THE GEOLOGY OF THE EASTERN PART OF THE PULUR MASSIF

Erkan TANYOLU*

ABSTRACT- It is hoped that important evidences about the formation and evolution of Eastern Pontid's can be obtained from a geological study of Pulur massif. Therefore, a total of 430 km area belonging to Pulur massif between Bayburt and Demirözü has been studied.The oldest basement of the area is the Pulur metamorphic complexes. At the base a metamorphic schists, which has a 600 m apparent thickness and at the top 200 m thick limestones of Permo-Carboniferous age are placed. Lower Jurassic formations, which consist of basal conglomerates and sandstones, disconformably overlie the Pulur massif. This formations, are conformably overlain by volcano-sedimentary series of Liassic age. During the Liassic time intensive volcanism yielded the dasite, andesite, diabase and basalt masses. Some basalts contain larce analcime phenocrysts that the rock can be named analcimite. Liassic volcano-sedimentary formations are conformably overlain by 50-60 m thick limestones of Dogger age. This limestones are also conformably overlain by another limestone facies of Malm-Lower Cretaceous age. At the top a melange facies of Aptien-Albien age are seen. These indicate that the Pulur area was a marine depositional environment at least since Carboniferous or since Devonian until Upper Cretaceous, except a time gap between Permian and Lias. This marine basin was shallow at the beginning but .gradually became deeper starring from Dogger time. In the region, two different type metamorphism, which are homblende-hornfels facies of contact metamorphism and Barrowian type amphibolite facies of dynamo-thermal metamorphism, are distinguished. The structural character of the region is similar to Alpine type tectonic. Anticlines, synclines, faults and large extention of a thrust faults, which are elongated in NNE—SWW direction, are the most important features of that tectonic style.

STRATIGRAPHY, STRUCTURAL GEOLOGY AND GEOTECTONIC EVOLUTION OF AMANOS MOUNTAINS WEST OF TÜRKOĞLU (K.MARAŞ)

Cavit DEMIRKOL**

ABSTRACT.- The studied area is situated to the northern part of Amanos mountains, to the west of Türkoğlu. It contains sediments ranging from Lower Palcozoic to Miocene. Lower Paleozoic units are exposed within the core of a large anticline and are parallel to the general trend of the main orogenic belt. Lower Paleozoic consists of a continuous succession ranging from Lower Cambrian to Upper Ordovician and Devonian is represented by Hasanbeyli formation. Mesozoic succession is characterized by thick platform carbonates which are, due to extensive dolomitization, very difficult to subdivide. The carbonate deposition was continuous until the Lower Maastrichtian. Koçali complex was emplaced post-Lower Maastrichtian, causing imbrication of the platform carbonates. The studied area was uplifted during the Eocenc and was later inundated by extensive Miocene transgression. The area has been uplifted at the end of Miocene and has gained the present day topography. N—S compressional regime in the Southeastern Anatolia, which was caused because of the collision of Arabic and Anatolian Plates, was responsible for the emplacement of ophiolites. This event resulted in a number of thrusting in the basement, generating extensive rock cleavage at the bottom of Mesozoic succession. This is, in effect, caused the easily crumbling nature of carbonates. Fold axis, vertical ant thrust fault planes are approximately parallel to each other. Neo-tectonic development of the region is thought to have been effected by the on going Southeast Anatolian compressional regime, yet the control of East Anatolian and Dead Sea faults were evident.

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SOME STRAT1GRAPHIC AND TECTONIC CHARACTERISTICS OF THE AREA AROUND HINIS (SOUTHEAST OF ERZURUM)*

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ABSTRACT.— The study area is situated between Hınıs-Tekman and Karayazı southeast of Erzurum. The aim of this study was to investigate some geological characteristics of the region. The rock units constituting the lowermost sections of the study area consist of alternating gneiss, amphibolite, schist and marble, from bottom to top, are exposed beneath the ophiolitic complex, as a tectonic window. The ophiolitic complex consists essentially of diabase, gabbro, and locally of serpentinite and peridotite. The acidic intrusions cut both the metamorphics and ophiolitic complex. The rocks ranging from Maestrichtian to Pliocene in age are made up of regular alternations of transgressive and regressive sequences, and rest upon the older rocks, unconformably. The study area became a piece of land during the Middle(?) — Upper Miocene. The Oligocene sequence of the sedimentary cover contains thin intercalations of andesitic basalt. In the Miocene-Pliocene sequence of the cover, firstly, volcanic rocks varying from dacite to andesite in composition, and then, in turn, andesitic basalt and basaltic pyroclastic rocks and lavas took place. Overthrusts which are of pre-Maestrichtian and Late Eocene age were also defined in the study area. Along the Late Eocene overthrusts the ophiolitic complex has thrusted over the Eocene rocks which are situated to the south of the complex. Parallel and oblique fault zones which have NW-SE striking right lateral-slips and NE-SW striking left lateral-slips were developed, as well as approximately E-W trending folds and overthrusts under the control of compressional tectonic regime, during the Late and post-Miocene. In addition, some other structures have been also developed in the Plio-Quaternary formations lying within both of two fault zones, by the effects of approximately E-W trending compressional forces.

INTRODUCTION

The study area is situated approximately between Hınıs, Tekman and Karayazı, southeast of Erzurum (Fig.l). The aim of this study was to investigate some geological characteristics of the region. Particularly, the stratigraphic relationships of the study area are controversial.

The pioneer geological works in the region were carried out by Mercier (1948), Erentöz (1949, 1954 a, 6) and Altınlı (1963, 1964). As a result of these works, the significant rock units and stratigraphy of the region were defined, in general. Later, various studies were made in the region, on the basis of 1:25,000 scale mapping (Rathur, 1965, 1969; Tokel, 1979; Şenalp, 1966; İlker, 1966a, *b;* Yılmaz, 1967; Erdoğan, 1966,1967,1972; Sungurlu, 1967; Özcan, 1967; Tütüncü, 1967; Özocak, 1967; Tanrıverdi, 1971; Havur, 1972). The studies which were made by Soytürk (1973) and Erdoğan and Soytürk (1974), constitute a synthesis of above mentioned 1:25,000 scale studies. However, the controversial stratigraphic problems, in particular, started to be determined and to be cleared, step by step, by contribution of new data obtained from the recent studies. On the other hand, the important structural elements such as the metamorphics constituting the lowermost section of the region and ophiolitic complex and relationships between them were examined insufficiently. In this paper, the present writers present some new stratigraphic and tectonic data obtained from an extensive area, by examining the characteristics and positions of the rock units constituting the lowermost section also.

STRATIGRAPHY

In the study area, the pre-Maestrichtian rocks

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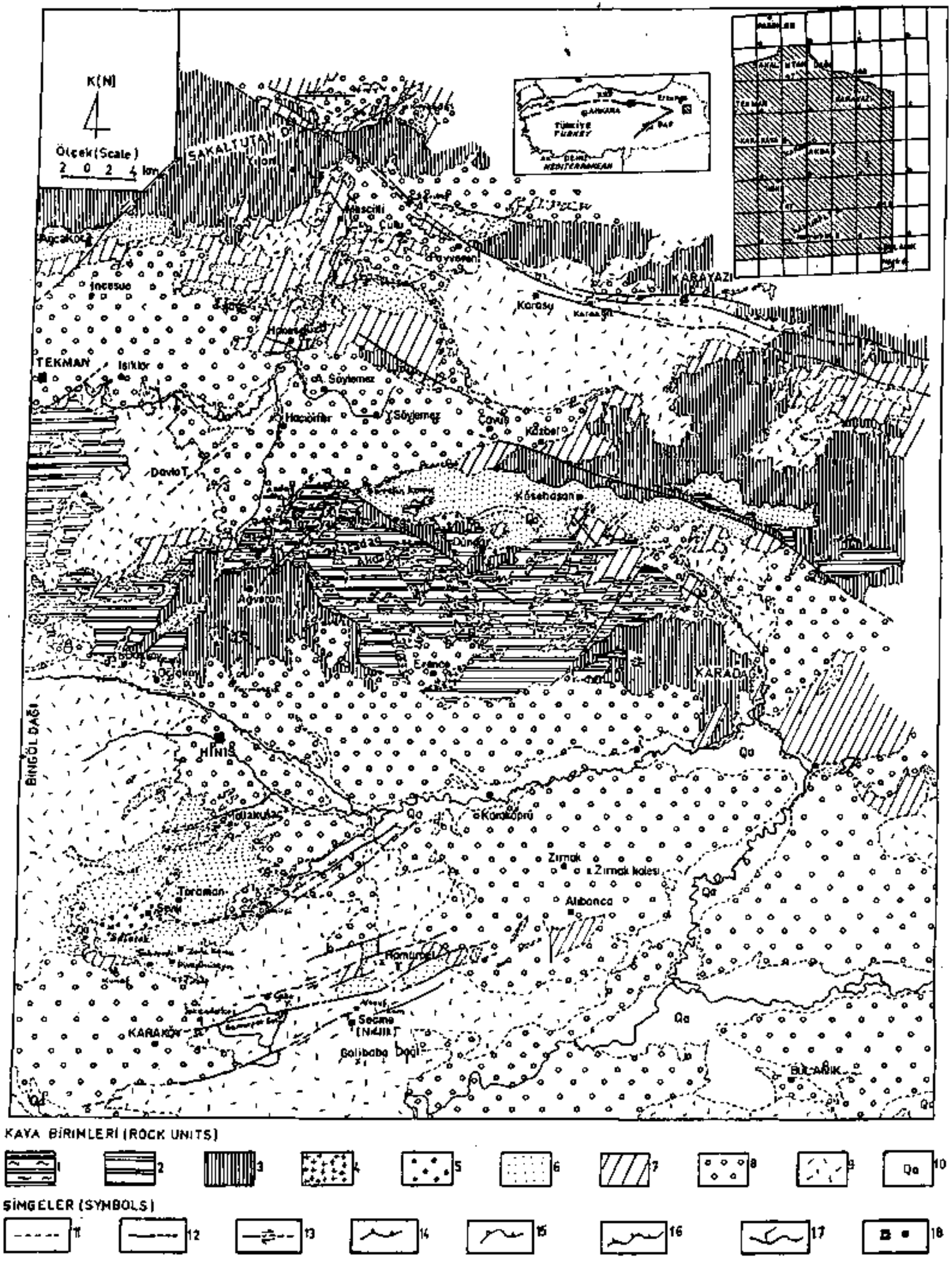

Fig.1— Location and geologic map of the study area (All previous studies were used for preparation of the map). Rock Units : 1,2 — Akdağ metamorphics (1- Gneiss amphibolite, schist; 2- Calcschist and marble), 3— Ağören complex and Sakaltutan group; 4— Tozlu yayla granitoids; 5— Maestrichtian - Paleocene rocks; 6— Eocene and Oligocene rocks; 7— Lower - Middle Miocene rocks; 8— Middle (?) - Upper Miocene - Pliocene terrigenous sediments; 9— Upper Miocene - Pliocene volcanics; 10— Quaternary deposits. Symbols : 11—Contact; 12—Fault, probable fault; 13—Strike - slip fault; 14—The Pre-Maestrichtian - Paleocene overthrust; 15— Late Eocene - Pre Miocene overthrust; 16— Late Miocene and subsequent overthrust; 17— Stream; 18— Human settlements.

and Maestrichtian-Pliocene rocks resting upon them, unconformably crop out.

Pre - Maestrichtian rocks

The rock units constituting the lowermost parts of the study area are subdivided into four units. These are the Akdağ metamorphics lying at the base, the Ağören complex which was thrusted over the metamorphics and the Sakaltutan group and the Tozlu Yayla granitoids.

Akdağ metamorphics — The metamorphic rocks which crop out in the central part of the study area and extend in E-W direction, are named "the Akdağ metamorphics" (Erdoğan, 1972). However, the rock units presenting the all levels of the metamorphics, are best exposed in the vicinity of Karadağ to the west of Akdağ and especially on the southwestern slopes of Karadağ mountain (Figs. 1 and 2).

The metamorphics consisting of gneiss, amphibolite and mica-schist in the lowermost levels and marble and locally calc-schist in the uppermost levels, are generally well foliated and intensively folded. On the basis of petrographic determinations, some rocks such as amphibole-gneiss, amphibole-pyroxene gneiss, garnet-biotite gneiss, biotite-muscovite gneiss, cordierite-almandine-sillimanite-quartz gneiss, garnet - sillimanite - biotite gneiss, quartz - oligoclase-diopside - actinolite gneiss, amphibolite, garnet mica-schist, quartz - cordierite micaschist, calcschist and sugary textured marble were defined in the metamorphics. On the basis of index minerals contained in the metamorphics, it has been concluded that metamorphics lying within the study area underwent a predominantly medium to high grade metamorphism (under the conditions of approximately 500° - 600°C temperature and 4 - 5 kb pressure) in amphibolite facies, according to the criteria of Winkler (1976), or a metamorphism in Barrowian type sillimanite - almandine - orthoclase sub-facies, according to the criteria of Turner and Verhoogen (1960) and Winkler (1967).

The Akdağ metamorphics crop out as a tectonic window beneath the Ağören complex. Uncorifownity showing regional character was not observed, between the metamorphics. Nevertheless, an unconformity which has local character or coincides with a tectonic phenomena, is observed, to the south of Karadağ. The marble contains some pebbles and blocks derived from the underlying amphibolite and gneiss along this local unconformity. The marble which contains pebbles and blocks, conformably passes into the alternations of marble and schist, upwards.

The Akdağ metamorphics, have been probably produced by the high grade metamorphism of a sequence of volcanics?, quartz sandstone, claystone, siltstone and limestone, in ascending order.

It may be suggested that the Akdağ metamorphics show similarities to the section of the Central Anatolian massif, in the vicinity of Akdağmadeni-Yıldızeli(Yılmaz, 1980,1982; Özcan et al., 1980), on the basis of rock types, stratigraphic succession, degree of metamorphism and their relationships to the Ağören complex and the other rock units. For instance, it has been defined that both two sequences consist predominantlyof gneiss and schist in the lower levels and marble in the upper levels, and underwent a metamorphism in amphibolite facies and are covered by the Maestrichtian sediments, unconformably.

The Maestrichtian-Palaeocene formations rest upon the metamorphics and ophiolitic complex in the region with angular unconformities. On the basis of these observations the metamorphics may be the pre-Maestrichtian-Palaeocene in age, at least.

Ağören complex.— The Ağören complex rests upon the Akdağ metamorphics, tectonically and consists essentially of serpentinite, peridotite, gabbro and diabase. It is termed after Ağören village, where it is best cropped out. The rock units constituting this complex, were thrusted over each other, too.

Serpentinite and peridotite are found in small outcrops, gabbro is locally common and diabase is dominant rock unit.

Serpentinite and peridotite occur as small lenses. Peridotite consists of serpentinized olivine, clinopyroxene, orthopyroxene and spinel, and has wehrlitic character. Gabbros are found as cumulates and have intersertal texture. Gabbro consists of olivine in lesser amounts, altered plagioclase (albitised), locally clinopyroxene, orthopyroxene in large amounts and rarely titanite and magnetite. Diabase has ophitic texture and contains hornblende and plagioclase (labradorite), both in the same proportions, with locally rutile in large amounts, titanite and opaque materal.

A dynamo metamorphism took place in some places where the ophiolitic complex had been thrusted over the metamorphics. In such places the units of ophiolitic assamblage may be called as metagabbro, metadiabase or metadolerite. The rock units belonging to the ophiolitic melange, i.e. gabbro and diabase were locally subjected to metamorphism, as mentioned above. For instance, amphiboles were altered to chlorites pyroxenes were uralitized, plagioclases were sericitized, fractures were filled with locally prehnite or iron-bearing formations.

Tozlu Yayla granitoids.— The acidic intrusions varying from granite to diorite in composition which cut both the Akdağ metamorphics and Ağören complex in the study area, and are of pre-Maestrichtian age, were named the Tozlu yayla granitoids (Fig. 1 and 2).

The acidic intrusions are found as small outcrops in the central part of the study area. The dykes and sills of the intrusions are very common in both the metamorphics and ophiolitic complex. The outcrops of the acidic rocks are distinguished from the adjacent rocks on the basis of yellowish colour, and cover an area of a few sq. km or less. The texture of granite is holocrystalline. It consists essentially of quartz, plagioclase, biotite and tourmaline. Granodiorite has holocrystalline porphyric texture and consists of quartz, orthoclase, plagioclase, biotite and muscovite; it has locally cataclastic texture and contains chloritized biotite, epidote, zircon, apatite and opaque material. The dioritic rocks which are generally called as quartz-diorite or monzodiorite have holocrystalline granular texture, and contain quartz, albite-oligoclase in large amounts, pyroxene (diopside-augite) and in places titanite.

The acidic intrusions constituted an extensive contact metamorphism at the contacts of the metamorphics and ophiolitic complex. In some sections, where the metamorphics were cut by the intrusions, calc-silicate fels and calcite-diopside fels were formed. In these zones, skarn rocks are characterized by banded structure and consist essentially of diopside, quartz and calcite. In addition, sericitized plagioclase, skapolite, muscovite, chlorite, datholite and titanite are observed in these zones. In the sections, where the rocks of the ophiolitic complex were cut by the acidic intrusions, the major minerals were undergone variations. For instance, plagioclases were sericitized, pyroxenes were altered to amphiboles. The edge parts of the acidic intrusions have porphyric texture.

The Tozlu yayla granitoids maybepre-Maestrichtian-Palaeocene in age as the Akdağ metamorphics and Ağören complex, probably Upper Cretaceous. The fact that the Maestrichtian - Palaeocene conglomerates and sandstones contain some fragments derived from the acid intrusive rocks, are in favor of this age.

Sakaltutan group.— The present writers have preferred to name the formations as the Sakaltutan group, been named firstly as the Sakaltutan ophiolites (Erdoğan and Soytürk, 1974), and then a part of them as the Anatolian ophiolitic melange (Koçyiğit, 1985), on the basis of the consideration that they have a heterogenous structure, and the ophiolites, melange, metamorphics and, the Upper Cretaceous pelagic limestones resting upon the melange as a cover may be differentiated from each other and named separately. The Sakaltutan group with its heterogeneous structure, were also differentiated from the Ağören complex consisting wholly of ophiolitic rock - units.

The Sakaltutan group consists essentially of serpentinite, peridotite gabbro, diabase, pillow structured volcanic rocks and pelagic limestone as well as metamorphics which underwent metamorphism in greenschist facies, marble and recrystallized limestone. The rock units corresponding to the ophiolitic sequence, were thrusted over one another and the other rocks. Additionally some olisthostromal levels are locally observed in this group. The matrix of these formations was derived from the volcanoclastic sandstone and claystone. The blocks of them were derived from the ophiolites or the other rock units constituting the group. The age of the limestone olistolithes ranges from Permian to Cenomanian, according to the determinations (Fig.2).

The Sakaltutan group constitutes the basement of the other rocks, to the north and further north of the study area, were thrusted over the Eocene olisthostromal rocks, in the south and north. The Oligocene red coloured terrigeneous and clastic rocks are the oldest formations resting upon this overthrust and complex, transgressively.

Maestrichtian and younger formations

In this chapter, the Maestrichtian - Palaeocene, Eocene - Oligocene, Lower - Middle Miocene and Upper Miocene - Pliocene units, in turn will be examined. The rocks which will be defined in turn, consist generally of regular alternations of transgressive and regressive sequences. The study area completely became a piece of land in the final regressive stage during the Middle? — Upper Miocene.

Maestrichtian - Palaeocene rocks

The Maestrichtian Dündar formation, Palaeocene Sevik formation and Merttepe formation will be defined in turn.

Diindar formation . — The Maestrichtian sequence consisting of conglomerate, sandstone and claystone alternations in the lowermost parts, sandstone, sandy-clayey limestone alternations in the central parts and clastic rocks with intercalations of rudistids-bearing limestone in the uppermost parts, are firstly named the Dündar formation by Erdoğan (1968), Erdoğan and Soytürk (1974).

The formation crops out typically to the south of Dündar village.

The grey-greenish coloured, thickly bedded polygenic conglomerate constituting the lowermost part of the formation, generally overlies the Ağören complex. The pebbles which were derived from" the rocks of the complex and metamorphics are well rounded and ill-sorted. By decreasing of grain size, gradually towards the central and uppermost levels, the alternations of sandstone and claystone, and intercalations of locally rudistids and algae bearing thick bedded limestone become predominate.

The Dündar formation rests upon the metamorphics and ophiolitic complex transgressively. It is represented by rudistids-bearing limestones, in the southwest and west of Ağören.

On the basis of palaeontological determinations (Fig.2), Maestrichtian age is assigned to the Dündar formation. It has probably deposited in a shallow-sea environment with high energy.

Sevik formation and Merttepe formation. — The rock units consisting of alternating Middle Palaeocene sandstone, limestone and marl in the lowermost levels, and the Upper Palaeocene rocks, mostly limestone in the uppermost levels, were named the Sevik formation and the Merttepe formation, respectively (Soytürk,1973).

Both formations crop out typically in the vicinity of Sevik village and in the eastern part of Seferek village.

The unit which consists predominantly of elastics, in the lowermost parts and predominantly of carbonates in the uppermost parts, is generally medium-thick bedded and regularly startified. Sole markings, gradation, parallel and convolute laminations are common in clastic rocks. The micritic carbonates contain abundant foraminiferas.

The base of the Sevik formation is not seen in the Hınıs basin. The levels of clastic rocks corresponding to the Upper Palaeocene-Lower Eocene rock units resting upon the Dündar formation conformably, are observed in the southern part of the Tekman - Karayazı basin. These levels may be equivalents to the Sevik formation. So, it may be suggested that the Sevik formation, conformably and passes into the overlying Merttepe formation, conformably.

On the basis of palaeontological determinations (Fig.2), the Palaeocene Sevik formation and Merttepe formation have been probably deposited in a shallow-sea environment, with low energy, in which the turbidity currents had predominated.

Eocene - Oligocene rocks

Here, the Toraman formation, the Sancaktar formation and the Kösehasan formation, the Ahlat formation and the Ağcakoca formation will be defined in turn. The sequence which started (as being) transgressive in the Maestrichtian, gained regressive character durirtg the end of Eocene and firstly transgressive and then regressive characters during the Oligocene (Fig.2).

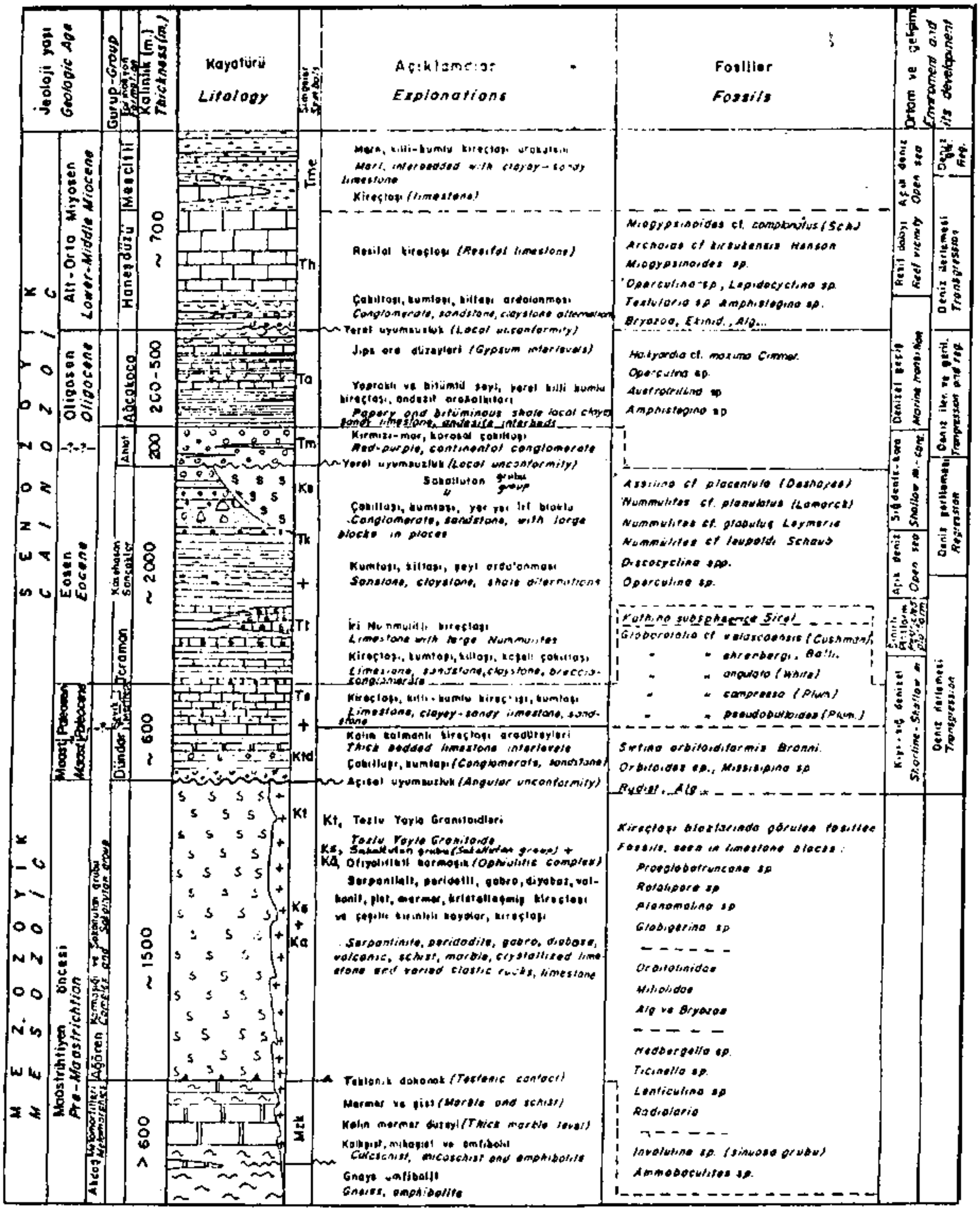

Fig.2— Generalized columnar section of the study area, up to Lower - Middle Miocene.

Toraman formation, —. The Eocene regular sequence consisting of alternating marine sandstone, claystone, marl and limestone was originally named the Toraman formation by Soytürk (1973) with its name of type location.

This unit consists of alternations of grey, greenish, yellowish coloured, thin to medium thick bedded and regularly stratified clayey limestone, claystone, sandstone, conglomerate and marl. The primary sedimentary structures such as sole structures and gradation, parallel and convolute laminations are clear in some sections where clastic rocks predominate.

The Toraman formation rests upon the Upper Palaeocene formations (Merttepe formations), conformably and transitionally, in the Hınıs basin (Soytürk, 1973). In the transition levels, some lenses of brecciated limestone lenses are found. The pebbles and matrix of the limestones were derived from the same carbonate. The above mentioned lenses may be regarded as products developed by high energy in the environment, during the transition from Palaeocene to Eocene.

On the basis of palaeontological determinations (Fig. 2), the Eocene Toraman formation has been probably deposited in *a!* shallow-sea environment, with low energy, in which the turbidy currents had predominated.

Sancaktar formation and Kösehasan formation.— The Eocene rocks consisting of regularly stratified clastic rocks and limestones in the lowermost levels, and coarse grained clastic rocks having olisthostromal and regressive characters in the uppermost levels in the Tekman-Karayazı basin were named the Sancaktar formation and the Kösehasan formation by Erdoğan and Soytürk(1974), on the basis of their type locations. However, the Eocene limestones are found as lenses in the section where the clastic rocks predominate, and locally correspond to a member.

Both of the formations crop out typically in the southern and western parts of Kösehasan village. These units consisting of alternating mostly sandstone, claystone, shale and locally conglomerate, contain some lenses of nummulites bearing bioclastic limestone, particulary in the sections where they rest upon the metamorphics, unconformably. The clastic rocks are grey, yellowish, greenish coloured, generally thin to medium thick bedded, whereas the limestones are generally thick bedded and regulary stratified. The unit having regular stratification in the lowermost part, show a blocky successions whose materials were derived from the complex, in the places where the Sakaltutan group was thrusted over it. This succession is observed to the western of Kösehasan village.

The Kösehasan formation overlies the Maestrichtian-Palaeocene clastic rocks, conformably. It overlies the basement which is constituted by the metamorphics and ophiolitic complex, with an angular unconformity. On the basis of palaeontological determinations (Fig. 2), it may be suggested that the Eocene units were deposited in a marine environment which had been shallow at the begining, became deepining and later became more shallow gradually towards the end of Eocene.

The lowermost parts of the Kösehasan formation resemble the Toraman formation, in terms of age, environment of deposition and rock types, but the middle and uppermost parts of the formation differ from the Toraman formation, on the basis of its blocky and regressive character.

Ahlat formation . — The terrigeneous unit consisting of alternating red coloured polygenic conglomerate, sandstone and locally siltstone, rests upon the Eocene rocks, was named the Ahlat conglomerate (Demirtaşlı and Pisoni, 1965), the Mollakulaç formation(Erdoğan, 1967 ; Özcan, 1967) and the Ahlat formation (Soytürk, 1973). Although it doesn't crop out typically in the vicinity of Ahlat, since it is the known and originally used name, the present writers preferred to use the geographic name as Ahlat for this formation. This formation crops out typically in the vicinity of Mollakulaç, village and southwestern part of it.

The reddish polygenic and blocky conglomerate as a product of terrestrial environment, is unbedded or thick bedded, its grains being unsorted, sub - rounded and poorly graded. The sandstone and siltstone are thin to medium thick bedded. Wedging and cross -

bedding are common in this unit. The coarse -grained materials show channel structure.

The Ahlat formation overlies the Toraman formations, conformably in the Hınıs basin, according to Özcan (1967). Soytürk (1973) and, Erdoğan and Soytürk (1974) suggested that the Ahlat formations contains some fragments of Lower - Upper Eocene, and underlies the Oligocene rocks, may be Upper Eocene in age. According to the observations, the relationship between the Ahlat formation and the Eocene rocks is unconformable in the Hınıs basin and conformable, in the vicinity of Ağcakoca situated within the Tekman - Karayazı basin. On the basis of available data the age of the Ahlat formation probably ranges from Upper Eocene to Lower Oligocene. The relationship of the formation to the underlying Lower-Middle Eocene formations, may be only a local unconformity ofa regional phenomenan.

Ağcakoca formation — Various names were used for the Oligocene rocks consisting essentially of pebbly sandstone, marl, foliated shale and limestone and gypsum - bearing levels in upwards. It has been named the Mollakulaç, Dere formation by Erdoğan (1967) and Özcan (1967), the Çığılgan formation by Rathur (1965), İlker (1966a), Erdoğan (1966, 1972) and Havur (1972), the Ağcakoca formation by Aziz (1971), the Gümüşali formation by Erdoğan and Soytürk (1974). Since it crops out typically in the vicinity of Ağcakoca village, in the northwestern part of the study area and hence, the lower - upper contact relations are best seen there, the name Ağcakoca formation (Aziz, 1971) was preferred. The name of Çığılgan formation wasn't preferred since the landslide made the section measurement impossible, in the vicinity of Çığılgan. Whereas Ağcakoca village and neighbouring areas are the most typical locations, where the all characteristics of the formation are best examined.

The formation consists predominantly of shale, claystone and marl, in places, levels of pebbly sandstone and beds of clayey - sandy limestone. The uppermost parts of the formation, contains gypsum bearing intercalations, in places. The Ağcakoca formation is generally greenish-bluish coloured and thin to medium-thick bedded. The levels of pebbly sandstone are thick-bedded. Sole structures, gradation, parallel and convolute laminations, and sand - concretions are common in this, unit. The Ağcakoca formation contains locally remnants" of plant and abundant macro or microfossils. Andesitic basalt interlayers are observed rarely.

