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STRATIGRAPHIC AND STRUCTURAL FEATURES OF THE YEŞİLBARAK NAPPE IN WESTERN TAURUS RANGE AND ITS COMPARISON WITH THE SIMILAR UNITS IN SE ANATOLIA AND NORTH CYPRUS

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ABSTRACT.- The Yeşilbarak nappe is situated in between the Lycian nappes and Beydağları autochthon in southeast Turkey; it is a continuous unit for long distance and has an intermediate zone character. It is generally made up of turbiditic elastics of Upper Lutetian-Lower Miocene age and comprises of more or less different structural units such as Gömbe and Yavuz units. The latter is observed as overlying the former unit and in many places it is overturned. At the base of the Gömbe unit, the Gebeler formation which is made up of Upper Cretaceous neritic carbonates takes place. The Gömbe unit is represented by two formations: a) Upper Lutetian-Lower Miocene Elmalı formation which comprises sandstone with limestone intercalations, siltstone and claystone, and b) Upper Burdigalian-Lower Langhian Ucarsu formation which is made up of sandstone with limestone bands and lenses, and conglomerates. The second structural unit of the Yeşilbarak nappe, the Yavuz unit is represented by Yavuz formation which comprises limestone-interbedded sandstone, siltstone and claystone of Upper Lutetian-Priabonian age. The Gebeler and Uçarsu formations of Gömbe unit are observed only in limited locations. The Yeşilbarak nappe has undergone intensive deformation related to the southward movement of the Lycian nappes at the end of the Lower Miocene that resulted in a structure of folded, fractured and overthrusted. The unit has been thrusted along a distance of tens of kilometers southward together with the Lycian nappes on the Beydağları autochthone. In southeast Anatolia, in between the Bitlis-Pötürge-Malatya nappes and Southeast Anatolian autochthone, Eocene-Lower Miocene Çüngüş-Hakkari nappe, bearing the features of turbiditic character, is observed in a long distance continuously, with an intermediate zone character. This nappe, as well as in the case of the Yeşilbarak nappe in Western Taurus range comprises of two structural units, the Çüngüş formation and the Hakkari complex. The Eocene-Lower Miocene Çüngüş formation is the lower unit which is made up of sandstone with occasional blocks, siltstone and claystone and has similarities with the Elmalı formation of the Gömbe unit in the west. The Hakkari complex, on the other hand, is the upper structural unit and is composed of two more or less different structural units; the Urşe formation of Eocene-Oligocene age made up of sandstone, claystone, limestone, and the Durankaya formation of Lower-Middle Eocene age made up of sandstone with occasional blocks, shale and conglomerate. These formations that belong to the Hakkari complex may, even if partially, be correlated with the Yavuz formation in the west. These above-mentioned formations of the Çüngüş-Hakkari nappe have undergone intensive deformation related to the southward movement of the Bitlis-Pötürge-Malatya nappes in Miocene and have been thrusted on the Southeast Anatolian autochthone for tens of kilometers. Similar formations with that of the Yeşilbarak nappe and the Middle Eocene-Lower Miocene clastic rocks of the Çüngüş-Hakkari nappe can be observed widespreadly in northern Cyprus. Allochthonous masses have been emplaced on these clastic rocks in Cyprus during Miocene. The Middle Eocene-Lower Miocene elastics have been thrusted by the Ovgos fault southward in the region, however, no large-scale thrusting as observed in Anatolia has not occurred here, in Cyprus. All these data indicate that the results of large-scale nappe tectonics in southern Turkey reveal the occurrence of more or less similar structural styles.

Key words: Yesilbarak Nappe, Stratigraphy, Correlation, Western Taurides, SE Anatolia

INTRODUCTION

In southern Turkey (western Taurus ranges) the tectonic units known as Mende res massif, Beydağları autochthone and Antalya nappes are situated from northwest to southeast (Fig 1). To the northwest of the region, rocks of Precambrian to Eocene age that have undergone low to medium and high metamorphism, Menderes massif, is situated. Between the Menderes massif and the Beydağları autochthone, the Lycian nappes made up of rocks of platform, slope, basin and oceanic crust origin are observed. Southwest and west of the area, Antalya nappes, which are made up of rocks of platform, slope, basin and oceanic crust origin crop out. Beneath the Antalya and the Lycian nappes the Beydağları autochthone is observed, as a southwest-northeast trending uplifted dome comprising platform type sediments. In southwest Turkey, except for the above mentioned tectonic units, allochthonous Yeşilbarak nappe which is represented by Upper Lutetian-Lower Miocene rock units lies between the Beydağları autochthone and the Lycian nappes as an intermediate zone along long distances (Fig. 2).

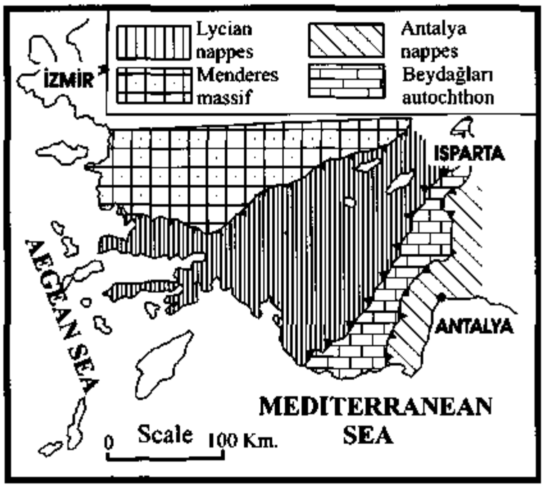

Fig. 1 – Tectonic units of SW Anatolia (West Taurides).

The Yeşilbarak nappe, the subject of this paper, has been studied by many researchers in various locations in Southwest Turkey. While Colin (1955; 1962), Yılmaz (1966), Bassaget (1967), Richard (1967), Maitre (1967), Graciansky (1968; 1972), Akbulut (1977; 1980), Selçuk et al. (1985), Yalçınkaya et al. (1986), Yalçınkaya (1989), Altunsoy (1999) describe the unit as autochthone, some other researchers have proposed that it is allochthone (Gutnic, 1971), Poisson (1977), Gutnic et al. (1979), Erakman et al. (1982), Şenel et al. (1986; 1987; 1989; 1994), Ersoy (1989; 1992), Özkaya (1990; 1991), Collins and Robertson (1997; 1998), Bilgin et al. (1997), Şenel (1997a, *b, c, d, e, f, g, h, i, j, k).* The researchers have proposed different age ranges for the unit, too.

In this paper the stratigraphic and structural features of the Yeşilbarak nappe which is observed under the Miocene nappes in southwestern Turkey (western Taurus ranges) will be discussed, also, they will be compared with the rock units comprised by the allochthonous masses under the Miocene nappes (Bitlis-Poturge-Malatya nappes) in southeast Anatolia.

DESCRIPTION OF THE YEŞİLBARAK NAPPE AND THE PREVIOUS STUDIES

The Yeşilbarak nappe lying in between the Beydağları autochthone and the Lycian nappes in southeast Turkey (western Taurus ranges) was first named by Önalan (1979). The unit was interpreted to be

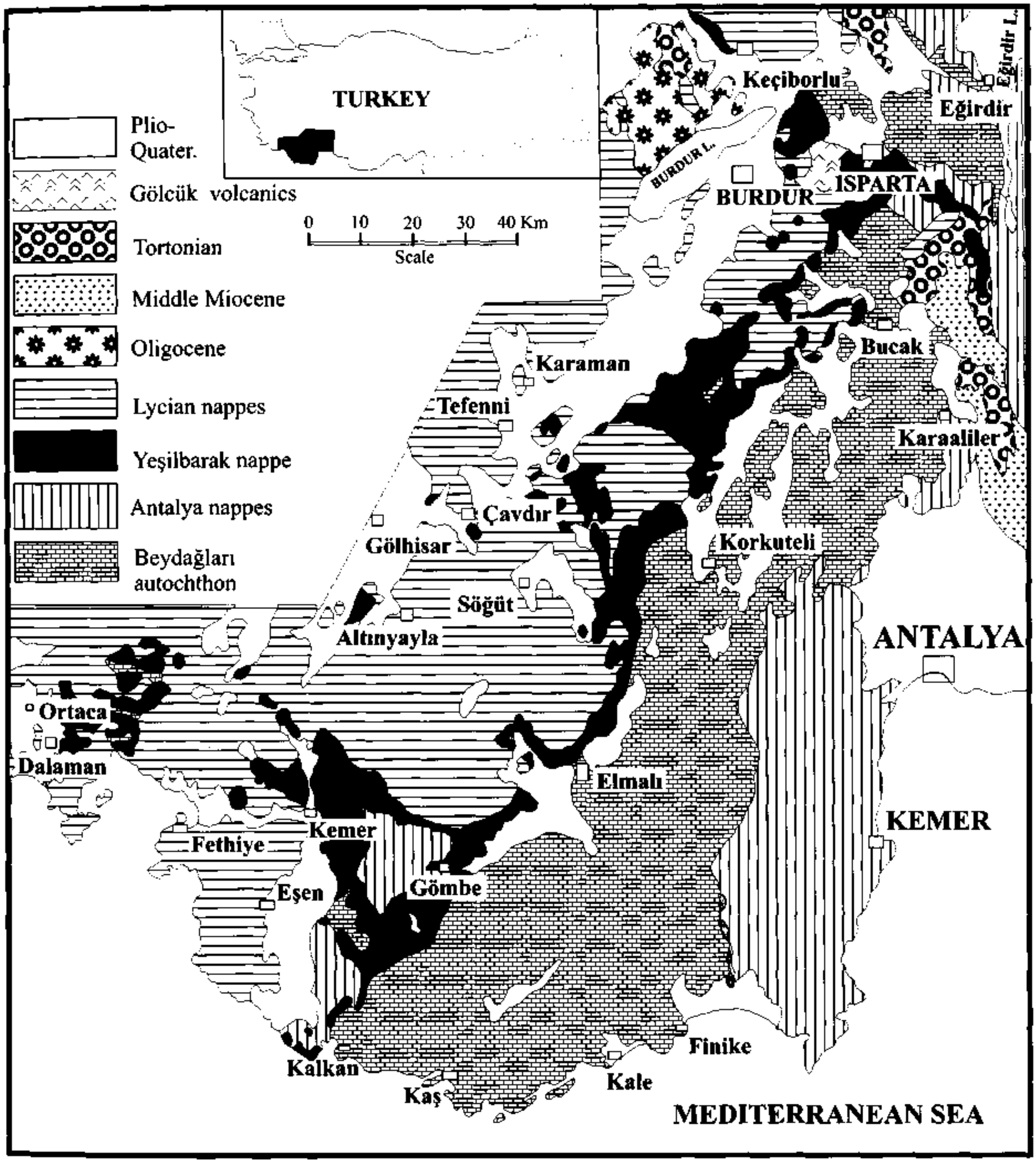

Fig. 2 – The map showing distribution of Yeşilbarak nappe in West Taurides.

autochthonous by Colin (1955, 1962) west of Elmalı as a thrusted, folded, chaotic flysch of Eocene-Miocene age. Graciansky (1968, 1972) proposed that the unit is an olistostrome of Langhian age that belongs to the Beydağları autochthonous west and northwest of Fethiye. Poisson (1977), in Korkuteli area, named the rocks as Yavuz unit as an allochthone of Upper Lutetian-Priabonian age in this region but of Oligocene age in Kemer region. Önalan (1979) who named the unit, stated that the unit is made up of the Elmalı formation (Lutetian), the Deliktaş shale (Oligocene), the Sinekçi formation (Burdigalian) and the Kasaba formation (Helvetian-Tortonian) west of Elmalı. Studying around Dinar-Akdağ, Gutnic (1971) proposed that the unit is an allochtonous flysch of Eocene age. These clastic rocks, on the other hand were defined as Oligocene-Lower Miocene Güneyce formation by Akbulut (1977; 1980) and were included in the Lower Miocene rocks of Beydağları autochthonous. In the western section of the western Taurus ranges Erakman et al. (1982) named the unit as Kemer flysch of Oligocene age. Bölükbaşı (1987*a,b)* differentiated these clastic rocks as two units: the Upper Eocene-Lower Miocene Kemer tectonic unit and the Upper Paleocene-Oligocene Sülekler formation. Around Isparta-Burdur region, Yalçınkaya et al. (1986) and Yalçın kaya (1989) proposed that these clastic rocks were included in the autochthone and named as Ağlasun formation (Lower Miocene) and as Yavuzlar formation (Eocene) in different locations. Altunsoy (1999), similarly proposed that this unit is to be considered in Lower Miocene rocks of the Beydağları autochthone and named it as Ağlasun formation. The Yeşilbarak nappe was defined as Elmalı thrust fault slice of Upper Paleocene-Oligocene age by Özkaya (1990, 1991). Colins and Robertson (1997. 1998) named the unit as Yavuz unit and Yavuz thrust sheet, while Bilgin et al. (1997) proposed the age of the unit was Upper Lutetian-Lower Miocene and they named it as Elmalı formation.

Şenel et al. (1986, 1987, 1989, 1994) defined the unit as an intermediate zone, but later on (Şenel, 1997a, *b, c, d, e, f, h)* they accepted the earlier definition of Önalan (1979) and used the term Yeşilba rak nappe and its definition for the whole western Taurus ranges. Şenel et al. (1986, 1987, 1989, 1994) and Şenel (1997a, *b, c, d, e, f, h)* studied many rock sections sampled from these clastic rocks all around the western Taurus ranges and concluded that the unit comprises two structural units; 1) the lower Gömbe unit of Upper Lutetian-Lower Miocene age and 2) the upper Yavuz formation of Upper Lutetian-Priabonian age. The researchers state that at the bottom of the Gömbe unit 60 m thick neritic carbonates of Upper Cretaceous age is situated.

THE STRATIGRAPHIC FEATURES OF YEŞİLBARAK NAPPE

The Yeşilbarak nappe is observed as tectonic windows in the front borders of the Lycian nappes and beneath the Lycian nappes in the western Taurus ranges (Fig. 2). These allochthonous masses were made up of clastic rocks in general and having intermediate zone character were divided into two structural (tectono-stratigraphic) units as Gömbe and Yavuz (Şenel et al., 1986, 1987, 1989, 1994). The generalized stratigraphic features of these structural units were given in figure 3.

The Gömbe unit

The Gömbe unit (Fig. 3) is the lower structural unit of the Yeşilbarak nappe and is represented by the Cenomanian-Santonian Gebeler formation made up of neritic carbonates, the Upper Lutetian-

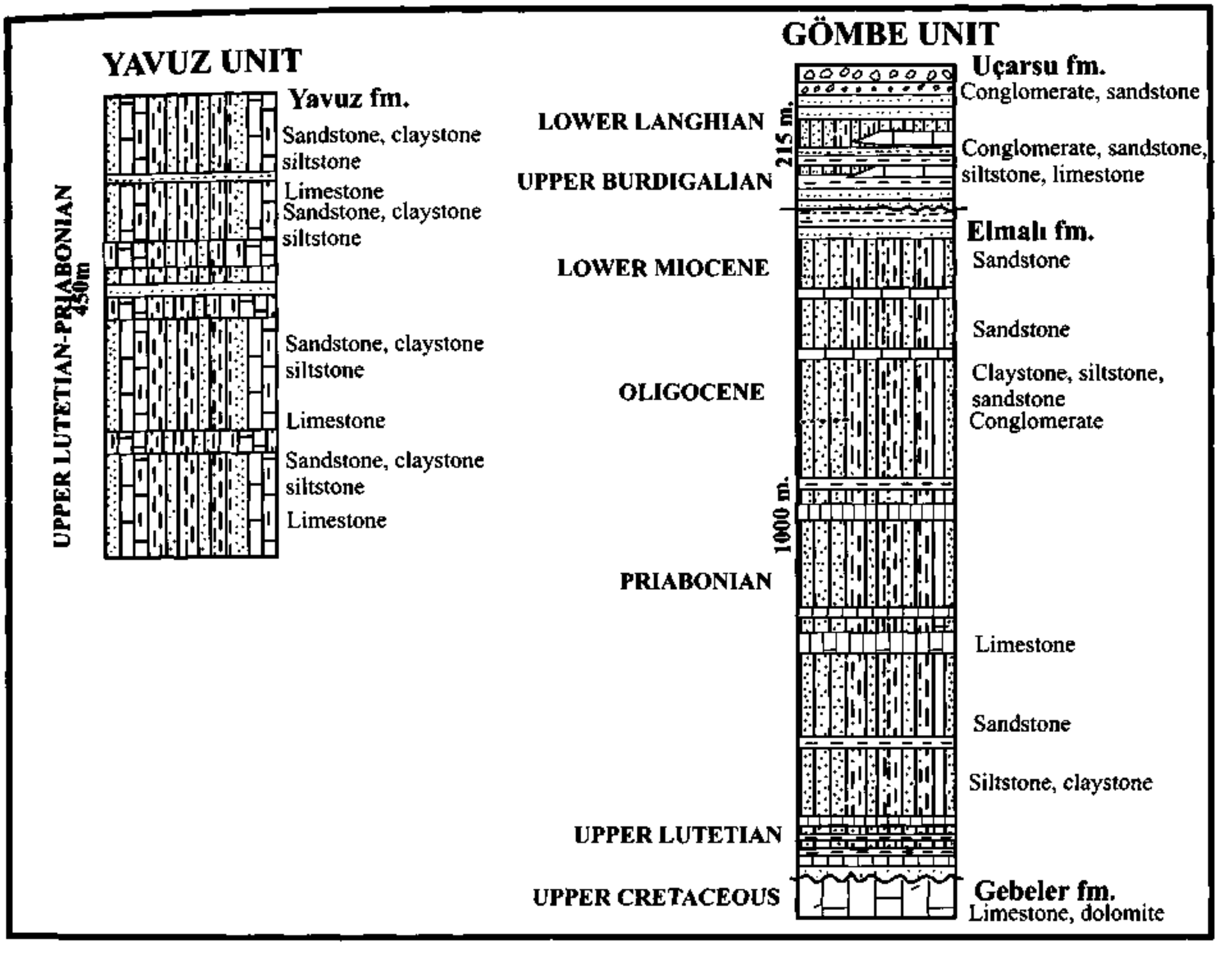

Fig. 3 – Generalized columnar section of structural units of Yeşilbarak nappe.

Lower Miocene Elmalı formation which is made up of sandstone and shale and the Upper Burdigalian-Lower Langhian Uçarsu formation bearing sandstone and conglomerate (Şenel et al., 1989). Of these, the Elmalı formation is widespreadly observed along the belt. On the other hand, the Gebeler formation can only be observed in the Gebeler district east of Fethiye (Fig. 4). Similarly, the Uçarsu formation crops out on the eastern slope of the Akdağ (Fig. 5), situated in between Fethiye-Elmalı.

The Gebeler formation. - This formation which is situated at the base of the Gömbe unit is named by Şenel et al. (1989). All along the western Taurus range it only crops out near a hot spring (Fig. 4) in the Gebeler district, approximately 25 km east of Fethiye. It is made up of massive, medium to thick bedded, dark gray, blackish gray, black and dark brown colored, stinky, hard, highly fractured and jointed karstic limestone, dolomite and dolomitic limestone. The limestones bear milliolides in places and are of biomicrite, biosparite and intrabiosparite character.

The lower contact of the Gebeler formation can not be observed but it is overlain by the Elmalı formation with angular unconformity. It is almost 60 m thick and poor in fossils. Depending on the fossils such as *Thaumatoporella parvoesessiculifera* Rannier, *Raadoshouuenia?* sp., *Biblanata* Sp., *Sgrossoella* sp., *Cuneolina* sp. its age is accepted as Cenomanian-Santonian The rocks similar to these carbonates which were deposited in shallow carbonate shelf environment can be observed in the synchronous units of Beydağları autochthonous and Dumanlıdağ nappe (Şenel, 1994).

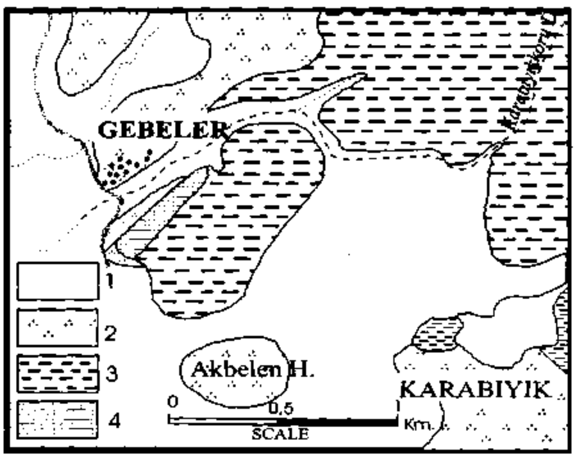

Fig. 4– The map showing exposure of Gebeler formation at Gebeler province southeast of Fethiye; 1-Quaternary, 2-Pliocene, 3-Elmalı formation, 4-Gebeler formation.

The Elmalı formation. - This formation, formed by sandstone, siltstone and claystone in general, has first been named by Önalan (1979). It is the most widespreadly outcropping unit of the Yeşilbarak nappe all along the belt and is highly folded and thrusted.

The Elmalı formation is represented by thin to medium, thick bedded gray, green, dark gray, light brown colored sandstones, siltstone and claystones with limestone intercalations. The dominant lithology of the unit is sandstone which grain size vary between coarse to fine and composed of various lithologic origins. They display turbiditic character and in places pillow structure and the grading may vary in between well to poor. The basal structures were well developed in these sandstones. The siltstones and claystones which are relatively less abundant than the sandstones are darker in color, foliated and have gained the appearance of shale. Sometimes these are thick enough to differentiate as layers. The limestones which are observed as interbeddings and lenses are as sandy limestone, calcarenite, micrite and clayey micrite and generally situated in the lower levels (in Upper Lutetian-Priabonian levels), and rarely in the uppermost levels (in Lower Miocene levels) with thicknesses varying in between 7-8 m. They contain nummulites and planctonic foraminifera in places, and have blocky appearances due to intensive deformation. In the Elmalı formation, rare multi-component conglomerates and debris flows can be seen.

The Elmalı formation is generally situated on the Lower Miocene elastics of Beydağları autochthone with tectonic contact. The only exception is that, in the Gebeler district, east of Fethiye, it rests on the Gebeler formation with angular unconformity (Fig. 4). The unit is technically overlain by the Yavuz unit, the upper structural unit of the Yeşilbarak nappe or the Lycian nappes. On the eastern flanks of the Akdağ, between Fethiye and Elmalı, the Elmalı formation is overlain by the Uçarsu formation of Upper Burdigalian-Lower Langhian age, conformably and unconformably, in places. It is difficult to measure the thickness of the formation since it is highly thrusted and fractured, but in places it was measured to be exceeding 1000 m. However, it is thought to be more thicker (Şenel etal., 1989).

