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HISTORICAL EVOLUTION OF THE GEOLOGICAL RESEARCHES IN THE MENDERES MASSIF

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ABSTRACT.- The Pan-African basement of Menderes Massif is made up of homogenous paragneiss and schist units (metaclastic sequence) which are intruded by metagabbros and gneisses derived from different types of granites. The basement is unconformably overlain by lower Palaeozoic metaclastic series consisting of quartzite and metaconglomerate at the lowest level. They show a transition into schists, and the Palaeozoic sequence ends with Permo-Carboniferous black marbles of Göktepe formation. Both basement and Palaeozoic sequence are intruded by lower Triassic leucocratic granites which were converted into orthogneisses by Alpine metamorphism. The Mesozoic series of the Menderes Massif begins with upper Triassic meta-sandstone/metaconglomerate intercalation and continues with Jurassic to Cretaceous dolomites and massive marbles. Platform-type massive marbles with metabauxite lenses and well-preserved rudist fossils at the uppermost levels are conformably overlain by late Campanian - late Maastrichtian reddish pelagic marbles. Flysch-type middle Paleocene metaolistostrome forms the uppermost unit of cover series. The protoliths of clastic sediments of the Pan-African basement consisting of paragneiss and conformably overlying schist units were deposited on a passive continental margin. The zircon ages of the granitoids intersecting this clastic sequence are restricted to a time range between 570 and 520 Ma with an average of about 550 Ma. The polyphase metamorphic evolution of the basement under granulite (583 ± 5,7 Ma), eclogite (529,9 ± 22 Ma) and Barrowian-type medium-pressure conditions (average 540 Ma) are related with the Pan-African Orogeny. The isotopic data including the ages of detrital zircons (592-3229 Ma) of paragneiss and schist units, the intrusion ages of granitoids and the age of granulite facies metamorphism constrain an age for the deposition of protoliths of metaclastic sequence between 592-580 Ma, Lower Late Neoproterozoic. In Eocene time, Pan-African basement and cover series were affected by Barrowian-type Alpine metamorphism under greenschist, lower amphibolite facies conditions, traditionally called as the 'Main Menderes Metamorphism. However, new HP/LT evidence found in Mesozoic cover series reveals that this metamorphism is more complex than it was considered. It is generally accepted by many researchers that the exhumation of the Menderes Massif as a core complex is related with the extensional tectonic regime. It is assumed that there is a genetic relation between the intrusion of Eðrigöz - Koyunoba granites and the detachment fault in the northern part of the Menderes Massif. Furthermore; to the South of the Massif, it is suggested that the old thrust fault between the Menderes Massif and Lycian Nappes reworked as a detachment fault during the exhumation. The reason of extension in the region is the thermal weakness in the thickened crust made by the magmatism developed after the the collusion of Anatolide-Tauride platforms and the Sakarya continent. The intrusion ages of Eðrigöz and Koyunoba granitoids are 20-21 Ma and the detachment fault was active during 25-19 Ma. In addition, the symmetric core complex formation in the central submassif was carried out in the middle-late Miocene. Pliocene to recent active graben faults intersect the detachment faults.

Key words: Menderes Massif, Pan-African and Alpine metamorphism, young exhumation.

INTRODUCTION

The Menderes massif, which has an important role during the shaping of geological structure of the western Anatolia, takes place within the main subjects of the geological researches carried out in Anatolia for 150 years. Within this context; the colloquium held in Izmir, between 5th-10th of November 2007 regarding the stratigraphical and tectonical structures, condition of metamorphism, kinematics and ages, latest compressional and extensional tectonical regimes caused to gaining of the recent form of

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the massif that gave the recent outlook to massif, gave chance to the presenting of the latest findings about the massif. In the Colloquium, in which invitational conferences took place, common sides of geological evidences were specified that has been put forward so far in the Menderes Massif and future oriented approaches were exhibited to solve the problems being discussed. Within this paper, which is aimed at presenting the last scene of the researches made in the massif, documents that have been prepared within years and contributed to the geology of the massif will be mentioned first. Later on; which geological problems of the massif have nowadays been solved will be investigated. And in the last section, some proposals about the future geological studies will be presented.

HISTORICAL EVOLUTION OF RESEARCHES IN THE MASSIF

First historical definitions of the rocks of Menderes Massif belong to Hamilton (1841), Tchichatcheff (1867), in the map of Geology of Aegean Region attached to his book named as "Asia Mineure", has approximately localized the NW-SE boundaries of the region, known as the Menderes Massif at present, but has extended the massif to the south of the Sea of Marmara at the north. He defines that the massif consists of micaschist, gneiss, marble, amphibolites, some undefined rocks and less granite. He also points out that the same rocks present in Cyclades (Figure 1). Philipson (1911-1915), uses the name "Lydia Caria" Massif for the region which is named as Menderes Massif at present. He also mentions about the crystalline rocks at the center of the massif and of marbles and semi crystalline limestones towards the western southern regions. In the geological map, a zone is shown at the NW boundary of the massif, which approximately corresponds to the present Izmir-Ankara zone composing of clayey schist, graywacke, serpentine and diabase veins (Figure 2). It is emphasized that the fractures forming the young grabens in the massif might be Late Pliocene-Quaternary. Menderes Massif was first named by Parèjas (1940) and appeared in the 1/1.000.000 scaled geological map of Turkey prepared by Egeran and Yener (1944) as MTA publication. The Menderes Massif is shown as a distinct tectonical unit in Izmir sheet of this map.

Schuiling, who made the first systematic geological-petrographical investigation of the massif in 1962, defines the lithostratigraphical sequence and divides it into two main units as "core" and "cover". With keen eye detection, he also distinguishes the fine grained basic gneisser from classical augen gneisses of the massif which are defined as paragneiss in the recent maps of the "Çine submassif". However, locating the Caledonian Orogeny into core and cover series, aging the uppermost units of the Massif as Permo-Carboniferous and last metamorphism as Hercynien totally disagrees with actual evidences. Graciansky (1965), mentions about the presence of an unconformity between the basement and cover series in Menderes Massif on the basis of lineation, schistosity and inclusion measurements. However, the age of Triassic Milas marbles shown by question marks has been approved as Santonian-Campanian.

Brinkman (1967), extends the southern boundary of the Menderes Massif to the Gulf of Gökova. Whereas, this boundary ends with a unit distinguished as Kazýklý at present, and Kurin, Karaova and Gereme units in the southernmost part of the region are included into the Lycian Nappes. Besides, in recent studies different than Brinkman (1967), Milas, Kýzýlaðaç and Kazýklý units are aged as Santonian-Campanian, Late Campanian-Maastrichtian by fossil content, (Özer et al., 2001).

The first studies aiming the investigation of mineral facies in Bozdað and in its western parts were made by Ýzdar (1971) and Evirgen (1979). Evirgen (1979), claims that regional metamorphism has occurred under a pressure of 3,5-6,5 kb and at a temperature of 400-700 °C by basing

Figure 1- Geological map of the Aegean Region (Tchichatcheff, 1867). Explanations: th-m: Tertiary; th: Terrace and diluvian, pl: Marine pliocene, l: Levantine, p: Pontian, s: Sarmatian, m: Marine lower Miocene, e: Eocene; cr-f: Mesozoic, cr: Cretaceous, ti: Titonian, t: Triassic; c-de: Palaeozoic, c: Carboniferous, de: Lower Devonian, x: Metamorphic schist, mica schist, gneiss, marble, amphibolite (granite and non differentiated rocks), g: granite and granulite, sy: Syenite, diorite, Serpentinite, Andesite, trachyte, basalt etc., +: lignite basins, sea depth up to 500 m, sea depth up to 1000 m., Dip and strike of bed and general delineation of folds.

on index minerals in this study, although cataclastic rocks in the southern part of Gediz Graben were mapped, a direct relation of deformation with the north dipping and a low angle detachment fault could not be established.

Dürr 1975 determined the first rudist fossils in Milas marbles which is extending along the southern boundary of Menderes Massif. Based on the fossil content, he defined the age of the platform type marbles belonging to the cover

Figure 2- Geological map of the Western Anatolia (Philippson, 1911). Explanations: 1- Holocene and Pliostecene, n2-Young Tertiary, 3- Tuff bearing Tertiary, 4- Tuff and andesite bearing Tertiary, e5-Lower Tertiary, k6- undated limestone, kk7-Cretaceous formation limestone (partly upper Jurassic), t8- Triassic, ck9-Carboniferous and Permo - Carboniferous calcerous, S10- clayey schist, greywacke, diabase. Partly definite, partly probable Paleozoic, S+D11- clayey schist plenty of diabase and serpentine with veins, gl12- Mica schist and other crystalline schists (very little schistose gneiss), 913- Gneiss (very few micaschist), m14- Marble and semi-marble, B15-Basalt, A16- Andesite, trachyte and rhyolite, T17-Andesitic, trachytic and rhyolitic tuff, AYT18- Andesite and tuff, D,Se19- Diabase, gabbro, porphyrite, green porphyry serpentine, G20-Granite

series of the Menderes massif which was determined as Late Cretaceous, and the overlying unit with the reddish pelagic marbles named as Kýzýlaðaç unit as Early Paleocene. Besides, he located the boundary of Menderes Massif and Lycian Nappes between Kýzýlaðaç and Kurin units as it is in Graciansky's section (Figure 3).

Dora (1976) published the 1/500.000 scaled geological map and the generalized geological map of the whole Menderes Massif based on his observations dividing the Massif into Eðrigöz, Gördes, Ödemiþ and Çine submassifs. In the publication, high temperatured metamorphic cores (gneisses and migmatites) were distinguished. Monocline tricline latticework in K-feldspars based on the transformation temperature and index minerals, shows the boundaries between greenschist and almandine amphibolite facies (Figure 4), but these boundaries were subjected to significant changes by new investigations.

Kun and Dora (1984) located leptites (metavolcanites) among the core and cover series, lithostratigraphically. Þengör et al., (1984), interpreted these metavolcanites as volcanics of the Pan-African Orogeny rich in silica, and was thought that this layer might be the continuity of the Pan-African suture belt determined in Karacahisar dome and symbolized the Main Upper Pan-African discontinuity.

Dora et al., (1990), published an evolution scenario shown in schematic figures belonging to the geological history of the Menderes Massif which was prepared on the basis of literature background and their findings (Figure 5a,b). The publication which presents many different interpretations with the recent evidences caused to the increasing interest of the researchers to the geology of the Menderes Massif, and interpretations from location and settlement mechanism points of view.

Erdoðan (1992) and Erdoðan and Güngör (1992) different than the previous researchers, claimed that gneissic granites syntectonically settled at the time of Menderes Main Metamorphism (MMM), and therefore, augen gneisses in the massif originated from Upper Cretaceous-Lower Eocene granitoids. This new hypothesis was later explained on schematic sections by Erdoðan and Güngör (2004). Besides, it is claimed in this study that Lycian Nappes synchronologically thrusted with the main metamorphism of Menderes Massif from south to north and therefore the Menderes platform has been subjected to folding plunged to the north. These papers gave rise to many other investigations from different perspectives for the Massif.

Bozkurt et al., (1993) and Bozkurt and Park (1994), pointed out that, the protolithes of gneisses in the Çine submassif are the granitoides intruded by the extensional collapsing of the thickened crust of the western Anatolia in Late Oligocene which were originated by the partial melting of the greywackes during the Main Menderes Metamorphism (MMM) formed under greenschist-upper amphibolites facies conditions. They also claim that, after the intrusion, granitoids have been rised under ductile conditions along a big shear zone dipping to the north which was the product of an extensional tectonics, and transformed in to gneisses. Therefore, according to these investigators gneiss-schist contact at the south of Çine submassif symbolizes a typical detachement fault. Later, Bozkurt (2004) accepted the presence of two different types of gneisses as leucocratic metagranites much younger than Eocene aged Menderes Main Metamorphism, and Precambrian augen gneisses in the Menderes Massif. They declare that the first interpretation they made in their previous publication is valid for the young leucocratic metagranites, but many investigators claim that gneiss schist contact at the southern margin of Çine submassif does not have any traces of brittle and ductile deformation following one another which is a characteristic for a detachement fault and this contact symbolizes one of the pervasive intrusional contacts belong-

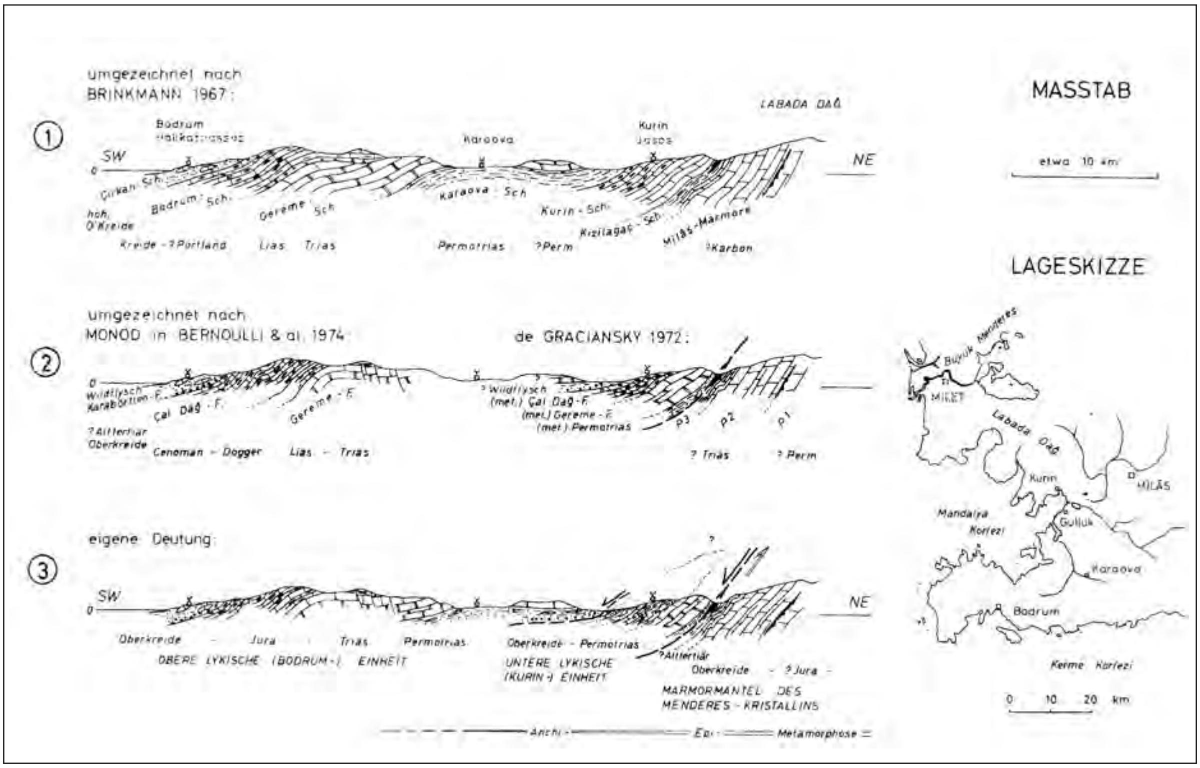

Figure 3- Semi schematic Bodrum-Kulin geological sections of Lycian units, SW Milas (Dürr, 1975). Explanations: 1: Brinkmann (1967), Uppermost Creatceous, Cretaceous-? Portlandian, Liassic, Triassic, Permo - Triassic, ?Permian, ?Carboniferous, 2: Bernoulli et al., (1974), Monod and De Graciansky (1972): Lower Tertiary-Upper Cretaceous, Senomanian-Dogger, Liassic-Triassic; 3: own interpretation; Upper Lycian (Bodrum) unit: Upper Cretaceous-Jurassic- Triassic, Permotriassic; Lower Lycian (Kurin) unit: Upper Triassic-Permo - Triassic; Marble mantle of the Menderes Massif: ?Lower Tertiary, Upper Cretaceous-? Jurassic.

ing to Pan-African granitoids (Candan et al., 2006; Konak et al., 1987).

Candan et al., (1992) illustrated napped structures in the Menderes Massif in Aydýn Mountains. Base cover series in this region composed of augen gneiss, leptite and schists take place on different units. Later on, it was understood that some of these klippes, have taken their recent forms by means of detachement faults. Widespread nappe structures in Menderes Massif were defined in many regions after 1994 (Konak et al., 1994; Dora et al., 1994). Dora et al. (1995) have shown some of these structures in his paper as attached maps and schematic sections (Figure 6). This study is interesting in terms of collecting the new evidences discovered so far in the Massif (granulites and eclogites in the Pan - African basement, the first Cambrian ages of the rocks named as leptite gneiss, Early Triassic metagranites and the polyphase metamorphism in the Massif). Granulite and eclogite relics at the Pan-African basement of Menderes Massif were first defined in 1994 by Candan et al., Later on, the formation of these relics, metamorphic conditions and relative ages were established with respect to each other (Oberhänsli et al., 1997; Candan et al., 2001).

Candan and Dora (1998), compiled generalized geological map of the Menderes Massif in 1/750 000 scale and presented it at a workshop

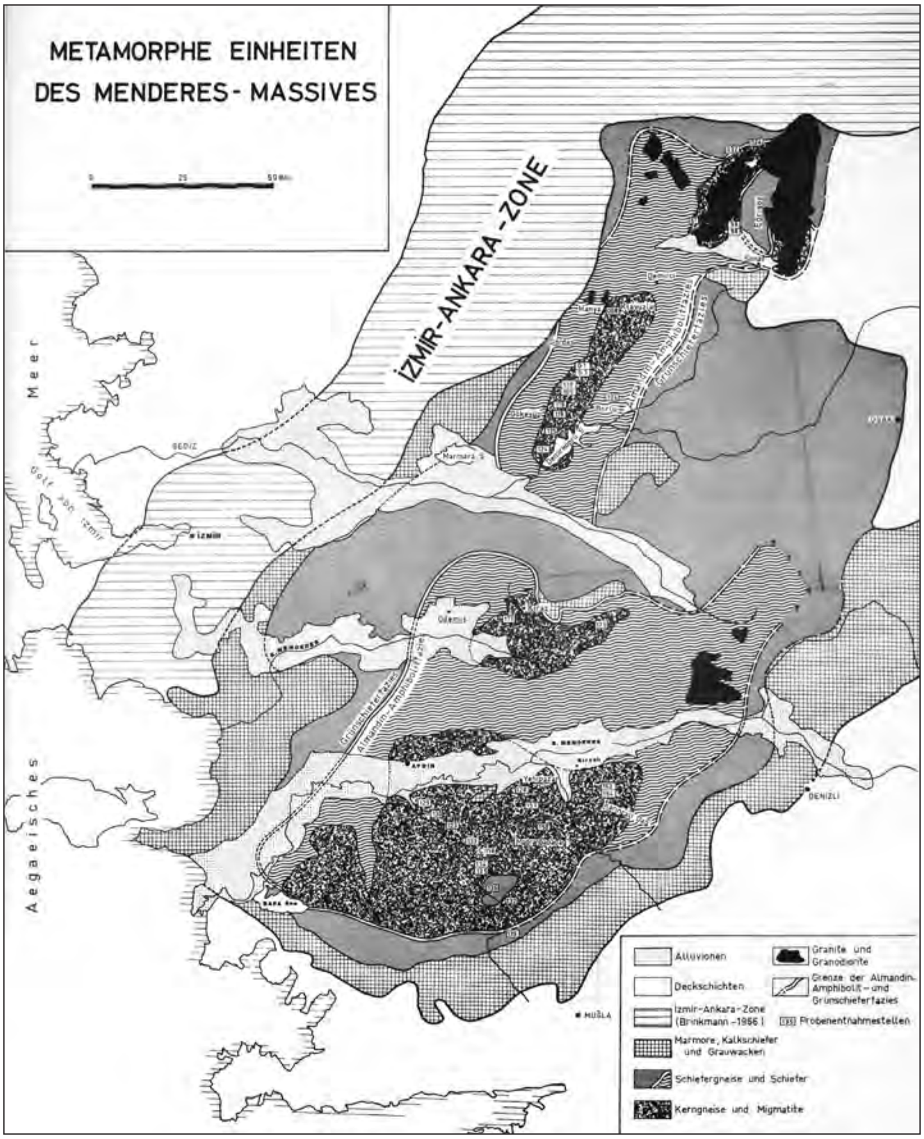

Figure 4- Metamorphic units of the Menderes Masif (Dora, 1976). Explanation: 1-Aluvials; 2-cover series; 3- Ýzmir-Ankara zone; 4- Marbles, Calcshists and graywackes; 5- Gneissic schists and schists; 6- Core gneisses and migmatites; 7- Granite and granodiorites; 8- The boundary between Almandine, amphibolites and greenschist metamorphism, 9-Sample Locations.

held in Mainz University. However, this map was published at a very less amount but, it still has been used by many investigators. Detailed studies on nappe packages in the Menderes Massif were investigated by Partsch et al., (1998), Ring et al., (1999) and Gessner et al., (2001*a*; 2001*b*). It can be recordet that Ring et al., (1999)'s suggestion about the formation of Dilek peninsula and Selçuk region which are formed by a nappe deposition drifted from Cycladic complex and the Menderes Massif has generally been formed from 4 nappe packages

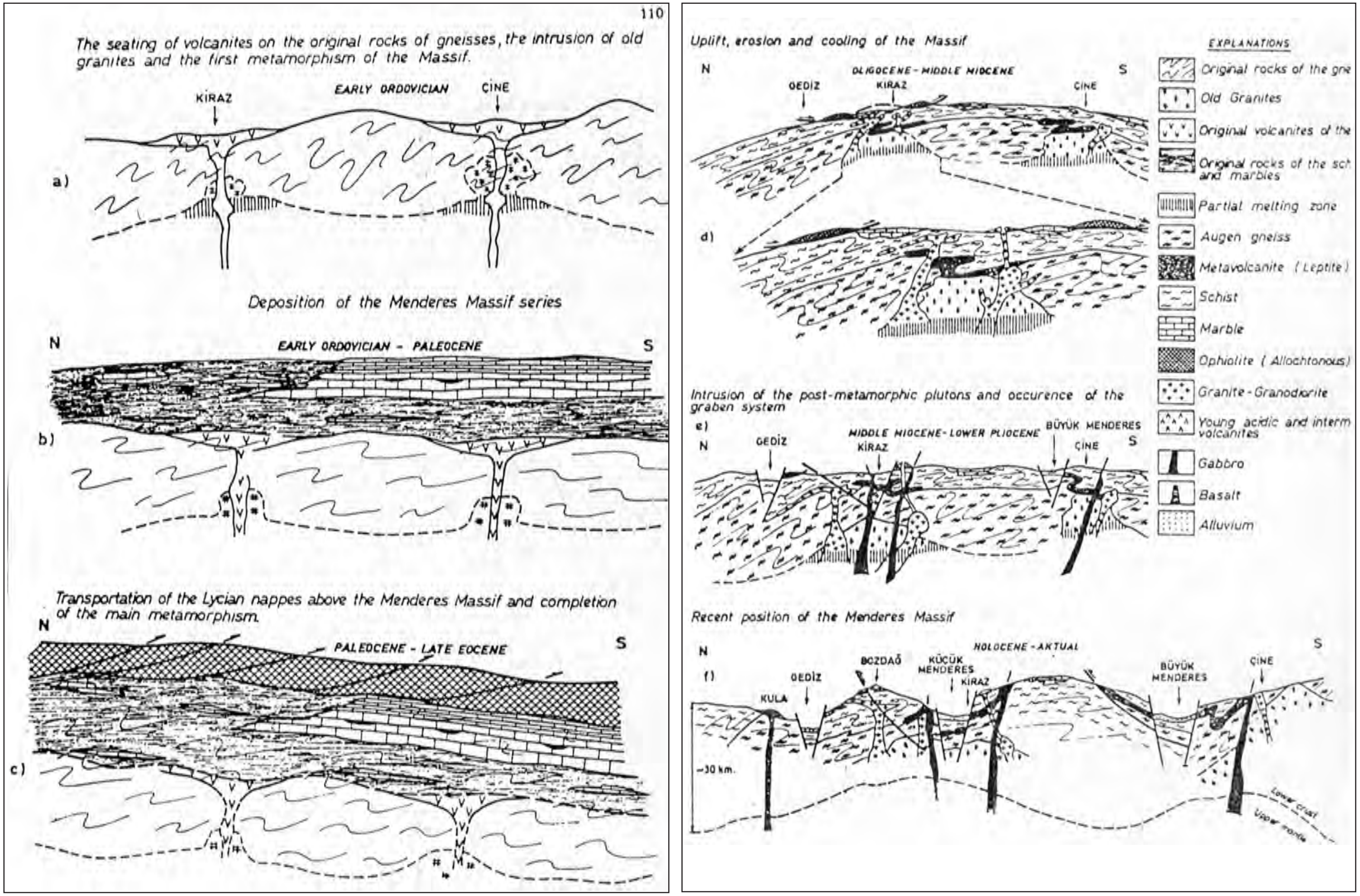

Figure 5ab - Schematic cross sections showing the tectono metamorphicevolution of the Menderes Massif from Devonian to recent a-f (Dora et al., 1990).

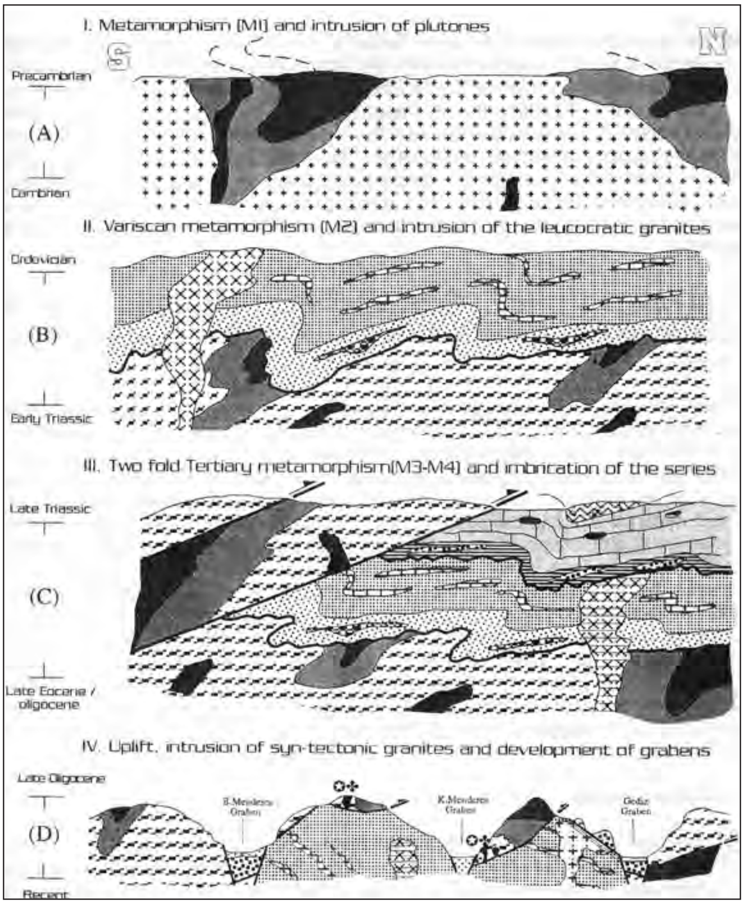

Figure 6- Drafts of the schematic sections of some important stages regarding the geological evolution of the Menderes Massif (Dora et al., 1995).

are the new interpretations. However, extending the shelf series (Dilek Nappe) belonging to Cycladic Complex along the southern margin of the Menderes Massif until Göktepe disagrees with the recent interpretations (Dora et al., 2005). At the bottom of Mesozoic series of the Menderes Massif, Mg-carpholites were detected different than Cycladic complex (Rimmelé et al., 2003). Besides, metaolisthostrome unit forming the uppermost unit of the series does not contain metagabbro and eclogite relics again different than Cycladic Complex (Figure 7).

After 1995, suspects arise about volcanic origins of leptites in the work group in Dokuz Eylül University. Especially, the layers of these rocks have many successions with schists ranging from cm to hundreds of meters in Demirci-Gördes submassif. On the other hand, it was determined that all zircons acquired from these rocks for radiometric dating have lost their primary smooth crystalline forms because of transportation and are highly in rounded shape. Radiometric dating (610-3229 Ma), sedimantological and geochemical studies have revealed that these rocks named as leptite are the paragneisses derived from the metamorphism of sediments alternating with sub arkose mudstone with abundant litarenite belonging to Pan-African basement. Field data show that paragneisses

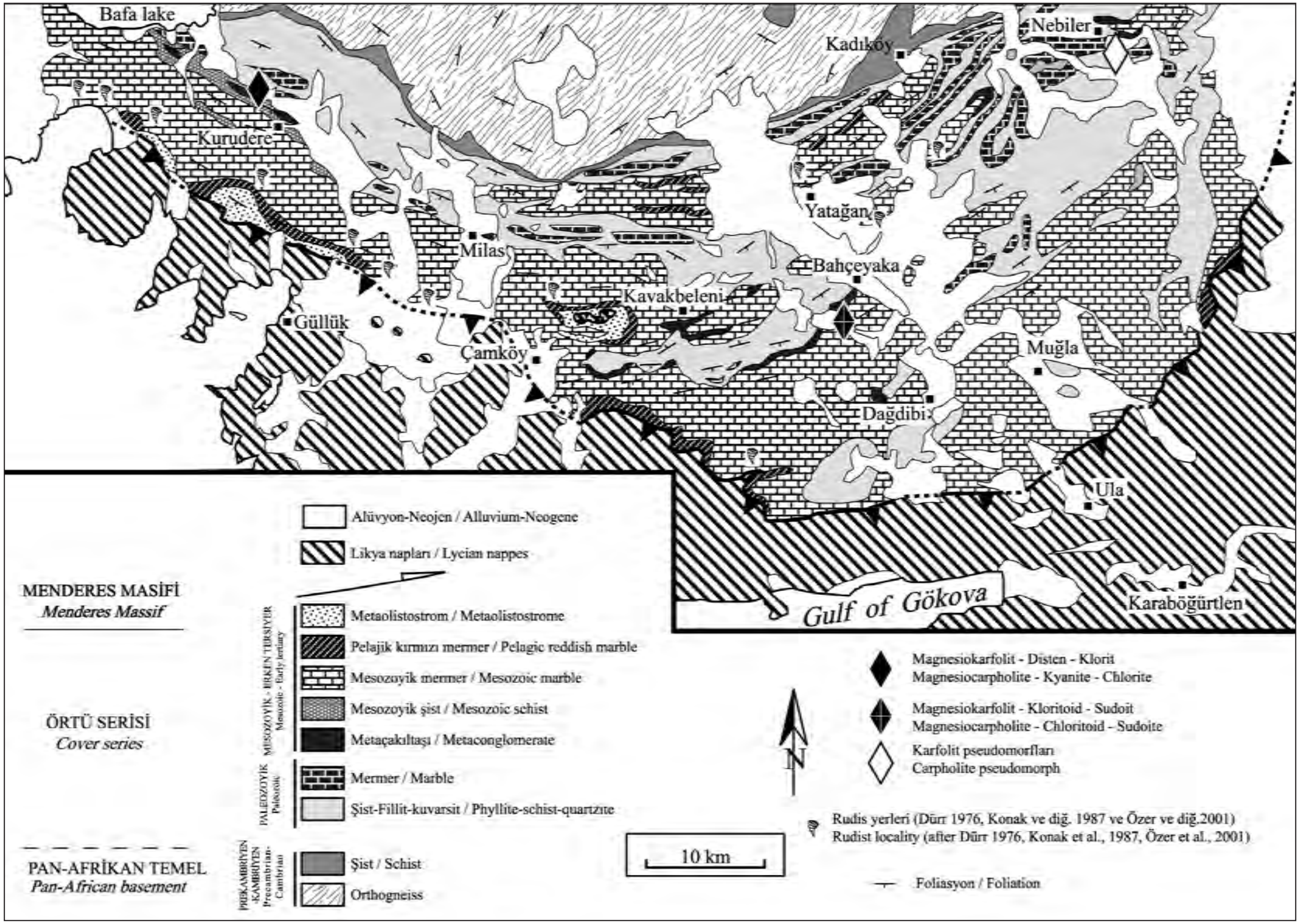

Figure 7- Generalized geological map of the Çine submassif (southern part) and the locations of Alpine HP/LT metamorphism groups (The map was obtained from Çaðlayan et al., (1980), in main details. Rudist locations belong to Dürr, 1975, Konak et al., 1978 and Özer et al., 2001, carpholite locations belong to Rimmelé et al., (2003).

are comformably and transitionally overlain by schists. The youngest detrital zircons in these schists were determined as 592 Ma. It is assumed that the protolithes of the series composed of paragneiss and schists are the clastic sediments derived from the source area composing of cratonic crystallines (Dora et al., 2001).

Okay (2001) states that Eocene aged structures in the Menderes Massif could be explained by a northward dipping recumbent fold. Thus, he asserts that the stratigraphical and mineralogical inversion in Bozdað and Aydýn Mountains could be understood easier. He also emphasises that, the Cycladic Metamorphic complex overlies the Menderes Metamorphics along a big thrust zone after the Menderes folding and internal imbricating. This claim which tries to explain the structural position of the Middle Menderes Massif out of nappe packages should also explain some other points with respect to top and bottom relationships of the units in lithostratigraphy.

Dora et al., (2005) stated that the age of all gneiss types in Çine submassif is around 550 Ma by means of radiometric data. These investigators also emphasize that the primary contact between the Pan-African basement and the Palaeozoic cover symbolizes an unconformity plane depending on the metagranite pebble bearing metaconglomerates derived from the Pan-African basement located at the bottom of Palaeozoic series in the Menderes Massif (over Pan- African Unconformity).

Many studies have been established and various models have been created concerning about the cropping out of the Massif after 1995 (Bozkurt and Park, 1997; Iþýk et al., 2003; Seyitoðlu et al., 2004; Thomson and Ring, 2006).

Many of the publications state that rapid erosion have taken place in the Massif following the uplift along the slip faults in northern and southern boundaries since Early Miocene.

Dating magmatic and metamorphic evolution of the Menderes Massif has rapidly increased in recent years. It has been approved that augen gneisses belong to Pan-African (Hetzel and Reischmann, 1996; Koralay et al., 2002; Gessner et al., 2004) and leucometagranites which cut the Palaeozoic series belong to some parts of the Massif in Lower Triassic (Dannat, 1997; Koralay et al., 2001). There is almost a concensus that, polymetamorphic evolution of core series is related to Pan-African (Hetzel et al., 1998; Oberhänsli et al., 2002; Koralay et al., 2006). It is also stated that Alpine metamorphism affecting the core together with cover series is Eocene (Satýr and Friedrichsen, 1986; Hetzel and Reischmann, 1996; Lips et al., 2001).

PICTURE OF TODAY

The geological investigations systematically carried out in the Menderes Massif since 1962 has served a lot to the solution of the problems up to a certain limit, such as the stratigraphy of the Massif, tectonic structure, age of the magmatic activities, conditions and ages of the metamorphisms, cropping out and correlation of neighboring tectonic units. Fine grained paragneisses are observed at the bottom of Pan-African basement in the generalized lithostratigraphical succession of the Menderes Massif. Paragneisses are often successively intercalated with schists then completely grades into schists towards upward. Dora et al., (2005) stated that the ages ranging between 592-3239 Ma were obtained for detrital zircons in these rocks. Due to this fact, it is assumed that these clastic series were deposited at the passive continental margin looking at the northern slope of Gondwanaland in Late Proterozoic. The thickness of the monotonous paragneiss and schist series reach 7-8 meters in Bozdað region. Both paragneisses and schist series are cut by gabbros and various granitoids of which the crystallization age varies between 570-5212 Ma mainly around 550 Ma (Koralay et al., 2007; Loos and Reischmann, 1999). The whole sequence has first undergone to a granulitic metamorphism, and then is affected by an eclogitic first, and then a Barrowian type, metamorphism under upper amphibolite facies conditions in Upper Proterozoic time. The Pan-African basement of the Massif, although not observed in all regions, is covered with Lower Palaeozoic series presented by an unconformity surface (quartzite and metaconglomerates). The thickness of the clastics in Palaeozoic series in the Menderes Massif reach 2-3 km. and become rich in graphite in upward levels and is intercalated with black marbles. The age of the black marbles was determined as Permocarboniferous (Onay, 1949; Schuiling, 1962). The bottom of the Upper Palaeozoic sequence is overlain by an Upper Triassic purple colored metaconglomerates and metasandstones (Konak and Çakmakoðlu, 2007). The unconformity between Palaeozoic and Mesozoic times is not possibly observable in all parts of the Massif. However, metaconglomerates abundant in quartz components are sparse and comprises typical disthene - chloritoid - Mg - carpholite assemblage. Again, Na-amphibolite relics are observed in metabasics which are located at the bottom. The platform type massive dolomites called as Milas marbles in upper levels of the sequence overlies the metasandstone and metaconglomerates. Emery lenses are observed at lower levels of the Milas formation and rudist fossils are recognized at top levels. Cenomanian age was given to emery levels and Santonian-Campanian age was given to rudistic levels by Özer et al., (2001). Red colored, thin bedded pelaj marbles of the Kýzýlaðaç formation overlies the platform type gray marbles. Due to foraminifers and nanno planktons, the age of these marbles is defined as Late Campanian-Late Maastrichtian (Özer et al., 2001). Kazýklý Formation is the uppermost unit made up of flysch type olistostromal rocks overlying the Kýzýlaðaç formation and the age was given as Paleocene. However, according to Özer et al., (2007), there is also a possibility that the Kazýklý unit unconformably overlies Kýzýlaðaç unit and therefore, Kazýklý unit is included into Lycian Nappes. Na-amphibole relics are also observed in Kazýklý unit including blocks such as marble, serpentinite and etc. On the other hand, the oldest deposits with nonmetamorphic and olisthostromal in character which overlies the Massif is Upper Miocene in age (Konak, 2007).

Today, it is accepted that the Menderes Massif is made up of a nappe package by many investigators. It is possible to trace the nappe packages starting from Demirci - Gördes submassif in the north, to the Çine submassif in the south (Figure 8). In some cases, inner imbrication of cover units and Precambrian basement and in some cases, the overlapping of two main units to each other are observed. Cycladic complex thrusted on the other units of the Massif along its west and northwest margins. According to Gessner et al., 2001*a*, Cycladic Menderes Thrust has occurred after the high pressure/low temperature (HP/LT) metamorphism of the Cycladic blueschists and at a period following the main Alpine metamorphism of the Massif. There are many different opinions about the deformation periods, strike and directions of the compressive tectonic. The Pan-African basement which was subjected to polymetamorphism generally shows very complicated movements in directions. Only in the cover series affected from the Alpine aged metamorphism, the northward lineaments are abundantly developed under high temperature conditions and are directly related with internal imbrication. Southward movements, on the other hand, has been developed under greenschist facies conditions and overlies the structures at the north. At present, all of the investigators agree with the idea that nearly E-W directed grabens in the Menderes Massif formed by the active normal faults since Pliocene.

The most significant magmatic activity in the Menderes Massif develops with the Pan-African Orogeny. The collision type calc alkaline and per alkaline granitoids are intruded (between 570-528 Ma) (Loos and Reischmann, 1999; Koralay et al., 2007). Today, the rocks that come up in two groups named as the augen gneisses and tourmaline leucocratic orthogneiss are the magmatic masses which were developed under the

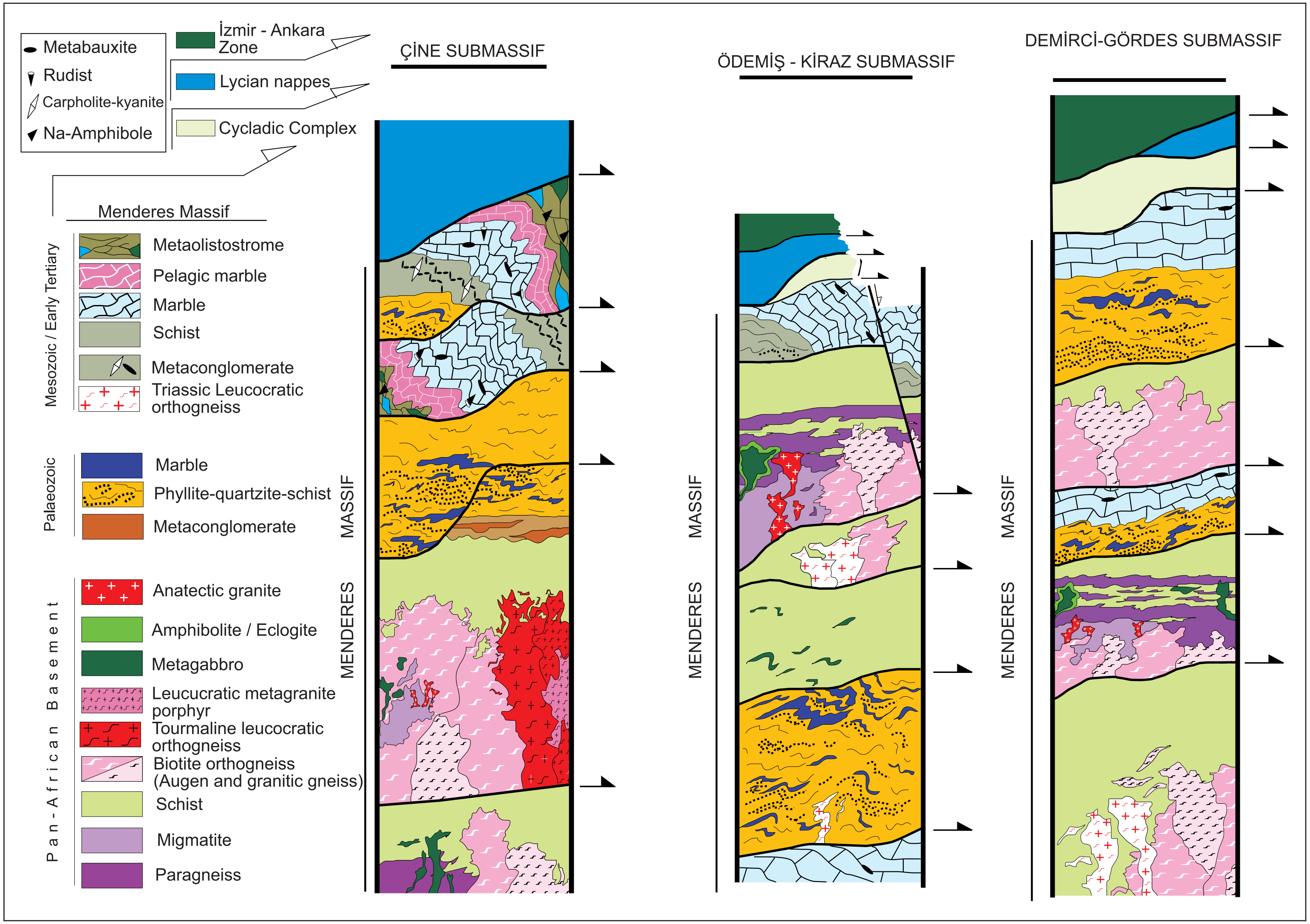

Figure 8- Generalized rock beddings belonging to the submassifs of the Menderes Massif (Dora et al., 2005).

same tectonism, crystallized from one source rock in different partial melting periods and were subjected to crustal contamination in differentrations. Synorogenic granitoids (generally augen gneisses) display less deformation than post orogenic granitoids (leucocratic metagranites and aplites). However, it is not possible to make a distinction among granitoid types with respect to deformation stage. The second big magmatic activity in the Menderes Massif was observed in Lower Triassic time. These leucocratic masses, very light colored and lack in biotite are 235-246 Ma. (Dannat, 1997; Koralay et al., 2001). The magmatism of this period has not yet been associated with a distinct orogenic phase in West Anatolia. However, Akkök (1983) and Koralay et al., (2001) state that these magmatites are related with the closure of Paleotethys. The youngest magmatic activity in the Massif are the young granitoidic intrusions that are mylonitic in character and synchronically uplifted by detachement faults as a product of dilation tectonism. Granitic stocks 21 Ma in age in Simav region and 13 Ma in age in Bozdað region, which show definite directional traces are closely related with the extension of the Menderes Massif.

Menderes Massif has experienced a polyphased metamorphic evolution. The mineral assemblages defining the granulite, eclogite and amphibolite facies, are respectively observed in Pan African basement. Oelsner et al., (1997) and Koralay et al., (2006), determined the metamorphism age of the granulite facies as 660+61/-63 Ma and as 583±5,7 Ma respectively. As for the metamorphic conditions was estimated 730 °C in temperature and 6 kb in pressure (Dora et al., 2001). Eclogitic metamorphism follows the granulitic metamorphism. It is accepted that this metamorphism of which its effects are observed especially in metagabbros, has occurred 529±22 Ma ago (Oberhänsli et al., 2010) under 644 °C in temperature and in 15 kb pressure. Granulite and eclogite facies assemblages show traces of widespread reverse processes because of the overlying Barrowian type medium pressure metamorphism. Migmatites derive from granulites and amphibolites generate from garnet. Amphibolite facies metamorphism has developed under 628 °C in temperature and 7 kb in pressure (Candan et al., 2007). The age of anatectic granites were defined as derived from paragneisses as 551±1,4 Ma (Hetzel et al., 1998) and as 540 Ma (Dannat ve Reischmann, 1998). The outermost circle of the zircons of granulites that was subjected to recycling gives an age about 560.0±5,6 Ma (Koralay et al., 2006). When the 528-570 Ma aged orthogneisses and polyphase metamorphism of the Pan-African basement are taken in to consider together, it is understood that the granitoids intruded synorogenically during Pan-African Orogeny and partly post orogenically.

The basement and cover series of the Menderes Massif were affected from an Alpine aged regional metamorphism so called as the Main Menderes Metamorphism. There are various opinions about the age of this Barrowian type metamorphism which has occurred under conditions of greenschist of lower amphibolite facies. It is stated that the first non metamorphic sedimentary rocks covering the Massif is Upper Eocene (Konak and Çakmakoðlu, 2007). The Rb/Sr white mica giving an approximate age of 56 ±1 Ma which ranges 63-48 Ma agrees with this geological evidence. The 37 Ma Ar/Ar biotite (Satýr and Friedrichsen, 1986) and muscovite approximate ages (Hetzel and Reischmann, 1996) are interpreted as the cooling age. Geological and geochronological data show that the age of the Alpine metamorphism in the Menderes Massif is Upper Paleocene to Middle Eocene. The formation conditions of this metamorphism was estimated as 5-8 kb in pressure and 430-550 °C in temperature (5 kb and 530-550 °C: Ashworth and Evirgen, 1984; 8 kb and 530 °C: Okay, 2001; 6 kb and 430-550 °C: Whitney and Bozkurt, 2002). By finding the carpholite-disthene assemblages in Triassic quartz-metaconglomerates in the Menderes Massif, it was understood that there is an Alpine aged HP/LT metamorphism in cover series of the Massif (Rimmelé et al., 2003). However, the location of the Barrowian type

Alpine metamorphism and this HP/LT metamorphism with respect to each other could not clearly be understood.

Many hypotheses were developed about collection of Anatolide-Tauride tectonic units. Generally, this process is associated with the closure of Northern branch of Neotethys and collision of Anatolide-Tauride platform with Sakarya continent in Paleogene. A schematic model is proposed by a group of investigators including the present author as shown below (Figure 9). Starting from Albian, the Anatolide-Tauride platform subducts northward, below the Sakarya continent along the Izmir-Ankara suture zone. In Campanian, because of low geothermal gradient, the upper series and the lower series of the Tavþanlý Zone reaching to 60 km depth, are affected by a greenschist metamorphism under 5-8 kb pressure, 250-300 °C temperature and 20±2 kb pressure, 430 °C tempeture conditions, respectively. The Afyon zone reaching to a subduction-obduction zone between Early Paleocene-Late Paleocene times undergoes HP/LT metamorphism under 6-9 kb pressure and 350 °C in temperature conditions. And the HP/LT metamorphism of the Menderes Massif has become under 10-12 km pressure and 440 °C in temperature, most probably before Main Alpine Metamorphism in Eocene time. According to data above, the HP/LT metamorphism ages of Anatolides become markedly younger from north to south. K/Ar dating that was estimated as 40 Ma from Na-amphiboles of Dilek Nappe belonging to Cycladic complex (Oberhänsli et al., 1997) show that the HP/LT metamorphism of this unit has become in Lower Eocene too. Lycian nappes which caused the regional metamorphism

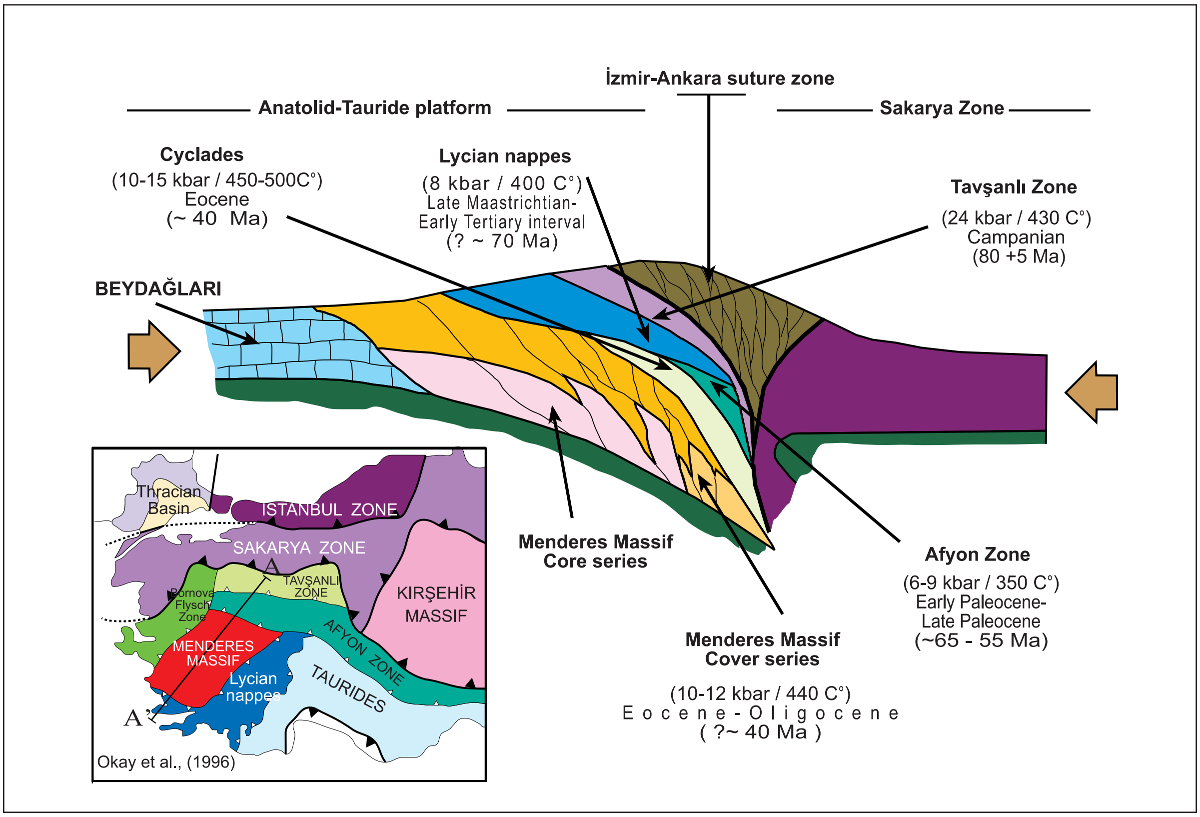

Figure 9- The tectonical environment, age and the condition of the Alpine metamorphism in the Menderes Massif (modified from Rimmelé et al., 2003).

burying the Menderes Massif into deep levels during its passage from North to South has passed its own HP/LT metamorphism (probably in Late Maastrichtian- Eocene times) under of 8 kb pressure and 400 °C temperature conditions.

In West Anatolia, at the beginning of Early Oligocene, various hypotheses were produced about the transformation of the compressional tectonics into the extensional tectonics. These probabilities put forwed mainly by Seyitoðlu and Scott, (1996) are given below:

1) The collapse of the thickened crust as a result of continental collision under gravity (orogenic collapse model); 2) Lithospherical slab break off or break away; 3) Dilatational environment developed as a result of the migration of Hellenic subduction zone to the south (regression and back arc rifting model); 4) The dilatation formed by the movement of Anatolian micro-continent eastward along East and North Anatolian faults as a result of the compressional stage following the continental collision in East Anatolia (tectonic escape model); 5) the combination of some mechanisms mentioned above. There is a common agreement about the beginning of transformation of compressive regime in West Anatolia into extensional regime in NS direction (Bozkurt and Satýr, 2000; Gessner et al., 2001*a*; Iþýk et al., 2003; Ring et al., 2003; Seyitoðlu et al., 2004; Thomson and Ring, 2006).

Many kinematical studies have been made to explain the dilatation and related exhumation of the Menderes Massif along the gently sloping normal faults in West Anatolia in the recent years. All researchers agree that, the kinematic data in the massif define the presence of two movements to the north (old) and to the south (young). However, there are great differences in the idea of meaning of these movements. According to Hetzel et al., (1998), Gessner ve et al., (1998), Bozkurt and Park (1999), Bozkurt and Satýr (2000), Rimmelé et al., (2003), the northward movement is the production of the compressional tectonics which is contem poraneous with the Main Menderes Massif metamorphism happened in Eocene time. This deformation has been developed by the back thrust of Lycian Nappes which caused to the internal imbrications of the massif . Ring et al., (1999) and Gessner et al. (2001*a*); interprete that northward movements are pre Alpine which is associated with the internal imbrications of the Pan-African basement of the massif in Precambrian. On the other hand, the same northward movement is interpreted as the structures that belong to dominant deformation observed in the massif and defines the extensional tectonism related with the exhumation of the massif by Seyitoðlu et al., (2004). Similarly, there are great differences in interpretations regarding the northward movements. According to Bozkurt and Park (1994, 1997, 1999), these movements are the production of Oligocene-Early Miocene aged extensional tectonism and are closely related with the exhumation of the massif. As a contradiction to this opinion, Ring et al., (1999), and Gessner et al., (2001*a,b*) interpret the southward movements as the production of a compressional tectonism and make a connection with the internal southward imbrications of the Massif in this stage and the passage of Lycian nappes to south. Seyitoðlu et al., (2004), explain that southward movements are secondary structures related with doming.

Many studies have been made showing that the Massif is a core complex and Oligocene-Miocene aged exhumation has occurred along the normal faults (slip faults) with low angle. At present, there are many different opinions about the location of these slip faults and about the effects of the exhumation of the massif. In preliminary studies, Emre and Sözbilir (1995), suggested that the normal fault with low angle extending between Kemalpaþa - Alaþehir played an important role on this exhumation and the beginning age of extensional tectonism that caused the exhumation of the massif is 19 Ma (Early Miocene) based on syntectonic granitic intrusions Hetzel et al., (1995*a*). Based on the kinematic data obtained from Bozdað region, it was interpreted as this side of the massif is a symmetrical core complex by Hetzel et al., (1995*b*). In the following years, the presence of a similar type fault has been detected on the southern part of the Aydýn Mountains associating with its northern conjugate and suggested that Ödemiþ-Kiraz submassif represents a typical symmetrical core complex (Bozkurt, 2001). However, in recent studies, the aforesaid two slip faults developed in the second period are young structures and are associated with the late period exhumation of Ödemiþ-Kiraz submassif (Seyitoðlu et al., 2004; Ring and Collins, 2005).

Bozkurt and Park (1994, 1997), Bozkurt and Satýr (2000) explained that the gneiss-schist contact extending between Bafa-Yataðan is a tensile shear zone that has played an important role especially on the exhumation of southern part of the Menderes Massif. According to the investigators, Tertiary granites have transformed into gneisses along this shear zone and have exhumed from an approximate depth of 15 km with a continuous period. Hetzel ve Reisschman (1996) suggest that this zone is a shear zone that has been active in Eocene but, this does not have an effective role on the exhumation of the massif. They claim that the exhumation has occurred with passive erosion. On the other hand, Ring et al., (1999) and Regnier et al., (2003) suggest that tectonic zone has no relation with the extension but is a product of compressional tectonism and have the character of a south vergent thrust fault. Bozkurt (2004) revisioned the tectonic model related with the exhumation of this zone and the southern part of the massif. He also stated that, massif was transported to shallow depths along the crustal scaled zone, and the last exhumation occurred by young graben faults. Bozkurt et al., in (2006), again claimed that, this zone is originally a northward thrust and reworked as southward shear zone as dilatational production in the following period.

In recent years, in some tectonic models, it has been suggested that detachement fault in north of Simav had played an important role on the exhumation of the Menderes Massif (Iþýk and Tekeli, (2001); Iþýk et al., (2003), 2004; Seyitoðlu et al., (2004); Ring et al., (2003); Ring and Collins, (2005); Thompson and Ring, (2006). Eðriboz and Koyunoba granites have an approximate age of 20-23 Ma and located on the footwall synchronously with the extensional tectonism and undergone ductile deformation along detachement fault. According to Seyitoðlu et al., (2004) Simav slip fault is genetically related with Datça-Kale detachment fault in south and these two faults have accomplished the main exhumation process in two phased exhumation model of the Menderes Massif. Ring et al., (2003), Ring and Collins (2005), Thompson and Ring (2006), claim that, apart from the model briefly mentioned above the old thrust zone between the Menderes Massif and Lycian Nappes in the south worked as a detachement fault during exhumation. The authors also state that while the massif exhumed along this line at the south, exhumed along the Simav slip fault at the north. The reason of the extension in the region is proposed as post collisional magmatism in the area and thermal weakening of the crust caused by the magmatism. As a result of this, the plane of old thrust faults among the main units reactivated as detachement faults. The Simav detachement fault following the thrust plane between the Cycladic units and the Menderes Massif has been active between the ages of 19-25 Ma in a fault character dipping at low angle. Koyunoba and Eðrigöz syntectonic granites were intruded into formerly activated fault plane. The entrance of the granite caused doming and locked and stopped the movement of the detachement fault. The rocks aged 16.4 Ma unconformably deposited on the exposed footwall. The formation of symmetrical core complex core bounded by the Big Menderes detachement faults at the south and Alaþehir at the north in the central Massif, has continued until Middle to Late Miocene (Hetzel et al., 1995*a*; Gessner et al., 2001*a*; Lips et al., 2001).

The normal faults developed in Pliocene and formed the young grabens in the western Anatolia cut all the detachement faults.

RESULTS AND RECOMMENDATIONS

Within the framework of data obtained during the last 40 years, there is a need for the development of the subjects described below for the geology of Menderes Massif and a detailed investigation on the problems of which have not clearly been solved yet,

a. The genetical relation of the polymetamorphic and magmatic evolution of the Pan African basement of the Massif with the integration of Gondwanaland in Late Proterozoic has not clearly been established. It has a great importance for the facies of primary sediments, the age and geochemistry of magmatics, and coeval tectonic period and metamorphism.

b. The relation of acidic magmatics defined with gneiss and metagranites in various compositions in the Pan-African basement of the Menderes Massif with the poly phase Pan-African basement is one of the problems to be investigated in regional scale and should be solved with respect to age and magmatic differentiation.

c. The primary unconformable contact relationship between the base and cover series has been established only in Upper Mesken Village to the southeast of Yataðan. The determination of the primary relationship which should be in regional scale will give rise to the complete solution of one of the main problems. It will also help get a good starting point regarding the evolution of the massif.

d. The evidences about the character of the Palaeozoic and Mesozoic contact in the massif are still in question. Although the bottom of Mesozoic deposit is composed of metaconglomerates in metaclastics and channel fillings character, the unconformable character of the contact could not clearly be established in regional scale. It is necessary to spread out this observation to whole massif and points where certain relations could be observed should be determined.

e. The association of the Early Triassic magmatism in the massif with the oldest magmatism in other tectonic units of West Anatolia and the definition of the meaning of tectonic environment in regional scale will illuminate the geological evolution of Mesozoic time.

f. High pressure mineral assemblages in the Mesozoic series of the Massif require a different interpretation for the Alpine metamorphism in the Menderes Massif rather than it was accepted so far. Besides, to reveal the relationship of the tectonic environment among the Alpine HP/LT metamorphisms of Cycladic Complex and massif and time is another question to be answered.

g. Nappes belonging to Cycladic Complex are located on the western and northwestern sides of the Menderes Massif. The study of the tectonical relationship and metamorphic stage of the massif with Cyclades in regional scale and an establishment of a common stratigraphic, metamorphic and tectonical history will add a great value to the solution of the problems related with the illumination of crustal evolution in Western Anatolia and to the closure of the northern branch of Neotethys.

h. To discuss different tectonical models suggested for the exhumation of the massif and to make these in detail, is another important point which will conclude to develop a common model will be able to answer to all geological questions and considered by all the investigators.

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REFERENCE

Akkök, R., 1983. Structural and metamorphic evolution of the northern part of the Menderes Massif. New data from Derbent area and their implications fort he tectonics of the massif. Journay Geology, 91, pp.342-350

Ashworth, J. R. and Evirgen, M., 1984. Mineral chemistry of regional chloritoid assemblages in the Chlorite Zone, Lycian Nappes, south-west Turkey, Mineralogical Magazine, 48, pp. 159-165.

Bozkurt, E., 2001. Late Alpin evolution of the central Menderes Massif, western Turkey. International Journal Earth Science, 89, pp.728-744.

_____, 2004. Granitoid rocks of the southern Menderes Massif: field evidence for Tertiary magmatism in an extensional shear zone. International Journal of Earth Science, 93, 52-71.

Bozkurt, E., Park, R.G. and Winchester, J.A., 1993. Evidence against the core/cover interpretation of the southern sector of the Menderes Massif, west Turkey. Terra Nova 5, pp.445-451.

_____ and _____, 1994. Southern Menderes Massif: an incipient metamorphic core complex in Western Anatolia, Turkey. Journal Geological Society London, 151, pp.213-216.

_____ and _____, 1997. Evolution of a mid-tertiary extensional shear zone in the southern Menderes Massif, western Turkey. Bulletin Society Geological France, 168,1, pp.2-14.

_____ and ____, 1999. The structure of the Paleozoic schists in the Southern Menderes Massif, western Turkey: a new approach to the origin of the main Menderes Metamorphism and its relation to the Lycian Nappes. Geodinamica Acta 12, pp.25-42.

Bozkurt, E. and Satýr, M., 2000. The southern Menderes Massif (western Turkey): geochronology and exhumation history. Geology Journay., 35, pp.285-296.

_____, Winchester, A. J., Mittwede, S. and Ottley, C., 2006. Geochemistry and tectonic implications of leucogranites and tourmalines of the southern Menderes Massif, southwest Turkey. Geodin. Acta, 19/5, pp.363-390.

Brinkmann, R., 1967. Die Südflanke des Menderes-Massivs bei Milas, Bodrum und Ören, Scient. Represantive Faculty Science, Ege University., Izmir, Turkey.

Candan, O., Dora, O.Ö., Kun, N., Akal, C. and Koralay, E., 1992. Aydýn Daðlarý (Menderes Masifi) güney kesimindeki allokton metamorfik birimler. Türkiye Petrol Jeologlarý Derneði (TPJD) Bülteni, C4/1, pp.93-110.

_____, _____, Dürr, St., and Oberhänsli, R., 1994. Erster Nachweis von Granulit und Eklogit - Relikten im Menderes - Massif / Türkei. Göttingen Abr. Geol. Paläont. Sb.1 5. Symposium TSK, pp.217-220.

_____ and ____, 1998. Menderes Masifinin genelleþtirilmiþ jeoloji haritasý. Dokuz Eylül Üniversitesi Jeoloji Mühendisliði Bölümü Bornova-Ýzmir (unpublished).

Candan, O., Dora, O.Ö., Oberhänsli, R., Çetinkaplan, M., Partzsch, J.H., Warkus, F. and Dürr, S., 2001. Pan-African high-pressure metamorphism in the Precambrian basement of the Menderes Massif, Western Anatolia, Turkey. International Journal of Earth Science (Geologische Rundschau), 89, 4, pp.793-811.

_____, Koralay, E., Dora, O., Chen, F., Oberhänsli, R., Akal, C., Satýr, M. and Kaya, O., 2006. Menderes Masifi'nde Pan-Afrikan Sonrasý Uyumsuzluk: Jeolojik ve Jeokronolojik Bir Yaklaþým. 59. Türkiye Jeoloji Kurultayý Bildiri özleri, Ankara, pp.25-27.

_____, _____, Dora, Ö., Chen, F., Oberhansli, R., Çetinkaplan, M., Akal, C., Satýr, M. and Kaya, O., 2007. Menderes Masifi'nin Pan-Afrikan temelin stratigrafisi ve örtü - çekirdek serilerinin ilksel dokanak iliþkisi. Menderes Masifi Kolokyumu, Ýzmir. pp.8-14.

Dannat, C., 1997. Geochemie,geochronologie und Nd-Sm Isotopie der granitoiden Kerngneiss des Menderes Massivs, SW-Turkey. PhD thesis, Johannes Gutenberf Universitat Mainz, 85p (unpublished).

_____ and Reischmann, T., 1998. Single zircon ages of migmatites from the Menderes Massif, SW Turkey. Program des Workshops 'Das Menderes Massif (Türkei) und seine nachbargebiete'. Mainz, Germany.

Dora, O.Ö., 1976. Menderes Masifi'nde alkali feldspatlarýn yapýsal durumlarý ve bunlarýn perojenetik yorumlarda kullanýlmasý. Türkiye Jeoloji Bülteni, 24, pp. 91-94.

_____, Kun, N. and Candan, O., 1990. Metamorphic history and geotectonic evolution of the Menderes Massif. Proc. of. International Earth Sciences Congress on Aegean Regions, Izmir/ Turkey, Vol. 2, pp.102-115.

_____, Candan, O, Kun, N. and Akal, C., 1994. Menderes Masifi'nin metamorfik evrimi ve orta kesiminin (Ödemiþ - Kiraz Asmasifi) 1 / 500.000 ölçekli jeoloji haritasýnýn yapýlmasý. TBAG - 937 nolu Türkiye Bilimsel ve Teknolojik Araþtýrma Kurumu (TÜBÝTAK) projesi, 124p. (unpublished).

Dora, O.Ö., Candan, O., Dürr, S. and Oberhänsli, R., 1995. New evidence on the geotectonic evolution of the Menderes Massif. International Earth Sciences Colloquium on the Eagean Region, Izmir-Turkey, V.1, pp.53-72.

_____, _____, Kaya, O., Koralay, E. and Dürr, S., 2001. Revision of the so-called "leptite-gneisses" in the Menderes Massif: A supracrustal metasedimentary origin. International Journal Earth Science, (Geologische Rundschau), 89/4, pp. 836-851.

Dora, O.Ö., Candan, O., Kaya, O., Koralay, E. and Akal, C., 2005. Menderes Masifi Çine Asmasifi' ndeki Koçarlý - Bafa - Yataðan - Karacasu arasýnda uzanan gnays / þist dokanaðýnýn niteliði: Jeolojik, tektonik, petrografik ve jeokronolojik bir yaklaþým. YDABÇAG - 101 Y 132 nolu Türkiye Bilimsel ve Teknolojik Araþtýrma Kurumu (TÜBÝTAK) projesi, 197p. (unpublished).

Dürr, St. 1975. Über Alter und geotektonische Stellung des Menderes Kristallins / SW - Anatolien und seine Äquivalente in der Mittleren Aegean. Habilitation thesis University of Marburg, 107p (unpublished).

Egeran, N. and Yener, H., 1944. Notes explicatives de la Carte Géologique de la Turquie, Feville Izmir. Maden Tetkik ve Arama Genel Müdürlüðü

Emre, T. and Sözbilir, H., 1995. Field evidence for metamorphic core complex, detachment faulting and accomodation faults in the Gediz and Büyük Menderes grabens, western Anatolia. International Earth Science Colloquim On the Aegean Region-Izmir, proceedings, pp.73-98.

Erdoðan, B., 1992. Problem of core - mantle boundary of Menderes Massif. In proceedings of the international Symposium of Eastern Mediterranean Geology., Geosound, pp.314-315.

Erdoðan, B., and Güngör, T., 1992. Menderes Masifi'nin kuzey kanadýnýn stratigrafisi ve tektonik evrimi. Türkiye Petrol Jeologlarý Derneði (TPJD) Bulletin - C.4/I,S. pp.9-34.

_____ and Güngör, T., 2004. The problem of the core -cover boundary of the menderes Masif and an emplacement mechanism for regionally extensive gneissic granite, Western Anatolia Turkey. Turkish Journal of Earth Science, 13, pp.15-36.

Evirgen, M.M., 1979. Menderes Masifi'nin metamorfizmasýna petroloji, petrokimya ve jenez açýsýndan yaklaþýmlar (Ödemiþ - Tire - Bayýndýr - Turgutlu yöresi). Hacettepe Üniversitesi doctorate thesis, Ankara, 185p (unpublished).

Gessner, K., Ring, U., Lackmann, Passchier, C.W. and Güngör, T., 1998. Structure and crustal thickening of the Menderes Massif, southwest Turkey, and consequences for large-scale correlations between Greece and Turkey. Bull. of the Geological Society. Greece, XXXII, pp.145-152.

Gessner, K., Piazolo, S., Güngör, T., Ring, U., Kröner, A. and Passchier, C.W., 2001*a*. Tectonic significance of deformation in granitoid rocks of the Menderes nappes, Anatolide belt, southwest Turkey. International Journal of Earth Science, 89, pp.766-780.

_____, Ring, U., Johnson, C, Hetzel, R., Passchier, C.W. and Güngör, T., 2001*b*. An active bivergent rolling-hinge detachment systemral Menderes metamorphic core complex in western Turkey. Geological Society American Bulletin, 29,7, pp.611-614.

_____, Collins A., Ring, U. and Güngör, T., 2004. Structural and thermal history of poly-orogenic basement: U-Pb geochronology of granitoid rocks in the southern Menderes Massif, Western Turkey. Journal Geological Society London, 161, pp.93-101.

Graciansky, P., 1965. Menderes Masifi'nin güney kýýýsý boyunca (Türkiye'nin GB'sý) görülen metamorfizma hakkýnda açýklamalar. Maden Tetkik ve Arama Genel Müdürlüðü Bülteni, 64, pp.8-22.

Graciansky, P., 1972. Recherches géologiques dans le Taurus Lycien occidental (Turquie du SW). Thése University Paris-Sud (Orsay), 896p. (unpublished).

Hamilton. W., 1841. Researches in Asia minor, Pontus and Ermenia, London.

Hetzel, R. and Reischmann, T., 1996. Intrusion age of Pan-African augen gneisses in the southern Menderes Massif and the age of cooling after Alpine ductile extensional deformation. Geological Magazine, 133(5), pp.565 - 572.

_____, Ring, U., Akal, C. and Troesch, M., 1995 *a*. Miocene NNE-directed extensional unroofing in the Menderes Massif, southwest Turkey. Journal Geological Society London, 152, pp.639-654.

_____, Passchier, C., Ring, U. and Dora, Ö., 1995 *b*. Bivergent extension in orogenic belts: The Menderes Massif (southwestern Turkey). Geology, 23, 5, pp.455-458.

Hetzel, R. Romer, R., Candan, O. and Passchier, C.W., 1998. Geology of the Bozdað area, central Menderes Massif, SW Turkey: Pan - African basement and Alpine deformation. Geologische Rundschau, 87, pp.394-406.

Iþýk, V. and Tekeli, O., 2001. Late orogenic crustal extention in the northern Menderes Massif (western Turkey):evidence for metamorfik core complex formation. International Journal of Earth Science, 89, pp.757-765.

_____, Seyitoðu, G. and Çemen., Ý, 2003. Ductile-Brittle transition along the Alaþehir detachment fault and its structural relationship with the Simav detachment fault, Menderes Massif, western Turkey. Tectonophysics, 374, pp.1-18.

_____, Tekeli, O. and Seyitoðu, G., 2004. The Ar/Ar age of extentional deformation and granite intrusion in the northern Menderes core complex: implications for the initiation of extentional tectonics in western Turkey. Journal Asian Earth Science, 23, pp.555-566.

Ýzdar, E., 1971. Introduction to geology and metamorphism of the Menderes Massif of western Turkey. Petroleum Exploration Society of Libya, pp.495-500

Konak, N., 2007. Menderes Masifi'nin Prekambriyen - Paleozoyik istiflerinin tektonik üniteler bazýnda tartýþýlmasý. Menderes Masifi Kolokyumu, Ýzmir, Geniþletilmiþ bildiri özleri kitabý, pp.17-23.

_____, Akdeniz, N. and Öztürk, E.M., 1987. Geology of the south of Menderes Massif, I.G.C.P. project no:5, Correlation of Variscan and pre-Variscan events of the Alpine Mediterranean mountain belt, field meeting, Mineral Research and Exploration Institute Turkey, pp.42-53.

_____, Çakmakoglu, A., Elibol, E., Havzoglu, T., Hepþen, N., Karamanderesi, I.H., Keskin , H., Sarikaya, H., Sav, H. and Yusufoðlu, H., 1994. Development of thrusting in the median part of the Menderes Massif. Abstracts 47th Geology Congress Turkey-Ankara, 34p.

Konak, N. and Çakmakoðlu, A., 2007. Menderes Masifi ve yakýn çevresinin Mesozoyik-Alt Tersiyer istiflerinin tektonik üniteler bazýnda tartýþýlmasý. Menderes Masifi Kolokyumu, Ýzmir, Geniþletilmiþ bildiri özleri kitabý, pp.56-65.

Koralay, O.E., Satýr, M. and Dora, O.Ö., 2001. Geochemical and geochronological evidence for Early Triassic calc-alkaline magmatism in the Menderes Massif, Western Turkey. International Journal of Earth Science, 89, pp.822-835.

______, Candan, O.,Dora, O.Ö., Chen, F., Satýr, M. and Kaya, O., 2002. Single zircon Pb-Pb ages for para- and orthogneisses in the Menderes Massif, western Turkey: An approach to the original deposition age of the paragneisses. 1st International Symposium of faculty of mines (Ýstanbul Teknik Üniversitesi) on Earth Sciences and Engineering. Abstracts,105p.

______, Chen, F., Oberhänsli, R., Wan, Y. and Candan, O., 2006. Age of Granulite Facies Metamorphism in the Menderes Massif, Western Anatolia / Turkey : SHRIMP U-Pb Zircon Dating. 59. Türkiye Jeoloji Kurultayý Bildiri özleri, Ankara, pp.28-29.

Koralay, E., Candan, O., Dora, Ö., Satýr, M., Oberhansli, R. and Chen, F., 2007. Menderes Masifi'ndeki Pan-Afrikan ve Triyas yaþlý metamagmatik kayaçlarýn jeolojisi ve jeokronolojisi, Batý Anadolu, Türkiye. Menderes Masifi Kolokyumu, Ýzmir, pp.24-31.

Kun N. and Dora, O., 1984. Menderes Masifi'ndeki metavolkanitler. 38. Türkiye Jeoloji Kurultayý, Bildiri özleri, pp.131-132.

Lips A., Cassard D., Sözbilir H., Yýlmaz H., and Wijbrans, J., 2001. Multistage exhumition of the Menderes Massif, western Anatolia (Turkey). International Journal of Earth. Science, (Geologische Rundschau), 89, pp.781-792.

Loos, S. and Reischmann, T., 1999. The evolution of the southern Menderes massif in SW Turkey as revealed by zircon dating. Journal Geological Society London, 156, pp. 1021-1030.

Oberhänsli, R., Candan, O., Dora, O.Ö. and Dürr, St., 1997. Eclogites within the Menderes Crystalline Complex / western Turkey / Anatolia. Lithos, 41, pp.135-150.

______, Warkus, F. and Candan, O., 2002. Dating of eclogite and granulite facies relics in the Menderes Massif. 1st International Symposium of faculty of mines (Ýstanbul Teknik Üniversitesi) on Earth Sciences and Engineering. Abstracts, 104p.

______, Candan, O. and Wilke, W., 2010. Geochronologic Evidence of Pan-African Eclogites from the Menderes Massif, Turkey. Turkish Journal of Earth Science, (in print).

Oelsner, F. Candan, O. and Oberhänsli, R., 1997. New evidence for the time of the high-grade metamorphism in the Menderes Massif, SW-Turkey. Terra Nostra, 87. Jahrestagung der Geologischen Vereinigung e.v., Fundamental geologic processes, 15p.

Okay, A., 2001. Stratigraphic and metamorphic inversions in the central Menderes Masif: a new structural modal. International Journal of Earth Science, 89, pp.709-727.

Önay, T., 1949. Über the smirgelgesteine SW-Anatoliens. Schweiz. Mineral. Petrol. Mitt, 29p.

Özer, S. Sözbilir, H. Özkar, I. Toker, V. and Sarý, B., 2001. Stratigraphy of Upper Cretaceous-Palaeogene sequences in the southern and eastern Menderes Massif (Western Turkey). International Journal of Earth Science, 89(4), pp.852-866.

______, Sarý. B, Özkar-Öngen. Ý and Toker, V., 2007. Menderes Masifi'nin Üst Kretase - Alt Tersiyer rudist, foraminifer ve nannoplankton biyostratigrafisi: metamorfizma yaþý ve kaya birimlerinin iliþkisine bir yaklaþým. Menderes Masifi Kolokyumu, Ýzmir, Geniþletilmiþ bildiri özleri kitabý, pp.44-50.

Paréjas, E., 1940. La tectonique transversale de la Turquie. Review Faculty Science University Ýstanbul Seri B,5, pp.133-244, Partzsch, J. H. Oberhänsli, R. Candan, O. and Warkus, F., 1998. The Menderes Massif, W-Turkey: A complex nappe pile recording 1.0 Ga of geological history. Third International Turkish Geology Symposium., Middle East Technichal University (M E T U) - Ankara, 281p.

Philippson A., 1911-1915. Reisen und Forschungen im westlichen Kleinasien. Patermanns, Mitt. Helf, pp.1-5, Gotha.

Regnier, J.L., Ring, U.,Paschier C.W., Gessner, K. and Güngör, T., 2003. Contrasting metamorphic evolution of metasedimentary rosks from Çine and Selimiye nappes in the Anatolide belt, western Turkey. Journal Metamorphic Geology, 21, pp.699-721.

Rimmelé, G., Oberhänsli, R., Goffe, B., Jolivet L., Candan O. and Çetinkaplan, M., 2003. First evidence of high-pressure metamorphism in the "cover series" of the southern Menderes Massif: Tectonic and metamorphic implications for the evolution of SW Turkey. Lithos, 71, pp.19-46.

Ring, U. Gessner, K. Güngör, T. and Passcchier., C., 1999. The Menderes Massif of western Turkey and the Cycladic Massif in the Aegean-do they really correlate? Journal Geological Society London, 155, pp.3-6.

______, Johnson C., Hetzel, R., and Gessner, K., 2003.Tectonic denudation of a late Cretaceous-Tertiary collusional belt-regionally symetric cooling Geological Magazine, 140, pp. 421-441.

______ and Collins A.S., 2005. U-Pb SIMS dating of syn-kinematic granites:Timing of core-complex formation in the northern Anatolide belt of western Turkey. Journal Geological Society London, 162, pp.289-298.

Satýr, M. and Friedrichsen, H., 1986. The origin and evolution of the Menderes Massif, W-Turkey: A rubidium/ strontium and oxygen isotope study. Geologische Rundschau, 75/3, pp.703-714.

Schuiling, R.D., 1962. On petrology, age and structure of the Menderes migmatite complex (SW - Turkey). Bulletin Mineral Research Exploration Institute Turkey, 58, pp.71-84.

Seyitoðlu, G. and Scott, B., 1996. The Cause of N-S extensional tectonics in western Turkey: tectonic escape vs back-arc spreading vs orogenic collapse. Journal Geodynamics, 22, 1/2, pp.145-153.

Seyitoðlu, G. Iþýk, V. and Çemen, Ý, 2004. Complete Tertiary exhumation history of the Menderes massif, western Turkey: an alternative working hypotesis. Geological Magazine, 139/1, pp.15-26.

Þengör, A.M.C. Satýr, M. and Akkök, R., 1984. Timing of tectonic events in the Menderes Massif, western Turkey. Implications for tectonic evolution and evidence for Pan-African basement in Turkey. Tectonics 3(7), pp.693-707.

Tchihatcheff, P.De., 1867. Asie Mineure, Description Physique de cette contree, quatrieme partie, Geologie I, Paris.

Thomson and Ring, U., 2006. Thermochronologic evaluation of postcollision extension in the Anatolide orogen, western Turkey. Tectonics, 25p.

Whitney, D.L. and Bozkurt, E., 2002. Metamorphic history of the southern Menderes massif, western Turkey. Geological Society American Bulletin, 114(7), pp.829-838.

BOS SAYFA

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STRATIGRAPHY OF THE PAN - AFRICAN BASEMENT OF THE MENDERES MASSIF AND THE RELATIONSHIP WITH LATE NEOPROTEROZOIC/CAMBRIAN EVOLUTION OF THE GONDWANA

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ABSTRACT.- The Menderes Massif, exposed in the Western Anatolia, is tectonically overlain to the north by the Afyon Zone and to the south by the Lycian Nappes. In the northwest, two high-pressure units, the Cycladic Complex and overlying the Lycian Nappes, as well as the nappes of Ýzmir - Ankara Zone tectonically overlay the Menderes Massif. The metamorphic rock succession of the Massif can be divided into two main units: 1-Pan-African basement (core series) and 2-Palaeozoic - Early Tertiary metasedimentary rocks (cover series). The Pan-African basement shows a stratigraphy consisting of a partly migmatized metaclastic sequence and polymetamorphic basic and acidic igneous rocks that intruded into these metaclastics. This metaclastic sequence, which is composed of paragneiss and conformably overlying schist units reaches a minimum thickness of eight kilometers and forms the oldest rocks of the Pan-African basement of the Menderes Massif. Field studies and geochronological data suggest that the protoliths of the paragneisses are predominantly clastic sediments of litharenitic composition. Frequently, the paragneisses alternate and interfinger in all directions with micaschists and biotite-albite schists, originating from mudstone and subarkosic sandstone, respectively. The paragneisses are conformably overlain by a schist unit. The originally gradual sedimentary contact is represented by a paragneiss and schist intercalation. The schists are dominated by micaschists with biotite-albite schist layers, derived from mudstone and subarkose, respectively. In the Menderes Massif, the paragneisses are generally extensively migmatized. Widespread anatectic granites occur as irregular-shaped bodies with migmatized margins. The detrital zircons of the paragneisses yielded ages scattered between 610 - 2558 Ma. Detrital zircons of the schist unit of the Pan-African basement were dated at 592 - 3239 Ma. The intrusion ages of the orthogneiss showing clear intrusive contact relationship with metaclastic sequence range from 570 to 520 Ma. The granulite facies metamorphism of the paragneiss unit was dated at 583±5.7 Ma. These geochronological evidence and the contact relationships clearly reveal that the deposition age of the protoliths of the metaclastic sequence of the Pan-African basement can be constrained to Late Neoproterozoic between ca 590 - 580 Ma. The metaclastic sequence is intruded by large granitoid bodies and gabbroic stocks. The intrusion ages of the granitic precursors now represented by orthogneisses with changed mineralogical compositions and primary textures range from 520 to 570 Ma with a major event at about 550 Ma. These granites, which can be divided into three main groups (biotite orthogneiss, amphibole orthogneiss and tourmaline leucocratic orthogneiss) are the products of a poly-phase Pan-African acidic magmatic activity that intruded the metaclastic sequence. They are syn- to post-metamorphic intrusions with respect to the Pan-African orogeny. The basic meta-igneous rocks occurring in the Pan-African basement have gabbroic to noritic composition. They display massive cores and foliated margins that consist mainly of garnet amphibolites with relics of eclogites revealing the polymetamorphic history of the Pan-African basement. Field evidence and radiometric age data indicate that the primary contact relationship

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between the Pan-African basement and the Palaeozoic cover series in the Menderes Massif was an unconformity (Supra-Pan-African unconformity) and the cover series were sourced from the Pan-African basement. The protoliths of the paragneiss and schist of the Pan-African basement were deposited on the passive continental margin of a basin occurring between East and West Gondwana during the Late Neoproterozoic time (Mozambique ocean). All the magmatic and metamorphic ages obtained from Pan-African basement coincide with the closure of this ocean and the final assemblage of the Gondwana supercontinent during Late Neoproterozoic-Cambrian time.

Key words: Menderes Massif, Mozambique Belt, Pan-African orogeny, Gondwana, Turkey.

INTRODUCTION

NE - SW trending Menderes Massif (200 x 300 km) forms one of the biggest crustal segments of the Western Anatolia. This crystalline complex is tectonically underlain by Lycian nappes to the south, by Ýzmir - Ankara zone and the extension of Cycladic complex in Turkey to the north and northwest. The massif is covered by Neogene sedimentary / volcanic units to the east. The Menderes Massif is divided into three sub massifs as the Demirci - Gördes sub massif (northern sub massif), the Ödemiþ Kiraz sub massif (central sub massif) and as the Çine sub massif (southern sub massif) by E - W trending graben systems which is still active at present (Figure 1). In previous studies the Menderes Massif was considered as an onion shell. However, today, it is clearly revealed that the Massif tectonically has a complex internal structure defined as thrust faults produced by the Late Alpine compressional tectonism (Konak et al., 1994; Partzsch et al., 1998; Hetzel et al., 1998; Candan and Dora, 1998; Ring et al., 1999; Gessner, 2000; Candan and Çetinkaplan, 2001; Dora et al., 2001).

During studies made in the Massif for a century, prevalent evidences have been obtained indicating the presence of a Precambrian basement exposed in large areas and reshaped by Alpine event (Figure 1). The rock assemblage which is named as "core series", "Precambrian basement" or as the "Pan - African basement" is overlain by the Palaeozoic - Early Tertiary metasedimentary deposit called the "cover series". Evidences show that the primary contact relationship between these two series is an unconformity in regional scale (Supra Pan-African unconformity; Þengör et al., 1984) (Konak et al., 1987; Dora et al., 2005; Candan et al., 2006). It has been known for a long time that the basement in the Menderes Massif is affected by a polyphase metamorphism related with the Pan-African Orogeny (Candan et al., 1994, 2001, 2007; Oberhänsli et al., 1997) and consists of widespread acidic / basic magmatics associated with this period (Hetzel and Reischmann, 1996; Loos and Reischmann, 1999; Koralay et al., 2004; Candan, 1996a). In studies which aimed at establishing the general geological and tectonical structure of the Menderes Massif (Erdoðan and Güngör, 1992; Bozkurt et al., 1993), it is claimed as an opposing idea that gneisses belonging to the basement as the most typical rocks are Upper Cretaceous-Lower Tertiary intrusions (Erdoðan and Güngör, 2004; Bozkurt et al., 1995) and so, there can not be a 'core - cover' relation in the Massif.

In the light of previous studies and original new evidences / interpretations, it is considered that the analysis of basic properties of the units of the Menderes Massif which is considered as the Pan-African in age in Massif scale, and discussion of problems may guide to the investigators who will study on this topic. The article prepared within this scope aims at presenting and discussing: 1- the stratigraphy of metaclastics forming the oldest unit of the Menderes Massif, 2- the migmatization of this sequence,

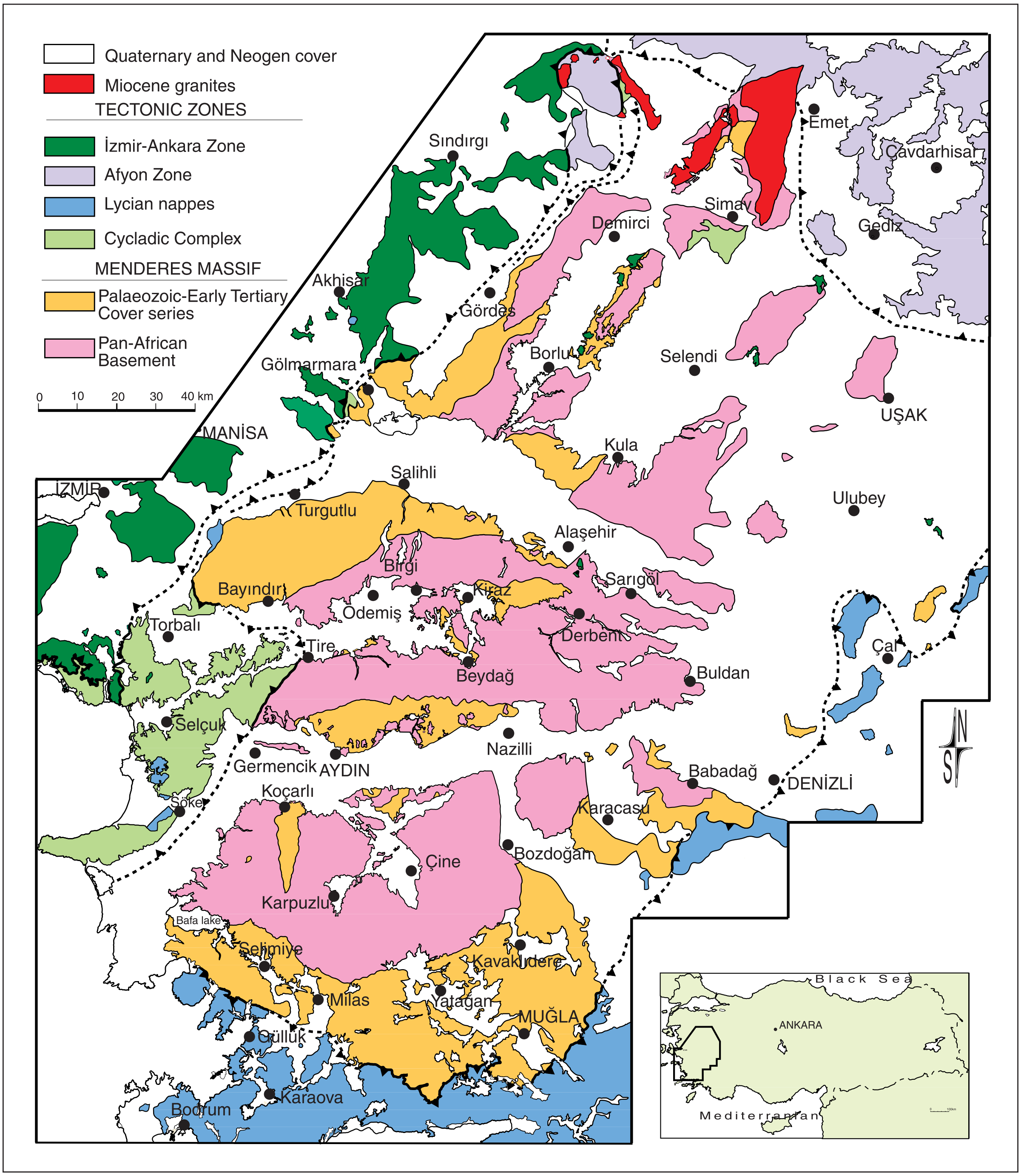

Figure 1- The generalized geology map showing the distribution of the Pan-African basement, Palaeozoic - Lower Tertiary cover series and tectonic zones surrounding the massif.

3- types of acidic / basic magmatics observed within metaclastics and their primary contact relationships with metaclastics, 4- the primary contact relationship between the Pan-African basement and cover series and 5- the temporal and spatial relations of the Pan-African base units with the stage of Gondwana in the Late Neoproterozoic time.

LITHOSTRATIGRAPHY

Recent studies made towards the tectonical structure of the Menderes Massif revealed that the original stratigraphy of the massif has largely been deteriorated by the Precambrian and the Alpine age deformations (Konak et al., 1994; Partzsch et al., 1998; Ring et al., 1999). Nowadays, in throughout the Menderes Massif, the units of the Pan - African basement exposing in large areas are shown in the form of tectonic layers strongly imbricated with Palaeozoic-Early Tertiary cover series near their internal napping. The generalized columnar section obtained by the correlation of tectonic layers made up of different Pan-African basement units is shown in figure 2.

As seen in the figure, the oldest rock units of the Menderes Massif are made up of regular and continuous metaclastic sequence. This metaclastic sequence is cut by granitoidic and gabbroic rocks which intruded at different stages of the Pan-African orogeny. The metaclastic sequence is divided into two sub units as i) paragneiss unit, and ii) schist unit from bottom to top. Rock successions of four regions belonging to northern, central and southern sub massifs are given in figure 3. These two sub units are observed as in primary contact relationships within these regions. At lower parts of the metaclastic sequence, there are gigantic granitic intrusions and / or cut by the Alpine aged thrust faults. Thus, the true thickness of primary sediments of this metaclastic sequence and by which assemblages it is underlain can not be detected. However, it is understood the minimum thickness of the sequence is 8 km, when the general stratigraphical and metamorphic properties of tectonic layers consisting the metaclastic sequence are considered throughout the Massif. The absence of carbonaceous layers is the most characteristic distinguishing feature of this thick metaclastic sequence derived from mudstone, siltstone and sandstone. The presence of emery lenses were detected only in two locations within the schist unit at the Pan-African basement.

The paragneiss unit which is the lower unit of the metaclastic sequence is formed by two lithologies that have both vertical and lateral transitions to each other. These high graded lithologies consist of sillimanite, disthene and garnet (± orthopyroxene) and are made up of schists in different compositions. Litharenitic sandstones are the predominant primary rock type in paragneiss unit. The original layer thicknesses of these sandstones which paragneisses were derived range in between 0.5 - 1 m. Thicknesses of homogenous sandstone layers which form fine grained massive paragneisses may reach 800 meters and can laterally be traced several kilometers. Paragneisses vertically and laterally grade into schists. Although there are many lithologies, the mica schist, biotite and albite schists form the dominant schist types. The protoliths of these schists are mudstone and subarkosic sandstone. Transitions into these schists originate from the lateral and vertical change of primary sand / clay ratios. The thickness of interlayers of schist generally ranges in between 300 - 500 m. However, the thickness of schists where these laterally transit into paragneisses may reach up 2 km. The widespread presence of calcsilicate rocks is another characteristic of the paragneiss unit. These fine grained, massive rocks with a zoning mineralogical composition show severe boudinage. Thickest exposures of the paragneiss unit are observed in the tectonic layer in Kula region. The 4 km apparent thickness observed in this tectonic layer can be accepted as the minimum thickness of the paragneiss unit in the massif.

There has been detected many tectonical layers consisting of the paragneiss unit with the overlying schist unit in the massif (Figure 3). Field data indicate that there is a conformable and transitional contact between these two units. The location of the contact between these two units which are fully derived from the clastic sediments is described by the latest litharenitic layer observed at the deposit. The original thickness of the related schist unit can not be interpreted, because it is present in the form of isolated layers from top and bottom and presence of probable inner nappes and isoclinal folds. For instance, 6 and 7 km apparent thicknesses were estimated in layers of Bozdağlar and Aydýn Mountains, respectively. However,

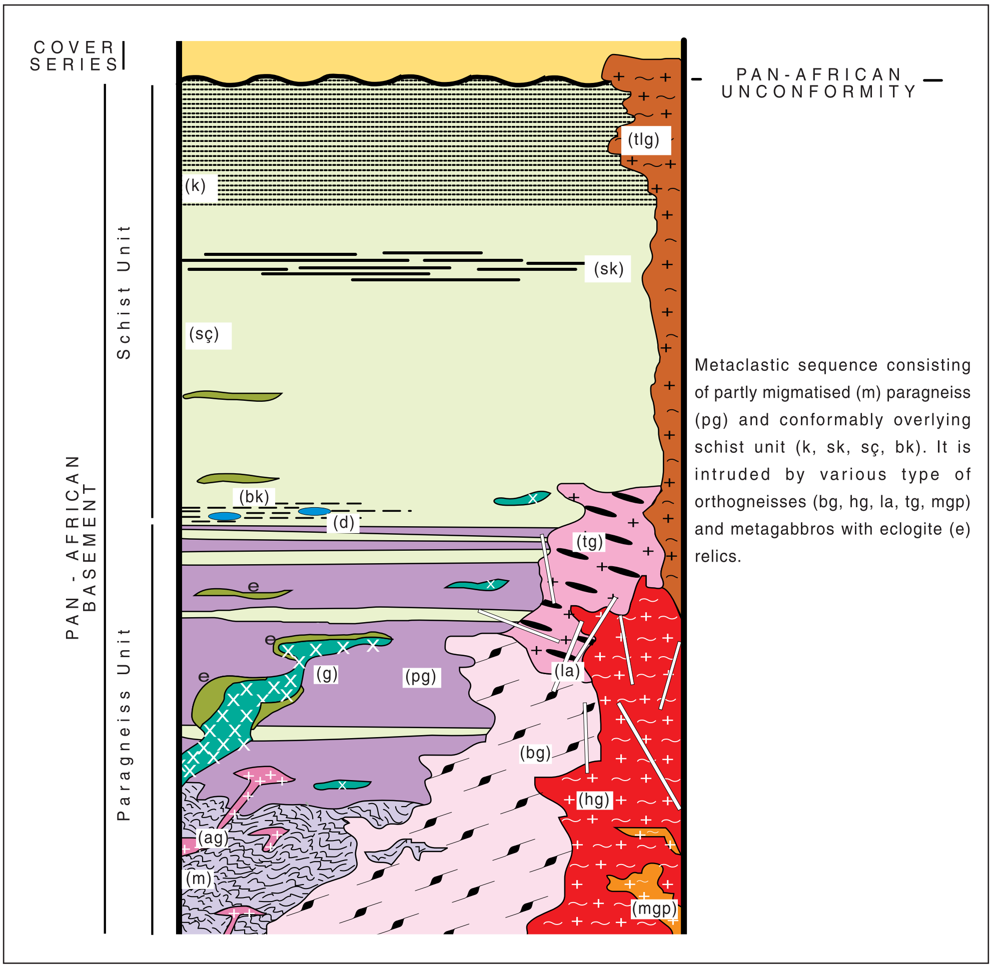

Figure 2- The generalized columnar section of the Pan-African basement of the Menderes Massif. Paragneiss (pg) which is observed partial migmatization (m) and anatectic granite (ag) and the metaclastic sequence which is made up of schists and conformably overlies it form the oldest units of the basement. Schist units are derived from subarkose - mudstone intercalations (sç) at lower layers and sandstones rich in quartz (k) at upper layers. Rare dolomite (d), lensoidal, white quartzite (bk) and black quartzite (sk) layers are observed within schist unit. Orthogneisses are derived from granites rich in biotite (bg), hornblende (hg) and tourmaline (tg) compositions and textures (la: leucocratic metaaplite, mgp: metagranite porphyry). The basement is also cut by Triassic leucocratic granites (tlg).

schists in these two tectonic units have different degree of metamorphisms (Bozdaðlar: garnet, staurolite and disthene zones; Aydýn Mountains: biotite and garnet zones). Therefore, it can be considered that primary thickness for the schist unit could be higher than layers indicated above. The schist unit is predominantly made up of mica schist and the intercalation of biotite - albite schists which its protoliths are equivalent of mudstone and subarkose. Black quartzite layers are also rarely observed within these schists. These graphite rich layers with a thickness of less than 0.5 m may show an intercalation within a zone of 1 km. It is considered that these quartzites are recognized at upper levels of subarkose - mudstone intercalation although their stratigraphical positions have not yet been determined. On the other hand, the schist unit composed of mica schist and biotite - albite schist intercalation at south of Çine sub massif (southwest of Bozdoðan and east of Karýncalý Mountain) is conformably and transitionally overlain by a deposition made up of muscovite schist / biotite muscovite quartz schist intercalation. These schists most probably represent the uppermost levels of schist units belonging to the Pan- African basement.

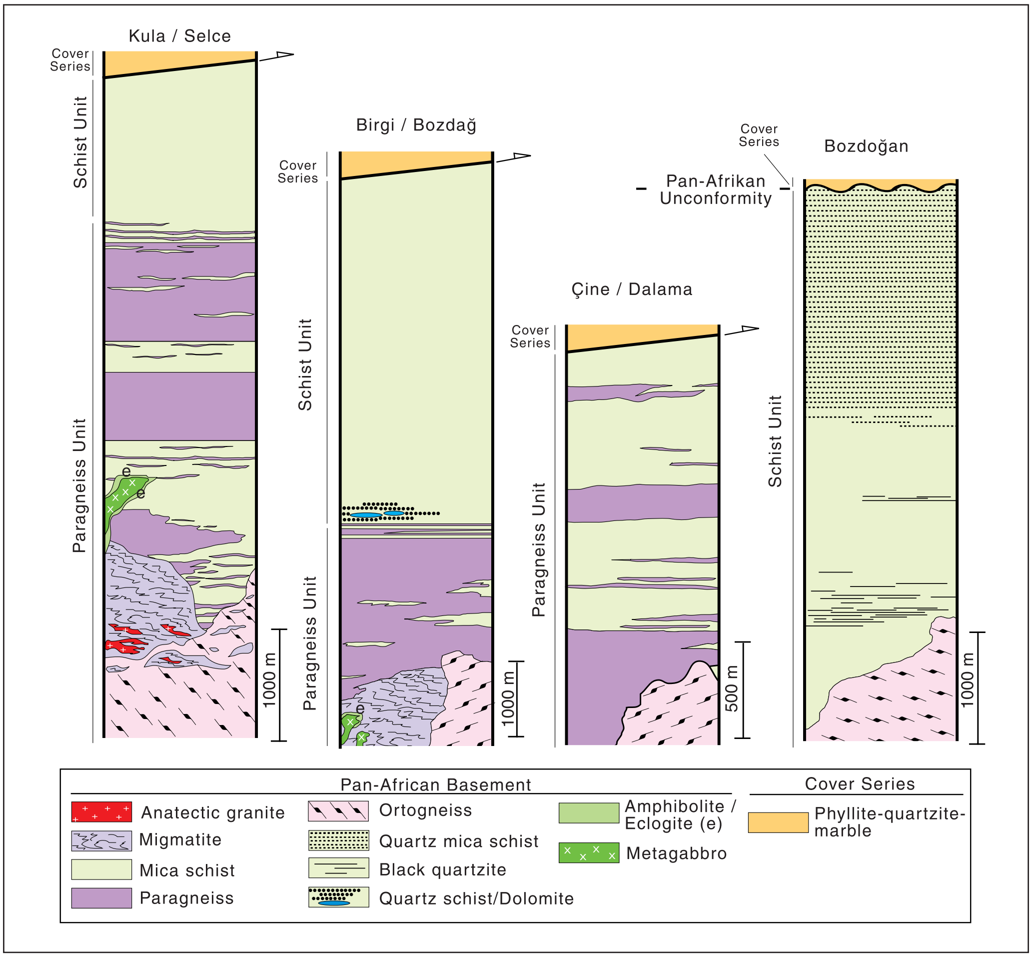

Figure 3- The generalized columnar section of metaclastics of the Pan-African basement observed in Demirci - Gördessubmassif, south of Kula, in Ödemiþ - Kirazsubmassif, north of Birgi and in Çine submassif, south of Dalama and southwest of Bozdoðan.

Another quartzite / muscovite - quartz schist / mica schist intercalation within schist unit are recognized in Bozdaðlar, between Birgi - Alaþehir. Within this quartz arenite / quartz rich sandstone derived intercalation, dolomitic emery lenses which are very rarely observed at the Pan-African basement take place with a dimension of 80 x 200 km. The paragneiss unit which forms the lower parts of the metaclastic sequence show widespread migmatization throughout the massif. In many places, these migmatites are accompanied by anatectic granites which are the product of partial melting. The great majority of these granites which consist of sillimanite and garnet were subjected to in situ crystallization and some of them have reached non migmatized upper levels of paragneisses.

Both schist and paragneiss units are cut by many basic magmatic rocks in vein and stock character with dimensions reaching up 1.5 km. The predominant lithology is the biotite gabbro. These are accompanied by olivine gabbro, noritic gabbro and norites. Petrological evidences show that the primary features of these extremely well preserved rocks in undeformed sections were polymetamorphosed under granulite, eclogite and amphibolite facies conditions. These so called 'eclogitic metagabbro' rocks are observed around the wall zones of gabbro stocks and show widespread transformations into garnet amphibolites. Apart from these, there are eclogites as well in the form of lenses and layers which probably derived from basaltic source rock in many locations within schist and paragneiss. In addition to these, garnet amphibolites derived from vein rocks in gabbroic composition are recognized in almost all parts of the metaclastic sequence.

Gneisses which are granitic origin form one the most prevalent rock types belonging to Pan-African basement. Orthogneisses crop out on an area of thousands of square kilometers. These are made up of plutons intruded each other with diameters reaching up tens of kilometers. Orthogneisses in the massif based on the primary textural and mineralogical compositions of its protoliths can be divided into many sub types such as; granoblastic textured biotite orthogneiss, amphibole orthogneiss, tourmaline leucocratic orthogneiss, metagranite porphyry and metaaplitic vein rocks rich in albite and quartz. These rocks can also be named as granitic, augen and banded gneiss depending on the intensity of ductile deformation. All orthogneiss types present a distinct intrusive contact relationship with paragneiss and schist units which are the oldest rocks of the Pan-African basement. These rocks can be observed as partly assimilated inclusions in orthogneisses with dimensions reaching up several kilometers. Age determinations based on the single zircon evaporation method show that the intrusional ages of granites forming the protoliths of orthogneisses are Precambrian / Cambrian (570-550 Ma; average 550 Ma).

GEOLOGY OF THE PAN-AFRICAN BASEMENT UNITS

The Pan-African basement is made up of Late Neoproterozoic metaclastics which form the oldest units of the Menderes Massif and this basement is intruded by acidic and basic magmatics associated with the Pan-African orogeny.

METACLASTIC SEQUENCE

The metaclastic sequence is divided into two sub units consisting of paragneiss and schist reflecting the facies change in the primary sedimentary rock. Basic geological properties of these partly migmatized typical lithologies of the Pan- African basement are given below.

Paragneiss unit

Macroscopic properties of Paragneisses.- Paragneisses are pinkish gray to brownish, fine grained, massif and / or coarsely foliated rocks. The pink color is originated from the homogenously distributed biotite and garnet in the rock. The primary layer thickness of litharenitic sandstones ranging between 0.5 - 1 m which paragneisses derived from, can still be recognized in paragneisses in spite of this high grade metamorphism (Figure 4a). Within a fine grained homogenous groundmass the presence of widespread mineral dwellings (speckle) ranging from several millimeters to several centimeters is the most characteristic properties of paragneisses (Kun and Candan, 1987*a*; Dora et al., 2001, 2002). In mineralogical studies these speckles were not formed from only one mineral. On the contrary, these speckles were made up of an assemblage defining high temperature (HP) conditions such as sillimanite, disthene and garnet formed by replacing of a previous porphyroblastic mineral. When evidences such as the integration of dwellings are evaluated in the evolution of the Pan-African basement, they have been interpreted as the old cordierite porphyroblasts of the product of granulite facies metamorphism (Dora et al., 2001, 2002). Speckles in paragneisses are divided into three groups based on the color and mineralogical compositions. These are; i) black, ii) green and iii) white speckles. It was identified that the discoidal black speckles reaching up 1 cm in maximum are controlled by the compositional change of the primary sedimentary rock and preferably developed on clay rich levels of the rock (Figure 4b). Much rarely observed green speckle formations could reach 4 to 5 cm and are characterized by structures of zoning composition (Figure 4c). White speckles, 4 - 5 mm in length that have homogenous distribution within a thin crystal pink groundmass (Figure 4d) have been interpreted as relic feldspar phenocrystals belonging to original porphyritic texture. Based on this, protoliths of these paragneisses were considered as volcanites in andesitic composition (Kun, 1983; Kun and Candan, 1987*a*). Whereas, detailed textural observations revealed that these were transformed from black speckles.

Widespread presence of calcsilicates is another characteristic of paragneisses (Kun, 1983; Kun and Candan, 1987*b*; Dora et al., 2001, 2002). Calcsilicates which show very intensive boudinage, are 1 x 0.4 m in dimension in discoidal form appearing several meters apart from each other sometimes (Figure 4e). These very fine grained and massif rocks have a diopsite rich green outer zone, albite rich white colored intermediate zone and pinkish core rich in garnet and zoisite (Figure 4f) (Barbol, 2005). Geochemical data show that these rocks were derived from quartzose-feldspatic levels rich in carbonate among litharenitic sandstones.

The general distributions and internal structures of paragneisses in the Massif.- The distribution of the paragneiss unit belonging to metaclastic sequence in the Pan - African basement throughout the Menderes Massif is given in figure 5. As seen, paragneisses crop out in each of three sub massifs, as being the largest exposure to be in the central sub massif.

In Demirci- Gördes sub massif the primary sedimentary internal structures of paragneisses in the Menderes Massif can clearly be observed in the southern part of Kula. This region is between Alaþehir and Kula, in dimensions of 15 x 17 km and exhibits a character of nappe pile (Figure 6). The layer which consists of the paragneiss unit is tectonically underlain by schist unit belonging to the Pan-African basement and overlain by cover series consisting of Palaeozoic-Mesozoic sequences (Candan, 1994; Dora et al., 2002). The rock succession of the tectonic layer consisting of paragneiss where all the units belonging to the Pan - African basement are observed as in clear contact relationships, is given in figure 3. Oldest units in the region are composed of metaclastics forming a continuous sequence made up of paragneiss and schist. The paragneiss unit possesses an apparent thickness of 4 km and is made up of paragneiss layers rich in sillimanite. This unit has a thickness dominantly varying between 400-600 m and a lateral continuity of 15 km. It also consists of sillimanite-garnet mica schist / sillimanite - biotite - albite schists that show both vertical and lateral transitions with sillimanite rich paragneiss layers. Sections showing the internal structures of

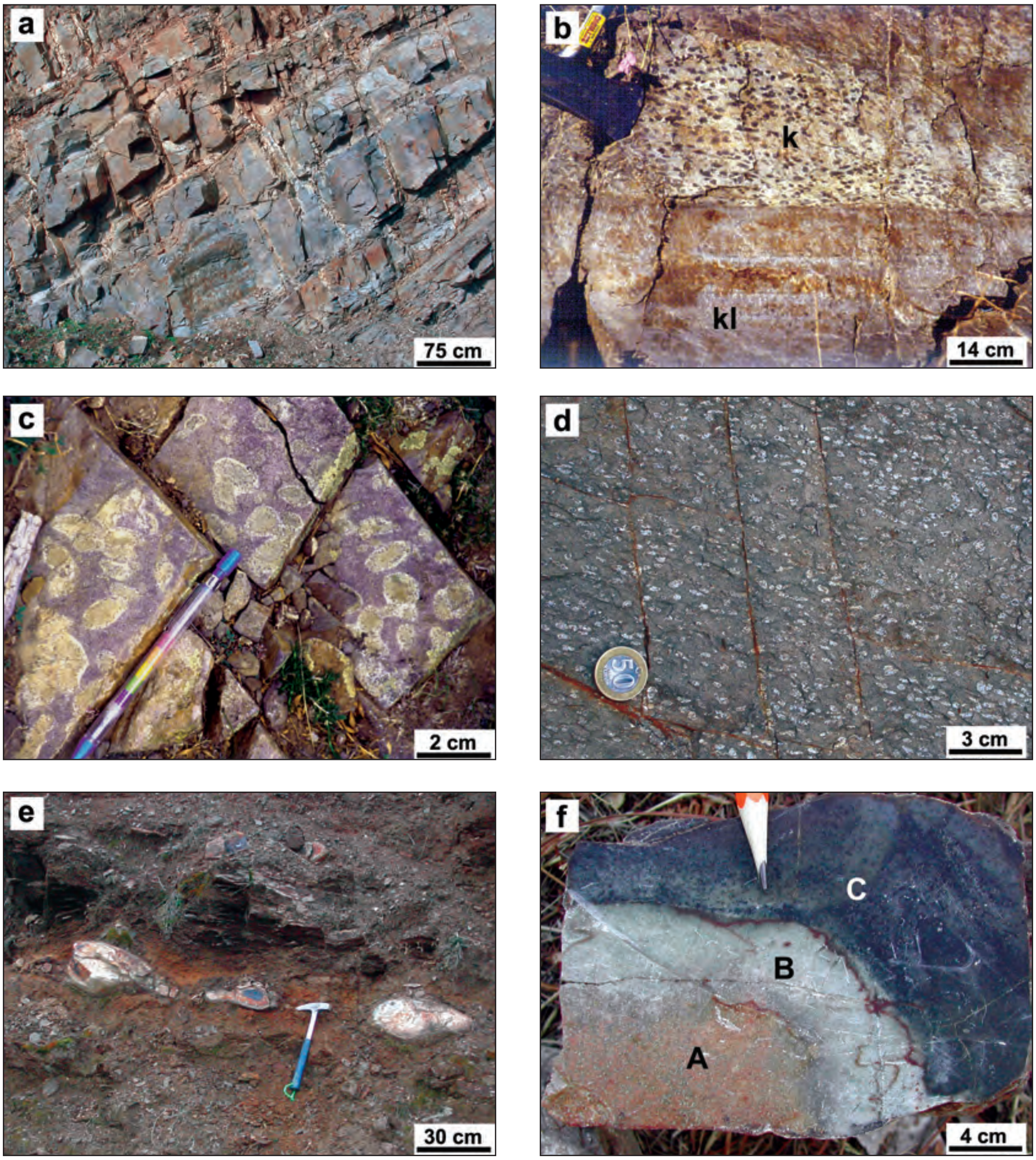

Figure 4- A: Preserved primary coarse layers observed within paragneissesin massive structure derived from litharenitic sandstones, B: Formations of black speckle derived from the probable cordierite porphyroblasts and the product of granulite facies metamorphism affecting paragneisses. Speckling markedly prefers clay rich layers of the primary sedimentary rock (k: sand rich, kl: clay rich), C: Green speckling observed in paragneisses. These speckles are derived from probable cordierite and are pseudomorphically replaced by the zoning mineral assemblage produced from high pressure metamorphism, D: White speckling formed by the formation of retrogradation of black speckles in paragneisses, E: Calcsllicate rocks characterized by boudinage, derived from carbonate rich sediments and widely observed in paragneisses, F: Mineral zoning observed in calcsilicates (A: garnet + zoisite / clinozoisite, B: quartz + plagioclase, C: Clinozoisite + plagioclase) (A - B - D: south of Kula, C: south of Alaþehir, E - F: north of Birgi).

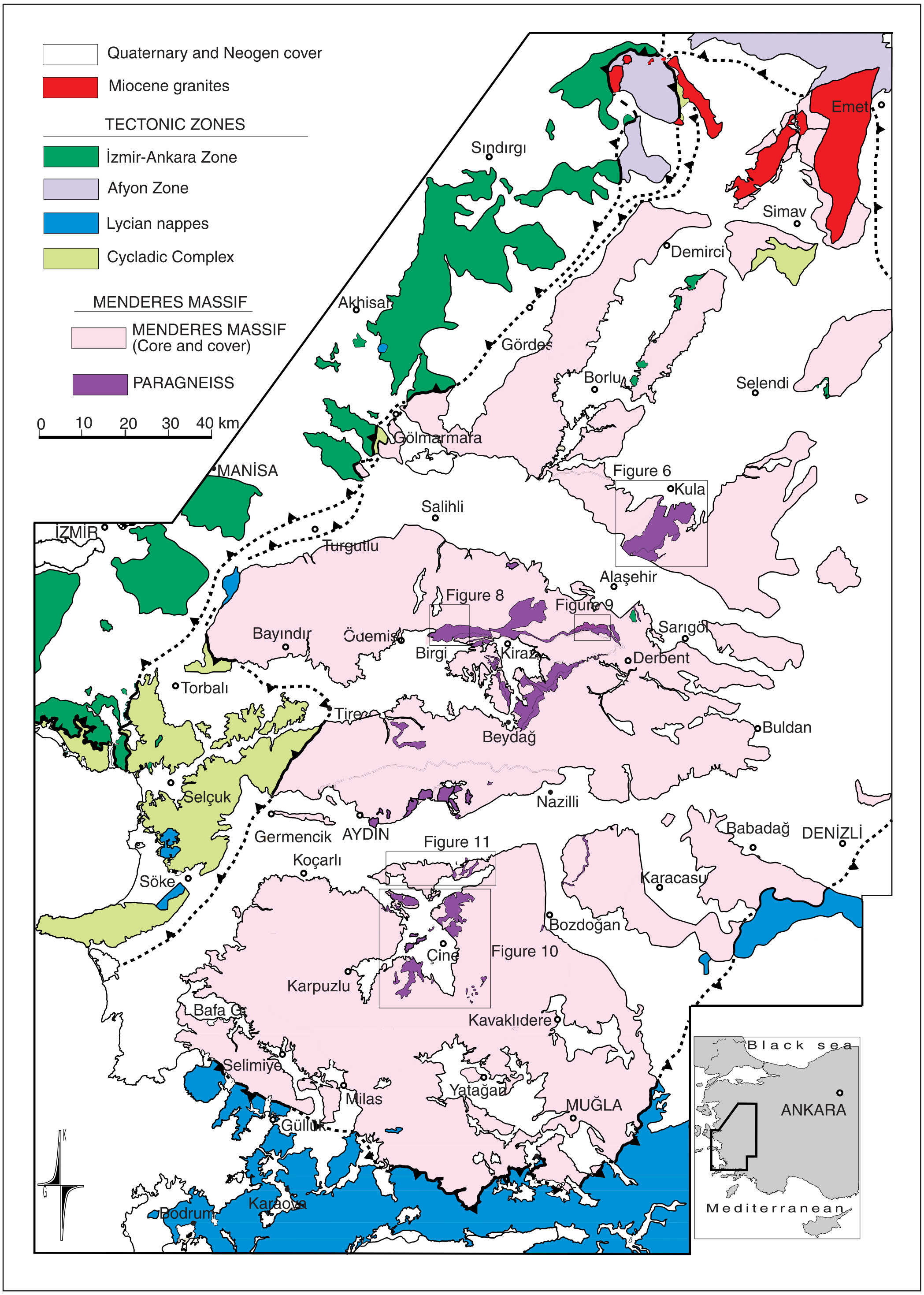

Figure 5- The distribution of paragneiss unit throughout the Menderes Massif. Paragneissic regions of which the geology maps are in detail given in the article are shown.

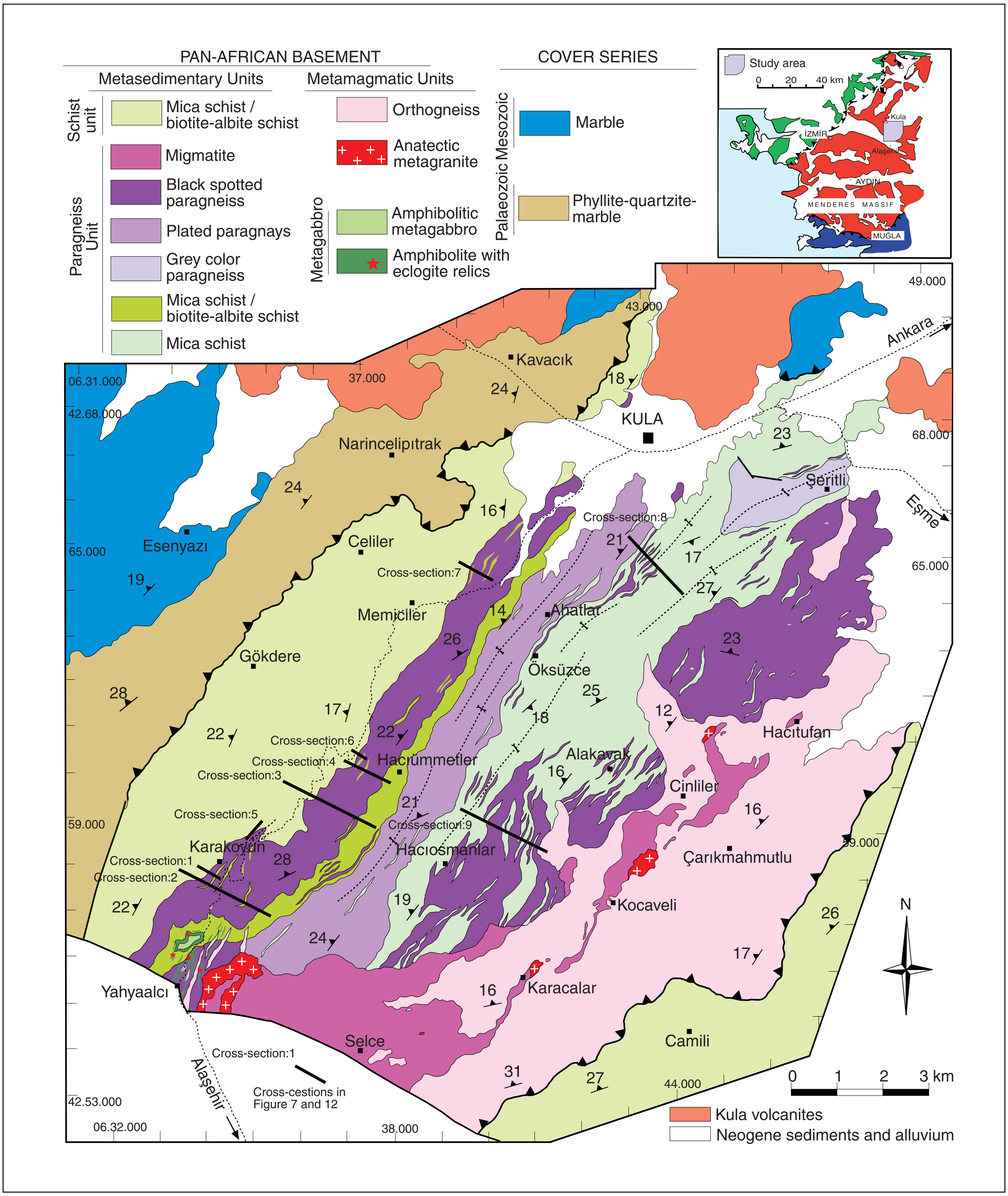

Figure 6- Detailed geology map of the metaclastic sequence of the Pan-African basement which is observed in the Demirci Gördes sub massif of the Menderes Massif, south of Kula (Location is shown in figure 5). Cross lines in figures 7 and 12 are shown on the map.

massif paragneiss layers are given in figure 7a. There are also many paragneiss layers with a thickness not exceeding several meters in schists intercalating with thick paragneiss layers (Figure 7b). In addition to these, intermediate rocks are pervasively recognized as well originating from continuous change in clay / sand ratio in the primary sediment within schists. The paragneiss unit is conformably and transitionally overlain by metaclastic schist unit. Paragneisses which were intensely migmatized at their lower parts are intruded by anatectic granites related to this migmatization. In many locations, basic magmatic rocks are present which intruded into paragneisses in stock and vein character. In amphibolitic circumferential zones of these rocks which are gabbroic in composition rarely eclogite relics are observed (Candan, 1994). The whole sequence is cut by big orthogneissic intrusions emplaced at the last stage of the Pan-African orogeny. Within the orthogneissic mass, partly assimilated migmatite and paragneissic inclusions with dimensions reaching up 2 - 3 km are extremely widespread.

In Ödemiþ - Kiraz sub massif, the units of the Pan-African basement and cover series which were determined as Palaeozoic-Mesozoic in age by fossil evidences, show strong imbrication by the Alpine age compressional tectonism. When the stratigraphy of these imbricated slices are studied, it is clearly observed that paragneisses take place in only one tectonic layer in the Ödemiþ-Kiraz sub massif (Figure 5). The paragneiss unit can laterally be traced 70 km in approximate and was mapped detailed in two locations, to the north of Birgi and south of Alaþehir.

The Bozdað region, the north of Birgi is one of the rarest area which the metaclastic sequence of the Pan-African basement continuously crops out. It has been known for many years that the sequence showing regular southward dipping is overturned by stratigraphy and the degree of metamorphism (Ýzdar, 1971; Kun et al., 1988; Okay, 2001). The thickness of the deposits in the region reaches 8 km (Figure 8). Paragneisses in the region have homogenous internal structure and were derived from litharenitic sandstones with a thickness of 1m. Black dwellings defining old cordierite porphyroblasts are widely observed in paragneisses. Sillimanite rich paragneisses show widespread migmatization at lower parts of their primary positions. To the east of the region, the effect of the partial melting has increased and caused evolution of granitic melt largely, so numerous anatectic granite entrances into migmatites have occurred within dimensions of 6 km.

Kestane river area, the south of Alaþehir is one of the regions where the nappe tectonism in the Menderes Massif and probable Miocene-Recent extansional tectonic structures are all observed clearly (Candan et al., 2001; Gökten et al., 2001). The Pan-African aged homogenous schist unit and cover series form the lower and the upper contacts of the layer which consists of paragneiss in the region respectively (Figure 9). Cover series at the bottom begin with a thin Palaeozoic sequence composed of phyllites. These series grade into Mesozoic age platform type marbles consisting of metabauxite and rudist fossils towards upper layers. The paragneiss unit is being intercalated with schist layers that reach a thickness of 600 m. This is one of the basic properties of paragneiss unit in this area, similar to Kula region. Schists are dominantly made up of garnet mica schists derived from mudstones. Biotite - albite schists originating from feldspar rich sandstones are observed among these garnet mica schists as intermediate layers with thicknesses not exceeding 20 m. In addition to these end members, many intermediate rocks are often observed originating from an infinite change of clay / sand ratio in the primary sediment. When this intercalation is evaluated with the homogenous internal structure in Birgi region, the sight of lateral facies change in paragneisses is frequently encountered as it has been seen in Kula region as well. Purple / pink colored, sillimanite rich, massive paragneisses form the predominant schist type observed in the

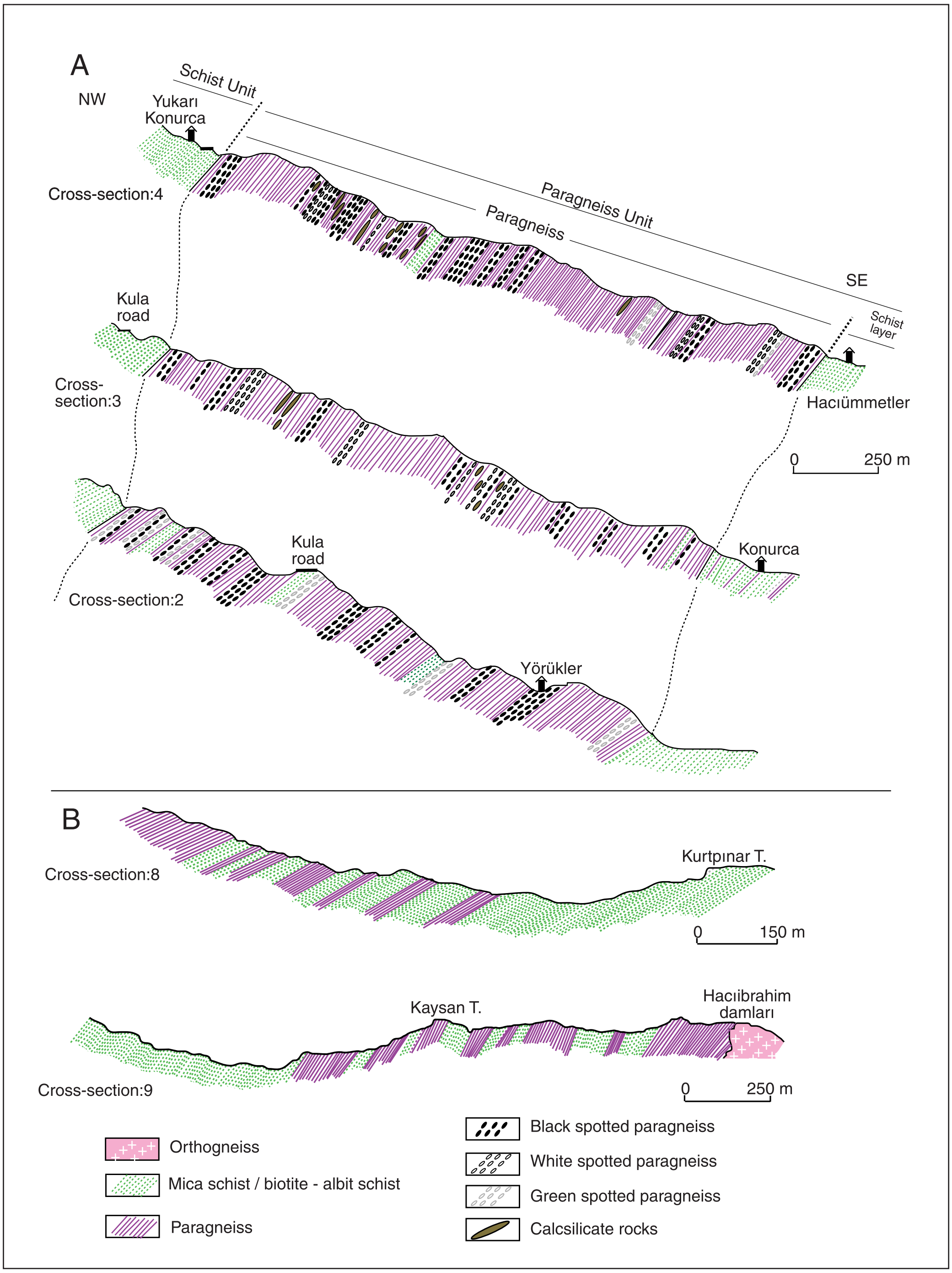

Figure7- A: Sections showing interiors of paragneiss surfaces exposed at south of Kula. B: The internal structure of schist layers within the paragneiss unit. Many paragneiss layers are observed in schistswith a thickness not more than several meters. (section locations are shown in figure 6).

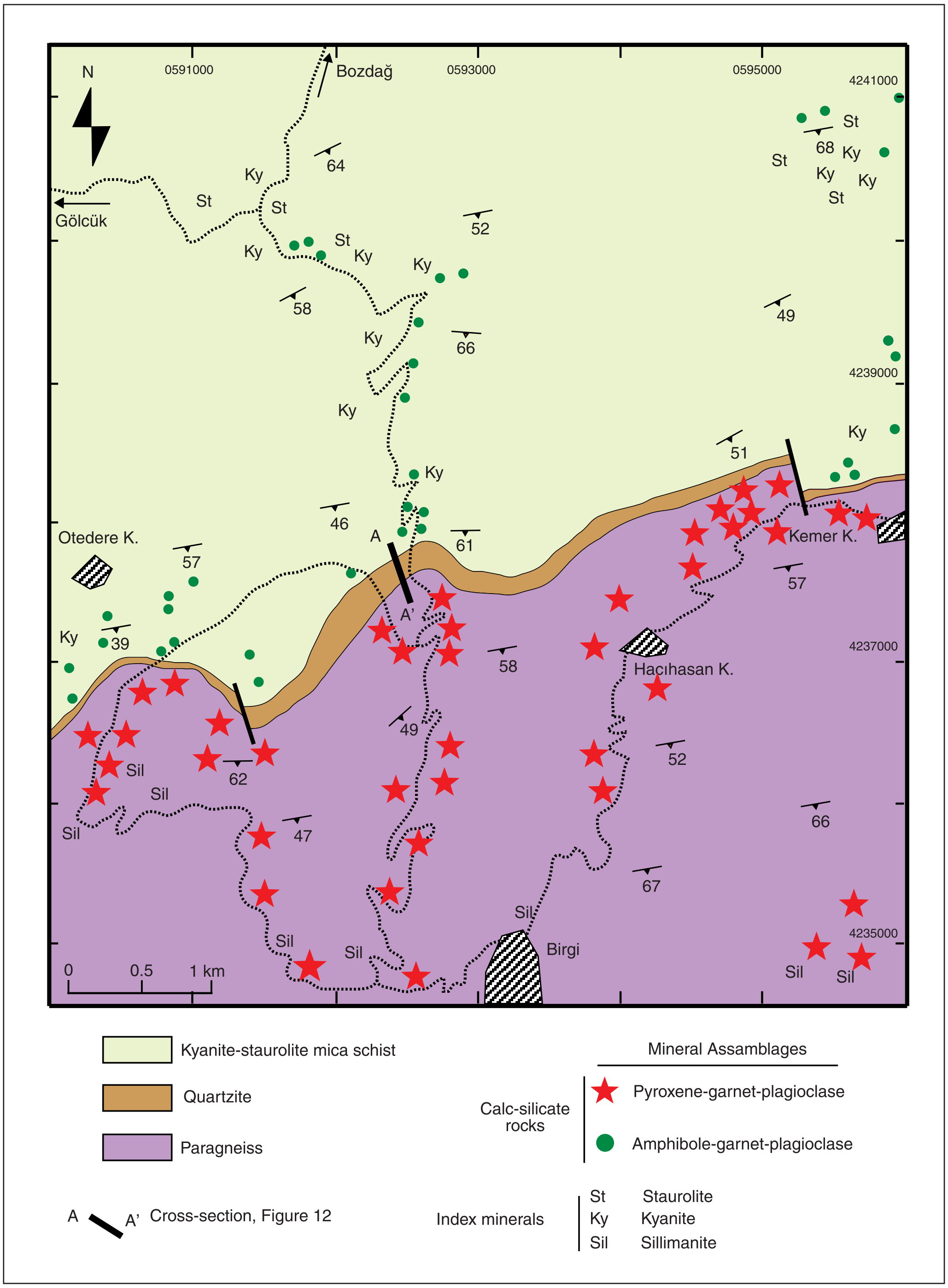

Figure 8- Geological map of Pan-African aged metaclastic sequence observed in Bozdaðlar, north of Birgi. Southward dipping sequence shows an inversion in terms of both stratigraphy and the degree of metamorphism. The succession begins with garnet schists at the bottom and the degree of metamorphism reaches the degree of migmatization at uppermost layer (Dora et al., 2001) (Location is shown in figure 5).

