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STRATIGRAPHY OF KARABURUN PENINSULA

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ABSTRACT. — In the Karaburun peninsula, a tectonic bell with a thick Mesozoic succession is located bordering the Izmir-Ankara zone on its western side. In the stratigrapnic column of the Karaburun belt, the oldest unit is the Lower and Middle Carboniferous Alandere formation consisting dominantly of fossiliferous limestones. Lower Triassic rocks rest directly above this unit and Upper Carboniferous and Permian sections are missing. The Lower Triassic is represented by rock units showing facies changes in short distances along vertical and horizontal directions. In this part of the section, the Karareis and Gerence formations are differentiated which are collectively named as the Denizgiren group. The Karareis formation is composed of intercalations of sandstones, bedded-black cherts, pelagic limestones and mafic volcanics, whereas the Gerence formation composed dominantly of ammonitic red limestones, thinly-bedded gray limestones and chcrty limestones. The Karareis and Gerence formations grade laterally into each other and range in age from Scythian to Late Anisian. The Camiboğazı formation resting with a gradational contact on each of these units, consists of massive limestones with reefal facies in places and yields an age of Ladinian-Camian. The Güvercinlik formation lies conformably above the Camiboğazı formation, and consists of algal stromatolites, megalodon-bearing limestones and quartzitic sandstone intervals. This unit ranges in age from Norian-Rheatian. The Nohutalan formation which is dominantly represented by thick-bedded ncritic limestones, overlies the Güvercinlik formation gradationally and yields an age ranging from Liassic to Albian. This unit appears to be lithologically continuous in the field without any indications of a break in the stratigraphic or structural record, however the Dogger age has not been determined which may probably indicate a presence of a hiatus. Above this Mesozoic comprehensive succession with an age spanning from Early Triassic to Early Cretaceous, The Balıklıova formation of Campanian-Maastrichtian age rests unconformably, which consists of carbonate rocks and sandstones of flysch facies. The Karaburun belt with the stratigraphy outlined above, is surrounded from all sides by a blocky unit which is called the Bomova melange. This blocky unit with highly sheared flysch matrix, was formed in the Izmir-Ankara zone during a Maastrichtian-Danian interval. The boundary relations between the Karaburun belt and the Bomova melange indicate that the Karaburun platform was transported tectonically as a nap into the İzmir-Ankara zone during its opening. In this study we have also found that, the stratigraphy of the Karaburun belt is completely different from that of the so-called Sakarya continent, and they can not be correlated with each other as suggested by the earlier workers.

INTRODUCTION

Three tectonic belts are located in the western Anatolia, that have taken their present position during the Middle to Late Eocene thrusting (Fig. 1). The easternmost of these is the Menderes massif, the middle one İzmir-Ankara zone and the western one is the Karaburun belt. During an extensional tectonics since Middle Miocene, grabens cutting obliquely the strike of these older belts have been formed and, in which lacustrine and continental sediments have been deposited.

The Menderes massif consists of very thick mica schists and quartz mica schists in the lower and marbles in the upper parts. The marbles give an age of Devonian-Permian in the lower horizons and Triassic to Late Cretaceous in the upper continuous sections. In recent studies, the youngest age from the metamorphic massif, has been reported as Early Eo*cene,* indicating that the main metamorphism took place after this period (Boray and others, 1973; Çağlayan and others, 1980; Konak and others, 1987).

The İzmir-Ankara zone, that forms general NW-SE trending outcrops in the west of the Menderes massif, is represented by a rock unit called the Bornova melange in the western Anatolia (Erdoğan, 1985). The Bornova melange with strongly deformed internal fabric, is composed of a matrix of flysch and mafic volcanic rocks of Campanian-Danian age and blocks of neritic limestones up to 20 km in length. Partial sections measured from these blocks indicate that they the broken parts of a platform-type succession with an age range from Triassic to Late Cretaceous and this succession was similar on lithological, paleontological and stratigraphic aspects to the Karaburun carbonate association (Erdoğan, 1990). The İzmir Ankara zone represented by the Bornova melange, thrustcd in the Late Eocene on top of the Menderes massif along low angle faults.

The Karaburun belt lies in the westernmost side.of the paleotectonic belts of Anatolia and in which a thick and continuous carbonate succession with an age spaning from Triassic to Albian crops out. It constituted the platform of the İzmir-Ankara zone and moved first as blocks and later as a large nappe into the Bomova melange in the Late Cretaceous (Erdoğan, 1990). Later on, together with the rock units of the İzmir-Ankara zone, the Karaburun nappe thrusted over the Menderes massif in the Late Eocene. This compressional tectonics, which started with the dislocation of the Karaburun platform in

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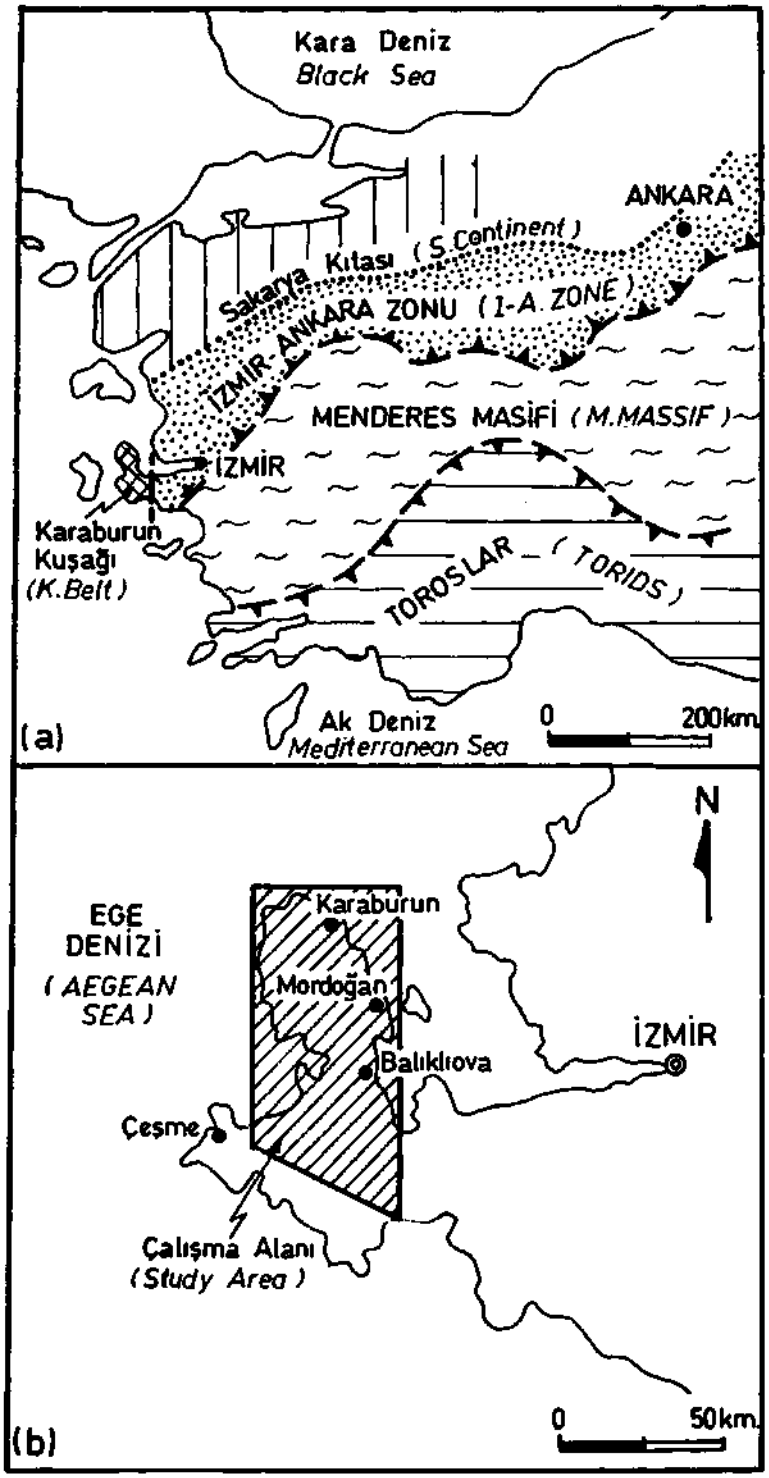

Fig. 1 – Paleotectonic belts of the western Anatolia; In the east Menderes massif, in the middle İzmir-Ankara zone and in the west Karaburun belt are located. The İzmir-Ankara zone thrust over the Menderes massif along low angle thrust faults.

the Maasirichtian and ended in the Late Eocene, strongly erased tracts of the early history of the tectonic evolution of the region. Detailed examination of the stratigraphy of the Karaburun carbonate succession will probably disclose the pre-Maastrichtian evolution of the western Anatolia. The stratigraphy of this succession was constructed on the basis of geological mappings of small areas, in the previous studies. However, the rock units of the area show facies changes in short distances and thus too many units were differentiated in each small areas studied, and mistakes were made in their stratigraphic correlations.

In this study, the geological map of the entire peninsula is prepared on 1:25 000 scale (Fig.2) and by measuring sections, stratigraphic relations of the differentiated units are determined. In namming the formations, earlier recommended

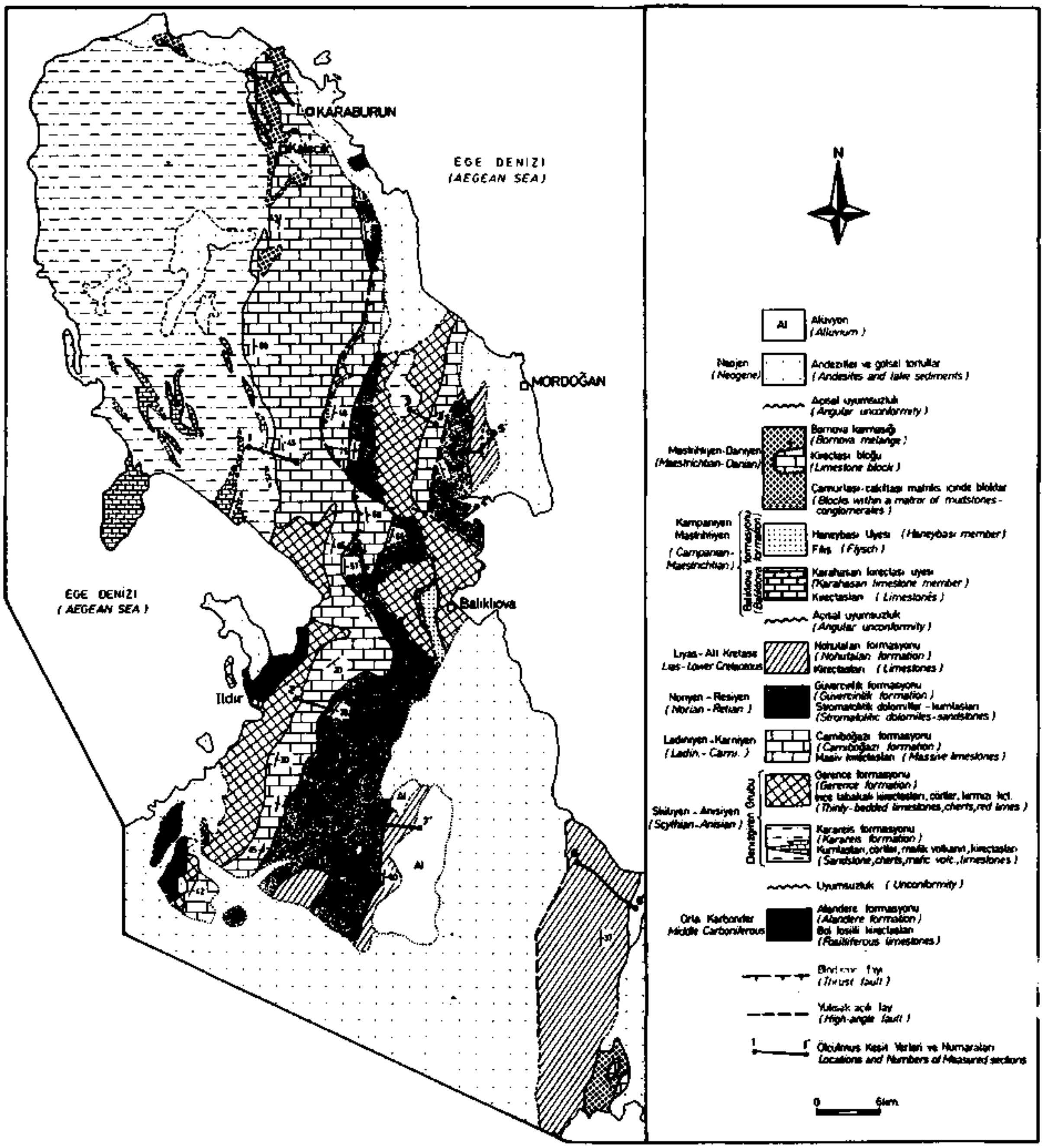

Fig. 2 - Simplified geological map of the Karaburun peninsula; locations of the measured sections discussed in the text are identified with numbers

names have been conserved as much as possible, but their boundaries arc redefined. Some new names arc given and number of units is reduced based on their map ability.

The geological mapping and writing of this paper were made by Burhan Erdoğan. Talip Güngör has joined to the first author in the geological mapping of the Balıklıova-Ildır area. The palconlological determinations were made by Demir Altıner and the stratigraphic sections were measured by Burhan Erdoğan and Sacit Özer.

PREVIOUS STUDIES

The first study on the stratigraphy of the Karaburun peninsula is that of Phillipson who prepared in 1911 a 1:300 000 scale map of the region. He called the Paleozoic basement to the unit forming large outcrops in the northwestern part of the peninsula, that is composed of sandstones, black cherts (lidites) and mafic volcanic rocks, and he suggested its age as Carboniferous on the basis of *Fusulinella* sp. found in the limestones around Ildır. He noted that above this detrital Paleozoic laid along an unconformity a thick Mesozoic carbonate succession.

Ktenas (1925) examined Chios Island and Karaburun together and stated that the oldest part of the succession is located in Oinoussai Island (Koyun adası) to the east of Chios and it consists of low-grade mctamorphic rocks, slates and sandstones with a probable age of Ordovician. He proposed the name Denizgiren group to the intercalations of black cherts and sandstones forming extensive outcrops in the northwest of the Karaburun peninsula, and assigned their age as the Early Devonian. The limestones and detrital rocks around Kalecik village were called by him as the Kalecik beds and they were thought to be the Devonian-Carboniferous. Above the Kalecik beds, he distinguished the Ildır and Yayla groups which were described as consisting of black cherts, graywackes, mafic volcanics and conglomerates with assigned age of the Middle Carboniferous. Above these various detrital Paleozoic he differentiated a thick Mesozoic carbonate succession.

Kalafalçıoğlu (1961) prepared a 1:100 000 scale geological map of the peninsula in the later period. He defined the detrital rocks of the northwestern part of Karaburun as the Devonian graywackes, and recognized on top of these units a continuous carbonate succession of Jurassic to Cretaceous age.

Höll (1966) and Lehnert-Thiel (1968) examined the mercury deposits in the detrital rocks which were considered Paleozoic in the previous studies. Correlating with the similar units in Chios they assigned their age as ranging from Gothlandian to Carboniferous.

In a relatively recent work, the stratigraphy of Karaburun has been examined by Brinkmann and his students between Ildır and Balıklıova regions (Brinkmann and others, 1972, 1977; Gümüş., 1971). In these works, the detrital rocks covering extensive areas in the northwestern part of the peninsula have been omitted all together and not been shown in their stratigraphic column. In their stratigraphic sequence, the oldest units are the Alandere and Tınaztepe formations of Early Carboniferous age. Above this old basement as stated so, they have differentiated a continuous series from Lower Triassic to Lower Cretaceous. Their defined stratigraphy consists of the following units from the bottom to the top: the Domuzçukuru, Koyutepe and the Laleköy formations of Scythian-Anisian; the Camiboğazı formation of Ladinian; the Hanaylı and Güvercinlik formalions of Camian; the Nohutalan formation of Rhaetian-Liassic; the Cladocoropsis limestones of Malm and the Aktepe formation of Early Cretaceous.

The northwestern part of the peninsula has been examined by Konuk (1979). He has determined the age of the detrital units which are composed of graywackes, cherts, olistostromal intervals and limestone lenses as Triassic, and considered as a separate tectonic belt from the Mesozoic carbonate succession of the Karaburun peninsula. He has given the name of the Karareis belt for these detrital and pelagic Triassic and the Karaburun bell for the Mesozoic carbonate succession. He has interpreted the Karaburun belt as a nappe overlaying the Karareis belt along a low-angle thrust fault.

STRATIGRAPHY

PALEOZOIC

The lowermost part of the Karaburun succession is made up of Lower to Middle Carboniferous limestones and above this there is a continuous section with an age range from Early Triassic to Early Cretaceous. On top of this continuous section, a Campanian-Maastrichtian unit lays along an unconformity.

Alundere formation

The oldest unit of the Karaburun succession is the Lower-Middle Carboniferous Alandere formation. This unit, which crops out in a region between Ildır and Reisdere, is composed of dark brown and dark gray limestones with abundant fossils and crinoid stalks. This unit is equivalent of both the Tınaztepe and Alandere formations of Gümüş (1971). Brinkmann and others (1972), however, have called the stratigraphic equivalent of this unit as the Lower Carboniferous massive limestones in their stratigraphic column.

The Alandere formation, besides in the close vicinity of Ildır, forms small outcrops in the northeast of Ildır and near the Balıklıova village. Around Reisdere, this unit is represented by massive limestones with gray and buff color. The most pronounced feature of these limestones is their abundance in coral fossils and in crinoid stalks. In the north of Ildır, this unit consists of dark-gray fossilliferous limestones, limestone conglomerates and green sandstone intervals. The detrital facies of the Alandere formation resembles lithologically to the Lower Triassic rocks and for this reason these different units have been confused by Gümüş. (1971) and Brinkmann and others (1972) in their geological mappings. The marked difference of the Alandere formation from the Lower Triassic rocks is their limestones being dark-gray in color and rich in fossils relative to the Triassic counterparts. In addition, the red and green cherts and red limestones which characterize the Lower Triassic units arc lacking in the Alandere formation.

The lower contact of the Alandere formation is not seen in the Karaburun peninsula and the total thickness of the outcropping part is more than 300 m. This unit forms the lowest part of the Karaburun succession and as contrary to the opinions of earlier workers (Ktenas, 1925; Höll, 1966; Lehnert-Thiel, 1968) there is no older unit than the Alandere formation in the peninsula. The upper contact is observed along the road connecting Ildır to the Gerence bay and, dark-gray fossilliferous limestones of this unit is directly overlain by light-gray limestones of the Lower Triassic. Along the contact zone, there is a karstic and oxidized horizon suggesting a subaerial exposure. To the east of Reisdere village, this contact can again be seen but 3-4 m wide area along the boundary is covered. At this last location, massive limestones of the Alandere formation are covered by thinly-bedded limestones and red-green chert intercalations of the Lower Triassic Gercnce formation. Close to the boundary, a conglomerate horizon, that can be traced laterally 15 to 20 m, is present but its clasts consist entirely of Triassic material suggesting its being an intraformational breccia rather than a basal conglomerate. In the north of Ildır, close to the Alantepe location, thinly-bedded cherty limestones and pelecypoda-rich limestones of the Gerence formation rest directly above the Alandere formation. This contact is also seen to the west of Eski Balıklıova village where the limestones of the Lower Triassic overlay the Carboniferous; at this last location lithologically there is no distinction between the two different units and only by palcontologic determinations they can be distinguished from each other.

As summarized above, observations from different areas suggest that, the Lower Triassic rocks directly rest upon the Lower-Middle Carboniferous Alandere formation and, in between. Upper Carboniferous and Permian are missing. The presence of a zone of karstic solutions along the boundary may suggest a period of subaerial exposure between the two units.

Point samples collected from different levels of this formations and, series of samples from the upper parts, have yielded the following list of fossils that give the early Middle carboniferous (Baskhirian) age for the formation.

Eostafella postmosquensis, E. postmosquensis acutiformis, E. pseudostruvei, E. varvariensis, E. ex gr. *ikensis, Pseudoslafella antiqua, P. compressa, Eostqfellina protvae, Eos. paraprotvae, Plectostaffella inconstans, Globivalvulina moderala, G. scaphoidea, G. bulloides, Bradyina cribrostomata, Earlandia elegans, Endothyra baschkirica, E. spirilliniformis, Pscudoendothyra* aff. *struvei. Planoendothyra* sp., *Eotuberitina reitlingerae, Glomospira subquadrata, Pseudoglomospira* sp., *Palaeonubecularia fluxa, Asteroarchaediscus gregorii, A. postrugosus, Monotaxinoides donbassicus, Diplosphaerina Inacqualis. Tetraxis conica." Endotexis "* sp., *Milerella* sp., *Trepeilopsis* sp., *Haplophragmina* sp., *Clymacammina* sp., *Turrispiroides* sp., *Deckerelta* sp., *Mediocris* sp.

Brinkmann and others (1972) have described the Alandere formation of this paper as the Paleozoic massive limestones and given their age as Early Carboniferous. Gümüş, (1971) suggested the age of the equivalent units as Early Carboniferous including Viscan. In these two papers below the Carboniferous units, Devonian black cherts and graywakes have been mentioned. We have found that, their Devonian, partly belongs to the detrital facies of the Lower Triassic and partly to the Carboniferous Alandere formation.

This formation is rich in coral and crinoid fossils, and in places the limestones of reefal facies contain intraformational conglomerate intervals. It was formed in a shallow marine environment with moderate climatic conditions.

MESOZOIC

In the Karaburun peninsula the Mesozoic is represented by a continuous succession from Lower Triassic to Albian. The lower part of this succession, which is Scythian-Anisian in age, shows facies changes even in short distances. These different facies have been named as formations in the earlier studies (Brinkmann and others, 1972; Konuk, 1979), and thus their stratigraphic relations have created problems.

In our study, in the *Lower*-Middle Triassic, the Denizgiren group has been differentiated as it was first suggested by Ktenas (1925). In this group, the Karareis and the Gerence formations have been recognized. The Karareis formation forms extensive outcrops in the northwestern corner of the peninsula and consists of bedded black cherts, sandstones, mudstones, mafic volcanics and pelagic limestones. The Gerence formation is the lateral equivalent of the Karareis formation and consists dominantly of thinly-bedded gray limestones, marls, ammonitic red limestones and red-green cherts. This unit is found in the southwest and east of the peninsula.

The Camiboğazı formation rests with a gradational contact on both of the formations of the Denizgiren group. Upward in the sequence above the Camiboğazı formation, first the Güvercinlik and second the Nohutalan formations arc present. The Upper Cretaceous Balıklıova formation unconformably overlies the Mesozoic sequence.

Denizgiren group

The Denizgiren group includes the detrital-rich Karareis and the carbonate-rich Gerence formations.

Karareis formation . _ This unit is composed of buff sandstones, mudstones, thinly-bedded black cherts and lenses of pelagic limestones. In the upper part of the unit, there are intervals of mafic tuffs and mafic volcanics.

This unit first has been examined by Konuk (1979) and interpreted as a separate tectonic belt in the peninsula and named as the Karareis tectonic belt. In our study, the Karareis formation is equivalent of 6 out of the total 7 formations of the Karareis tectonic belt of Konuk, excluding only his Tuzluk formation. This unit is heterogeneous in lithology and various lithologics grade laterally into each other. These various lithologies are named together as one formation, rather than distinct units as proposed by Konuk. Besides that, in our study, the Karareis formation is not considered as a separate tectonic belt but only a unit in the Karaburun succession.

The lower parts of the Karareis formation, as seen in the northwestern corner of the peninsula near Sarpıncık village, consist of a thick succession of mudstones. Upward in the sequence buff sandstones and thinly-bedded black cherts become dominant. In the middle and upper parts, yellow and red coloured thinly-bedded limestone lenses with micritic facies are abundant. These lenses include thin-shelled pelecypods and rarely ammonites and reach up to 300 m in thickness. They can be faced laterally one or two kilometers. These pelagic limestone lenses grade laterally into black chert intervals and mudstones.

In the Karareis formation, there are intervals of olistostromes with limestone blocks of various sizes and ages set in a matrix of mudstones. In rare cases, these blocks reach up to 500 m in length and contain abundant crinoid stalks. In some of these blocks Carboniferous age has been determined and they are found to be similar in lithology to the limestones of the Alandere formation.

In the uppermost parts of the Karareis formation, there arc hyaloclastites and mafic lavas interbedded with mudstones. The mafic lavas are composed of albite and augite fenocrysts of up to 0.5 cm set in an aphanitic matrix. They have affected by strong alteration and amgydules and veins are filled with chlorite, calcite and epidote minerals.

Although the Karareis formation shows heterogeneous lilhology, the internal structure is rather regular. For example, the black chert and pelagic limestone lenses thin and thicken along the strike but still show lateral continuity, so that their map patterns discern very tightly folded internal structure of the Karareis formation.

Because of the tightly folded internal structure, the true thickness of the unit can not be predicted definitely but from geological cross sections it is estimated to be more than 2000 m.

The lower contact of the formation is now here seen in the peninsula but its equivalent Gerence formation rests on the Carboniferous Alandere formation. The upper contact is gradational with the Camiboğazı formation and as seen along the İdecik Çeşmesi section on Figure 3, there is a 19 m thick gradational zone between these two units characterized with thinly-bedded yellow limestones. The gradational nature of these two formations can also be followed in the paleontological determinations as there is a continuous range of an age from Anisian to Ladinian.

The Karareis formation is very poor in fossil content and, because of strong secondary dolomitization, only radiolaria remains and sponge spines arc found in few places. In only a few of the samples collected from various levels of the formation, *Spirorbis phlyctaena* and, pelecypoda and echinodermata remains have been recognized which have yielded a probable Early Triassic age (Scythian).

In the upper part of the formation, in the vicinity of the İdecik Çeşmesi section (Fig. 3), a badly-preserved ammonite has been found, which docs not give a precise age. The samples collected for conodont determinations (Fig. 3), however have yielded Late Schythian-Early Anisian age. The foraminifera contents, on the other hand, indicate a range of age from Late Anisian to Early Ladinian between the Karareis and overlaying Camiboğazı formation. Along this transitional zone, about 3 km far from the İdecik Çeşmesi. in thinly-bedded limestones and cherty limestones *Daonella elongata, D. semicordiformis* and D. *moussori* have been determined suggesting Late Anisian to Early Ladinian age.

Konuk (1979) has found *Naticella acuticostata* in the lower parts of this formation, which indicates Late Scythian age. The paleontologic determinations from various parts as summarized above, indicate that the age of the Karareis formation ranges from Scythian to Late Anisian.

The Karareis formation has been deposited in a marine environment in which detrital input was very high. From time to time, pelagic limestones and bedded black cherts were formed in this basin. As olistostromal intervals and old limestone blocks of up to 500 m in length suggest, the depositional site was technically active and in this opening and foundering Early Triassic basin, mafic volcanism took place during the Anisian period. In Ladinian, the opening had stopped and a platform condition, in which a thick neritic carbonate succession builded up, had prevailed.

Gerence formation . _ If the Karareis formation is followed from the northwestern part of the peninsula, where it forms large outcrops, southward to the Gerence bay, it gradually changes into a unit composed dominantly of thinly-bedded limestones, marls, ammonilic red limestones, green-red cherts and sandstones. This carbonate-rich equivalent of the Karareis formation is called, in this study, the Gerence formation.

This formation is lithologically heterogeneous and its various facies have been named as formations by Brinkmann and others (1972) and the Domuzçukuru, Koyutepe and the Laleköy units have been differentiated in upward sequence. In our study, the Gerence formation encompases their three formations. The ammonitic red limestones, for example, which have been named as the Laleköy formation by Brinkmann and others (1972) are found repeatedly in different levels of the Gerence formation and thus can not be considered a separate unit but rather a change in lithology.

The Gerence formation forms large outcrops in two regions, one in the western side of the peninsula between Ildır and Gerence bay and the other as a belt starting from Balıklıova village extending northward from the western side of Küre mountain and ending near to Eylenhoca village.

This unit is composed of light-gray cherty limestones and light-gray marls. In the lower parts of the unit, there are sandstone intervals with plant fragments and red to green cherts. In the detrital facies of the Gerence formation, there are fossil-rich bioclastic limestone lenses. The detrital facies of this unit is confused in places with the Carboniferous Alandere formation. In the upper parts of the formation, ammonitic red limestone intervals are present. In the middle and upper parts of the unit, conglomerate horizons with limestone and chert particles are abundant These horizons laterally and vertically grade into massive limestones which are lithologically similar to the clasts of the conglomeratic intervals. They seem to be intraformational breccias 'inside the massive limestone. Around the Sıcakbük location, 4-5 km to the north of Balıklıova, interbeddcd with the conglomerate intervals there are massive reefal limestones up to 50 m in thickness composed entirely of coral, bryozoa and algae colonies. These reef and fore reef facies of the Gerence formation gradually pass upward to the massive limestones of the overlaying Camiboğazı formation as observed around Küre mountain.

The minimum thickness of the Gerence formation is about 150-200 m but it measures more than 500 m in the north of Balıklıova.

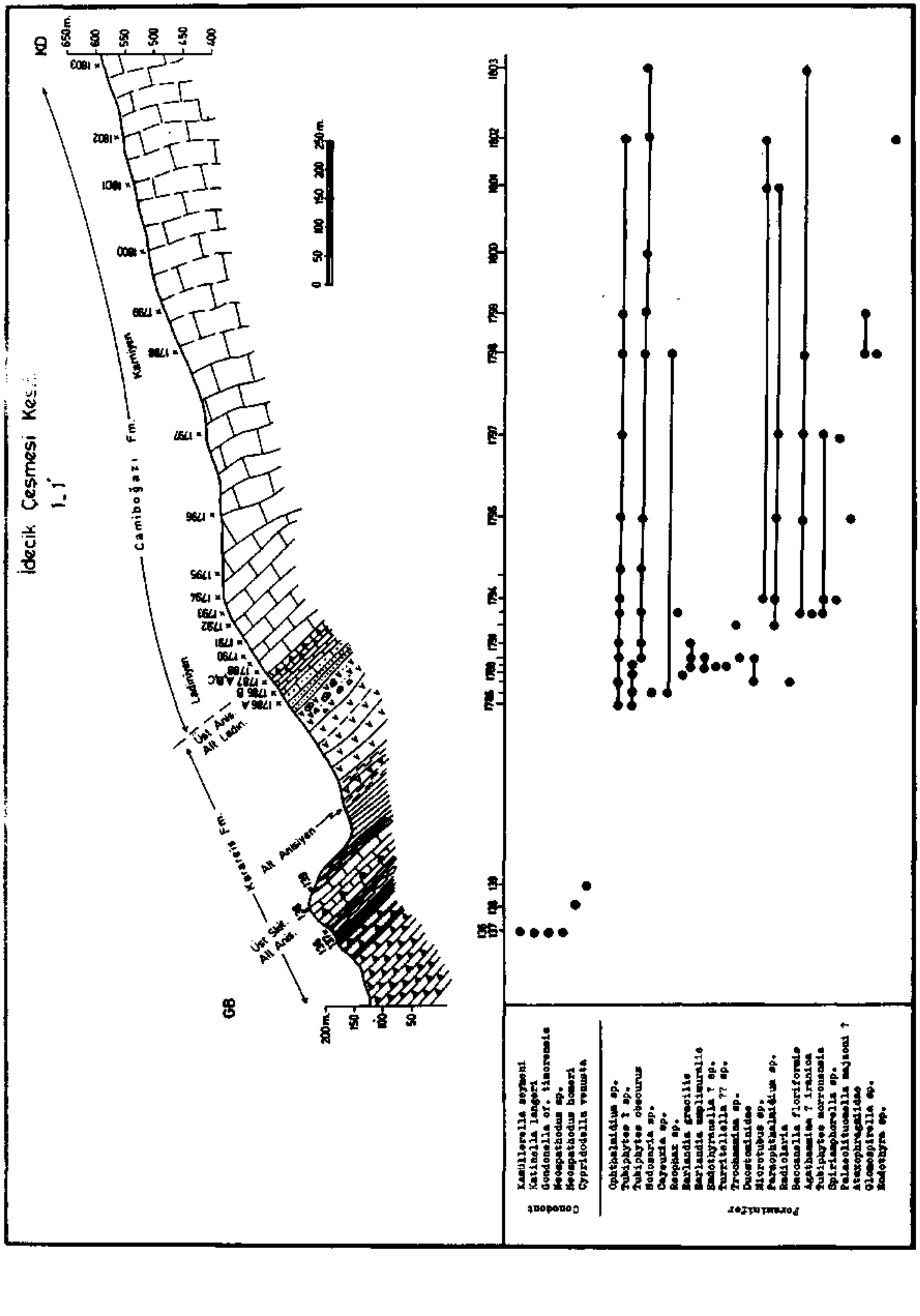

Fig. 3 – İdecik Çeşmesi section; Karareis formation lays below the Camiboğazı formation and the boundary between them is gradational. Sample numbers and fossil contents are given below the section. See Figure 2 for the location of the section.

This unit is laterally gradational with the Karareis formation and this relation can be observed along the road between Gerence bay and Küçük Bahçe village. At this location thinly-to moderately-bedded gray limestones of the Gerence formation change gradually along the strike to the red pelagic limestones with volcanic intercalations, bedded-black cherts and, the sandstones of the Karareis formation.

The Gerence formation overlies the Lower-Middle Carboniferous Alandere formation and this relation is described in that section. Along the boundary, although there is no angular unconformity, Upper Carboniferous and Permian are not present and in the contact zone karstic cavities are abundant.

On top of the Gerence formation, rests with a gradational relation, the Camiboğazı formation (Fig. 4). In between the two units, there is a zone characterized by thinly-bedded yellow limestones which gradually passes upward into the massive limestones of the overlaying unit. Similar gradational boundary is seen to the west of Küre mountain and the transition zone is 15-20 m in thickness.

The uppermost parts of the Gerence formation arc determined to be Late Anisian in age (Fig. 4). Below these parts, there are ammonite-bearing red pelagic limestones which have been determined by Brinkmann and others (1972) as Late Anisian by ammonite faunas.

The middle and upper parts of the unit arc observed in the Küre mountain section (Fig. 5) and their age arc probably Scythian. Near Ildır, close to the lower contact, samples have yielded thin-shelled pelecypods, echinid particles and *Ophihalmidium* sp., which may only suggest a range of Triassic.

The Domuzçukuru and Koyutepe units of Brinkmann and others (1972), which arc interpreted as different facies within the Gerence formation of our study, have been reported as Scythian and Late Anisian.

In summary, the age of the Gerence formation probably starts from the Scythian and continuous to the Late Anisian. Above this unit, along a transitional zone, that yields Late Anisian-Early Ladinian age, the Camiboğazı formation is present.

This unit was formed in a relatively deep marine environment in which cherts and ammonitic pelagic limestones were formed. Sporadic submarine volcanic activities had produced thin mafic tuff intervals. Upward in the unit, reefal massive limestones and intraformational conglomerate horizons arc present, that suggest a gradual shallowing of the dcpositional site. As the lateral interfingering relations with the Karareis formation suggest, from the southern carbonate-forming marine environment of the Gerence formation, it was passed northwestward into the tectonically active deep trench of the Karareis formation.

Camiboğazı formation

The Camiboğazı formation consists dominantly of light-gray massive limestones that form relatively sharp topography comparing to the surrounding units. It overlies both the Gerence and Karareis formations of the Denizgiren group. This unit was first named and described by Brinkmann and others (1972) and Gümüş (1971).

The most open outcrops of this formation arc seen to the cast of Ildır, near the Camiboğazı location (Fig. 4). From this area, the thickness of the unit increases to the north and it forms the Akdağ Range which is topographically highest parts of the peninsula (Fig. 2).

The lower parts of the unit arc veined with pink color and these parts were excavated for marbles in historic times. The Camiboğazı limestones appear to be lack of fossils with naked eyes but in thin sections, abundant algae colonies arc seen, and bioclasts of foraminifers, ostracods, crinoids and gastropods arc common.

At the type location, along the Camiboğazı section, the formation measures 400 to 500 m in thickness, but along the Akdağ Range to the north, it is more than 1000 m.

The lower, contact of this unit is gradational with the Gerence and the Kararcis formations (Fig. 2, 4 ,5) and the upper contact is conformable with the Güvercinlik formation (Fig. 4).

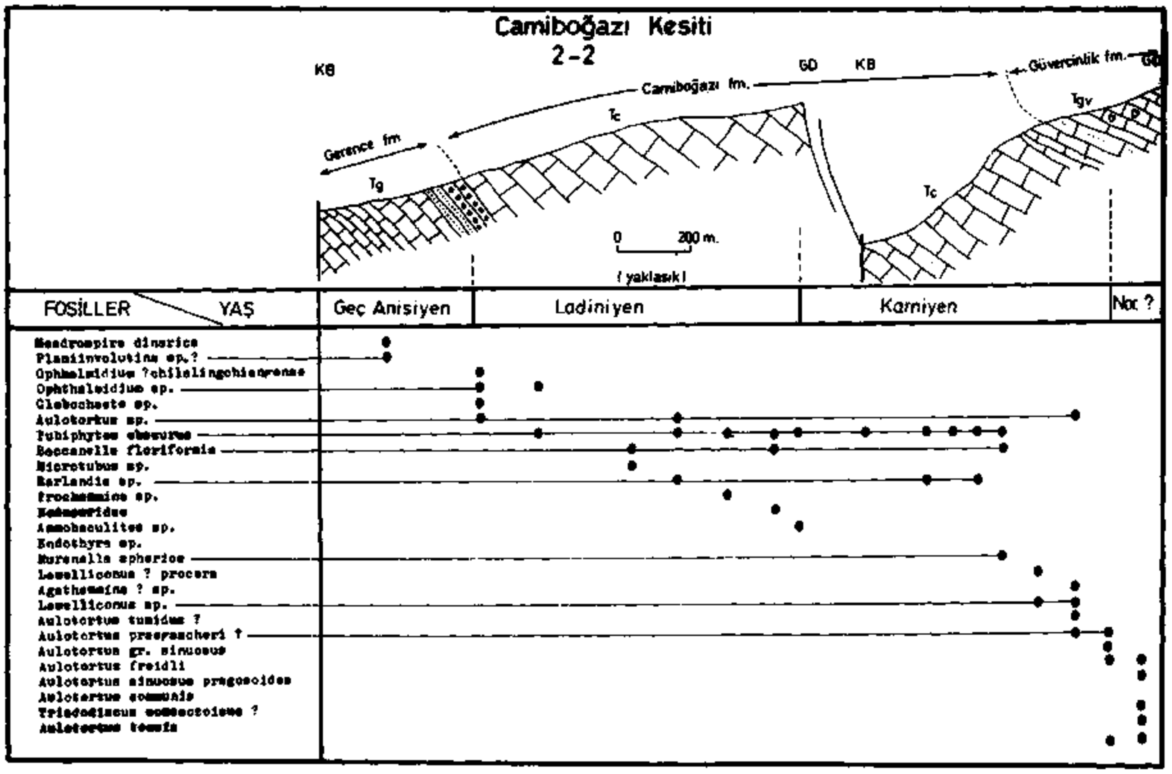

Fig. 4 - Camiboğazı section; Stratigraphic relations of the Camiboğazı formation with the Gerence formation below and the Güvercinlik formation above are seen. See Figure 2 for the location of the section.