The clastic rocks of the formation are poor in fossil content contrary to the limestone interbeddinqs. In the lower sections of the. formation *Nummulites millecaput* Boubee, *Naturicus* (Joly-Lrymerie), *N, holveticus* (Kaufman), *N.* cf. *fabianii* (Prever), *N.* cf. *munieri* Ficheur, *Chapmanina gassiensis* Silvestri, *Fabianina cassis* (Oppenheim), *Eorupertia magna* Le Calvez, *Linderina brugesi* Schlumberger, *Sphaerogypsina globus* Reuss, *Halkyardia minima* 'Liebus, *Globorotalia centralis* Cushman-Bermudez, *Globigerosistf. kugleri* Loeblich-Topp, *Assilina* sp., *Alveolina* sp., *Globigerina* sp., *Discocyclina* sp., and in the upper levels *Lepidocyclina* sp. *(Nephrolopidin* type), *Miogyssinoides complanatus* (Schlumberger), *Amphistegina* cf. *lessoni* D'Orbigny, *Globoquadrina* cf. *dehicens* (Ch:-Parr.-Col.), *Catapsydrax* cf. *dissimilis* (Cushman-Bermudez), *Globigerinoides* cf. *trilobus* Reuss, G. cf. *bisphericus* Todd, G. cf. *diminitus* Bolli, *Globigerina* sp., *Globigerinata* sp., *Globigerinatella* sp., *Operculina* sp. forms were determined. The fossils in the lower levels indicate Upper Lutetian-Priabonian, whereas the fossils in the upper levels indicate Lower Miocene age. The fossils collected from the central sections are nannoplanktons such as *Spherolithus distendus* (Martini), S. *Predistendus* Bramlette-Wilcoxon, *Cyclicargolithus abisenctus* (Muller), *C. Floridanus* (Roth.-Hay), *Helicopontosphaera intermedia* (Martini), *H. Recta* (Hao), *H. Seminilum* (Bramlette-Sullivan) which indicate Oligocene age. These data show that the Elmalı formation is of Upper Lutetian - Lower Miocene age. No unconformity in the formation has been observed. However, in the-Western Taurus ranges, and moreover in the Central Taurus ranges a continuous sequence between Upper Lutetian and Lower Miocene has not been identified. Therefore, the presence of an unconformity at post-Upper Lutetian-Priabonian and pre-Lower Miocene is possible and it may not have been observed due to intensive deformation and also the similarity of the elastics in the lower and upper levels.

The Elmalı formation displays transgressive features at the base, it has been deposited on shelf slope-basin medium.

The Elmalı formation has -at least partial- similarities with the Varsakyayla formation of Upper Lutetian-Priabonian age (Poisson, 1977; Şenel et al., 1989) and the Yavuz formation which constitutes the upper structural unit of the Yeşilbarak nappe of the same age and the Küçükköy formation (Poisson, 1977; Şenel, 1997h,j) that is observed on the upper levels of the Beydağları autochthonous. In the above-mentioned formations, the carbonate interbeddings are much more compared to the Elmalı formation. In the Western and Central Taurus ranges there are no elastics that are similar to the Oligocene-Lower Miocene elastics of the unit.

The Uçarsu formation - This formation is comprised of sandstone with abundant macro fossils and conglomerate and first named by Şenel et al. (1989). It can only be observed in the Western Taurus ranges, on the eastern flank of the Akdağ in between Fethiye and Elmalı (Fig. 5) and on the northeast of the Deliklitaş Hill east of the Döğüş district.

The Uçarsu formation can easily be differentiated from the Elmalı formation by its abundant content of macro fossils and high content of coarse elastics. The unit is

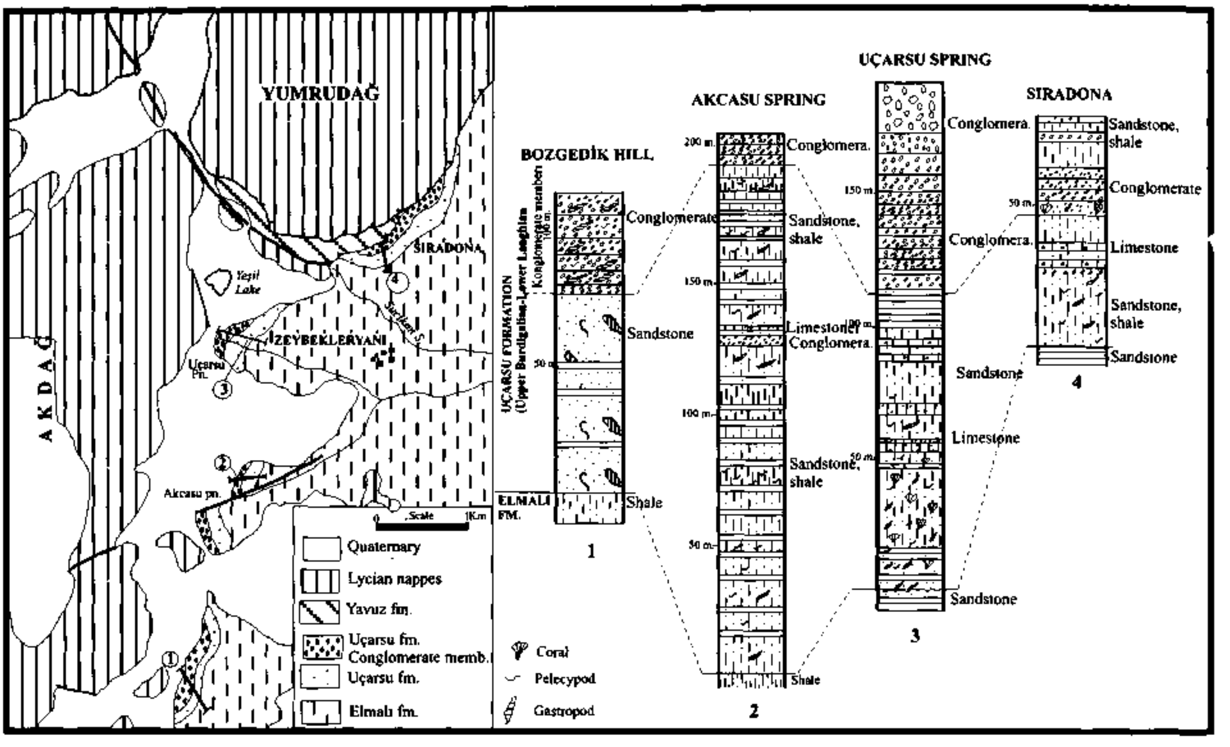

Fig. 5 – Geological map and columnar section of Uçarsu formation

composed of thin-medium-thick bedded, green, greenish gray sandstone, claystone and siltstone, and multi-component conglomerates (Fig. 5.b1). The formation rests on the shales of the Elmalı formation conformably as thick bedded, light gray to green colored, poorly sorted coarse sandstones including gastropods, corals, echi nides. This sandstone is overlain by thick bedded, well sorted, conglomerates with rounded gravels with abundant macro fossils. The elastics that may appear as shales due to foliation may be differentiated from the Elmalı formation by their abundant macro fossil content. The elastics contain 30 cm- thick sandy limestone level including benthic foraminifera and macro fossils. There is a very thin grained conglomerate level below this unit. In the Uçarsu section (Fig. 5.b3) the unit comprises thin-medium-thick bedded, green, greenish gray colored sandstone including gastropoda, lamelli branch, corals and macro fossils, claystone, siltstone, and thick overlying conglomerates. In the lower levels where the sandstone is dominant, limestone intercalations of bioherm nature including abundant ben thic foraminifera and macro fossils can be observed. These limestone intercalations that may contain sand in places, pinch out in elastics. The overlying conglomerates are thick bedded, but in the upper levels they are not bedded. They may contain sand lenses and pebble fills. The gravels which are small in size and angular in the lower levels grow in size in the upper levels (up to 70 cm) and they have sharp corners; the sorting is well in the lower levels but poor in the upper levels. There are no macro fossils in these coarse elastics but instead relicts of plants can be seen. In the measured Sıradona section (Fig. 5.b4) the Uçarsu formation comprises, in the lower levels, massive, thick bedded, green in color sandstone, claystone and siltstone with macro fossils. Above these elastics there are conglomerates and an intercalation of conglomerate, sandstone and claystone. Another outcrop of the Uçarsu formation can be seen in the Döğüş district, in the east but here the formation is represented with poligenic conglomerates.

The Uçarsu formation is in turn overlain by the Lycian nappes and the Yavuz formation with tectonic contact. The maximum thickness of the formation is 215 m. It frequently changes lithology in lateral direction.

The formation is rich in macro and micro fossils. Micro fossils such as *Miogy psina irregularis* (Micheloss), *M.* cf. *intermedia* Droger, *Miogypsinoides dehaartii* (Van Der Klerk, *M.* cf. *bantamensis* Tan, *Amphistegina lessoni* D'Orbigny, *Operqu lina complanata* Defrance, *Spiroplectammina carinata* D'Orbigny, Nonion *pompilioides* D'Orbigny, *Globigerinoides* cf. *trilobus* Reuss, *Globigerina* sp., *Ditrupa* sp., *Acervolina* sp., *Gypsina* sp., *Victoriella* sp., *Lithohammium* sp., *Cibicides* sp., *Robulus* sp., corals such as *Thegiostraea crassi coslata* (Michelotti), *Heliastraea oligophyllia* Reuss, *Aquitanastraea quetterdi* (Michelin), *Siderastraea miocenica* Osasca, *Stylophera* cf. *reussiana* Montanara-Galitelli, *Acanthocyathus trasiluencus* Reuss, *Balanophyllia conconna* Reuss, *Leptomussa? Faloti* Chevaier, gastropoda such as *Turitella* (Turitella) cf. *terebralis* Lamarck, *T. terebralis terebralis* Lamarck, *Ancilla* (Baryspira) *glandiformis* (Lamarck), *Conus* cf. *betulinoides* Lamarck, *Strombus* sp., *Natica* sp. and pelecypods such as *Pecten* cf. *josslingi* Smith, *Pecten fushsi styriacus* Hilbe'r, *Glycymeris* (Glycymeris) *inflatus* Brocchi, *Venus* cf. *multilamella* Lamarck, the age of the formation is determined as Upper Burdigalian-Lower Langhian.

The Uçarsu formation was deposited on the Elmalı formation in shelf environment with slow regression. However, since the terrestrial input to this high-energy medium is too intense, in general, the formation of a wide reef is hindered but instead small reefs in patches were formed. Related to the emplacement of the Lycian nappes the basin has become shallower and finally has closed, giving way to accumulation of coarse material and hence the fans. Moreover, as in the case of the Uçarsu section, alluvial fans in the upper levels have developed.

The Uçarsu formation can be correlated with the Kasaba formation of the Beydağları autochthonous (Şenel et al., 1989, 1994) and with the transgressive rock units (Becker-Platen, 1970; Hakyemez and 6rgen, 1982; Şenel et al., 1989) observed as small outcrops over the Lycian nappes.

The Yavuz unit

The Yavuz unit, which is the upper structural unit of the Yeşilbarak nappe, is represented by the sandstone, claystone and limestones of Upper Lutetian-Priabonian age. It tectonically overlies the Lower Miocene elastics of the Beydağları autochthonous and the Gömbe unit, which is the lower structural unit of the Yeşilbarak nappe and is overlain by the Lycian nappes. In the fore front of the Lycian nappes it is generally observed as overturned.

The Yavuz formation.- This formation comprises of limestone, claystone and sandstones and is named by Poisson (1977). It is difficult to differentiate this formation with the Elmalı formation due to the similarity of the lithology. The formation is quite widespread around Korkuteli area and is made up of an intercalation of claystone with dominant limestone, limestone, siltstone and sandstone in the lower levels and that of sandstone, claystone and siltstone in the upper levels. The limestone has micritic texture and is thin to medium bedded, beige and light gray in color with planktonic foraminifera. It sometimes contains chert nodules. These are as interbeddings reaching up to 20 m. The micrites pass into the clayey limestone and marls in the upper levels. There are thin to medium and rarely thick, beige and light gray, light brown in color, calcarenites and clayey limestones in the formation. These are below or in between the micrites, sandstones and claystones. There are small nummulites in calcarenites and limestones. Flow structures can be seen at the base of the calciturbidites. The sandstones, siltstones and claystones, which are thin-medium-thick bedded with gray, light gray, green, greenish gray in color are of turbidite nature. The claystones and siltstones are sometimes foliated and appears as shale. The formation sometimes observed to comprise conglomerates as thin layers, some limestone layers appear as apart blocks due to intensive deformation. In the lower levels of the Yavuz formation red clayey limestone and claystones as marker beds are present and they extend laterally.

The Yavuz formation technically overlies the Lower Miocene elastics of the Beydağları autochthonous and the lower structural unit of the Yeşilbarak nappe, the Gömbe group and is overlain by the Lycian nappes technically. The thickness of the unit is measured as 450 m, however, Poisson (1977) proposes that its thickness may exceed 750 m. The unit does not display lateral change in lithology.

The biostratigraphic features of the unit was discussed in detail by Poisson (1977). According to its fossil content, such as *Nummulites* cf. *millecaput* Boubee, *Sphaerogypsina globulus* Reuss, *Globorotalia* cf. *bulbrooki* Bolli, *Eorrupertia magna* Le Calvez, *Nummulites* sp., *Discocyclina* sp., *Alveolina* sp., *Globorotalia* sp., *Globigerina* sp., *Truncorotaloides* sp., etc., the age of the formation is Upper Lutetian-Priabonian (Poisson, 1977; Şenel, 1989).

The unit comprises similar lithology with the Varsakyayla (Poisson, 1977; Şenel et al., 1989) formation which is transgressive over the Lycian nappes.

THE OCCURRENCE OF THE YEŞİLBARAK NAPPE AND ITS STRUCTURAL SETTING

The Yeşilbarak nappe which is observed below the Miocene nappes in southwest Turkey and occurring as an intermediate zone between the Lycian nappes and the Beydağları autochthonous, can continuously be observed in the fore front of the Lycian nappes and as tectonic windows at the back of the fore front of the Lycian nappes. The drillings made by TPAO shows that the Yeşilbarak nappe taking place between the Lycian nappes and the Beydağları autochthonous display various thickness in the area.

The Yeşilbarak nappe can be observed in between Dalaman (east of Köyceğiz) and southeast of Isparta and is represented by the Elmalı formation that belongs to its lower structural unit (Gömbe) west of the area Fethiye-Akdağ. The westernmost outcrops of the unit can be observed in Karadere (Fig. 6), Kargın and Günlük (Fig. 7) tectonic windows, 5 km southeast, 4 km east and 10 km north of Dalaman, respectively. The Yeşilbarak nappe, which is represented only by the Elmalı formation in these tectonic windows, is overlain by the Marmaris ophiolitic nappe and the Tavas nappe. In Günlük and Kargın tectonic windows, the nappe can not be observed. In Karadere tectonic window, the Yeşilbarak nappe overlies the Burdigalian claystones (Sinekçi formation-Çayboğazı member) of the Beydağları autochthonous. The area where the Yeşilbarak nappe crops out widely Is the Göcek-Aygırdağı tectonic window. (Fig. 8). In this area where the Beydağları autochthonous is uplifted by normal faults, only the Elmalı formation of the nappe can be observed. The Yeşilbarak nappe which tectonically overlies the Burdigalian claystones (Sinekçi formation-Qaybogazi member) of the Beydağları autochthonous, is tectonically overlain by the Tavas nappe, the Marmaris ophiolitic nappe and partly by the Bodrum nappe. In Eldirek tectonic window (Fig. 9), 11 km east of Fethiye,. the lower contact of the Yeşilbarak nappe can not be observed. The unit is overlain by the Marmaris ophiolitic nappe, the Bodrum nappe and the Gülbahar nappe-Middle-Upper Triassic Çövenliyayla volcanics of the Ağla unit. The Yeşilbarak nappe which lower contact can not be

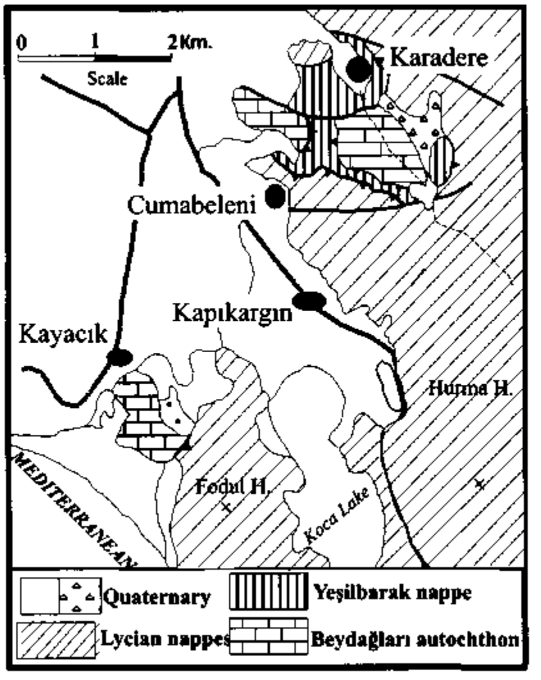

Fig. 6 – Geological map of Karadere tectonic window and surrounding area.

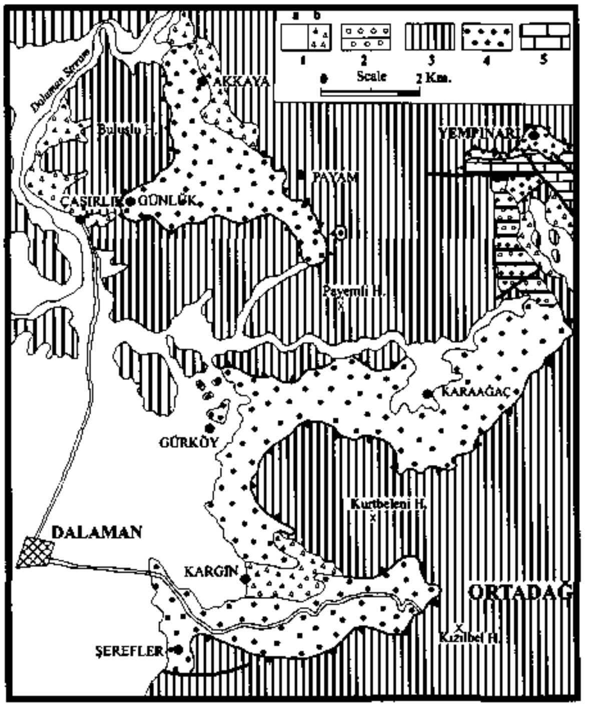

Fig. 7 – Geological map of Kargın and Günlük tectonic windows and surrounding areas; 1- Quaternary- a) Alluvium,- b) Slopesorce deposits, 2- Pliocene, 3- Lycian nappes, 4- Yeşilbarak nappe, 5- Beydağları autochthone.

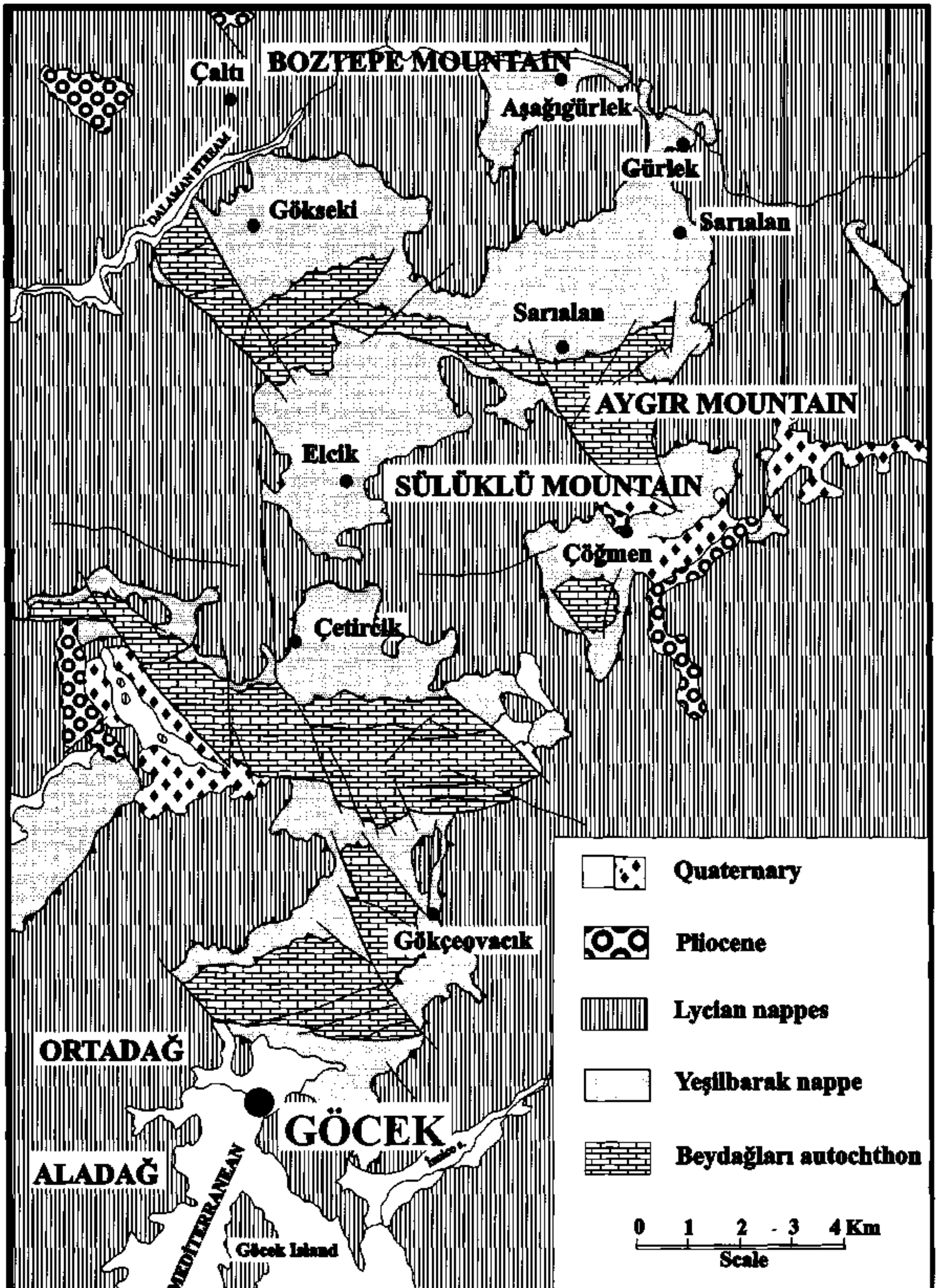

Fig. 8 – Geological map of Göcek-Aygır dağı tectonic Windows and surrounding areas.

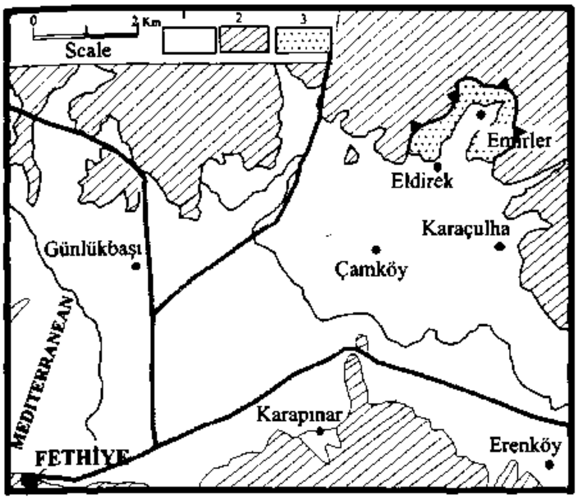

Fig. 9 - Geological map of Eldirek tectonic window and surrounding area; 1- Quaternary, 2- Lycian nappes (Marmaris ophiolite nappe), 3. Yeşilbarak nappe (Elmalı fm.).

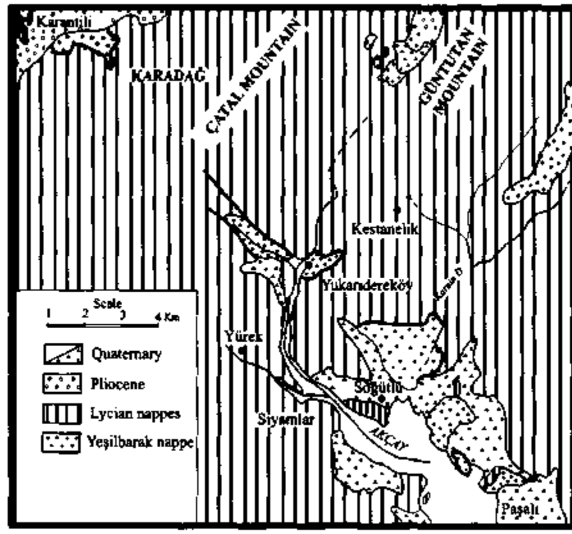

Fig. 10 – Geological map of Söğütlüdere, Karantili tectonic windows and surrounding areas.

