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Architecture of Kinematics and Deformation History of the Tertiary Supradetachment Thrace Basin: Rhodope Province (NE Greece)

Adamantios Kilias, George Falalakis, Aristides Sfeikos,
Eleftheria Papadimitriou, Agni Vamvaka and Chara Gkarlaouni
Aristotle University of Thessaloniki, School of Geology
Greece

1. Introduction

The Rhodope province in Northern Aegean region where the Thrace basin, one of the largest Tertiary basins in this region, is located (Fig. 1) belongs to a highly extended continental terrain during Tertiary time. Recent studies indicate that extensional deformation began in Eocene (Kilias et al., 1999; Burchfiel et al., 2003, Brun & Sokutis, 2007) or earlier in Early Paleocene (Bonev et al., 2006). Dinter & Royden (1993), Dinter et al. (1995) and Wawrenitz & Krohe (1998) suggest that the onset of extension can be set in the Miocene. Extension took place simultaneously with compression at the more external parts of Hellenides (Fig. 1; Schermer et al. 1990; Kilias et al. 1991, 1999; Jolivet & Brun 2010).

Extensional basins and crustal subsidence are often temporally and spatially associated with uplift/exhumation of metamorphic rocks (Friedmann & Burbank, 1995; McClaughry & Gaylord, 2005). Two end-members of extensional basins can be recorded during a continuous extensional tectonism, which are: rift basins and supradetachment basins (Friedmann & Burbank, 1995). The differences in magnitude and rate of extension, volcanism, heat flow and structural architecture define the basin style and suggest the tectonic setting related to the basin evolution (Friedmann & Burbank, 1995; McClaughroy & Gaylord, 2005).

The Thrace basin formed on the metamorphic rocks of the Rhodope massif in Northern Greece and the Strandja and Sankarya massifs in NW-ern Turkey (Fig. 1, 2). In Southern Bulgaria and Western Former Yugoslavian Republic of Macedonia (FYROM) a series of equivalent basins with similar deposits infilling are also developed between the metamorphic rocks of the Bulgarian Rhodope massif (Fig.1, 2; Burchfiel et al., 2003; Dumurtzanov et al., 2005). The basins evolution is related to an intensive calc-alkaline and locally shoshonitic magmatism forming the NW-SE trending Eocene-Oligocene volcanic arc of the Hellenides. Its SW-ward progressive migration until the present day active Hellenic volcanic arc has been recorded by several works (Fig. 1; Innocenti et al., 1982; Fytikas et al., 1984). The position of the Thrace basin on the highly extended Rhodope metamorphic sequence, behind of the SW-ward contemporaneous nappes stack of Hellenides, give the Thrace basin an important role to understand better the geodynamic context of the Hellenides.
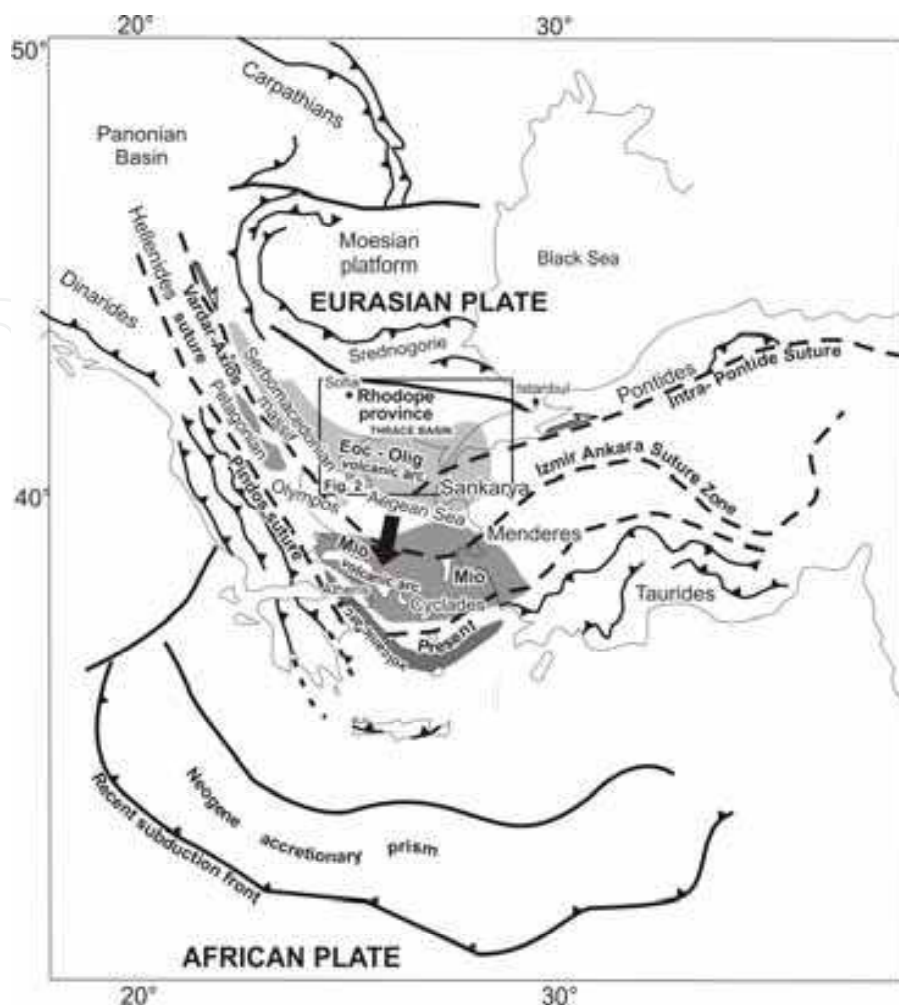
Fig. 1. Position of the Rhodope province and Thrace basin with respect to the main tectonic units of southeastern Europe and eastern Mediterranean region. The SSW ward migration of the Tertiary magmatic activity in the Hellenic arc is also shown

Our study is focused on the Greek part of the Thrace basin stretched from the Pangaion mountainous range until the Greek borders with Turkey along the Evros river. Metamorphic units of the Rhodope massif constitute the Northern tectonic boundary of the Greek Thrace basin, while its Southern parts are lost under the Aegean Sea and are exhumed again as small remnants on some islands of the Northern Aegean (e.g. Limnos, Samothraki; Fig. 1, 2). The Thrace basin in NW-ern Turkey has been intensely studied because of its importance as a natural hydrocarbon reservoir (Okay et al., 1990; Goeruer & Okay, 1996; Coskun, 1997; Turgut & Eseller, 2000; Siyako & Huvaz, 2007).

For the Greek Thrace basin published results are also available (e.g. Christodoulou, 1950; Kopp, 1965; Lescuyer et al., 2003; Kilias et al., 2006; Tranos, 2009). However, details on the basin evolution and its tectonic position in the geodynamic context of North Aegean region, as well as a relationship to the exhumation history of the Rhodope massif are lacking.

The aim of the current study is to present detailed information on the structural evolution of the Greek Thrace basin and its kinematics of deformation, as well as to clarify its geodynamic setting in the frame of the Hellenic orogenic belt. Furthermore, the calculation of the paleostress tensor for each deformational event recorded in the overall tectonic history of the basin is performed. We interpret the Tertiary Thrace basin as a

supradetachment basin related to the extensional exhumation history of the footwall Rhodope metamorphic rocks.

2. Geological setting

2.1 The Rhodope metamorphic basement

The Rhodope metamorphic province consists of high to moderate grade metamorphic units forming a complicated metamorphic nappe pile complex. It records a series of tectonic events of pre-Alpine and Alpine age and a different Tertiary exhumation age of the several metamorphic units. Alpine nappe stacking and crustal thickening were resulted from the Cretaceous-Paleocene SW-ward accretion of a number of structural units of mainly continental origin, sandwiched between Moesian continental block (Eurasia) in the North and Adriatic-Apulia continental block (Africa) in the South (Burg et al., 1996; Zagorchev, 1998; Bonev et al., 2006; Georgiev et al., 2010; Jahn-Awe et al., 2010).

According to their composition and tectonometamorphic setting the several metamorphic units are distinguished into a Lower and an Upper structural plate (Fig. 2; Krohe & Mposkos, 2002; Marchev et al., 2005; Bonev et al., 2006; Georgiev et al., 2010). The Upper plate contains the Kimi unit and the Lower plate includes the Sidironero, Kardamos-Kesebir and Kechros units (Fig. 2). At the westernmost Rhodope massif, the Sidironero unit overthrusts the lowermost Rhodope unit, the Pangaion unit (Fig. 2; Papanikolaou & Panagopoulos, 1981; Kilias & Mountrakis, 1990; Dinter et al., 1995).

Lower and Upper units are divided by low angle normal detachment fault systems, resulting in a progressively extensional exhumation of the several tectonic units from top to the bottom and the building of several metamorphic core complexes and tectonic windows (Fig. 2; e.g. Kessebir-Kardamos, Kechros and Pangaion domes). The structurally uppermost Kimi unit was the earliest exhumed Rhodope part (cooling age 65-60 Ma), followed by the exhumation of the Sidironero, Kessebir-Kardamos and Kechros units of the Lower plate (cooling age 42-30 Ma). Last to be exhumed was the lowermost Pangaion unit (cooling age 26-10 Ma), forming a well defined metamorphic core complex (Kilias et al., 1999; Krohe & Mposkos, 2002; Mposkos & Krohe, 2006; Georgiev et al., 2010).

Three main metamorphic events of alpine age from Cretaceous to Oligocene-Miocene time are identified: An earlier ultra-high to high pressure eclogite facies event, a second medium pressure amphibolite facies event and a late greenschist facies retrogressive event. The lowermost Pangaion unit was affected by lower P-T metamorphic conditions in the greenschist facies (Kilias & Mountrakis, 1990; Wawrenitz et al., 1998; Liati & Gebauer, 1999; Krohe & Mposkos, 2002; Mposkos & Krohe, 2006).

Acid to basic magmatism of Paleogene to Neogene age, associated with syn- to late-orogenic extension and thinning of the former over-thickened continental crust during Cretaceous-Paleocene time, characterizes the Rhodope metamorphic nappe pile (Koukouvelas & Pe-Piper, 1991; Dinter et al., 1995; Burg et al., 1996; Kilias & Mountrakis, 1998; Christophides et al., 2001, 2004; Marchev et al., 2005; Bonev et al., 2006).

To the West the Rhodope massif is separated from the Serbomacedonian massif by a mylonitic, low angle, normal detachment fault zone defining very well the Rhodopian Pangaion metamorphic core complex (Fig. 2; Dinter & Royden, 1993; Sokoutis et al., 1993; Kilias et al., 1999). The Serbomacedonian massif forms also a metamorphic nappe pile complex of Paleozoic or older age's rocks, including the Upper Vertiskos and the Lower Kerdylia metamorphic units (Fig. 2; Kockel et al., 1971; Papadopoulos & Kilias, 1985; Kilias et al.,

