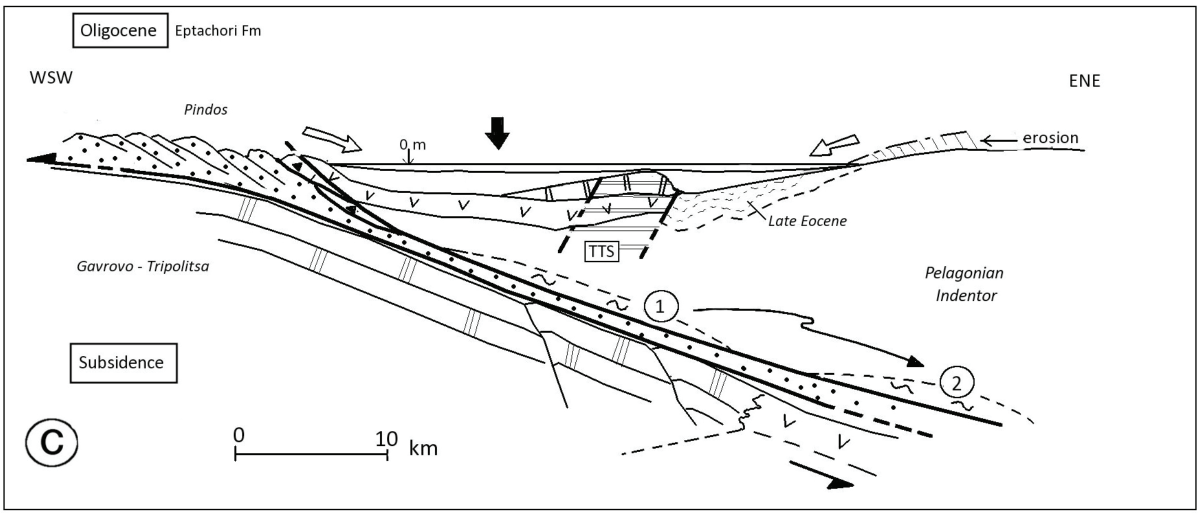


Figure 20C

Oligocene (Eptachorion Formation). The elongated wellknown MHB is initiated at this time (Greece and Albania). The development of subsidence while a relatively thick crust (Gavrovo-Tripolitsa zone) is underthrusted has to be explain. We consider that a basal tectonic erosion (numbers 1 to 2) is the mechanism at the origin of the subsidence (see also Fig. 19).

The collision is therefore marked in the overriding plate, even though the Gavrovo-Tripolitza crust may have been thinner than a normal continental crust, as suggested by its continuous subsidence during the Mesozoic and early Paleogene at the margin of the Pindos basin [Aubouin, 1959].

From this observation, it appears that the time required for the subduction of the whole Pindos basin is about 10 Ma long (45-43 Ma to 35-33 Ma). The former total width of the Pindos basin, 300 to 600 km in the northern Hellenides, is reconstructed from: i) mapping of the isopic zones; ii) the amount of tectonic shortening estimated from outcrops and balanced cross-sections [Skourlis and Doutsos, 2003], and iii) the estimated extension/spreading rate of this basin made of thin continental crust [Thie-

