GEOLOGY OF THE WESTERN PART OF THE EASTERN TAURUS BELT (SSE OF TURKEY)

Sait METİN*; Abdulkadir AYHAN* and Ibrahim PAPAK*

ABSTRACT. — The western part of the Eastern Taurus Belt comprise three different groups that include sedimentary, metamorphic and magmatic rocks. These groups are composed of four main rock units according to their structural anel stratigraphic characteristics. Göksun metamorphics occur in the eastern part of the Göksu fault. Andinn complex occurs in the southeastern part of the Göksu fault. The Taurus autochthonous sequence occurring in the eastern part of the Göksu fault is tectonically covered by the Kire9likyayla allochthonous ophiolite complex. The Taurus autochthonous sequence and the Göksun metamorphics show a uniform stratigraphic sequence in contrast to the ununiform stratigraphy of Kireçlikyayla and Andırın complexes comprising ophiolites and Mesozoic limestones of different age and characteristics. The autochthonous sequence is represented by a uniform and thick sedimentation ranging from Cambrian to the Quaternary. It is fosilliferous throughout. It consists essentially of well-bedded platform carbonates with minor elastics. The region is extremely tectonized resulting in numerous overthrusts trending NE and extending for 50 to 100 km. The data obtained suggests a NW-SE compression. The intensity of deformation of the Göksun metamorphics shows a proressive diminution from base the top. Is composed mainly of schists with lenses of marbles. The uppermost section is incipiently deformed comprising seldom occurrences of Jurassis-Createceous fossils. Kire9likyayla complex was emplaced during the Maastrichtian. The emplacement of the Andırın complex possibly postdates Eocene. These rock units are unconformably overlain by Oligocene-Pliocene sediments.

STRATIGRAPHY OF THE UPPER CRETACEOUS AND PALEOGENE IN YIĞILCA-BOLU (NW TURKEY)

Orhan KAYA*; Atife DİZER**; İzver TANSEL** and Sacit ÖZER*

ABSTRACT. — The post-Triassic rocks of the Yığılca area extend in age from late Santonian/early Campanian to Ypresian or probably Lutetian, and constitute a marine sequence intervened by submarine and subaerial unconformities. The sequence is represented primarily by volcaniclastic sandstone and conglomerate, derived from a provenance of mafic volcanic rocks, and subordinately by epiclastic sandstone. The Cretaceous and Paleogene rocks of the Yığılca area have a marked lithic and sequential similarity to those in the northerly-lying Ereğli area. The Cretaceous is suggested to have been deposited by the progressive onlap of an apparently southward transgressing sea. The coeval but differing rocks of the uppermost Cretaceous and Paleogene in the Yığılca, and southerly-lying Bolu and Mengen areas, suggests a structural divide to the south of the Yığılca area, formed by the latest Cretaceous. The totally 200 m thick Cretaceous olistostromal interval of the Yığılca sequence, which has a 4 m thick correlative in the Ereğli sequence, has apparently a significant bearing on the stratigraphy of the so-called «Ankara melange». The arcal structure is characterized by a southward recumbent syncline whose northern limb is thrusted on the southern. High-angle faults are suggested to be originally shear fractures related to a nearly northeast-trending horizontal acute bisector, and later to have acted as extensional. High-angle faults postdate the thrust, and both are probably post-Lutetian in age.

INTRODUCTION

The map area is situated in G26-b2 and bl sheets of 1:25,000 scale (Fig. 1). The pioneer regional work was done by Blumenthal (1948). Batum (1968), Görmüş (1980, 1982a, *b),* Bürkan and others (1982) and Kaya (1982) have studied the map and surrounding areas.

The geology of the Yığılca area has a significant bearing on the understanding of the north-south stratigraphic variations in the Cretaceous and Paleogene sequences of the western parts of northern Anatolia.

The symbols used in the graphic presentations are explained in Figure 2. The terminologies used herein for sandstones and mudrocks follow the classifications of Gilbert (1954: in Williams and others, 1954), and Lundegard and Samuels (1980), respectively. The term «limy» is used to qualify the clastic rocks with high carbonate content, which may have the deceptive appearance of a carbonate rock in the field.

O. Kaya is responsible for the field data and text. A. Dizer, İ. Tansel and S. Özer contributed by studying the benthic and pelagic foraminifers, and rudistids.

STRATIGRAPHY

The generalized rock succession of the Yığılca area is given in Figure 3. The age of the poorly fossiliferous or non-fossiliferous units, and classification of some rock units with poorly exposed contact relationships are based on the correlations with the northerly-lying Ereğli area (Fig. 4). The geologic map and representative cross-sections are given in Figure 5 a,b. The Paleozoic and older rocks (Kaya, 1982) have not been dealt with in this study.

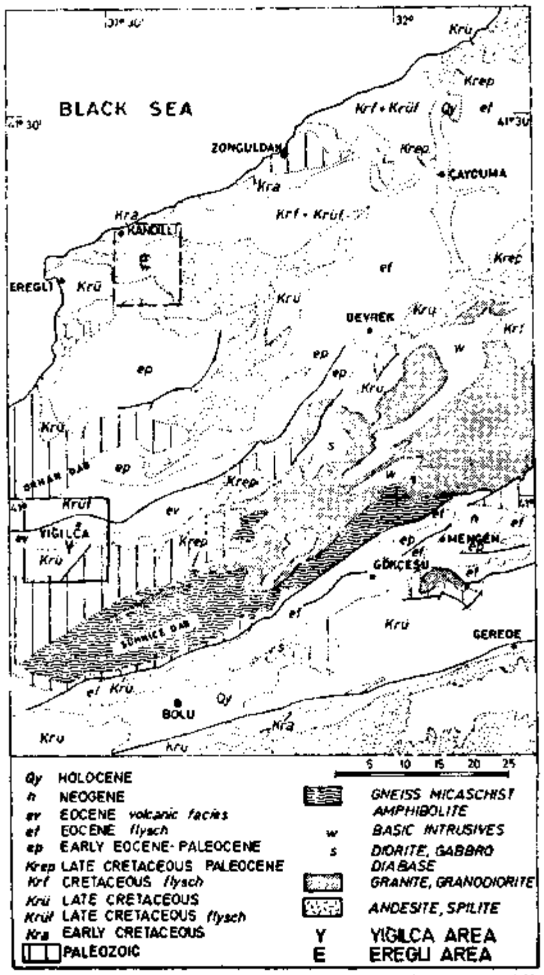

Fig. 1 - Geologic setting of the Yığılca (study) and Ereğli (reference): areas in 1:500, 000 scale geologic map of Turkey (Zonguldak sheet).

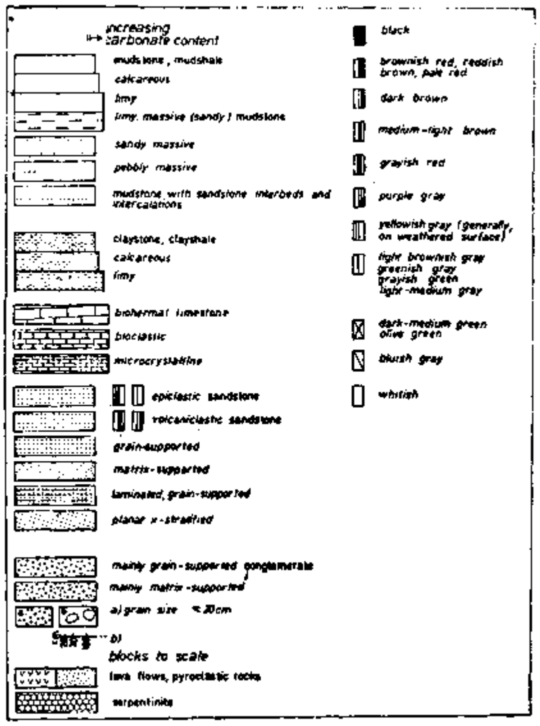

Fig. 2 - Symbols for rock types and colors.

Kırık formation

The name Kırık formation is here applied to a sequence of red beds consisting of slate, slaty lithic sandstone and conglomerate. The partial type section is exposed between 66.00:29.55 and 65.90:27.75. Reference sections are situated at 63.50:31.23 and 67.85:30.17. The Kırık formation corresponds to Görmüş's (1982a) probably late Devonian «Değirmendere formation».

The slate is purplish red to reddish brown, moderately indurated, thinly bedded, and originally mudshale and clayshale. The slaty sandstone is fine-grained quartzose lithic arenite, and lithic wacke. The basal conglomerate (69.47:31.73) is primarily gray, poorly stratified, and is divisible into three parts. The bottom part is poorly consolidated and matrix-supported lithic conglomerate. The constituents are angular to subround and with little sorting in size, and include shale, sandstone and limestone of Devonian age. The middle and thicker part of the basal conglomerate is well consolidated, massive and carbonate-cemented limestone-pebble conglomerate. The clasts are round to

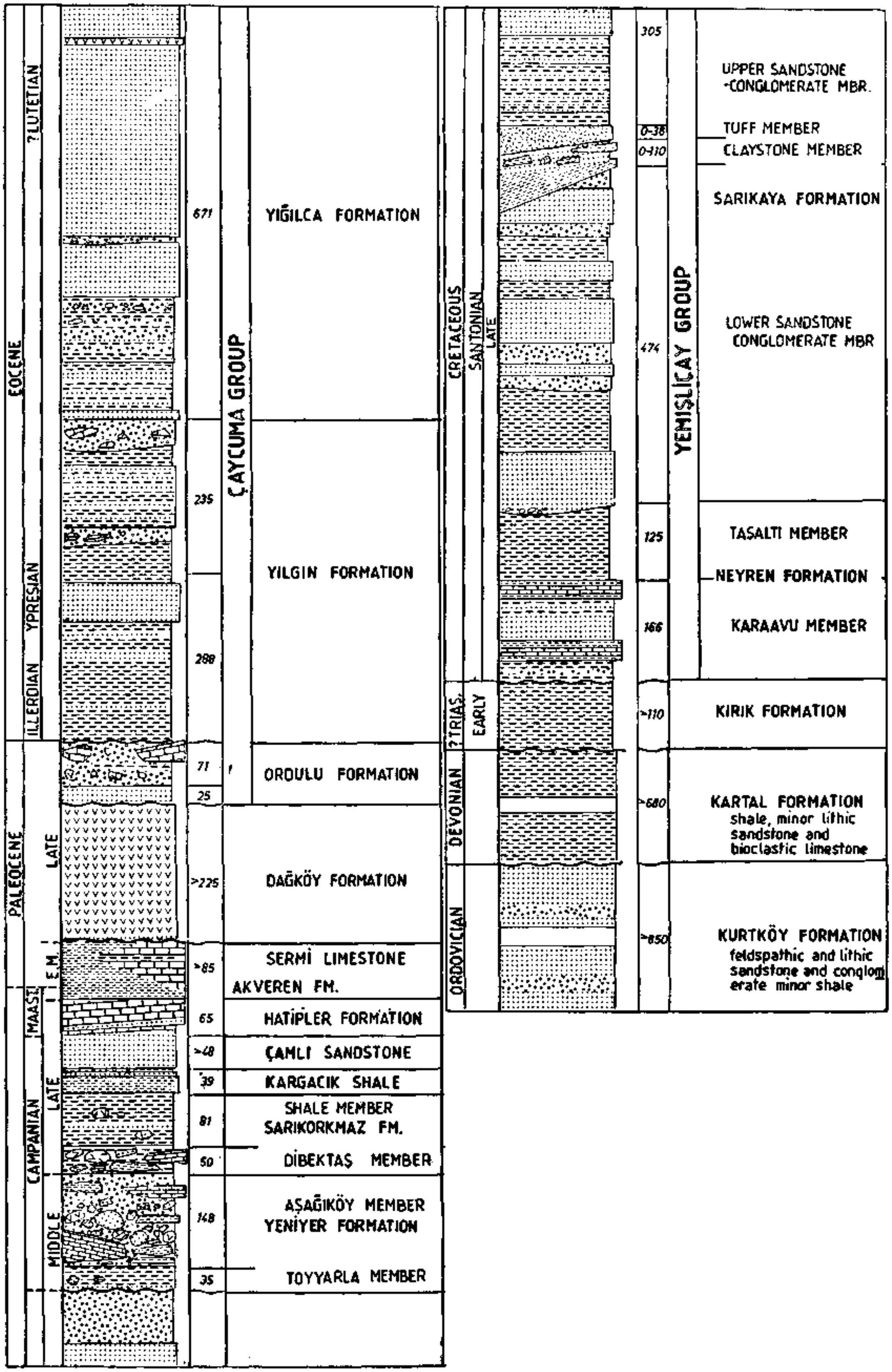

Fig. 3 - Generalized rock succession of the Yığılca area. The Peleozoic and older rocks (Kaya, 1982) have not been dealt with in this study.

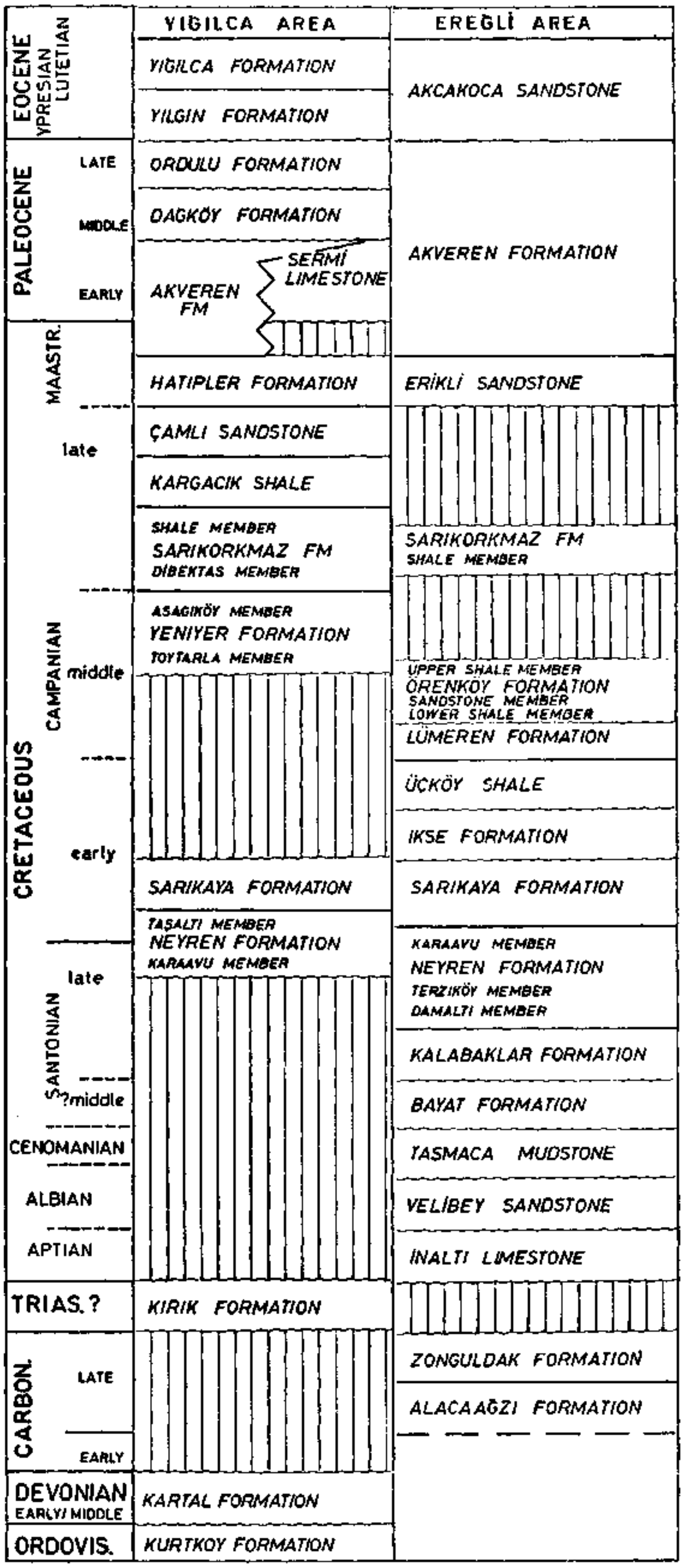

Fig. 4 - Correlation of the rocks units of the Yığılca and Ereğli areas. The Ereğli succession is from Kaya and others (1984*c*).

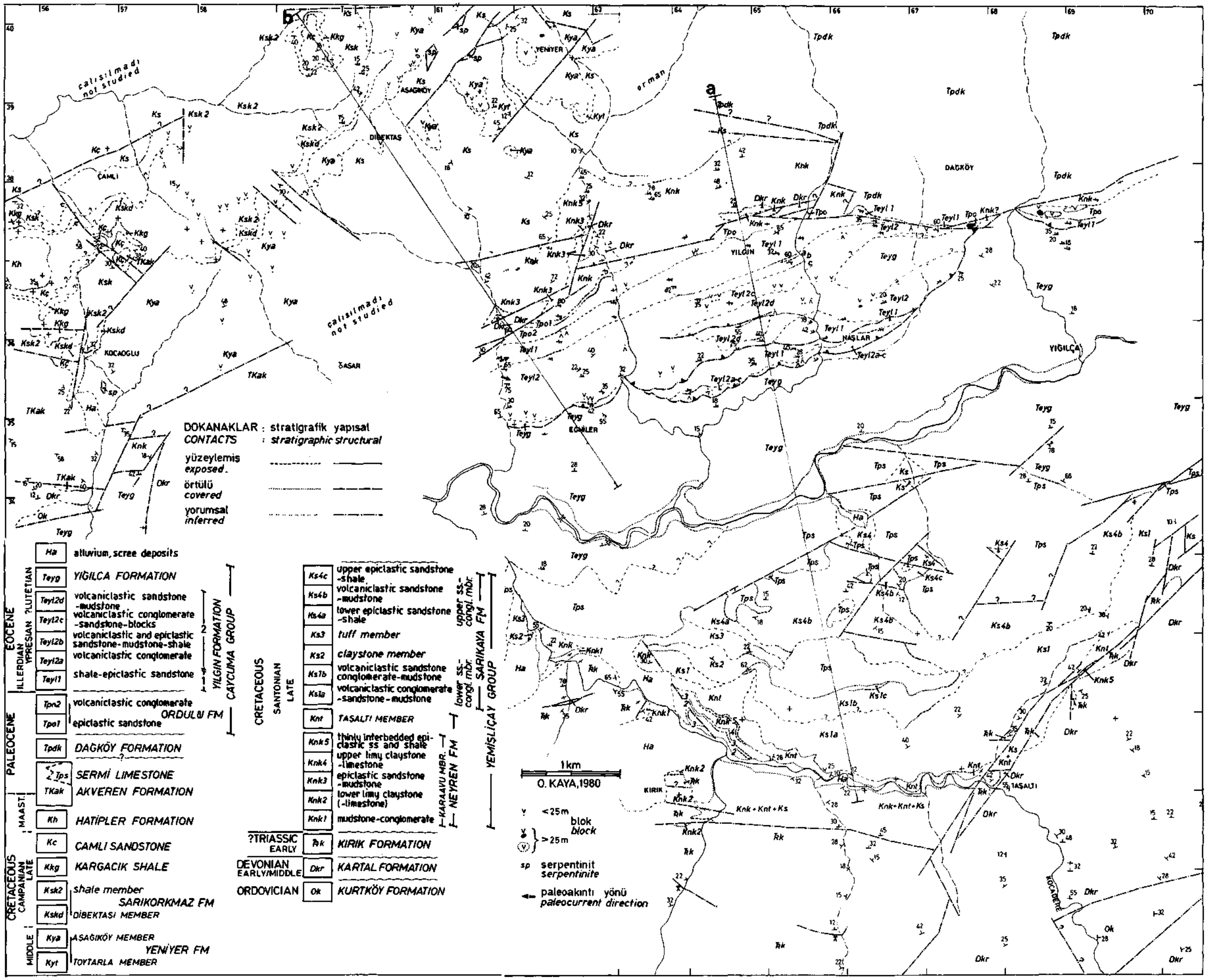

Fig.5 a – Geologic map of the Yığılca area.

subround, moderately sorted in size, and include mainly Devonian limestone up to 35 cm in size. The upper part is poorly to moderately indurated and grain-supported lithic conglomerate. The constituents include mainly Devonian rocks, minor volcanic rocks, chert and vein quartz. A widely extending lenticular bed of conglomerate exists in the apparently middle part of the formation. It is up to 8 m thick, gray, in places well indurated, massive, grain-supported and carbonate-cemented limestone-pebble conglomerate. The clasts include dark gray Devonian limestone, white marble, light gray and brownish gray bioclastic limestone. The grains and carbonate cement are pervasively recrystallized. Thin interlayers of brownish gray, fine pebbly clayey limestone occur in the lower part of the conglomerate.

The basal conglomerate of the Kırık formation rests directly on the Lower Devonian rocks. The unconformity is well exposed at the locality 69.49:31.65. Deformed contacts occur at 67.90:30.13 and 69.07:31.50. The slate, which makes up the bulk of the formation, lies conformably and abruptly on the basal conglomerate.

The Kırık formation is barren of fossils. It underlies the late Santonian and overlies the Middle Devonian strata. Because the Kırık formation is in part lithically comparable with the probably Permian to early Triassic rocks in İstanbul (Asserato, 1972), it is tentatively suggested to be early Triassic in age (Kaya, in prep.). A lithic correlation with the late Devonian rocks of the Istanbul area (Kaya, 1973), as considered by Görmüş (1982a), is not adequate.

YEMİŞLİÇAY GROUP

The name Yemişliçay group has been applied by Kaya and others (1984c) to a heterogeneous assemblage consisting primarily of volcaniclastic sandstone and conglomerate. It contains subordinate but significant amounts of mafic tuff, blockstone, agglomerate, lava, pelagic limy mudrocks and limestone. The assemblage which is characterized by volcaniclastic rocks, contains minor epiclastic rocks at the basal part. The Yemişliçay group corresponds to Ketin and Gümüş's (1963) Senomanian-Campanian «Yemişliçay formation», and Tokay's (1952) Turonian-Coniacian «volcanic flysch».

The Yemişliçay group is divisible into four formations: in ascending order, (1) The Bayat formation consisting of epiclastic and volcaniclastic sandstones; (2) The Kalabaklar formation consisting of mudrocks and mainly epiclastic sandstone; (3) The Neyren formation consisting of volcaniclastic sandstone, mudrocks, and minor limy claystone and microcrystalline limestone; (4) The Sarıkaya formation consisting of volcaniclastic conglomerate, sandstone, mudrocks, and minor mafic tuff. The Bayat and Kalabaklar formations are restricted to the Ereğli area. The Neyren and Sarıkaya formations are widely exposed in the study area.

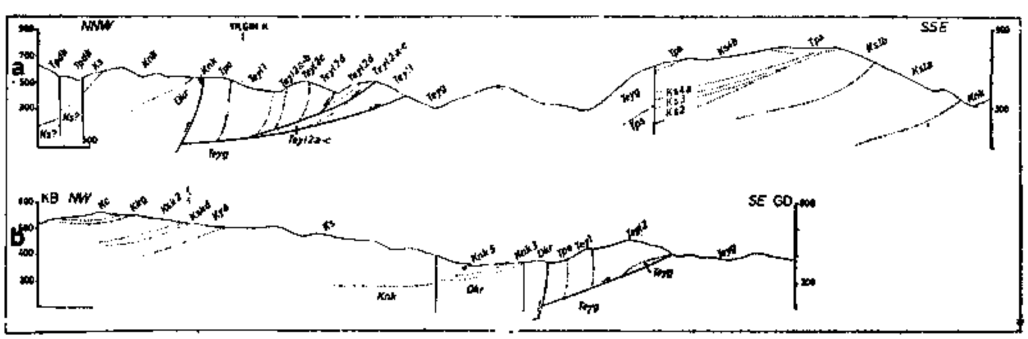

Fig. 5b - Cross-sections of the Yığılca area.

The entire Yemişliçay group is suggested to extend in age from probable Middle Santonian to early Campanian (Kaya and others, 1984c).

Neyren formation

The Neyren formation was designated by Kaya and others (1984c) for a heterogeneous sequence consisting primarily of volcaniclastic sandstone, mudrocks, and minor but stratigraphically significant limy claystone, microcrystalline limestone and mafic tuff. The type section is situated in Ereğli. The formation is divided into four formal members: in ascending order, (1) The Damaltı member consisting of volcaniclastic sandstone; (2) The Terziköy member consisting of mudrocks with local thin interlayers of volcaniclastic sandstone; (3) The Karaavu member consisting mainly of volcaniclastic sandstone, mudrocks, and minor limy claystone and microcrystalline limestone, and (4) The Taşaltı member consisting primarily of mudrocks. The Damaltı and Terziköy members are confined to the Ereğli area. The Karaavu member is widely exposed in the Ereğli and Yığılca areas. The Taşaltı is restricted to the Yığılca area. The Karaavu and Taşaltı members correspond to parts of Görmüş's (1982a) «Hızardere formation of Senonian-Campaman-Maastrichtian and partly Turonian age».

The uppermost part of the Neyren formation contains pelagic foraminifers indicating the turn in age from late Santonian to early Campanian. Accordingly, the main bulk of the formation appears to be late Santonian in age.

Karaavu member. — The name Karaavu member was designated by Kaya and others (1984c) for a heterogeneous sequence of volcaniclastic sandstone, mudrocks, and minor limy claystone, microcrystalline limestone, epiclastic sandstone and mafic tuff. The type section was established in the Ereğli area. In the map area, for practical field purposes, the Karaavu can be divided into 5 units: in ascending order, (1) mudstone-conglomerate, (2) lower limy claystone (-limestone), (3) epiclastic sandstone-mudstone, and (4) upper limy claystone-limestone. The type sections of this units are exposed, in the same order, at 62.50:31.85, 64.70:30.58, 64.70:30.60 and 64.98:30.52 (Fig. 6C). In the northernmost part of the map area the Karaavu member is mainly represented by a (5) thinly interbedded epiclastic sandstone-shale unit typically exposed between 62.97:37.55 and 62.80:38.15.

In the mudstone-conglomerate unit of the member, the mudstone is greenish gray, and contains sporadic thin interlayers of turbiditic lithic arenite. The conglomerate is up to 9 m in thickness, laterally discontinuous, moderately indurated, matrix-supported and polymictic. The clasts are angular to subrounded, poorly sorted in size, and include Kırık rocks, which are up to 60 cm in size, Paleozoic limestone and mudrocks, probably Mesozoic microcrystalline limestone and lithic arenite and vein quartz. The limy claystone is pale red, purplish gray and locally greenish gray, well indurated and brittle, and contains thin interlayers or sets of thin beds of microcrystalline limestone, in the same colors. The sandstone of the epiclastic sandstone-mudstone unit is dark gray, medium to massively bedded, medium to coarse-grained and carbonate cemented quartzose lithic arenite. The sandstone locally contains angular intraclasts of mudstone and horizontal burrowings. In the northern part of the study area the sandstone representing the base of the Karaavu is thickly bedded, coarse to very coarse-grained and carbonate cemented lithic arenite. It is mostly planar cross-stratified, and locally contains pebble to small block-sized Devonian limestone, minor pebbles of vein quartz and altered volcanic rocks, and large leaf prints. The sandstone of the uppermost unit is thinly bedded, fine-grained lithic arenite and wacke, interbedded with shale. The tuff of this unit is greenish gray, massive, medium to very coarse-grained and is related to a mafic source rock.

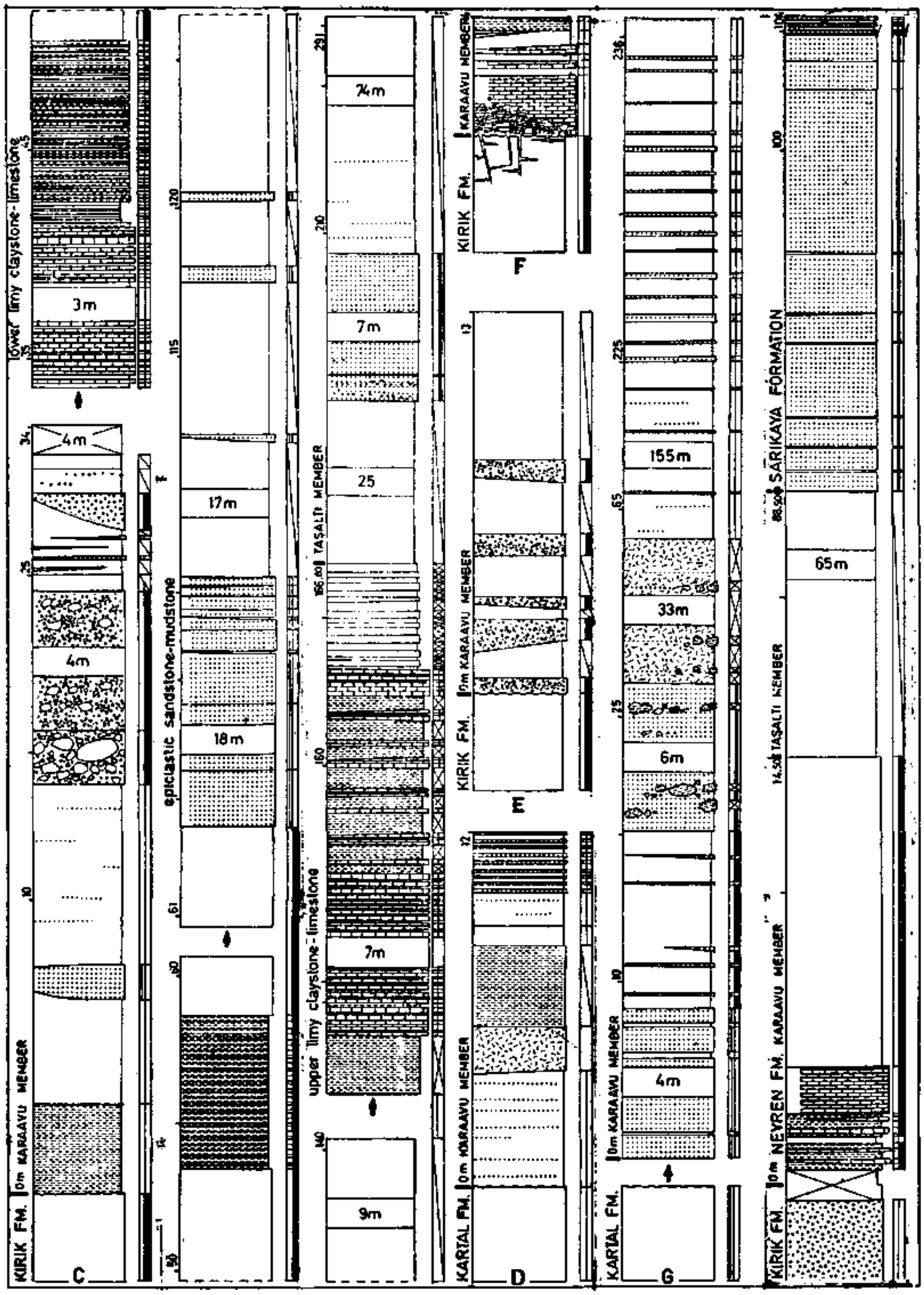

Fig. 6 - C- Complete reference section of the Karaavu member exposed in the southern parts of the map area, partial type section of the Taşaltı member; D,E,F - Lower contact of the Karaavu member with the Devonian, and probably early Triassic Kırık formation; G-Reference section of the Karaavu member exposed in the northernmost part of the study area.

The contact between the lower limy claystone (-limestone) unit of the Karaavu member and the Kırık formation is an unconformity. At 62.40:31.88 it is defined by basal mudshale and conglomerate (Fig. 6E). At 63.38:30.36 and 64.17:29.78 *Trocholina-bearing* basal bioclastic limestone is penetrated into the regolithic material of the Kırık formation. The contact between the lower limy claystone (-limestone) and overlying sandstone-mudstone units of the Karaavu member is gradational (Fig. 6C).

The lower limy claystone (-limestone) unit contains *Globotruncana bulloides* Vogler, G. *coronata* Bolli, G. *lapparenti* Brotzen, G. *linneiana* (d'Orbigny), G. *tricarinata* (Querau), G. *ventricosa* White and abundant *Heterohelix* sp., *Hedbergella* sp. and *Praeglobotruncana* sp., which indicate the turn from Santonian to early Campanian age. The upper limy claystone-limestone unit of the member carries G. *coronata* Bolli, G. *lapparenti* Brotzen, G. *linneiana* (d'Orbigny), G. cf. *bulloides* Vogler and G. cf. *area* (Cushman) supporting the above age assignment.

Taşaltı member. — The name Taşaltı member is here applied to the uppermost part of the Neyren formation, consisting primarily of mudrocks, and minor interlayers of volcaniclastic sandstone. The partial type section is exposed around 68.85:31.56 (Fig. 6H). The reference sections are situated between 65.28:30.50 and 65.32:30.63 (Fig. 6C), and 64.77:31.50 and 64.85:31.75.

The mudrocks are bluish gray, generally massive mudstone, clayshale and mudshale. The sandstone occurs in the middle part of the member, and includes thin to thick-bedded, laterally discontinuous volcaniclastic lithic arenite and lithic wacke. The latter locally contains coalified plant material in the form of plant twigs and stems, up to 25 cm wide and 60 cm long (64.69:31.02).

The lower contact of the Taşaltı member has been taken arbitrarily as the top of the upper limy claystone-limestone unit of the Karaavu member at 68.85:31.56 (Fig. 6H), 65.05:30.52, 64.42:31.05. The contact is gradational.

Sarıkaya formation

The name Sarıkaya formation was first used by Kaya and others (1984c) for a heterogeneous sequence of volcaniclastic sandstone, conglomerate, mudstone, and minor mafic tuff, claystone and shale, typically exposed in the Yığılca area. The Sarıkaya formation corresponds to Görmüş's (1982a) late Cretaceous «Hızardere formation». The proportions and thicknesses of the constituent rocks show great lateral variation over short distances. However, on its type section, the Sarıkaya formation is divisible into four informal members: in ascending order, (1) A lower sandstone-conglomerate member; (2) A claystone member; (3) A tuff member; (4) An upper sandstone-conglomerate member. These are practical lithologic subdivisions established for a better understanding of the detailed stratigraphy of the Sarıkaya in the type area.

The age of the Sarıkaya formation cannot be determined directly. According to its superjacent position on the upper part of the Neyren formation representing the turn from late Santonian to early Campanian, it may be early Campanian in age.

Lower sandstone-conglomerate member. — This member consists primarily of volcaniclastic sandstone, conglomerate, mudstone, and minor mafic tuff. A nearly complete composite type section, and a reference section of the member are established between 65.32:30.63 and 65.39:31.27 (Fig. 7), and between 68.83:31.56 and 68.35:31.97, respectively. On its composite type section the member is divisible into two units: a lower volcaniclastic conglomerate-sandstone-mudstone and an upper volcaniclastic sandstone-conglomerate-mudstone unit.

The conglomerate is mainly brownish gray, very thick-bedded to massive, grain-supported and without fabric elements. Locally, it is scour channel filling up to tens of meters in thickness and a few kilometers in lateral extension. The clasts are rounded to well rounded, fine to very coarse-grained and with little sorting in size. They include mainly mafic volcanic rocks, locally up to 7 m across, and intragenetic mudstone and sandstone. Many of the conglomerate beds show a remarkable fining-upward and terminate with matrix-supported conglomerate, pebbly sandstone and mudstone in turn. The massive conglomerate bodies appear to represent turbidite facies A. The sandstone is brownish gray, fine to very coarse-grained volcaniclastic and feldspathic lithic arenite. The sandstone beds are either stacked, or interlayered with mudstone of nearly the same thickness. Several massive beds of sandstone occur in association with the conglomerates. Sole markings, termination with laminated siltstone and fining-upward, and mudstone intraclasts and stringers of volcaniclastic fine pebbles at the base are common in the sandstones, assigning them to turbidite facies B. However, the very local presence of symmetrical ripples suggests the presence of sandstones of traction current origin (66.48:30.57).

The conglomerate and sandstone sequences mostly occur as large scour-and-fill deposits; e.g., the conglomerate and sandstone in the upper part of the lower unit cut out the mudstone predominating lower part and the top beds of the Neyren formation, and the conglomerate and sandstone of the upper unit cut down into the lower unit and, in places, amalgamate with it.

The contact with the underlying Taşaltı member of the Neyren formation is abrupt (65.57:-30.27). The basal volcaniclastic sandstone contains remarkably well rounded, and sometimes perfectly polished epiclastic pebbles. The abrupt contact between the higher beds of the member and the Neyren formation (65.62:30.50, 65.32:30.63, 64.82:31.63) characterizes a huge channel scour.

Claystone member. — This member consists of claystone and volcaniclastic sandstone with abundant claystone blocks. The type section is exposed between 64.44:31.52 and 64.39:31.88.

The claystone is grayish green, massive and locally siliceous or opalized. The volcaniclastic sandstone is brownish gray, moderately indurated, massive, medium to very coarse-grained and fine pebbly lithic arenite. It contains abundant intragenetic clasts of claystone.

The claystone member overlies abruptly the thinning out lower sandstone-conglomerate member (64.91:31.72). At 63.78:31.93 and 63.52:31.93 a 2 m thick mafic tuff separates both members. Synsedimentary deformation occurred before, during and after the deposition of the claystone member.

Tuff member. — This member consists of a unique layer of mafic tuff and minor volcaniclastic sandstone. The type section is exposed between 63.75:32.03 and 64.35:32.17.

The tuff is dark greenish gray, massive and very coarse to fine-grained. It becomes finer upwards, and grades into the mudstone of the overlying unit. The volcaniclastic sandstone occurs at the basal part of the member. It grades upward into the tuff and toward the east into fine pebbly volcaniclastic conglomerate (64.67:31.82). Both the sandstone and conglomerate contain abundant clasts of claystone, varying in size from pebble to very large block, which are derived from the underlying claystone member.

The lower contact of the tuff member with the claystone member is defined by a large-scale erosional channel (64.78:31.87, 63.90:32.03, arid 63.42:31.97).

Upper sandstone-conglomerate member. — This member consists of volcaniclastic sandstone, conglomerate, mudstone, and subordinate but significant epiclastic sandstone and mudshale. The type and reference sections are exposed between 65.27:31.73 and 65.80:32.05 (Fig. 7), and 66.55:32.50 and 66.85:32.53, respectively. The member can be divided into three units: (1) Lower epiclastic sandstone-shale; (2) Volcaniclastic sandstone-mudstone; (3) Upper epiclastic sandstone-shale.

The volcaniclastic conglomerate, sandstone and mudstone are lithically identical to those in the lower parts of the Sarıkaya formation. The epiclastic sandstone, which weathers yellowish gray, is thinly bedded turbiditic lithic arenite interlayered with thicker mudshale.

The member overlies gradationally the tuff member (64.85:31.96). The contact interval is represented by the continuous upward diminution in the grain size of the underlying tuff (Fig.7).

Yeniyer formation

The name Yeniyer formation is here used for a heterogeneous sequence of gray and red mudshale and clayshale, with floating or intimately admixed blocks. The Yeniyer formation corresponds to a section of Görmüş's (1982a) «Hızardere formation» of late Cretaceous age. Partial type section of the formation is situated between 62.82:38.64 and 62.40:39.12. The formation is divided into two formal members: a lower Toytarla member consisting mainly of gray mudshale with minor blocks, and an upper Aşağıköy member consisting of mudshale and clayshale, with abundant blocks.

The Yeniyer formation is barren of fossils. According to its stratigraphic position a Middle Campanian age can be suggested.

Toytarla member. — The name Toytarla member is here applied to a sequence consisting primarily of greenish gray mudshale with floating blocks; and minor sandstone, at the base. The type and reference sections of the member are located at 62.75:39.42, and 62.82:38.75, respectively.

The mudshale is in part slightly calcareous, and contains sporadic thin interbeds of turbiditic (Ta-Tb) lithic arenite. In its lower part the basal sandstone is gray, poorly stratified and coarse-grained lithic arenite (62.75:39.45). It is carbonate cemented, without sorting in size, and made up primarily of volcanic rocks and limestone. The lowermost beds contain bioclastic sandy limestone as lenses and intraclastic angular to rounded pebbles, and mafic volcaniclastic pebbles and cobbles. The upper part of the basal sandstone consists of slightly calcareous lithic wacke and sandy musdstone. All of the lower beds of the member contain abundant prisms of broken Inoceramus and rudist shells. In places, the lower part of the member contains isolated or intimately admixed blocks including pink to pale red microcrystalline limestone and limy claystone, gray microcrystalline limestone, and volcaniclastic conglomerate (62.82:38.64).

The basal sandstone (with mafic volcaniclastic pebbles, cobbles and abundant Inoceramus rudist fragments) of the Toytarla member overlies unconformably the Sarıkaya formation (62.75: 39.42, 61.99:40.5, and less distinctly 63.17:39.52). Gray mudshale (61.30:39.00) and a row of limestone blocks (62.82:38.64) locally define the unconformity.

The extrabasinal red micritic limestone blocks contain *Globotruncana lapparenti* Brotzen, *G. linneiana* (d'Orbigny), G. *coronata* Bolli, and acoompanying *Hedbergella* sp., *Heterohelix* sp. and *Praeglobotruncana* sp., indicating a late Santonian-early Campanian age. They are possibly derived from İkse formation lying in the Ereğli area. Accordingly, a Middle Campanian age can tentatively be suggested for the Toytarla member.

Aşağıköy member. — The Aşağıköy member is here named for a poorly stratified red shale and pebbly mudstone with isolated and intimately admixed blocks (olistostrome). The type section of the member is exposed between 60.38:39.05 and 60.35:39.36 (Fig. 8N).

The olistostrome is of debris-flow depositional origin. The mudrocks constituting the matrix are brownish red clayshale and sandy to fine pebbly mudstone. The blocks are upto several hundred meters across and, most commonly, intimately admixed. They include pink to pale red massive micro-

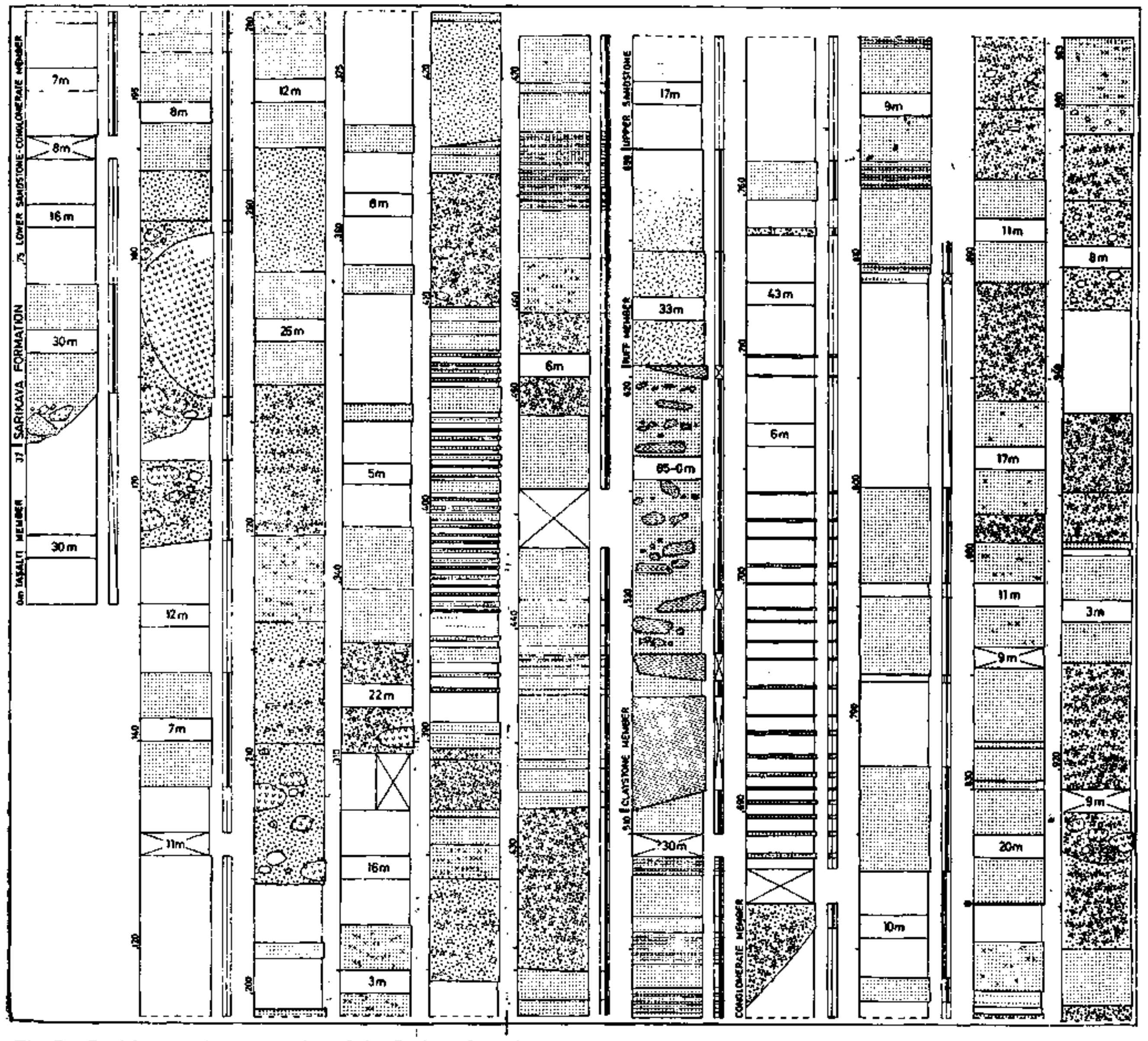

Fig. 7 - Partial composite type section of the Sarıkaya formation.

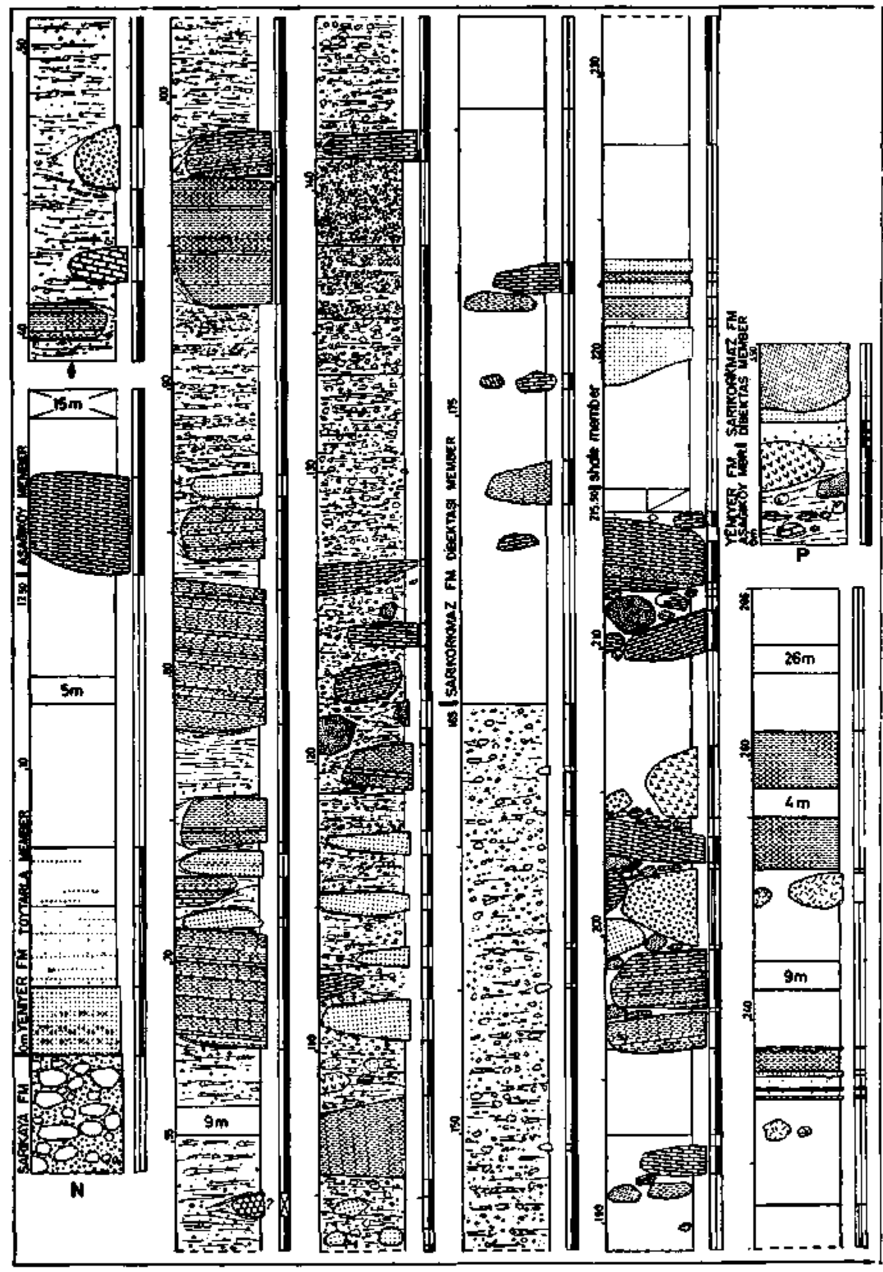

Fig. 8 - N - Complete type section of the Toytarla member, partial type section of the Aşağıköy member, nearly complete type section of the Sarıkorkmaz formation; P - Unconformity between the Sarıkorkmaz and Yeniyer formations.

crystalline limestone, pale red to brownish red thinly interbedded limestone and limy claystone, gray to red lithic wacke and mudstone, greenish gray thick-bedded and coarse-grained lithic arenite with thin interbeds of pale red mudstone, gray recrystallized limestone of probably Paleozoic age, epiclastic rounded pebble conglomerate, volcaniclastic rocks lithically similar to those of the Sarıkaya formation and manganiferous rocks (Atabek, 1940) which might have been derived from the Örenköy formation in the Ereğli area (Fig. 4). The serpentinetes, which indicate a derivation from the ultramafic tectonites, are suggested to be blocks, because (1) their occurrence in the area is confined to the distribution of the member, (2) they are not related to recognizable faults, and (3) the contact with matrix rocks is not intensively deformed.

At the type locality the matrix of the olistostrome has undergone a pervasive shearing typical of debris flow deposits. The evidence for the synsedimentary origin of the shearing is the following: (1) Areally the shearing is confined to the olistostrome part of the Aşağıköy member. It fades out toward the bottom and top of the olistostrome; (2) Discontinuous planes of cleavage in the matrix parallel the bedding; (3) The outer surfaces of the clasts are polished and slickensided while the near -surface parts of the clast are unaffected. The cleavage planes in the matrix are bent toward the clasts.

The contact between the Aşağıköy and the underlying Toytarla member is everywhere covered. At 61.90:38.35 and 61.66:38.37, below a thin soil cover, the contact appears to be abrupt and beds on both sides show a structural conformity. In most places a row of blocks of pale red microcrystalline limestones defines the contact.

The matrix rocks of the Aşağıköy are apparently barren of fossils. The reddish brown to pale red limestone and limy claystone blocks contain pelagic foraminifers ranging in age from late Santonian to the turn from early to Middle Campanian. This age interval and lithic peculiarities of the red limestone and limy claystone blocks may suggest that they are derived from the northerly-lying İkse and Örenköy formations of the Ereğli area (Fig. 4). According to its stratigraphic position the Aşağıköy member can be suggested'to be Middle Campanian in age.

Sarıkorkmaz formation

The name Sarıkorkmaz formation, following Tokay's (1952) nomenclature of the «Sarıkorkmaz series», was applied by Kaya and others (1984c) to a unit of gray mudshale with minor interbeds of epiclastic sandstone and isolated blocks, widely exposed in the Ereğli area. In the study area the Sarıkorkmaz formation contains and olistostrome at the base. The Sarıkorkmaz corresponds to a part of Görmüş's (1982a) late Cretaceous «Hızardere formation». The nearly complete type section of the formation is exposed between 60.25:39.37 and 55.82:39.40 (Fig. 8N). A reference section is situated between 56.65:38.88 and 56.15:39.37. In the Yığılca area the Sarıkorkmaz is divided into a formal Dibektaş member and an overlying informal shale member.