1999). Moreover, recent works distinguish a normal detachment fault zone also along the tectonic contact between the two Serbomacedonian units (i.e. Kerdylia and Vertiskos units), regarding the lower Kerdylia unit as a part of the Rhodope Massif (Brun & Sokoutis, 2007).
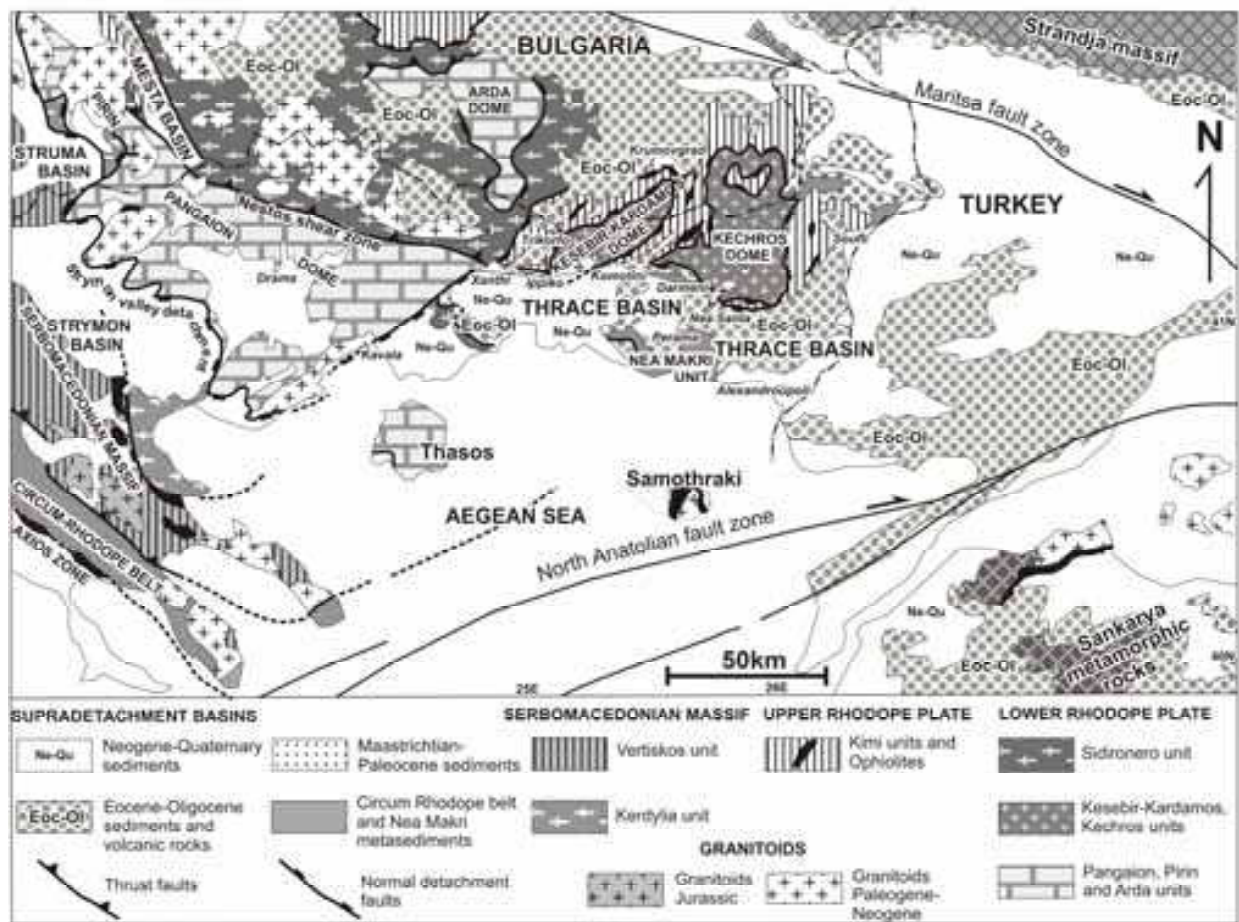

Fig. 2. Simplified geological map of the Rhodope province. The Thrace basin and the main structural units of the Rhodope and Serbomacedonian massifs with their tectonic relationships, as well as the Sankarya and Strandja metamorphic rocks in northwestern Turkey (Pontides-Anatolides belt) are shown (modified after Kilias et al., 1999; Turgut & Eseller, 2000; Krohe & Mposkos, 2002; Bonev et al., 2006)

A low grade metamorphic volcano-sedimentary unit of Upper Paleozoic-Mesozoic age, named Nea Makri unit, forming possibly the E-ward continuation of the Circum Rhodope Belt (Fig. 2; Kockel et al., 1971; Kaufmann et al., 1976), occurs in some places as small rests over the high grade Rhodope metamorphic rocks. The contact between the Nea Makri metasediments and the underlying Rhodope metamorphic units usually develops as a normal detachment fault zone with sense of shear towards S to SW (Fig. 2), related to the Tertiary extensional collapse history of the Rhodope basement rocks (Kilias et al., 1999; Bonev et al., 2006; Georgiev et al., 2010).

2.2 The deposits of the Thrace basin
The Tertiary sediments in the Greek part of the Thrace Basin are overlaid with a total well as in a few cases over the Nea Makri metasediments bounded usually with the

underlying units with big fault zones. In the Turkish part, a thickness up to 9 km is reported (Fig. 2, 3; 4; Huvaz et al., 2007; Siyako & Huvaz, 2007; Mainhold & BouDagher-Fadel, 2010; Jolivet & Brun, 2010).

The deposits of the Thrace basin consist of Paleogene molassic type sedimentary rocks, as well as Neogene and Quaternary sediments (Christodoulou, 1950; Kopp, 1965; Lalechos, 1986; Mainhold & BouDagher-Fadel, 2010). The Paleogene molassic sediments are intercalated with a lot of calc-alkaline and partly shoshonitic type, volcanic products (Fig. 3). The latter form lava flows, debris-flows, hydroclastites, domes, dykes and numerous pyroclastics. Their chemical composition ranges from basaltic pyroxene andesite to biotite rhyolith through pyroxene or biotite-hornblende andesite, pyroxene trachyandesite, biotite-hornblende dacite and pyroxene-biotite trachydacite (Innocenti et al., 1982; Christofides et al., 2004). K/Ar data show ages from 33.4 to 20 Ma, establishing an Early Oligocene to Early Miocene volcanic activity (Christofides et al., 2004). However, intercalations of volcanic products with the earliest Middle-Late Eocene clastic sediments of the Thrace basin indicate that volcanic activity started earlier than Early Oligocene, in Middle-Late Eocene times. Furthermore, the Oligocene-Miocene Xanthi granite (Koukouvelas & Pe-Piper, 1991) intrudes in some places into the Paleogene strata of the basin (Fig. 2, 4), forming a contact metamorphic aureole of a few meters thickness.
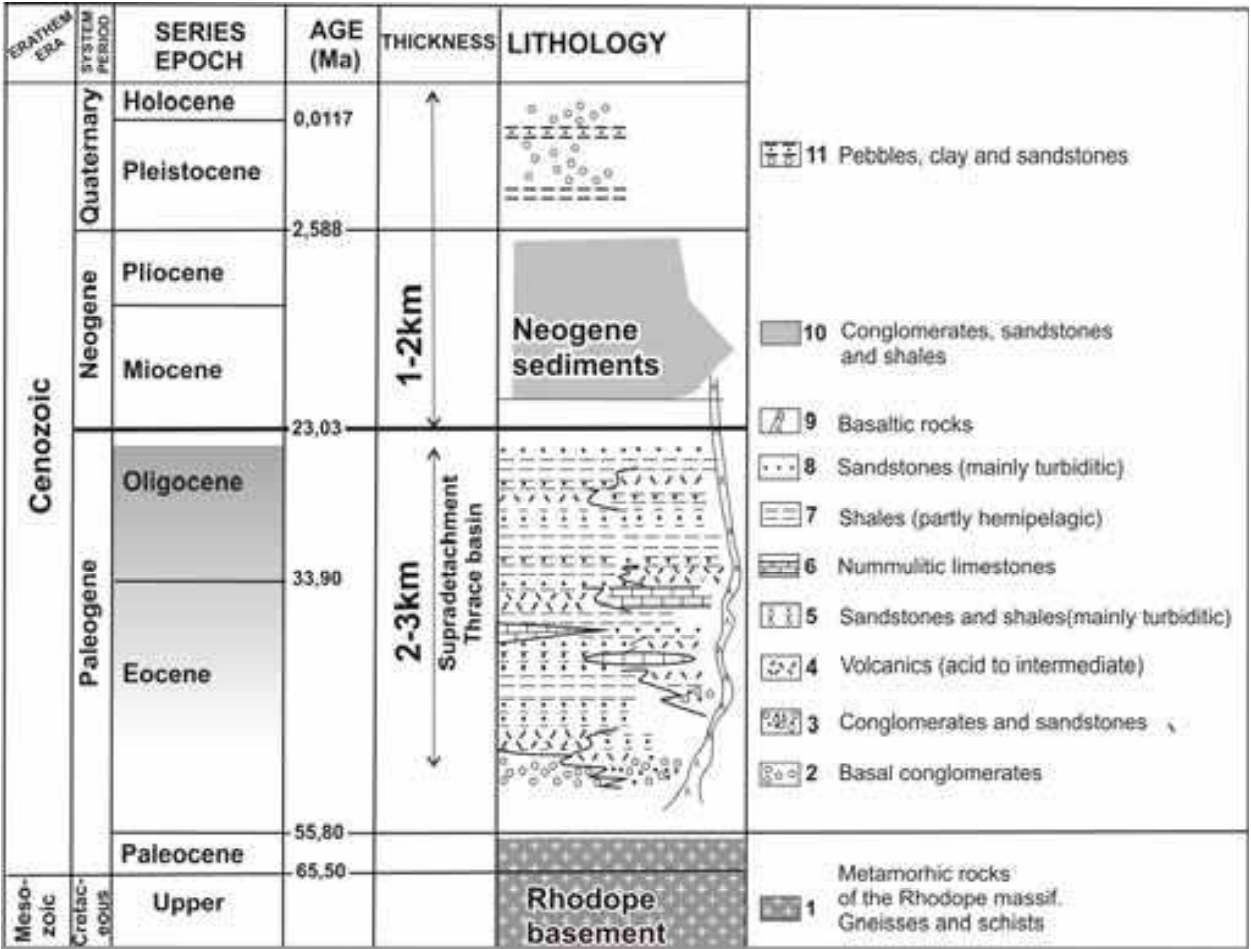

Fig. 3. Schematic stratigraphic column of the Thrace basin compiled after data in Christodoulou, 1950; Kopp, 1965; Lescuyer et al., 2003; Burchfiel et al., 2003

The Paleogene sediments of the Thrace basin show an age from Middle-Late Eocene to Oligocene (Christodoulou, 1950; Dragomanov et al., 1986; Zagorcev, 1998; Burchfiel et al., 2003; Meinhold & BouDagher - Fadel, 2010). They constitute a complicated stratigraphic sequence composed by intercalations of bedded conglomerates, breccia conglomerates, sandstones, numulitic limestones, turbiditic layers, and shales (Fig. 3). Sedimentation starts during Lutetian- Bartonian time with initial deposition of continental sediments (mainly breccias conglomerates and sandstones), followed during Late Eocene-Oligocene by marine turbiditic type deposits and limestones interbedded by the volcanogene products of the Late Eocene-Oligocene volcanic activity in the area (Fig. 3, 4; Dragomanov et al., 1986; Zagorcev; 1998; Burchfiel et al., 2003).
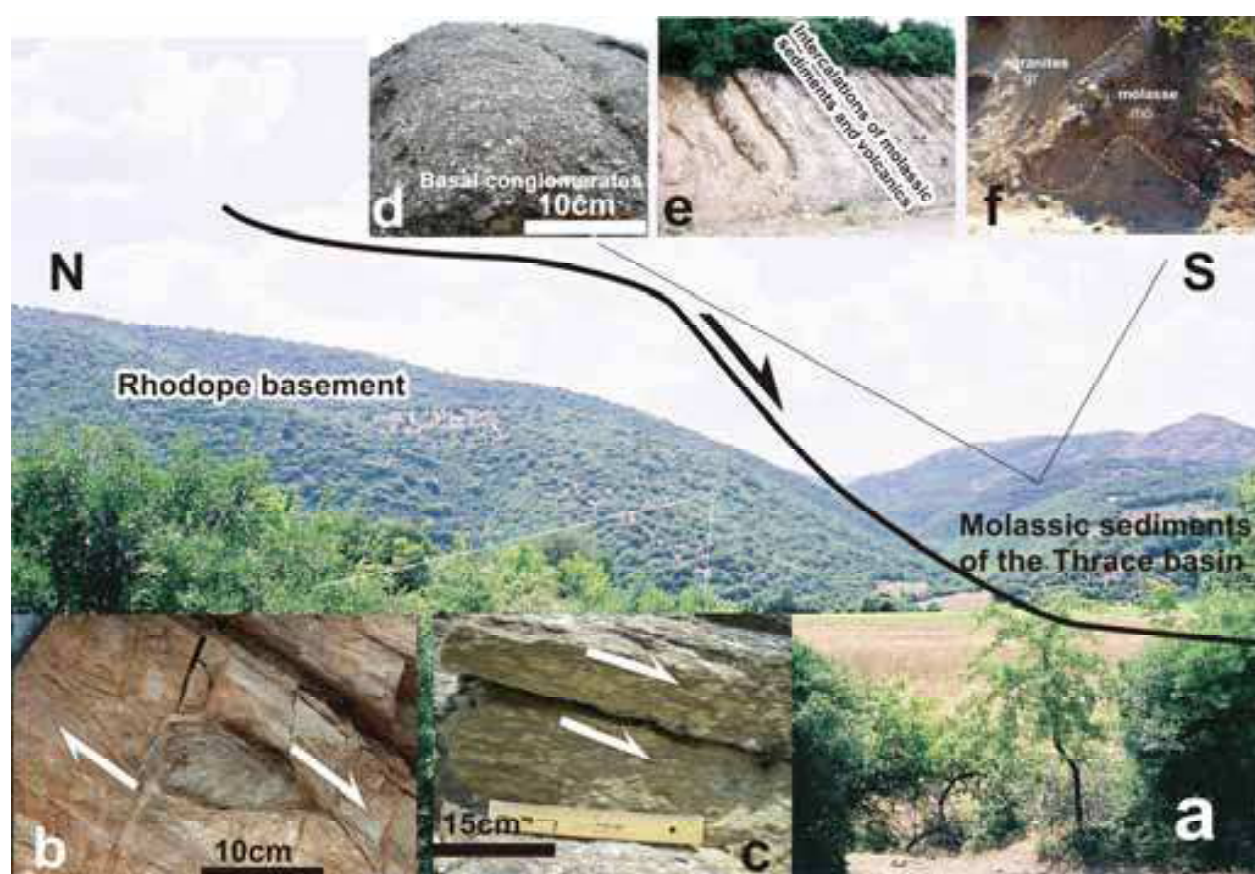