' All the earlier workers of the region, suggested that this formations rests upon the underlying terrigeneous formations, conformably and transitively. Oligocene age is assigned to the uppermost part of this formation by ilker (1966a) and Lower Miocene by Havur (1972) in the northern part of the Tekman-Karayazı basin. The lowermost and uppermost levels of the formation were deposited in a shallow - sea environment, whereas the central levels were deposited in a relatively deep - sea environment. On the basis of its fossil content, the Ağcakoca formation is probably of Oligocene age.

Lower - Middle Miocene rocks

The Haneşdüzü formation and the Mescitli formation will be defined, in turn, here. The sequence which had continued its development, regressively, towards the end of Oligocene, gained a transgressive character, at the begining of the Lower Miocene and regressive character again during the Middle Miocene. The study area became a piece of land, as a whole, as a result of development of final regressive sequence (Fig.2).

Haneşdüzü formation.— This formation, occurring very widespread in Eastern Anatolia and consisting predominantly of platform type of carbonates was named the Adilcevaz calcareous by Demirtaşlı and Pisoni (1965), the Haneşdüzü formation by Rathur (1957), İlker (1966a), Erdoğan (1966) and Koçyiğit (1985), the Alibonca formation by Soytürk (1973) and the Güzelbaba formation by İlker (1966b). The outcrops and lower and upper contacts of the formation which is of Lower Miocene and consists predominantly of limestone, are best seen near Haneşdüzü hill and in its eastern part inside the study area (Fig.l).

The unit, starting with alternation of fine - grained conglomerate, sandstone and claystone, passes into the reef carbonates, conformably. The uppermost levels of this formation is locally represented by reef carbonates. The clastic rocks lying within the lowermost levels are greenish, yellowish coloured, and mostly, thin to medium - thick and cross-bedded.The carbonates are grey, whitish and yellowish coloured, very thick bedded, and contain abundant macro or microfossils. The sequence of bioclastic limestone generally containing fragments of reef - forming organisms, in places has oolithic texture and contains intraclasts.

According to Rathur (1965) and Soytürk (1973), an unconformity exists between this formations and the Oligocene formations. On the basis of variations and characteristics of rocks lying at the lower contact, it may be suggested that the sea had become rather shallow toward the end of Oligocene and become terrestrial, locally. However the contact relation doesn't point out a significant regional unconformity. This evidence reflects that a regressive sequence which had developed at the end of Oligocene, probably gained a transgressive character during the Lower Miocene.

On the basis of palaeontological determinations (Fig. 2), the Haneşdüzü formation, Aquitanian — Burdigalian in age, was deposited in a shallow - sea environment, which was neritic and the development of reefs was continuing in.

Mescitli formation.— The sequence which crops out typically within a limited area in Eastern Anatolia and particulary in the vicinity of Mescitli, consists predominantly of marl and rests upon the Lower Miocene carbonates, conformably and transitionally, was named the Mescitli formation. It corresponds to the marl series of the Mescitli formation of Rathur (1969), and the Mescitli member of Erdoğan and Soytürk (1974). İlker (1966a), Tanrıverdi(1971), Erdoğan (1972) and Havur (1972) named the same unit as the Mescitli formation.

All the characteristics and lower-upper contacts of the formation are best seen in the vicinity of Mescitli.

Marl is the dominant rock. It consists of limestone, in the lowermost levels, claystone and clayey limestone in the middle levels, and interlayers of light coloured dacitic and andesitic pyroclastic and epiclastic rocks and lavas in the uppermost levels. In general, it is light grey, whitish beige coloured and thin to medium - thick bedded and regularly stratified. The thick beds are common in lower and uppermost levels. Gradation and lamination are observed in some levels of the wackestone and packstone. Pelagic formations are seen in the middle levels gastropoda and lamellibranchiata fossils in the uppermost levels.

It has been accepted that the Mescitli formation rests upon the Haneşdüzlü formation, conformably and transitively (İlker, 1966a; Rathur, 1966). On the basis of palaeontological determinations (Fig. 2 and 3), it may be suggested that the Mescitli formation, Lower - Middle? Miocene in age was deposited in a marine environment which had been shallow at the beginning, deep and pelagic later and, subsequently become more shallow again.

Upper Miocene - Pliocene rocks

The Middle - Upper Miocene - Pliocene formations show some remarkable differences in places, in terms of their settings, terrestrial facies and relationships to the synchronized volcanics. For this reason, separate columnar sections were prepared for each location (Fig. 3, 4 and 5). Soytürk (1973), Erdoğan and Soytürk (1974) named this deposits as the Zırnak formation and subdivided it into several members. The Pliocene terrigenous units were mapped separately and defined that the units were rested upon the Zırnak formation, unconformably, in this study. However, some relationships which were mentioned above to be unconformable, were observed to be conformable. The Middle?—Upper Miocene—Pliocene sequence was regarded as separately for different places. In this chapter, Mescitli area, the area between Hınıs and Tekman, and Hamurpetdağı area will be defined, in turn.

Mescitli area .— The Mescitli formation is underlain by the Çullu formation, conformably in the vicinity of Mescitli. The İncesu formation and Karayazı volcanics follow the Çullu formation, respectively (Fig.3).

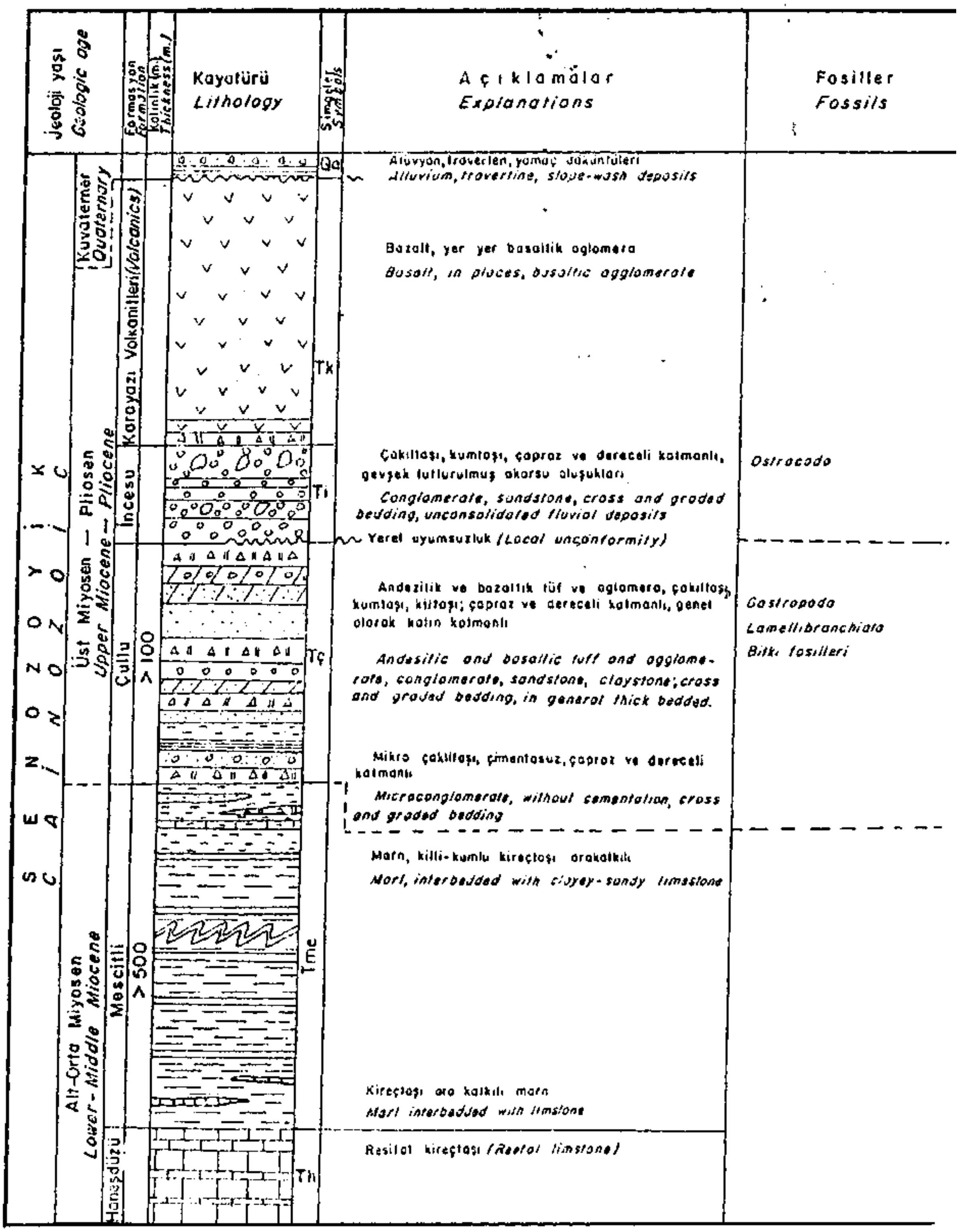

Fig.3— Columnar section of the Mescitli area.

Çullu formation: The rock units, which had been named the agglomerate series of the Mescitli formation by Rathur (1969) and the Çullu agglomerates by Erdoğan and Soytürk (1974), were named the Çullu formation, on the basis of the fact that they consist of alternations of terrigenous sedimentary deposits and pyroclastic rocks, Koçyiğit (1985) named the terrigenous sedimentary rocks as the Payveren formation, and the pyroclastic rocks as the Çilligöl formation in the eastern part of Çullu village. The Çullu formation probably corresponds to the lateral transitive sections of these two formations.

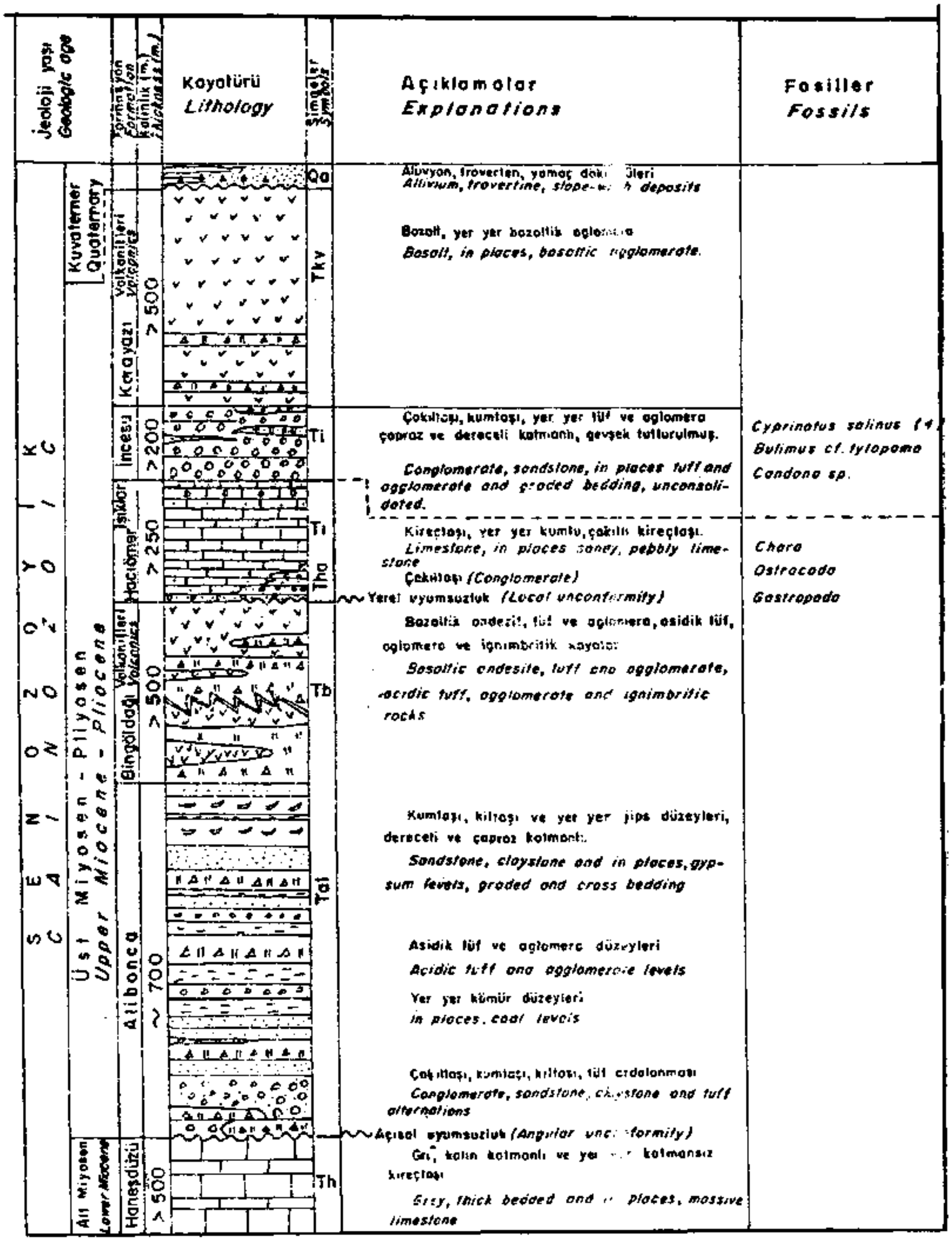

Fig.4— Columnar section of the Hınıs - Tekman area, (1) after Erdoğan and Soytürk (1974).

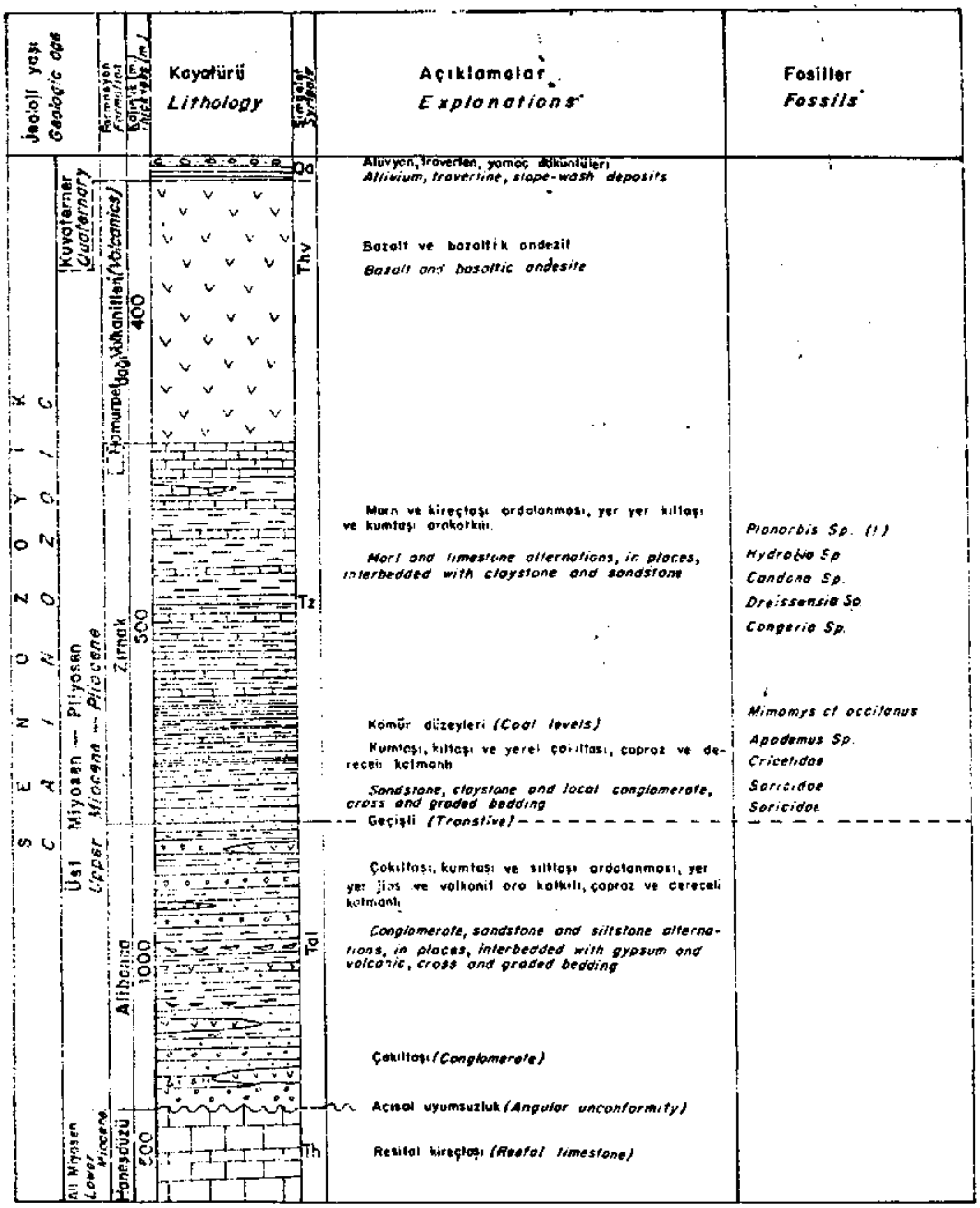

Fig.5— Columnar section of the Hamurpet dağı area, (1) after Sungurlu (1967).

The Çullu formation, consisting essentially of alternating polygenic conglomerate, sandstone, claystone, andesitic, pyroclastic and epiclastic rocks is grey - dark grey, in places, brown and greenish coloured and medium - thick to thick bedded and in some places very thick bedded. Coal fragments, plant fossils and well elutriated sandstones are common in some levels, especially in the lowermost levels. Cross and graded bedding, channel and load structures, sliding deposits synchronized with deposition, oblique-slip faults are often observed. The size and quantity of pebbles belonging to the conglomerates gradually increase from the lowermost levels to uppermost levels. The rock units mostly show lateral and vertical transitions.

The above mentioned rocks rest upon the Mescitli formations, conformably, according to Rathur (1965) and Koçyiğit (1985), unconformably, according to Erdoğan and Soytürk(1974). On the basis of the observations, this unit rests upon the Mescitli formation, both conformably and transitively. The transitive relationship represents the phenomena that the study area changed as a whole from the marine environment to a piece of land as a result of the last regressive development which started during the Middle? Miocene.

Since this unit which was formed in various terrestrial environments such as deltaic, fluvial and lacustrine environments, rests conformably, on to the Middle Miocene Mescitli formations, the Çullu formation may be Middle? — Upper Miocene in age.

İncesu formation: The terrigenous deposits consisting of conglomerate and locally sandstone and claystone, was named the İncesu formation (Aziz, 1971; Tanrıverdi, 1971; Erdoğan, 1972; Havur, 1972; Erdoğan and Soytürk, 1974).

The unit, consisting of alternating poorly - bedded conglomerate, sandstone, mudstone and claystone is generally orange, grey - brown coloured, uncemented or weakly cemented, its grains being well - rounded and unsorted In the some places, blocks are seen as much as 1m in size.Cross and graded bedding have been developed rather well. The unit rarely contains interlayers of vitric - crystalline pyroclastic rocks.

On the basis of the above mentioned preliminary studies, the unit which is of Pliocene age, overlies all the formations unconformably. On the basis of the observations, in the vicinity of Tekman, the İncesu formation rests upon the Işıklar formation which was formed in a lacustrine environment, conformably and transitionally. Nevertheless the relationship regarded as an unconformity must be a local unconformity.

Karayazı volcanics: The rocks" consisting predominantly of olivine basalts and, in places, andesitic rocks were named as the Karayazı volcanics, on the basis of the fact that they crop out typically in the vicinity of Karayazı. They partly correspond to the Karasu basalt and Kaletepe andesite of Koçyiğit (1985). The volcanics varying from andesite to basalt in composition, are regarded as different products of the same phenomena. It may be suggested that andesitic lavas solidified in higher places whereas basaltic lavas solidified in lower places, since basaltic lavas have much lower viscosity.

The lavas which are dark - black and brown coloured and with abundant vesicles and columnar joints, have porphyric or intersertal texture with flow structure. Olivine, hornblende, augite, plagioclase varying from andesine to labradorite in composition and opaque minerals are found in a groundmass composed of plagioclase microlites. These volcanics are with abundant greenish lichens.

The Karayazı volcanics overlie the older units, approximately horizontal. They are probably of Pliocene or Plio - Quaternary age.

The area between Hınıs and Tekman — The above mentioned Haneşdüzü formation is overlain by the Upper - Miocene Alibonca formations which is a terrigenous volcanosedimentary unit, with an angular unconformity, to the northwest of Hınıs. The Alibonca formation is followed by the Bingöldağı volcanics, Hacıömer formation, Işıklar formation and the İncesu formation which crops out in the vicinity of Mescitli, to the north of the study area, and the Karayazı volcanics, respectively (Fig.4)

Alibonca formation: The Upper Miocene terrigenous volcanosedimentary unit consisting of alternating conglomerate, sandstone, mudstone, marl, clayey limestone, gypsum, pyroclastic rocks and lavas, was named as the Alibonca formation (İlker, 1966b; Sungurlu, 1967). This formation corresponds to the lowermost levels of the Zırnak formation of Soytürk (1973).

The unit is orange, locally grey and greenish coloured, thin, medium-thick and thick bedded and contains coaliferous levels and fragments of plants. Cross and graded bedding, channel and load structures, sliding formations synchronized with deposition, oblique-slip faults are common in this unit. This unit contains more volcanic intercalations, in the eastern part of Bingöldağı, than those near Alibonca village.

The Alibonca formation resting upon the Haneşdüzü formation with angular unconformity, was deposited in a terrestrial environment between fluvial and lacustrine, in which a volcanic activity took place.

Bingöldağı volcanics: The volcanics, a part of which was named as the Tekman basalt (Erdoğan and Soytürk, 1974), are mostly of Upper Miocene age and varying from andesite to andesitic basalt in composition, originated from Bingöldağı caldera. So, these volcanics were named as Bingöldağı volcanics.

They crop out typically in the vicinity of Bingöldağı, particularly near Ortaköy - Başköy, on the eastern slope of the Bingöldağı mountain.

The volcanics, which are grey, blackish grey, generally thin medium-thick, in places thick bedded, and contain a large number of regular and parallel joints, greenish lichens in some places, and have concentric flow structure, and parasitic cones, consists generally of alternating andesitic basalt lavas, ignimbrite, tuff and agglomerate. In thin sections, flow texture is prominent. Andesine laths and opacitized hornblende phenocrysts showing parallel arrangement are found in a groundmass composed of plagioclase microlites. In some places, iron - oxide compounds were exposed. As a result of this, the volcanics became reddish.

The Bingöldağı volcanics rest upon the Lower Miocene carbonates with an angular unconformity, in the northwest of Hınıs, but the Alibonca formation conformably and transitionally. The lowermost part of the Bingöldağı volcanics alternates with the terrigenous formation. The Bingöldağı volcanics are overlain by the Pliocene lacustrine formations with angular unconformity. On the basis of these data, the volcanics are most probably of Upper Miocene in age.

Hacıömer formation: The terrigenous formations consisting predominantly of mottled and orange-gray coloured conglomerate were named the Hacıömer formation (Erdoğan, 1966; İlker 1966a). It was regarded as the Kırmızıtuzla member of the Zırnak formation by Erdoğan and Soytürk (1974).

The formation crops out typically in the southern part of Hacıömer village and in the vicinity of Şertefin komu.

The unit consists mostly of polygenic conglomerate and passes into the alternation of conglomerate, sandstone, claystone and mudstone, upwards. The pebbles of conglomerate are well-rounded ill-sorted and poorly-consolidated. The transitions as lateral wedges are observed in this formation as well as poorly, thin-medium thick-thick bedding, gradation changing laterally and vertically and cross-bedding. The gypsum beds are seen in some levels.

According to Erdoğan and Soytürk(1974), the Hacıömer formation which rests upon the Lower Miocene marine carbonates with angular unconformity, shows conformable relationship to the Bingöldağı volcanics. On the basis of the observations, the unit rests upon the volcanics with an angular unconformity, to the west of Hacıömer.

The Hacıömer formation was deposited in fluvial environment. It is of Upper Miocene (Erdoğan and Soytürk, 1974) or Upper Miocene - Pliocene in age (Erdoğan, 1966).

Işıklar formation: The unit, consisting mostly of lacustrine carbonates, crops out typically, in the eastern part of Tekman and in the vicinity of Işıklar village. This unit was named the Işıklar limestone member by Erdoğan and Soytürk (1974). However, since it contains, in places, clastic-pyroclastic rock levels, it was regarded as a formation.

The lacustrine carbonates are dominant rock unit in this formation. The carbonates are yellowish, white and beige coloured, and thin to medium-thick, in places thick bedded. They contain a large number of pores. Cross and graded bedding are common. The unit contains, in places, intercalations of conglomerate, and travertine in the lowermost levels, and intercalations of pebbly and sandy limestone, conglomerate and pyroclastic rocks in the uppermost levels.

The unit rests upon the Bingöldağı volcanics with angular unconformity but the Hacıömer formation conformably. The lower and uppermost levels of the formation reflect the conditions of fluvial environment, whereas the middle levels typically reflect the conditions of lacustrine environment. The formation is of Pliocene in age (Erdoğan and Soytürk, 1974).

The Işıklar formation is overlain by the incesu formation and Karayazı volcanics conformably.

Hamurpet dağı area.— The Alibonca formation which is a terrigenous volcano-sedimentary unit,overlies the Lower Miocene Haneşdüzü formation with angular unconformity, in the vicinity of Hamurpet dağı and near Hınıs-Tekman. The Alibonca formation is overlain by the Zırnak formation and Hamurpet dağı volcanics (Fig. 5).

Zırnak formation: The coaliferous terrigenous formations resting upon the Alibonca formation which has terrestrial character and consists generally of sandstone, claystone and predominantly limestones upwards, were name as the Zımak formation (İlker, 1966b). Soytürk (1973) preferred to use the same name for the Upper Miocene terrigenous formations and named the Pliocene terrigenous clastic rocks as the Bulanık formation and the volcanic terrigenous rocks as the Elmakaya formation. As a result of the observations, the name given by İlker (19666) was preferred for this formation.

The formation crops out typically in the vicinity of Zırnak village and on the slopes of Hamurpet dağı mountain.

Some levels of the formation which is grey, yellowish coloured and consists of sandstone, claystone, marl in the lowermost levels and limestone in the uppermost levels, contain pyroclastic interlayers- and coaliferous formations. The rocks is thin to medium-thick bedded. Locally well developed graded and cross-bedding, oblique sliding structures are the same age with deposition are common in the lowermost levels. The limestones which are dominant in the uppermost levels, contain ooliths and abundant lacustrine macrofossils in some places.

İlker(1966b) and Tütüncü (1967) claimed that this formation rests upon the Alibonca formation, unconformably, whereas Sungurlu (1967) proposed a paraconformity between them. The same relationship was examined again near Zırnak and Alibonca villages and it was concluded that the existing relationship may be a local unconformity, only.

According to Erdoğan (1966) and Sungurlu (1967) and on the basis of mammalian fauna determined, Pliocene age is assigned to this formation (Fig. 5). The formation was deposited in a lacustrine environment.

Hamurpet dağı volcanics: The volcanics which consist mostly of basalt and crop out near Hamurpet mountain, were named as the Hamurpet dağı volcanics.

They crop out typically on the northern and southern slopes of Hamurpet dağı mountain.

They are dark-grey and blackish coloured, unbedded, locally thick bedded and contain locally greenish lichens and a great number of joints intersecting each other. The joints were affected by the movements of the oblique faults. Basalts having prominent flow structure, contain large plagioclase laths (varying from andesine to labradorite) and hornblende and in places pyroxene and olivine.

The Hamurpet dağı volcanics rest upon the Zırnak formation and were probably formed after the formation had been deposited during the Pliocene or Plio - Quaternary.

Correlation of Middle ? — Upper Miocene — Pliocene rocks

The examination and correlation of the units which are exposed in the vicinity of Mescitli, in the area between Hınıs and Tekman, and near Hamurpet dag area, will lead to the following conclusions : The Çullu formation cropping out in the vicinity of Mescitli, probably corresponds to the Alibonca formation cropping out in the area between Hınıs and Tekman. Both formations are terrigenous volcano-sedimentary sequences and probably of Upper Miocene age, and represent the lowermost levels of the younger formations. The Bingöldağı volcanics rest upon these formations, locally and show lateral transitions also.

The Pliocene Işıklar and incesu formations which show conformable relationships to each other, rest upon the Upper Miocene formations,unconformably, in the Tekman-Karayazı basin, whereas the Pliocene Zırnak formation rests upon the Upper Miocene formations, conformably, near Hamurpet dağı area. However, the Pliocene Işıklar formation rests upon the Hacıömer formation ranging from Upper Miocene to Pliocene in age, conformably in the northern basin. Therefore, unless any remarkable deformational and erosional features are recognized, the relationship between the Upper Miocene and Pliocene formations must be regarded as a local unconformity.

The Pliocene Işıklar and incesu formations cropping out in the Tekman-Karayazı basin can be correlated with the Pliocene Zırnak formation cropping out near Hamurpet dağı mountain. For example, while the lacustrine carbonates were dominant at the beginning and fluvial deposits later, in the northern basin, lacustrine conditions prevailed, wholly and lacustrine carbonates predominated towards the end of this deposition, in the southern basin.

The Pliocene formations are overlain by the basaltic Karayazı volcanics in the northern basin, and basaltic Hamurpet dağı volcanics in the southern basin. These basaltic volcanics whose relationships are not seen, directly are probably products of the same phenomena.

Consequently, two significant volcanic stages were differentiated, one being of Upper Miocene age (Bingöldağı volcanics) and one being of Pliocene - Plio - Quaternary age (Karayazı and Hamurpet dağı volcanics).

The above mentioned differences are possibly related to the phenomena that the study area which was developed as a unique basin during the Maestrichtian - Lower Miocene time interval, were subdivided into several basins during the Middle - Upper Miocene - Pliocene.

TECTONIC CHARACTERISTICS OF THE REGION

In the study area, three significant tectonic stages were differentiated, in turn, pre-Maestrichtian-Palaeocene, Late Eocene-pre-Miocene and one that from Middle? - Upper Miocene to the present day in age," and the structures related to these stages were mapped (Fig. 1).

Pre — Maestrichtian — Palaeocene stage

The Akdağ metamorphics constituting the basement, are exposed beneath the Agoren complex as a tectonic window. The ophiolitic complex rests upon the metamorphics, at an angle varying between 30° and 80°. The angle of overthrusting is higher, in the southern part of the metamorphics. But, the metamorphics rest upon the ophiolitic complex, as being bounded by the oblique - slip faults only. The Maestrichtian - Palaeocene sediments rest upon the two units which show tectonic relationships to each other, unconformably. The sedimentary unit contains well rounded pebbles of the formations belonging to the basement. In other words, a definite erosive stage took place during the pre-Maestrichtian. On the basis of these data, it may be concluded that the ophiolitic complex and hence the ophiolites were emplaced in their secondary position, during the pre-Maestrichtian.

Late Eocene stage

In this stage, the Sakaltutan group was thrusted over the Eocene Kösehasan formation, in the southern part of the group. The olisthostromal levels of the Eocene units may be suggested that overthrusting took place during the Eocene. The continuations of the overthrust are covered by the Lower Miocene marine carbonates, in the northern part of Kösehasan and in the southwestern part of Sakaltutan mountain (Fig.l). For this reason, this stage differs from the recent tectonic stage which began during the Middle? — Upper Miocene.