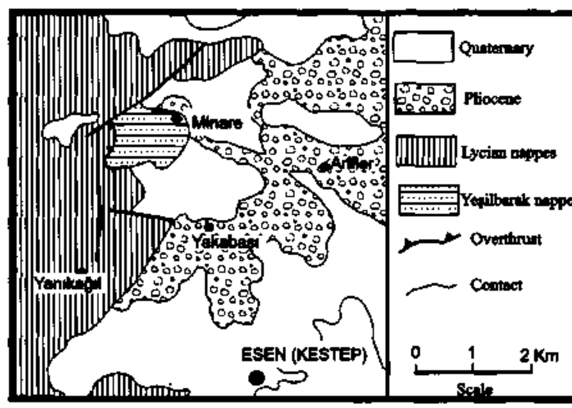

Fig. 11 – Geological map of Minare tectonic window and surrounding area.

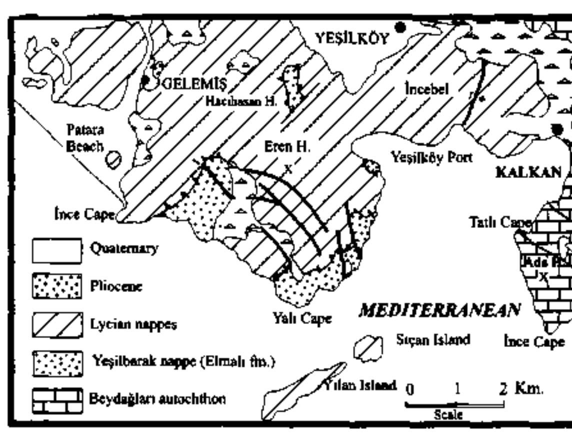

Fig. 12 – Geological map of Yalıburnu tectonic window and surrounding area.

observed in Söğütlüdere and Karantili tectonic windows (Fig. 10), is only represented by the Elmalı formation and is overlain by the Tavas nappe, the Gülbahar nappe and the Marmaris ophiolitic nappe. In Minare tectonic window (Fig. 11), 4.5 km north of Eşen (Kestep), between Fethiye and Kalkan, the Yeşilbarak nappe can be observed in a very narrow area and its lower contact can not be observed. It, here, is represented by the Elmalı formation and is overlain by the Tavas nappe, technically. In Yalıburnu tectonic window (Fig. 12), 7 km southwest of Kalkan, the Yeşilbarak nappe is represented only by the Elmalı formation again, and its lower contact can not be observed. In this area the nappe is overlain by the Dumanlıdağı nappe. In Keller tectonic window (Fig. 13), west of Burdur-Antalya (Dirmil), the Elmalı formation has a wide outcrop. However, the lower contact can not be observed in this area, too. In this tectonic window, generally the Lower Miocene rocks of the Elmalı formation are well displayed (Selçuk et al., 1985) as well

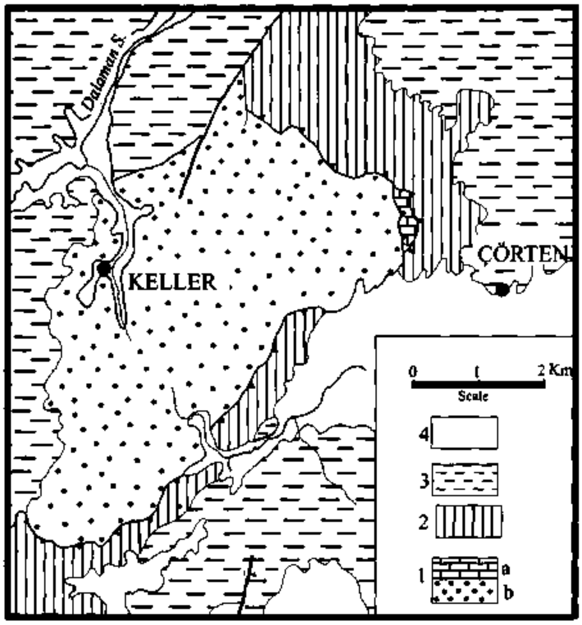

Fig. 13 - Geological map of Keller tectonic window and surrounding area; 1. Yeşilbarak nappe- a)Elmalı formation, b)Yavuz formation,. 2- Lycian nappes, 3- Pliocene, 4- Quaternary.

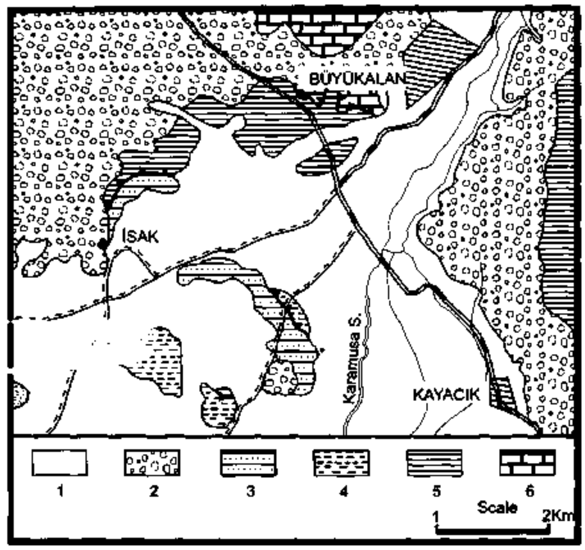

Fig. 14 - Geological map of İsak tectonic window and surrounding area; 1-Quaternary, 2-Pliocene, 3-Elmalı formation, 4-Yavuz formation, 5-Marmaris ophiolite nappe, 6-Gülbahar nappe.

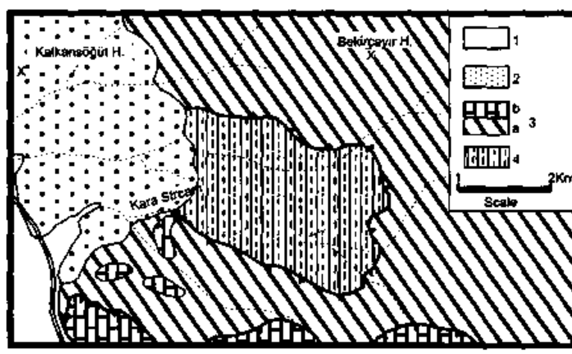

Fig. 15 - Geological map of Çavdır tectonic window and surrounding area; 1-Quaternary, 2-Pliocene, 3-Lycian nappes-a)Marmaris ophiolite nappe, b)Domuzdağ nappe, 4-Yeşilbarak nappe (Elmalı formation).

as a small slice of the Yavuz formation of the Yavuz unit. The Yeşilbarak nappe is overlain by the Marmaris ophiolitic nappe and by the Pliocene terrestrial elastics. In İsak tectonic window (Fig. 14), northwest of Burdur-Çavdır, both of the structural units (the Gömbe and Yavuz units) of the Yeşilbarak nappe can be observed. They are covered by alluviums and terrestrial Pliocene. North of the İsak village, the nappe is technically overlain again by the Marmaris ophiolitic nappe. In this area subophiolitic meta-morphics (amphibolite schist, etc.) can be seen as a thin tectonic slice (Şenel et al., 1989). In Çavdır tectonic window (Fig. 15), 3 km southeast of the Çavdır, only the Elmalı formation can be observed and its lower contact is not visible. The unit is technically overlain by the Marmaris ophiolitic nappe.

The units of the Yeşilbarak nappe which widespeadly crop out in between the Lycian nappes and the Beydağları autochthonous in vicinity of Isparta and Bucak (Fig. 16) has not yet been investigated sufficiently. In the area, the rock units of the Yeşilbarak nappe have been studied with the formations of the Beydağları autoch-

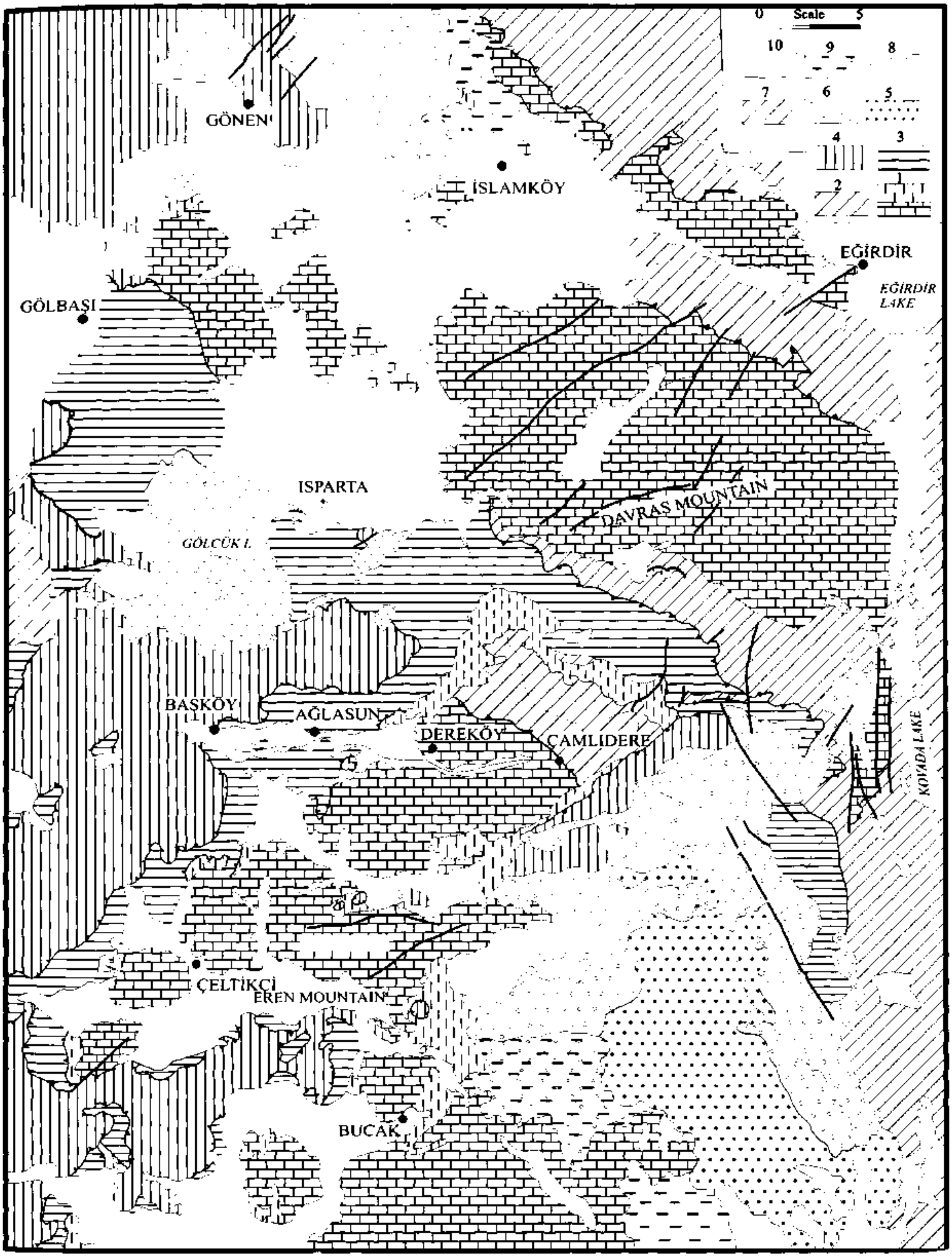

Fig. 16 - Map of structural units of Isparta and surrounding; 1- Beydağları autochthone-a)Pre-Miocene, b)Miocene, 2-Antalya nappes, 3-Yeşilbarak nappe, 4-Lycian nappes, 5- Middle Miocene, 6-Upper Miocene, 7-Pliocene, 8-Gölcük volcanics, 9-Pleistocene, 10-Quaternary.

thonous by researchers such as Gutnic (1971), Poisson (1977), Akbulut (1977), Yalçınkaya et al. (1986), Yalçınkaya (1989) and Altunsoy (1999). The structural features in the area where the Elmalı formation of the lower structural unit of the Yeşilbarak nappe (the Gömbe unit) crops out have not been differentiated. Akay and Uysal (1985), Akay et al. (1985) report the presence of similar units southeast of Lake Kovada-northwest of Bucak (Fig. 2). They are really similar to the Elmalı formation and must be belonging to the Yeşilbarak nappe. Gutnic (1971) reports the presence of allochthonous Eocene flysch in vicinity of Dinar-Kegiborlu area (Fig.2). This flysch also must be belonging to the Yeşilbarak nappe.

The widest outcrops are displayed by the Elmalı formation of the Yeşilbarak nappe can be divided into two structural units. The Elmalı formation has wide outcrops in between Elmalı and Dalaman (Fig. 2). Similar wide outcrops can be observed in between north of Korkuteli and Isparta (Fig. 2). The other formations of the Gömbe unit, the Gebeler and the Uçarsu formations are observed in very limited areas. The Yavuz unit which can only be represented by the Yavuz formation is observed in between northwest of Korkuteli and west of Elmalı. The Yavuz unit, on the other hand, can be observed in northeast of Fethiye-Kemer, to the north of Akdağ and Yumrudağ, in north and northwest of Akçay as thin, small overturned slices. Between Bucak and Isparta the unit can not be observed due to lack of detailed investigation, however, it can be observed in the east and north of Gökçebağ and northwest of Keçiborlu.

In the fore front of the Lycian nappes, the Yeşilbarak nappe can be observed on the Burdigalian Çayboğazı member (comprised of claystones) of the Sinekçi formation in between Dalaman and Elmalı, and the Lower Miocene Karakuştepe formation of (comprised of sandstone, siltstone and claystone) and rarely on the Upper Burdigalian-Lower Langhian Kasaba formation (comprised of conglomerate and sandstone). The nappe has emplaced on the Beydağları autochthonous in the end of Lower Miocene and in the beginning of Middle Miocene (in Lower Langhian) along the fore front of the Lycian nappes. In the same period, around Isparta-Çay (Fig. 16) the Yeşilbarak nappe and the Lycian nappes were emplaced on the Antalya nappes in Danian and as well as on the Karakuştepe formation overlying the Lower Miocene Karabayır formation, both with tectonic contact. Some researchers (Poisson, 1977; Yalçınkaya et al., 1986), however, include the Elmalı formation in the Lower Miocene Karakuştepe formation which is made up of elastics. The formations belonging to the Yeşilbarak nappe has been mapped, even if partially, by Bölükbaşı (1987b). The nappe, on the southern flanks of the Davras mountain, east of Isparta, has been technically overlain by the Tertiary elastics of the Beydağları autochthonous, the Kızılcadağ melange and the olisthostromes of the Lycian nappes and the structural units of the Antalya nappes. The Yeşilbarak nappe has been thrusted over the Tortonian Aksu formation south of Lake Kovada and related to this thrusting, the Antalya nappe has been thrusted over the Yeşilbarak nappe. This thrusting of rocks of the Beydağları autochthonus and the Antalya nappe over the Elmalı formation of the Yeşilbarak nappe in the abovementioned areas must be related to the Aksu thrust (Aksu phase; Poisson, 1977) that took place north of the Gulf of Antalya (Şenel et al., 1996).

REGIONAL COMPARISON OF THE YEŞİLBARAK NAPPE

Southeast Anatolia (western Taurus ranges) has been exposed to emplacement of large scale allochthonous masses in the end of Lower Miocene and in the beginning of Middle Miocene (Lower Langhian). Almost at the same time, similar emplacements (the Bitlis-Pötürge-Malatya nappes) were observed in southeast of Turkey (Ricou, 1979; Şengör and Yılmaz, 1981; Aktaş and Robertson, 1984; Göncüoğlu and Turhan, 1984; Perinçek and Kozlu, 1984; Yılmaz and Yiğitbaş, 1990; Perinçek, 1990). These Miocene nappes in southeast Turkey have been differentiated and named as the Keban metamorphics, the Malatya metamorphics, the Pötürge metamorphics, the Bitlis metamorphics (Tolun, 1954), the Baskil magmatites, the Yüksekova complex (Özkaya, 1977), the Bitlis-Pötürge nappe (Aktaş, and Robertson, 1984), the Upper nappe and the Lower nappe (Yılmaz et al., 1991), the Bitlis-Pötürge-Malatya nappes (Şenel, 1999), the Mordağ metamorphics (Özkaya, 1977; Perinçek, 1990), the Hakkari complex (Maxson, 1937), the Maden complex (Ketin, 1948), and the Çüngüş formation (Sungurlu, 1974; from Yılmaz and Duran, 1997). Of these allochthonous masses that emplaced in Miocene, the Çüngüş formation and the Maden complex, both by their structural setting and by their stratigraphic features, are very similar to the Yeşilbarak nappe observed in southeastern Turkey. These masses that have been studied in detail by Özkaya (1977), Perinçek (1990), Yılmaz and Duran (1997) were observed as an intermediate zone in between the autochthonous rock units and the allochthonous rocks that emplaced in Miocene and were observed along a thrust zone (Bitlis thrust zone) as similar as that observed in west (Fig. 17). The Çüngüş formation observed in between the Miocene

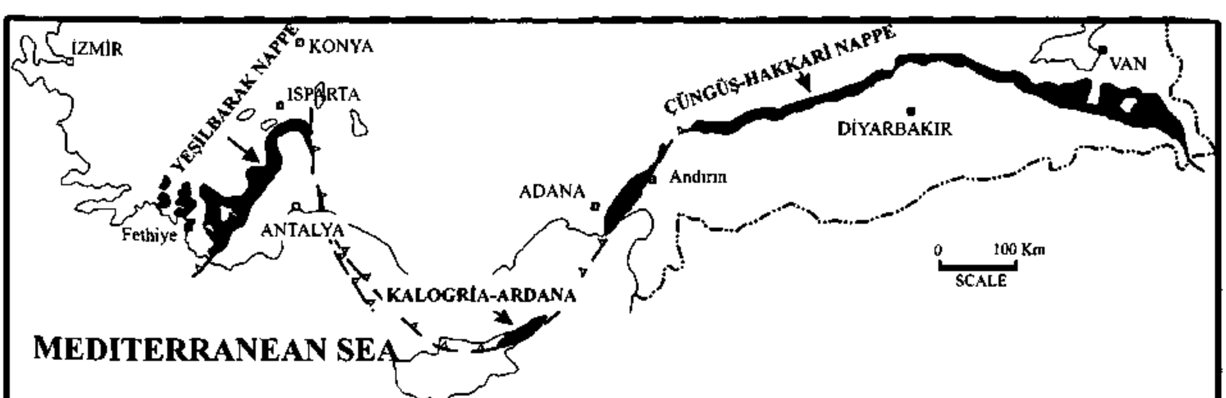

Fig. 17 - The alignment of Eocene-Lower Miocene allochthonous clastic rocks (Yeşilbarak nappe, Çüngüş-Hakkari nappe etc.) as intermediate zone beneath Miocene nappes (Lycian nappes, Bitlis-Pötürge-Malatya nappes) at southern Turkey.

nappes in southeast Anatolia and southeatern Anatolian autochthonous, and the Hakkari complex will be discussed as the Çüngüş-Hakkari nappe in this paper. The Çüngüş formation, the lower structural unit of the Çüngüş-Hakkari nappe was first named by Sungurlu (1974; from Yılmaz and Duran, 1997). It can be accepted as the first allochthonous structural unit (below the Miocene nappes) on the southeast Anatolian autochthone represented by Eocene-Lower Miocene elastics. It is lithologically and structurally similar to the Upper Lutetian-Lower Miocene Elmalı formation of the lower structural unit (the Gömbe unit) of the Yeşilbarak nappe and is continuous along the Bitlis suture zone.

The Çüngüş formation (Fig. 18) cornrises thin-medium-thick bedded, green, gray, greenish gray, yellowish gray sandstones, siltstones and marl (Perinçek, 1990; Yılmaz and Duran, 1997). Local conglomerates and thin limestone intercalations may also be seen in the unit which has underone intensive deformation and hence appears as to be thrusted, folded and fractured. It may contain blocks, in places. The Çüngüş formation has turbiditic character ana debris flows may be seen on the formation. The lower and upper contact of the formation is tectonic. Its thickness varies in between 200-1500 m. The age of the formation is Eocene-Lower Miocene.

The Hakkari complex situated under the Miocene nappes (the Bitlis-Poturqe-Malatya nappes) in southeast Anatolia has first been named by Maxson (1937). It is the upper structural unit of the Çüngüş-Hakkari nappe and generally is of Eocene age but sometimes reaches up to Oligocene. The Hakkari complex (Fig. 18) comprises more or less different Urşe and Durankaya formations which have tectonic contacts in between (Perinçek, 1990; Yılmaz and Duran, 1997). The Urşe formation has first been named by Perinçek (1977; from Yılmaz and Duran, 1987). The age of the formation is Eocene-Oligocene and its lower and upper contacts are tectonic (Yılmaz and Duran, 1997). It comprises thin-medium-thick, gray, green, greenish gray, reddish sandstones, shale and limestones (Yılmaz and Duran, 1997). Its thickness may reach up to 2075 m. The Durankaya formation named by Perinçek (1978; from Yılmaz and Duran, 1997) is of Lower-Upper Eocene age and comprises of sand-stone with serpantinite, gabbro, basic volcanics, marble, limestone, amphibolite etc. blocks, shale and conglomerate. The blocks are in an olisthostromal fades. There are red pelagic limestone and gray limestone blocks and lenses in the unit. Many of the limestones were broken and in form of blocks due to intensive deformation. The red limestones bear planctic foraminifera and may reach up to thickness (150-200 m) to form hills. There are chert nodules in these limestones which do not have lateral continuity. The grav limestones have abundant amounts of reworked and broken nummulites. The Durankaya formation comprises of lithologies that have undergone low grade metamorphism. The Urşe and Durankaya formations may be correlated with the Yavuz formation with respect to their lithologies and structural setting. However, the Yavuz formation does not contain onsthostrome facies and blocks. Kozlu (1997) reports about the existence of Tertiary allochthonous elastics to the south of Engizek mountain, although

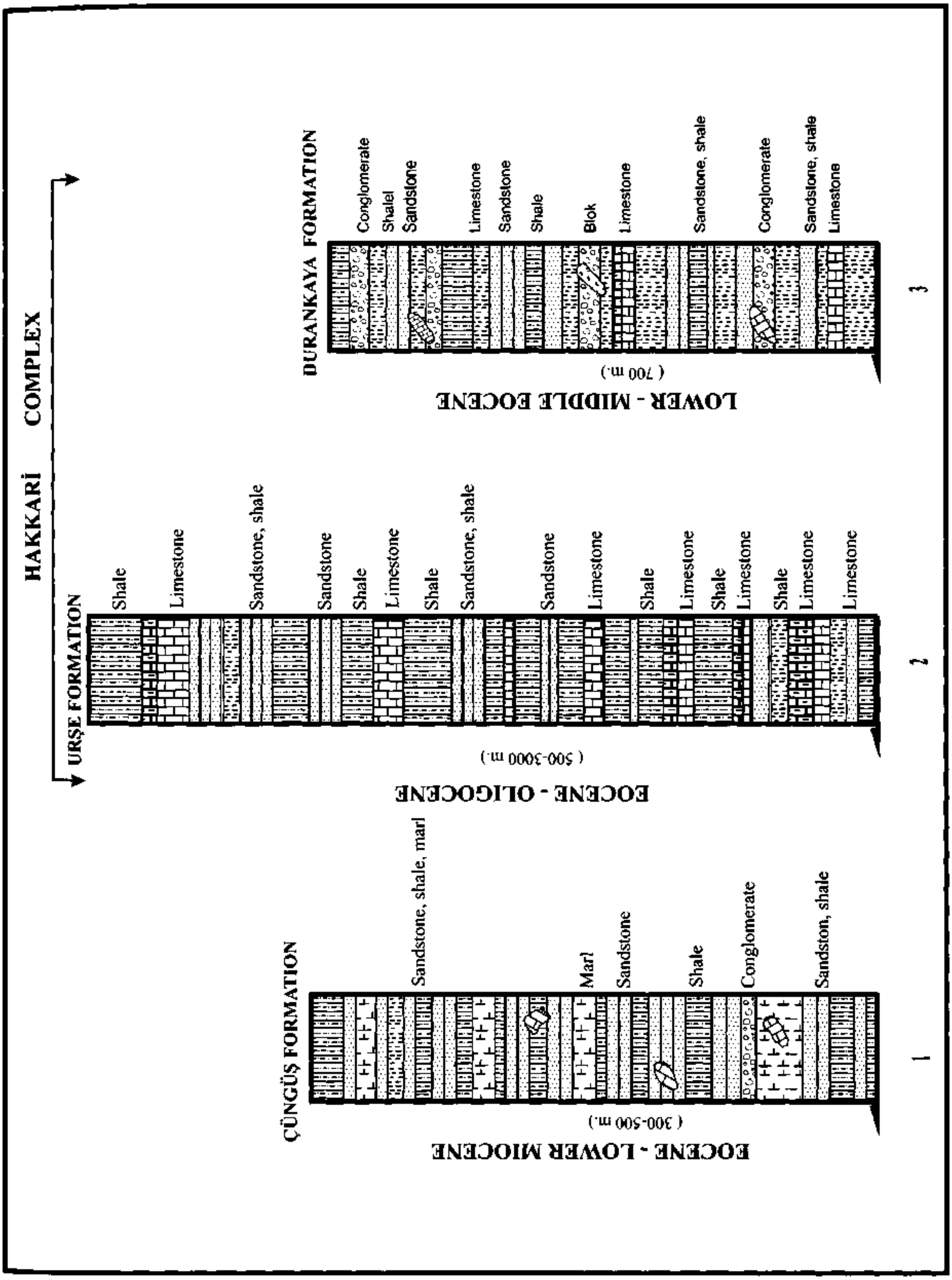

Fig. 18 Generalized columnar sections of the units of Çüngüş-Hakkari nappe

the features and the origin are not known well. In the area around Andırın similar rocks were observed (Kozlu, 1997). The elastics resembling the Çüngüş formation overlying the Lower Miocene around Adana-Andırın were observed during our studies but no detailed investigation was carried out.

