Bull. Min. Res. Exp. (2015) 151: 47-106

Bulletin of the
Mineral Research and Exploration

http://bulletin.mta.gov.tr

LATE CENOZOIC EXTENSIONAL TECTONICS IN WESTERN ANATOLIA: EXHUMATION OF THE MENDERES CORE COMPLEX AND FORMATION OF RELATED BASINS

Gürol SEYİTOĞLU[a*] and Veysel IŞIK[a]

[a] *Ankara University, Department of Geological Engineering, Tectonics Research Group, 06100 Tandoğan, Ankara*

Keywords:
Extensional Tectonics,
Core Complex,
Neogene, Detachment
Fault, Graben

ABSTRACT

The Aegean region (Western Anatolia, Aegean Sea and Greece) is one of the areas of the earth under the effect of extensional tectonics and includes the typical features of core complexes in this type of region. The Menderes Massif in Western Anatolia was exhumed initially as an asymmetric core complex in the Early Miocene due to extension beginning in the Oligocene and then the central Menderes Massif was further exhumed as a symmetric core complex. This article discusses the exhumation mechanisms of the Menderes Massif and development of surrounding sedimentary basins in light of new findings. The proposed model successfully explains, the location of the Oligocene Kale basin, different movement directions of the Lycian nappes in northern and southern parts of the Datça-Kale Main Breakaway Fault and the top-to- the NNE directed shearing dominantly observed in the whole Menderes Massif.

1. Introduction

The Aegean region (Western Turkey, Aegean Sea and Greece) is one of the rapidly extending areas of the earth. The determination of a metamorphic core complex in the Aegean region (Lister et al., 1984), similar to that previously described in the "Basin and Range" region of North America with comparable tectonic characteristics, has created results deeply affecting tectonic perceptions of the region.

The similarity of structures, sedimentary basins, volcanism and underground sources developing under extensional tectonic regimes in two different and distant regions of the earth is noteworthy. Earth scientists working in these two different regions have encountered similar problems with mechanisms and initiation age of extensional tectonics, basin stratigraphy, and relationship of volcanism and tectonism. Currently in the Aegean region, where many national and international research groups continue to study, the search for answers to scientific problems continues.

This article comprises the viewpoint and assessment of the Ankara University Tectonics Research Group on the Cenozoic geology of Western Anatolia, located east of the Aegean region, containing underground sources in the form of borax, lignite, uranium, oil, gold and geothermal energy and where frequent destructive earthquakes are experienced. There are many research papers and tectonic hypotheses in the Aegean region making it impossible to explain each in detail and provide all related references. By drawing the reader's attention to this point, it may be appropriate for the reader to perform their own literature search.

1.1. Detachment Faults and Core Complex Formation

Metamorphic core complexes are the main tectonic features under the effect of large-scale crustal extension (Coney, 1980; Wernicke, 1981; 1985). Described for the first time in the "Basin and Range" region of North America, core complexes are known as "Cordilleran" metamorphic core complexes in the literature. Metamorphic core complexes comprise one or more tectonic slices represented

* *Corresponding author: Gürol Seyitoğlu, seyitoglu@ankara.edu.tr*

structurally from top to bottom by metamorphic and/or non-metamorphic lithologies, low angle normal fault(s) and basement rock units (Figure 1). The tectonic slices in the hanging wall block of the fault carry widespread effects of brittle deformation. Again metamorphic rocks have very low-low grades of metamorphism. Detachment faults (low angle normal faults), the main structural element of the core complex, and related ductile shear zones separate the non-metamorphic and/or low grade metamorphic rocks from crystalline (plutonic, high grade metamorphics) basement rocks. The basement units are generally Precambrian metamorphic rocks accompanied by Tertiary granitoid intrusions. In the literature movement of the detachment fault and related ductile shear zone are mentioned as compatible with Tertiary granitoid intrusions (Hetzel et al., 1995*b*; Işık et al., 2003*b*; 2004*a; b*).

Metamorphic core complexes are common extensional mechanisms in regions that have experienced thickening of continental crust in the past. Typical examples are described from different regions of the world (Cordillera, Aegean region, Himalayas, Alps). Metamorphic core complex studies in previous years have been completed on the areas of special interest of these basic concepts: (1) Metamorphic core complex models, (2) Low angle normal faulting (detachment faults) mechanics and geometry and (3) relationship between magmatism and extensional deformation (Fletcher et al., 1995).

Metamorphic core complex models: When the geometry and kinematics of regional scale extensional shear zones are considered, two different metamorphic core complexes are found. These are symmetric and asymmetric core complexes (Malavieille 1993) (Figure 2). Symmetric metamorphic core complexes are represented by two detachment faults with opposing directions and the ductile shear zones related to them. Middle and lower crustal rocks symmetrically outcrop along these zones. Contrary to symmetrical ones, exhumation of asymmetric metamorphic core complexes occur along a single detachment fault and related ductile shear zone.

Mechanics and geometry of detachment faults: Used by Armstrong (1972) to describe Tertiary-age normal faults, detachment faults control exposure of middle-lower plate rocks (Lister and Davis, 1989). According to Davis and Lister (1988) detachment faults have these basic characteristics: (1) Detachment faults bring non-metamorphic or low grade metamorphosed upper plate rocks into contact with high grade metamorphosed lower plate rocks. (2) Detachment faults separate uppermost younger rocks from older rocks below. (3) Detachment faults have regional scale. (4) Normal faults in the upper plate may link to detachment faults with different geometries. (5) Detachment faults show large displacements; this displacement may be on the scale of tens of kilometers. (6) Detachment faults are commonly ductile deformation structures and these are overlain by brittle deformation structures (Figure 3).

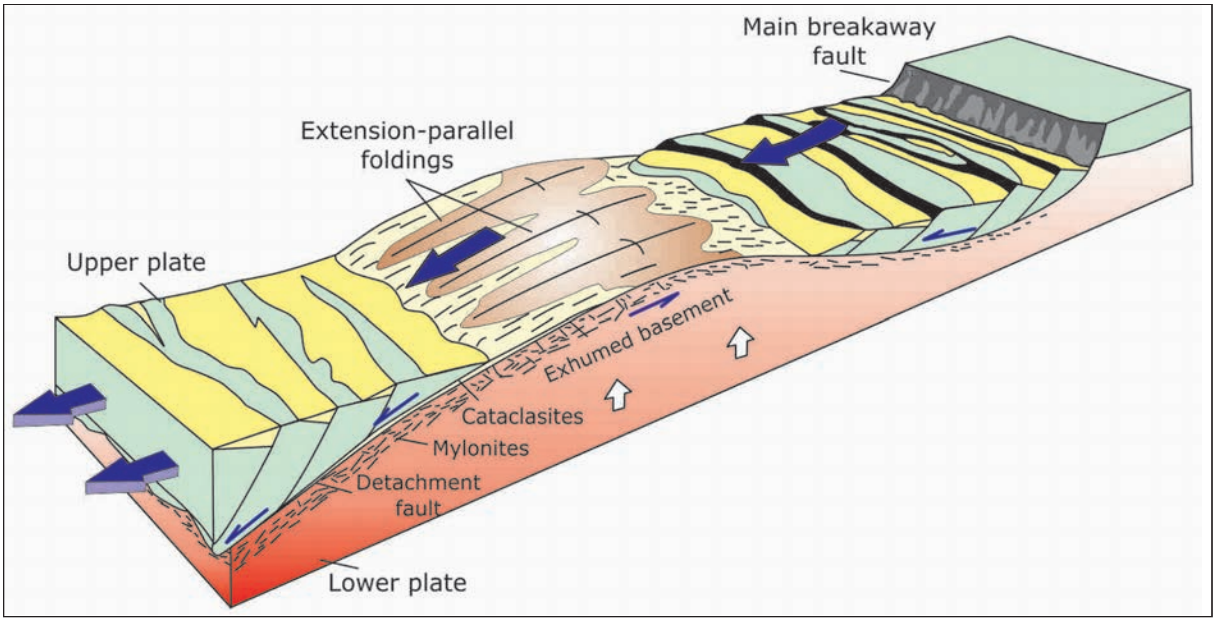

Figure 1- Typical metamorphic core complex and related structures (redrawn after Fossen 2010).

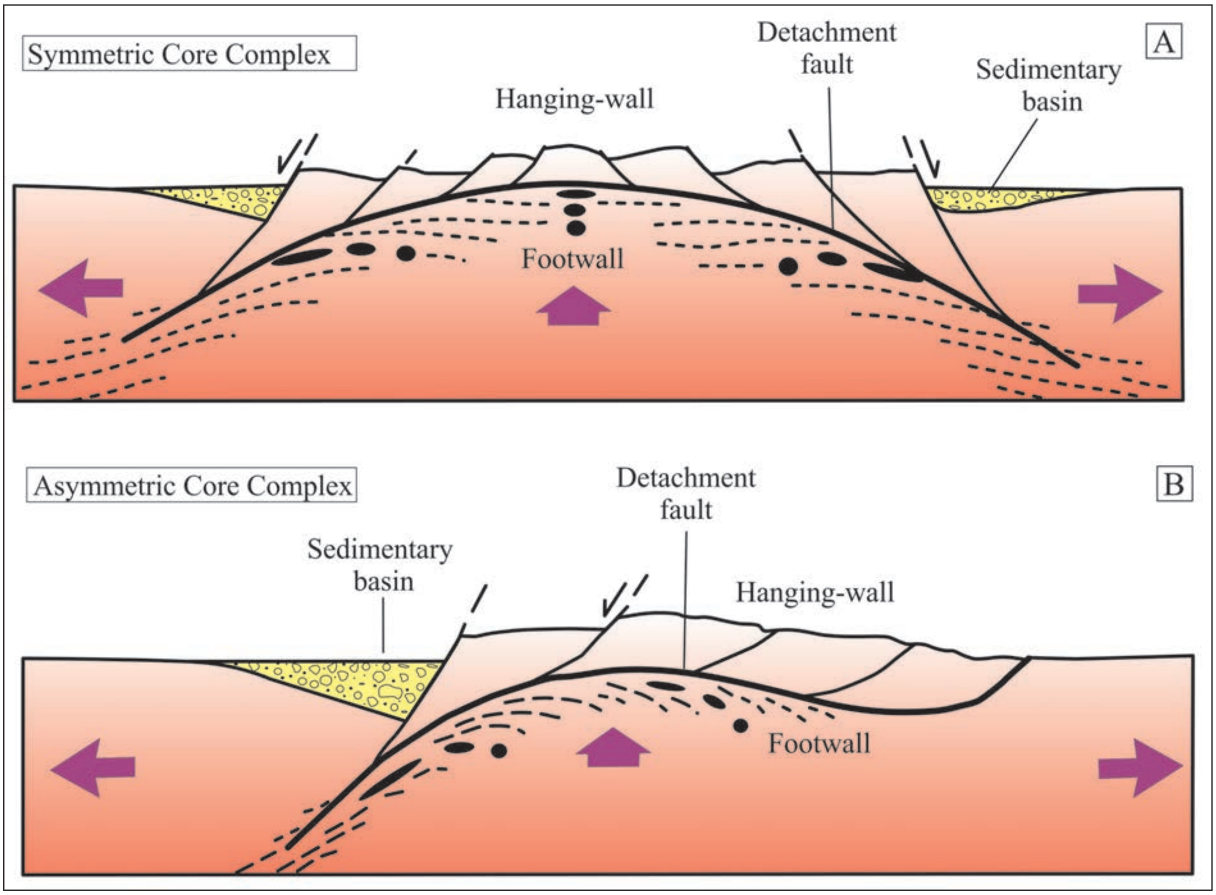

Figure 2- Metamorphic core complex models. A) Symmetric and B) asymmetric core complexes.

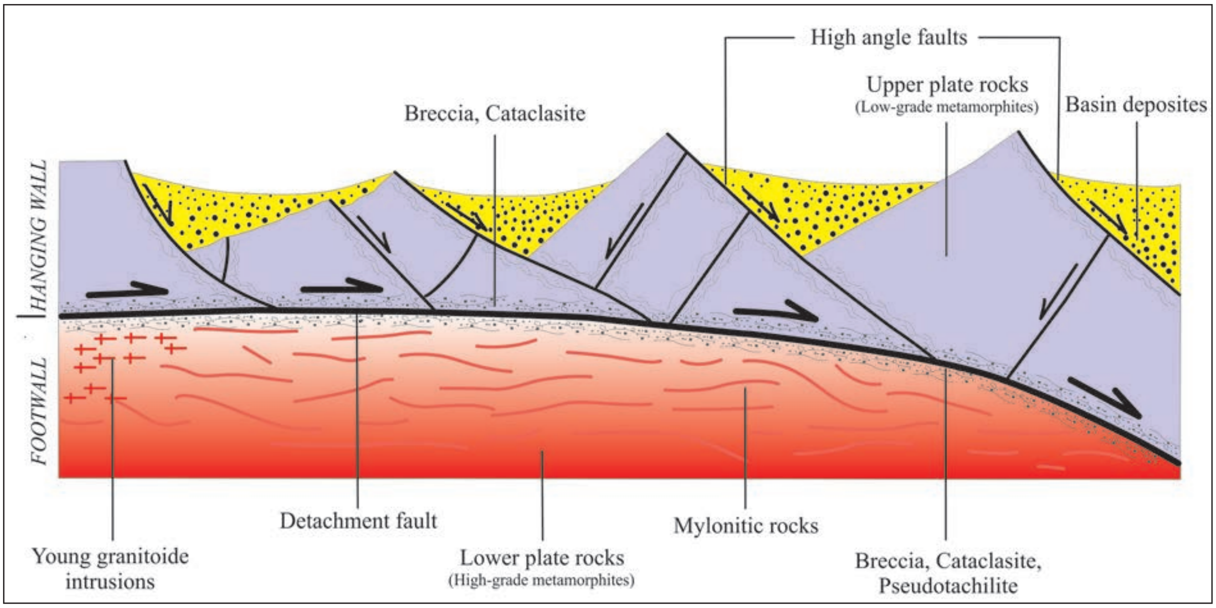

Figure 3- Cross-section of a typical detachment fault and internal structural characteristics.

While these characteristics of detachment faults are largely accepted, there is still discussion on the formation of the faulting and whether it is low-angle throughout development, in other words there is controversy over their origins. The first model of detachment faults proposed that these faults had low angles (dip lower than 30°) (Davis and Lister, 1988; Lister and Davis, 1989); however if the largest stress direction (σ_1) had a vertical direction during extension, the formation of a low-angle normal fault is not mechanically possible and thus it was reported that not every seismicity could produce such faulting. The latest developments on this topic may be seen in studies by Collettini (2011) and Prante et al. (2014).

Contrary to the "detachment faults develop with low angle" model, Buck (1988) and Wernicke and Axen (1988) recommended that detachment faults are initially high-angle normal faults and that due to doming of the footwall linked to the extensional regime, the faults become low-angle. In the flexural rotation/rolling hinge model, detachment faults in situations where the possibility of developing a sliding motion are not present, high-angle faulting develops in the hanging wall block. Thus within the system while motion continues along high-angle faults, non-active faults rotate to become low-angle. According to Buck (1988) and Wernicke and Axen (1988) the youngest faults are those with the most vertical dip.

Relationship between magmatism and extensional deformation: Many studies of metamorphic core complexes have proposed that crustal extension forms during magmatic activity and there is a genetic relationship between the two processes. According to this, rising mantle heat flow and/or mafic magma located at the bottom of the crust causes thermal softening and this initiates extension (Rehring and Reynolds, 1980; Reynolds, 1985; Gans et al., 1989; Lister and Baldwin, 1992).

Detachment faults in metamorphic core complexes typically present a "ridge and groove" structural topography (Spencer, 1982; 1984; Davis and Lister, 1988; Yin, 1991; Yin and Dunn, 1992). The undulated appearance of detachment faults and sedimentary deposition is related to (1) macroscopic geometry of banded rocks in upper and lower plates, (2) mesoscopic finite strain in upper and lower plates and (3) relative time of mylonitization. The structural characteristics representing the transition from ductile to brittle deformation in the footwall block of the detachment fault are generally related to evolution of the shear zone (Işık et al., 2003*b*). Accordingly, rocks in the ductile shear zone at middle crustal levels are carried into faulting zone in the upper section of the crust (Davis and Lister, 1988).

1.2. Extensional Tectonic Models in the Aegean Region

In the Aegean region with N-S extensional tectonics dominant, there are differing opinions on the initiation time of extension and its causes. Plate tectonic concepts were used in an attempt to explain the current structure of the Eastern Mediterranean (McKenzie, 1970) and focal mechanism solutions of earthquakes in the region were presented, forming the basis of a tectonic escape model (McKenzie, 1972; Figure 38). The tectonic escape model develops a cause-effect relationship and describes a process triggered by collision along the Bitlis Suture in Southeast Anatolia in the Middle Miocene resulting in development of the North Anatolian and East Anatolian Faults, western escape of the Anatolian plate, and concluding with the N-S extensional tectonic regime displayed in E-W grabens in Western Anatolia in the Late Miocene (Dewey and Şengör, 1979; Şengör, 1982; Şengör et al., 1985).

However, the N-S extensional tectonics in the Aegean is also explained by a back-arc extensional model. The time of subduction of the Aegean arc has key importance for back-arc extension, with the trench migrating south and southwest causing extension in the back-arc region. The initiation of subduction has been proposed as Middle Miocene (13 Ma) by Le Pichon and Angelier (1979; 1981) and as Pliocene (5 Ma) by McKenzie (1978) and Jackson and McKenzie (1988). Taking note of Cretan geology according to Meulenkamp et al. (1988) and tomographic images obtained from the subducting plate by Spakman et al. (1988), the initiation of subduction in the Aegean arc must have begun before at least 26 Ma; however for back-arc opening to form related to subduction the subducting plate must reach a certain length and accordingly the extension of the back-arc in the Aegean was stated to have begun in the Middle-Late Miocene. However, Thomson et al. (1988) have proposed that the roll-back process of the Aegean arc began to have effect in the Early Oligocene (Figure 4).

Among the reasons for acceptance of the initiation of extensional tectonics in Western Anatolia as "Late Miocene" (Şengör et al., 1985), the Late Cenozoic stratigraphy developed during lignite exploration by Turk-German cooperation in Western Anatolia played an important role (Becker-Platen, 1970;

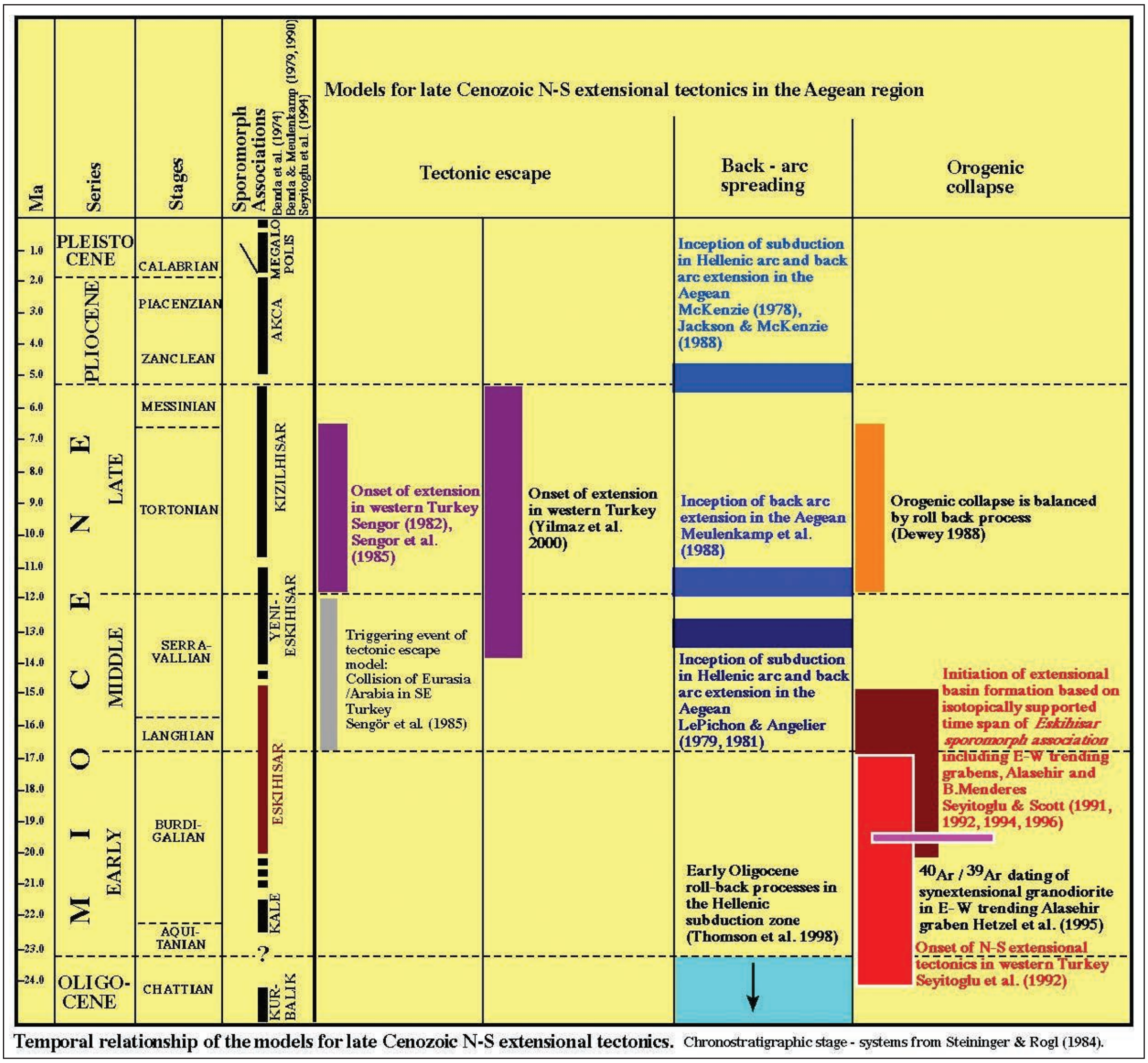

Figure 4- Temporal relationship of models proposed for Late Cenozoic N-S extensional tectonics in Western Anatolia. Chronological divisions from Steininger and Rögl (1984), age intervals of sporomorph assemblages from Benda et al. (1974), Benda and Meulenkamp (1979; 1990) and Seyitoğlu et al. (1994). The chart was adapted from Seyitoğlu and Scott (1996), by including data from Yılmaz et al. (2000) and Thomson et al. (1998).

1971). This study described the Neogene sequence in Western Anatolia as, from bottom to top, Upper Miocene Turgut member, Upper Miocene-Lower Pliocene Sekköy member, Lower Pliocene Yatağan member and Upper Pliocene-Lower Pleistocene Milet member. At the same time palynologic studies defined the sporomorph associations in Western Anatolia and related them to the above stratigraphy. According to Benda (1971) the Eskihisar sporomorph association is found between the base of the Turgut member and the lowest portion of the Sekköy member. The Yenieskihisar sporomorph association is found in the upper portion of the Sekköy member. With the possibilities presented by radiometric age dating, age dating began to be completed on volcanic rocks in Western Anatolia and the true placement of volcanic rocks and interlayered lignite levels and sporomorph associations began to be determined. From this point of view, Benda et al. (1974) represents the beginning, Benda and Meulenkamp (1979) the development and Benda and Meulenkamp (1990) the final stage. According to these studies, the age of the Eskihisar sporomorph association is determined as 20-14 Ma (Early-Middle Miocene), the age of the Yenieskihisar sporomorph association is 14-11 Ma (Middle-Late Miocene), the age of the Kızılhisar sporomorph association is 11-5 Ma (Late Miocene), the age of the Akça sporomorph association is 5-2 Ma (Pliocene), and the Megalopolis sporomorph association is younger than 2 Ma (Pleistocene) (Benda et al., 1974; Benda and Meulenkamp, 1979; 1990) (Figure 4).

The presence of Early Miocene basins in Western Anatolia were revealed by Kaya (1981) using radiometric age dates for volcanic rocks in the region; however taking note of the NNE strike of the basins, a cross-graben model (Şengör, 1987) was proposed. According to this model, north-trending basins developed similarly to Tibet-type grabens during N-S compression in the Early Miocene and from the Late Miocene the E-W trending graben system continued to develop by cutting the north-trending basins. Sediments from the north-trending basins should be found as remnants in the E-W trending graben systems. The cross-graben model (Şengör, 1987) explains both the locations of the basins developing in the Early Miocene and the opinion that the N-S extensional tectonic regime began to develop in the Late Miocene in Western Anatolia. As a result, extensional tectonics in the Aegean continues to be linked to the tectonic escape model.

By using the establishment of the sporomorph associations age intervals, especially the position of Eskihisar sporomorph association on the west Anatolian stratigraphy, Seyitoğlu and Scott (1991) stated that "Late Miocene" timing of the initiation of basin development in Western Anatolia should change, because basin development is largely beginning in the Early Miocene and it is linked to the orogenic collapse (Seyitoğlu and Scott, 1991). Required for the cross-graben model (Şengör, 1987), as described in the short explanation of the hypothesis above, it is clear that finding Early Miocene age rocks in north-trending basins is not sufficient to refute to opinion that the cause of the N-S extension in Western Anatolia is the tectonic escape model. Here the important thing is Early Miocene age data from E-W trending basins and showing that these sediments are controlled by E-W trending faults. This data was obtained from Eskihisar sporomorph association in Hasköy, noted in brief by Becker-Platen (1970; p. 174). A field study of the sedimentary unit including the Hasköy lignites checked whether it was within the Büyük Menderes graben or not and showed the E-W graben system developed in the Latest Oligocene-Early Miocene. It was clearly stated that the tectonic escape model could not be shown as the reason for N-S extension in western Anatolia due to timing inconsistencies (Seyitoğlu and Scott, 1992*a*) (Figure 4).

In the Gördes basin, one of the north-trending basins, Early Miocene central volcanics cut both the ophiolitic basement of the Izmir-Ankara suture zone and the sedimentary sequence. This situation requires that compression due to the Izmir-Ankara suture zone end before the Early Miocene, if it is correct that the Lycian nappes in southwest Anatolia are originated from the Izmir-Ankara suture zone (Ricou et al., 1975). The Menderes Massif must be free of Lycian nappe cover before the Early Miocene due to Early Miocene basins observed at Gördes and Dalama. As a result the last Lycian nappe movements, documented as continuing until the Late Miocene in southwest Anatolia (de Graciansky 1970; Besang et al., 1977), must have developed as rootless gravity slides (Seyitoğlu et al., 1992).

It has been stated that there is a relationship between extensional tectonics, thought to begin in the Late Miocene period in Western Anatolia, and the character of volcanism with calc-alkaline volcanism dominant in the Early-Middle Miocene attributed to a compressional regime and alkaline volcanism becoming widespread in the Late Miocene related to extensional tectonics (Yılmaz, 1989; 1990; Savaşçın and Güleç, 1990; Güleç, 1991; Savaşçın, 1991). However, when the beginning of extensional tectonics in Western Anatolia was pulled back to the Latest Oligocene-Early Miocene, the volcanism in the region was reexamined and the geochemical signature of calc-alkaline volcanism developing during the extensional tectonic regime was found to be inherited from previous subduction events while in the advanced period of the continuing extensional tectonics alkaline volcanism developed with thinning of the crust. The situation of calc-alkaline volcanism at the beginning of the extensional tectonic period followed by later alkaline volcanism has also been observed in the "Basin and Range" province (Seyitoğlu et al., 1992, Seyitoğlu and Scott, 1992b).

E-W grabens in Western Anatolia contain sediments from the Early Miocene period based on palynological data (Seyitoğlu and Scott, 1992a; 1996*a*; Seyitoğlu, 1992); together with radiometric age data from volcanic rocks and palynological analyses in north-south basin fill which indicate that north-south basins began to develop in the Early Miocene (Seyitoğlu et al., 1992; Seyitoğlu and Scott, 1994; Seyitoğlu et al., 1994; Seyitoğlu, 1997; Seyitoğlu and Benda, 1998), all reveal that these two differently trending basins developed simultaneously. Noting that the tectonic escape and back-arc extension models could not explain the sedimentary basin development in Western Anatolia due to age inconsistencies, an orogenic collapse model was proposed for the Early Miocene (Seyitoğlu, 1992; Seyitoğlu and Scott, 1996*b*). This proposal is different in terms of age from the Late Miocene orogenic collapse proposed by Dewey (1988) for the Aegean (Figure 4).

Koçyiğit et al. (1999) determined different structural and temporal relationships for extension in Western Anatolia and explained their findings with an episodic two-stage model. According to the model, extension in the region developed in two separate stages. The first event is related to orogenic collapse in the Early-Middle Miocene. The second event encompasses the Plio-Quaternary to the present day and is represented by normal faulting and graben formation. These two extensional stages are separated by a N-S crustal compressional period (Late Miocene-Early Pliocene). The two-stage graben model has been supported by a variety of later studies (Bozkurt, 2000; 2001; 2003; Sözbilir, 2002; Bozkurt and Sözbilir, 2004; Kaya et al., 2004; Beccaletto and Stenier, 2005; Bozkurt and Rojay, 2005).

Yılmaz et al. (2000) related the basis of extension in Western Anatolia to the tectonic escape model. According to the researchers the region was under the effects of a N-S compressional regime until the Early-Middle Miocene period. The Late Miocene-Early Pliocene (?) is a period of peneplanation. Then the region felt the effects of a N-S extensional regime. Similarly Gürer et al. (2009) advocated the presence of a NE-SW trending compression and E-W extensional regime in the Early-Middle Miocene. In the Pliocene-Quaternary period the basin formation in the region is explained within a N-S extensional regime related to the tectonic escape model.

1.3. Metamorphic Core Complex formation in the Aegean Region

Since the beginning of the 1980s the formation of metamorphic core complexes has been propounded to explain the exhumation of crystalline massifs in an extensional regime in the “Basin and Range” region of North America (Coney, 1980; Wernicke, 1981; Norton, 1986; Hill, 1987; Hodges et al., 1991; Malavielle, 1993). After Lister et al. (1984) proposed the formation of a metamorphic core complex similar to the regional extension in the “Basin and Range” province for the Aegean islands, many similar formations have been described in the Aegean region (Gibson, 1990; Gautier and Brun, 1994; Bozkurt and Park, 1994; Dinter et al., 1995; Jolivet et al., 1996; Vandenberg and Lister, 1996; Hetzel et al., 1995; Okay and Satır, 2000; Işık et al., 2001; Gessner et al., 2001). These are the core complexes of Menderes, Kazdağ, Rhodope, Cyclades and Crete (Işık et al., 2004) (Figure 5).

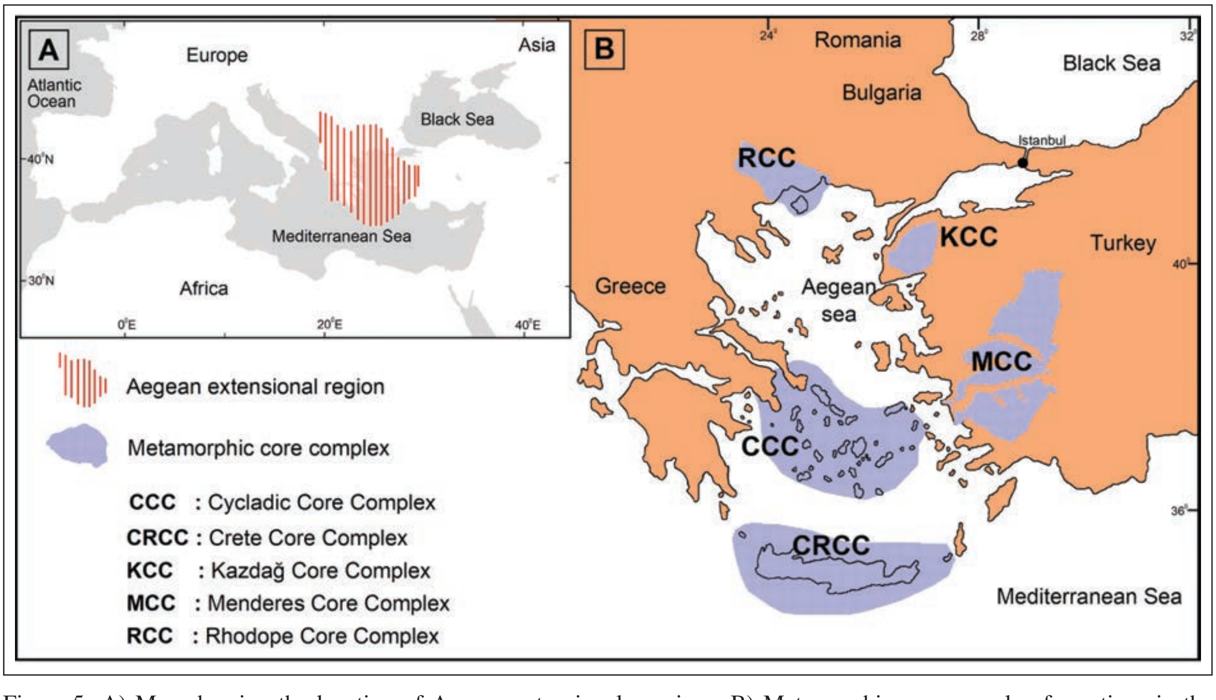

Figure 5- A) Map showing the location of Aegean extensional province. B) Metamorphic core complex formations in the Aegean region (Western Turkey, Aegean Sea, Greece) (adapted from Işık et al., 2003; 2004).

2. Menderes Core Complex

In the literature the Menderes Massif, defined as the Menderes core complex, is one of the areas with crystalline basement in the Aegean (Figure 5B). The massif is bounded to the north by the Izmir-Ankara-Erzincan zone and by the Lycian nappes to the south. While it is covered by Neogene sequences to the east, in the west it passes to units of the Cyclades massif. The Menderes Massif has a NE-SW broad dome-shaped exposure in Western Anatolia and is a complex formed of mainly metamorphic and granitic rocks.

The first geological information related to the massif was presented by Tchihatcheff (1867-1869). The first detailed geological mapping and investigation was performed by Philipson (between 1910-1915) and it was named the "Lydia-Caria massif". Philipson considered the massif a core unaffected by the Alpine orogeny and interpreted the gneisses as being Precambrian. In later years while Akyol (1924) described the massif as the "Saruhan-Menteşe massif", Ketin (1966) investigated it under the name "Western Anatolian massif". Today the widely-used nomenclature of "Menderes Massif" was used by Pajares (1944). The placement of the massif within the framework of plate tectonics and description of its evolution were completed by Şengör et al. (1984).