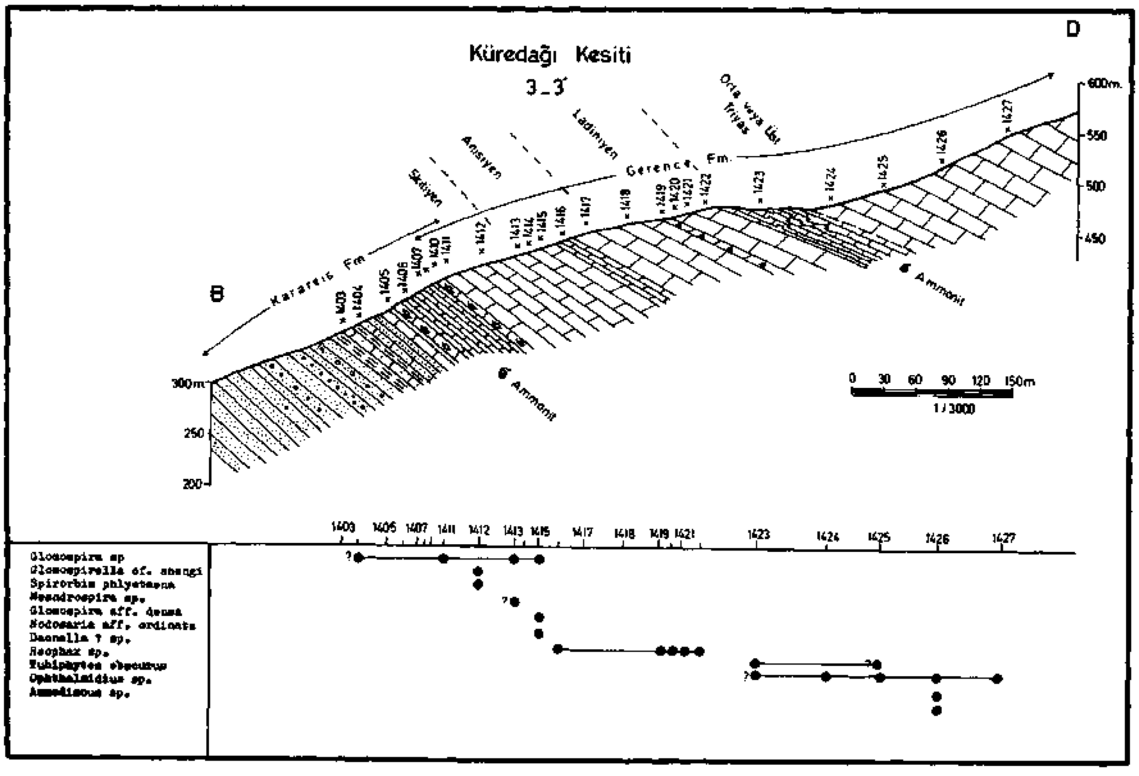

Fig. 5 - Küre mountain section; Stratigraphic relations between Gerence and Kararcis formations are observed. See Figure 2 for the location of the section.

The age of this unit ranges from Ladinian to Carnian, as observed in the Camiboğazı and İdecik Çeşmesi sections in Figure 3 and 4.

The limestones of the unit arc light gray in color and generally massive-bedded with abundant bioclasts and algae colonies which suggest a shallow marine depositional environment with sporadic patch reefs.

Güvercinlik formation

The Güvercinlik formation consists of intercalations of dolomitized algal stromatolites, megalodon-bearing limestones and quartzitic red sandstones.

In this study, this formation comprizes both of the Hanaylı and Güvercinlik formations of Brinkmann and others (1972). In their definitions, these workers have put dolomites under the Hanaylı unit and sandstone with dolomite intervals under the Güvercinlik unit. These different lithologies interdigitate along lateral and vertical directions and so that it is considered to be better to put them into one formation.

The open outcrops of this unit are seen along the road-cuts between Ildır and Barbaros villages and near the Tahtaiskele location, which is in the north of Balıklıova (Fig. 4,6).

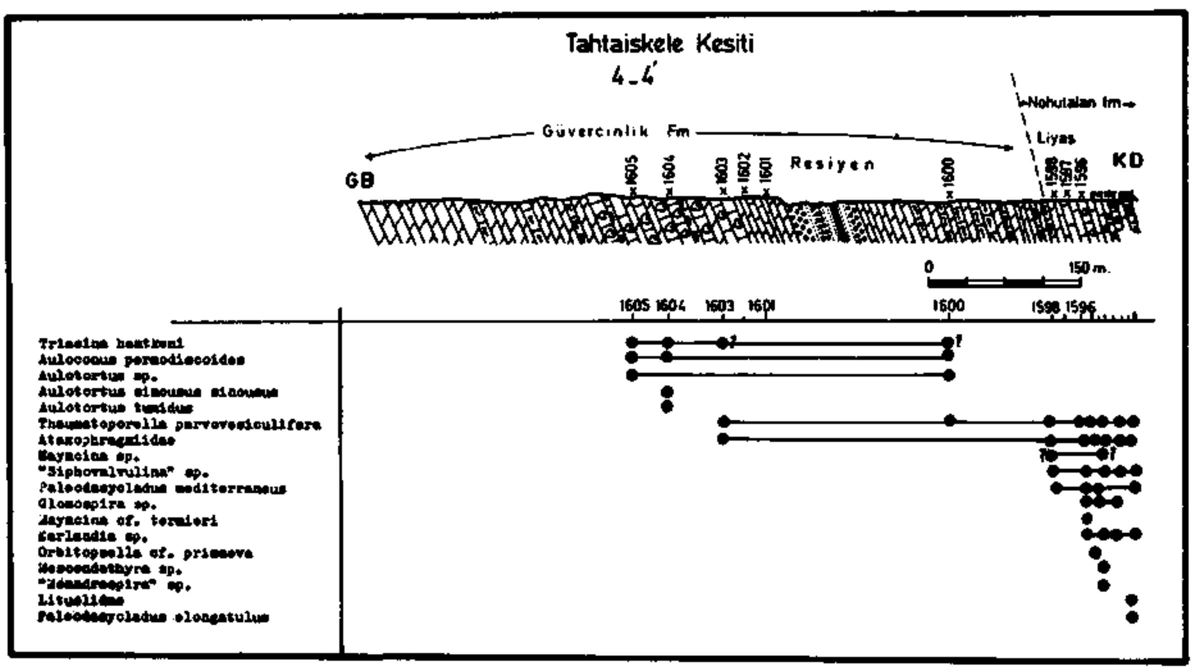

Fig. 6 - Tahtaiskele section; Stratigraphic relations between the Güvercinlik and the Nohutalan formations are observed. See Figure 2 for the location of the section.

In the lower section, the unit consists of thinly bedded, megalodon-rich yellow limestones. Upward in the sequence, laminated dolomites and algal stromatolites become dominant. Interbedded with laminated dolomites, there are 1 -5 m thick massive limestone intervals which contain big megalodons up to 30 cm in diameter. Associated with algal stromatolites, lithoclastic conglomerate horizons are present, and the particles of these intervals arc cemented by red carbonate and sandy materials. In the algal stromatolites, birds-eye structures are abundant and their voids are filled with calcite.

In the Güvercinlik formation, claystone intervals with red and green colors and quartzitic red sandstone lenses are seen and, they laterally interfinger with laminated dolomite horizons. In the claystone beds, loferite structures are present in places, that can be recognized with skeletal calcite vugs and with high porosity.

The lower contact of the Güvercinlik formation, as seen in Figure 4, is gradational with the Camiboğazı limestones. The upper contact is also gradational and at the Tahtaiskele location (Fig. 6) the Nohutalan formation rests conformably above this unit. Samples collected along two sections (Fig. 4,6) indicate the age of this formation as Norian and Rhaetian.

This unit was deposited in a shallow carbonate-depositing basin. The laminated dolomite intervals suggest a tidal-flat environment in which algal stromatolites deposited as primary dolomites. Intercalation of the algal stromatolites and massive megalodon-rich limestone intervals indicate sporadic change in the depth so that patch-reefs formed when the sea-level rised and the environment become suitable for prolification of the big megalodons. The loferite horizons in the green claystone beds suggest dominance of a evaporitic conditions from time to time and, lithoclastic conglomerates were formed from the disintegration of subaerially exposed dolomites.

Into the carbonate depositing sedimentary environment of the Güvercinlik formation, detrital materials were carried sporadically and the cross-bedded sandstone lenses were formed.

Nohutalan formation

The Nohutalan formation consists of well-bedded gray limestones and dolomitic limestones. This unit is equivalent of both the Nohutalan unit and "the Cladocoropsis limestones" of Brinkmann and others (1972).

In five different areas, sections have been measured and samples have been collected from this unit; three of the measured sections will be discussed in this paper. Mapping of the entire Karaburun peninsula has shown us that there is nolithological variation to divide this unit into two different formations as done by Brinkmann and others (1972).

This unit can be examined in open outcrops in an area to the north of Balıklıova, in the vicinity of Barbaros village and around the Urla İçmeler region (Fig. 2).

Along the Tahlaiskele section (Fig. 6), the lower part of the unit is seen, which is composed of thick-bedded gray limestones with abundant small megalodon fossils. Upward in the sequence, dolomitic limestones and well-bedded limestones become predominant. In the middle and upper parts of the formation, as seen along the Çatalkaya sections (Fig. 7,8),

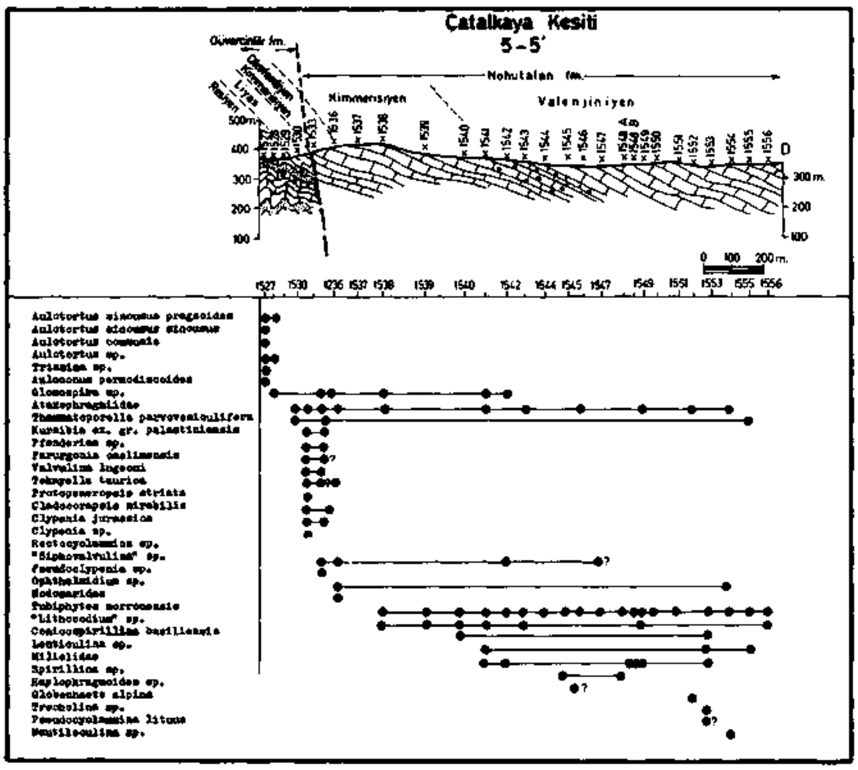

Fig. 7 - Çatalkaya section; Age of the middle and upper parts of the Nohutalan formation is observed. See Figure 2 for the location of the section.

there are chert nodules and beds within gray limestones. Along the İçmeler section in Figure 8, pelecypoda-rich horizons arc found that can be followed as key beds along their strikes. In the upper parts of this last section bryozoa, coral and algae-rich

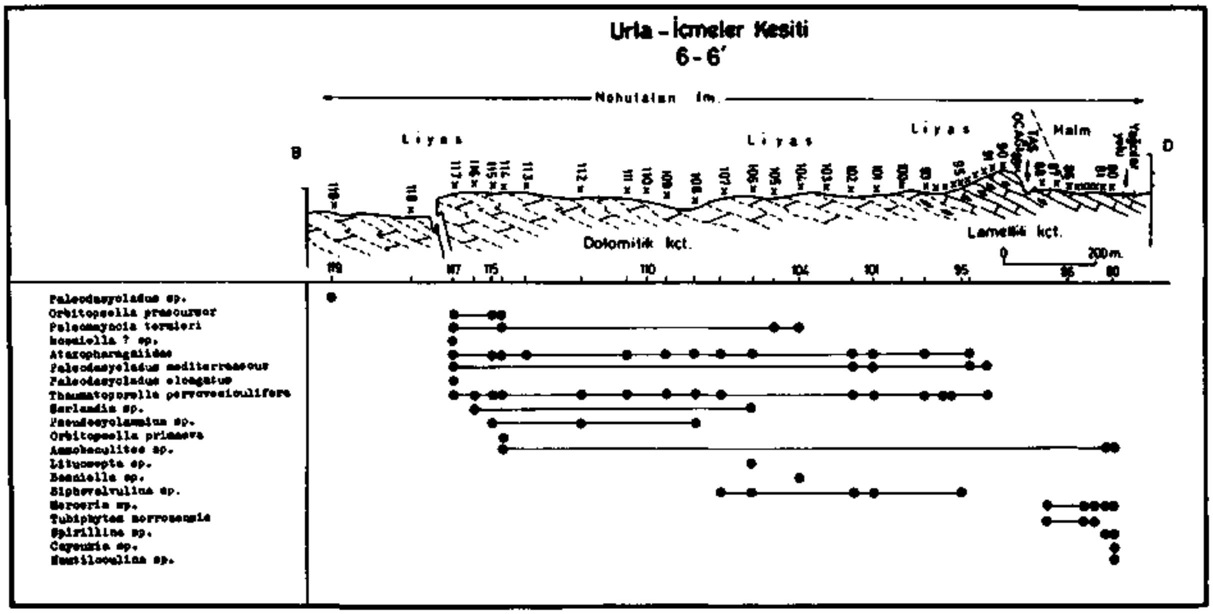

Fig. 8 - Urla İçmeler section; Lower and middle parts of the Nohutalan formation arc observed. See Figure 2 for the location of the section.

reefal limestones predominate. Along the Barbaros section in Figure 9, nearly the complete thickness of the Nohutalan formation can be seen. In the lower parts at this location dolomites arc common, in the middle parts buff limestones with abundant miliolids and gastropodas are found and in the upper part of the section there are laterally discontinuous bauxite pockets which are only a few meter in length.

The thickness of this formation is estimated to be more than 500 m. The lower contact is gradalional and can be seen along the Tahtaiskele section (Fig. 6). The Balıklıova formation of Campanian-Maastrichtian age rests unconformably above the Nohutalan formation.

The lower part of the Nohutalan formation yields the Liassic age as observed in Figure 6. Along the Çatalkaya, Barbaros and the Urla sections (Fig. 7,8,9) without any lilhological changes, the unit is represented by neritic shallow-marine limestones. In these three sections, Liassic Malm and Early Cretaceous ages have been determined, but Dogger is absent. In addition to these three sections, along two more which arc not shown in this paper, it has been observed a continuous carbonate succession from Liassic to Albian but still no characteristic fossils suggesting the Dogger age has been found. It is probable that there is a Dogger hiatus in the Nohutalan formation.

Our study shows that the age of the Nohutalan formation ranges from Liassic to Albian. However, in an earlier paper, Brinkmann and others (1972) have separated a Liassic Nohutalan formation and a Malm "Cladocoropsis limestone unit" although they have not mentioned any lithological distinction.

The Nohutalan formation deposited in a shallow marine platform environment. The lower part of the unit, similar to the underlaying Güvercinlik formation, was formed in a tidal-flat and a sublidal environment. In the middle part of the unit abundant bryozoa, coral and pelecypoda fossils arc found which form reefs in places. The uppermost parts of the unit, as the presence of bauxite lenses suggest, were formed in a very shallow carbonate-depositing platform area which was subacrially exposed from time to time. The subacrial conditions were probably short-lived and not extensive laterally, so that above the laterally discontinuous bauxite lenses, carbonate rocto with similar lithology to the underlaying part of the unit, were deposited. The bauxite lenses are located within the Nohutafan formation, not at the base of a overlaying Upper Cretaceous unit as proposed by Brinkmann and others (1972). Thus, they do not indicate a well-marked unconformity, but rather sporadic sea level changes during the Albian time.

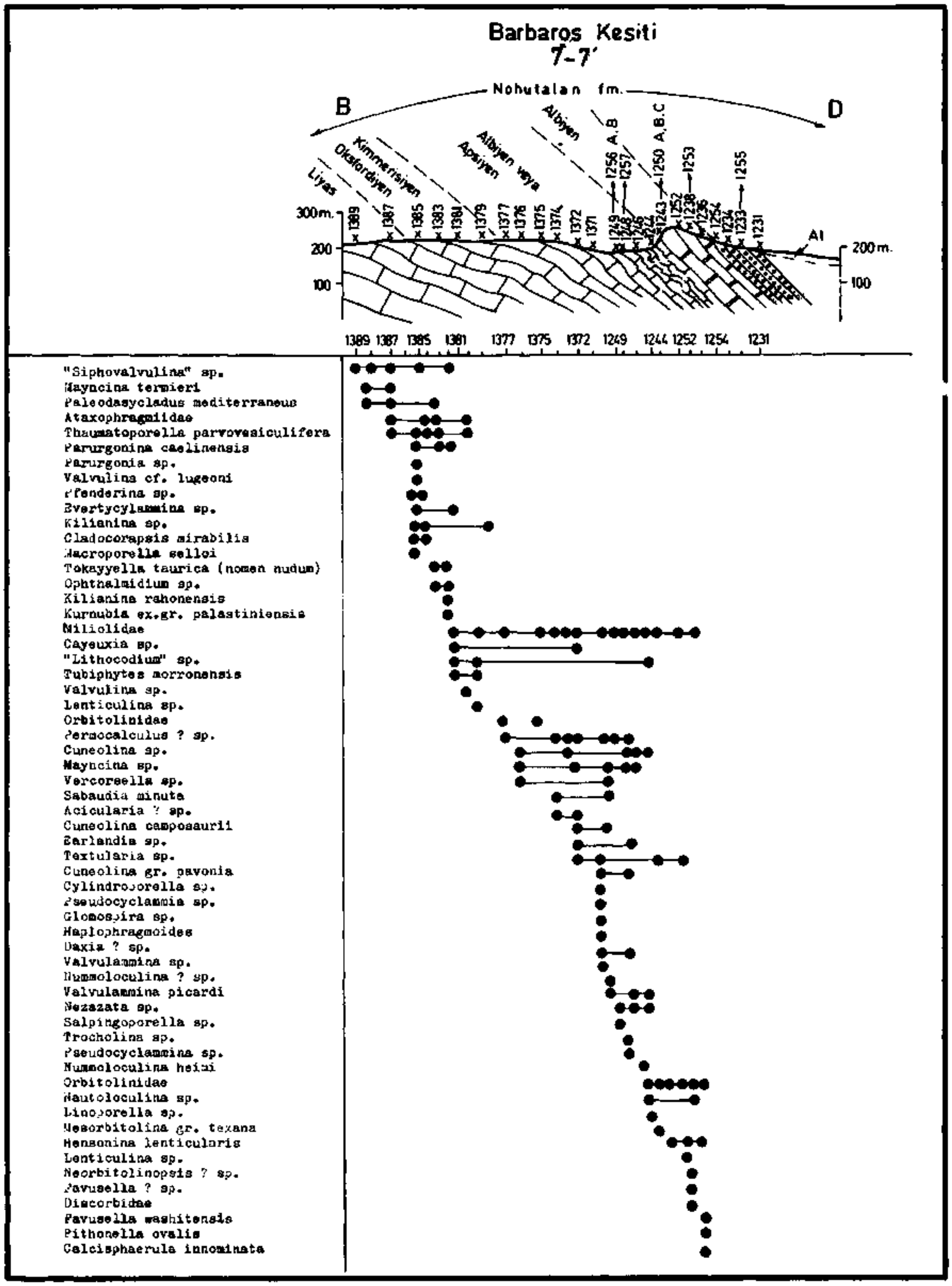

Fig. 9 - Barbaros section; Nohutalan formation is observed. See Figure 2 for the location of the section.

Balıklıova formation

In the Karaburun peninsula, above the continuous Mesozoic carbonate succession, lays along an unconformity the Balıklıova formation of the Campanian-Maastrichtian age. The stratigraphy of the Upper Cretaceous in the area was described by the first author in a separate paper (Erdoğan, 1990) and the tectonic relations between the Karaburun belt and the İzmir-Ankara zone was discussed. In this paper only a short summary of this formation will be given.

The Balıklıova formation consists of two members, which are named the Karahasan limestone and Haneybaşı member in ascending order. The Karahasan limestone is composed of thick-bedded gray limestones at the lower parts and pelagic cherty and thinly-bedded limestones in the upper sections, which together range from 5 to 100 m in thickness. The pelagic red-to pink-coloured uppermost part of the member gradually passes to the flysch-type detrital rocks of the Haneybaşı member above.

The Balıklıova formation overlays various parts of the Karaburun succession; in places directly resting over the Lower Triassic unit and in others over the Nohutalan formation of Liassic to Albian age. There is a marked angular unconformity at the base of this formation.

The Karahasan limestone is fossilifcrous and yields an age of Campanian-Maastrichlian, whereas the Haneybaşı member is Maastrichtian as determined from a rare thin carbonate lenses in the flysch-type sandstone and mudstone intercalations. The detrital rocks of this upper member of the Balıklıova formation measure at least 300 m in thickness, but the upper contact is a thrust fault.

The Balıklı3ova formation was formed in a shallow-marine carbonate-depositing environment at its lowermost part but as the lithological changes in the middle and upper part of the unit indicate, the shallow environment suddenly changed into a deep basin, in which first the pelagic limestones and later the flysch-type detrital rocks were formed. This sudden and fast subsidence of the basin might be related to the openning of the İzmir-Ankara zone that was probably reached on an oceanic extend from place to place in the western Anatolia.

The Karaburun succession formed the platform of the İzmir-Ankara zone and when it was openning, small and large slices of this platform were technically transported into this nearby basin, so that the blocky unit called the Bornova melange was formed. The Bornova melange includes large and small limestone blocks derived from the Karaburun succession, and it was formed during the Campanian-Danian interval, as determined from the age of the matrix of this chaotic unit.

This blocky unit crops out both in the northern and eastern parts of the peninsula, and the contact relations with the Karaburun succession indicate that, this belt is a large nappe in the chaotic unit and the tectonic transport occurred during the opening of the İzmir-Ankara zone (Erdoğan, 1990).

STRATIGRAPHIC AND STRUCTURAL EVOLUTION OF THE KARABURUN PLATFORM

The generalized stratigraphic section of the Karaburun succession, and the ages and lateral relations of the outlined rock units arc shown in Figure 10. The Alandere formation of the Early-Middle Carboniferous represents the oldest unit in the succession. After the deposition of the reefal limestones of this unit and before Scythian, the platform was uplifted and as the absence of the Upper Carboniferous and Permian suggests, these scries were either not deposited or erroded in the peninsula.

The Gerence formation deposited over the Carboniferous unit with the Scythian transgression and there is a hiatus at the base of this unit.

The basin, in which the Gerence formation was formed, continued as a deeper trench to the north of the peninsula and in this teclonically active tantional furrow, the pelagic limestones, bedded cherts, mafic volcanics of the Karareis formation were formed. In this relatively deeper basin, which was spreading and producing the mafic volcanic intervals, very thick mudstone and sandstone deposition took place that arc found nowinterdigitatingwith the pelagic sediments.

Due to this tantional spreading and opening in the northward direction, the Karareis formation had reached more than 2000 m in thickness while the equivalent Gerence formation measuring only 200 m in the south. Besides that, the presence of more than 100 m thick interval of the mafic volcanic rocks, olistostromal horizons with older limestone particles, and lime-

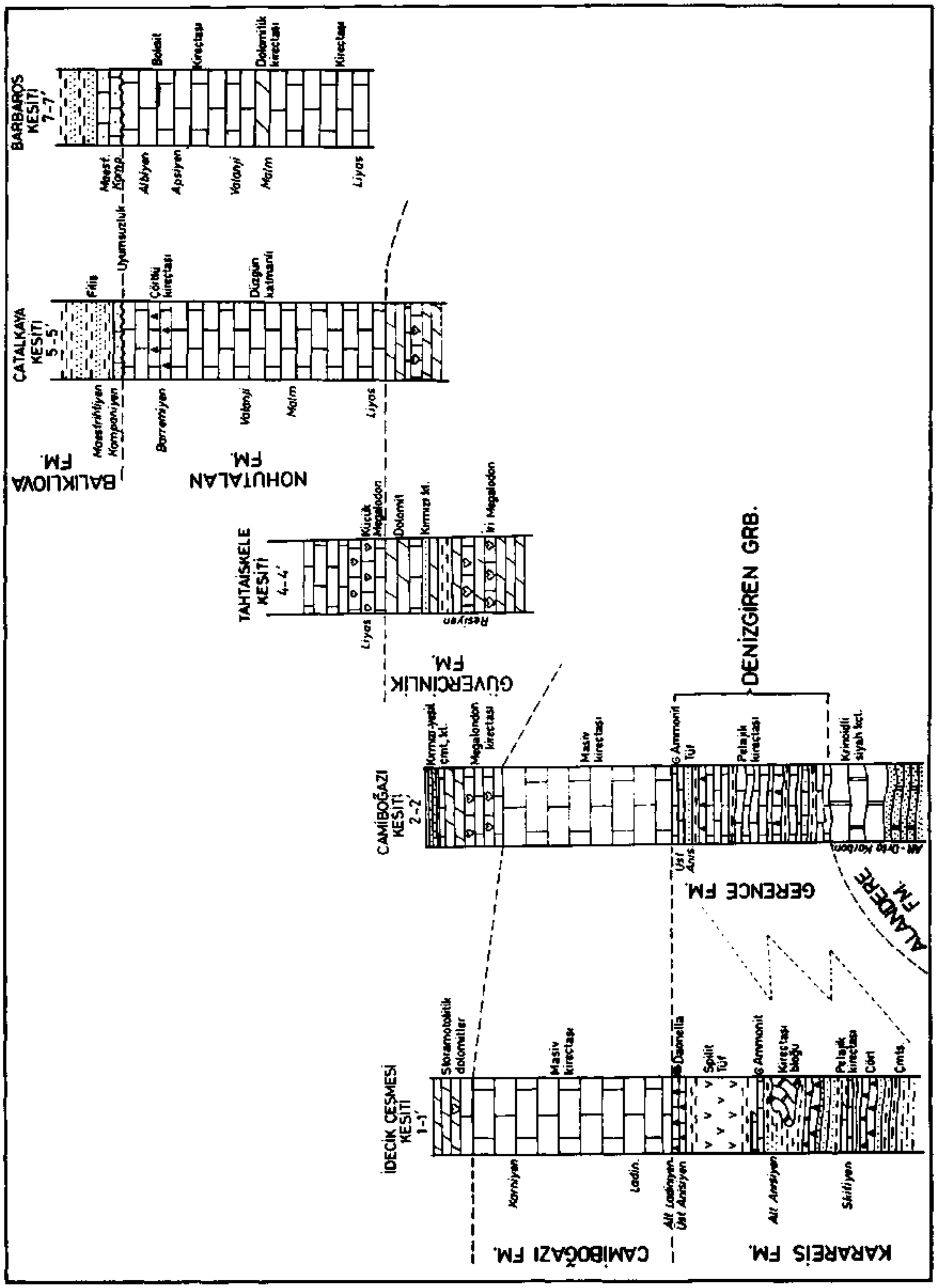

Fig. 10 – Stratigraphy of the Karaburun succession and correlation of the measured sections.

stone blocks of up to 500 m in length, are all indicative of continuous tectonic activities in the basin. Due to this synsedimantary tectonic activities, the chert intervals in the Karareis formation become brecciated and in places, formed chert olistostromes bound with mudstone matrix.

In the Gerence formation, short-distance facies changes are common and especially in its upper parts, the unit is represented by carbonate rocks of reefal facies that gradually pass upward into the Camiboğazı formation, that was deposited in a neritic environment.

During the Late Anisian and Early Ladinian, the spreading in the Karareis trench had seized and a platform condition had prevailed up to the Albian time. It is discemable that, the Early Triassic opening in the Karaburun platform took place on a limited extend and a relatively narrow aulacogen was only formed.

The Camiboğazı formation, which is Ladinian-Carnian in age, is represented by shallow marine limestones with abundant reef colonies. Overlaying this unit, the Güvercinlik formation, which consists of dolomitic algal stromatolites and megalodon-bearing limestones, were deposited. The lithological characteristics of this unit indicate a tidal-flat environment that turned to a evaporitic condition from time to time. Into this very shallow carbonate sedimentary environment, fluxes of detrital materials occurred and the sandstone intervals with cross-stratification were formed as discontinious lenses. In the region, this sedimentary condition was dominant from the Norian to the end of Rhactian. From the Liassic to Albian, in a shallow and sporadically reefal marine environment, the Nohutalan formation which consists of well-bedded limestones and dolomites, were formed. This unit contains horizons rich in algae, corals, bryozoa and pelecypoda colonies. The measured sections from this unit have yielded Liassic, Malm and Early Cretaceous (Valanginian to Albian) ages, but Dogger has not been determined, which may suggest a short-lived nondeposition during this interval.

During the Albian time, the Karaburun platform became very shallow and in a sporadic subaerial conditions the bauxite pockets were formed.

A tectonic activity took place in the region some time after the Albian and before the Campanian time and the angular unconformity at the base of the Balıklıova formation was formed. This tectonic activity was probably related to the initiation of the opening of the İzmir-Ankara zone which become pronounced during the Maastrichtian time. With this rapid subsidance, the Balıklıova formation that consists of a shallow marine limestones at the base and pelagic limestones and flysch-type sediments in the upper sections, was formed. This unit overlaid the various formations of the Karaburun succession.

Following this rapid subsidance in the Campanian-Maastrichtian interval, the Karaburun platform broke into large and small slices and was transported into the Izmir-Ankara zone. Some of these large slices measure up to 20 km in length in the sedimentary units of the Izmir-Ankara zone. In the final period of this tectonic deformation, the platform thrusted as a huge nappe into the nearby basin and surrounded by the blocky unit of the Maastrichtian-Danian age, which was called the Bomova melange.

The Bornova melange thrusted above the Menderes massif during the Late Eocene and it was carried on its back the already transported Karaburun nappe, and with this last episode, the paleotectonic evolution of the western Anatolia has ended.

Şengör and Yılmaz (1983) explained the tectonic evolution of Anatolia by opening and closing of the Tethyan ocean and delineated continents which controlled the large-scale deformational history of the region. They named the Sakarya continent to a small one located in the northwestern part of the Anatolia and stated that its southern boundary was first defined by Brinkmann (1966). In their map, they showed this continent to narrow toward Sakarya in the east and to continue southwestward enclosing the Karaburun Peninsula.

In the correlation of different tectonic belts, the most reliable attribute appear to be unconformity surfaces in the related stratigraphic colurhns. In the Karaburun succession, there is an unconformity between the Carboniferous and the Scythian units and above this horizon there is a continuous section including the Albian rocks on the top. In the stratigraphic column of the Sakarya continent, however, there is a marked angular unconformity at the base of the Liassic and with a basal conglomerate it overlies the deformed Triassic Karakaya formation or older units. Şengör and Yılmaz (1983) interpreted this Liassic unconformity with the closing of the Karakaya marginal basin.

The Karaburun stratigraphy can not be correlated with that of the Sakarya continent. There is no unconformity between the Triassic and Liassic sections and also no equivalence of tectonic deformation is noted which may suggest closing of a Karakaya-type ocean, in the same time-span. For this reason. It would not be correct to extend the boundary of the so-called Sakarya continent southwestward to include the Karaburun peninsula into the same micro continent.

The correlation of the Karaburun stratigraphy with the Menderes massif and with the Lycian nappes in the south is subject of a different paper in preparation.

CORRELATION OF KARABURUN STRATIGRAPHY OF THIS PAPER WITH THOSE OF EARLIER STUDIES

The stratigraphy of the Karaburun Peninsula was studied for the first time in detail by Brinkmann and others (1972). In examining Figure 10 and 11, their stratigraphic column can be correlated with this study. The Devonian-Carboniferous limestones of these workers are equivalent of the Alandere formation of our study which is found to be Early to Middle Carboniferous in age. The Domuzçukuru, Koyutepe and the Laleköy formations of them with an age range from Scythian to Anisian, are put together under the Gerence formation in this study and they are found to be laterally and vertically digitating facies changes in our formation. The Karareis formation, which is the lateral extention of the Gerence formation in the north-eastern part of the peninsula is not present in their stratigraphic column.

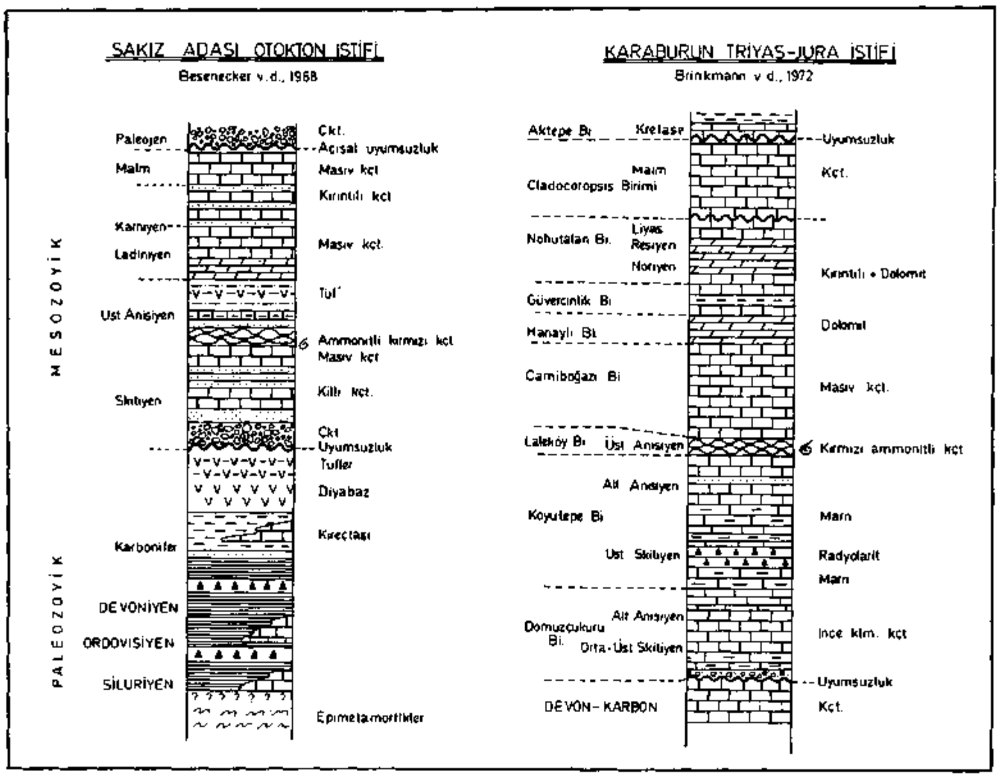

Fig. 11 — Stratigraphy of the Karaburun peninsula according to Brinkmann and others (1972), and stratigraphy of Chios autockthonous section accordng to Besenecker and others (1968).

The Camiboğazı formation of Brinkmann and others is the same in our study. The Hanaylı and Güvercinlik formations in their column are grouped under one unit in our study and called the Güvercinlik formation. We have also determined the age of this newly defined unit as Norian-Rhaetian. Similarly, the Nohutalan formation of our work is represented in their paper by two units named as Nohutalan unit and "Cladocoropsis unit". We have found that there is only one formation with an age range from Liassic to Albian. In this continuous section, we have not determined Dogger but still no lithological interruption or change have been noted.

In our study, the uppermost unit in the Karaburun succession is called the Balıklıova formation, which is separated into two members.

The Karareis formation in our column, which crops out in the northwestern part of the peninsula, is equivalent of the Karareis tectonic belt of Konuk (1979). We have found that it is only a stratigraphic unit below the Camiboğazı formation rather than a separate tectonic slice as proposed by this worker. We have mapped N-S trending reverse faults that cause rcpetaiion of the same stratigraphy from one compartment to another in the peninsula, but they are not found to be large-scale thrust faults juxtaposing completely different tectonic belts.

In our work, the Gerence and Karareis formations grade laterally into each other and are collectively called the Denizgircn group. Further detailed studies in the future may necessitate separation of members in these two formations.

