THE STRUCTURAL AND STRATIGRAPHIC POSITION OF DAĞKÜPLÜ (NORTH OF ESKİŞEHİR) OPHIOLITHIC COMPLEX AND PETROGRAPHY OF CUMULATES

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ABSTRACT.— Dağküplü ophiolithic complex which is situated around Yakakayı-Gündüzler villages to the north of Eskişehir indicates southerly overturned structural position. Mesozoic ophiolithic rocks, from bottom to top present a sequence of ophiolithic melange, mafic and ultramafic cumulates and tectonites. Cumulate sequence begins with gabbroes at the bottom passes to the dunite interlayered pyroxenites towards to the top.

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GEOLOGYAND STRATIGRAPHYOF THE CAINOZQIC SEDIMENTARYROCKS IN THE KALE-KURBALIK AREA, DENİZLİ, SOUTHWESTERN TURKEY

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ABSTRACT.— In this work, the geological and stratigraphical characteristics of the Cainozoic sedimentary rocks exposed at the Kale-Kurbalık (SW Denizli) area have been investigated. In the study area the Palaeozoic and Mesozoic rocks form the basement rocks and are overlain by the Oligocene to the Quaternary aged sedimentary rocks. The Tertiary rocks have been considered in two groups, namely the Akçay group and the Muğla group. The Oligocene to the Burdigalian aged Akçay group is represented by the Karadere, Mortuma, Yenidere, Künar and Kale formations. During this time interval, mostly the terrestrial fine and coarse clastics were deposited, but in the last stage of this time shallow marine carbonates were sedimented on the some parts of the region. There is an angular unconformity between the Mortuma and the Yenidere formations of the Akçay group; others are conformable to each other. The Akçay group is 4100 meter thick. The Upper Astarasian (Middle Miocene) to Pliocene aged Muğla group which overlies the Akçay group uncorformably is formed of the Sekköy, Yatağan and Milet formations. During this time interval, the lacustrine siltstones and carbonates and the terrestrial coarse elastics were deposited. The formations of the Muğla group are conformable and gradational to each other. The Muğla group is 550 meter thick. The Quaternary deposits have been considered in two units, namely "Lower" and "Upper" Quaternary sediments. Only the Mortuma formation of all units of the study area is gently folded, but others have low degree dips. Tectonic activity has played a big important role in the forming of the various sedimentary basins which have been generated since the beginning of the Oligocene up to the present.

INTRODUCTION

This study was carried out on the quadrangles M 21- c4d3 and N 21-a2,b1 of the Kale-Kurbalık area (SW Denizli). This area is of considerable importance to evaluate the Cainozoic geology and stratigraphy of the SW Anatolia (Fig. 1). The fieldwork was undertaken between 1979 and 1981. In this region, the previous researchers confirmed important geological data, i.e. Altınlı (1955), Dizer (1962), Becker-Platen (1970), Lüttig and Steffens (1976), Becker-Platen et al. (1977), Benda and Meulenkamp (1979), Gökçen (1982) and Hakyemez and Örçen (1982) (Fig.2). The first geological research in the area was carried out by Altınlı (1955). Becker - Platen (1970) contributed much to the understanding of the whole region. Lüttig and Steffens (1976) played a significant role in the interpretation of the palaeogeographic evolution of the region. Hakyemez and Örçen (1982) studied the region in detail. The investigators were concerned mainly with the palaeontological or chronostratigraphical aspects of the study area.

STRATIGRAPHY

PALAEOZOIC AND MESOZOIC

Basement rocks

Since the basement rocks are not the subject of this study, only their lithologies, which are important to define the source areas, were shortly described. The main lithologies are quartzite, marble, metamorphic schist, limestone, radiolarite and ophiolite. According to Altınlı (1955) these rocks were formed during the Palaeozoic and Mesozoic time interval.

CAINOZOIC

The Cainozoic aged rock units, which form the subject of this study, overlie the Palaeozoic and Mesozoic basement rocks. The Cainozoic aged rock units are composed of the Akçay and the Muğla groups, and the Quaternary sediments (Fig. 3).

Akçay group

The Akçay group, mainly composed of continental and partly lagoon and marine elastics and car-

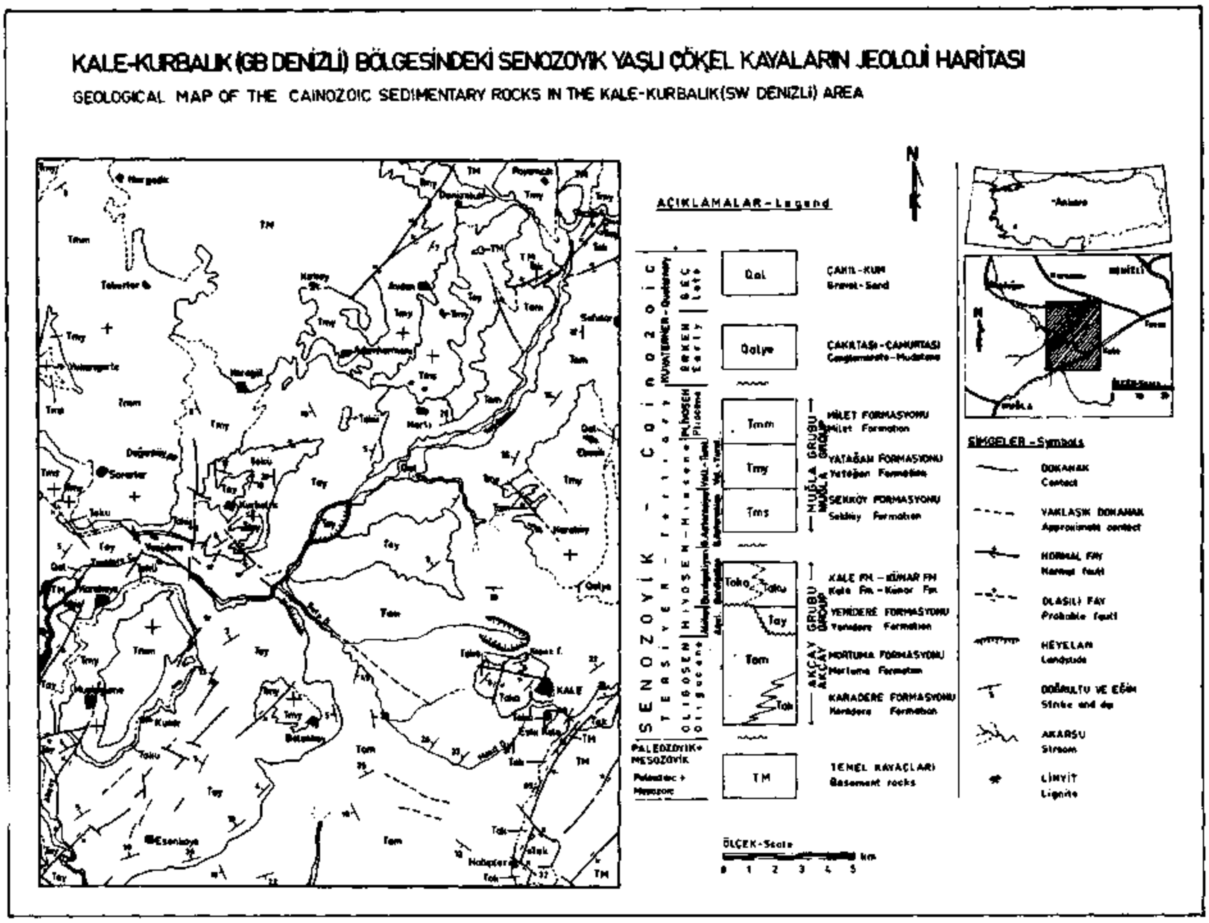

Fig.1— Geological map of the study area.

bonates deposited during the Otigocene to the Burdigalian time interval, consists of the Karadere, Mortuma, Yenidere, Künar and Kale formations. The Akçay group is 4100 m thick. It has a large outcrop in an area between Akçay, Karagöl village, Kırtaş hill and Habipler village, and it crops out in the region between Mugla and Denizli out of the studied area.

Karadere formation.— This formation was named as the Karadere member by Hakyemez and Örçen (1982) formerly. The type section of the formation is located along the Masit creek, 1 km east of the old Kale town. The Karadere formation is dark red and olive green in colour. The conglomerates in the formation were derived from mainly ophiolites and some marbles and limestones. The Karadere formation is formed of five different fades. The first fades is found in the lower part of the formation; the second one appears in the middle and upper parts; the third facies characterizes the middle part; and the last two facies belongs to the upper part:

1. Very poorly sorted conglomerate and mudstone: This facies is formed of matrix-supported, very poorly sorted, thick to very thick of massive bedded conglomerates and mudstones with similar properties. The bases of beds are erosive or straight and sharp, and the gravels are subangular to subrounded. This fades represents the debris flow deposits (Bull, 1972).

2. Cross— and parallel-bedded conglomerate and sandstone: These are grain-supported, poorly to moderately sorted, planar cross-to parallel-bedded conglomerates and trough cross-bedded pebbly sandstones. Gravels are subrounded to subangular, and the gravel im-

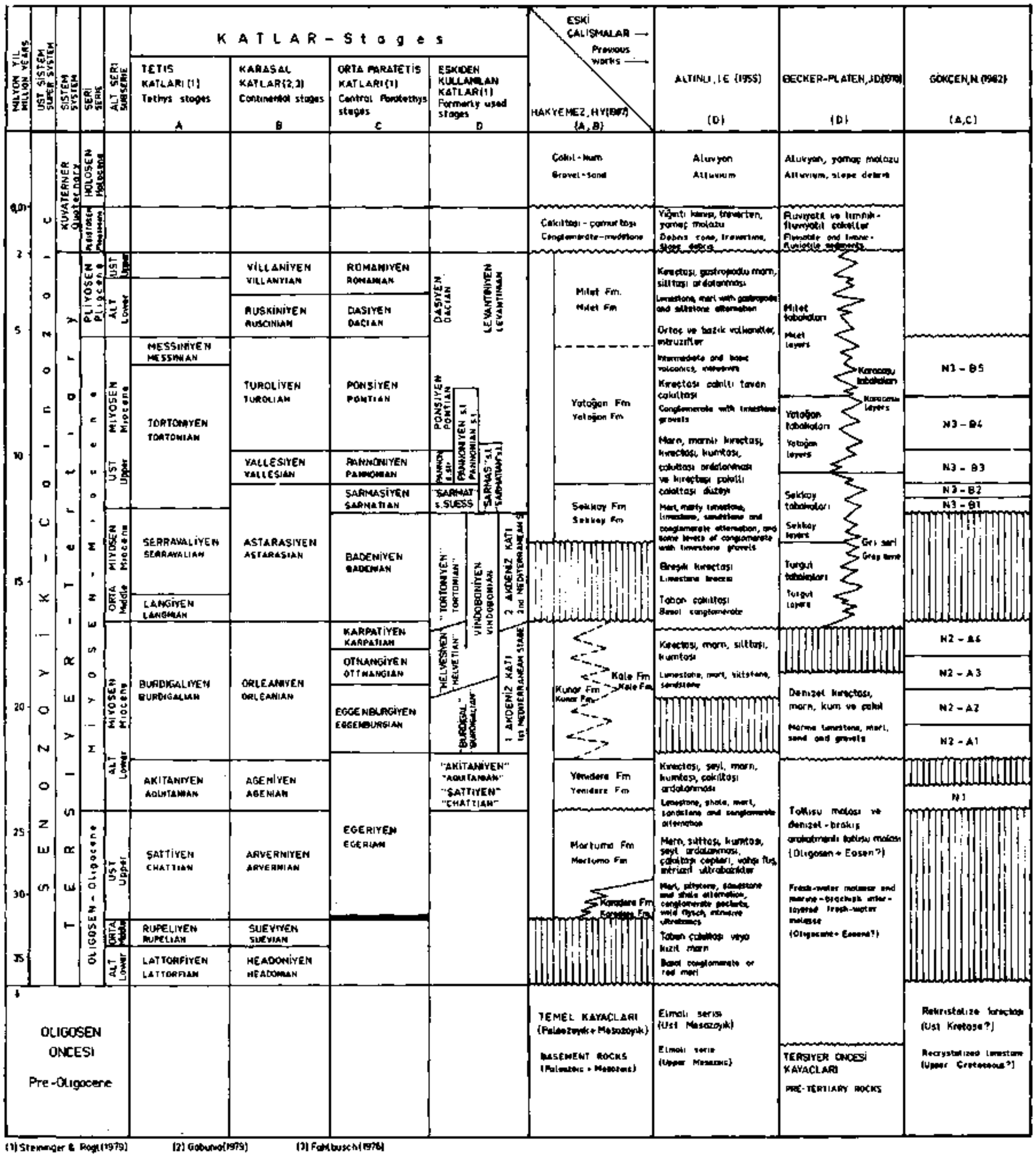

Fig.2— Stratigraphical correlation chart.

brication is common in the parallel beds. The conglomerates grade into sandstones laterally and they alternate vertically. This fades represents the braided channels and bars.

3. Parallel-bedded conglomerate: This facies is made up of grain-supported, moderately to well sorted, parallel and wide lenticular bedded conglomerates. The gravels are subangular to subrounded and the gravel imbrication is absent. This facies has been interpreted as sieve deposits (Hooke, 1967).

4. Parallel-laminated and cross-bedded sandstone : This facies is formed of coarse sandstones with cross-beds or upper flow regime parallel lamina. The cross-bed sets are solitary in general. In any case, the

thin bedded and moderately sorted mudstones overlie the sandstone beds. These are interpreted as sheet flood deposits (Rahn,1967; Bull, 1972).

5. Thin bedded mudstone: The moderately sorted and thin parallel-bedded mudstones form this fades. The facies contains scattered gypsium crystals. They are interpreted as overbank deposits.

The Karadere formation has a fining upwards sequence. This formation covers the basement rocks unconformably; but transitionally grades into the Mortuma formation with which it interfingers laterally. Its thickness is up to 425 m at maximum. There is no fossil in the formation except some destroyed plant fossil traces; so the ages of the formation was defined as Oligocene in relation to that of the Mortuma formation. The depositional environment is considered as a semi-arid and retrograding alluvial fan.

Mortuma formation,— This formation was designated by Hakyemez and Örçen(1982) formerly. The formation has its typical outcrop along the Mortuma creek bcated to the west of the study area. The type section of the formation is seen along the Masit creek. The lowermost part of the formation, which is some 180 m,is composed of mainly yellowish brown and grey coloured, parallel-and cross-bedded conglomerate and sandstone alternation. The properties of this part of the formation is similar to the 2nd facies of the Karadere formation. The rest of the formation, which is about 2200 m thick and yellowish brown, gray and green coloured, is formed of the cyclic units, beginning with conglomerates, continuing upwards with cross to parallel-bedded and laminated sandstones, ending up with thin lignite-bearing siltstones and claystones intercalated with sandstones. Five facies were differentiated in this cyclic alternation :

1. The first facies is formed of 1.5-6.0 m thick cycles. In the lowermost part, on an erosive base, there is a poorly sorted, extra- and ultra- formational lag congiomerate(Laury, 1971) containing a coarse sandstone matrix. This conglomerate is overlain by trough cross-bedded pebbly sandstones with some parallel-laminated levels. Finally, the cycle ends with the cross-laminated silty sandstones and parallel-laminated silty sandstones respectively. This facies represents the meandering-rfver point-bar deposits (Allen, 1964, 1965a, 1965b,1968,1970; Simons et al., 1965).

2. The second facies is formed of 2-4 m thick cross-bedded pebbly sandstones and pebblestones overlying cross-laminated sandstones. This facies grades into the first facies laterally and has been interpreted as intra-channel bars and sand-waves(Harms, 1975; Allen, 1968; Smith, 1970,1971).

3. This facies consists of strongly bioturbated, fine to very fine sandstone and siltstone alternation with plant fossils. The sandstones are characterized by thin layers, cross and parallel-lamination and climbing-ripples whereas the siltstones are very thin layered and parallel-laminated. This facies is interpreted as natural levee and crevasse-splay deposits (Coleman, 1969).

4. This facies is composed of parallel-laminated siltstones and claystones with some parallel-laminated very fine sandstone intercalations. It also contains thin lignite beds, caliche nodules, laminated caliches, thin-walled pelecypods and plant fossils. The bioturbation is strong. This facies was deposited in a flood plain (Allen, 1964,1965a).

5. Strongly bioturbatedsiltstones,claystones and silty very fine sandstones with abundant plant remains form this facies. The facies always overlies a trough cross-bedded sandstone sequence and it grades into first facies laterally. The depositional environment of this facies is assumed to be ox-bow lake(Allen, 19656; Bernard and Major, 1963).

This formation is characterized by a fining trend in terms of grain size vertically as well as laterally in a northeast direction. The Mortuma formation has faulted boundaries with the basement rocks; the lowermost part of the Mortuma formation interfingers with the Karadere formation but the rest of the formation overlies the latter gradationally upwards. In the study area, some thin walled pelecypods and some fragments of plant fossils together with pollens were found only in the lithologies of siltstones and claystones of the formation; but in the lagoonal part of the forma-

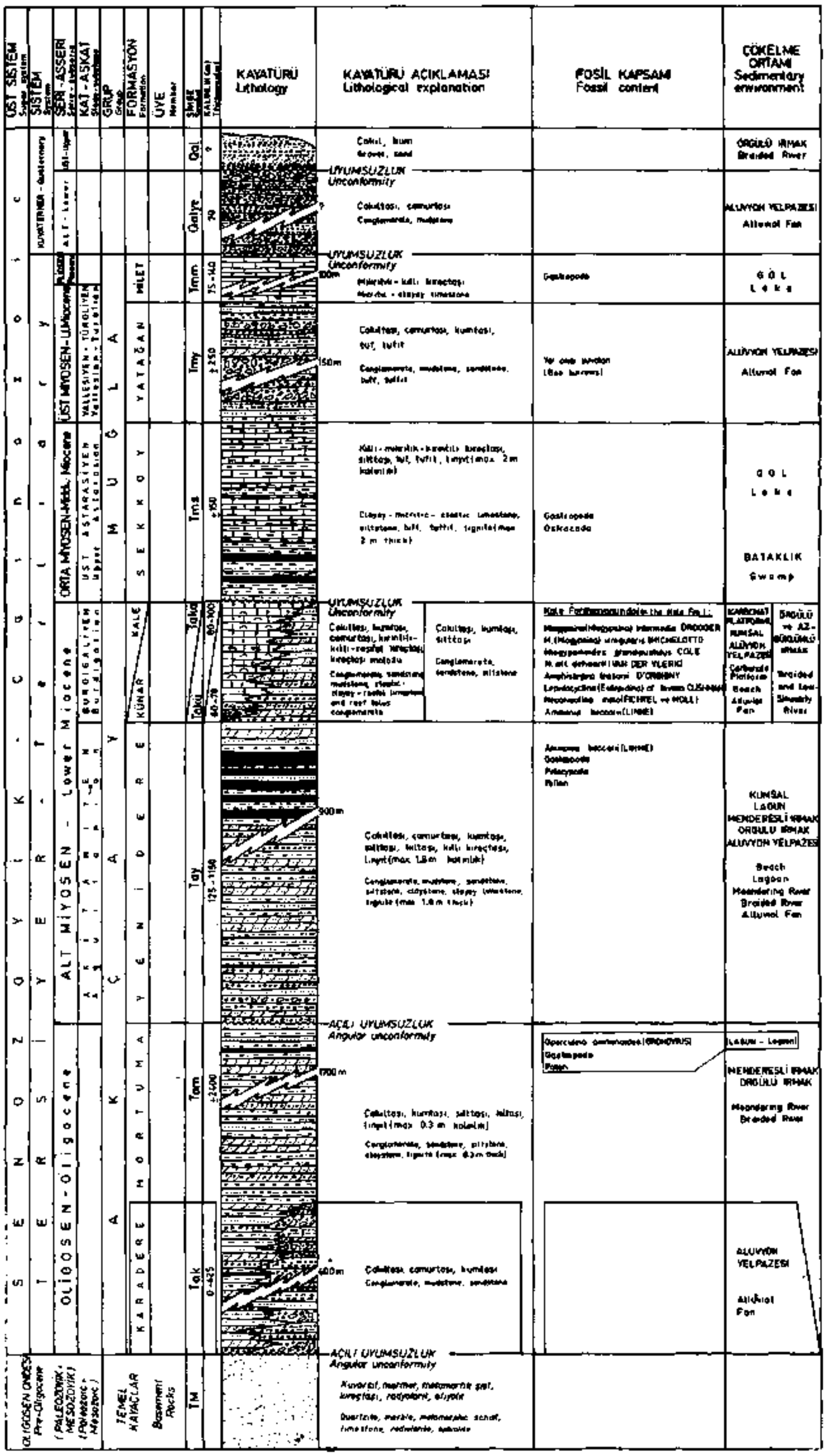

Fig.3— Generalized stratigraphical section of the study area.

tion out of the study area, which lies 4 km south of Çukurköy located to the east of Tavas town, the following fossils were found: Foraminifers such as *Operculina ammonoides* and Miliolidae, gastropods such as *Ampullina (Ampullinopsis)* cf. *bourcarti, Barbatia (Barbatia) albanica, Tympanotonus* sp., *Potamides* sp. together with some ostracods and fish teeth. According to this fossil content, the age of the formation is the Upper Oligocene. The lower 180 m part of the formation was deposited in a braided-river environment, whereas the main part was sedimented in a meandering-river.

Yenidere formation.— Hakyemez and Örçen (1982) differentiated and named this formation formerly. The typical outcrop of the Yenidere formation is found along the Yenidere creek between the Narlı and Yenidere villages. The type section of the formation is exposed along an intermittent stream running from the south of the Kuzlualan hill to the Yenidere creek. The Yenidere formation consists of five levels which have different characteristics:

1. The first level contains three facies. These are (a) very poorly sorted conglomerates and mudstones, (b) cross- and parallel-bedded conglomerates and (c) parallel-laminated and cross-bedded sandstones. The first facies is formed of very poorly sorted, matrix-supported and massive or thick to very thick parallel-bedded conglomerates with subangular gravels, and very poorly sorted and parallel-bedded mudstones with small boulders. These are debris flow deposits. The second facies is composed of poorly sorted, grain-supported, and planar cross-bedded and parallel-bedded conglomerates with subrounded to subangular and occasionally imbricated gravels. The bases of the beds are erosive. These are braided-stream deposits. The last facies consists of parallel-laminated and trough cross-bedded sandstones with some thin mudstone interlayers. These are sheet flood deposits.

2. The second level is similar to the 2nd facies of the first level. But the second level is dominantly formed of cross- and parallel-bedded conglomerate whereas the 1st facies forms the first level mainly. The sediments of the second level are braided-river deposits.

3. The third level contains,five facies: (a) The first facjes is made of 2.5-8.0 m thick cycles which show a rhytmic alternation. In the lowermost part of the cycles, poorly sorted lag deposits with coarse sandstone matrix overlie the underlying cycle with an erosive base. The lag deposits consist of intraclasts in the upper part of this third level whereas they are mainly composed of extraclasts in the lower part. Intraclasts are sandstone, siltstone and claystone gravels, silisified plant fragments and lignite clasts. The trough cross-bedded sandstones with some parallel-laminated medium to coarse sandstones overlie this unit. The uppermost part of the cycle is formed of trough cross-laminated, silty, fine to very fine sandstones, sandstones with climbing-ripple laminae, and parallel-laminated and bioturbated fine to very fine sandstones. This facies represents the meandering-river point bar deposits, (b) The second facies is composed of 1-4 m thick planar cross-bedded pebbly sandstones. It grades into 1st facies laterally. These are intra-channel sand-wave deposits, (c) This facies is rare in the sequence of the formation. It is formed of trough cross-bedded, parallel-laminated and fine to very fine grained silty sandstones with some climbing-ripples, and parallel-laminated siltstones. These are natural levee and crevasse-splay deposits, (d) The fourth facies is made up of thin, clayey limestone interlayered, parallel-laminated siltstones and claystones. It contains some lignite beds ranging from a few mm up to 1.8 m, caliche nodules, plant remains and some fresh-water gastropod fossils. The bioturbation is common. The depositional environment is a flood plain, (e) This facies is characterized by strong bioturbation and is composed of silty, fine sandstones and siltstones. It is interpreted as ox-bow lake deposits.

4. This level composed of thin elastics containing lagoon fossils is formed of four facies: (a) The first facies is fine sandstone interlayered, bioturbated siltstones and claystones with gastropods, pelecypods, rare foraminifers *(Ammonia beccarii),* fish teeth and plant fossils. The fossils are lagoon forms and the sandstones are storm sand layers (Hayes, 1967). This facies characterizes the bottom of a lagoon, (b) The wave-rippled

sandstones made up of 30-80 cm thick layers with thin heavy mineral lamina form the second fades and overlie the first facies with an alternating gradation. This facies was deposited in the subaqueous part of a lagoon beach above the wave base, (c) This facies contains the units composed of flaser-bedded siltstone - wave and current ripple cross-bedded fine to medium sandstone alternation, and coarse sandstone-siltstone alternation. The coarse sandstones are characterized by primary current lineation. The facies was deposited in the subaqueous part of a lagoon beach during and after the storm waves flooding into the lagoon, (d) Fine to medium sandstones made up of parallel-beds or cross-beds dipping 10° at maximum form this facies. The facies contains macrofossil shells and shell fragments. The thickness of the set of cross-beds is 1 m at maximum. These are interpreted as the deposits of the upper shoreface of a lagoon (Elliott, 1978).

5. This level is composed of sandstones with marine macrofossils and contains two facies: (a) This facies is formed of planar cross-bedded fine to medium sandstones which contains some marine macrofossils and heavy mineral lamina. The dips of the cross-beds change from 2° up to 8°. These deposits characterize marine upper shoreface. (b) The second faeies is represented by trough cross-bedded and high-angled planar cross-bedded sandstones. The bedding was destroyed by roots in some places. The erosion surfaces between the beds contain iron oxide. These are assumed as sand dune deposits (McKee,1957).

The Yenidere formation overlies the Mortuma formation unconformably and the overlying Künar formation covers the formation conformably. The thickness of the unit is 1150 m at maximum . The Yenidere formation contains the following fossils above the 830th meter: *Ammonia beccarii* and gastropods such as *Melanopsis* cf. *bonelli bonelli, Terebralia* cf. *bidentata bidentata, T.* cf. *subcorrugata, Tympanotonus (Tympanotonus) margaritaceous* cf. var. *tabana, Turritella (Turritella)* cf. *gradata* and *Galeodes lainei* .together with some pelecypods like *Gryphea (Crassbstrea) gryphoides crassissima, Ostrea* cf. *fimbriata* and *Anadara (Anadara)* aff. *turonica* and pollens. According to the gastropods and pelecypods, the age of the formation is the Aquitanian. The described levels in the formation characterize different depositional environments these are from bottom to upwards alluvial fan, braided-river, meandering-river, lagoon and beach.

Künar formation.— This formation was named as the Karakaya member by Hakyemez and Örçen (1982) formerly. The Karakaya member was designated in a later work as the Künar formation by Hakyemez (1987). The type section of the formation is in Yenidere village. The Künar formation is yellowish grey coloured and is mainly formed of three facies:

1. Poorly sorted conglomerate: These deposits are matrix-supported, poorly sorted, parallel-bedded conglomerates. The bases of the beds are sharp or erosive. Gravels were derived from quartzite, marble and radiolarites. Boulders are common in this facies. These are interpreted as debris flow deposits.

2. Cross- and parallel-bedded conglomerate and sandstone: The trough and planar cross-bedded conglomerates and sandstones form this facies. The conglomerates are medium to well sorted and occasionally imbricated. Gravels are derived from quartzite, marble and radiolarites. The conglomerate and sandstone beds are gradational laterally, and they alternate vertically. Sandstones are dominant in the facies. These deposites characterize the braided and low-sinusoity river channels and bars (Collinson, 1978).

3. Cross -laminated sandstone and parallel-laminated siltstone: The trough cross-laminated fine sandstones form the lower part of this facies. The upper part of the facies consists of parallel-laminated siltstones. The facies is a few tens of cm in thickness. It was deposited in the abondoned braided-river channels and/or on the bars before the lower flow stages(Coleman, 1969).

The Künar formation overlies the Yenidere formation conformably and it has been stated that the Künar formation is the continental equivalent of the Kale formation. Although the Sekköy formation unconformably rests on the Künar formation according to chronostratigraphical setting, it is the Yatağan formation that overlies the Künar formation unconformably in the study area as the first unit on the Sekköy formation. The thickness of the Künar formation changes from 50 to 70 meters. The formation contains no fossil. Its age is defined as Burdigalian correlating with the Kale formation. Although the first facies represents debris flows at small alluvial fans; this formation was formed in a fluvial environment, apparently characterized by an anostomosed (braided and low-sinusoity) river type.

Kale formation.— This formation was formerly named by Altınlı (1955) as the "Kale Marine Helvetian". Later, Hakyemez and Örçen (1982) designated the unit as the "Kale formation". The formation has typical outcrops in Kale town and its type section is along NW slope of the Kepez hill. The Kale formation is yellowish white in colour. In the lower part of the formation, sometimes there is a few meter thick conglomerate-sandstone alternation, but in general, the lower part is composed of a thin sandstone-conglomerate alternation overlying the 30-70 cm thick transgressive lag deposits. The rest of the formation is formed of limestones. Various facies are defined in these three levels:

1. The first level composed of conglomerate-sandstone alternation contains two facies: (a) Thin and parallel-bedded conglomerates in granule size, and trough cross-bedded pebbly sandstones in the form of solitary sets form the first facies. It contains medium and parallel-bedded and reddish brown coloured mudstones occasionally. This facies is interpreted as sheet flood deposits, (b) Matrix-supported, medium to poorly sorted, planar cross-bedded and parallel-bedded conglomerates, and trough cross-bedded and parallel-bedded sandstones form the second facies. The bases of beds are erosive and gravels are subrounded. Both lithologies are gradational laterally and alternate vertically. These are braided-stream deposits.

2. Two facies were defined in this level: (a) The first facies is transgressive lag deposits forming a 30-70 cm thick level in the lowermost part of the formation which is composed of macrofossiliferous sandstones including subangular to angular boulders and gravels. (b)The parallel-bedded and trough or low-angled planar cross-bedded , macro and microfbssiliferous fine to medium sandstones, and low-angled planar cross-bedded conglomerates form the second facies. These are beach deposits.

3. Five facies are defined in the limestone sequence: (a) The first facies is formed of thin to medium bedded clayey limestones with ahermatypic corals, abundant ostracods, bentbnic and rare planktonic foraminifers, gastropods and pelecypods. This facies is very limited and was probably deposited in a nearshore part of a carbonate platform close to a river mouth effected by cold currents(Stanley, 1979) where terrestrial silt and clay influx was introduced into the carbonate sedimentation, (b) The parallel- and cross-bedded bioclastic limestones with abundant Miliolidae, Neoalveolina and some Miogypsina form this facies. It also contains gastropods, pelecypods, hermatypic corals, red algae and echinid spicules. These limestones are packstones in general, but the cross-bedded limestones are grainstones. Cement is microsparite. This facies was deposited in a nearshore area partly protected from wave effect and is similar to the 8th facies of Wilson (1975)'s standart facies belts. The cross-bedded limestones imply that the longshore currents were also effective during deposition, (c) The third facies is thin to thick parallel-bedded clastic limestones with abundant Miogypsina, Operculina and Amphistegina. These limestones are packstones and wackestones, and contains red algae, bryozoa, hermatypic corals, gastropods, pelecypods and annelids. The cement is microsparite. This facies was deposited in a relatively deep open platform (7th facies belt of Wilson, 1975). (d) The fifth facies is characterized by thick to very thick parallel-bedded reef limestones mainly composed of hermatypic corals and algae. It also contains some bryozoa, binding foraminifers (Acervulinidae) and echinid spicules. According to Dunham (1962)'s classification, these are boundstones and characterize the patch reefs in a semirestricted platform . (e) This facies is characterized by limestones

which are completely formed of fragments of reefs bounded by a packstone matrix. It contains echinids, corals, algae, pelecypods and less foraminifers, and could be correlated with 4th facies belt of Wilson (1975).

In the study area, the Kale formation overlies the Mortuma formation with an angular unconformity. But in terms of its chronostratigraphical setting, it is considered to be overlying the Yenidere formation; and it has been accepted as the shallow marine equivalent of the Künar formation. According to chronostratigraphical setting, the Sekköy formation unconformably rests on the Kale formation, however there is not any other unit on the Kale formation in the field. The thickness of the Kale formation is about 100 m. The Kale formation contains abundant foraminifers, ostracods, corals, algae, gastropods and pelecypods. Most importantimes of these fossils follow as :

Foraminifers: *Neoalveoline melo, Miogypsina (Miogypsina) irregularis, M.(Miogypsina) intermedia, Miogypsinoides* aff. *dehaarti, M. grandipustulus, Amphistegina* cf. *lessonn, Lepidocyclina (Eulepidina)* cf. *favosa, Ammonia beccarii.*

Gastropods : *Terebralia bidentata bideritata, Galeodes lainei, Tympanotonus (Tympanotonus) margaritaceous, Ficus (Fulgoroficus) conditus.*

Pelecypods: *Ostrea (Ostrea) lamellosa boblayei, Ostrea edulis* var *adriatica.*

Corals: *Paleoplesiastraea desmoulinsi, Defrancia irregularis, Favia melitae, Aquitanastraeaguettardi, Tarbellastraea* cf. *eggerburgensis, Acropora* cf. *exarata, Porites* cf. *collegnian, Acanthocyathus versicostatus, A. verrucosus, A. transilvanicus, Balanophyttia varians, B. concinna.*

According to these fossils and especially occurrence of *M. (Miogypsina) intermedia* species, the age of the formation is defined as Burdigalian.

The unfossiliferous conglomerates and sandstones in the lower part of the succession were deposited on a lower alluvial fan whereas the fossiliferous sandstones and conglomerates were sedimented on a beach transgressively. The limestone facies are the product of a shallow carbonate platform.

Muğla group

The Muğla group, composed of lacustrine and continental sediments deposited during Late Astarasian-Pliocene, consists of the Sekköy, Yatağan and Milet formations. The Muğla group is 550 m thick. It crops out around Tekerler, Sararlar, Muslugüme, Belenköy, Adamharmanı, Avdan, Payamcık and Karaköy villages, and it covers extensive areas in the region between Muğla and Denizli.

