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A CRITICAL REVIEW OF THE ANATOLIAN GEOLOGY: A DIALECTIC TO SUTURES AND EVOLUTION OF THE ANATOLIAN TETHYS AND NEOTETHYS

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ABSTRACT.- A progressively improving hypothesis of evolution for the Anatolian Tethys and Neotethys will be presented in this paper. The Tethyan, the Western (Bursa-İzmir-Antalya zone) and the southern Neotethyan sutures will be substantiated after a discussion on the controversial aspects of the Anatolian geology. The initiation of the Tethys in Cambrian is supported on the basis of continous Palaeozoic sedimentation on north-facing Gondwanian and south-facing Eurasian platforms. A thin continental sliver, the Apula-Anatolia, started to rift off northern Gondwana in the Early Triassic coevally with the onset of the northward Tethyan subduction. The marginal ophiolites have obducted onto the Pontian active margin by the Middle Triassic as the consequence of the dextral rotation of Western Pontides. The compressional and dilatational fields have shifted oceanward due to a recess (?) of the subduction zone. The positive area covering most of the central and northern Sakarya has collapsed progressively and has been onlapped peripherically from the Liassic onwards. The Anatolian microcontinent detached off Africa in the Upper Triassic-Liassic and drifted northward during the Jurassic and the Cretaceous and collided incipiently with the Eurasian margin (Pontides) in the Upper Cretaceous. Unsubducted pockets of ocean floor have closed with consequent syn-collisional magmas in the Paleogene. The Salt Lake pocket, the East Anatolia and the Western Neotethys, the Intra-Gondwanian rift separating the Anatolia from Apulia-Greece, have survived until the Late Miocene.

Key words: Controversial aspects of the Anatolian geology; sutures of Anatolia; geological and geophysical constraints; possible evolutionary frames; geologic evolution.

INTRODUCTION

The stratigraphy of Anatolia is characterized by many Paleozoic - Mesozoic sections overlying unconformably overlying the Precambrian basement of the Pangea. Had there been no Alpine events, Anatolia would present sceneries similar to the grand canyon of USA. There are many sections implying such a stratigraphy. Southern Menderes Massif, Central Taurids (Cambrian-Miocene) and Bitlis yield examples of this stratigraphy.

The sutures of Anatolia, based essentially on the teachings of Brinkmann (1972) and of many others, has first been published by Sengun et al (1990), when the East Anatolia was suggested to be a continental fragment, in the sense that it is not an accretionary prism, and the Tethyan suture was suggested to tie to the Sevan-Akerra considering the Munzur-Taurus connection. The Bursa-Antalya zone has later been designated as the Intra-Gondwanian zone separating the Aegean plate from the Anatolian. This has been the most important revision on the sutures of Anatolia.

The Tethyan frames, based essentially on the Atlantic Ocean data (Smith, 1971; Pitmann and Talwani, 1972 and Dewey et al., 1973), agree on the theory that slivers of continental crust have rifted off northern Gondwana and drifted north to collide with Eurasia (Stocklin, 1974, 1977; Adamia et al., 1977; Biju-Duval et al., 1977 and Dercourt et al., 1986). The essence of this theory is applicable also to the Anatolian segment of the Tethyan belt in agreement and in liaison with the neighbouring areas (Stocklin, 1977 and Biju-Duval et al., 1977). The post-Liassic part of the evolutionary frame presented in this paper is almost world-wide accepted.

* The author unfortunately has passed away during the processing of the manuscript. The corrections and the changes were carried out by his colleagues and very close friends who know him very well in a way to be consistent with his thoughts and beliefs . The author has completed his work span in MTA and performed invaluable scientific performances in Geology. We remember him with a great mercy.

On the other hand, the very diverse disputes on Anatolia will be discussed in this paper towards a plate tectonic model for the Anatolian Tethys/Neotethys by mounting the discussed evidence on a basic and generalised frame of evolution. Unfortunately, what has been published about the Anatolian geology is an intermingled and living bundle of imaginary hypotheses. Many quests have naturally arisen on the existing theories/interpretations with consequential revisions and corrections as the field evidence has progressed. Some of the controversies will be discussed in this paper, hoping that a step will be taken towards the final solutions.