In the Chios Island (Fig. 11) a similar stratigraphy with the Karaburun succession has been described in the autochthonous association by Besenecker and others (1968). The heterogeneous unit in the lower part of their stratigraphic column with lithologies of cherts, mudstones diabases and limestones, resemble closely to our Karareis formation of the Scythian Anisian age. However, they have indicated the age of this unit as ranging from Silurian to Carboniferous, which are probably given the ages of the included blocks. The similarity between these rocks in Chios with the Karareis formation has been first noted by Konuk (1979).

Likewise, the carbonate-rich unit of Besenecker and others (1968) with tuff intervals and red ammonitic limestones, that yielded the age of Scythian-Anisian, is similar to our Gerence formation.

In the same paper, the massive limestones of the Ladinian age, are similar to the Camiboğazı formation and the formation designated as the dethtal unit appears to be the equivalent of the Güvercinlik formation. The massive limestones with the Malm age in their column match with the Nohutalan formation of the Karaburun succession.

The correlation of the Chios Island and the Karaburun peninsula requires a field checking.

As discussed in the earlier sections, the stratigraphy of the Karaburun peninsula does not match with that of the Sakarya continent, so that in delineating tectonic belts of the western Anatolia this dissimilarity should be taken into account.

ACKNOWLEDGEMENT

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GEOLOGY OF THE SİVAS-ERZİNCAN TERTIARY BASIN

H.Tahsin AKTİMUR*; M.Ender TEKİRLİ* and M.Emin YURDAKUL*

ABSTRACT. - The Tertiary basin situated between Sivas and Erzincan in Eastern Anatolia display complex sedimentary and tectonic features. Munzur limestones lies to the south of the basin whereas, in the northwest lies the Karaçayır formation. Intrusion of the Refahiye ophioliiic complex took place between Lower Campanian-Lower Maastrichtian and are overlain unconformably by Upper Maastrichtian aged neritic carbonate sediments. These carbonates conformably pass into Paleocene- Eocene aged deep-sea sediments with olistostromal flysch characteristics. Oligo-Miocene and Lower-Middle Miocene aged elastics and carbonates unconformably cover the older units. A fore said elastics and carbonates were deposited in marine, lagunar and continental environmental conditions and the environs in the region are transitional laterally and vertically with each other. Upper Miocene aged evaporitic elastics unconformably cover the lower units. Plio-Quatemary is represented by continental deposits. Lateral movement of the Refahiye ophiolitic complex intruded in the region between Lower Campanian-Lower Maastrichtian took place intermittently and repeated few times with interruptions. As a result of lateral movement (transfer) of the Refahiye ophiolitic complex, a large pan of the Eocene aged Gülandere formation was thrusted over itself in north-south direction and pan of it was enclosed by the ophiolitic complex lower beds were thrusted over the Oligo- Miocene and Lower-Middle Miocene aged units, thus forming overturned folds with East-West strike. During Neotectonic period strike-slip faults of Tecer and Düzyayla were formed.

INTRODUCTION

The Tertiary basin situated between Sivas and Erzincan in East Anatolia, display complex sedimentary and tectonic features. This basin was studied by several researchers and valuable data were gathered (Baykal, 1953, 1966; Blumenthal, 1937; Okay, 1952; Kurtman, 1973; Arpat, 1964; Tatar, 1974; Akkuş. 1964; Norman, 1964; Pisoni, 1965; Demirmen, l%5; Bulut,1965; Koşal.1970; Sestini, 1968; Gökçen, 1976). However as these studies were not directed to the whole basin, satisfactory solutions could not be given to the important problems of the region.

Study area includes Sivas, Ulaş, Divriği, Kemaliye, Kemah, Erzincan, Refahiye, İmranlı and Zara regions (Fig.l). Tertiary basin situated to the north of Erzincan, Kemah, Divriği and Ulaş line is represented by deep sea, shallow sea.

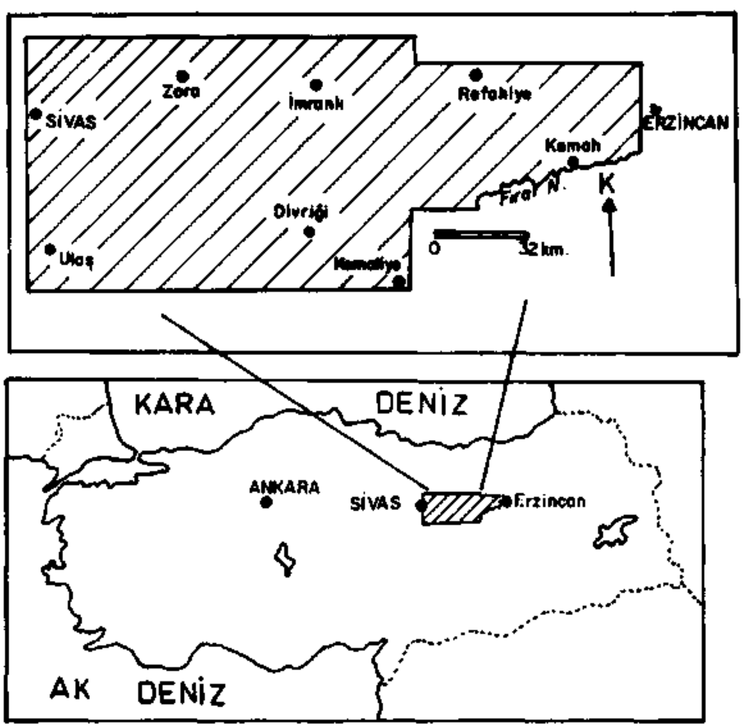

Fig. 1 - Map of the study area.

lagunar and continental sediments between Upper Maastrichtian; and Upper Miocene. The Basin to the south of this line is represented by continental sediments after Maastrichtian. As objective of this study is to exhibit basic geological characteristics of the Tertiary basin situated between Sivas and Erzincan units in the southern Basin were only briefly described. Information for this brief description were taken from the works of Aktimur, 1986; Aktimur and others, 1988; Aktimur, 1988 *a,b* and Tütüncü and Aktimur, 1988.

GEOLOGICAL SETTING OF SİVAS-ERZİNCAN TERTIARY BASIN

On the base of the Tertiary basin covering most of the study area lies the Mesozoic aged Munzur limestones in the south and in the northwest lies the Paleozoic-Mesozoic aged Karaçayır formation. The Refahiye ophiolitic complex which was intruded in the region between Lower Campanian-Lower Maastrichtian overlies these units with tectonic contacts. North Anatolian Fault borders the region in the northeast.

Refahiye ophiolitic complex are overlain by Upper Maastrichtian - Paleocene aged Tecer limestone and Çerpaçindere formation. This formation passes to Eocene aged Gülandere formation with olistostromal flysh character. Oligo-Miocene and Miocene aged marine, lagunar and continental detritics with evaporites and carbonates, are deposited.unconformably on all these units. Plio-Quatemary is represented by continental deposits and by volcanites (Fig.2).

PRE-TERTIARY STRATIGRAPHY OF THE BASIN

Karaçayır formation, Munzur limestone and Refahiye ophiolitic complex occur in the base of the Sivas-Erzincan basin.

The Karaçayır formation

The unit consists of white, brown coloured medium to thick bedded, foliated calc schist, quartz-epidote-albite schist, marble and quartzite conglomerate interlevels, developed in low degree metamorphism. It's thickness is 800 m (Yılmaz, 1981).

Index fossil could not be found in the unit But by comparing Tokat massive, accepted as Paleozoic in age by Yılmaz (1981), with the formation, the formation could be taken as between Upper Paleozoic-Lower Mesozoic in age (Aktimur, 1988).

Munzur limestone

The unit generally consisting of shelf type carbonates is located in the south of the Sivas-Erzincan Tertiary basin Fig.3). According to Özgül (1981), Munzur limestone from bottom to top includes neritic carbonates, oolitic limestones, flint, limestone, limestone with Lamellibranchiate and Gastropoda, reef limestone and pelagic limestone reaching 1200 m's thinckness, and it is deposited between Upper Trias-Upper Cretaceous. Refahiye ophiolitic complex overlies these with tectonic contacts (Özgül, 1981; Aktimur, 1986).

Refahiye ophiolltic complex

The unit, pan of which was also studied by Yılmaz (1975), outcrops in the environs of Erzincan, Refahiye, Kızıldağ, Çavuşdağ, Tecer dağı, Söğütlü, Divriği and in the base of Tertiary basin in the north foot of the Munzur mountains. Refahiye ophiolitic complex generally having mixtures of dark green, dark brown, dark grey coloured, hard, blocky disintegrated dunite, peridotite, serpentinite, amphibolite, gabro also has mixtures of: a) Metamorphites both in green and blue schist facies represented by multi coloured (light brown, grey, dark grey, dark green), thinly foliated, platy, altered, jointed chlorite-muscovite schist, epidote schist, granat schist, calc schist, serizite schist, actinolite-chlorite-albite schist and glaucophane schists; b) Multi coloured (white, whitish yellow, paleblue, dark grey) with breccia structure in lower pans; c) Pink, red coloured, hard, thin to medium bedded, folded radiolarites; d) Multi coloured (green, red, dark grey), in places pillowy structured spilite mixed in place by pelagic limestones; e) Multi coloured (light brown, dark grey, purple, dark red, dark green), pillowy structured, thick bedded, meta-volcanite outcropping with Jurassic limestones; f) Dark grey coloured, hard, jointed, blocky altered metadiabase; g) Dark grey, grey coloured hard, thick bedded, Triassic-Jurassic pseudosparitic limestones bearing fossils of. *Trochammina* sp., *Endothyra* sp., *Doustominidae* sp., *Involutina* sp., *Miliolipora* sp.; h) Cream, grey, light

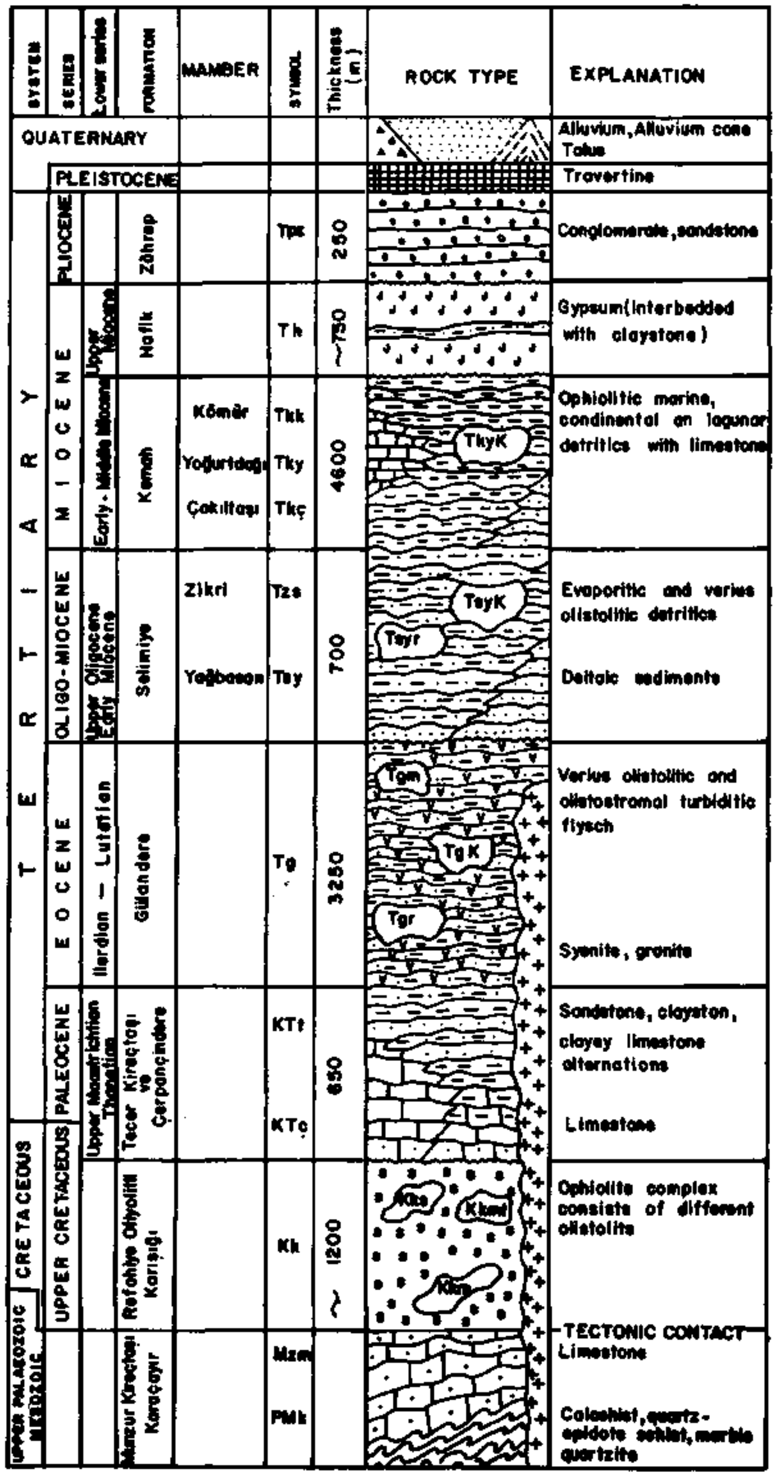

Fig. 2 - Generalized stratigraphic section of the Tertiary basin between Sivas-Erzincan.

gray coloured, thick bedded, calcite veined, colitic, siliceous Jurassic-Lower Cretaceous microsparitic limestone bearing fossils *of Trocholina* sp., *Textularia* sp., Pseudocyclammina, and *Robulus* sp.; i) Black-grey coloured, thin to thick bedded, silica-nodular, platy altered, calcite veined biomicritic Cretaceous limestones bearing fossils of Textularidae, Alg, Opthalmidiidae, *Orbitolina* sp., *Cuneolina* sp., Litualidae, Miliolidae, *Globotruncana tricarinata* (Quereau), *Globotruncana* sp., *Globigerlna* sp., *Globotruncana raselta* (Rarsey); j) Pink-red coloured, thinly bedded, pelagic limestones bearing fossils of *Globotruncana* sp., *Globigerina* sp. Gabros in place cut the rock ur. s they are associated with. Regular slip surfaces are seen in the ophiolitic complex. These surfaces where serpentinization and frequent crushed zones are seen are also important in mineralization aspects:

Refahiye ophiolitic complex was intruded in the region between Lower Campanian-Lower Maastrichtian (Özgül, 1981; Aktimur 1986). But repeated their movement few times until to the end of Miocene (Arpat and Tütüncü, 1978; Aktimur, 1986; Tütüncü and Aktimur, 1988). The unit, overlaying Munzur limestone and Karaçayır formation by tectonic contacts are overlain unconformably by Tecer limestone and Çerpaçindere formation.

TERTIARY STRATIGRAPHY OF THE BASIN

Sivas-Erzincan Tertiary basin in the region, begin with Upper Maastrichtan-Paleogene aged Tecer limestone and Çerpaçindere formation. Çerpaçindere formation gradually pass into Eocene aged Gülandere formation of olistostromal flysh character. Oligo-Miocene aged Selimiye formation and Lower-Middle Miocene aged Hafik and İslamkenti formations deposited unconformably on Kemah formation. Old units are all covered with Pliocene aged continental detritics (Fig.2).

Tecer limestone

The unit outcropping near Tecer and Gürleyik mountains are named as Tecer limestones by Blumenthal (1937). They consist of gray-dark grey, black coloured, medium to thick bedded, jointed, micritic textured limestones in places with alg and fossil fragments.

Thickness of Tecer limestone changes frequently and although seen in places with bedding outcrops of 50 cm's can also be seen with 700 to 750 m's thicknesses. The unit bearing fossils *of Cuneolina* sp., *Discocyclina* sp., *Alveolina* sp., Dasycladacea, *Volvulammina* sp., *Laffitteina bibensis* Marie (A form), *Laffitteina* sp., is aged as Upper Maastrichlian-Paleocene. Furthermore, in a detailed study in Tecer mountain; İnan and İnan (1988) determined the unit to be deposited between Upper Maastrichtian and Thanetian.

Tecer limestone deposited unconformably on Refahiye ophiolitic complex reached it's present location by secondary transfer of the ophiolites in Eocene (Aktimur, 1988). Therefore contact of both Tecer limestone and the underlying ophiolitic complex with the Eocene aged Gülandere formation in places is represented by slip surfaces (Aktimur and others, 1988). This contact was mapped as thrusts by other researchers. Laterally uncontinuous Tecer limestones discontinue under young deposits (Aktimur, 1988).

Çerpaçindere formation

The unit outcrops near Fıdıl mountain, Karadağ, Gülandere and Çerpaçindere. Formation is examined in two sections: Çerpaçin member having successions of conglomerate, sandstone, sandy limestone, clayey limestone and Karadağ basalt having andesite, basalt, tuff and agglomerates.

Çerpaçin member. - The unit comprising successions of sandstone, conglomerate, claystone, clayey limestone, sandy limestone and tuff is generally gray, green, reddish coloured, calcite veined and cross-bedded. Although variable, unit has thickness of 500-600 m's. In upper parts, Çerpaçin member passes gradually into Eocene aged Gülandere formation. This contact in some places are represented by slip surfaces.

In the examples collected from the unit from bottom to top fossils of *Vaccrites ultimus, Pseudopolyconites* sp., *Globorotalia* cf. *pseudobulloides, Globorotalia* cf. *compressa, Discocyclina* sp., *Heterohelix* sp., *Duberretia Kelleri* were determined. According to these fossils units age is Upper Maastrichtian-Paleocene.

Karadağ basalts. - Comprises andesites, basalts, tuff and agglomerates which were extruded in Çerpaçindere formation during deposition parallel to the deposition. Dominant rock is basalt and the unit outcrops near Karadağ.

The unit has a thickness of 250-300 m's and it's contact with Çerpaçin member is represented by a slip surface.

Gülandere formation

Gülandere formation described by Aktimur (1986) consists of the Kozluca formation, Bahçecik conglomerates, Bozbel formation and Kösedağ formation as distinguished by Kurtman (1973). This unit is also the equivalent of Akıncılar and Karataş formations described in the northeast of the basin by Yılmaz and others (1985).

The formation generally having sedimentary rocks of turbiditic flysh character consists of successions of sandstone, claystone, conglomerate, siltstone, tuffs and agglomerates. Unit having andesitic and basaltic lavas also bears olistostromal levels resembling blocks of Refahiye ophiolitic complex.

The formation streching in the Tertiary basin between Sivas and Erzincan in some places reaches a thickness of 4500m's. The multicoloured (gray, brown, red, green, blue), thin medium to thick bedded, folded, faulted and jointed unit deposited under different conditions in different localities, as indicated by is the abrupt ending of conglomerate and agglomerate levels, intraformational local discordances and repetitions caused by slip surfacess.

The Gülandere formation transgresses gradually to the Çerpaçindere formation which was deposited in deep sea environment suitable for flysh deposition. Horizontal movements of slip causing repetitions in the sequence and the andezitic-basaltic volcanism, is contemporaneous with deposition. These horizontal movements also dragged the large units (olistolites) broken from Refahiye ophiolitic complex to the deposition basin. According to the fossils of *Discocyclina* sp., *Nummulites* spp., *Ranikothalina* spp., *Alveolina* spp., *Cuvillierina* sp., *Sphaeergypsina* sp., *Globorotalia* sp., *Distichoplax biserialis* (alg) described from various levels of the formation, the units age is Lower-Middle Eocene.

Söğütlü conglomerate

It was defined as Söğütlü complex by Arpat (1964). But due to the unit having only conglomerate, it was named as Sögütlü conglomerate by the authors. In this study area the unit only outcrops in some places near Söğütlü where it is blackish green, red coloured and thickly bedded. Having a thickness of 50m's, pebbles and matrix of the Söğütlü conglomerate consist of serpentinites. Serpentinite cement is replaced by carbonate sediments in upper levels.

The unit overlaying Refahiye ophiolitic complex which was deposited in a mobile medium consists of fossils of *Nummulites* sp., *Operculina* sp., *Discocyclina* sp., *Orbitolites* sp., *Gypsina* sp. According to these fossils, the unit, is Eocene aged.

Selimiye formation

This formation overlaying unconformably Gülandere formation and gradually transgressing in to Kemah formation was first described by Kurtman (1973). Aktimur and others (1988), studied this formation by dividing into Yağbasan and Zikri members.

Yağbasan member. - The member outcropping near Ulaş, Beypınarı, Beydağı and Yağbasan is represented by successions of sandstone, claystone, mudstone, gypsum and conglomerate. It is multicoloured (red, gray, greenish, white), thin to medium bedded, jointed and in places overturned folded. The member deposited in shallow sea, lagunar and continental environs, varies in thickness. The unit having a thickness of 800-900m's begins with conglomerate in some places and with gypsum in others. It's contacts beginning with gypsums, generally include strontium.

The Yağbasan member was deposited in a mobile environment. The unit over thrusted upon itself by slip surfaces formed by the result of the movements which occured during deposition and these movements formed intraformational local discordances. Depending upon these conditions, olistolites of lower beds similar to Refahiye ophiolitic complex, entered the deposition basin.

Fossils of *Peneroplis* sp., *Operculina* sp., *Amphistegina* sp., *Textularia* sp., *Alveolina* sp., Miliolidae were gathered from the examples compiled from Yağbasan member. With these fossils a definite age can not be given. But the unit passing gradually into Lower Miocene aged Kemah formation is accepted as Upper Oligocene-Lower Miocene (Aktimur and others, 1988).

The Zikri member. - The member generally consists of red-dark red coloured, thin to thick bedded, conglomerate, sandstone and siltstone. Frequent cross -beddings and cross- lamination is observed and the unit has a thickness of 200m's.

Index fossils could not be found in the member which was deposited in deltaic conditions. However it was taken as Oligo-Miocene as it transgresses horizontally to Yağbasan member (Tütüncü and Aktimur, 1988).

Kemah formation

The unit was described by Özgül (1981) and generally consist, of conglomerates siltstone, sandstone and limestones. The formation have some blocky levels, as a result of over thrusting along slip surfaces and then by bending on itself, intra-formational local discordances were formed. Correlated with these characteristics by Selimiye formation, Kurtman (1973) defined the unit as Karacaören, unconformably overlying Selimiye formation.

According to various researches made in the basin, the Kemah formation, was observed to be transgressing horizontally to Selimiye formation and was unconformably covered by Hafik formation. Kemah formation was studied by the members of Çakıltaşı, Kömür and Yoğurtdağı (Aktimur and others, 1988).

The Çakıltaşı member. - The member mostly outcropping near Kemah and generally consisting reddish, green, in places grayish coloured, medium to thick bedded, clay and carbonate cemented, good sorted conglomerate and sandstones transgress vertically and horizontally to The Kömür member.

The Kömür member. - Outcrops near Kemah, Kömür, Kuruçay, Çengellidağ, Beydağı and Sivas. The unit formed by successions of sandstone, claystone, mudstone, clayey limestone and siltstone is multicoloured (red, yellow, white, gray, green) thin to thick bedded, folded, jointed, over turned folded in places, friable and soily altered. It also contains thin carbonate and coal levels.

The Kömür member deposited in continental, lagunar and marine environments has variable thicknesses. In places where repetitions caused by slip surfaces, are seldomly seen, thickness reaches 4500m's.

The unit having fossils of Lepidocyclina *(L. Eulepidina cliatata), Lepidocyclina* sp., *Miogypsina* sp., *Heterostegina* sp., *Miolepidocyclina* sp., *Textularia* sp., *Rotalia beccarii, Globigerinoides trilobus, Robulus vortex, Amphistegina* cf. *hauerina, Borelis melo, Asterigerina* sp., *Siphonina* sp., and *Ophthalmidium* sp. is aged between Aquitanian- Langhian.

The Yoğurdağı member.- Outcrops near Kemah, Çerpançindere, Kızıldağ, Çengellidağ, Beydağı, the unit formed by limestones is white, dark white coloured, medium to thick bedded, calcite veined, jointed and conchoidal fractured. Limestone in places sandy and clayey has characteristics of intrasparite and biomicrite with alg. Cement is sparry calcite. The Yoğurtdağı member deposited in shallow marine evironment transgresses horizontally to Kömür member. It's thickness is about 700m's.

The member having remains of echinoid and lamellibranchiate also has fossils *of Lepidocyclina sp., Miogypsina* sp., *Amphistegina* sp., *Textularia* sp., *Globigerina* sp., *Peneroplis* sp., *Miogypsina* cf. *globulina* and *Miolepidocyclina* sp. Age of the member is between Aquitanian-Burdigalian.

The Hafik formation

The formation was described by Kurtman (1973) generally outcrops near Kızılırmak valley and Kuruçay. The unit having dominant rock unit of gypsum is formed by white coloured gypsum and successions of multicoloured (red, dark red, green, blue) claystone and sandstone.

The Hafik formation, having a thickness of about 700-750m's, was deposited in a lagunar environment. The unit is overlain unconformably by Zohrep formation.

Fossils *of Amphistegina* spp., *Rotalia beccarii, Robulus vortex, Aurilina* sp., *Krithe* sp. and *Bairdia* sp. were detected in the Hafik formation (Aktimur and others, 1988). Formation is probably Upper Miocene (Tortonian) aged.

The İslamkenti formation

The unit outcropping in northeast of Kuruçay in the west of Erzincan and in the south of Refahiye generally consists of red coloured, medium to thick bedded, conglomerate having pebbles of 2 to 20 cm's in diameter and yellow coloured thin to medium bedded sandstone and tuffs.

Index fossil coult not be found in the formation which unconformably overlies the Kemah formation. But the units age is thought to be Upper Miocene because it is overlaying the Kemah formation and it's correlation with Pliocene aged Zöhrep formation (Aktimur, 1986).

The Divriği formation

The unit only outcrops in the south of Tertiary basin near Divriği. The formation generally starting with pink coloured, thick bedded, good sorted, clay cemented conglomerate is formed by the successions of clayey limestone, claystone, siltstone and sandstone. The unit seen in places with gypsum intercalations and having lake gastropod's and plant remains, is horizontally bedded. An index fossil could not be detected in the Divriği formation which has a thickness of 250m's. But formation is thought to be of Lower Miocene in age (Aktimur and others, 1988).

The Kavak formation

The unit outcrops in the south of the basin near Kavak and Ula$. The Kavak formation approximately having a thickness of 200m's consists of successions of conglomerates, sandstone, clayey limestones and is coaly in places.

An index fossils could not be detected in the unit which is horizontally bedded and deposited in calm lake environments. But with the formation overlain unconformably by Pliocene aged Zöhrep formation, it is thought to be of Middle-Upper Miocene aged.

Zöhrep formation

The unit is represented by grey-greyish coloured loosely cemented conglomerate and sandstones interbedded in places with clay and carbonate.

The formation having a thickness of 150-200m's overlies unconformably the Hafik and the Kavak formations. The unit having fossils of *Candona angulata G.* W.Müller, *İlgocypris gibba* (Ramd), *Condona* sp., *Valvata Atropidina* cf. *pulchella* (Studer) is Pliocene aged.

The Dumluca formation

The unit having pinkish gray tuffs at the bottom has dark coloured basalts at the top.

Due to tuffs, lateraly transgression to Zöhrep formation, units age is taken as PIio-Quaternary.

MAGMATISM AND VOLCANISM

Magmatism

A unit with rocks of syenite, granite and granodioritc, outcrops near Divriği, in the south of the Tertiary basin. This unit, which is a product of Eocene magmatism and having mineralized zones, effected the Munzur limestones, Refahiye ophiolitic complex, Çerpaçindere formation and Gülandere formations.

Volcanism

From time to time; In the stratigraphy section andesitic and basaltic volcanism products are mixed in to the depositional environment.

TECTONIC FEATURES OF THE REGION

In the Tertiary basin between Sivas-Erzincan four structural stages, namely the Pre-Upper Maastrichtian, the Pre-Upper Lutetian, the Pre-Tortonian and the Upper Miocene-Present were observed.

The Pre-Upper Maastrichtian stage

Between the Lower Campanian-Lower Maastrichtian the Refahiye ophiolitic complex overlies the Mesozoic aged Munzur limestone situated in the south of the Tertiary basin and the Paleozoic-Mesozoic aged Karaçayır formation situated in the north west of the basin (Özgül, 1981; Aktimur, 1986). After the intrusion of the ophiolite to the region, detritics of depositional period beginning with Upper Maastrichtian were developed by post-tectonic events. The overlies the lower units with angular unconformity.

The Pre-Upper Lutetian stage

After the intrusion of the ophiolite, in the region, in the beginning, a shallow basin, where rudist could live, was formed. Later this basin was transformed into deep sea basins, suitable for flysh deposition. This basin is probably a depression in the front of an ophiolite nappe advancing from north to south. The carbonate deposition coritinued in the basin's margins between Upper Maastrichtian-Thanetian period (Tecer limestone).

Together with the development of the above-mentioned basin, and as a result of significant gravity slides an olistostrom carrying all kinds of olistolites from the formerly deposited formations in the vicinity began to develop. In the end Refahiye ophiolitic complex in the region was again thrusted probably in Lutetian. As a result of this over thrust Tecer limestone together with the underlaying ophiolites moved in to the Eocene basin to be located at its present location and most of the Gülandere formation was over thrusted over itself by sliding in the north-south direction and a part of the formation was incorporated in the ophiolitic complex.

The Pre-Tortonian stage

At this stage, Upper Oligocene, Lower-Middle Miocene aged marine, lagunar, continental detritics and carbonates were deposited unconformably on the lower series. The units deposited in all three environments transgresses vertically and horizontally with each other and processes with characteristics resembling the ones of Pre-Lutetian stage were also developed at this stage. Third over thrust of the ophiolitic complex started probably between Aquitanian-Burdigalian and entered the deposition basin as an olistolite. Rock associations in the basin over thrusted over themselves resulted in repetitions. In the end, Refahiye ophiolitic complex and with the successions deposited between Maastrichtian Lutetian, thrusted over Oligo-Miocene and Lower-Middle Miocene aged deposits along east-west trends and caused, over turns in the northern limbs of the Miocene synclines.

The Upper Miocene-Present stage

At this stage Upper Miocene aged marine, lagunar, continental detritics and carbonates were deposited unconformably on the lower units and Pliocene aged continental detritics were also deposited unconformably on these. As a result of probably Pre-Pliocene aged North-South compression, Upper Miocene aged Hafik formation was folded with east-west strike. After these events, transform fault of North Anatolia and East Anatolia were formed and started the movement of Anatolian continent in the west direction (Şengör, 1980; Şaroğlu and others, 1987). All these events caused to form strike slip faults of Tecer and Düzyayla and travertine deposition started as a result of these faults (Fig. 3).

CONCLUSIONS

Ophiolitic complex was intruded in the region between Lower Campanian-Lower Maastrichtian. But repetitive over thrusting of the ophiolitic complex with interruptions continued until the end of Lower Miocene.

The vertical transgressing of the Upper Maastrichtian-Paleocene-Eocene aged detritics overlying ophiolitic complex were determined.

Oligo-Miocene and Lower-Middle Miocene aged detritics with carbonates deposited in marine, lagunar and continental environmental conditions were determined to be transgressing vertically and horizontally in themselves.

In Pre-Tortonian stage the lower units were thrusted upon Oligo-Miocene and Lower-Middle Miocene deposits in the east-west direction.

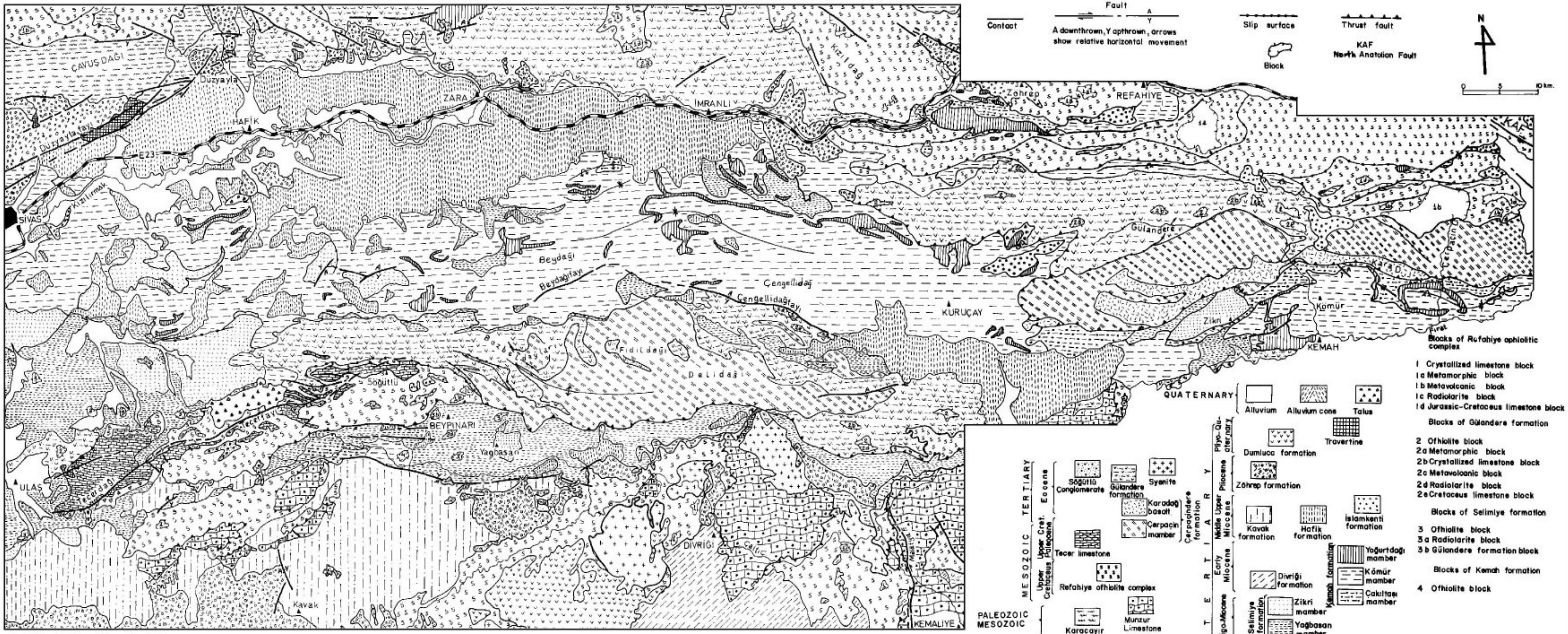

Fig. 3 - The geological map of between Sivas-Erzincan

Land forming in the region starting partly in Upper Lutetian, completed it's development in Upper Miocene.

Strike-slip faults of Tecer and Düzyayla were formed during neotectonic period and travertines relating to these faults were deposited.

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ORIGIN OF THE CONCRETIONARY LIMESTONES IN THE ISTANBUL DEVONIAN SUCCESSION

Mehmet ÖNALAN*

ABSTRACT. - Nodular limestone occurrences are widely present in the Devonian - aged sediments of the İstanbul region. This nodular units have been developed in the blue-grey coloured micritic limestones and marls alternation. The occurrences seen in the Lower-Middle Devonian strata are larger and the ones observed in the Upper Devonian sediments are smaller. The laboratory and field findings indicate that the nodules were developed within the. limestone-marl intercalation in which the thickness of the individual beds varies from 2-3 mm. to 5-6 cm. during the early diagenesis compaction, and loading and, later, by means of presure-solulion effects and, in part by tectonic deformation. No effects of organic activities and transportation have been detected. The size, shape and the order of the nodules are found depending on the thickness ratio of the individual beds in the limestone-marls intercalations.

Mineral Res. Expl. Bull., 111, 31-42, 1990

THE FAULT TRACE OF 1953 YENİCE-GÖNEN EARTHQUAKE AND THE WESTERNMOST KNOWN EXTENSION OF THE NAF SYSTEM IN THE BİGA PENINSULA

Erdal HERECE*

ABSTRACT.- Northeast-striking en-echelon faults in the Biga peninsula represent the westernmost known extension of the NAF system. The main faults exhibit a clockwise en-encelon pattern from Manyas-Danişment in the east, Yenice-Gönen in the center and Sarıköy-İnova in the west The NAF system in eastern Turkey developed about 11 my ago and has accumulated a total dextral separation of about 30-40 km. The apparent decrease in displacement to 8 km in the southern segment east of the Marmara Sea corresponds to a change from a single to multiple fault zones. The nature of grabens, as pull-apart basins, during Late Pliocene to Quaternary time in an essentially strike-slip regime has been quantified in the Bursa-Gönen graben apparently moved 3.4 km west-southwest along these faults, while east-west normal faults developed in the graben that account for some 8.2 km of north-to-south extension. No data are available for displacement on the Sarıköy-İnova and Manyas-Danişment faults.

INTRODUCTION

There has been considerable debate over westward continuation of the North Anatolian Fault (NAF) zone east of Marmara Sea. From western termination of the Mudurnu valley through westward; some workers believe that the NAF zone consists essentially of a single strand (Alptekin, 1973; Bingöl, 1976) while others indicate that it sprays into two (Şaroğlu et al., 1987) or three active strands (Dewey and Şengör, 1979; Şengör and Canıtez, 1982; Şengör et al., 1983, 1985; Hancock and Barka, 1983). Some other workers, however, point out that the NAF zone branches and loses its continuity east of Marmara Sea (Koçyiğit, 1984; Kıyak, 1987) or the right-lateral slip of the fault is distributed over a number of grabens with a large component of normal faulting (McKenzie, 1978; Jackson and McKenzie, 1984; Crampin and Evans, 1986).

The Gaziköy-Saros bay fault ** is generally known as the northern arm of the NAF system. The fault extends towards west through the Lake Sapanca, İzmit bay and northern Marmara trough and reaches to northern Aegean trough of the Saros bay. Another branch of the NAF, the middle branch, also extends toward west from the Lake İznik and Gemlik bay to the southern shore-line of Marmara Sea. In addition to the northern and middle branch of the NAF, the third branch from Bursa to Yenice, has been considered. This branch is the result of the structures induced by the 1953 Yenice-Gönen and 1964 Manyas earthquakes.