Sekköy formation.— This formation was firstly named by Becker-Platen (1970) as "Sekköy layers" and later it was designated by Atalay (1980) as "Sekköy member". Finally, Hakyemez and Örçen (1982) redefined this unit as the "Sekköy formation". The typical outcrop of the formation is located on aslope in the north of Narlı village, the type section is exposed along an intermittent stream in the north of the same village. Three facies are defined in the Sekköy formation. From bottom to upwards these follow as :

1. Lignite-bearing siltstones: This facies forms the lowermost 50 m part of the formation. It is composed of gray coloured, thin to medium parallel-bedded and laminated siltstones containing 1 to 200 cm thick lignite interbeds. Plant remains and gastropod shells form abundance zones in some places, and the bioturbation is high.

2. Clastic limestone: These are white coloured, thin to medium parallel-bedded and occasionally cross-laminated limestones. This facies is 1-2 m thick and overlies the previous facies.

3. Clayey limestone-micritic limestone-calcareous siltstone alternation: This facies forms the uppermost and thickest part of the formation; and it is white coloured, thin to medium parallel-bedded and tuff and tuffit interlayered. It contains abundant organic material.

Although the Sekköy formation rests on the Kale and Künar formations according to chronostratigraphical setting, it unconformably overlies the Yenidere formation which is the first underlying unit of these formations in the study area. On the other hand, it conformably overlies the Middle Astarasian aged meandering river deposits out of the study area, which was named as "Turgut formation" by Hakyemez and Örçen (1982). The Yatağan formation covers the Sekköy formation conformably; its thickness is 150 m at maximum. Some ostracods such as *Candona* cf. *neglecta* and Cytheridae, gastropods such as *Pseudoam nicola sp.,Valvata* sp. and Planorbidae,and pollens were found in the formation. Since these fossils are not characteristic to determine the age of this formation, the age determination has been based on the vertebrate fossil data of Atalay (1980) and radiometric measurements of Becker-Platen et al. (1977) (11.1 + 0.2 my and 13.2 + 0.35 my). The writer agrees with Atalay (1980) that the age of this formation is Upper Astarasian (Latest Middle Miocene). The first facies which forms the lower part of the formation was sedimented in a swamp; the second one characterizes a lacustrine beach; and the last facies was deposited in lacustrine environment. The source of the tuff and tuffit layers was probably located in the Bodrum region (Ercan et al., 1981).

Yatağan formation.— Becker-Platen (1970) designated the formation as Yatağan layers" formerly. Later, Atalay (1980) used the name of the "Yatağan formation" in his work. But Atalay's Yatağan formation also includes the Milet formation (in this paper) and he named the unit the "Bayır member", which is equivalent of the Yatağan formation in this study. Finally, Hakyemez and Örçen (1982) differentiated and designated the unit as the Yatağan formation. The Yatağan formation has its typical outcrop around Adamharmanı village; the type section is along an intermittent stream running to the south about 1.2 km east of the same village. The Yatağan formation is reddish brown in colour. Five facies are defined in the formation:

1. Very poorly sorted conglomerate and mudstone : The matrix-supported, very poorly sorted, massive or thick to very thick parallel-bedded conglomerates and mudstones with similar properties form this facies. The bases of beds are erosive or straight and sharp, and the gravels are angular to subrounded. These are interpreted as debris flow deposits.

2. Cross and parallel bedded conglomerate and sandstone: This facies is formed of grain-supported, poorly to medium sorted, planar cross-bedded and parallel-bedded sandstones. Gravels are subrounded and occasionally imbricated. The bases of beds are generally erosive. Both lithologies are gradational laterally and alternate vertically. This facies represents the braided-stream channels and bars.

3. Parallel-bedded, well sorted conglomerate: These are matrrix-supported,weli sorted and parallel-bedded conglomerates. Gravels are subangular and the gravel imbrication is absent. This facies represents sieve deposits.

4. Parallel laminated and cross-bedded sandstone: Thick parallel-laminated and trough cross-bedded coarse sandstones with some conglomerate sheets in granule size form this facies. In a single level, conglomerate sheets grade from parallel-laminated sandstones to cross-bedded sandstones respectively in the direction of downstream.

5. Thin bedded mudstone: This facies is formed of moderately sorted and thin bedded mudstones with dessication cracks, bee burrows, scattered gypsium crystals, caliche nodules and laminated caliches. It generally alternates with 4th facies, but sometimes is found as interbeds in the pebbly parts of second facies. This facies is interpreted as overbank deposits. In the vertical section, coarse and fine grained sequences alternates frequently. The formation also contains some tuff and tuffit interlayers. It is underlain by the Sekköy formation and overlain by the Milet formation respectively with a conformable and gradational contact. Its thickness is 250 m at maximum. In the moderately sorted and parallel-laminated mudstones, bee burrows and an Hipparion tooth were found. So the age definition has correlatively been based on the vertebrate fossil data of Atalay (1980) and radiometric age determination (9.25 + 0.2 my and 10.2 + 0.15 my) of Becker-Platen et al. (1977); and it is stated that the formation was deposited during the Vallecian to Turolian (Late Miocene) time interval. The depositional environment of the Yatağan formation resembles an arid to semi-arid

alluvial fan complex; the fans are considered to characterize gradational and retrogradational periods during the deposition time. The source of the tuff and tuffit layers was probably located in the Bodrum region and on Kos Island (Ercan et al., 1981).

Milet formation.— This formation was firstly named by Becker-Platen (1970) as "Milet layers". Later, Atalay (1980) renamed it as "Bozarmut member". Finally, Hakyemez and Örçen (1982) designated the same unit as the "Milet formation". The type section of the formation is around Yukarıgörle village. The Milet formation is mainly formed of white coloured micritic limestones and it contains some thin clayey limestone interlayers. At some locations, there is a lignite-bearing, 1 to 2 meters thick siltstone level in the lowermost part of the formation. The Milet formation overlies the Yatağan formation conformably and gradationally. The Early Quaternary sediments overlie the formation. Its thickness is 140 m at maximum. Since it was only some crystallized lacustrine gastropods that were found in the formation, the age of the formation is based on the vertebrate fossil data of Atalay (1980) and stratigraphical relationships. Therefore it is accepted that the formation was sedimented in the Turolian (Very Late Miocene) to the Pliocene time interval. The Milet formation characterizes a lacustrine environment.

QUATERNARY DEPOSITS

Early Quaternary deposits.— These deposits, located on the western part of the study area, are formed of reddish brown coloured and poorly sorted conglomerates and mudstones. The thickness of the Early Quaternary deposits is up to 700 meters. They were deposited in an alluvial fan environment.

Late Quaternary deposits.— These sediments are gravels and sands of the Akçay and the Yenidere creek mainly. They characterize a braided-river environment.

GEOLOGICAL EVOLUTION OF THE BASIN

The sedimentation of the units which are the subject of this study began in the Oligocene. Lüttig and Steffens(1976) has discussed that a SW-NE trending continental basin was formed in the SW Anatolia after the regression had started towards the end of Late Eocene. It was in this basin, that the Karadere and the Mortuma formations were deposited. The sediments of a NE running braided-river system of this basin which later evolved into a meandering-river gradationally passes into lagoonal deposits around Çukurköy and marine units (Dizer, 1962) to the NE of Denizli finally. Alluvial fan sediments deposited at the margins of the river basin are represented by the Karadere formation.

At the end of the Oligocene, a southeastward tilting probably resulting from an uplift of the region at the north or northwest led to formation of gently folding of the Oligocene sediments. The sedimentation during the Aquitanian started with the development of alluvial fans at the northern part of the study area. On this southward inclined continental area, a braided-river and a meandering-river has developed respectively. At the same time, the Tethyan sea started to transgress to the northwards. This event firstly caused to the formation of the lagoonal deposits and consequently to the deposition of the beach sediments over the lagoonal sediments. Shelf sediments crop out to the south of the study area (Poisson, 1977).

Because of the tectonic uplift of the continental area at the end of the Aquitanian, the continental coarse elastics of the Künar formation were deposited on the exposed former beach area from the beginning of the Burdigalian;whereas on the southern part of the study area, a new transgression started and so the marine carbonates of the Kale formation were deposited.

From the end of the Burdigalian up to the Late Astarasian, there was no deposition in the study area. However in the neighbouring districts, the first deposition after the Burdigalian was started by a meandering-river at the beginning of the Middle Astarasian (Hakyemez and Örçen, 1982). The non-depositional stage between the Burdigalian and the Middle Astarasian, which is the equivalent of the Langhian, is the stage of the movement of the Lician nappes (Poisson, 1977). During the Late Astarasian the coal stams formed in the swamps in the centre of the closed basins, and later, the basin changed into a lake environment. As a result of the increasing tectonic movements, the arid type alluvial fans developed along the margins into the Sekköy lake basin which had already dried up at the end of the Late Astarasian, and the alluvial fan deposition continued during the Late Miocene. The alluvial fans of the Yatağan formation extensively developed all around the SW Anatolia (Becker-Platen, 1970, Hakyemez and Örçen, 1982). The tectonic activity ended at the end of the Late Miocene. .In wetter climatic conditions probably resulting from the Pliocene marine transgression on the south of the study area, the basin changed into a lake again and lacustrine carbonate sedimentation continued till the end of the Pliocene.

The existence of the Early Quaternary aged alluvial fan deposits in the study area shows that the tectonic movements became effective in the Quaternary. The young grabens which were formed by NE—SW trending extensions (Dumont et al., 1979) give further support to this idea.

CONCLUSIONS

In this study, the geology and the stratigraphy of the Cainozoic sedimentary rocks cropped out in four quadrangles were investigated, and the facies characteristics of the formations were defined with the aim of interpreting sedimentary environments.

The geological evolution of the region is also explained primarily based on the environmental development together with the geological and the stratigraphical data.

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STRATIGRAPHY OF THE PRE-JURASSIC BLOCKY SEDIMENTARY ROCKS TO THE SOUTH OF BURSA, NW TURKEY

Orhan KAYA*; Orhan ÖZKOÇAK and Alvis LISENBEE*****

ABSTRACT.—Two of the major pre-Jurassic units of northwest Turkey consist of blocky sedimentary rocks (Dışkaya formation in this report) and low-grade metamorphic rocks (glaucophanitic) greenschist facies . In the study area, the Dışkaya formation is divisible into laterally continuous stratigraphic units of olistostromes and shale-lithic sandstone sequences. The blocks include variably recrystallized limestones, some with Late Paleozoic faunal elements, marble-like recrystallized limestones, submarine mafic volcanic rocks, quartzo-feldspathic sandstones, gray and red bedded chert. The complex internal structure, which is characterized by large sandstone pseudo-boudins up to several meters across, is the product of soft-sediment deformation. The basal olistostrome unit of the Dışkaya formation rests with a slightly deformed contact on the metamorphic rocks, and contains at its base blocks derived from the immediately underlying metatuff unit. The Dışkaya formation has slope characteristics. It appears to have been deposited on an older structural system comprising primarily low-grade metamorphic rocks, which has also constituted source area. The field data is not directly indicative of an accretionary wedge origin for the pre- Jurassic blocky sedimentary rocks (the Dışkaya formation) which is suggested in all recent tectonic syntheses.

INTRODUCTION

The oldest four major rock units of the southern parts of northwest Anatolia and northern parts of west Anatolia (Fig.l) include:a- ultramafic rocks;b- medium -grade amphibolite-banded gneiss; c- low-grade metamorphic rocks, d- pre-Jurassic sedimentary rocks characteristically containing Late Paleozoic limestone blocks.

In early studies (Erk, 1942; Ketin, 1947; Brinkmann, 1976) the blocky nature of the pre-Jurassic sedimentary rocks was overlooked, and they were considered to be a regularly stratified graywacke-shale-limestone succession. The latter was generally designated the "Permo-Carboniferous graywacke series". This series was believed to lie unconformably on the low-grade metamorphic rocks, although no sound confirmation existed.

Özkoçak (1969) recognized the blocks in the pre-Jurassic sedimentary terrain in the study area, but he classified it as a Late Cretaceous megabreccia. Bingöl (1974) and Bingöl et al. (1975) were the first to recognize the blocky nature of the pre-Jurassic sedimentary rocks throughout their distribution in northwest and west Anatolia (Karakaya formation) and to establish a mainly Early Triassic age. However, they considered the blocky assemblage to be variably metamorphic and defined the Karakaya formation as consisting of metabasic rocks and metagraywackes, containing large blocks of limestone, and basic and ultrabasic rocks (Bingöl, 1978). It is said that associated schists often display the characteristic features of a medium pressure greenschist facies with glaucophane present only locally. Şengör et al. (1980) redefined the Karakaya formation as an ophiolitic melange consisting of "blocks of Permian limestones, various members of now-disrupted ophiolitic suit and blueschists jumbled in an extremely highly sheared meta-pelite matrix".

In all recent works, accepting plate tectonic implications, the ultramafic rocks, low-grade metamorphic

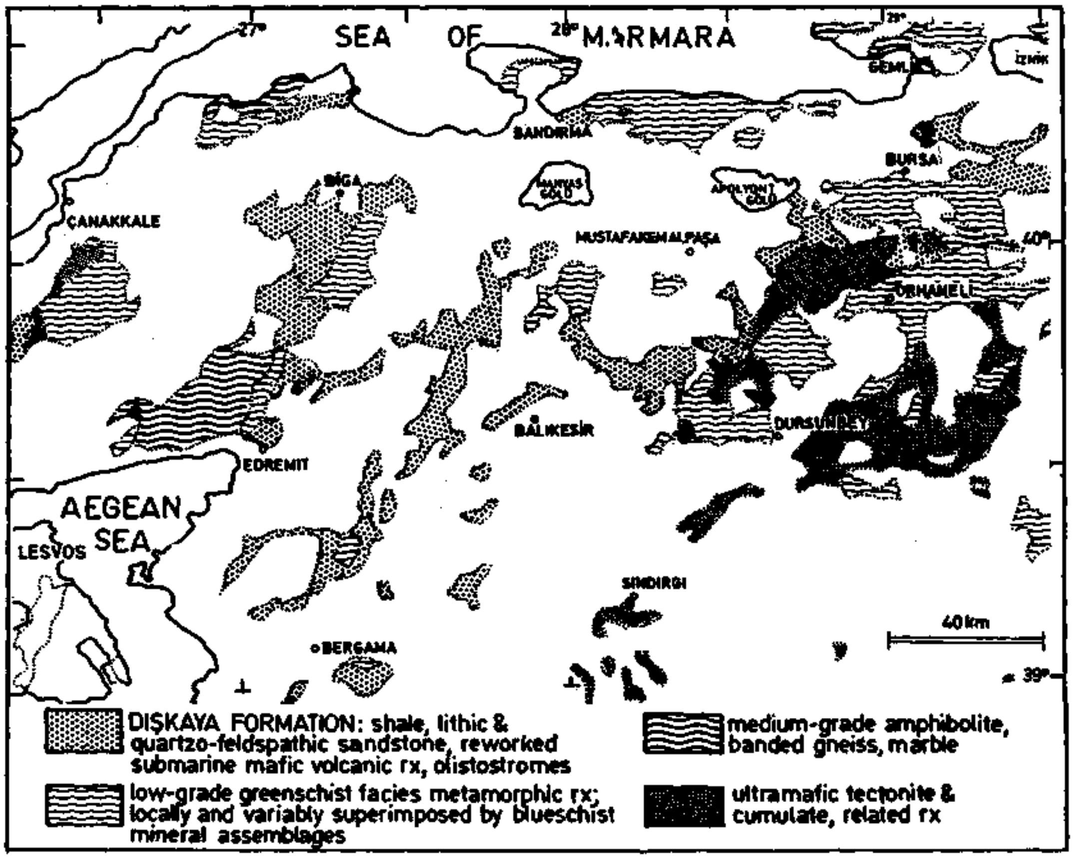

Fig. 1— Distribution of the pre-Jurassic major units of northwestern and western Turkey. Modified after 1:500,000 scale Geologic Map of Turkey, and Bingöl et al. (1975). 1- Study area; 2- Dışkaya dağları.

rocks and blocky sedimentary rocks have been considered as coeval segments of the Late Palaeozoic and/or Triassic oceanic (or semi—oceanic) lithosphere which were incorporated into convergent margin deformation (Bingöl, 1974,1978, 1983; Şengör et al., 1980,1982; Şengör and Yılmaz, 1981; Tekeli, 1981; Üşümezsoy, 1987).

Kaya et al. (1986) have shown that the pre—Jurassic blocky sedimentary rocks were incorporated in stratigraphic successions; that their apparently complex internal structure is the product of the synsedimentary deformation; and that they rest unconformably on the immediately underlying low—grade metamorphic rocks. The name Dışkaya formation was proposed by Kaya et al. (1986) for the blocky sedimentary rocks to replace the Karakaya formation. Broadly, the Dışkaya formation corresponds to the blocky sedimentary part of the "Karakaya formation" suggested by Bingöl (1974).

This report presents further criteria to distinguish and delimit the Dışkaya formation.

UTHOSTRATIGRAPHY

The data on the ultramafic rocks and Tertiary igneous rocks are taken from Ldsenbee (1971,1972). The information about the Jurassic—Early Cretaceous rocks is adopted from Özkoçak (1969).

Ultramafic-mafic layered suit

The ultramafic-mafic layered suit corresponds to Özkoçak's (1969) "Massif ultrabasique d'Orhaneli" and Lisenbee's (1971, 1972) "Ultramafic-gabbro complex

(Orhaneli ultramafic complex) ". It consists of four major rock types which recur in vertical extent; dunite, harzburgite, gabbro and clinopyroxenite. Lherzolite and wehrlite occur in subordinate amounts. Dunite and harzburgite constitute over 90 percent of the layered suit . The mappable units range in thickness from 25 to 3500 meters. They recur in vertical extent several times at all scales, totalling nearly 13,000 m in thickness. The dominantly north-south and subvertical contacts of the major dunite, harzburgite and gabbro units are paralleled by an internal fabric of thin layers of clinopyroxenite and chromitite, and individual grains of elongate enstatite and chromite. Serpentinization has affected much of the primary units. Smaller amounts of jasperoid silica, silicified listwanite and magnesite are also present.

Tankut (1982) recorded the relict cumulative features of stratiform type, and chemical properties of Alpine-type complexes. Özkoçak (1969) suggested that the Orhaneli ultramafic massif was a Late Cretaceous intrusion. Lisenbee (1971, 1972) argued that the massif was emplaced as a solid mass during the Late Cretaceous.

The presence of detrital chromite in the nearby Late Jurassic basal clastic rocks (Özkoçak,1969) may suggest that the ultramafic-mafic layered suit is pre-Late Jurassic in age. The recent recognition of the Late Jurassic unconformity between a Late Jurassic slate unit and ultramafic rocks, which is defined by serpentinite-derived basal conglomerate and pebbly (slaty) mudstone, in Gemlik (Bursa) (Kaya and Kozur, 1987) and Almacıkdağ (Bolu) (Kaya, 1987), may support a pre-Late Jurassic tectonic setting for the ultramafic rocks.

Metatuff unit

This unit consists of bluish to olive-gray, homogeneous, fine to very coarse-grained mafic metatuff with subordinate interlayers of pervasively recrystallized limestone and metalava. The metatuff unit corresponds to Özkoçak's (1969) "Le serie metamorphique superieure" (The upper metamorphic serie). The unit exhibits a well developed foliation, which becomes more pronounced in the weathered-out exposures. The metatuff is apparently basaltic in composition, and consists of chlorite, albitic plagioclase, tremolite, actinolite, epidote, white mica, biotite, relict titaniferous augite, quartz, and glaucophane. Secondary minerals include clinozoisite, grossular-weighted garnet, sphene, apatite, tourmaline, magnetite and calcite. The marble-like recrystallized limestone interlayers are 5 to 100 m in thickness, light gray to reddish gray and fine to medium-grained, and have a gradational contact relationship with the metatuff. The thicker ones are traceable for considerable distances.

The metatuff unit represents the top of the low-grade (glaucophanitic) greenschist facies metamorphic sequence which is widely distributed in the southern parts of northwest Anatolia (Özkoçak, 1969, Lisenbee, 1971). Ketin et al. (1947), v.d. Kaaden (1959) and Brinkmann (1976) have suggested that the contact between the metamorphic rocks and overlying Permo-Carboniferous graywacke series is an unconformity. The only field evidence recorded for the unconformity is the presence of crystalline rock pebbles in the graywacke series, and the so-called rubefaction (weathering) of the metamorphic rocks at the contact, prior to the deposition of the graywacke series (Özkoçak, 1969).

In the study area, the blocky sedimentary unit (Dışkaya formation) contains blocks of metatuff which are identical in all aspects to the immediately underlying metatuff unit. The presence of these blocks at the very base of the Dışkaya formation is the most conclusive evidence so far for an unconformity bounding the metamorphic sequence at the top (Dışkaya formation, lower contact). Thus a pre-Lute Triassic age for the metamorphic sequence is evident.

Dışkaya formation

Stratigraphy .— The Dışkaya formation (Kaya et al., 1986) consists of shale (facies E and G), lithic sandstone—shale (facies C and D, and less commonly B and A), quartzo—feldspathic sandstone ("gully sandstone": Surlyk, 1987) and olistostromes. The latter include a matrix of the above rock types, reworked submarine mafic volcanic rocks and, typically, pebbly mudstone. The partial composite type section is exposed in the Dışkaya area, north of Bursa. In the study D

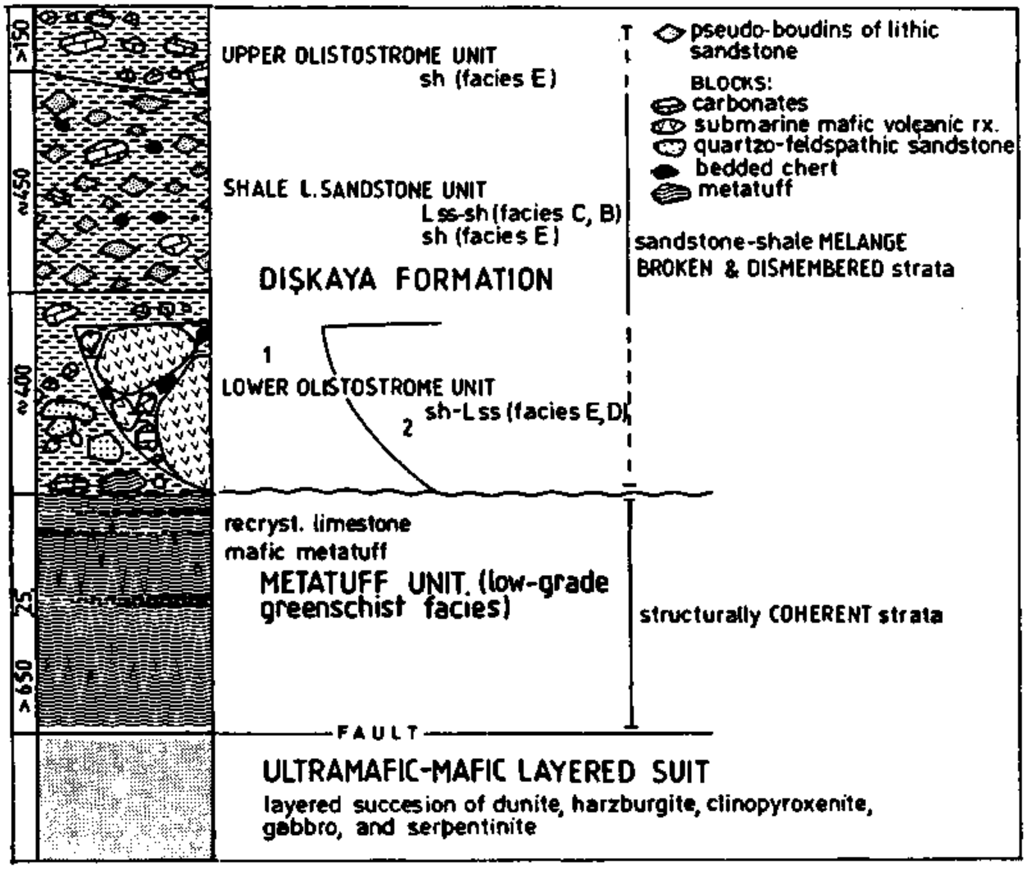

Fig. 2— Composite reference section of the Dışkaya formation exposed in the study area.

area the Dışkaya formation is divisible into three strattgraphic units (Figs. 2 and 3) in ascending order: a- lower olistostrome unit, b- shale-lithic sandstone unit, c- upper olistostrome unit. The units are delimited by arbitrary boundaries on the basis of the traceable distribution of blocks. Because many of blocks are probably undetected or covered by surface deposits, the boundaries may locally be subject to further modification.

Lower olistostrome unit: This unit consists, of isolated and intimately admixed blocks and a matrix of shale, and shale—lithic sandstone sequences primarily of facies C. With respect to the predominating types of blocks, it is divided into two parts: olistostrome 1A and olistostrome 1B.

The olistostrome 1A characteristically contains isolated blocks of gray and red-gray recrystallized limestone quartzo- feldspathic and Iithic sandstone, bedded cherts, submarine mafic volcanic rocks and minor metatuff, all floating in a matrix of shale and shale-sandstone (Appendix 1,l). The olistostrome is well exposed in the surroundings of Göktepe Köyü (Fig.3) where it was first recognized and mapped by Özkoçak (1969), who called it, however, a Late Cretaceous megabreccia.

The olistostrome 1B is dominated by chaotically admixed blocks of submarine mafic volcanic rocks (tuff, lava, reworked volcanic rocks with small limestone blocks, etc.) displaying differences in color, internal stratification and depositional structure. Other blocks are limestone, red and gray bedded chert and sandstone, which are interspersed among the volcanic blocks. The olistostrome is partly mappable in detail on the Belentarla Sırtı where favorable outcrops exist

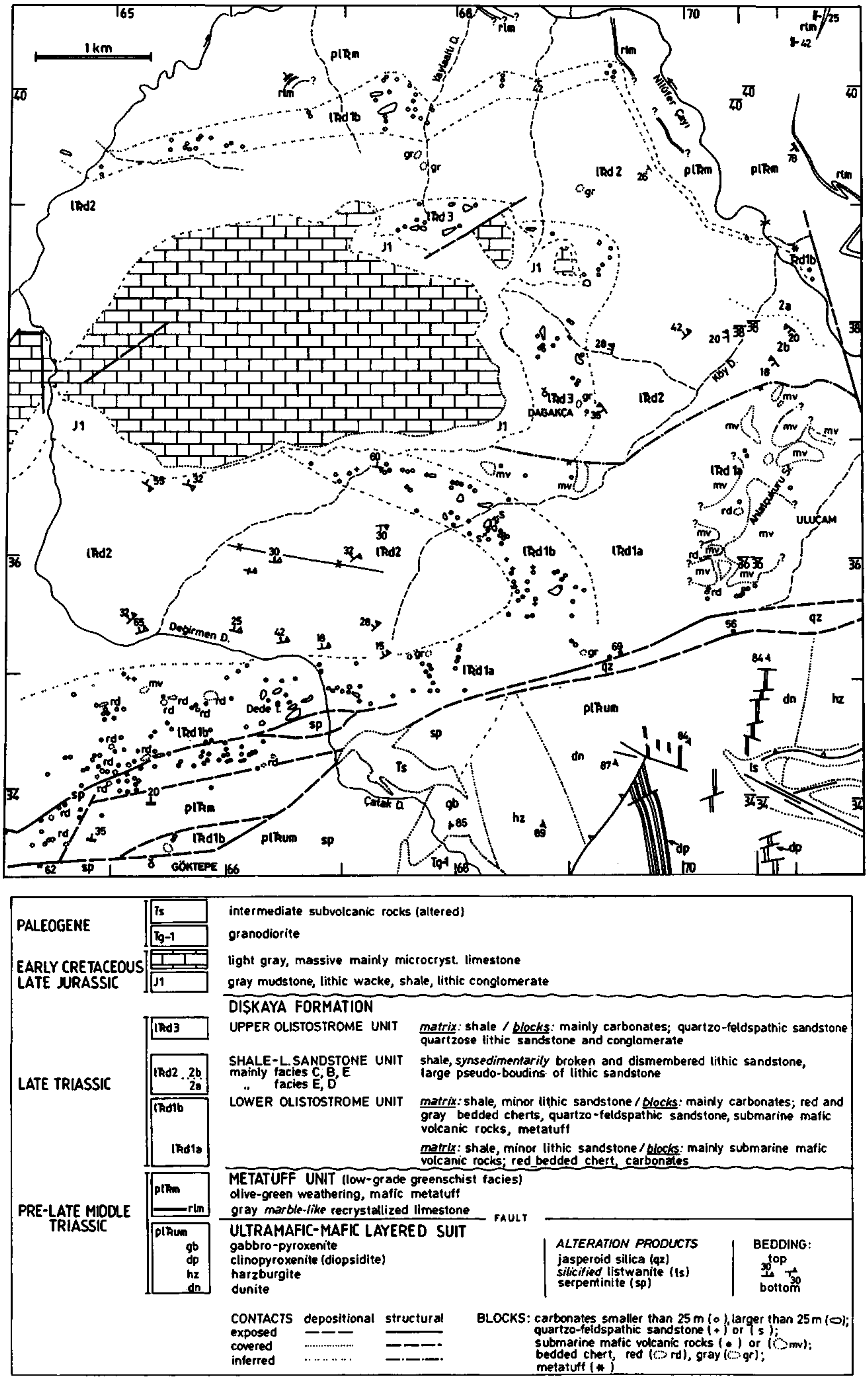

Fig. 3— Geological map of the study area.

(Fig. 3). The matrix consists of synsedimentarily deformed lithic-sandstone sequences primarily of facies D and C, shale, and minor amounts of pebbly shale. The latter contains small blocks of volcanic rocks with clear outlines (Appendix 1,2).

Shale-lithic sandstone unit: This unit consists primarily of shale, and turoiditic lithic sandstone-shale sequences characteristically containing sandstone pseudo-boudins up to 10 m in size. The unit is divisible into a lower shale (Appendix 1,3) and an upper sandstone dominating part (Appendix 1,4). The shale is facies E. The lithic sandstone-shale sequences originally represent facies D,C and B. They are synsedimentarily deformed to the extent of sandstone-shale melange which is distinguished by its large pseudo-boudins of sandstone representing parts of facies B and A (Fig. 4). The unit locally contains gray bedded chert blocks incorporated in an olistostromal interlayer. Lithic conglomerate, a probable cut-and-fill deposit, occurs locally. It is matrix-supported and characterized by perfectly round clasts of lithic sandstones, and minor metaquartzite and altered granitoid (or gneissoid) rocks.

Upper olistostrome unit: This unit consists primarily of isolated small blocks of gray limestone, quartzo—feldspathic sandstone, quartzose lithic sandstone and, conglomerate, gray bedded chert, and minor submarine mafic volcanic rocks, enclosed in a primarily shale matrix (Appendix 1, 5).

Deformation— Abundant broken and dismembered strata, to the extent of sandstone melange, give the Dışkaya formation its complex structural appearance. The deformation features of the sandstones include pull-apart and pinch-and swell structures in particular pseudo-boudins. The latter range in size from a few centimetres to several metres. The smaller ones are lens to lozange-shaped bodies with polygonal outline and sometimes smooth polished surfaces. Larger ones exhibit slab-like to subround blocky shapes and consist of either massive sandstone or interbedded sandstone and shale which have been cut along the bedding at the top and bottom and bounded by curved fault planes at the sides. The long axis orientation of the pseudpboudins, together with the scaly cleavage of the shale matrix, presents a planar fabric. The criteria indicating a soft- sediment origin for the pseudo-boudins in the sandstone-shale melanges include the following (Lash, 1985, Cowan, 1985, Barber et al., 1986).

— The pseudo-boudins exhibit surface irregularities looking like load casts and deformed scour-and-fill structures (Fig.4A). Flame structures and shale penetrations across the bedding are common occurrences (Fig. 4A,B).

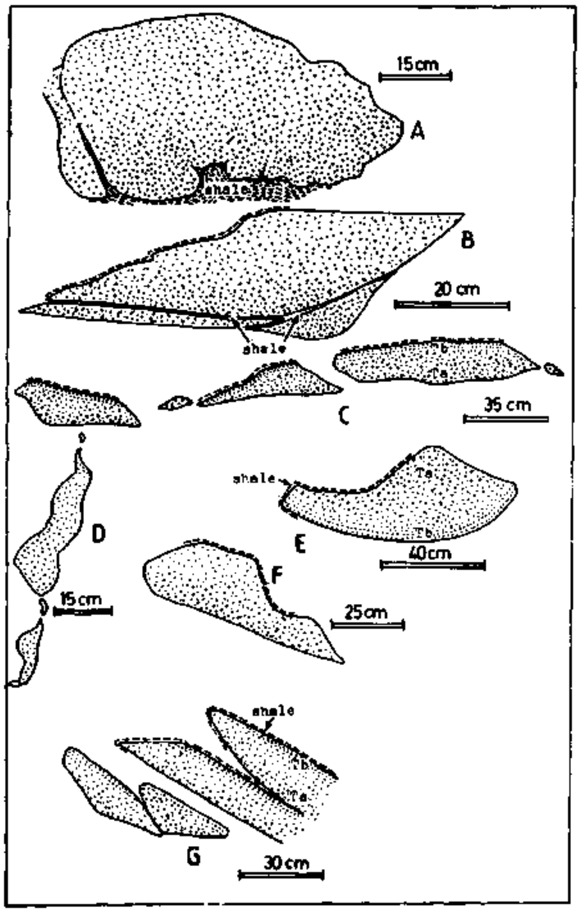

Fig. 4- Representative soft-sediment deformation features in the Dışkaya formation. A- Bulbous protrusions of sand associated with flames of shale, and planar penetration of shale (left side); B- discordant planar shale penetrations; C,D- trains of pseudo-boudins related to synsedimentary stratal disruption; E, F- representative pseudo-boudins; G- synsedimentarily imbricated pseudo-boudins as a part of slide mass. On the exposed surfaces pseudo-boudins exhibit encrustations of thin veneer of shale. A and C are drawn from photographs. Localities for the structures are in turn (Appendix 1,5-12).