New tectonic stage ranging from Middle - Upper Miocene to the recent

It has been proposed that the Eastern Anatolia is being controlled by compressional regime, from the

Middle? — Upper Miocene to the recent and as a result of this, a crustal thickening took place (Saroğlu et al., 1980; Şaroğlu, 1980). On the other hand, it has been claimed that the compressional and extensional regimes followed each other in the same area (Barka, 1984) or a tectonic regime which took place as compression-narrowing during the Upper Miocene and compression . extending during the Pliocene (Koçyiğit et al.,1985).

The relationships between the Middle? — Upper Miocene terrigenous formations and Lower - Middle Miocene marine formations were examined again. For instance, the Lower Miocene carbonates pass into the Lower? — Middle Miocene Mescitli formation, conformably in the vicinity of Mescitli and the Mescitli formation passes into the Çullu formation, conformably. Nevertheless, the Upper Miocene formations rest upon the Lower Miocene marine carbonates with angular unconformity near Hınıs and Hamurpet dağı area. On the basis of this, the study area was affected by the compressional regime, at least, during the Middle? — Upper Miocene. Extensional forces have probably affected the study area at right angle to compressional forces. It may be concluded that angular unconformity is seen in the areas uplifted under the compressional regime, whereas local conformities are observed in the extensional part of the areas.

In new tectonic stage, the study area was affected by an appoximately N-S striking compression, as in the other parts of Eastern Anatolia (Şaroğlu and Yılmaz, 1984; Şaroğlu, 1986) and as a result of this, approximately NW-SE striking right lateral-slip and NE-SW striking left lateral-slip oblique fault zones took place, as well as E-W trending folds and overthrusts (Fig.l). Along these zones, the northern or southern blocks were uplifted, in places. Therefore, in addition to strike-slip, vertical-slip has occured, also. The fault planes which were developed in the Miocene and older rocks, confirm the oblique character of these faults. The fault traces which were developed on fault planes, make an angle varying between 20° and 45° with a horizontal plane. Some characteristics of the oblique fault zones are presented below.

NW—SE striking right lateral-slip oblique faults.— The Karayazı faults which are observed, in the northwestern part of the study area, and the Kazbel faults and Akdağ faults which are seen in the south of the Karayazı faults, respectively are the most significant faults. A part of the Karayazı fault was originally defined and mapped by Koçyiğit (1985). On the basis of the observations, it has been determined that this fault was separated into a great number of small faults which are disconnected from each other, but following one another, the northern of southern blocks were uplifted and the faults have rather small displacements. The maximum visible lateral displacement of the Kazbel fault which is observed further to tire south, is 4.5 - 5 km (Fig.l). In other words, the rock units lying on both sides of these faults don't show significant differences. So, in the studies of these fault belts, it is not necessary to seperate the rock units as northern or southern blocks of the faults, always.

NE—SW *striking left lateral-slip oblique faults.—* The conjugated faults of the above mentioned faults are observed in the eastern part of Ağcakoca village to the southwest of Sakaltutan mountain, in the western part of Karadağ, around Hamurpet hill, particularly in the northern or southwestern part of the hill(Fig.l).

The fact that some lakes and swamps are arranged in parallel, some streams were displaced, the significant morphological lines are observed and the younger formations are found as being bounded along these faults, may suggest that these faults are active at present and may be moved as a result of energy discharge.

Examination of the relationships between the oblique faults and younger terrigenous formations (Fig.l) and folds and overthrusts which were formed in these formations, may show that both the Hınıs basin and Tekman-Karayazı basin were developed under the control of movements of oblique faults which predominantly have strike slips, as a result of compression, during the Upper Miocene-Pliocene. However, some fault planes and displacements of the faults which are incongruous to the general strike, have been determined in the Lower Miocene marine carbonates and particularly Plio-Quaternary formations and travertines, within some fault zones (Fig. 6A and B). Two small outcrops of travertine which are found.

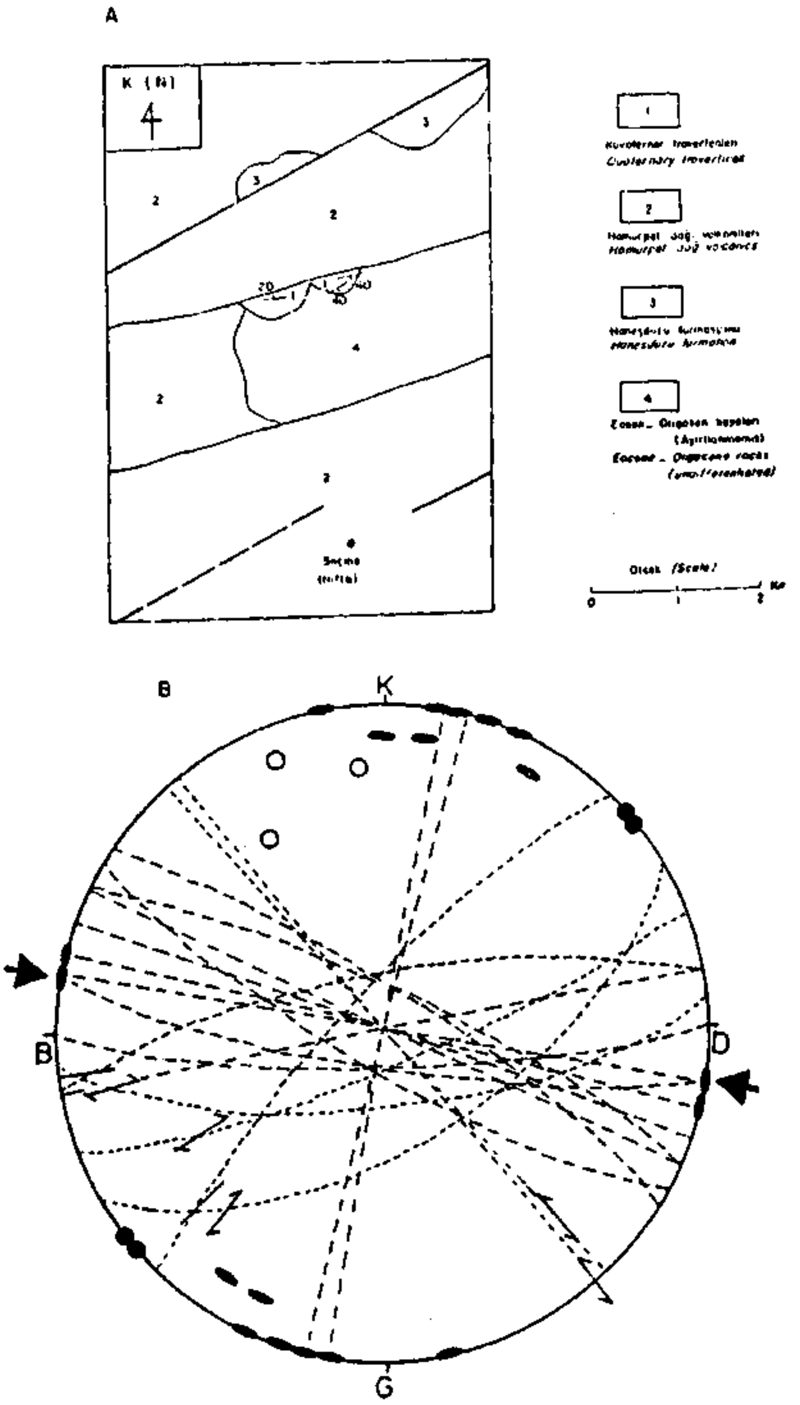

Fig.6— The direction of compression and Schmidt diagram (B) of measured fractures in travertine, to the north (A) of Seçme (Niftik), (18 micro strike-slip faults and cracks have been evaluated).

3 km north of Seçme (Niftik) village, were mapped, one in the east (striking of dip, 140° - 180° and dipping 40°) and another in the west (striking of dip, 350° and dipping 40°), However, evaluation of 18 micro-scale strike-slip faults and cracks in the same outcrops, point out an approximately E-W striking compression (Fig.6B). On the diagram, the short lines represent faults, the long ones cracks, the closed circles left lateral-slip oblique faults, the open ones right lateral-slip oblique faults and the lens-like symbols poles of cracks. In addition, the E-W striking micro-scale normal faults, which have oblique slips also, have been determined in lacustrine carbonates of probably Plio-Ouaternary in age, in Cako upland, 1 km north of Hamurpet lake. These structures are products of E-W striking compression.

Consequently, examination of positions of faults and their relations to the basins, suggests that both the Tekman-Karayazı basin and Hınıs basin were developed under the control of particularly oblique-slip faults and some incongrous structures were formed in the youngest deposits.

CONCLUSIONS

As a result of the studies, the geological map of the area was simplified and made again. The conclusions obtained from these studies are as follows :

The general stratigraphic sequence of the metamorphics in the region, has been determined. In addition, it has been emphasized that the metamorphics had undergone a metamorphism in amphibolite facies and resemble the Central Anatolian massifs (Yılmaz, 1980, 1982; Özcan et al., 1980) on the basis of rock type, the sequence and degree of metamorphism. Furthermore, it has been concluded the metamorphics are exposed beneath the ophiolitic complex as a tectonic window and the both units were cut by acidic intrusions and subjected to contact metamorphism.

The Maestrichtia'n-Pliocene cover, which rests unconformably upon the basement consisting of the metamorphics and ophiolitic complex, is composed essentially of regular transgressive and regressive sequences. This cover started to gain a terrestrial character, as a result of development of final regression during the Middle? — Upper Miocene. Additionally, some data related to the fact that the Ağören complex and hence ophiolites were emplaced in the study area during pre-Maestrichtian, and the transitions between the Maestrichtian-Palaeocene and Oligocene-Miocene rocks, are presented in this paper, for the first time.

The Oligocene levels of the sedimentary cover contain thin interlayersofandesiticbasalt. Additionally, some levels of dacitic and andesitic pyroclastic and epiclastic rocks, and lavas are found in the Middle Miocene formations. The lowermost levels of the Upper Miocene-Pliocene sequence consist of andesite, andesitic basalt and pyroclastics (Bingöldağı volcanics), whereas the uppermost levels consist of andesitic basalt, olivine basalt and pyroclastics (Karayazı and Hamurpet dağı volcanics). So, the Middle-Upper Miocene-Pliocene volcanics varies from dacitic - andesitic volcanics to olivine basalts.

The pre-Maestrichtian overthrusts and Late Eocene-pre-Miocene overthrusts were defined in the study area also.

The NW-SE striking right lateral-slip oblique faults and NE-SW striking left lateral-slip oblique faults were defined as well as the E-W trending folds and overthrusts. These structures were formed as a result of compressional tectonic, during the Late and post-Miocene. The Tekman-Karayazı basin and Hınıs basin were a single basin until the Lower Miocene. The last compressional tectonic regime caused this basin to be separated into two parts during the Middle? — Upper Miocene. Both basins were developed under the control of oblique faults. In addition, some structures were developed locally in the Miocene and Plio-Ouaternary rocks by the effects of the approximately E-W striking compressional forces.

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POST-EOCENE TECTONICS OF THE CENTRAL TAURUS MOUNTAINS

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ABSTRACT.— In post-Eocene time, the Central Taurus mountains have been subjected to four episodes of compression in probably Upper Eocene — Lower Oligocene, Langhian, Upper Tortonian, and Upper Pliocene to recent times. In the Upper Eocene — Lower Oligocene compressional period, Ecemiş, and Beyşehir conjugate faults which have both vertical and lateral components have been formed after a N - S compression. In the Langhian compression period, the Lycian nappes were emplaced from the NW to SE and this tectonic movement has also effected the Antalya and the Adana Miocene basins. In the Upper Tortonian compression period, firstly a WSW-ENE compression has resulted in the formation of Aksu thrust, Kırkkavak oblique reverse fault, Köprüçay syncline, Beşkonak anticline, Radyoring anticline, Taşağıl syncline and Kargı reverse faults. In this period a later phase of N — S compression has formed Çakallar folds, Gökçeler normal fault, the smooth anticline in Mut-Karaman and the syncline in Ulukışla. In the latest compressional period from Upper Pliocene to recent, first on E — W compression which can be recognized by some mesoscopic faults, has been developed and later a N — S compression resulted in the formation of the active faults on Ecemiş and Gökçeler faults, and the Antalya bay graben.

INTRODUCTION

Central Taurus mountains which extend from Ecemiş fault (Blumenthal, 1952), to Antalya Miocene basin and to the western end of Antalya Upper Miocene-Pliocene basin (Akay et al., 1985) ; are not very well known in terms of post-Eocene tectonic characteristics although this part of the Taurus belt has long been subject of intense studies. Structural implications of the field works carried out in 1981 - 1983 have been evaluated together with some tectonic and stratigraphical features related to the other regions in the Central Taurus in order to establish the post - Eocene structural evolution of the region. Evolution described here is mainly based on macrotectonics with few exceptional microtectonic considerations.

Poisson (1977) described the Lycian phase but his data about the Aksu thrust were limited. Dumont and Kerey (1975) studied the Kırkkavak fault. On the other hand, the Beyşehir fault (Akay 1981a; Özgül, 1976), the Ecemiş fault (Metz,1956; Arpat ve Şaroğlu, 1975; Yetiş, 1984), the Ulukışla basin (Demirtaşlı et al .,,1984) and rather smooth anticline in Mut are (Şaroğlu et al., 1983) studied for the evolutionary tectonic history of the dentral Taurus. N-S compression of the Central Taurus following a E-W compression is also discussed in order to contribute the debate on the Anatolian plate movements.

PRE - OLIGOCENE STRUCTURAL FEATURES OF THE REGION

Central Taurus region is formed by several units featuring different stratigraphic, lithologic, tectonic and metamorphic characteristic (Fig. 1). Autochthonous Geyikdağı (Özgül, 1976) rock units range in age from Infra - Cambrian (Dumont, 1978) to Eocene (Özgül, 1976; Monod, 1977). On the other hand, other autochthonous unit, Beydağlan, comprises platform carbonates ranging in age from Jurassic to Miocene (Poisson, 1977). Antalya nappes lying between these two autochthonous units comprise mostly deep sea sediments and are originated from a basin between two autochthons (Poisson et al., 1984). Antalya nappes exposed between the Geyikdağı unit and Alanya metamorphits however, are originated from a basin between these two tectonic units Lycian nappes are transported from the NW to SE and emplaced in the Miocene time (Poisson, 1977), The Bozkır unit - Beyşehir Hoyran nappes, which are composed of pelagic limestones of Triassic - Cretaceous and the ophiolite nappe, are emplaced on the partlj metamorphic Bolkar unit comprising rocks of Devonian - Upper Cretaceous (Özgül, 1976 ;Demirtaşlı, 1983) in the Upper Cretaceous (Özgül, 1984; Tekeli et al., 1984; Demirtaşlı et al., 1984). These four, units are retransported and emplaced on the Geyikdağı unit in Upper Lutetian - Lower Priabonian (Özgül, 1976 ; Monod, 1977). The Niğde massif exhibiting high grade metamorphism is composed of Paleozoic-Mesozoic rocks (Göncüoğlu. 1981).

To the east of Ecemiş, fault, Aladağ nappes comprise an ophiolitic nappe and sedimentary rocks ranging in age from Devonian to Cretaceous (Tekeli etal.,1984).

In the Aladağ region, thin, volcanics-free Lutetian deposits transgressively overlie the basement rocks (Yetiş, 1984). On the other hand, to the west of Ecemiş fault, a thick volcano - sedimentary sequence of Upper Maastrichtian - Middle Eocene age, which is quite different from the eastern sequence, is widespread (Oktay, 1982; Demirtaşlı et al., 1984).

Furthermore, to the south of Ulukışla basin, the Bolkar unit is thrusted northwards over the Paleocene - Lower Eocene deposits where an Upper Lutetian sequence covers the thrust contact uncomformably (Demirtaşlı et al., 1984).

POST - EOCENE TECTONIC FEATURES

Central Taurus belt is subjected to four compressional episodes in Upper Eocene - Oligocene, Langhian, Upper Tortonian and Upper Pliocene.

In the Upper Eocene - Oligocene compressional period, Beyşehir and Ecemiş, faults moved considerably. The lack of any sedimentary sequence indicative of a basin existed before the Upper Oligocene - Miocene basin, shows that the region was uplifted during this compressional period.

In the Langhian period, Lycian and Davraz mountain thrusts, which are covered by the Tortonian sediments, were developed.

In the Upper Tortonian period, the whole region was subjected to an intense compression and gained important structural features. In the post - compressional stage, Messinian:- Pliocene deposits of the Antalya basin covered, these structural features.

The Upper Pliocene-recent compressional period which is defined by some mesoscopic faults exhibits a rather weak activity in the Central Taurus as a whole.

Ecemiş fault

This fault zone, which is also known as Ecemiş. corridor in the literature, is said to have an important lateral offset (Blumenthal, 1952).

Yetiş, (1984) reports that the Ecemiş fault zone has been developed in post-Paleocene but pre-Lutetian times. However, his data to omit the probability of the development of the fault zone in later times is not sufficient.

To the east of Ecemiş fault, there exist a nonmetamorphic ophiolite and several sedimentary nappe.s formed by disintegration of a platform (Tekeli et al., 1984). Overiying these nappes in a limited area, is the sedimentary sequences of the Lutetian transgression (Yetiş, 1984). On the other hand, to the immediate west of the fault, there are, high grade metamorphic Niğde massif (Göncüoğlu, 1981), low grade metamorphic rocks of Bolkar mountain and an Upper Cretaceous ophiolitic melange (Demirtaşlı et al., 1984). Volcano-sedimentary rocks of Ulukışla basin ranging in age from Upper Maastrichtian to Lutetian overlie both Bolkar mountain metamorphics and the ophiolitic melange (Oktay, 1982; Demirtaşlı et al., 1984). Juxtaposition of metamorphic and nonmetamorphic units nessecitates several tens of lateral offset along the fault zone. Additionally, thin, volcanits free Lutetian deposits (Yetiş, 1984) in the east completely differ from the basinal characteristics of the western Upper Maastrichtian - Lutetian volcano - sedimentary sequences (Oktay, 1982; Demirtaşlı et al., 1984). This situation, therefore, implies that the basins may be juxtaposed by virtue of the fault movement. That is to say that a lateral offset of tens of kilometers should have occurred in or post-Upper Eocene times.

Metz (1956) proposed that the structural evolution of the Ecemiş, corridor occurred after depo-

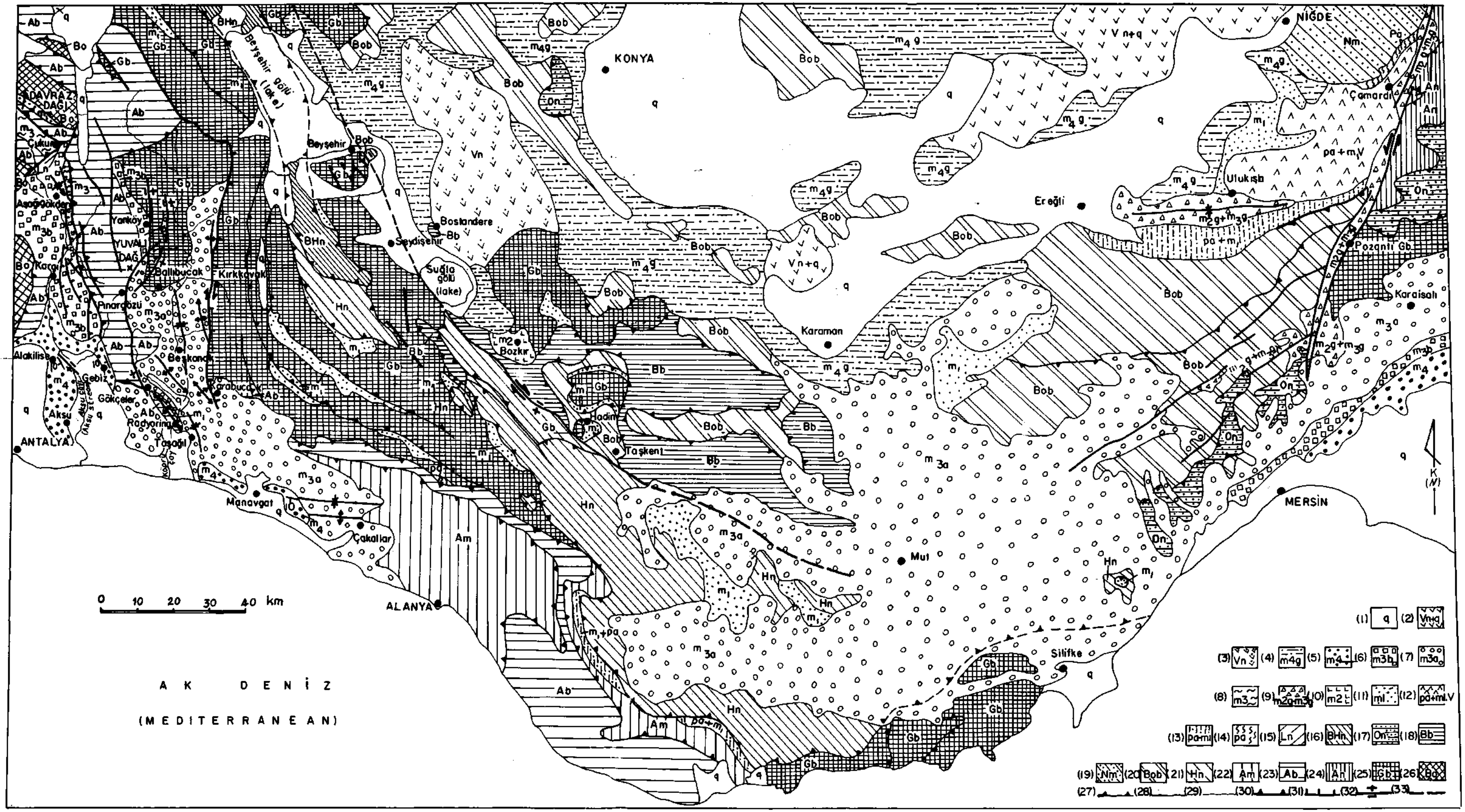

Fig.1 – Geologic map of the Central Taurides. 1 – Quaternary deposits; 2 – Neogene-Quaternary volcanits; 3 – Neogene volcanits; 4 – Continental Pliocene deposits; 5 – Marine Pliocene deposits; 6 – Tortonian deposits; 7 – Miocene deposits; 8 – Lower-Middle Miocene deposits; 9 – Oligocene-Miocene deposits; 10 – Oligocene deposits; 11 – Eocene deposits; 12 – Paleocene-Eocene volcanits; 13 – Paleocene-Eocene deposits; 14 – Paleocene deposits; 15 – Lycian nappes; 16 – Beyşehir-Hoyran nappe; 17 – Ophiolite nappe; 18 – Bozkır unit; 19 – Niğde massif; 20 – Bolkar unit; 21 – Hadim nappe; 22 – Alanya metamorphits; 23 – Antalya unit; 24 – Aladağ nappes; 25 – Geyikdağı unit; 26 – Beydağları autochthon; 27 – Pre-Oligocene nappe contact and overthrust; 28 – Pre-Oligocene faults; 29 – Contact; 30 – Post-Eocene overthrust; 31 – Post-Eocene reverse fault (A down; Y, up); 32 – Post-Eocene normal fault (+) upthrow, (–) downthrow block; 33 – Post-Eocene covered structures. (General geology : Göncüoğlu, 1981; Tekeli et al., 1984; Demirtaşlı, 1983; Demirtaşlı et al., 1984; Gedik et al., 1979; Demirtaşlı, 1984; Özgül, 1984; Monod, 1977; Akay, 1981 *a*; Akay and Uysal, 1985; Akay et al., 1985; Gutnic et al., 1979; Poisson, 1977; Akbulut, 1977; Dumont et al., 1980.)

sition of the Cretaceous sediments but before the Paleogene deposition and Oligocene sediments covered the structural imprints. These Oligocene sediments are equivalent to the Kılan group of Oktay (1982) which are continental sediments overlying the Ulukışla formation (Paleocene - Upper Eocene) with an angular unconformity. Oligocene sediments are dated according to that of Blumenthal (1955) proposed which is Chattian-Aquitanian, Furthermore, Oktay (1982) reports that sedimentation is likely to be continued even in the Lower Miocene. According to these data, the Ecemiş, fault zone should have experienced its major activity at least before the deposition of the Oligocene - Lower Miocene basinal sediments.

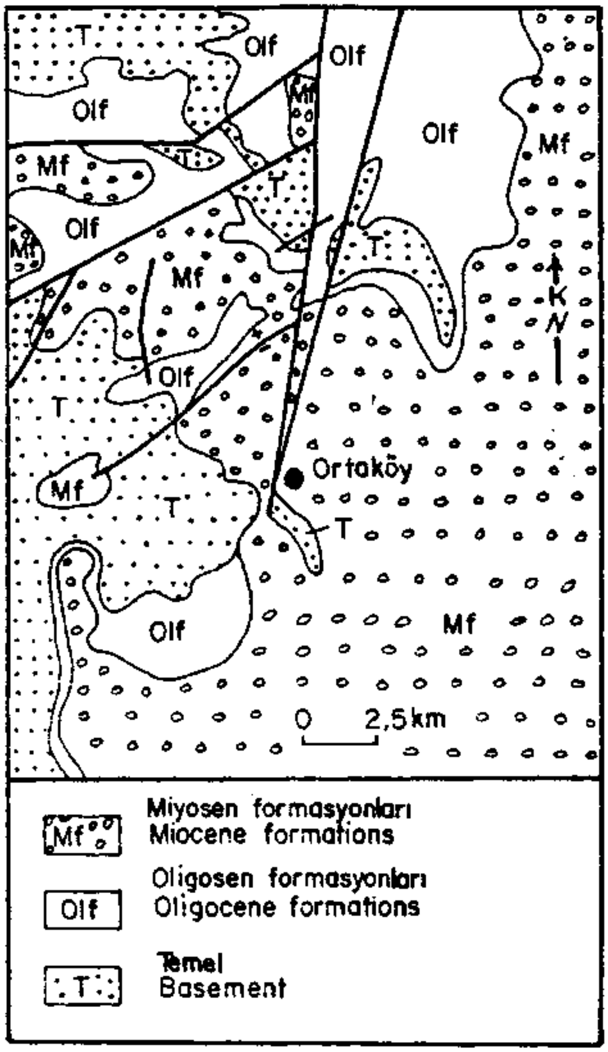

Fig.2 - Geologic map of the vicinity of Ortaköy (Demirtaşlı, 1983).

In the geological map of Ortaköy region to the south of Ecemiş, fault (Fig. 2), the fault activity is seen to be confined in the shallow-marine Miocene limestone formations. Activity of the fault could be more intense in the north. In this case, the major movement of the fault should have occurred before the deposition of Oligocene formations which are overlain conformably by the Miocene formations.

On the other hand, the Miocene formation in the Ulukışla basin are effected by this movement. The Ecemiş, fault which is oblique to the E-W trending fold (Demirtaşlı et al., 1984) in the Aktoprak formation, dated by Blumenthal (1955) as Chattian-Aquitanian, should have been reactivated dextrally deforming the Miocene formations during this folding. In fact, in Figure 2, the Miocene formations to the south of the fault are seen to be deformed.

Along the fault zone, 10 km to the east of Çamardı, an active fault is found to be lying in the fault trend direction for 10 km. The fault plane is dipping towards the mountains and its mountain side is downthrown without any indication of movement direction (Arpat and Şaroğlu, 1975). Therefore, the Ecemiş, fault zone is a still active zone.

Beyşehir fault

Beyşehir fault which extends from Beyşehir to Taşkent — Mut in Central Taurus, is an important structural element of the Upper Eocene - Lower Oligocene compressional period.

The fault juxtaposes allocthonous and autochthonous lithologies in Beyşehir region (Fig.3)ln the Konya sheet of the 1:500.000 scale geological map of Turkey, around Bostandere 10 km to the north of Seydişehir, the map unit which is shown as of Dogger-Malm, is composed of Mesozoic pelagic limestones and green volcanics. They should belong to the Bozkır unit (Özgül, 1976). Beyşehir fault should, therefore, be between these units and the autochthone to the south of Seydişehir. On the other hand, the normal fault which was located in the south of Bozkır - Hadim-Taşkent by Özgül (1976) is the prolongation of Beyşehir fault.

In Beyşehir region, the important offset of the fault is defined by nearly side by side position of the nappes (Monod, 1977) on the Eocene deposits and the autochthonous Ordovician sediments. The Cambrian dolomites of the autochthone extends for about 4 km to the west of another fault to the west of the Beyşehir fault (Fig. 3). However, there is no Cambrian dolomite between these two faults. This, henceforth, shows that the autochthonous slice between the faults has been transported there from elsewhere at a distance of at least 4 km. On the other hand, in the Hadim region cross section of Özgül (1976), the northeastern block of the fault is seen to be considerably downthrown.

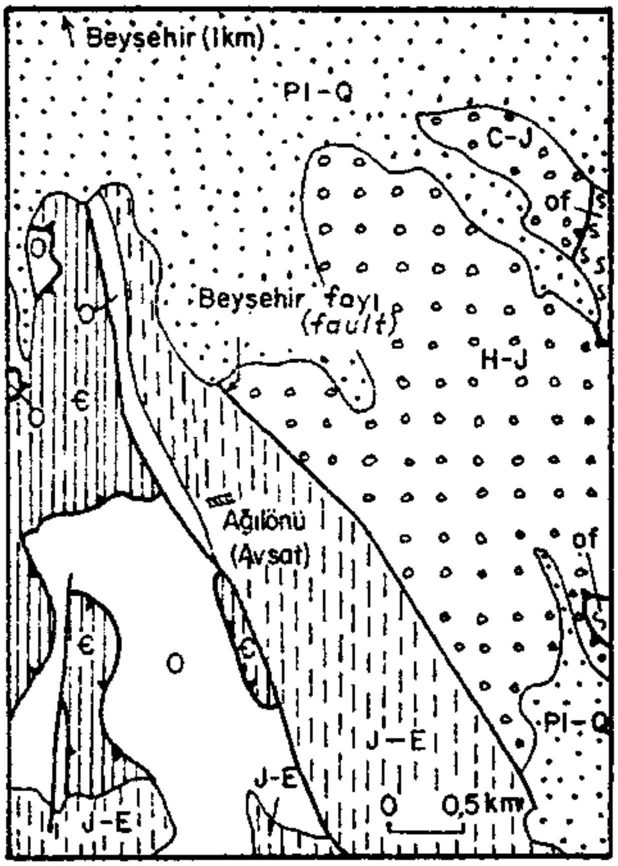

Fig.3— Geologic map of south of Beyşehir : Autochthon: Cambrian dolomite; O- Ordovician detritus; J-E- Deposits from Jurassic to Eocene. Allochthonous : H-J- Deposits from Carboniferous to Jurassic; Of- Ophiolite; PL-Q- Pliocene - Quaternary deposits.

The fault, in Beyşehir region, has been developed after the emplacement of Beyşehir - Hoyran nappes in Upper Lutetian - Lower Priabonian (Monod, 1977) and thus, juxtaposed allochthons and autochthons by its normal slip. The fault itself is covered by Upper Pliocene deposits (Blumenthal:, 1947; Akay, 1981a). Therefore, the fault has been formed during/after the Upper Eocene and before the Upper Pliocene. In the Konya sheet of 1:500,000 scale geological map of Turkey, Miocene limestones cover the fault 14 km to the southeast of Taşkent (N. Özgül 1980 pers. comm.; A.Z. Bilgin 1986, pers. comm.). Actual age of the Miocene deposits has been reported to be (Gedik et al., 1979) as Langhian-Serravalian in the south. Thus, the fault has been developed before the Langhian in this region.