The main purpose of this paper is to discuss the structural importance of the 1953 Yenice-Gönen earthquake in the regional neotectonic events, based on the field observations along the main earthquake fracture and its northeastward and southwestward extension.

THE FAULT TRACE OF 1953 YENİCE-GÖNEN EARTHQUAKE

A destructive earthquake occurred between Yenice and Gönen on March 18, 1953. The shock was felt over a large area in northwest Turkey. The epicenter of the shock was located approximately 12 km east of Yenice, the magnitude of P wave was 7.4 and focal depth determinations varied from 10 to 12 km (Herece, 1985).

A 50 km-long surface break formed during earthquake. From a study of the aftershocks, Pınar (1953) and Ketin and Roesly (1953) investigated the epicenter area in detail. The Yenice-Gönen fault zone, which includes the fault trace of 1953 event, has been investigated in detail for the first time by Herece (1985).

The fault trace bf 1953 Yenice-Gönen earthquake is shown in Figure 1. The break begins in andesite to the west of Çakmak, then proceeds west displaying continental Neogene and Quaternary deposits to the east of Koru değirmeni, where the fault scarp is still observable. During the earthquake the southern block was displaced approximately 80 cm downward along this fault scarp. The fault can be traced through the alluvial plain of Gönen and westward into the Neogene deposits north of Muratlar. The fault trace is not visible in the alluvial plain. However, approximately 2 km west of the Neogene de-

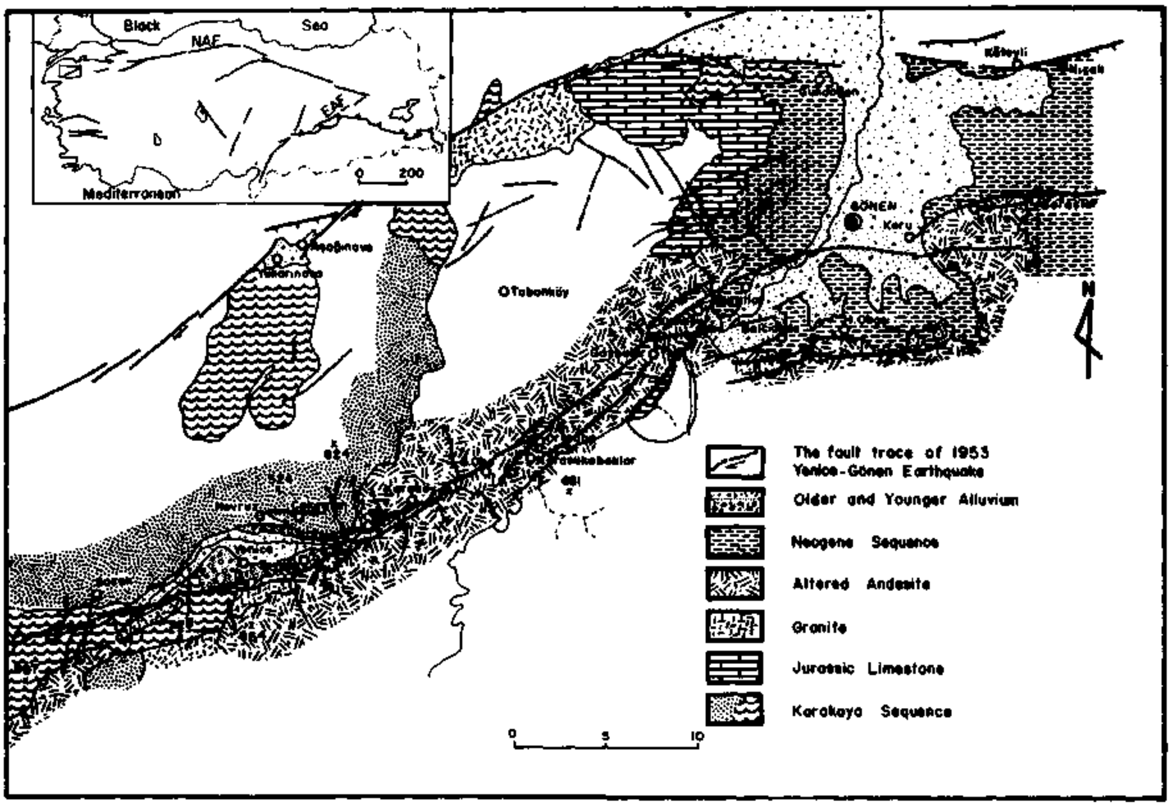

Fig.1- The fault trace of 1953 earthquake between Yenice and Gönen.

posits and the road between Yenice and Gönen, the fault break apparently is visible in the poplar forest and thickets. The trace continues in Neogene deposits trending approximately west-southwest about *2.5* km to the weathered andesites, northwest of Muratlar.

North of Muratlar and Kumköy villages the fault trace is not distinct due to high erosion rates on the mountain slopes and agricultural activities in the valleys. An excavation for dam construction, however, exhibits the fresh fault trace in weathered andesites, and northwest of Gaybular it cuts a small Jurassic limestone outcrop. The fault trace again displaces andesite in continuing southwest from north of Oba. The fault trace there occurs on the north-facing slope of a mountain where the fault scarp apparently developed because of normal dip-slip displacement of the southern block during the earthquake. It forms the boundary between an alluvial plain and andesite east of Çakır.

Northeast of Çakır the fault trace continues for the most pan between alluvium and andesite along the southern border of the valley. It extends to the southwest, cutting the road between Yenice and Kalkım and continues south-southwest of Seyvan.

Between Seyvan and Yenice the fault trace is well exposed as a scarp appoximately 1 m high. Here, the fault break indicates that, in addition to dextral movement, the northern block was displaced downward during the earthquake. The fault trace continues to the west-southwest in the andesite as a very distinctive scarp. It then enters granite, changing its southwesterly direction to westerly south of Yenice (the old town of Yenice originally was located on the fault some 800 m from its present site). The distinct fault trace then makes a southwesterly bend to the south of Yenice and extends through the valley south of Sazak. Here the fault trace marks the southern border of the valley, which is underlain by metamorphic rocks. After almost 4 km, the fault crosses diagonally to the northern side of the valley, displacing the Quaternary deposits in the valley. The fault trace in the northern side of the valley strikes south-southwest into metamorphic rocks for about 5.5 km, and seems to terminate somewhere between Sazak and Zeyberçayırı.

In the more rapidly eroding areas such as the weathered andesites and the rougher terrains associated with higher elevations not all of the faults could be mapped because scarps had been removed by erosion. Mapping faults between Yenice and Gönen was rendered even more difficult by the dense forest cover at higher elevations, and contour plowing and cultivation of the lower slopes. As a result of these factors many fault traces were not continued laterally.

During the mapping of the 1953 earthquake fracture zone between Yenice-Gönen a number of scarp in different stages of erosion were noted. These scarps indicate multiple faulting events during the Quaternary (Herece, 1985).

I would like to discuss about northeast and southwest extension of the Yenice-Gönen fault zone which reactivated by the 1953 Yenice-Gönen earthquake. Neotectonic studies on the northeast extension of the fault zone is utilized by the fault trace developed during 1964 Manyas earthquake. The northeast extension has been considered as a right-lateral strike-slip fault with a minor dip-slip component after the 1964 Manyas earthquake (Ketin, 1966). The field works conducted by Erentöz and Kurtman (1964) and Herece (1985) and the fault plane solutions utilized by McKenzie (1972) and Öcal et al. (1968), however, indicate that the northeast extension of the fault zone is not right-lateral strike-slip motion but it is normal dip-slip movement. This conclusion supports the consideration of the closed depression area between Bursa and Gönen is being a graben (Herece. 1985; 1988).

BURSA-GÖNEN GRABEN

The location of Bursa-Gönen graben is shown in Figure 5. This closed depression begins east of Bursa where it is connected to the NAF system by northeasterly lineaments and closes around Sarıköy and Gönen in the west The graben is as much as 160 km in length and 5-30 km in width.

The depression is bounded by fracture system of different directions to the north and south. In the eastern part of the graben northeast rupture directions predominate while west-northwest trends are seen mostly around Manyas lake in the western sector. The main faults along the borders appear to be parallel in the area between Karacabey and west of Bursa. Both the northern and southern escarpments here trend northeast, but within the graben and in the easternmost and western sectors, as well as on their edges, numerous tilted segments were formed with different orientation.

The graben, in genera, subsided along inward-dipping faults to form a down-faulted, wedge-shaped block (Ekingen, 1972). The main faults along the margins of the graben are normal dip-slip (Fig. 2). Although dips of approximately 70° are indicated on the gravimetric cross-section, scarps on the graben flanks show widely different slope angles due to mass wasting denudation, erosion, and rejuvenation (Herece, 1985).

The elevation of the graben floor is about 0 to 50 meters above sea level. A broad arc may have developped concurrently with graben formation because the shoulders of the bordering block have gently outward dipping slopes. Relief is variable with maximum elevations of 250 and 680 m for the shoulders in the eastern sector and 350 to 450 m in the west. Amounts of subsidence in the graben area apparently differ with shoulder elevation and with thickness of the Quaternary deposits. Present morphotectonic features of the graben have a similar profile to the asymmetric depression (Herece, 1985).

When and how the depositional basin in the Bursa-Gönen graben formed is questionable. Is the stress system (s) that caused the northeast- southwest and northwest-southeast trending faults visible on the Landsat images also responsible for forming this depression? What king of tectonic environment formed this closed basin? Is the basin a dropped keystone block or graben, an extensional depression on listric normal faults, or a "sag-pond" type block associated with an earlier strike- slip regime similar to that prevailing along the NAF today?

Unfortunately, a lack of adequate field data on the Neogene sequences within and around the Bursa-Gönen graben precludes an answer to these questions.

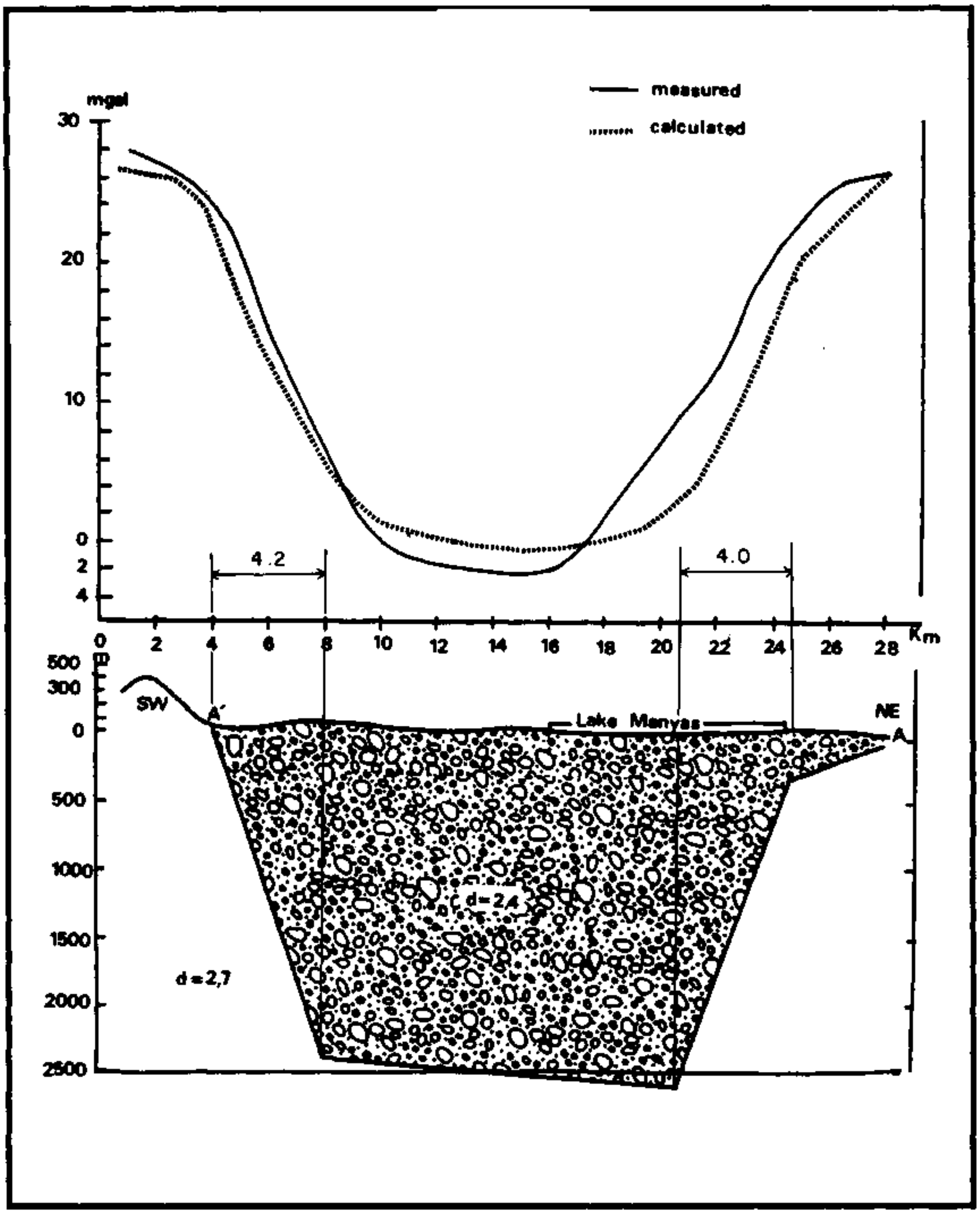

Fig. 2– The separation in the Bursa-Gönen graben, deduced from a gravimetric model.

AGE OF THE NEOGENE SEQUENCE

The age of Neogene sequence in northwest Turkey is questionable because of lack of correlation of Tethys and Paratethys depositional environment. Attemp at correlation of Tethys and Paratethys rocks involves radiometric dating of units interlayered with mammalian faunas in these local sequences (Mein, 1975; Fahlbush, 1976). The Tethys stages have been dated radiometrically (Vaas, 1978) and paleomagnetically (Ryan et al.. 1974; Vaas.1978) and, therefore, can be used for correlation. A tentative correlation of Tethys and Paratethys stages is given in Figure 3 for Neogene deposits on the Biga peninsula.

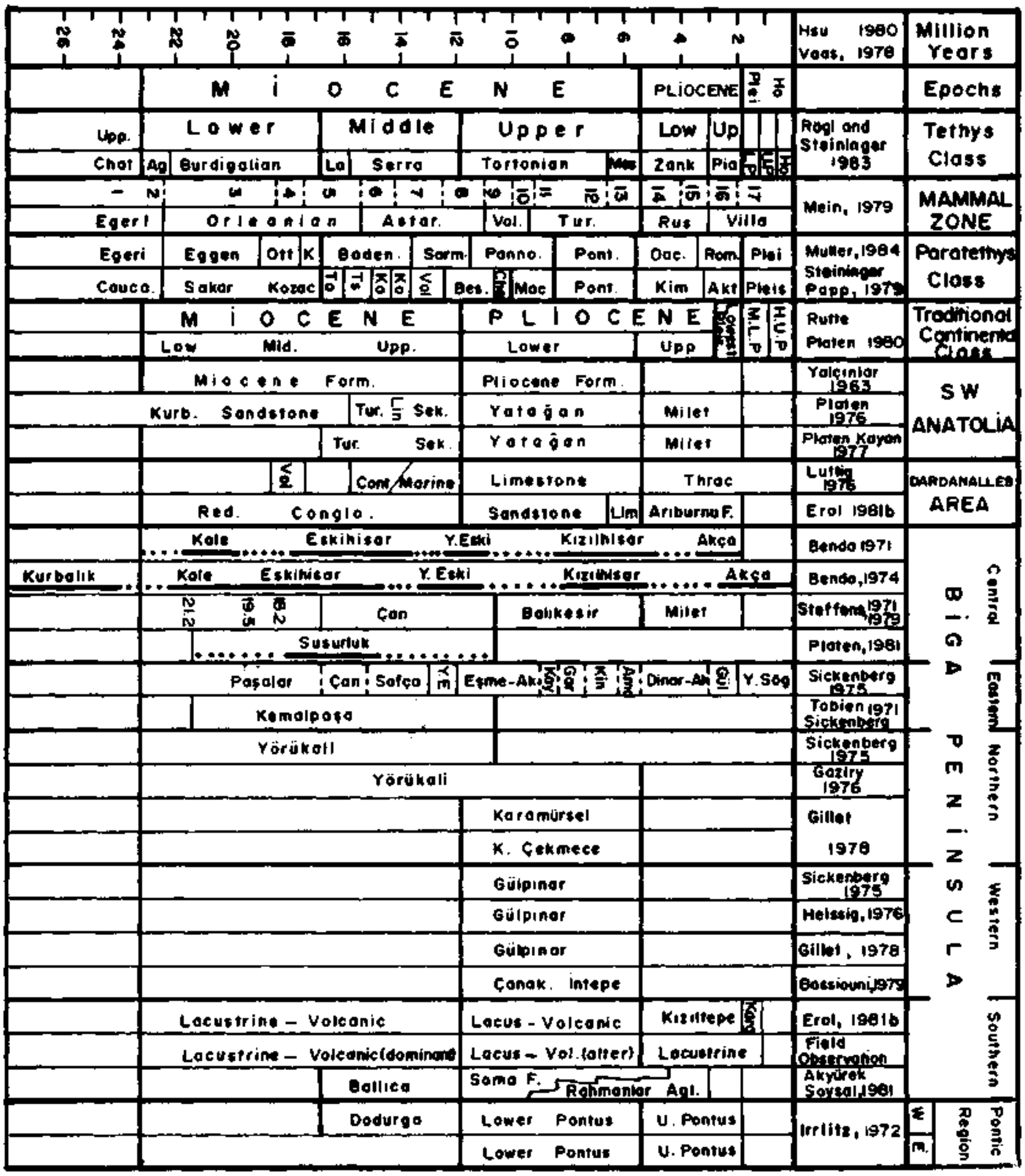

Fig. 3– The age of Neogene sequences in Biga peninsula (Modified from Herece, 1985).

The lignite layers in the Neogene deposits of Biga peninsula are assigned to the "Eskihisar Pollen Assenv age ranging in age from Burdigalian to Lover Serravallian. The viability of the pollen horizons is supported by several other distinctive fossil assemblages and radiometric dates of lavas (Benda et al., 1974). The pollen bearing sediments in northwest Turkey correspond to the lignite between Turgut and Sekköy members of western Turkey.

The fossil assemblages in the Neogene sequence on the northern and western boundaries of the Biga peninsula indicate an exclusively Paratethys (Chersonian) environment in the north. Mixed assemblages are widespread in the Bosphorous and Dardanelles area indicating a transition region between Tethys and Paratethys seas. The sedimentation of Neogene sequence started in Langhian, however, the Late Miocene was a time of extensive tectonic activities. These tectonic movements caused breakup of the landmass and connected Tethys and Paratethys. The Upper Miocene was also a time of extensive volcanic activity during which north- south distension began, thus deepening the sedimentary basin.

On the other hand, there are Upper Miocene lacustrine and fluvial deposits with widespread lignite in the basins (Havza, Ladik, Taşova- Erbaa) located in the area east of Marmara Sea along the NAF zone. To the west of these basins, the Çerkeş-Kurşunlu, Ilgaz and Tosya were developing as parts of final closure of northern Neo- tethys sea along the intra - Pontide suture zone (Şengör and Yılmaz, 1981; Barka, 1984). Andesite, tuff and agglomerate (Devrez and Sivricek formations) were the first deposits of these basins and they were formed as the result of continuing north-south compression during Early Miocene (Barka, 1984). During the deposition of Middle Miocene lacustrine sequences between Kurşunlu and Ilgaz and south of Tosya, the Havza-Ladik and Taşova-Erbaa basins were started to form by the north-south compression at the end of Middle Miocene (Barka, 1984).

During Tortonian time, the lacustrine deposits of lower Pontus formation (Irrlitz, 1972) began to accumulate in all of these basins. The interpretation of syn-sedimentary structures in the lower Pontus formation (Hancock and Barka, 1983) indicates that a broad right- lateral strike-slip motion is initiated at the end of Tortonian (Barka, 1984). These data, however, require that the Anatolian plate should move westward during Upper Miocene. Is the origin of the Upper Miocene basins in the Biga peninsula different than that of the basins in the central portion of the NAF zone. Are those basins formed by the earlier strike-slip regime similar to that prevailing along the NAF today? The answers of these questions are out of scope presently because of lack of some additional field work. The development of the basins in the central portion of the NAF also is related to the westward motion of the Anatolian plate (Hancock and Barka, 1981, 1983; Barka, 1984; Şengör et al., 1985).

No matter what the origin of the Neogene basins, the faults can be recognized from their topographic expression as young scarps that bound the Bursa-Gönen graben. The Bursa-Gönen graben at the least can be explained by the fault systems that surrounded it. It is evident from Landsat images that the diamond-like shape of the graben area is due to east-northeast and west- northwest sets of lineaments. Because only some of the lineaments associated with these Neogene basin margins can be shown to be recent fault scarp, it is suggested that the rest were present as Neogene fractures providing structural control for the developing sedimentary basins. This postulated movement on some lineaments appears to be responsible for deformation of the Neogene sequence in the Biga peninsula.

A tectonic model for the Quaternary (Recent) events in the Eastern Mediterranean incorporates the NAF, EAF, and Hellenic subduction zones, and calls for the Pontide unit, (on the northern side of NAF), to be stationary while the Anatolian plate to the south moves westward (McKenzie, 1972, 1978; Dewey and Şengör, 1979; Şengör and Canıtez, 1982). Assuming this model to be correct, with the northern side relatively fixed with respect to the northeasterly trending Yenice-Gönen lineament, the block south of this lineament must move southwestward if it is to be part of the NAF system. Such a movement was confirmed during the earthquake of Yenice-Gönen in 1953 (Ketin and Roesly, 1953; Herece, 1985).

Besides the Yenice-Gönen line, the Sarıköy-İnova and the Manyas-Danişment lineaments also must represent right - lateral strike-slip faults. The southern side of the Bursa -Gönen graben, therefore, moves west-southwest along these strike-slip traces while the graben area is enlarged north to south by normal faults as shown in Figure 5. This latter movement also is verified by the normal faults of 1964 event (Herece, 1985).

Apparently this closed depression, in which some 3000 m of Neogene sediments are deposited, is bounded by normal faults; elsewhere along the NAF system, right-lateral strike-slip faults are dominant. These contrasting styles can be accomodated in second-order shear models of Lensen (1958).

TOTAL LATERAL DISPLACEMENT IN THE SOUTHERN BRANCH OF THE NAF'S

There are no available data to suggest any displacement for the NAF on Biga peninsula. It is essential to know the depth of the subsidence and the thickness of Quaternary deposits (When and how the graben floor started subsiding). The gravimetric cross-section of the Manyas depression area, however, implies a lateral displacement of the southern branch of the NAF in the Biga peninsula.

As can be seen in Figure 2, the separation in the graben area is about 20.5 km. This represents the maximum separation in the graben, and includes the cumulative strain from both Neogene and Quaternary times. It is difficult to assign incremental displacement with time, because neither the tectonic style nor strain history has been worked out for the Neogene deposits in the graben (Herece, 1985). The Yenice -Gönen, Sarıköy-İnova and Manyas-Danişment faults, therefore, should not be responsible for the total separation if the Bursa-Gönen graben used different fault systems during the Neogene sequence.

The graben has subsided and enlarged since Late Upper Pliocene due to the distinct young faults which bound these sedimentary basins to the north and south. However, there is a possibility to find out what the separation is in the depression area. If the profile of the graben is restored, one can calculate the absolute separation in the graben as 8.2 km since Late Upper Pliocene time (Herece, 1985). It is also possible to determine that there has been combined dextral movement of 8.2 km on the Yenice-Gönen, Sarıköy-İnova and Manyas-Danişment faults. This value of 8.2 km should represent the cumulative lateral displacement on all the faults that constitute the southern branch of the NAF during Quaternary time.

LATERAL DISPLACEMENT ON THE YENİCE-GÖNEN FAULT

By matching up the weathered andesite and granite intrusions on either side of the Yenice valley (Fig. 4) a total horizontal displacement of 1.7-2.8 km is inferred on the Yenice-Gönen fault line. The weathered andesite and younger volcanic rocks overlying the Neogene sequence east of Yenice apparently have been displaced southwestward about 3.4 km, east-west offset of 2.8 km. In contrast, the contact of the granite intrusion in the Yenice basin appears to have been separated (right-lateral) about 2.8 km with an east-west offset of 1.7 km. Assuming the higher of these values as a maximum, lateral displacement can be taken as 2.8 km on the fault line in the Yenice valley.

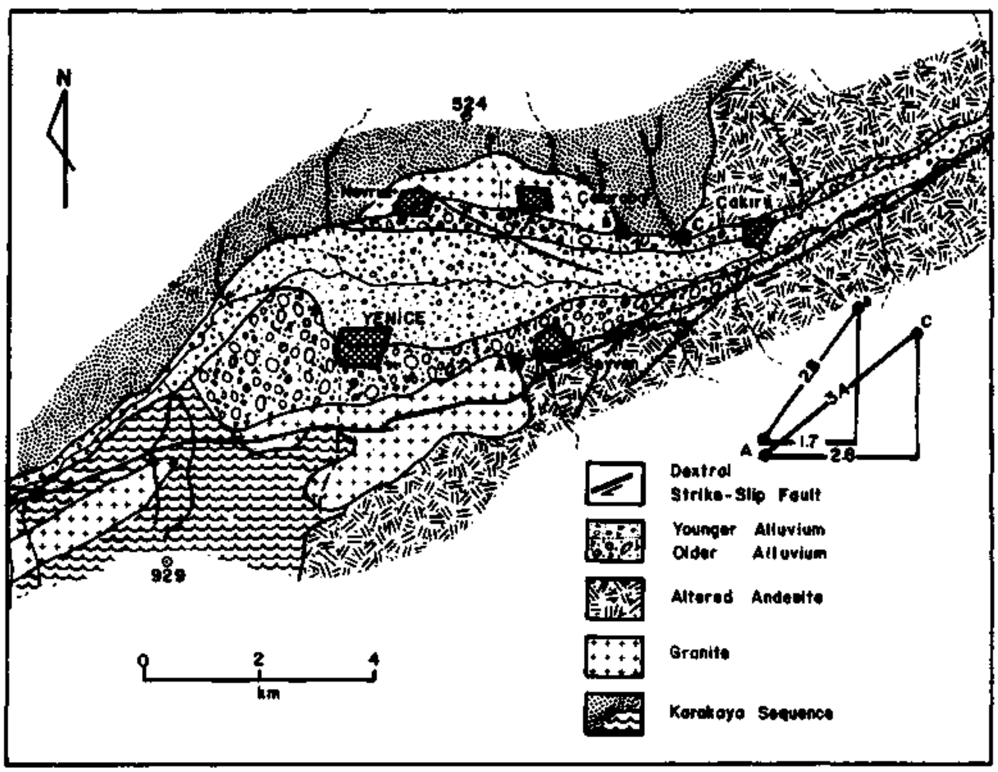

Fig. 4 - The lateral displacement of volcanic units in the Yenice valley (Herece, 1985).

The rest of 5 km offset should be used by Sarıköy-İnova and Manyas-Danişment faults. No data, however, are available for displacement on these faults.

In the current literature, there are discussions on the amount of the lateral displacement of the fault zone as in the westward extension of the NAF zone. Seymen (1975) and Tatar (1975) state a 85±5 km lateral displacement obtained from displaced suture zone in the eastern part of the NAF. This Figure has been widely used in the modelling works. Yılmaz (1985) does not agree with this offset (85±5 km) and he states that this amount of offset must be reconsidered.

Hancock and Barka (1983) indicate that there has been 25±5 km total dextral separation along the NAF since Lower Pliocene. The 25-27 km offset of the Kızılırmak river is in harmony with the value given above. The new data come from the field work around Gerede in 1987. It demonstrates that dextral offset of the NAF in this area is about 18-22 km (Herece, in preparation).

The change in direction of the Sakarya river near Geyve from south-north to easterly, may correspond to the offset of the NAF. This offset is roughly measured as 13-18 km (Herece, 1985).

8 km of the total offset has been used by the faults in the Biga peninsula since the Late Upper Pliocene (in Quaternary). The rest of the offset, the location and the time interval of the remaining displacement of the NAF must be studied. As stated in the current literature the NAF connects the northern Marmara trough to the Saros bay and the Lake İznik-Gemlik bay to the Marmara depression areas. In other words, the Marmara depression area and/or the parts of the enlargement of this area may correspond to the strike- slip component of the NAF. The Gaziköy-Saros bay fault, which runs from the Sea of Marmara to the Saros bay, is a very distinct right-lateral strike-slip fault and it is currently active. The activity of this fault zone is indicated by 1912 and 1928 earthquakes. For the above mentioned reasons, most of the displacement of the NAF might be located in the northern part of the Bursa-Gönen graben.

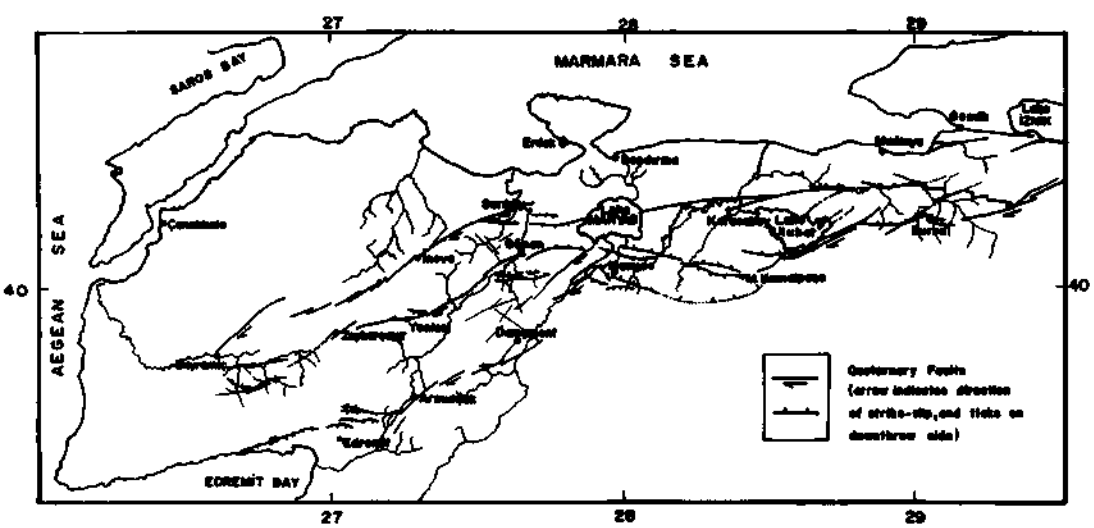

Fig. 5– The major fault lines in Biga peninsula.

WESTWARD CONTINUATION OF THE YENİCE-GÖNEN FAULT

The fault trace of the 1953 earthquake line and other fault breaks caused by previous tectonic activity in the Yenice valley have been mapped as far southwest as Sazak. They appear to diminish in size and even lose their surface expressions farther to the southwest In particular, the lateral movement on the Yenice-Gönen fault zone which appears to be responsible for the opening of the Bursa-Gönen graben needs to be addressed. From here westward, the continuation of the Yenice-

Gönen fault zone is discussed in detail by Herece (1985) and the fault zone is terminated somewhere to the west of Eskiyayla village.

DISCUSSION AND CONCLUSION

The east and central portion of the NAF zone is very distinct morphologically. It branches and becomes less distinctive in Marmara Sea region. Well known dextral strike-slip component of the fault is distributed over a number of depressional areas with a large component of normal faulting. The Bursa-Gönen graben is one of those areas.

The bounding faults, formes the Bursa-Gönen graben, is connected to the NAF zone by northeasterly lineaments in the east of Bursa. Westward, near Gönen and Sarıköy, this graben closes. The main faults along the northern and southern margins of the graben are normal dip-slip with dip angles of 68° - 70° . These dip angles also are very close to the theoretical dip angles of grabens (Heiskanen and Vening Meinesz, 1958). The graben, formed by normal dip-slip faults, is filled with 3 km thick deposits. There is a possibility that these normal faults may extend to a depth of 8 km.

The normal faults along the southern border of the Bursa- Gönen depression, which includes the Lake Manyas, are very distinct; the northern border coincides with the northern shore-line of the Lake. The western and southern parts of the Bursa- Gönen graben apparently move toward west-southwest along dextral strike-slip faults, while the graben area is enlarged by east-west trending normal faults.

It is essential to know the age and nature of the deposits in the graben if deductions on how the graben formed and the graben floor started subsiding are to be made. As stated in the literature (Şengör et al., 1985; Barka, 1984) the acceptance of the tectonic events in the Upper Miocene basins are caused by westward movement of the Anatolian plate brings the possibility that these basins were developed by an earlier dextral strike-slip regime similar to the NAF of today. Because Lower Pontus formation is overlain by the Upper Pontus formation with angular unconformity (Irrlitz, 1971; Barka, 1984). This unconformity may suggest that the deformative tectonic events during deposition of the Lower Pontus formation slowed down or stopped for a while (Herece, 1985). The NAF of today has been active since Late Upper Pliocene time.

In the Bursa-Gönen graben the actual separation in north to south direction has been about 8 km (4 mm/year) during and since the Quaternary. The possible total horizontal displacement along the Yenice-Gönen fault zone, which is one of the key component of the extension, has been about 2.8 km (1.4 mm/year). No data, however, are available for displacement on the Sarıköy-İnova and Manyas-Danişment faults.

The Sarıköy-İnova fault starts from Sarıköy and continues southwestward. The eastwaru continuation of this distinct fault trace is thought to represent the northern boundary of the Bursa-Gönen graben. In other words, this fault does not represent the middle branch of the NAF which runs (Barka, 1983; Barka and Kadinsky-Cade, 1988) from Gemlik bay to Bandırma and Sarıköy to Bayramiç and Ezine and Aegean Sea (Herece, 1985; 1988). The westward continuation of the Yenice-Gönen and Sarıköy-İnova faults terminate somewhere to the east of Bayramiç (Herece, 1985).

The Manyas-Danişment fault caused extensive and complex deformation in its western part. This fault might run over Armutluk to the Edremit bay. The Yenice valley (3-4x7-9 km), which is cut by the fault trace of 1953 event, and İnova depression area (1.5x2.5 km) are "pull-apart" basins (Herece, 1985).

In conclusion, the northeast-southwest trending dextral strike-slip faults in the Biga peninsula are the westernmost known extension of the NAF. Based on this observation the Yenice-Gönen fault zone starts from southwest of Gönen and extends toward Yenice. In other words, northeast-southwest trending fault system, from Muratlar, Kumköy to Yenice, is right-lateral in character. This character implies that it is related to the NAF. East-west trending fault from Muratlar to Çakmak, which has been developed by the 1953 earthquake, is the latest connection of the Yenice-Gönen fault zone to the Bursa-Gonen graben. The fault trace of 1964 Manyas earthquake is not the eastern extension of the fault zone developed partly by the 1953 Yenice-Gönen earthquake. This conclusion debates that of Şaroğlu et al. (1987). The fault zone developed by the 1964 Manyas earthquake is not a right-lateral strike-slip fault but it is a normal dip-slip fault belonging to the Bursa-Gönen graben (Herece, 1985).

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DİE MİNERALOGIE DER Pb-Zn-LAGERSTAETTE VON GÜMÜŞHANE (TÜRKEİ)

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ZUSAMMENFASSUNG. - Die wichligsten Blei-Zinklagersttetien bei Gümüşhane sind Hazine Mağara und Kırkpavili. Die Lagerstaetten sind durch Zufuhr der mezolhenmalen Erzlösungen in die Oberkreide-Kalkstein hydrolhermal melasomatisch entstanden. Die Vererzung sieht höchstwahrscheinlich mit der tertiaeren Granitoiden im Zusammenhang. Die zahlreichen Erzproben, die mineraiogisch untersucht wurden, stammen von Hazine Mağara und Kırkpavili Vorkommen. Nach dieser Untersuchungen wurden im Erz neben der bis heute bekannten Mineralien, zahlreiche unbekannle Mineralien festgestellt. Die beobacnteten Erzmineralien sind wie folgend: Pyrit, Galenit (Bleiglanz), Zinkblende, Fahlerz, Aikinit, Gedigenes Gold, Elektrum, Gedigenes Silber, Kupferkies, Boumonit, Boulangerit, Luzonit, Bornit, Mawsonit, Klaprothil, Galenowismutit, Hessit, Arsenopyrit, Emplektit, Gedigenes Wizmut, Wittichenit, Altait, Telradymit, Magnelkies, Rutil, Anaias, Zirkon, Titanit und Graphit Durch die Umwandlung dieser primaeren Mineraben wurden kalkosin, Covellin, Smithsonit, Malachyt, Azurit, Psilomelan, Pyrolusit enlslanden. Die Gangmineralien der Hazine Mağara und Kırkpavili Vorkommen sind Quarz, Calcit, Dolomit, Baryt, Ankerit, Siderit, Chlorit und Sericit.

ABSTRACT . - The most important lead-zinc deposits of Gümüşhane, Hazine Mağara and Kırkpavili ore deposits were formed by the mesothermal solutions ascended through the faults and metasomatosed the Upper Cretaceous massive limestones. The ore mineralization is thought to be closely related to the Tertiary granitoids. Many samples which were taken from the waste of Hazine Mağara and Kırkpavili ore deposits have been mineralogically examined in detail and beside the known ore minerals up to the present, many new minerals have been determined as a result of this study. The ore minerals are; pyrite, galena, sphalerite, fahl-ore, aikinite, native gold, enargite, luzonite, bomite, mawsonite, klaprotite, galena-bismuthine, hessite, arscnopynte, emplectite, native bismuth, vitishenite, altaite, tetradymite, pirotite, rutile, anatase, zircone, Utanite and graphite. The surfacial alteration and weathering of these primary minerals resulted secondary minerals such as chalcocite, covelline, limonite, arsenic-antimony ochres, anglesite, cerussite, smithsonite, malachite, azurite, psilomelane, pyrolusite. The gangue minerals of Hazine Mağara and Kırkpavili ore deposits are quartz, calcite, dolomite, barite, ankerite, siderite, chlorite and serisite.