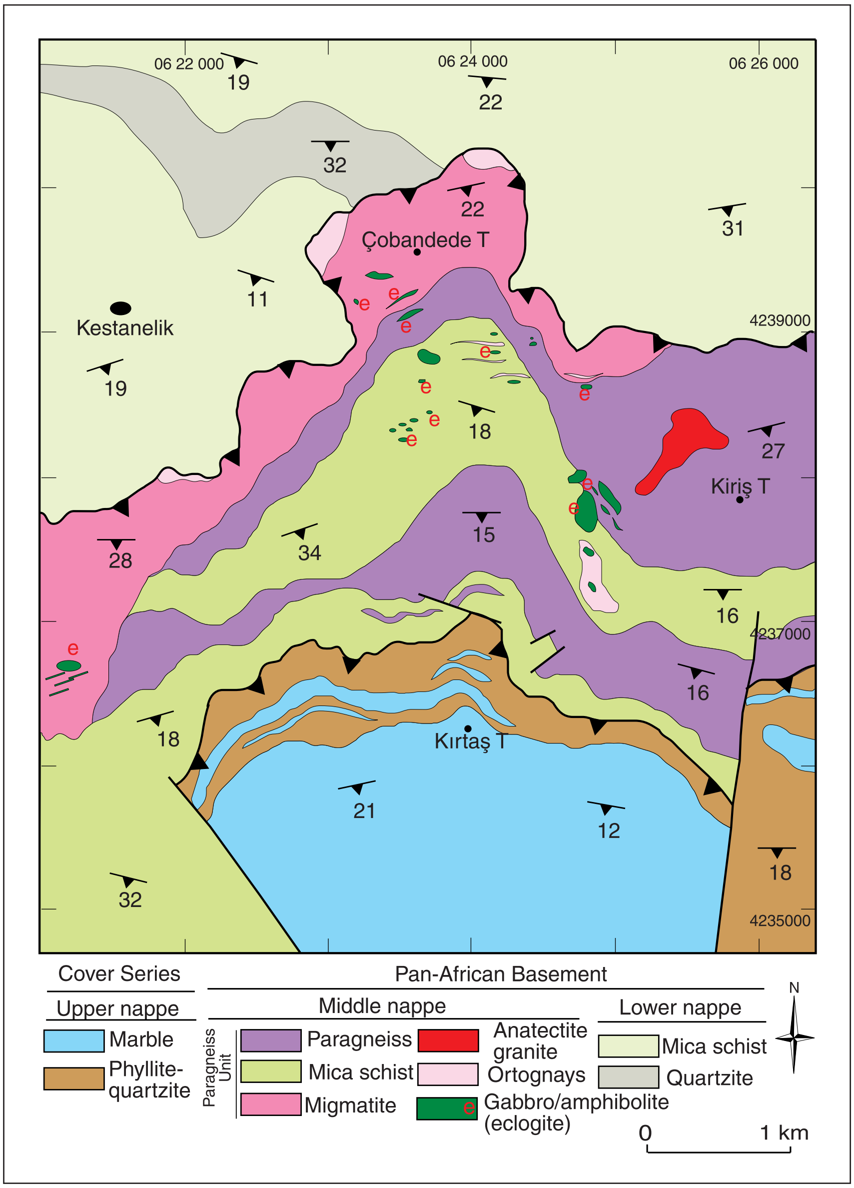

Figure 9- Geological map of paragneiss unit made up of eclogitic lens paragneiss - schist intercalation observed in Kestane River, south of Alaþehir (location is shown in figure 5).

region. Similar to other region, black and green speckles are widely encountered in paragneisses. Lower parts of the paragneiss unit were widely migmatized and cut by the intrusion of anatectic granitic masses associated with these. It is difficult to recognize these structures since migmatitic levels have been subjected to intense ductile deformation. Both in paragneiss and schist layers, there are basic intrusive rocks in vein and stock character. These are gabbroic and noritic in composition with dimensions reaching 300 meters. It was detected that related basic magmatics were transformed into eclogites along their wall zones in more than 20 locations (Candan et al., 2001).

Çine sub-massif is characterized by the presence of big sized granites transformed into augen and banded gneisses at present. Paragneisses do not crop out along the southern border of Bafa - Denizli region of this submassif. Many paragneiss exposures are observed at central and northern parts of the Massif. Paragneisses in this region are in the form of inclusions within dimensions of 7 - 8 km which floats in very big sized orthogneissic masses. On the other hand, the intensive migmatization observed in these rocks cause difficulty in gathering information about the primary internal stratigraphy of Pan-African aged metaclastic sequence (Figure 10).

To the east of Eski Çine, around Ovacýk village there are many inclusions dominantly made up of schists. These schists have lateral and vertical transitions into paragneisses (Baþarýr, 1975). Paragneisses are observed in schists as layers with thicknesses not exceeding several hundreds of meters. However, the dimensions of inclusions within orthogneisses may reach up 6 to 7 km at southwest of Çine. These inclusions are generally formed by high grade migmatized paragneisses. Especially at east of Çine, granitic masses with sizes of 1 km are encountered within migmatites. These fine grained and massif granites consist of garnet and sillimanite and have many inclusions related to paragneisses where these granites were derived from. Dalama region located at the northeastern part of Çine submassif is the best region which the paragneiss / schist intercalation is clearly observed (Çolak, 1985; Dora et al., 2002; Þengül et al., 2006). The region is made up of paragneiss in which orthogneiss intrusions took place, the overlying Pan-African basement made up of schist unit and of tectonically overlying Palaeozoic - Mesozoic aged cover series (Figure 11). Purple / pink colored paragneiss unit which rarely bears black speckles crops out at south of Dalama. Paragneisses consist of schist layers in variable thicknesses ranging from several meters to several hundreds of meters. Schists are transitionally in contact with paragneisses and are predominantly made up of mica schists derived from mudstones. These are accompanied by horizons of biotite plagioclase schist derived from subarkoses similar to other regions of the Menderes Massif.

Contact relationships between paragneiss and schist units.- As described above, the metaclastic sequence which is the oldest units of the Pan-African basement is divided into two sub units as paragneiss and schist units. Schist horizons are encountered in paragneiss unit as well and these two lithologies show both lateral and vertical transitions due to facies change in the primary sediment. The protoliths of schist layers within paragneiss unit and rocks in schist unit mentioned above show a great similarity. So, it makes difficult to determine the character and to find the position of the contact between these two sub units of the metaclastic sequence in many places of the Massif. Despite all these problems, this contact relationship on either regions of the Menderes Massif is clearly observed. The first of these is the region between Kula - Alaþehir. The contact between paragneiss and schist units show a lateral continuity of 15 km (Figure 6) and can be recognized in many locations. The contact between two units is defined by the latest litharenitic sandstone layer observed at the deposition. Four measured sections taken along this contact are given in figure 12a. As seen in figures, the gradation from paragneiss to schist occurs in intermediate zones of maximum 100 meters. Within this zone there is a distinct increase in intermediate schist levels. In addition, intermediate rocks which can neither be defined as paragneiss nor as schist are frequently encountered. After the latest paragneiss level, schist units are encountered with a transition defined by the gradational increase in mica schist in a narrow zone less than a meter. After this contact, it is not observed any layer which

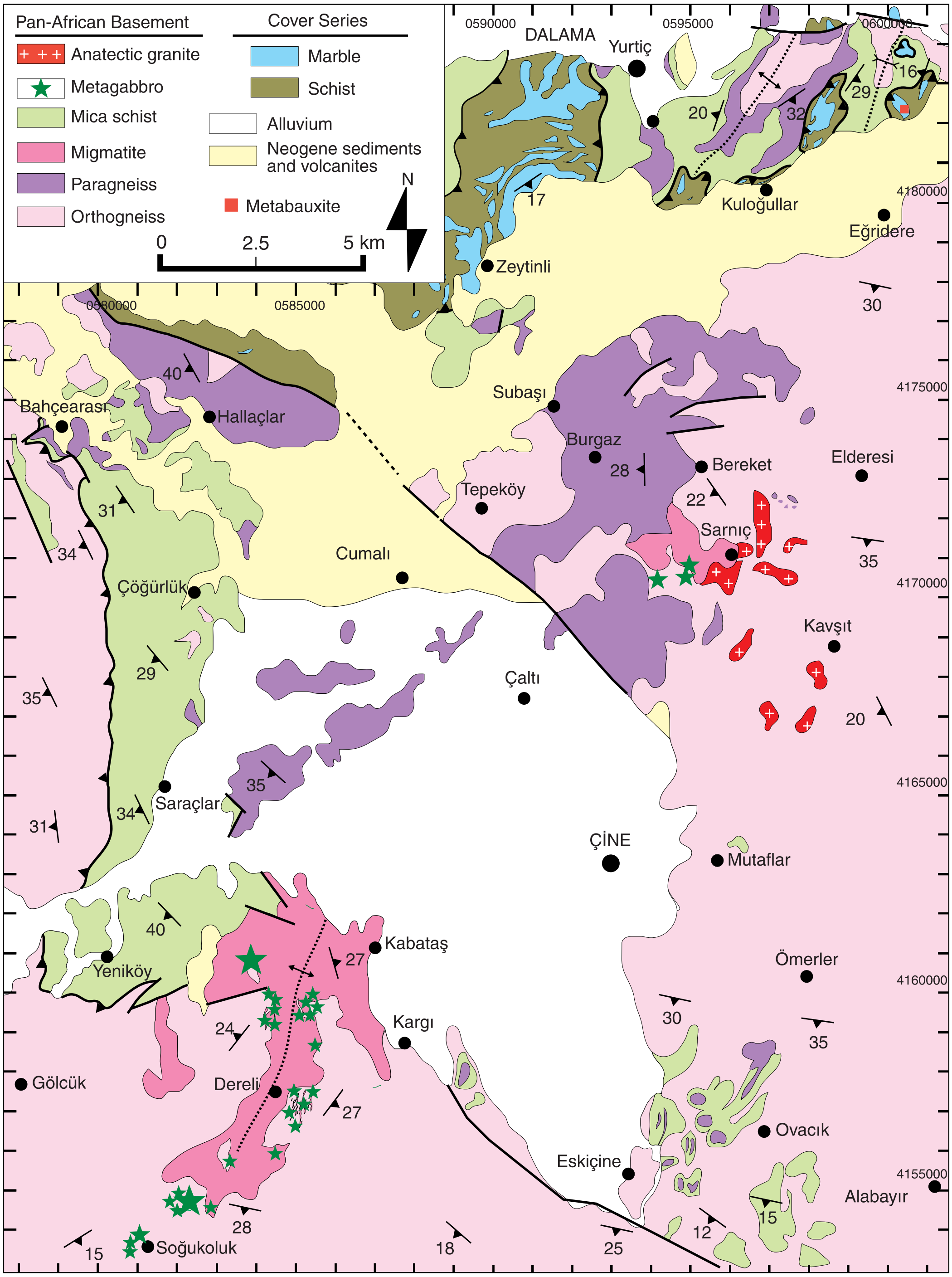

Figure 10- The general distributions of migmatizedparagneisses. These are exposed in the form of large and small inclusions in granitic gneisses at the middle and northern part of the Çinesubmassif (the map was compiled from Schulling, 1962; Baþarýr, 1975; Kun, 1983; Çolak, 1985; Candan, 1996*a-b*; Candan and Dora, 1998 and Þengül et al., 2006) (location is shown in figure 5).

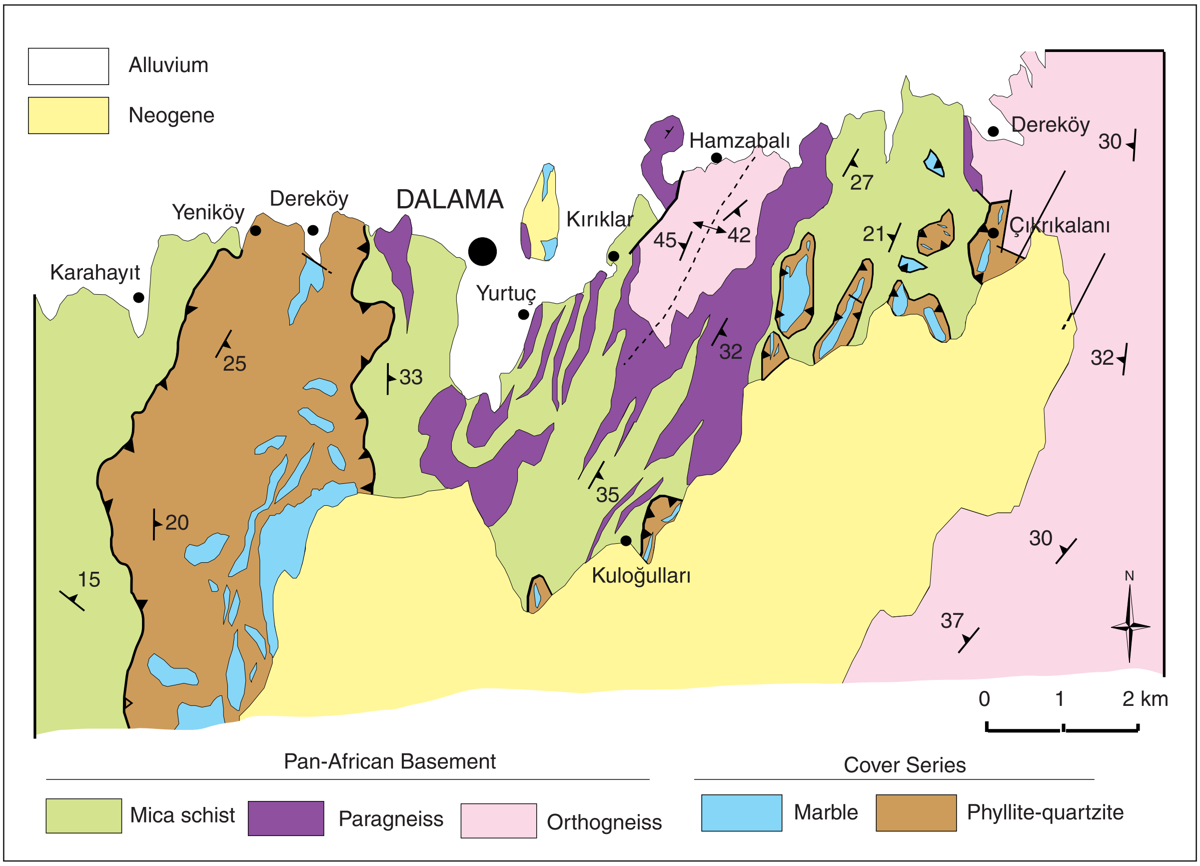

Figure 11- The geological map of Dalama surround where the paragneiss unit is observed at the northern part of Çine submassif (modified from Çolak, 1985; Dora et al, 2002; Þengül et al., 2006).

might be defined as paragneiss within the schist unit having a thickness of 2 km's in the region.

The second region where the contact relationship is markedly observed is the northern part of Birgi (Figure 8). The contact relationship in this region was previously defined by Dora et al. (2001). The contact between the two units can be traced approximately 2 km's. The distinct relationship between two overturned units can clearly be observed at a section where Bozdað - Birgi road cuts the contact (Figure 12b). The contact is conformable and transitional in this region and consists of some differences than in Kula region. The transition from paragneiss that have homogenous internal structure with a thickness of 2.5 km into schist occurs at 20 m zone. This zone is described by widespread presence of intermediate rocks in addition to frequent intercalations of schist - paragneiss horizons. After the latest paragneiss layer 80 m homogenous schist is traversed and begins an intercalation of metaquartzite - quartz schist - mica schist. This intercalated zone is defined by white colored, pure metaquartzite layers (0.5 - 5 m in thickness). It has a maximum thickness of 170 m and can laterally be traced about 35 km up to Alaþehir. Besides, amphibolitic layers and metaaplitic sills are encountered within this intercalating sequence. In addition, emery lenses are present in quartzites in two locations in dimensions of 70 x 150 m. These rare carbonate lenses at the Pan-African basement of the Menderes Massif are located at north of Birgi, in Yýlanlý Kale

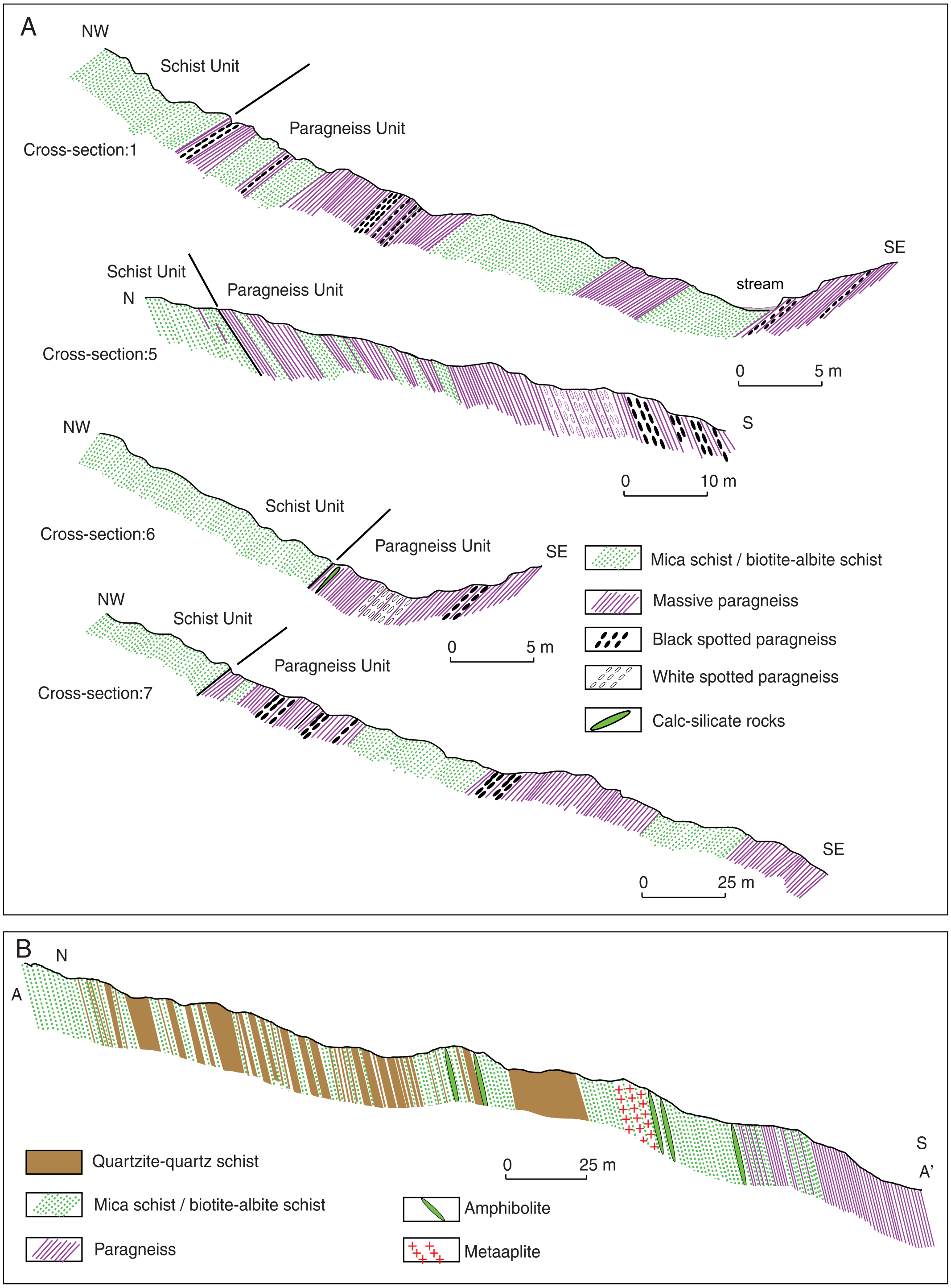

Figure 12- Geological cross sections showing conformable and transitional contact relationship between paragneiss and schist units forming the metaclastic sequence belonging to the Pan-African basement. A: south of Kula, B: north of Birgi (locations are shown in figures 6 and 8).

and at south of Alaþehir / Azýtepe. This intercalating sequence is overlain by disthene - staurolite schists derived from subarkose - mudstone intercalation that have a thickness of 6 km.

Microscopic properties of paragneisses.- Fine grained, purple colored paragneisses without speckle form the most prevalent paragneiss type in the Menderes Massif. These rocks are generally massif and gain a platelet structure in ductile shear zones. Massif structured paragneisses are characterized by their fine to medium crystalline granoblastic - polygonal textures. In petrographical studies, related paragneisses are divided into three subgroups based on the contents of sillimanite and orthopyroxene. Non sillimanite massif paragneisses composed of 'biotite + plagioclase (±orthoclase) + garnet + quartz ± muscovite ± rutile ±zircon ± opaque mineral (ilmenite)' and derived from aluminum poor sediments present a polygonal textural structure in a way to characterize the high temperature metamorphism. Fine to medium grained, dark brown biotite crystals within a groundmass of euhedral garnet and quartz (+ plagioclase) crystals show a random distribution in these paragneisses. The ratio of sillimanite may reach up 15% in massif paragneisses consisting of sillimanite. Sillimanites in these rocks could be observed in two different positions. Sillimanites which were developed by the pseudomorphic replacement of biotite crystal form the prevalent type of sillimanites. Other sillimanite formations are the fibrolitic sillimanite crystals developed at two feldspar or feldspar / quartz contacts. In these formations thin sillimanite crystals present a comb texture located at the contact of two minerals. Foliation planes developing in samples of this type which were subjected to ductile deformation are defined by needlelike sillimanite crystals showing parallel growth to each other (Figure 13 a). Orthopyroxene bearing massif paragneisses are the most rarely observed type. In orthopyroxene paragneisses extremely complex textural relations reflecting the polymetamorphism were observed. These rocks are extremely rich in sillimanite with a ratio of up to 20%. Sillimanites are generally observed in the form of pseudomorphic dwellings developed by replacement of coarse biotite crystals (Figure 13b). In addition to these, fibrolitic sillimanites are also widely observed which were developed in the form of comb between the contacts of two feldspars (Figure 13c).

Speckle bearing paragneisses are divided into three groups based on the composition of speckles. Black speckled paragneisses form the most prevalent type. These are made up of a groundmass composed of fine grained crystals and speckle structures presenting zoning composition changes in it. The general mineral paragenesis of the groundmass consisting 70% of the rock is 'biotite + plagioclase + quartz + garnet ± sillimanite (± disthene) ± muscovite ± orthoclase ± rutile ± zircon'. Speckles on the other hand are made up of 'biotite + sillimanite + garnet + quartz + muscovite'. Hundreds of thin sections were prepared and a relic that could directly define primary mineral in dwellings was encountered in none of them. It is suggested that these speckles have been derived from cordierite porphyroblasts characterizing an earlier metamorphism by their macroscopic pictures, the rarely seen primary crystal forms avoided from deformation, and the speckles developed in granulite facies conditions during multiphase metamorphic evolution of the Pan-African basement. The zoning mineral composition is the most typical properties of black speckles (Figure 13d). In this structural type, there is a euhedral garnet crystal at the center of dwelling. Garnets are also encountered as in the form of crystal assemblages showing interstitial growth. The central part is made up of garnet which is surrounded by an intermediate zone with a composition of 'biotite + sillimanite + (± disthene) + quartz'. Sillimanites in this zone are made up of extremely thin fibrolitic crystals. Since crystal sizes are very small, it is often too difficult to make a distinction between disthene - sillimanite. In some samples, the presence of late stage muscovites in thin crystals was encountered. These muscovites are thought as crystals retrograded from sillimanite which is the product of overlying retrograded metamorphism. At the outermost part of the speckle, there is a white / gray colored circumferential zone composed of 'sillimanite + quartz'.

The green speckled paragneisses are characterized by ellipsoidal speckles in 7-8 cm dimensions presenting a homogenous distribution in a fine-grained groundmass as similar to the ones with black speckles. The general mineral paragenesis of the groundmass forming the 60% of the rock is made up of 'biotite + plagioclase + quartz + garnet ± disthene ± sillimanite ± muscovite ± rutile ± zircon'. Speckles are composed of 'biotite + disthene + (± sillimanite) + garnet + quartz'. As clearly seen in figure 13, a zonal replacement texture is observed in green speckles. The main body of this structure is made up of green colored 'biotite + disthene + garnet + quartz (±muscovite)' with a homogenous distribution. The homogenous ensemble forming the main body is surrounded by a partly developed intermediate zone made up of a coarse single biotite crystal. This biotite zone can easily be distinguished from other fine grained biotites forming the body with its dark red colors. In macroscopic observations, it is markedly seen that speckles are surrounded by a white colored outer zone. Although it can not be discriminated by distinct borders, this mica poor outer zone originates from relative enrichment of quartz and feldspar.

White speckled paragneisses are described by the formations of ellipsoidal shaped, white colored speckles with dimensions of 0.5 cm within a pink colored fine grained groundmass. In addition to biotite, the muscovite is also encountered in fine grained matrix. The mineral composition of the matrix that has distinct foliation is 'biotite + quartz + muscovite + plagioclase (± garnet)'. White colored dwellings are made up of 'muscovite + quartz + garnet + (± sillimanite ± biotite)'. At centers of white dwellings garnet takes place made up of one or two crystals (Figure 13f). In subhedral garnets diffuse quartz inclusions are encountered. In some dwellings one or two biotite crystals may rarely accompany garnets. The rest of the dwelling is made up of fully quartz and muscovite. In some dwellings the presence of irregular sillimanite patches was determined. Textural data indicate that white colored mineral dwellings are formed by the replacement of the muscovite with sillimanites in former black speckles, as a result of the overlying low temperature metamorphism.

Schist unit

As explained above, the metaclastic sequence is formed by paragneiss with schist intercalation and the overlying schist unit. Schists within these two units show big similarities with each other in terms of source rock and mineral compositions. Therefore, to avoid repetition, the all schists in metaclastic sequence were considered under one title.

Macroscopic properties of Schists and their general distributions.- Schists form the predominant rock type in the Pan African basement. Almost all parts of NE-SW trending Gördes Demirci highlands with dimensions of 60 x 10 km are made up of partly migmatized schists in Demirci Gördes sub massif at north. In distant sections of schists from migmatization, the disthene, staurolite and garnet porphyroblasts reaching 4 - 5 cm in dimensions are pervasively encountered (Candan and Dora, 1993). The size of disthene crystals reach up 40 - 50 cm in quartz rich pegmatoids that show parallel growth to foliations of schists. It is considered that these pegmatoids were formed by the emplacement of Si and Al which were migrated from the country rock along the foliation plane during metamorphism (Candan, 1991). Sillimanites accompany these minerals at lowermost parts of schist series. These biotite and muscovite rich schists were dominantly derived from claystones.

Ödemiþ - Kiraz sub massif is made up of tectonic layers belonging to Pan African core and cover series overlapped on each other. The layer

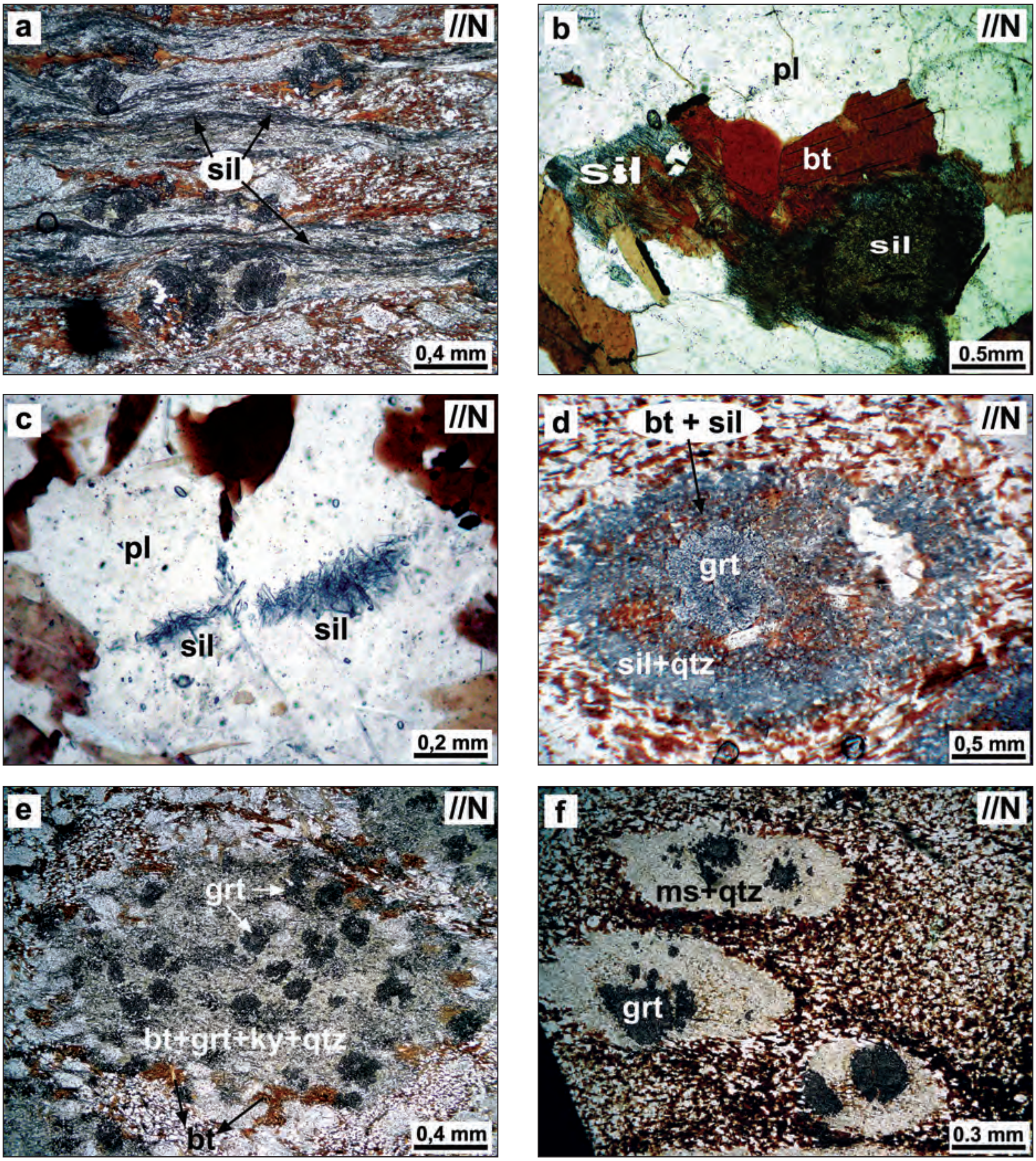

Figure 13- A: Sillimanite rich paragneissessubjected to ductile deformation. Foliation plane is defined by sillimanite needles. B: Sillimanites developed by rthe replacement ofbiotite crystals, C: Sillimanite crystals developing at walls of feldspar. D: Black speckles made by the assemblage of upper amphibolite facies that present mineralogical zoning and replaced cordierite porphyroblastswhich is the granulite facies product. E: Zonal interior structures of dwellings in green speckled paragneisses. F: Microscopic views of white speckles derived by the replacement of sillimanite in black speckles by muscovite (Bt: biotite, sil: sillimanite, pl: plagioclase, grt: garnet, qtz: quartz, ms: muscovite, ky: disthene).

which is observed in Aydýn Mountains crops out in a region of 15 x 90 km and is fully composed of schists. This schist sequence has homogenous internal structure and was derived from the intercalation of mudstone - subarkosic sandstone. The most characteristic lithology of the schist unit, the biotite - albite schists were produced from subarkosic sandstones. These are also defined by the presence of single and coarse biotite crystals showing homogenous distribution within a white colored quartz feldspatic groundmass. Petrographical / petrological data indicate that this sequence which is composed of biotite and garnet schists is in overturned position in terms of the degree of metamorphism (Okay, 2001). Bozdaðlar which form the northern part of Ödemiþ - Kiraz submassif is made up of paragneisses and of conformably overlying schist units (Figure 3). This schist series are derived from a source rock similar to that of Aydýn Mountains. However, these schists consist of disthene and staurolite minerals defining lower - middle amphibolite facies conditions. Petrographic data show that schist series as well are in overturned position (Ýzdar, 1971; Dora et al., 2001). The abundant presence of layers composed of 'hornblende - garnet - quartz - plagioclase' is one of the most characteristics of schist unit. Barbol (2005) stated that these layers which are described by the presence of amphibolites were transformed into boudinaged calcsilicates consisting of pyroxene within paragneisses as a result of the increasing degree of metamorphism.

Pan African units within Çine submassif are made up of metaclastics cropping out in narrow areas and of gigantic orthogneisses intruding into them. Schists belonging to metaclastic sequence observed in Dalama and to the west of Karýncalý Mountain are generally composed of garnet - mica schist and biotite albite schist (Kun, 1983; Þengül et al., 2006). Despite that, schists located at west of Karacasu and south of Bozdoðan consist of rarely observed lithologies in the remnant part of the Menderes Massif. These rocks define the uppermost levels of the schist unit and present a regular schist deposit at south of Bozdað (Figure 3). At upper levels of schists derived from the subarkose - claystone intercalation black colored quartzite interlayers are observed. These quartzite layers which have variable thickness in 5 - 50 cm are intercalated with mica schists in a 1 km zone and can laterally be traced 3 km's. Similar black quartzite layers are encountered at south of Tire and south of Adagide as well. This deposit is conformably and transitionally overlain by quartz rich schists. These rocks form uppermost layers of schist unit reaching a thickness of 2 km. These are silver to white in color and have varying muscovite / quartz ratios. These rocks were derived from quartz and / or clay rich sediments and were made up of muscovite schist and muscovite - biotite - quartz schists with vertical and lateral transitions in cm scale. In addition to these, schist unit of the Pan-African basement lies as a thin line along the contact of orthogneiss - schist of Çine submassif between Bafa Lake and Yataðan. These schists show an intrusive contact relationship with orthogneiss and are predominantly made up of biotite - albite schists derived from subarkosic sandstones.

Microscopic properties of Schist unit.- As also described above, primary sediments of schists of the Pan-African basement can be divided into 3 main lithologies although these consist of continuous interlayer. These are; 1) subarkosic sandstones, 2) quartz rich sandstones and 3) mudstones.

Rocks originating from subarkosic sandstones form 'biotite - albite schist' and 'sillimanite - garnet - biotite - albite schists'. The general mineral composition of these plagioclase rich rocks was determined as 'quartz + plagioclase (albite) + biotite + garnet ± muscovite ± sillimanite ± rutile ± ilmenite ± zircon'. A continuous schistosity, made up of homogenously distributed coarse individual biotite crystals, is observed in these rocks (Figure 14a). Garnets in biotite - albite schists are generally in the form of subhedral small crystals. Rocks which show lateral transitions with paragneisses consist of very fine crsytalline fibrolitic sillimanite developed between the two feldspar contacts.

Quartz rich sandstones were transformed into 'muscovite schist / garnet muscovite schist' and 'muscovite - biotite - quartz schist' as these sandstones had been metamorphosed under conditions of upper greenschist facies. Main components of these rocks are quartz and muscovite. Muscovite and quartz ratios in these rocks vary between 30 - 60% and 40 - 70%, respectively. Besides, biotite, garnet and hornblende are observed in these rocks as well. Hornblendes form porphyroblasts with size of 3 cm and show poikiloblastic texture because of dense quartz inclusions.

Mica schists originating from mudstones are the most widespread schist type in the Massif and show quite different compositions as a function of degree of metamorphism. These schists consist of ensembles of Barrowian type medium pressure metamorphism and define development conditions extending from biotite to sillimanite. Biotite schists which are observed at south of Aydýn Mountains and forming the lowest graded rocks of the deposit are composed of 'quartz + plagioclase + biotite + muscovite ± zircon'. By the addition of garnet to this assemblage, it is passed through the most prevalent schist type which is garnet mica schists. These rocks are composed of 'quartz + plagioclase + muscovite + garnet + biotite ± zircon'. Polyphase growing structures are pervasively observed reflecting the polymetamorphic stage in garnets (Figure 14b,c). Staurolite which defines the transition into almandine - amphibolite facies conditions is widely observed in Bozdaðlar and around Demirci - Gördes. These rocks are 'quartz + plagioclase + staurolite + garnet + biotite + muscovite ± zircon ± apatite' in composition and majority of staurolites are in syn tectonic structure (Figure 14d). Disthene schists are appeared in ensembles in which staurolite disthene accompanies and staurolite disappears (Figure 14 e). The general mineral composition of these rocks can be given as 'quartz + plagioclase + disthene + garnet + biotite + muscovite ± zircon ± apatite'. Disthene crystals which might reach 5 - 6 cm in size give a characteristic porphyroblastic texture. Schists which intercalate especially with paragneisses include sillimanite. The ratio of these minerals is less than 1% and observed in the form of thin, fibrolitic crystals between the contacts of two feldspars. The mineralogical contents of these rocks are 'quartz + plagioclase + garnet + biotite + sillimanite ± zircon ± apatite'.

Migmatization and anatectic granite.- The metaclastic sequence of the Pan African basement consist of prevalent data related to partial melting and the development of anatectic granite in many places of the Massif. Schists with disthene that belong to schist unit cropping out in large areas at north, around Gördes - Simav region has been known for many years (Ayan, 1971; Dað and Dora, 1991; Akdeniz and Konak, 1979). These rocks are defined as layered migmatites and are intruded by many pegmatites and granites. However, in central and southern parts of the Menderes Massif, all migmatized rocks were derived from the intercalation of paragneiss - schist which corresponds to lower layers of the metaclastic sequence (Figure 15a). The relation of migmatization - granite development is best observed in region between Alaþehir - Kula (Figure 6). The migmatitic front in the region obliquely cuts primary stratigraphy of paragneiss unit. Migmatization begins with ptigmatic leucocratic zones and continues until the development stage of anatectic masses with dimensions of 1- 2 km's. Great majority of these granites makes transitional contacts with migmatites in a way that shows in situ crystallization. It is still possible to observe partly assimilated relics of migmatite even at highest graded melted parts of these masses. Various migmatite types are widely observed in the region. Masses formed by the ensembles of migmatite and / or migmatite - anatectic granite may be seen as non assimilated inclusions with dimension of 6 km, in gigantic orthogneiss intrusions (Figure 6).

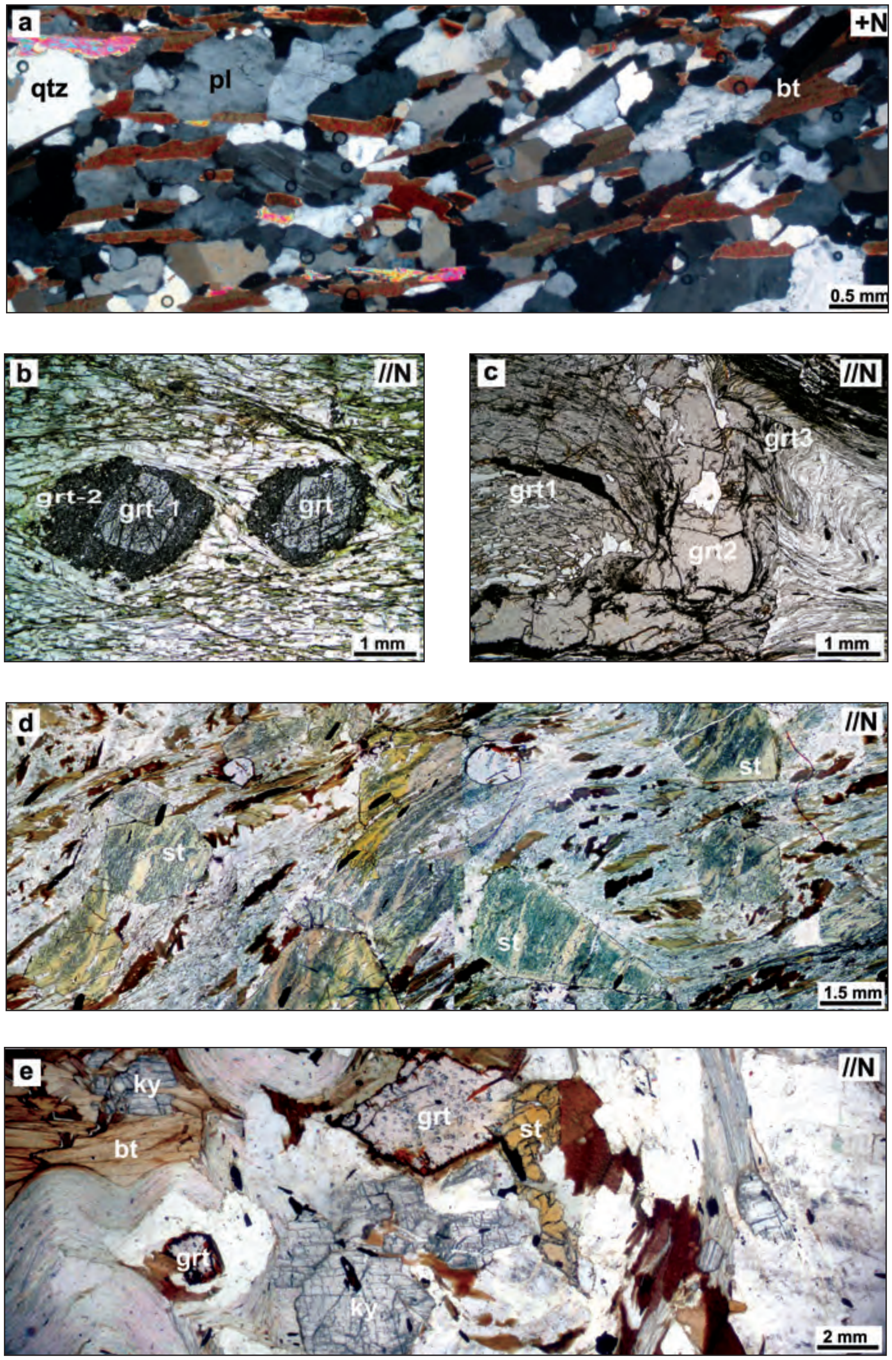

Figure 14- A: biotite - albiteschists made up of homogenously distributed single biotite crystals within a quartz and feldspar rich groundmass, B-C: Zoning garnet crystals defining polyphase growth and recognized by the change in quartz inclusion proportions, D: Stauroliticporphyroblasts showing syntectonic growth in schists, E: Assemblage of 'staurolite + disthene + garnet' in mica schists (grt: garnet, st: staurolite, ky: disthene, bt: biotite, qtz: quartz, pl: plagioclase).

Migmatites and associated granites are pervasively observed in Ödemiþ - Kiraz submassif as well. These cropping out rocks around Ödemiþ - Kiraz towns are only observed in tectonic layer consisting of metaclastic sequence of the Pan-African basement within the napped structures of this submassif. The migmatization generally occurs at lower parts of the paragneiss unit. It exposes as in broken focals with a lateral continuity not exceeding 7 - 8 km's, since the deposit is cut by thrust faults. Very different structural types were developed in migmatites similar in Kula region. Migmatites are accompanied by big masses of anatectic granites and more than 10 masses of granite were determined in this region. The biggest one of these granitic masses is located between Birgi - Kiraz within a size of 6 x 3 km's. This fine grained, rock in granoblastic texture, largely shows a massif structure (Figure 15b). Despite that, this rock shows a high graded mylonitization along rarely ductiled shear zones which cut the mass within thicknesses ranging from one to tens of meters. These granites microscopically consist of garnet crystals with a size of 4 cm in addition to sillimanite content (Figure 15c). However, some granitic masses at the southern part of Kiraz were completely transformed into strongly lineated and foliated ultra mylonites (Figure 15d).

Great majority of paragneiss exposures show high graded migmatization in Çine sub massif. These migmatites are best observed at south / north of Çine and at south of Karpuzlu. Migmatites which are accompanied by masses of anatectic granite with a size of 1 km are observed in the form of inclusions generally floating in intrusions of gigantic orthogneiss. The size of these inclusions reach up 5 km's (Candan, 1996b). Both migmatite and granites associated with them have structural, textural and compositional properties in a way that reflects a common origin similar to Ödemiþ, Kiraz and Kula regions.

Metamagmatic rocks

Partly migmatized metaclastic sequence belonging to Pan-African basement of the Menderes Massif is intruded by widespread acidic / basic magmatics associated with the Pan-African Orogeny. The distinguishing features of these magmatics are summarized below.

Acidic metamagmatics

In a broad sense, orthogneisses which are the well known rocks of the Menderes Massif have been described as 'augen' or as 'granitic' gneiss in many studies. Besides, the same rocks have been named in different ways in order to define the degree of metamorphism (sillimanite gneiss), properties of deformation (mylonitic gneiss) and source rocks (orthogneiss). Evidences obtained in recent years have shown that these intrusions can be gathered in to 3 groups based on mineralogical compositions (Bozkurt, 2004; Dora et al., 2005). The type and the amount of the mafic mineral form the main parameter in this classification. Within this scope, orthogneisses can be classified as; 1- biotite orthogneisses, 2- amphibole orthogneisses and 3- tourmaline leucocratic orthogneisses. These rocks are the differentiation products of the same Pan-African aged magmatic activity following each other. The structural, textural and mineralogical properties of these are given below.

Biotite orthogneisses.- This type is the widely observed orthogneiss type in the Menderes Massif. These coarse crystalline rocks can be divided into 2 sub types according to primary textural properties of the granite which these were derived from. The most prevalent type is the porphyritic granite in porphyry texture which was originated from 6-7 cm euhedral orthoclase crystals within a medium to fine grained groundmass. This groundmass is made up of quartz and plagioclase. These rocks traverse into blastomylonitic, augen / banded orthogneisses showing strong foliation and lineation developed

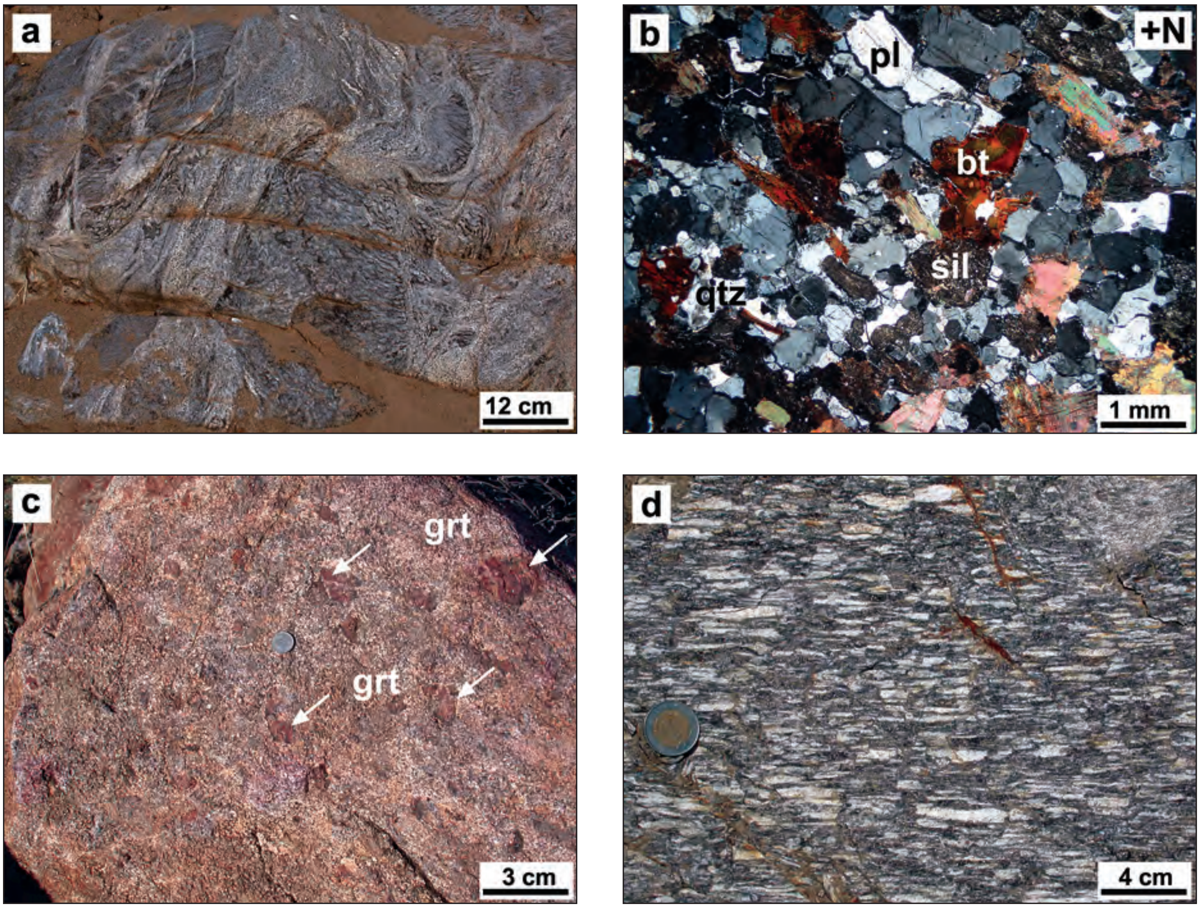

Figure 15- A:migmatites derived from paragneisses observed in Kula / Selce area, B: granoblastic texture observed in anatectic granites associated with migmatites, C: garnet porphyroblasts observed in anatectic granites (Çine), D: biotite and feldspar lineations in anatectic granites which were subjected to ductile deformation in Kiraz (pl: plagioclase, sil: sillimanite, grt: garnet, qtz: quartz).

along the shear zones under ductile deformation. Shear bands in this deformation prefer especially equally sized quartz rich groundmass whereas, coarse orthoclase porphyroblasts transform into porphyroclastics surrounded by recrystallized orthoclase zones (Figure 16a). The second type of these rocks is medium to coarse crystalline, granoblastic in texture, equally sized granites. These are described as granitic gneiss in undeformed sections. However, these rocks turn into banded gneisses in ductile deformation zones. The ratio of biotite in both types varies in between 15 - 25 %. Blastomylonitic type of these rocks is easily recognized with its strong biotitic lineation. These orthogneisses are observed in the form of plutons that have intruded each other with diameters of 8 - 10 km in the Massif. Country rock fragments widely observed in orthogneisses clearly reveal intrusive character of intrusive contacts of these rocks (Figure 16b).

Amphibole orthogneisses.- These are granoblastic in texture and fine to medium crystalline massif rocks described by the presence of hornblende and garnet porphyroblasts. These rocks are observed only in Karýncalý Mountain and around Buldan area located at north, in the Menderes Massif. These masses are stock in character within diameters of 500 - 600 m and have been intruded into augen gneisses and schists of the Pan-African Basement. Dark green colored and non oriented amphibole crystals have a homogenous distribution in massif orthogneiss. These crystals have a ratio of 20% and a dimension of 1 - 2 cm (Figure 16c). Rocks have been transformed into ultra mylonitic gneisses that could be named as gray to silver colored hornblende schist along internal shear zones and the walls of masses.

Tourmaline leucocratic orthogneisses.- In studies performed around Çine submassif, it has been established that gray to white colored leucocratic orthogneisses consisting of tourmaline (±biotite) as mafic mineral spreads in very large areas (Bozkurt et al., 2006; Dora et al., 2005). These rocks are completely white in color and the ratio of tourmaline may reach up 20%. Tourmalines are generally in the form of individual crystals having a size of 2 - 3 mm and as dispersed in the texture. In addition to this, tourmalines may be observed in the form of nodules as 6-8 cm in length and tourmaline minerals as 2 - 3 cm in length which is the characteristics of these orthogneisses. Leucocratic orthogneisses show variable contact properties ranging from transitional to sharp intrusive relationship with biotite rich augen / granitic gneisses. These leucocratic rocks can be divided into 3 groups among them based on the shape of their masses, textural properties and mineralogical compositions. These are; 1) tourmaline leucocratic orthogneisses observed in the form of plutons, 8 - 10 km in diameter, 2) leucocratic porphyritic metagranites derived from vein or stock like small intrusions and 3) leucocratic metaaplites defining the latest stage which shows distinct vein character.

Tourmaline leucocratic orthogneisses are distinctive with their marked white colors and relatively more massif structures. These are characterized by high tourmaline (10 - 20 %) and muscovite contents especially despite the non presence of biotite (max. 5%). Leucocratic orthogneisses may show some textural differences due to magmatic facial changes in them. These are; 1) leucocratic orthogneisses with a mealy view, defined by gray colored heaps made up of muscovite (±tourmaline) heaps, 2) tourmaline leucocratic orthogneisses, granoblastic in texture made up of equally sized feldspar crystals (Figure 16d) and 3) porphyritic leucocratic orthogneisses consisting of coarse orthoclase phenocrystals within a gray colored, relatively fine grained quartz feldspatic groundmass. Leucocratic orthogneisses are observed in very large areas especially between Bafa - Bozdoðan, in Çine submassif in the form of gigantic plutons interferencing with each other. Leucocratic orthogneisses show distinct intrusive contact relationship with schists (Figure 16f) both derived from biotite rich, augen orthogneisses (Figure 16e) and subarkosic sandstones. These belong to the schist unit of the Pan-African basement forming the country rock at all points along 150 km gneiss - schist contact. On the hand the related granitic intrusions are emplaced in schist and biotite rich augen gneisses in the form of stocks reaching up 3 - 4 km in size. These rocks are located at 100 - 150 km to the north of this region, south of Koçarlý, the northern part of Karýncalý Mountain, south of Kula and north of Sarýgöl. This data show that intrusions of leucocratic orthogneiss can not be restricted with any region of the Massif.

Leucocratic porphyritic metagranites are exposed in the form of amorphous masses, 50 - 150 meters in dimension or in lensoidal structures within leucocratic orthogneiss intrusions. It is considered that fine grained, white colored rocks which are transitionally in contact with tourmaline leucocratic orthogneisses were stock / vein type products of the leucocratic magmatic activity in many places. The pervasive presence of tourmalines in many samples supports that these have a primordial relationship with tourmaline leucocratic orthogneisses.

Leucocratic metaaplites form the latest stage of the Pan-African age acidic magmatic activity. These rocks are substantially composed of 'albite + quartz + (± tourmaline ± rutile) and are generally observed in the form of veins not exceeding several meters in thickness. These rocks are in white color and have also distinct or transitional contacts with tourmaline leucocratic orthogneisses. Despite that, thicknesses of these vein rock types which are managed as albite deposits in Çine sub massif may reach 100 meters in thickness and 4 - 5 km in length. These vein rocks can be observed in a very large region from Selimiye-Bafa surround at the south, to Karpuzlu at the north and to the Karýncalý Mountain at the east. Leucocratic metaaplites are described by; 1) consisting of tourmaline nodules in zonal structure (Figure 16 g,h), 2) their fine granularity, 3) being completely white in color, 4) consisting of pink colored titanium rich zones made up of rutile and sphene around the circumference of many veins throughout the Massif.

Basic metamagmatics

The presence of basic metamagmatics in the Menderes Massif (metagabbro - metanorite) has been known since Schuiling (1962). In the following years, around Çine, Ödemiþ-Tire and at south of Kula, the widespread presence of similar composite basic magmatics have been determined (Kun, 1983; Kun and Candan, 1991; Candan, 1992, 1994, 1996*a,b*). The basic geological properties of these rocks closely related with eclogites (Candan et al., 1994, 2001; Oberhänsli et al., 1997) are given below within the framework of sub massifs.

Gabbroic rocks in Çine sub massif are densified at southwest of Çine and around Karýncalý Mountain. Hundreds of metagabbroic exposures have been determined at southwest of Çine (Kun, 1983; Candan, 1996a). Great majority of these rocks are approximately in dimensions of 2 x 30 m and are in vein or lensoidal masses. The sizes of stock type masses rarely reach 300 meters. The country rock is made up of migmatized paragneiss and augen / granitic gneisses. Metagabbros show quite different textural features ranging from fine grained ophitic texture (crystal sie 2-3 cm) to holocrystalline granoblastic texture (Figure 17a). It is extremely difficult to observe metamorphic characters of these massif rocks on the field. Only in their peripheral zones and rarely in their internal shear zones, transformations into amphibolites not exceeding 20 - 30 meters in thicknesses can be observed (Figure 17b). Garnet occurrences and replacement of pyroxenes by amphibole widely develop along these zones (Figure 17c). These rocks are made up of biotite gabbro and olivine gabbro and their general mineral composition is 'plagioclase + clinopyroxene + biotite + ilmenite (± olivine) (Figure 17d).

Basic magmatics around Karýncalý Mountain consist of more widespread metamorphic effects than in Çine region. Hundreds of vein rocks have been determined in the region not exceeding 50 m in size. The size of masses may very rarely reach 1 km. Basic rocks in the region are predominantly formed by garnet amphibolites. Dark green colored garnet porphyroblasts may reach 0.5 cm in size and are composed of 'hornblende + plagioclase + epidote / zoisite + sphene ± garnet'. Small sized vein rocks have totally been transformed into amphibolite. These formations in big masses whereas, densify around peripheral zones. Basic magmatics in the region in undeformed zones have completely preserved coarse crystalline holocrystalline textures and have statically been recrystallized. These rocks are named as amphibolitic metagabbro and while primary clinopyroxenes were consumed by hornblende the plagioclases were consumed by clinozoisite too.

The most dense gabbro formations are observed at the north of Birgi (Candan, 1996*b*) south of Kiraz (Candan et al, 2001) and at the southwest of Tire in the Ödemiþ - Kiraz submassif. The presence of eclogite zones in their

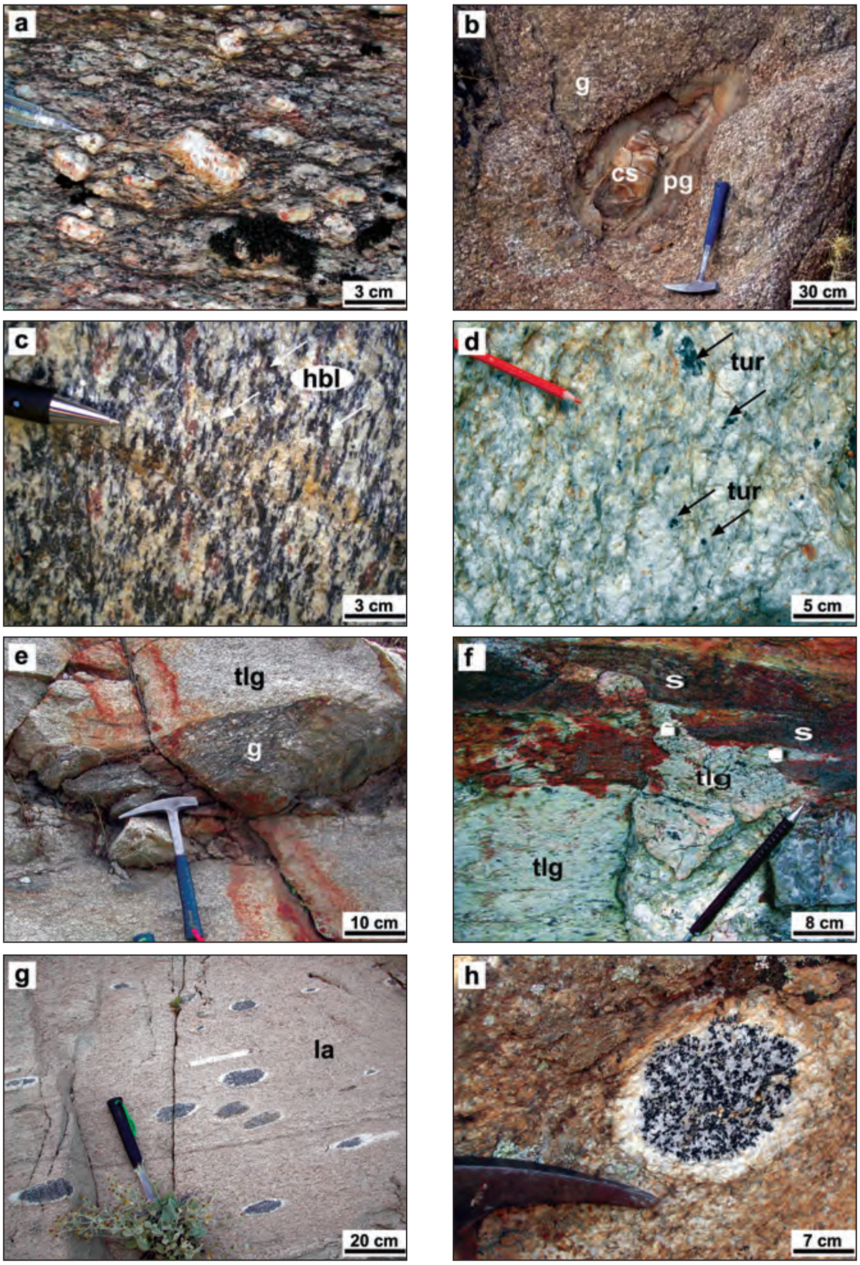

Figure 16- A: Blastomyloniticbiotiteorthogneiss forming by the ductile deformation of source granitic rock with porphyroblastic texture (Çine - Selce), B: paragneiss and calcsilicate inclusions within biotite-orthogneisses (Alaþehir / Karacalar), C: Amphibole orthogneisses enriched by hornblende crystals, D: Equidimensional, granoblastic textured, tourmaline rich leucocratic orthogneisses (Karýncalý Mountain), E: Distinct contact relationship between biotite rich orthogneiss and tourmaline leucocratic orthogneisses (Çine - Selce), F: Preserved primary intrusive contact relationship between tourmaline leucocratic orthogneiss and schist, G: Leucocratic metaaplites made by zoning tourmalinite nodules within fine grained quartz feldspatic groundmass (north of Selimiye), H: Close up view of tourmalinite nodules (Çine - Selce) (g: biotiteorthogneiss, cs: calc silicate, pg: paragneiss, tur: tourmaline, la: leucocratic aplite, hbl: hornblende, s:schist, tlg: tourmaline leucocratic orthogneiss).

circumferences is the basic characteristics of metagabbros. To the north of Birgi, around Cevizalan village 3 stocks with a size of 1.5 km and many vein rocks are observed. In one of them, relics of eclogite have been detected around circumferential zones of amphibolite and along internal shear zones (Candan, 1996*b*). Inner parts of stocks have been very well preserved thus, it is macroscopically impossible to detect the effects of metamorphism. Rocks in the region are made up of olivine gabbro and have a mineralogical composition of 'olivine + plagioclase + clinopyroxene + ilmenite ± biotite'. In sections where gabbros were subjected to ductile deformation the mylonitic texture has been developed widely (Figure 17e). The metamorphism effects in these sections can be described by formations of garnet and multi coronal structures developed around olivine crystals. In southeast of Tire numerous metagabbro stocks have been determined in a clippe, 5 x 5 km in size reflecting metamorphism under lower crust conditions (Çetinkaplan, 1995). Sizes of these stocks are 300 x 500 m and around these stocks well preserved eclogitic - metagabbroic zones are observed (Candan et al., 2001). Metabasic rocks in the region have a composition varying in between gabbro - norite. The mineralogical compositions of norites and gabbros are 'plagioclase + orthopyroxene + biotite + ilmenite' and 'plagioclase + clinopyroxene + olivine + biotite + ilmenite' respectively.

The biggest metagabbroic mass is observed at north of Alaþehir, in Yahyaalcý village in Demirci Gördes submassif (Candan, 1994). This stock is in a size of 300 x 200 m and is accompanied by many vein rocks. Towards the center of stock continuous textural and mineralogical variations are observed related to polyphase metamorphism. Badly preserved eclogite relics are present within the amphibolitic circumferential zone of the stock. At the center, gabbro has been statically recrsytallized and primary holocrystalline texture has totally been preserved in low tensile regions. Transformations of pyroxenes into amphibole, the consumption of plagioclase by clinozoisite and partial garnet coronas around magmatic phases are the fundamental metamorphic effects.

THE CONTACT RELATIONSHIP BETWEEN THE PAN-AFRICAN BASEMENT / COVER SERIES

Many investigators have emphasized that there should be an unconformity between the Pan-African basement and Palaeozoic - Early Tertiary cover series in the Menderes Massif (Pan-African unconformity) (Þengör et al., 1984; Dora et al., 1995). Objective evidences towards this problem have been obtained around Mesken village at north of Yataðan (Konak et al., 1987; Dora et al., 2005; Candan et al., 2006). In the map prepared around Mesken village, it was determined that the unconformity plane is defined by muscovite-quartz schists derived from quartz arenite and the presence of meta conglomerates in character of channel fills among them has been established. These conglomeratic channel fills are in the form of crop outs broken from each other and can be traced 35 km towards Bozdoðan at the east. Conglomerate bearing quartz schist unit has a minimum thickness at north of Mesken village but reaches maximum thickness of 1.5 km towards the south. Quartzite forming the lowermost part of Palaeozoic cover series resides on different units such as orthogneiss and schist which belong to the basement in such a way to describe a deep abrasion. Quartzites are overlain by probable Carboniferous black phyllites with a transitional contact. The conglomerate layers are best observed in Gökçen River and Kale Tepe at the north of Mesken. This meta-conglomerate layer is laterally traced 7 km approximately and is important in order to determine especially the source rock of cover series. The compounds of conglomerates are made up of fully leucocratic in character, tourmaline rich granite, aplite and quartzite pebbles. The primary rock of this conglomerate has an intermediate sandy material. In these conglomerates black tourmalinite pebbles are also encountered.

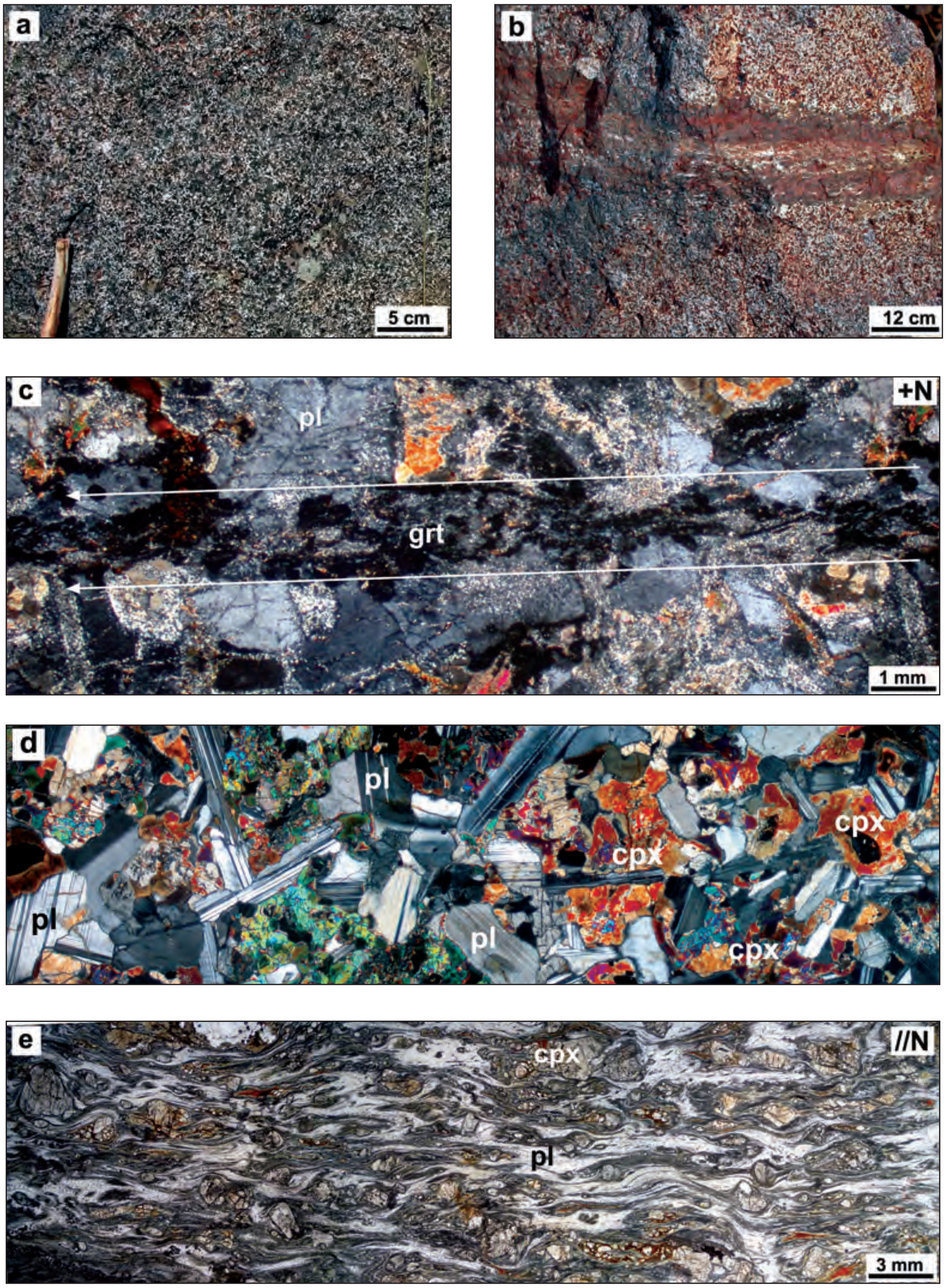

Figure 17- A: Primary ophitic texture of gabbros in sections preserved from metamorphism (BirgiCevizalan), B: amphibolite transformations forming along shear zones cutting gabbros (Birgi), C: garnet crystals forming along ductile shear zones cutting gabbros, D: microscopic view of primary ophitic texture in gabbros which were preserved from metamorphism, E: Mylonitic texture in gabbros subjected to ductile deformation (pl: plagioclase, cpx: clinopyroxene, grt: garnet).

The unconformity between the Pan-African basement and Palaeozoic cover series is clearly observed in N-S directed cross section passing through Kale Tepe (Figure 18a,b). Here, tourmaline leucocratic orthogneiss and porphyritic metagranites forming the basement show a clear intrusive contact relationship with muscovite schists which belong to metaclastic sequence forming the country rock (porphyritic metagranite sample taken here gave an intrusion age of 551.5 ±2.9 Ma (Dora et al., 2005). Palaeozoic deposit begins with 60 m thick quartzites in homogenous structure derived from coarse quartz sandstones. Quartzites are overlain by 30 m thick conglomerates made up of huge blocks at lower parts. Above meta-conglomerate layer, 18 m thick muscovite quartz schist layer takes place. Above quartzites Carboniferous black colored garnet - chloritoid phyllites / calcschist intercalation is recognized with a distinct contact. The matrix of conglomerate layer is made up of green, coarse muscovite crystals which can easily be recognized on field. Porphyritic metagranite, leucocratic orthogneiss, quartzite, aplite and tourmalinites are the main rock types forming metaconglomerate pebbles. Porphyritic meta-conglomerate pebble taken here and from the bedrock that has similar textural / structural, mineralogical and geochemical features was dated as 550.4 ± 2.6 Ma (Dora et al., 2006; Candan et al., 2006). The other composites forming conglomerates as well present very similar properties with that of magmatics at the basement in such a way that it supports the unconformable character of the contact.

DISCUSSION

During the last 20 years many ideas have been put forward towards the stratigraphy of the Pan-African basement and source of protoliths of main lithologies in the Menderes Massif.

The source, primary depositional age and provenance of paragneisses

These rocks were distinguished as a distinct unit by Schuiling (1962) in Çine sub massif of the Menderes Massif and named as 'fine grained gneiss in basic character'. In the following years, similar rocks in the southern part of the same region were named as 'hornfels like rocks' by Baþarýr (1975). The first study directly towards the source of paragneisses was made by Kun (1983). The investigator named these rocks in Çine region as 'leptite'. Kun (1983) interpreted these rocks as 'island arc volcanites, composed of rhyolite - andesite and calcalkaline in character'. In the following years, the presence of rocks that have similar mineralogical, petrographical and geochemical features was detected around Ödemiþ - Kiraz (Kun and Candan; 1987a; Kun et al., 1988; Candan and Kun, 1991; Candan, 1996*a,b*), in Nazilli Karýncalý Mountain regions (Kun and Candan, 1991) in Çine, Yenipazar (Çolak, 1985; Þengül et al., 2006), at southern slopes of Aydýn Mountains (Candan et al., 1992), at north of Alaþehir (Candan, 1994), at south of Tire (Candan, 1995) and at north of Çine Kun (1983) and these were interpreted as high graded metamorphic derivatives of continental volcanites in accordance with the data obtained by Kun (1983). The formation of white speckles within the fine grained groundmass in paragneisses was wrongly interpreted as relic hypocrystalline porphyritic texture belonging to old volcanites and that has been effective in this opinion. Dora et al. (1988, 1990) interpreted related rocks as continental volcanites developed on the Precambrian basement in the following metamorphism stage of core series within the metamorphic stage of the Menderes Massif.

Loos (1995) determined the presence of abrasions of crystal face and distinct roundness' to show the detritic origin in zircon which he picked in paragneisses. Besides, these zircons were dated as ranging between 585 - 1871 Ma by means of the single zircon evaporation method supporting the detritic origin. Using these evidences, Dora et al. (1995) stated that these leptites could be sedimentary in origin. In the following years, many studies towards the determination of source rocks of paragneisses have been made and to the contrary of volcanic origin,

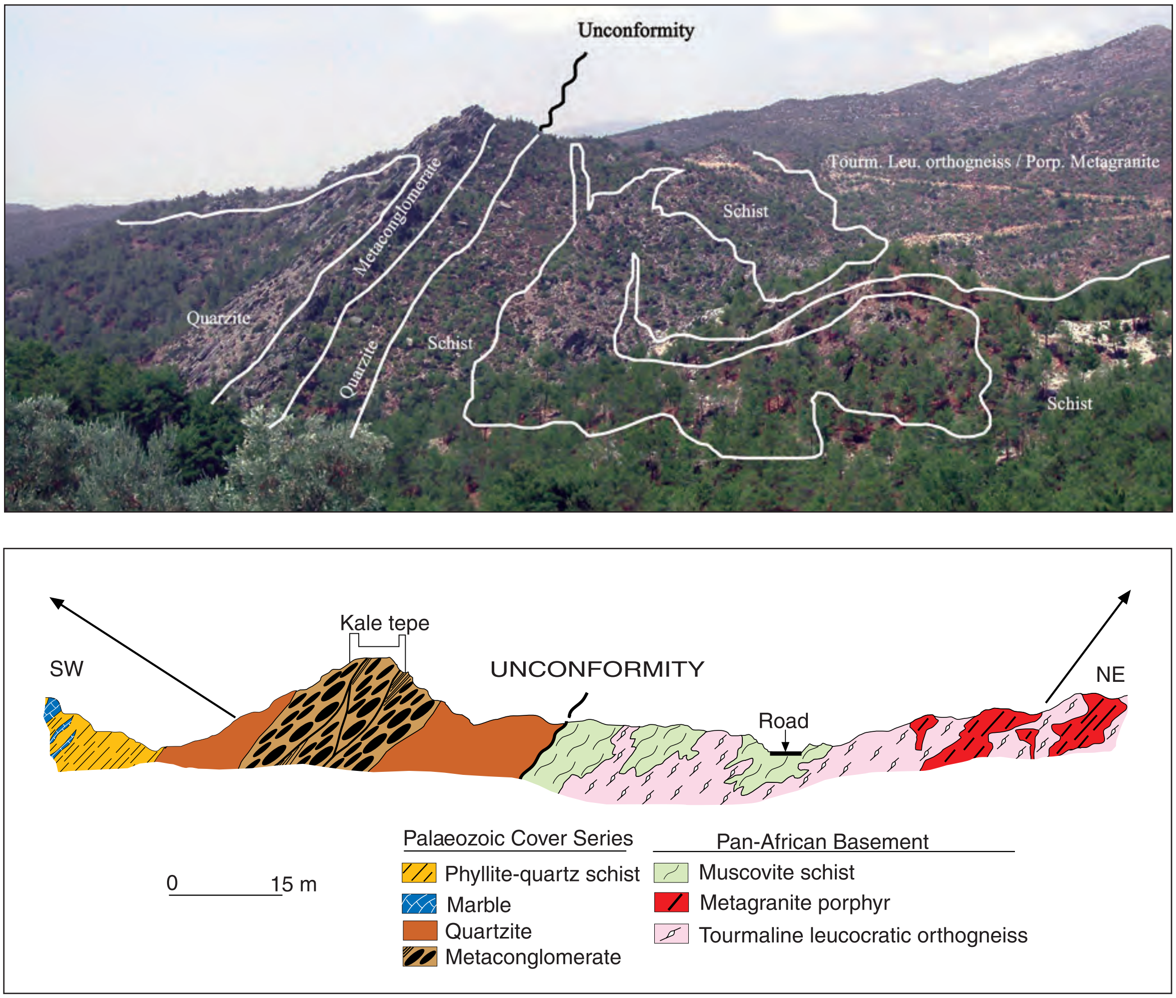

Figure 18- A: Field view and B: Geological section of unconformable contact relationship between the Pan-African basement and Palaeozoic cover series in the Menderes Massif.

new evidences indicating that these were derived from clastic sedimentary rocks have been obtained (Dora et al., 1998, 2000, 2001, 2002; Koralay et al., 2002, 2003). Investigators stated that the data such as; i) the presence of no textural data to be preserved belonging to the primary volcanics, ii) the absence of clastic facies' such as lava, agglomerate and tuff in different compositions that might be found in a volcanic sequence at this thickness and iii) the non visualization of structures such as dyke and dome that could be observed in a volcanic complex in this size decreases the probability of volcanic source in these rocks. Despite that, evidences such as; i) the constitution of the mica - schist unit of the Pan-African basement a conformable and continuous sequence with the paragneiss unit, ii) the case of schists derived from mudstone and subarkosic sandstone layers in the paragneiss unit are gradually transitional with paragneisses both in vertical and lateral directions, iii) the widespread occurrence of intermediate rocks which can not be defined as mica schist and paragneiss that originated from continuous changes in ratio of clay / detritic grain in primary sedimentary rock within paragneiss unit,

iv) the presence of mica or quartz (+feldspar) rich layers originating from the rhythmic compositional change of the primary clactic sediment in paragneisses which are the equivalent of clay and sand rich layers at millimeter - centimeter thickness and v) zircons in the paragneiss containing abrasioned crystal faces or their presence to be as fully rounded grains, and their scattered age data support feeding from a cratonic field which clearly reveal the sedimentary source of paragneisses according to the investigators (Dora et al., 2001, 2002; Koralay et al., 2002).

In order to determine the depositional age of primary clastic sediments which paragneisses derived from, zircon clastics were picked and dated from paragneisses at each of three sub massifs. On single heating step, zircon ages ranging between 613 - 2558 Ma were obtained from these samples. Only one zircon grain was detected as 587 Ma which is not robust (Dora et al., 2002; Koralay et al., 2002, 2003, 2005). The crystallization age of primary granites in orthogneisses that have intrusive contact relationship with paragneisses were dated as 549.3 ± 13.4 Ma in Karacalar Village, north of Alaþehir. On the other hand, ages of zircon clastic of schists belonging to Pan-African basement in Çine submassif were dated as 592-3239 Ma (Dora et al., 2005). The metaclastic sequence to be a continuous deposit made up of paragneiss and schist units foresees a time range of 590-550 Ma as for the depositional age of primary sediments of these metaclastics forming the oldest units of the Menderes Massif (Dora et al., 2002; Koralay et al., 2005; Candan et al., 2007). When it is considered that the intrusional age of primary granites of orthogneisses vary between 570 - 520 Ma (Loos and Reischmann, 1999) throughout the Massif and schists forming country rocks with the age of 570 Ma belong to metaclastic sequence, the depositional age lies between 590-570 Ma (Late Neoproterozoic). On the other hand, it is known that core series of the Menderes Massif were affected by the polyphase metamorphism associated with the Pan-African Orogeny (Candan et al., 2001). The granulite facies metamorphism affecting paragneisses and forming the first stage as well was dated as 583 ± 5.7 Ma by Koralay et al. (2006). The time interval for the depositional age of primary sediments is foreseen as 590-580 Ma when all these geochronological data are evaluated for the metaclastic sequence forming the oldest rocks of the Menderes Massif. According to Dora et al. (2002) paragneiss unit derived dominantly from litharenitic sandstone, bears the character of grain flow of shoreline drift with high energy. According to investigators, provenance conditions are suggested as 'planar, humid - arid climate, gneisoid - granitoid rock in composition and gradually uplifted as an example structural behavior everywhere'.

The source and formation age of protoliths of gneisses

In studies made for more than 50 years, different opinions have been put forward about protoliths of gneisses which are the most characteristic of the Massif. These opinions can be gathered in three groups; 1- gneisses were formed under high graded metamorphism of sedimentary rocks (Schuiling, 1962; Þengör et al., 1984; Akkök et al., 1984; Dora et al., 1990; Satýr and Friedrichsen, 1986), 2- gneisses in both the sedimentary and magmatic in origin are observed in the Massif (Baþarýr, 1970, 1975; Scotford, 1969; Graciansky, 1965; Konak, 1985; Konak et al., 1987) and 3- all gneisses were formed as a result of metamorphisms of magmatic rocks in granitic composition (Erdoðan, 1992; Bozkurt, 2004; Bozkurt and Park, 1994; Bozkurt et al., 1995, 2006; Loos, 1995; Dora et al., 1995, 2005; Hetzel and Reischmann, 1996; Dannat, 1997; Dannat and Reischmann, 1998; Loos and Reischmann, 1999; Gessner et al., 2001, 2004; Koralay et al., 2004, 2007; Erdoðan and Güngör, 1992, 2004). The solution of this problem lies under the assessment of field data and geochronological / geochemical evidences together. Recent studies have revealed that these rocks were derived from a granitic source rock due to basic properties of gneisses. These properties are as follows; i) gneisses have homogenous internal structure, ii) the preserved holocrystalline structure peculiar to protolith can be observed, iii) gneisses show clear intrusive contact relationship with schists forming the country rock, iv) gneisses diffusively have zircon peculiar to magmatic rocks describing the crystallization from a melt, v) these zircons give extremely close results describing magmatic crystallization and vi) source rock diagrams in geochemical analyses clearly define a magmatic rock (Erdoðan and Güngör, 1992; Bozkurt and Park, 1994; Bozkurt et al., 2006; Dora et al., 2005; Hetzel and Reischmann, 1996; Dannat, 1997; Loos and Reischmann, 1999; Gessner et al., 2004; Koralay et al., 2004).

Ideas about the ages of orthogneisses in the Menderes Massif can be gathered in to two groups except for different ideas on source rock. These are; i) Precambrian - Cambrian and ii) Upper Cretaceous / Tertiary. In this article, this problem will be explicated as a brief summary since detailed data were given by Koralay et al. (in this ýssue) regarding the geochronology of orthogneisses. So far, many different groups have estimated the intrusional age of primary granites for orthogneisses in laboratories based on total rock, single zircon evaporation, classical zircon and SHRIMP methods and obtained age of Precambrian / Cambrian (Schuiling, 1973, 548 Ma; Dora, 1975, 490 ± 90 Ma; Satýr and Friedrichsen, 1986, 502 - 471 Ma; Hetzel and Reischmann, 1996, 546 Ma;, Dannat and Reischmann, 1997, 540 Ma; Koralay et al., 2004, 560 - 570 Ma; Loos and Reischmann, 1999, 520 - 570 Ma; Gessner et al., 2004, 541-566 Ma; Dora et al., 2005, 545 - 552 Ma). The idea of orthogneisses to be young however, is based on field data and deformational properties (Erdoðan, 1992; Erdoðan and Güngör, 1992, 2004; Bozkurt and Park, 1994, 1997*a,b*, 1999; Bozkurt and Park, 2001; Bozkurt, 2000; Bozkurt et al., 2006). Investigators who supported this idea interpret that all the dated zircons in orthogneisses are relic zircons.

The age of gabbros

The first data related to the presence of basic magmatic rocks in the Menderes Massif are seen in Schuiling (1962). These rocks located in the Çine submassif were interpreted as Miocene aged post metamorphic intrusions following vertical tectonic lines by Kun (1983). In following years, presence of gabbros has been detected in other parts of the Massif as well and these rocks have similarly been assessed as young plutons (Dora et al., 1988, 1992; Kun and Candan, 1991). However, as an opposition to those opinions new field data and petrological studies were clearly revealed that gabbros are not young and were not affected by polyphase metamorphism (Candan, 1994, 1996*a,b*; Candan et al., 1994, 2001; Oberhänsli et al., 1997, 2010; Dora et al., 2001; Çetinkaplan, 1995).

The relative Precambrian age was suggested for gabbros in articles which the metamorphic character of these rocks was first defined (Candan, 1994, 1996 *a,b*). Gabbros are only located in the Pan-African basement and do not cut cover series by intrusion. These data were used as basic geological features. K/Ar ages estimated from micas in gabbros were dispersed much and gave geologically meaningless results (Candan, 1996). Oberhänsli et al. (2010) performed a study on a sample which was well preserved from metamorphic effects and dated zircons in magmatic origin as 540 ± 3.5 Ma. This age is compatible with basic geological data and was interpreted as the crystallization age of gabbro by investigators.

The primary contact relationship between the Pan-African basement and cover series

Many of the articles regarding the general geological structure of the Massif confirm that the contact between the Pan-African basement and the overlying cover series accepted as Palaeozoic - Early Tertiary in age is an unconformity plane which traces were greatly erased by latter metamorphisms (Schuiling, 1962; Graciansky, 1965; Baþarýr, 1970; Dora, 1975, Þengör et al., 1984; Dora et al., 1988, 1990, 1995; Konak et al., 1987). Many investigators define and name this unconformity plane as the 'Pan-African unconformity'. First objective data on this subject were obtained by Konak et al. (1987). The unit which was defined as basal conglomerate in some studies by investigators and as channels fills in cover series (Erdoðan and Güngör, 2004; Bozkurt et al., 2006) was retreated and studied by Dora et al, (2005). Investigators claim that units of the Pan-African basement is unconformably overlain by a deposit of cover series (Pan-African unconformity) (Dora et al., 2005; Candan et al., 2007, 2010) that begins with quartzite / conglomerate, continues with phyllite - quartzite - marble. These are based on evidences given below.

i) The main components of conglomerates are made up of tourmaline rich, granite, leucocratic in character and of tourmalinites, ii) tourmaline rich granites that have the same mineralogical composition, textural feature and chemical composition with these pebbles are pervasively observed in the Pan-African basement, iii) granite pebbles and their equivalents give identical ages (Pebble: 552 ± 3.1 Ma, Basement: 551.5 ± 2.9 Ma), iv) quartzites containing conglomerate layers cover different units of the Pan-African basement, v) probable Late Devonian aged quartzite / conglomerate unit is placed at the lowermost part of the Palaeozoic cover series.

The Palaeogeographical position of the Pan-African basement in Late Neoproterozoic time

The Pan-African Orogeny comprising the events of the integration of Gondwana ranges in between 950 - 450 Ma (Kröner, 1984). The distribution of continents in Neoproterozoic time indicates that East and West Gondwana lands were separated by an ocean named the Mozambique Ocean (Stern, 1994; Wilson et al., 1997; Daiziel, 1991). This ocean is as big as the Pacific Ocean and is considered as it was formed by the disintegration of Rodinia Super Continent 800 - 850 Ma ago. The closure of this ocean and collision of east and west Gondwana lands with each other caused the formation of orogenic belt extending in N-S directions along the eastern margin of the African continent. This orogenic belt which is also defined as the 'Mozambique belt' was named as the East African Orogeny by Stern (1994).

In the final integration stage of Gondwana in the Latest Neoproterozoic time, it is considered that Anatolia was located at the northeast of the African-Arabian peninsula and at the northernmost part of Mozambique belt (Stern, 1994; Wilson et al., 1997). Þengör et al. (1984) states that the northern continuity of suture belts associated with the closure of this ocean can be traced with the Pan-African events in Bitlis and Menderes Massifs. Similar paleogeographic position is foreseen by different investigators based on various geological data and correlations (Dora et al., 1995, 2002; Stampfli and Borel, 2002; Gessner et al., 2004; Gürsu et al., 2004; Monod et al., 2003; Neubauer, 2002; Koralay et al., 2005; Candan et al., 2007; Oberhänsli et al., 2010).

Within this context, it is seen that assemblages of units belonging to the Pan - African basement of the Menderes Massif could be associated with Mozambique belt from Egypt, Red Sea and Arabian Peninsula to South Africa along East Africa. Thus, it can be concluded that the metaclastic sequence made up of paragneiss and schist units forming oldest units of the Pan-African basement were deposited on the passive continental margin of Late Neoproterozoic Mozambique Ocean that is in between East and West Gondwana lands. In this issue of the periodical, the original relation between the Late Neoproterozoic evolution of the Mozambique Ocean and the polymetamorphic evolution of the Pan-African basement in the Menderes Massif is being discussed in the article given by

Candan et al. (in this issue). In this study it is emphasized that the Mozambique belt was defined by the metamorphism of granulite facies with ages of 715-650 Ma and 620-520 Ma (average 550 Ma) (Stern, 1994) and besides, rare 530-500 Ma aged eclogites were observed in Malawi as well (Ring et al., 2002). Investigators claim that age of granulite (583.0±5.7 Ma; Koralay et al., 2007) and eclogite (529.9±22 Ma; Oberhänsli et al., 2010) facies metamorphisms in the Menderes Massif show big similarity with the metamorphisms and events in the Massif could be associated with closure of the Mozambique ocean and with final collisional stage of East and west Gondwana lands. On the other hand, again in issue, Koralay et al. (in this issue) interprets that Late Neoproterozoic / Cambrian aged granitic intrusions within the Pan-African basement of the Menderes Massif ranging between 570-520 Ma (average 550 Ma) (Loos and Reischmann, 1999) are syn to post African intrusives associated with the closure of this ocean. Finally, data obtained from the Pan-African basement indicate that sedimentary, magmatic and metamorphic stages of core series of the Menderes Massif could be associated with final integration period of the Gondwana Continent resulted by closure of the Mozambique Ocean and collision of East and West Gondwana lands. Exposures belonging to the Pan-African Orogeny in Anatolia by plate tectonics in Palaeozoic and Mesozoic times were napped with younger units and acquired their present positions in the form of isolated tectonic layers.

RESULTS

The stratigraphy of the Pan African basement of the Menderes Massif obtained by the studies in recent years and the results related to properties of main lithologies are given below:

1- The Pan-African basement consists of a thick metaclastic sequence and numerous acidic / basic magmatics that have intruded into the basement.

2- The metaclastic sequence begins with a litharenite dominant sequence at the bottom (paragneiss unit) and continues with sandstone - mudstone intercalation sequence (schist unit).

3- The thickness of this metaclastic sequence reaches 8 km and the depositional age of primary sediments occurred in a time interval of 590 - 580 Ma.

4- This metaclastic sequence was migmatized at the last stage of the Pan- African orogeny and pervasive development of anatectic granite took place associated with this event.

5- Orthogneisses in the Massif derived from a granitic source rocks that have various textural / mineralogical features and have an intrusive primary contact relationship. Orthogneisses which are the differentiation products of the same magmatic activity can be divided into three groups as; i) biotite orthogneiss, ii) tourmaline leucocratic orthogneiss and iii) amphibole orthogneisses.

6- Geochronological and petrological evidences indicate that rocks of which have an intrusion age ranging between 570 - 52 Ma (averaging at 550 Ma) could be granites synchronous with the Pan-African orogeny and have intruded in the following stage.

7- Gabbros which are observed only in the Pan-African basement are Precambrian / Cambrian aged intrusions and consist of polyphase metamorphic data associated with the Pan-African orogeny.

8- The sedimentary, metamorphic and magmatic evolutions of the Pan-African basement in the Menderes Massif are associated with closure of the Mozambique Ocean in Late Neoproterozoic - Cambrian and with the collision period of East and West Gondwana which results in the final assemblage of the Gondwana super continent.

9- The primary contact relationship of the Pan-African basement - Palaeozoic - Early

Tertiary cover series which was reshaped by the effect of Alpine age compressional and the following extensional tectonism are unconformable (Upper Pan-African unconformity). The unconformity is described by (?) Upper Devonian quartzite metaconglomerate sequence.

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REFERENCE

Akdeniz, N. and Konak, N. 1979. Menderes Masifi'nin Simav dolayýndaki Kayabirimleri ve Metabazik, metaultramafik kayalarýn konumlarý. Türkiye Jeoloji Bülteni, 22, pp.175-183.

Akkök, R., Satýr, M. and Þengör, AMC. 1984. Timing of tectonic events in the Menderes Massif and its implications. 38th Scientific and Technical Congress of the Geological Society of Turkey, pp.9-11.

Ayan, M. 1971. Gördes migmatitleri ve güneydoðu yöresindeki uranyum zuhurlarý oluþumu. Lectorship thesis, Ankara Üniversitesi Fen Fakültesi, 127p. (unpublished).

Barbol, D. 2005. Menderes Masifi'nin Pan-Afrikan temeline ait metakýrýntýlýlar içerisinde gözlenen kalksilikat türü kayalarýn mineralojisi, petrografisi ve metamorfizmasý. Dokuz Eylül Üniversitesi Fen Bilimleri Enstitüsü, MSc. Thesis, 118p. (unpublished).

Baþarýr, E., 1970. Bafa gölünün doðusunda kalan Menderes Masifi güney kanadýnýn jeolojisi ve petrografisi. Ege Üniversitesi Fen Fakültesi Ýlmi Raporlar Serisi, No: 102, pp.1-44.

Baþarýr, E. 1975. Çine güneyindeki metamorfiklerin petrografisi ve bireysel indeks minerallerin doku içerisindeki geliþimleri. (Lectorship thesis), Ege Üniversitesi YBÝ, Ýzmir (unpublished).

_____ 2000. Timing of extension on the Büyük Menderes Graben, western Turkey, and its tectonic implications. In: Bozkurt, E., Winchester, J.A. and Piper, J.D.A. (eds) Tectonic and magmatism in Turkey and its surrounding areas. Journal Geological Society London Special Publication 173, pp.385-403.

_____ 2004. Granitoid rocks of the southern Menderes Massif: field evidence for Tertiary magmatism in an extensional shear zone. International Journal of Earth Sciences, 93, pp.52-71.

_____, Park, R.G. and Winchester, J.A., 1993. Evidence against the core/cover interpretation of the southern sector of the Menderes Massif, west Turkey. Terra Nova 5, pp.445-451.

_____ and ____ 1994. Southern Menderes Massif: an incipient metamorphic core complex in Western Anatolia, Turkey. Journal Geological Society London, 151, pp.213-216.

_____ Winchester, J.A. and Park, R.G.v1995. Geochemistry and tectonic significance of augen gneisses from the southern Menderes Massif (West Turkey). Geological Magazine, 132, pp.287-301.

_____ and Park, R.G. 1997*a*. Evolution of a mid-Tertiary extensional shear zone in the southern Menderes Massif, western Turkey. Bulletin Society Geological France, 168,1, pp.2-14.

_____ and _____ 1997*b*. Microstructures of deformed grains in the augen gneisses of southern Menderes Massif (western Turkey) and their tectonic significance. Geologische Rundschau, 86, pp.103-119.

_____ and ____ 1999. The structure of the Palaeozoic schists in the Southern Menderes Massif, western Turkey: a new approach to the origin of the main Menderes Metamorphism and its relation to the Lycian Nappes. Geodinamica Acta, 12, pp.25-42.

Bozkurt, E. and Park, G. 2001. Discussion on the evolution of the Southern Menderes Massif in SW Turkey as revealed by zircon dating. Journal Geological Society London, 158, pp.393-395.

_____ Winchester, A. J. Mittwede, S. and Ottley, C. 2006. Geochemistry and tectonic implications of leucogranites and tourmalines of the southern Menderes Massif, southwest Turkey. Geodinamica Acta, 19/5, pp.363-390.

Candan, O. 1991. Demirci-Gördes Asmasifi'nde (Menderes Masifi) gözlenen disten-andalusit pegmatoidlerin oluþum þekli ve oluþum þekli ve oluþum evreleri. Selçuk Üniversitesi Mühendislik - Mimarlýk Fakültesi Dergisi, 1, pp.12-29.

_____ 1992. Menderes Masifi / Demirci - Gördes Asmasifi'nde Kula - Yeþilyurt kasabalarý arasýnda kalan bölgenin jeolojisi, petrografisi ve metamorfik evrimi. Dokuz Eylül Üniversitesi Rektörlüðü, 0.908.90.05.02 no'lu proje, 129 p. (unpublished).

_____ 1994. Alaþehir kuzeyinde (Menderes Masifi, Demirci -Gördes Asmasifi) gözlenen metagabrolarýn petrografisi ve metamorfizmasý. Türkiye Jeoloji Bülteni, 37, pp.29-40.

_____ 1995. Menderes Masifi'ndeki kalýntý granulit fasiyesi metamorfizmasý. Turkish Journal of Earth Sciences, 4, pp.35-55.

_____ 1996*a*. Aydýn - Çine Asmasifi'ndeki (Menderes Masifi) gabrolarýn metamorfizmasý ve diðer asmasiflerle karþýlaþtýrýlmasý. Turkish Journal Earth Sciences, 5, pp.123-139.

_____ 1996*b*. Kiraz - Birgi çevresindeki (Menderes Masifi / Ödemiþ-Kiraz Asmasifi) metagabrolarýn petrografisi ve metamorfizmasý. Yerbilimleri, 18, pp.1-25.

_____ and Kun, N. 1991. Ödemiþ Asmasifindeki (Menderes Masifi) olasýlý Pan-African metavolkanitleri. Maden Tetkik ve Arama Dergisi, 112, pp.27-40.

_____ Dora, O.Ö., Kun, N., Akal, C. and Koralay, E. 1992. Aydýn Daðlarý (Menderes Masifi) güney kesimindeki allokton metamorfik birimler. Türkiye Petrol Jeologlarý Dergisi Bülteni, C4/1, pp.93-110.

Candan, O. and Dora, O. 1993. Application of schreinemakers method to a metamorphic area located at the northern flank of the Menderes Massif (Western Turkey). Bulletin of the Geological Society of Greece, Vol XXVIII/2, pp.169-186.

_____ , ____, Dürr, St. and Oberhänsli, R. 1994. Erster Nachweis von Granulit und Eklogit - Relikten im Menderes - Massif / Türkei. Göttingen Abr. Geol. Paläont. Sb.1 5. Symposium TSK, pp.217-220.

_____ and _____ 1998. Menderes Masifi'nde granulit, eklojit ve mavi þist kalýntýlarý: Pan-Afrikan ve Tersiyer metamorfik evrimine bir yaklaþým. Türkiye Jeoloji Bülteni, 41/1, pp.1-35.

_____ and Çetinkaplan. M. 2001. Menderes masifi'ndeki eklojit / epidot-mavi þist fasiyesi metamorfizmasý ve Kikladik kompleksle karþýlaþtýrmasý. YDABÇAG-495 nolu The Scientific and Technological Research Council of Turkey project. 182p. (unpublished).

_____, Dora, O.Ö., Oberhänsli, R., Çetinkaplan, M., Partzsch, J.H., Warkus, F. and Dürr, S. 2001. Pan-African high-pressure metamorphism in the Precambrian basement of the Menderes Massif, Western Anatolia, Turkey. International Journal Earth Sciences (Geologische Rundschau), 89, 4, pp.793-811.

_____, Koralay, E., Dora, O., Chen, F., Oberhänsli, R., Akal, C., Satýr, M. and Kaya, O. 2006. Menderes Masifi'nde Pan-Afrikan Sonrasý Uyumsuzluk: Jeolojik ve Jeokronolojik Bir Yaklaþým. 59. Türkiye Jeoloji Kurultayý Bildiri özleri, Ankara, pp. 25-27.

_____, _____,_____,_____,_____, Çetinkaplan, M., Akal, C., Satýr, M. and Kaya, O. 2007. Menderes Masifi'nin Pan-Afrikan temelin stratigrafisi ve örtü - çekirdek serilerinin ilksel dokanak iliþkisi. Menderes Masifi Kolokyumu, Ýzmir, pp.8-14.

_____, _____, Akal, C., Kaya, O., Oberhansli, R., Dora, O.Ö., Konak, N. and Chen, F. 2010. Supra-Pan-African unconformity between core and cover series of the Menderes Massif / Turkey and its geological implications. Precambrian Research (in print).

Çetinkaplan, M. 1995. Geochemical, mineralogical and petrographical investigation of the eclogites in southern part of Tire area, Ödemiþ-Kiraz submassif of the Menderes Massif. Unpublished MSc. Thesis, Dokuz Eylül Üniversitesi, Ýzmir, 92p. (unpublished).

Çolak, M. 1985. Yenipazar - Aydýn dolayýnýn (Menderes Masifi) jeolojisi ve metamorfizmasý. Dokuz Eylül Üniversitesi Jeoloji Mühendisliði Bölümü Graduation project, 67p. (unpublished).

Dað, N. and Dora, Ö. 1991. Gördes (Menderes Masifi Kuzeyi) pegmatoidleri. Türkiye Jeoloji Bülteni, 34, pp.1-8.

Daiziel, I.W.D. 1991. Pasific margins of Laurentia and east Antartica-Australia as a conjugate rift pair: Evidence and implications for an Eocambrian supercontinent. Geology, 19, pp.598-601.

Dannat, C. 1997. Geochemie, geochronologie und Nd-Sm Isotopie der granitoiden Kerngneiss des Menderes Massivs, SW-Turkey. PhD thesis, Johannes Gutenberf Universitat Mainz, 87p (unpublished).

_____ and Reischmann, T. 1998. Geochronological, geochemical and isotopic data on granitic gneisses from the Menderes Massif, SW Turkey. Abstract 3rd International Turkish Geology Symposium, 282p.

Dora, O.Ö. 1975. Menderes Masifi'nde alkali feldspatlarýn yapýsal durumlarý ve bunlarýn petrojenetik yorumlarda kullanýlmasý. Türkiye Jeoloji Bülteni, 24, pp.91-94.

_____, Kun, N. and Candan, O. 1988. Metavolcanics (leptites) in the Menderes Massif: a possible paleoarc volcanism. Middle East Technical University Journal of Pure and Applied Sciences, 21, 1-2, pp.413-445.

_____, ____ and ____, 1990. Metamorphic history and geotectonic evolution of the Menderes Massif. Proceedings of International Earth Sciences Congress on Aegean Regions, Ýzmir/Turkey, 2, pp.102-115.

_____, ______ and ______, 1992. Menderes Masifi'nin metamorfik tarihçesi ve jeotektonik konumu. Türkiye Jeoloji Bülteni, 35/1, pp.1-14.

Dora, O.Ö., Candan, O., Dürr, S. and Oberhänsli, R. 1995. New evidence on the geotectonic evolution of the Menderes Massif. International Earth Sciences Colloquium on the Eagean Region, Izmir-Turkey, 1, pp.53-72.

_____, _____, Kaya, O., Koralay, E. and Dürr, S. 1998. Revision of the leptites in the Menderes Massif: a supracrustal metasedimentary origin. Third International Turkish Geology Symposium., Middle East Technical University (METU) - Ankara, 283p.

_____, ______, _______ and ______, 2000. New data relating to the origin and polymetamorphism of the paragneisses in the Menderes Massif. International Earth Sciences Colloquium on the Eagean Region (IESCA), Ýzmir, pp.133.

_____, ____, _____, ____ and Dürr, S., 2001. Revision of the so-called "leptite-gneisses" in the Menderes Massif: A supracrustal metasedimentary origin. International Journal of Earth Sciences (Geologische Rundschau), 89/4, pp.836-851.

_____, ____, _____, ____ 2002. Menderes Masifi'ndeki Leptit-Gnayslarýn Kökenlerinin Yeniden Yorumlanmasý, Metamorfizmalarý ve Jeotektonik Ortamlarý. YDABÇAG - 554 nolu The Scientific and Technological Research Council of Turkey project, 165p (unpublished).

_____, ____, _____, ____ and Akal, C. 2005. Menderes Masifi Çine Asmasifi'ndeki Koçarlý - Bafa - Yataðan - Karacasu arasýnda uzanan gnays / þist dokanaðýnýn niteliði: Jeolojik, tektonik, petrografik ve jeokronolojik bir yaklaþým. YDABÇAG - 101Y132 nolu The Scientific and Technological Research Council of Turkey project, 197p (unpublished).

Erdoðan, B. 1992. Problem of core - mantle boundary of Menderes Massif. In proceedings of the international Symposium of Eastern Mediterranean Geology, Geosound, pp.314-315.

_____ and Güngör, T. 1992. Menderes Masifi'nin kuzey kanadýnýn stratigrafisi ve tektonik evrimi. Türkiye Petrol Jeologlarý Derneði Bülteni, 4/I, pp.9-34.

_____ and ____ 2004. The problem of the core - cover boundary of the menderes Masif and an emplacement mechanism for regionally extensive gneissic granite, Western Anatolia Turkey. Turkish Journal of Earth Sciences, 13, pp.15-36.

Gessner, K. 2000. Eocene nappe tectonics and late Alpine extension in the central Anatolide belt, western Turkey- Structure, kinematics and deformatiýn history. PhD thesis, Mainz 74p (unpublished).

_____, Piazolo, S. Güngör, T. Ring, U. Kröner, A. and Passchier, C.W. 2001. Tectonic significance of deformation in granitoid rocks of the Menderes nappes, Anatolide belt, southwest Turkey. International Journal of Earth Sciences, 89, pp.766-780.

_____, Collins A. Ring, U. and Güngör, T. 2004. Structural and thermal history of poly-orogenic basement: U-Pb geochronology of granitoid rocks in the southern Menderes Massif, Western Turkey. Journal of Geological Society London, 161, pp.93-101.

Gökten, E, Havzaoðlu, T, and Þan, Ö. 2001. Tertiary evolution of the central Menderes Massif based on structural evolution of metamorphics and sedimentary rocks between Salihli and Kiraz (western Turkey). International Journal of Earth Sciences, 89, pp.745-756.

Graciansky, P. 1965. Menderes Masifi'nin güney kýyýsý boyunca (Türkiye'nin GB'sý) görülen metamorfizma hakkýnda açýklamalar. Maden Tetkik ve Arama Dergisi, 64, pp.8-22.

Gürsu S., Göncüoðlu, C. and Bayhan, H. 2004. Geology and geochemistry of Pre-early Cambrian rocks in the Sandýklý area: Implication fort he Pan-African evolution of NW Gondwana. Gondwana Research, 7, pp. 923-935.

Hetzel, R. and Reischmann, T. 1996. Intrusion age of Pan-African augen gneisses in the southern Menderes Massif and the age of cooling after Alpine ductile extensional deformation. Geological Magazine, 133, 5, pp.565 - 572.

_____, Romer, R., Candan, O. and Passchier, C.W. 1998. Geology of the Bozdað area, central Menderes Massif, SW Turkey: Pan - African basement and Alpine deformation. Geologische Rundschau, 87, pp.394-406.

Ýzdar, E. 1971. Introduction to geology and metamorphism of the Menderes Massif of western Turkey. Petroleum Exploration Society of Libya, pp.495-500.

Konak, N. 1985. Menderes Masifi'nde çekirdek - örtü iliþkilerinin yeni gözlemler ýþýðýnda tartýþýlmasý. Türkiye Jeoloji Kurultayý, bildiri özleri, 33p.

_____, Akdeniz, N. and Öztürk, E.M. 1987. Geology of the south of Menderes Massif, I.G.C.P. project no:5, Correlation of Variscan and pre-Variscan events of the Alpine Mediterranean mountain belt, field meeting, Mineral Research and Exploration Instutite Turkey, pp.42-53.

_____, Çakmakoðlu, A., Elibol, E., Havzoðlu, T., Hepþen, N., Karamanderesi, I.H., Keskin , H., Sarýkaya, H., Sav, H. and Yusufoðlu, H. 1994. Development of thrusting in the median part of the Menderes Massif. Abstracts 47th Geological Congress Turkey-Ankara, 34p.

Koralay, O.E., Candan, O.,Dora, O.Ö., Chen, F., Satýr, M. and Kaya, O. 2002. Single zircon Pb-Pb ages for para- and orthogneisses in the Menderes Massif, western Turkey: An approach to the original deposition age of the paragneisses. 1st.International Symposium of faculty of mines (Istanbul Technical University) on Earth Sciences and Engineering, 105p.

_____, Dora, Ö., Candan, O., Chen, F. and Satýr, M. 2003. Menderes masifindeki paragnayslarýn ilksel çökelme yaþýna tek zirkon Pb/Pb evaporasyon jeokronolojisi yöntemi ile yaklaþým. 56. Türkiye Jeoloji Kurultayý, pp.64-65.

_____, Dora, O.Ö., Chen, F., Satýr, M. and Candan, O. 2004. Geochemistry and geochronology of orthogneisses in the Derbent (Alaþehir) area, Eastern part of the Ödemiþ - Kiraz submassif, Menderes Massif: Pan-African magmatic activity. Turkish Journal of Earth Sciences, 13, pp.37-61.

_____, Chen, F., Candan, O., Dora, O.Ö., Satýr, M. and Oberhänsli, R. 2005. Pb-Pb geochronology of detrital zircons from Neoproterozoic paragneisses in the Menderes Massif, Turkey.

International Earth Sciences Colloquium on the Aegean Regions, Ýzmir-Turkey, 69p.

Koralay, E., Chen, F., Oberhänsli, R., Wan, Y. and Candan, O. 2006. Age of Granulite Facies Metamorphism in the Menderes Massif, Western Anatolia / Turkey: SHRIMP U-Pb Zircon Dating. 59. Türkiye Jeoloji Kurultayý Bildiri özleri, Ankara, pp.28-29.

_____, Candan, O., Dora, Ö., Satýr, M., Oberhansli, R. and Chen, F. 2007. Menderes Masifi'ndeki Pan-Afrikan ve Triyas yaþlý metamagmatik kayaçlarýn jeolojisi ve jeokronolojisi, Batý Anadolu, Türkiye. Menderes Masifi Kolokyumu, Ýzmir, pp.24-31.

Kröner, A. 1984. Late Precambrian plate tectonics and orogeny: a need to redefine the term Pan-African. In: J Klerkx, J Michot (eds). African Geology, Tervuren: Musee. Royal I'Afrique Centrale, pp.23-28.

Kun, N. 1983. Petrography of the Çine region and petrologic findings of southern part of Menderes Massif. PhD Thesis Dokuz Eylül Üniversitesi Izmir, 125p. (unpublished).

_____ and Candan, O. 1987*a*. Menderes Masifi'ndeki Erken Paleozoyik yaþlý bazik damar kayalarý. Yerbilimleri, 14, pp.121-132.

_____ and Candan, O. 1987-*b*. Ödemiþ Asmasifi'ndeki leptitlerin daðýlýmlarý, konumlarý ve oluþum koþullarý. TBAG - 688 nolu TÜBÝTAK (The Scientific and Technological Research Council of Turkey) project, 133p. (unpublished).

Kun, N. and Candan, O. 1991. Menderes Masifi güneydoðusunda kalan Karýncalýdað çevresinin jeolojisi, petrografisi ve metavolkanitlerin (leptit) kökeni. Selçuk Üniversitesi Mühendislik Fakültesi Dergisi, 1. pp.31-44.

____, ____ and Dora, O.Ö. 1988. Kiraz-Birgi yöresinde (Ödemiþ-Menderes Masifi) metavolkanitlerin (leptitlerin) varlýðý. Türkiye Jeoloji Bülteni, 31, pp.21-28.

Loos, S. 1995. Alterbestimmungen im SW Menderes Massiv, Türkei, mit der einzirkon - Pb/Pb evaporatýosmethode. Diplomarbeit, Johannes Gutenberg Universitat Mainz, 95p (unpublished).

Loos, S. and Reischmann, T. 1999. The evolution of the southern Menderes massif in SW Turkey as revealed by zircon dating. Journal of Geological Society London. 156, pp. 1021-1030.

Monod, O., Kozlu, H., Ghienne, W., Dean, W.T., Günay. Y., Herisse, A. and Paris. F. 2003. Late Ordovician glajiation in southern Turkey. Terra Nova, 15, 4, pp.249-257.

Neubauer, F. 2002. Evolution of Late Neoproterozoic to early Palaeozoic tectonic elements in central and southern European Alpine mountain belt: review and synthesis. Tectonophysics 352, pp.87-103.

Oberhänsli, R., Candan, O., Dora, O.Ö. and Dürr, St. 1997. Eclogites within the Menderes Crystalline Complex / western Turkey / Anatolia. Lithos, 41, pp.135-150.

______,______ and Wilke, W. 2010. Geochronologic Evidence of Pan-African Eclogites from the Menderes Massif, Turkey. Turkish Journal of Earth Sciences, (in printhouse).

Okay, A. 2001. Stratigraphic and metamorphic inversions in the central Menderes Masif: a new structural modal. International Journal of Earth Sciences, 89, pp.709-727.

Partzsch, J. H., Oberhänsli, R., Candan, O. and Warkus, F. 1998. The Menderes Massif, W-Turkey: A complex nappe pile recording 1.0 Ga of geological history. Third International Turkish Geology Symposium, Middle East Technical University, Ankara, 281p.

Ring, U., Gessner, K., Güngör, T. and Passcchier., C. 1999. The Menderes Massif of western Turkey and the Cycladic Massif in the Aegean-do they really correlate? Journal of Geological Society London, 155, pp.3-6.

_____, Kröner, A., Buchwaldt, R., Toulkeridis, T. and Layer, P.W. 2002. Shear-zone patterns and eclogite-facies metamorphism in the Mozambique belt of northern Malawi, east-central Africa: implications for the assembly of Gondwana. Precambrian Research, 116/1-2, 19-56.

Satır, M. and Friedrichsen, H. 1986. The origin and evolution of the Menderes Massif, W-Turkey: A rubidium/ strontium and oxygen isotope study. Geologische Rundschau, 75/3, pp.703-714.

Schuiling, R.D. 1962. On petrology, age and structure of the Menderes migmatite complex (SW - Turkey). Bulletin of Mineral Research and Exploration Institute of Turkey, 58, pp.71-84.

_____ 1973. The Cyclades; an early stage of oceanisation? Bulletin of Geological Society Greece, 10, pp.174-176.

Scotford, D.M. 1969. Metasomatic Augen gneiss in greenschist facies, Western Turkey. Geological Society America Bulletin, 80, pp.1079-1094.

Stampfli, G.M. and Borel, G.D. 2002. A plate tectonic model for the Palaeozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrones. Earth and Planetary Sciences Letters, 196, pp.17-33.

Stern, R.J. 1994. Arc assembly and continental collusion in the Proterozoic east African orogen: Implications for the consolidation of Gondwanaland. Annual Review Earth Planetary Sciences, 22, pp.319-351.

Şengör, A.M.C., Satır, M. and Akkök, R. 1984. Timing of tectonic events in the Menderes Massif, western Turkey. Implications for tectonic evolution and evidence for Pan-African basement in Turkey. Tectonics, 3(7), pp.693-707.

Şengül, F., Candan, O., Dora, O.Ö. and Koralay, E. 2006. Petrography and geochemistry of paragneisses in the Çine Submassif of the Menderes Massif, Western Anatolia. Turkish Journal of Earth Sciences, 15, pp.321-342.

Wilson, T.J., Grunow, A.M. and Hanson, R.E. 1997. Gondwana assembly: The view from southern Africa and east Gondwana. Journal of Geodynamics, 23, ¾, pp.263-286.

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THE GEOLOGY AND GEOCHRONOLOGY OF THE PAN-AFRICAN AND TRIASSIC METAGRANITOIDS IN THE MENDERES MASSIF, WESTERN ANATOLIA, TURKEY

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ABSTRACT.- The Menderes Massif is a metamorphic complex cropping out on a large region in the Alpine orogenic belt in Western Anatolia. The massif mainly, is made up of a Precambrian basement and the overlying Palaeozoic-Early Tertiary cover series. The basement comprises Late Proterozoic metaclastics composed of paragneiss and high grade micaschists, syn to post-tectonical Pan-African orthogneisses that have intruded into them, and metagabbros which have partly turned into eclogitic form. Cover series unconformably overlying the basement are divided into two units, in Palaeozoic and Mesozoic-Early Tertiary ages. The basement and cover series were influenced by an effective Alpine contractional deformation and a regional metamorphism in Tertiary time. Geological and geochronological data in the Menderes Massif, indicate the presence of three acidic magmatic activities of i) Pan-African, ii) Triassic and iii) Miocene in ages. The acidic magmatics forming the protolithes of Pan-African orthogneisses can be divided into three main types according to their textural and mineralogical compositions as; 1) Biotite orthogneiss, 2) Tourmaline leucocratic orhogneiss and 3) Amphibole orthogneiss. The relations of primary granites of these orthogneisses present a clear intrusive contact relationship among them and with the metaclastics of Late Proterozoic age. The contact relationships show that these orthogneiss types can be ordered as biotite orthogneisses, tourmaline leucocratic orthogneisses and amphibole orthogneisses ranging from old to young, with respect to relative aging. Although there has been some problems originating from the definition of samples dated in previous studies, it is noted that radiometric data mainly show a consistency with this relative relationship (biotite orthogneiss: 550-570 Ma; tourmaline leucocratic orthogneiss: 541-547 Ma and amphibole orthogneiss: 531 Ma). These radiometric data indicate that different orthogneiss types in the Massif are differentiated products of the same Late Proterozoic - Early Cambrian acidic magmatic activity. When the paleogeographical position of the Massif in Early Cambrian and the close temporal relation of the metamorphic stage of cover series and the acidic magmatism are assessed with the geochemical character of orthogneisses, this widespread magmatic activity with an average age of about 550 Ma which is related to the Pan-African orogenesis can be attributed to proccesses of closure of the Mozambique Ocean, collision of East and West Gondwanaland, crustal thickening and partial melting of the lower crust. Triassic leucocratic orthogneisses constitutes the second effective acidic magmatic activity in the Menderes Massif with dimensions of 6-7 km. These plutonic leucocratic orthogneisses in 6-7 km dimensions are exposed in Ödemiþ-Kiraz and Demirci-Gördes submassifs. They show well-preserved intrusive contact relationships with Late Proterozoic metaclastics of the Pan-African basement and with metasediments of Late Palaeozoic cover series. Geochemical data show that protolithes of leucocratic orthogneisses are in calc alkaline and S type in character. Based on single zircon Pb/Pb evaporation method, these were radiometrically dated as ranging from 227 to 246 Ma. These ages are interpreted as the age of emplacement of protolithes of orthogneisses in Middle Triassic. The existence of Early-Middle Triassic magmatic activity is widely known not only in the Massif but also at the tectonical zones of the Anatolides, at the Karaburun peninsula, in Cyclades and at the inner Hellenides. When the regional character of Triassic magmatic activities is considered, it is suggested that there is a close genetical relationship between these leucocratic orthogneisses and Triassic magmatics. It is also considered that it can be attributed to the opening of Neothethys Ocean.

Key words: Menderes Massif, Pan-African, Cadomian, Triassic, magmatism, geochronology

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INTRODUCTION

The Menderes Massif which is extending in NE-SW direction and presenting strong interior nappes produced by Alpine compressional tectonics, has a great importance in the geological evolution of Western Anatolia. The Menderes Massif is tectonically overlain by the extension of Izmir-Ankara suture zone including also the Bornova Flysch zone in the west and northwest, by the extension of Cycladic complex in Turkey by the Afyon zone containing high pressure/low temperature (HP/LT) metapelite and metacarbonates at the north, and by high pressure metasediments and Lycian nappes with thick ophiolitic slices at the south (Figure 1) (Þengör and Yýlmaz, 1981; Okay, 1984; Dora et al., 1995). In previous studies, there is a general acceptance that the Cycladic Complex in Aegean Sea is the eastern extension of the Menderes Massif (Dürr et al., 1978; Jacobshagen, 1986; Candan and Dora, 1998). In recent studies, although there have been many objections in dimension and distribution, it was suggested that, Cyclades were exposed as a tectonic slice in Western Anatolia and present clear differences from the Menderes Massif (Candan et al., 1997; Ring et al., 1999; Gessner et al., 2001; Okay, 2001).

The Menderes Massif is mainly divided into two main rock groups as; i) Pan- African basement and ii) Palaeozoic-Early Tertiary cover series (Figure 2) (Dora et al., 1995). The Pan-African basement is predominantly composed of metaclastic sediments of which the depositional age of protolithes is Late Proterozoic (Koralay et al., 2003). Cover series on the other hand, are divided into two sub groups as; Palaeozoic and Mesozoic-Early Tertiary rocks. Palaeozoic series are dominantly composed of quartzite, phyllite and marbles (Çaðlayan et al., 1980; Konak et al., 1987). Mesozoic-Early Tertiary series start with metaconglomeratic schist at the bottom and grade into the platform type thick metacarbonates including emery lenses with a transitional zone. Carbonates are overlain by pelagic marble and the deposition ends with metaolisthostrome (Dürr, 1975; Konak et al., 1987; Dora et al., 1995; Özer et al., 2001).

Metaclastic series made up of partly migmatized paragneiss and micaschists forms the oldest unit of the Pan-African basement of the Menderes Massif. These Late Proterozoic rocks were affected by poly-metamorphism under granulitic, eclogitic and amphibolitic facies conditions related with the Pan-African Orogeny in Precambrian-Cambrian (Dora et al., 1995, Oberhänsli et al., 1997; Candan and Dora 1998; Candan et al., 1994, 2001, 2007). Basement series are cut by widespread acidic/basic magmatics related with the Pan-African Orogeny (Hetzel and Reischmann 1996; Loos and Reischmann 1999; Koralay et al., 2004; Candan, 1996). As a contrast, in recent years, in tectonical studies, especially in the southern part of the Çine submassif, it has been claimed that the most typical rocks belonging to the basement and the Pan-African aged acidic magmatics (orthogneisses) that were obtained by geochronological studies are i) Late Cretaceous (Erdoðan and Güngör, 2004) or ii) Early Tertiary intrusives (Bozkurt et. al., 1995). Geochronological data show that, the main acidic magmatic activity phase forming the protolith of Pan-African orthogneisses has occurred between 520-570 Ma (Late Proterozoic-Cambrian) with a strong emphasis for 550Ma (Hetzel and Reischmann 1996; Loos and Reischmann 1999; Dannat, 1997; Hetzel et al., 1998; Gessner et al., 2001, 2004; Koralay et al., 2004).

Triassic leucocratic orthogneisses form the second prevalent magmatism observed in the Menderes Massif. In eastern part of Ödemiþ-Kiraz submassif, at the south of Alaþehir, at the east of Aydýn-Köþk, at the southeast of Demirci-Gördes submassif and at the south of Kula, NNE directed leucocratic orthogneiss masses cut Precambrian basement series and Permo-Carboniferous units by intrusive contacts (Akkök, 1983; Candan, 1994; Koralay et al., 2001). From these, by the single zircon Pb/Pb evaporation

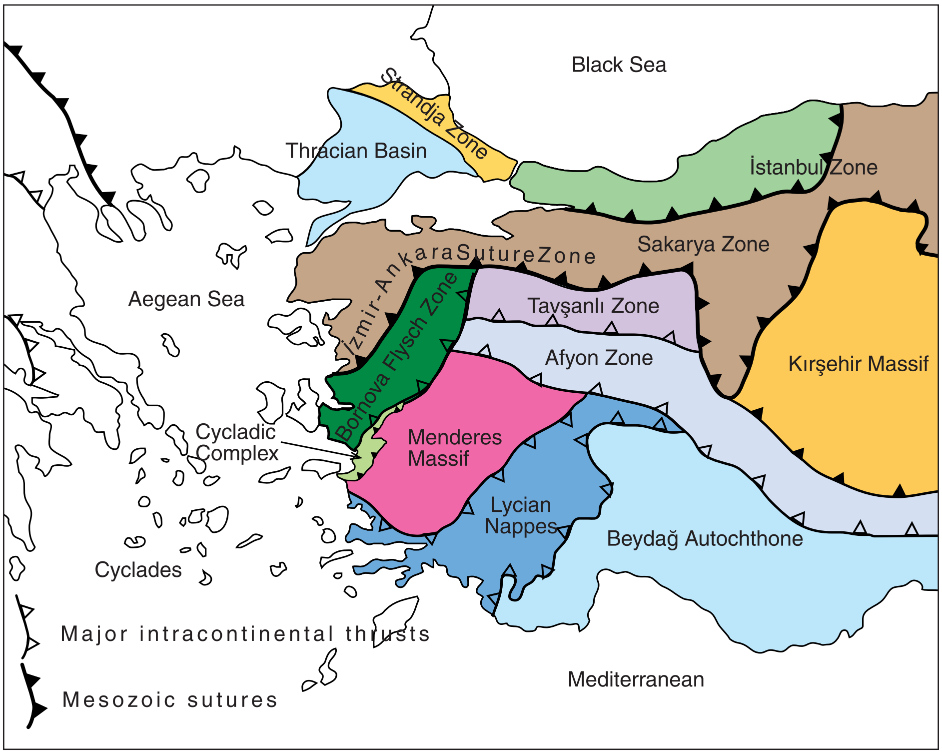

Figure 1- Generalized tectonical map of the Eastern Anatolia and tectonical zones surrounding the Menderes Massif (modified from Okay et al., 1996).

method 230-245 Ma (Early-Middle Triassic) ages have been dated (Dannat, 1997; Dannat and Reischmann 1998; Koralay et al., 2001). Koralay et al. (2001) claimed that these granites were intruded in the period that follows Early Kimmerian metamorphism related with the closure of Paleothethys. Non metamorphic, 12-25 Ma aged granites and Kersantites represent the third magmatic activity (Öztunalý, 1973; Bingöl et al., 1982; Reischmann et al., 1991; Hetzel et al., 1995; Delaloye and Bingöl 2000; Iþýk and Tekeli 2000; Iþýk et al., 2004; Lipps et al., 2001; Catlos and Çemen, 2005; Thompson and Ring, 2006; Glodny and Hetzel, 2007).

In the studies published during the last 10 years, there has been a contradiction in the interpretations of geochronological data and field results and on kinematical data regarding the intrusion ages of orthogneisses except for Triassic intrusions. It is considered that, representation of the available data towards the solution of this problem and associating these with geological data will illuminate the future studies. The paper was prepared in this purpose and aims at presenting; i) the geological/petrographical properties and the distributions of orthogneisses, found in the core series, in the Menderes Massif, ii) the geochronological data obtained from these orthogneisses and the association of these data with geological data, iii) the distribution of Triassic leucocratic orthogneisses in the Menderes Massif, iv) the geochronological ages obtained from Triassic aged leucocratic orthogneisses and probable tectonical environments.

ACIDIC MAGMATISM IN THE MENDERES MASSIF

As it has been mentioned briefly above, the geological and geochronological data in the

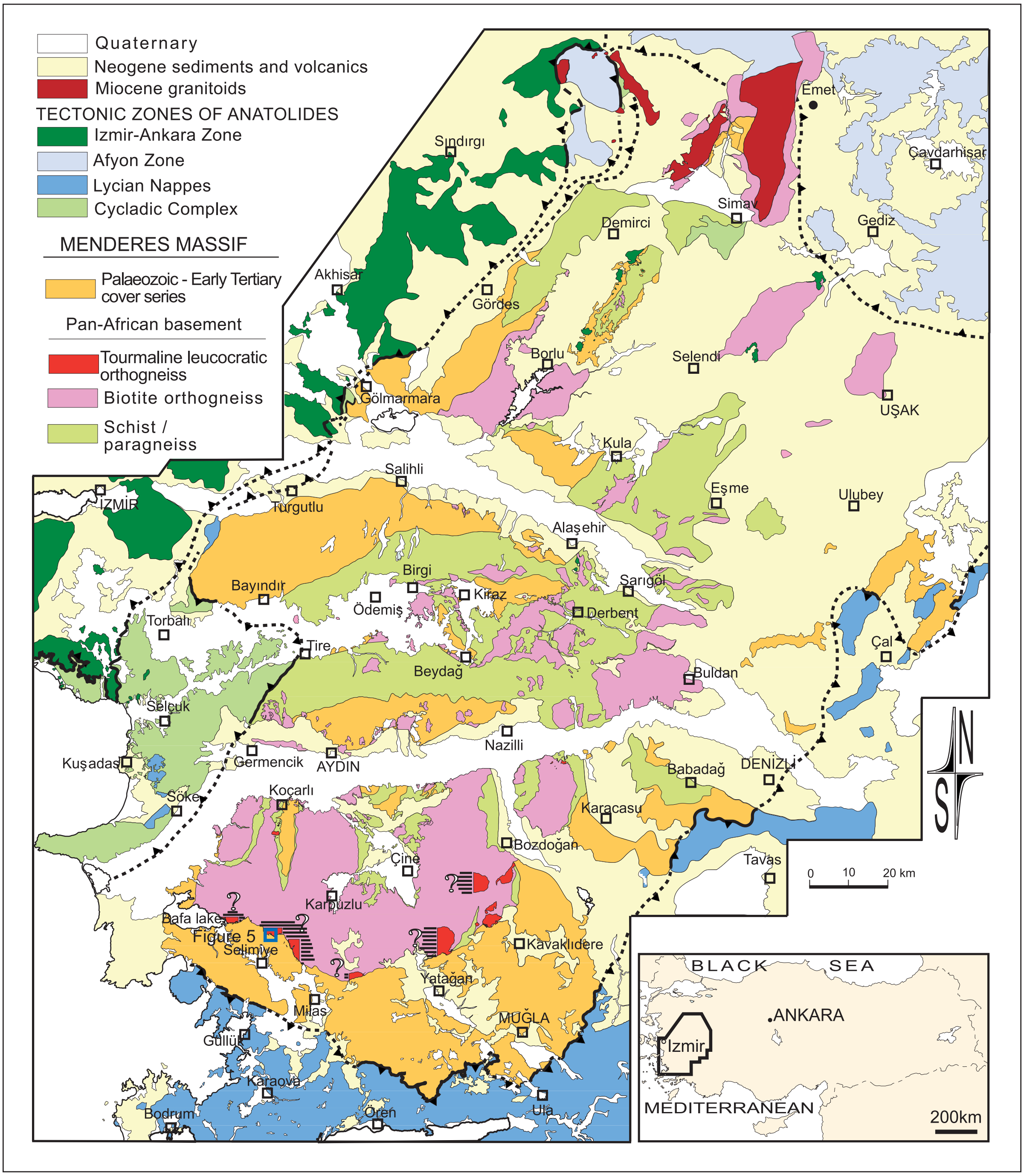

Figure 2- The general distribution of the Pan-African aged acidic magmatics in the Menderes Massif. Areas drawn in horizontal lines represent partly mapped probable extensions of these intrusives around the masses of tourmaline leucocratic orthogneisses (simplified from Candan and Dora, 1998).