Fig. 4. a. Looking E the detachment zone between Rhodope basement rocks and molassic sediments near Nea Santa village. b & c. ductile fabrics at the footwall of the detachment fault basement rocks, σ-clasts and shear bands, showing a top to SSW sense of shear. d. basal breccia conglo-merate of the basin deposits. e. molassic sediments interbedded with volcanics with high angle dipping (>65 dip angle). f. Oligocene granite intrusion into the molassic sediments

A thick sedimentary sequence of Neogene age overlies unconformably the Paleogene strata of the basin containing conglomerates, sandstones and siltstones. Pebbles (breccio-conglomerates), clay and sandstones form the youngest Quaternary sediments of the whole stratigraphic column (Fig. 2, 3).
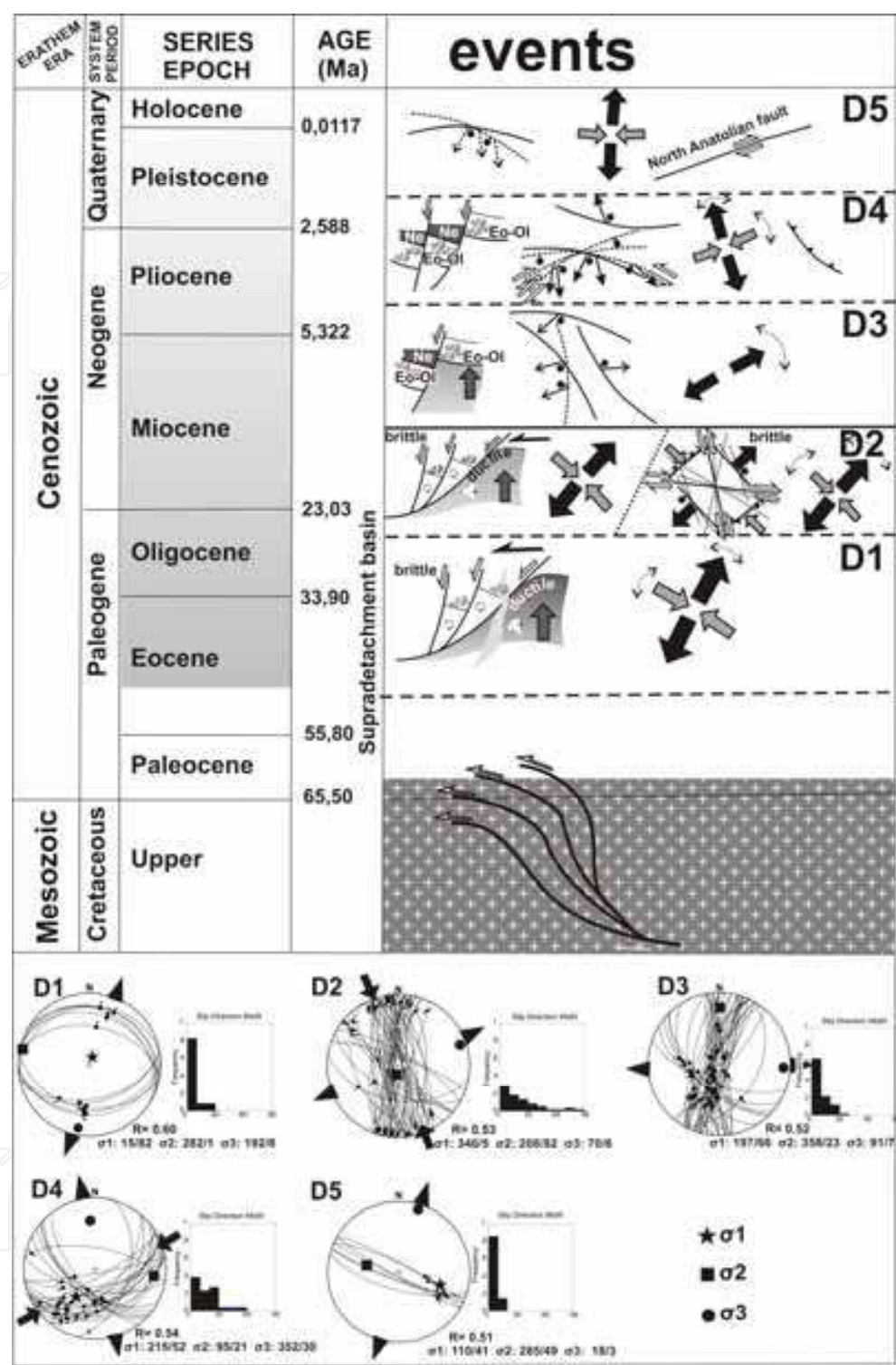
Fig. 5. Distinctive features and kinematics of the five main deformational processes (D1 to D5) related to the evolution of the Thrace basin from Eocene to recent. Mesozoic-Early Paleogene contractional tectonic and nappe stacking in Rhodope massif (Burg et al., 1996; Jahn-Awe et al., 2010) were replaced during Paleogene to Neogene by extensional tectonic, exhumation of deep crustal metamorphic sequences and coeval magmatism, as well as subsidence of the Thrace basin. Active tectonic dominates during Quaternary time. Paleostress analysis diagrams (σ1>σ2>σ3) for each tectonic event are also shown

3. Data and observations. Structural analysis

The Thrace basin shows a complicated structural evolution from its initial stages until present. Several successive deformational events interacted making their distinction difficult in several cases. The deformation progression documents semi-ductile to brittle conditions. Subhorizontal extension characterizes the overall deformation history.

The geometry and kinematics of deformation, cross cutting relationships, kinematic indicators and slickenslides overprinting criteria on the fault planes record the progressive activity of five (D1 to D5) tectonic events. They took place from Middle-Late Eocene up to present and are related to the basin evolution and the unroofing of the Rhodope metamorphic units (Fig. 5). Fault-slip data were used in order to calculate the paleostress tensor, for each tectonic event following the direct stress inversion method (Angelier, 1979, 1990). The solution is considered satisfactory if more than 80% of the fault-slip data from a site show a misfit angle less than 30o between the theoretic and real slip vector. The program My fault, Version 1.03 (Pangea scientific, 2005) and StereoNett (Duyster, 2000) were used for the graphical presentation of the tectonic data.

3.1 D1 event

In places where the contact between Paleogene sediments of the Thrace and the metamorphic basement rocks of the Rhodope massif or the low grade metasediments of the Nea Makri belt has not been reworked by more recent events, this is developed as a low angle, normal detachment fault (Fig. 5, 6, 7).

In many places, the entire Nea Makri belt is omitted or exists as small remains in between Paleogene sediments and underlying Rhodope metamorphic sequence, indicating the extensional regime of the tectonic contact and the basin evolution (Fig. 5, 6). The omission of several tectonic units of the whole Rhodope metamorphic nappe stack and the extensional unroofing and exhumation of the underlying deep crustal units during the Tertiary deformation of the Rhodope province have been described in detail by several authors (Kilias & Mountrakis, 1990; Sokoutis et al., 1993; Kilias et al., 1999; Bonev et al., 2006; Brun & Sokoutis, 2007; Georgiev et al., 2010).

The detachment fault zone indicates brittle to semi-ductile conditions and consists commonly of a cataclastic domain of a few meters thickness (ca. 2-5m; Fig. 7, 8b). Furthermore, pseudotachylites between the cataclastic fault structures are observed in some cases. On the fault plane, a well-developed striation, some times with an important strike-slip component, is recorded. Kinematic indicators (Hancock, 1985) show a main top to SSW-S sense of movement (Fig. 8b). In some places (e.g., Perama area), the tectonic contact is evolved with opposite top to the N-NNE sense of movement. It is related to secondary, conjugate faults with opposite N-NNE-ward dip direction to the main fault zones. These faults are evolved on the hangingwall fault segment of the main detachment fault zone but rooted into it. (Fig. 6, 9).

The strike of the fault zone is not constant. It is usually developed corrugated or folded with a SW to SSW dipping fold axis (Fig. 6), following the dipping of the fault plane. The hangingwall Paleogene strata dips in some cases opposite to the main dipping of the detachment fault plane or with a low angle conformably to the fault dipping (Fig. 7).

At the footwall of the detachment fault zone, the metamorphic basement rocks contain a S-SW-ward dipping mylonitic fabric with the same kinematic symmetry (top to the S-SSW, through ductile kinematic indicators, Simpson & Schmid, 1983) to the fault zone (Fig. 4).

Antithetic top to the N-NNE sense of movements are interpreted as secondary, of local important movements, due to strain inhomogenity or as a result of an important coaxial component of deformation (Kilias at al., 1999; Bonev et al., 2006). A NE-SW to NNE-SSW micro folding of the mylonitic fabric parallel to the stretching lineation (X- axis of the finite strain ellipsoid in the field) is also mapped, following the described folding geometry of the detachment fault plane.

Paleostress analysis based on fault-slip data from the detachment zone indicates NE-SW to NNE-SSW trending subhorizontal minimum $\sigma3$- and subvertical maximum $\sigma1$-axes of the stress tensor (Fig. 5, 10).