A CRITICAL REVIEW OF MAJOR DISPUTES

The Eastern Mediterranean and the Tethyan disputes will be discussed below with some of the entailed arguments, which are crucially related to Tethyan/Neotethyan evolution of Anatolia.

The Eastern Mediterranean dispute

The main controversial issues have been the age of rifting of the Eastern Mediterranean and the origin of 'Antalya nappes'. It has been suggested that the Eastern Mediterranean has not rifted until the Cretaceous and the 'Antalya nappes' have originated from northern Anatolia, the northern margin of the Gondwanaland (Ricou et al., 1974-1986; Dercourt et al., 1986). This theory is in debate with the theory of Triassic age of rifting of the Eastern Mediterranean and southern origin of Antalya nappes (Robertson and Woodcock, 1981; Robertson and Dixon, 1984; Poisson, 1984; Özgül, 1984 and Yılmaz, 1984). Structural and stratigraphic evidence is in full compliance with the Early Triassic rifting and the author is in full agreement with those who defend the Antalya complex to consist of Tertiary imbrications of the continental margin and the marginal Neotethyan ocean floor of the Antalya region.

The Anatolia fragment must have rifted off northern Gondwana harmoniously with what has been suggested by Stocklin (1974, 1977) and Biju-Duval et al (1977) respectively for Central Iran and Apulia-Greece. It is tied to Central Iran in the east and bounded by the Menderes massif in the west while the Karaburun-Biga must be the eastern margin of Apulia-Greece. The Western

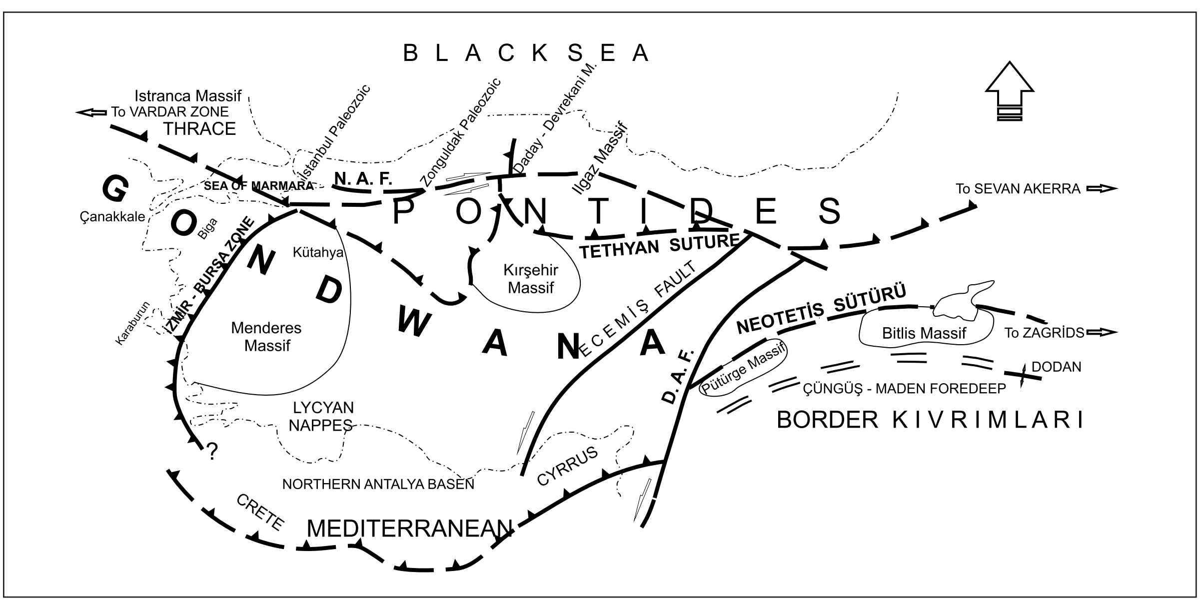

Figure 1- Sutures of Anatolia (Revised after Şengün et al,1990).