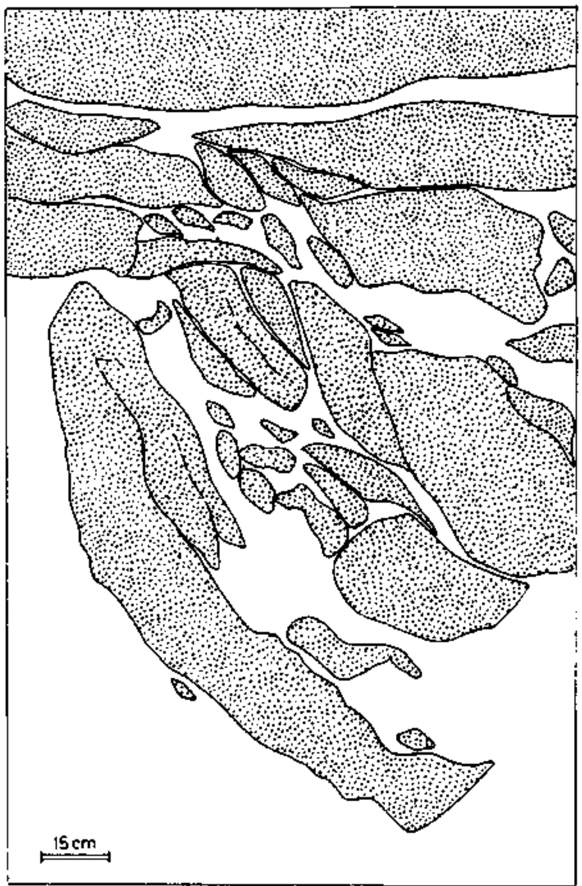

Fig. 5— Slump deposit with accompanying pseudo-boudins of different sizes and orientations. Drawn from photograph (Locality, Appendix 1, 13).

— Trains of pseudo-boudins with the same Bouma divisions and thickness mark the traces of disrupted sandstone strata (Fig. 4C,D).

— Shearing and grain deformation in the peripheral as well as the inner parts of the pseudo-boudins are absent. Sandstone pseudo-boudins are delicately encrusted by shale (Fig. 4A-G).

— Pseudo-boudins, also including the typically phacoid-shaped ones, are the constituents of the slide (Fig. 4G) and slump (Fig.5) masses.

Lower contact.— The lower olistostrome unit (1A) of the Dışkaya formation rests directly on the low-grade greenschist facies metatuff unit, however, the contact is slightly deformed (Appendix 1,14, southern road-cut). In the same place (northern road-cut)the shale matrix encloses blocks of metatuff (Fig.6), up to 8. m across, exactly identical, in their lithic, mineralogic and structural aspects, to the immediately underlying metatuff unit or to that most continuously exposed 250 m westward (Appendix 1,15). The original depositional contact between blocks and matrix rocks is Intact.

The presence of metatuff blocks in the very base of the Dışkaya formation is the first piece of conclusive evidence for the unconformity bounding the so-called Permo-Carboniferous graywacke series.

The polymictic conglomerates recorded by Özkoçak (1969) as the basal elastics of the Permo-Carboniferous graywacke series do not in fact show a traceable outcrop connection with the Dışkaya formation. They are of a lower diagenetic grade when compared with those in the Dışkaya formation, and lithologically resemble the Tertiary deposits outside the map area.

Age.— The matrix of the Dışkaya formation is everywhere barren of fossils. The presence of Late Scythian to Early Norian blocks (Dışkaya Dağları) and the clear-cut unconformity with the early Middle Triassic low-grade greenschist facies metamorphic rocks (Bergama) indicates a Late Triassic age for the Dışkaya formation. In the study area, Early Carboniferous *Densosporites* sp. and *Lophotrilites* sp. recorded by Özkoçak (1969) in fact come from the coaly shale interbeds of gray bedded chert blocks (Appendix 1,16). Restudy of the so-called Globotruncana fragment recorded by Özkoçak (1969) in the Late Cretaceous megabreccia (herein, olistostrome 1A), has shown that the fragment is not informative.

Jurassic—Cretaceous rocks

The Jurassic to Cretaceous rocks include two distinct units, a basal clastic unit, and an overlying limestone unit.

The basal clastic unit consists, in a broadly ascending order, of gray lithic conglomerate, lithic sandstone, mudstone and shale. The conglomerate is grain to matrix-supported, thickly bedded to massive, and contains round pebbles moderately sorted in size. The pebbles include lithic sandstone and shale derived from

Fig. 6— Block of metatuff in the lowermost part the Dışkaya formation. It is exactly identical in all respects to the nearby continuously exposed low-grade Metatuff unit. Drawn from photograph. Locality (Appendix 1, 14).

the immediately underlying Dışkaya formation; and gneiss, metaquartzite, mica schist,metatuff,marble, granite and vein quartz. The sandstone, primarily lithic wacke, contains accessory minerals such as detrital micas, chromite, zircon, tourmaline, apatite,hematite, and spinel. The mudstone and shale consist of siliciclastic material including minor detrital white mica, biotite and ohlorite.

The basal clastic unit shows rapid change in thickness, suggesting that it leveled off the pre-Late Jurassic topography before the deposition of the limestone unit.

The limestone is gray, thickly bedded to massive, and primarily microcrystalline. Algal and foraminiferal detritus and ooids as large as 2.5 cm in size seem to be major constituents. Detrital quartz is locally present. The limestone is locally pervasively recrystallized,and dolomitized and silicified.

The clastic and limestone units can be considered, to be of Late Jurassic and Late Jurassic to Early Cretaceous age, respectively, as several early workers have already recorded. The basal clastic unit has only sparse fossils. Molluscan *Pleuromya alduini* and *P.* aff. *tellina* indicate Bathonian-Oxfordian and Callovian-Portlandian ages, respectively. Foraminifera sampled from the different parts of the limestone unit include Marinella *lugeoni, Trocholina* cf. *alpina, Pseudocyclammina* sp., *Cayeuxia* sp., *Valvulinella* sp., *Valvulina* sp., and Verneulinidae, as a whole indicating a Late Jurassic-Early Cretaceous age.

Tertiary rocks

The Paleocene granodioritic complex exposed as several separate plutons, intrudes the ultramafic-mafic layered suit. Dikes are abundant, both in the surrounding ultramafic-mafic rocks and plutons themselves.

The Neogene dacite, rhyolite and andesite intrude the ultramafic-mafic layered suit and the Dışkaya formation, as dikes and small plugs. In places, they are totally altered to quartz and sericite.

GENERAL CONSIDERATIONS

In all recent tectonic syntheses, the pre-Jurassic blocky sedimentary rocks, together with the coeval low-grade metamorphic rocks, have been viewed as being a subduction-related melange. The suggested tectonostratigraphic units include the Karakaya formation or group (Bingöl, 1974,1978,1983); the Paleo-Tethyan ophiolitic melanges or Karakaya orogen(Şengör et al., 1980,1982; Şengör and Yılmaz, 1981; Şengör et al., 1985), the North Anatolian Melange (Tekeli, 1981), the middle Sakarya melange group (Şentürk and Karaköse, 1981), the Carboniferous and Permo-Carboniferous accretionary prisms (Üşümezsoy, 1987), etc. In addition, Şengör and Yılmaz (1981) and Şengör et al. (1982) established the "root zones" of the Karakaya and ?Bursa sutures comprising the Paleo-Tethyan and Hercynian ophiolites, ophiolitic melange and deep-sea sediments which were superimposed in the surroundings of Bursa, apparently including the study area. Okay (1985) considered the Karakaya as a medium to high pressure greenschist facies complex. Bergougnan and Fourquin (1980) suggested an allochthonous assemblage of diabases, spilites, radiolarites, Halobia-limestones, Triassic clastic rocks and some serpentinites on a Hercynian basement, representing the Triassic opening of the Tethys.

Major field data which does not give a direct support to the recent tectonic interpretations are the following:

1— The Dışkaya formation consists of laterally continuous stratigraphic units. The rock categories include primarily shale (facies E) and lithic sandstone-shale (facies B to D) sequences, their synsedimentarily deformed versions (sandstone-shale melange), and olistostromes are atypical of Flores (in. Hsü, 1974) original definition in not having a matrix of debris-flow origin. Submarine volcanic rocks occur as isolated or intimately admixed blocks which are incorporated in olistostromal interlayers. They are supported by epiclastic matrix and reworked volcanic matrix and are associated with blocks of different rock types. Pelagic rocks are present only as blocks.

The lithologic and sedimentary characteristics of the Dışkaya formation are indicative of a slope apron.

2— Throughout the Dışkaya formation top directins are available and are consistent with an open fold system (Fig.3). Where exposed, blocks show sedimentary contacts with the matrix rocks. The apparently complex internal structure (the sandstone-shale melanges) of the Dışkaya formation is the product of synsedimentary deformation. In other words, the sandstone-shaje melange interlayers are related to submarine sliding of thick piles of semi-lithified sediments. There are no critical accretionary deformation features, such as thrusts and folded packets, which should occur in significant numbers, and refolding structures, penetrative cleavage and elongation lineation. Those existing locally can be best explained as a product of post-Triassic tectonics, primarily because of the absence of transition from ductile to brittle deformation, and conformity with the post Triassic structures, etc.

A great variety of tectonic settings, including passive margins, give rise to soft- sediment deformation to the degree of sandstone-shale melange (Jacobi,1984).

3— The low-grade (glaucophanitic) greenschist facies metamorphic sequence (the metatuff unit) is structurally and stratigraphically coherent. The erosional unconformity between the metatuff unit and overlying Dışkaya formation indicates the latter, at least originally, to be structurally autochthonous.

CONCLUSIONS

— Lithologic, sedimentary and structural characteristics of the Dışkaya formation indicate that it was deposited on a slope floored primarily by low-grade metamorphic rocks. The metamorphic sequence also constituted a nearby provenance.

— There is no direct evidence that the Dışkaya formation itself is of an oceanic origin and it is a tectonic melange representing a part of an accretionary prism.

— Before a Triassic plate tectonics reconstruction is attempted a thorough understanding of the geology of the low-grade metamorphic rocks and ultramafic rocks, within the framework of classical field surveying, seems to be necessary.

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APPENDIX-1

(map coordinates)

1. 70.93:38.64 - 70.96:38.60 (H22d4)
2. 70.87:37.20 (H22d4)
3. 70.80.:38.33 - 71.10:37.90 (H22d4)
4. 70.54:38.10 - 69.20:37.34 (H22d4)
5. 67.58:36.81 - 67.31:36.66 (H22c3)
6. 69.21:37.33 (H22c3)
7. 67.90:36.29 (H22c3)
8. 69.37:37.31 (H22c3)
9. 70.87:37.20 (H22d4)
10. 65.29:36.62 (H22c3)
11. 68.06:36.45 (H22c3)
12. 67.22:35.66 (H22c3)
13. 69.81:37.71 (H22c3)
14. 70.93:38.64 (H22d4)
15. 70.60:38.75 (H22d4)
16. 67.64:39.34 (H21c3)

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THE STRATIGRAPHY AND GEOLOGICAL DEVELOPMENT OF THE CARBONATE PLATFORM IN THE POZANTI-KARSANTI-KARAİSALI (EAST TAURUS) AREA

Cavit DEMİRKOL*

ABSTRACT.— The basement of the investigated area is composed of the Karahamzauşağı formation of Paleozoic (Permo-Carboniferous) age. The Mesozoic is represented mainly by calcareous Demirkazık formation (Jurassic-Upper Cretaceous) and pelagic foraminifera bearing Yavça formation (Campanian-Maastrichtian). The Tertiary succession developed on an irregular paleotopographyof the Paleozoic-Mesozoic aged lithostratigraphic units. In the Tertiary succession, lateral and vertical transitions are quite common within short distances the terrestrial Gildirli formation (Oligocene-Lower Miocene) is found at the base of the Tertiary succession. The lacustrine Karsantı formation overlies the Gildirli formation. In the Lower Miocene, the Kaplankaya formation of shallow water-beach elastics and the Karaisali formation of reefal carbonates developed during a progressive transgression of the sea from the south. Fore reef facies of the reefal Karaisalı formation is represented by the Güvenç formation which is composed mainly of deep marine shales and marls.Turbiditic Cingöz formation developed due to the high sediment influx to the investigated area. The compressional forces were dominant during the evolution of the study area through the considered period. The Kızıldağ melange and the Faraşa ophiolite were emplaced on the platform which was stable till Maastrichtian, during which northern and northeastern parts of the platform had uplifted and the sea had retreated towards the south and southwest. Today, tectonic activity still is going on in the form of a normal block faulting.

INTRODUCTION

The study area, located at the Eastern part of the Taurides, is bordered by Pozantı in the northwest, Karsantı in the north, Karaisalı in the southwest and İmamoğlu in the southeast (Fig.l). Paraautochthonous, neoautochthonous and allochthonous rock bodies are present in the region.

The most important structural property of the region, which is outlined by Blumenthal (1952), is the existence of the nappe structure. The most common rock bodies of this structure ranging from the Upper Devonian to the Early Senonian are paraautochthonous sequences in which carbonates are dominant; allochthonous ultramafic-mafic typed ophiolithes and the units of melange character which are less common relatively to the others. Neoautochthonous units were deposited on both the paraautochthonous and allochthonous rock bodies with an angular disconformity while overthrusts were dominant in the middle and western Taurides after Maastrichtian age (Brunn et ah, 1971; Özgül, 1976), in the studied area no tectonic structure of this type was developed. Instead of this tectonic movements were dominant all along the Ecemiş fault, which played an important role in the structural formation of the investigated area and its vicinities after Maastrichtian age. The Ecemiş fault, which crosses the eastern Taurides in the NE—SW direction both tectonically and morphologically, has been interesting for geologist up to now. Arpat and Şaroğlu (1975) stated that this zone is active. The Ecemiş fault is defined as a sinistral strike-slip fault.

In the study area, while Paleozoic and Mesozoic aged units form the basement, the Tertiary age units of the Adana basin are wide spread. The sequence representing the basement of Carboniferous-Upper Permian age generally exhibits a steadiness, in respect of facies features and depositional environment, whereas there is, in this terms, an obvious difference in the units of continental margin which were deposited at the interval of the Upper Jurassic-Cretaceous age.

STRATIGRAPHY

In the studied area, the units of Paleozoic, Mesozoic and Cenozoic age are present.

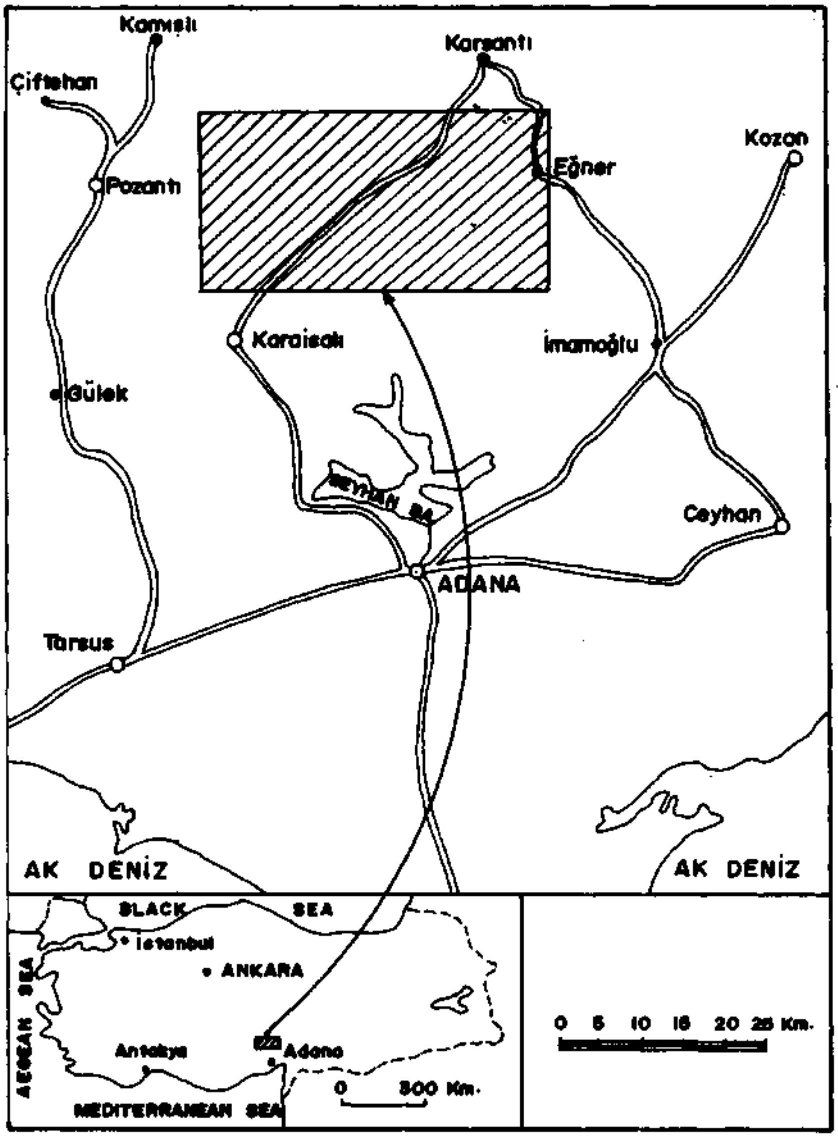

Fig.1 – Location map.

PALEOZOIC

Karahamzauşağı **formation** (PKbk)

The upper part of the unit is composed of limestone of wackestone-packstone type which is yellow-grey, brown-grey, dark-grey, thin-middle thick bedded, clayey, abundant fossiliferous (brachiopoda, foraminifera, echinoids), in some place dolomitized and neomorphized. Lamination and stilolitization are observed at some horizons. This unit is overlain by thin-middle thick bedded limestone including terrigeneous detritus, and is green-brown grey in colour. Limestone of wackestone-grainstone type comprises thin interiayers of mudstone-marl. In this part, some levels of grainstone-wackestone formed by the envoloping of fossils grain by alga are observed. Girvanella in algal shells is common around fusilinoids. The upper part is represented by blue, green, brown quartz sandstone and mudstone-marl interbedded fine-middle thick bedded limestone. At the same time, grey thin-thick bedded limestone and

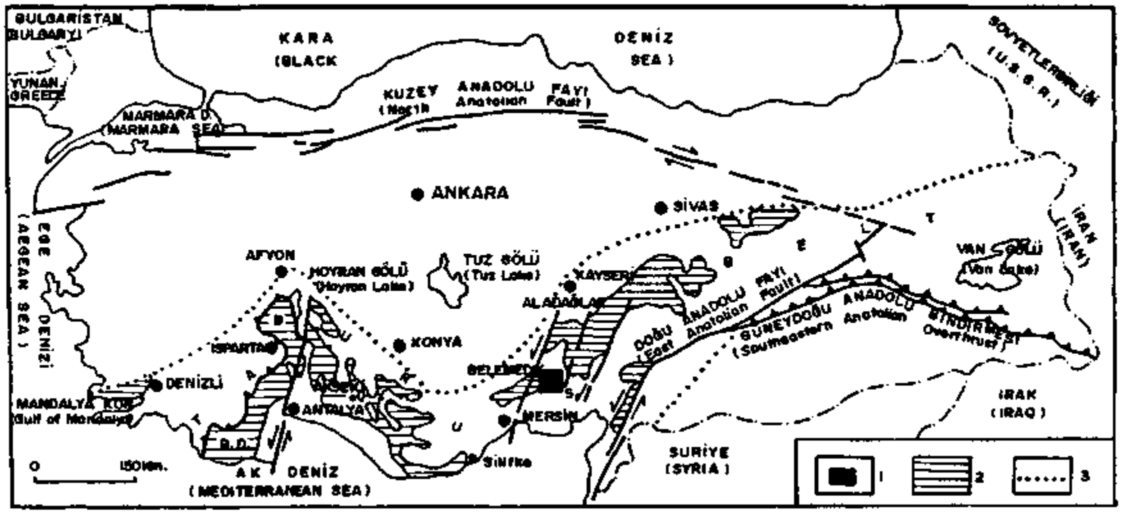

Fig.2 — Map showing the distrubition of para—autochthonous platform sediments of the Taurus belt and main structural lines (Simplified from Adamia and others, 1980).
1- Studied area, 2- Paraautochthonous platform sediments of Taurus belt, 3- Aproximate boundary of Taurus belt. B- Barladağ, BD- Beydağları.

marl levels are also observed. The limestone include echinoids, crinoids, ostracods, algae, bryozoa, gastropods, brachiopods, corals and foraminifera. Ironish marl and silty levels are present. Bioturbation and stilolitization are common. Cross-bedding in quartzite, chert and dolomitization in limestone are present.

The Karahamzauşağı formation: the dominance of colours of green, brown, grey, the beds having different thicknesses, the existence of abundant terrigeneous elastics (quartzite, sandstone, mudstone, marl), the observation of quartz and microfossils widely in carbonates indicate that a shallow environment deposition. The sequence deposited in shelf-shelf lagoon environmental conditions is suggested by the majority of limestone including terrigeneous materials, grainstone interbedded wackestone-packstone together with abundant echinoids, gastropoda, brachiopoda, fragments of foraminifera in some parts, the enveloping grains of oncoid shape by alga and iron material.

In the study area; the lateral changes of the unit with fault-boundered outcrops can not be traced. The basement can also not be observed. The Carboniferous-Permian aged unit is overlain unconformably by younger units.

MESOZOIC

Demirkazık formation (JKd)

In the investigated area the unit composed mainly of limestone, dolomitic limestone, dolomitic and pelagic foraminiferous micritic limestone, is named as the Demirkazık formation (Yetiş, 1978). In the bottom of the unit is grey, hard, angular fractured, locally containing chert, calcite-filling and limonitized, thick-very thick bedded micritic limestone. This limestone is overlied by a dark grey, yellowish grey, thick bedded, hard, locally calcite-filling, scarcely chert-banded, fossiliferous limestone. Neomorfic alteration and dolomitization prevent both the primary matrix and some grains being recognized. Neomorphic spar formations around the pellet grains are observed. The upper unit of the sequence is grey-yellowish, thick bedded, dolomitic and thin-middle crystallized. There are locally yellow-brown mudstones in the dolomite.

The unit, which is unconformably on the Karahamzauşağı formation, is concordant in many places with the Yavça formation of the Upper Cretaceous age in the extending of Köpekdağ (A—5),Çilgurliz Dağı, western part of the studied area (Fig.3). The unit re-

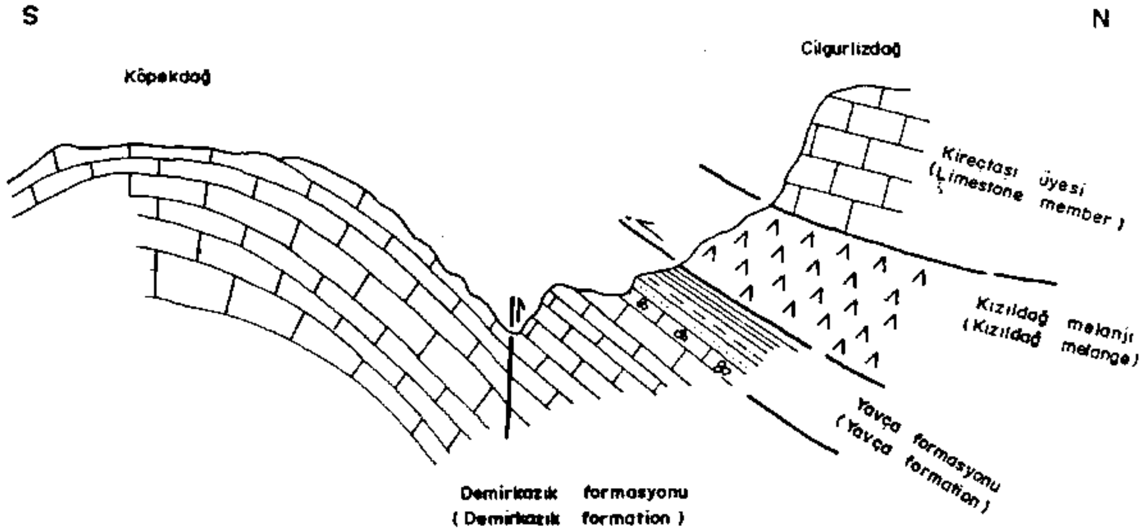

Fig.3 — The shematic cross section of the Köpekdağ—Cilgurliz Dağ area and the contact relations between the Kızıldağ melange and the Yavça formation, the Demirkazık formation.

fleets tidal lakes and tidal plainlagoons in limetted platform facies-belt. The Demirkazık formation was deposited in Jurassic-Upper Cretaceous age (Ünlügenç and Demirkol, 1988).

Yavça formation (Ky)

The unit which starts at the bottom with pink, pelagic foraminiferous micritic limestone and passes up into abundant ophiolite and limestone derived turbiditic sediments is distinguished as the Yavça formation (İlker, 1975; Monod and Erdoğan, 1981). The limestone at the bottom of the sequence comprises reddish-pink coloured, thin-medium bedded and abundant foraminiferous biomicritic limestone, and is overlain by light-dark green, yellow-brown elastics. These are made mainly of an alteration of angular limestone, radiolarite, chert, basalt, quartzite, and serpantinite-derived thick-medium bedded, unsorted, scarce gravel-mudstone and greenish-grey, black, thin bedded, locally dominated shales.

The unit overlying conformably the Demirkazık formation exhibits lateral and vertical facies changes. This is overlain by the Kızıldağ melange and limestone member with a tectonic contact. The deposition of the unit can be considered between Campanian and Upper Maastrichtian age (Unlligenc, and Demirkol, 1987).

Kızıldağ melange (Kk)

In the south of the studied area, the unit which comprises radomly scattered massive and big rock blocks of various sizes in a good relief is named as Kızıldağ melange (Ünlügenç and Demirkol,1988). Spilitic lavas are dominant and flysch, volcano-sedimentary material, volcanic sediments, radiolarite, terrigenous mass movements are common. It includes serpantinite lenses of various size and position, granodiorite and gabro blocks of various kinds, ophiolitic rocks and various sedimentary rocks. The radiolarite blocks are red-brown, often with small folds and interbedded by clay having 30 cm thickness. The serpantinites have a blocky structure. The orientation of long axes of the rock fragments displays the existence of a tectonic control. Additionally, individual rock fragments have their own internal deformational structure.

The Kızıldağ melange was thrusted over the lithostratigraphic units of pre-Upper Maastrichtian age and overlain tectonically by the Faraşa ophiolite where the Tertiary aged sequence was not exposed (Fig.4).

The limestone blocks of various sizes from a few meters to kilometers were seperately mapped (Ünlügenç and Demirkol, 1987). Block-shaped, massive, cherted-limestone, massive limestone, radiolarite, volca-

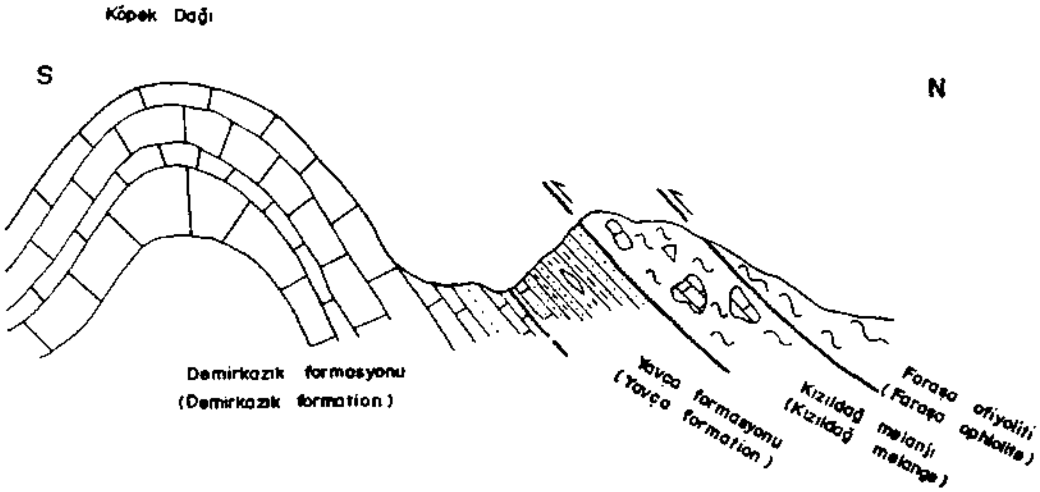

Fig.4 — The shematic cross section of the northern part of the Köpekdağ which shows the contact relations between Farasa ophiolite and the Kızıldağ melange, the Yavça formation.

nic breccia, gabro, diabase, granodiorite which form a morphological highs are present. All rock blocks have a caotic relation with each other, and their long axes extend nearly in E—W direction. The orientation of their long axes indicates the existence of a tectonic control. The blocks also reflect their own internal deformational structure. The limestone blocks are similar to the Mesozoic carbonates. The blocks of limestone exhibit locally dolomitic limestone and dolomitic character.

The Kızıldağ melange is thrusted over the Demirkazık formation of Jurassic - Upper Cretaceous age and the Yavça formation of the Upper Cretaceous (Campanian - Maastrichtian) in the boundaries of the investigated area (Çalapkulu, 1976; Yılmaz, 1984). Thus the Kızıldağ melange may have emplaced in the region during and after the Upper Maastrichtian age (Ünlügenç and Demirkol, 1987).

Faraşa ophiolite (Kf)

The unit has two seperate outcrops in the north-northwest of the region, which are serpantinized ultramafic and mafic rocks assemblage. Harzburgite, dunite, pyroxenite gabro and diabase dykes are common. It is observed that harzburgites have dunite interlayers-and isoclinal folding within well developed foliation and li neation-contacts. It is informed that there is diorite and granite in a little amount in the north of the studied area (Çakır, 1972; Juteau, 1979). The unit is thrusted over the Kızıldağ melange and is overlain with a heterolitic discordance by the Karsantı formation of lake character, composed of clastic carbonates of Tertiary sequence of the Adana basin. The Aladağ ophiolite complex, distinguished by Tekeli et al. in the Aldağ, is similar to the plutonic assemblage and dolerite, diabase dykes and volcanic rocks, defined but unnamed by Anıl et al. (1986) in the Gerdibi-Pozantı area.

Within the investigated area the Kızıldağ melange is thrusted latest over the Demirkazık and Yavça formations of the Upper Cretaceous. This is overlain with a second overthrusting by the Faraşa ophiolite (Fig.4). The Oligocene-Lower Miocene aged Tertiary deposits are unconformably on the ophiolite nappe.

Since the Faraşa ophiolite was placed tectonically on the Yavça formation of Campanian-Upper Maastrichtian age within the investigated area and the units deposited outside it after the emplacement of the ophiolite nappes are of or younger than Maastrichtian age, it should have emplaced during or after the Upper Maastrihtianage.

CENOZOIC

Gildirli formation (Tgi)

Schmidt (1961) described Miocene aged conglomerate, sandstone, shale which deposited in the southwest part of the investigated area. Gildirli formation crops out around Gildirli village (B-6) and consists mainly of conglomerate, pebbly sandstone, siltstone and mudstone which are characteristically pink-pale red to reddish brown. Poorly sorted and imbricated conglomerate is quite thick at the base of the formation and it contains detritics derived from limestone, ophiolite, radiolarite, metamorphic rocks, subrounded and in same places it contains intraformational mud balls. Every cycle of the Gildirli formation generally has erosional bases at the base, cross bedded conglomerate transits to pebbly sandstone to sandstone with grain size decreases. Sandy layers transits to fine sandstone-siltstone alternations and at the top of each cycle reddish mudstone is found: 25-35 cm thick mudstone layers are poorly consolidated, and displays parallel lamination.

The basement of the Senozoic succession has an irregular topography and this trough filled by basal conglomerate of Gildirli formation. The formation normally disconformably overlies Paleozoic and Mesozoic aged rock units. It transits to Kaplankaya formation at the topographic depressions; by contrast at the topographic highs it is overlain by Karaisalı formation. At the base of the formation Jurassic-Upper Cretaceous aged Demirkazık formation are found and Gildirli formation conformably is overlain by Burdigalian-Langhian aged Kaplankaya formation. On the other hand, Lutetian aged rock units are outcrupped around the investigated area (Schmidt, 1961; Abdüsselamoğlu, 1962; İlker,1975; Yetiş.,1978; Yetiş, and Demirkol,1984). During the Lutetian-Paleogene transgression has reached to a maximum in Turkey and its surrounding area. In spite of unfossiliferous character, Gildirli formation has to be deposited in Oligocene-Lower Miocene time period.

Karsantı formation (Tk)

Karsantı formation which outcrops in the northern part of the investigated area, mainly consists of marl and mudstone with some pebbly sandstone and sandstone intercalations at the base. Bedding thickness and grain size of pebbly layers decrease towards the top. Mudstone and marl are dominant at the upper level of the succession and have some sandstone intercalations, coal beds and plant debris. Karsantı formation is generally green-grey, medium to thick bedded, poorly sorted, and it sometimes shows spheroidal alteration surfaces. Slump and sliding structure, current ripples are. common. Karsantı formation reaches to amaximum thickness in the northern part of the investigated area, but its thickness diminishes by the reason of being nearer to the edge of the basin (Schmidt, 1961; Abaci, 1986;Yurtmen,1986).

Karsantı formation unconformably overlies the Kızıldağ melange and Faraşa ophiolite at the base and it form a transition between Miocene deposits of Kaplankaya and Karaisalı formations. Through basinal relationship and vertical-horizontal transition with the terrestrial Gildirli formation, Karsantı formation started to deposite during Oligocene. The upper age limit of formation is still unclear. But some fossil descriptions (Cyprinotus, Eucyris, Loxoconcha, Costa, Heterocypris, Thyrrenocythere, Viviparus, Planorbis) from the upper level of marl-mudstone layer indicate the period of Middle-Upper Miocene (Abacı, 1986;Yurtmen, 1986).