The rock units similar to that forming the Yeşilbarak nappe and the Çüngüş-Hakkari nappe were observed in Cyprus, around the Beşparmak mountains (Fig. 17). The Middle-Upper Eocene turbiditic clastic rocks around the Beşparmak mountains that were named as the Kalograia-Ardana formation and the Mavri Skala flysch by Knup and Kluvyer (1969) and Baroz (1979) were re-defined by Hakyemez et al. (2000) as the Ardahan and Kantara formations. The Mavri skala flysch (Baroz, 1979) or the Ardahan formation (Hakyemez et al., 2000) is reported to unconformably overlie different formations and its age is reported as Upper Lutetian-Priabonian. It is made up of turbiditic elastics and has completely similar features with the Elmalı and Çüngüş formations. The Oligocene-Lower Miocene turbiditic clastic rocks on which the Ardahan formation rests on (Hakyemez et al., 2000) are very similar to those equivalent units in the Elmalı and Çüngüş formations. The Kalogria-Ardana (Baroz, 1979) or the Kantara (Hakyemez et al., 2000) formation is of Upper Lutetian-Priabonian age and is made up of clastic rocks including various blocks. The Andırın formation, a member of the Misis group (Schmidt, 1961; Bilgin et al., 1981; Ayhan and Bilgin, 1986; Ayhan et al., 1988) observed around Adana, is of Upper Lutetian-Priabonian age and is represented by elastics bearing various blocks (Bilgin et al., 1981; Ayhan et al., 1988). With this feature, the formation is very similar to the Middle-Upper Eocene Kalograia-Ardana formation (Knup and Kluvyer, 1969; Baroz, 1979) in northern Cyprus. However, it is known that the Andırın formation includes Lower Miocene, too. The Kantara and Andırın formations may be correlated with the Durankaya formation, the upper structural unit of the Çüngüş-Hakkari nappe in southeast Anatolia. The micritic, clayey micritrc, calciturbiditic, marly and claystone lithologies of Upper Lutetian-Priabonian age that observed around Sangarbulakçeşme and Yenicerisırtı area (south of Sipahili, east of the Beşparmak mountains) have similar character with the lower levels of the Yavuz formation of the Yeşilbarak nappe.

RESULTS AND DISCUSSIONS

The Lycian nappes in the western Taurus ranges have emplaced on the Beydaglari autochthonous and the Yeşilbarak nappe in Lower Langhian (Graciansky, 1972; Poisson, 1977; Şenel et al., 1989; 1994). The lithological features of the Yeşilbarak nappe indicate the presence of a very wide basin (Fig. 19) in between the Beydağları autochthone and the Lycian nappes in which the deposition of turbiditic elastics from the Lycian nappes were dominant in Lutetian-Lower Miocene. When the para-allochthonous transgressif Varsakyayla formation of Upper Lutetian-Priabonian age (Poisson, 1977), the Küçükkoy formation (Poisson, 1977; Şenel et al., 1989) of the same age situated on the Beydağları autochthonous, the Susuzdağ formation (Önalan, 1979; Şenel et al., 1989; 1994) and the synchronous units in the Yeşilbarak nappe are studied, it is understood that the development of this basin started in the beginning of Upper Lutetian.

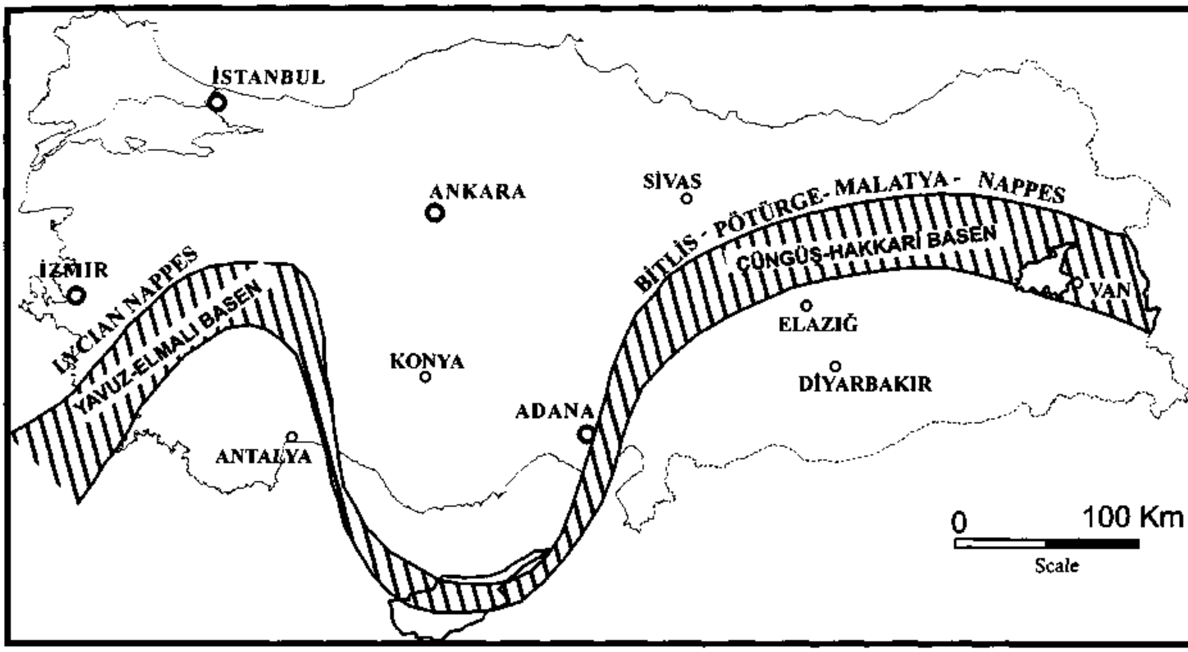

Fig. 19 - Schematic map indicating basins of Yeşilbarak and Çüngüş-Hakkari nappes observed between autochthonous masses and Miocene nappes in Tauride belt.

This basin, known as the Yavuz-Elmalı basin, is the source of the Yeşilbarak nappe which has been thrusted on the Beydağları autochthonous for kilometers-long distance along with the Lycian nappes. This basin was closed in the end of Lower Miocene, at the beginning of Middle Miocene (Lower Langhian).

Similar phenomena have been observed in southeast Anatolia. The lithological features of the Çüngüş formation forming the Çüngüş-Hakkari nappe and the Durankaya and Urse formations of the Hakkari complex (Perinçek, 1990; Yılmaz and Duran, 1997) indicate the presence of a basin where the deposition of the elastics were dominant in between the Bitlis-Poturge-Malatya nappes and the southeast Anatolian autochthonous during Eocene-Lower Miocene (Fig. 19). While the deposition of shelf-type carbonates were prevailing during Eocene-Miocene in the southeast Anatolian autochthone, in the north, in the same period, we can talk about the presence of a deeper basin that was fed by the Bitlis-Poturge-Malatya nappes. These elastics comprising blocks locally imply that the basin is instable. This basin which can be defined as Çüngüş-Hakkari basin (Fig. 19), has been thrusted for tens of kilometers during Eocene-Lower Miocene on the Southeast Anatolian autochthonous, related to the southward transfer of Bitlis-Poturge-Malatya nappes. As related to this thrusting, The Çüngüş-Hakkari basin was closed in the east, in the area reaching down to iskenderun Bay, most possibly in the end of the Lower Miocene.

In northern Cyprus, the existence of the rocks similar to and synchronous with the Yeşilbarak nappe and the Qungus-Hakkari nappe (Baroz, 1979; Robertson and Woodcock, 1986; Hakyemez et al., 2000), and the emplacement of the nappes in the

Beşparmak mountains in Miocene implies the connection of the basins located in the western Taurus ranges and in the southeast Anatolia, and also their similar geodynamic evolution.

In the southern Aegean, the Hellenide nappes overlie the Upper Eocene-Oligocene flysch (Hall et al., 1984; Bonneau, 1984). The Ida zone (Bonneau, 1984) or the Plattenkalk series as defined by Hall et al. (1984) which is considered as autochthonous relative to the Hellenide nappe and shows similarities with the Lycian nappes and the Tripolitza nappe that (Bonneau. 1984, Hall et al., 1984) end up with Eocene-Oligocene flysch. It is not well known, however, if these flysch covers Lower Miocene or not. They show, at least partial, similarities with the rocks forming the Yeşilbarak nappe. During or after the deposition of these flysch, large scale emplacement (in Lower Miocene, Bonneau, 1984) of the Hellenide nappes in the south Aegean was observed as seen in the western Taurus ranges.

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ACTIVE SLUMPING OFFSHORE AMASRA (SOUTHWEST BLACK SEA) AND ITS RELATION WITH REGIONAL TECTONICS

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ABSTRACT.- The Bartın earthquake of 3rd September, 1968 provided the first evidence for active thrust faulting at the southern margin of the Black Sea. 500 km of 2D seismic reflection profiles acquired in this area show the presence of large scale slumps and landslips due to oversteepening of the shelf sediments, as a result of thrust faulting. The onshore drainage pattern is affected by interaction of the faulting with the geology, and this in turn affects the offshore shelf. Kilometre-scale bathymetric features on the sea-floor indicate a previous mass movement of a relatively coherent block of sediment, which has subsequently been partially buried. The source location for the block is visible on the geological map of the area. Incisional and erosional features in Pliocene sediments far from today's coastline and drainage systems suggest that water level was lower at the time they were formed than at present day.
Key words: The Black Sea, seismicity, seismic reflection study, slumping.

INTRODUCTION

The seismicity and tectonics of Turkey have long been interpreted in terms of the plate movements in between Africa, Arabia, Eurasia and Anatolia (McKenzie, 1972; Alptekin, 1973; Dewey, 1976; Şengör, 1979; Şengör and Canitez, 1982; Şengör et al., 1985). Most of the large earthquakes occur on two fault systems, the North Anatolian fault (NAF) and the East Anatolian fault (EAF), along which Turkey is moving westward. The focal mechanisms of these events are generally consistent with the relative plate motions in these locations. The focal mechanism of the earthquakes that are away from plate boundaries, however, are difficult to relate with the tectonic movements and this type of earthquakes occur on the Black Sea coast of Turkey.

In Anatolia, three major tectonic provinces have been recognized on the basis of their predominant structural styles and associated strain patterns (Şengör et al., 1980; 1983; 1985) (Fig. 1). They are 1) the East Anatolian high plateau, 2) the Central Anatolian ova regime, 3) the Aegean qraben system. There are three secondary neotectonic regions except for these major tectonic provinces: 1) Black Sea region north of the North Anatolian fault (The North Turkish Province), 2) Thrace, 3) Adana basin - Isparta angle.

The neotectonic regime of the areas lying north of the North Anatolian fault and comprising a major portion of the Pontide paleotectonic unit (Ketin, 1966) has not been studied in detail; largelv because of the neotectonic structures of the region are neither as active, nor as spectacular, nor as abundant as in other parts of Turkey.

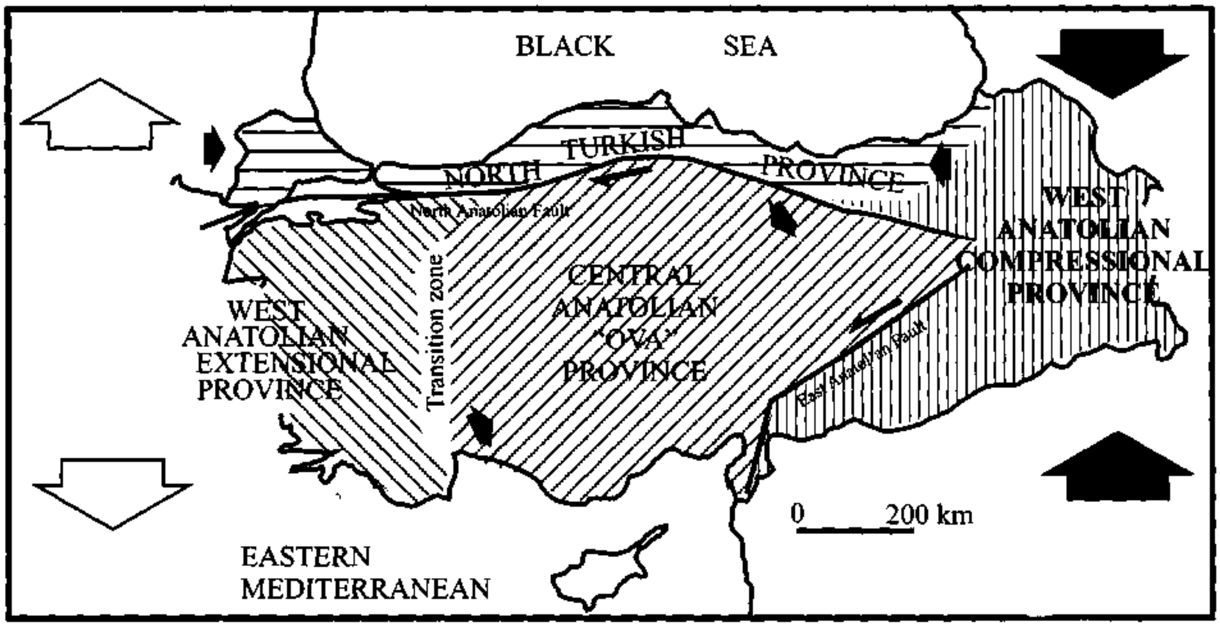

Fig. 1 - Neotectonic provinces of Turkey. Black arrows denote compressional areas while the white arrows denote extensional areas (after Şengör et al., 1985).

Indeed, the areas lying between the Black Sea shore and the North Anatolian fault have long been viewed as inactive. However, a study of the neotectonics of the North Turkish region is of critical importance for the solutions to a number of problems such as the interpretation of the weak regional seismicity and the understanding of numerous, discontinuous, scattered neotectonic features recognizable on land, on aerial photographs and in the field.

In September 1998, seismic reflection data were acquired in the southwestern Black Sea. The cruise was planned to image a possible scarp or fault which might have been associated with repeated slip on the Bartın fault. It is also aimed that, the data acquired will be useful to understand the regional tectonics as well as the new findings on the above-mentioned fault.

MORPHOLOGY AND SEISMICITY OF THE BLACK SEA

The Black Sea is a large (423 000 km^3), deep (2000 m) semi-isolated marine system situated to the north of Turkey, between two Alpine mountain ranges, the Caucasus and the Pontides (Fig. 2). It is thought to be a remnant of the Tethys oceanic system, which existed between Eurasia and Anatolia. It is linked to the Mediterranean by the Strait of Istanbul (the Bosphorus), the Sea of Marmara and the Strait of Çanakkale (the Dardanelles), and contains water of below normal marine salinity, a result of freshwater input from major rivers such as the Dnieper and the Danube. Extreme changes in the sedimentation regimes of the Black Sea occur as a result of eustatic changes in sea level. The stratigraphic history of the Black Sea is stil a

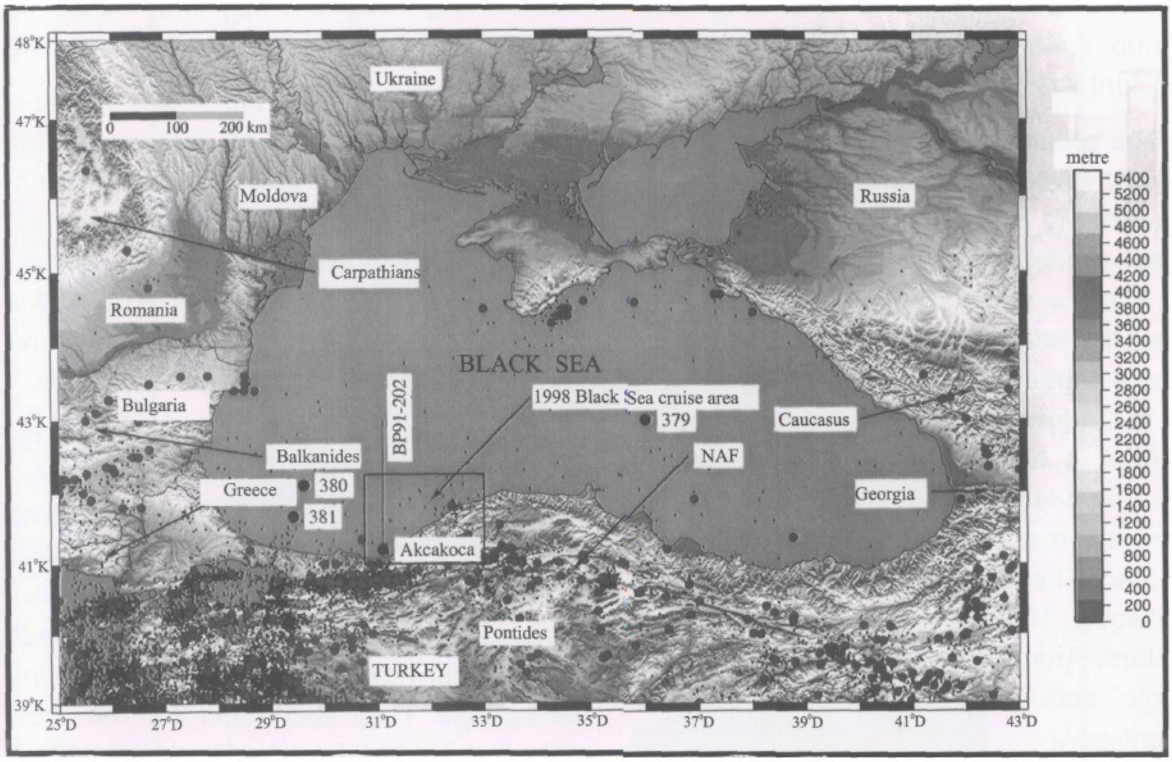

Fig. 2 - Map showing the seismicity in and around the Black Sea in between 1900-2000 (from NEIC catalogues). Small black dots are for epicenters Mw>3.0. Dairk gray circles are for Mw>5.0. DSDP wells (379, 380, 381) and Akgakoca-1 are the major well drilled in the area. KAF denotes the North Anatolian fault. The study area is shown in rectangle.

subject of debate due to the effects of complex extensional and compressional tectonic regimes (Letouzey et al., 1977; Okay et al., 1994; Spadini et al., 1996). Although the Black Sea is an intra-continental sea, it has geological features of a small ocean.

From morphological point of view, there are two types of margins in the Black Sea. There are no wide shelves on the eastern and southern Black Sea, the sea suddenly deepends after a narrow shelf. On the other hand, there are wide shelves on the north and west, following the basin slope, abyssal plain is reached. These features, from only morphological point of view, resemble the Pacific and Atlantic type continental margins (Erinç, 1984).

The development and evolution of the shelves are related to the rivers and their material input. Along the Black Sea coast, the sedimentation rate is not high, this is due to a) the presence of rather small but highly incising rivers, b) high topographic relief, c) lack of estuaries to preserve the deposits, d) the presence of narrow shelves, and e) the presence of canyons along which the sediments were transported to the deeper sections of the basin. There is a rather flat morphology in between İğneada and Ereğli. On Thrace peninsula, there are small rivers carrying the material from

Istranca Massif The most significant river in this region is Sakarya. East of Ereğli, up to Sinop, the shore is steeper and the most significant river in this area is Filyos.

The seismic activity in the Black Sea is relatively weak. In the central part of the sea the seismicity is negligible, however, on the shores moderate earthquakes were recorded. There are two important seismic belts around the Black Sea, these are situated in northern Turkey (the North Anatolian fault) and in Caucasus region. The North Anatolian fault is an east-west trending, highly active, right-lateral strike-slip fault. In Caucasus region active folding and thrusting is observed. The distribution of the epicenters in between 1900-2000 (Fig. 2), shows that the North Anatolian fault has a remarkable seismic activity in historical times and present-day.

There is some tectonic activity which is not located on the major controlling structures in the area, the most notable example being the shallow focus seismicity along the Turkish margin of the Black Sea. Eight moderate sized historical earthquakes have been reported in the area (Soysal et al., 1981). The Bartın earthquake of 3rd September, 1968 (Ms=6.6) provided the first seismological evidence that active thrusting is occurring at the southern margin of the Black Sea.

THE BARTIN EARTHQUAKE OF 3rd SEPTEMBER, 1968

On 3rd September 1968, at 10h 20min 36 sec GMT, the two small townships of Amasra and Bartın were shaken by an earthquake of magnitude 6.6. The earthquake has caused casualty (official statistics: 24 dead and hudreds of wounded) and heavy damage in these towns and the surrounding villages. In Ankara, Bursa, Istanbul and Samsun the earthquake was felt.

The epicentre of the main shock is located 10 km north of Amasra, in the Black Sea by ISC (International Seismological Center) (Fig. 3a). ISC recorded nine aftershocks following the main shock, magnitudes varying between $4.0<m_b<4.6$ in the area, five of these in the same day and the remaining in the following four months.

Ketin and Abdüsselamoğlu (1969), reported that the epicenter was around Akpınar and Kirlik (Fig. 3b) although no fresh fault breaks were observed in the field following the earthquake. The researchers concluded that the rather younger fault family must have been responsible for the earthquake. They also noted that the. coast near Amasra had been uplifted some 35-40 cm during the earthquake as a proof the movement of these faults, as well as some observed cracks on the alluviums and landslides near the epicentral area. Landers (1969) reported a tsunami in the Amasra Bay just after the earthquake.