Neotethys or the Bursa-İzmir-Antalya zone comprises Triassic sediments as the base of a Mesozoic sequence (the Antalya nappes). This shows that the rifting is not directional but scattered in northern Gondwana with the implication that initiation of rifting of Apulia and Anatolia is contemparaneous although the detachment may be not. The rifting of the southern Neotethys has been ascribed to the drag caused by the northward diving Tethys (Robertson, 1990). This is agreeable and seems an effective factor in addition to the essential control, rifting of the Atlantic.

The Antalya complex is the prototype of Antalya nappes and is well studied (Lefevre, 1968; Robertson and Woodcock, 1981; Poisson, 1984; Özgül, 1984 and Yılmaz, 1984). Unfortunately, rifting margins of northern Gondwana have also been defined as Antalya nappes as in the cases of the central and eastern Alanya massif, Biga, Karaburun-İzmir or Kütahya. The Mesozoic fining-upward sediments of these areas are unconformable/gradational on the Palaeozoic, with the implication that a continental crust bases them. The Palaeozoic sequences of the rifting margins are similar to that elsewhere in the Taurids, while the Mesozoic sections reflect unstable conditions caused by the neighbouring rifting. The Lower Palaeozoic is represented by an alternating series of carbonates, quartzites and pelites while thick carbonate sections are encountered in the Devonian and the Permo-Carboniferous. There is generally a sharp facies change into turbiditic sediments in Schytian along rifting margins. The Antalya nappes are characterised by a flyschoid sequence with intermittent basic volcanism during Carnian-Norian (Antalya and Karaburun) followed by deep marine carbonates of Jurassic-Cretaceous age.

The allochtoneity of the Antalya nappes, including the rifting margins of the Neotethys, has almost been unquestioned and the discussions have been focused on where they had come from (Brunn et al., 1975). Southern origins have been defended for Alanya and Antalya (Poisson, 1984; Özgül, 1984) disputing the northern origin defended by Ricou et al., (1974-1986).

The Alanya debate

Özgül's (1976) structural analysis, a widely accepted tectonic model for the Taurids, defends allochtonous entities or tectono-stratigraphic units that are piled up on one another. These are Geyikdağ, Bozkır, Bolkardağ, Antalya, Alanya and Aladağ. The structural setting of the Alanya with respect to the Antalya unit will be discussed below for description of the ongoing dispute. Evidence will be presented to defend that the Antalya and Alanya units are in situ, versus floatation of Alanya on an also allochtonous Antalya (Ricou et al., 1974; Özgül, 1976-1984; Ulu, 1983). A northeast section (Figure 2) from Demirtaş to the flyschcorridore (the Antalya unit) is described below essentially through the author's window with occasional reference to and discussion on the views of the contraveners.

Carbonates and slates, in the southern part of the section (Point 1 in Figure 2), dated respectively as Cambrian and Ordovician (Özturk et al., 1995), constitute the overturned southern flank of an asymetric anticline in the core of which garnetiferous micaschists are exposed. The micaschist has the following generalised paragenesis: Quartz+Muscovite+Garnet+Mg-chlorite+Albite/Oligoclase±Biotite. The physical conditions of almandine-amphibolite facies or of the medium grade (Winkler, 1974) implied by this paragenesis are certainly incomparable with those of the overlying incipiently deformed Palaeozoic rocks that are exposed in rest of the section. The Cambrian carbonates have been locally sheared out in the northern flank, whereupon the garnetiferous micaschist has come into tectonic contact with the Ordovician-Silurian shales/slates. The sheared contact zone between the micaschist and the overlying shales/slates displays a conformable fabric. However, the compositional change is very sharp, implying the impossibility of a gradation in physical conditions of metamor-