3.2 D2 event

The D2 event is related to the further opening and reconstruction of the Thrace basin and is associated with intense magmatic activity guided by the D2 crustal structures. D2 is characterized by a conjugate pair of major strike-slip faults forming very important structures in the Paleogene strata of Thrace basin. The first set is developed with NNE-SSW to NNW-SSE strike with a sinistral component. The second set strikes WNW-ESE to NW-SE with a dextral horizontal component (Fig. 5, 6, 11, 8a).
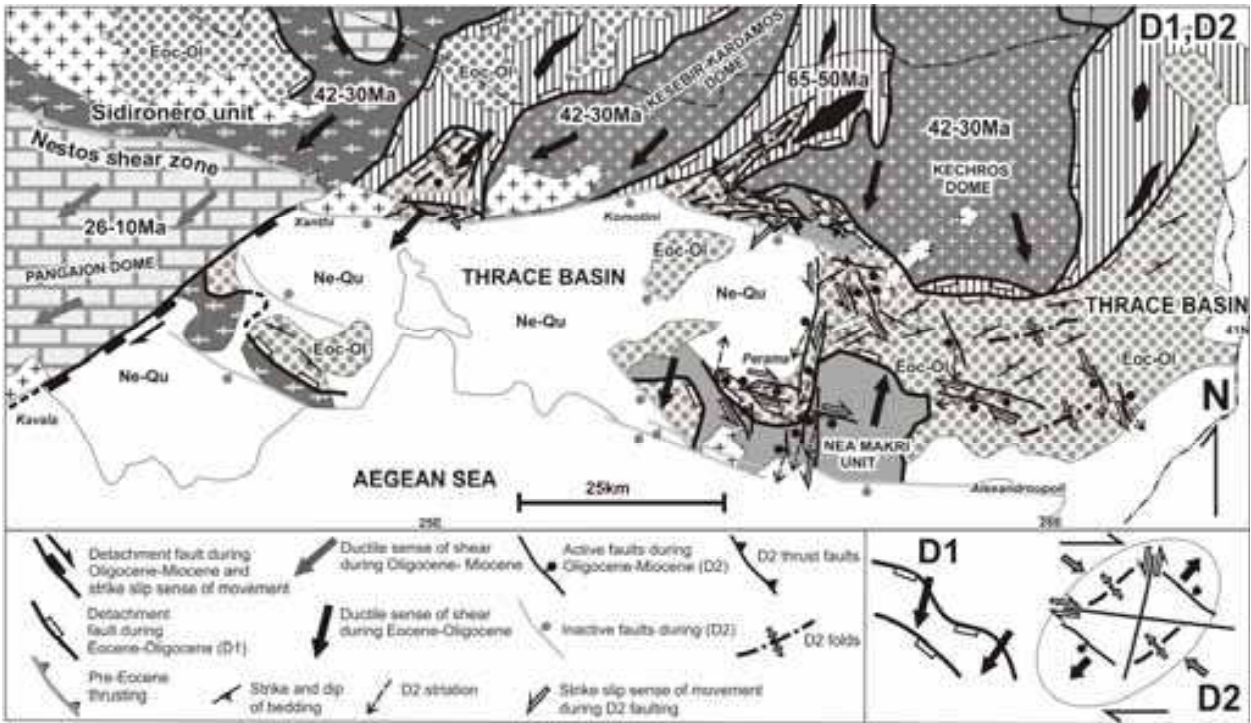

Fig. 6. Main structures and kinematics of D1 and D2 events. Ductile and brittle conditions take place simultaneously during D1 and D2 events at several tectonic crustal levels; ductile at deeper and brittle at the higher crustal levels. Legend as in Fig. 2

Reverse faults and open to tight folds in between the Paleogene deposits of the Thrace basin (Fig. 12) were interpreted from us as simultaneous to the D2 strike-slip structures as concluded from their geometry of kinematics, as well as structural and stratigraphic relationships. They are developed parallel to D2 extension and are also related to horizontal shortening perpendicular to D2 stretching (Fig. 5, 6). Overthrusting of the metamorphic

basement rocks (Rhodope basement or Nea Makri metasediments) on the Paleogene strata of the Thrace basin along NE-SW to E-W trending thrust faults associates usually the D2 compressional structures (Fig. 5, 6).

Vertical or high angle dipping Paleogene layers of the basin should be regarded as a result of D2 compressional overprinting (Fig. 6, 4). Thus, a series of NE-SW trending rest basins are formed as a result of the NW-SE to N-S directed compressional component (Fig. 2, 6).

Furthermore, NW-SE directed veins and normal faults with an important downward dip-slip movement in between the Palaogene sediments develop simultaneously with the others described D2 structures. The kinematic and dynamic compatibility of all these brittle structures show their contemporaneous origin. Moreover the infilling of the NW-SE directed veins with products of the Oligocene magmatism, some of which are also gold bearing, indicates clearly that at least D2 extension was contemporaneous with magmatism (Kilias & Mountrakis, 1998; Christofides et al., 2001). The geometry and kinematics of the described complicated network of the D2 structures suggest that during the D2 event NE-SW to ENE-WSW stretching took place simultaneously with NW-SE to NNW-SSE shortening (Fig. 5, 6).

The paleostress analysis of D2 fault-slip data estimated the orientation of the lower most $\sigma3$-stress axis to a NE-SW to ENE-WSW strike, dipping with low angle and the maximum $\sigma1$-axis subhorizontal, acting perpendicularly (Fig. 5, 13).
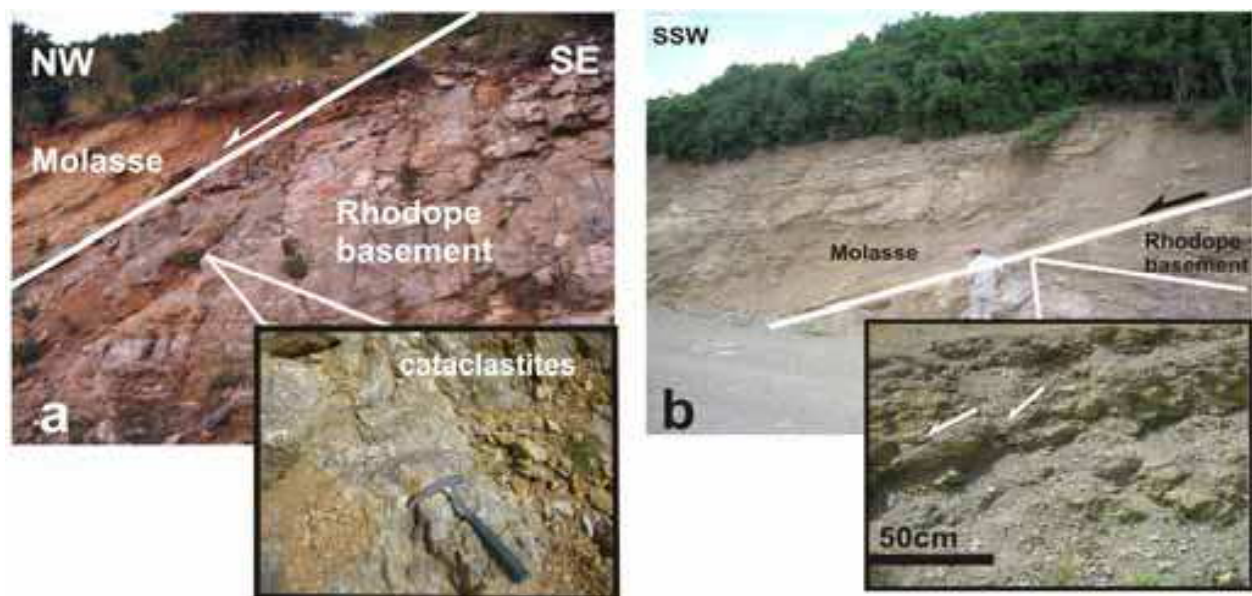

Fig. 7. a. The detachment fault zone between Rhodope basement rocks and molassic sediments with the development of a strong cataclastic domain at the footwall basement rocks (inset photo). Road to Trikorfo. b.Looking W, the low angle normal fault zone between Rhodope basement rocks and molassic sediments northern of Darmeni village. Shear bands in the footwall basement rocks showing a top to SSW sense of shear (inset photo)

3.3 D3 event

D3 is recognized from the occurrence of a second younger, oblique to ca. dip-slip striation on the planes of the D2 strike-slip faults (Fig. 5,11, 8a), showing a reactivation of the D2

structures during D3. On the other hand, D3 is characterized by large WNW-ESE to NNW-SSE trending, oblique to dip-slip, normal faults that dismember the Paleogene basin and are related to the evolution of the Neogene deposits of the Thrace basin (Fig. 11, 14). The D3 normal faults were developed subparallel to the D2 dip-slip normal faults and veins but are distinguished from that because D3 faults bound the Neogene subbasins forming usually steep slopes in between Neogene sediments and the Paleogene strata or the metamorphic basement rocks. (Fig. 11, 14).

Fault plane analyses yielded subvertical σ1- and subhorizontal ENE-WSW to E-W trending σ3-axes (Fig. 5, 15). In some places, D3 kinematics coincides with the kinematics of the D2 event. Nevertheless, D3 and D2 can be easily distinguished because the D2 strike-slip faults exhibit a progressively oblique to about dip-slip D3 component of slide, as well as by the fact that D3 normal faults border the Neogene basins (Fig. 5, 11, 14).

3.4 D4 event

The D4 event is related to large WNW-ESE to NE-SW fault zones some of which are older and were reactivated during D4. This is indicated by the existence of at least two striations generations on their fault planes (Fig. 5, 11, 14), with the younger one to be compatible with the D4-kinematics. The Kavala-Xanthi, Xanthi-Komotini, Komotini-Nea Santa, Aisymi-Soufli and Maronia-Alexandroupoli fault zones belong to the main fault structures of the D4 event (Fig. 5, 11, 14). On their fault planes the older D2 or D3 striations were clearly overprinted by the D4 kinematics. The D4 striation is mainly developed as oblique lineation with an important dextral or sinistral strike-slip component. Pitch angle was measured equal to 50° -75° to the E or W, being dependent on the dip direction of the fault plane (Fig. 5, 11, 14).
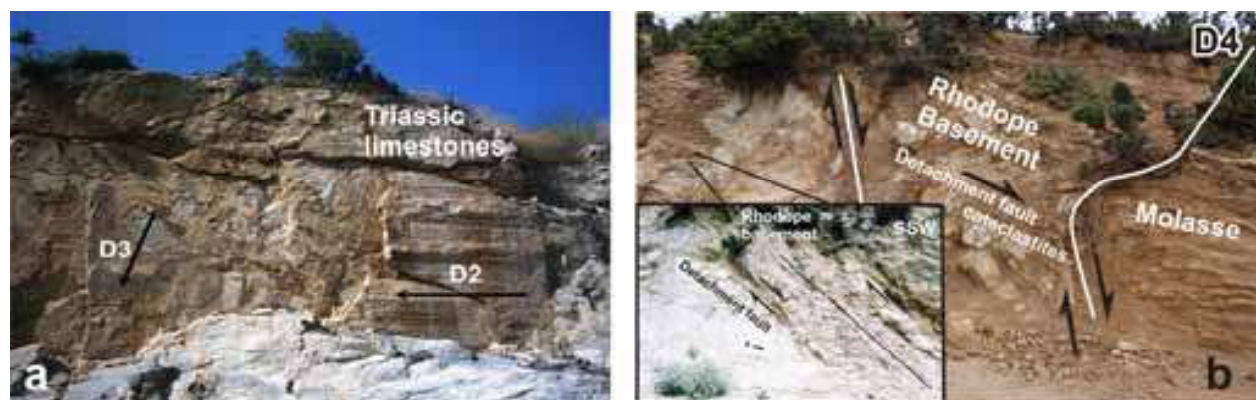

Fig. 8. a. D2 strike-slip and D3 oblique to dip-slip slickenlines on the same fault plane. (Egnatia road from Komotini to Alexandroupoli). b. D4 high angle normal faults cut the cataclastic domain of the low angle normal detachment fault zone related to initial opening of the Thrace basin. Inset photo: the cataclastic domain of the detachment fault zone. Well developed shear bands and kinematic indicators show a downward top to SSW sense of movement (near Ippiko village)

In many places, the D4 normal faults cut with a high angle the cataclastic, pseudotachylite bearing domain related to the D1 low angle normal detachment fault (Fig. 8b). In these cases, due to the uplift of the footwall fault segment and the intensive erosion of the overlying strata, the related detachment fault zone cataclastites are exhumed. The latter

form a cataclastic gouge-bearing cap, below of which the top to the SW-SSW D1 mylonitic, downward shearing of the metamorphic basement rocks dominates (Fig. 4, 9,14; Kilias & Mountrakis, 1990; Kilias et al., 1999; Krohe & Mposkos, 2002). NW-SE trending oblique thrust faults, as well as NW-SE trending, open, commonly angular folds, younger than the D2 contractional structures, are interpreted as structures related to the D4 kinematic setting. They are linked with a ca. ENE-WSW compressional component that coincides with the D4 dynamics.
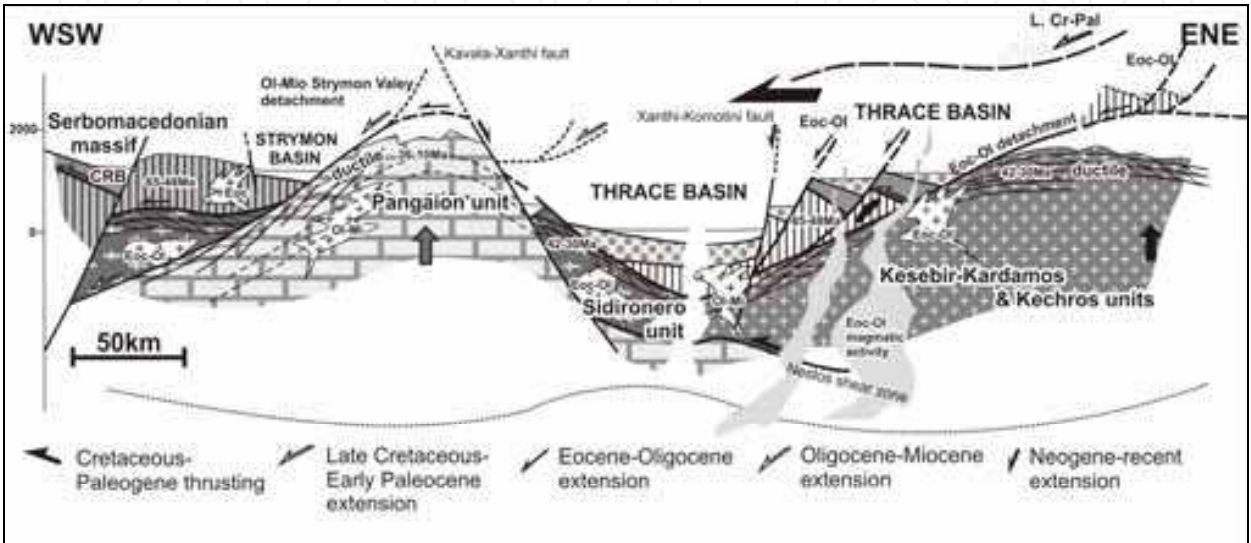