Kaplankaya formation (Tkp)

Kaplankaya formation was differentiated by Yetiş, and Demirkol (1986). It mainly consists of pebbly sandstone, sandstone, pebbly-sandy limestone and siltstone. Sandstone and siltstone alternations are found at the upper level which contains lamellibranchia and gastropoda. Carbonate ratio increase to the top of the formation. Kaplankaya formation unconformably overlies Paleozoic and Mesozoic rock units at the paleotopographic highs. There is a conformably contact relation between Gildirli and Kaplankaya formation. It has lateral and vertical transitions with Karaisalı, Cingöz and Güvenç formations at the top (Fig.5). The outcrops of Kaplankaya formation are scattered and parallel to each other in the southern of Paleozoic and Mesozoic aged rocks in the investigated area. Its thickness mainly related to the paleotopographic situations. Pebbly-sandy limestone of the upper level of this unit is fossiliferous and it contains echinid, corall, coralline algae, small bentic algae etc. Fossil description indicates

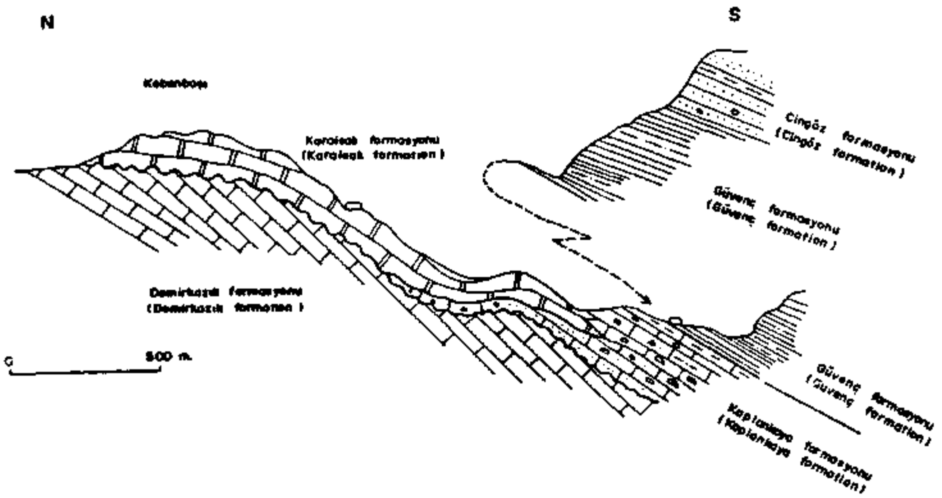

Fig.5 — The shematic cross section of the southwestern part of the Mustafaağalar illustrating the facies relations between Kaplankaya formation. Karaisalı and Güvenç formations.

Lower-Middle Miocene age for Kaplankaya formation (Yetiş, and Demirkol, 1986). With the transitional contact relation of Burdigalian-Langhian aged Karaisalı formation, Kaplankaya might have been deposited during Burdigalian-Langhian (?) time space.

Karaisalı formation (Tka)

Karaisalı formation is mainly composed of reefal carbonates and dolomitic limestone at the investigated area. It generally white to pale grey, medium to thick bedded and it contains coralline algae, echinoderms, bryozoa, corals, mollusca and foraminifera. Karaisalı formation accumulated on the pre-Miocene topographical highs and in the adjacent areas it formed reef and associated deposits. The unit has a lateral and vertical facies relationship with Kaplankaya and Gildirli formations at the base, and Güvenç, Cingöz formations at the top (Fig. 5,6). According to the related fossil descriptions, Karaisalı formation might have been deposited during Burdigalian-Langhian time space (Yetiş and Demirkol, 1986).

Cingöz formation (Tc)

Cingöz formation crops out at the eastern-southeastern side of the investigated area. This unit begins with conglomerate, pebbly sandstone and sandstone on the fine grained sediments of Güvenç formation at the base. This basal section of the Cingöz formation comprises subrounded grains of granule to block sized limestone, ophiolite, chert,.etc.,and forms transition between sandstone and sparse pebbly sandstone to the south. The subrounded clasts are composed mainly of quartz, feldspars, limestone, ophiolite, etc. The sandstone is generally light grey to green, subrounded, coarse-very fine grained, and well sorted. Sandstone layers have occasionally bases with some poorly developed bottom structures such as flutecast, scour and fill, tool marks. At the top, Cingöz formation is covered conformably with shale of the Güvenç formation. The contact zone is identified with sand/shale ratio. Cingöz formation has a turbiditic character. Normally, Cingöz formation has a vertical and lateral transition with transgressive marine succession of the Kaplankaya and Karaisalı formation (Fig. 5,6). There are no fossils were determined for this unit. Burdigalian-Langhian age are applied to the Kaplankaya, Karaisalı and Güvenç formations which are found at the base of the Cingöz formation (Yetiş and Demirkol,1986). On the other hand, Cingöz formation transits to Burdigalian-Serravalian aged Güvenç formation at the top. According to this information, Cingöz formation was deposited in Burdigalian-Langhian period (Yetiş and Demirkol, 1986).

Güvenç formation (Tgü)

Güvenç formation emerges in the southern part of the investigated area. The unit mainly consists of

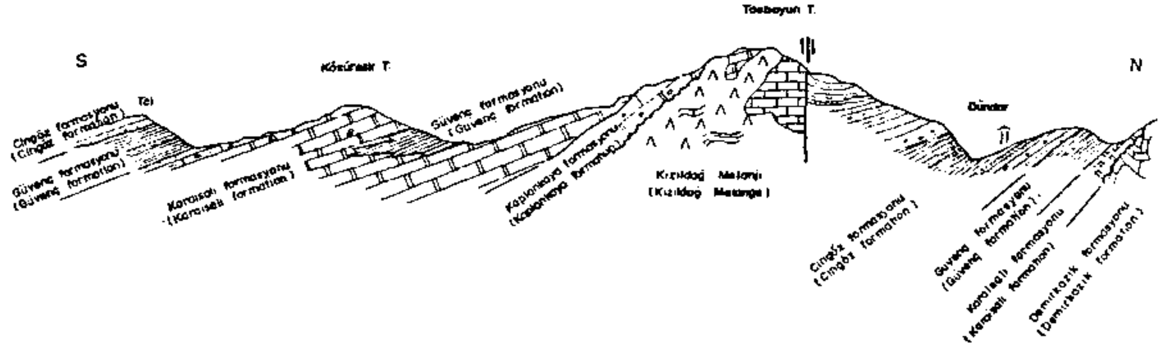

Fig.6 — The shematic cross section of the Dündar region illustrating lateral and vertical facies relations of the marine Miocene units deposited on the Kızıldağ melange.

dark grey, greenish grey, grey colour of sandstone interbedded siltstone and clayey limestone with abundant pelagic microfauna. Güvenç formation starts with fossiliferous sandstone-siltstone and shale alternations. Shale layers contain pelagic and sandy layer with bentic forams. Quartz, feldspar and lithoclast are common in the siltstone and sandstone. The Cingöz and Güvenç formations has lateral and vertical transition with the Karaisalı and Kaplankaya formation at the base (Fig. 5,6). According to the fossil determinations of the Güvenç formation age was Burdigalian-Serravalian time space (Yetiş, and Demirkol, 1986).

QUATERNARY

Terrace (Qt)

The terrace formation was developed along the bed of Körkün, Eğlence Suyu, Seyhan river and has different width. Terrace mainly consists of rough conglomerate with rare pebbly sandstone interbeds. Subrounded grains are made up by ophiolite, radiolarite, quartzite, limestone of different source rocks, cherts, etc.

Alluvium (Qal)

Developed along streams,consisted of poor sorted consolidated gravel,sand and clayey material. The grains depending on the basement are derived from ophiolite, limestone, radiolarite, chert and quartzite.

DEVELOPMENT OF CARBONATE PLATFORM

In the study area and its surroundings while Carboniferous-Upper Permian aged sequence reflecting intracratonic basin conditions shows steadiness in terms of facies features and sedimentary environment, in the Jurassic-Upper Cretaceous aged sequence of continental margin having reefal and back-reef character no obvious differentation in these terms were observed.

In the investigated area, the sediments of Upper Paleozoic age are generally composed of terrigeneous elastics interbedded liniestone. Terrigeneous elastics are, in order of their distribution; quartzite, quartz-sandstone, mudstone-marl and shale. During this period, in the Belemedik sequence, the western part of the investigated area, there is a stratigraphic discontinuity (Tekeli et al., 1981; 1984). In the study area, the Karahamzauşağı formation of Carboniferous-Permian age bears a resemblance to cyclic shelf carbonates. The reason for cyclic deposition should have been eustatic changes of sea level repeated with a constant and regular deposition, which can be attributed to tectonic and glacial reasons (Wilson, 1975). It can be said that the unit was developed in shelves and platforms present in open circulation intracratonic basins. At the end of Permian age, block faulting started (Fig. 7a).

The Paleozoic (Upper Devonian-Permian) aged Belemedik sequence, similar to the Karahamzauşağı formation in terms of lithological features and cronostratigraphy, was deposited in shallow intracratonic basins and reflects eustatic changes of sea-level repeated with a regular deposition (Tekeli et al., 1984).

In the interval of Jurassic-Upper Cretaceous age, the Demirkazık formation, made of limestone and dolomitic limestone, overlies the units of the Upper Pale-

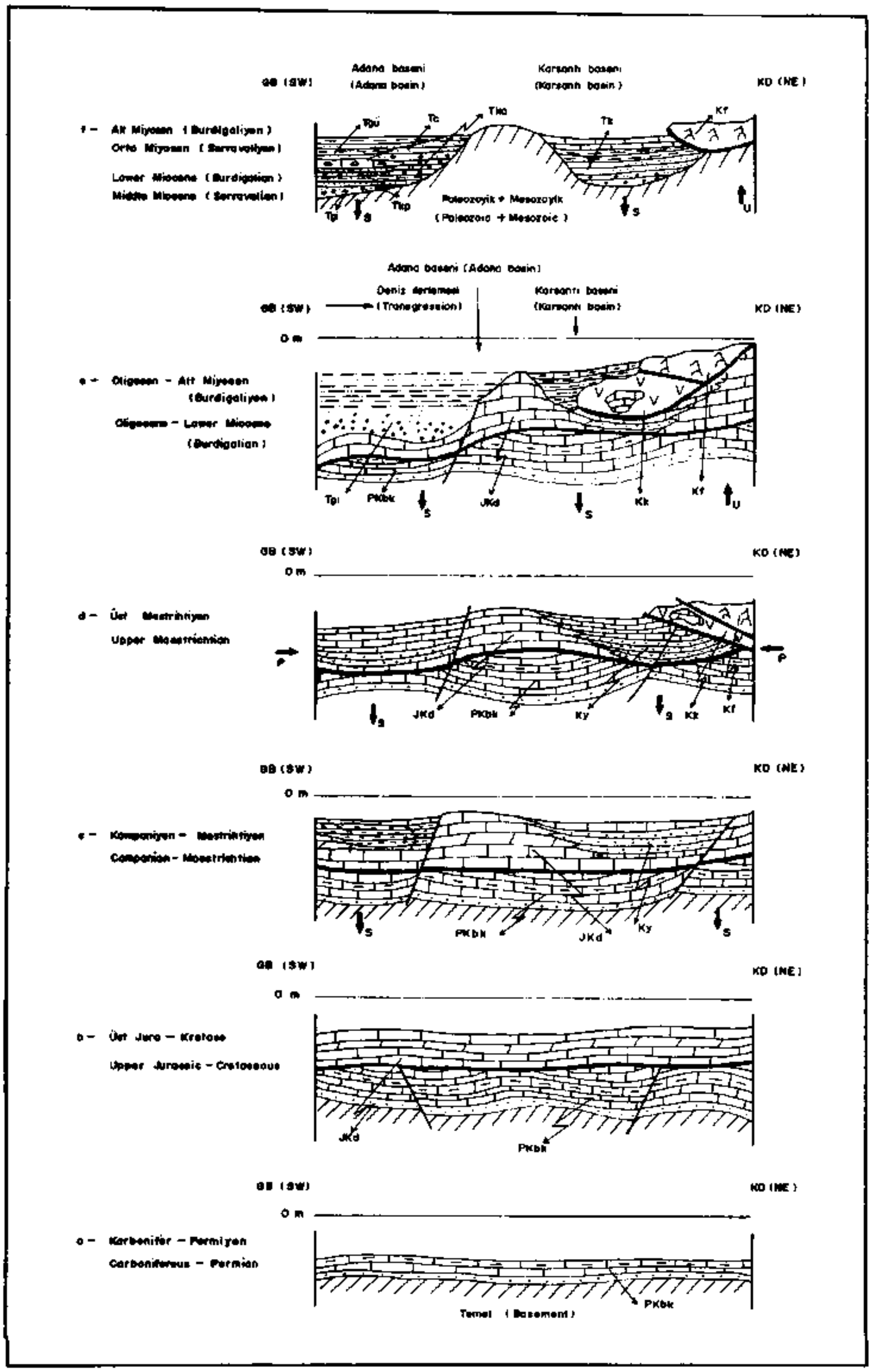

Fig.7 - Serial sections illustrating the evolution of the northern Adana Basin. PKbk- Karahamzauşağı formation; JKd- Demirkazık formation; Ky- Yavça formation; Kk- Kızıldağ melange; Kf- Faraşa ophiolite; Tgi- Gildirli formation; Tkp- Kaplankaya formation; Tk- Karaisalı formation; Tka- Karaisalı formation; Tc- Cingöz formation; Tgü- Güvenç formation; S- Subsidence; P- Compressional stress; U- Uplifting.

ozoic age with an unconformity. It often includes such fossils as Miliolidae, etc. The unit which reflects platform/bank environments and organic deposition in marginal areas of the platform and reaches a great thickness should have developed in lagoons on the platform of limited circulation and on a tidal flat and it should have developed in subtidal and intertidal conditions. This suggests that, during Mesozoic, basinal conditions belonging to shelf were developed on the continental margin (Fig. 7b). The Demirkazık formation starts with foraminiferous micritic limestone, and overlain by the Yavça formation formed from abundant ophiolite and sandstone derived turbiditic sediments in Campanian-Lower Maastrichtian age (Fig. 7c).

The stable position of the region lasted till Upper Cretaceous (Campanian - Maastrichtian) age. The evidence of this stable position is that carbonates pass up into pelagic limestone upward getting deeper.

According to Tekeli et al. (1981), who have interpreted the evolution of the Aladağ in various ages in the northern part of the studied area, the sequence of Triassic-Cretaceous age was deposited in environment of continental margin. It is considered that some relations reflecting stable continental conditions in between Tethys Ocean and Arab-African continent, in the eastern part of Mediterranean.

The stable position of the region was broken by Laramide orogenic phase during and after Maastrichtian age. The Kızıldağ melange was superimposed for the first time by compressional tectonic and the Faraşa ophiolite unit was thrusted over to continental crust (Fig. 7d).

The units in the region were folded and emerged from the sea due to Laramide orogenic phase, then the weathering process has started. A transgression took place in Eocene age. Clastics were deposited on the bottom but some units of limestone in the upper parts, which none of them can be observed in the investigated area (İlker, 1975; Yetiş, 1978). Outside the study area, the Lutetian aged rock, in the northern parts suggest that the Lutetian sea lasted till Late Lutetian (İlker, 1975; Yetiş, 1978; Gedik et al., 1979). In the Late Lutetian the regression resulted from Pironian orogenic phase coused continental environment conditions to redominate in the region.

The Gildirli formation, which overlies unconformably the Mesozoic aged units in the south of the studied area, was deposited at the interval of Oligocene-Miocene (Burdigalian) age as various facies relative to an irregular topography (Yetiş and Demirkol, 1984). In many parts this units is overlain by the Kaplankaya formation, which reflects a character of Shallow sea beach and reefal Karaisalı formation but in the north (the Karsantı basin) by Karsantı formation (Fig.7e).

The Gildirli formation comprises elastics having terrestrial character and cyclic alternation. It is mainly made up conglomerate, sandstone, siltstone and mudstone. The unit starts, on an erosional surface, with an abrupt bottom and channel-filling mudstone and passes up into the alternation of sandstone-siltstone-mudstone by fining upwards of grains. On the Gildirli formation, the Kızıldağ melange and Faraşa ophiolite which are in the bottom of the Karsantı basin is the Karsantı formation composed of sandstone, siltstone, claystone and shale, and the Karsantı formation also comprises coal seams in the lower-middle horizons.

In the Early Miocene (Burdigalian age) during a transgression from south-southwest should be land, on the one hand the elastics of beach-shallow sea character (the Kaplankaya formation) was developed in areas where terrigeneous sediments are dominant, on the other hand the Karaisalı formation of reefal character developed on topographic highs. The Kaplankaya formation overlies terrestrial Gildirli formation and passes up into the Karaisalı formation of which the upper contact is reefal and it is overlain by the Cingöz and Güvenç formations in some areas where environmental conditions are not favourable for the deposition of the Karaisalı formation. The Karaisalı formation to the southward transits into the Güvenç formation of fore-reefal facies which is composed of grey, locally thin sandstone interbedded, deep marine marl and shale (Fig 7f).

Since the transgression of the sea level in the Miocene advenced to the northward until Tortonian, shoreline advanced towards Karaisalı formation and its fore-reefal facies progressed to the northward, e.g. allunits from the Burdigalian to the end of Langhian (?) and early Seravillian occasionally pass over each other to the northward.

During the deposition of the Güvenç formation, elastics were transported via some paleotopographic lows, which coused sedimentation of the Cingöz formation of proximal-distal type. The Cingöz formation is southwestwardly transitional with the Güvenç formation, and shales of the Güvenç formation overlies the Cingöz formation.

CONCLUSIONS

1. In the Upper Paleozoic cyclic sediments derived from an epiric sea basin were deposited in the intracratonic basin. These are generally interbedded with terrigeneous elastics and composed of shallow marine carbonates of a steady platform.

2. In the study area, the Kızıldağ melange was thrusted onto the continental platform for the first time by the Laramide orogenic phase with a continuity of compressional tectonic and then the Faraşa ophiolite slice was emplaced on the continental crust.

3. The Taurid carbonate platform deposits, except some properties, bear important resemblance to those of Africa-Arab platform, specially during Mesozoic and Cenozoic age, which suggests that a marginal sea was deposited between the Anatolian plate and Africa-Arab plate during the Mesozoic age.

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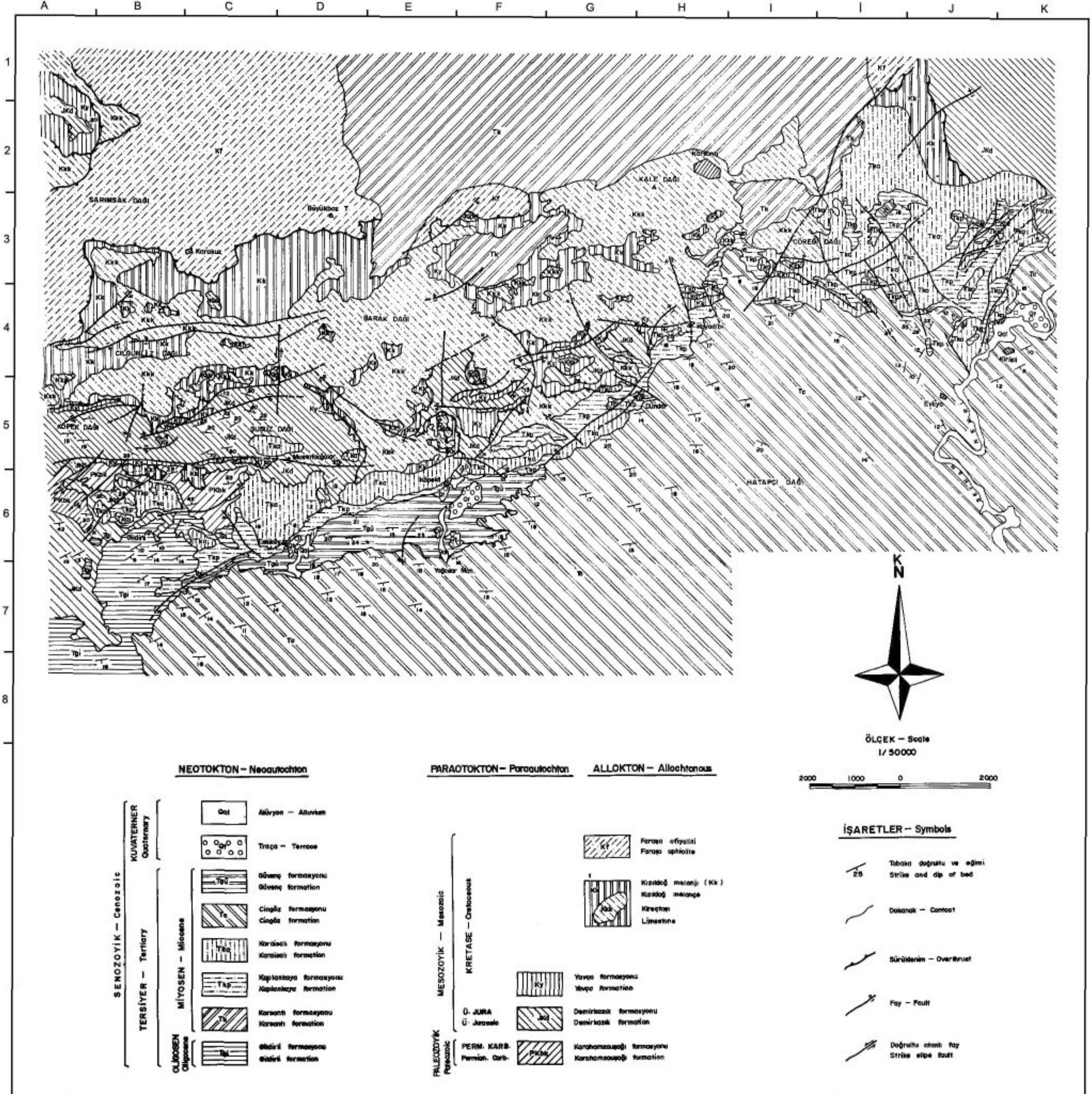

Appendix 1 : Geological map of the Eastern Toros carbonate platform on the northern Adana basin

Mineral Res.Expl. Bull., 109, 37-51, 1989

GEOLOGY OF THE MENDERES MASSIF AND THE LYCIAN NAPPES SOUTH OF DENİZLİ, WESTERN TAURIDES

Aral İ. OKAY*

ABSTRACT.—Mount Honaz region in the Western Taurides is made up of superimposed several tectonic units. In the west, monotonous green metashales called the Honaz shale occur at the base of the tectonic stack. Honaz shale is tectonically overlain by weakly metamorphosed, massif white limestone, thinly bedded cherty limestone and shale which covers the Mendercs massif. Menderes massif is in turn tectonically overlain by the Sandak complex of the Lycian nappes made up of Mesozoic dolomites and limestones and ophiolite mainly of harzburgite lies oven the Sandak complex. All these tectonic units form an eastward overturned major anticline called the Honaz anticline; a pelagic sedimentary sequence of Late Cretaceous—Middle/Late Eocene age, called the Göbecik Tepe complex, occurs tectonically beneath the overturned limb of the Honaz anticline. The Göbecik Tepe complex constitutes the relative autochthonous in the region of the Mount Honaz. Different tectonic units in the Mount Honaz region show effects of Late Cretaceous, Middle Eocene and Late Eocene/Oligocene tectonics. Obduction of the ophiolite over the Sandak complex probably occurred during the Late Cretaceous, while the age of thrusting of the Sandak complex over the Menderes massif is probably Middle Eocene. Emplacement of these tectonic units over the Göbecik Tepe complex and the formation of the Honaz anticline is of Late Eocene/Oligocene age.

INTRODUCTION

The stratigraphy of the Menderes massif and the Lycian nappes and their tectonic relationship in the region between the Bafa lake and Muğla (Fig. 1) are known fairly well through studies carried out during the last twenty years (Graciansky, 1968; Dürr, 1975; Çağlayan and others, 1980; Erakman and others, 1986; Konak and others,1987). In contrast, farther northeast in the Kale-Tavas region little is known on the eastward extension of the Menderes massif and its relation to the allochthonous units. An area where these regional problems can be solved is the mountainous terrain south of Denizli which lacks the post-tectonic Tertiary cover. The Mount Honaz and the neighbouring areas south of Denizli are mapped on a scale of 1:25,000 with the aim of solving the internal structure and relation of the Menderes massif and the allochthonous units (Okay, 1986). This paper describes the geology of the region and summarizes the important results.

A geological definition of the Menderes massif is necessary in order to map its extent and its relation to other units. In this definition the region between the Bafa lake and Muğla, where the geology of the Menderes massif is best known, should be taken as a reference area (Fig. 1). The main features which characterize the Menderes massif in this region are (Graciansky, 1966; Başarır, 1970; Dürr, 1975; Alkanoğlu, 1978; Çağlayan and others, 1980; Okay, 1985; Konak and others, 1987) : (1) The Menderes massif is made up from the base upwards of Precambrian gneisses; Lower Paleozoic micaschists; Permo-Carboniferous metaquartzite, black phyllite and dark recrystallized limestone; Mesozoic, thickly bedded, recrystallized neritic limestones with bauxite horizons; recrystallized pelagic limestone and flysch of Paleocene to Early Eocene in age; (2) The Lycian nappes overlie tectonically the Eocene flysch of the Menderes massif. The emplacement age of the Lycian nappes over the Menderes massif is regarded as Mid-Eocene; (3) A Barrovian-type regional metamorphism of Eocene age has affected the Menderes massif; the metamorphic grade shows a gradual decrease upwards in the sequence (Ashworth and Evirgen, 1984; Okay, 1985; Satır and Friedrichsen, 1986). This regional metamorphism was related to the emplacement of the Lycian nappes over the Menderes massif (Şengör and others, 1984), and has also affected the lower

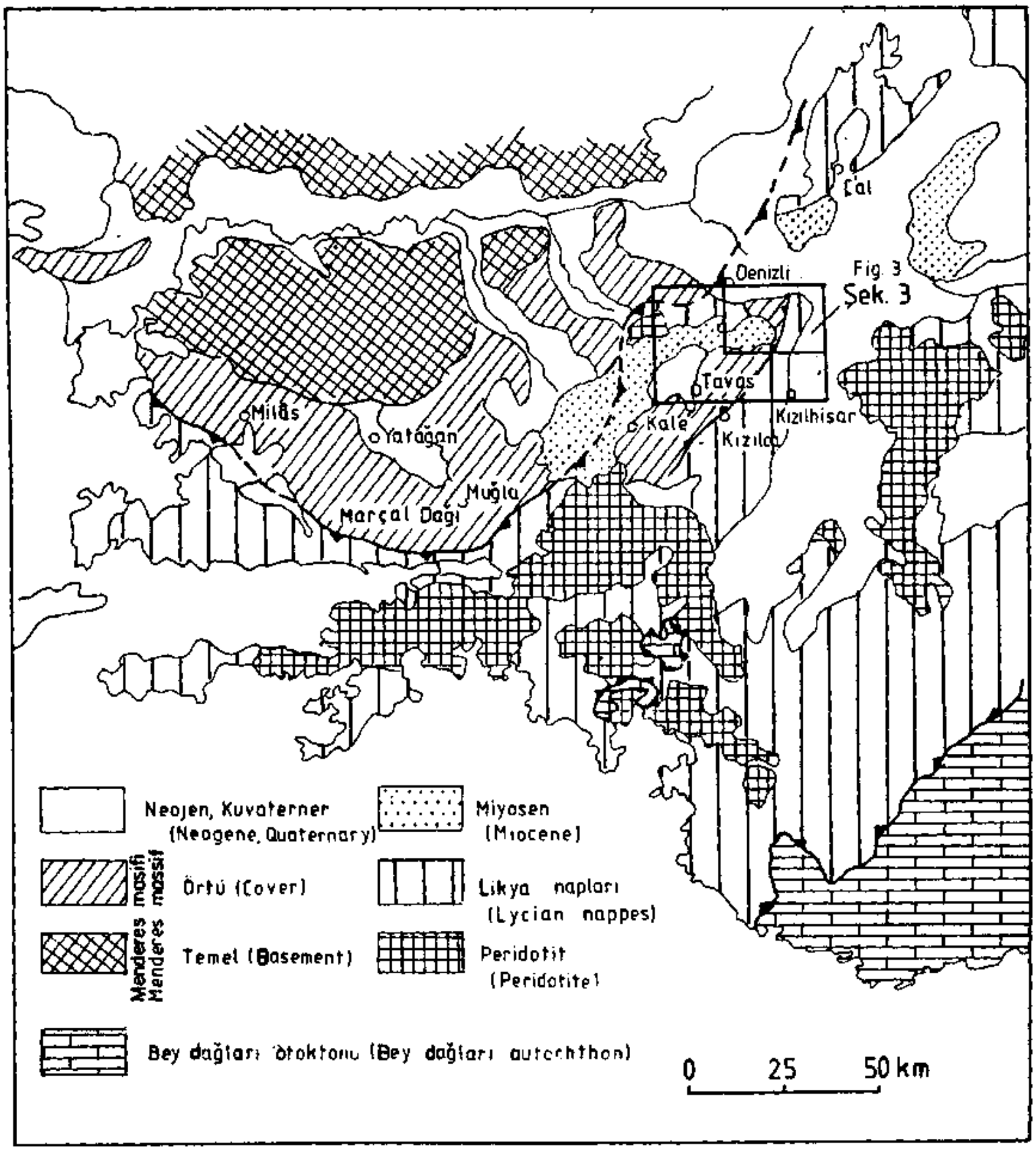

Fig.1 -- Simplified tectonic map of the Western Taurides showing the location of the studied area.

parts of the Lycian nappes, such that there is no metamorphic discontinuity between the Menderes massif and the overlying nappes. Apart from the Eocene mebamorphism, traces of an older Pan-African metamorphism have been documented in the Precambrian gneisses of the Menderes massif (Satır and Friedrichsen, 1986); (4) The Menderes massif forms the lowest tectonic unit in the region between the Bafa lake and Muğla and is the relative autochthon; (5) The Menderes massif has a relatively simple south and southeast dipping regional structure formed as a result of the regional uplift in the Oligocene. Large scale thrusts and isoclinal folds are not observed in the massif.

As can be deduced from the above features, the Menderes massif is a tectono-stratigraphic unit. The main features which characterize the Menderes massif are its stratigraphy, tectonic setting and regional metamorphism.

TECTONICFRAMEWORKOFTHEREGION

South of Denizli there are several major and secondary tectonic units forming a nappe stack. The relation between these tectonic units are shown schematically in Figure 2. At the top of the nappe stack is the Honaz ophiolite made up largely of ultramafic rocks, and restricted to the eastern part of the studied region. Sandak unit of the Lycian nappes lies tectonically below the Honaz ophiolite and above the Mesozoic-Early Tertiary cover sequence of the Menderes massif. The Menderes massif has an allochthonous position in the investigated area and is tectonically underlain by the Honaz shale made up of green phyllites which form the core of the Mount Honaz. In the east of the Mount Honaz this nappe stack is thrusc over the Göbecik Tepe unit made up of sedimentary rocks up to Mid/Late Eocene in age. The Zeytinyayla formation, which outcrops in the west can be correlated with the Göbecik tepe unit (Fig. 2).

main mass of the Menderes massif lie the Kale-Tavas Tertiary basin and allochthonous units of the Lycian nappes (Fig. 1 and 3). In the studied region the Mesozoic sequence of the Menderes massif tectonically overlies the Honaz shale, and is tectonically overlain by the Gereme formation of the Sandak unit. This relationship can be clearly observed in the Cemal damı locality (Fig. 3). The three uppermost formations of the Menderes massif which occur in the region are the Pınarlar formation, the Yılanlı formation and the Zeybekölen Tepeformation.

Pınarlar formation.— The Pınarlar formation outcrops around the Pınarlar village south of Tavas, and outside the area of Figure 3. It is made up mainly of slightly metamorphosed, pink, grey, bluish-grey, clean, medium to thickly bedded quartzite, fine to medium grained, red sandstone, red and green shale, yellowish-white dolomite, poorly sorted conglomerate with me-

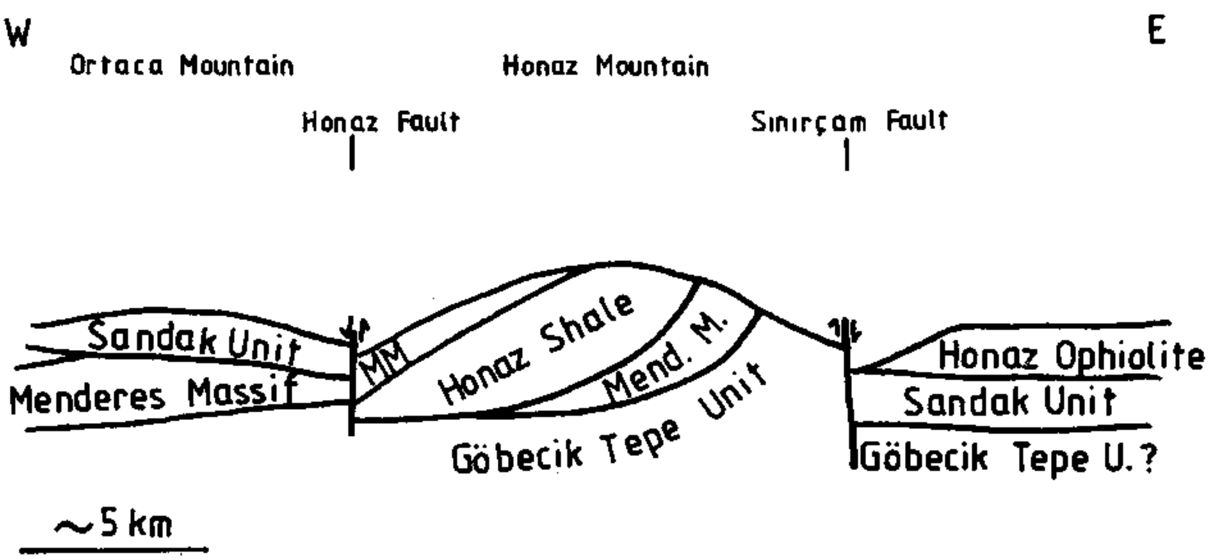

Fig.2— Schematized E–W section showing the relationship of the tectonic units in the studied region.

STRATIGRAPHY

The Menderes massif

Regionally metamorphosed sedimentary rocks outcropping over a large area extending from southeast of Tavas to the Mount Honaz are included into the Menderes massif sequence. Between these rocks and the

dium rounded, white and pink quartzite pebbles, and rare limestone lenses. The minimum apparent thickness of the Pınarlar formation is around 1000 meters. Quartzites with thicknesses of several hundred meters are the dominant lithology in the basal parts of the Pınarlar formation, abundant fusulinide forms were discovered in a dark grey limestone lens within the quartzites immediately east of the Pınarlar village, indicating a Late Permian age for the quartzites.

The Permo-Carboniferous sequence from the main mass of the Menderes massif is known for a long time from the regions of Göktepe, Karıncalıdağ and Babadağ (Phillipson, 1918; Onay, 1949; Kaaden and Metz, 1954; Schuiling, 1962; Dürr, 1975; Okay, 1985; Konak and others, 1987). This Permo-Carboniferous sequence is lithologically similar to the . Pınarlar formation, however, it includes abundant dark limestone and shale besides quartzite, and probably represents a slightly deeper marine environment during the Permo-Triassic.