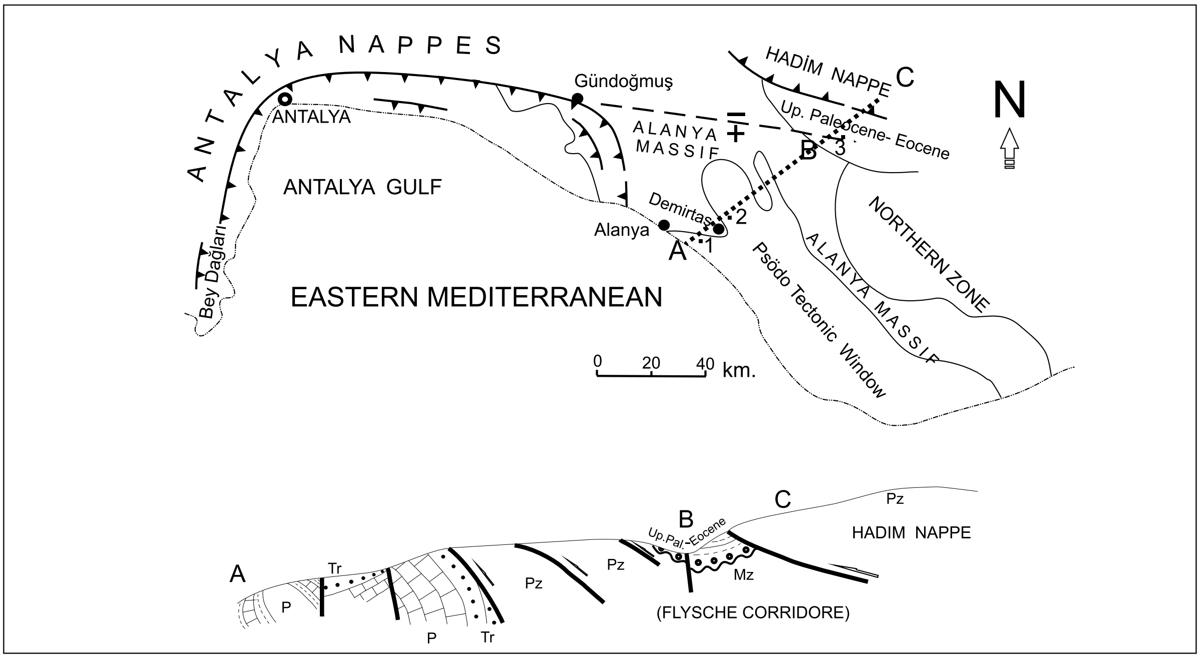

Figure 2- A tentative sketch map and cross-section from Alanya to the Hadim nappe via the flysch corridore, to show the relations between the tectonic (?) units of Alanya and Antalya. The westernmost Alanya, until the town of Gündoğmuş, is the stratigraphicand the structural continuation of the Antalya complex, the present morphology of the Antalya gulf being the result of the 30° sinistral rotation of the Beydağları (Robertson, 1990) and dextral rotation (?) of the eastern side.

phism. Therefore, an unconformity, in analogy to other Precambrian basements, seems to be the plausible relation.

The micaschist comprises lenses of amphibolites consisting essentially of pyroxene, amphibole and pyrope-almandine rich garnets. Block formation is presumably due to the extreme incompetency of the micaschist. The amphibolite lenses have been interpreted as alpine eclogites on the basis of the omphacitic (very close to diopsitic augite) composition (Okay and Özgül, 1982) of the pyroxene. However, lenses with the assemblage of pyroxene+amphibole+almandine are well known and commonly encountered in the Precambrian basements of Anatolia. They are also peculiar to Eurasian-Gondwanian Precambrian basements in Europe as well as Africa or Arabia. These rocks are hydrous and have a high plagioclase-amphibole content, which is not acceptable for eclogites. It is customary to define these rocks as garnetiferous amphibolites.

However, the more important question is whether or not the micaschist-amphibolite assemblage is part of a basement underlying unconformably a Palaeozoic-Mesozoic sequence. An analysis is attempted below on the relations of pseudo-tectonic units, Alanya and Antalya, in search of an answer to this question.

A vertical fault on the Demirtaş road juxtaposes the Triassic sandstones of the Antalya unit to the Lower Palaeozoic rocks of the Alanya unit of Özgül (1976), a few km northeast of the micaschists (point 2 in Figure 2). This fault has been evaluated as the base of the Alanya unit of Özgül (1976), while it is a normal (vertical) fault of minor importance to the author. The downthrow is not more than a few meters. The Ordovician shales-slates appear randomly below the soil cover on the downthrown side. The Triassic sandstones of the tectonic window lie horizontally on these rocks. These observations show that the Triassic sandstones overlie the Lower Palaeozoic slates unconformably and this fault causes the juxtaposition of these rock units. The base of this section comprises Cambro-Ordovician rocks (Öztürk et al., 1995), overlying the garnetiferous micaschists. These cannot be included in the Antalya unit, which is devoid of Palaeozoic sediments by definition. This section is not significantly different from the Geyikdağ or Aladağ unit of Özgül (1976) except that it has a Precambrian base. The deformations are restricted to Alpine shear zones. The Palaeozoic sequence is almost the same as that of any other location in the Taurids.