Fig. 9. Geometry of kinematics during Paleogene-Neogene extension of Rhodope Massif and formation of the Paleogene volcano-sedimentary Thrace basin on top of the Paleogene extensional detachment fault system, simultaneously with uplift of deep crustal metamorphic Rhodope units. Extension migrates towards SW-SSW.
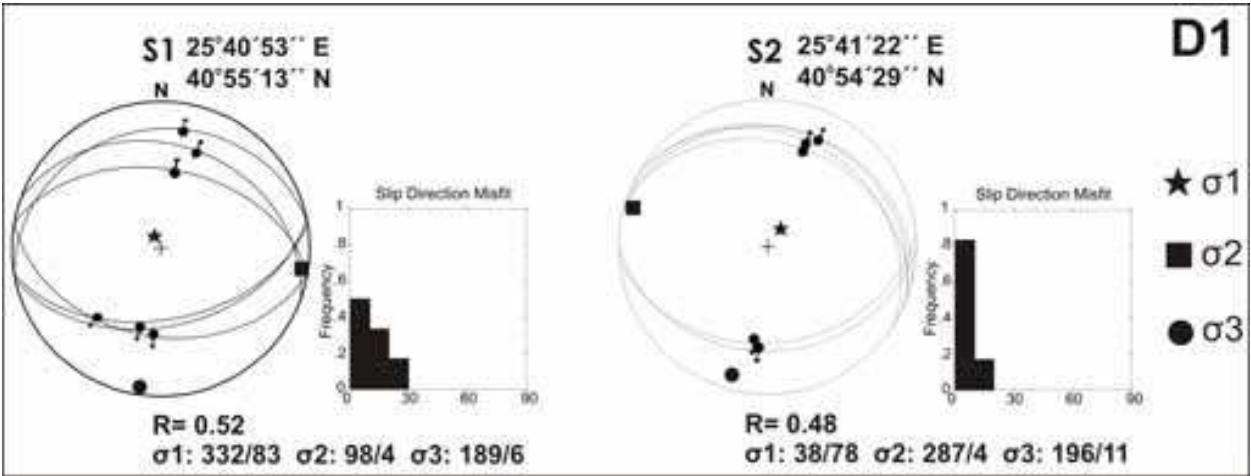Legend as in Fig. 2

Fig. 10. Faults–striae data and deduced paleostress field (σ1>σ2>σ3) orientation during D1 (equal area, lower hemisphere). Fluctuation histograms of deviation angles (angle between the calculated slip vector and the measured slickenside) and stress ratio R=σ2-σ3/σ1-σ3 are shown. The geographic coordinates for each station are also indicated

Paleostress analysis from the D4 fault-striae data revealed the minimum σ3-axis with NNW-SSE to NNE-SSW strike and the maximum σ1-axis with ENE-WSW to ESE-WNW strike both with a low to moderate dip angle, respectively (Fig. 5, 16).

3.5 D5 event

Some of the D4 fault zones remain active until recent time as it is shown by the international accepted geological and seismotectonic criteria (Hancoock, 1985; Kilias et al., 2006; Mountrakis et al., 2006; Gkarlaouni et al., 2007). They form large active faults reactivated during the present stress field of the area (Fig. 5, 11, 14). On their fault planes, the old generations of slickensides together with the younger ones of the D5 striations are usually recognized. The latter dip mainly to the SE with a high to moderate angle of dipping (i.e., pitch 50o-80o to the E on fault planes dipping to the S-SSW; Fig. 5, 11, 14).
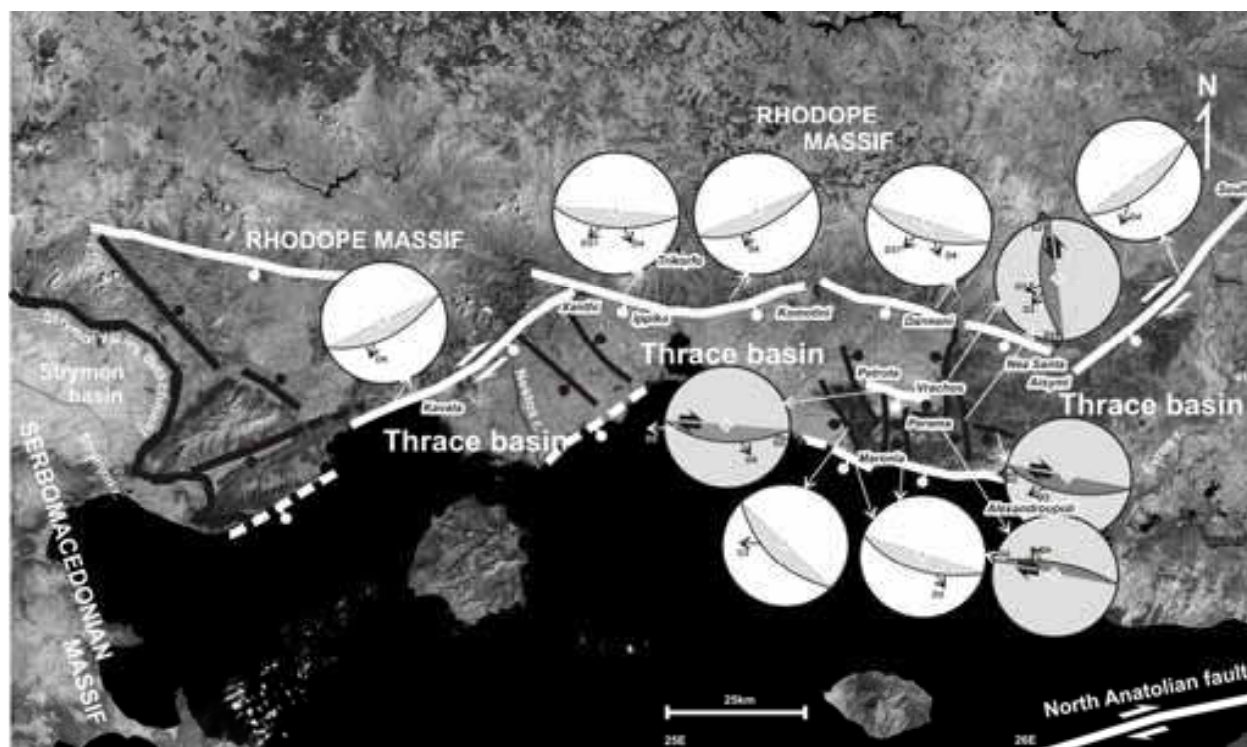

Fig. 11. Shaded relief map with the surface traces of the main fault zones mapped in the study area; black lines, faults related to the D2 event with their possible reactivations, white lines, faults related to the D3, D4 and D5 events. The main geometry of kinematics for each fault zone is shown in the stereographic diagrams (equal area, lower hemisphere). Arrows in the diagrams indicate the main development of the slickensides on the fault planes and their relative dating

Representative D5 active fault zones form the Maronia – Alexandroupoli fault zone, consisting of two main subparallel ca. E-W trending fault segments, and the Petrota-Vrachos fault with the same geometry of kinematics with the Maronia-Alexandroupoli fault (Fig. 11). According to their visible length and taking into account the relationship between fault length and earthquake magnitude (Papazachos et al., 2004), the most probable earthquake magnitude of the reactivation of the active fault zones, either as continuous structures or

separated into smaller segments, as they are shown on the geological maps of Fig. 11, 14, is ranged between M=3,9 and M=5,3.
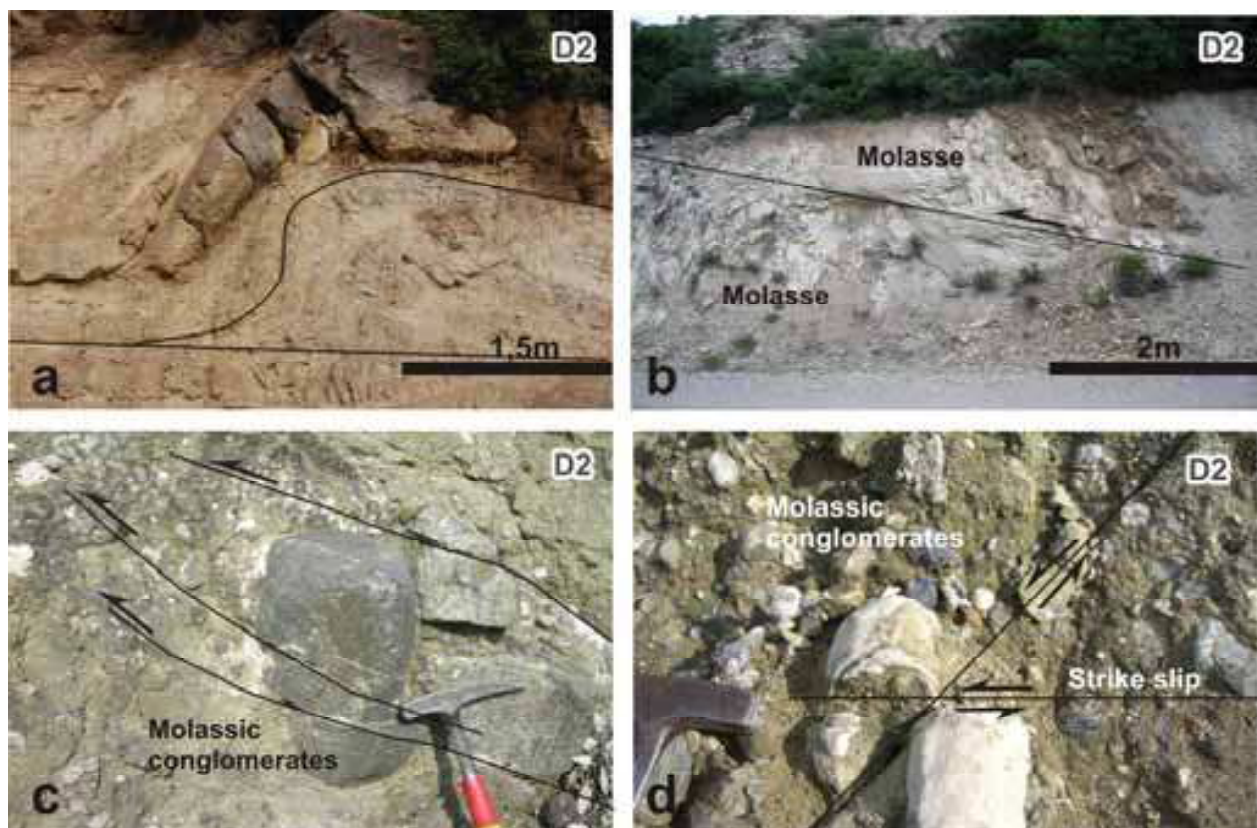

Fig. 12. During D2 kinematics; contractional structures (a,b & c) and strike–slip movements (d)

The paleostress analysis of the D5 faults revealed a minimum σ3-stress axis subhorizontal with NNE-SSW strike (i.e., 10° to 30° strike; Fig. 5,17) which coincides exactly with the active extension in the broader area, as it is concluded from the focal mechanisms of the strong earthquakes (Papazachos et al., 1998).