Yılanlı formation.— A grey, light grey, thickly bedded-massif, locally laminated, fine-grained, locally gastropoda bearing apprx. 1500 m thick, monotonous carbonate sequence overlies the Pınarlar formation with a probable discordance (Neşat Konak, 1987, pers. communication). This limestone sequence, which represents the Mesozoic neritic carbonate cover of the Menderes massif, is named as the Yılanlı formation (Meşhur and Akpınar, 1984). It forms a major NE—SW striking mountain chain southeast of Tavas and extends to the slopes of the Mount Honaz (Fig. 1 and3). In the uppermost parts of this carbonate sequence west of the Kızılhisar rudist shell fragments are cautiously identified indicating that the age of the Yılanlı formation extends, as in other regions, up to the Late Cretaceous.

A typical feature of the Mesozoic neritic carbonate cover (Yılanlı formation) of the Menderes massif from the Bafa lake to Denizli is the presence of bauxite horizons within the carbonates. These bauxite horizons, which can be traced semi-continuously up to Denizli (Onay, 1949), are not found in the main mass of the Menderes massif southwest of Denizli and in the studied region, suggesting that the Mesozoic carbonate platform was deeper in this region and escaped the periodic subareal exposure inferred in regions farther southwest.

Zeybekölen Tepe formation.— An over 1000 m thick sequence of recrystallized pelagic limestone and shale occur above the Yılanlı formation. This sequence forming stratigraphically the highest part of the Menderes massif in the studied region, shows important lithological and thickness changes from the corresponding sequence in the main part of the Menderes massif, and was therefore assigned to a separate formation. The Zeybekölen Tepe formation is named after the Zeybekölen Tepe east of Tavas, and the reference section is the Sarp stream valley between Tavas and Büyükkale Viran hill (Okay, 1986). The Zeybekölen Tepe formation outcrops' extensively east of Tavas and in the slopes of the Mount Honaz (Fig. 3).

The Zeybekölen Tepe formation is made up of slightly metamorphosed, thinly to medium bedded, grey, light grey, red limestone with frequent chert nodules, calciturbidite, shalely limestone and friable, fine-grained, light green shale. In the Mount Honaz the Zeybekölen Tepe formation begins with thinly to medium bedded, grey pelagic limestone with 2-3 cm thick white chert bands overlying conformably the neritic limestones of the Yılanlı formation. These cherty limestones pass upwards to thinly bedded pink limestones, shalely limestone and shale. The Gereme formation of the Sandak unit overlies the shales with a nappe contact (Fig. 3).

In the main mass of the Menderes massif the pelagic limestone and flysch of Paleocene to Early Eocene age forms the equivalent of the Zeybekölen Tepe formation and can be traced from the Bafa lake to Denizli as a marker horizon defining the uppermost part of the Menderes massif. In the region between the Bafa lake and Milas the variegated pelagic limestones are called the Kızılağaç formation (Brinkmann, 1967) and the overlying flysch the Kazıklı formation (Dürr, 1975). The palaeontological age data on these formations all come from the Marçal mountains southeast of Milas where the effects of the regional metamorphism have been very slight (Gutnic and others, 1979; Konak and others, 1987). Tectonically the Kızılağaç formation represents the foundering of the Mesozoic carbonate platform of the Menderes massif in front of the advancing Lycian nappes, while the Kazıklı formation represents the overriding of the foundered carbonate platform by the Lycian nappes. Therefore, the ages of these formations will be strongly time-transgressive and related to their geographic positions with respect to the

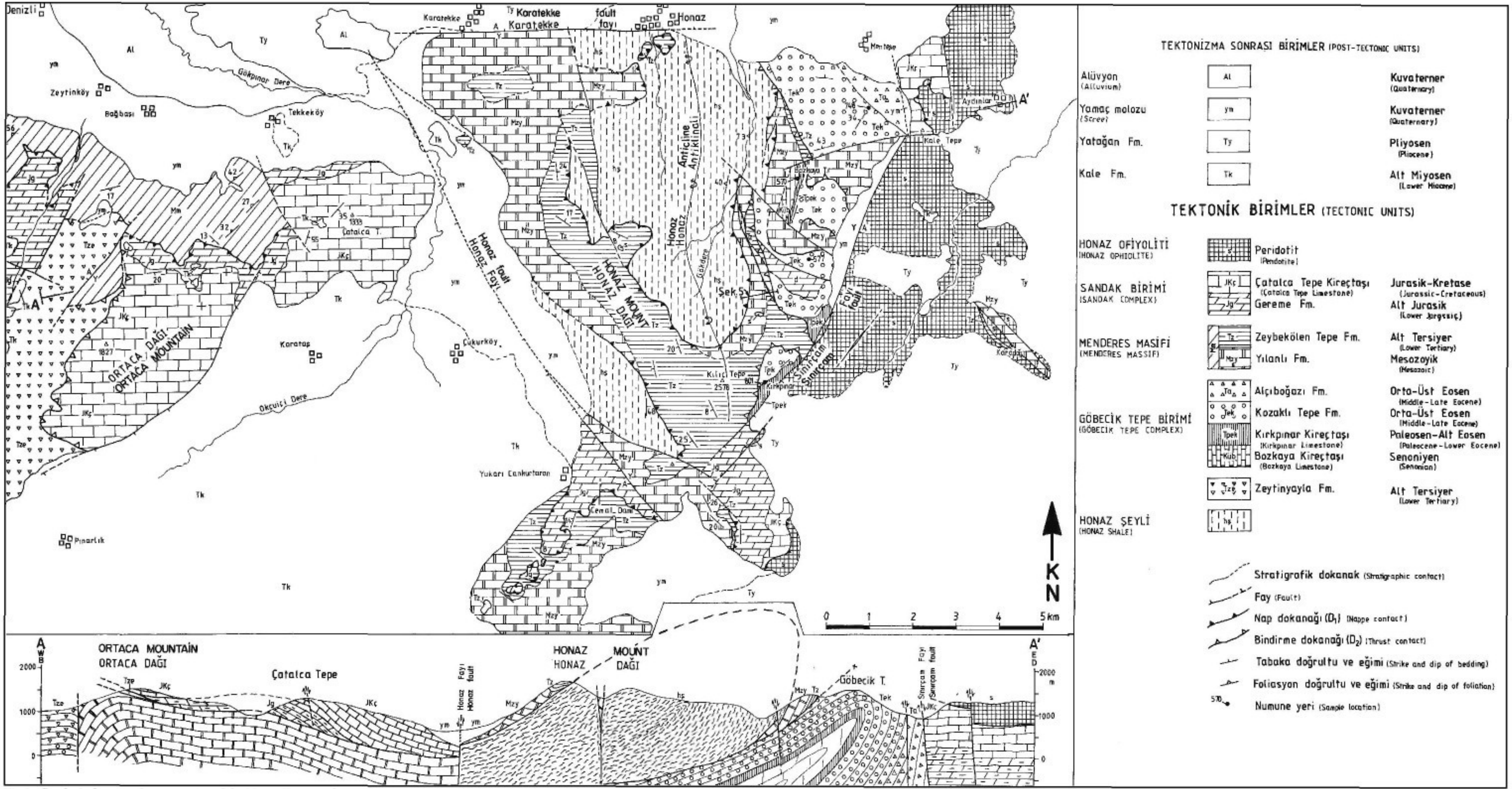

Fig 3 - Gerological map and cross-section of the Mount Honaz and surrounding area. For location see Fig 1 advancing Lycian nappes. However, the termination of the neritic carbonate cover of the Menderes massif with rudist-bearing limestones both in the region of Milas and Denizli indicates that in both of these regions the Lycian nappes were emplaced over the Menderes massif at a similar period. Therefore the age of the Zeybekölen tepe formation is provisionally taken as Paleocene-Early Eocene. The strong recrystallisation in the Zeybekölen tepe formation prevents any direct dating through fossils.

The allochthonous position of the sequence assigned to the Menderes massif in the studied area contrasts with the autocthonous position of the Menderes massif in the main mass. However, as described above this sequence is otherwise very similar, in terms of its stratigraphy, metamorphism and the overlying tectonic units,to that of the main mass of the Menderes massif. Its present allochthonous position is due to major post-metamorphic thrusting not observed in the main mass of the Menderes massif. For these reasons this sequence was assigned to the Menderes massif rather than to a new tectonic unit.

Sandak unit

Allochthonous units between the Menderes massif and the Bey Dağları authochthon are collectively known as the Lycian nappes. The Lycian nappes include various tectonic units ranging from continental margin deposits to ophiolites (Graciansky, 1972; Erakman and others, 1986). In the 250 km long region between the Bafa lake and Denizli a rootless tectonic unit of Mesozoic continental margin deposits overlies tectonically the Lower Eocene flysch of the Menderes massif. Recent studies have shown that this unit is in fact composite, and consists of two similar units called the Sandak and Haticeana units (Erakman and others, 1986). The main differences between these units are that the age of the Haticeana sequence ranges up to the Early Eocene, as in the Menderes massif, while the Sandak sequence is terminated at Late Cretaceous, and the post-Liassic rocks of the Haticeana unit are more pelagic than rocks of the corresponding age in the Sandak unit (Erakman and others, 1986). In the region between Muğla and Gölhisar, where these units are differentiated, the thrust stack consists from the base upwards: Menderes massif, Haticeana unit, Sandak unit, ophiolitic melange and ophiolite (Erakman and others, 1986). The unit tectonically overlying the Menderes massif south of Denizli is similar to the Sandak unit in terms of its stratigraphy and its tectonic position directly below the ophiolite, and was therefore assigned to this unit. In this region the Sandak unit consists of three formations: the Karaova formation, the Gereme formation and the Çatalca Tepe limestone.

Karaova formation.— The Karaova formation, which outcrops northwest of the Kızılcabölük outside the area of Figure 3, consists of slightly recrystallized, red, purple, bluish-grey shale, sandstone, conglomerate with quartz pebbles, quartzite and rare limestone horizons. It has a minimum thickness of 500 m and passes upwards gradually to the Gereme formation.

The Karaova formation has a striking appearance in the field with its multicoloured shales, and can be traced in its typical lithology from Bodrum to Uşak (Okay, 1985). The formation was first described and named by Phillipson (1918) from the Bodrum Peninsula. Its characteristic lithology in the Tauride Scythian lithofacies and the conformable overlying Late Triassic-Liassic Gereme formation indicate a Triassic age for the Karaova formation.

Gereme formation.— The Gereme formation is made up of monotonous, massively to thickly bedded, generally grey, dark grey dolomites; its maximum thickness is 500 meters; the cavernous surface wheathering is very characteristic for the dolomites of the Gereme formation. In the studied region the Karaova formation has acted as a decollement horizon during the thrusting so that the Gereme formation lies tectonically directly over the Menderes massif; the Gereme formation is overlain by the Çatalca Tepe limestone. These relationships can be clearly observed south of the Mount Honaz (Fig.3).

The Gereme formation can be traced with its characteristic dark dolomites within the Sandak and Haticeana units from the Bodrum peninsula to the region of Uşak. In fact, the Gereme formation was initially named by Blumenthal (1918) from the region of Milas, and is described in detail by Graciansky (1968) and Bernoulli and others (1974). In the studied region no fossils have been determined in the Gereme formation which consists completely of dolomites; its Late Triassic-Liassic age is based on fossils determined in undolomitised limestones from the region of Bodrum (Bernoulli and others, 1974); the same age range is cautiously accepted for the Gereme formation in the studied region.

Çatalca Tepe limestone.— Grey, dark grey, massive to thickly bedded limestone with rare small chert nodules which overlies the Gereme formation is named as the Çatalca Tepe limestone. The name of the formation derives from the Çatalca Tepe in the Ortaca mountain (Fig.3). The maximum thickness of the Çatalca tepe limestone is 750 meters. The Çatalca Tepe limestone is tectonically overlain by the Honaz ophiolite, which can be observed around the Kale tepe northeast of the Mount Honaz, and south of the Mount Honaz (Fig.3).

The Çatalca Tepe limestone is made up of slightly recrystallized micrite/dismicrite, sparsely packed biomicrite. Ostracoda, Milliolidae, Opthalmidiidae, Gastropoda, Brachiopoda, *Clodocoropsis* sp. have been determined in the collected specimens. Among these forms *Clodocoropsis* sp. indicates a Late Jurassic-Early Cretaceous age range. Based on this form, and the general stratigraphy of the Sandak unit (Erakman and others, 1986), the age range of the Çatalca Tepe limestone is taken as Dogger-Late Cretaceous.

An Upper Cretaceous flysch is generally present at the top of the Sandak unit (Erakman and others, 1986). This flysch, representing the emplacement of the ophiolite and ophiolitic melange over the Sandak unit, is not observed in the studied region probably because of tectonic omission, and the ophiolite sits directly on the Çatalca Tepe limestone.

Honaz ophiolite

The ophiolitic rocks outcropping extensively east of the Mount Honaz and consisting dominantly of serpentinised peridotite are called as the Honaz ophiolite. As can be clearly seen north of the Kale tepe (Fig.3), the Honaz ophiolite lies along a subhorizontal tectonic contact over the Çatalca Tepe limestone and constitutes the highest tectonic unit in the region. The Honaz ophiolite is cut by several subvertical faults; its contact with the Göbecik Tepe unit along the eastern margin of the Mount Honaz is also a major normal fault.

The major part (> 98%) of the Honaz ophiolite is made up of dark green, blocky, partially serpentinised and locally silicified harzburgite. Apart from the ultramafic rocks, there are also minor gabbro and chromite bodies.

Honaz shale

Slightly metamorphosed, dark bluish-green friable shale and siltstone sequence forming the core of the Mount Honaz is called as the Honaz shale. The Honaz shale has a very monotonous lithology and generally shows no bedding or a regular schistosity. It underlies tectonically the Yılanlı, and Zeybekölen Tepe formations of the Menderes massif (Fig.3). East of the Mount Honaz it is thrust along with the Menderes massif over the Eocene sediments of the Göbecik Tepe unit.

Apart from the ubiquitous dark green shales, the Honaz shale also includes reddish siltstone, sandstone and conglomerate with quartz pebbles intercalated with the green shales in the eastern side of the Gökdere. Cross-cutting the shales are rare thick (>10m) dark andesite dykes. No fossils have been found in the Honaz shale.

Göbecik Tepe unit

The Göbecik Tepe unit comprising Mesozoic-Lower Tertiary sedimentary rocks, is a newly discovered unit in the studied region. It tectonically underlies the Menderes massif or the Honaz shale along the eastern margin of the Mount Honaz (Fig.3). The Göbecik tepe unit has got an imbricated internal structure and its contact with the Honaz ophiolite in the east is faulted. The name of the unit derives from the Göbecik tepe south of the Menteşe village. Göbecik Tepe unit is divided into four formation (Fig.4); these are from the base upwards : the Bozkaya Tepe limestone, the Kırkpı-

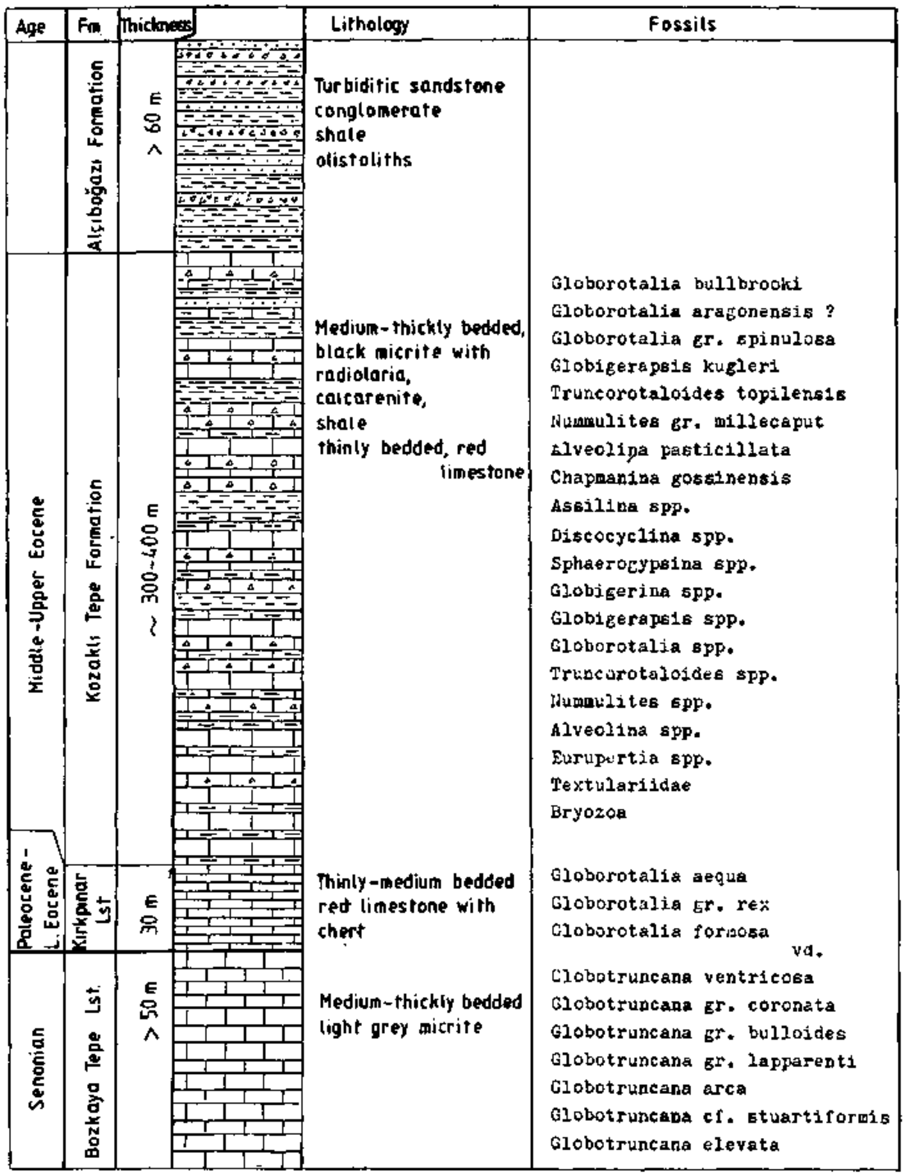

Fig.4— Stratigraphic section of the Göbecik tepe unit.

nar limestone, the Kozaklı Tepe formation and the Alçıboğazı formation.

Bozkaya Tepe limestone.—The Bozkaya Tepe limestone consists of medium to thickly bedded, cream coloured, microfossiliferous micritic limestone with infrequent chert lenses. It forms a ridge extending southward from the Bozkaya hill south of the town of Honaz. The reference and type sections are in the western side of the Bozkaya ridge. The Bozkaya tepe limestone has a minimum thickness of 80m; its lower contact with the Menderes massif is tectonic while it passes upwards conformably to the Kırkpınar limestone.

A specimen (570—A) collected from the middle part of the type section includes the microfossils of *Globotruncana ventricolosa, Globotruncana* gr. *coronata, Globotruncana* gr. *bulloides,* which indicate a Senonian (Campanian) age. Specimens collected higher up in the section include *Globotruncana elevata, Globot-*

runcana area, Globotruncana ventricosa, Globotruncana calcarata ?,forms characteristics for Late Campanian. The palaeontological data show that the Bozkaya Tepe limestone is of Late Senonian age. Subsequent studies have indicated that the age of the Bozkaya Tepe limestone may go down to Jurassic (Neşat Konak, 1987, personal communication).

Kırkpınar limestone.— The brick-red, thinly to medium bedded microfossiliferous micrite overlying the Bozkaya tepe limestone is called as the Kırkpınar limestone (Fig.4). The Kırkpınar limestone is overlain stratigraphically by the Kozaklı Tepe formation and tectonically by the Menderes massif. The thickness of the Kırkpınar limestone is 30 to 40 meters, the type and reference sections are along the Bozkaya ridge. The Kırkpınar limestone occurs, apart from the Bozkaya ridge, farther south in the Kırkpınar mevkii east of the Honaz summit. The name of the formation comes from this locality where its concordant stratigraphic contact with the overlying Kozaklı Tepe formation can be clearly observed.

The red, thinly bedded limestones of the Kırkpınar formation lie with a sharp but conformable contact on the grey limestones of the Bozkaya tepe limestone on the Bozkaya hill. A specimen (570—B) of the Kırkpınar limestone from this region contains *Globorotalia aequa, Globorotalia* gr. *rex, Globorotalia formosa, Globorotalia* gr. *laevigata, Globoratalia* gr. *convexa, Globigerina triloculinoides* indicating a Late Paleocene age. A specimen (801) collected from the uppermost part of the Kırkpınar limestone section in the Kırkpınar mevkii has the following microfossils *Globorotalia gracilis, Globorotalia rex, Globorotalia* gr. *mckannai ?, Globigerina* spp. of Late Paleocene ?-Early Eocene age. The palaeontological data and the age of the underlying Bozkaya tepe limestone indicates a Paleocene-Early Eocene age for the Kırkpınar limestone.

Kozaklı Tepe formation.— The Kozaklı Tepe formation consists of medium to thinly bedded, black micritic limestone with radiolaria, yellowish, thinly bedded sandy limestone, shalely limestone, red and green shale, calcarenite locally with abundant nummulites and thinly bedded red limestone. It lies conformably on the Kırkpınar limestone and is tectonically overlain by the Menderes massif (Fig.3). The name of the formation comes from the Kozaklı hill north of the Honaz summit. The reference section is the Kozaklı Tepe ridge. The thickness of the Kozaklı Tepe formation is 300-400 meters.

The Kozaklı Tepe formation constitutes the thickest and most widespread formation of the Göbecik tepe unit, and occurs along the eastern margin of the Mount Honaz. Its lower contact with the Kırkpınar limestone can be observed in the Kırkpınar mevkii where the thinly bedded red limestones of the Kırkpınar limestone are conformably overlain by the grey to black limestone-shale intercalation of the Kozaklı Tepe formation.

The major part of the Kozaklı Tepe formation consists of an intercalation of medium to thickly bedded, black, dark grey micritic limestone and yellowish grey shalely limestone. Within this sequence there are rare thinly bedded purple, red limestone, calcarenite locally with abundant nummulites and especially in the higher levels red and green shales. Calcarenites locally approaching bioclastic limestone contain apart from the nummulites reaching up to 2cm in length, quartz, green chert, chlorite-schist, shale and limestone fragments.

A specimen (748—A) from the red limestones from the Kozaklı Tepe formation contains *Globigerapsis kugleri, Truncorotaloides topilensis, Globorotalia bullbrooki, Globorotalia aragonensis ?* and indicate a Middle Eocene age. Several specimens collected from the nummulite-bearing calcarenites contain transoorted microfossils of Middle-Upper Eocene age. For example, a specimen (577) from the Kozaklı Tepe has *Discocyclina* spp., *Alveolina pasticillata, Nummulites millpcaput, Assilina* spp., and *Sphaerogypsina* spp., which indicate a Middle-Late Eocene or younger age. Interestingly the same specimen also contains transported Late Cretaceous microfossils, *Globotruncana* gr. *lapparenti, Gümbelina* spp. The palaeontological data indicate a Middle-Late Eocene age for the Kozaklı Tepe formation.

Alçıboğazı formation.— The Alçıboğazı formation consists of shale, sandstone, coarse sandstone and conglomerate with quartz, chert, serpentinite, limestone and volcanic rock fragments, and locally shows a wild-flysch character. The Alçıboğazı formation occurs southeast of the town of Honaz in the Alçıboğazı mevkii where it shows faulted contacts with the Kozaklı tepe formation. However, the lithological features of the Alçıboğazı formation suggest that it should overlie stratlgraphically the Kozaklı Tepe formation and thus should constitute the uppermost formation of the G6-becik tepe unit. Especially the upper parts of the Alçıboğazı formation have the features of an olistostrome; poorly sorted blocks of limestone, marblp, shale, red radiolarian chert reaching up to a few meters in size lie in a matrix of dirty green siltstone and sandstone. The minimum thickness of the Alçıboğazı formation is 60 meters.

No fossils have been found in the Alçıboğazı formation. However, its lithology indicates continuation of the rapid flysch deposition, which has already started in the Kozaklı Tepe formation, thus suggesting an Eocene age for the Alçıboğazı formation.

The Göbecik Tepe unit can be correlated in terms of its stratigraphy and tectonic setting with the Kızılca sequence described by Poisson (1977). The Kızılca sequence consists of Liassic to Middle Eocene sedimentary rocks with the post-Liassic sediments in pelagic facies. The Kızılca sequence which occurs in a small area south of Tavas has an imbricated internal structure and lies tectonically beneath the Mesozoic cover rocks of the Menderes massif.

Zeytinyayla formation

The flysch sequence locally with large limestone and sprpentinite blocks outcropping west of the Ortaca mountain is called the Zeytinyayla formation. The name of the formation comes from the Zeytinyayla mevkii northwest of the Ortaca mountain. The Zeytinyayla formation has an intermediate tectonic position between the Menderes massif below and the Sandak unit above (Fig.3).

The major part of the Zeytinyayla formation consists of slightly sheared preen, brown shale, siltstone sandstone and rare red shale. Within this clastic sequence there are occasional horizons of basic volcanic rock, calciturbidite and limestone, marble, radiolarite and serpentinite olistoliths up to 500 m in size. A limestone block (917—B) from the Zeytinyayla mevkii contains *Globotruncana area, Globotruncana* gr. *lapparenti, Globotruncana stuartiformis* of Campanian-Early Maastrichtian age. Thus, the depositional age of the Göbecik tepe formation is post-Maastrichtian.

The relation of the Zeytinyayla formation to the other tectonic units in the region is not well known. In terms of its lithology and the absence of regional metamorphism the Zeytinyayla formation is similar to the Göbecik Tepe unit and especially to the Alçıboğazı formation. However, it differs from this unit through its tectonic position above the Menderes massif. Nevertheless, the Zeytinyayla formation is here correlated with the Göbecik Tepe unit with its tectonic position above the Menderes massif attributed to a later thrusting event. However, it is equally possible that the Zeytinyayla formation constitutes the top part of the Sandak unit.

Neogene units

Neogene rocks overlie unconformably all the tectonic units in the studied region and are grouped into two lithostratigraphic units : the Kale (Meşhur and Akpınar, 1984) and the Yatağan (Becker-Platen, 1970) formations.

Kale formation.— In the studied region the major part of the Kale formation is made up of reddish green conglomerate with well rounded, polished serpentinite pebbles in a sandy and silty matrix Intercalated with the conglomerate are brownish green sandstone, siltstonp. shale and thin discontinuous coal beds. The Kale formation has a thickness of above 800 m and represents post-tectonic molasse facies deposits. It is overlain unconformably by the Yatağan formation.

Macrofossils collected by Becker-Platen (1970) 4 km south of Çukurköy (Fig.3) have given a Rupelian-Helvetian (Early Oligocene-Middle Miocene) age range. A more precise age range for the Kale formation comes from east of the town Kale outside the studied area; palaeontological data from this region indicate an Aquitanian age (Becker-Platen, 1970, Gökçen, 1978). In the Kale region the Kale formation is overlain through an angular unconformity by the Burdigalian marine limestones (Nebert, 1961, Becker-Platen, 1970, Gökçen, 1978). Thus, the data suggest that the Kale formation in the studied area is of Aquitanian age.

Yatağan formation.— The Yatağan formation, which outcrops east and north of the Mount Honaz (Fig.3) consists of white porous, hard, lacustrine limestone, grey, greyish green limely siltstone, sandstone, mudstone, basalt and coal. Yatağan formation has been assigned a Pliocene age by Becker-Platen (1970) based on molluscs, ostracoda and palinological determinations.

STRUCTURE

The studied region is a typical thrust/nappe belt. The structures in the region can be classified into three types based on their features and times of formation.

Nappes (D_1)

The first structures observed in the region is the stacking of the tectonic units as nappes. At the base of the nappe stack lies the Honaz shale. Honaz shale is tectonically overlain by the Mesozoic cover units of the Menderes massif. Farther up in the nappe stack lies the Sandak unit (Fig.2). The Sandak unit is thrust over by the Honaz ophiolite, which forms the highest unit in the nappe stack. In areas not directly affected by the later movements, the nappe contacts between these tectonic units are subhorizontal.

There are no data on the age of emplacement of the Honaz ophiolite over the Sandak unit. However, it is known that in the Taurides the ophiolite obduction over the carbonate platforms occurred during the Late Cretaceous (Özgül, 1976). Likewise, the age of the Sandak unit farther south is documented to range up to the Late Cretaceous (Erakman and others, 1986). Therefore, the emplacement of the Honaz ophiolite over the Sandak unit in the studied region is thought to have occurred during the Late Cretaceous.

The thrusting of the Sandak unit over the Menderes massif probably occurred during the Mid-Eocene. This emplacement age is based on the stratigraphy of the Menderes massif in the Margal mountains where it is shown to extend up to the Early Eocene (Gutnic and others, 1979; Konak and others, 1987).

The thrusting of the Menderes massif over the Honaz shale has occurred before the D1 movements; the D_2 structures of Late Eocene-Oligocene age truncate the tectonic contacts between the Honaz shale and the Menderes massif (Fig.3).

Thrusts and overturned isoclinal folds (D_2)

Following the emplacement of the nappes into the region, NNE—SSW striking and westward dipping thrusts and eastward overturned close-isoclinal folds developed as a result of NW-SE directed compression. The thrusts developed penecontemporenously with the overturned isoclinal folds, and frequently the isoclinal folds were transported over their overturned and sheared limbs. The D_2 structures have affected the D_1 nappe contacts and resulted in their folding and truncation; for example the D_1 nappe contact between the Honaz shale and the Menderes massif is folded and locally truncated by the D_2 structures (Fig.3).

The important D_2 structures shown on Figure 3 are the Honaz anticline and the Honaz thrust. These structures extend southward towards Tavas as a major SW—NE striking anticline with an axial length of over 20 km (Tavas anticline) and as a major thrust (Tavas thrust; Fig. 1, Okay, 1986).

Honaz anticline and Honaz thrust.— The Honaz anticline is a N—S striking, eastward overturned close anticline with an axial length of about 10 km (Fig. 2 and 3). Honaz anticline is sheared off along its overturned limb and is thrust over the Göbecik Tepe unit along the Honaz thrust. The nappe contact between the Honaz shale and the Menderes massif is folded during the D_2 phase and acquired the silhouette of the Honaz anticline (Fig. 3). Honaz shale occurs in the core of the Honaz anticline while the Mesozoic cover units of the Menderes massif occur in its flanks.

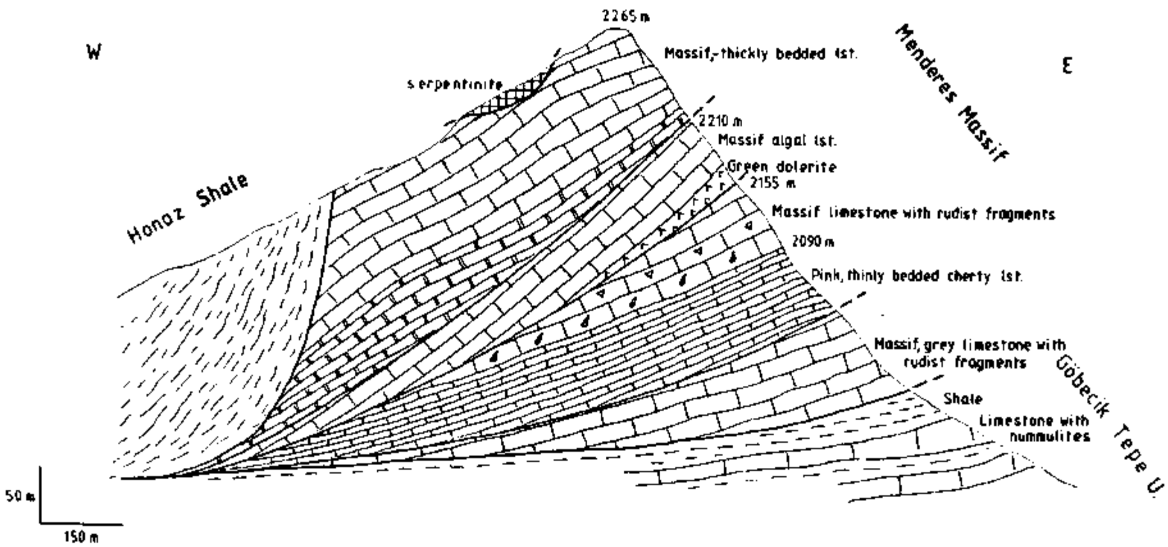

Fig.5 — Imbricated tectonic slices of the Mesozoic cover units of the Menderes massif along the Honaz thrust, Baymanlı ridge, the eastern side of the Mount Honaz. For location see Fig.3.

In some areas the Honaz thrust is not represented by a single thrust plane but by several closely spaced subparallel thrusts; for example in the eastern side of the Baymanlı ridge, the upper parts of the Menderes massif sequence is repeated four times in a 300 m thick section (Fig . 1 and 5).

The age of the D_2 structures is post-Middle Eocene based on their trans-cutting relationships to the D_1 structures; the Aquitanian Kale formation in mollasse facies gives an upper age limit for the D_2 structures. Therefore the age of the D_2 structures are constrained to the Late Eocene- Oligocene time interval.

Normal faults (D_3)

Following the NW—SE directed compression, the region was affected by a tensional regime with the formation of WNW—ESE and NE—SW striking major normal faults. These faults with important vertical throws cut the rocks of the Kale formation suggesting a post-Aquitanian age for their latest movements; their relation to the Yatağan formation is not known. Some of these important faults in the studied region are the Honaz, Sınırçam and Karatekke faults which surround the Mount Honaz like the sides, of a triangle (Fig.3). The Honaz fault is a major tectonic line limiting the southward and westward extension of the Honaz shale; a minimum of 1000 m throw is estimated along the Honaz fault. The Sınırçam fault can be traced for 11 km along the eastern side of the Mount Honaz; it constitutes the western limit of the Honaz ophiolite.

METAMORPHISM

All the tectonic units in the region with the exception of the Göbecik Tepe unit and the Zeytinyayla formation, have been affected by a low-grade regional metamorphism. The regional metamorphism was syn- to post-nappe emplacement (D_1), so that no discontinuity in metamorphic grade is observed across the nappe contacts, such as that between the Sandak unit and the Menderes massif; the metemorphic grade shows a regular decrease upwards in the tectonic sequence. However, the metamorphic grade in the studied region does not exceed that of the greenschist facies; biotite and garnet do not occur even in the Honaz shale which lies at the base of the nappe stack. The regional metamorphism in the studied region is thought to be of Eocene age similar to that in the southern margin of the Menderes massif.