The Sarıkorkmaz formation is barren of fossils, in the study area. A late Campanian age can tentatively be suggested, on the basis of its stratigraphic position.

Dibektaş member. — The name Dibektaş member is here applied to an olistostronial unit at the base of the Sarıkorkmaz formation. The type and reference section are exposed at 60.25:39.37 (Fig. 8N), and 56.65: 35.87 and 60.00:38.65, respectively.

The olistostrome is a nonstratified, chaotic unit consisting of closely packed to matrix-supported, angular to subrounded blocks. The matrix is gray mudstone and lithic wacke. The clasts are round to subrounded, and range in size from pebble to large block up to 10 m on one axis. They include primarily brownish red to pale red limy claystone and limestone, which are lithically similar to those of the Örenköy and İkse formations of the Ereğli area (Fig. 4), gray sandstone with different textural and diagenetic grades, and volcanic rocks derived from the Sarıkaya and Lümerli formations, the latter being confined to the Ereğli area.

The contact between the Dibektaş member and the underlying Aşağıköy member of the Yeniyer formation is everywhere abrupt. At 56.65:35.88 the basal lithic wacke contains angular fine pebbles of red mudstone derived from the Aşağıköy. Across the contact, the composition and size of the coarse clasts show an abrupt change, and the shearing typical of the Aşağıköy matrix material disappears.

Shale member. — The shale member, which represents the bulk of the.Sarıkorkmaz formation, consists of mudshale with sporadic epiclastic sandstone interlayers. The type and reference sections are exposed between 60.18:39.17 and 59.97:39.43, and at the surroundings of 56.20:35.90 and 57.25: 37.32, respectively.

The mudshale is medium gray, thickly bedded to massive, and weathers light brownish gray. The epiclastic sandstone occurs as thin to thick-bedded channel-fills at wide intervals. Thinly bedded sandstone represents Ta and Tb Bouma divisions. In the lower part of the member the mudshale contains widely isolated blocks of red and gray limy clayshale and claystone up to 80 cm in size, and lava blocks (60.03:39.20, 57.48:38.27), as much as 175 cm in diameter, derived from the Lümeren formation in the Ereğli area.

The contact between the shale member and the matrix rocks of the Dibektaş member is gradational (56.80:37.65, 60.20:39.40). At 60.25:39.37 gray mudshale with floating blocks overlies abruptly the Aşağıköy member (Fig. 8P).

Kargacık shale

The name Kargacık shale is here used for a unit of primarily reddish gray, thinly bedded to laminated, limy clayshale and mudshale. The Kargacık shale corresponds to a part of Görmüş's (1982a) late Cretaceous «Hızardere formation». The complete type section is exposed at 57.20:37.18. The reference sections for the lowermost and uppermost parts of the Kargacık are situated at 56.12; 36.00 (Fig. 9B) and 59.37:40.02 (Fig. 9D), respectively.

The clayshale and mudshale are reddish gray, grayish brown, pale red, thinly bedded to laminated and calcareous to limy. The main body consists of rhythmically interbedded carbonate-poor and rich layers. The mudshale is predominant in the lower part of the Kargacık. The upper part is represented mainly by fucoidal clay shale with sporadic interlayers of thinly bedded fine-grained lithic arenite and mudshale.

The gradational contact of the Kargacık shale with the underlying shale member of the Sarıkorkmaz formation is well exposed at 56.15:35.97 (Fig. 9B), 57.10:37.10, 59.35:40.02. The contact interval consists of a sequence of thinly interlayered red and gray claystone, mudstone and sandstone. At 55.62:37.08 the Kargacık shale overlies unconformably the Sarıkaya formation, with a nearly 35 cm thick greenish gray mudstone (Fig. 9A).

The Kargacık shale is barren of identifiable fossils. A late Campanian age can be suggested depending on its stratigraphic position.

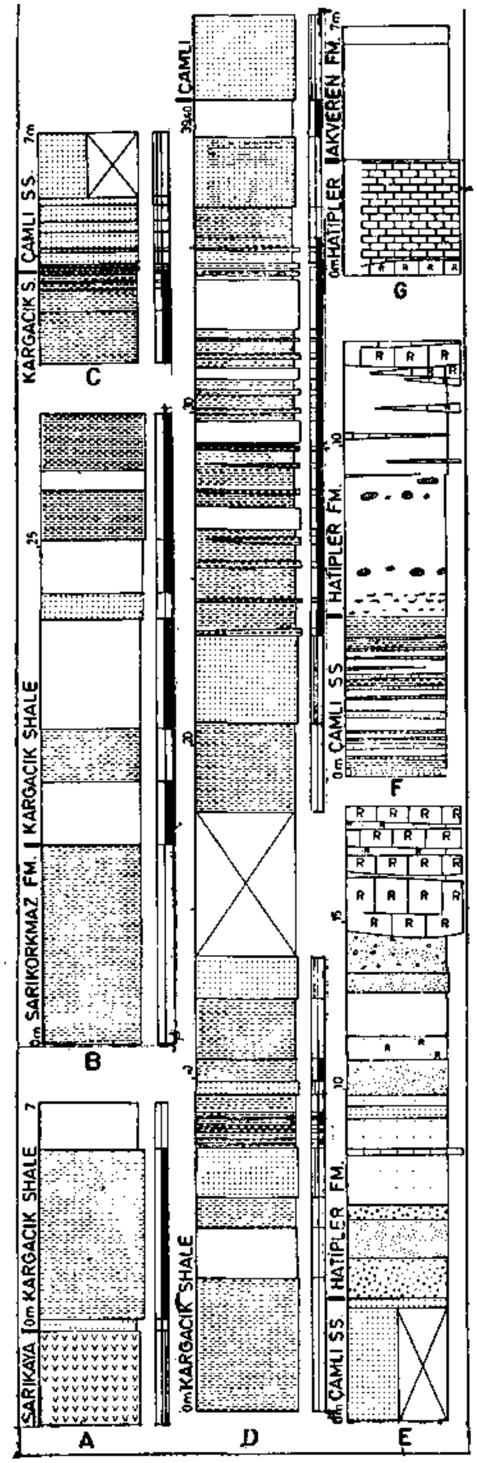

Fig. 9 - A- Unconformable contact between the Kargacık shale and Sarıkaya formation; B- Reference section for the lower part of the Kargacık shale, and its gradational contact with the Sarıkorkmaz formation; C- Gradational contact between the Çamlı sandstone and Kargacık shale; D- Reference section for the upper part of the Kargacık shale; E,F- Gradational contact between the Hatipler formation and Çamlı sandstone; G- The abrupt but conformable contact between the Akveren and Hatipler formations, which is suggested to represent a hiatus related to an abrupt lateral shift of lithotope.

Çamlı sandstone

The name Çamlı sandstone is here applied to a primarily epiclastic sandstone sequence with minor mudstone interlayers. The Çamlı sandstone eorresponds to a section of Görmüş's (1982a) late Cretaceous «Hızardere formation». The partial type section representing the main sandstone body of the Çamlı is exposed at 56.68:37.27. Partial section of the mudstone rich upper part of the Çamlı is at 57.15:37.13.

The sandstone is medium to thick-bedded and medium to coarse-grained quartzose lithic arenite that is structureless except for the sporadic lamination. It weathers characteristically yellowish gray to grayish orange.

At 56.15:35.98 the contact between the Çamlı sandstone and the underlying. Kargacık shale is gradational over a wide interval (Fig. 9D). At 56.85:37.14 the contact is abrupt in respect to sandstone and red shale. There, the presence of fucoidal elements in the lowermost thin sandstone beds of the Çamlı, which are characteristic for the underlying formation, and the termination of the Kargacık with a few gray claystone beds, as much as 6 cm in thickness, indicate an interbedded gradational contact (Fig. 9C).

The Çamlı sandstone is apparently non-fossiliferous. According to its stratigraphic position a late Campanian age can be assigned to it.

Hatipler formation

The name Hatipler formation is here applied to a sequence of rudistid-bearing limestone, and mudrocks and lithic sandstone. The type section of the Hatipler formation is situated outside the map area, in the G26-a2 sheet, between the localities 45.45:31.36 and 45.45:31.20. In the map area the reference section of the formation is exposed between 56.30:36.63 and 56.30:36.73 (Fig. 9E). In the type locality the Hatipler consists of a lower mainly clastic and an upper mainly carbonate part.

The rudistid-bearing limestone is gray, unevenly and thickly bedded and fragmental detrital. The mudrock and lithic sandstone weather grayish yellow and are calcareous to limy. The rudistids are reworked, and decrease in size and abundance upward in the formation.

The contact with the underlying Çamlı sandstone is abrupt and is suggestive of a submarine stratigraphic break. It is exposed in the map area (55.82:36.62, Fig. 9E) and type locality (45.94:31.20, Fig. 9F). The contact relationship between the underlying (Çamlı) and overlying (Akveren) formations indicates the Hatipler formation to be a large-scale lenticular body. The large-scale cross-bedded internal structure of the entire Hatipler, as it is very distinct in the map area, supports the lenticular shape of the Hatipler. In the map area, the lee-sides of the cross-beds have an inclination up to 30°, apparently toward the west.

The Hatipler formation contains *Hippurites radiosus* Des Moulins, *H. colliciatus* Woodward, *Vaccinites ultimus* Milovanovic, *V. loftusi* Woodward, *Joufia reticulata* Boehm, *Radiolites* sp. and *Biradiolites* sp., which are as a whole indicative of a Maastrichtian age.

Akveren formation

The name Akveren formation was used by Ketin and Gümüş, (1963) for the sequence of «interlayered clayey limestone-marl, and minor lava, tuff and sandstone». In the study and surrounding areas the Akveren formation consists of greenish gray limy claystone, mudstone, and sporadic allodapic limestone. Görmüş. (1982) considered the Akveren rocks as the late Cretaceous-early Eocene «Sarıkaya formation» which is herein called the Sermi limestone.

To the west of the Yığılca area (G26-a2 sheet, 49.94:31.20) the massive, calcareous to limy mudstone of the Akveren formation overlies abruptly the Hatipler formation. In the Yığılca area the basal massive sandy limestone (56.29:35.25), and clayey limestone and limy mudstone (57.05:37.07) of the Akveren formation rest conformably on the Çamlı sandstone. At 56.40:34.09 the Akveren rests on the Lower Paleozoic rocks. The above contact relationships indicate an apparent southward onlap of the Akveren formation.

Dizer (1971) established the foraminiferal biostratigraphic zonation in the Akveren formation, and proposed a Maastrichtian to late Paleocene age.

Sermi limestone

The name Sermi limestone is here used for a homogeneous unit of bioclastic and patchily coralgal limestone, and minor limy mudstone at the top. The Sermi limestone corresponds to Görmüş's (1982) late Cretaceous-early Eocene «Sarıkaya formation». Partial type section is exposed between 61.80:32.87 and 61.80:33.07.

The limestone is white, homogeneous and massive. Small-sized coralgal buildups and related bioclastic limestone recur in vertical and lateral extensions, however, without a distinct stratification. A very restricted exposure of light greenish gray, poorly indurated, thinly bedded limy mudstone occurs at the top of the unit.

The Sermi limestone lies unconformably on the different parts of the Sarıkaya formation (66.80:32.65, 67.40:31.40, 65.10:31.57, 65.00:32.38, 69.57:33.62, 62.20:32.30). At many localities the Sermi limestone rests directly on the Sarıkaya but without clasts derived from the latter. In the outside area (G27-al, 82.75:35.50) the Sermi overlies the Lower Paleozoic rocks, with basal elastics up to 80 cm thick.

The limy mudstone, the top bed of the Sermi, contains *Globigerina triloculinoides* Plummer, *Neodiscocyclina barken* Waughan and Cole, *Ranikothalia* cf. *soldadensis* Waughan and Cole, *Discocyclina* sp. (gr. *D. seuneusi)* and *D.* sp. (gr. *D. nummulitica),* which indicate a late Paleocene age. Other foraminifers which are not age diagnostic include *Globorotalia* sp., *Globigerina* sp., *Planorbulina* sp. and *Asterigerina* sp. The Sermi type limestone blocks found in the younger detrital units (e.g. Yılgın formation) carry *Planorbulina cretacea* Marsson, *Alveolina ovulum, Miscellanea* sp., *Glomaheolina* sp., *Lockhartia* sp., *Globorotalia* sp., *Discocyclina* sp., *Distichoplax* sp. and *Asterocyclina* sp., which as a whole indicate a lower age limit of Middle Paleocene for the Sermi.

The Sermi limestone appears to be age equivalent with the upper part ofthe Akveren formation. Both formations lie at least 5 km apart, the Sermi representing a compound carbonate buildup developed on a paleomorphological high to the south.

Dağköy formation

The name Dağköy formation is here applied to a homogenous unit of massive subaerial lavas of intermediate composition. The Dağköy formation corresponds to Görmüş's (1982a) «Keltepe volcanics» of Neogehe age.

The lower contact and the underlying unit (s) are not exposed. The well exposed unconformity between the Dağköy and the overlying earliest Eocene Yılgın formation indicates the Dağköy formation to be pre-Eocene in age. The volcaniclastic rocks of the Ordulu formation lithically correspond to the volcanics ofthe Dağköy formation. A Paleocene age for the Dağköy can tentatively be suggested.

ÇAYCUMA GROUP

The name Çaycuma group is applied by Kaya and others (1984c) to a sequence of epiclastic and volcaniclastic rocks, and minor volcanic rocks. Broadly, it corresponds to the «Çaycuma formation» of Saner and others (1979). The group is divided into four formations: in ascending order,

1. The Ordulu formation consisting of volcaniclastic conglomerate;

2. The Akçakoca sandstone consisting primarily of epiclastic sandstone, and being confined to the coastal parts of the Black Sea;

3. The Yılgın formation, the time-equivalent of the Akçakoca sandstone in the study and surrounding areas, consisting of mudrocks, epiclastic sandstone and minor volcaniclastic sandstone;

4. The Yığılca formation consisting uniformly of volcaniclastic sandstone, conglomerate and minor mafic lava. In the study area the Ak9akoca sandstone is" not exposed.

Ordulu formation

The name Ordulu formation is here applied to a blocky unit consisting primarily of volcaniclastic rocks, and epiclastic sandstone and mudstone at the base. The Ordulu formation corresponds to a part of Görmüş's (1982a) late Cretaceous «Hızardere formation». The Ordulu is divisible into a lower epiclastic sandstone and an upper volcaniclastic conglomerate unit. The complete type section of the formation is situated between 63.12:36.67 and 63.17:36.35 (Fig. 10A). Reference sections are exposed between 62.46:36.13 and 62.46:35.95, and 64.80:37.37 and 64.81:37.11.

The volcaniclastic rocks are dark greenish gray to reddish gray, poorly consolidated, poorly stratified conglomerate and very coarse-grained, fine pebbly sandstone. The clasts include primarily intermediate volcanic rocks. They are angular and with no sorting in size. Their size varies from fine pebble to cobble, althouhg large blocks up to 150 cm in diameter are also common. The upper half the volcaniclastic unit is finer grained than the lower. The epiclastic basal sandstone unit consists primarily of sandstone, pebbly sandstone and mudstone. The sandstone is greenish gray, poorly to moderately indurated, thin-bedded to massive, fine- to very coarse-grained feldspathic lithic arenite and lithic wacke. Thinly bedded sandstones interbedded with mudshale represent facies C to E. Massive beds contain swirls and stringers of volcaniclastic and/or epiclastic rounded pebbles, up to 8 cm in diameter. From base to top the volcaniclastic clasts replace the epiclastic material in both percentage and size. In the uppermost part of the unit floating blocks of mafic volcanic rocks, up to 250 m on one dimension, occur sporadically. The blocks include limestone, limy mudstone, and mudstone, up to 250 m in length and 75 m in width, and ranging in age from probably late Cretaceous to Paleogene. Locally exposed matrix rocks are typically yellowish gray weathering mudshale with thin interbeds of sandstone (64.80:37.24).

The Ordulu formation rests unconformably on the Karaavu member of the Neyren formation (62.46:36.13,62.90:36.48). The missing older Paleocene and Cretaceous rocks indicate a deep-reaching unconformity (Fig. 10A). The contact between the volcaniclastic bulk of the formation and the lower sandstone unit is conformable (63.10:36.65; 62.45:36.07; and 62.93:36.43). It is abrupt with respect to epiclastic clasts, but gradational with respect to volcaniclastic constituents. At the locality 62.48: 36.07 the basal volcaniclastic conglomerate bed contains mudstone fragments derived from the sandstone unit that may reflect a short hiatus.

The Ordulu formation is barren of fossils. The limestone blocks, lithically similar to the Sermi limestone, contain *Globogerina* sp., *Globorotalia* sp. and *Discorbis* sp., which may suggest a latest Paleocene age for the Ordulu.

Yılgın formation

The name Yılgın formation is here used for a sequence consisting primarily of shale and typically yellowish gray weathering lithic sandstone, which show interlayering at all scales, and minor volcaniclastic sandstone and conglomerate. It corresponds, in parts, to Görmüş's (1982a) Middle Eocene «Alaptura formation» and late Cretaceous «Hızardere formation». The Yılgın is divisible into two units: a lower shale-epiclastic sandstone, and an upper volcaniclastic and epiclastic sandstone-shale unit. The latter can further be divided into four parts, for practical field purposes (Fig. 5A). The complete type section of the lower unit is exposed between 64.86:37.10 and 65.36: 36.85, and that of the upper unit between 66.54:37.30 and 66.80:37.10 (Fig. 10C).

The shale is greenish gray, thinly bedded to massive, yellowish gray weathering mudshale and clayshale. The epiclastic sandstone is yellowish gray weathering, thin to thick-bedded lithic arenite of turbidite facies B to D. The volcaniclastic sandstone is brownish gray, medium to coarse-grained feldispathic lithic arenite with a salt-and-pepper appearance. It locally shows sedimentary structures implying a turbidity origin. The conglomerate is laterally discontinuous, matrix and grain-supported, poorly sorted in size, and is internally unorganized. It has an overall gradation into blocky pebbly mudstone. The clasts range from fine pebble to large block in size. Extragenetic clasts are mafic volcanic and related volcaniclastic rocks, Sermi-type limestone, late Cretaceous limestone and Ordovician sandstone. Intragenetic clasts are foraminiferal sandy limestone and limy mudstone, which are mostly tabular in shape. The matrix rocks include feldispathic and volcaniclastic lithic arenite, sandy and finely pebbly mudstone and lithic wacke. Synsedimentary folding, faulting and distruption in the conglomerate and underlying strata are common. The conglomerate and pebbly mudstone are debris-flow deposits scouring and filling the channels.

The lower epiclastic part of the Yılgın formation overlies abruptly the volcaniclastic conglomerate unit of the Ordulu formation (62.46:35.98, 63.17:36.35, 64.81:37.11; in the same order Figs. 10A, 10B, 10C). Widespread limestone penetrations into fissures of the Ordulu volcaniclastic rocks in the form of neptunian dykes (64.81:37.11) indicate the contact to be an unconformity (Fig. 10C). Because the yellowish gray weathering shale typical of the Yılgın formation occurs as early as in the underlying Ordulu formation, the contact suggests a limestone deposition and subsequent erosion, within a short duration. The epiclastic unit of the Yılgın formation overlies unconformably the Dağköy formation consisting of volcanic rocks (66.35:37.52, 67.36:37.24). The contact between the epiclastic and epiclastic-volcaniclastic units of the Yılgın formation is conformable. It is abrupt with respect to the first occurrence of volcaniclastic sandstone, brownish red mudstone (65.35:36.84, 62.40:35.73, 65.06:35.85) and conglomerate (64.10:35.15) in the upper unit.

The bulk rocks of the Yılgın formation are barren of fossils. The intragenetic pebbles and small blocks of limestone and limy mudstone in the conglomerate (66.62:35.97, 66.70:35.95, 66.80: 36.50, 66.70:36.47) contain *Nummulites* cf. *planulatus* (Lamarck), *N.* cf. *solitarius* de la Harpe, *Discocylina* sp. (gr. *D. archiaci),* which indicate an early Ypresian age. The limestone penetrations into the fissures of the Ordulu rocks contain *Globorotalia* sp. supporting a post-Paleocene age.

Yığılca formation

The name Yıgılca formation is here applied to a thick sequence of typically brownish gray weathering volcaniclastic sandstone, conglomerate tuff, mudrocks and minor basaltic lava. The Yığılca formation corresponds to Görmüş's (1982a) «Melendere formation» of early Eocene age. The formation, with little change in the lithology of the strata, is extensively widespread in and outside the study area. The partial composite type section is compiled between 68.62:33.98 and 68.72:34.50, 67.60:35.23 and 67.33:35.98, and 68.95:36.96 and 68.82:37.16, which represents the middle, upper, and uppermost parts of the formation, respectively (Fig. 11).

The sandstone is primarily a feldispathic lithic arenite of the salt-and-pepper type, containing dark volcaniclastic fragments, feldspar and quartz. It is thin-bedded to massive, and in part, interbedded With mudstone. Many of the very thick bedded to massive sandstones, in the lower part of the Yığılca, contain stringers of fine volcaniclastic- pebbles. Nearly all of the sandstones show fining upward, and in part grade into mudstone. Most peculiarly the sandstone exfoliates with large ellipsoidal or spheroidal cores, sometimes up to a radius of several meters.The light brownish gray weathering mudstone contains interlayers of subfeldspathic arenite. The olive gray to greenish gray weathering shale occurs in the lowermost part of the formation, on the Sermi limestone. Some of the thick mudstone beds contain floating pebbles of limestone of the Sermi-type, and most commonly, intrabasinal mudstone. The conglomerate is poorly indurated, poorly stratified, and matrix and grain-supported. The constituents include almost entirely mafic lavas of a large variety of lithology. However, mudstone, volcaniclastic sandstone and limestone of the Sermi-type are locally abundant. Large detrital calcite crystal and chlorite-muscovite-schist occur locally. Most clasts are subangular to round and up to 16 cm in diameter, although the size varies from fine pebble to large cobble. Blocks of Sermi-type limestone occur locally. Many of the conglomerate beds are widespread channel fills, in some of which synsedimentary deformation is prevalent. They contain mudstone, shale and volcaniclastic sandstone fragments up to 6 m in size, as intragenetic products.

The contact between the Yığılca and the underlying Yılgın formations is conformable. It is abrupt with respect to the disappearance of epiclastic turbiditic (Ta-Tc) sandstone, thick sections of interbedded sandstone and mudstone, and yellowish gray and reddish weathering mudstones of the Yılgın (Fig. 10C). In most places the contact is defined by the lowermost massive volcaniclastic sandstone bed and/or block-bearing conglomerate of the Yığılca formation (62.97:35.10; 66.76:37.12). The lower contact of the Yığılca formation with the underlying Sermi limestone is well exposed at the locality 68.62:33.93. There, it is a sharp mudshale-on-limestone break, and the limestone pebbles and blocks of the Sermi-type first occur about 26 m above the contact. A similar but less well exposed contact is at 61.94:33.03. The contact is considered to be an erosional unconformity, because the detritus of the Sermi limestone occur abundantly in the lower part of the Yığılca, and the Ordulu and Yılgın formations are missing.

No fossil material was obtained from the Yığılca formation, so the age of the beds connot be determined directly. An Ypresian and/or Lutetian age can tentatively be placed on the Yığılca formation by the fact that the Sermi-type limestone pebbles and blocks contain *(Planorbulina cretacea* (Marsson), *Miscellanea* sp. and *Verneuillina* sp., indicating a Middle to late Paleocene age.

AREAL STRUCTURE

The structure defining the distribution of the Cretaceous and Paleogene rock units includes the following major elements, in order of their relative age (Fig. 5a, b):

1. The Yığılca thrust fault extends nearly east-west, and dips northward with an angle of 30° to 40°. It apparently coincides with the axial plane of a southward recumbent syncline, bringing the northern limb onto the southern (i.e., in places it has brought the Yılgın rocks on the Yığılca, the older parts of the Yığılca on the younger parts).

2. High-angle faults have deformed the Paleozoic to Paleogene rocks into numerous blocks, on various scales. The fault planes are either not exposed or poorly exposed, and movement (s) on

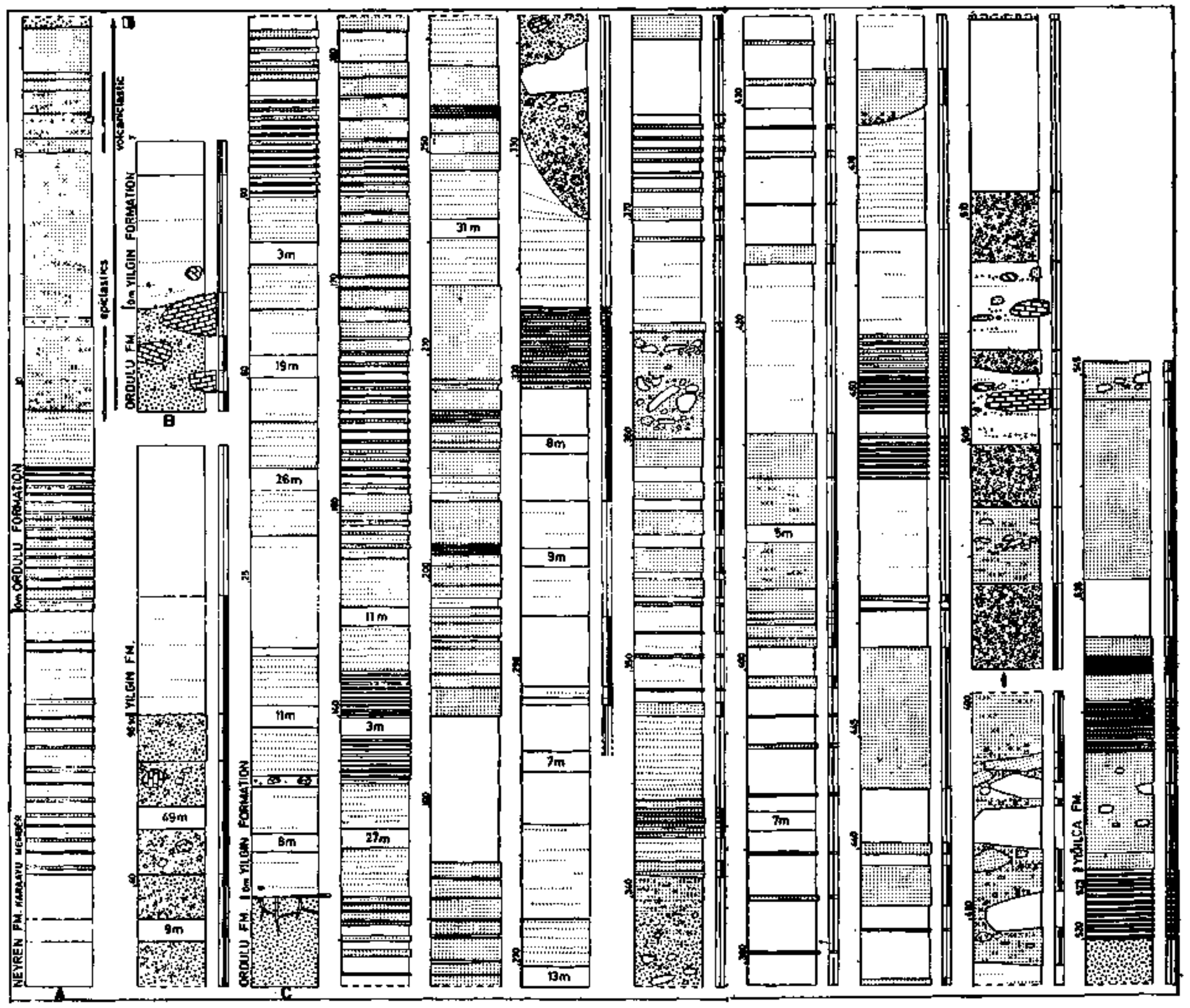

Fig. 10 - A - Complete type section of the Ordular formation; B, C - unconformity between the Yılgın and Ordulu formations, complete composite type section of the Yılgın formation.

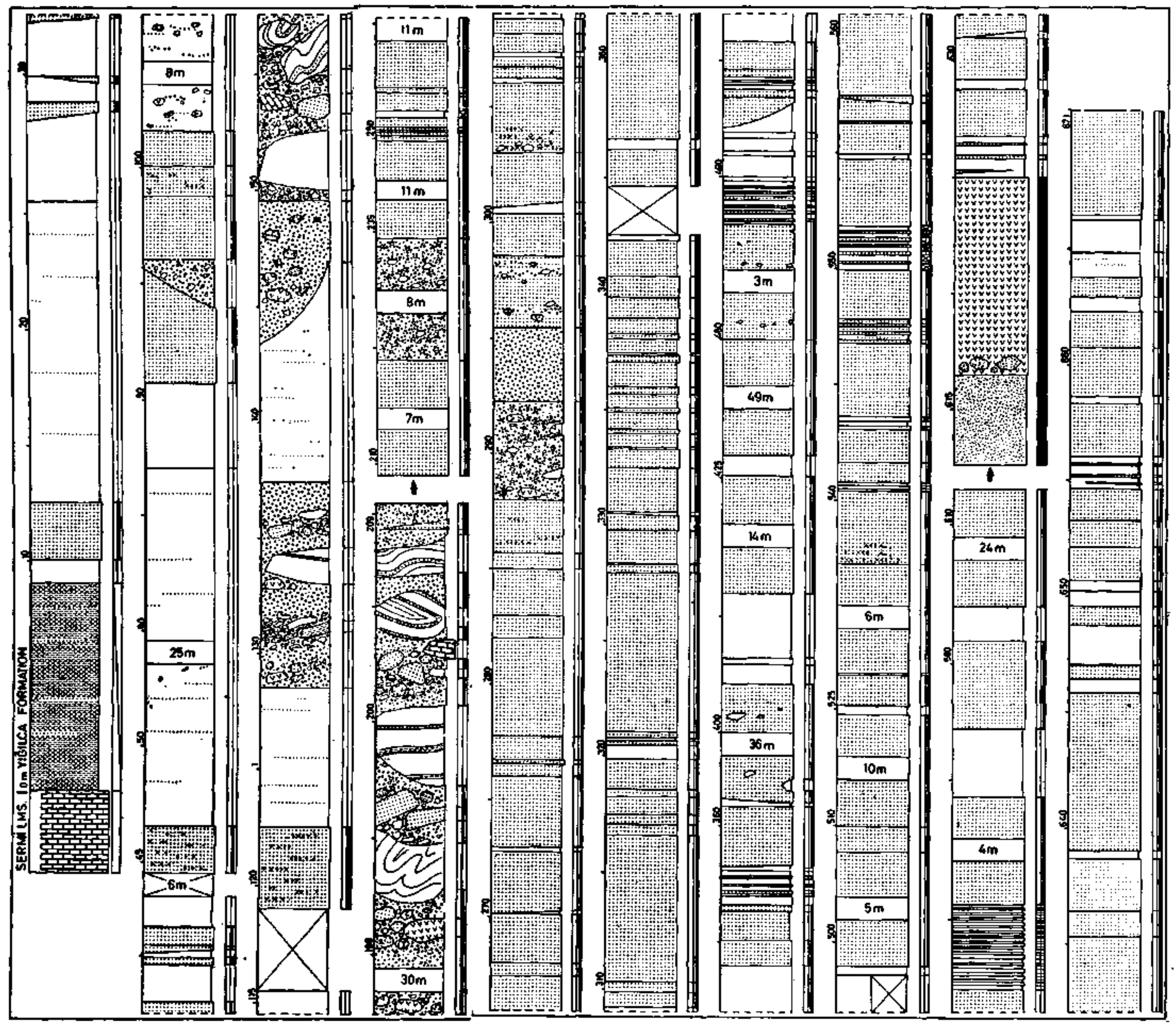

Fig. 11 - Partial composite type section of the Yığılca formation.

them are not recognizable. However, the conjugate vertical sets of the faults are suggested to have been originally shear fractures related to a northeast-trending, horizontal acute bisector, which later acted as extensional. The original shear fractures postdate the Yığılca thrust fault, and both are probably post-Lutetian in age.

INTERPRETATIONS

The comparison of the Cretaceous and Paleogene sequences of the Yığılca area with those of the northerly-lying Ereğli (Kaya and others, 1984c), and the southerly-lying Bolu and Mengen areas (Kaya and Dizer, 1984a, *b)* indicates the following stratigraphic and structural significance of the relevent rock units:

1. The lower part of the Ereğli Cretaceous sequence, the İnaltı limestone to Terziköy member of the Neyren formation in vertical extent, is missing in the Yığılca area. The Neyren formation was deposited by the progressive onlap of an apparently southward transgressive sea.

2. The Yemişliçay group is a primarily volcanogenic sequence with minor epiclastic constituents at the top and base. The ash-fall deposits are very subordinate (totally about 150 m) and lava-flows are absent. The predominating volcaniclastic sandstone and conglomerate imply reworking of loose pyroclastic ejecta of subaerial explosive volcanic eruptions, variably mixed with fine epiclastic debris, under unconfined mass flow depositional mechanisms. Large leaf prints and large-scale planar cross-stratification (Karaavu member), coalified plant fragments in non-turbiditic sandstone (Taşaltı member), symmetrical ripple marks (lower sandstone-conglomerate member of the Sarıkaya formation), and the progressive onlap of the entire group imply a broad shelf environment adjoining a subaerial volcanic apron.

3. The upper part of the Cretaceous sequence in the Ereğli area, the İkse to Örenköy formations in vertical extent, is not represented in the Yığılca area. The pebble to large block-sized clasts derived from this part of the Ereğli Cretaceous occur in the olistostromes of the Yığılca sequence (i.e., Aşağıköy member of the Yeniyer formation and Dibektaş member of the Sarıkorkmaz formation). The totally 200m thick olistostromal interval of the Yığılca sequence has a 4m thick lithic correlative in the Örenköy formation in the Ereğli area (Kaya and others, 1984c, Fig. 7N). Because the olistostromal interval of the Yığılca sequence contains blocks of serpentinite and has a certain time equivalency with a part of the «Ankara melange» of Bailey and McCallien (1954), it may have important structural and stratigraphic bearings on the understanding of the so-called «melange» (Kaya, in prep.).

4. The Maastrichtian Hatipler formation is primarily a rudistid bank, a large-scale cross-stratified lenticular body recurring toward the west. The Hatipler and the correlative carbonate buildups of organic detritus, in the other parts of northwestern and northern Anatolia (e.g. the lower part of the Erikli formation in the Ereğli area), overlie the older rocks with either submarine or subaerial stratigraphic break. In the Yığılca area the massive sandy limestone (56.29:35.25), classed as the basal beds of the Akveren, appears to represent the coeval filling of the areas between the banks.

5. As is the general case throughout nortwhestern Anatolia, the Akveren formation overlies conformably and abruptly the Maastrichtian Hatipler formation and the older Çamlı sandstone, and laps onto the basement. The overall abrupt basal boundary appears to be related to a strong shift of facies.

6. The Sermi limestone appears to be the marginal coralgal buildup corresponding to the upper part of the Akveren onlapping onto a paleomorphological high of the basement rocks, in the southern part of the Yığılca.

7. The Maastrichtian to Eocene sequence of the Yığılca area, extending from the Hatipler to the Yığılca formation, is lithically and sequentially comparable with the Ereğli area rather than the southerly-lying Bolu and Mengen areas. The marked contrast with the Bolu and Mengen sequences may suggest either a structural separation of a previous basin, occurring as late as during the earliest Maastrichtian, or a post-Eocene crustal shortening. The available evidence may support the alternative interpretatipn of a structural separation (Kaya, in prep.).

ACKNOWLEDGEMENT

This report is a part of a research project supported by Turkish Petroleum Corporation (TPAO). E. Akyol (İzmir) has critically reviewed the manuscript. M. Gürle (İzmir) has assisted in the preparation of the illustrations.

Manuscript received, July, 15, 1986

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GEOLOGY AND PETROLOGY OF THE KIZILDAĞ OPHIOLITE (HATAY)

Okan TEKELİ* and Murat ERENDİL*

ABSTRACT. — In the south of Amanos mountains, a continuous ophiolite succession, the Kızıldağ ophiolite, is exposed from tectonite peridotites, through layered and isotropic gabbros to the sheeted dike complex and pillowed volcanics. The tectonites are composed mostly of harzburgites with minor dunite. The plutonic section comprises a layered gabbroic sequence of mainly wehrlite-gabbro alternations (cpx-ol and ol-cpx-pl cumulates) with minor dunite and Iherzolite, planar laminated nonlayered gabbro (chiefly ol-cpx-pl cumulates) section, an isotropic gabbro section of massive noncumulus hb-px-gabbros with plagiogranite bodies. Therefore, the plutonic sequence with minor ultramafic cumulates differs from that of the other Tauric ophiolites. The sequence also shows drastic lateral thickness and lithological differences. The sheeted dike complex is well developed and its internal structure indicates that additional spreading axes were active during the spreading process. The volcanics are composed of pillowed and massive basaltic lava flows. Cumulates show a limited cryptic variation without any considerable cryptic evolution indicating repeated primitive liquid replenishment of the magma chamber. Chemically the tholeiites of the dike and volcanic complexes are transitional between island arc and mid-ocean ridge basalts and are highly depleted in incompatible elements. Thus the Kızıldağ ophiolite is proposed to represent a kind of oceanic lithosphere produced in a slow spreading center with multiple small magma chambers developed over an already depleted mantle.

INTRODUCTION

Ophiolites of the Southeastern Turkey occur along the border folds belt forming the northern boundary of the Arabian plate. They extend along an ophiolitic zone between the Troodos (Cyprus) and the Semail ophiolite (Oman) which, at least at the first glance, poses implications about the existence of a southerly ocean, with respect to the major North Anatolian ophiolite belt, in the Eastern Mediterranean. In the western end of this zone, the second ophiolite after the Troodos is the Kızıldağ ophiolite. It is the most complete and the best preserved ophiolite among the Turkish ophiolites. The Kızıldağ ophiolite nappe has been emplaced upon the thick autochthonous Cambrian to Cretaceous shelf section that characterizes the Arabian Peninsula. Previous regional geologic works (Dubertret, 1953; Aslaner, 1973; Selçuk, 1981) dated the emplacement age of the ophiolite as Campanian-Early Maastrichtian. Delaloye et al. (1979, 1980a, 1980b) carried out several geochemical and geochronological studies and obtained an early Upper Cretaceous age for the formation of the Kızıldağ ophiolite.

Therefore the Kızıldağ ophiolite provides a valuable opportunity to study a part of the Neo-Tethyan oceanic crust and to derive important clues to contribute on the history of the Eastern Mediterranean.

REGIONAL GEOLOGY

In order to provide a reasonable framework for the general geology of the Amanos mountains, the rock units are summarized in three groups: (1) The autochthonous Arabian platform rocks; (2) The ophiolite nappes and (3) The neo-autochthonous cover rocks (Fig. 1).

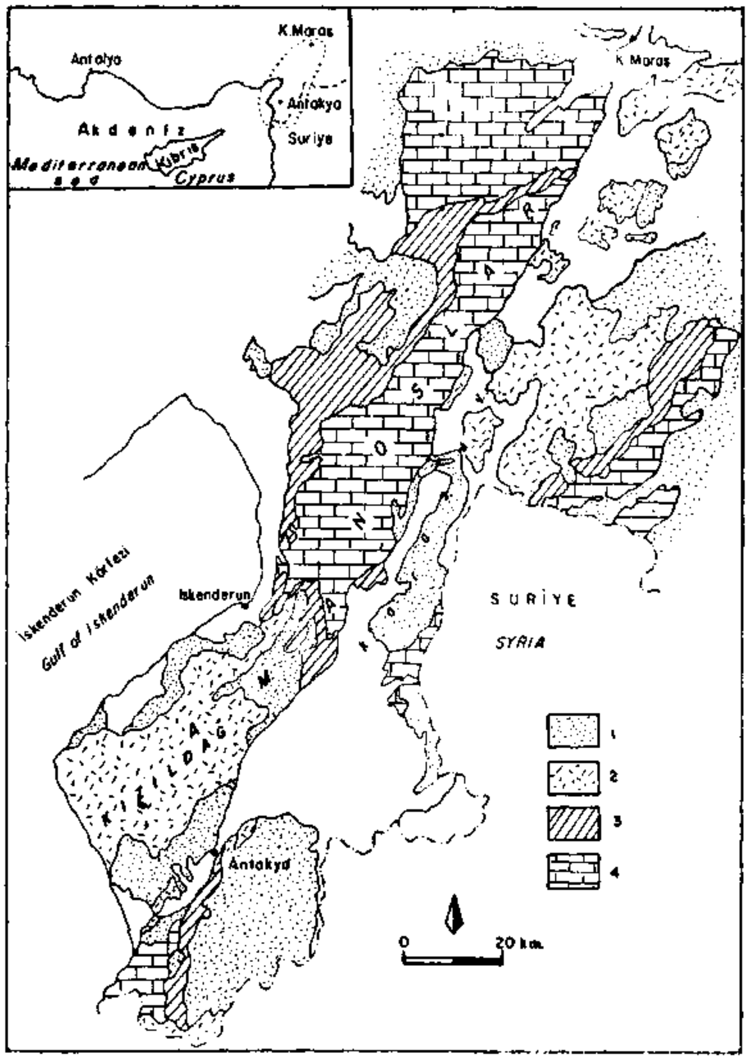

Fig. 1 - Simplified geological map of the Amanos mountains.
1 - Neo-autochthonous cover (U. Maastrichtian-Pliocene); 2 - Ophiolite; 3 - Amanos olistostrome; 4 - Autochthonous Arabian platform sediments (Paleozoic-Mesozoic).

The Arabian platform

The autochthonous Arabian platform rocks of the Amanos mountains represent the north-western edge of the Arabian continent. The rocks of the platform sequence range from the Lowest Cambrian up to Upper Cretaceous.

The Paleozoic sequence. —The Amanos Paleozoic sequence starts with Cambrian and extends up to the Lower Carboniferous (Fig. 2). The majority of the Cambrian is of fine and coarse elastics. They include a dolomite level in the upper parts which is overlain by a section of graywacke-slate alternation. This uppermost section is the unique fossil bearing part of the Cambrian sequence (Middle Cambrian Trilobites: Dean and Krummenacher, 1961). Ordovician sequence starts with a quartzite level at the base which passes into green-brown sandstone-slate alternation in which fossil traces (Crusiana) are common (Dean and Monod, 1985). The uppermost part is of siltstone-mudstone-quartzite alternation with trilobites, brachiopoda, crinoids and tentaculitids. It is conformably overlain by the Silurian sequence of quartzite, conglomerate, siltstone and mudstone alternation rich in brachiopoda fauna. Silurian rocks are overlain by a 20-30 m thick quartzitic key horizon which indicates the beginning of the Devonian-Lower Carboniferous sequence. The rest of the sequence consists of quartzite, limestone, sandy limestone and mudstone alternations with occasional radiolarian chert beds. Any Paleozoic formation younger than Lower Carboniferous has not been differentiated yet. Therefore,the Permian period is accepted as a regional stratigraphic gap.

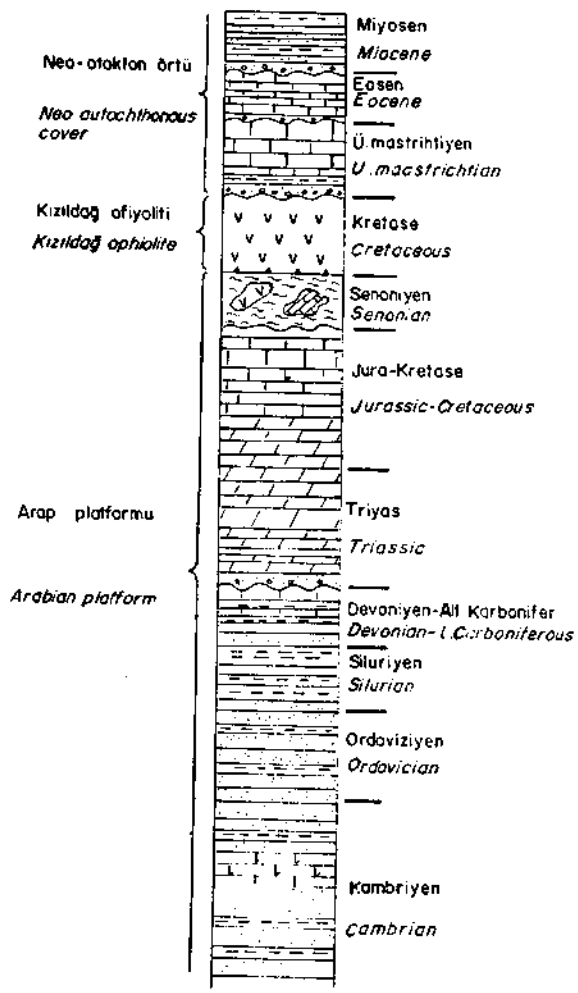

Fig. 2 - Syntethic columnar section of the Amanos mountains (See text for lithological explanations).

The Mesozoic sequence. — Transgressive across the Paleozoic elastics is an approximately 1500 m thick Mesozoic marine carbonate sequence (comprehensive serie: Blumenthal, 1938). In the south, the sequence starts with a quartzite level including a quartzite pebble bearing conglomerate at the base (Anlık quartzite: Atan,-1969). In the northern part of the Amanos mountains, the Mesozoic sequence starts with carbonates directly on the Paleozoic elastics. Although the basal clastic level is fossil-free, the immediately overlying carbonates contain Scythian fossils (A. Işık, unpublished data). The Triassic part consists of dolomites whereas the rest of the sequence is of limestones. Between the dolomites and the Jurassic limestones, frequently seen bauxite levels indicate the transition. The youngest age obtained from the limestone exposures which are surely belonging to the Mesozoic carbonate sequence is Cenomanian-Turonian (Dubertret, 1953). The Mesozoic carbonate sequence as a whole reflects tidally influenced zone of a broad shallow shelf environment (A. Aksay, personel communication, 1984).

The Amanos olistostrome. — The carbonate platform rocks are, in turn, covered by an olistostromal unit (Amanos olistostrome) containing ophiolitic blocks. This unit is exposed at only a few localities in the northern part of the Kızıldağ ophiolite as small exposures through tectonic windows. But along the Amanos range, it crops out over wide areas on both western and eastern flanks.

The matrix of the unit is of mostly sheared serpentinite. At least majority of these serpentinites should be of sedimentary origin since in thin sections most of the samples show clear clastic origin made up of serpentine particles. The serpentinite may locally alternate with or contain carbonate interbeds containing serpentine clasts. In some places serpentine debris flows with rounded pebbles of almost totally serpentinized peridotitic rocks may locally alternate with volcanic or volcanoclastic levels. These levels are found to be discontinuous or boudinized due to the intensive shearing occurred due to the nappe emplacement on top.

The blocks within the matrix show extreme dimensional variety ranging from a few tens centimeters up to kilometers. Majority of the blocks are ophiolitic rocks including mostly harzburgite, dunite and minor gabbro and pillowed lava. Olivine bearing rocks are extensively serpentinized. Apart from the ophiolitic blocks, sedimentary rocks, limestones and sandstones, occur as blocks. Limestone blocks are mostly recrystallized or dolomitic. They may also contain serpentine particles indicating that they have a different origin than the autochthonous limestones. The age of the blocks are scattered in Senonian up to Campanian (Atan, 1969; Aslaner, 1973).

The Amanos olistostrome developed on the carbonates indicates subsidence of the platform during the Senonian. Its development also indicates that the oceanic lithosphere has been uplifted and ophiolitic nappe movements have begun at the beginning of the Senonian epoch. After the development of the olistostrome with a rather regular lithological arrangement with several blocks and serpentinitic-calcerous matrix, emplacement of the huge ophiolitic nappes causes its internal deformation and chaotic structure.

Ophiolite emplacement

The Arabian platform carbonates and the Amanos olistostrome have been subjected to an intensive ophiolitic nappe emplacement during the late Senonian. This overthrusting event is one of the examples of the late Cretaceous ophiolite emplacement onto the stable Arabian platform (Stoneley, 1975) which is characteristic all along the southern ophiolitic zone up to the Semail ophiolite in Oman (Glennie, 1974; Coleman, 1981). In the Amanos region, the Amanos olistostrome with its serpentinitic matrix and ophiolite blocks witnesses beginning of the ophiolite obduction since the early Senonian. The nappe emplacement into the region should be in the late Senonian since the ophiolite nappes are transgressively covered by the Upper Maastrichtian sediments.

The neo-autochthonous cover rocks

The neo-autochthonous sediments, which were deposited after the nappe emplacement has been ceased, starts with Upper Maastrichtian (Selçuk, 1981). These shallow water marine sediments comprise a conglomeratic level at the base, with pebbles derived from all the underlying ophiolite units.

The upper levels are of sandy limestones which grade up into fossiliferous Paleocene rocks (Aslaner, 1973). Open marine sedimentation during the Eocene-early Miocene with several limestone sequences, continues with a flysch type sedimentation during the late Miocene.

THE KIZILDAĞ OPHIOLITE

The ophiolite nappes of the Amanos mountains are seen as dispersed klippes and nappes covering an area of approximately 1300 km^2. The Kızıldağ ophiolite is the biggest nappe to the south ofthe Amanos range. It provides excellent exposures over wide areas especially along the Mediterranean coastal cliffs. The ophiolite comprises all the ophiolitic igneous rocks and exhibits a classical ophiolite stratigraphy which consists of, from bottom to top: (1) Tectonite peridotite, predominantly harzburgite with minor dunite; (2) Layered gabbro, mainly wehrlite-gabbro alternations; (3) Isotropic gabbro, nonlayered and noncumulus cpx-gabbro; (4) Sheeted dike complex that consists of 100% diabase dikes and (5) Volcanic complex, of pillowed and massive basaltic lava flows (Fig. 3).

The Kızıldağ ophiolitic tectonites show only the high-temperature foliation. Therefore, the high-stress deformation related to the ophiolite obduction, which is quite evident in the other ophiolites of this belt (Semail ophiolite, Boudier and Coleman, 1981), is lacking in Kızıldağ. The sub-ophiolitic metamorphic rocks, which are found under most of the other Tauric ophiolites, do not exist in Kızıldağ. Also, the isolated diabase dikes, which are believed to be intruded in some intra-oceanic initial subduction zones (Parrot and Whitechurch, 1978), are not encountered in Kızıldağ ophiolite peridotites.

Tectonite peridotites

Tectonite peridotites of the Kızıldağ ophiolite occur in the highlands constituting the rugged middle parts of the massif (Fig. 4). The peridotite is the dominant rock type of the exposed Kızıldağ ophiolite and makes up about 85% of its surface exposures.

Tectonite peridotites consist of two main rock types, harzburgite (70%) and dunite (30%). Harzburgite consists of olivine (Fo_{90}), orthopyroxene (En_{90}) and spinel. Dunites are serpentinized more than the harzburgite. The tectonites exhibit a low-stress, high-temperature subsolidus deformation and a banded appearance that locally grades into massive zones. The banded tectonites consist of orthopyroxene-rich zones alternating with olivine-rich zones. Orthopyroxene foliation in the banded harzburgites is well developed (Plate I, fig. 1). Within the tectonites, there are irregular dunite zones of varying size (some are kilometric). These dunite bodies irregularly cut the harzburgite foliation. Spinel lineation is generally parallel to the foliation and banding. In the harzburgite, orthopyroxene foliation is parallel to banding, generally striking NE-SW or E-W plunging to the SE (45°). Orthopyroxene-rich bands may show isoclinal folding. In the northern part of Kızıldağ, foliation and banding show complex relations due to the intense neotectonic movements. The sense of shear is such that the upper parts of the grains are displaced northward relative to the lower parts, indicating a southward mantle flow.