Menderes Massif establish the existence of the acidic magmatic activity in three different ages. These are; i) Pan-African (Late Proterozoic-Cambrian), ii) Triassic, and iii) Tertiary rocks. Orthogneisses forming the Pan-African aged acidic magmatism are widely observed in the core series of the Menderes Massif and forms the most characteristic rock type of this basement (Hetzel and Reischman, 1996; Dannat, 1997; Hetzel et al., 1998; Loos and Reischman, 1999; Gessner et al., 2001; Koralay, 2001; Gessner et al., 2004; Koralay et al., 2004). Granitic protolithes of Triassic leucocratic orthogneisses form the second main magmatic activity in the Massif (Dannat 1997; Koralay 2001; Koralay et al., 2001). The Miocene aged third group is the intrusion in post metamorphic character and is observed in the central and in the northern parts of the Massif, in Tavþanlý zone located at north of the Massif, and in Sakarya Zone (Öztunalý, 1973; Bingöl et al., 1982; Reischmann et al., 1991; Hetzel et al., 1995; Delaloye and Bingöl, 2000; Iþýk and Tekeli, 2000; Iþýk et al., 2004; Lipps et al., 2001; Catlos and Çemen, 2005; Thompson and Ring, 2006; Glodny and Hetzel, 2007). The reason of these granites are not limited within the Menderes Massif and is genetically associated with the general evolution of northwest Anatolia makes it out of the scope of this paper. The geological and petrographical properties of the Pan-African and Triassic acidic metamagmatics are given below.

PAN-AFRICAN MAGMATISM

The distribution of the Pan-African Magmatism in the Menderes Massif

In recent studies performed in the Menderes Massif, the tectonical structure of this crystalline basement has been shaped by the Precambrian and Alpine contractional deformations and presents a complex internal structure. During the investigations, it was determined that the Pan-African base rocks are tectonically intercalated with Palaeozoic-Early Tertiary cover series in addition to its own intrinsic imbrications (Konak et al., 1994; Partzsch et al., 1998; Ring et al., 1999). The primary stratigraphy of the Massif has almost lost its original structure as a result of this Alpine compressional tectonics. The sequence belonging to Pan-African basement obtained by the correlation of stratigraphies of different tectonic slices is given below.

When the Menderes Massif is generally considered, Late Proterozoic metaclastic series forms the oldest rocks of the Pan-African basement in each of three submassifs. Paragneisses which are the oldest rocks of this series conformably grade into schist unit in upward direction (Dora et al., 2002). In many tectonical slices, paragneiss unit and conformably and transitionally overlying schist unit are observed together. The metaclastic series composed of paragneisses and schists are widely cut by granitoids with acidic composition that were intruded in different periods of the Pan-African Orogeny. When Demirci-Gördes, Ödemiþ-Kiraz and Çine submassifs which were made up of many tectonical slices were studied, it was noticed that orthogneisses took place in different tectonical slices. It is assumed that, the orthogneisses acquired their structural and textural properties as a result of the Alpine Orogeny and metamorphism. The orthogneisses named as augen and granitic gneiss in many of the previous studies are divided into two sub groups in recent investigations (Bozkurt, 2004; Dora et al., 2006; Candan et al., 2010). These are the; i) orthogneisses rich in biotite and ii) leucocratic orthogneisses rich in tourmaline. Orthogneisses rich in biotite are remarked as in big plutons of which were intruded into each other, whereas, leucocratic orthogneisses rich in tourmaline are in the form of stocks and veins in various dimensions.

The distribution of the orthogneisses mentioned above, over the Menderes massif is given in figure 2. The tourmaline rich leucocratic orthogneiss masses described so far by the above classification are shown in this map. Areas drawn in horizontal lines show the probable extensions of partly mapped tourmaline rich leucocratic orthogneisses. And the other areas described by orthogneiss in the same map, comprises the all biotite and tourmaline rich leucocratic orthogneisses. The orthogneisses which are seen most widespread in Çine submassif, are exposed in large areas in other submassifs too.

Demirci-Gördes submassif.- It is one of the region in which the all units belonging to the Pan-African basement are observed in clear contacts. The lowest unit in this submassif showing internal nappings is dominantly composed of homogenous micaschists of the Pan-African basement. These rocks having garnet- mica-schist in composition are cut by Pan-African orthogneisses and by amphibolitic metagabbro stocks at the southwest of Eþme and by Triassic leucocratic orthogneissic intrusives at the north-east of Alaþehir. On these, the gigantic ortho-gneisses present at the lowermost part of the slice extending between Alaþehir-Kula were intruded into conformably overlying micaschists (Candan, 1994, Dora et al., 2002). Within orthogneisses, migmatite and paragneiss traps which were partly assimilated, are widely observed in a dimension of 2-3 km. This Pan-African aged slice is overlain by phyllite-quartzite-marble intercalations and by Palaeo-zoic-Mesozoic slice composed of platform type marbles overlapping on these.

Ödemiþ-Kiraz submassif.- The stratigraphy of this submassif has been established in more detailed compared to the other submassifs (Partzsch et al., 1998; Ring et al., 1999; Candan et al., 2001; Dora et al., 2002, 2006). The lower-most slice of the metamorphic deposit is made up of Upper Cretaceous marbles defined by the existence of emery deposits and by well pre-served rudist fossils (Koralay, 2001; Özer and Sözbilir, 2003) which are observed in Aydýn-Eðrikavak village. The platform type marbles are tectonically overlain by Palaeozoic series along the southern margin of Aydin mountains. Phyllites found in the chlorite-chloritoid bearing series in Aydin Mountains and in staurolite-garnet bearing series in Bozdaglar are cut by Triassic (235 Ma) vein rocks and by leucocratic metagranitic stocks. Palaeozoic series is overlain by three different slices belonging to the Pan-African basement. The lowermost slice is composed of homogenous micaschists which are exposed in large areas in Aydin mountains and Bozdaglar. South of Alaþehir is characterized by the othogneiss that was intruded into medium slice micaschists observed in Derbent and by the Triassic leucocratic orthogneissic intrusives. The paragneiss which was migmatised at lower the parts and the uppermost Pan-African slice made up of conformably overlying micaschists, are cut by the masses of metagabbro. The uppermost layer of the Menderes massif metamorphites is made up of Palaeozoic(?) clastics with the composition of staurolite-phyllite-marble and of the overlying Mesozoic platform type marbles containing emery beds (Koralay et al., 2001). The phyllite-marble community forming the uppermost tectonic unit at south of Derbent is located on the middle tectonic layer with a tec-tonic contact interpreted by a slip fault.

Çine submassif.- The lowermost tectonic unit of the Çine submassif is exposed in eastern and western parts of Çine-Bozdoðan. This slice is dominantly composed of garnet micaschists containing thin paragneiss layers. This clastic series is cut by orthogneissic and metagabbroic intrusives in variable compositions. The tectoni-cal slice overlying this clastic series and present-ing a very large propagation in the middle part of the submassif is dominantly made up of Pan-African granites with different characters that were intruded in periods following one another. Orthogneisses crop out in Yenipazar and Koçarlý at the north, in Karacasu and Bozdoðan at the east, in Bafa Lake at the west and in a large region between Milas and Yataðan at the south (Figure 2). In orthogneisses, there are mig-matised paragneiss and schist traps with a dimension of about 5-6 km. Garnet micaschists observed especially along the southern boundary of the submassif form the country rock of which the orthogneisses were intruded. In this slice, the primary unconformable contact relationship between the Pan-African basement and Palaeozoic cover series can be observed (Konak et al., 1987; Candan et al., 2006, 2007; Dora et al., 2006). The Palaeozoic deposit starting with quartzite which contains metaconglomeratic layers is made up of porphyritic metagranite, tourmaline leucocratic orthogneiss, metaaplite and tourmaline pebbles in channel fillings character. This deposit continues with a series of garnet-chloritoid-phyllite, marble and quartzite intercalations. The Palaeozoic deposit showing imbrication is overlain by metacarbonates produced from Mesozoic- Early Tertiary aged Anatolide-Tauride platform. Red colored pelagic marbles and Paleocene-Eocene aged Na-amphibol bearing blocky series constitute the youngest units of the Çine submassif.

THE CONTACT RELATIONSHIP BETWEEN ORTHOGNEISSES AND PAN-AFRICAN METASEDIMENTS

The contact relationship of orthogneisses (biotite orthogneiss and tourmaline leucocratic orthogneiss) with thick metaclastic deposit forming the country rock presents common properties in each of three submassifs. The primary contact relationship of granitic orthogneisses with metaclastics forming the country rock is in intrusive character in all regions. Late evolution products of protolithes of orthogneisses that were intruded into schists such as pegmatite, aplite and tourmaline belong to Pan-African basement and are widely observed in both units close to contact areas. The contact relationship of orthogneiss/metaclastic well observed in the various regions of Demirci-Gördes, Ödemiþ-Kiraz and Çine submassifs is described below in detail.

Candan (1994) stated that, in south of Kula in Demirci-Gördes submassif, the granites of which are the protolithes of orthogneisses revealed an intrusive contact relationship with paragneiss unit. Many apophysis of orthogneisses with dimensions of several kilometers intrude into paragneisses. Besides, gneiss stocks in large and small dimensions are observed in paragneiss, close to gneissic contact. In gneisses close to contact zone, the existence of country rock traps range from a few cm. to a couple of km. and this is another significant data supporting the intrusive contact relationship. Especially, in undeformed orthogneisses, the detection of paragneiss, migmatised paragneiss and calc silicate rocks showing a strong foliation, in the form of partly assimilated traps (Figure 3a,b), indicate that granites of which are the protolithes of orthogneisses has been intruded after the metamorphisms of Pan- African paragneisses.

Ödemiþ-Kiraz submassif.- to the south of Derbent of Ödemiþ-Kiraz submassif, orthogneisses intrude into schists (Koralay, 2001, Koralay et al., 2004). An intrusive contact relationship between orthogneisses and schists is clearly observed in intermediate slice. Many dikes and sills belonging to orthogneisses are located in schists forming the country rock. It is clearly observed that some of these vein rocks cut the schistosity of schists (Figure 3c). Similar to other regions, schist traps are available in various sizes among orthogneisses, close to schist contacts. The original intrusive contact relationships of orthogneisses with paragneisses in the size of submassif can be detected in almost everywhere and be depicted with similar data. In addition to this, as in the vicinity of Semit and Cevizalan villages, strong ductile deformation evidences can be detected at contacts of paragneisses and orthogneisses that have planar geometry and thickness reaching even 1 km. It is considered that these orthogneiss masses settled into paragneisses in the form of tectonic slices under ductile conditions during the intrinsic imbrication of the units belonging to the Pan-African basement.

Çine submassif is characterized by the existence of gigantic granites in the structure of orthogneiss which represents varieties in composition. The character of orthogneiss/schist contact has been studied by many investigators. It has lateral extend of 100 km and located at the south of the submassif. Field data brings up that, the primary relationship along the contact between the Upper Proterozoic metaclastics forming the country rock and all orthogneiss types is intrusive in character. This relationship is not only limited to this contact but can also easily be detected at northwest of the submassif, south of Koçarlý (Dora et al., 2005). In this region, there is a clear intrusive contact relationship between gigantic sized, biotite rich orthogneiss plutons and tourmaline leucocratic orthogneiss stocks and Late Proterozoic schists. This contact is described especially by the existence of widespread vein rocks (Figure 3d,e,f). It can also be detected that, there is a primary intrusive contact relationship between the orthogneisses and paragneisses, in the form of absorbed traps at the north and southwest of Çine where there is less deformation.

As mentioned above, east of Bafa Lake is one of best places where intrusive relationships of orthogneisses are observed with Late Proterozoic schists forming the country rock. In addition, to Bademyaný point located at the western part of Bucak village, the intrusive relationships of tourmaline leucocratic orthogneisses with Upper Proterozoic schists can clearly be detected at the north of Bucak and Karahayýt villages. At Bademyaný point, there are many vein rocks in schists and near the tourmaline leucocratic orthogneisses widespread schist traps are available. This case complicates the determination of the main contact boundary between the orthogneiss and schist (Figure 4a,b). Although the contact was subjected to ductile deformation, similarly the primary intrusive relationship is observed fairly between schists and orthogneisses around Viranköy and Ekiztaþ villages located at the eastern most part of the contact (Figure 5). At the south of Katrancý village, biotite orthogneiss and tourmaline leucocratic orthogneisses present a well preserved intrusive contact relationships with Late Proterozoic schists forming the country rock (Figure 6a). Schist traps in biotite orthogneisses and aplitic vein rocks in those schists are widely observed in this region. Besides, the intrusive contact relationship between biotite orthogneisses and tourmaline leucocratic orthogneisses is also clearly observed which represent two different phases of Pan-African magmatic activity (Figure 6b).

Around Seykel village located at the south of Çine, the primary intrusive contact relationships of biotite orthogneisses and tourmaline leucocratic orthogneisses is well preserved. This relationship type between the orthogneisses and schists belonging to the Pan-African basement which forms the country rock is also clearly detected. This intrusive contact relationship is noticeable in cuts along Yataðan-Çine road. Field data reveal that, granites of which are the protolithes of biotite orthogneisses in the primary stage were intruded into schists. Along contact of these rocks, gneissic veins in schists and schist traps in gneiss are widely remarked (Figure 7a, c). The tourmaline leucocratic orthogneiss mass belonging relatively to a latter stage of the same Pan African magmatic activity shows a clear intrusive contact relationship that can be detected in biotite orthogneisses and schists (Figure 7d). There are many leucocratic vein rocks with a thickness of 1.5 m within schists forming the country rock and in a zone of 200 m (Figure 7e). At contacts of those veins, mica minerals were developed in the form of thin zones as a result of contact metamorphism (Figure 7f). The parts shown by nonexistence of any orientation in mica minerals are the parts protected from the Alpine contractional deformation in which the original contact relationship is observed.

Around Mesken village located at the north of Yataðan, many biotite orthogneisses, tourmaline leucocratic orthogneisses and porphyritic metagranites belonging to poly-phased Pan-African magmatic activity show an intrusive contact relationship conformable with relative age relations (Figure 8a,b). In the region, to the north of Yukarý

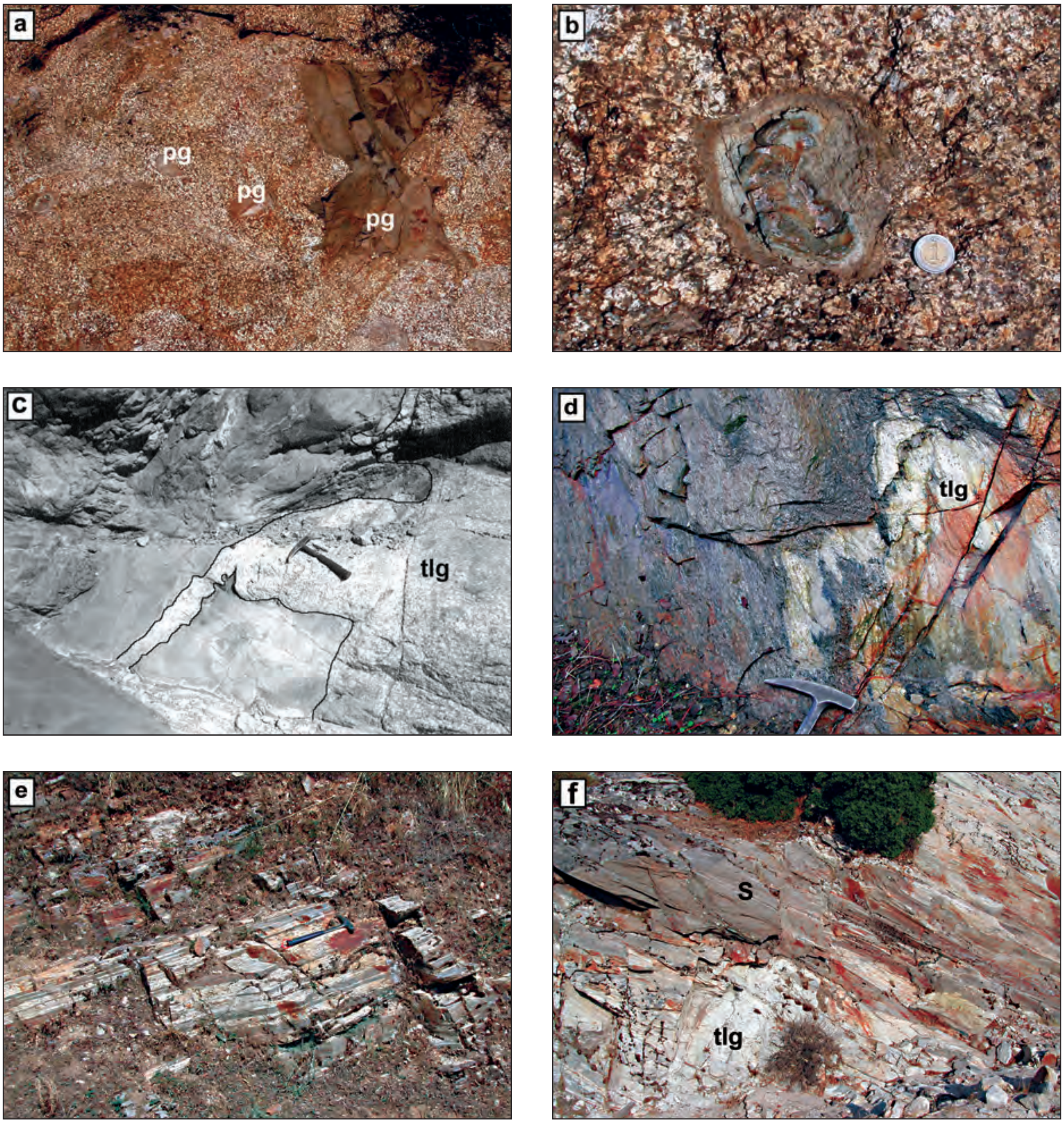

Figure 3- a) Paragneiss showing strong foliation in non oriented biotite orthogneisses, b) country rock traps formed by massif calc silicate rocks, c) the intrusive contact relationship of biotite orthogneisses located in NE Derbent with garnet micaschists forming the country rock, d) the intrusive contact relationship schists belonging to Base series with tourmaline leucocratic orthogneisses, Çulhalar village, e) tourmaline leucocratic veins extending along the foliation planes of schists, f) schist traps observed within tourmaline leucocratic orthogneisses (pg: paragneiss, ks: calc silicate, s: schist, bg: biotite orthogneiss, tlg: tourmaline leucocratic orthogneiss).

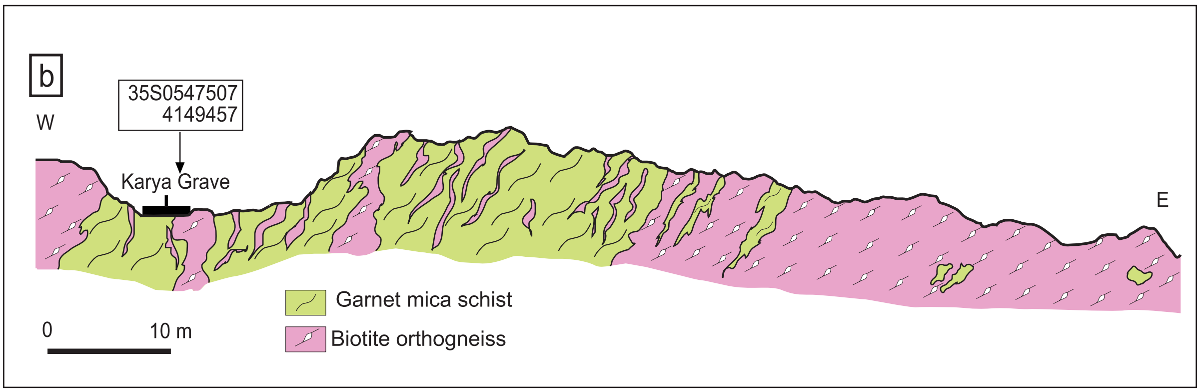

Figure 4- a) Veins formed by biotite orthogneisses and aplites within Late Proterozoic micaschists and garnet micaschist forming the country at Bademyaný foreland, b) the intrusive contact relationship observed between biotite orthogneiss and schists around Bademyaný peninsula near Bafa lake (Dora et al., 2006).

Mesken village, both biotite orthogneisses and tourmaline leucocratic orthogneisses form inter-tongues intruding into Late Proterozoic schist and that reaches several hundreds of meters. Similar to other regions, many country rock traps that are partly assimilated are available in biotite orthogneisses.

MACROSCOPIC AND MICROSCOPIC PROPERTIES OF ORTHOGNEISSES

Orthogneisses in the Menderes Massif were defined typically as augen and/or granitic gneiss by the investigators in previous studies. And, in recent studies, it was determined that it was possible to differentiate the mineralogical composition and textural properties of primary granites of these rocks (Bozkurt, 2004; Dora et al., 2005).

These rocks of which are the consecutive products of Pan-African acidic magmatic activity can be divided into three groups based on mafic phase composition and textural properties (Dora et al., 2005). These orthogneiss types and the basic petrographic properties of their primary granites can be described as follows;

Biotite Orthogneisses.- Rich in biotite (± tourmaline), coarse grained crystalline, granite with equally sized and/or in porphyritic texture.

Tourmaline Leucocratic Orthogneisses.- Rich in tourmaline and muscovite (± biotite), light colored (leucocratic), medium to coarse grained, generally equally sized, rarely porphyritic textured granite.

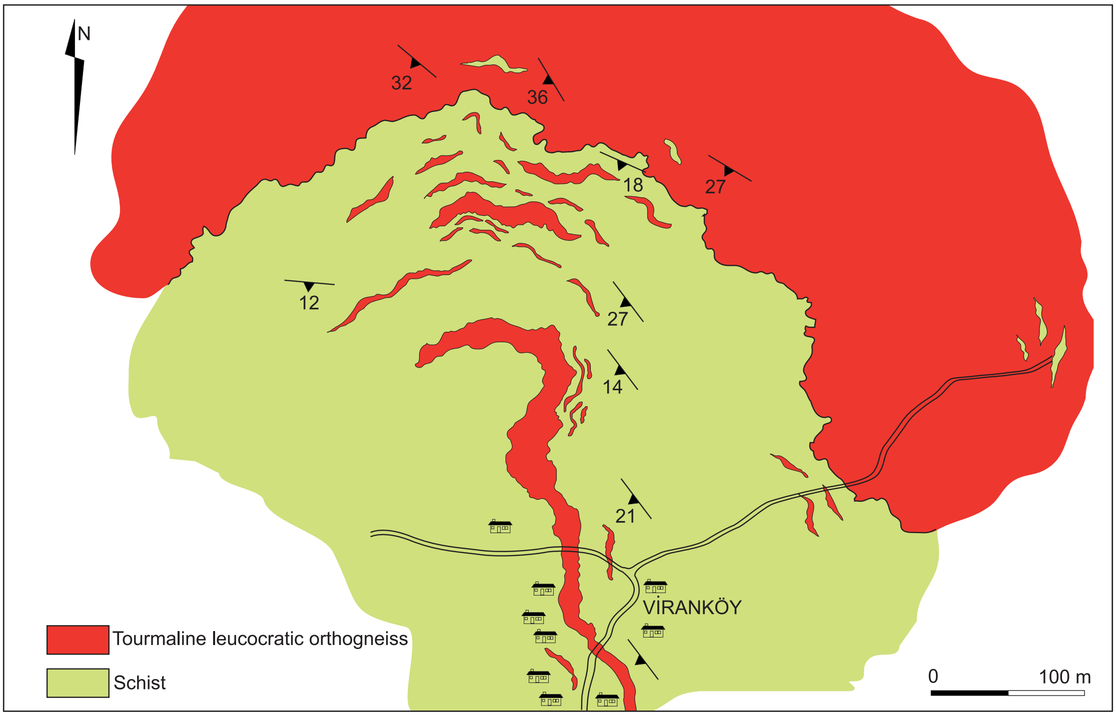

Figure 5- The geological map of the region in which the primary intrusive contact relationship is observed between the Upper Proterozoic schists belonging to the Pan African basement and the tourmaline leucocratic orthogneisses. The primary intrusive property between the two units are well preserved although these were deformed in Alpine metamorphism (Dora et al., 2006). Coordinates of Viran village: Aydýn N 19-a2 4146150: 059000.

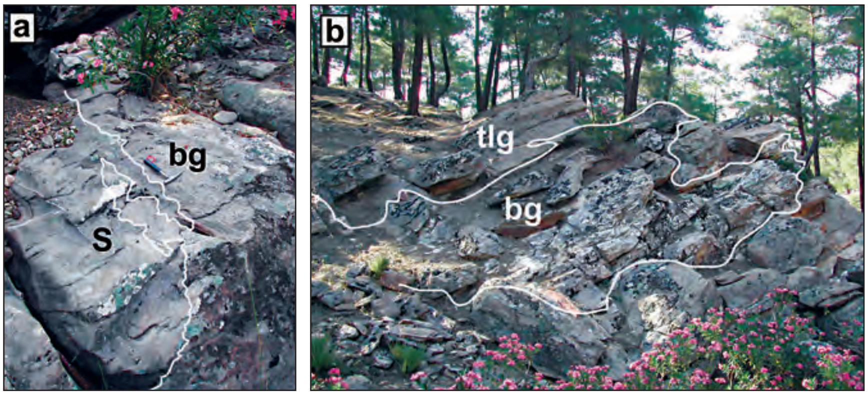

Figure 6- a) The intrusive contact relationship of biotite orthogneisses with schists, b) leucocratic orthogneiss vein cutting biotite orthogneisses by intrusive contacts, south of Katrancý village (s: schist, bg: biotite orthogneiss, tlg: tourmaline leucocratic orthogneiss).

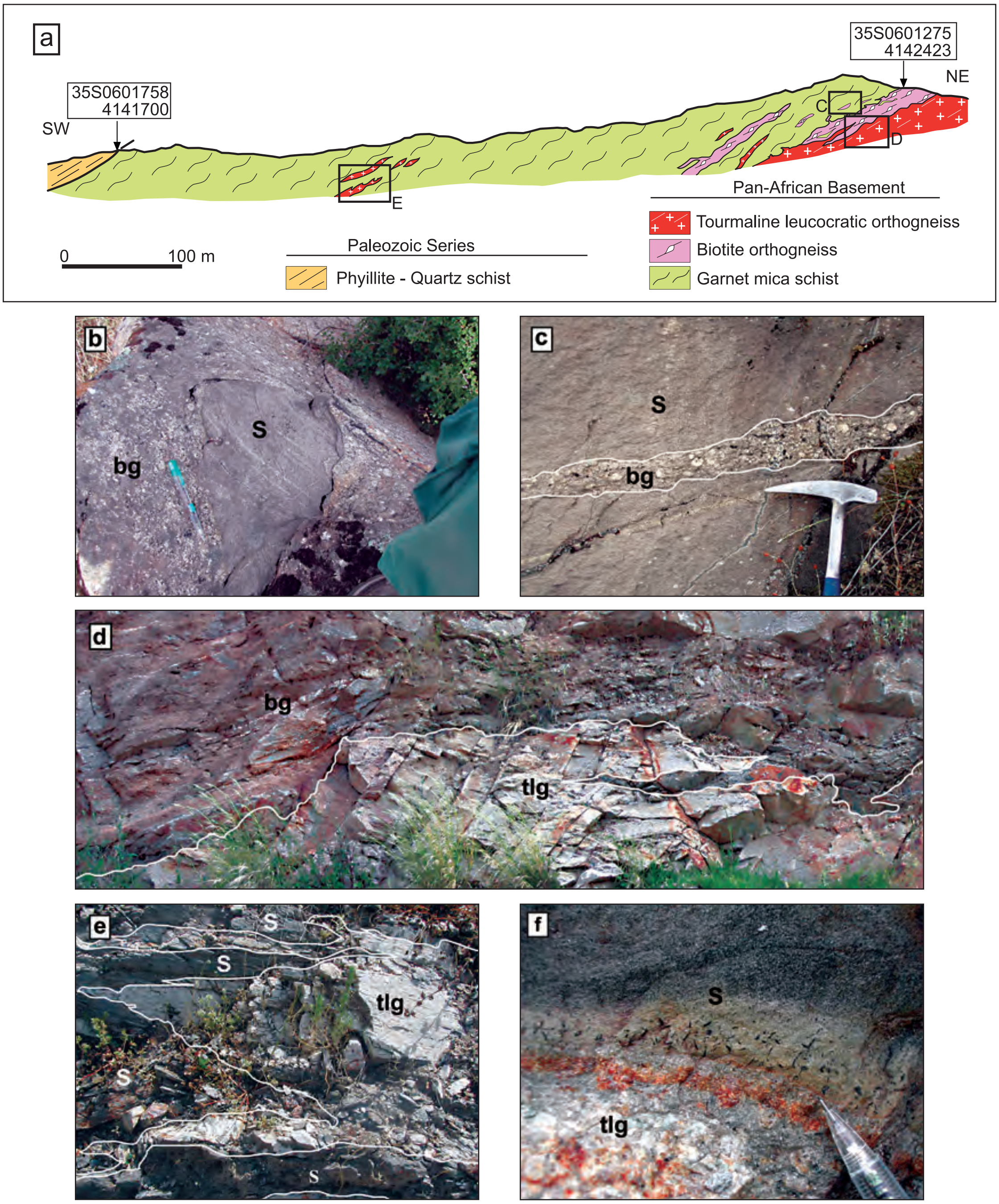

Figure 7- a) Geological cross section observed in Çine-Yataðan road cut and the contact relationships among Pan-African units (Dora et al., 2006), b) schist traps in biotite orthogneisses, c) biotite orthogneiss vein in micaschist forming the country rock, d) the sharp intrusive contact relationship between the biotite orthogneiss and tourmaline leucocratic orthogneiss, e) vein rocks belonging to tourmaline leucocratic orthogneisses in the Upper Proterozoic schist, f) contact metamorphism product preserved from Alpine overlap which is observed in contacts of leucocratic vein rocks with schists, randomly oriented mica crystals (s: schist, bg: biotite orthogneiss, tlg: tourmaline leucocratic orthogneiss).

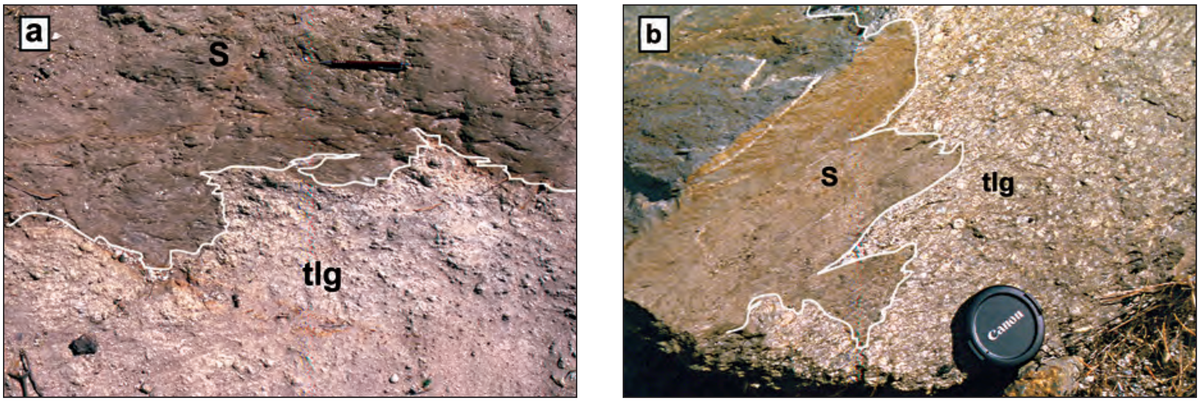

Figure 8- a) Intrusive contact relationship between the tourmaline leucocratic orthogneiss and Upper Proterozoic schist, b) the intrusive contact relationship observed between biotite orthogneisses and Upper Proterozoic schists, Mesken Village (s: schist, bg: biotite orthogneiss, tlg: tourmaline leucocratic orthogneiss).

Amphibolic orthogneiss.- Hornblende bearing (± biotite), fine to medium grained, equally sized holocrystalline textured granite.

1. Biotite Orthogneisses.- These are the most characteristic orthogneiss types that are widely observed in the Menderes Massif. The ratio of biotite mineral in these orthogneisses varies between 15 to 25 %. These are divided into two subgroups based on the primary textural properties in which the sectors were avoided from the deformation of primary granites. Porphyritic granites in phanerocrystalline texture are the most frequent orthogneiss type that is observed. These granites are described as orthoclase crystallines with a less lineation, medium grained, in a groundmass composed of quartz and plagioclase with a size of 8 to 10 cm, in euhedral to subhedral form (Figure 9a). In these rocks, based on the intensity of deformation that develop under ductile conditions, crystal sizes decrease and there occurs a traverse into blasto mylonitic gneiss (Figure 9b), generally named as augen gneiss in previous studies and ultra mylonitic banded gneisses (Figure 9c) which show a strong foliation and lineation.

Equally sized quartz and plagioclase/orthoclase crystals and grano blastic textured coarse grained granites, characterized by randomly distributed, well formed biotite crystals, form the second biotite orthogneiss type (Figure 9d). There is a widespread transformation into banded gneisses of this biotite orthogneiss type also named as granitic gneiss in sections where there occurred ductile deformation.

General mineral composition of biotite orthogneisses were defined as quartz, plagioclase, orthoclase, biotite, muscovite (± microcline, garnet, zircon, tourmaline, sphene, zoisite and opaque mineral). Quartz, plagioclase and orthoclase constitute the main components of biotite orthogneisses. Orthoclases generally cover pertitic structures and are observed in the form of porphyroblasts together with polysynthetic twinned plagioclases. In pertitic orthoclases, microcline transformations are widely observed which is accepted as the deformation product. On the other hand, the secondary twinning which is the deformation product in plagioclases and recrystallization that occur along fractures reflect in similar conditions (Figure 9e). Although coarse grained porphyroclastics originate from orthoclases in gneisses, ductile deformation product in high grade temperature, medium to fine grained feldspar rich layers are made up of microclines and this supports the idea cited above (Figure 9f). Muscovites and biotites depending on the

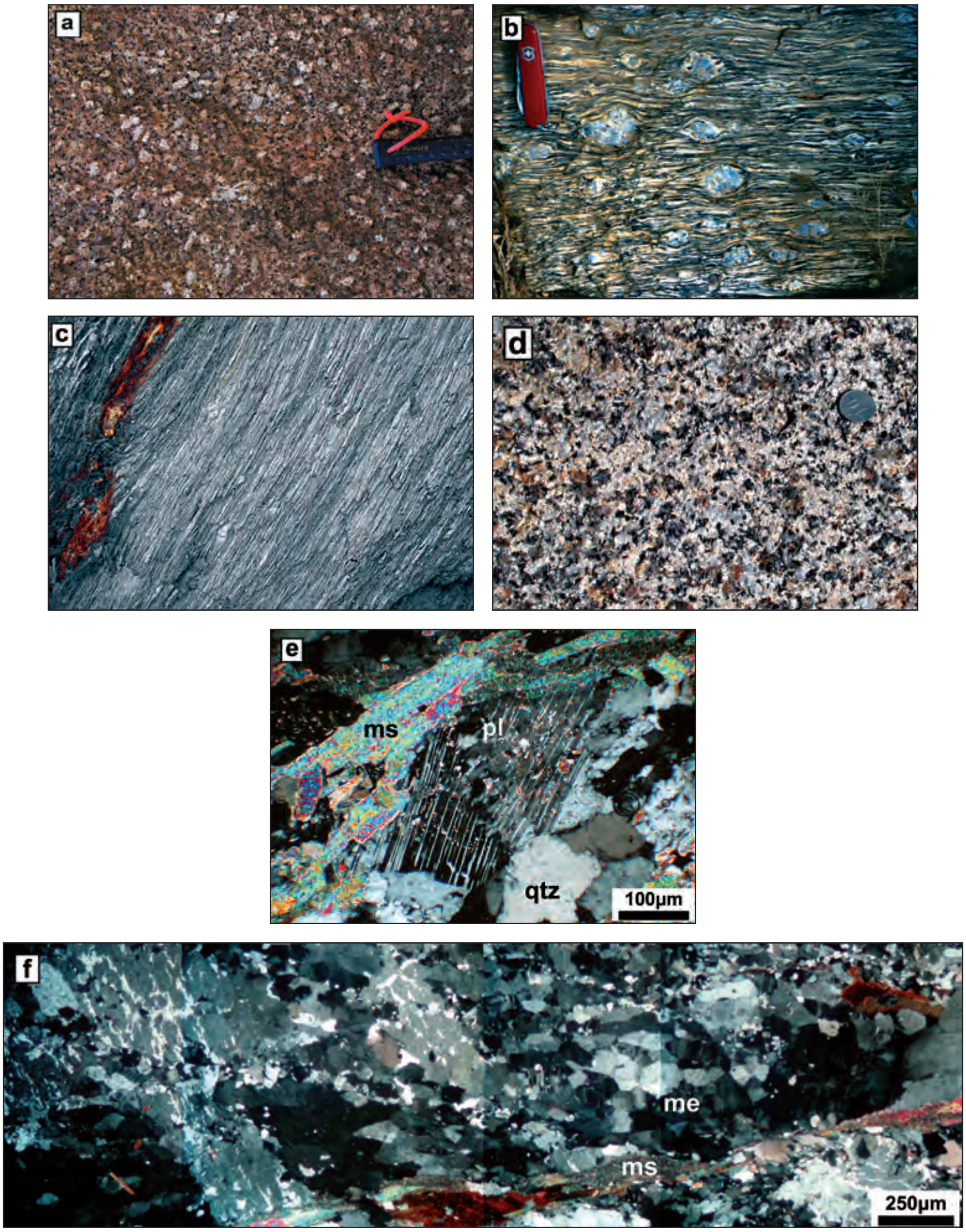

Figure 9- a) Biotite orthogneisses derived from coarse crystallined porphyritic granites where the primary texture was greatly preserved, North of Seykel village, b) the blasto mylonitic texture, product of ductile deformation which is observed in biotite orthogneisses, Dedebelen hill, c) the general view of ultra mylonitic orthogneisses formed by intense ductile deformation of biotite rich porphyritic granites, North of Katrancý, d) massif biotite orthogneisses derived from coarse grained granites with granoblastic texture characterized by equally sized, randomly distributed quartz, plagioclase, orthoclase and biotite crystals, e) cracks developed in plagioclases by deformation and new recrystallizations occurred in these cracks (+nikol), f) microcline crystals produced by the ductile deformation and developed from orthoclase porphyroblasts, ms: muscovite, pl: plagioclase, or: orthoclase, me: microcline.

deformation in rock show a distinct lineation and form foliation planes.

2. Tourmaline Leucocratic Orthogneisses.- These orthogneisses are in dirty white colors, leucocratic in character and contain abundant tourmaline (± biotite) as mafic mineral and are widely observed especially in Çine submassif (Alkanoðlu 1978; Bozkurt 2004; Dora et al., 2005; Bozkurt et al., 2006). This orthogneiss type is encountered in Ödemiþ-Kiraz submassif, at south of Kula and at west of Alaþehir. These rocks show a clear intrusive contact relationship with orthogneisses (augen/granitic gneisses) in many regions. Although these rocks are Pan-African aged, these represent a relatively young magmatic stage. The most typical property of these very light gray to white colored rocks is the content of tourmaline in high rates as mafic phase (Figure 10a). Many tourmaline formations were detected in various structures and textures (Bozkurt, 2004; Dora et al., 2006). The most typical tourmaline formations among these are "tourmaline-quartz" formations in nodular structures. In undeformed zones, the aforesaid nodules present a distinctive spherical or ellipsoidal shape. But, in regions where shear zones are effective, these gain a disc or lensoid shape making a parallel elongation to mineral lineation by lengthening and flattening. The most typical examples of tourmaline-quartz nodules are observed along the road of Çine and in Karpuzlu area. The core of nodule is composed of tourmaline and quartz at a percentage of more than 80% and the outer shell is formed by a feldspar and quartz with a thickness of 1-1.5 cm (Figure 10 b,c). The other tourmaline formation in these rocks is in the form of vein, band and amorphous masses and do not represent any lateral continuity. Besides, in this orthogneiss type, homogenously distributed tourmaline crystals in euhedral form and rosette shaped tourmaline formations are widely observed.

According to the structural and textural properties of primary granites and the size and geometry of emplaced masses, these leucocratic rocks show some differences. These tourmaline leucocratic orthogneisses are recognized as in big plutons reaching to tens of kilometers in diameter and interfingered to each other, especially in Çine submassif, between Bafa-Bozdoðan. These masses represent distinctive intrusive contact relationships both with Late Proterozoic schists belonging to core series and with biotite orthogneisses forming the country rock. Furthermore, these leucocratic granites are found in as smaller sized stocks at the south of Koçarlý and north of Karýncalý Mountain in the Çine submassif and at the south of Kula, north of Sarýgöl in the Ödemiþ-Kiraz submassif (Figure 2). These large masses in plutonic geometry might illustrate some textural differences depending on primary magmatic facial changes. Orthogneisses having granoblastic texture with coarse grained, equally sized feldspar crystals form the most widely observed leucocratic tourmaline orthogneiss type (Figure 10d). Besides, the porphyritic textured, coarsely crystallined rocks described in a groundmass of coarse orthoclase phenocrystals in gray colored and fine grained quartz-feldspar groundmass form the second common tourmaline leucocratic orthogneiss type (Figure 10 e). It is too difficult to map these granites in detailed which are mainly in leucocratic character, since intervals belonging to these magmatic facies are too narrow.

During field studies, it was observed that these aforesaid gigantic plutons found as in shapeless masses or in the form of lensoid structures with an approximate thickness of 50 to 150 meter were again cut by leucocratic rocks. These rocks are described by the existence of feldspar crystals with a size of 4-5 mm within a relatively fine grained groundmass and are named as "leucocratic metagranite porphyry" based on primary textural properties (Figure 10 f). The excess existence of tourmaline and distinctive leucocratic character shows that those rocks have a common origin with tourmaline leucocratic orthogneisses. In addition, in Çine submassif, leucocratic aplitic veins are widely recognized which cut biotite orthogneiss and tourmaline leucocratic orthogneiss plutons (Figure 10 g). These rocks are white in color,

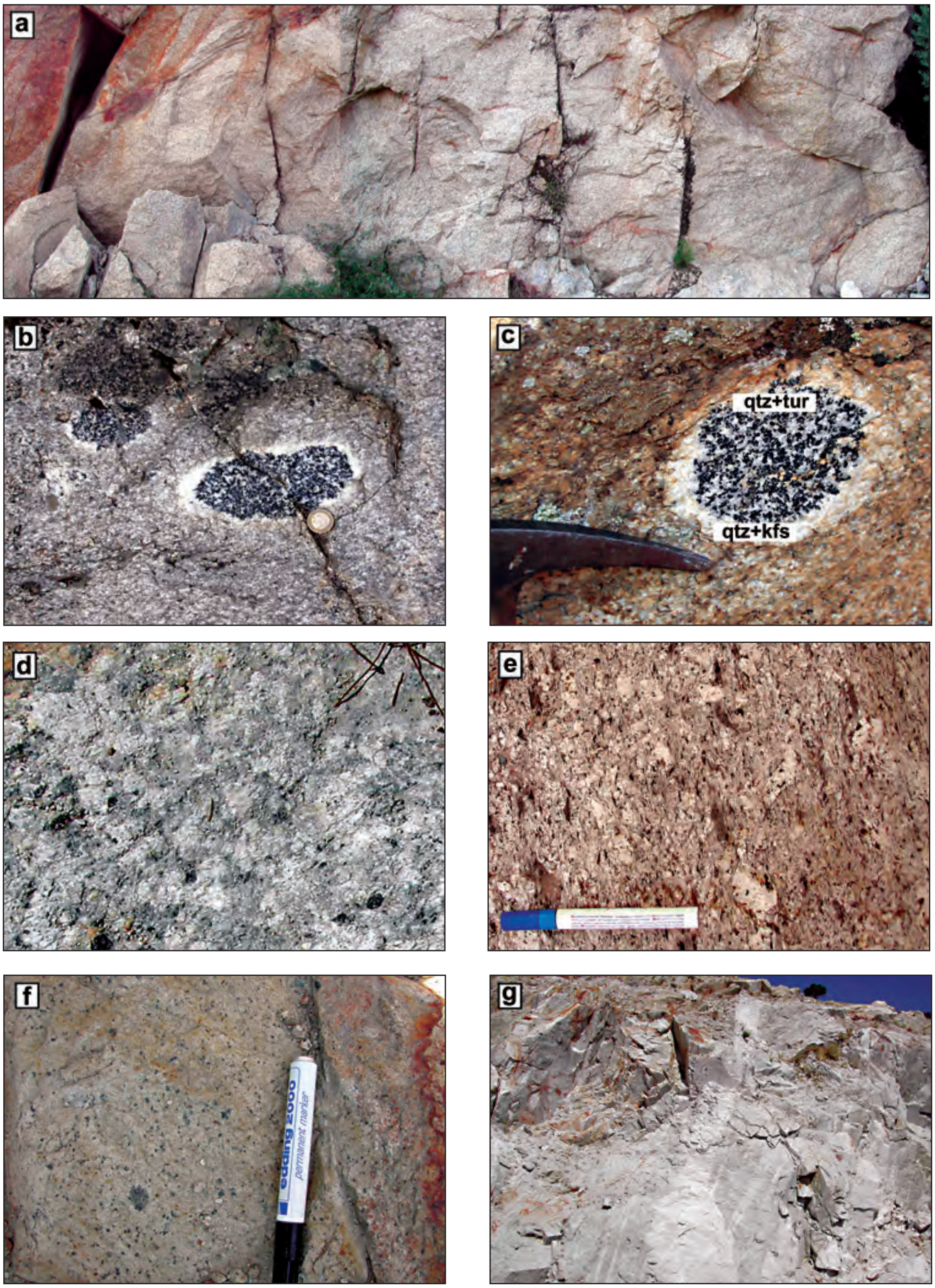

Figure 10- a) General view of tourmaline leucocratic orthogneisses described by its distinctive gray-white colors, North of Viranköy, b-c) tourmalinite nodules showing textural/mineralogical zoning, d) tourmaline leucocratic orthogneisses showing equal sized granoblastic texture, e) tourmaline leucocratic orthogneisses having porphyritic texture, f) fine grained, massive structured, tourmaline rich leucocratic porphyritic metagranites, north of Kale Hill/Mesken village, g) vein rocks fully in leucocratic character which cuts tourmaline leucocratic gneisses, Çomakdaðý village (qtz: quartz, tur: tourmaline, kfs: K-feldspar).

very fine grained which are composed of albite and quartz with more than 98% and are processed as albitic deposits. The rocks of which are the last products of leucocratic magmatic activity is characterized by; 1- the constitution of tourmalinite nodules in zonal structure, 2- fine grained formation, 3- distinctive vein characteristics, 4- the presence of rutile and/or sphene rich sections at circumferences of host rock.

Based on petrographical studies, the mineral composition of tourmaline leucocratic orthogneisses were detected as quartz, plagioclase, orthoclase, tourmaline, muscovite (± microcline, sericite, biotite, zircon, opaque). In undeformed and approximately equally sized sections of tourmaline leucocratic orthogneisses, the ratio of quartz, plagioclase and K-feldspar (orthoclase + microcline) may reach 90% that could reflect the leucocratic character of primary granite. On the other hand, textural evidences, such as the platy plagioclase traps in K-feldspars, the graphical texture formed by the growth of K-feldspar and quartz and the process of partially melted plagioclase crystals being trapped during the crystallization of primary magma by K-feldspars which has grown in the following stage are the evidences that support the magmatic origin of mentioned rocks. Quartz of which are the products of the last stage are observed in rocks equally sized, and in anhedral to subhedral crystals. Plagioclases observed in the form of various crystals anhedral to subhedral in form show widespread primary polysynthetic twinning. Orthoclase crystals showing Carlsbad twinning and/or non twinning present strip and patch type perthitifications. Besides, transformations from orthoclases into microclines are widely recognized especially in deformed samples. Involvement of individual tourmaline crystals or the tourmalinite nodules formed by the combination of all these is one of the differentiating properties of orthogneisses in study areas. Anhedral shaped, tourmaline rich core in tourmalinite nodules are surrounded by an outer shell made up of quartz, microcline and orthoclases (Figure 11 a).

In field observations and petrographical studies, it was determined that tourmaline leucocratic orthogneisses has the characteristic of ductile deformation varying between protomylonite to ultramylonite along shear zones although they mainly represent a massive and non oriented structure. Substantially, textural data belonging to each stages of this ductile deformation under amphibolitic conditions, the amphibolite facies that can be defined as the decrease in the crystal size, the recrystallization and acquisition of planar/linear structure can be observed in rocks (Figure 11 b,c,d).

General mineral composition of leucocratic metagranite porphyries is described as quartz, orthoclase, muscovite (± tourmaline, biotite, zircon). The basic property of these rocks is the availability of feldspar crystals of which its euhedral structure has been protected in a fine grained quartz-feldspar groundmass. Feldspar porphyroblasts have highly preserved their euhedral shapes and have different orientations. This shows that this texture was ordered partly by metamorphism and belongs to primary porphyritic texture related to precursor magmatic rock. Relic graphic textures belonging to primary magmatic rock are common in feldspar. Orthoclases representing Carlsbad twinning form the majority of feldspars in the rock (Figure 11 e). Sparse muscovite development is observed along the planes where there is a distinct foliation in rocks where ductile deformation took place. Tourmaline ratios reaching high values reinforce the probability of connection of aforesaid rocks with tourmaline leucocratic orthogneisses in origin.

Leucocratic metaaplites are the last products of the Pan-African magmatic activity and of rocks rich in quartz and plagioclase in vein character. In thin section studies, the mineral composition of these rocks were determined as quartz, albite, rutile (± orthoclase, sphene, zircon). Plagioclases were detected as in fine grained, equally sized crystals. These show polysynthetic twinning and albitic composition (Figure 11 f).

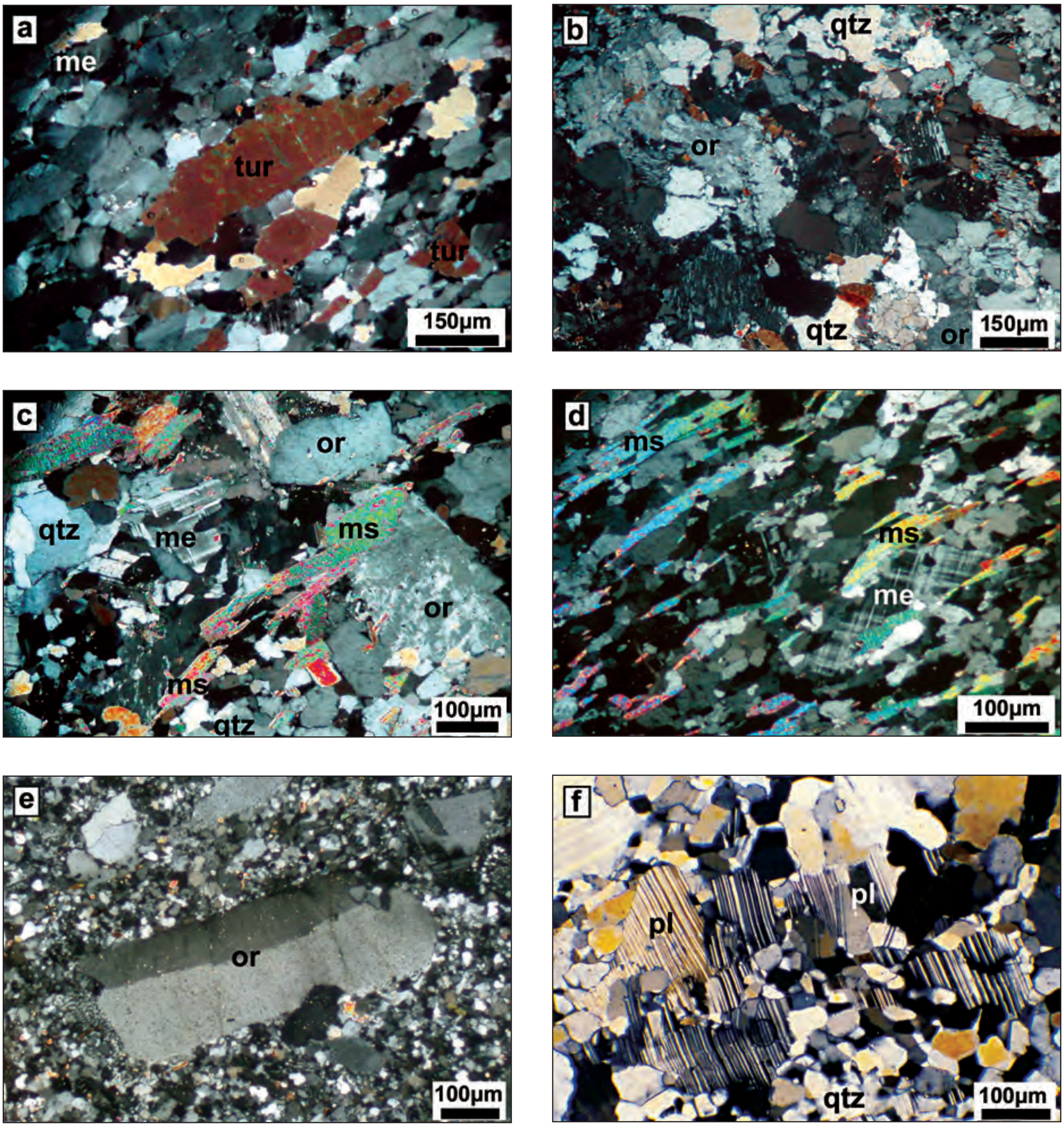

Figure 11- a) core and quartz enriched by tourmaline crystals, tourmalinite nodule made up of a leucocratic exterior zone formed by microcline and orthoclases, b,c,d) textural changes developed in tourmaline leucocratic orthogneisses depending on the intensity of increasing ductile deformation, e) primary orthoclase crystals that has mostly preserved its euhedral shape in a fine grained groundmass of leucocratic metagranitic porphyries, f) microscopic view of metaaplitic rocks composed of albite and quartz (±rutile) me: microcline, tur: tourmaline, or: orthoclase, qtz: quartz, ms: muscovite, pl: plagio-clase.

Titanium phases concentrated especially in lateral zones of aplites are made up of rutile and sphene. Rutile and sphene reaching mostly 6% are composed of homogenously distributed small crystals among quartz and albite crystals.

3. Amphibolitic orthogneisses.- These are located only in Karýncalý Mountain and around Buldan area and are described by the existence of hornblende and garnet porphyroblasts, granoblastic in texture and in fine to medium crystallized massif rocks. These masses are observed in the form of small stocks being intruded into biotite orthogneisses and schists which belong to Precambrian basement. Dark green colored, non oriented amphibole crystals with a size of 2 cm present a homogenous distribution in equally sized, granoblastic texture (Figure 12 a,b). At circumferences of the mass and internal shear zones, rocks turned into gray to silver colored ultra mylonitic gneisses that show intensive planar separation property. Partly designed texture belonging to protolithes can be noticed in zones where the intensity of deformation is much less. General mineral composition of amphibolitic orthogneisses was described as orthoclase, plagioclase, quartz, hornblende, biotite, garnet and muscovite (± tourmaline, zircon, sphene, epidote). Amphibolitic porphyroblasts characterizes this type of orthogneisses. Dark green colored amphiboles are in hornblende composition and have a poikilitic texture originated from quartz, feldspar and epidotic inclusions. Garnet porphyroblasts in these gneisses have anhedral crystal forms similar to amphiboles and in some grains, inclusion amount reaches the ratio of major minerals (Figure 12 c,d).

GEOCHRONOLOGY OF THE PAN-AFRICAN VOLCANISM

Zircon Morphology

Basic morphological properties of zircons enriched by biotite and tourmaline rich orthogneisses in the Menderes Massif, features related to internal structures and study methods applied to these are briefly summarized in this part. Koralay (2001), presented detailed descriptions related to these studies. Zircons selected under binocular microscope, were analyzed under Scanning Electron Microscope (SEM) before age determination. SEM and cathode luminesans (CL) photos of zircons were taken in order to classify according to their morphologies and to analyze internal structures. In SEM analysis, it was determined that majority of zircons differentiating from biotite orthogneisses reflected the magmatic origin and has similar morphological features. These are euhedral, sometimes asymmetrical, generally colorless, rarely in pink and brown colors transparent, in short (2:1) or long (3:1, 4:1), prismatic form and sometimes in the form of grains including inclusion and old cores of the former grains (Figure 13). According to Pupin and Turco (1974) classifications, zircons were crystallized dominantly in S6, S7, S11 and S12 types and in less amounts in L1, S2, S13 and S17 types. Depending on these dominant types, the crystallization temperature of primary magma was determined as 700-750 C (Loos and Reischmann, 1999; Koralay, 2001). Pupin (1980) stated that, this type of zircons is typical for two mica granites. Cathode-luminesans studies established that zircons in orthogneisses were magmatic in origin (Figure 14). The crystal faces in zircons which represent a typical zoning peculiar to magmatic rocks make a distinct parallelism to zoning pattern observed in internal parts of the grain. This data show that the whole grain was crystallized in similar conditions. Some zircon grains present textural data related with poly phase growth. While old dendritic cores are not recognized in long, needlelike grains (Figure 14 a,d), one or two old cores in various sizes could be detected in short and thick grains (Figure 14 e). It was also determined that these old grains show magmatic zonation. In order to determine the crystallization ages of primary granites of orthogneisses, nonbearing relic core, thin and long prismatic grains that have magmatic zoning patterns defining the crystallization from a melt

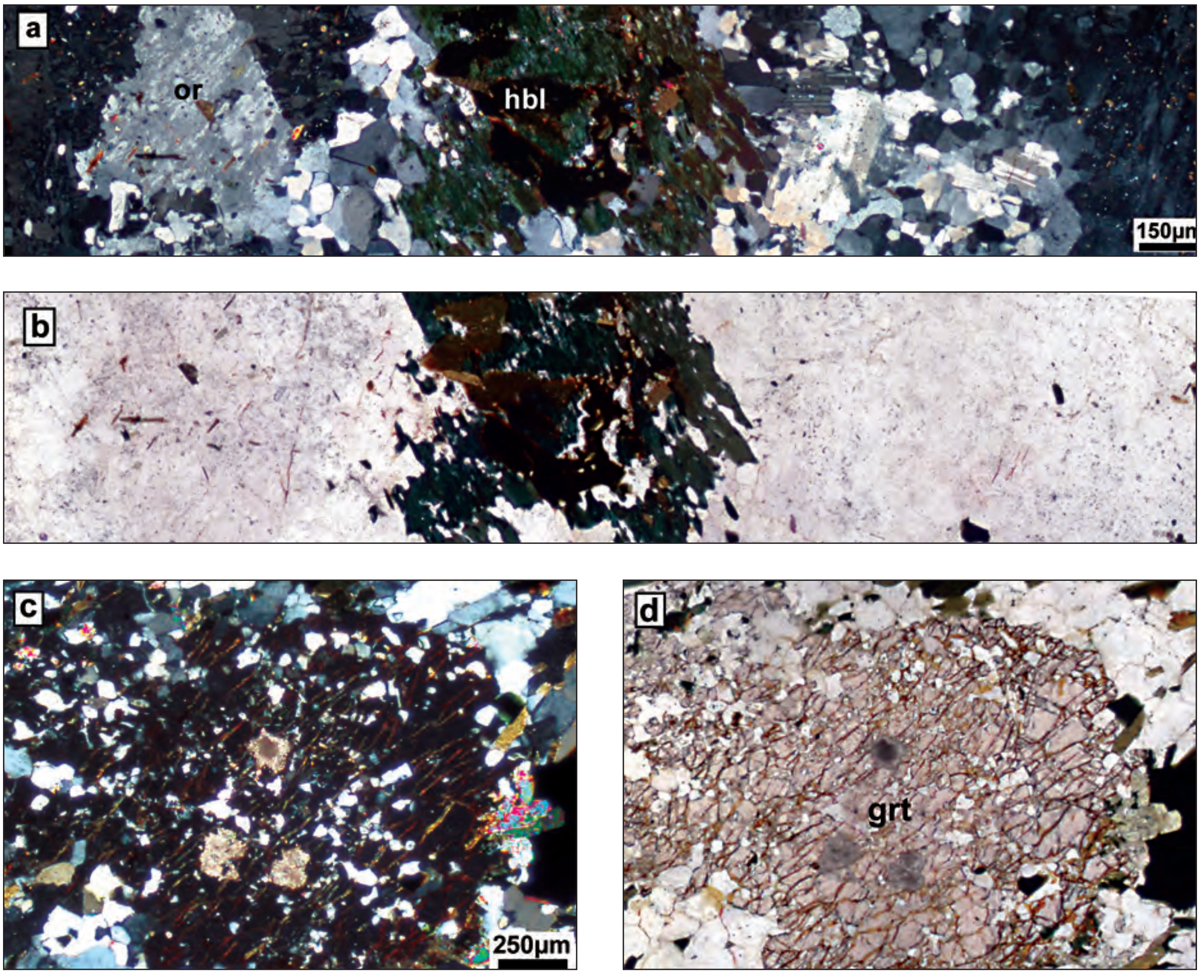

Figure 12- a-b) General view of orthoclase rich amphibole orthogneisses showing granoblastic texture, c-d) anhedral garnet crystal containing much inclusions in amphibole orthogneisses (a-c: cross nicoles, b-d: parallel nicoles) or: orthoclase, hbl: hornblende, grt: garnet.

were selected in age determinations. It is observed that zircons in tourmaline leucocratic orthogneisses are magmatic in origin, similarly.

Ages Obtained

As has been mentioned in former sections, orthogneisses were not differentiated into other types and all gneisses were defined by investigators as augen or granitic gneisses without any distinction in previous studies. Whereas, the evidences acquired in recent years have obviously established the necessity of the classification of gneisses in the Massif according to its mineral composition and rock dating studies should be set on this base. Up to now, majority of geochronological studies performed in the Massif concentrated on orthogneisses cropping out only in Çine submassif. In preliminary geochronological studies, orthogneisses in Çine submassif, south of the Menderes Massif, were dated as 490 ±90 Ma (Dora, 1975, 1976) and 470±9 Ma (Satýr and Friedrischsen, 1986) using Rb-Sr method (Table 1). After mid 90's, rock dating studies have accelerated for throughout the

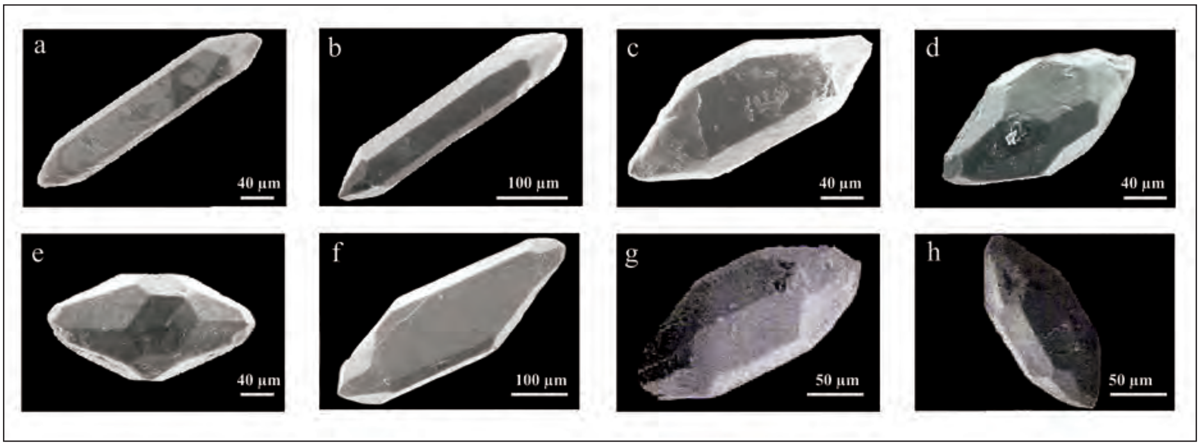

Figure 13- SEM views of typical zircons obtained from biotite orthogneisses (a-f: by Koralay et al., 2004, g-h: by Loos and Reischmann, 1999).

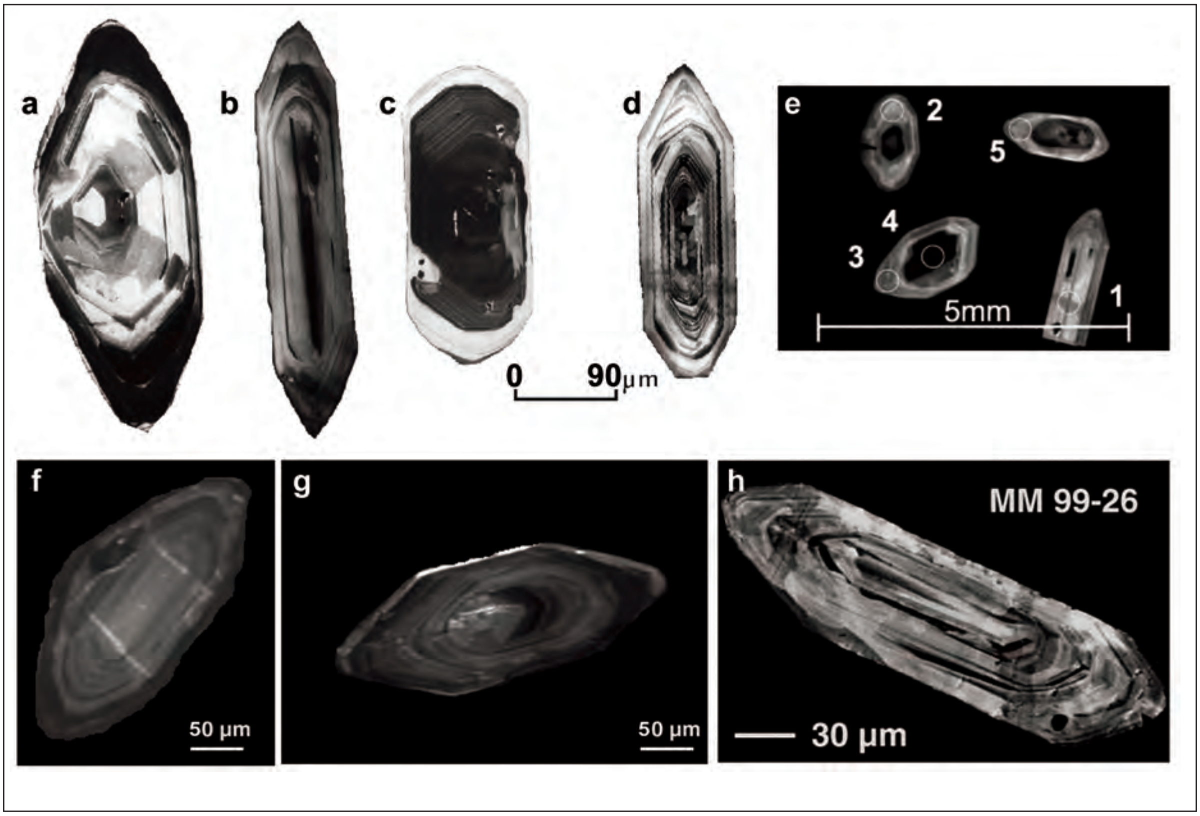

Figure 14- a-e) biotite orthogneiss and f-h) cathode luminance (CL) views of typical long prismatic magmatic zircons in tourmaline leucocratic orthogneisses (a-d-f: by Koralay et al., 2004 e: Gessner et al., 2004, h: Gessner et al., 2001).

Massif. In these studies, new methods have been used and more sensitive ages were determined conformable with geological relationships.

In this paper, it was remained true to the definition of gneisses of the authors of the previous studies. Probable equivalent names of gneisses in our newly proposed classification were given as italic in parentheses. Furthermore; orthogneisses which were dated in our studies have been named based on the criteria defined in previous sections (Table 1). As seen in Table 1 (Dannat, 1997; Dora et al., 2002; Koralay et al., 2004; Dora et al., 2005), while gneisses named as augen/granitic gneiss are equivalent of biotite rich gneisses in the new classification, in many studies (Bozkurt and Park, 1994; Bozkurt et al., 1995; Hetzel and Reishmann, 1996; Loos and Reischmann, 1999; Gessner et al., 2001; Gessner et al., 2004; Erdoðan, 1992; Erdoðan and Güngör, 2004; Bozkurt, 2004; Bozkurt et al., 2006), proposed augen/granitic gneiss and metagranite definitions are given as the equivalent of tourmaline leucocratic orthogneisses. Results of the studies performed about the orthogneiss dating in the Massif are given below.

Demirci-Gördes submassif.- In the Demirci-Gördes submassif, first studies about rock dating were established by Dannat in 1987. The investigator dated 541.4±2.5 Ma from augen gneisses (biotite orthogneiss) samples at southeast of Simav, 537±2.4 Ma at southwest of Demirci, and 544.1±4.3 Ma at the vicinity of Demirköprü dam (Figure 15, Chart 1). These ages were accepted as the intrusion age of protolithes of augen gneisses (biotite orthogneiss), by the investigator. Dora et al. (2002), dated the intrusion age of protolithes of granoblastic textured orthogneisses (biotite orthogneiss) that has intruded into paragneisses as 549.7± 7.6 Ma at south of Kula (Figure 15, Chart 1).

Ödemiþ-Kiraz submassif.- There are also rock dating studies on orthogneisses in this submassif. Augen gneiss (biotite orthogneiss) samples were taken from three different regions by Dannat (1997) and dated as 528±4.3 Ma at Northeast of Kuyucak, 528.1±1.6 Ma in the mass located at west of Buldan and 538.1±1.6 Ma in the gneissic mass located at southwest of Kiraz (Figure 15, Table 1). These ages were inferred by the author as the granitic intrusion ages forming the protolithes of augen gneisses (biotite orthogneiss). Koralay (2001) and Koralay et al. (2004) dated 561.5±0.8 Ma and 570.5±2.2 Ma for the intrusion age in augen gneisses (biotite orthogneiss) located at south of Alaþehir (Figure 15, Table 1).

Çine Submassif.- Gneisses in this submassif were geochronologically investigated in detail. Hetzel and Reischmann (1996), dated the intrusion ages of primary granites of the rocks as 546.0±1.6 and 546.4±0.8 Ma named as less deformed and exposing at north of Selimiye. These rocks were defined as leucocratic tourmaline gneisses in the same region by Dora et al. (2005). In the following years, Loos and Reischmann (1999) dated the intrusion ages of gneisses located at the south of Selimiye in a range between 521±8 - 572±7 Ma (Figure 15, Table1).

Gessner et al. (2001) stated that orthogneisses along the road between Eskiçine and Akçova (Southeast of Çine) were cut by metagranites and intrusion ages of relatively younger metagranites were dated as 547.2±1.0 Ma based on single zircon Pb/Pb evaporation method (Figure 15, Chart 1). In recent studies it was detected that metagranites which cut Augen gneisses (biotite gneisses) are equivalent to tourmaline leucocratic orthogneisses (Dora et al., 2005). In the following years, from the samples taken in the same region, by U-Pb SHRIMP method, augen gneisses (biotite gneisses) and tourmaline leucocratic orthogneisses that cut augen gneisses were dated as 566±9 Ma and 541±14 Ma, respectively (Gessner et al., 2004). Along the southern border of Çine submassif, Dora et al. (2005) stated that, field data showed biotite orthogneisses presented granites of the oldest stage and these orthogneisses were cut by the

Table 1- Geochronological and relative ages of orthogneisses suggested in previous studies which are emplaced in the Pan-African basement of the Menderes Massif.

	Location	Reference	Type of Orthogneiss		Method	Age (Ma)
			Original nomenclature	This study		
Geochronological Ages	**Demirci-Gördes Submassif**					
	SE of Simav	Dannat (1997)	Augen Gneiss	---	Pb/Pb evapor.	541.4±2.5
	SW of Demirci					537.2±2.4
	Demirköprü Dam					544.1±4.3
	S of Kula	Dora et al(2002)	Granitic Gneiss	Amphibole Gneiss	Pb/Pb evapor.	549.7±7.6
	Ödemiş-Kiraz Submassif					
	NE of Kuyucak	Dannat (1997)	Augen Gneiss	---	Pb/Pb evapor.	528.0±4.3
	W of Buldan		Augen Gneiss	---		528.1±1.6
	SD of Kiraz		Augen Gneiss	Amphibole Gneiss		538.1±2.6
	S of Alaşehir	Koralay et al (2004)	Ortognays	Amphibole Gneiss	Pb/Pb evapor.	561.5±0.8
						570.5±2.2
	Çine Submassif					
	---	Dora (1975, 1976)	---	---	Rb-Sr tot.rock	490±90
	SW of Çine	Şengör et al (1984); Satır and Friedrichsen (1986)	Metagranite	---	Rb-Sr tot.rock	470±9
	N of Selimiye	Hetzel and Reischmann (1996)	Augen Gneiss	Tourmaline leucocratic orthogneiss	Pb/Pb evapor.	546.0±1.6
						546.4±0.8
	N of Selimiyei	Loos and Reischmann (1999)	Granitic Gneiss	Tourmaline leucocratic orthogneiss	Pb/Pb evapor.	563±3, 536±9
						572±7, 521±8
						556±4, 546±5
						551±5
	SW of Çinei	Gessner et. al. (2001)	Metagranite	Tourmaline leucocratic orthogneiss	Pb/Pb evapor.	547.2±1.0
	N of Yatağan	Gessner et. al. (2004)	Metagranite	Tourmaline leucocratic orthogneiss	U-Pb SHRIMP	541±14
	SW of Çine		Metagranite	Amphibole Gneiss	U-Pb SHRIMP	566±9
	N of Yatağan	Dora et. al. (2006)	Granitic Gneiss	Amphibole Gneiss	Pb/Pb evapor.	552.1±2.4
	NW of Kavaklıdere		Porpyritic Metagranite	Leucocratic Metagranite porphyre	Pb/Pb evapor.	551.5±2.9
	N of Yatağan		Tourmaline leucocratic orthogneiss	Tourmaline leucocratic orthogneiss	Pb/Pb evapor.	545.6±2.7
	N of Yatağan		Tourmaline leucocratic orthogneiss	Tourmaline leucocratic orthogneiss	U-Pb isotop dilution	549±26
	NW of Karacasu	Koralay et. al. (2007)	Amphibole Gneiss	Amphibole Gneiss	Pb/Pb evapor.	530.9±5.3
Relative Ages	N of Selimiye	Bozkurt and Park (1994)	Augen Gneiss	Tourmaline leucocratic orthogneiss		Late Oligocene
	N of Selimiye	Bozkurt et. al. (1995)	Augen Gneiss	Tourmaline leucocratic orthogneiss		Late Oligocene / Early Miocene
	N of Yatağan	Erdoğan (1992); Erdoğan and Güngör (1994)	Augen Gneiss	Tourmaline leucocratic orthogneiss		Late Cretaceous/Early Cenozoic
	N of Yatağan	Bozkurt (2004)	Tourmaline leucocratic orthogneiss	Tourmaline leucocratic orthogneiss		Late Oligocene / Early Miocene
	S of Çine submassif	Bozkurt et al (2006)	Tourmaline leucocratic orthogneiss	Tourmaline leucocratic orthogneiss		Tertiary

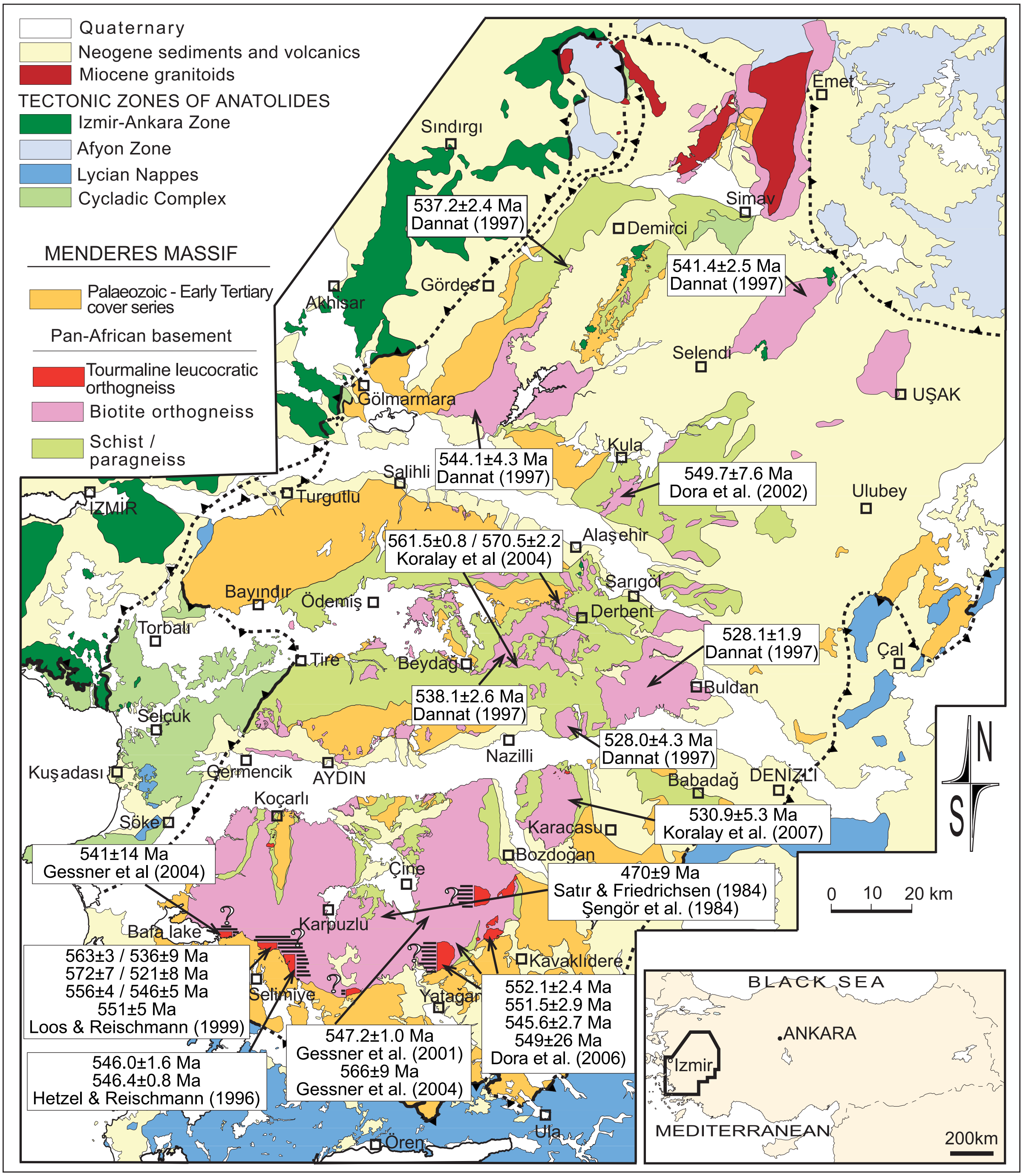

Figure 15- Location and ages of dated orthogneisses emplaced in the Pan-African basement of the Menderes Massif.

other granite types. Geochronological studies also support this fact. The intrusion ages of biotite orthogneisses located at north of Yataðan and north of Kavaklýdere were dated as 552.1±2.4 Ma and 551.5±2.9 Ma, respectively. From tourmaline leucocratic orthogneisses cutting biotite orthogneisses, at north of Yataðan, around Seykel village were dated as 549±29 and 545.6±2.7 Ma defining younger intrusion ages, using U-Pb and single zircon Pb/Pb evaporation method (Dora et al., 2005). In region located at west of Karacasu county in Çine submassif, the intrusion age of amphibolitic orthogneisses observed in small intrusions into Precambrian schists were dated as 530 .9±5.3 Ma (Koralay et al., 2007).

TRIASSIC MAGMATISM

THE DISTRIBUTION OF TRIASSIC MAGMATISM IN THE MENDERES MASSIF

The presence of Middle Triassic granites widely observed in the Menderes Massif demonstrates the second main tectonic activity in the Massif. Triassic leucocratic orthogneisses are seen in various regions along a NE-SW directed zone in the Massif (Þengör et al., 1984), in the Ödemiþ-Kiraz submassif, at south of Alaþehir around Derbent and around Demirhan village, north of Atça (Koralay et al., 2001). Besides, these granites are observed in Demirci-Gördes submassif, around Kýrcaali village, south of Kula in the form of masses in various sizes (Candan, 1994) (Figure 16).

Around Derbent, leucocratic orthogneisses are located with intrusive contacts within schists and orthogneisses belonging to the Pan-African basement that presents internal nappe structure (Figure 17). There are four tectonic slices in the region. The first three slices belonging to the Pan African basement are overlain by a Palaeozoic aged tectonical slice composed of phyllite, quartzite and marble. The first layer belonging to the Pan-African basement is completely made up of garnet micaschists. Triassic leucocratic orthogneisses located in intermediate slices make an intrusive contact mainly with Precambrian aged garnet-micaschists around Dede Mountain and Derbent (Figure 18 a). The biggest orthogneiss mass located around Sarýpýnar village at south cut Precambrian schists and 560-570 Ma aged (Koralay et al., 2004) biotite orthogneisses. In Derbent region, leucocratic orthogneisses located at west of Karacaali village contacts with marble lens, probably in Palaeozoic age. This zone presents distinct signatures of contact metamorphism and has a mineral assemblage of wollastonite, grossular, vesuvianite and epidote (Koralay, 2001). Similar leucocratic orthogneisses in different sizes and its metaaplites are recognized at the north of Alaþehir, the region between Balcýlar and Kýrcalý villages (Candan, 1994). Triassic leucocratic orthogneisses observed in this region were intruded into schists of basement series and phyllites of Palaeozoic cover series with intrusive contacts (Figure 16). Especially, in marble intercalations the scarn zones were developed made up of epidote and amphibole. The contact metamorphic effects in phyllites are defined by the formations of massif garnet.

Triassic leucocratic orthogneisses are observed around Demirhan village at north of Atça (Aydýn) (Figure 19), in the form of small masses within Permo Carboniferous cover series (Koralay et al., 2001). Numerous aplitic veins are also recognized within a couple of meters in thickness related with this magmatism around leucocratic orthogneisses reaching a size of 600 m Dannat (1997), mentions about the existence of rocks in similar ages at north and south of Simav in the Demirci-Gördes submassif. In the investigation made recently, new leucocratic orthogneiss exposures were detected in Palaeozoic cover series in small sized sills in the Kiraz submassif. These orthogneisses are observed around Bayramlýk village at southeast of Kemalpaþa and 4 km to the south of Allahdiyen village, south of Salihli (Figure 18b).

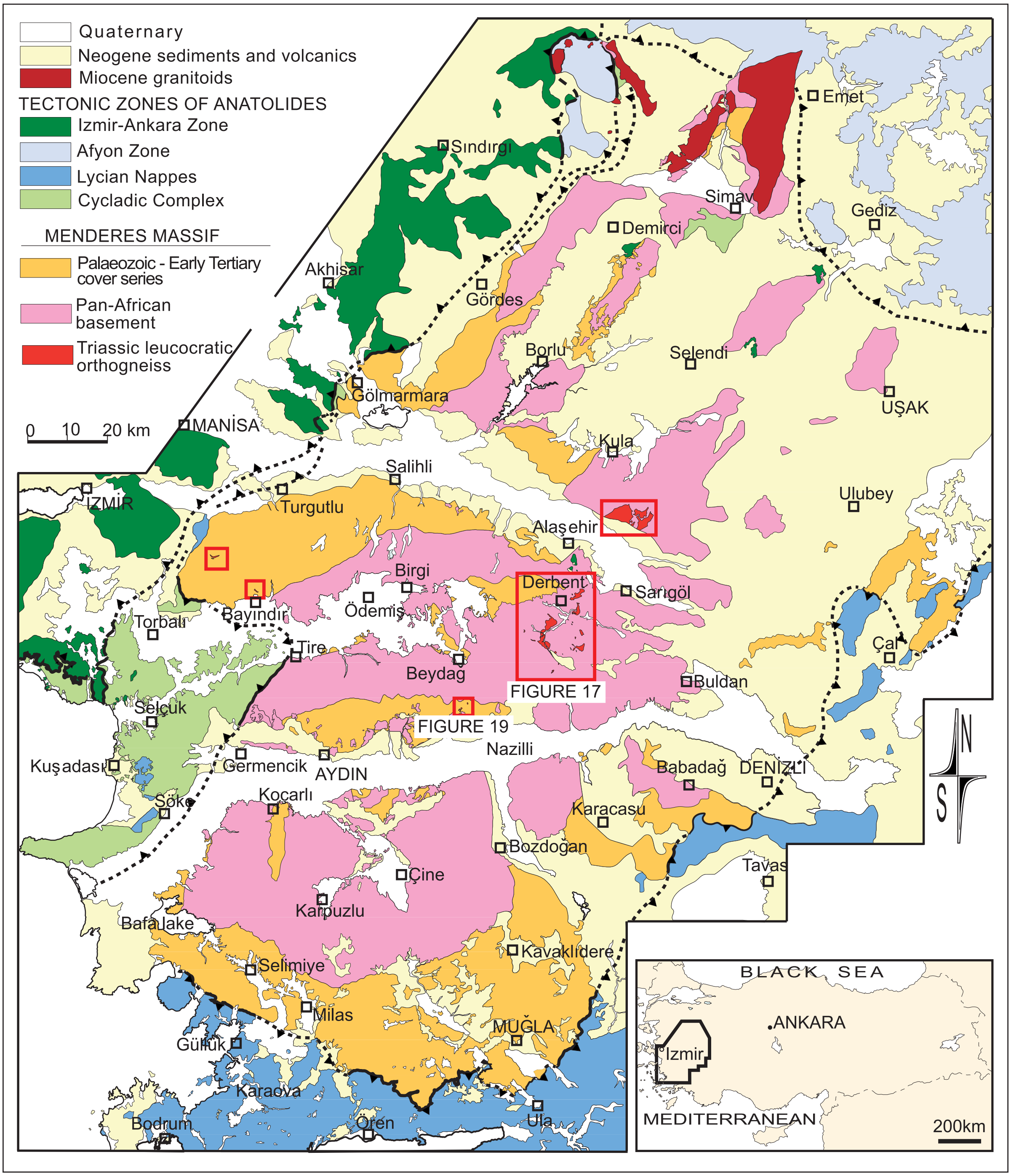

Figure 16- The distribution of Triassic leucocratic orthogneisses in the Menderes Massif.

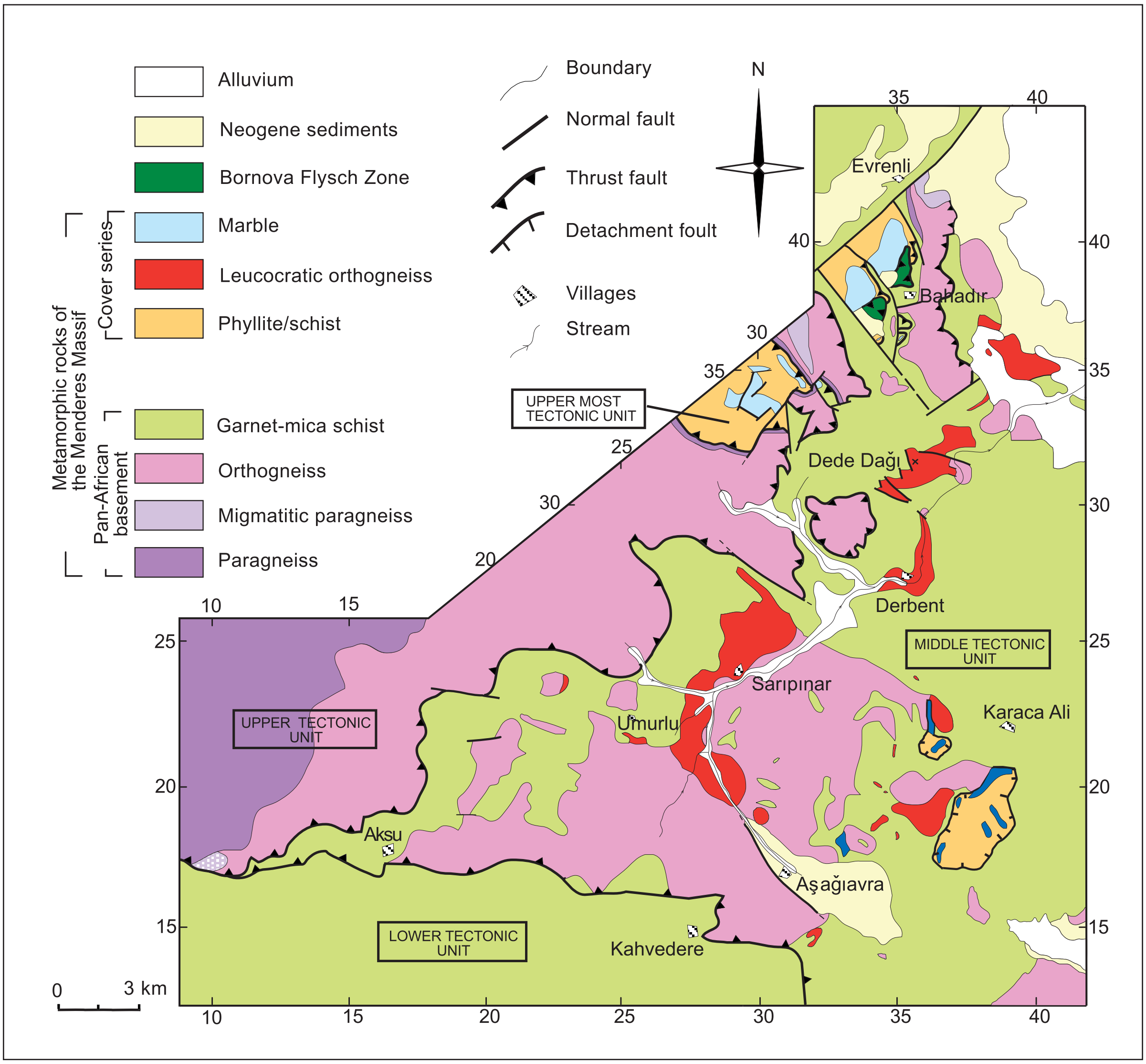

Figure 17- The geological map of Derbent region located in western part of Ödemiþ-Kiraz submassif (by Koralay et al., 2001).

MACROSCOPIC AND MICROSCOPIC FEATURES OF LEUCOCRATIC ORTHOGNEISSES

Triassic leucocratic orthogneisses are white to dirty white and greenish white in color, equally sized, fine grained and have a distinct foliation (Figure 20a). Massif structure is observed in undeformed sections. The rock becomes proportionally more in greenish color as the muscovite rate increases. Magnetite bearing orthogneiss types on the other hand are in grayish color (Figure 20b). Muscovites reaching 10-15% in amount determines the foliation planes in the rock. The quartz, plagioclase and orthoclase rates indicating the primary granitic composition reach the percentage of 88-95%.

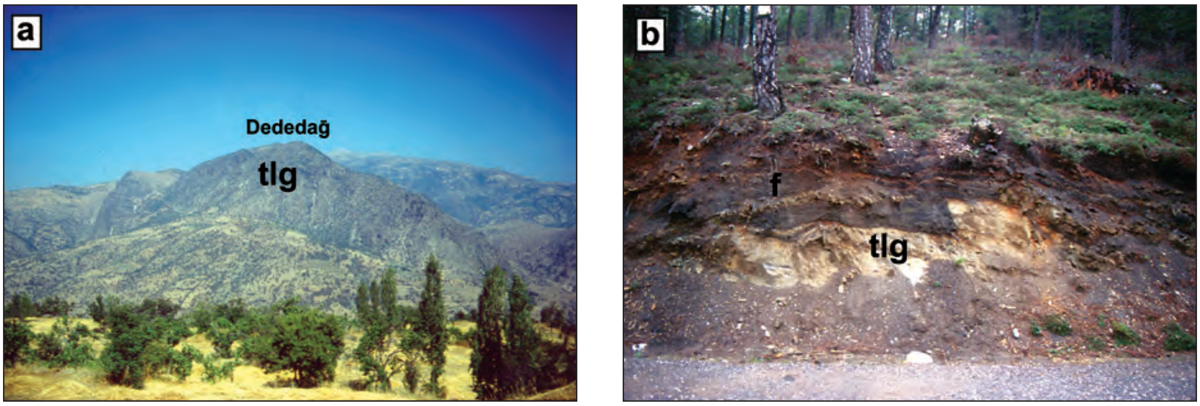

Figure 18- a) Dededað intrusive made up of Triassic leucocratic orthogneisses that cuts Precambrian schists, south of Derbent, b) Triassic leucocratic orthogneiss vein in Palaeozoic phyllites, Bayramlýk village/Kemalpaþa, (tlg: Triassic leucocratic orthogneiss, f: phyllite).

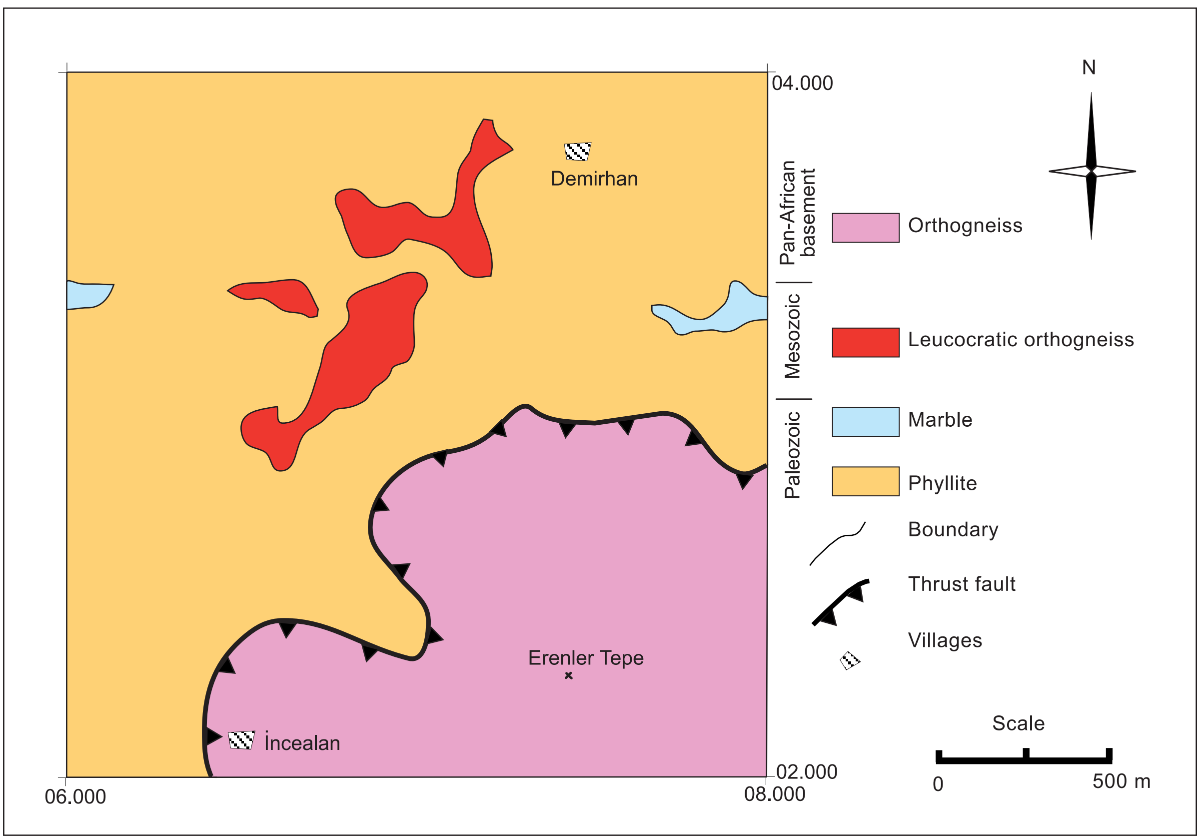

Figure 19- The geological map of Triassic leucocratic orthogneisses located around Demirhan Village/Aydýn, south of Ödemiþ-Kiraz submassif (Koralay et al., 2001).

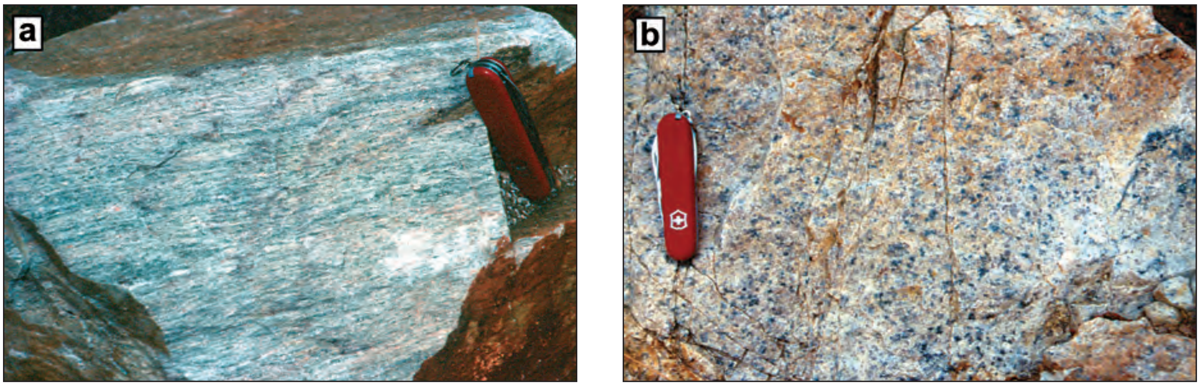

Figure 20- a) Triassic biotite-muscovite leucocratic orthogneisses showing distinct orientation, b) massif structured magnetite leucocratic orthogneisses.

According to mineralogical compositions, Triassic leucocratic orthogneisses are divided into three different groups. These are; i) Biotite-muscovite Leucocratic orthogneiss, ii) Magnetite leucocratic orthogneiss and iii) Biotite leucocratic orthogneiss. The general mineral composition of leucocratic orthogneisses is made up of quartz, orthoclase, plagioclase and muscovite (± biotite, tourmaline, garnet, apatite, zircon, magnetite). Rate of quartz, orthoclase and plagioclase are 42-54%, 10-38% and 6-15%, respectively. K-feldspars (orthoclase and microcline) which is the main component of the rock, is observed in subhedral to anhedral forms and grains reach 6-7 mm in size. Depending on the intensity of deformation in rocks under ductile conditions, crystal sizes decrease and strongly foliated, mylonitic leucocratic orthogneisses are recognized (Figure 21 a,d). Muscovites in masses located at north of Derbent and Alaþehir show a pale green pleochroism at 21%, whereas this decreases below 10% around Demirhan. Magnetite percentage in magnetite leucocratic orthogneisses can even reach up to 15%. Magnetite crystals are observed individually or as in accumulations (Figure 21 e,f).

GEOCHRONOLOGY OF TRIASSIC MAGMATISM

Morphology of Zircon

Zircons differentiating from orthogneisses were examined under SEM before the radiometric rock dating studies. In order to classify zircons according to its morphology and analyze their internal structures SEM and CL photos were taken. As a result of the SEM analysis, it was determined that most of zircons have a similar morphology reflecting the magmatic origin. These are euhedral, generally non transparent, bearing old cores and in the form of grains that have cavities on its plane (Figure 22). Zircons, according to the classification of Pupin and Turco (1974), were crystallized dominantly in types of P3, P4, P5 type and less dominantly in P1, P2, D, S7, S17 types. Based on these dominant types, the crystallization temperature is given as 750-850 C (Koralay, 2001). According to Pupin (1980), zircons in this type are observed as granites in alkaline origin derived from hybrid magmas in subvolcanic and anorogenic complexes.

In CL analysis made, zircons that present a typical magmatic zonation indicates the magmatic origin of these rocks (Figure 22). Old dendritic cores are not observed as in long, spindle shaped grains and in some grains old cores subjected to erosion which reflects the melting from a source rock of sedimentary in origin are recognized. For rock dating long prismatic grains bearing no old cores were selected.

Ages obtained

Although there is not any geochronological data, based on the regional geology, the age of

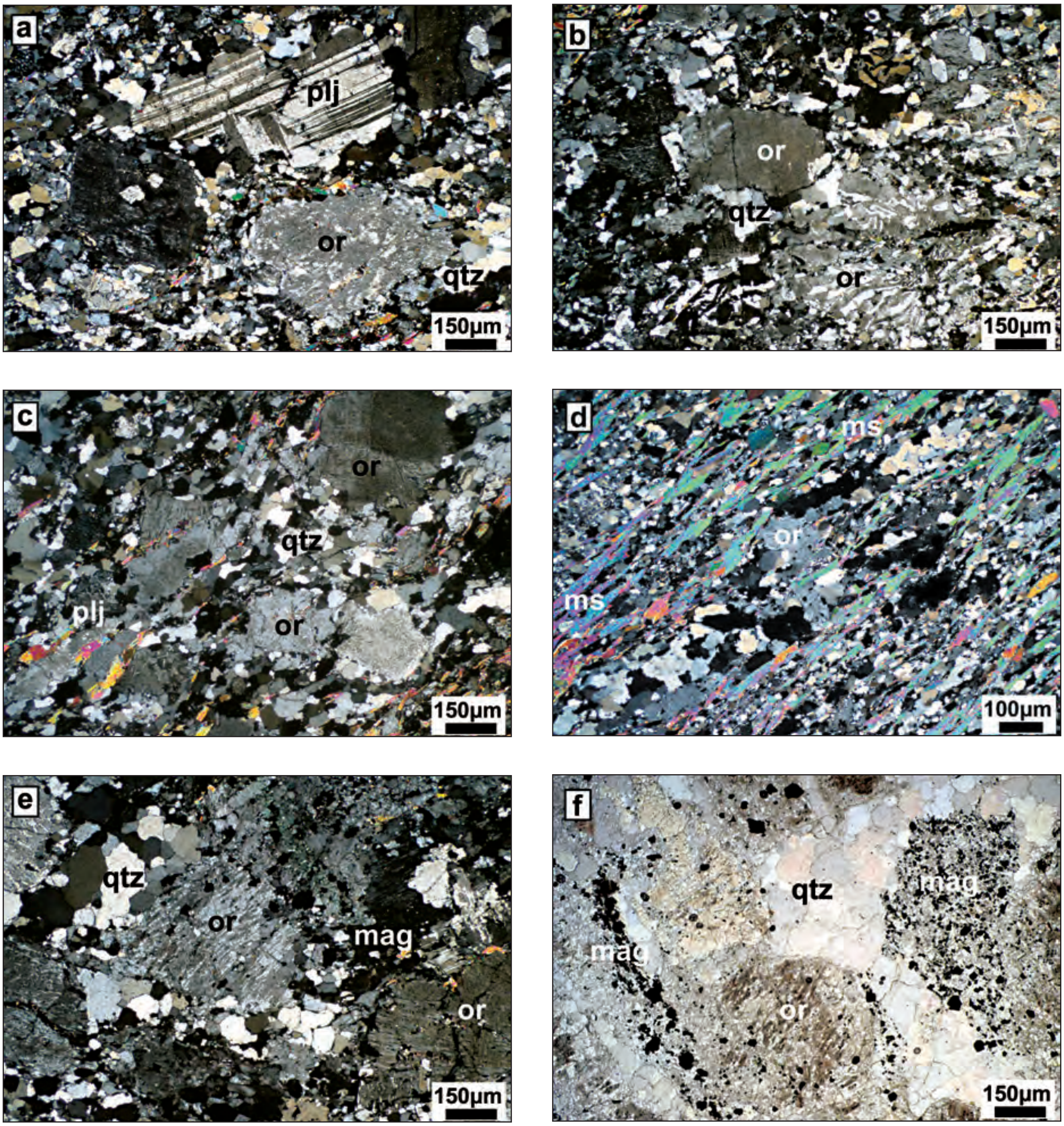

Figure 21- a-d) Textural variations developed in orthogneisses based on the intensity of increasing ductile deformation, e-f) single magnetite crystals and magnetite accretions in magnetite leucocratic orthogneisses, (e: cross nicole, f: parallel Nicole, pl: plagioclase, or: orthoclase, qtz: quartz, ms: muscovite, mag: magnetite).

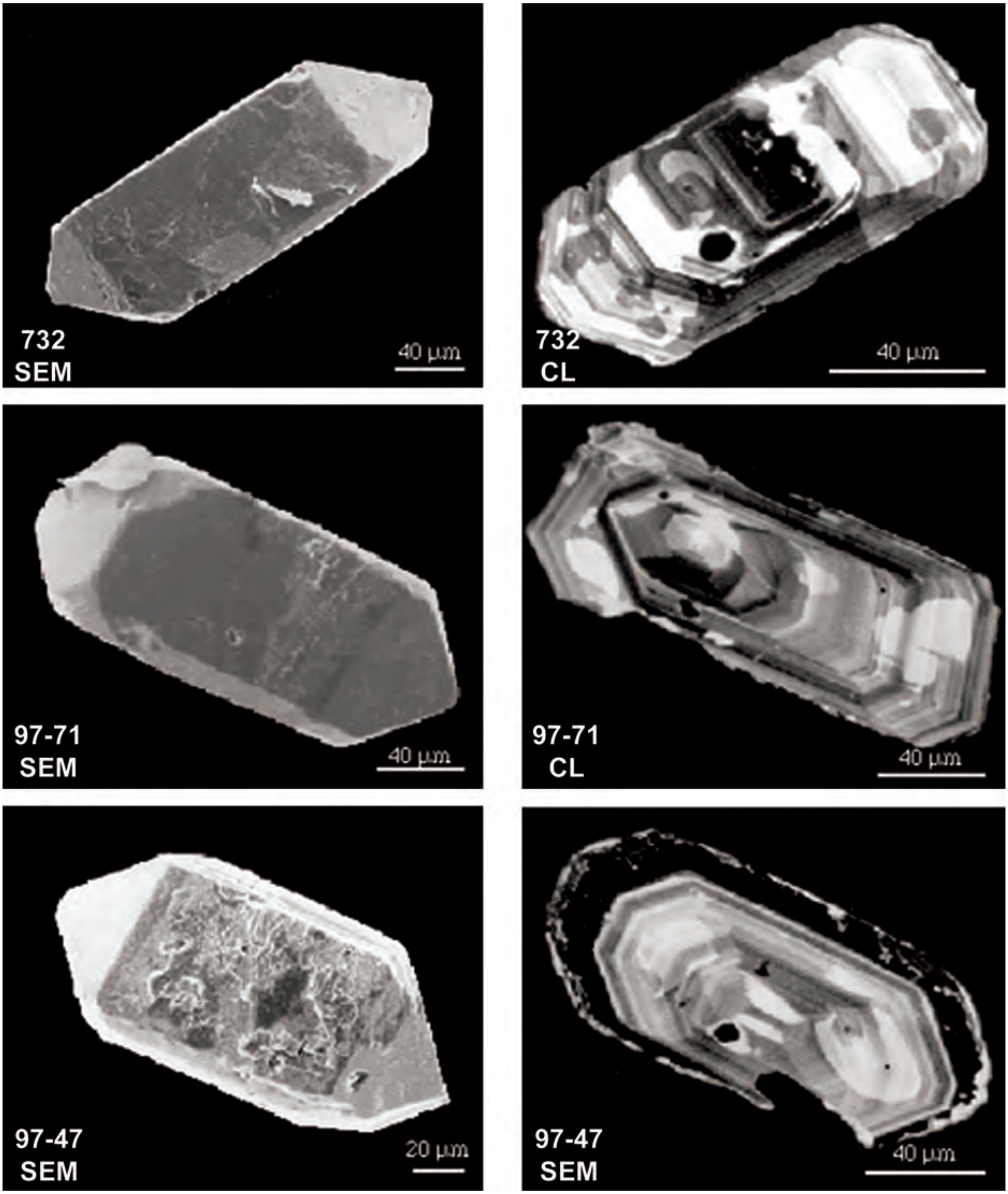

Figure 22- SEM and CL views of zircons belong to Triassic leucocratic orthogneisses (Koralay et al., 2001).

leucocratic orthogneisses until the end of 90's were accepted as Late Triassic and this age was associated with the closure of Karakaya Ocean in Late Triassic (Akkök, 1983, Þengör et al., 1984). After the widespread existence of leucoratic orthogneisses were recognized in the Menderes massif, geological, geochronological and geochemical studies were intensified onto these rocks. Dannat (1997) and Dannat and Reischmann (1998) dated leucocratic orthogneisses in Ödemiþ-Kiraz and Gördes submassifs ranging in between 227-240 Ma (Table 2, Figure 22). These ages were interpreted by the authors as the intrusion age of protolithes of orthogneisses. In Ödemiþ-Kiraz submassif, leucocratic orthogneisses that cut Precambrian schists around Derbent region were dated and 245.7±4.6 and 241.1±5.2 Ma intrusion ages were obtained by single Zircon Pb/Pb evaporation method. Same investigators determined an intrusion age for the leucocratic orthogneisses as 234.9±5.8 Ma. that cut Palaeozoic cover series located around Demirhan village, north of Atça (Table 2, Figure 23).

DISCUSSION

There have been many opinions about the age and tectonical environments of base rocks for gneisses belonging to two different magmatic activities which are widely observed in the Pan-African basement and Palaeozoic cover series in the Menderes Massif. Main ideas related to these are discussed within the framework of previous evidences and personal data of the author of this paper.

PAN-AFRICAN MAGMATISM

Protolithes of Orthogneisses

Protolithic rocks of the Pan-African orthogneisses forming the oldest magmatic activity in the Menderes Massif have always been controversial so far. Although not expressed exactly, opinions on protolithes of gneisses are grouped in two topics. These are;

i) Sedimentary origin.- Schuilling (1962) stated that orthogneisses were defined as migmatite in general and these were derived from sedimentary rocks as a result of high grade metamorphism by the study made in Çine submassif. Öztürk and Koçyiðit (1983), Akkök et al. (1984), Þengör et al. (1984) and Satýr and Friedrichsen (1986) explain that great majority of orthogneisses in the Menderes Massif originated from clastic sedimentary rocks and gained migmatitic character by partial melting. This opinion was widely supported in many of the studies performed in 70's and 80's, as well (Akat et al., 1975; Akdeniz and Konak, 1979; Akkök, 1983). Baþarýr (1970) claims that rocks defined

Table 2- Radiometric ages obtained from Triassic leucocratic orthogneisses in the Menderes Massif.

	Method	Age (Ma)	Reference
S of Simav NE of Simav	Single Zircon Pb/Pb evaporation	230-240	Dannat (1997) Dannat and Reischmann (1998)
SW of Derbent N of Alaşehit	Single Zircon Pb/Pb evaporation	227-240	Dannat (1997) Dannat and Reischmann (1998)
SW of Derbent N of Derbent (Dededağı) N of Atça (Demirhan Village)	Single Zircon Pb/Pb evaporation	245.7±4.6 241.1±5.2 234.9±5.8	Koralay (2001) Koralay et. al. (2001)

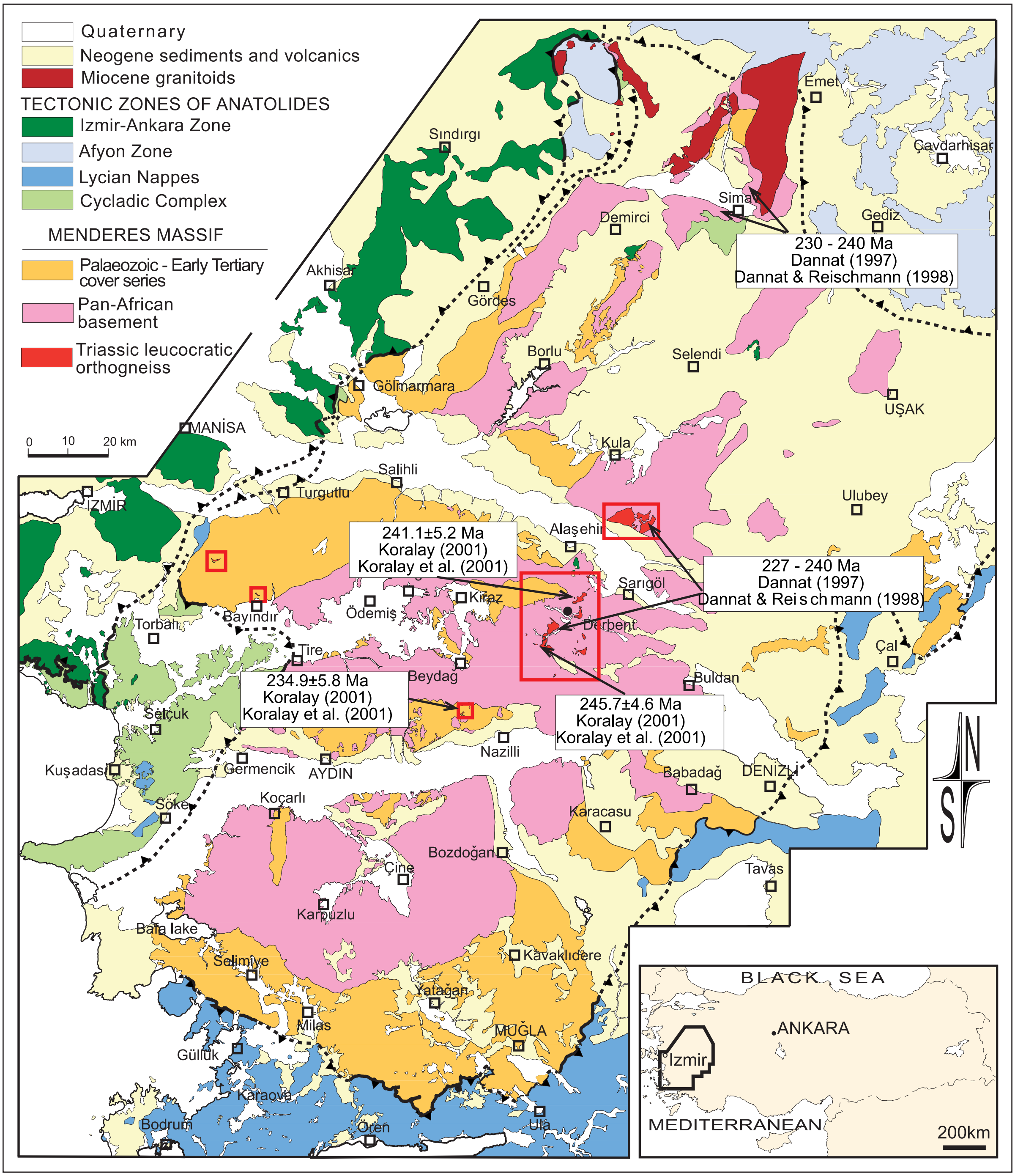

Figure 23- Locations and ages of dated Triassic leucocratic orthogneisses within the Pan-African basement and Palaeozoic cover series in the Menderes Massif.

as in fine grained in Bafa Lake are the sedimentary in origin. This opinion was also supported by Dora et al. (1990) in a study about the evolution of the Menderes Massif.

ii) Magmatic origin.- As an alternative to sedimentary origin, many investigators suggested that gneisses were magmatic in origin (Graciansky, 1965; Baþarýr, 1970; Konak, 1985; Konak et al, 1987). Great majority of studies made in the Menderes Massif after 1990 has shown that, there has been extremely strong geological, geochemical and geochronological evidences that all gneisses derived from magmatic origin contrarily to the sedimentary origin (Erdoðan, 1992, 1993; Bozkurt et al., 1992, 1993, 1995; Bozkurt, 1994; Bozkurt and Park,1994; Loos, 1995; Dora et al., 1995; Hetzel and Reischmann, 1996; Dannat, 1997; Dannat and Reischmann, 1998; Loos and Reischmann, 1999; Bozkurt, 2004; Gessner et al., 2001, 2004; Erdoðan and Güngör, 1992, 2004; Koralay et al., 2004; Bozkurt et al., 2006). Dora et al. (1994, 2005), suggested that protolithes of orthogneisses with different structures are the granites that have settled down related to Pan-African Orogeny and syn to post-metamorphics. Similar opinion is accepted by many researchers (Hetzel ve Reischmann, 1996; Loos and Reischmann, 1999; Dannat, 1997; Hetzel et al., 1998; Gessner et al., 2004; Koralay et al., 2004). On the other hand, geochemical data indicate that these granites are in calc-alkaline origin, per-aluminuous in character, S-typed and as syn to post-tectonic granite/granodiorite (Bozkurt et al., 1992, 1993, 1995; Dannat, 1997; Koralay and Dora, 1999; Koralay et al., 2004; Bozkurt et al., 2006). In addition to these, morphologies of zircons related to orthogneisses which are investigated intensely in recent years and internal zoning patterns clearly display the magmatic origins of these rocks (Hetzel and Reischmann, 1996; Loos and Reischmann, 1999; Dannat, 1997; Gessner et al., 2004; Koralay et al., 2004; Dora et al., 2005).

As a result, contemporary field data and petrographical, geochemical and geochronological evidences clearly display that protolithes of Pan-African gneisses in the Menderes Massif are deep rocks in character with granitic composition.

Intrusion Ages of Primary Granites of Orthogneisses

When papers published in recent years have been overviewed, it has been clearly observed that big differences in opinions about the intrusive/crystallization ages of the primary granites of orthogneisses had been observed. Opinions about these can be gathered in two main topics. In the first opinion, orthogneisses in the massif is defined as young intrusives (Tertiary/Late Oligocene: Bozkurt et al., 1992, 1993, 1995; Mittwede et al., 1995*a*, 1995*b*, 1997; Bozkurt and Park, 1994, 1997*a*, 1997*b*, 2001; Cretaceous/ Early Senozoic: Erdoðan, 1992, 1993; Erdoðan and Güngör, 2004). Substantially, this opinion is put down to regional geology, field and kinematical data. It is emphasized that zircon ages obtained so far presents a large spectrum. It is also interpreted that these zircons are well preserved in melt and clastics found in sediments where the granites originated from belong to relics grains of Zircon. Although there are some differences in opinions in detail, the judgement which brought investigators to this result mainly depends on five field observations. These are; 1) metaclastic and metacarbonate deposit in the Menderes Massif is Palaeozoic-Early Tertiary, 2) this deposit is continuous and does not have any structural or stratigraphical discontinuity, 3) the primary granites of orthogneisses cut this continuous series extending until Early Tertiary by intrusive contacts, 4) kinematical data related to gneiss and country rocks show that these granites are post metamorphic intrusives according to Alpine metamorphism, 5) transformation of granites into gneiss is related with the cropping out of the Massif in Late Oligocene. Depending on this argument, investigators interprete orthogneisses substantially as the crustal thickening in last stage of Alpine metamorphism and formed by the reason of young acidic intrusives formed by partial melting of high grade clastic sediments.

And in the second opinion, it is suggested that orthogneisses are the granites that are related with the Pan-African Orogeny and intruded into Late Proterozoic gneisses and schists in Late Precambrian-Early Cambrian (Þengör et al., 1984; Satýr and Friedrichsen, 1986; Dora et al., 1990; Hetzel and Reischmann, 1996; Danat, 1997; Dannat and Reischmann, 1997; Hetzel et al., 1998; Loos and Reischmann, 1999; Gessner et al., 2001; Okay, 2001; Dora et al., 2002; Gessner et al., 2004; Koralay et al., 2004; Dora et al., 2005). This opinion of the investigators substantially relies upon; 1) the tectonical relationships of Palaeozoic units that belong to cover series of orthogneisses with metaclastics reaching 6 km. in thickness and included into Pan-African basement, and 2) radiometric data obtained from Palaeozoic metaclastic belonging to cover series, metaclastic series forming its the country rocks and orthogneiss.

Country rock.- As previously explained above, the country rocks of the orthogneisses in the Menderes Massif are made up homogenous metaclastics and does not contain carbonaceous levels. In studies performed at the north of Alaþehir around Birgi-Kiraz, it was detected that related metaclastic deposits were formed by paragneiss derived by litarenitic sandstones at the bottom and schist units originated by the intercalation of sub arkosic sandstone-mudstone (Dora et al., 2001, 2002). This clastic series reach a thickness of 6 km, and cut by biotite orthogneisses with a clear contact relationship (Candan, 1994; Dora et al., 2002; Candan et al., 2010). The depositional age of primary sediments of this clastic series ranges between 550-600 Ma (Late Proterozoic) based on the youngest zircon ages in metaclastic series and Pb-Pb zircon evaporation ages obtained by those granites (Dora et al., 2002; Koralay et al., 2005). Petrographical data and field studies show that these metaclastics are widely observed in the Massif. For example, at south of Çine submassif, schists that have intruded into orthogneisses along 100 km long gneiss boundary can be associated with the same schists belonging to Late Proterozoic metaclastic series. Besides, the ages of the clastic zircon obtained from metaclastics extending along this boundary (the youngest clastic zircon age 594 Ma, Dora et al., 2005) and the cutting relationships of orthogneisses clearly show that related rocks are Late Proterozoic schists belonging to the Pan-African basement. However, there is no strong evidence at any place in the Menderes Massif that an orthogneiss extists (except the Triassic leucocratic orthogneisses) which cuts a metaclastic in Palaeozoic age by means of fossil dating or other data. On the contrary, 550-560 Ma clastic zircon ages were obtained showing that source rock is orthogneiss using Palaeozoic aged (?) metaclastics of cover series (Loos, 1995; Dora et al., 2005). On the other hand, Palaeozoic cover series related to this magmatic activity to be sterilized makes an age limitation for intrusives of the primary granites of orthogneisses.

Radiometric data.- Radiometric ages obtained from the Menderes Massif so far are given in table 1. As can be obviously seen, zircons were dated between 570-520 Ma, averaging at 550 Ma. All investigators state that zircons dated describe the crystallization from a melt based on the morphology and zonation pattern and these ages obtained are interpreted as the crystallization/intrusion ages of the primary granites of orthogneisses. According to investigators, ages observed as spread originates from the constitution relic core belonging to sedimentary base rock of some zircon grains that were dated. As clearly emphasized in previous parts of the paper, there are some confusion in the classification and nomenclature of orthogneisses in the Massif. However, the field evidences and petrographical data obtained in recent years indicate that biotite, tourmaline and amphibole orthogneisses based on mineralogical compositions in the Massif can be divided in to three main groups based on mineralogical compositions. Field data and contact relationships show that these gneiss types based on the relative aging can be ordered from oldest to youngest as biotite orthogneisses, tourmaline leucocratic orthogneisses and amphibole orthogneisses. In table 1, although there have been some problems in the nomenclature of samples dated in previous studies, it is observed that radiometric dating values substantially show conformity with this relative relationship. The intrusion ages of biotite orthogneisses show variation between 550-570 Ma, in general (Dora et al., 2002; Koralay et al., 2004; Gessner et al., 2004; Dora et al., 2005). Although tourmaline leucocratic orthogneisses give an age similar to that of biotite orthogneisses throughout the Massif, relatively much younger ages ranging between 541-547 Ma were dated in the sections where clear contact relationships were observed on the field (Hetzel and Reischmann, 1996; Gessner et al., 2001, 2004; Dora et al., 2005). Amphibole orthogneisses were dated as 531 Ma (Koralay et al., 2007) pointing that this granite type indicates a younger magmatic phase. When the information given above is evaluated together, it is noticed that radiometric datings show a general consistency and fits with field data.

As a result, when all geochronological ages, contact relationships and petrographical/mineralogical data are assessed together, it is understood that the metaclastics in which the orthogneisses intrude are Late Proterozoic units that belong to the Pan-African basement. It is also seen that these data define the differentiation product intrusives belonging to stages of the same magmatic activity following one another related with the Pan-African Orogeny of orthogneisses derived from granites of different characters.

TECTONIC ENVIRONMENTS OF THE PRIMARY GRANITES OF ORTHOGNEISSES

Data given above, obviously reveal the presence of a magmatic activity mostly varying between 520-570 Ma, with an approximate date of 550 Ma, throughout Menderes Massif. The other magmatic activities with similar ages in Turkey (except the Menderes Massif) were determined as; 1- Istanbul Zone (Chen et al., 2002; Ustaömer et al., 2005), 2-Sandýklý region (Kröner and Þengör, 1990; Gürsu et al., 2004), 3-Afyon Zone (Gürsu et al., 2005) and 4-Bitlis Massif (Ustaömer et al., 2009) (Figure 23). Besides, the diffuse existence of similar aged magmatic activities in Europe (Neubauer, 2002) and in Africa are clearly observed in regional scale. The magmatic activity in related age range is assocciated with two main orogenies in regional scale. These are; 1- the Cadomian Orogeny and 2- the Pan-African Orogeny. Thoughts related to magmatic activity with an age of approximately 550 Ma in the regions given above and the probable location of magmatic activity in the Menderes Massif under this tectonism is discussed below,

THE MAGMATISM ASSOCCIATED WITH CADOMIAN OROGENY

Around Karadere, located at east of Istanbul Zone, the intrusion ages of granites settling in base rocks were dated varying between 560-590 Ma (Chen et al., 2002) (Figure 24). Investigators, suggest that according to low $^{87}Sr/^{86}Sr$ rates and high ε Nd values, these granites are the product of an arc magmatism formed by the subduction of an old ocean (Iapetus?) under Gondwanaland. In Bolu Massif, the east of this region, 576±6 and 565±2 Ma. ages were dated for Kapýkaya and Tülükiriþ plutonics, respectively (Ustaömer et al., 2005) (Figure 24). Geochemical data reveal that these calc-alcaline and I-type granitoids are the products of Andes type arc magmatism. In recent years, it has been suggested that Palaeozoic basement of West Pontides was the continuity of Cadomian belt observed in West Europe based on these dating values and geochemical data (Ustaömer and Kipman, 1998; Ustaömer and Rogers, 1999; Ustaömer, 1999; Ustaömer et al., 2005). In a similar way, Okay et al. (2008) dated 570 Ma in granitoids, located in west of Armutlu

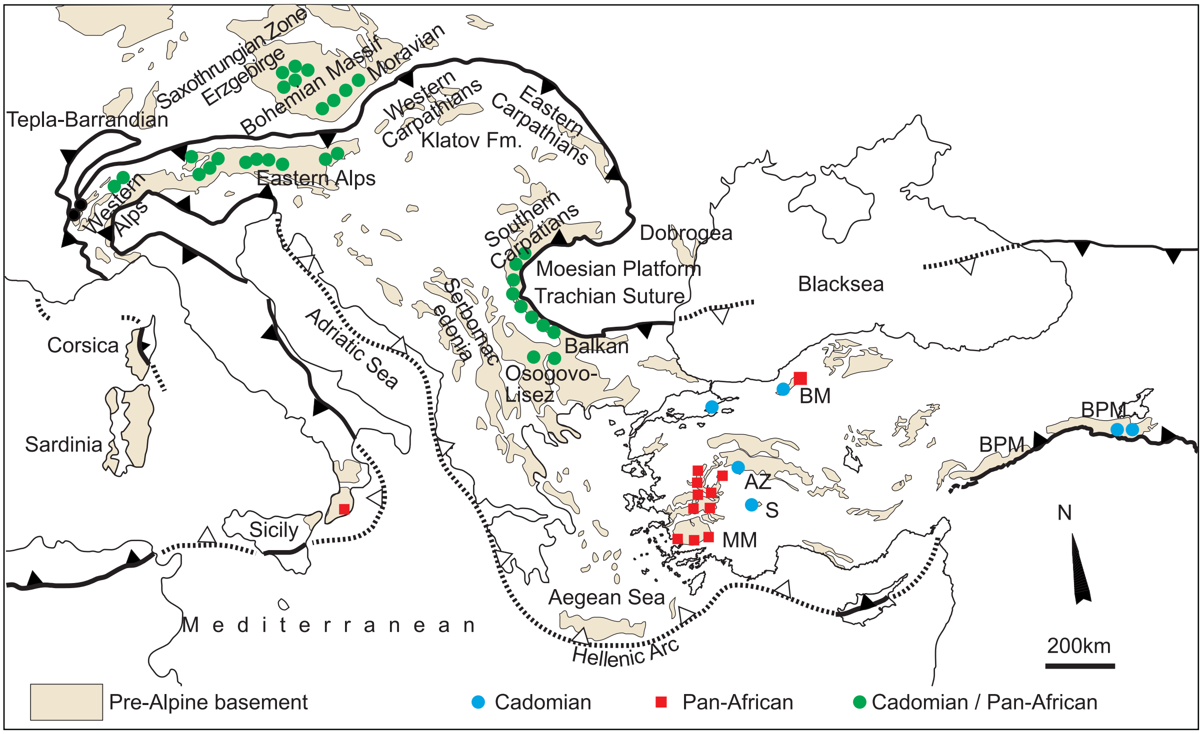

Figure 24- Locations of Pan-African and Cadomian magmatic activities observed in Turkey (modified from Neubauer, 2002). Cadomian location was taken from Kröner and Þengör, (1990), Gürsu and Göncüoðlu (2005), Gürsü et al. (2005), Ustaömer et al. (2005), Okay et al. (2008), Ustaömer et al., (2009); Pan-African locations were taken from Dannat (1997), Hetzel and Reischmann (1996), Loos and Reischmann (1999), Gessner et al. (2001, 2004), Chen et al. (2002), Dora et al. (2002), Koralay et al. (2004), Dora et al. (2006) and Cadomian- Pan-African locations were taken from Neubauer (2002) AZ: Afyon Zone, BM: Bolu Massif, BPM: Bitlis-Pötürge Massif, MM: Menderes Massif, S: Sandýklý.

Peninsula. Investigators include these granitoids into the Cadomian basement of Istanbul Zone.

The intrusion ages of rocks defined as mylonitic granite in Sandýklý Region were dated as 543±7 Ma. by single zircon evaporation method (Kröner and Þengör, 1990) (Figure 24). Gürsu et al. (2004) names these rocks as meta quartz porphyry and defines these are geochemically post orogenic and I-type granites. In recent years, these metaquartz porphyries have been dated as 541±9 Ma similarly, by single zircon evaporation method (Gürsu and Göncüoðlu, 2005). Granite and its accompanying mineral metarhyolites are considered as I-magmatism products associated with extension that has occurred in the evolution following the Pan-African Orogeny. These are observed in Taurides being considered as it is located at north of Africa (Gürsu et al., 2004; Gürsu and Göncüoðlu, 2005, 2006). In the suggested model of investigators, it was stated that, the oceanic plate subducting into the northern margin of Gondwana caused dilation in the northern part of the continent and thus, widespread granitic intrusions along the continental margins covering Taurides and South Europe is between 550-540 Ma. ages have occurred. It is considered that all these metamorphisms and magmatic activities observed in these regions are the products of Cadomian

Orogeny (Bozkaya et al., 2006, Gürsu and Göncüoðlu, 2006).

Similarly, the presence of metaquartz porphyries was determined and dated as 541.1±3.6 Ma. by the single Zircon evaporation method in Afyon zone (Figure 24). These I-typed alkaline magmatic rocks are interpreted as the intrusives developed by the post orogenic crustal expansion in Cadomian Orogeny (Gürsu et al., 2005).

Mutki granite located in Bitlis Massif and the accompanying dikes were dated as 545.5±6.1 Ma and 531.4±3.6 Ma respectively (Ustaömer et al., 2009), and these ages were interpreted as intrusion age of related magmatics (Figure 24). Geochemical and Nd istopic analysis results show that these granites and dikes have originated from mantle and are Arc-type granites associated with subduction zone. By the same authors, all of the Ediacaran-Early Cambrian magmatic activities observed at Bitlis Massif, Menderes Massif and at the Palaeozoic basement of Istanbul Zone are the products of Cadomian arc type magmatism (Ustaömer et al., 2009).