The Alanya unit of Özgül (1976) has been suggested to float on an entirely undeformed sequence. The Upper Permian pelmicrites, with well-preserved fossils, display a broad and symetric anticline in the suggested tectonic window, with a clear gradation to variegated shales of Schytian age as the base of a continous Mesozoic sequence (Ulu, 1983). This sequence is unconformable on Ordovician shales/slates that sit on the garnetiferous micaschists. The described sequence proves very clearly that the Antalya unit of the pseudo-tectonic window has a continental character.

The northern boundary of the tectonic window is a thrust fault dipping 30 to the north on the road exposure. This fault dies out on both sides according to observations of the author and is one of the several northward dipping ecailles of the Alanya. It dissects a gradational and continous Permian- Lower-Middle Triassic sequence. The Alanya massif (Palaeozoic rocks with incipient deformation) is overlain by the Triassic rocks of the tectonic window a few kms north of the northern boundary of the tectonic window. This outcrop has been defended as evidence against the allochtonous nature of this sequence. However, the contraveners interpreted it (Özgül, 1984) as an upthrow of Antalya onto the Alanya versus the earliest interpretation of a sedimentary contact (Blumenthal, 1951). The deformation in this region is almost nil and hardly different from that of the tectonic window. Furthermore, there are other exposures (Öztürk et al., 1995), Middle and Upper Triassic in age, sitting on Palaeozoic sediments in the northwesternmost Alanya, incompatible with the stratigraphy defended (Özgül, 1976 -1984) for the Alanya unit.

The examination of the northern boundary of the Alanya massif (Figure 2) yields crucial evidence. The Alanya massif has been thrusted imbricately onto the Cretaceous-Eocene section of the flysch corridore or the Antalya unit of Özgül (1976) in the westernmost segment. The boundary is a thrust trending E-W until the town of Gündoğmuş where it assumes a southward trend. Although the southward turn has been recognised informally by some, the boundary between the Alanya and the flysch corridore continues to be accepted as the previously defined EW trending thrust. This boundary continues from Gündoğmuş eastward, to the author, without any thrusting but with discontinuous normal faults with the northern blocks downthrown (Point 3 in Figure 2). There is Upper Paleocene-deposition throughout the eastern half of the northern boundary covering mutually the Alanya, the flysch corridore and the northern zone of Demirtaşlı (1984). On the other hand, the westernmost part of Alanya, as the eastern continuation of the Antalya Complex, has been thrusted imbricately onto the flysch corridore after Eocene. There has been no objection to this evidence, which shows very clearly that the central/eastern part of the Alanya is in situation. Being more elaborate, it proves that:

1- The Alanya could not have been transported relative to the flysch corridore (Antalya unit) after Paleocene with Paleocene-Eocene clastics on its back, because the cover is mutual on the Alanya and the northern zone (Geyikdağ unit).

2- The Antalya nappes of the tectonic window cannot be connected to that of the flysch corridore because of emplacement ages of pre and post Upper Paleocene. This means that the Alanya needs to have been emplaced after the youngest sedimentation of the pseudo-tectonic window, which is Upper Cretaceous in age (Ulu, 1983), and prior to the Upper Paleocene deposition. However, there is no evidence, a reason or an implication for a pre-Upper Paleocene event.

3- The absence of the Antalya unit between the Alanya and the northern zone of Demirtaşlı (1984) and the gradation between the Alanya and the Geyikdağ tectonic units is indicative of a normal stratigraphic order rather than piled up tectonic units.