Harzburgites exhibit a granoblastic texture. Orthopyroxenes usually include diopside exolution lamellae. Subsolidus deformations such as folding, kinking or ondulose extinction are evident. Linear grain boundaries forming triple junctions suggest a syntectonic recrystallization of orthopyroxenes. Brown spinels are seen as wormy grains surrounding orthophyroxenes. Olivines exhibit syntectonic recrystallization traces beside deformation larqellae and kink bands. Thus, the high temperature deformation of the tectonite peridotite is evident.

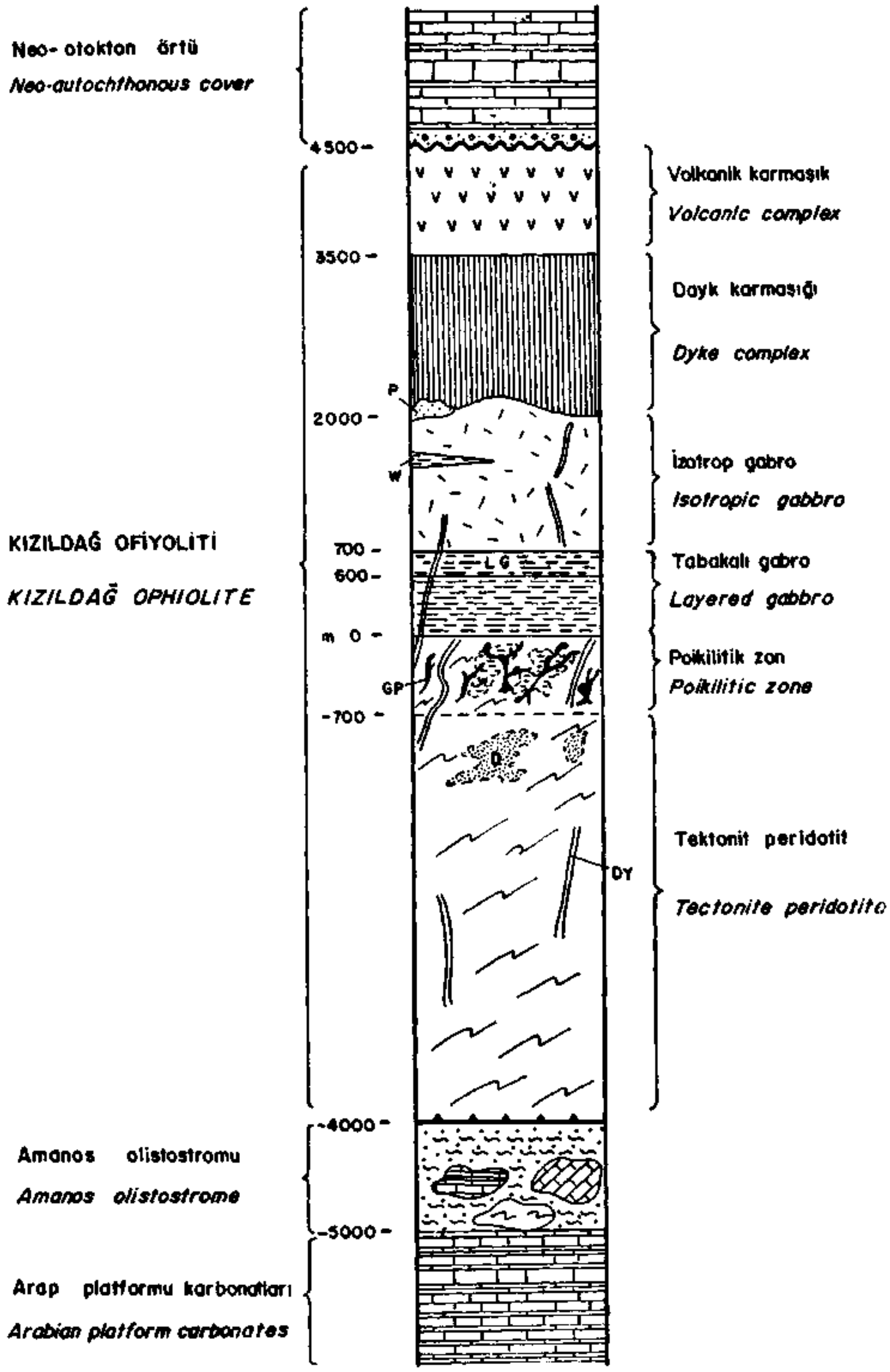

Fig. 3 - Columnar section of the Kızıldağ ophiolite.
LG - Laminated gabbro; P - Plagiogranite; W - Wehrlite; GP - Gabbro pegmatite;
D - Dunite; DY-Diabase dike.

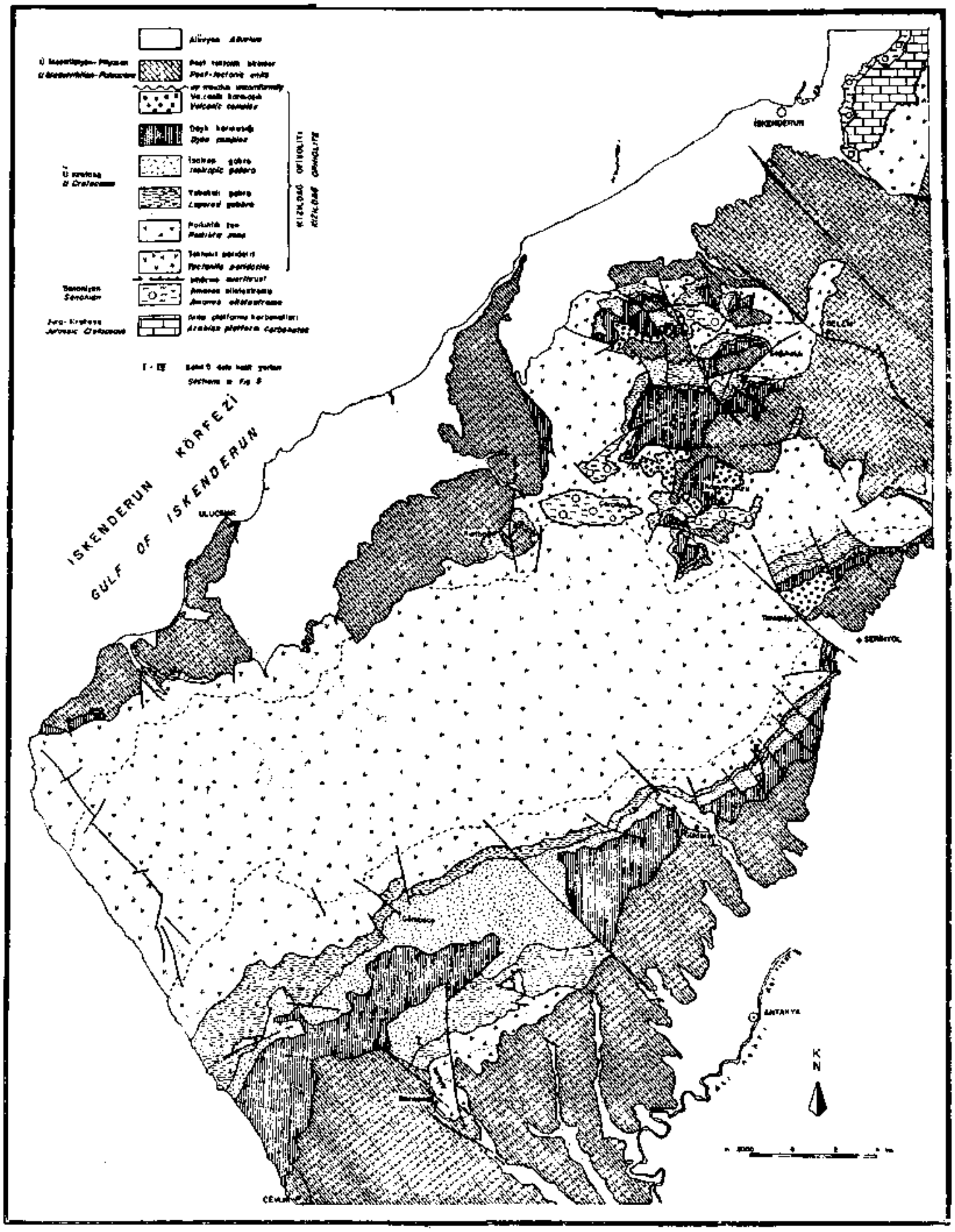

Fig. 4 - Geological map of the Kızıldağ ophiolite.

The tectonites are cut by gabbro dikes. They are usually a few cm thick (rarely 20-30 cm) with straight and sharp contacts cutting the foliation at various angles. Rarely encountered pyroxene segregations consist of concentration of orthopyroxenes (up to 2 cm) surrounded by a spinel-rich dunite cover. In the tectonite, gabbroic zones are found as irregular veins and diffuse segregations. These are generally surrounded by a dunitic cover.

Chromite in the tectonites usually occur as chramitite or as disseminated within the dunites. About 80% of the chromite bearing rocks is in the tectonites, most of which occurring in the upper levels, and the rest is in the poikilitic zone.

Poikilitic zone

The uppermost part of the peridotite tectonites forming the transition to the overlying plutonic suite is called the poikilitic zone (Fig. 4). Poikilitic texture of the rocks of this zone is striking. Lower limit of this zone is obscured but the upper contact is quite sharp. Thickness of this zone varies between 100 and 600 m depending on the overlying plutonic section thickness. This unit is widely exposed along the southeastern flanks of Kızıldağ.

This zone includes the cumulates of poikilitic texture in association with the tectonite harzburgite and dunite. It consists approximately 60% tectonite and 40% cumulate rocks and exhibits a «poikilitic» appearance with cumulate patches enclosed by the peridotite tectonites. Harzburgites may show pyroxene-rich and poor banding. Orthopyroxene foliation is not as clear as it is in the lower levels and therefore they exhibit a rather massive appearance. Zones of poikilitic texture occur as irregular bodies with diffuse boundaries mostly cutting the harzburgite banding (Plate I, fig. 2). They may also occur as strips parallel to the bands. The poikilitic rocks within the harzburgite are Iherzolite, wehrlite and websterite containing little plagioclase whereas within the dunites they are mostly wehrlite and minor melatroctolite. The other characteristic of this zone is the abundant in situ gabbro pegmatite dikes. The density of these dikes increases upwards. In the field, a genetic relation between the poikilitic cumulate pockets and the gabbro pegmatite dikes is clearly visible.

The harzburgites of this zone contain 1-2 % clinopyroxene and plagioclase occurring interstial between or around the olivine and orthopyroxene crystals. Olivine and orthopyroxene boundaries with clinopyroxene are corroded. Orthopyroxenes occur as single crystals or multigrained porphyroclasts which are intensely altered. They show deformation traces such as bending or poligonization. The dunites of this zone are of two types. The first type exhibit a typical tectonite fabric with olivine possessing kink bands. The second type includes interstial clinopyroxene and plagioclase with cumulate texture. Some of the Iherzolites develop due to the increase of clinopyroxene and plagioclase content of the harzburgite. This type of Iherzolites show cumulate and tectonite texture in association. The other type of Iherzolite is of pure cumulate origin, being olivine and some clinopyroxene representing the cumulus phase while plagioclase-clinopyroxene is the postcumulus phase. The typical rocks of this zone are cumulate wehrlite and websterite in which the texture is poikilitic. Within this zone, chromitites form discrete lenses and bodies.

Plutonic sequence

The Kızıldağ ophiolite plutonic sequence is well exposed especially along the eastern flanks of the massif. The sequence forms two regionally mappable units: layered gabbro and isotropic gabbro. In the upper parts of the layered gabbro, in fact, rocks do not show layering but lamination. This part is considered as a seperate unit to examine. Therefore the plutonic sequence comprises three zones of different features.

Layered gabbro. — The plutonic sequence of the Kızıldağ ophiolite show considerable lateral thickness and facies variation. In the southern part, for example along the Karaçay valley (Fig. 4), the 1800 m thick unit consists of 600 m thick layered gabbro. But in the northern part it is 100-200 m thick along the northeastern flanks and it continuously thins towards northwest where it is about 100 m thick (Fig. 5). Thickness variation is accompanied by lithological changes. The unit gains a striking homogeneity and loses its layered nature while it thins. Megascopic features such as ratio layering, size layering, graded bedding and slump structures result in a sedimentary appearance of the layered gabbros (Plate I, fig. 3-4). In the southern part, the unit consists mainly of gabbro and wehrlite alternation besides the interbeds of norite, gabbronorite and ultramafic cumulates such as dunite, Iherzolite, ol-websterite and websterite. Ultramafic cumulates are restricted to the basal part of the sequence. The layered sequence is characterized by well developed milimetric to metric layering. Layering results either from gradation ofthe cumulus phases (ol, cpx, pl) or from alternation of isomodal levels which are in phase and ratio contacts with each other (Plate I, fig. 4).

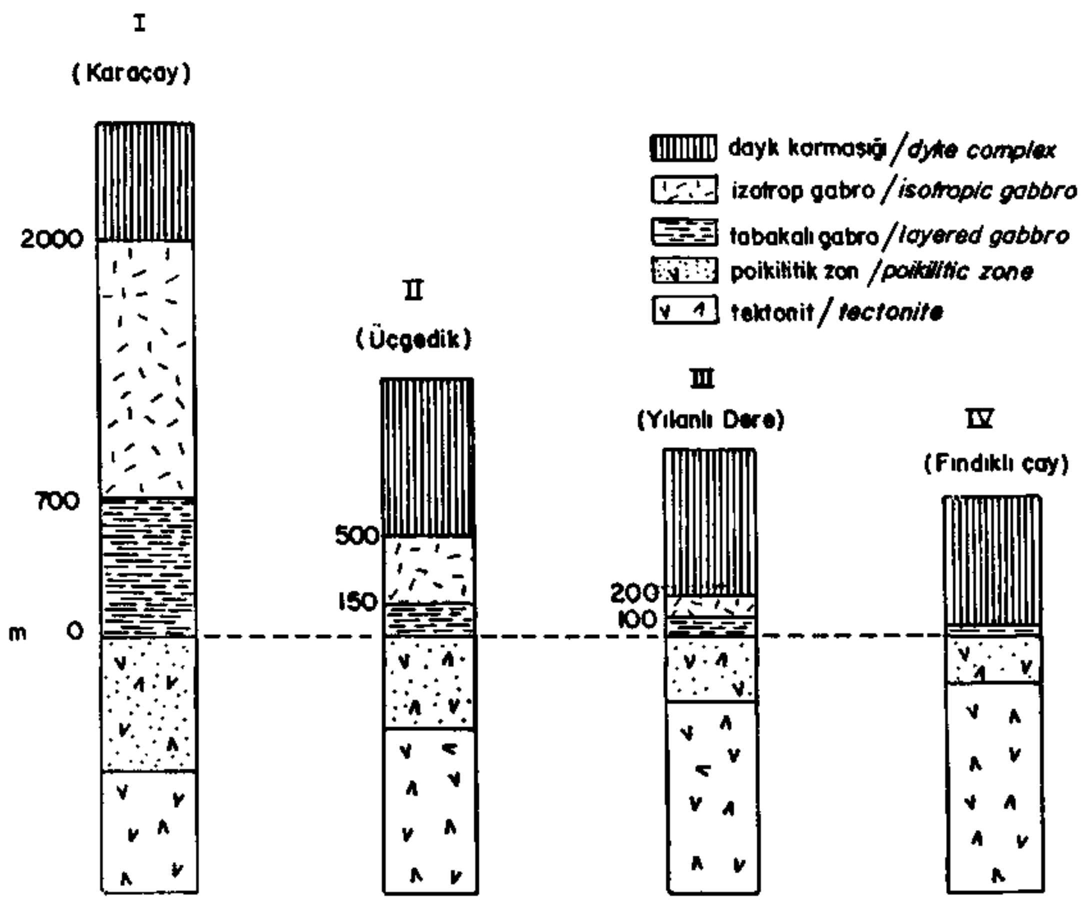

Fig. 5 - Sections showing the local differences of the plutonic sequence (Section numbers are indicated on the map, Fig. 4).

Ol-cpx-gabbro is the most abundant rock type. Cpx-gabbro and opx-cpx-gabbro occur in lesser amounts. Texture is adcumulus in the gabbros and is heteradcumulus in the olivine rich rocks. Planar plagioclase and clinopyroxene show generally well developed magmatic lineation and sometimes size grading.

Wehrlite occurs as 5-25 m thick massive interbeds along the base or as cm-dm thick layers alternating with gabbros. Thick wehrlite zones, in turn, comprise ol-and cpx-rich zones including some plagioclasev Texture is either poikilitic or granular. Ol±cpx is the cumulus, cpx+pl is the post-cumulus phase being clinopyroxene is poikilitic and plagioclase is intersertal.

Dunite occurs generally in the lower levels as irregular bodies in the thick wehrlite zones or as interbeds alternating with gabbros. Texture is ol-adcumulus and poikilitic clinopyroxene may also be present. Therefore olivine constitutes the early cumulus phase.

Lherzolite, containing little plagioclase, occurs as rare interbeds associated with gabbros. Oliopx is the cumulus whereas cpx+opx+pl is the postcumulus phase. Cpx-opx is poikilitic and plagioclase is intersertal. Texture is ol-adcumulus.

Troctolite is a rare rock type in the layered gabbro. It contains little poikilitic clinopyroxene. Texture is ol-heteradcumulus.

Along the northeastern flanks, the plutonic sequence loses its thickness considerably (Fig. 5). Layering is poorly developed and the rocks are more or less massive. Dunites and wehrlites constitute almost all the sequence. Lherzolites are rare associates. These rocks are in irregular and diffuse contacts with each other. In situ gabbronorite veins crosscut these rocks in anastomosing pattern.

In the northern and northwestern part of the massif, the layered gabbro section is locally as thin as 50 m. Layering is poorly developed. Poikilitic Iherzolite, granular wehrlite and poikilitic wehrlite-ol-gabbro alternation constitute the major lithologies. These are cut by in situ norite veins. In the Fındıklı Çay valley (section IV in Fig. 5), the sheeted dike complex merges directly into the layered gabbro section. This section consists of brown hornblende and plagioclase bearing massive Iherzolites.

Laminated gabbro. — Layered gabbros pass upwards into a zone in which the apparent layering of the lower levels of the plutonic sequence gradually dissapears. This zone is characterized by a pronounced lamination generally parallel to the underlying layering, but sometimes irregular. Lamination is a consequence of parallel arrangement of plagioclase and clinopyroxene. This laminated section develops where the gabbroic section is thick. Along the Karaçay valley, the laminated gabbro is 100 m thick and provides a transition from the layered part to the overlying isotropic gabbro. The rocks are ol-and cpx-gabbros with cpx+pl-mesocumulus and ol-cpx-pl-adcumulus textures. The unit continuously thins towards north and terminates.

Isotropic gabbro. — The uppermost part of the Kızıldağ ophiolite plutonic section is composed of gabbros showing neither layering nor lamination. This part reflects the higher levels of the magma chamber, solidified from top to bottom.

The isotropic gabbro features an extreeme grain size variation from microcrystalline to pegmatitic. Texture is hypidiomorphic and minerals show well developed compositional zoning. Therefore, a noncumulus crystallization is apparent.

In the Kara9ay valley, the isotropic gabbro reaches its maximum thickness (1100 m) and forms 2/3 of the plutonic sequence. Majority of the unit is of hb-cpx-gabbro, including hb-gabbro in the upper levels. The unit as a whole, has been subjected to an intense hydrothermal alteration. Clinopyroxene and brown hornblende are replaced by fibrous white-green actinolitic hornblende and plagioclase is saussuritized. Intensity of this alteration decreases from dike-gabbro contact downwards. Plagiogranite bodies concentrated along the dike complex-gabbro contact may include several isotropic hb-gabbro and diabase xenoliths. Some plagiogranite chonoliths intruding the base of the dike complex, with xenoliths, are also cut by later diabase dikes. Dikes derived from the underlying plagiogranite are locally abundant in the sheeted dike complex and they are crosscut by other diabase dikes. These complex intrusive relations between these rocks indicate that the roof of the chamber was in a dynamic, unstable state throughout the crystallization history of the plutonic sequence. In situ plagiogranite bodies and dikes in the lower levels of the isotropic gabbro define the sandwich horizon (Wager and

Brown, 1967). They represent the latest fractionation liquids. Quartz, feldspar, hornblende and magnetite are the main constituents. Locally, although rarely, the isotropic gabbro includes wehrlite chonoliths or sills of varying size (Plate I, fig. 5).

Along the Karaçay valley where the plutonic sequence is complete and well exposed, several samples have been collected in order to study the mineral chemistry of the constituent rocks. Sampling transect includes the gabbroic rocks of the poikilitic zone and extends up to the sheeted complex. Microprobe analyses of olivine, plagioclase, clinopyroxene and orthopyroxene from the Kara9ay samples were done using an automatized electron microprobe CAMEBAX at Nancy 1 University (accelerating voltage: 15 kv, speciment current: 20 mA, counting time: 6 sec, HENOC correction programe, natural silicate standards).

In this section of the plutonic sequence, high MgO contents of olivine and pyroxenes are characteristic (Table 1). Olivine compositions show gradual decrease in Mg content up to the layered gabbro base and limited variation in the layered part (Fo 85 - 90)- In the isotropic gabbro Fo_{75} is the mean value. Olivine and the other mineral compositions are stratigraphically depicted in Fig 6. In these profiles, mineral chemistry show only a limited cryptic variation. The limited range of this variation indicates continuous primitive magma supply during the spreading process. Olivine, pyroxene and plagioclase variation profiles exhibit the same pattern and thus indicate that these three minerals were crystallized in the same sector of the magma chamber. This feature, which was previously reported in the other ophiolites, is interpretted that gravitational crystal settling has not been occurred considerably (Pallister and Hopson, 1981).

The sheeted dike complex

The Kızıldağ ophiolite includes a well developed dike complex composed of subparallel meta-diabase dikes. The complex is a mappable unit with a consistent stratigraphic position. Along the dikes no wall rock is present so that the complex is of 100% dike rocks (Plate I, fig. 6). The upper contact of the dike complex is exposed in only a few localities in the northern part of the massif where the volcanics are discontinuously preserved. The diabase dikes cutting the volcanics increase in number towards the dike-volcanic contact to form small irregular dike complexes within the volcanics. This 30-40 m thick irregular zone provides transition from volcanics to dikes. Isotropic gabbro generally underlies the dike complex. This contact is also gradual but rather rapid with respect to the upper contact. But although the passage from dikes to gabbro occurs along a smoother contact in between, the contact, in regional sense, is quite irregular. In addition to local thickness variations of the isotropic gabbro unit, the isotropic gabbro as well as an important part of the layered gabbro may be regionally absent so that several plutons along the isotropic gabbro unit can be considered.

Dikes are chilled against the cooler host dikes. These chilled margins of the dikes are darker colored and less susceptible to weathering than their middle parts. The dikes cut each other in complex relations. But two main dike generations can be roughly distinguished simplifying this complexity (Erendil, 1984).

The dikes are fine to medium grained, sometimes phyric but mostly aphyric and nonvesicular meta-diabases. Plagioclase, clinopyroxene (augite) and magnetite are the main constituents being the second group of dikes are richer in pjagioclase content. The complex is metamorphosed hydrothermally. Greenschist facies is characteristic', but upper parts of the complex are in zeolite facies. Metamorphic facies boundaries show local irregularities (Fig. 7) depending on the intensity of the hydrothermal activities (Erendil, 1984).

Table 1 - Selected microprobe analyses of the Karaçay samples

	Olivines					*Clinopyroxenes*			
	KC8	*KC35*	*KC38*	*KC44*		*KC8*	*KC35*	*KC38*	*KC44*
SiO_2	39.41	40.66	39.67	40.27		53.05	52.53	52.62	52.53
TiO_2	0.04	0.01	0.01	0.00		0.10	0.20	0.24	0.12
Al_2O_3	0.00	0.02	0.03	0.00		1.01	2.37	2.67	2.22
FeO	20.46	11.86	14.64	14.84		4.51	3.31	3.87	4.71
MnO	0.14	0.11	0.23	0.30		0.36	0.03	0.17	0.03
MgO	40.14	47.49	44.30	44.66		16.91	17.98	17.39	17.34
CaO	0.18	0.09	0.15	0.15		23.16	21.60	21.95	22.99
Na_2O	0.06	0.00	0.03	0.01		0.27	0.08	0.32	0.09
K_2O	0.00	0.00	0.00	0.00		0.02	0.00	0.00	0.00
Cr_2O_3	0.00	0.00	0.00	0.00		0.11	0.99	0.76	0.26
NiO	0.14	0.21	0.00	0.16		0.00	0.16	0.00	0.00
Total	100.57	100.44	99.061	100.39		100.38	99.25	99.99	100.29
Fo %	77.76	87.71	84.36	84.28	En%	46.82	50.81	49.17	47.46
Fa %	22.24	12.29	15.64	15.72	Wo%	46.17	43.95	44.69	45.31
					Fs%	7.01	5.25	6.14	7.23

	Orthopyroxenes					*Plagioclases*			
	KC8	*KC24*	*KC31*	*KC48*		*KC8*	*KC24*	*KC31*	*KC44*
SiO_2	53.65	55.63	55.18	55.77		44.71	44.79	44.49	45.70
TiO_2	0.11	0.05	0.05	0.11		0.00	0.01	0.00	0.00
Al_2O_3	1.21	1.37	1.84	1.37		35.16	34.02	35.49	34.63
FeO	11.76	11.06	8.12	10.02		0.40	0.52	0.26	0.40
MnO	0.42	0.23	0.00	0.00		0.06	0.00	0.00	0.00
MgO	29.57	29.35	31.08	30.17		0.04	0.08	0.05	0.08
CaO	1.83	2.03	2.10	1.67		18.77	19.52	20.53	19.44
Na_2O	0.04	0.01	0.02	0.02		1.02	0.88	0.56	0.99
K_2O	0.00	0.02	0.00	0.00		0.05	0.00	0.00	0.00
Cr_2O_3	0.08	0.24	0.49	0.00		0.00	0.00	0.00	0.00
NiO	0.00	0.06	0.00	0.02		0.09	0.00	0.13	0.00
Total	98.67	100.05	98.88	99.15		100.30	99.82	101.51	101.24
En%	78.88	79.29	83.66	81.55	Ab%	8.91	7.53	4.70	8.43
Wo%	3.56	3.95	4.07	3.25	Or%	0.29	0.00	0.00	0.00
Fs%	17.60	16.77	12.27	15.20	An%	90.80	92.47	95.30	91.57

The general orientation of the complex appears to be EW. Chill margin statistics show that there are local anomalous one-way chilling percentages in the both north and south directions (Fig. 7). These local zones are interpreted as subcomplexes indicating secondary intrusion axes (Erendil, 1984). The overall one-way chilling percentage is to south but considerably low (3.8%). Southward chilling preference is in favour ofa northerly spreading axis with respect to the present position ofthe Kızıldağ ophiolite within the limits of this low percentage.

The rocks of the sheeted dike complex are tholeiitic basalts very poor in trace elements. They show a transitional chemistry between the mid-ocean ridge basalts and island arc tholeiites (Erendil, 1984). MORB normalized trace element and REE profiles indicate that the diabases have been originated from a highly depleted source. Low Ti and Zr values are indicative of a low spreading rate (Fig. 8). This low spreading rate deduced from chemical properties is compatible with the low crustal thickness measured in the massif (Rheid and Jackson, 1981).

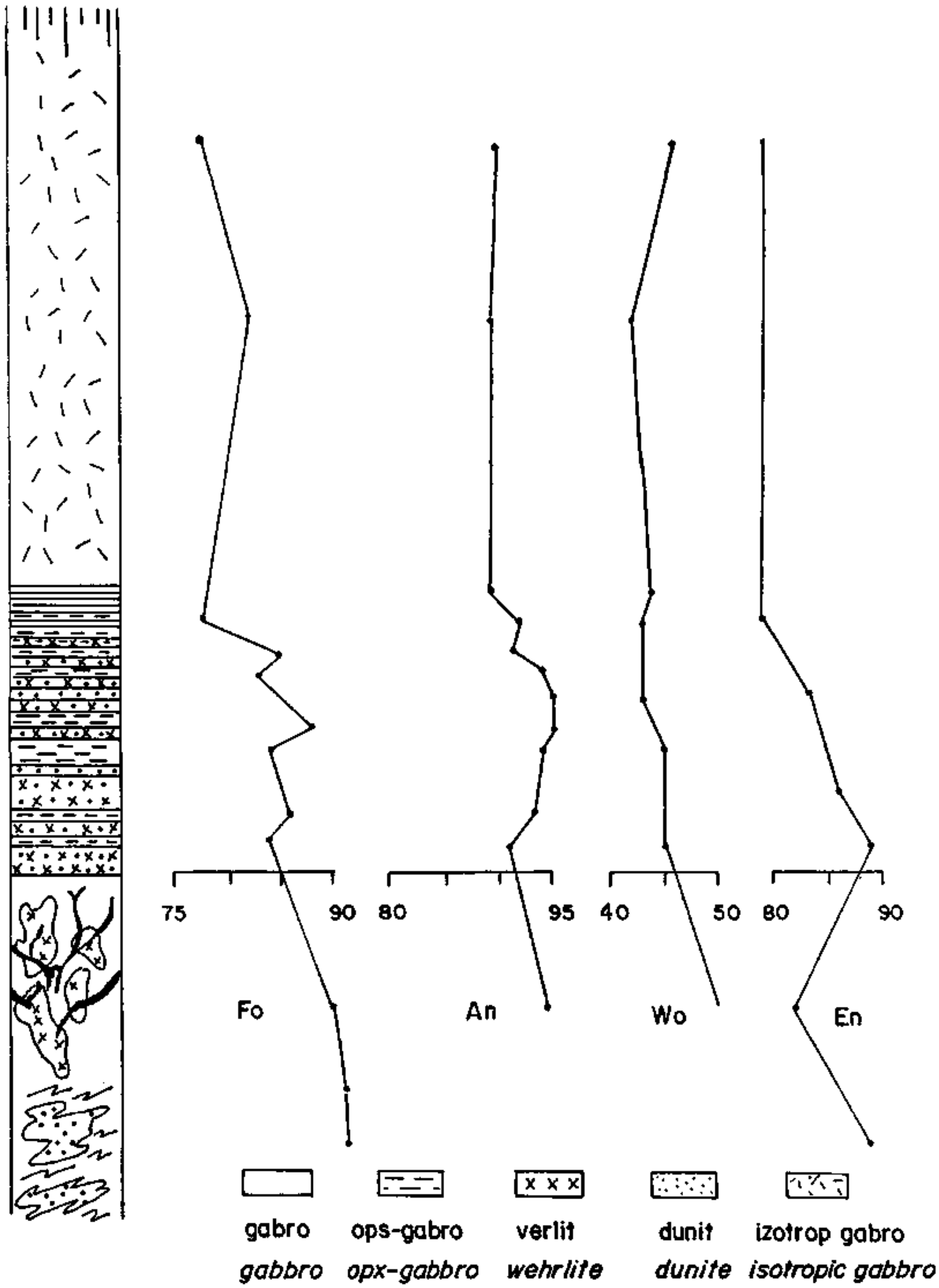

Fig. 6 - Mineral composition variations along-the Karaçay section.

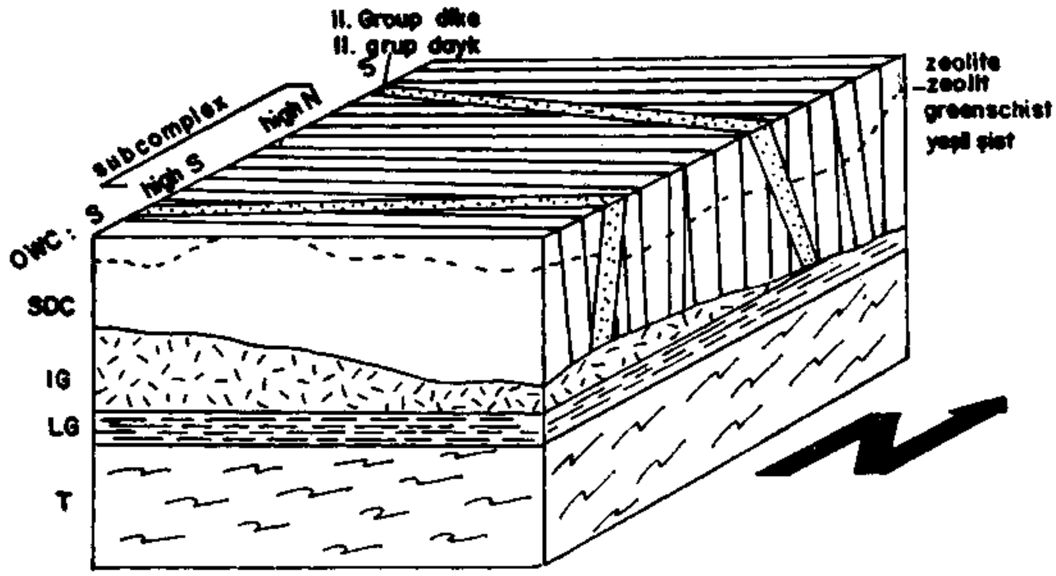

Fig. 7 - Block diagram illustrating internal characteristics of the dike complex.
OWC - One-way chilling percentage; SDC - Sheeted dike complex; IG - Isotropic gabbro; LG - Layered gabbro; T - Tectonites.

The volcanic complex

The volcanic complex of the Kızıldağ ophiolite is poorly preserved and exposed only in the northern part around Kömürçukuru village (Fig. 4). It comprises pillowed and massive lava flows with interflow and intrapillow sediments. Volcanic rocks are basalts with plagioclase, clinopyroxene (augite) and glass as their main constituents. They are hydrothermally metamorphosed in zeolite facieswith increasing temperatures from top to bottom (Erendil, 1984).

The volcanic sequence does not contain any carbonate interbeds. Instead, partly or fully manganese enriched red-brown chert beds. They occur as discrete lenses concentrated at two levels. Manganese occurs as oxide or hydroxide phases mainly as manganite or pyrolusite. Apart from the manganese, sulfides are found as concentrated in certain zones oblique or perpendicular to the volcanic layering. In these zones, forming the former conduits of the hydrothermal circulations, pyrite, chalcopyrite and malachite mineralizations are apparent.

The chemical properties of the volcanic complex is quite similar to those of the sheeted dike complex. They are transitional between island arc tholeiites and the mid-ocean ridge basalts. The complex as a whole is poor in trace and REE so that the volcanics also have a depleted magma source. Low concentrations of Ti, Zr and Y confirm the previously stated slow spreading nature of the paleospreading center (Fig. 8).

DISCUSSION AND CONCLUSIONS

The Kızıldağ ophiolite exhibits a full, except the deep sea sediments usually overlying the volcanics, ophiolite stratigraphy. Of these units, the tectonite peridotites represent the uppermost mantle during the Upper Cretaceous and records a history of mantle deformation under subsolidus conditions associated with the spreading axis environment. High-stress deformation related to the obduction mechanism or to any transform fault activity is lacking. Any subophiolitic metamorphic rock along the basal thrust contact related to the oceanic tectonics prior to emplacement upon the Arabian continental margin, do not exist in Kızıldağ. This feature, together with the lacking of isolated diabase dikes, distinguishes the Kızıldağ ophiolite from the other Tauric ophiolites. Assuming that the high-temperature deformation corresponds to asthenospheric flow related to the spreading ridge, foliation and banding as well as the sense of shear in deformed minerals in the tectonites correspond a southward mantle flow. If the asthenopsheric flow occurs perpendicular to the spreading ridge (Boudier and Coleman, 1981), the ridge axis should be oriented roughly east-west. This configuration is supported by the spreading geometry deduced from the sheeted dike complex.

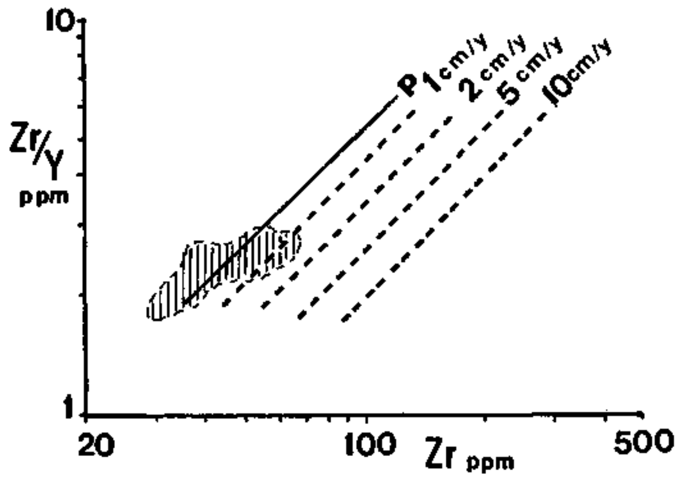

Fig. 8 - Plots of the dike and volcanic complex samples on the trace element-spreading rate diagram (Pearce, 1980).

The plutonic sequence show spatial lithological variations. The layered gabbro section is composed mainly of gabbro-wehrlite alternation with minor ultramafic cumulates at the base. Layering is well developed in the sections where the plutonic section is thick. Thin plutonic sections exhibit massive and monotonous lithological features. The isotropic gabbro section, overlying a transitional laminated zone, represents crystallization that occurred at the top of the chamber and more differentiated than the underlying cumulates. Plagiogranite is volumetrically insignificant. It occurs along the dike complex-gabbro contact and along the sandwich horizon. Mineral chemistry of the plutonic sequence show that mineral compositions vary in a limited range with a slight cryptic variation but without a significant cryptic evolution. This feature and the stratigraphy of the cumulates necessitate multiple melt injections into the magma chamber.

The sheeted dike complex formed in a 100% extensional environment and it consists of olivine-free meta-diabase dikes. Internal structure of the complex is more complex than an idealized seafloor spreading product. There are small complexes defined by the occurrence of symmetrical one-way chilling percentages. Therefore the spreading axis should be either moving during the process of spreading or accompanied by several injection axes. The low one-way chilling percentage can be a natural consequence of a spreading with multiple or moving injection axes in a true oceanic environment. But this feature may well indicate a marginal basin spreading with diffuse dike injection axes. The diabase is characteristically poor in trace and REE. Their parental magma should originate from an already depleted source.

The volcanic complex consists of pillowed and massive basaltic lava flows, now pervasively altered by low-temperature hydrothermal activity. The volcanic sequence includes only manganiferous deep sea chert and mudstones, but not carbonates, concentrated at two certain levels. Therefore the eruption should have occurred as continuous phenomenon with two passive episodes. Also, the lack of any breccia and pyroclastic zones indicates that the eruption was not vigorous. Chemical properties, being similar to those of the sheeted dike complex, indicate a transitional affinity between arc tholeiites and mid-ocean ridge basalts, with a depleted source.

The above petrochemical features of the volcanic and sheeted dike complexes can be interpreted to propose two possible oceanic spreading environments: (1) An ocean-floor spreading occurred within an already formed oceanic crust over an already depleted mantle as a new and weak ridge formation; or (2) A spreading occurred within an environment over a subduction zone.

ACKNOWLEDGEMENTS

This study was carried out in the Tauric Ophiolites Project organized by the Fundamental Research Department of the Mineral Research and Exploration General Directorate (MTA) during the period of 1979-1982. We are particularly indebted to Hubert Whitechurch and Thierry Juteau both for valuable discussion and help in the field and for analyses of the samples at Nancy 1 University.

Manuscript received December 9, 1985

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GEOLOGY OF THE BASKİL (ELAZIĞ) AREA AND THE PETROLOGY OF BASKİL MAGMATICS

H. Jerf ASUTAY*

ABSTRACT. — The study area which covers the region around Baskil on Eastern Taurus Range comprises of Keban metamorphics and Baskil magmatics overlain by a Tertiary sedimentary cover. The Keban metamorphics are represented by regional and contact metamorphic rocks in the study area. Calc schist and marble associations are widespread on regional scale. Between Baskil granite and Keban metamorphics exomorphism and endomorphism zones have been developed. Metasomatic effects are observed in the contact metamorphic rocks which reflect the pyroxene-hornfels facies. The sedimentary sequence begins with Middle Paleocene (Thanetian) aged rocks in the study area. The same sequence, however, has been deposited starting in Santonian-Campanian in the surrounding area. The sedimentary rock sequence which is composed of Kuşçular conglomerate, Seske formation, Kırkgeçit formation (Paleocene-Plio-Quaternary) are represented by conglomerate, carbonates and flysch kind of sedimentary rocks. Baskil magmatics are an association of plutonic, hypabyssal and volcanic rocks. Of this association, Baskil granite contains dioritic, monzonitic and tonalitic kind of magmatic rocks which are mostly observed as transitional. Baskil granite, in the study area, is frequently cut across by basic and acidic dykes which locally intrudes between the granite and the basaltic, andesitic rocks overlying the granite and are transitional with the volcanics. Chemically, Baskil granite is of calc-alkaline type. It is rich in silica and alkaline. Trace element distribution is quite regular. Baskil granite which is determined as of type 'I' is generally rich in hornblende but poor in muscovite and biotite. It shows the features of continental margin magmatism and is an example of systematic differentiation. Considering their features and under the light of plate tectonics concept, Baskil magmatics may be said to be a product of continental margin magmatism. They are, presumably, the products of an oceanic lithosphere existing between Keban microplate and Arabian platform which later on subducted under Keban microplate.

INTRODUCTION

The study area lies in the west of Eastern Anatolian region. It is surrounded by Keban in the north, Malatya-Elazığ highway in the east and Euphrates River in the west and south. The purpose of the study is to reveal the geology around Baskil and the petrological characteristics of the magmatic rocks in the region. For this purpose, three sheets of 1:25 000 scale maps have been completed in the region (Fig. la, 1b) and the petrographic descriptions and the chemical analyses of the collected samples have been done.

STRATIGRAPHY

In the study area which contains mainly metamorphic, magmatic and sedimentary rocks, the stratigraphic sequence is as follows, from bottom to top: (1) Keban metamorphics; (2) Baskil magmatics; (3) Kuşçular conglomerate; (4) Seske formation; (5) Kırkgeçit formation.

Keban metamorphics

Keban metamorphics are mainly composed of regional and contact metamorphic rocks. Contact metamorphic rock associations are observed in the localities where Keban metamorphics are in contact with plutonic and semi-plutonic rocks of Baskil magmatics. Kipman (1976) divides them into three groups, namely lower schists, Keban marble and upper schist which crop out extensively out of the study area, around Keban and in the surrounding area. Depending on his fossil findings, Glomospira, Ammodiscus, Hemigordius, he proposes the age of the deposition of the metamorphics as Permo-Carboniferous. On the other hand, Özgül (1976, 1981) states that the age of the Keban metamorphics is Permo-Triassic after studying around Munzur mountains and in the surrounding area.

These metamorphics, exhibiting their rough topography and dark colors in the study area, have been tectonized especially during Miocene and have been thrust onto the younger formations.

Under the microscope, in lower and upper schist thin sections, mainly calcite, chlorite, sericite, quartz and locally K-feldspar minerals have been observed. In calcite crystals pressure twinnings and elongation along schistosity are seen. Chlorite and sericite are generally lepidoblastic and show kink band structures. Quartz minerals have also been elongated along schistosity and their wavy extinctions are clearly observable. K-feldspars are locally observed as porphyroblasts.

For these rocks, as protoliths, carbonaceous sandstones may be assumed. As paragenesis, they are in quartz-albite-chlorite subfacies of low degree metamorphism (greenschist) (Winkler, 1974).

Baskil magmatics

This unit which covers the largest part of the study area is represented mainly by plutonic, hypabassal and volcanic rocks. Baskil magmatics, in frame of Eastern Anatolia, has been called Yüksekova formation or Elazığ complex by different researchers (Perinçek, 1979a, 1979b; Naz, 1979; Tuna, 1979; Perinçek and Özkaya, 1981; Bingöl, 1982, 1984; Hempton and Savcı, 1982; Hempton, 1984). During the investigation, a systematic magmatic sequence has been observed rather than a complex, therefore, «Baskil magmatics» which is named first by Yazgan and Asutay, 1981, is preferred. Later, during another investigation (Asutay, 1985), magmatic rocks have been named as Baskil magmatics and the plutonic equivalents of them have been treated as Baskil granite.

Baskil granite has weakly been altered and this is very clear in the hand specimens. This granitic series, in which medium and coarse grained rocks are observed, has been frequently cut by joint systems in NNE and NNW directions. Granitic rocks, especially around Baskil, have been surrounded by dark colored semi-hypabyssal and volcanic rocks. In the study area and in the close vicinity granitic rocks cut Keban metamorphics (Asutay, 1985; Asutay and Turhan, 1986) and contact metamorphism zone is observed in between the rocks. Baskil batholith is an example of shallow-emplaced granite and reflects the epizonal characteristics which have been defined by Read (1957) and Buddington (1958). The most frequent hypabyssal rock in Baskil granite is diabase which is very clear with its dark color especially where they cut white tonalites. Acidic hypabyssals have been emplaced later in the granite and they cut diabases.

All the inclusions in the granitic rocks have cropped out as a kind of granite. This kind of formation is a proof of differentiation in the batholith.

One of the most important features of the Baskil granite is the widespread occurrence of hornblende as melanocratic mineral. Biotite is a rare mineral contrarily to hornblende which is observed almost in every kind of granitic rocks.

Kuşçular conglomerate

Kuşçular conglomerate is a wholly conglomeratic unit which is observed as narrow outcrops in the study area. This unit has been defined as Kuşçular formation since it is very thick around Keban upon the studies made by E.İ.E. Department (1972). The same unit has been named Medik formation by Hakyemez and Örçen (1982) in NW of Malatya. Kuşçular conglomerate overlies Baskil magmatics in the study area. This unit consists mainly of wine colored conglomerates with carbonaceous matrix. It is generally devoid of fossils. The color is due to the iron content of the matrix. Balçık et al. (1978) states that the unit contains low percent manganese, too. The bad sorted and low graded pebbles within the conglomerate wholly belongs to Keban metamorphics. The age of the unit is Middle Paleocene (Asutay, 1985).

Seske formation

This unit which has been named and described by Erdoğan (1975) is quite widespread around Elazığ. It is wholly limestone and is Middle Paleocene (Thanetian) aged. This unit is vertically transitive to Kuşçular conglomerate and is generally medium to thick bedded, light gray and yellowish in color. In the upper levels, there are karstic cavities. Microfossils are abundant in this formation. Under microscope it appears as biomicrite, containing fossils and shell fragments in a micritic matrix. Many specimens have been collected in and around the study area. E. Sirel has determined the following fossils: *Kathina* cf. *selveri* Smout, *Operculina* cf. *heberti* Munier-Chalmas, *Daviesina* sp., *Discocyclina* sp., *Ranikothalia* sp., *Miscellanea* sp., *Rotalia* sp., *Planorbulina* sp., *Kathina* cf. *subsphaerica* Sirel, *Globorotalia* sp., Algae-Bryozoa. According to the above fossil association, the age of the Seske formation, in and around the study area is Middle Paleocene (Thanetian).

Kırkgeçit formation

This formation which is extensively widespread in Eastern Anatolian region is represented mainly by conglomerates, carbonate rocks and flysch in the study area. The distribution and the lower contact relations are locally different. According to Turan (1984), it starts with basal conglomerates but in the study area the formation starts with carbonate rocks which disconformably overlies the Seske formation. On the other hand, the same formation, around Keban transgressively overlies the Seske formation with conglomerates (Asutay and Turan, 1986). Kırkgeçit formation, in the study area, contains a member with olistoliths which overlies the basal carbonate rocks. This member is observed clearly around Marik village and is named after that (Asutay and Turan, 1986). Marik member together with the clasts of Baskil magmatics, contains the blocks of Keban metamorphics. After the Marik member, Kırkgeçit formation gains a typical flysch appearance.

The time interval in which Kırkgeçit formation is observed in Eastern Anatolian region is Middle Eocene (Lutetian)-Oligocene (Perinçek, 1979a). Oligocene materials are mostly in limestone facies and are seen in the upper levels. They are, generally not thick, around Keban maximum 30 m thickness is observed. In the study area, the oligocene deposits are represented by thin to medium bedded, white-grayish sandy carbonate rocks which overlies the Baskil magmatics. For this formation, around Şelil Mountain and Karameşe hill, NW of the study area, samples have been collected and the following fossils have been determined from the lowermost parts of the unit: *Europertia magna* (Le Calvez), *Nummulites* sp., *Discocyclina* sp., *Sphaerogypsina* sp.
Also, the following fossils have been determined from the samples collected from the upper parts of the unit: *Pellatispira* sp., *Heterostegina* sp., *Nummulites* sp., *Discocyclina* sp. *(D. discus* group). According to the above fossils, Lutetian for the lower units, and Upper Lutetian-Upper Eocene for the upper units have been assumed by E. Sirel.

The upper deposits of the Kırkgeçit formation, in the study area, are observed overlying the Baskil magmatics. The fossils collected from these deposits are dated Upper-Middle Oligocene. The following fossils have been determined from the Oligocene unit in the sandy limestone facies: *Nummulites fichteli, Eukpidina* sp., *Amphistegina* sp., *Nephrolepidina* sp.

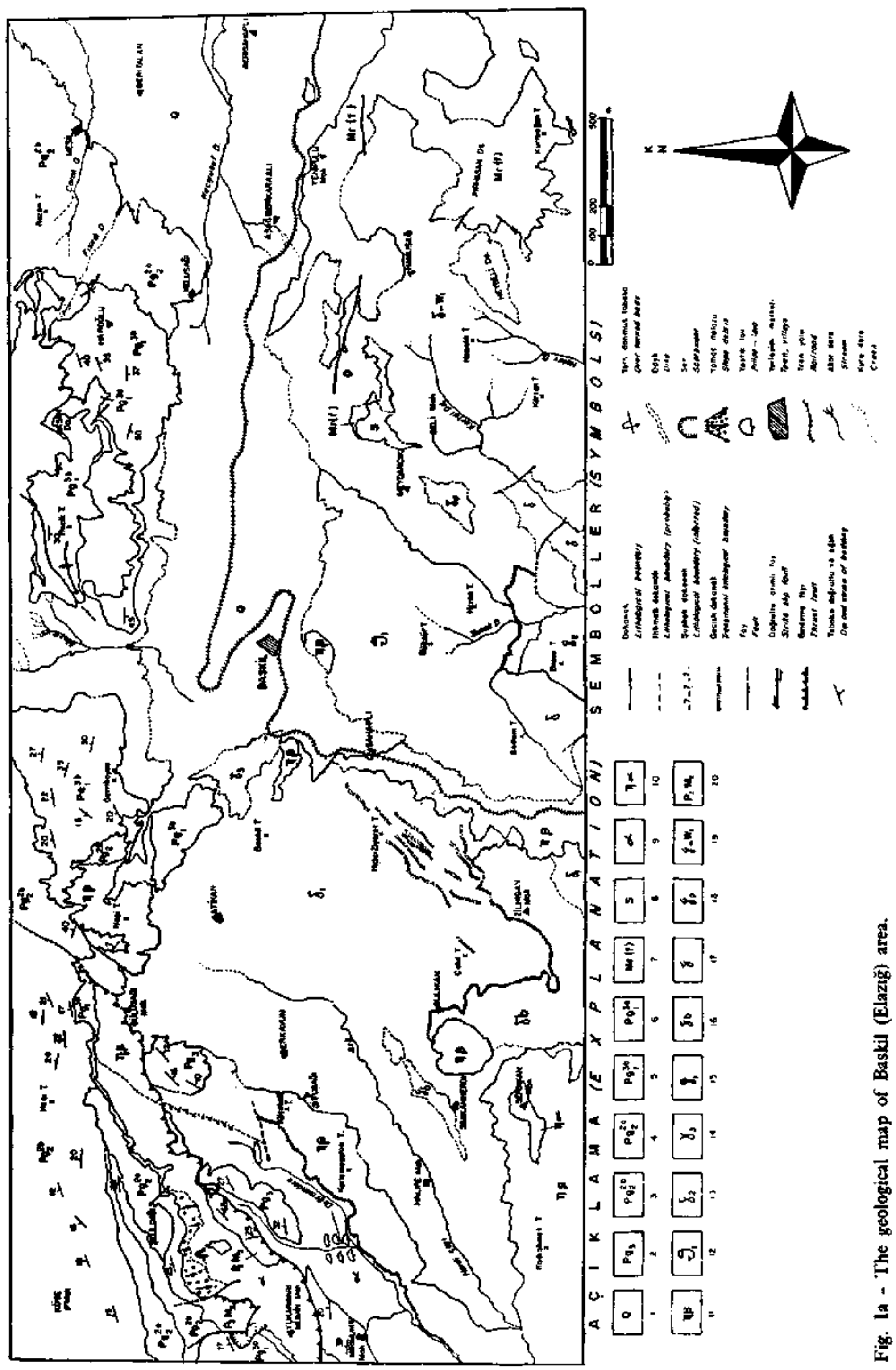

Fig. 1a - The geological map of Baskil (Elazığ) area.
1 - Quaternary, 2,3,4 - Kırkgeçit formation; 5 - Seske formation; 6 - Kuşçular conglomerate; 7 - Marble, contact metamorphic; 8 - Skarn; 9 - Volcanic rocks; 10 - Asidik hypabassal rock; 11 - Basic hypabassal rocks; 12 - Quartz monzonite; 13 - Quartz monzonite; 14 - Granodiorite; 15 - Tonalite; 16 - Transitional rocks; 17 - Pegmatitic monzodiorite; 18 - Quartz diorite; 19 - Diorite, monzodiorite; 20 - Keban metamorphics.