Looking at Europe in general, Cadomian tectonical units (Figure 24), are observed within various regions of Late Neoproterozoic- Early Palaeozoic base units cropping out in Alpine-Mediterranean mountain belt extending from Alpines to Turkey (Neubauer, 2002). Ustaömer et al. 2009, state that Cadomian tectonical units reach up to Iran and India. It is also stated that the easternmost extension of this belt in Europe is defined by the Cadomian magmatic activity dated as between 590-510 Ma in the Iberian Massif (Ochsner, 1993; Bandres et al., 2004, and related references). Cadomian tectonic units are located in Alpine basement constitutes Helvetic, Penninic, Austro Alpine and South Alpine tectonic units (Müller et al., 1995; Neubauer, 2002). The ages of the intrusive rocks in this zone show a variation between 520-570 Ma. Besides, in Erzgebirge, located at north of this zone, in Saxonia and in Bohemian massif 550 Ma, 555 Ma and 567 Ma magmatism ages were dated, respectively (Kröner et al., 1995; Müller et al., 1995; Friedl et al., 2004). At south Carpathians-Balcanians 567-563 Ma and in Sirbian-Macedonian Massif 545-568 Ma aged magmatic intrusives were detected (Neubauer, 2002). Similarly, the existence of 533±9 Ma, 540±10 Ma, 548±9 Ma aged granitoidic intrusives related to Cadomian orogeny are known in Poland and Czech Republic (Zelazniewicz et al., 2004). In recent studies, the intrusion age of gneisses in the Pelagonian Zone have been dated as 546±10 Ma (Anders et al., 2007) (Figure 24).

In summary, majority of granitoids between 520-570 Ma in Europe are associated with Cadomian Orogeny. It is also accepted that at Andes-type continental margin located at north of Gondwana super continent of this orogeny, the granitoids are represented by the products of a calc alkaline, I-type magmatism which were located in subduction-obduction zone towards south (Neubauer, 2002).

Magmatism associated with Pan-African orogeny

In general, the Pan-African Orogeny defines a chain of poly-phase orogenic events which covers the defragmentation of Gondwana Super Continent, subduction, collision and connection periods associated with it and conclude the formation of orogenic belts. The age of Pan-African Orogeny is defined differently by many investigators as; 950 - 450 Ma (Kröner, 1984), 1000-540 Ma (Stern, 1994), 650-550 Ma (Wilson et al., 1997; Veevers, 2004) and 1000-550 Ma (Unrug, 1996). This orogeny is not only limited to African continent, but also covers the whole Gondwana continent and the events occurred at south America, in Madagascar, Sri Lanka, south India, Antarctica and Australia (Kröner and Stern, 2005). There is a general agreement on the separation of East and West Gondwana by a big ocean named as Mozambique Ocean (Dalziel, 1991) in Neoproterozoic (Stern 1994; Wilson et al.,1997). The closure of this ocean formed by the fragmentation of Rodinia continent 800-850 Ma ago, the collision of East and West Gondwana continents caused the formation of N-S extending belt along the eastern margin of today's African continent (Figure 25). This belt is defined as Mozambique belt or East African Orogeny (Stern, 1994; Kröner and Stern, 2005).

During the Pan-African Orogeny, many small plates collisioned and formed different belts. Along these belts, pervasive existence of intrusives showing similar ages and features of orthogneisses in the Menderes Massif is observed (Figure 25). When the paleogeographical position of the Menderes Massif in Late Proterozoic-Early Cambrian is assessed, the Mozambique belt occupies a special place. Stern (1994) states the presence of 540 Ma aged granites in south Somalia. Wilson et al., (1997) claims the presence of S typed granites and granitoids as 500-538 Ma in Tanzania, as 550 Ma in India and Sri Lanca and as 550 Ma in Pyrdz Bay/Antarctica. S typed 560 Ma aged granites in Eastern Ethiopia is known and Teklay et al. (1998) stressed that these granites can be correlated with western Ethiopia and Arabic-Nubian Shield. Furthermore, in southern Ethiopia the existence of post-tectonical granitic and tonalitic intrusions with ages varying between 529-557 Ma were determined by many investigators (Yibas et al., 2002, and the references there in). Magmatic activities with varying ages between 563-611 Ma were reported in northeastern desert of Egypt, one of the closest area to the Menderes massif (Gessner et al., 2004). Kröner and Stern (2005) mentions about the presence of granite derived from the crust of 550 Ma aged in Kaoko belt. In Calabria region, at southern Italy, out of Africa, calc alkaline, post collisional augen gneisses were determined with ages varying between 526-562 Ma (Micheletti et al, 2007). It is considered that these gneisses are the Pan-African post collisonal granites and show big similarities with West African Craton. Goodenough et al. (2010) dated post collisional granitoids as 522-537 Ma which have formed in the stage following the East African Orogeny, north of Madagascar.

TECTONIC ENVIRONMENTS OF ORTHOGNEISSES IN THE MENDERES MASSIF

In the pioneering geochronological studies made in the Menderes Massif, deformation/metamorphism were dated as 500±10 Ma and tonalitic and granitic intrusives were dated as 471±9 Ma (Satýr ve Friedrichsen, 1984). Þengör et al. (1984) were suggested an opinion that the magmatic and metamorphic stages of core series in the Menderes Massif can be associated with the Pan-African Orogeny. Investigators suggest a location for the Menderes Massif in the western part of the Arabian Peninsula, at the northernmost part of Mozambique belt during the Early Cambrian continental arrangement. This positioning is also supported with other geological evidences (Late Silurian glaciation, Monod et al., 2003; clastic zircon source rock, Gessner et al., 2004) and the location of the Menderes massif is positioned to the North of Anatolia-Arabian Peninsula in general maps (Stern, 1994; Kröner and Stern, 2005). In the following years Dora et al. (1994) stated that protolithes of orthogneisses observed in the Menderes Massif based on the regional geology were syn to post metamorphic granites intruded by the Pan-African Orogeny. This opinion has been accepted by many researchers as a result of the studies performed in recent years (Hetzel and Reischmann, 1996; Dannat, 1997; Dannat and Reischmann, 1997; Loos and Reischmann, 1999; Hetzel et al., 1998; Gessner et al., 2001; Okay, 2001; Dora et al., 2002, 2006; Gessner et al., 2004; Koralay et al., 2004). Geochemical evidences show that these granites are calc alcaline in origin, per aluminuous in character, S typed, syn/post orogenic granite/granodiorites (Bozkurt et al.,1992, 1993, 1995; Dannat 1997; Koralay and Dora, 1999; Koralay et al., 2004; Bozkurt et al., 2006).

In addition to orthogneisses, many studies have been performed in the Menderes Massif

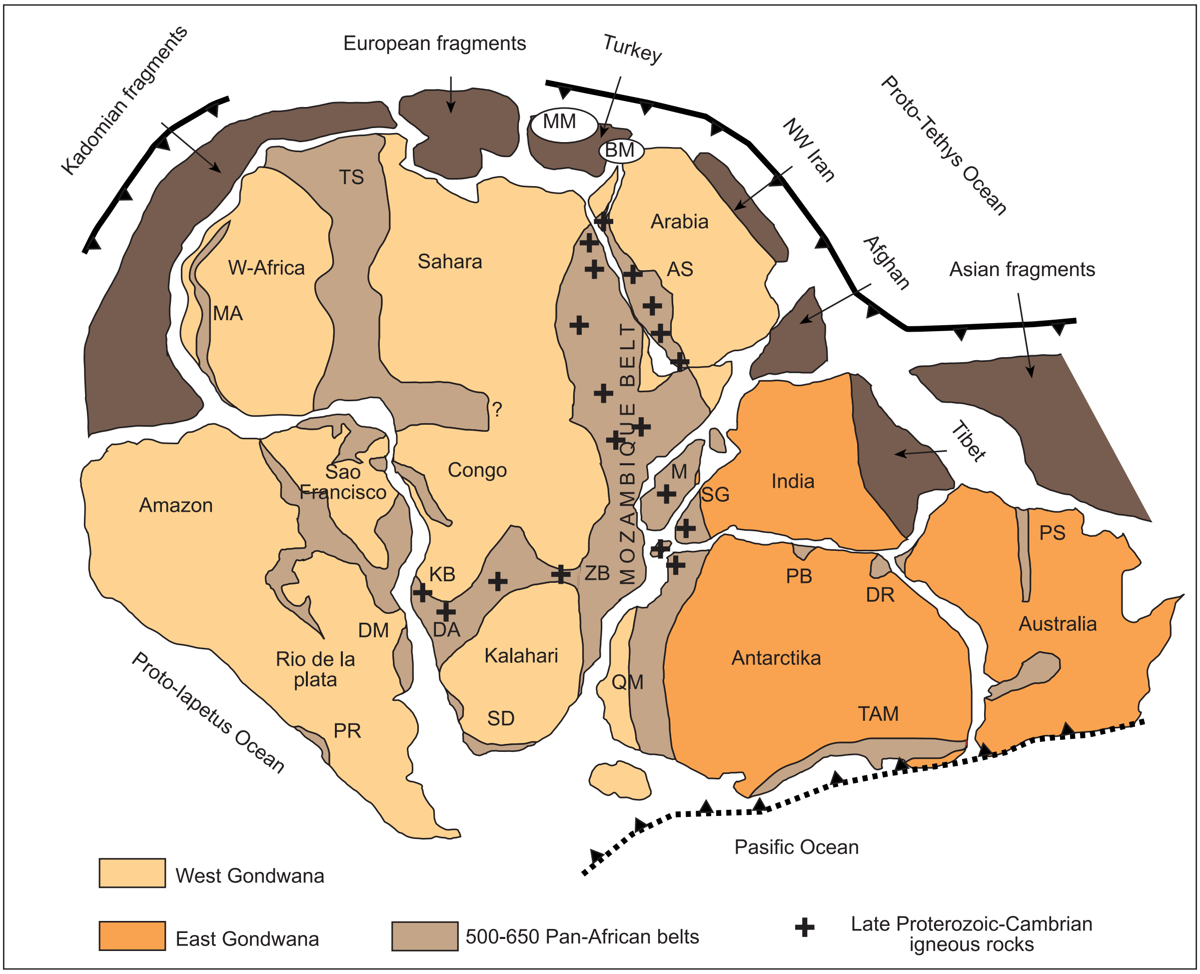

Figure 25- The paleogeographical map of Gondwana super continent in lower Late Neoproterozoic/Early Cambrian (modified from Wilson et al., 1997 and Kröner and Stern, 2005). The locations of the Mozambique belt and Late Neoproterozoic-Cambrian magmatic rocks are shown in the map. AS: Arabian Shield, DA: Damara, DM: Dom Feliciano, DR: Denman Darling, M:Madagaskar, MA: Mauretanides, MM: Menderes Massif, PB: Pryolz Bay, PR: Pampean Ranges, PS: Paterson, QM: Queen Maud Land, SD: Saldania, SG: Southern granulite region, TAM: Trans Antarctic Mountains, TS: Trans Sahara belt, ZB: Zambezi. Magmatic rock locations (from Unrug, 1996; Teklay et al., 1998; Hefferean et al., 2000; Yibas et al., 2002 and Kröner and Stern, 2005).

involving the dissertation of poly-metamorphic stage of core series petrologically and geochronologically in recent years. Petrological studies reveal the presence of a poly-metamorphic stage under granulite, eclogite and upper amphibolite facies conditions in core series (Çetinkaplan, 1995; Candan, 1995; Candan and Dora, 1998; Candan et al., 1994, 2001, 2006; Oberhänsli et al., 1997). In geochronological studies for dating these metamorphisms, metamorphisms of granulite, eclogite and upper amphibolite facies were dated as 583±5.7 Ma (Koralay et al., 2006), 529.9±22 Ma (Oberhänsli et al., 2009) and 551±1.4 Ma (Hetzel et al., 1998), respectively. Radiometric ages and relative textural relationships clearly display these metamorphisms are associated with the stages of an orogenic event following one another. There has not been any metamorphism under granulite and eclogite facies in orogeny involving metamorphic regions since the Cadomian Orogeny has not resulted with a collision of another continent. On the contrary, the stage related with the closure of Mozambique Ocean of the Pan African Orogeny in Late Proterozoic-Early Cambrian are defined by 610-520 Ma (with an average in 550 Ma) aged widespread granulite facies metamorphism (Paquette et al., 1994; Hölzl et al., 1994; Shiraishi et al., 1994; Ayalew and Gichile, 1990; Key et al., 1989) and 530-500 Ma aged eclogite facies metamorphism (Ring et al., 2002). The paleogeographical position of the Menderes Massif in Precambrian and Cambrian times and the stage of core series based on the temporal and spatial similarities of metamorphic evolution of Mozambique Ocean are associated with the closure of Mozambique Ocean and with the collision of East-West Gondwana continents (Candan et al., 2006, Oberhänsli et al., 2009).

The consistency of intrusion ages of orthogneisses varying between 570-520 Ma in the Menderes Massif with poly metamorphic stages of country rocks varying between 580-550 Ma, clearly displays that magmatic stage and the metamorphism is observed within the same orogenic event, genetically. When the paleogeographical position of Menderes Massif in Late Proterozoic- Early Cambrian and close temporal and spatial relationships are assessed with the character of metamorphism and the presentation of orthogneisses have a definite S type, the 570-520 Ma aged acidic magmatic activity in the Massif can be correlated with the closure of the Mozambique Ocean, collision of Gondwana, crustal thickening and the partial melting process occurred in sub crust.

TRIASSIC MAGMATISM

Intrusion age of primary granites of leucocratic orthogneisses

The Triassic magmatic activity in the Menderes Massif constitutes the second biggest magmatic activity observed after the Pan-African magmatism. Leaving aside the amphibolites defining probable Triassic basaltic volcanism located at the base of Mesozoic platform, the Triassic magmatism in the Massif are described by leucocratic orthogneisses. In preliminary studies, although there has been no geochronological evidence the orthogneiss mass cropped out in Dededag, south of Alaþehir was given Triassic age (Akkök, 1983; Þengör et al., 1984). As obviously explained above, detailed radiometric dates were obtained from these orthogneisses in 90's (Dannat, 1997; Dannat and Reischmann, 1998; Koralay, 2001; Koralay et al., 2001). Textural and structural data, contact relationships, ages of the country rock and the radiometric dating values by single Zircon Pb/Pb evaporation method clearly display that these leucoratic masses are magmatic in origin and their intrusion ages vary between 227-247 Ma (Middle Triassic - according to Gradstein et al., 2004).

Tectonic envrionmets of primary granites of leucocratic orthogneisses

As mentioned above, the Triassic magmatism is not only observed in the Menderes Massif (associating the phenomena with a regional tectonism) but also in Lycia belonging to Anatolides, in Afyon and Tavþanlý zones, in Karaburun, in Cyclades and in inner Hellenides.

The middle Carnian aged, inner plate MORB type transition basalts located in Gülbahar Nappe belongs to Lycian nappes and are interpreted as the products of the first formational evolution of the oceanic crust at northern branch of Neothethys (Göncüoðlu et al., 2003). Dacitic and rhyolitic metavolcanites accompanying with coarse clastics and located at the base of Mesozoic cover series of Afyon Zone were dated as 224-243 Ma. (Middle - Late Triassic) (Akal et al., 2007a; Akal et al., 2008). It is considered that these volcanites subducts southward and the Paleothethys ocean characterizes the rifting stage of Neothethys Ocean that was opened as backiarc basin located at the northern margin of Gondwana (Göncüoðlu et al., 2003; Akal et al., 2007*a*). It is known that blueschist metabazite stages and jadeite bearing magmatics are present in Tavþanlý zone, in metaaplites, at lower stages of platform type carbonates (Kulaksýz, 1978; Okay and Kelley, 1994; Çetinkaplan et al., 2008). There is not any investigation involving the geochemistry and its tectonical environment of these metabazites in Triassic (?) age. The Triassic magmatism in Karaburun peninsula show similarities with the Menderes Massif. Granodiorites cutting the Devonian-Carboniferous clastic series were dated as 229±3 Ma by single Zircon Pb/Pb evaporation method (Ercan et al., 2000). These are considered as the products of back arc magmatism developed synchronously by the rifting of Neothethys located at northern margin of Gondwana which is related with the Paleothethys subducting southward. The existence of magmatism in similar ages is known as Cyclades which is one of the closest magmatic regions to the Menderes Massif, except Anatolides and at Inner Hellenides at North (Reischmann, 1998; Himmerkus et al., 2009). Granites recognized in Nacsosia island were dated as 233±2 Ma by single Zircon Pb/Pb evaporation method (Reischmann, 1998). Leucocratic granites observed in the Serbo-Macedonia Massif located at inner Hellenides were dated varying between 221±2 Ma. and 241±3 Ma by the single Zircon Pb/Pb evaporation method. The average age is given as 228±6 Ma for this region (Himmerkus et al., 2009). Besides, the existence of Triassic granites at east of Vardar Zone and east of Pelagonian zones is also known (Himmerkus et al., 2009). As can be seen in the abstract, the Triassic magmatism can widely be observed in Turkey and in close tectonic units and is substantially associated with the opening of Neothethys ocean.

In pioneering investigations on Triassic granites in the Menderes Massif, these rocks have been interpreted as plutons that have intruded at the stage of deformation and the metamorphism by which the northern parts of the Massif was affected (Akkök, 1981, 1983; Þengör et al., 1984). This event was suggested as associative with the closure of Paleotehtys subducting southward which was located between Laurasia and Gondwana in Late Triassic (Akkök, 1983; Þengör et al., 1984). This scenario has been appropriated in following studies as well, and these rocks were interpreted as leucocratic orthogneisses that have settled in the stage following the Early Kimmerian Orogeny related with the closure of Paleotethys Ocean (Koralay, 2001; Koralay et al., 2001). However, although there is not detailed geochemical data, when the close relationship of the Menderes Massif with above described tectonic units in Triassic time and origin are considered, it is assumed that associating the Middle Triassic leucocratic granite intrusions with the rifting of northern branch of Neotethys ocean would be more realistic.

CONCLUSIONS

Results related to basic geological features of orthogneisses of which are the Pan-African and Triassic acidic magmatic activity products in the Menderes Massif are given below. These results have been obtained by investigations that have continued more than 50 years.

1. The Pan-African basement of the Menderes Massif is made up of Late Proterozoic metaclastics and by acidic - basic magmatics of which have intruded into. Orthogneisses form the most common rock type.

2. All gneisses in the Menderes Massif are granitic in origin and present well preserved intrusive contact relationships by metaclastics composed of paragneiss and schists that form the country rock. These rocks are in the form of plutons that have intruded one other and reach tens of kilometers, stocks reach in several kilometers and veins reach a few hundreds of meters.

3. The primary rocks of orthogneisses can be divided into three categories based on mineralogical composition and textural properties.

These are; i) biotite orthogneiss, ii) leucocratic orthogneiss and iii) amphibole orthogneisses.

4. Contact relationships relatively define that biotite orthogneisses are the oldest and amphibole orthogneisses are the youngest magmatic stages. Although there are some temporal contradictions, radiometric ages obtained so far support this relative relationship (biotite orthogneiss: 550-570 Ma; tourmaline leucocratic orthogneiss: 540-550 Ma., and amphibole orthogneiss: 530 Ma).

5. When poly metamorphic evolution of the Massif basement, intrusional ages of orthogneisses, geochemical properties of these rocks and the paleogeographical position of Turkey in Late Neoproterozoic-Early Cambrian are assessed altogether, it is noticed that these intrusives can be associated with the closure of Mozambique Ocean in Late Proterozoic and with the Pan- African Orogeny that caused the collision of East-West Gondwanaland.

6. The Triassic leucocratic orthogneisses observed in the Pan-African basement and Palaeozoic cover series are granitic in origin and represent the second effective acidic magmatic activity in the Massif.

7. Gneisses in the form of plutons and vein rocks reaching a 5-6 km in size present well preserved intrusive contact relationships with country rocks.

8. The intrusion ages of these granites were dated as 227-246 Ma (Middle Triassic) by geochronological studies.

9. When the Triassic magmatism in other tectonic zones belonging to Anatolides and the regional tectonism in Triassic are assessed together, the evolution and emplacement of these granites can be associated with the rifting mechanism of northern branch of Neotethys Ocean.

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REFERENCES

Akal, C., Candan, O., Koralay, O.E., Chen, F., and Oberhänsli, R., 2007*a*. Afyon Zonu'na ait olasýlý Erken Triyas yaþlý metavolkaniklerin jeokimyasý, jeokronolojisi ve tektonik ortamlarý. Türkiye Bilimsel ve Teknolojik Araþtýrma Kurumu Project No 103Y011.

_____, Koralay, O.E., Oberhänsli, R., Chen, F., and Candan, O., 2007*b*. Karaburun Yarýmadasý kýrýntýlý seri içindeki Triyas diyorit-granodiyoritinin jeokimyasý, jeokronolojisi ve tektonik ortamý. 60. Türkiye Jeoloji Kurultayý Bildiri Özleri, Ankara, 229-231 p.

_____, Candan, O., Koralay, O.E., Okay, A., Oberhänsli, R. and Chen, F., 2008. Afyon Zonu'nundaki Erken Devoniyen Asidik Magmatizmaya ait jeolojik, jeokimyasal ve jeokronolojik ön bulgular. 61. Türkiye Jeoloji Kurultayý Bildiri Özetleri Kitabý, 204 p.

Akat, U., Öztürk, Z., Öztürk, E. and Çaðlayan, A., 1975. Menderes Masifi güneyi-GB Toros Kuþaðý iliþkisi (ön rapor), Maden Tetkik ve Arama Genel Müdürlüðü Raporlarý, no:548, Ankara (unpublished).

Akdeniz, N. and Konak, N., 1979. Menderes Masifi'nin Simav dolaylarýndaki kaya birimleri ve metabazik, metaultramafik kayalarýn konumu. Türkiye Jeoloji Kurultayý Bülteni, 22, pp.175-183.

Akkök, R., 1981. Menderes Masifi'nin gnayslarýnda ve þistlerinde metamorfizma koþullarý, Alaþehir / Manisa. Türkiye Jeoloji Kurumu Bülteni, C 24, pp.11-20.

_____, 1983. Structural and metamorphic evolution of the northern part of the Menderes Massif: New data from the Derbent area and their implication for the tectonics of the massif. Journal of Geology, 91, 342-350.

_____, Satýr, M. and Þengör, A.M.C., 1984. Timing of tectonic events in the Menderes Massif and its implications. 38th Scientific and Technical Congress of the Geological Society of Turkey, 9-11.

Alkanaoðlu, E., 1978. Geologisch-Petrographische und Geochemische Untersuchungen am Südostrand des Menderes-Massivs in West Anatolien / Turkei. Dissertation, Universität Bochum, 166 p.

Anders, B., Reischmann, T. and Kostopoulos, D., 2007. Zircon geochronology of basement rocks from the Pelagonian Zone, Greece: constrains on the pre-Alpine evolution of the westernmost Internal Hellenides. International Journal of Earth Science, 96, 639-661.

Ayalew, T. and Gichile, S., 1990. Preliminary U-Pb ages from southern Ethiopia. In G.Rocci, M. Deschamps (eds), Recent data in African Earth Sciences, CIFEG Occ. Publ. 22, 127-130.

Bandres, A., Eguiluz, L., Pin, C., Paquette, J.L., Ordonez, B., Le Fevre, B., Ortega, L.A. and Ibarguchi, G.J.A., 2004. The northern Ossa-Morena Cadomian batholith (Iberian Massif): magmatic arc origin and early evolution. International Journal Earth Science, 93, 860-885.

Baþarýr, E., 1970. Bafa gölünün doðusunda kalan Menderes Masifi güney kanadýnýn jeolojisi ve petrografisi. Ege Üniversitesi Fen Fakültesi Ýlmi Raporlar Serisi No: 102, 1-44.

Bingöl, E., Delaloye, M. and Ataman, G., 1982. Granitic intrusions in western Anatolia: a contribution to the geodynamic study of this area. Eclogea Geologica Helvetica, 75, 437-446.

Bozkaya, Ö., Gürsu, S. and Göncüoðlu, M.C., 2006. Textural and mineralogical evidence for a Cadomian tectonothermal event in the eastern Mediterranean (Sandýklý-Afyon area, western Taurides, Turkey). Gondwana Research, 10, 301-315.

Bozkurt, E., 1994. Effects of Tertiary extension in the southern Menderes Massif, western Turkey. PhD Thesis, University of Keel, 395 p.

_____, 2004. Granitoid rocks of the southern Menderes Massif: field evidence for Tertiary magmatism in an extensional shear zone. International Journal of Earth Science, 93, 52-71.

_____, Park, R.G. and Winchester, J.A., 1992. Evidence against the core/cover concept in the southern sector of the Menderes Massif. International Workshop: Work in progress on the Geology of Turkey, Keel-England, 9-10 April 1992, Abstracts, 22 p.

_____, _____ and _____, 1993. Evidence against the core/cover interpretation of the southern sector of the Menderes Massif, west Turkey. Terra Nova 5, 445-451.

_____ and _____, 1994. Southern Menderes Massif: an incipient metamorphic core complex in Western Anatolia, Turkey. Journal Geologe Society London 151, 213-216

_____, Winchester, J.A. and Park, R.G., 1995. Geochemistry and tectonic significance of augen gneisses from the southern Menderes Massif (West Turkey). Geological Magazine, 132, 287-301.

_____ and Park, R.G., 1997*a*. Evolution of a mid-Tertiary extensional shear zone in the southern Menderes Massif, western Turkey. Bulletin de la Societe Geologique France, t.168,1, 2-14.

_____ and _____, 1997*b*. Microstructures of deformed grains in the augen gneisses of southern Menderes Massif (western Turkey) and their tectonic significance. Geologische Rundschau, 86, 103-119.

Bozkurt, E., Winchester, A. J., Mittwede, S. and Ottley, C., 2006. Geochemistry and tectonic implications of leucogranites and tourmalines of the southern Menderes Massif, southwest Turkey. Geodinamica Acta, 19/5, 363-390.

_____ and Park, G., 2001. Discussion on the evolution of the Southern Menderes Massif in SW Turkey as revealed by zircon dating. Journal Geological Society, 158, 393-395.

Candan, O., 1994. Alaþehir kuzeyinde (Menderes Masifi, Demirci -Gördes Asmasifi) gözlenen metagabrolarýn petrografisi ve metamorfizmasý. Türkiye Jeoloji Bülteni, 37, 29-40.

_____, 1995. Menderes Masifi'ndeki kalýntý granulit fasiyesi metamorfizmasý. Turkish Journal of Earth Sciences, 4, 35-55.

_____, 1996. Aydýn - Çine Asmasifi'ndeki (Menderes Masifi) gabrolarýn metamorfizmasý ve diðer asmasiflerle karþýlaþtýrýlmasý. Turkish Journal of Earth Sciences, 5, 123-139.

_____, Dora, Ö., Dürr, St. and Oberhänsli, R., 1994. Erster Nachweis von Granulit und Eklogit - Relikten im Menderes - Massif / Türkei. Göttingen Abr. Geol. Paläont. Sb.1 5.Symposium TSK, 217-220.

_____, _____, Oberhänsli, R., Ölsner, F. and Dürr, St., 1997. Blueschist relics in the Mesozoic cover series of the Menderes Massif and correlations with Samos Island, Cyclades. Schweiz Mineral. Petrol. Mitt., 77, 95-99.

_____ and _____, 1998. Menderes Masifi'nde granulit, eklojit ve mavi þist kalýntýlarý: Pan-Afrikan ve Tersiyer metamorfik evrimine bir yaklaþým. Türkiye Jeoloji Bülteni, 41/1, 1-35.

_____, _____, _____, Çetinkaplan, M., Partzsch, J.H., Warkus, F. and Dürr, S., 2001. Pan-African high-pressure metamorphism in the Precambrian basement of the Menderes Massif, Western Anatolia, Turkey. International Journal of Earth Science (Geologische Rundschau), 89, 4, 793-811.

_____, Koralay, E., Dora, O., Chen, F., Oberhänsli, R., Akal, C., Satýr, M. and Kaya, O., 2006. Menderes Masifi'nde Pan-Afrikan Sonrasý Uyumsuzluk: Jeolojik ve Jeokronolojik Bir Yaklaþým. 59. Türkiye Jeoloji Kurultayý Bildiri özleri, Ankara, 25-27.

Candan, O. Koralay, E. Dora, Ö. Chen, F. Oberhansli, R. Çetinkaplan, M. Akal, C. Satýr, M. and Kaya, O.,, 2007. Menderes Masifi'nin Pan-Afrikan temelin stratigrafisi ve örtü - çekirdek serilerinin ilksel dokanak iliþkisi. Menderes Masifi Kolokyumu, Ýzmir.

_____, Dora, O.Ö., Oberhaensli, R., Koralay, O.E., Çetinkaplan, M., Akal, C., Satýr, M., Chen, F. and Kaya, O., 2010. Menderes Masifi'nin Pan-Afrikan Temelin Stratigrafisi ve Gondvana'nýn Geç Neoproterozoyik/Kambriyen Evrimi Ýle Ýliþkisi. Menderes Masifi Kolokyumu, Ýzmir.

Catlos, E.J. and Çemen, Ý., 2005. Monazite ages and the evolution of the Menderes Massif, western Turkey. International Journal of Earth Science, 94, 204-217.

Chen, F., Siebel, W., Satýr, M. and Terzioðlu, M.N., 2002. Geochronology of the Karadere basement (NW Turkey) and implications for the geological evolution of the Ýstanbul Zone. International Journal of Earth Science, 91, 469-481.

Çaðlayan, A.M. Öztürk, E.M. Öztürk, Z. Sav, H. and Akat, U., 1980. Menderes Masifi güneyine ait bulgular ve yapýsal yorum. Jeoloji Mühendisliði, 10: 9-17.

Çetinkaplan, M., 1995. Geochemical, mineralogical and petrographical investigation of the eclogites in southern part of Tire area, Ödemiþ-Kiraz submassif of the Menderes Massif. Unpublished master theses, Dokuz Eylül Üniversitesi, Ýzmir, 92p.

_____, Candan, O. Oberhansli, R. and Bousquet, R., 2008. Pressure-temperature evolution of lawsonite eclogites in sivrihisar, Tavþanlý zone, Turkey. Lithos, 104, 12-32.

Dalziel, I.W.D., 1991. Pasific margins of Laurentia and east Antartica-Australia as a conjugate rift pair: Evidence and implications for an Eocambrian supercontinent. Geology, 19, 598-601.

Dannat, C., 1997. Geochemie, geochronologie und Nd-Sm Isotopie der granitoiden Kerngneiss des Menderes Massivs, SW-Turkey. PhD thesis, Johannes Gutenberg Universitat Mainz.

_____ and Reischmann, T., 1998. Single zircon ages of migmatites from the Menderes Massif, SW Turkey. Program des Workshops 'Das Menderes Massif (Türkei) und seine nach-bargebiete'. Mainz, Germany.

Delaloye, M. and Bingöl, E., 2000. Granitoids from Western and Northwestern Anatolia: Geochemistry and modelling of geodynamic evolution. International Geology Review, 42, 241-268.

Dora, O.Ö., 1975. Menderes Masifi'nde alkali feldspat-larýn yapýsal durumlarý ve bunlarýn perojenetik yorumlarda kullanýlmasý. Türkiye Jeoloji Kurumu Bülteni, 24, 91-94.

_____, 1976. Die Feldspäte als petrogenetischer Indikator im Menderes Massiv / Westanatolien. Neues Jahrbuch für Mineralogie Abh., 127, 289-310.

Dora, O.Ö., Kun, N. and Candan, O., 1990. Metamorphic history and geotectonic evolution of the Menderes Massif. Proc. of International Earth Sciences Congress on Aegean Regions, Ýzmir/Turkey, Vol. 2, 102-115.

_____, Candan, O., Kun, N., Koralay, O.E. and Akal, C., 1994. Ödemiþ-Kiraz Asmasifi'ndeki yeni jeolojik bulgular ve sorunlar. 47. Türkiye Jeoloji Kurultayý, pp. 32-33.

_____, _____, Dürr, S. and Oberhänsli, R., 1995. New evidence on the geotectonic evolution of the Menderes Massif. International Earth Sciences Colloquium on the Eagean Region, Izmir-Turkey, V.1, 53-72.

_____, _____,Kaya, O. Koralay, E. and Dürr, S., 2001. Revision of the so-called "leptite-gneisses" in the Menderes Massif: A supracrustal metase-dimentary origin. International Journal of Earth Science (Geologische Rundschau), 89/4, 836-851.

Dora, O.Ö., Candan, O., Kaya, O. and Koralay, E., 2002. Menderes Masifi'ndeki Leptit - Gnays - larýn Kökenlerinin Yeniden Yorumlanmasý, Metamorfizmalarý ve Jeotektonik Ortamlarý YDABÇAG - 554 nolu TÜBÝTAK projesi, 165p.

_____, ____, ____, ____ and Akal, C., 2005. Menderes Masifi Çine Asmasifi'ndeki Koçarlý - Bafa - Yataðan - Karacasu arasýnda uzanan gnays / þist dokanaðýnýn niteliði: Jeolojik, tektonik, pet-rografik ve jeokronolojik bir yaklaþým. YDABÇAG - 101 Y 132 nolu TÜBÝTAK projesi, 197p.

Dürr, St., 1975. Über Alter und geotektonische Stellung des Menderes Kristallins / SW - Anatolien und seine Äquivalente in der Mittleren Aegean. Habilitation thesis University of Marburg, 1-107.

_____, Alther, R., Keller, J., Okrusch, M. and Seidel, E., 1978. The median Aegean crystalline belt: Stratigraphy, structure, metamorphism, mag-matism. In: Closs, H., Roeder, D.R., Schmidt, E. (eds), Alps, Apennines, Hellenides. Schweizerbart, Stuttgart, 445-477.

Ercan, T., Türkecan, A. and Satýr, M., 2000. Karaburun Yarýmadasý'nýn Neojen volkanizmasý. Cumhuri-yetin 75. yýldönümü Yerbilimleri ve Madencilik Kongresi Bildiriler Kitabý, 1, 1-18.

Erdoðan, B., 1992. Problem of core - mantle boundary of Menderes Massif. In proceedings of the international Symposium of Eastern Medi-terranean Geology. Geosound, 314-315.

_____, 1993. Menderes Masifi'nin kuzey kanadýnýn stratigrafisi ve çekirdek örtü iliþkisi. Özetler, Türkiye Jeoloji Kurultayý, 56p.

_____ and Güngör, T., 2004. The problem of the core - cover boundary of the menderes Masif and an emplacement mechanism for regionally exten-sive gneissic granite, Western Anatolia Turkey. Turkish Journal of Earth Science, 13, 15-36.

Friedl, G., Finger, F., Paquette, J.L., von Quadt. A., McNaughton, N.J. and Fletcher, I.R., 2004. Pre-Variscan geological events in the Austrian part of the Bohemian Massif deduced from U-

Pb zircon ages. International Journal of Earth Science, 93, 802-823.

Gessner, K., Piazolo, S., Güngör, T., Ring, U., Kröner, A. and Passchier, C.W., 2001. Tectonic significance of deformation in granitoid rocks of the Menderes nappes, Anatolide belt, southwest Turkey. International Journal of Earth Science, 89, 766-780.

_____, Collins A. Ring, U. and Güngör, T., 2004. Structural and thermal history of poly-orogenic basement: U-Pb geochronology of granitoid rocks in the southern Menderes Massif, Western Turkey. Journal of the Geological Society London, 161, 93-101.

Glodny, J. and Hetzel, R., 2007. Precise U-Pb ages of syn-extensional Miocene intrusions in the central Menderes Massif, western Turkey. Geological Magazine 144(2), 235-246.

Goodenough, K.M., Thomas, R.J., De Waele, B., Key, R.M., Schofield, D.I., Bauer, W., Tucker, R.D., Rafahatelo, J.M., Rabarimanana, M., Ralison, A.V. and Randriamananjara, T., 2010. Post-collisional magmatism in the central East African Orogen: The Maevarano Suite of North Madagascar. Lithos, 116, 18-34.

Göncüoðlu, M.C., Turhan, N. and Tekin, K., 2003. Evidence for the Triassic rifting and opening of the Neotethyan Ýzmir -Ankara Ocean, northern edge of the Tauride-Anatolide Platform, Turkey. Bulletin of Geological Society Italy, Special Volume 2, 203-212.

Graciansky, P., 1965. Menderes Masifi'nin güney kýyýsý boyunca (Türkiye'nin GB'sý) görülen metamorfizma hakkýnda açýklamalar. Maden Tetkik ve Arama Genel Müdürlüðü Bülteni, 64, 8-22.

Gradstein, Ogg, J.G., Smith A.G.B. and Lourens, L.J., 2004. A new Geological Time Scale, with special reference to Precambrian and Neogene. Episodes, 27, 83-100.

Gürsu, S., Göncüoðlu, M.C. and Bayhan, H., 2004. Geology and geochemistry of Pre-early Cambrian rocks in the Sandýklý area: Implication for the Pan-African evolution of NW Gondwana. Gondwana Research, 7, 923-935.

Gürsu, S. and Göncüoðlu, M.C., 2005. Early Cambrian back-arc volcanism in the Western Taurides, Turkey: implications for the rifting along northern Gondwanan margin. Geological Magazine, 142(5), 617-631.

_____, _____ and Turhan, N., 2005. Geology adn petrology of Cadomian felsic magmatism in Afyon area, western central Turkey. International Earth Sciences Colloquim on the Aegean Regions, Abstracts, pp. 46-47.

_____ and _____, 2006. Petrogenesis and tectonic setting of Cadomian felsic igneous rocks, Sandýklý area of the western Taurides, Turkey. International Journal of Earth Science, 95, 741-757.

Hetzel, R., Ring, U., Akal, C. and Troesch, M., 1995. Miocene NNE-directed extensional unroofing in the Menderes Massif, southern Turkey. Journal Geological Society, London, 152, 639-654.

_____ and Reischmann, T., 1996. Intrusion age of Pan-African augen gneisses in the southern Menderes Massif and the age of cooling after Alpine ductile extensional deformation. Geological Magazine 133(5): 565 - 572.

_____, Romer, R. Candan, O. and Passchier, C.W., 1998. Geology of the Bozdað area, central Menderes Massif, SW Turkey: Pan - African basement and Alpine deformation. Geologische Rundschau, 87, 394-406.

Himmerkus, F., Reischmann, T. and Kostopoulos, D., 2009. Triassic rift-related meta-granites in the Internal Hellenides, Greece. Geological Magazine, 146 (2), 252-265.

Hölzl, S. Hofmann, A.W. Todt, W. and Köhler, H., 1994. U-Pb geochronology of the Sri Lankan basement. Precambrian Research, 66, 123-149.

Iþýk, V. and Tekeli, O., 2001. Late orogenic crustal extension in the northern Menderes Massif (western Turkey): evidence for metamorphic core complex formation. International Journal of Earth Science, 89, 757-765.

Iþýk, V., Tekeli, O. and Seyitoðlu, G., 2004. The Ar/Ar age of extentional deformation and granite intrusion in the northern Menderes core complex: implications for the initiation of extentional tectonics in western Turkey. Journal of Asian Earth Science, 23, 555-566.

Jacobshagen, V., 1986. Geologie von Griechenland. Beitrage zur regionalen Geologie de Erde Band 1, Gebrueder Bornrüder Berlin, 363 p.

Key, R.M. Charsley, T.J. Hackman, B.D. Wilkinson, A.F. and Rundle, C.C., 1989. Superimposed Upper Proterozoic collision-controlled orogenies in the Mozambique orogenic belt of Kenya. Precambrian Research, 44, 197-225.

Konak, N., 1985. Menderes Masifi'nde çekirdek - örtü iliþkilerinin yeni gözlemler ýþýðýnda tartýþýlmasý. Türkiye Jeoloji Kurultayý, bildiri özleri, 33p.

_____, Akdeniz, N. and Öztürk, E.M., 1987. Geology of the south of Menderes Massif, I.G.C.P. project no:5, Correlation of Variscan and pre-Variscan events of the Alpine Mediterranean mountain belt, field meeting, Mineral Research and Exploration Institute, Turkey, 42-53.

_____, Çakmakoðlu, A., Elibol, E., Havzoðlu, T., Hepþen, N., Karamanderesi, I.H., Keskin , H., Sarýkaya, H., Sav, H. and Yusufoðlu, H., 1994. Development of thrusting in the median part of the Menderes Massif. Abstracts 47th Geological Congress Turkey-Ankara: 34p. (Abstract).

Koralay, O.E., 2001. Geology, geochemistry and geochronology of granitic gneisses and leucocratic orthogneisses at the eastern part of Ödemiþ-Kiraz submassif, Menderes Massif: Pan-African and Triassic magmatic activities. PhD Thesiss, Graduate School of Natural Science, Dokuz Eylül University, Ýzmir, 191 p.

_____ and Dora, O.Ö., 1999. Menderes Masifi'nde Derbent (Alaþehir) yöresinin jeolojisi ve olasýlý Kimmeriyen metamorfizmasý. Yerbilimleri (Geosound), 34, 151-172.

_____, Satýr, M. and Dora, O.Ö., 2001. Geochemical and geochronological evidence for Early Triassic calc-alkaline magmatism in the Menderes Massif, western Turkey. International Journal of Earth Science, 89, 822-835.

Koralay, O.E., Ðora, O.Ö., Candan, O., Chen, F. and Satýr, M., 2003. Menderes masifindeki paragnayslarýn ilksel çökelme yaþýna tek zirkon Pb/Pb evaporasyon jeokronolojisi yöntemi ile yaklaþým. 56. Türkiye Jeoloji Kurultayý, 64-65.

_____, ____, Chen, F., Satýr, M. and Candan, O., 2004. Geochemistry and geochronology of orthogneisses in the Derbent (Alaþehir) area, Eastern part of the Ödemiþ - Kiraz submassif, Menderes Massif: Pan-African magmatic activity. Turkish Journal of Earth Science, 13, 37-61.

_____, Chen, F., Candan, O., Dora, O.Ö., Satýr, M. and Oberhänsli, R., 2005. Pb-Pb geochronology of detrital zircons from Neoproterozoic paragneisses in the Menderes Massif, Turkey. International Earth Sciences Colloquim on the Aegean Regions, Ýzmir-Turkey, 69p.

_____, Candan, O., Dora, Ö., Satýr, M., Oberhänsli, R. and Chen, F., 2007. Menderes Masifi'ndeki Pan-Afrikan ve Triyas yaþlý metamagmatik kayaçlarýn jeolojisi ve jeokronolojisi, Batý Anadolu, Türkiye. Menderes Masifi Kolokyumu, Ýzmir.

Kröner, A., 1984. Late Precambrian plate tectonics and orogeny: a need to redefine the term Pan-African. In: Klerkx, J. ve Michot, J. (eds) African Geology, pp.23-28. Tervuren: Musccc. R. I'Afrique Centrale.

_____ and Þengör, A.M.C., 1990. Archean and Proterozoic ancestry in the Late Precambrian to Early Palaeozoic crustal elements of southern Turkey as revealed by single zircon dating. Geology, 18, 1186-1190.

_____ and Stern, R.J., 2005. Pan-African Orogeny. Encyclopedia Geology, Volume-I, A to E, 1-12, Elsevier.

_____, Willner, A.P., Hegner, E., Frischbutter, A., Hofmann, J. and Bergner, R., 1995. Latest Precambrian (Cadomian) zircon ages, Nd isotopic systematics and P-T evolution of granitoid orthogneisses of the Erzgebirge, Saxony and Czech Republic. Geologische Rundschau, 84, 437-456.

Kulaksýz, S., 1978. Sivrihisar kuzeybatý yöresi eklojitleri. Yerbilimleri, 4, 89-94.

Lips, A.L.W., Cassard, D., Sözbilir, H., Yýlmaz, H. and Wijbrans, J.R., 2001. Multistage exhumation of the menderes Massif, western Anatolia (Turkey) International Journal of Earth Science, 89, 781-792.

Loos, S., 1995. Alterbestimmungen im SW Menderes Massiv, Türkei, mit der einzirkon - Pb/Pb evaporatýosmethode. Diplomarbeit, Johannes Gutenberg Universitat Mainz, 95p.

_____ and Reischmann, T., 1999. The evolution of the southern Menderes massif in SW Turkey as revealed by zircon dating. Journal Geological Society London, 156, 1021-1030.

Micheletti, F., Barbey, P., Fornelli, A., Piccarreta, G. and Deloule, E., 2007. Latest Precamrian to Early Cambrian U-Pb zircon ages of augen gneisses from Calabria (Italy), with inference to the Alboran microplate in the evolution of the peri-Gondwana terranes. International Journal of Earth Science, 96, 843-860.

Mittwede, S.K., Karamanderesi, Ý.H. and Helvacý, C., 1995*a*. Tourmaline-rich rocks of the southern part of the Menderes Massif, southwestern Turkey. International Earth Sciences Colloquium on the Aegean Region, 1995 Excursion Guide, Dokuz Eylül University, Department of Geological Engineering, Ýzmir, 25p.

_____, Sinclair, W.D., Karamanderesi, Ý.H. and Helvacý, C., 1995*b*. Geochemistry of quartz-tourmaline nodules from Irmadan (Muðla-Yataðan), Türkiye. Abstracts of the Second International Turkish Geology Wrokshop, September 6-8, 1995, Sivas, Turkey, 74p.

_____, Sinclair, W.D., Helvacý, C. and Karamanderesi, Ý.H., 1997. Quartz-tourmaline nodules in leucocratic metagranite, southern flank of the Menderes Massif, SW Turkey. Tourmaline 97, International Syposium on Tourmaline, Abstract Volume, Czech Republic, 57-58.

Monod, O., Kozlu, H., Ghienne, W., Dean, W.T., Günay. Y., Herisse, A. and Paris. F., 2003. Late Ordovician glajiation in southern Turkey. Terra Nova, 15, 4, pp.249-257.

Müller, B., Klötzli, U.S. and Flisch, M., 1995. U-Pb and Pb-Pb zircon dating of the older orthogneiss suite in the Silvretta nappe, eastern Alps: Cadomian magmatism in the upper Austro-Alpine realm. Geologische Rundschau, 84, 457-465.

Neubauer, F., 2002. Evolution of Late Neoproterozoic to Early Palaeozoic tectonic elements in central and southern European Alpine mountain belt: review and syntesis. Tectonophysics, 352, 87-103.

Oberhänsli, R. Candan, O. Dora, O.Ö. and Dürr, St., 1997. Eclogites within the Menderes Crystalline Complex / western Turkey / Anatolia. Lithos, 41, 135-150.

_____, _______ and Wilke, W., 2009. Geochronologic Evidence of Pan-African Eclogites from the Menderes Massif, Turkey. Precambrian Research (submitted).

Ocshner, A., 1993. U-Pb geochronology of the Upper Proterozoic-Lower Palaeozoic geodynamic evolution in the Ossa-Morena Zone (SW Iberia): constraints on the on the timing of the Cadomian Orogeny. PhD Thesis no. 10392, ETH, Zurich Switzerland, 430.

Okay, A.I., 1984. Distribution and characteristics of the northwestern Turkish Blue schists. In: Dixon, J.E. ve Robertson, A.H.F. (eds). The Geological Evolution of the Eastern Mediterranean. Geological Society Special Publication 17, 455-466.

_____, 2001. Stratigraphic and metamorphic inversions in the central Menderes Massif: a new structural modal. International Journal of Earth Sciences, 89, 709-727.

_____ and Kelley, S.P., 1994. Tectonic setting, petrology and geochronology of jadeite + glaucophane and chloritoid + glaucophane schists from northwest Turkey. Journal of Metamorphic Geology, 12, 455-466.

_____, Bozkurt, E, Satýr, M., Yiðitbaþ, E., Crowley, Q.G. and Shang, C.K., 2008. Defining the southern margin of Avalonia in the Pontides: Geochronological data from the Late Proterozoic and Ordovician granitoids from NW Turkey. Tectonophysics, 461, 252-264.

Özer, S., Sözbilir, H., Özkar, I., Toker, V. and Sarý, B., 2001. Stratigraphy of Upper Cretaceous-Palaeogene sequences in the southern and eastern Menderes Massif (Western Turkey). International Journal of Earth Science, 89(4), 852-866.

_____ and _____, 2003. Presence and tectonic significance of Cretaceous rudist species in the so-called Permo-Carboniferous Göktepe Formation, central Menderes metamorphic massif, western Turkey. International Journal of Earth Science, 92 (3), 397-404.

Öztunalý, O., 1973. Petrology and geochronology of Uludað (NW Anatolia) and Eðrigöz (W Anatolia) massifs. Ýstanbul University Science Faculty Monograph, 23, 1-115.

Öztürk, A. and Koçyiðit, A., 1983. Menderes grubu kayalarýnýn temel-örtü iliþkisine yapýsal bir yak laþým (Selimiye-Muðla). Türkiye Jeoloji Bülteni, 26, 99-106.

Paquette, J.L., Nedelec, A., Monie, B. and Rakotondrazafy, M., 1994. U-Pb single zircon Pb-evaporation and Sm-Nd isotopic of granulitic domain in SE Madagascar. Journal of Geology, 102, 523-538.

Partzsch, J. H. Oberhänsli, R. Candan, O. and Warkus, F., 1998. The Menderes Massif, W-Turkey: A complex nappe pile recording 1.0 Ga of geological history. Third International Turkish Geology Symposium, Middle East Technical Universýty - Ankara, 281p.

Pupin, J.P., 1980. Zircon and granite petrology. Contributions Mineralogy Petrology, 73, 207-220.

_____ and Turko, G., 1974. Application a quelques roches endogenes du massif franco-italien de l'Argentera-Mercantour, d'une typologie originale du zircon accessoire. Etude comparative avec la methode des R.M.A. Bulletin Society For Mineral Cristallography. 97, 59-69.

Reischmann, T., 1998. Pre-Alpine origin of tectonic units from the metamorphic complex of Naxos, Greece, identified by single zircon Pb/Pb dating. Bull. of the Geol. Soc. of Greece, Proceedings of the 8th International Congress, Patras, XXXII/3, 101-111.

_____, Kröner, A., Todt, W., Dürr, S. and Þengör, A.M.C., 1991. Episodes of crustal growth in the menderes Massif, W Turkey, inferred from zircon dating. Terra Absracts, 3, 34p.

Ring, U. Gessner, K. Güngör, T. and Passcchier, C., 1999.The Menderes Massif of western Turkey and the Cycladic Massif in the Aegean-do they really correlate? Journal of Geological Society London, 155, 3-6.

_____, Kröner A. Buchwald, R. Toulkeridis, T. and Later, P., 2002. Shear zone patters and eclogite-facies metamorphism in the Mozambique belt of northern Malawi, east-central Africa: implication for assembly of Gondwana. Precambrian Research, 116, 19-56.

Satýr, M. and Friedrichsen, H., 1986. The origin and evolution of the Menderes Massif, W-Turkey: A rubidium/ strontium and oxygen isotope study. Geologische Rundschau, 75/3, 703-714.

Schuiling, R.D., 1962. On petrology, age and structure of the Menderes migmatite complex (SW - Turkey). Bulletin Mineral Research Exploration Institut,Turkey, 58, 71-84.

Shiraishi, E.D.J., Hiroi, Y., Fanning, C.M., Motoyoshi, Y. and Nakai, Y., 1994. Cambrian orogenic belt in east Antartica and Sri Lanka: Implications for Gondwana assembly. Journal Geology, 102, 47-65.

Stern, R.J., 1994. Arc assembly and continental collusion in the Proterozoic east African orogen: Implications for the consolidation of Gondwanaland. Annual Review of Earth and Planetary Sciences, 22, 319-351.

Þengör, A.M.C. and Yýlmaz, Y., 1981. Tethyan evolution of Turkey: a plate tectonic approach. Tectonophysics, 75: 181-241.

_____, Satýr, M. and Akkök, R., 1984. Timing of tectonic events in the Menderes Massif, western

Turkey. Implications for tectonic evolution and evidence for Pan-African basement in Turkey. Tectonics, 3(7), 693-707.

Teklay, M., Kröner, A., Mezger, K. and Oberhänsli, R., 1998. Geochemistry, Pb-Pb single zircon ages and Nd-Sr isotope composition of Precambrian rocks from southern and eastern Ethiopia: implications for crustal evalution in East Africa. Journal of African Earth Science, 26/2, 207-227.

Thomson, S.N. and Ring, U., 2006. Thermochronologic evaluation of postcollision extension in the Anatolide orogen, western Turkey. Tectonics, 25, TC3005, doi: 10.1029/2005TC001833.

Unrug, R., 1996. The assembly of Gondwanaland. Episodes, 19, 1-2, 11-20.

Ustaömer, P.A., 1999. Pre-Early Ordovician Cadomian arc-type granitoids, the Bolu Massif, West Pontides, northern Turkey: geochemical evidence. International Journal of Earth Science, 88, 2-12.

_____ and Kipman, E., 1998. An example for a Pre-Ordovician Arc magmatism from northern Turkey: geochemical investigation of Çaþurtepe Formation (Bolu, W Pontides). Bulletin Mineral Research Exploration Institute, 120, 61-77.

_____ and Rogers, G., 1999. The Bolu Massif: remnant of a pre-Early Ordovician active margin in the west Pontides, northern Turkey. Geological Magazine, 136(5), 579-592.

Ustaömer, P.A., Mundil, R. and Renne, P.R., 2005. U/Pb and Pb/Pb zircon ages for arc-related intrusions of the Bolu Massif (W Pontides, NW Turkey): evidence for Late Precambrian (Cadomian) age. Terra Nova, 17/3, 215-223.

_____, Ustaömer, T., Collins, A.S. and Robertson, A.H.F., 2009. Cadomian (Ediacaran-Cambrian) arc magmatism in the Bitlis Massif, SE Turkey: Magmatism along the developing northern margin of Gondwana. Tectonophysics, 473, 99-112.

Veevers, J.J., 2004. Gondwanaland from 650-500 Ma assembly through 320 Ma merger in Pangea to 185-100 Ma break up: supercontinental tectonics via stratigraphy and radimetric dating. Earth-Science Reviews, 68, 1-132.

Wilson, T.J., Grunow, A.M. and Hanson, R.E., 1997. Gondwana assembly: The view from southern Africa and East Gondwana. Journal of Geodynamics. 23, ¾, 263-286.

Yibas, B., Reimold, W.U., Armstrong, R., Koeberl, C., Anhaeusser, C.R. and Phillips, D., 2002. The tectonostratigraph, granitoid gechronology and geological evolution of the Precambrian of southern Ethiopia. Journal of African Earth Science, 34, 57-84.

Zelazniewicz, A., Dörr, W., Bylina, P., Franke, W., Haack, U., Heinisch, H., Schastok, J., Grandmontagne, K. and Kulicki, C., 2004. The eastern continuation of the Cadomian orogen: U-Pb zircon evidence from Saxo-Thuringian granitoids in south-western Poland and the northern Czech Republic. International Journal Earth Science, 93, 773-781.

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POLYMETAMORPHIC EVOLUTION OF THE PAN-AFRICAN BASEMENT AND PALAEOZOIC-EARLY TERTIARY COVER SERIES OF THE MENDERES MASSIF

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ABSTRACT.- The Menderes Massif exposing in the Western Anatolia substantially presents a complex tectono-stratigraphy as a result of Late Alpine compressional tectonism. The lithostratigraphical succession of this crystalline complex can be divided into two units: 1- The Pan-African basement (core series) and 2-Palaeozoic - Early Tertiary metasedimentary rocks (cover series). The Pan-African basement of the Menderes Massif is made up of a Late Neoproterozoic metaclastic sequence consisting of paragneisses and conformably overlying micaschists. This high-grade metaclastic sequence is extensively migmatized and intruded by the syn- to post-Pan-African gabbros and granitoids. The primary contact relationship between the core and cover series is a regional unconformity in character. The Palaeozoic (?Upper Devonian - Permian) cover units, which are cut by the intrusion of Triassic leucocratic metagranites are consisted of phyllites, quartzites and marbles. The Mesozoic cover units are characterized by Triassic to Upper Cretaceous platform-type thick marbles at lower levels of the sequence. Upper Campanian - Upper Maastrichtian pelagic carbonates and the overlying Middle Paleocene - Eocene flysch-type blocky unit constitute the uppermost units of cover series. Relic mineral assemblages observed in the Pan-African basement reveal a complex polyphase metamorphic evolution of this basement under granulite, eclogite and amphibolite-facies conditions. The high temperature metamorphism developing under granulite facies is characterized by the presence of hypersthene type pyroxene. Pelitic granulites, orthopyroxene gneisses, orthopyroxene paragneisses and metagabbroic / metanoritic rocks form typical granulite-facies relics observed in the massif. Geothermobarometric estimations characterize an average temperature of 730 °C and pressure of 6 kbar for the granulite-facies metamorphism. By means of SHRIMP II method, clustering ages of 583±5.7 Ma were dated from the outer parts zircons in pelitic granulites which have no zoning but have overgrown under granulite facies. High grade metamorphism relics in the Pan-African basement are characterized by eclogite and eclogitic metagabbros. Fully recrystallized, fine grained massif eclogites, with non bearing relic texture belonging to protolith are composed of 'omphacite (jd_{40-52}) + garnet + clinozoisite + amphibole + quartz + rutile'. However; relic texture and minerals are extensively observed in metagabbros derived from eclogitic gabbros. The pressure-temperature (P-T) conditions of the Pan-African high-pressure metamorphism were estimated as 644°C with a minimum pressure of about 15 kbar, which corresponds to a burial depth of about 50 km. $^{206}Pb/^{238}U$ zircon ages obtained from eclogitic metagabbros by TIMS yield 529.9±22 Ma, reveal the high-pressure metamorphism as Pan-African in age. The Barrowian type medium pressure metamorphism reaching up migmatization stage in which anatectic granites developed caused extensive retrogradations. Geothermobarometric estimates from garnet amphibolites, retrograded from eclogites indicate that this metamorphism developed under P/T conditions of 7 kbar in pressure and 628°C in temperature. The crystallization ages of these anatectic granites range from 551 to 540 Ma. They were generated by migmatization of paragneisses and reveal that this medium-pressure event is related to the last stage of polyphase Pan-African metamorphism. All metamorphic ages obtained from the Pan-African basement are compatible with the latest stages of assemblage of Gondwana super continent. It is considered that protoliths of paragneisses and schists of the Pan-African basement were deposited on a passive continental margin of a basin occurring between East and West Gondwana during the Late Proterozoic time (Mozambique Ocean). The Pan-African basement of the Menderes Massif was deeply buried and metamorphosed under granulite, eclogite and amphibolite-facies conditions as a result of the closure of this basin and collision of East and West Gondwana during Late Neoproterozoic time. Both core and cover series of the Menderes Massif were affected by an Alpine aged old regional metamorphism. In the Palaeozoic sequence of

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cover series, this metamorphism is characterized by a Barrowian type medium-pressure metamorphism. This metamorphism developed under greenschist to lower amphibolite-facies (6 kbar in pressure / 430-550 °C in temperature) and described by the occurrences of garnet, staurolite and kyanite (disthene) in phyllites. Mesozoic-Early Tertiary cover series at the southern part of Çine submassif contain data associated with Alpine aged HP/LT metamorphism. Carpholite-kyanite assemblage within Triassic quartz metaconglomerates shows a metamorphism under a pressure of 10-12 kbar and temperature of 440 °C corresponding to a minimum depth of 30 km. So far, there has not been detected any data for an Alpine HP / LT metamorphism, neither in the Pan-African basement nor in the Palaeozoic sequence of the Menderes Massif. Based on the fossil content obtained from youngest unit of the cover series and from the oldest non metamorphic sedimentary cover on the massif, the Alpine metamorphism can biostratigraphically be constrained into Eocene and Oligocene time interval. Few isotopic data (37±1 Ma, Late Eocene Rb/Sr biotite age; 36±2 Ma, Middle Eocene Ar/Ar muscovite age; 43-37 Ma, Eocene Ar/Ar muscovite age) related to Alpine metamorphism are compatible with the related time interval. The Alpine metamorphisms of tectonical zones belonging to Anatolides is substantially associated with the closure of the northern branch of Neothethys Ocean and with the collision in Paleogene. In such a tectonic model, the segment of the Anatolide-Tauride platform corresponding to the Menderes Massif was subjected to intense internal imbrication during the subduction process of the northern branch of Neotethys and the following period in which the continental collision occurred. The tectonical slices being formed were buried at different depths and metamorphosed under varying conditions related with burial depths under the load of Afyon zone at north in Eocene-Oligocene times, and of Lycian nappes passing south and of ophiolites.

Key words: Menderes Massif, Pan-African Orogeny, metamorphism, granulite, eclogite, carpholite.

INTRODUCTION

NE-SW trending the Menderes Massif crops out in Western Anatolia pervasive and is one of the best studied crystalline regions of Turkey. This crystalline complex is tectonically overlain by Lycian nappes at south, by Izmir-Ankara suture zone at north and northwest and by the extension of Cycladic complex in Turkey. This massif is constrained by low grade metamorphics belonging to Afyon zone at the north and covered by the Neogene sedimentary /volcanic units at the east. E-W trending young graben systems divide the Menderes Massif into three sub massifs as; Demirci Gördes submassif (northern submassif), Ödemiþ-Kiraz submassif (central submassif) and Çine submassif (southern submassif) (Figure 1).

For many years, it has been considered by many investigators that the Menderes Massif is made up of a simple and uniform stratigraphy by Precambrian core and by the surrounding Palaeozoic-Early Tertiary cover series all around (Schuiling, 1962). However, recent studies have clearly revealed that thrust faults produced by the Late Alpine compressional tectonic have largely reshaped the primary stratigraphy of the Massif (Konak et al., 1994; Partzsch et al., 1998; Ring et al., 1999; Gökten et al., 2001). The primary contact relationship between the core and cover series can today be observed in very rare areas (Pan-African unconformity; Þengör et al., 1984; Konak et al., 1987; Dora et al., 2005; Candan et al., 2006). This contact relationship between these units is defined by thrust faults in many places. The stratigraphy of core and cover series at present have largely been brought to light by the determination of tectonostratigraphies of these tectonic units and by its correlation with each other (Konak et al., 1987; Dora et al., 2001, 2005). The general distribution of core and cover series in massif scale so called as the "Pan-African basement" and also as the "Precambrian basement" is shown in figure 1.

The overall metamorphic structure of the Menderes Massif was acquired by the Alpine age latest metamorphism which is also named as the Main Menderes Metamorphism and this opinion has generally been accepted by many investigators (Þengör et al., 1984). However, in recent years many new evidences has been obtained showing that the Menderes Massif has a more

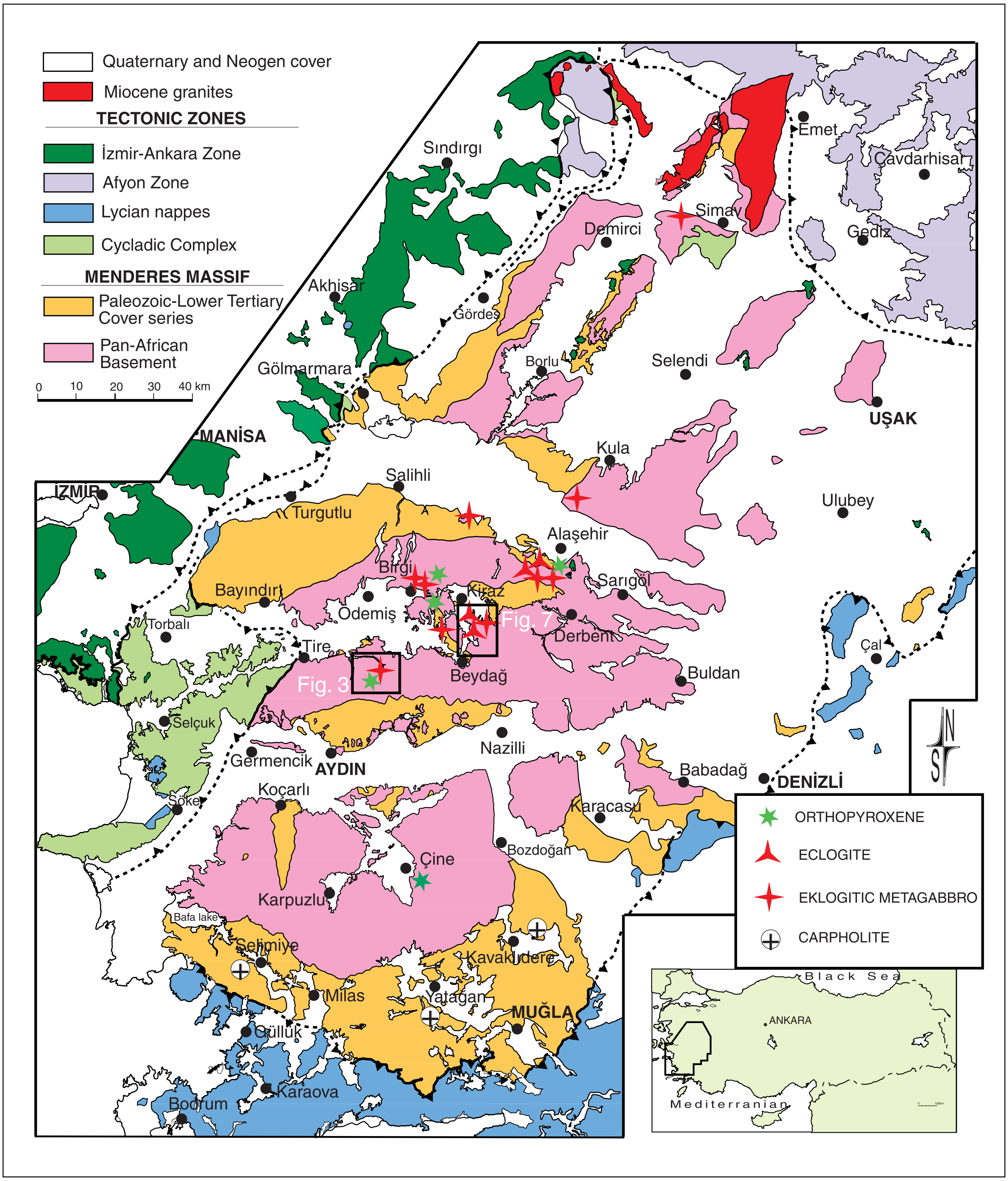

Figure 1- The distribution of the Pan-African Basement and Palaeozoic-Early Tertiary cover series throughout the Menderes Massif. Locations where relics of Pan-African aged granulite-eclogites facies metamorphism and Alpine age high pressure metamorphism are shown on map.

complex metamorphic history than had ever been considered. Petrological evidences indicate that the Precambrian basement in the Menderes Massif was affected by the polyphase metamorphism associated with the Pan-African orogeny under granulite, eclogite and amphibolite facies (Candan and Dora, 1998; Candan et al., 1994, 1995, 2001, 2007; Oberhänsli et al., 1995*a-b*, 1997). However in cover series, new evidences have been obtained defining HP / LT metamorphism conditions in addition to widespread Barrowian type medium pressure metamorphism assemblages in cover series (Rimmelé et al., 2002, 2003). In this article, the Pan-African basement and poly metamorphic evolutions of Palaeozoic-Early Tertiary cover series of the Menderes Massif has been reevaluated. The article prepared within this context mainly aims at presenting and discussing the data on; 1- petrographical / petrological properties, ages of polyphase metamorphism observed in the Pan-African basement of the Menderes Massif, and its relationships with the Pan-African orogeny and 2- the poly metamorphism of cover series and temporal and spatial relationships of these series with the evolution of Neothethys Ocean.

THE LITHOSTRATIGRAPHY OF THE MENDERES MASSIF

Recent studies have largely established that the massif has lost its original stratigraphy by the Alpine age compressional tectonism (Konak et al., 1994; Partzsch et al., 1998; Ring et al., 1999; Gökten et al., 2001). At present, units that belong to the Pan-African basement are observed as tectonical slices presenting imbrications with Palaeozoic-Early Tertiary cover series. The generalized columnar section of the Menderes Massif obtained by the tectonostratigraphy of tectonic slices and its correlation with each other is given in figure 2. As seen in the figure, the stratigraphy of the Menderes Massif is divided into two main units as; 1- Late Neoproterozoic Pan-African basement and 2- Palaeozoic-Early Tertiary cover series. The primary contact between these two units is in character of regional unconformity defining a deep erosion (Pan-African unconformity; Þengör et al., 1984; Konak et al., 1987; Candan et al., 2006).

The oldest rock units of the Pan-African basement are composed of metaclastics forming a regular and continuous sequence. Related metaclastic sequences are cut by granitoid and gabbroic rocks that have intruded at various stages of Pan-African orogeny. This metaclastic sequence has a minimum thickness of 8 km and is divided into two units as; i) paragneiss and ii) schist (from bottom to top) (Dora et al., 2001). The paragneiss unit forming the lower unit of metaclastic sequence is composed of two lithologies showing both lateral and vertical transitions. The dominant lithology is paragneiss derived from litarenitic sandstones and containing sillimanite (± orthopyroxene). The schists rich in garnet and sillimanite which show lateral and vertical transitions with paragneisses form the other lithology. Although there are many transitional type lithologies, the mica schist and biotite albite schists derived from the mudstone and sub arkosic sandstones form dominant schist types. The true thickness of the paragneiss is not known as its lower levels are cut by thrust faults and / or it includes granitic inclusions. However, 4 km the apparent thickness of a slice observed in Kula region is accepted as the maximum thickness of paragneiss unit in the Massif. Paragneiss unit is conformably and transitionally overlain by the schist unit. Schist unit is dominantly composed of mica schist and biotite albite schist intercalations of which its protoliths correspond to mudstone and sub arkose. Probably at the upper levels of these schists, black quartzite layers that have a thickness not exceeding 0.5 m are rarely observed. On the other hand, these schists are conformably and transitionally overlain by a deposit made up of muscovite schist / biotite, muscovite, quartz schist intercalation located at the southern part of Çine submassif. These schists show an approximate thickness of 2 km and present probably the uppermost levels of the schist unit belonging to the Pan-African basement. Within schist unit,

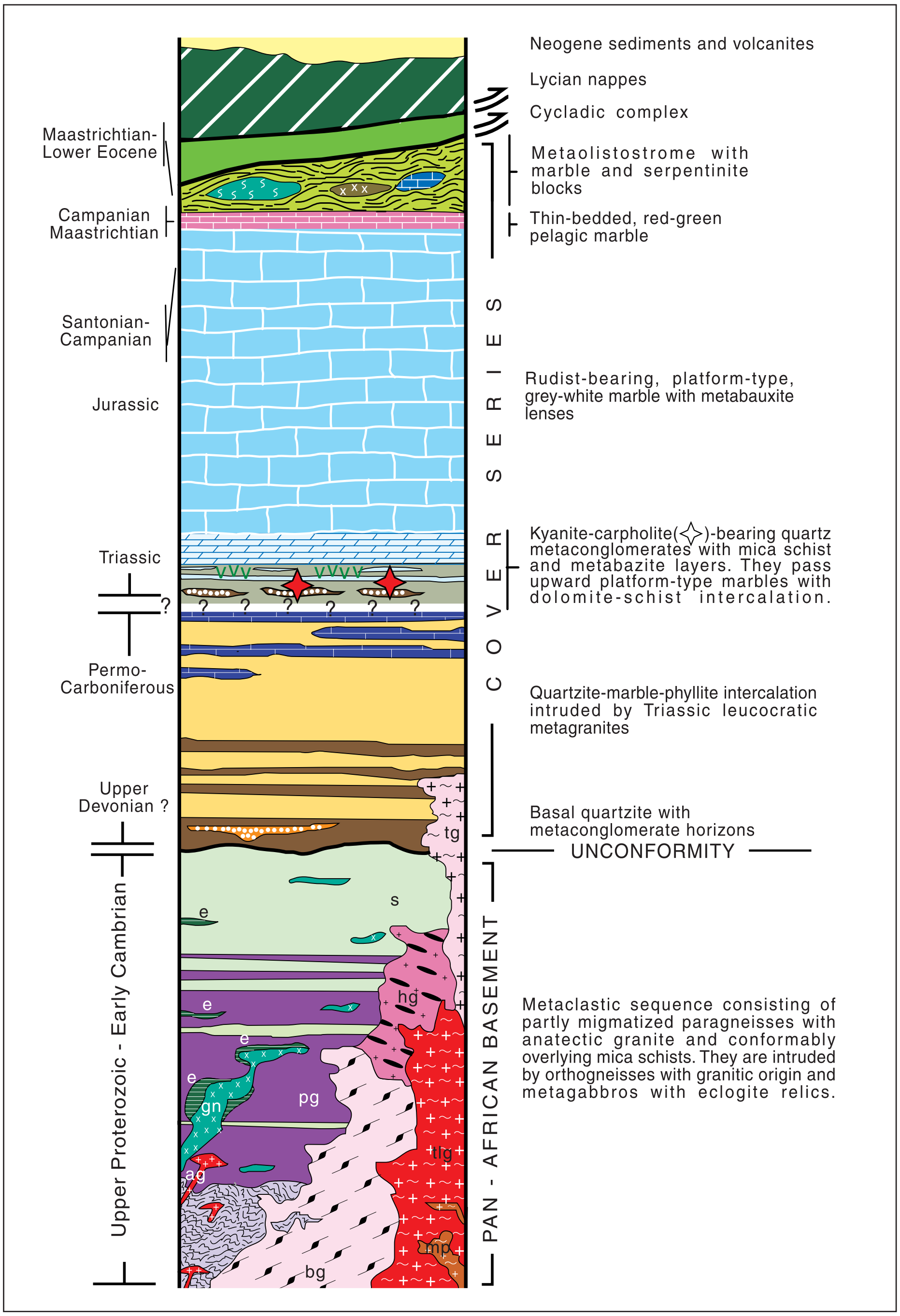

Figure 2- Generalized columnar section of the Menderes Massif (gn:Gabbro-norite, ag:Anatectic granite, tg:Triassic leucocratic orthogneiss, hg:Hornblende orthogneiss, bg:Biotite orthogneiss, tlg:Tourmaline leucocratic orthogneiss, pg:Paragneiss, s:Schist, e:Eclogite, modified from Dora et al., 2001).

dolomitic marble lenses can very rarely be observed with a dimension of 80x200 m. Paragneiss unit forms the lowermost layers of the metaclastic sequence and show a widespread migmatization throughout the Massif. These migmatites are accompanied by anatectic granites which were produced by partial melting in many places.

Both schist and paragneiss units are cut by numerous basic magmatic rocks with a character of stock and vein reaching up to a 1.5 km in dimension. These Precambrian / Cambrian rocks are composed of biotite gabbro, olivine gabbro, noritic gabbro, norite and display extensive transformations into eclogite and amphibolites along their peripheral zones (Candan et al., 2001). The orthogneisses made up of plutons that have intruded into each other are granitic in origin and form one of the most widespread rock type belonging to the Pan-African basement. Orthogneisses in the massif can be divided into many sub-types like, biotite orthogneiss, tourmaline leucocratic orthogneiss, amphibole orthogneisses, metagranite porphyry, albite and metaaplitic vein rocks rich in quartz in granoblastic textures by basing on the texture and mineralogical composition of the primary granites (Bozkurt, 2004; Dora et al., 2005). These orthogneiss types defining a Precambrian / Cambrian aged (varying between 570-520 Ma; average 550 Ma; Loos and Reischmann, 1999) acidic magmatic activity, present clear intrusive contact relationship with metaclastic sequence as the oldest rock of the Pan-African basement.

The Pan-African basement is overlain by Palaeozoic units with an unconformable contact. Palaeozoic sequence in Late Devonian(?) - Permo Carboniferous age (Çaðlayan et al., 1980; Konak et al., 1987) begin with muscovite-quartz schists derived from pure quartz arenite defining an unconformity at lowermost layers. These rocks reach a thickness of 1.5 km and meta-conglomerates in the form of channel fillings are observed close to the basement. These conglomerates can laterally be traced 35 km in the form of discontinuous exposures. The components of these conglomerates are composed of granite, aplite and of tourmaline pebbles in various mineralogical compositions which were derived from the Pan-African basement of the Massif. The quartzite is conformably and transitionally overlain by black colored phyllites. In these rocks bearing garnet, chloritoid, staurolite and disthene, there are gray to black colored marble layers. Fossil finding indicate that this black sequence is Permo-Carboniferous in age (Konak et al., 1987). Palaeozoic cover series are cut by leucocratic metagranites rich in quartz-feldspar minerals and rarely bear biotite. These rocks are stock and sill in character reaching a dimension of 5-6 km. Their crystallization ages were determined as 241-236 Ma (Early to Middle Triassic) by Pb/Pb method (Koralay et al., 2001). The Mesozoic units of cover series begin with a mica schist layer having a thickness of 200 m and bears meta conglomeratic channel fills and basic additives. The primary contact relationship of this assemblage of probably Upper Triassic with the Palaeozoic cover series is interpreted as a regional unconformity (Konak et al., 1987; Erdoðan and Güngör, 2004). Meta-conglomerates completely made up of quartz pebbles include disthene-chloritoid schist interbeddings derived from clay rich in aluminum and do not exceed a couple of meters in thickness. In quartz veins of pebbles, carpholite-disthene interval is observed as the product of Alpine age high pressure metamorphism Rimméle et al., 2003). Meta-conglomerate schist intercalation is conformably and transitionally overlain by platform type thick carbonates. The transition zone is defined by pink-yellow dolomite-quartzite-calc schist intercalations. This intercalation has a thickness of 50 m and overlain by gray colored massive dolomites. Meta bauxite layers are observed 150 meters above probable Upper Triassic-Lower Jurassic dolomitic carbonates. They are located in marbles with lense shaped and have a lateral continuity reaching tens of kilometers. Meta bauxites are repeated several times in massive marbles which reach a thickness of 1500 meter (Milas formation). There is not any clear data if these meta bauxites are different layers or tectonically repeated. Well preserved rudist fossils are widely observed 50-400 meters above the meta bauxite layer in the massive. These layers are Santonian-Campanian in age and can laterally be traced by tens of kilometers (Özer et al., 2001). Rudistic layers upwardly grade into intra formational limestones and cherty marbles. Cherty marbles are overlain by Upper Campanian-Upper Maastrichtian red pelagic marbles (Kýzýlaðaç formation) which transitionally reach a thickness of 150 meters. Kazýklý formation is made up of a sandstone and shale derived matrix and marble blocks, forms the uppermost unit of the Menderes Massif. The contact relationship of this unit with red marbles are still in discuss and ages varying between Middle Paleocene to Lower-Middle Eocene were obtained (Konak et al., 1987; Özer et al., 2001).

THE METAMORPHISM OF THE PAN-AFRICAN BASEMENT

Recent evidences support the idea that main metamorphic effect shaping the Pan-African basement of the Menderes Massif is associated with the Pan-African orogeny and the Alpine metamorphism caused limited retrogradations on this basement. Petrological / petrographical and geochronological data indicate that poly phase metamorphism of the basement occurred successively under granulite, eclogite and upper amphibolite facies conditions.

THE METAMORPHISM OF GRANULITE FACIES

Petrography and mineral chemistry

Relic mineral clusters and rocks related to high temperature metamorphism under granulite facies affecting the Pan-African basement are very rarely observed in the Massif (Figure 1). Great majority of these evidences were obtained from Ödemiþ-Kiraz submassif. Data defining the related metamorphism can be divided into four groups. These are; 1- Pelitic granulites, 2- Gneisses with orthopyroxene, 3- Orthopyroxene paragneisses, 4- Metagabbro / metatonalites (Candan, 1995).

The region where pelitic granulites are best observed is the southeast of Tire, around Küre village (Figure 3). The region is made up of two tectonical units belonging to the Pan-African basement (Çetinkaplan, 1995; Candan and Çetinkaplan, 2001). The lower slice is composed of biotite and garnet mica schist having a homogenous composition. The upper slice in the form of clippe and has a dimension of 6x4 km and contains all data related to a poly-metamorphic evolution of the Pan-African basement. Pelitic granulites crop out in a 1x1 km region within a clippe. Granulites are grayish rocks in massive structure. These medium to fine crystallized rocks in granoblastic texture contain black colored orthopyroxene crystals reaching 0.5 cm dimensions. Widespread calcsilicate rocks that show strong boudinage are observed within granulites. These calcsilicates are one the most characteristic lithologies that belong to the Pan-African basement and their presence indicate that paragneisses transformed into pelitic granulites in some places. Textural data reveal that these rocks have a complex metamorphic evolution. The mineral assemblage of granulite facies is composed of 'quartz + plagioclase + orthoclase + orthopyroxene + biotite + garnet +cordierite (?) + ilmenite + rutile'. Granoblastic polygonal texture with a grain boundary of 120° in angle and straight boundaries are observed defining the high temperature conditions in quartz and feldspars in these rocks (Figure 4a). Plagioclases in An_{29-31} composition contain antipertitic structures specialized to granulites. Garnets of granulitic stage are in anhedral and in the form of porphyroblasts with many quartz inclusions. Compositional homogenization developed in garnets because of the volumetric diffusion that occurred under high temperature conditions. Garnets of this stage are distinctively poor by spessartin and grossular end members but substantially rich in almadine and pyrop. The end member components of these garnets have

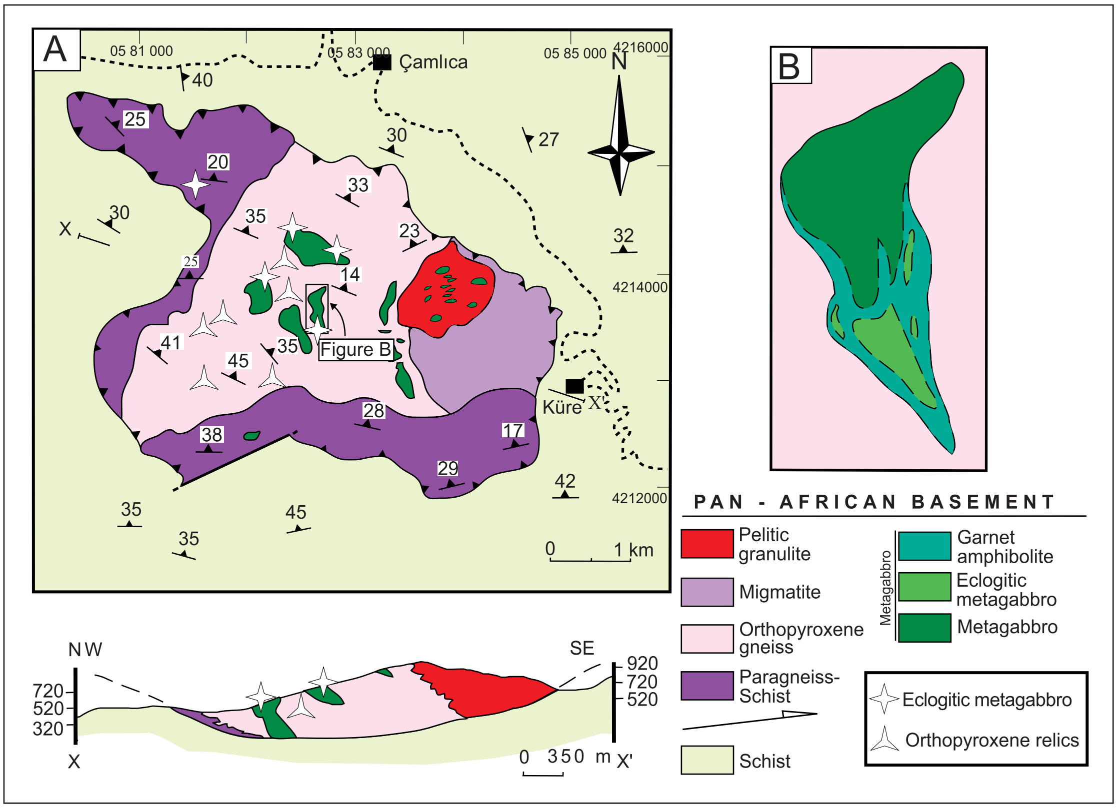

Figure 3- Geological map of Küre village and its surround, southeast of Tire. The region is made up of clippe and underlying low graded schists. Clippe is made up of high graded rocks where assemblages of granulite, eclogite and amphibolite facies belonging to the Pan-African Basement are observed. Region is shown is figure 1 (modified from Çetinkaplan, 1985).

been detected as 'Grs$_{0-1}$ - Alm$_{68-72}$ - Prp$_{21-24}$ - Sps$_{2-3}$ - And$_{1-2}$. The most characteristic mineral of pelitic granulites is orthopyroxene. FeO content is approximately %29 and MgO is %16 in equidimensional crystals observed granoblastic textures. These minerals are hypersthene in composition (Figure 5a) and have high Al_2O_3 amount specialized to orthopyroxenes in granulite facies. The amount of Al_2O_3 in minerals varies between 3.4%-3.8%. The metamorphic products that overlie pelitic granulites include pervasive retrogradation data. The most characteristics of these are; 1- garnet coronae developing around mafic phases, 2- replacement of orthopyroxene by biotite, 3- consumption of cordierites by high temperature ensemble, 4- replacement of biotites by sillimanite. When occurring phases are considered, it is understood that the overlying metamorphism had taken place under the conditions of upper amphibolite facies. Garnet corona developed into feldspar, along contacts of feldspars with orthopyroxene, biotite, rutile, ilmenite and garnet (Figure 4b). These garnets are rich in grossular composition and gradual increase in CaO towards plagioclase. In garnet corona the textural zoning is observed in addition to combined zoning. In general, there is a narrow intermediate zone made up of quartz between mafic phases and garnet corona. At contacts of orthopyroxene-biotite partial reaction zones developed in a way that expresses an inequality on the overlying metamorphism. Biotite porphyroblasts belonging to granulitic stage along these zones recrystallize in the form of finger type biotite - plagioclase syn - growth. However, accessory mineral which is the orthopyroxene is consumed by fine grained biotites (Figure 4c). Similar to those in paragneisses diffuse pseudomorphic replacement textures belonging to a former mineral are observed in pelitic granulites. This mineral was interpreted as cordierite based on general properties (Dora et al., 2001). These mineral holes have been filled by a high temperature ensemble made up of 'sillimanite / disthene - garnet - biotite - quartz' (Figure 4d). In addition to cordierites, biotite porphyroblasts which are the granulitic stage products are also consumed pervasively by sillimanite. The new mineral group is made up of sillimanite and of small biotites.

The great majority of orthopyroxene paragneisses were recognized in Ödemiþ-Kiraz submassif (Figure 1). Except the location at south of Alaþehir, all orthopyroxene paragneisses in the massif are present as intercalations with dimensions reaching several hundreds of meters. These paragneisses are dark gray in color and fine grained massive rocks. As for the rocks at south of Alaþehir are composed of migmatites as dark brown fine grained massive rocks. Besides, the widespread presence of black speckles is known in paragneisses throughout the Massif. These speckles are made by the mineral clusters of upper amphibolite facies and are considered as the cordierites which are the metamorphism product of granulite facies that affect paragneisses (Dora et al., 2001). The general mineral assemblages of paragneisses are "quartz + plagioclase + orthoclase + orthopyroxene + biotite + ilmenite / rutile". The consumption of biotites by sillimanite, the replacement of orthopyroxenes by biotite, the development of garnet ring around orthopyroxene are the main textural evidences related to retrogradation in these rocks under amphibolite conditions (Figure 4e).

Orthopyroxene gneisses are observed in Tire and Birgi regions in the Massif (Figure 1, Figure 3). These rocks were derived from coarse crystalline porphyritic granites. It is almost impossible to distinguish these rocks in the field from gneisses in granitic origin which have intruded at last and the post stage of the Pan-African Orogeny. However, the presences of greenish mineral dwellings rich in sillimanite and brown-black colored orthopyroxene speckles with a dimension of 2-3 mm are the main distinguishing feature of orthopyroxene gneisses. The exposure in Birgi region extends between Birgi and Cevizalaný village. Various textural features are recognized in this unit. Especially, samples that are much darker in color, coarse crystalline, equally sized granoblastic in texture, are rich in orthopyroxene. The primary granites of orthopyroxene gneisses in Tire region are in general porphyritic textured rocks including orthoclase porphyroblasts. In ductile deformed zones widespread retrogradations into blasto mylonites are observed. Intercalations of paragneiss with orthopyroxene that reach several hundreds of meters in dimension are extremely widespread within the gneisses.

The granulite facies mineral assemblage of orthopyroxene gneisses is made up of 'plagioclase + orthoclase + biotite + orthopyroxene + garnet + ilmenite / rutile' (Figure 4f). Orthopyroxenes are in composition of ferric hypersthene (En_{43-47}) (Figure 5a). Formations of garnet corona similar to pelitic granulites are widely observed in gneisses as well. Garnet coronae around biotite are divided into two zones. The inner zone is defined by plentiful quartz inclusions. Non inclusive outer zone is made up of euhedral garnet crystals advancing through plagioclase. Especially in mylonitic zones, the orthopyroxenes in shear zones are completely replaced by fine grained biotite cluster. These retrogradations make the discrimination of granulitic gneisses impossible from post Pan-African gneisses. The extension of sillimanite crystals parallel to shear zones which give rise to retrogradations indicate that these minerals belong to stage at upper amphibolite facies.

Metagabbro / Metanorites is another lithology which belongs to the Pan- African basement. The magmatic texture and mineralogy in these rocks

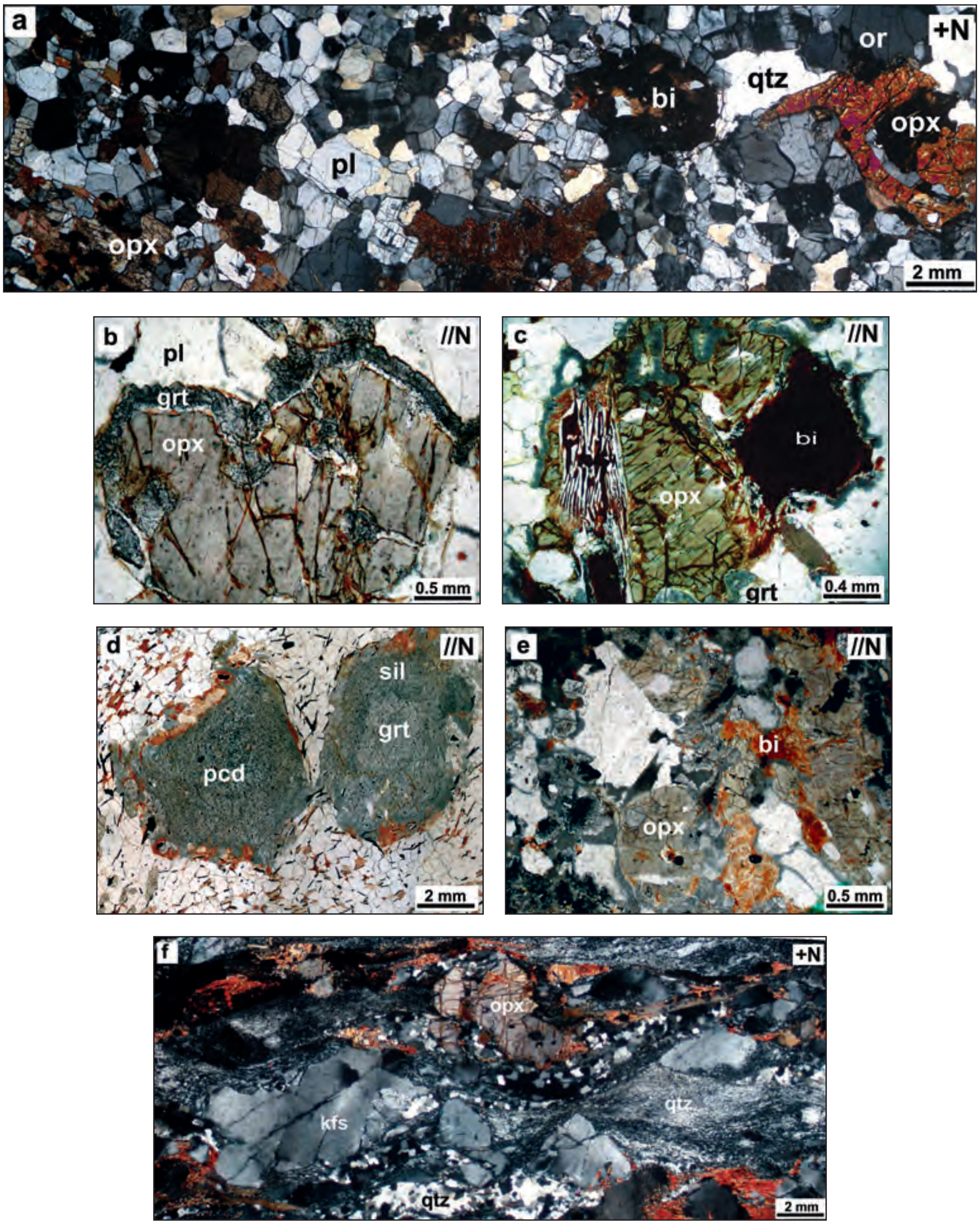

Figure 4- Microscopic views of granulites in the Menderes Massif (Tire / Küre). a: general view of pelitic granulites showing granoblastic polygonal texture and are composed of orthopyroxene, garnet, plagioclase and quartz, b: Garnet rings developing at contacts of orthopyroxene - plagioclase, c: Reaction zones developing at contacts of orthopyroxene - biotite. Orthopyroxenes are consumed by new, small biotites while coarse biotite crystals transforms into biotite - plagioclase symplectic growths, d: Assemblage of sillimanite, garnet, biotite and quartz replacing ?cordierite porphyroblasts belonging to granulite facies (pcd= pseudocordierite), e: the view of fine grained, granoblastic in texture orthopyroxene para-gneisses, f: The mylonitic texture in orthopyroxene gneisses (opx: orthopyroxene, pl: plagioclase, bi: biotite, or: orthoclase, qtz: quartz, grt: garnet, sil: sillimanite, kfs: K- feldspar).

are extremely well preserved and it is macroscopically impossible to examine the granulite facies evidences. Multi ring corona structures that developed at contacts of plagioclase / olivine were interpreted as the effects of granulite facies metamorphism over these rocks (Candan, 1995). The inner ring developing into olivine in these formations is made up of orthopyroxene. However, the outer ring advancing into plagioclase is formed by the spinal - hornblende symplectic growth. In some samples partial garnet ring can also develop at the outermost.

Pressure - temperature conditions

Miscellaneous lithologies in the Pan-African basement include ensembles of similar granulite facies. Classical geothermobarometric calibrations were applied in these ensembles. The geothermometer based on the Fe-Mg change between garnet / orthopyroxene suggested by Lal (1993) gave an average temperature of 730± 20 °C for pelitic granulites (Candan et al., 1998; Dora et al., 2001). Applying of the same geothermometer to orthopyroxene paragneisses around Çine an average temperature of 756 °C has been estimated. 12 different calibrations were applied to biotite - garnet pairs in same rocks. These estimations have shown similarity with values obtained from garnet / orthopyroxene pairs giving an average temperature of 714°C. During assumptions of pressure conditions of the granulite facies metamorphism, the calibration suggested by Newton and Perkins (1982) was used based on the 'garnet + plagioclase + orthopyroxene + quartz' ensembles. This geobarometer gave a pressure value of 6.1 kbar.

The age of metamorphism

The preliminary attitudes regarding the relative age of granulite facies metamorphism affecting the Pan-African basement have been based on the basic geological properties of the Massif (Candan, 1995). Granulite facies metamorphism was interpreted as a probable Precambrian / Cambrian aged event based on data such as; 1- The observation of the evidences related to this metamorphism only in some units that belong to Pan-African basement, and the case that the degree of metamorphism in cover series has not reached to this level in any place. 2- Retrogradation of the ensembles related to granulite and eclogites by an overlying metamorphism under the conditions of upper amphibolite facies most probably associated with widespread migmatization observed in the basement series of the Massif. 3- It has been interpreted that the granulite metamorphism is an Precambrian/ Cambrian event affected only the core series of the massif by basing on the lackness of data approving thet orthogneisses with an average age of 550 Ma intruded at a stage following the Pan-African Orogeny, and contrarily by including the rocks experienced from the granulite metamorphism as intercalations (Candan, 1995; Candan and Dora, 1998).

The first study to determine the age of high temperature metamorphism by radiometric methods has been made on gneisses with orthopyroxene belonging to granulite facies which was observed in Birgi region. U, Th and Pb concentrations of monozites which were picked over this rock were estimated by EMS and the metamorphism age was determined as 660 +61/-63 Ma (Oelsner et al., 1997; Warkus et al., 1998). Although there is a big error range in the estimation this dating is important as this is compatible with the suggested Pan-African age based on the basic geological data. In recent years, zircon was picked ovre pelitic granulites in Tire region and ion micro probe (SHRIMP II) method was applied to these in order to detect the age of this metamorphism (Koralay et al., 2006). Cathodoluminescence (CL) photos of zircons picked over these granulites showed that unzoned zircon overgrowths belonging to granulite facies are surrounded by planar zoned outer zones indicating crystallization from an anatectic melt. The U-Pb ion microprobe (SHRIMP II) analysis of unzoned overgrowths were clustered at 583.0±5.7 Ma and this late Neoproterozoic age has been interpreted as new zircon overgrowths occurred during the granulite facies metamorphism related with the Pan-African Orogeny (Koralay et al., 2006).

ECLOGITE FACIES METAMORPHISM

Petrography and Mineralogy

The basic properties of high pressure metamorphism affecting the Pan-African basement of the Menderes Massif have been established by studies that have continued nearly 15 years (Candan, 1994, 1996*a-b*; Candan and Dora, 1998; Candan et al., 1994, 2001; Oberhänsli et al., 1997, 1998). Field and petrographical/petrological data have shown that high pressure metamorphism rocks within the Pan-African basement of the Massif could be evaluated in two groups. These are; 1- Eclogitic metagabbro in which relic texture and phases could be observed belonging to uncompleted reactions and primary magmatism that are closely associated with metagabbros, and 2- Eclogites that do not bear relic texture and mineral related to protoliths and fully recrystallized.

The eclogitic metagabbros has been determined in more than 30 locations in the northern and central submassives of the Massif (Figure 1). However, almost no data have been obtained from Çine submassif. The most typical and well preserved outcrops are observed in the southeast of Tire, around Küre village (Figure 3), west of Birgi, and around Keþat region. There is a metagabbro stock in Keþat region which has intruded into orthogneiss with a dimension of 1.5x0.4 km. This stock is surrounded by a garnet amphibolite circumference zone not exceeding 20 m of thickness. In many locations within amphibolites, anhedral, partly preserved eclogitic metagabbro relics with dimensions varying in between 20-30 cm to 10-15 cm are present. In addition to these, relic high pressure rocks are frequently recognized in internal shear zones which cut metagabbro. The transformation of gabbro-coronitic metagabbro-eclogitic metagabbro occurs in intermediate areas characterized by low stress. In these intermediate zones primary magmatic phases are statically replaced by high pressure minerals in a way to preserve the original texture (Figure 6a). The tectonic slice in Tire/Küre region where the best preserved exposures of granulites in the Massif are observed, includes numerous metagabbro / metanorite stocks. Many of these stocks are surrounded by partly developed amphibolite zones. The width of eclogitic metagabbro zones exceed 30 m in the region where gabbro-eclogitic metagabbro-garnet amphibolite transformations could be observed with all textural / mineralogical intermediate terms belonging to prograding and retrograding metamorphisms. Even in eclogitic metagabbro samples that have been metamorphosed at the highest grade, there could still be observed relic magmatic phases. Eclogitic metagabbro- garnet amphibolite metamorphisms follow the late stage shear zones which are more effective especially at the outermost parts of masses. Apart from these regions, it was determined that gabbroic stock and vein rocks at south of Alaþehir in Kestane Deresi area and at south of Kula to the north of Yahyaalcý village have been metamorphosed into metagabbros. The eclogitic metagabbro exposure recognized at the northernmost part of the Massif takes place at Simav/south of Beyceköy. In this mass with dimensions of 20 x 5 m located in schists with disthene that show heavy migmatization, the high grade garnet amphibolite metamorphism is observed.

During thin section studies, it has been clearly detected that gabbro-eclogitic gabbro transformation occurred by gradual replacement of plagioclase and augitic clinopyroxene by garnet and omphacite. This transformation begins with a coronitic stage and ends by complete replacement of the phases. The first stage is described by the development of garnet rings located at contacts of clinopyroxene, biotite and ilmenite with plagioclases (Figure 6b). Rings made up of euhedral garnet ensembles progress into plagioclase. There is always observed chemical zoning in garnets that the highest CaO content would be towards plagioclase. While the grossular component would increase from 20% to 22%, the pyrope component would decrease from 28% to 25%. Plagioclases also get a cloudy view by inclusions made up of diffuse clinozoisite and by very rare disthene crystals. Omphacite transformation occurs in two different ways. Especially in areas of low tensile where static recrystallization occurs, the magmatic clinopyroxene would turn into omphacite in a way to preserve its primary crystal form. In this formation, the primary pyroxene show an omphacitic transformation in a way that jadeite component would decrease from outer to inner parts. The jadeite component of Na-pyroxene formed by this mechanism is directly proportional to the amount of reactions formed which is proportional to the consumption amount of plagioclase in gabbro. Clinopyroxenes in the original rock present an augitic composition possessing 1.3% moles of Jd component which contains approximately 0.48% Na_2O. As a result of this transformation, the jadeite component of clinopyroxene possessing Na-Aguite component reaches 10% (Figure 5b). The second type of transformation into omphacite realizes by the replacement of primary clinopyroxene by small mineral assemblage made up of omphacite. This type of polygonal crystals of which the Jadeite component reaches 23% presents a composition varying between Na-augite- omphacite (Figure 5b). In further stages of the transformation into eclogitic metagabbro individual garnet dwellings occur in plagioclases. These minerals are also replaced by the ensemble made up of anhedral garnets in a way to preserve their primary crystal forms (Figure 6c). These garnets possess a homogenous chemical composition and their average composition is 'Alm_{60} - Prp_{14} - $Sps_{1.5}$ - Grs_{24} - $And_{0.5}$'. Even in furthermost stages of the transformation, it is possible to observe relic plagioclases in eclogitic metagabbros.

Eclogites are rarely observed high pressure rocks at the Pan-African Basement. These are medium to fine grained, massive structured and are made up of red single garnet crystals that presents a homogenous dispersion over green groundmass composed of omphacite (Figure 6d). Dimensions of these lens shaped masses vary between 5 - 400 meters and are placed in paragneiss and schist units. Eclogites were recognized in two locations in the Massif. 12 different masses were detected in Kestane Deresi area, at south of Alaþehir. The height of these rocks does not exceed 20 meters and are in the form of lensoidic masses. The masses observed within mica schist unit are in dimensions of 80 x 5 meters and are planar rocks at south of Kiraz, the north of Yeniþehir Village. The largest one of the 21 various exposures has a dimension of 400 x 200 meters (Figure 7). The inner sides of the mass are well preserved and are transformed into garnet amphibolites along shear zones of circumferences of the mass (Figure 6e). In this transformation omphacites are consumed by symplectically growing augite-plagioclase ensemble (Figure 6f) and garnets are gnawn by plagioclase rings.

These rocks posses a homogenous composition and are composed of 'omphacite + garnet + ruilte + quartz + clinozoisite + amphibole' (Figure 6g). Needle like omphacite crystals have a jadeitic composition varying between 42-50% and are composed of omphacite (Figure 5b). Euhedral garnet crystals have several inclusions composed of Ca-amphibole that belong to former medium pressure metamorphism stage. Euhedral garnets show definite chemical zoning. While the grossular compound increases from 22% to 25% towards circumference, the pyrope compound decreases from 22% to 16%. Amphiboles in eclogites are texturally in equal with pyroxenes. Na_2O content of these amphiboles vary between 4-5% and are composed of barroicite / magneso-catoporite.

Similar retrogradation textures are exhibited both in eclogites and in eclogitic metagabbros. Omphacites are consumed by the assemblage made up of clinopyroxene (Jd_{14}) and plagioclase (An_{8-11}) which presents symplectic growth. These symplectic growths transform into Ca-amphiboles with magnesium hornblende compositions

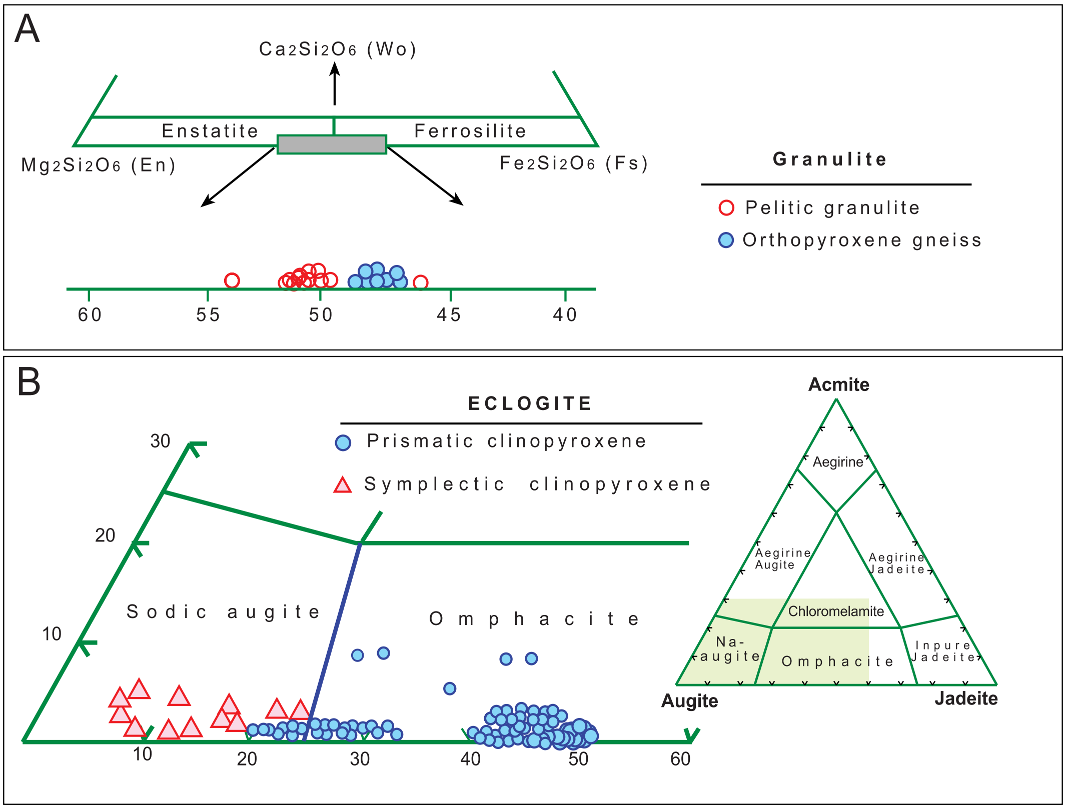

Figure 5- A: The composition of orthopyroxenes observed in granulites (Candan, 1995), B: The compositions of clinopyroxenes observed in eclogites in Essene and Fyfe (1967) diagram (Candan et al., 2001).

in further stages. In addition to these, garnets are surrounded and consumed by coronae made up of plagioclases which describe pressure decrease.

Pressure - temperature conditions

There are many calibrations related to geothermometer based on the Fe^{+2}-Mg change between garnet / clinopyroxene pair and these are widely used in eclogites. Estimations of Krogh (1988) and Ellis and Green (1979) were applied to eclogites in the Massif. Under the assumed 15 kbar pressure, the average temperature of the first eclogite was estimated as 596 °C and the second one was estimated as 644 °C. When the temperature interval of 100 mineral pairs were analyzed it was observed that the calibration of Ellis and Green (1979) gives compatible values clustering around 640-655 °C in temperature. This estimation however gives much dispersed values (±151 C) because of incomplete reactions, relic phases and distinct chemical zonings in eclogitic metagabbros. The estimation of Ellis and Green (1979) gave a temperature value of 633 °C under assumed 12 kbar in pressure.

Reaction of albite = jadeite + quartz which belongs to Holland (1980) is widely used in eclogites for the assumption of pressure. However the pressure values obtained from this calibration are accepted as minimum since albite is not in equilibrium in eclogites. Pressure of 15

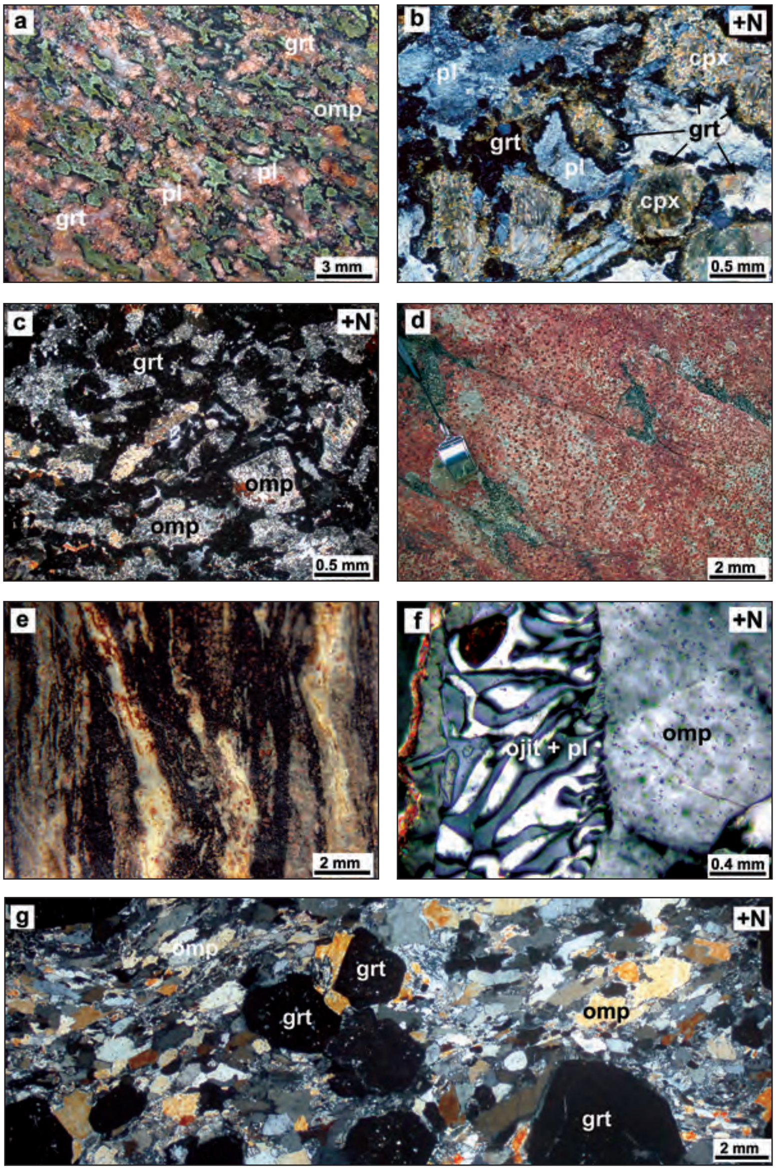

Figure 6- a: Relic texture belonging to primary gabbroic rock in eclogitic metagabbros which were statically recrystallized in low tensile regions. Plagioclases belonging to primary gabbro in rock can be observed, b: the development of garnet corona between primary clinopyroxene and plagioclase illustrating the beginning stage of gabbro - eclogitic metagabbro transformation, c: The relic primary ophitic texture in eclogitic metagabbros. Plagioclases and clinopyroxenes are pseudomorphically replaced by the assemblage of garnet and omphacite. Primary relic plagioclases among garnet ensembles can still be detected, d: Fresh eclogite made up of euhedral garnets within fine grained omphacitic groundmass. Textural / mineralogical evidences belonging to protoliths in fully recrystallized rocks were completely wiped away, e: Eclogite - garnet amphibolite transformations developing along shear zones that belong to retrogradation stage. Dark areas describe garnet amphibolite and light areas describe partly preserved eclogite layers, f: The consumption of omphacite by the symplectic growth augite - plagioclase ensemble, g: The microscopic views of fresh eclogites composed of prismatic omphacite and euhedral garnets (omp: omphacite, grt: garnet, pl: plagioclase, cpx: clinopyroxene, views from A to C belong to Tire / Küre region, from D to G belong to south of Kiraz).

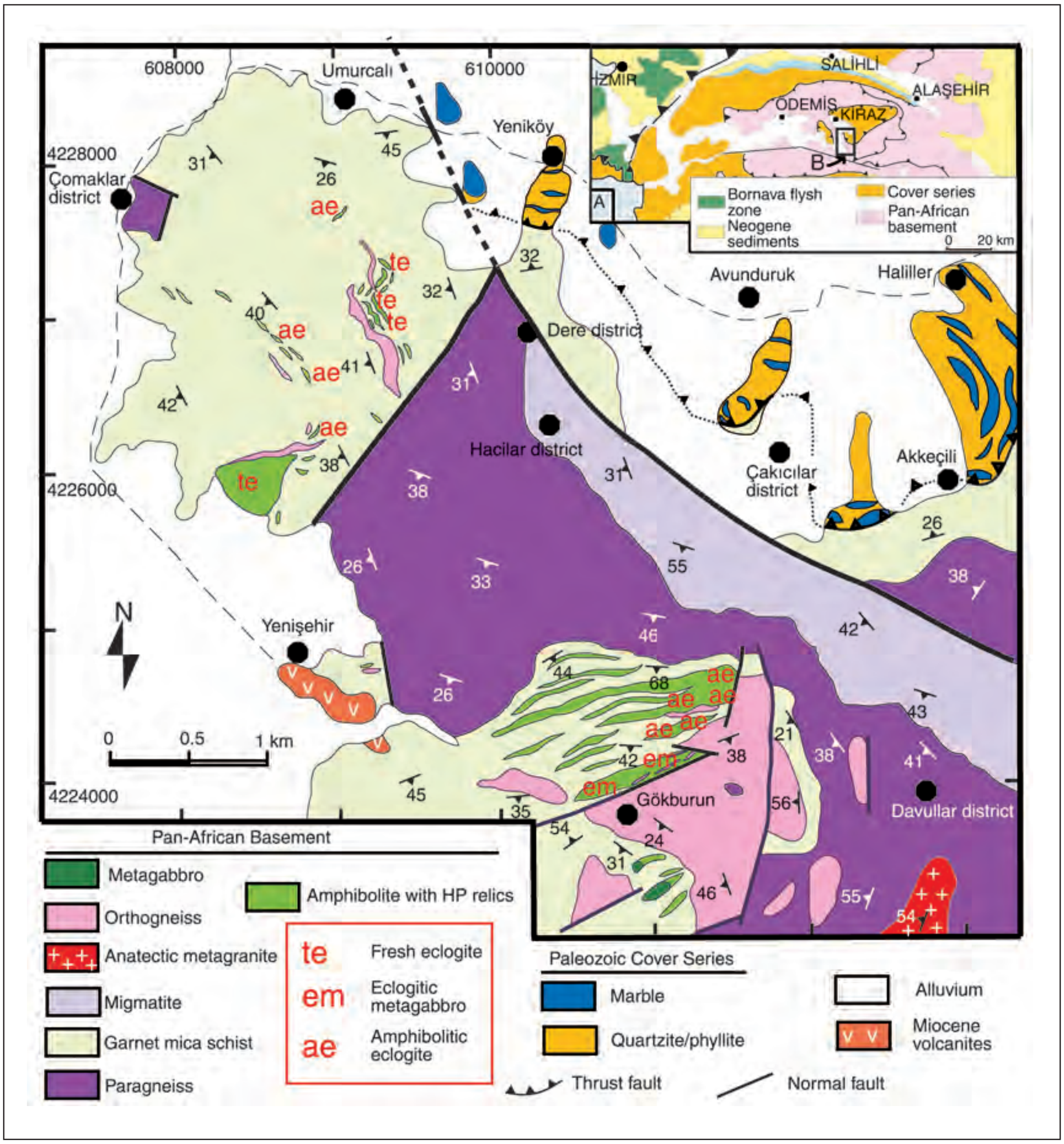

Figure 7- Geological map of Kiraz surround. It is composed of paragneisses that belong to the Pan-African basement, well preserved lens shaped eclogites and metaclastic serial vein made up of comformably overlying schist units (Candan et al., 2001).

kbar has been obtained for the temperature of 630 °C of omphacites in fresh eclogites. The same calibration gave a pressure value of 12 kbar in eclogitic metagabbros.

The age of metamorphism

After determining of the presence of eclogites in the Pan-African basement of the Menderes

Massif (Candan et al., 1994) the suggestions regarding the probable age of this high pressure metamorphism have generally been based on basic geological properties for many years (Oberhänsli et al., 1997; Candan and Dora, 1998; Candan et al., 2001). It was suggested that the high pressure metamorphism is Precambrian / Cambrian in age and can be associated with the Pan-African Orogeny (Oberhänsli et al., 1997; Candan et al., 2001), based on data that; 1- Eclogites are closely associated with gabbro stocks, 2- These gabbros are observed within the Pan African basement, 3- Eclogites are not present in cover series, and 4- Cover series are excluded from textural data although there are excess evidences related to pressure decrease in amphibolites within the basement.

However many attempts have been done for dating eclogites by classical zircon method, no age data can be obtained in them (Warkus, 2001). Oberhänsli et al. (2001), has applied TIMS method to zircons picked up from eclogites and eclogitic metagabbros. No result has been obtained since these are rather smaller than zircons in eclogites and have low uranium content. However, zircons in a sample of eclogitic metagabbro around Birgi region were dated as 529.9±22 Ma ^{206}Pb/ ^{238}U. This age is compatible with the basic geological properties of the Menderes Massif and has been interpreted as the age of high pressure metamorphism of the Massif by investigators.

AMPHIBOLITE FACIES METAMORPHISM

Petrography and mineralogy

To leave aside granulite and eclogites relics, the Pan-African basement of the Menderes Massif is described dominantly by the presence of the ensembles of Barrowian medium pressure metamorphism. A large range extending from the begining of greenschist facies to the development of migmatization - anatectic granite in metamorphic conditions is observed. Schists belonging to the basement widely crop out in Aydýn Mountains and Bozdaðlar sectors of Ödemiþ-Kiraz submassif. The metaclastics in Aydýn Mountains in which the lowest grade metamorphics of the Pan-African basement are observed, are composed of biotite schist and mica that is separated from each other by the appearing of isograd of garnet. Garnets are in the form of euhedral crystals not exceeding one or two millimeters. These rocks are composed of 'quartz + plagioclase + biotite + muscovite + garnet' and diffused syn tectonic growths are recognized in these garnets (Figure 8a). The isograd of staurolite and disthene can be observed in regional scale within a schist unit that reaches 6 km in thickness in Bozdaðlar (Ýzdar, 1971). The formation isograd of these minerals can also be detected in Demirci-Gördes Submassif, at south of Demirci (Candan, 1993). Staurolitic porphyroblasts within schists reach a dimension of 5-6 cm in Demirci region. Inclusional arrangements of staurolites indicate a syn tectonic growth in Bozdaðlar (Figure 8b). The general mineral composition of schists rich in biotite is 'quartz + albite + biotite + muscovite + garnet + staurolite + rutile'. Disthene is also added to this assemblage as a result of progressing metamorphism. The disthene crystals in disthene - staurolite schists around Demirci region reach a dimension of 7-8 cm (Figure 8c,d). Staurolite in these rocks gradually removes out and passes into disthene schist in 'quartz + albite + biotite + muscovite + garnet + staurolite + rutile' combination. In transition zones 'staurolite + disthene + sillimanite' assemblage can be detected. Sillimanite is diffused especially in paragneisses forming the lowermost part of metaclastic sequence and schists which intercalates with paragneisses. Cordierite and biotites belonging to granulite facies metamorphism in paragneisses are replaced by sillimanite bearing ensembles. This textural data indicate that the development of sillimanite is associated with the medium pressure metamorphism which occurred at the last stage of poly metamorphism.

Dispersive migmatization and anatectic granite development accompany to paragneisses rich in sillimanite (Figure 8e).The best migmatitic exposures in the Menderes Massif are recognized at south of Kula, around Selce village. This migmatitic focus is in dimension of 15 x 5 km and diagonally developed to primary stratigraphy. The transition into migmatite occurs by gradual increases of leucocratic sections within an approximate zone of 200 meters. Ptygmatic, schollen and schlieren structures are the most frequent migmatitic structures observed in the region. Migmatites in the region can also be recognized as inclusions reaching 4-5 km in dimension in orthogneiss masses. The formation of migmatization in Ödemiþ-Kiraz submassif can be detected within only one tectonical slice where eclogites and granulite relics are observed. Many migmatitic focal points that are broken apart from each other are present with a dimension not exceeding than 7 - 8 km in dimensions. Anyhow, all migmatitic paragneisses in Çine submassif are in the form of floating inclusions within gigantic orthogneisses. The dimensions of inclusions change from 10 cm to 6 km. At further stages of migmatization, anatectic granites are gradually passed through. Dimensions of granitic masses may reach up to 4 x 5 km. The anatectic granites in the Massif can be divided into two groups based on the places of crystallization. In situ crystallizing granites in these anatectic granites are dispersive in migmatites. In these granites which garnet porphyroblasts are widespreadly observed and made up of poly grained crystals with a dimension of 4-5 cm, have transitional contacts with migmatites which form the country rock (Figure 8f). These are fine to medium grained massive rocks and are composed of 'quartz + plagioclase + orthoclase + biotite + muscovite + garnet (±sillimanite)'. Granites which do not show migmatization and being crystallized as it rises upper layers form the second group. These granites present distinct intrusive contact relationships with surrounding rocks. The most typical exposures belonging to these granites are recognized in Ödemiþ - Kiraz submassif, the south of Kiraz. In this region, 6 independent granitic exposures with dimesions of 2 x 3 km have been observed within paragneisses. Anatectic granites are generally in massive character and were transformed into mylonites that show strong lineation and foliation along Alpine age ductile shear bands.

Pressure - temperature conditions

The Pan-African basement of the Menderes Massif is dominantly defined by ensembles of Barrowian type, medium pressure metamorphism. This basement has gained its main structure by Pan-African metamorphism and is overlain by the metamorphism developed during Alpine orogeny. Especially in homogenously composed, low grade schists, it is almost impossible to differentiate the effects of these two metamorphisms from each other. Classical index minerals show that conditions of Pan-African age medium pressure metamorphism has changed from greenschist facies to upper amphibolite facies which the partial melting had occurred. Okay (2001) made temperature estimations in garnet / biotite (Ferry and Spear, 1978) and garnet / hornblende (Graham and Powell, 1984) pairs from garnet schist and garnet amphibolites in Aydýn Mountains and pressure estimations from 'garnet + biotite + muscovite + plagioclase' ensemble (Ghent and Stout, 1984). In these classical estimations, the conditions of the lowest graded rocks belonging to the basement have been determined as 530±40 °C in temperature and 8 ± 2 kbar in pressure. According to Bucher and Frey (1994), the first occurrence of staurolite in pelitic rocks defines the beginning of amphibolite facies and expresses a temperature of more than 500 °C. However, Hoschek (1967) suggests a 540 °C temperatures at 4 kbar pressures and 565 °C temperatures for a pressure of 7 kbar for the occurrence of staurolite. In schists of the Massif that do not bear muscovite, the occurrence of sillimanite with the presence of quartz indicates that upper amphibolite conditions have been reached (600 - 650 °C). In order to estimate the conditions of the last stage, Dora et al. (2001), benefited from 'biotite + garnet + sillimanite' ensemble which replaces cordierites

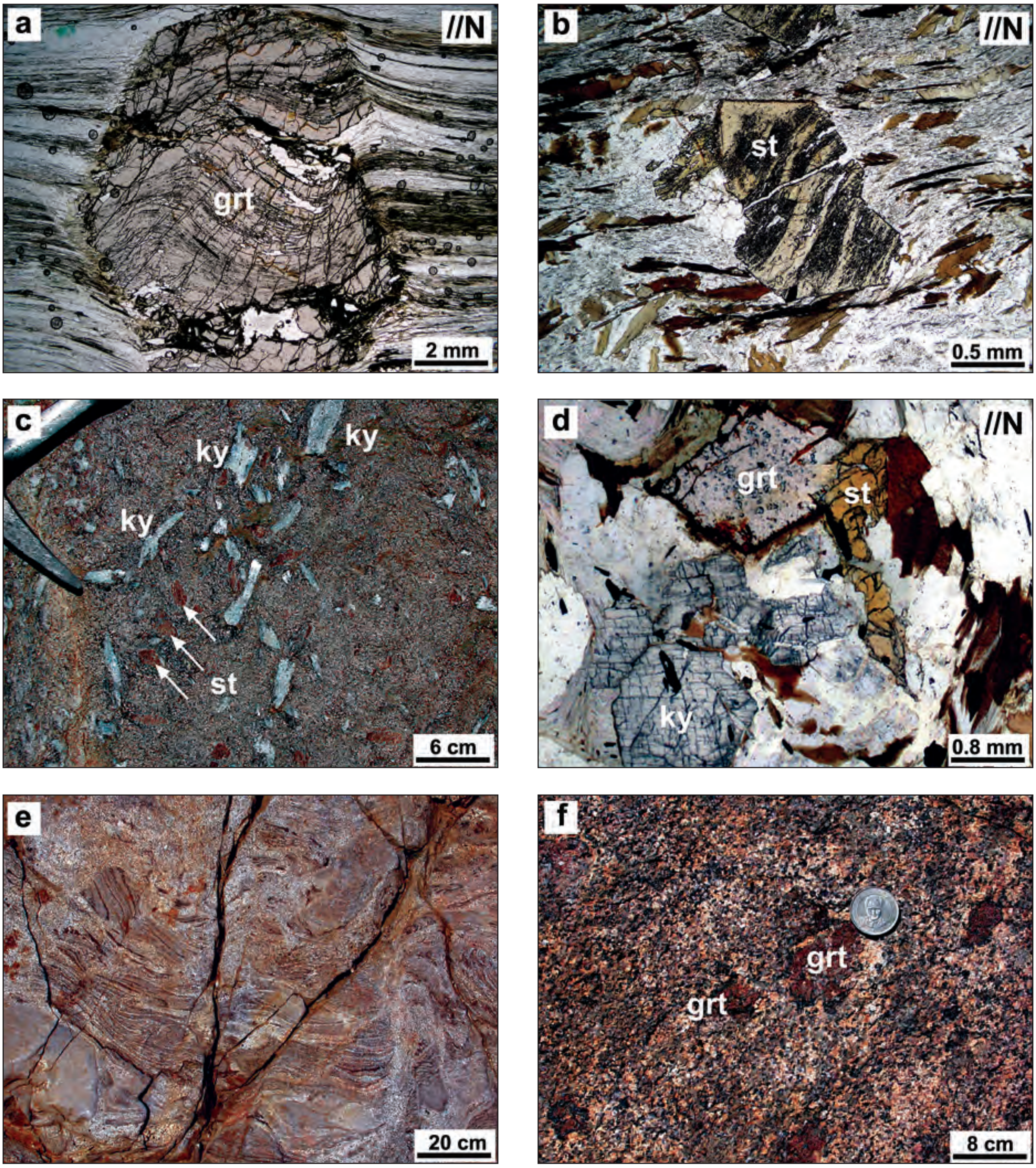

Figure 8- a-b: Syntectonic garnet and staurolite crystals defined by the patterns of quartz + opaque mineral inclusions, c: Staurolite and disthene crystals with dimension of 4- 5 cm which is observed in schists belonging to the Pan-African basement in Demirci region, d: The microscopic view of high graded schists made up of disthene + staurolite + garnet assemblages, e: Widespread migmatization observed in paragneisses cropping out in Selce Village / north of Alaþehir, f: Garnet crystals with dimensions reaching up 4 cm in anatectic granites around Solaklar village / Kiraz (grt: garnet, st: staurolite, ky: disthene).

belonging to granulite facies metamorphism. The temperature of garnet / biotite pairs were estimated at a lower value than assumed (588 °C) as they are fine grained and was affected from retrograde Fe-Mg equilibrium during cooling. It is known that the aforementioned last stage medium pressure metamorphism caused pervasive retrogradations in high pressure metamorphism ensembles at the basement (Candan et al., 2001). An approximate temperature of 623 °C were estimated for garnet / biotite pairs in garnet amphibolites around the wall zones of eclogites based on the calibrations made by Ferry and Spear (1978), Perchuk and Lavrent'eva (1983) and Battacharya et al. (1992). Garnet / hornblende pairs gave also similar temperature values based on Graham and Powel (1984) calibration (approximately 611 °C). 'Garnet + hornblende + plagioclase + quartz' ensemble suggested by Kohn and Spear (1989) were used for the assumption of pressure in garnet amphibolites. This geobarometer gave an approximate pressure of 7.3 kbar (Candan et al., 2001). In addition to these, as described above, pervasive melting in paragneisses at the basement and development of anatectic granite has occurred. This occurrence defines the maximum temperature conditions which the last metamorphic stage affecting the Pan-African basement has reached. In quartz and feldspar bearing rocks, the temperature values of partial melting in water pressure functions are accepted as 650 - 700 °C at water saturated zones (Bucher and Frey, 1994). When classical index minerals and geothermobarometric calculations given above in small amounts are assessed, the conditions of the last stage Barrowian medium pressure metamorphism can be given as 530 - 650 °C in temperature and 7 - 8 kbar in pressure. This last stage metamorphism has affected the Pan-African basement and caused diffuse retrogradations in ensembles belonging to former metamorphisms.

The age of metamorphism

Textural data show that ensembles of granulite and eclogites facies observed at the Pan-African Basement of the Menderes Massif were subjected to retrogradations under the conditions of upper amphibolite facies. Alpine age metamorphism made resetting the isotopic ratios of micas at the basement. Therefore, it seems impossible to designate this high temperature settlement related with the last stage of Pan-African Orogeny by ages of mica. When considered that the Alpine age metamorphism has affected the Pan-African basement as well, especially in schist series those of which have simple in composition, it is extremely difficult to distinguish the effects of the Pan-African and the overlying Alpine from each other. Therefore, to determine the age of the regarding metamorphism by indirect methods might give more realistic results. The most suitable method is to take into consider the migmatites and the associated anatectic granitic development at the basement. Field data shows that migmatization and granite development are restricted to the Pan-African basement. On the other hand, migmatites in question are observed in the form of floating inclusions within low graded and 550 Ma aged orthogneisses. This shows that, the migmatization has occurred before the intrusion of primary granites of orthogneisses and this high temperature can be associated with event causing retrogradation in granulite and eclogites. If this idea is relevant then the migmatization aging could play a critical role to determine the age of the last stage affecting the Pan-African basement.

Geological and petrographical data supports the presence of an original relation between migmatization and the development of anatectic granitic development in the Massif. Such granite observed in Ödemiþ-Kiraz submassif, east of Birgi was dated by Hetzel et al. (1998). Zircons which were selected from this granite and thought that crystallized since melting, have been dated as 551±1.4 Ma by classical U/Pb method. On the other hand, age determinations have been made by zircons picked up from neozones of migmatites observed in the middle and in the northern parts of the Menderes Massif

(Dannat and Resichmann, 1998). These zircons were dated in a range between 552 - 502 Ma, averaging at 540 Ma. Investigators have interpreted this age as the crystallizing age of granite associated with partial melting and migmatization in the Massif. When field, textural, petrographical and geochronological data are all assessed, it is understood that the Barrowian type, medium pressure metamorphism representing the last stage in polymetamorphical evolution of the Pan-African basement should be upper late Neoproterozoic (550 - 540 Ma) in age.