Since the 1950s many studies have increasingly been completed on the massif. In these studies the lithodemic nature of the units in the massif, timing of deformation and metamorphism and exhumation mechanism have all been topics of debate. Studies in the region specified the two main lithology groups that form the massif many years ago and these are widely accepted (Schuiling, 1962; Dürr, 1975; Şengör et al., 1984; Konak et al., 1987; Dora et al., 1995). However, some uncertainties still continue related to the concepts of core and cover and different opinions are common. One of these is related to the contact relationship between the two units. Some studies have determined the contact between the two units as unconformable (Çağlayan et al., 1980; Şengör et al., 1984; Konak et al., 1987; Dora et al., 1995; Candan et al., 2011). According to Şengör et al. (1984) this unconformity represents the Pan-African unconformity in the region. The character of the contact between the two units has also been defended as intrusive (Erdoğan, 1992; Bozkurt et al., 1993; 1995). Another of the uncertainties related to the units defined as core or cover is formed by their lithology, age and metamorphism. Core rocks are dominantly gneiss species and high grade metamorphic rocks. A significant portion of these gneisses are named "augen gneiss" due to their appearance. There are different interpretations of the primary rock for augen gneiss. While Schuiling (1958; 1962) determined they were originally sedimentary based on zircon morphology, Graciansky (1965) determined they were originally magmatic. Bozkurt et al. (1995) stated they were originally magmatic based on the geochemistry of the rocks. There are different opinions on the age of these gneisses. There are those who defend the gneisses as Precambrian (Şengör et al., 1984; Dora et al., 1995; Satır and Friedrichsen, 1986; Hetzel et al., 1998), in addition to those who consider them Tertiary (Bozkurt et al., 1993, 1995). Cover rocks, dominantly comprising lithologies such as schist and marble, are formed of low grade metamorphics. The limited fossil findings in these rocks are interpreted as from the interval between the Paleozoic and Tertiary (Dürr, 1978; Çağlayan et al., 1980; Şengör et al., 1984; Satır and Friedrichsen, 1986; Dora et al., 1995; Özer et al., 2001).

A variety of studies have examined the metamorphic characteristics of the massif. Just as some of these studies have interpreted that the massif was affected by a single metamorphism (Ashworth and Evirgen, 1984), some have determined characteristics of multiple metamorphism events (Schuiling, 1962; Akkök, 1983; Şengör et al., 1984; Candan, 1994; 1996; Oberhansli et al., 1997, Candan and Dora, 1998; Whitney and Bozkurt, 2002). The general view is that the character of the current metamorphism of the massif was shaped by Barrovian-type metamorphism in the Tertiary and that metamorphism developed under green schist-amphibolite facies conditions. This metamorphism is known as the "Main Menderes Metamorphism" in the literature (Şengör et al., 1984).

Since the 1990s studies in different regions of the massif have propounded the view that the massif is a core complex (Verge, 1993; Bozkurt and Park, 1993; 1994; Hetzel et al., 1995*a*, Emre and Sözbilir, 1997; Gessner et al., 2001; Işık and Tekeli, 2001; Ring et al., 2003; Seyitoğlu et al., 2004). Contrarily there are studies in the literature that do not concur that the formation is a metamorphic core complex (Okay, 2001; Erdoğan and Güngör, 2004; Westaway, 2006; Akay, 2009).

2.1. Lithological Characteristics

When we investigate the terminology of massif for the broad outcrop of the Menderes core complex in Western Anatolia, it appears to be based on metamorphic rocks with different metamorphisms and deformations and the young granitoid intrusions that cut them.

Taking note of the protolith stratigraphy of the metamorphics within this, defining them as core/Pan African basement and cover units has become a tradition. According to this, Pan-African basement is formed of Precambrian-Cambrian metasedimentary rocks and metamagmatites that have intruded them. Within this framework the basement rocks are paragneiss and schist and metamagmatite lithologies with a primary intrusive relationship to these. Cover units are Paleozoic-Mesozoic metapsammite, metapelite and metacarbonate rocks. The lower sections of the cover units are mainly represented by marble intercalated with schist and quartzite while upper sections commonly include thick marble lithologies. Within these sections it is possible to observe metabauxite levels with rudist fossils. The uppermost section is formed of pelagic marble levels (Şengör et al., 1984; Satır and Friedrichsen, 1986; Konak et al., 1987; Oberhansli et al., 1997; Candan et al.; 1998; 2011; Dora et al., 1990; 1995; 2001; Işık, 2004; Koralay et al., 2004). A variety of studies have determined the primary relationship between Pan-African basement and cover units as a regional scale unconformity (Çağlayan et al., 1980; Şengör et al., 1984; Candan et al., 2011).

Apart from metamorphics the other lithologies in the massif are local granitoid intrusions cutting the metamorphics and deformed with the metamorphics. Isotopic dating of these rocks, exposed at various scales especially in the upper and northern sections of the massif, have revealed them to be Miocene age (Hetzel et al., 1995; Delaloye and Bingöl, 2000; Işık et al., 2004*b*; Glodny and Hetzel, 2007). The intrusions are petrographically granodiorite, quartz monzonite and granite, with lesser amounts of quartz diorite and diorite. Geochemical data suggest the intrusions are generally subalkaline-peraluminum I-type (Işık et al., 2003; 2004*a, b*; Aydoğan et al., 2008; Akay, 2009).

All these lithologies were transformed at varying rates and into different types of shear zones during exhumation of the massif.

2.2. Structural Characteristics

In structural terms the Menderes Massif is a regional-scale Tertiary core complex. Within this framework, the massif represents mega, meso and micro structures developed with penetrative character in the extensional regime (Figure 6).

The megascopic structures in the Menderes Core Complex are formed of detachment faults and/or shear zones related to these (Figure 6) (Işık et al., 2003*a,b*; 2004*b*; Işık and Seyitoğlu, 2006; 2007). From south to north, these are (Figure 7): (1) Datça-Kale Main Breakaway Fault, (2) Lycian Detachment Fault, (3) Kayabükü Shear Zone, (4) Büyük Menderes Detachment fault (5) Alaşehir Detachment Fault and (6) Simav Detachment Fault. These structures are accompanied by high-angle normal faults that control the current topography of the core complex.

2.2.1. Datça-Kale Main Breakaway Fault

Datça-Kale Main Breakaway Fault is NE-SW trending normal fault zone extending from Gökova Gulf near Datça to the Denizli basin (Figure 7). It was interpreted by Seyitoğlu et al. (2004) as the fault playing a role in the initial exhumation of the Menderes core complex (see: Section 4). The Gökova graben at the southwest corner of Turkey is 150 km long and between 5 and 30 km wide. The north and south sides of the graben are bounded by faults; the south side is bounded by the Datça Fault (Kurt et al. 1999). On submarine seismic reflection profiles the Datça Fault (Kurt et al., 1999) appears to be a north-dipping listric normal fault with clear control of the sedimentary sequence on the hanging wall (Seyitoğlu et al., 2004). Investigations in the NE-SW-trending Kale basin have discovered that the basin begins with Upper Oligocene units (Hakyemez, 1989; Akgün and Sözbilir, 2001; Gürer and Yılmaz, 2002). During our investigations in the southeast section of the basin we observed the Kale-Tavas basin units are bounded by NE-SW-trending and NW-dipping normal faults. These faults are interpreted as having a genetic association with the Datça Fault located to the southwest. The footwall of the breakaway fault comprises dominant carbonate rocks of the Lycian nappes outcropping in the region and ophiolite and ophiolitic melange rocks. Basin fill is found in the hanging wall of the fault (Hakyemez, 1989; Akgün and Sözbilir, 2001; Gürer and Yılmaz, 2002).

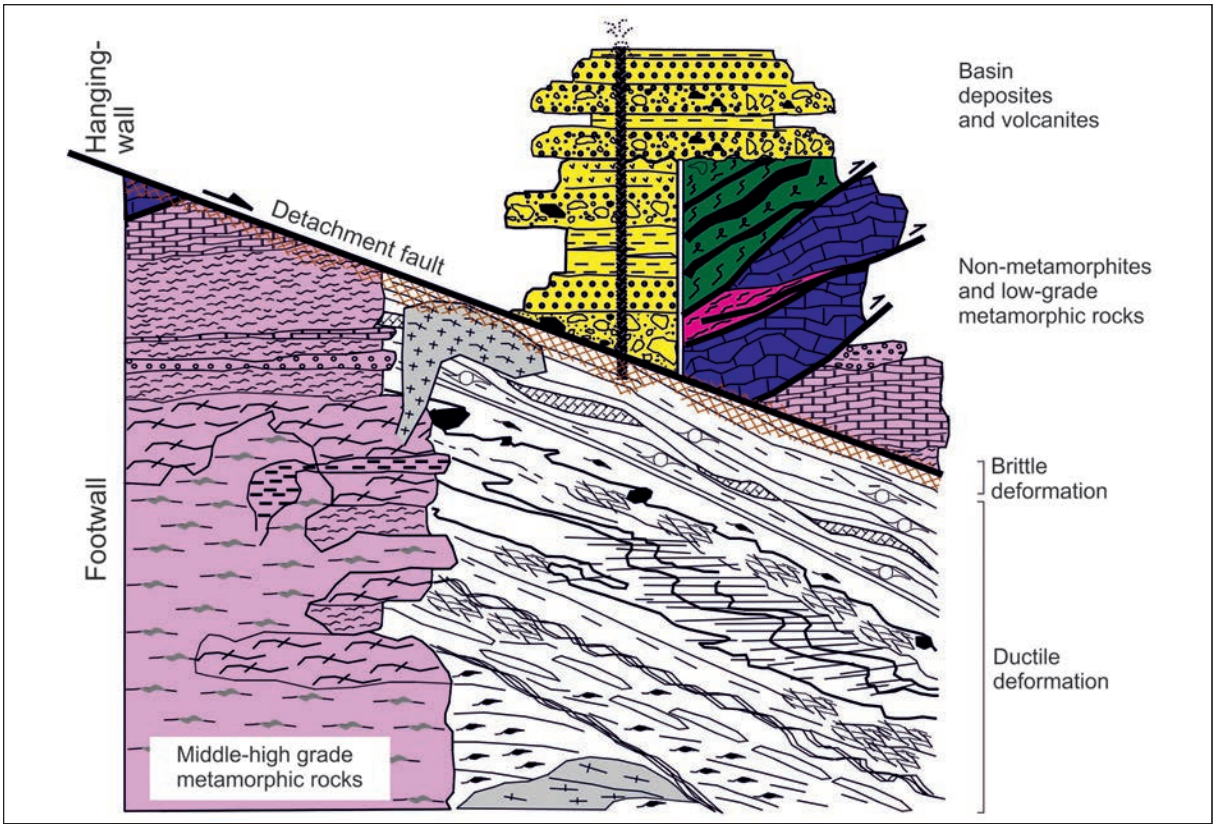

Figure 6- Cross section showing detail of a general detachment fault observed in the Menderes Metamorphic Core Complex.

2.2.2. Lycian Detachment Fault

The Lycian nappes, bordering the south and southeast of the Menderes core complex, have a special importance for the evolution of the Tethys ocean and Alpine orogeny. The Lycian nappes represent tectonic slices with different extents and tectonically overlie the relative Beydağı autochthon to the south. Basically the nappes are formed of three main units. These are; Lycian thrust slices formed of Upper Paleozoic-Tertiary aged sedimentary rocks; ophiolitic melange rocks called the Lycian melange and peridotite nappes called the Lycian ophiolite (Collins and Robertson, 1998). Though there is no full consensus related to the origin of the Lycian nappes, it is widely accepted that these tectonic slices represent the northern branch of the Neo-Tethys ocean located north of the Menderes Massif and moved from north over the massif to south (Şengör and Yılmaz, 1981; Şengör et al., 1984). It is proposed that emplacement of the nappes in the region occurred in the interval from the Upper Cretaceous to Late Miocene (Okay, 1989; Collins and Robertson, 1998; 1999). Studies in recent years which found carpholite mineral representing high pressure metamorphism at the base of the Lycian nappes (Oberhansli et al., 2001; Rimmele et al., 2003) have led to different interpretations relating to the geodynamics of the region. Rimmele et al. (2003) defined a shear zone between the Menderes Massif and the Lycian nappes. They determined three deformation events based on interpretations of the kinematics of deformation in the region. These are; (1) deformation represented by southern emplacement of the Lycian nappes; (2) main Alpine deformation and (3) deformation representing regional extension. Accordingly, the first deformation represents the transport of the Lycian nappes from the north and emplacement in the south. Later deformation represents Eocene main Alpine deformation and related metamorphism. Researchers explain northerly kinematics of this deformation as back thrusting. The deformation related to Oligo-Miocene extension is defined in the core and cover units of the Menderes Massif (Rimmele et al., 2003).

Ring et al. (2003) interpret the discontinuity between the Menderes Massif and Lycian nappes as reactivation of thrust faults forming the detachment fault. According to the researchers, the Menderes

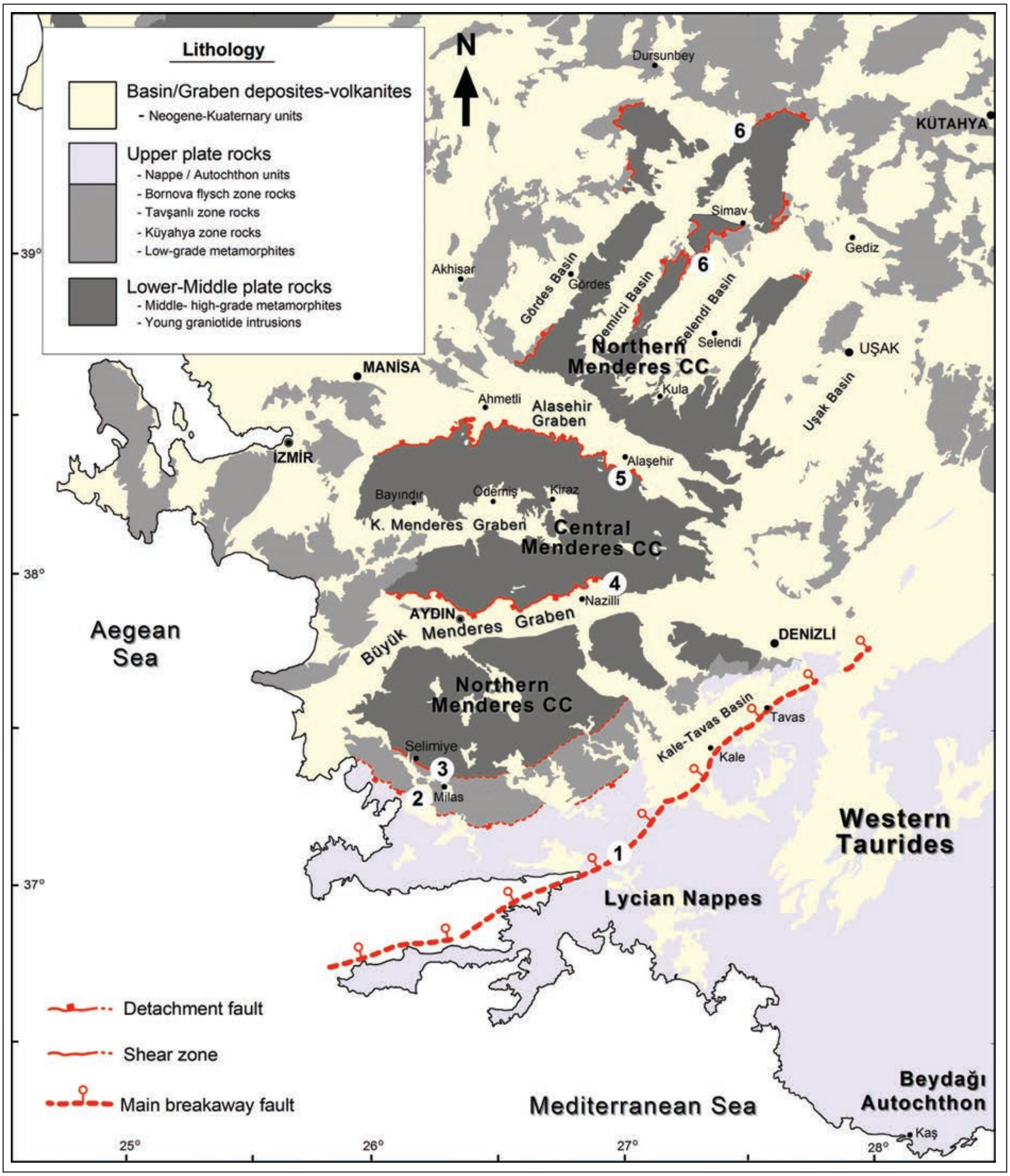

Figure 7- Detachment faults and ductile shear zones observed in the Menderes Core Complex (adapted from Işık et al. 2003, 2004). 1: Datça-Kale Main Breakaway Fault, 2: Lycian Detachment Fault, 3: Kayabükü Shear Zone, 4: Büyük Menderes Detachment fault, 5: Alaşehir Detachment Fault and 6: Simav Detachment Fault. See text for related references.

Massif is a symmetric core complex with the Lycian Detachment Fault controlling the southern section.

2.2.3. Selimiye (Kayabükü) Shear Zone

Though the Selimiye (Kayabükü) Shear zone was first described as an extensional detachment fault by Bozkurt and Park (1994), it was named the Kayabükü Shear Zone by Işık et al. (2003*a*, 2004). The extensional structures in this section have been examined in numerous studies (Bozkurt and Park, 1994; 1997*a*; 1997*b*; 1999; Bozkurt et al., 1995; Bozkurt, 2004).

The region of the Selimiye Shear Zone is important in some regards. A significant portion of the interpretations and assessments of the Menderes Massif are the result of studies in this section. The presence of the core-cover contact and character of the contact proposed within the basic stratigraphy of the massif is debatable. Many studies have interpreted the contact between the two units as the Pan-African unconformity (Şengör et al., 1984; Konak et al., 1987; Candan et al., 2011). According to Bozkurt and Park (1994), cover rocks of the Menderes Massif are separated from the core rocks by a south-dipping detachment fault with the extensional character of the contact examined in detail by later studies (Bozkurt and Park, 1997*a*; 1997*b*; Bozkurt et al., 1995; Bozkurt, 2004). The interpretation of the contact as again having intrusive character with Cenozoic-age metamagmatites representing the core rocks has been made (Bozkurt and Park, 1994). However, isotopic ages obtained from these rocks indicate the metamagmatites are Precambrian in age (Hetzel and Reischmann, 1996; Loos and Reischmann, 1999; Gessner et al., 2004).

Our field studies in the region have shown that there is no typical detachment fault between the two units and that the contact relationship has the characteristics of a ductile shear zone (Figures 7, 8). Stated differently, the contact was not affected by brittle deformation processes. The protolith rocks of the Selimiye Shear Zone comprise metamagmatite, metapsammite, metapelite and metacarbonate rocks. These rocks have been affected by upper green schist and amphibolite facies metamorphism. The multiple metamorphism features in the massif have been determined in a variety of studies (Candan and Dora, 1998; Whitney and Bozkurt, 2002; Gessner et al., 2004). Metamorphic rocks are cut by small outcrops of young granitoid intrusions.

The effects of extensional deformation are clearly observed in the rocks of the region. Our hand sample and microscopic investigations have shown widespread mylonitization in these rocks. Foliation and lineation structures are typical in mylonitic rocks. In outcrop and hand samples mylonitic foliation is represented by flattened feldspar minerals and quartz banding and mica mineral orientations. Though the development of this foliation is penetrative, it presents a heterogeneous development linked to lithology differences. Foliation planes are dominantly NW-SE trending and dip SW with dip amounts varying between 5° and 55°. These planes are found with E-W and ENE-WSW trends locally. These clear foliation planes are overprinted by secondary weak and poorly developed foliation planes. The secondary foliation has similarly oriented trend, however dips are from 30° to 70° southwest. Lineation is another typical structural component of extensional deformation. Stretching lineation is represented by lengthened minerals and mineral groups. Stretching lineation strikes NNE-SSW and NE-SW with dips SSW and SW.

Microscopic investigation has found that blastomylonite is dominant in metamorphic rocks in a wide area in the region. These rocks are accompanied by different mylonite formations. Porphyroclast composition is dominantly large feldspar grains in mylonitic rocks. Biotite, muscovite, quartz and tourmaline minerals are observed at varying sizes as porphyroclasts. Matrix composition of the rock is mainly recrystallized quartz and mica minerals. Recrystallized feldspar minerals are again observed in the matrix. In areas where crystal plasticity deformation mechanisms developed strongly, development of bands up to a centimeter thick parallel to foliation in quartz minerals is typical. In addition to band development kinking of recrystallized grains and core-shell textures are among other features of the crystal plasticity deformation mechanisms.

Mesoscopic and microscopic kinematic indicators reveal that exhumation of core rocks along the Selimiye shear zone occurred with NE oriented movement followed by SW oriented movement. Young intrusions outcropping in the region include structures representative of SW movement showing that these intrusions were affected by the late stages of extension.

Figure 8- Field view of Selimiye (Kayabükü) shear zone.

2.2.4. Büyük Menderes Detachment Fault

Büyük Menderes Detachment Fault bounds the northern side of the Büyük Menderes graben (Figure 7). This fault has also been named the Başçayır (Emre and Sözbilir, 1997) or Güney (Gessner et al., 2001*a*; Ring et al., 2003) detachment fault. With a curved geometry, the footwall of the Büyük Menderes Detachment Fault is composed of metamorphic rocks of the massif and young granitoids (Figure 9). Here the metamorphic lithologies and structural features may be correlated with the characteristics of the Selimiye shear zone. Common mylonitic orthogneiss with mylonitic paragneiss (mica gneiss, garnet mica gneiss, biotite gneiss), schist (mica schist, garnet mica schist, muscovite quartz schist, quartzitic schist, kyanite staurolite schist) and marble are contained in the footwall of the detachment fault. Mesoscopic and microscopic kinematic analysis data from mylonitic rocks have revealed two different kinematic orientations, similar to the Selimiye Shear Zone. Accordingly the footwall rocks were first affected by ductile deformation representing top-to-the NE movement linked to regional extension and then by deformation related to a top-to-the SSW movement overprinting textural features. The Büyük Menderes Detachment Fault is related to the second deformation. Hanging wall rocks are low-grade metamorphic rock masses of varying sizes and sedimentary units of the Büyük Menderes graben. Lithologies representing mylonitic gneiss are observed in local areas. The fill in the Büyük Menderes graben is dated as Early Miocene-Quaternary and consists of mainly clastics (Seyitoğlu and Scott, 1992*a*; Şen and Seyitoğlu, 2009).

There are outcrops where the detachment fault plane and slip surface are visible (Figure 10). The lithology which best preserves the plane is marble. The Büyük Menderes Detachment Fault trends NE-SW and NW-SE and dips from 10°-42°. In sections where fault striations have been preserved on the plane, the lineations are NE-SW; with a small amount NNE-SSW. The dip direction of the fault lineations is

Figure 9- Appearance in the field of the Menderes Detachment Fault. See text for details.

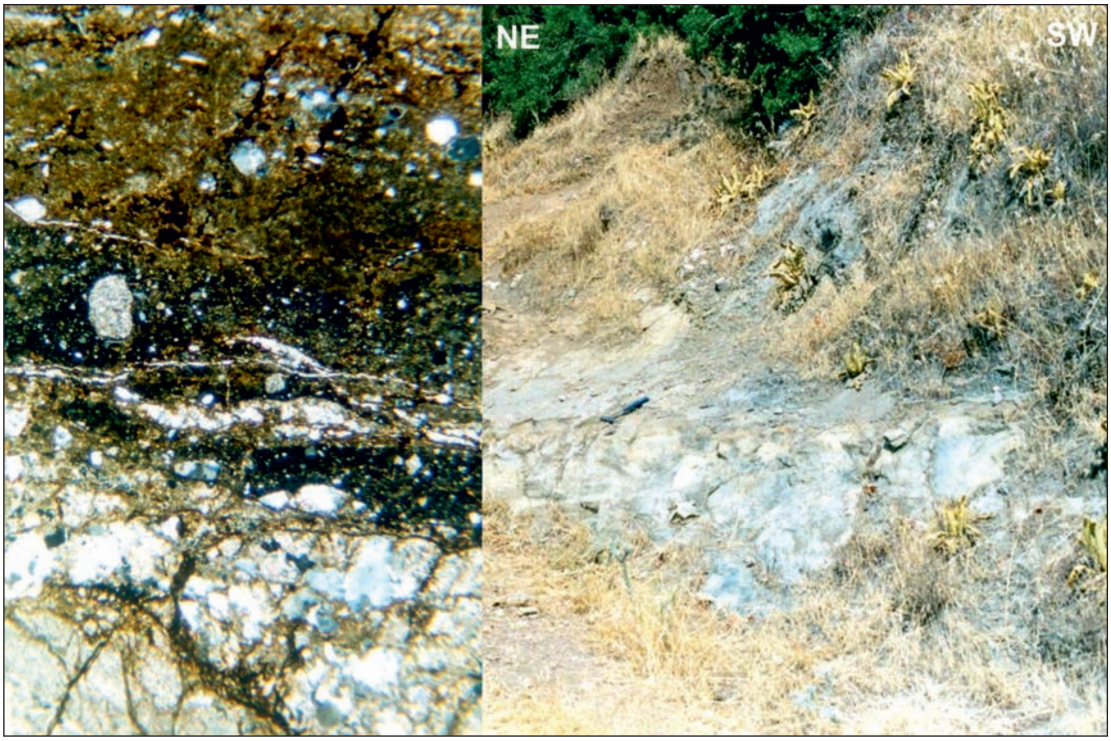

Figure 10- Appearance of brittle deformation of the hanging and foot walls of the Büyük Menderes detachment fault in the field (right). Thin section appearance of the slip plane of the detachment fault. Upper section of the photograph shows development of cataclasite on the thin section slip plane (left).

toward the SW. Along the detachment fault, rock above and below the fault plane comprises a cataclastic zone (Işık et al., 2003b) with thickness varying from 1 to 60 m. While the severity of brittle deformation is very intense in sections close to the fault plane, in areas distant from the plane this effect gradually reduces. The systematic development of the effects of brittle deformation is clearly observable, especially in the footwall. Accordingly, in sections close to the fault plane cemented (cohesive) breccia and cataclasite formations are dominant (Figure 10). The hanging wall of the fault plane has characteristic uncemented breccia and/or fault gouge.

2.2.5. Alaşehir Detachment Fault

The Alaşehir Detachment Fault is among typical detachment faults developed within the crust (Figure 11). In this regard, it has been a study topic for many research groups (Emre, 1992; Hetzel et al., 1995*a*; 1995*b*; Emre and Sözbilir, 1997; Koçyiğit et al., 1999; Gessner et al., 2001*a*; Seyitoğlu et al., 2002; Işık et al., 2003*b*; Seyitoğlu et al., 2004; Purvis and Robertson, 2005; Hetzel 2007, Öner and Dilek, 2011). The Alaşehir Detachment Fault has been described as the Allahdiyen (Emre, 1992), Karadut (Emre and Sözbilir, 1997), Çamköy (Koçyiğit et al., 1999) and Kuzey detachment fault (Gessner et al., 2001a; Ring et al., 2003) in the literature. The characteristics of the Alaşehir Detachment Fault have been determined in detail by Işık et al. (2003*b*).

The footwall of the fault is again various metamorphic rocks of the massif and syn-tectonic young granitoid intrusions (Işık et al., 2003*b*). The metamorphic rocks have similar lithological characteristics to the Selimiye Shear Zone and the footwall of the Büyük Menderes Detachment Fault. Along the detachment fault, outcrops of varying sizes of syn-tectonic intrusions are granodiorite, monzonite and granite rock types. These lithologies are accompanied by mafic inclusions in places. The main mineral composition of these rocks is feldspar, quartz and varying ratios of biotite and hornblende; secondary minerals include apatite, sphene, ilmenite and opaque minerals. Both metamorphic rocks and young intrusions show the effects of ductile and brittle deformation indicating exhumation linked to extension (Figures 12 and 13).

Ductile deformation in these rocks is characterized by widespread mylonitic formations. Apart from blastomylonite formations in metamorphic rocks, protomylonite, mylonite and ultramylonite formations are found within granitoid rocks. Typical structural elements in these rocks are mylonitic foliation and lineations, with similar textural characteristics to mylonitic foliation and lineations observed in other zones in the massif.

Along the detachment fault foliation has NE-SW orientation, with dominant dip direction to the NW. Fewer SE dip directions have been measured. Stretching lineations are typical lineation features in these rocks. They have NE-SW trend and NE dip. Kinematic data obtained along the zone (asymmetric porphyroblasts, S-C, -C' structure, oblique foliation,

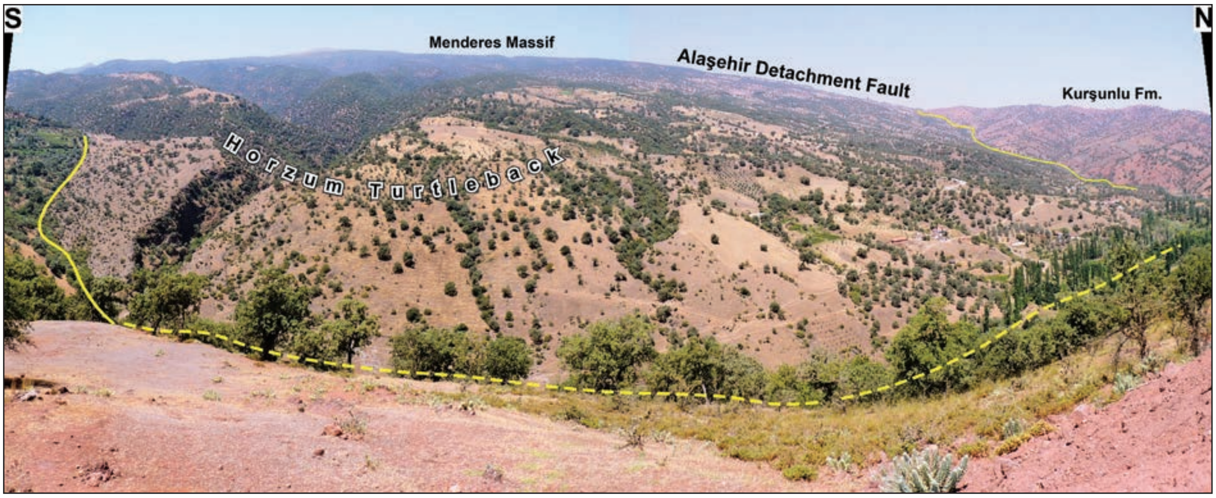

Figure 11- General appearance of the Alaşehir Detachment Fault (Kavaklıdere-Horzumkeserler road). For characteristics of the detachment fault see Işık et al. (2003b), for its role in graben development see Seyitoğlu et al. (2002) and for the "Horzum Turtleback" formation mechanism see Seyitoğlu et al. (2014).

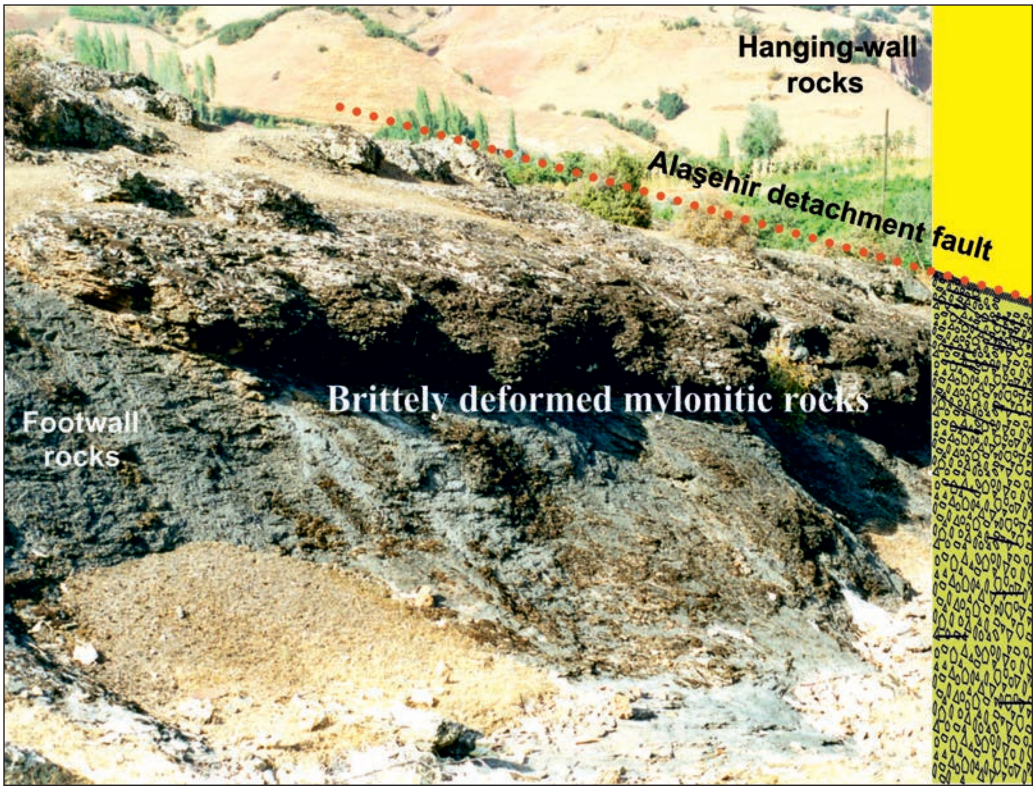

Figure 12- Appearance of brittle deformation features of the footwall of the Alaşehir Detachment Fault in the field.