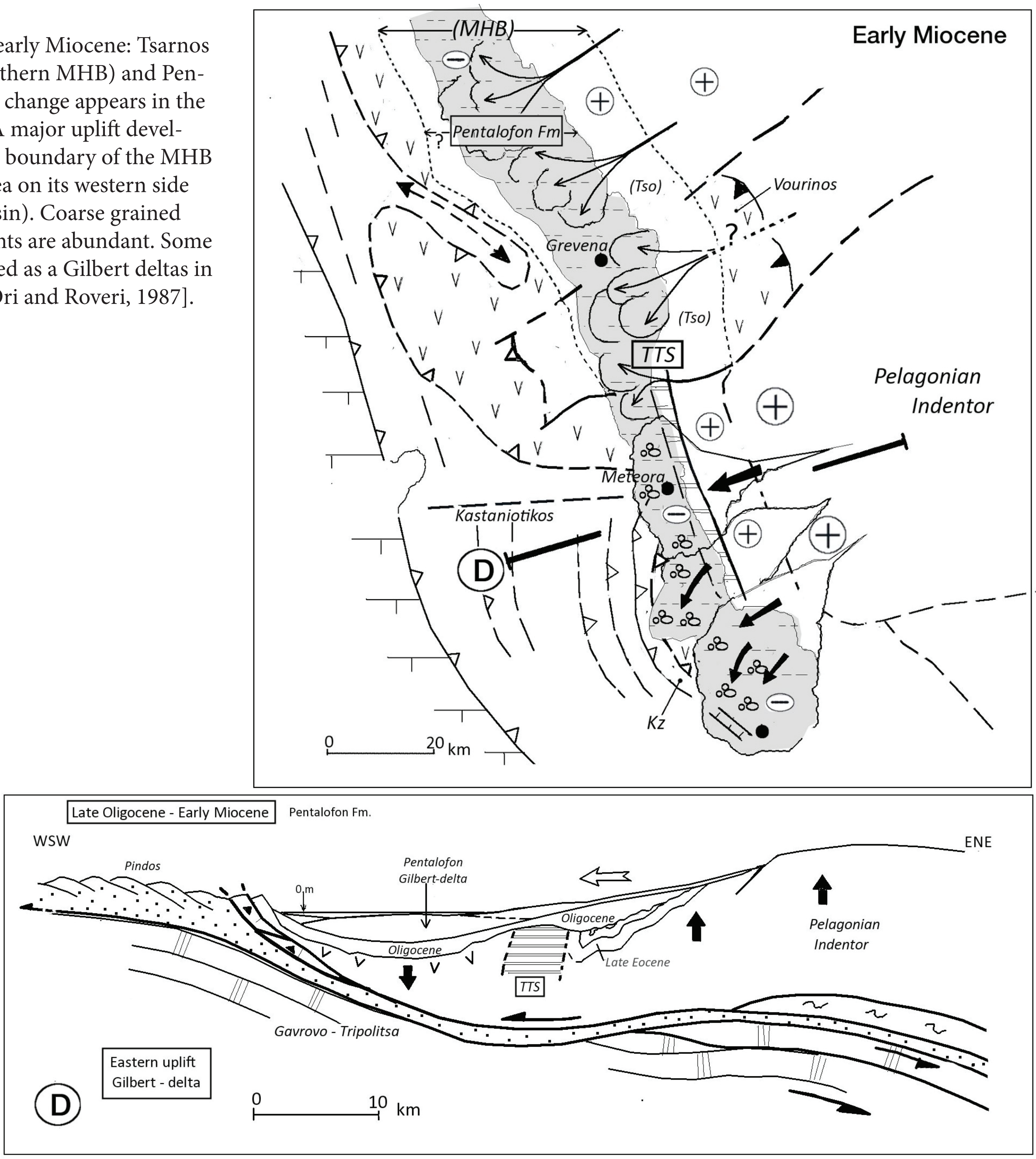


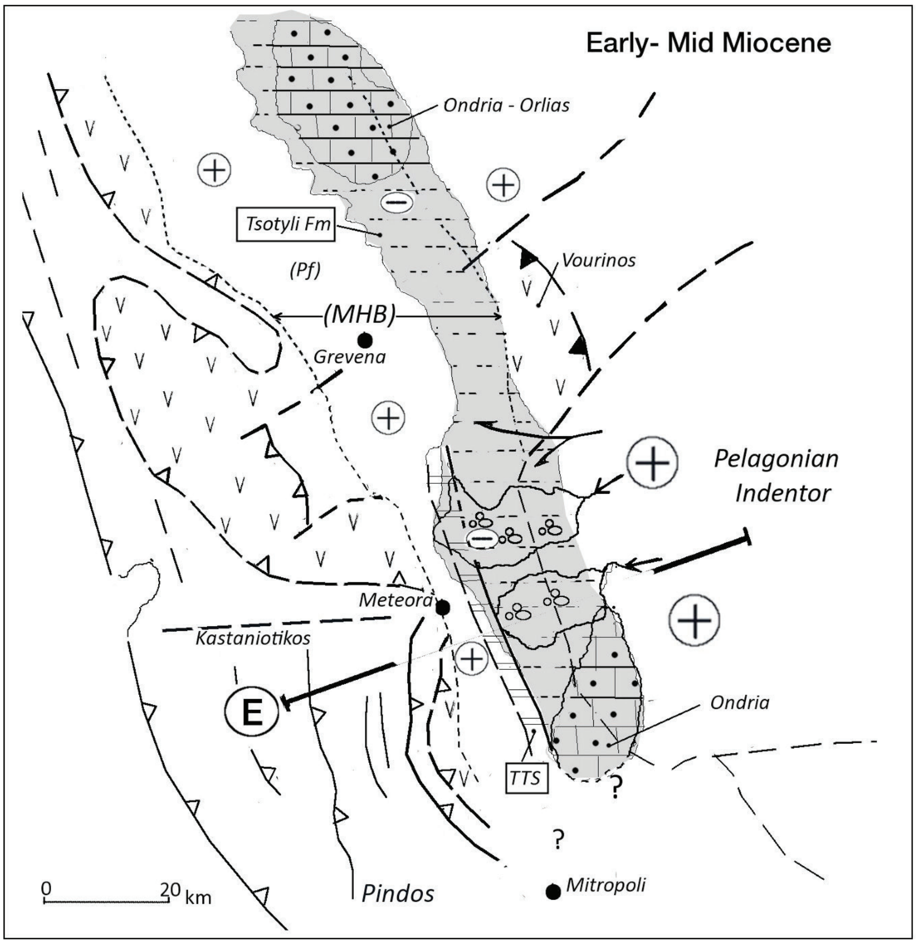
Figure 20D
Late Oligocene (?)-early Miocene: Tsarnos Fm (only in the northern MHB) and Pentalofon Fm. A main change appears in the MHB at this time. A major uplift developed on the eastern boundary of the MHB with a subsiding area on its western side (Pentalofon sub-basin). Coarse grained terrigenous sediments are abundant. Some of them are deposited as a Gilbert deltas in the Meteora area [Ori and Roveri, 1987].

bault, 1982] or oceanic crust [Bonneau, 1982]. These studies suggested an average Pindos Basin subduction rate of about 3 to 6 cm/year.

The first elongated major basin:the western MHB

Oligocene initial development of the large MHB

The basal Oligocene sediments are generally thick conglomerate beds resting unconformably on top of deformed basement, including late Eocene sub-basins. The Oligocene Eptachorion Formation is characterized by a very strong tectonic subsidence: the basal continental conglomerates are overlaid by turbidites and then by massive marls indicating an important deepening of the basin floor, up to 600m water depth according to Barbieri [1992]. Locally, subsidence is accompagnied by synsedimentary normal faults and by onlaps of relatively deep turbidites on top of subaerial to shallow-marine conglomerates or reefal limestones (e.g. north of Alatopetra) revealing a sharp steepening of depositional profiles. During that time, the Pindos Fold and Thrust Belt keep rising, as recorded by syntectonic fans in the lower part of turbiditic deposits in the western MHB (north of Alatopetra). Such a strong subsidence in the course of the major underthrusting of a thick crust has to be explained. As the MHB



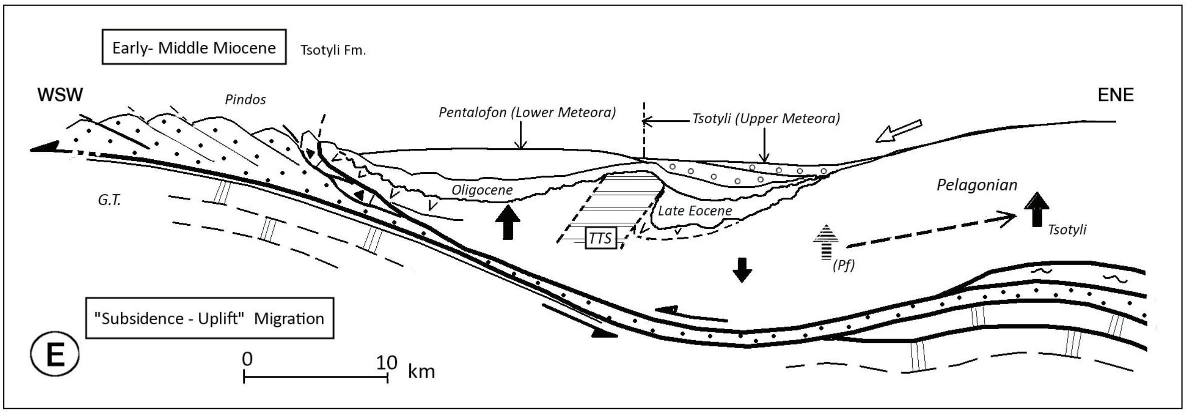


Figure 20E
Early-mid Miocene (Tsotyli and Ondria-Orlias Fms). The main event at this time is the migration toward the east of the subsiding and uplifted areas. In the southern MHB, a structural high (TTS) works as a boundary between the old Pentalofon and the new Tsotyli sub-basins. The Ondria-Orlias Fms are the last known marine sediments in the MHB.