THE METAMORPHISM OF THE COVER SERIES

The cover series of the Massif are stratigraphically divided into two sub groups as; 1- Palaeozoic (Upper Devonian (?) - Permo Carboniferous) and 2- Mesozoic - Early Tertiary (Upper Triassic - Eocene) units. Only ensembles of Barrowian type, medium pressure metamorphism are observed in Palaeozoic sequence when metamorphisms of these units are studied. Whereas, although Mesozoic - Early Tertiary sequences are dominantly made up of ensembles of medium pressure metamorphisms, they also contain relic groups of the previous LT/HP metamorphism too.

HIGH PRESSURE METAMORPHISM

Petrography and mineralogy

Data of Alpine high pressure metamorphism (HP) in the Menderes Massif have only been determined in Mesozoic cover series (Rimméle et al., 2003). Mesozoic cover series observed at the southern part of Çine submassif begin with Upper Triassic schists containing quartz and metaconglomeratic layers at the bottom and this unit is transitionally overlain by an Late Triassic-Late Cretaceous aged platform type thick carbonates. Maastrichtian red pelagic marbles indicating the collapse of platform overlies these neritic carbonates and the deposit ends with Paleocene - Eocene aged metaolisthostrom (Özer et al., 2001).

Metaconglomerates emplaced in Upper Triassic clastics has a significant importance in terms of Alpine metamorphism (HP). These quartz metaconglomerates presenting a maximum thickness of 150 meters with a lateral continuity of 3 km, are observed in a few layers (Baþarýr, 1970). These metaconglomerates are in the character of channel filling which can be traced 150 km from Bafa Lake at the west to Karacasu at the east and made up of a lithology by quartz pebbles and coarse sand with dimension reaching up 6 cm (Figure 9a). Mica rich layers with a thickness of 0.2 - 0.6 m are frequently recognized within conglomerates. These layers have been derived from aluminum rich clays. These clays are rich in rosetta typed, needle like disthene (±chloritoid) crystals in 1 - 1.5 cm (Figure 9b). Quartz veins reaching 20 m in size are presented as contemporaneously with the metamorphism in quartz metaconglomerates. In three locations of the southern part of Çine submassif the presence of carpholite has been detected in these veins (Rimmelé et al., 2003) (Figure 10).

In Selimiye / Kurudere village (the best location), quartz veins are present with dimensions of 20 x 1.5 m within 200 x 100 m quartz metaconglomeratic layers. Green colored fibers with 70 cm in length are observed in these veins (Figure 9c). The fibers which are partly turned into chlorite and made up of carpholite also contain bluish disthene crystals (Figure 9d). Disthenes are pervasively transform into prophyllites in veins where 'Mg - carpholite + disthene + chlorite + quartz' ensemble is observed. Carpholites in the massif are rich in magnesium end member and X_{Mg} ratio varies in between 0.60 - 0.90. Carpholites in Kurudere region are approximately X_{Mg}= 0.68 in composition. However the XMn component is 0.03 approximately. Chlorites are rich in magnesium (X_{Mg}=0.8) and has a composition similar to clinochlor end member. In the second location, the southeast of Yataðan, Bahçeköyü 'Mg - carpholite + chloritoid + sudoite + quartz' assemblage was determined again in quartz veins of

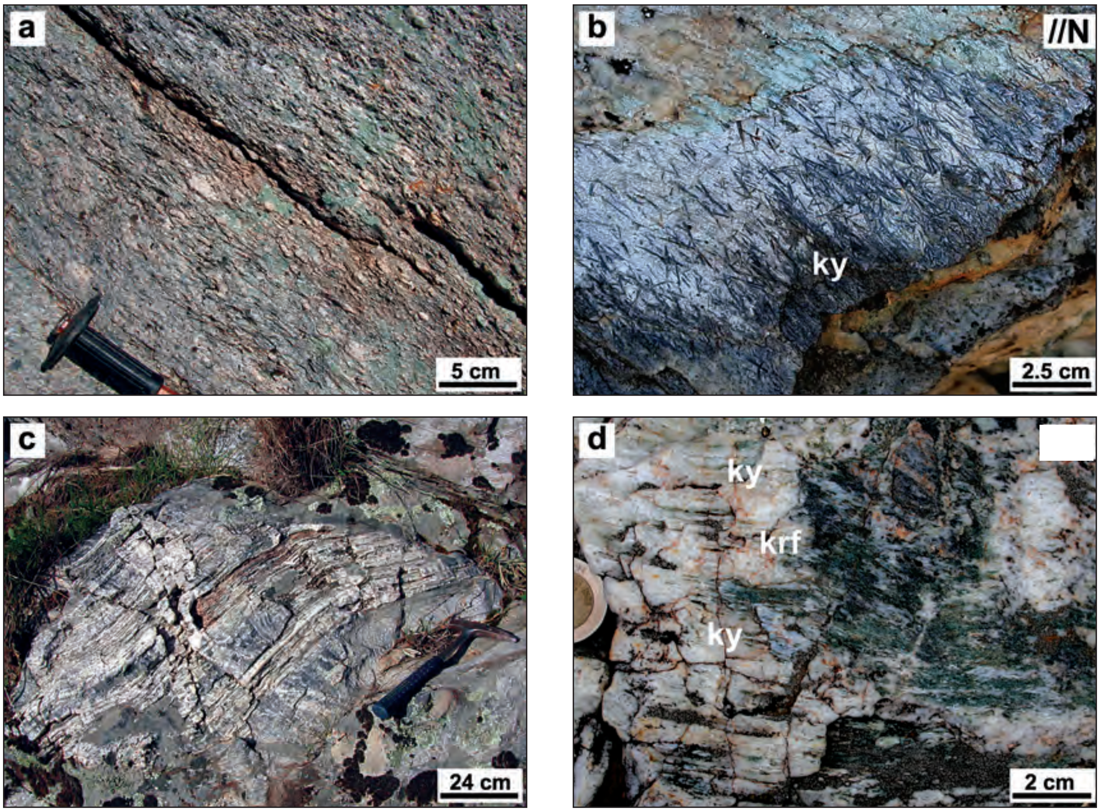

Figure 9- a: Upper Triassic metaconglomerate made up of quartz pebbles and cement of coarse sand, b: Needle like disthene crystals in phyllitic layers transformed from Al rich clays in metaconglomerates, c: Ligament like carpholite crystals reaching up 70 cm in length and observed in quartz veins within conglomerates, d: Green colored prismatic carpholite and blue colored disthene interval observed in quartz veins (krf: Carpholite, ky: disthene, images belong to north of Kurudere village).

quartz metaconglomerates. Carpholites are observed as well preserved in the form of thin fibers in quartz veins. Carpholites in X_{Mg}=0.7 composition are accompanied by phengite, sudoite, pyrophyllite and chloritoid and are rich in manganese (0.15 < X_{Mn} < 0.25) in the region. Textural data indicate that carpholites are the products of transformation of chlorites. Although there is not observed any dishtene in rocks textural evidences in veins show that pyrophyllites were derived from disthene. Chloritoids of X_{Mg}= 0.40 contain 0.40 X_{Mn} end member. In the last location exposed at east of Kavaklýdere, around Nebiler village, 'chloritoid + disthene + chlorite + quartz' and 'pseudocarpholite + disthene + chlorite + quartz' ensemble were detected. Textural data show that carpholites were consumed by the ensemble of 'disthene + chlorite'.

In addition to carpholite development in Triassic conglomerates, Na - amphibole formations have also been detected within Paleocene - Eocene aged metaolistostrome (Kazýklý Formation). Most extensive formation is recognized at west of Denizli / Çal, at Gözdek Hill. Very fine grained, blue colored layers are observed in thicknesses not exceeding 10 - 15 cm in matrix that belongs to olistostrome in this region. In

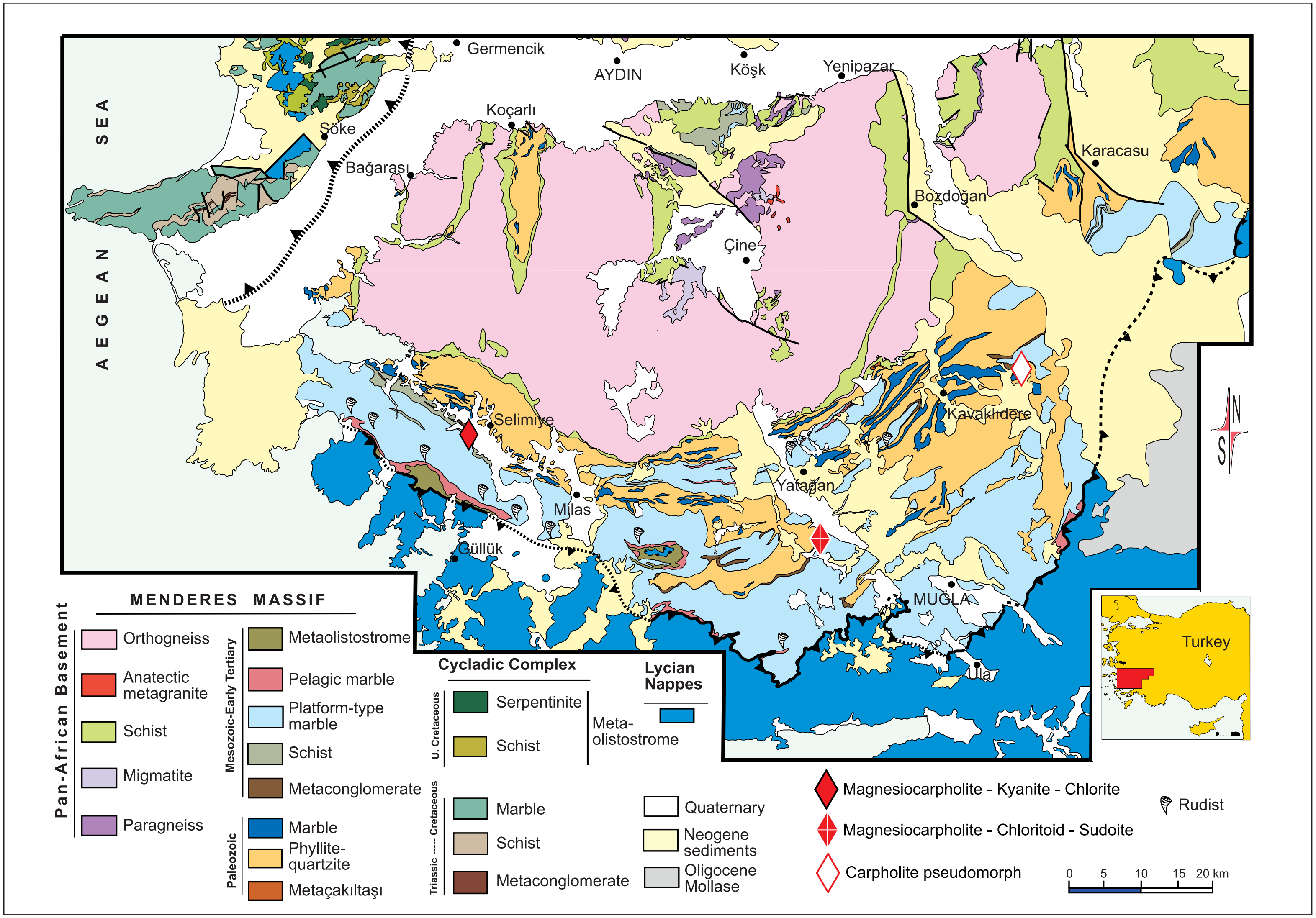

Figure 10- The generalized geological map of the southern part of Çine submassif and locations of Alpine high pressure / low temperature metamorphism ensembles (The map substantially was modified from Çaðlayan et al., 1980, locations of rudist belong to Dürr, 1975, Konak et al., 1978 and Özer et al., 2001 and the carpholite locations belong to Rimmelé et al., 2003).

addition to these, similar blue amphiboles were recognized in red pelagic marbles underlying the olistostrome. Al_2O_3 content is less than 2.57 % in these amphiboles and is in Mg - Riebeckite composition.

Pressure - temperature conditions

The P/T (pressure/temperature) conditions of Alpine high pressure metamorphism which affect the Mesozoic sequences of the Menderes Massif have been estimated by PTAX program using the compositions of phases in the region (Rimmelé et al., 2003). Different conditions have been established by the calculations made in three regions where carpholite is observed. The transformation of carpholite into disthene and chlorite and disthene into pyrophyllite define a retrogradation showing isothermal pressure decrease in Kurudere region. Minimum temperature of 430 °C and minimum pressure of 9 kbar were supposed for this retrogradation reaction of carpholite. In the second location (the Bahçekaya region) it was detected that carpholites were replaced by chloritoid. For this region, 440 °C temperature and 4.5 - 6.5 kbar pressure is estimated. The presence of only diaspores in metabauxites around the region shows that retrogradation has occurred in cooler conditions compared to Kurudere. For the 'chloritoid + disthene + chlorite' assemblage in Nebiler region, an average temperature of 450 °C and a pressure of 10-13 kbar is suggested. Consequently, Rimmelé et al. (2003) suggest the condition of this HP metamorphism affecting the Massif as 450 °C in temperature and 10 -12 kbar in pressure.

The age of metamorphism

HP metamorphism in the Massif has not yet radiometrically been dated. Stratigraphical and paleontological data show that the Mesozoic cover series in which HP metamorphism evidences are observed, begins with probable Upper Triassic metaconglomerates and ends with metaolistostrome which the age of it is under discussion (Kazýklý Formation). The Kazýklý Formation which is the uppermost unit of cover series of the Massif was aged as Middle Paleocene by Özer et al. (2001) and as Early Eocene by Konak et al. (1987) in Milas region. In Denizli /Çal region, Early - Middle Eocene age was dated in blocky matrix which overlies red pelagics and is considered most probably as the continuity of metaolistostrome in Milas (Özer et al., 2001). Early Miocene oil shales at the south of Alaþehir are the oldest units discordantly placed on the Menderes Massif. These data show that there has not been any deposition in the Menderes Massif between Eocene - Late Oligocene. Few isotopic age data regarding Alpine metamorphism show compatibility with related time interval. Rb/Sr analysis in white micas were dated in between 63 - 48 Ma with an average of 56 ±1 Ma (Late Paleocene). These ages are interpreted as the crystallization age of Alpine metamorphism. 37±1 Ma Rb/Sr values obtained from biotites are accepted as the cooling age (Satýr and Friedrichsen, 1986). New Rb/Sr and Ar/Ar ages obtained from micas show a correspondence with old data. It is considered that Rb/Sr mica ages varying between 62 - 43 Ma (Paleocene - Early Eocene) define the following stage of Barrowian type Alpine metamorphism and 36±2 Ma (Middle Eocene) Ar/Ar ages state the cooling stage (Bozkurt and Satýr, 2000). Similarly, Ar/Ar muscovite ages varying in between 43 - 37 Ma (Eocene: Hetzel and Reischmann, 1996) and 36±2 Ma (Middle Eocene: Lips et al., 2001) are interpreted as the cooling age by investigators. On the other hand, it is claimed that the main stage during the passage of Lycian nappes from north to south has occurred between Paleocene - Eocene time interval and this event has continued until Miocene (Collins and Robertson, 1999). It is accepted that the Massif has been buried under Lycian nappes during this tectonical period (Þengör et al., 1984; Dora et al., 1990, Bozkurt and Park, 1999; Rimmelé et al., 2003). When the age of youngest units in the Massif, the oldest sedimentary rocks which cover metamorphic rocks, the available geochronological data and time of passage of Lycian nappes to the south are all assessed together, it can be concluded that the Alpine aged high pressure metamorphism in the Massif and the overlap that caused retrogradation on these communities should occur in Eocene - Oligocene time interval.

MEDIUM PRESSURE METAMORPHISM

Petrography and mineralogy

As explained above, while Mesozoic - Early Tertiary sequences of the Menderes Massif have evidences regarding an overlap under conditions of Alpine aged high pressure and greenschist facies, the Palaeozoic cover series are only made up of assemblages of Barrowian type medium pressure metamorphism. The Palaeozoic cover series of the Menderes Massif has a stratigraphy of Upper Devonian quartzite and conformably and transitionally overlying Permo-Carboniferous phyllite - marble - quartzite intercalation. It has been known for many years that low graded marbles intercalated with phyllites are rich in fossils that have Permo-Carboniferous age in the Massif, in Göktepe region / the north of Muðla, Denizli / south of Babadað and in Aydýn Paþa Valley (Figure 11a) (Çaðlayan et al., 1980; Konak et al., 1987; Okay, 2001; Erdoðan and Güngör, 2004). Coral fossils in these marbles have been elongated and flattened as a result of ductile deformation (Figure 11 b). Quartzites derived from pure quartz arenite posses a simple mineral composition and made up of 'muscovite + quartz + apatite + zircon'. As for the black phyllites, that present a diffusion in very large areas in the Massif are rich in index minerals as total rock composites are suitable. At the southern part of the Çine submassif, mineral assemblages reflecting the conditions of greenschist facies are observed along a line extending from Bafa Lake to Denizli. Chloritic phyllites form the most dominant phyllitic type. These fine grained rocks are made up of 'muscovite + quartz + chlorite + opaque minerals'. Garnet phyllites and chloritoid - garnet phyllites are the other dominant phyllitic types. The lengths of the garnet porphyroblasts forming the 50 - 60 % of the rock may reach up to 1 cm. Euhedral garnets posses a composition rich in almandine. Chloritoids are in the form of dark green / black speckles and generally do not exceed 2 mm. The general mineral composition of these rocks is 'muscovite + chlorite + chloritoid + garnet + quartz'. Especially in phyllites around Karýncalý Mountain, the west of Karacasu, 'disthene + chloritoid' intercalation is pervasively observed (Figure 11c). Bow tie and rosetta shaped disthenes with 4 - 5 cm in length are completely in black as these are very rich in graphitic inclusions. Chloritoids could similarly form bow tie - bundle shaped crystals in 2 - 3 cm lengths. The general mineralogical compositions of these rocks are 'muscovite + chlorite + quartz + biotite + disthene + chloritoid + graphite'.

Palaeozoic age phyllites in Ödemiþ - Kiraz submassif are observed in two different tectonical slices. Phyllites in the lower tectonical slice make approximately 60 km lateral extension at the south of Aydýn Mountains and at the north of Bozdaðlar. The Palaeozoic deposit in Aydýn Mountains is dominantly made up of phyllites and in these, black marble bands and quartzite layers are observed that have several km's in lateral. Phyllites are composed of biotite phyllite and chlorite (±chloritoid). The Palaeozoic sequences in Bozdaðlar are dominant in phyllites and strongly show isoclinal foldings. This folding has a strike of N30°E and cause the repeating of the staurolitic isograd several times. Chloritoid phyllite and staurolite - garnet phyllites are observed in the region. Chloritoid phyllites are fine grained rocks and composed of 'muscovite + chlorite + biotite + chloritoid + quartz' (Figure 11d). With a transition zone where 'chloritoid + staurolite' assemblage (Figure 11 e) in the equilibrium is observed passes through staurolite - garnet phyllites. The staurolitic crystals in these rocks are needle like and 3 to 4 mm in length. Garnets form 2 to 3 mm in diameter and are euhedral crystals. The general mineralogical composition of these phyllites was defined as 'muscovite + quartz + biotite + garnet + staurolite'. The most typical characteristic of phyllites observed in the uppermost tectonical slice is the involvement of garnet porphyroblasts in dimensions of 1 -1.5 cm and 70%. Garnets are accompanied by staurolite and disthene crystals in dimensions of 3 - 4 mm (Figure 11f). In garnet porphyroblasts the textural zoning showing multi phase growth is observed that is made up of non inclusive wall zones and syntectonic core. The mineralogical composition of these phyllites is 'muscovite + quartz + garnet + staurolite (±chloritoid, ±disthene). Chloritoids are generally observed in the form of inclusions around cores of garnets. In Ödemiþ - Kiraz submassif, the units belonging to Palaeozoic sequences crop out between Kemalpaþa and Salihli. This region is made up of phyllites rich in graphites and bear black marble bands. Phyllites similar to ones in Ödemiþ contain 'chloritoid - staurolite - garnet' ensemble.

Pressure and temperature conditions

As also seen above, the presence of mineral assemblages associated with Barrowian type medium pressure metamorphism has been established in the Palaeozoic sequence of the Menderes Massif, so far. Chlorite, chloritoid, biotite and garnet minerals defining the conditions of progressing greenschist facies are pervasively observed in these rocks where exposed in various regions of the Massif. As for the phyllites in Bozdaðlar, the presence of 'staurolite - chloritoid' assemblage have been determined. This assemblage is equivalent of passage of greenschist - amphibolite facies which defines a narrow temperature interval (approximately 550 °C) in rocks possessing a special total chemical composition (Bucher and Frey, 1994). The companion of staurolitic disthene at south of Ödemiþ and Kula defines the progressing conditions of amphibolite facies. As a summary, the metamorphic conditions of the units petrographically equivalent to Palaeozoic sequences of the Massif present a change extending from lower greenschist facies to middle amphibolite facies.

The pressure/temperature calculations in ensembles of Palaeozoic sequences based on classical geothermobarometric methods have only been made in several locations. Ashword and Evirgen (1984) and Whitney and Bozkurt (2002) benefited from garnet/biotite pairs in Palaeozoic sequence that contains garnet and crops out at the southern part of the Menderes Massif. In these studies temperatures of 550 °C and 500 °C were obtained. In same studies, a pressure of less than 6 kbar was envisaged based on the assumption of the garnet + biotite + muscovite + plagioclase + quartz assemblage. For garnet - chloritoid phyllites in the same region Regnier et al. (2003) similarly suggests a pressure of 4 kbar and a temperature of 525 °C. The calculations in staurolite - garnet - disthene phyllites are a bit high and 8 - 11 kbar / 600 - 650 °C pressure and temperature values have been obtained, respectively. These phyllites are exposed at south of Koçarlý and have mistakenly been added to the Pan-African basement by investigators. Substantially, petrological / petrographical data are compatible to each other and these define an advancing medium pressure metamorphism and are changing from lower greenscihst to middle amphibolite facies.

The age of metamorphism

There is not any direct temporal / spatial evidence related to age of Palaeozoic sequence of the Menderes Massif and with Alpine HP/LT metamorphism observed in Mesozoic - Early Tertiary sequences. There are few cooling ages obtained from micas in units of the Pan-African basement in the Massif (Paleocene - Eocene). The primary contact relationship between Palaeozoic and Mesozoic - Early Tertiary cover series is stratigraphic (Konak et al., 1987). There is not any data associating with the effects of Variscan Orogeny in Anatolides. When all these evidences are evaluated together it is considered that the metamorphism of Palaeozoic sequences should be in Alpine age (Eocene-Oligocene) like Mesozoic - Early Tertiary sequences.

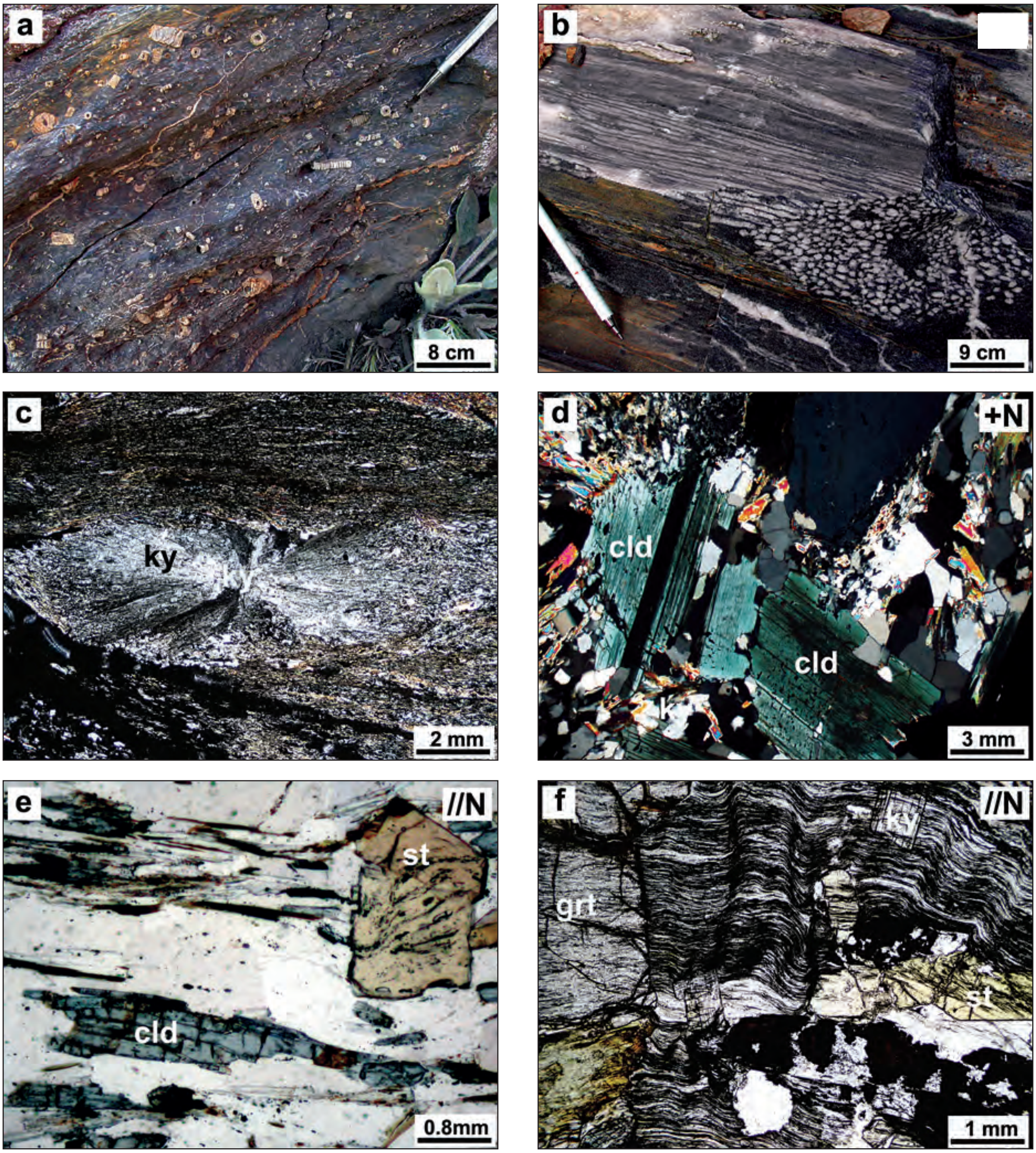

Figure 11- a: Crinoid fossils within black marbles belonging to Palaeozoic cover series in Göktepe / Muðla, b: coral fossils subjected to ductile deformation within black marbles in Karýncalýdað / Karacasu, c: Disthene crystals growing as bow tie observed in phyllites, Karýncalýdað / Karacasu. Disthenes are accompanied by small chloritoid crystals (within a mica rich groundmass) of in a way to describe greenschist facies to them, d: Chloritoid phyllites observed in Bozdaðlar, e: 'Chloritoid + staurolite' assemblage defining conditions of greenscihst - amphibolite facies transitions in phyllites, f: 'Garnet + staurolite + disthene' assemblage observed in phyllites at south of Kiraz (ky: disthene, cld: chloritoid, st: staurolite, grt: garnet).

DISCUSSION

Before discussing the metamorphic evolution of the Menderes Massif on the basis of core and cover series it will be useful to briefly deal with the opinions made on this subject. These opinions can be divided into two subgroups as i) the single phase Alpine metamorphism that yielded the present metamorphic character of the Massif and ii) the Pan-African and the Alpine age metamorphic evolutions of the base and cover series.

The majority of investigators defending the single phase model base their opinions on studies they have made on the southern part of Çine submassif. Ashword and Evirgen (1984); Erdoðan and Güngör (2004) and Bozkurt (1996); Whitney and Bozkurt (2002) are the chief investigators defending the single phase model in the Menderes Massif. These investigators substantially, accept orthogneisses belonging to basement as Tertiary granites and Pan-African metaclastics as units of cover series. As a result of this, it is defended that there is not an old basement in the Menderes Massif and thus the whole metamorphic evolution of the Massif can be explained only by the Alpine metamorphism.

Contrary to opinions stated above, there has been presented many evidences that the base and cover series of the Menderes Massif has a complex polymetamorphic evolution. Since there have been fewer evidences in radiometric and petrological data different opinions have been suggested about the metamorphic age and conditions of the Pan-African Basement in studies made before 1990. Schuiling (1962) suggests Pre-Hercynian age for the metamorphism of the basement while Brinkmann (1967) and Baþarýr (1970) accept this event had occurred in Precambrian. In 80's, based on the first age data obtained from gneisses which accumulates around 550 - 500 Ma this metamorphism was dated as Cambrian - Ordovician in age (Dora, 1975; Þengör et al., 1984; Satýr and Friedrichsen, 1986; Dora et al., 1990, 1992). Nowadays, it is heavily considered that the age of the poly phase metamorphism of the Pan-African basement is Precambrian and is associated with the Pan-African Orogeny (Þengör et al., 1984; Oberhansli et al., 1997; Candan and Dora, 1998; Candan et al., 2001).

By many investigators, it is accepted that the last metamorphism which affected the Pan-African core and Palaeozoic - Early Tertiary cover series and gave its recent structure is in Alpine age. Because of insufficient paleontological data in previous studies the age of this metamorphism was suggested as Devonian - Mesozoic (Schuiling, 1962); Jurassic (Baþarýr, 1975) and Liassic (Dora, 1975). Nowadays, it was determined that the depositional age of units of cover series shows continuity up to Eocene (Konak et al., 1987; Özer et al., 2001). Therefore, the age of Alpine metamorphism has shifted to Tertiary. This Alpine metamorphism was dated as Early Eocene - Oligocene (Þengör et al., 1984), Paleocene - Late Eocene (Dora et al., 1990, 1992), Late Cretaceous - Early Eocene (Erdoðan and Güngör, 2004), Eocene (Rimmelé et al., 2003), Early Eocene - Early Oligocene (Bozkurt et al., 1995).

THE METAMORPHISM OF THE PAN-AFRICAN BASEMENT

As briefly stated above, in many of the studies it has been claimed that the core series of the Massif should have been metamorphosed in Precambrian. Þengör et al. (1984), suggested that the metamorphic evolution of the basement could be associated with the Pan-African Orogeny without presenting any detailed data. The first petrological evidences regarding the complex metamorphic history of the Pan- African basement were defined by Candan (1995) and Candan et al. (1994). The high temperature metamorphism in granulite facies defined by the presence of orthopyroxene has rarely been preserved in units of the Pan-African Basement (Çetinkaplan, 1995; Candan, 1995; Candan and Dora, 1998). However, the presence of extremely well preserved eclogite and eclogitic metagabbro which defines the HP (high pressure) metamorphism conditions crop outs in nearly 100 locations throughout the Massif except the Çine submassif (Candan et al., 1994, 2001; Candan and Çetinkaplan, 2001; Candan and Dora, 1998; Oberhänsli et al., 1997, 1998; Çetinkaplan, 1995). These rocks are observed within the metasediments of high graded continental crust and represent an environment of crustal thickening by means of continental collision. The textural data show that HT/HP metamorphisms of the Massif restricted to the Pan-African basement has been overlain by Barrowian type medium pressure metamorphism which is defined by migmatization and anatectic granite development and were subjected to retrogradation. Migmatization and granitic intrusions are restricted to basement units. Thus, the related medium pressure metamorphism should also be evaluated within the metamorphic evolution of the basement.

As described above, the temporal evidences clearly reveal that the Pan - African basement were affected from the metamorphism that had occurred under conditions of granulite, eclogite and upper amphibolite facies. Geological, mineralogical and textural data show that the medium pressure metamorphism defines the last event within this polymetamorphic evolution (Candan and Dora, 1998). However, there has not been obtained any clear evidence regarding the relative ages of granulite and eclogite facies metamorphisms. Candan et al. (2001) established that high pressure ensembles were retrograded under conditions of isothermal pressure decrease. He has obtained these results by pressure / temperature studies made on eclogites and amphibolites of which are the medium pressure retrogradation products. There is not any textural / petrological data defining the temperature increase between the high pressure and the medium pressure metamorphism which causes retrogradation in high pressure. Therefore, investigators suggest that the relative order of metamorphism affecting the Pan- African basement is 'granulite - eclogites - amphibolite'.

Nowadays, it has been started to obtain some data about the absolute ages of related stages. The first evidences regarding the granulite facies have a broad error range (660+61/-63 Ma; Oelsner et al., 1997, Warkus et al., 1998) and is important in a sense that this high pressure event is Precambrian in age that is associated with the metamorphic stage of the basement. Despite that, analysis made on metamorphic zircons in granulites by U-Pb ion microprobe (SHRIMP II) method in recent years gave 583.0±5.7 Ma age cluster around Late Neoproterozoic age (Koralay et al., 2006). In order to date the high pressure metamorphism in the Massif TIMS method was applied to zircons in eclogite and eclogitic metagabbros. By means of zircons in a sample of eclogitic metagabbro, again 529.9±22 Ma $^{206}Pb/^{238}U$ age was obtained associated with the Pan-African event (Oberhänsli et al., 2010). These two age data are compatible with the suggested relative age between granulite - eclogite metamorphisms. As mentioned in previous parts, dating anatectic granites associated with the migmatization seems as the most reliable method which could be used in determining the age of medium pressure metamorphism defining the last stage of polymetamorphic evolution of the basement. So far, the age data has been obtained only from one of these granitic samples. By classical U/Pb method, Hetzel et al. (1998) estimated 551±1.4 Ma crystallization age for zircons in such a type of granite around Birgi region. On the other hand, zircon ages which define the partial melting period taken from Neozones of migmatites vary in between 502 - 552 Ma (average at 540 Ma) (Dannat and Reischmann, 1998). As seen, while the restricted age data obtained from eclogites and anatectic granite / migmatites clearly show the relation of these metamorphisms with the Pan-African Orogeny. It also presents an interfingering with each other within error ranges. Basic geological data show that the migmatization belongs to the last stage in the poly-phase metamorphism. However, it is clear that radiometric data should be increased in order to make the ages of these two events to be more sensitive.

Þengör et al. (1984) suggested that the evolutions of the Menderes Massif at the west and the old basements of the Bitlis Massif at the east could be associated with the Pan-African Orogeny. The Pan-African Orogeny describes poly-phased orogenic chain of events comprising the subduction, collision and suturing periods associated with the assemblage of the Gondwanaland between 950 - 450 Ma (Kröner, 1984). Actually this event is not only restricted to the African continent but also comprises the all phenomena occurred within the Gondwanaland. The vision of separation of East Gondwana (which is made up of Antarctica, Australia, Madagascar, Sri Lanka and India) from the West Gondwana (mainly made up of Africa and South America) by a big ocean called "the Mozambique Ocean" (Daiziel, 1991) in continental distribution in Neoproterozoic is generally accepted by many investigators (Stern, 1994; Wilson et al., 1997). This ocean is as big as the Pacific Ocean and is believed to have formed by the break up of Rodinia super continent 800 - 850 Ma years ago in Mezo Proterozoic time. The closure of this ocean and the collision of East - West Gondwana continents caused an orogenic belt development in NS trend which extends along the Eastern side of the African continent (Figure 12). This belt is named as the Mozambique belt or as the East African Orogeny (Stern, 1994).

There are many thoughts regarding that the final connection of Gondwana Super Continent was formed by the closure of large and small chains of oceans or by one big ocean (Unrug, 1996). On the other hand, the closure of the Mozambique Ocean and the final connection time of East and West Gondwana are in doubt. The Mozambique belt is substantially characterized by the formation of high graded metamorphism and by extensive granulite formation. In general terms, it is claimed that these granulitic formations are associated with the collision of East and West Gondwana Lands and the crustal thickening by the closure of Mozambique Ocean. It is also considered that the determination of granulite facies metamorphism could play a key role in solving the problem of time of assemblage (Stern, 1994). The age of granulite facies metamorphism in this belt is divided into two groups. Generally, granulite ages of the West Gondwana vary between 715-650 Ma. These granulites crop out at large areas in Kenya, Tanzania, Malawi, Sudan and in Mozambique. 710 Ma (Kröner et al., 1987) and 650-710 Ma (Maboko et al., 1989) ages were taken in Sudan and Tanzania, respectively. However, the granulites located on the same line especially in Madagascar, India, Sri Lanka and East Antarctica gives much younger ages varying in between 620-520 Ma (averaging at 550 Ma) (Madagascar: Paquette et al., 1994, 570-580 Ma; India: Collins et al., 2007, 513 Ma; Sri Lanka: Hölzl et al., 1994, 550-610 Ma; East Antarctica: Shiraishi et al., 1994, 550-520 Ma). In addition to these, similar ages of granulite facies metamorphism were also detected in southern Ethiopia as 545 Ma (Ayalew and Gichile, 1990) and 570 - 620 Ma (Key et al., 1989), in Erithrea as 593 Ma (Andersson et al., 2000). This two various age groups are interpreted in different ways. According to Stern (1994), the first group ages indicate the time of maximum thickness of the crust in continental collision period that is; the final assemblage age of the Gondwana. The investigator also interprets younger ages as a second young continent to continent collisional stage or the development of crustal thickening along a collision zone from west to east. However, Wilson et al. (1997) suggest that the granulites that belong to older group in the Mozambique belt might have developed either in deeper parts of arc regions at approaching plate margins or in collisional environments of small continents with island arcs, before the main collision stage. Wilson et al. (1997) claims that the final collision time between East and West Gondwana presents more compatibility with the ages of granulite facies metamorphism between 600 - 550 Ma determined in Madagascar, India, Sri Lanka and Antarctica as suggested by Kröner (1993) and Kröner et al. (1994) as well.

In recent years, the presence of eclogites has been detected describing the high pressure

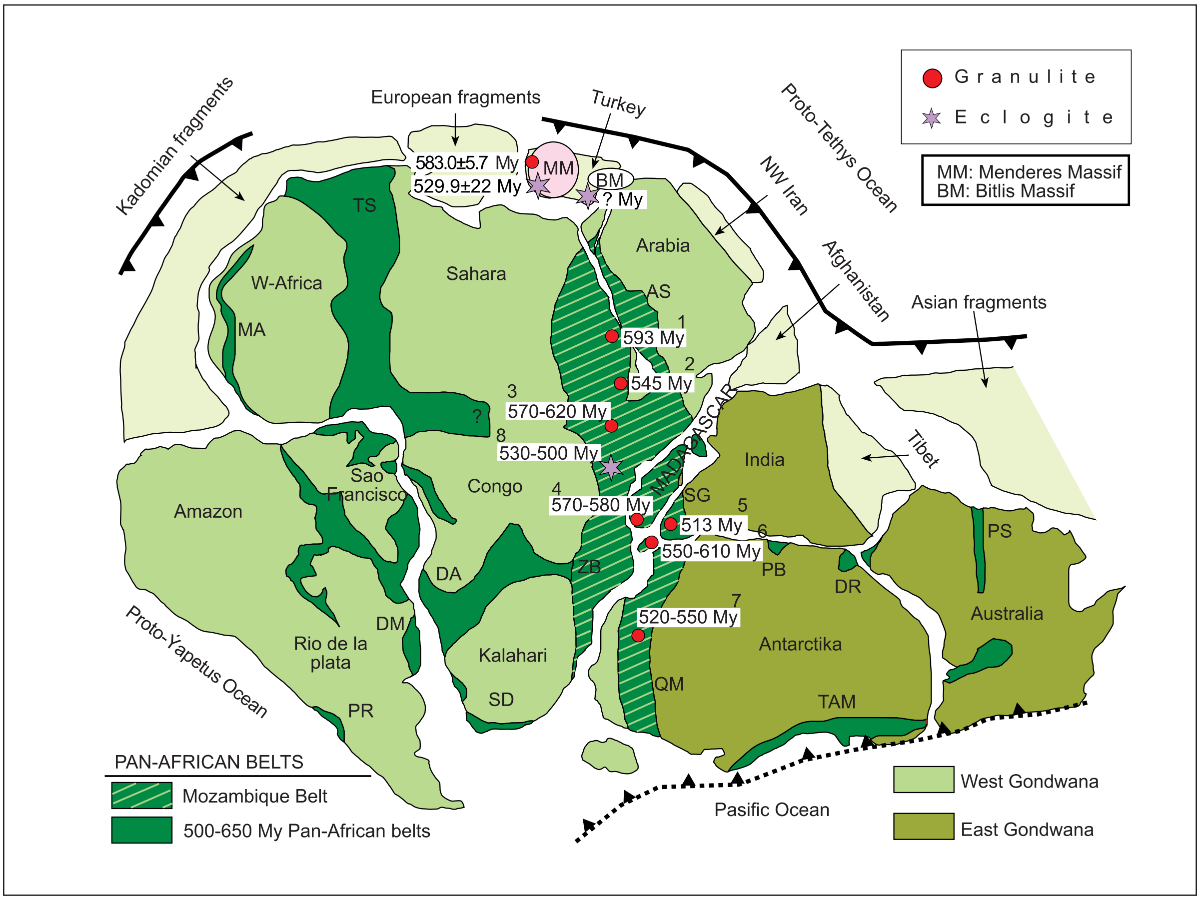

Figure 12- The paleogeographical map of the Gondwana Super Continent in Upper Late Neoproterozoic / Early Cambrian. The Mozambique belt and Late Neoproterozoic - Cambrian aged granulite and eclogite locations are shown on the map (AS: Arabian Shield, BM: Bitlis Massif, DA:Damara, DM:Dom Feliciano, DM:Denman Darling, MA:Mauretanides, MM: Menderes Massif, PB: Pryolz Bay, PP:Pampean Ranges, PS:Paterson, QM:Queen Maud Land, SD:Saldania, SG: South granulite terrane, TAM:Transantarctic Mountains, TS:Trans Sahara belt, ZB:Zambezi, Age determinations were taken from: 1-Andersson et al., 2000, 2-Ayalew and Gichile, 1990, 3-Key et al., 1989, 4-Paquette et al., 1994, 5- Collins et al., 2007, 6- Hölzl et al., 1994, 7- Shiraishi et al., 1994, 8- Ring et al., 2002).

metamorphism under medium temperature conditions in North Malawi located within the Mozambique belt (Ring et al., 2002). These rocks were transformed into high graded garnet amphibolites and are present in the form of lenses and tectonic slice. Zircons selected from these were dated as 530 - 500 Ma according to Pb / Pb method and were interpreted as the age of high pressure metamorphism. Ring et al. (2002) states that these eclogites indicate an environment of crustal thickening that have occurred at subduction and at the following continental collision period. Thus, these eclogites can be used in determining the closure time of Mozambique Ocean. So according to investigators, the collision of East - West Gondwana and the final connection times of the Mozambique Ocean and / or related other oceanic basins are in Cambrian age.

As seen in data given above, the ages of the granulite and eclogite facies metamorphisms within core series of the Menderes Massif show compatibility in temporal and spatial with the periods in the Mozambique belt. On the other hand, the paleogeographical position of Anatolide in Late Neoproterozoic/Early Cambrian is spatially compatible with this metamorphic history. Based on various geological evidences and correlations the western part of Arabian Peninsula is suggested for the position of Anatolia in this period (Þengör et al., 1984; Dora et al., 1995, 2002; Chen et al., 2002; Stampfli and Borel, 2002; Gessner et al., 2004; Gürsu et al., 2004; Monod et al., 2003; Neubauer, 2002; Koralay et al., 2005; Candan et al., 2001; Kröner and Stern, 2005; Oberhänsli et al., 2010). As it is known, Anatolia and the Menderes Massif took place in Alpine - Himalayan belt and were greatly reshaped by the Alpine age deformation and with the associated poly-phase metamorphism. At present, the direct relation between the Menderes Massif and the Mozambique belt has broken by the Alpine deformation. However, the granulite and eclogite ages obtained from the Massif make possible the correlation of the areas in the Pan-African basement like the Menderes and Bitlis massifs (Okay et al., 1985) in Turkey with the Mozambique Ocean. In addition to these, the same data also support that the closure of the Ocean and final collision of East - West Gondwanalands occurred at Precambrian-Cambrian boundary.

Based on the data given above, the tectonic and metamorphic evolution of the Pan-African basement in the Menderes Massif between 600 - 520 Ma time interval can be summarized in main topics as below;

1- Approximately 590 - 580 years ago, the protoliths of paragneiss and schists which are the oldest rocks of the Menderes Massif were deposited. Those are made up of litarenite, subarkose and mudstone and define the environment of passive continental margin (Dora et al., 2001; Koralay et al., 2005). This basin is fed from a provenance composed of crystalline rocks and probably is placed at a location closer to North of Gondwana in the Mozambique Ocean.

2- The closure of the related basin between 580 - 540 Ma time interval within the scope of Pan-African Orogeny caused the development of poly-phase Pan- African Metamorphism. At the first stage of this metamorphic period sediments belonging to core were buried to a depth of 20 km and were metamorphosed under conditions of granulite facies at 730 °C in temperature and 6 kbar in pressure. At this depth, an extra heat source is needed to cause the development of related high temperature metamorphism. It is considered that this heat source is provided from the basaltic magma underlying the crust. The intrusions of gabbro / norite in composition within the Pan-African basement support this idea.

3- In the latter stage of the closure of Mozambique Ocean, the collision of East and West Gondwana caused an excess crustal thicknening and the base rock and gabbros that intruded into were buried to a depth of 50 km. At this stage, the basement was metamorphosed under a temperature of 644 °C and minimum 15 kbar in pressure and gabbro - eclogite transformation occurred in some basic rocks.

4- At stage following the high pressure metamorphism the Pan-African basement got rid of the overlying mass without having a thermal release and was retrograded under Barrowian type medium pressure metamorphism (625 °C / 7 kbar) with a isothermal pressure decrease. This pressure decrease caused diffuse migmatization and development of anatectic granite in rocks.

5- At last stage of the Pan-African Orogeny, 550 Ma in age (Loos and Reischmann, 1999; 570-520 Ma), S type gigantic granite intrusions (protoliths of orthogneisses) occurred by the partial melting of lower crust.

THE METAMORPHISM OF PALAEOZOIC - EARLY TERTIARY COVER SERIES

It will be useful to summarize the most basic geological properties of cover series in the Menderes Massif before discussing the probable tectonic model in order to explain their metamorphisms. These are;

1- The high pressure metamorphism defined by the presence of eclogites within the Pan-African Basement is related with the Pan-African Orogeny and the Alpine metamorphism caused retrogradations only in conditions of lower amphibolite facies on this basement.

2- The Pan-African basement is unconformably overlain by Palaeozoic cover series.

3- Only medium pressure metamorphism related data were observed in Palaeozoic series, so far.

4- There is a stratigraphical primary contact relationship between Palaeozoic and Mesozoic - Early Tertiary sequence.

5- In Mesozoic-Early Tertiary units of cover series data of high pressure metamorphism and medium pressure metamorphism that caused retrogradation were detected.

6- The deposit in cover series continues until Eocene by probable Triassic discordance.

Considering these basic properties, it is needed to discuss the probable reasons of the difference of metamorphism which seem as contradictional between Palaeozoic and Mesozoic - Early Tertiary sequence. The first of these; i) although Palaeozoic sequences were subjected to high pressure metamorphism in Alpine, there is a small probability in the case of not having any related evidence yet. Apart from this, when the presence of disthene and staurolite are considered, it might be thought that Palaeozoic sequences were also affected from Alpine high pressure / low temperature (HP/LT) metamorphism and these assemblages were completely retrograded by the overlying medium pressure metamorphism reaching up lower amphibolite facies conditions. Palaeozoic sequences made up of extremely low graded rocks and bearing only chlorite in regions of the Massif such as Aydýn Mountains and Yataðan area do not contain any high pressure metamorphism evidence. Thus, the chance of the first probability abruptly decreases. Another probability is the effect of total rock chemistry on mineral occurrence. It is considered that total Al, K and Na contents of rocks are very effective on carpholite development (Rimmelé et al., 2003). Generally the carpholites are rich in Al and develop in prophyllite and disthene bearing sedimentary rocks and / or in syn metamorphic quartz veins within these sedimentary rocks. The state of being poor in Al or rich in Na-K for rocks obstructs the carpholite formation although there are suitable pressure/ temperature conditions (Rimmelé et al., 2003). Within this scope, it can be considered that phyllites in the Massif are very rich in albite and the carpholite formation in Palaeozoic sequence is obstructed since total rock chemistry is not suitable. Today, there is not any sufficient data to approve which probability is valid. However, there is also another data that could match with available data. In recent studies, the Massif was subjected to strong intra napping during Alpine metamorphism (Konak et al., 1994; Partzsch et al., 1998; Ring et al., 1999; Gökten et al., 2001). The core and cover series of the Menderes Massif were subjected to severe intra napping during collision. Various tectonic slices were buried at different depths and so were metamorphosed under different conditions. All these seem as realistic in probability. In tectonic slices where Palaeozoic sequences are observed, many different degree of metamorphisms are recognized ranging from lower greenschist facies to staurolite - disthene bearing amphibolite facies conditions. Cover series ranging from the Pan-African basement to Tertiary originally overlie each other by stratigraphical contacts. However, many tectonic slices belonging to basement are deprived of data of Mesozoic high pressure metamorphism or Palaeozoic medium pressure metamorphism that reaches the amphibolite facies. All these cases support the second probability.

Another subject that should be discussed is the relative ages of Alpine high pressure and medium pressure metamorphisms. A direct relation between the Alpine medium pressure metamorphism in Palaeozoic sequence and high pressure metamorphism in Mesozoic sequence has not been observed in any places. There are two possibilities in these metamorphisms of which should have periods of tectonical event following the other. If the high pressure metamorphism occurs first, then ensembles associated with this event should have been preserved at low graded layers of the overlying medium pressure metamorphism. If the medium pressure metamorphism occurred first, then it might be considered that the overlying high pressure metamorphism might be effective at low graded layers of the first metamorphism. However, it is not possible to assume that the core and cover series within thicknesses of thousands of meters in the Massif have been buried at the same depth as in one piece during the closure stage of Neothethys Ocean. In case of severe intranapping within the tectonic Alpine evolution of the Massif, it can be considered that each slice has an original metamorphic evolution associated with its burial depth and these have no obligation in resembling each other in character. In conclusion, in order to reach a definite result there is not any sufficient evidence. However, when available geological data are evaluated it is inferred that core and cover series during Alpine tectonometamorphic stage of the Massif were subjected to intranapping. Tectonic slices being formed were buried at various depths and experienced their own metamorphic evolutions and merged together during overlapping.

Within the framework of general tectonical structure of Turkey (Ketin, 1966; Okay et al., 1996), the southern part of the Ýzmir - Ankara - Erzincan suture zone can be divided into two tectonical units as Anatolides and Taurides. The temporal geological properties of these units which derived from Anatolide - Tauride platform are directly related with the Albian - Eocene stage of the northern branch of Neothethys Ocean (Þengör and Yýlmaz, 1981; Okay et al., 2001; Rimmelé et al., 2003; Candan et al., 2005). Anatolides which are the metamorphic equivalents of Taurides are divided into tectonic zones mainly as; Tavþanlý Zone, Afyon Zone, the Menderes Massif and the Lycian nappes from north to south. Under this consideration, the Alpine tectonometamorphic evolution of the Menderes Massif should be deliberated with the evolutions of other zones forming the Anatolides. Two basic properties are observed at common stages of these zones associated with the closure of the Neotethys Ocean. There is a systematic rejuvenation from north to south during the collapse of platform in tectonic zones and the development of pelagic environments and in the high pressure metamorphisms of these zones (Okay et al., 2001; Candan et al., 2005).

Þengör and Yýlmaz (1981); Okay et al. (1998) suggest that closure of the northern branch of Neothethys Ocean began in Albian along the intra oceanic subduction zone and the majority of oceanic lithosphere were subducted between Albian - Turonian time intervals (Figure 13 A). The beginning of pelagic carbonate deposition at Tavþanlý zone in Cenomanian shows that the platform began to subside under the load of overlapping oceanic lithosphere progressing southward. At the stage following the subduction of lithospheric mantle, deposits of the passive continental margin were buried under the load made up of oceanic mantle wedge and accretional prism (Figure 13 B). These deposits will later turn into metamorphites of Tavþanlý zone and forms the northern most part of the Anatolide - Tauride platform. The pressure and temperature conditions obtained from Tavþanlý zone (24±3 kbar / 430±30°C; Okay, 2002) and Ar-Ar phengite ages (80±0.5 Ma; Sherlock et al., 1999) indicate that the Tavþanlý zone buried at a depth of minimum 60 km in Campanian and subjected to high pressure metamorphism (Okay et al., 1998, Çetinkaplan et al., 2008). While the neritic carbonate deposition occurred in Turonian - Early Santonian in Afyon zone and the Menderes Massif, the Karaböðürtlen flysch began to deposit in Lycian nappes located at northern parts at that time. The petrological data defined by the presence of carpholite indicate that Lycian nappes were subjected to high pressure / low temperature metamorphism under 8 kbar in pressure and 400 °C in temperature which is equal to a burial depth of 30 km (Oberhänsli et al., 2001; Rimmelé et al., 2005). It is considered that the Lycian nappes were metamorphosed as a result of intranapping of the platform in early Maastrichtian-Early Tertiary (? Early Paleocene) and its burial under the slice of oceanic lithosphere passing southward. The pelagic carbonate deposition in Afyon zone which was derived from the southern part of Anatolide -Tauride platform begins in early Maastrichtian and is overlain by Early Paleocene flysch. The stratigraphical data indicate that intranapped Afyon zone was buried under the Lycian nappes and peridotite slice which passes southward in Middle Paleocene and were subjected to high pressure metamorphism. Metamorphic conditions were determined as 6-9 kbar in pressure and 350 °C in temperature, which equal to a burial depth of 35 km (Candan et al., 2005). The pelagic carbonate depositon in the Menderes Massif forming the southernmost tectonic zone of Anatolides began in late Campanian - late Maastrichtian and this deposition continues with the deposition of Paleocene-Eocene age flysch (Özer et al., 2001). This data shows that ophiolites and Lycian nappes reached this part of the platform. There is a temporal and spatial relation on southward passage between the Alpine metamorphism of the Pan-African basement and cover series of Menderes Massif and Lycian nappes (Þengör et al., 1984; Dora et al., 1990, Bozkurt and Park, 1994, 1999; Rimmelé et al., 2003). The stratigraphical / paleontological evidences and mica ages show that the Barrowian type medium pressure and synchronous high pressure metamorphism are Eocen - Oligocene in age. Palaeozoic and Mesozoic cover series show different metamorphic conditions since Lycian nappes and the platform buried under the overlying peridotite slice were severely intranapped and forming tectonic slices were buried at different depths. For example, pressure of 10 -12 kbar and a temperature of 440 °C values determined for carpholite - disthene ensemble in Triassic metaconglomerates show that some slices were buried at a depth of 35 km. However, slices which were buried at shallower depths were metamorphosed in Barrowian type medium pressure conditions. There has not been any sedimentation on the Massif between Eocene - Late Oligocene time interval. This indicates that the massif was overburden by Lycian nappes in this time interval. Lower Miocene shales are the oldest sedimentary cover on the Massif and indicate that the Massif got rid off the load on it and cropped out at this time (Figure 13 C).

CONCLUSION

The results based on data related to polymetamorphic stages of the Pan- African basement and Palaeozoic - Early Tertiary cover series are given below:

1- The Pan-African basement is defined by the presence of poly metamorphism under high temperature granulite, high pressure eclogites and under the medium pressure upper amphibolite facies conditions.

2- Geochronological data show that these metamorphisms occurred in time intervals of 580 - 540 Ma (Late Neoproterozoic) (granulite facies: 583.0±5.7 Ma; eclogites facies: 529.9±22 Ma; Upper amphibolite facies: 551 - 540 Ma).

3- This poly-metamorphism is associated with the closure of Mozambique Ocean and the collision of East - West Gondwana lands forming the very last stage of assemblage of Gondwana super continent. In this stage the basement was

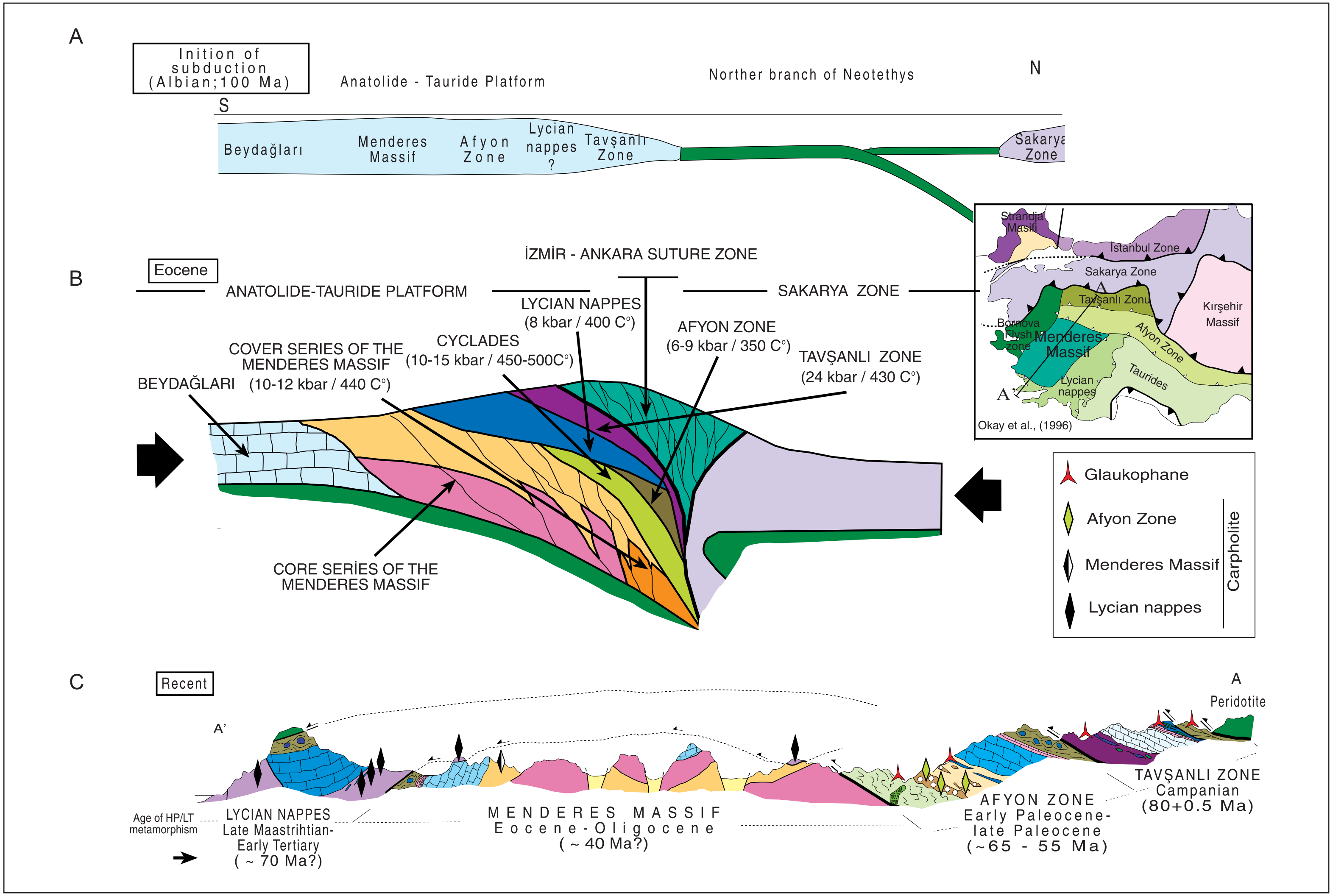

Figure 13- A: the location of Anatolide - Tauride platform and the northern branch of Neothethys Ocean, B: Accretional complex developed on the northern branch of Neothethys Ocean in Eocene. In schematical section, the relative structural positions of HP tectonic zones belonging to Anatolide are shown, C: The geological cross section showing relations of tectonic zones of Anatolides at present (Figures were modified from Rimmelé et al., 2003 and Candan et al., 2005).

subjected to a deep burial and metamorphisms of high temperature and high pressure during overlapping were effectively retrograded under medium pressure conditions.

4- The Alpine metamorphisms of Palaeozoic - Early Tertiary cover series of the Massif are Eocene - Oligocene in age.

5- This metamorphism is associated with the closure of northern branch of Neothehtys Ocean, the continental collision, intranapping of Anatolide - Tauride platform, the burial stage of the part of Menderes massif of the platform below Afyon Zone at the north and the ophiolitic slices of the Lycian nappes passing southward.

6- Some Mesozoic slices were buried at an average depth of 35 km which might cause change in high pressure / low temperature conditions.

7- The effects of Alpine metamorphism in Palaeozoic sequence and the Pan- African Basement dominantly define greenschist facies conditions of Barrowian type medium pressure metamorphism and rarely reach lower amphibolite facies condition.

8- There is not any sufficient data showing if the Pan-African basement and its Palaeozoic sequences were effected from Alpine high pressure metamorphism.

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REFERENCES

Andersson, U.B., Ghebreab, W., Teklay, M. and Whitehouse, M.J., 2000. Ages and character of crust-forming and metamorphic events in east-central Eritrea:new convensional and SIMS U-Pb and mineral P-T data. Journal African Earth Science, 18th Colloquium of African Geology, vol. 30. Special Abstract Issue, 7p.

Ashworth, J. R. and Evirgen, M., 1984. Mineral chemistry of regional chloritoid assemblages in the Chlorite Zone, Lycian Nappes, south-west Turkey, Mineralogical Magazine, 48, pp.159-165.

Ayalew, T. and Gichile, S., 1990. Preliminary U-Pb ages from southern Ethiopia. In: G.Rocci, M Deschamps (eds). In recent data in African Earth Sciences, CIFEG Occ. Public 22, pp.127-130.

Baþarýr, E., 1970. Bafa gölünün doðusunda kalan Menderes Masifi güney kanadýnýn jeolojisi ve petrografisi. Ege Üniversitesi Fen Fakültesi Ýlmi Raporlar Serisi, 102, pp.1-44.

_____, 1975. Çine güneyindeki metamorfiklerin petrografisi ve bireysel indeks minerallerin doku içerisindeki geliþimleri. (Doçentlik tezi), Ege Üniversitesi YBÝ, Ýzmir 87p, (unpublished).

Bhattacharya, A., Mohanty, L., Maji, A., Sen, S.K. and Raith, M., 1992. Non-ideal mixing in the phlogopite-allinite binary: contrasts from experimental data on Mg-Fe partitioning and a formulation of the biyotite - garnet geothermometer. Contributions Mineralogy Petrology, 111, 87-93.

Bozkurt, E., 1996. Metamorphic conditions in the Palaeozoic schists of the southern Menderes Massif: field, petrographic and textural evidence. Turkish Journal of Earth Science, 5, 105-121.

_____, 2004. Granitoid rocks of the southern Menderes Massif: field evidence for Tertiary magmatism in an extensional shear zone. International Journal Earth Science, 93, 52-71.

_____ and Park, R. G., 1994. Southern Menderes Massif: an incipient metamorphic core complex in Western Anatolia, Turkey. Journal Geological Society London, 151, 213-216.

Bozkurt, E., Winchester, J.A. and Park, R.G., 1995. Geochemistry and tectonic significance of augen gneisses from the southern Menderes Massif (West Turkey). Geological Magazine, 132, 287-301.

_____ and Park, G., 1999. The structure of the Palaeozoic schists in the Southern Menderes Massif, western Turkey: a new approach to the origin of the main Menderes Metamorphism and its relation to the Lycian Nappes. Geodinamica Acta, 12, 25-42.

_____ and Satýr, M., 2000. The southern Menderes Massif (western Turkey): geochronology and exhumation history. Geological Journal, 35, 285-296.

Brinkmann, R., 1967. Die Südflanke des Menderes-Massivs bei Milas, Bodrum und Ören, Scient. Representative Faculty Science, Ege University, Izmir, Turkey, 43p.

Bucher, K. and Frey, M., 1994. Petrogenesis of metamorphic rocks. Springer-Verlag, 318p.

Candan, O., 1993. Menderes Masifi'nin kuzeyinde Demirci - Borlu arasýnda kalan bölgenin petrografisi, petrolojisi ve metamorfizmasý. Doða - Türk Yerbilimleri Dergisi, 2, 69-87.

_____, 1994. Alaþehir kuzeyinde (Menderes Masifi, Demirci -Gördes Asmasifi) gözlenen metagabrolarýn petrografisi ve metamorfizmasý. Türkiye Jeoloji Bülteni, 37, 29-40.

_____, 1995. Menderes Masifi'ndeki kalýntý granulit fasiyesi metamorfizmasý. Turkish Journal of Earth Science, 4, 35-55.

_____, 1996*a*. Aydýn - Çine Asmasifi'ndeki (Menderes Masifi) gabrolarýn metamorfizmasý ve diðer asmasiflerle karþýlaþtýrýlmasý. Turkish Journal of Earth Science, 5, 123-139.

_____, 1996*b*. Kiraz - Birgi çevresindeki (Menderes Masifi / Ödemiþ-Kiraz Asmasifi) metagabrolarýn petrografisi ve metamorfizmasý. Yerbilimleri, 18, 1-25.

Candan, O. and Dora, O.Ö., 1998. Menderes Masifi'nde granulit, eklojit ve mavi þist kalýntýlarý: Pan-Afrikan ve Tersiyer metamorfik evrimine bir yaklaþým. Türkiye Jeoloji Bülteni, 41/1, 1-35.

_____, _____, Dürr, St. and Oberhänsli, R., 1994. Erster Nachweis von Granulit und Eklogit - Relikten im Menderes - Massif / Türkei. Göttingen Abr. Geol. Paläont. Sb. 1 5. Symposium TSK, 217-220.

_____, _____, Oberhänsli, R. and Dürr, St., 1995. Relicts of a high - pressure metamorphism in the Menderes Massif: Eclogites. International Earth Sciences Colloquim on the Aegean Regions, Güllük, 8-9.

_____, _____, _____, Çetinkaplan, M., Partzsch, J. H. and Dürr, S., 1998. Pan-African high-pressure metamorphism in the Precambrian basement of the Menderes Massif, Western Anatolia-Turkey. Third International Turkish Geology Symposium., Middle East Technichal University, Ankara, 275p.

_____ and Çetinkaplan, M., 2001. Menderes masifi'ndeki eklojit / epidot-mavi þist fasiyesi metamorfizmasý ve kikladik kompleksle karþýlaþtýrmasý. YDABÇAG-495 nolu Türkiye Bilimsel ve Teknolojik Araþtýrma Kurumu projesi. 182p. (unpublished).

_____, Dora, O.Ö., Oberhänsli, R., Çetinkaplan, M., Partzsch, J.H., Warkus, F. and Dürr, S., 2001. Pan-African high-pressure metamorphism in the Precambrian basement of the Menderes Massif, Western Anatolia, Turkey. International Journal of Earth Science (Geologische Rundschau), 89, 4, 793-811.

_____, Çetinkaplan M., Oberhansli R, Rimmelé, G. and Akal C., 2005. Alpine high-pressure / Low temperature metamorphism of Afyon Zone and implication for metamorphic evolution of western Anatolia, Turkey. Lithos, 84, 102-124.

_____, Koralay, E., Dora, O., Chen, F., Oberhänsli, R., Akal, C., Satýr, M. and Kaya, O., 2006. Menderes Masifi'nde Pan-Afrikan Sonrasý Uyumsuzluk: Jeolojik ve Jeokronolojik Bir Yaklaþým. 59. Türkiye Jeoloji Kurultayý Bildiri özleri, Ankara, 25-27.

Candan, O., Koralay, E., Dora, Ö., Chen, F., Oberhansli, R., Çetinkaplan, M., Akal, C., Satýr, M. and Kaya, O., 2007. Menderes Masifi'nin Pan-Afrikan temelin stratigrafisi ve örtü - çekirdek serilerinin ilksel dokanak iliþkisi. Menderes Masifi Kolokyumu, Ýzmir, 8-13.

Chen, F., Siebel, W., Satýr, M. and Terzioðlu, M.N., 2002. Geochronology of the Karadere basement (NW Turkey) and implications for the geological evolution of the Ýstanbul Zone. International Journal of Earth Science, 91, 469-481.

Collins, A. S. and Robertson, A. H. F., 1999. Evolution of the Lycian Allochthon, western Turkey, as a north-facing Late Palaeozoic to Mesozoic rift and passive continental margin. Geological Journal, 34, 107-138.

______, Santoshb, M., Braunc, I. and Clarka, C., 2007. Age and sedimentary provenance of the Southern Granulites, South India: U-Th-Pb SHRIMP secondary ion mass spectrometry. Precambrian Research, 155, 1-2, 125-138.

Çaðlayan, A.M., Öztürk, E.M., Öztürk, Z., Sav, H. and Akat, U., 1980. Structural observations on the southern Menderes Massif. Publication Chamber Geology Engineering. Turkey (in Turkish with English summary), 10, 9-17.

Çetinkaplan, M., 1995. Geochemical, mineralogical and petrographical investigation of the eclogites in southern part of Tire area, Ödemiþ-Kiraz submassif of the Menderes Massif. Yüksek Lisans tezi, Dokuz Eylül University, Ýzmir, 92p (unpublished).

______, Candan, O., Oberhansli, R. and Bousquet, R., 2008. Pressure-temperature evolution of lawsonite eclogitein sivrihisar, Tavþanlý zone, Turkey. Lithos, 104, 12-32.

Daiziel, I.W.D., 1991. Pasific margins of Laurentia and east Antartica - Australia as a conjugate rift pair: Evidence and implications for an Eocambrian supercontinent. Geology, 19, 598-601.

Dannat, C. and Reischmann, T., 1998. Single zircon ages of migmatites from the Menderes Massif, SW Turkey. Program des Workshops 'Das Menderes Massif (Türkei) und seine nachbargebiete'. Mainz, Germany, 12p.

Dora, O.Ö., 1975. Menderes Masifi'nde alkali feldspatlarýn yapýsal durumlarý ve bunlarýn perojenetik yorumlarda kullanýlmasý. Türkiye Jeoloji Bülteni, 24, 91-94.

______, Kun, N. and Candan, O., 1990. Metamorphic history and geotectonic evolution of the Menderes Massif. Proc. of International Earth Sciences Congress on Aegean Regions, Ýzmir/ Turkey, 2, 102-115.

______, ______ and ______, 1992. Menderes Masifi'nin metamorfik tarihçesi ve jeotektonik konumu. Türkiye Jeoloji Bülteni, 35/1, 1-14.

______, Candan, O., Dürr, S. and Oberhänsli, R., 1995. New evidence on the geotectonic evolution of the Menderes Massif. International Earth Sciences Colloquium on the Eagean Region, Izmir-Turkey, 1, 53-72.

______, _____, Kaya, O., Koralay, E. and Dürr, S., 2001. Revision of the so-called "leptite-gneisses" in the Menderes Massif: A supracrustal metasedimentary origin. International Journal of Earth Science (Geologische Rundschau), 89/4, 836-851.

______, ______,______ and ______, 2002. Menderes Masifi'ndeki Leptit - Gnayslarýn Kökenlerinin Yeniden Yorumlanmasý, Metamorfizmalarý ve Jeotektonik Ortamlarý .YDABÇAG - 554 nolu Türkiye Bilimsel ve Teknolojik Araþtýrma Kurumu projesi, 165 (unpublished).

______, _____, Kaya, O., Koralay, E. and Akal, C., 2005. Menderes Masifi Çine Asmasifi'ndeki Koçarlý - Bafa - Yataðan - Karacasu arasýnda uzanan gnays / þist dokanaðýnýn niteliði: Jeolojik, tektonik, petrografik ve jeokronolojik bir yaklaþým. YDABÇAG - 101 Y 132 nolu Türkiye Bilimsel ve Teknolojik Araþtýrma Kurumu projesi, 197p. (unpublished).

Dürr, St., 1975. Über Alter und geotektonische Stellung des Menderes Kristallins / SW -Anatolien und seine Äquivalente in der Mittleren Aegean. Habilitation thesis University of Marburg, 107p (unpublished).

Ellis, D.J. and Green, D.H., 1979. An experimental study of the effects of Ca upon garnet - clinopyroxene Fe - Mg exchange equilibria. Contributions Mineralogy Petrology, 71, 13 - 22.

Erdoðan, B. and Güngör, T., 2004. The problem of the core - cover boundary of the Menderes Masif and an emplacement mechanism for regionally extensive gneissic granite, Western Anatolia Turkey. Turkish Journal of Earth Science, 13, 15-36.

Essene, E.J., and W.S. Fyfe., 1967. Omphacite in Californian metamorphic rocks. Contributions Mineralogy Petrology. 15 1-23.

Ferry, J.M. and Spear, F.S., 1978. Experimental calibration of the partitioning of Fe and Mg between biotite and garnet. Contributions Mineralogy Petrology, 66, 113-117

Gessner, K., Collins, A., Ring, U. and Güngör, T., 2004. Structural and thermal history of poly-orogenic basement: U-Pb geochronology of granitoid rocks in the southern Menderes Massif, Western Turkey. Journal Geological Society London, 161, 93-101.

Ghent, E.D. and Stout, M.Z., 1984. Geobarometry and geothermometry of plagioclase-biyotite-garnet-muscovite assemblages. Contributions Mineralogy Petrology, 76, 92-97.

Gökten, E, Havzaoðlu, T, and Þan, Ö., 2001. Tertiary evolution of the central Menderes Massif based on structural evolution of metamorphics and sedimentary rocks between Salihli and Kiraz (western Turkey). International Journal of Earth Sciences, 89, 745-756.

Graham, C.M. and Powell, R., 1984. A garnet - hornblende geothermometer: calibration, testing and application to the Pelona schist, Southern California. Journal Metamorphic Geology, 2, 13-31.

Gürsu, S., Göncüoðlu, C. and Bayhan, H., 2004. Geology and geochemistry of Pre-early Cambrian rocks in the Sandýklý area: Implication fort he Pan-African evolution of NW Gondwana. Gondwana Research, 7, 923-935.

Hetzel, R. and Reischmann, T., 1996. Intrusion age of Pan-African augen gneisses in the southern Menderes Massif and the age of cooling after Alpine ductile extensional deformation. Geological Magazine, 133, 5, 565 - 572.

_____, Romer, R. Candan, O. and Passchier, C.W., 1998. Geology of the Bozdað area, central Menderes Massif, SW Turkey: Pan - African basement and Alpine deformation. Geologische Rundschau, 87, 394-406.

Holland, T.J.B., 1980. The reaction albite = jadeite + quartz determined experimentally in the range 600 - 1200 °C. American Mineralogist, 65, 129-134.

Hoschek, G., 1967. Untersuchungen zum Stabilitatsbereich von chlorioid und staurolith. Contrubutions Mineralogy Petrology, 14, 123-162.

Hölzl, S., Hofmann, A.W., Todt, W. and Köhler, H., 1994. U-Pb geochronology of the Sri Lankan basement. Precambrian Research, 66, 123-149.

Ýzdar, E., 1971. Introduction to geology and metamorphism of the Menderes Massif of western Turkey. Petroleum Exploration Society of Libya, 495-500.

Ketin, Ý., 1966. Tectonic units of Anatolia (Asia Minor). (In Turkish), Maden Tetkik Arama Enstitüsü Dergisi, 66, 20-34.

Key, R.M., Charsley, T.J., Hackman, B.D., Wilkinson, A.F. and Rundle, C.C., 1989. Superimposed Upper Proterozoic collision-controlled orogenies in the Mozambique orogenic belt of Kenya. Precambrian Research 44, 197-225.

Kohn, M.J. and Spear, F.S., 1989. Emprical calibration of geobarometers for the assemblage garnet+ hornblende+plagioclase+quartz. American Mineralogist, 74, 77-84.

Konak, N., Akdeniz, N. and Öztürk, E.M., 1987. Geology of the south of Menderes Massif, I.G.C.P. project no:5, Correlation of Variscan and pre-Variscan events of the Alpine Mediterranean mountain belt, field meeting, Mineral Research and Exploration. Institute, Turkey, 42-53.

Konak, N., Çakmakoðlu, A., Elibol, E., Havzoðlu, T., Hepþen, N., Karamanderesi, I.H., Keskin , H., Sarýkaya, H., Sav, H. and Yusufoðlu, H., 1994. Development of thrusting in the median part of the Menderes Massif. Abstracts 47th Geological Congress. Turkey-Ankara, 34p.

Koralay, O.E., Satýr, M. and Dora, O.Ö., 2001. Geochemical and geochronological evidence for Early Triassic calc-alkaline magmatism in the Menderes Massif, western Turkey. International Journal of Earth Science, 89, 822-835.

_____, Chen, F., Candan, O., Dora, O.Ö., Satýr, M. and Oberhänsli, R., 2005. Pb-Pb geochronology of detrital zircons from Neoproterozoic paragneisses in the Menderes Massif, Turkey. Int. Earth Scie. Coll. on the Aegean Regions, Ýzmir-Turkey, 69p.

_____, _____, Oberhänsli, R., Wan, Y. and Candan, O., 2006. Age of Granulite Facies Metamorphism in the Menderes Massif, Western Anatolia / Turkey: SHRIMP U-Pb Zircon Dating. 59. Türkiye Jeoloji Kurultayý Bildiri özleri, Ankara, 28-29.

Krogh, E.J., 1988. The garnet - clinopyroxene Fe-Mg geothermometer- a reinterpretation of existing experimental data. Contributions Mineralogy Petrology, 99, 44-48.

Kröner, A., 1984. Late Precambrian plate tectonics and orogeny: a need to redefine the term Pan-African. In: J Klerkx, J Michot (ed). African Geology, Tervuren: Musccc. R. I'Afrique Centrale, 23-28.

_____, 1993. The Pan-African Belt of northeastern and eastern Africa, madagascar, southern India, Sri Lanka and Eastern Antarctica: terrane amalgamation during formation of the Gondwana supercontinent. In: U. Thorweihe and H. Schandelmeier (eds). Geoscientific Research in Northeast Africa. Rotterdam, Balkema, 3-9.

_____, Greiling, R.O., Reischmann, T., Hussein, I.M., Stern, R.J., Durr, J. St., Kruger and Zimmer, M., 1987. Pan African crustal evolution in northeast Africa. In: Kroner, A. (ed.). Proterozoic Lithospheric Evolution. Am. Geophysical Union, Washington, DC, 69-94.

Kröner, A. and Stern, R.J., 2005. Pan-African Orogeny. Encyclopedia Geology, Volume-I, A to E, 1-12, Elsevier.

_____, Jaeckel, P. and Williams, I.S., 1994. Pb loss patterns in zircons from a high-grade metamorphic terrain as revealed by different dating methods: U-Pbi Pb-Pb ages for igneous and metamorphic zircons from northern Sri Lanka. Precambrian Research, 66, 151-181.

Lal, R.K., 1993. Internally consistent recalibrations of mineral equilibria for geothermometry involving garnet-orthopyroxene-plagioclase-quartz assemblages and their application to the South Indian granulites. Journal Metamorphic Geology, 11, 855-866.

Lips, A.L.W., Cassard, D., Sözbilir, H., Yýlmaz, H. and Wijbrans, J.R., 2001. Multistage exhumation of the Menderes massif, western Anatolia (Turkey). International Journal of Earth Science, 89, 781-792.

Loos, S. and Reischmann, T., 1999. The evolution of the southern Menderes massif in SW Turkey as revealed by zircon dating. Journal Geological Society London. 156, 1021-1030.

Maboko, M.A.H., McDougall, I. and Zeitler. P.K., 1989. Dating Late Pan-African cooling in the Uluguru granulite complex of eastern Tanzania using +0Ar-39Ar technique. Journal African Earth Science, 9, 159-167.

Monod, O., Kozlu, H., Ghienne, W., Dean, W.T., Günay. Y., Herisse, A. and Paris. F., 2003. Late Ordovician glajiation in southern Turkey. Terra Nova, 15, 4, 249-257.

Neubauer, F., 2002. Evolution of Late Neoproterozoic to Early Palaeozoic tectonic elements in central and southern European Alpine mountain belt: review and syntesis. Tectonophysics, 352, 87-103.

Newton, R.C. and Perkins, D., 1982. Thermodynamic calibration of geobarometers based on the assemblages garnet-plagioclase-orthopyroxene (clinopyroxene)-quartz. American Mineralogist, 67, 203-222.

Oberhänslý, R., Candan,O., Mezger, K., Dora,Ö. and Dürr, St., 1995*a*. Eclogites and granulites in the Menderes Massif, Western Turkey. Strasburg. EUG 8, Terra abstracts, 8p.

_____, ______,______,______ and______, 1995*b*. High pressure relics in the Menderes Massif, Turkey. Bochumer Geol. und Geotech. Abr. 44, 132-133.

_____, ____, Dora, O.Ö. and Dürr, St., 1997. Eclogites within the Menderes Crystalline Complex / western Turkey / Anatolia. Lithos, 41, 135-150.

_____, Partzsch, J. H., Çetinkaplan, M. and Candan, O., 1998. HP record in the Lycian nappes (Western Turkey). Third International Turkish Geology Symposium, Middle East Technical University, Ankara, 274p.

_____, Candan, O. and Wilke, W., 2010. Geochronologic Evidence of Pan-African Eclogites from the Menderes Massif, Turkey. Turkish Journal Earth Science, (in printhouse).

Oelsner, F., Candan, O. and Oberhänsli, R., 1997. New evidence for the time of the high-grade metamorphism in the Menderes Massif, SW-Turkey. Terra Nostra, 87. Jahrestagung der Geologischen Vereinigung e.v., Fundamental geologic processes, 15p.

Okay, A.I., 2001. Stratigraphic and metamorphic inversions in the central Menderes Masif: a new structural modal. International Journal of Earth Science, 89, 709-727.

_____, 2002. Jadeite-Chloritoid-glaucophane-lawsonite blueschists in North-west Turkey: unusually high P/T ratios in continental crust. Journal Metamorphic Geology, 20, 757-768.

_____, Arman, M.B. and Göncüoðlu, M.C., 1985. Petrology and phase relations of the kyanite-eclogites from eastern Turkey. Contributions Mineralogy Petrology, 91, 196-204.

Okay, A., Satýr, M., Maluski, H., Siyako, M., Monie, P., Metzger, R. and Akyüz, S., 1996. Palaea- and Neo-Tethyan events in northwestern Turkey: Geologic and geochronologic constraints. In: Yýn, A and Harrison M (eds). Tectonic of Asia. Cambridge University Press, 420-441.

Okay, A., Harris, N. and Kelley, S., 1998. Exhumation of blueschists along a Tethyan suture in northwest Turkey. Tectonophysics, 285, 275-299.

_____, Tansel, I. and Tüysüz, O., 2001. Obduction, subduction and collusion as reflected in the Upper Cretaceous - Lover Eocene sedimentary record of western Turkey. Geological Magazine, 138, 2, 117-142.

Özer, S., Sözbilir, H., Özkar, I., Toker, V. and Sarý, B., 2001. Stratigraphy of Upper Cretaceous-Palaeogene sequences in the southern and eastern Menderes Massif (Western Turkey). International Journal Earth Science, 89, 4, 852-866.

Paquette, J.L., Nedelec, A., Monie, B. and Rakotondrazafy, M., 1994. U-Pb single zircon Pb-evaporation and Sm-Nd isotopic of granulitic domain in SE Madagascar. Journal Geology, 102, 523-538.

Perchuk, L.L. and Lavrent'eva, I.V., 1983. Experimental investigation of exchange equilibria in the system cordierite - garnet - biyotite. In: Saxena SK (ed). Kinetics and equilibrium in mineral reactions. Advances in physical geochemistry 3. Springer, 199-239.

Regnier, J.L., Ring, U., Paschier, C.W., Gessner, K. and Güngör, T., 2003. Contrasting metamorphic evolution of metasedimentary rosks from Çine and Selimiye nappes in the Anatolide belt, western Turkey. Journal Metamorphic Geology, 21, 699-721.

Rimmelé, G., Oberhansli, R., Goffe, B., Jolivet L., Candan, O. and Çetinkaplan, M., 2002. High-pressure rocks from the Lycian nappe complex and the southern Menderes Massif: Implication for tectonic evolution of Southwest Turkey. I[st] int. Symp of Istanbul Technical University, 102p.

_____, _____,_____,_____,_____ and _____, 2003. First evidence of high-pressure metamorphism in the "cover series" of the southern Menderes

Massif: Tectonic and metamorphic implications for the evolution of SW Turkey. Lithos, 71, 19-46.

Rimmelé, G., Para, T., Goffe, B., Oberhansli, R., Jolivet, L. and Candan, O., 2005. Exhumation paths of high-pressure - low- temperature metamorphic rocks from the Lycian Nappes and the Menderes Massif (SW Turkey): a multi-equilibrium approach. Journal Petrology, 46, 3, 641-669.

Ring, U., Gessner, K., Güngör, T. and Passcchier. C., 1999. The Menderes Massif of western Turkey and the Cycladic Massif in the Aegean-do they really correlate? Journal Geological Society London, 155, 3-6.

_____, Kröner, A., Buchwald, R., Toulkeridis, T. and Later, P., 2002. Shear zone patters and eclogite-facies metamorphism in the Mozambique belt of northern Malawi, east-central Africa: implication for assembly of Gondwana. Precambrian Research, 116, 19-56.

Satýr, M. and Friedrichsen, H., 1986. The origin and evolution of the Menderes Massif, W-Turkey: A rubidium/ strontium and oxygen isotope study. Geologische Rundschau, 75/3, 703-714.

Schuiling, R.D., 1962. On petrology, age and structure of the Menderes migmatite complex (SW - Turkey). Mineral Research and Exploration Instute, 58, 71-84.

Sherlock, S., Kelley, S., Inger, S., Harris, N. and Okay, A., 1999. Ar-Ar and Rb-Sr geochronology of high-pressure metamorphism and exhumation history of the Tavþanlý Zone, NW Turkey. Contributions Mineralogy Petrology, 137, 46-58.

Shiraishi, E.D.J., Hiroi, Y., Fanning, C.M., Motoyoshi, Y. and Nakai, Y., 1994. Cambrian orogenic belt in east Antartica and Sri Lanka: Implications for Gondwana assembly. Journal Geology, 102, 47-65.

Stamfli, G.M. and Borel, G.D., 2002. A plate tectonic model for the Palaeozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrones. Earth Planetary Science Lett, 196, 17-33.

Stern, R.J., 1994. Arc assembly and continental collusion in the Proterozoic east African orogen: Implications for the consolidation of Gondwanaland. Annual Review Earth Planet Science, 22, 319-351.

Þengör, A.M.C. and Yilmaz, Y., 1981. Tethyan evolution of Turkey: a plate tectonic approach. Tectonophysics, 75, 181-241.

_____, Satýr, M. and Akkök, R., 1984. Timing of tectonic events in the Menderes Massif, western Turkey. Implications for tectonic evolution and evidence for Pan-African basement in Turkey. Tectonics, 3, 7, 693-707.

Partzsch, J. H., Oberhänsli, R., Candan, O. and Warkus, F., 1998. The Menderes Massif, W-Turkey: A complex nappe pile recording 1.0 Ga of geological history. Third International Turkish Geology Symposium, Middle East Technical University, Ankara, 281p.

Unrug, R., 1996. The assembly of Gondwanaland. Episodes, 19, 1-2, pp.11-20.

Warkus, F., 2001. Untersuchungen an hochdruckrelikten im zentral Menderes Massiv, W-Turkey. PhD thesis, Institut für Geowissenschaften, Universitat Potsdam, 80p, (unpublished).

_____, Partzsch, J. H., Candan, O. and Oberhänsli, R., 1998. The tectono-metamorphic evolution of the Birgi - Tire nappe in the Menderes Massif, SW-Turkey. 7. Syposium tektonik-sutrukturgeologie - kristallingeologie, Freiberger Forschungsheft, C-471, 237-238.

Whitney, D.L. and Bozkurt, E., 2002. Metamorphic history of the southern Menderes massif, western Turkey. Geological Society America Bulletin, 114, 7, 829-838.

Wilson, T.J., Grunow, A.M. and Hanson, R.E., 1997. Gondwana assembly: The view from southern Africa and east Gondwana. Journal Geodynamics. 23, (3/4), 263-286.

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EMPLACEMENT CHARACTERISTICS OF THE GNEISSIC GRANITES IN THE MENDERES MASSIF (WESTERN ANATOLIA) AND THEIR IMPLICATIONS ON THE TECTONIC EVOLUTION OF THE MASSIF: NEW FIELD OBSERVATIONS AND RADIOGENIC AGE DETERMINATIONS

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ABSTRACT.- Understanding the ages and emplacement mechanism of the gneissic granites cropping out in large areas of the Menderes Massif has a critical importance in its tectonic evolution. Based on some radiogenic age data, the gneissic granites and the surrounding high-grade micaschists have been advocated to be the Precambrian "Core Succession" that was undergone high-grade metamorphism during the Pan-African Orogenesis. The micascists and marbles of Palaeozoic-Mesozoic age have been defined as the "Cover Succession" unconformably overlying the core assemblages. It has also been indicated that during the Alpine Orogenesis and by the Main Menderes Metamorphism the core and cover successions were metamorphosed together in relatively lower grade conditions. Although the Menderes succession underlies a large region in the western Anatolia and display uncovered outcrops, in nowhere structural evidences of the unconformity between Pan-African core and the Palaeozoic-Mesozoic cover series has been reported. It is very difficult to expect the Alpine Orogeny to erase the older Pan-African structures to a point of undefinable state. In this study, we have new mapped and examined the boundary relations between the so-called Pre-Cambrian core and Palaeozoic-Mesozoic country succession around Dibek Mountain, Çine-Yataðan and Ýncirliova Dam site. In Dibek Mountain gneissic granites were emplaced, as migmatitic fronts, into the marble lenses-bearing micaschists parallel to their foliation planes. Along the Çine-Yataðan road, migmatitic syn-tectonic granitic fronts injected into and engulfed the Palaeozoic black marble, black chert and micaschist alternation. In this region, the Palaeozoic units pass upward into the Triassic metadetritals with mafic volcanic intervals and they in turn grade into the Mesozoic platform-type marble succession. In this location the granites intrude into the Palaezoic and Mesozoic series which were paleontologically dated in some other areas of the masif. Similarly around the Ýncirliova Dam site the augen gneisse-migmatitic granite complex intruded into a complete stratigraphic section from Palaeozoic to Triassic and the Mesozic marble succession. New field data indicate that the high-temperature-type Barrowien Main Menderes Metamorphism caused a rejuvenation in the crust and granites with large migmatitic fronts emplaced syntectonically into the entire section of the masif from thick metadetrital units below and the Palaeozoic-Mesozoic cower series above. Precambrian and Alpine zircon ages determined from the gneissic granites could be explained by the rejuvenation and migmatism during the Main Menderes Metamorphism.

Key words : Menderes Massif, granite emplacement, core-cover series relation

INTRODUCTION

Granites with augen texture cover large areas in the Menderes Massif. In the literature their emplacement age are stated to be Precambrian (540-560 Ma) based on the radiogenic ages (Candan, 1994*a*; 1994*b*; 1995; 1996; Hetzel and Reischmann, 1996; Candan and Dora, 1998; Loos and Reischmann, 1999; Candan et al., 2001; Koralay et al., 2001; Gessner et al., 2004;

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Koralay et al., 2007). It has also been stated that the oldest succession of this metamorphic complex, named as the "Core" association, was metamorphosed under migmatitic conditions during the Pan-African Orogenesis (Schulling, 1962; Dürr, 1975; Dora et al., 1992; Candan, 1994a; 1994*b*; 1995; 1996; Candan and Dora, 1998; Candan et al., 2001; 2007). Micaschist and marble association of Palaeozoic-Early Eocene age has been named as the "Cover" and presence of a pronounced unconformity between core and cover has been inferred although in no where this boundary defined by structural evidence.

If the above mentioned core and cover definition is correct it would be assumed the core was deformed earlier under high-grade conditions, later, erroded and is overlain by Palaeozoic-Mesozoic succession. The Alpine tectonics was suggested, in earlier studies, to be caused by nappe emplacement in associate with relatively lower metamorphic conditions and without any granite emplacement (Dürr, 1975; Þengör et al., 1984; Candan, 1994*b*; 1995; Okay, 2001; Whitney and Bozkurt, 2002; Rimmele et al., 2003). In this case the unconformity between core and cover would be clearly seen and be mappable on 1/25.000 scale along the Menderes Massif which form extensive and open outcrops in the western Anatolia. Along the Taurus range in the eastern Anatolia, regionally metamorphic Bitlis Massif displays similar metamorphic characteristics to the Menderes Massif. In Bitlis Massif, the unconformable contact between the Precambrian and Palaeozoic-Mesozoic sequences is open and clearly definable by map pattern (Erdoðan, 1982; Erdoðan and Dora, 1983) although they were metamorphosed together during the Alpine Orogeny.

In some publications carried out between Bafa Lake and Yataðan region in the Menderes Massif, the core-cover boundary was defined as southerly-dipping crustal-scale detachment fault (Bozkurt, 1996; Bozkurt and Park, 1994; 1997*a*; 1997*b*; 1999; 2001; Hetzel and Reischmann, 1996; Loos and Reischmann, 1999; Bozkurt and Satýr, 2000; Bozkurt and Oberhansli, 2001; Lips et al., 2001; Whitney and Bozkurt, 2002). However, Erdoðan and Güngör (2004) described and mapped that boundary as a normal intrusive boundary of a syntectonic granite as no missing part between the gneissic granites and the marble lenses bearing schists is present. Besides that, the map pattern of this boundary forms a crescent concave to the north just opposite to what would be expected to the southerly-dipping detachment zone in which case must be concave to the south (Erdoðan, 2006).

In this study, the boundary of the gneissic granites are examined and re-mapped in three different regions. To the North of Salihli, the area along western side of the Demirköprü Dam in the Dibek Mountain region is mapped on 1/25.000 scale and the boundary of the gneissic granites with the country rocks are examined in detail (Figure 1). In the Dibek Mountain region, gneissic granites emplaced into micaschists with marble lenses. The similar marble lenses-bearing country schists have also been mapped around Koçarlý, in the north of the Bafa Lake. However, in the previous studies, Precambrian core association with marble lenses has not been described in the massif.

Besides the Dibek Mountain along the southern border of the masif the second region is studied along the Çine Yataðan road (Figure 1). In this area the gneissic granites directly intruded into the well known Palaeozoic and Mesozoic successions dated by fosil content in the southern parts of the masif (Önay, 1949; Konak et al., 1987; Güngör and Erdoðan, 2001). The third area mapped is the Ýncirliova Dam site recently under constructionto the North of Ýncirliova (Aydýn) (Figure 1). Along the newly opened cross-cuts of the excavation areas there are open and clealry observed outcrops between the gneissic granites and the country rocks. We mapped and examined this area on 1/10000 scale. By detailed field photographs and pre-

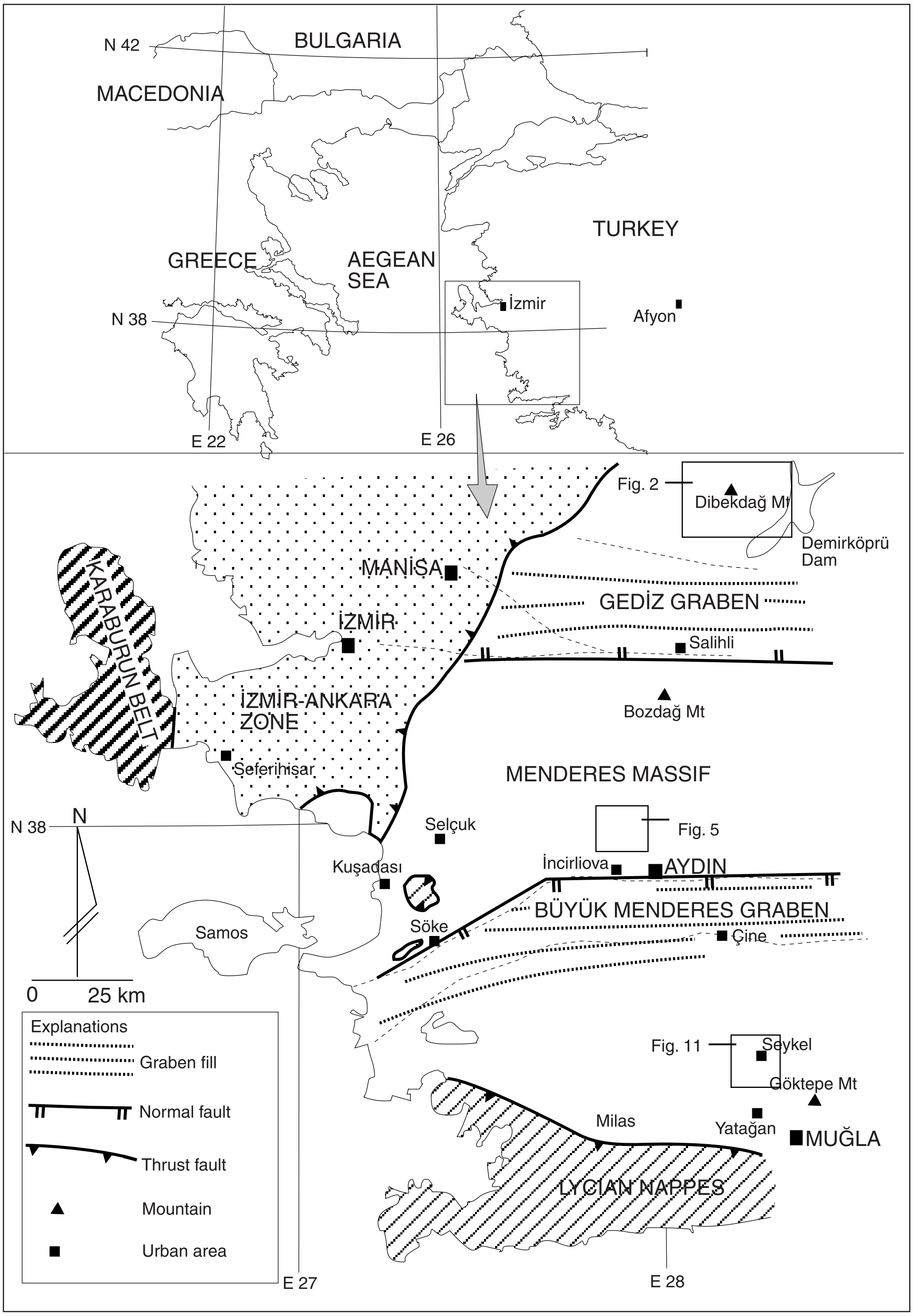

Figure 1- Major tectonic belts cropping out in western Anatolia and location map of the areas studied.

pared thin sections in these three regions both the internal structure and the boundary characteristics of the augen gneisses are described.

Comparing the Menderes succession with the various parts of the Tauride-Anatolide Platform is importent in understanding the tectonic evolution of the masif. In this study, the Menderes successions are compared with the Tauride successions in Sandýklý (Afyon) region and the Bitlis Massif in Avnik (Bingöl) region.

BOUNDARY RELATIONS BETWEEN THE GNEISSIC GRANITES AND THE COUNTRY ROCKS IN DÝBEK MOUNTAIN

Dibek Mountain where the three-dimentional forms of the gneissic granites form open outcrops, is located in the western side of the Demirköprü Dam and forms the eastern limp of the NE-SW-trending antiform (Figures 2, 3). In this region gneissic granites were emplaced as NE-SW-trending and SE-dipping granites zones paralel and concordantly into the foliation planes of the hosting mica schists (Figures 2, 3). The core and the limp of the antiform are made up of quartz mica schists and biotite muscovite schists. In the metadetrital succession there are white marble lenses with lengths and thickness of mappable on 1/25.000 scale. One of these marble lenses shown on figure 2 and 3, was engulfed by the gneissic granites along its northern extension and converted into coarse grained calcite marble with 3-4 mm- large crystals. The map pattern, cross section characteristics, structural properties and diffuse boundary relations are typical granitic emplacement as migmatitic fronts. Toward the migmatitic fronts and close to the boundary, up to 2-3 km in distance to gneissic granites, micaschists become cleaved and shiny with 3-4 mm-large biotite flakes and red transparent almandine porphyroblasts. Kyanite and sillimanite accompany with this paragenesis in some places. In the field along road cuts almandine biotite schists include 1-2 cm-long quartzo-feldspatic lenses and augens. As the granitic bodies approach eastern, their numbers and thickness increase and become septums parallel to the foliation of the biotite schists. After the contact and inside the granitic body, there are numereous biotite schist restites partly to completely assimilated. Eastward, toward the Demirköprü Dam these restites decrease in number and homogenous augen textured granites become dominant (Figures 2, 3). The foliation of the granites and the country schists are completely conformable to each other. During the granite emplacement in situ assimilation of the country schists caused this conformable relation of centimeters to outcrops scale. The migmatitic gneissic granite fronts, in places, cover an area of several 1/25.000 scale topographic sheets. In situ assimilation of the country schists caused gradational changes from leucocratic granites to biotite-rich melanocratic gneisses and into biotite schists. Geological map patterns, relations in the third dimension in cross-sections, and structural concordance with the country schists are typical for the emplacement mode and mechanism of the granites in the Menderes Massif and are repeatedly observed in differenet regions (Bozdað, Kiraz) in western Anatolia.

In some rare areas of the massif such as Bafa Lake region, an intrusive contact between the gneissic granites and the country schists is found along which enclaves from the schists are found in the granites. These rare and very narrow intrusive parts might be shallower regions in the crust that the melt formed in deeper parts, moved upward and formed injections in the shallower parts. The Bafa Lake region of the Menderes Massif is an exeptional area where the gneissic granites do not display the typical geometry of the migmatitic syn-tectonically emplaced granites.

THE ÇÝNE-YATAĐAN ROAD

The Çine-Yataðan road is one of the rare areas were the gneissic granites are in a close contact with the Palaeozoic-Mesozoic succes-

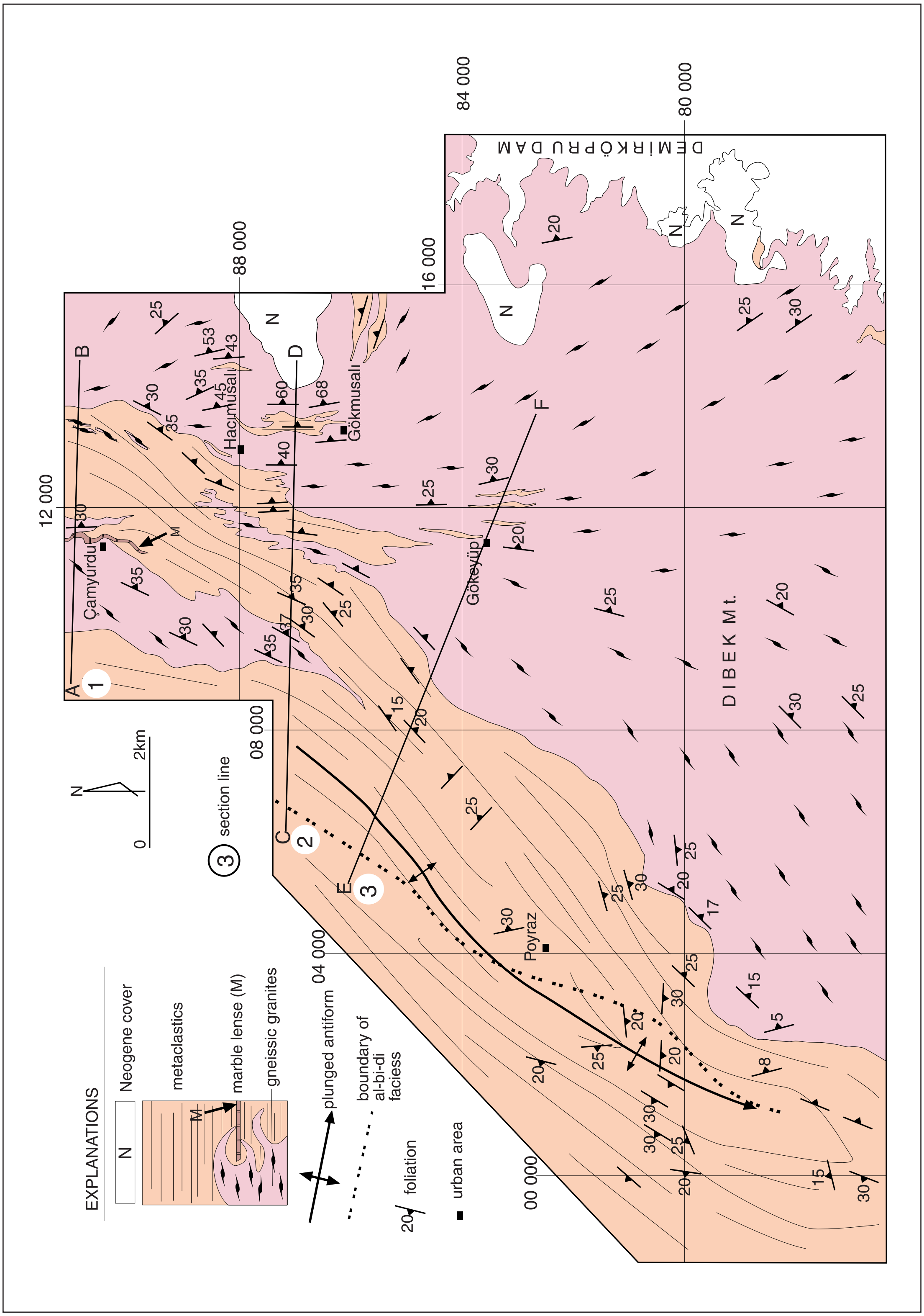

Figure 2- Geological map of the Dibek Mountain area.

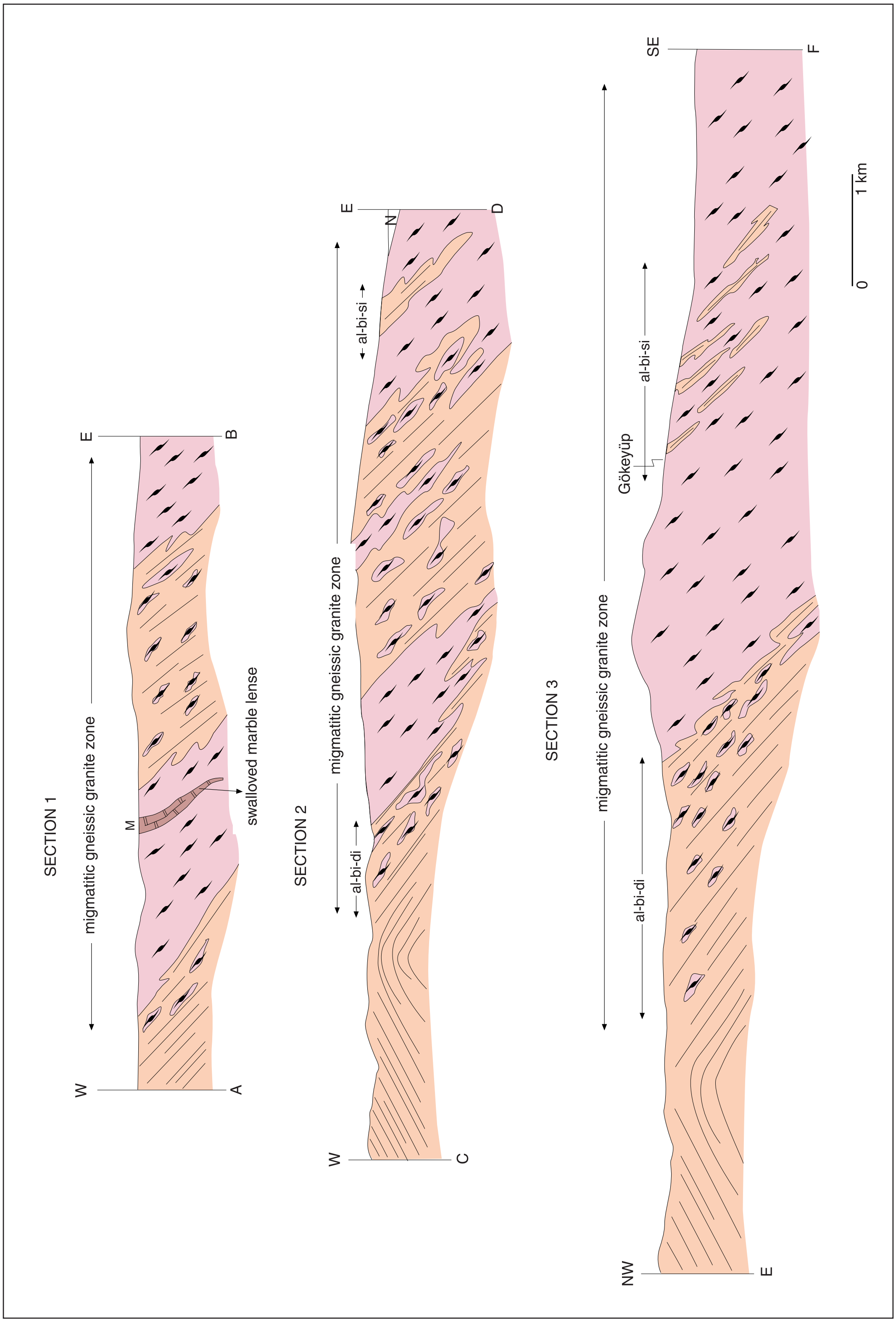

Figure 3- Cross-sections showing the contact relations between the gneissic granites and marble-bearing mica schists in Dibek Mountain Area (for section lines see figure. 2, al: almandine, bi: biotite, di: kyanite, si: sillimanite).

sions of the massif. To the South of Yataðan region the Palaeozoic and Mesozoic series were only metamorphosed under low-grade conditions and they were paleontologically aged (Önay, 1949; Konak et al., 1987; Güngör and Erdoðan, 2001). In the map area around Kafacakaplancýk, Irmadan and Seykel villages (Figures 1, 5) the gneissic granites intrude into this Palaezoic-Mesozoic succession which named previously as "the Cover"

Along the road, gray phyllite, black chert and metaquartzite intercalations are in contact with the gneissic granites (Figure 5). This intercalation is known around Göktepe forest fire lookout tower with fossil content of fusulinids, corals and crinoidal fragments as Carboniferous-Permian in age and named as Göktepe Formation (Önay, 1949; Konak et al., 1987). The fosil-rich Göktepe Formation has been mapped and found to extend to the map area along the Çine-Yataðan road (Önay, 1949; Konak et al., 1987; Güngör and Erdoðan, 2001). In the map area, to the South, this Palaeozoic unit is overlain by an intercalation of yellow marbles, metadetritals and mafic metavolcanics. Around the Göktepe region their age is determined as Triassic by their fossil content (Güngör and Erdoðan, 2001).

In the study area the Triassic metadetrital succession passes into the Mesozoic massif white marbles including emery lenses in the upper parts. Upward, these platform-type marbles are overlain by thin, red laminated pelagic marble, and metaserpentinite blocks-bearing metadetritals in ascending order which are known in various parts of the massif as Campanian-Meastrichtian in age (Konak et al., 1987; Güngör, 1998; Güngör and Erdoðan, 2001; Özer et al., 2001). Therefore along the Çine-Yataðan road the Entire Palaeozoic-Mesozoic succesion crops out.

In the northern part of the map area around Irmadan and Seykel villages the gneissic granites cut the Palaeozoic units. A serious of photographs show, both on the map and cross-section (Figures 5, 6, 7, 8, 9), that the gneissic granites are injected as migmatites into the foliation planes of micaschist and marble intercalation. Along the boundary zone between the marble lenses and and the migmatitic granites, the preserved magmatic texture is observed in thin sections although the entire boundary zone is strongly deformed (Figure 9). At the contact zone in the black marbles large almandine-bearing calc-schists are common. Along the road of Seykel village granitic seams are found both above and below the black marble lenses (Figure 10).

The boundary zone is several kms wide in the Dibek Mountain region whereas between Irmadan and Seykel the same zone is only 100 m in width. But similar to the Dibek Mountain area granitic seams and surrounding foliation are always conformable and in situ melting along the boundary of the granitic seams are common. The numbers of quartzo-feldspatic seams increase northward approaching to the granitic body and finally graduated into the homogenous granites. In the granitic body strongly assimilated mica schist restites with a length of several meters are common.

Formation of foliation in the country rocks and emplacement of the granites occured together. So that the entire boundary zone display regional foliation. This relation has been wrongly interpreted as a crustal-scale detachment zone by Bozkurt (1996), Bozkurt and Park (1994; 1997*a*; 1997*b*; 1999; 2001), Hetzel and Reischmann (1996), Loos and Reischmann (1999), Bozkurt and Satýr (2000), Bozkurt and Oberhansli (2001), Lips et al. (2001), Whitney and Bozlurt (2002) and into the shear zone leucocratic granitic seams have been suggested to injected (Bozkurt and Park, 1994; 1997*a*; 1997*b*). On the other hand, the foliation is not restricted a particular zone but the entire granitic body and the country rocks are deformed and attained penetrative foliation. In situ melting along granitic seams in the marble-schist intercalation is recognized as diffuse, centimeter- to

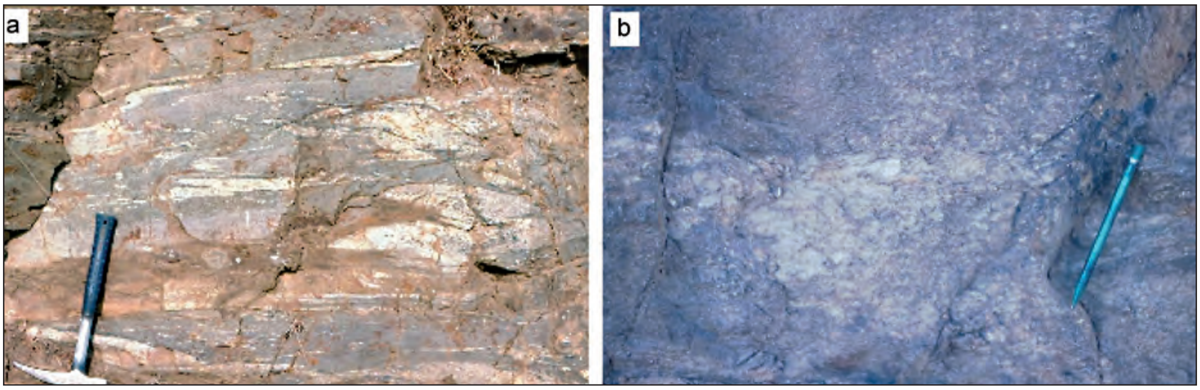

Figure 4- Field photographs showing the gradational diffusive contact relations between the gneissic granites and mica schists formed by in situ melting: a) Toward the gneissic granites mica schists include granitic seams in increasing number and thickness. b) Close-up view of granitic seams in mica schists.

milimeter-scale gradational passes from granitic texture to quartzo-feldspatic layers to surrounding schists.

ÝKÝZDERE DAM SITE

Five km far and in the north of Ýkizdere town (Aydýn), along the recently opened cuts, the intrusive boundary relation between the gneissic granites with pronounced augen texture and the Palaeozoic-Mesozoic succession is observed (Figure 11). The Palaeozoic unit consist of black cherts, quartz mica schists, black phyllites and dark gray marbles (Figure 12). Above this intercalation yellow marble, mica schist and quartz metaconglomerate (Figure 12) association is present and in the uppermost parts there are thick, white massive marbles. Black chert and black marble associaiton is lithologically similar to the Palaeozoic parts of the Menderes Massif and yellow marbles, quartz metaconglomerates and overlying white marbles are the Triassic units of the masif. Only difference with the Triassic sections are the absence of mafic volcanic intervals in the Ýkizdere dam site. But the Triassic units show lateral facies changes in large area of the masif as described by Güngör (1998). Neogene conglomerates, sandstones and claystones overlay the metamorphic complex unconformably in the map area.

In the Ýkizdere Dam site along two seperate outcrops the gneissic granites lay inside the Palaeozoic and Mesozoic parts of the masif. In the SW corner of the map (Figure 11) the granites display augen texture with K-Feldspar porphyroblasts of 2-3 cm length. These augen gneisses intrude both the Palaeozoic and Mesozoic units (Figure 11). The augen gneiss - country rock boundary, starting from the spillway of the dam, intruded into the Palaeozoic units and westward into the Mesozoic white marbles (Figure 11).

Along the spillway cuts, granitic seams, ranging from centimeters to meters in thickness, injected conformably into the black cherts and black marbles (Figure 13). In thin sections in some areas granitic texture are stil recognized in the foliated gneisses (Figure 14). These granitic seams are rich in tourmaline crystals and texturally similar to the augen gneisses in the SW corner of the map area (Figure 13a and b).

In the E-NE corner of the map area, there is another gneissic granite body (Figure 11). A NE-SW trending derivation tunnel, excavated for the dam, starts in the north from the granites and after about 200 m it enters into the Palaeozoic black marble, calcschist, mica schist intercalation

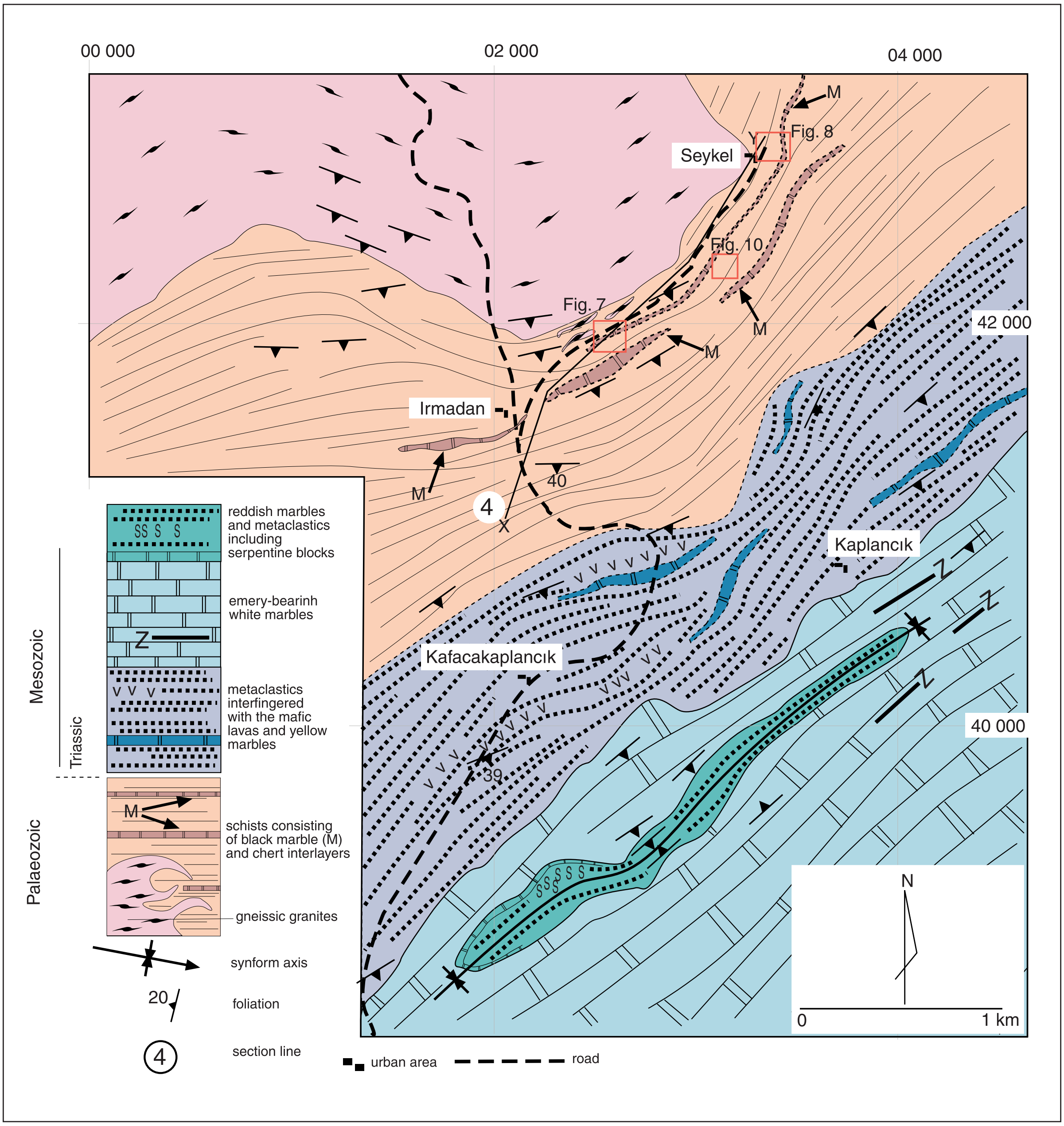

Figure 5- Geological map of the Çine-Yataðan road.

along a sharp contact (Personal communication with A. Rýza Özdemir, 2007). In the SE part of the map in Figure 11, at Ballýkaya site, metamorphic succession consists of yellow marbles and meta-conglomerates that are overlain by white massif Mesozoic marbles. In this Mesozoic section, there are numerous granitic seams and where the marbles are in contact with these granitic seams, marbles are turned into calc schists with large garnet porphyroblasts. Around the Ýncir-

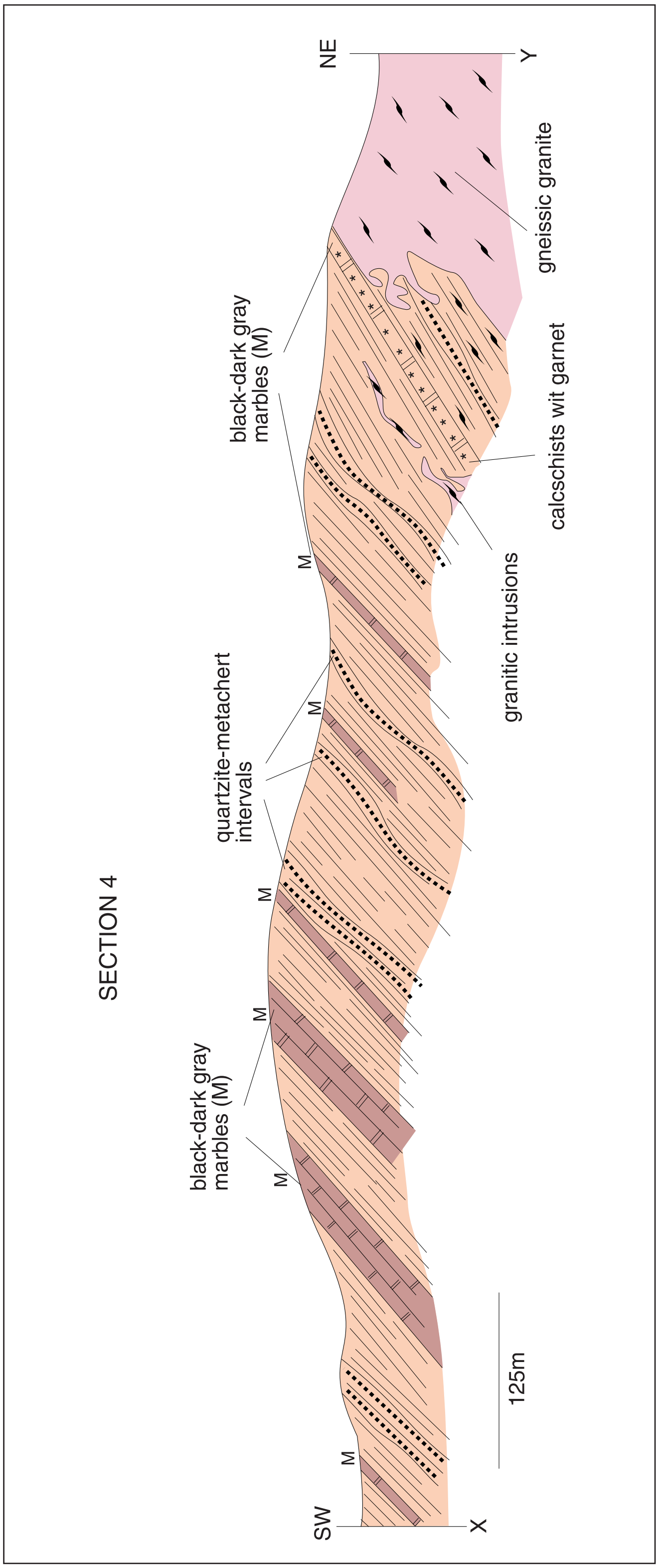

Figure 6- Cross-section showing the relation between gneissic granites and metadetritals in Çine-Yatađan road-Seykel area. For section line see figure 5.

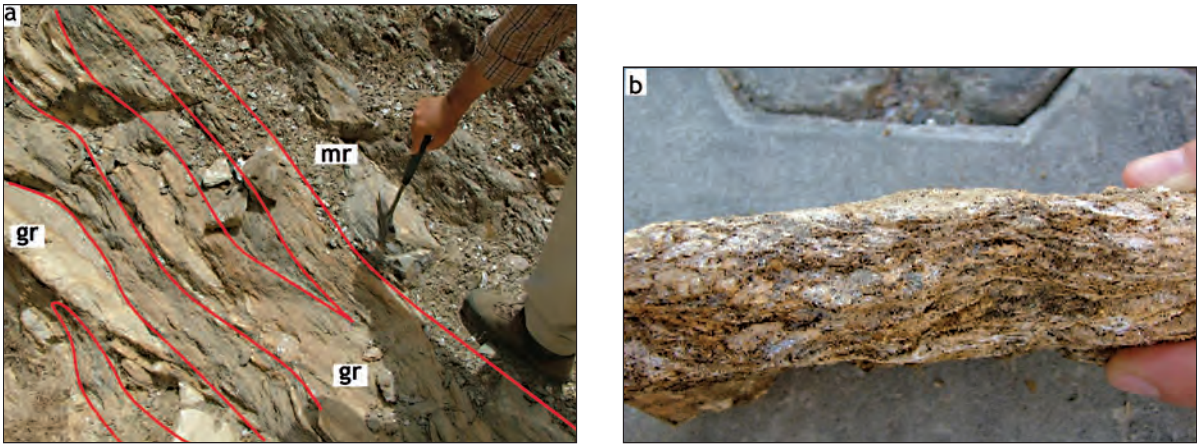

Figure 7- Granitic seams in Palaeozoic schists at Seykel village (Figures 1, 5) road cuts: a) Granites in different thickness emplaced parallely along the foliation planes of marble lenses-bearing schists, b) Just below the marble lenses, typical magmatic texture with quartz and feldspars is seen in granitic seams. Photomicrograph taken from this handspecimen is given in Figure. 9 (Dotted lines: contacts, solid lines: foliation. Mr: marble, gr. Granite, for location of photographs see Figure. 5, UTM 602564 / 4141888).

liova Dam site, both the map pattern and close-up views of the gneisses and country rocks indicate that they were emplaced into the Palaeozoic-Mesozoic succession during formation of their foliation.

COMPARISON OF THE MENDERES SUCCESSION WITH THE PALAEOZOIC AND PRECAMBRIAN UNITS OF THE TAURUS BELT

The western part of the Tauride-Anatolide Platform in the Menderes Massif and eastern side in the Bitlis Massif were metamorphosed during the Alpine Orogeny. But large parts along the Taurus range were not affected by this metamorhism. The oldest parts of the platform crop out around the Sandýklý region and they are not metamorphic (Erdoðan et al., 2004). In the Sandýklý region there is the Kocayayla Group of more than 3 km in thickness in the lowermost part of the succession (Figure 15). In the lowermost parts, the Kocayayla Group consists of quartzite and phyllite association and passes upward into the mudstones and rare limestone intercalations-bearing detritals with felsic lava and pyroclastics. In the mudstones, there are trace fossils that have yielded Early Cambrian age. The Kocayayla Group ends in its upper parts with quartzite and phyllite association. The Early Cambrian Kocayayla Group is unconformably overlain Hudai quartzites passing upward into the Çaltepe formation of dolomite and nodular limestones. The Çaltepe formation contains abundant trilobite and conodont fosil assemblage of Middle-Late cambrian age (Gedik, 1977; Dean and Özgül, 1994). Above the Çaltepe formation there is the Late Cambrian-Early Ordovician Seydiþehir formation of thick mudstone succession. In the Sandýklý region a platform succession starting from the detrital Triassic unit to Eocene carbonates overlays unconformably the older units. Thick metaquartzite, mica schist and rare marble succession cropping out in extensive areas in the Menderes Massif that have been mapped and defined as the Pan-African core association (Dora et al., 1992; Candan, 1994*a*; 1994*b*; 1995; 1996; Hetzel and Reischmann, 1996; Candan and Dora, 1998; Loos and Reischmann, 1999; Candan et al., 2001; Koralay et al., 2001; Gessner et al., 2004; Koralay et al., 2007; Bozkurt, 1996; Candan et

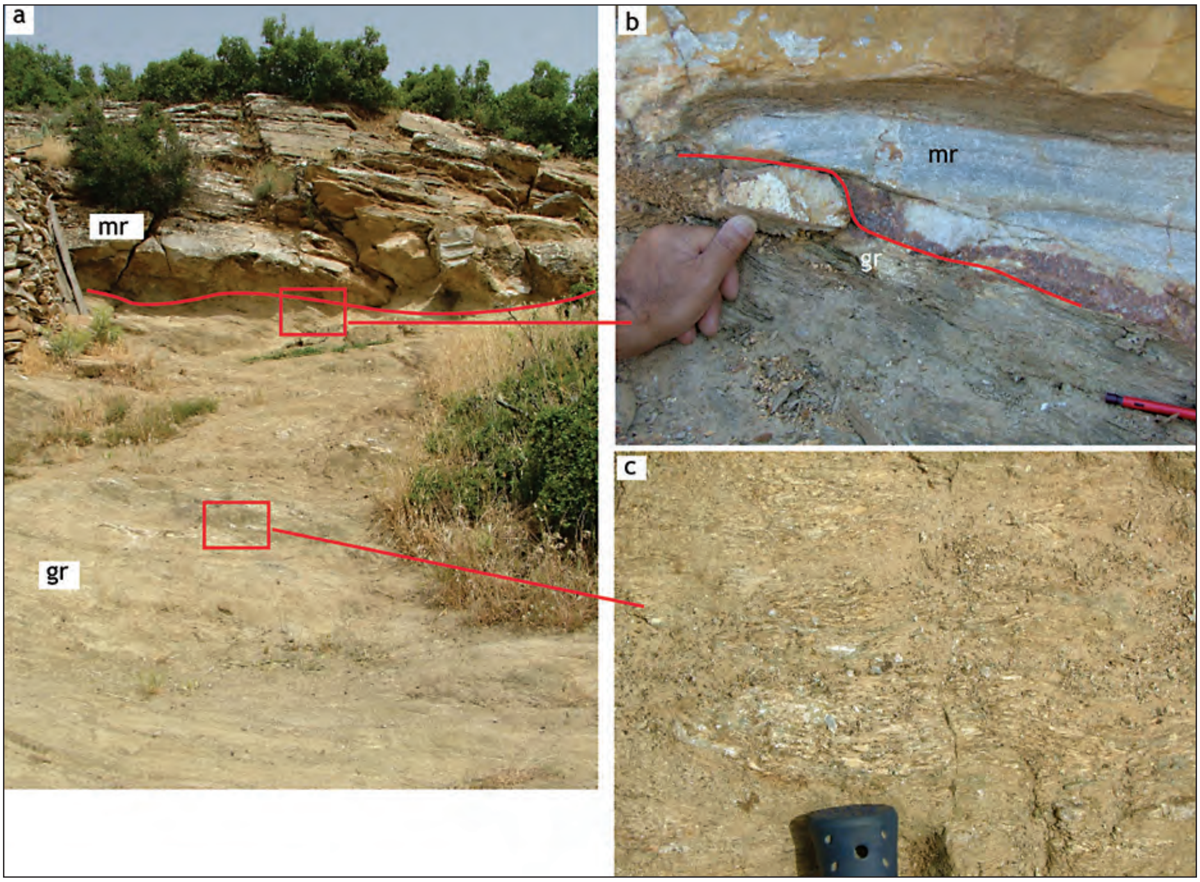

Figure 8- Near Seykel village (Figures. 1, 5) main granitic body engulfes the marble lenses where partly migmatized schists and granitic seams are in contact with the marbles including garnet porphyroblasts of 0,5-1 cm in length. a-b) Marble lenses intercalating with the schists which are engulfed and assimilated by migmatitic granites (in Figure b, finger shows gneissic granite seam touching marble lense). c) Gneissic granites with magmatic texture in schists. In situ migmatization is seen in the upper part of the hammer (Dotted line: contact, mr: marble, for location of photographs see figure. 5, UTM 603656 / 4142751).

al., 2006; 2007) resemble in facies and thickness to the Kocayayla Group and might be Early Cambrian in age.