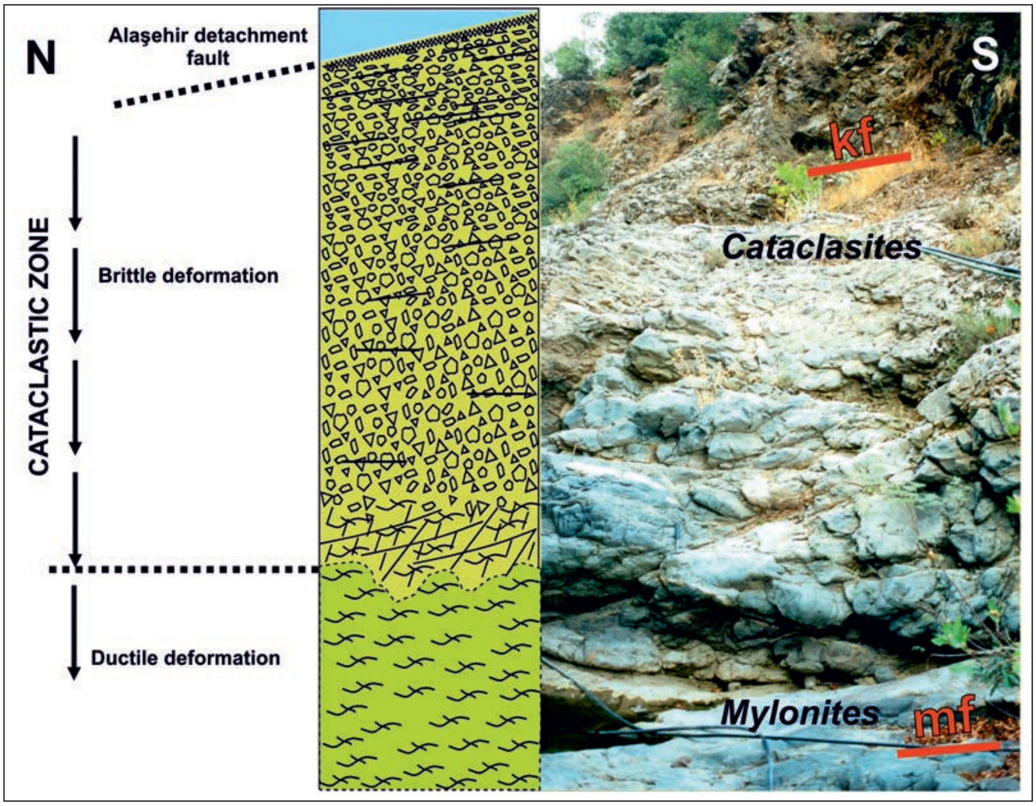

Figure 13- Appearance of ductile-brittle deformation transition on the Alaşehir Detachment Fault in the field and schematic sections of this relationship.

mica fish) indicate ductile deformation developing as a result of a top-to-the NE directed extensional regime.

In the field ductile-brittle transition is clearly observed along the Alaşehir Detachment Fault (Figure 13). Işık et al. (2003*b*) described a cataclastic zone within the detachment fault, which incorporates products of brittle deformation. The cataclastic zone is 1 to 20 m thick, and represents different brittle deformation events (Figure 12). The zone begins with systematic and non-systematic fracturing at the base, before transitioning to breccia toward the top. The textural features of the breccia levels change toward the top and a transition is seen to cataclasite-type fault rocks (Figure 12). Especially in cataclasite rocks lateral continuity is limited and rough cataclastic foliation is locally observed. Generally this foliation appears parallel to mylonitic foliation. The top of the zone is bounded by the detachment fault surface. The slip surface is locally preserved and fault striations are clear. The slip lineations are NE-SW trending and dip NE, in accordance with the stretching lineations linked to ductile deformation. This situation indicates that the Alaşehir Detachment Fault was formed by ductile and brittle stages within the same regime (Işık et al., 2003*b*).

The hanging wall of the Alaşehir Detachment Fault is mainly formed of sedimentary rocks of the Alaşehir graben. This sedimentary sequence of different formations is Early Miocene-Quaternary in age (Seyitoğlu et al., 1996*a*; Seyitoğlu et al., 2002; Şen and Seyitoğlu, 2009).

2.2.6. Simav Detachment Fault

The Simav Detachment Fault is at the very north of the Menderes core complex. The fault was first described by Işık et al. (1997) and Işık and Tekeli (2001) (Figure 14). The Simav Detachment Fault divides the moderate-high grade metamorphic rocks of the massif and young granitoid rocks in the region, from low-grade metamorphic rocks, ophiolitic melange rocks and Neogene-Quaternary sedimentary and volcanic basin sediments.

Moderate-high grade metamorphic rocks in the footwall of the fault are mainly composed of gneiss

Figure 14- General appearance of the Simav Detachment Fault in the field.

(banded gneiss, orthogneiss, biotite gneiss) and schists. Migmatitic banded gneiss outcrops in the lowest sections of the region structurally. Sequences of light- and dark-colored bands are clear, with thickness of bands from a few mm to a few cm. These metamorphics structurally transition to biotite gneiss in upper sections. Schist interlayering in the gneiss and marble, amphibolite bands and lenses at varying levels are present. Orthogneiss has similar features as those mentioned in other sections of the massif. Other lithologies in the footwall of the Simav Detachment Fault are granitoid intrusions (Eğrigöz and Koyunoba plutons). Granitoids are medium-grained, have holocrystalline granular texture in hand samples and are formed of granodiorite, granite and monzonite rock types. Occasional felsic, with lesser amounts of mafic, dikes and pegmatites are among other rocks found in the footwall.

Along the Simav Detachment Fault these rocks have been affected by differing amounts of ductile deformation, and may be compared to the Alaşehir Detachment Fault, especially. Mylonitization has affected metamorphics and granitoids by varying amounts. Mesoscopic and microscopic investigation has provided important information related to mylonitization along the Simav Detachment Fault (Işık and Tekeli, 1998; 1999; 2001; Işık, 2004; Işık et al., 2004). Mylonitic foliation and lineation are characteristic. In gneiss and schist rocks mylonitic foliation is represented by quartz banding, recrystallized quartz and mica minerals and preferential orientation of biotite, plagioclase, sometimes muscovite, sillimanite, kyanite and tourmaline porphyroblasts. In marbles foliation is visible as recrystallized calcite and/or calcite prophyroclasts with preferential stretching. In pegmatites a foliation formed by stretched recrystallized quartz and muscovite grains is typical. In granitoids mylonitic foliation is generally represented by recrystallized quartz and biotite minerals. The measured foliation planes are dominantly NW-SE striking and slope to the SW. Other sections of the foliation measurements are NE-SW striking and slope to the SE. The mean dip amount is 27°. Stretched quartz, feldspar, kyanite, tourmaline, hornblende and mica are common minerals and groups that form the mylonitic lineation (Figure 15). The lineation direction is NE-SW. The dominant dip direction of these lineations is SW, with fewer NE dips measured. In this section of the massif, there is NW-NE stretching lineation found. The dip

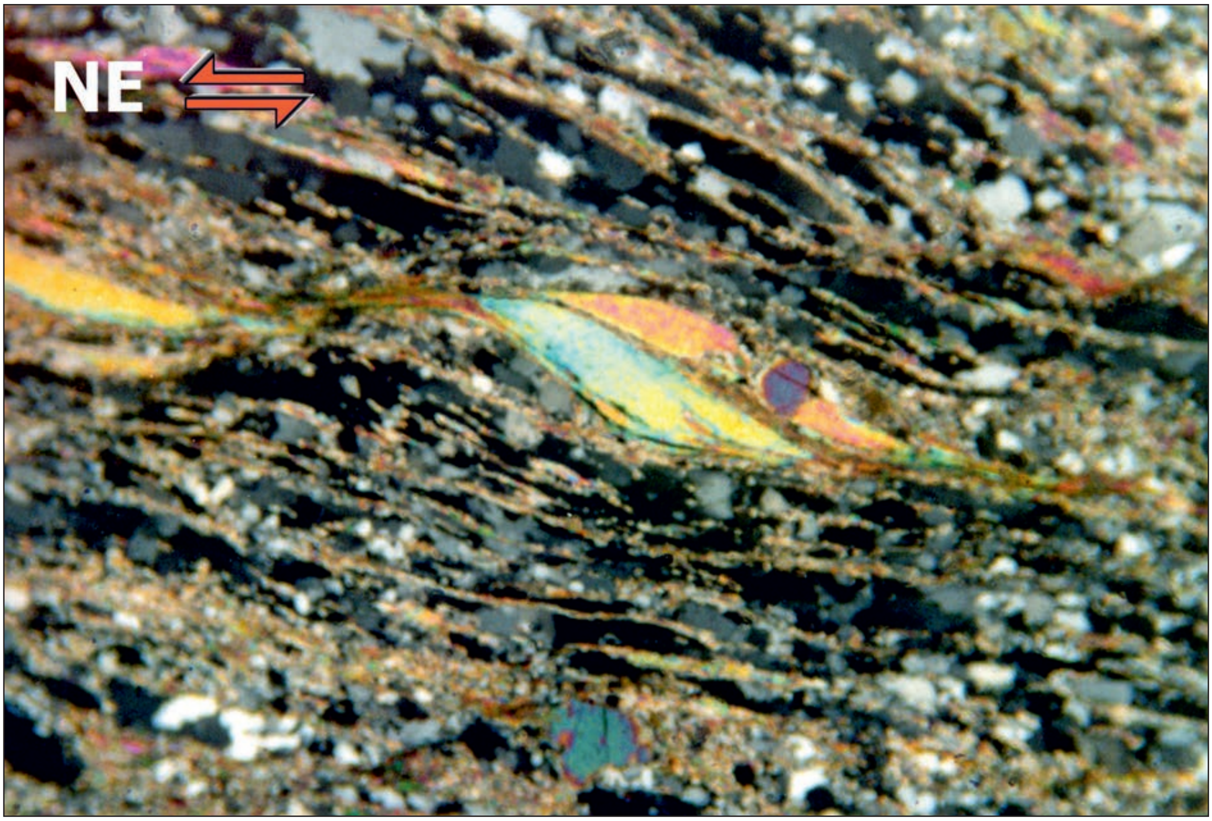

Figure 15- Typical mica fish appearance on oriented thin section. Movement direction top-to-the NE.

direction of these lineations is SW. Perpendicular to mylonitic foliation and on parallel surfaces to lineation, there are common kinematic indicators (asymmetric porphyroclasts, oblique recrystallization, mica fish, S-C and S-C' fabrics, shifted grains and V-pull apart micro textures). These indicators show the extensional tectonics forming the Simav Detachment Fault represented a top-to-the N-NE directed movement (Figures 15 and 16).

Brittle deformation structures in the Simav Detachment Fault are defined by a cataclastic zone. The apparent thickness of the zone reaches 100 meters, but typical features are observed within 30 meters. Mylonitic rocks within the cataclastic zone have fractured outcrops and hand samples, small-scale faulting, fragmentation, crushing and alteration features. In sections close to the fault plane there are high rates of crushing and milling with cataclasite and local breccia fault rocks observed with fracturing, mesoscopic faulting and breccia formation in sections slightly further from the plane.

The hanging wall of the Simav Detachment Fault within the north of the Menderes core complex is represented by metamorphic and non-metamorphic lithologies, generally formed of allochthonous rock units and Neogene-Quaternary basin sediments. Especially in low grade metamorphics and non-metamorphic rocks, the effects of varying amounts of brittle deformation are commonly observed. These rocks are characterized by high-angle faults of mappable scale with WNW-ESE and NW-SE directions. The normal component is clear on surfaces with relative movement visible. The movement direction is mainly toward the NE. However, there are a few SW dipping faults found. Some faults have clear listric geometry.

3. Neogene Basins in Western Anatolia

3.1. East-West-Trending Grabens

One of the dominant morphological features in Western Anatolia is E-W-trending grabens (Figure 7). The most comprehensive explanation of the basis of Turkey's neotectonics at time of publication, in the article by Şengör et al. (1985) the E-W grabens represented the initiation of N-S extensional tectonics in the Aegean (revolutionary structures). After Early Miocene ages were obtained from E-W graben fill (Seyitoğlu and Scott, 1992*a*; Seyitoğlu, 1992),

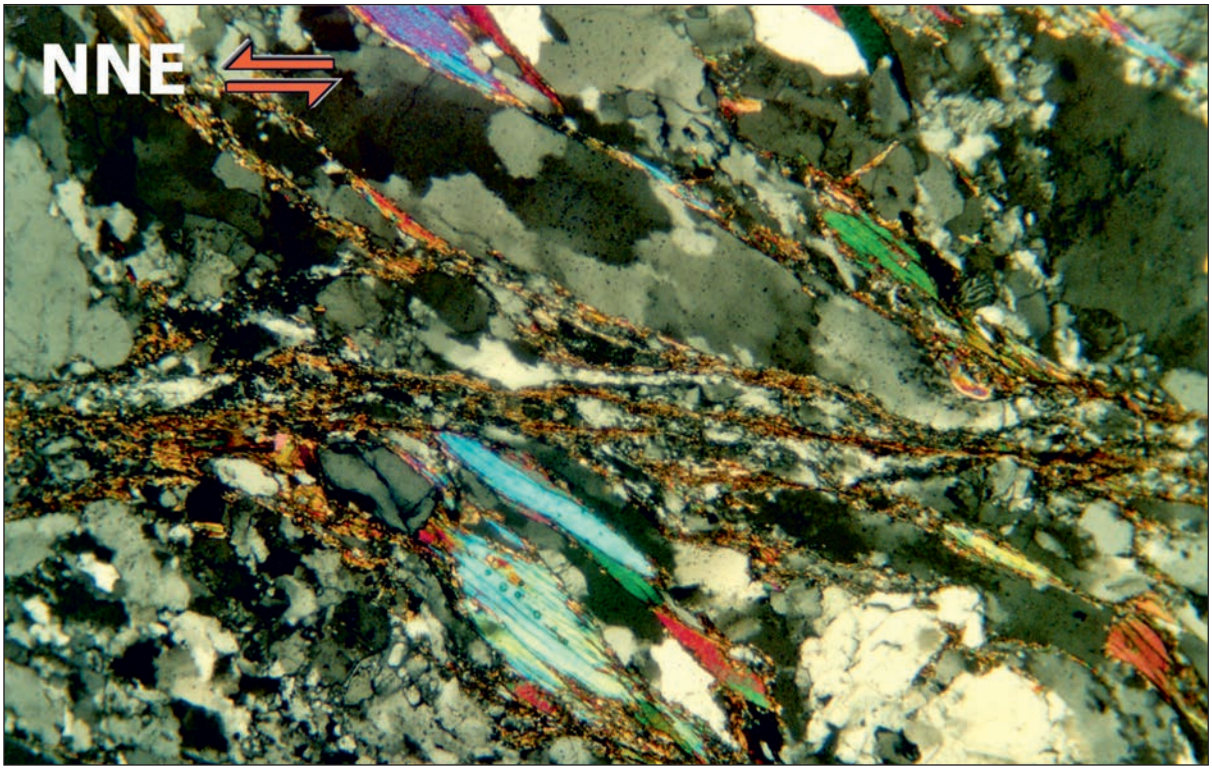

Figure 16- S-C fabric on oriented thin section. Horizontal foliation represents C-plane, diagonal foliation represents the S-plane. Movement direction is top-to the NNE.

sedimentological investigation determined whether graben fill in E-W grabens belonged to them and not pre-existing structures, a matter significantly emphasized by Cohen et al. (1995). This was because the data obtained from this fill determined the initiation time of extensional tectonics in Western Anatolia and was used to test the validity of proposed tectonic models (Seyitoğlu, 1992). If a section of the sedimentary sequence within the grabens is revealed to belong to cross grabens as proposed in the cross graben model, it would question the validity of age data obtained from the fill for initiation of E-W grabens. In a sedimentological study of both Alaşehir (Gediz) and Büyük Menderes sediments by Cohen et al. (1995), the fill in both grabens were coeval with the E-W graben system and they determined that the ages obtained from graben fill could be used for graben formation (Cohen et al., 1995; p. 637). Contrary to this, interpretation of the deepest graben fill as belonging to north-trending basins continues in some studies (Yazman, 1997; Yılmaz et al., 2000; Yılmaz and Gelişli, 2003; Gürer et al., 2009).

3.1.1. Alaşehir Graben

Grabens are generally named after the river they contain; as a result the name Gediz graben is commonly used. However, it is thought that using the name Alaşehir graben is more appropriate than Gediz graben (Seyitoğlu, 1992) because; (1) the town of Alaşehir is located within the graben while the town of Gediz is not within the graben, (2) the Gediz river does not run along the whole graben but enters in the middle of the graben at Adala, (3) the surface ruptures of the 28 March 1970 Gediz earthquake (M=7.0) developed outside the graben near the town of Gediz, and (4) Alaşehir stream, Alaşehir town and surface ruptures of the 28 March 1969 Alaşehir earthquake (M=6.1) are within the graben (Arpat and Bingöl, 1969; Ketin and Abdüsselamoğlu, 1969; Eyidoğan and Jackson, 1985).

While studies related to basin fill in the Alaşehir graben have proposed many formation names, it is not possible to observe or map some of these throughout the whole graben. In the field three different formations are clearly distinguished and their relationships to faults are observable (Figure 17). The lowest Alaşehir formation and the Kurşunlu formation above it are unconformably followed by the Sart formation and the sequence ends in current alluvial sediments (Figure 18) (Seyitoğlu et al., 2002).

The first two sedimentary sequences in the Alaşehir graben, the Alaşehir and Kurşunlu formations, are bounded by the Alaşehir Detachment Fault (Fault I). In the lower sections of the Alaşehir formation (İztan and Yazman, 1990) angular blocky conglomerates are found. The clasts are formed of mylonitic rocks. Toward the top of the unit it transitions to yellow sandstone-mudstone intercalations. Generally within 50 m toward the top the clast size rapidly reduces and angular blocky conglomerates up to 1.5 m are found within very fine-grained lacustrine sediments. The upper section of the formation passes to very well lithified laminated mudstone rich in organic material passing into sandstone and limestone with conglomerate levels observed. Grey sediments indicating underwater environmental conditions intercalate with red sediments, before passing into a unit dominated by

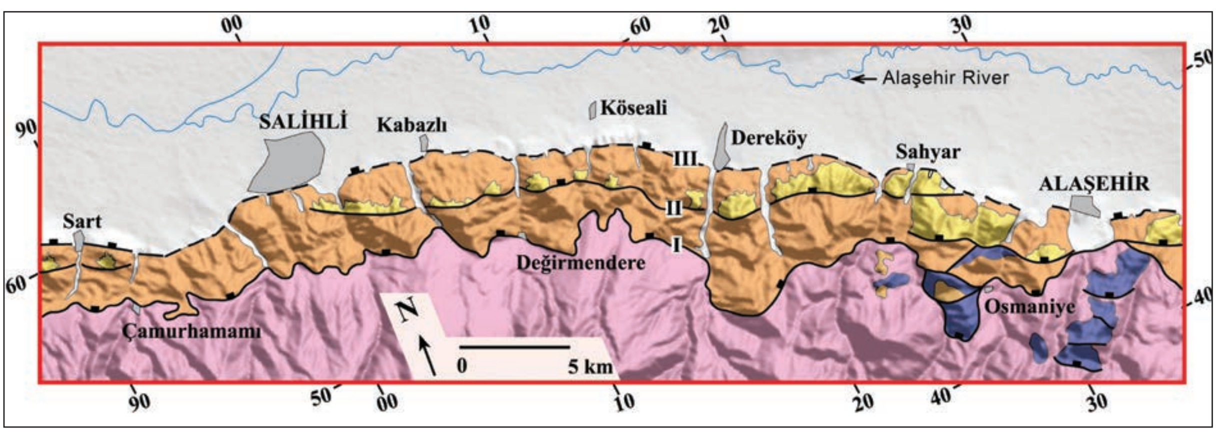

Figure 17- Geological map of the Alaşehir graben (taken from Seyitoğlu et al., 2000). Pink represents metamorphic basement, blue represents the Alaşehir formation, dark yellow-orange represents the Kurşunlu formation, light yellow represents the Sart formation and gray represents Quaternary alluvium.

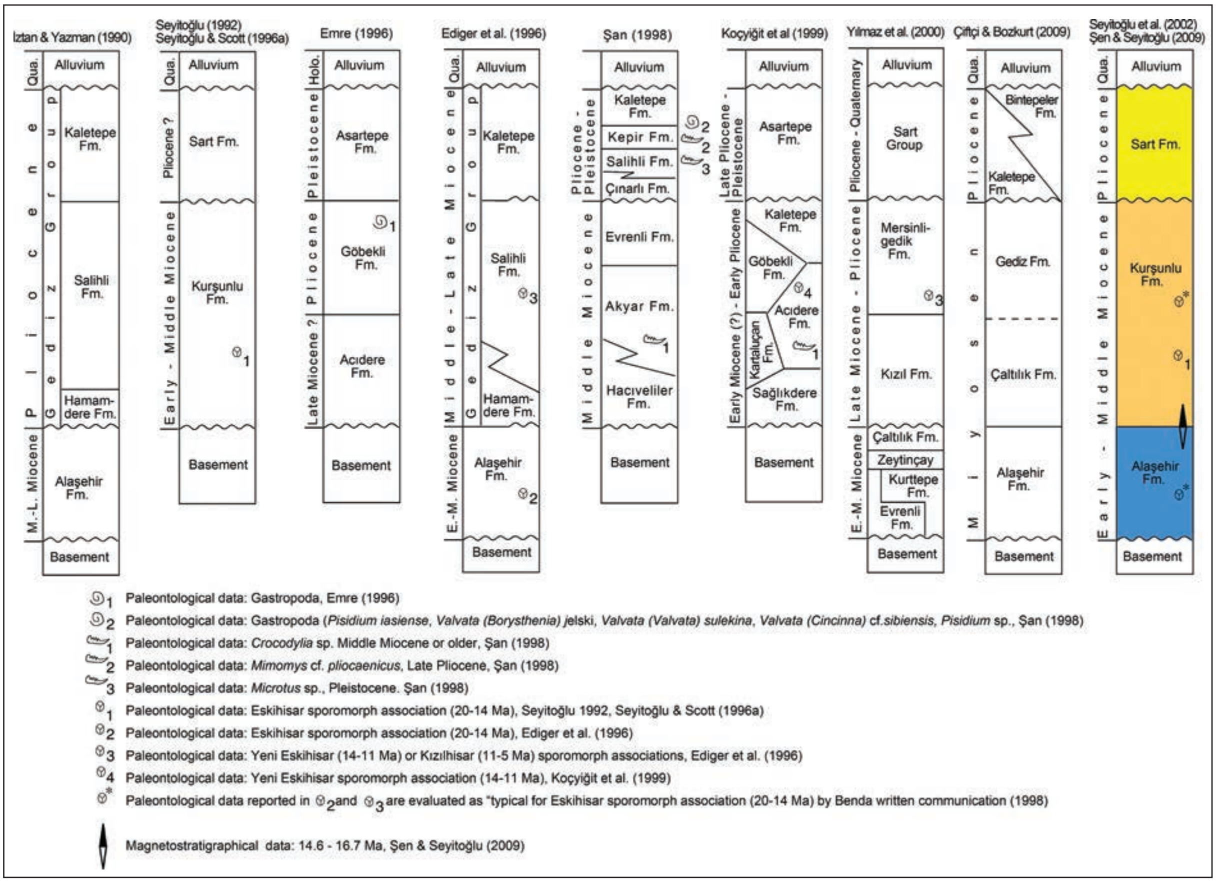

Figure 18- Stratigraphy of the Alaşehir graben (modified from Demircioğlu et al., 2010).

red color. This unit is described as the Kurşunlu formation (Seyitoğlu, 1992; Seyitoğlu and Scott, 1996*a*; Seyitoğlu et al., 2002). The Kurşunlu formation comprises angular conglomerates followed by coarse and fine sandstone in a cyclical sedimentary sequence. The typical red color of the Kurşunlu formation is seen in the lower levels while in the upper levels the color lightens to pink with gray intercalations.

The Eskihisar sporomorph association (20-14 Ma) has been defined in both the Alaşehir formation (Ediger et al., 1996) and the Kurşunlu formation (Seyitoğlu and Scott, 1996*a*). Taking note of magnetostratigraphic age data obtained from the transition between the Alaşehir and Kurşunlu formations (14.6-16.6 Ma) and the thickness of the lower Alaşehir formation, the age of initiation of the graben is determined as Early Miocene (Şen and Seyitoğlu, 2009).

The Sart formation which overlies the Alaşehir and Kurşunlu formations above an unconformity comprises light yellow, slightly consolidated conglomerates and sandstones. The name Sart formation is taken from the ancient city of Sardis (Seyitoğlu, 1992; Seyitoğlu and Scott, 1996*a*; Seyitoğlu et al., 2015*a*). The Pliocene Sart formation contains microfossils (Şan, 1998). Within the graben this formation developed on the hanging wall of Fault II (Seyitoğlu et al., 2002). Quaternary alluvium is found on the downdropped block of Fault III forming the graben plain and surface rapture from the 28 March 1969 Alaşehir earthquake developed along this fault (Arpat and Bingöl, 1969; Eyidoğan and Jackson, 1985) (Figures 17 and 18).

When studies of the tectonic evolution of Alaşehir graben are investigated, it is possible to observe traces of the "low angle vs. high angle" argument about normal faults in structural geology in assessments of the faults controlling graben formation. There are two different views on formation of the Alaşehir graben. The first group of researchers says the graben bounding faults were low-angle from the time of basin formation and these are cut by the higher-angle faults which young toward the north (Hetzel et al., 1995; Emre and Sözbilir, 1997; Sözbilir, 2001; Öner and Dilek, 2011). The second group of researchers proposes that initiation of the Alaşehir graben involved high-angle normal faults which gradually became low angle over time (Seyitoğlu and Şen, 1998; Gessner et al., 2001; Bozkurt, 2001; Seyitoğlu et al., 2002; Purvis and Robertson, 2005; Çiftçi and Bozkurt, 2009; 2010; Demircioğlu et al., 2010; Seyitoğlu et al., 2014).

According to Öner and Dilek (2011), from the first group of researchers who define the Alaşehir graben as a supradetachment basin developing on the Alaşehir Detachment Fault, at the start of basin development from the Early Miocene low-angle detachment faults were active, accompanied by scissor faults and from the Late Pliocene (3.5 Ma) high-angle faults cut the detachment fault leading to back-tilting of the sequence (Öner and Dilek, 2011). The lack of validity of this model may be shown with a few points. (1) Comparing basin fill in "rift basins" and "supradetachment basins" the clearest difference is the distance of the lacustrine depocenter from the main fault (Friedmann and Burbank, 1995). In supradetachment basins due to the giant alluvial fan deposits, lacustrine basins develop far from the basin edge whereas in rift basins due to high-angle normal faults the largest depocenter is very close to the edge of the basin and alluvial fans are of limited size (Friedmann and Burbank, 1995). When the geologic map of Öner and Dilek (2011) is examined, it is possible to observe the oldest rocks in basin fill are organic-rich mudstone and lacustrine limestone [Gerentaş fm. and Kaypaktepe fm. in Öner and Dilek, (2011)] located very close to the basin-edge fault. This feature was mentioned by Seyitoğlu et al. (2002) and the necessity of high-angle faults bordering the basin/graben initially was stated. (2) The Alaşehir Detachment Fault is one of the most dated faults on the earth. With the age dating completed to date it is possible to access relative spatial locations of the Alaşehir Detachment Fault (Seyitoğlu et al., 2014). Age dates on the Alaşehir Detachment Fault vary from 1.75±0.62 Ma to 21.70±4.50 Ma (Lips et al., 2001; Gessner et al., 2001; Catlos and Çemen, 2005; Glodny and Hetzel, 2007; Catlos et al., 2010; Buscher et al., 2013; Hetzel et al., 2013). In the supradetachment basin model of Öner and Dilek (2011), the Alaşehir Detachment Fault is cut by high-angle faults from 3.5 Ma and must cease activity. However, the age data mentioned above identified movement on the fault up to 1.75±0.62 Ma. (3) In the supradetachment basin model, the crosscutting relationships of faults in the Alaşehir graben are ordered relatively (Öner and Dilek, 2011; Figure 12c). The faults are ordered from old to young as high angle, low angle, high angle, low angle. When it is considered that the proposed model involves development of first low angle and then high angle normal faults, it appears that this situation is not observed in nature.

The second group of researchers who believe the Alaşehir graben began with high-angle faults have adapted the flexural rotation/rolling hinge model to the Alaşehir graben (Seyitoğlu et al., 2002). In the original "flexural rotation" model (Buck, 1988; Wernicke and Axen, 1988) initially high-angle normal faults reduce in angle due to isostatic rebound. New normal faults developing in the hanging wall of the first fault undertake the task of the first fault which cannot cope with the extension. This situation implies that movement should end on the rotating first fault and primary throw on faults formed pre-rotation will remain unchanged. At the end of the process, faults and related sediments young towards the center of the graben (Buck, 1988; Wernicke and Axen, 1988; Manning and Bartley, 1994; Axen and Bartley, 1997).

In the initial stages of the Alaşehir graben, the Alaşehir and Kurşunlu formations were deposited in front of high-angle faults (Fault I) in the Early Miocene. This conclusion is reached due to the proximity of the lacustrine facies to the graben edge in the mapped area of the Alaşehir formation (Seyitoğlu et al., 2002) (Figure 19). Additionally, geological maps by Cohen et al. (2005), Purvis and Robertson (2005) and Öner and Dilek (2011) include similar results. As observed in gravity data (Akçığ, 1988; Ateş et al., 1999) and seismic profiles parallel to the graben (Çiftçi and Bozkurt, 2010) in front of the towns of Alaşehir and Salihli, there are two separate sub-basins separated by possible "relay ramps" (Seyitoğlu et al., 2002). Thermochronological data from the Alaşehir graben (Gessner et al., 2001; Figure 3f) show that rocks in the footwall of Fault I were rapidly exhumed since about 5 Ma. This data supports the view that Fault II formed in the downdropped block of Fault I and while the Sart formation was being deposited ahead of this, Fault I rotated to become low angle (Figure 19).

As Fault I rotated and became low angle, different to the original "rolling hinge" model, activity continued and field data show this caused a larger amount of metamorphic basement to be exposed (Seyitoğlu et al., 2002; Figure 11). Here as a result of

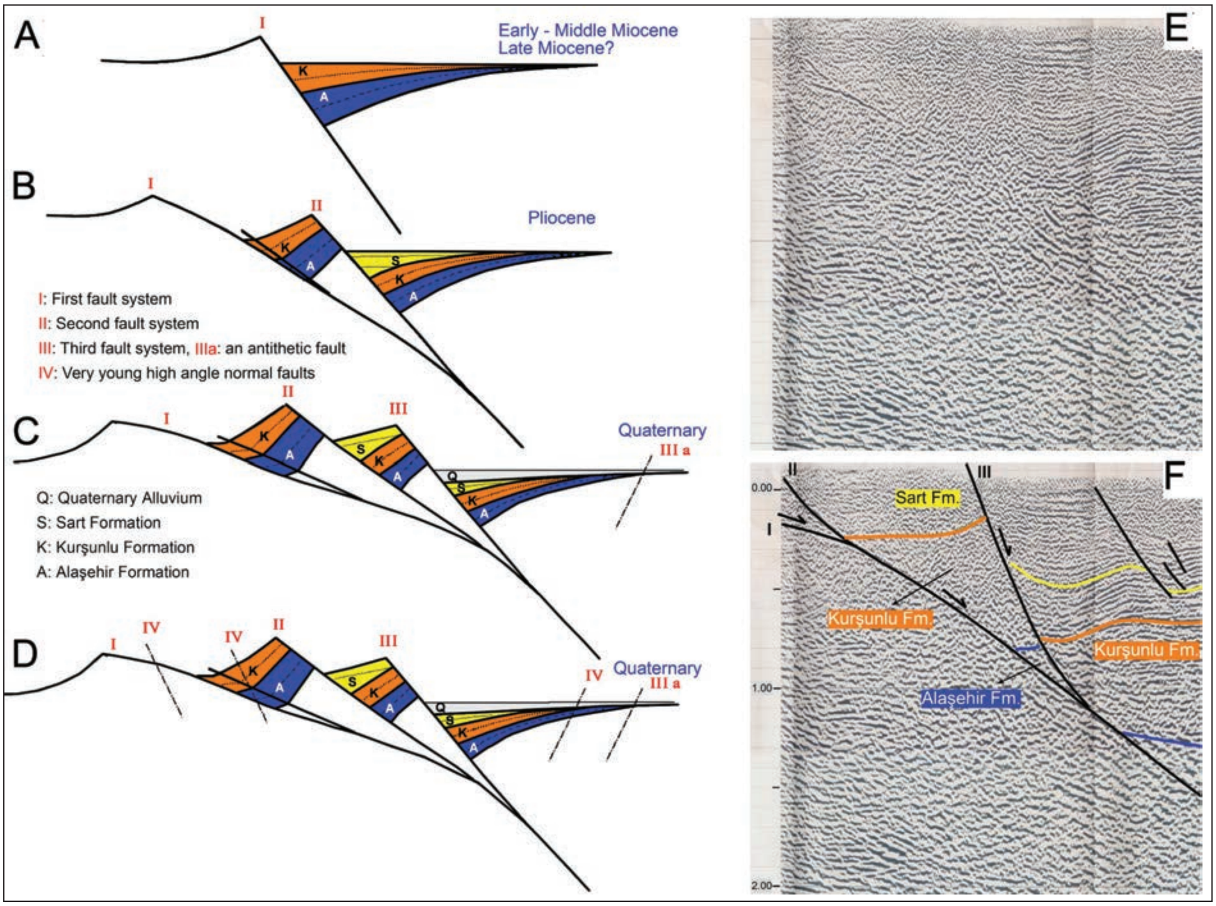

Figure 19- “Alaşehir-type rolling hinge” model for the Alaşehir graben (A-D) and seismic reflection profiles of the graben showing Fault II and Fault III merged to Fault I (E-F) (Adapted from Seyitoğlu et al., 2002; Demircioğlu et al., 2010; Seyitoğlu et al., 2014).

the low angle normal faults affecting the Kurşunlu formation with activity on these low angle normal faults occurring after deposition of the formation, it is understood that there was activity on the rotated low angle fault. Age dating (9.2±0.3 - 3.7±0.2 Ma) from the same location confirms this (Hetzel et al., 2013; Samples 10Me09 and 10Me10, Figure 3b). Preservation of activity as Fault I rotates to low angle is different to the original model. This difference is described as the “Alaşehir type rolling hinge model” (Seyitoğlu et al., 2014). This model explains (1) the extensional tectonic regime continuing from the Early Miocene to Quaternary without major disruption, (2) the large interval of dates obtained from the Alaşehir Detachment Fault (1.75±0.62 Ma to 21.70±4.50 Ma), and (3) the formation mechanism of the Horzum Turtleback (Seyitoğlu et al., 2014).