A number of isoseismal maps were prepared for the Bartın earthquake (Albers and Kalafatçıoğlu, 1969; Ketin and Abdüsselamoglu, 1969; Ergünay and Tabban, 1983). The maximum intensity on these maps, despite some small variations, is VIII MMS.

McKenzie (1972), Kudo (1983), Şengör et al. (1983) and Jackson and McKenzie (1984) provided fault plane solutions for the earthquake. Except for Şengör et al. (1983), the researchers indicated strike-slip faulting with reverse fault component for the earthquake. Şengör et al. (1983), using the

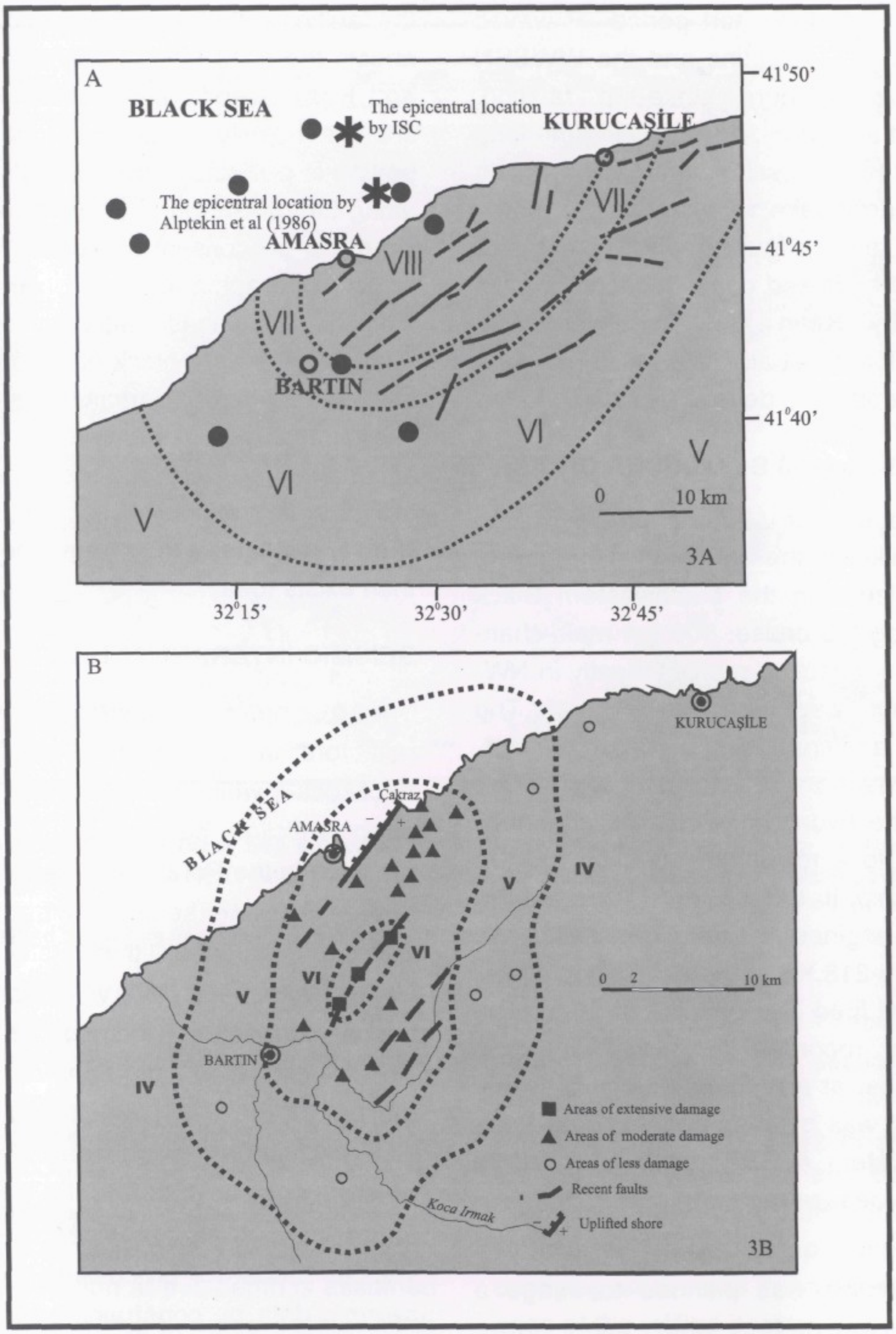

Fig. 3 - A) The epicenters proposed for the Bartın earthquake. Isoseismals are shown by broken lines. The stars denote the epicenters proposed by ISC and Alptekin et al. (1986). Black circles show the epicenters of the aftershocks of Alptekin et al. (1986).

B) Seismotectonic map of the epicentral region of the Bartın earthquake (after Ketin and Abdüsselamoğlu, 1969). Contours are isoseismal lines enclosing areas of varying intensities according to the Mercalli-Sieberg scale (Roman numerals indicate intensities).

first motion of the short period P waves reported in ISC bulletins and the WWSSN data, indicated pure strike-slip faulting. Alptekin et al. (1986), re-examining the long and short period seismograms, proposed that the earthquake was due to the movement of a reverse faulting which is compatible with the raised coast observations in Amasra by Ketin and Abdüsselamoğlu (1969). Alptekin et al. (1986) also has relocated the epicentre defined by ISC.

SEPTEMBER 1998 BLACK SEA CRUISE

In September 1998, a cruise using MTA Sismik-1 in the area shown in figure 2 was realized in the southwestern Black Sea. During the cruise, 500 km multi-channel seismic reflection profiles mostly in NW-SE direction were acquired (Fig. 4). The data were acquired using a 10-gün, 23 litre, tuned generator-injector compressed air source. The hydrophone had 96 channels, extending to a maximum ofset of 1500 m from the ship. In order to reduce noise from the ship's engines, the data were bandpass filtered at 8-218 Hz at the acquisition stage. Shots were fired at a spacing of 50 m. The data were recorded to a two-way time (TWT) of 5s, at a sample interval of 4ms. Navigation was by Differential Global Positioning System (DGPS), with an onshore VHF radio beacon, allowing a location accuracy of 5 m.

The cruise was planned to image a possible scarp or fault which might have been associated with repeated slip on the Bartın fault. The data show evidence for large scale slumping and landslip due to oversteepening of the continental shelf, as a result of thrust faulting. The onshore drainage pattern is affected by the interaction of the faulting with the geology, and this in turn affects the shelf offshore. Where the Filyos and Koca rivers enter the Black Sea, the shelf is heavily incised and no overstee pening is possible. Several kilometers east, along the coast, the shelf oversteepens and slumping processes can occur. Kilometre-scale topographic features on the sea-floor indicate a previous mass movement of a relatively coherent block of sediment, which has since been partially buried. Other incisional and erosional features in Pliocene sediments far from today's coastline and drainage systems suggest that the water level at the time of their formation was lower than exists today.

SEISMIC INTERPRETATION

An important constraint on dating reflectors in the seismic profiles is the Akcakoca-1 well. This well drilled on a small ramp anticline caused by Tertiary compression in the Pontides. It penetrated Quaternary-Pliocene sediments above a thick Eocene sequence, thin Paleocene and Upper Cretaceous (UC) volcanics and tuffaceous carbonates (Robinson et al., 1995). The Akgakoca-1 stratigraphy can be tied to the profile Bla98-04, which helps to correlate stratigraphic sequences between the offshore seismic data and the onshore geological units. It also provides a good initial velocity model for processing the reflection seismic data by constraining the depths of strong reflectors. Nonetheless, it lies close to the shore, in a zone affected by Tertiary compression and submarine to subaerial erosion, both of which hinder extrapolation of stratigraphy away from the wells into the Black Sea basin centre.

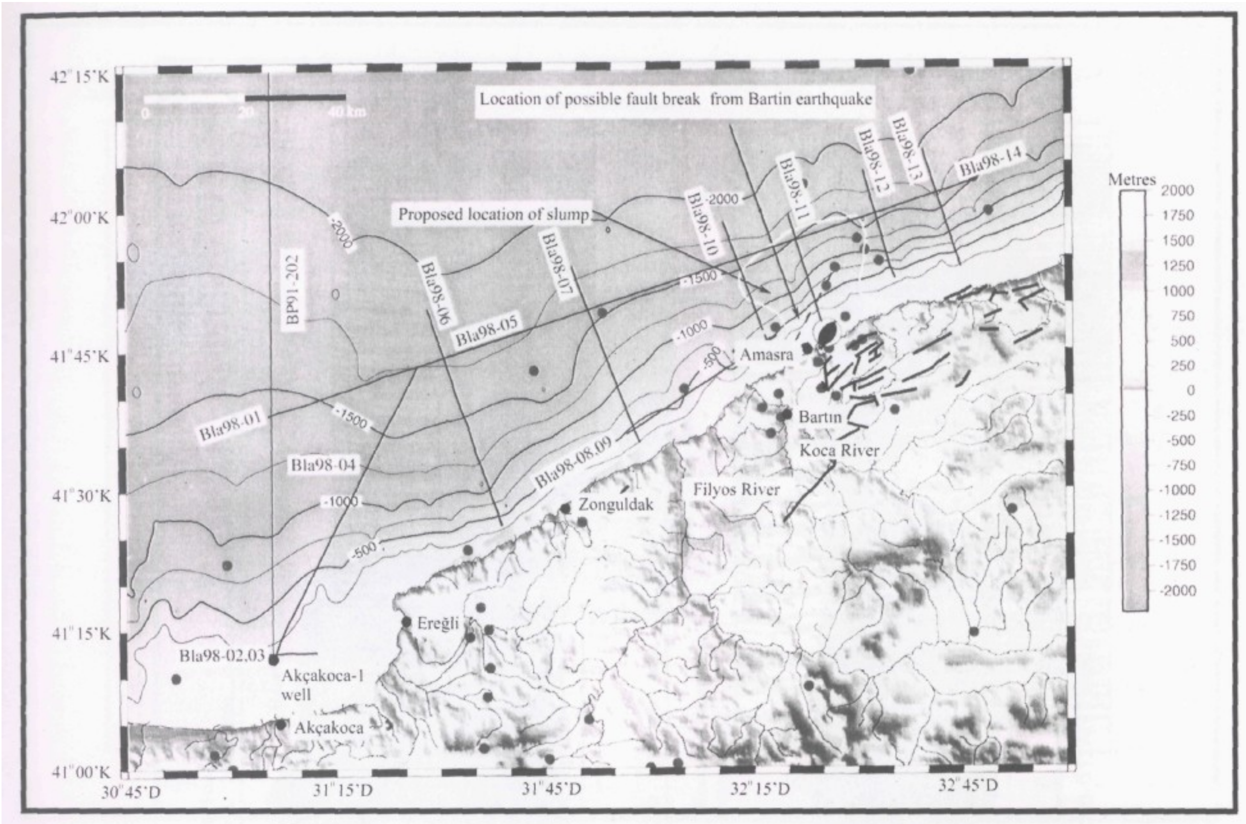

Fig. 4 - Map of the 1998 Black Sea survey area. Thin black lines show the reflection profiles. Earthquake epicenters from the NEIC catalogue are plotted as black dots. The Bartın fault plane solution is plotted, along with the theoretical location of a fault break (dashed line). The location of the proposed slump is marked by a white line. Onshore faulting is taken from Alptekin et al. (1986).

The well data are presented in fig. 5, along with the relevant time section of the seismic profile through the borehole, a depth migrated section and a plot of the interval velocities used to stack the data. The thin Paleocene horizon is not visible in the section at the location of the well, and the main reflection at this depth is assumed to be from the UC carbonates. Further north along the section, where a unit appears between the base of the Eocene and the top of the Cretaceous, it is interpreted as a thickening Paleocene unit.

Line Bla98-04 ties the Akçakoca-1 well to the main regional seismic profile (Bla98-01) (Figs. 4 and 6). The Cretaceous basement is visible in the section, and has undergone extensional faulting with evidence of syn- and post-rift deposition in the overlying Paleocene and Eocene units. The section between CDP 8200 and 9400 exhibits a classic onlap pattern with the overlying sediments terminating along the edge of the infilled basin.

The deepest identifiable reflection in this dataset is interpreted as the top of the Upper Cretaceous. This geological unit is also widely observed outcropping onshore, and is discussed later. At the southern end of the seismic section the UC reflection dips

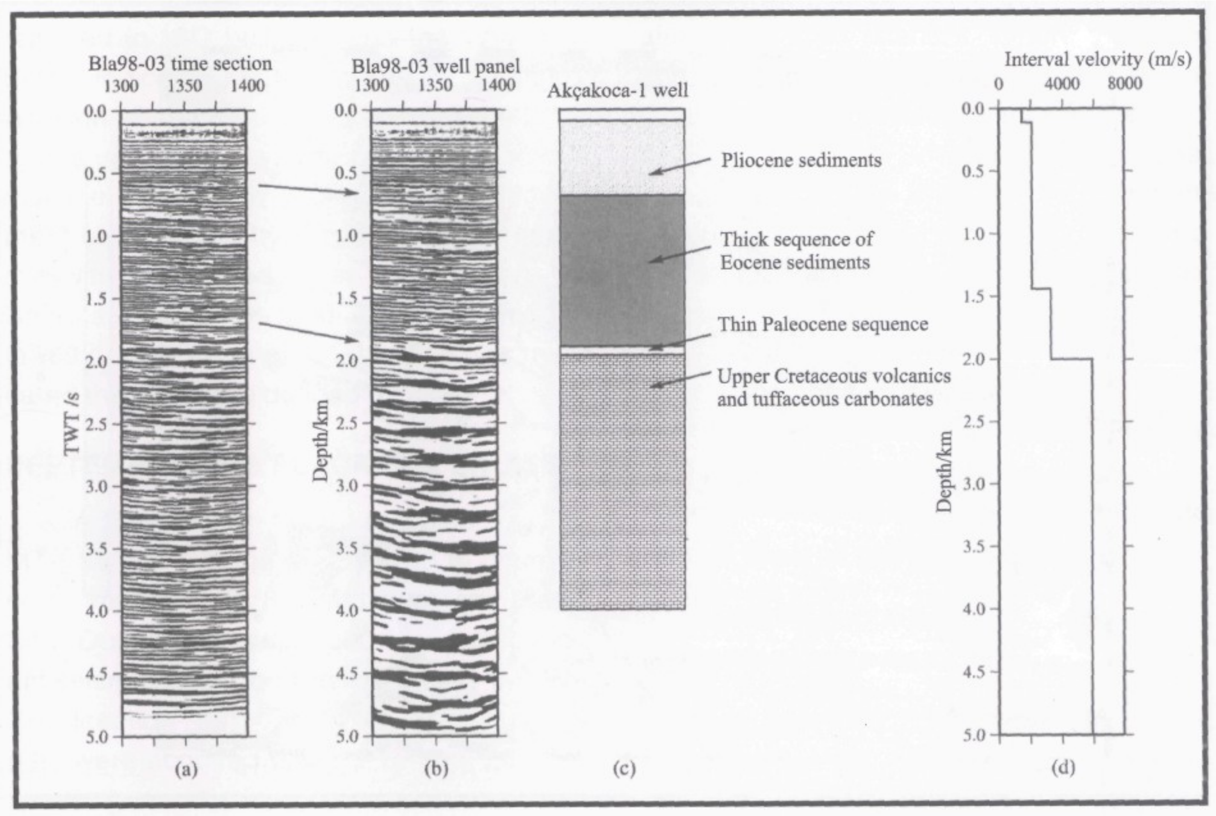

Fig. 5 - The Akgakoca-1 well. Shown here are a panel of: (a) Time-migrated reflection seismic profile data through the Akgakoca-1 well, (b) Depth conversion, (c) A sketch of the stratigraphy in the well, and (d) The interval velocities used to stack and depth convert the data.

down to the north, away from the coast. The Akgakoca-1 well was drilled into a small anticline in a compressional area in the shelf, and the northern limb of this anticline is observed from CDP 1200-2400 in the section. Further north, the UC horizon becomes flat lying, and enters an extensional regime. The northernmost end of the line shows a buried normal fault block which is also visible on the regional survey of the area (Robinson et al., 1995).

The sediments overlying the UC basement have had a complicated history. The well enables identification of the base of the Eocene and Pliocene sequences, neither of which is conformable with their respective underlying units. A thickening unit is observed further north between the UC and the base of the Eocene, and is interpreted as the Palaeocene basin fill at the northern end of the line. The top of the Eocene terminates against the UC at CDP 9200, in a characteristic onlap style, so that when the intersection with the regional line (Bla98-01) is reached, there remains only Pliocene or post- Pliocene deposits, unconformably overlying the Cretaceous basement. There are no Oligocene or Miocene horizons at the well site. The DSDP wells (Fig. 2) further offshore show that the region was subaerial during the late Miocene (Stoffers et al., 1978). It seems likely there-

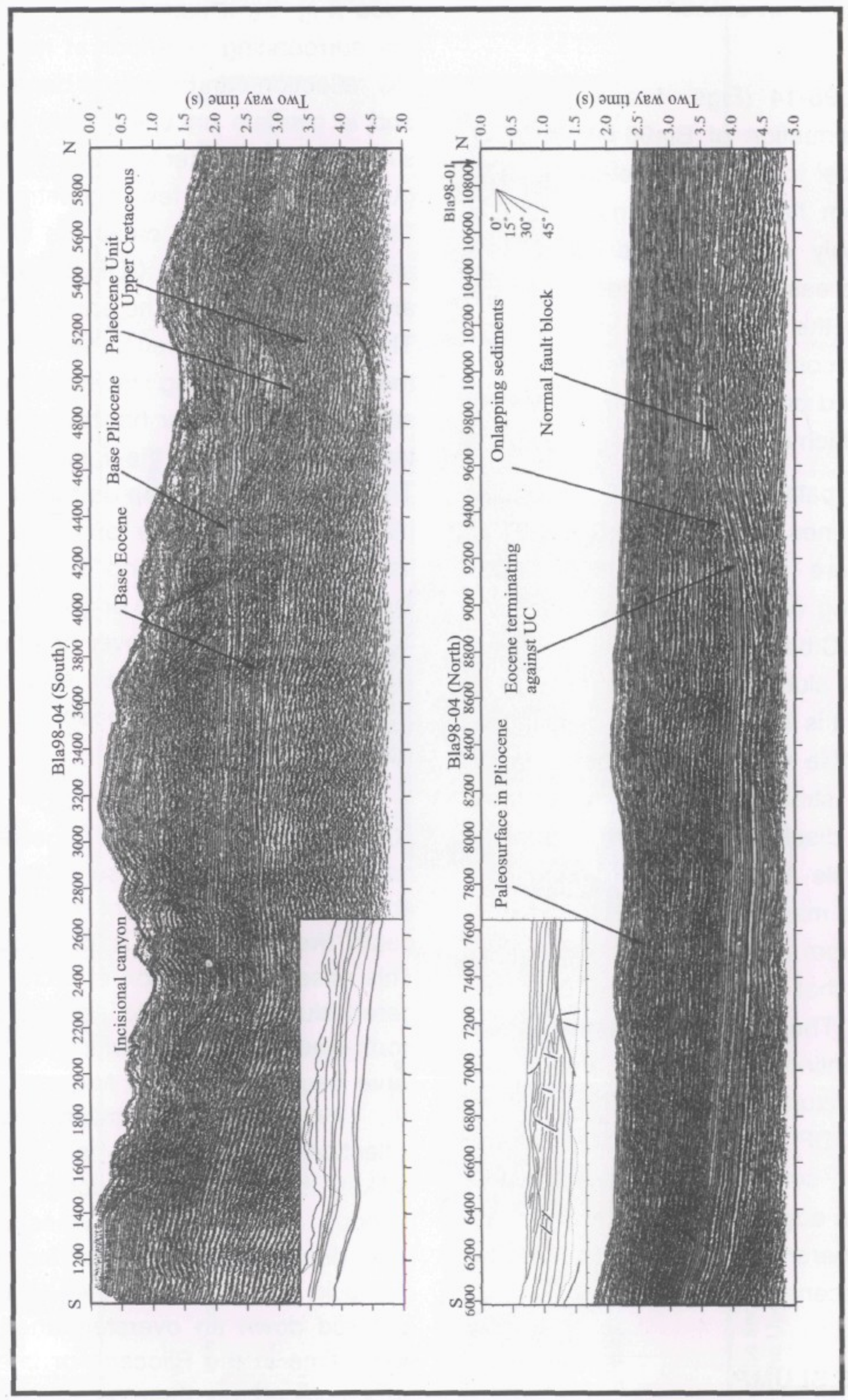

Fig. 6 - Line Bla98-04, tying the Akçakoca-1 well to the regional line Bla98-01. An incised canyon is marked between CDP 1400-3000.

fore that any Oligocene-Early Miocene deposits have been eroded from the top of the anticline.

Line Bla98-14 (Figs. 4 and 7) is an easterly continuation of Bla98-01, approximately parallel to the coastline. It shows a transition from flat lying sediments in the west to highly disturbed sediments and slumps in the east. The UC reflector can be traced along the entire length of this line, giving good control on the depth of the basement, and providing a useful tie to the other lines which cross this one.

The palaeo-surface previously identified on lines Bla98-04 and Bla98-01 is continued here from CDP 19800-14800. The underlying unit has a similar reflective style to the UC basement in the area. Given the nature of slumps and mass movement of sediment, it is likely that some units in the reflection profile will not be in their correct place in the stratigraphic column, as they have been displaced from their original location. While the material forming the paleo-surface may be displaced Cretaceous carbonates from a landslide, we know from the well tie that it is resting on Pliocene sediments. The relatively undeformed basement continues to the east, underlying different structural features. The section between CDP 15000-10000 shows examples of small kilometre-scale listric faulting at the edges of incised canyons, as well as coherent slumped and rotated blocks of Pliocene sediment.

A PLIOCENE SLUMP

The most noticeable bathymetric feature visible on profile Bla98-14 is the large hill, some 19 km wide, between CDP 8000 and 5000 (Fig. 7). It rises almost 750 m above the surrounding sea-floor at its peak. The UC reflection can be traced beneath the hill and is overlain unconformably by deposits similar in character to the Pliocene seguences visible a few kilometres laterally. These deposits first onlap the sides of the small basin, between CDP 6800 and 6000, and then drape over the entire UC reflector. The large hill has a reflective style similar to that of the underlying limestone, though it is sitting on a particular horizon which can be traced laterally into the Pliocene deposits. The deposits also onlap up the flanks of the hill, on both sides. The top of the hill exhibits some canyoning, and it is thought that the canyons are due to ongoing incision, as no obvious sediment cover has been deposited since this unit was emplaced. The flanks, however, have been buried under some 300 m of sediment.

This feature, between CDP 8000 and 5000, is interpreted as a largely coherent, massive slump. The base of the slump is above the last coherent reflector underneath the pile, which is of Pliocene age. This raises the question of technically triggered slumping, due to seismic activity in north-western Turkey, which is thought to have begun in the late Miocene (Görür et al., 1995). From its similarity with dated reflection packages elsewhere in the data, it is likely that the slump is composed of Cretaceous limestone. This limestone would have decoupled from the underlying Jurassic rock, somewhere further to the south. and slid down an oversteepened shelf, at some time in the Pliocene or later, ending up emplaced on then recent sedimentary deposits. It has since been partly buried by up to 300 m of younger sediments.