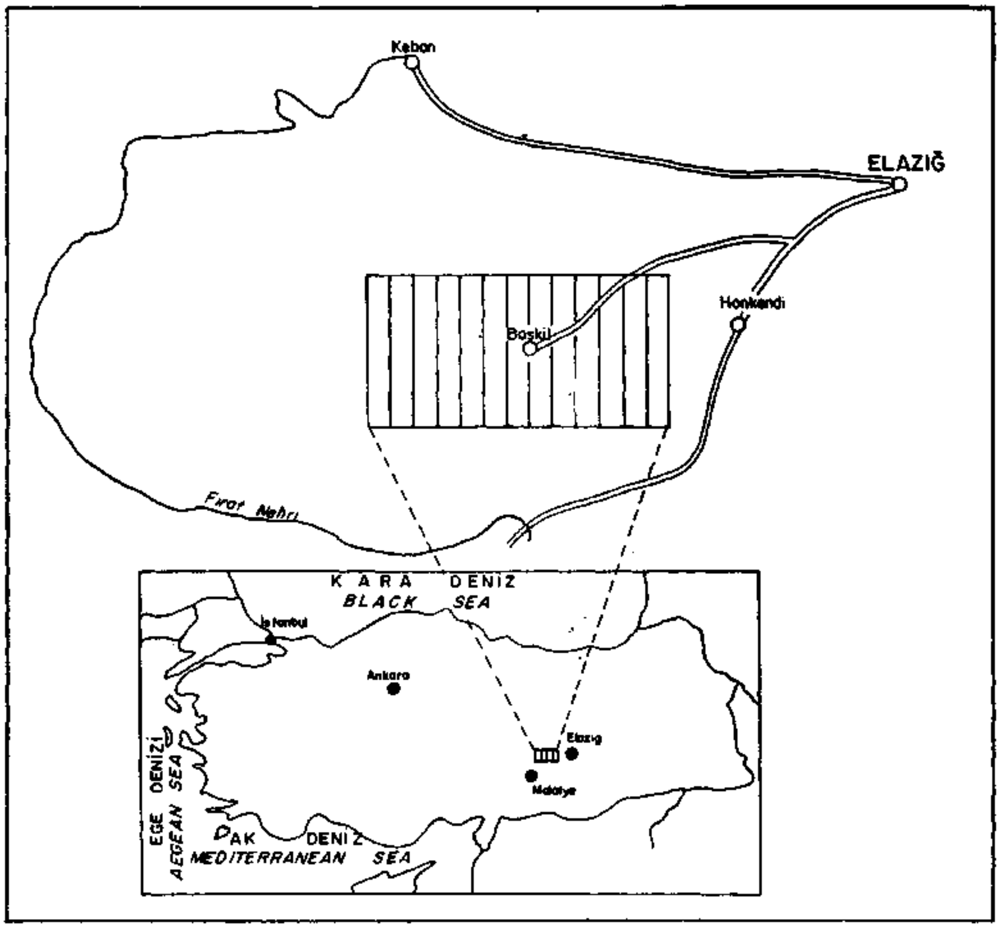

Fig. 1b - Location map.

PETROGRAPHY OF BASKİL MAGMATICS

Baskil magmatics are mainly composed of plutonic, hypabyssal and volcanic rocks. The plutonics, Baskil granite, are divided into four according to some certain characteristics. These are, from older to younger in the emplacement order, are as follows: (1) Diorite-monzonite group; (2) Transition rocks; (3) Tonalite-granodiorite group; (4) Monzonite group (Fig. 2).
In the close vicinity of the study area, Baskil magmatics start with gabbros and end up with syenites (Asutay and Turan, 1986). This implies that magmatism has started rich in plagioclase and has enriched in quartz and ended up with an increase in K-feldspars.

Baskil granite

7. *Diorite-monzonite group.—* This is the first group emplaced in the study area. It is subdivided into two groups:

a. Monzodiorite-quartz monzodiorite: When a monzodiorite from the study area is studied by naked eye, it is seen that the rock is dark gray and includes some mineral associations having different grain sizes. These minerals are plagioclases, 2-3 mm in length and whitish, K-feldspars, 1-2 mm in length and pink, and dark green-black amphiboles dominant all over the rock. Quartz is less abundant than the other minerals, 2-3 mm in length and glassy in appearance.

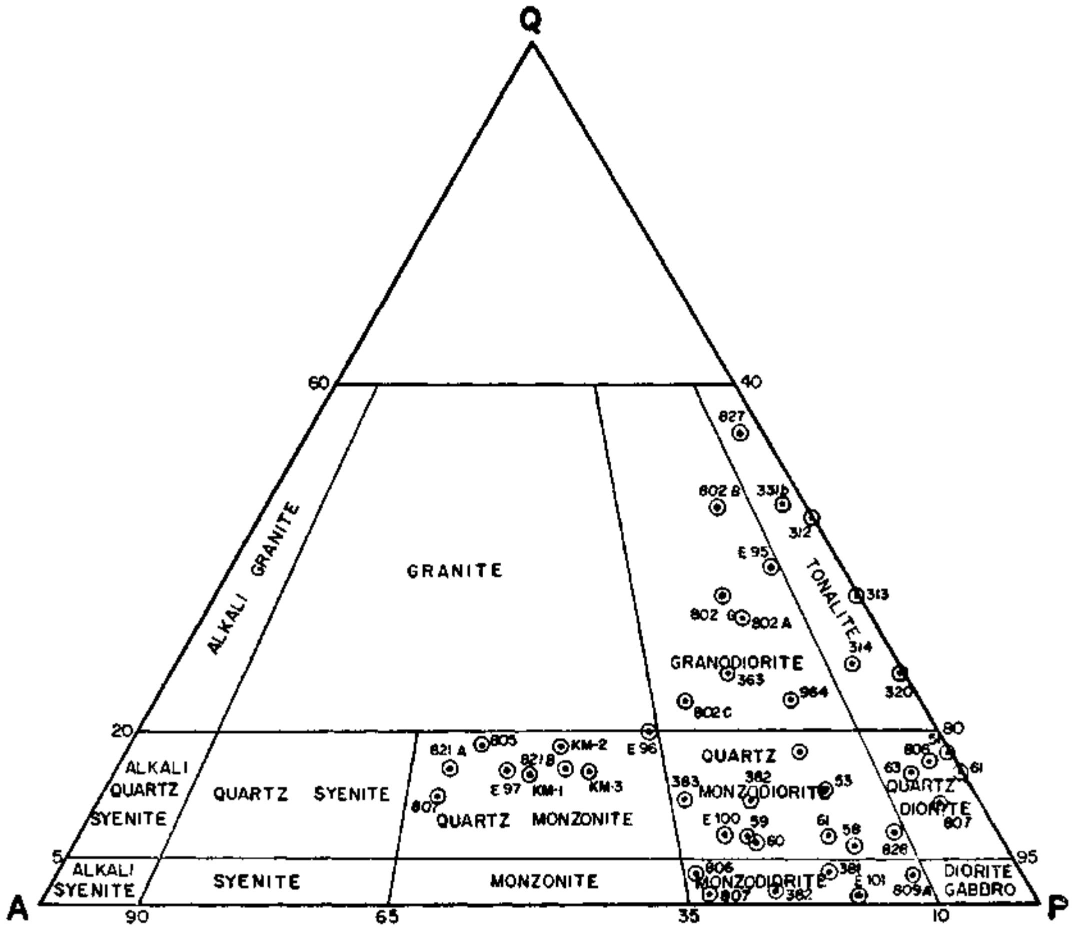

Fig. 2 - Distribution of Baskil granitic rocks in Streckeisen (1976) diagram.

Under the microscope, a hypidiomorph granular texture is observed. The results of the modal analyses of the rock are as follows:

	%
Quartz	1.9
Plagioclase	49.5
K-feldspar	5.8
Amphibole	40.3
Other minerals	2.5
Total	100

Quartz content in the rock is always changing, therefore, it is difficult to differentiate monzodiorite and quartz-monzodiorite during mapping. In the quartz-monzodiorites, the increasing quartz rate is clearly visible in the result of the modal analysis:

	%
Quartz	4.2
Plagioclase	37.8
K-feldspar	5.7
Amphibole	45.0
Other minerals	7.3
Total	100

b. Quartz diorite: They are located in the south of the study area and are easily distinguished from the other rocks by their darker colors. When their hand specimens are studied amphiboles, 1-2 mm in length and dark green in color, white-beige plagioclases and less abundant quartz minerals are seen. Amphiboles, which give their color to the rock, are the minerals with the highest percentage. The results of the modal analyses are as follows, averagely:

	%
Quartz	7.6
Plagioclase	40.6
K-feldispar	2.9
Amphibole	47.0
Other minerals	1.9
Total	100

2. Transition rocks. — This group of rocks are in between diorite-monzodiorite and tonalite-granodiorite groups. The best outcrops can be seen south of study area and locally in tonalites like small islands. The quartz-rich rocks of Baskil granite starts with this unit. Their appearance also is different than the other group of rocks. They are generally gray and dark gray in the study area but locally include less dark parts which are the result of differentiation in magma chamber.

When the hand specimens or the outcrops of the transition rocks are studied, two main different parts are differentiated. First part, as in quartz-diorites, is composed of plagioclase needles and mafic minerals in between them and light colored or transparent quartz. Second part is composed of elliptic or round quartz crystals whose radii or long axes are approximately 1 cm (Fig. 3). No apparent orientation is observed in elliptic quartz crystals and they are randomly distributed. On the outcrops quartz minerals surrounding the plagioclases gather together and appear as white spots on tonalites which are situated on top of transition rocks.

These small islands locally get denser and surrond the melanocratic parts and therefore make them appear as «autoliths». This formation is especially observed in a zone passing south of Zilhigan quarter.

The results of the modal analyses are different. This is simply because of the coarse quartz grains. In the thin sections which are taken from the quartz rich parts quartz content is very high. If the thin sections which are not rich in quartz were studied, it would have been seen that the rocks would fall into the quartz-diorite area in Streckeisen (1976) diagram. The results of such thin sections are as follows:

	%
Quartz	8.2
Plagioclase	42.5
K-feldspar	1.0
Amphibole	40.0
Other minerals	8.3
Total	100

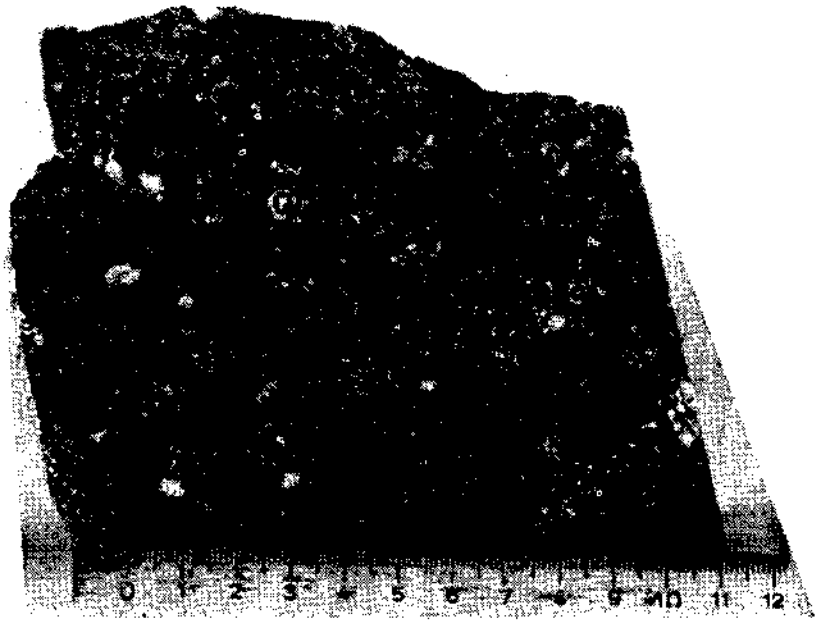

Fig. 3 - Quartz minerals within transitional rock.

It is deduced from these results that the transition rocks are mainly composed of quartz -diorites. When the coarse quartz crystals are abundant the rate changes and the rock falls into tonalite group.

3, *Tonalite-granodioritegroup.* — This group is divided into two groups:

a. Tonalite: Tonalites are the most extensively widespread unit in Baskil plutonic rocks. Both their appearance in the field and their features under microscope are quite different than the other group of rocks. The round and elliptical quartz grains in the rock provide a different appearance. The striped appearance of light colored tonalites when they were cut by diabase dykes is another characteristic of the rock.

Of the plutonic rocks in the study area, tonalites are the rocks in which leucocratic minerals are highly dense. This unit is transitive to granodiorites with a local increase in the rate of K-feldspars. In hand specimens and in the outcrops they are grayish-white or whitish in color. Coarse, mostly round and elliptical quartz grains are striking with their glassy appearance. Sericitized and clayed parts indicate plagioclases. Between all of these leucocratic minerals, dark colored epidotesand chlorites are the secondary minerals. Tonalites are richer than the other rocks in having secondary minerals since they have frequently been cut by basic dykes. Also, amphibole may be observed as a melanocratic mineral. In some specimens amphibole is very scarce, therefore they are very light in color.

Tonalites which are observed in hypidiomorphic texture locally have porphyritic features. The results of the modal analyses are as follows:

	%
Quartz	48
Plagioclase	39
K-feldspar	1
Amphibole (chl. + epi.)	12
Total	100

b. Granodiorite: Tonalites, with increasing K-feldspar rate pass into granodiorites. In the study area granodiorites cover a limited area with respect to the other units. They are quite similar to tonalites in appearance. If hand specimens are carefully studied according to the following features of granodiorites, they can be distinguished from tonalites: No porphyritic texture, bright biotite crystals, existence of pink K-feldspars. The rates of-mineral distribution in the rock are as follows:

	%
Quartz	33.7
Plagioclase	49.0
K-Feldspar	12.7
Biotite	2.8
Amphibole	1.0
Other minerals	0.8
Total	100

4. Monzonite group. — This group which forms the last members of Baskil plutonic rocks is mainly divided into two sections:

a - Quartz-monzodiorite

b - Quartz-monzonite

The quartz-monzodiorites which are included in monzonite group are different than previously mentioned quartz-monzodiorites. These differences, in turn, are:

— The rocks in monzonite group have been formed after the other group of rocks. This has been verified by field observations.

— The color, texture and the structural features of the rocks of this group are quite different.

— The K-feldspar content and the rate of these rocks are higher than Baskil plutonic rocks and they show idiomorphic crystals. As a result, the quartz-monzodiorites in diorite-monzodiorite group and the quartz-monzodiorites in monzonite group characterize the different emplacement phases of Baskil granite.

a. Quartz-monzodiorite: They are observed in north of the study area. The following result is obtained from the modal analysis:

	%
Quartz	3.0
Plagioclase	40.0
K-feldspar	15.0
Amphibole	35.0
Other minerals	7.0
Total	100

b. Quartz-monzonite: It is the last member of Baskil plutonic rocks in the study area. It has clear contacts with the other rock groups. Its pinkish color is a distinguishing feature than the other rocks. It is the richest rock in K-feldspar in the study area. The coarse idiomorphic K-feldspar minerals are easily observed in hand specimens. The distribution of minerals is as follow:

	%
Quartz	15.0
Plagioclase	38.0
K-feldspar	40.0
Amphibole	4.0
Other minerals	3.0
Total	100

Hypabyssal rocks

One of the characteristic features of Baskil magmatics is the abundance of hypabyssal rocks. They are observed in various forms and compositions of basic and acidic rocks. Except for their mineral paragenesis and texture, another characteristic feature is their appearance in the field.

Around Baskil, the geommetrical system of hypabyssal rocks is of two types. First of them is basic dykes (locally acidic) which cut the plutonic rocks and observed up to one km in length (especially in tonalites). In the field, almost all the dykes show chilled margins. These dykes trend in NNE direction (Fig. 4).

Fig. 4 - Diabase dike, cutting granodiorites.

The second type is the cover rocks which overlie almost all the granitic rocks and are very easily seen by their dark colors. They are transitive to the overlying volcanic rocks.

Main hypabyssal rocks observed in the area are as follows:

a. Orbicular gabbro. — These rocks have formed in rare places in the world, including Baskil and have the features of natural monuments. In the study area, they are observed on the left flank of Hısırlık Dere, east of Haroğlu village and they are not mappable in 1:25 000 scale. The main outcrop is few meters wide and 5 m long. Orbicular gabbros are mainly altered and have soil appearance. Locally they are seen as nodules as big as egg or fist if not altered yet.

Some fragments which can reflect the overall features of the rock can be found in Hısırlık Dere as blocks. When observed closely, they are seen consisting two parts. First part is the matrix which contains coarse grains and the second part consists spherical nodules whose longitudinal axes are 5-15 cm (Fig. 5). in the matrix, melanocratic minerals, green amphiboles and yellowish-greenish pyroxene crystals are more abundant.

Fig. 5 - Orbicular gabbro.

When the nodules are observed a different structure is seen. In the cores of what called orbicule, generally mafic and coarse minerals are seen. Around the core there is a surrounding radial ring. In this ring extended plagioclase and olivine minerals are observed (Fig. 6). With respect to the volume of the orbicule, the number of the rings may increase. The composition of the ring is troctolite whereas the core is olivine-gabbro-norite. The results of the modal analysis of the core is as follows:

	%
Plagioclase	38.0
Orthopyroxene	12.0
Clinopyroxene	18.0
Olivine	21.0
Other minerals	11.0
Total	100

The matrix has pegmatitic features. Asutay (1985) states that the orbicules may reach the temperature up to 535±20°C, after the formation, related to metamorphism.

b. Diabase. — Most of the hypabyssal rocks in the study area are diabases. They are mainly observed as cutting the granites and as overlying the granites. They are transitive to overlying volcanic rocks when they are observed as parallel dykes. Chilling margins can clearly be observed in the dykes. The general trend is in NNE direction and sometimes they cut each other.

Under the microscope, diabases show ophitic, subophitic and porphyritic textures. Besides, plagioclase, augite, amphibole, secondary minerals such as chlorite, epidote and also magnetite as an opaque mineral can be seen.

Fig. 6 - The outer circulars texture of orbicular gabbro.

c. Granite porphyry. — Granite porphyry is locally observed in the study area. Its red color, resulted from the alteration is a distinguishing feature. In hand specimens, in a fine grained matrix, coarse quartz crystals are easily observed. Under the microscope, in a cryptocrystalline silicified matrix, clayish K-feldspar and also quartz minerals effected by magmatic corrosion is seen. Compositionally it can be classified as a semi-hypabyssal derivative of dacitic rocks.

d. Granophyry. — This is the youngest dyke system in the study area. It is seen everywhere in the region but is not frequently. Generally, they cut the diabase dykes. They are rich in leucocratic members and their grain sizes are variable. Coarse grained hand specimens are similar to granites whereas the fine grained ones resemble aplites. K-feldspars and quartz can be studied by naked eye. Some of them does not contain melanocratic minerals and some contains biotite even if it is in less amounts.

In thin sections granophyric texture is typical which is formed by K-feldspars and quartz.

e. Quartz veins. — Quartz veins are observed less than the other rocks in the study area. They are denser in the contacts of the quartz monzonites and the other rocks. Under the microscope, together with hypidiomorphic quartz crystals, chlorite and idiomorphic pyrite formations can be observed. No economical mineral formation is observed.

Volcanic rocks

Volcanic rocks which are not widespread around Baskil region are generally seen in forms andesitic lava flows. Sometimes the lava flows are seen as pillow lavas. When examined, they are seen to be highly altered. In some localities where they are in contact, it is difficult to distinguish the plutonic and volcanic rocks. Under the microscope, they are seen to be carbonatized and containing clays.

CHEMISTRY OF BASKİL GRANITIC ROCKS

From some selected samples of Baskil granitic rocks main element oxides (Table 1), trace elements (Table 2) and CIPW norms of the rocks (Table 3) have been determined. Only the sample No. 64 belongs to the gabbros of Kömürhan ophiolite (Yazgan and Asutay, 1981) which is out of the study-area. This sample has been analysed to show the difference between.

Baskil granitic rocks are wholly subalkaline rocks (Fig. 7) and they are rich in aluminium (Fig. 8) but not rich in iron as in the tholeiitic rocks (Fig. 9). The Baskil granitic rocks which are poor in femic minerals, fall into calc-alkaline field in diagrams (Fig. 10).

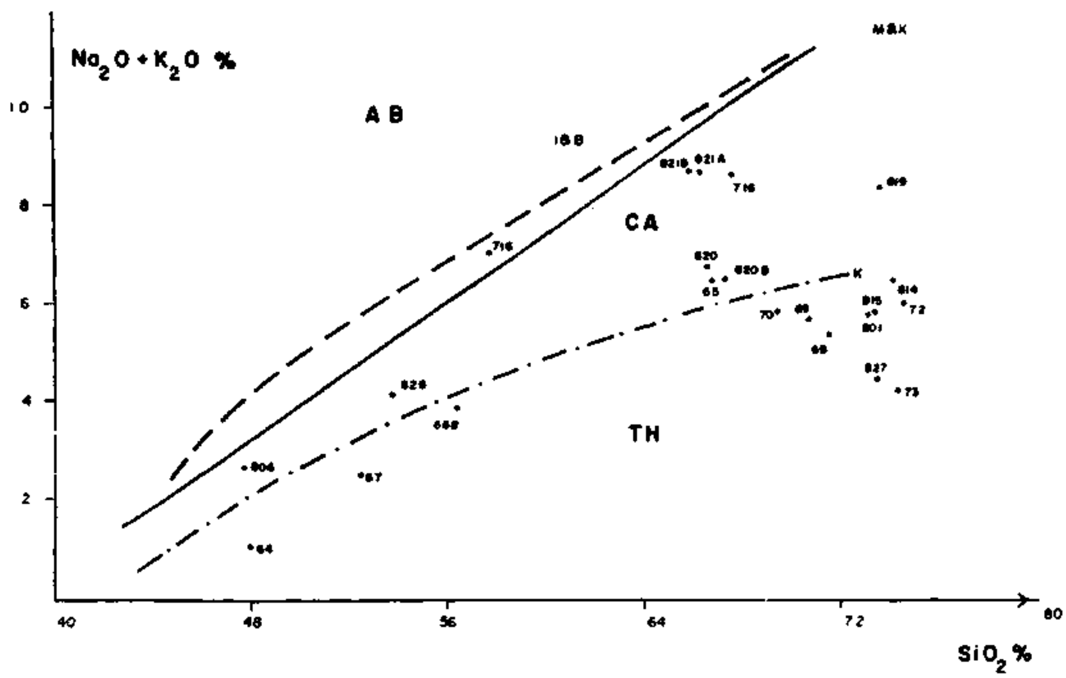

Fig. 7 - Distribution of Baskil magmatic rocks in $Na_2O + K_2O/SiO_2$ diagram.
I & B - Irvine and Baragar (1971); M & K - MacDonald and Katsura (1964); K - Kuno (1960).

PETROGENESIS OF BASKİL MAGMATIC ROCKS

Baskil magmatism is a very rare magmatic event so that in Turkey any similar magmatism may not be present. It reflects some characteristic features. There is a clear and regular interrelation between the plutonic, hypabyssal and volcanic rocks. These relations are shown in the field, under the microscope, and by chemical analyses.

Most of the contacts between the plutonic rocks are transitive. Clear contacts between the units are not observed. They are seen only between monzonite group and the adjacent rocks. Between diorite-monzonite and the tonalite-granodiorite groups, the transition is observed in a special group of rocks which are called transitive rocks. Mineralogical features of this group changes in between

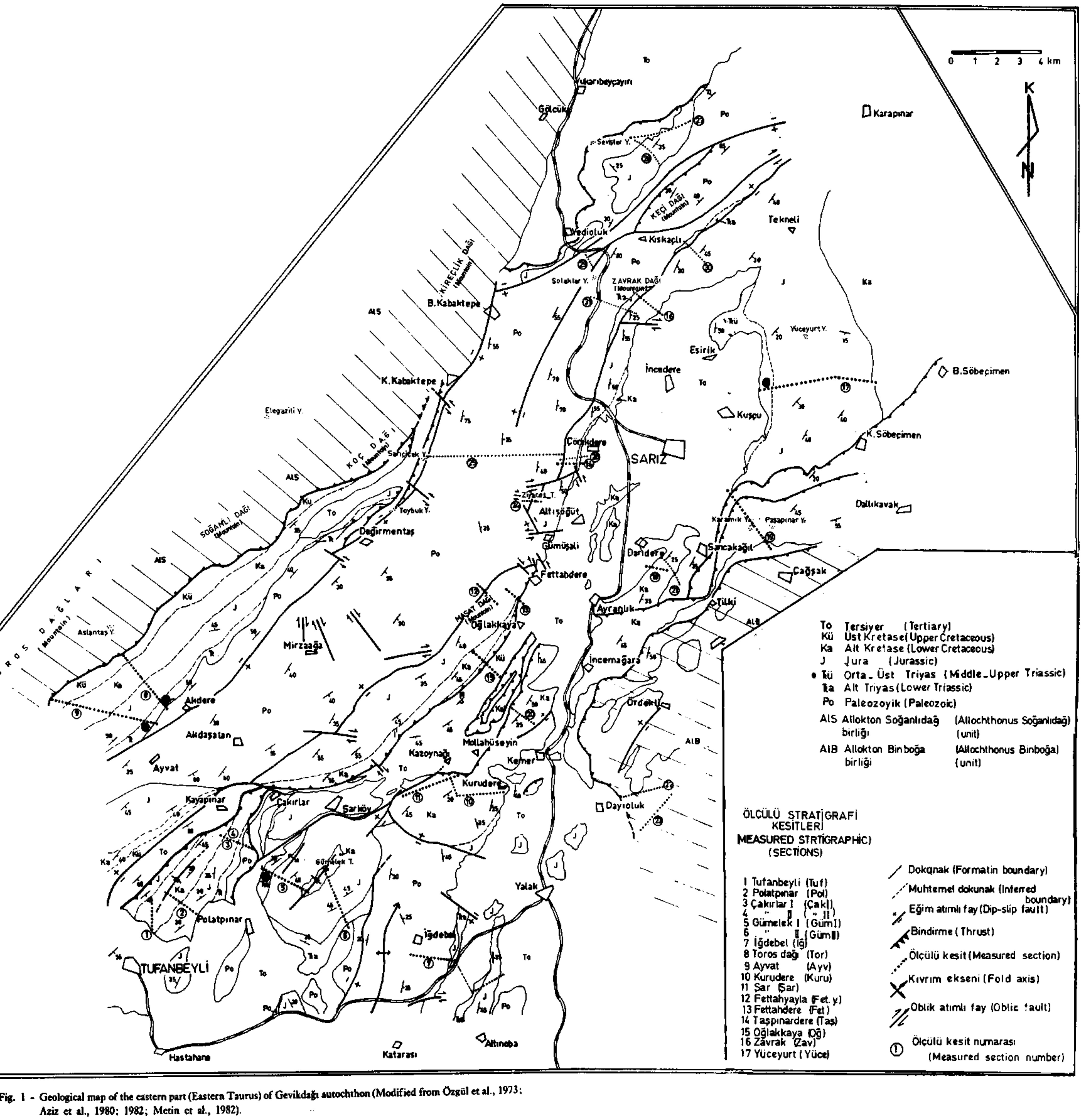

Fig. 1 - Geological map of the eastern part (Eastern Taurus) of Gevikdağı autochthon (Modified from Özgül et al., 1973; Aziz et al., 1980; 1982; Metin et al., 1982).

Table 1 - Variation of major elements in Baskil magmatic rocks

Sample no.	*64*	*65*	*66 B*	*67*	*68*	*69*	*70*	*71 G*	*71 C*	*72*	*73*	*802*	*806*	*814*	*815*	*819 A*	*820*	*820 B*	*821 A*	*821 B*	*826*	*827*
Rocks	*Gabbro*	*Gr.Dio.*	*Tonalite*	*Diorite*	*Granophyre*	*Granophyre*	*Gr. Dio.*	*Q. Mon.*	*Q. Mon.*	*Gr. Dio.*	*Tonalite*	*Gr. Dio.*	*Q. Dio.*	*Granophyre*	*Tonalite*	*Granophyre*	*Gr. Dio.*	*Gr. Dio.*	*Q. Mon.*	*Q. Mon.*	*Diabase*	*Tonalite*
SiO_2	48	66.8	56.4	52.5	71.6	70.7	69.4	67.6	57.7	74.6	74.3	73.41	47.64	74.06	73.22	73.75	66.54	67.15	66.71	65.94	53.90	73.57
Al_2O_3	21.09	15.3	15.3	10.8	14.2	13.8	14.3	14.5	15.6	13.0	12.1	13.19	16.96	13.73	13.93	14.92	15.67	16.23	15.42	15.16	14.58	12.87
F_2O_3	2.7	3.5	9.9	9.4	1.4	1.2	2.5	9.6	8.7	3.0	2.4	3.19	9.99	1.99	3.09	0.65	3.25	3.37	4.43	4.57	8.92	2.70
MgO	8.39	1.07	4.74	11.0	0.76	0.69	0.35	0.86	2.05	0.75	0.85	0.82	7.35	0.39	1.10	0.34	1.05	0.99	1.03	0.97	3.98	0.84
CaO	15.9	4.7	8.9	13.3	5.6	5.6	4.8	3.2	5.8	3.0	3.3	2.85	11.55	2.25	2.08	1.32	3.96	4.24	3.30	3.29	4.51	3.29
Na_2O	1.14	4.13	3.08	2.14	5.14	4.84	5.58	4.21	5.85	4.05	3.91	3.96	2.30	6.11	5.10	6.98	4.01	4.02	4.21	4.09	3.36	4.14
K_2O	0.17	2.36	0.75	0.37	0.29	0.23	0.33	4.43	1.34	1.91	0.28	1.89	0.88	0.38	0.57	1.38	2.71	2.54	4.43	4.52	0.85	0.26
Mn_3O	—	—	—	—	—	—	—	—	—	—	—	0.18	0.22	0.10	0.07	0.05	0.09	0.09	0.13	0.15	0.15	0.09
Mn_3O_4	0.064	0.071	0.186	0.145	0.028	0.024	0.031	0.096	0.226	0.087	0.051	—	—	—	—	—	—	—	—	—	—	—
TiO_2	0.07	0.21	0.66	0.34	0.26	0.24	0.26	0.22	0.51	0.24	0.25	0.26	0.62	0.10	0.44	0.10	0.23	0.24	0.30	0.29	0.87	0.27
BaO	>0.01	0.23	>0.01	>0.01	>0.01	>0.01	>0.01	0.04	>0.01	0.02	>0.01	—	—	—	—	—	—	—	—	—	—	—
S_2O	0.01	0.09	0.01	0.01	0.04	0.04	0.04	0.02	0.08	0.01	0.01	—	—	—	—	—	—	—	—	—	—	—
H +	2.62	0.77	0.94	1.34	0.59	1.10	0.74	0.22	0.69	0.27	1.02	—	—	—	—		—	—	—	—	—	—
P_2O_5	—	—	—	—	—	—	—	—	—	—	—	0.05	0.08	0.03	0.07	0.08	0.07	0.11	0.96	0.09	0.12	0.04
Total	**100.97**	**99.23**	**100.87**	**101.39**	**99.91**	**98.47**	**98.69**	**104.99**	**98.49**	**100.93**	**99.08**	**99.75**	**97.04**	**99.14**	**99.67**	**98.57**	**97.57**	**98.98**	**100.42**	**99.07**	**91.24**	**98.07**

Table 2 - Variation of trace elements in Baskil magmatic rocks

								TRACE ELEMENTS														
Sample no.	*Nb*	*Zr*	*Y*	*Sr*	*U*	*Rb*	*Th*	*Pb*	*Ga*	*Zn*	*Cu*	*Cr*	*Ni*	*Co*	*Nd*	*Sm*	*Pr*	*Ce*	*Ba*	*La*	*Ti*	*V*
64	—	—	3	132	—	—	—	—	—	44	36	969	—	—	—	—	—	—	58	—	0.07	48
65	—	—	17	893	—	—	—	—	—	10	47	18	—	—	—	—	—	—	2527	—	0.21	46
66 B	—	—	41	153	—	—	—	—	—	78	107	96	—	—	—	—	—	—	87	—	0.66	269
67	—	—	20	153	—	—	—	—	—	51	5	518	—	—	—	—	—	—	43	—	0.34	194
68	—	—	20	374	—	—	—		—	2	10	2	—	—	—	—	—	—	58	—	0.26	26
69	—	—	18	351	—	—	—	—	—	2	5	2	—	—	—	—	—	—	54	—	0.24	25
70	—	—	22	387	—	—	—	—	—	21	11	16	—	—	—	—	—	—	79	—	0.26	27
716	—	—	22	204	—	—	—	—	—	43	12	2	—	—	—	—	—	—	394	—	0.22	39
71 C	—	—	43	206	—	—	—	—	—	94	68	154	—	—	—	—	—	—	99	—	0.51	105
72	—	—	20	104	—	—	—	—	—	25	10	2	—	—	—	—	—	—	208	—	0.24	34
73	—	—	33	115	—	—	—	—	—	11	7	2	—	—	—	—	—	—	76	—	0.25	41
802	2	109	24	103	0	59	3	7	14	33	7	122	5	8	0	0	0	20	216	12	—	—
806	2	30	20	237	0	19	0	9	16	91	113	148	25	32	0	0	0	0	113	6	—	—
814	8	141	11	257	7	17	27	15	19	25	16	174	5	11	18		0	33	53	26	—	—
815	2	96	37	168	0	12	0	5	15	19	6	117	0	9	0	0	0	30	82	10	—	—
819 A	7	61	20	174	0	20	7	10	17	7	64	84	15	6	0	0	0	33	464	22	—	—
820	4	147	18	798	16	88	28	37	21	25	26	79	5	8	0	0	0	45	2212	41	—	—
820 B	3	154	20	849	5	84	30	41	19	28	33	84	3	9	0	0	0	41	2312	40	—	—
821 A	19	165	36	202	4	183	18	22	19	49	8	70	0	10	29	0	0	42	411	26	—	—
821 B	19	168	36	192	5	190	30	23	20	53	10	67	6	10	17	0	0	67	399	54	—	—
826	2	66	33	124	0	18	0	14	17	106	74	34	0	22	15	0	0	0	107	0	—	—
827	2	99	33	123	0	5	0	5	13	21	4	130	0	11	0	0	0	15	71	0	—	—

Table 3 - CIPW norms which are calculated from chemical analyses of samples

No.	*IL*	*OR*	*AB*	*AC*	*AN*	*TN*	*HT*	*WO*	*DI*	*HY*	*OL*	*NE*	*CS*	*LC*	*PF*	*Q*	*HM*	*A*	*F*	*M*	*CI*	*AD*
802	0.49	11.17	33.51	0.00	12.63	0.00	1.73	0.00	1.24	3.32	0.00	0.00	0.00	0.00	0.00	35.28	0.00	61.32	30.09	8.59	6.78	SACA
814	0.19	2.25	51.70	0.00	8.92	0.00	0.46	0.00	1.91	2.32	0.00	0.00	0.00	0.00	0.00	31.09	0.00	74.87	20.63	4.50	4.89	SACA
82.08	0.46	15.01	34.01	0.00	18.74	0.00	2.69	0.00	1.79	1.64	0.00	0.00	0.00	0.00	0.00	23.98	0.35	62.01	28.64	4.36	6.91	SACA
821/A	0.57	26.18	35.62	0.00	10.09	0.00	2.23	0.00	5.22	2.97	0.00	0.00	0.00	0.00	0.00	16.15	0.00	63.27	29.19	7.54	11.00	SACA
826	1.65	9.02	28.43	0.00	22.19	0.00	2.68	0.00	0.15	18.55	0.00	0.00	0.00	0.00	0.00	11.58	0.00	25.96	49.49	24.55	23.04	SATH
827	0.51	1.54	35.03	0.00	15.77	0.00	1.23	0.00	0.46	3.77	0.00	0.00	0.00	0.00	0.00	37.45	0.00	57.41	31.64	10.96	5.77	SACA
806	1.18	1.95	19.46	0.00	34.78	0.00	4.51	0.00	17.99	12.29	3.69	0.00	0.00	0.00	0.00	0.00	0.00	13.87	47.39	38.75	39.66	SATH
815	0.84	3.37	43.15	0.00	10.32	0.00	2.52	0.00	0.00	2.80	0.00	0.00	0.00	0.00	0.00	35.25	0.00	39.40	29.08	11.52	6.16	SACA
819 A	0.15	8.15	59.06	0.60	5.30	0.00	0.00	0.00	0.91	0.43	0.00	0.00	0.00	0.00	0.00	24.81	0.57	10.06	6.28	3.66	2.05	SACA
820	0.44	11.01	33.93	0.00	11.32	0.00	2.60	0.00	0.00	3.14	0.00	0.00	0.00	0.00	0.00	24.00	0.00	62.92	27.35	9.74	6.17	SACA
821 B	0.55	26.71	34.61	0.00	9.66	0.00	2.44	0.00	5.55	2.56	0.00	0.00	0.00	0.00	0.00	16.48	0.00	52.88	30.03	7.08	11.09	SACA

SA — Subalkaline; CA — Calc-alkaline; TH — Tholeiitic.

diorites and tonalites for which quartz is responsible. Together with the anhedral minerals which have clear contacts with the other minerals there are round or elliptic quartz crystals whose roundness increase towards tonalites. The inclinations from formerly emplaced plutonic rocks to the rocks rich in quartz is a feature of Baskil magmatism. This inclination ceases from tonalite and granodiorite and passes into a magma rich in K-feldspar. The forms seen in the quartz crystals which gives a porphyritic texture to the tonalites are related to the events in the magma chamber. These forms may be due to the movements during the formation and these movements are not in large scale and may be because of the changes in the inclination in the magma chamber. The quartz-feldspathic solution whose density increases with the advance of differentiation locally surrounds the dioritic and quartz-dioritic remnants (Fig. 3). This event is the increase of quartz rich solution in the magma chamber and it, as an autolith, surronds the previously formed magma products which are rich in melanocratic minerals. No heat transfer between remnant magmatic rocks is seen. Their contact is not clear. In some outcrops the relics of the newly formed rocks are observed in remnant rocks.

The events we have observed in transitive rocks may not be considered as assimilations but the proofs of the changing conditions in the same magma chamber.

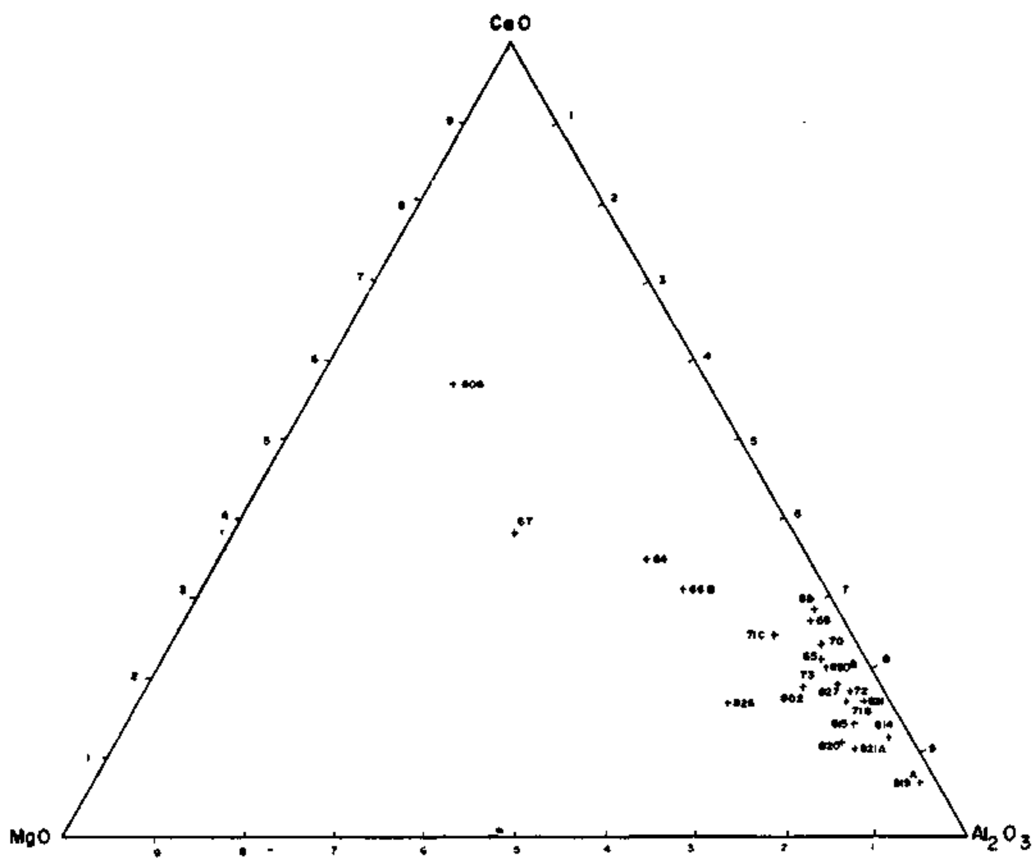

Fig. 8 - Distribution of Baskil magmatic rocks in CaO-MgO-Al_2O_3 triangular diagram.

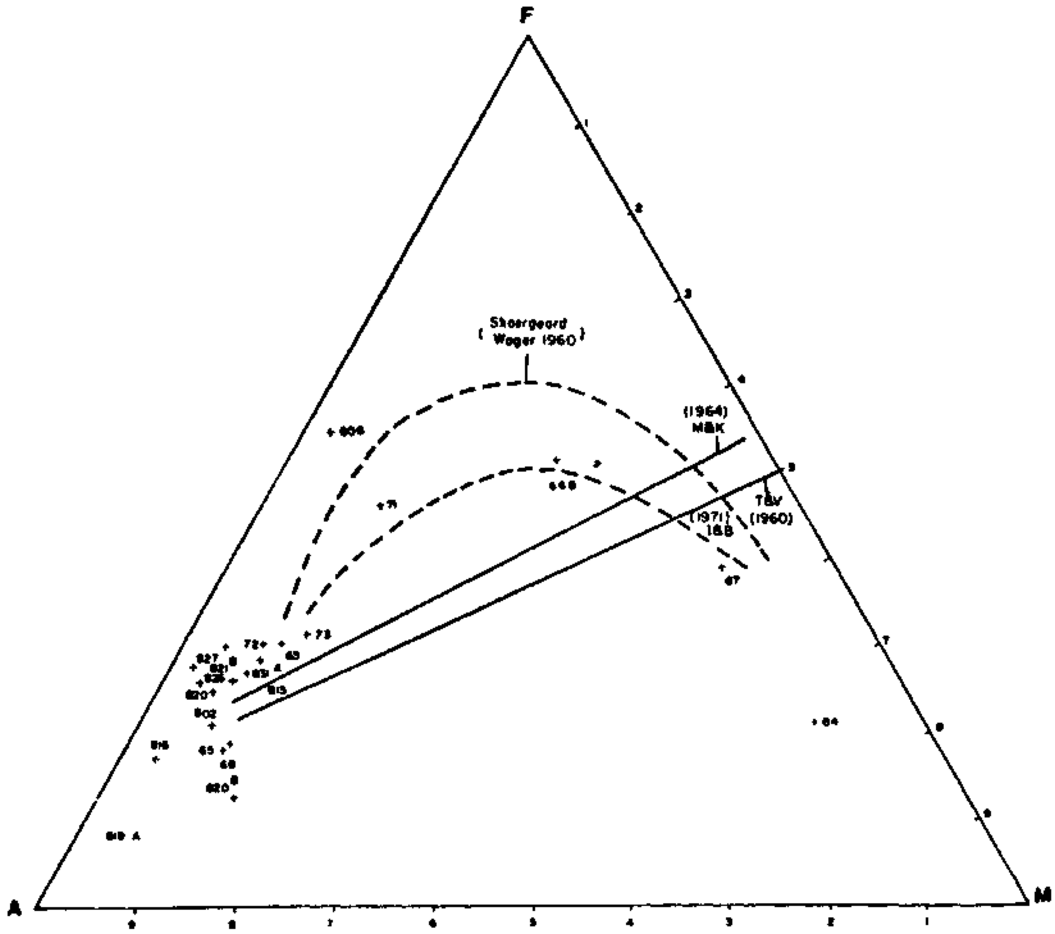

Fig. 9 - Distribution of Baskil magmatic rocks in AFM diagram. After, Wager (1960); M & K (1964) - MacDonald and Katsura (1964); I & B - Irvine and Baragar (1971); T & V - Turekian and Vedepahl (1961).

Quartz-monzonites are the last emplaced rocks in the Baskil plutonic rocks and they have quite clear contacts with the other rocks. In regional scale (especially around Keban) the last emplaced rocks are syenites (Asutay and Turan, 1986). This shows the increase of K-feldspars in magma after a certain time. This feature is also visible in the granitic rocks of the Divriği region. The results of the modal analyses of Gysin (1943), when applied to the Streckeisen diagram shows that Divriği plutonic rocks are almost, 99 %, under the line of 20 %. When the granitic rocks of the two regions are compared, it is seen that the Baskil granites are richer in quartz. When the increase of K-feldspar in Baskil magmatics after a certain time and the regional positions of Baskil and Divriği are considered, it may be concluded that both events may have the same genesis.

THE TYPE AND AGE OF BASKİL GRANITE

The features of the Baskil granite has been studied and has been compared with the descriptions (Appendix 1) of the various researchers and its type has been determined. These features are as follows:

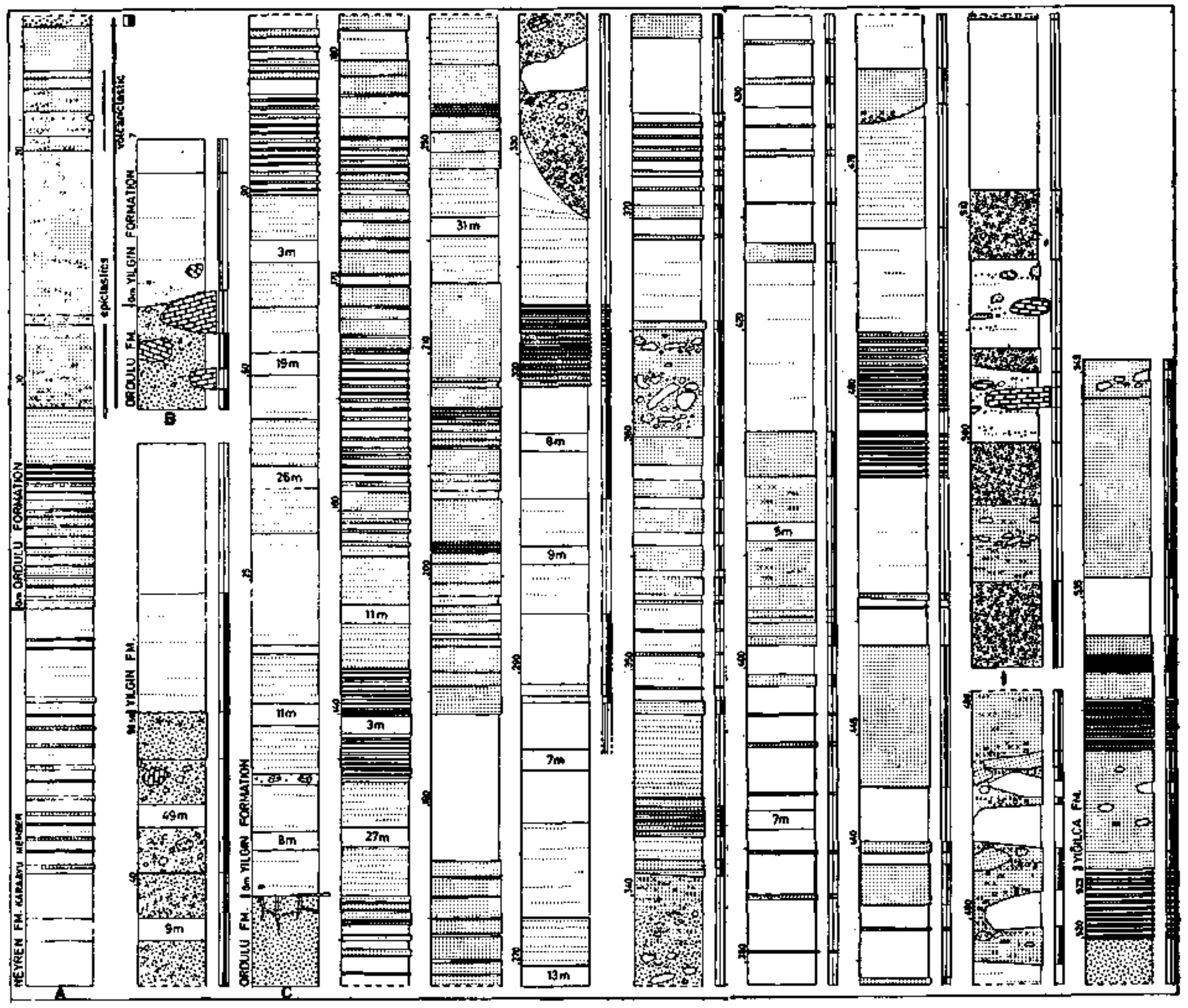

Fig. 10 - A - Complete type section of the Ordular formation; B, C - unconformity between the Yılgın and Ordulu formations, complete composite type section of the Yılgın formation.