In the hanging wall of Fault II, Fault III developed and Fault I and II continued to rotate to lower angles and Quaternary alluvium was deposited on the downdropped block of Fault III. At the same time faults developing on the northern side of the Alaşehir graben made the graben symmetric. High angle faults (Fault IV) developing from the Quaternary to the present appear to cut low angle faults (Seyitoğlu et al., 2002) (Figure 19).

Faults identified on seismic reflection profiles completed by TPAO match faults mapped on the surface and named as Fault I, II and III (Demircioğlu et al., 2010). Accordingly at depth Faults II and III merge to Fault I but do not cut it (Figure 19). This finding supports the “rolling hinge” model of evolution of the Alaşehir graben. Seismic reflection profiles also illustrate the wedge geometry of the Alaşehir formation. This data shows that the Alaşehir formation was deposited simultaneous to E-W striking faults (Demircioğlu et al., 2010) and are not the product of north-trending basins as advocated by Yılmaz et al. (2000).

One of the most striking models for the evolution of the Alaşehir graben is the two-stage extensional model. Koçyiğit et al. (1999) examined sequences in the Alaşehir graben and found folds in rocks of the Salihli group and assuming a horizontal position for the Karataş group above an angular unconformity, they determined a N-S oriented short regional compressional stage in the Late Miocene-Early Pliocene. Seyitoğlu (1999) mentioned the radiometric age dates and palynologic analyses (Seyitoğlu, 1997*b*; Seyitoğlu et al., 1997; Seyitoğlu and Benda, 1998) from the Selendi and Uşak-Güre basins immediately to the north of the Alaşehir graben along with the horizontal placement of the İnay Group with age interval in the Lower-Middle Miocene and lack of effects from regional compression as contradicting this view. Seyitoğlu et al. (2000) investigated folds in the Alaşehir graben and found they were drag folds or rollover anticlines related to normal faults showing they were folds related to extensional tectonics, given a theoretical basis in Janecke et al. (1998). Purvis and Robertson (2005) concur with this observation. Sözbilir (2002) introduced the interpretation that the folds in graben fill developed in an extensional regime and formation was related to the ramp and flat geometry of detachment faults.

3.1.2. Büyük Menderes Graben

The Büyük Menderes graben is an east-west structure where the main fault is located on the north side. The base of the graben fill is blocky conglomerate, sandstone and mudstone with lignite levels of the Hasköy formation and it is reported that the formation includes E-W trending normal growth faults (Sözbilir and Emre, 1990). The Hasköy formation was determined to have been deposited in the Middle-Late Miocene by Sözbilir and Emre (1990) and this age was supported in palynological studies by Akgün and Akyol (1999). Among our reservations about this age data are the lack of first hand correlation of the proposed age with isotopic age data, and mammal or marine biochronological ages (Seyitoğlu and Şen, 1999). The response to this debate states the lack of confidence in isotopic age data (Akyol and Akgün, 2001).

It is known that Leopold Benda's Hasköy locality is where the Eskihisar sporomorph association was recorded (Becker-Platen, 1970). Taking this information together with the new age interval for the Eskihisar sporomorph association (20-14 Ma), the age of initiation of the E-W-trending Büyük Menderes graben, and as a result the N-S extensional tectonics in the Aegean, is determined to be Early Miocene (Seyitoğlu and Scott, 1992*a*).

The Gökkırantepe formation, conformably overlying the Hasköy formation (Sözbilir and Emre, 1990), is formed of red conglomerate, sandstone and mudstone and is accepted as the second sedimentary sequence in the graben (Figure 20). The transition from the Hasköy formation to the Gökkırantepe formation was dated to 14.88 - 15.97 Ma by magnetostratigraphy (Şen and Seyitoğlu, 2009). According to a mainly sedimentological study of the Büyük Menderes graben, each of the formations is a product of the E-W graben (Cohen et al., 1995) and the initiation age of the graben is described as Early Miocene (Seyitoğlu and Scott, 1992*a*; Şen and Seyitoğlu, 2009).

The Asartepe formation in the Büyük Menderes graben unconformably overlies previous units (Sözbilir and Emre, 1990; Şen and Seyitoğlu, 2009). At the Şevketin Dağı location within the formation, micromammalian fossils with Late Pliocene-Pleistocene ages are found (Ünay et al., 1995; Ünay and De Bruijn, 1998) (Figure 21).

Gürer et al. (2009) proposed the presence of a Late Pliocene-Pleistocene sequence according to micromammalian data from between Aydın – Ortaklar in the Büyük Menderes graben. According to age data, there is a high possibility that the position of this sequence in Büyük Menderes stratigraphy is equivalent to the Asartepe formation with certain position. To confirm this it is necessary to complete careful geological mapping laterally from the Şevketin Dağı location where the Asartepe formation is observed toward Aydın. The lack of validity of the claim of inconsistency between palynological ages near Nazilli and the micromammalian ages proposed by Bozkurt (2000) has been shown by similar geological mapping (Şen and Seyitoğlu, 2009). Here while the Eskihisar sporomorph association (20-14 Ma) was recorded in the Hasköy formation, the Late Pliocene-Pleistocene micromammalian age findings are within the Asartepe formation at different stratigraphic levels (Şen and Seyitoğlu, 2009) (Figure 21).

The newest study to support the cross graben model proposed by Şengör (1987) is by Gürer et al. (2009) who propose that the Lower-Middle Miocene sedimentary sequence in the Büyük Menderes graben was deposited under a N-S compressional regime in north-trending basins (Tibet-type grabens). Based on the Late Pliocene-Pleistocene micromammalian ages obtained from the sequence between Aydın – Ortaklar in the Büyük Menderes graben, mentioned

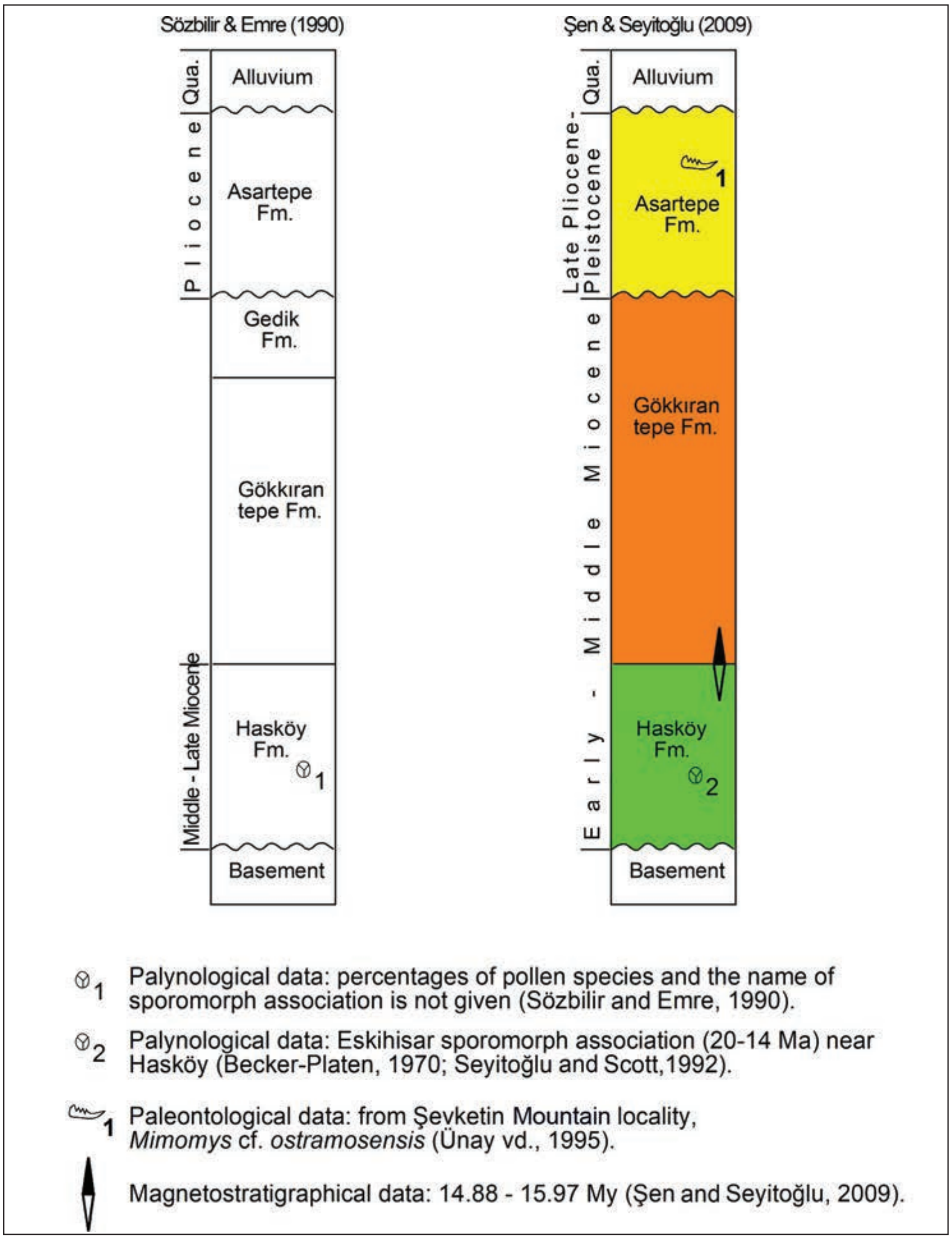

Figure 20- Generalized stratigraphy of the Büyük Menderes graben (prepared after Şen and Seyitoğlu, 2009).

above, it is advocated that the extensional regime in Western Anatolia began in this period and evidence of the compressional regime continuing to the Late Pliocene is shown by the last movement of the Lycian nappes in SW Anatolia. This model does not comply with several observations and studies in Western Anatolia. (1) Sedimentological studies of both the Alaşehir and Büyük Menderes grabens have stated that the sediments are coeval to tectonism and the age of these sediments may be used to identify the age of formation of E-W grabens (Cohen et al. 1995). The lowest sequence in the E-W grabens of Alaşehir and Büyük Menderes has been proven to have Early Miocene age by magnetostratigraphical studies (Şen and Seyitoğlu, 2009). (2) As discussed in the previous section, data from studies based on seismic reflection profiles in the Alaşehir graben show the sequence at the bottom of the graben fill (Alaşehir formation) developed simultaneously to the east-west normal fault system (Demircioğlu et al., 2010; Çiftçi and Bozkurt, 2010). In the study by Çiftçi et al. (2011) assessing seismic reflection data from the Büyük Menderes graben, seismic reflection profiles in an E-W direction showed the lowest two sequences in the

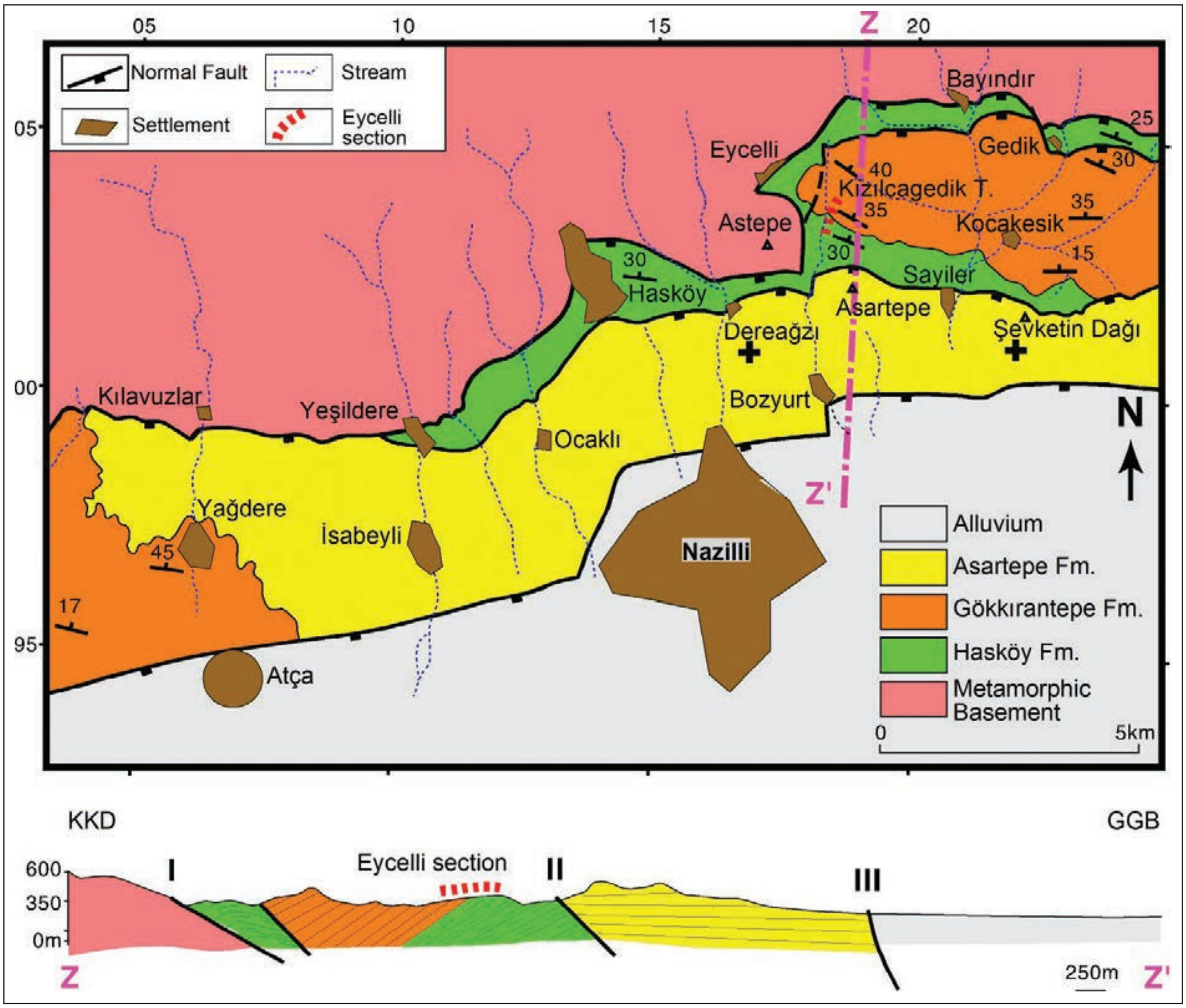

Figure 21- Geological map and cross section north of Nazilli in the Büyük Menderes graben. Dashed red line indicates location of magnetostratigraphic section (Taken from Şen and Seyitoğlu, 2009).

graben fill (equivalent to Hasköy and Gökkırantepe formations) have lens-shaped geometries, while on N-S seismic reflection profiles E-W normal faults acting as growth faults for the first two sequences. North-trending faults, observed at the surface and proposed by Gürer et al. (2009) as controlling units equivalent to the lowest unit of graben fill, the Hasköy formation, developed after deposition and are reported to be transfer faults frequently observed developing in basins with normal faults (Çiftçi et al., 2011). These findings make the proposed assessment by Gürer et al. (2009) invalid. (3) Gürer et al. (2009) used movement of the Lycian nappes to the south as data showing continuation of the compressional tectonic regime until the Late Pliocene. However, the ages of basins in the Menderes Massif (Gördes and Dalama-Kuloğulları; Seyitoğlu et al., 1992) and thermochronological ages (Gessner et al., 2001*b*) show that the massif was freed of Lycian nappe cover before the Early Miocene, requiring the last movement of Lycian nappes toward the south, thought to have been sourced at the Izmir-Ankara suture zone, to be rootless and for this crustal shortening is not required. This was stated by Seyitoğlu et al. (1992) and this opinion was supported by Collins and Robertson (2003). (4) The deformed units and reverse fault images used to support the view of Gürer et al. (2009) that the compressional tectonic regime continued to the Late Pliocene. Firstly it must be determined whether the deformation is due to slump folding. Additionally, it must not be forgotten that folds and reverse faulting in sequences above detachment faults may develop linked to the ramp-flat geometry of the detachment fault (McClay, 1989; 1990). In recent times data from within the Alaşehir graben was published in a study assessing similar structures as products of progressive deformation (Şengör and Bozkurt, 2013). Another dimension of the topic are the regional effects of the proposed compression continuing to the Late Pliocene (Gürer et al., 2009) or short-duration compression between the Late Miocene-Pliocene (Koçyiğit et al., 1999). As discussed in an earlier section, the İnay Group, with definite Lower-Middle Miocene age based on palynology and isotopic age data in basins immediately north of the E-W trending grabens, is in a horizontal position and was not affected by the proposed regional compression (Seyitoğlu, 1999). However, compressional data published in papers to support the regional compression of Gürer et al. (2009) has been refuted. For example, the folds mentioned by Koçyiğit et al. (1999) have been shown to be drag folds or rollover anticlines (Seyitoğlu et al., 2000). The reverse fault proposed by Koçyiğit et al. (2000) in the Akşehir-Afyon graben was not found on seismic data (Kaya et al., 2014). Folds in Neogene units of Eskişehir plain in Koçyiğit (2005) were determined to be related to active strike-slip faulting (Seyitoğlu et al., 2015*b*). All these studies show the compressional regime in Western Anatolia did not continue to the Late Pliocene and there was no short-term compression between the Miocene-Pliocene.

3.1.3. Denizli Graben

The WNW-ESE-trending Denizli graben is located in the SE of the Menderes Massif. The graben is 70 km long and 50 km wide, with the southern edge of the graben forming the northern slope of the Babadağ reaching 2000 m. On these slopes the southern edge of the graben is bordered by the Babadağ Fault Zone with 45-50^{o} north-dipping normal faults. The Buldan horst in the NW of the graben divides the graben fill in two. In the northern section of the graben, the Denizli graben with NW-SE normal faults developed in the Quaternary nearly unites with the Alaşehir graben. The southern section is linked to the Büyük Menderes graben by E-W Quaternary normal faults (Figure 22). Seismic activity in the region is very high due to the Babadağ and Pamukkale fault zones (Kaypak and Gökkaya, 2012).

Denizli graben fill is reported as Pliocene age (Taner, 1974*a*, *b*; 1975), so have attracted less attention of the studies dealing with the initiation of extensional tectonics. However, after the identification of micromammalian fossils showing Early Miocene age (Saraç, 2003), as the Denizli graben aroused more interest as it houses sedimentary

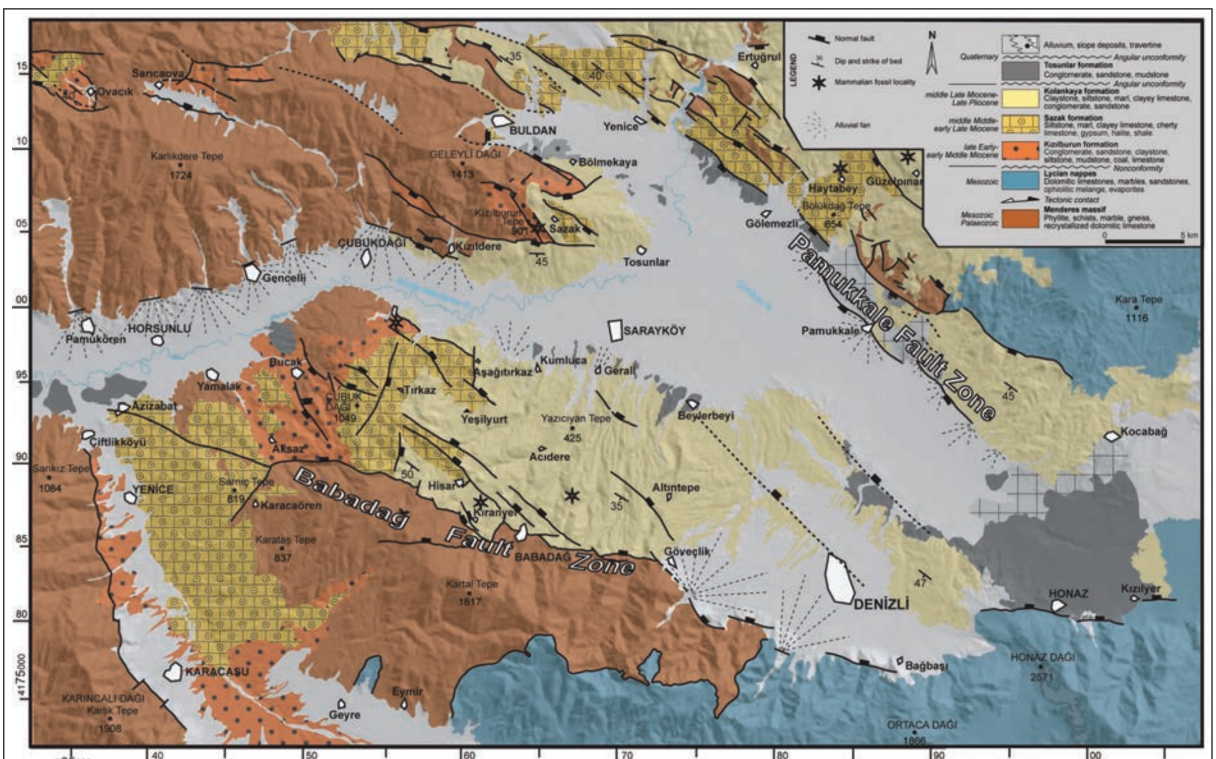

Figure 22- Geological map of the Denizli graben (Taken from Sun, 1990 and Alçiçek, 2007).

fill from a very broad time interval. Koçyiğit (2005), Westaway et al. (2005) and Kaymakçı (2006) do not accept the Lower-Middle Miocene sedimentary sequence as a product of the Denizli graben, but defend the opinion that these sediments developed outside the graben system.

According to Koçyiğit (2005) the Denizli graben developed under two stages of extensional regime separated by a compressional phase. The first extensional regime was in force from the Middle Miocene-Middle Pliocene, and then in the Latest Middle Pliocene a compressional regime reigned and from the Latest Pliocene to the present the second stage of the extensional regime developed. Westaway et al. (2005) proposed that current continental crustal extension in the Denizli region developed about 7 Ma at the beginning of the Late Miocene, revising the proposal of Westaway (1993). Kaymakçı (2006) interpreted Denizli graben fill as Upper Miocene to present sediments advocating the extension had been effective since the Late Miocene.

Alçiçek et al. (2007) revised the previously created stratigraphy (Şimşek, 1984) and geological maps (Sun, 1990) and using micromammalian findings determined the sedimentary characteristics of Denizli graben fill. The lowest level of Denizli graben fill begins with the Kızılburun formation (Figure 23). Matrix-supported coarse conglomerates

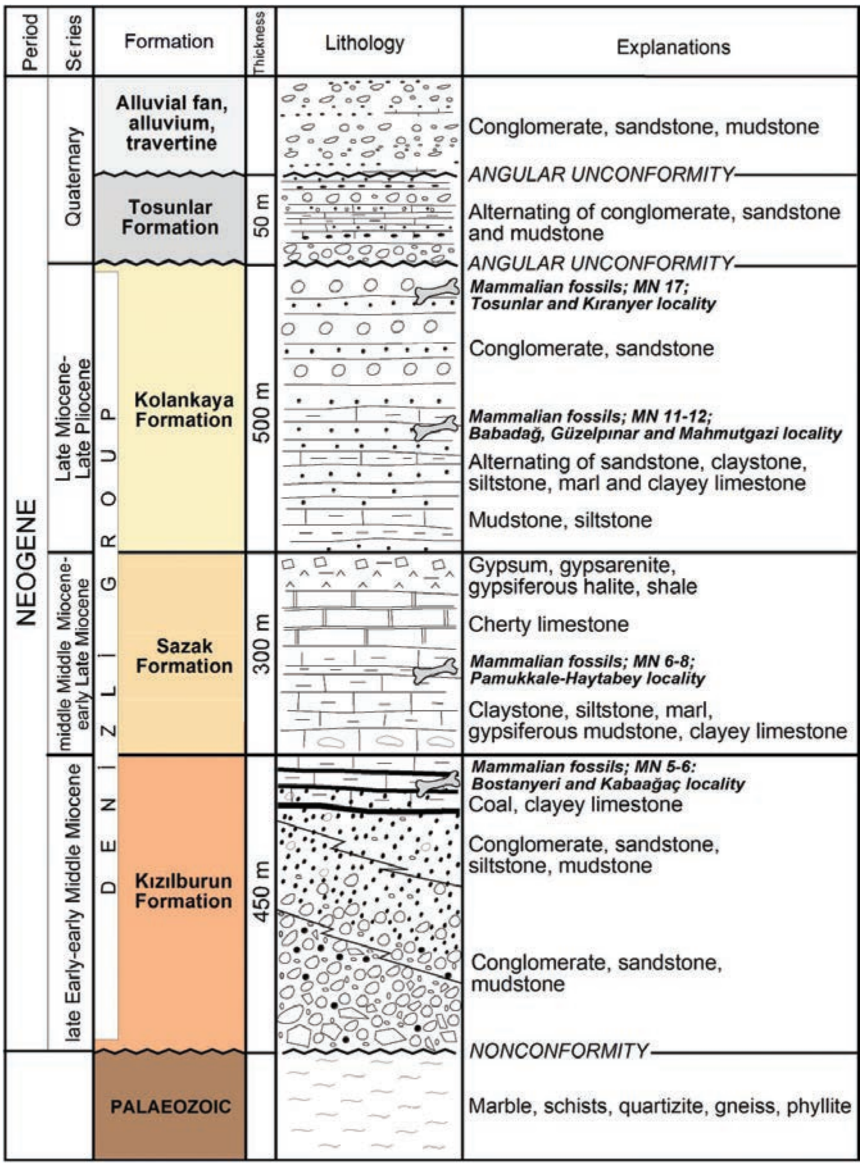

Figure 23- Generalized stratigraphy of the Denizli graben (Taken from Şimşek 1984; Alçiçek, 2007).

and red mudstone laminations pass upwards into clast-supported channel-fill conglomerates and fine-grained sediments including coal. At the localities of Bostanyeri and Kabaağaç MN5-MN6 (Late Burdigalian-Early Serravallian) fossils have been found (Saraç, 2003). The lower sections of the formation have been interpreted as proximal-medial alluvial fans while the upper sections are thought to be distal alluvial fans (Alçiçek et al., 2007).

The Sazak formation comprises limestone containing gypsum, green marl laminated siltstone-mudstone, clay limestone and fine mudstone with gypsum intercalations. Toward the top of the formation it passes into cherty limestones, gypsarenites, gypsum levels and shales. The Sazak formation includes fossils representing MN6-MN8 (Langian-Serravallian) levels (Saraç, 2003). The sequence in this formation is interpreted as lake-edge, shallow lacustrine and playa lake (Alçiçek et al., 2007).

The Kolankaya formation conformably overlies the Sazak formation and overlies the metamorphic basement at the north edge of the graben (Şimşek, 1984; Sun, 1990; Alçiçek et al., 2007). The formation, comprising laminated mudstone-siltstone, marl and clayey limestone, passes to laminated sandstones, cross-bedded conglomerates and sandstones and has been assessed as lacustrine, coastal, offshore and alluvial fan sediments (Alçiçek et al., 2007). The Kolankaya formation includes MN11-MN12 level mammalian fossils (Babadağ, Güzelpınar and Mahmutgazi locations: Sickenberg and Tobien, 1971; Saraç, 2003). Based on mammalian remains at Tosunlar and Kıranyer the Kolankaya formation was evaluated as Late Pliocene by Kaymakçı (2006).

The Tosunlar formation comprises loosely consolidated yellow-brown conglomerates, sandstone, siltstone and mudstone. It includes claystone and marl intercalations. This formation unconformably overlies previous units and is thought to be Pleistocene in age. The sequence ends with recent alluvium (Alçiçek et al., 2007) (Figure 23).

The WNW-ESE Babadağ Fault Zone bounding the Denizli graben controls sedimentation of graben fill nearly continuously deposited from the Early Miocene to the present. The relationship between the Babadağ Fault Zone and Kızılburun formation is observed in a deeply gouged valley 2 km south of Aksaz (Figure 22). Here blocky conglomerates of the Kızılburun formation overlie synthetic faults of the Babadağ Fault and it is seen that the Kızılburun formation thickens towards the Babadağ Fault with very well-developed wedge geometry (Figure 24).

Geotraverse from the Kabaağaç fossil locality to the Babadağ Fault Zone have shown clearly that though the original positions are disrupted by young normal faults, in the Early – early Middle Miocene period, the Kızılburun formation was controlled by the Babadağ Fault Zone (Figure 25).

Near Babadağ village (60588N-85191E) blocky conglomerate layers belonging to the Kolankaya formation have progressively less dip in the upper levels in the downdropped block of the Babadağ Fault Zone. This observation shows that the Babadağ Fault acted as a growth fault during deposition of the Kolankaya formation in the Late Miocene-Late Pliocene (Figure 26).

Recent activity on the Babadağ Fault Zone is clearly observed in development of alluvial fan deposits in the Quaternary (Figure 22) and seismicity (Kaypak and Gökkaya, 2012). Sarı and Şalk (2006) showed that basin fill thickened toward the Babadağ Fault Zone and had a wedge geometry using gravity data. All this data shows that the Babadağ Fault Zone played a significant role in the development of the Denizli graben from Early Miocene to Quaternary. Folds observed in the Neogene sequence in the Denizli graben and attributed to a compressional phase by Koçyiğit (2005) are the anticlines and synclines of drag folds developing on footwal or hanging wall of normal faults with an extensional origin, similar to those in the Alaşehir graben (Seyitoğlu et al., 2000).

Quaternary faults border the Buldan horst and the northern edge of the graben. A series of high-angle normal faults have created a stepped topography on the south slope of the Buldan horst and metamorphic basement together with Mio-Pliocene sequences have been uplifted. The Pamukkale Fault Zone borders the north of the Denizli graben in which travertine development and archeoseismology of Hieropolis are well known features (Altunel, 1996; Uysal et al., 2009).

3.1.4. Küçük Menderes Graben

The Küçük Menderes graben, slightly less prominent compared to the Alaşehir and Büyük Menderes grabens, has a basement of metamorphic

Figure 24- Coeval relationship between faulting on the Babadağ Fault Zone and the Kızılburun formation south of Aksaz.

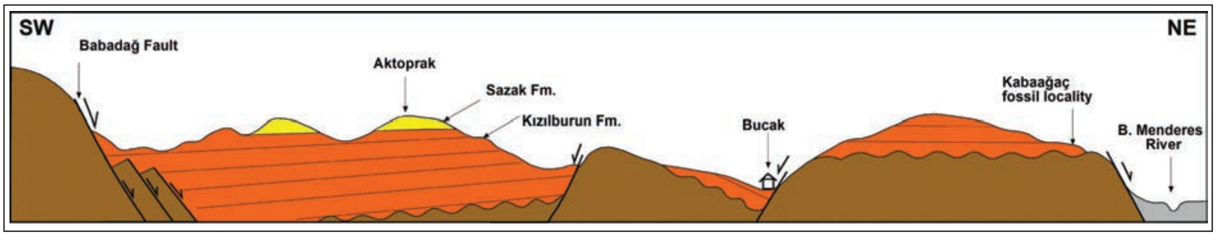

Figure 25- Scale-less cross section between Babadağ fault zone and Kabaağaç fossil locality.

Figure 26- Reduction in dip of upper layers of the Kolankaya formation, wedge geometry thickening toward the Babadağ Fault, SW of Babadağ village.

rocks namely mica gneiss, mica schists, garnet mica schist, calc-schist, quartzitic schist and marble (Seyitoğlu and Işık, 2009). Başova andesites cut the metamorphic rocks and have been dated to 14.3-14.7 Ma (Emre et al., 2006). Graben fill begins with the Suludere formation. Unconformably overlying the metamorphic basement and Başova andesites (14.3-14.7 Ma), this formation includes conglomerates, sandstone, mudstone and lacustrine limestones and is interpreted as flow-dominated alluvial fan and lacustrine sediments (Emre et al., 2006). The Aydoğdu formation unconformably overlies the Suludere formation and comprises conglomerates with light brown, reddish brown weakly lithified sandstone and mudstone intercalations. Recent alluvial sediments cover previous units (Emre et al., 2006).

At the north edge of the Küçük Menderes graben the boundary between metamorphic basement and graben fill was mapped as a reverse fault by Bozkurt and Rojay (2005), with this relationship indicating the presence of the N-S compressional phase of the two-stage extensional model. Contrary to this, Emre et al. (2006, Figure 4) mapped the same boundary as a normal fault. However, later studies by the same researchers determined that the sedimentary sequence was deposited in a compressional system in the late Middle Miocene-Early-Middle Pliocene and that normal faults developed in the Plio-Quaternary (Emre and Sözbilir, 2007). With different characteristics of the tectonic contact shown by different researchers, the north edge of the Küçük Menderes graben was carefully mapped by Seyitoğlu and Işık (2009). The results of this mapping revealed that the contact between the metamorphic basement and graben fill had characteristics of a brittle deformation zone and was bounded by an east-west-trending normal fault dipping >45º to the south. Overturned layers of the sedimentary sequence were not observed, asymmetric synclines observed on the downdropped block were interpreted as a drag fold syncline (Seyitoğlu and Işık, 2009) (Figure 27).