Emplacement of the Lycian nappes

Lycian nappes, in Isparta region, have been emplaced onto Beydağları and Antalya nappes from the northwest to southeast and they are later covered transgressively by Tortonian deposits (Gutnic et al., 1979). In the same study, it is proposed that the Aksu thrust passes through the south of Kargı - Çukur - Davraz and Isparta region.

However, in the geological map of Çukur region (Fig. 1) the Tortonian deposits of Antalya Miocene basin cover the Davraz mountain sediments, Beydağları Miocene basin deposits and the Lycian nappes. This shows that the Aksu thrust does not continue northwestwards after Çukur. Therefore, the thrust fault passing through the northern end of Beydağları Miocene basin deposits has not been developed as a consequence of Aksu fault as proposed by Gutnic et al. (1979) but as a consequence of Lycian nappe emplacement from the northwest to southeast. Davraz mountain is, thus, partly carried together with the Lycian nappes and emplaced on the Langhian sediments (Poisson, 1977) of Beydağları Miocene basin (Akay et al., 1985).

Kırkkavak fault

Kırkkavak fault (Dumont and Kerey, 1975) is a N — S trending tectonic line with which the Geyikdağı unit (or Anamas — Akseki autochthone; Monod, 1977) Mesozoic sediments and Antalya Miocene basin (Akay et al., 1985) sediments are juxtaposed. The fault terminates to the south of Kızıldağ before reaching the sea.

Field observations reveal that the Mesozoic sediments overlie Antalya Miocene basin deposits with a high angle fault plane (Fig. 1). There are en-echelon folds has a southwards knee shape, the middle one is symmetrical and the northern fold has high angle flanks. The southern and middle folds make an angle of 35° to the fault whereas the northern fold makes 10°. There is another fold to the east of the fault in Karabucak extending for 2 km parallel to the fault.

On the other hand, to the immediate east of Kırkkavak fault on the northern flank of Yarımkavak stream passing through Kırkkavak, there are north vergent Miocene polygenetic deposits transgressively overlying the Triassic elastics. These deposits have nothing to do with the monogenetic lithology which was reported as the basal breccia of the Miocene basin by Dumont and Kerey (1975). Therefore, the breccia levels should be older.

The fault is formed by the tectonics which is responsible for termination of the sedimentation in Antalya Miocene basin in Tortonian (Akay et al., 1985). Therefore, as a consequence of the tectonic regime (ENE-WSW compression) which is responsible for the development of the Köprüçay folds, the fault has been developed as an oblique fault with both right lateral strike - slip and reverse components.

Dumont and Kerey (1975) report that the fault initially, in or during the Burdigalian, has moved with a right lateral strike - slip and then, in Tortonian, has been subjected to compression.

Dumont and Kerey (1975) assume the brecciated limestones with monogenetic fragments observed along the Kırkkavak fault as the basal lithology of the Miocene basin. The right lateral movement of the fault is proposed according to the structural study of this breccia level. However, as it is indicated before, this level should be older. Therefore, the tectonics effected the breccia level may be older than the proposed age. On the other hand, transgressive occurrence of the Miocene deposits on the Triassic elastics necessitate a considerable erosion of the Jurassic - Cretaceous - Paleogene sediments of the Geyikdağı unit before the Miocene transgression. However, except the Upper Lutetian - Lower Priabonian tectonics (Monod, 1977), any important tectonic episode which could be responsible for this erosion is not known yet in this region. Furthermore, the position of the Kırkkavak fault (Fig. 1) which is compatible with the Upper Lutetian - Lower Priabonian macrostructures (Monod, 1977) indicates that the development of the right lateral movement in Upper Lutetian - Lower Priabonian is very likely if it is compared to the Dumont and Kerey's (1975) proposal.

Occurrence of the Triassic elastics on the Miocene deposits with a high angle fault plane, compatible position of the overturned Miocene deposits along Yanmkavak stream and existance of a 2 km long fold parallel to the fault indicate the reverse movement along the fault. The en-echelon folds related to the fault development are, in turn, indicative of a strike-slip component. Therefore, the Kırkkavak fault which lies at a high angle to the compression direction, should have been active as an oblique fault with a right lateral strike-slip and a reverse components. The compression direction, deduced from the Köprüçay folds, is from ENE to WSW.

Köprüçay folds

The Köprüçay folds comprise Köprüçay syncline, Beşkonak anticline, Radyoring anticline and Taşağıl syncline (Fig. 1).

Radyoring anticline has an axis plunging in both directions and high angle flanks. Its axial trend is NW - SE for 15 km. There are also some small scale folds observed along its continuity towards the sea.

The Taşağıl syncline is 13 km long in NNW-SSE direction. Its eastern flank is gentle whereas the western flank is steep.

The syncline along Köprüçay is an approximately N-S trending, 30 km long and a rather flat syncline.

If the axial trends of all these folds are considered, an ENE-WSW compression can be deduced for this region. The tectonic features mentioned above are, in turn, the results of the tectonic episode which destroyed the Tortonian deposits of the Antalya Miocene basin (Poisson, 1977).

Aksu thrust

The thrust lies along a paleosuture zone between the Geyikdağı unit and the Beydağları autochthon, from which the Antalya nappes are believed to have been originated (Poisson et al., 1984).

Gutnic et al. (1979) report that the Aksu thrust, passing through Kargı - Çukur - Isparta region, has been developed with a left lateral movement. However, after some macroscopic observations, it is now known that its extention and the nature of its movement show some differences than previous descriptions.

The thrust can be well traced from Gökçeler in the south to Çukur in the north. It terminates in the south of Gökçeler and does not continue southwards anymore. Gutnic et al,(1979) join the Aksu thrust to another thrust fault in the south of Davraz mountain to the immediate north of Çukur. However, as it can be easily seen on the map (Fig. 1), the Davraz mountain thrust is covered by the Tortonian deposits of the Antalya Miocene basin (Akay et al.,1985) and the Aksu fault which deforms these deposits continues towards north.

The thrust, as observed in Fig. 1, has been developed together with the other thrusts and reverse faults. Between Gebiz and Gökçeler, the rock sequences belonging to the Antalya nappes occur as thrusted over the Antalya Miocene basin sediments. In the south, there is at least 10 km of transportation. It is likely, although without significant data, that the fault is transformed into a reverse fault towards north. The oblique position of a NW-SE fold in Asağı - gökdere in the north indicates that the thrust, in this region, probably moved with a right lateral component. As a consequence of the compression from ENE to WSW which has formed the Kırkkavak fault and the Köprüçay folds, the Aksu thrust has been developed. Existence of a northwards right lateral component is probable. In the light of this interpretation, a left lateral component of the thrust, as proposed by Gutnic et al. (1979), can not be considered.

Kargı reverse faults

In the west of Aksu thrust, the lower levels of the Antalya Miocene basin are of conglomerates while in the upper levels they are of sandstone-siltstone alternation (Akay et al., 1985). Therefore, lithological difference and together with bedding position is an important criterian in distinguishing the reverse faults. There are five reverse faults distinguished in Kargı region. There is a 2 km long peridotite exposure along the fault plane in the reverse fault to the north of Gebiz, Antalya nappes lithologies are exposed beneath the hanging wall. Furthermore, in the southernmost parts of this fault, garnet schists together with serpentinite and the other rock units belonging to the basement.

In Kargı region, the thickness of the Antalya Miocene basin deposits is at least 1500 m. Exposure of the basement rocks beneath such a thick cover after reverse faulting necessitates a considerable vertical displacement.

Therefore, the Kargı reverse faults are structural features lying in NNW—SSE direction and which are formed by deformation of Tortonian deposits (Poisson, 1977) and whose eastern blocks are uplifted.

Çakallar folds

An anticline and a syncline are observed between Manavgat and Çakallar (Fig. 1). Although the folded Miocene formations are eroded at a great extend, a folding in the order of 25 km is observed.

An anticline of flat and symmetrical folds has a burried prolongation up to the south of Manavgat (Canadian Superiour Oil Ltd., 1973). This, in turn, indicates an E - W trending folding for at least 50 km.

Folding should have developed in the Antalya Miocene basin after the basinal sedimentation conditions which lasted up to Tortonian (Akay et al., 1985). Also, the Upper Miocene - Pliocene basin deposits in the western continuation of the folds (Fig. 1) form a monoclinal towards the south. The lack of the effects of Çakallar folds in these deposits indicates that the Pliocene sediments should have deposited after the folding. Therefore, the Çakallar folds are structures showing considerable E-W elongation and which are formed before the deposition of Upper Messinian - Upper Pliocene sediments of the Antalya Neogene basin.

G6k9eler fault

This vertical fault, trending north to south, passes through Gökçeler and Pınargöz and extends to Yanköy (Fig.l).

The Miocene deposits in Gökçeler, to the east of fault was approximately displaced downward 1000 m. Along the footwall block, some small discontinues Miocene deposites can be traced laterally (Akay and Uysal,1985). To the east of Yuvalı mountain on the north; this fault has been formed as the result of hundreds of meters downward displacement of the eastern block drawing a sharp boundary between basement rocks of Yuvalı mountain and Miocene deposites. The fault to the west of Ballıbucak, in and to the east of Yanköy, however, show different features. Here, the normal faults indicate that the western side of the fault is downthrown.

There is even a fault plane, situated on the extension of fault 4 km east of Yanköy, trends N60°W and dips62°southwest. The latter can be observed in a very young scree deposits located on the western side of Sarpdağ and 1 km south of Katrandiş, area at the elevation of 1700 m. The slickensides on the fault plane make angles of 10° — 20° with the dip direction. There is also a possibility that all the faults east of Yanköy have been active during Quaternary.

North - south trending Gökçeler fault has been generated by the deformation of the Tortonian deposits in the Antalya basin (Akay et al.,1985). This fault dips to the west on the north, and to the east on the south and has been active locally on the north.

Gebiz - Aksu monocline and Manavgat monocline

These monoclines can be seen in the Upper Messinian and Upper Pliocene deposits of the Antalya. basin (Akay et al.,1985). The Gebiz - Aksu monocline dips 5° - 10° to the southwest around Alakilise, Gebiz and Aksu. The fault passing through Alakilise indicates that northwestern side of the break has displaced* downward 50 to 100 meters. The Pliocene deposits around and especially to the east of Manavgat dip 5° - 10° to the south. These structural features do not give much clues about the tectonic which results in the development of Upper Pliocene regressive deposites (Akay et al.,1985). Around Gebiz some small scale strike - slip and thrust faults in the Messinian limestone, however, indicate east - west compressional strain (Fig. 4). According to the micro-tectonic studies of Dupoux (1983) ; there is a young east - west shortening in the large area to the north of Antalya Upper Messinian - Upper Pliocene basin. This latter compressional event is younger than the Tortonian tectonic resulted in Aksu thrust. This contractional province also accounts for the formation of the regressive beds of the basin deposits during the Upper Pliocene time (Akay et al.,1985). On the Aksu - Antbirlik road, small scale faults in the Upper Pliocene deposits, on the other hand, show north-south extension (Fig.4). The area around Aksu and probably in the sea basin to the south indicate that this region developed under north-south extensional regime during the late stage of contractional province.

All the Quaternary deposits in the Antalya Neogene basin are, however, flat and do not give any compressional clues.

TECTONIC EVOLUTION

The compressional strain of post- Lutetian in the Centra] Taurus mountains might be divided into four main phases; namely a Late Eocene-Early Oligocene, a Langhian, a Late Tortonian and a Late Pliocene to present.

The compressional cycle of Late Eocene-Early Oligocene

During this compressional period, Beyşehir and Ecemiş faults, which are now known, have been formed and all the Central Taurus region have been influenced.

One of the most important fault of this cycle is Ecemiş fault and the movement on it has occurred in different times.

In or post-Eocene, before the deposition of Oligocene-Miocene basin, this fault must probably

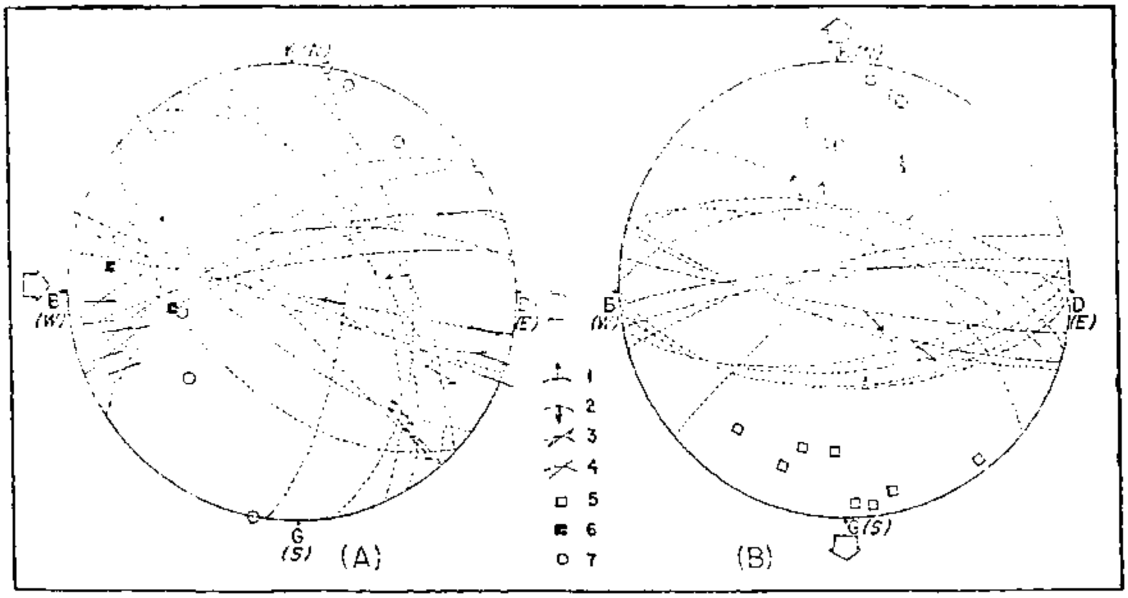

Fig.4 – The fault plane diagrams have been measured in the Messinian limestone on south-east of Gebiz (A) and in detritic deposits of the Upper Pliocene on the Aksu Antbirlik road (B).
1 – Normal fault; 2 – Reverse fault; 3 – Sinistral fault; 4 Fault line; 5 Normal fault pole trace; 6 Reverse fault pole trace; 7 – Sinistral fault pole trace.

have experienced its major sinistral displacement as the first movement. Beyşehir fault, extending along Beyşehir - Taşkent, started to move just before the deposition of Langhian sequence and had at least 4 kilometers lateral movement after the emplacement of Upper Lutetian - Lower Priabonian Beyşehir-Hoyran nappes. The geometric position of Ecemiş, and Beyşehir faults look like they are conjugate faults and they must have moved in the same kinematic. The explanations given above may, therefore, indicate that these faults have been formed by the north-south compression. This contractional regime appears to have developed after the nappes emplacement stage occurred during the Late Lutetian-Early Priabonian times and might be considered as the result of Late Eocene-Early Oligocene orogenic phases which do not have any clues in the study area. Furthermore, the thrusting to the north, covered by Upper Lutetian sequence and on the south of Ulukışla basin, seem to have developed during this time as northerly removement (Demirtaşlı et al, 1984).

Beydağları, Antalya, Mut, Adana Miocene and Ulukışla Oligocene-Miocene basins could be generated just after the relaxation period of Late Eocene-Early Oligocene phase.

Langhian compressional period

This contractional cycle results in emplacement of Lycian nappes and has also affected Beydağları Miocene basin completely (Akay et al. 1985) and Adana Miocene basin partly (Yalçın and Görür, 1984).

The deposition of Beydağları Miocene basin lasted up the Upper Langhian (Poisson, 1977) by the emplacement of Lycian nappes (coming) from north-west to southeast recarried Antalya nappes and replacement of Davraz mountain sequence. During this time, the deposition was continued in Antalya Neogene (Akay et al, 1985), possibly in Ulukışla Oligocene - Miocene (Oktay, 1982; Demirtaşlı et al, 1984) and Mut Miocene basins (Gedik et al, 1979). The deposition in Adana Miocene basins, however, stopped partly (Yalçın and Görür, 1984).In other ward, Langhian phase affected Beydağları Miocene basin completely and Adana Miocene basin partly while deposition was continued in other basins.

After the Langhian phase, affected the area partly, a relaxation cycle developing Antalya Miocene (Akay et al.,1985) and Adana Miocene basins (Yalçın and Görür, 1984; Gürbüz et al,,1985) and deposited transgressive Tortonian deposits. During this time, the deposition (conditions) were continued to the east of Antalya Miocene basin (Akay et al, 1985), in Mut, in Ulukışla and in the south of Adana basin (Yalçın and Görür, 1984).

Upper Tortonian compressional period

The deposits Antalya Neogene basin, caused by the relaxation cycle of Serravalian (Poisson, 1977) — Early Tortonian, were deformed by the strong Late Tortonian phase. The WSW-ENE compressional system has resulted in the formation of Kırkkavak oblique - slip reverse fault, Köprüçay syncline, Beşkonak anticline, Radyoring anticline, Taşağıl syncline, Aksu thrust and Kargı reverse faults. During the late stage of east-west compression, this area has been under north-south working contractional regime and has, therefore, started to develope Çakallar folds and Gökçeler fault. East-west extending folds of Oligocene-Lowpr Miocene deposits in Ulukışla basin (Demirtaşlıet al, 1984) and large scale smooth fold in Miocene limestone (Şaroğlu et al, 1983) around Mut-Karaman were developed during this time. Meanwhile, Ecemiş fault has started to move.and deformed Miocene deposits partly on the south.

After the Upper Tortonian phase, the area has partly been under the relaxation cycle. During this time, transgressive basine sequence has deposited in Antalya Upper Messinian-Pliocene (Akay et al, 1985) and in Adana Pliocene basin (Yalçın and Görür, 1984; Evans et al.,1979). Continental deposits has accumulated on the north of Central Taurus as the result of this period (Fig.l). Upper Miocene - Pliocene volcano-sediments has also developed around Hoyran lake area (Koçyiğit, 1983).

Upper Pliocene to recent compressional period

Upper Pliocene deposits in Antalya basin (Akay et al., 1985) dip 5° to 10° to southwest (Fig.L). Pliocene sequence around Manavgat also dip 5° to 10° to south. Furthermore, observed mesoscopic faults in the Gebiz limestone, which is the base of Upper Messinian-Upper Pliocene deposits in Antalya basin, indicate east-west compressional system. Dupoux (1983), on the other hand, points out that there is a east-west compressional system, which is younger then Tortonian phase, in the northern side of Antalya basin. It is, therefore, possible to conclude that after the relaxation cycle developed during post-Tortonian events, there has been a rather weak E-W compressional deformation influenced the Central Taurus area in Late Pliocene time. During the later stage of this young compressional event, the area has been under the north-south working system. As a result of these, some recent movements have occurred on the Gökçeler and Ecemiş faults and have formed a north-south trending graben in Antalya basin (Özhan, 1983).

Besides, Evans et al (1979) indicates that the southerly thrusting of Messinian evaporites and knee-like folds within the zone, which are controlled by the east-west trending structural features in Adana basin south of Silifke, have been developed as the results of the sliding events within evaporite bedding planes.

In summary, the Central Taurus and the sea basin between Anatolia and Cyprus have been under N — S compressional system since Late Pliocene as indicated before by McKenzie(1972).

RESULTS

Ecemiş, and Beyşehir faults are two conjugate faults, and their first main movements occurred during the Late Eocene - Early Oligocene orogenic phase.

Lycian nappes were emplaced onto Beydağları in Langhian (Poisson, 1977), and these events affected Antalya and Adana Miocene basins weakly.

During the Late Tortonian compressional period, firstly east northeast-west southwest compression has resulted in the formation of Aksu thrust, Kırkkavak oblique-slip reverse fault and the other structures associated with it as first events. Şengör (1980), on the other hand, indicates that the Central Anatolia has been under east-west compressional system because of the contractional province in Eastern Anatolia.

Antalya Miocene basin has been affected as the result of this compression. Furthermore, North Anatolian Fault (NAF) and East Anatolian Fault (EAF) appear to have developed during this time due to observed convergence in Central Anatolia. After a ENE — WSW compressional regime in Central Taurus, a N — S (working) compression has affected this area and has resulted in the formation of Çakallar fold, smooth fold in Mut (Şaroğlu et al, 1983) and syncline in Ulukışla (Demirtaşlı et al, 1985).

After the short relaxation cycle of post - Tortonian, the deformation in the area, has been east-west compressional again, but rather weak activity. During this time, NAF and EAF, which were formed in Late Tortonian, have experienced their second movements. The latest stage of the east - west compression in Antalya Neogene basin points out that all the area have been under north-south working system. As the result of this north-south compression, some recent movements have occurred on the Gökçeler and Ecemiş, faults, and formed a graben in Antalya bay (Özhan, 1983). It is, therefore, concluded that the Central Taurus region is now north-south compressional system as proposed earlier by McKenzie (1972).

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STRATIGRAPHY AND TECTONICS OF THE LATE CRETACEOUS - PLIOCENE SERIES IN THE NORTHWEST OF ANKARA(BAĞLUM - KAZAN)

Ergun GÖKTEN*; Nizamettin KAZANCI* and Şükrü ACAR**

ABSTRACT.- The series outcropped between Bağlum and Kazan towns in the northwest of Ankara are the sandstone rich flysch representing the interval of Late Campanian - Maestrichtian, fluvial and limnic sediments of Paleocene epoch, shallow marine limestones of Lutetiari age, and limnic and fluvial type continental series of Late Eocene and Mid Miocene period. The only uncomformity in the series of Late Campanian and Miocene is seen between Maestrichtian and Paleocene in the area. The continental occurrences of Pliocene, cover the older units by an angular uncomformity. The formations representing the period of Late Campanian and Miocene had folded together related to the continental collision commencing by Mid Miocene. The region bears the features of a fore-arc depositional basin in the mentioned period.

AGE OF THE MURMANO PLUTON AND ITS RELATIONSHIP WITH THE OPHIOLITES

H.P.ZECK*** and Taner ÜNLÜ****

ABSTRACT.— The composite Murmano pluton (a few km NNW of the town of Divriği, province of Sivas, Eastern Central Anatolia), ranging from quartz-syenitic to dioritic in composition, is intrusive into serpentinhes belonging to the Divriği ophiolite complex. The pluton gives a 110±5 Ma (IV) Rb-Sr whole rock isochron date which is interpreted to represent the age of the intrusion. A series of 7 samples, one micro quartz syenite, 5 monzonites and one hydrothermal scapolite rich dyke rock, together representing a c. 100 by 200 m large area at the southern margin of the pluton, defines a 112 ±8 Ma(IV) Rb-Sr whole rock isochron with an initial ^{87}Sr/^{86}Sr ratio of 0.7068. Another series of 7 samples representing ac.200 by 500 m large area at the southwestern margin of the pluton gives a 5 point 109±5 Ma (IV) isochron with an initial ^{87}Sr/^{86}Sr ratio of 0.7058. The two concordant dates are interpreted as the age of intrusion which is thus given as 110±5 Ma (IV). The two sample series represent two separate magma batches which may have originated by anatexis of different source rock complexes with different Rb-Sr isotopic ratios. The predominantly silicic character of the pluton combined with its considerable size c.25 km excludes an origin in the oceanic realm. The magma was intruded subsequent to the obduction of the ophiolite complex which is thus suggested to have taken place more than 110±5 Ma ago.

Mineral Res. Expl. Bull., 108. 35 42.1988

HELIUM - 3 DISTRIBUTION IN WESTERN TURKEY

Nilgün Güleç*

ABSTRACT . — In this study, to investigate the mantle-crust interaction in Western Turkey, geothermal fluids from various locations are analyzed for their ^{3}He/^{4}He ratios. The results reveal a mixing between mantle - and (continental) crustal - helium components. The distribution of mantle-helium, which is characterized by high ^{3}He/^{4}He ratios, does not show any correlation with the spatial and/or temporal distribution of the surface volcanics in the region, but appears to be governed by the distribution of tectonic deformation. The lack of any correlation between the distribution of mantle-helium and surface volcanism suggests that helium, now degassing from mantle, has most probably entered the crust in association with the melts emplaced at deeper levels. The fault systems of the present extensional tectonics are thought to have had an efficient role in the escape of helium to the surface through the brittle parts of crust.

INTRODUCTION

The determination of ^{3}He/^{4}He ratios in continental fluids provides valuable contributions to the understanding of mantle - crust interactions.^{3}He is essentially primordial, that is to say, trapped in the Earth's interior at the time of its accretion from the solar nebula (Ozima and Podosek, 1983). Therefore, the presence of any excess ^{3}He at the surface indicates that volatiles have been transported from mantle. Mantle-helium is characterized by ^{3}He/^{4}He ratios of about 10-^{5} (Mamyrin and Tolstikhin, 1984). On the other hand,^{4}He is produced by the a -decay of ^{238}U, ^{235}U and ^{232}Th. Since U and Th are mostly concentrated in the continental crust relative to mantle and littlte3He is retained in the crust, continental crust is enriched in ^{4}He and (continental) crustal - helium is characterized by^{3}He/^{4}He ratios of about 10-^{7} - 10-^{8} (Andrews, 1985). The isotopic composition of helium in the atmosphere is rather constant over the globe (^{3}He/^{4}He =1.4 xlO-^{6} (Craig and Lupton, 1981; Lupton, 1983)). Because of this uniform composition atmospheric - helium is used as a standard in most laboratories and it has become convention to express He-isotope compositions relative to that of atmospheric - He (R/R$_a$=(^{3}He/^{4}He)samble (^{3He}/^{He}atm.

The bulk of^{3}He loss from mantle takes place in ocean basins through processes associated with the generation and cooling of oceanic lithosphere. However, a small proportion of ^{3}He is also lost through continental crust undergoing contemporary deformation. Although the mechanism of transport of helium from mantle into crust is not well-understood, the spatial association of mantle-helium and volcanics in various settings (Craig et al., 1978; Condomines et al., 1983; Kurz et al., 1983) suggests a relationship between mantle melts and transport of mantle - He.

The aim of the present study is the investigation of the mantle-crust interaction in Western Turkey through the He-isotope analyses of geothermal fluids. With the exception of one sample taken from a drilling well (Ömerbeyli - Germencik), all the other samples are collected from natural springs either as water and/or a gas phase bubbling through the water The majority of samples are groundwater discharges associated with the faults bounding the major grabens in Western Turkey. These samples are likely to represent mixtures of a deep cfrculating hydrothermal system and cold, meteoric waters infiltrating through the graben filling sediments (IUTF, 1971, 1975). The temperature of the fluids range from 35°C to 100°C. The sampling locations are shown in Figure 1 along with the measured He-isotope compositions.

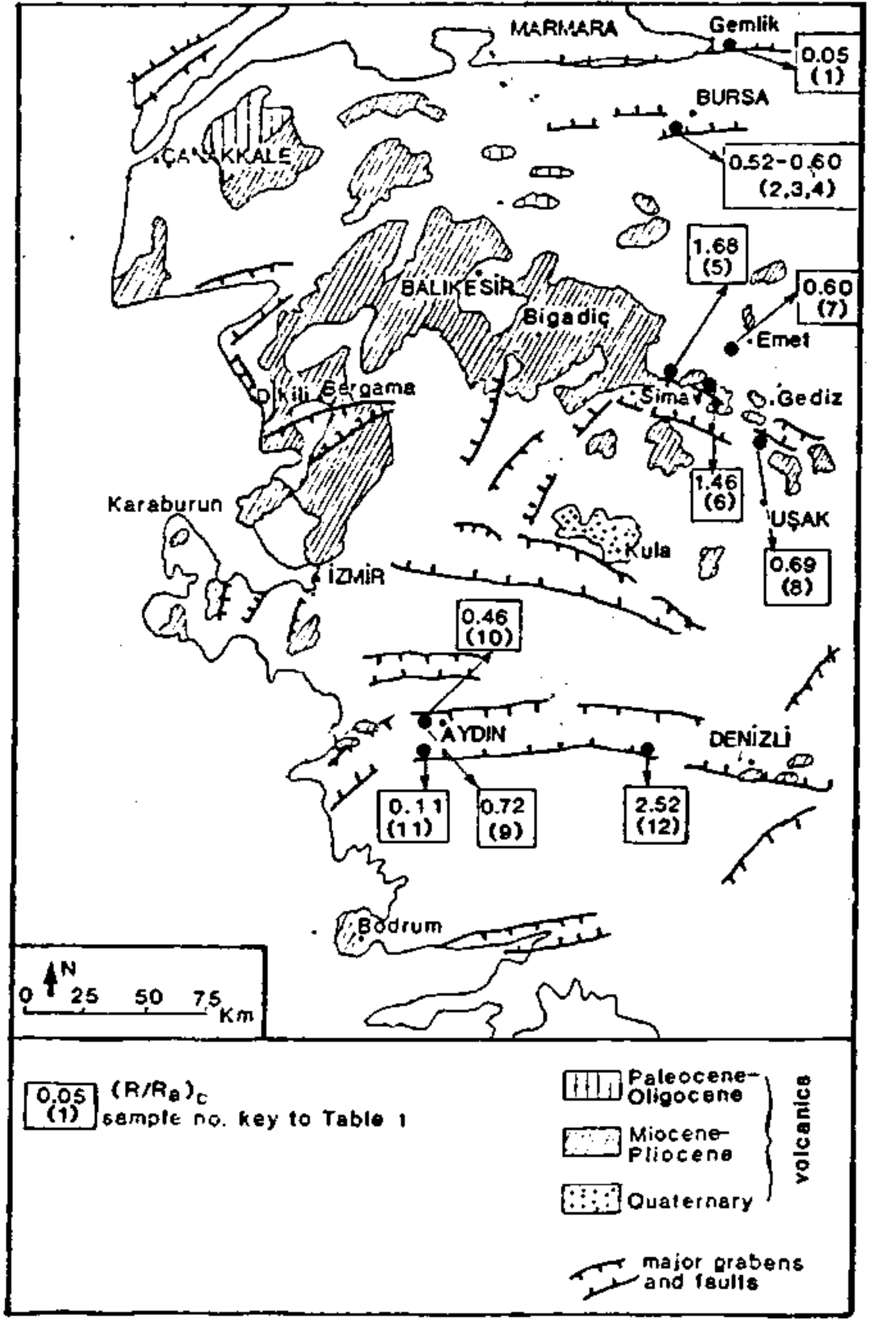

Fig. 1 — Distribution of ^{3}He in Western Turkey in relation to distribution of fault systems and surface volcanics.

ANALYTICAL METHODS

Sample collection

Samples were collected in 10 ml "Weiss" type copper tube devices (Weiss, 1969). The copper tubing is clamped at each end with hinged knife-edged clamps. Copper sample tubes were connected to the spring sources using PTFE tubing and flushed with several volumes of water to ensure that any trapped air bubbles were released. Gas samples were collected at open bubbling pools by means of an inverted funnel which was attached to the copper sample tube and immersed to the maximum possible depth.