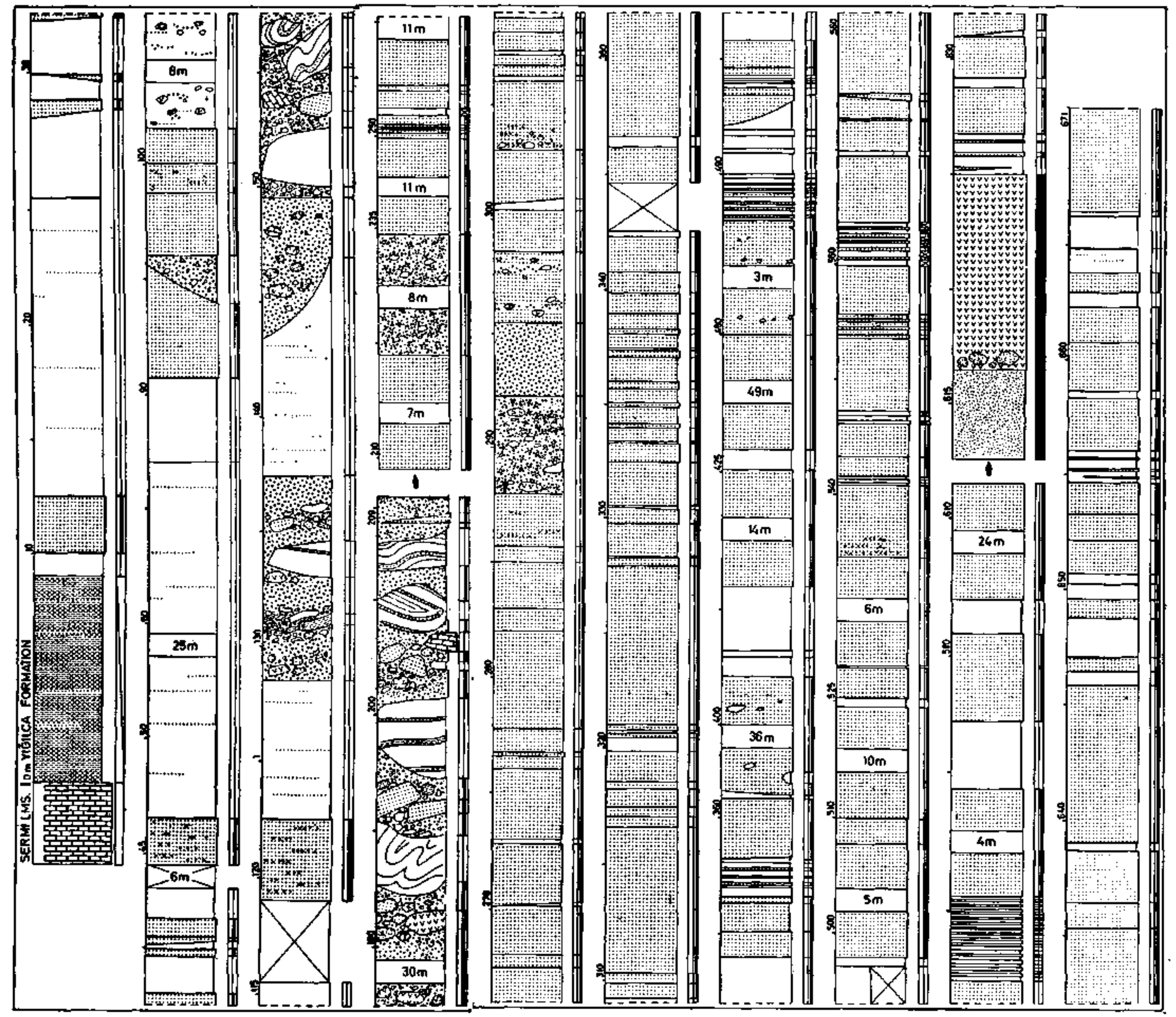

Fig. 11 - Partial composite type section of the Yığılca formation.

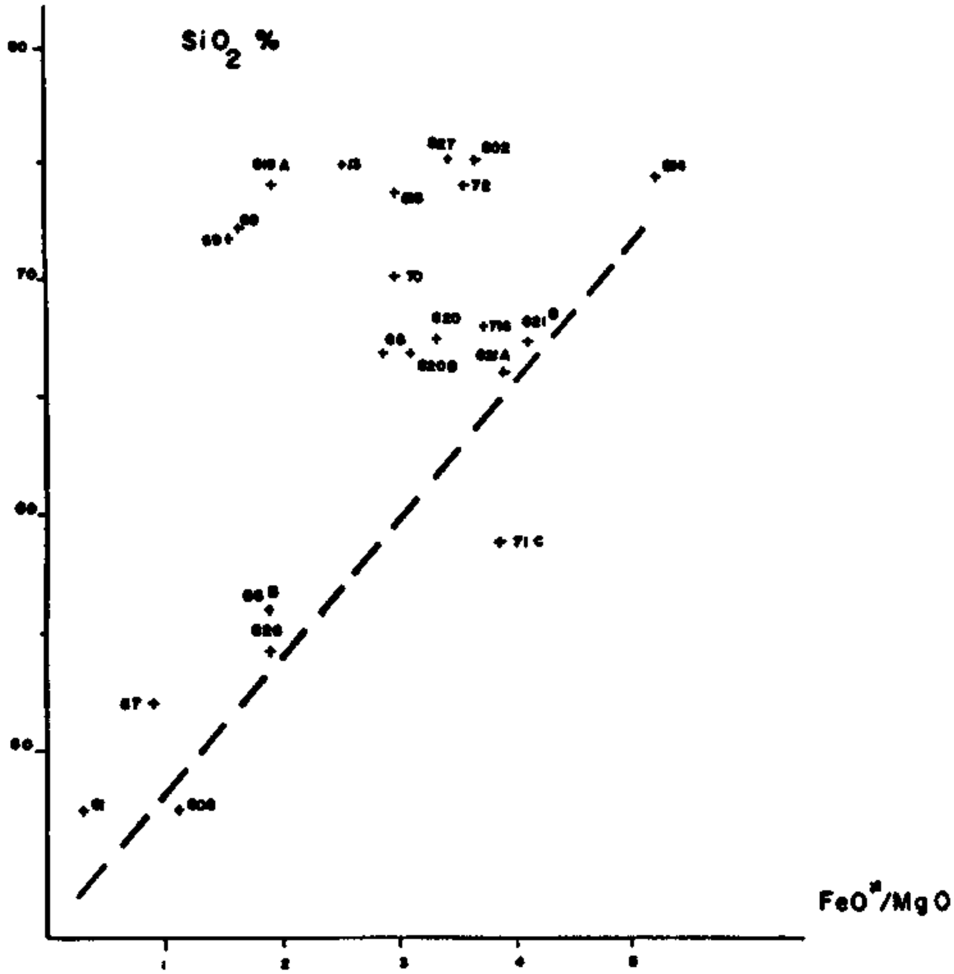

Fig. 10 - Distribution of Baskil magmatic rocks in $SiO_2/FeO/MgO$ diagram. (After, Miyashiro, 1975). FeO - Total iron.

1. Baskil granite is of type 'I', because the descriptions made for this type is suitable (+marked features in Appendix 1).

2. According to Nb/SiO_2 diagram (Fig. 13), low Nb magma is dependent to subduction zones (Pearce and Gale, 1977).

3. According to geochemical data, Baskil granite is calc-alkaline, it has no similarities alkaline or per-alkaline granites (Fig. 13a).

4. It has been formed in a compactional environment, because in the region, a continental collision has taken place in Upper Cretaceous (Yazgan, 1984).

5. The distribution of trace elements points out a regular magmatic crystallization for Baskil granitic rocks (Asutay, 1985).

6. A regular relation with the andesites which are only seen in subduction zones is observed. The oldest unit with which Baskil granite has contact is Keban metamorphics. In these contacts well developed skarn zones are always observed. Now that the sedimentation age of the metamorphic rocks

is Upper Paleozoic-Lower Mesozoic (Özgül, 1976, 1981) it can be said that Baskil granite is younger than Triassic age. However, the age of the granitic rocks is determined only when the age of the Baskil magmatic rocks are studied. The volcanics related to Baskil magmatic rocks are the distal flysch deposits and its interbeds of Sağdıçlar formation which crops out well around Hamzikan village and Ciban quarter (Asutay and Turan, 1986). The age of the flysch is determined as Santonian-Campanian by M. Serdaroğlu. The samples from the formation includes the following fossils: *Dicarinella concavata* (Brontzen), *Globotruncana lapparenti* (Qr), *Globotnmcana area* (Cushman), *Marginotruncana marginata* (Reuss).

Fig. 11 - Diyoritic restits in transitional rocks.

When the above data and the relation between the granitic rocks and hypabyssal rocks are considered, it is deduced that the age of the Baskil granite is older than Santonian. The granitic pebbles of the conglomerate in the base of the Maestrichtian limestones present around Keban supports the age of magmatism (Asutay and Turan, 1986). On the other hand, Yazgan (1984) determines the age of the granitic rocks as Coniacian (Appendix 2).

CONTACT METAMORPHISM

Keban metamorphics has experienced contact metamorphism due to presence of Baskil plutonic rocks. In the study area, contact metamorphism is observed in two ways:

a - Endomorphism: Metamorphism occurred inside the magmatic rock.

b - Exomorphism: Metamorphism observed in the rock which is intruded by magmatic rock. Different minerals have been determined in both kind of metamorphisms.

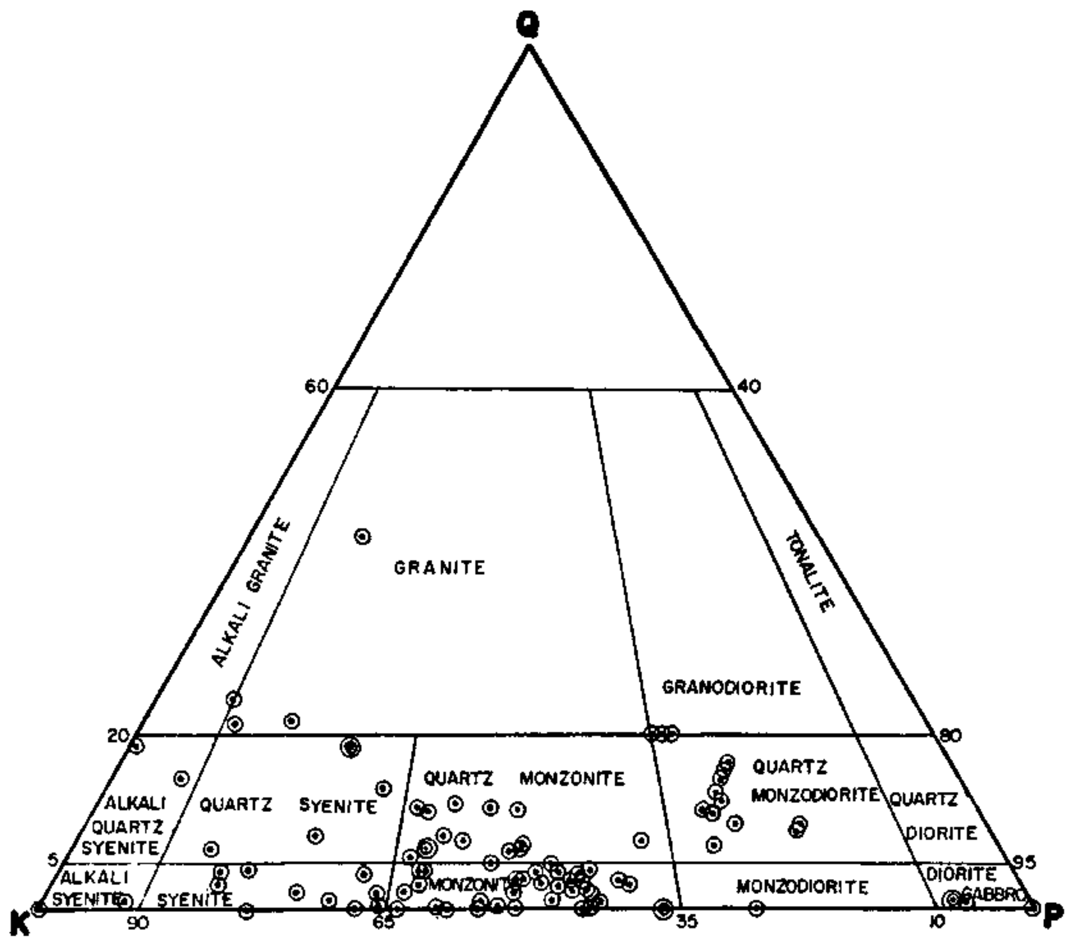

Fig. 12 - Distribution of Divriği granitic rocks in Streckeisen diagram (1976).

Endomorphism

Endomorphism is observed in the plutonic rocks when they are close to the Keban metamorphics. The rocks experienced this metamorphism are generally diorite-monzodiorite group and quartz-monzonites. The following minerals have been determined from the samples in which endomorphism is observed:

Garnet. — Generally idiomorphic, yellowish-light brown under single nicol and hexagonal. It has clear contacts with the other minerals. As a result of the qualitative analyses (microprobe) «grandite» composition has been determined (Asutay, 1985).

Clinopyroxene (diopside). — Pyroxenes, together with short prismatic sections locally display (001) sections and are generally green and shows pale green pleochroism. In the thin sections whose double birefringence is high, extinction angle is determined as 37°. 2V is 60° and it is optically positive.

Amphibole., — It is generally in prismatic forms and green-brownish, green in color. Pleochroism formula is (determined as light green (x), green (y) and dark green (z). Its pseudohexagonal sections are observed with their typical cleavages.

Sphene. — It is easily known by its very high double birefringence and idiomorphic minerals. Its rhombic sections are widespread.

Appendix 1 - Differences of 'I' type and 'S' type granites

Data	*I type granites*	*S type granites*
Field	+ Generally large-scale intrusions	Generally small-scale intrusions
	+ Wide composional distribution. Gabbro-diorite 15 %; Granodiorite 50 %; Granite 35 %	Limited compositional distribution. Gabbro-diorite 2 %; Granodiorite 18 %; Granite 80 %
	+ Genetic and spatial relation with volcanites	No synchronologic volcanism.
Mineralogic	+ Hornblende more dominant than biotite	Biotite more dominant than hornblende
	+ Little muscovite (in too felsic rocks)	Muscovite and double-mica granite dominant
	+ Magnetite being the most dominant iron-oxide	Ilmenite being the most dominant iron-oxide
	+ Allanite and sfen. No accessory cordierite, garnet, andalusite or sillimanite	Monozite and cassiterite. Accessory cordierite, garnet, andolusite and sillimanite may be present
Chemical	+ Mol. $Al_2O_3/Na_2O.K_2O.CaO$ 1.1 SiO_2 65% very widespread	Mol. $Al_2O_3/Na_2O.K_2O.CaO$ 1.1 SiO_2 65%
	+ Na_2O in normal levels (3.2% in felsic rocks and 22% in mafic rocks)	Na_2O may be low. (in rocks bearing 5% K_2O 3.2%)
	+ Regular element distribution. Normative diopside or normative corundum less than 1% in CIPW norms.	Distribution diagrams irregular. Always contains normative corundum.
	+ $Fe^3/(Fe^3+Fe^2)$ generally high and bigger than 0.2	$Fe^3/(Fe^3+Fe^2)$ generally low.
Isotopic	Low $Sr^{87/86}$ ratio (<0.706)	High Sr^{87}/Sr^{86} ratio (<0.706)

After, Chappell and White (1974), Coleman (1980), Hine and others (1978), Ishira (1978) O'neil and others (1977), Pankhurst (1980), Pitcher (1979-1980), White and Chappell (1977).

Appendix 2 - Some radiometric ages of Baskil magmatics

Sample no. *Rock type* *Location*	*System*	*K_2O (wt%)*	*^{40}Ar. rad. (10^{-11} mol/g)*	*$100 \times \frac{^{40}Ar.\ rad.}{^{40}Ar.\ total}$*	*Numbering age ($\pm\sigma$) (Million year)*
Sanidine-bearing microsyenite intruded into Keban marble (Northern Keban) 30/80	Sanidine and other feldspars	3.98	45.98	79	78.5 ± 2.5
31/80	Sanidine	11.90	133.2	82	76.0 ± 2.5
Amphibole-gabbro intruded into Keban marble (Elazığ-Keban highway) 124/81	Hornblende	274	3.129	26	77.5 ± 4.5
Granodiorite (SW Baskil) 95/79	Biyotite	6.12	78.00	90	86.5 ± 2.5
Quartz-monzodiorite (Southern Baskil) 96/79	Hornblende	941	11.55	87	83.5 ± 2.5
Quartz-monzodiorite (Southern Baskil) 100/79	Hornblende	11.080	13.71	87	86.0 ± 2.5
Quartz-monzodiorite (Southern Baskil) 101/79	Hornblende	901	11.16	87	84.0 ± 2.5
Diorite (Southern Baskil)	Amfibol	1.50	16.71	92	76 ± 2.5
9/2-78	Biotite	6.68	73.94	94	75.5 ± 2.5

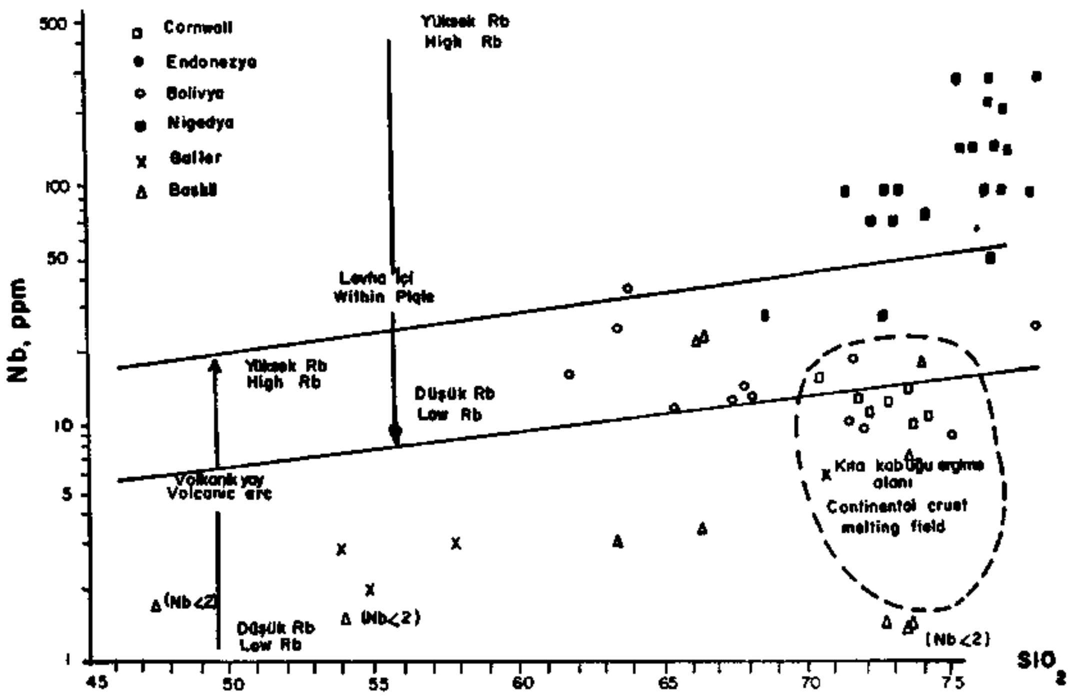

Fig. 13 - Distribution of Baskil magmatic rocks in Nb/SiO_2 diagram and comparison of some other granites of the world.

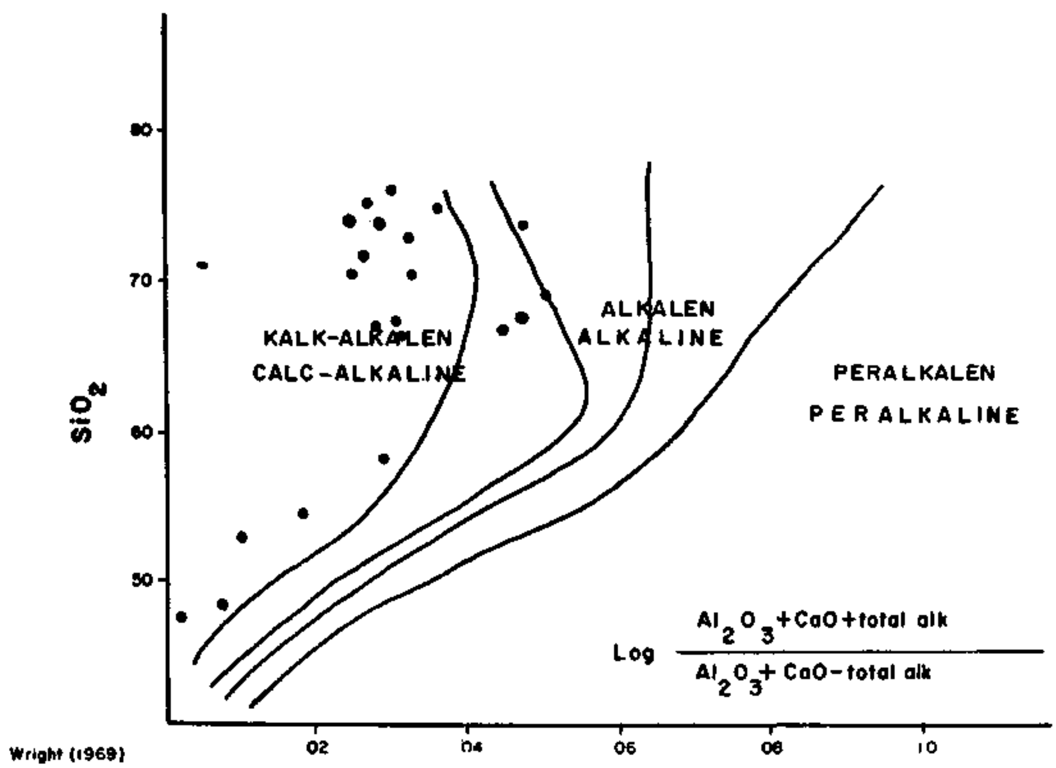

Fig. 13a - Distribution of Baskil magmatic rocks in Wright (1969) diagram.

As will be seen from the above mentioned minerals an endomorphism rich in Ca and Mg is displayed. In the surrounding rocks there are no minerals which would experience this metamorphism except for the Keban metamorphics. The existing ones are younger than the magmatism. The minerals in the rocks reflecting endomorphism have formed as a result of the assimilation of the marbles of Keban metamorphics, containing dolomite and calcite, by the Baskil magmatic rocks. The mineral paragenesis in the rocks corresponds to the pyroxene-hornfels facies conditions. Together with this paragenesis, alteration in the same magmatic rocks and sausaritization in feldspars are observed. This formation, in a way, is a clear result of Ca metasomatism. A similar endomorphism is observed also in Akdağmadeni by Sağıroğlu (1982).

Exomorphism

Exomorphism is observed in Keban metamorphics in which magmatic rocks are intruded. Around Baskil, both plutonic and hypabyssal rockshave experienced contact metamorphism. Of these, the hypabyssal ones are orbicular gabbros. In Keban metamorphics which have contacts with orbicular gabbros minerals such as quartz, garnet (andradite), clinopyroxene (diopside) are observed.

In the contacts of plutonic rocks with Keban metamorphics locally the melanocratic minerals of contact metamorphism and mostly calcite are observed. The size of calcite minerals are smaller when they are away from the contact. In such rocks saccaroid texture has developed, they are easily breakable into pieces. These rocks are like islands in the magmatics around Keban and Elazığ. In Keban metamorphics such marbles are not seen. Some researchers, since they have not seen any mafic mineral in these marbles, have not differentiated them from Keban metamorphics. These marbles are the raw materials of the limestone quarries. As mafic minerals, in the contact metamorphism, minerals such as magnetite, olivine (forsterite), spinel have been determined (Asutay, 1985). Aşvan iron ore displays contact metasomatic iron mineralization which is a result of the exomorphism between Keban metamorphics and Baskil magmatics.

RESULTS AND DISCUSSION

The main results obtained from the above study are as follows:

a. Baskil magmatic rocks form a regular sequence and cannot be interpreted as a complex;

b. Baskil magmatics have been derived from a calc-alkaline magma which has continental margin magmatism characteristics;

c. Between Baskil magmatics and Keban metamorphics always contact metamorphism is observed;

d. No ophiolitic rocks are observed in Baskil magmatics.

Together with these results the following results obtained from some observed phenomena around the area studied:

The first sedimentary unit deposited on the Baskil magmatics is Santonian-Campanian aged Sağdıçlar formation (Asutay and Turan, 1986). Upper Maestrichtian aged Harami formation overlies both Keban metamorphics and Sağdıçlar formation in the close vicinity of the study area. The trend of the axes (NNE) of the great asymmetrical folds observed in the Keban metamorphics are coinciding with the diabase dykes cutting the Baskil granite. It is known that the first compaction in the region has started in Turonian (Yazgan, 1984), therefore Keban carbonates have been metamorphized possibly in Turonian-Lower Maestrichtian. When the structural features with granitic rocks are regarded, this age may change up to Coniacian-Lower Maestrichtian. In Lower Maestrichtian the tectonic style of the region has greatly changed. Because, the carbonates of Harami formation overlies the folded Sağdıçlar formation in a quiet and uniform way (Asutay and Turan, 1986).

Baskil magmatics may be said to be a product of continental margin magmatism. They are presumably, the products of an oceanic lithosphere existing between Keban microplate and Arabian platform which later on subducted under Keban microplate.

ACKNOWLEDGEMENTS

Author thanks E. Sirel and M. Serdaroğlu who made the fossil determinations and datings.

Manuscript received January 20, 1986

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GEOLOGICAL EVOLUTION AND BASIN MODELS DURING NEOTECTONIC EPISODE IN THE EASTERN ANATOLIA

Fuat ŞAROĞLU* and Yücel YILMAZ**

ABSTRACT. —In the Eastern Anatolia, neotectonic regime beginning in Middle Miocene has considerably affected the geological evolution of the region. During the neotectonic episode, compressional tectonic regime, characteristic for the region, resulted in formation of folds, thrust and strike-slip faults, and large-scale extentional fractures. Under the control of all these structural elements, basically two types of basins (intermountain and pull-apart) are formed. Among these, Muş, Ahlat-Adilcevaz and Karayazı-Tekman basins are the intermountain basins. Kağızman-Tuzluca basin, however, has been evolved as a pull-apart type. The Erzurum-Pasinler-Horasan is another type of intermountain basin which was also affected by strike-slip faults. The general features of the new episode deposits are to be in nonmarine facies and with the coeval volcanites their accumulation in separate basins.

INTRODUCTION

This paper is an attempt to introduce the main affects of the neotectonism to geological evolution of the Eastern Anatolia. The Eastern Anatolia is a tectonic region that is characterized by unique deformational style during neotectonic episode (McKenzie, 1972; Şengör, 1980; Şaroğlu and Yılmaz, 1984). The area, which is here introduced under the name of the Eastern Anatolia region, is to the further east of the intersection point (around the east of Karlıova) of North and South Anatolian Faults (Allen, 1969; Apart and Şaroğlu, 1972; Şengör, 1979). As in Figure 1, the Eastern Anatolia lies between the Pontides on the north, folded and thrust belt on the south, and extends to the Turkish-Iranian and Turkish-Russian state boundary to the east (Ketin, 1966).

Some earlier works have been carried but dealing with the general features of the neotectonism in the Eastern Anatolia. These works primarily discussed the structural, morphological and volcanic events in the region (Şaroğlu et al., 1980; Şengör, 1980; Şaroğlu and Güner, 1981; Yılmaz, 1984; Şaroğlu and Yılmaz, 1984; Yılmaz et al., 1986; Şaroğlu, 1985). These studies have also provided information on the tectonically related structures and their subsequent deformational geometries that are shaped up in the period between the last change in tectonic regime and the present time.

Depending on the closure of Neotethys, this neotectonic evolution is the result of the continent-continent collision that is evident along the Bitlis suture belt (Şengör et al., 1979). The continent-continent collision initiated a new tectonic episode mainly characterized by compressional tectonic regime in the Eastern Anatolia. During this new episode, folds, thrust and strike-slip faults, and large-scale pull-apart type of extentional fractures are formed. These structures led to the narrowing and subsequent widening of the region in N-S and E-W directions, respectively. These structures also caused the thickening of continental crust, thus subsequent uplifting of the region.

In general, synclines and anticlines trend east-west in the region. They correspond to and are overlain by east-west trending basins and elongate ridges. Different type of basins are also formed along the north-south trending extentional fractures and in areas where strike-slip faults step up in en-echelon character. The young volcanism in the region has displayed some changes depending on the evolution of the continental crust. Eruptions have mostly followed the extentional fractures and chose them as paths to get out. N-S trending deep valleys and E-W trending meandering rivers are among other features formed during this neotectonic episode.

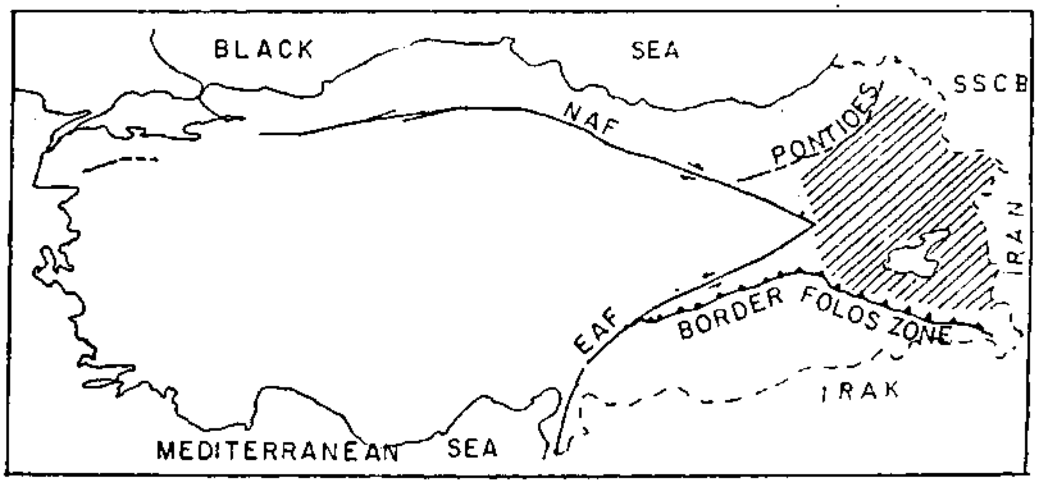

Fig. 1 - Location map for the study area.

In the region, sedimentary rocks and coeval volcanites cover very wide areas. In this paper, first the stratigraphy of new episode will be introduced, and then, to a great extent, the geological evolution of the region will be discussed under the light of the stratigraphic relations.

GENERAL ASPECTS OF THE STRATIGRAPHY OF THE NEOTECTONIC EPISODE IN THE EASTERN ANATOLIA

Geological evolution of the Eastern Anatolia can be analized in four structural stages (Şaroğlu and Güner, 1981; Şaroğlu and Yılmaz, 1984). From oldest to youngest these can be arranged as in the following.

The first stage covers the Palaeozoic to Lower Mesozoic metamorphic rocks that are the oldest strata in the region (Boray, 1975; Perinçek, 1980; Perinçek and Özkaya, 1981; Yılmaz et al., 1981; Göncüoğlu and Turhan, 1983; Çağlayan et al., 1983). The second stage rocks consist of ophiolitic melange type of strata that was structurally pushed over the first group in Upper Cretaceous (Demirtaşlı and Pisoni, 1965; Ketin, 1977; Yılmaz et al., 1981). The third stage rocks cover a sequence of Eocene to Lower Miocene sedimentary rocks. They unconformably overlie the first and second group strata. The fourth stage strata consist of the Upper Miocene to present day deposits. They are nonmarine in character and are strongly affected by both volcanism and neotectonism. The last stage sediments unconformably sit on the all of these older rocks and have some diastems and stratigraphic onlaps. These are the products of neotectonic episode.

Overall, the tectonic evaluation (including paleotectonic) of the Eastern Anatolia was earlier presented by Şengör and Yılmaz (1981). This paper, however, only deals with the tectonic events that has occured during the neotectonic episode.

First, the stratigraphy will be introduced within the eight areas where the sediments of the neotectonic episode have very widespread distribution. Second, the stratigraphic units will be considered in terms of time and space. Third, on the basis of facies, facies changes and distributions in each basin, an attempt will be made to explain the evolution of the entire region. Among these areas, first the ones that are on the south and then, the ones on the north will be introduced. These areas are in turn, (1) Karlıova-Bingöl, (2) Muş, (3) Ahlat-Adilcevaz, (4) Karayazı-Tekman, (5) Hınıs, (6) Zırnak, (7) Erzurum-Pasinler-Horasan, and (8) Kağızman-Tuzluca. Among them, there are several small and large basins, as well. Along with the 8 areas, evolution of the other basins will be considered eventhough their stratigraphy is not included in this paper.

Bingöl-Karlıova area

A very large area, between Muş basin and the East Anatolian strike-slip fault, is covered with widespread volcanites (Fig. 2). Here, Tertiary deposits are exposed at several localities. The stratigraphic relations between the Tertiary rocks and underlying basement strata are well discernable. Such relations are important because they lead us to know the areal distribution of the Tertiary sediments deposited following the paleotectonic events. The Lower Miocene strata are primarily overlying the older rocks thin to the west.

In the Bingöl-Karlıova area, the Lower Miocene consists of Adilcevaz limestone. The Adilcevaz unconformably overlies the basement strata and have sandy limestones at its very base. The Adilcevaz limestone is exposed at very small localities along the North Anatolian fault. Fossils found in the limestones are characteristic for Burdigalian. Very lower portion of the Adilcevaz can be designated to Aquitanian (Seymen and Aydın, 1972). Solhan volcanites, showing very wide distribution in the area are Upper Miocene in age. They are intercalations of lavas that are continental in origin. Well exposed outcrops of the Solhan volcanites are on the both sides of the East Anatolian fault. In the lower portion of the Solhan volcanites, gravels collected from the conglomerate have fossils thought to be Upper Miocene in age. Zırnak formation is another stratigraphic unit with widespread distribution in the area. It is Pliocene in age. Samples collected from coal layers of the Zırnak formation have fossils of Middle-Upper Pliocene (Nakoman, 1968). The lower contact with the underlying Solhan volcanites is probably unconformable. Quaternary in Bingöl-Karlıova area comprises Boran formation, landslides, travertine and unnamed ancient and modern alluvial deposits (Fig. 3). The Boran formation bears the characteristics of alluvial fan deposits with a lateral extend limited to the Karlıova plateau. The Boran formation is old Quaternary, thus is assumed to be Pleistocene in age.

As inferred from the overall sequence, the area of Bingöl-Karlıova has become continent after the Lower Miocene. The oldest strata of the neotectonic episode is the Solhan volcanites. They are intercalated with nonmarine deposits. The presence of Pliocene Zırnak formation on both sides of the East Anatolian fault indicates that faulting occured after the deposition of Zırnak formation. Presence of the Pliocene Boran formation indicates the opening of the Karlıova basin.

Muş area

The area of Muş is in the southeast portion of the study area. This area is interesting because age relations between the widespread Tertiary deposits can clearly be seen. The Muş area, with an approximate east-west extention is bordered in the south by Bitlis mountain, in the east by Nemrut volcano, and in the north by Bingöl volcano and Hamurpet uplift (Fig. 2). The stratigraphy in the area is showrt; in Figure 4. The Lower Miocene Adilcevaz limestones have gradual contact with underlying Aquitanian Abulbahar formation (Ünal, 1970), and has very widespread distribution in the northern part of the Muş plain. The Adilcevaz is dominated by limestones in the west, clayey and sandy limestones in the east of Muş plain. The limestones of the Adilcevaz characterize the Burdigalian.

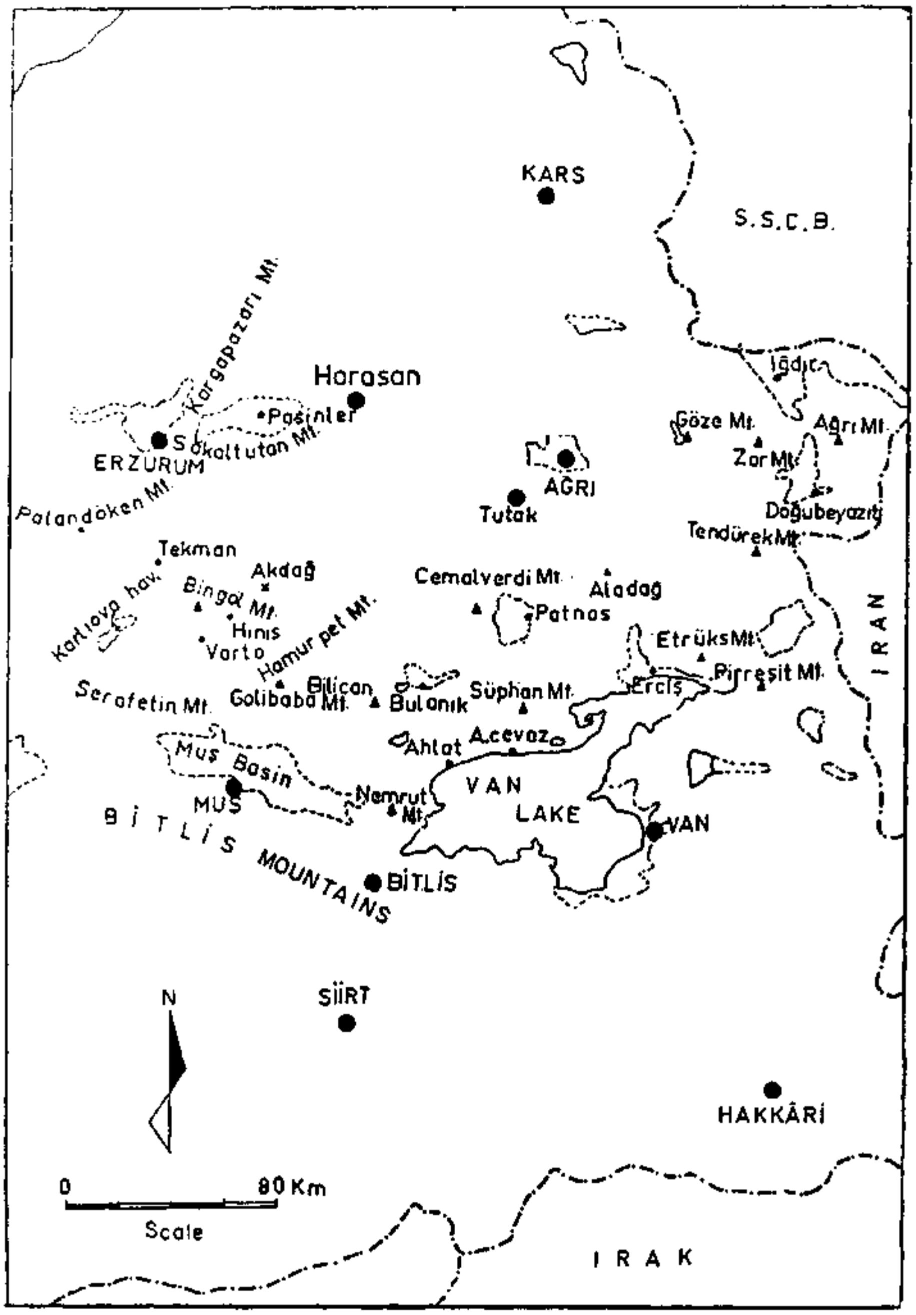

Fig. 2 - Simplified map for locating the basins and areal distribution of Quaternary rocks in the Eastern Anatolia.

The Solhan volcanites, unconformably overlying the Adilcevaz limestones, have wide distribution in the western part of the area (Yılmaz et al., 1986). Type sections for the Solhan volcanites are along the both sides of the Murat river valley. Lower Miocene gravels have been found in the lower portion of the Solhan volcanites. The Solhan volcanites are different from the Middle Miocene sediments of the Ahlat-Adilcevaz area, thus the Solhan volcanites are thought to be Upper Miocene in age.

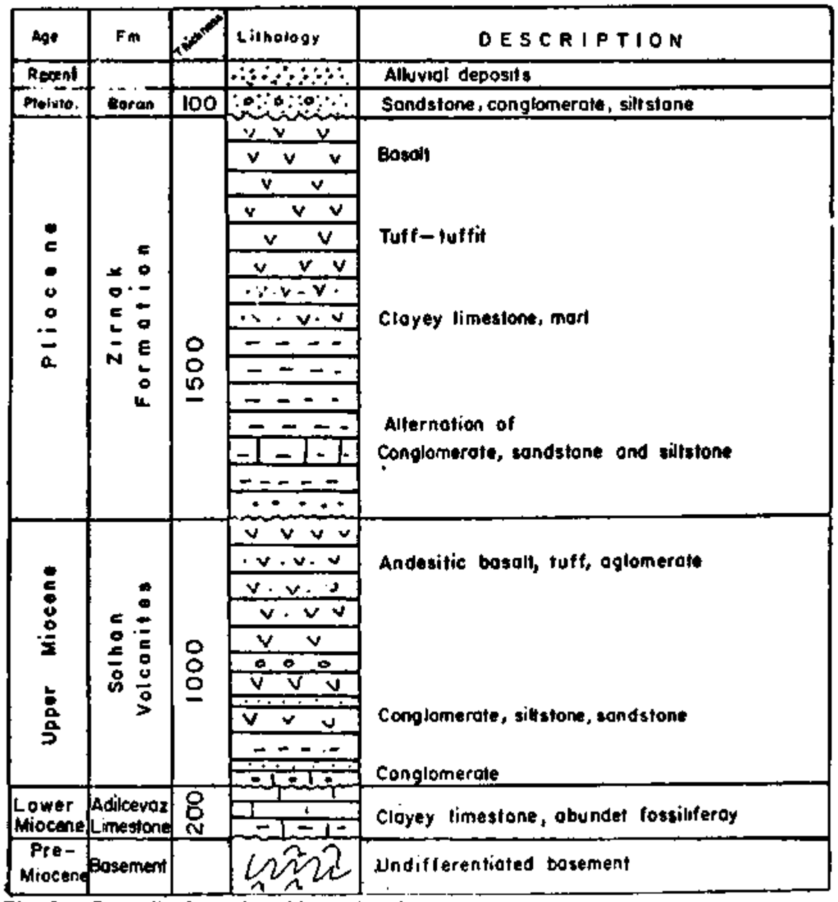

Fig. 3 - Generalized stratigraphic section for Bingöl-Karlıova area.

The Zırnak formation, lying in the Solhan volcanites with possible unconformity, has a widespread areal distribution in the northern part of the Muş basin. The Zırnak is deposited in nonmarine environment and has some limestone interbeds that are characteristic for lake deposits. Fossils collected from different layers of the Zırnak formation are designated to Upper Miocene to Pliocene time span.

The Anzar formation, which was formed in the boundaries of today's Muş basin (Yılmaz et al., 1986), consists mainly of deposits of a lake, that was present at the beginning of Quaternary. The possible age for the Anzar formation is Pleistocene.

Ignimbrites, whose distribution is conformable with the drainage pattern of the Murat river and adjacent streams, are thought to be Pleistocene in age. Basalts, tuffs and dasites of Nemrut volcano are designated to Quaternary (Güner, 1984).

According to the overall sequence of strata, the present sea regressed from the area towards the end of the Lower Miocene. Middle Miocene strata has not been found in the area. There is an angular unconformity between the strata of neotectonic and paleotectonic episode. The presence of widespread volcanites in the area, and the unconformity between the Solhan volcanites and overlying Zırnak formation indicate that the tectonism had a great effect in neotectonic episode.

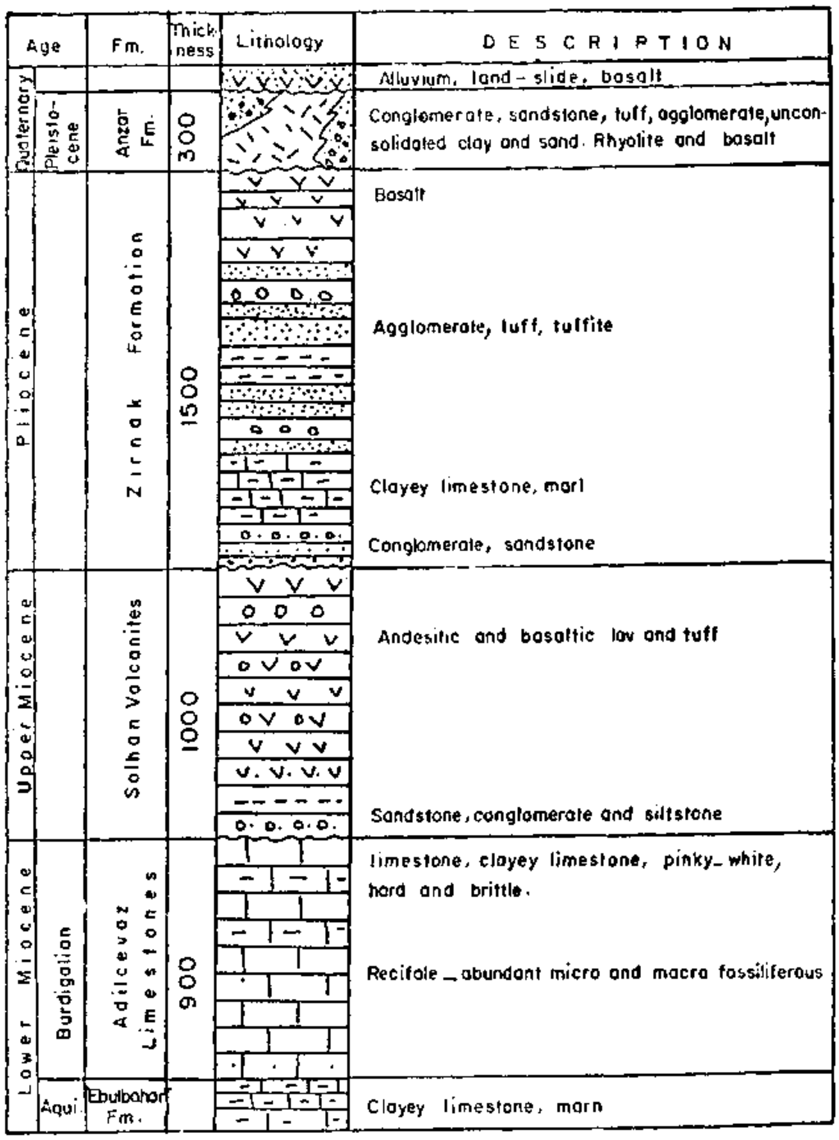

Fig. 4 - Generalized stratigraphic section of Muş area (adapted from Yılmaz et al., 1986).

Ahlat-Adilcevaz area

In the northern part of the Van lake, there are widely distributed volcanites of neotectonic episode (Fig. 2). Tertiary deposits interstratified with these volcanites are well exposed at some localities between Ahlat and Adilcevaz. From bottom to top, the lithostratigraphic units in the area are the Lower Miocene Adilcevaz limestone, Middle-Upper Miocene Develi formation with Aktaş conglomerate at its very base, Pliocene Çukurtarla limestone and unnamed volcanites (Fig. 5). Age of the Pliocene for the Çukurtarla limestone is still debateble. The stratigraphic units in the area were first named by Demirtaşlı and Pisoni (1965).

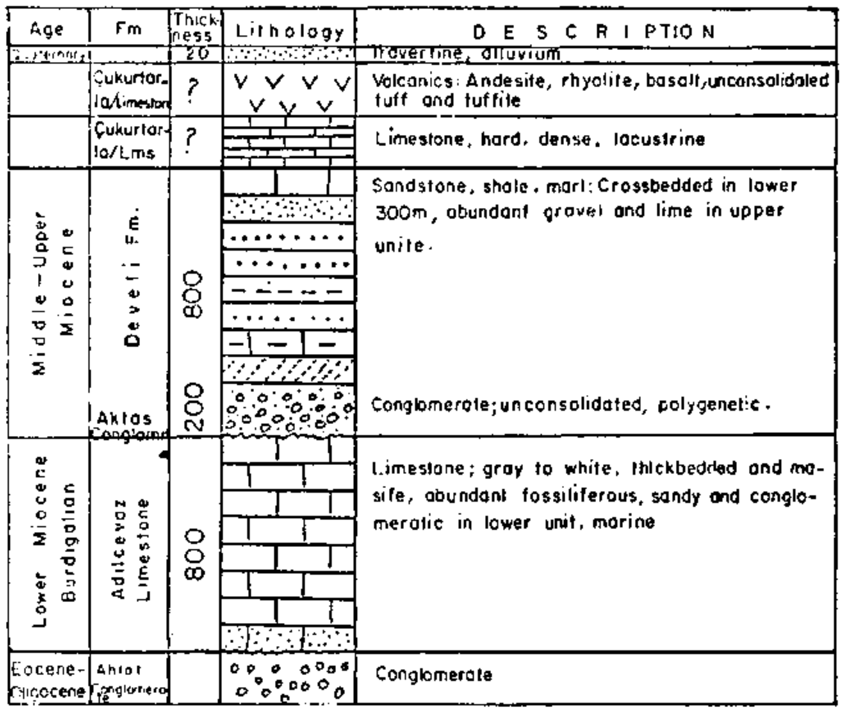

Fig. 5 - Generalized stratigraphic section of Ahlat-Adilcevaz area (modified from Demirtaşlı and Pisoni, 1965).

The Adilcevaz limestone is exposed at the outcrops in the south portion of the area. It is in marine fades and shows the same fades characteristics around Erciş, in the east, and Muş, in the west. The Aktaş conglomerate, which comprises the lowest portion of the Adilcevaz formation, do not have any fossil to designate it to a particular age. The Aktaş contains the gravels of the underlying Adilcevaz limestone, thus it is younger than the Lower Miocene. Owing to gradual contact between the conglomerate and the overlying Middle-Upper Miocene Develi formation, the age for the Aktaş conglomerate should be older than Middle Miocene. With respect to the fossils found in the Develi formation, it is Middle Miocene in age. The Develi formation gradually pass upward into Çukurtarla limestones. The Çukurtarla limestone is in lake facies and has distribution in the north of the basin. Along with the overlying volcanites, the Çukurtarla limestones are thought to be Pliocene in age. In terms of facies characteristics, the Çukurtarla limestone resembles the Zırnak formation of the Muş basin.

The present sea regressed from the Ahlat-Adilcevaz during the Middle-Upper Miocene time. The facies gradation from marine deposits of the Develi formation up into lake limestones of the Çukurtarla formation indicates that the marine environment was gradually replaced by the lake environment. Such a regression is probably related to the local uplifting resulted from neotectonic events.

Karayazı-Tekman area

The Karayazı-Tekman basin nearly extends from east to west. It is bounded in the north by an east-west trending ophiolitic ridge, in the south by Akdağlar, that is made up of the metamorphic basement rocks, in the west by volcanites of neotectonic episode. In the east, the basin opens up into Zırnak basin. The basin has a different position among others because of the unconformity between different stratigraphic units, widespread existence of the deposits of neotectonic episode and presence of definitive basin boundaries in the north and south at the beginning of neotectonic episode. Figure 6 represents the observed strata in the basin.

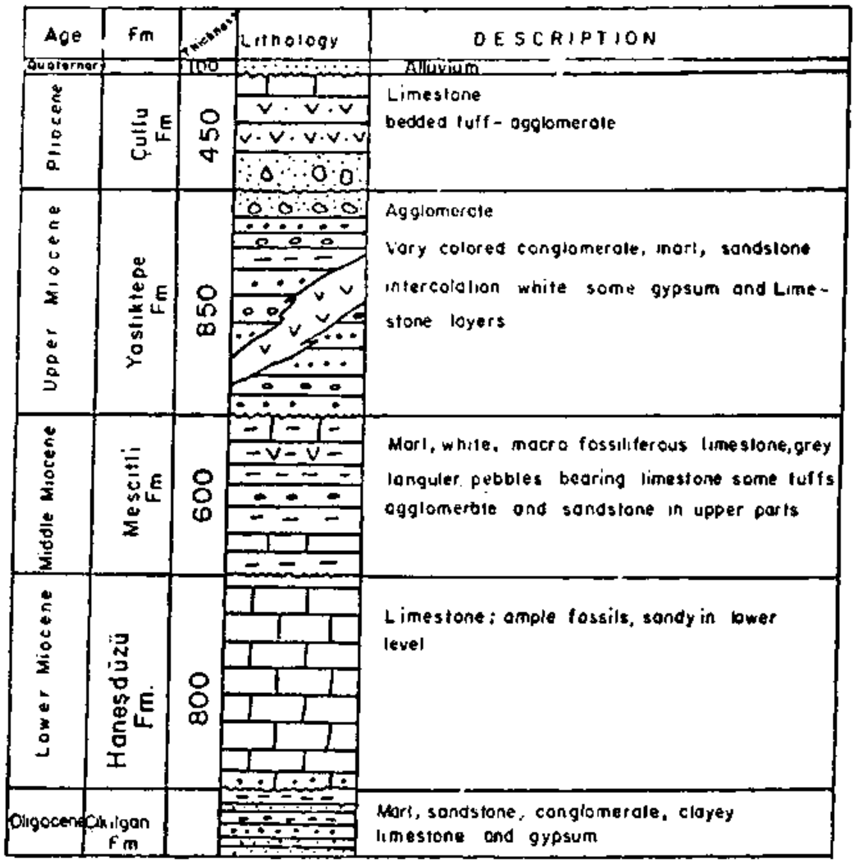

Fig. 6 - Generalized stratigraphic section of Karayazı-Tekman area (adapted from İlker, 1966 *b*; Erdoğan, 1966 and Tanrıverdi, 1977).