The Küçük Menderes graben has a unique tectonic position (Figure 28). In the north normal faults bounding the south edge of the Alaşehir graben and in the south normal faults bounding the north

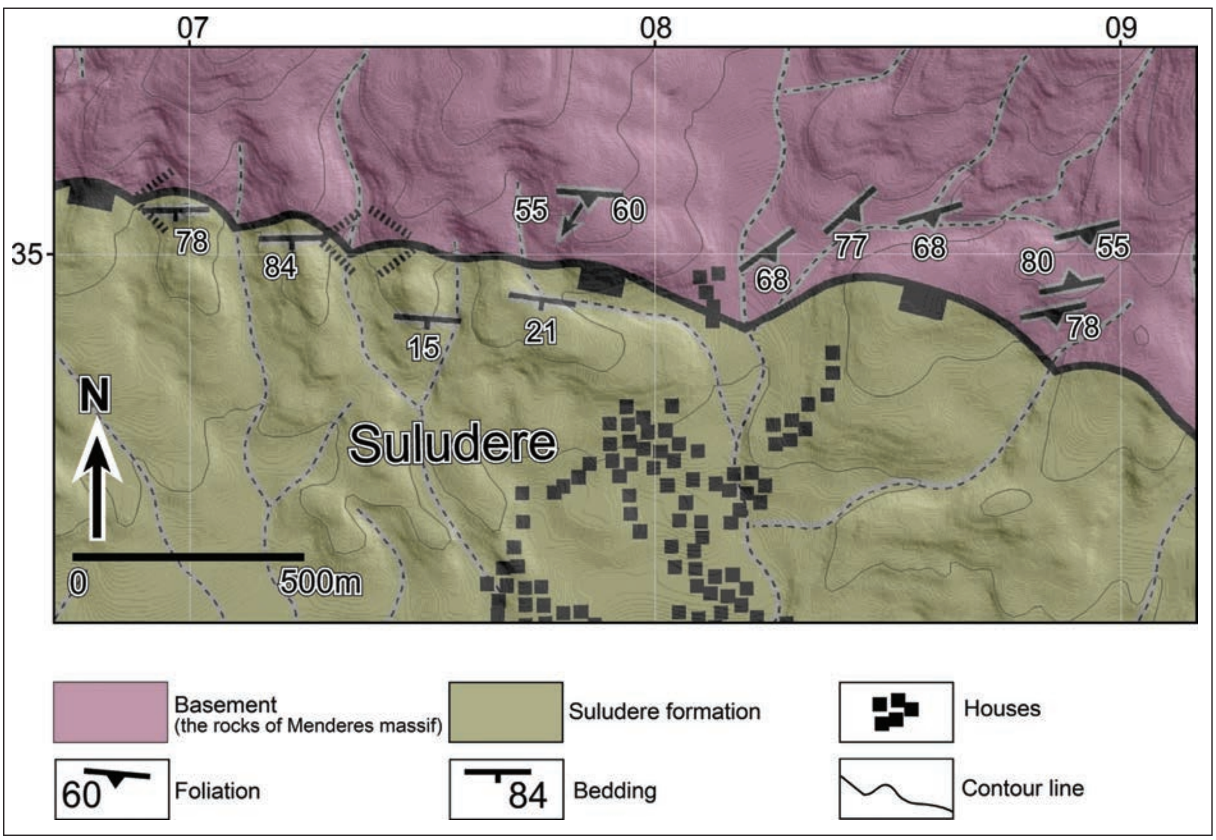

Figure 27- Geological map of the north edge of the Küçük Menderes graben near Suludere (Taken from Seyitoğlu and Işık, 2009).

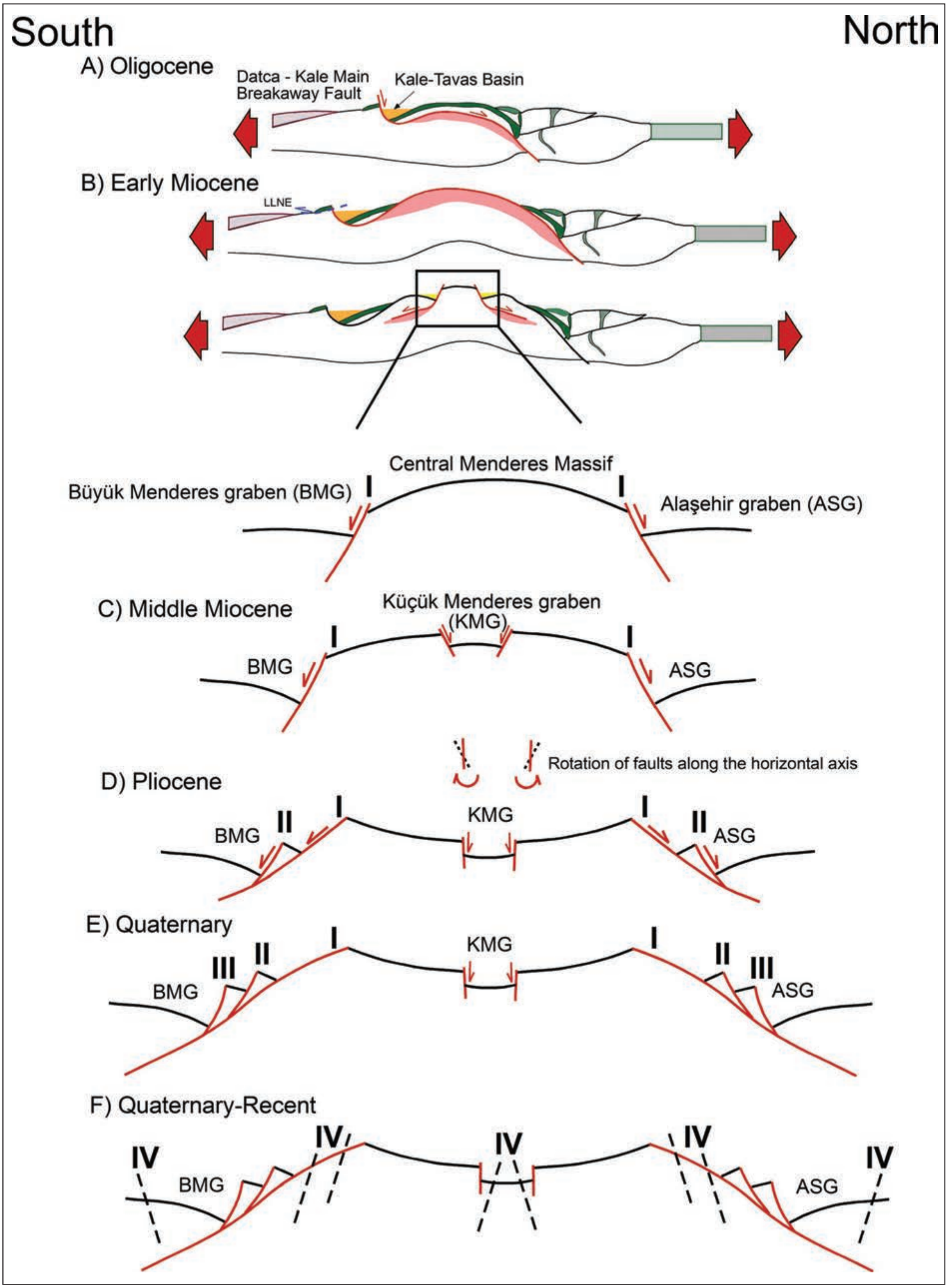

Figure 28- Unique tectonic position of the Küçük Menderes graben (Taken from Seyitoğlu and Işık, 2009).

edge of the Büyük Menderes graben have developed in accordance with the rolling hinge model and a gigantic synclinal with wave length of 45 km and width of 10 km has developed. The Küçük Menderes graben is located in the axial region of this syncline. According to kinematic analysis of folds, in the center of the fold reverse faults developed parallel to the fold axis linked to narrowing (Ramsey and Huber, 1987; Davis and Reynolds, 1996). If this basic kinematic rule is applied to the Menderes Massif between the Alaşehir and Büyük Menderes detachment faults rotating in opposite directions to each other, it may be expected that reverse faults be observed around the Küçük Menderes graben (Bozkurt and Rojay, 2005; Emre and Sözbilir, 2007). However, field studies have shown that only unusually high-angle normal faults are found north of the Küçük Menderes graben (Seyitoğlu and Işık, 2009). Possibly the high-angle faults in the Küçük Menderes graben initially were about 45º. It is thought that the current angles greater than 45º may be due to rotation of the horizontal axis. This rotation may mean results of rotation of the Alaşehir and Büyük Menderes detachment faults observed in the Pliocene of the central Menderes Massif may be related to rotation on the limbs of a large syncline developing due to the effects of an extensional tectonic regime (Seyitoğlu and Işık, 2009) (Figure 28). If compressional structures had been successfully demonstrated near the Küçük Menderes graben or even if they are in the future, it is not possible to attribute these types of structures to a regional compressional regime because the giant synclinal structure of the central Menderes Massif was created by rotational flexure of the Alaşehir and Büyük Menderes detachment faults; in other words a result of extensional tectonic processes. For this reason, it is not possible to attribute local contractions observed in the Küçük Menderes graben located in the axial region of the giant syncline to a regional compressional regime (Seyitoğlu and Işık, 2009) (Figure 28).

3.1.5. Simav Graben

The Simav graben is one of the important E-W trending grabens in Western Anatolia with a topographic difference between the south edge of the graben and the floor of 1100 m. One of the first studies in the graben is Zeschke (1954). Konak (1979) advocated that the Simav Fault is a strike slip fault active since the Early Miocene and determined that since the late Miocene 6 km of displacement has occurred based on offset metamorphic zones (Konak, 1982). The effect of this opinion may be observed on the current MTA active fault map (Emre et al., 2011). While Eyidoğan and Jackson (1985) advocated that the north and south edges of the Simav graben are bordered by faults with the northern edge currently dominant, Westaway (1990) mentioned that the main normal fault is found on the southern edge.

The nearly E-W trending and 65-70º north-dipping normal fault on the southern edge of the Simav graben is known as the "Simav Fault" in the literature, different to the Simav Detachment Fault (Seyitoğlu, 1997*a*). The Simav Fault is a high-angle fault and is post-tectonic compared to the Simav Detachment Fault (Işık, 2004). Unconsolidated blocky conglomerate and coarse sandstone is found on the hanging wall of the Simav Fault and the contact with the metamorphic basement is clearly seen. However, in the uplifted block of the Simav Fault it becomes difficult to observe the Simav Fault within basin sediments of the north-trending Demirci basin. In this situation the only criterion to separate conglomerate with blocks derived from metamorphic basement in the Demirci basin from blocky conglomerate on the downdropped block of the Simav fault is consolidation. One of the clearest morphological effects of the eastern extension of the Simav Fault is the drop of the northern section of Kıbletepe. Further east it is difficult to observe the trace of the Simav Fault in the Hacıbekir Group sediments of the Selendi basin (Seyitoğlu, 1997a). The Simav Fault is active, with focal mechanism solutions of recent earthquakes showing dominant pure normal faulting (Yolsal-Çevikbilen et al., 2014). North of the Simav Fault, the north-trending Akdere basin has semi-consolidated blocky conglomerates, coarse sandstones and white tuff units, grading up to pink conglomerates, sandstone and tuff layering covered by Naşa volcanics. According to the determination of the age of Naşa volcanics (15.8±0.3 Ma and 15.2±0.3 Ma) (Ercan et al., 1997) the sedimentary fill in the Akdere basin must be older than 15.8 Ma. The Akdere basin began to develop as a symmetric graben with activity on the eastern edge lasting longer than activity on the western edge (Seyitoğlu, 1997*a*).

3.1.6. Early Miocene-Quaternary Paleogeographic Development of East-West Grabens in the Central Menderes Massif

The paleogeographic development of the Alaşehir, Büyük Menderes, Denizli and Küçük

Menderes grabens in the interval from the Early Miocene to Quaternary is summarized in figure 29. After the first exhumation of the Menderes massif with a dome shape (see: section 4) as a result of continuing north-south extension the proto Alaşehir, Büyük Menderes, Denizli and Sarıcaova grabens began to develop in the Early Miocene (Figure 29a). Proof of sedimentation in this period in the Alaşehir, Büyük Menderes and Sarıcaova grabens comes from the Eskihisar sporomorph association (Becker-Platen, 1970; Seyitoğlu and Scott, 1992; 1996; Ediger et al., 1996). In the Denizli graben sedimentation data comes from mammalian findings from the Early Miocene at Bostanyeri and Kabaağaç localities (Saraç, 2003; Alçiçek et al., 2007). In accordance with descriptions by Gawthorpe and Hurst (1993) the remaining area between the Büyük Menderes graben and Denizli graben may be assessed as a antithetic transfer zone while the area between the Denizli graben and Alaşehir graben may be a synthetic transfer zone (Figure 29a). In the Middle Miocene the antithetic transfer zone between the Büyük Menderes and Denizli grabens was semi-parallel to the Babadağ Fault controlling the Denizli graben forming the Bozdoğan and Karacasu grabens and the proto Küçük Menderes graben formed (Figure 29b). In the Late

Figure 29- Early Miocene-Quaternary Paleogeographic Development of Alaşehir, Büyük Menderes, Denizli and Küçük Menderes grabens. Pink areas represent flexural rotation of the Alaşehir and Büyük Menderes detachment faults.

Miocene the Karacasu graben and Denizli grabens became linked (Figure 29c). In the Pliocene second faults developed in the downdropped blocks of the first faults in the Alaşehir and Büyük Menderes grabens, the first faults became low-angle in accordance with the rolling-hinge model and they began to exhume as detachment faults. The Denizli graben continued to develop under control of the Babadağ Fault (Figure 29d). In the Quaternary the third faults developed in the Alaşehir and Büyük Menderes grabens and then antithetic faults in all half-grabens activated and they became full grabens (Figure 29 e). In the interval from the Quaternary to the present, young faults and the Buldan Horst developed leaving sediments from the Denizli graben suspended above it. The Büyük Menderes graben and Denizli graben joined and the north of the Denizli graben and east of the Alaşehir graben morphologically approached each other (Figure 29f).

3.2. NE-SW-Trending Basins

North of the E-W trending Alaşehir graben the presence of parallel NE-SW trending basins attracted the attention of researchers many years ago (Kaya, 1981; Şengör, 1987). As mentioned briefly in the introduction and to be further explained in the next sections, their role within the extensional tectonics of Western Anatolia is still debated (e.g., Ersoy et al., 2011; Karaoğlu and Helvacı, 2014).

3.2.1. Gördes Basin

The first observations on the basin are in studies by Nebert (1961) and Yağmurlu (1986). Seyitoğlu et al. (1992), Seyitoğlu et al. (1994) and Seyitoğlu and Scott (1994) investigated the isotopic age dating of volcanic rocks, palynology and stratigraphy in the basin. According to these studies, in the northwest section of the basin, the basement is composed of Izmir-Ankara suture zone and Dağdere formation starts with rounded cobble conglomerate derived from the ophiolitic basement and pebble conglomerates derived from metamorphic basement interlayers before passing up into sandstone, mudstone, lignite levels (Çıtak lignite) and marls. Lignite samples from three locations in the Dağdere formation were investigated by L. Benda and Eskihisar sporomorph association (20-14 Ma) was identified (for detailed pollen list see Seyitoğlu, 1992; Seyitoğlu et al., 1994) (Figure 30).

The lower levels of the Tepeköy formation, outcropping on the south and east edges of the Gördes basin, have been uplifted by the effect of central volcanics and are observed near Azimdağı (Figure 31). After a thin conglomerate derived from ophiolites, blocky conglomerates derived from metamorphics, coarse sandstone and red lithified sandstone form the lower levels of the Tepeköy formation. The upper levels of the Tepeköy formation

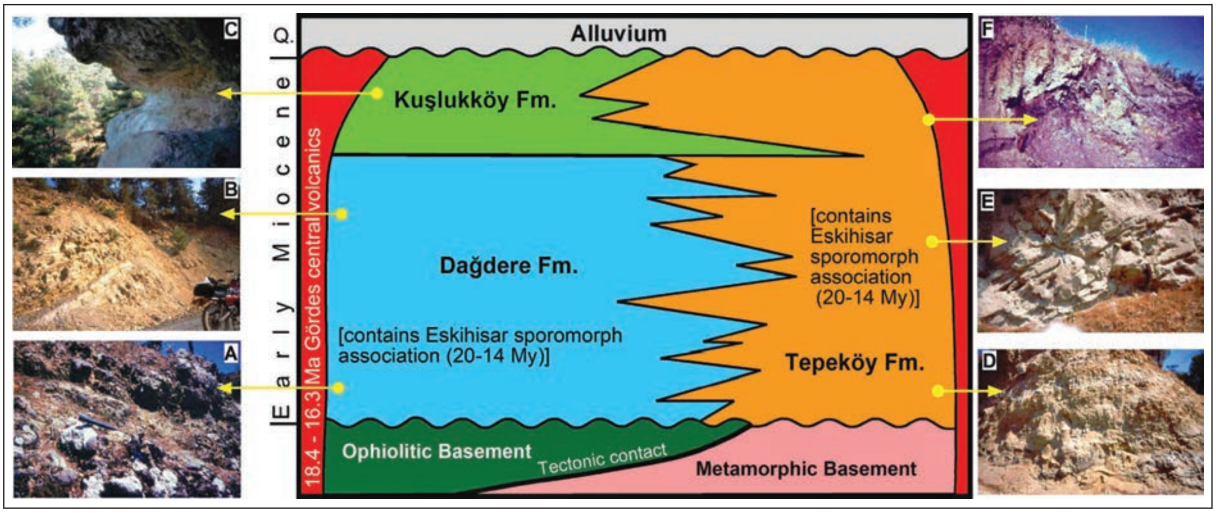

Figure 30- Stratigraphy of the Gördes basin (Taken from Seyitoğlu, 1992; Seyitoğlu and Scott, 1994). A) Conglomerate with recrystallized limestone blocks derived from ophiolitic basement which dominates the Dağdere formation. B) Upper levels of the Dağdere formations comprise fine-grained sandstone and mudstone and include lignite levels. C) The Kuşlukköy formation is distinguished by tuffs, comprising tuff, marl and limestone layers. D) Tepeköy formation includes conglomerates with blocks derived from metamorphic basement, blocks include mylonitic rocks. E) Yellow sandstone forming the upper levels of the Tepeköy formation. F) Cross cutting relationship of Gördes central volcanics and basin fill.

interfingers with the Kuşlukköy formation. Their general lithology is dark yellow sandstone and mudstones including conglomerate levels. The Tepeköy formation includes two locations with the Eskihisar sporomorph association (20-14 Ma) (Seyitoğlu, 1992; Seyitoğlu et al., 1994). The Kuşlukköy formation conformably overlies the Dağdere formation and shows lateral transition to the upper levels of the Tepeköy formation (Figure 30). The general lithology is tuff, sandstone, marl and silicified limestone intercalations. The tuff levels are the distinguishing element of the formation (Seyitoğlu and Scott, 1994). The central volcanics (18.4±0.8 Ma - 16.3±0.5 Ma) cut all basin fill and in the eastern margin of the basin a dated leucogranite dyke (24.2±0.8 Ma -21.1±1.1 Ma) provide its pebbles to the basin fill. As a result, the Gördes basin fill is broadly dated to the Early Miocene from 24 to 16 Ma (Seyitoğlu et al., 1992) (Figure 30).

Recent dating studies using new techniques for the Gördes basin have provided roughly the same results as isotopic age dating of volcanic rocks (Purvis and Robertson, 2005; Ersoy et al., 2011). Purvis and Robertson (2005) assessed the sedimentary facies of the basin fill while Ersoy et al. (2011) differentiated on the basis of formation. The Kızıldam formation at the bottom of the basin has been mapped near the basin margin faults. However, in the north of Dağdere village, the overlapped section is composed of limestones belonging to the upper levels of Dağdere formation. It is not a clastic unit of so-called Kızıldam formation. The Kızıldam formation described by Ersoy et al. (2011) has a conglomeritic unit with rounded clasts derived from ophiolitic basement and a conglomeritic unit with angular clasts derived from metamorphic basement. This situation means the Kızıldam formation is equivalent to the lower levels of the Dağdere and Tepeköy formations named in Seyitoğlu and Scott (1994) (Ersoy et al., 2011, p. 166). Contradictory to this statement, the type locality for the Kızıldam formation was chosen as the upper stratigraphic section where it interfingers with the Kuşlukköy formation (Ersoy et al., 2011). These clastic units found immediately east of Gördes are the upper sections of the Tepeköy formation, interfingering with the Kuşlukköy formation, found both above and below this formation (Seyitoğlu and Scott, 1994). However, in the general stratigraphic sequence of Ersoy et al. (2011) the Kızıldam formation is shown below the Kuşlukköy formation. The name used for the Kuşlukköy formation above the Kızıldam formation is the same as in Seyitoğlu and Scott (1994), but the descriptions are different (Ersoy et al., 2011). The contact of the original Kuşlukköy formation was drawn as the first observed tuff unit in the sequence (Seyitoğlu, 1992), in this situation the Çıtak coals remain within the Dağdere formation. However, Ersoy et al. (2011) include the Çıtak coals in the Kuşlukköy formation. According to Ersoy et al. (2011) the Kuşlukköy formation covers a large area including blocky conglomerates of the lower sections of basin fill cut and steepened by the central volcanics. The geological map in Ersoy et al. (2011, Figure 4) does not show a faulted/overlapped relationship in the southwest section of the basin. Seyitoğlu (1992) and Seyitoğlu and Scott (1994) showed that basin fill overlies metamorphic basement in the east section of the basin. These data and effects of the newly-described Kızıldam formation on the proposed regional model will be discussed later (See; Section 5.2).

3.2.2. Demirci Basin

Demirci basin sediments were examined by İnci (1984) and two sequences separated by an unconformity were determined. At the bottom of the basement fill, contrary to the original description of the Kürtköyü formation (Ercan et al., 1978), conglomerates including clasts derived from dominantly metamorphic rocks were reported to occur (İnci, 1984). Basin stratigraphy was examined by Yılmaz et al. (2000) and Early-Middle Miocene sequence and Pliocene limestones were distinguished. Ersoy et al. (2011) proposed the presence of two sequences separated by an unconformity. Here the Kürtköyü formation found in the lower section and outcropping in the north of the basin was determined to be formed of blocky conglomerate derived from metamorphic basement. It is necessary to emphasize this situation as it will be used as data in the discussion in Section 5.2 (Ersoy et al., 2011; Seyitoğlu, 1997*a*).

3.2.3. Selendi and Uşak-Güre Basins

The classic basin stratigraphy in Western Anatolia was created near Uşak by Ercan et al. (1978) and near Selendi by Ercan et al. (1983). The lower Hacıbekir Group comprises the Kürtköyü, Yeniköy and Küçükderbent formations (Figure 32). While the Kürtköyü formation is formed of conglomerates with a single source from ophiolite basement below, above schist and marble fragments are observed and are interpreted as alluvial fan deposits. The conformable

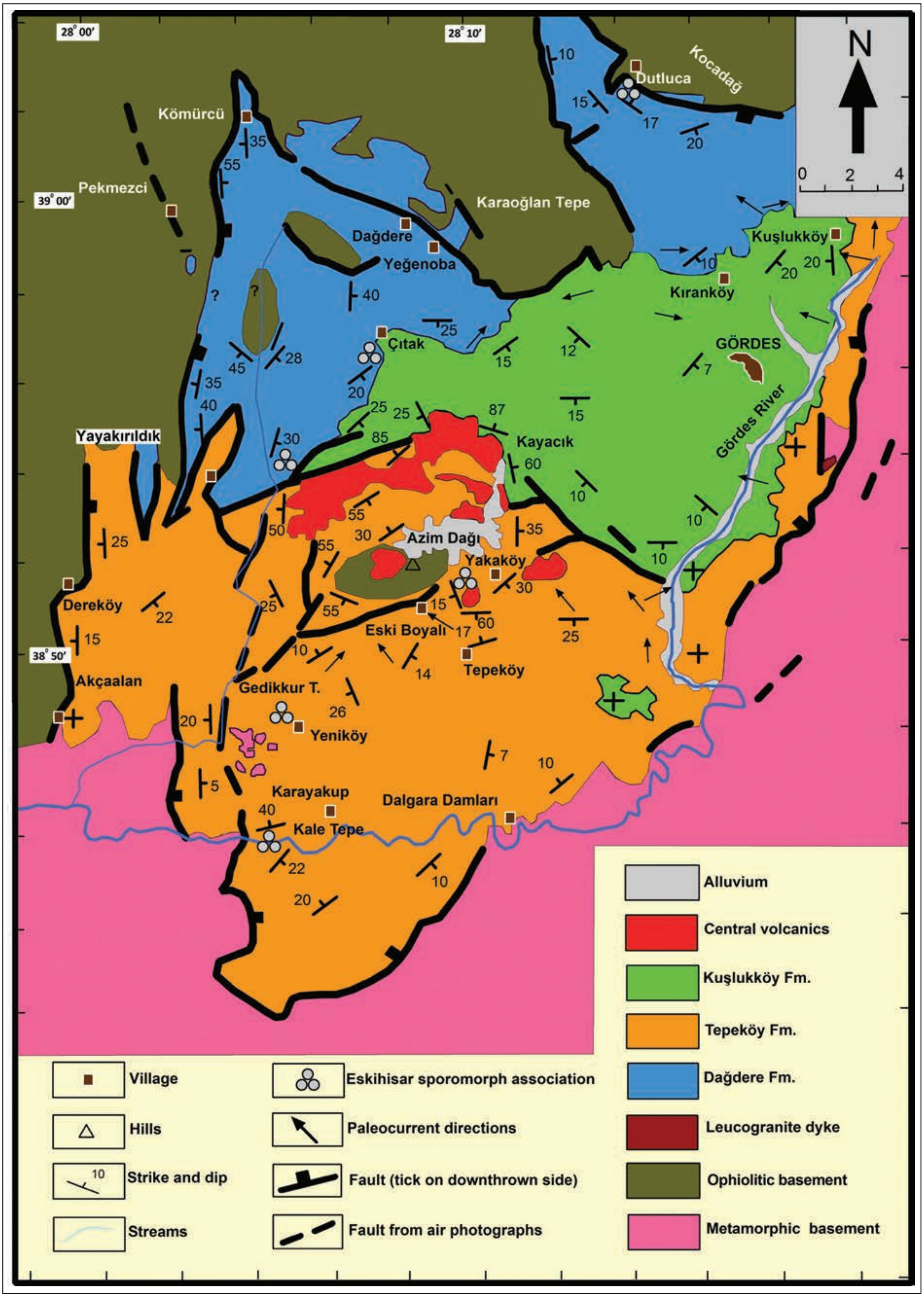

Figure 31- Geological map of the Gördes basin (Taken from Seyitoğlu, 1992; Seyitoğlu and Scott, 1994).

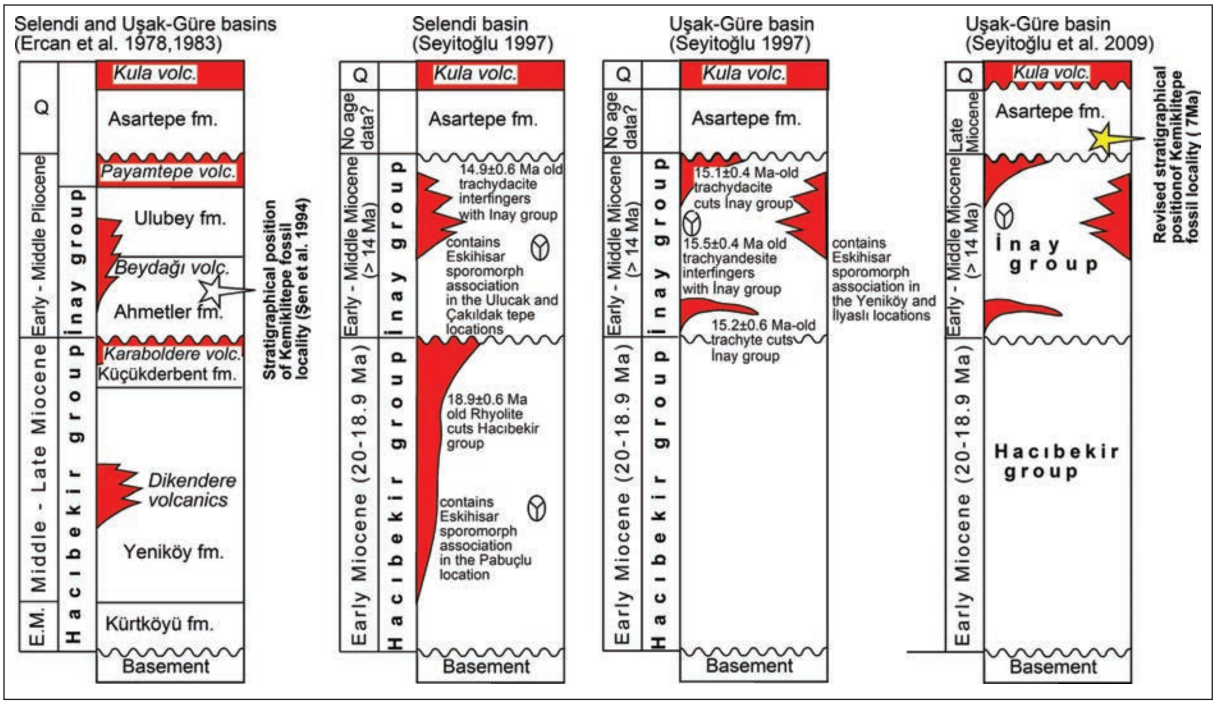

Figure 32 – Stratigraphy of the Selendi and Uşak-Güre basins and revised position of the Kemiklitepe fossil location (Taken from Seyitoğlu et al., 2009).

Yeniköy formation is formed of dark yellow-colored conglomerates, sandstone, claystone, tuffite and clayey limestone layers. Conformably overlying the Yeniköy formation, the Küçükderbent formation is composed of sandstone, claystone, tuffite and marly limestone and includes levels of organic-rich mudstone and gypsum. The İnay Group, unconformably overlying the Hacıbekir Group, is formed of the lower Ahmetler and upper Ulubey formations (Figure 32). The Ahmetler formation is divided into three members, generally formed of light-colored conglomerate, sandstone, claystone, tuffite and marls. The Ulubey formation is composed of lacustrine limestones (Ercan et al., 1978; 1983). Overlying both the Hacıbekir Group and the İnay Group unconformably, the Asartepe formation has red and orange/brown conglomerates and sandstone intercalations. This stratigraphy, along with simultaneously developing volcanic products, was named and stratigraphic positions were determined. It was reported that the Hacıbekir Group was deposited in the Early – Late Miocene, while the İnay Group formed during the Early – Late Pliocene and the Asartepe formation was assessed as Quaternary (Ercan et al., 1978; 1983).

Later studies identified the relationship between volcanic units and sedimentary units in the Uşak-Güre and Selendi basins. Isotopic age dating has been completed and palynological age dates obtained for lignite levels within the sequence (Seyitoğlu, 1997*b*; Seyitoğlu et al., 1997; Seyitoğlu and Benda, 1998).

Within the Hacıbekir Group in the Selendi basin, the Eskihisar sporomorph association (20-14 Ma) was identified, while the age of rhyolites cutting the group was determined as 18.9±0.6 Ma. Within the İnay Group, with an angular unconformity above the Hacıbekir Group, in both the Selendi and Uşak-Güre basins the Eskihisar sporomorph association has been determined. In the Selendi basin 14.9±0.6 Ma trachyandesites interfinger with the İnay Group, in the Uşak-Güre basin 15.2±0.6 Ma and 15.1±0.4 Ma trachite and trachyandesites cut the İnay Group and 15.5±0.4 Ma trachyandesite interfingers the İnay Group (Figure 32). While all of these age data show the Hacıbekir Group was deposited in the Early Miocene, they show the İnay Group formed in the early Middle Miocene. As the İnay Group includes the Eskihisar sporomorph association it cannot be younger than 14 Ma and as a result the Kemikliktepe fossil location (Şen et al., 1994) within the İnay Group and given an age of Late Miocene needs to be re-evaluated. This study once again reveals, with comprehensive data, that the E-W grabens and north-trending basins in Western Anatolia began to form simultaneously under a N-S extensional tectonic regime (Seyitoğlu, 1997*b*).

After using the early Middle Miocene age of the İnay Group and its horizontal position to disprove (Seyitoğlu, 1999) regional compression in the proposal of a two-stage extensional model by Koçyiğit et al. (1999), discussions of the age of the İnay Group are found in a series of articles researching the regional uplift, especially. Westaway et al. (2003; 2004) investigated the uplift history of Western Anatolia by using terraces of the Gediz River covered by Kula volcanics. Using the mammalian fossil content of the Kemiklitepe fossil location, shown within the İnay Group, and magnetostratigraphy (~7 Ma) it was proposed that Gediz river erosion began after the end of deposition of the İnay Group in the Late Pliocene about 3 Ma ago (Westaway et al., 2005; Westaway et al., 2006). Uplift calculations in the region are based on this acceptance and it is proposed that as edge faults of the Uşak-Güre basin are not found, and there is a problem between the isotopic ages of the İnay Group and mammalian fossil ages, palynological ages need to be reworked (Westaway et al., 2005; 2006). Seyitoğlu et al. (2009) reviewed the stratigraphic position of the Kemiklitepe fossil location and as stated in the study first defining the age (Şen et al., 1994) determined it was not within the İnay Group, but contrarily it is in the Asartepe formation unconformably overlying the İnay Group. Additionally the Kemiklitepe fossil locality within the Asartepe formation was compared with the Karabeyli fossil location newly found in the same formation deposited in front of NE trending normal faults in the Uşak-Güre basin and it was determined that the Asartepe formation was deposited in the Late Miocene. This data confirms studies reporting the age of the İnay Group as early Middle Miocene and the isotopic age data and palynological findings (Seyitoğlu, 1997*b*; Seyitoğlu et al., 1997; Seyitoğlu and Benda, 1998) and disproves studies proposing a contradiction between mammalian fossils, and palynological and isotopic age data in Western Anatolia (Figure 32). As a result, it is necessary to reconsider studies (Westaway et al., 2003; 2004; 2005; 2006) extending the deposition of the İnay Group to the Pliocene and beginning erosion around 3 Ma, as well as all uplift models ignoring NE-trending faulting (see Seyitoğlu et al., 2009 for detail).

Other data showing the Asartepe formation deposited in the Late Miocene is found in the Selendi basin (Ersoy and Helvacı, 2007). Here the Kocakuz formation, accepted as equivalent to the Asartepe formation, is covered (Ersoy and Helvacı, 2007) by trachybasalts with ages from 8.5±0.2 Ma and 8.37±0.07 Ma (Ercan et al., 1996; Innocenti et al., 2005).

The volcanic rocks in the Selendi and Uşak-Güre basin have been dated by a more sensitive method (Ar/Ar) (Purvis et al., 2005) and obtained similar values to K-Ar results in Seyitoğlu et al. (1997). Ersoy et al. (2008) defined calcalkaline and alkaline volcanic products with two different compositions from the Early Miocene (20.03-17.87 Ma) interfingering the Hacıbekir Group and determined the presence of bimodal volcanism. The tectono-sedimentary development proposed by this study will be discussed in the latest developments section about exhumation mechanisms of the Menderes Massif (See: Section 5.2).