The eastern part of the Tauride-Anatolide Platform had been metamorphosed as the Menderes Massif and forms the Bitlis Massif. In the north of Selvi town and around Avnik village (Bingöl), the Bitlis metamorphics consist of Lower and Upper associations (Figure 16a). The Upper association is represented by 2 km-thick white and gray marble succession of Palaeozoic-Mesozoic age characterizing the platform of the Taurus range. Below the platform marbles there is more than 1 km-thick lithologically uniform garnet mica schists unconformably overlying the Lower association along a thin metaquartzite basal level. These uniform mica schists resemble to the Ordovician-Silurian Seydiþehir formation of the Taurus range.

The Lower association is more than 2,5 km in thickness around Avnik village and consists of mafic volcanics, amphibolites and leucocratic

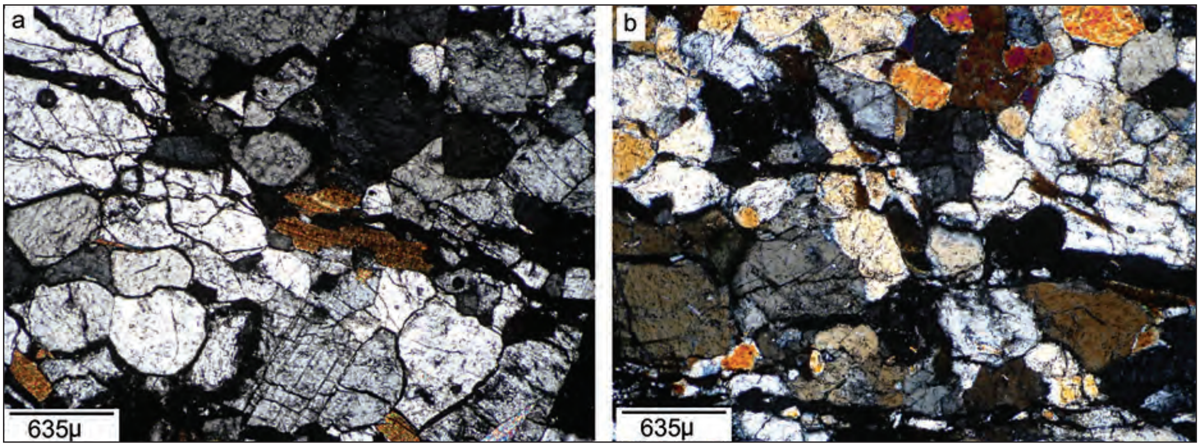

Figure 9- Typical magmatic texture and mineral assemblage are seen in granitic seams along the Seykel Village road (Figures 1, 5). (Hand specimen: Figure. 7b. Photographs were taken in cross-polarized light. This sample was collected below the marble lense shown in map).

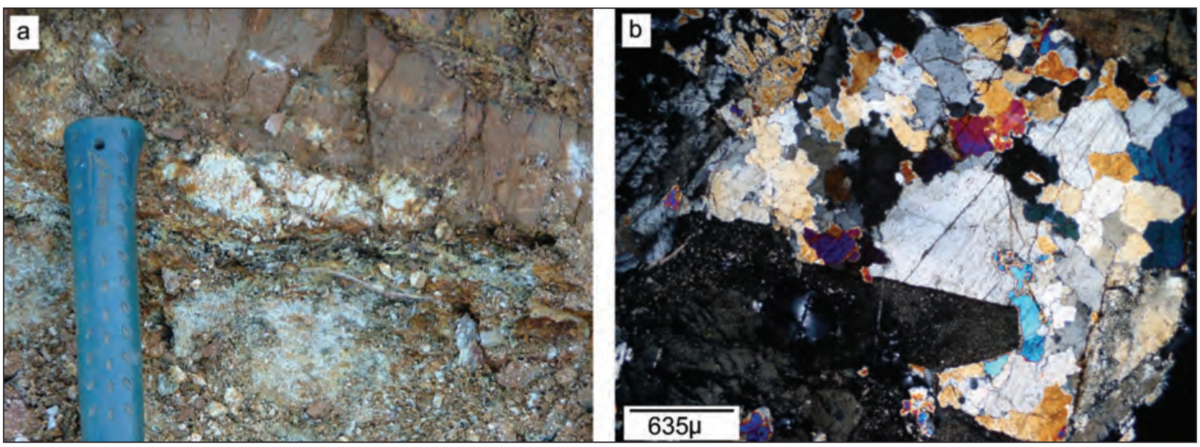

Figure 10- Along Seykel village (Figures. 1, 5) granitic seams are found also above the marble lense. Deformation did not errased the granitic texture. a) Field view, b) Deformed granitic texture in thin sections (For location of photograph and thin sections, see figure. 5, UTM 603003 / 4142130, photo-micrograph was taken in cross-polarized light).

gneisses (Figure 16a, b). In the metavolcanic succession there are laterally contunious banded iron formations. Apatite-bearing banded iron formations in the north of Avnik village along the Mur valley (Figure 17) interlayered with the metatuffs, mafic and felsic metavolcanics and formed as lenses paralel to the original stratigraphic layers in the volcano-sedimentary succession (Erdoðan et al., 1981; Erdoðan, 1982). The Lower association is cut by migmatitic leucocratic and fine grained Avnik granitoid. The heterogenous pile of the Lower association, cropping out below the Ordovician-Silurian Upper association and separated from it by a pronounced unconformity, has been interpreted as the Precambrian in age (Erdoðan et al., 1981; Erdoðan, 1982). Banded volcanostratigraphic iron formations are characteristics in the Precambrian units all around the world (Bankes, 1973; Goodwin, 1973; Banerji, 1977; Kimberley,

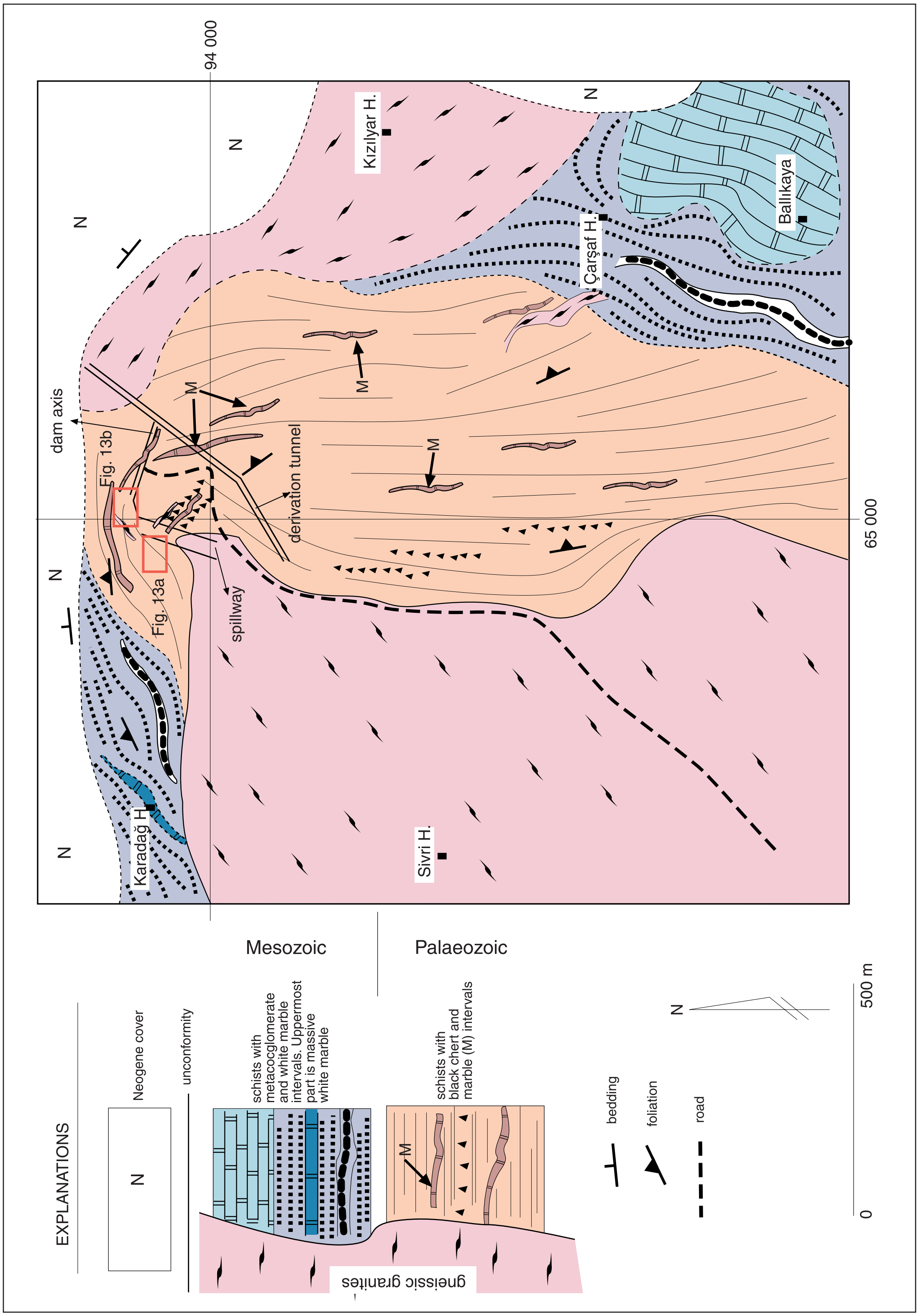

Figure 11- Geological map of the Ýkizdere Dam site

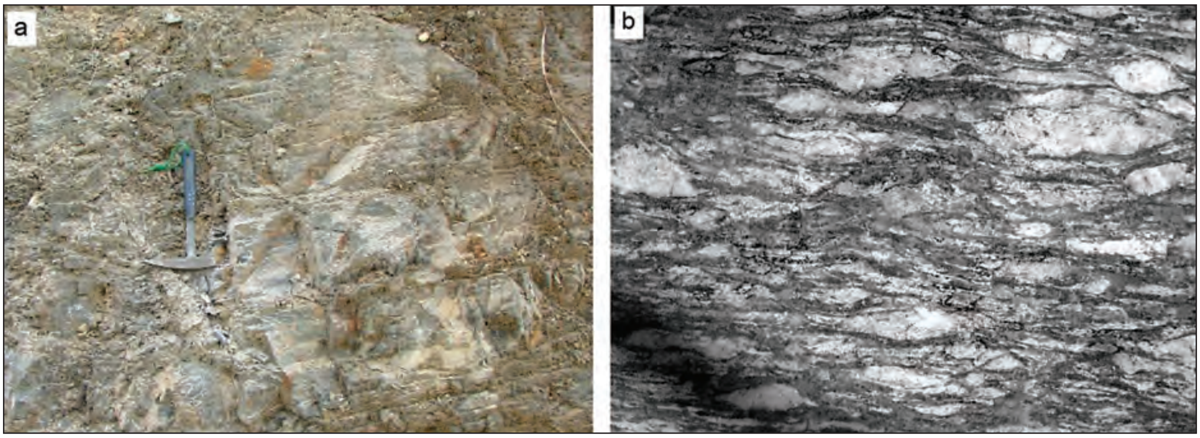

Figure 12- Characteristics of the Palaeozoic-Mesozoic sequence around Ýkizdere Dam site. a) Black chert s (UTM 564972 / 4193988), b) Conglomerate intercalations in Mesozoic units overlying the Palaeozoic succession (UTM 564591 / 4194107) (Lines on photographs show foliation).

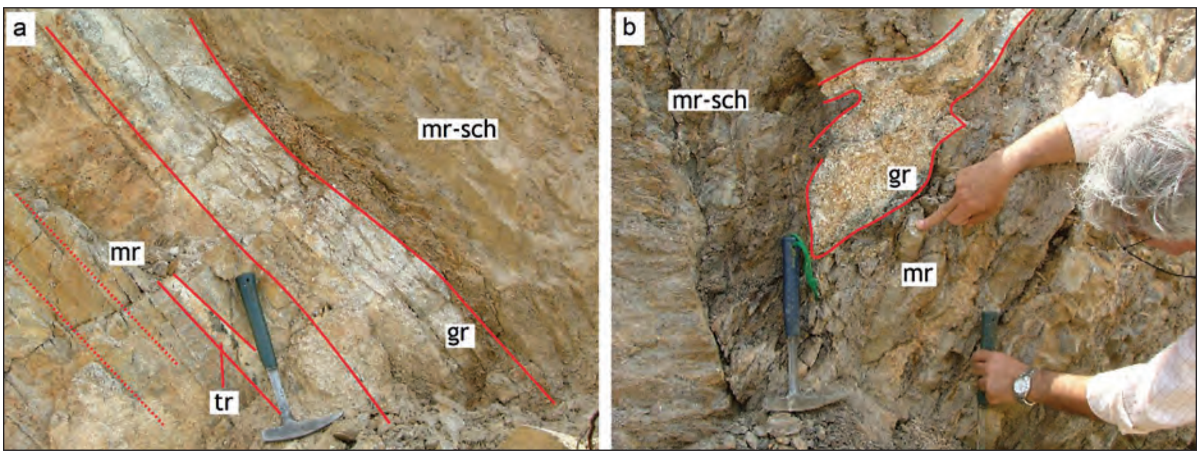

Figure 13- a-b) At the spillway cuts of the Ýkizdere Dam, granites with tourmaline crystals emplaced into the marbles parallely to the foliation. In these outcrops quartz and feldspar-bearing mineral association and magmatic texture are clear in granitic seams. Tourmaline layers are found (Dotted lines: foliation, solid lines: contacts, mr: marble, gr. granite, tr. tourmaline, for location of photographs see Figure 11. a: UTM 565038 / 4194140, b: UTM 565038 / 4194156)

1978; Gole, 1981). In the core-succession of the Menderes Massif, however, there is no economically important banded iron formations and besides that they are almost devoid of any metallic mineralizations. For this reason the thick metadetrital succession of the Menderes which is considered to be Pan-African basement do not have any resemblance with internally continuous banded iron and apatite-bearing thick Precambrian metavolcanic succession of the Bitlis Massif.

In the Bitlis Massif the Lower association is cut by migmatitic granites and deformed. The

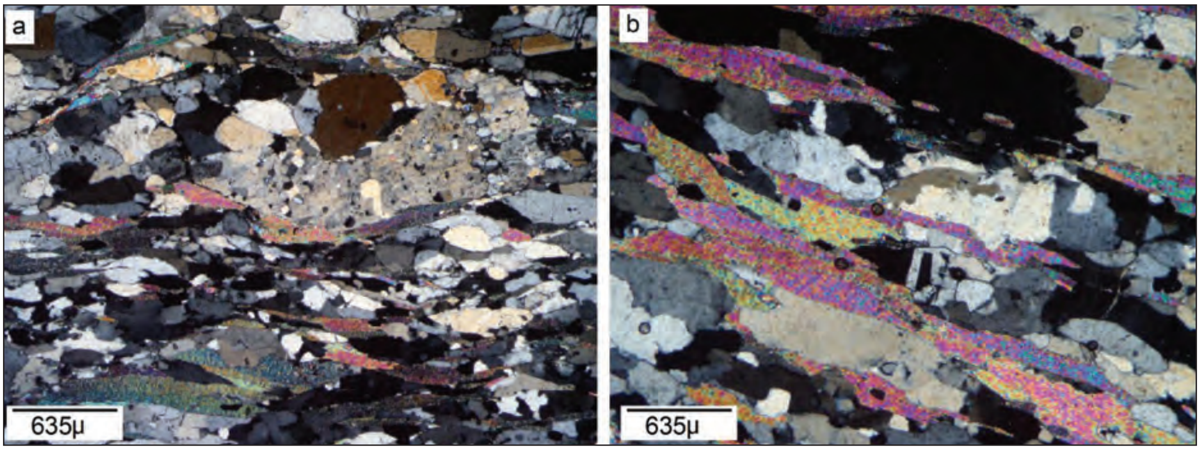

Figure 14- Granitic texture is clealrly determined in thin sections from gneissic granites in marble-gneissic granite alternation. Granites are made up of quartz, feldspar and biotie (Photomicrographs are cross-polarized in light).

Upper association unconfomably overlies these units. Later during the Alpine Orogeny, both groups deformed and metamorphosed. However, the unconformity between the Precambrian units and the Palaeozoic Upper association is still clearly definable and mappable. Along the unconformity surface (shown on Figure 16a by arrows) the basal quartzites of the Upper association directly overlay different units of the Lower association.

Stereographic projection of the mesoscopic fold axis which were measured on a relatively not complicated area covering both the Upper and Lower associations in the North of Avnik Village is given in figure 18. In the Upper association prominent fold axis is N40-50W/30-60 NW in setting. In the Lower association though, two different groups of fold axis are plotted, one is conformable with those found in the Upper association coinciding the Alpine deformation, and the other (EW/10-30W) is inherited from the older deformation before the Ordovician-Silurian unconformity. Although the region had experienced strong Alpine deformation and metamorphism the old deformation were not errased and stil recognizable.

DISCUSSION AND CONCLUSIONS

In the Dibek Mt, Çine-Yataðan and Ýncirliova regions the gneissic granites emplaced into the surrounding rocks syntectonically and display wide migmatitic zones. The granites and the country rocks deformed together attaining concordant foliation. On regional-scale foliation zones channalized migmatitic fronts form quartzofeldspatic augens and septums in thickness ranging from cm to map scale. Along these three regions there are always transitional passing from the country rocks into the granitic bodies. In the direction to the granitic bodies grade of metamorphism increases and almandine biotite kyanite sillimanite paragenesis appears in shiny biotite quartz mica schists. In the biotite schists with red clear almandine porphyroblasts, quartzofeldspatic lenses are observed in the field. These granitic seams and augens increase in number and thickness toward granitic bodies in the migmatitic fronts. There are abundant partly or completely engulfed and digested restites of biotite schist as it is entered into homogenous granitic bodies in which sillimanites join into almandine-biotite-kyanite paragenesis. The sillimanite-bearing restites display polygonal texture under microscope suggesting partial melting and

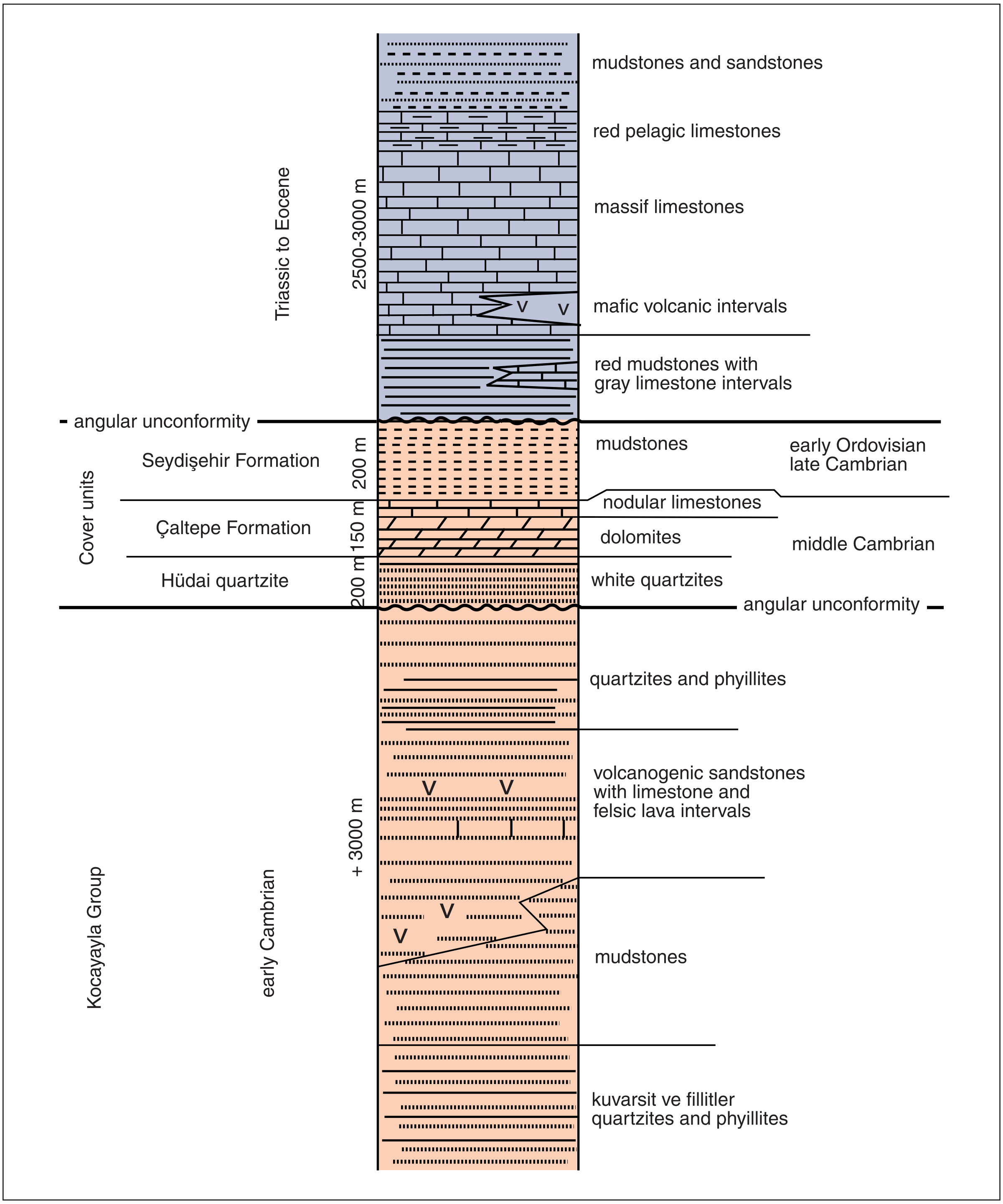

Figure 15- Generalized stratigraphic column of the Sandýklý region (After Erdoðan et al., 2004). Palaeontologically aged Palaeozoic units are lithologically resemble to the metadetrital units of the Menderes Massif.

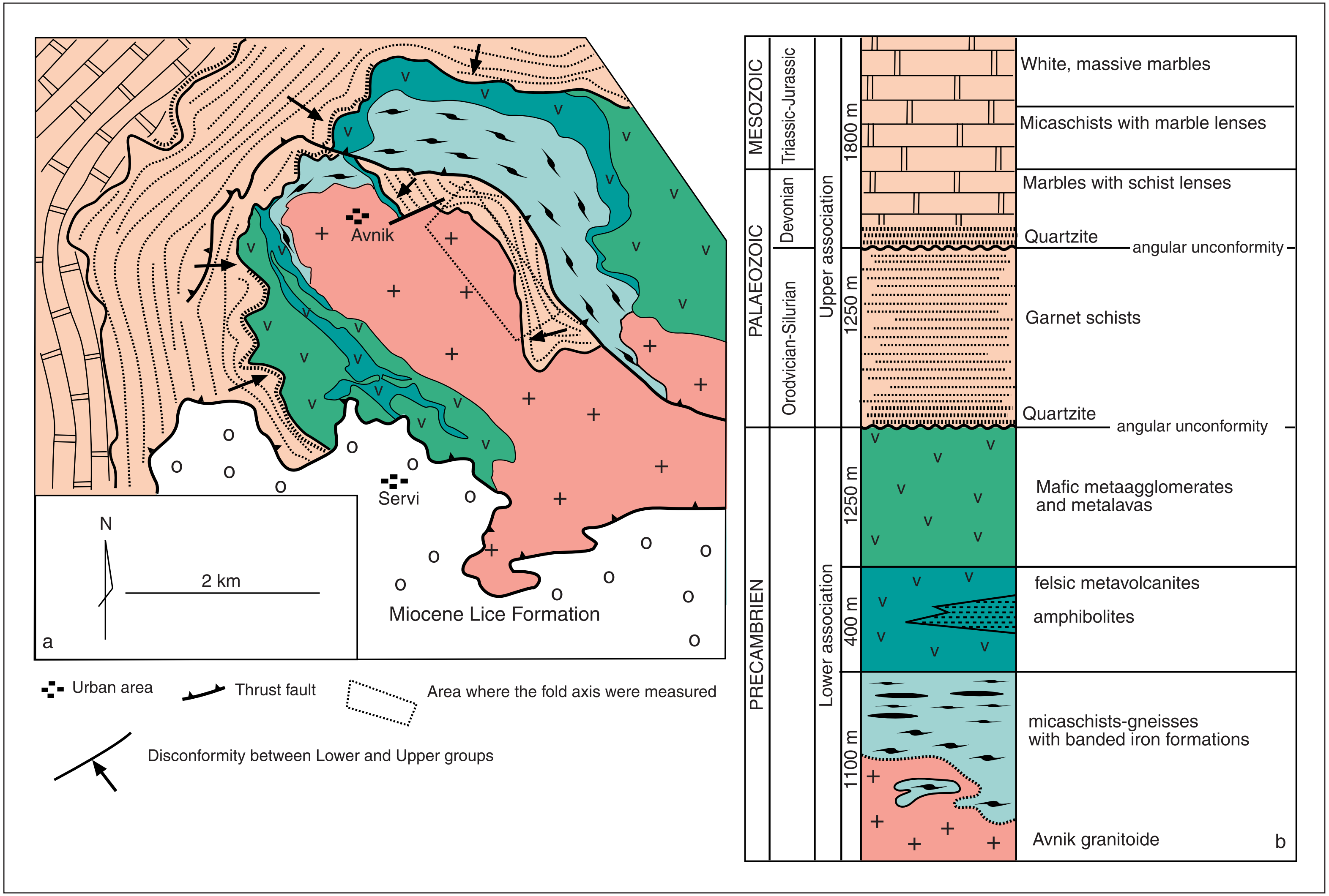

Figure 16- a) Geological map of the Bitlis Massif around Avnik Village (Bingöl). Unconformity between the precambrian and Palaeozoic-Mesozoic units is definable and mappable although the Alpine metamorphism (Arrows in map show the unconformity surface). The mesozcopic fold axis given in figure 17 were measured in the squarred area (For explanations of map, see figure 16b) b) Generalized stratigraphic column of the Avnik region. Lateral equivalents of the Precambrian units are not found in the Menderes Massif (Simplified after Erdoðan, 1982 and Erdoðan and Dora 1983).

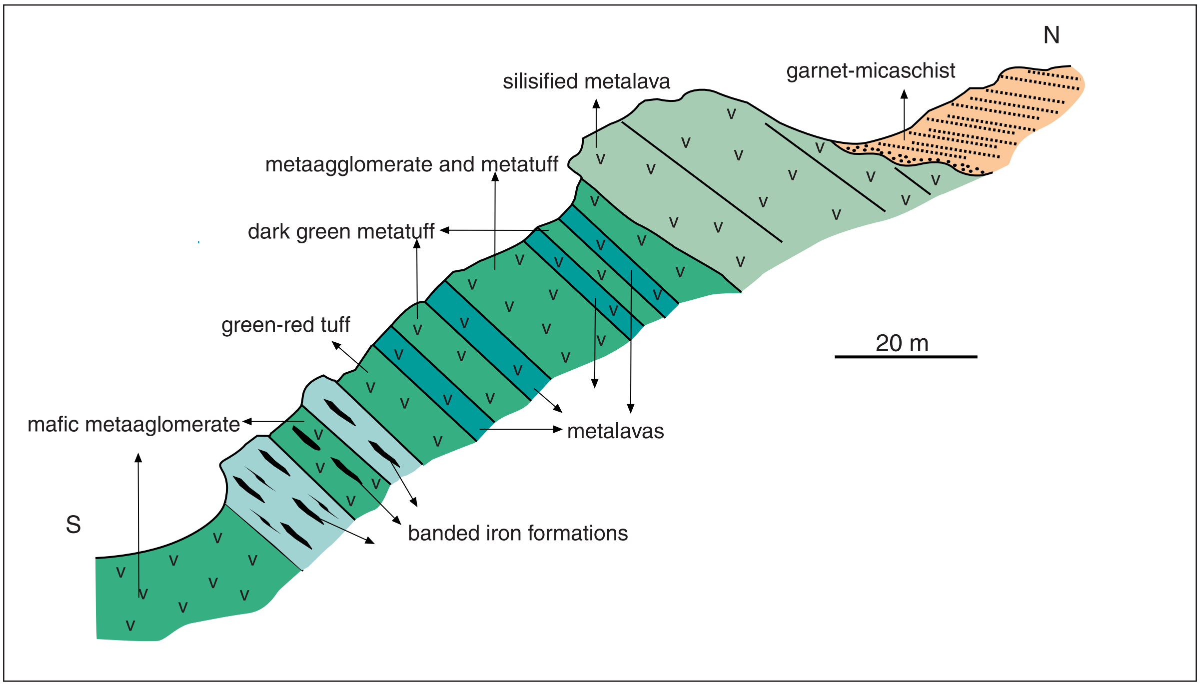

Figure 17- Cross-section showing stratigraphy of the Precambrian units in the Bitlis Massif along Mur valley (north of Avnik) (After Erdoðan et al., 1981). Banded iron formations are found concordant intercala tions in mafic metacolcanics.

they are common in gneissic granites in various parts of the masif and mapped as leptite gneisses in previous works (Dora et al., 1988; 2001).

In the previous studies based on the dips of foliation and an assumption that the gneisses would always be structurally lower attitude, the boundaries with foliation dipping toward the migmatitic fronts were interpreted as thrust faults. Besides, increase in grade of metamorphism along the dip of foliation toward migmatitic granites were wrongly interpreted as overturned limbs. On the other hand, in the map areas the country schists were melted in situ and strongly assimilated during the formation of foliation and the foliation surfaces channalized the migmatization (Figure 4). So that while foliation was forming on a regional scale migmatization fronts prefered the shear surfaces that caused in situ melting. Dibek Mt region is an example of this type of syntectonic emplacement of migmatitic fronts. In the direction of increase in metamorphic grade it is gradually entered into migmatitic front where lithologically homogenous but still strongly foliated granitic bodies crop out. Thus, the central parts of large granitic bodies would be a magmatic realm. During such a syntectonic granite emplacement gneissic granites structurally overlying the mica schists do not mean an overturned sequence or tectonic imbrication. Around Ödemiþ and Kiraz region small diabasic or gabbroic dykes and stocks are found completely digested in the magmatic realm. These strongly assimilated and digested ghosts of diabasic rocks were defined as eclogite or granulites and interpreted as the older metamorphic phases (Candan, 1994*a*; 1994*b*; 1995; 1996; Candan and Dora, 1998; Candan et al., 1998; 2001). Defining metamorphic phases and granites in magmatic centers of the gneissic granitic fronts would lead us wrong conclusions

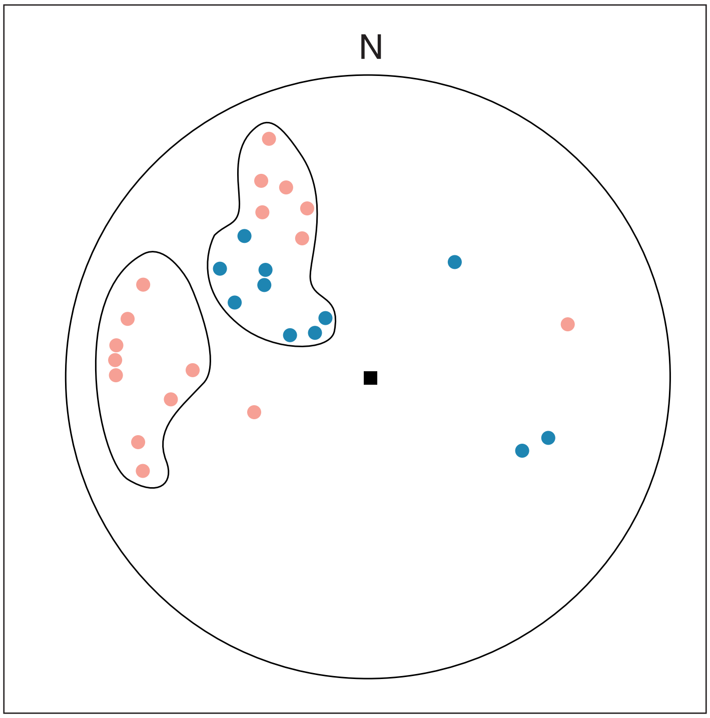

Figure 18- Stereographic projection of the Mesozcopic fold axis measured from both the Lower and Upper associations in Avnik region (Red dots: axis measured from the Lower association, blue dots: axis measured from the Upper association)

and be carefully accepted. In earlier studies under microscope magmatic phases with charnockites were defined in central parts of magmatic fronts (Candan, 1995). They would also be accepted carefully.

At the border zones of the gneissic granite bodies there are various magmatic phases and types of granites. For example near Bafa Lake there are aplitic dykes, muscovite-rich granites, biotite-rich phases, tourmaline-bearing leucocratic gneissic granites and metagranites. These different phases, on the other hand, show close geochemical similarity on nomenclature geochemical character and tectonic phase diagrams (Erdoðan and Güngör, 2004) (Figure 19). In this area granites varying from leucocratic to melanocratic character also form a single geochemical association. They are S-type and Syn-Collisional in character (Figure 19) and compared to the alkalines they are all rich in Al content indicating that the different phases were evolved from one parent magma and do not differ in phase and formation as stated by Bozkurt and Park (1994; 1997*a*; 1997*b*) some phases injected into the detachment zone.

As described in the Dibek Mt these various phases ranging from leucocratic to melanocratic granites were formed in the deep crustal environment by partial to complete assimilation of country schists inside the migmatitic fronts without any spatial relation with extensional shear zones. Clearly intrusive contacts with mica schist enclaves in the gneissic granites, on the other hand, are rare in the Menderes Massif and might

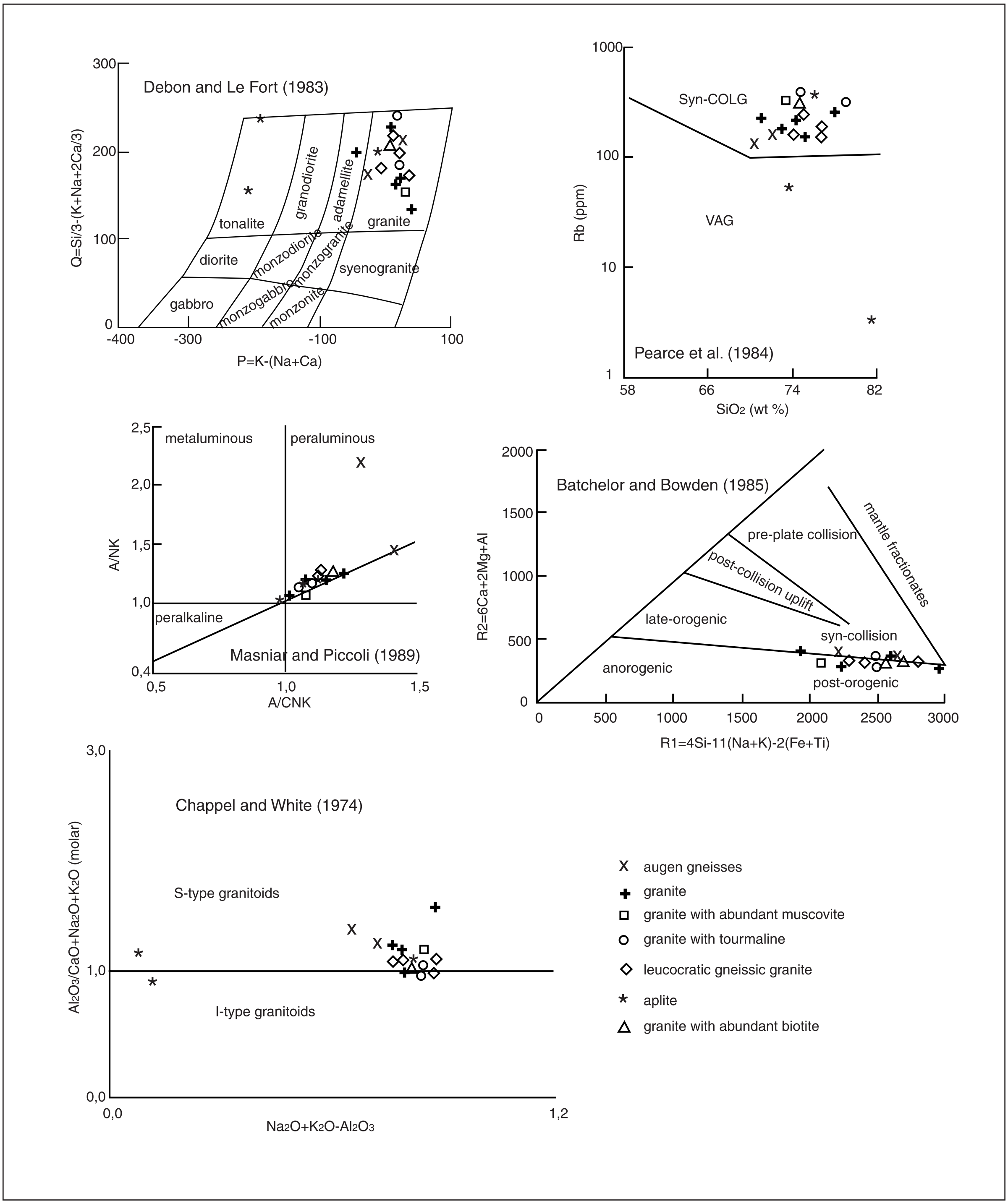

Figure 19- Geochemical nomenclature and tectonic discrimination diagrams of different-phase gneissic rocks from the Bafa Lake region. Different-phase gneisses show similar geochemical characteristics (After Erdoðan and Güngör, 2004).

correspond the shallower parts of the granite emplacemet.

Along Çine-Yataðan road and near Ýkizdere Dam, the gneissic granites injected directly into the well-known Palaeozoic and Mesozoic succession during the Main Menderes Metamorphism as concordant syntectonic granites. Along their boundaries no structural unconformities are present. The marble intercalations at the contact zone were inverted into calc-fels with garnet porphyroblasts.

The gneissic granite emplacement occured during the Alpine Orogeny. 500-560 Ma Precambrian radiogenic ages reported in some earlier publications are not in accord with the structural and geological data. In recent years metagranites, emplaced structurally concordant to the gneisses, have been mapped along the northern border of the Menderes Massif (Akay, 2009) and 30 Ma zircon ages have been determined in these metagranites (Hasözbek et al., 2010*a*). The Alpine metagranites and leucogranites from different parts of the masif were also documented in the literature (Bozkurt and Satýr, 2000; Bozkurt and Park, 2001). These metagranites geochemically and structurally resemble to the gneissic granites (Figure 19; Erdoðan and Güngör, 2004). 550 Ma upper intercept points obtained from the zircons in gneissic granites, beside, indicate that the Alpine metagranites were originated from the older granites in the crust (Hasözbek et al., 2010*a*).

New field mapping and radiogenic studies indicate that during the Alpine orogeny the crust was strongly melted and rejuvenated. Remobilized magma, almost completely, assimilated the Precambrian-Palaeozoic unconformity and directly injected into the cover succession. The Precambrian ages from the rejuvenated Alpine migmatites would inherited from the parent rocks. In the Çine-Yataðan region in the structurally lowermost parts of the granitic terrane from probable remobilized leucocratic granites we are presently trying to do new radiogenic age determinations and expecting to obtain Alpine traces of remobilization.

Evirgen (1979; 1981), Evirgen and Ashworth (1984) to the north of Ödemiþ city examined and mapped the facies characteristics of the Main Menderes Metamorphism and defined a high-temperature Barrowian-type metamorphism as represent by appearence of sillimanite. Erdoðan and Güngör (2004) tried to explain this high-temperature by southward dipping subduction zone along the northern magrin of the Menderes Platform. On the other hand metamorphic facies of the extensive tectonic zones that were defined along the northern edge of the Menderes Massif by Okay et al. (1996; 2005); Candan et al. (2005); Okay (2007), indicate northward diving of the platform edge up to the 80 km in Tavþanlý and Afyon Zones. Therefore, from the northern border any southward dipping subduction would not be valid. In recent years we mapped southery imbricated thrust packages along the northern side of the massif (Hasözbek et al., 2010*b*) which also preclude the model of Erdoðan and Güngör (2004). But it is stil a problem to explain high-temperature-type Barrowian metamorphism and migmatitic emplacement of granites during the Alpine Orogeny. Burried by tectonic slices would not be able to explain this controversy and needs a better explanation.

In this study we compared the Menderes succession with the Taurus Range in two different areas; one is metamorphosed during the Alpine Orogeny and the other is not. In the Sandýklý region there is a thick detrital sequence (Kocayayla Group) which is Early Cambrian in age. The metamorphic equivalent of this unit would be the thick metadetrital succession around Ödemiþ-Alaþehir examined by Evirgen (1979; 1981) and Evirgen and Ashworth (1984).

The Bitlis Massif around Avnik region is similar to the Menderes Massif and affected by Alpine metamorphism. There is a thick mafic-felsic volcanic succession with banded iron formation of probably Precambrian age around

Avnik and is overlain by Palaeozoic-Mesozoic platform succession along an unconformity. The Alpine metamorphism was as high as that of the Menderes Massif but still the Palaeozoic unconformity and old deformations are recognizable and can be easily discovered by mapping on 1/25000 scale. The Precambrian succession of the Bitlis Massif is completely different with thick mafic-felsic volcanic succession and including banded iron formation of economic size that are characteristics of the Precambrian successions in other parts of the world.

As a conclusion the Precambrian outcrops are either completely engulfed by gneissic granites or limited in areal extend. They would be searched in areas of below the Palaezoic schist sequence and away from the granitic fronts.

ACKNOWLEDGEMENTS

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REFERENCES

Akay, E. 2009. Geology and petrology of the Simav Magmatic Complex (NW Anatolia) and its comparison with the Oligo-Miocene granitoids in MW Anatolia; implications on Tertiary tectonic evolution of the region. International Journal of Earth Sciences 98, 1655-1675. DOI 10.1007/s00531-008-0325-0

Banerji, A.K. 1977. On the Precambrian banded iron-formations and the manganese ores of the Singhbhum region, eastern India. Economic Geology 72, 90-98.

Bankes, N.J. 1973. Precambrian iron-formations of southern Africa. Economic Geology 68, 960-1004.

Batchelor, R.A. and Bowden, P. 1985. Petrogenetic interpretation of granitoid rock series using multicationic parameters. Chemical Geology 48, 43-55.

Bozkurt, E., 1996. Metamorphism of Palaeozoic schists in the southern Menderes Massif: field, petrographic, textural and microstructural evidence. Turkish Journal of Earth Sciences 5, 105-121.

_____ and Park, R.G. 1994. Southern Menderes Massif: an incipient metamorphic core complex in western Anatolia, Turkey. Journal of the Geological Society, London 151, 213-216.

_____, _____ 1997*a*. Evolution of a mid-Tertiary extensional shear zone in the southern Menderes Massif, western Turkey. Societe Geologique de France Bulletin 168, 3-14.

_____, _____ 1997*b*. Microstructures of deformed grains in the augen gneisses of southern Menderes Massif and their tectonic significance. Geologische Rundschau 86, 103-19.

_____, _____ 1999. The structure of the Palaeozoic schists in the southern Menderes Massif, western Turkey: a new approach to the origin of the main Menderes metamorphism and its relation to the Lycian Nappes. Geodinamica Acta 12, 25-42.

_____ and Satır, M. 2000. The southern Menderes Massif (western Turkey): geochronology and exhumation history. Geological Journal 35, 285-296.

_____ and Park, R.G. 2001. Discussion on the evolution of the southern Menderes Massif in SW Turkey as revealed by zircon dating. Journal of Geological Society, London 158, 393-395.

_____ and Oberhansli, R., 2001. Menderes Massif (western Turkey): Structural, metamorphic and magmatic evolution: a synthesis. International Journal of Earth Sciences 89, 679-708.

Candan, O. 1994*a*. Petrography and metamorphism of the metagabbros at the northern part of the Menderes Massif, Demirci-Gördes submassif of the Menderes Massif. Geological Bulletin of Turkey 37, 29-40.

Candan, O. 1994*b*. Metamorphism of the gabbros in the Aydýn-Çine submassif and their correlation with those in the related submassifs of the Menderes Massif. Turkish Journal of Earth Sciences 3, 123-129.

_____ 1995. Relict granulite-facies metamorphism in the Menderes Massif. Turkish Journal of Earth Sciences 4, 35-55.

_____ 1996. Petrography and metamorphism of the gabbros around Kiraz-Birgi region, Ödemiþ-Kiraz submassif of the Menderes Massif. Yerbilimleri 18, 1-25.

_____ and Dora, O.Ö. 1998. Granulite, eclogite and blueschist relics in the Menderes Massif: An approach to Pan-African and Tertiary metamorphic evolution. Geological Society of Turkey Bulletin 41, 1-36 [in Turkish with English abstract].

_____, ______ Oberhansli, R., Çetinkaplan, M., Partzsch, J.R., Warkus, W.C. and Durr, S.H., 2001. Pan-african high-pressure metamorphism in the Precambrian basement of the Menderes Massif, western Anatolia, Turkey. International Journal of Earth Sciences 89, 793-811.

_____, Koralay, E., Dora, O.Ö., Chen, F., Oberhansli, R., Akal, C., Satýr, M. and Kaya, O., 2006. Menderes Masifi' nde Pan-Afrikan sonrasý uyumsuzluk: Jeolojik ve jeokronolojik bir yaklaþým. Türkiye Jeoloji Kurultayý, Bildiri Özleri, 25

_____, _____, _____, _____, _____, Çetinkaplan, M., Akal, C., Satýr, M. and Kaya, O. 2007. Menderes Masifi' nin Pan-Afrikan temelinin stratigrafisi ve örtü-çekirdek birimlerinin ilksel dokanak iliþkisi. Menderes masifi Kollokyumu, Bildiri özleri, 8-13.

Chappel, B.W. and White, A.J.R. 1974. Two contrasting granite types. Pacific Geology 8, 173-174.

Dean, W.T and Özgül, N. 1994. Cambrian rocks and faunas, Hudai area, Taurus Maountains, southwestern Turkey. Bulletin Royal des Sciences Naturelles de Belgique. Science de la Terre 64, 5-20.

Debon, F. and Le Fort, P. 1983. A chemical-mineralogical classification of common plutonic rocks and associations. Transactions of Royal Society, Edinburgh Earth Sciences 73, 135-149.

Dora, O., Kun, N. and Candan, O. 1988. Metavolcanics (leptites) in the Menderes Massif: a possible paleoarc volcanism. Middle East Technical University Journal of Pure and Applied Sciences 21, 413-445.

_____, _____ and _____, 1992. Metamorphic history and geotectonic evolution of the Menderes Massif. In: Savaþçýn, M.Y. and Eronat A.H. (eds), Proceedings of International Earth Sciences Congress on Aegean Regions 1990, Dokuz Eylül University Publications 2, 107-115.

_____, Candan, O., Kaya, O. and Koralay, E. 2001. Revision of the so-called leptite-gneisses in the Menderes Massif: A supracrustal metasedimentary origin. Geological Rundschau 89, 836-851.

Durr, S.H. 1975. Iber alter und geotektonische stellung des Mendereskristallins/SW- Anatolien und seine aequivalente in der mittleren Aegaeis. Habil.-Schr. Philipps - Univ. Marburg / Lahn, 107 p.

Erdoðan, B. 1982. Bitlis Masifinin Avnik (Bingöl) yöresinde Jeolojisi ve yapýsal özellikleri: Ege Üniv. Yerbilimleri Fak. Ýzmir, doçentlik tezi, 106s (unpublished).

_____, 2006. Menderes Masifi' nin çekirdek kompleksi modeli olarak evrimi ve yüzeylenmesinde kýtasal ölçekli sýyrýlma faylarýnýn rolünün tartýþýlmasý. 59. Türkiye Jeoloji Kurultyý, Bildiri Özleri, 15-16.

_____, Dora, O. and Helvacý, C., 1981. Avnik (Bingöl) yöresi apatitli demir yataklarýnýn jeolojisi ve oluþumu: Ege Üniv. Yerbilimleri Fak. Ýzmir, rapor, 122 s (unpublished).

_____ and ______ 1983. Geology and genesis of the apatite-bearing iron deposits of the Bitlis Massif. Bulletin of the Geological Society of Turkey: 26, 133-144.

Erdoðan, B. and Güngör, T. 2004. The problem of the core-cover boundary of the menderes massif and an emplacement mechanism for regionally extensive gneissic granites, western Anatolia (Turkey). Turkish Journal of Earth Sciences 13, 15-36

______, Uchman, A., Güngör, T. and Özgül, N. 2004. Lithostratigraphy of the Lower Cambrian metaclastics and their age based on trace fossils in the Sandýklý region, southwestern Turkey. Geobios, 37/3: 346-360

Evirgen, M. 1979. Menderes Masifi metamorfizmasýna petroloji, petrokimya ve jenez açýsýndan yaklaþýmlar (Ödemiþ-Tire-Bayýndýr-Turgutlu yöresi). Doktora tezi, Hacettepe Üniversitesi. 190 (260) E 93m, L19 paftasý

______, 1981. Menderes Masifinin gnayslarýnda ve þistlerinde metamorfizma koþullarý, Alaþehir-Manisa: Tartýþma ve Yanýt. Türkiye Jeoloji Kurumu Bülteni, 24, 91-94.

______ and Ashworth, J.R. 1984. Andalusitic and kyanitic facies series in the central Menderes Massif, Turkey. Neues Jahrbuch Miner. Monatshefte, H5, 219-227.

Gedik, Ý. 1977. Conodont biostratigraphy in the Middle Taurus. Geological Society of Turkey Bulletin 20, 35-48.

Gessner, K., Collins, A.S., Ring, U. and Güngör, T. 2004. Structural and thermal history of poly-orogenic basement: U?Pb gecohronology of granitoid rocks in the southern Menderes Massif, Western Turkey. Journal of the Geological Society, London 161, 93-101.

Gole, N.J. 1981. Archean banded iron-formations, Yilgran Block, western Australia. Economic Geology 76, 1954-1974.

Goodwin, A.M. 1973. Archean iron formations and tectonic basins of the Canadian Shield. Economic Geology 68, 915-937.

Güngör, T. 1998. Stratigraphy and Tectonic Evolution of the Menderes Massif in the Söke-Selçuk Region. PhD Thesis, Dokuz Eylül University, Graduate School of Natural and Applied Sciences, 147 p [unpublished].

Güngör, T. and Erdoðan, B. 2001. Tectonic significance of the Mesozoic mafic volcanic rocks in the Menderes Massif, west Turkey. Internatinal Journal of Earth Sciences 89, 874-882.

Hasozbek, A., Akay, E., Erdoðan, B., Satýr, M. and Siebel, W., 2010*a*. Early Miocene granite formation by detachment tectonics or not? A case study from the northern Menderes Massif (Western Turkey). Journal of Geodynamics. doi:10.1016/j.jog.2010.03.002

______, Satýr, M., Erdoðan, B., Akay, E. and Siebel, W. 2010*b*. Early Miocene post-collisional magmatism in NW Turkey: geochemical and geochronological constraints. International Geology Review. DOI: 10.1080/00206810903579302.

______ and Reischmann, T. 1996. Intrusion age of Pan-African augen gneisses in the southern Menderes Massif and the age of cooling after Alpine ductile extensional deformation. Geological Magazine 133, 565-572.

Kimberley, M. M. 1978. Paleoenvironmental classification of iron formations. Economic Geology 73, 215-229.

Konak, N., Akdeniz, N. and Öztürk, E.N. 1987. Geology of the South of Menderes Massif. IGCP Proj. 5: Guide Book field excursion Western Anatolia, Turkey. Mineral Research and Exploration Institute of Turkey Publication, 42-53.

Koralay, E., Satýr, M. and Dora, O.Ö., 2001. Geochemical and geochronological evidence for Early Triassic calc-alkaline magmatism in the Menderes Massif, western Turkey. International Journal Earth Sciences 89, 822-835.

______, Candan, O., Dora, O.Ö., Satýr, M., Oberhansli, R. and Chen, F. 2007. Menderes Masifi' ndeki Pan-Afrikan ve Triyas yaþlý magmatik kayaçlarýn jeolojisi ve jeokronolojisi, Batý Anadolu, Türkiye. Menderes Masifi Kollokyumu, Bildiri Özleri, 24-31.

Lips, A.L.W., Cassard, D., Sözbilir, H., Yýlmaz, H. and Wijbrans, J.B. 2001. Multistage exhumation of the Menderes Massif, Western Anatolia (Turkey). International Journal of Earth Sciences 89, 781-792.

Loos, S. and Reischmann, T., 1999. The evolution of the southern Menderes Massif in SW Turkey as revealed by zircon dating. Journal of the Geological Society, London156, 1021-1030.

Maniar, P.D. and Piccoli, P.M. 1989. Tectonic discrimination of granitoids. Geological Society of America Bulletin 101, 635-643.

Okay, A.I. 2001. Stratigraphic and metamorphic inversions in the central Menderes Massif: a new structural model. International Journal of Earth Sciences 89, 709-727.

_____, 2007. The Tavþanlý Zone-The subducted northern margin of the Taurides. Colloquium on Menderes Massif, Extended Abstracts, 34-38.

_____, Satýr, M., Maluski, H., Siyako, M., Monie, P., Metzger, R. and Akyüz, S. 1996. Paleo- and Neo-Tethyan events in northwestern Turkey: Geologic and geochronologic constraints. In: Yin A. and Harrison T.M. (Eds.) The Tectonic Evlution of Asia, Cambridge University Press, 420-441.

_____, Tansel, Ý. and Tüysüz, O. 2005. Obduction, subduction and collision as reflected in the Upper Cretaceous-Lower Tertiary sedimentary record of western Turkey. Geological Magazine 138, 117-142.

Önay, T.S. 1949. Ber die Smirgelgesteine SW-Anatoliens. Schweizerische Mineralogische und Petrographische Mitteilungen 29, 359-484.

Özer, S., Sözbilir, H., Özkar, Ý., Toker, V. and Sarý, B. 2001. Stratigraphy of Upper Cretaceous-Paleocene sequences in the southern and eastern Menderes Massif, western Turkey. International Journal of Earth Sciences 89, 852-866.

Pearce, J.A., Harris, N.B.W. and Tindle, A.G. 1984. Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. Journal of Petrology 25, 956-983.

Rimmele, G., Oberhansli, R., Goff, B., Jolivet, L., Candan, O. and Çetinkaplan, M. 2003. First evidence of high-pressure metamorphism in the Cover Series of the southern Menderes Massif. Tectonic and metamorphic implications for the evolution of the SW Turkey. Lithos 71, 19?46.

Schulling, R.D. 1962. On petrology, age and structure of the Menderes migmatite complex (SW Turkey). General Directorate of Mineral Research and Exploration Institute of Turkey (MTA) Bulletin 58, 71-84.

Þengör, A.M.C., Satýr, M. and Akkök, R. 1984. Timing of tectonic events in the Menderes Massif, western Anatolia. Implications or tectonic evolution and evidence for Pan-African basement in Turkey. Tectonics, 3, 693-707

Whitney, D.L. and Bozkurt, E. 2002. Metamorphic history of the southern Menderes Massif, western Turkey. Geological Society of America Bulletin114, 829-38.

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TAVÞANLI ZONE: THE NORTHERN SUBDUCTED MARGIN OF THE ANATOLIDE-TAURIDE BLOCK

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ABSTRACT. - The Tavþanlý Zone constitutes the northern margin of the Anatolide-Tauride Block that has undergone high pressure-low temperature metamorphism during the Cretaceous. It is bounded in the north by the Ýzmir-Ankara suture and in the south by the rocks of the Afyon Zone. The Tavþanlý Zone is subdivided into four tectonic units. At the base there is the Orhaneli Group, which shows a regular stratigraphic succession that has underdone metamorphism at ~24 kbar pressure and 430-500 °C temperature during the Late Cretaceous (~80 Ma). From the base upward the Orhaneli Group consists of micaschist, marble and metabasite-metachert-phyllite, and is tectonically overlain by ophiolitic mélange or directly by ophiolite. The ophiolitic mélange consists of basalt, chert, pelagic shale and limestone, and has undergone an incipient blueschist facies metamorphism. The ophiolite constitutes the topmost member of the tectonic stack. It consists mainly of peridotite (>%90) with minor gabbro and pyroxenite, and is cut by isolated diabase dykes. In the western part of the Tavþanlý Zone all these tectonic units are intruded by Lower to Middle Eocene granodiorites, and in the eastern part of the Tavþanlý Zone the blueschists and ophiolite are overlain by Lower Eocene marine limestones. The northern margin of the Anatolide-Tauride Block was buried in an intra-oceanic subduction zone during the Campanian and Paleocene underwent HP/LT metamorphism. The blueschists were exhumed during ongoing subduction and prior to continental collision through a thrust fault at the base and a normal fault at the top. In terms of tectonic setting and the timing of the geological events, the Tavþanlý Zone exhibits close similarities to the Semail ophiolite and the underlying blueschists.

Key words: Tavþanlý Zone, blueschist, ophiolite, northwest Turkey, subduction.

INTRODUCTION

The Tavþanlý Zone constitutes the northern margin of the Anatolide-Tauride Block, which has undergone high pressure/low temperature (HP/LT) metamorphism during the Cretaceous (Figure 1). It extends for 280 km from Mustafakemalpaþa in the west to Mihaliççýk - Yunak in the east, the extension of the Tavþanlý Zone farther southeast is difficult to follow because of the extensive Neogene cover in central Anatolia (Figure 2). Isolated exposures of blueschists within the Neogene of the central Anatolia suggest that the Tavþanlý Zone extends towards the Bolkar Mountains following the southern margin of the Haymana-Ulukýþla basin. The Tavþanlý Zone is in contact in the north with the Sakarya Zone along the Ýzmir-Ankara suture, in the west with the Bornova Flysch and in the south with the Afyon zones.

Stratigraphic, petrologic and geochronological arguments to be discussed below indicate that the Tavþanlý Zone represents a promontory of the Anatolide-Tauride Block that was subducted and exhumed during the Late Cretaceous. In this respect the Tavþanlý Zone can be compared with the continental HP/LT metamorphic belts such as Oman (e.g., Lippard et al., 1986; El-Shazly et al., 1990; Warren et al., 2005), Cyclades in the Aegean (e.g., Okrusch and Bröcker, 1990) and Alaska (Forbes et al., 1984;

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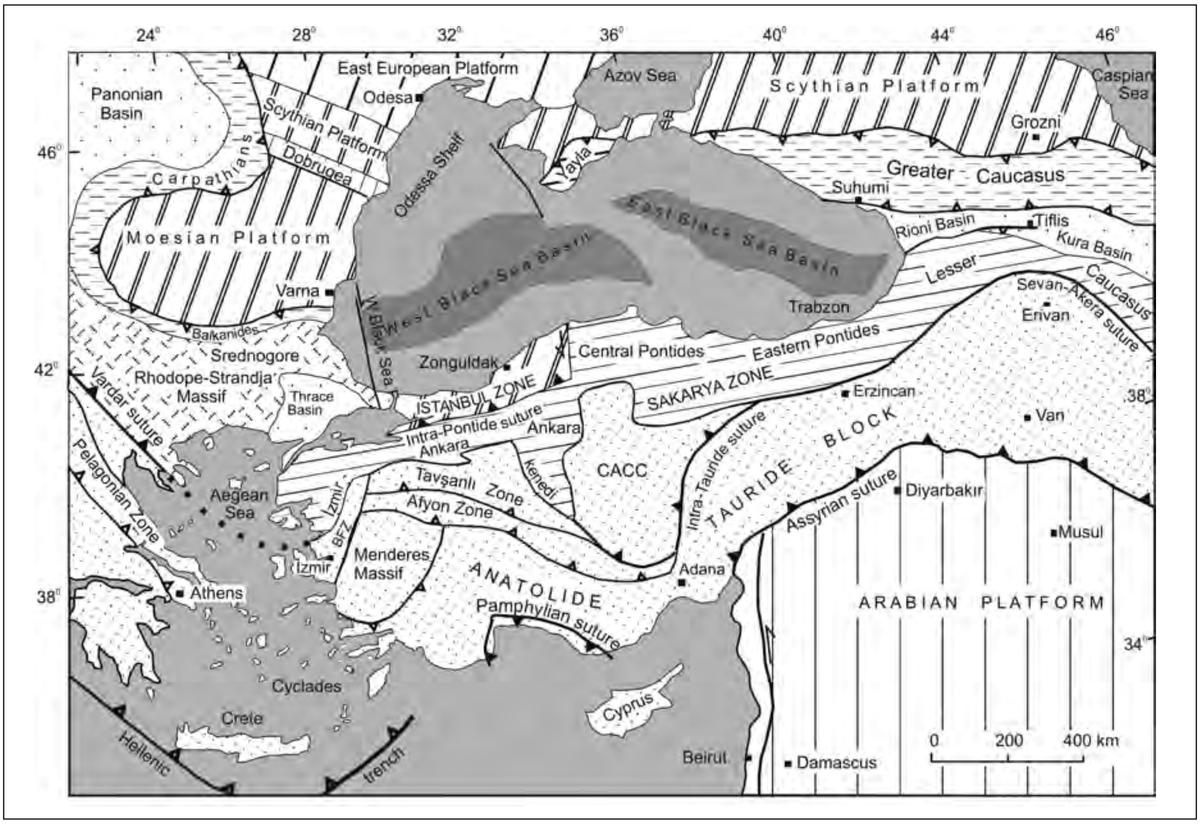

Figure 1 - Tectonic map of Turkey (modified from Okay and Tüysüz, 1999). CACC, Central Anatolian Crystalline Complex; BF, Bornova Flysch Zone.

Patrick and Evans, 1989). In terms of its size and the preservation of the HP/LT mineral assemblages, the Tavþanlý Zone is one of the largest and best preserved blueschists belts in the world (Okay, 1989).

TECTONOSTRATIGRAPHY

The Tavþanlý Zone is subdivided into four tectonostratigraphic units (Figure 3) (Okay, 1984). These are from base upwards: 1) the Orhaneli Group made up of a coherent continental upper crustal stratigraphic sequence, 2) ophiolitic mélange, 3) ophiolite, 4) Eocene sedimentary rocks and Eocene granitoids (Figures 4 and 5). These units are described below.

ORHANELÝ GROUP

The Orhaneli Group constitutes a coherent stratigraphic section made up predominantly of metasedimentary rocks (Okay, 1985). It is particularly well described from the western part of the Tavþanlý Zone and is subdivided into three formations. These are from base upwards the Kocasu formation consisting of micaschists, Ýnönü Marble and the Devlez formation made up mainly of metabasites (Figure 3). Apart from these three formations, the micaschists and marbles, which crop out at around Sivrihisar, are known as the Sivrihisar formation. The stratigraphic base of the Orhaneli Group is not observed, only in a small area around Orhaneli an Ordovician metagranitoid crops out, which

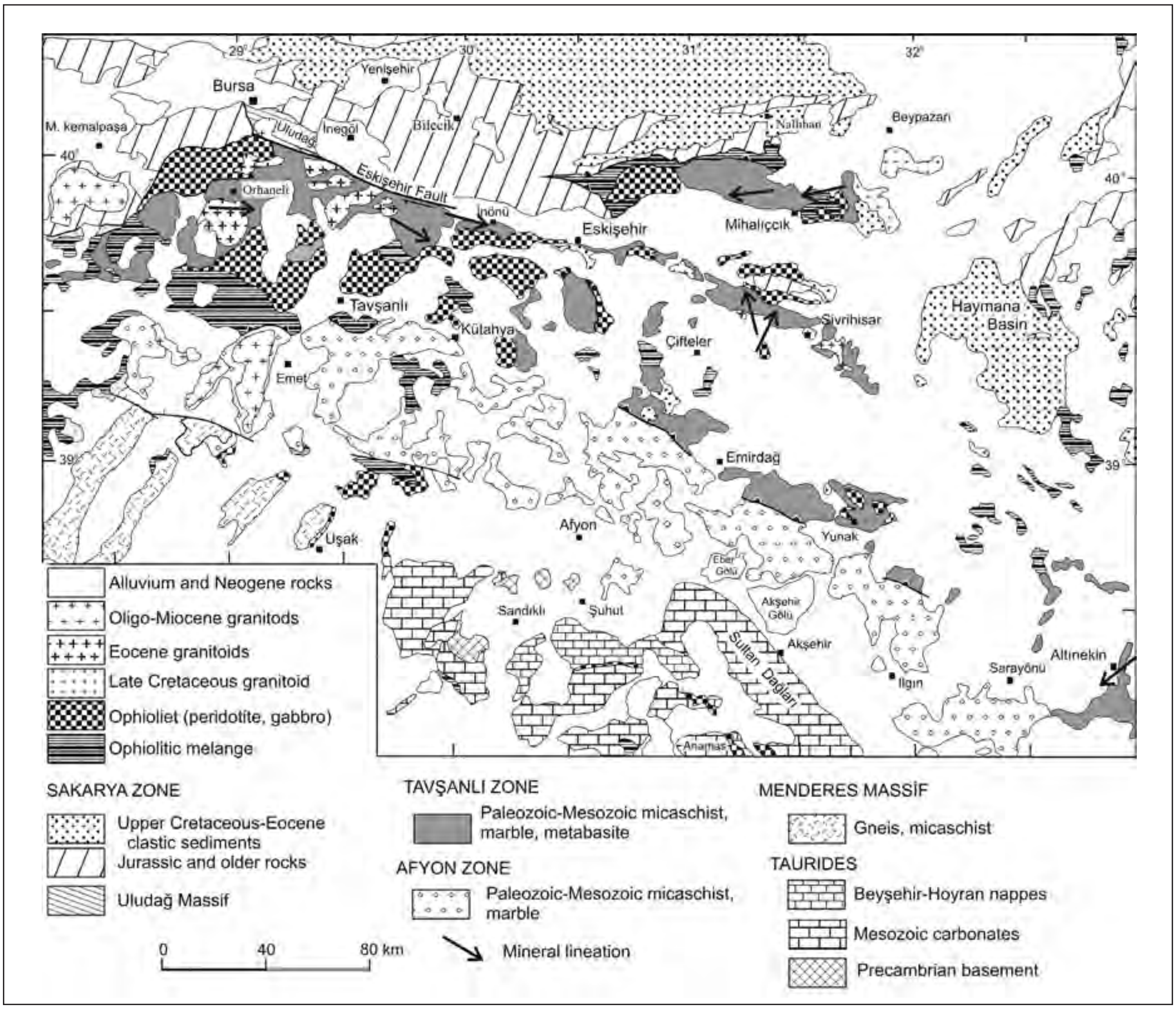

Figure 2- Geological map of the Tavþanlý Zone and the surrounding region. The map is based on Konak (2002) and Turhan (2002).

may represent a tectonic slice from the basement.

Kapanca Metagranitoid - an old granitic basement of Ordovician age

Rocks of the Orhaneli Group must have been deposited on a continental granitic-metamorphic basement; however this basement does not crop out in the Tavþanlý Zone. Only a small metagranitoid, interpreted as a tectonic slice from the basement, crops out in the core of a synform south of Orhaneli (Figure 4, Okay et al., 2008). This Kapanca metagranitoid has a thickness of about 400 meters and takes up an area of 1.5 km^2 (Figure 6). Despite complete recrystallization, the granitic texture is generally well preserved (Figure 7). The granitoid is underlain by the Triassic micaschists and marbles of the Kocasu formation. The Kapanca metagranitoid is made up essentially of quartz and jadeite with minor chloritoid, lawsonite, glaucophane and phengite. Similar HP/LT paragenesis observed in the surrounding micaschists and the subparallel

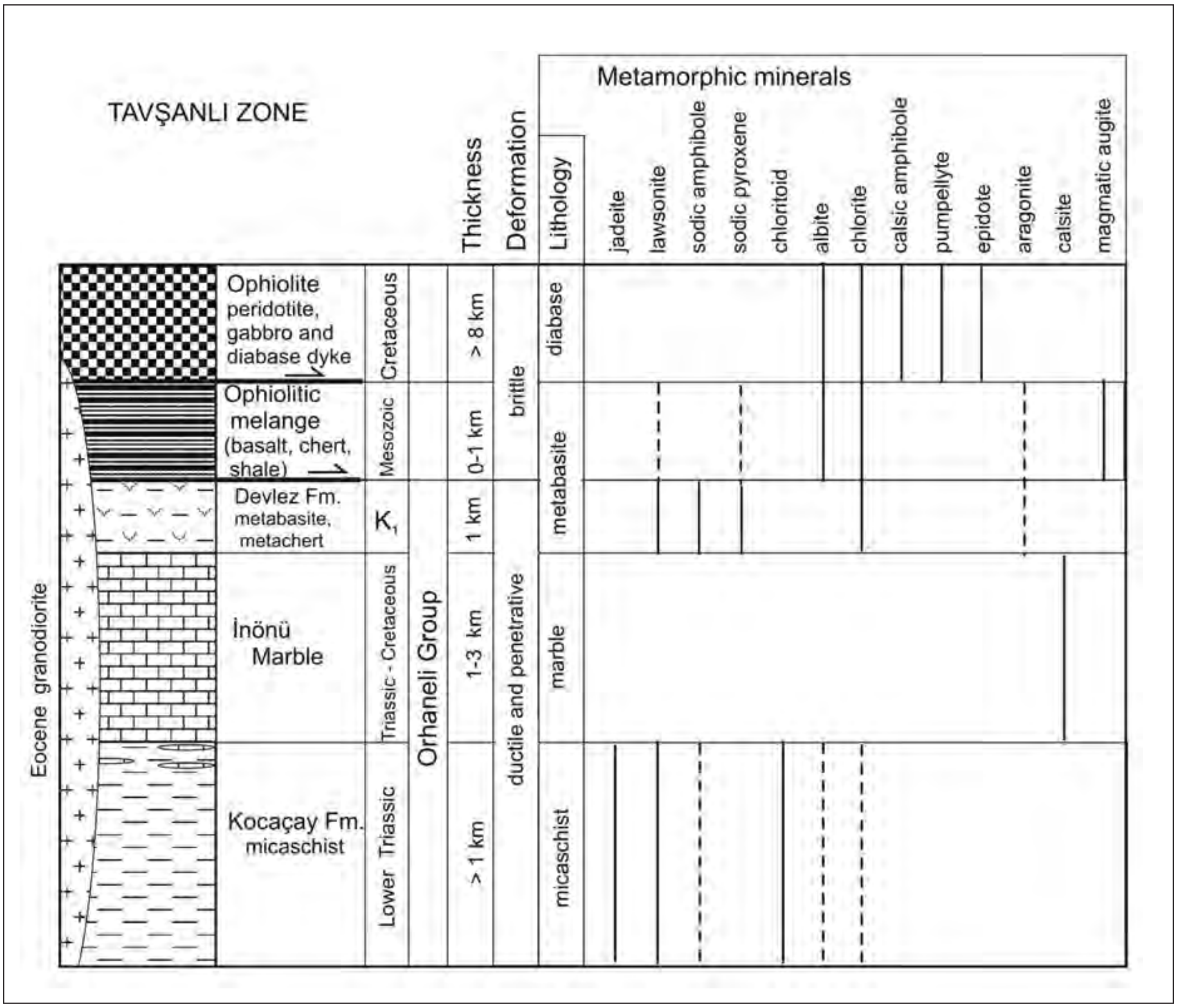

Figure 3- Tectonostratigraphy of the western part of the Tavþanlý Zone.

attitude of the foliation indicate that both units share a common metamorphic and deformational history. The micaschists surrounding the metagranitoid has yielded clastic zircons as young as Permo-Carboniferous showing that the granitoid cannot have intruded into the micaschists. It must have been tectonically emplaced in the Kocasu Formation before the HP/LT metamorphism. Single zircon U-Pb evaporation analysis on zircons from two metagranitoid samples have yielded Middle Ordovician (467.0 ± 4.5 ma) ages (Okay et al., 2008*a*). These ages are interpreted as the crystallization age of the granitic magma.

The crystalline basement of the Anatolide-Tauride Block is made up generally of Pan-African (570-520 ma) granitoids (Satýr and Friedrichsen, 1986; Kröner and Þengör, 1990; Hetzel and Reischmann, 1996; Loos and Reischmann, 1999; Gürsü and Göncüoðlu, 2006). These granitoids of Late Proterozoic-Early Cambrian ages crop out over large areas in the Menderes Massif. Granitoids of similar age cover large areas in the northern parts of Africa

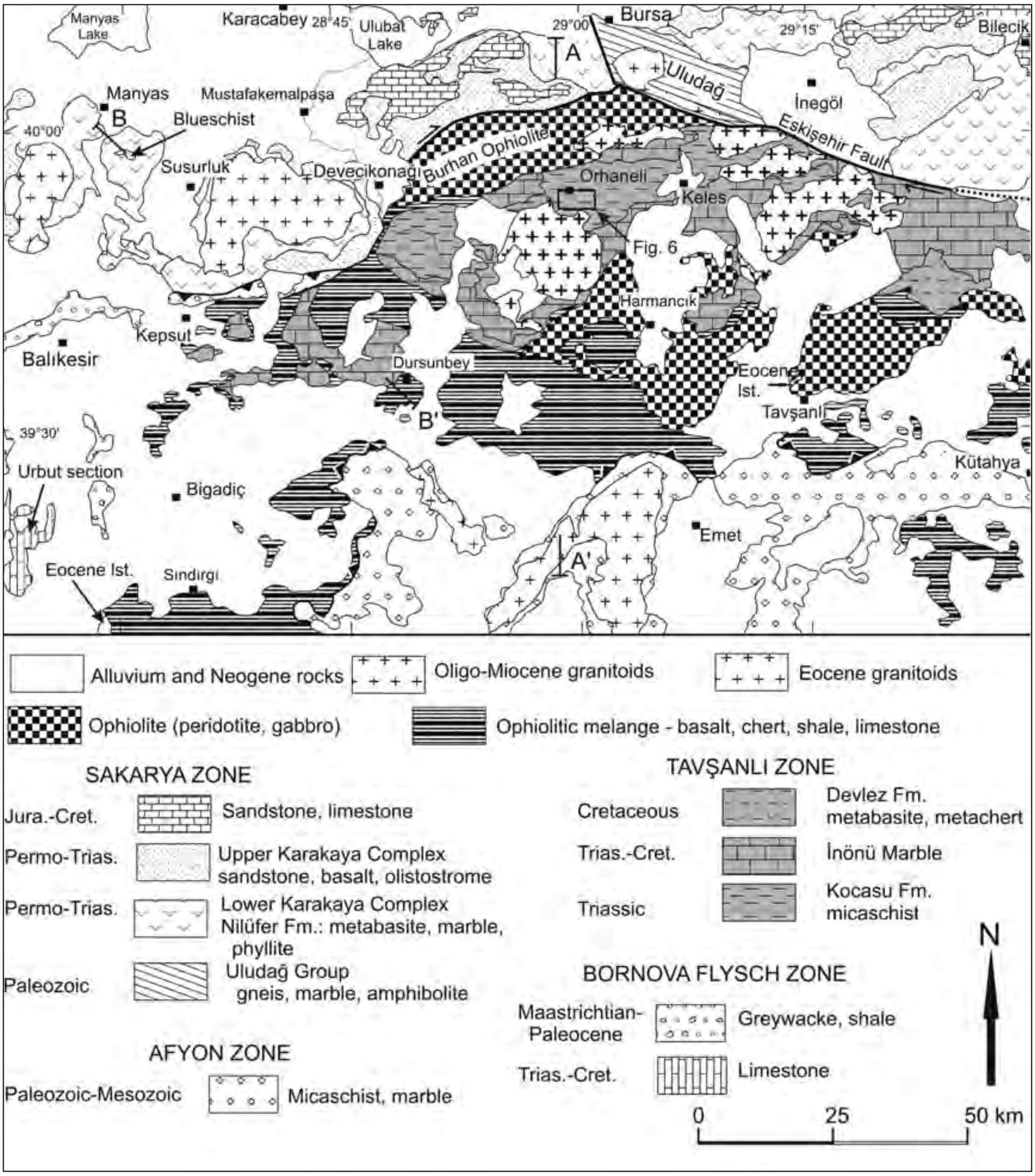

Figure 4- Geological map of the western part of the Tavþanlý Zone.

and Arabia. On the other hand, Ordovician magmatic rocks are described from Western Europe in small continental plates, which have rifted off from the northern margin of Gondwana during the Early Palaezoic (e.g., von Raumer et al., 2002). The Ordovician granitoid in the Tavþanlý Zone represents the eastward extension of this Ordovician magmatism. The Ordovician acidic magmatism is probably related to the rifting of continental terranes, such as the Ýstanbul Zone, from the northern margin of Gondwana (Okay et al., 2008*a*).

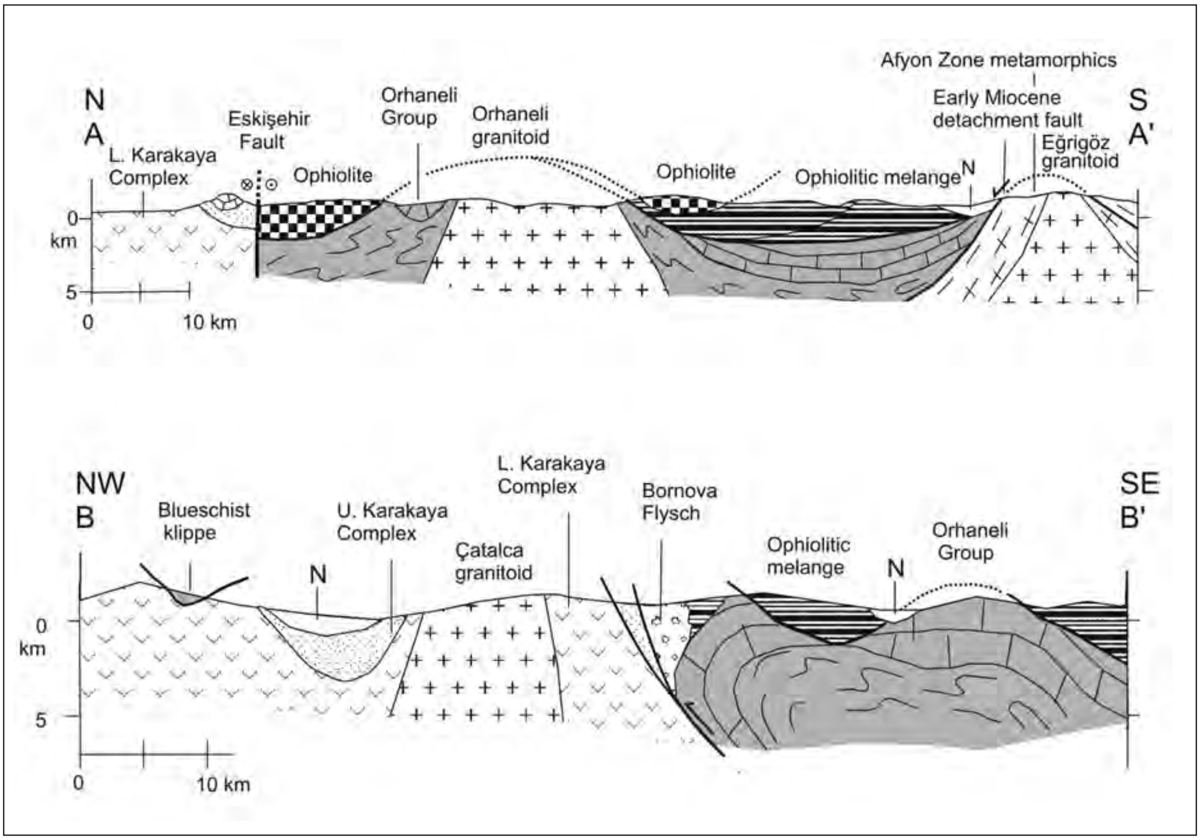

Figure 5- Geological cross-sections from the western part of the Tavþanlý Zone. For location see figure 4.

THE KOCASU FORMATION

The Kocasu Formation is a coherent sequence of quartz-micaschists, with a minimum thickness of 800 meters, at the base of the Orhaneli Group (Okay, 2004). Micaschists form medium-grained, hard, finely banded, grey to light grey rocks. Quartz-rich micaschist bands with a gneissic texture, 0.1 to 2 m in thickness, alternate with finer grained and more mica-rich micaschists (Figure 8, Okay and Kelley, 1994; Okay, 2002). Metaconglomerates with quartz clasts occur rarely within the micaschists. The percentage of mica increases upwards in the metamorphic sequence. South of the town of Devecikonaðý there are metaaplitic sills and dykes, 0.5 to 3 m in thickness, within the micaschists (Figure 9). These acidic vein rocks, which were emplaced prior to the HP/LT metamorphism consists of jadeite, quartz and secondary albite (Okay and Kelley, 1994).

The lithological characteristics of the Kocasu formation indicate that before the metamorphism the sequence consisted of an alternation of sandstone and shale, however, the rocks are now completely recrystallized and have lost their primary lithological features. The Kocasu formation passes upwards gradually to the Ýnönü Marble. The transition zone consists of micaschists intercalated with marble layers.

The Kocasu formation crops out over a large area between Devecikonaðý in the west and south of Uludað in the east (Figure 4). It also crops out northwest of Mihaliççýk under the Ýnönü

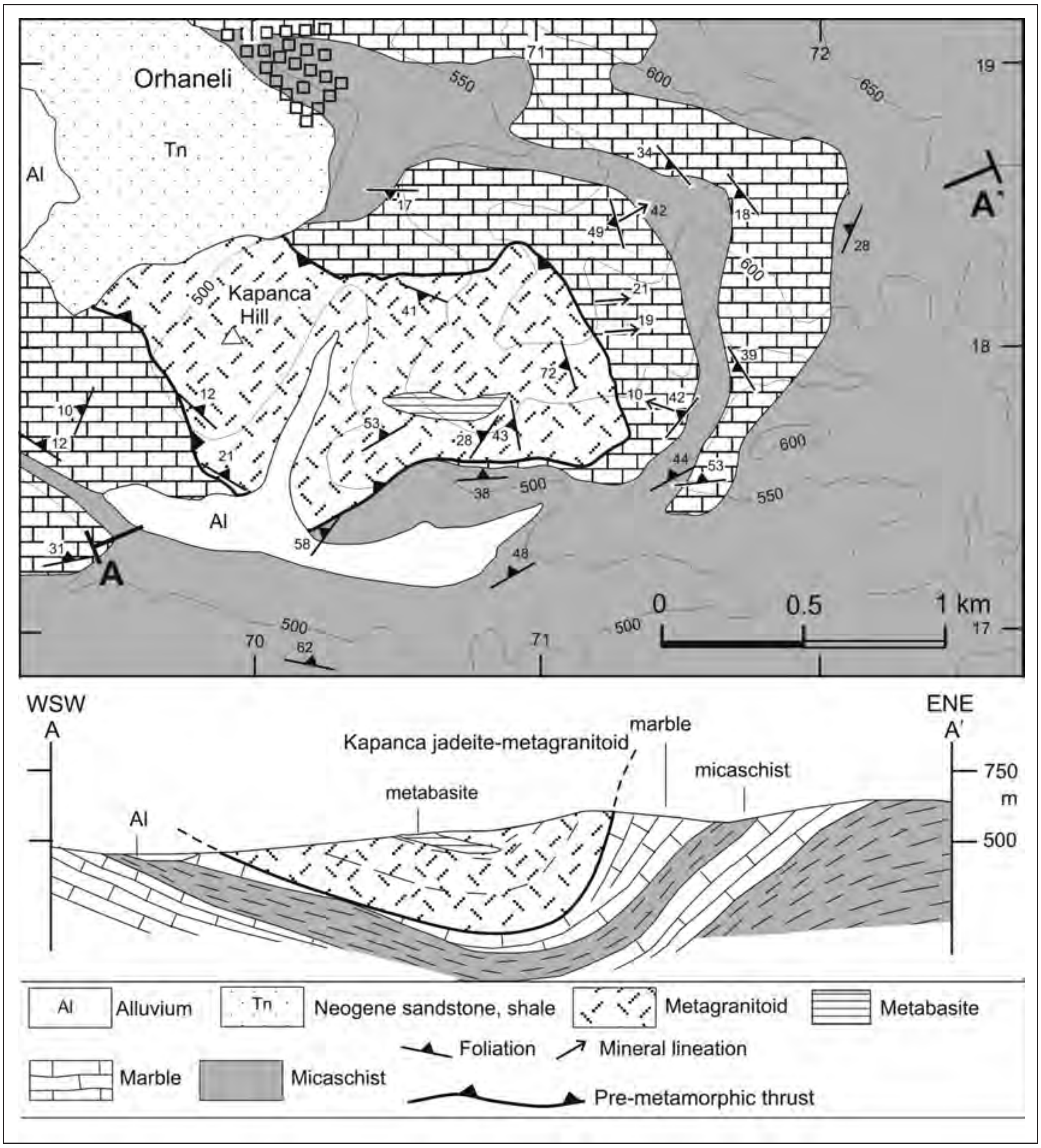

Figure 6- Geological map and cross-section of the Orhaneli region showing the setting of the Kapanca metagranitoid (Okay et al., 2008). For location see figure 4.

Marble. In this latter region, the metaclastic sequence has been called as the Göktepe metamorphics by Göncüoðlu et al. (2000). The stratigraphic base of the Kocasu formation is not exposed; however, as discussed above it is probably underlain by a Lower Palaezoic granitic basement. In order to constrain the depositional age of the Kocasu formation clastic zircons from the micaschists have been dated in the regions of Orhaneli and Keles. The clastic zircons give Ordovician and Permo-Carboniferous age peaks (Okay et al., 2008*a*), which indicate that the

Figure 7- Kapanca metagranitoid. The dark regions are quartz and the white ones jadeite.

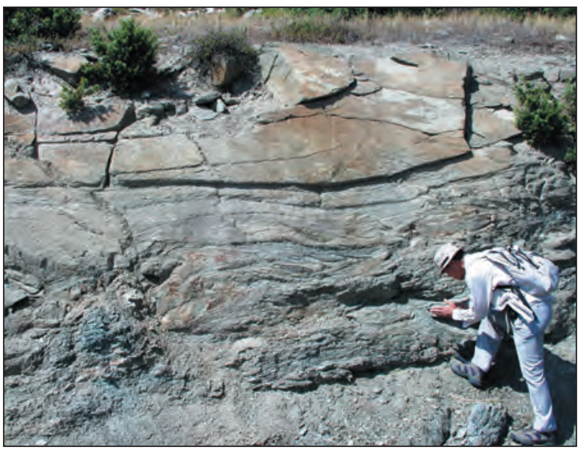

Figure 8- Isoclinally folded micaschists of the Kocasu Formation. The thick horizon in the upper part of the photograph is quartz-micaschist of psammatic origin, the lower horizon is finer grained metapelitic micaschist. The jadeite + chloritoid + quartz paragenesis in these rocks show that they were metamorphosed at 80 km depth. East of Orhaneli, the road between Orhaneli and Kabaklar.

Kocasu formation is younger than Carboniferous. A comparison with the general stratigraphy of the Taurides (e.g. Gutnic et al., 1979; Özgül, 1976) suggests a Permo-Triassic, and probably Early to Mid-Triassic age for the Kocasu formation.

The micaschists of the Kocasu formation consists essentially of quartz and phengite. South of

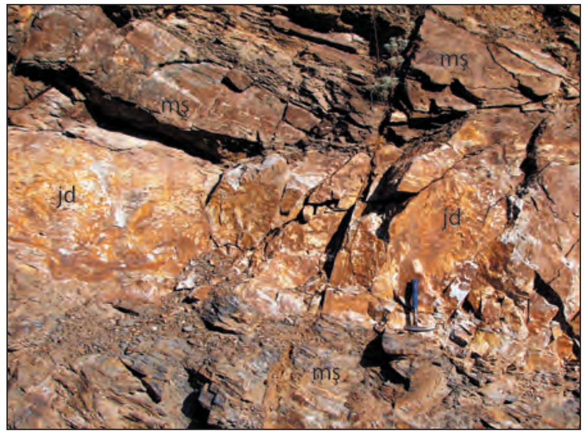

Figure 9- Metaacidic vein rock (jd) within the micaschists (mþ) of the Kocasu Formation. The rock is composed of jadeite and quartz. The shaft of the hammer marks the contact between the two rock types. South of Devecikonaðý along the Kocasu valley.

Bursa these minerals are accompanied by chlorite, jadeite, chloritoid, lawsonite and albite. The critical HP/LT mineral paragenesis in the micaschists is quartz + phengite + jadeite + chloritoid + glaucophane + lawsonite. Jadeite, locally pseudomorphed by sericite and albite, is common in the Orhaneli region, which possibly is the most jadeite-rich area in the world.

ÝNÖNÜ MARBLE

The marble series with a structural thickness of several kilometers, which overlie the Kocasu formation, is called as the Ýnönü Marble (Servais, 1982). The Ýnönü Marble crops out over large areas in the western part of the Tavþanlý Zone in the regions of Dodurga, Tahtaköprü and Ýnönü (Figure 4, Konak, 2002). Ýnönü Marble is made up of white, light grey, massif locally banded marble with occasional chert bands. A characteristic microstructural feature of the Ýnönü Marble is a strong mineral lineation defined by the parallel alignment of elongate calcite grains. The Ýnönü Marble consists of calcite, although it must have been made up of aragonite during the metamorphism.

The Ýnönü Marble represents the metamorphosed equivalent of the Tauride Mesozoic carbonate platform. This is supported by the description of Late Triassic (Late Norian) conodonts from the lower parts of the Ýnönü Marble from east of Orhaneli (Kaya et al., 2001). The upper parts of the Ýnönü Marble probably extends into the Cretaceous.

DEVLEZ FORMATION

The sequence of metabasite, metachert and phyllite, which lie over the Ýnönü Marble, is called as the Devlez formation (Okay, 1981; 2004). Metabasites constitute the bulk (more than 80%) of the Devlez formation. They are represented by submarine lavas, pyroclastic rocks and tuffs, which are, however, completely recrystallized with the development of a penetrative metamorphic fabric and new minerals. The typical mine-ral paragenesis in the metabasites is sodic amphibole + lawsonite + chlorite + sodic pyroxene + phengite. Relict magmatic augite is occasionally found in the metabasites. In the metacherts the HP mineral paragenesis is quartz + sodic amphibole + lawsonite + spessartine-rich garnet + phengite + hematite (Okay, 1980*a*). The structural thickness of the Devlez Formation in the region northeast of Tavþanlý is one kilometers. Metabasites and metacherts of the Devlez formation show a strong foliation and a strong mineral lineation defined through the parallel alignment of the sodic amphibole grains. The Devlez formation shares a common metamorphic and deformational history with the underlying Ýnönü Marble, however, it is possible that it represents an exotic tectonic slice emplaced on the Mesozoic limestones of the Ýnönü Marble prior to regional metamorphism.

SÝVRÝHÝSAR AND HALÝLBAÐI FORMATIONS

In the eastern part of the Tavþanlý Zone, especially in the region of Sivrihisar, The Orhaneli Group is represented by an intercalation of marble and micaschist (Figure 10). These metasedimentary rocks, called as the Sivrihisar formation, have a structural thickness of over three kilometers (Kulaksýz, 1981; Gautier, 1984; Monod et al., 1991). Marbles form bands within the micaschists, whose thickness ranges from a few meters to several hundred meters. There are approximately equal amounts of marble and micaschist in the Sivrihisar formation. Apart from these two dominant rock types, there are rare metabasite layers within the micaschists. The Sivrihisar formation probably represents a lateral facies variation of the Kocasu and Ýnönü formations.

Unlike the western part of the Tavþanlý Zone, HP/LT mineral paragenesis is not well preserved in the Sivrihisar formation. The common mineral assemblage in the micaschists is quartz + albite + chlorite + phengite, and in the metabasites albite + chlorite + actinolite + epidote (Gautier, 1984). However, the occasional presence of lawsonite in the metabasites and in the calcium-rich micaschists and relict sodic amphibole in the metabasites indicates that the Sivrihisar formation has undergone a regional HP/LT metamorphism (Monod et al., 1991), which was subsequently overprinted by a greenschist facies metamorphism.

South of the village of Halilbaðý, the Sivrihisar formation is overlain by metabasite, metachert, marble and metaserpentinite, called the Halilbaðý formation (Figure 10). The HP/LT mineral paragenesis including lawsonite-eclogites are well preserved in the Halilbaðý region, which has been the subject of several detailed petrological studies (Kulaksýz, 1978; Monod et al., 1991; Davis and Whitney, 2006, 2008; Whitney and Davis, 2006; Çetinkaplan et al., 2008). In the Halilbaðý region, the metabasic rocks consist of an interesting intercalation of blueschist, garnet-blueschist and lawsonite-eclogite. The Halilbaðý Formation passes downwards into the Sivrihisar Formation. The nature of the contact between the two formations is difficult to map, as petrologically uninformative marbles crop out at the contact zone. The Halilbaðý formation can be correlated with the Devlez formation from the western part of the Tavþanlý Zone.

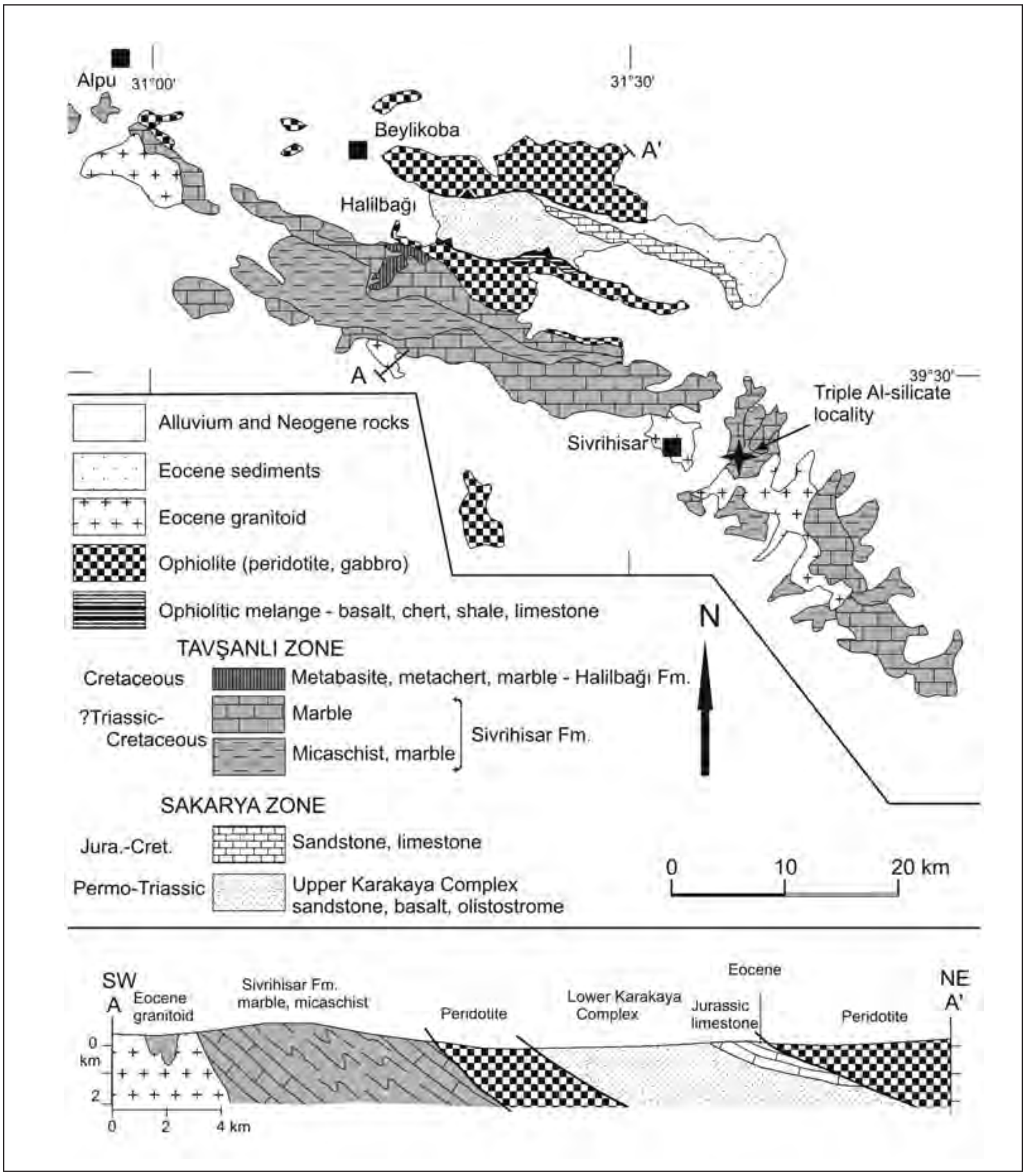

Figure 10- Geological map and cross-section of the Sivrihisar region modified from Monod et al., 1991).

Tectonostratigraphy of the southeastern part of the Tavþanlý Zone

Because of the extensive Neogene cover, rocks metamorphosed in blueschist facies form small isolated outcrops south of the Haymana basin. Within this large area blueschists of the Altýnekin region northeast of Konya and those from Yunak have been studied in detail. A metamorphic sequence consisting of marble and phyllite with subordinate metabasite, similar to the Sivrihisar Formation, crops out in the Yunak region (Yeniyol, 1979). Sodic amphibole + lawsonite paragenesis is widely described from the metabasites. This sequence is tectonically overlain by an ophiolitic mélange.

A metamorphic sequence of marble, calcschist, phyllite and metaquartzite, probably Permian to Mesozoic in age, crops out in the Altýnekin region northeast of Konya (Karaman, 1986; Özgül and Göncüoðlu, 1999; Eren, 2000; Droop et al., 2005). This sequence metamorphosed in blueschist facies is tectonically overlain by an ophiolitic mélange and ophiolite. The ophiolitic mélange in the Altýnekin region shows a distinct blueschist facies metamorphism and associated deformation. The grade of metamorphism and intensity of deformation in the Altýnekin region increases upwards in the sequence and from south to north. The typical paragenesis in the metabasites of the ophiolitic mélange is sodic amphibole + epidote ± lawsonite + albite + chlorite + phengite.

Sodic amphibole-bearing metamorphic rocks in the Bolkardað region south of the Ulukýþla basin (Blumenthal, 1956; Çalapkulu, 1980) may represent the eastward extension of the Tavþanlý Zone. Information on the tectonic setting and mineral paragenesis of these rocks are limited.

OVACIK COMPLEX - OPHIOLITIC MELANGE

Slices of ophiolitic mélange rest with tectonic contacts on the Orhaneli Group. The ophiolitic mélange, called as the Ovacýk Complex (Kaya, 1972*a, b*), crops out over large areas and consists mainly of basalt, radiolarian chert, pelagic shale and limestone with lesser amounts of serpentinite, talc, greywacke and bedded manganese deposits. The basaltic rocks, which make up at least 60% of the Ovacýk Complex, consist generally of pyroclastic rocks and agglomerates; pillow lavas are rare. All the basaltic rocks are spilitized and locally show an incipient high pressure metamorphism. Radiolarian cherts are red, thinly to medium bedded and are intercalated with thin black shale layers; rarely do they show stratigraphic contacts with the basalts. The carbonate rocks are found as clasts and olistoliths in the basalts, 2 cm to several hundred meters across, and as thinly bedded pelagic limestones intercalated with the basalts. The metaphonolites, which today occur as clasts and blocks in the Neogene deposits south of Bursa, were probably once part of the Ovacýk Complex (Okay, 1997). They have an interesting mineralogy of jadeite, quartz and K-feldspar and are marketed as a semi-precious stone under the name of purple jade.

The Ovacýk Complex and equivalent ophiolitic mélanges crop out widely outside the Tavþanlý Zone within the Afyon Zone, in the Menderes Massif and in the Taurides (Figure 2). As discussed below the Ovacýk Complex and the Anatolian Ophiolite probably formed once a continuous tectonic cover over the Anatolide-Tauride Block.

Although the Ovacýk Complex is generally known as a mélange, it lacks an easily definable matrix and instead consists of strongly deformed rock types juxtaposed along tectonic contacts. Stratigraphic relation between the different rock types are generally not preserved and it is difficult to follow bedding for more than ten meters without encountering a fault or shear zone. With its lithological features, internal structure and with its incipient HP metamorphism, the Ovacýk Complex represents a Tethyan subduction-accretion complex.

Radiolaria from the cherts within the Ovacýk Complex from the Tavþanlý and Bornova Flysch zones have given Late Triassic (late Carnian, late Norian), Jurassic, Early Cretaceous (Berriasian-Hauterivian) and Late Cretaceous (Cenomanian, Turonian) ages (Bragin and Tekin, 1996; Tekin et al., 2002; Göncüoðlu et al., 2006). These radiolarian ages indicate that the Neo-Tethyan oceanic crust north of the Anatolide-Tauride Block has a minimum age range from Late Triassic to Late Cretaceous. A 179 ± 15 Ma zircon U-Pb age from a plagiogranite from an ophiolitic mélange near Ankara also indicates the presence of an Early Jurassic oceanic crust in the northern Neo-Tethys (Dilek and Thy, 2006). The geochemistry of the basaltic rocks in the Ovacýk Complex shows the presence of several magma types with a dominance of oceanic island and mid-ocean ridge type basalts (Tankut et al., 1998; Rojay et al., 2004; Göncüoðlu et al., 2006; Gökten and Floyd, 2007).

ANATOLIAN OPHIOLITE

Large ophiolite massifs lie tectonically over the Ovacýk Complex or directly on the blueschists of the Orhaneli Group. Over 90% of the ophiolitic masses in the Tavþanlý Zone are made up of peridotites; the rest is represented by pyroxenite, gabbro, chromite and diabase dykes. Sub-ophiolitic metamorphic rocks are described from the base of the ophiolites.

The peridotites mainly consist of harzburgite and dunite; lenticular chromite deposits occur within the dunites. The best studied ophiolite is the Burhan ophiolite north of Orhaneli; it consists predominantly (> 90%) of harzburgite and dunite (Figure 4), the rest is made up of gabbro, pyroxenite, chromite and diabase dykes (Lisenbee, 1971; 1972; Tankut, 1980). The harzburgites and dunites form bands, ~2 km thick, intercalated with thinner bands of gabbro and pyroxenite. The bands are transitional over a width of 50 m. The peridotites show a tectonic foliation subparallel to the lithological layering. The thickness of the Burhan ophiolite, measured perpendicular to lithological layering, is over 13 kilometers. The lithological and structural features of the Burhan ophiolite show it to be a deformed cumulate sequence.

The mineral paragenesis in the peridotites of the Tavþanlý Zone is olivine + orthopyroxene + clinopyroxene + chrome-spinel (Lisenbee, 1971; Okay, 1985, Lünel, 1986; Asutay et al., 1989; Önen, 2003). This mineral assemblage is stable at pressures of less than 14 kbar; garnets start to form at higher pressures (Perkins et al., 1981). The paragenesis in the gabbros is plagioclase (An_{89-100}) + clinopyroxene + orthopyroxene + spinel (Önen, 2003). Based on the petrography and geochemistry of the ophiolitic rocks around Kütahya, Önen (2003) argues that they are similar to those formed in the oceans and in the back-arc basins.

Between the Ýzmir-Ankara suture and the Mediterranean there are several ophiolitic massifs. The ophiolites in this large region show many common lithological and structural features: a) they are generally made up of harzburgite and dunite, b) the peridotites are cut by diabase dykes, c) all of the isotopically dated sub-ophiolite metamorphic rocks in Anatolia are Albian in age, d) the sub-ophiolite metamorphic rocks show a low-grade HP/LT metamorphism that overprints the earlier HT metamorphism (Dilek and Whitney, 1997; Okay et al., 1998; Önen and Hall, 2000). These common features suggest that these ophiolitic massifs formed part of a very large ophiolite obducted over the Anatolide-Tauride Block during the Late Cretaceous (Dilek et al., 1999; Önen 2003). In terms of its size and its association with the HP/LT metamorphic rocks this Anatolian ophiolite nappe can be compared with the Semail ophiolite in Oman.

SUB-OPHIOLITE METAMORPHIC ROCKS

Most ophiolites have a metamorphic sole produced during the intra-oceanic thrusting stage (e.g., Williams and Smyth, 1973; Woodcock and Robertson, 1977). The sub-ophiolite metamorphism forms during the intra-oceanic thrusting through frictional heating and through the downward convection of the heat from the overlying ophiolite body. As the heat is conducted downward, the sub-ophiolite metamorphic rocks show an inverted metamorphic zonation with the metamorphic grade increasing upwards in the sequence. As the sub-ophiolite metamorphism occurs in the oceanic lithosphere, the metamorphic rocks are naturally of oceanic crustal origin.

Sub-ophiolite metamorphic rocks are described from several localities in the Tavþanlý

Zone (Gautier, 1984; Monod et al., 1991; Önen and Hall, 1993, 2000; Okay et al., 1998). The garnet-amphibolites at the base of the Burhan ophiolite have peak P-T values of 8.5 ± 3.5 kbar and ~700 °C (Okay et al., 1998). The pressure values indicate an initial ophiolite thickness of 25 ± 10 km. The amphibolites at the base of the Burhan ophiolite also exhibit a second phase of low-grade HP/LT metamorphism marked by the development of sodic amphibole and very fine-grained aggregates of lawsonite. A similar case has been described from the sub-ophiolite metamorphic rocks in the Central Taurides (Dilek and Whitney, 1997). Hornblende Ar-Ar cooling ages of the sub-ophiolite metamorphic rocks at the base of the Burhan ophiolite are 101.1 ± 3.8 Ma (Harris et al., 1994), and those from the Kütahya region 93 ± 2 Ma (Önen, 2003). These ages are similar to the 95-90 Ma Ar-Ar hornblende and mica ages from the sub-ophiolite metamorphic rocks of the Lycian, Beyþehir, Aladað, Kýzýltepe and Mersin ophiolites in the Taurides (Dilek et al., 1999; Parlak and Delaloye, 1999; Çelik et al., 2006).

DIABASE DYKES

The ophiolites in the Tavþanlý Zone are cut by generally east-west trending diabase dykes. The diabase dykes, which have an average thickness of 1-2 meters, cannot generally be followed along strike for more than 100 meters. The dyke frequency is quite variable and ranges from a single dyke within a stretch of several hundred meters of peridotite to 10 dykes within 30 m of peridotite. The dyke-peridotite contacts are generally faulted due to later brittle deformation, however, in some localities the original intrusive contact are preserved and the dykes can be seen to have narrow chilled margins. This observation indicates that the dykes intruded into cold peridotite (Okay, 1981). The mineral assemblage in the diabase dykes is augite, partly replaced by magmatic hornblende, and altered plagioclase. Plagioclase has commonly altered into very fine grained aggregates of pumpellyite and albite. The mineral assemblage in the diabase dykes indicates that they and by implication the ophiolitic rocks have not undergone the HP/LT metamorphism observed in the underlying Orhaneli Group.

Diabase dykes very similar to those described above occur in the peridotites of the Lycian nappes (Whitechurch et al., 1984) and of the Central Taurides (Lytwyn and Casey, 1995; Parlak, 2000). Whole rock Ar-Ar ages of the diabase dykes cutting the Mersin ophiolite and the sub-ophiolitic metamorphic rocks of the Mersin ophiolite range between 90 Ma and 64 Ma (Parlak and Delaloye, 1996). Considering the large errors associated with the whole rock Ar-Ar dating, and that the emplacement of the ophiolite over the Anatolide-Tauride Block occurred during the Campanian, the crystallization ages of the diabase dykes are expected to be at around 90 Ma.

LOWER EOCENE SEDIMENTARY ROCKS AND EOCENE PLUTONS

In the eastern part of the Tavþanlý Zone ophiolitic rocks, ophiolitic mélange and the blueschists of the Orhaneli Group are unconformably overlain by the Lower Eocene marine sedimentary rocks. The westernmost exposure of the Eocene sediments is found north of Tavþanlý, where the peridotites are unconformably overlain by a 60-m-thick shallow marine sequence of Lower Eocene (Cuisian) sandy and pebbly limestones (Figure 4, Baþ, 1986). Farther east, south Eskiþehir a Lower Eocene (Cuisian) marine sequence of conglomerate, sandstone, shale and shaley limestone with abundant nummulites, ~300-m-thick, lies unconformably over the ophiolitic gabbros (Figure 2, Gözler et al., 1985). South and west of Çifteler a 300-m-thick sequence of limestone, shaley limestone and marn of Early Eocene age (early Ilerdian-middle Cuisian) lies uncomformably over the metamorphic and ophiolitic rocks (Göncüoðlu et al., 1992; Özgen-Erdem et al., 2007). The stratigraphic data indicate that the Tavþanlý Zone was covered by a shallow sea at the beginning of the Early Eocene (50 Ma).

Apart from the marine Eocene sedimentation, a series of plutons have intruded into the Tavþanlý Zone during the Eocene. These intrusions form a WNW-ESE trending linear belt between Sivrihisar and Bursa (Figure 2, Okay and Satýr, 2006). The plutons have generally a granodioritic composition and comprise hornblende and biotite; their ages range from 53 Ma and 45 Ma (Early-Middle Eocene) (Table 1). The 48-Ma-old Topuk pluton farthest west cuts the blueschists of the Orhaneli Group as well as the overlying peridotites indicating that the tectonic contact between these units is pre-Eocene in age (Okay et al., 1998).

The Eocene plutons in the Tavþanlý Zone have a metaaluminous composition and the amount of SiO_2 varies between 63 and 69 %. Enrichment in the LIL elements in the granodiorites indicates a calc-alkaline magma. Relatively low Y and HREE ratios suggest that garnet was not an important restite phase in the region of magma generation and hence the magma was generated at depths of less than 30 kilometers (10 kbar). The Eocene granodiorites have formed by fractionation of mantle-derived magmas in shallow magma chambers and through crustal melting induced by mantle derived basic magmas (Harris et al., 1994; Altunkaynak, 2007; Karacýk et al., 2008).

METAMORPHISM AND METAMORPHIC AGES IN THE TAVÞANLI ZONE

The three tectonostratigraphic units of the Tavþanlý Zone show different metamorphic characteristics. The Orhaneli Group has undergone a regional blueschist facies metamorphism (Çoðulu, 1965; 1967; van der Kaaden, 1966; Lünel, 1967; Okay, 1980*a, b*, 1981, 1984, 2002; Servais, 1981; Kaya, 1981; Gautier, 1984; Monod et al., 1991; Davis and Whitney, 2006; Çetinkaplan et al., 2008). The minerals of the blueschist metamorphism are particularly well preserved in the western part of the Tavþanlý Zone. The characteristic HP/LT mineral para-

Table 1- Isotopic ages of the Eocene granitoids in northwest Anatolia

Sivrihisar	53.0 ± 3.0[1]		
Gürgenyayla	45.0[2]		
Tepeldağ	44.7 ± 0.4[3]	45.0 ± 0.2[4]	
Topuk	47.8 ± 0.4[5]		
Orhaneli	52.6 ± 0.4[6]	52.4 ± 1.4[7]	49.8 ± 1.3[8]
Karabiga	52.7 ± 1.9[9]	45.3 ± 0.9[10]	
Kuzey Kapıdağ	39.9 ± 0.8[10]	38.3 ± 0.8[11]	35.5 ± 0.3[11]
Güney Kapıdağ	36.1 ± 0.8[10]	38.2 ± 0.8[10]	35.3 ± 0.3[11]
Avşa	40.9 ± 1.1[12]	44.4 ± 0.4[12]	
Fıstıklı	48.2 ± 1.0[10]	35.4 ± 0.8[10]	

1) Ar-Ar hornblende, Sherlock et al. (1999); 2) Rb-Sr biotite, Ataman (1973 *a,b*);
3) Rb-Sr biotite, Okay and Satır, 2006; 4) U-Pb zircon, Okay and Satır, 2006;
5) Ar-Ar hornblende, Harris et al. (1994); 6) Ar-Ar biotite, Harris et al. (1994);
7) Ar-Ar biotite - in hornfels, Harris et al. (1994);
8) Rb-Sr biotite, Ataman (1972); 9) U-Pb xenotime, Beccaletto et al. (2007);
10) K-Ar hornblende, biotite, Delaloye and Bingöl (2000);
11) Rb-Sr biotite, M. Satır (unpublihed data);
12) K-Ar biotite, Karacık et al. (2008).

genesis in the metabasites is sodic amphibole + lawsonite + chlorite ± sodic pyroxene + phengite + sphene (Çoðulu, 1967; Okay, 1980*a*). Apart from these minerals some metabasites contain garnet. In the metabasites the sodic amphibole is generally of glaucophane and crossite in composition and the sodic pyroxene is chloromelanite. The common mineral paragenesis is the metacherts is quartz + garnet + sodic amphibole + lawsonite + phengite + hematite. In the metacherts the garnet is rich in spessartine and sodic amphibole is magnesio-riebeckite and crossite in composition.

The characteristic HP/LT mineral paragenesis in the micaschists of the Kocasu Formation and of the Kapanca metagranitoid is: jadeite + chloritoid + lawsonite + glaucophane + quartz + phengite (Okay and Kelley, 1994; Okay, 2002; Okay and Satýr, 2006; Okay et al., 2008). This mineral assemblage in the metapelitic rocks shows that the peak pressure and temperature values of the metamorphism in the western part of the Tavþanlý Zone is 24±3 kbar and 430 ± 30 C° (Okay, 2002).

In the Halilbaðý region, where the temperature during the metamorphism was slightly higher, the metabasic rocks contain the mineral assemblages characteristic for the lawsonite-eclogite: sodic pyroxene + garnet + sodic amphibole + lawsonite (Kulaksýz, 1981; Monod et al., 1991; Davis and Whitney, 2006, 2008; Whitney and Davis, 2006; Çetinkaplan et al., 2008). The lawsonite eclogites in the Halilbaðý region are intercalated with blueschists and garnet-bearing blueschists. The peak P-T conditions in the Halilbaðý region have been estimated as 22-24 kbar pressure and 520 C° temperature (Davis and Whitney, 2008).

The P-T conditions of the HP/LT metamorphism in the Sivrihisar Formation is not well constrained because of the strong overprint by the greenschist facies metamorphism. In the southeastern part of the Tavþanlý Zone in the Konya-Altýnekin region the peak P-T conditions of the blueschist facies metamorphism has been estimated as 9-11 kbar pressure and 375-400 °C temperature (Droop et al., 2005).

Metamorphism, in terms of recrystallization and associated deformation, is not apparent in the field in the rocks of the Ovacýk Complex. However, close petrographic examination of the basalts usually reveals HP minerals such as lawsonite, sodic pyroxene and aragonite in the veins and amygdales of the rock (Okay, 1982). The magmatic clinopyroxene in the basalts has commonly been partly or totally pseudomorphed by aegerine-rich sodic pyroxene. Another interesting feature of the Ovacýk Complex is the replacement of primary micrite in the pelagic limestones by several centimeters large aragonite crystals (Topuz et al., 2006). This case of prograde aragonitization from northeast of Tavþanlý, unique in the world, points to the very low temperature and relatively high pressure values in the subduction-accretion complex.

The Ovacýk Complex is made up of numerous tectonic slices buried to different depths; therefore it is not possible to give a single peak P-T value for the unit. The general absence of recrystallization indicates that the temperatures were below 200 C°. The HP minerals in the basalts and their composition suggest that the pressure was in the range of 4 to 7 kbar (Okay, 1982; Topuz et al., 2006).

In some parts of the Ovacýk Complex metamorphism is more apparent, foliation has started to develop in finer grained rocks and the colour of the red cherts has become pale as a result of recrystallization. In such basaltic rocks the magmatic texture is still largely preserved, however the magmatic mineral assemblage is replaced by sodic pyroxene + lawsonite + chlorite + sphene. With an increase in penetrative deformation and in the intensity of foliation, sodic amphibole forms at the rims of the sodic pyroxenes through the reaction: sodic pyroxene + chlorite + quartz = sodic amphibole + lawsonite (Okay, 1980 *b*).

The Anatolian Ophiolite does not show any regional metamorphism. The magmatic mineral assemblage of plagioclase and pyroxene is well preserved in the ophiolitic gabbros (Önen, 2003). The diabase dykes in the peridotites consists of magmatic hornblende, which has partly or totally replaced augite, and altered plagioclase. The plagioclase in the diabase dykes is altered to fine grained aggregates of pumpellyite and albite. The secondary mineral assemblage in the diabase dykes is indicative for very low-grade metamorphism and shows that the Anatolian Ophiolite has not been affected by the HP/LT metamorphism observed in the Orhaneli Group.

The tectonic contact between the Orhaneli Group and the overlying ophiolitic mélange-ophiolite represents a major jump in the metamorphic grade. Rocks below the contact have undergone metamorphism at pressures of ca. 24 kbar and those above at pressures below 8 kbar; this difference in metamorphic pressures indicates that a rock column of 50 km in thickness has been excised along the contact.

Some regions in the Tavþanlý Zone show an Eocene high temperature - low pressure metamorphism related to the granitic magmatism. Such metamorphism has developed in the margins of the Eocene granodiorites at around Uludað, and is characterized by the formation of andalusite + cordierite + biotite + muscovite + K-feldspar + plagioclase mineral assemblage in the micaschists of the Kocasu formation (Okay and Satýr, 2006). In this region the development of a new foliation defined by biotite, cordierite and muscovite shows that the metamorphism is not just static contact metamorphism. The peak P-T conditions of this metamorphism overprinting the blueschist facies metamorphism is estimated as 2 ±1 kbar and 575 ± 50 °C Rb-Sr muscovite and biotite isotopic analyses from a single specimen gave cooling ages of 46 ± 3 Ma and 39 ±1 Ma, respectively (Okay and Satýr, 2006).

A different type of metamorphism in the Orhaneli Group has been described from southeast of Sivrihisar. The metamorphic belt extending southeastward from Sivrihisar is made up of an intercalation of marble, calc-schist and mica-schist (Figure 10, Türkay and Kuþçu, 1992). Lithostratigraphically the sequence resembles the Sivrihisar formation, however, Whitney (2002) has described from this region micaschists and quartzites with andalusite, kyanite, sillimanite, staurolite and garnet. The relation of this amphibolite facies metamorphism with the blueschist facies one is not known. However, considering the widespread presence of Eocene granitoids in this region, it is likely that the amphibolite facies metamorphism is of Eocene age.

K-Ar and Ar-Ar isotopic determinations in HP/LT metamorphic rocks give frequently incompatible and contradicting ages due excess argon and incomplete equilibration at these low temperatures (e.g., Arnaud and Kelley, 1995; Scaillet, 1996; Sherlock and Kelley, 2001; Warren et al., 2005). The K-Ar and Ar-Ar ages from the blueschists of the Tavþanlý Zone range between 175 Ma and 60 Ma (Çoðulu and Krummenacher, 1967; Okay and Kelley, 1994; Harris et al., 1994; Sherlock et al., 1999). A detailed study by Sherlock (1998) on this topic has shown that this spread in Ar-Ar ages has no geological meaning but is due to excess argon. In contrast Rb-Sr phengite ages from four blueschists sampled between Tavþanlý and Sivrihisar are coherent and range between 78.5 ± 1.6 Ma and 82.8 ± 1.7 Ma (Sherlock et al., 1999). The relatively low peak metamorphic temperatures in the Orhaneli Group (430-450 °C) imply that the Rb-Sr ages reflect the age of the HP/LT metamorphism. Similar Rb-Sr phengite ages from the blueschists of the Tavþanlý and Sivrihisar regions, separated by a distance of 130 km, indicates that the Orhaneli Group has undergone the HP/LT metamorphism during the Campanian (80 ± 2 my). The isotopic ages from the blueschists of the Konya-Altýntekin region also correspond to Campanian (Giles Droop, oral communication).

A firm upper age limit for the HP/LT metamorphism is given by the 53 Ma Sivrihisar (Sherlock et al., 1999) and Orhaneli granodiorites (Harris et al., 1994), which are intrusive in the blueschists and shows that the Orhaneli Group has reached upper crustal levels during the Early Eocene.

The Ovacýk Complex has undergone a low-grade HP/LT metamorphism. The age of this metamorphism is not known. Furthermore, as the Ovacýk Complex represents an accretionary complex, one is not dealing with a single age of metamorphism. Considering that the northward subduction of the Neo-Tethys has started in the Albian, the metamorphism in the Ovacýk Complex could have encompassed the whole of the Late Cretaceous.

STRUCTURAL FEATURES OF THE TAVÞANLI ZONE

The metamorphic rocks of the Orhaneli Group show penetrative foliation and isoclinal folding, and have lost all traces of their primary sedimentary protolith. The foliation is gently dipping except at some fault zones. A mineral stretching lineation has developed in marble, metachert and metabasites defined mainly by calcite and sodic amphibole (Monod et al., 1991; Okay et al., 1998; Masuda et al., 2004). The mineral lineation trends approximately east-west in the Orhaneli, Tavþanlý and Mihaliççýk regions subparallel to the Ýzmir-Ankara suture (Figure 2), and N-S and NE-SW in the Sivrihisar (Monod et al., 1991) and Konya-Altýnekin regions (Eren, 2000), respectively. The axis of the isoclinals folds are subparallel to the mineral lineation. These observations indicate a very strong stretching subparallel to the present trend of the Ýzmir-Ankara suture. The shapes of the clasts in the metaconglomerates of the Orhaneli Group also indicate that the finite strain ellipsoid is of constrictional type (Figure 11). An analysis of the crystallographic orientation of the quartz crystals in the blueschists of the Halilbaðý Formation also produced a finite strain ellipsoid that falls in the field of apparent constriction ($0.2 < k < 0.8$, Monod et al., 1991). This deformation is coeval with the HP/LT metamorphism.

REGIONAL CONTACTS OF THE TAVÞANLI ZONE

TAVÞANLI ZONE - BORNOVA FLYSCH ZONE

The Tavþanlý Zone is in contact with the Bornova Flysch Zone south of Mustafakemalpaþa. The Bornova Flysch Zone consists of blocks and tectonic slices in a highly deformed Maastrichtian - Paleocene clastic matrix. In the western part of the Bornova Flysch Zone these blocks and slices are mainly tectonized Mesozoic limestone olistoliths, and in the eastern part mainly basalt, radiolarian chert and rare serpentinite. The Early Eocene (Cuisian, Akdeniz, 1980) marine limestones, which lie unconformably over the Bornova Flysch Zone indicate that the deformation is Paleocene in age (Figures 4 and 12). In contrast, the HP/LT metamorphism and the associated deformation in the Tavþanlý Zone are of Campanian age (~80 Ma), and the blueschists of the Tavþanlý Zone were on or near the surface before the Early Eocene and probably by Maastrichtian.

South of Susurluk and Mustafakemalpaþa three important tectonic belts - Bornova Flysch Zone, Tavþanlý Zone and Sakarya Zone are in contact (Figure 4, Akyüz and Okay, 1996). South of Çataldað the ophiolitic mélange passes westward into the greywacke and shale of the Bornova Flysch Zone. Both the ophiolitic mélange and the greywackes of the Bornova Flysch Zone are underlain by the Ýnönü Marble of the Tavþanlý Zone in the Kepsut-Dursunbey region, and by the Nilüfer Formation of the Sakarya Zone in the Çataldað area (Figures 4 and 5). Blueschist metabasites and greywacke crop out as a klippe over the Nilüfer Formation in the region south of Manyas (Akyüz and Okay, 1999). This relation indicate that the Orhaneli Group is thrust northwestward over the Sakarya Zone during the Late Paleocene-Eocene.

TAVÞANLI ZONE - SAKARYA ZONE

The Tavþanlý Zone is in contact with the Sakarya Zone along the Ýzmir-Ankara suture. In

Figure 11- Metaconglomerate with quartz clasts in the Kocasu Formation. The photographs are taken parallel (a) and at right angles to the finite axis of elongation and define a prolate finite strain ellipse.

the west the suture is represented by the dextral Göktepe and Eskiþehir faults. The Eskiþehir Fault, with a total dextral strike-slip displacement of over 100 km, constitutes the Ýzmir-Ankara suture and the northern boundary of the Tavþanlý Zone between south of Bursa and Eskiþehir (Figure 2). Most of the displacement along the Eskiþehir Fault occurred during the Oligocene (Okay et al., 2008 *b*). East of Eskiþehir the contact between the Tavþanlý and Sakarya zones jumps to the north. In this region and around Sivrihisar the ophiolitic mélange of the Tavþanlý Zone and the Karakaya Complex of the Sakarya Zone form south-vergent imbricate tectonic slices (Figures 10 and 13, Göncüoðlu et al., 2000; Okay et al., 2002). Farther east in the region south of Nallýhan the Nilüfer Formation (Lower Karakaya Complex) has been thrust southward over the ophiolitic mélange. These thrusts affect Middle Eocene rocks and hence their latest movements are younger than Middle Eocene (Yýkýlmaz, 2002).

The Ýzmir-Ankara suture, which trends east-west between Bursa and Mihaliççýk, makes a sharp southward bend east of Mihaliççýk and extends south-southeastwards towards Konya. In this region the Sakarya Zone is represented by the Haymana basin, the contact of the Haymana basin and the blueschists of the Tavþanlý Zone lies under the Neogene sediments.

TAVÞANLI ZONE - AFYON ZONE

The Tavþanlý Zone is in contact in the south with the Afyon Zone (Figure 2). The Afyon Zone, which is also part of the Anatolide-Tauride Block, is constituted mainly of metasedimentary rocks that have undergone low grade regional HP/LT metamorphism characterized by the common presence of carpholite (Candan et al., 2005). The rocks in the Afyon Zone commonly contain greenschist facies mineral paragenesis, the typical mineral assemblage in the metabasites is "actinolite + chlorite + albite + epidote; sodic amphibole is found rarely and lawsonite has not been described from the metabasites of the Afyon Zone (Candan et al., 2005). Carpholite, which is not described from the Tavþanlý Zone, has widely developed in the phyllites of the Afyon Zone. The peak P-T conditions of metamorphism in the Afyon Zone is 6-9 kbar and 350 °C, whereas in the Tavþanlý Zone they are 24 kbar and 440 °C.

Marble crops out in the contact zones between the Tavþanlý and Afyon zones, as the carbonate rocks are insensitive to metamorphic pressures it is difficult to map a contact between the two tectonic zones in such areas. An exception to this occurs in the Altýnekin region north of Konya. Here the blueschists are thrust over low-grade metamorphic rocks of the Afyon Zone known as the Ladik metamorphics (Eren, 1996; Eren et al., 2004; Droop et al., 2005). The contact between the Ovacýk Complex and the Afyon Zone in the region south of Tavþanlý has been reworked by the Early Miocene extensional tectonics (Figures 4 and 5, Iþýk and Tekeli, 2001; Iþýk et al., 2004).

GEOLOGICAL EVOLUTION

EARLY CRETACEOUS - BEGINNING OF SUBDUCTION AND FORMATION OF THE BACK-ARC OCEANIC CRUST

In the Early Cretaceous the northern part of the Anatolide-Tauride Block was a passive margin facing north to the Ýzmir-Ankara Neo-Tethyan ocean (Figure 14a). The Pontides, consisting of the Sakarya and Ýstanbul zones, lay north of the Ýzmir-Ankara ocean. Paleontological data from the radiolarian cherts from the ophiolitic mélange show that the age of the Ýzmir-Ankara ocean extends from Mid Triassic to Cretaceous (Bragin and Tekin, 1996; Tekin et al., 2002).

Data on the age of the initiation of northward subduction of the Ýzmir-Ankara Neo-Tethyan ocean are not clear. The arc magmatism in the Pontides has started in the Turonian, however, isotopic ages of the Elekdað blueschists and eclogites in the Central Pontides indicate ongoing subduction during the Albian (~105 ma, Okay et al., 2006). The geochemical features of the Pontide arc magmatism (Keskin et al., 2003) and the observation that the magmatism was wholly submarine are indicative of an extensional tectonic regime. This extensional tectonic setting might have resulted in the development of an oceanic back-arc basin over the northward subducting Neo-Tethyan ocean (Figure 14c-d). The Anatolian ophiolite represents not the Ýzmir-Ankara Neo-Tethyan ocean but rather this back-arc oceanic lithosphere formed during the Cretaceous. There is no isotopic or paleontological data on the age of this back-arc type Anatolian ophiolite. However, studies in the ophiolites worldwide have shown that the age of the sub-ophiolite metamorphism is close to the age of the ophiolite (e.g., Spray et al., 1984; Hacker et al., 1996). The 95-90 Ma ages from the sole of the Anatolian Ophiolite (Dilek et al., 1999; Parlak and Delaloye, 1999; Çelik et al., 2006) suggest that the Anatolian Ophiolite is of Cenomanian age.

LATE CRETACEOUS (CAMPANIAN) - CONTINENTAL SUBDUCTION AND METAMORPHISM

Following the complete subduction of the Ýzmir-Ankara Neo-Tethyan oceanic lithosphere, the northern margin of the Anatolide-Tauride Block entered in an intra-oceanic subduction zone and was metamorphosed at HP/LT conditions (Figure 14e-f). The Rb-Sr ages from the Orhaneli Group blueschists indicate that the continental crust was subducted to a depth of 80 km at 80 Ma (Campanian). The obduction of the ophiolite over the continental crust must have started with an intra-oceanic slicing (Figure 14e). The 90-95 Ma isotopic ages from the sub-ophiolite metamorphic rocks shows that the intra-oceanic thrusting began in the Cenomanian-Turonian. Biostratigraphic data from the blocks of the Bornova Flysch Zone also indicate that the foundering of the Anatolide-Tauride carbonate platform as a result of compression started at the late Cenomanian (Figure 12, Okay and Altýner, 2007).

The lithostratigraphic features of the Orhaneli Group imply that these rocks were not deposited on a continental margin but rather on a shelf or on a shallow-marine platform. This suggests that during the continental subduction the continental margin deposits were detached from their substratum and thrust southward (Figure 14f-g). These continental margin sequences are probably represented by the Lycian Nappes in the Taurides. The Mesozoic carbonate stratigraphy within the blocks of the Bornova Flysch Zone shows great similarity to some of the units in the Lycian Nappes (Okay and Altýner, 2007).

LATE CRETACEOUS (MAASTRICHTIAN) - PALEOCENE - EXHUMATION

The Lower Eocene marine deposits in the Tavþanlý Zone show that by the end of the Paleocene the Orhaneli Group was on the surface or very close to the surface. The marine character of the Early Eocene sediments indicates that the crust was of normal thickness in the Tavþanlý Zone at this time. The post-tectonic Eocene granodiorites, which intrude the Orhaneli Group, the ophiolitic mélange and the ophiolite in the Tavþanlý Zone also indicate that the main tectonism in the Tavþanlý Zone was completed by the end of the Paleocene.

Stratigraphic and sedimentological data from the Sakarya Zone indicate that the blueschists of the Orhaneli Group were locally on the surface by the Maastrichtian. In the southern part of the Sakarya Zone and in the Haymana basin the Maastrichtian deposits are represented by thick flysch-type clastics (Figure 12). In contrast, in the northern parts of the Sakarya Zone and in the Ýstanbul Zone the Maastrichtian-Paleocene interval is represented by deposition of marine limestone and marl. This shows that the source of the Maastrichtian clastics lay south of the Sakarya Zone. Although continental collision between the Sakarya Zone and the Anatolide-Tauride Platform had not started by the Maastrichtian, the two terranes must have been in close proximity. During the Maastrichtian the Sakarya Zone was receiving detritus from the uplifted and eroding Tavþanlý Zone. Upper Campanian - lower Maastrichtian debris flows with glaucophane-lawsonite pebbles have been described within the 5000-m-thick flysch sequence of the Haymana basin, (Batman, 1978), furthermore, serpentinite and blueschist clasts are common in the Paleocene-Eocene sandstones and conglomerates (Norman and Rad, 1971). Maastrichtian sandstones of the Gölpazarý Group from the region of Nallýhan contain clastic glaucophane grains (Yýkýlmaz, 2002).