4. Age of deformation

Aiming to determine the time of deformation and its progression, the stratigraphic succession of the basin's strata and overprinting structural criteria, as well as the record of the structures evolution in certain layers in the whole stratigraphic column of the basin, have been considered. Additionally, the existing geochronological data for the ductile deformation recognized on the footwall segment of the detachment fault zone were also taken into account (Liati & Gebauer, 1999; Dinter et al., 1995; Wawrenitz & Krohe, 1998).

4.1 D1 event
For the D1 event related to the initial opening of the Thrace basin and the orogenic collapse along normal detachment faults (Kilias et al., 1999; Krohe & Mposkos, 2002; Bonev et al., 2006; Brun & Sokoutis, 2007) an age of Middle-Late Eocene is concluded (Fig. 5, 9). This age is indicated by the syndetachment deposition of the basal Middle-Late Eocene clastic sediments of the basin, supporting indeed the initiation of the extension during the Middle-Late

Eocene time (Kilias et al., 1999; Burchfiel et al., 2003; Brun & Sokoutis, 2007). Ductile fabrics of the metamorphic basement rocks at the footwall segment of the detachment fault with the same kinematic sense to the hangingwall have also been dated of Eocene age (Wawrenitz & Krohe, 1998; Liati & Gebauer, 1999; Krohe & Mposkos, 2002).
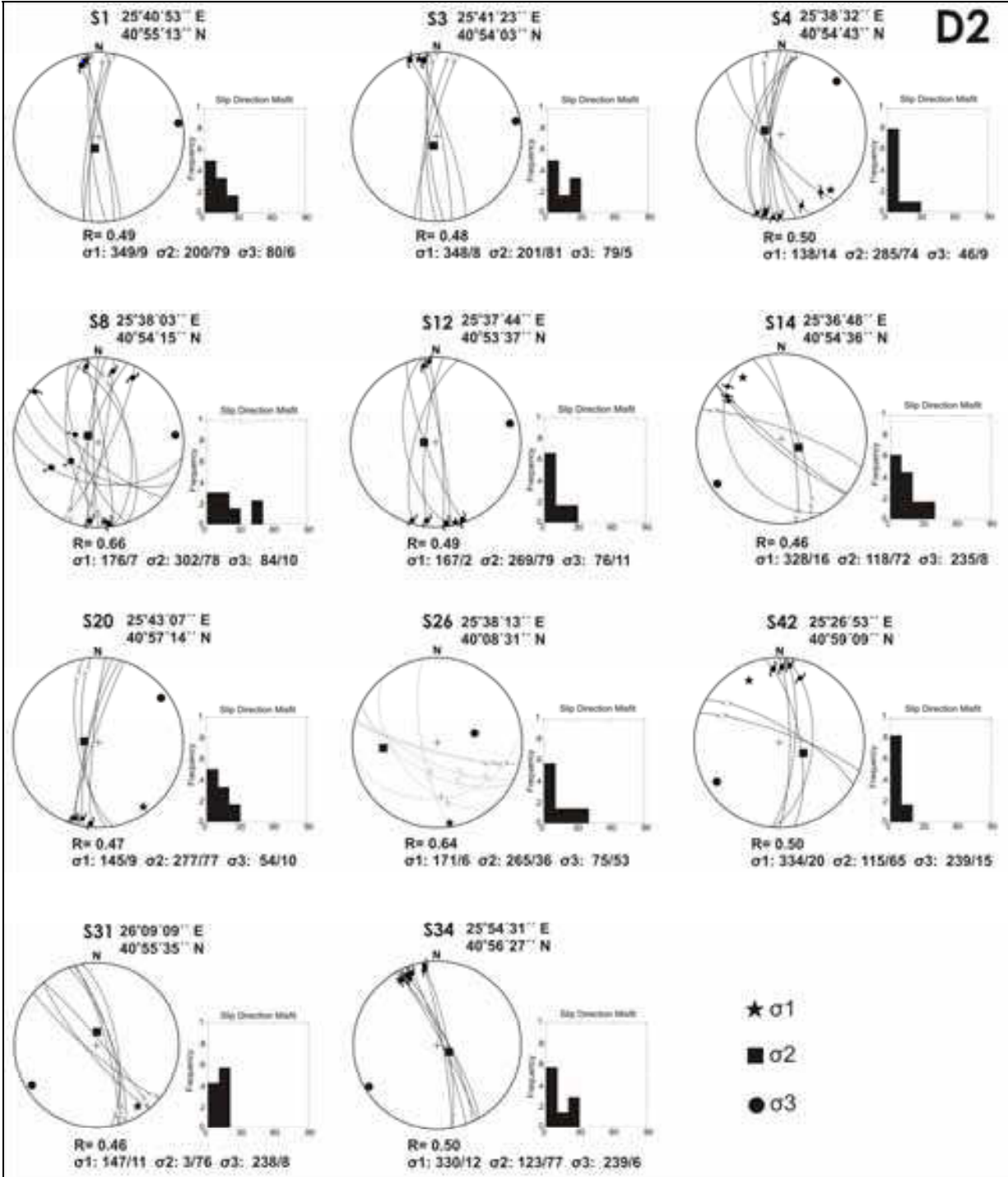

Fig. 13. Faults–striae data and deduced paleostress field (σ1>σ2>σ3) orientation during D2 (equal area, lower hemisphere). Fluctuation histograms of deviation angles (angle between the calculated slip vector and the measured slickenside) and stress ratio R=σ2-σ3/σ1-σ3 are shown. The geographic coordinates for each station are also indicated

However, at the same time other parts of the Rhodope nappe pile (e.g. Kimi and Nea Makri units) were already exhumed (Mposkos & Krohe, 2006; Bonev et al., 2006) and behaved as brittle segments on which the sedimentary sequence of the Thrace basin initially deposited (Fig. 5, 9).
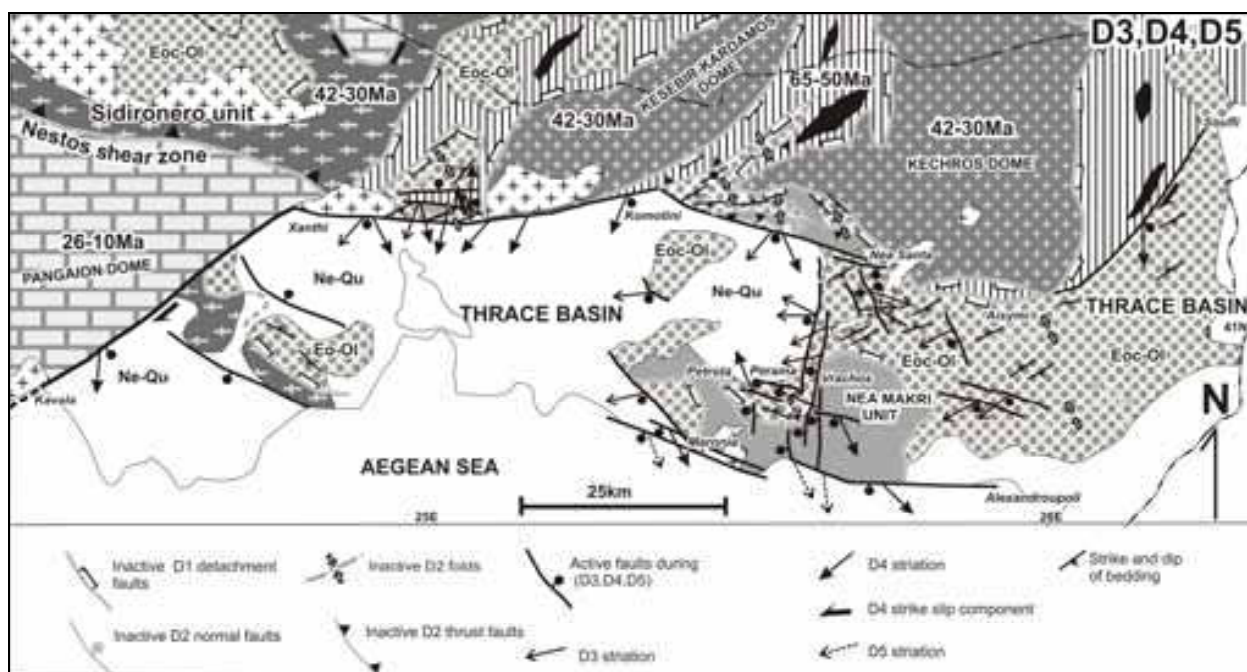

Fig. 14. Main structures and kinematics of D3, D4 and D5 events. Brittle conditions dominate in whole Rhodope region (including metamorphic basement and volcano-sedimentary series of the Thrace Basin). Legend as in Fig. 2

4.2 D2-event

D2-structures cut or overprint clearly the Eocene-Oligocene strata of Thrace basin but do not affect the Neogene sediments of the basin that lie unconformably on the Paleogene suite. In this case, a Late Oligocene-Early Miocene age could be supposed for D2 deformation having taken place under a brittle conditions regime. Moreover, ductile mylonitic fabric of the same Oligocene-Miocene age is referred for the lowermost Pangaion unit at the western edge of the Rhodope province (Kilias & Mountrakis, 1990; Dinter et al., 1995; Wawrenitz & Krohe, 1998; Kilias et al., 1999). This is related to the Oligocene-Miocene SW-ward detachment of the Serbomacedonian massif and the exhumation of the Pangaion metamorphic core complex (Dinter & Royden, 1993; Sokoutis et al., 1993; Dinter et al., 1995; Kilias et al., 1999).

4.3 D3-event

D3 structures dismember the Paleogene volcanosedimentary sequence of the Thrace basin into NW-SE to WNW-ESE trending Neogene basins (Fig. 5, 9, 14). D3 normal faults evolved as synsedimentary structures in relation to the Neogene basins deposition, as it is shown by the tilting of the older Miocene strata towards the fault planes and the overlying horizontal to subhorizontal younger Pliocene sediments which occur in direct contact with the fault planes. Thus, the age of D3 structures should range in between Middle Miocene and

Pliocene time and certainly before D4-structures as the latter clearly overprint the D3-structures.

4.4 D4 event

The complicated network of the D4 structures is recognized to affect the deposits up to the Pliocene and is related to different geometry of kinematics from the former D3 deformation. Therefore, at least a Pliocene age can be concluded for the activation of D4 structures and clearly just subsequent to the D3 event (Fig. 5, 14).

4.5 D5-event

This event concerns the last up to the present active structures, some of which are related to historical earthquakes in the broader Thrace area (Papazachos et al., 1998) and thus its age is ranged between Pleistocene up to present, postdating the D4 deformation.

5. Geodynamic implications-basin evolution

Our detailed tectonic analysis and geological mapping, on both meso- and micro-scale, of the Thrace basin in Greek mainland allowed to constrain the geometry and kinematic of deformation and the tectonic position of the basin, as well as its relationships to the Cenozoic unroofing and exhumation history of the Rhodope metamorphic deep crustal rocks.

The stratigraphic and structural data indicate initially, during Middle-Late Eocene time clastic deposition above a detachment fault zone, continued until Late Oligocene time by molassic type sedimentation. It is linked with the Tertiary NE-SW extension and collapse, which affected the whole Rhodope province (Dinter & Royden, 1993; Kilias et al., 1999; Krohe & Mposkos, 2002; Brunn & Sokoutis, 2007; Bonev et al., 2006; Georgiev et al., 2010) after Cretaceous-Paleocene nappe stacking and crustal thickening (Fig. 5, 9; Burg et al., 1996; Liati & Gebauer, 1999; Jahn-Awe et al., 2010).

The position of the Thrace basin deposits above the Eocene detachment fault zone of the collapsed Rhodope nappe stack allows us to define the Thrace basin as a Paleogene supradetachment basin or as basin above an asymmetrical collapsed volcanic arc, taking in to account the abundant magmatic products in between the basin deposits (Fig. 5, 9; Einselie, 2000; Frisch & Meschede, 2007). As a supradetachment basin, it could be also seen as a type of "piggy-back" basin formed above the primary bounding detachment fault. According to our data the interpretation of the Thrace basin as a forearc basin (Görür & Okay 1996; Tranos, 2009) should be challenged. The interpretation of the Thrace basin as a forearc basin was mainly based on the occurrence along its southern margin, of a sequence of chaotic deposits interpreted as a tectonic melange developed in an accretionary wedge. However, this tectonic melange may represent olistoliths in the Paleogene sediments of the basin.