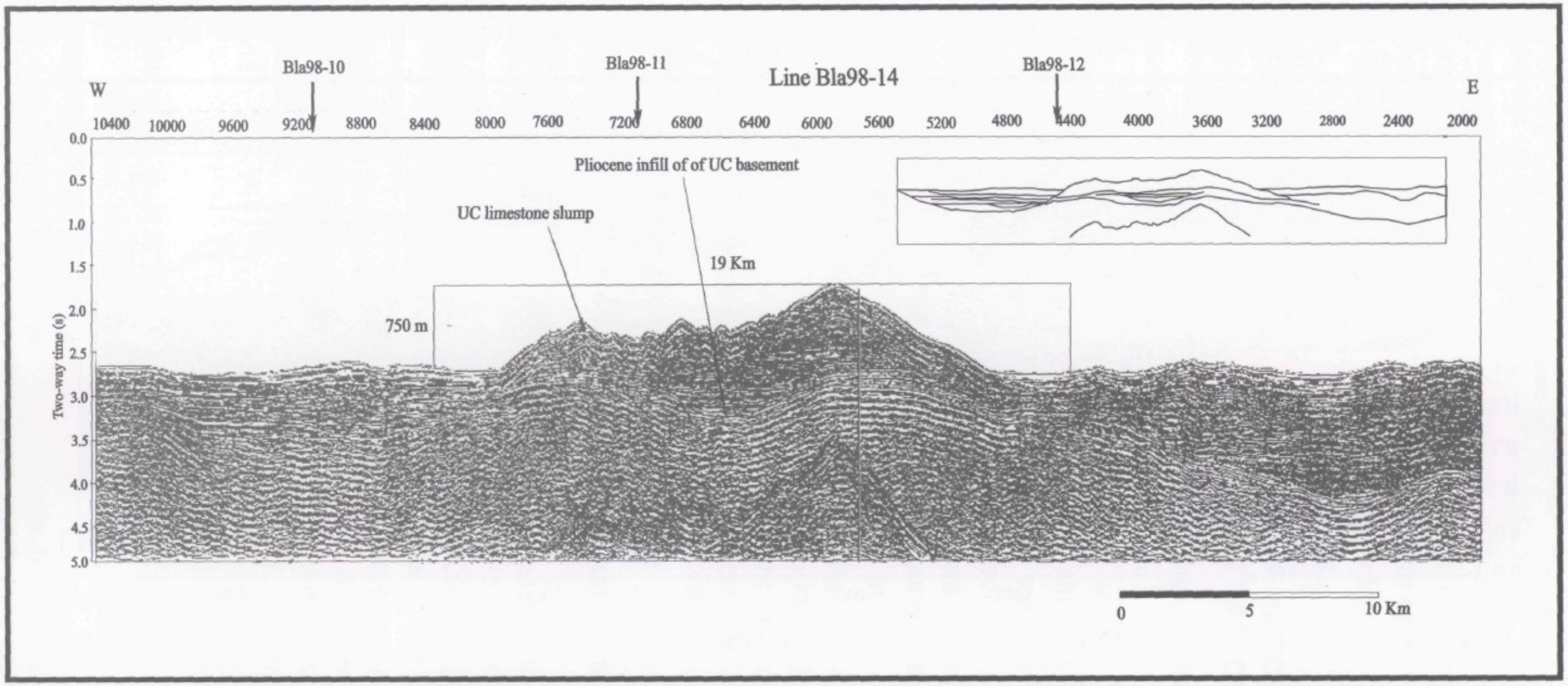

Fig. 7 - The section of Line Bla98-14 relevant to the slumping. The dotted lines on the seismic profile mark the boundaries of the slump. The Upper Cretaceous (UC) basement reflector is visible across the entire length of this line.

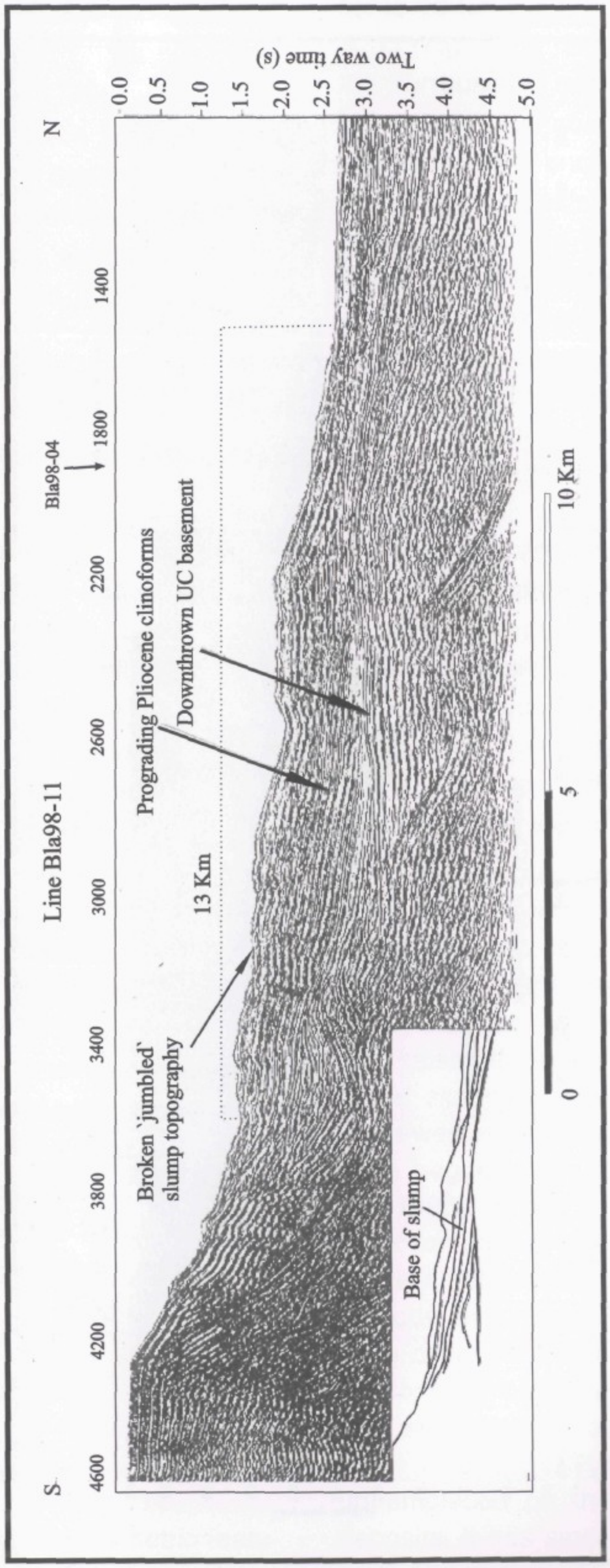

Fig. 8- Line Bla98 -11, shot over the theoretical location of the surface break of the bartın earthquake. Although the fault plane is not visible. the slumping and the over are cut by the profile.

Bla98-11 (Figs. 4 and 8) intersects Bla98-14 at CDP 7100, on the proposed slump. The UC basement on this line dips down to the south, between CDP 3200 and 2200. The unconformable overlying sediments dip down to the north. Both of these units fit with the observed stratigraphy interpreted in figure 7. The basement shows evidence of extensional faulting.

This reflection profile also exhibits all the classic characteristics of a slump caused by oversteepening of the shelf sediments. The jumbled, blocky nature of the topography starts at CDP 3800 and extends out to CDP 1800, giving a good indication of the thickness and offshore extent of the feature. Again, the toe of the slump extends under the sea-floor, where it has been partially buried by more recent sediments. Underneath the slump structure, Pliocene sediments are observed, terminating against the basement. The Pliocene deposits form a series of dipping, prograding clinoforms which downlap onto the basement.

Two other perpendicular lines can be used to delineate the lateral extent of the slump. Bla98-10 (Figs. 4 and 9) lies 12 km to the west, but has a significantly shallower slope. Between CDP 2000 and 3400 is the highly disturbed, blocky region associated with the slump. The 'blocks' in the slump are about 1 km in size. The basement reflector is also less deformed here, but is still unconformably overlain by the Pliocene deposits.

Sixteen kilometres to the east is Bla98-12 (Figs. 4 and 10). This line shows evidence of oversteepening, and has the steepest slope in the data, at -38°, probably related to the seismicity in the area. The UC basement has been downthrown by extension at the basin margin, but is folded in places, which may be the result of the current compressional tectonics. There is no evidence of slump related topography. The toe of the slope is at CDP 3000, and as before, is onlapped from the north by the basin filling sediments.

SLUMP KINEMATICS AND THE AFFECTING FACTORS

Using the constraints provided by the seismic data, and the bathymetry, it is possible to define the perimeter of the slump, and thus to estimate its area and volume. The slump area is marked in Fig. 9. It covers a surface area of 400 km^2. The slump reaches a maximum height of 750 m above the surrounding sea-floor, from the data visible in the sections, although appears to extend some 250 m beneath the current sea-floor, giving it a total topographic height of about 1 km. Estimating an average thickness of 500 m across the whole feature gives a total slump volume of 200 km^3.

The volume, though significant, is low compared with the largest reported submarine slide, the Agulhas slide off southern Africa. This has an estimated volume of 20000 km^3 (Dingle, 1977). Slumps on the flanks of the Hawaiian islands have recorded volumes that exceed 5000 km^3, as does the Storegga slide off the continental slope of the Norway (Hampton and Lee, 1996). The area covered by the Black Sea slump is not large, particularly when compared within other slumps such as the 1080 km^2 coverage of a slump from the Malaspina Glacier, in Alaska (Carlson, 1978).

Data from current slumps and landslides show that the ratio of their run-out

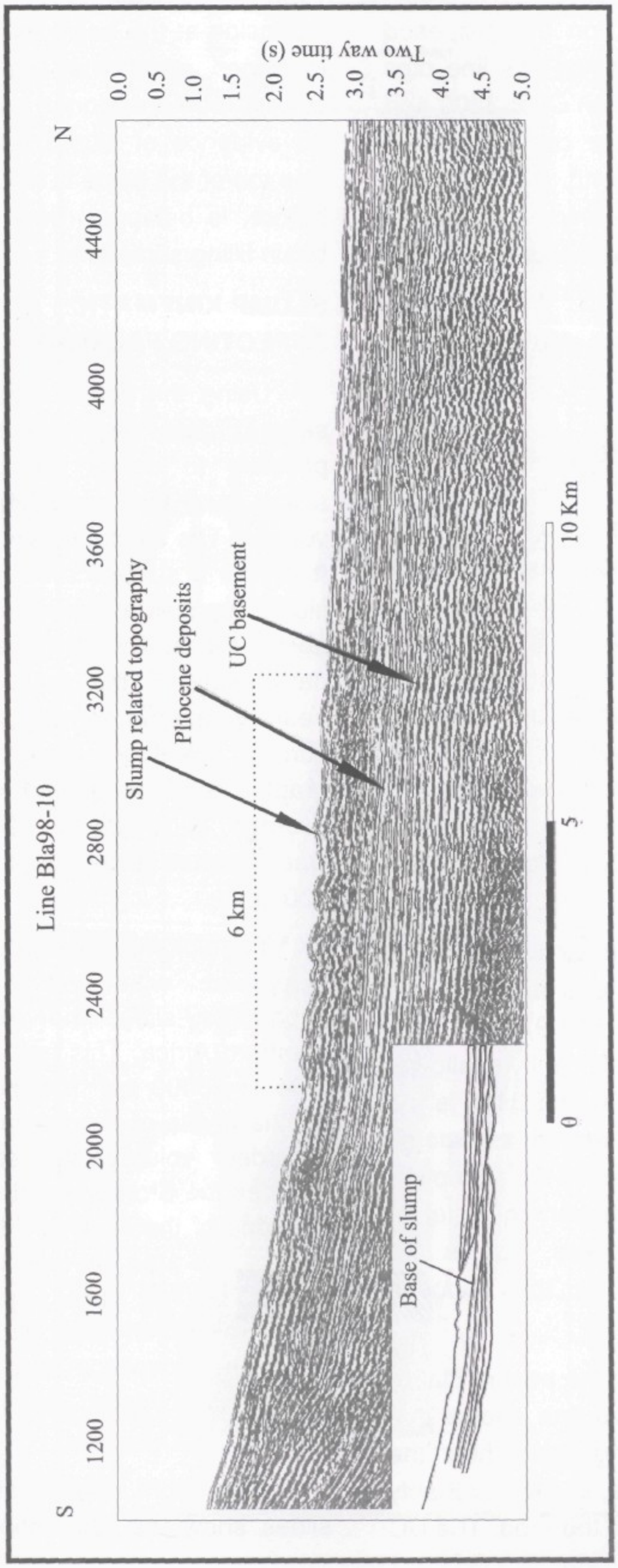

Fig. 9- Line Bla98 -10, shot 12 km west of Line Bla98-11. The characteristic topography related to slumping is visible on the line

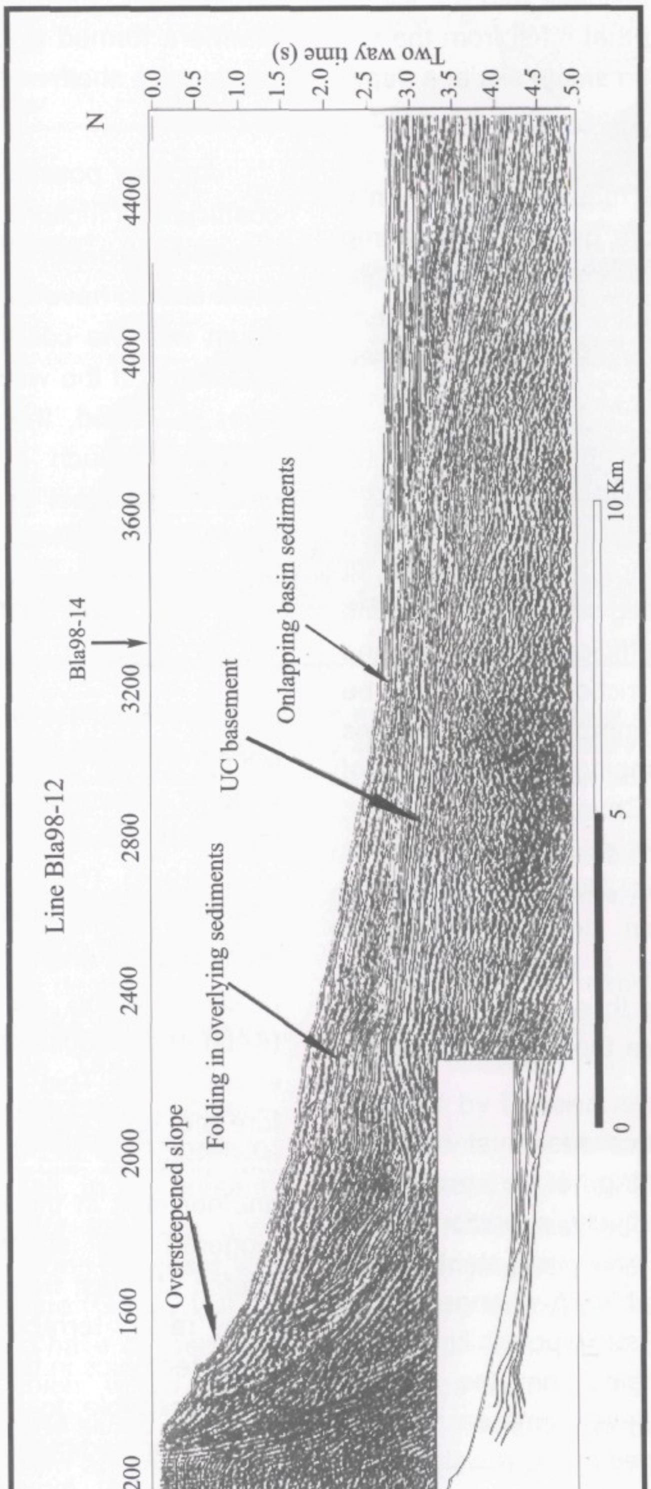

Fig. 10- Line Bla98 -12, shot 16 km east of Line Bla98-11. No slumping has occurred at this part of the coast, though the shelf is severely oversteepened.

length (L) to drop height (H) varies considerably and systematically with the slump mass (M). Assuming that it fell from the top of the shelf (the Bartın slump sits at a depth of 1500 m), then it could have travelled a distance between 20-60 km depending on the volume of the mass (250±50 km^3). Based on the onshore geology, this slump does not seem to have moved much more than 20 km, it appears to have had less than average mobility, based on observations of other slumps.

The above mentioned model based on a mass sliding down a slope is, however, too simplistic to explain all the phenomena associated with sumps and landslides. It does indicate that long runouts correspond to small effective coefficients of friction, and suggests that a low friction model might be a more appropriate approach. The reasons for initiation of slumping is a subject of ongoing research (Chuang and Greeley, 2000; Chamberlain et al., 2001). The force of gravity alone is not likely to be the sole cause of failure on continental margin slopes (Ross, 1971). If it were, then the large slumps such as those observed would not be possible, since the sediments would fail much earlier.

Some other processes must occur to enable oversteepening of a previously stable slope. Alternatively, conditions must change such that a previously stable slope then becomes unstable. A change in the water depth is one such possibility. It is known that depending on the drastic change in the water level some parameters (i.e. density of the sediment, resistance to failure, pore pressure of sediment) may change. Most authors agree there was a substantial marine flooding approximately 8000 years ago, when a connection to the Aegean via the Bosphorus and Sea of Marmara formed and over 100 000 km^3 of continental shelf was submerged (Ryan et al., 1997).

Another possibility is a change in the coefficient of friction. A layer of sediments at a particular depth in a submarine environment should have a pore pressure in equilibrium with the confining water pressure at that depth. If the water level drops, and the layer is sealed, then it will become overpressured. Such a layer could become unstable and act as a low-friction horizon upon which slip could occur. Some previously observed submarine slumps have been associated with rapid changes in sea level (Hampton and Lee, 1996).

The most likely trigger for this slump, once the sediments had become unstable is tectonic movement in the area. Such effects have been observed elsewhere by Hazlett et al. (1991). The Bartın earthquake of 1968 occurred directly between this slump and the coastline, and an uplift of 30-40 cm was recorded along the coast near Amasra (Ketin and Abdüsselamoğlu, 1969; Alptekin et al., 1986). This slump occurred well before the 1968 earthquake, but it is possible that previous slip on the same fault, or another fault in the same area could have triggered the submarine landslide. The stretch of coast from Ereğli to Amasra has many raised terraces, probably due to the thrust tectonics in the area, and is known to be susceptible to landslides. The sparse post-siump sedimentary drape on the slump suggests that movement occurred in the Pliocene to holocene, which is consistent with the tectonic activity in the area today.

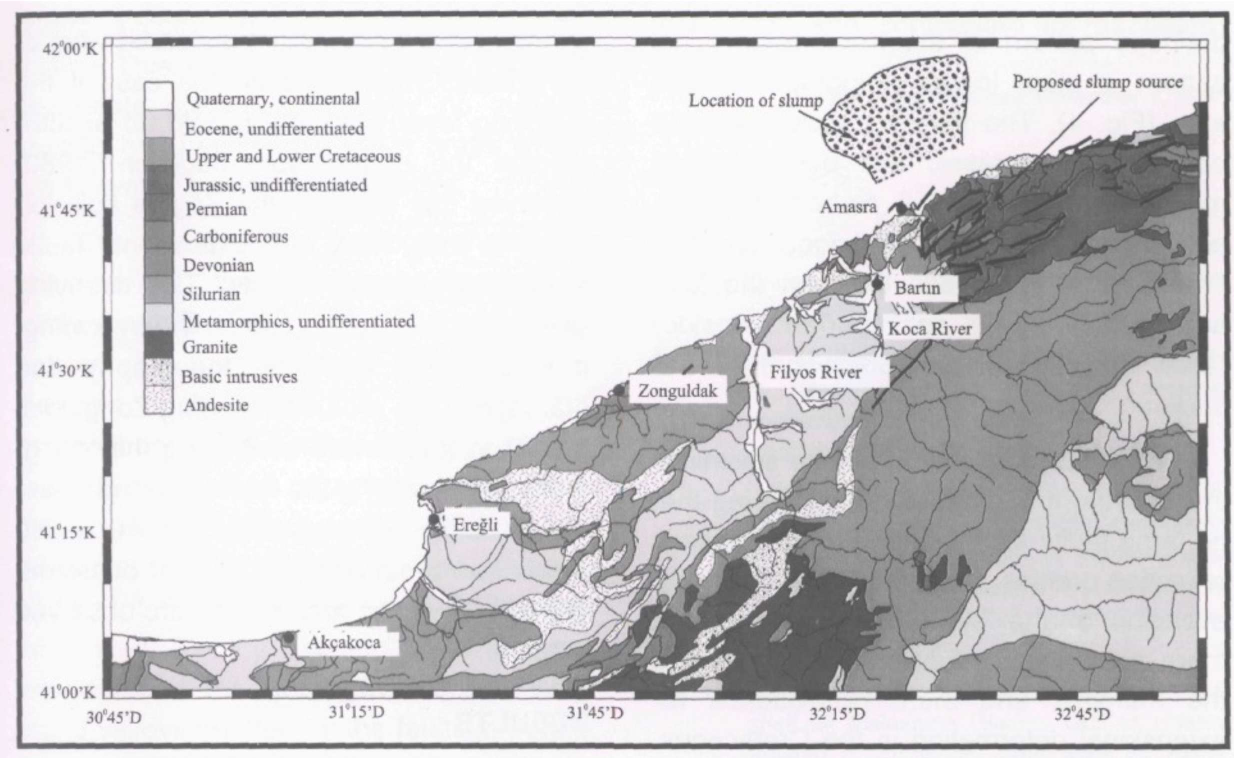

Fig. 11-1/500.000 scale geological map of the Zonguldak area (Tokay, 1964). Most of the area is covered by Cretaceous and younger sediments except for the area east of Bartın where underlying Permian and Jurassic rocks are exposed: this is assumed to be the source of the offshore slump.

DISCUSSIONS

The reflection profile evidence suggests that at some time during the Pliocene, a large volume of material destabilised and slumped to a position some 20 km away from the present coastline. The trigger for the event was probably tectonic, but consideration must also be given to the reason for the oversteepened shelf. In the present day, there is evidence for thrusting and folding in both directions along the coast, as well as uplift of the coastline (Görür, 1997). Folding and faulting could have caused the oversteepened shelf, which would have subsequently collapsed. The slump material is probably Cretaceous from an analysis of the reflection seismic data. This means that in the source area, the underlying Jurassic rocks would have been exposed by its removal. Examination of the onshore geology shows a likely source location for the slump (Fig. 11).

The dominant exposure in the Zonguldak area is of Upper Cretaceous carbonates, and volcanics (Fig. 11). This is overlain in places by Eocene and Quaternary deposits, as well as having been intruded by later igneous material. To the east of Amasra the Cretaceous has been removed, and the underlying Jurassic and Permian rocks have been exposed. This exposure covers an area of some 1000 km^2, and is affected by surface faulting. The faults have a trend which matches that of the instrumental fault plane solutions, and are parallel to the current coastline. The different lithology means that faults in this area are able to break to the surface, and are controlling the drainage pattern of the Koca River. This pattern is clear in the topography of the area (Fig. 4). The Permian rocks exposed beneath the Jurassic are being eroded much faster than is the carbonate cover, and show up the faults in good contrast. This type of interaction between the fault and the lithology is an important consideration in active areas (Goldsworthy and Jackson, 2000).

With the high level of diffuse seismicity in this area, it is at first sight surprising that no large faults are observed on the seismic reflection profiles. It seems likely, given the evolution of the western Black Sea, that there should be extensional faults around the margins, and there is evidence for extensional deformation in the Cretaceous. The later thrusting associated with the closure of the Tethys Ocean, and continuing today, could be accommodated on reactivated normal faults. If this is the case, then since most of the faults on the southern margin of the Black Sea are north dipping, and related to previous extension, we might not see any large fault breaks on the reflection profiles as they would be too close to the shore to be covered by the acquisition. Another possibility is that whichever way the fault plane is dipping, it may not have broken along a plane to the sea-floor. On land, when thrusts reach the surface, they diverge from the fault plane dip at depth, and break closer to the vertical. The nose of the hanging wall often breaks off and masks the true location of the fault, causing a blind thrust (Dunne and Ferrill, 1988).