On a rough asessment of the temporal dimension, the argument presented above excludes all but the Uppermost Cretaceous-Upper Paleocene interval for a possible allochtony. However, the Upper Cretaceous cap in northernmost part of central Alanya brings on a further constraint. The conglomeratic base is horizontal and has been interpreted as a faulted contact (Ulu, 1983), while it is a sedimentary one to the author. Whatever the relation is, the presence of Upper Cretaceous sedimentation on Alanya indicates dilatation in the Upper Cretaceous, quite incompatible with an overthrusting phenomenon. In fact, it marks the onset of a grabenization (the flysch corridore). This location needs to be observed by a third party for the crucial conclusion that the Alanya cannot be transported after the Upper Cretaceous, implying there is no interval for transportation.

The idea of allochtonous Alanya is a hyphothesis asserted in search of an answer to how Alanya has been metamorphosed. A chronological sequence of the theories for Alpine metamorphism of the Alanya massif is listed below to contribute to a further understanding of the Alanya debate. The Alanya unit is:

1- An allochtonous entity that has been deformed at northernmost Gondwana (Ricou et al., 1974).

2- An autochtonous asemblage with an alpine HP/LT deformation of the westernmost part (Şengün et. al. 1978).

3- An allochtonous assemblage of southern origin, a deformed active continental margin with glucophane bearing assemblages (Okay and Özgül, 1982)

4- An allochtonous entity metamorphosed in the island arc setting in the vicinity of Cyprus (Özgül, 1984).

5- An autochtonous mass in the central/eastern segment while the western part, suffering a HP/LT alpine metamorphism, is the eastern part of the Antalya nappes (Figure 2).

The author's perspective for the Alpine metamorphism is as follows. The Central Taurids, rotated dextrally by the Ecemis fault (Figure 1), compressed the marginal ophiolites in the vicinity of the Antalya gulf during the Paleogene. This has resulted in northward thrusting of the marginal ophiolites of the Antalya gulf onto the Beydağları and the westernmost Alanya. The northern margin of the Antalya basin has been imbricated with these thrust sheets in the area lying west of the town of Gündoğmuş. On the other hand, there is a gradual twist, from Gündoğmuş eastward, of north vergent folds to the south vergent of the central and eastern Alanya. This observation implies a dextral torsion that has been caused by the southward push of the western block of the Ecemis fault versus the northward thrusting of the western Alanya. It has folded and twisted the central Alanya so as to expose the Precambrian basement in the vicinity of the Alanya town. The post-Eocene obduction of ophiolites and imbricate northward thrusting of westernmost Alanya have caused a very-low grade (sensu Winkler, 1974) metamorphism (blue schist facies) in that region. The contemporary deformation appears as northward dipping shear zones in the central and eastern Alanya. The convertion of the Palaeozoic pelites to slates or phyllites are restricted to these shear zones. The alpine deformation on the Precambrian micaschists appears as imprints of shear planes, which intersect mutually the Ordovician shales/ slates. They appear as quartz-chlorite veins, which have a discordant relation to the Precambrian paragenesis in thin sections of the micaschists. Pseudomorphs of chlorite after garnet may also be considered as a frequently encountered symptom of the alpine imprint.

The westernmost Alanya is tectonically and stratigraphically similar to the western side of the Antalya gulf. On the other hand, the central and eastern part of the Alanya is not different stratigraphically from the other tectonic units of Özgül (1976), which differ from one another by facies changes and deformational variations only. In conclusion, the central and eastern Alanya is in situation.

The alpine metamorphism of the Tauric belt is in close relation to rotational processes. The related strike-slip faults of the western Taurids have not yet been recognised formally although many scientists have declared that rotations have played a very important role (personal communication with Dr. Robert Hall) in the structural evolution and the present configuration of the Neotethyan ophiolites. They have created dilatation on one side while causing thrusting and deformation on the other, resulting in appearance of deformed and undeformed rocks side by side throughout the western Taurids.