Two coeval tectonic processes were responsible for the exhumation of the Orhaneli Group blueschists that were buried to 80 km depth (Okay et al., 1998). The first one is the detachment of the Orhaneli Group from its crystalline basement and its exhumation within the subduction channel bounded by a thrust at the base and a normal fault at the top (Figure 14g). The other process is the rupture of the subducting oceanic lithosphere from the continental one.

PALEOCENE - CONTINENTAL COLLISION

The continental crust thickens in the region of continental collision, undergoes uplift and erosion. Consequently an upward coarsening and regressive clastic sedimentation is observed in such regions, which is followed by uplift and erosion. The first clastic sedimentation on top of pelagic carbonates in the southern parts of the

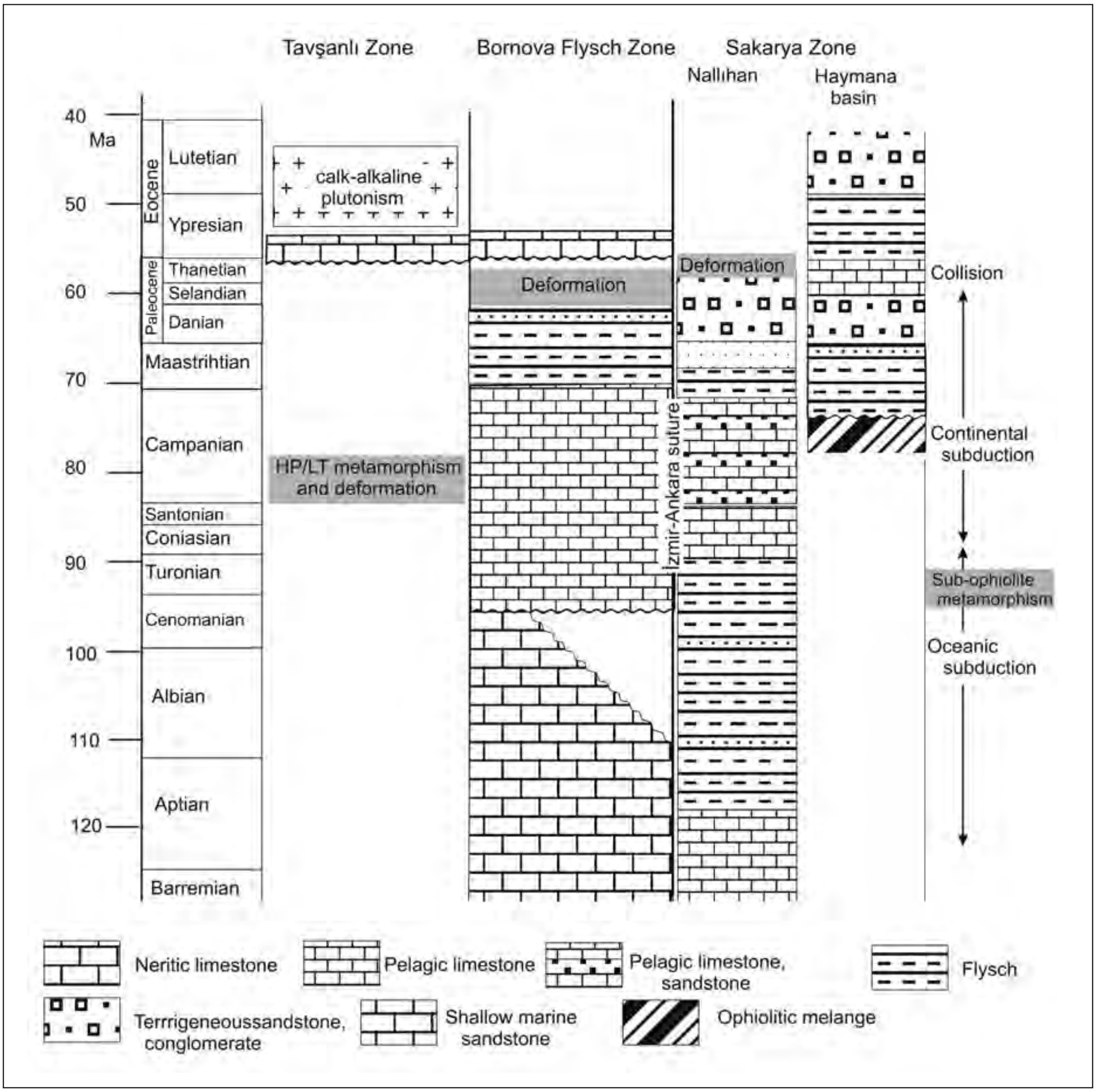

Figure 12- Stratigraphic and tectonic evolution of the Tavþanlý, Sakarya and Bornova Flysch zones.

Sakarya Zone starts in the Middle to Late Albian, and the siliciclastic turbidites intercalated with pelagic carbonates continue into the Campanian and Maastrichtian. In the late Maastrichtian the flysch deposition gives way to sedimentation of shallow marine sandstone, and in the Paleocene fluviatile sandstone and conglomerate are deposited (Tansel, 1980; Yýlmaz, 2008). The sedimentary data suggests that the collision between the Tavþanlý and Sakarya zones started in the Paleocene. A transition from the deep sea Maastrichtian turbidites to Paleocene continental clastics is also observed in the Haymana basin (Ünalan et al., 1976).

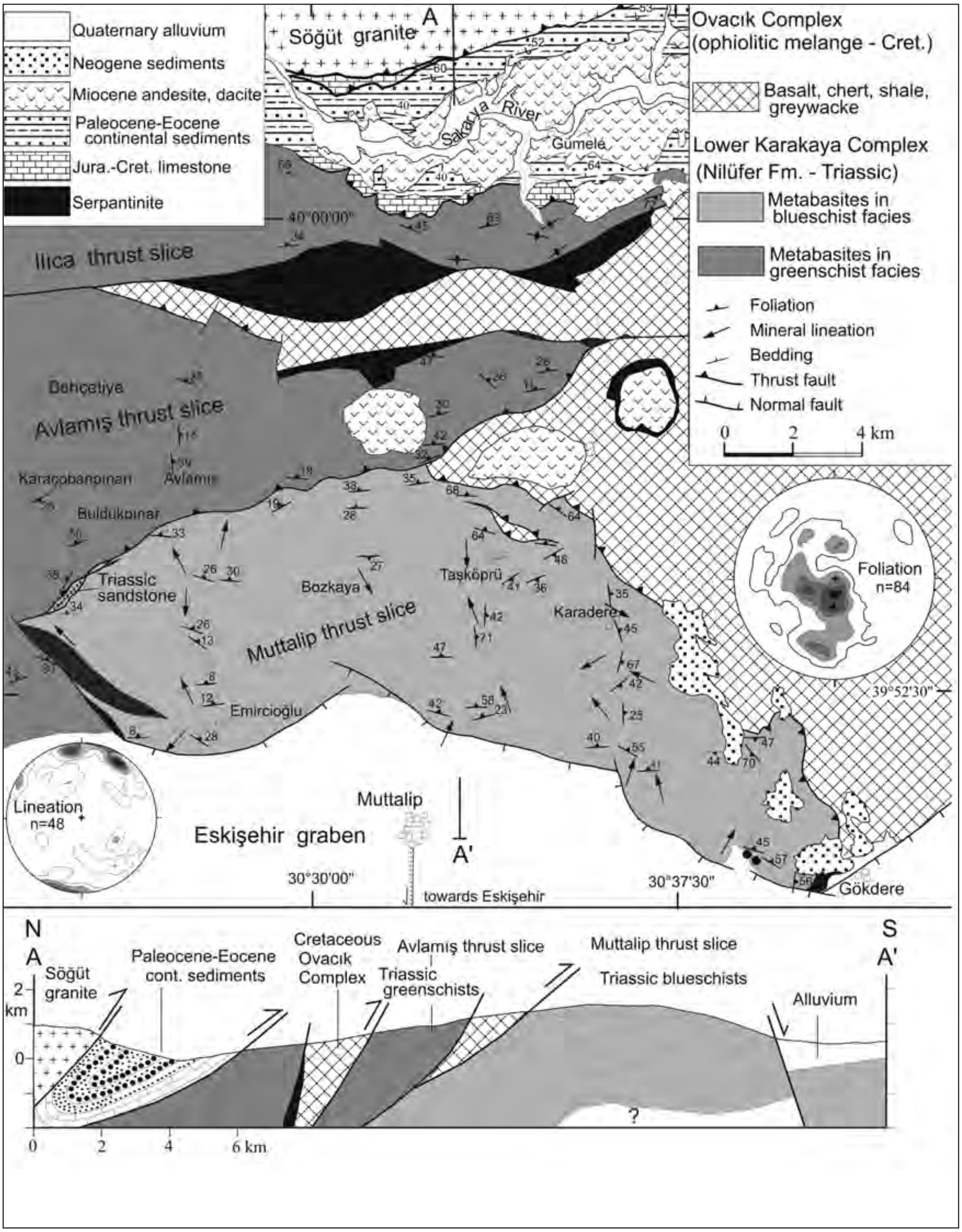

Figure 13- Geological map and cross-section of the region north of Eskiþehir illustrating the imbricate tectonic contacts between the Tavþanlý and Sakarya zones (modified from Okay et al., 2002).

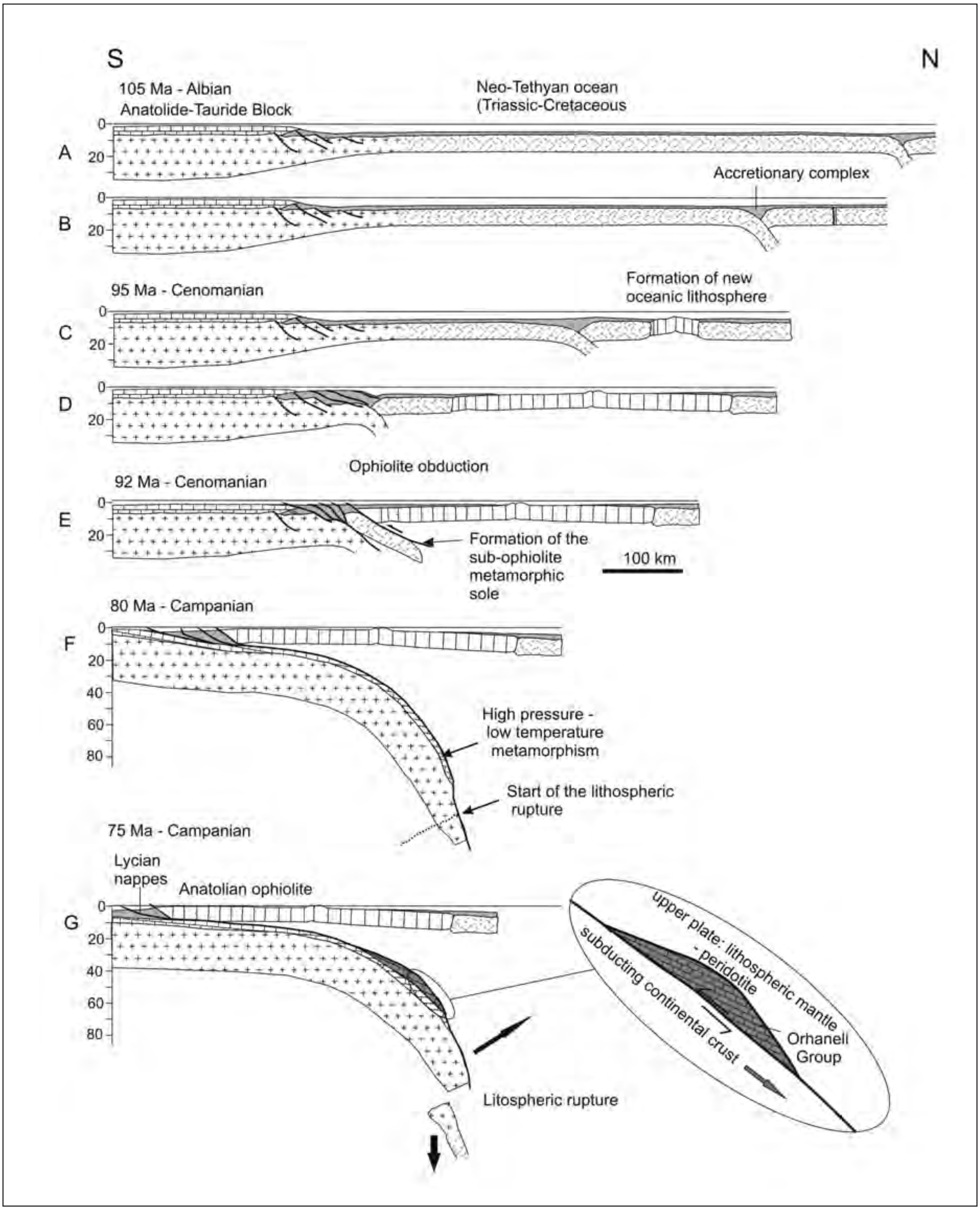

Figure 14- Gelogical evolution of the Tavþanlý Zone. For explanation see the text. The Figure is inspired from Lippard et al. (1986).

Sedimentary sequences on both sides of a suture are expected to show features of deposition on continental margins. However, such sequences are not recognized north and south of the Ýzmir-Ankara suture in northwest Turkey. Shallow marine Jurassic limestones crop out within 1.5 km of the Ýzmir-Ankara suture south of Bursa (Figure 4). In contrast in the Eastern Pontides the Jurassic-Cretaceous sediments show an increasingly deeper marine character towards the Ýzmir-Ankara-Erzincan suture. The absence of such deep marine Jurassic-Cretaceous sequences in the western part of the Sakarya Zone suggests major strike-slip faulting following continental collision (e.g. Okay et al., 2008).

EOCENE - CALK-ALKALINE MAGMATISM AND ASSOCIATED LOW-PRESSURE METAMORPHISM

Rocks of the Tavþanlý Zone are unconformably overlain by the Lower Eocene shallow marine limestones. This provides an upper age limit for the deformation related to continental collision in the Tavþanlý Zone. Furthermore, several Early to Mid-Eocene calc-alkaline plutons have intruded the Tavþanlý Zone between 53 and 45 Ma. These Eocene plutons are of post-tectonic character and intrude the Orhaneli Group, ophiolitic mélange and the ophiolite and cut the tectonic contacts between these units. A low pressure dynamo-thermal metamorphism has developed around the Eocene plutons south of Uludað characterized by the andalusite + cordierite + biotite + muscovite + K-feldspar + plagioclase paragenesis in the metapelitic rocks (Okay and Satýr, 2006).

Two hypotheses have been suggested for the genesis of Eocene magmatism: slab-break off (Altunkaynak, 2007; Karacýk et al., 2008) and arc magmatism (Okay and Satýr, 2006). In the slab-break off hypothesis asthenospheric mantle intrudes into the rupture between the oceanic and continental lithosphere and the additional heat brought by the asthenospheric mantle leads to partial melting and magmatism. Both hypotheses result in the generation of magmas with similar geochemical and petrographic features. Nevertheless, the extension of the Eocene plutonic belt 140 km northwest of the Ýzmir-Ankara suture to the Marmara island and Karabiga makes the magmatic arc hypothesis more probable. Furthermore, as discussed above, the rupture between the oceanic and continental lithosphere most probably occurred during the Maastrichtian and not in the Eocene.

DISCUSSION AND CONCLUSIONS

The Tavþanlý Zone provides one of the best examples of continental crust subducted to a depth of 80 km and exhumed, while preserving to a large extent the HP/LT mineral assemblages. Most probably a major part of the subducted continental crust, including the lower crust, will never be exhumed. This has major implications in terms of the heterogeneity of the mantle and magma genesis.

The deformation and metamorphism in the Tavþanlý Zone is of Campanian age and constitutes the beginning Alpide orogeny in the western Anatolia. The deformation migrated southward and affected the Menderes Massif and most of the Taurides during the Mid Eocene; during the Miocene the Lycian nappes were thrust over the Tauride autochthon (e.g. Gutnic et al., 1979).

The Tavþanlý Zone shows strong similarities to the Semail ophiolite and the underlying HP/LT metamorphic rocks in terms of tectonic setting, geological evolution and in the timing of the deformational and metamorphic events. The Semail ophiolite has a length of over 400 km and a width of 150 km, its thickness prior to emplacement is thought to be 15-20 km (Lippard et al., 1986; Hacker et al., 1996); the thickest part of the ophiolitic sequence (8-12 km) is made up of peridotites (Boudier and Coleman, 1981; Lippard et al., 1986). The zircon ages from the plagiogranites of the Semail ophiolite (95.4-

94.5 my) and the radiolarian ages from the cherts overlying the ophiolite show that the Semail ophiolite is Cenomanian in age. The ~93.5 Ma hornblende Ar-Ar ages from the subophiolite metamorphic rocks show that within 2 ma following the formation of the Semail ophiolite at a mid-ocean ridge, it was emplaced, probably along a transform fault, over the neighbouring oceanic crust (Hacker et al., 1996; Warren et al., 2005). The Ar-Ar ages from the base of the Anatolian Ophiolite are also in the range of 95-90 Ma. The Semail ophiolite was first emplaced over an oceanic crust and then over the continental margin of Arabia. During its emplacement over the Arabia margin, it bulldozed the continental margin sequences in its front. These continental margin sequences, which crop out southwest of the Semail ophiolite, are known as the Hawasina nappes and show close similarities to the Lycian Nappes in the Taurides. The Arabian continental crust under the Semail ophiolite underwent HP/LT metamorphism during the Campanian (82-79 my, e.g. Warren et al., 2005), which is of the same age as the HP/LT metamorphism in the Tavþanlý Zone. The blueschists and eclogites under the Semail ophiolite are unconformably overlain by the marine Eocene deposits, as in the Tavþanlý Zone (Poupeau et al., 1998). The main difference between the Tavþanlý Zone and Oman is that in the Tavþanlý Zone the ophiolite emplacement was followed by the continent-continent collision, whereas Indian ocean lies north of the Semail ophiolite. Another difference is that the Semail ophiolite is thrust over the Arabian platform, whereas the Anatolian ophiolite over the Anatolide-Tauride Block. The northern and southern branches of the Neo-Tethys, which bound the Anatolide-Tauride Block, join up in Iran and continue southeastward as the Zagros Neo-Tethyan ocean. The Anatolian ophiolite is derived from the Ýzmir-Ankara Neo-Tethyan ocean (the northern branch of the Neo-Tethys), and the Semail ophiolite from the Zagros ocean.

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REFERENCES

Akdeniz, N. 1980. Baþlamýþ Formasyonu. Jeoloji Mühendisliði, 10, 39-47.

Akyüz, S. and Okay, A.I. 1996. A section across a Tethyan suture in northwest Turkey. International Geological Review, 38, 405-418.

Akyüz, H.S. and Okay, A.I. 1999. Manyas güneyinin jeolojisi (Balýkesir) ve maviþistlerin tektonik konumu. Maden Tetkik ve Arama Dergisi, 120, 105-120.

Altunkaynak, Þ. 2007. Collision-driven slab breakoff magmatism in northwestern Anatolia, Turkey. Journal of Geology, 115, 63-82.

Arnaud, N.O. and Kelley, S.P. 1995. Evidence for excess argon during high-pressure metamorphism in the Dora Maira Massif (western Alps, Italy), using an ultra-violet laser ablation microprobe ^{40}Ar-^{39}Ar technique. Contributions to Mineralogy and Petrology, 121, 1-11.

Asutay, H.J., Küçükayman, A. and Gözler, M.Z. 1989. Daðküplü (Eskiþehir kuzeyi) ofiyolit karmaþýðýnýn stratigrafisi, yapýsal konumu and kümülatlarýn petrografisi. Maden Tetkik ve Arama Dergisi, 109, 1-8.

Ataman, G. 1972. L'age radiometrique du massif granodioritique d'Orhaneli. Türkiye Jeoloji Kurumu Bülteni, 15, 125-130.

Ataman, G. 1973 *a*. Gürgenyayla (Domaniç) granodiyoritik kütlesinin radyometrik yaþý. Türkiye Jeoloji Kurumu Bülteni, 16, 22-26.

_____ 1973 *b*. Mihallýççýk (Eskiþehir) granit kütlesinin radyometrik yaþý üzerine bir çalýþma. TÜBÝTAK IV Bilim Kongresi, 1-5.

Baþ, H. 1986. Domaniç-Tavþanlý-Kütahya-Gediz yöresinin Tersiyer jeolojisi. Jeoloji Mühendisliði, 27,11-18.

Batman, B. 1978. Haymana kuzeyinin jeolojik evrimi ve yöredeki melanjýn incelenmesi I: Stratigrafi birimleri. Yerbilimleri, 4, 95-124.

Beccaletto, L., Bonev, N., Bosch, D. and Bruguier, O. 2007. Record of a Paleogene syn-collisional extension in the north Aegean region: evidence from the Kemer micaschists (NW Turkey). Geological Magazine, 144, 393-400.

Bingöl, E., M. Delaloye, and Genç, Þ. 1994. Magmatism of northwestern Anatolia. Excursion Guide Book. International Volcanological Congress, pre-congress excursion September 06-11, 1994, 56 s.

Blumenthal, M.M. 1956. Geologie des hohen Bolkardað, seiner nördlichen Randgebirge und westlichen Auslaufer. MTA Enstitüsü yayýnlarý, D7, 153 s.

Bragin, N.Y. and Tekin, U.K. 1996. Age of radiolarian chert blocks from the Senonian ophiolitic melange (Ankara, Turkey). The Island Arc, 5, 114-122.

Candan, O. Çetinkaplan, M., Oberhansli, R., Rimmele, G. and Akal, C. 2005. Alpine high-P/low-T metamorphism of the Afyon Zone and implications for the metamorphic evolution of Western Anatolia, Turkey. Lithos, 84, 102-124.

Çalapkulu, F. 1980. Horoz granodiyoritinin jeolojik incelemesi. Türkiye Jeoloji Kurumu Bülteni, 23, 59-68.

Çelik, Ö.F., Delaloye, M. and Feraud, G. 2006. Precise ^{40}Ar-^{39}Ar ages from the metamorphic sole rocks of the Tauride Belt Ophiolites, southern Turkey: implications for the rapid cooling history. Geological Magazine, 143, 213-227.

Çetinkaplan, M., Candan, O., Oberhänsli, R. and Bousquet, R. 2008. Pressure-temperature evolution of lawsonite eclogite in Sivrihisar; Tavþanlý Zone-Turkey. Lithos, 104, 12-32.

Çoðulu, E. 1965. Remarques sur les schistes a glaucophane et lawsonite de la region de Mihalliçcik (Turquie). Arch. Sc. Soc. Phys. His. Nat. Geneve, 18, 126-131.

Çoðulu, E. 1967. Etude pétrographique de la région de Mihaliççik (Turquie). Schweizerisch mineralogische und petrographische Mitteilungen, 47, 683-824.

_____ and Krummenacher, D. 1967. Problèmes géochronométriques dans le partie NW de l'Anatolie Centrale (Turquie). Schweizerisch mineralogische und petrographische Mitteilungen, 47, 825-831.

Davis, P.B. and Whitney, D.L., 2006. Petrogenesis of lawsonite and epidote eclogite and blueschist, Sivrihisar Massif, Turkey. Journal of Metamorphic Geology, 24, 823-849.

Davis, P.B. and Whitney, D.L. 2008. Petrogenesis and structural petrology of high-pressure metabasalt pods, Sivrihisar, Turkey. Contributions to Mineralogy and Petrology, 156, 217-241.

Delaloye, M. and Bingöl, E. 2000. Granitoids from western and northwestern Anatolia: geochemistry and modelling of geodynamic evolution. Int. Geol. Rev., 42, 241-268.

Dilek, Y. and Whitney, D.L. 1997. Counterclockwise P-T-t trajectory from the metamorphic sole of a Neo-Tethyan ophiolite (Turkey). Tectonophysics, 280, 295-310.

_____, Thy, P., Hacker, B. and Grundvig, S. 1999. Structure and petrology of Tauride ophiolites and mafic dike intrusions (Turkey): Implications for the Neotethyan ocean. Geological Society of America Bulletin, 111, 1192-1216.

_____ and ________ 2006. Age and petrogenesis of plagiogranite intrusion in the Ankara melange, central Turkey. Island Arc, 15, 44-57.

Droop, G.R., Karakaya, M.C., Eren, Y. and Karakaya, N. 2005. Metamorphic evolution of blueschists of the Altýnekin Complex, Konya area, south central Turkey. Geological Journal, 40, 127-153.

El-Shazly, A.E-D., Coleman, R.G. and Liou, J.G. 1990. Eclogites and blueschists from northeastern Oman: Petrology and P-T evolution. J. Petrol., 31, 629-666.

Eren, Y. 1996. Ilgýn-Sarayönü (Konya) güneyinde Bozdaðlar masifinin yapýsal özellikleri. Türkiye Jeoloji Bülteni, 39, 49-63.

_____ 2000. Tuzgölü havzasý güneybatýsýndaki (Altýnekin-Konya) temel kayaçlarýnýn jeolojisi. Haymana-Tuzgölü-Ulukýþla Basenleri Uygulamalý Çalýþma, Türkiye Petrol Jeologlarý Derneði Özel Sayý, 5, 113-126.

_____ Kurt, H., Rosselet, F. and Stampfli, G.M., 2004. Palaeozoic deformation and magmatism in the northern area of the Anatolide block (Konya), witness of the Palaeotethyan active margin. Eclogae geologica Helvetica, 97, 293-306.

Forbes, R.B., Evans, B.W. and Thurston, S.P. 1984. Regional progressive high-pressure metamorphism, Seward Peninsula, Alaska. Journal of Metamorphic Geology, 2, 43-54.

Gautier, Y. 1984. Déformations et métamorphismes associés à la fermeture téthysienne en Anatolie Centrale (Région de Sivrihisar, Turquie). Ph.D. thesis, University Paris-Sud, s 236 (unpublished).

Gökten, E. and Floyd, P.A. 2007. Stratigraphy and geochemistry of pillow basalts within the ophiolitic melange of the Izmir-Ankara-Erzincan suture zone: implications for the geotectonic character of the northern branch of Neotethys. International Journal of Earth Sciences, 96, 725-741.

Göncüoðlu, M.C., Özcan, A., Turhan, N. and Iþýk, A. 1992. Stratigraphy of the Kütahya region. "A geotraverse across Tethyan suture zones in NW Anatolia" excursion guidebook, for the International Symposium on the Geology of the Black Sea Region, Ankara, 3-11.

_____ Turhan, N., Þentürk, K., Özcan, A., Uysal, Þ. and Yalýnýz, M.K. 2000. A geotraverse across northwestern Turkey: tectonic units of the Central Sakarya region and their tectonic evolution. "Tectonics and Magmatism in Turkey and the Surrounding Area" (eds: Bozkurt, E., Winchester, J.A. and Piper, J.D.A.), Geological Society, London, Special Publications, 173, 139-162.

_____ Yalýnýz, M.K. and Tekin, U.K. 2006. Geochemistry, tectono-magmatic discrimination and radiolarian ages of basic extrusives within the Izmir-Ankara suture belt (NW Turkey): time constraints for the Neotethyan evolution. Ofioliti, 31, 25-38.

Gözler, M.Z., Cevher, F. and Küçükayman, A. 1985. Eskiþehir civarýnýn jeolojisi ve sýcak su kaynaklarý. Maden Tetkik ve Arama Dergisi, 103/104, 40-54.

Gutnic, M., Monod, O., Poisson, A. and Dumont, J.F. 1979. Géologie des Taurides Occidentales (Turquie). Mémoire Societe geologique de France no. 137, 109 s.

Gürsü, S. and Göncüoðlu, M.C. 2006. Petrogenesis and tectonic setting of Cadomian felsic igneous rocks, Sandýklý area of the western Taurides, Turkey. International Journal Earth Sciences, 95, 741-757.

Hacker, B.R., Mosenfelder, J.L. and Gnos, E. 1996. Rapid emplacement of the Oman ophiolite: Thermal and geochronological constraints. Tectonics, 15, 1230-1247.

Harris, N.B.W., Kelley, S.P. and Okay, A.I. 1994. Post-collision magmatism and tectonics in northwest Turkey. Contributions to Mineralogy and Petrology, 117, 241-252.

Hetzel, R. and Reischmann, T. 1996. Intrusion age of Pan-African augen gneisses in the southern Menderes Massif and the age of cooling after Alpine ductile extensional deformation. Geological Magazine, 133, 565-572.

Iþýk, V. and Tekeli, O. 2001. Late orogenic crustal extension in the northern Menderes massif (western Turkey): evidence for metamorphic core complex formation. International Journal Earth Sciences, 89, 757-765.

_____, _____and Seyitoðlu, G. 2004. The $^{40}Ar/^{39}Ar$ age of extensional ductile deformation and granitoid intrusion in the northern Menderes core complex: implications for the initiation of extensional tectonics in western Turkey. Journal of Asian Earth Sciences, 23,555-566.

Karacýk, Z., Yýlmaz, Y., Pearce, J.A. and Ece, Ö.I. 2008. Petrochemistry of the south Marmara granitoids, northwest Anatolia, Turkey. International Journal of Earth Sciences, DOI 10.1007/s00531-007-0222-y (in press).

Karaman, E. 1986. Altýnekin (Konya) çevresinin jeolojisi ve tektonik evrimi. Türkiye Jeoloji Kurumu Bülteni, 29, 157-170.

Kaya, O. 1972*a*. Tavþanlý yöresi ofiolit sorununun ana çizgileri. Türkiye Jeoloji Kurumu Bülteni, 15, 26-108.

_____ 1972*b*. Aufbau und Geschichte einer anatolischen Ophiolith - Zone. Z. deutsch. geol. Gesellschaft, 123, 491-501.

_____ 1981. Preliminary study on the paragenetic relationships in the polymetamorphic blueschist rocks of the Tavþanlý area, West Anatolia. Aegean Earth Sciences, 1, 27-43.

Kaya, O., Kozur, H., Sadeddin, W. and Helvacý, H. 2001. Late Norian conodont age for a metacarbonate unit in NW Anatolia, Turkey. Geobios, 34, 527-532.

Keskin, M., Ustaömer, T., and Yeniyol, M. 2003. Ýstanbul kuzeyinde yüzeyleyen Üst Kretase yaþlý volkano-sedimenter birimlerin stratigrafisi, petrolojisi ve tektonik ortamý. Ýstanbul'un jeolojisi sempozyumu Bildiriler Kitabý, 23-35.

Konak, N. 2002. Türkiye jeoloji haritasý, Ýzmir paftasý 1:500 000. MTA Genel Müdürlüðü, Ankara.

Kröner, A. and Þengör, A.M.C. 1990. Archean and Proterozoic ancestry in late Precambrian to early Palaeozoic crustal elements of southern Turkey as revealed by single-zircon dating. Geology, 18, 1186-1190.

Kulaksýz, S. 1978. Sivrihisar kuzeybatý yöresi eklojitleri. Yerbilimleri, 4, 89-94.

_____, 1981. Sivrihisar kuzeybatý yöresinin jeolojisi (Geology of the region of northwest of Sivrihisar). Yerbilimleri, 8, 103 124.

Lippard, S.J., Shelton, A.W. and Gass, I.G. 1986. The Ophiolites of Northern Oman. The Geological Society of London, Memoir 11, 178 p.

Lisenbee, A. 1971. The Orhaneli ultramafic-gabbro thrust sheet and its surroundings. In A.S. Campbell (Editor), Geology and History of Turkey. Petroleum Exploration Society of Libya, Tripoli, s 349-360.

_____ 1972. Structural setting of the Orhaneli ultramafic massif near Bursa, northwestern Turkey. Ph.D. thesis. Pennsylvania State University, s157 (unpublished).

Loos, S. and Reischmann, T. 1999. The evolution of the southern Menderes Massif in SW Turkey as revealed by zircon dating. J. Geol. Soc. London, 156, 1021-1030.

Lünel, T. 1967. Geology of Sübren, Karacaalan-Yukarý Çaðlayan area, Eskiþehir county, Turkey. Ph.D. Thesis, University of Bristol.

Lünel, T. 1986. Petrology of Gümele ultramafic suite of Eskiþehir Complex. METU Journal of Pure and Applied Sciences, 19, 167-195.

Lytwyn, J.N. and Casey, J.F. 1995. The geochemistry of postkinematic mafic dike swarms and subophiolitic metabasites, Pozantý-Karsantý ophiolite, Turkey: Evidence for ridge subduction. Geological Society of America Bulletin, 107, 830-850.

Masuda, T., Nakayama, S., Kimura, N., Onodera, K. and Okamoto, A. 2004. Triaxial stress state deep in orogenic belts: an example from Turkey. Journal of Structural Geology, 26, 2203-2209.

Monod, O., Andrieux, J., Gautier, Y. and Kienast, J.R. 1991. Pontides Taurides relationships in the region of Eskiþehir (NW Turkey). Bulletin of the Technical University of Istanbul, 44, 257 277.

Norman, T. and Rad, M.R. 1971. Çayraz (Haymana) civarýndaki Horhor (Eosen) Formasyonunda alttan üste doðru doku parametrelerinde ve aðýr mineral bolluk derecelerinde deðiþmeler. Türkiye Jeoloji Kurumu Bülteni, 14, 205-225.

Okay, A.I. 1980 *a*. Mineralogy, petrology and phase relations of glaucophane-lawsonite zone blueschists from the Tavþanlý region, northwest Turkey. Contributions to Mineralogy and Petrology, 72, 243-255.

_____ 1980 b. Lawsonite zone blueschists and a sodic amphibole producing reaction in the Tavþanlý region, northwest Turkey. Contributions to Mineralogy and Petrology, 75, 179-186.

_____ 1981. Kuzeybatý Anadolu'daki ofiyolitlerin jeolojisi ve maviþist metamorfizmasý (Tavþanlý - Kütahya). (Geology and blueschist metamorphism of ophiolites in northwest Anatolia (Tavþanlý-Kütahya). Türkiye Jeoloji Kurumu Bülteni, 24, 85-95.

_____ 1982. Incipient blueschist metamorphism and metasomatism in the Tavþanlý region, northwest Turkey. Contributions to Mineralogy and Petrology, 79, 361-367.

_____ 1984. Distribution and characteristics of the northwest Turkish blueschists. In: The Geological Evolution of the Eastern Mediterranean (eds. J.E. Dixon and A.H.F. Robertson), Geological Society Special Publication No. 17, 455-466.

Okay, A.I. 1985. Kuzeybatý Anadolu'da yer alan metamorfik kuþaklar. Ketin Simpozyumu Kitabý'nda, Türkiye Jeoloji Kurumu Yayýný, 83-92.

_____ 1989. Alpine-Himalayan blueschists. Annual Reviews of the Earth and Planetary Sciences, 17, 55-87.

_____ 1997. Jadeite-K-feldspar rocks and jadeitites from northwest Turkey. Mineralogical Magazine, 61, 835-843.

_____ 2002. Jadeite - chloritoid - glaucophane-lawsonite schists from northwest Turkey: unusually high P/T ratios in continental crust. Journal of Metamorphic Geology, 20, 757-768.

_____ 2004. Tectonics and High Pressure Metamorphism in northwest Turkey. Field trip guide book - P01, 32nd International Geological Congress, APAT, Italy, 56 pp.

_____ and Kelley, S.P., 1994. Tectonic setting, petrology and geochronology of jadeite + glaucophane and chloritoid + glaucophane schists from northwest Turkey. Journal of Metamorphic Geology, 12, 455-466.

_____ Harris, N.B.W. and Kelley, S.P. 1998. Exhumation of blueschists along a Tethyan suture in northwest Turkey. Tectonophysics, 285, 275-299.

_____ and Tüysüz, O. 1999. Tethyan sutures of northern Turkey. In "The Mediterranean Basins: Tertiary extension within the Alpine orogen" (eds. B. Durand, L. Jolivet, F. Horváth and M. Séranne), Geological Society, London, Special Publication 156, 475-515.

_____ and Satýr, M. 2006. Geochronology of Eocene plutonism and metamorphism in northwest Turkey: evidence for a possible magmatic arc. Geodinamica Acta, 19, 251-266.

_____ Tüysüz, O., Satýr, M., Özkan-Altýner, S., Altýner, D., Sherlock, S. and Eren, R.H. 2006. Cretaceous and Triassic subduction-accretion, HP/LT metamorphism and continental growth in the Central Pontides, Turkey. Geological Society of America Bulletin, 118, 1247-1269.

Okay, A.I. and Altýner, D. 2007. A condensed Mesozoic section in the Bornova Flysch Zone: A fragment of the Anatolide-Tauride carbonate platform. Turkish Journal of Earth Sciences, 16, 257-279.

_____ Monod, O. and Monié, P. 2002. Triassic blueschists and eclogites from northwest Turkey: andstiges of the Paleo-Tethyan subduction. Lithos, 64, 155-178.

_____ Satýr, M. and Shang, C.K., 2008. Ordovician metagranitoid from the Anatolide-Tauride Block, northwest Turkey -geodynamic implications. Terra Nova, 20, 280-288.

______,______,Zattin, M., Cavazza, W. and Topuz, G. 2008. An Oligocene ductile strike-slip shear zone: Uludað Massif, northwest Turkey - implications for the escape tectonics. Geological Society of America Bulletin, 120, 893-911.

Okrusch, M. and Bröcker, M. 1990. Eclogites associated with high-grade blueschists in the Cyclades archipelago, Greece: A review. European Journal of Mineralogy, 2, 451-478.

Önen A.P. 2003. Neotethyan ophiolitic rocks of the Anatolides of NW Turkey and comparison with Tauride ophiolites. Journal of the Geological Society, London, 160, 947-962.

_____ and Hall, R. 1993. Ophiolites and related metamorphic rocks from the Kütahya region, north-west Turkey. Geological Journal, 28, 399-412.

_____ and _______ 2000. Sub-ophiolite metamorphic rocks from NW Anatolia, Turkey. Journal of Metamorphic Geology, 18, 483-495.

Özgen-Erdem, N., Akyaz, M. and Karabaþoðlu, A. 2007. Biostratigraphic interpretation and systematics of *Alveolina* assemblages from the Ilerdian-Cuisian limestones of southern Eskiþehir, Central Turkey. Journal of Asian Earth Sciences, 29, 911-927.

Özgül, L. and Göncüoðlu, M.C., 1999. Koçyaka metamorfik Complex'nin metamorfik evrimi: Batý

Orta Anadolu'da HP/LT metamorfizmalý tektonik bir birim. 52. Türkiye Jeoloji Kurultayý Bildiriler Kitabý, 279-286.

Özgül, N. 1976. Toroslarýn bazý temel jeoloji özellikleri. Türkiye Jeoloji Kurumu Bülteni, 19, 65-78.

Parlak, O. 2000. Geochemistry and significance of mafic dyke swarms in the Pozantý-Karsantý ophiolite (southern Turkey). Turkish Journal of Earth Sciences, 24, 29-38.

_____ and Delaloye, M. 1996. Geochemistry and timing of post-metamorphic dyke emplacement in the Mersin Ophiolite (southern Turkey): New age constraints from $^{40}Ar/^{39}Ar$ geochronology. Terra Nova, 8, 585-592.

_____ and _______ 1999. Precise $^{40}Ar/^{39}Ar$ ages from the metamorphic sole of the Mersin ophiolite (southern Turkey). Tectonophysics, 301, 145-158.

Patrick, B.E. and Evans, B.E. 1989. Metamorphic evolution of the Seward Peninsula blueschist terrane. J. Petrol., 30, 531-555.

Perkins, D. III, Holland, T.J.B. and Newton, R.C. 1981. The Al_2O_3 contents of enstatite in equilibrium with garnet in the system $MgO-Al_2O_3-SiO_2$ at 15-40 kbar and 900-1600°C. Contributions to Mineralogy and Petrology, 78, 99-109.

Poupeau, G., Saddiqi, O., Michard, A., Goffe, B. and Oberhansli, R. 1998. Late thermal evolution of the Oman Mountains subophiolitic windows: Apatite fission-track thermochronology. Geology, 26, 1139-1142.

Rojay, B., Altýner, D., Altýner, S.Ö., Önen, A.P., James, S. and Thirwall, M.F. 2004. Geodynamic significance of the Cretaceous pillow basalts from North Anatolian Ophiolitic Melange Belt (Central Anatolia, Turkey): geochemical and paleontological constraints. Geodinamica Acta, 17, 349-361.

Satir, M. and Friedrichsen, H. 1986. The origin and evolution of the Menderes Massif, W Turkey: a rubidium/strontium and oxygen isotope study. Geologische Rundschau, 75,703-714.

Scaillet, S. 1996. Excess ^{40}Ar transport scale and mechanism in high-pressure phengites: a case study from an eclogitised metabasite of the Dora-Maira nappe, Western Alps. Geochimica Cosmochimica Acta, 60, 1075-1090.

Servais, M. 1981. Donnees pre'liminaires sur la zone de suture medio-tethysienne dans la region d'eskiþehir (NW Anatolie). C.R. Acad. Sc. Paris, 293, ser. II, 83-86.

_____ 1982. Collision et suture téthysienne en Anatolie Centrale, étude structurale et métamorphique (HP-BT) de la zone nord Kütahya. Ph.D. Thesis, Universite de Paris-Sud, Centre d'Orsay, 374 s.

Sherlock, S.C. 1998. Exhumation of blueschist-facies assemblages from western Turkey: The significance of ^{40}Ar-^{39}Ar ages and excess argon in a HP/LT terrain. Ph.D. Thesis, Open University, UK.

Sherlock, S., Kelley, S.P., Inger, S., Harris N. and Okay, A.I. 1999. ^{40}Ar-^{39}Ar and Rb-Sr geochronology of high-pressure metamorphism and exhumation history of the Tavsanli Zone, NW Turkey. Contributions to Mineralogy and Petrology, 137, 46-58.

_____ and _______ 2001. Excess argon evolution in HP-LT rocks: a UVLAMP study of phengite and K-free minerals, NW Turkey. Chemical Geology, 182, 619-636.

Spray, J.G., Bebien, J., Rex, D.C. and Roddick, J.C. 1984. Age constraints on the igneous and metamorphic evolution of the Hellenic-Dinaric ophiolites. In: J.E. Dixon and A.H.F. Robertson (Editors), The Geological Evolution of the Eastern Mediterranean. Geological Society of London, Special Publications, 17, 619-627.

Tankut, A. 1980. The Orhaneli Massif, Turkey. In: A. Panayiotou (Editor), Proceedings International Ophiolite Symposium, Cyprus 1979, 702-713.

_____ Dilek Y. and Önen P., 1998. Petrology and geochemistry of the Neo-Tethyan volcanism as revealed in the Ankara melange, Turkey. Journal of Volcanology and Geothermal Research, 85, 265-284.

Tansel, Ý. 1980. Nallýhan ve dolayýnýn biyostratigrafi incelemesi. Yerbilimleri, 5/6, 31-47.

Tekin, U.K., Göncüoðlu, M.C. and Turhan, N. 2002. First evidence of Late Carnian radiolarians from the Izmir-Ankara suture complex, central Sakarya, Turkey: implications for the opening age of the Izmir-Ankara branch of Neo-Tethys. Geobios, 35, 127-135.

Topuz, G., Okay, A.I., Altherr, R., Meyer, H.P. and Nasdala, L. 2006. Partial high-pressure aragonitization of micritic limestones in an accretionary complex, Tavþanlý Zone, NW Turkey. Journal of Metamorphic Geology, 24, 603-613.

Turhan, N. 2002. Türkiye jeoloji haritasý, Ankara paftasý 1:500 000. MTA Genel Müdürlüðü, Ankara.

Türkay, O. and Kuþçu, M. 1992. Atlas-Çaykoz (Sivrihisar-Eskiþehir) dolayýnýn jeolojisi ve mermer yataklarý. C.Ü. Mühendislik Fakültesi Dergisi, Yerbilimleri, 9, 59-65.

Ünalan, G., Yüksel, V., Tekeli, T., Gönenç, O., Seyirt, Z. and Selahi, H. 1976. Haymana-Polatlý yöresinin (güneybatý Ankara) Üst Kretase-Alt Tersiyer stratigrafisi ve paleocoðrafik evrimi. Türkiye Jeoloji Kurumu Bülteni, 19, 159-176.

Van der Kaaden, G. 1966. The significance and distribution of glaucophane rocks in Turkey. Bulletin of the Mineral Research and Exploration Institute of Turkey, 67, 37-67.

von Raumer J.F., Stampfli, G.M., Borel G. and Bussy, F. 2002. Organization of pre-Variscan basement areas at the north-Gondwanan margin. International Journal of Earth Sciences, 91, 35-52.

Warren, C.J., Parrish, R.A., Waters, D.J. and Searle, M.P. 2005. Dating the geologic history of Oman's Semail ophiolite: insights from U-Pb geochronology. Contributions to Mineralogy and Petrology, 150, 403-422.

Whitney, D.L. 2002. Coexisting andalusite, kyanite, and sillimanite: Sequential formation of three Al_2SiO_5 polymorphs during progressive metamorphism near the triple point, Sivrihisar, Turkey. American Mineralogist, 87, 405-416.

Whitney, D. L. and Davis, P. B. 2006. Why are lawsonite eclogites so rare?: Metamorphism and preservation of lawsonite eclogite, Sivrihisar, Turkey. Geology, 34, 473-476.

Williams, H. and Smyth, R. 1973. Metamorphic aureoles beneath ophiolite suites and Alpine peridotites: tectonic implications with west Newfoundland examples. American Journal of Science, 273, 594-621.

Woodcock, N.H. and Robertson, A.H.F. 1977. Origins of some ophiolite-related metamorphis rocks of the "Tethyan" belt. Geology, 5, 373-376.

Yeniyol, M. 1979. Yunak (Konya) magnezitlerinin oluþum sorunlarý, deðerlendirilmesi ve yöre kayaçlarýnýn petrojenezi. Ýstanbul Yerbilimleri, 3, 21-51.

Yýkýlmaz, M.B. 2002. Eosen yaþlý bir kývrým-bindirme kuþaðýnýn yapýsý (Nallýhan-Ankara). Yüksek Lisans Tezi, Ýstanbul Teknik Üniversitesi, Avrasya Yerbilimleri Enstitüsü, 85 s.

Yýlmaz, Ý.Ö. 2008. Cretaceous pelagic red beds and black shales (Aptian-Santonian), NW Turkey: Global oceanic anoxic and oxic events. Turkish Journal of Earth Sciences, 17, 263-296.

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GEOLOGY OF THE KÜTAHYA-BOLKARDAĐ BELT

M. Cemal GÖNCÜOĐLU*

ABSTRACT.- Kütahya-Bolkardað Belt is one of the subunits of the Tauride-Anatolide Terrane extending from the Aegean Sea to the Hýnzýr Mountains. It includes numerous tectonic slices, formed during the closure of the Ýzmir-Ankara Oceanic branch of the Neotethys. The tectonic slices are mainly derived from three different tectonic settings: i- rocks representing the oceanic lithosphere and subduction- accretion prism of the Ýzmir-Ankara Ocean (ophiolites and ophiolitic mélanges), ii- flysch-type deposits that were formed in foreland-basins on the northern and passive edge of the Tauride-Anatolide platform in front of the southward advancing nappes (olistostromes with olistoliths, sedimentary mélanges), and iii- successions, in some cases with HP/LT metamorphism, representing the slope margin and external platform of the northern Tauride-Anatolide margin. Rock-units of the Kütahya-Bolkardað Belt surround the HT/LP Menderes Core Complex and are also observed as slices or klippen in the massif, or as nappes to the south of it. The rocks of the Ýzmir-Ankara oceanic lithosphere occur as huge allochthonous bodies/tectonic slices and blocks within the mélange and olistostromes. The fossil data and geochemical data obtained suggest the following: The earliest "oceanic" volcanism commenced during middle Carnian, the generation of ocean island-type (OIB) volcanics lasted from Bajocian to Abtian, where as the MOR-basalts spread from Aalenian to Turonian. Supra-subduction- and island-arc type basalts of Albian to Cenomanian age indicate an intra-oceanic subduction within the Ýzmir-Ankara Ocean. The mélanges are characterized by HP/LT metamorphism with a LP/LT overprint. Middle Maestrichtian olistostomes with olistoliths formed in foreland basins in front of the nappes include blocks of all kind of tectonic settings mentioned above. The flysch rocks are in depositional contact with the underlying platform and/or slope rocks of the Tauride-Anatolide passive margin. The Tauride-Anatolide slope and external platform deposits are partly affected by HP/LT metamorphism and occur as slices along the belt and as blocks within the flysch-basins. In Afyon area the Late Permian transgresses onto the Precambrian basement, whereas in Konya, more internal in regard to the platform, the Devonian carbonate platform is drowned and covered by back-arc-type sediments and volcanism of Carboniferous age. All along the belt, early Late Permian unconformably covers a slightly metamorphosed and deformed basement, attributed to a Variscan event within the Tauride-Anatolide platform. The Lower Triassic sequences unconformably covering the older units and starting with volcanogenic continental clastics, pass into the marine carbonates by Anisian. In the allochtonous belonging to the more internal platform, the Ladinian-Lower Cretaceous sequences are represented with thick platform carbonates. The first deep marine sediments take place in some slices of these sequences which is interpreted as the initial rifting of the Ýzmir-Ankara oceanic branch. While only the slope sediments accompanying with transitional ocean crust volcanics are observed in the allochtonous derived from the northernmost part of the Taurid-Anatolid platform, the Ladinian-Lower Cretaceous sequences are represented by thick platform carbonates in the allochtonous of the inner platform. The transition from platform to slope-type deposits is in Malm in the allochtonous of the external platform, but Abtian in more internal parts. This indicates a stepwise deepening of the platform-margin. The presence of HP/LT metamorphic platform-margin sediments is indicative for a deep subduction of the attenuated continental-crust of the Tauride-Anatolide margin. The initial compression-slicing and nappe-emplacement must have realized prior to Middle Paleocene. Middle Paleocene-Middle Eocene in the Kütahya-Bolkardað Belt is characterized by shallow-marine or continental molasse-type deposition in the remnant basins on the platform.

Key Words: Kütahya-Bolkardað Belt, Tauride-Anatolide, geological evolution.

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INTRODUCTION

Tauride-Anatolide terrane is one of the main Alpine tectonic units of Turkey that was formed by opening and closure of oceanic branches of Neotethys. It represents a continental crust. In SW Greece (Gavrovo-Tripolitza zone) in the west of the Aegean Sea and together with extensions in central Iran in the east, this unit can be considered as a micro-continent reaching today's Sumatra Island in size. Taurides are defined as an independent unit since the earliest classification of the Anatolian tectonic units. In the context of plate tectonics (Þengör and Yýlmaz, 1981) and previous tectonic classifications (Ketin, 1966), it is divided into Tauride and Anatolide units.

Kütahya-Bolkardað Belt (KBB) is one of the subunits of the Tauride-Anatolide tectonic unit that was suggested by Özcan et al. (1989). KBB is located in the south of Ýzmir-Ankara suture and extends from Karaburun to Kütahya and from there to Bolkar and Hýnzýr Mountains, also surrounding the Menderes Core Complex (Figure 1). Allochthonous high pressure metamorphic units in the south of Menderes Core Complex and Lycian nappes are primarily segments of KBB. Goncuoglu et al. (1997*a*) has subdivided Tauride-Anatolide terrane into three components and redefined the KBB. According to this definition, KBB is composed of various tectonic slices of continental and oceanic crust origin displaying different type metamorphisms. These tectonic slices include:

I) Rocks representing the oceanic lithosphere and subduction-accretion prism of the Ýzmir-Ankara Ocean (ophiolites and ophiolitic mélange)

II) Flysch-type deposits that were formed in foreland-basins on the northern and passive edge of the Tauride-Anatolide platform in front of the southward advancing nappes.

III) Successions, in some cases with HP/LT metamorphism, representing the slope margin and external platform of the northern Tauride-Anatolide margin.

The units of KBB that surround northern and eastern margin of the Menderes Core Complex are observed as slices within the massif and as klippen and nappes on it, they can be shown as nappe fragments in the south of it.

Common features of the KBB and overall differences between the Tauride unit and KBB are:

I) KBB units have undergone poly-phase metamorphism. As a common feature, they are more or less affected by Alpine HP/LT metamorphism.

II) In all slices, Late Middle Permian marine transgression and/or Early Triassic unconformity are observed over a basement which was affected by Variscan deformation.

III) Mesozoic platform successions in relation to their original position in the northern margin of Tauride-Anatolide platform, gradually deepen during Jurassic to Early Cretaceous.

In this study, field-data from the Kütahya-Bolkardað Belt obtained by the projects of General Directorate of Mineral Research and Exploration carried out in Kütahya and Konya areas between 1982 and 1987, by the Turkish Petroleum Corporation projects carried out in the west of Tuz Lake between 1995-1996, and finally by TUBÝTAK projects carried out on the same belt between 1998 and 2003 will be presented briefly.

TAURIDE-ANATOLITE CONTINENTAL MARGIN UNITS

In KBB, continental margin successions of Tauride - Anatolide terrane are partially affected by subduction. They occur as HP/LT metamorphosed tectonic slices or as huge blocks in Upper Cretaceous olistostrome deposits. In these successions, a package that starts with the Middle Permian unconformity generates the first common reference plane. Fairly monotonous Middle Permian successions are angular uncon-

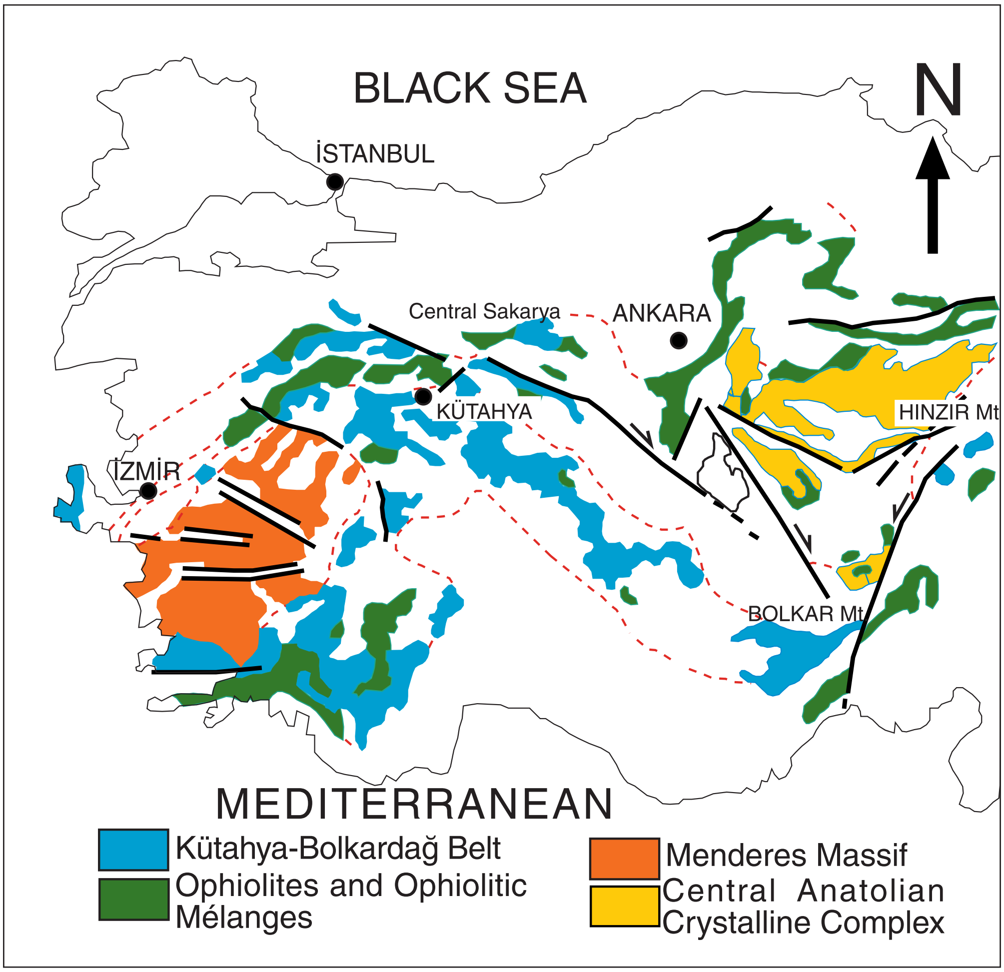

Figure 1- Location of the Kütahya-Bolkardað Belt within the Tauride-Anatolide Unit (simplified after Göncüoðlu et al., 1997*a*).

formable over various units through the belt. Second reference unit that is unconformable over several units, starts with Triassic terrestrial deposits and continues with Middle-Triassic-Cretaceous carbonate platform deposits.

After Cretaceous, transition to deep marine sediments can be seen in varying ages through the belt. At the end of Cretaceous, the depositing of the flysch sediments over the continental margin successions of KBB and starting of the slicing can be seen.

LOWER PERMIAN BASEMENT UNITS

In KKB, Lower Permian basement is represented by more than one unit. Lower contacts of all these units are thrust-faults.

Afyon-Type Late Neoproterozoic Basement

This type of basement outcrops at the northernmost part of the belt in the east of Eskisehir on Sömdiken Mountains (Göktepe Metamorphics; Göncüoðlu et al., 1996, 2000*a*), in the South of Kütahya on Yellice Mountains (Ýhsaniye Metamorphic Complex; Özcan et al., 1984, 1989) and in the north of Afyon in Köroðlu Mountain (Afyon Basement Complex; Gürsu and Göncüoðlu, 2008).

Metamorphic rocks that constitute the lowermost visible unit in Sömdiken Mountains (Figure 2) are described by Göncüoðlu et al. (2000*a*) as Göktepe Metamorphics.

Göktepe Metamorphics is constituted by graphite schist, garnet mica schist, quartz schist with rare marble bands, and metarhyolite, meta-quartz porphyry and metabasics among them as irregular outcrops. Since the unit has undergone poly-phase metamorphism and deformation, it is not possible to make detailed determinations about original successions of the unit. In the lower part of the unit, micaschists, para- and orthogneiss are common. Micaschists are described as quartz-muscovite- albite schist, quartz-muscovite-biotite schist, cholorite-muscovite-chlorite schist, biotite-albite-muscovite-quartz schist, garnet-biotite-muscovite-quartz-albite schist. Micaschists generally exhibit quite monotonous outcrops and they are probably of felsic volcanic or volcanoclastic origin. In relatively upper part of the section besides mica rich bands; thin marble bands, graphite-rich laminae and quartzite bands are observed. By considering these features it can be said that the mica-schists have sedimentary origin. Orthogneiss are observed in various masses in mica schists and have blastomylonitic structure. Main constituents are quartz and feldspar porphyroclasts, muscovite and red-brown colored biotite minerals. Further constituents are sphene, tourmaline, zircon, apatite and hematite as accessory minerals. It is considered that orthogneiss are originated from granitic-rhyolitic rocks; and muscovite-rich quartzo-feldspathic gneiss are originated from aplitic-pegmatitic. In the low and middle part of Göktepe Metamorphites in addition to mica schists and orthogneiss, 30-35 meters thick, green schist bands and lenses are commonly observed. Microscopically, various rocks like glaucophane albite-epidote-chlorite schist, glaucophane and phengite titanite-epidote-chlorite-albite schist, garnet and phengite epidote-chlorite-albite schist can be described. Glaucophanes are observed as long needle-like crystals and coarse crystals that possibly replaced coarse prismatic actinolite. Phengites occur as thin long crystals and in some sections grows within Mg-chlorite. It is considered as part of green schist forming massive lenses representing lava domes and dykes; whereas those intercalated with the mica schist represents basic volcanoclastics.

By examining macroscopic and microscopic features of Göktepe Metamorphics, it is understood that they are undergone two different metamorphic phases. In the first of these phases, parallel to dominant foliation in clastic and felsic originated rocks blastomylonitic texture and biotite + chloritoid + muscovite + garnet + chlorite + quartz + plagioclase paragenesis has been formed. As for in basic volcanic and volcanoclastic rocks appropriate to local foliation again chlorite + actinolite + epidote + garnet + plagioclase paragenesis has been formed. This paragenesis indicates that first metamorphic phase has happened in green schist facies conditions. In the second metamorphic phase, deformation is locally effective only. In this phase, in clastic and felsic volcanic rocks: muscovite + chlorite + stilpnomelane + albite paragenesis and in basic rocks: glaucophane + phengite + stilpnomelane + chlorite + albite paragenesis has been formed. A third phase paragenesis that represented by actinolite and white mica formation overprints HP/LT metamorphic paragenesis. The same paragenesis are observed in diabase dikes that intersects Göktepe Metamorphics. The last two phases are considered as a product of Alpine HP/LT metamorphism (Göncüoðlu et al., 2000*a*). Göktepe Metamorphics are overlain by Kayapýnar Marbles including a basal unit with quartz conglomerates and quartzite. This unit can be correlated with Ýhsaniye Metamorphic Complex (Özcan et al., 1989) or Afyon Basement Complex (Gürsu and Göncüoðlu, 2008) that are going to be defined below.

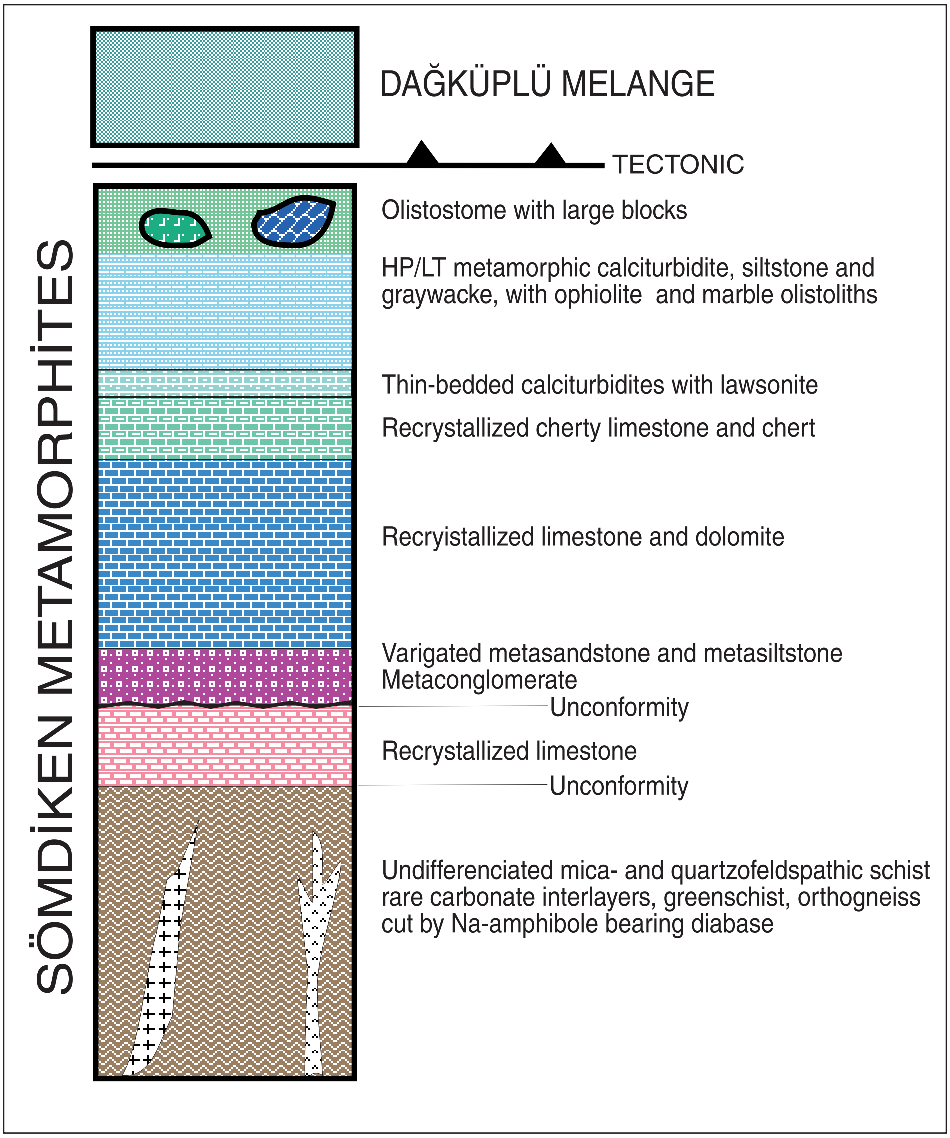

Figure 2- Stratigraphy and lithologies of the Afyon-type basement in Sömdiken Mountains to the NE of Eskiþehir (simplified after Göncüoðlu et al., 2000*a*).

In the north of Afyon between Bayat and Ýhsaniye, outcrops of KBB basement can be observed commonly (Figure 3).

Ýhsaniye Metamorphic Complex represents the lower part of metamorphic successions that located between Kütahya and Afyon. The unit includes mica schist and meta granitic rocks that were affected by poly-phase metamorphism, and marble, graphite-schist and quartzite in less amount (Figure 4).

Mica schists are generally represented by garnet - biotite- muscovite and quartz - albite - muscovite schist. It is interpreted that thick schist packages having homogeneous composition were originated from magmatic rocks. In between marbles, mica schists and graphite chlo-

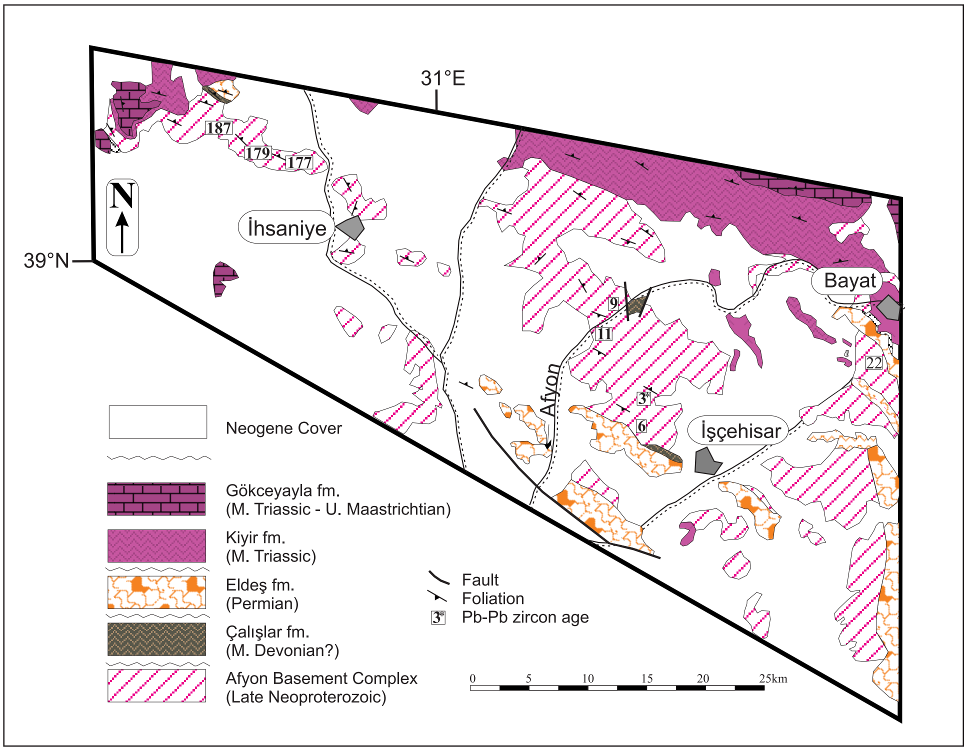

Figure 3- Geological map of the KBB units to the N of Afyon (simplified after Gürsu and Göncüoðlu, 2007).

ritoid schists occur as thin layers. Ýhsaniye Metamorphic Complex, likewise the Göktepe Metamorphics, includes basic green schist intercalations of basic origin. These metamorphics of the earliest phase (Cadomian?) are overprinted by Na-amphibole-bearing phases which are products of Alpine HP/LT metamorphism (Özcan et al., 1989; Candan et al., 2005). Ýhsaniye Metamorphic Complex is unconformably overlain by Eldeþ formation, or by the Lower Triassic Kýyýr formation. Eldeþ formation starts with quartz conglomerates and passes into fossiliferuos Permian limestones. Kýyýr formation includes red colored conglomerates (Özcan et al., 1989).

Afyon Basement Complex outcrops on first major tectonic slice that placed over non-metamorphic Tauride type units. Outcrops of this unit can be followed from Aslanapa to Bolvadin. The lowermost part of the unit comprises mica schists and metafelsic rocks as in Sömdiken and Kütahya areas (Göncüoðlu et al., 2001; Turhan et al., 2003, 2004; Candan et al., 2005, Gürsu and Göncüoðlu, 2008). They include pre-Alpine (Cadomian) paragenesis in micaschists; garnet, biotite, muscovite. Felsic magmatic rocks that have undergone deformation and metamorphism together with micaschists, have rhyodacite-dacite composition and show blastoporphyritic texture. Zircons obtained from these felsic rocks are dated by single zircon evaporation method. The age of this intrusion is 541 + / - 4 Ma (Gürsu and Göncüoðlu, 2008). This data shows that the basement of KBB is of Late Neoproterozoic age.

These Cadomian magmatics are characteristic for north of Gondwana. They constitute the basement of not only KBB, and also Taurides (Erdoðan et al., 2004, Gürsu and Göncüoðlu, 2005, 2006*a*, 2008), Menderes Core Complex

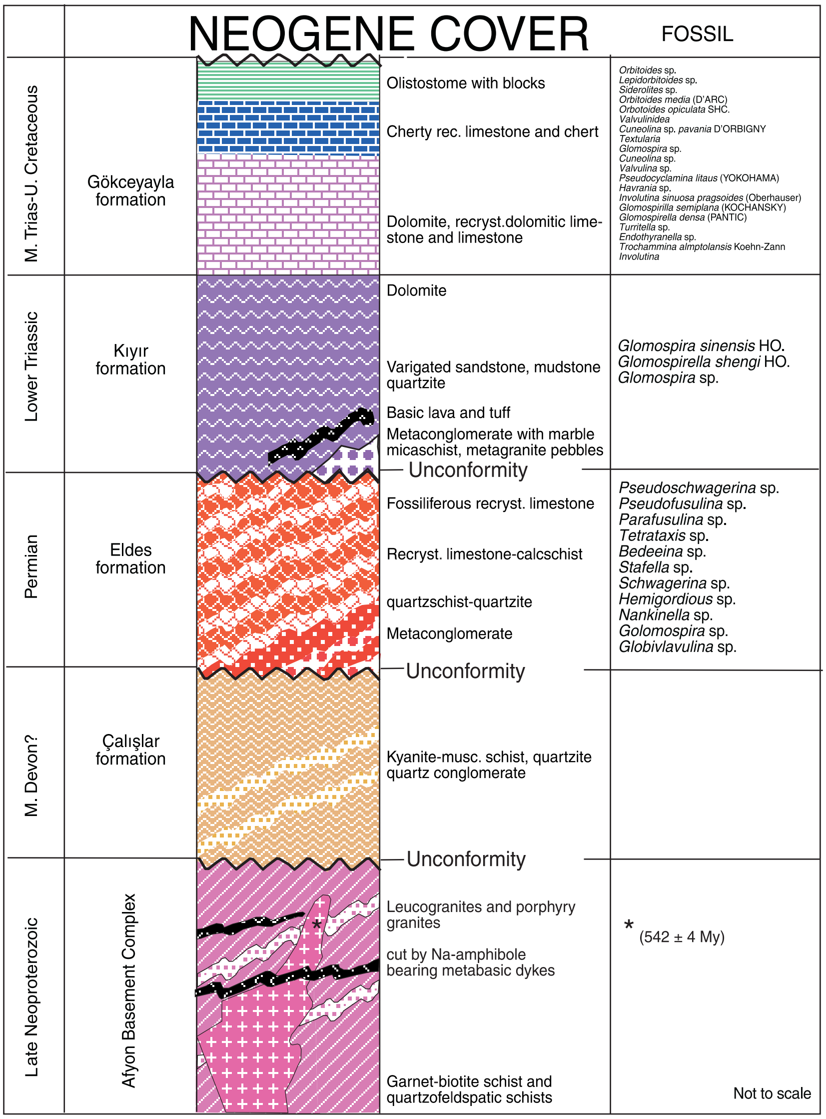

Figure 4- Generalized columnar section of the KBB units in the N of (simplified after Gürsu and Göncüoðlu, 2007).

(Dora et al., 2001) and Ýstanbul-Zonguldak units (Ustaömer, 1999; Chen et al., 2002).

In the pre-Permian basement of Afyon region, there is one more metamorphic unit that unconformably overlies the Cadomian units. This unit is named as Çalýþlar formation by Gürsu et al., (2004). It is composed of quartz conglomerates including deformed granite and schist pebbles, quartzite and quartz mica schists and it transgressively overlies Late Neoproterozoic basement. Locally this non fossiliferuos unit has 250 m thickness. It is correlated with Devonian quartzites in Sultandað (Gürsu and Göncüoðlu, 2008).

Konya-type Palaeozoic Basement

In the region from North of Konya (Figure 5) to Kulu area, pre-Permian low-grade metamorphosed successions are found in the basement of KBB (Özcan et al., 1987, 1990*a*, 1990*b*; Eren, 1993, 1996; Göncüoðlu et al., 2000*b*, 2007).

This basement outcrops commonly in the tectonic slices in the Konya Bozdað Mountains. In lower most visible part of it, quite thick meta-siliciclastic succession is found (Figure 6). Black colored meta-siltstone, black laminated lydite (Figure 7), dark gray silicified shale / tuff and nodular chert bands are included in the unit. They are cut by diabase and quartz porphyry. In the upper part of the unit there are thin and brown-black colored limestone bands. In the succession 1-2 m thick gray-black nodular chert is found locally.

Göncüoðlu et al. (2000*b*) named this unit unofficially as "Siliciclastic Turbidite Unit". Towards top, unit is composed of thin layered, brown-black limestone bands and 3 m thick massive black chert. Over the cherts, pink colored nodular limestone that belongs to lowermost part of the Bozdað Limestone is found.

Besides the Middle Silurian Muellerisphaerid finding from samples collected by Kozur (1999) from nodular cherts in the middle part of the succession, sample T8-26 from thin limestone layers from top of the unit includes conodonts *Dapsilodus obliquicostatus* (Branson and Mehl), *Panderodus recurvatus* (Rhodes), *Ozarkodina excavate*, *Pseudooneotodus beckmanni* and *Panderodus unicostatus*. By this a Late Silurian age is given to unit. In sample T8-28 (Figure 8) conodonts such as *Coryssognathus dubius* (Rhodes), *Pseudooneotodus bicornis* Drygant, *Dapsilodus obliquicostatus* (Branson and Mehl), *Pseudooneotodus beckmanni* (Bischoff and Sannemann) and *Papinochium* sp. (Müllerispherid) belonging to Ludlow-Pridoli (Late Silurian) are observed. In addition, in sample T8-29 Wenlok-Pridoli (middle-late Silurian) conodonts such as *Dapsilodus obliquicostatus* (Branson and Mehl) and *Pseudooneotodus bicornis* Drygant are determined (by Y. Göncüoðlu and H. Kozur). These findings indicate that the age of the Siliciclastic Turbidite Unit is Middle-Late Silurian.

Bozdað Limestone, 800 m thick, is composed of recrystallized limestone and dolomites. In the lowermost part of the unit, pink colored nodular limestone is observed. After non-fossilliferuos black-white colored, thin-medium layered section that composed of dolomites, a band with 3-8 cm long nautiloid (Orthoceras) and crinoids is found. Black colored massive- thick bedded limestone with Amphiphora limestone and dolomites, constitutes the main body of the Bozdað Limestone. Especially in the northern part of Bozdað Massive, limestones are cut by diabase dikes with NE- SW extension.

Pink- black colored nodular limestones in the lower-most part of the unit, includes conodonts such as *Ancyrodelloides kutscheri* Bischoff and Sannemann, Icriodus sp., *Panderodus unicostatus* and Ozarkodina sp. (Figure 9) that belong to A. Delta Zone of upper Lochkovian (Early Devonian). Also in nautiloid limestone, Early Devonian conodonts such as *Panderous unicostatus*, *Ozarkkodina excavate* and *Oulodus* sp. that gives Lochkovian-Pragian (Early Devonian) are found (determined by Y. Göncüoðlu). In Amphiphora carbonates in the middle and upper part of the succession, rare solitare corals are observed. These limestones constitute the common rock type of Middle Devonian in Taurides. In massif limestones in the upper part of the formation no fossils are determined yet.

Over the Bozdað Limestone, Halýcý mélange commences with a sedimentary contact. This unit sometimes overlies karstic, cavity filling coarse siliciclastic, and mudstone and limestones. It includes olistolithes of various magnitude and olistostromal conglomerates which are products of mass flow, in the fine grained matrix that is transformed into greywacke and slate (Figure 10).

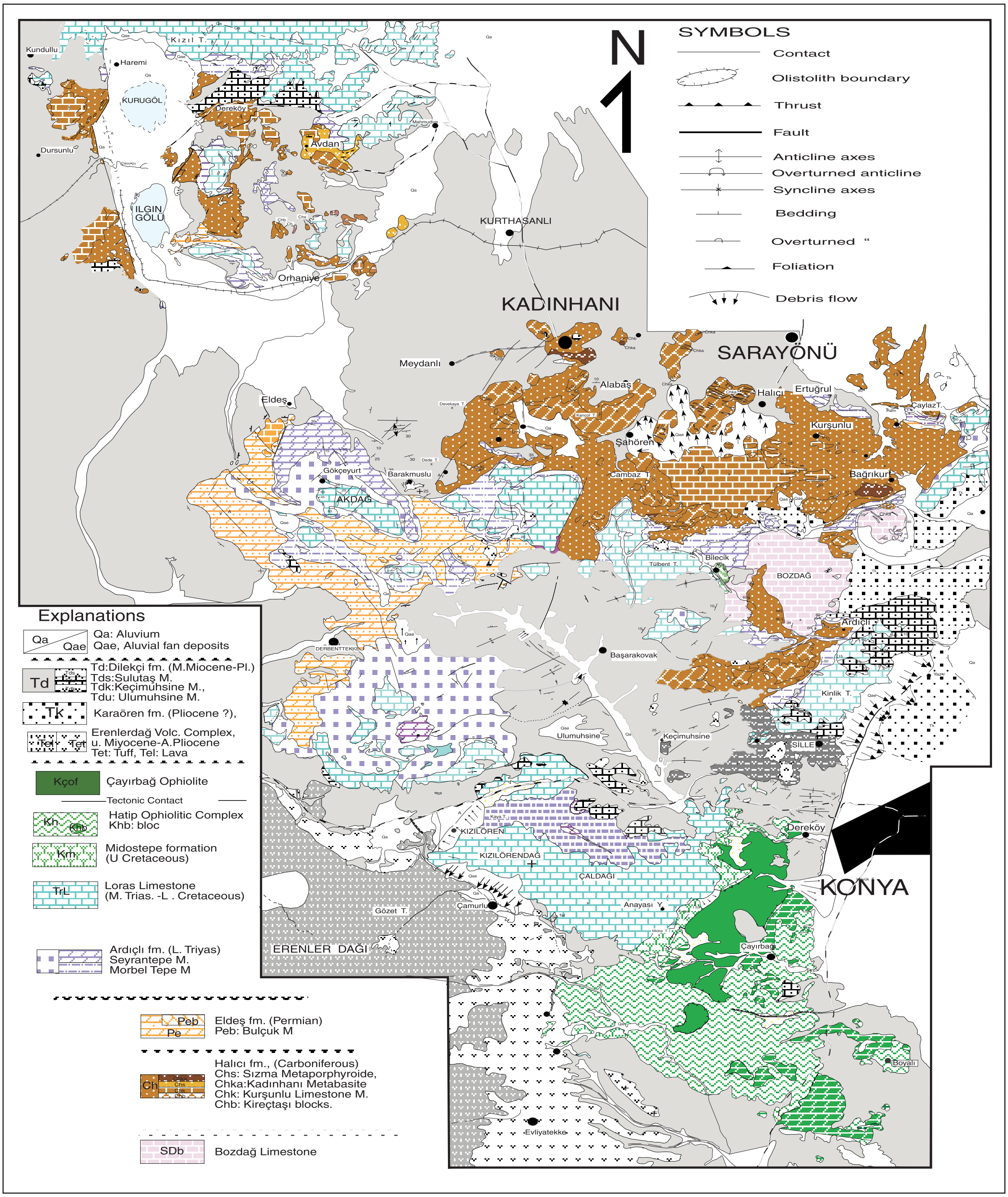

Figure 5- Distribution of the main geological units in Konya and surrounding regions (Özcan et al.,1990*a*).

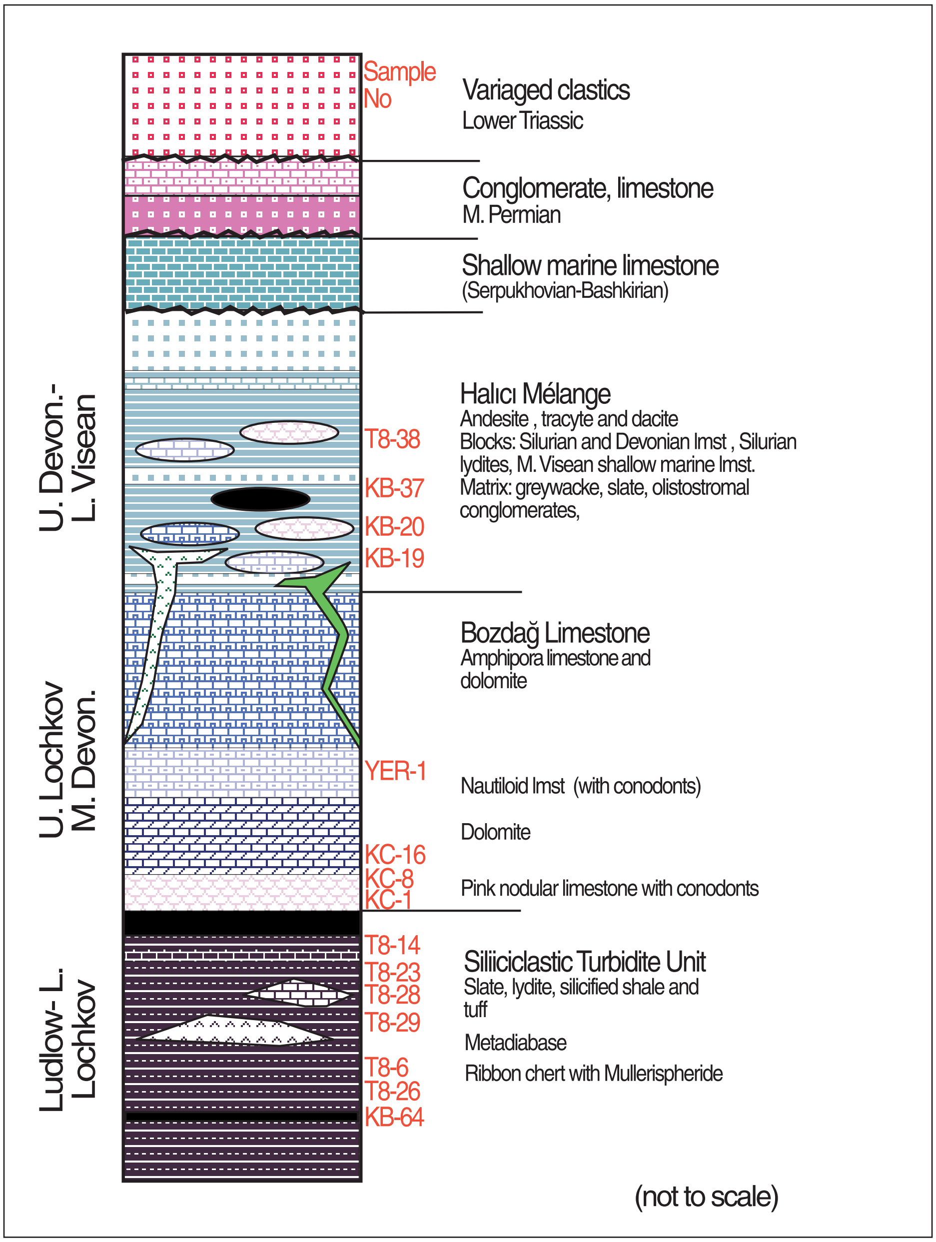

Figure 6- Stratigraphy of Konya-type basement rocks in KBB (simplified after Göncü-oðlu et al., 2007).

In the unit, both siliciclastic turbidite unit of Silurian age and blocks of Bozdað limestone are found as olistoliths. Apart from these, black, crinoidal Lower Carboniferous olistolithes are also observed in the unit. In the Halýcý mélange, both syn-sedimentary lava flows and olistolithes are observed. In the parts that olistoliths are not found, unit presents broken formation feature. The volcanic rocks of that unit include trachyandesite (Figure 11) and rhyolites (Bayiç, 1968;

Figure 7- Lydite levels within the Siliciclastic Turbidite Unit to the N of Konya.

Kurt, 1996; Eren et al., 2004; Göncüoðlu et al., 2007) geochemically defined as subalcali basalts. The shallow marine carbonates overlaying the siliciclastics of the unit with a sedimentary contact include Serpukhovian - Bashkirian foraminifers (determined by D. Altýner). In the region, no fossils are found besides given here. The formation is younger than Lower Carboniferous and older than the unconformably overlying late Middle Permian. By considering the geochemical and lithostratigraphic features of the volcanic rocks, it is widely accepted that unit is developed in a Carboniferous back arc basin (Özcan et al., 1990*a*; Göncüoðlu et al., 2007). Nevertheless, there is no consensus on paleogeographic position of the arc (Robertson and Ustaömer, 2009).

KBB units that represent similar features with Konya-type Palaeozoic basement are located in Karaburun Peninsula and western Taurides. One of these, the Karaburun unit, is defined by Erdoðan et al., (1995), Kozur (1998), Robertson and Pickett (2000), Rossalet and Stampfli (2002). In these studies, there is no consensus neither on their lithological features, nor their structural relations and the ages of the units.

It is supposed that Tavas Nappe is one of the Lycian Nappes. It is one of the units that belong to Tauride-Anatolite platform like units of KBB (Þenel et al., 1994). Halýcý mélange like Carboniferous units recently defined in Konya, are named (Kozur et al., 1998; Kozur and Þenel, 1999; Stampfli and Kozur, 2006) as Teke Dere Unit (Figure 12).

Teke Dere Unit or Teke Dere Slice of Collins and Robertson (1999) is actually composed of more than one tectonic slice. From bottom to top the slices include following units (Göncüoðlu et al., 2000*c*):

A- Early Upper Permian limestones vertically transitional to blocky flysch (Triassic),

B- Volcano-sedimentary slice (Figure 13) starts with approximately 20 m thick pillow lava and conglomerate. Most of its pebbles were originated from volcanic rocks. Upwards follow sandstone, 1-2 m thick lava flow, brachiopod and crinoid- rich limestone, sandstone- sandy limestone, tuffite, black shale and mudstone intercalation and ends with gray-beige and pinkish, medium-thick layered vertical cliff forming limestones. The fossils that are detected in this unit were dated as Moscovian-Kasimovian (Upper Carboniferous). Geochemical analysis of alkali basalt constituting vast parts of the unit, of trachyandesite, pillow lava (Figure 14, samples T4A-H in Figure 13) which is originated from trachyte, lava breccia (Figure 14; samples T1A-B in Figure 13) and pebbles. They are forming a co-magmatic sequence and show 'oceanic island basalt (OIB)' feature (Göncüoðlu et al., 2000*c*).

C- Approximately 20 m thick lavas, from time to time with pillow-structure is cut by diabase dikes, intra pillow carbonate filled olivine basalt slice. In this slice, Middle-Late Carboniferous, badly preserved fossils in carbonate and chert intra-pillow fillings are detected. Samples that are taken from this unit display Mid-Oceanic Ridge Basalt (MORB) character (Göncüoðlu et al., 2000*c*).

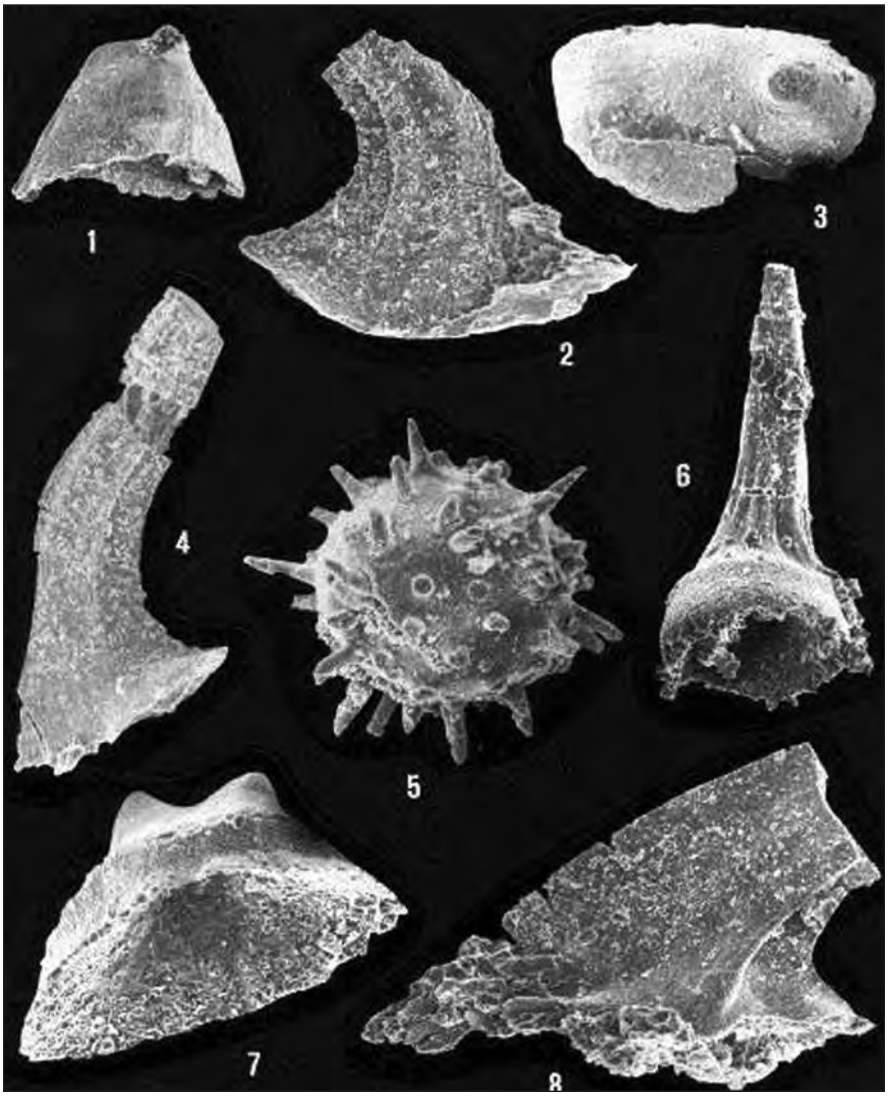

Figure 8- Conodonts and Muellerisperids from the micritic limestone-chert bands of the upper Siliciclastic Turbidite Unit: 1, 7- *Pseudooneotodus bicornis drygant*, 2, 4, 8- *Dapsilodus obliquicostatus* (Branson and Mehl), 3- *Pseudooneotodus beckmanni* (bischoff and sannemann), 5- *Papinochium* sp., 6- *Coryssognathus dubius* (Rhodes).

D- Shallowing upward successions with sandstone- siltstone and shale with the appearance of a flysch is unconformably covered by Upper Permian limestone and variegated terrestrial siliciclastics of Triassic Çenger formation. In the flyschoidal unit that constitutes lower part of the slice, conodonts of Early Carboniferous (Visean) are found by Kozur et al., (1998).

The Carboniferous 'oceanic island' and 'mid-oceanic ridge' type volcanic rocks together with Carboniferous distal flysch within the Tavas

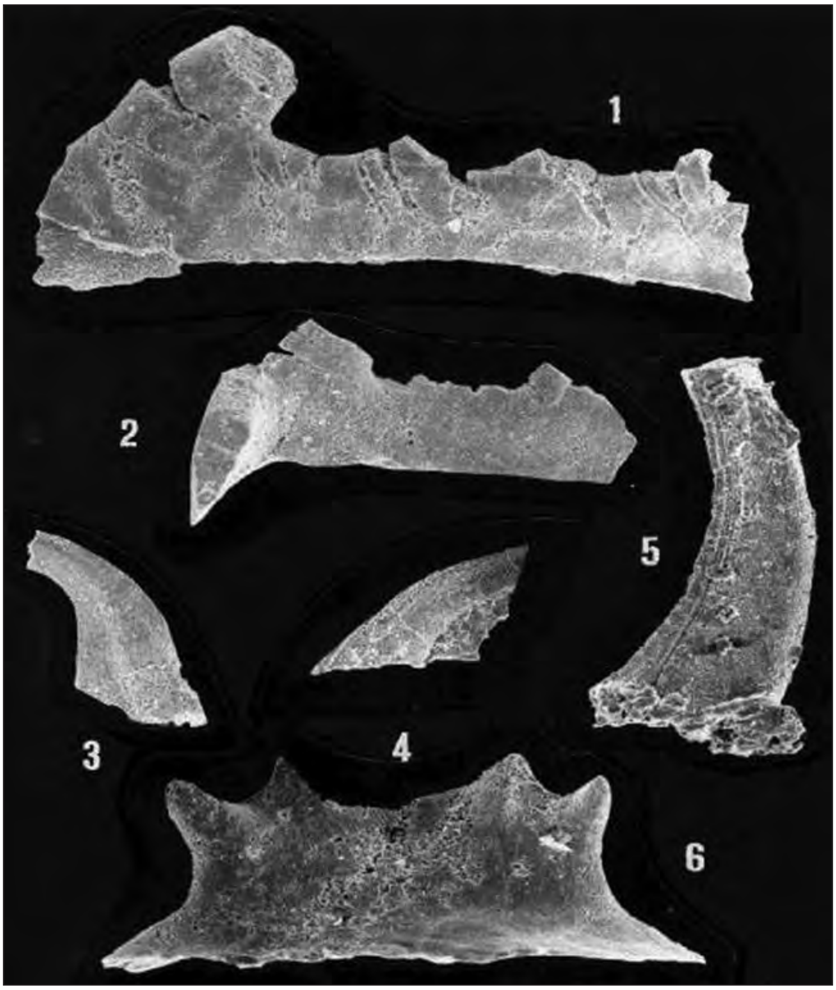

Figure 9- Early Devonian (Lochkovian) conodonts from the lower part of Bozdađ Limestone: KC 8A 1: *Ozarkodina* sp., Sc element, 2: *Ozarkodina* sp., Sc element, 5: *Panderodus unicostatus* (Branson and Mehl), 6:(KC 8C): *juvenil Icriodus* ? sp. Age: Silurian - Early Devonian. (KC 1) 3, 4: *Panderodus* ? sp., Age: A. delta Zone, Upper Lochkovian.

Nappe are evaluated together with the coeval back arc basin units found in Konya. It is claimed that these units are related with the Variscan event that occurred in pre-Permian at the Northern margin of Palaezoic Tauride-Anatolite platform (Göncüoðlu, 1989).

Middle Permian Cover

All along the KBB, middle Permian units overlie the older units defined above with an angular unconformity. Considering Tauride Units that are defined by Özgül (1976) this unconformity is the

Figure 10- Crinoidal limestone blocks within the olistostromal matrix of the Halýcý Melange to the N of Konya-Ardýçlý.

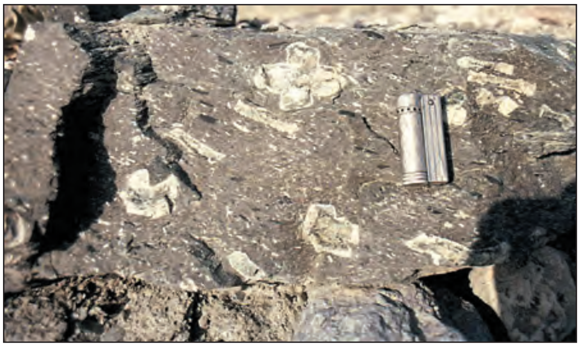

Figure 11- Field occurence of Sýzma Metaporphyroid with large sanisine crysts and well-developed trachyitic texture.

characteristic feature. Early Upper Permian is a common unit that overlies various units in different regions of KBB by transgression. This transgressive unit in some places (ex. North of Afyon) overly a variably eroded substratum that may reach down to the Neoproterozoic basement. This erosion is related with a rapid rising phase of Tauride-Anatolite Platform before Upper Permian and it is evaluated as representative of the Variscan event (Özcan et al., 1989).

In the bottom of the Permian successions always quartz rich conglomerates or white, cream or black colored quartzites with quartz-pebbles are observed. Pebbles are mostly composed of well rounded quartz, quartzite, mica schist, meta-quartz porphyry grains. The succession continues with light gray, greenish gray, beige-light brown colored quartzite that shows lamination, and cross bedding. Towards top it is composed of calcschists and medium-thin layered sugary textured recrystallized limestone bands. Upper part of the succession is represented by medium-thick-bedded, gray, white and black colored, crinoid, *Mizzia* and fusulinid- bearing limestones. Slices in the north only include deformed fusulinids and *Mizzia* where the internal structures are erased because of recrystallization. On the other hand in the south Konya region (Figure 15) *Tetrataxis* sp., *Staffella* sp., *Hemi-gordius* sp., *Nankinella* sp., *Globivalvulina* sp., *Verbeekina* sp., *Neoschwagerina* sp., *Kahlerina* sp., and algae (*Pseudovermiporella* sp.) are determined. *Verbeekina* sp., *Neoschwagerina* sp., and *Kahlerina* sp, are of Wordian-Capitanian (Guadalupian) age. By referring these data it is claimed that the unit is deposited in late Middle Permian (revisied data from Özcan et al., 1989, 1990*a* by Dr. C. Okuyucu; Eren, 1993; Göncüoðlu et al., 2003, 2007)

In the NE of Konya the Permian unit has a special position in KBB; it is determined as one of the slices of KBB by Özcan et al., (1989), on the other hand Eren (1993) determined it as Gökçeyurt Group. It exhibits similar metamorphic features to KBB but differs from them in its stratigraphy. This difference is observed especially in lithostratigraphy and relations of Permian and Triassic successions. In typical KBB successions, between Permian carbonates and Triassic continental successions there is an important

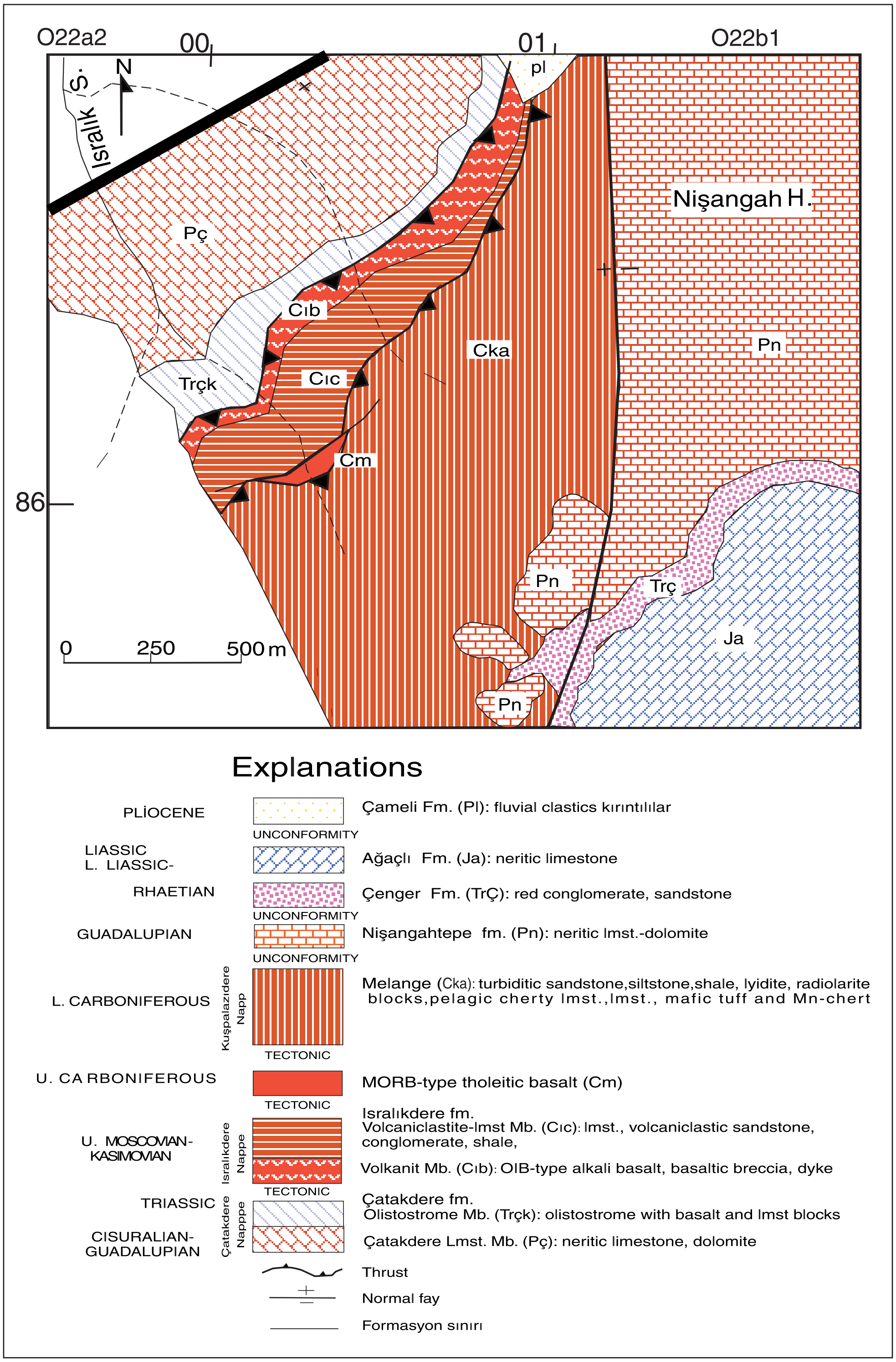

Figure 12- Geological map of the Teke Dere slice of the Tavas Nappe in KBB (Göncüoðlu et al., 2000*d*).

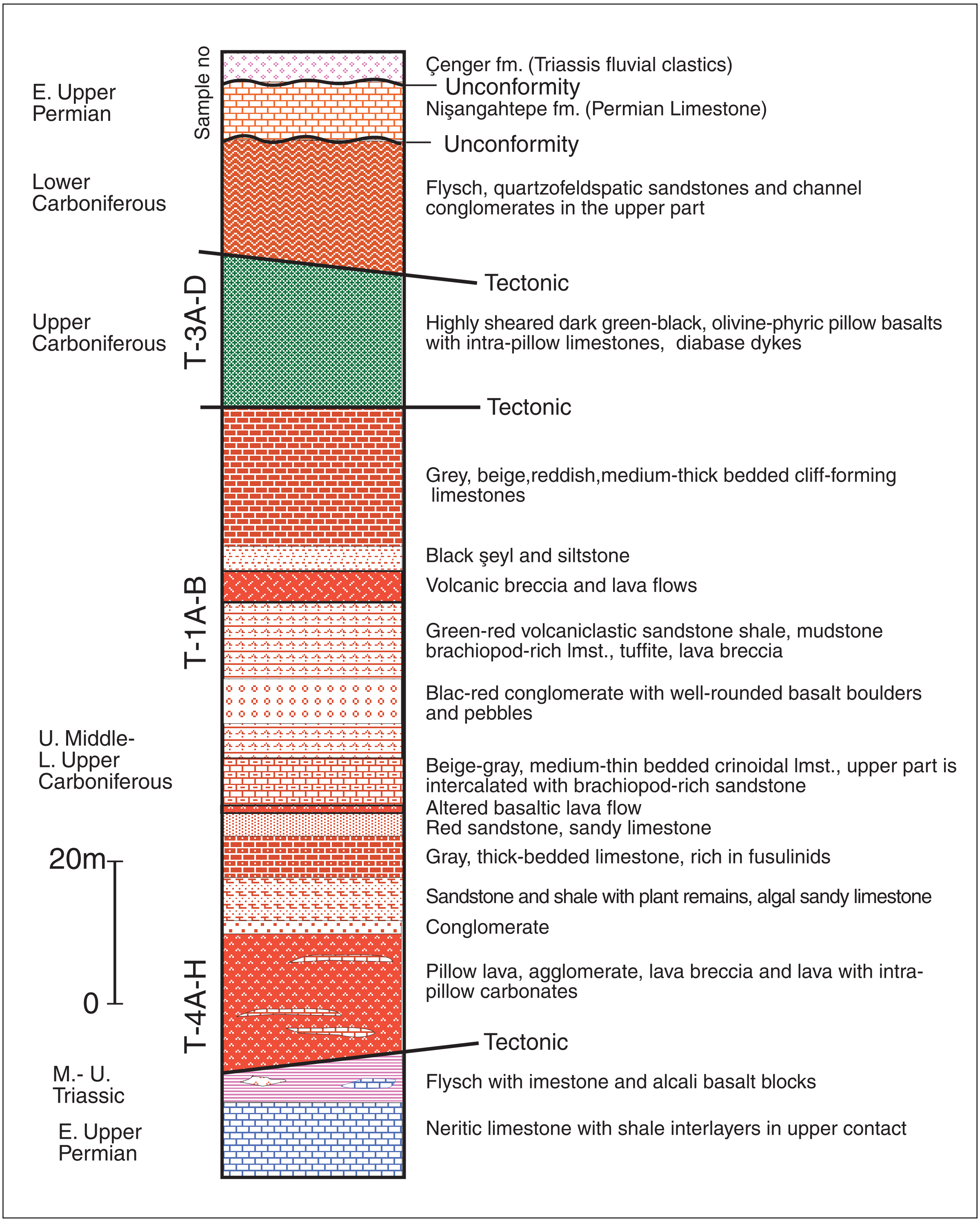

Figure 13- Generalized stratigraphic sections and fossil locations of Teke Dere slice of the Tavas Nappe.

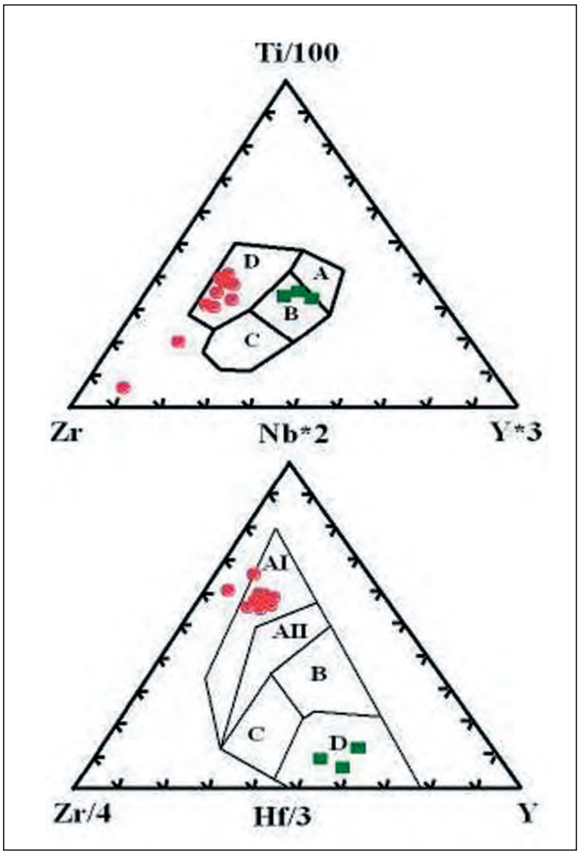

Figure 14- Tectonomagmatic classification of the volcanic rocks in different slices of the Tavas Nappe in Teke Dere. Circles are OIB lavas that alternate with fusulinid limestones, squers represent MORB lavas from olivine basalt slices (Göncüoðlu et al., 2000*d*).

unconformity but in this area relation of these two units are conformable, as in Aladað Unit of Özgül (1976). Because of that, interpretation (Eren, 1993, 1996) that Gökçeyurt Group belongs to another Tauride Unit is more realistic.

MESOZOIC PLATFORM SUCCESSIONS

Mesozoic successions are the most fundamental units of KBB. They overlie different type basements and can be correlated easily in every slice. In these successions from bottom to top pretty thick sediments are deposited. Lower Triassic starts with continental clastics and passes to platform carbonates in Middle Triassic and Jurassic, in different slices between end of Jurassic and Early Cretaceous successions are transitional to slope-type pelagic sediments and followed in Late Cretaceous by thick, ophiolite-bearing flysch-type sediments (Figure 16).

Lower Triassic Terrigenous Units

Lower Triassic rocks that are characteristic with their variegated colors in all tectonic slices of KBB are named as Kýyýr formation (Özcan et al., 1989; 1990*a, b*; Göncüoðlu et al., 1992) in Kütahya region (Figure 17), Ardýçlý formation (Özcan et al., 1990*a*; Göncüoðlu et al., 2003) in Konya region and, Otluk Metaclastite (Göncüoðlu et al., 1996, 2000*a*) in Central Sakarya region (Figure 18).

Name of the Kýyýr formation will be used in terms of priority all along the belt. The formation includes Morbel Tepe and Seyrantepe members which are commonly observed in Konya region (Özcan et al., 1990*b*, 1992).

Morbel Tepe member is composed of red, purple, pink, violet conglomerate, sandstone and mudstone. In the unit, rarely dirty yellow colored dolomitic interbeds are noticed. Places where Morbel Tepe member reaches its maximum thickness are: SW of Kütahya Kocadere, and NE of Afyon Kýyýrderesi. The reference sections of this unit are located in Afyon-Altýntaþ antique marble quarry exposures, north of Sevdiðin Village on the southern slope of Kulaksýz Mountain and Meydan Village located 32 km NNE of Konya. Morbel Tepe member overlies pre-Triassic units with an angular unconformity. In the lower part of the unit, the thickness of the basement conglomerate from time to time exceeds 100 meters. It includes red-brown colored, grain supported, subrounded- rounded-angular pebbles.

It is composed of pebbles from the Neoproterozoic basement and boulders of the metase-

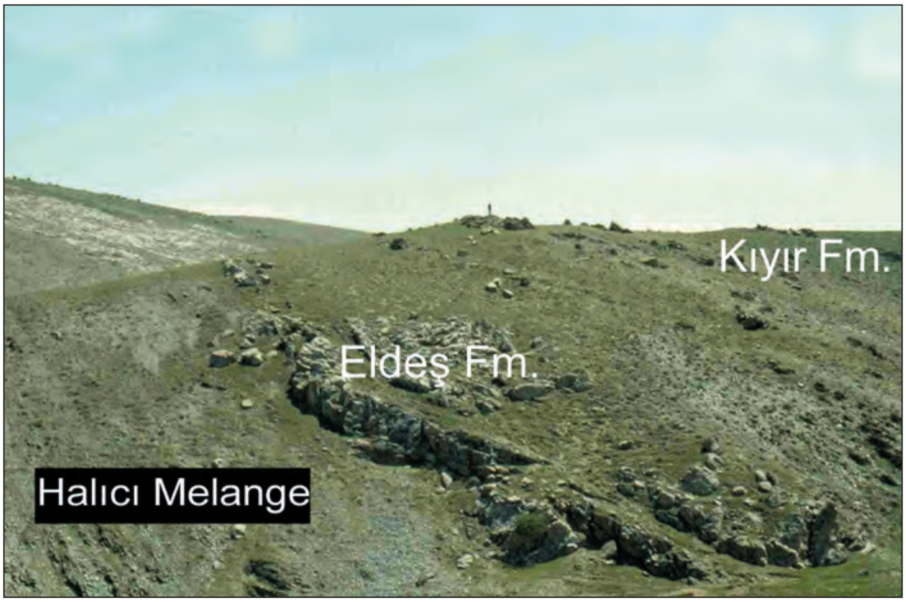

Figure 15- View showing the Permian and Early Triassic unconformities in Çaylaz Dere NE Konya.

dimentary (quartz-mica schist, chlorite-muscovite schist) and metaigneous (meta-quartz porphyry, metarhyolite, metabasic) basement rocks. Black lydite, sandstone, quartzite, *Mizzia*-bearing recrystallized limestone clasts are considered as transported from the Palaeozoic basement (Figure 19). In a marble quarry in Altýntaþ, pockets and 1 m thick vertical fillings are observed. These pockets are formed by the same material inside the paleokarstic cavities of Permian limestones that located in the basement of the unit. Same images are also seen in south of Ýþçehisar, where Permian recrystallized limestone is excavated under the name "Afyon marble".

Red, green, gray colored quartzarenite as intercalations with conglomerate and some mottled mudstone are seen. Over this unit, fining upward, cyclic, brown-red-purple colored, medium-coarse grained, laminated sandstone-siltstone-mudstone packages are placed. The top of each cycle, mud cracked and bioturbated purple mudstone intervals are characteristic. In the uppermost part of the unit, greenish-brown and pinkish gray colored dolomites, dolomitic limestones, and oolitic limestones are found as discontinuous bands and lenses. They are oosparites and oomicrites and include undetermined bivalves.

By considering its depositional characteristic and the absence of marine fossils, it is claimed that lower and middle part of this member are formed in fluvial environment (Özcan et al., 1984). In more detail, the conglomerate dominated lower part formed in proximal alluvial fan and flood plain, whereas intercalated variegated siliciclastics represent meandering river sediments alternating with flood plane and beach sediments. The upper part of the unit represents subtidal and intertidal environment in oscillating platform margin. It is also claimed that dolomitic and oolitic limestones are sabhka sediments (Göncüoðlu et al., 2003).

Five km north of Afyon-Altýntaþ between Ýncebel Tepe and Obruk Tepe, first limestone bands overlying the varigated clastics include *Glomospira sinensis* HO and *Glomospirella*

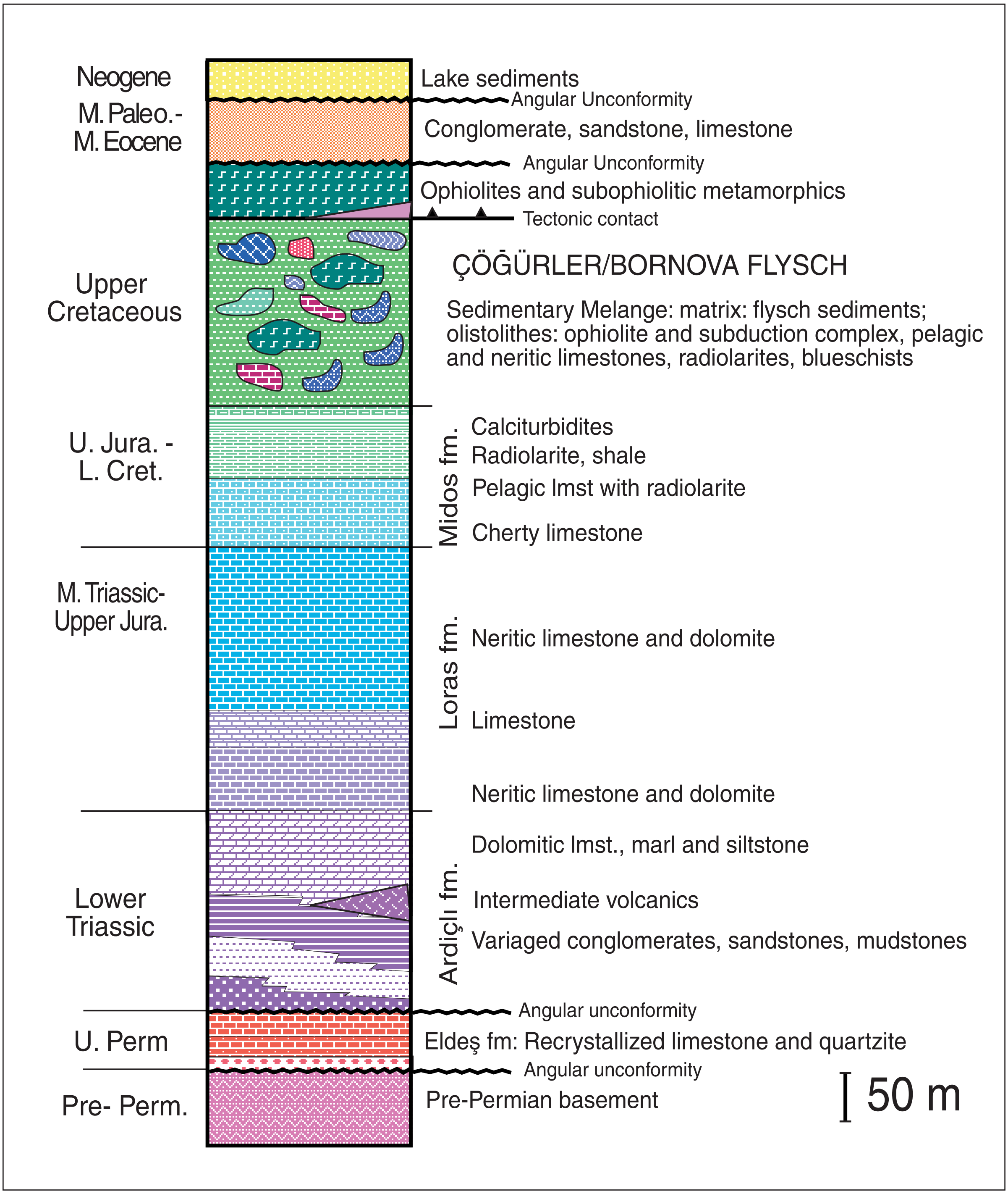

Figure 16- Generalized Mesozoic columnar sections in KBB (Göncüoðlu et al., 2002).

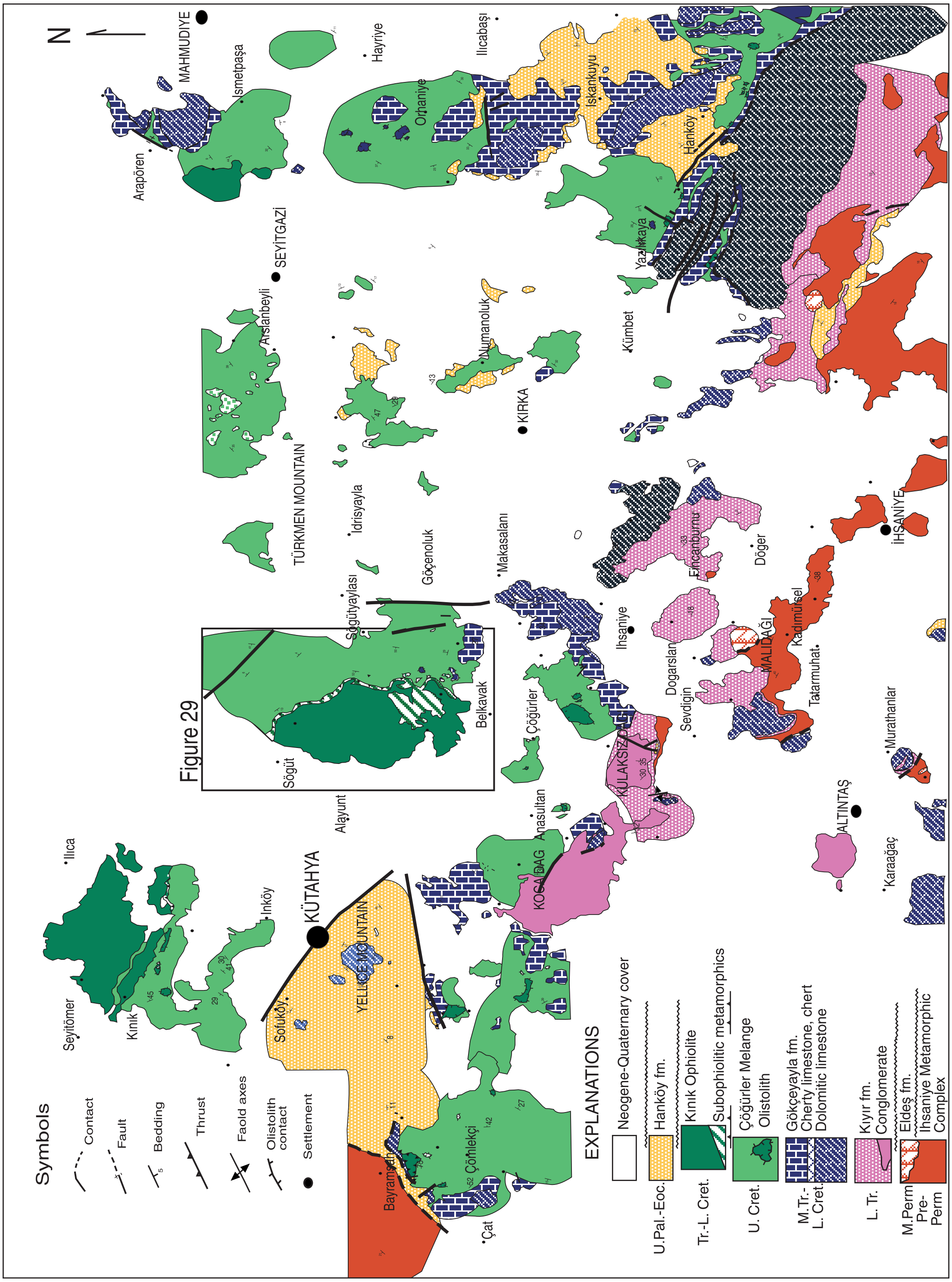

Figure 17- Geological map of Kütahya area (simplified after Özcan et al., 1989). Inserted field shows the locality of figure 29.

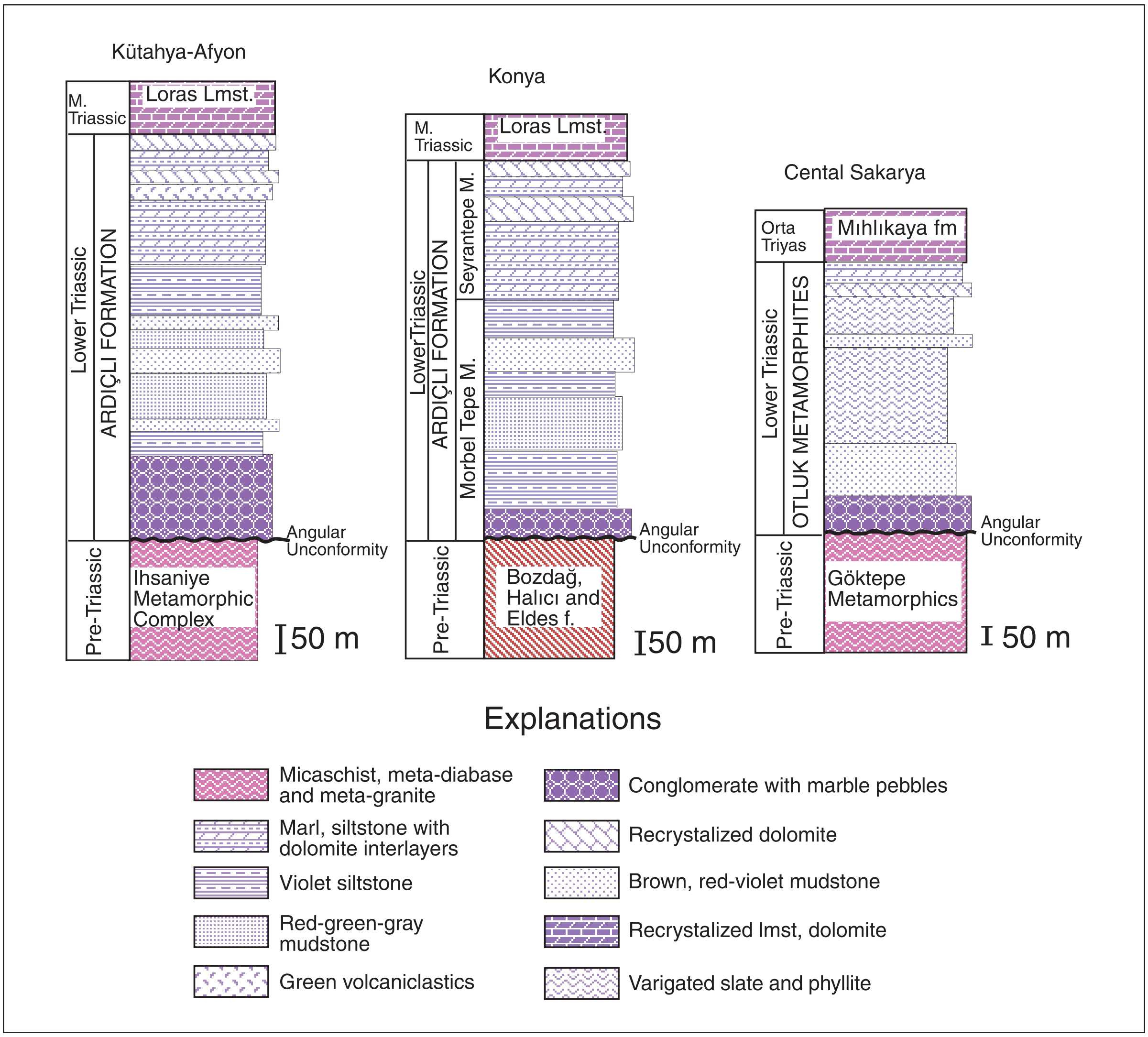

Figure 18- Lithology and correlation of the Triassic rock-units in KBB (Göncüoðlu et al., 2003).

shengi HO. In the section between Tepecik Hill-Sýzma, located 15 km NNW of Konya, first limestone bands and lenses in the outskirts of Gökçeyayla Village, 13 km SW of Konya-Ilgýn also include *Glomospira sinensis* HO, *Glomospirella shengi* HO, *Glomospira* sp., *Meandrospira pusilla* HO, *Nodosinella* sp., *Earlandia* sp., algae and gastropods. According to these data, Mobel Tepe member was deposited in Early Triassic (Induan- Olenekian) and before. By considering sedimentary features of the unit, conglomeratic levels present fluvial facies and sandy, silty, muddy sections present meandering river facies. Banded and lensoidal mudstone-siltstone levels present flood plain and coastal beach. Towards the top of the unit, laminations, cross laminations, wavy lamination and bioturbation indicate intratidal sediments, and shallow marine affect. In general terms, this unit presents an environment that starts with fluvial and passes to shallow marine conditions.

Figure 19- Polymictic basal conglomerates of Kýyýr formation.

The upper member of the formation, composed of carbonates, is named as the Seyrantepe member (Özcan et al., 1990*a*). This member from bottom to top includes greenish brown colored, medium-thick bedded, dolomitic limestone, oolitic limestone, dolomite, rarely marn, calcarenite, and siltstone intercalations (Figure 20). In these rocks recrystallization is common. Around Ýþçehisar -Afyon in between and under the limestones, altered lava flows and volcanoclastics as discontinuous bands are observed together with mudstones. Thickness of this unit varies from 10 m to 200 m. It is vertically and laterally transitional with Morbel Tepe member. These volcanic rocks are attributed to Jurassic by Candan et al. (2005) without any evidence. Petrographically limestones are defined as bioclastic grainstone and oolitic grainstone and it is claimed that they are deposited between tidal barriers (Wilson, 1975) in shelf margin (Özcan et al., 1990*a*). Towards top, where thick carbonates dominate, bioturbated bioclastic wackestone represents transition to marginal platform facies belt. Fossils obtained from this part indicate that its age is Lower Triassic as the Morbel Tepe member. Seyrantepe member passes to the thick carbonates of Middle Triassic (Loras formation) through the top of the section.

Ardýçlý formation, which is defined in Konya region, is affected less by metamorphism than the KBB units in the north. In this area KBB successions in N and NE of Konya, display a distinct Lower Triassic disconformity. On the other hand, in Aladað area (NW of Konya) more different Lower Triassic succession crop out. This unit sometimes has been correlated with the Halýcý complex by mistake (Eren, 1993). However, its Triassic age is provided by foraminifera and conodonts.

In Central Sakarya region where northernmost outcrops of KBB are seen, metamorphic equivalents of the Kýyýr formation are named as Otluk metaclastite (Göncüoðlu et al., 1996, 2000*a*). The basement of this formation starts with red-purple colored, thick bedded metaconglomerates. It is composed of orthogneiss, micaschist and marble pebbles. Through top, after variegated metaclastic intercalations, recrystallized carbonates comprise the dominant rock type. This part of the formation reaches to 160 m thickness and it is overlain by Loras-type massive limestones.

Around Konya, in the N and NE of Bozdaðlar massive, typical successions of KBB which are defined in Afyon and Kütahya regions crop out commonly. In Aladaðlar, NW of the massive, more different successions are observed (Eren, 1993). In these successions, in contrast with other slices of KBB, the distinctive Lower Triassic angular unconformity is not observed. In this area, over the carbonates of Eldeþ formation, after a hardground section, a succession starting with oolitic and algal limestone and passing into red-purple-green colored siltstone and mudstone, are observed. Towards top, this succession is overlain by cream- gray colored medium bedded, Middle Triassic (Anisian)-Jurassic limestones (Eren, 1993, 1996). Most significant feature of this unit is the presence of algal and oolitic limestones in the lower part of the Triassic units. In this unit, also in contrast to other KBB units, no volcanic-volcanoclastic and olistostromal Carboniferous; but shales, quartz-arenites and *Girvanella*-bearing limestones are found

Figure 20- Transition of the Kýyýr formation into Gökçeyayla (Loras) formation in Afyon Gökçeyayla valley.

below the Permian quartzites. These successions are found also in NW of Konya and North of Ilgýn and corresponds to Özgül (1976)'s Aladað unit as defined by Eren (1993) correctly. This Aladað-type unit includes Lower Triassic mottled siliciclastics, dolomite and dolomitic limestones of various thicknesses, and volcanic-volcanoclastic (Akal et al., 2003) intercalations. In NW of Konya (Uzunyayla Stream), limestones have blocky appearance because of folding and rupture. In these limestones conodonts, such as *Neogondolella balcanica*, *N. oerti* and *N. polignathiformis* are detected and it is evidenced that the carbonate sedimentation reaches to Middle Triassic.

It is evident that the KBB contains not only Bolkar Daðý-type rocks (sensu Özgül, 1976), but includes also tectonic slices which were deposited in the more internal parts (e.g. Göncüoðlu et al., 2007) of the Tauride-Anatolite platform (e.g. Aladað unit, sensu Özgül, 1976).

MIDDLE TRIASSIC-LOWER CRETACEOUS PLATFORM CARBONATES

The most distinctive unit of KBB is platform carbonates that cover a large portion of Mesozoic. In Kütahya region this unit is described as Gökçeyayla formation also covering the Midos Tepe formation which bears slope sediments (Özcan et al., 1989). In the Central Sakarya region, recrystallized limestones are named as Mýhlýkaya metacarbonate (Göncüoðlu et al., 1996). They have the same lithostratigraphic position as the Gökçeyayla formation.

Limestones are the most dominant rock type in the unit. In the lower part of the succession, the formation is in transition with dolomitic limestones, and it is composed of gray colored, thin medium bedded micrites. Just above these, middle-thick bedded, gray-beige colored, algal and oolitic, pelecypod shell bearing limestone is observed. Overall in KBB, the carbonates are dolomitized and intensively affected by deformation and recrystallization. Because of this deformation and tectonic slicing the observed thicknesses of the unit varies between 200-700 m in different parts of the belt. The type section of the unit is on the road between Bayat and Gökçeyayla to the E of Afyon. Reference sections are in Loras Mountain, Konya; West of Kütahya in Kocasu Stream and in the West of Bayat.