EİNLEİTUNG

Die Gold und Silber enthallende Blei-Zinklagerstaette bei Gümüşhane liegt 2-3 km westlich von der Stadt Gümüşhane und innerhalb der geologischen Karte Blatt TRABZON H 42 b_2 (Abb. 1). Die Stadt Gümüşhane wurde nach dem Silber der Lagerstaetten benannt.

Mit der ersten Bergbautaetigkeiten sollen in den Blei-Zinklagerstaetten bei Gümüşhane, nach Angaben des "Milliyet Türkiye İller Ansiklopedisi, 1982" in den Jahren 1238-1268 begonnen worden sein. Nach der gleichen Quelle wurde die Stadt Gümüşhane im Jahr 1243 von Selçuken in den Haenden von Ilhanen übertreten. Spaeter hat der Sultan von dem osmanischen Reich, Fatih Sultan Mehmet nach dem Krieg "Otlukbeli" diesen Ort von Akkoyunlular zurückgenommen. In der Blei-Zinklagerstaette bei Gümüşhane soll die erste Bergbautaetigkeit waehrend des osmanischen Reich unter die Führung von IV. Murat begonnen sein.-Nach dieser Taetigkeit wurden diese Gruben einige Zeitlang stillgelegt und in der gleichen Zeit die Stollen der Lagerstaette mit dem Grundwasser geflutet. Waehrend der Herrschaft von III. Mustafa wurde versucht, die Stollen zuentwaessern und die Gruben wieder aufzumachen. Der Versuch für diese Bergbautaetigkeit war aber erfolglos.

Nach Angaben von Kraus (1889) hat die Firma Daniel Pappa im Jahr 1860 mit dem Bergbau begonnen. Im Jahr 1894 wurde aber die Bergbaugenehmigung von der damaligen Regierung abgesagt. Danach gab es unter die Bergleute viele Arbeitlosen, die in verschiedenen Orten des Anatoliens wanderten und zu der Entwicklung der anderen Bergbautaetigkeiten des Anatoliens beigetragen haben.

Nach dem ersten Weltkrieg hatten Fuat Bey und seine Teilhabern die Lagerstaetten von Gümüşhane übernommen. Im Jahr 1921 haben in diesem Provinz englische Militaergeelogen gearbeitet.

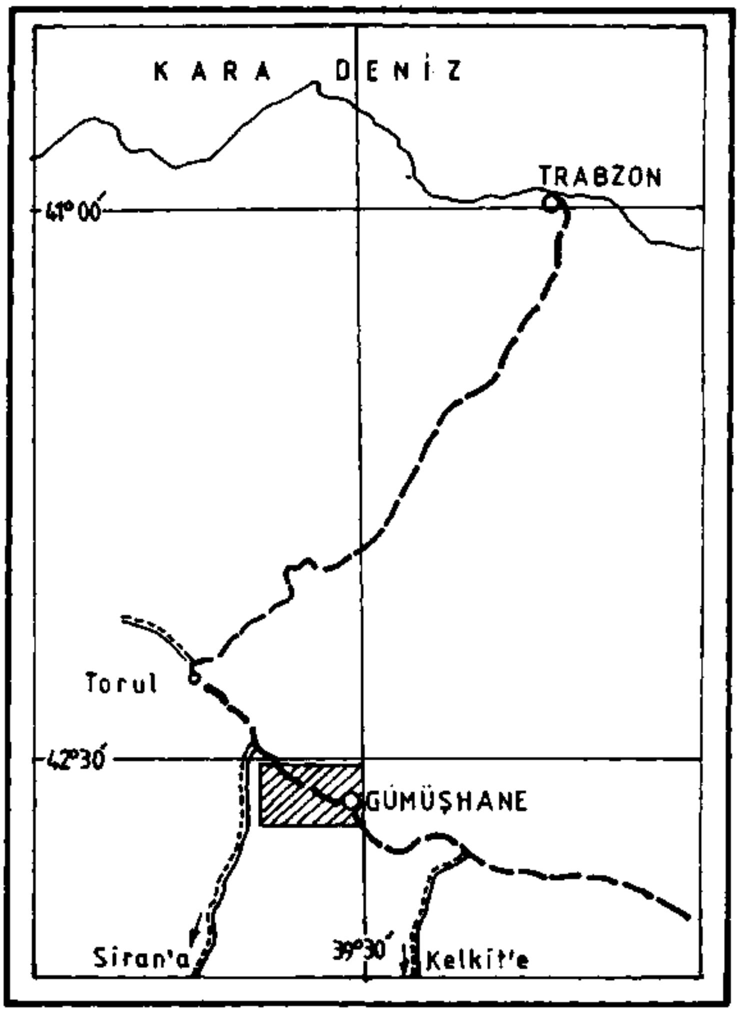

Abb. 1- Übersichtskarte des Arbeitsgebietes

Nach der Begründung der türkischen Republik wurde die Untersuchungen in der Blei - Zinklagerstaette bei Gümüşhane von Ölsner (1935) durchgeführt. Er hat die Lagerstaette genelisch in zwei Gruppen geteilt. Nach Ölsner ist die erste Gruppe dieser Lagerstaette in Kalksteinen metasomatisch entstanden. Die zfeite Gruppe besteht aus der hydrothermalen Gaengen in Graniten.

Kovenko (1937) hat in dieser Lagerstette drei verschiedene Vererzungstypen beobachtet. Dem ersten Typ gehoren die im Kalkstein eingelagerten Erzlinsen, wie bei Hazine Mağara und Kırkpavili zu beobachten sind. Der zweite Typ der Vererzung bestehet aus hydrothermalen Erzgaengen und Adem wie in Deremaden auftreten. Der dritte Typ der Vererzung ist an der Storungszonen gebunden und besteht aus der sekundaeren Mineralien.

Schumacher (1937), der diese Lagerstaette sehr kurz besucht hat, hat die Vererzung als hydrothermale Gaenge angesehen.

Gysin (1938) hat auch geologische Untersuchungen über die Lagerstaette bei Gümüşhane durchgeführt.

Nach Angaben von Dandria (1940) wurden die Vorkommen von Hazine Mağara und Kırkpavili durch die in Oberkreide - Kalkstein eingedrungene mezothermale Erzlösungen als hydrothermal - metasomatische Vererzung gebildet.

Pejatovic und et al. (1970) haben die Lagerstaette bei Gümüşhane in zwei Gruppen geteilt. Zur ersten Gruppe gehören die Gaengen, die in den magmatischen Gesteinen auftreten. Die zweite Gruppe besteht aus der Vererzungen, die in den Kalkstein metasomatisch eingelagert sind.

Çoğulu (1970) hat petrographische Untersuchungen über die Granitoiden durchgeführt. Er hat das Granitoid von Gümüşhane mit dem Granitoid von Rize vergliechen.

Bosch und et al. (1974) haben im Rahmen des UNESCO- Projekt im Lagerstaettenbereich eine geologische Untersuchung durchgefuhrt.

Yılmaz (1976) hat das inhomogen aussehende Granitoid von Gümüşhane petrographisch in verschiedenen Granitoidarten eingeteilt. Diese Abarten sind Granodiorit, Çamlıca - Adamelith, Gümüşhane - Ademelith, porphyrische Mikrogranit, Er hat auch die mit der Granitoiden von Gümüşhane eng verbundenen Aplit, Pegmatit und Quarzadem festgestellt.

Kamitani und et al. (1977) haben die Granitoiden, die Oberkreide - Kalksteine und Eozaen - Vulkaniten von Gümüşhane untersucht und danach festgestellt, dass die Gümüşhane Vererzungen hauptsaechlich in den Oberkreide - Sedimenten eingelagert sind.

Erbayar und Ödevci (1979) haben an einigen Erzproben geochemische Analysen durchgefuhrt, die aus der Gruben der Gümüşhane - Lagerstaetten stammen (Tabb. 1).

SAMPLE LOCATION	ELEMENTS						AUTHORS
	Pb %	Zn %	Cu %	Fe %	Ag gr/t	Au gr/t	
Hazine Mağara	—	—	—	—	2368	3,9	Ölsner, 1935
Kırkpavili Maden	48.3	—	—	—	500	—	
Dere Maden	60.5	—	—	—	20,7	—	
Dere Maden	20.5	—	—	—	580.4	—	
Canca Zuhuru	—	—	—	—	4.0	0.2	
Dere Maden	80 17	—	—	—	—	—	Kovenko, 1937
Hazine Mağara ya ait 5 m	8.04	8.46	—	—	1600	13,0	Dandria, 1940
karot ortalaması	3.04	2,80	0,80	17.4	89	2.55	
Hazine Mağara	14.30	35.0	0,20	—	243,5	2.5	Erbayar und Ödevci, 1979
Kırkpavili Maden	0.07	0.30	0.43	—	—	—	
Hazine Mağara 1	4.75	—	17.26	—	3739,3	48,7	Güner et. al. 1985
2	—	—	0.13	—	29,8	2,2	
3	6.98	—	0.33	—	103,0	2.9	
Kırkpavili Maden	—	—	0.25	—	4.2	1.0	

Tabelle 1- Die chemischen Analysen der einigen Proben die von dem Arbeitsgebiet entnommen wurden.

Nach Angaben von Öztunalı (1983) hat die Vererzung hauptsaechlich die WNW-OSO verlaufende Störungen ausgewaehlt und durch die in Oberkreide - Kalkstein eingedrungenen Erzlosungen hyprothermal - Metasomatisch entstanden.

Çınar und et al. (1983) führten in der Umgebung von Gümüşhane eine geologische Untersuchung durch und danach fertigten eine geologische im Masstab von 1:25 000 von der Umgebung der Gümüşhane - Lagerstaette an.

Güner und et al. (1985) führten in der Gebieten von Gümüşhane ebenfalls geologische Arbeiten durch.

Nach dieser oben erwaehnten Arbeiten stellt man fest, dass die zahlreiche, gründliche Untersuchungen über die Blei - Zinklagerstaette bei Gümüşhane durchgeführt wurden. Dagegen fehlt aber eine gründliche mineralogishe Untersuchung dieser Lagerstaette. Die Autoren dieser Arbeit besuchten im September 1986 die Lagerstaette bei Gümüşhane, um eine gründliche mineralogische Untersuchungen durchzuführen. Waehrfend dieses Besuches wurden viele Proben, hauptsaechlich von der Hazine Mağara und Kırkpavili genommen. Bei der Probenahme auf der Gelaende haben uns die Geologen von MTA (Trabzon) sehr geholfen.

GEOLOGIE

Die aeltesten Gesteinen der Blei - Zinklagerstaette bei Gümüşhane sind die permokarbonischen Granitoide und die metamorphen Gesteine (Çoğulu, 1970). Darüber liegen Konglomerate, die eingeschalted Jura (Lias) aeltrige Andezit-Bazaltlaven und ihre pyroklastische Gesteinskomponente enthalten (Yılmaz, 1976). Die liegenden Konglomeraten überlagern die dünnschichtigen, fossil enthaltenden, roten Kalksteine (Abb. 2). Darüber folgen Sandsteine, Mergel und mergelige,

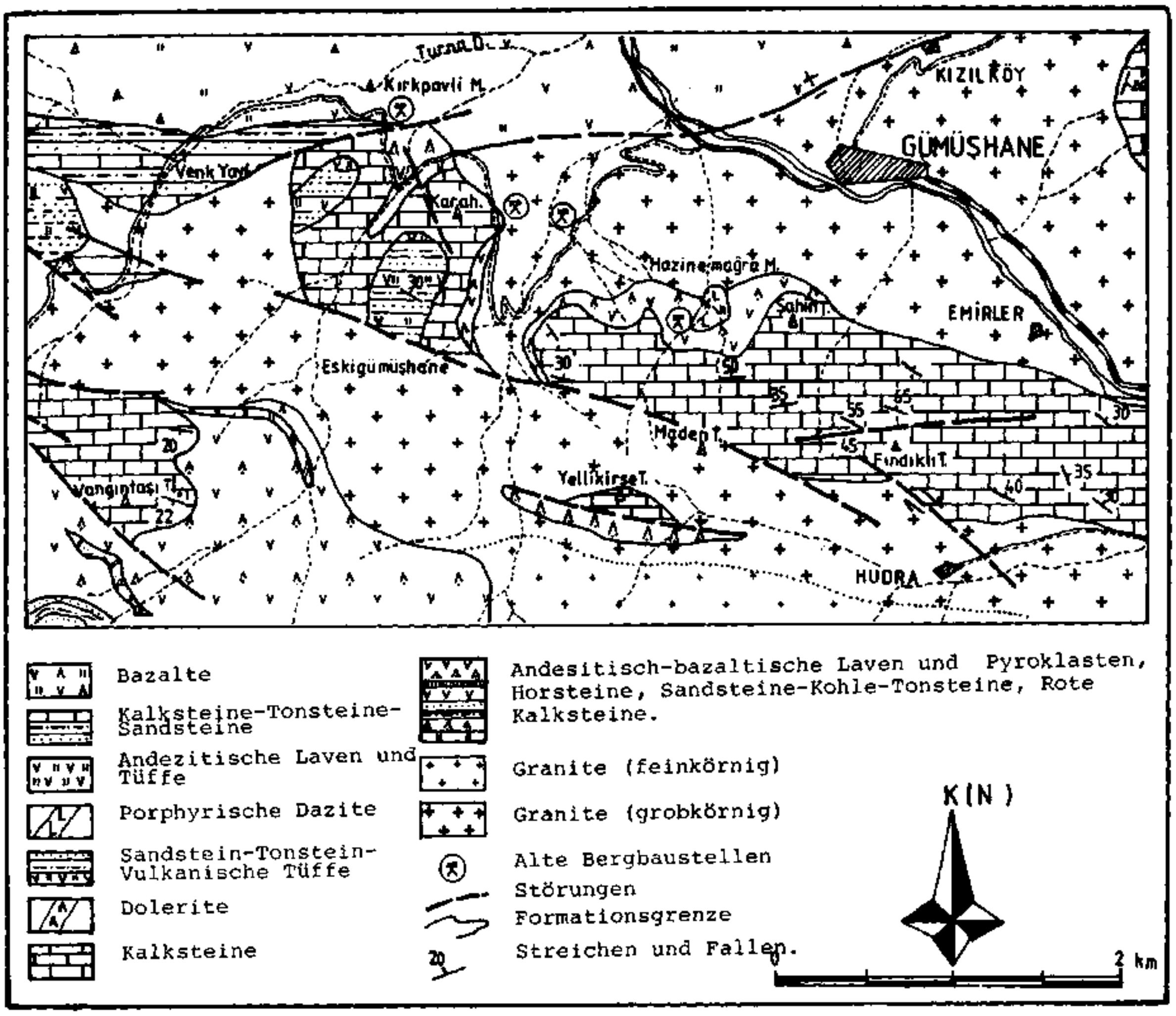

Abb. 2- Die geologische Karte des Arbeitsgebietes (nach Güner und et al., 1985).

kohlige Wechsellagerungen. Über dieser Serien liegen 3 - 5 m maechtige Hornstein - Serien (Güner und et al.,1985). Die massiv aussehende Oberkreide - Kalksteine, die über Dogger - Malm liegen und hellgraue Farbe zeigen, sind dolomitisiert und schwach rekristallisiert. Darüber sind 400 - 500 m maechtige tonige, rot farbige Oberkreide - Kalksteine, vulkanische Tuffe, Sandsteine gelagert. Über dieser Serien kommen Oberkreide - Vulkaniten und Produkte, die aus verwitterten andezitischen Tuffen bestehen.

Nach Angaben von Güner und et al. (1985) sind die tertiaeren Granitoide in die oben genannten Serien eingedrungen. Das Eozaen beginnt mit den dünnen Konglomeratlagen und diese werden von vulkanisch - sedimentaeren Serien nachgefolgt. Die im Eozaen abgelagerten Serien, die aus Kalksteine, Tonsteine, Sandsteine und andezitisch - bazaltischen Laven und ihren piroklastischen Komponenten bestehen, wurden von Granitoid - Intrusionen nicht veraendert. In dieser Umgebung gibt es auch junge Andezit - Diabaz - Gaenge.

Das Gümüşhane - Gebiet befindet sich im südlichen Teil des ostpontitischen Bereiches (Ketin, 1966) und erlebte die Entwicklung der herzinischen und alpidischen Orogenese mit. Deshalb trift man sehr oft im Arbeitsgebiet zahlreiche Störungen, Verfaeltungen und Überschiebungen.

VERERZUNGEN

Bei der Blei - Zinklagerstaette von Gümüşhane sind die hydrothermalen Erzlösungen in die Störungszonen eingedrungen und haben im Kalkstein linsenartige, hydrothermal - metasomatische Erzlagen gebildet. Für diese linsenartige Vererzung bilden die Hazine Mağara und Kırkpavili Vorkommen sehr gute Beispiele (Kovenko, 1937). Nach Angaben von Kovenko (1937) und Dandria (1940) liegt Hazine Mağara Vorkommen innerhalb der dolomitischen Kalksteine und an dem Kontakt von der Kalkstein - Mergeln. In der naehe von dem Hazine Mağara - Vorkommen liegt eine Verwaerfung, die in der Richtung 38° SW einfaellt und höchstwahrscheinlich mit der Vererzung in enger Beziehung steht. Da die tektonische Taetigkeit an dieser Verwerfung auch nach der Vererzung angedaurt hat, trift man hier dadurch zerbrochene und zerstückelte Erze. Die wichtigsten Erzmineralien von Hazine Mağara sind in den Kompakt - Pyriterzen ausgebildet. Makroskopisch kennt man in den Erzproben neben Bleiglanz auch Zinkblende. Kupferkies und Fahlerz sehr deutlich.

MİKROSKOPISCHE UNTERSUCHUNGEN

Die An-und Dünnschliffe, die aus der Erzproben von der Hazine Mağara und Kırkpavili Vorkommen herrgestellt worden sind, wurden unter dem Mikroskop gründlich untersucht. Die Erzproben, die hauptsaechlich aus Bleiglanz bestehen, sind an Erzparageneze reicher als die Erzproben, die mehr Pyrit enthalten. Deshalb wurden bei der mineralogischen Untersuchungen meistens die Erzproben vorgezogen, die an Bleiglanz reich sind.

Pyrit. -Es ist das haeufigste, weit verbreitete und wahrscheinlich aelteste Sulfidmincral der Blei-Zinklagerstaette Gümüşhane. Dieses Mineral ist fast überall in den Gaengen oder zerstreut als "Impraegnationen" in den magmatischen und sedimentaeren Gesteinen zu finden. Neben der idiomorphen und hypidiomorphcen Pyriten kommen auch in Bleiglanz und Zinkblende als skelettartige Aggregattcn vor. Solche Pyriten wurden manschmal von der Wirtmineralien verdrengt. Pyrit zeigt fast immer kataklastische Risse und Sprünge. Entlang dieser Risse und Sprünge wurde Pyrit von jüngeren Mineralien wie Bleiglanz, Zinkblende, Bornit, Kalkosin, Covellin und Kupferkies gefüllt. Selten tritt radialstrahlige Aggregaten von Pyrit auf, die wahrscheinlich Markazitpsödomorphosen sind.

Bleiglanz. -Nach dem Pyrit ist Galenit das nacchst haeufigste und verbreitete Sulfidmineral der Blei-Zinklagcrstaette Gümüşhane. Bleiglanz bildet hier idiomorphe und hypidiomorphe Kristalle aus, die manschmal um ihre Rande herum und entlang ihre Spaltflaechen, Risse und Sprüngen in Anglesit und Cerussit umgewandelt wurden. Die dreiekigen Spaltbrücke des Bleiglanz zeigen oft Biegung, Krümung und Windung auf. Die tektonische Tactigkeit des Gebietes wurde höchst wahrscheinlich auch nach der Entstehung des Bleiglanzes weiter angedauert. Die Spaltflaechen und die Risse-Sprünge der Bleiglanz-Kristallen wurden gelegentlich von zwei verschiedenen Fahlerz-Typen und Kupferkies angefüllt (Tafel I und II).

Zinkblende. -Das Zinkblende ist nach dem Pyrit und Bleiglanz das dritt haeufigste end verbreitete, primaere Erzmineral der Lagerstaette Gümüşhane, das oft zahlreiche Entmischungen und Kupferkies, selten auch Bornit und Bleiglanz enthaclt. Die Zinkblende-Kristallen sind hypidiomorph und zeigen öfter kataklastische Struktur. Die Spaltflaechen, Sprünge und Risse der Zinkblende-Kristallen wurden durch jüngere Bleiglanz, Fahlerz und Kupferkies angefüllt. Sie sind an der Randen, entlang der Spaltflaechen, Sprüngen und Rissen in Smithsonit umgewandelt worden. Im Allgemein hat die Zinkblende von Gümüşhane-Vorkommen rote innenreflekse, die reich an FeS-lnhalt andeuten.

Fahlerz . - Femer ist Fahlerz in der Blei-Zinklagerstaette fast überall in geringer Menge als erste Generation vorhanden. Fahlerze treten hier in drei verschiedenen Arten auf. Die erste Art ist Wismut-Blei-Fahlerz. Die zweite ist Tennantit. Selten tritt in Bleiglanz Ag-Fahlerz auf. Im Allgemein kommen Wismut-Blei-Fahlerz und Tennantit zusammen vor (Tafel I, abb. a, b; Tafel II, abb. a). Die Fahlerze, die kataklastische Struktur aufweisen, wurden entlang der Rissen und Sprüngen von Bleiglanz, Karbonate und Quarz angefüllt. Diese Fahlerze treuen teilweise in Bleiglanz und manchmal zwischen Quarz und Pyriten auf. Wismut-Blei-Fahlerze und Tennantit kommen mit Kupferkies als ineinander Verwachsen vor. Sie sind an der Rande der Kristalle in Enargit-Luzunit umgewandelt worden. Diese Fahlerze besitzen manchmal Entmischungen des Kupferkieses. Sie bilden ab und zu auch mit dem Kupferkies eine mirmekitischen Verwachsungen. Tennantit verdringt das Bi-Pb-Fahlerz als kleine Gaengchen. Durch die oberflaechliche Umwandlung der Fahlerze entstehen Sb-As-Oker, Kalkosin und Covelling. Ag, Fahlerze kommen in Bleiglanz als kleine runde Einschlüsse vor. Durch die Zersprung und Zersetzung der Bi-Pb-Fahlerzen entstehen Wismutil, Wittichenit, Klaprothit, Emplektit und sekundear Bleiglanz (Tafel I, abb. a und b; Tafel II, abb. a). Diese Zersetzung der Bi-Pb-Mineralien wurde auch von Ramdohr (1975) oft beobachtet.

Aikinit. - Im Allgemein tritt dieses Mineral in geringen Mengen mit der Kupferkies, Bi-Fahlerze und Klaprothit ineinander Verwachsen auf. Aikinit hat meistens die Grosse von 100 mikron. Sie sind in Formen von idiomorph, hypidiomorph und kleinen Staebchen kristallisiert und zeigen mit der Bleiglanz und Kupferkies mirmekitische Verwachsungen auf. Daneben wurde in Aikinit die kleinen Teilchen von Pyrit, Bi-Pb-Fahlerz, Kupferkies als Einschlüsse beobachtet. Durch die oberflaechliche Umwandlung wurden Aikinite gelegentlich in Covellin, Cerussit und Wizmutit umgewandelt.

Gold und Elektrum . -Oft wurden diese Mineralien im Bleiglanz, in den Spaltflaechen der Bleiglanzen und zwischen den Bleiglanzkristallen beobachtet. Die Farbe von Gold ist nach Ag-Inhalt von Weis nach gelb wechselnd. Manchmal tritt Gold zwischen Fahlerz, Quarz und Pyrit auf (Tafel I, abb. b, c und d). Das Gold in diesen Lagerstaetten ist immer noch als "Freigold" und manchmal Elktrum vorhanden. Wenn der Silbergehalt des gediegenen Gold höher und seine Farbe lichter ist, so nennt man ihn als Elektrum.

Silber . Gediengenes Silber wurde in den Rissen und an Sprüngen von den Fahlerzkristallen beobachtet. Sehr selten treten auch silberaderchen auf. Die Hauptmenge des Silber steckt hier in gediegenen Silbern, in Sulfiden und Sulfosalzen.

Kupferkies . - Die in dieser Mineralparagenese sehr wenig auftretende Kupferkiese kommen in Bleiglanz und Zinkblende als Einschlüsse vor. Ausserdem kann man kleine Gaengchen beobachten, die im Bleiglanz eingedrungen sind (Tafel I, abb. a; Tafel II, abb. b).

Bournonit und Boulangerit . - Diese beiden Mineralien treten an den Kontakten von Kupferkies, Fahlerz und Bleiglanz sehr selten auf. Die manche Bleiglanzen enthalten Boumoniteinschlüsse. Sie sind rundliche allotriomorphe Körner, die manchmal von Fahlerzen kokardenförmig umhüllt werden. In einzelfaellen wird Boumonit und Boulangerit von Bleiglanz schön und typisch verdraengt. Die Körner von Boulangerit sind meist stengelig. Die isolierte Kristalle haben rhombischen Querschnitt.

Enargit-Luzonit. - Sie treten sehr selten an den Randen von Fahlerzen und Kupferkiesen, manchmal in Fahlerzen als Einschlüsse auf. Manche Enargil-Luzonit-Kristallen werden von kleiner Gaengchen Fahlerzs verdraengt. Enargit hat grosse Achnlichkeit mit Luzonit und bildet gedrungen prismatische Körner inmitten von anderen Erzen. Durch die oberflaechlichen Alteration wurden die Enargit-Luzonit in Kalkosin, Covellin und As-Sb-Oker umgewandelt.

Bornit. - Im Allgemein wurde Bomit in sehr kleineren Mengen und Kömer in Zinkblende und Bleiglanz beobachtet. Sehr oft kommt mit Kupferkies und Fahlerz ineinander verwachsen und manchmal in Kupferkies als Entmischungen vor. Kupferkieslamellen in Form dünner Tafeln oder Linsen sind sehr hacufig entmischt. Durch die oberflaechlichen Umwandlung wurde Bomit in Kalkosin, Covellin und Limonit umgewandelt.

Mawsonit . - Dieses Mineral tritt sehr selten in Bleiglanz, Kupferkies und Fahlerz, manchmal auch mit der Luzonit verzahnt auf. Mawsonit sieht unter dem Mikroskop dem Bomit sehr achnlich aus. Der Unterschied ist nur die höhere Anizotrophic, die Mawsonit besitzt.

Klaprothit. - Dieses Mineral tritt sehr selten mit den anderen Bi-Mineralien zusammen in Bleiglanz auf (Tafel I, abb. a). In Form von strahlig bis saeulig aussehende Klaprothit-Kristalle zeigen mit Aikinit mirmekitische Verwachsungen auf. In Klaprothit kann man auch sehr selten kleine Einschlüsse von Bi-Fahlerze beobachten. An den Kontakten von Bleiglanz-Klaprothit kommen sehr selten gedingene Wizmut- und Galenowizmut-Kristalle vor, die unter sich mirmekitische Verwachsungen zeigen. Klaprothit kann auch ein Zerfallsprodukt von Wittichenit sein, der selbst aus Bi-Fahleiz entsteht

Galenobismutit. - Nadelige bis leistenförmige Kristalle von Galenobismutit treten sehr selten und in kleineren Mengen in Bi-Fahlerzen auf. Manchmal kommt es auch sehr selten im Quarz und Calcit vor. Die Verwachsung mit Wizmutglanz ist manchmal skelettartig oder baumförmig, dass man vielleich an Zerfall eines komplexen Sulfosalzes denkt

Hessit. - Die Komgrösse von Hessit in vorliegenden Proben sind 25-30 mikron und enthalten kleinere Einschlüsse von Elektrum, Altait und Tetradimit. Im Allgemein wird Hessit in Fahlerzen als hypidiomorphe Kömer zerstreut beobachtet

Arsenkies . - Im Bleiglanz tritt dieses Mineral sehr selten als sehr kleinere idiomorphe Kristalle auf, das unter sulfidischen Minerale, das aelteste ist.

Emplektit, - Im Allgemein entsteht Emplektit durch die Zersetzung der Bi-Pb-Fahlerzen. Dieses Mineral tritt sehr selten mit Bi-Pb-Fahlerz, Klaprothit, Wizmutit und gedigener Wizmut zusammen. Emplektit zeigt manchmal eine mirmekitische Verwachsung mit der Aikinit und Klaprothit auf.

Wittichenit. - Dieses Mineral tritt wie andere Sulfosalzen sehr selten und mit Klaprothit und Emplektit verwachsen auf, das durch die Zersetzung von Bi-Pb-Fahlerz entstanden ist (Tafel II, abb. a).

Altait. - Er wurde im Allgemein sehr selten und in sehr kleineren Mengen in Bleiglanz als idiomorphe und hypidiomorphe Kristallen beobachtet.

Tetradymit. - Dieses Mineral tritt auch in kleineren Mengen mit der anderen Bi-Sulfosalzen zusammen in Galenit

Magnetkies . - Dieses Mineral tritt in kleineren Mengen mit der Klaprothit, Bomit, Bleiglanz, Zinkblende und Fahlerz zusammen in der idiomorphen Pyriten auf. Unter diesen Mineralien ist nur das Magnetkies aelter als Pyrit und bilden im Pyrit Einschlüsse. Magnetkies wird von Bleiglanz, Kupferkies, Zinkblende, Fahlerz und vielen anderen verdraengt

Rutil-Anatas . -Waehrend der hydrethermal-jnetasomatischen Vorgaenge wurden Rutil-Anatas von dem Nebengestein (Kalkstein) aufgenommen. Sie treten als kleine idiomorphe, hypidiomorphe Körner in kleineren Mengen auf. Sie können fast in allen hydrothermal-ausgebildeten Mineralien als Einschlüsse beobachtet werden. Sie zeigen manchmal kataklastische Struktur. Die Risse, Spalte und Brüche von Rutil-Anatas wurden durch junge sulfidischen Mineralien ausgefüllt. Die Kristalle zeigen manchmal Druck- und manchmal auch Verwachsungszwillinge. Die im Erz auftretenden Rutilkristalle haben mit den Rutilkristallen, die im Nebengestein beobachtet wurden, grosse Aehnlichkeit.

Zirkon . - Zirkon wurde im Erzparagenese selten beobachtet Die 25-30 mikron grösse, idiomorph-hypidiomorphe Zirkonkristalle treten hauptsaechlich in Bleiglanz und Quarz auf. Zirkon wurde auch wie Rutil-Anatas aus der Nebengestein aufgenommen.

Titanit. - Titanit kommt in der Mineralparagenese selten als idiomorphe, kleinere Kristalle in Quarz, Calcit und Bleiglanz vor. Titanit ist ein detrisches Mineral, das in Kalkstein transportiert wurde. Bei der Vererzung wurde Titanit aus Nebengestein mobilisiert

Graphit. - Dieses Mineral wird mit den Gangmineralien (Quarz, Dolomit und Calcit) zusammen in Bleiglanz beobachtet Heksagonal-Plattrige Graphitkristallen wurden bei der Vererzung aus dem Nebengestein übernommen. Wahrsci- -in-lieh wurde die kohlige Subtanz, die im Kalkstein in gerinen Mengen vorhanden ist waehrend der Vererzung graphitisiert.

Kohle . -In den Gangmineralien des Erzes wird gelegentlich kohlige Subtanz beobachtet, die auch im Nebengestein (Kalkstein) zu beobachten ist

Kalkosin-Cavellin . - Diese Mineralien entstehen durch die Umwandlung der kupferhaltigen Mineralien wie Kupferkies, Bomit, Fahlerz, Enargit-Luzonit. In Erzmineralien tritt Covellin immer mehr als Kalkosin auf. Diese beiden Mineralien wurden am Rande und entlang der Sprung- und Spaltrissen von primaeren Kupfermineralien beobachtet.

Limonit . - Diese sekundaeren Mineralien entstehen durch die Umwandlung aus Eisen enthaltende Mineralien wie Pyrit, Kupferkies, Bornit. Oft werden die Limonitmodifikasionen wie götit, Lepidokrosit zusammen ineinander und nebeneinander beobachtet.

Arsenik-Antimuan-Oker . —Sehr selten auftretende diese sekundaeren Mineralien entstehen durch die auseren Umwandlung aus Fahlerz, Enargit-Luzonit und Aikinit. Sie werden auch sehr selten von Covellin begleitet.

Anglesit-Cerusşit . - Sie enstehen durch die Umwandlung aus Bleiglanz. Sie treten an den Racnden, Spaltflaechen und entlang der Sprüngen und Rissen von Bleiglanz auf. Zuerst wurde der Anglezit, danach Cerussit ausgebildet. In diesen Mineralien beobachtet man oft die Reste von Bleiglanz und ab und zu Mal auch sehr wenig Covellin. Die bei der Umwandlung von Bleiglanz entstandenen Anglesit-und Cerussit-Minerale zeigen sehr schöne konsantrisch schalige Struckturen.

Smithsonit . - Dieses sekundacre Mineral entsteht durch die Umwandlung von Zinkblende. Er tritt an Racnden und entlang der Spaltflaechen, Sprünge und Rissen von Zinkblende auf. Die in Zinkblende auftretenden Mineralien wie Bornit, Pyrit, Kupferkies wurden bei der Smithsonit-Entstehung in Covellin und Limonit umgewandelt.

Malachyt-Azurit . - Sie wurden durch die Umwandlung der kupferenthaltenden Mineralien wie Kupferkies, Fahlerz und Enargit-Luzonit entstanden. Im Handstück und auch unter dem Mikroskop kann man die radialstrahlige Aggregaten von Malachyt-Azurit leicht erkennen.

Psilomelan-Pyrolusit . - Sehr selten auftretende Mineralien wurden durch die Umwandlung aus Siderit-Ankerit entstanden. Diese Mineralien treten sehr wenig an Raenden, entlang der Spaltflaechen, Sprüngen und Rissen von Siderit und Ankerit auf.

Gangmineralien

Quarz . - Er tritt als das haeufigste Gangmineral in kleineren idiomorph-hypidiomorphen Kristallen auf. Manchmal verdringt Quarz als Aederchen die Erzmineralien, die xenomorph und hypidiomorph und miteinander verzahnt verwachsen sind.

Calcit-Dolomit . - Sie treten als kleine meist rhombische Kristalle auf. Die Calcit-Adem und Gaengchen dringen die ganzen Erzmineralien durch. Manchmal füllt Calcit als matriks den Zwischcnraum zwischen den Quarzkristallen aus.

Schwerspat (Baryt) . - Baryt tritt als idiomorphe, stenglige Aggregaten sehr wenig auf. Der Zwischenraum der Barytkristallcn wurden häufig durch Karbonate, Bleiglanze und Quarze ausgefüllt.

Siderit-Ankerit . - Sie werden in wenigen Mengen als kleine rhombische Kristalle beobachtet. Sie sind an Raenden und entlang der Spaltflaechen in Limonit, psilomelan-Pyrolusitumgewandelt.

Chlorit-Sericit. - Diese Mineralien wurden oft zwischen den Quarzkörnem, manchmal auch zwischen den Erzmineralien in kleineren Mengen beobachtet.

Chemische Analysen

Die chemische Analysen, die von der verschiedenen Autoren in verschiedenen Zeitabstaenden durchgeführt worden sind, wurden Tabb. 1 angegeben. Wie daraus zu entnehmen ist, zeigen manche Proben höhere Werte von Gold und Silber. Die Autoren wie Ölsner (1935) und Dandria (1940) behaupten, dass die Silbergehalle der Erze aus dem Fahlerz und dem gediegenen Silber entstammen. Dagegen aber konnten die Autoren dieser Arbeit unter dem Mikroskop das gediegene Silber sehr selten feststellen.

SCHLUSSFOLGERUNGEN

Die Gold-Silber haltigen Vererzungen der Blei-Zinklagcrstactte in der naehe von Gümüşhane werden in den ganzan magmatischen und sedimentaeren Gesteinsfolgen beobachtet, die in Zeitabstaenden zwischen Paleozoikum und Eozaen entstanden sind. Die hydrothermale Vererzung hat hauptsachlich die Störungszonen ausgcwachlt. Die Hazine Mağara und Kırkpavili Vorkommen wurden in Oberkreide-Kalksteinen, entlang der Störungszonen metazomatisch entstanden. Die Vererzung weist mit tertsieren Granitoiden naehere Beziehungen auf. Unter der Erzmineralien der Blei-Zinklagerstaette Gümüşhane gibt es kein Erzmineral, das als geologisches thermometer gelten kann. Deshalb kann man die Entstehungstemperatur der Blei-Zinklagerstaette nicht genau bestimmen. Aber mineralogischen Untersuchungen deuten darauf hin, dass die Erzmineralien dieser Lagerstaette waehrend der mezothermalen Phase entstanden sein können. Nach der mikroskopischen Untersuchungen der Erzproben, die aus Hazine Mağara und Kırkpavili Vorkommen stammen, wurde 3 verschiedene Erzmineralgruppen festgestellt.

1- Rutil-Anatas, Zirkon, Titanit, Graphit, Serisit, Chlorit, ein Teil von Pyrit, Quarz und kohle stammen aus der Nebengestein (Kalkstein). Diese Minerale wurden auch in den Nebengestein beobachtet.

2- Der grosste Teil von Pyrit, Bleiglanz, Zinkblende, Fahlerz, Aikinit, gediegenes Gold, Elektrum, gediegenes Silber, Kupferkies, Boumonit, Boulangerit, Enargit-Luzonit, Bornit, Mawsonit, Klaprothit, Galenobizmutit, Hessit, Arsenkies, Emplektit, Wittichenit, Altait, Tetradymit, Magnetkies, ein Teil von Quarz, Calcit, Dolomit, Baryt, Siderit und Ankerit wurden aus den hydrothermalen Lösungen entstanden.