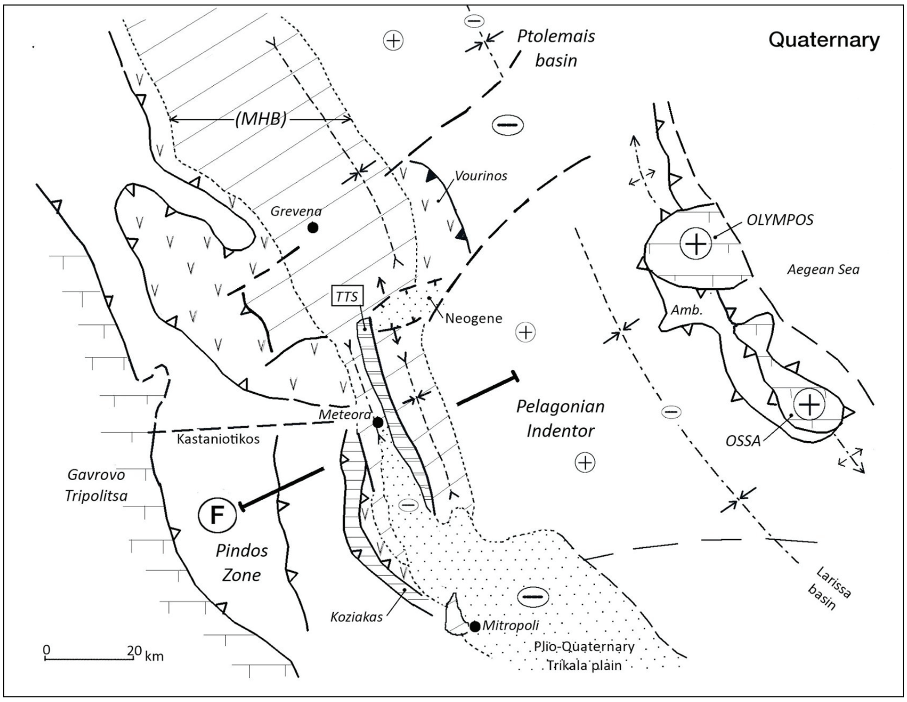
evolves into a major NW- SE marine trough, the process is necessarily at the scale of the whole margin.

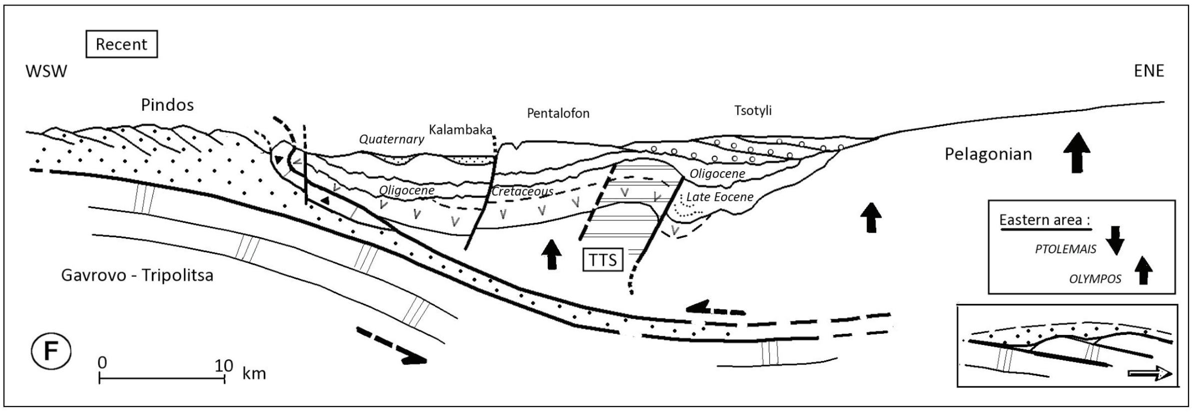
The presence of thick dense ophiolitic bodies west of the Pelagonian unit, forming the basement of the MHB from this area of Greece up to Albania [Robertson and Shallo, 2000], could partly control the location of the main subsiding areas. However, the NW-SE direction of the Oligo- Miocene MHB trough, parallel to the Hellenic frontal thrust and to the external zones, argues for a main control by the external underthrusted units. Crustal delamination at the base of the Pelagonian overriding unit (basal tectonic erosion) during the eastward underthrusting of Gavrovo-Tripolitza crust beneath the basin (Fig. 20C) has to be considered as the most probable model to acount for this major synsubduction subsidence during Oligocene times.

The first major sedimentary change in the main MHB: coarse grained deposits linked to an eastern uplift (Pentalofon Fm, Latest Oligocene – Early Miocene)

The stratigraphic transition from Eptachorion marls to Pentalofon coarse sediments,

Figure 20F
The Quaternary map of the MHB with the Olympos tectonic window on the east which argue in favor of the piggyback nature of the MHB. The MHB has been largely uplifted. The migration toward the east of the subsiding (Ptolemais basin) and uplifted (Olympos-Ossa) areas is the result of a Neogen-Quaternary tectonic activity.





notably the Lower Meteora Conglomerates, corresponds to a major event in the MHB evolution. Because of the lack of angular unconformity between these two formations, we propose that two main forcing factors have to be considered: eustatic sea-level variations and tectonically induced vertical motions. At that stage, the migration of depocenters toward the inner domains is not really effective.

The stratigraphic transition from turbidites and marls to coarse-grained deltas (e.g. Taliaros Formation or Lower Meteora Conglomerates) reflects an overall progradation trend during the late Oligocene to early Miocene times. Whereas conglomerates of Pentalofon Formation overlie Taliaros deltas in the northern MHB, they directly truncate Eptachorion marls in the south.

On one hand, the imprint of the Rupelian-Chattian eustatic sea-level fall [Haq et al., 1988] could explain this sharp Eptachorion-Pentalofon boundary. Nevertheless, this event seems less prominent on some of the most recent charts edited for the European basins [Abreu and Haddad, 1998] (Fig. 12). On the other hand, we have demonstrated that the conglomeratic fan-deltas, especially the Lower Meteora Conglomerates (Pentalofon Formation), that dominate deposition during the late Oligocene - early Miocene times are mainly due to tectonic activity [Ferrière et al., 2011]. This has been demonstrated by: i) the evolution of topsets geometry of the Meteora Gilbert-deltas that implies progradation over a steep and active basin slope; ii) the reactivation of existing tectonic structures at the time of Pentalofon Formation deposition (deposition of Meteora Gilbert-delta on the side of the Theopetra-Theotokos Structural-high, Pl. III-B); iii) the existence of syn-deposition Early Miocene normal faults on the south-western side of the MHB (Mitropoli area, no 6 on Fig. 13B ; Pl. III-E).