In terms of microfacies, the lower part of the unit is protected from dolomitization. It is composed of ostracod and algae bearing, pelletoidal limy mudstone and stomatoporitic limy wackestone. According to Wilson (1975) these are deposited in low circulated, shallow marine conditions. They are overlain by megalodont-bearing micritic packstone. It shows rarely dolomitic and fine-coarse intraclastic wackestone-packstone-pelletic limestone-mudstone characteristics and may have been deposited in low circulated, shallow marine conditions with limited faunal development. Above these formations, the succession continues with packstone-boundstone-grainstone-algal- pelletic packstone and dolomitic limestone. This part represents back reef sedimentation in limited shallow marine environment. In North of Altýntaþ, just above this unit mass flow-type, purple colored micrites including intraformational conglomerate levels are observed. They are late Ladinian-early Carnian in age. (Kaya et al.,1995). Over these levels, again pelletic grainstone, calcisphere limestone, algal dolomite, crystalline limestone is observed. It is interpreted that this part deposited in restricted marine facies belt of Wilson (1975). After this level the affect of dolomitization increases. In rare isolated areas, presence of fine intraclastic dolomite and dolostone indicates restricted shelf belt deposition. As approaching uppermost levels of the unit, chert bands (Figure 21), micrite pockets and shale laminations increase.

In summary, it is suggested that the unit started with tidal environment conditions. It

Figure 21- Occurence of the cherty limestones of Loras formation.

progressively changed to restricted shelf and finally to open shelf conditions (Özcan et al., 1989). Thereafter, deposition of siliciclastics, pink, violet and green micrite, radiolarian micrite and chert are increased and the unit passes into slope sediments, as described below.

The age of the unit is introduced by the help of foraminifers (determined by Ahmet Iþýk and Ayþe Turþucu) and conodonts (determined by Asuman Keskin and Heinz Kozur). Samples are taken from various sections and points from West of Kütahya to East of Konya (Göncüoðlu et al., 1992). According to these data the age of the succession starts with Anisian. In various part of the KBB the age ends with Late Jurassic and Early Cretaceous. Within the succession each stage starting from Anisian to Malm are determined. After this interval, because of heavy dolomitization the age determination could not be done properly. However, any findings that may correspond to a considerable hiatus or erosion could not be observed in this formation.

Middle Triassic - Lower Cretaceous platform carbonates constitutes the basement of unit which is named as Bornova Mélange (Erdoðan, 1990) or Bornova Flysch Zone (Konuk, 1977; Okay and Siyako, 1993, Okay et al., 1996). This unit is located at NW of the Menderes Core Complex (Figure 22).

In Manisa - Akhisar and Simav regions as seen in generalized stratigraphic section (Figure 23) Akdeniz and Konak (1979), Akdeniz et al. (1980), Akdeniz (1985) the thick Mesozoic carbonate succession including Gökbel, Hasköy, Kocakýr, and Görenez formations equalizes the carbonates of KBB. Lower Cretaceous recrystallized neritic limestones of Görenez formation in the uppermost part of the unit is transitional first to slope-type pelagic limestone and chert and then to the turbiditic successions of "Bornova Flysch" (Yalýnýz et al., 2005; Tekin et al., 2006), as in the Kütahya region.

UPPER CRETACEOUS SLOPE SEDIMENTS

Successions that transitionally overlie the platform carbonates of KBB are named as Midos Tepe formation in Konya region. The formation is accepted as a member of Gökçeyayla formation in Kütahya-Afyon region and Mýhlýkaya formation in Central Sakarya. It starts with reddish-pink

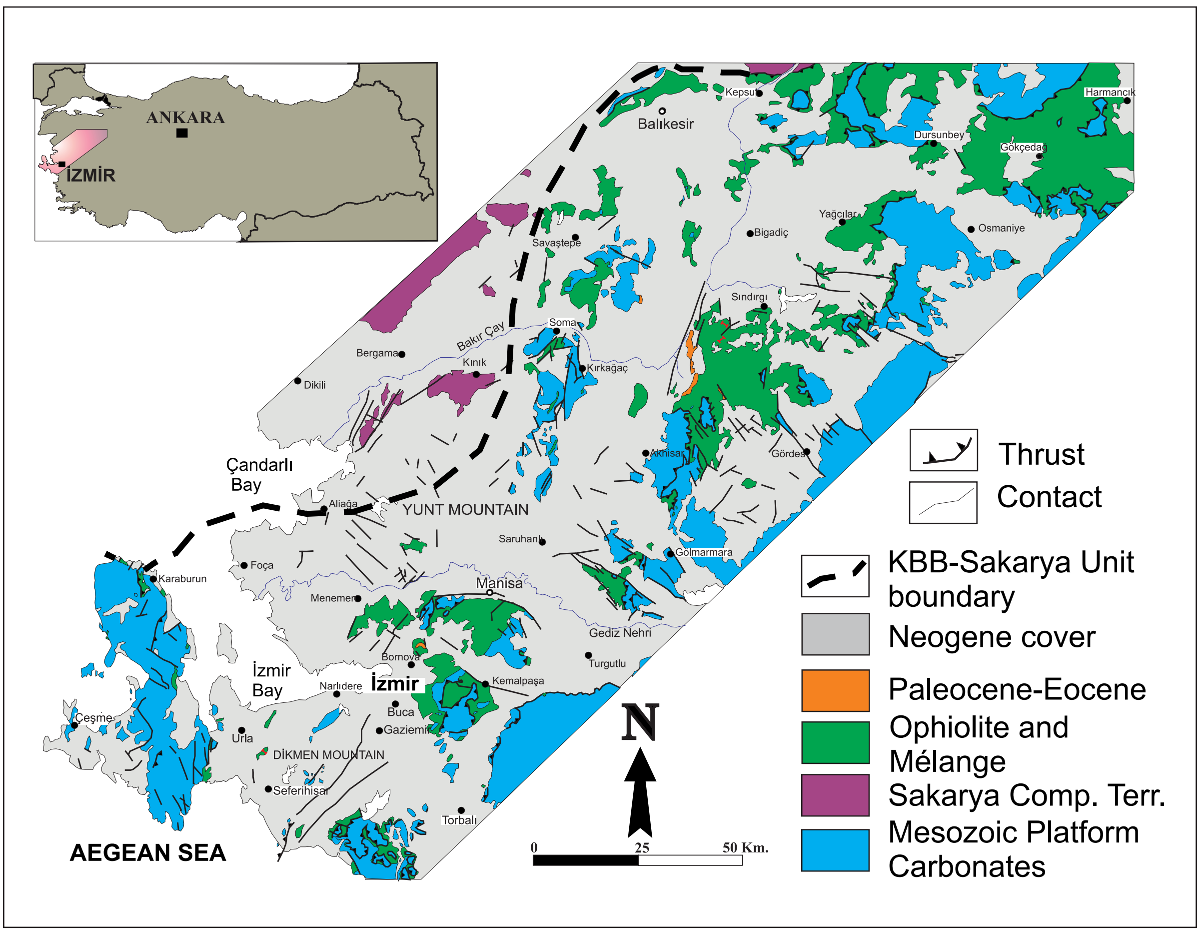

Figure 22- Simplified geological map of Bornova Flysch Zone and sample locations. (a) Paleozoic-Mesozoic carbonates of the Tauride-Anatolite Platform, (b) Sakarya Composite Terrane, (c) Ophiolite and ophiolitic melange, (d) Eocene carbonates and clastics, (e) Post-Eocene units, (f) Boundary between Sakarya and Tauride-Anatolide units, (g) Normal contact, (h) Fault, (i) Thrust, (j) Rivers, (k) Localities of studied sections, (simplified after MTA, 2002).

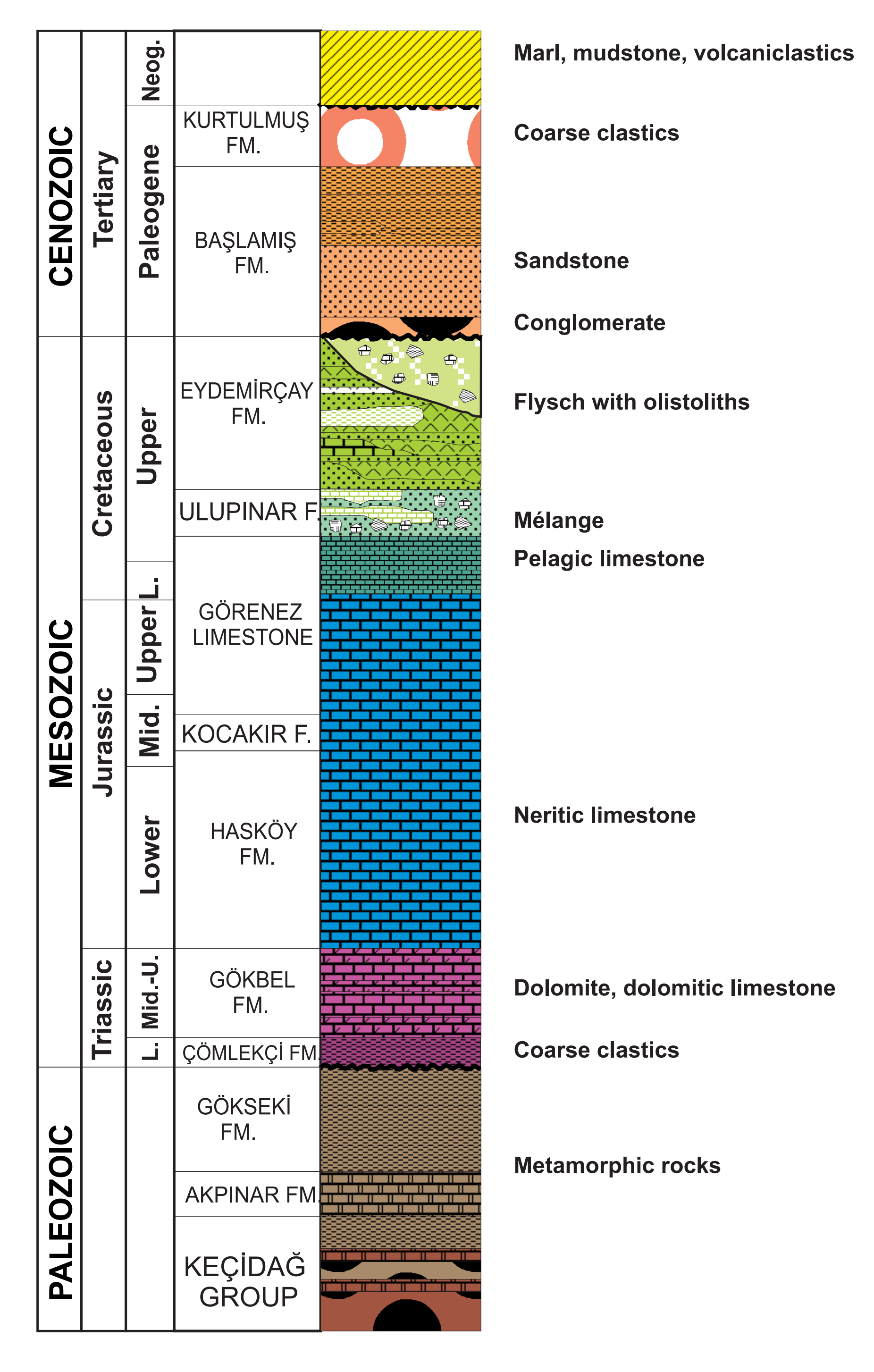

Figure 23- Generalized stratigraphic section of Bornova Flysch Zone (after Akdeniz et al., 1980; Akdeniz, 1985).

colored, thin bedded and lensoidal semi-pelagic limestones with chert nodules. From time to time it continues with an intercalation of chert-micritic limestone and pelagic fossil bearing red mudstone-radiolarite. In Bayat region the unit starts with 5 m thick quartzarenite and continues with shale-micrite-chert alternation. In this zone radial/prismatic calcite after aragonite is a typical feature of carbonates (Özcan et al., 1989; Candan et al., 2005) which were generated because of HP/LT metamorphism. In Tavþanlý-Ovacýk region, Midos Tepe formation from bottom to top includes the following rock types: micrite, pelletoidal-fosilliferous packstone, calcisphere Orbitoid-bearing micrite, mylonitic, ostrocoda bearing wackestone, ostracoda-pellet-calcisphere bearing wackestone, limy mudstone, recrystallized fossil rich (radiolaria) micrite, radiolarian chert, radiolarite, shale (Figure 24). According to the microfacies analyses, Midos Tepe formation is in general deposited in continental slope facies belt of Wilson (1975), but its uppermost part presents typical characteristics of continental slope-oceanic basin facies.

Figure 24- Occurence of Midos Tepe formation in Konya Ýpekler section.

In East of Konya, in Koçkaya tectonic slice (Figure 25) - the most characteristic HP/LT slice of KBB - thin layered pink micrites and radiolarian cherts of Midos Tepe formation was metamorphosed to fine grain marble and Mn-silicate bearing quartz-schists (Özgül and Göncüoðlu, 1997; Floyd et al., 2003). Deposition of Midos Tepe formation in different slices of KBB commences in various times. In its type locality, deposition of the unit starts in Barriasian-Valanginian and continued until late Campanian-early Maastrichtian. It includes all stages of Cretaceous without a considerable gap. On the other hand in East of Yunak the deposition started in middle Malm (Göncüöðlu et al., 1997*b*). In Midos Tepe formation towards top, calciturbiditic and clayey sandstone intercalation increases. Then, it passes to foreland sedimentary mélange in late Cretaceous.

Upper Cretaceous Foreland Sedimentary Complexes

KBB slope sediments are overlain by olistostromes by a primarily transitional contact. It includes material from the Ýzmir-Ankara Ocean, sand-size clasts to giant (> 10 km) blocks. Locally, it is olistostromal and turbiditic, sometimes tectonically mixed, sometimes in classical occurrence of a proximal flysch. A non-genetic term "complex (mélange)" is applied to this unit. In order to differentiate these from metamorphic units produced by tectono-sedimentary processes in an accretion prism, the name "sedimentary complex" is preferred. In these units the main process is mass-flows and slides related with sedimentation. As their generation is realized in a compressional environment after their sedimentation, folding, faulting and rupture by shearing is common. As these formations were subsequently incorporated together with their continental crust into the subduction and metamorphosed, they sometimes display similarities to the subduction-accretion prism of the Ýzmir-Ankara Ocean. During the geological mapping campaign from Bornova to Konya, and then to Bünyan and Hýzýr mountains, the main criteria used for differentiation was: a) transitional contact between platform-slope successions and ophiolite bearing olistostrome, b) metamorphism.

First criteria can be used in every distinctively metamorphic (e.g. Koçkaya in the NE of Konya

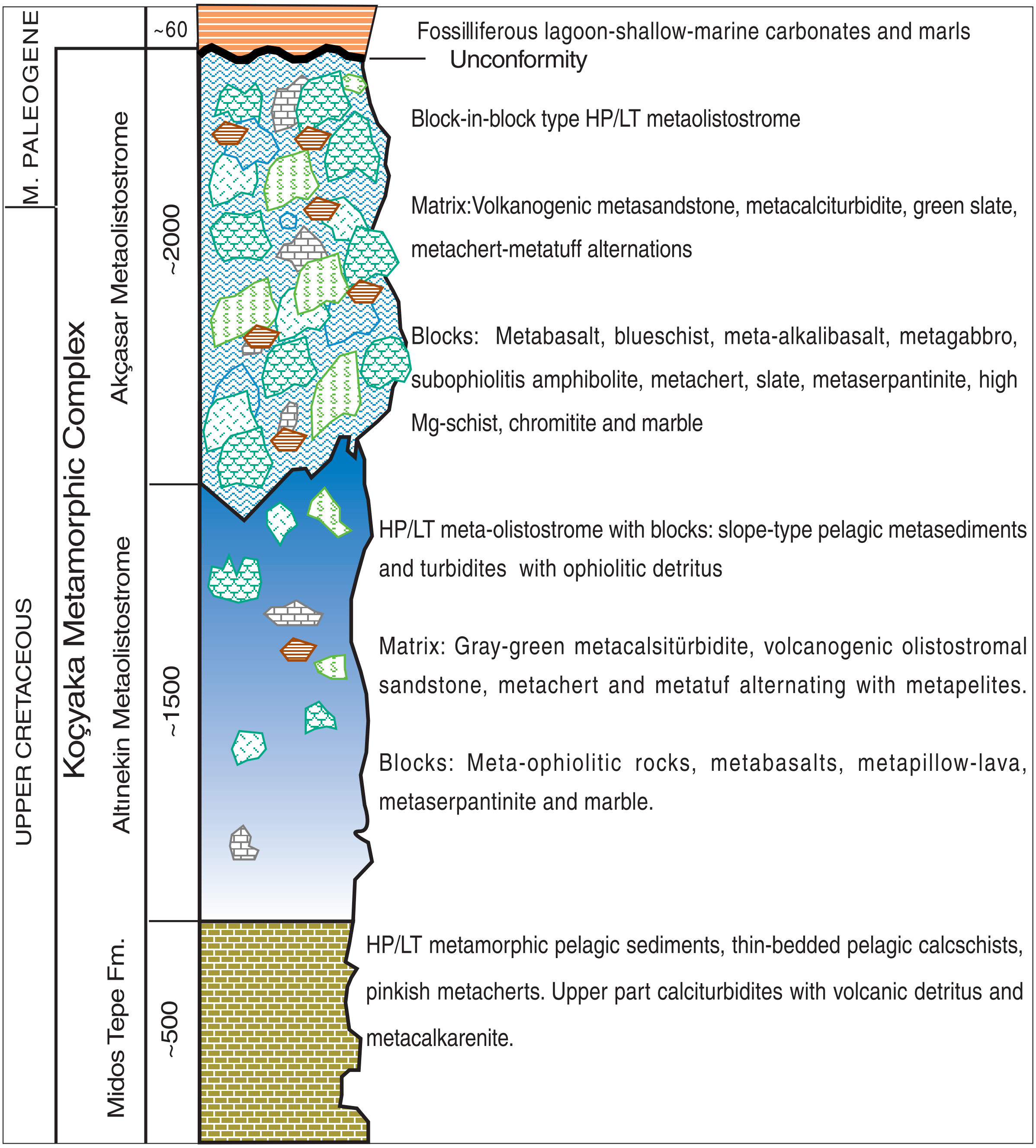

Figure 25- Generalized stratigraphy of the Koçyaka slice (after Özgül and Göncüoðlu, 1999).

(Figure 25, 26) and Girdopdere in Central Sakarya) or non-metamorphic (Çakmak and Çöðürler in Kütahya region, Yüreðil in N of Emirdað, Hanköy section in the N of Bayat, Figure 27) slice of KBB.

Each of these sections is composed of different rock types. However their common feature is the presence of well-developed calcitubidites. In the metamorphic slices (Girdopdere and Koçkaya) these rocks are transformed into aragonite marble that contain Na amphibole bearing mafic rock fragments. In slices that display low-grade metamorphism, they include mafic and ultramafic rocks that belong to mélange, red chert and rock fragments. Sandstones are the most dominant rock types of the unit. The sandstone includes clasts of basalt, diabase, blueschist, green and red chert, red-purple micritic limestone and serpantinite. They have rounded pebbles -reaching 20-25 cm size- and mass flow deposits. They also contain black, purple and pink colored mudstone and clayey limestone bands. The complex includes olistolithes of various sizes: few meters to 10 km. They are incorporated into the clastic matrix by mass gliding and include rock types as: recrystallized Mesozoic limestones of the platform margin and slope, all units of ophiolite sequence and their equivalents that have undergone blueschist metamorphism, mélange blocks affected by blueschist metamorphism, amphibolites and andesitic-dacitic volcanics (Göncüoðlu et al., 2000*a*). In slices that do not show distinct foliation and metamorphism, greywackes and sandstones include clasts of distinctively foliated glaucophane- lawsonite bearing blueschist pebbles. They indicate that some fragments of the mélange have undergone subduction and subsequently transported into the foreland basin. Moreover, limestone blocks reaching a few kilometers in size are equivalents of the Tauride-Anatolite platform. This also indicates that platform-margin/slope units are also transported into the basin by gravity sliding. Locally, not only the blocks but the matrix itself is also affected by deformation and metamorphism.

The thickness of the unit is more then 3000 m in Kütahya region (Özcan et al., 1989) and 5000 m in Bornova region (Yalýnýz et al., 2005; Tekin and Göncüoðlu, 2007).

In Kütahya-Yüreyir and Tavþanlý - Ovacýk areas, in the lower part of unit middle - late Maastrichtian fossils: *Globotruncana* cf. *conica*, *G. linneinae, G.* aft. *gannseri, Hedbergella* spp. (determined by Dr. T. Çoruh) are found (Özcan et al., 1989). On the other hand, the youngest blocks are dated as Turonian-Campanian. The oldest units that unconformably overlie the unit in Konya region yielded Thanetian fossils (Göncüoðlu 1992; 1997*b*). According to these data the age of the unit is post Campanian-pre Thanetian (Göncüoðlu et al., 1997*b*).

In the easternmost edge of the belt the foreland units are defined as "Yeþiltaþ Yayla Complex" by Erkan et al. (1978). The characteristics features of the succession and the transition facies to siliciclastics are very similar to the western areas (Figure 28). In northern slope of Hýzýr Mountain, in Göktaþlýyurt and Büyükbileyik areas recrystallized limestones of Hýnzýrdaðý Metamorphics also occur as olistolithes reaching a few kilometers in size. They are embedded in a matrix composed of green colored mudstone (Göncüoðlu et al., 1994). Sometimes, the matrix is calciturbiditic and alternate with olistostromal sandstones and conglomerates. Within the matrix, blocks and clasts of limestone, serpentinized ultramafics, radiolarian mudstone and HP/LT metabasic rocks are included. The fossils obtained from pelagic limestone blocks are indicative for Campanian. By this a post-Campanian formation age is guarantied.

SUBDUCTION-ACCRETION COMPLEXES AND OPHIOLITES OF THE ÝZMÝR-ANKARA OCEAN

OPHIOLITE SLICES

This unit differs from ophiolite blocks within the mélanges by their structural position and by

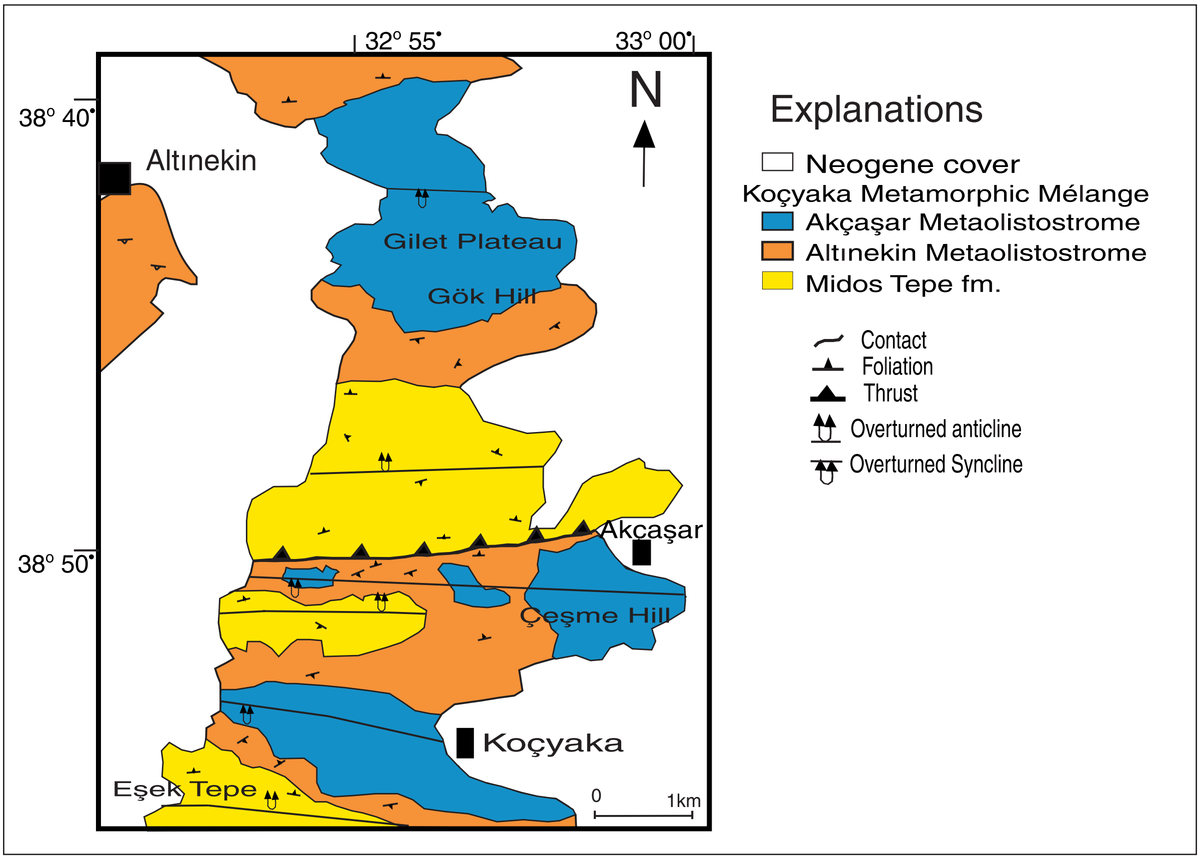

Figure 26- Geological map of the type-locality of Koçyaka slice to the NE of Konya.

including partial sections of an ordered ophiolite sequence. The ophiolites tectonic slices are of a few kilometers in size, and represent dismembered successions of the oceanic lithosphere. Ophiolitic massifs of Bursa-Orhaneli (e.g. Tankut, 1991), Bursa-Harmacýk (Manav et al., 2004), Kütahya-Daðardý (Bacak et al., 2003), Central Sakarya-Daðküplü (Asutay et al., 1989; Göncüoðlu et al., 1996), Ankara-Edige/Kalecik (Tankut, 1984) are relatively well-known units. These massifs include mostly metamorphic tectonites and partially cumulate sections. They are generally highly serpentinized and intersected by isolated dykes.

Daðküplü Ophiolite is located in Central Sakarya area. The name is used by Þentürk and Karaköse (1981) for describing an ophiolitic slice which is supposed to have pre-Liassic age. Peridotites with chromites are the most common rock types. Through shear zones they are highly serpentinized and intersected by rodengitic dykes. The unit includes pyroxenite bands and lenses in some places. In the lower part of the unit typical metamorphic texture is observed. In addition to dunite and harzburgite, orthopyroxenite with large (ca 3 cm long) orthopyroxene crystals and phlogopite-bearing peridotite are observed. In this massif, from bottom to top following tectonic slices were determined by Asutay et al., (1989): gabbro, clinopyroxenite, mafic and ultramafic cumulates and tectonites.

In Orhaneli and Harmancýk ophiolitic slices are generally thrust over the HP/LT subduction-accretion prism units. Their visible thickness is a

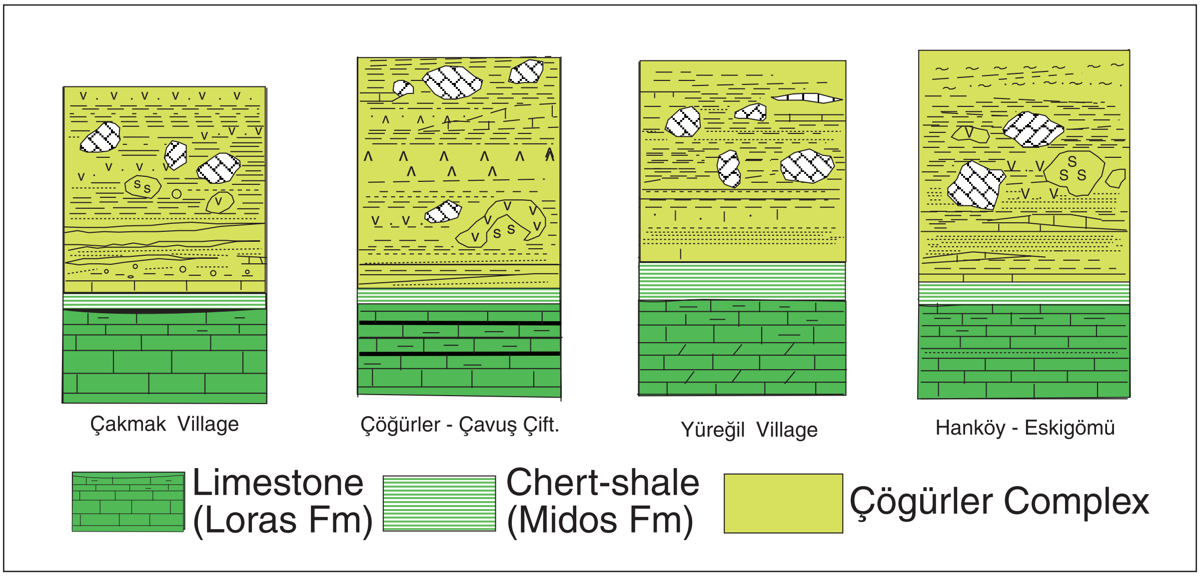

Figure 27- Sections with transitional contacts between platform slope and olistostromal flysch sediments in: (a) Kütahya-Çakmak, (b) Kütahya-Çöðürler, (c) Emirdað -Yüreðil and (d) Bayat-Hanköy (simplified after Özcan et al., 1989).

few thousand meters. In regions like Sakarya, Yunak and Konya, beneath the ophiolite slices different units of the mélange complexes are found.

Subophiolitic Metamorphic Rocks

Throughout KBB, metamorphic rocks are found in the basement of ophiolite slices. They are described in North of Kütahya Baþdeðirmen and Kaynarca (Göncüoðlu 1990 *a,b*; Figure 29), Central Sakarya (Göncüoðlu et al., 2000*a*) and Konya Altýnekin (Özgül and Göncüoðlu, 1999). In both areas in the Kütahya region, metamorphic rocks outcrop as a single slice but sometimes they have more than one slice. They are attached to the basement of the metamorphic ultramafics and their thickness varies between 10 m and 150 m. Rock types of these metamorphics and their structural relations in Kaynarca area are shown in figure 30. Generally the ophiolite is composed of peridotite intersected by micro gabbro dikes, serpentinized harzburgite and dunite. At the ultramafic-mélange contact, 30 meters thick, black colored, well foliated amphibolites are observed (Figure 31a,b). The minerals of the unit are: hornblend + garnet + sphene and secondary Mg-chlorite and epidote. Away from the contact diopside-green amphibole-plagioclase-sphene schist; actinolite-biotite-oligoclase schist, pinkish fine crystallized, thin bedded marble intercalated with rutile-Mn-garnet-piemontite muscovite quartzite are observed. It is suggested that the rocks in that last part represent micritic limestones and Mn- rich cherts. The 30 meters thick lowermost slice that overlies mélange unit comprises greenschists showing less deformation and partly preserved basaltic texture and pillow structure; fine grained recrystallized limestone intercalations, Mn-rich radiolarian chert, slates and quartz-schists are observed. It is concluded that this slice includes the same rock types as the overlying slice. The overlying slice includes metabasics that are metamorphosed to garnet-bearing amphibolite. It is argued that these metavolcanic and metasedimentary successions represent oceanic sediments. During intra-oceanic subduction they

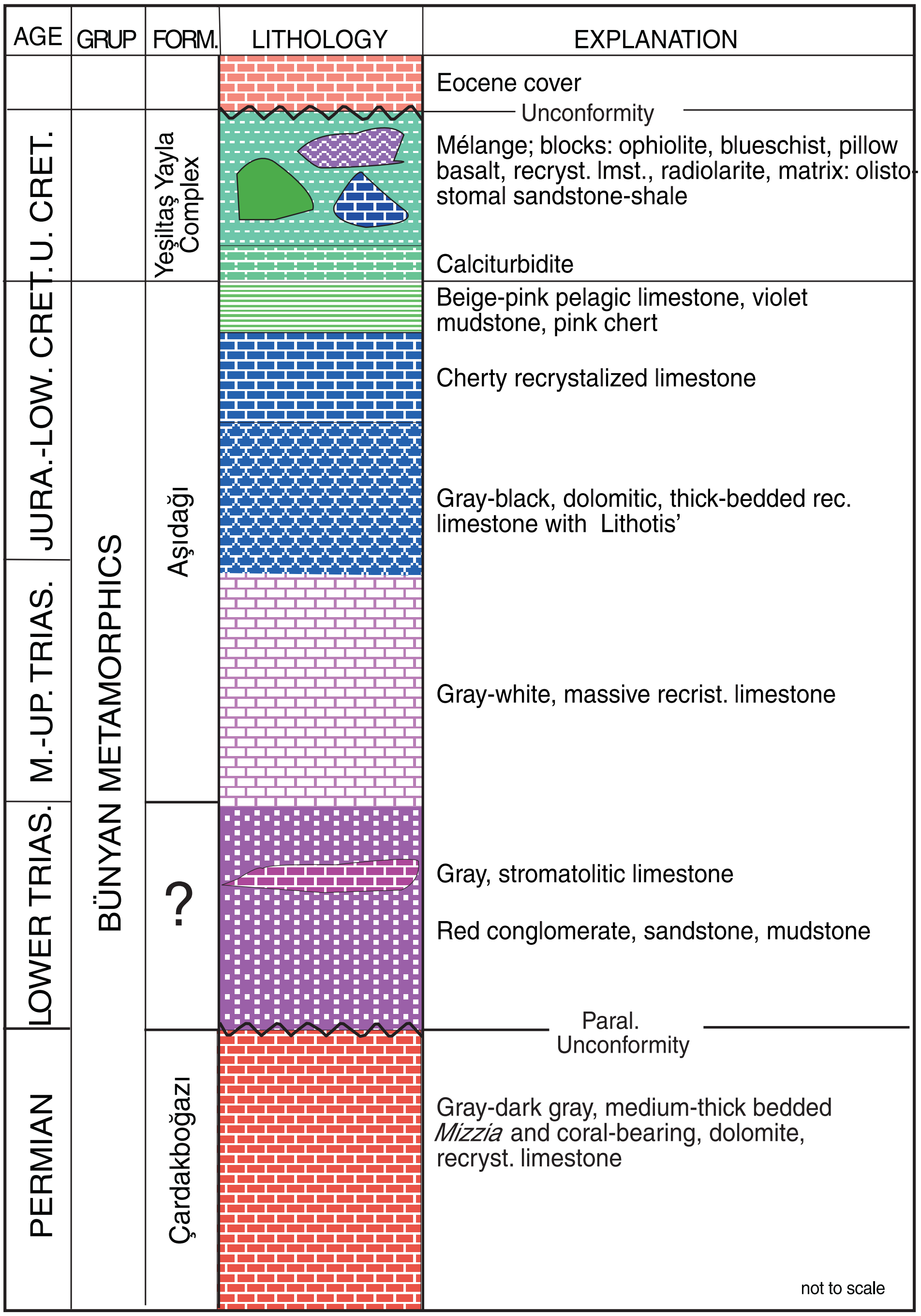

Figure 28- Generalized stratigraphy of the Hýnzýr Daðý Metamorphics and their relations with the ophiolite-bearing melange (Göncüoðlu et al., 1994).

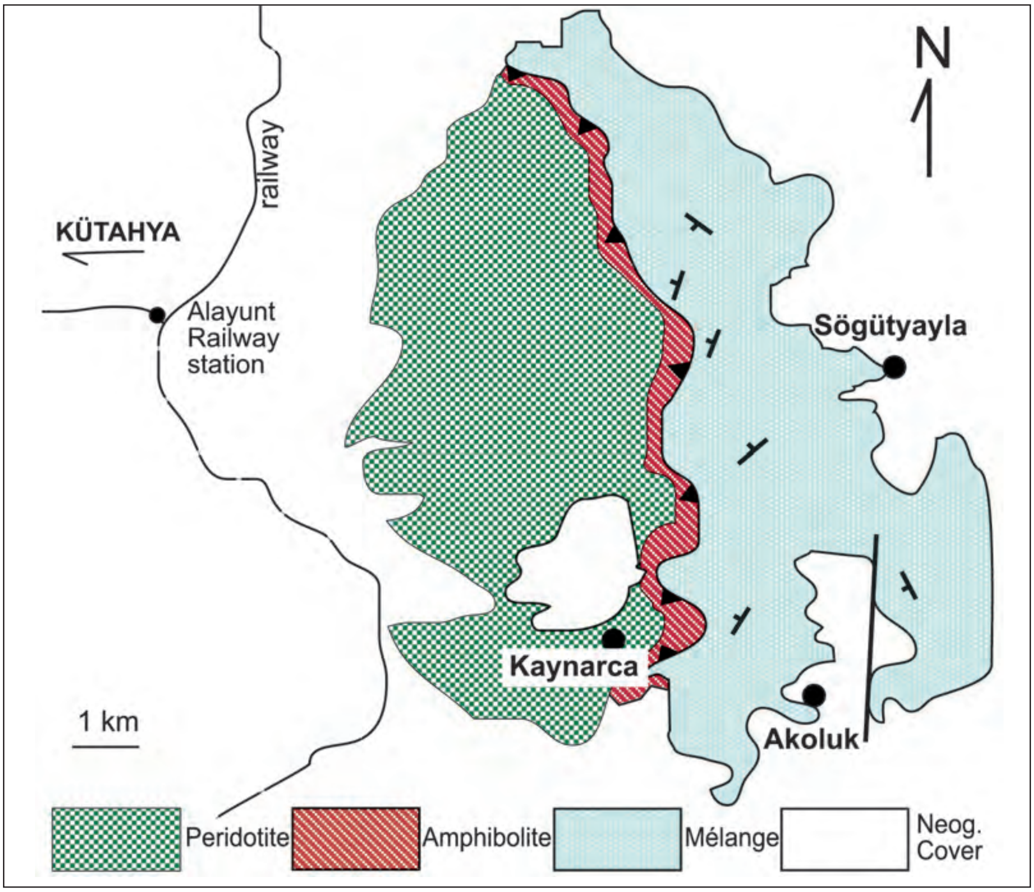

Figure 29- Structural setting of the sub-ophiolitic rocks and melange units with the ophiolite to the N of Kütahya-Kaynarca. The location of the map-area is shown in the regional map in figure 17.

came into contact with the hot lithospheric mantle and affected by high temperature metamorphism (Göncüoðlu 1990 *a,b*; Önen and Hall, 1993). In both slices HP/LT Na-amphiboles partly replace brown and green amphiboles which are product of HT metamorphism. This shows that the sub-ophiolitic rocks were also subducted and affected by HP/LT metamorphism (Özcan et al., 1989).

Subduction-Accretion Mélanges

Apart from more or less uniform ophiolitic slices that belong to the Ýzmir-Ankara oceanic lithosphere, most of the oceanic material that belongs to this ocean, is mixed or accreted in subduction zone.

In this mélange, units that belong to the upper mantle and oceanic crust, oceanic islands and their platforms and slope sediments that related with them, island arc and related pyroclastics, fore-arc and back-arc oceanic crust fragments or basin sediments developed in this conditions, various rocks that belong to slope of the continental crust or transitional rocks between continental-oceanic crust units and their tectonically mixed equivalents that were affected by HP/LT or medium P- LT are observed.

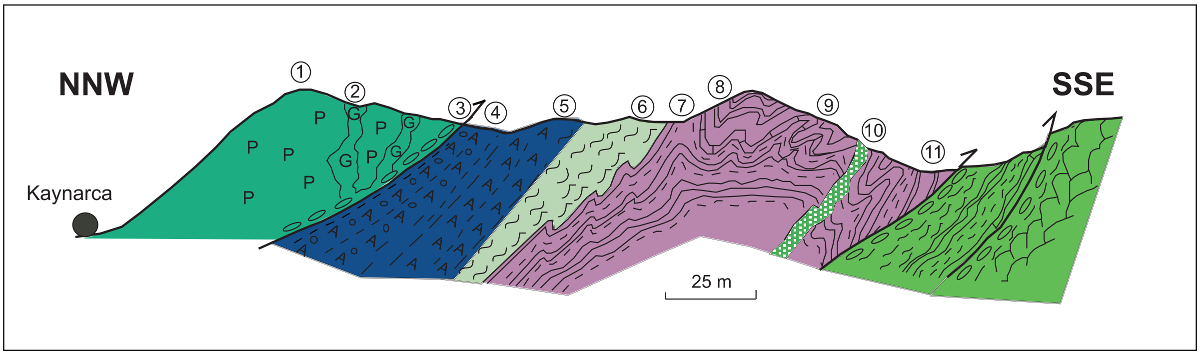

Figure 30- Contact relations of the ophiolitic units with the underlying subophiolitic amphibolites.

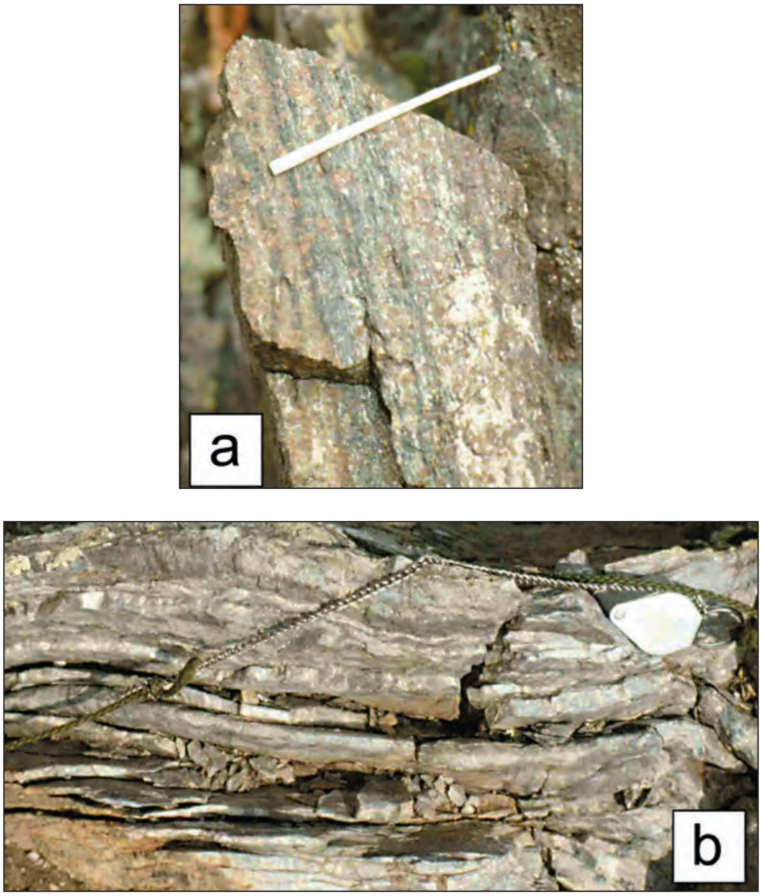

Figure 31 a- Garnet amphibolites from the Kaynarca subophiolitic metamorphics, b- greenschist metamorphic pelagic sediments (more resistant meta-radiolarite bands alternating with recrystallized micritic limestones).

As different from the foreland sedimentary complexes, subduction-accretion mélange units always display moderate to high deformation and metamorphism. These units are mostly tectonically mixed. Even primarily accumulated by sedimentary processes and show characteristics of mass flow and olistostromal features, they are variably sheared. So that their deformed matrix is represented by disordered alternation of green-gray- red coloured, conglomerate-sandstone-siltstone. Sandstones are the most common rock types. They are generally green and gray colored, fine- medium grained and thin-medium bedded. They include microscopic grains-pebbles- blocks of ophiolitic lithologies. The grains are subrounded and non-graded. In places where sandstones are green, basic volcanic grains are dominant. In places where sandstones are yellow and red, radiolarite and marble clasts are dominant. Siltstones and mudstones are generally highly sheared. Dark gray-green and yellow colored siltstones are thin bedded and laminated.

All units of mélange show metamorphism. Metamorphism sometimes erased all sedimentary and magmatic textures, and completely eliminated the original mineral compositions. So, the rocks have commonly gained schistose texture. As it can be seen in the microscopic analysis of incipiently metamorphosed metavolcanics with poorly developed planar fabrics, the new metamorphic minerals are formed in the veins or replace primary minerals along cleavage planes.

Blocks

In KBB the mélange includes several blocks with variable sizes: ultramafics, gabbros, basalt, dolerite dike fragments, radiolarite, pelagic limestones, mafic tuffs, blueschists and recrystallized limestone blocks.

Mafic-Ultramafic Rocks are observed as large allochthonous masses or blocks within the mélange. They include all members of a dismembered ophiolite sequence.

Amphibolites are 5-20 m long blocks and observed in outcrops extending from Bornova to East of Konya. They are massive- weekly foliated and banded Under microscope the rock includes hornblende, plagioclase, rare garnet and epidote.

Blueschist Blocks and Slices occur either as olistolithes that were incorporated into the accretion prism material by tectonic or sedimentary processes. They include differences in terms of their metamorphism and matrix. Slices are homogeneous HP/LT metamorphosed rocks that are traceable laterally through several kilometers within a proper belt with several slices. No matter what rock type is included, they present sheared contacts with their surrounding units. Their character as a tectonic sliver is obvious by differences in metamorphism with surrounding units. So, although they are integral parts of the mélange, they can be handled separately as "blocks" and "slices". Blueschists are common blocks of mélange. They form outcrops from few ten cm to several hundred meters in size. These blocks are mostly composed of basic volcanic and volcanoclastic rocks. They include rocks with different HP/LT metamorphic paragenesis. Original magmatic texture of basic rocks in some of the blocks is preserved. Only in veins, prismatic lawsonite and needle like Na-amphibole is generated. On the other hand, in some blocks basic rocks are well-foliated. Some minerals like glaucophane and lawsonite reached textural equilibrium. In the most common blueschists the first metamorphic phase has variably erased the original mineral composition and parallel to the S1 plane, chlorite-epidote- albite and actinolite are formed. In the rock the remnant texture and relic pyroxene can be still preserved. In the second phase, blocks less affected from metamorphism/deformation include needle-like Na amphibole crystals. In blocks where textural equilibrium is established, typical violet glaucophane phenocrysts are formed parallel to S2 plane. Blueschist metamorphism is not limited to rocks of basic volcanic origin blocks. It is also seen in coarse grained gabbros, serpentinites, and cherts.

Important units that are described as slices are: Yeniþehir (Okay, 1980, 1986), Sivrihisar (Çetinkaplan et al., 2008), Yunak and Koçkaya HP/LT. It is possible to see similar metamorphism features in units around Menderes Core Complex (e.g. Rimmele et al., 2005). Koçkaya slice (Figure 25 and 26) has typical feature among them. It is examined by Floyd et al. (2003) and Droop et al. (2005).

Andesite and Dacite blocks are not very common in the mélange. Their size varies from pebble size to 60-70 meters. Blocks are placed in a volcanoclastic matrix. It has green-yellow color, and contains conglomerate-sandstone intercalation mostly with grains of volcanic origin. In hand specimen andesites are green, mylonitic and include large plagioclase phenocrysts. Under the microscope the rock includes plagioclase, biotite and hornblende phenocrysts in a highly altered and sheared matrix. Along shear planes metamorphic phengites are formed. Dacitic rocks look like andesites but they include corroded quartz in phenocrysts phase.

Radiolarian Chert and Mn-Cherts are the most common block types which are very distinctive with their red-green-purple-black colors and bedded-laminated structures. Chert blocks reach their maximum size in North of Akhisar and Central Sakarya. Radiolarian cherts include purple-pink-green colored shale interlayers In smaller blocks deformation is observed distinctively. Under the microscope, radiolarian cherts are composed of very fine grains of quartz and opaque minerals. In deformed blocks, foliation is distinctive because of presence the white mica flakes.

Radiolarian cherts found in the mélange units have been studied comprehensively in the last few years (Bragin and Tekin, 1996; Rojay et al., 2001; Tekin et al., 2002; Tekin and Göncüoðlu, 2007; Göncüoðlu et al., 2006*a,b;* Tekin et al., 2006; Tekin and Göncüoðlu, 2009). Provided information shows that in Ýzmir-Ankara Basin the deposition of radiolarian cherts first start in Upper Carnian. As it can be seen in figure 32, apart from a gap in Jurassic, radiolarian chert deposition continued up to Upper Cretaceous.

Massive and Pillow Lavas are together with the radiolarian cherts, the most common rock types of KBB. These rocks constitute blocks -few centimeters to 100 meters in size- in mélange. In areas where the matrix of the mélange is visible, blocks are surrounded with an olistostromal matrix. Lava clasts and pebbles of are the dominant ingredients of the matrix. In areas of block-block contact, rocks and contacts are sheared. Blocks of lava are different in terms of their appearance (pillow lava, lava breccia, massive flow, pillow breccia etc.), pillow size (10-80 cm), amygdale distribution, metamorphism and most importantly chemical properties. In numerous studies done through the belt (Göncüoðlu et al., 2006 a,b; Aldanmaz et al., 2008 and references there in Gökten and Floyd, 2007) it is noticed that the geochemical character of the pillow lava changes from normal mid oceanic ridge belt (N-MORB), enriched mid oceanic ridge belt (E-MORB), oceanic island belt (OIB), island arc tholeiites (IAT), supra subduction zone (SSZ) types (Figure 33). The SSZ types further include fore-arc (FABB) and back-arc (BABB) subgroups. In figure 34 the geochemical characterization together with the age of volcanism is presented. In the light of data obtained from the blocks in the mélange, it is suggested that the earliest formation of the oceanic crust in Ýzmir-Ankara Ocean started in early Late Triassic, seafloor spreading continued in Jurassic-Early Cretaceous, and the ocean started to close by intra-oceanic subduction by late Early Cretacous.

Recrystallized Limestones are also common constituents of the mélange and represented by a grate variety of properties. The most common blocks are gray-black colored, medium-thick bedded recrystallized limestones. These blocks may form 200-300 meters long to 70-100 meters large outcrops. Another frequently outcropping recrystallized limestone block type is gray-yellow-pink colored, thin bedded, and well-foliat-

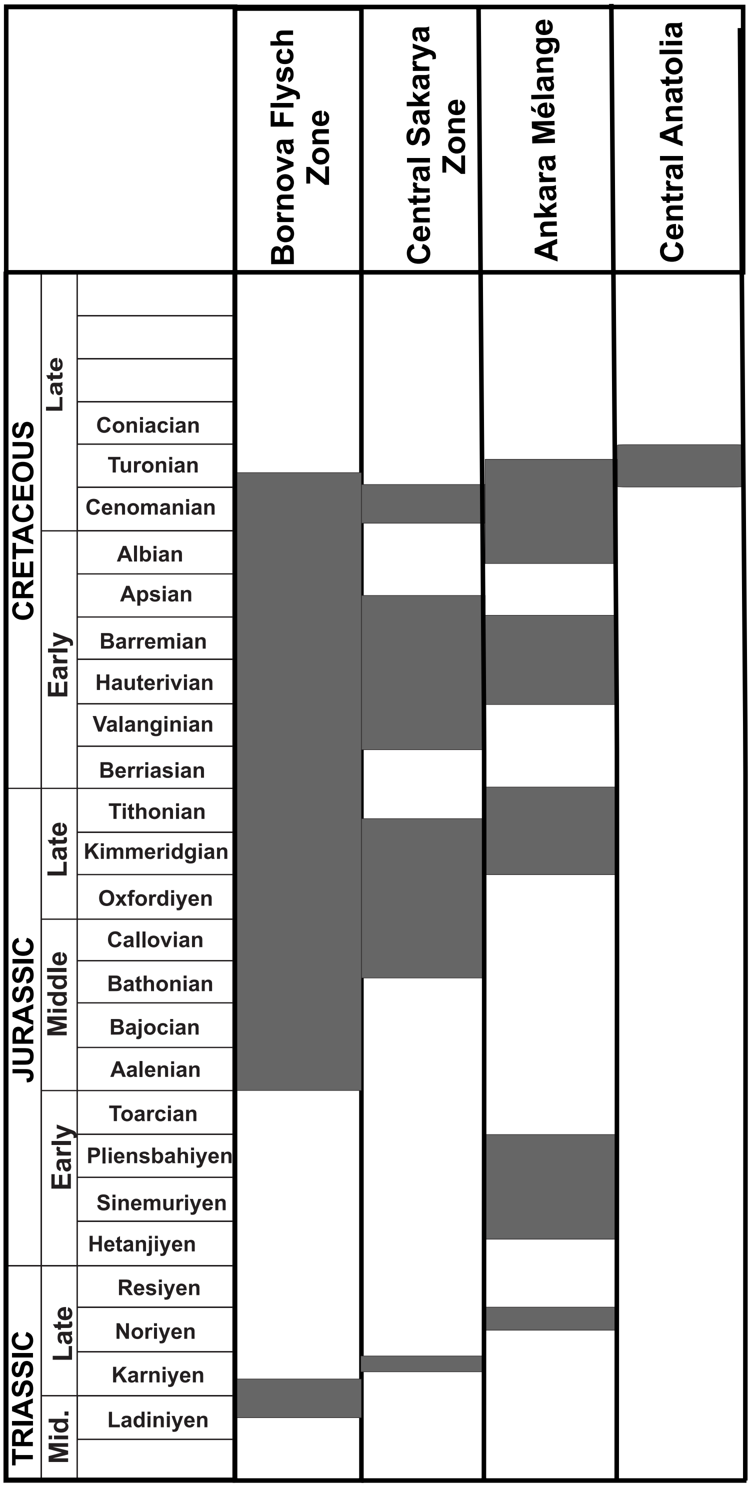

Figure 32- Radiolarian ages obtained from different parts of the KBB (for data see text).

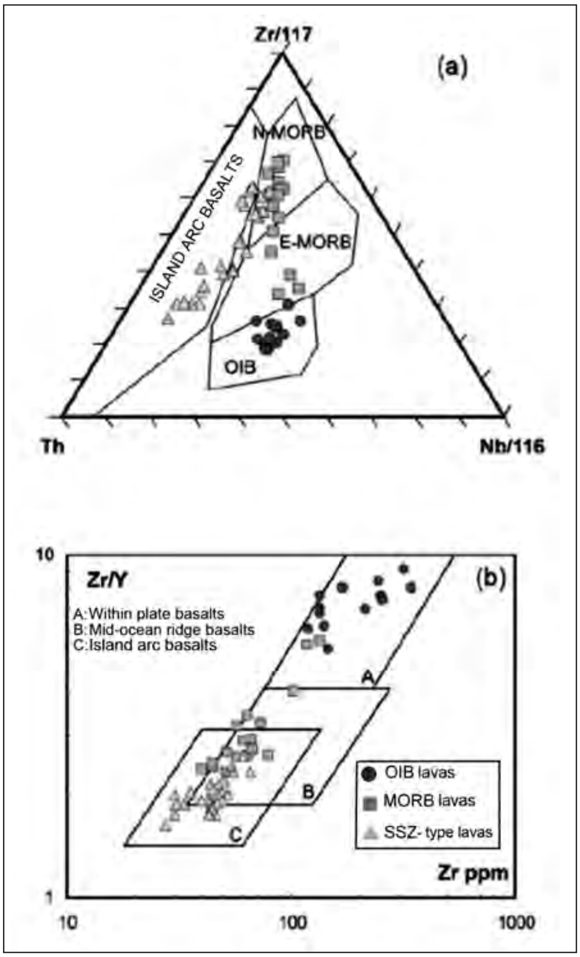

Figure 33- Tectonomagmatic discrimination of volcanic rocks from the KBB melange, flysch and ophiolitic nappe (for data see Göncüoðlu et al., 2006*b*).

ed pelagic limestones. Similar lithologies to these limestones are observed in the upper part of the platform successions. Generally these limestones are recrystallized and include badly preserved Lower Cretaceous pelagic fossils. Micritic limestones and radiolarian cherts are in many cases associated with basalts and yielded foraminifers in chert intercalations. In previous work, Late Cretaceous (Campanian) (Erdoðan, 1990), late Santonian- Maastrichtian (Akdeniz et al., 1986) ages are determined by using planktic foraminifers in these lenses. Samples from different parts of the matrix in the mélanges of the Bornova Flysch Zone yielded Campanian, Maastrichtian-Danian (Erdoðan, 1990; Akdeniz et al., 1986) ages. By this, it is proposed that the deposition in this basin is Maastrichtian-Early Paleosen (Erdoðan, 1990; Akdeniz et al., 1986).

PALEOSEN-EOSEN COVER UNITS

The oldest overstep sequence disconformably covering products of the Late Cretaceous ophiolite emplacement and related tectonic imbrication is described as Kýzýlçay Group in Central Sakarya and Kartal formation in NE of Konya (Göncüoðlu et al., 1997*b*; Çemen et al., 1999). Both of these formations are composed of dark maroon-red, locally green-yellowish-green colored, thick bedded, non-graded or badly graded conglomerates with rounded-sub angular pebbles. Compounds of the conglomerate ranging from pebble to block in size are quartz, red-black chert, quartzite, andesite, monzonite, gabbro, gray limestone, white recrystallized limestone, red chert, and metamorphic rock fragments. Towards the top conglomerate-sandstone alternation is more distinctive.

In the North of Altýnekin (Figure 35) the cover unit starts with red colored conglomerate overlying highly sheared serpentinites of the Koçkaya Metamorphic Complex. Locally the pebbles reach to block size. They are dominantly derived from the radiolarite, pelagic limestone, gabbro and rarely syenites of the underlying mélange rocks. In lower part pebbles are cemented with carbonate cement. In the cement, plenty byrozoa and algae (*Melobesia*) fragments are found. Above, macrofossil rich gray colored clayey limestones, and conglomerates with limestone and radiolarite pebbles are deposited. In some of these pebbles, *Globotruncana*-rich limestones of Cenonian age are found. Limestones of this age and fossil content are also noticed in the mélange of the basement. In clayey limestones constituting the matrix of Kartal formation plenty of algal flocs, corals, *Haddonia* sp., and *Planorbulia create* (determined by Dr. E. Sirel); *Micocodium* sp., *Planorbulina* sp., *Ethalia* sp.,

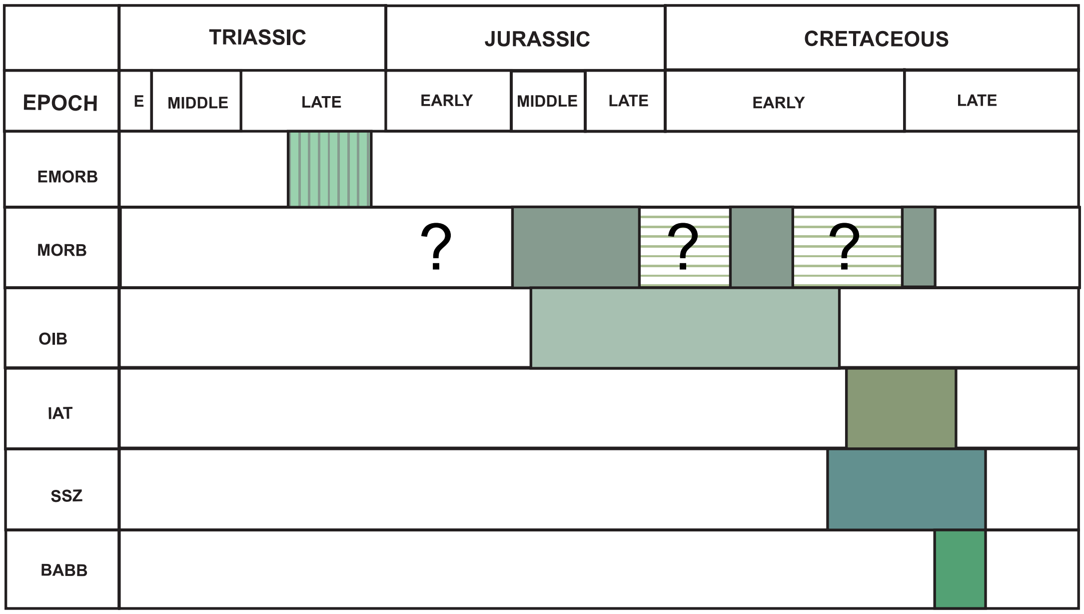

Figure 34- Evolution of the volcanism within the Ýzmir-Ankara ocean through time (after Göncüoðlu et al., 2006*a*).

Milliolidae and Discorbidae (det. S. Erk) are found. By this, the age of the unit is given as Danian- ?early Tanesian. Towards top in marly sections fossils like *Chiasmolithus bidens*, *Fasciculithus tympaniformis, Fasciculithus involutus, Ellipsolithus macellus, E. distichuc, Discoaster multiradiatus, D. Aster, D. mohleri, Neoschiastozygus perfectus, Sphenolithus anarrhopus, Zygodiscus herlynii ve Coccolithus* sp. were determined. The age of this level is accepted as Thanetian. Here and also in the neighbouring Tuz Lake basin, upper level limestones that probably deposited in lagoonal environment and include the fossils *Laffiteina* sp., *Broeckinella* cf. *Arabica*, *Glomalveolina primaeva, Alveolinidae, Asterigerina* spp., *Hottingerina* cf. *lucasi*, *Mississippina* spp., *Ethelia* sp., *Corallinacea, Dasycladacea, Distichoplax biserialis*. The age of this limestone level is also Thanetian (Dr. E. Sirel, oral communication).

In NW of Cihanbeyli, in Sülüklü-Sarýkaya area another cover succession starts with a sedimentary breccia with angular fragments (ophiolite, radiolarite, Loras- and Midos-type recrystallized limestones) from the basement over the Loras-type limestones. Its includes 5-10 meters thick intervals of purple-red- brown sandstones, carbonate cemented limestones, marls, evaporites (sabkha sediments) and conglomerates. The pink and gray colored algal limestones in this succession yielded Danian fossils.

The Early Tertiary cover rocks are encountered in Afyon-Bayat and in its North (Figure 17). They are the oldest rocks we found yet covering several thrust slices with an angular unconformity. These rocks are defined as the Hanköy formation by Özcan et al. (1989). The lithostratigraphic features and fossiliferous levels of this formation in N KBB are shown on figure 36. In the lowermost part of the successions in several sections algae (*Melobesia*) rich sabkha-type carbonates are found within red colored fluvial clastics. Alveolina-bearing samples taken from carbonate-dominated parts indicate an Upper Paleocene age.

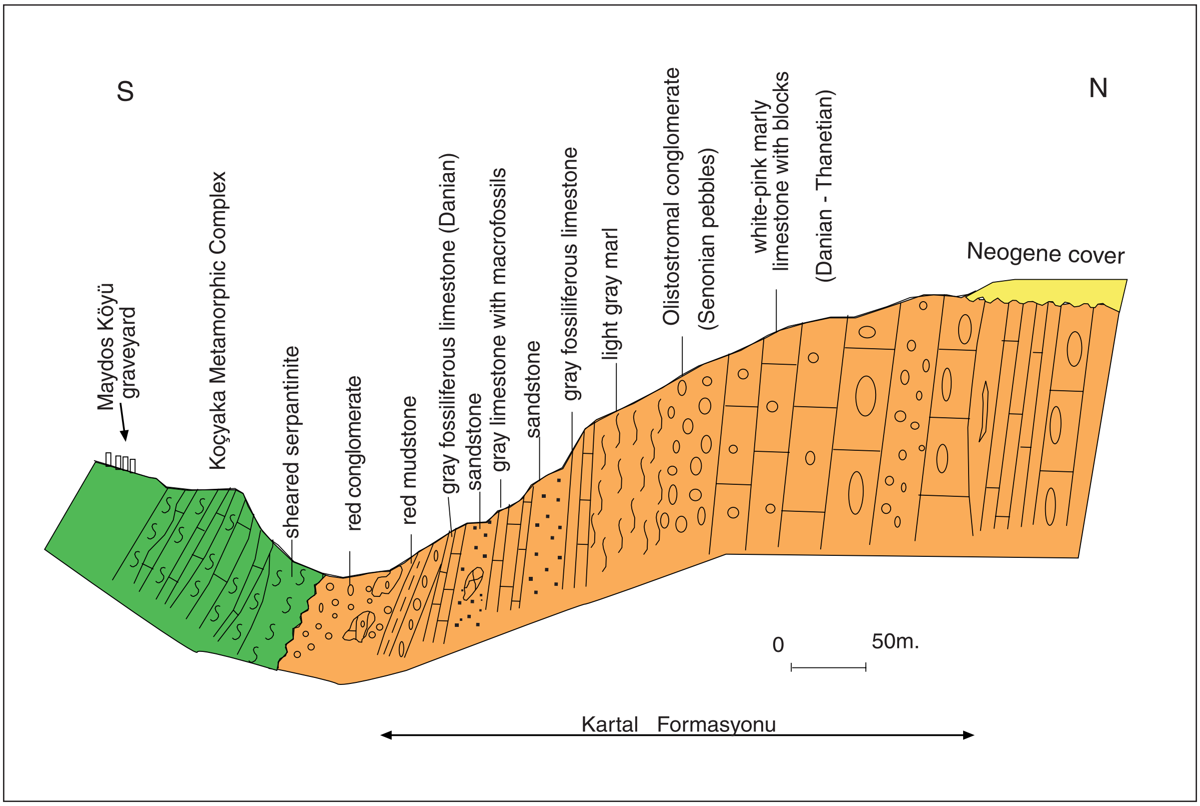

Figure 35- Cross-section of the Upper Paleocene cover of the melange units to the N of Altýnekin (Göncüoðlu et al., 1997*b*).

Also in the Bornova Basin, Tertiary sediments starting over the Upper Cretaceous flyschoidal sediments with an angular unconformity are named as Baþlamýþ formation by Konak et al. (1980). The unit starts with red conglomerates. The overlying shallow sandstones and limestones of brackish water environment include Upper Paleocene fossils.

In Central Sakarya region, the first unconformable unit over the ophiolitic rocks and mélange units are pink colored algal carbonates. Towards top, red colored mudstone-sandstone and carbonate alternation is correlated with the Upper Palaeocene rock units mentioned in the other regions (Göncüoðlu et al., 1997b). In this area the cover sediments are sliced with the underlying mélange units during the compressional events of the end Eocene.

In terms of their depositional environment, Paleocene units have the characteristics of alluvial fan deposits that reach a very shallow lagoon. Considering the blocky character and the internal order, they were probably formed in front of a rapidly uplifting block.

The Eocene units sometimes transgressively sedimented on the peneplained topography of the eroded Palaeocene units. They start with a few meters thick carbonate-cemented, green-cream colored marine conglomerates-sandstones and continue with cream colored, medium bedded locally nodular limestones. Towards top volcanic rocks as domes and volcanoclastic material is observed in carbonate-clastic successions. All along KBB in samples taken from the lower part of the unit, Middle Eocene (Lutetian) fossils are determined (e.g. Konak et al., 1980;

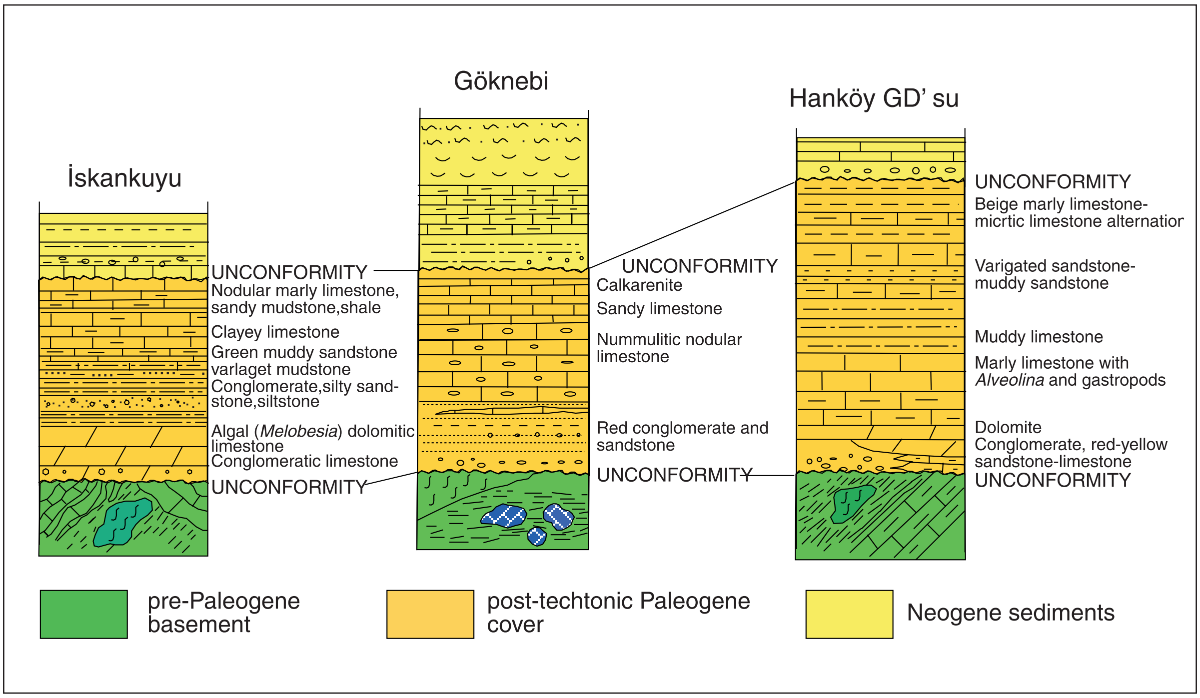

Figure 36- Stratigraphic features of the Hanköy formation in Kütahya area.

Özcan et al., 1987, 1989; Göncüoðlu et al., 1997b).

After the Middle Eocene no marine deposition is reported yet in KBB. The Eocene sediments, especially in the north of KBB are incorporated in Miocene tectonic slices (e.g. Central Sakarya area).

The formation of the Neogene basins and related volcanism and the evolution of the region in the Neotectonic period are beyond the scope of this present review but can be found in several recent publications (e.g. Özsayýn and Dirik, 2008).

GEOLOGICAL EVOLUTION

Geological evolution of KBB in a way includes the geological evolution of Tauride-Anatolide Platform and Menderes Core Complex. This evolution will be examined in the frame of major geological events and data obtained from the KBB.

PAN-AFRICAN/CADOMIAN PERIOD

The Precambrian units are commonly represented by low grade metamorphism in the Taurides. In Anatolides they include rocks of ortho- and para origin and their outcrops of proven age are found in Menderes Core Complex, and in KBB. Afyon-Ýhsaniye Basement Complex, their equivalents -Sandýklý Basement Complex in the S and Göktepe Metamorphics in North in Sömdiken Mountains have similar properties. Considering Afyon and Sandýklý units, it is observed that a clastic-dominated unit with rare carbonates is intersected by post-collisional felsic magmatic rocks (Gürsu and Göncüoðlu, 2008). Zircon U/Pb ages (542 Ma) obtained from this unit are coherent with the age of the core gneisses of Menderes Core Complex (Koralay et al., 2004; Candan et al., 2005). Göncüoðlu (1997) and Gürsu and Göncüoðlu (2005, 2006*a*) propose that this magmatism was formed in northern margin of Gondwana (Figure 37), above the southward subducting oceanic lithosphere. According to this model, the Late Proterozoic subduction and related arc magmatism that had began at 600-575 Ma. It stopped between 575-550 Ma by collision of North margin of Gondwana-arc-trench collision. A new magmatic phase has started in between 550-525 Ma because of post collision and/or back arc extension. The products of this succession of events are generally attributed to the late events of the Pan-African Orogeny (e.g. Þengör et al., 1984). However a number of Late Neoproterozoic magmatic/metamorphic events in the Avalonian, Southern European and North African terranes, seem to be unrelated to the Pan-African events in terms of time and space. Because of these differences such events were ascribed to a proper igneous-metamorphic event, named as Cadomian Magmatism (Murphy, 2002). Göncüoðlu (1997) pointed out that this magmatism in North Gondwana between 600-550 Ma has not only affected the Avalonian-S European-N African active margin but also the Tauride-Anatolide and Ýstanbul-Zonguldak terranes (Figure 37).

Detailed clay mineralogical studies (Bozkaya et al., 2006) performed on the Precambrian sediments in the Sandýklý area proved that these Cadomian events are not limited with magmatism but affected the pre-Tommotian (< 530 Ma) metasediments and the granitoids intruding them by low grade metamorphism.

VARISCAN PERIOD

Except the undated quartz-rich clastics of Ýhsaniye Metamorphic Complex (Gürsu et al., 2004) no Lower-Middle Palaezoic sediments are described from the KBB and hence from the northern part of the Tauride-Anatolide platform. In the inner part of the Palaezoic platform, however, disregarding the deepening characterized by the Silurian ribbon-cherts in Konya, the deposition between Cambrian to Devonian is represented by platform-type siliciclastics and carbonates. Within this platform, outcrops of an Early Carboniferous back-arc basin extents from Konya to Karaburun (Özcan et al., 1990*b*). The within-plate alkalen magmatism represented by intensive dike-swarms cutting the Devonian platform carbonates are interpreted as an evidence of continental extension that were formed at the first step of this basin opening. By considering the geological features of Halýcý mélange in the North of Konya together with the petrographic character of bi-modal volcanic rocks, Göncüoðlu et al. (2007) claimed that the opening of this basins may be due to southward subduction of Paleotethyis as marginal or more probably as a back-arc basin (Figure 38) in the North of Tauride-Anatolide platform. The presence of such an oceanic basin could also explain the Middle Carboniferous oceanic rocks observed in the Tavas Nappe (Göncüoðlu et al., 2000*c*). As mentioned above, in this nappe, also Middle Carboniferous MOR-basalts and representatives of a Moscovian-Kasimovian oceanic island is discovered., Considering that Ýzmir-Ankara Ocean has not been opened yet during this period and that Lycian Nappes are originated from the northern margin of the Tauride-Anatolide platform, the source of this oceanic crust material must be Paleotethys or more probably a basin to the South of it. What led the closure of this basin is arguable (Robertson and Pickett, 2000; Göncüoðlu et al., 2007; Moix et al., 2008). However it is clear that this closure accompanied by deformation of Variscan time (Göncüoðlu, 1989). In northern part of KBB, Middle Permian clastics are transgressively overlying various units with a weak angular unconformity. The Permian transgression (Figure 39) generally starts with deposition of shallow-marine quartzites. Locally the deposition continued with thick platform carbonates. The fact that Middle-Upper Permian platform deposited unconformably over both the Tauride-Anatolide Units and the Sakarya Composite terrane, indicates

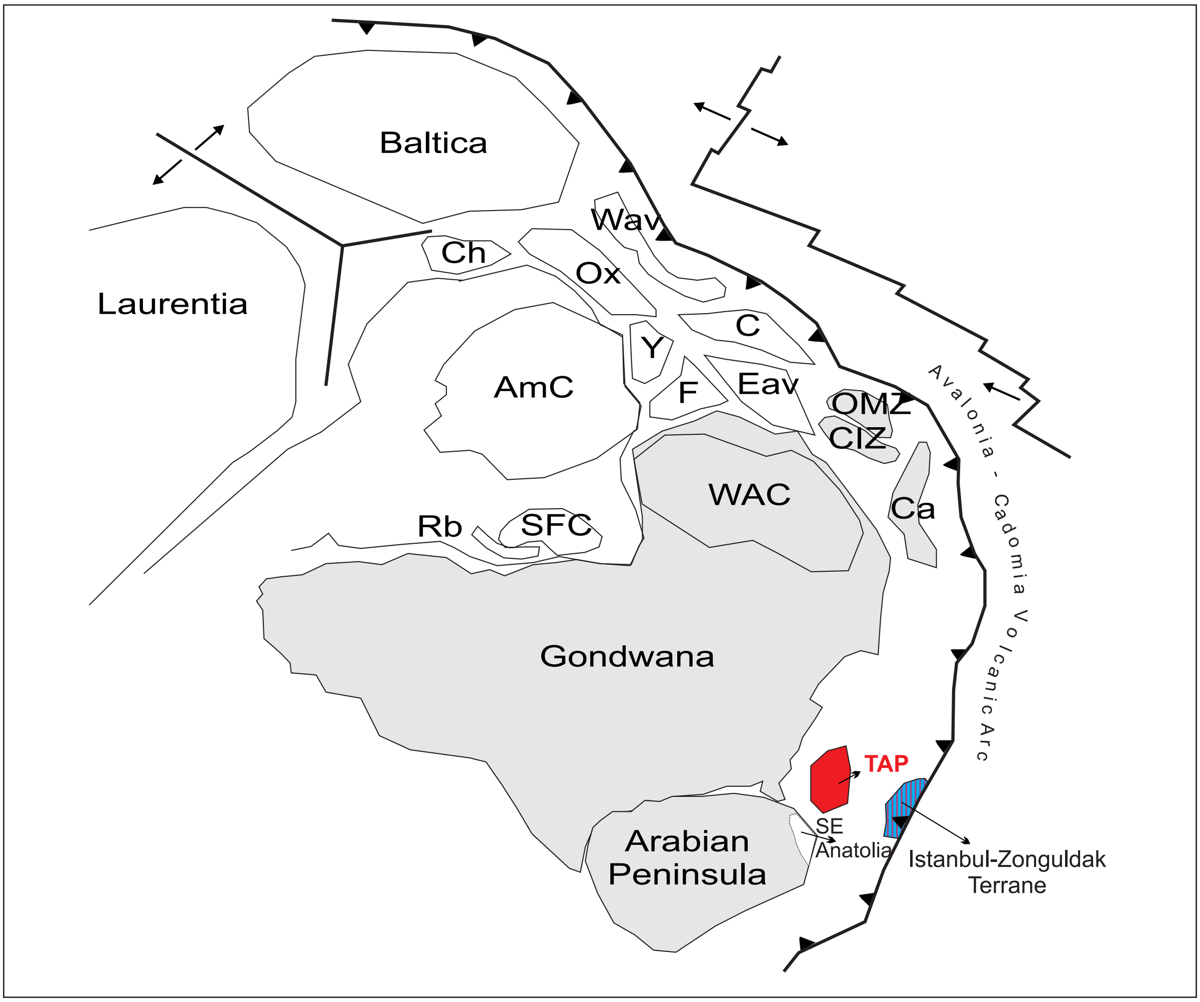

Figure 37- Paleogeographic setting of the Tauride-Anatolide Platform during the Late Neoproterozoic. AmC: Amazonia Craton, C: Carolina, Ca: Cadomia, Ch: Chortis Block, CIZ: Central Iberian Zone (Iberia), Eav: D Avalonia, F: Florida, OMZ: Ossa-Morena Zone (Iberia), Ox: Oaxaquia, Rb: Ribeira, SFC: San Fransisco Craton, WAC: B Africa Craton, Wav: West Avalonia, Y: Yukatan. (Gürsu and Göncüoðlu, 2006*b*).

that both units were affected by the same tectonic regime in Middle Permian.

ALPINE PERIOD

The first step of the Alpine cycle in KBB is generated by Early Triassic continental deposition. The fluvial clastics of Kýyýr formation overlie the KBB units with angular unconformity. In some slices this unit was deposited only in Permian, in some others, it disconformably overlies the Late Neoproterozoic basement, indicating deep erosion that removed a very thick sedimentary pile of Palaezoic rocks. This is evaluated as an important indicator of extension and uplifting in Tauride-Anatolide Platform that resulted in rifting and subsequent opening of the Ýzmir-Ankara Ocean at the end of Middle Triassic (Göncüoðlu et al., 2003; Mackintosh and Robertson, 2008). This interpretation is confirmed by late Ladinian

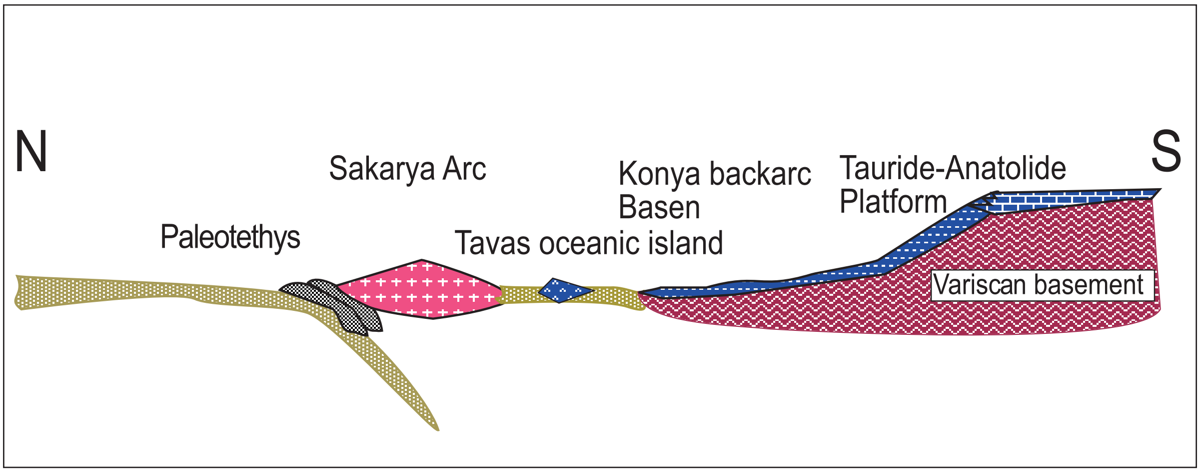

Figure 38- Carboniferous reconstruction of the Tauride-Anatolide northern margin (Göncüoðlu et al., 2004).

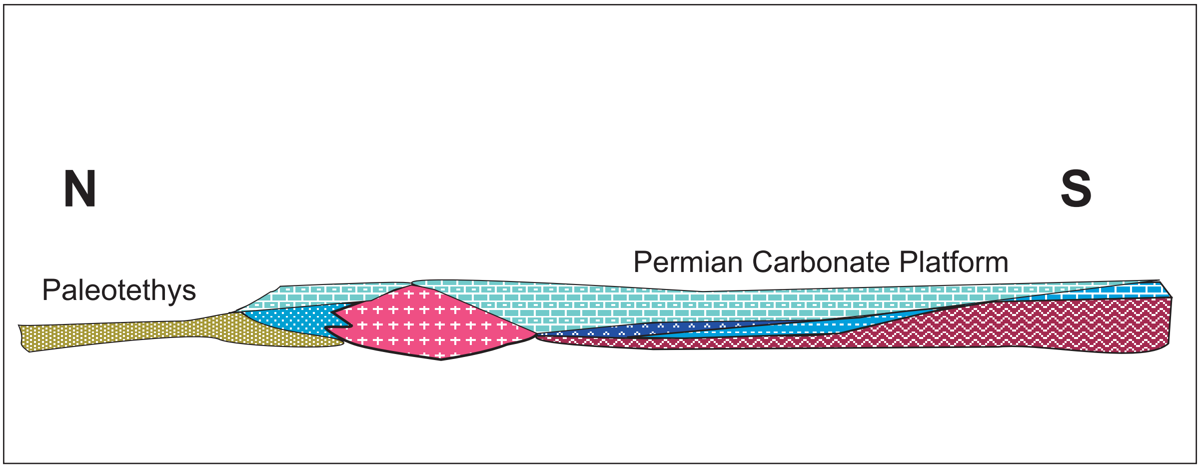

Figure 39- Middle Permian reconstruction of the Tauride-Anatolide northern margin (after Turhan et al., 2004).

(Tekin and Göncüoðlu, 2007) and Carnian (Tekin and Göncüoðlu, 2002; Tekin et al., 2002) radiolarian ages obtained from intra-pillow cherts of the Ýzmir-Ankara Ocean. Also the Triassic age (Koralay et al., 2007) obtained from granitoids intruding Menderes Core Complex and its cover, must be related with the crustal melting related to this uplifting. In the internal Tauride platform, traces of this event are lavas, intercalated with the Middle Triassic fluvial sediments (Candan et al., 2005) and Ladinian (Kaya et al., 1995) olistostromes in shallow platform carbonates, that indicate the role of extension in this deepening .

In most parts of the Tauride-Anatolide Platform during Middle Triassic (Anisian) to Late Jurassic- Early Cretaceous restricted platform, open platform, and finally slope environments respectively dominated (Figure 40). In different tectonic slices, lateral facies changes related to depositional environment and their ages show some differences. Deepening of the platform-margin generally increases from North to South in time. This leads to differences in the age of contacts of lithostratigraphic units. For example, transition to deeper pelagic sediments (Midos formation) from platform carbonates (Loras - Gökçeyayla formations) is of Malm age in the external platform. In the inner platform however, this transition is during the Abtian.

During the Middle Triassic- Cretaceous time interval, in North, between the Sakarya Continent and Tauride-Anatolide Platform, Ýzmir-Ankara Ocean is evolved (Figure 41). Although the oldest transitional type volcanism in this oceanic basin started in Carnian, the earliest Mid Oceanic Ridge basalts dated yet are late Early Jurassic-beginning of Middle Jurassic in age. Formation of MORB and hence sea-floor spreading in Ýzmir-Ankara Ocean continued until the end of Cretaceous without any interruption. Ocean island type volcanics of different ages may indicate the presence of mantle plumes under the Ýzmir-Ankara oceanic lithosphere since Triassic.

At the end of late Early Cretaceous-Late Cretaceous (Turonian-Campanian) slope sediments of KBB facing to Ýzmir-Ankara Ocean should have replaced by oceanic basin sediments. In the same time period in Ýzmir-Ankara Oceanic important changes are taken place. Yet, starting from Albian in general "supra subduction zone (SSZ)", specifically "island arc" and "back arc basin"-type volcanic rocks were formed (Göncüoðlu et al., 2006a). These formations indicate that Ýzmir-Ankara oceanic lithosphere started to break and subduct along an intra-oceanic subduction zone (Figure 41). The fact that the youngest products of the SSZ volcanism are of Cenomanian age, this intra oceanic subduction has been continuing at least since the early Late Cretaceous. Another data for the age of the intra-oceanic subduction is derived from the sub-ophiolitic amphibolites that are found in mélange units and in basement of ophiolite slices. These amphibolites of ocean island origin are metamorphosed in contact with the mantle rocks in such intra-oceanic subduction zone. The radiometric ages obtained from the amphibolites range from Albian to Campanian (Önen and Hall, 1993) that are in accordance with other findings on the initiation of intra-oceanic subduction. .

The common feature of Kaynarca, Beþdeðirmen and Koçkaya amphibolites is that they are overprinted by HP/LT metamorphism. It has been proved by Sherlock et al. (1999) that in Tavþanlý area this HP/LT metamorphism occurred in about 80 Ma. In this case, subduction in Ýzmir-Ankara Ocean and related accretion prism generation (Figure 42) must have happened at the end of Cretaceous. Blueschist metamorphosed blocks and lawsonite-glaucophane clasts are found in Maastrichtian foreland sediments. Therefore all these events related to closure were realized in about 10 Ma between the middle Campanian and middle Maastrichtian. HP/LT metamorphism is observed in all slices including Tavþanlý (Okay, 1980), Sünnüce Mountain (Göncüoðlu et al., 2000a), Yunak (Yeniyol, 1982), Koçkaya (Özgül and Göncüoðlu, 1999) and also the relatively thin continental crust slivers representing the northern margin successions of the Tauride-Anatolide platform margin. It shows that these continental units were also deeply subducted and obducted onto the platform margin, together with the subduction-accretion prism material and foreland basin deposits by forming a structural complex with allochthonous bodies derived from different tectonic settings. By this, the geographic subdivision of the KBB in a HP/LT metamorphic northern belt (Tavþanlý Zone) and a MP/LT metamorphic southern belt (Afyon Zone) by Okay (1980) contradicts with the available geological data.

The limited data yet obtained on the initial emplacement of ocean-derived material on Tauride-Anatolide continental margin is Maastrichtian in age. This data is obtained from the oldest ophiolite-bearing olistostromal sediments in Kütahya region. However by new data this age could be changed into an older age. The progression of the emplacement of oceanic crust,

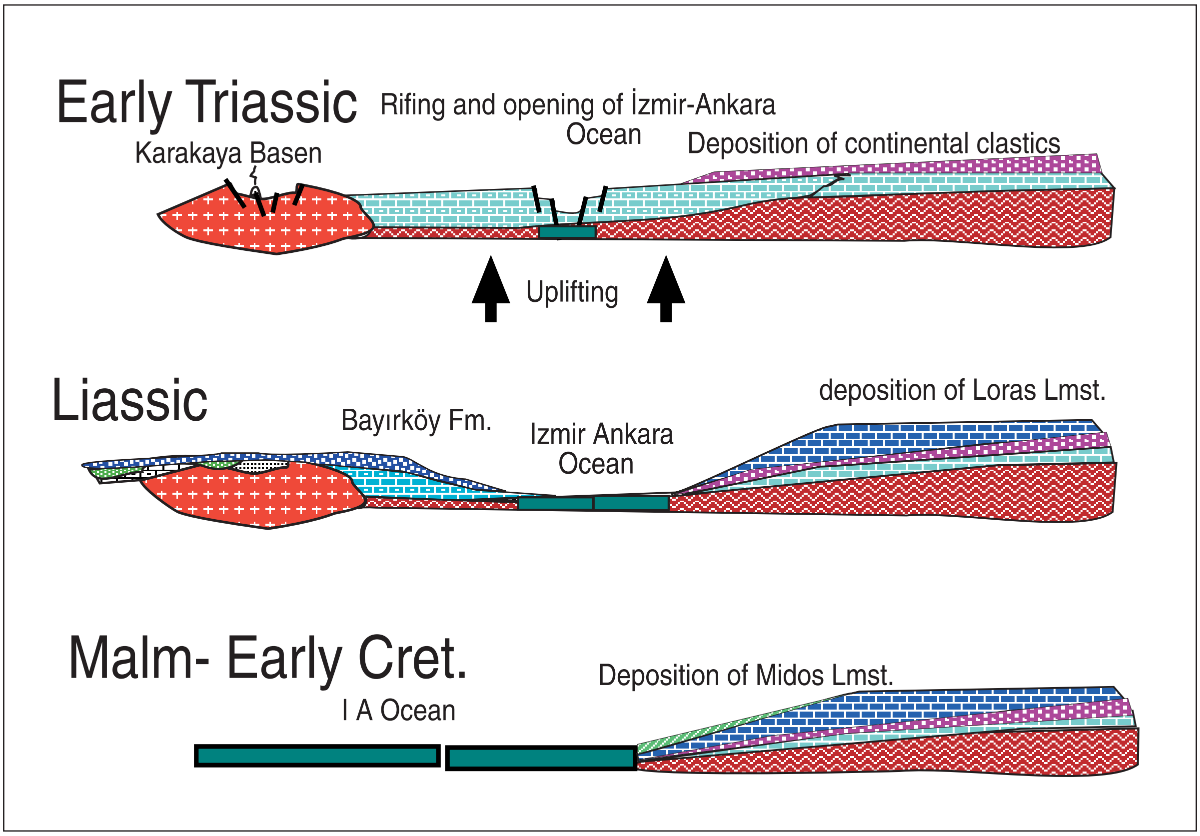

Figure 40- Triassic- Early Cretaceous reconstruction of the Tauride-Anatolide northern margin.

accretion prism and thin continental crust slices into the flyschoidal foreland basins could be continued in Early Paleocene. The first common cover of foreland, mélange and ophiolite slices is Middle-Upper Paleocene. According to this data the Alpine compression, slicing and nappe emplacement in KBB should have stopped before Middle Paleocene. During the Middle Paleocene-Middle Eocene period in remnant basins on the Tauride Anatolite Platform, terrigenous and shallow marine molasse- type sediments are deposited. The fact that the basal units are thrust over the Middle Eocene carbonates indicates another compressional episode through the belt.

RESULTS

The Neoproterozoic basement of KBB, similar to other parts of the Tauride-Anatolide Platform, includes sedimentary and volcano-sedimentary rocks and post-collisional felsic magmatic rocks that intrude them. These units are affected by deformation and low grade metamorphism prior to Lower Cambrian. The Lower-Middle Palaeozoic units are not preserved in the northern KBB. They are represented by discontinuous outcrops through W Central Anatolia in Konya, Ýzmir -Karaburun and in Tavas nappes. They include volcanic and sedimentary remnants of a marginal/back-arc basin which opened in the North of the Tauride-Anatolian Platform in Carbo-

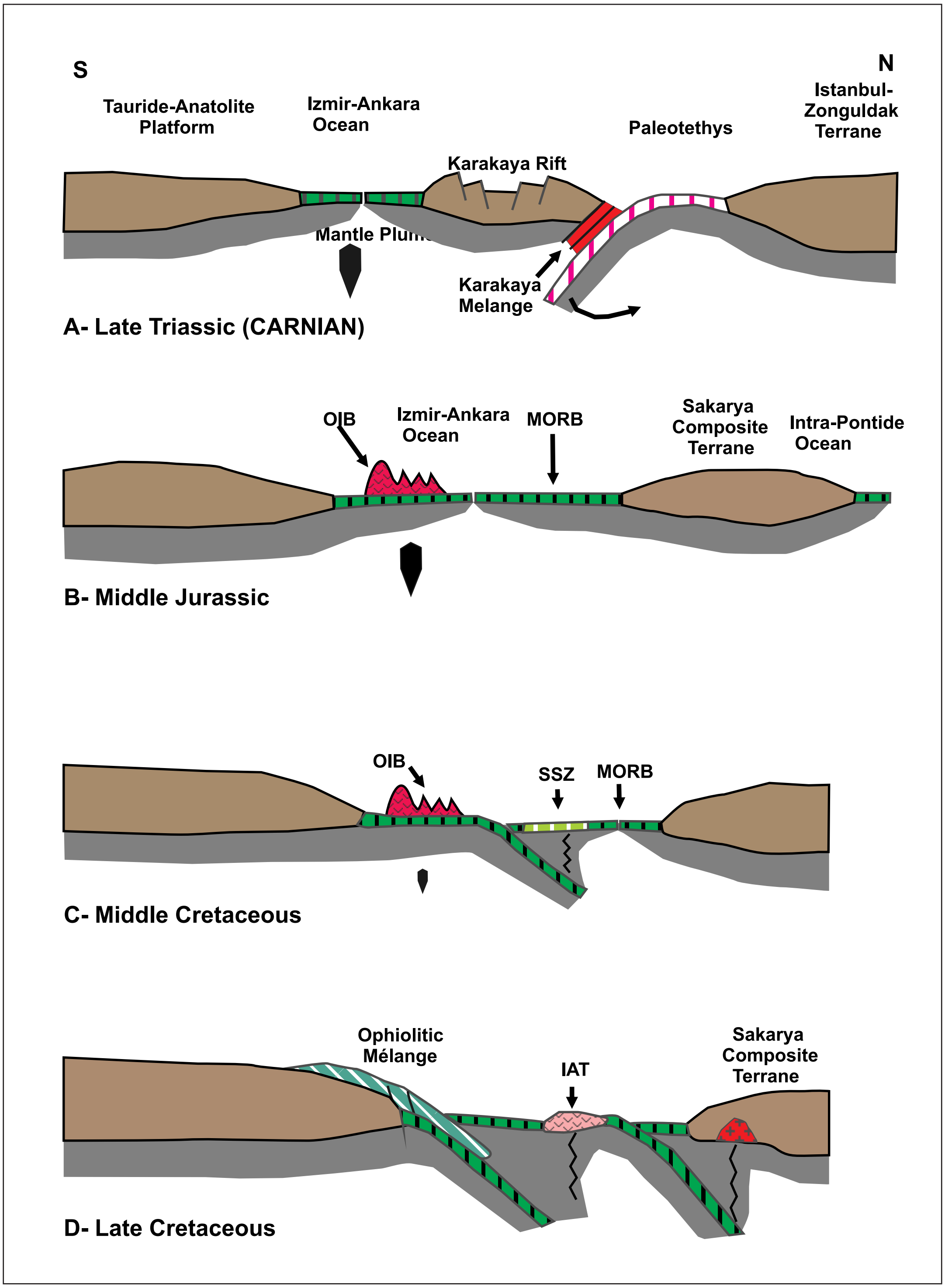

Figure 41- Mesozoic evolution of the Ýzmir-Ankara Ocean Mesozoyik evrimi (simplified after Göncü-oðlu et al., 2006*a*).

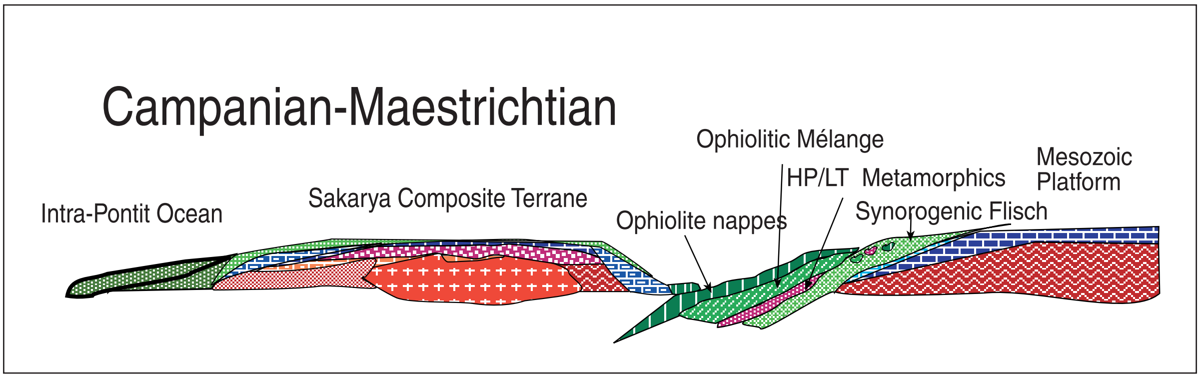

Figure 42- Late Cretaceous evolution of KBB.

niferous. Regional Middle Permian transgression observed through the belt, indicates restoration of an extensive carbonate platform by closure of this Carboniferous basin. In KBB the onset of the Alpine cycle is characterized by rapid uplifting of the basement rocks and deposition of fluvial clastics with volcanics during the Early Triassic. At the end of Middle Triassic rifting of Sakarya microcontinent from the Tauride-Anatolide platform and opening of the Ýzmir Ankara Ocean followed this uplifting. On the platform margin during the Middle Triassic - Early Cretaceous interval typical passive continental margin development has occurred. At the same time-interval the Ýzmir Ankara Ocean must have continued to spread by generating MORB-type volcanics. At the end of Early Cretaceous this convergence is replaced by divergence giving way to a N-directed intra-oceanic subduction. This subduction resulted in formation of subduction/accretion prisms, HP/LT metamorphisms, supra-subduction-type volcanism, etc. Emplacement of these oceanic material on the N margin of the Tauride-Anatolide Platform, formation of peripheral foreland basins, their closure by ongoing compression , napping and slicing of all these units and their emplacement towards S onto the southern Anatolides above what is today the Menderes Core Complex should have happened in Late Cretaceous-Early Paleocene interval. The oldest late/post orogenic overstep sequence is represented by terrigenous sediments of Middle Paleocene age.

In conclusion, KBB represents a napped/sliced belt generated by the closure of the Ýzmir Ankara Ocean and collision of the northern margin of Tauride-Anatolian Platform with the Sakarya continent. Rock-units of this belt surround Menderes Core Complex from North, East and South. They also constitute the pre-Miocene structural cover of Menderes. In NW Anatolia, tectonic units of KBB are known as: Bornova Flysch Zone, Tavþanlý Zone, Afyon Zone, Lycian Nappes, Cycladic Nappes etc. KBB is obviously not affected or partially affected from the exhumation of the Menderes Core Complex as a core complex in the Neotectonic period. So, its structural order and successions provides first hand data for understanding the alpine compressional period on the Tauride-Anatolide Platform. Thus, detailed studies that will be done on these units would shed light to the evolution of the High-grade metamorphic "massifs" such as Menderes and Central Anatolian crystalline complexes.

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REFERENCES

Akal, A., Candan, O., Koralay, E., Chen, F., Oberhaensli R., Satýr M. and Dora O.Ö., 2003. Geochemistry, geochronology and metamorphism of the Early Triassic metavolcanics in the Afyon Zone. TÜBITAK Report, 59s (unpublished).

Akdeniz, N., 1985. Akhisar-Gölmarmara-Gördes-Sýndýrgý arasýnýn jeolojisi. Ýstanbul Üniversitesi Fen Bilimleri Enstitüsü, PhD theses, 254 s. (unpublished).

_____ and Konak, N., 1979. Simav, Emet, Tavþanlý, Kütahya dolaylarýnýn jeolojisi. Maden Tetkik ve Arama Genel Müdürlüðü Report No. 6547 Ankara, (unpublished).

_____, _____ and Armaðan. F., 1980. Akhisar (Manisa) güneydoðusundaki Alt Mesozoyik kaya birimleri. Türkiye Jeoloji Mühendisliði Kongresi Bülteni, 2, 77- 90.

_____, ______, Öztürk, Z. and Çakýr, J., 1986. Ýzmir-Manisa dolaylarýnýn jeolojisi. MadenTetkik ve Arama Genel Müdürlüðü Report no: 7929, Ankara (unpublished)

Aldanmaz, E., Yalýnýz, M.K., Güçtekin, A. and Göncüoðlu, M.C., 2008. Geochemical characteristics of mafic lavas from the Tethyan ophiolites in western Turkey: implications for heterogeneous source contribution during variable stages of ocean crust generation. Geological Magazine, 145, 37-54.

Asutay, H.J., Küçükayman, A. and Gözler, Z., 1989. Daðküplü (Eskiþehir Kuzeyi) Ofiyolit Karmaþýðýnýn stratigrafisi, yapýsal konumu ve kümülatlarýn petrografisi. Maden Tetkik ve Arama Genel Müdürlüðü Bulletin, 109, 1-6.

Bacak, G. and Uz, B., 2003. Daðardý güneyi (Kütahya) ofiyolitinin jeolojisi ve jeokimyasal özellikleri. Ýstanbul Teknik Üniversitesi Bulletin, 2, 86-98 .

Bayiç, A., 1968. On the metaporphyroids of the Sýzma region, province of Konya. Maden Tetkik ve Arama Genel Müdürlüðü Bulletin, 70, 142-156.

Bragin, N.Y. and Tekin, U.K., 1996. Age of radiolarian chert blocks from the Senonian ophiolitic mélange (Ankara, Turkey). Island Arc, 5, 114-142.

Bozkaya, H., Gürsu, S. and Göncüoðlu, M.C., 2006. Mineralogical evidence for the Cadomian tectonothermal event in the western Central Taurides (Sandýklý-Afyon area), Turkey. Gondwana Research 10, 301-315.

Candan, O., Cetinkaplan, M., Oberhansli, R., Rimmele, G. and Akal, C., 2005. Alpine high-P/low-T metamorphism of the Afyon Zone and implication for the metamorphic evolution of Western Anatolia, Turkey. Lithos, 84, 102-124.

Chen, F., Siebel, W., Satýr, M., Terzioðlu, N. and Saka, K., 2002. Geochronology of the Karadere basement (NW Turkey) and implications for the geological evolution of the Istanbul zone. International Journal of Earth Sciences, 91, 469-481.

Collins, A.S. and Robertson, A.H.F., 1999. Evolution of the Lycian Allochthon, western Turkey, as a north-facing Late Palaeozoic to Mesozoic rift and passive continental margin. Geological Journal, 37, 197-138.

Çemen, I., Göncüoðlu, M.C., and Dirik, K., 1999. Structural evolution of the Tuzgolu (Salt Lake) basin: evidence for Late Cretaceous extension and Cenozoic inversion in Central Anatolia, Turkey Journal of Geology, 107, 693-706.

Çetinkaplan, M., Candan, O., Oberhänsli, R. and Bousquet, R., 2008. Pressure-temperature evolution of lawsonite eclogite in Sivriihisar; Tavsanli Zone-Turkey. Lithos, 104, 12-32.

Dora, O.Ö., Candan, O., Kaya, O., Koralay, E. and Dürr, S., 2001.Revision of "Leptite-gneisses" in the Menderes Massif: a supracrustal metasedimentary origin. International Journal of Earth Sciences, 89, 836-851

Droop, G.R., Karakaya, M., Eren, Y. and Karakaya, N., 2005. Metamorphic evolution of blueschists of the Altýnekin Complex, Konya area, south central Turkey. Geological Journal, 40, 127-153

Erdoðan, B., 1990. Ýzmir-Ankara Zonu'nun, Ýzmir ile Seferhisar arasýndaki bölgede stratigrafik özellikleri ve tektonik evrimi.Türkiye Petrol Jeologlarý Derneði Bülteni, 2/1, 1-20.

_____, Güngör, T., Özer, S. and Altýner, D., 1995. Stratigraphy and deformational style of Karaburun Belt and Ýzmir-Ankara Zone., International Earth Sciences Colloquium on the Aegean Region 1995, Gezi Kitabý, pp. 1-31.

_____, Uchman, A., Güngör, T. and Özgül, N., 2004. Lithostratigraphy of the Lower Cambrian metaclastics and their age basen on trace fossils in the Sandýklý region, southwestern Turkey. Geobios, 38, 346-360.

Eren, Y., 1993. Konya kuzeybatisinda Bozdaðlar Masifinin otokton ve örtü birimlerinin stratigrafisi. Geological Bulleten Turkey, 36, 7-23.

_____, 1996. Structural features of the Bozdaðlar Massif to the south of Ilgýn and Sarayönü (Konya). Geological Bulletin Turkey 39, 49-64.

_____, Kurt, H., Rosselet, F., and Stampfli, G.M., 2004. Palaeozoic deformation and magmatism in the northern area of the Anatolide block (Konya), witness of the Palaeotethys active margin. Eclogae Geological Helvetica, 97, 293-306.

_____, Özer, S., Sümengen, M. and Terlemez, Ý., 1978. Sarýz-Þarkýþla-Gemerek-Tomarza arasýnýn temel jeolojisi: Maden Tetkik ve Arama Genel Müdürlüðü Report no: 6546, 121p, Ankara (unpublished).

Floyd, P.A., Özgül, L., Göncüoðlu, M.C., Yalýnýz, M.K. and Winchester, J.A., 2001. Konya HP Belt Metabasalts: aspects of petrology and geochemistry. 4.Int. Geological Symposium, 24-28 September, 2001, Adana. Abstracts, 95.

_____, Özgül, L., and Göncüoðlu, M.C, 2003. Metabasite blocks from the Koçyaka HP-LT metamorphic rocks, Konya, central Anatolia: geochemical evidence for an arc-back-arc pair?. Turkish Journal of Earth Sciences, 12, 157-174.

Gökten, E. and Floyd, P.A., 2007. Stratigraphy and geochemistry of pillow basalts within the ophiolitic mélange of the Ýzmir-Ankara-Erzincan suture zone: implications for the geotectonic character of the Northern branch of Neotethys: International Journal of Earth Science 96, 725-741, (Geol. Runch.)

Göncüoðlu, M.C., 1989. Structural framework of the Anatolian Hercynides: 28th International Geological Congress, Washington, Abstracts, 1, 563-564

_____, 1990*a*. Sub-ophiolitic metamorphics at the Kütahya-Bolkardað Belt: Northern Margin of the Menderes Massif, NW Anatolia. International Earth Science Congress Aegean Regions, Ýzmir, Abstracts, 61-62.

_____, 1990*b*. Mesozoic Platform evolution of the northeastern edge of Menderes Massif, Kütahya Region, NW Anatolia: International Earth Science Congress Aegean Regions, Ýzmir, Abstracts, 162-163.

_____, 1997. Distribution of Lower Palaezoic Units in the Alpine Terranes of Turkey: paleogeographic constraints. In: Göncüoðlu, M.C. and Derman, A.S.(Eds), Lower Palaezoic Evolution in northwest Gondwana, Turkish Association of Petroleum Geologists Special Publication, Special Publication No:3, 13-24, Ankara.

_____, 2000. Restoration of an alpine foreland-thrust belt: Kütahya-Bolkardað Zone of the Tauride-Anatolide Platform, NW Turkey. Geology 2000, Wienna, April 14-17, 2000, Terra Nostra, Schriften der Alfred Wegener Stiftung, 2000/1, p 47.

_____, Özcan, A., Turhan, N., and Iþýk, A., 1992*a*. Stratigraphy of the Kütahya Region. Guide Book: A Geotraverse Across Suture Zones In NW Anatolia, 3-8, Maden Tetkik ve Arama Genel Müdürlüðü Publication, Ankara.

_____, _____, _____, Þentürk, K. and Uysal, Þ., 1992*a*. Pre-alpine events at the northern edge (Kütahya-Bolkardað Belt) of the Tauride-Anatolide Platform: 6th Geological Congress of Greece, Athens, 25-27 Mayýs 1992, Abstracts, 13-14.

_____ and Türeli, K., 1993. Petrology and geodynamic setting of plagiogranites from Central Anatolian Ophiolites (Aksaray-Turkiye). Turkish Journal of Earth Sciences, 2, 195-203.

Göncüoðlu, M.C., Erler, A., Dirik, K. and Yalýnýz, K., 1994. Sivas Baseninin batýsýndaki temelin jeolojisi ve basen birimleri ile iliþkisi. Türkiye Petrolleri Anonim Ortaklýðý Report no: 3535, 135 p.

_____, Turhan, N., Sentürk, K., Uysal, Þ.,Özcan, A. and Iþýk, A., 1996. Orta Sakaryada Nallihan-Sarýcakaya arasindaki yapýsal birliklerin jeolojik özellikleri Maden Tetkik ve Arama Genel Müdürlüðü Report no: 10094, 173 p.

_____, Dirik, K. and Kozlu, H. 1997*a*. General characteristics of pre-Alpine and Alpine Terranes in Turkey: explanatory notes to the terrane map of Turkey. Annales Ge´ologiques des Pays Helle´nique, 37, 515-536.

_____, ______, Erler A.,Yalýnýz, K., Özgül, L., and Çemen I., 1997*b*. Tuz Gölü Havzasý batý kesiminin temel jeolojýk sorunlarý: Türkiye Petrol Arama Ortaklýðý Report no: 3753, 114p.

_____, Turhan, N., Özcan, N., Þentürk, K., Uysal, Þ., Göncüoðlu, Y., Iþýk, A., and Kozur, H.W., 1998. Kütahya-Bolkardað Kuþaðýnda (Konya Kuzeyi, Orta Anadolu) Alpin öncesi olaylar: Cumhuriyetin 75. Yýldönümü Yerbilimleri ve Madencilik Kongresi, 2-6 Kasým 1998, Ankara, Abstracts, 45-46.

_____, ______, Þentürk, K., Özcan, A., Uysal, S. and Yalýnýz M.K., 2000*a*. A geotraverse across NW Turkey: tectonic units of the Central Sakarya region and their tectonic evolution. Bozkurt, E., Winchester, J. and Piper, J.A., (Eds.) Tectonics and magmatism in Turkey and the Surrounding Area. Geological Society London Special Publication 173, 139-161.

_____, ______ and Göncüoðlu, Y., 2000*b*. Vestiges of Late Palaezoic ("Variscan") events within the Tauride-Anatolide Belt, Turkey: Implications for the Paleotethyan evolution in NW Peri-Gondwana. ESF Europrobe Meeting, 30 Sept-2 Oct., 2001, Ankara. Abstracts, 24-26.

_____, Kozur, H., Turhan, N. and Göncüoðlu, Y., 2000*c*. Stratigraphy of the Silurian-Lower Carboniferous rock units in Konya area (Kütahya-Bolkardað belt, Central Turkey). VIII International Meeting of IGCP 421, Evora, 12-14 Oct., 2000, Abstracts, 227-228.

Göncüoðlu, M.C., Yalýnýz, M.K. and Floyd, P.A., 2000*d*. Petrology of the Carboniferous volcanic rocks in the Lycian Nappes, SW Turkey: implications for the Late Palaezoic evolution of the Tauride-Anatolide Platform. International Earth Science Congresson Aegean Regions 2000, Ýzmir, September.25-29, 2000, Abstracts, 213.

_____, Tekin, U.K. and Turhan, N., 2001. Geç Kretase yaþlý Orta Sakarya Ofiyolitli Karmaþýðý (KB Anadolu) içerisinde yeralan Geç Karniyen yaþlý radiyolaritli basalt bloklarýnýn jeolojik anlamý. Jeo 2000, Proceedings, CD-54-56, 6s.

_____, Turhan, N. and Tekin, U.K., 2003. Evidence for the Triassic rifting and opening of the Neotethyan Ýzmir-Ankara Ocean, northern edge of the Tauride-Anatolide Platform, Turkey. Bulletin Geological Society Italy, Special Volume 2, 203-212.

_____, Göncüoðlu, Y., Kozlu, H. and Kozur, H., 2004. Geological evolution of the Taurides during the Infra-Cambrian to Carboniferous period: a Gondwanan perspective based on new biostratigraphic findings. Geolocgica Carphatica, 55/6, 433-447.

_____, Yalýnýz, K. and Tekin, U.K., 2006*a*. Geochemistry, tectono-Magmatic discrimination and radiolarian ages of basic extrusives within the Ýzmir-Ankara Suture Belt (NW Turkey): Time constraints for the Neotethyan evolution. Ofioliti, 31, 25-38.

_____, ____ and Tekin, U.K., 2006*b*. Geochemical features and radiolarian ages of volcanic rocks from the Ýzmir-Ankara Suture Belt, Western Turkey. Proceed. Int Symp. Mesozoic Ophiolite Belts of the N Balkan Peninsula (Belgrade-BanjaLuka, 11 May-6 June, 2006) 41-44.

_____, Çapkýnoðlu, Þ., Gürsu, S., Noble, P., Turhan, N., Tekin, U.K., Okuyucu, C. and Göncüoðlu, Y., 2007. The Mississippian in the Central and Eastern Taurides (Turkey): constraints on the tectonic setting of the Tauride-Anatolide Platform. Geologica Carpathica, 58, 427-442.

Gürsu, S. and Göncüoðlu, MC., 2005. Early Cambrian back-arc volcanism in the western Taurides, Turkey: implications for rifting along the northern Gondwanan margin. Geological Magazine, 142, 617-631.

Gürsu, S. and Göncüoðlu, M.C., 2006*a*. Petrogenesis and tectonic setting of Late Pan-African metafelsic rocks in Sandýklý area (Western Turkey). Int. Journ Earth Science, 95, 741-757.

_____ and Göncüoðlu, M.C., 2006*b*. Batý Toroslarýn (Sandýklý GB'sý, Afyon) Geç Neoproterozoyik ve Erken Paleozoyik yaþlý birimlerinin jeolojisi ve petrografisi. Maden Tetkik ve Arama Genel Müdürlüðü Bulletin, 130, 29-55.

_____ and Göncüoðlu, M.C., 2008. Petrogenesis and geodynamic evolution of the Late Neoproterozoic post-collisional felsic magmatism in NE Afyon area, Western Central Turkey. In: The boundaries of the West African craton. In: ENNIH, N. and LIE´ GEOIS, J.-P. (eds) The Boundaries of the West African Craton. Geological Society, London, Special Publications, 297, 409-431.

_____, ____ and Bayhan, H., 2003. KB Gondwana'da izlenen yay-gerisi volkanizmaya bir örnek: Sandýklý (Afyon GB'sý) yöresinde yüzeylenen Erken Kambriyen yaþlý mafik volkanitlerin petrolojisi ve petrojenezi. Süleyman Demirel Üniversitesi, Mühendislik-Mimarlýk Fakültesi. 20.Yýl Jeoloji Sempozyumu, 14-16 Mayýs. 2003, Proceedings, 107-108.

_____, ____ and Turhan, N., 2005. Geology and petrology of Cadomian felsic magmatism in Afyon Area (NW), Western Central Turkey. International Earth Sciences Colloquium on the Aegean Regions 2005- Abstracts, 46-47.

_____, ______, ______ and Kozlu, H., 2004. Characteristic features of Precambrian, Palaezoic and Lower Mesozoic successions of two different tectono-stratigraphic units ýn the Taurides in Afyon area, western Central Turkey. Chatzipetros, AA and Pavlidis, SB. (eds) Proceed. 5. International Symposium on Eastern Mediterranean Geology Thessaloniki, 14-20 April 2004, 80-83.

_____, _____, Kozlu, H. and Besbelli, A., 2005. Toros-Anatolit Platformu'nda Kambriyen yaþlý metamagmatik kayaçlarýn petrografisi, petrolojisi ve petrojenez özelliklerinin belirlenmesi. Maden Tetkik ve Arama Genel Müdürlüðü Report Nr: 10759, 72.s, Ankara (unpublished).

Gürsu, S., Göncüoðlu, M.C., Kozlu, H. and Turhan, N., 2006*a*. Toros-Anadolu Kýtacýðý ve yakýn çevresinin Geç Neoproterozoyik evrimi. Proceedings 59. Geological Congree, 48-50.

_____, _____, _____ and _____, 2006*b*. Toros Kuþaðýnda Geç Neoproterozoyik-Erken Kambriyen yaþlý birimlerin litolojik özellikleri. Türkiye Stratigrafi Komitesi. 6. Workshop Proceedings, 4-5.

Kaya, O., Saadeddin, W., Altýner, D., Meric, E., Tansel, I. and Vural, A., 1995. Stratigraphic and structural setting of the anchimetamorphic rocks to the south of Tavsanlý (Kutahya, western Turkey): relation to the Ýzmir-Ankara Zone. Maden Tetkik ve Arama Genel Müdürlüðü Bulletin, 117, 5-16.

Ketin, I., 1966. Türkiyenin tektonik birlikleri. Maden Tetkik ve Arama Genel Müdürlüðü Bulletin, 66, 23-34.

Koralay, E., Candan, O., Dora, O., Satýr, M., Oberhansli, R., and Chen, F., 2007. Menderes Masifi'ndeki Pan-Afrikan ve Triyas yaþlý metamagmatik kayaclarýn jeolojisi ve jeokronolojisi, Batý Anadolu. Menderes Masifi Colloqium Ýzmir, 18-24.

Konak, N., Akdeniz, N. and Armaðan, F., 1980. Akhisar-Golmarmara -Gördes - Sýndýrgý dolaylarýnýn jeolojisi. Maden Tetkik ve Arama Genel Müdürlüðü Report no: 6916, 177p, Ankara (unpublished)

Konuk, T., 1977. Bornova Filiþinin yaþý hakkýnda. Ege Üniversitesi Fen Fakültesi Bulleten, B 1, 65-74.

Koralay, E., Dora, O., Chen, F., Satýr, M. and Candan, O., 2004. Geochemistry and geochronology of orthogneisses in the Derbent (Alaþehir) area, eastern part of Ödemiþ-Kiraz submassif, Menderes Massif: Pan-African magmatic activity. Turkish Journal of Earth Science, 13, 37-61.

Kozur, H., 1998. The age of the siliciclastic series ("Karareis Formation") of the western Karaburun peninsula, western Turkey. Paleontologia Polonica 58, 171-189.

Kozur, H.W, 1999. A review of the systematic position and stratigraphic value of Mullerisphaerida. Bollettino della Societa Paleontologia Italia, 38, 197-206.

_____, Þenel, M. and Tekin, K., 1998. First evidence of Hercynian Lower Carboniferous flyschoid deep-water sediments in the Lycian Nappes, southwestern Turkey. Geologia Croatica 51 (1), 15-22.

_____, and ____,1999. Carboniferous oceanic sequences in the Lycian nappes of southern Turkey. XIV ICCP, International Congress on the Carboniferous-Permian, Calgary, p. 79.

Kurt, H., 1996. Geochemical characteristics of the metaigneous rocks near Kadýnhaný (Konya), Turkey. Geosound, 28, 1-22.

_____ and Arslan, M., 1999. Geochemistry and petrogenesis of Kadýnhaný (Konya) K-rich metatrachyandesite: The evolution of Devonian (?) volcanism. Geological Bulletin Turkey, 41, 57-69

Manav, H., Gültekin, A.H. and Uz, B., 2004. Geochemical evidence for he tectonic setting of Harmancýk ophiolites, NW Turkey: Journal of Asian Earth Sciences, 24, 1-9.

Mackintosh, P.W. and Robertson, A.H.F., 2008. Structural and sedimentary evidence from the northern margin of the Tauride platform in south central Turkey used to test alternative models of Tethys during Early Mesozoic time. Tectonophysics, 473, 149-172

Moix, P., Beccaletto, L., Kozur, H.W., Hochard, C., Rosselet, F. and Stampfli, G.M., 2008. A new classification of the Turkish terranes and its implication for paleotectonic history of the region. Tectonophysics, 451, 7-39.

MTA, 2002. 1/500.000 ölçekli Türkiye Jeoloji Haritasý, Ýzmir Paftasý, Düzenleyen: Neþat Konak, Maden Tetkik ve Arama Genel Müdürlüðü, Publications.

Murphy, J.M., 2002. Cadomian Orogens, peri-Gondwanan correlatives and Laurentia Baltica connections. Tectonophysics, 352, 1-9.

Okay, A.I., 1980. Mineralogy, petrology, and phase relations of glaucophane-lawsonite zone blueschists from the Tavþanlý Region, Northwest Turkey. Contributions Mineral. Petrol., 72, 243-255.

Okay, A.I. and Siyako, M., 1993. The new position of the Ýzmir-Ankara Neo-Tethyan suture between Ýzmir and Balýkesir. In: (S. Turgut ed.) Tectonics and Hydrocarbon Potential of Anatolia and Surrounding Regions, Proceedings of the Ozan Sungurlu Symposium, Ankara, 333-355.

_____, Satýr, M., Maluski, H., Siyako, M., Metzger, R. and Akyüz, S., 1996. Paleo and Neo-Tethyan events in northwest Turkeey: geologic and geochronological constraints. In: Yin A. and Harrison T. M. (eds.). The Tectonic Evolution of Asia, Cambridge University Press, 420-441.

Önen, P. and Hall, R., 1993. Ophiolites and related metamorphic rocks from the Kütahya region, north-west Turkey. Geological Journal, 28, 399-412.

Özcan, A., Turhan, N., Göncüoðlu, M.C., Þentürk, K., Iþýk, and Keskin, A., 1984. Kütahya-Çifteler-Bayat-Ýhsaniye yöresinin temel jeolojisi. Türkiye Jeoloji Kurultayý 38. Bilimsel ve Teknik Kurulu Abstracts, 135-136.

_____, Göncüoðlu, M.C., Turhan, N., Uysal, Þ. and Þentürk, K., 1987. Late Palaezoic evolution of the Kütahya-Bolkardað Belt: Melih Tokay Geology Symposium, Ankara, Abstracts, 23-24.

_____, _____ and _____, 1989. Kütahya-Çifteler-Bayat-Ýhsaniye Yöresinin Temel Jeolojisi: Maden Tetkik ve Arama Genel Müdürlüðü Report no: 8974(8188), 142 s, Ankara (unpublished).

_____, _____, _____, Þentürk, K., Uysal, Þ. and Iþýk, A., 1990*a*. Konya-Kadýnhaný-Ilgýn Dolayýnýn Temel Jeolojisi: Maden Tetkik ve Arama Genel Müdürlüðü Report no: 9535, 132 p., Ankara (unpublished).

_____, _____, _____, Uysal, Þ. and Þentürk, K., 1990*b*. Late Palaezoic evolution of the Kütahya-Bolkardaðý Belt. Middle East Technical University Journal of Pure and Applied Sciences, 21/1-3, 211-220.

_____, Turhan, N. and Göncüoðlu, M.C., 1992. Field Guide to Kütahya Region. A Geotraverse

Across Suture Zones In NW Anatolia, 9-11, Maden Tetkik ve Arama Genel Müdürlüðü Publication, Ankara.

Özer, S. and Ýrtem, O., 1982. Bornova Güneyi (Ýzmir) Üst Kretase kireçtaþlarýnýn stratigrafi ve fasiyes incelemesi: Türkiye Jeoloji Kurultayý, Ankara, Bildiri Özetleri,

Özgül, N., 1976. Toroslarýn bazý temel jeoloji özellikleri. Geological Society Turkey Bulletin, 9, 65-78.

_____ and Kozlu, H., 2002. Kozan-Feke-Mansurlu arasýndaki bölgenin stratigrafisi ve tektonigine ait yeni veriler. Turkish Association Petroleum Geologists Bulletin, 14, 1-36.

Özgül, L. and Göncüoðlu, M.C., 1997. A HP/LT Neo-Tethyan sliver in the Northern Central Taurides, Turkey: Koçyaka Metaophiolitic Complex; remnant of a subducted passive continental margin: 20th Anniversy of Geological Education in Çukurova Universitesi, 30 April-3 May 1997, Adana, Abstracts, 3-4.

_____ and _____ 1998. Geology and petrology of HP/LT metamorphic rock units in Koçyaka Metamorphic Complex, Altýnekin Area, Konya: a HP/LT Neotethyan sliver in the northern Central Taurides. 3. International Turkish Geology Symposium, 31 August-4 September, 1998, Ankara, Abstracts, 277.

_____ and _____ 1999. Kocyaka metamorfik kompleksinin metamorfik evrimi: Batý Orta Anadoluda YB/DS metamorfizmalý bir tektonik birim. 52. Türkiye Jeoloji Kurultayý. Abstracts, 279-286.

Özsayýn, E. and Dirik, K., 2008. Quaternary activity of the Cihanbeyli and Yeniceoba fault zones: Ýnönü-Eskiþehir fault system, Central Anatolia. Turkish Journal Earth Science, 16, 471-492

Rimmelé, G.T., Parra, B., Goffé, R., Oberhänsli, L., Jolivet, and Candan, O., 2005. Exhumation paths of high-pressure-low-temperature metamorphic rocks from the Lycian Nappes and the Menderes Massif (SW Turkey): a multi-equilibrium approach. Journal of Petrology, 46, 641-669.

Robertson, A.H.F. and Pickett, E.A., 2000. Palaeozoic-early Tertiary Tethyan evolution in the Karaburun Peninsula (western Turkey) and Chios Island (Greece). In: Buzkurt, E., Winchester, J.A., Piper, J.D. (Eds.), Tectonics and Magmatism in Turkey and the Surrounding Area: Geological Society, London Special Publication, 173, 25-42.

Robertson, A.H.F. and Ustaömer, T., 2009. The Palaeozoic-early Mesozoic development of the Konya Complex, south central Turkey: testing of alternative subduction/accretion versus intra-continental marginal basin settings. Tectonophysics 473, 113-148.

Rojay, B., Yalýnýz, M.K. and Altýner, D., 1995. Age and origin of some spilitic basalts from 'Ankara Mélange' and their tectonic implications to the evolution of northern branch of Neotethys, Central Anatolia. International Earth Sciences Congree Aegean Regions Abstracts, Ýzmir.

Rosselet, F. and Stampfli, G., 2002. The Karaburun Units, a remnant of the Paleotethys fore-arc basin, in 1st International Symposium of the Faculty of Mines (ITU) on Earth Sciences and Engineering, 2002, Istanbul, Turkey.

Sherlock, S., Kelley, S.P., Inger, S., Harris, N., and Okay, A.I., 1999. ^{40}Ar-^{39}Ar and Rb-Sr geochronology of high-pressure metamorphism and exhumation history of the Tavþanlý Zone, NW Turkey. Contribution to Mineralogy and Petrology 137, 46-58.

Stampfli, G.M. and Kozur, H.W., 2006. Europe from the Variscan to the Alpine cycles. In: Gee, D.G., Stepherson, R.A. 2006. European Lithosphere Dynamics, Geological Society, London, Memoir 32, 43-56.

Þenel, M., Akdeniz, N., Öztürk, E.M., Özdemir, T., Kadýnkýzýn, G., Metin, Y., Özal, H., Serdaroðlu, M., and Örçen, S., 1994. Fethiye (Muðla)-Kallkan (Antalya) ve kuzeyinin jeolojisi. Maden Tetkik ve Arama Genel Müdürlüðü Report no: 9761, Ankara (unpublished).

Þengör, A.M.C. and Yýlmaz, Y., 1981. Tethyan evolution of Turket: a plate-tectonic approach. Tectonophysics, 75, 181-241.

_____ Satýr, M., and Akkök, R., 1984. Timing of tectonic events in the Menderes Massif, western

Turkey: Implications for tectonic evolution and evidence for pan-African basement in Turkey. Tectonics, 3, 693-707.

Þentürk and Karaköse, 1981. Orta Sakarya bölgesindeki Liyas öncesi ofiyolitlerin ve maviþistlerin oluþumu ve yerleþmesi. Geological Society Turkey Bulletin., 24, 1-10.

Tankut, A., 1984. Basic and ultrabasic rocks from the Ankara Melange, Turkey. Geological Society of London, Special Publications, 17, 449-454.

_____ 1991. The Orhaneli masif, Turkey.Ofioliti, 61, 702-713.

Tekin, U.K., 1999. Biostratigraphy and systematics of Late Middle to Late Triassic radiolarians from the Taurus Mountains and Ankara region. Geologisch-Paläontologisch Mitteilungen, 5, 297p.

_____ and Göncüoðlu, M.C., 2002. Middle Carnian radiolarians from the Intra-pillow limestones of the Turunc Unit, within the Gülbahar nappe (Lycian nappes, Marmaris, S Turkey: geodynamic implications, 1.Int Symp. Fac. of Mines (ITU) on Earth Sci. and Eng.16-18 May, 2002, Istanbul, Abstracts, 84p.

_____, _____ and Turhan, N., 2002. First evidence of Late Carnian radiolarian fauna from the Ýzmir-Ankara Suture Complex, Central Sakarya, Turkey: Implications for the opening age of the Ýzmir-Ankara branch of Neotethys. Geobios, 35, 127-135.

_____, _______ Yalýnýz, M. K., and Altýner-Özkan, S., 2006. Neotetis Volkanitlerinin Planktonik Fosil Faunasý (Radyolarya ve Planktonik Foraminifera) ile Yaþlandýrýlmasý, Bornova Filiþ Zonu, KB Anadolu. Final Report TUBITAK Project 103Y027, 229p.

_____ and _____ 2007. Discovery of the oldest (upper Ladinian to middle Carnian) radiolarian assemblages from the Bornova Flysch Zone in western Turkey: Implications for the evolution of the Neotethyan Ýzmir-Ankara Ocean. Ofioliti, 32, 131-150.

Tekin, U.K. and Göncüoðlu, M.C., 2009. Late Middle Jurassic (Late Bathonian-Early Callovian) radiolarian cherts from the Neotethyan Bornova Flysch Zone, Spil Mountains, Western Turkey. Stratigraphy and Geological Correlation, 17/ 3, 298-308.

Turhan, N., Gürsu, S., and Göncüoðlu, M.C., 2003. Afyon yöresinde Prekambriyen temel ve Üst Paleozoyik-Alt Mesozoyik örtüsünün Stratigrafisi ve Jeolojisi. Mersin Üniversitesi 10.Yýl Sempozyumu., Abstract, 26-27.

_____ Okuyucu, C. and Göncüoðlu, M.C., 2004. Autochthonous Upper Permian (Midian) Carbonates in the Western Sakarya Composite Terrane, Geyve Area, Turkey: Preliminary Data. Turkey Journal Earth Science, 13, 215-229.

Ustaömer, P.A., 1999. Pre-Early Ordovician Cadomian arc-type granitoids, the Bolu Massif, West Pontides, Northern Turkey: Geochemical evidence. International Journal Earth Science, 88, 2-12.

Wilson, J.L., 1975. Carbonate facies in geologic history. Springer Verl., NewYork, 265s.

Yalýnýz, M.K. and Göncüoðlu, M.C., 2005. Bornova Fliþ Zonu ve doðusunda yer alan ofiyolitik birimlerin petrojenezi. Final Report TUBITAK Project 199Y100, 74p.

_____, ____ and Floyd, P.A., 1998. Geochemistry and geodynamic setting of basic volcanics from the northernmost part of the Ýzmir-Ankara branch of Neotethys, Central Sakarya Region,Turkey. 3. International Turkish Geology Symposium, 31 August-4 September, 1998, Ankara, Abstracts, 174.

Yeniyol, M., 1982. Yunak (Konya) magnezitlerinin oluþum sorunlarý, deðerlendirilmeleri ve yöre kayaçlarýnýn petrojenezi. Ýstanbul Yerbilimleri, 3/ 1-2, 21-51.

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