In the Karayazı-Tekman area, the Lower Miocene is characterized by Haneşdüzü formation. The name «Haneşdüzü» which unconformably sits on the Çığılgan formation, was firts used by İlker (1966b). It consists of marly, marine limestones with some breccia and has characteristic fossils of Burdigalian. The Haneşdüzü is unconformably overlain at the top by Mescitli formation. The Mescitli appears to have features of the lake deposits and has intercalations of volcanites towards its top. Fossils collected from the Mescitli formation are designated to Middle Miocene. Up in the section, Mescitli formation is conformably overlain by Yastıktepe formation that consists of conglomerate, with varigated color, sandstones and marl interbeds. The name «Yastıktepe» was first used by Akkuş (1965). The upper portion of the Yastıktepe formation is intercalated with the volcanoclastic rocks. Fossils of Upper Miocene are found in the Yastıktepe formation. Yastıktepe deposits were thought to be accumulated in lagoonal environment, following the regression of the Lower Miocene sea.

Pliocene in the area is characterized by the Çullu formation that conformably lies on the Yastıktepe formation. The Çullu formation consists of interbedded agglomerate, tuff and limestones. The limestones show the characteristics of the lake deposits. At some localities, Quaternary rocks of alluvium and basalt lavas are present.

In the area, general regression probably occured in Middle Miocene because the shallow marine sediments of the Lower Mescitli pass upward into shallower lagoonal sediments. The Lower Miocene Haneşdüzü formation is present in basins on the north and south. This indicates that the basin boundary was not definite in the Lower Miocene; on the other hand, the formation of Upper Miocene-Pliocene strata in the Karayazı-Tekman basin, but not in the adjacent basins indicates that the Karayazı-Tekman basin was present in the Upper Miocene.

Hınıs area

The area of Hınıs, which is located in the north of the Muş basin, is bounded in the west by Bingöl volcano, in the north by Akdağ metamorphics, in the south by Hamurpet uplift (Fig. 2). Within these boundaries, the Hınıs is a seperate basin and directly opens up into the Zırnak basin in the east. In the Hınıs basin, the stratigraphic relations between neotectonic strata and underlying paleotectonic strata are well seen. This basin has a seperate importance in terms of establishing the relations between the Muş basin in the south and the Karayazı-Tekman basin to the further north.

Neotectonic episode deposits of the Hınıs basin disconformably overlie the Lower Miocene marine carbonates of Güzelbaba limestone (Fig. 7). The Güzelbaba limestone gradually passes downward into Oligocene Aktuzla marls. On the south part of the Hınıs basin, the exposures of the Güzelbaba limestone are at the near Niftlik area around the Hamurpet uplift and Divanhüseyin village along the Hınıs-Varto state road.

The first deposits «Alibonca formation» of the neotectonic episode consist mainly of conglomerate, sandstone, tuff and mudstone. These deposits with marine character also include basalts and trachytic lavas. The gravels in the conglomerates have fossils of the Lower Miocene, thus the Alibonca formation should be Upper Miocene in age.

The Alibonca formation is unconformably overlain by the Pliocene Zırnak formation. Up in the section, the Zırnak formation is covered by unconsolidated Quaternary deposits. The Zırnak formation with various lithology mainly consists of marl, limestone, tuff, tuffite, basalt and andesite lavas. These basalts and andesites are thought to be products of Bingöl and Golibaba volcanoes.

Zırnak area

The Zırnak area is not a seperate from the Hınıs basin. But, because of the very widespread extention of the Hınıs basin to the east, both the area and the stratigraphy of the Zırnak will be considered seperately, Such a way of introduction will perhaps help better understanding the lateral changes in the stratigraphic units. The Zırnak basin is bounded in the north by Akdağ, in the east by Cemalverdi mountain, in the south by Bilican mountain (Fig. 2). The type section of the Zırnak formation, with the characteristic, fossils, is in this area.

In the Zırnak area, Aquitanian Aktuzla formation at the base gradually pass upward into the Burdigalian Güzelbaba limestones with marine character. There are andesites on the Güzelbaba limestones exposed at the outcrops in the northeast and southwest Zırnak basin (Fig. 8).

The Güzelbaba formation is unconformably overlain by the Upper Miocene Alibonca formation. The Alibonca formation consists of alternated beds of clayey limestones, conglomerate, mudstone and basalts. The name «Alibonca» was first used by İlker (1966a) in this area. There is no fossil in these nonmarine strata. Inferred age for the Alibonca is probably the Upper Miocene.

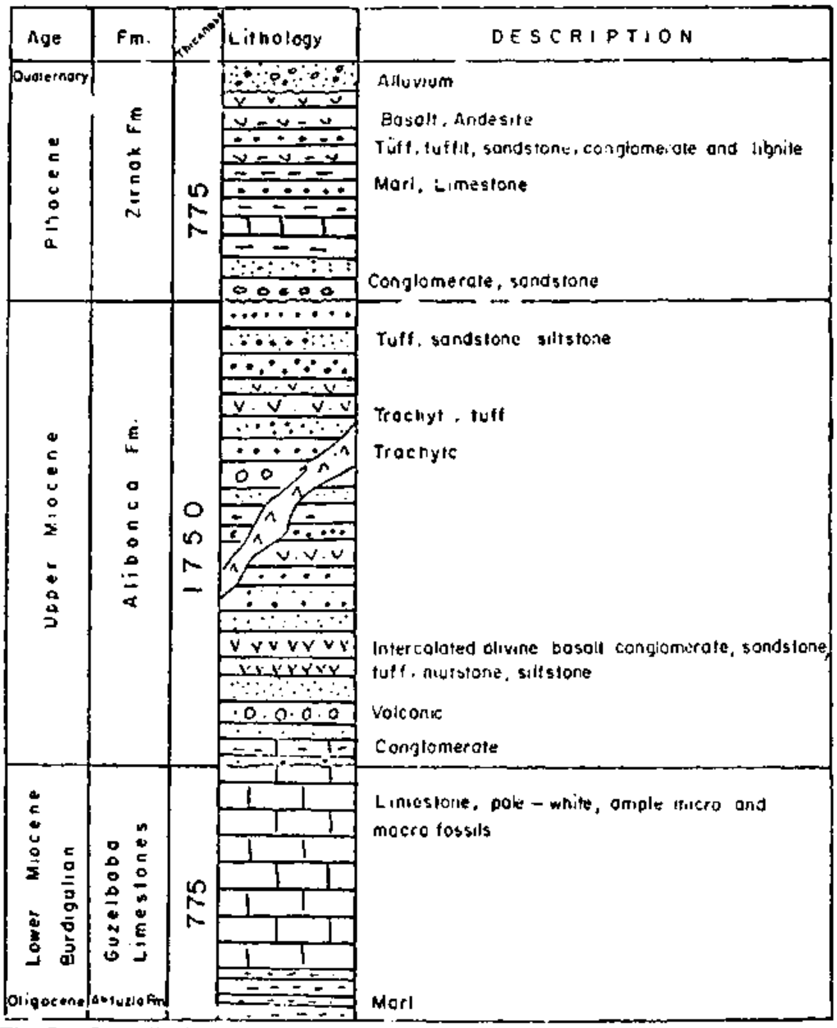

Fig. 7 - Generalized stratigraphic section of the Hınıs area (collected from Tütüncü, 1967; Özcan, 1967; Erdoğan, 1967 and Sungurlu, 1967).

As seen in the other areas, the Zırnak formation starts at the lower portion with the same lithology of sandstone, conglomerate and clayey limestone, and continue upward with sandstones and clayey limestones with coal layers. The very upper portion of the Zırnak formation is dominated by conglomerate, limestone and travertine-like limestones. The name «Zırnak» was also first used by İlker (1966a) for such deposits whose type section is around the Zırnak village. The limestones in the Zırnak formation are in lake facies and include the characteristic fossils of Pliocene (Dacian). The Pleistocene Karaali formation, which consist of conglomerate and sandstone, was also first named by İlker (1966a) in this area.

Erzurum-Pasinler-Horasan area

In the Eastern Anatolia, the Erzurum-Pasinler-Horasan basin is the northeastern basin where the deposits of the neotectonic episode are present. Nearly the east-west extending basin of the Erzurum-Pasinler-Horasan is bounded in the north by Kargapazarı mountains. Today, Erzurum, Pasinler

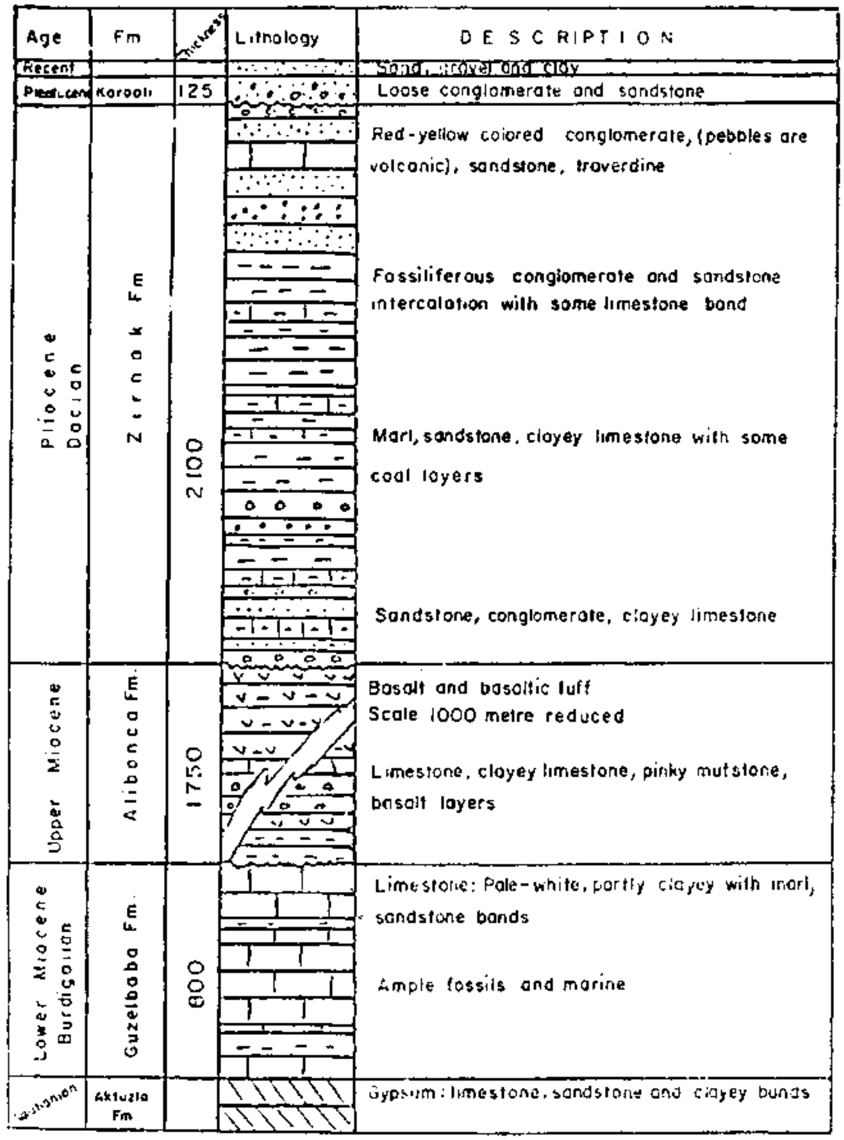

Fig. 8 - Generalized stratigraphic section of the Zırnak area (collected from Birgili, 1968; Şenalp, 1966 and İlker, 1966*a*).

and Horasan are seperate basins, however, there are strong evidences that they were interconnected at the beginning of neotectonic episode. As a result of effective neotectonic deformation and neovolcanism, these basins were become seperated. Because they have a similar neotectonic evolution, here these three basins are analized together (Fig. 2).

The stratigraphic sequence in this area is as in following (Fig. 9); Oligocene Çığılgan formation at the base is unconformably overlain by marine limestones of the Haneşdüzü formation that gradually passes upwards into the Mescitli formation. The Mescitli formation was subdivided into two members of marl and agglomerate. The Mescitli formation is Middle Miocene in age and is unconformably overlain by nonmarine deposits of Upper Miocene Yastıktepe formation. Pliocene in this area is characterized by Horasan formation that unconformably lies on the underlying units. The Horasan formation consists mainly of repetafive beds of sandstone, conglomerate, clayey mudstone and marls. The name «Horasan» was first used by Rathur (1966). Time stratigraphic equivalent of the Horasan formation in the area is the Gelinkaya formation. The Gelinkaya consists of tuff, tuffite, basalt lavas and interbeds of lime to clayey rocks. The youngest strata in the area is the Plio-Quaternary Karatepe volcanites characterized by basalt lavas.

As inferred from the overall stratigraphy in the area, the present sea probably regressed in the Middle Miocene. The present nonmarine deposits of the Upper Miocene are related to the neotectonic episode but Pliocene deposits are not because the Pliocene deposits are different from those in other basins, therefore they are unique to the Erzurum-Pasinler-Horasan basin. This indicates

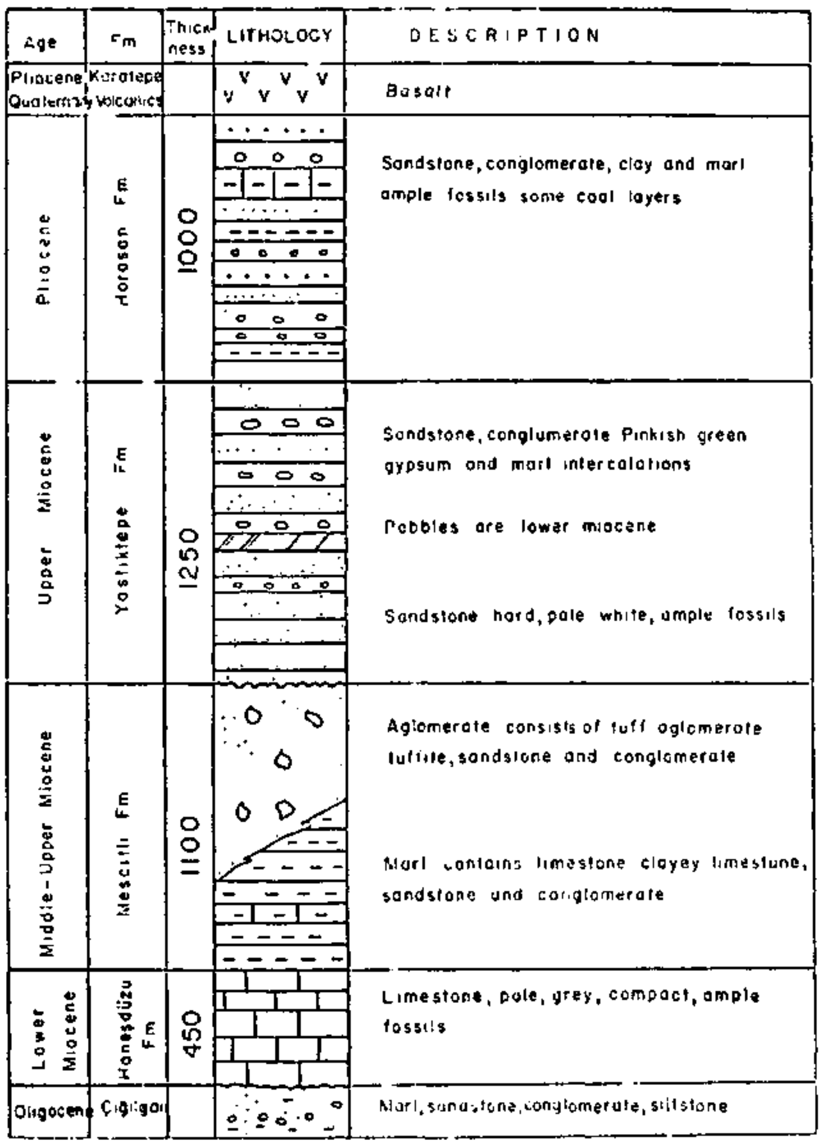

Fig. 9 - Generalized stratigraphic section of the Erzurum-Pasinler-Horasan area (Akkuş, 1965; Toker, 1965; Rathur, 1966 and Tütüncü, 1966).

that the marginal area of the basin was probably uplifted thus the connection between the Erzurum-Pasinler-Horasan and other basins were cut. Upwards in the section, sediments alternated with lavas do not have widespread areal distribution. This probably indicates that the basin was more effected by the compressional stress that resulted in subsequent narrowing of the basin.

Kağızman-Tuzluca area

The Kağızman-Tuzluca basin with approximate east-west extend is in the northeastern part of the Eastern Anatolia. It lies between Kars plateau in the north, Tendürek and Ağrı mountains in the south (Fig. 2). This basin is characterized by very thick Pliocene strata. Absence of the Lower Miocene marine deposits is important because that assist to identify the paleogeographic boundaries at the beginning ofthe neotectonic episode. The Kağızman-Tuzluca basin has also another importance because it is a pull-apart type of basin that began to evolve in the neotectonic episode and is still effected by the active strike-slip faults.

The youngest strata of the paleotectonic episode in this basin is Eocene in age and consist of conglomerate, sandstone, marl, siltstone, and limestone. The deposits of neotectonic episode are in Pliocene in age and comprise at the lower portion repetative beds of salt, gypsum and mudstone, and in the upper portion conglomerate and sandstone.

The Pliocene Tuzluca formation was first named by Şenalp (1969). Şenalp (1969) suggested that some beds in the Tuzluca formation can be assigned to shallow marine or lake deposits. The Tuzluca formation in this basin is overlain by unnamed Pliocene volcanites of mainly tuff, tuffite and basalts. As seen from overall sequence of Eocene-Pliocene, this area has stayed as a land on which no sediment was deposited. During the Lower Miocene the northern boundary of present peneplaine did not reach to the Kağızman-Tuzluca area. In the Pliocene, abundant volcanoes were present and divided the basin into its components. At the very beginning of the Pliocene, Gürbulak and Kağızman-Tuzluca basin were together but they were become seperated by volcanoes such as those above (Fig. 10).

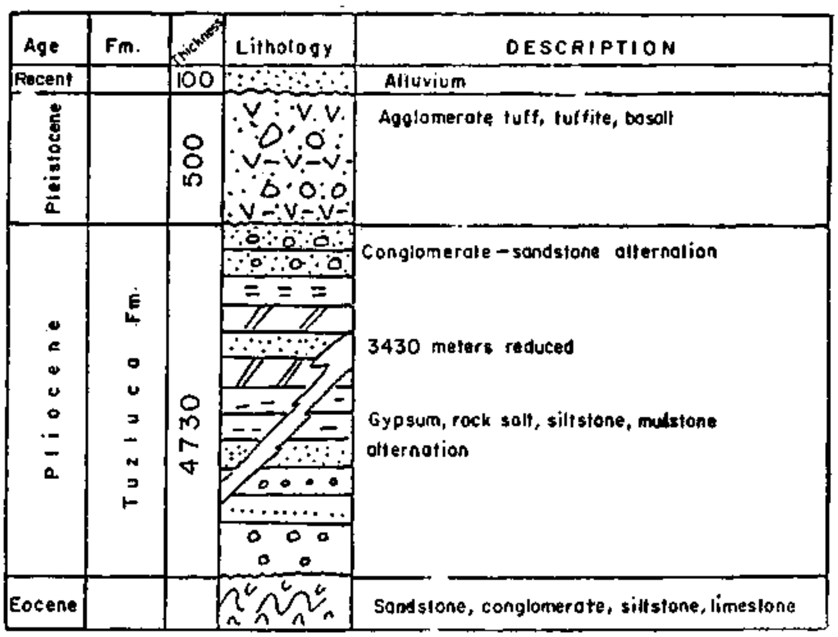

Fig. 10 - Generalized stratigraphic section of the Kağızman-Tuzluca area (taken from Havur, 1968; Şenalp, 1969 and Yurdakul, 1971).

GENERAL FEATURES OF REGIONAL STRATIGRAPHY

On the basis of stratigraphy in different areas or basins, the general stratigraphic features of the Eastern Anatolia will be introduced below. In this chapter, in order f6r the better understanding of neotectonic stratigraphy, the top unit of paleotectonic episode will be described first

Lower and Middle Miocene

In the Eastern Anatolia, the deposits of the latest paleotectonic episode are the Lower Miocene in age, and show the characteristics of the marine strata in the region. They generally bear features of a reef environment and have characteristic fossils for Burdigalian. The sedimentation is continuous from Oligocene up to Aquitanian. But in many areas, the Lower Miocene strata unconformably sit on the underlying units. The main lithology types for the Lower Miocene strata are generally the limestones and the clayey limestones. Towards the northern Anatolia, elastics associated with the limestones are also present and from place to place elastics are predominant. In such places, interbeds of evaporites are also seen. The Lower Miocene sea regressed from the region towards the end of the Lower Miocene.

In the Eastern Anatolia, Middle Miocene strata are found in restricted areas. They are in marine facies and show the characteristics of a regressive sequence. Regression beginning towards the end of the Lower Miocene finally led to the deposition of lagoon sediments. The Middle Miocene strata consist mainly of clayey limestone, marl, sandstone and siltstone. Fossils are rare in the Middle Miocene deposits, thus characteristic fossils are not found. The Middle Miocene has a pradual contact with the underlying Lower Miocene.

Despite the Lower-Middle Miocene strata is well laced with marine facies, their nonmarine equivalents are not found yet. The evaporite-bearing nonmarine type of deposits are present in northern part of the region (these deposits are shown as Oligocene on the geologic map of the Kars section with scale of 1:500 000). Some parts of these deposits are considered as Miocene strata, and if so, they correspond to the Miocene marine deposits. There are also coavel volcanites associated the Lower-Middle Miocene deposits. They mainly consist of lavas and pyroclastic rocks. Within these volcanites, basalt, trachyte, andesite and pyroclastic rocks are predominant. They should be the Lower Miocene in age because in localities between Patnos and Tutak, these volcanites are interstratified with the Lower Miocene limestones. This is true because on the south of Taşlıçay town of Ağrı, and on the north of Aladağ, from place to place these volcanites underlie the Lower Miocene limestones. In also some places they cook the very bottom portion of limestones and causes them to alter. They are mostly overlain by Middle Miocene deposits; thus reexistance of these volcanites has probably continued until the end of the Lower Miocene. The Lower Miocene volcanites cannot be distinguished across the region, thus their detailed analysis have not been done yet. However, under the light of regional tectonic, these volcanites are assigned to the island-arc type of volcanites in origin (Şengör and Yılmaz, 1981). The morphological features of the Lower Miocene volcanites are eroded, therefore no relict morphologic features are well seen. The trends of the faults are not parallel to the general trend of the volcanoes, instead they cut across the volcanoes. Such evidences show theexistance of volcanoes before the neotectonic activities.

Upper Miocene

In the Eastern Anatolia, the Upper Miocene strata begins with sandstone, siltstone and conglomerate, and upward continues with clayey limestone, tuff, agglomerate and volcanic lavas. The Upper

Miocene rocks unconformably sit on the underlying strata. The followings are the evidences that show the presence of such an unconformity: (a) the presence of the basal conglomerate at the very base of the Upper Miocene, (b) an angularity between the conglomerate beds and the underlying Lower-Middle Miocene strata, eventhough it is not very discernable, and (c) the presence of the Lower Miocene fossils in the gravels of these conglomerate, (d) without Lower-Middle Miocene in the most localities, the Upper Miocene directly sits on the older strata. However in some places, the Upper Miocene conformably overlie the Lower Miocene strata. But the Middle Miocene is not determinate in between. Therefore, there should be an hiatus between the Upper Miocene strata and those of the underlying units. The Lower Miocene strata shows the characteristics of the marine facies whereas the Upper Miocene units are in nonmarine facies. In addition, the Upper Miocene volcanites are different from those of underlying units.

Fossils in the Upper Miocene are rare, and none is characteristic for age determination; the present ones are lamellibranch, gastropoda and plant fragments. The age of the Upper Miocene is not definite, thus stratigraphically inferred.

Pliocene

In the Eastern Anatolia, Pliocene strata consist mainly of sandstone, siltstone, marl, conglomerate, tuff, tuffite, agglomerate and lake limestones. The limestone beds contain very fossilliferous levels; some beds contain large amount of shells thus the term «coccina» may be appropriate for such limestones. Fossils collected from these limestones are characteristic for Pliocene (Dacian). There are some economically important coal layers interbedded with the Pliocene deposits. The age of Pliocene was also supported by determined spors in the samples from these coals. The present volcanic rocks in the Pliocene are the basalts, andesites or trachy-andesites. They unconformably lie on the underlying units.

Pleistocene

Pleistocene in the Eastern Anatolia is characterized by nonmarine deposits with mostly lake or fluvial in character. Along with the unconsolidated clay, sand and gravels, well indurated sandstone, gravelstone and siltstone are the main deposits. Eventhough some fossiliferous levels are present, there is no characteristic fossil present in the Pleistocene rocks. The configuration of the Pleistocene rocks reflects the drainage pattern of ancient lake and streams. The Pleistocene strata contain the gravels of older rocks and unconformably overlie them. Depending up on the type of extrusive materials scattered around the volcanoes. The Pleistocene strata can contain tuff, tuffite, agglomerate, basalt, andesite and rhyolite lavas. In addition, Ağrı, Tendürek, Süphan and Nemrut volcanoes showed activities in the past and acted as a seperate source for the Pleistocene rocks. In places where there was not any fossils, age determination was made with respect to stratigraphic position of the strata in the section.

The Anzar formation was attributed to the Pleistocene in the Muş basin where the Pleistocene rocks give one of the best exposures. The Anzar is about 300 m thick in the Muş basin. In the Hınıs-Zırnak area, however, the Karaali formation forms the Pleistocene strata, and is in the order of 125 m thick. In the localities around the Doğubayazıt-Kars area, unnamed Pleistocene is 500 m thick. In the Tutak-Patnos area and around Van lake, Pleistocene show the characteristic of the lake deposits. 200 m thick Pleistocene deposits in the Tutak-Patnos basin was earlier attributed to the Lower Quaternary.

Holocene

Holocene strata mainly covers modern stream and lake deposits. Minor amount of landslides, glacier deposits and morens are also present in the Holocene strata. From place to place the modern deposits generally has gradual contact with the Pleistocene deposits. At some outcrops, the Holocene deposits are not even be distinguished from the Pleistocene deposits. At such localities, flood plain are assumed to be the boundary between the Holocene and Pleistocene deposits.

THE RELATIONS BETWEEN STRUCTURAL ELEMENTS AND THE BASINS FORMED IN THE NEOTECTONIC EPISODE

In the Eastern Anatolia, folds, reverse faults, large-scale extentional fractures, right and left lateral strike-slip faults are formed in the neotectonic episode (Fig. 11). As in Figure 11, in places where strike-slip faults step up in en echelon character, there are some volcanoes existed as a result ofthe effective tectonism. The major structural features on the neotectonic map ofthe Eastern Anatolia are the east-west extending folds, reverse faults, north-south striking extentional fractures, left (northeast-southwest striking) and right lateral strike-slip (northwest-southeast striking) faults (Fig. 12). In the region, pull-apart type of basins are formed between strike-slip faults. Another different type of basins in the Eastern Anatolia are the intermountain basins which generally extend east-west and correspond to the synclines (Fig. 13). Such basins can be bounded on one side by a trust fault. Along with these two major type of basins, there are also basin-like localities between compressional features or along the extentional fractures. But these are relatively small in scale. Thus the basin types in the Eastern Anatolia can be summarized as two types; (a) pull-apart, and (b) intermountain.

As we glanced over the volcanism during the neotectonic episode (Fig. 14), there is a number of volcanoes existed in different time. The existence of these volcanoes is related to the structural elements of the Eastern Anatolia. The nature of the volcanism, however, has changed with time as continental crust has been evolved by the persistent tectonic effect (Yılmaz et al., 1986).

GEOLOGICAL EVOLUTION OF NEOTECTONIC EPISODE

The stratigraphy in each basin, stratigraphic correlation between the basins (Figs. 3,4,5, 7,8,9), present volcanic activities and major structural elements are considered together and combined to analize the geologic evolution of the Eastern Anatolian region.

Towards the end of the Lower Miocene the Eastern Anatolia had a peneplain-like paleomorphologic feature. This peneplain was lying between Bitlis mountain on the south and Tuzluca-Kağızman-Tortum line on the north (Fig. 2). The east and west boundaries of this peneplain were outside of the region. Over the Middle Miocene, the region was compressed under the influence of north-south directed tectonism. This led to the flactuation of the peneplain, subsequent formation of the folds and fractures. Thus the peneplain, covering very large area, was turned into high mountainous area. As a result of this uplifting, the, present sea began to regress from the region. The

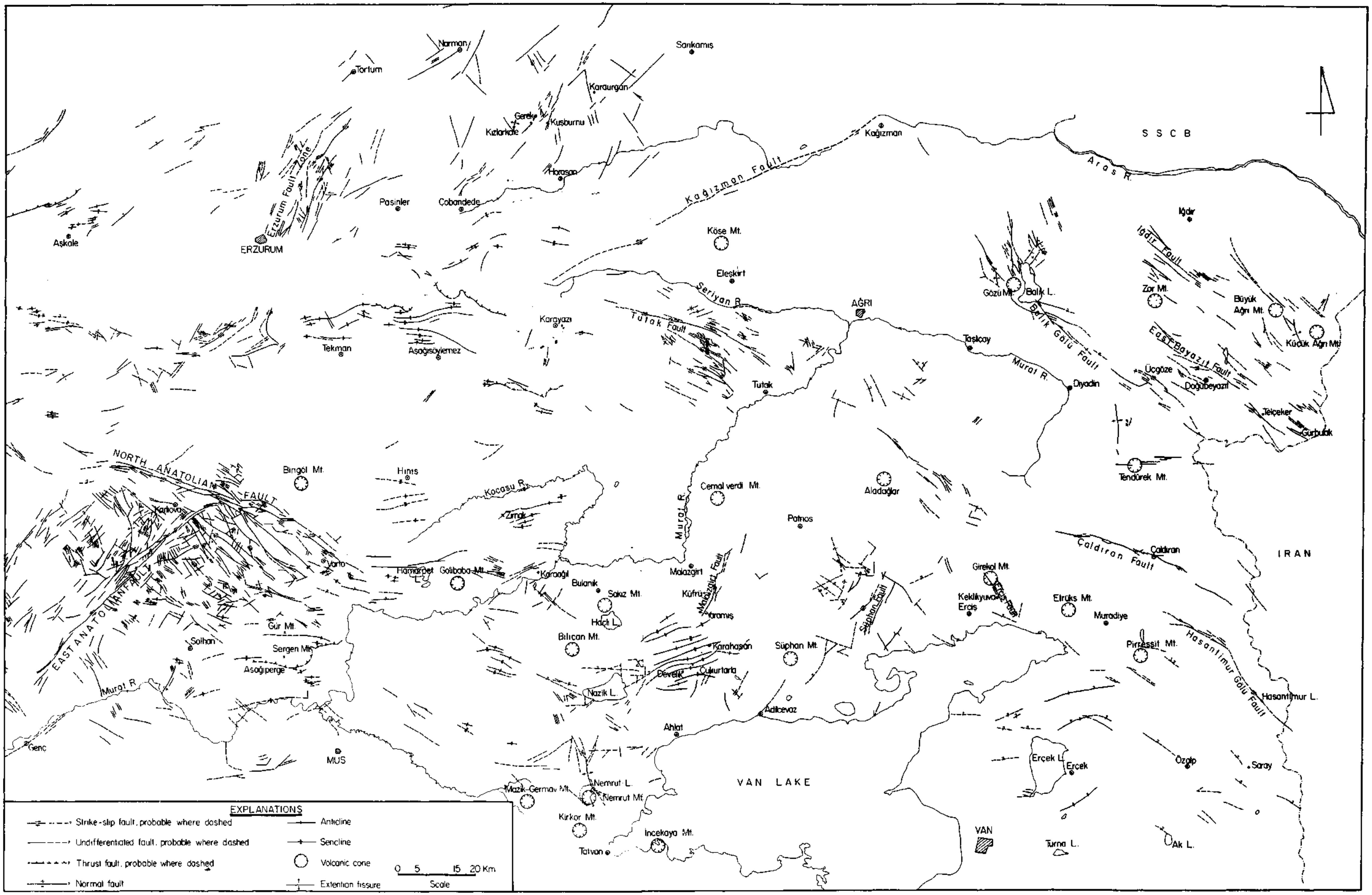

Fig. 11 - Neotectonic map of the Eastern Anatolia.

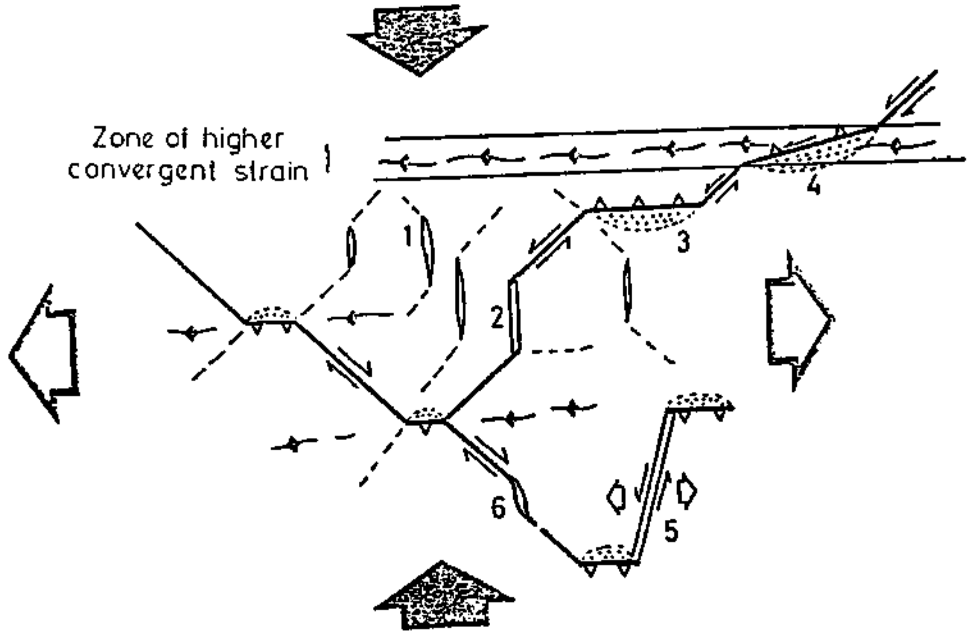

Fig. 12 - Schematized diagram showing the deformational features along the narrowing area in the Eastern Anatolia. Numbers show the basin types evolved in the region. 1-Extentional fracture; 2 and 6-Intermountain basin; 4 and 5-Other basins formed under the control of more than 1 structural effect (Şengör et al., 1985).

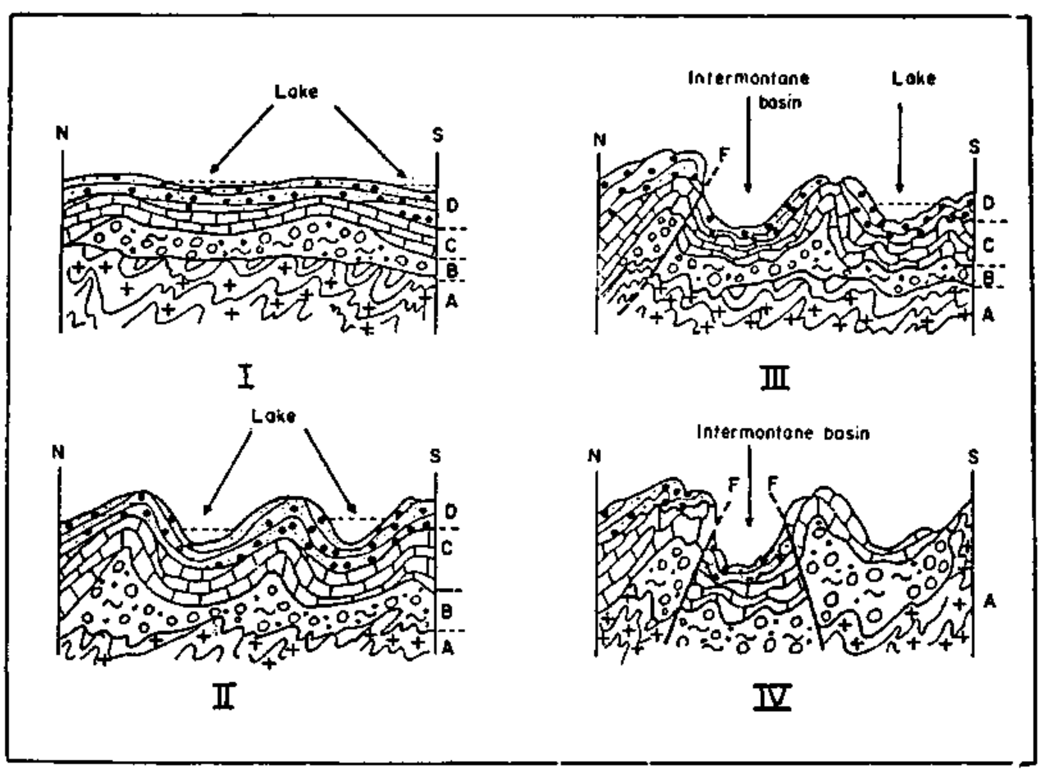

Fig. 13 - Schematized cross section showing the evolution of the intermountain basin in the Eastern Anatolia.

elongated ridges, that formed as response to the uplifting, led to the basins to form in between. The Develi formation in the Ahlat-Adilcevaz, Yastıktepe formation in the Erzurum-Pasinler-Horasan and Mescitli formation in the Karayazı-Söylemez basin are the stratigraphic units deposited right after the formation of the basins. In the Upper Miocene, the present sea completely regressed from the region. As parallel to the regression, the volcanoes started errupting through the extentional fractures. The lakes and streams were also existed at this time. The Solhan volcanites on the south, the Alibonca formation in the Hınıs-Zırnak area, Yastıktepe formation between the Erzurum-Pasinler-Horasan-Tekman-Karayazı are the products of the Upper Miocene and include intercalated materials. The Akdağ and the Sakaltutan mountain ridges played as dividing highs between described basins. New morphologic features, formed at the end of the Upper Miocene, led to the existence of lakes covering very large area in the region. The lake deposits have volcanoclastic materials derived from the Aladağ, the Bingöl and the Pirreşit volcanoes. The lake on the south dies among Bitlis mountain, Akdağ, Ahlat-Erciş and Bingöl-Karlıova. The Zırnak formation in this lake, the Çullu formation between the Akdağ-Sakaltutan mountains on the further north, the Horasan formation between Sakaltutan and Kargapazarı mountains, the Çukurtarla limestone in the Ahlat-Adilcevaz area were deposited. As a result of the compression and the subsequent folding, the continental crust were thickened. This led to the formation of strike-slip faults. As response to this type of faulting, in the northeast part of the region (area surrounded by Tuzluca-Kağızman-Iğdır-Doğubayazıt-Gürbulak) a new basin was formed under the control of obliquely displaced fault blocks. The Tuzluca formation was deposited in this basin. At the same time, Etrüsk, Bilican, Cemalverdi, Gözü, Zor, Köse and Sakız mountain volcanoes were existed. Thus by these volcanoes, present large basins were divided into smaller ones; while Bilican and Sakız together divided Muş-Varto area from the Ahlat-Adilcevaz; Cemalverdi mountain seperated the Patnos-Tutak basin from Malazgirt in the north. First Gözü, later Zor mountains divided Kağızman-Tuzluca from Doğubayazıt-Gürbulak and Iğdır basins. Persisting tectonic regime led to the narrowing of previously compressed basins. Tectonic regime also caused the present ridges to gain more relief. Such an uplifting was associated with the volcanic activities that resulted in more subdivision of mentioning basins. Towards the Middle Pliocene, strike-slip faults are become important structural elements of the region. On the west, Northern and Eastern Anatolian faults were merged. As a result of that, Karlıova and on its further south Bingöl plains were existed. On the north, however, in the Kars area where strike-slip faults come to a proximal position, intensive volcanism caused the formation of Kars plateau. Nearly at the same time, Hamurpet and Şerafettin mountains were existed and gave way to the formation of the new basins in between. Muş-Van basin was shaped up in the Upper Pliocene. Towards the beginning of Quaternary Ağrı, Süphan, Nemrut, Tendürek volcanoes were existed and helped the basins to gain today's configuration. At the beginning of the Quaternary, the Anzar formation in Muş basin, Boran formation in Karlıova, the Karaali formation in the Zırnak area and other unnamed units were deposited. As proved via earthquakes, today's morphologic features have still been altered to a great extent.

As the basins are analized under the light of the major structural elements in the Eastern Anatolia, they may be classified as follows: The Muş-Ahlat-Adilcevaz and the Karayazı-Tekman basins appear to be intermountain type. The Tuzluca-Kağızman-Iğdır-Doğubayazıt-Gürbulak basin, which was first divided by Gözü, second by Zor and third by Ağrı mountain, can be considered as the pullapart types. The Erzurum-Pasinler-Horasan basin is an intermountain basin that was formed under the supplementary influence of the left lateral strike-slip faults. The Karlıova-Bingöl basin was formed under the control of both left lateral (Eastern Anatolia) and right lateral (Northern Anatolia) strikeslip faults. In fact, the Hınıs and the Zırnak area, each of which is a seperate basin but together, it is a wide basin formed under the control of several structural elements.

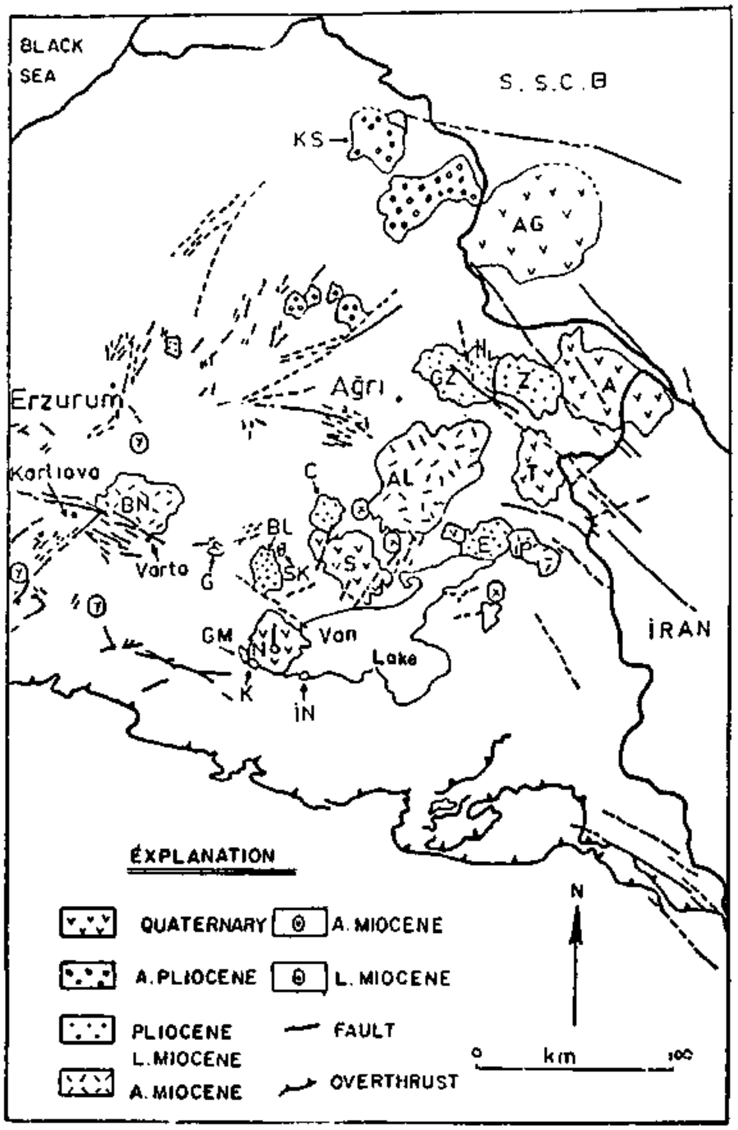

Fig. 14 – Existed volcanoes during the neotectonic episode in the Eastern Anatolia. AG- Alagöz mountain; KS- Kısır mountain; BN- Bingöl mountain; T- Tendürek mountain; A- Ağrı mountain; AL- Alagöz mountain; İN- İncekaya mountain; C- Cemalverdi mountain; P- Pirreşit mountain; E- Etrüsk mountain; S- Süphan mountain; SK- Sakız mountain; BL- Bilican mountain; G- Golibaba mountain; GZ- Gözü mountain; GR- Germav mountain.

This study covers very large area in the Eastern Anatolian region. In a gross sense, an attempt was made to define the position, type and evolution of the basins. In this respect first opinions and approaches that were made up to this point can be a guide for detailed works will probably be made in the future.

THE RELATIONS BETWEEN THE GEOLOGIC EVOLUTION AND THE PETROLEUM POTENTIAL OF THE REGION IN NEOTECTONIC EPISODE

In the Eastern Anatolia, at some localities some oil seepages have been known to us. Such oil seepages are the main reason for many workers to initiate investigations in the region. In preceding chapters, most unpublished reports and cite references are about these works made in the Eastern Anatolia. There are some wells drilled to test the determined structural trapsin the region. Well site reports are not present, thus it is obvious that there is no opportunity for us to make a detail evaluation about the reached results. However we know that in all these drilling projects, the deposits of the paleotectonic episode, especially the Lower Miocene limestones were the main target but the deposits of neotectonic episode were not considered to have potential for oil production. In nowadays, petroleum potential of the Eastern Anatolia has been discussed to a great extent. Some tectonic results derived from this study will be presented below so that we think they may help better understanding the problems of the petroleum-related studies. In some places of the Eastern Anatolia there are five thousand meter thick neotectonic deposits. This fairly thick sediment is due to effective tectonism and volcanism. It is known that during the neotectonic episode, the thermal conductivity was very high in the continental crust (Dewey et al., 1986). This value should be higher around the mouth of the volcano. Also the lavas are both thermally conductive and has a potential to be cap rock. Within the Pliocene rocks some intervals are rich in fossils. Some petroleum seepages are from the deposits of the neotectonic episode or related to them. Within the Pliocene strata, there are some layers of bituminous shale. These direct or indirect evidences show that the deposits of the neotectonic episode in the Eastern Anatolia have a tendency to generate the petroleum. If this idea was correct, the basins developed during neotectonic episode should be taken into account in terms of having a potential for petroleum.

RESULTS

1. The youngest deposits of the paleotectonic episode in the Eastern Anatolia are the Lower Miocene in age. The shallow marine limestones characteristic for the Lower Miocene are widespread in the entire Eastern Anatolia.

2. The neotectonic episode in the Eastern Anatolia began in the Middle Miocene.

3. Middle Miocene is characterized at the lower portion by marine, towards the top by nonmarine deposits. The regressive sequence of the Middle Miocene is related to the compressional tectonic, thus the uplifting of the region.

4. The deposits of the neotectonic episode are continuous from the Lower Miocene to present. They show the characteristic of the nonmarine facies and are formed in intensive tectonic regime associated with volcanism. Local unconformities, lateral gradations and hiatus are usual.

5. In the Eastern Anatolia the peneplain, formed at the beginning of neotectonic episode, is bounded on the south by Bitlis mountain, on the north by Tuzluca-Kağızman-Karaurgan-Tortum. Eastern and western boundaries of the peneplain are the outside of the region.

6. Among the basins formed in neotectonic episode, Muş Ahlat-Adilcevaz and the Karayazı-Tekman basins are intermountain in type; whereas, Kağızman-Tuzluca, Doğubayazıt-Gürbulak and Iğdır basins are pull-apart type. Erzurum-Pasinler-Horasan basin is another type intermountain basin that was also effected by strike-slip faults. Hınıs and Zırnak basins have been stayed under the control of various structural elements. The Karlıova-Bingöl basin, however, was opened under the effect of the strike-slip faults.

ACKNOWLEDGEMENTS

The study was carried out under the project of «Turkey Neotectonic and active fault mapping» conducted by the General Directorate of Mineral Research and Exploration Institute. We thank Directorate of Mineral Research and Exploration for giving us this opportunity to study. We also thank to O. Sungurlu, Y. Güner and A.M.C. Şengör for their appreciable advise and comments.

Manuscript received May 7, 1986

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EVOLUTION AND ORIGIN OF GRANITOID MAGMAS RELATED TO THE CLOSURE OF NEO-TETHYS IN EASTERN TAURUS (TURKEY)

Niyazi TARHAN*

ABSTRACT. — Granitoids cutting the sheeted dyke complex (Göksun metaophiolite), the Elbistan ensimatic island arc volcanoclastic sequence of Neocomian age and the Kabaktepe metamorphites (Bitlis/Pötürge metamorphites) of Paleozoik-Lower Triassic age respectively, outcrop around Afsin-Elbistan-Göksun. The units deposited during Upper Cenonian-Eocene overlie these older units with an angular unconformity starting with a basal conglomerate. In the study area, outcropped granitoids have not been developed during island arc eruptions. These granitoids take place on the collision belts of the subduction zone formed by the N-S directed compressional forces of late Cretaceous started in Lower Cretaceous and increased after Neocomian. Their development is due to the increasing crustal thickness and sinking of the island arc deposits and its basement which is oceanic crust during Coniacian-Upper Santonian. Granitoids are the differentiated products of anatectic magma (diorite, monzonite, syenite, tonalite, granodiorite, granite, alcaline granite and silexite). Which is formed from melting of oceanic (Göksun metaophiolite) and crustal (Kabaktepe metamorphites) rocks together or sepertly at high temperatures. This anatectic magma is named and defined as Afsin magmatism for the first time. However, except these wide spread granitoids intrusions in Eastern Taurus, it was also observed that seme granitic rocks (diorite, monzonite, syenite, tonalite, granodiorite, granite and alcaline granite). Which don't show any intrusive features and magmatic phase. They formed by recrystallization from metamorphites rocks, island arc volcano-clastic deposits and ophiolites (gabbro, sheeted dyke complex and volcanosedimentary rocks) insitu. Granitoids were subjected the low grade regional metamorphism (greenschist facies) probably in Upper Santonian-Campanian.

INTRODUCTION

The granitoids outcrop within the Afsin-Elbistan-Göksun triangle (Fig. 1). Aim of this study is to interpret the petrology and the geotectonic evolution of these rocks and to make a correlation between them and the granitoids outcropping in Eastern Taurus.