Given that the seismicity is distributed throughout this region, it is likely that faults exist all along the coast, even though no surface breaks have been observed west of Zonguldak. This would be the case if the overlying layer was able to fold, to accommodate the deformation. Jackson (1980) observed this effect in the Zagros collision zone, in Iran. Here, old extensional faults are reactivated as thrusts. The. overlying sediments take up the shortening by folding. It is apparent from the topography and drainage in the area surrounding Zonguldak that blind thrusts are controlling the topography, and causing the folding. In this case, the thrusting is responsible for the uplift, which has produced the present coastline, and this is why no major faults are observed offshore.

RESULTS

A large recent slump, of estimated area 400 km^2 and volume 200 km^3, has moved material approximately 20 km offshore, east of Amasra. The slump is interpreted as being of Cretaceous material, and this analysis is consistent with an examination of the onshore geology. The aspect ratio and run-out distance of the slump are comparable with other features observed elsewhere, though the slump was less mobile and retained its internal structure better than have most other slumps: we suggest that this is due to its composition of competent Cretaceous limestones.

The pattern of onshore faulting, drainage and geology shows that the dominant recent tectonic deformation in the area is thrust faulting. These thrusts are responsible for oversteepening of the shelf, which then collapses, most likely because of a tectonic trigger. It is likely that some of the thrusting is being taken up on reactivated extensional faults, with the overlying sedimerits accommodating the shortening by folding. This explains the folding of Pliocene sediments offshore, and why no surface faulting has been observed in areas which are known to be seismically active.

The present day geology shows an absence of the dominant Cretaceous formation to the east of Bartın. The faulting pattern in the exposed Jurassic and Permian rocks highlights the important contribution which surface lithology makes towards recognition of active faulting. It is only in this area that surface faulting has been mapped on the ground. The drainage pattern of the rivers has been significantly influenced by the faulting. The eastern tributaries of the Koca River are running along valleys created by the faulting, before turning north and draining into the Black Sea. This means that for much of the coast to the east of Bartın, there is very little sedimentary input. When the major rivers enter the Black Sea, incision and canyoning are visible in the sea-floor relief of the reflection profiles, and no oversteepening has occurred.

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THE CONTROL OF THE FORCED - REGRESSION, TRANSGRESSION AND SEDIMENT SUPPLY ON THE SEDIMENTOLOGICAL AND SEQUENCE STRATIGRAPHICAL DEVELOPMENT OF THE BASIN MARGIN DEPOSITIONAL SYSTEMS; ERMENEK BASIN, MIDDLE TAURIDES

Ayhan ILGAR *

ABSTRACT.- The Ermenek basin located on the Bozkır and Aladağ nappes in the central Taurides, is filled with Early Miocene lacustrine elastics (Yenimahalle fm.) and Middle Miocene reefal shelf carbonates. This study concentrates on the facies analysis and sequence stratigraphic framework of clastic basin fill sediments on southern margin of the Ermenek basin. Sedriments deposited in alluvial fan, Gilbert-type delta, beach and shoreface environments are transitional laterally and vertically in the succession. These depositional systems are repeated in the vertical section. The control of the lake-level changes in relation with tectonics, climate and sediment supply are also considered besides sedimentary processes on the sedimentological and sequence stratigraphic development of the basin margin depositional systems. Sequence along fault-bounded southern margin of the basin consists of stream-dominated alluvial fan and delta plain deposits, mass-flow dominated-delta foreset deposits and high energy beach and shoreface deposits. Retrogradational or progradational stacking patterns of these facies associations indicate high frequency lake-level changes. Four unconformity-bounded sequences represented by forced regressive erosional surfaces have been identified within the Yenimahalle formation. Lowstand systems tracts of alluvial fan deposits overlie the sequence boundaries. Retrogradational stacking pattern of Gilbert-type delta, shoreface and beach deposits onlapping alluvial fan surface constitute transgressive systems tracts of 1. 2. and 3. sequences. Progradational Gilbert-type delta of sequence 4 represents hiahstand systems tract over lowstand systems tract. The sequences show paleogeographic changes that developed with lake-level changes during depositional evolution. The development of the facies, systems tract and seauences were controlled by tectonism, climate and sediment supply besides sedimentary processes. Tectonism resulted in the change of accommodation space (increasing or decreasing) by controlling basin subsidence or rise. The amount of water and sediment supply into basin was controlled by climate whereas the distribution and lateral variation of facies within the sequence is attributed to sediment supply and the source area.

MIDDLE - UPPER MIOCENE STRATIGRAPHY OF ÇANAKKALE, NW TURKEY

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ABSTRACT.- Middle-Upper Miocene terrigenous and marine sedimentary rock units deposited in Truva basin outcrop throughout the eastern shore of Dardanelles. These deposits overlie unconformably Paleozoic schists, marbles, quartzites, Permian-Triassic ophiolites and Eocene volcanic rocks between Çanakkale and Truva; only Eocene volcanics and volcanoclastic deposits in the vicinity of Lapseki, north of Çanakkale. The sedimentary rock units are composed of Middle Miocene Sarıyar formation and Upper Miocene Çanakkale formation. Sarıyar formation overlying metamorphic and magmatic basement units with an angular unconformity contains red-dark red colored alluvial deposits. Çanakkale formation containing marine deposits overlies Sarıyar formation with an unconformity. Çanakkale formation consists of Güzelyalı, Tekkedere and İntepe members that have distinctive lithological components, sedimentary features and depositional environments. These are transitive laterally and vertically to each other. Güzelyalı member that contains mainly fine to coarse sandstones, and lesser amounts of mudstones, siltstones and conqlomerates deposited on beach and shoreface. İntepe member is composed of mudstones and siltstones deposited and sandstones. Sandstones developed with the tidal processes have bidirectional cross-stratifications, planar and trough cross-stratification, flaser and lenticular beds. Tidal channel deposits that have erosional bases in lagoon mudstones are observed. These are made of medium to coarse sandstones, conglomerates and abundant broken shell fragments. Tekkedere member contains algal mat limestones, oolitic limestones being a product of shoals, and conglomerates and coarse sandstones beach. Upper Miocene Çanakkale formation is overlain by Pleistocene marine terraces and Pleistocene-Recent alluvial dreposits.

Key words: Middle-Upper Miocene, Stratigraphy, Çanakkale.

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SEDIMENTARY INFILL AND GEOLOGICAL EVOLUTION OF ÇAMELİ NEOGENE BASIN, DENİZLİ-SW TURKEY

Mehmet Cihat ALÇİÇEK, Nizamettin KAZANCI", Mehmet ÖZKUL and Şevket ŞEN"

ABSTRACT.- Çameli basin is one of the western Anatolian grabens formed during the neotectonic period. The basin contains data which may enlighten neotectonic stage of the region due to its setting and successions. Characteristics of the basin fill are determined by the fades analysis and kev data related to time interval of the deposition are provided by using mammalian faunas. NE-SW trended Çameli basin begins to open as *a* graben under the control of the Dirmil fault at the east and Bozdağ fault at the west during the Late Miocene (10.8-8.5 Ma). Growth faults seen intensively in the preliminary sediments of the basin composed of alluvial fan, river and lacustrine deposits point out an effective extension. After this period, the basin is divided into two parts by an intensive faulting at the Early-Middle Pliocene (3.8-3.2 Ma). Later, the impression of extensional tectonics relatively decrease and the basin turn into a large lacustrine environment. The lacustrine deposits overlap both fault that separate the basin into two parts, and marginal faults and this stage continues until Middle to Late Pliocene (3.5-2.5 Ma). While the lake basin become shallow by filling of alluvial fan and river delta progradation. lacustrine carbonates precipitate in the central part of the basin. After this stage the basin is broken again in Late Pliocene (2.6-1.8 Ma) by two fault systems that are parallel to marginal basin faults indicated with a travertine layer. The final deposits of the basin are alluvial sediments deposited by this faulting stage. According to the growth faults seen in the sediments accumulated after the latest faulting stage, the extension has been reactivated and the Çameli basin more or less has taken recent form.

Key words: Çameli, Neogene, graben infill, Neotectonics, facies analysis, southwest Turkey

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STABLE ISOTOPES (δ^{18}O, δ^{13}C) OF MOLLUSK SHELLS IN ENVIRONMENTAL INTERPRETATIONS; AN EXAMPLE FROM SİNOP MIOCENE SUCCESSION (NORTHERN TURKEY)

Baki VAROL

ABSTRACT.-.Stable isotope values of mollusk shells representing centain levels in Sarmatian and Tshokrakian units <ithin Miocene succession in Sinop region have been respectively measured as δ^{13} C= between -0.32 and 0.78 ‰, and δ^{18} O = between -2.18 and -2.38 ‰, δ^{13} C= between-2.61 and 0.79 ‰, and ‰, δ^{18} O= between -6.46 and -0.10 ‰. When these values have been compared with values of mollusk fauna from recent Black Sea (δ^{13} C=between - 0.99 and 0.39 ‰, and δ^{18}O = between -2.32 and -0.41 ‰) and with values of surface waters of Black Sea (δ^{13} C = between 0 and 1 ‰, and δ^{18} O= -2.84 ‰ (in average)), it is shown from the stable isotope analysis point of view that Tshokrakion Sea resembles to recent Black Sea and the Sarmatian Sea displays a wide range from normal sea water to brackish waters.

INTRODUCTION

Application of stable isotope (δ^{18}O and δ^{13}C) analysis to shell-froming calcite and aragonite of organisms is a long-lasting practice in investigation of temperature and salinity changes in ancient seas (Epstein and Lowenstom, 1953). The main principle of this application is to have records of isotope values of the water in which organisms lived is reflected by the isotopic values within the shells. However, this is not a simple phenomenon and a series of complex events can take place. Therefore, following points must be taken in account in interpretation and usage of data obtained in these studies. The origin of calcite or aragonite, either inorganic or organic, constituting the mineralogical composition of a shell may differ in its isotope composition. For example, it has been recorded that aragonite inorganically crystallized at 25°C displays an increase of ‰ 0.6 in its isotope content with respect to inorganic calcite. However, stable isotope values in biogenic aragonite can have more positive or more negative values with respect to biogenic calcite (Arthur et al. 1983). Grosman and Ku (1986) similarly stated that rate of increase in δ^{18}O values of aragonite in foraminifera and mollusk shells with respect to calcite was not related to the temperature of the water. Moreover, these authors investigated that difference in &13C values of biogenic aragonite and dissolved inorganic carbon increased with decreasing temperature and recommended the application of this relationship in studying the paleo-temperatures of ancient oceans.

Isotope changes controlled by climatic and hydrological cycles are among the subjects that had been frequently studied. Isotopic changes in ocean water are mainly controlled by natural events such as glaciation, rain and flooding. For example, effects of Tertiary continental glaciation are reflected by important changes in δ^{18}O isotope values of deep marine sediments of the same time-span (Shackleton and Kennett, 1975). Periods of large floods enrich the waters in nutrients causing low salinity. Increasing organic production by this way causes the formation of sapropels as seen in Mediterranean basin (Calvert, 1983; Abrojen et al. 2002). Effects of this formation are recorded by enrichment of heavy isotopes (^{13}C) in surface waters and carbonates, although organic rich bottom sediments display enrichment in light isotopes (^{12}C). Within this process, it is observed that surface waters have lower $d^{18}O$ values but higher $d^{13}C$ values (Abrojen et al. 2002).

The escape of the light 16O isotope from the system causes enrichment of ^{18}O heavy isotope in global or seasonal warming periods, periods of increased evaporation and drought (Craig et al. 1963). Therefore evaporitic beach facies are characterized by sediments with higher $d^{18}O$ values with respect to temperate belts (Arthur et al. 1983).

Diagenesis is one of the most important factors effecting isotope values in the study of depositional environments by isotopes. Especially, in the isotope studies on shells of organisms, secondary calcite which can be formed depending on the chemistry of pore fluid in every stage of diagenesis, can effect the isotope character of the primary calcite (shell) and lead to misinterpretation of the results or mask the values obtained (Geary et al. 1989). Therefore, shells of planktonic organisms in deep marine sediments exhibiting properties of closed diagenetic environment have been frequently preferred due to balance between sediment and pore fluid (Resales et al. 2001). These authors emphasize to have a control on factors such as the disintegration of organic matters by bacterial effects or transformation of volcanic particles to clay minerals in the diagenetic environment that may result in unexpected low values of stable isotopes.

The subjects mentioned above are among the most important factors that affect the isotopic character of the carbonate association or of the shells which have gained their isotopic properties from their living environments. Moreover, many factors such as the vital effect and ontogenetic factor can also be affective on the isotopic character of the shells. For example, considareble increase in heavy isotope (^{13}C) values of shell structure in juvenile stage having high rate of growth and metabolic activations is frequently recorded (Romanek and Grossman, 1989). Similarly, metabolic carbonate which can enter to the shell structure can cause enrichment of shell with respect to light isotope (^{12}C) and deplete the shell with respect to heavy isotope (^{13}C). In this point of view, organisms having an effective respiratory system (athropods, mollusks, annelids) do not reflect the changes mentioned above as they do not including metabolic CO_2 in their skeletons (Weber, 1968).

In this study, isotope values obtained from shells representing the Sarmatian and Tshokrakian units have been correlated with those of the recent Black Sea fauna and with isotepe values of marine water in which they lived. So, by establishing the similarities or differences between recent Black Sea and the Sarmatian and Tshokrakian seas with respect to isotope values, approaches for depositional environments have been made. Application of these types of studies are very limited. Paleo-temperature records belonging to Romanian - Chauvda stages of Dardanelle Straight (Taner, 1996), isotopic characters of sediments with planktonic foraminifers of Campanian-Maastrichtian age around Hekimhan (Yıldız and Özdemir, 1999) are the studies published yet. That the Miocene succession of Sinop had been studied relatively well in regard to sedimentology and paleontology has facilitated this present study. The Miocene succession of Sinop region has quite different depositional environments within the scope of sedimentological and paleontological definitions. According to betterknown details of Eastern Tethyan chronostratigraphy the Sarmatian and Tszhokrakian sediments are found to be very suitable for the scope of this study. In the previous scientific reports, it is stated that most of the Tshokrakian fauna display very similar properties close to the salinity values of recent Black Sea (Özsayar, 1977). However, it has been recorded that Sarmatian was deposited in a depositional environment having a broad variety of salinity changes, ranging from hyper saline to brackish waters and include fauna associations reflecting this conditions (G6-rur et al. 2000; Varol et al. 2001).

MATERIAL AND METHOD

This study is carried out to test applicability of isotope values obtained from shells of organisms in interpretation of depositional environments. Number of samples for isotope analysis is compulsorily minimized due to financial restriction. Therefore, the more definite expression of the results obtained will be performed by another study including a larger number of samples in the furture. In sampling, great care was given to obtain samples from only a single group of organisms (Mollusks) in the levels which are rich in shells of sedimentologically well defined units. Whole shells that can be easily extracted from the base rock and cleaned were preferred. The extracted shells are washed with distilled water in an ultrasound tank to remove clays, oxidized matter, etc, and checked under the binocular microscope. The shells within the limestone in upper part of the Sarmatian succession are extracted only in fragments due to extensive diagenesis. Within these shells, samples which have no effects of diagenetic changes and are avoid of basement rock fragments were chosen. For comparison of the isotope values and referencing of the samples, recent samples from Black Sea fauna are obtained and especially certain different genus has been chosen within the beach sands from outer seaport region. All isotope analysis were carried out in the laboratories of the department of geochemistry in Tubingen University. SMOW and PDB standards were measured together with all samples and PDB is used for our anaysis to obtained more sensitive results for paleosalinity and paleotemperature interpretations.

SEDIMENTOLOGY

Sedimentology and paleogeography of Miocene succession of Sinop region is reported in detail by Görür et al (2000). From bottom to top, some local unconformities with low angle are observed in this succession represented by Paratethyan stages of Tarkhanian, Tshokrakian, Karaganian and Sarmatian. In this study, sedimentology of Tshokrakian and Sarmatian sediments in the studied region is described. Generalized geological map and stratigraphic section of Sinop Penninsula are illustrated in figure 1 and 2.

Tshokrakian: The rock units of this age are best observed around Kurtkuyusu in the west of Sinop Penninsula. The thickness of the unit ranges between 20-50 m. It is characterized by dark colored mud stone, rich in pyrite and including carbonatized plant fragments at the bottom, yellow colored weakly cemented massive sandstone with wave ripples and some large-scale cross-bedding in the middle and at the top (Fig. 3a). Properties of semi-closed depositional environment dominated by anoxic conditions rich in organic matters at the bottom turn in to high-energy environments with deposition of high amount of sand in middle-upper portions. Large -scale cross-bedden sandstones representing a large portion of the Tshokrakian succession display high energy sand accretion. Some preserved ripples within the succession reflect the same depositional development. The shell samples for isotope analysis are recovered from "Canbula sp.", which is widely found in thick sandstone levels forming an important portion of the Tushokrakian succession and from *"Acteocina* sp.", which is concentrated within silty-sandy mudstone forming intercalations in this unit.

Sarmatian: Thickness of this unit that outcrops in the seaward slopes and coastal margins around Kayıkbaşıburnu in Southwest of Sinop Peninsula ranges between 40-600 m. Moreover, this unit can be more than 1300 m thick in the reaion. It is divided into two formations, Kavıkbaşı Burnu and Yavkıl formations (Görür et al. 2000V The Sarmatian succession is Kavıkbaşı burnu and Yaykıl formations (Görür et al. 2000V The Sarmatian succession is described according to the facies types due to the purpose of this study and the sampling is described according to the facies types due to the purpose of this study and the sampling is realized within the same framework. The facies descriptions and sampling levels are described as followings.

Lensoidal Conglomerate: Conglomerates cropping out on the basement of Sarmatian occurrences in Sinop Penninsula are poorly sorted, clast-supported and block sized; Limestone pebbles and large pebbles derived from the basement mud stones are present in the constituent of the conglomerates. Internal structure of these conglomerates characterizes mass flow and lensoidal distribution of this conglomerate support basin magrin depositional system.

Shell bearing sandy limestone- This unit cropping out close to sea level in between Yaykıl and Kayıkbaşı headlands diplays planar and though cross-bedding and shell accretions (Fig. 3b). Shells are intensively cemented by carbonate and mixed embedded fragments or whole shells within the levels of sandy conglomerates and sandstones. Samples for isotope analysis are obtained from association characterized by Card/urn sp., and *Dreissena* sp.

This unit does not display long-distance lateral continuity and presents lensoidal out crops along the shore. In thes unit, low-angle unidirectional planar cross-beddings ana tnrough cross-oeddings support presence of wave case and nign-energy currents periodically activated in shore line belt respectively (Scholle and Spearing. 1988).

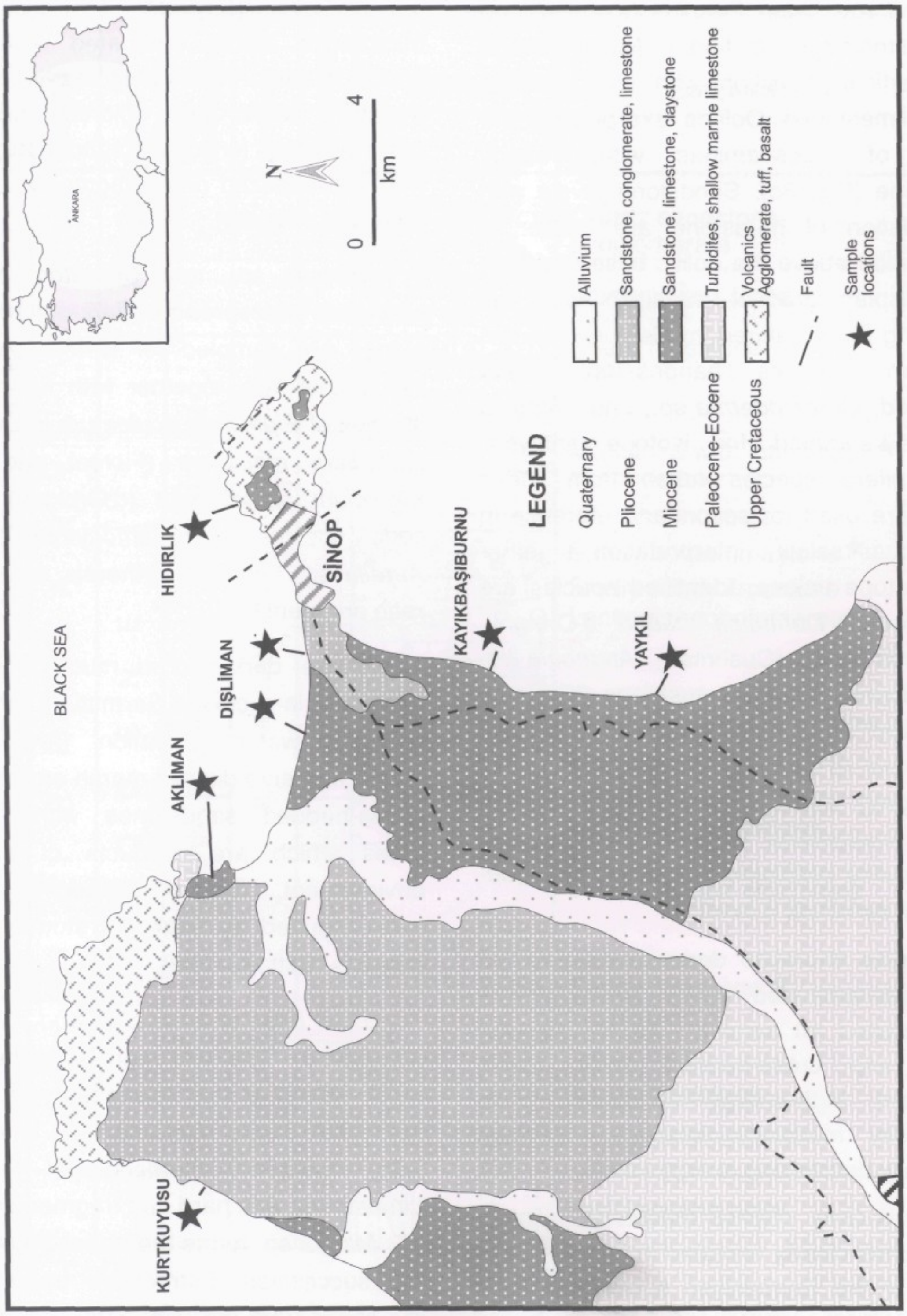

Fig. 1 - The generalized geological map of the study area and the sample locations

Sandy *oolitic limestone.*- This unit is a few meters thick and crops out in shore line between Kayıkbaşı Penninsula and Sinop with restricted distribution. It displays lateral and vertical transition with shell bearing sandy limestones. Oolites take place in the levels of cross-laminae within sandy limestone (Fig. 3c). Sandstone levels with intercalation of mudstone are thicker in succession above the oolitic base (> 15 m) and display gradual transition with the overlying dark colored mudstones. In this part where facies changes are clearly observed, *Cerestoderma* sp., and *Gibbula* sp., is sampled for isotope analysis. Foraminifera species taken from these levels are used for secondary reference in sea water salinity interpretation together with isotope values. Identified species are as follows: *Dentritina haueri* d'Orbigny, *Ammonia tepida* (Cushman), *Ammonia beccarii* (Linne), *Spirolina austriaca* d'Orbigny; *Elphidium reginum* (d'Orbigny), *Elphidium hauerinum* (d'Orbigny), *Elphidium rugosum* (d'Orbigny), *Elphidium macellum* (Fichtel et Moll), *Sinuloculina mayehana* (d'Orbigny), *Sinuloculina cyclostoma* (Reuss), *Ouinqueloculina seminula* (Linne).