In the Western Taurides the regional metamorphism decreases gradually and eventually disappears with increasing distance from the Menderes massif. The studied region lies in the threshold of this metamorphism. East of the Mount Honaz no effects of the metamorphism is generally discemable while in the west an increasing degree of metamorphism is observed.

The absence of metamorphism in the Göbecik Tepe unit, which is believed to constitute palaeogeographically the eastward extension of the Menderes massif, may be due for two reasons: (a) the nappe cover over the Göbecik Tepe unit during the Eocene may have been thinner than that above the Menderes massif, or may have been non-existent, and/or (b) the heat flux during the Eocene may have been much higher in the region of the Menderes massif compared to the region of the Göbecik Tepe unit.

The Do structures in the studied region were post-metamorphic and resulted in the juxtaposition of the metamorphic and non-metamorphic units, for example the Menderes massif was thrust over the non-metamorphic Göbecik Tepe unit.

THE TECTONIC EVOLUTION OF THE REGION

The Taurides are characterized by sedimentation without important orogenic breaks from the Early Paleozoic to the Late Cretaceous. The orogenic movements started in the Late Cretaceous, and have continued with several phases up to the present. Within this framework the tectonic evolution of the region is shown in Figure 6 for different time periods and is explained briefly below.

Early Cretaceous (Fig. 6a)

The relative position of the tectonic units in the studied region are shown in Figure 6a for the Early Cretaceous when the orogenic movements have not started, and quiete carbonate sedimentation was continuing. The important features are the position of the Göbecik Tepe unit in a pelagic basin (Kızılca basin, Poisson,1984) south of the Menderes massif, and the position of the Sandak unit to the north of the Menderes massif. Farther north lay the Tethys ocean now represented by the ophiolites south of the Izmir-Ankara suture.

The Kızılca basin is a pelagic trough thought to have lain between the Menderes massif and the Bey Dağları autochthon (Poisson, 1984). This basin, bordered by two shallow carbonate platforms, was initiated during the Lias and preserved its basinal character till the Late Cretaceous; in this respect it can be correlated with the Ionian basin in Greece also lying between two carbonate platforms (Poisson, 1984).

Late Cretaceous (Fig. 6b)

During the Senonian the ophiolites were emplaced on the continental margin and platform sediments of the Taurides. The emplacement of the Honaz ophiolite over the Sandak unit is similarly thought to have occurred in the Late Cretaceous.The effects of this major obduction event were felt in the Anatolide-Tauride carbonate platform: the platform was initially uplifted by elastic rebound and partially eroded, then as the ophiolite nappe approached the platform foundered and became a transient pelagic basin (Fig. 6b). The ophiolite nappe did not reach the Menderes massif and the Göbecik Tepe unit, where the sedimentation continued up to the Eocene.

Middle Eocene (Fig. 6d)

Following the ophiolite emplacement during the Late Cretaceous, the orogenic movements started again in the Middle Eocene after a quiete period during the Palaeocene. During the Middle Eocene the carbonate platform was internally imbricated and the Sandak unit with its cover of the Honaz ophiolite was thrust over the Menderes massif. The Menderes massif became buried under the nappe cover and underwent a Harrovian-type regional metamorphism.

Late Eocene—Oligocene (Fig. 6e)

After the nappe emplacement and regional metamorphism, major eastward overturned megascopic folds and thrusts developed during the D_2 period due to NW—SE compression. The Mesozoic cover units of the Menderes massif became detached from their basement and were thrust over the Göbecik Tepe unit during these D_2 movements.

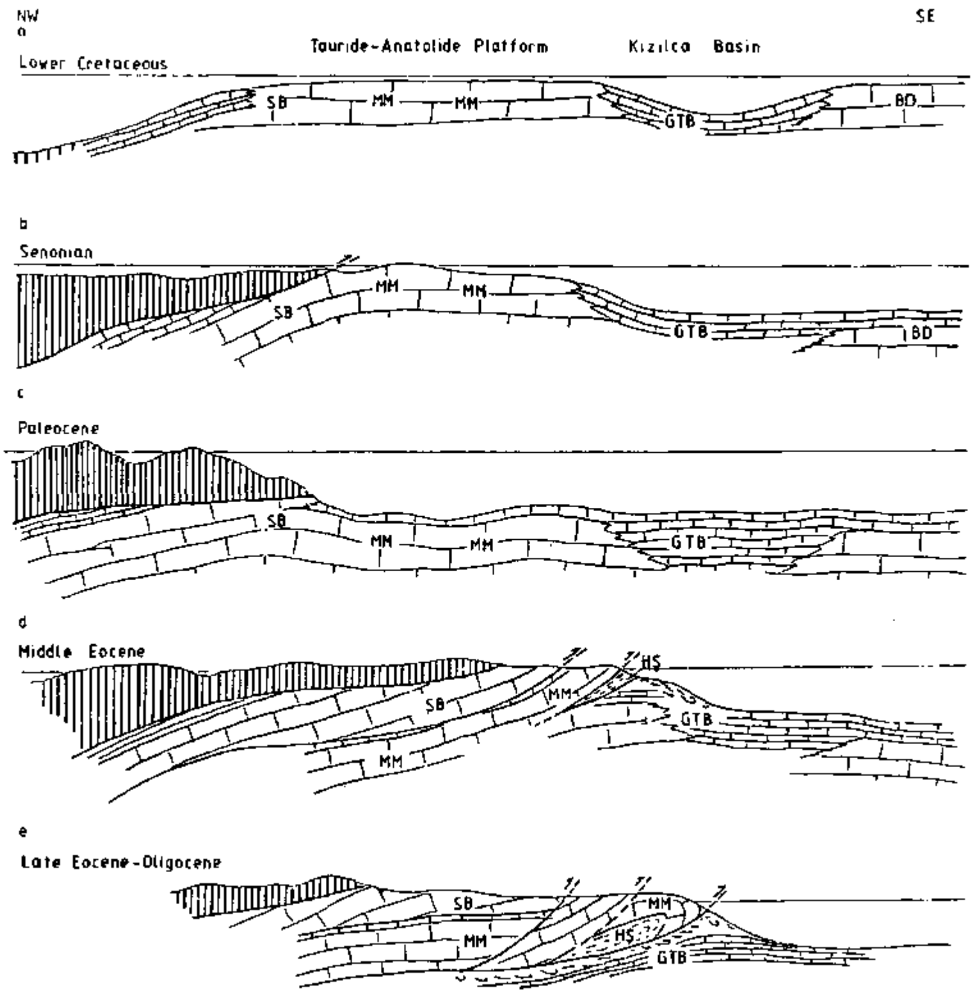

Fig.6—**Schematized paleogeographic sections illustrating the tectonic evolution of the region.**
SB—**Sandak unit; MM**—**Menderes massif; GTB**—**Göbecik tepe unit; BD**—**Bey dağları autochthon; HS**—**Honaz shale.**

Late Miocene—Pliocene

Fluvial sediments of the Kale formation were deposited over the whole area during the Aquitanian. Normal faulting following this sedimentation resulted in a block uplift of the Honaz and Ortaca mountains and subsequent erosion. This tectonic regime is still continuing today.

CONCLUSIONS

The important conclusions and results of this study are listed below :

Five major tectonic units are present in the region of the Mount Honaz. These are from the base upwards: Göbecik Tepe unit, Honaz shale, Menderes massif, Sandak unit and the Honaz ophiolite.

The Göbecik Tepe unit, which is first described in this study, comprises a Mesozoic-Tertiary sedimentary sequence ranging up to Middle/Upper Eocene in age, and constitutes the relative autochthon in the studied region.

The Menderes massif sequence represented by the Mesozoic cover rocks, has an allochthonous position and is thrust over the Honaz shale and the Göbecik Tepe unit.

The Mount Honaz has an eastward overturned anticlinal structure. The Honaz anticline is sheared along its overturned limb and is thrust eastward.

Three deformational phases are differentiated in the region: D_1, nappe emplacement and metamorphism during the Middle Eocene; D_2, eastward overturned folds and thrusts during the Late Eocene—Oligocene; D_3, post-Aquitanian normal faulting.

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STRATIGRAPHY OF THE MUŞ TERTIARY BASIN

Ergün AKAY*; Emin ERKAN* and Engin ÜNAY*

ABSTRACT.— The Muş Tertiary basin is composed of independently developed basins which are the Middle-Late Eocene, Uppermost Eocene-Early Miocene, Middle-Late Miocene, Pliocene and Uppermost Pliocene-Quaternary in age. The Middle-Late Eocene basin is represented the detritics of Kızılağaç formation. The Uppermost Eocene-Early Miocene basin comprises the transgressive detritics of Uppermost Eocene Ahlat formation at the bottom. The upper part of this formation is composed of intercalated continental and marine detritics. Upwards the calcareous detritic rocks of Gerisor formation and detritics of Norkavak formation are seen. Above them Middle-Late Oligocene Yazla formation composed of marine detritic and calcareous rocks is exposed. The sequences varies upwards, to rhyolitic volcanic rocks of the Uppermost Oligocene of Sergen formation of which the exposures are few. On top of the above mentioned rocks, regressive detritic and calcareous rocks of Adilcevaz formation of Uppermost Oligocene-Eaiiy Miocene are exposed. In the Middle-Late Miocene basin, the rocks are developed under the conditions which caused the deformation of the previous basin and is represented by acidic volcanics of Elçiler formation. In the Pliocene basin, basaltic andesites, agglomerates and tuffites of Solhan formation and lacustrine deposites of Zırnak formation are developed. The Uppermost Pliocene-Quaternary basin represented by lacustrine and fluvial deposites of Bulanık formation, tuffites of Nemrut formation, sandstone and conglomerates of Muş Ovası formation and Holocene alluvial deposites.

Mineral Res.Expl.Bull., 109, 53-64, 1989

EVOLUTION OF THE POST-COLLISIONAL CRATONIC BASINS IN EASTERN TAURUS

Ergün AKAY*

ABSTRACT. — A number of cratonic basins developed as a result of continent—continent collision in Eastern Taurus. During this collision, some changes in the rate of continental convergence considerably affected the evolution of these cratonic basins. Following the Late Eocene collision, the Oligocene transgression started as a result of a decrease at the rate of continental convergence. The weaker effect of the convergence rate during the Middle—Late Oligocene resulted in lithospheric deformation forming some troughs and uplifts. Further decrease in the effect of the convergence rate during the Uppermost Oligocene, caused the widespread Early Miocene transgression. During Langhian, a sudden increase in the effect of the convergence rate gave rise to the uplifting of the whole region above the sea level. In Serravalian—Tortonian, continental and marine transgressions took place, due to the relatively decreasing effect of the convergence rate. However, convergence rate which was considerably effective in the Uppermost Tortonian, played an important role in the formation of Anatolia. A partial decrease in the convergence rate during the Early—Late Pliocene provided the deposition in continental basin conditions. In the Uppermost Pliocene, the convergence rate increased again and as a result of this, the tectonic lines representing the Uppermost Tortonian phase were displaced. The effect of convergence rate which was relatively low during the Early—Late Pleistocene provided restricted continental depositional conditions. Increasing effect of convergence rate from the Late Pleistocene to the present—day is responsible for the active faults. This evolutionary scheme of the post—Eocene basins proves that the collision has been continuing since Late Eocene.

INTRODUCTION

Up to the present—day, some investigations were made on the geodynamic evolution of Eastern Taurus (Baştuğ, 1980; Dewey et al., 1986; Şaroğlu et al., 1980; Şaroğlu, 1985; Şengör, 1980; Şengör and Yılmaz, 1983; Y.Yılmaz et al., 1987). Eastern Taurus is bounded by the East Anatolian Fault(EAF) in the west and the Southeastern Anatolian overthrust in the south. The interpretation of the results obtained from the detailed stratigraphic and tectonic studies of the Muş Tertiary basin (Akay et al., 1989) with the other previous investigations in. Eastern Taurus and palaeomagnetic data obtained from Taurid belt, led to the reconstruction of the post-Eocene geodynamic evolution of Eastern Taurus.

GENERAL GEOLOGICAL CHARACTERISTICS OF EASTERN TAURUS IN PRE-OLIGOCENE

In Eastern Taurus, the various units showing different stratigraphic, lithological, tectonic and metamorphic features came together in pre—Oligocene. Of these units, the Bitlis massif is made up of Palaeozoic—Mesozoic rock assemblages which underwent metamorphism in pre—Maastrichtian and an ophiolitic unit which was emplaced in the Late Cretaceous (Baştuğ, 1980; Göncüoğlu and Turhan, 1983, Çağlayan et al., 1983). On the other hand, the northern part of the massif is made up of an accretionary prism (Şengör and Yılmaz, 1983).

The Munzur mountain is made up of platform type rock assemblages which span from Triassic to Campanian and overlain by an ophiolitic melange which was emplaced in the Late Cretaceous (Özgül and Turşucu, 1983). On the other hand, the Keban platform has greenschist character and comprises rock units which were deposited in a period ranging from Permian to Upper Cretaceous (Özgül and Turşucu, 1983; Yazgan, 1983). The rock assemblages of Malatya metamorphic platform resemble to those of the Keban metamorphic platform (Yazgan, 1983).

POST-EOCENE GEODYNAMIC EVOLUTION

Before we consider the tectonic events which happened in a time interval from Oligocene to the present—day, we will briefly describe some events before and during the Eocene. Şengör (1980) claims that the Southeast Anatolian suture zone formed as a result of the continent—continent collision, during the Middle Miocene. However Michard et al. (1985) puts forwards two hypothesis concerning the region. According to the first hypothesis, in a part of Eastern Taurid belt which is baunded by the Ecemiş, fault, the continent-continent collision took place in the Uppermost Cretaceous. Afterward, until Eocene, melting of the subducting oceanic crust remnants gave rise to a volcanism. The Arabian continent plunged northwardly underneath the Anatolides in a time interval of Eocene to the Middle Miocene and as result of this, a volcanism took place. The continent—continent collision has been continuing up to the present—day. According to the second hypothesis, in Eastern Taurus, the northward dipping subduction of the oceanic crust, lasting until Eocene gave rise to a calc—alkaline volcanism. Afterwards, the volcanism which took place in a period lasting until Miocene, is caused by either melting of the remnants of the oceanic crust as a result of subduction or melting of the lithosphere as a result of deformation.

As mentioned above, the geodynamic evolution of Eastern Taurus which is questionable in the period ranging from Oligocene to the present—day was tried to be reconstructed under the light of some evidences such as the geometry of the basins in the region, time of the basin formation—closure, time of the basin, deformation, type and age of volcanism.

Uppermost Eocene-Late Oligocene depositional period

In a section of Muş Tertiary basin, where the sediments representing this period are the thickest (Akay et al., 1989), the approximately 1000 m. thick fluvial sediments of the Uppermost Eocene Ahlat formation are found at the base. The approximately 500 m. thick marine sediments of Lower Oligocene Norkavak formation overlie them conformably. These marine sedimepts are conformably overlain by the approximately 3500 m. thick Yazla formation of Middle—Upper Oligocene which was deposited below the wave base. In eastern part of the basin, the thickness of the fluvial sediments belonging to the Ahlat formation is unknown (assumed to be very thick). In this section, the 170 m. thick limestone beds of the Lower—Middle Oligocene Gerisor formation, which were deposited within the wave base, rest conformably on these fluvial sediments (Ahlat formation). These limestones are overlain by the 60 m. thick sediments of the Upper Oligocene Yazla formation, which were deposited below the wave base. The thickness of the sediments which were deposited below the wave base, considerably decreases to the east of the Muş, basin. On the other hand, a westward decrease in the thickness of the same sediments is also assumed to be considerable.

In the Hınıs basin, located to the north of the Muş, basin (A.Yılmaz et al.,1988), the unit corresponding to the Uppermost Eocene fluvial sediments of the Muş, basin is 1200 m. thick. It is transitionaly overlain by about 300 m. thick Oligocene marine sediments which are interbedded with andesite in the upper levels.

In the vicinity of Palu, the Middle—Upper Oligocene sediments, composed of detritic rocks at the bottom and carbonates at the top, overlie the basement with an angular unconformity (Sirel et al., 1975).

The Lower—Upper Oligocene carbonates which were deposited on the continental shelf show transition into the underlying Upper Eocene marine carbonates, around Elazığ (Sirel and Gündüz, 1979).

On the other hand, the Lower—Upper Oligocene marine sediments show transition into the underlying Upper Eocene marine sediments in the Kahramanmaraş Tertiary basin (Uysal et al., 1985).

Consequently, a post-collisional restricted eratonic basin which was extending from the Kahramanmaraş, basin to the Hınıs basin or further to northeast

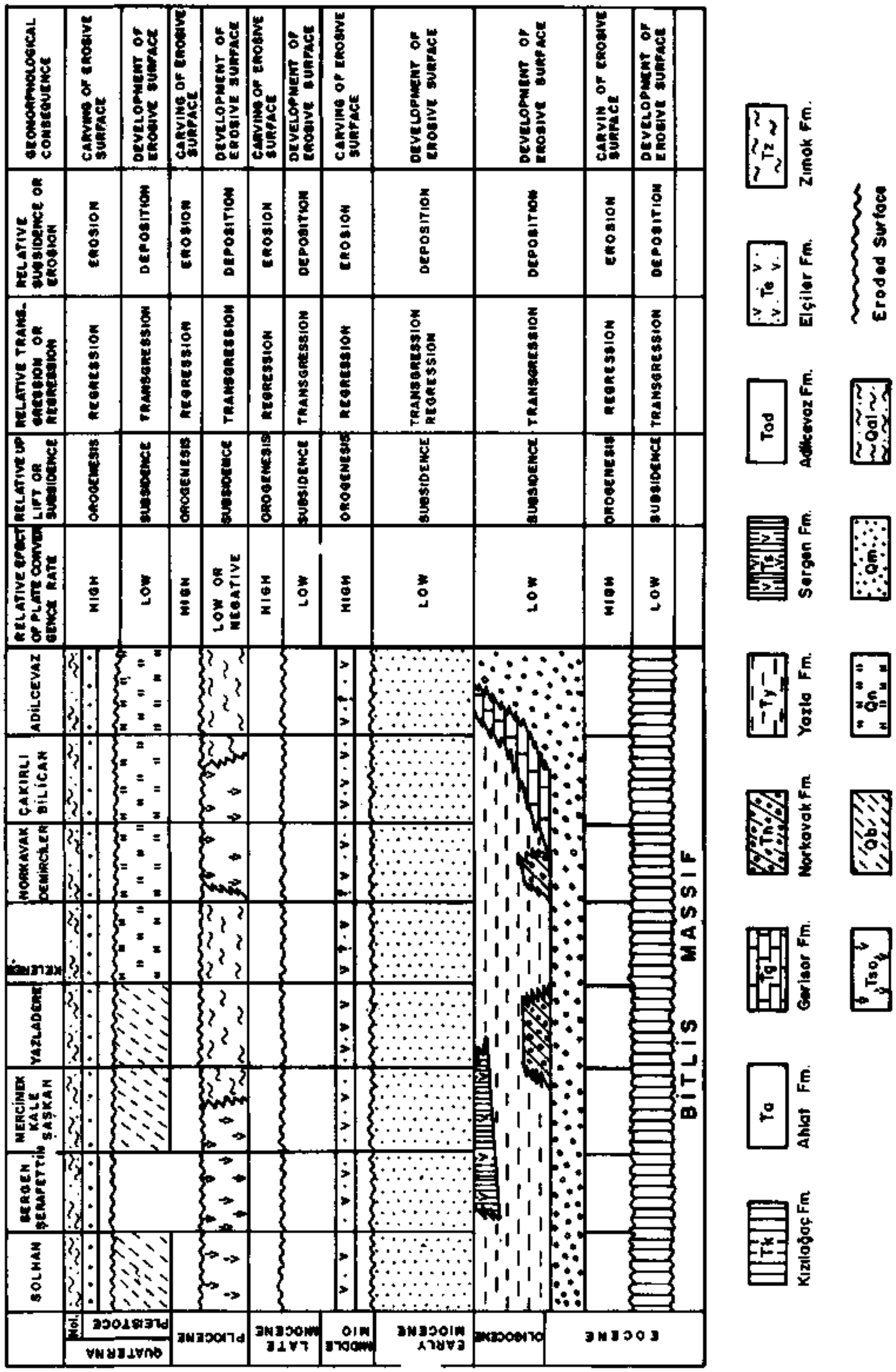

Fig.1 Table of comparative composite columnar sections of Muş Tertiary basin (Akay et al., 1989) and geodynamic events coinciding with periods at the table.

and lying more or less along the present East Anatolian Fault, had existed in the Oligocene.

In the Muş Tertiary basin (Akay et al., 1989), the marine claystone—siltstone beds of the Upper Oligocene Yazla formation show transition into the rhyolitic volcanic rocks of the Sergen formation. These beds are nonconformably overlain by the Miocene sediments (Fig-1).

Livermore and Smith (1983) further reported that the Anatolia has been undergoing continuous compression about in N—S direction during the Oligocene—Quaternary. On the other hand, according to Dercourt et al. (1986), the rate of continental convergence suddenly decreased at the boundary between Eocene and Oligocene and this decrease continued during the Oligocene. Due to the high rate of continental convergence during the Late Eocene. The Arabian and northern continents juxtaposed and as a result of this, an orogenic belt developed. Afterwards, decreasing rate of continental convergence during the period of Late Eocene to the end of Early Oligocene gave rise to deposition of fluvial sediments of the Muş—Hınıs basin and the overlying sediments which were deposited within the wave base, the fluvial sediments of the Adilcevaz area (Demirtaşlı and Pisoni, 1965) and the shelf carbonates of Elazığ—Palu area. The effect of compressional forces was slightly weak and continuous during the Middle-Late Oligocene and as a result of this, lithospheric deformation developed. The fact that the thickness of the sediments which were deposited below the wave base in the Muş basin, decreases from 3500m. to 60 m. at a distance of 30 km., can be explained by this event. The litosphere underwent an intensive deformation in the Late Oligocene, while as the sediments of the basins were not deformed. The deformation induced melting of the litosphere led to the development of rhylotic volcanic rocks in the Muş basin (Akay et al., 1989), quartz porphyries in the vicinity of Palu (Sirel et al., 1975) and andesitic lavas in the Hınıs basin. On the other hand, shelf depositional conditions prevailed in the vicinity of Elazığ—Palu during the Early—Late Oligocene while the fluvial sediments were deposited in the vicinity of Adilcevaz. The East Anatolian Fault,which is now oblique to that compression direction, can be the present-day continuation of a fault system which has been active since that time. Also it is very likely that the Ecemiş fault which is assumed to have moved mainly during the Late Eocene (Akay and Uysal, 1987) and the ancestor of the East Anatolian Fault moved together within the same system in the late phase of the Late Eocene collision. Meanwhile, a fault which is about in the eastern extension of the North Anatolian Fault should have operated as an conjugate of the East Anatolian Fault.

Aquitanian—Burdigalian depositional period

The Early Miocene transgression took place over a very extensive area in Anatolia.

In the Muş basin (Akay et al., 1989) the Lower Miocene Adilcevaz formation (Fig. 1),shows transitions into the underlying Oligocene Yazla formation in marine conditions in the eastern part of river Murat and in continental conditions in the western part of river Murat. The Lower Miocene formations consisting of carbonates and clastic rocks are generally 1000 m.thick. These formations which were deposited within the wave base show regressive character with respect to the underlying Oligocene sediments. In the Hınıs basin, the Lower Miocene sediments sometimes show transition into the underlying Oligocene sediments and sometimes transgressively rest upon the basement (A.Yılmaz et al., 1988). In the vicinity of Adilcevaz, the Adilcevaz formation shows transition into the red fluvial sediments of the underlying Ahlat formation (Akay et al., 1989) and is 500 m. thick. The Lower Miocene sediments show transition into the underlying Oligocene sediments, as shelf sediments around Elazığ—Palu (Sirel et al., 1975; Sirel and Gündüz, 1979). The Lower Miocene sediments overlie the Upper Eocene sediments with angular unconformity to the northwest of Malatya (Örçen, 1986). In the Kahramanmaraş, basin, the Oligocene sediments show transition into the Lower Miocene sediments (Uysal et al., 1985). The transgressive limestone and sandstone beds over the basement rocks to the north of lake Van which yield Aquitanian and Burdigalian ages are a few hundred meter thick and probably have an age reaching the Lower Langhian (Gelati, 1975). The Early Miocene transgression took place not only in this region but in the Adana (Yalçın and Görür, 1984), Antalya and Beydağları basins (Akay et al.,1985). Therefore, it is concluded that the existing post collisional cratonic basin became wider.

Which mechanism gave rise to this widening of the basins during the Early Miocene and why?

In this period regressive sediments were deposited in marine basins and transgressive-regressive sediments in continental basins. It is apparent that the continuous slow compression resulted in lithospheric deformation. The phenomenon which happened during the Early Miocene can be explained by the fact that the effect of continental convergence rate which had resulted in this deformation, considerably decreased during the Uppermost Oligocene (Dercourt et al., 1986). As a result of this, the subsidence of continental uplifts gave rise to transgression, while the uplift of subsided basins gave rise to regression.

More or less invariable thickness of sediments and the lack of volcanic intercalations in the Early Miocene basins, indicate that the effect of continental convergence rate is much weaker or lacking, as reported by Dercourt et al. (1986). On the other hand, the development of D I erosive surface described by Erol (1983) coincides with this period (Fig. 1).

Langhian folding period

The Miocene depositional period was interrupted probably during the Lower Langhian to north of lake Van (Gelati, 1975). The depositional period ended in the Muş basin (Akay et al., 1989), Hınıs basin (A.Yılmaz et al., 1988), Elazığ-Palu basin (Sirel et al., 1975; Sirel and Gündüz, 1975) at the end of Lower Miocene and in the Kahramanmaraş, basin (Uysal et al., 1985) in Middle Miocene. On the other hand, a considerable overthrusting is apparent over the Lower Langhian se: diments in the Beydağları basin (Akay et al., 1985). This phenomena coincides with the end of depositional period in Eastern Taurus. In the vicinity of Şarkışla—Gemerek, the Miocene continental depositional conditions ceased in the lower stage of Middle Miocene (Sümengenetal., 1987). ;

In the area between Bilican mountain—Muş—Bingöl, the volcanic rocks of the Elçiler formation, consisting generally of andesites were (Akay et al.,1989) formed after the Lower Miocene depositional period.

In the Hınıs basin, the Lower Miocene sediments forming E—W trending folds, are overlain by the Pliocene sediments with an angular unconformity (A.Yılmaz et al., 1988). The same position is seen in the vicinity of Çakırlı—Yünören within the Muş basin (Akay et al., 1989). The folds effecting the Lower Miocene sediments, strike NE—SW in the Palu basin. These sediments are overlain by the Pliocene sediments with an angular unconformity (Sirel et al., 1975).

In the Misis—Andırın basin, the rock units belonging to the basement are embedded in the Middle Miocene sediments as giant blocks (N.Turhan, 1988, oral communication).

Y.Yılmaz et al. (1988) claimed that a continuous compression from Oligocene—Early Miocene to the Middle Miocene gave rise to the closure of the Miocene basin in the vicinity of Kahramanmaraş. If it is so the regressive beds which formed in the northern nappe area must be older and the beds in the southern nappe area must be younger. However, the regressive beds in the northern nappe area are of at least Middle Miocene age. The regressive beds of the Miocene basin are of Middle Miocene age around Kahramanmaras.. On the other hand, the contact between the Lower Miocene Aslantaş formation and the Middle Miocene Karataş formation is conformable in the vicinity of Andırın This indicates that no considerable deformation took place in the region until the formation of the Middle Miocene regressive beds. Therefore, the effect of continental convergence rate which is assumed to be low in Eastern Taurus during the Early Miocene, must also be low or even much lower in this region. Therefore the closure of the Miocene basin around Kahramanmaraş was caused by the Langhian compressive regime whose traces can be observed everywhere in Anatolia, but not by a series of continuous events occured during the Early—Middle Miocene.

In conclusion, the effect of continental convergence rate which was very low or lacking during the

Early Miocene, considerably increased during the Langhian, and as a result of this, the lithosphere underwent a deformation, inter-mountain basins developed and the Arabian continent was forced to plunge underneath the northern continent. The andesitic volcanic rocks such as the Elçiler formation (Akay et al., 1989) resulted from these events. The northward overthrusting determined by Şengör et al. (1985) by means of seismic profile, cuts the Oligocene— Lower Miocene sediments in the southern part of Muş plain. This overthrust should have moved also during the Middle Miocene phase.

Additionaly. the fracture system which is oblique to N-S trending compressive forces and assumed to be ancestor of the East Anatolian Fault and thought to have moved probably during the Oligocene, should have also moved in this period.

Serravalian—Tortonian depositional period

The Langhian compressional period was followed by a relaxation period whose traces are observed in the neighbouring areas, but not in the region. As a result of this, the Middle Late Miocene marine transgression in the north of lake Van (Demirtaşlı and Pisoni, 1965), Tortonian marine transgression in the vicinity of the Adıyaman (Meriç, 1987), Tortonian continental-marine transgression in the Misis Andırın basin (N.Turhan, 1988, oral communication), Tortonian marine transgression in the Adana basin (Yalçın and Görür, 1984), Tortonian marine transgression in the southern part of iskenderun (Y.Yılmaz et al., 1988), continental transgression starting in the lowermost stage of Late Miocene in the Şarkışla-Gemerek basin (Sümengen et al., 1987) and Serravalian -Tortonian marine transgression in the Antalya basin (Akay et al., 1985) took place. This period represents a transition from the period of deposition in pre-existing, post-collisional marine eratonic basin to the period of deposition in inter-mountain basin. On the other hand, the formation of D II erosive surface described by Erol (1983) coincides with this period (Fig.l).

Uppermost Tortonian compressional period

Akay and Uysal (1987) suggested that the 11 km. long westward overthrust in the Antalya basin probably developed by the westward movement of the portion between the North Anatolian Fault and East Anatolian Fault, following the formation of these faults. For this reason the effect of continental convergence rate during the Uppermost Tortonian is maximum in Eastern Taurus. At the same time, the sediments in Adana (Yalçın and Görür, 1984), Antalya (Akay and Uysal, 1987), Misis—Andırın (N.Turhan, 1988, oral communication) marine cratonic basins and inter-mountain basins in Eastern Taurus were deformed. On the other hand, the fact that the Late Miocene sediments and Pliocene sediments show transition into each other in the Şarkışla—Gemerek basin (Sümengen et al.. 1987) indicates that the region was well-preserved as inter-mountain or extensional basin in this period. During the initial stage of the Uppermost Tortonian phase, the region and its surroundings were affected firstly by an approximately N—S trending compression and then the East Anatolian Fault moved again and the arc-like shape of the North Anatolian Fault formed in the late phase of compression.

There is an overthrust which developed during the Uppermost Pliocene in the northern margin of Muş plain (Soytürk, 1973; Akay et al., 1989). This overthrust probably moved during the Uppermost Tortonian, because the lower level of the Pliocene sediments underlying the overthrust consists completely of pebbles derived from the Oligocene formations and these pebbles indicate a transportation from north. Hence, if the present uplift in the north of Muş, plain formed during the Uppermost Pliocene compressional period, a probable overthrust might have formed an uplift and caused the formation of Muş basin as an inter-mountain basin originally during the Uppermost Tortonian. Therefore the clastic materials derived from this uplift were transported into the inter-mountain basin in the south. Afterwards, due to the decrease of compressional effect during the Pliocene relaxation period, the sediments derived from the Bitlis massif were deposited, which rest conformably on the underlying beds and overlapping this uplift. This type of other basins probably developed in Eastern Taurus. On the other hand, a buried fault in the northern part of Bitlis massif, which is assumed to have moved during the

Langhian compressional period might have moved essentially during this period.

Furthermore, the structures which formed during the Langhian were deformed again during this period.

Early—Late Pliocene depositional period

Related to the sudden decrease in the effect of continental convergence rate or reversal in some regions during the Uppermost Tortonian, andesitic—basaltic volcanic rocks of the Solhan formation and terrigenous sediments of the Zırnak formation were deposited together in subsided areas (Fig. 1). These formations show lateral transition into each other in the Muş Pliocene basin (Akay et al., 1989). The volcanic rocks of the Solhan formation which has been considered to be of Late Miocene age at first and the volcanic rocks of the Zırnak formation which is assumed to have formed in an immediate period following the Late Miocene (Y. Yılmaz et al., 1987) are interpreted to be contemporaneous according to Early—Late Pliocene ages obtained from the sediments showing lateral transitions into both of them. Early and Late Pliocene ages have been obtained in the Hınıs (A.Yılmaz et al., 1988) - Muş (Akay et al., 1989) basin.

Y.Yılmaz et al. (1987) reported that the volcanic rocks of the Solhan formation are the initial products of the neomagmatic period and later the volcanic rocks of the Zırnak formation an Sergen rhyolite developed, respectively. Since the volcanic rocks of the Solhan and Zırnak formations are contemporaneous and the rhyolitic magmatism of the Sergen formation is of Oligocene age, the neomagmatic period needs to be rewieved.

In the Muş—Hınıs basin, no deformation wok place during the depositional period ranging from the lowermost stage of Lower Pliocene to the uppermost stage of Upper Pliocene. This situation disagrees with the hypothesis of Barka (1987) that the movement of the North Anatolian Fault is maximum during the Early Pliocene. On the contrary, it is likely -that the fault might have moved reversely in a short distance, as a result of the removal or extreme lessening of the force which caused the formation of the fault.

On the other hand, the marine transgression belonging to the same period in the Adana (Yalçın and Görür, 1984) and the Antalya basins (Akay et al., 1985) supports a regional event which occured during the relaxation.

Atalay (1983) mentioned about three different erosive surfaces which formed in different levels, in Bitlis mountains to the south of Muş plain. These erosive surfaces one of which is at the altitude of 2250 m. formed during the Oligo—Miocene and the other is at the altitudes of 2000-2050 m formed during the Mio-Pliocene are shown as a single erosive surface in the geomorphological map. However, the Pliocene Solhan formation constitutes the Kozmadağı mountain at the altitude of 2100 m. located to the south of the western margin of Muş plain (Akay et al., 1989). This altitute approximately corresponds to the above mentioned two erosive surfaces. Thus, both erosive surfaces developed during the Pliocene (Fig. 1). The correlan deposits on this surface belong to the Pliocene basin (Fig. 1). On the other hand, the formation of I) 111 erosive surface described by Erol (1983) coincides with this period.