Basin subsidence during the Paleogene took place simultaneously with the exhumation of the metamorphic core complexes of the Eastern Rhodope massif (e.g., the Kesebir-Kardamos and Kechros domes; Fig. 5, 9; Krohe & Mposkos, 2002; Bonev et al., 2006; Georgiev et al., 2010). A series of smaller Paleogene basins that extended in the broader Rhodope-Serbomacedonian province in Bulgaria FYROM or Turkey (Fig. 2) should be interpreted as rift basins related to the same Paleogene kinematic setting and normal detachment faulting

(e.g., Mesta half graben in Bulgaria; Buchfiel et al., 2003) or other basins of Paleogene age in FYROM, Bulgaria and Turkey (Okay et al., 1990; Turgut & Eselle 2000; Dumurdzanov et al., 2005; Bonev et al., 2006).
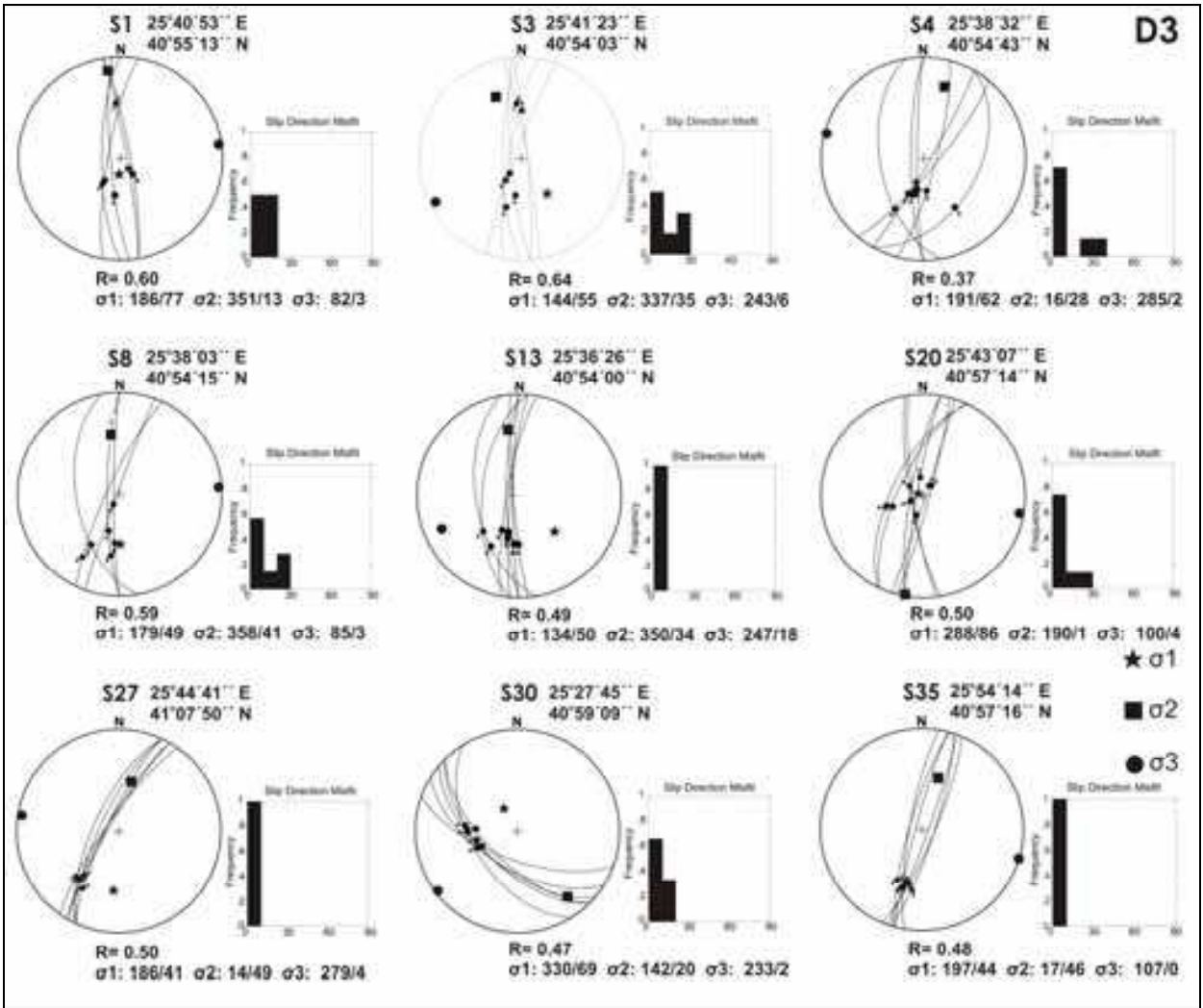

Fig. 15. Faults–striae data and deduced paleostress field (σ1>σ2>σ3) orientation during D3 (equal area, lower hemisphere). Fluctuation histograms of deviation angles (angle between the calculated slip vector and the measured slickenside) and stress ratio R=σ2-σ3/σ1-σ3 are shown. The geographic coordinates for each station are also indicated

The second D2 deformation affected the Paleogene strata of the Thrace basin, dismembering the basin during Oligocene-Miocene time. Contraction perpendicular to the NE-SW to NNE-SSW trending main extension also continued during that time. Compressional structures parallel to the extension (X-axis of the finite strain ellipsoid) associated with strike-slip fault zones and tension fractures are interpreted as contemporaneous formed structures. Compression can act parallel to the Y-axis of the finite strain ellipsoid simultaneously with a main subhorizontal extension. In this case, the 3-D deformation is thought to be a constrictional type. On the other hand, the described geometrical relationships between reverse, normal and strike-slip faults of the same age could be explained by a transpressional type of deformation in accordance with which a strike-slip movement acts

simultaneously with a pure shear component or simultaneous action of a non-coaxial and coaxial component of deformation (Sanderson & Marchini, 1984).
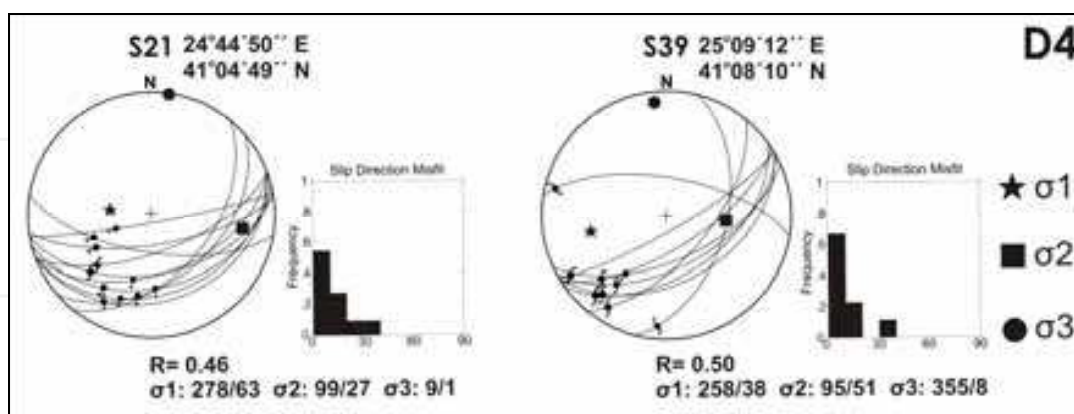

Fig. 16. Faults–striae data and deduced paleostress field (σ1>σ2>σ3) orientation during D4 (equal area, lower hemisphere). Fluctuation histograms of deviation angles (angle between the calculated slip vector and the measured slickenside) and stress ratio R=σ2-σ3/σ1-σ3 are shown. The geographic coordinates for each station are also indicated

The Oligocene-Miocene D2 brittle tectonics of the upper crustal parts in Eastern Rhodopes took place simultaneously with ductile deformation in the lower- most Pangaion metamorphic core complex in Western Rhodopes related to the collapse and tectonic denudation of the upper Rhodope tectonic nappes along the Strymon-Valey normal detachment fault zone (Dinter & Royden, 1993; Kilias et al., 1999). Oligocene-Miocene ductile deformation indicates the same kinematic symmetry with D2 brittle event at the Eastern structurally higher Rhodope nappes (Kilias & Mountrakis, 1990; Sokoutis et al., 1993; Dinter & Royden, 1993; Dinter et al., 1995; Kilias et al., 1999; Brun & Sokoutis 2007).The Strymon basin at the western edge of the Pangaion metamorphic core complex evolved as a syndetachment rift basin (Dinter & Royden, 1993; Dinter et al., 1995; Sokoutis et al., 1993) or as an Oligocene-Miocene supradetachment basin, above the Strymon-Valey detachment fault zone (Fig. 2, 9; Friedmann & Burbank, 1995).

According to our descriptions, a progressive migration of the exhumation towards W-SW through time can be concluded during Tertiary time due to successive Eocene and Oligocene-Miocene stages of extension and metamorphic core complexes formation. Firstly, the structurally higher Rhodope metamorphic units were exhumed during the Eocene, and then the structurally lower units during Oligocene-Miocene time. Therefore, a multi stage extensional unroofing is suggested for the several Rhodope metamorphic units.

The same conclusions for a progressive formation and subsidence of the supradetachment basins in space and time are extracted. Initially, the Paleogene Thrace basin at the Eastern Rhodopes was opened, and then followed the opening of the Neogene Strymon basin at the Western margin of the Rhodope province. This gradual evolution is fairly consistent with the main W-SW-wards sense of movement of the tectonic hangingwall during the Tertiary orogenic collapse of the Rhodope nappe stack, at least for the parts in the Greek mainland (Kilias & Mountrakis, 1990; Sokoutis et al., 1993; Dinter & Royden, 1993; Kilias et al., 1999; Brun & Sokoutis, 2007).

The next D3 deformational stage, during the Miocene-Pliocene involved intense dismemberment of the Thrace basin into Neogene basins, filled with terrigene sediments,

covered unconformably the Paleogene strata. NE-SW to E-W extension, under brittle conditions, common for all Rhodope province, was related to the Neogene basins subsidence, as well as the finally exhumation of the lowermost Pangaion metamorphic core complex.

During the Pliocene the extension continued with NNW-SSE to NNE-SSW strike (D4 event) with simultaneous strike-slip and dip-slip movements taking place. Shortening structures in the Pliocene strata, related to the D4 kinematics, are of local importance. The same complicated kinematic geometry and faults network is also revealed for the older D2 deformation, indicating that an important strike-slip component was active during the overall Tertiary structural evolution of the Rhodope province. Co-existence of strike-slip faults with normal and reverse fault zones is common for many areas (Hancock, 1985; Ratschbacher et al., 1989; Kilias et al., 1999).
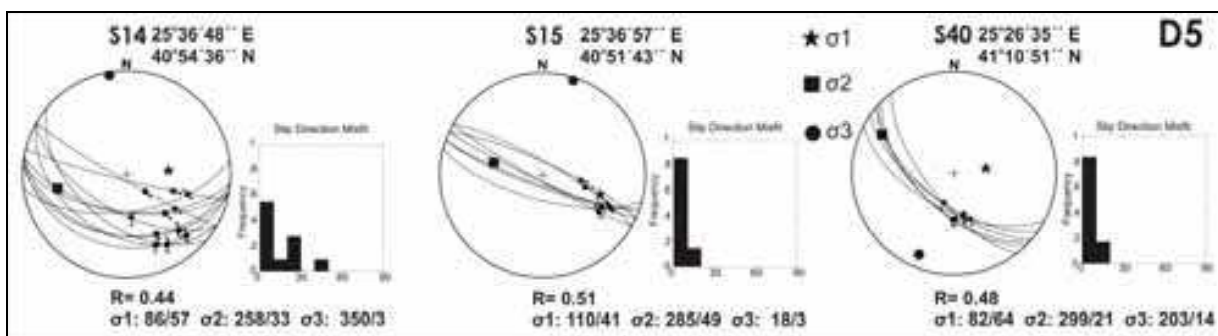

Fig. 17. Faults–striae data and deduced paleostress field (σ1>σ2>σ3) orientation during D5 (equal area, lower hemisphere). Fluctuation histograms of deviation angles (angle between the calculated slip vector and the measured slickenside) and stress ratio R=σ2-σ3/σ1-σ3 are shown. The geographic coordinates for each station are also indicated

Extension remains active with a little change of its trend until present, now with N-S to NNE-SSW trend (D5 event), related to the active faults of the study area. D5 dynamic is well coincided with the dynamic of the large active dextral strike-slip Anatolian fault zone (Papazachos et al., 1988).