THE TETHYAN DISPUTE

The evidence that the Upper Jurassic conglomerates of Central Pontides contain serpentinite pebbles (Yılmaz, 1979) triggered another major dispute, a very complex one, on the evolution of the Anatolian Tethys. The oceanic crust has been dated unquestionably as Jurassic-Cretaceous in the Ankara-Ilgaz part of the İzmir-Ankara zone of Brinkmann (1972) and the adjacent Triassic deformation has been covered by an undeformed sedimentary wedge of Liassic-Lutetian age along the southern Sakarya. Şengör and Yılmaz (1981) have challenged the previous models accordingly, with a radically different plate tectonic model. They have presumed a two-stranded Tethys, the consecutive Palaeotethys and the Neotethys, with respective positions of north and south of the Cimmerian tectonites. This model has been a benchmark in the history of plate tectonic interpretations for the Anatolian segment. It complies with the following chain of reasoning made up with evaluation of today's knowledge.

1- There is an ocean along the Ankara-Ilgaz zone, the rift separating the Western and Eastern Pontides, during the Jurassic-Cretaceous, and a continous Mesozoic on the Gondwanian side in the vicinity of Kütahya (Özcan et al., 1988), implying the Palaeotethys to be located north of the Cimmerian tectonites.

2- The Palaeotethys must have had a southward polarity for compliance with the Paleotethyan / Neotethyan evolution of Pontides.

3- When the southward polarity is accepted, the Pontian magmatism, being on the passive margin, cannot be of the IA type. Thus, it has to be ascribed to crustal thickening or some other phenomenon.

4- A passive margin cannot suture with entire undeformation so that the European margin has to consist of allochtonous entities to hide the Paleotethyan suture and comply with the crustal thickening.

The kinematics of evolution, as suggested by Şengör and Yılmaz (1981), seems readily acceptable once one accepts the need or the obligation for an ocean responsible for the Cimmerian deformation. However, there has to be something wrong with this chain, which is, to the author, the negligence of the possibility of marginal ophiolite obduction onto the active European margin without a continent-to-continent collision. The model of Şengör and Yılmaz (1981 was confronted with several other objections immediately (Bergougnan and Fourquin, 1982; Robertson and Dixon, 1984) and after the work in the 1980's (Üşümezsoy, 1987).

Şengün et al., (1990) have objected by claiming that the northern branch of Neotethys is actually the Tethys and has been consumed under the Pontides through northward subduction. The marginal ophiolites have been emplaced onto the active margin in the initial stage of convergence, in Lower-Middle Triassic. The obduction is ascribed to the dextral rotation of Western Pontides during the Lower-Middle Triassic, as implied by the marked southward offset of the Eastern Sakarya. This is in analogy with what has happened in the case of the Antalya gulf. The obducted ophiolites on both sides of the Antalya gulf are covered by unfolded sediments of Miocene age, although the Eastern Mediterranean has pockets of unsubducted ocean floor south of Cyprus and Crete, implying that ophiolites can be emplaced onto active margins before completion of a continent-to-continent collision. A recessed (?) subduction, ascribed to a high rate of Tethyan convergence, occurred in the Early Upper Triassic. The consequence of the recess has been dilatation on the upwarped continental margin, when an island arc was set up in the Upper Triassic. The upwarped continental margin or the island arc has become a progressively collapsing terrain, being onlapped by the Tethys and the Black Sea.

On the other hand, most of the tectonic models dealing with the pre-Liassic events accept generally the closure of a marginal ocean to explain the Cimmerian deformation. One of these is the idea of closure of a local (Küre) marginal European basin in the Early Mesozoic (Ustaömer and Robertson, 1992). The latest proposal (Göncüoğlu, et al, 2000), with an evolutionary scenario similar to that of Şengör and Yılmaz (1981), defends a Triassic-Tertiary İzmir-Ankara Ocean geographically coincident with the İzmir-Ankara zone of Brinkmann (1972) instead of the Liassic northern Neotethys.

GEOLOGICAL AND GEOPHYSICAL CONSTRAINTS OF THE TETHYAN EVOLUTION

The dialectic analysis of the Tethyan convergence can only be complete after assesment of the existing evidence and arguments.

1- There is no crustal thickening in the Sakarya fragment.

The Bougeur anomaly maps of Turkey, prepared by the geophysical department of the Turkish Geological Survey (MTA), display contours that run very smoothly through the entire Western Pontides. A uniform continental crust, 35 km. thick, has been estimated for western Pontides. This evidence justifies the objections (Bergougnan and Fourquin, 1982) to the hypothesis of Liassic crustal thickening (Şengör et al., 1980).