Hatay (1966), Polat (1970), Karul (1971), Gökalp (1972), Akkoca and Bahçeci (1972) and Atasever (1978) are the earlier workers in the region. These workers mention the acid plutonic rocks in the study area. According to their view, these rocks intruded during the Eocene time in the area. Perinsek and Kozlu (1983) called the association consisting of ophiolite, island-arc rocks and acid plutonic rocks which are cropping out in the study area «Yüksekova complex». They report that these rocks were thrusted over Mardin complex with the Berit group. Present author, Tarhan (1984) prepared the detailed geological map of the granitoids in the scale of 1:25,000 for the first time. There it is reported that these granitoids are Coniacian-Santonian in age.

GENERAL GEOLOGY

Göksun metaophiolite constitutes the basement of the units cropping out in the study area (Tarhan, 1984). These are overlain by Elbistan ensimatic island-arc volcanoclastic sediments of Neocomian age (spilitic basalt, basaltic andesite, andesite, agglomerate, dacite, alkaline rhyolite and deep-sea pelagic sediments) (Tarhan, 1985). The Kabaktepe metamorphic nappe (Bitlis/Pötürge metamorphites) overlies the Göksun metaophiolite (cumulate, ultramafic rocks of metamorphic tectonite

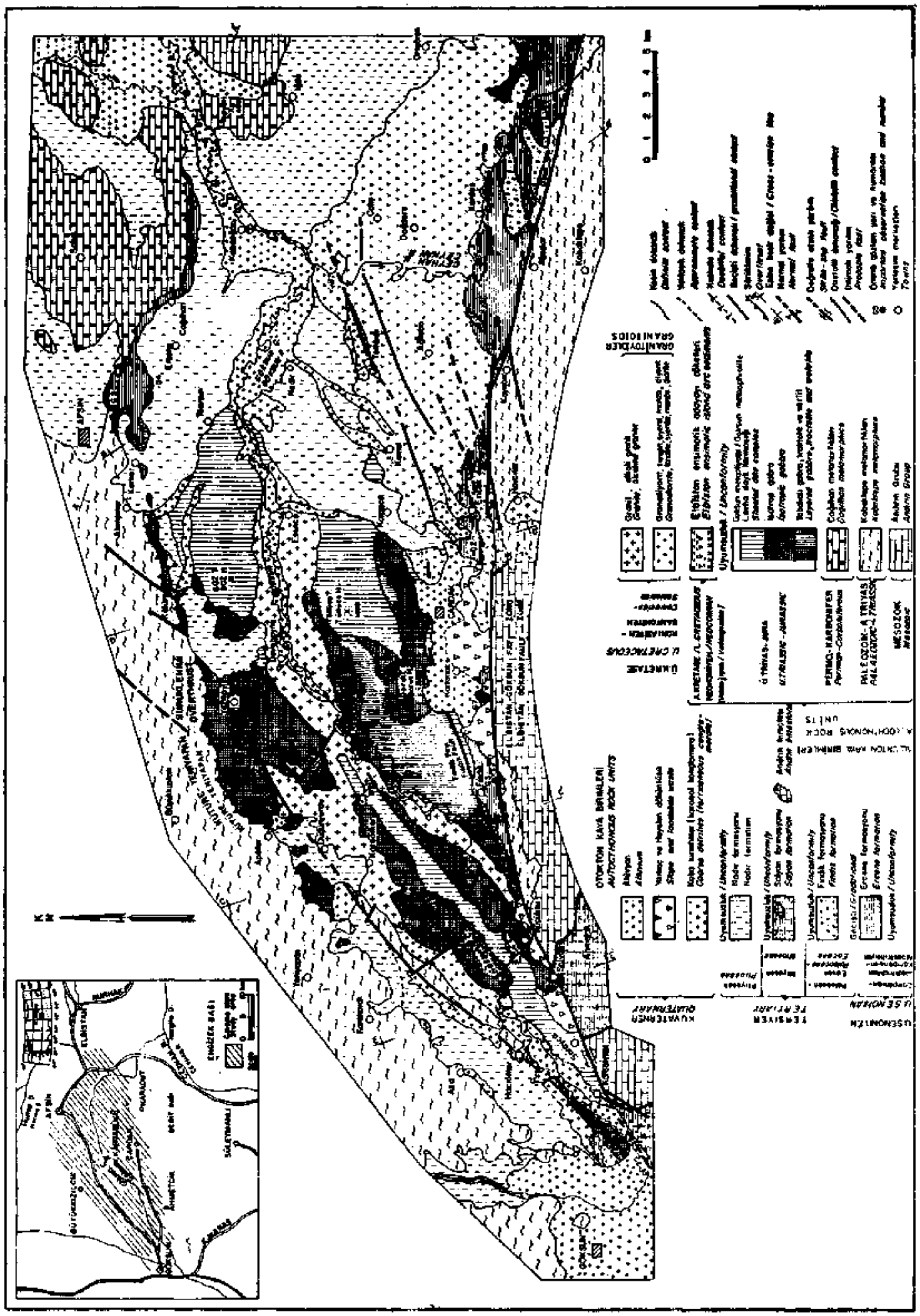

Fig. 1 - The geological map of the Göksun-Elbistan region.

structure, layered and isotropic gabbros and sheeted dyke complex) and the island-arc sequence during the post-Neocomian periods. The Kabaktepe metamorphites (garnet-two mica-gneiss, amphibole-schist, micaschist, quartzite-schist, actinolite-schist, sericite-quartz-chlorite-albite-schist, phyllite, calcschist and marble) form a crushed zone at the contacts with these units (Fig. 1,2,3). The acid plutonic rocks (Çardak association, Tarhan, 1984) cutting the sheeted dyke complex and the Elbistan ensimatic island-arc sequence (intra-oceanic island-arc sequence) and the Kabaktepe metamorphites which overlie these units as nappes, crop out in the area. Isolated diabase dykes cutting the acid plutonic rocks and the other rock units which are cut by the acid plutonic rocks, took place after the intrusion. These dykes extending in NW-SE direction, have variable thickness (30-150 cm), dark-black color and basic composition (Fig. 3). They were subjected to the metamorphism in green-schist facies.

The Ergene formation of Campanian-Maestrichtian age overlies the allochthonous rock units (Göksun metaophiolite, island-arc sediments and Kabaktepe metamorphites) cropping out in the study area. This formation consists mainly of detritic materials derived from these rocks units and begins with a basal conglomerate. It is subdivided into two members represented by flysch and volcanic (andesite, pyroclastic rock and mudstone) facies. The Erçene formation passes into the overlying Fındık formation of Paleocene-Eocene age gradually. The Fındık formation shows flysch character. However, it locally contains lenses, wedges and intercalations of andesitic lavas and pyroclastic rocks. In addition, it also contains pebbles and blocks of ophiolites and metamorphites. The Salyan formation of Oligocene-Miocene age overlies the Ergene formation and the Findik formation with angular unconformities. This formation showing flysch character contains the Andırın limestone olistolithes of Jurassic-Cretaceous age. All the formations cropping out in the study area are covered by the Nadir formation of Pliocene age (Fig. 2,3). The Nadir formation consists of sandstones, claystones, marls, lacustrine limestones, conglomerates, tuffites, mudstones and coaliferous levels (Afsin lignites).

GRANITOIDS

The granitoids are observed in continuous outcrops extending in about E-W direction in the neighbourhoods of Kitiz, Deveboynu and Hacıömer amongst Afsin-Elbistan-Göksun (Fig. 1). In addition, they are seen in discontinuous outcrops near Havcılar and Ambarköy, outside of the study area.

The granitoids cropping out in the study area are composed mainly of diorites, tonalites, granodiorites, granites, alkaline granites and their derivatives, obviously showing gradual transitions to eachother from bottom to top. These intrusive rocks are cut by dykes of aplite, pegmatite and silexite (quartzolite) in form of nets and veinlets. As a rule, monzonites and syenites are scarce, but they occur very widespread in marginal facies. The granitoids contain some fragments of all rock units (15x10 cm to 50x100 cm) as xenoliths (enclaves). Some of these xenoliths are derived from the rocks cut by granitoids and the others are derived from the rocks which granitoids must be originated from. Some xenoliths have preserved their original rock textures, whereas the other ones have gained metamorphic textures and structures due to the effects of metamorphism and have been metamorphosed to amphibolites. The xenoliths show alignments in certain directions.

The granitoids cropping out in the study area were subdivided into two groups and mapped on the basis of this. The first group consists mostly of granodiorites and locally of diorites, monzonites, syenites and tonalites. The second group consists of granites, alkaline granites and silexites. The granites are predominant rocks of this group. The granitoids (first group) usually cropping out in gabbros, in the western part of the study area show gradual transitions to gabbros. By increasing of regional metamorphic grades gabbros pass into amphibolitized gabbros, amphibolites, migmatites, granulites and granitoids (first group). Thus, for this relationship which hasn't been reported in literature, to the present-day, the boundaries between the granitoids and the rocks (ophiolites and metamorphites) from which the granitoids should be derived, were indicated by transitional contacts, for the first time (Fig. 1,2,3). The brief field and petrography descriptions of these rocks are as follows:

Gabbro

The gabbros are layered and show isotropic structures. Levels, lenses and parts of wehrlite and troctolite are locally observed in isotropic gabbros. The gabbros are composed mainly of pyroxenes, plagioclases and opaque minerals (magnetite and chromite), especially in sections where metamorphic effects are lacking or negligible. The effects of cataclasis are insignificant in these minerals.

Amphibolitized gabbro

The amphibolitized gabbros are blackish-green in color. They occur in gabbros and show transitions, the original texture of the gabbro was preserved in the amphibolized gabbros. The plagioclases with albite twinning are partly albitized. The pyroxenes were partly replaced and altered to fibrous actinolite. Increasing of regional metamorphism cause gradual, increase in the hornblende content. In some places, the large xenomorphic pyroxene crystals are partly or wholly altered to the smaller crystal aggregates composed of actinolite and hornblende, in situ. The traces of cataclastic fracturing appear to be more obvious in the crystals of the minerals constituting the rock, during this stage. In other words, by appearance of hornblende, the degree of dislocation increases, also. This relationship confirmed by field observations, as well, suggests that the rocks softened by increasing of temperature was subjected to the plastic deformation in an environment in which the moderate-high pressure/ temperature conditions probably prevailed.

Amphibolite

The hornblende gradually appears instead of actinolite, by increasing of regional metamorphism. Thus, the amphibolitized gabbros are converted into amphibolites, by development of metamorphic minerals and textures. These amphibolites are composed of xenomorphic phenocrysts of, primary pyroxenes and plagioclases constituting the gabbros. These phenocrysts have traces of cataclastic deformation. The cataclastic fracturing (fractures and fissures) in phenocrysts of pyroxene and plagioclase developed, during regional metamorphism, but no traces of cataclastic fracturing are seen in amphibolitized sections enclosing them. The proportion of these phenocrysts clearly indicates a contrast to which amphibolitic parts (levels, bands, lenses and wedges) occur very widespread in. gabroic rocks, depending on grade of metamorphism. In other words, these phenocrysts show a gradual decrease in amount and size, by increasing of amphibolitization grade in gabbros. Therefore, it is obvious that these relict phenocrysts which were larger and angular at the beginning, reacted with the minerals developed by increasing of regional metamorphic grade and as a result of this, they not only became smaller in size, but decreased in amount and gained spherical shapes. They were replaced by metamorphic minerals and completely disappeared during the later stage of the metamorphism. Bingöl (1968) has proved that the amphibolites were magmatic (ortho) in origin, on the basis of their mineralogical and geochemical compositions, and were derived from gabbros and their basic equivalents.

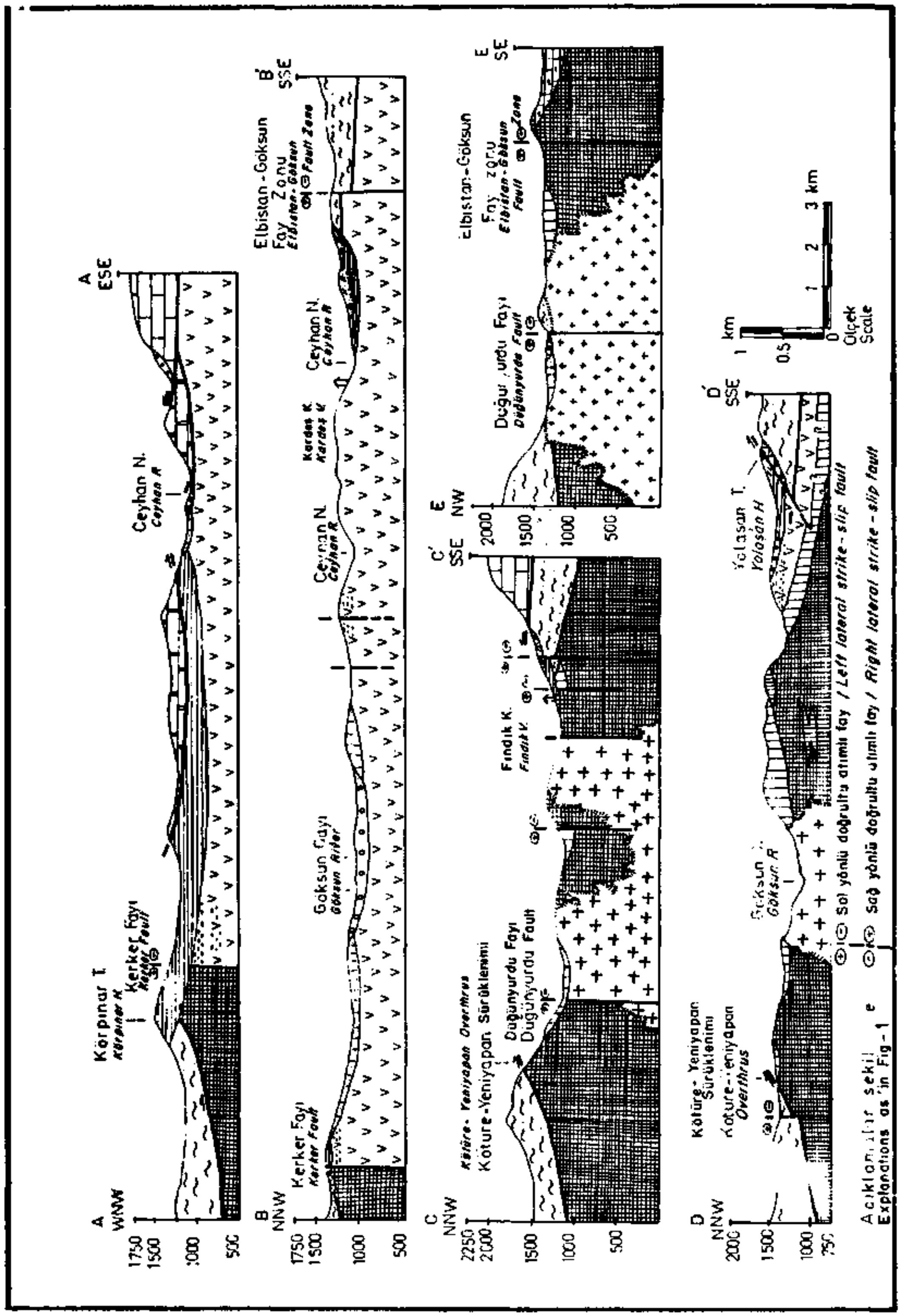

Fig. 2 - The geological cross-sections of the Göksun-Afşin-Elbistan region.

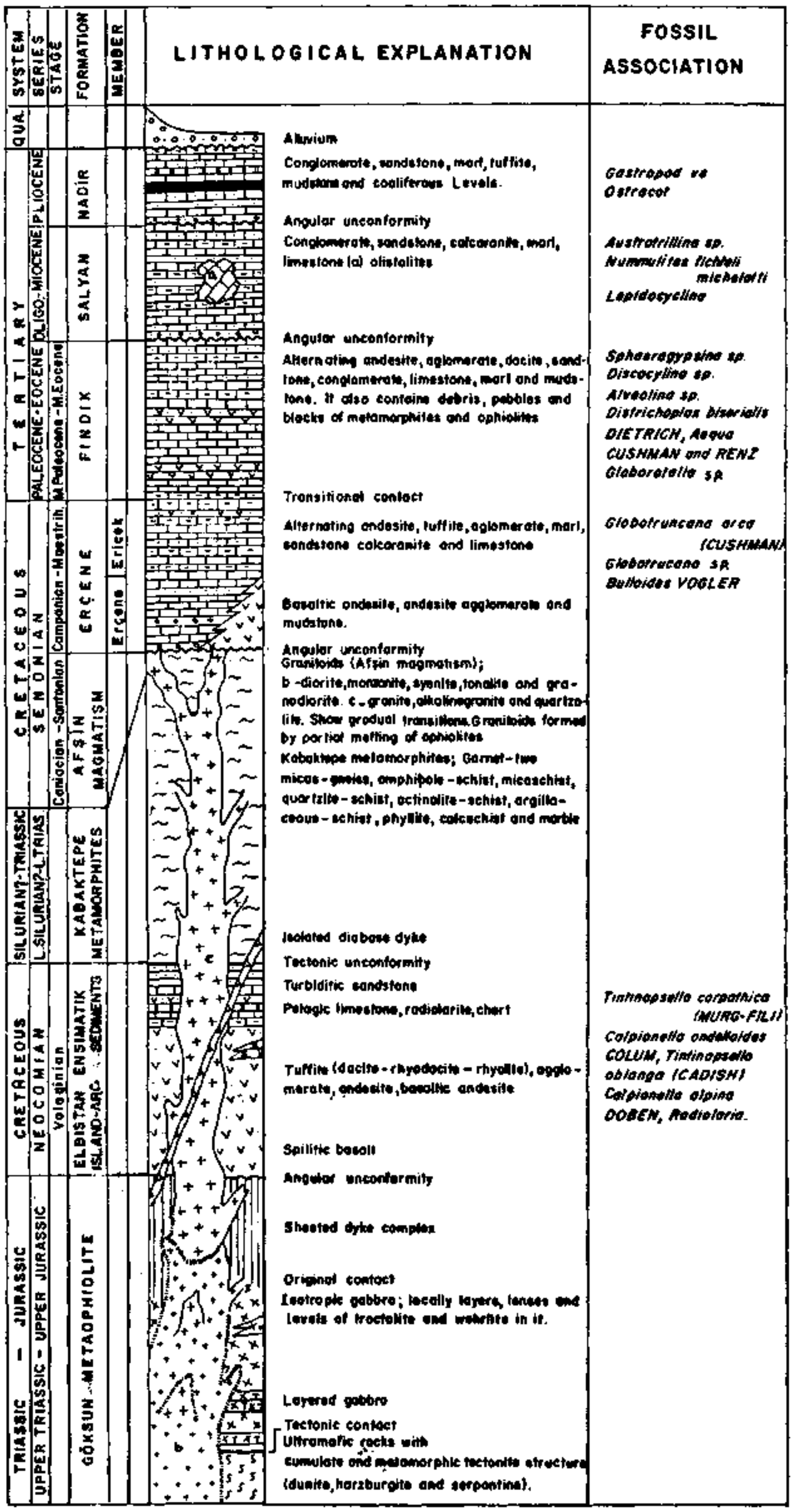

Fig. 3 - The schematik columnar section showing stratigraphic and tectonic relations among the units in the study area.

Based on the field and petrological studies, it is suggested that the amphibolites were subjected to partial melting, during the later stage of the metamorphism and the continental crust material which had existed in natural environment played an important role in this melting, also and as a result of this, the alkaline feldspars were formed, for the first time. Due to enrichment of these alkaline feldspars, banded, lensoidal, wedge-like and augen structures are developed in rocks. Alkaline feldspars are followed by quartz formations. By increasing alkaline feldspars and quartzs, the amphibolites gradually pass into migmatites (Fig. 1).

Migmatite

The migmatites occur at the contacts between the granitoids (first group) and the gabbros (Plate I, fig. 1,2,3,4). They are in appearance of metatexite, as well as diatexite (Mehnert, 1968). In the area, it is observed that the rocks which are light coloured and acidic and were developed in amphibolites and generally extended in parallel direction to the contact of granodiorite, show a transition to the granodiorite in a metatexitic zone characterized by bedded, banded, folded, ptygmatic veined, wedge-like, lensoidal and augen structures. They occur more widespread in the places close to the contact between isotropic gabbro and sheeted-dyke complex. The diatexitic migmatites usually occur at the contacts between the lower parts of the gabbros converted in to amphibolite, and diorites-tonalites-granodiorites (Plate-I, fig. 2). Due to the progressive melting developed in them, it is difficult to distinguish the paleosomes and the neosomes from one another, however, these parts of migmatites gradually pass into eachother. The presence of migmatites is seen at the contacts of the same granitoids (first group) with the rocks belonging to the oceanic crust and with the rocks belonging to the continental crust. The granites, agmatites and migmatites occur very widespread in marbles, quartzite-schists, amphibolites and gneisses all of which form the continental crust (Plate I, fig. 4). They gradually pass into granitoids. The molten parts (neosomes) of the migmatites which were developed in gabbros altered to amphibolites, consists primarily of quartz and alkaline feldspar (Plate I, fig. 1). The prolonged debris and fragments of amphibolites which are conformable to the foliation, and the debris and fragments of plagioclases, pyroxenes and hornblendes are embedded in a matrix (groundmass) composed generally of prolonged crystals of quartz. By increasing leucocratic minerals, these xenoliths decrease in amount.

Diorite-quartz diorite-tonalite

These are generally found in marginal facies of granodiorites which were derived from gabbros. They form transitional rocks between gabbros and granodiorites. They crop out within a small area.

The diorite is rich in leucocratic spots, and dark-green and gray-buff coloured. It is hypidiomorphic and granular in texture. With the increase in grade of reginal metamorphism which is identified with the appearence of zoned plagioclase (oligoclase and andesine) and small amounts of K- feldspar (microcline), the amphibolitized gabbros grade into diorites. The diorites locally contain myrmekite in insignificant quantities. Pyroxenes are partly or wholly in some places altered to hornblendes. The traces of recrystallisation could be observed in plagioclases and pyroxenes. The diorites grade into quartz-diorites and tonalites, with increasing alkaline feldspars and quartzs gradually.

Granodiorite

The granodiorites form the majority of the granitoids cropping out in the study area. The presence of granitic parts could be observed in them. In fact, in some cases, the presence of granitic and alkaline granitic dykes cutting the granodiorite and the xenoliths is seen also (Fig. 3).

The granodiorites are light-buff and greenish in color. They are hypidiomorphic and granular in texture. Quartz is very common. Pyroxenes are partly in some cases or wholly in some cases altered to hornblendes. Plagioclases show zoned structures albite twinnings. In addition, they contain microcline and opaque minerals. Owing to the cataclastic deformations, the individuals of twins belonging to the xenomorphic phenocrysts of pyroxene and plagioclase were bent, twisted, broken and fractured. As a results of this, fractures and fissures were developed in them. In the region, the granodiorites were derived from only oceanic crust or continental crust (Plate I, fig. 4), as well as from hybrid anatectic magmas formed by melting of ophiolites and rocks belonging to continental crust together. In such types of rocks, plagioclases show zoned structures and albite twinnings. Pyroxenes are partly altered to hornblendes. Hornblendes are locally chloritized. The granodiorites contain micas (biotite and muscovite) in addition to opaque minerals. Micas are mostly alteration products of pyroxenes.

Granite

The granodiorites cropping out in the study area, grade into granitic rocks, with decreasing plagioclases and mafic minerals gradually and with increasing K-feldspars and quartz. The granites are hypidiomorphic and granular in texture. The aphanitic texture can be noticed at the contacts between the granites and the rocks units cut by them. The granites contain partly amphibolitized pyroxenes and hornblendes in small amounts. Hornblendes are mostly chloritized. These granites contain abundant dark-black coloured opaque minerals (magnetite), but no micas.

Alkaline granite

These are in the forms of small stocks, veins and dykes, locally cutting the granites and the granodiorites, and are pinkish, in color. Quartz and K-feldspar are commonly dominant components in them. The granites formed by partial melting of only continental crust, contain micas also, in addition to quartz, K-feldspar and opaque minerals. Saussuritisation can be noticed in most of the granitoids (first and second groups) discussed above. The secondary minerals developed at the latest stage and the micro-faults intersecting at right angles, typically represent the cataclastic deformations.

Isolated diabase dykes

These dykes cutting the granitoids and all the rock units cut by granitoids are blackish-gray, greenish-black in color. They are locally broken away, boudinaged and sheared. Ophitic texture was preserved in them. Pyroxenes are partly altered to fibrous actinolites. The laths of plagioclases are partly albitized. In addition, there are sphene, epidote, calcite and chlorite in them. They contain opaque minerals as well (Fig. 3).

Metamorphic grade and age of granitoids

Chlorite, epidote, sericite and calcite were secondary developed in granitoids. Actinolite, albite, epidote, chlorite, sphene and calcite are observed in isolated diabase dykes cutting the granitoids. These mineral paragenesis clearly indicate that the granitoids were subjected to regional metamorphism in green-schist facies (Winkler, 1967). The micro-faults intersecting at right angles and the traces of fracturing developed in crystals of minerals are observed in granitoids and diabase dykes cutting them. These micro-faults also intersect the mineral paragenesis generated under regional metamorphic conditions.

The granitoids are covered by Maestrichtian sediments beginning with a basal conglomerate (Tarhan, 1983-1984). In this case, the age of metamorphism observed in granitoids is probably pre-Campanian or Campanian.

Evolution of Afsin magmatism

There are metamorphic fields subjected to conditions of high temperature/low pressure in Göksun metaophiolite constituting the basement of the rock units cropping but in the study area and the overlying island-arc volcanoclastic sediments of Neocomian age (Tarhan, 1985). The presence of metamorphic fields generated under conditions of low/high temperature and the presence of migmatites developed by anatexis, can be clearly recognized in gabbros. Due to the N-S directed compressional forces probably started in Lower Cretaceous and increased after Neocomian, the Kabaktepe metamorphites have been thrusted over the Göksun metaophiolite and the intra-oceanic island-arc sequence as nappes, during the late Cretaceous. A collision (between two continents) has probably resulted from this compressional regime. It is obvious that the granitoids in the region, cut these metamorphites. Since the crustal thickening wihich was caused by the post-Neocomian compressions, both maintained the isostasy of the region (Moore, 1971) and as a result of this possibly, the oceanic crust sunk into depth. It is likely that the progressive metamorphism of high temperature/ low pressure, depending on depth, caused the partial melting of the rocks belonging to the Göksun metaophiolite at the bottom.

Şengör and Kidd (1979) reported that the crustal thickening and the tensional joints could be formed by the effects of compressional forces. Turcotte (1983) expressed that the intense lateral tensional forces caused by crustal thickening related to the collision and by the thinning of mantle lithosphere, probably surpass the regional compressional forces in a mathematical model.

In the region, the crust which has become thick as a result of the collision which took place over the Benioff zone (Tarhan, 1985), sank into depth depending on the isostatic balance. Since the regional compressional forces probably gave rise to the development of the regional tensional forces, during this collision, the weak zones were formed. The anatectic magma generated by the partial melting of the rocks forming the Göksun metaophiolite at depth, has risen upward along these zones and cut the sheeted dyke complex (Göksun metaophiolite) and the overlying Elbistan ensimatic island-arc sequence and the Kabaktepe metamorphites which had been thrusted over both of them, as nappes during the late Cretaceous.

Mitchell and Bell (1973) thought that the stratigraphic relationships in island-arcs infrequently indicate that the granitic plutons were probably emplaced in the region, during the eruption of volcanic rocks or later. Şengör and Yılmaz (1983) suggested that both the granitic intrusion and the extensive areas in which the metamorphism of high temperature took place along the suture zone, show striking similarities to Tibet plateau-Himalayas which had been considered to be developed, as a result of collision, by Dewey and Burke (1973).

It is concluded that the granitoids cropping out in the study area developed after the collision which had taken place over the subduction zone. In other words, the island-arc sequence and the acidic plutons cutting its basement are not seen before the metamorphites have been thrusted, as nappes.

Origin of Afşin magmatism

The amphibolitisation locally takes place in the rock units forming the Göksun metaophiolite, whereas amphibolite occurs widespread in them. The gabbros gradually pass into amphibolitized gabbro, amphibolite, migmatite, granulite and granitoids respectively in increasing order of metamorphic grade. The rocks of Göksun metaophiolite in which partial melting took place, indicate the presence of conditions of the regional metamorphism which took place in facies of low/high grade almandine-amphibolite (Winkler, 1967) and granulite (Escola, 1939). The original textures and structures were partly preserved in Elbistan ensimatic island-arc sediments overlying the Göksun metaophiolite with an angular unconformity, whereas the original minerals were partly converted into hornblende, actinolite, albite, sericite, epidote and calcite. In fact, some porphyroblasts of albite were developed in some samples (Tarhan, 1985). These mineral paragenesis indicate the presence of regionally metamorphosed fields which were developed in facies of green-schist and low-grade amphibolite in volcanic-arc sediments. Miyashiro (1961, 1972a) and Zwart (1967, 1969) suggested that these fields probably correspond to the volcanic belts which were developed in island-arcs and continental margins. An ophiolitic melange crops out in the southeastern part of the study area. According to the geological and petrological data, probably a pair of metamorphic belts were formed in the region. The study area probably corresponds to a metamorphic belt, which was formed by regional metamorphism of high temperature/low pressure. In other words, both the Göksun metaophiolite and the overlying Elbistan ensimatic island-arc sequence probably represent the region over the oceanic plate plunging along the Benioff zone. Miyashiro (*1972b)* suggested that both the metamorphites which were formed under the conditions of high temperature/low pressure and the granitic rocks, form the basement of a volcanic-arc which was developed in front of the plunging lithosphere. Oxburgh and Turcotte (1970) point out that the characteristic volcanism of subduction zones take place as a result of increasing heat currents. According to the opinions of these authors and the geological evidences obtained from this study, the presence of northwards (or northwesterly) subduction zone is apparent in the southern part of the study area.

Many hypothesis have been offered about the origin and the formation of acidic magmas. According to the dominant common opinion, the only continental crust may be subjected to partial melting and the acidic magma which has resulted from this process, form the granitic plutons. One of them, Winkler (1967) suggested that it is unlikely that acidic magmas are derived from any magma of gabroic composition by differentiation. For this reason, he proved experimentally that the rocks which have granitic, granodioritic and tonalitic compositions have been derived from metamorphic rocks (continental crust) by anatexis. However, some authors suggested that ultramafic and ultrabasic rocks might be subjected to partial melting, as a result of anatexis (Tuttle and Bowen, 1958; Rittmann, 1965; Winkler, 1967 and Bingöl, 1968).

In the study area, the presence of migmatite which was developed between gabbro transformed to amphibolite, and diorite-tonalite-granodiorite is observed. The migmatites occurred also at the contacts of the same granitoids with" both the rocks of continental crust and ophiolites, both of which were metamorphosed in the facies of almandine-amphibolite and granulite (Tarhan, 1986). On the basis of these relationships, the present-writer thinks that basic or ultrabasic rocks might be subjected to partial melting just like a continental crust and the anatectic magma which was formed by partial melting, might accumulate in situ or in the upper zones by filter-pressing prevailing in the environment and finally as a result of this accumulation, the intrusions of acid plutonic rocks might take place.

The probable sources of heat by which anatexis may take place in the rocks of oceanic crust are as follows: (a) Fields of high heat current which were generated in oceanic lithosphere forming the basement of the Elbistan ensimatic island-arc which was developed over the subduction zone (Oxburgh and Turcotte, 1970); (b) The sinking of the oceanic crust into depth, depending on the isostatic balance caused by the metamorphites which were thrusted as nappes, during the late Cretaceous. As to the probable sources of water needed for anatexis, they are as follows: (a) Ocean-floor metamorphism which provides considerable amounts of into the oceanic crust (Melson, Thompson and Andel, 1968; Miyashiro et al., 1971); (b) Chlorite (with content of H_2O 15 %) biotite and amphibole (with content of H_2O 2-3%) all of which were formed, as a result of regional metamorphism of high temperature/low pressure (Miyashiro, 1961, 1972b; Zwart, 1967, 1969) which took place in oceanic lithosphere forming the basement of island-arc and as a result of ocean-floor metamorphism.

The granitoids cropping out in the study area were formed by melting of the Göksun metaophiolite, the island-arc and the Kabaktepe metamorphites together or separately, depending on the crustal thickening, during the late Cretaceous. This anatectic magma defined as Afsin magmatism caused the development of granitoids by solidification in situ or by intrusion into the upper parts of the crust, depending on the pressures prevailing in the environment.

Age of Afşin magmatism

The granitoids cropping out in the study area, cut the Neocomian aged island-arc volcanoclastic sediments. They are covered by the Upper Cenonian (Campanian-Maestrichtian) sediments. The granitoids cut also the allochthonous Kabaktepe metamorphites which were emplaced in the region, after Neocomian According to the geological evidences, the age of Afşin magmatism is the time interval ranging from post-Neocomian to pre-Campanian (Fig. 2,3). The Baskil magmatism defined by Yazgan and Asutay (1981) and Yazgan (1983) in the vicinity of Malatya, forming the eastern extension of the study area is the magmatic equivalent of the Afşin magmatism. Coniacian-Santonian age (82-86 my.) obtained from K/Ar dating of these rocks coincides with the upper boundary of the acid plutonic rocks in the study area.

Based on these data, Coniacian-Santonian ? age (or Coniacian-Upper Santonian) can be assigned to Afşin magmatism.

Correlation between the rocks related to Afşin magmatism and the granitoids of Eastern Taurus

The granitoids cropping out to the north of Doğanşehir show similar relationships and features to the rocks related to Afşin magmatism. The rocks of Coniacian-Santonian age related to Baskil magmatism (granodiorite, tonalite, quartz-monzonite, monzonite, monzo-diorite, diorite and gabbro) which crop out in the vicinity of Malatya are considered to be products of magmatism of continental margin by Yazgan and Asutay (1981) and Yazgan (1983). These rocks are included in Yüksekova complex by them. They point out that these magmatic rocks out the Pötürge metamorphites and are covered by a volcanoclastic flysch of Campanian-Maestrichtian age. Many investigators have studied on the metamorphic rocks and the granitic rocks of Bitlis. One of them, Göncüoğlu (1983), claimed that the Muş-Kızılağaç granite is Middle Devonian-Lower Permian in age and was subjected to dynamothermal metamorphism, during the Lower Turonian. The Bitlis metamorphites and granitic rocks are overlain by Maestrichtian aged sediments with an angular unconformity (Meri9, 1973).

In the study area, there are from bottom to top, the Göksun metaophiolite, the Elbistan ensimatic island-arc sequence and the Kabaktepe metamorphic nappe which is thrusted over both of them. These rock units are cut by the rocks of Afşin magmatism of Coniacian-Upper Santonian? age. The isolated diabase dykes cut the granitoids and the rock units which are cut by the granitoids (Fig.3). All these rock units cropping out in the study area are covered by Campanian-Maestrichtian aged sediments with angular unconformities (Tarhan, 1984, 1985). The Kabaktepe metamorphites are the equivalents of Bitlis/Pötürge metamorphites (Tarhan, 1985). This structural relationship observed in the study area (see; the 1:500,000 scale geological map of Turkey) is observed also in the neighbourhoods of Malatya, Doğanşehir and Bitlis, all of which are located in the same belt. The

Muş-Kızılağaç granite cuts also the lower and upper associations (Boray, 1975) forming the Bitlis metamorphites (on the basis of the observations of the present writer). For this reason, the present writer thinks that it is unlikely that the Muş-Kızılağaç granite described by Göncüoğlu (1983) is Middle Devonian-Lower Permian in age.

Consequently, the rocks related to Baskil magmatism and the granitoids cropping out around Doğanşehir, Elazığ and Bitlis are the magmatic equivalents of the rocks related to Afşin magmatism. They were probably formed by intrusions of anatectic magma in different places and times, the generation of anatectic magma being as a result of melting of oceanic and continental crusts, together or seperately, as is seen in the rocks related to Afşin magmatism (Fig. 3).

GRANITIC ROCKS

A second type of rocks which have granitic composition and crop out extensively in Taurid belt and differ from the granitoids cropping out in the study area in some respects, such as mode of occurrence and place of occurrence, were named as granitic rocks and described, for the first time. These granitic rocks were derived from the same rocks consisting of ophiolite association (gabbro, sheeted dyke complex and volcanosedimentary deposits), island-arc volcanoclastic sediments and metamorphic rocks of sedimentary origin, as the granitoids which intruded in the area, and gradually pass into them. However, they don't show any intrusive features. They were formed by recrystallisation of the older rock units which were present then, with increasing grade of metamorphism. They contain relict bands, wedges, lenses, beds and fragments of the rocks from which they were derived. They show gradual transitions to relict formations which are included in them. Since the granitic rocks were developed along the bedding and foliation planes of the older rocks which were present then, due to the recrystallisation, and didn't pass through any magmatic stage, the bedded and foliated structures of parent rocks as well as the original tectonic and stratigraphic relationships between the parent rocks were preserved identically. These rocks formed by recrystallisation show similar compositions to those of parent rocks. In addition to the presence of dykes and veins of quartz observed locally in them, the presence of dykes and veins of acidic and basic compositions, cutting the granitic rocks is noticeable. In other words, due to the probable inefficiency of P/T conditions of regional metamorphism, any anatexis didn't take place in parent rocks (ophiolites, metamorphites, volcanosedimentary deposits) transformed to granitic rocks. Since they didn't pass through any magmatic and crystallisation stages, they were transformed to granitic rocks (diorites, monzonites, syenites, granodiorites, granites, alkaline granites and their derivatives) by recrystallisation in a solid phase.

In the study area, from the metamorphic rocks of pelitic origin in which the lower grade metamorphism took place towards the granitic rocks, the presence of metamorphic rock units metamorphosed in the facies of Barrovian green-schist, low grade amphibolite, high grade amphibolite and granulite, respectively, is observed. In places where these metamorphic rocks crop out, it was observed in both petrographic and field studies that these metamorphic rocks gradually pass into granitic rocks. The paragenesis of hornblende-plagioclase, disthene-sillimanite-orthose, almandine-sillimanite-orthose and monocline pyroxene-quartz are seen in metamorphic rocks of these transitional zones. These mineral paragenesis generated under conditions of high pressure/temperature are followed by the generation of paragenesis of cordierite-orthose-sillimanite, andalusite-orthose-plagioclase, cordierite-orthose-plagioclase, all of which characterize the conditions of low?-moderate pressure/high temperature. However, the principal stable minerals are blue-green amphiboles and orthose in this field. It is noticeable that sillimanite, garnet arid monoclinic pyroxene completely disappear in this field where granitisation starts taking place. As a result of this, andalusite, biotite, muscovite, chlorite, epidote, tremolite, calcite and quartz arise depending on metamorphic grade. Plagioclases are partly or wholly replaced by albite, orthoclase, microclane and quartz. These replacements gradually take place. According to these mineral paragenesis, the progressive Barrovian type metamorphism which took place under the conditions of high pressure/temperature is followed by Abukuma type metamorphism in which the conditions of low?-moderate pressure/high temperature prevailed.

In field studies, it is observed that the metamorphic rocks which were subjected to high grade metamorphism, pass into the granitic rocks which are medium-coarse grained, granulitic (mosaic) textured and enriched in leucocratic minerals, as a result of gradual disappearance of metamorphic textures and structures, and migmatite was developed at the contacts between them. The fact that the blue-green amphiboles maintained their stabilities, during the metamorphism of cordierite-amphibolite facies following the facies of almandine-amphibolite and granulite, indicates that the hydraulic pressure of the environment was high, then (Winkler, 1976). In spite of the presence of sufficient water and temperature in the environment, following the regional metamorphism in the facies of almandine-amphibolite and granulite, the absence of partial melting can be possibly attributed to decrease in pressure. The mineral paragenesis generated, as a result of the regional metamorphism of high pressure/temperature, lose their stabilities, and change with decreasing pressure. They interact in solid state and are transformed to the other mineral groups, being stable in new conditions. Although the effects of pressure are insignificant, the presence of temperature which is high enough, probably caused the crystal grains of the metamorphic rocks to separate from eachother, by growing, as a result of recrystallisation and at the same time, to be transformed to granitic rocks in situ by enrichment in leucocratic minerals. The granitisation took place also in the rocks of ophiolite association in similar conditions.

Biotite is generally observed in metamorphic rocks of sedimentary origin granitoids and granitic rocks derived from them. According to the petrography studies, biotites generally occurred in genetic relationship with opaque minerals. The minute opaque minerals (iron oxides) which were developed along the cleavages of chlorite derived from pyroxenes and amphiboles, were altered to reddish-brown opaque material in green-schist facies. It is observed that this material entered into the lattice of chlorite and made it convert into biotite, partly or wholly (depending on the alteration of opaque material). Biotite isn't seen in granitoids derived from rocks of ophiolite association by granitisation or by partial melting. However, it can be seen in granitoids derived from hybrid anatectic magmas which were generated by melting of oceanic and continental crusts, together (H type granitoids).

Consequently, according to the view of the present writer, the regional metamorphism which took place in cordierite-amphibolite facies, following the regional metamorphism which took place in almandine-amphibolite and granulite facies, and probably caused the granitisation, resulted in the formation of granitic rocks which have different mineralogical compositions and show lateral and vertical transitions to eachother and present alternating and napped structures, depending on the original compositions, stratigraphy and tectonic relationships of metamorphic rocks.

CONCLUSIONS AND DISCUSSION

The fact that the Neocomian ensimatic island-arc sequence overlies the Göksun metaophiolite cropping out in the study area and that the rifting related to the opening of the southern branch of Neo-Tethys begins, during the Upper Triassic (Friedman et al., 1971; Goldberg and Friedman, 1974; Şengör and Yılmaz, 1983), reveals that the Göksun metaophiolite is a part of a Middle?/

Upper Triassic-Jurassic aged oceanic lithosphere which continued its spreading during the Jurassic and reached its maximum spreading probably during the Middle Jurassic. However, Whitechurch et al., (1983) emphasized the absence of the Jurassic oceanic crust in Taurid belt and offered two alternative solutions as follows: (a) Ocean-floor spreading couldn't take place during this period; (b) The part of Neo-Tethys in this period has been consumed, as a result of subduction. The geological evidences obtained from the study area, reveal that an oceanic crust exists in Taurid belt extending in the southern branchof Neo-Tethys and it hasn't been wholly consumed by subduction (Tarhan, 1983).

Arabian-African plate probably continued moving northwards after the Neocomian. The north-south compressional forces which have resulted from this phenomenon, caused the Kabaktepe metamorphites to be thrusted over both the Göksun metaophiolite and the Elbistan ensimatic island-arc sediments, during the late Cretaceous. The phenomena which happened, during the late Cretaceous, probably caused the crustal thickening and the closure of the southern branch of Neo-Tethys.

Many investigators who have studied in Taurid belt (Blumenthal, 1952; Aslaner, 1973; Sungurlu, 1974; Lapierre, 1975; Özgül, 1976; Gutnic et al., 1979; Özgül et al., 1981; Yazgan 1983; Whitechurch et al., 1985) suggested that the ophiolite and the metamorphic nappes were emplaced in the region during the Maestrichtian or later. Tarhan (1983, 1984) point out that the Upper Cenonian (Campanian-Maestrichtian) sediments have transgressive features and had obtained detritic materials from the underlying rock units (ophiolites, metamorphites and granitoids). According to this evidence although the presence of ophiolites and metamorphic nappes emplaced, during the Maestrichtian is known in Taurid belt, at least pre-Upper Cenonian (Campanian) aged ophiolites, metamorphites, island-arc sediments and an acidic magmatism cutting all of them, had existed in the study area. According to these data, there are two alternative solutions as follows (Fig. 1,2,3):

1. The closure of the southern branch of Neo-Tethys took place during the late Cretaceous. The metamorphic and ophiolites nappes of Taurid belt were emplaced during the late Cretaceous (at least pre-Cenonian), for the first time, but not during the Maestrichtian.

2. The Göksun metaophiolite forms the basement of the rock units cropping out in the study area. It is unknown that what types of rocks are present at the bottom of this ophiolite. Since both the Göksun metaophiolite and the overlying island-arc sequence have preserved their original relationships and internal structures well, against the tectonic deformations, they are considered to be autochthonous (or parautochthonous) units.

There are two groups of grankic rocks which their place of formation and mode of formation are different from eachother in Taurid belt. The rocks belonging to the first group are granitoids which crop out widespread in Eastern Taurus and in the study area and intrude the enclosing rocks. They aren't formed by differentiation of a magma having gabroic (or basaltic) composition. They were generated by gradual anatexis of rocks of oceanic crust, island-arc sediments and rocks of continental crust, with increasing of high temperature regional metamorphism (almandine-amphibolite and granulite facies) accompanying the Medium/high pressures. Granitoids commonly show gradual transitions to eachother and form regular series. Monzonite, syenite and their derivatives are generally seen in marginal facies of the granitic rocks generated by partial melting of metamorphic rocks of sedimentary origin. Except for quartz, feldspar and plagioclase, the mafic minerals which are included in granitoids are related to parent rocks (Fig. 1,2).

1. The granitoids formed by partial melting of the rocks of oceanic rock, contain no mica, whereas they contain hornblende, amphibolitized pyroxene and opaque minerals. They show similar characteristics as I type granitoids (Didier et al., 1982; White and Chappell, 1977).

2. The granitoids derived from hybrid anaitectic magma by melting of rocks of oceanic and continental crusts together, contain hornblende, amphibolized pyroxene, pyroxene, mica (biotite, muscovite) and opaque spinel group of minerals. For this reason, they are named as hybrid granitoids (H type granitoids) by the present writer, for the first time.

3. The granitoids formed by partial melting of metamorphic rock of sedimentary origin, contain only mica of mafic minerals. They show similar characteristics as S type granitoids (Didier and et al, 1982; White and Chappell, 1977).

The second group of rocks are called as granitic rocks. Their outcrops are scarce in Eastern Taurus. They were derived by recrystallisation of the same parent rocks (ophiolite, volcanic-arc sediments and metamorphic rocks) with increasing regional metamorphism, as is seen in granitoids. Since they didn't pass through any magmatic stage, they don't show any intrusive and contact metamorphic features. Aplite, pegmatite and isolated diabase dykes aren't seen as cutting this type of granitic rocks. They show gradual transitions and similar compositions to parent rocks, and have some relict structures.

Consequently, the oceanic crust should sink into depth, depending on the crustal thickening and isostatic balance which were developed by collision over the subduction zone, for the formation of rocks related to Afşin magmatism. Since the regional compressional forces probably cause the developement of regional tensional forces, during the collision, the weak zones are formed (Turcotte, 1983). The same granitoids cropping out in the study area were formed by partial melting (or complete) of rocks belonging to continental and oceanic crusts, both together and separately. Therefore, the meltings which had been derived by anatexis of oceanic and continental rocks (metamorphites, island-arc sediments), together or separately in different places and depths, caused the development of an anatectic magma. This anatectic magma has risen upwards along the weak zones, by the effects of high pressures in the environment, intruded the upper parts of the crust in different places and depths and finally formed local contact metamorphic zones in country rocks. The high grade differentiation of this anatectic magma defined as Afşin magmatism, caused the granitoids (diorite, monzonite, syenite, tonalite, granodiorite, granite, alkaline granite, silexite and their derivatives) which intruded in different periods, to be formed.

The fact that the intrusive granitoids and non-intrusive granitic rocks occur so widespread, both in Turkey and in the other parts of the world, can't be explained only by melting of continental crust, as a result of anatexis and by differentiation of a primary gabroic (or basaltic) magma. It is concluded that they were mostly formed either by recrystallisation (granitic rocks) or by melting (granitoids) of continental and oceanic rocks and volcanic-arc sediments, in situ, with increasing of regional metamorphic grade.

ACKNOWLEDGEMENT

The present-study was carried out within the framework of the project of «Taurid ophiolites», directed by the Department of Basic Researches, MTA.

The author is deeply indebted to TPAO for providing camp facilities, to Dr. O. Tekeli, Dr. D. Perinçek and H. Kozlu for their assistance, during the field studies and to Dr. M. Şengün and A. Çağlayan for reading this article critically.

Manuscript received December 27, 1985

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PLATE

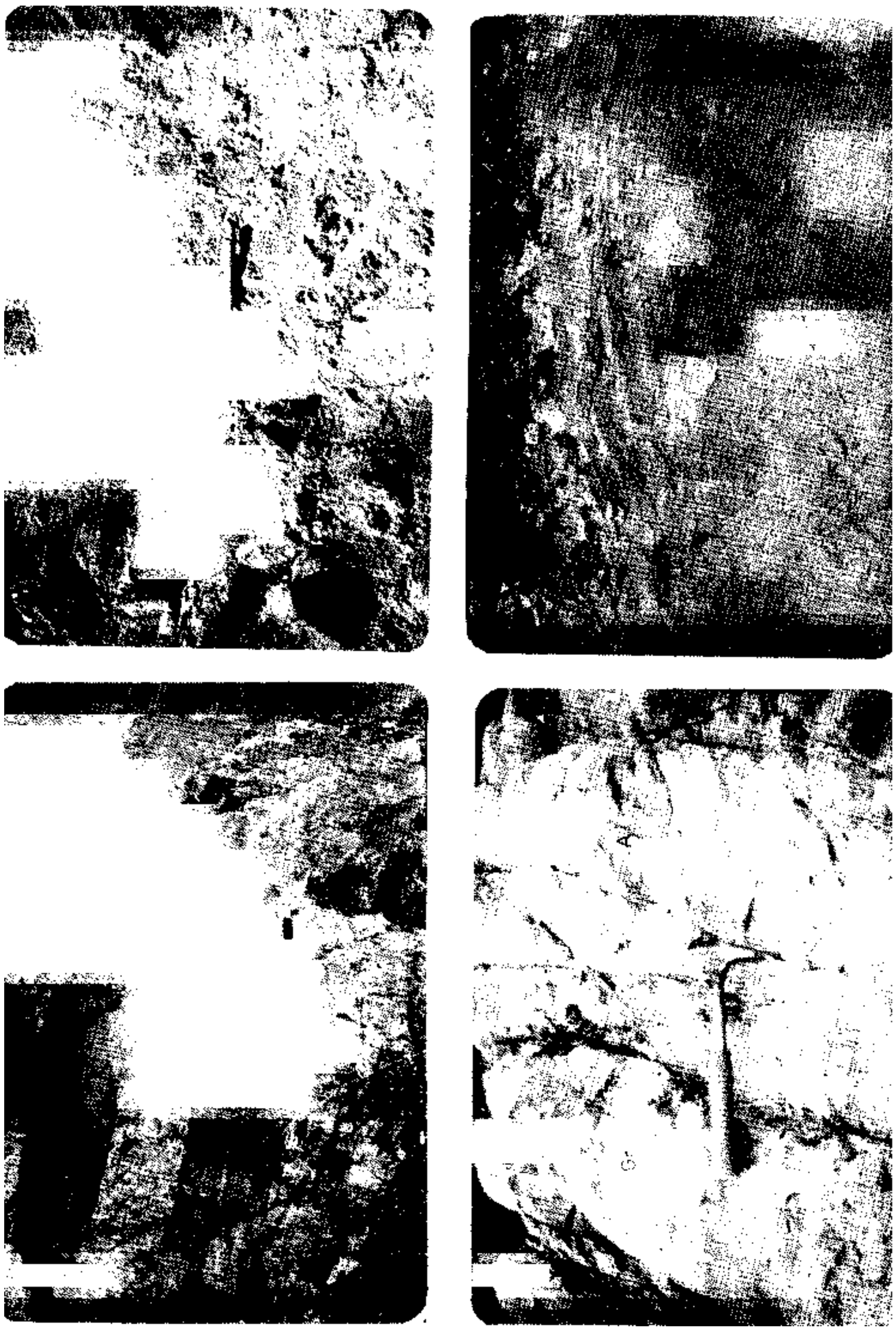
PLATE -I

Fig. 1 — Metatexites migmatite observed in gabbros
Gr—Neosome (granite); A—Paleosome (amphibolite).

Fig. 2 — Diatexites migmatite observed in gabbros
Gr—Neosome (granite); A—Paleosome (amphibolite).