Basin subsidence and deposition during the Paleogene were associated with intense magmatic activity (Kilias & Mountrakis, 1998; Marchev et al., 2005; Bonev et al., 2006). The origin of the Late Eocene-Oligocene (34-26 Ma) syndepositional magmatic activity (Christofides et al., 2001; Kilias et al., 2006) could be attributed to the subduction processes evolved during Paleogene time further to the W-SW, in the Olympos-Ossa and Cyclades areas, due to subduction of a thin continental crust or parts of the Pindos oceanic lithosphere beneath the Pelagonian continent (Fig. 1, 18; Schermer et al., 1990; Kilias et al., 1991; Robertson et al., 1996; Lips et al., 1998). However, the remnant of the subducted slab of the Eastern margin of the opened, until the Tertiary time, Vardar-Axios ocean (Brown & Robertson, 2004; Sharp & Robertson, 2006) under the Internal Hellenides could also be the driving mechanism for the Tertiary magmatic activity in the Rhodope province, simultaneously with the opening and sedimentation of the Thrace basin (Fig. 1, 18).

However, Marchev et al., (2005) explain the origin of the Paleogene magmatism and the simultaneous extension and crustal thinning of the Rhodope continental crust as having been due to convective removal of the lithosphere and mantle diapirism. A close analogy

was found to the magmatic activity and orogenic collapse taken place in the Menderes massif in SW-ern Turkey during Eocene-Miocene time that resulted in a series of exhumation events (Lips et al., 2001). Furthermore, Bonev et al., (2006) suggest that extension in the Rhodope metamorphic belt initiated a little earlier than the Eocene, during Early Paleogene time, with syndetachment deposition of Paleocene-Early Eocene clastic sediments in the Krumovgrad half graben basin at the SW Bulgaria (Fig. 2, 9).
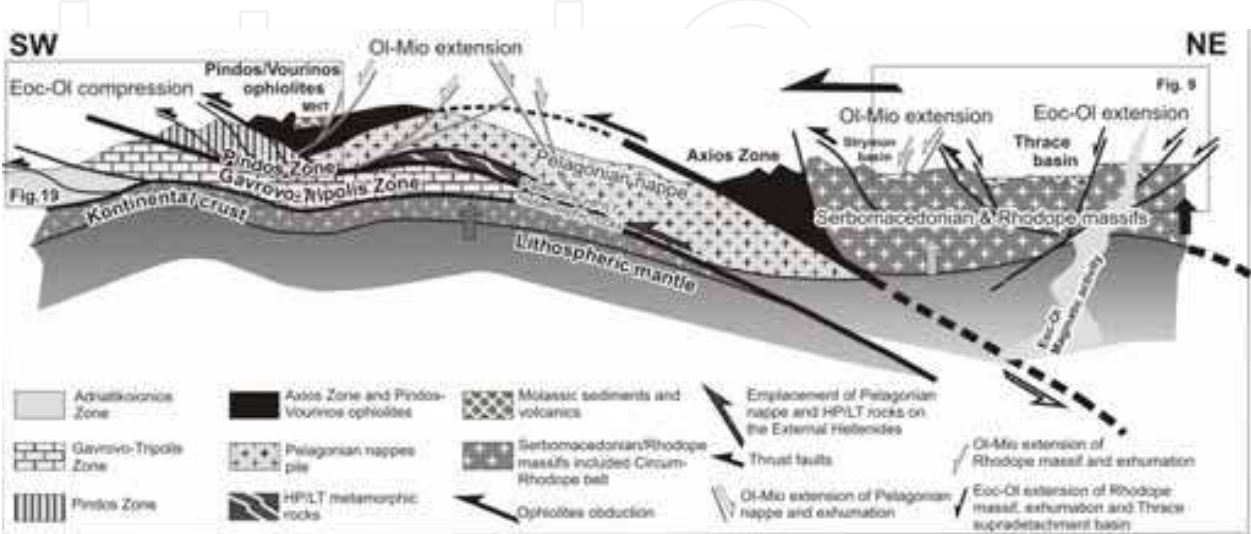

Fig. 18. Schematic lithospheric-scale cross-section through the Hellenides during Paleogene. The future extensional structures leaded to the finally Oligocene-Miocene exhumation of the Hellenides units and basins subsidence are respectively shown. Contraction tectonic and nappe stacking associated with HP/LT metamorphism dominate during Paleogene in External Hellenides and Pelagonian nappes pile whereas extension and orogens collapse take place simultaneously in Internal Hellenides (Serbomacedonian and Rhodope massifs)
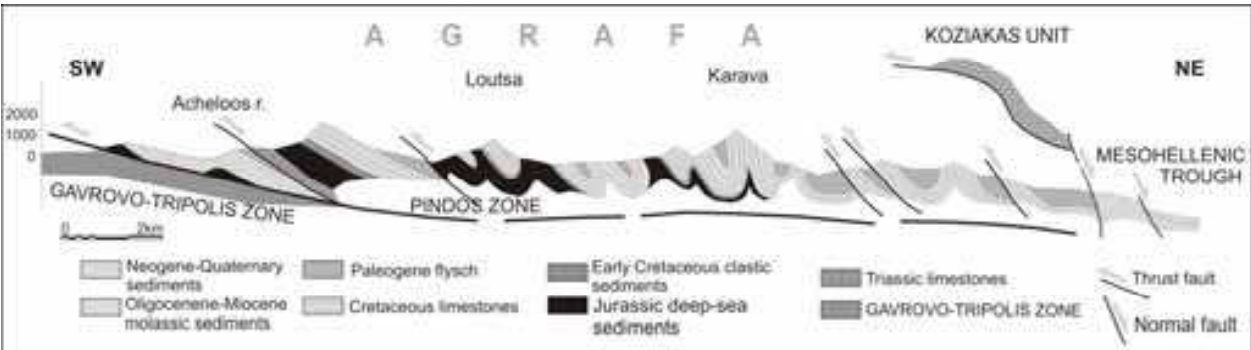

Fig. 19. SW ward thrusting during Paleogene thin skinned tectonic of External Hellenides thrust belt takes place simultaneously with extension and orogenic collapse at the inner parts of Hellenides

Contraction, nappe stacking and crustal thickening associated with HP/LT metamorphism take place simultaneously with the described Rhodope Tertiary extension at the suture zone between External Hellenides and Pelagonian nappe pile more W-wards (Fig. 18; Godfriaux, 1968; Schermer et al., 1990; Kilias et al., 1991). This structural setting at the specific orogen part of Hellenides was further followed by extensional orogenic collapse and crustal

thinning during the Oligocene-Miocene, while compression migrated towards the SW to the more external parts of Hellenides at the Hellenic foreland (Fig. 18, 19; Schermer et al., 1990; Kilias et al., 2002, 1999).

In conclusion, in the Hellenic orogen during Tertiary time, a progressive migration towards W to SW, of compression vs. extension under a continuous plate convergence regime, is well established. As a result, the deep crustal levels were successively exhumed during extensional tectonics (e.g., Rhodope metamorphic core complexes, Olympos-Ossa unit, Pelion, Cyclades and Crete complexes), simultaneously with burial of lithospheric material and nappe stacking at the deformed orogenic front towards SW (Fig. 18, 19; Kilias et al., 1999, 2002; Jolivet & Brun, 2010; Ring et al., 2010). Supradetachment basins were successively formed at the internal Hellenides above the extensional detachment fault zones and the footwall segment of the collapsed nappe stack.

The dynamic migrating system extension vs. compression in the Hellenides during the Tertiary time, could be ascribed to, (I) successive subduction processes towards the SW-SSW under oblique plate convergence conditions (Karfakis & Doutsos, 1995; Kilias et al., 1999, 2002; Vamvaka et al., 2006; Schmid et al., 2008), (II) periodic changes in the convergence rate between Africa and European plates, combined with a retreat of the subduction zone towards SW (Royden, 1993, Kilias & Mountrakis, 1998; Dumurdzanov et al., 2005; Papanikolaou, 2009; Jolivet & Brun, 2010; Ring et al., 2010), as well as (III) dynamic instabilities and gravitational collapse of the over-thickened Hellenic accretionary prism (Platt, 1993) or the clockwise rotation of Western External Hellenides away from more Eastern parts of the Internal Hellenides (Kontopoulou, 1986; Kissel & Lai, 1998).

6. Conclusions

The structure and sedimentation of the Thrace basin in Northwestern Greek mainland indicate that NE-SW extension initiated during Middle-Late Eocene time. The Thrace basin evolved during the Eocene-Oligocene as a supradetachment basin above a normal detachment zone with a main W-SW direction of movement of the hangingwall (D1).

The Paleogene Thrace basin was affected during Oligocene-Miocene time by contemporaneous activity of large strike-slip, reverse and normal faults, as well as folds (D2). Extension was trended NE-SW to NNE-SSW almost subhorizontally while compression NW-SE to WNW-ESE, also subhorizontally. During Miocene-Pliocene, Neogene sub-basins developed unconformably over the Paleogene strata bounded by high angle dipping normal faults linked to NE-SW to E-W extension (D3). The next deformational stage (D4) produced simultaneously during the Pliocene local oblique shortening structures and normal faults related to NNW-SSE to NNE-SSW extension and ca. ENE-WSW compression. Some of the D4 normal faults remain active until present (D5) coinciding with the active tectonic dynamics of the broader Rhodope province and the North Aegean region. Tertiary basin subsidence and structural evolution was associated by unroofing and successively, multistage extensional exhumation of the deep crustal metamorphic Rhodope levels, as well as intense calc-alkaline and locally shoshonitic magmatic activity. Exhumation and basin subsidence ages are recorded younger towards the W-WSW and they were linked to a general SW-ward migration of syn-orogenic extension related to an oblique plate convergence regime. Finally, from our descriptions about the structural evolution and geotectonic position of the Thrace basin is clearly concluded, apart from the industrial

interesting of the basin (gold bearing deposits, hydrocarbon reservoir), its important contribution to the general geodynamic knowledge.

7. Acknowledgments

8. References

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New Frontiers in Tectonic Research - At the Midst of Plate Convergence

Edited by Dr. Uri Schattner

Ocean closure involves a variety of converging tectonic processes that reshape shrinking basins, their adjacent margins and the entire earth underneath. Following continental breakup, margin formation and sediment accumulation, tectonics normally relaxes and the margins become passive for millions of years. However, when final convergence is at the gate, the passive days of any ocean and its margins are over or soon will be. The fate of the Mediterranean and Persian Gulf is seemingly known beforehand, as they are nestled in the midst of Africa-Arabia plate convergence with Eurasia. Over millions of years through the Cenozoic era they progressively shriveled, leaving only a glimpse of the Tethys Ocean. Eventually, the basins will adhere to the Alpine-Himalaya orogen and dissipate. This book focuses on a unique stage in the ocean closure process, when significant convergence already induced major deformations, yet the inter-plate basins and margins still record the geological history.

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Adamantios Kilias, George Falalakis, Aristides Sfeikos, Eleftheria Papadimitriou, Agni Vamvaka and Chara Gkarlaouni (2011). Architecture of Kinematics and Deformation History of the Tertiary Supradetachment Thrace Basin: Rhodope Province (NE Greece), New Frontiers in Tectonic Research - At the Midst of Plate Convergence, Dr. Uri Schattner (Ed.), ISBN: 978-953-307-594-5, InTech, Available from: http://www.intechopen.com/books/new-frontiers-in-tectonic-research-at-the-midst-of-plate-convergence/architecture-of-kinematics-and-deformation-history-of-the-tertiary-supradetachment-thrace-basin-rhod

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