Uppermost Pliocene compressional period

The traces of this period can be observed in the basins located to the north of Bitlis mountains. The approximately SE—NW trending overthrusts and folds developed in the Muş basin (Akay et al., 1989). At first, the Muş plain and its surrounding formed as an inter-mountain basin as a result of compression during the Uppermost Tortonian. Afterwards the uplifts in the north of the inter-mountain basin subsided due to the subsidence during the Pliocene depositional period and took part in to the depositional basin. The inter-mountain basin was formed again by the effect overthrusting in the north of the plain during the Uppermost Pliocene (Akay et al., 1989). The Pliocene sediments dip northward at an angle of 10° in the westernmost part the southern margin of the plain. The dips of these strata increase eastwards. On the other hand, a buried fault extending along the southern margin of the plain, which is assumed to have moved slightly during the Langhian compressional period and considerably during the Uppermost Tortonian, might have been thrusted over northwards without affecting the strata in the western part of the plain during this period. Furthermore, in the Hınıs basin to the north of the region, folding axes in the eastern part of Zırnak strike in NE—SW direction, where those in the western part of Zırnak extend in NW—SE direction (K. Sulu, 1988, oral communication).

The volcanoes such as Nemrut, Süphan, Tendürek and Ağrı which are assumed to have erupted along the N—S trending fractures produced by the N—S trending compressive forces (Şaroğlu et al., 1980; Güner, 1984) are seen in a NE—SW trending line. This line represents a fracture zone which developed probably during the Uppermost Pliocene and which is responsible for the formation and of these volcanoes.

On the other hand, Şaroğlu and Yılmaz (1987) further reported that the East Anatolian Fault and North Anatolian Fault, cutting this formation were formed during the Late Pliocene time on the basis of Early Pliocene age obtained from the Zırnak formation. The Middle—Late Pliocene age was obtained from the regressive strata of the Muş Pliocene basin (Akay et al., 1989). In the same way, the Middle of Late Pliocene age was obtained from the regressive strata of the Pliocene sediments in the Antalya basin (Akay et al., 1985). Hence, this orogenic phase is synchronous everywhere in Anatolia and has been occured during the Uppermost Pliocene.

In the Antalya basin, there is a 11 km.long westward overthrust which developed during the Tortonian phase. However, the Pliocene strata, dipping at an angle of 40° maximum, generally dip 5°—10° (Akay and Uysal, 1985). If the Uppermost Tortonian and Uppermost Pliocene phases gave rise to deformations in the same way (it must be so), the first phase was 11 times more effective than the second phase.

As is seen on a 1:60,000 scale aerial photograph, at the west of Bilican mountain there are 4 different terraces whose altitudes varying from 1550 m. to 1700m. The present dip of slope increases more and more above 1700 m. These terraces developed during the regression of the Pliocene lake in Eastern Anatolia due to the Uppermost Pliocene deformation. On the other hand, the base of the Pliocene sediments are seen at the altitude of 1950 m.further to the north of Muş plain. When the thickness of sediments is taken into consideration, the phenomenon indicates that uplifts of at least 500 to 600 m.occured in the region.

The present morphology and drainage affecting the morphology started to form after the Uppermost Pliocene compressional period in Eastern Taurus and its surroundings, even all over Anatolia.

On the other hand, the erosive surfaces which developed during the Lowei—Upper Pliocene depositional period have been carved as a result of the uplifting of the region.

Early-Middle Pleistocene depositional period

Following the Uppermost Pliocene compressional period, a relaxation period during which the effect of continental convergence rate decreased, took place in the Early—Middle Pleistocene. This period is less important than the Early—Late Pliocene relaxation period. As a result of this the further subsidence of intermountain basins which developed during the Uppermost Pliocene compressional period led to the deposition of sedimentary rocks of the Bulanık formation (Fig.l). Late Pleistocene age was obtained from the upper level of these sediments in Muş plain (Akay et al., 1989). The coarse grained sediments were deposited in the margins of the basins, whereas the fine grained sediments were deposited in the central parts of the basins. These sediments are 200 m. thick maximum and consist generally of sandstone, siltstone and claystone. The sandstones are different from the sediments belonging to upper and lower depositional periods, because they contain abundant white small fragments of lamellibranchiata and are cross-bedded. These sediments are observed also around Hınıs—Tutak—Ağrı (K. Sulu, 1988, oral communication).

The NE—SW trending fracture which had developed during the Uppermost Pliocene and activated Nemrut, Süphan, Tendürek and Ağrı volcanoes, partially moved during this period and gave rise to the development of volcano-sedimentary deposits around the volcanoes.

Erol (1983) reported that the D IV erosive surface formed during the Early Pleistocene. The erosive surfaces at the altitude of 1500 - 1800 m in the north and at the altitude of 1750 - 1800 m.in the south of Muş plain (Atalay, 1983) formed during the Early-Middle Pleistocene (Fig.l). The erosive surface in the north the plain is sometimes covered by ignimbrite—tuffite, being final products of the Nemrut volcano (Özpeker, 1973). On the other hand, the same ignimbrites are found on the floor of the Bitlis valley. This valley developed prior to the formation of ignimbrites which formed in the first stage of the Late Pleistocene (Akay et al, 1989).

Maxon (1936) considered the Bitlis valley as the extension of the ancient river Murat. He suggested that the Bitlis valley and its surroundings were filled up by the volcanic rocks of Nemrut and as a result of this lake Van developed. Özpeker (1973), Şaroğlu and Güner (1981), Güner (1984) supported this hypothesis. However, the floor of the Bitlis valley, which is made up of metamorphic rocks is at the altitude of at least 1700 m. The deepest site of the Van basin which didn't contain any water 100,000 years ago (Wong and Finslch, 1978) is at the altitude of 1200 m. The meandering river sediments of the Middle Pleistocene are at the altitude of 1250 -1300 m.in the Muş basin. If it is supposed that the compressive regime lasting from the Late Pleistocene to the present day did not disturb the settings of the rocks in this area, it is unlikely that the Middle Pleistocene meandering river and the other rivers reaching lake Van which did not contain any water at this time, drained by Bitlis creek. Consequently, the Bitlis creek should be the present extension of a river which was active prior to the Early—Middle Pleistocene depositional period and probably during the Early—Late Pliocene.

The Muş plain and basin of lake Van are inter-mountain basins which were formed and disturbed by the Uppermost Tortonian tectonic regime first and than formed again by the Uppermost Pliocene compressive regime, but not by Pleistocene tectonic regime as reported by Şaroğlu and Güner (1981). These two basins were joint during the initial volcanism of Nemrut. Afterwards, these basins were seperated by the Nemrut volcanic rocks.

The subsidence-deposition during the Early-Middle Pleistocene indicates that the effect of continental convergence rate was slightly low. Meanwhile, it is probable that the movements of North Anatolian Fault and East Anatolian Fault decreased, stopped or were reversed.

Late Pleistocene present day compressional period

No remarkable deformation is observed in the Lower—Middle Pleistocene sediments. The strata dip at an angle of 3°-5° or horizontal over an area extending from Muş to Ağrı However, in some places where the strata lean against the mountains, angles of 15°-25° can be measured (K. Sulu, 1988, oral communication). On the other hand, the Değirmen stream draining into the Muş plain carved its bed for 70 m. following the depositional period in which the Bulanık formation developed (Akay et al., 1989). The bottom of tuffites on the erosive surface which developed during the Early-Middle Pleistocene, on the sides of Norkavak and Çaksor streams draining into the Muş plain was carved for 200 m. maximum by the streams. The similar examples are seen in northern edge of the Bitlis massif. On the other hand, the terrace systems described by Erol (1983) probably developed in this period.

Consequently, the effect of continental convergence rate increased again following the partial relaxation during the Early—Middle Pleistocene. As a result of this, the region was slightly uplifted but no effective deformation took place yet. All the active faults in Eastern Taurus were activated in this period. Additionaly. the sediments such as the Muşovası formation (Akay et al., 1989), were deposited as inter-mountain basin sediments in the region; following the slight erosion related to the uplifting. On the other hand, the recent eruptions of Nemrut and the other volcanoes, which had started to erupt in the Uppermost Pliocene along the NE—SW trending fractures are controlled by the N—S trending fractures formed as a result of the N—S trending compression which is still effective, as reported by Şaroğlu et al. (1980).

The age of the water of lake Van was determined to be 60,000 years on the basis of major element content but argued to be of 100,000 years (Wong and Finckh, 1978). On the other hand, the deepest site of the lake is at the altitude of 1200 m. If it is supposed that the Bitlis creek whose highest site is at the altitude of 1700 m.on the metamorphic rocks was not subjected to any change in structural setting, it is unlikely that this creek was drained by the rivers reaching lake Van which did not contain any water during the Upper Pleistocene. As discussed in the Early—Middle Pleistocene depositional period, it is also impossible that the Bitlis creek was drained by the rivers in Muş plain in this period. On the other hand, the fact that no terraces resembling to those of on the side of lake Van are found in the margin of Muş plain, indicates that the Van-Muş basins were not joint in this period.

The Muşovası formation was carved (Akay et al., 1989) as a consequence of the final lowering of sea-level and partly by the effect of tectonics, but later, these troughs are filled up by the Holocene (actual) sediments related to the rising of sea-level again.

CONCLUSIONS

Şengör et al. (1985) and Dewey et al. (1986) report that the Eastern Anatolia has been undergoing a shortening of the continental lithosphere as a result of the continent-continent collision since the Late Serravalian. However, stratigraphic data obtained by Gelati (1975) indicate that the compressive regime probably lias been taken place since the Langhian. On the other hand, Michard et al. (1985) claims that the continent-continent collision took place during the Late Eocene. The facts that the Oligocene basin did not develope widespread, the geometry of sediments belonging to this basin frequently changes due to the lithospheric deformation, very limited acidic volcanism took place in the final phases of the basin and syncronous occurrence of the Early Miocene transgression in the both southern and northern parts of Southeast Anatolian suture zone (Baştuğ, 1980; Meriç., 1987) supports the Late Eocene collision. Therefore, the neotectonic period started in the Late Eocene. The shortening of continental lithosphere has been continuing since Late Eocene. In the region, the changes in the fate of continental convergence gave rise to the formation of depositional basins and chains of uplifts. The relative rate of continental convergence can be deduced depending on whether these formations developed widespread or not. This rate was maximum during the Uppermost Tortonian, and lower during Late Langhian, Uppermost Pliocene, Early—Middle Pleistocene, Serravalian—Tortonian, Early—Late Pliocene, Oligocene, Early Miocene in a decreasing order. The rate of continental convergence must be considerably high from Late Pleistocene to the present-day.

The frequent occurrence of basin uplifting indicates that the continental crust has been progressively thickening since Langhian.

On the other hand, the neomagmatic period which is assumed to have started in the Late Miocene by Y.Yılmaz et al. (1987) has been continuing since the Late Oligocene volcanism.

Although Şaroğlu and Güner (1981) dated the Late Miocene lithologies and Pleistocene lithologies in the Muş basin without any fossil evidence, they reported that the Muş inter-mountain basin gained its characteristics in the Pleistocene in one place and at the end of Pliocene in another place. However, the Muş inter-mountain basin firstly developed during the Uppermost Tortonian compressional period and was disturbed during the Early—Late Pliocene depositional period and gained its present characteristics during the Uppermost Pliocene compressional period. It underwent no remarkable structural change in the Quaternary.

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TECTONIC ZONES OF THE CAUCASUS AND THEIR CONTINUATIONS IN THE NORTH-EASTERN OF TURKEY : A CORRELATION

Ali YILMAZ*

ABSTRACT.— The study area covers the Caucasus and the northeastern Turkey. Tectonic zones of the Caucasus and their continuations in the northeastern Turkey, and also the relationships, lateral variations, similarities and differences of the both district, will be presented. On the basis of the main geologic characteristics, the rock units of the Caucasus are divided into the tectonic zones. Each zone has pre-Liassic, Liassic and post Liassic units reflecting different geotectonic environment. The northern part named as the Great Caucasus, the southern part as the Lesser Caucasus, median part as the Transcaucasus of the Caucasus was bordered by the Scythian platform to the north and by the Iranian platform to the south. The tectonic zones of the Great Caucasus lie from north to south are presented below: the Laba-Malka zone (the Bechasinian Subzone and the Forerange Subzone), the Main Range zone, the Southern Slope zone. The Gagra-DJava zone and the Drizula massif and its covers are situated to the north, the Somcheti-Kafan (Karabakh) zone to the south and the Adjara-Trialetian and Talysh zones which are the continuation of each other are between the zones of the Transcaucasus. The ophiolitic belt (the Sevan-Akeran Ophiolitic zone to the north, the Vedi Ophiolitic zone to the south) of the Lesser Caucasus and the Miskhan-Zangezur zone and the Araks zone of the northern part of Iranian platform have been differentiated. The Caucasus tectonic zones are bordered by the overthurst planes dipping 70-80 degrees to the north. The Oligocene-Recent molasse showing enormous lateral and vertical facial changes, sits upon the rocks of the tectonic zone conformably or unconformably, in places. The results, presented below, can be obtained by the correlation of the Caucasus tectonic zones and tectonic zones of northeastern Turkey: 1- The tectonic zones of the Great Caucasus and northern part of the Transcaucasus can not be followed in the northeastern Turkey. 2- The Adjara-Trialetian zone continues along the Black Sea Shores. 3- The Somcheti-Kafan (Karabakh) zone, which is southern part of the Transcaucasus corresponds to the Pontian zone. But, considerable differences on the basis of the stratigraphic sequence and facial changes are observed in the both sides of the zone. 4- The Lesser Caucasus ophiolitic belt corresponds to the North Anatolian ophiolitic belt. There are two subzones, one of them is to the north and another is to the south, showing similarities in both side of the belt. 5- Iranian platform of the Lesser Caucasus corresponds to the Taurus platform, in general. Pre-Liassic Miskhan-Zangezur zone of Iranian platform corresponds to the Central and East Anatolian massifs the Araks zone to the Taurus zone respectively. If the corresponding tectonic zones of the Caucasus and northeastern Turkey are correlated, considerable facial changes as well as the similarities are observed. A lot of the differences result in the lateral and vertical changes of the zones.

PETROLOGICAL INVESTIGATION OF LOWER TERTIARY AGED DETRITAL SEQUENCE AROUND BURDUR

Emel BAYHAN*

ABSTRACT.— In the study area, starting with Triassic-Jurassic series and including Upper Pliocene-Quaternary series as well, Lower Tertiary series show the characteristics of turbidite fades. It has been determined through quantitative analyses of light, heavy and clay minerals of this detrital sequence, that the sandstones in the region are of moderately and poorly sorted greywacke characters, and that the principal constituents consist of mono and polycrystalline quarty, plagioclase, igneous and metamorphic rock fragments. The most abundant group of heavy minerals is the group of amphiboles. Pyroxene, epidote, garnet and mica, apatite, zircon and tourmaline are found in lesser amounts. Smectite is the most important clay mineral thin the clay fraction in the region. Apart from dioctahedral smectite, illite and chlorite occur sparsely. Under the light of the present information, it is seen that the detrital material in the region is derived principally from a source consisting of igneous and metamorphic rocks.

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SEDIMENTARY PETROGRAPHY AND ORIGIN OF PHOSPHATE PELOIDS OF THE MAZIDAĞ-DERİK AREA (MARDİN, SOUTHEAST TURKEY)

Baki VAROL*

ABSTRACT.— The Mazıdağ phosphate beds of the Upper Cretaceous carbonate sequence were deposited in an area of upwelling water which supplied phosphorous was deposited on the sea bottom as biogenic detritus of zoo— and phytoplankton, fish bones and scales. The biogenic accumulation with high content of organic matter and phosphorous was a prime source of the phosphate which formed the phosphate deposits of the Mazıdağ—Derik area. Commonly, currents transported the phosphatic sediments laterally, forming parallel and low angle cross laminations. In some cases, the sea floor was uplifted after upwelling periods. Consequently, some phosphate beds underwent karstification, giving rise to phoscretes and silcretes. Ground water percoalating into the karstic realm caused phosphate replacement of many carbonate grains of the shallow water limestone facies. These complex events resulted in the formation of three basic phosphate peloid types in the Mazıdağ-Derik phosphate sequence, which are classified to the following origin of peloids : (1) in situ precipitation; (2) abrasion of phosphatized intraclast and bioclast and (3) phosphate micritization of fish bones. The phosphate peloids show the following different microstructures under the electron microscopy : (1) amorphous; (2) microglobular; (3) semi-crystallized; (4) microcrystallized and (5) cementing.

INTRODUCTION

The phosphate assemblages of the southeast Turkey (Fig. 1) occurred in the Upper Cretaceous carbonate sequence, which have the different depositional and textural characteristics. Previous studies have described the geochemistry, general petrology, and genesis of the Mazıdağ—Derik phosphate beds and reached general agreement about the origin of the phosphates related to upwelling waters (Sheldon, 1964, Köksoy, 1977; Lucas et al., 1979). In fact, upwelling is the most reasonable mechanism for the

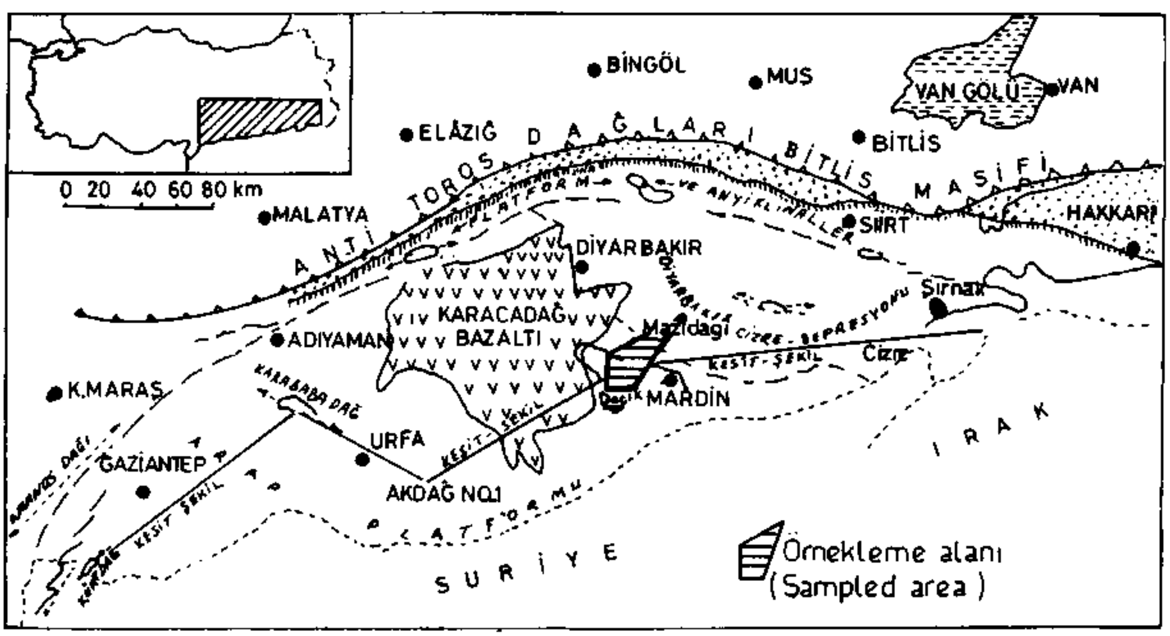

Fig. 1 — Location map of the Mardin (Mazıdağ—Derik) phosphate deposits (taken from Köksoy, 1977).

triggering of the phosphate precipitation, because cold deep sea waters can carry much dissolved phosphate and rich nutrients to the shelf zone (Kazakov, 1939; Baturin, 1982). Therefore, this hydrological regime has been applied to the formation of many ancient and modern sedimentary phosphate occurrenceses of the world (Arthur and Jenkyns, 1981; Baturin, 1982; Slansky, 1986). The model has been discussed in detail in the literature, and therefore is most treated further in this paper.

In this study, detailed sedimentary petrographic and electron microscopy examination of the phosphate rocks are presented and the genesis of the different types of the phosphate peloids within the beds is discussed. The samples bearing the phosphate peloids were collected from three sections (the Taşıt, Kasrık and Şemikan sections) in the Mazıdağ—Derik area, southwest Turkey (Fig.2).

The Upper Cretaceous sequence is basically diveded into three major lithological groups, carbonate, siliceous and phosphate rocks. The carbonates mainly consist of foraminiferal—algal mudstones, wackestones and dolostones. In some beds, biogenie packstone are present, occurring as brachiopod shell "lumachelle limestone" and whole rudist bivalves patch reef. The lithological associations indicate deep —water to shallow— water to supratidal carbonate environments.

The phosphate peloids were laid down in the limestone facies which show different depositional condition, both shallow and deep water. Likewise, the type of the peloids and associated grains reflect these major differences in depositional environment. For

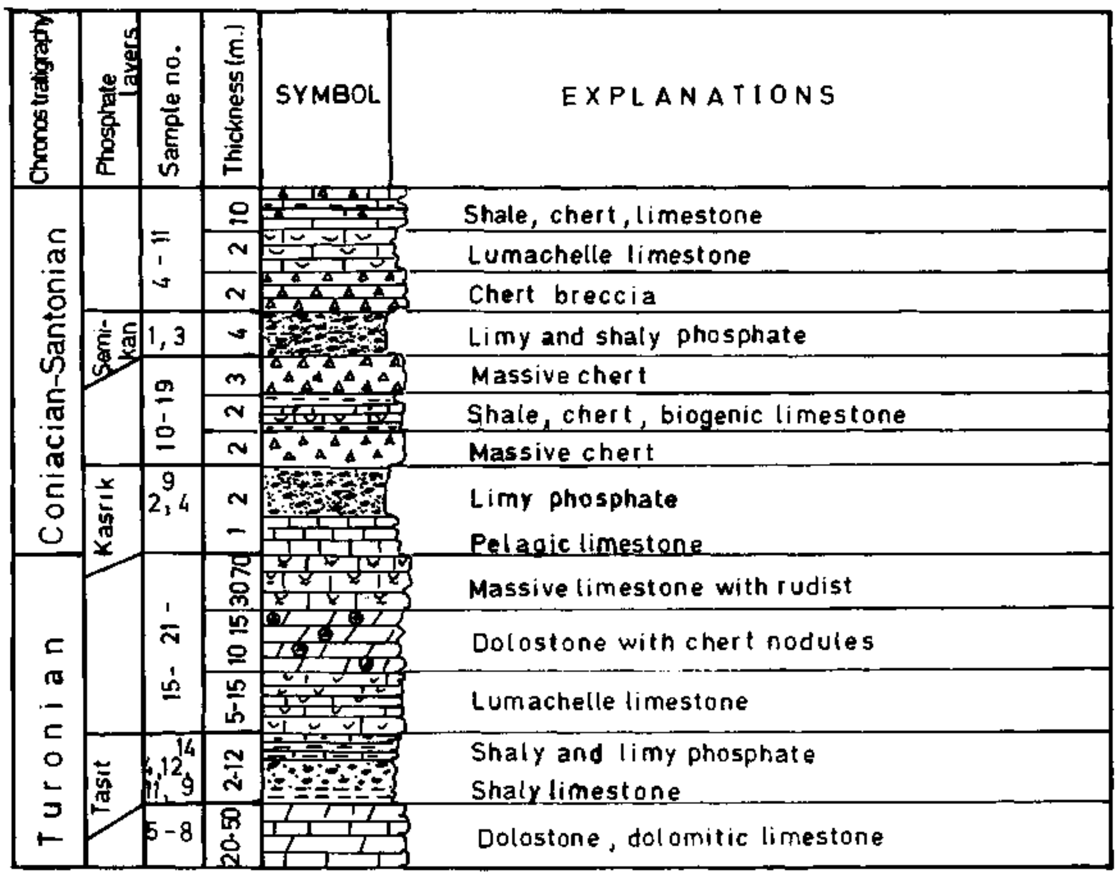

Fig. 2 — Generalized columnar section of the Mazıdağ—Derik phosphate layers.

instance, pseudoolitic phosphate grains, which were assosiated with the phosphate peloids in the some beds, were only restricted to deep—water limestones which exhibit parallel and cross laminations, whereas uniform phosphate peloids with regular internal structure are only found in the shallow water carbonates, along with the phosphatized biogenic grains. Fish bones, teeth, spines, and phosphatized intraclasts are commonly in every phosphate beds, showing different environmental habitats.

Francolite (carbonate-fluorapatite) is the predominant phosphate minerals in the examined peloid samples. Additionally, some intraclasts derived from nonpelloidal phoscrete layers contain collophane, and also the peloids including bone remnants are rich in dahlite.

The phosphate peloids are the dominant component of the phosphate—bearing layers, and they are discussed in detail below.

PHOSPHATE PELOIDS

In this study, all spherical and oval phosphate grains ranging from 0.15 to 0.50 mm in diameter are considered to be phosphate peloids (McKee and Gutschick, 1969) regardless of their origin and internal structures. Grains larger than 0.50 mm in size are termed phosphate ovoids if rounded, and phosphate intraclasts if angular.

The phosphate peloids usually appear isotropic under the polarizing light microscope, because they are composed of very fine crystals, showing the usual character of marine phosphates (Baturin, 1982). Some peloids also may exhibit a weak anisotropy. The microinternal structures of the different types of the phosphate peloids were examined by scanning electron microscopy (SEM), under 1000—3000 magnification. Additionally, phosphate grains were scanned with an EDAX system in order to reveal the phosphate distribution.

The combination of the normal and electron microscopy examinations has led to defmation of the following types of phosphate peloids.

In situ precipitated peloids (zonal and nucleate or pseudoofitic)

This type of peloids refers to in situ phosphates precipitated from phosphate—rich solutions. They are diveded into two types, zonal and nucleate peloids, based on texture. The zonal peloids are between 0.15—0.20 mm in size, and have central parts which are blurry and dark colored (mostly brown). They include numerous non—phosphatized inclusions, such as iron oxide pigments, organic matter and carbonate mud. In rare case, the dark colored material covers the entire surface of the peloids. Marginal part of the peloids are transparent and more clear, and are mostly yellowish in color because dark inclusions are absent.

Nucleate peloids have a nuclei in the centers and transparent rims. Skeletal calcitic grains, minute fish bones and teeth formed peloid nuclei, and the peloids themselves served centers for the growth of the transparent phosphate rims. The internal structure is roughly similar to the concentric laminae of ooid grains; however, true concentric accretion laminae are absent and is interpreted as a phosphate pseudoolite.

The zonal and pseudoolitic assosiations are mostly found in a pelagic mud matrix showing parallel and cross laminations, and well—rounded phosphate intraclasts. The intraclasts were derived from the phosphatized mud, and suggest that bottom erosion succeeded the phosphate formation, induced by currents. The current activity concurrently caused the mixing of the different type of the phosphate grains in the same level, which lead to deposition of biopelphosphoarenite (Slansky, 1986) in the Derik-Mazıdağ phosphate beds. The rock type (Plate I, fig.l) is typical in the whole sequence of Kasrık phosphate and below levels of the Şemikan phosphate.

The zonal phosphate peloids present an inhomogeneous phosphate distribution, as seen well with EDAX scanning (Plate I, fig. 3,4). The EDAX scanning revealed that the dark and turbid central portions of the peloids have less phosphate than clear peripheral zone, and high organic matter and iron content. In the electron microscope, organic mud can be easily distinguished as its surface was covered by micro-swelling, fistules, and nodules, whereas the clear and transparant peripheral zone consisted of well—pocked micronodules without organic mud. According to the electron microscope data, the formation of the zonal peloid took place in two main stages, as follows.

In the initial stage, the phosphatization began as microcenters or swelling, which made up an uneven and bumpy surface on the organic mud matrix (Plate I, fig. 5). In the second stage, the bumpy surfaces were completely turned into micronodules (Plate I, fig. 6,7). The different degree of the phosphatization stages affected the colored character of the zonal peloids. For instance, where the volume of the organic mud was very high, the peloid is blurry and dark. To the contrary, the high content of phosphate gives rise to the transparent character of the peloid which is devoid of organic mud. In addition, non—nodular and/or amorphous and semi—crystallized (fusiform) phosphates have been observed in the electron microscope. The amorphous type is a homogenous mass which shows no sign of granulation and crystallization. However, in rare case, some ultra—micronodulations are present on homogenous phosphate masses (Plate I, fig.8). The semi—crystallized type mostly shows rosettes with satellite appearance (Plate II, fig.2). The fusiform habit has been attributed to an intermediate stage of crystallization between amorphous and semi—crystallized (Baturin, 1982). In the examined samples, these types were mostly found on the transparent surfaces of the peloids.

Pseudoolitic phosphate grains include both semi-crystallized and micronodular varieties. The micronodules are commonly scattered as free grains on the semijcrystallized phosphate background (Plate II, fig-1)- This microtexture suggests that some mobile phosphate solutions were developed during the formation of the pseudoolitic phosphate grains. The mobile phosphate solutions may have evolved at the Sediment/sea water interface. The abundance of the pseudoolitic grains in the laminated (cross and parallel) biopelphosphoarenite indicate the depositional environment was influenced by current activity, and led to physical mobilization of the phosphate gel around some nuclei of the pseudoolites.

Peloids derived from intraclasts (non-internal structure)

These are formed as a result of physical abrasion of the undurated or semiplastic phosphate intraclasts. The rounded grains can be easily differentiated from in situ precipiated peloids by lack of internal structure. Generally, they are yellow or brown in color. Grain-size is always coarser than the other type of the phosphate peloids, ranging from 0.20 mm to 0.60 mm. Oval shapes are predominant and very diagnostic for the interpretation of this type. Similar forms have been described as egg shaped—phosphate grains "ovules" by the study of Cook (1972). In the Mazıdağ-Derik phosphate sequences, is type of peloid was largely accumulated in the biopelphosphoarenites which were deposited by currents. Consequently, they were more rounded than the ones in the shallow—water carbonate facies. On the other hand, the phosphate peloids derived from intraclasts show different color properties, probably depending on the environmental conditions. The types with light colors (generally yellow) are found within the deep—water limestone, and have a high phosphate content. Shallow water types are darker (brownish) and less phosphatized. The latter types are commonly associated with phosphate intraclasts derived from phoscrete layers. Their internal structures are very similar to each other, which imply genetic relationship between the phosphate intraclasts and peloids in some shallow water limestones.

In the electron microscope examinations, the microinternal structures of this group are characterized by an organic mud which was subjected to different degrees of phosphate replacement. The samples belonging to deep sea facies were made up of diatom mud.

Generally, the microholes of the diatom tests can be discernible within the phosphatized mud. In some cases, the microholes were filled with a cement of microapatite crystals (Plate II, fig.3). Samples taken from shallow water limestone do not contain any trace of diatom valves, and they are mostly recognized by microcrystallized (tabular or elongated) apatites, replacing the micrite matrix (Plate II, fig. 4).

Peloids derived from bioclasts

This type of the phosphate peloids usually makes up non—economic phosphate beds and consequently it has not been previously studied in detail. The content of phosphate is low. However, they exhibit some interesting diagenetic properties, which help to define the phosphatization conditions and processes within the shallow water limestones.

The peloids are often angular and coarse—grained (0.25—0.75 mm). In most cases, traces of shell structures are retained within the peloids. The original shell structures can be easily observed by light microscope examination with high magnifications in case of low-grade phosphatization. This suggests that disaggregated shell fragments in the intertidal—subtidal limestones underwent phosphate replacement. Commonly brachiopod shells and outer walls of gastropods were favored the phosphate replacement.

Neomorphic sparite (pseudosparite) is always associated with this kind of phosphate peloids. The association implies diagenetic effects in the fresh water phreatic zone during phospatization (Longman, 1980). Phosphatization through ground water, which is rich in dissolved phosphate, has been noted by Shalkowitz (1972). The worker considered that phosphate bearing ground water must be a prime agency for the phosphate replacement of the biogenic grains in the shallow-water limestone facies.

Peloids formed by phospho—micritization of fish bones (with regular internal structure)

This type of the phosphate peloids may be present within some level of the Mazıdağ—Derik phosphates, but they have been commonly developed as the massive phosphate deposits in the Şemikan phosphate layers. They are dark in color, and show a gradualy lightening towards the exterior parts of the peloid grains (Plate I, fig.2). Their shape, petrographic character, and host rock texture are entirely different from other types. Grain—size distribution of the peloids is uniform (averaging 0.15 mm) and sorting is very good. The rock is composed mostly of pure phosphate peloids, and the other usual constituents such as pseudoolite, bioclast, ihtraclast and various carbonate grains are either diminished or completely absent in these layers. Compaction is very weak, and also the ratio of cementing is very low. Overall, phosphate peloids are bounded by either grain contact or a thin calcite cement. Porosity is very high, up to 20—30 percent, and consequently the rock is friable, and disaggregated into sandy. In the light microscope, the peloids are distinguished by a regular internal structure, mostly including a micritic central zone and a light yellow a peripheral zone which shows weak anisotropy and some remnants of the fish bones.

In the electron microscope, the phosphate peloids exhibit microrelief within the phosphatized mass. These structures are very similar to elongated micro-channels of the fish bones (Marshall and Cook, 1980). The probable channels were covered with microphosphate granules or apatite crystals grown perpendicularly on the wall of the channel (Plate II, fig.6,7,8). This structural character implies that fish bones were phosphatized, creating the phosphate peloids. Similar textures have been observed by Cook (1980), and the processes have been attributed to phosphomicritization of the fish bones, induced by microalgae. Indeed, in the examined samples, in addition to the channel—like microstructures, the micritic appereance and weak anisotropy resulting from partial phosphatization of the fish bones supports the hypothesis that the fish bones were converted to the phosphate peloids. Seyhan et al. (1973) also reported phosphate grains with bone-brecciated texture in the Şemikan phosphate layers.