He — analyses

He — isotope analyses (together with the measurements of He and Ne abundances) were carried out on VG MM 3000 mass spectrometer in the Rare Gases Laboratory of the Department of Earth Sciences-University of Cambridge. The ^{3}He/ He ratios (= R) were measured against an air standard (= R) and presented as R/R_a values in Table 1. He and Ne abundances were determined by comparison against the air standard. Taking into account the possible air (atmospheric-He) contamination during collection or measurement of the samples, R/R_a ratios were corrected using He/Ne ratios. The correction procedure assumes that all the Ne in the samples is of atmospheric origin; the reason behind this assumption is that the crustal radiogenic processes generate neon in negligible proportion to radiogenic helium and mantle sources are characterized by He/Ne ratios more than 10^3 . The equation used in correction is given in the foot-note of Table 1. The errors associated with the analyses were determined by the consideration of the following factors : 1) the precision related to the performance of the mass spectrometer during the measurements; 2) reproducibility of the measurements; 3) blank corrections An uncertainty of ± 10% is assigned to He — isotope measurements and of ± 20% is assigned to the measurement of He and Ne abundances.

He - ISOTOPE COMPOSITION OF THE FLUIDS

The isotopic composition of He in the geothermal fluids of Western Turkey is given in Table 1 along with the He/Ne ratios and He and Ne abundances. The distribution of ^{3}He in Western Turkey is shown in Figure 1 in relation to the distribution of the fault systems and surface volcanics. Figure 2 compares the He — isotope composition of the Turkish fluids with those from a variety of volcanic settings.

The recorded $(R/R_a)_c$ values range from 0.05 to 2.52. A comparison with mantle-helium (typified by the samples associated with mid-ocean ridges and hot-spot sites) and (continental) crustal-helium, (Fig.2) reveals the presence of a mantle-helium component in Western Turkey. However, the isotopic composition of this mantle-helium is modified as a result of mixing with crustal-helium. Assigning average R/R_a values of 8 and 0.02 for mantle and crustal-helium components, respectively, the involvement of mantle-derived helium in Western Turkey is found not to exceed 30% of the total-He content of any single sample. As far as the comparison with different tectonic settings is concerned the values obtained for Western Turkey are within the documented range of active rifting zones and are rather similar to those reported for the Rhine (graben and the Pannonian basin (Fig. 2).

From Figure 1, ^{3}He appears to be widely distributed in Western Turkey paralleling the distribution of the fault systems of the present extensional regime. In this respect,^{3} He distribution in Western Turkey seem to be similar to that observed in the Pannonian basin (Fig. 2 of Oxburgh et al., 1986) in contrast to the highly localized distribution pattern seen in the Rhine graben (Fig. 2 of Oxburgh et al., 1986). With regard to the relationship between ^{3}He distribution and tectonic deformation, an important point to note is the low R/R_a values obtained from Gemlik which is located along the extension of the North Anatolian Fault Zone where the dominant strike-slip motion is accompanied by normal faulting. High R/Ra values are concentrated in the normal fault systems of Western Turkey where the N — S extension is much more prominent. On the other hand, there does not seem to be any relationship between the distribution of ^{3}He and surface volcanism. In general, the geothermal systems associated with the youngest volcanic activities are expected to have the highest ^{3}He/^{4}He ratios, because the ratio should decrease with time due to the accumulation of radiogenic^4 He. However, this is not the case in Western Turkey ; the highest R/R_a value is in the vicinity of Pliocene age Denizli volcanics rather than Quaternary Kula volcanics. In this respect, either Denizli represents an area of high mantle-helium leakage (coupled with low crustal degassing in order to keep this helium in the crust since Pliocene) or the distribution of He is not linked to the surface volcanism.

Table 1– He isotope measurements on Western Turkish fluids

Sample no.	*Locality*	*Key*$^{(1)}$	*Year*	*Type*	$(^4He)^{(2)}$	$(^{20}Ne)^{(2)}$	$(^4He/^{20}Ne)^{(3)}$	R/R_a	$(R/R_a)_c$
T1	Gemlik	1	1984	water	3.4×10^{-7}	3.6×10^{-7}	0.95 ± 0.2	0.29 ± 0.03	0.05
T3	Çekirge-Bursa	2	1984	water	2.7×10^{-7}	8.1×10^{-9}	45 ± 12	0.52 ± 0.05	0.52
T4	Çekirge-Bursa	3	1984	water	4×10^{-7}	1.2×10^{-8}	106 + 230/−70	0.52 ± 0.05	0.52
T85-1	Çekirge-Bursa	4	1985	water	6.4×10^{-8}	1.5×10^{-7}	0.42 ± 0.09	0.83 ± 0.03	0.60
T7	Eynal-Simav	5	1984	gas	------	------	300	1.68 ± 0.17	1.68
T8	Ilıcalar	6	1984	water/gas	------	------	100 ± 26	1.46 ± 0.15	1.46
T6	Emet-Kaynarcâ	7	1984	water	1.5×10^{-7}	9×10^{-9}	34 + 24/−17	0.60 ± 0.06	0.60
T9	Banaz	8	1984	gas	------	------	17 ± 3	0.70 ± 0.07	0.69
T14	Çamur-Germencik	9	1984	gas	------	------	1.5 ± 0.3	0.77 ± 0.08	0.72
T13	Bozköy - Germencik	10	1984	gas	600 ppb	80 ppb	78 ± 16	0.46 ± 0.05	0.46
T16	Ömerbeyli-Germencik	11	1984	gas	------	------	2.3 ± 0.5	0.22 ± 0.02	0.11
T85-15	Tekke Hamam-Denizli	12	1985	gas	3.1 ppm	39ppb	79 ± 16	2.52 ± 0.08	2.52
Air saturated water 0°C$^{(5)}$					4.9×10^{-8}	2×10^{-7}	0.24		

(1) Designation on Fig. 1.

(2) (^4He) and (^{20}Ne) concentrations are expressed as ccSTP/g H_2O for water samples and as volume fractions for gas samples. Since details of volumetric splitting were not recorded, (^4He) and (^{20}Ne) concentrations could not be obtained for the gas samples collected in 1984. All determinations ± 20%.

(3) Errors denote scatter typical of successive air runs.

(4) $(R/R_a) \simeq ((R/R_a)X - 1)/(X - 1)$

$X = (^4He/^{20}Ne)_{sample} \,/\, (^4He/^{20}Ne)_{atm}$

$(^4He/^{20}Ne)_{atm} = 0.288$ (for gas samples)

$(^4He/^{20}Ne)_{atm} = 0.24$ (for water samples)

(the factor, 0.24/0.288 = 0.83 accounts for the differing solubilities of He and Ne in water).

(5) Benson and Krause (1976).

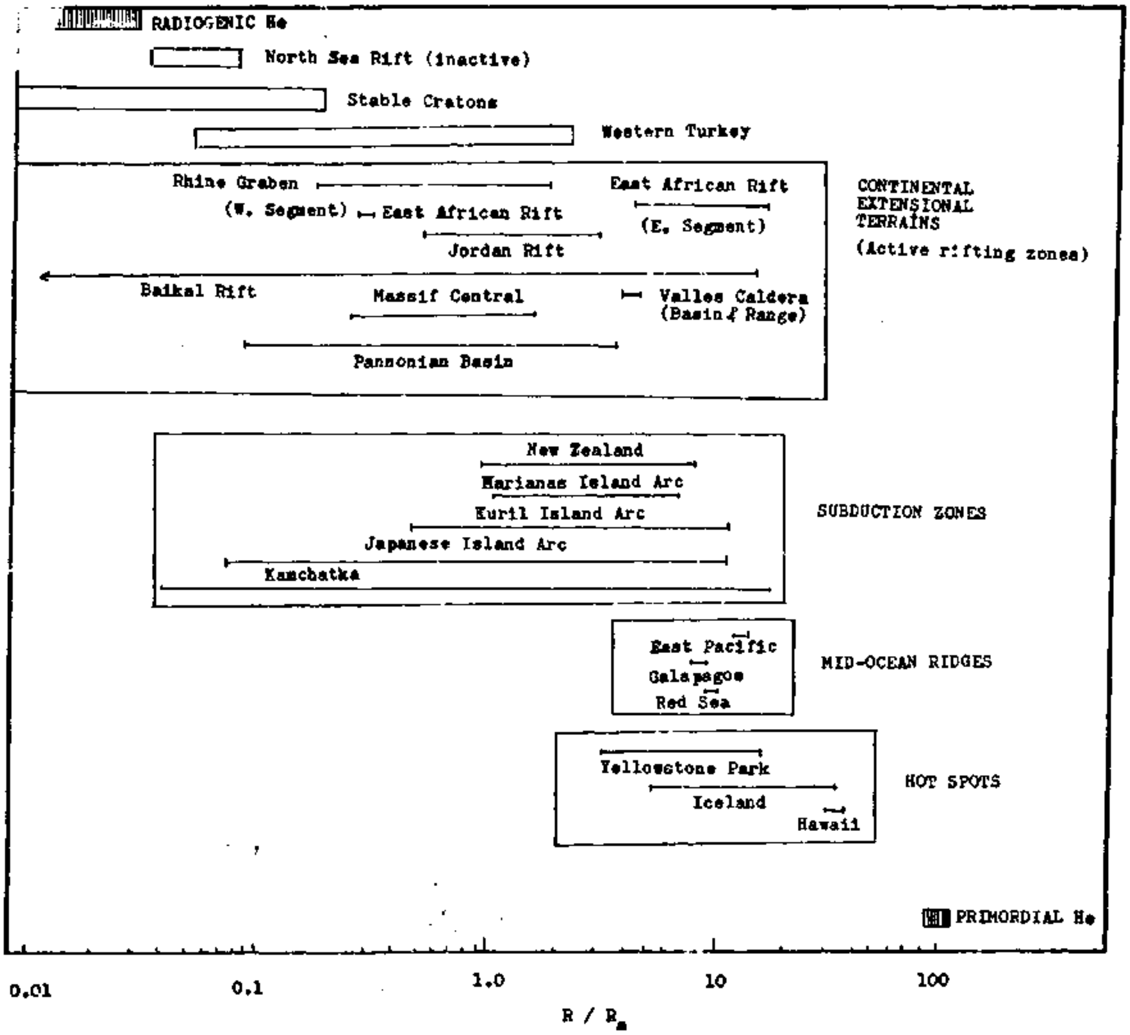

Fig.2— The isotope composition of He from different tectonic settings compared with results obtained for Western Turkish fluids.

Data sources : 1) Kennedy et al., 1985 (Yellowstonepark); 2) Hooker et al., 1985(Rhine Graben and North Sea rift); 3) Oxburgh et al., 1986 (Pannonian basin); 4) Smith and Kennedy, 1985 (Valles caldera); 5) Polyak and Tolstikhin, 1985 (the others).

Bars represent the ranges of R/R_a values for 1, 2, 3 and 4 and 95% limits of log-distribution for 5.

DISCUSSION

The present study confirms the presence of ^{3}He enrichments in Western Turkey over that expected from normal crustal lithologies. The discussion presented here are directed towards answering the following questions:

1-Do the ^{3}He enrichments necessarily indicate a mantle contribution ?

2- If so, what are the possible mechanisms responsible for the transportation of mantle-helium to the surface?

^{3}He and possible sources

Although ^{3}He is essentially primordial, it can also be produced by the radioactive decay of ^{6}Li and ^{3}H isotopes. ^{3}He produced from ^{6}Li and ^{3}H is known as radiogenic ^{3}He and tritiugenic ^{3}He, respectively. The rate of radiogenic ^{3}He production is dependent on the Li content of the material. However, since most rocks have Li abundances less than 50 ppm, the R/R_a values of the helium produced by ^{6}Li decay does not exceed 0.02 (Mamyrin and Tolstikhin, 1984; Andrews, 1985). Therefore, radiogenic ^{3}He can not be an adequate source for the ^{3}He observed in the Western Turkish fluids.

Tritiugenic ^{3}He production occurs through the decay of tritium (^{3}H) introduced into the atmosphere during the nuclear testing. If the groundwaters are supplemented by post - 1950 (i.e. after the commencement of nuclear testing) recharges from surface waters, then they are subject to the accumulation of ^{3}He produced by the decay of atmosphere-derived ^{3}H in surface waters. Since there is no information regarding the distribution of bombproduced tritium in Western Turkey or about the groundwater ages, it is difficult to asses the effects of tritiugenic ^{3}He production It should be noted that because the ^{3}He contents of water samples are low (10 ccSTP g), the decay of ^{3}H could potentially increase the He/ He ratios. On the other hand, a comparison between the He contents of water and gas samples reveals that the He contents of gas samples are about an order of magnitute higher than those of water samples. The reason for the low He contents of water samples is the separation of a gas phase into which most of the noble gases partition. However, although the He contents of gas and water samples are different, their ^{3}He/^{4}He ratios are similar to each other. If tritium were introduced into the groundwaters, then as a result of the separation of water and gas phases, the ^{3}He/^{4}He ratios of waters would be enhanced relative to the gas phase (He partitions into the gas phase, whereas ^{3}H is retained in water as a part of water molecule (^{3}H^{1}HO). The retained ^{3}H would produce, in time, ^{3}He in waters). Therefore, in the case of tritiugenic ^{3}He contributions, lower He contents in waters are expected to be accompanied by higher ^{3}He/^{4}He ratios. In Western Turkey, ^{3}He/^{4}He ratios of water samples are not necessarily higher than those of gas samples ; hence, the ^{3}He enhancements can not be due to H decay.

Transport mechanism of mantle - helium

The transport of helium from mantle into crust is-in general - not well understood, but it is thought to be associated with mantle - derived melts or fluids. If mantle-helium in Western Turkey is transported into the crust by mantle related magmatism, then the fact that no relationship exists between the distribution of ^{3}He and volcanism suggests that plutonic activity may be widespread in Western Turkey. If mantle - derived fluids are responsible for the transportation of mantle-helium, then they must have been injected into the crust over a large area. The association of hot-springs with major fault structures suggests that the fault zones might have acted as conduits for mantle-derived fluids.

One of the problems in determining how mantle-helium is transported into the crust is that it is not known when it was added to the crust. Miocene (or older) volcanism may represent a possible mechanism only if helium can be stored in the crust long enough (i.e. 10^{7} years). The relationship between heat flow and He-isotope compositions (Polyak and Tolstikhin, 1985) and the observation that the thermal relaxation time for continental crust is of the order of 10^{8} years (Sclater et al., 1981) suggests that crustal storage of helium over such a period may be possible. However, there is not any evidence for deciding when mantle-helium was added to the crust; it might have been added very recently in some areas.

As has already been stated in the foregoing section, the observation of high R/R_a values in the vicinity of Pliocene age Denizli volcanics indicates that either Denizli is a site of prominent mantle - He leakage and that this helium is still being retained in the crust, or that the distribution of ^{3}He is not linked with surface volcanism. In the latter case, the transport of mantle-helium into the crust might result from plutonic activity (much younger than volcanism) or present day loss of fluids from the mantle.

There is an accumulating evidence for mantle-derived melt additions to the continental crust in extensional areas. Geophysical surveys carried out on some of the extensional basins (e.g. Rhine graben , Baikal rift, Rio Grande rift) reveal the presence of highly conductive, anomalously low—velocity zones at sub—crustal (and/or intra—crustal) levels which are interpreted as the zones of magma accumulation (Hermance, 1982 and the references terein). On the basis of these evidences, mantle—derived melts are believed to have had an efficient role in transporting mantle—helium into the crust of Western Turkey where extension is prevailing since about Upper .vliocene. Since mantle — He is widely distributed in Western Turkey, rather than localized plutonic activities, a widespread melt addition to the crust is possibly indicated. The fault systems are thought to have aided the escape of helium to the surface through the brittle parts of crust.

CONCLUSIONS

The main conclusions drawn from the present study can be stated as follows :

1 — The He — isotope composition of the geothermal fluids reveal mixing between mantle —He and crustal—He components.

2 — The distribution of mantle—He does not show any relation to the distribution of surface volcanics but rather appears to be governed by the distribution of the main fault structures.

3 — The lack of any evidence between the distribution of mantle—helium and surface volcanism suggests that helium is degassed from the melts emplaced deeper in the crust.

4 — The fault systems of the present extensional regime are thought to have acted as channelways for the escape of helium to the surface through the crust.

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UNE NOUVELLE ESPECE DU GENRE DE *BOURNONIA* FISCHER (RUDISTE, BIVALVIA) DANS LE MAESTRICHTIEN DE L'ANATOLIE CENTRALE (TURQUIE)

Sacit ÖZER*

RESUME.—Dans cette article nous donnons l'etude systematique d'une nouvelle espece du genre *Bournonia* Fischer qui a ete trouvee dans le Maestrichtien de I' Anatolie Centrale.

ABSTRACT.—In this article, the systematic study of a new species of the genera *Bournonic* fischer which is collected from the Maestrichtian of the Central Anatolia is given.

INTRODUCTION

Nos travaux sur les Rudistes des bassins de Haymana-Polatlı, Tuz Gölü et Çankırı-Çorum qui se trouvent dans l'Anatolie Centrale, nous ont permis de mettre en Evidence l'existence d'une riche faune de Rudistes d'age maestrichtien(Özer, 1985).

Les echantillons de Rudistes qui ont ete' recueillies de la localite de Malıboğazı,situee au Sud de Çandır, a l'Quest du bassin de Çankırı-Çorum (Fig.l) nous offrent l'occasion de decrire une nouvelle espece de *Bournonia* Fischer qui est une genre trouvee pour la premiere fois en Turquie.

Dans cette localite, la formation a Rudiste est constitute par des altemances de calcaires, calcaires greseux et gres. Les Rudistes forment une association appartenant a la sous-famille Biradiolitinae. D'autre

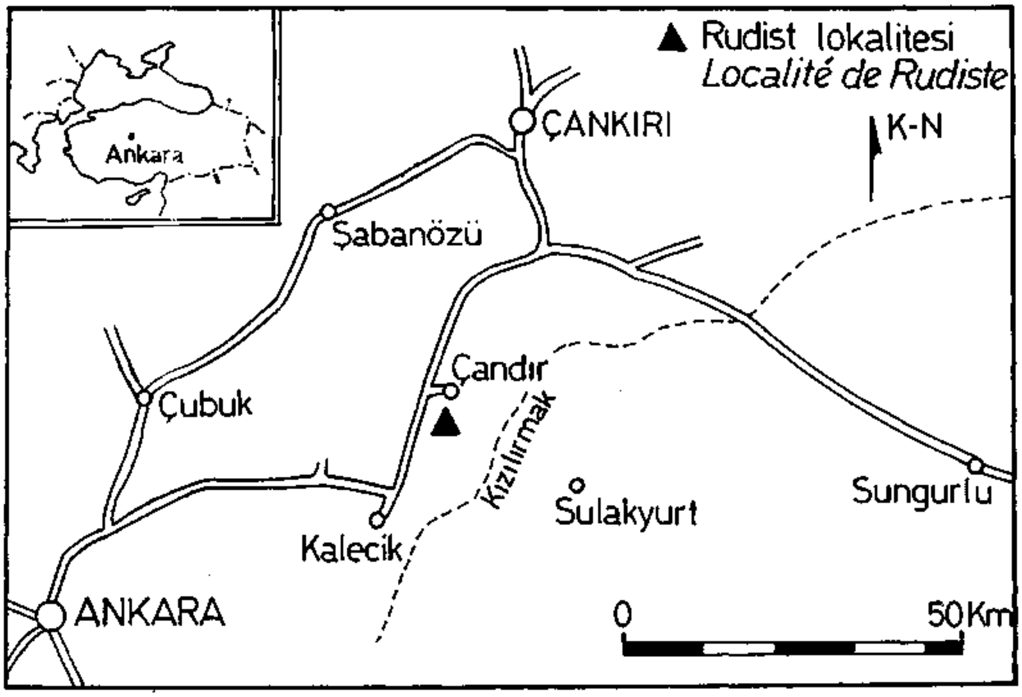

Fig . 1 — Situation géographique de la localité de Rudistes.

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part, nous avons ramasse des Lamellibranches, Gastropodes, Madreporaires, Algue rouges et grands Foraminiferes benthiques.

ETUDE SYSTEMATIQUE

Classe : Bivalvia
Ordre : Hippuritoida Newell, 1965
Super-famille : Hippuritacea Gray, 1848
Famille : Radiolitidae Gray, 1848
Sous - famille : Biradiolitinae Douville, 1902
Genre : *Bournonia* Fischer, 1887
Genotype : *Sphaerulites bournoni* Des Moulins, 1826

Bournonia anatolica n. sp.

(Fig.2 **anche I et II, fig.1,2)**

Materiel fossile : 2 exemplaires a deux valves et 6 exemplaires de la valve inferieur.

Holotype : L'echantillon est conserve sous le No. SM36 dans les laboratoires de l'Universite' de Dokuz Eylül(İzmir), sa photographie est donne'e sur la Planche I, fig.1 et Planche II, fig. 1,2.

Diagnose : Section transversale de la valve, subtriangulaire ou ellipsoidale. Valve inferieure rectangulaire et robuste. Bande siphonale posterieure se trouvant au milieu de la region siphonale et se recourbant legerement vers la partie dorsale. Bande siphonale anterieure deplacee vers le cote ventrale. Interbande en forme de sillon simple.

Description : La valve inferieure est rectangulaire et robuste (Planche I, fig. 1). La section transversale est subtriangulaire ou ellipsoidale (Planche II, fig. 2).

La bande siphonale posterieure (S) se trouve au milieu de la region siphonale. Elle est un peu saillante et se recourbe legerement vers le cote dorsale et se continue jusqu'au bas de la coquille. Elle est large au niveau de la commissure et etroite dans l'extremite de la valve. La bande siphonale anterieure (E) est deplacee du cote ventrale. Elle a la forme d'une cote longitudinale dont la largeur ne change approxitivement pendant toute sa longeur. Elle offre une structure relativement saillante a son extremite. Entre les deux bahdes siphonales se trouve un sillon simple, c'est a dire sans cotes, correspondant a l'interbende etroite au niveau de la commissure et large a l'extre'mite' de la coquille. La profondeur de l'interbande est plus grande a la base de la valve a cause de bandes siphonales S et E qui y sont tres saillantes (Planche I, fig. I)-

La surface exteme de la valve inferieure est toujours lamelleuse.

La valve superieure est a peu pres plane (Planche I et II, fig. 1).

L'epaisseur de la couche externe est variable. Le tissu du test est cellulo-prismatique dont les prismes sont sous forme de cellules rectangulaires (Planche I, fig. 2). Dans les sections transversales, la structure interne des bandes siphonales consistent toujours de lamelles, paralleles a la surface externe de S et E. On remarque l'existence d'une structure convexe (Planche II, fig. 2) qui est remplie par la calcite, ce qui s' observe dans la section transversale des bandes S et E.

La couche interne est probablement plus mince (1-2 mm).

L'arete ligamentaire n'est pas representee. L'appareil cardinale de l'holotype n'est pas en bon etat de conservation. II semble que les dents sont relativement robustes (Fig. 2). En effet, elles sont tres visibles sur un paratype (Fig. 3). L'apophyses myophore posterieure est plus large que l'apophyses myophore anterieure et s'e'tende jusqu'au devant de la bande siphonale posterieure. Le test est invejine au niveau de l'interbande.

Les mesures prises sur l'holotype et des paratypes sont portees sur le Tableau 1.

Rapports et differences : Cette nouvelle espece est voisine de *Bournonia bournoni* (Des Moulins) Bayle, par la section transversale des valves subtriangulaire. Elle en differe nettement par l'absence de cotes ventrale et dorsale (Douville', 1910; Pejovic, 1968). D'autre part, elle a d'affinite avec *Bournonia*

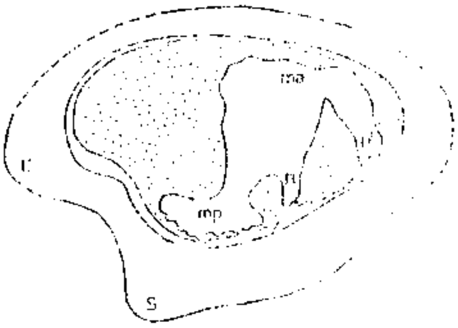

Fig. 2 — ***B. anatolica*** **n. sp., section transversale de la valve inférieure, holotype, l'exemplaire sur la Planche II, fig.2.**

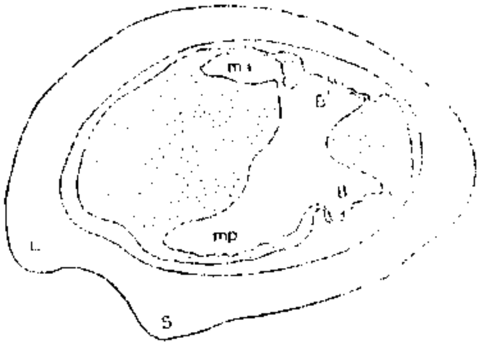

Fig. 3 — ***B. anatolica*** **n. sp., section transversale de la valve inférieure, paratype, SM 34.**

dinarica Sliskovic par la place de la bande anterieure et par la section transversale de valves (Sliskovic, 1971) et s'en distinguent par 1'ornementation tres simple de 1'interbande.

On constate une grande ressemblance des regions siphonales entre *B. anatolica* n. sp. et *Biradiolites* cf. *leychertensis* decrit par Plenicar (1977).

Bournonia anatolica n. sp. se distingue des autres especes connues du genre de *Bournonia* Fischer par les caracteristiques specifiques de la structure de la region siphonale.

Origine du nome : Cette espfece est trouvee pour la premiere fois dans 1'Anatolie.

Localite type : Holotype, Malıboğazı-Çandır (Ankara). Coordinates sur 1 : 25 000 carte d'echelfe (carte de reference H 30 c2) sont de 42.06:52.14.

Strato type = Etage type : Dans la localite de Malıboğazı, la formation a Rudiste se trouve au dessus d'une formation flyschoide d'age maestrichtien, et elle est surmontee par une formation clastique d'age paleocene (Ünalan et Harput, 1983).

Les Rudistes y sont associes avec des macrofossiles *(Cyclolites, Pycnodonta, Exogyra* etc.) et egalement des Foraminiferes benthiques *(Orbitoides medius, Siderolites calcitropoides, Lepidorbitoides* sp.) qui sont abondants dans le Maestrichtien de 1'Anatolie Centrale.

On y observe une association de Rudistes de *Biradiolites bulgaricus* Pamouktchiev (trfes abondants), *Sphaerulites solutus* Pethö,*Vaccinites loftusi* Woodward. Ces Rudistes ont ete identifies dans le Maestrichtien de Yougoslavie, de Bulgarie et d'Anatolie :*B. bulgaricus* en Bulgarie (Pamouktchiev, 1967, 1981); S. *solutus* en Yougoslavie (Pethö, 1906, Klinghardt, 1921; Veselonovic et Pejovic, 1958; Sladic-Trifunovic, 1972) et a 1'Est du bassin de Tuz Gölü-Anatolie Centrale (Özer, 1985);*V. loftusi* en Bulgarie (Pamouktchiev, 1961, 1981), en Yougoslavie (Sladic-Trifunovic, 1979) et en Turquie (Karacabey, 1974; Karacabey - Öztemür, 1979; Özer, 1985).

Le strato type de *Bournonia anatolica* n. sp. doit etre Maestrichtien, d'apres sa position stratigraphique et les fossiles cites ci-dessus.

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Je remercie sincerement M. le Professeur Jean Philip pour ses critiques au cours de mes etudes sur les determinations de Rudistes de 1'Anatolie Centrale dans son laboratoire (Univ. Provence, Marseille) et particulierement pour ses conseils sur I'identification de cette nouvelle espece.

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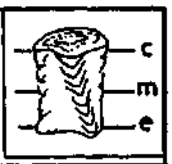
Tableau 1 — Les mensurations de *B. anatolica* n. sp.

Ölçüler *Mensurations* (mm.)			Holotip *Holotype*	Paratipler *Paratypes*						
Çap - *diamètre*			115 x 80	120x80	110x80	105x80	110x75	95x65	80x55	85x55
Alt kavkı uzunluğu - *longeur de la valve inf.*			170	110	140	100	60	90	80	145
Genişlik - *largeur*	S	c	20	35	30	25	25	20	20	20
		e	10	–	–	–	–	15	–	–
	E	c	20	25	25	20 ?	20	20	18	13
		e	15	–	15	–	–	15	–	–
	I	c	20	15	15	15 ?	20	15	18	15
		m	35	25	25	–	30	18	25	–
		e	55	–	30	–	–	20	–	–
Dış tabaka kalınlığı - *épaisseur de la couche externe*			10-20	13-20	15-25	10-20	10-20	10-18	6-15	10-15
Örnek No *no d'exemplaire*			SM36	SM40	SM37	SM39	SM41	SM42	SM26	SM30

Largeurs des bandes (S, E, I) ont été mesurees aux trois niveaux (cf. fig. au coin du tableau): c - niveau de commissure; m - milieu de l'interbande; e - extremite de la valve inférieure.

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PLANCHES

PLANCHE-I

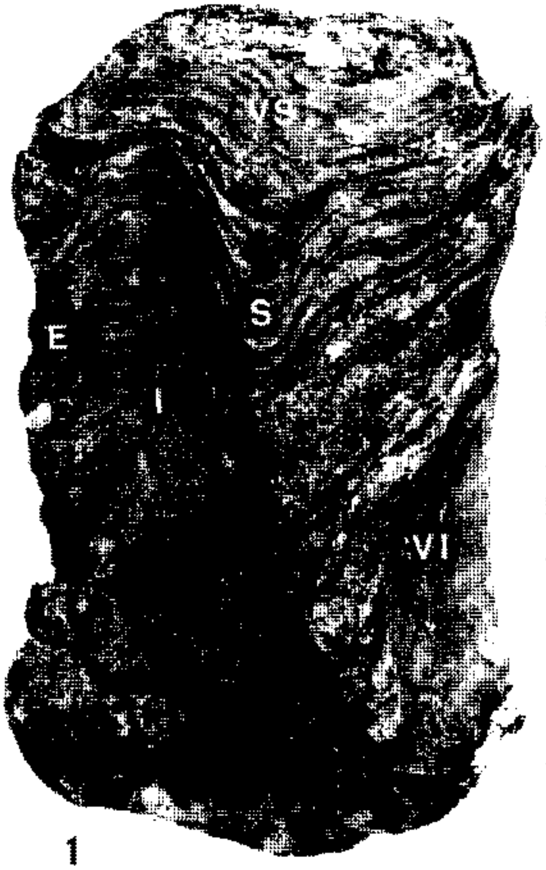
Bournonia anatolica n. sp.

Fig.l— Valves inferieure (VI) et superieure (VS), vues du cote siphonal, x 3/4, holotype.

S, E -- bandes siphonales posterieure et anterieure,
I — interbande.

Fig.2— Section transversale de la valve inferieure et structure de la couche externe, x 1, paratype.

1 — lamelles,
r — cellules rectangulaires.