4. Exhumation Mechanism of the Menderes Core Complex

According to nearly all thermochronological data obtained from the Menderes Massif (Gessner et al., 2001; Ring et al., 2003), the Menderes Massif reached the surface nearly 25-20 Ma ago. Under the control of the Alaşehir and Büyük Menderes detachment faults, the central Menderes Massif appears to have been rapidly exhumed a second time since 5 Ma. Based on this data Ring et al. (2003) determined the Menderes Massif was exhumed as a symmetrical core complex between the north-dipping Simav Detachment Fault and south-dipping Lycian Detachment Fault in the Late Oligocene-Early Miocene. In the Miocene-Pliocene period the Alaşehir and Büyük Menderes detachment faults worked in accordance with the flexural rotation model in the central Menderes massif and it was determined that the symmetric core complex was uplifted once more.

Seyitoğlu et al. (2004) used previous studies and microtectonic data in the massif to propose an alternative model for the complete exhumation history of the Menderes Massif (Figure 33). Accordingly the Menderes Massif was first exhumed as an asymmetric core complex.

The main breakaway fault extends from west to east from south of the Gulf of Gökova following south of the Kale basin toward the northeast (Figure 33). This north-dipping normal fault is named the

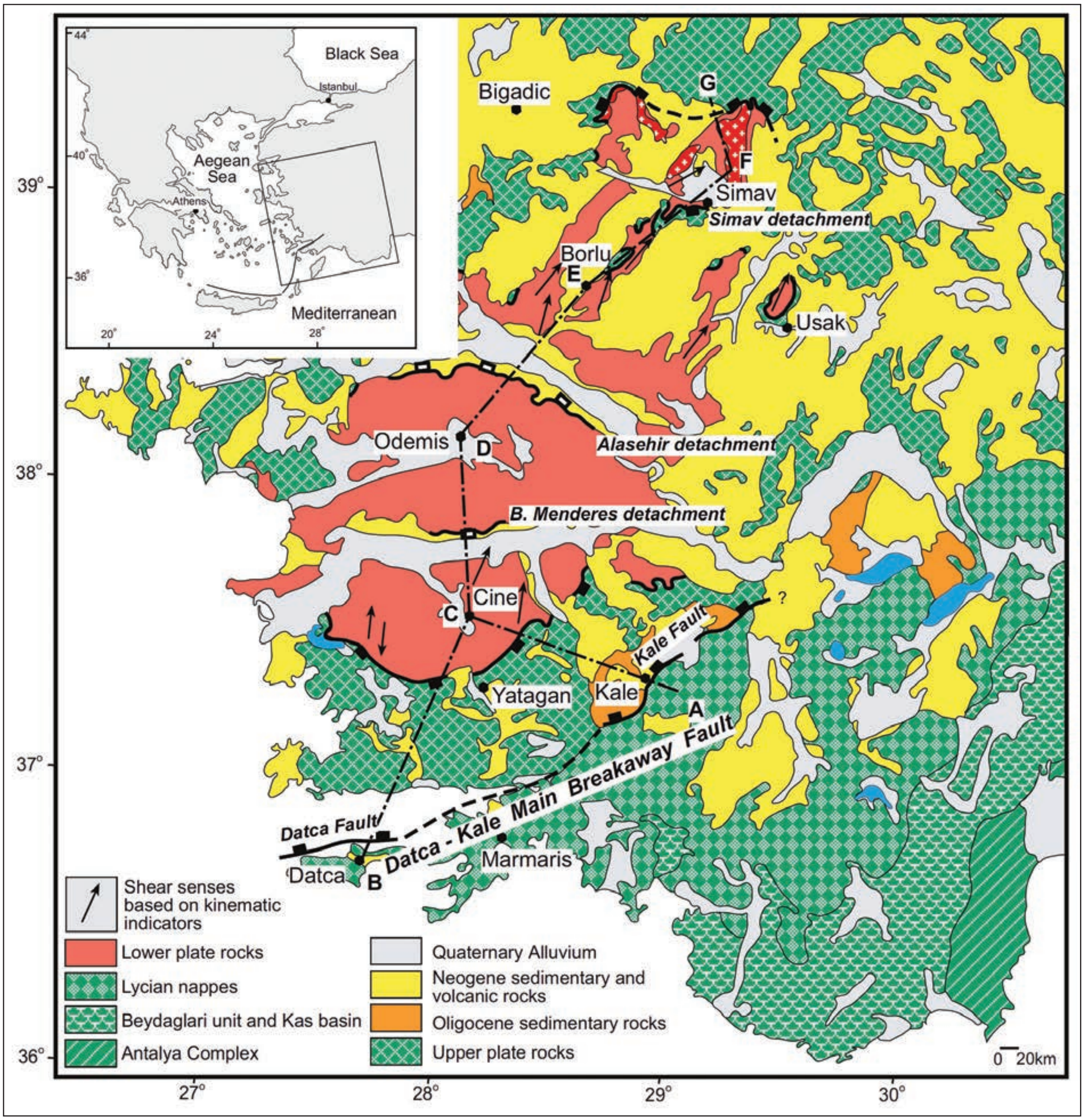

Figure 33- Menderes core complex and location of main tectonic elements in Western Anatolia (Taken from Seyitoğlu et al., 2004).

"Datça – Kale Main Breakaway Fault" (Seyitoğlu et al., 2004) and is clearly seen on submarine seismic reflection profiles in the Gulf of Gökova (Kurt et al., 1999). The north-dipping listric normal fault observed on seismic reflection profiles continues in the land toward the northeast (Çağlar and Duvarcı, 2001). A wedge geometry on the downdropped block thickening toward the main fault is clearly observed and contains a sedimentary sequence that may be said to have deposited simultaneously with faulting (Kurt et al., 1999) (Figure 34).

Though the age of this sequence was interpreted as Late Miocene-Quaternary by Kurt et al. (1999), there is no definite data on this topic. Oligocene conglomerates have been mapped north of the Gulf of Gökova (Gürer and Yılmaz, 2002) controlled by antithetics of the Datça Fault; as a result the sequence observed on the downdropped block of the Datça Fault may possibly be Oligocene (Özerdem et al., 2002). Toward the northeast the Kale basin developed in the Oligocene-Early Miocene (Dürr, 1975; Yılmaz et al., 2000; Akgün and Sözbilir, 2001) begins with coarse conglomerates derived from ophiolite

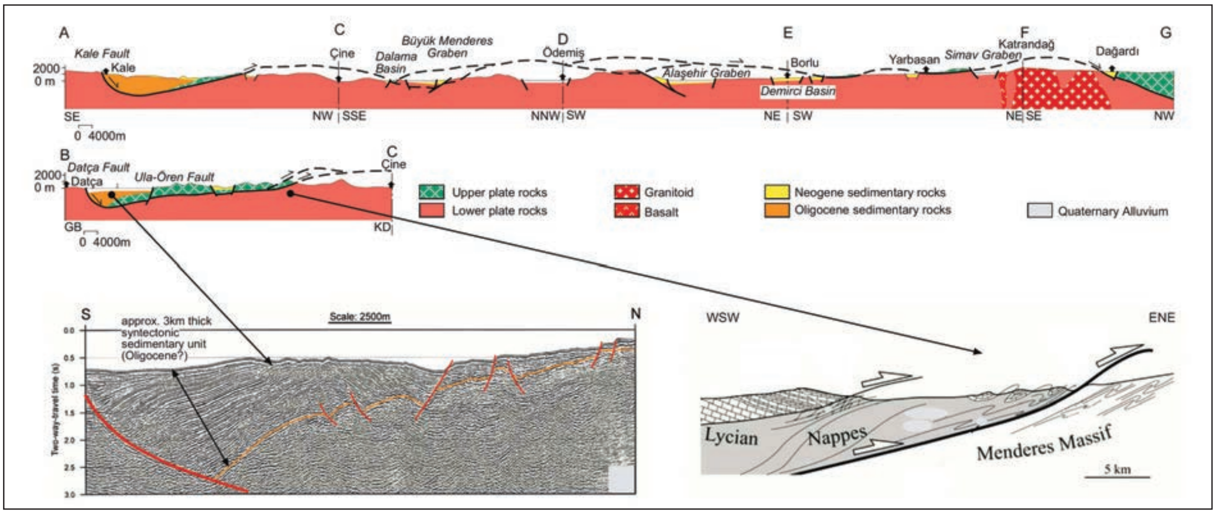

Figure 34- At top, geological cross section including the Simav Detachment Fault north of the Datça-Kale Main Breakaway Fault (Seyitoğlu et al., 2004). Lower left, reinterpretation of seismic reflection profile of the Datça Fault in the Gulf of Gökova from Kurt et al. (1999). Lower right, cross section of the region interpreted as where the Datça-Kale Main Breakaway Fault flexes and comes to the surface (Rimmele et al., 2003). See text for details.

basement and continues with conglomerates, sandstone, siltstone, shale and limestone intercalations. The coarse clastic debris flows at the base of the sedimentary sequence, fluvial deposits show paleocurrent directions from southeast to northwest and are controlled by the Kale Fault in the south of the basin (Gürer and Yılmaz, 2002). Thinning toward the top of the Kale basin, the Oligocene – Lower Miocene sequence is unconformably overlain by the Upper Miocene – Pliocene sequence. The section where the Datça-Kale Main Breakaway Fault flexes upward and reaches the surface is found 6 km north of Yatağan on the Yatağan-Çine road (Figure 35).

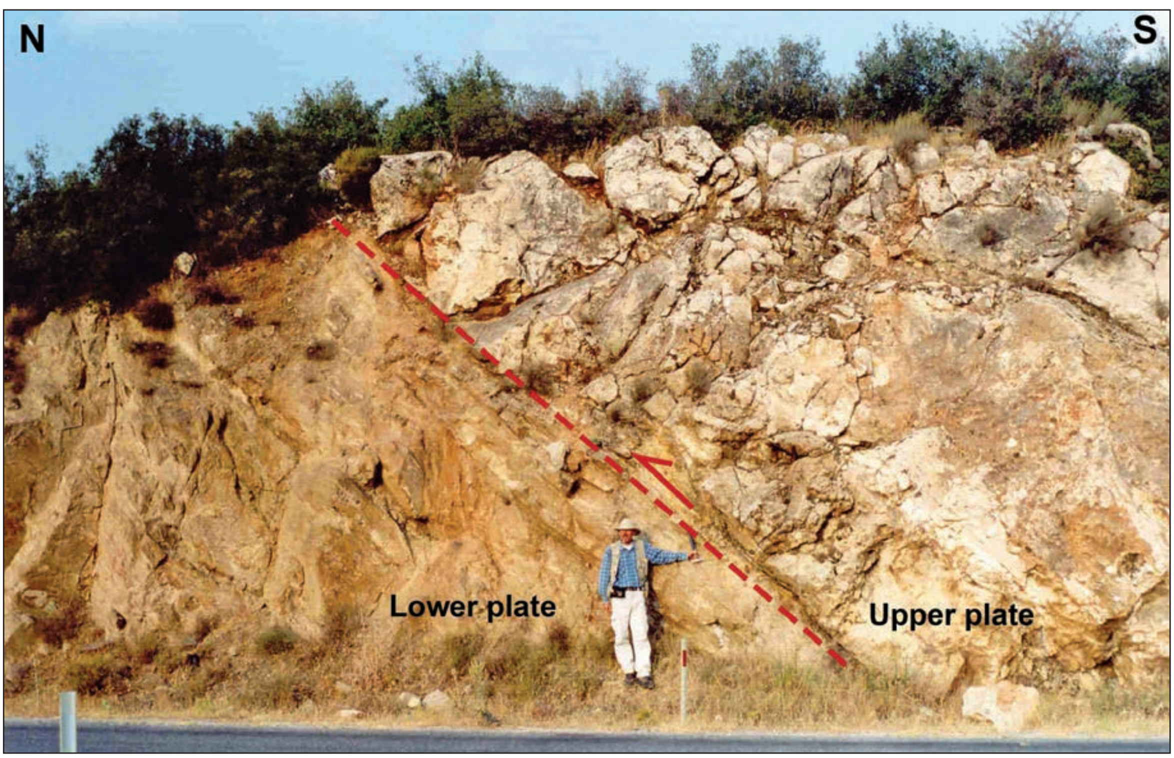

Figure 35- Section where the Datça-Kale Main Breakaway Fault flexes to reach the surface on the Yatağan –Çine road.

The footwall block is overprinted by scattered brittle structures of top-to the NNE shearing. A similar shearing direction is observed between the Lycian ophiolites and the Menderes Massif (Bozkurt and Park, 1999; Rimmele et al., 2003). It is thought that the Datça-Kale Breakaway Fault which flexes up toward the top-to the NNE shearing was formed by the main shear zone (Seyitoğlu et al., 2004) (Figure 34). According to this assessment, Gökova and Kale basin fill, Lycian ophiolites and Menderes Massif cover rocks form the upper plate above the main breakaway fault. From north of the Yatağan-Çine road to beyond Mount Simav, the whole Menderes Massif are lower plate rocks of the Datça-Kale Main Breakaway Fault and its northern continuation, the Simav Detachment Fault. Apart from this, remaining upper plate pieces from the first detachment above the massif are found currently at Dalama south of Aydın, in the southeast of the Gördes basin, in the east of the Demirci basin and south of Simav. When all data are evaluated, according to this model (Seyitoğlu et al., 2004), which is in accordance with thermochronological data, explains basin development, provides a logical explanation for the contradictory north and/or south movement within the Lycian nappes, and does not contradict geological observations, the exhumation of the Menderes Massif occurred in the following way (Figures 36 and 37).

In the Late Paleocene-Early Eocene continental collision between the Menderes-Taurus block and the Sakarya continent occurred along the Izmir-Ankara suture zone and the Lycian nappes were emplaced above the Menderes Massif causing main Menderes metamorphism in the Late Eocene (Şengör et al., 1984) (Figure 36a).

After completion of Lycian nappe emplacement, the orogeny developing as a result of collision began extension in a N-S direction. In the Oligocene at the surface the north-dipping Datça-Kale Main Breakaway Fault controlled deposition in the Gökova and Kale basins, while in the middle crust top-to-the N directed shearing occurred. This shearing is dated to 43-30 Ma in the South Menderes Massif (Hetzel and Reischmann, 1996; Lips et al., 2001; Catlos et al., 2002); however debates continue about this date (Gessner et al., 2004; Bozkurt, 2004; Erdoğan and Güngör, 2004). On the Simav shear zone, the northern continuation of the Datça-Kale Main Breakaway Fault, syntectonic intrusion of the Eğrigöz granitoid occurred at 22 Ma (Işık et al., 2003; 2004) (Figure 37a,b). This age data (Ar-Ar and apatite fission-track) generally is observed to become younger toward the north. Flexure of the Datça-Kale Main Breakaway Fault upwards began around 25 Ma according to thermochronological data. This flexure caused the development of apatite fission-track ages which young to the south in the southern Menderes Massif. Finally the flexure of the main breakaway fault brought lower plate rocks to the surface (Figures 36c and 37b) at 20 Ma as shown by thermochronological data (Gessner et al., 2001). Micro tectonic data obtained from the Menderes Massif show that post Eocene movement direction was top-to-the NNE. However, the top-to-the NNE structures in the southwest of the massif were overprinted by weaker top-to-the SSW structures (see also: Hetzel et al., 1998; Lips et al., 2001; Bozkurt, 2004). This situation may be related to the Menderes Massif having a dome shape and the main breakaway fault slipping slightly south (Seyitoğlu et al., 2004).

In the interval from the Oligocene-Early Miocene uplift of the footwall of the Datça-Kale Main Breakaway Fault caused movement of the Lycian nappes to the south due to gravity sliding and the final Lycian nappe emplacement in the Burdigalian (Seyitoğlu et al., 1992; Collins and Robertson, 1998; 2003; Seyitoğlu et al., 2004).

The dome-shaped uplift of the Menderes Massif was fragmented by the E-W Alaşehir, Büyük Menderes and Denizli grabens and north-trending basins in the Early Miocene (Seyitoğlu, 1997; Seyitoğlu et al., 2002; Şen and Seyitoğlu, 2009; Alçiçek et al., 2007). Due to the flexural rotation/rolling hinge of the Alaşehir and Büyük Menderes grabens, the central Menderes Massif was exhumed for a second time, this time as a symmetric core complex (Gessner et al., 2001; Seyitoğlu et al., 2002) (Figures 36c,d and 37c,d,e). In the Pliocene to Quaternary young grabens developed (e.g., Simav), other main grabens became symmetric and high-angle faults fragmented older structures masking the previous extensional history of the Menderes Massif (Seyitoğlu et al., 2004) (Figure 37f).

5. Discussion

5.1. Movement of the Lycian Nappes in Southwest Turkey and First Exhumation of the Menderes Massif

In Southwest Anatolia in the area between Lake Bafa and the Gulf of Gökova, movement of the Lycian nappes above the Menderes Massif was

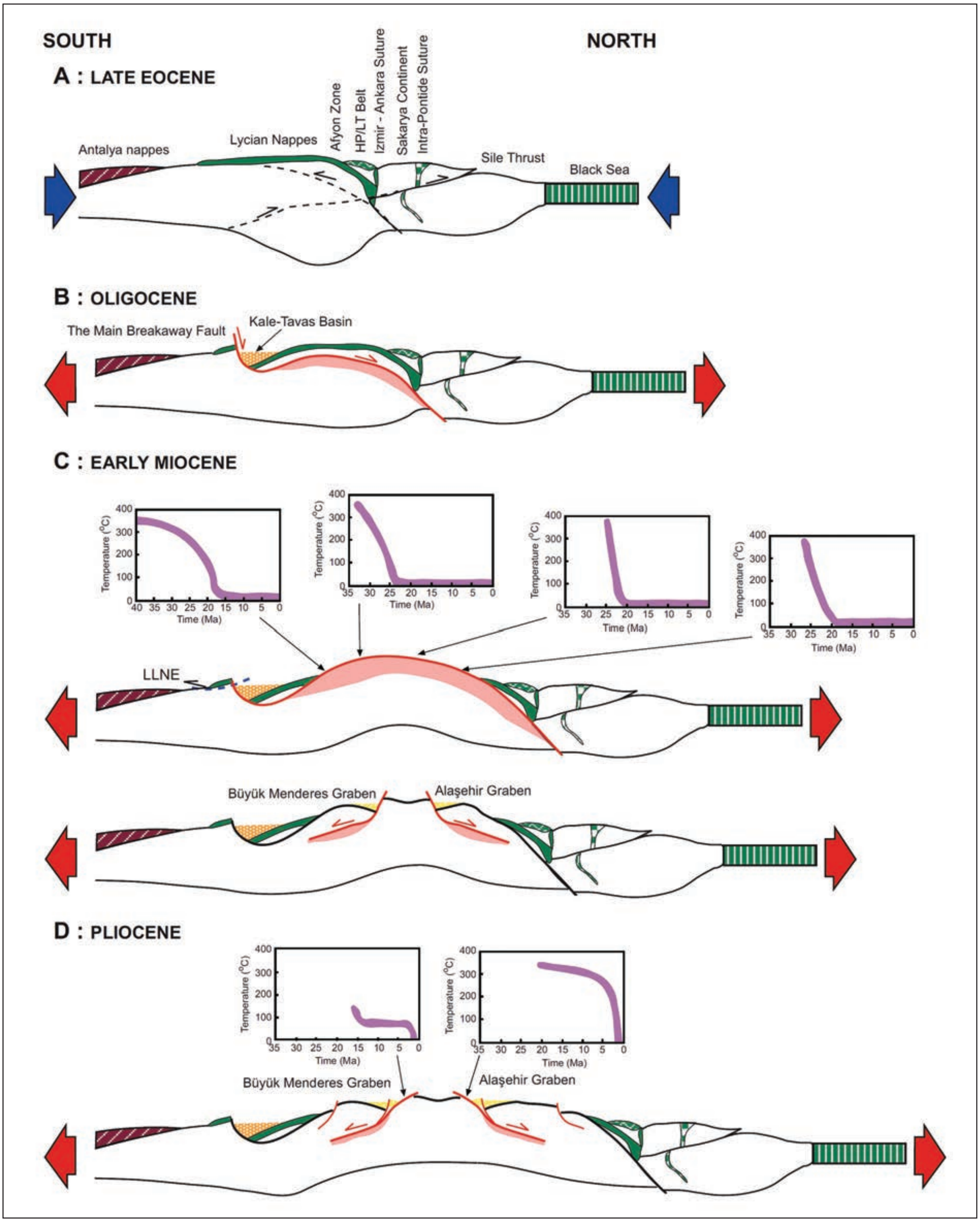

Figure 36- Two stage exhumation model of the Menderes Massif as asymmetric and symmetric core complex (Taken from Seyitoğlu et al., 2004). Thermochronologic ages belong to Gessner et al. (2001) and Ring et al. (2003). LLNE: Last Lycian Nappe Emplacement.

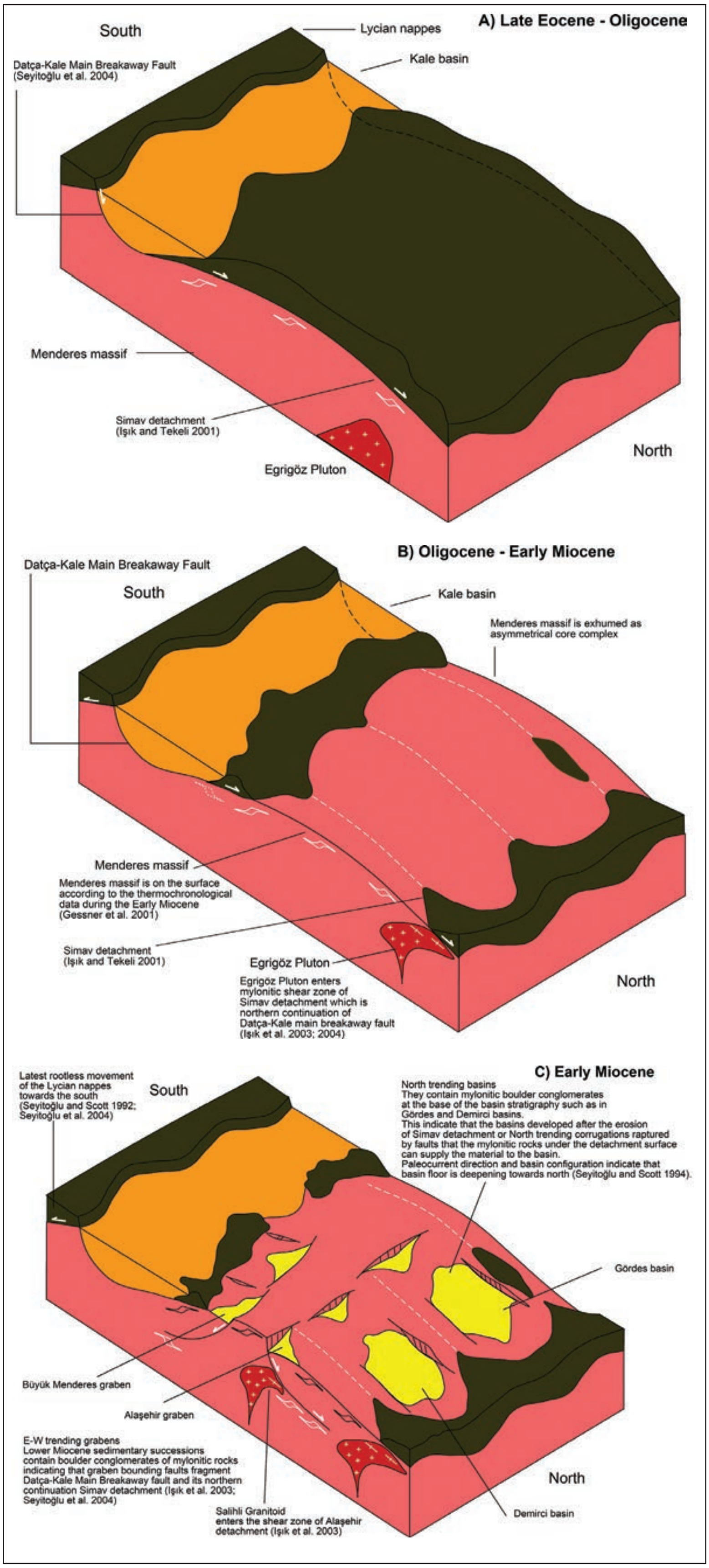

Figure 37- (A-B-C) Three-dimensional representation of exhumation of the Menderes Massif as asymmetric and symmetric core complexes.

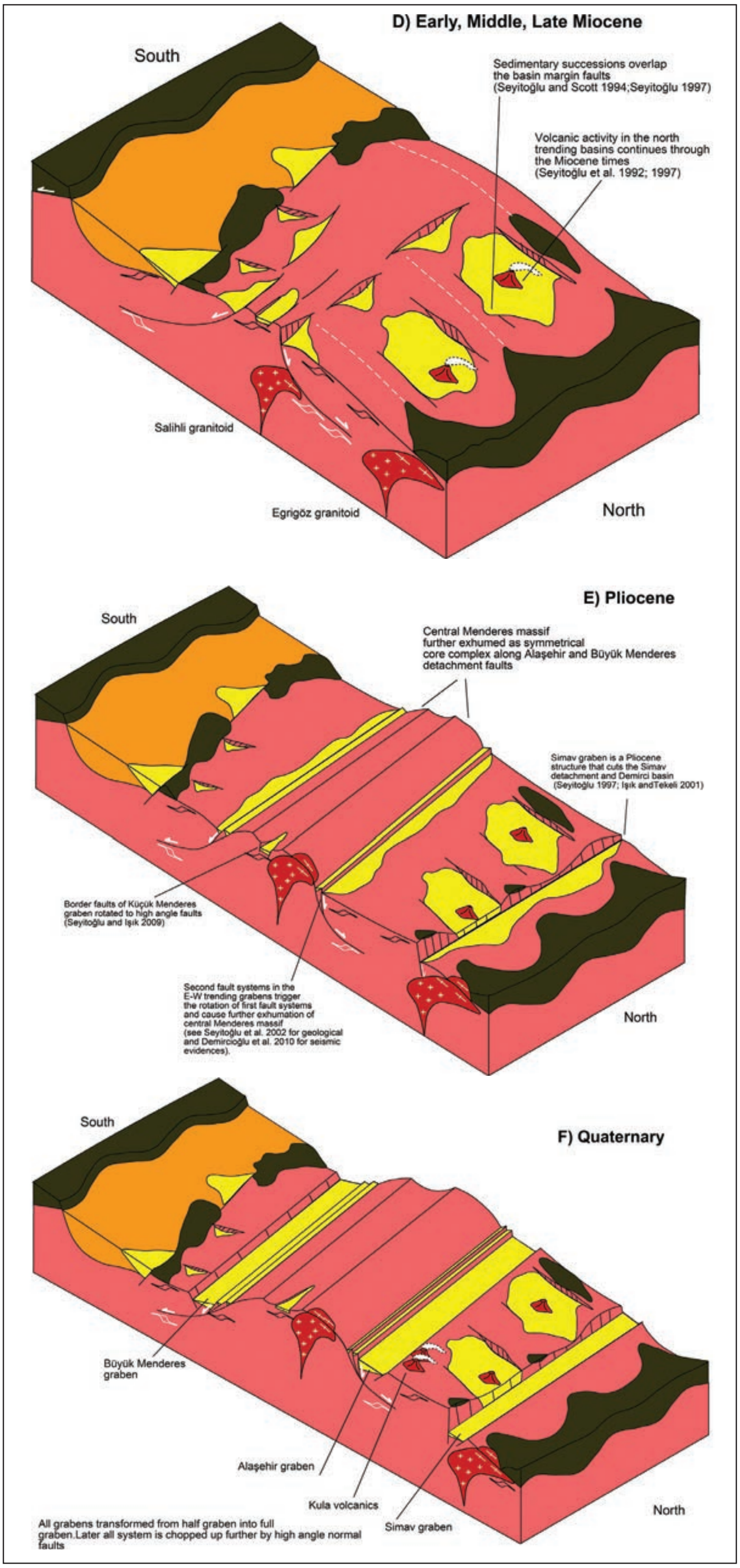

Figure 37- (D-E-F) Three-dimensional representation of exhumation of the Menderes Massif as asymmetric and symmetric core complexes.

toward the north (Bozkurt and Park, 1999; Rimmele et al., 2003). This data does not comply (Seyitoğlu et al., 2004) with the assumption that the Menderes Massif was sheared with top-to the S shearing (Ring et al., 2003) by the south-dipping Lycian Detachment Fault in the Oligocene-Early Miocene. On the other hand, the S-SE movement (Collins and Robertson, 2003) of the Lycian nappes between the Menderes Massif and Beydağları gives the impression of the contradiction about movement direction of the Lycian nappes in southwest Anatolia (van Hinsbergen, 2010). In fact the model of the first exhumation of the Menderes Massif as asymmetric (Seyitoğlu et al., 2004) provides a logical explanation of the two opposite directional movements of the Lycian nappes. In the area remaining north of the Datça-Kale Main Breakaway Fault, the movement to the north (Rimmele et al., 2003) observed for the Lycian nappes between Lake Bafa and the Gulf of Gökova is in accordance with shearing as the Datça-Kale Main Breakaway Fault reached the surface through upward flexure. The south-southeast movement of the Lycian nappes south of the Datça-Kale Main Breakaway Fault is related to the rootless gravity sliding to the south of the Lycian nappes as a result of uplift of the footwall of the main breakaway fault (Seyitoğlu et al., 2004).

5.2. Latest Developments about Exhumation Mechanisms for the Menderes Massif

On the exhumation of the Menderes Massif, van Hinsbergen (2010) proposed that as the Menderes Massif was exhumed with a top-to-the NE direction on the Simav Detachment Fault, Lycian nappes were detached above the massif toward the southeast. The element of this proposal that requires explanation is that while the movement developing on the Simav Detachment Fault in the north Menderes Massif left traces, the Lycian detachment proposed by van Hinsbergen (2010) left no trace on the massif. The dominant kinematic data from the south Menderes Massif has top-to the N direction, slightly overprinted by a top-to the S directed movement (Seyitoğlu et al., 2004). The proposal of van Hinsbergen (2010), similar to that of Ring et al. (2003), does not appear to comply with kinematic indicators in the south Menderes Massif.

A recent study investigating the complete exhumation history of the Menderes Massif (Gessner et al., 2013) proposed that the first exhumation of the massif occurred with top-to-the N directed unilateral shearing and shows it is one step closer to the asymmetric exhumation model than the symmetric exhumation model (Ring et al., 2003) by the same researchers. However, this proposal does not include any recommendations related to the necessary sedimentary basin forming in the hanging wall of the main breakaway fault (Gessner et al., 2013). In this study, Gessner et al. (2013) proposed a left directed shearing west of the Menderes Massif named the "Western Anatolian Transfer Zone". Seyitoğlu et al. (2004) stated the necessity for a transfer zone linking the main breakaway fault north of Crete proposed by Faure et al. (1991) with the Datça-Kale Main Breakaway Fault, and advocated that the Cyclades and Western Anatolia had a common extensional history (Seyitoğlu et al., 2004) (Figure 33, attached small sketch).

Recent studies encompassing only the north Menderes Massif and aiming to explain the Simav and Alaşehir detachment faults and north-trending basin development will be discussed below in terms of compliance with field observations.

The proposed model for the evolution of the northern Menderes Massif in the article by Ersoy et al. (2010) occurs in advanced stages within an extensional tectonic regime. In the Late Oligocene-Early Miocene the Simav Detachment Fault developed along with deposits of the Hacıbekir Group and bimodal volcanism above it. In the Middle Miocene, the Gediz (Alaşehir) detachment fault formed and the deposits of the İnay Group above it. On the hanging wall of the Gediz (Alaşehir) detachment fault, cross grabens bounded by oblique normal/strike slip faults developed and controlled the deposition of the İnay Group and Kocakuz (Asartepe) formation. At this stage, the noteworthy elements of the proposed model may be listed as: (a) the Gediz (Alaşehir) graben (basin) begins as a low angle detachment fault, (b) contains coeval sediments to the Middle Miocene-age İnay Group, and (c) in the Plio-Quaternary period high-angle normal faults form E-W trending symmetric Gediz (Alaşehir) and Simav grabens. The opposing opinions to the above model proposed in the article by Ersoy et al. (2010) may be summarized as follows.

(1) First exhumation of the north Menderes Massif by the Simav Detachment fault the stage where the massif is cleared of ophiolite cover (Lycian nappes) above the Simav Detachment Fault is not well reflected to the sediments of the Hacıbekir Group. The measured stratigraphic section given in Ersoy et al. (2010) has thickness of ophiolite-source

clasts as around 10 m. After such a small amount of ophiolitic material, there are conglomerates derived from rock fragments from the massif below.

The impression Ersoy et al. (2010; 2011; 2014) wish to create in their articles is the presence of the well-developed Kürtköyü formation derived from ophiolitic basement at the bottom of all north-trending basins. If this can be shown it supports deposition of the Hacıbekir Group simultaneous to the Simav Detachment Fault.

The Kürtköyü formation was first investigated in the Uşak-Güre basin in an article by Ercan et al. (1978) and was described as dominantly clastic material derived from ophiolitic basement (majority conglomerates, sandstone). However, in the Selendi basin the Kürtköyü formation is not differentiated on the geological map of the Selendi Basin in Ersoy et al. (2010) (or has such a small area that it cannot be shown on that scale of map), while the text says the conglomerate at the base of the Kürtköyü formation is metamorphic origin and emphasizes that the conglomerates in the Uşak-Güre basin mentioned in passing have ophiolitic origin. On the other hand, measured stratigraphic sections in both basins show ophiolitic-origin conglomerates. Their thickness is too thin for them to be the first detachment material considering the gigantic presence of Lycian nappes causing main metamorphism in the Menderes Massif.

In Ersoy et al. (2011) the Kürtköyü formation with proposed presence in the north of the Demirci basin is blocky conglomerate derived from metamorphic basement (according to our personal observations these blocks include mylonitic rock fragments observed below the detachment fault) which does not fit the description of the original Kürtköyü formation. This formation is more in accordance with the description of the Borlu formation by Yılmaz et al. (2000). The Borlu formation shown by Yılmaz et al. (2000) at the base of the Demirci basin, is carried to the upper sections of the sequence by Ersoy et al. (2011) (See Ersoy et al., 2011; Figure 3). Shown underneath the Yeniköy formation in the Demirci basin (Ersoy et al., 2011; Figure 9b) the Kürtköyü formation is not the original Kürtköyü formation described by Ercan et al. (1978) formed of ophiolitic-origin rock fragments.