This basin stage, starting with Pentalofon Formation, was mainly controlled by uplift of the eastern border of the MHB which became the main drainage area (Fig. 20D). The high sedimentation rate associated with conglomeratic deposition led to rapid overfilling of the basin on the west side of the present TTS (Fig. 20D).

The second stage of large basin: the western MHB

The sedimentological transition from Pentalofon to Tsotyli Formation is sharp in the southern MHB but is not really marked in the northern part of the basin. However, the main characteristic of this transition at the whole basin scale is the significant eastward shift of the depocenter. This shift seems to correspond to a brief event. The two successive areas of sedimentation, Pentalofon to the West and Tsotyli to the East, are separated by the TTS in the southern MHB (Fig. 20E).

This 10 to 15 km brutal eastward shift of the subsidence area to the other side of the Theopetra- Theotokos structure (TTS) can be interpreted as an eastward displacement of the zone of tectonic erosion beneath the basin crust, coupled with an increasing uplift rate in the hinterland because of underplating. These eastward migrations of subsidence and uplifted areas give probably rise to major normal faults well expressed in the MHB (Fig. 4).

The final stage of MHB development: Ondria – Orlias Formations

The latest stratigraphic Formations are only preserved on both southern and northern tips of the MHB. Up North, they only rest on top of Tsotyli Formation whereas in the South, they are lying unconformably on various lithological units ranging from Tsotyli Formation to Mesozoic and Paleozoic basement (Pelagonian zone). The sedimentological trend is a significant increase in carbonate content and a general decrease of water-depth of deposition (shallow water Orlias Formation).

The present-day MHB

The end of deposition in the basin is linked to the general uplift of the MHB area (Fig. 20F) possibly emphasized by the eustatic sea-level drop proposed for the mid-late Miocene [Abreu and Haddad, 1998].

Flexuration between Rizoma and Lagadia (southeastern MHB, Fig. 16A) and folding east of Theotokos carry on during and after basin uplift, illustrate renewal of compressional tectonic activity at this eastern side of the basin. The uplift affects the whole MHB and its amplitude can be estimated to about 1000 m in the North and about 700 to 800m in the South. A Late Miocene compressional tectonic event is generally accepted [e.g. Vamvaka et al., 2006; Tranos et al., 2010] and happened before the well-known Plio-Quaternary extensional period in this part of the Hellenides [Aubouin, 1959; Mercier et al., 1989].

During that post-MHB period, the related subsidence and uplift axis migrated toward the East : Plio-Quaternary subsidence in Ptolemais Basin and uplift in Olympos and Ossa domain (Figs. 20F and 21).

CONCLUSIONS

1) Amongst large orogenic, intermontane syntectonic basins, the MHB is an important one, the development of which is at the transition between subduction (fore-arc stage) and collision (piggyback stage). It might be basically controlled by underthrusted units. Its large size (300 km in length), thickness (up to 4.5 km), duration (30 Ma, and 20 Ma for its piggyback stage) and dominantly marine siliciclastic infill are to be related to lithosphere-scale processes. These processes (underplating, tectonic erosion, thermal relaxation) can be quantified based on the stratigraphic record of the basin (paleobathymetry, subsidence) together with the thermochronology of the surrounding source areas (rock uplift). Ongoing numerical modeling using these data will allow reconstructing the evolution of relief of this part of the Hellenic chain.

The sedimentary record of the MHB is dominated by fluvially supplied gravity deposits (fan- deltas, slope and basin-floor fans), the most emblematic being the Meteora conglomeratic Gilbert deltas (Pentalofon Fm).

The "molassic" deposits are resting unconformably on the internal zones of the Hellenides, which were intensely tectonized during the lower to middle Eocene. These deposits are also syntectonic, as they record several extensional but also compressional