1

2

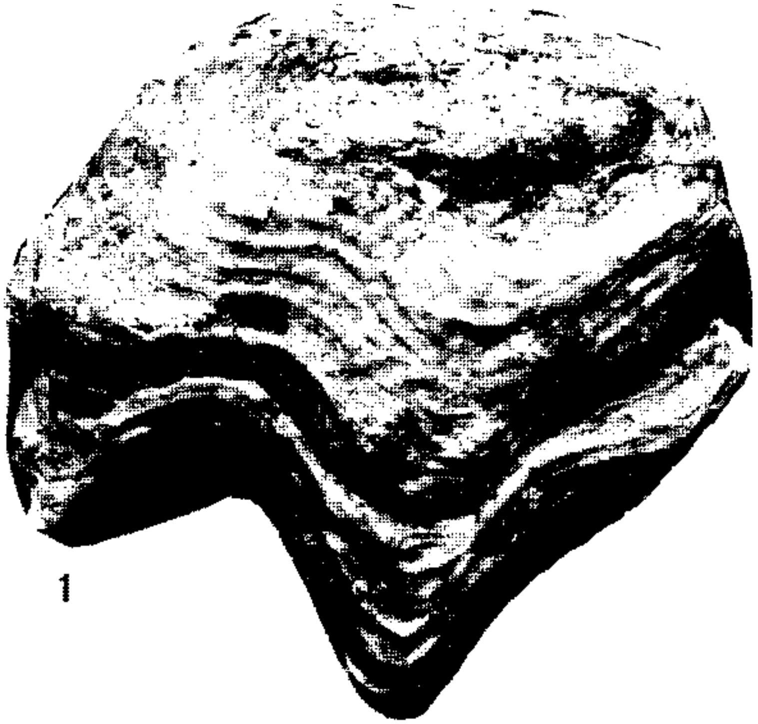
PLANCHE-II

Bournonia anatolica n. sp.

Fig.1— Valve superieure, vue du haut, x 1, holotype.
On observe une partie de la valve inferieure.

Fig.2— Section transversale de la valve inferieure, 5 mm au dessous de la commissure, x 1, holotype.

B, B' — dents posterieure et anterieure,
mp, ma — apophyes myophores posterieure et anterieure,
c — structure convexe.



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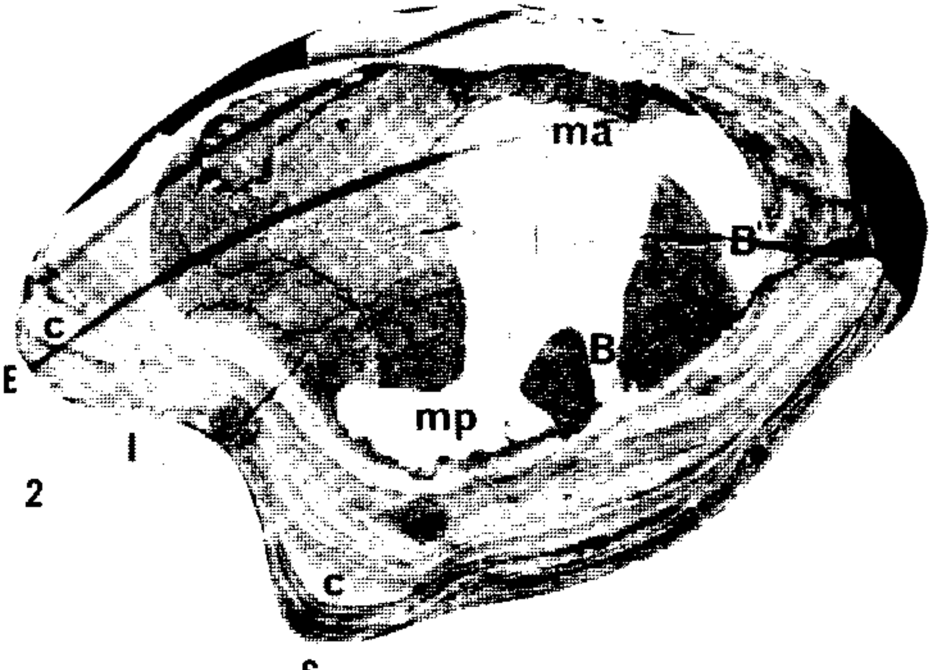


2

Mineral Res. Expl. Bull., 108, 49-56, 1988

DASYCLADACEAN ALGAE FROM THE MESOZOIC CARBONATE FACIES OF THE SARIZ-TUFANBEYLİ AUTOCHTHON (KAYSERİ, SE TURKEY)

Baki VAROL* ; Demir ALTINER** and Yavuz OKAN*

ABSTRACT .— Some levels of the Mesozoic carbonates of the Sanz-Tufanbeyli area yield abundant dasycladacean algae which are useful parameters to define the chronostratigraphic position and depositional environment of these sequences. The following forms identified in the carbonate sequences of the Mesozoic are typical to subdivide these levels : *Griphoporella curuata* Gümbel, *Diplopora* sp. in the Upper Triassic; *Selliporella donzellii* Sartoni and Crescenti in the Dogger; *Macroporella sellii* Crescenti, *Clypeina jurassica* Favre and *Campbelliella striata* (Carozzi) in the Malm; *Salpingoporella annulata* Carozzi, *Salpingoporella steinhauseri* Conrad, Praturlon and Radoicic, *Salpingoporella pygmaea* (Gümbel), *Munieria baconica* (Carozzi), *Clypeina? solkani* Conrad and Radoicic in the Lower Cretaceous (Neocomian); *Salpingoporella dinarica* Radoicic, *Salpingoporella hasi* Conrad, Radoicic and Rey in the Aptian-Albian and *Heteroporella lepina* Praturlon in the Upper Cretaceous (Cenomanian). In addition to these forms, many other algal species have been sporadically recognized in these levels.

INTRODUCTION

In the eastern prolongation of the taurids, the autochthonous Mesozoic carbonates of the Sariz - Tufanbeyli region (Fig. 1), limited in the NW by Soğanlıdağ and in the SE by Binboğa Dagları, preserve their shallow water depositional characteristics in a 1000 to 1500 m thick sequence of Early Trjassic to Late Cretaceous age. The facies variations in this sequence were mostly controlled by movements of tides and sea level changes due to the epirogenic movements of the platform. In the time intervals when these movements were accelerated, sudden changes in the sea level and salinity variations controlled the mechanism of dolomitization and highly limited the distribution of foraminifers. On the contrary, the dasycladacean algae showed an increase in abundance and became more frequent especially at Jurassic-Cretaceous boundary and in the Neocomian of the study area. In general, these units represented by the same environmental characteristics in the whole taurus carbonate platform, except few studies carried out up to now (Jaffrezo et al., 1978; Altıner and Septfontaine, 1979 ; Altıner and Decrouez, 1982), the absence of detailed studies on dasycladacean algae led to the incomplete interpretations on these units, treatment of the Neocomian as a whole, undivided chronostratigraphic unit and incomplete description of the Jurassic-Cretaceous boundary.

Approximately 100 samples coming from 11 stratigraphic sections (Fig. 2) measured in the autochthonous Mesozoic carbonates of the Sariz - Tufanbeyli region have been analyzed and dasycladacean algae have been identified. Nearly 30 species of dasycladacean algae have been listed in the stratigraphic column according to the character of being biostratigraphic zonal marker or of local occurrence (Fig. 3).

GEOLOGIC SETTING

The studied Mesozoic unit belongs to the autochthonous Geyikdağı unit of Özgül (1976). The main unconformity surfaces recognized in the sequence He at Lower — Middle — Upper Triassic and Triassic — Jurassic (Dogger) boundaries. The unconformity observed within the Triassic system is only local importance. On the other hand, the unconformity seperating the Triassic and Jurassic systems is characteristic for whole region but difficult to observe because of local dolomitization. In the areas where the Jurassic seems to overlie the Permian, the boundaries are mainly tectonic in origin.

The continuous sedimentation through the Jurassic-Cretaceous boundary can only be followed in the areas where the interval corresponding to this boundary is not masked by dolomitization. The sedimentation of the Mesozoic sequence shows a hiatus between Cenomanian - Turonian and Santonian in the upper levels in the uppermost stages of the Upper Cretaceous, the carbonate sequence is either represented by rudistid or pelagic limestones depending on the variation in the paleogeographic setting of the region. In the studred carbonates, the fundemental geologic studies

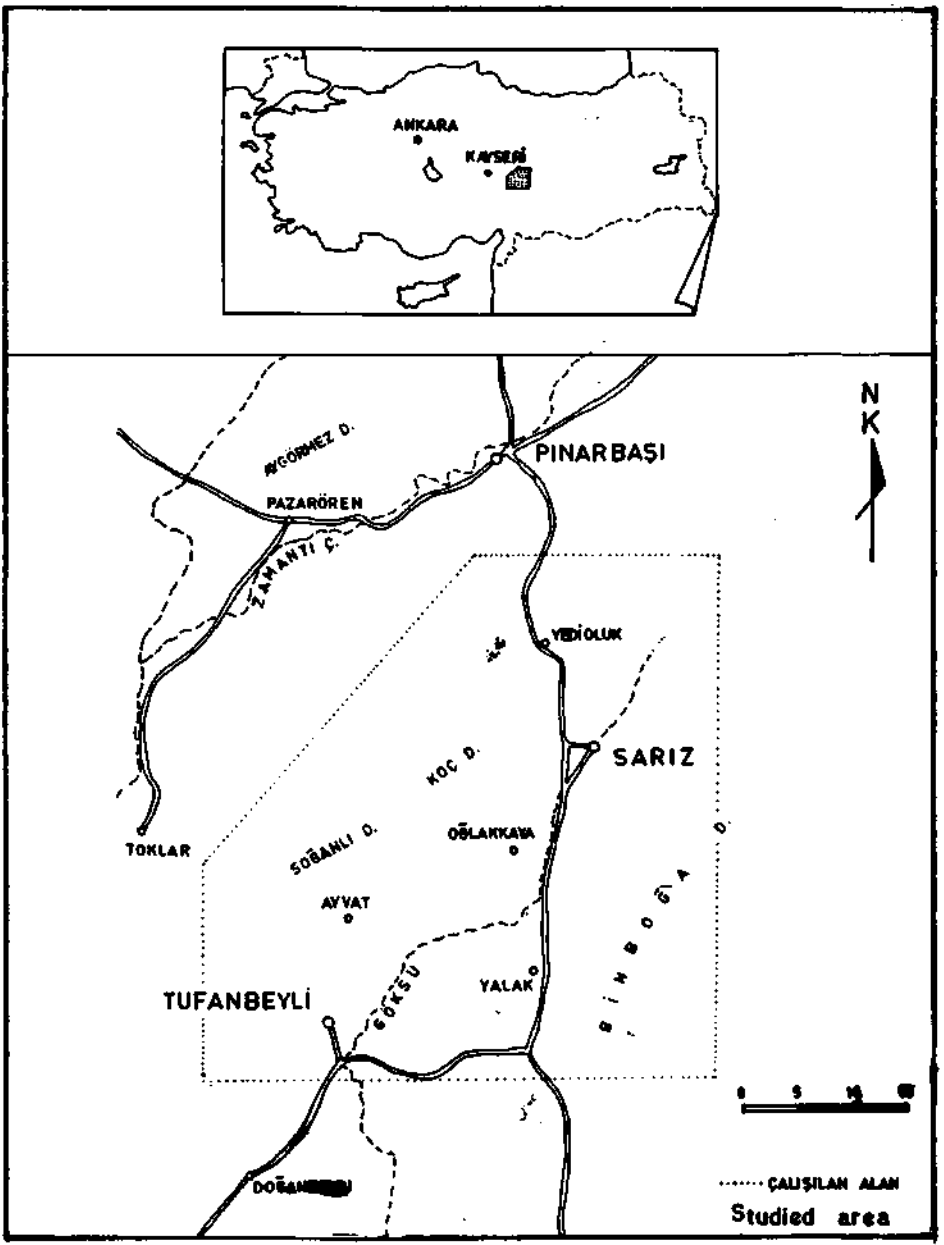


Fig.1 – Location map of the study area.

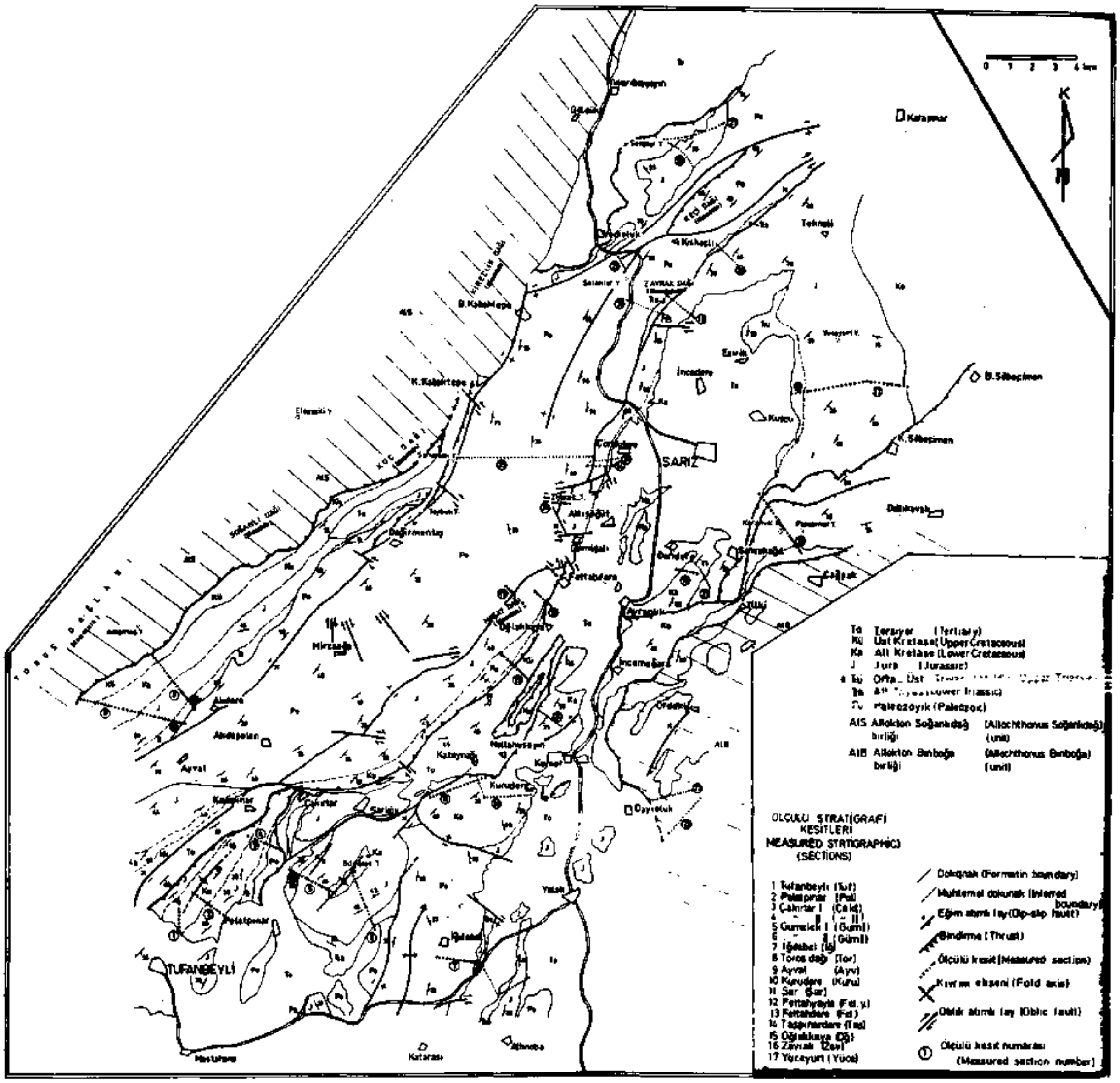


Fig.2— Location of the stratigraphic sections and generalized geologic map of the study area. Geologic map of the eastern part of the autochthonous Geyikdağı unit in the eastern taurids has been prepared after Özgül (1976) and Metin et al. (1982).

and lithostratigraphic subdivisions have been carried out for the first time by Demirtaşlı (1967). More recent studies, such as Metin et al. (1982) presented detailed maps and geological investigations of the area. The Jurassic - Lower Cretaceous limestones containing rich dasycladacean algal levels have been previously described as the Köroğlutepe formation (Demirtaşlı, 1967) or the Yüceyurt formation (Aziz-et al., 1980).

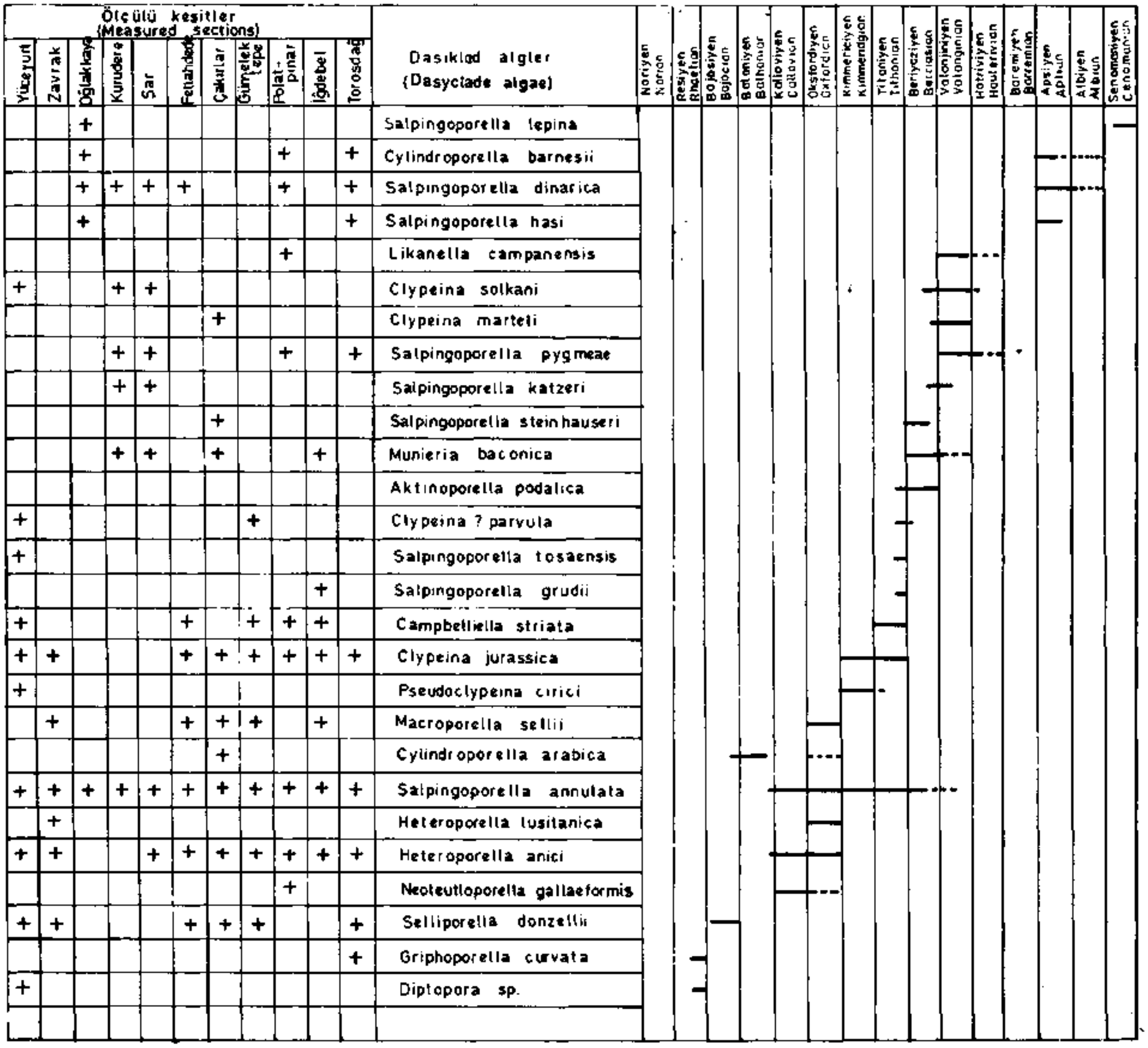
TRIASSIC ALGAE

The Lower Triassic carbonates of very shallow marine (infralittoral) character do not yield any dasycladacean algae. Although the Middle - Upper Triassic is also characterized by shallow water carbonates, the depositional characteristics differ from those of the Lower Triassic sequence. The dasycladacean algae and other bioclastic elements are found in the cavities of green algal biolithites or in the back-reef mudstones which were laid down in lagoonal environment of relatively normal marine conditions. *Griphoporella curvata* Gümbel and *Diplopora* sp. (Plate I, fig. 1,2) are the algal species recognized among the dasycladacean associations of the Upper Triassic.

JURASSIC ALGAE

In the Jurassic limestones, the dasycladacean algae are the major constituens of the lime -mudstone-wackestone and dolomitic mudstone facies. They constitute 4 important biostratigraphic zones and are associated with many species, some of which are recognized for the first time in the study area.

Selliporella donzellii .— *Selliporella donzellii* Sartoni and Crescenti is the major algal species recognized in the lowermost limestone levels of the transgressive Jurassic sequence. The presence of the species in two distinct textural types, such as lime-wackestone and grainstone, indicates variable energy conditions in a shallow littoral belt. The whole or fragmented nature of the specimens generally follow the character of the depositional texture (Plate I, fig. 3).



Ölçülü kesitler (Measured sections)											
Yüceyurt	Zavrak	Oğlakkaya	Kurudere	Sar	Fettahdede	Çakırlar	Gümelek tepe	Polat-pınar	İğdebel	Torosdağ	Dasıklad algler (Dasyclade algae)
		+									Salpingoporella lepina
		+						+		+	Cylindroporella barnesii
		+	+	+	+			+		+	Salpingoporella dinarica
		+								+	Salpingoporella hasi
								+			Likanella campanensis
+			+	+							Clypeina solkani
						+					Clypeina marteli
			+	+				+		+	Salpingoporella pygmeae
			+	+							Salpingoporella katzeri
						+					Salpingoporella steinhauseri
			+	+		+			+		Munieria baconica
											Aktinoporella podalica
+							+				Clypeina ? parvula
+											Salpingoporella tosaensis
									+		Salpingoporella grudii
+					+		+	+	+		Campbelliella striata
+	+				+	+	+	+	+	+	Clypeina jurassica
+											Pseudoclypeina cirici
	+				+	+	+		+		Macroporella sellii
						+					Cylindroporella arabica
+	+	+	+	+	+	+	+	+	+	+	Salpingoporella annulata
	+										Heteroporella lusitanica
+	+			+	+	+	+	+	+	+	Heteroporella anici
								+			Neoteutloporella gallaeformis
+	+				+	+	+			+	Selliporella donzellii
										+	Griphoporella curvata
+											Diplopora sp.

Fig.3— **Mesozoic dasycladacean algal species and their stratigraphic distributions in the measured stratigraphic sections.**

The levels containing *Selliporella donzellii* Sartoni and Crescenti have been defined for the first time in Turkey as the *Selliporella donzellii* Cenozone by Altıner and Septfontaine (1979) and these authors placed this zone in the Bajocian stage. The same zone has been previously recognized by Farinacci and Radoicic (1964) in the Lower Dogger of the apennines and dinarids in Europe. In the study area, the presence of *Teutloporella gallaeformis* Radoicic (Plate 1, fig.4) has been noted in the upper part of this zone. The typical examples of this form were reported from the Upper Dogger ?— Lower Malm sequences of Yugoslavia, Italy and Greece (Radoicic, 1964, 1966) and of the Katran Dağı (Antalya) in Turkey (Bassoulet and Poisson, 1975).

In the Çakırlar section, an interesting algal population has been recognized in the *Selliporella donzellii* zone. This population which has comparable dimensions to those of *Cylindroporella arabica* Elliott is found in association with some *Valvulina* sp., gastropods and plant roots in the dark coloured limestones deposited in reducing conditions (Plate I, figs. 5, 6). The absence of diagnostic marker fossils present some difficulties to determine the chronostratigraphic position of these levels since *Cylindropnrella arabica* Elliott has been usually noted from the Upper Jurassic sequences elsewhere. In the studied samples, the presence of *Mesoendothyra croatica* Gusic in the upper levels of *Cylindroporella arabica* bearing sequence indicates that the stratigraphic distribution of this form should be shifted to older chronostratigraphic levels. Probably represented in the Bajocian? — Bathonian stages this population has also been reported from the same chronostratigraphic levels by some previous studies (Elliott, 1975).

Macroporella sellii .— This form recognized in the algal - foraminiferal wackestones of the Lower Malm series has been used as the biostratigraphic zonal marker of the Lower Malm in apennines and dinarids (Sartoni and Crescenti, 1962; Nikler and Sokac, 1968; Velic, 1977). In the studied area, *Salpingoporella annulata* Carozzi and *Heteroporella anici* Nikler and Sokac are recorded in the Upper Bathonian substage, below this zone, and frequently found in the limestones limited at the top by the Kimmeridgran stage (Plate I, figs. 7, 8). In this zone, *Heteroporelta lusitanica* (Ramalho) and *Cylindroporella? arabicc* Elliott identified in the Zavrakdağı and Gümelektepe sections respectively, are the other important dasycladacean algal species noted in the region (Plate I, fig.9; Plate II, fig. 1).

Clypeina jurassica .— In the studied region, the *Clypeina jurassica* zone corresponds to the *Clypeina* cenozone of Altıner and Septfontaine (1979) and *"Clypeina jurassica* Favre bearing Kimmeridgian beds" of western taurids (Bassoulet and Poisson, 1975). In apennines and dinarids this form is used as biostratigraphic zonal marker in the recognition of the Upper Malm series (Sartoni and Crescenti, 1962; Farinacci and Radoicic, 1964; Nikler and Sokac, 1968; Gusic, 1969; Velic, 1977).

In the studied area, the limestones containing *Clypeina* are massive in character in the lower levels but later dark coloured, medium bedded, alternating with dolomites. An increase in the amount of algae and hydrozoa *(Cladocorapsis mirabilis* Felix) is observed in the levels which show a considerable increase in bed thickness and these levels are characterized by algal packstone facies. Associated with *Clypeina jurassica* Favre, *Pseudoclypeina cirici* Radoicic has been also frequently noted in several stratigraphic sections (Plate I, fig. 10).

Campbelliella striata .— *Campbelliella striata* (Carozzi), known as the diagnostic form of the Tithonian stage, has been used to define the uppermost Malm (Tithonian) zonal marker in apennines and dinarids (De Castro, 1962; Farinacci and Radoicic, 1964; Velic, 1977).

This zone, which also contains *Clypeina jurassica* Favre, is characterized by the formation of brecciation, internal cavities, vadose silt and intense bioturbation. The slow rate or discontinuities in sedimentation sometimes reflect the atmospheric influence. As in the whole Tethys region, in contrast to the rare occurrence of the foraminifers at Jurassic-Cretaceous boundary, a definite increase in the dasycladacean algae is noted by the appearance of the interesting algal species in the study area. In addition to *Campbelliella striata* (Carozzi) and *Clypeina jurassica* Favre, *Salpingoporella annulata* Carozzi recorded in all stratigraphic sections, *Salpingoporclla tosaensis* Yabe and Toyama, *Clypeme parvula* Carozzi, *Aktinoporella pudalica* Praturlon and Radoicic (Yüceyurt section), *Aktinoporella* sp. and *Salpingoporella* cf. *grudii* Conrad, Praturlon and Radoicic (İğdebel section) are the most frequently observed forms (Plate II, figs. 2-6).

LOWER CRETACEOUS ALGAE

Neocomian

It is quite difficult to draw the Jurassic-Cretaceous boundary because of dolomitization and facies similarities of the limestones. In dolomitized Upper Jurassic and Lower Cretaceous series it is quite impossible to recognize the Jurassic and Cretaceous systems. However, in the dolomitic limestone lithologies intercalated within the dolomites, the original texture is partially preserved and the facies analysis make the recognition of this boundary possible. The dasycladacean algae are the most important elements in the recognition of these facies because the textural properties of the limestones are highly monotonous at Jurassic - Cretaceous boundary. Two main dasycladacean groups whose definitions are given below are used to define the lowermost Cretaceous and subdivide the Neocomian.

Salpingoporella annulata - Munieria baconica . — The most remarkable facies types at Jurassic - Cretaceous boundary are the laminated blue-green algal mudstones, coprolithic packstones and dolomites intercalated with the typical Upper Jurassic algal mudstone-wackestone facies. Following the rapid disappearance of *Clypeina jurassica* Favre, *Campbelliella striata* Carozzi, new dasycladacean algal species accompanied by *Salpingoporella annulata* Carozzi indicate the presence of the Berriasian stage. These species show some differences in distribution in the different stratigraphic sections. Among these, *Munieria baconica* (Carozzi), *Aktinoporella podalica* Praturlon and Radoicic and *Clypeina* sp. in the pelletic mudstones of the İğdebel, Çakırlar, Kurudere and Şar sections (Plate II, figs. 7-9), *Clypeina? solkani* Conrad and Radoicic and *Salpingoporella katzeri* Conrad and Radoicic in the section (Plate II, fig. 10; Plate III, fig. 6) and *Salpingoporella steinhauseri* Conrad, Praturlon and Radoicic in the dolomitic mudstone facies of the Çakırlar section (Plate III, figs. 1-3) yield many characteristic sections. *Salpingoporella steinhauseri* Conrad, Praturlon and Radoicic is an imfportant chronostratigraphic marker of the Middle Berriasian in France and Switzerland (Conrad et al., 1973).

Salpingoporella pygmaea. — The lime-mudstone-wackestone facies, characteristic and frequent in the lower part of the Neocomian are overlain by a new facies belt characterized by pelletic and intraclastic grainstones. The most frequent dasycladacean algal species recognized in the grainstone facies is *Salpingoporella pygmaea* (Gümbel) and this form is usually accompained by *Salpingoporella katzeri* Conrad and Radoicic which has already appeared in the underlying mudstone facies (Plate HI, figs. 4-6). This assemblage and the other dasycladacean algal species identified in the various stratigraphic sections of the study area indicate that this sequence is in Valanginian - Hauterivian age. The other accompanying species recognized in the study of stratigraphic sections are *Likanella campanensis* Azema and Jaffrezo (Polatpınar section), *Clypeina marteli* Emberger and *Linoporella?* sp. (Kurudere and Şar sections) (Plate III, figs. 7-10; Plate IV, fig. 1). Among these, *Salpingoporella katzeri* Conrad and Radoicic and *Clypeina marteli* Emberger are the diagnostic fossils of the Berriasian-Valanginian strata in Yugoslavia and Switzerland (Conrad and Radoicic, 1978).

The algal species found in the grainstone facies dissappear in the Barremian and dismicritic mudstone facies rich in algal structures of schizophytoid type become predominant in this stage.

Aptian - Albian

The Aptian and Albian stages are recognized by a thick sequence of frequently occuring blue-green algae lamination and stramatolithic limestones. These facies are usually intercalated with miliolid packestone-grainstone and dasycladacean algae mudstone - wackestone facies whose definition is given below. The overall facies characteristics generally reflect an intertidal-supratidal depositional environment.

Salpingoporella dinarica.— In the eastern taurids, the presence of this algal species has been recognized by Altıner(1981) and later Altıner and Decrouez (1982) defined this form and the accompanying foraminiferal species *Vercorsella scarsellai* (De Castro) as the zonal markers in this belt. The same zone has been previously introduced in Italy and dinarids as corresponding to the Barremian and Lower Aptian stage and substage (Chiocchini et al., 1979; Velic, 1977). In the study area this form has been found together with an assemblage which mainly characterize the Aptian stage. It is usually accompanied by *Salpingoporella hasi* Conrad and Radoicic?*Salpingoporella melitae* Radoicic, *Cylindroporella barnesii* Johnson, *Acicularia antiqua* Pia and the problematic and complex *Bacinella irregularis* Radoicic (Plate IV, figs. 2-7).