3- Kalkosin, Covellin, Limonit, As-Sb-Oker, Anglesit, Cerussit, Smithsonit, Malachyt, Azurit, Psilomelen und *Pyr*olusit sind Sekundaermineralien, die durch Umwandlung aus primaeren Mineralien entstanden wurden.

Hier sind ntürlich die Mineralien wichting, die aus hydrothermalen Lösungen entstanden sind. Nach dieser Untersuchungen wurden viele unbekannte Bizmosulfosalzen und Tellur-Mineralien festgestellt, die der Erzparageneze von Blei-Zinklagerstaette Gümüşhane angehören. Sie sind: Aikinit, Bi-Pb-Fahlerz (Annivit), Tetradymit, Wittichenit, Emplektit, Klaprothit, gediegenes Wizmut und Hessit. Der Hessit ist gelegentlich ein wichtiges Silbermineral der Lagerstaette von Gümüşhane. Fast gleiche Mineralien wurden auch in den Lagerstaetten von Ost-Schwarzmeerkuste festgestelldt (Çağatay, 1979; Arman und Altun, 1983; Altun, 1984; Özgür und et al., 1989). Die Wizmut-Tellurmineralien, die im Erzparagenese der Lagerstaette von Gümüşhane vorkommen, wurden in Bleiglanz beobachtet. Dagegen die in der Lagerstaette von Ost-Schwarzmeerkuste auftretende Bi-Te-Mineralien findet man immer in Kupferkies.

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TAFELN

TAFEL - I

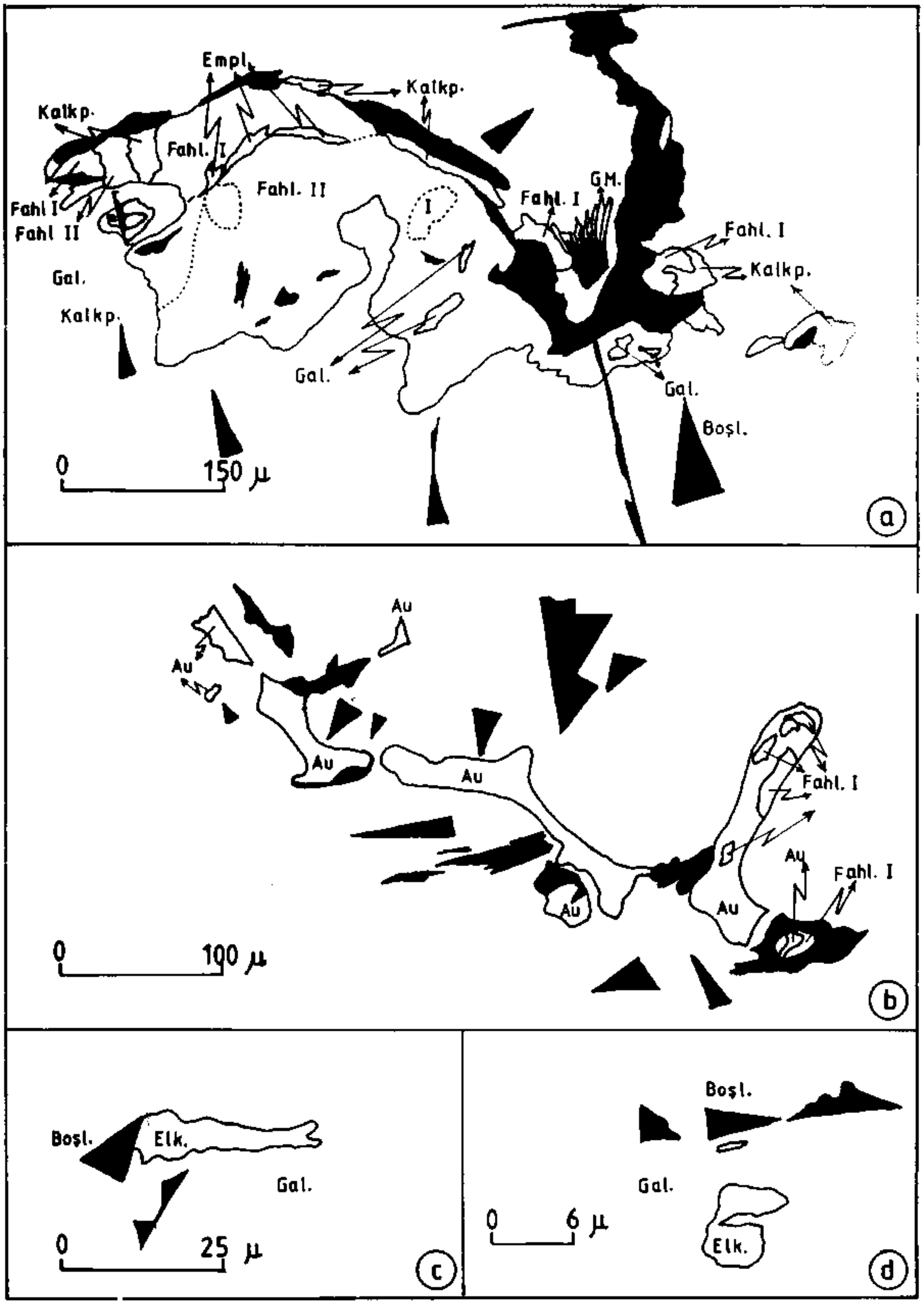
Die Mineralparagenese des Blei-Zinkvorkommen von Gümüşhane

Abb. a - Tennantit-Tetraedrit (Fahl I),
Pb-Bi-Fahlerz (Fahl II),
Kupferkies (Kalp.),
Emplektit (Empl.) in Bleiglanz.
Lokation : Hazine Mağara-Vorkommen

Abb. b - Gediegenes Gold (Au) und
Tennantit-Tetraedrit (Fahl. I) in Bleiglanz.

Abb. c - Elektrum (Elk.) in Bleiglanz und
Spaltbruche (Boşl.).

Abb. d - Elektrum (Elk.) in Bleiglanz.

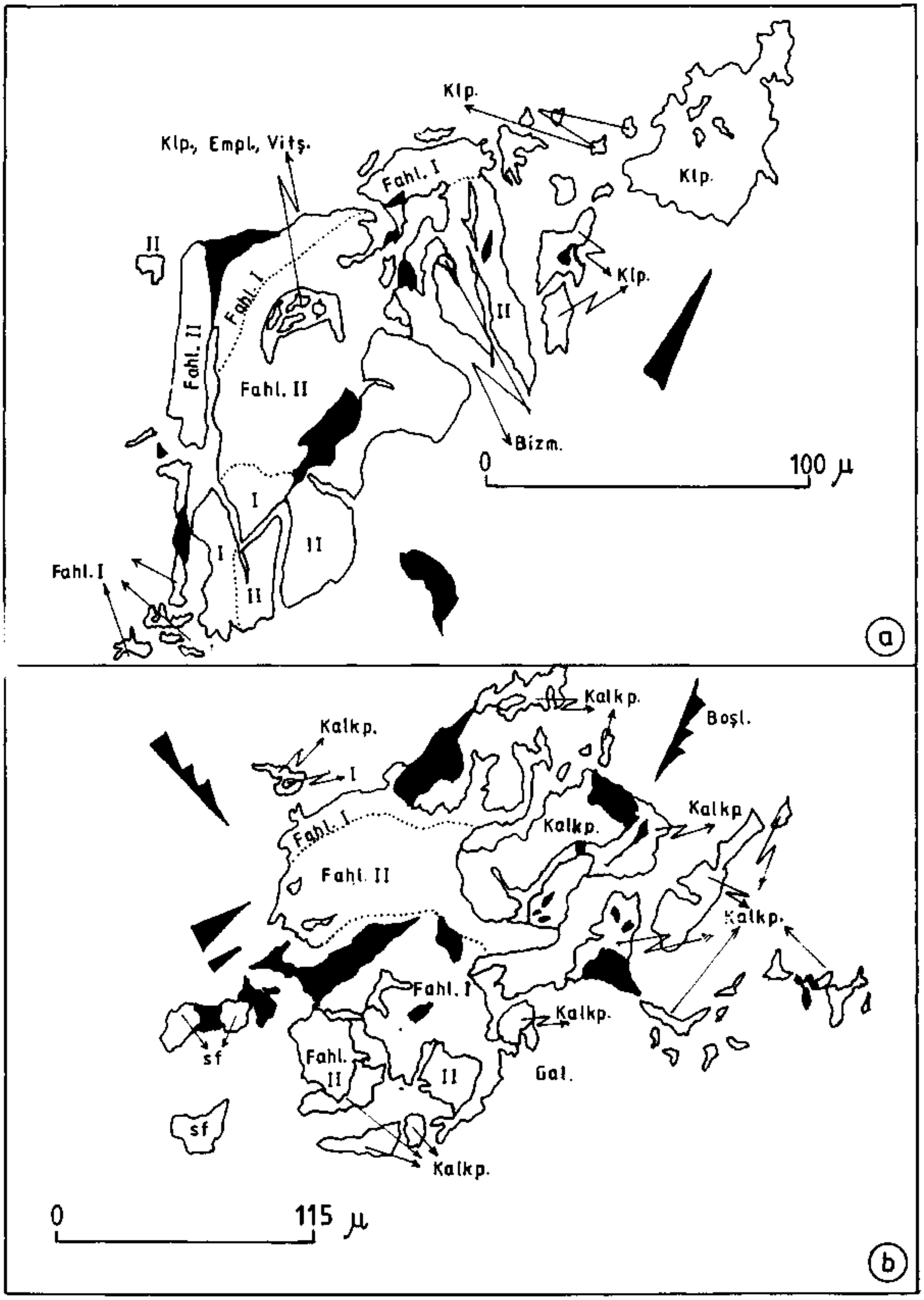

Empl.
Kalkp.
Kalkp.
Fahl. I
Fahl. II
Fahl I
Fahl II
I
G.M.
Fahl. I
Fahl. I
Kalkp.
Gal.
Kalkp.
Gal.
Gal.
Boşl.
0 150 µ
a
Au
Au
Au
Au
Au
Au
Au
Fahl. I
Au
Fahl. I
0 100 µ
b
Boşl.
Elk.
Gal.
0 25 µ
c
Boşl.
Gal.
0 6 µ
Elk.
d

TAFEL-II

Die Mineralparagencse Des Blei-Zinkvorkommen Von Gümüşhane

Abb. a - Klaprothit (klp.),
Emplektit (Empl.),
Witlishenit (Vits.) und
Pb-Bi-Fahlerz (Fahl. II) in Bleiglanz.

Abb. b - Tennantit-Tetraedrit (Fahl. I),
Bi-Pb-Fahlerz (Fahl, II),
Klaprothit (Klp.) und
Zinkblende (sf) in Bleiglanz.


Klp.
Klp., Empl., Vitş.
Fahl. I
Klp.
II
Fahl. I
Fahl. II
Klp.
II
Fahl. II
Bizm.
0
100 μ
I
I
II
II
Fahl. I
a
Kalkp.
Boşl.
Kalkp.
I
Kalkp.
Fahl. I
Kalkp.
Fahl. II
Kalkp.
Fahl. I
Kalkp.
sf
Fahl. II
II
Gal.
sf
Kalkp.
0
115 μ
b

ROCK AND CLAY MINERALOGY OF THE UPPER CRETACEOUS-TERTIARY SEDIMENTARY SEQUENCE AROUNDBURDURLAKE

Emel BAYHAN** and Hüseyin YALÇIN***

ABSTRACT. - Rocks and clay mineralogy of the Upper Cretaceous marine and Neogene shallow water units around Burdur lake have been studied. Characteristic minerals of the Upper Cretaceous are dolomite and corrensite, Paleocene-Lower Eocene is smectite, Neogene are aragonite, analcime and smectite. Calcite, quartz, feldspar, illite and chlorite are commonly found in all these units. It has been shown that rock units of different age and environment differ from one another in respect to their rock and clay mineralogy. Rock and clay mineralogy of the units have been studied and their mode of occurrence discussed.

MINERALOGIES OF THE METEORITES FALLEN IN TURKEY

Ahmet ÇAĞATAY**** and İbrahim ÇOPUROĞLU****

ABSTRACT. - In this study mineral compositions, textures, structures and their relationships with one another of three meteorites which have fallen in various parts of Turkey have been examined. The Şeyhhalil meteorite, being the one of these, displays similar mineral paragenesis, textures and composition with those of the Bursa meteorite. Minerals which occur in both of these meteorites have been differentiated as silicate and ore minerals. Silicate minerals that common in both are orthopyroxene, olivine, plagioclase, serpentine, talc, seriate and clay. Ore minerals, in addition, are kamacite, troilite, chromite, taenite, native copper, ilmenite, mackinawite, and limonite. Trace and very small amounts of rutile, chalcopyrrhotine, whitlockite and apatite arc also observed in spongy-like Şeyhhalil meteorite. According to their mineralogical compositions, both of the meteorites might be considered within the "siderolite" class. Ağrı meteorite, on the other hand, might be included within the "octahedral" class and is composed mainly of very scarce amount of troilite; besides, kamacite, taenite and plessite of three different iron-nickel minerals.

PETROLOGICAL COMPARISON OF THE UPPER CRETACEOUS - LOWER TERTIARY BASINS OF THE "ANKARA VIRGATION"

Emel BAYHAN* and Sungu L. GÖKÇEN**

ABSTRACT. - In the present work, the Upper Cretaceous-Lower Tertiary sequence in the zone named as Ankara Virgation, covering Haymana-Polath, Northern Tuz Gölü, and Kırıkkale-Yahşihan regions in SW-NE direction has been studied from the sedimentary - geological point of view, and the sedimentological and sedimentary petrological properties of those three basins have been correlated. It has been determined that detritic sequences in all three basins have similar petrographical properties with the exception of certain micro - mineralogical differences, and that the arenites and other detritic materials in the region have been derived from the magmatic and metamorphic rocks situated roughly in the North and the South. Petrographic study of the elastics carried out on sandstones (Dickinson, 1982) have shown that these rocks had a "recycled orogen provenance", and that the sequences in the region had been deposited during the Upper Cretaceous-Lower Tertiary around an active collision zone of plates in relation with a subduction facies.

SULFUR ISOTOPE STUDY OF KURŞUNLU (ORTAKENT-KOYULHİSAR-SİVAS) VEIN TYPE Pb-Zn-Cu DEPOSITS

Ahmet GÖKÇE ***

ABSTRACT. - Kurşunlu Pb-Zn-Cu deposits are the typical examples of the vein type deposits which are widely seen all over the southern and western parts of Eastern Black Sea region. They are deposited along the fault zone which cut the Upper Cretaceous, mostly andesitic, partly dacitic volcanic and volcanosedimentary rocks. Galena, sphalerite, pyrite, chalcopyrite, calcocite and hematite are seen as ore minerals while quartz, calcite and small amount of barite occur as gangue minerals. Isotopic composition ($d^{34}S_{CDT}$) of the galena, sphalerite, pyrite and chalcopyrite mineral separates from these mineralizations, are as follows; -6.6 - -8.4 °/°°, -4.6 -- 7.6 °/°°, -4.3 - -6.0 °/°° and -3.7--6.3 °/°°. These are the negative values range between -3.7 and -8.4 °/°°. A possible isotopic equilibrium seems to be developped between sphalerite and galena, and it suggests an average formation temperature of 327°C according to the sulfur isotopes fractionation thermometer. According to these isotopic composition; It is very difficult to identify the source of the sulfur in these ore veins as magmatic, marinal or biologic. But it may be suggested that the sulfur in this composition have been produced as follows; A magmatic sulfur with an isotopic composition nearly zero ($d^{34}S$) which dissolved from the surrounding volcanic and volcanosedimentary rocks by deep circulated suificial water was shared between sulfates (which useheavier isotopes; such as barite) and sulfides (which use lighter isotopes) of the veins as parallel to the isotopic fractionation trend.

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GEOCHEMICAL PROXIMITY INDICATORS OF THE MURGUL VOLCANOGENIC COPPER DEPOSIT, EAST PONTIC METALLOTECT NE TURKEY

Nevzat ÖZGÜR* and Carlos M. PALACIOS**

ABSTRACT . - A study on the behaviour of F, Tİ, Mn, Cu, Au, and REE during hydrolhermal mineralization at the Murgul volcanogenic copper deposit reveals that Ti, Mn, and REE are strongly depleted in altered host rocks whereas F, Cu, and Au show remarkable positive anomalies in the altered mineralized areas. We propose the use of the elements F, Ti, and Mn as proximity indicators for exploration of concealed ore deposits of the same type in the East Pontic metallogenetic province of Turkey.

INTRODUCTION

The Murgul volcanogenic Cu-deposit comprises one of the principal copper ore districts of Turkey (Fig. 1). The open pits are located 7 km SE of the town Murgul. The area belongs to the East Pontic metallogenetic province in which a considerable number of base metal deposits are located (Çağatay and Boyle, 1977, 1980; Akın, 1979; Akıncı, 1980; Dieterle, 1986). This zone, however, has not been investigated intensively to find new ore deposits.

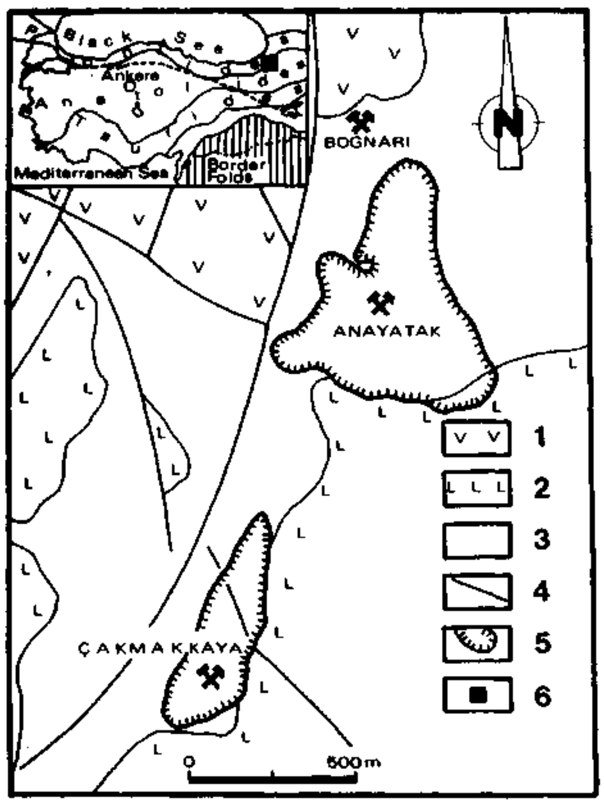

Fig. 1 - Geological sketch map of the Murgul deposit.
1 - Andesitic lava flows of the uppermost Cretaceous; 2- Hanging-wall felsic volcanics; 3- Pyroclastic host rocks; 4- Main faults, generally vertical movements; 5- Limits of the open pits (1983); 6- Investigated area of Murgul.

Detailed results of the geochemical investigations on the chemical solubility and depletion of Ti in the altered rocks under thermodynamic conditions will be reported in a separate paper. The aim of our paper is to discuss the geochemical behaviour of F, Ti, Mn, Cu, Au, and REE at the Murgul deposit during hydrothermal mineralization. We attempt to develop with these data an exploration model which could be applicable for practical use in this area. Previous geological investigations (Özgür, 1985; Dieterle, 1988; Schneider et al. 1988) of the deposits support the interpretations presented in this paper.

GEOLOGICAL SETTING

The East Pontic metallogenetic province represents a volcanic island arc system of Jurassic through Miocene age which hosts a great number of base metal deposits (Akın, 1979; Akıncı, 1980; Dieterle, 1986; Özgür and Schneider, 1988; Schneider et al., 1988). The East Pontides extend over an area of more than 350 km E-W and 60 km N-S and represent the mobile belt between the Pontic and Anatolian plate. The ratio of economically important base metal deposits changes along the general strike of the metallogenetic province from east (Cu >>Pb+Zn) to west (Pb+Zn >>Cu). The East Pontides consist of a 2,000 to 3,000 m thick sequence of volcanic rocks with minor intercalations and lenses of marine sediments (Fig. 2) which have been divided into three stratigraphic cycles (Maucher, 1960; Maucher et al., 1962):

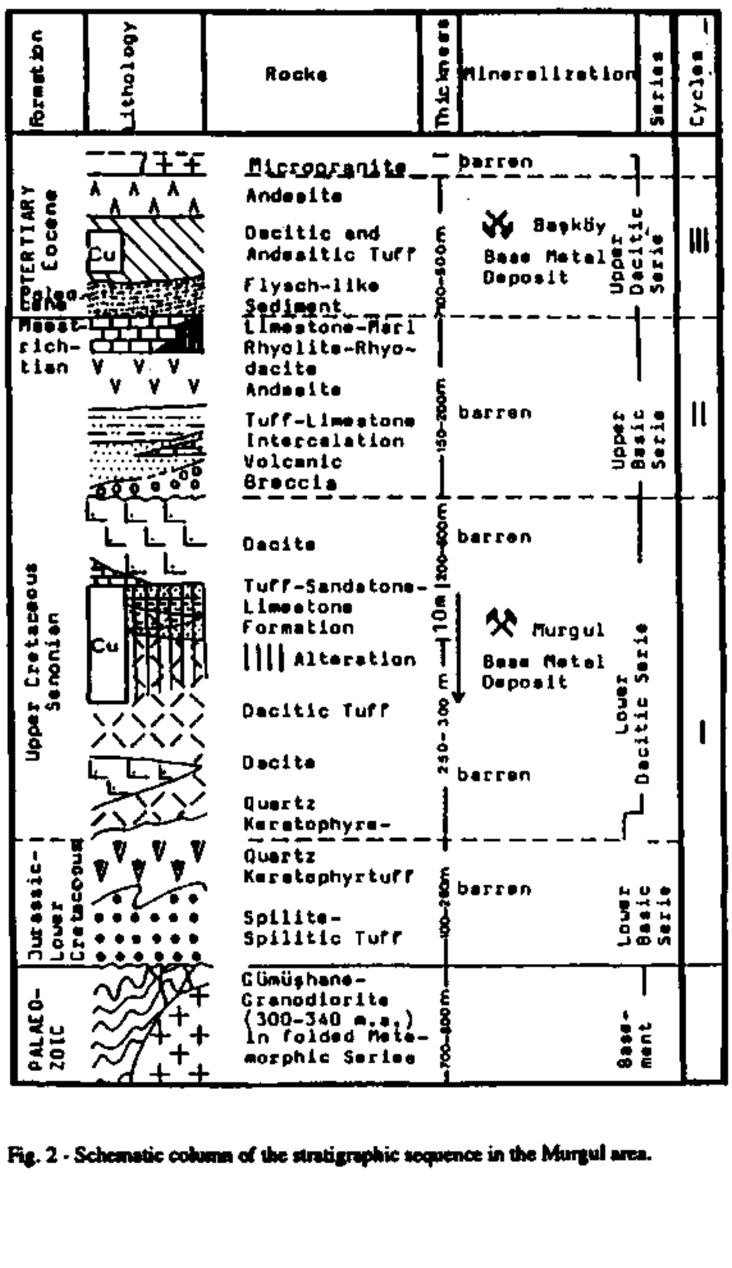

Fig. 2 - Schematic column of the stratigraphic sequence in the Murgul area.

1. The first cycle comprises a volcanic pile deposited between Jurassic and Upper Cretaceous. It is represented by initial basaltic activity (spilites) which changes progressively to felsic lava flows and thick pyroclastics in the middle and upper part.

2. The second cycle starts with volcanic breccias, tuffs and minor intercalations of marine sediments overlain by andesitic and rhyolitic lava flows, followed by limestones of uppermost Cretaceous age (Maastrichtian).

3. The last cycle consists of a basal sequence of marine sediments of Paleocene age which are overlain by andesitic and basaltic lava flows representing Tertiary volcanic activity.

The Murgul deposit is linked to the upper part of the first volcanic cycle and is associated with a 250 m thick felsic pyroclastic sequence. The top of the deposit is marked by a thin layer of marine sediments (Sawa and Sawamura, 1970; Mado, 1972; Buser and Cvetic, 1973) and is characterized by intense erosion and weathering (Özgür, 1985). This sequence is overlain by 200-500 m thick barren felsic volcanites. The age of mineralization in the pyroclastic sequence is pre-Maastrichtian according to paleontological observations (Buser and Cvetic, 1973).

THE MURGUL ORE DEPOSIT

The Murgul deposit consists of at least two primary orebodies (Anayatak and Çakmakkaya; Fig. 1) hosted in the same volcanic member, spanning a horizontal distance of about 500 m. According to former descriptions (e.g. Sawa and Sawamura, 1970; Mado, 1972) and our observations and interpretations of drilling profiles, the mineralization of both orebodies shows exactly the same feature. A third minor orebody, Bognari, came into production recently (Fig. 1) which has been interpreted by Mado (1972) as an erosional product of the upper part of the Anayatak orebody. The sulfide mineralization of both orebodies contains predominantly pyrite and lesser chalcopyrite. Minor quantities of galena, sphalerite, and fahlore occur locally only. Additionally, minor occurrence of aikinite, hessite, tetradymite, clausthalite, and free gold have also been determined by electron microprobe analysis (Willgallis et al., 1989).

The copper deposit consists of (1) widespread disseminated ore with varying Cu contents ranging from 0.2 to 0.7 percent, (2) stockwork ore with average Cu contents between 1.0 and 2.5 percent, and (3) small ore lodes with Cu contents from 5.0 to 10.0 percent (Schneider et al., 1988). The recoverable ore reserves are estimated at 40 million metric tonnes with an average content of 1.25 percent Cu, 0.1 percent Zn, 25 ppm Ag, and 0.2 ppm Au.

The ore mineralization may be divided into an early stage associated with a phyllic zone surrounded by a peripheral argillic zone, and a late stage related to a central pervasive silicification (Fig. 3). According to Schneider et al. (1988), the first stage of alteration led to destruction of the primary paragenesis of the pyroclastics and replacement of the host rock by quartz and pale greasy sericite. This stage reveals poor mineralization of disseminated pyrite and chalcopyrite (type one). The late stage of hydrothermal activity in this area of the deposit is represented by silicic alteration that appears as quartz replacement of the volcanic host rock, as cryptocrystalline varieties of jasper, and later on, as open-space fillings (quartz- ore veins). The sulfide mineralization of this stage represents the principal commercial ore (types two and three). The surrounding country rocks show pervasive argillization which characterized by an alteration assemblage containing quartz, montmorillonite, illite, dickite, and pyrite only.

SAMPLING AND ANALYTICAL METHODS

Various rock samples have been obtained from altered and mineralized zones; 53 samples from the surface (Tab. 1 and Fig. 4) and 87 from deep drilling holes (Tab. 2). For comparison, the less altered background rocks of the pyroclastic flows (18 samples) have been analyzed too (Tab. 1), which were taken between 500 and 700 m outside of the mineralized and altered area. The background pyroclastics include host member in which the alteration is on a regional scale generally weak.

Rare earth elements (La, Ce, Sm, Eu, Tb, Yb, and Lu) and gold were determined by instrumental neutron activation at the Hahn-Meitner Institut fur Kernforschung, Berlin, with a routine precision better than ± 9 % for most elements (Dulski and Moller, 1975) using GSP-1 of the U.S. Geological Survey as the reference standard.

Ti, Mn, and Cu were determined by atomic absorption spectrometry, and F by ion-sensitive electrode at the Institut fur Geologie, Geophysik und Geoinformatik, Freie Universitat Berlin, with a precision better than ± 5 %. For all analyses, BCR-1 and GSP-1 rock standards have used.

Element Sample	F (ppm)	Ti (ppm)	Mn (ppm)	Cu (ppm)	Au (ppb)	ΣREE (ppm)	Element Sample	F (ppm)	Ti (ppm)	Mn (ppm)	Cu (ppm)	Au (ppb)	ΣREE (ppm)
1	740	850	50	30	n.d.	n.d.	37	410	1250	150	10	3	33.8
2	655	900	100	90	n.d.	n.d.	38	430	1500	205	10	4	66.8
3	545	3650	550	45	n.d.	n.d.	39	425	500	40	4500	n.d.	n.d.
4	555	550	50	2600	n.d.	n.d.	40	575	800	50	1650	n.d.	n.d.
5	860	700	100	75	n.d.	n.d.	41	1100	1100	50	185	n.d.	n.d.
6	1035	650	50	55	n.d.	n.d.	42	965	650	50	60	n.d.	n.d.
7	490	1300	100	90	n.d.	n.d.	43	180	500	50	190	80	15.1
8	265	650	100	310	n.d.	n.d.	44	n.d.	n.d.	n.d.	n.d.	30	37.9
9	490	550	50	300	n.d.	n.d.	45	n.d.	n.d.	n.d.	n.d.	70	35.0
10	665	650	100	305	n.d.	n.d.	46	225	500	50	275	60	19.4
11	55	125	50	160	n.d.	n.d.	47	135	500	100	30	30	16.0
12	625	800	110	320	n.d.	n.d.	48	165	1150	10	5	2	45.6
13	1965	n.d.	50	4950	n.d.	n.d.	49	n.d.	n.d.	n.d.	n.d.	20	7.0
14	1690	200	100	6750	n.d.	n.d.	50	1280	800	950	10	2	37.2
15	85	500	35	3700	n.d.	n.d.	51	745	550	50	115	6	28.7
16	335	300	50	13000	n.d.	n.d.	52	30	500	40	225	80	1.9
17	1650	100	50	80	n.d.	n.d.	53	415	550	50	7250	4	20.0
18	140	150	100	10000	n.d.	n.d.	54	240	3000	965	35	3	37.3
19	460	700	100	70	n.d.	n.d.	55	380	5750	810	15	3	31.2
20	630	1000	120	60	n.d.	n.d.	56	345	3550	1370	40	2	33.7
21	930	300	50	175	n.d.	n.d.	57	285	3100	4800	70	2	45.6
22	165	2300	400	5	n.d.	n.d.	58	215	3750	1000	15	2	38.6
23	415	2600	370	80	n.d.	n.d.	59	345	4250	1700	20	2	40.7
24	1810	400	355	355	n.d.	n.d.	60	230	4250	1130	50	2	43.2
25	1030	500	90	90	n.d.	n.d.	61	300	2400	550	10	3	49.4
26	340	250	300	2300	n.d.	n.d.	62	360	225	35	45	2	21.5
27	315	350	100	465	n.d.	n.d.	63	305	1000	35	370	2	22.4
28	2515	700	60	385	n.d.	n.d.	64	700	2000	600	45	3	44.6
29	1370	1250	75	110	n.d.	n.d.	65	455	3000	545	15	4	58.9
30	1235	1400	210	15	n.d.	n.d.	66	290	1750	40	15	2	26.5
31	435	600	55	380	2	18.4	67	800	n.d.	n.d.	n.d.	n.d.	n.d.
32	1980	500	100	400	n.d.	n.d.	68	595	1000	40	25	2	20.1
33	770	1100	565	6	2	52.8	69	153	n.d.	n.d.	n.d.	n.d.	n.d.
34	440	250	100	210	n.d.	n.d.	70	176	3750	1700	15	2	34.9
35	230	700	50	85	2	16.7	71	265	2500	925	610	2	49.2
36	535	1400	40	120	2	31.9							

* for locations see Figure 3.
n.d. not determined.

Table 2 - Element data of the altered drill holes (A and B) from both ore bodies of Anayatak (samples 70 to 118) and Çakmakkaya (samples 119 to 158) *

Element Sample	F (ppm)	Ti (ppm)	Mn (ppm)	Cu (ppm)	Au (ppb)	ΣREE (ppm)	Element Sample	F (ppm)	Ti (ppm)	Mn (ppm)	Cu (ppm)	Au (ppb)	ΣREE (ppm)
72	425	375	100	95	n.d.	n.d	116	480	1250	n.d.	11	n.d.	n.d.
73	n.d.	10	n.d.	24400	430	1.5	117	800	2640	300	16	n.d.	n.d.
74	460	460	1550	55	n.d.	n.d.	118	710	2500	n.d.	29	20	33.1
75	395	1230	n.d.	100	n.d.	n.d.	119	140	430	50	45	130	8.0
76	465	550	n.d.	85	n.d.	n.d.	120	75	n.d.	350	305	n.d.	n.d.
77	420	1600	100	1580	130	21.2	121	141	340	50	360	n.d.	n.d.
78	415	660	n.d.	15	n.d.	n.d.	122	50	80	150	430	n.d.	n.d.
79	260	730	300	35	n.d.	n.d.	123	260	100	50	300	n.d.	n.d.
80	605	1340	n.d.	12	n.d.	n.d.	124	50	10	n.d.	22500	n.d.	n.d.
81	235	1300	100	46	8	29.5	125	45	100	n.d.	21500	n.d.	n.d.
82	130	570	n.d.	930	n.d.	n.d.	126	30	10	n.d.	11000	1700	n.d.
83	175	750	50	850	n.d.	n.d.	127	95	120	n.d	23700	n.d.	n.d.
84	340	1560	n.d.	20	n.d.	n.d.	128	45	30	n.d.	3800	n.d.	n.d.
85	450	1000	50	150	n.d.	n.d.	129	40	10	100	660	160	n.d.
86	400	1180	n.d.	44	28	30.3	130	270	400	50	380	n.d.	n.d.
87	540	1620	n.d.	160	n.d.	n.d.	131	20	30	n.d.	8400	n.d.	n.d.
88	320	700	100	40000	n.d.	n.d.	132	225	200	50	22000	n.d.	n.d.
89	675	1150	350	75	15	31.8	133	210	380	n.d.	3400	n.d.	n.d.
90	420	920	n.d.	460	n.d.	n.d.	134	35	10	n.d.	10000	300	n.d.
91	850	1350	500	335	n.d.	n.d.	135	335	780	n.d.	1270	n.d.	n.d.
92	435	840	n.d.	13500	n.d.	n.d.	136	300	390	n.d.	9400	780	15.0
93	520	910	250	9000	160	21.8	137	125	260	n.d.	13000	390	9.8
94	1235	5200	n.d.	42	n.d.	n.d.	138	190	380	n.d.	8300	n.d.	n.d.
95	710	1100	250	23000	n.d.	n.d.	139	125	200	n.d.	8100	n.d.	n.d.
96	635	1180	900	1450	n.d.	n.d.	140	270	500	n.d.	6200	130	12.4
97	620	460	n.d.	152	n.d.	n.d.	141	265	560	n.d.	11800	6000	15.0
98	960	1620	n.d.	26	17	36.6	142	180	470	500	12000	260	13.2
99	780	750	300	80	n.d.	n.d.	143	170	270	650	1040	n.d.	n.d.
100	680	1160	n.d.	31	n.d.	n.d.	144	420	400	350	675	n.d.	n.d.
101	1100	1300	250	55	n.d.	n.d.	145	140	220	250	2300	n.d.	n.d.
102	635	2030	n.d.	30	n.d.	n.d.	146	305	305	50	39500	n.d.	n.d.
103	265	3040	250	19	n.d.	29.8	147	180	170	n.d.	920	n.d.	n.d.
104	795	1830	n.d.	10	n.d.	n.d.	148	90	100	50	13500	n.d.	n.d.
105	880	800	100	55	n.d.	n.d.	149	30	40	n.d.	1990	n.d.	n.d.
106	475	1200	n.d.	20	n.d.	n.d.	150	200	500	150	2200	n.d.	n.d.
107	495	1700	n.d.	16	n.d.	n.d.	151	565	1570	100	60	n.d.	16.1
108	930	3390	650	420	11	39.7	152	640	1580	100	200	n.d.	n.d.
109	625	n.d.	n.d.	n.d.	n.d.	n.d.	153	150	n.d.	50	3550	n.d.	n.d.
110	690	950	200	75	n.d.	n.d.	154	180	400	100	13500	n.d.	n.d.
111	695	690	n.d.	74	n.d.	n.d.	155	420	1320	350	40	n.d.	19.6
112	435	1380	n.d.	22	n.d.	n.d.	156	120	240	n.d.	34000	n.d.	n.d.
113	620	1160	n.d.	25	n.d.	n.d.	157	230	470	n.d.	60	n.d.	19.6
114	820	1250	250	90	27	31.1	158	100	70	n.d.	3200	n.d.	n.d.
115	510	1380	n.d.	3700	n.d.	n.d.							

* for locations see Figure 3.
n.d. not determined.

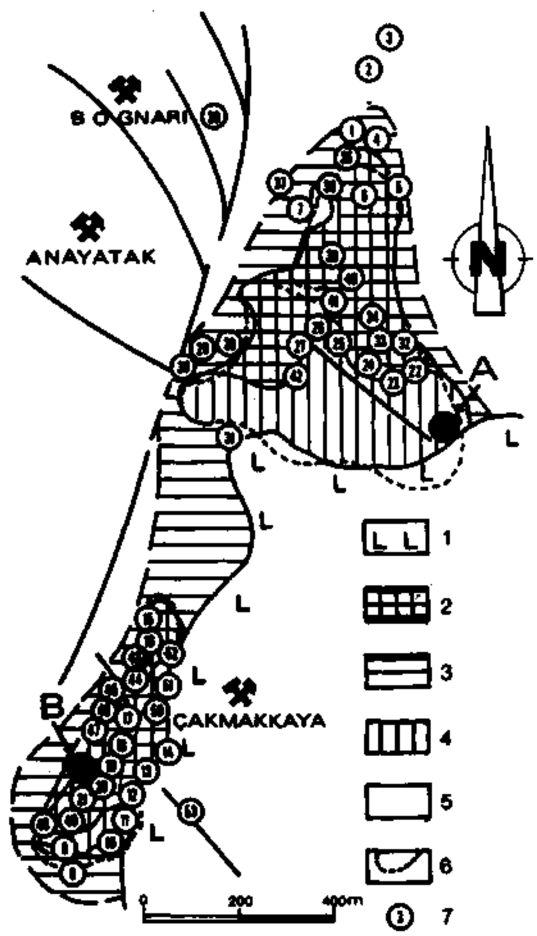

Fig. 3 - Map showing hydrothermal alteration zones and sample locations from the surface of both orebodies.
1- Hanging-wall felsic volcanics; 2- Silicic alteration; 3- Argillic alteration; 4- Phyllic alteration; 5- Pyroclastic host rocks; 6- Limits of the open pits (state of mining: 1983); 7- Location of the samples for the analyses.