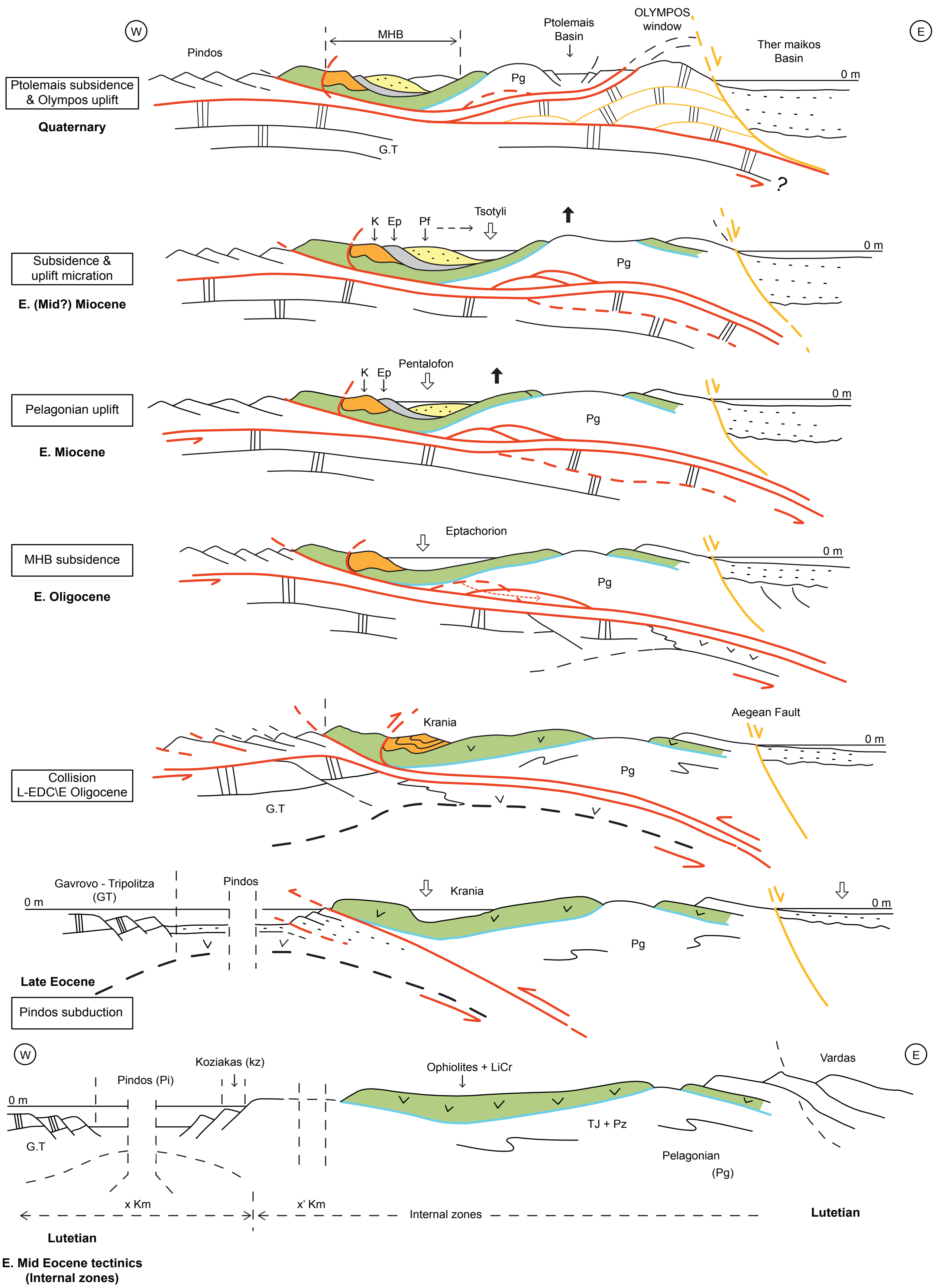


Figure 21

Main stages of MHB evolution in the context of the Tertiary geodynamic evolution of the domain between the Gavrovo-Pindos external zones and the eastern Pelagonian unit above the Olympos window. See text and Figures 19 for detailed discussion on mechanisms.

faulting stages (associated to strike-slip motions, the importance of which remains to be determined), as well as olistolites or sharp coarsening-upward shifts pointing to stages of rejuvenation of the border reliefs.

2) The deposits are organized in at least 5 transgressive-regressive supercycles commonly dominated by conglomerates at the base and sometimes at the top, and by finer-grained, turbiditic or pelagic facies during the maximum flooding. These supercycles correspond to the successive stages of basin tectonic evolution.

The first stage of the MHB corresponds to the first supercycle and dates to latest Eocene (ca 45-34 Ma). It is preserved in the isolated Krania and Rizoma sub-basins. The related syn-tectonic facies are contrasted. Close to the subduction front (Krania), they are dominated by turbidites recording slope instabilities (olistostomes, slumps, mass-wasting unconformities). Backwards, they point to shallower-water, deltaic systems (Rizoma).

The second stage of the MHB, which corresponds to four supercycles, dates to Oligo-Miocene (ca 34-15 Ma). The first super-cycles in this stage are well differenciated. They are, from base to top: the Eptachorion Fm (Oligocene), the Pentalofon Fm (Late Oligocene- Lower Miocene), the Tsotyli Fm (Lower Miocene). These super-cycles are followed by a last one, corresponding to the Ondria-Orlias Fms, only outcropping at the tips of the MHB. Within the super-cycles, smaller-scale sequences are preserved, likely controlled by eustatic sea-level changes. This could be for example the case for the wedges forming the Meteora fan-deltas.

3) These successive depositional stages record synsedimentary deformations from which stress fields can be determined, in order to relate the basin evolution to the changing geodynamic context.

Compression is recorded mostly in late Eocene, bringing about folds, reverse faults and larger- scale structures which we called "faulted-flexures".

Normal faults are also numerous through the MHB evolution, some being cleraly syn- sedimentary (e.g. Mitropoli), and other ones, less well dated, more likely formed during the generalized Plio-Quaternary extension.

Some authors, based on seismic profiles and horizontal slickensides on plane faults, stress on the role of major strike-slips in the Oligocene to Lower Miocene [i.e. Vamvaka et al., 2006], but the importance of these motions in the basin evolution are not established and thus may be overestimated.

These tectonic deformations control the basin geometry and rather the location of depocentres, for instance the dissymmetry of the MHB syncline (location of late Eocene sub-basins, steeper flank of the syncline to the west, narrower to the south in front of the Pelagonian Indentor) and the migration toward the east of the depocenters.

4) Contrasted interpretations have been published regarding to the geodynamic control of the MHB (Fig. 18). These different successive interpretations have been discussed above. The basin has been first interpreted as a retro-arc basin [Doutsos et al., 1994] but the eastward thrusts are of limited extent and the source areas shift from west to east at the end of Oligocene, then it has been interpreted as a strike-slip hemi-graben or pull-apart basin but no major strike-slip faults have been clearly prooved.

In our interpretation, the MHB is basically a piggyback basin, because sedimentation and accommodation take place during the underthrusting of tectonic units which now outcrop to the east of the basin (Olympos). The two main stages of basin tectonic evolution can be related to two geodynamic stages of the subduction of the external zones of the Hellenides beneath the internal zones:

- During the first stage (fore-arc setting), the maximum subsidence area is located close to the subduction front and the scattering of depocentres reflect heterogeneities of the upper plate. In our interpretation, it is because at this pre-collision time, there is little plate coupling and therefore little influence of the thin-crusted subducting plate (Pindos basin).

- During the second stage (piggyback setting), the maximum subsidence is generalized to the great Albano-Thessalian basin and moves stepply eastward. In our interpretation, this is the consequence of the less easy subduction (collision) of a thicker-crusted plate (the Gavrovo- Tripolitsa platform).

The processes controlling the vertical motion of the upper plate during collision are deep processes and so are relatively speculative. The underthrusting of the Gavrovo-Tripolitsa block would likely promote localized tectonic erosion and underplating

(Fig. 19).

5) The final inversion of the MHB is not the end of the intermontane basin system of the Hellenides, as it corresponds to the onset of the Ptolemais basin further eastward, on the western border of the Olympos relief. Finally, the Trikala and Larissa plains still accommodate sediments during the Plio-Quaternary above former Miocene (MHB) and Pliocene (Ptolemais) depocentres.

ACKNOWLEDGEMENTS

This work has been supported by EC Projects (PLATON), by CNRS (UMR 8217) and Agricultural University of Athens fundings. We also thank the persons who help for the technical and drawing works (Annie Chini-Ferriere, Sandra Ventalon).

REFERENCES

Abreu V.S. and Haddad G.A. (1998) Glacioeustatic fluctuations: the mechanism linking stable isotope events and sequence stratigraphy from the early Oligocene to middle Miocene. In de Graciansky P.C., Hardenbol J, Jacquin T.and Vail P (Eds.), Mesozoic and Cenozoic sequence Stratigraphy of European Basins, SEPM Special Publication n°60, pp. 245259.