UPPER CRETACEOUS ALGAE

In the studied Upper Cretaceous units, the dasycladacean algae have been identified only in one level of the brecciated limestones of the Oğlakkaya section. In this facies, accompained by *Cuneolina pauonia* d'Orbigny, *Heteroporella lepina* Praturlon is easily identified by its typical yellowish calcareous wall. In Apennines and Yugoslavia, this form has been reported from the Cenomanian - Turanian stages (Praturlon, 1966; Berthou and Poignant, 1969).

CONCLUSIONS

By the study of many samples collected from the Mesozoic autochthonous sequence of the Sanz - Tufanbeyli region (Geyikdağı unit), several dasycladacean algal species have been identified in the platform deposits and these identifications made the recognition of the Jurassic-Cretaceous boundary and the subdivision of the Neocomian possible. Thus, in the subdivision of the Neocomian which is considered to be an undivided series in the Taurids, the use of dasycladacean algae seems to be an excellent tool.

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PLATES

PLATE -1

Fig.1— *Griphoporella curvata* Gümbel, Toros dağı section, Tor. 7, X 40.

Fig.2- *Diplopora* sp.. Yüceyurt section, Yü.18, X 10.

Fig.3— *Selliporella donzellii* Sartoni and Crescenti, Zavrak dağı section, Zav. 11a, X 40.

Fig.4— *Teutloporella gallaeformis* Radoicic, Polatpınar section, Pol. 2, X 30.

Fig.5 and 6— *Cylindroporella arabica* Elliott, Çakırlar section, Ça.11, Ça.12, X 40.

Fig.7— *Macroporella sellii* Crescenti, Taşpınar section, Taş.39, X 40.

Fig.8— *Heteroporella anici* Nikler and Sokac, Gümelek Tepe section, Gü. 29a, X 30.

Fig.9— *Heteroporella lusitanica* (Ramalho), Zavrak dağı section, Zav.16, X 40.

Fig.10— *Pseudoclypeina cirici* Radoicic, Yüceyurt section, Yü.44, X 30.

Baki VAROL; Demir ALTINER and Yavuz OKAN PLATE-I


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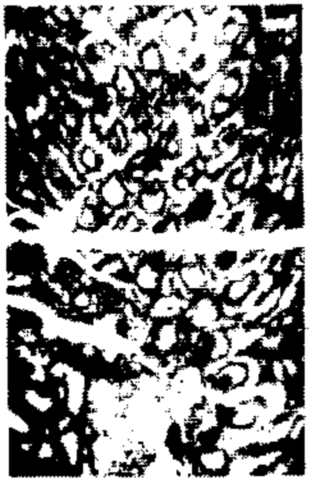
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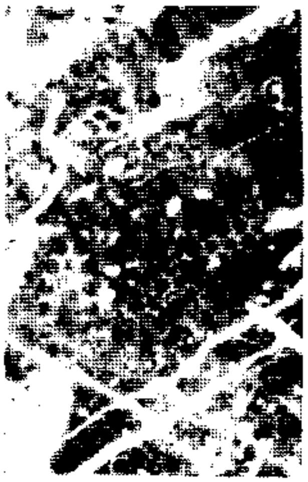
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PLATE - II

Fig.1— *Cylindroporella ? arabica* Elliott, *Salpingoporella annulata* Carozzi, Gümelek tepe section, Gü.29, X 3

Fig.2— *Clypeina jurassica* Favre, Çakırlar section, Ça.21, X 100.

Fig.3— *Campbelliella striata* (Carozzi), *Clypeina jurassica* Favre, *Salpingoporella annulata* Carozzi, İğdebel section, İğ.90, X 20.

Fig.4— *Salpingoporella tosaensis* Yabe and Toyama, Yüceyurt section, Yü.64, X 10.

Fig.5— *Clypeina jurassica* Favre, *Clypeina parvula Carozzi, Aktinoporella podalica* Praturlon and Radoicic, Yüceyurt section, Yü.64a, X 10.

Fig.6— *Salpingoporella* cf. *grudii* Conrad, Praturlon and Radoicic, İğdebel section, İğ.92, X 40.

Fig.7— *Munieria baconica* (Carozzi), İğdebel section, lg.93a, X 30.

Fig.8— *Clypeina* sp., İğdebel section, İğ.93a, X 40.

Fig.9— *Aktinoporella podalica* Praturlon and Radoicic, İğdebel section, lg.93a, X 30.

Fig.10— *Clypeina ? solkani* Conrad and Radoicic, Kurudere - Şar section, Şar. 18, X 30.

Baki VAROL; Demir ALTINER and Yavuz OKAN PLATE-II


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PLATE - III

Fig.1,2 and 3— *Salpingoporella steinhauseri* Conrad, Praturlon and Radoicic, *Salpingoporella annulata* Carozzi, Çakırlar section, ga.26, X 100.

Fig.4 and 5— *Salpingoporella pygmaea* (Gümbel), Polatpınar section, Pol.32 and 33, X 100.

Fig.6— *Salpingoporella ? katzeri* Conrad and Radoicic, Kurudere - Şar section, Şar.12, X 40.

Fig.7— *Clypeina marteli* Emberger, Kurudere - Şar section, Kur.13, X 50.

Fig.8— *Linoporella* sp., Kurudere - Şar section, Kur.13a, X 40.

Fig.9 and 10— *Likanella campanensis* Azema and Jaffrezo, Polatpınar section, Pol.32 and 33, X 50.


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PLATE - IV

Fig.1— *Salpingoporella pygmaea* (Gümbel), Kurudere - Şar section, Kur.13, X 40.

Fig.2— *Salpingoporella dinarica* Radoicic, Toros dağı section, Tor.72, X 10.

Fig.3 and 4— *Salpingoporella hasi* Conrad, Radoicic and Rey, Oğlakkaya section, Oğ.19, X 70.

Fig.5— *Cylindroporella barnesii* dohnson, Oğlakkaya section, Oğ.40, X 70.

Fig.6— *Salpingoporella ? melitae* Radoicic, Toros dağı, section, Tor.72, X 70.

Fig.7— *Acicularia antiqua* Pia, Toros dağı section, Tor.74, X 30.

Fig.8— *Heteroporella lepina* Praturlon, Oğlakkaya section, Oğ.10, X 30.


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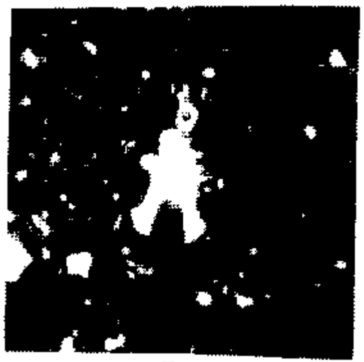
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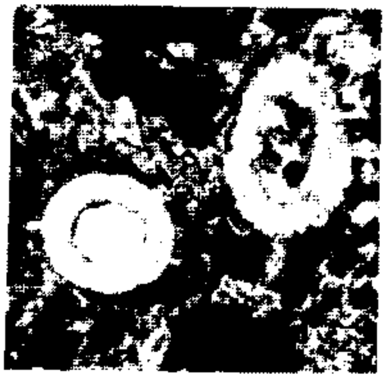
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Mineral Res. Expl. Bull., 108, 57-63, 1988

CONTRIBUTION TO THE STRATIGRAPHY OF THE MIDDLE DEVONIAN IN THE SURROUNDINGS OF ADAPAZARI, NORTHWEST TURKEY

Orhan KAYA* and Rudolf BIRENHEIDE**

ABSTRACT.— The basal part (Alabalık member, new name) of the Middle Devonian limestone unit (Yılanlı formation) exposed in the surroundings of Adapazarı contains a well preserved prolific fauna of rugosa and tabulata (corals). The majority of the faunal elements is conspecific or comparable with such species which have been recorded mainly in the Middle Eifelian of Eurasia and North Africa. Thus the Alabalık member is very probably of the same age. The following species identified by Birenheide are new : *XystriphyHum kayai* n. *sp.,Dohmophyllum bulbosum* n. sp., *Favosites dorotheae* n. sp., *Mariusilites osmanicus* n. sp., *Mesolitesinterruptus n.* sp. and *Heliolites usiaeminoris* n. sp.

INTRODUCTION

In northwestern Turkey the Lower and Middle Devonian are the oldest well-known stratigraphic units with respect to their prolific fauna (Fig. 1).

In Istanbul Penck (1919) and Paeckelmann (1925, 1938) established the present outlines of the Devonian stratigraphy, and important contributions then came from numerous authors. An alternative stratigraphic classification of the Lower and Middle Devonian rocks was proposed by Kaya (1973) on the basis of faunal descriptions given by Sayar (1962), Haas (1968), Kullmann (1973), Gandl (1973) and Carls (1973). This constituted a basis for the generalized stratigraphic succession of the Paleozoic of Adapazarı (Fig.2).

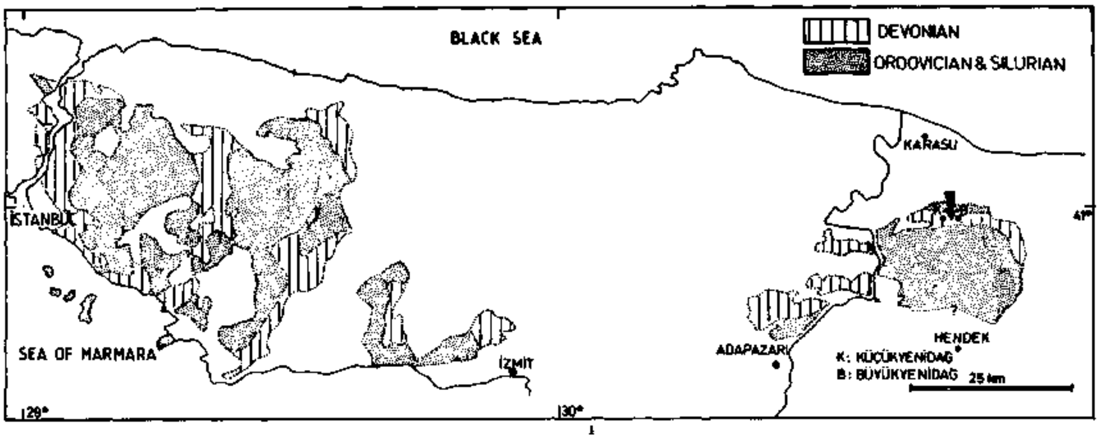

Fig. 1 — Generalized geographical distribution of Ordovician - Silurian and Devonian rocks of the İstanbul and Adapazarı areas : modified from the geological map of Turkey, 1:500,000. Black arrow = position of the road cut between K = Küçük yenidağ and B = Büyük yenidağ (G25 -a2, 05.72 : 38.78).

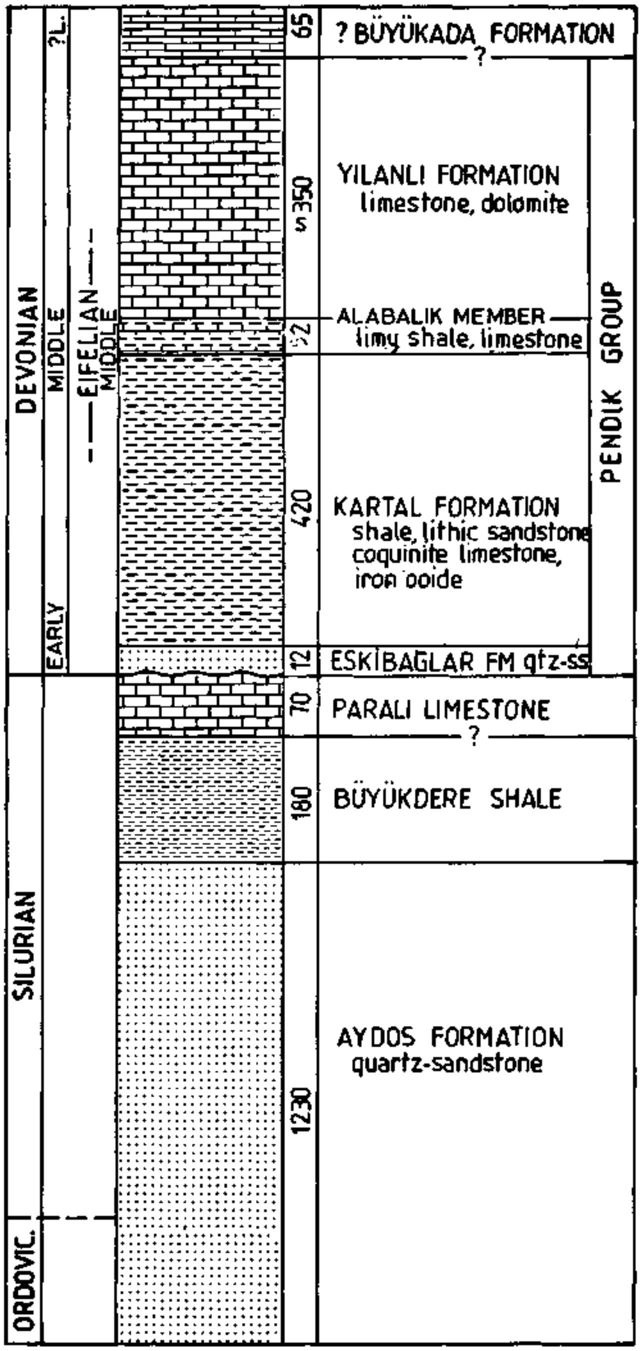

Fig. 2 — Generalized succession of the Paleozoic rocks in the Küçükyenidağ - Büyükyenidağ study area to the east of Adapazarı.

Outside Istanbul the Devonian rocks have attracted less attention. In the study and contiguous areas reconnaissance surveys of the Devonian have been made by Kleinsorge and Wijkerslooth (1940),Baykal (1955) and Türkünal (1957). Kipman (1974) proposed a stratigraphic succession for the study area, which is widely modified in the present paper.

Over extensive areas in northwestern Turkey the Devonian comprises primarily an Early to Middle Devonian mudrock sequence and a Middle Devonian carbonate sequence. However, both sequences show strong lateral changes in lithology and stratigraphic relations. The thick and pervasively dolomitized carbonate sequence of the western Pontides (Saner et al., 1980) has its westernmost exposure in the study area. The basal part of the carbonate sequence, overlying the primarily Early Devonian mudrock, contains a coral fauna which may serve as a satisfactory time marker for the boundary between the mudrock and overlying carbonate sequences.

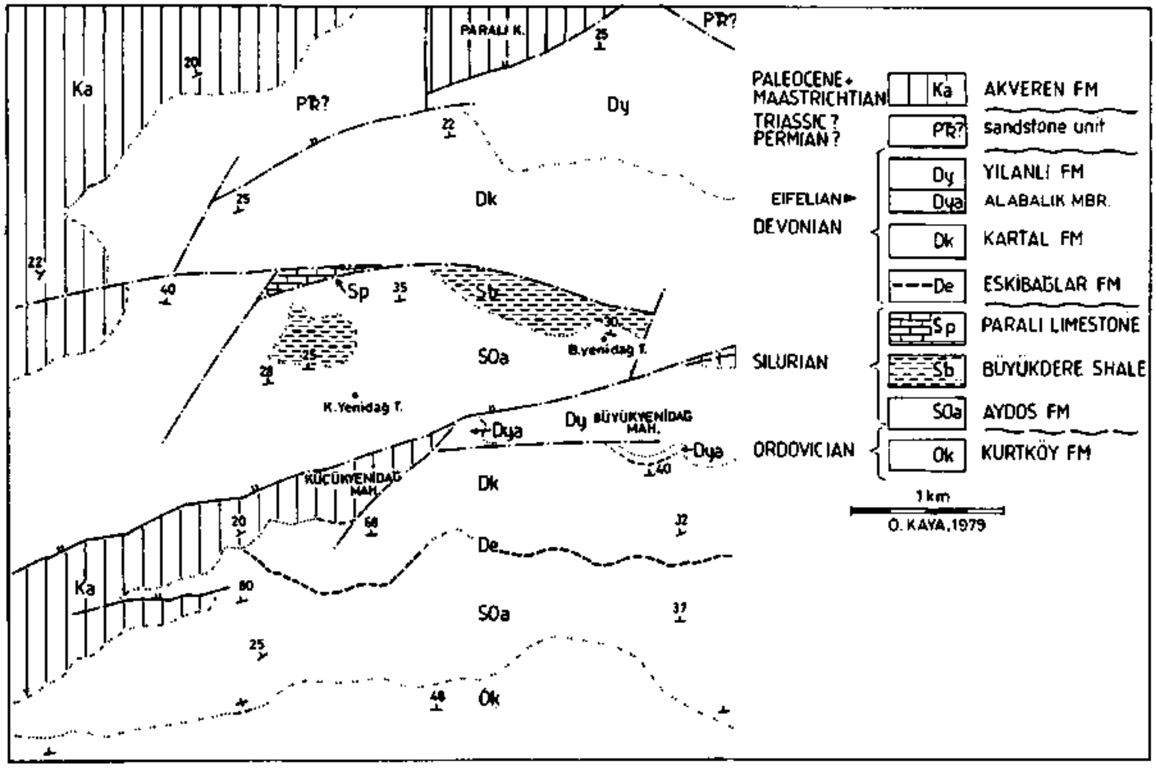

Fig. 3 — Geological map of the study area.

The coral-bearing beds are exposed between Küçükyenidağ and Büyükyenidağ villages (Fig. 1), at the road cuttings. The stratigraphical position of the coral-bearing beds within the Alabalık member of the Yılanlı formation is presented on the column section of Fig.4.

The coral fauna given herein by Birenheide was collected by Kaya during recent years. The thin sections were prepared and are stored in the Geological Department of the Senckenberg Institute, Frankfurt am Main (SMF — numbers). Detailed descriptions of the new species are in preparation.

LITHOSTRATIGRAPHY

Eskibağlar formation

The name Eskibağlar was used by Kaya (1973, p. 9) for the lower member of the Gedinnian İstinye limestone mainly composed of white quartz - sandstone rocks. The reference section is exposed in İstanbul.

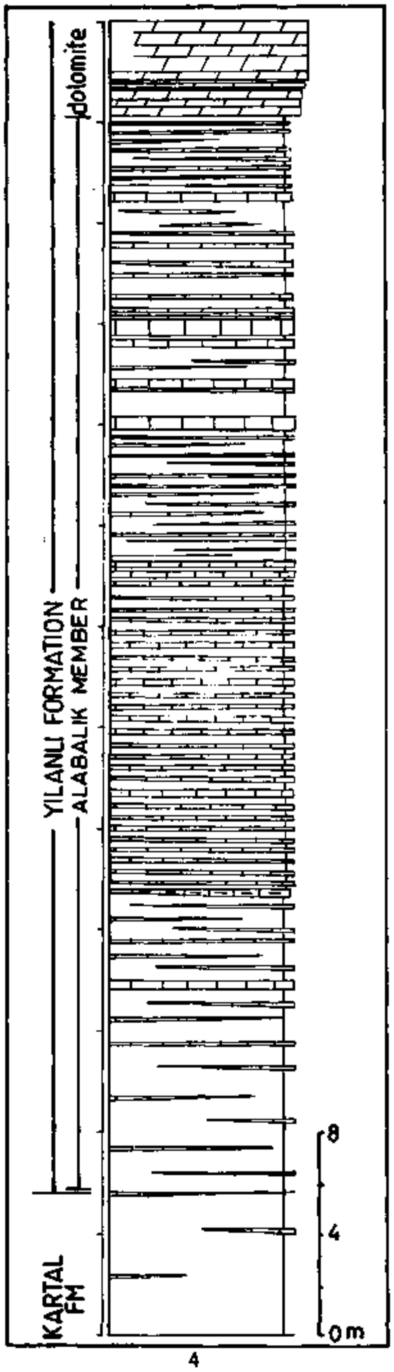

Fig.4 — Type section of the Alabalık member at the base of the Yılanlı formation. The coral - bearing beds are situated within the middle part of the sequence.

The Eskibağlar is a diachronous unit occurring beneath the Kartal formation from the study area further westward up to the east of Istanbul (Kaya, in prep.). Thus,-the rank of this unit is here elevated to a formation. In the study area the Eskibağlar formation consists uniformly of white, massive and homogeneous, recrystallized quartz arenite and fine — grained quartz conglomerate. The Eskibağlar formation rests unconformably on the Silurian rocks.

Kartal formation

The name Kartal formation was designated by Kaya (1973, p. 12) for the thick sequence of grey, variably calcareous mudrocks with intercalations of thin - bedded coquinite limestones and lithic sandstones. In Istanbul, at its type locality, it corresponds to the lower clastic part of the old Lower Devonian classification (Paeckelmann, 1938). In the study area it replaces the Bıçkı formation of Kipman (1974, p. 17) and parts of the underlying Yayla and of the overlying Kabalak formations.

The mudrock consists of variably yellowish-grey weathering, thickly stacked illitic mudshale and clayshale. Abundant intercalations of lithic and quartzitic sandstone occur in the lower part. The coquinite limestone interlayers are present abundantly in the upper parts of the formation. Sedimentary iron ore layers, up to 6m thick, occur about 60m below the top of the formation (Kipman 1974, p.190).

In the study area the Kartal formation is only sparsely fossiliferous. Because it is gradationally overlain by the coral - bearing beds of the Yılanlı formation, the object of this study, an upper age limit of Early Eifelian is suggested for the Kartal formation.

Yılanlı formation

The name Yılanlı formation was applied by Saner et al. (1980, p.113) for the sequence consisting of grey limestone, dolomite and limy mudrocks, being widely exposed in the western Pontides. It is partly the lithic equivalent of Kaya's (1973, p.14) Kozyatağı limestone and the time equivalent of the limestones of the Büyükada formation in İstanbul.

Table 1— List of the coral species (left) and related or same species described elsewhere (middle), with reference to their stratigraphical ranges (right)

Coral species of Adapazarı	*Same or similar species recorded elsewhere*	*Stratigraphical range of compared species*
Cyathophyllum (C.) spongiosum (Schulz,1883)	*Cyathophyllum (C.) spongiosum*	Middle Eifelian (abundant) to Early Givetian (rare)
Spongophyllum sedgwicki Mine - Edwards &Haime,1851	*Spongophyllum sedgwicki*	Late Eifelian to Givetian (rare)
Xystriphyllum kayai Birenheide n. sp.	*Xystriphyllum glinskii*	Middle Eifelian (Junkerberg Fm.)
Dohmophyllum bulbosum Birenheide n. sp.	*Dohmophyllum helianthoides*	Early Eifelian (rare), Middle Eifelian (abundant), Early Givetian
Acantophyllum cf. *filosum* (Wedekind,1923).	*Acanthophyllum filosum*	Early to Middle Eifelian (rare spec., Nohn to Ahrdorf Fm.)
Acantophyllum heterophyllum (Mine - Edwards & Haime,1851)	*Acanthophyllum heterophyllum*	Eifelian (abundant) to Early Givetian (very rare)
Favosites dorotheae Birenheide n. sp.	*Favosites regularissimus*	Late Emsian (rare) to (Early) Eifelian (? abundant)
Favosites aff. *styriacus* Penecke,1984	*Favosites styriacus*	Emsian (? abundant)
Favosites cf. *saginatus* LeCompte,1939	*Favosites saginatus*	Middle to Late Eifelian (abundant)
Mariusilites osmanicus Birenheide n. sp.	*Mariusilites elegans*, e.p. *M. germanicus*	Eifelian Eifelian, Ahrdorf Fm.
Caliapora cf. *venusta* Janet,1972	*Caliapora venusta*	Givetian
Mesolites interruptus Birenheide n. sp.	*Mesolites multiperforatus*	Eifelian
Alveolites straeleni LeCompte,1939	*Alveolites straeleni*	Early to Middle Eifelian (abundant), Late Eifelian (rare)
Squameoalveolites sp., aff. *fornicatus* (Schlüter,1889)	*Squameoalveolites fornicatus*	Eifelian (abundant in Middle Eifelian Junkerberg Fm.)
Heliolites asiaeminoris Birenheide n. sp.	*Heliolites barrandei; H. Porosus lindstroemi ; H. vulgaris*	Givetian ; Late Emsian to Middle Eifelian ; Eifelian

The limestones are composed of dark gray, massively to thickly bedded medium to coarse grained skeletal — fragmental micrites and sparites. The dolomitization, which is controlled by structure, is pervasively developed in the upper part of the formation.

The Yılanlı formation rests gradationally on the Kartal formation; the Alabalık member represents the interval of the transitional beds.

On the basis of the coral fauna of the Alabalık member a lower age limit of Middle Eifelian can be suggested for the Yılanlı formation.

Alabalık member.— The name Alabalık member is here applied to the limy mudrock sequence with intercalations of limestone and small - sized coral build-ups (Fig.4) at the base of the Yılanlı formation. The type section is exposed at the geographical coordinates 05.72 : 38.78 of the topographic sheet G25-a2, at the eastern flank of the Alabalık river between Küçükyenidağ and Büyükyenidağ; villages (Fig. 2). The member was designated by Kipman (1974, p.20) as the Manastır nodular limestone member.

The limy mudrocks are dark grey and unevenly bedded. Laterally discontinuous skeletal-fragmental limestones and individual colonies or small-sized build-ups of corals occur abundantly in the middle part of the member. The limestone intercalations increase in thickness and abundance upward in the Alabalık member, grading into the main body of the Yılanlı formation.

The Alabalık member overlies gradationally the Kartal formation.

The rugose and tabulate corals indicate a predominantly Middle Eifelian age for the Alabalık member. For species and their stratigraphic range see Table 1.

STRATIGRAPHICAL CONCLUSIONS

From the compilation given in Table 1, it is immediately evident that the majority of the species are widespread in the Middle Eifelian deposits of Eurasia and North Africa, with special relations to the Middle and West European realm. A relation to Devonian species of the Eastern Alps might at first be expected, but this does not seem to be present. This may have as its reason the poor development of Eifelian marine deposits within that realm. It is an open question as to whether only the Alabalık member represents Eifelian age within the Devonian succession of Adapazarı, or whether this geological time unit is continued in the overlying parts of the Yılanlı formation as well as in the underlying upper parts of the Kartal formation. We have some reason to assume that the greatest part of Eifelian time is represented only by the Alabalık member; this is evident by the occurrence of a small number of coral species which have their bulk in Givetian time, as *Spongophyllum sedgwicki* and *Caliapora* cf. *venusta,* or in Late Emsian to Early Eifelian time, as the *Favosites styriacus* species group.

At present we have only very little knowledge of fossils of the thick-banked Yılanlı limestones which are dolomitized throughout; compared with the similar development of the Givetian reef limestones of the Rhenish Mountains it cannot be excluded that at least their majority is of Givetian age, too.

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Mineral Res. Expl. Bull., 108, 65-75, 1988

AIRPHOTO INTERPRETATION AS AN AID TO LITHO - STRUCTURAL MAPPING IN TROPICAL TERRAIN - THE NEW FEDERAL CAPITAL CITY SITE, ABUJA, NIGERIA

Murat AVCI*

ABSTRACT . — The Federal Capital Development Authority of Nigeria requested that an engineering geological investigation of the proposed New Federal Capital City Site at 1: 25.000 scale be carried out for suitability purposes. Mapping with conventional field methods alone proved to be insufficient due to the concealed terrain conditions in the area. Two different scales (1: 40.000 and 1:10.000) of aerial photographs were used which have revealed three main bedrock units and the entire structural framework of the area. Airphoto interpretation data shows lithological banding running NNE — SSW and fractures developed mainly in NW — SE and NE — SW directions. Through this information a major tectonic event has been inferred, resulting from a principal stress acting WNW—ESE.The directional frequency and the length of fractures show variations according to rock type. The metaigneous rocks gave high fracture density and frequency values, whereas metasedimentary rocks yielded low values of the same parameters.

INTRODUCTION

A detailed engineering geological study of the southern section of the New Federal Capital City Site was carried out in 1979 by the Geology Department of the University of Ife, Nigeria. A similar study was performed during the same period by the University of Ahmadu Bello, Zaria, Nigeria for the northern section of the city site. Because of the particular usefelness of airphoto interpretation in the area, only the photogeological part of the entire study is presented in this paper. The other aspects of the investigation are explained in detail, in the UNIFE project report (1979). The main features, investigated in this study are surface structures, regional lithological units and fracture traces.

The work of Blanchet (1957), Lattman (1958), Norman (1976) illustrate the use and capabilities of large scale remotely sensed data for engineering applications. Most of these publications emphasize the importance, of a fracture interpretation in concealed terrain. Ray (1960) provides several stereograms as examples pertinent to terrain conditions of the type found in Nigeria. Thomas (1974) discovered that the lineaments traced from aerial photographs were the framework of large structural block patterns in Williston Blood Creek Basin (USA) and that these features were not evident on the conventional geological maps.

This paper illustrates a case in which aerial photographs have been of particular value for foundation engineering. It also provides relevant information on photo interpretation for site investigation Upon which other kinds of studies can be effectively implemented.

The study area covering approximately 200 km^2 stretches between Latitudes 8° 45' and 9°00' North and Longitudes 7° 15' and 7°30' East. The topography is mature and has a low relative relief except for hills of various sizes which rise above the general topography. Several mature profile rivers drain the area westward to the river Usuman which runs south to join the river Niger.

THE ABUJA AREA

The new Federal Capital Site lies in the tropical savannah vegetation zone where there is fairly complete soil and vegetation cover. This presents problems for geological field investigations. For example, conditions such as delineation of lithological boundaries, trace of fractures and general structural trends are difficult to solve solely by field traversing.

Initially, five geologists were sent to the field to start the preliminary geological mapping without airphoto interpretation data. Thus all that could be done was together some local geological information in accessible localities. As a result, only scattered information was assembled and no conclusion could be drawn about the main structural framework of the area. At this point, it was realized that the remotely sensed data were needed.