In the Gördes basin the Tepeköy formation, coeval with the Hacıbekir Group as revealed by isotopic age dating and explained in detail in Section 3.2.1, is formed fully of mylonitic material derived from metamorphic rocks while the Dağdere formation which interfingers with it, contains material derived from dominantly ophiolitic rock (Seyitoğlu, 1992; Seyitoğlu and Scott, 1994). This observation indicates that in some areas of the massif, north-trending basins formed after being denuded of ophiolitic rock with the basins filling with clasts derived from whatever lithology was located as the basement locally. However, Ersoy et al. (2011) ignored this distinction in the Gördes basin and combined formations containing conglomerates with different composition and named them the Kızıldam formation. This formation is correlated with the Kürtköyü formation observed in other north-trending basins (Ersoy et al., 2010). When it is considered that the above explanation is incompatible with the original Kürtköyü formation in the Demirci basin, it is clear that the correlations made will create confusion in basin stratigraphy.

In the geological map of the Gördes basin presented by Ersoy et al. (2011) the conglomeratic units of the lower levels of the basin deformed by volcanics near the central volcanics are shown as the upper levels of the Kuşlukköy formation in basin stratigraphy. Apart from this the tuff levels used as a characteristic marker for the original Kuşlukköy formation definition (Seyitoğlu, 1992; Seyitoğlu and Scott, 1994) appear to not to have been taken into account by Ersoy et al. (2011). The clastic Kızıldam formation, the lower part of basin fill, overlaps ophiolitic basement on the north of Dağdere (Ersoy et al., 2011), in fact this overlapping occurred with limestone in the upper levels of the Dağdere formation (Seyitoğlu, 1992). The faulted/overlapped relationship of the Tepeköy formation, with palynological samples from lignite levels in the southwest of the Gördes basin identified to contain the Eskihisar sporomorph association, was determined to have deposited under control of high angle faults since the beginning of the Gördes basin (Seyitoğlu, 1992; Seyitoğlu and Scott, 1994); however maps for this section of the basin are not observed in the study by Ersoy et al. (2011).

In conclusion, the Hacıbekir Group or equivalent sedimentary units alleged to have been deposited above the Simav Detachment Fault do not contain significant thicknesses of sedimentary material derived from the upper plate (mainly ophiolitic rocks), with the initial sediments in some basins observed to contain mylonitic rock fragments from below the detachment fault. In the location where the

Kürtköyü formation is found in accordance with the original description (Ercan et al., 1978), there are sections with patches of ophiolitic fragments belonging to the upper plate of the Simav Detachment Fault. As a result, this casts doubt on the claim that the Hacıbekir Group was deposited simultaneously to the Simav Detachment Fault.

(2) The photograph showing the relationship between the Simav Detachment Fault and the proposed coevally deposited Hacıbekir Group (Ersoy et al., 2010; Figure 9b) does not show slipped sediments above the detachment fault but shows a buttress unconformity.

(3) In the model by Ersoy et al. (2010), it is proposed that the first stage of the Gediz (Alaşehir) basin is controlled by low angle detachment faults. The counter argument (See: Section 3.1.1) to the similar proposal by Öner and Dilek (2011) is valid for the model by Ersoy et al. (2010). Additionally the first sedimentary unit in the Alaşehir (Gediz) graben of the Alaşehir formation was deposited in the Early Miocene, which is inconsistent with the proposed Gediz (Alaşehir) graben in Ersoy et al. (2010) presented as beginning in the Middle Miocene. The Early Miocene age data (Catlos et al., 2010) obtained above the Alaşehir Detachment Fault is not explained in the Ersoy et al. (2010) model.

(4) In the last stage of the Ersoy et al. (2010) model (Plio-Quaternary), high angle faults cut the low angle Gediz (Alaşehir) detachment fault. As a result, movement is not expected on this low angle fault. However, age data above the Alaşehir Detachment Fault (Gessner et al., 2001; Buscher et al., 2013) shows activity on the Alaşehir Detachment Fault in the Plio-Quaternary period. Age data from above the Alaşehir Detachment Fault in the Early Miocene to Quaternary interval show the initial high-angle normal faults of the Alaşehir (Gediz) graben rotated and activity continued to develop in a “rolling hinge” model (Seyitoğlu et al., 2002), or in other words an “Alaşehir type - rolling hinge model” developed. Readers can reach collected age data above the Alaşehir Detachment Fault in the article by Seyitoğlu et al. (2014).

Karaoğlu and Helvacı (2012) proposed a similar model to the one discussed above by Ersoy et al. (2010). The tectono-stratigraphic sequence presented by Karaoğlu and Helvacı (2012; Figure 3) shows a physical contact between the Ahmetler and Ulubey formations and the Gediz (Alaşehir) detachment fault (Ersoy et al., 2010). Such a physical contact is not observed in the field, is not reflected in field relationships and is based on assumption. The Menderes Massif did not reach the surface in the Early Miocene near Uşak-Güre as asserted by Karaoğlu and Helvacı (2012) and it is stated that the Hacıbekir Group does not contain rock fragments from the Menderes Massif. But even in the upper levels of the Kürtköyü formation at the bottom of the Hacıbekir Group, there are metamorphic pebbles reported (Ercan et al., 1983). In the Gördes and Demirci basins, containing coeval sediments to the Hacıbekir Group, blocky conglomerates derived from the Menderes Massif are clearly observed. Similarly in the Selendi basin at Pabuçlu village metamorphic rock fragments are clearly seen in the Hacıbekir Group (Figure 38). Additionally within the Uşak-Güre basin the typical Yeniköy formation within the Hacıbekir Group observed along the Uşak-Kütahya road contains metamorphic pebbles within conglomeratic levels. At Eynehan village, the tilted Yeniköy formation is overlain above an unconformity by the İnay Group and metamorphic pebbles are clearly seen in the conglomeratic levels (Figure 39).

All of these observations show the Menderes Massif was already exhumed under the Simav Detachment Fault during sedimentation of the Yeniköy formation (Hacıbekir Group) and provided material for the Yeniköy formation. The fault shown between the Menderes Massif and the Yeniköy formation NE of Kurtçamı on the geological map presented in Karaoğlu and Helvacı (2012) does not have “detachment” features but is a moderate-angle normal fault. The cover unit of this fault is not the Merdivenlikuyu member as shown on the map, but belongs to the Asartepe formation unconformably overlying the İnay Group. The most striking relationship mapped by Karaoğlu and Helvacı (2012) is the presence of sediments belonging to the Yeniköy formation dipping toward a very low angle detachment fault near Kadiroğlu village (Karaoğlu and Helvacı, 2012). Foliation of metamorphic rocks and layering of sedimentary units above them are nearly parallel in this area with an overlapping relationship. Near to the Kadiroğlu village, İnay Group sediments have turned yellow from the heating effect of the Zahman volcanics. Moving from here, the sedimentary units colored yellow due to the heating effect may be confused with the Yeniköy formation. The true Yeniköy formation outcrops on the lower altitudes inside the valley to the east of Kadiroğlu village.

Figure 38- Metamorphic blocks in tilted blocky conglomerates of the Hacıbekir Group at Pabuçlu village in SW Selendi basin overlain above an unconformity by the nearly-horizontal İnay Group. Clear examples shown by red arrows. Length of the pickaxe is 80 cm.

The presence of metamorphic pebbles within sediments in the lower sections of the north-trending basins; in the Tepeköy formation of the Gördes basin (Seyitoğlu and Scott, 1994), in the Borlu formation of the Demirci basin (Yılmaz vd., 2000), and in the Hacıbekir Group of the Selendi and Uşak-Güre basins (Seyitoğlu, 1997) show that during deposition the Menderes Massif was already at the surface and as a result indicate that the model presented by Karaoğlu and Helvacı (2012; Figure 13) is invalid.

6. Conclusions and Recommendations

The exhumation of the Menderes Massif in the Oligocene as an asymmetric core complex caused the formation of the Oligocene Kale basin in the hanging wall of the Datça – Kale Main Breakaway Fault (Seyitoğlu et al., 2004). The observed field relationships and other observations shed doubt on the claim that the coeval sediments of Simav Detachment Fault are sediments supposedly belonging to the Hacıbekir Group observed in the north-trending basins. This situation should lead to a search for an answer to the question: if coeval sediments to the Simav Detachment Fault are not the Hacıbekir Group, what group are they? Outcropping in a very small area of the north Menderes Massif the Başlamış formation is described as an Eocene-Oligocene (?) sedimentary unit (Akdeniz, 1980) and no other Oligocene-age sedimentary outcrop is known.

Detailed kinematic analysis of the Kazdağ core complex in the Biga Peninsula has shown the first exhumation of the Kazdağ core complex was top-to-the N directed (Kurt et al., 2010). This observation allows us to speculate that the Simav Detachment Fault passed under the ophiolites of the Izmir-Ankara Suture Zone and over the Kazdağ core complex to reach Marmara. If this is true, investigation of the relationship of the Oligocene magmatism in south Marmara with the detachment is necessary and

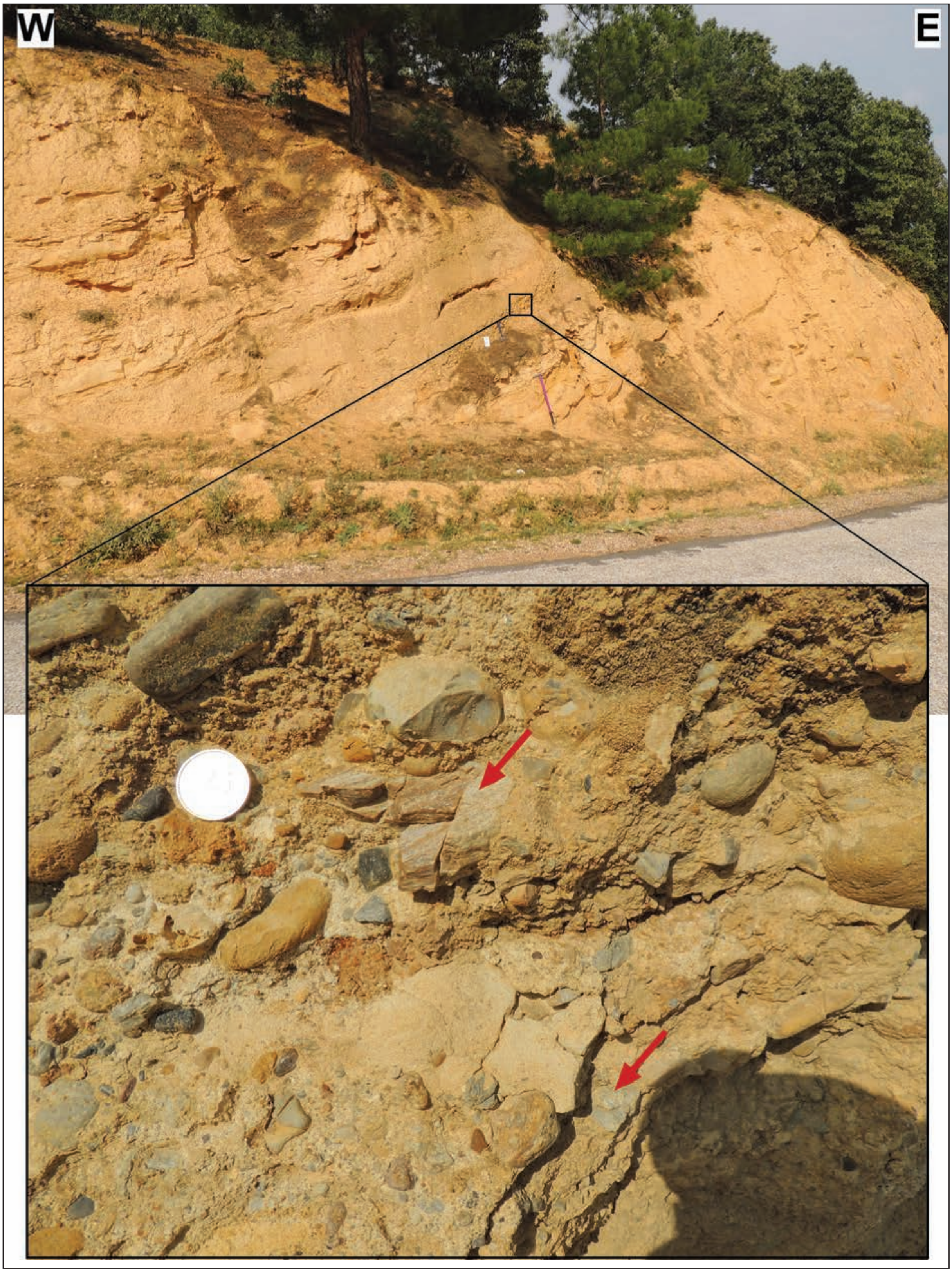

Figure 39- Metamorphic pebbles, shown by red arrows, within conglomerate levels in the tilted yellow-colored Hacıbekir Group on the Eynehan road in the Uşak-Güre basin. Length of the pickaxe is 80 cm. Scale on the close up is 25 kuruş coin.

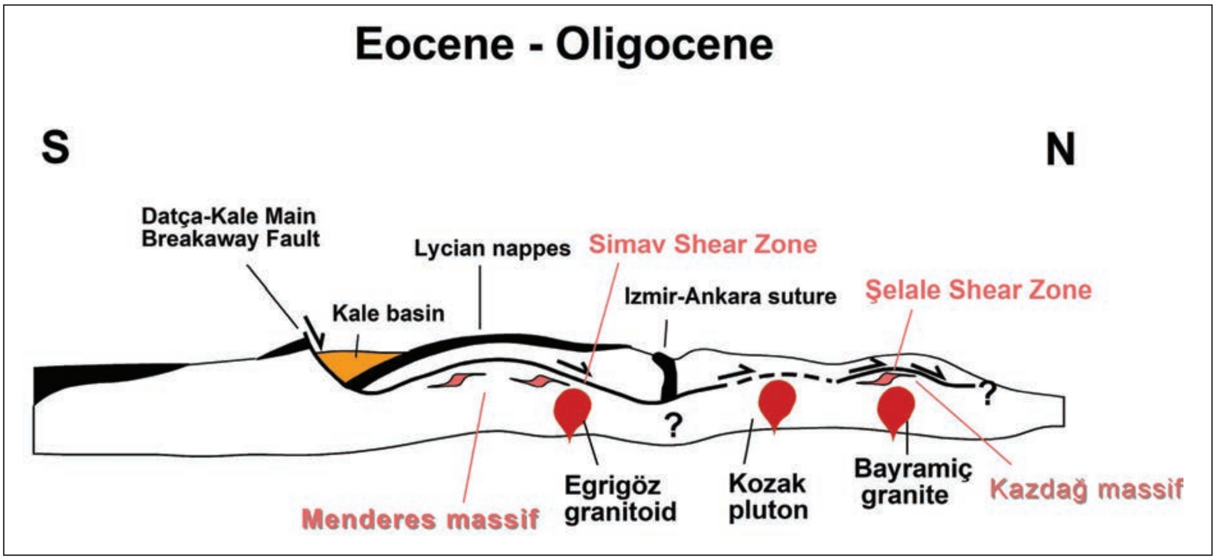

Figure 40- Sketch showing possible relationship between the Menderes core complex and the Kazdağ core complex.

possible candidates for remnants of the lower plate of the Uludağ Massif and Marmara Island granites should be reinvestigated. In this situation, the candidate for the sedimentary basin developing further north of the Simav Detachment Fault is the Eocene-Oligocene Thrace basin. If this is proven the indication that the ophiolites known as the Izmir-Ankara suture zone is a true suture zone will come under suspicion and these will have to have moved above the Simav Detachment Fault (Figure 40). Testing this hypothesis in the field, based on the data that the first detachment direction was top-to-the N in the Kazdağ core complex, will help us better understand the late Cenozoic extensional tectonics in Western Anatolia.

Acknowledgements

We gratefully remember our supervisors Barry C. Scott and Okan Tekeli for illuminating our path with their direction during our doctoral studies in Western Anatolia. We bow to the respected memories of Tuncay Ercan, Orhan Kaya, Erol Akyol, Okan Tekeli and Leopold Benda who labored on the geology of Western Anatolia and have left our presence. We thank all our colleagues working hard in Western Anatolia producing geological data and making scientific discussions possible, and Taner Ünlü who suggested this article for this special publication.

Received: 08.05.2015
Accepted: 25.07.2015
Published: December 2015

References

Akçığ, Z. 1988. Batı Anadolu'nun yapısal sorunlarının gravite verileri ile irdelenmesi. *Türkiye Jeoloji Bülteni,* 31, 63-70.

Akdeniz, N. 1980. Başlamış Formasyonu. *Jeoloji Mühendisliği,* 10, 39-47.

Akgün, F., Akyol, E. 1999. Palynostratigraphy of the coal bearing Neogene deposits in Büyük Menderes graben, western Anatolia. *Geobios,* 32, 367-383.

Akgün, F., Sözbilir, H. 2001. A palynostratigraphic approach to the SW Anatolian molasse basin: Kale-Tavas molasse and Denizli molasse. *Geodynamica Acta,* 14, 71-93.

Akyol, E., Akgün, F. 2001. Response a Seyitoğlu G. and Sen S. 1999. 'Discussion on Akgün F. and Akyol E. 1999, Palynostratigraphy of the coal bearing Neogene deposits in Büyük Menderes graben, western Anatolia. Geobios, 32,6:915. *Geobios,* 34, 109-112.

Alçiçek, H. 2007. Denizli havzası (Sarayköy-Buldan bölgesi, GB Türkiye) Neojen çökellerinin sedimantolojik incelemesi. Doktora Tezi, Ankara Üniversitesi.

Alçiçek, H., Varol, B., Özkul, M. 2007. Sedimentary facies, depositional environments and palaeogeographic evolution of the Neogene Denizli Basin of SW Anatolia, Turkey. *Sedimentary Geology,* 202, 596-637.

Altunel, E. 1996. Pamukkale travertenlerinin morfolojik özellikleri, yaşları ve neotektonik önemleri. *Maden Tetkik ve Arama Dergisi,* 118, 47-64.

Armstrong, R.L. 1972. Low-angle (denudation) faults, hinterland of the Sevier orogenic belt, eastern Nevada and western Utah. *Geological Society of America Bulletin,* 83, 1729-1754

Arpat, E., Bingöl, E. 1969. The rift system of the Western Turkey; thoughts on its development. *Maden Tetkik ve Arama Genel Müdürlüğü Dergisi,* 73, 1-9.

Ateş, A., Kearey, P., Tufan, S. 1999. New gravity and magnetic anomaly maps of Turkey. *Geophysical Journal International,* 136, 499-502.

Axen, G.J., Bartley, J.M. 1997. Field test of rolling hinges: Existence, mechanical types and implications for extensional tectonics. *Journal of Geophysical Research,* 102, 20515-20537.

Becker-Platen, J.D. 1970. Lithostratigraphisce Untersuchungen im Kanozoikum Südwest- Anatoliens (Türkei). *Beihefte zum geologischen Jahrbuch* 97.

Becker-Platen, J.D. 1971. Stratigraphic division of the Neogene and oldest Pleistocene in Southwest Anatolia.*Newsletters on Stratigraphy,* 1-3, 19-22.

Benda, L. 1971. Principles of the palynologic subdivision of the Turkish Neogene. *Newsletters on Stratigraphy* 1-3, 23-26.

Benda, L., Innocenti, F., Mazzuoli, R., Radicati, F., Steffens, P. 1974. Stratigraphic and radiometric data of the Neogene in Northwest Turkey. *Zeitschrift der Deutschen Geologischen Gesellschaft* 125, 183-193.

Benda, L., Meulenkamp, J.E. 1979. Biostratigraphic correlations in the Eastern Mediterranean Neogene. 5. Calibration of sporomorph associations, marine microfossils and mammal zones, marine and continental stages and the radiometric scale. *Annales Geologiques Des Pays Helleniques* (hors ser.) 1, 61-70.

Benda, L., Meulenkamp, J.E. 1990. Biostratigraphic correlations in the Eastern Mediterranean Neogene.9.Sporomorph associations and event stratigraphy of the Eastern Mediterranean. *Newsletters on Stratigraphy,* 23, 1-10.

Besang, C., Eckhardt, F.J., Harre, W., Kreuzer, H., Müller, P. 1977. Radiometrische altersbestimmungen an neogenen eruptivgesteinen der Türkei. *Geologisches Jahrbuch,* B25, 3-36.

Bozkurt, E. 2000. Timing of extension on the Büyük Menderes Graben, western Turkey, and its tectonic implications. Bozkurt, E., Winchester, J.A. & Piper, J.D.A. (eds) Tectonics and Magmatism in Turkey and the Surrounding Area. *Geological Society of London. Special Publication,* 173, 385-403.

Bozkurt, E. 2003. Origin of NNE-trending basins in Western Turkey. *Geodinamica Acta,* 16, 61-81.

Bozkurt, E. 2004. Granitoid rocks of the southern Menderes Massif (southwestern Turkey): field evidence for Tertiary magmatism in an extensional shear zone. *International Journal of Earth Sciences,* 93, 52–71.

Bozkurt, E, Park, R.G. 1993. Menderes Massif: A Cordilleran type metamorphic core complex in western Turkey. EUG VII, Strasbourg, *Terra Abstracts* v.5, p.255.

Bozkurt, E., 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.

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

Bozkurt, E., 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 de France,* 168, 1, 3-14.

Bozkurt, E., Park, R.G. 1997*b*. Microstructures of deformed grains in the augen gneiss of southern Menderes massif, western Turkey and their tectonic significance. *International Journal of Earth Sciences,* 86, 109-119.

Bozkurt, E., Park, R.G. 1999. The structure of the Paleozoic schists in the southern Menderes Massif, western Turkey: a new aproach to the origin of the Main Menderes metamorphism and its relation to the Lycian Nappes. *Geodinamica Acta,* 12, 25-42.

Bozkurt, E., Sözbilir, H. 2004. Tectonic Evolution of the Gediz Graben: Field Evidence for an Episodic, Two-Stage Extension in Western Turkey. Geological Magazine, 141, 63–79.

Bozkurt, E., Rojay, B. 2005. Episodic, two-stage Neogene extension and short-term intervening compression in Western Turkey: field evidence from the Kiraz Basin and Bozdağ Horst. *Geodinamica Acta,* 18 299-316.

Buck, W.R. 1988. Flexural rotation of normal faults.*Tectonics,* 7, 959-973.

Buscher, J.T., Hampel, A., Hetzel, R., Dunkl, I., Glotzbach, C., Struffert, A., Akal, C,. Ratz, M. 2013. Quantifying rates of detachment faulting and erosion in the central Menderes massif (western Turkey) by thermochronology and cosmogenic 10Be. *Journal of Geological Society London,* 170, 669-683.

Candan, O., Dora, Ö.O. 1998. Menderes Masifi'nde granulit, eklojit ve mavişist kalıntıları: Pan-Afrikan ve Tersiyer metamorfik evrimine bir yaklaşım. *Türkiye Jeoloji Bülteni,* 41/1, 1-35.

Candan, O., Koralay, E., Akal, C., Kaya, O., Oberhansli, R., Dora, O.Ö., Konak, N., Chen, F. 2011. Supra-Pan-African unconformity between core and cover series of the Menderes Massif / Turkey and its geological implications. *Precambrian Research,* 184, 1-23.

Catlos, E. J., Çemen, I., Işık, V., Seyitoğlu, G. 2002. In-situ timing constraints from the Menderes Massif, western Turkey. Geological Society of America, Abstracts with Programs, 180.

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

Catlos, E.J., Baker, C., Sorensen, S.S., Çemen, İ., Hançer, M. 2010. Geochemistry, geochronology and cathodoluminescence imagery of the Salihli and Turgutlu granites (Central Menderes Massif, western Turkey): Implications for Aegean tectonics. *Tectonophysics,* 488, 110-130.

Cohen, H.A., Dart, C.J., Akyüz, H.S., Barka, A.A. 1995. Syn-rift sedimentation and structural development of Gediz and Büyük Menderes graben, western Turkey. *Journal of the Geological Society London,* 152, 629-638.

Collettini, C. 2011. The mechanical paradox of low-angle normal faults: Current understanding and open questions. Tectonophysics, 510, 253-268.

Collins, A.S., Robertson, A.H.F. 1998. Processes of Late Cretaceous to Late Miocene episodic thrust-sheet translation in the Lycian Taurides, SW Turkey. *Journal of the Geological Society London,* 155, 759-772.

Collins, A.S., Robertson, A.H.F. 2003.Kinematic evidence for Late Mesozoic - Miocene emplacement of the Lycian allochthon over western Anatolide belt, SW Turkey.*Geological Journal,* 38, 295-310.

Coney, P.J. 1980. Cordilleran metamorphic core complexes; an overview. In: Crittenden, M.D., Coney, P.J. and Davis, G.H. (Ed.). Cordilleran Metamorphic Core Complexes. *Geological Society of America. Memoir,* 153, 7-31.

Çağlar, I., Duvarcı, E. 2001. Geoelectric structure of inland area of the Gökova rift, southwest Anatolia and its tectonic implications. *Journal of Geodynamics*, 31, 33-48.

Çağlayan, M.A., Öztürk, E.M., Öztürk, Z., Sav, H., Akat, U. 1980. Menderes masifi güneyine ait bulgular ve yapısal yorum. *Jeoloji Mühendisliği,* 10, 9-19.

Çiftçi, G., Pamukçu, O., Çoruh, C., Çopur, S., Sözbilir, H. 2011. Shallow and deep structure of a supradetachment basin based on geological, conventional deep seismic reflection sections and gravity data in the Büyük Menderes graben, western Anatolia. *Surveys in Geophysics,* 32, 271-290.

Çiftçi, N.B., Bozkurt, E. 2009. Evolution of the Miocene sedimentary fill of the Gediz Graben, SW Turkey. Sedimentary Geology, 216, 49-79.

Çiftçi, N.B., Bozkurt, E. 2010. Structural evolution of the Gediz Graben, SW Turkey.temporal and spatial variation of the graben basin. *Basin Research,* 22, 846-873.

Davis, G.H., Reynolds, S.J. 1996. Structural Geology of Rocks and Regions. John Wiley & Sons Inc., 776s.

Davis, G.A., Lister, G.S. 1988. Detachment faulting in continental extension: Perspectives from the south wetern U.S. Cordillera. Clark, S.P. Processes in continental lithospheric deformation. *Geological Society of America Special Publication,* 218, 133-159.

de Graciansky, P.C., Lorenz, C., Magne, J. 1970. Sur les étapes de la transgression du Miocene inferieur observée dans les fenetres de Göcek (sud-quest de la Turquie). *Bulletin de la Societe Geologique de France,* 12, 557-564.

Demircioğlu, D., Ecevitoğlu, B., Seyitoğlu, G. 2010. Evidence of a rolling hinge mechanism in the seismic records of hydrocarbon-bearing Alaşehir graben, western Turkey. *Petroleum Geoscience,* 16, 155-160.

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

Dewey, J.F. 1988. Extensional collapse of orogens. *Tectonics* 7, 1123-1139.

Dewey, J.F., Şengör, A.M.C. 1979. Aegean and surrounding regions: complex multiple and continuum tectonics in a convergent zone. *Bulletin of Geological Society of America,* 90, 84-92.

Dora, Ö.O., Kun, N., Candan, O. 1990. Metamorphic history and geotectonic evolution of the Menderes massif. I.E.S.C.A. Proceedings, 2, 102-115.

Dora, Ö.O., Candan, O., Dürr, S., Oberhansli, R. 1995. New evidence on the geotectonic evolution of the Menderes massif. I.E.S.C.A. Proceedings, 1, 53-72.

Dürr, S. 1975. Über Alter und geotektonische Stellung des Menderes-Kristallins SW-Anatolien und seine Aequivalente in der mittleren Aegaeis. Doctoral Dissertation, University Marburg Lahn.

Dürr, St., Allherr, R., Keller, J., Okrusch, M., Seidel, E. 1978. The median Aegean crystalline belt: Stratigraphy, structure, metamorphism, magmatism. In closs, H. Poeder, D.H. and Schmidt, K. (Eds), Alps, Apennines, Helienides., 455-477, Shweizerbart, Stuttgart.

Ediger, V., Batı, Z., Yazman, M. 1996. Paleopalynology of possible hydrocarbon source rocks of the Alaşehir - Turgutlu area in the Gediz graben (western Anatolia). *Turkish Association of Petroleum Geologists Bulletin,* 8, 94-112.

Emre, Ö., Duman, T.Y., Özalp, S. 2011. Kütahya (NJ 35-4) Paftası, 1:250.000 ölçekli Türkiye diri fay haritası serisi no: 10. *Maden Tetkik ve Arama Genel Müdürlüğü*, Ankara-Türkiye.

Emre, T. 1992. Gediz grabeni'nin (Salihli-Alaşehir arası) jeolojisi. 45. Türkiye Jeoloji Kurultayı Bildiri Özleri, s.60.

Emre, T. 1996. Gediz grabeni'nin jeolojisi ve tektoniği. *Turkish Journal of Earth Sciences*, 5, 171-185.

Emre, T., Sözbilir, H. 1997.Field evidence for metamorphic core complex, detachment faulting and accomodation faults in the Gediz and Büyük Menderes grabens, western Anatolia. İç: IESCA Proceedings (Ed: Ö. Pişkin, M. Ergün, M.Y. Savaşçın, G. Tarcan), s. 74-93.

Emre, T., Sözbilir, H., Gökçen, N. 2006. Neogene-Quaternary stratigraphy of Kiraz – Beydağ vicinity, Küçük Menderes graben, west Anatolia.

Maden Tetkik ve Arama Genel Müdürlüğü Dergisi, 132, 1-32.

Emre, T., Sözbilir, H. 2007. Tectonic evolution of the Kiraz basin, Küçük Menderes graben: Evidence for compression/uplift-related basin formation overprinted by extensional tectonics in west Anatolia. *Turkish Journal of Earth Sciences,* 16, 441-470.

Ercan, T., Dinçel, A., Metin, S., Türkecan, A., Günay, A. 1978. Uşak yöresindeki Neojen havzalarının jeolojisi. Bulletin of the Geological Society of Turkey, 21, 97–106.

Ercan, T., Türkecan, A., Dinçel, A., Günay, E. 1983. Kula - Selendi (Manisa) dolaylarının jeolojisi. Jeoloji Mühendisliği, 17, 3-28.

Ercan, T., Satır, M., Sevin, D., Türkecan, A. 1996. Batı Anadolu'daki Tersiyer ve Kuvaterner yaşlı volkanik kayaçlarda yeni yapılan radyometrik yaş ölçümlerinin yorumu. *Maden Tetkik ve Arama Genel Müdürlüğü Dergisi*, 119, 103-112.

Erdoğan, B., 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.

Ersoy, Y., Helvacı, C. 2007. Stratigraphy and geochemical features of the Early Miocene bimodal (ultrapotassic and calc-alkaline) volcanic activity within the NE-trending Selendi basin, western Anatolia, Turkey.*Turkish Journal of Earth Sciences,* 16, 117-139.

Ersoy, Y., Helvacı, C., Sözbilir, H., Erkül, F., Bozkurt, E. 2008. A geochemical approach to Neogene-Quaternary volcanic activity of western Anatolia: An example of episodic bimodal volcanism within the Selendi basin, Turkey. *Chemical Geology,* 255, 265-282.

Ersoy, Y.E., Helvacı, C., Sözbilir, H. 2010. Tectono-stratigraphic evolution of the NE-SW trending superimposed Selendi basin: Implications for late Cenozoic crustal extension in western Anatolia, Turkey. *Tectonophysics,* 488, 210-232.

Ersoy, Y.E., Helvacı, C., Palmer, M.R. 2011. Stratigraphic, structural and geochemical features of the NE-SW trending Neogene volcano-sedimentary basins in western Anatolia: Implications for associations of supra-detachment and transtensional strike-slip basin formation in extensional tectonic setting. *Journal of Asian Earth Sciences*, 41, 159-183.

Ersoy, Y.E., Çemen, İ., Helvacı, C., Billor, Z. 2014. Tectono-stratigraphy of the Neogene basins in western Turkey: Implications for tectonic evolution of the Aegean extended region. *Tectonophysics*, 635, 33-58.

Eyidoğan, H., Jackson, J. 1985. A seismological study of normal faulting in the Demirci, Alaşehir and Gediz earthquakes of 1967-70 in western Turkey: implications for the nature and geometry of deformation in the continental crust. *Geophysical Journal of the Royal Astronomical Society London,* 81, 569-607.

Faure, M., Bonneau, M., Pons, J. 1991.Ductile deformation and syntectonic granite emplacement during the late Miocene extension of the Aegea (Greece). *Bulletin de la Societe geologique de France,* 162, 3-11.

Fletcher, J. M., Bartley, J. M., Martin, M. W., Glazner, A. F., Walker, J. D. 1995. Largemagnitude continental extension: An example from the central Mojave metamorphic core complex. *Geological Society of America Bulletin,* 107/ 12, 1468-1483.

Fossen, H. 2010. Structural Geology. *Cambridge University Press*. 463p.

Friedmann, S.J., Burbank, D.W. 1995. Rift basins and supradetachment basins: intracontinental extensional end-members. *Basin Research,* 7, 109-127.

Gans, P.B., Mahood, G.A., Schermer, E.R. 1989. Syn-extensional magmatism in the Basin and Range province; a case study from the eastern Great Basin.*Geological Society of America.Special Publication* 233, 60 p.

Gawthorpe, R.L., Hurst, J.M. 1993. Transfer zones in extensional basins: their structural style and influence on drainage development and stratigraphy. *Journal of the Geological Society London* 150, 1137-1152.

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

Gessner, K., Ring, U., Johnson, C., Hetzel R., Passchier, C. W., Güngör, T. 2001*b*. An active bivergent rolling-hinge detachment system: Central Menderes metamorphic core complex in western Turkey. *Geology,* 29, 611-614.

Gesner, K., Collins, A. S., Ring, U., 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.

Gessner, K., Gallardo, L.A., Markwitz, V., Ring, U., Thomson, S.N. 2013. What caused the denudation of the Menderes Massif: Review of crustal evolution, lithosphere structure, and dynamic topography in southwest Turkey. Gondwana Research, 24, 243-274.