GENESISOFPELOIDS

In the Mazıdağ—Derik phosphate beds, in situ zonal peloids always occur in organic mud which is rich in zoo—phytoplankton and fish remnants. This indicates that there is a causal relationship between in situ peloids and the organic mud. Indeed, many ancient and phosphate beds were laid down in diatom mud, and phosphate peloids were diagenetically formed within the organic mud, instead of by direct precipitation from sea water (Baturin, 1971; Baturin and Bezrukov, 1979; Balson, 1980). A few samples are interpreted as having directly precipiated from sea water as oolithic phosphate beds (Arthur and Jenkyns, 1981). According to Brice and Calvert (1978), in the actively upwelling zones of the continental margin of southwest Africa, Peru, and Chile interstital waters include dissolved phosphate much more than ten times the concentration in the surrounding sea water. Likewise, Baturin (1982) reported that the initial phosphate formation began within organic mud when the phosphate concentration reached 5 to 9 mgl^{-1} in the interstital water. At this stage, all of the non—phosphatized remnants, such as very minute fish bones, diatom tests and other carbonate particles were precipitated, along with phosphate. The mixing material gave rise to form a dirty, dark and blurry central zone of the phosphate peloids. At the later stage, pure phosphate solutions cleaned from non—phosphatized remnants led to precipitation of the transparent exterior part of the peloids. The different stages and the different solutions during the formation of phosphate peloids should be effective to evolve the in situ zonal peloid of the Mazıdağ-Derik phosphate beds. Indeed, examined samples of this type of peloids generally contain an inner zonation. The central portions contain organic mud and less phosphate (bumpy surface); this suggests that the initial phosphate growth started from these organic mud surface. Then, the granulation up to full micronodules created the clear and transparent exterior zones of the peloids; this processes was probably enhanced by current movements which led to roll of peloids on the sea bottom. The presence of free micronodules on the pseudoolite surfaces support the conclusion that a mobile phosphate solution caused the growth of the in situ peloids. Hence, sedimentological and hydrological conditions such as current activity should be an important controlling factor for the evolution of the phosphate peloids, along with geochemical equilibrium of the interstital water. On the other hand, Bromley (1967) and Shalkowitz (1973) have reported that bacterial oxidation served as the prime agency for the release of phosphate from organic carbon, during the uplift of the sea water at the interface of the sediment and sea water.

That the sea level fluctuated greatly during deposition of the phosphate succession of the Mazıdağ-Derik is cleary indicated by the alternations of the deep —and shallow water—phosphate and carbonate facies. In this light of this data, in situ peloids have probably been formed in the following steps : (1) Deposition of an organic mud, containing abundant zooplankton and phytoplankton, during period of upwelling water; (2) Starting of the initial granulation on the surface of organic mud, in slightly reducing conditions; here, diatom tests might have served as nuclei for the triggering of the phosphate precipitation and/or initial nodulations; (3) Uplifting of the sea bottom, with current activity enhancing the growth of the in situ peloids under weak oxiding conditions, in which the phosphate is released from organic matter much faster than in reducing condition. At this latter stage, some phosphate solutions could be mobile on the sea bottom, induced by currents. The physical activity might be a prime agency for the formation of the phosphate pseudoolite material around the nucleis as well as the growth of the in situ zonal peloids.

The phosphate peloids derived from the intraclasts resulted from the rolling of both soft intraclasts and hard intraclasts. The soft material was probably turned upypartly lithified, phosphatized mud. Current activity favored the formation of these types, as cross and parallel laminations were associated with the layers which had the soft intraclasts and the relavent phosphate peloid. Hard intraclasts are mostly found in the shallow water limestone or near the karstic phosphate beds, suggesting they are derived from the phoscrete surfaces. In fact, the internal structures of these phosphate peloids are very similar to those of the phoscrete samples. On the other hand, some burrows on these grains can be attributed to biogenic erosion, which likely aided in rounding the hard phosphate fragments.

The peloids derived from bioclasts have been only observed in the shallow water limestones. This situation suggests that a part of the phosphatization process could have occured in the coastal area. As a matter of fact, Ames (1959) experimentally calculated that even a very low phosphate concentrations (0.1 ppm PO_4^{-3}) in solutions can achieve the replacement of a calcitic shell. Likewise, Cook (1982) has proved that the tidal—flat limestones underwent phosphatization through replacement by ground water percoalating from nearby phosphate karst. Karstic surfaces, which are common in the Mazıdağ—Derik phosphate sequence, could have been a major source

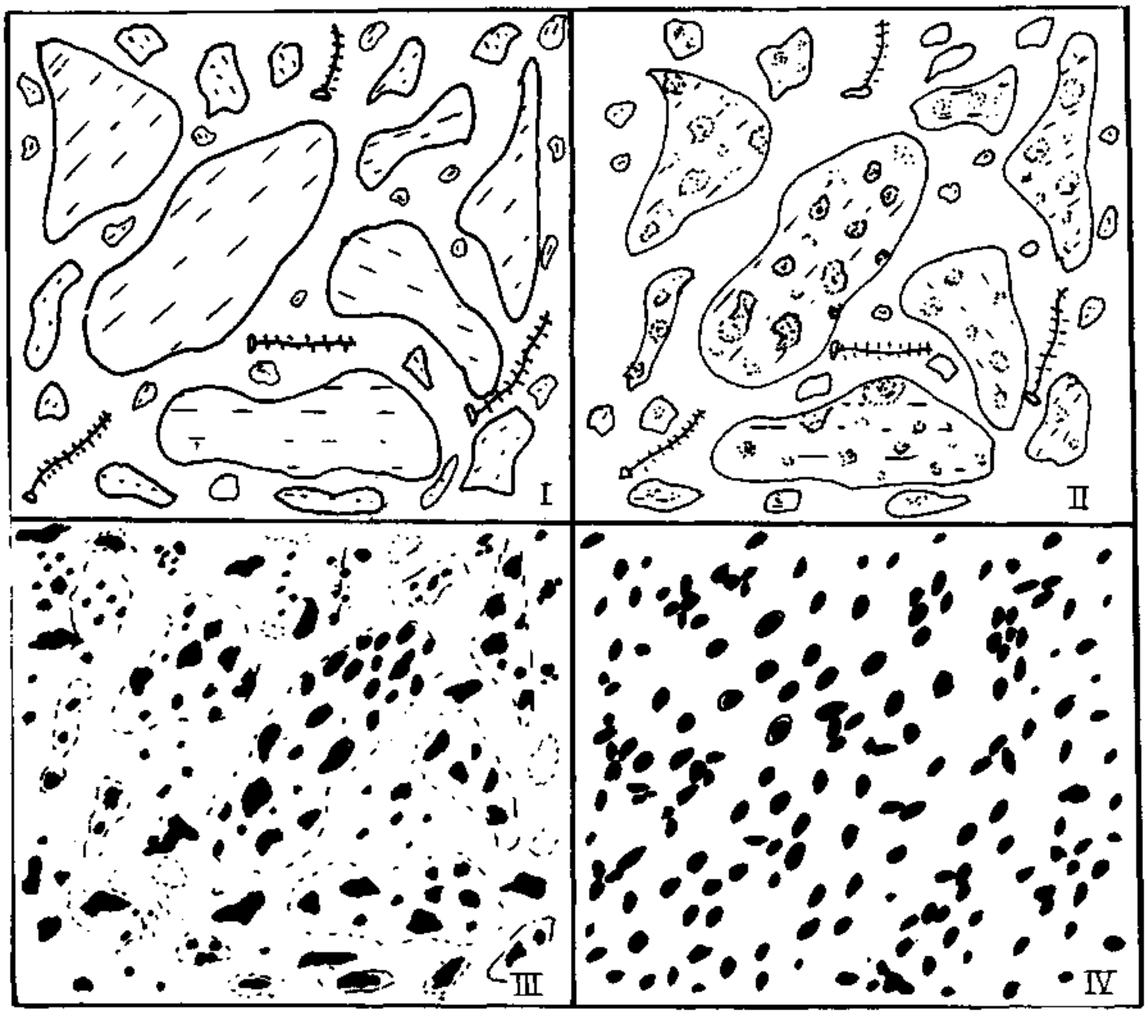

Fig. 3 — Diagrammatic model showing development of phosphate peloids from fish bones by phospho—micritization and physical abrasion. I—Bone piles; II—The pile's phospho—micritization of the bone fragments, by microbiological processes *(Schizomycophyta)*. This phase is shown on Plate II, fig.5; III-Disappearance of bone textures with increase of the phosphate content; IV-The conversion of the disintegrated fish bones, which have undergone the physical alteration by microbiological processes, to phosphate peloids.

for phosphate, which could have replaced biogenic grains in coastal limestone that were in contact with groundwater.

The last phosphate pelloid type associated with fish bone beds, can be attributed to microbiological activity. Similar examples have been interpreted as phosphatization of the fish bones via the activity of the microalga *(Schizomycophyta)* (Soundry and Nathan, 1980). The processes is a phospho—micritization in association with physical abrasion, leading to peloid formation. This occurs in the same way as micritization of a calcitic shell through microboring algae (Soudry, 1979; Friedman et al.,1971; Bathurst, 1976). Especially, physical abrasion with microalgal activity could be a prime agency to construct the pelloidal forms in the studied samples.

In the electron microscope, regular microgrooves and longitudinal microrelief reflect the remnants of the bone texture in the pelloidal phosphate mass. Moreover, micritic inner zones should result from a phospho—micritization process, and also minor bone fragments within the weak anisotropic phosphate peloids suggest that this phosphate peloid type was generated from fish bone material by phosphomicritization under the influence of microalgae. On the other hand, the low content of the non—phosphatized material and calcite cement indicate a closed system during phosphatization of the fish bones. The following sequence is proposed: (1) The piles of the fish bones experienced the microbiogenical attact in the coastal area. The period might have been long, with non and/ or slow sedimentation; (2) The microbiogenic activity increased with time and some portions of the fishbones were converted to the phosphate (Plate II, fig. 5); (3) The original fish bone textures were severely disturbed by the processes of phosphomicritization; (4) The phosphatized fish bones disintegrated into the pelloidal grains. The conversion from fish bones to the phosphate oeloids is illustrated diagramatically in Figure 3.

CONCLUSIONS

The phosphate peloids in the Mazıdağ—Derik area resulted from a chain of complex events. The first type occured under the open marine conditions and precipftated as microswellings and nodules from interstital water rich in dissolved phosphate. Sometimes, microcrystalline to semi—crystallized- types were associated with the nodular phosphate formation. The micronodules were perhaps mobile onto the sea bottom during the formation of the pseudoolitic phosphate, which suggests active currents during growth of the phosphate peloids. In this study, this type is interpreted as in situ zonal peloids.

The other types of the phosphate peloids were mostly concentrated in the shallow water carbonate facieses, and they formed through different processes than the former type. Ground water rich in dissolved phosphate and microbiogenic activity were the prime agencies for the. phosphatization and the production of the pelloid grains. Ground water, which percolated into the phosphate karst nearby the limestone realms led to phosphate replacement of biogenic grains. Microbiological activity through prosphomicritization *by Schizomycophyta* or a similar algae, aided by physical abrasion, formed phosphate peloids from fish bones as the condensed phosphate deposits.

The environment in which phosphate was formed underwent sea level fluctuions during or soon after phosphate formation, and consequently different types of phosphate peloids were often mixed each other in the same bed in the Derik-Mazıdağ phosphate succession.

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PLATES

PLATE-I

Fig. 1 — The zonal internal texture of light—colored, in situ formed peloid grains. The dark colored central parts show organic material and the remnants of diatomaceous mud inclusions which exhibit a range phosphatization.
Ta-12 (Taşıt), X 30.

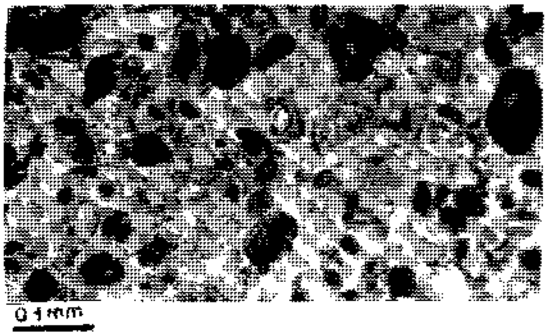
Fig. 2 — Homogenous, dark colored phosphate peloids, formed by phospho micritization of fish bone.
Şe-3 (Şemikan),X 30.

Fig. 3,4 - Scanning with the EDAX, the zonal peloids show an inhomogenous phosphate distribution.
Ka-9 (Kasrık).

Fig. 5,6 Phosphate within the interior of a phosphate peloid, showing micro-swellings and micro-nodules. The dark colored botton part is diatomaceous mud.
Ta-14.

Fig. 7 Pure phosphate micronodular accumulations, the organic mud was greatly disappeared within themicronodularmass. This material represents the pure and clear rim of the peloids.
Ta-12.

Fig. 8 - Amorphous phosphate, apart from very small and free micronodules, no granulation has been developed.
Ka-4.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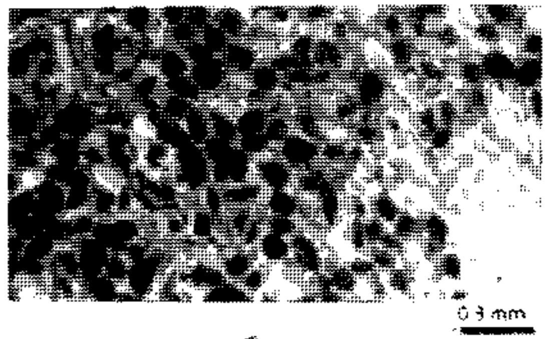
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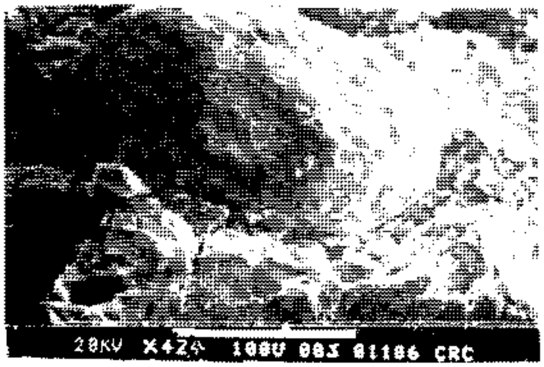
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3


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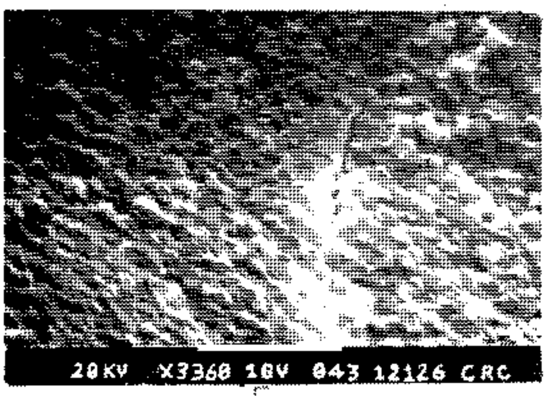
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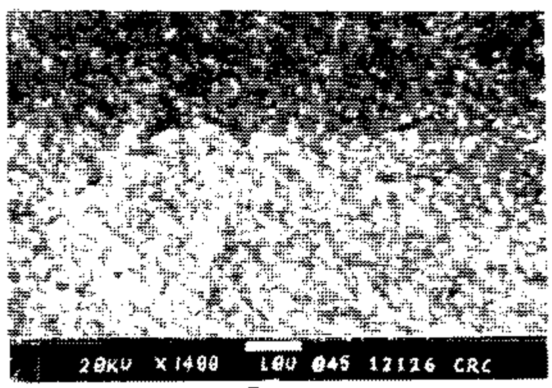
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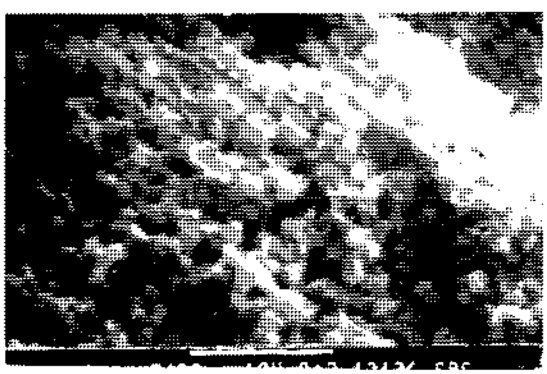

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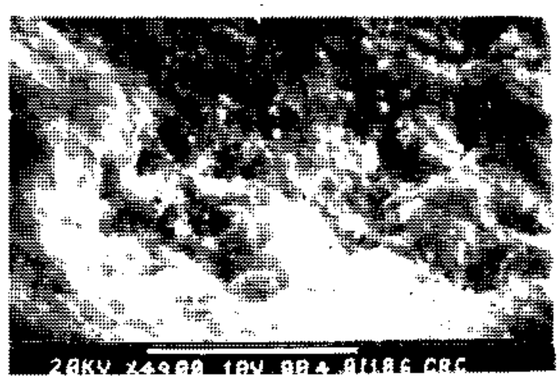
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8

PLATE-II

Fig. 1 — The free micronodules in the interior of the pseudooid—type phosphate peloids. This reveals the water's chemical transport at the interface of the sediment and sea water.
Ta-11.

Fig. 2 — Rosette and fusiform segregations of phosphate in a light-colored peloids.
Şe-l.

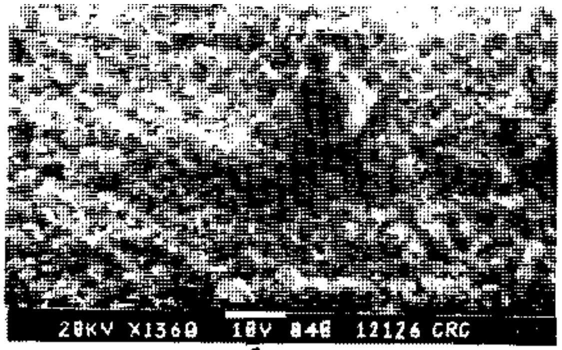
Fig. 3 — The developing microapatite crystals cementing the micropores of diatom tests.
Ka-2.

Fig. 4 — The cylindrical to flat microcrystalline apatite crystal aggregates which replaced the carbonate mud matrix. The internal texture of a typical peloid derived from intraclasts.
Ta-4.

Fig. 5 — A fish bone fragment in early stage of conversion to a phosphate peloid by phospho—micritization.
Ta-9, X 25.

Fig. 6 — An ordered bone texture, showing typical internal structure of fish bone fragment.
Şe-3.

Fig. 7,8— The internal texture of a peloids formed by phospho—micritization. The micro—granules and crystals have been developed within the ordered framework of the primary texture. Microrelief on the peloid surfaces suggests relict primary bone texture.
Şe-3.



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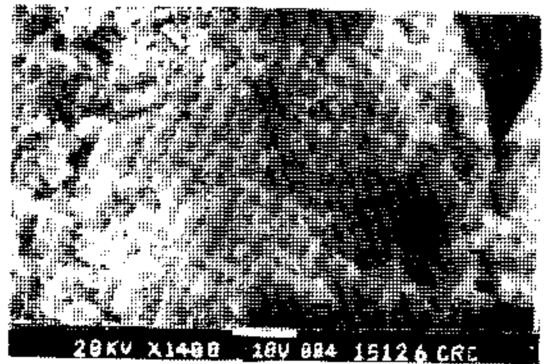

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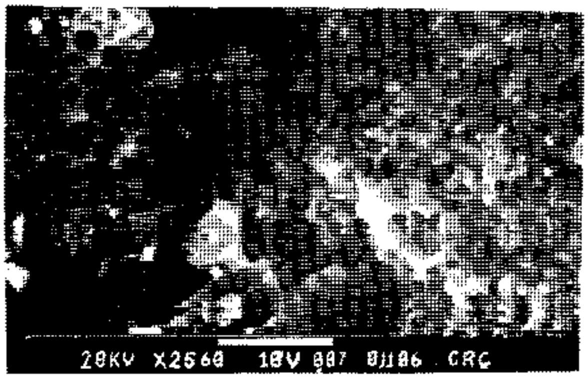

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4

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AN EXAMPLE FOR THE MULTIVARIATE GEOSTATISTICAL ANALYSES OF GEOCHEMICAL DATA: IRON MINES OF DİVRİĞİ AREA, CENTRAL TURKEY

Taner ÜNLÜ* and Henrik STENDAL**

ABSTRACT.— Geostatistical analyses were carried out on 160 rock samples for 24 elements from the Divriği iron ore region. The samples were initially treated as one population. Thereafter the individual rock types were divided into several groups and geostatistically analysed. The geostatistical methods are described shortly for univarite and bivariate analyses and, most importantly, the multivariate methods such as Discriminant, Cluster and Factor analyses. The results of the geostatistical analyses yield a division into different rock groups (Discriminant analysis), and several elementassociation (Cluster and Factor analyses) which reflect the different rock types. In the individual groups the elementassociation tells more about the geological processes e.g. serpentinization and hydrothermal alteration. The difference between Cluster and Factor analyses is seen in the Factor analysis, which is a little more differentiated, enabling a more subtle interpretation of the possible geological environment. The interpretation of the elementassociation suggests that the iron ores are closely associated with mafic to ultramafic rocks, their serpentinization and also later hydrothermal events.

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A RAPID DECOMPOSITION METHOD FOR ANALYZING ZIRCONIA

Bahattin AYRANCI*

ABSTRACT.— A rapid method is presented for the decomposition of zirconia using a suitable combination of (NH4) F + $(NH_4)_2SO_4$ as fusion agent at about 350°C. The cooled melt (cake) is easily soluble in acidified water in 15-20 minutes. The sample solution may be used for recovery of many components using several analytical methods. During the composition silica is volatilized completely.

INTRODUCTION

Due to its particular physical and chemical properties (e.g. high melting point and resistance to acids and alkalis) the element zirconium and also its compounds have a wide variety of applications in industrial and nuclear technology. The oxide of zirconium (zirconia) is one of the main component of glazed enamels and is largely used for opacity of ceramics as well as in ceramic color technology. Furthermore, zirconia is commonly used in manufacture of alkali resistant ceramics and is one of the fundamental components of refractory materials, briquetts etc.

Technical applications of zirconium compounds are extensive in the electronic and electrotechnical industries. Zirconium is also used in high vacuum technology, catalyst techniques, in the pharmaceutical industry as well as in the manufacturing of special glasses and synthetic gems (Gmelin, 1958; Hathaway, 1984).

In nature zirconium mainly occurs as silicate (e.g. zircon) and oxide (e.g. baddeleyite) which are commonly associated with magnetite ilmenite, monazite, rutile, garnet, sillimanite, quartz etc.

Of the twenty seven commonly known zirconium-bearing minerals which contain variable amounts of actinides and rare earth elements (Vlasov,1966),zircon and baddeleyite are the most interesting for science and technology.

The mineral zircon is not only for interest because of its zirconium concentration, but also because of its application in geochronological studies, whereas baddeleyite is commonly used in industrial purposes.

In the concentration of zirconium and manufacture of zirconia several methods (e.g. magnetic—, wet-mechanical separations, floatation technique)are used. These procedures do not seek to obtain absolutely pure zirconium concentrates. It is, however, necessary to know the amount of impurities (e.g. silica, alkali, aluminium, titanium, iron, manganese) before it may be used for industrial applications.

The determination of zirconium concentration in the dissolved samples may be performed easily by the usual analytical methods (e.g. by titrimetric-, gravimetric-, or spectrometric-methods). However, not only the determination of impurities from the same sample solution may involve difficulties in analysis (due to the higher zirconium concentration in the matrix) but also because the complete decomposition of zirconium minerals is difficult to ensure.

This paper deals with the disintegration problem of zirconia and represents a part of the paper "A Rapid Decomposition Method for Analyzing Zirconium Minerals and Zirconia" read in 8. Spectrometer Meeting, Baden - Baden FRG. 1986.

ANALYTICAL PROCEDURES

The decomposition technique commonly used to dissolve zirconia depends on the composition of minerals and also the preliminary thermal treatment. The disintegration of samples can be carried out by acid attack as well as by fusion methods.

Acid attack method

The disintegration of samples may be performed by reaction with conc. H_2SO_4, H_2SO_4+HF, HNO_3+HF. In order to dissolve one of the highly resistant zirconium materials, zirkite (consisting of a mixed fibrous baddeleyite, zircon, altered zircon (orwillite), and other minerals) requires four times the sample weight of conc. H_2SO_4 with a decomposition time of 2 h at about 400°C. On the other hand, the mineral cyrtholite is attacked using two times the sample weight of excess conc. H_2SO_4 at about 210-220°C for 30 minutes (Gmelin, 1958). A H_2SO_4 acid attack on zirconium material produces water soluble zirconium sulphates.

HF is also one of the efficient disintegration agents used for the decomposition of zirconium minerals and acts by converting zirconium to a soluble zirconium fluoride (1.388 g/100 ml). However, due to its low boiling point (max. 112°C) HF is rather inefficient during open vessel acid attack procedures. Instead of HF, an efficient acid attack can be performed using CaF_2 (m.p. 1360°C) with conc. H_2SO_4 (max. boiling temperature of conc. acid d 339°C). If the decomposition is carried out by CaF_2+H_2SO_4 (using 2 parts of CaF_2+2.5 parts of conc. H_2SO_4) either $H_2[ZrF_6]$ + $CaSO_4$ or basic zirconium-sulphate is formed (Weiss, L.; Marden, J.W. and M.Rich, in Gmelin, 1958). This procedure causes the loss of silica as SiF_4, and a part of titanium as TiF_4.

The common acid attack disintegration of zirconia at high temperature in a platinium vessel is not a suitable procedure because of the volatilisation and splashing of acids before the sample is completely dissolved. Due to incomplete decomposition of zirconia, the procedure is better performed in an autoclave using HF+H_2SO_4, HF+HNO_3 as used for zircon (Ito, 1962; Krough, 1973).

One of the very promising acid attack procedures used for the analysis of zirconia and titania is presented by Bastius (1984): Zirconia and titania are attacked by HF+H_2SO_4 in presence of ammonia. The more resistant materials are decomposed with an acid attack of H_2SO_4+.$(NH_4)_2SO_4$, so that the boiling point of H_2SO_4 is elevated. Furthermore, Bastius uses a two-step treatment: The highly resistant samples are attacked initially by H_2SO_4+$(NH_4)_2SO_4$ and then these are decomposed by HF+(NH_4)F. Possible contamination, which may occur using $(NH_4)_2SO_4$ is minimized by using high quality NH_3, H_2SO_4, HF.

Fusion method

Fusion methods commonly used for zirconium material analysis are carried out with a variety of fluxes, such as NaOH, Na_2O, Na_2O+NaOH, Na_2O+ Na_2CO_3, NaF+$Na_2B_4O_7$, $NaHSO_4$, KF, KHF_2, $NaHF_2$ (Bock, 1979, Dolezal, et al., 1968; Gmelin, 1958).

The cooled melt is dissolved by selected acids. These fusion procedures of zirconia mentioned above have several disadvantages due to the various fluxes used, which yield undesirable components (overloading) in the solution of sample. In addition, if the sample decomposition is carried out at low temperatures, the sample may not be dissolved completely, whereas at temperatures that are too high insoluble zirconium compounds (e.g zirconium—oxide) are produced.

The ammonium salts (e.g. (NH_4)F, $(NH_4)_2SO_4$, (NH_4)C1) have also been used as fusion agents to decompose ores and silicates (Bock, 1979; Dolezal, et al., 1968, Liteanu and Paniti, 1972, Milner, et al., 1967; Verbeek, et al., 1970).

Shead and Smith (1931) performed ammonium fluoride fusion for the decomposition of refractory silicates (e.g. sillimanite), and the analysis of silica from glass sand.

Bayer, et al. (1982) reported sulphate reaction of Si—, Al—, Fe—, Mg—, Ti—bearing silicate minerals, where reactions were carried out with ammonium sulphates at 350—550°C. Due to the particular affinity of zirconium for ammonia slightly soluble, double ammonium—zirconium sulphate compounds may be produced during the disintegration of zirconium material by H_2SO_4 in presence of ammonia.

Devillebichot (1983) decomposed zirconia by ammoniumbifluoride for the producing of ammoniumheptafluorozirconate. If NH_4HF_2 (m.p. 125°C) is used to attack zirconia complete decomposition of sample can not be obtained. This is due to the lower boiling temperature of the flux and is also dependence on the preliminary thermal treatment of starting material. After adding $(NH_4)_2SO_4$ m.p.235°C to the NH_4HF_2, the melting point of ammonium bifluoride can be elevated, this aims to additionally minimalize the volatilisation of zirconium and titanium as fluorides and the formation of all zirconium as zirconium-fluorides. The use of $(NH_4)_2SO_4$ as fusion agent for the decomposition of zirconia is connected with some difficulties, due to the impurities of silicates, which may be present in the sample and do not react during the disintegration procedures.

The addition of a complementary compound such as NH_4HF_2 (or NH_4F) to the $(NH_4)_2SO_4$ produces a potential flux, which is more efficient than alone using $(NH_4)_2SO_4$ or NH_4HF_2 in disintegration procedures of zirconia as fusion agent.

Decomposition of zirconia using $(NH_4)_2SO_4$+ $(NH_4)F$ as a fusion agent can be carried out also for the production of slightly soluble double—sulphates and fluorides of Zr, Ti, Fe, Al. During the disintegration the silica is volatilized completely. This procedure may be considered similar to an acid attack carried out by H_2SO_4+HF at 350°C, but it is less dangerous and more effective than the H_2SO_4+HF disintegration.

RAPID DISINTEGRATION PROCEDURE

A rapid decomposition of zirconium material may be obtained if the material is fused with a suitable combination of $(NH_4)F+(NH_4)_2SO_4$. This method is preferred because of its convenience compared to acid attack and the use of fluxes because of potential excess components being included. The fusion of zirconia is performed by $(NH_4)F+(NH_4)_2SO_4$ in a 25 ml covered platinum crucible using a tubular furnace.

Tubular furnace

This is made using a ceramic tube which is closed at one end and is 300—350 mm in lengh and about 100 mm in diameter. The heating is performed by a heating band, which should be well isolated to prevent temperature gradients. The oven is heated to 350°C and the temperature is measured by an inserted thermoelement (or thermometer). An outlet is needed for the escape of fumes that evolve during the decomposition of the sample.

Sample decomposition

The experimental procedure presented in this paper, was carried out on a 1000 mg sample (zirconia) which was mixed with 2500 mg $(NH_4)F$+5000 mg $(NH_4)_2SO_4$ in a closed platinum crucible and disintegrate for 45-60 minutes at 350°C.

At about 350°C the sample (with flux) is placed in the tubular oven and left for 45—60 minutes for the fusion procedure. The crucible is quenched after removal from the oven and transferred to a beaker containing 50—100 ml cold water + 1—1.5 ml conc. HC1 (37%). Instead of HC1, H_2SO_4 may also be used. Heating of the solvent which is used to dissolve the cooled melt (fusion cake) is not necessary and it should be avoided to prevent the formation of insoluble zirconium compounds. The cake is dissolved very quickly (during 15—20 minutes) by stirring using a magnetic-bar and stirrer. After dissolving the cake in acidified water a clear solution is obtained. Experience shows that for the decomposition of samples with a high content of silica (e.g. 15 %) the relative amount of $(NH_4)F$ must be increased up to 5000 mg. In a forthcoming paper the application of this flux combination, to analyse silicate rock samples will be discussed in detail (Ayrancı, in prep.).

The reproducibility of this procedure is checked by applying it simultaneously to three aliquots of a sample. It was found that the samples gave the same results within the analytical error, which itself depends on the method being used. The components of the sample solutions were analyzed by AAS and ICP. The analytical results of this experimental work and the recovery of individual element concentrations will follow.

Table 1

	Al		Fe		Ca		Ti		Remarks
	Cert.	Recov.*	Cert.	Recov.**	Cert.	Recov.*	Cert.	Recov.*	
Aliquot 1	<500	470	<40	45	<15	25	<15	12	Recovered values are
Aliquot 2	<500	460	<40	55	<15	20	<15	18	mean of AAS and ICP
Aliquot 3	<500	490	<40	50	<15	20	<15	17	analyses

Starting Material: Zirconia, U.P.H. (grain size 0.04-0.5 m., contains 2.5 % HF), Cricerom, France.
Certified impurities: Al < 500, Ca < 15, Fe <40, Mg <2, Na < 15, Si <30, Ti <15 ppm.3 Aliquots of zirconia simultaneously decomposed with (NH.) F + $(NH)_2SO_4$ in closed platinum crucibles at about 350° C for 55 minutes. Aliquot 1.2.3: 1000 mg zirconia + 2500 mg (NH_4) F + 5000 mg $(NH_4)_2SO_4$ After dissolving cooled melt in an acidified water (containing 1.5 ml conc. H_2SO_4 in 100 ml H2O), the given elements were analized.

* Measured by AAS and ICE

** Recovered by AAS and Spectrometric procedures.

CONCLUSIONS

The disintegration of zirconium materials using $(NH_4)F+(NH_4)_2SO_4$ as a fusion agent and dissolving the cooled melt (fusion cake) by dilute acids (e.g. HC1, H_2SO_4) yields negligible concentration of extra components that are undesirable for the recovery of components in analytical procedures. Due to the absence of several components (e.g. KF, NaOH, $Na_2B_4O_7$, CaF_2, Na_2O) it is also possible to analyse these components themselves in the sample solution.

The sample disintegration by $(NH_4)F+(NH_4)_2SO_4$ in a closed platinium crucible is rapid and it is also less dangerous than a common H_2SO_4+HF acid attack carried out in open vessels. After the disintegration of sample using such ammonium fluxes, mostly double ammonium sulphates of the elements Al, Ti, Fe, Zr and (perhaps) also ammonium zirconium fluorides are formed. These products are slight soluble in acidified water. The sample disintegration is completed quickly and at low temperatures (about 350°C).

During the fusion of zirconium minerals using $(NH_4)F+(NH_4)_2SO_4$ silica is completely volatilized as SiF_4, H_2SiF_6. The mothod may be used for the routine analysis of zirconia in samples of various weights, and in runs of various sample.

The known chemical affinity of zirconium for ammonium in presence of H_2SO_4 to produce water-soluble double ammonium—zirconium sulphates and ammonium zirconium fluoride may be utilized if the zirconium material is decomposed by $(NH_4)F$ + $(NH_4)_2SO_4$ or $(NH_4)HF_2+(NH_4)HSO_4$ in a closed platinium crucible. This procedure is similar to a H_2SO_4+HF acid attack method, which is well known in "rapid silicate analysis" to prepare a "solution B" (Maxwell, 1968). The procedure may also be used for the disintegration of silicate rocks and minerals as an alternative decomposition method.

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PRESENCE OF UPPER TRIASSIC BY CONODONTS IN ARMUTLU PENINSULA (WESTERN PONTIDES)

Fuat ÖNDER*** and M. Cemal GÖNCÜOĞLU****

ABSTRACT.— At the vicinity of Elmalı village to the NE of İznik (Armutlu Peninsula, Western Pontides) recent data have shown the presence of Upper Triassic conodonts in the limestones on the upper part of metaclastics, which were previously considered to be of Paleozoic age. New techniques were utilised in the obtaining and determination of the conodonts in this study, as the conventional methods have failed.