The field team was not able to identify the major fractures and litho-structural units but the study of aerial photographs described below identified 276 fractures and three main litho-structural divisions

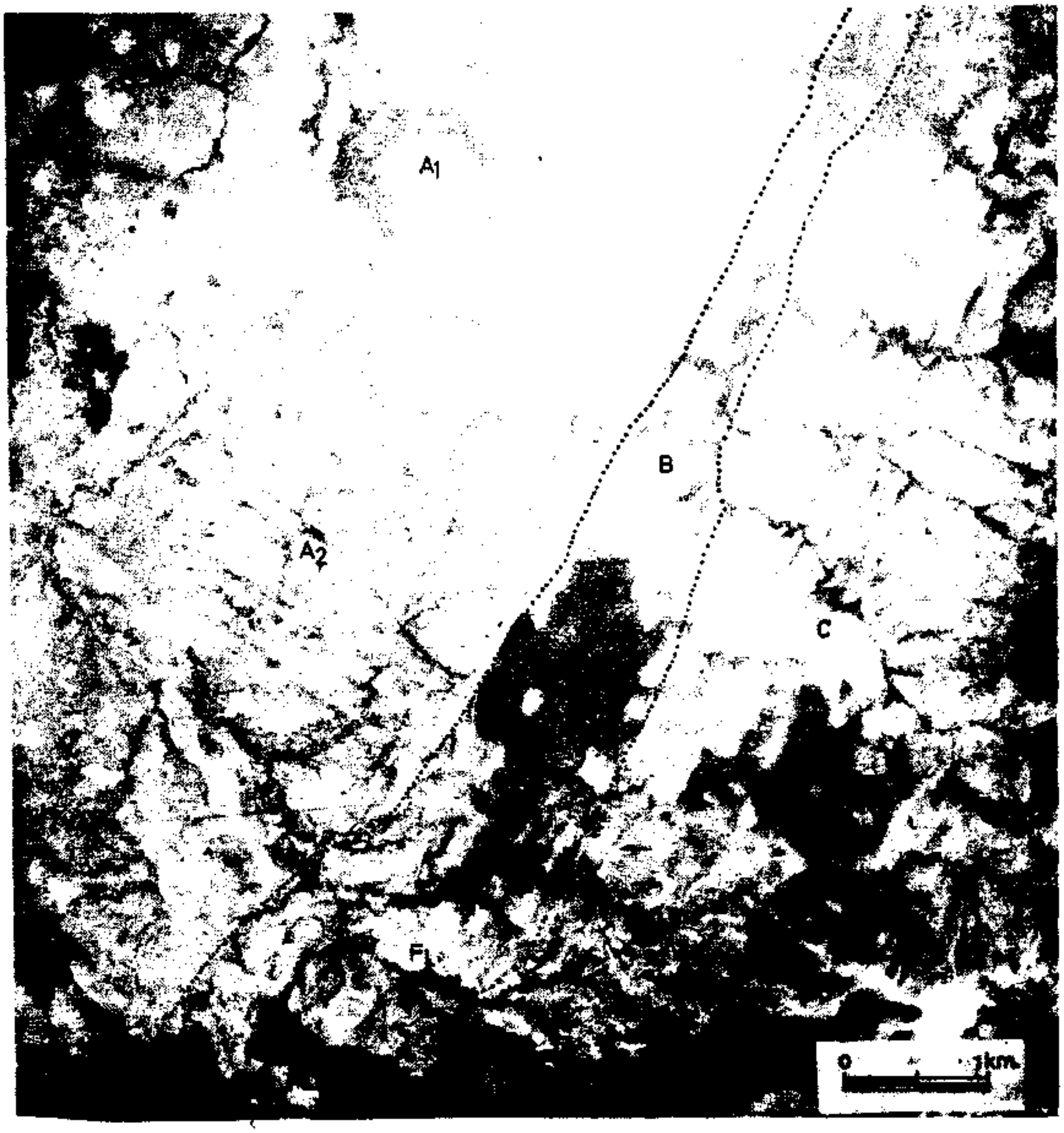

Fig. 1 — Aerial Photograph (1: 40.000) showing three Lithological units, A_1 Migmatite, A_2, Granite gneiss, B, Quartzite, Amphibolite and Banded gneiss and C, Schists. At FL, the beds show drag effect due to faulting. Line H—J is the southern boundary on the map (Fig.5). The Northern boundary of the photograph is shown on the map (K— L).

in the area. This information then enabled the field geologists both to collate the scattered field observations and to prepare a traverse plan which formed the oasis for the geological site investigation.

GEOLOGICAL SETTING AND REGIONAL TECTONIC CONTEXT

The area is underlain by Nigerian Basement Complex rocks of Precambrian age. Studies by Grant (1971), McCurry (1976), and Ball (1980)show that the Nigerian Basement Complex rocks occur in the Mobile Zone of Pan-African reactivation between the West African Craton, to the west, and Congo Craton to the southeast. Radiometric ages obtained on different rock types using several radiochronometers show that the Nigerian Basement Complex is polycyclic Grant (1971), Rahaman et al. (1983), Rahaman and Lancelot (in press). The evidence of deformation in the rocks observed in the field also confirm the polycyclic nature of this Basement Complex

Most workers agree that the Pan-African orogeny and the associated plutonic phase which is approximately 600 million years have extensive influence on the rock characterization and deformational structures in the region (Grant, 1971; Traswell and Cope, 1963; Oversby, 1975; Van Bremeen et al., 1976).

Two distinct rock groups are distinguished in the area:

1— Metaigneous and older granite group (i.e. granite gneiss, migmatite, biotite granite, porphyritic granite and granodiorite) and

2 — Metasedimentary group(i.e.quartzite,amphibolite and banded gneiss, mica schist, quartz schist and amphibolite schist).

A sharp escarpment clearly visible on both scales of aerial photographs separates the two rock groups which has been interpreted as a sheer zone (USGS, 1977) passing through the west of Kusaki (Fig. 5) in the South extending in a NNE—SSW direction (approximately N30°E). This shear zone is the extension of the shear zones observed by Ball (1980) in the Hoggar region North Central Africa, which the trend of these reasonably distant shear zones is very probably related to the deformational mechanism that is affected by the proximity and location of the cratons with respect to both areas.

Both the shear zone and the granitic plutons are cut by NW - SE and NE - SW shear faults in the area (Figs. 3 and 5). This was also observed by Ball (1980) and McCurry (1976) in the northern extent of the Pan-African Orogenic Belt. The granitic plutons

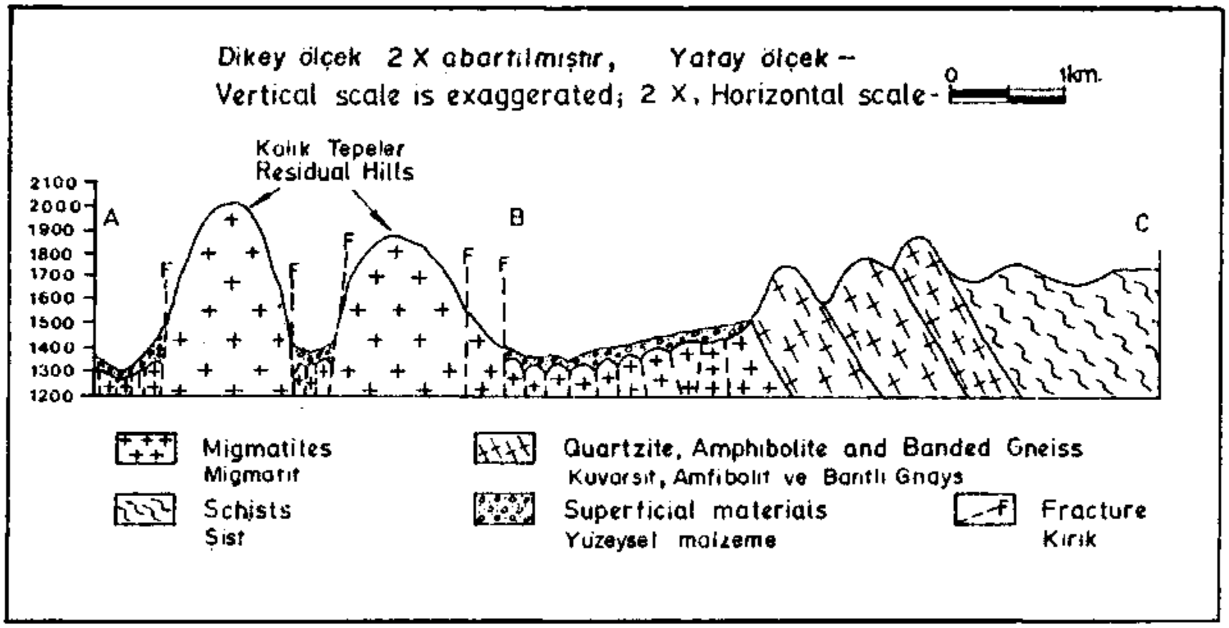

Fig. 2 — Geological cross-section along A, B and C on the map (Fig.5).

are dated as syn-to Post-Pan—African Orogeny (Grant, 1971; Traswell and Cope 1963 and Bremeen et al., 1976). Using this dating, an inference of more or less 600 my can be made as the age of the fractures. However, elsewhere in Nigeria, it was observed that the Cretaceous sediments are cut by some major faults which may have been the result of the reactivation of post Pan-African fractures (Avcı, unpublished). This faulting is the latest activity affecting the Basement Complex since the late Cretaceous period

MATERIAL AND METHODOLOGY

The remotely sensed data employed in this study are of two types :

1 1 : 40. 000 scale black and white infra-red poor quality photographs covering a large regional area, and

2. 1 : 10.000 scale black and white panchromatic good quality photographs which covered confines of the City site.

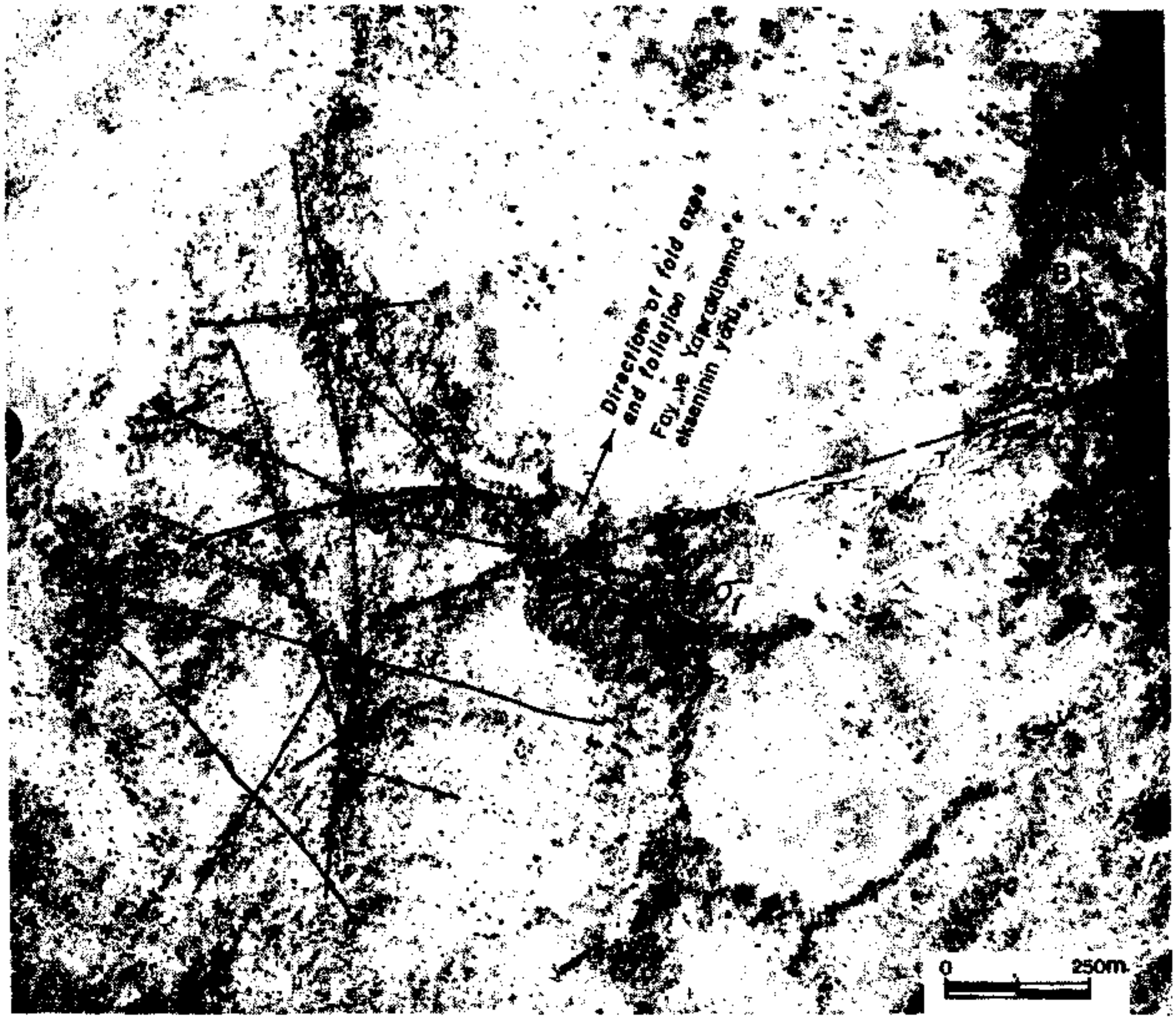

Fig. 3 — Aerial photograph (1 : 10.000) showing fractures cutting accross both the metaigneous (A) and metasediments (B) F_1 is a Shear fracture along which tin mineralization has developed. The trenching is clearly visible at T F_2 and others are also shear faults developed as a result of the same principal stress O_1

The 1 : 10. 000 scale photographs helped in solving many local scale problems. Several kinds of topographic maps also proved to be reasonably reliable when used in combination with the aerial photographs.

Geological information about the area is rather scarce. However, since the area was selected as the New Federal Capital City Site, some investigations have been carried out. These are the reports by USGS(1977), University Ahmadu Bello, Zaria (U.A.B. 1979) and the University of Ife (UNIFE 1979).

The preliminary photo interpretation started with small - scale photographs (1:40,000) in order to make a quick regional analysis of the area. This led to the identification of two major litho-structural units which have been the fundamental information of the entire study. At the 1:40,000 scale, it was also possible to discriminate some internal characteristics of each unit, and many fractures could also be traced.

The interpretation of 1:10. 000 photographs considerably improved the quantity and quality of information obtained. For example, it was possible to identify another unit within the metasedimentary group which could not be distinguished on the 1:40.000 scale photographs.

On the other hand, recognition of the. patterns representing the litho-tructural units on 1:10. 000 scale was not possible as the pattern development is poor on large scale photographs. However, the most comprehensive analysis of the geologic condition resulted from a complementary study of both photographs at both scales. For example, when an observation was made on 1:10. 000 photographs, it was necessary to examine the 1:40.000 scale photographs as well in order to correlate features with the regional structure in the area. And conversely, interpretations made with the 1:40. 000 scale were checked by 1:10.000 scale photographs for the final interpretation

AIRPHOTO PATTERN AREAS AND LITHO - STRUCTURAL UNITS

Morpho-structural units as ground features are very large. In the field it is not easy to observe and to study them, particularly when the terrain is poorly exposed on aerial photographs, the scale of these features is such that patterns are visible and reflect areas of similar morpho-structural characteristics.

In the study area, these morpho-structural patterns have evolved as a consequence of lithology and structure and determined the litho-stiwctural units. On the basis of these units, a programme for traverses and field work was prepared in such a way that each unit was appropriately field checked and sampled.

Two major litho-structural units were identified and several other subunits were distinguished as follows :

1. Area of metaigneous and older granite rocks

Granitic rocks cover 60% of the study area. The area is characterized by residual hills which are in places clustered but mostly occur as isolated individual hills. They are controlled by fractures and have slopes that are very steep (Fig.2). The tops of the hills are rounded and the process of conchoidal weathering (exfoliation) is active on all of them. The fractures across the hills have two dominant orientations (NE — SW and NW — SE) and appear as rectilinear narrow depressions which are readily identified by concentration of vegetation, a grey tonal quality on aerial photographs and by the presence of straight water courses (Fig. 1).

Three types of residual hills were distinguished which correlate with their underlying lithology :

a. Well formed, residual hills which occur in the northern area, developed in porphyritic granite, granodiorite and biotite granites all of which have similar characteristics in terms of erosion.

b. Clustered and sometimes elongated, poorly formed residual hills in the South which formed in the granite gneiss area. The formation of these hills apparently reflects the gneissic structure and are elongated parallel to the general foliation of the rock.

c. Well formed, larger, massive residual hills which have formed in migmatitic rocks. The relief of the hills in this rock type is related to the physical characteristics of the rocks which are the most resistant rock type in the area.

2. Area of metasedimentary rocks

These rocks cover the remaining part of the area and are divided into two subunits :

a. Ouartzite, amphibolite and banded gneiss : The unit consists of alternating bands of quartzite, incorporated in amphibolite and banded gneiss which are considered as one lithologic group. The thickness of the unit varies between 50 m to 1500 m.

The metasedimentary rocks display a typical trellis type drainage with a NNE—SSW trending parallel ridge and valley pattern and is readily recognised on the aerial photographs (Fig.l).

Few fractures cut across the structural trends at high angles which are the extensions of fractures developed in the metaigneous area which die out or transform into another type of deformation in this unit as a result of the different physical characteristics of the rocks. An example of this is observable along the Wosika River as a flexure at location FL in Figure 1. In this area, it was easier to trace the faults due to the off-sets produced by faulting along the structural trends (Figs.3 and 5).

b. Schists: Towards the East, beyond the unit (a), the area is underlain by mica schists and amphibolite schists which are structurally associated with unit (a) as the strike and dip of both are similar. Field measurements along a traverse from Kusaki to Buze village have shown a strike of N30°E and an average dip angle of 60°. However, deformation in these rocks has changed the trends of strike and the attitude of dips dramatically as observed in the field.

The metasedimentary unit was investigated by traverses accross the structural trends. Three traverse lines were planned along routes that were designed to retrieve the appropriate information and to collect samples.

FRACTURE ANALYSIS AND GEOTECTONIC INTERPRETATION

There is a NNE — SSW orientation of the lithological banding observable both in the aerial photographs and in micro texture of all the rock types in the area. This orientation is clearer on the aerial photographs in meta-sedimentary formations which are more readily distinguishable from the metaigneous rocks (Fig.l). In the metaigneous rock area there are only few hills with elipsoid shape and valleys which are elongated parallel to the direction of the general structural trend.

The dense fracture system and the prominent lithological orientation show that the area has been affected by a major tectonic event (Pan - African Orogeny). This event caused deformation of the rocks and produced certain strtictural features. An indication of the polycyclic nature of the Basement Complex is the chaotic micro structure imprinted in the rocks, but some minor folds are preserved with their axial planes parallel to the general lithological banding within the gneissic rocks. Major foldings, mentioned by Benkhelil in UNIFE (1979), was not observed in the area. According to the present deformations on the rocks the magnitute of Pan-African orogeny was too high to expect to find primary structural features in the area.

The analysis of photo-interpretation data and observations in the field made it possible to determine the principal stress direction in the area. Strike slip faults, oriented in NW - SE and NE - SW directions were used to elucidate this notion (Figs. 3 and .5). The strike slip sense of displacement of the faults is observable where they cut and displace the NNE — SSW lithological bandings. The proposed principal stress model, inferred from the present fracture system in the area is comparable to the models proposed both by Moody and Hill (1956) and Hobbs et al. (1976). These fractures are shown on the photogeological map (Fig. 5).

Experiments have shown that faults can represent shear fractures which develop at an angle of less than 45° to the direction of principal stress with the fold axes normal to it (Hobbs et al., 1976) . However, considering the uncertainity and the variability of the conditions (Cook, 1969; Friedman, 1972) in this principal stress, the direction of foliation and the fold axes were also used in identifying the direction of principal stress field in the area (Fig.3).

Figure 4a is a directional frequency diagram plot of 276 fractures for the whole area. The fractures oriented 140° — 165° , all of which represent the NW — SE shear fault directions, are the longest and most frequent. The small peak at 70° is the NE — SW shear faults. In this case, the very small cluster along 110° represents the extension fractures developed along the direction of principal stress (O_1). The clusters at 20° and 35° directions are the extension fractures formed along O_1 and parellel to the fold axes and foliation plane which postulates the presence of principal stress direction in a WNW — ESE trend.

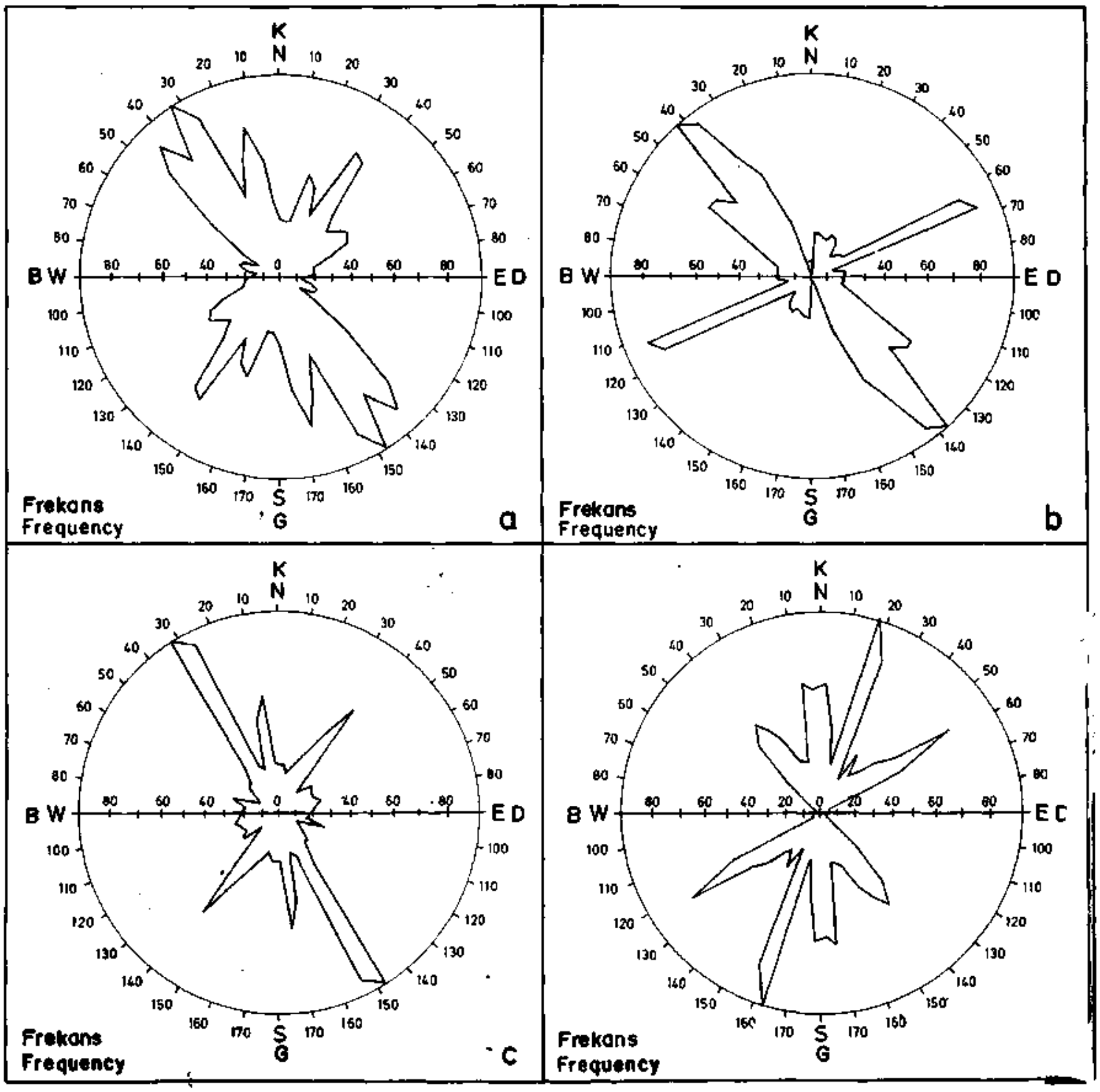

Fig. 4 — Rose diagrams plots of variable amount of fractures from different parts of the study (The class interval for each figure is 5 degrees.)

Considering the formation of Pan - African Orogeny by the movements of the two rigid craton masses as mentioned earlier, the principal stress field obtained is in concordance with the geographical location of the cratons as the sources of compression which are also in a WNW — ESE direction with respect to the Nigerian Basement Complex.

Rose diagrams (Figs. 4b, c and d) were prepared from data from the different parts of the area to observe the effect of the deformation characteristics on different rock types. A comparison of these diagrams with each other shows visible differences which are attributable to the different rock characteristics. For example, Figure 4b is a frequency diagram of 40 fractures occurring in the migmatite area. These are shown to occur in 2 clusters at 65° and 140°. However, NW — SE oriented fractures are dominant and frequent and the two main clusters on the diagram are the shear fractures reflecting the criss-cross fracturing. These indicate an unimodal rock character. The (O_1) principal stress direction can be inserted along 100° cluster set.

Figure 4c is a frequency diagram of 50 fractures from a porphyritic granite area where major faults have clustered at 150°. Clusters at 40° appears to be the representation of NE — SW shear fractures. The (O_1) direction is around 105°. The irregular distribution of the frequency directions reflects a near unimodal rock characteristic.

Figure 4d represents 51 fractures mapped in the southwest corner of the area that is underlain by granite gneiss and banded gneiss. The diagram illustrates a different trend of fracturing in this part of the area. The main frequency directions of fracture in this area are in NNE — SSW direction, the highest cluster is around 20° NE and also a great concentration in a northern direction. This change in frequency direction of fracture, is due to the strong foliation direction which parallels the direction of weakness in these rocks. The clusters at 60° and 140° are the shear fractures; hence, (O_1) principal stress direction at 100° southeast.

Directional frequency diagrams of fractures alone are not adequate information for analysis. Thust, other parameters, like fracture density and fracture frequency calculations, were also carried out.

For example, relation ship of the number of fractures and their length to surface area depends on the strength of the substrate. By calculating the fracture density and fracture frequency, the relative visible state of brittleness of an area can be more realistically determined.

The fracture density is defined as the total length of all the recorded fractures divided by the area that include these fractures.

$$\mathbf{Fd} = \frac{\Sigma L}{A}$$

Where Fd = Fracture density
L = Total length of all fractures
A = Area

One hundred and seventy four kilometers of fractures were identified in an area of 142 km^2 within igneous rocks that underlie the area of the City Site.

$$\textbf{Therefore : FD} = \frac{174}{142} = 1.225$$

This fracture density value falls in the range of high fracture density (Vielon et al., 1976).

Another important quantitative measurement is the fracture frequency, defined as the number of visible fractures per unit area.

$$\textbf{i.e. Ff} = \frac{\Sigma F}{A}$$

Where Ff=fracture frequency
F = total number oı fractures
A = Area

$$\textbf{Thus : Ff} = \frac{276}{142} = 1.943.$$

This is a medium to high fracture frequency according to Vielon et al. (1976). The density and frequency calculations for metasedimentary area showed very low values that reflect their plastic characteristics.

These simple statistical analyses outlined above show that the area is fairly densely fractured. However,

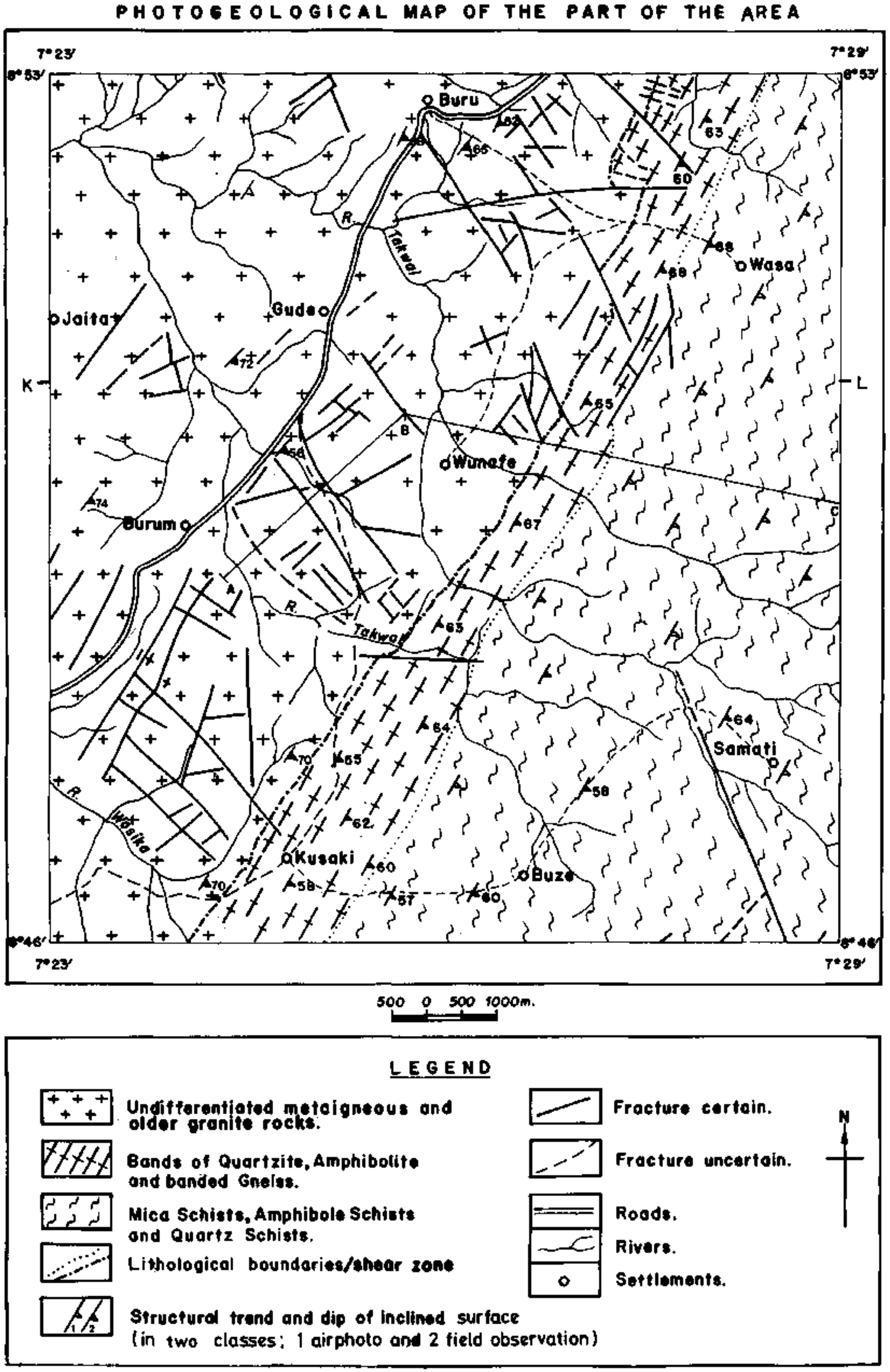

Fig. 5 — Photogeological map of the part of the area.

information on the tectonic history of the area shows that Nigerian Basement Complex rocks were stabilized after the Pan-African Orogeny (600 million years ago or less) and related plutonic phase(Grant,1971). There are no written or oral records of earthquakes in the region.

The area underlain by undifferentiated metaigneous rocks is stable, bears high strength parameters (Krynine and Judd, 1969) and has a gentle topography with occassional residual hills forming an attractive landscape.

The metasedimentary area, on the other hand, is topographically rather rugged and the section underlain by schists have a low shearing capacity (UNIFE, 1979) and thus may create unstable conditions for founding structures.

Therefore, the area underlain by undifferentiated metamorphic igneous rocks (Fig. 5) is recommended as the development site for the proposed Capital City. The infrastructure of Abuja City is now nearly completed. The Governmental offices are expected to move into the residents soon.

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PREHISTORICAL RESEARCHES AROUND ANTALYA BAY AND ITS GEOMORPHOLOGICAL CONTEXT

Angela Minzoni DEROCHE* and Nuri GÜLDALI**

ABSTRACT.— Up to the present, comparatively very little work on prehistory has been undertaken in Turkey. In the neighbouring countries of Turkey, such as Syrian, Lebanon and Israelis, much work has been done and large number of remains of obsidian and flint artifacts, made for the purposes of cutting and breaking into pieces, of the Neolithic period have been found. Assuming that the similar objects may also be found in Turkey, the formulation of a comprehensive research in this country has been undertaken. As a start, research work has been carried out at the mediterranean coasts of Turkey ; this work will also, be extended to the southeastern Anatolia in due course. In the mediterranean region of Turkey, the research work has particularly conducted in three main areas. The first area that has been chosen as a case study lies between the River Alara and the River Karpuz and situated near to the coast. The next area that has been studied includes the terraces and alluvial fan area of the River Burhan, northwest of Antalya. In these areas very little evidence of flint artifacts have been found. The third area that has been selected for a study comprises the surroundings of Kocapınar village, which is situated to the north of town of Elmalı. It is in this area that large quantities of sileks nodules and flint tools have been found. The preliminary examination of this findings imply that the tools are of Late or Middle Neolithic period.