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

Güleç, N. 1991. Crust-mantle interaction in western Turkey: Implications from Sr and Nd isotope geochemistry of Tertiary and Quaternary volcanics. Geological Magazine 128, 417-435.

Gürer, F., Yılmaz, Y. 2002. Geology of the Ören and surrounding areas, SW Anatolia. *Turkish Journal of Earth Sciences,* 11, 1-13.

Gürer, Ö.F, Sarıca-Filoreau, N., Özburan, M., Sangu, E., Doğan, B. 2009. Progressive development of the Büyük Menderes Graben based on new data, western Turkey. *Geological Magazine,* 146, 652-673.

Hetzel, R., Passchier, C.W., Ring, U., Dora, Ö.O. 1995*a*. Bivergent extension in orogenic belts: The Menderes massif (southwestern Turkey). Geology, 23, 455-458.

Hetzel, R., Ring, U., Akal, C., Troesch, M. 1995*b*. Miocene NNE-directed extensional unroofing in the Menderes massif, southwestern Turkey. *Journal of Geological Society London*, 152, 639-654.

Hetzel, R., 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.

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

Hetzel, R., Zwigmann, H., Mulch, A., Gessner, K., Akal, C., Hampel, A., Güngör, T., Petschick, R., Mikes, T., Wedin, F. 2013. Spatiotemporal evolution of brittle normal faulting and fluid infiltration in detachment fault systems: a case study from Menderes massif, western Turkey. *Tectonics,* 32, 1-13.

Innocenti, F., Agostini, S., Di Vincenzo, G., Doglioni, C., Manetti, P., Savaşçın, M.Y., Tonarini, S. 2005. Neogene and Quaternary volcanism in Western Anatolia: Magma sources and geodynamic evolution. *Marine Geology,* 221, 397-421.

Işık, V., Tekeli, O., Çemen, İ. 1997. Mylonitic fabric development along a detachment surface in northern Menderes massif, western Anatolia, Turkey. Geol Soc Am., Annual Meeting, Abstracts with programs 29, A-220.

Işık, V., Tekeli, O. 1998. Structure of Lower Plate Rocks in Metamorphic Core Complex; Northern Menderes massif, Western Turkey. 3th Int. Turkish Geology Sym., METU, Ankara, Turkey, p. 268.

Işık, V., Tekeli, O. 1999. Microstructure of deformed grains in the granitic mylonites of lower plate: northern Menderes massif, western Turkey. EUG 10, J. Conference Abstracts, 4 (1), 730.

Işık, V., Tekeli, O. 2001. Late orogenic crustal extension in the northern Menderes massif (western Turkey): Evidences for metamorphic core complex formation. *International Journal of Earth Sciences,* 89, 757-765.

Işık, V., Seyitoğlu, G., Çemen, İ. 2003*a*. Extensional structures of the Menderes core complex, western Turkey. GSA, Annual Meeting Abstracts with Programs, Seattle, USA, v.35, p.27-28.

Işık, V., Seyitoğlu, G., Çemen, İ. 2003*b*. Ductile-brittle transition along the Alasehir shear zone and its structural relationship with the Simav detachment, Menderes massif, western Turkey. *Tectonophysics,* 374, 1-18.

Işık, V., Gürsu, S., Göncüoğlu, C., Seyitoğlu, G. 2004*a*. Deformational and geochemical features of syn-tectonic Koyunoba and Egrigöz granitoids, western Turkey. Chatzipetros, A.A., Pavlides, S.B. (Ed.), 5th International Symposium on Eastern Mediterranean Geology, Thessaloniki, Greece, 3, pp. 1143–1146.

Işık, V., Tekeli, O., Seyitoğlu, G. 2004*b*. The $^{40}Ar/^{39}Ar$ age of extensional ductile deformation and granitoid intrusions in the northern Menderes core complex: Implications for the initiation of extensional tectonics in western Turkey. *Journal of Asian Earth Science,* 23, 555-566.

Işık, V., Seyitoğlu, G. 2006. Menderes metamorfik çekirdek kompleksinde sıyrılma fayları ve ilişkili makaslama zonları, Batı Anadolu. 59. Türkiye Jeoloji Kurultayı Bildiri Özleri, s. 21-24

Işık, V., Seyitoğlu, G. 2007. Menderes Masifinin Geç Senozoyikte Asimetrik ve Simetrik Çekirdek Kompleksi Olarak Yüzeyleme ve Parçalanma Tarihçesi. Menderes Masifi Kolokyumu, Genişletilmiş Bildiri Özleri Kitabı, 88-97.

Işık, V., Seyitoğlu, G. 2010. Cenozoic Exhumation and Sedimentary Basin formation in the Menderes Massif, Western Turkey. Tectonic Crossroads: Evolving Orogens of Eurasia-Africa-Arabia. Abstracts with Programs (36-1), p.70.

İnci, U. 1984. Demirci ve Burhaniye bitümlü şeyllerinin stratigrafisi ve organik özellikleri. Türkiye Jeoloji Kongresi Bülteni, 5, 27-40.

İztan, H., Yazman, M. 1990. Geology and hydrocarbon potential of the Alaşehir (Manisa) area, western Turkey.Proceedings to International Earth Sciences Congress on Aegean regions, İzmir, 327-338.

Jackson, J.A., McKenzie, D. 1988. The relationship between plate motions and seismic moment tensors and rates of active deformation in the Mediterranean and Middle East. *Geophysical Journal,* 93, 45-73.

Janecke, U.S., Vandenburg, J.C., Blankenau, J.J. 1998.Geometry, mechanisms and significance of extensional folds from examples in the Rocky Mountain Basin and Range province, USA. *Journal of Structural Geology,* 20, 841-856.

Jolivet, L., Faccenna, C., D'agostina, N., Fournier, M., Worrall, D. 1999. The kinematics of back-arc basins, examples from the Tyrrhenian, Aegean and Japan Seas.Mac Niocaill, C., Ryan, P.D. (Ed.).

Continental Tectonics. *Journal of the Geological Society London Special Publication,* 164, 21-53.

Jolivet, L., Faccenna, C. 2000. Mediterranean extension and the Africa–Eurasia collision.*Tectonics,* 19, 1095-1106.

Jolivet, J., Faccenna, C., Huet, B., Labrousse, L., Le Pourhiet, L., Lacombe, O., Lecomte, E., Burov, E., Denèle, Y., Brun, JP., Philippon, M., Paul, P., Salaün, G., Karabulut, H., Piromallo, C., Monié, P., Gueydan, F., Okay, A.I., Oberhänsli, R., Pourteau, A., Augier R., Gadenne, L., Driussi, O. 2013. Aegean tectonics: Strain localisation, slab tearing and trench retreat. *Tectonophysics,* 597–598, 1-33.

Karaoğlu, Ö., Helvacı, C. 2012. Structural evolution of the Uşak-Güre supradetachment basin during Miocene extensional denudation in western Turkey. Journal of the Geological Society London, 169, 627-642.

Kaya, O. 1981. Miocene reference section for the coastal parts of west Anatolia.*Newsletters on Stratigraphy,* 10, 164-191.

Kaya, S., Esat, K., Ecevitoğlu, B., Işık, V., Kaypak, B., Uyar Aldaş, G., Can, A.Z., Baksı, E.E., Akkaya, İ. Seyitoğlu, G. 2014. Afyon-Akşehir grabeni batı kenarının tektonik özellikleri üzerine jeolojik ve jeofizik gözlemler: İki evreli genişleme modeli hakkındaki tartışmalara bir katkı. *Yerbilimleri,* 35, 1-16.

Kaymakçı, N. 2006. Kinematic development and paleostress analysis of the Denizli Basin (Western Turkey): implications of spatial varition of relative paleostress magnitudes and orientations. *Journal of Asian Earth Sciences,* 27, 207-222.

Kaypak, B., Gökkaya, G. 2012. 3-D imaging of the upper crust beneath the Denizli geothermal region by local earthquake tomography, western Turkey. *Journal of Volcanology and Geothermal Research,* 211, 47-60.

Ketin, İ., Abdüsselamoğlu, Ş. 1969. 23 Mart 1969 Demirci ve 28 Mart 1969 Alaşehir-Sarıgöl depremleri hakkında makro-sismik gözlemler. *Maden Mecmuası,* 4, 21-26.

Koçyiğit, A., Yusufoğlu, H., Bozkurt, E. 1999. Evidence from the Gediz graben for episodic two-stage extension in western Turkey. *Journal of the Geological Society London*, 156, 605-616.

Koçyiğit, A., Ünay, E., Saraç, G. 2000. Episodic graben formation and extensional neotectonic regime in west central Anatolia and the Isparta Angle: a case study in the Akşehir-Afyon graben, Turkey. İç: Tectonics and Magmatism in Turkey and the Surrounding Area (Ed: E. Bozkurt, J.A. Winchester, J.D.A. Piper), Geological Society London, Special Publications 173, s. 405-421.

Konak, N. 1979. Simav grabeni ve getirdiği kentleşme sorunları. Türkiye Jeoloji Mühendisliği 1. Bilimsel ve Teknik Kongresi Bildirileri, 157-164.

Konak, N. 1982. Simav dolayının jeolojisi ve metamorfik kayaçlarının evrimi. *İ.Ü.M.F. Yerbilimleri Dergisi*, 3, 313-337.

Konak, N., Akdeniz, N., Öztürk, E.M. 1987. Geology of the south of Menderes massif, correlation of Variscan and pre-Variscan events of the Alpine Mediterrenean Mountain Belt (Guide book for the field exursion along western Anatolia, Turkey) IFCP Project No. 5, 42-53.

Koralay, E., Dora, O.Ö., Chen, F., Satır, M., Candan, O. 2004. Geochemistry and geochronology of orthogneisses in the Derbent (Alaşehir) area, Eastern part of the Ödemiş-Kiraz submassif, Menderes Massif: Pan-African magmatic activity. *Turkish Journal of Earth Sciences*, 13, 37-61.

Kurt, H., Demirbağ, E., Kuşçu, İ. 1999. Investigation of submarine active tectonism in the Gulf of Gökova, southwest Anatolia-southeast Aegean sea, by multi-channel seismic reflection data. *Tectonophysics,* 305, 477-496.

Kurt, F.S., Işık, V., Seyitoğlu, G. 2010. Alternative Cenozoic exhumation history of the Kazdağ core complex. Tectonic Cross Roads: Evolving Orogens of Eurasia-Africa-Arabia. Abstracts with Programs (36-3), METU Ankara, s.70.

Le Pichon, X., Angelier, J. 1979. The Hellenic arc and trench system: a key to the neotectonic evolution of the eastern Mediterranean area. *Tectonophysics,* 69, 1-42.

Le Pichon, X., Angelier, J. 1981. The Aegean Sea. *Philosophical Transactions of the Royal Society London,* A300, 357-372.

Lips, A.L.W., Cassard, D., Sözbilir, H., Yılmaz, H. 2001. Multistage exhumation of the Menderes Massif, western Anatolia (Turkey). *International Journal of Earth Sciences,* 89, 781-792.

Lister, G.S., Banga, G., Feensta, A. 1984. Metamorphic core complexes of Cordilleran type in the Cyclades, Aegean Sea, Greece. *Geology*, 12, 221-225.

Lister, G.S., Davis, G.A. 1989. The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the northern Colorado River region, U.S.A. *Journal of Structural Geology,* 11, 65-94.

Lister, G.S., Baldwin, S.L. 1992. Plutonism and the origin of metamorphic core complexes. *Geology*, 21, 607-610.

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

Malavieille, J. 1993. Late orogenic extension in mountain belts: insights from the Basin and Range and the Late Paleozoic Variscan belt. *Tectonics,* 12, 1115-1130.

Manning, A.H., Bartley, J.M. 1994. Postmylonitic deformation in the Raft River metamorphic core complex, northwestern Utah: Evidence of a rolling hinge. *Tectonics,* 13, 596-612.

McClay, K.R. 1989. Physical models of structural styles during extension. İç: Extensional tectonics and stratigraphy of the north Atlantic margins (Ed: A.J. Tankard, H.R. Balkwill). AAPG Memoir 46, s. 95-110.

McClay, K.R. 1990. Deformation mechanics in analogue models of extensional fault systems. İç: Deformation Mechanisms, Rheology and Tectonics (Ed: R.J. Knipe, E.H. Rutter). Geological Society Special Publication 54, s. 445-453.

McKenzie, D.P. 1970. Plate tectonics of the Mediterranean region. *Nature,* 226, 239-243.

McKenzie, D.P. 1972. Active tectonics of the Mediterranean region. *Geophysical Journal of Royal Astronomical Society,* 30, 109-185.

McKenzie, D. 1978. Active tectonics of the Alpine - Himalayan belt: The Aegean sea and surrounding regions. *Geophysical Journal of Royal Astronomical Society,* 55, 217-254.

Mercier J.L. 1981. Extensional-compressional tectonics associated with the Aegean arc: comparison with the Andean Cordillera of south Peru-north Bolivia. *Phil. Trans. Royal Society London,* A 300, 337-355.

Meulenkamp, J.E., Wortel, M.J.R., van Wamel, W.A., Spakman, W., Hoogerduyn Strating, E. 1988. On the Hellenic subduction zone and the geodynamic evolution of Crete since the late middle Miocene. *Tectonophysics,* 146, 203-215.

Nebert, K. 1961. Gördes bölgesindeki Neojen volkanizması hakkında bilgiler. *Maden Tetkik ve Arama Genel Müdürlüğü Dergisi*, 57, 50-54.

Oberhansli, R., Partzsch, J., Candan, O., Çetinkaplan, M. 2001. First occurrence of Fe-Mg-carpholite documenting a high-pressure metamorphism in metasediments of the Lycian Nappes, SW Turkey. *International Journal of Earth Sciences,* 89, 867–873.

Okay, A.I. 1989. Geology of the Menderes Massif and the Lycian Nappes south of Denizli, western Taurides. *Miner. Res. Explor. Bull.*, 109, 37–51.

Öner, Z., Dilek, Y. 2011. Supradetachment basin evolution during continental extension: The Aegean province of western Anatolia, Turkey. *Geological Society of America Bulletin,* 123, 2115-2141.

Özerdem, C., Çemen, İ., Işık, V. 2002. The conglomerate member of the Gökçeören formation, Ören basin, western Turkey: its age, sedimentology, and tectonic significance. Geological Society of America, Abstracts with Programs, 250.

Pamir, H.N., Erentöz, C. 1974. 1/500.000 ölçekli Türkiye Jeoloji Haritası Denizli Paftası İzahnamesi. *Maden Tetkik ve Arama Genel Müdürlüğü Yayını*, 83 s.

Phillipson, A. 1915. Reisen und Forschungen in Westlichen Kleinasien. Pet. Mitt. Erg. M., 167, 173, 177, 180 Cotha.

Purvis, M., Robertson, A.H.F. 2005. Sedimentation of the Neogene - Recent Alaşehir (Gediz) continental graben system used to test alternative tectonic models for western (Aegean) Turkey. *Sedimentary Geology,* 173, 373-408.

Prante, M.R., Evans, J.P., Janecke, S.U., Steely, A. 2014. Evidence for paleoseismic slip on a continental low-angle normal fault: Tectonic pseudotachylyte from the west Salton detachment fault, CA, USA. *Earth and Planetary Science Letters*, 387, 170-183.

Ramsay, J.G., Huber, M.I. 1987. The techniques of modern structural geology (Volume 2: Folds and Fractures). Academic Press London, 700s.

Ricou, L.E., Argyriadis, I., Marcoux, J. 1975. L'Axe calceire du Taurus un alignement de fenetres arabo-africans sous des nappes radiolaritiques, ophiolitiques et metamorphiques. *Bulletin de la Societe Geologique de France*, 17, 1024-1044.

Reynolds, S.J. 1985. Geology of the South Mountains, central Arizona. *Arizona Bureau of Geology and Mineral Tecnollogy, Geological Survey Bulletin* 195, 61 p.

Reynolds, S.J., Rehring, W.A. 1980. Mid-Tertiary plutonism and mylonitization, South Mountains, central Arizona. *Geological Society of America. Memoir* 153, 159-175.

Rimmele, G., Jolivet, L., Oberhansli, R., Goffe, B., 2003. Deformation history of the high-pressure Lycian Nappes and implications for tectonic evolution of SW Turkey. *Tectonics,* 22, 1007-

Ring, U., Johnson, C., Hetzel, R., Gessner, K. 2003. Tectonic denudation of a Late Cretaceous-Tertiary collisional belt-regionally symmetric cooling patterns and their relation to extensional faults in the Anatolide belt of western Turkey. *Geological Magazine*, 140, 1-21.

Ring, U., Collins, A.S. 2005. U–Pb SIMS dating of synkinematic granites: timing of core-complex formation in the northern Anatolide belt of western Turkey. *Journal of the Geological Society London,* 162, 289–298.

Saraç, G. 2003. Türkiye Omurgalı Fosil Yatakları. *Maden Tetkik ve Arama Genel Müdürlüğü Rapor* No: 10609, 208s. Ankara (unpublished).

Sarı, C., Şalk, M. 2006. Sediment thicknesses of the western Anatolia graben structures determined by 2D and 3D analysis using gravity data. *Journal of Asian Earth Sciences,* 26, 39–48.

Sarıca, N. 2000. The Plio-Pleistocene age of Büyük Menderes and Gediz grabens and their tectonic significance on N-S extensional tectonics in west Anatolia: mammalian evidence from the continental deposits. *Geological Journal,* 35, 1-24.

Savaşçın, M.Y. 1991. Magmatic activities of Cenozoic compressional and extensional tectonic regimes in western Anatolia. Proceedings of International

Earth Sciences Congress on Aegean Regions, İzmir Turkey, s. 420-434.

Savaşçın, M.Y., Güleç, N. 1990. Neogene volcanism of western Anatolia-Field excursion B3.International Earth Sciences Congress on Aegean Regions, İzmir Turkey.

Seyitoğlu, G. 1992. Late Cenozoic Crustal Extension, Basin formation and Volcanism in West Turkey. PhD Thesis, University of Leicester, England, UK.

Seyitoğlu, G. 1997*a*. The Simav graben: An example of young E-W trending structures in the Late Cenozoic extensional system of western Turkey. *Turkish Journal of Earth Sciences,* 6, 135-141.

Seyitoğlu, G. 1997*b*. Late Cenozoic tectono-sedimentary development of Selendi and Uşak-Güre basins: a contribution to the discussion on the development of east-west and north- trending basins in western Turkey. *Geological Magazine,* 134, 163-175.

Seyitoğlu, G. 1999. Discussion on evidence from the Gediz Graben for episodic two-stage extension in western Turkey. *Journal of the Geological Society London* 156, 1240.

Seyitoğlu, G., Scott, B.C. 1991. Late Cenozoic crustal extension and basin formation in west Turkey. *Geological Magazine,* 128, 155-166.

Seyitoğlu, G., Scott, B.C. 1992*a*. The age of the Büyük Menderes graben (west Turkey) and its tectonic implications.*Geological Magazine,* 129, 239-242.

Seyitoğlu, G., Scott, B.C. 1992*b*. Late Cenozoic volcanic evolution of the NE Aegean region. *Journal of Volcanology and Geothermal Research,* 54, 157-176.

Seyitoğlu, G., Scott, B.C., Rundle, C.C. 1992. Timing of Cenozoic extensional tectonics in west Turkey. *Journal of the Geological Society London,* 149, 533-38.

Seyitoğlu, G., Scott, B.C. 1994. Late Cenozoic basin development in west Turkey. Gördes basin: tectonics and sedimentation. *Geological Magazine,* 131, 631-637.

Seyitoğlu, G., Benda, L., Scott, B.C. 1994. Neogene palynological and isotopic age data from Gördes basin, West Turkey. *Newsletters on Stratigraphy,* 31, 133-142.

Seyitoğlu, G., Scott, B.C. 1996*a*. Age of Alaşehir graben (west Turkey) and its tectonic implications. *Geological Journal,* 31, 1-11.

Seyitoğlu, G., Scott, B.C. 1996*b*. The cause of N-S extensional tectonics in western Turkey: Tectonic escape vs. Back-arc spreading vs. Orogenic collapse. *Journal of Geodynamics,* 22, 145 - 153.

Seyitoğlu, G., Anderson, D., Nowell, G., Scott, B.C. 1997. The evolution from Miocene potassic to Quaternary sodic magmatism in western Turkey: implications for enrichment processes in the lithospheric mantle. *Journal of Volcanology and Geothermal Research,* 76, 127-147.

Seyitoğlu, G., Benda, L. 1998. Neogene palynological and isotopic age data from Selendi and Uşak-Güre basins, western Turkey: a contribution to the upper limit of Eskihisar sporomorph association. *Newsletters on Stratigraphy*, 36, 105-115.

Seyitoğlu, G., Şen, Ş. 1998. The contribution of first magnetostratigraphical data from E-W trending graben fill to the style of late Cenozoic extensional tectonics in western Turkey. Third International Turkish Geology Symposium, Abstracts 188.

Seyitoğlu, G., Şen, Ş. 1999. Discussion on "Akgün F., Akyol E. 1999. Palynostratigraphy of the coal bearing Neogene deposits in Büyük Menderes graben, Western Anatolia. Geobios, 32(3), 367-383".*Geobios,* 32, 934.

Seyitoğlu, G., Çemen, İ., Tekeli, O. 2000. Extensional folding in the Alaşehir (Gediz) graben, western Turkey. *Journal of the Geological Society London,* 157, 1097 -1100.

Seyitoğlu, G., Tekeli, O., Çemen, İ., Şen, Ş., Işık, V. 2002. The role of the flexural rotation / rolling hinge model in the tectonic evolution of the Alaşehir graben, western Turkey.*Geological Magazine,* 139, 15-26.

Seyitoğlu, G., Işık, V., Çemen, İ. 2004. Complete Tertiary exhumation history of the Menderes massif, western Turkey: an alternative working hypothesis. *Terra Nova*, 16, 358-364.

Seyitoğlu, G., Alçiçek, M.C., Işık, V., Alçiçek, H., Mayda, S., Varol, B., Yılmaz, I., Esat, K. 2009. The stratigraphical position of Kemiklitepe fossil locality (Eşme, Uşak) revised: Implications for the Late Cenozoic sedimentary basin development and extensional tectonics in western Turkey. *Neues Jahrbuch für Geologie und Palaeontologie*, 251, 1-15.

Seyitoğlu, G., Işık, V. 2009. Meaning of the Küçük Menderes Graben in the tectonic framework of the central Menderes metamorphic core complex. *Geologica Acta,* 7, 323-331.

Seyitoğlu, G., Işık, V., Esat, K. 2014. A 3D model for the formation of Turtleback surfaces: The Horzum Turtleback of Western Turkey as a case study. *Turkish Journal of Earth Sciences,* 23, 479-494.

Seyitoğlu, G., Cahill, N.D., Işık, V., Esat, K. 2015*a*. Morphotectonics of the Alaşehir graben with a special emphasis on the landscape of the ancient city of Sardis, western Turkey. İç: Landforms and Landscapes of Turkey (Ed: C. Kuzucuoglu, N. Kazancı, A. Çiner), Springer, in press.

Seyitoğlu, G., Ecevitoğlu, B., Kaypak, B., Güney, Y., Tün, M., Esat, K., Avdan, U., Temel, A., Çabuk, A., Telsiz, S., Uyar Aldaş, G. 2015*b*. Determining the main strand of the Eskişehir strike-slip fault zone using subsidiary structures and seismicity: a hypothesis tested by seismic reflection studies. *Turkish Journal of Earth Sciences*, 24, 1-20.

Sickenberg, O., Tobien, H. 1971. New Neogene and Lower Quaternary vertebrate faunas in Turkey. *Newsletters on Stratigraphy,* 1, 51-61.

Spakman, W., Wortel, M.J.R., Vlaar, N.J. 1988. The Hellenic subduction zone; a tomographic image and its geodynamic implications. *Geophysical Research Letters,* 15, 60-63.

Spencer, J.E. 1982. Origin of folding of Tertiary low-angle fault surfaces insoutheastern California and western Arizona, U. S. Geological Survey Professional Paper, Report: P 1375, pp.66.

Spencer, J. E., 1984, Role of denudation in warping and uplift of low-angle normal faults. *Geology* 12, 95-98.

Steininger, F.F., Rögl, F. 1984. Paleogeography and palinspastic reconstruction of the Neogene of the Mediterranean and Paratethys. In: The Geological Evolution of the Eastern Mediterranean (Dixon, J.E. and Robertson, A.H.F., Eds). Geological Society London, Special Publication, 17, 659-668.

Sözbilir, H. 2001. Extensional tectonics and the geometry of related macroscopic structures: field evidence from the Gediz detachment, western Turkey. *Turkish Journal of Earth Sciences,* 10, 51-67.

Sözbilir, H. 2002. Geometry and origin of folding in the Neogene sediments of the Gediz Graben, western Anatolia, Turkey. Geodinamica Acta, 15, 277-288.

Sözbilir, H., Emre, T. 1990. Neogene stratigraphy and structure of the northern rim of the Büyük Menderes graben.Proceedings to International Earth Sciences Congress on Aegean regions, İzmir, 314-322.

Sun, S. 1990. Denizli-Uşak Arasının Jeolojisi ve Linyit Olanakları. *Maden Tetkik ve Arama Genel Müdürlüğü Rapor* No: 9985, 92. Ankara (unpublished).

Şan, O. 1998. Ahmetli (Manisa) güneyinde Menderes masifi ve Tersiyer örtü kayalarının jeolojisi. Yüksek Lisans Tezi, Ankara Üniversitesi.

Şen, Ş, De Bonis, L., Dalfes, N., Geraads, D., Koufos, G. 1994. Les gisements de mammifères du Miocène supérieur de Kemiklitepe, Turquie: 1. Stratigraphie et magnétostratigraphie. İç: Les gisements de mammifères du Miocènesupérieur de Kemiklitepe, Turquie (Ed: Ş. Şen), *Bulletin du Muséum National d'Histoire Naturelle,* 4e sér.,16, sec. C, no. 1, s. 5-17.

Şen, Ş., Seyitoğlu, G. 2009. Magnetostratigraphy of early - middle Miocene deposits from E-W trending Alaşehir and Büyük Menderes grabens in western Turkey, and its tectonic implications. İç*:* Geodynamics of Collision and Collapse at the Africa – Arabia-Eurasia subduction zone. (Ed: D.J.J. van Hinsbergen, M.A. Edwards, R. Govers), *The Geological Society London, Special Publications* 311, 321–342.

Şengör, A.M.C. 1979. The North Anatolian Transform Fault: its age, offset and tectonic significance. *Journal of the Geological Society London,* 13, 269-282.

Şengör, A.M.C. 1980. Türkiye'nin Neotektoniğinin Esasları. Türkiye Jeoloji Kurumu Konferanslar Dizisi:2, Ankara, 40 s.

Şengör, A.M.C., 1982. Ege'nin neotektonik evrimini yöneten etkenler. Batı Anadolu'nun Genç Tektoniği ve Volkanizması Paneli, *Türkiye Jeoloji Kurultayı* 59-71.

Şengör, A.M.C. 1987. Cross-faults and differential stretching of hanging walls in regions of low-angle normal faulting: examples from western Turkey. İç: Continental Extensional Tectonics. (Ed: M.P. Coward,. J.F. Dewey, P.L. Hancock),*The Geological Society of London Special Publications* 28; 575-589.

Şengör, A.M.C., Yılmaz, Y. 1981. Tethyan evolution of Turkey: a plate tectonic approach. *Tectonophysics,* 75, 181– 241.

Şengör, A.M.C., Satır, M., 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.

Şengör, A.M.C., Görür, N., Şaroğlu, F. 1985. Strike-slip deformation, basin formation and sedimentation: Strike-slip faulting and related basin formation in zones of tectonic escape: Turkey as a case study. Biddle, K.T. and Christie-Blick, N. (Ed.).Strike-slip Faulting and Basin formation.*Society of Economic Paleontologists and Mineralogist*. *Special Publication* 37, 277-264.

Şimşek, Ş. 1984. Denizli - Kızıldere - Tekkehamam - Tosunlar - Buldan - Yenice alanının jeolojisi ve jeotermal enerji olanakları. *Maden Tetkik ve Arama Genel Müdürlüğü Rapor* No: 7846, 85s. Ankara (unpublished).

Taner, G. 1974*a*. Denizli bölgesi Neojeni'nin paleontolojik ve stratigrafik etüdü. *Maden Tetkik ve Arama Genel Müdürlüğü Dergisi,* 82, 89-126.

Taner, G. 1974*b*. Denizli bölgesi Neojeni'nin paleontolojik ve stratigrafik etüdü. *Maden Tetkik ve Arama Genel Müdürlüğü Dergisi,* 83, 145-177.

Taner, G. 1975. Denizli bölgesi Neojeni'nin paleontolojik ve stratigrafik etüdü. *Maden Tetkik ve Arama Genel Müdürlüğü Dergisi,* 85, 45-66.

Tchihatcheff, P., 1867-1869. Asie Mineure (Description Physique) Paris.

Thomson, S.N., Stöckhert, B., Brix, M.R. 1998. Thermochronology of the high-pressure metamorphic rocks on Crete, Greece: implications for the speed of tectonic processes. *Geology,* 26, 259–262.

Ünay, E., Göktaş, F., Hakyemez, H.Y., Avşar, M., Şan, Ö. 1995. Dating of the sediments exposed at the northern part of the Büyük Menderes graben (Turkey) on the basis of Arvicolidae (Rodentia, Mammalia). *Geological Bulletin of Turkey,* 38, 75-80.

Ünay, E., De Bruijn, H. 1998. Plio-Pleistocene rodents and lagomorphs from Anatolia.*Mededelingen Nederlands Instituut voor Toegepaste Geowetenschappen TNO* 60, 431-466.

van Hinsbergen, D.J.J. 2010. A key extensional metamorphic complex reviewed and restored: The Menderes massif of western Turkey. Earth-Science Reviews, 102, 60-76.

Wernicke, B., 1981. Low-angle normal faults in the Basin and Range Province: nappe tectonics in an extending in an extending orogen. *Nature,* 291, 645-648.

Wernicke, B.P. 1985. Uniform-sense normal simple shear of the continental lithosphere. *Canadian Journal of Earth Sciences,* 22, 1, 108-125.

Wernicke, B., Axen, G.J. 1988. On the role of isostasy in the evolution of normal fault systems. *Geology,* 16, 848-851.

Westaway, R. 1990. Block rotation in western Turkey. 1.Observational evidence. *Journal of Geophysical Research* 95/B12, 19857-19884.

Westaway, R. 1993. Neogene evolution of the Denizli region of western Turkey. *Journal of Structural Geology,* 15, 37-53.

Westaway, R. 2006. Cenozoic cooling histories in the Menderes Massif, western Turkey, may be caused by erosion and flat subduction, not low-angle normal faulting. *Tectonophysics,* 412, 1-25.

Westaway, R., Pringle, M., Yurtmen, S., Demir, T., Bridland, D., Rowbotham, G., Maddy, D. 2003. Pliocene and Quaternary surface uplift of western Turkey revealed by long-term river terrace sequences. *Current Science,* 84, 1090–1101.

Westaway, R., Pringle, M., Yurtmen, S., Demir, T., Bridgland, D., Rowbotham, G., Maddy, D. 2004. Pliocene and Quaternary regional uplift in western Turkey: the Gediz river terrace staircase and the volcanism at Kula. *Tectonophysics,* 391, 121–169.

Westaway, R., Guillou, H., Yurtmen, S., Demir, T., Scaillet, S., Rowbotham, G. 2005. Constraints on the timing and regional conditions at the start of the present phase of crustal extension in western Turkey, from observations in and around the Denizli region. *Geodinamica Acta,* 18, 209-238.

Westaway, R., Guillou, H., Yurtmen, S., Beck, A., Bridgland, D., Demir, T., Scaillet, S., Rowbotham, G. 2006. Late Cenozoic uplift of western Turkey: Improved dating of the Kula Quaternary volcanic field and numerical modelling of the Gediz River terrace staircase. *Global and Planetary Change,* 51, 131–171

Yağmurlu, F. 1984. Akhisar doğusundaki kömür içeren Miyosen tortullarının stratigrafisi, depolanma ortamları ve tektonik özellikleri. *Türkiye Jeoloji Kongresi Bülteni* 5, 3-20.

Yazman, M. 1997. Western Turkey poses exploration challenges on/offshore. *The Leading Edge,* 16, 897-899.

Yılmaz, M., Gelişli, K. 2003. Stratigraphic – Structural interpretation and hydrocarbon potential of the Alaşehir Graben, Western Turkey. *Petroleum Geoscience,* 9, 277 – 282.

Yılmaz, Y. 1989. An approach to the origin of young volcanic rocks of western Turkey. İç: Tectonic Evolution of the Tethyan Region (Ed: A.M.C. Şengör), Kluwer Academic Publishers, Dordrecht, s.159-189.

Yılmaz, Y. 1990. Comparison of young volcanic associations of western and easternAnatolia formed under a compressional regime: a review. *Journal of Volcanology and Geothermal Research,* 44, 69-77.

Yılmaz, Y., Genç, S. C., Gürer, F., Bozcu, M., Yılmaz, K., Karacık, Z., Altunkaynak, S., Elmas, A., 2000. When did the western Anatolian grabens begin to develop? İç: Tectonics and Magmatism in Turkey and the Surrounding area. (Ed: Bozkurt, E., Winchester, J.A., Piper, J.D.A.). *The Geological Society London Special Publication* 173, s. 353-384.

Yolsal-Çevikbilen, S., Taymaz, T., Helvacı, C. 2014. Earthquake mechanisms in the Gulfs of Gökova, Sığacık, Kuşadası and the Simav region (western Turkey): Neotectonics, seismotectonics and geodynamic implications. *Tectonophysics,* 635, 100-124.

Yin, A. 1991. Mechanism for the formation of the domal and basinal detachment faulting : A three-dimensional analysis. *Journal of Geophysical Research,* 96, 14577-14594.

Yin, A., Dunn, J.F. 1992.Structural and stratigraphic development of the Whipple-Chemehuevi detachment fault system,, southeastern California: Implications for the geometrical evolution of domal and basinal low-angle faults. *Geological Society of America Bulletin,* 104, 659-674.