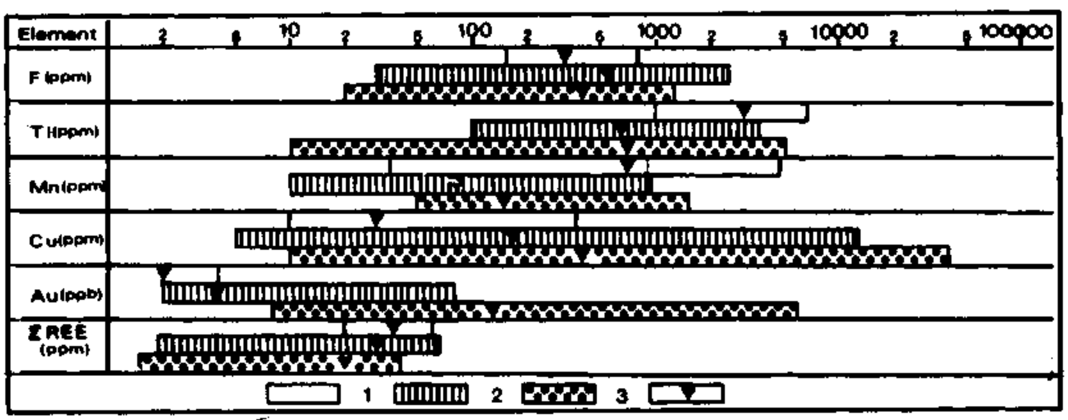

Fig. 4 - Background and range of the content of the elements in 1- Less altered background rocks, 2- Host pyroclastics, and 3- Drill holes; ▼ - background value.

The less altered samples formed the basis for determinating regional backgrounds of each element For the statistical evaluation, a computer program "Geo-500" together with "Stasy" and "Easy" of Company PIC, Munich/West-Germany, was applied to determine geochemical parameters and anomalous values. Additionally, the method of Lepeltier (1969) was constituted to establish anomalous populations.

RESULTS

Data on the elements F, Ti, Mn, Cu, Au, and REE in the altered volcanic host rocks indicate extensive geochemical dispersion halos and anomalies within the area of both orebodies (Figs. 5 and 6). The contents of F, Cu, and Au increase in the altered areas remarkably whereas Ti, Mn, and REE exhibit a distinct depletion.

Cu is enriched in phyllic and silicic alteration areas with concentrations greater than 220 ppm. This element is present with a background of about 30 ppm in the less altered pyroclastics (Fig. 4) and shows a higher value of 60 ppm in argillie zones (Fig. 5 and 6).

The two and three dimensional distributions of fluorine in altered pyroclastic host rocks are shown in Figs. 5 and 6. Fluorine in less altered pyroclastics has a background value of 325 ppm (Fig. 4). Geochemical halos in the phyllic and silicic alteration zones exhibit values of 320 to 500 ppm and more than 500 ppm F are observable. Locally, the fluorine contents in both altered zones reach extreme values of up to 2515 ppm (Fig. 4).

Gold has a background value of 2 ppb in the less altered pyroclastic country rocks (Fig. 4). Gold distribution within the both orebodies displays anomalous areas (Fig. 7) represented by values between 2-80 and more locally values greater than 80 ppb (Fig. 4). Particularly remerkable is the distribution of higher Au values which are linked to parts of silicic alteration in a greater distance to the surface. Some sectors have shown local economic concentrations.

In contrast to the positive anomalies represented by F, Cu, and Au; Ti, Mn, and REE were intensely depleted in the altered and mineralized areas. The less altered pyroclastics show a Ti background value of 3000 ppm (Fig. 4). In the mineralized areas, Ti is obviously depleted (Figs. 5 and 6), especially in the areas of phyllic and silicic alteration which indicates a estimated background value of 650 ppm in the host rocks. Similar behaviour is shown by Mn (Figs. 5 and 6) which can reach concentrations of about 10 ppm whereas the regional background was estimated at 705 ppm (Fig. 4).

As reported by Schneider et al. (1988), the REE have been leached from altered host rocks. Fig. 8 shows the distributions of the REE values of the investigated areas. It is notable that the silicified host rocks display the more important negative REE anomalies.

DISCUSSION

The geochemical data from the Murgul deposit indicate that fluorine, titanium, and manganese are excellent indicators of volcanogenic sulphide deposits in the East Ponticmetallogeneticprovince. The Murgul deposit has been genetically interpreted as a subvolcanic type associated with Upper Cretaceous island arc volcanic activity (Akın, 1979; Özgür, 1985; Özgür and Schneider, 1988; Schneider et al., 1988).

The dispersion halos of F, Ti, Mn, Au, and REE outline perfectly the presence of a hydrothermal mineralization and alteration pattern in altered host pyroclastics. This has-been corroborated additionally by Çağatay and Boyle (1977) and Dietcrle (1986) in Madenköy, Sırtköy, and Kutlular ore deposits in the western part of the East pontic metallogenctic province which arc linked to similar type of pyroclastic host rocks in nearly the same stratigraphic horizon. They belong genetically to the subvolcanic hydrothermal mineralization too.

The increase of fluorine during hydrothermal alteration is a well known phenomenon: the fluorine contents of the primary host rock are increasing within the mineralized area because the element is concentrated by the ascending hydrothermal fluids. Due to the similar radii of F and (OH), fluorine can replace (OH) in the lattice of micas and clay minerals. This is well documented at the Murgul deposit in which the greater F values are concentrated within the phyllic and parts of the argillic zones (Figs. 5 and 6). The size of the deposit in the pyroclastic host rock stratigraphy seems to dictate the magnitude of the elemental haloes as evidenced by the fluorine distribution around Anayatak orebody in contrast to that

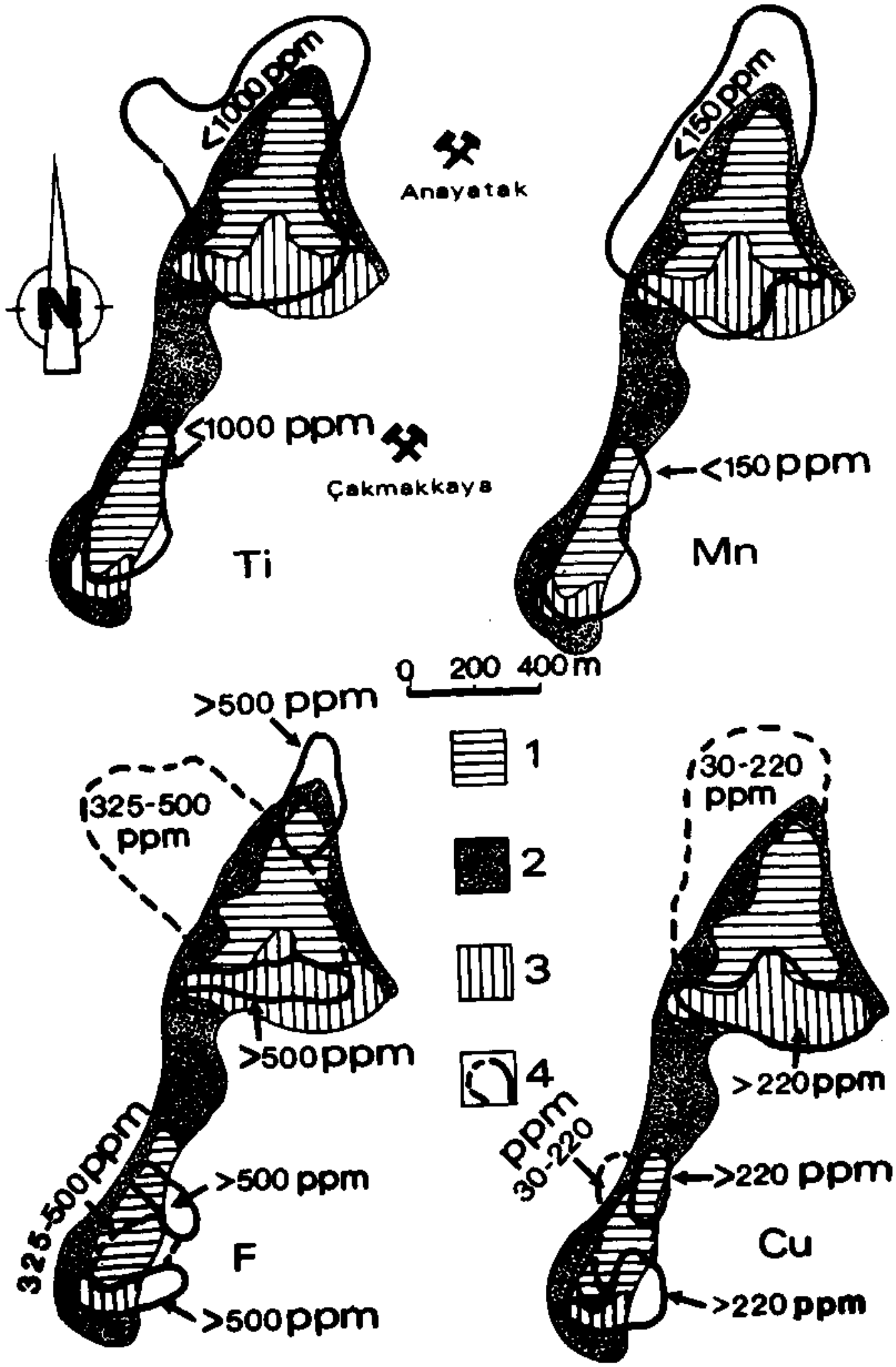

Fig. 5 - The distributions of Ti, Mn, F, and Cu in the Murgul deposit.
1- Silicic alteration; 2- Argillic alteration; 3- Phyllic alteration; 4- Boundary of anomalies.

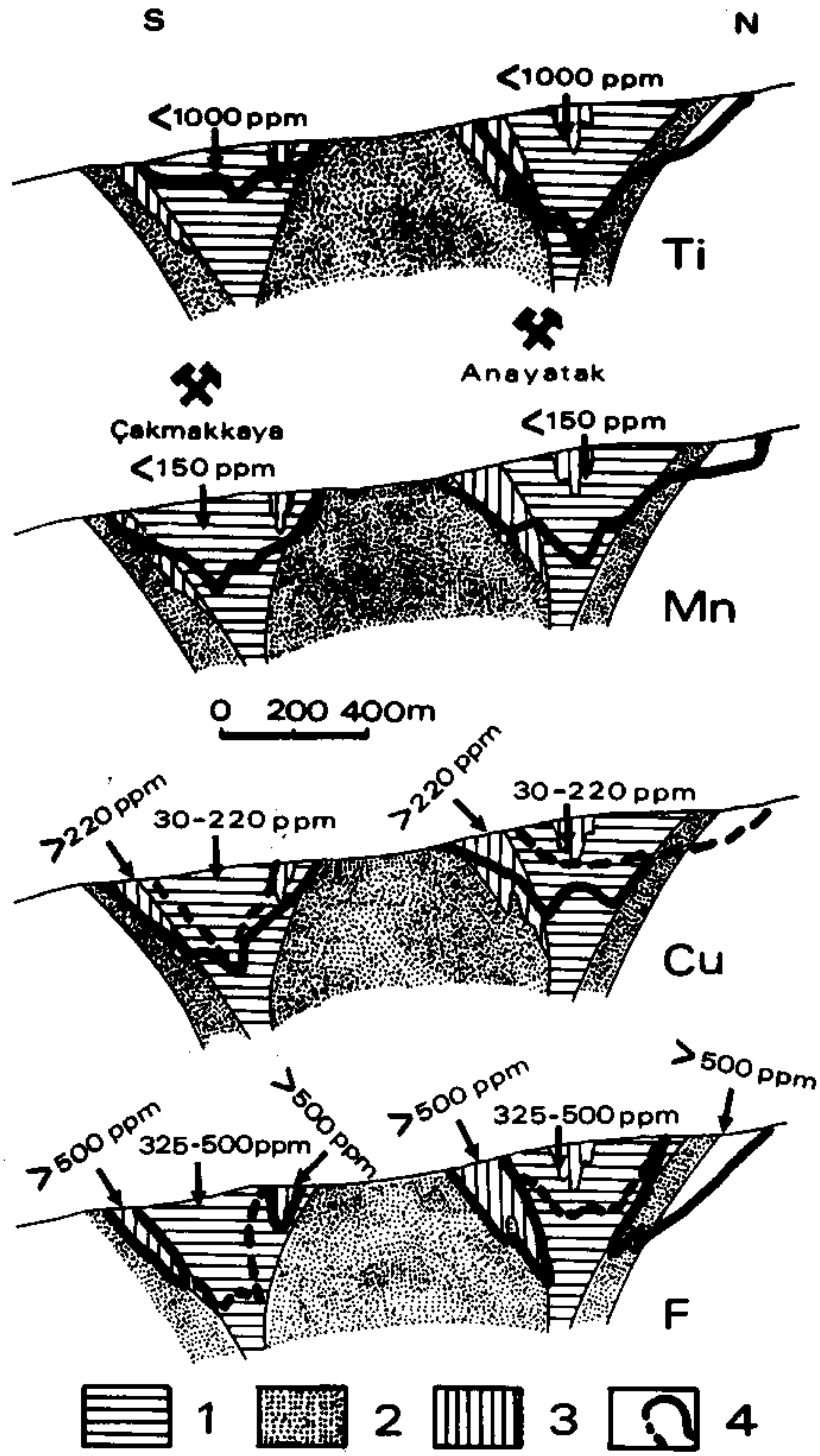

Fig. 6 - Cross sections showing the distributions of Ti, Mn, F, and Cu in the Murgul deposit.
1- Silicic alteration; 2- Argillic alteration; 3- Phyllic alteration; 4- Boundary of anomalies.

around Çakmakkaya which is to led the lack of the rock samples from Çakmakkaya due to hanging wall volcanics and soils. In addition to above, the analyzed host pyroclastics are derived from the silicic alteration area of Çakmakkaya. Therefore, these rock samples indicate low fluorine contents in comparison to Anayatak orebody.

The two and three dimensional distributions of manganese (Figs. 5 and 6) show strong negative anomalies within both ore bodies. This could be generated by the breakdown of Mn-bearing minerals (biotite, feldspars, and possibly glass). Thus, manganese was released from the rocks during alteration.

In the original pyroclastics, titanium is present in sphene and rare rutile or anatase. These minerals are not stable under the thermodynamic conditions during hydrothermal alteration. Therefore, titanium is leached from the altered areas too. Gold was especially enriched in silicic alteration zones (Fig. 7) which could be attributed to a hydrothermal remobilization of the Au contents of the host rocks.

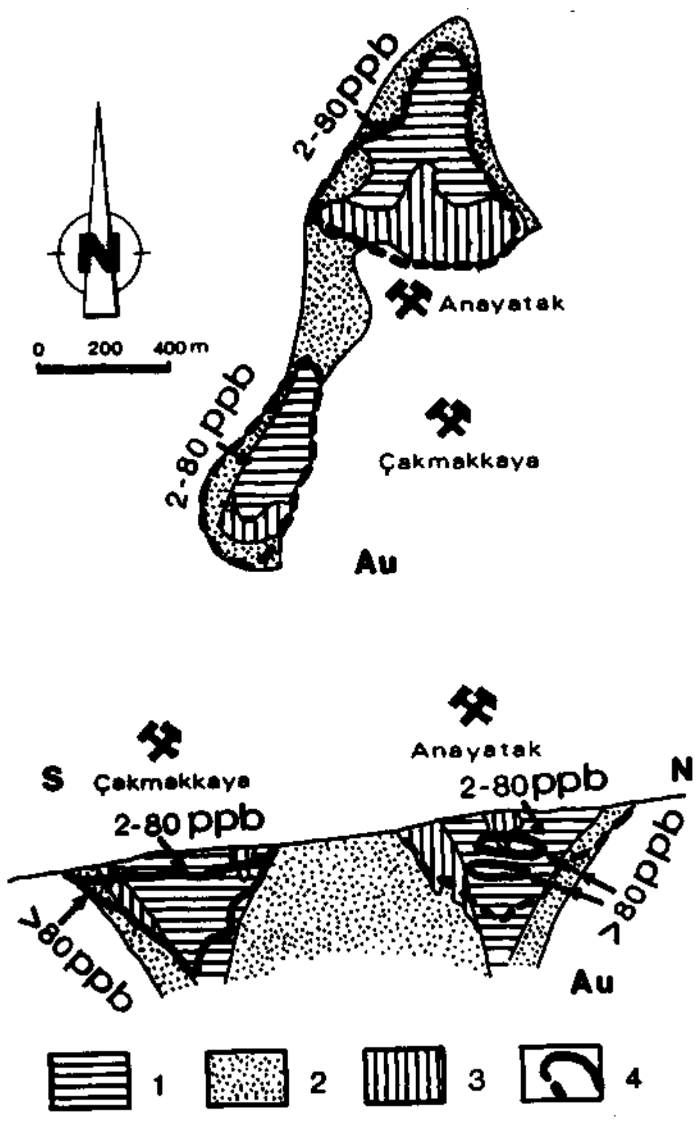

Fig. 7 - Gold distributions at the Murgul ore deposit.
1- Silicic alteration; 2- Argillic alteration; 3- Phyllic alteration; 4- Boundary of anomalies.

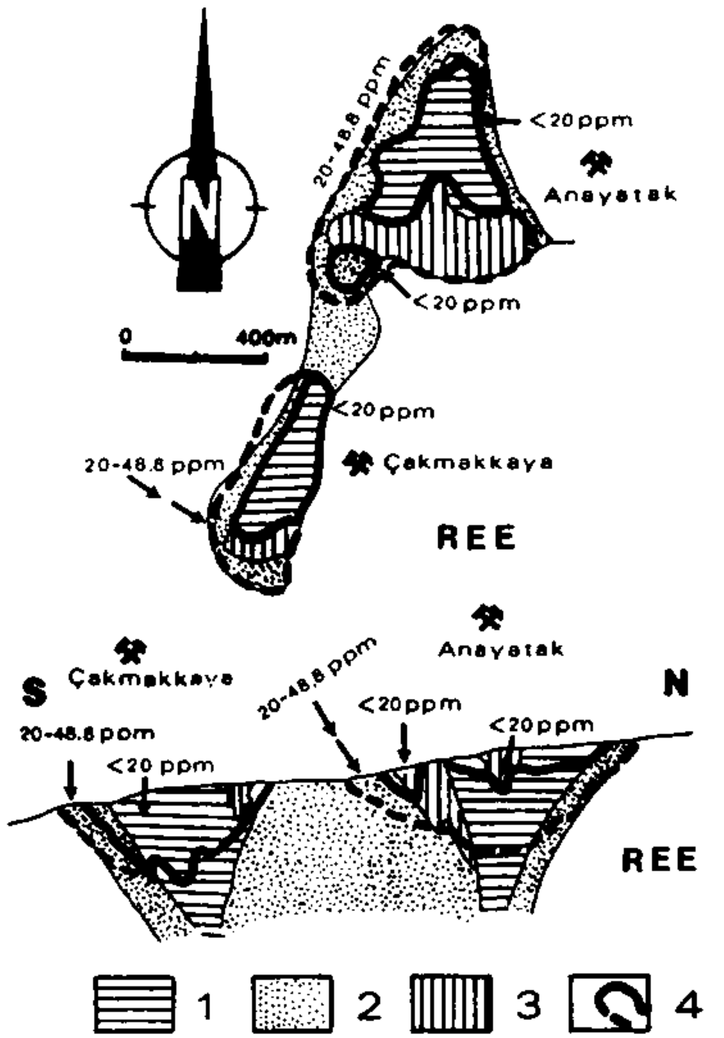

Fig. 8 - REE distributions at the Murgul ore deposit.
1- Silicic alteration; 2- Argillic alteration; 3- Phyllic alteration; 4- Boundary of anomalies.

The mineralization of Murgul seems to be strata-bound observing the semistratigraphic position of all deposits in the East Pontic metallogenetic province which arc linked to the volcanic sequence of Senonian age. Therefore, manganese, titanium, and fluorine could be applied as proximity indicators for the concealed deposits of the same type throughout the entire East Pontides.

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QUANTITATIVE TOTAL IRON ANALYSIS USING XRD - FLUORESCENCE RADIATION INTENSITY

Doğan AYDAL *

ABSTRACT. - Fluorescence radiation is being known as "unwanted incident", while XRD determination, and various methods arc in use to solve this secondary radiation. In this study, fluorescence radiation is especially created in order to show the possibility of quantitative analysis with the help of this secondary radiation. Some selected standards, which have different quantity of iron content, were chosen and Cu-target was intentionaly used during XRD determination, in order to cause fluorescence radiation in different intensity. Finally the positive relation was detected between the total iron content and fluorescence radiation intensity. It takes only 40 seconds to find the quantity of total iron in the analysed sample with the suggested method. In addition, another important result of this study is that the radiation intensity have direct relation with the total iron content in the sample, regardless of iron combination with other elements.

Mineral Res. Expl. Bull., 111.65-74. 1990

METAMORPHISM AND FISSION-TRACK AGE DETERMINATION OF APATITE CRYSTALS FROM DEMİRCİ-BORLU REGION, GÖRDES SUBMASSIF OF THE MENDERES MASSIF-WESTERN TURKEY

Osman CANDAN*; Cahit HELVACI*; G. BÖHLER**; G. WALDER** and T.D. MARK**

ABSTRACT. - Investigated area is located in the Gördes submassif of the Menderes massif, the metamorphic basement consists of the following lithologics in ascending order: Sillimanite-gamet gneiss. sillimanite-gamel-kyanitc schist. sillimanite-staurolite-gamet-kyanite schist, staurolite-garnet schist and garnet mica schist. Kyanite-andalusite pegmatoids which occur within the kyanite-bearing schists were formed in the course of the last major metamorphism giving the final stage to the Menderes massif. The metamorphic basement is overlain by the allochthonous units which are relicts of the Lycian nappes which caused the last major metamorphism during the Eocene-Oligocene time in Menderes massif. The age of the apatite crystals obtained from the pegmatoids arc determined by the fission-track method. The cooling age of the apatite crystals ranging from the Early Oligocene to Early Miocene are in good agreement with the field observations in the study area and geological evidences relating to the Menderes massif.

INTRODUCTION

Menderes massif is a large area of metamorphic rocks lying in the western part of Turkey. The study area is approximately 200 km northeast of İzmir city (Fig. 1). Although investigations on the Menderes massif have continued since the pinoeering work of Phillipson (1911), the polyphase history and timing the metamorphic events within the basement remain unclear. This paper provides new petrological and age date to the PT(t) pathways of the schists.

In particular, we report fission-track age determinations on apatites from the pegmatoids within the kyanite-schists from the northern part of the Menderes massif. We have also attempted to interpret previous geologic and radiometric studies of the Menderes massif in the light of our investigations.

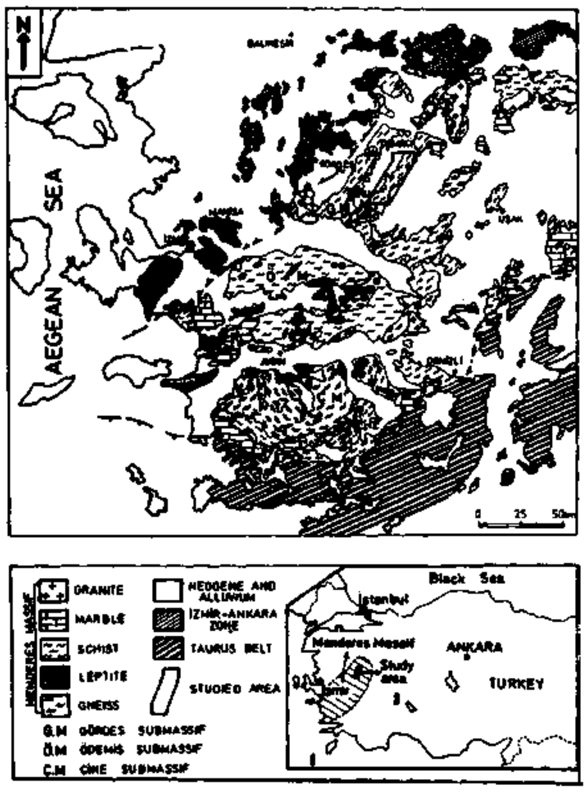

Fig. 1 - Location map of the study area (Candan,1988).

GEOLOGICAL SETTING OF THE REGION

Although considerable areas of the Menderes massif have now been mapped in detail and studied by several workers, problems still to be solved include the depositional ages of the core and cover rock units of the massif, the metamorphic phases affecting these rock units, and the ages of the metamorphism. There are some different opinions on these subjects.

Menteşe marbles, which contain emery deposits, occur at the. southern boundary of the Menderes massif and the Göktepe formations which-overly these marbles, were deposited in Permian time as suggested by Önay (1949), Kaaden and Metz (1954) and Schuiling (1962). The same authors suggested that these formations were metamorphosed during the Hercynian orogeny. Akdeniz et al. (1980) and Akdeniz and Konak (1979) argues that the protoliths for the metamorphic sequence are Precambrian-Paleozoic in age, and they are overlain by unmetamorphosed Triassic-Upper Cretaceous rock units with an angular disconformity. They also suggested that the metamorphism took place during the Hercynian orogeny.

On the other hand, Brinkmann (1966, 1967) suggested that the core of the Menderes massif was metamorphosed in Late Precambrian-Early Cambrian time. The same author also argued that the cover rock units of the sequence in the massif continued until Early Jurassic times, and were metamorphosed during Middle Jurassic times. According to Wippem (1964), the cover rock units and the overlying Göktepe formation are Devonian and Permo-Carboniferous in age respectively, whereas the emery bearing marbles are probably Triassic in age. Wippem (1964) also noted that the age of the metamorphism is Jurassic. An another opinion widely accepted presently on the major metamorphism of the Menderes massif will be treated in detail in the following sections.

LITHOSTRATIGRAPHY

On the basis of field observations and petrographic studies, the rocks in the study area can be divided into tree main groups: Metamorphic units of the Menderes massif, allochthonous units overlying the metamorphic basement with tectonic contacts, and Neogene volcanic and sedimentary rocks resting on both metamorphic and allochthonous units with an angular unconformity (Fig. 2).

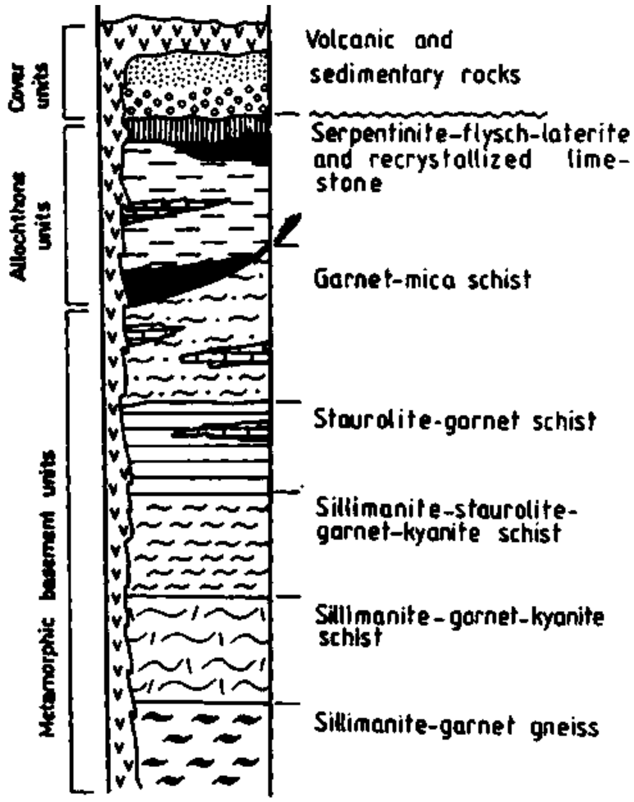

Fig. 2 - Generalized columnar section of the study area (Candan, 1988).

In the region, the metamorphic sequence consists of the following successions in ascending order Sillimanite-gamet gneiss; sillimanite-garnet-kyanite schist; sillimanite-staurolite-gamet-kyanite schist with abundant kyanite-andalusite pegmatoids; staurolite-garnet schist and garnet-mica schist with marble intercalations containing emery deposits. The units and the main mineral assemblages observed in these rocks are shown on Figure 3.

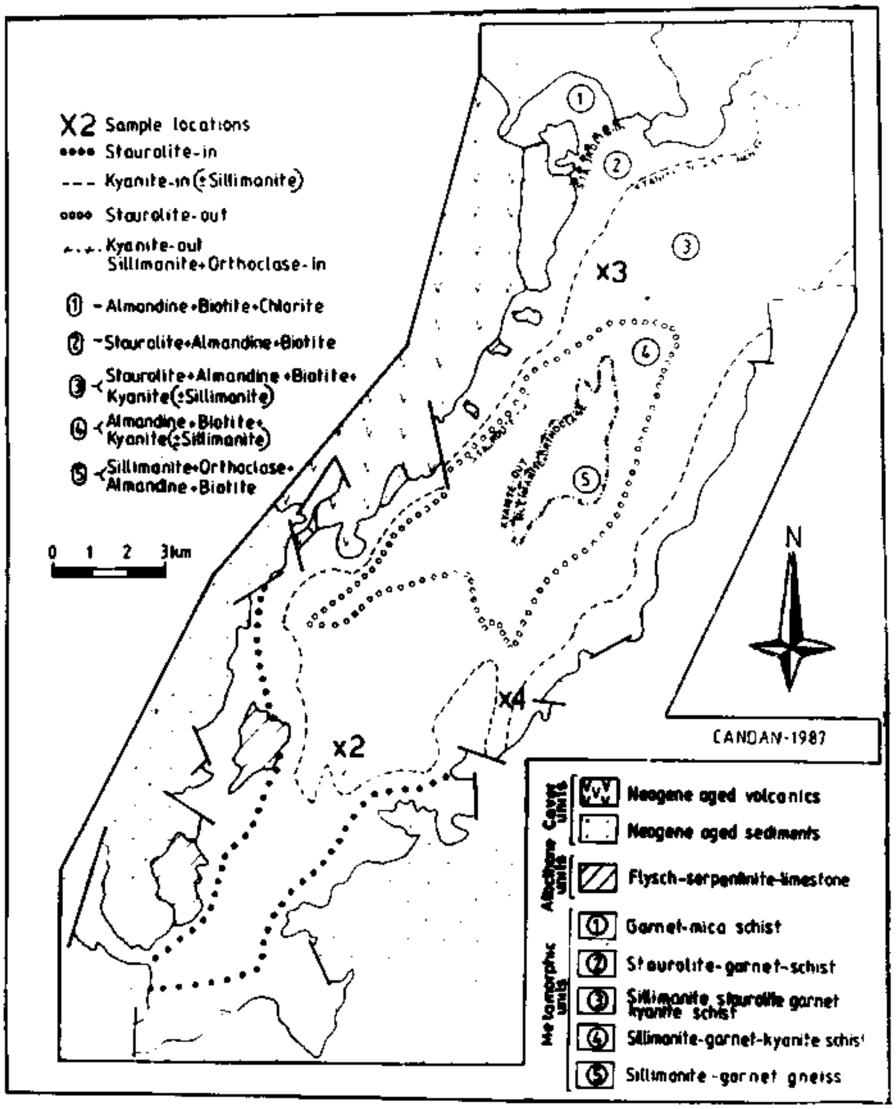

Fig. 3 - Main units and stable mineral assemblages observed within these rocks, and the isograd map of the study area (simplyfied from Candan, 1988).

The allochthonous units were emplaced over the metamorphic basement units along nearly horizontal thrust planes and occur as klippe on the metamorphic rocks. The allochthonous units consist mainly of limestone, conglomerate, sandstone, shale and ophiolitic rocks. Large recrystallized muscovite and chlorite crystals are observed within deformed pelites. These new minerals were formed in the course of the emplacement of the allochthonous units. Other rock types identified within the allochthonous units are harzburgite, which *is* often totally altered to serpentinite, recrystallized limestone and pink-coloured thin-bedded limestone. Lateritic soils are commonly developed above mentioned ultrabasic rocks. According to Dürr (1975), Dürr et al. (1978) and Şengör et al. (1984), these allochthonous units are relicts of the Lycian nappes which were transformed front the north to the south on the Menderes massif during the Eocene-Oligocene time.

Middle Miocene rock units consisting of volcanic and volcano-sedimentary rocks rest on both metamorphic and allochthonous units with angular unconformity.

PETROGRAPHY OF THE METAMORPHICS

Sillimanite-garnet gneiss

These rocks are the lowest rock units observed in the sequence and are composed of the following mineral assemblages: quartz+plagioclase+orthoclase+biotite+muscovite±sillimanite+garnet±chlorite+apatite+tourmaline and zircon. These rocks show mortar texture and contain fibrolitic sillimanites forming at the feldspar boundaries.

Kyanite-bearing schists

These rocks are widely distributed and can be divided into two subgroups according to their mineral composition as either sillimanite-garnet-kyanite schist or sillimanite-staurolite-garnet-kyanite schist. These schists have kyanite, staurolite and garnet porphyroblasts reaching up to 7-8 cm in size (Fig. 4).

Fig. 4 – Staurolite and kyanite crystals reaching up to 7-8 cm in size observed in the sillimanite-staurolite-garnet-kyanite schists. Stau-staurolite; ky-kyanite.

The mineral assemblage of the sillimanite - garnet - kyanite schists is quartz + plagioclase (An 27 - 30) + kyanite ±sillimanite+garnet+biotite+muscovite±chlorite+apatite+zircon+tourmaline and sphene. Sillimanites are grouped in two different types according to their formation. The most abundant type is fibrolitic sillimanite growing between two feldspar boundaries, and the other is the polymorphic transformations from kyanite as a result of temperature increase during progressive metamorphism (Fig. 5). Garnets are represented by almandine ($Alm_{77}Prp_{19}Grs_4$), and all the chlorites in these rocks were formed after biotite by the last retrograde metamorphism.

The second group of the kyanite-bearing schists consist of quartz+plagioclase (An 23-25) ± sillimanite + kyanite ± andalusite + staurolite + garnet + biotite + muscovite ± chlorite+apatite + zircon + tourmaline and sphene. These rocks also contain staurolite and andalusite in contrast to the other kyanite-bearing schists. The staurolite prophyroblasts are extremely fractured and altered to chlorite and sericite.

Although all the Al_2SiO_5 polymorps (kyanite-andalusite-sillimanite) exist in these schists, the petrographical observations and textural evidences suggested that these polymorps are not in equilibrium. The polymorphic transformations from kyanite to sillimanite and andalusite due to increasing temperature and decreasing pressure respectively, are very common.

Fig. 5 - Polymorphic transformation from kyanite to sillimanite as a results of increasing temperature during the progressive metamorphism. Plane polarized light, 10 x. Sil-sillimanite; Ky-kyanite.

Kyanite-andalusite pegmatoids

Kyanite-andalusite pegmatoids reaching up to 10-12 m in size occur within the kyanite-bearing schists. These pegmatoids are often lens shaped and parallel to the schistosity (Fig. 6). The pegmatoids do not show internal zoning, and are enclosed by a biotite-rich envelope.

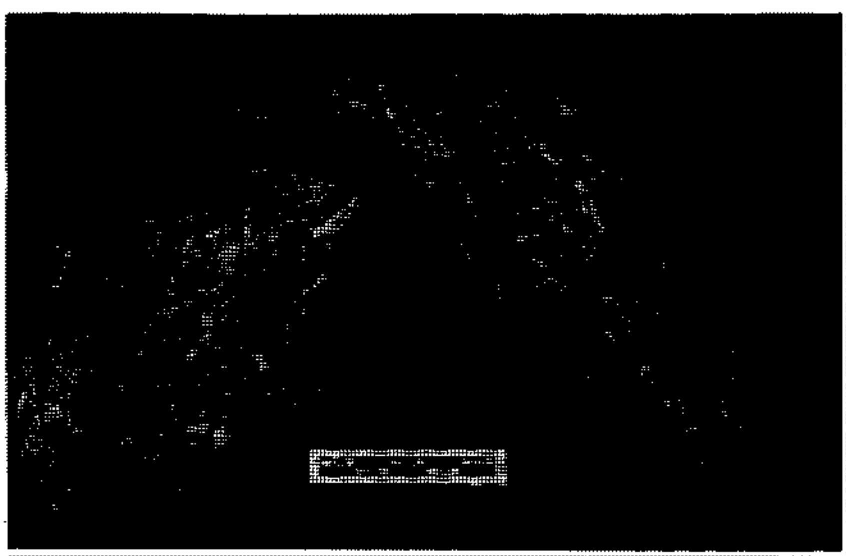

Fig. 6 - General appearance of kyanite-andalusite pegmatoids.

The general mineral assemblage of the pegmatoids is composed of quartz+plagioclase (An 6) + kyanite ± sillimanite + andalusite ± staurolite ± garnet + biotite + muscovite ± chlorite + apatite + zircon ± graphite + rutile and diaspore. The most abundant mineral of the pegmatoids is kyanite which occurs often in white, pale-blue, dark-blue and green colours and is often deformed.

The second most abundant mineral is dark pink andalusite altered widely to sericite. These minerals are formed after kyanite by polymorphic transformation (Fig. 7). Trace amounts of sillimanite are also formed after kyanite due to increasing temperature.