Fig. 3 — Metatexites migmatite observed in gabbros
Gr—Neosome (granite); A—Paleosome (amphibolite).

Fig. 4 — Diatexites migmatite observed in marbles
Gd—Neosome (granodiorite); Mr—Paleosome (marble).



TECTONICS OF SULTANDAĞ AND ITS SURROUNDINGS

Cavit DEMİRKOL**

ABSTRACT.— In the tectonic evolution of Sultandağ and its surroundings, the compressional and tensional phases seem to have taken place alternately. The Paleozoic aged basement rocks were folded by the of Caledonian and post Caledonian phases. These phases affacted the structure of the basement rocks to some considerable scale. From the morphological extend of Sultandağ, it is apparant that the fold axes are in the direction of northwest-southeast. However, in the western side of Çay-Hoyran area these axes are in the direction of northeast-southwests: The investigated area was probably quite stable during Upper Jurassic-Upper Cretaceous, but it was increasingly affacted by the compressional stresses at the end of Lutetian, and as a result, the Hoyran ophiolitic complex unit moved over the older rocks by a tectonic contact. The investigation of the sediments of Upper Miocene-Pliocene shows that the region has neotectonically developed by compression. The Upper Miocene-Pliocene aged fluviatile-lacustrine and terrestrial formations of Bağkonak, Göksöğüt and Yarıkkaya, unconformably overlain the rock units that had previously existed. Although these units show an upward lithology of conglomerate-sandstone and siltstone-argillaceous limestones-limestone, they also show lateral and vertical transitions. The folds, lying nearly in the direction of northwest-southwest, and reverse faults were formed by a tectonic deformation. The sediment adjacent to the basement rocks were strongly folded and also broken. The explanations given so far seem to indicate that the area has once undergone a compression in the eastnortheast-westsouthwest direction. Finally, it's possible to say that the eastnortheast-westsouthwest directed compression regime, which looks effective in the area, has taken places by the resistance of the Aegean plate to the Anatolian plate during a westernly movement.

CENOZOIC VOLCANISM OF CENTRAL ANATOLIA

Tuncay ERCAN*

ABSTRACT.— Volcanism in Central Anatolia, the regional extent of which is defined by 1:500 000 scale sheets of Ankara-Konya-Kayseri and Adana, is widespread. In this paper the volcanism, which have been active during the various periods ranging from Paleocene to recent, has been subject of study and the related volcanics have been differentiated into 6 different groups. In Central Anatolia, the volcanics of Paleocene and Eocene age are of calc-alkaline in nature and are arc type-volcanism, the genesis of which is related to former subduction zones. The other 4 groups of volcanics, ranging from Oligocene to Miocene, Pliocene and Quaternary, are basically calc-alkaline and are considered to have been originated from continental crust during the period following the subduction of the oceanic crust and the collision of continental crusts resulting from the convergence of the plates in this region. There are also some, mantle originated hybriditic volcanics in this area.

FAUNE DE RUDISTES MAESTRICHTIENNE DE L'ENVIRON DE KAHTA-ADIYAMAN (ANATOLIE SUD-EST)

Sacit ÖZER*

ABSTRACT. — In this paper, the Rudist fauna of the Kahta-Adıyaman, which contain the forms of *Dictyoptychus euphratica, Dictyoptychus striatus, Dictyoptychus leesi, Sabinia klinghardti, Caprina* sp., *Vautrinia syriaca, Lapeirousia* sp., *Pseudopolyconites* sp., *Pironaea anatolica, Pironaea syriaca, Vaccinites braciensis, Hippurites* cf. *cornucopiae, Bournonia* sp., *Radiolites* sp., *Biradiolites* sp. and which form an association during the Maastrichtian, are examined and their geographic and stratigraphic repartitions in Turkey and Mediterranean province are given.

RESUME. — Dans cette note, il est etudie la faune de Rudistes de Kahta-Adıyaman qui constitue une association Maestrichtienne comprenant les formes suivantes: *Dictyoptychus euphratica, Dictyoptychus striatus, Dictyoptychus leesi, Sabinia klinghardti, Caprina* sp., *Vautrinia syriaca, Lapeirousia* sp., *Pseudopolyconites* sp., *Pironaea anatolica, Pironaea syriaca, Vaccinites braciensis, Hippurites* cf. *cornucopiae, Bournonia* sp., *Radiolites* sp., *Biradiolites* sp. et donne leurs repartitions geographiques et stratigraphiques en Turquie et dans la region mediterraneenne.

INTRODUCTION

Les formations a Rudistes de la region d'Anatolie du Sud-Est offrent une vaste distribution aux environs de Kahta-Adıyaman. La formation de Besni y contient essentiellement des Rudistes et comprend egalement des grands foraminiferes benthiques tels que *Orbitoides medius, Siderolites calcitropoides, Omphalocyclus macroporus, Loftusia morgani* et des Gastropodes, Echinides et Madreporaires. Les indications feiostratigraphiques tirees des micro et macrofossiles nous permettent d'attribuer cette formation au Maestrichtien, plus definitivement au Maestrichtien superieur (Meriç, 1965; Sungurlu, 1974; Yalçın, 1977; Meriç, Oktay et Özer, 1985; Meriç et al., 1986).

Ces Rudistes qui n'ont ete signales que tout dernierement (Özer, 1985a) forment l'object de la presente note du point du vue de leurs repartitions geographiques et stratigraphiques dans toute l'Anatolie et la province mediterraneenne.

FAUNE DE RUDISTES

Les Rudistes de l'environ de Kahta-Adıyaman ont ete etudiee dans six localites (Fig. 1):

Localite du Nord d'Alidamı: La faune la plus riche se trouve dans cette localite. Nous y avons determine *Vautrinia syriaca* (Vautrin) Milovanovic (tres abondants), *Vaccinites braciensis* Sladic-Trifunovic (abondants), *Pironaea syriaca*. Vautrin, *Pironaea anatolica* Karacabey, *Sabinia klinghardti* Böhm, *Dictyoptychus euphratica* Karacabey-Öztemür, *Dictyoptychus leesi* Kühn, *Caprina* sp. et *Radiolites sp.* (Planche I et II).

Il faut remarquer que *V. syriaca* et *V. braciensis* s'y observent specialement aux niveaux inferieures et *D. euphratica* au contraire particulierement aux niveaux superieures de la formation de Besni (Fig. 2).

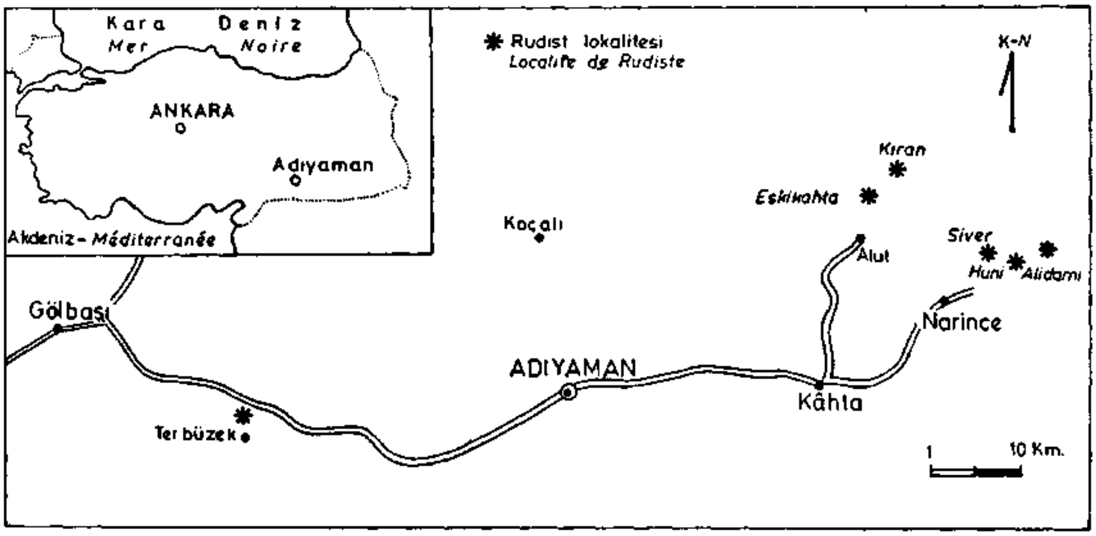


Fig. 1 - Situation géographique des localités de Rudistes de l'environ de Kâhta-Adıyaman.

Localites de Siver et de Huni: Dans ces localites nous avons constate l'existence tres abondante de *D. euphratica* (Planche II).

Localite d'Eskikahta: La formation de Besni y contient deux niveaoix de Rudistes. Au niveau inferieur *Dictyoptychus striatus* Douville (abondants), *Dictyoptychus* sp., *Hippurites* sp., *Vaccinites* sp. et *Caprina* sp. (Planche II et III) et au niveau superieur *Lapeirousia* sp. (abondants), *Pseudopolyconites* sp., *Biradiolites* sp. et *Bournonia* sp. (Planche III) sont determines.

II faut noter que la determination specifique des Rudistes du niveau superieur est impossible a cause de leur mauvaise conservation.

Localite du Sud de Kıran: Les calcaires a Rudistes de cette localite comprennent des fragments appartenant aux genres de *Lapeirousia, Bournonia* et *Biradiolites* (Planche III).

D'autre part, nous avons pu identifier *Hippurites* cf. *cornucopiae* (Planche III).

Localite du Nord de Terbüzek: Cette localite presente le type localite de la formation de Besni. On y rencontre des fragments appartenant en particulier a la famille de Radiolitidae comme *Biradiolites* sp., *Bournonia* sp., *Lapeirousia* sp. mais specifiquement indeterminables (Planche III). Hippuritidae sont tres rares.

La composition de Rudistes de ces six localites nous a permis d'annoncer l'existence de deux differentes associations de Rudistes dans la region examinee dont l'une se caracterise par les especes de *V. syriaca, V. braciensis* et *D, euphratica* qui ont ete observees a Alidamı, Huni et Siver et l'autre se represente par les formes de *D. striatus, Lapeirousia, Pseudopolyconites, Bournonia, Hippurites* qui ont ete rencontrees a Eskikahta, Kıran et Terbüzek.

REPARTITION GEOGRAPHIQUE ET STRATIGRAPHIQUE

Les travaux effectues jusqu'a maintenant sur les Rudistes de la Turquie nous montrent que les Rudistes de Kahta-Adıyaman offrent une repartition geographique tres vaste. On les trouve dans l'Anatolie Sud-Est, Est, Centrale et a la Peninsule de Kocaeli.

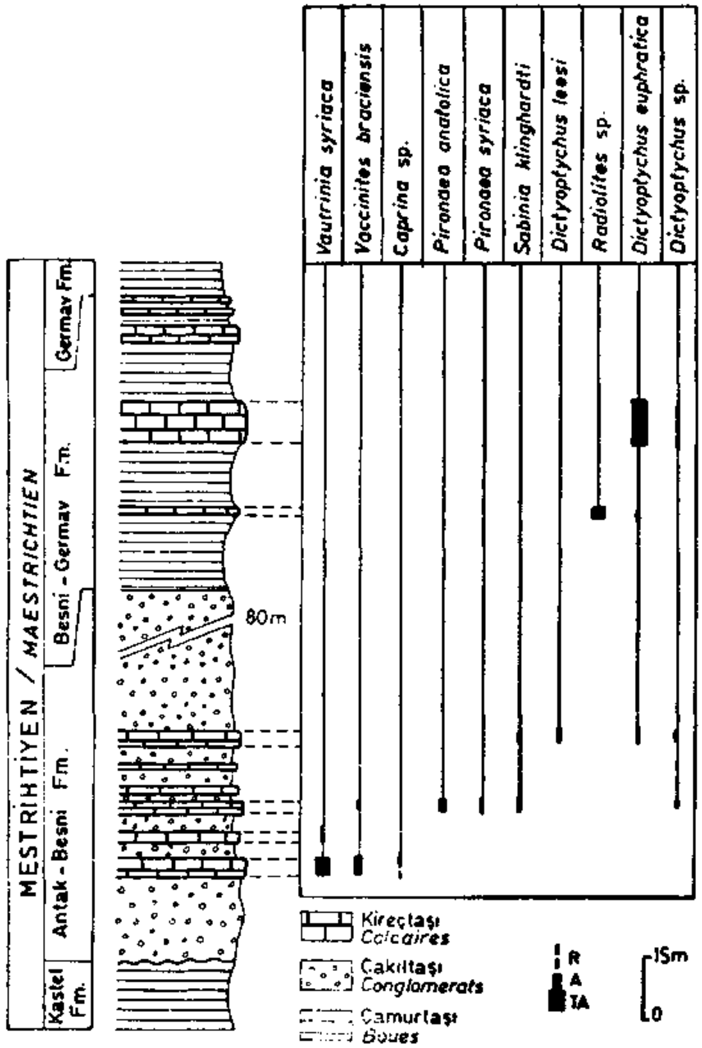


Fig. 2 - Distribution des Rudistes dans la coupe stratigraphique d'Alidamı. Echelle d'Abondance-Dominance: TA-Très abondant; A- Abondant; R- Rare.

Certains especes determinees ont ete trouvees dans d'autres localites maestrichtiennes de la Turquie: *V. syriaca* a Gaziantep (Stchepinsky, 1946) et au sud d'Antakya (Dechaseux, 1954; Dubertret, 1966; Karacabey-Öztemür et Selçuk, 1981); *V. braciensis* ä Hekimhan (Sladic-Trifunovic, 1967), ä Darende-Malatya (Karacabey-Öztemür, 1976) et ä Hereke-Kocaeli (Özer, 1986); *P. syriaca* ä Gaziantep (Stchepinsky, 1946) et au sud d'Antakya (Dubertret, 1966; Karacabey-Öztemür et Selçuk, 1981); *P. anatolica* ä Yazıhan-Malatya (Karacabey, 1970); *S.klinghardti* au sud d'Antakya (Karacabey-Öztemür et Selçuk, 1981) et ä Hereke-Kocaeli (Özer, 1986); *H. cornucopiae* dans l'Anatolie de l'Est (Karacabey-Öztemür, 1976) et Centrale (Özer, 1985b).

Certaines autres sont signalees dans le Maestrichtien des pays voisins: *D. striatus* en Iran (Douville, 1910); *V. syriaca* en Syrie sepfenttionale (Dechaseux, 1954; Dubertret, 1966); *H. cornucopiae* en Iran (Kühn, 1933; Douville, 1910); *D. leesi* ä Oman-Peninsule d'Arabe (Kühn, 1929).

Parmi les especes determines de la region etudiee, seules *V. braciensis* et *H. cornucopiae* offrent une repartition dans les pays balcans, en Italie, en Sicile et dans les Alpes Orientales; *V. braciensis* a ete trouve dans les formations Maestrichtiennes de Yougoslavie (Sladic-Trifunovic, 1967,1968; Plenicar, 1971; Sliskovic, 1971) et des Alpes Orientales (Sladic-Trifunovic, 1978). Tandis que *H. cornucopiae* a ete signale dans le Maestrichtien de Yougoslavie (Nedela-Devifle et Polsak, 1961; Sladic-Trifunovic, 1972), d'ltalie (Parona, 1900, 1916) et de Sicile (Matteucci et al., 1982; Camoin, 1983).

La faune etudiee nous a permis d'identifier quelques especes *(P. anatolica, V. braciencis, D. striatus, D. leesi)* et quelques genres *(Pseudopolyconites, Bournonia* et *Biradiolites)* qui sont signales pour la premiere fois dans l'Anatolie Sud-Est.

La faune determined aux environs de Kahta-Adıyaman nous mene a la dater definitivement du Maestrichtien.

REMERCIEMENTS

Je remercie sincerement M. Ozan Sungurlu grace a qui j'ai pu me servir de toutes les possibilites de travaille sur le terrain de TPAO, M. le Professeur Engin Meriç (İTÜ) qui a bien voulu collaborerpendant mes travaux de terrain et M. le Professeur Erol Akyol (DEÜ) pour ses conseils et critiques.

Mes remerciements vont aussi a M. Erol Şanlı qui a pris soigneusement les photes et a M. Mualla Gürle qui a dessine les figures.

Manuscript received April 2, 1986,

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PLATES

PLANCHE - I

Vautrinia syriaca (Vautrin) Milovanovic

Fig. 1 - Section transversak de la valve inferieur, commissure inconnue, X 3/4, Alidamı.

Ep, Sp - pseudo-piliers

ma, mp - apophyes myophore antcrieure et posterieure

br - hranche rayonnante de tissu prismatique

c - colonettes

Pironaea syriaca Vautrin

Fig. 2 - Section transversale de la valve inffrieure, 20 mm au dcssous de la commissure, X 3/4, Alidamı.

Vaccinites braciensis Sladic-Trifunovic

Fig. 3, 4 - Sections transversale de la valve inferieure, 10 mm au dessous de la commisure. x 1, Alidamı.

L - arete ligamentaire

S, E - premier et second piliers

B, B' - dents de la valve supcrieure

N - dent de la valve inferieure

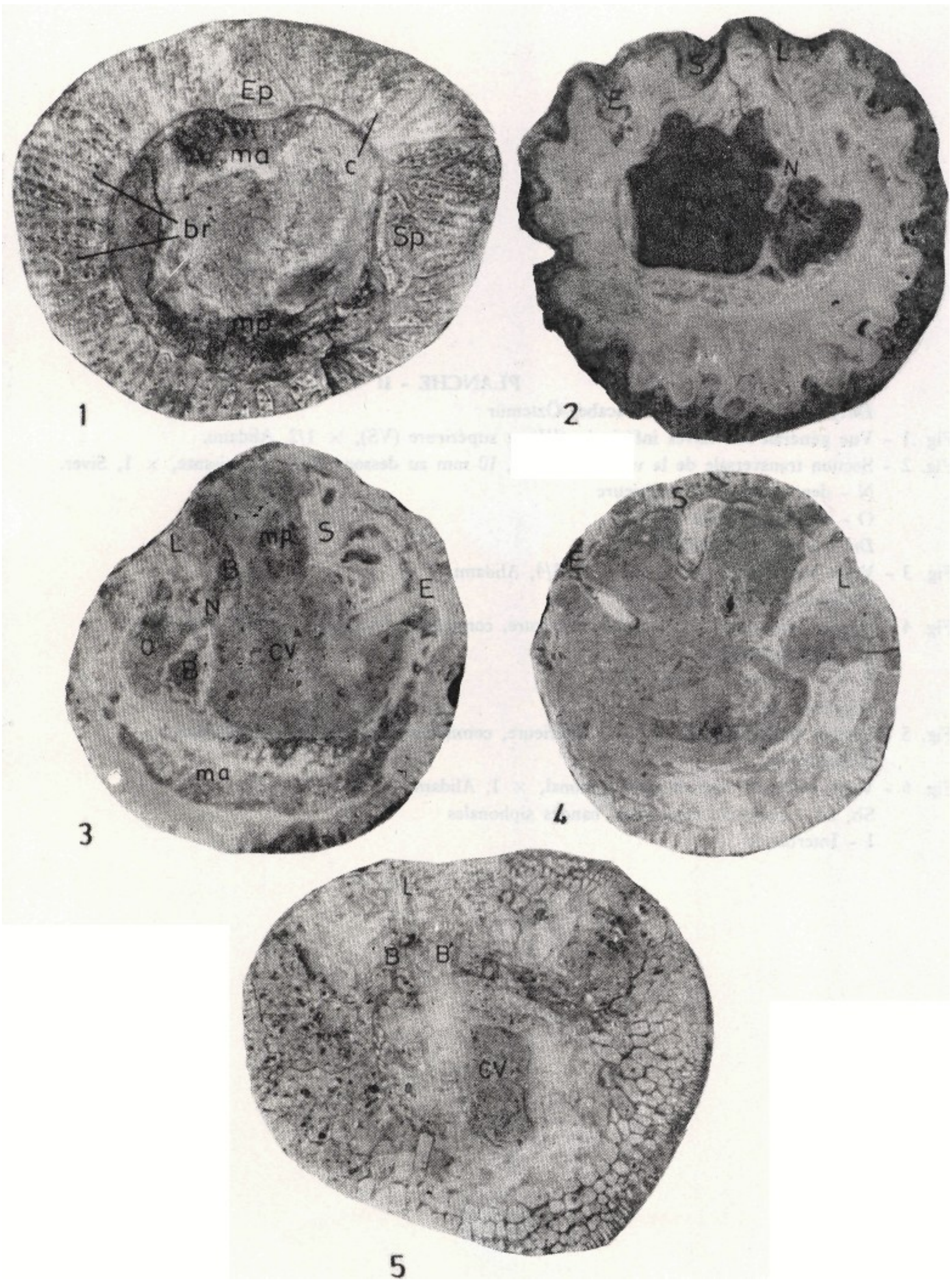
CV - cavite principale

O - cavite accessoire

Sabinia klinghardti Böhm

Fig. 5 - Section transversale de la valve superieure, vuedubas, 15 mm au dessus de la commissure, x 3/4, Alidamı.

On distingue deux type de canal (Polygonale irregulier et radiale).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CV
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PLANCHE - II

Dictyoptychus euphratica Karacabey-Öztemür

Fig .1 - Vue generale des valves inferieure (VI) et superieure (VS), x 1/2, Alidamı.

Fig. 2 - Section transversale de la valve inferieure, 10 mm au dessous de la commissure, x 1, Siver.

N - dent de la valve inferieure

O - cavite accessoire

Dictyoptychus leesi Kühn

Fig. 3 - Valve superieure, vue du haut, x 3/4, Alidamı.

Pironaea anatolica Karacabey

Fig. 4 - Section transversale de la valve inferieure, commissure inconnue, x 3/4, Alidamı.

L - aret ligamentaire

S, E - premier et second piliers

Hippurites sp.

Fig. 5 - Section transversale de la valve inferieure, commissure inconnue, x 1, Eskikahta.

Radiolites sp.

Fig. 6 - Valve inferieur, vue du cote siphonal, x 1, Alidamı.

Sb, Eb - premiere et seconde bandes siphonales

I - Interbande

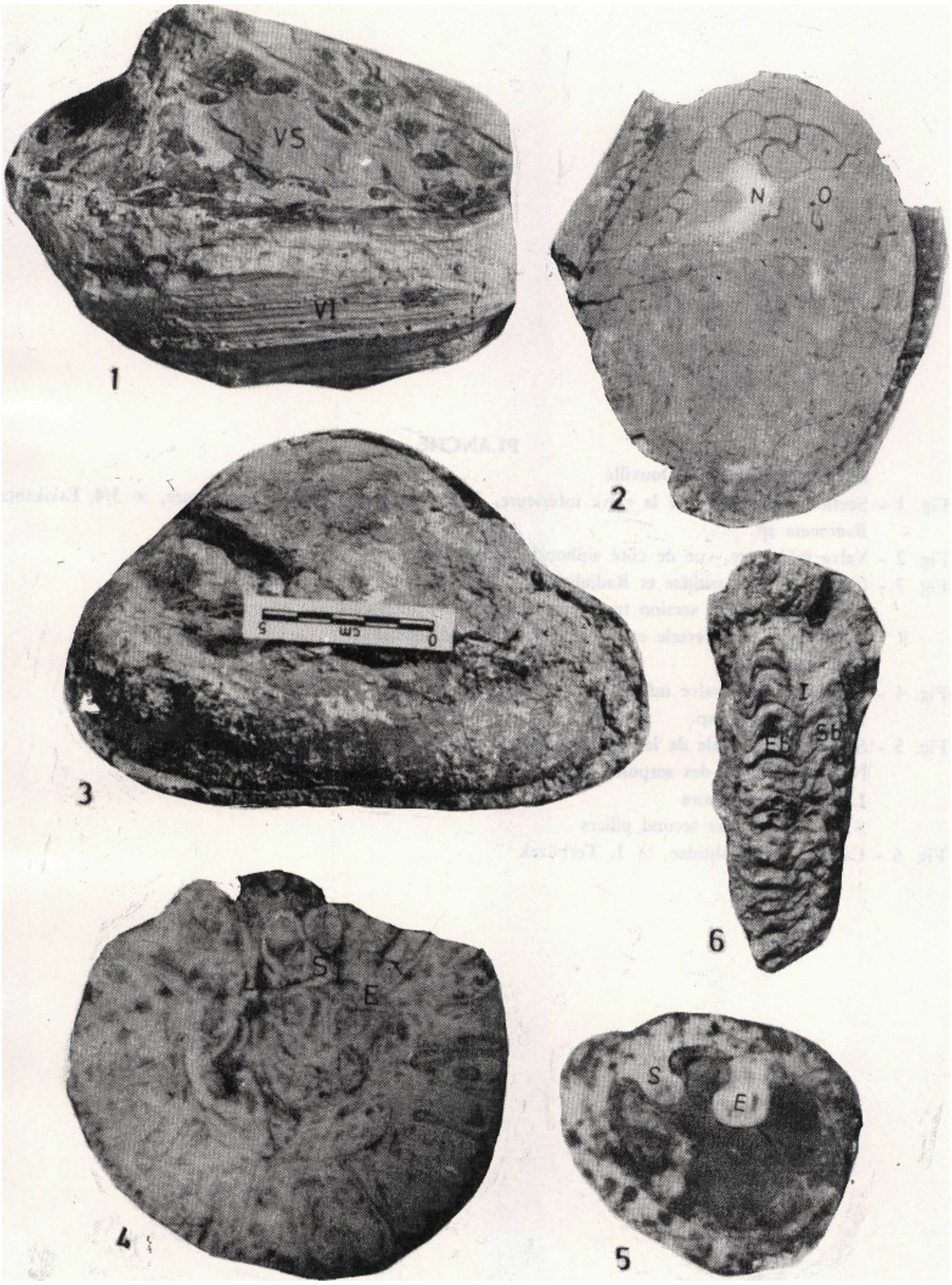
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PLANCHE - III

Dictyoptychus striatus Douville

Fig. 1 - Section transversale de la valve inferieure, 5 mm au dessous de la commissure, X 3/4, Eskikahta.

Bournonia sp.

Fig. 2 - Valve inferieure, vue de cote siphonal, x 3/4, Eskikahta.

Fig. 3 - Calcaire a Hippuritidae et Radiolitidae, x 3/4, Kıran.

a - *Hippurites* sp., section transversale de la valve inferieure

b - sections transversale et longitudinale de Radiolitidae

Lapeirousia sp.

Fig. 4 - Fragment de la valve inferieur, x 3/4, Kiran.

Pteudopolyconites sp.

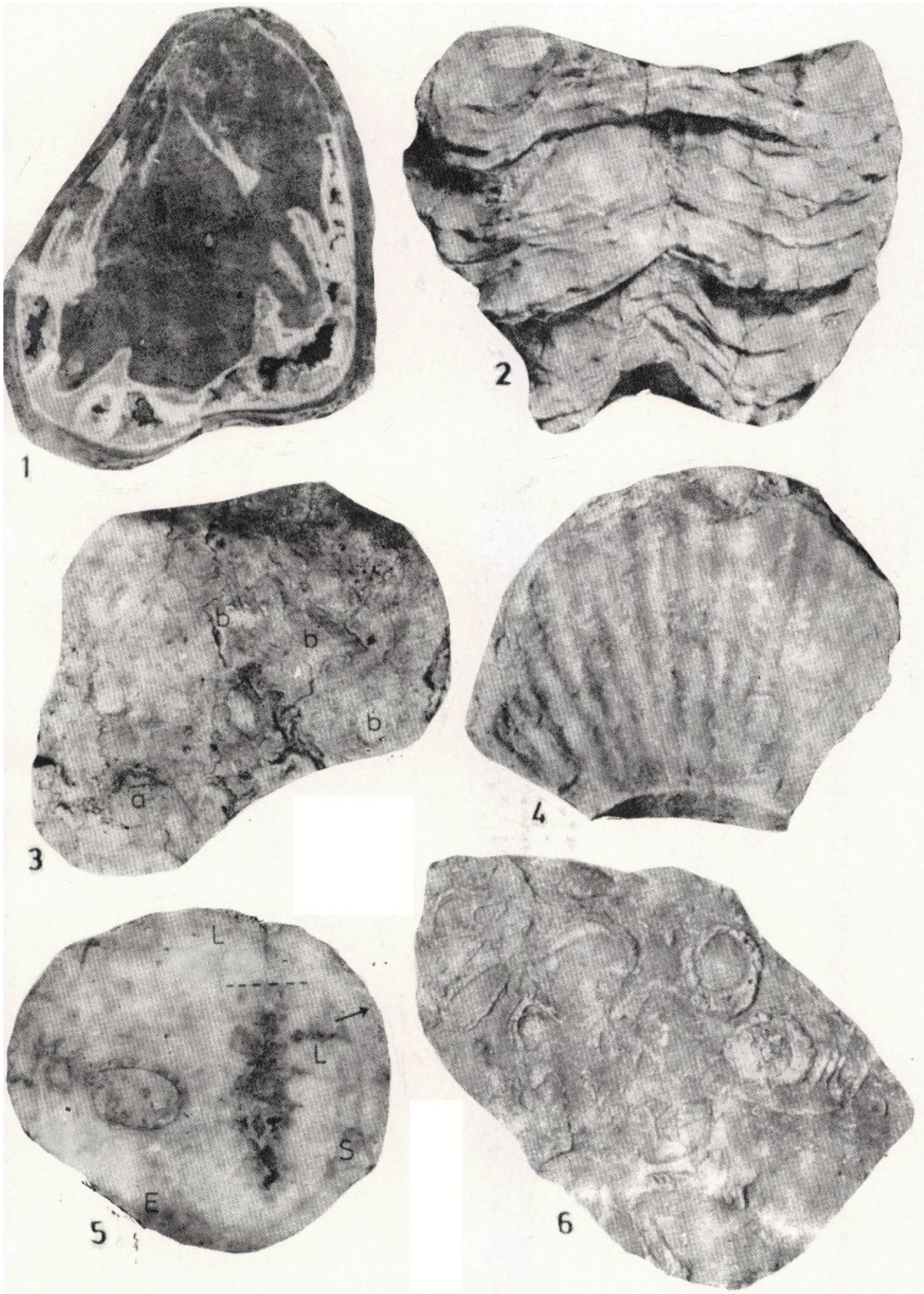
Fig. 5 - Section transversale de la valve inferieure, commissure inconume, X 1, Eskikahta.

Noter les coupes des serpules (fleche).

L - aret ligamentaire

S, E - premier et second piliers

Fig. 6 - Calcaire a Radiolitidae, X 1, Terbüzek.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A NEW DISCOVERY OF THE LOWER CRETACEOUS IN ISTANBUL, TURKEY

Orhan KAYA*; Jost WIEDMANN**; Heinz KOZUR***; Ülker ÖZDEMİR****; Sacit ÖZER* and Louise BEAUVAIS*****

ABSTRACT. — Limestone-pebble conglomerates in the Kocaeli Peninsula have so far been widely accepted as either Triassic carbonates or basal elastics of the Upper Cretaceous. New evidence of age and stratigraphic relations indicate that part of the conglomerate, together with some shale and coralgal limestone units, formerly considered as being Triassic, is early Cretaceous in age.

INTRODUCTION

Previous authors (e.g., Endriss, 1926; Böhm, 1927; Erguvanlı, 1949; Altınlı, 1968; Altınlı et al., 1970; Özdemir et al., 1975; Zaninetti and Dağer, 1978) considered the post-Triassic rocks of the Kocaeli Peninsula, with particular reference to the surroundings of Gebze (Fig. 1), as being late Cretaceous in age. The only exception is Arthaber's (1914) record of «probably Liassic» brachiopods *(Spiriferina moeschi* Haas and *Terebratula* cf. *punctata* Sowerby) in gray limestones intervening between the Triassic and Hippurites-bearing limestones».

In the present paper, the limestone-pebble conglomerate, shale and coralgal limestone units younger than Carnian Halobia-shales and older than rudistid-bearing Upper Cretaceous beds, have been assembled in a heterogeneous rock unit, namely the Çerkesli formation.

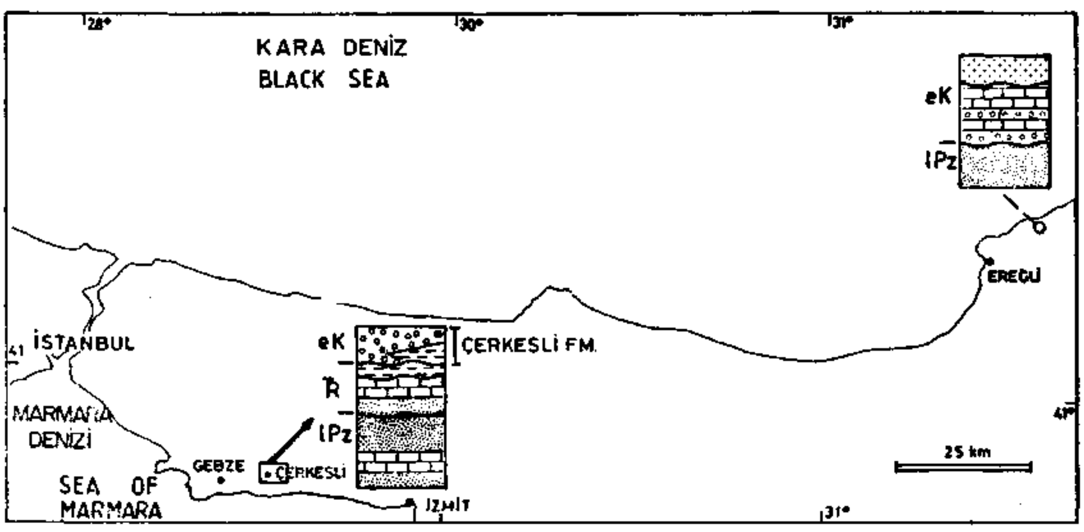

Fig. 1 - Lower Cretaceous sequences in the study and Ereğli areas.

STRATIGRAPHY

Çerkeşli formation

The name Çerkeşli formation is here used for the sequence of gray limestone-pebble conglomerate, shale and minor coralgal limestone (Fig. 2-4). Each major rock division of the formation is established as an informal member: the conglomerate, shale and limestone members. The composite type section is based on the type sections of the members.

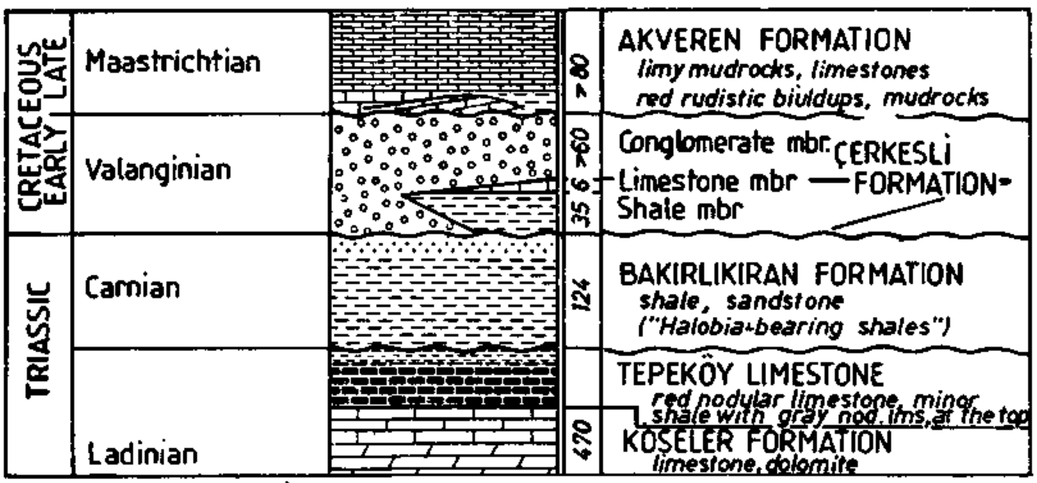

Fig. 2 - Generalized columnar section of the study area. The Triassic stratigraphic elements introduced here are based on Kaya and Kozur (in prep.).

The Çerkeşli formation rests unconformably on the Triassic rocks, with limestone-pebble conglomerate derived almost entirely from the Triassic carbonates.

The foraminifers in the matrix of the limestone-pebble conglomerate and an ammonite in the shale indicate an early Cretaceous (Valanginian) age for the Çerkeşli.

The limestone-pebble conglomerate is lithologically identical to the conglomerate interlayers in the early Cretaceous, İnaltı limestone typically exposed in the Ereğli area. Tokay (1952) and Kaya et al. (1984) suggested the Valanginian-Aptian and Aptian ages for the İnaltı limestone, respectively.

Conglomerate member. — The conglomerate member consists of gray limestone-pebble conglomerate and very subordinate pebbly limestone. It is divisible into a lower and an upper part. The type section representing the lower part of the member is exposed at 16.15:21.65 (Fig. 4).

The conglomerate is moderately to well indurated, massive to thick-bedded, grain-supported, closely packed, and moderately to well sorted in size. The clasts are round to subrounded, in average 4 to 8 cm in size, and include primarily Triassic limestones. The interstitial material is mainly gray microcrystalline limestone, which locally occurs as thin interlayers, and less commonly yellowish gray weathering limy claystone.

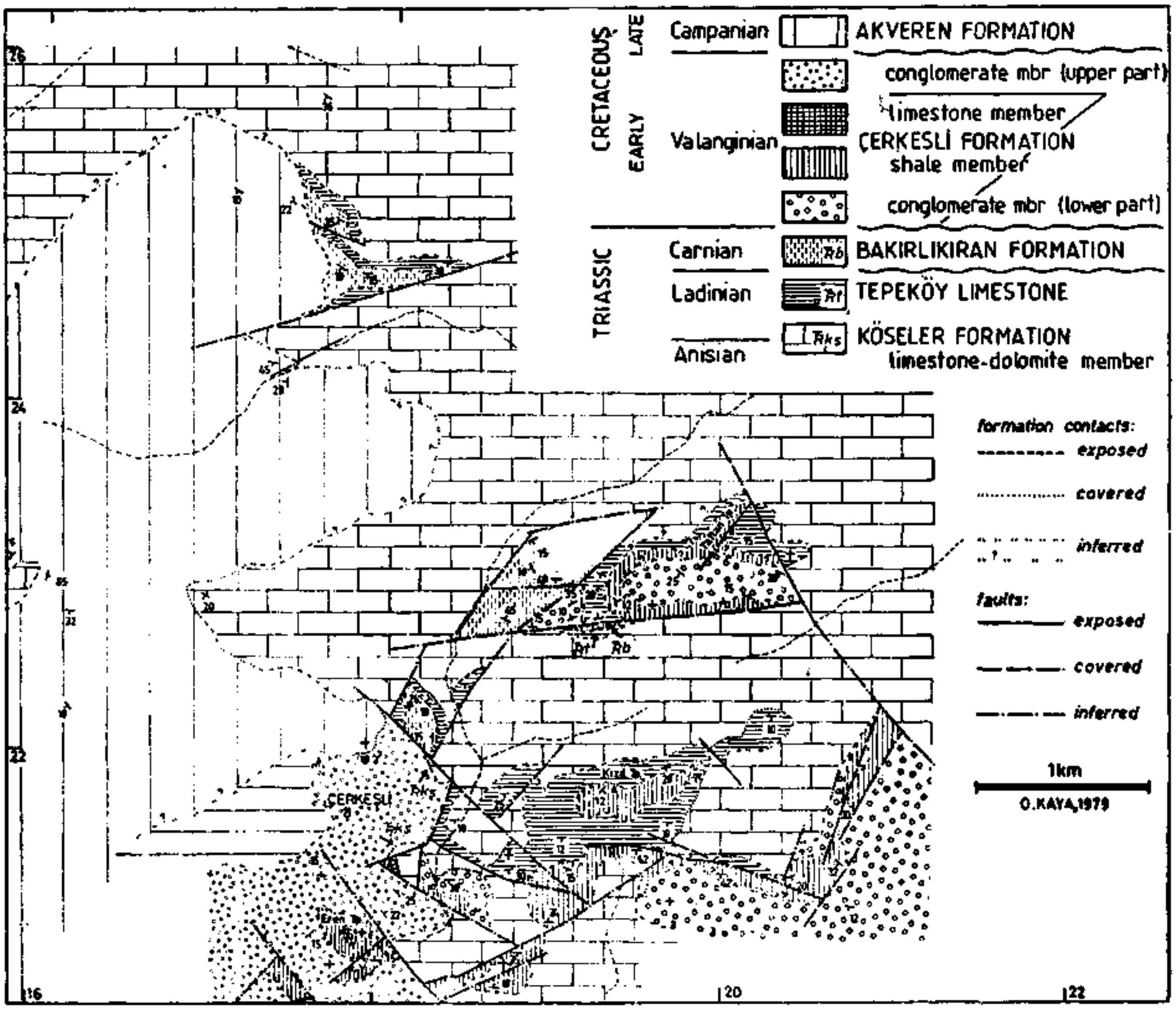

Fig. 3 - Geologic map of the study area.

The lower part of the conglomerate member rests on the Middle Triassic carbonates (18.12: 21.42; 17.90:20.76; 18.61:21.38). It is locally overlain by the shale member. The upper part of the conglomerate member overlies both the shale member (17.90:20.76) and the limestone member (17.88:20.98). The contacts are abrupt, structurally conformable, and represent stratigraphic breaks.

The limestone matrix of the upper part of the conglomerate member yields foraminifera *Cyclammina* sp., *Gavellinella* sp. and *Trocholina* cf. *valdensis* (Reichel), the latter indicating a Valanginian age. The coral *Cladophyllia dichomota* (Goldfuss) occurs as reworked masses up to small block in size. It is known only from the Upper Jurassic of Germany and France.

In the map area, the conglomerate member is indicated by Erguvanlı (1949) as «semi-conglomeratic limestone», and by Altınlı et al. (1970) as the Hereke formation, both being early to Middle Triassic in age.

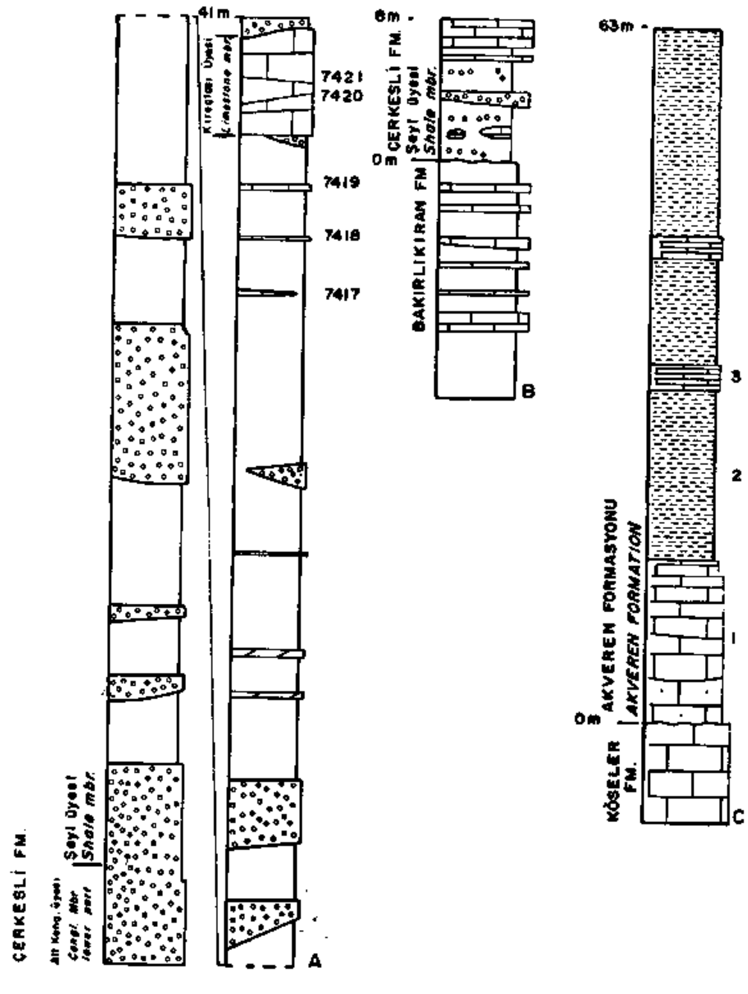

Fig. 4 - Composite type section of the Çerkeşli formation (A, B); reference section of the Akveren formation, and its basal interval on the Triassic rocks (C).

Shale member. — The member consists of gray shale with widely spaced thin interbeds of limestone, and laterally discontinuous interlayers of limestone-pebble conglomerate. The composite type section is located between 18.03:20.97 and 18.00:20.98, and at 17.91:21.00 (Fig. 4). A reference section is situated between 20.40:21.35 and 20.62:21.32.

The shale is homogeneous and moderately indurated. The limestone-pebble conglomerate is poorly indurated and lithically similar to the conglomerate member. However, it contains abundant reworked clasts of limestone -pebble conglomerate. The limestone is dark gray, thinly bedded, fetid, clayey and fine-grained. The limestone beds in the lowermost part of the member contain fine-grained limestone pebbles.

The shale member rests on the Triassic carbonates (19.25:22.80; 19.38:21.90) and the shales of the Triassic Bakırlıkıran formation (18.27:22.80; Fig. 4B). The contact between the shale member and the lower part of the conglomerate member is abrupt and structurally conformable (18.02:20.80, 18.12:21.42), and suggests a synsedimentary, erosion.

An early Cretaceous lytoceratid *Protetragonites quadrisulcatus* (d'Orbigny) indicates an upper age limit of Valanginian.

The early workers referred to the shale member as the Triassic Halobia-shales.

Limestone member. — The member consists of small-sized coralgal buildups and related detritus, randomly scattered in the shale bulk of the Çerkeşli formation. The type outcrop of the member is situated at 17.90:20.76 (Fig. 4A). The other exposures are located at 19.02:21.18; 20.75:18.40; 18.85:22.87.

At the type exposure the core is about 3 m in vertical extent, and has an algal and coral frame supported by bioclastic limestone. Vertically and laterally it changes into stratified, fine-grained and bioclastic limestone. In other localities the core may be represented by a unique colony of corals, or by structureless dolomite. The corals of the genus *Cladophyllia* d'Orbigny seem to be the most common and identifiable elements.

The contact between the limestone member and the enclosing shales is abrupt (17.91:21.00). It may be represented between the limestone-conglomerate interlayer of the shale member and the limestone member (17.88:22.90; Fig. 4A).

Akveren formation

Ketin and Gümüş (1963) applied the name Akveren formation to a sequence of mainly white, thin to thick-bedded, calcareous to limy mudrocks and limestone. The type section is outside the study area. A reference section of the Akveren is exposed between 16.63:23.05 and 16.10:22.95 (Fig. 4C).

The basal section of the Akveren formation is characterized by different and laterally interchanging rocks, lying directly on the Triassic rocks: (1) yellowish gray-weathering, thickly-bedded to massive bioclastic limestone (18.03:23.05; Fig. 4C); (2) pink to pale red rudistid patch reefs (17.42: 25.38); (3) light gray to grayish green, and pale red mudrocks (17.25:24.35) and (4) pale red limestone conglomerate with interlayers of rudistid debris (Hereke conglomerate: Erguvanlı, 1949).

Böhm (1927) documented in detail the faunal content of the basal beds of the Akveren formation in the investigated area, and concluded a Campanian age. Erguvanlı (1949), Altınlı (1968), and Altınlı et al. (1970) adopted this age assignment. In the map area, the rudistid patch reefs and detritus include *Pironaea timacensis* Milovanovic *Vaccinites braciensis* Sladic-Trifunovic, *Sabinia klinghardti* Böhm, *Schosia bilingius* Böhm, *Hippurites* cf. *cornucopiae* Defrance and three new species of the *Gorjanovicia* Polsak (G. *akyoli, G. kayae, G. polsaki:* Özer, 1982), which indicate a Maastrichtian age.

Manuscript received July 14, 1986

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GEOPHYSICAL STUDIES ON THE RESEARCH OF GEOTERMAL FIELDS IN TURKEY

Çetin ŞENEL**; Remzi ERDOĞAN** and Mehmet Emin ÖZGÜLER**

ABSTRACT. — Turkey is located on the orogenic, active, tectonic Mediterranean belt where frequently young volcanic activities are observed during the Alpin orogenic phase strong fracturing, tectonics, magma uprise followed by magma chamber locating in the earth crust played important role in the existence of the geothermal systems. Geophysical surveys have very important role in the exploration of geothermal fields in Turkey. Very impotant information is obtained about the location depth and formation of the re servoir systems by applying several basic geophysical methods according to the fields characteristics. Since the electrical resistivity values decrease in the rock formation incontact with saline water and hot water in a geothermal field electrical resistivity methods are widely used in Turkey too. For reconnaissance of the location and the dimensions of shallow geothermal reservoirs shallow temperature gradient method which is directly applicable like electrical resistivity method is proven to be a fast and cheap method according to the studies carried on in the field. With the equipment available in Turkey geophysical surveys carried out up till date makes it possible to get information from a limited depth which is in the order of few kilometers. Therefore no geophysical data is obtained dealing with the neat source which is much deeper. Application of geophysical surveying methods such as heat flow and' much deeper magnetotelluric methods is necessary for both improving the existing proven geothermal system and also exploration of consealed geothermal systems with no surface manifestation like hydrothermally altered rocks, hot spring ete.

PRESENCE OF THE MIDDLE-UPPER TRIASSIC IN THE AUTOCHTONOUS GEYİKDAĞI UNIT OF THE EASTERN TAURUS (SARIZ-TUFANBEYLİ REGION, KAYSERİ)

Baki VAROL*; Nizamettin KAZANCI* and Demir ALTINER**

Mesozoic carbonates of Sarız-Tufanbeyli area located in the eastern extension of the Geyikdağı unit which represents the autochton of the Taurus belt. These carbonates form a NE-SW trending belt of 20 km width limited by Binboağa and Soğanlı dağı allochtons. In the studies of Demirtaşlı (1967), Özgül et al. (1973), Erkan et al. (1978), Aziz et al. (1980, 1982), Metin et al. (1982) which present the stratigraphy and geologic maps of the region, the Lower Triassic corresponding to the Katarası formation has been recognized and the Middle to late Triassic has been interpreted as a period of emergence and erosion (Fig. 1). However, there are some facies developments of Middle to Upper Triassic age interrelated between the Lower Triassic (Katarası formation) and the overlying Köroğlutepe formation of Jurassic to Lower Cretaceous age. Although some parts of these facies are difficult to recognize because of the intense dolomitization, the parts which escaped from dolomitization exhibit mappable outcrops in the region. The Middle-Upper Triassic recognized in three measured stratigraphic sections (Fig. 1) shows the following facies characteristics:

— In Yüceyurt section, *Endothyra* sp. and dasycladacean algae in the cavities of the green algae biolithids, *Endothyranella* sp., *Trochammina* sp. and Duostominidae in the foraminiferal mudstones and Sphaerecodium oncoids in the oncoidal packestones-wackestones represented by nodular limestones;

— In Gümelek tepe section, traces or well preserved specimens of *Aulotortus* gr. *sinousus* Weynschenk, *Aulotortus* sp., *Trochammina* sp., Nodosaridae in the dolomitic limestones alternating with massive dolomites;

— In Ayvat section, *Nodosaria ordinata* Trifanova in the foraminiferal packestones, *Endothyranella* sp., *Griphoporella cur-vat a* Gümbel and Prostromate algae with a complex internal structure in the algal-foraminiferal mudstones-wackestones.

Similar reef facies which are recognized among these various facies types defined by uppermost Middle and Upper Triassic associations (Zaninetti, 1976) have been also described from the Panaromi carbonate platform in Sicily (Abbate et al., 1977) and Sphaerocodium oncoids have been illustrated from the German and Austrian Alps (Peryt, 1977).

In conclusion, the presence of a reefal and dolomitic Middle-Upper Triassic sequence has been proved in the Triassic section of the Geyikdağı autochthon which is thought to be composed of only the Lower Triassic rocks.

Manuscript received September 17, 1986

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