Allen P.A., Homewood P. and Williams G.D. (1986) Foreland basins: an introduction. In: Foreland Basins (Ed. by P.A. Allen and P. Homewood), Spec. Publ. Int. Ass. Sediment., 8, 312.

Aubouin J. (1959) Contribution à l'étude géologique de la Grèce septentrionale: les confins de l'Epire et de la Thessalie. Ann. Géol. Pays hellén., 10, 1484.

Barbieri R. (1992) Foraminifers of the Eptachori Formation (early Oligocene) of the Mesohellenic Basin, northern Greece. J. Micropalaeontology, 11, 7384.

Bizon G., Lalechos N. and Savoyat E. (1968) Présence de l'Eocène transgressif en Thessalie. Incidences sur la paléogéographie régionale. Bull. Soc. Géol. Fr., 10, 3638.

Bornovas J. and RondogianniTsiambaou T. (1983) Geological Map of Greece, 1:500,000. IGME, Athens.

Bonneau M. (1982) Evolution géodynamique de l'arc égéen depuis le Jurassique supérieur jusqu'au Miocene. Bull. Soc. Geol. Fr.,7,t. XXIV, n°2, 229242.

Bourcart J. (1925) Observations nouvelles sur la Tectonique de l'Albanie moyenne. Bull. Soc. Géol. Fr., 25, 391428.

Brunn J.H. (1956) Etude géologique du Pinde septentrional de la Macédoine occidentale. Ann. Géol. Pays hellén., 7, 1358.

Brunn J.H. (1969) Geological Map Pentalofon Sheet, 1:50,000. IGME, Athens.

Cashman, S. M., Kelsey, H. M., Erdamn, C. F., Cutten, H. N.C, and Berryman, K. R., 1992, Strain partitioning between structural domains in the forearc of the Hikurangi subduction zone, New Zealand, Tectonics, 11, No. 2, 242-257

Celet P. and Ferrière J. (1978) Les Hellénides internes : le Pélagonien. Eclogae geol. Helv. 71/3, p. 467-495.

Chanier F., Ferrière J. and Angelier J. (1992) Extension et Erosion tectonique dans un prisme d'accrétion : l'exemple du Prisme Hikurangi (Nouvelle-Zélande). C. R. Acad. Sci. Paris, C. R. Acad. Sci. Paris, t. 315, Série II, p.741-747.

Chanier F., Ferrière J. and Angelier J. (1999) Extensional deformation across an active margin, relations with subsidence, uplift and rotations: The Hikurangi subduction, New Zealand. Tectonics, vol. 18, 862-876.

Desprairies A. (1979) Etude sédimentologique des formations à caractère flysch et molasse, Macédoine, Epire (Grèce). Mém. Soc. Géol. Fr., 136, 180.

Dilek Y. and Altunkaynak S. (2009) Geochemical and temporal evolution of Cenozoic magmatism in western Turkey: mantle response to collision, slab break-off, and lithospheric tearing in an orogenic belt. In: van Hinsbergen, D.J.J., Edwards, D.J.J., Govers, R. (Eds.), Collision and Collapse at the Africa–Arabia–Eurasia.

Ditbanjong P. (2013) Stratigraphy and tectonics of Cenozoic deposits at the southwestern border of the Mesohellenicbasin (Continental Greece). Unpublished Ph. D. thesis, Lille University.

Doutsos T., PePiper G., Roronkay K. and Koukouvelas I. (1993) Kinematics of the central Hellenides. Tectonics, 12, 936953.

Doutsos T., Koukouvelas I., Zelilidis A. and Kontopoulos N. (1994) Intracontinental wedging and postorogenic collapse in Mesohellenic Trough. Geol. Rundsch., 83, 257275.

Dubois M., Buret C. and Chanier F. (2000) " SUBSILOG ", a " C " program for decompaction and subsidence computation. Application to the NewZeland forearc basin. Ann. Soc. Geol. du Nord, 8, 1924

Faugères L. (1977) Recherches géomorphologiques en Grèce septentrionale (Macédoine), Thesis Univ. Paris IV, 950 p. [Unpublished data].

Faugères L. (1977) Naissance et développement du relief de l'Olympe (Grèce) : une manifestation éclatante de la tectonique récente. Rev. Géog. phys. et Géol. dyn., 19, 726.

Ferrière J., Reynaud J.Y., Migiros G., Proust J.N., Bonneau M., Pavlopoulos A. and Houzé A. (1998) Initiation d'un bassin transporté : l'exemple du "sillon mésohellénique" au Tertiaire (Grèce). C. R. Acad. Sci. Paris, 326, 567574.

Ferriere J., Reynaud JY., Pavlopoulos A., Bonneau M., Migiros G., Chanier F., Proust JN. And Gardin S. (2004) Geological evolution and Geodynamic controls of the Intramontane Piggyback MesoHellenic Basin, Greece. Bull. Soc. Geol. France.

Ferriere J., Chanier F., Reynaud J.-Y., Pavlopoulos A., Ditbanjong P., Migiros G., Coutand I. and Bailleul J. (2011) Tectonic control of the Meteora conglomerates (Meso-Hellenic Basin, Greece). Bull. Soc. Geol. France, vol.182, p. 437-450

Ferrière, J., Chanier F. and Ditbanjong P. (2012) The Hellenic ophiolites: eastward or westward obduction of the Maliac Ocean, a discussion. International Journal of Earth sciences, vol. 101, p.1559-1580. (doi: 10.1007/s00531-012-0797-9)

Fleury J.J. and Godfriaux I. (1975) Arguments pour l'attribution de la série de la fenêtre de l'Olympe (Grèce) à la zone de GavrovoTripolitza : présence des fossiles du Maastrichtien et de l'Eocene inférieur (moyen ?). Ann. Soc. Geol. du Nord, 94, 149156.

Godfriaux I. (1968) Etude géologique de la région de l'Olympe (Grèce). Ann. Géol. Pays hellén., 19, 1281.

Haq B.U., Hardenbol J. and Vail P.R. (1987) Chronology of fluctuating sealevels since the Triassic (250 Myrs ago to present). Science, 235, 11561167.

Kilias A., Fasoulas C., Priniotakis M., Sfeikos A. and Frisch W. (1991) Deformation and HP-LT Metamorphic conditions at the tectonic Window of Kranea (WThessaly, Northern Greece). Z. dt. geol. Ges., 142, 8796.

Kontopoulos N., Fokianou T., Zelilidis A., Alexiadis C. and Rigakis N. (1999) Hydrocarbon potential of the middle Eocene middle Miocene Mesohellenic piggyback basin (central Greece): A case study. Marine and Petroleum Geology, 16, 811824.