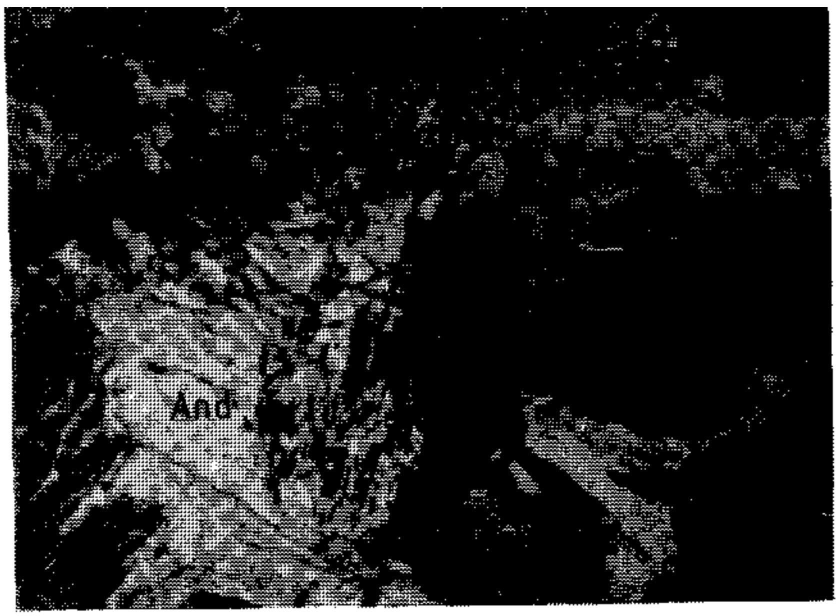

Fig. 7 - Polymorphic transformation from kyanite to andalusite in kyanite-andalusite, pegmatoids. Cross polar, 4x. and-andalusite; ky-kyanite.

Idimorphic Fe-staurolite and garnet ($Alm_{76}Prp_2Grs_{22}$) are also important phases in the pegmatoids and occur with biotite, muscovite and chlorite (ripidolite and pycnochlorite) after biotite. The other important mineral of pegmatoids is pale green apatite with crystals often reaching 0.5 to 1 cm and rarely 7 to 8 cm in size (Fig. 8). Optic and difractometric observa-

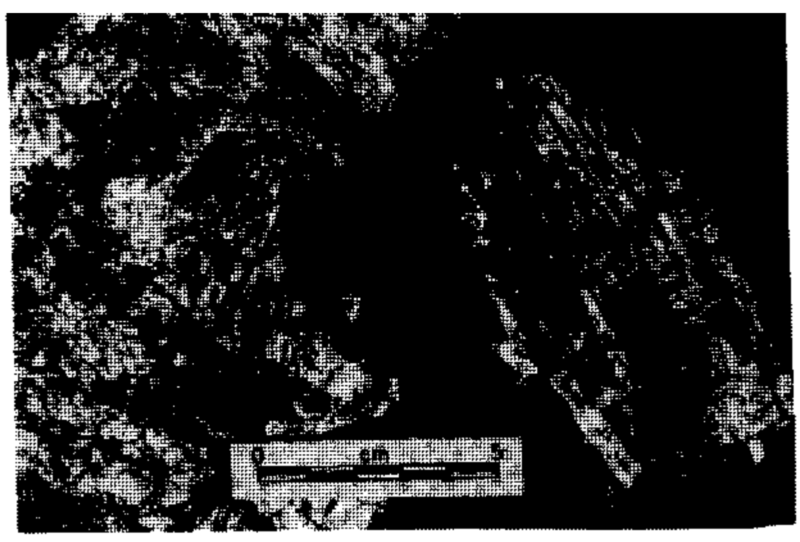

Fig. 8 - Euhedral apatite crystals intergrowing with kyanites in kyanite-andalusite pegmatoids.Ap-apatite.

tions indicate that these apatites are fluor-apatite which shows intergrowth with kyanite, and extend parallel to the mineral lineation of the wallrocks. These observations suggest that the crystallization of the apatites are synmetamorphic. For this reason the apatites are used to determine the age of the last main metamorphic phase affecting the Menderes massif.

Kyanite-andalusite pegmatoids were formed by the lateral migration of some elements, such as Al, Si, K, etc. from the wallrock of the Al-rich kyanite schists during the high-grade regional metamorphism affecting the region. The occurence of the kyanite-andalusite pegmatoids only within the kyanite-bearing schists indicates that element migration took place for only short distences (Candan ve Dora, 1984; Candan, 1988). The large and pure kyanite crystalls reaching up to 50-60 weight percentage in some pegmatoids are of economic importance. These rocks widely exploiding are utilized as the raw material of the refractory industries.

Staurolite-garnet schist

Sillimanite-staurolite-garnet kyanite schists are overlain by the staurolite-garnet schists. These schists consist of mainly quartz + plagioclase + biotite + muscovite + staurolite + garnet + apatite + zircon, and they often alternate with muscovite-quartz schist and marbles containing emery deposits.

Garnet-mica schist

These rocks alternating with marble and muscovite-quartz schist in the region form the highest level of the metamorphic sequence in the region, and contain mainly quartz + plagioclase + biotite + muscovite + chlorite + garnet and zircon.

EXPERIMENTAL RESULTS

The basic theory and the experimental technique of fission- track dating has been reviewed up to 1975 by Fleischer et al. (1975). Certain points pertinent to the present determination are summarized in the following.

As fission-tracks in minerals become partly annealed under the influence of increased temperature, it is necessary to correct the uncorrected fission-track age formula. One method of correction is based on the observation that as a mineral sample is heated, the fission-track lengths decrease at a rate related to the track density decrease. The corrected fission-track age for apatite is given according to Mark et al. (1973) by

$$t_{1_{min}} = \frac{1}{\lambda_\alpha} \ln\left(1 + \frac{\lambda_\alpha \cdot \sigma_f \cdot I \cdot n}{\lambda_f} \cdot \frac{P_s}{P_i} \cdot K\right) \qquad (1)$$

or

$$t_{1_{min}} = \frac{\sigma_f \cdot I \cdot n}{\lambda_f} \cdot \frac{P_s}{P_i} \cdot K \qquad (2)$$

$$(\text{for } t < 3 \times 10^8 a)$$

The symbols have the following meanings: λ_α 1.54×10^{-10} a^{-1} time constant for the α-decay of ^{238}U(Spadavecchia and Hahn, 1967), λ_f 0.84×10^{-16} a^{-1} time constant for spontaneous fission of ^{238}U (Spadavecchia and Hahn, 1967), σ_f 582 barn induced fission cross section of ^{235}U by thermal neutrons (Hanna et al. 1968), I: 7.26×10^{-3} isotope abundance ratio $^{235}U/^{238}U$ (De Wet and Turkstra, 1968), n integrated neutron flux. P_s and P_i measured area densities of spontaneous and induced fission-tracks, K correction factor with K:l_i/l_s in apatite (Mark et al., 1973), l_i / l_s ration of average length of induced and spontaneous fission-tracks.

Mark et al. (1973) have shown that this corrected age t l_{min} is the amount of the necessary to reduce the length of fission (produced in full length) by geothermal annealing to l_{min} is the minimum track length-horizontal projection-which can be observed in the microscope; in apatite 2 mm). In other words, t l_{min} is the age of tracks with the shortest length that can be detected today. If the track length ratio K is omitted in equ. (1) or (2), i.e. if K: l_i / l_s is set equal to 1, a value of t is obtained called uncorrected fission-track age.

Following Mark et al. (1973, 1981 *a,b)* and Mark and Mark (1982) it is possible to determine by means of a theoretical calculation a temperature value, Tl_{min} corresponding to the measured and corrected fission-track age, t l_{min} assuming a linear or exponential cooling rate. This calculated relationship is given Mark et al. (1981 *a,b)* and Mark and Mark (1982) and can be used to obtain a complete temperature-age determination (see also the discussion by Bertagnolli et al., 1981).

The preparation of the present apatite samples was done in the usual way (Koark et al., 1978). The track densities were counted with a texture analyzing system of Leitz combined with a phasecontrast Leitz Orthoplan microscope. In order to evaluate the reliability of the present determination (measurement of neutron flux etc.) a Durango apatite sample was measured simultaneously with the present apatite samples, yielding a value in excellent agreement with previous high precision determinations (Mark et al., 1981b).

Table 1 gives the present results of the counted track densities (the induced tracks were obtained by irradiating the annealed samples with a thermal neutron dose n of 9.36 x 10^{14} n/cm^2) and track length ratios. Also shown in Table 1 are the determined corrected fission-track ages and the corresponding temperatures. It is interesting to note that these age determinations were confirmed by an additional independent measurement in our laboratory, yielding the values of 21, 27 and 35 my, respectively. We have also determined the uranium concentration (using the fission-track method described my Mark et al., 1974), yielding uranium concentrations of 22.5, 41.5 and 26 ppm for the samples A2, A3 and A 4, respectively.

Table 1 - Experimental fission-track results for apatite crystals from the Menderes massif

Sample number	*Uranium contentration in ppm*	*Absolute number of counted tracks spontaneous*	*induced*	*Fission-track areal densities in $10^5/cm^2$ spontaneous*	*induced*	P_s/P_i	$\bar{l}_i/\bar{l}_s$	$t_{l_{min}}$ *in 10^6a*	$T_{l_{min}}$ *in °C*
A2	22.5 ± 5	410	1214	2.1 ± 0.1	4.7 ± 0.15	0.45 ± 0.04	1.09 ± 0.5	23 ± 5	165
A3	41.5 ± 10	1051	1544	5.4 ± 0.2	7.9 ± 0.2	0.68 ± 0.05	1.16 ± 0.08	37 ± 7	163
A4	26 ± 5	678	1380	3.5 ± 0.15	5.3 ± 0.15	0.66 ± 0.06	1.18 ± 0.07	36 ± 7	163

CONCLUSIONS

The sedimentation in the Menderes massif continued until the beginning of Tertiary is now widely accepted, although previous workers had suggested Jurassic time for the upper limit of the sequence of the Menderes massif. According to Çağlayan et al. (1980), the sedimentation reaches to Upper Paleocene time, and as result of paleontological evidence, the emery-bearing marbles range from Triassic to Upper Cretaceous in age. The red marbles overlying the emery-bearing marbles are Upper Paleocene in age, whereas the same red coloured marbles are Eocene in age according to Gutnic et al. (1979). Dürr et al. (1978) suggested that the sedimentation in the Menderes massif continued until the Lower Tertiary time, and the main Harrovian metamorphism reached up to migmatization grade and took place during Eocene times.

Andriessen et al. (1979) subdivides three major metamorphic phases according to their radiometric studies by using K-Ar method on the emery-bearing series of the Menderes massif in Naxos island. The medium pressure/high temperature type metamorphism reaching up to anatexitic stage gives an age of 25 ± 5 my (Oligocene - Early Miocene). Şengör et al. (1984) and Akkök et al. (1984) also suggested that the age of the sequence in the Menderes massif reaches up to Eocene, and the main metamorphic phase is in between Eocene-Oligocene. This result coincidences with the radiometric age of 35 ± 5 my. This main metamorphism of the Menderes massif was a product of the latest Paleocene collision across Neo-Tethys and the consequent internal imbrication of the Menderes-Taurus blok that resulted in the burial of the Menderes massif area beneath the Lycian nappe complex (Şengör et al., 1984).

In summary, there are three main opinions regarding the age of the major metamorphism which affected the Menderes massif: Hercynian (Önay, 1949; Kaaden and Metz, 1954; Schuiling, 1962; Akdeniz and Konak, 1979), Middle Jurassic (Brinkmann, 1966, 1967; Wippern, 1964), and Eoccnc-Oligocene (Andriessen et al., 1979; Şengör et al., 1984; Akkök et al., 1984) in age as mentioned in the previous works.

As far as metamorphic paragenesis in concerned, the study area located in the northern part of the Menderes massif shows similarity to the Ödemiş submassif. In addition to this similarity, the presence of the emery bearing marbles interlayercrcd with the schists in the region can be clearly correlated with the scries of the southern limb of the Menderes massif.

In the petrography section, it is already mentioned that the kyanite-andalusitc pegmatoids were formed by the lateral migration of some elements from the wallrocks of the kyanite schists during the high-grade regional metamorphism that affected the region. Thus, the kyanite-andalusite pegmatoids were formed penecontemporaneously with the kyanite schists; which resulted from the major metamorphism giving the present appearance of the massif.

Apatite crystals intcrgrowing with kyanites can give the formation age of the schists. The experimental fission-track results for apatite crystals are shown in Table 1. As it is understood from these data that these cooling ages range between Early Oligocene and Early Miocene. These results are in good agreement with the geological evidence in Turkey and the radiometric age determination obtained from the Greek islands (Andriessen et al., 1979).

In the study area, the allochthone units consisting of serpantinite, flysch and limestones, rest on the metamorphic basement with tectonic contacts. These allochthone units are relicts of the Lycian nappes which caused the last major metamorphism during the Eocene-Oligocene time. These geological interpretations support the radiometric cooling ages obtained from the apatite crystals. Şengör et al. (1984) suggest that the final major metamorphism of the Menderes massif is completed during Middle Eocene-Early Oligocene, and the uplifting and cooling of the massif took place between Late Oligocene and Early Miocene with respect to the geotectonic evolution of the Western Anatolia. The obtained cooling ages ranging between Early Oligocene and Early Miocene of the apatite crystals from the study area are in good agreement with above mentioned concept of the cooling and uplifting stages which took place after the last major metamorphism of the massif.

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PALYNOLOGY OF THE BORATE BEARING NEOGENE SEDIMENTS IN BİGADİÇ, KESTELEK, EMET AND KIRKAREGIONS

Erol AKYOL ** and Funda AKGÜN **

ABSTRACT. - The Neogene graben systems of western Anatolia are filled by clastic, carbonate and volcanic material. In some places these continental deposits contain economically potential coal, bituminous shale, uranium, clay and berates. Borate beds in Bigadiç, Kestelek, Emet and Kırka Neogene basins are accompained by thin layers and lenses of coal and coaly shale which are distinguished by a rich microflora. The microflora of these basins consists of two pollen assemblages "a lower and an upper pollen assemblages". These assemblages enable us to correlate the sedimentary sequences of the basins and the understand the palcoclimatic and paleogeographic conditions prevailed during the Neogene. The lower pollen assemblage which is Early Serravallian in age are recognized in Çan, Orhaneli, Soma, Selendi, Şahinali. It is indicative of widespread forests during this time. Mammalian fauna studied in Tire and Sarıçay is also conclusive of forest environment. The sedimentary sequences which contain coal-bearing sand, clay and calcareous shale were deposited in lakes surrounded by mountains with a dense vegetation. The upper pollen assemblage is Late Tortonian in age. Although it broadly represents a moderate humid climate, a relatively dry and coal environment can be suggested when compared with that of the Early Serravallian time. Widely distributed Late Tortonian mammalian fauna, which is well known in western Anatolia, indicates wide steppes. However, as it is the case in four basins with upper pollen assemblage, the Late Tortonian steppes were studded by savannah parks.

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FINDINGS RELATED TO THE HISTORY OF MINING IN TURKEY

Ergun KAPTAN*

ABSTRACT . - The greatest innovation in the history of civilization is the use of minerals. Anatolia, owes its rich and long history, to the varied non-depleeting mineral wealth. Nine thousand years ago copper was the first metal the habitants of Anatolia made use of. The oldest underground mining in Anatolia has been discovered around Tokat-Erbaa, Kozlu which belongs to the beginning of 5 thousand years B.C. In this study, for the first time in Turkey, using archaeometric methods, in two locality around Niğde tin mineralizations have been discovered. This is an important discovery for the mining history of Turkey.

INTRODUCTION

Ancient man has worked metals and minerals distributed in the earths crust for thousands of years. The technological level achieved in todays world began with experimentation leading to a growing knowledge of metals and how they were worked. For this reason the greatest breakthrough in the history of civilization is the discovery of metals. Man began to take greater advantage of the metal resources around him through experience. Metals triggered the birth and growth of new cultures based on increasing needs. The evolution of these ancient metal industries has led to the high levels of todays civilizations.

From the Paleolithic period to the modern era, Anatolia has its place in world history as the home to inestimable cultural treasures. Anatolia's rich and well-rooted cultures and their variegated expressions are due in part to its many metal resources. Other ancient Near Eastern cultures exploited these metal resources throughout the millennia in part because of the Eurasian nature of Anatolia, that is, as being both a part of Asia as well as Europe. For this reason, Anatolia served as a bridge between the eastern and western civilizations for thousands of years. Trade between these areas rich with metal resources in Anatolia and ancient Near Eastern civilizations was thus bom four thousand years ago.

A GENERAL PERSPECTIVE OF ANATOLIAN METALLURGY

This chronological breakdown of the history of Anatolian metallurgy is based on years of research:

Chalcolithic period	**5000 - 3000 B.C.**
Early Bronze Age	**3000 - 2000 B.C.**
Middle Bronze Age	**2000 - 1550 B.C.**
Late Bronze Age	**1550 - 1200 B.C.**
Iron Age	**1200 - 550 B.C.**

The oldest worked metal made of metallic ore was found at the excavations of Çayönü tepesi mound near Diyarbakir-Ergani (Çambel and Braidwood 1970). These materials were made of native copper and malachite and were worked cold and shaped. These materials are the first finds proving ancient man in Anatolia had knowledge about and worked ores 9000 years ago.

The first finds signifying metallurgical work was found at the excavations of Çatalhöyük near Konya-Cumra (Mellaart, 1966). It is interesting and exciting to find such precocious evidence at Çatalhöyük in 6000 B.C. On the basis of these findings, this date should be noted as the beginning of Anatolian metallurgy.

Copper was the first metallic ore known to ancient man in Anatolia. However, aside from the knowledge that copper had better malleability than rocks, early experimentation never really reached high levels in technology. The Chalcolithic period (5000-3000 B.C.), which means the stone-metal era, is accepted as the transition and developmental phase between the Neolithic and the Early Bronze Age. Copper implements in this period are less evident than stone tools. Copper was a precious commodity in Anatolia in 5000 B.C. For this reason it was used only for important tools and weapons such as arrowheads, axes, adzes, and awls. Finds at Mersin-Yümüktepe indicate that casting was known during this period. In addition, objects that indicate tin was used as an alloying material, was found at Yümüktepe dating to 4300 B.C. (Esin, 1969). The tin content measures 2,9 % in these materials, however, this amount is very low for a true tin-bronze. If the ancient metallurgists knowingly, perhaps experimentally, were able to add 2,9 % tin to copper, then these can be considered the oldest, but most primitive bronzes in Anatolia.

The Bronze Age (3000-1200 B.C.) is a period when bronze has gained precedence over stone. In the third millennium B.C. Anatolian metallurgists and craftsmen were capable of casting metal and making alloys such as bronze and electrum (Koşay and Akok, 1966). In the Early Bronze Age (3000-2000 B.C.) metallurgists and craftsmen produced metals which indicate high levels achieved in casting. These are high quality examples of copper, lead, gold, silver, bronze (coppertin), electrum (gold-silver) objects. The best examples of these have been found in Alacahöyük (Çorum), Troy (Çanakkale), Horoztepe (Tokat-Erbaa), Eskiyapar (Çorum), Mahmatlar (Amasya), Karataş (Elmalı-Semayük). For example, at Alacahöyük, important finds include gold, silver electrum, bronze and copper objects. Analyses of these objects obtained at MTA Institute laboratories have yielded tin levels from 9-17 %. Since no other impurities were found in these metals, they are good examples of high quality tin bronze alloying (Koşay, 1938). In addition, latest analyses of Alacahöyük bronze objects, conducted at the Darmstad Technical University in Germany, indicated that tin content was 10 % (Koşay and Akok, 1966). The worlds oldest examples of the sword, which is the most efficient weapon known to man, was again found at Alacahöyük. All of these finds indicate, that by the third millennium B.C., Alacahöyük had achieved a high level of metallurgical skills and technology of manufacturing metal.

During the Early Bronze Age iron was a comparatively less known metal. It is assumed that the experimental techniques tried on other metals were also attempted with iron, with little success. This is due to the high temperatures needed for iron metallurgy. The product of this type of smelt, would have required new processing techniques, and for this reason, iron objects in this period are quite rare. The best example is a gold handled iron blade 18.5 cm. long from the second half of the third millennium B.C. found at Alacahöyük. MTA Institute analyses on group three materials from the 1937 season of this site yielded iron with 4-5 % nickel content (Koşay, 1938). It is highly possible that this was produced from meteoric iron. In the Early Bronze Age (3000-1200 B.C.) iron, which was considered the most precious metal, was probably manufactured out of meteoric iron. Research (Bjorkman, 1973) has indicated that meteoric iron was used in Anatolia and the Near East.

Research on tin sources is continuing to be an important topic in Turkey. It has always been recognized that tin was a crucial alloying metal for bronze, but its source was as-yet unknown. The source of the tin used first in the Early Bronze Age in Anatolia was not known. However, in the second millennium B.C., tin was imported as tubes into Anatolia as 300 kilogram weights from Mesopotamia (Bilgi9, 1943). During this period Anatolian cities had knowledge of its resources and had a well developed metal industry. Central and north central Anatolia especially are the earliest regions to develop underground mining and high levels of metallurgy (Kaplan, 1983). As a result, a metal trade belween Assur in Mesopolamia and Analolia conlinued for 200 years. Donkey caravans slarling from Mesopolamia would carry tin to the industrial centers of Anatolia, which had a high demand on this product. In relum they would oblain copper, argenliferous lead, silver and gold (Bilgiç, 1948).

Research into the sources of tin have continued in the present day. The first investigations were in 1899 with the permission of the Ottoman Imperial government (MTA report, 1900). After the birth of the Republic, tin was searched for beiwecn 1932-1939. Tin mineralizalion was first found in the Bursa-Keleş, Soğukpınar area (Çağatay el al, 1981). Ancient mines were also found in this area. In 1985 a tin-rich Pb-Zn ore was discovered al Bolkardağ-Sulucadere (Niğde-Ulukışla) (Yener and Özbal, 1986, 1987). This same investigation was exiended in 1987 to the Niğde-Çamardı area. An ancieni mining complex of galleries, which was mosl probably exploiting tin, was discovered in the Kestel-Sarıtuzla area of Celaller village (Kaplan, 1988; Yener, 1988; Yener el al., 1989).

Most of the metal objects made during the Old Hittite period (1750-1450 B.C.) were mostly copper and bronze. In addition to gold and silver objects, some rare iron objects were also found. The rarity of iron was due to the lack of melallurgical knowledge of iron working on the part of the Hittites. During the Hittite Empire period (1450-1200 B.C.), Anatolian melallurgy developed, became widespread, and took the characleristics of an industry. In spile of the fact that Anatolia had entered the Iron Age circa 1200 B.C., iron was less common than other metals.

The Urartians were highly successful in melallurgy in eastern Analolia. The success displayed by the metallurgical indusiry during the Urartian kingdom (900-600 B.C.) even influenced the Etruscans in northern Italy. During this period, malerials of bronze were exported from Anatolia to Greece and Italy (Akurgal, 1959). Anatolian melallurgy achieved great heights during the Phrygian period (700-550 B.C.). However, iron was less common than other metals. Sardis (Manisa), the capital of the Lydian kingdom (700-550 B.C.), is an imporlant industrial and metallurgical center located in westem Analolia. The splendor of the Lydian kingdom during this period was due to the exploitation of placer deposits of gold in the Pactolus River (Sart Çayı), which flows nearby (Şükun, 1943). The Lydian people called the Pactolus the "river that flowed with golden waves".

Anatolia, which possessed these rich mineral sources, later came under the hegemony of the Achacmenid Persians, the Hellenistic, Roman and Byzantine Empires. The Great Selçuk and Otoman Empires continued working these metal sources, expanded upon the long accumulation of metallurgical skills and added new techniques, as fitting an old Anatolian tradition.

ANCIENT MINING SITES

The MTA Institute project on the "History of Metallurgy in Turkey" has yielded important and interesting results. New finds of ancient mining have especially placed Turkey into an important place in the history of metallurgy.

The oldest mining remains in Turkey were found in the Tokat-Erbaa (Kozlu Eski Gümüşlük) and were substantiated by soundings (Giles and Kuijpers, 1974). This mine is 50 meters deep. Excavations were conducted in the mine and in the tailings, which represent thousands of years of exploitation, and reached bedrock. Wood from an ancient gallery was found at 8.20 meters deep on bedrock (Kaptan, 1986). Radiocarbon dates obtained from the Physics Department of the Middle Eastern Technical University in Ankara gave a date of 3789±109(4650±109 B.C. calibrated date). This ancient mine was the source of copper for the region for thousands of years through a multitude of periods. The bronzes of Horoztepe (Erbaa), dated to the Early Bronze Age, display superior casting workmanship. Perhaps the source of the copper was this mine at Erbaa-Kozlu Eski Gümüşlük (Fig. 1).

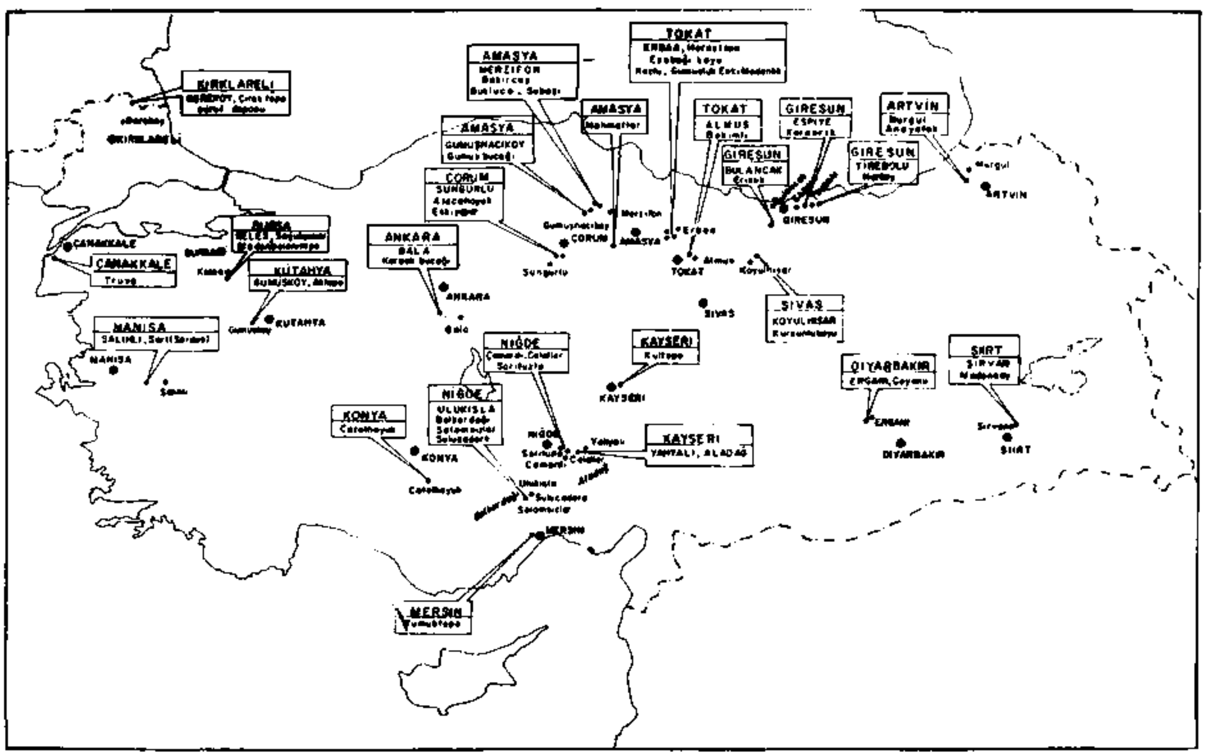

Fig. 1

The second oldest mine (Kaptan, 1988; Yener, 1988) was found in 1987 at the mining complex at the Niğde-Çamardı Celaller village al Kestel-Sarıtuzla mine (Fig. 1). The first workings were dated to the Early Bronze Age by pottery and radiocarbon dates. Four charcoal samples gave radiocarbon determinations from 40201880 to 3830±65 years B.P., calibrated to 2874-2133 B.C. (Yener et al., 1989). The mine continued in use into the Byzantine period. Research continues at this site.

The third oldesl mine was found in 1979 at Kütahya-Gümüşköy (Aktepe Mevkii). Torches were used for illumination by the ancient miners. Radiocarbon dales were obtained from burnt torch specimens at Aaechen University in Germany. The

C14 dates were 1950±85 B.C. (Kaptan; 1984). The calibrated dales are 2425±85. This mine is important not only for Anatolian metal history but for other areas as well.

The oldest remains as yet from the copper mines at Artvin-Murgul date to the second half of the fust millennium B.C. This dale was established with the help of an ancient mine found during open-pit mining in 1967 (Kaptan, 1977). This mine was destroyed due to the on-going mining operations. Another date abte ancient mine gallery was found at Sivas-Koyulhisar near Kusunlu village. The mine dates to the beginning of the first millennium B.C. (Kış and Güler, 1984). The date of the Siirt-Şirvan ancient mine was established by the wooden support beams in the mine during modem soundings at 66 meters deep. Radiocarbon dates yielded 590±45 A.D. (Kış and Güler. 1984). However. this does not represent the earliest and latest stages of the mine.

An ancient mine gallery found at Niğde-Ulukışla Madenköy Bulkardağ Selamsızlar area al 2000 meters altitude was 190-200 meters long and dated to the 8 th century A.D. (Kaptan. 1988). Another mine a Bolkardağ Sulucadere was dated to the 9th century A.D. (Fig. 1).

Another mine was accidently exposed by Etibank at Giresun-Espiye Karaerik area. This mine was 100 meters long, is dated to the 11th century A.D. and yields chalcopyrite (Kaptan, 1980). Again at Giresun-Bulancak, Eriklik village area a mine gallery was found that was 17 meters long. This mine was dated to the 11 th century A.D. and was worked for copper (Kaplan, 1980). While soundings were being conducted by the MTA General Directorate to search for increased reserves of Cu-Pb-Zn. an ancient mine was discovered in 1980 al Giresun-Tirebolu Harköy. The mine is 28 meters deep. Radiocarbon dales from the gallery timbers yielded 1550±42 A.D. (Kış and Güler, 1984).

While investigating reserves at Kayseri-Yahyalı. Aladağ Pb-Zn veins, another ancient mine was discoveicd al 2645 meters altitude. Radiocarbon dates from the timbers gave 1050+73 A.D. (Kış, 1985). In addition to these examples, mines dating to the Roman-Byzanline and Otoman periods were found in a variety of areas, especially in eastern Black Sea regions.

ANCIENT MINING TOOLS AND EQUIPMENT

Ancient tools and equipment relating to underground mining aclivities arc rare in Anatolia. The first finds stem from the Murgul-Anayatak area. A mining shovel, carved from a tree trunk, was found in the galleries (Fig. 2). Radiocarbon dales based on 5730 half-life yielded 316±170 B.C. (Kaptan, 1977).

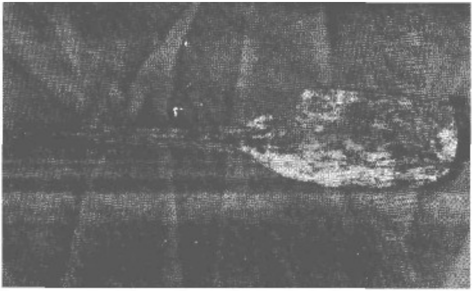

Fig. 2

Among the oldest mining finds was a wooden ladder which was from a gallery near Koyukhisar-Kurşunlu village. Radiocarbon dales yielded 948±56 B.C. (Kış and Güler, 1984). A second ladder was also found in the Kurşunlu village area and dated to 350±5G A.D. This example was maxle with more care than the ladders found in this area dating 10 earlier periods (Fig. 3). It was made by carving out steps on one face of the tree trunk and is unique in Anatolia.

Fig. 3

A stationary ladder was found in the raise between lower levels in the Selamsızlar mine in Ulukışla-Madenköy at Bolkardağ (Fig. 1). Three separate ladders measuring 130-150 cm were found in the three raises. Three steps were carved on one face of one of the ladders which was made of black pine (Pinus nigra Arnold). The ladder was dated by C14 to 777±55 A.D. (Kaptan. 1988). Again at Bolkardağ Sulucadere, another ancient mine yielded a black pine shovel found 31 meters deep inside the gallery. Radiocarbon dates yielded 836±70 A.D. (Kış. 1985; Kaptan 1988).

C14 analyses yielded dates of 1161+74 A.D. (Kaptan, 1980). which calibrates to 1070+.74 A.D. An ore carrier, which is 17 cm. long was found in the galleries al Giresun-Bulancak, Eriklik village area (Fig. 1). Made from the trunk of a chestnut tree, radiocarbon dates it to 992±75 A.D. (Kaptan, 1980). which calibrates to 1050+75 A.D. It can be considered a prototype of mine wagons. This ore carrier is characteristic to Anatolia and is now unique in the world. (Fig. 4).

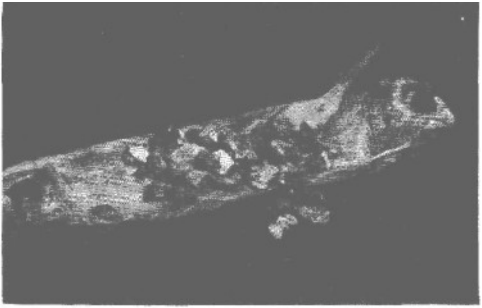

Fig. 4

ANCIENT ORE BENEFICIATION TOOLS

Tools used to process ores prior to smelling have been found, researched, and preserved in this country. The first time an ore beneficiation tool was identified was one found al the entrance of an ancient mine at Kütahya-Gümüşköy Aktepe mining area (Fig. 1). This ore processing tool was used for middle-to-fine pulverization. It dates to the second century A.D. (Kaptan. 1984).

The second find, a stone mortar used for breaking larger chunks of ore. was found al Merzifon-Bakırçay (Fig. 1).

mortar with multiple depressions (Fig. 5). 28 holes for breaking the ore are on me surface. It was used in the phase of enriching ore between middle-sized chunks and fine powder. This ore enrichment tool from Bakırçay dates to the 1-2 centuries A.D.. the Roman period (Kaptan. 1987). Another multi-holed stone mortar was found in the copper slag mounds at Tokat-Erbaa near Ezebağı village. Made of andesite, both faces of this mortar were used. 33 depressions are on its obverse, and 9 on its reverse (Kaptan, 1986). A similar exasimple was found al Kırklareli-Dereköy (Fig. 1) at the Çitak tepe slag mound. The polygonal shaped find was used as a mortar, but its date has not been ascertained.

Similar mulli-holed mortars were found al the Rio Tinto mining region in Spain (Blanco and Luzon. 1969). Another

complex (Kaptan. 1988, Yener. 1988). This open air workshop had 213 depressions on its marble surface (Fig. 6). The quantity of these depressions will probably increase as more work is done at this site. This is the first such mulli-holed-ore-beneficiation-workshop found in Turkey and is unique in the world. Ancient mining began here in the Bronze Age (3000-1200 B.C.) and continued through the Roman and Byzantine periods.

Fig. 5

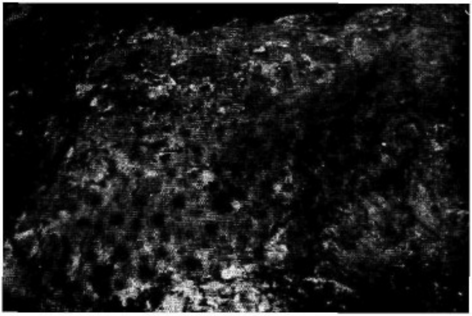

Fig. 6

ANCIENT METALLURGY FINDS

It is normal that ancient metallurgical tools, which are the continuation of the mining process, are found in Anatolia. The oldest find relating to ancient metallurgy was found *at* a small slag mound 1 kilometer to the north of Tokal-Almus Bakımlı village (Fig. 1), Furnace lining fragments were Found inside the slag heap. And arc a result of copper smelting. When the slag is light weight, it indicates that metallurgists were successful in their smelting. This slag mound (Kaptan. 1986) is the only find so far of a successful and knowledgeable copper smelting process in Anatolia dating to the Early Bronze Age (3000-2000 B.C.).

The second oldest metallurgical finds are from the slag mound at Subaşı (Çesme başı) at Büklüce village, overlooking the Merzifon-Bakırçay valley. A large amount of tuyeres were found in the copper slag mounds (Fig. 7). In addition, wood charcoal used during smelling gave radiocarbon dates of 1887 B.C. (Ergin and Güler. 1985). Therefore, this find makes evident that the Anatolian metallurgists were successfully smelting copper at the beginning of the second millennium B.C.

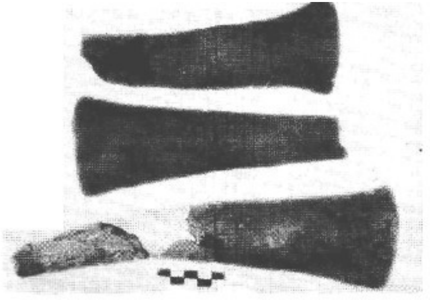

Fig. 7

Another copper slag mound at Ankara-Bala. Karaali village measure 70.000 tons and represents the Roman-Byzantine, Selçuk-Ottoman periods in the first millennium A.D. A 1.5 million ton slag deposit at Gümüş village near Amasya-Gumiyhacikoy represents the remains of ancient lead smelling. It is possible that metallurgical activities which began in the first millennium B.C. in this area, continued into the next millennia. These workings were continued in the Roman-Byzanline and Ottoman periods. In addition, smelting furnaces in Gümüş were used until 1886 (Kaptan. 1975).

There are more than 200 slag deposits in Turkey. 40 of these slag heaps in central and north central Anatolia have been established as belonging to the Roman-Byzantine periods. Eastern Black Sea slag heaps represent the smelling of the Cu-Pb mineralization and represent the Roman-Byzantine and Ottoman periods primarily.

CONCLUSION

The first metal that was recognized and smelted by ancient man in Anatolia was copper. The greatest find of Anatolian industry was the achievement of making bronze alloys by adding tin to copper. These are the developmental stages which led to modem high technology. Research into the history of metallurgy in Anatolia began in 1973. As a result, tools and equipment relating to mining, ore beneficiation and metallurgy have increased appreciably. These have conlribuled to the understanding of the history Anatolian metal industry and its developmental slages. In addition. due to the research into the history of metals, tin was discovered in two separate areas in Niğde. For this reason, the combination of the geological search for metals and archaeomelrical research, as it is in other countries, will yield important results.

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