These shoreline deposits represented by cross-laminated oolitic limestones is accumulated on sand shoals. Cross-laminated and cross-bedded oolitic levels represent high-energy conditions. Thin sandy-sity mudstone intervals represent gradually decreasing energy levels.

Dark-Gray colored mudstone- These mudstones clearly revealed in middle-upper levels of the Sarmasiyen succession form the most widespread and the thickest facies association. In the bedded portion between massive sections, many wave ripples, bioturbation structures, and score structures are observed (Fig. 3d). Syn-depositional microfaults within laminated levels are typical in this unit. In outer headland beach trenches, within dark colored mudstones cross-bedded lensoidal sandstone levels with carbonatized plant fragments are also observed (Fig. 3e).

Gibbula sp., and *Cerastoderma* sp., which are widespread in dark-colored mudstones are sampled for isotope analysis. Within this unit, together with foraminifera species of *Elphidium aculeatum* (d'Orbigny), *Elphidium macellum* (Fichtel and Moll), *Elphidium fichtelianum* (d'Orbigny), ostracoda species of *Cyamocytheridea* sp., *Auralia* sp. Leguminocythereis and *Cytheretta* are identified.

These dark-colored mudstones were deposited in a closed Sarmatian basin with restricted water circulation. Carbonatized plant fragments display marsh environment. Cross-bedded sandstones within mudstones which are products of stagnant environment, display fluctuating energy levels caused by seasonal storms or tide currents (Clifton, 1988).

Coal bearing limestone and mudstone alternation- This unit represents the upper most portion of the Sarmatian succession. Thin coal bands alternate with limestones with mudstone intercalations (Fig. 3f). Limestones with hard and fragmented shells of? *Mactra* sp. forms the upper boundary of the succession. Extractable fragments of the shells from the rock body are userd for isotope analysis.

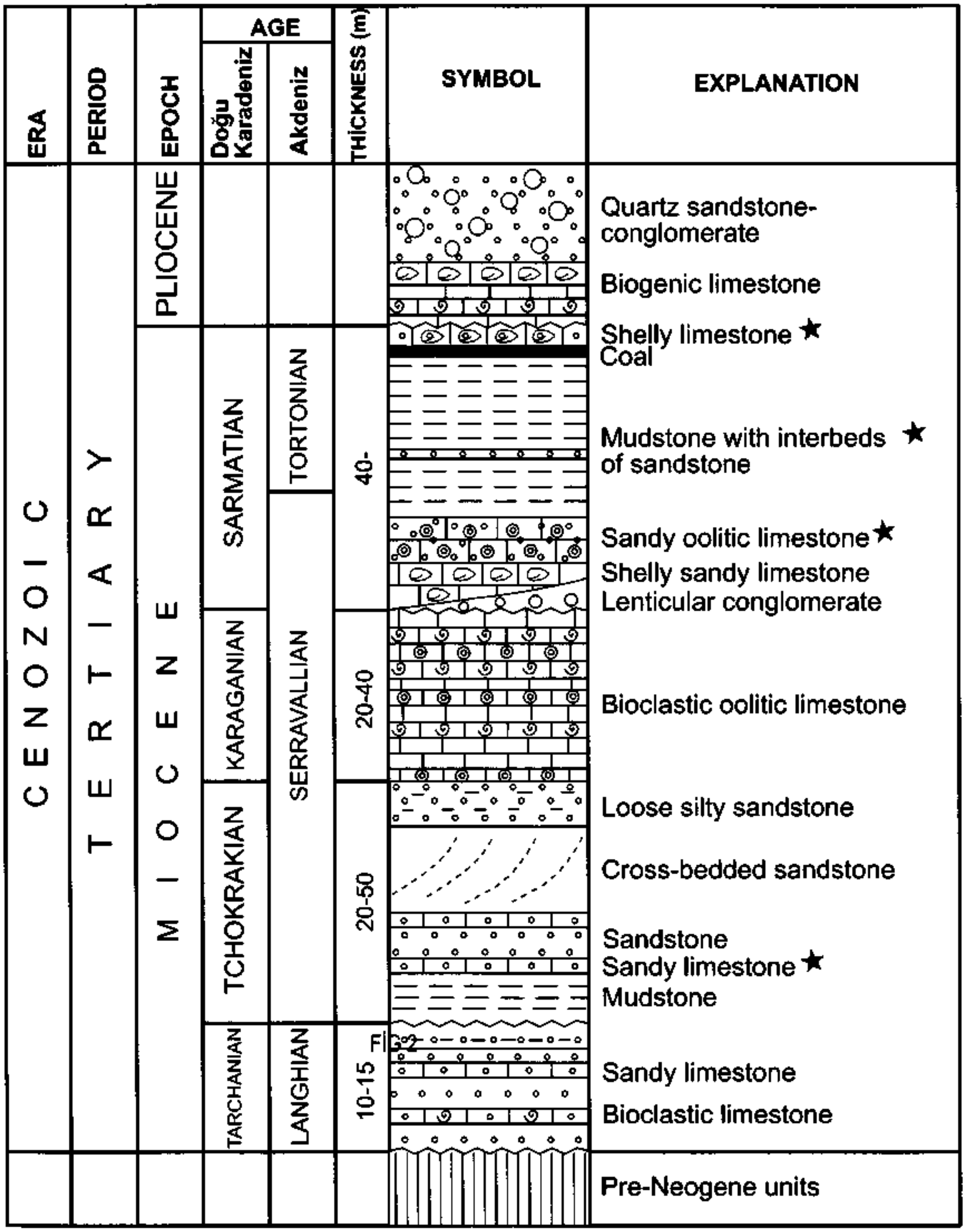

Fig. 2 - Generalized stratigraphic section of the Miocene succession; black stars represent sampled levels (modrified after Özsayar, 1977).

Except a few ostracoda, the limestone with mudstone intercalations forming the larger portion of the facies association does not include paleontological evidence. Coal bearing levels overlying these levels display coastal marsh development and increasing fresh water influences in Sarmatian marine environments.

STABLE ISOTOPES

The values belonging to shells are displayed in table 1. These values are combined in three groups as followings:

Recent Shells: $d^{13}C$= between - 0.99 and 0.48 ‰; $d^{18}O$= between -2.32 and -0.41

Sarmastian Shells: $d^{13}C$= between -2.61 and 0.79 ‰; $d^{18}O$= between -6.46 and -0.10‰;

Tshokrakian Shells: $d^{13}C$= between -0.32 and 0.84 %o; $d^{18}O$= between -2.18 and -2.38 ‰;

In addition to these, the values of recent Black Sea surfaces waters;$d^{13}C$= between 0 and 1 ‰; $d^{18}O$= -2.84 ‰ are obtained from Deuser (1972) and Rank et al. (1999).

When these isotope data are plotted in a diagram, the values for Sarmatian display distribution on a wider area with respect to values of reent Black Sea and values of Tshokrakian shells (Table 1; Fig. 4). However, the values of recent Black Sea and values of Tshokrakian shells display close aerial distribution to each other. The values representing Tshokrakian Sea are grouped in the same area and but the values representing Sarmatian Sea are spread in the diagram. Only one of the values of Sarmatian lies within the recent Black Sea zone, the others are distributed in very wide areas.

The isotope values of shells in the cross-bedded sandy-pebbly limestones with abundant shell accretions overlying the Sarmatian basal conglomerates ($d^{18}O$= between - 3.24 and -3.07 ‰) display considerable decrease in heavy oxygen isotope (^{18}O) with respect to the isotope values derived from oolitic limestones and mudstones overlying them. This kind of decrease in heavy oxygen isotope (^{18}O) in marine environments is due to mixing of fresh water or surface waters with marine water. Fresh waters enriched in light oxygen isotopes (^{16}O) causes dipletion of heavy oxygen isotope (^{18}O) in marine waters (Matyas et al. 1996). Being sedimentologically identified, gullied beach sediments associated with fresh water fluxes in the lower parts of the Sarmatian (Varol et al. 2001), are supporting evidences for isotopic variations in the shells of these units. The second succession representing the middle-upper part of the Sarmatian is composed of sandstone, oolitic limestone and mudstone and interpreted as deposited in shore and backshore environment (Görür et al. 2000; Varol et al. 2001). Heavy oxygen isotope (^{18}O) values of mollusk shells recovered from these levels display an increase with respect to previous levels. This indicates that fresh water influx to the system is stopped and probably that the marine conditions returned. Moreover, the isotope values concentrates around $d^{18}O$= between -0.1 and -0.47 ‰ includes heavier values than other species ($d^{18}O$= between -232 and -187) and the value measured in recent Black Sea waters (^{18}O= -2.84 ‰) except *Mytilus* sp. ($d^{18}O$= -0.41 ‰) living in recent

Table 1 - Table for the list of measured d[18]O and d[18]C isotope values of Tchokrakian and Sarmatian and recent Black Sea faunas.

Sample Number	Speices	d[13]C	d[18]O	Age
1	*Mytilus* sp.	-0,99	-0,41	Black Sea-Recent
2	*Cerastoderma* sp.	0,48	-1,87	Black Sea-Recent
3	*Dona* sp.	0,37	-2,26	Black Sea-Recent
4	*Chlayms* sp.	0,39	-2,32	Black Sea-Recent
5	*Mactra* sp.	-2,61	-6,46	Sarmatian
6	*Gibbula* sp.	-1,5	-0,16	Sarmatian
7	*Cerastoderma* sp.	-1,38	-0,2	Sarmatian
8	*Cerastoderma* sp.	-1,2	-0,1	Sarmatian
9	*Gibbula* sp.	-1,48	-0,16	Sarmatian
10	*Gibbula maeotica*	-1,78	-0,31	Sarmatian
11	*Cerastoderma* sp.	-1,69	-0,47	Sarmatian
12	*Dreissena* sp.	0,41	-3,07	Sarmatian
13	*Cardium* sp.	0,79	-3,24	Sarmatian
14	*Corbula* sp.	-0,78	-2,38	Tshokrakion
15	*Corbula* sp.	-0,78	-2,27	Tshokrakion
16	*Acteocina* sp.	-0,84	-2,26	Tshokrakion
17	*Acteocina* sp.	-0,32	-2,18	Tshokrakion

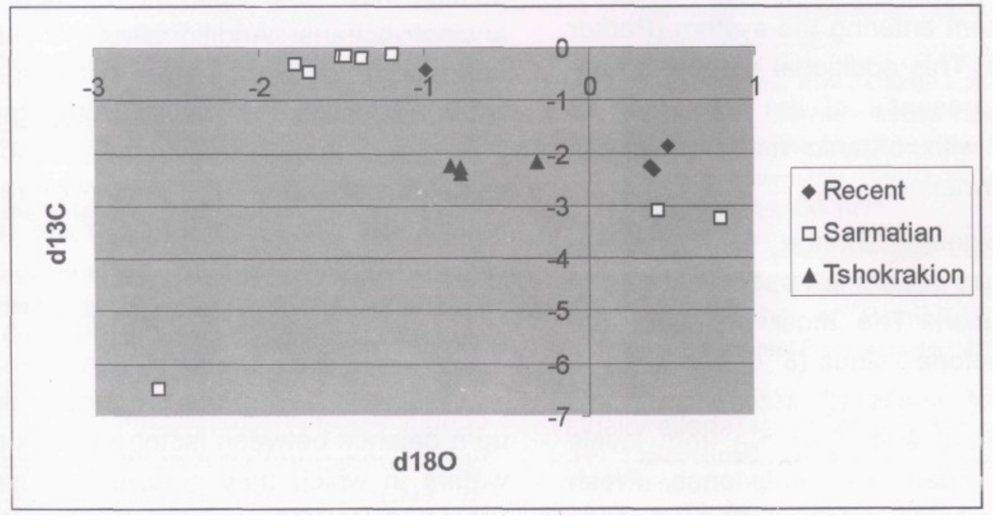

Fig. 3 - Correlation diagram for d[18]O and d[13]C isotope values between Tshokrakian and Sarmatian Tauna ang recent Black Sea surface waters

Black Sea. Additionally, *Dentritina* sp., and *Spirolina* -sp.., concentrated in sandy limestone intercalated with oolitic limestones which were sampled for isotope analysis are known as an association of organisms living in conditons, above the normal marine salinity (A. Poignant, 2000, personal communication). The relative enrichment of heavy oxygen isotope (^{18}O) in these levels in combination of paleontological and isotope data indicates that Sarmatian Sea is more saline and closer than recent Black Sea as indicated in sedimentological studies.

In addition to this, there are other ideas indicating that different organisms living in the same environments with habitats of $CaCO_3$ binding rate, different than marine waters, can cause differences in heavy oxygen isotope ^{18}O fractionation (Arthur, 1983). Depletion of heavy carbon isotopes ^{13}C and enrichment of ^{18}O in shells representing the lower-middle and partially upper portion of the Sarmatian succession can be interpreted as a increase in organic carbon content entering the system (Parker et al. 1972). This additional carbon is supported by presence of the thickening of mudstones with organic matter in these levels of Sarmatian.

The biggest change in Sarmatian isotope values occurs in upper most part of the succession. The most important decrease in isotope values ($d^{18}O$ and $d^{13}C$) of fragments of shells (? *Mactra* sp.), are recorded (-6.46 and -2.61‰) from levels with marls, coals and limestones. Fresh water with light isotopes derived from continental environments and transported to marine conditions decrease the concentration of heavy isotopes in marine environments (Arthur et al., 1983; Matyas et al. 1996). Enrichment in light isotope (^{12}C) in the analysed samples is the evidence of development of fresh water condition in the basin within time. Coal bands in these levels, which are indicators of marsh development, are sedimentological evidences supporting these environmental changes.

RESULTS

The isotope values of shells recovered from Tshokrakian and Sarmatian sediments of Miocene successions in Sinop region ($\mathbf{d}^{18}O$ and $d^{18}C$) is in conformity with the paleogeographic and sedimentological data. When isotope values of shells from recent Black Sea and surface waters and shells from Tshokrakian and Sarmatian Sea are considered, Tshokrakian Sea displays similarities with the recent Black Sea, but the Sarmatian Sea data presents mixing of fresh water fluxes within time.

Stable isotope analysis of shells is a good application for the interpretation of paleosalinity and paleo-temperature of ancient oceans. Additionally, the number and type of the samples are very important factors affecting the results. Although it is considered that the restricted number of samples can be a risk factor for reliable results, the close conformity of the results with previously defined sedimentological properties of Tshokrakian and Sarmatian units decreases this risk factor.

Mollusks having the property of setting up a balance between isotope values of the waters in which they survive with isotope values of CaCO3 secreted for their shells, rae ideal organisms for this kind of studies. It must be considered that the usage of organisms with unknown isotopic fractionation of their shells in environments with indefinite sedimentological framework would be high risky for stable isotope analysis obtained from these organisms.

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PLATE

PLATE - 1

Fig. 1 - Yellow and thick bedded sandstone mudstone is a typical facies for the Tchokrakian succession (carta headland) in the Sinop region.

Fig. 2 - Gravelly and sandy limestone with shell accumulation is a characteric deposite for the basal part of the sarmatian succession (Kayıkbaşı heai dland).

Fig. 3 - Sandy oolitic beds within the sarmatian succession (Gelincik district, SW Sinop)

Fig. 4 - Black silty miudstone with wave ripples typify the middle and upper parts of the sarmatian succession (Yaykıl shore)

Fig. 5 - Cross - bedded sandstones is an indication of high energy deposition within the black mudstone unit with environmental quiescence (Disliman beach)

Fig. 6 - Yellow shelly sandstones and clayey limestones interbedded with coal-bearing layers indicate the upper most part of the sarmatian succession (Disliman beach)


Fig.1

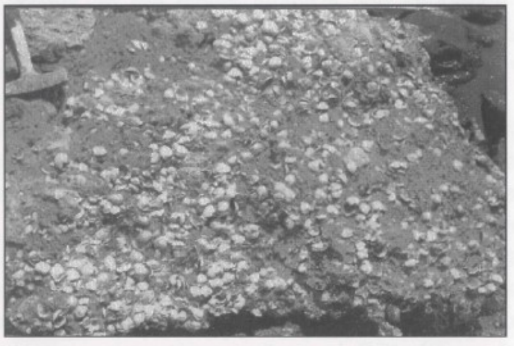
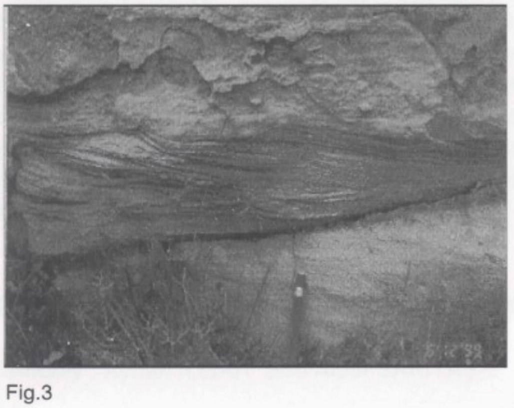


Fig.2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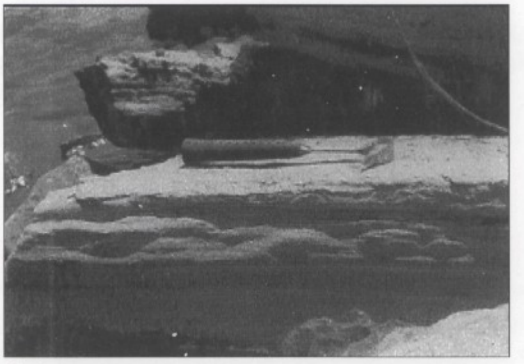
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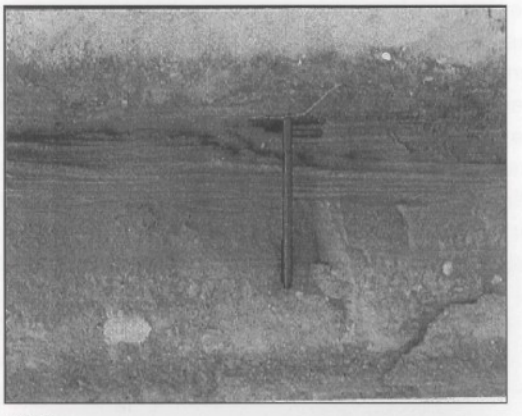
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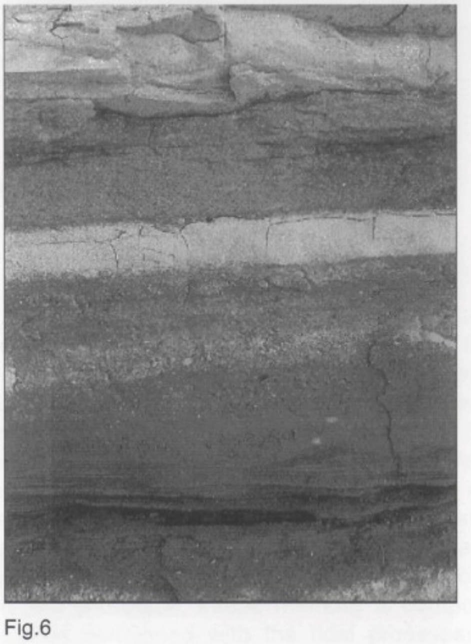
Fig.6

GASTROPODA FAUNA OF KASABA MIOCENE BASIN (WESTERN TAURIDS, SW TURKEY)

Yeşim İSLAMOĞLU**'*

ABSTRACT.- In this study, 37 numbers of gastropoda species which were found in the Uçarsu and Kasaba formations outcroping in the Kasaba Miocene basin, have been determined systematically and their paleo-geographic distributions have been told. Species, which were found into the Uçarsu formation such as *Turntella terebralis turritissima* Sacco, *Turritella terebralis subagibbosa* Sacco, *Turritella (Peyrotia) desmarestina* Basterot, *Turbonilla (Mormula) aturensis* (Cossmann ve Peyrot), *Cassidaria tauropomum* (Sacco) and *Vexillum (Uromitra) pluricostata percostulata* (Sacco) belonging to Early Miocene, they haven't been known in the Middle Miocene. *Cerithium zejszneri* Pusch which was found into the Kasaba formation is a species peculiar to Middle Miocene. According to this, the age of Uçarsu formation as Upper Burdigalian and the age of Kasaba formation as Langhian have been accepted. A great deal of the investigated gastropod fauna distributed both in Tethys and Central Paratethys during early and middle Miocene. Most of the fauna such as *Turritella (Turritella) tricarinata* (Brocchi 1814), *Turritella terebralis turritissima* Sacco, *Turritella terebralis subagibbosa* Sacco, *Turritella (Haustator) striatellatus* Sacco, *Turritella (Peyrotia) desmarestina* Basterot, *Turbonilla (Mormula) aturensis* (Cossmann ve Peyrot), *Cassidaria tauropomum* (Sacco), *Mitrella (Macrurella)* cf. *nassoides grateloupi* (Peyrot). *Vexillum (Uromitra) pluricostata percostulata* (Sacco), *Clavatula (Clavatula) calcarata francisci* (Toula). *Conus conoponderosus* (Sacco) and *Conus clavatulus* d'Orbigny only displays widespreading only in Tethys Tethian origine fauna and and typical species demonstrate that the study area is part of the Proto- Mediterranean - Atlantic biogeographic provence. In the study region, the species which are known only Central Paratethian marine stages have been found. *Cerithium zejszneri* Pusch found in Kasaba formation, is peculiar to Lower Badenian and *Murex (Bolinus) subtorularius* Homes ve Auinger is a species characteristic for Karpatian and Badenian. This situation shows wide connection and faunal immigrations also from western central paratethvs to mediterrane

Key words: Antalya, Kasaba, Miocene, Gastropoda, Systematic, Paleontology

GRAVIMETRIC MODELING OF BASIN AND RELATION WITH EARTHQUAKE DAMAGES

Hakkı ŞENEL

ABSTRACT.- After the 17 August 1999 earthquake, as the regional distribution of the damaged buildinas at Gölcük-İzmit and surroundings are examined, it can be observed that some of the buildings in some regions that have the same quality and type with other buildings in other areas have more damage. This work attempts to point out that the concerned situation is closely related with the alluvium and basement topography over lined by the deposit layer as well as the rigidity of the soil. After 17 August 1999 earthquake, considering the allocation of heavy damaged buildings in the study area it can be seen that, they are close to the half basins whose southern sides are open or buried valley walls. This direction is in the way of the Earthquake waves. This situation clearlv demonstrates the focusing effects of the earthquake waves. In order to show the focusing effect of the earthquake waves, 785 gravity data in 4 km2 area are measured in the settlement of İzmit municipality and. their modelling with inverse solution techniques were done to obtain the shape of the basement topoaraphy which is covered with alluvium. The location of the collapsed, heavy and intermediate damaged buildings was plotted on the calculated basement map and accordingly basin depth and damage distribution were observed as correlated. This relation will be preliminary information for the locations where the damage would be more concentrated in probable further earthquakes. The basement rock shows a deepening attitude towards the southern part of the study area. Basement topography is clearly formed by the buried sequential valley-hillside structures and slope of buried basin is close to right angle over some places. It is interesting that the damage density on the high stored buildings increases on the alluvial areas that are over the buried valley shape basement topography. The vertical or sub vertical hillsides of the basement topography covered by alluvium can be considered as buried fault planes. The surface topography that is seen as valley-hillside in the northern of the study area shows continuity to the west with the alluvial cover in southern of the study area needs to be taken into consideration of the oblique faulting.

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