Koumantakis J., Matarangas D., TsaïlaMonopolis S. and Georgiadou E. (1980) Geological Map Panayia Sheet, 1:50,000, IGME, Athens.

Lacassin R., Arnaud N., Leloup P.H., Armijo R. and Meyer B. (2007). - Exhumation of metamorphic rocks in N Aegean : the path from shortening to extension and extrusion. - eEarth Discuss., 2, 1-35.

Lecanu H. (1976) La région du HautPénée, Thessalie, Grèce. Thesis University of Paris 6. [Unpublished data.]

Lekkas E., (1988) Geological structure and geodynamic evolution at the Koziakas mountain range (western Thessaly). National and Kapodistrian University of Athens, Dept of Geology. Geol Monogr.1, 281p.

Mavridis A., Kelepertzis A.K., TsaïlaMonopolis S. and SkourtsiKoroneou V. (1993) Geological Map Knidi Sheet, 1:50,000. IGME, Athens.

Mavridis D., Matarangas D., TsaïlaMonopolis S. and Mostler H. (1979) Geological Map Ayiofillon Sheet, 1:50,000. IGME, Athens.

Mercier J.L., Simeakis K., Sorel D. and Vergely P. (1989) extensional tectonic regimes in the Aegean basins during the Cenozoic. Basin Research 2, 49-71

Migiros G., Bathrellos G., Skilodimou H., Karamousalis Th. (2011): Pinios (Peneus) river (Central Greece): Hydrological - Geomorphological elements and changes during the Quaternary. Cent. Eur. J. Geosci. 3(2) 215228.

Nicol A., Mazengarb C., Chanier F., Rait G., Uruski C. and Wallace L. (2007). Tectonic evolution of the active Hikurangi subduction margin, New Zealand, since the Oligocene. Tectonics, Vol.26, TC4002.

Ori G.G. and Roveri M. (1987) Geometries of Gilberttype deltas and large channels in the Meteora Conglomerate, MesoHellenic basin (OligoMiocene), Central Greece. Sedimentology, 34, 845859.

Ori G.G. and Friend P.F (1984) Sedimentary basins formed and carried piggyback on active thrust sheets. Geology, 12, 475478.

Papanikolaou D., Lekkas E., Mariolakos H. and Mirkou R. (1988) Evolution of the Mesohellenic Basin. Bull. Soc. Géol.Gr., 20, 1736.

Pe-Piper, G., Piper, D.J.W. (2006). Unique features of the Cenozoic igneous rocks of Greece. In: Dilek, Y., Pavlides, S. (Eds.), Postcollisional Tectonics and Magmatism in the Mediterranean Region and Asia: Geological Society of America Special Paper. Geological Society of America, pp. 259–282. http://dx.doi.org/10.1130/ 2006.2409(14).

Pe-Piper, G., Piper, D.J.W., (2007). Neogene back-arc volcanism of the Aegean: new in- sights into the relationship between magmatism and tectonics. In: Beccaluva, L., Bianchini, G. (Eds.), Cenozoic Volcanism in the Mediterranean Area: Geological Society of America Special Paper. Geological Society of America, pp. 17–31. http:// dx.doi.org/10.1130/2007.2418(02).

Qirjaku Kaleshi E. (2000) Analyse structurale de la sigmoïde albanaise par imagerie satellitaire et modèles numériques de terrain. Unpublished Ph. D. thesis, Paris 6 university.

Robertson A.H.F. and Shallo M. (2000) MesozoicTertiary evolution of Albania in its regional Eastern Mediterranean context. Tectonophysics 316, 197254.

Roussos N. (1994) Stratigraphy and paleogeographic evolution of the Paleogene molassic basins of the north Aegean Sea. Bull. Soc. Greece, 30, 275294.

Savoyat E., Lalechos N. and Bizon G. (1969a) Geological Map Trikala Sheet, 1:50,000. IGME, Athens.

Savoyat E., Lalechos N. and Bizon G. (1969b) Geological Map Karditsa Sheet, 1:50,000. IGME, Athens

Savoyat E., Lalechos N., Philippakis N. and Bizon G. (1972a) Geological Map Kalabaka Sheet, 1:50,000. IGME, Athens.

Savoyat E., Monopolis D. and Bizon G. (1971a) Geological Map Nestorion Sheet, 1:50,000. IGME, Athens.

Savoyat E., Monopolis D. and Bizon G. (1972b) Geological Map Grevena Sheet, 1:50,000. IGME, Athens

Savoyat E., Verdier A., Monopolis D. and Bizon G. (1971b) Geological Map Argos Orestikon Sheet, 1:50,000. IGME, Athens.

Schermer E. R., Lux D. and Burchfiel B. (1989) Age and tectonic significance of metamorphic events in the mt Olympos region, Greece. Bull. geol. Soc. Greece, 23, 1327.

Schermer E.R. (1993) Geometry and kinematics of continental basement deformation during the Alpine orogeny, Mount Olympos region, Greece. J. Struct. Geol., 15, 571591.

Sinclair H.D. (1997a) Flysch to molasse transition in peripheral foreland basins: the role of the passive margin versus slab breakoff. Geology, 25, 11231126.

Sinclair H.D. (1997b) Tectonostratigraphic model for underfilled peripheral foreland basins: an Alpine perspective. Geol. Soc. Am. Bull., 109, 324346

Skourlis K. and Doutsos T. (2003) The Pindos Fold-and-thrust belt (Greece): inversion kinematics of a passive continental margin, Int. J. Earth Sci. (Geol. Rundsch.), vol.92, p.891–903. (doi: 10.1007/s00531-003-0365-4).

Soliman H.A. and Zygojiannis N. (1980) Geological and Paleontological studies in the Mesohellenic Basin, Northern Greece. I Oligocene smaller Foraminifera. II Eocene smaller Foraminifera Geol. Geophys. Research, XXII, 1: 166, 27 plates; IGME Athens

Thiebault F. (1982) Evolution géodynamique des Hellenides externes en Peloponnèse méridional (Grèce). Soc. Geol. Nord, Publication n° 6.

Tranos M.D., Mountrakis D.M., Papazachos C.B., Karagianni E. and Vamvakaris D. (2010) Faulting Deformation of the Mesohellenic Trough in the Kastoria-Nestorion Region (Western Macedonia, Greece). Bulletin of the Geological Society of Greece. Proceedings of the 12th International Congress, Patras, May, 2010.

Vamvaka A., Kilias A., Mountrakis D. and Papaoikonomou (2006) Geometry and structural evolutionof the Mesohellenic trough (Greece): a new approach. Bull. Geol. Soc. Greece, 32, 1, 209-219.

Vidakis M., Papazeti E. and TsaïlaMonopolis S. (1998) Geological Map Nestorion Sheet, 1:50,000, IGME, Athens.

Wilson J. (1993) The Anatomy of the Krania Basin, NW Greece. Bull. Geol. Soc. Greece, 28, 361368.

Zelilidis A. and Kontopoulos N. (1996) Significance of fan deltas without toesets within rift and piggy-back basins: examples from the Corinth graben and the Mesohellenic trough, Central Greece. Sedimentology, 43, 253262.

Zelilidis A., Kontopoulos N., Avramidis P. and Bouzos D. (1997) Late Eocene to early Miocene depositional environments of the Mesohellenic basin, NorthCentral Greece: implications for hydrocarbon potential. Geologica Balcanica, 27, 4555.

Zelilidis A., Piper D. and Kontopoulos N. (2002) Sedimentation and basin evolution of the Oligocenemiocene Mesohellenic Basin, Greece. AAPG Bulletin, 86, 161182.

Zygojiannis N. and Müller C. (1982) NannoplanktonBiostratigraphie der tertiären Mesohellenischen Molasse (NordwestGriechenland). Z. dt. geol. Ges., 133, 445455.

Zygojiannis N. and Sidiropoulos D. (1981) Schwermineralverteilungen und paläogeographische Grundzüge der tertiären Molasse in der Mesohellenischen Senke, NordwestGriechenland. N. Jahrb Geol. Paläontol. Mh., 100128.

copper-gold deposit, Chalkidiki, Greece. In; Porter TM (ed) Porphyry hydrothermal Cu-Au deposits. A global perspective Australian Mineral Foundation Conference Proceedings. Glenside, South Australia, 159-168.

Tolstykh, N. D., Sidorov, E.G., Krivenko, A.P., 2005. Platinum-group element placers associated with Ural-Alaska type complexes. Mineralogical Association of Canada Short Course , 35, 113-143.

Tosdal, R.M. and Richards, J.P., 2001. Magmatic and structural controls on the development of porphyry CuMoAu deposits, SEG Reviews, 14, 157-181.

Tompouloglou, C., 1981. Les mineralisations Tertiaires, type cuivre porphyrique, du massif Serbo-Macedonien (Macedoine Greece) dans leur contexte magmatique (avec un traitment geostatistique pour les donnees due prospect d' Alexia). Rh.D thesis presented at the L' Ecole Nationale Superieure des Mines de Paris, France, pp. 204.

Thompson, J.F.H., Lang, J.R. and Stanley, C.R., 2001. Platinum group elements in alkaline porphyry deposits, British Columbia. Exploration and Mining in British Columbia, Mines Branch, Part B, 57-64.

Tsoupas, G., Economou-Eliopoulos, M., 2005. Extremely abundant platinum-group minerals in chromitites hosted in the Veria ophiolite complex, northern Greece. In: Törmänen, T.O., Alapieti, T.T. (Eds.), 10th International Platinum Symposium, Oulu, Finland, 592-595.

Tsoupas, G., Economou-Eliopoulos, M. 2008. High PGE contents and extremely abundant PGE-minerals hosted in chromitites from the Veria ophiolite complex, northern Greece. Ore Geology Reviews 33, 3–19.

Von Quadt, A., Peytcheva, I., Kamenov, B., Fanger, L., Heinrich, C.A. and Frank, M. 2002. The Elatsite porphyry copper deposit in the Panagyurishte ore district, Srednogorie zone, Bulgaria: U-Pb zircon geochronology and isotope-geochemical investigations

of magmatism and ore genesis. in: Blundell, D.J., Neubauer, F. and von Quadt, A. (eds): The Timing and Location of Major Ore Deposits in an Evolving Orogen. Geological Society, London, Special Publications 204, 119–135

Vergely P. 1976. Origine "vardarienne", chevauchement vers l'Ouest et re´trocharriage vers l'Est des ophiolites de Macédoine (Grèce) au cours du Jurassique supérieur-Eocrétacé. C R Acad Sci Paris 280, 1063–1066

Vergely, P. 1984. Tectonique des ophiolites dans les Hellenides internes. Conséquences sur l'évolution des régions téthysiennes occidentales. These Université Paris sud Orsay 2, 250-411.

Walker, R., Brandon, A.D. Bird, J.M. Piccoli, P.M. McDonough, W.F. Ash, R.D. 2005. 187Os-186Os systematics of Os-Ir-Ru alloy grains from southwestern Oregon, Earth Planet. Sci. Lett. 230, 211 – 226.

Walker, D.A., Cameron, W.E., 1983. Boninite primary magmas: evidence from the Cape Vogel Peninsula, Papua New Guinea. Contributions to Mineralogy and Petrology 83, 150–158.

Werle, J.I., Ikramuddin, M. and Mutschler, F.E. 1984. Allard stock, La Plata Mountains, Colorado — an alkaline rock hosted porphyry copper precious metal deposit. Can. J. Earth Sciences 21, 630–641.

Whiteley, J.D., Murray, F., 2003. Anthropgenic platinum group element (Pt, Pd and Rh) concentrations in road dusts and roadside soils from Perth, Western Australia. The Science of Total Environment 317, 121-135.

Wilde, A.R., Edwards, A.C., Yakubchuk, A., 2003. Unconventional deposits of Pt and Pd: a review with implications for exploration: SEG Newsletter, 52, pp. 1-18.

Wood, S.A., Mountain, B.W., and Pujing Pan, 1992. The aqueous geochemistry of platinum, palladium and gold: recent experimental constraints and a re-evaluation of theoretical predictions, Canadian Mineralogist, 30, 955-982.

Wood, S.A., 1990. The interaction of dissolved platinum with fulvic acid and simple organic acid analogues in aqueous solutions: The Canadian Mineralogist, 28, 665-673.

Wood, S. A. 2005. The effect of organic ligands on the mobility of the PGE in soils and natural waters: Implications for exploration and the environment. Geochimica et Cosmochimica Acta 69, A329.

Wood, B.W., 2002. The aqueous geochemistry of the platinum-group elements with applications to ore deposits, in Cabri, L.J., ed., Geology, Geochemistry, mineralogy, metallurgy and beneficiation of PGE: Canadian Institute of Mining and Metallurgy Special Paper, 54, 211–249.

Xiong, Y. and Wood, S.A., 2000. Experimental quantification of hydrothermal solubility of platinum-group elements with special reference to porphyry copper environments. Mineralogy and Petrology 68, 1-28.