2- Sakarya fragment has a Palaeozoic sedimentation unconformably capped by the south-facing Karakaya formation.

Paleozoic sedimentation in northern segments (İstanbul and Zonguldak Palaeozoic) of Western Pontides is of the shallow marine type covered by the Kocaeli Triassic comprising of continental and shallow marine sediments. The Carboniferous is only partly shallow marine in northern Western Pontides, and generally consists of coal bearing sediments. Red sandstones, possibly fluviatile, were mapped to represent the Permian of this region. Towards the south, the Permo-Carboniferous rocks are represented by shallow marine carbonates and are the gradational base of the south facing Triassic flysch. Thus, the Permo-Carboniferous of Western Pontides, or the Eurasian Karakaya formation is south-facing. There are two sections one between Daday and Azdavay (Şengün et al, 1990) and the other in the Ankara region where the Karakaya is unconformable on the Palaeozoic (Figure 3).

3- Detachment of the Anatolian fragment was not complete prior to the Liassic.

Rifting in northern Gondwana started in Late Permian/Early Triassic on the basis of stratigraphic and sedimentologic evidence. However, the Anatolian microcontinent has been suggested (Şengün, 1993) to detach off Africa not in the Early Triassic but in the Late Triassic- Liassic. Because, time is needed for crustal thinning and the related Carnian-Norian volcanism must have occurred prior to the detachment. This suggestion is to point out that there is no discrepancy between the kinematic (Westphall et al., 1986) and the structural/ stratigraphic/sequential evidence.

4- The allochtoneity of the Pontian Palaeozoic is hyphothetical.

İstanbul Palaeozoic is the term commonly used to denote a vaguely defined area essentially on the eastern side of the İstanbul strait in northwesternmost Western Pontides. The active margin of the Tethys in Thrace is the Strandjha massif, the domed westward continuity of the İstanbul Palaeozoic (Cağlayan and Yurtsever, 1999). The suggestion of the allochteneity of the 'İstanbul Palaeozoic' is a means of elimination of the contradiction that this domain is of European origin but located south of the Palaeotethyan suture of Şengör and Yılmaz (1981). The suggested thrust (Şengör et al., 1980, 1984) and strike-slip faults, which revise the former theories of thrusting (Okay et al., 1994), have been neither substantiated nor there has been a hint pointing to a locality where the thrusting or the strike-slip faulting has been observed. The only location that may be considered as the base of the proposed nappes comprising the Palaeozoic rocks of the northern Western Pontides is the Devrekani charriage, which is 400 and 600 km respectively to the İstanbul Palaeozoic and the Strandjha. Furthermore, the age of this charriage is not Liassic as suggested by Yılmaz (1979) and Şengör et al., (1980), but is post Upper Cretaceous as it captures globotruncana bearing sediments on its front (Şengün et al., 1990). This overthrust comprises of the Precambrian-Mesozoic sequence as the overriding block, capturing on its front, many ophiolite outcrops with unconformable Cretaceous sediments. It continues not only to the Ankara region southwards, but is also continuous in the north causing formation of the ridge (Zonenshain and Le Pichon, 1986) that separates the Eastern and the Western Black Sea.

5- Northern branch of Neotethys has been annulled.

Northern strand of Neotethys of Şengör and Yılmaz (1981) has never existed. Instead, there existed an ocean, the İzmir-Ankara zone of Brinkmann (1972) whose suture is of the Tethys from Ilgaz to the city of Bursa where it splits to İzmir as the Western Neotethys (Figure 1) and continues to the Vardar zone via the Sea of Marmara. The Tethyan suture connnects to the Sevan Akerra through the Eastern Pontides-Eastern Anatolia boundary. The following are some of the reasons for annulment of the northern Neotethys.

There is a continuous Lower Triassic-Cretaceous fining upward sequence, Kütahya, south of the northern strand of Neotethys (Özcan et al., 1988). This evidence shows on its own that this suture belongs to an ocean that existed at least for the entire Mesozoic (Şengün, 1990; Göncüoğlu, et al., 1994), in fact, to the Palaeozoic- Mesozoic Tethys on the basis of north facing Gondwanian and south facing Eurasian Palaeozoic platforms.

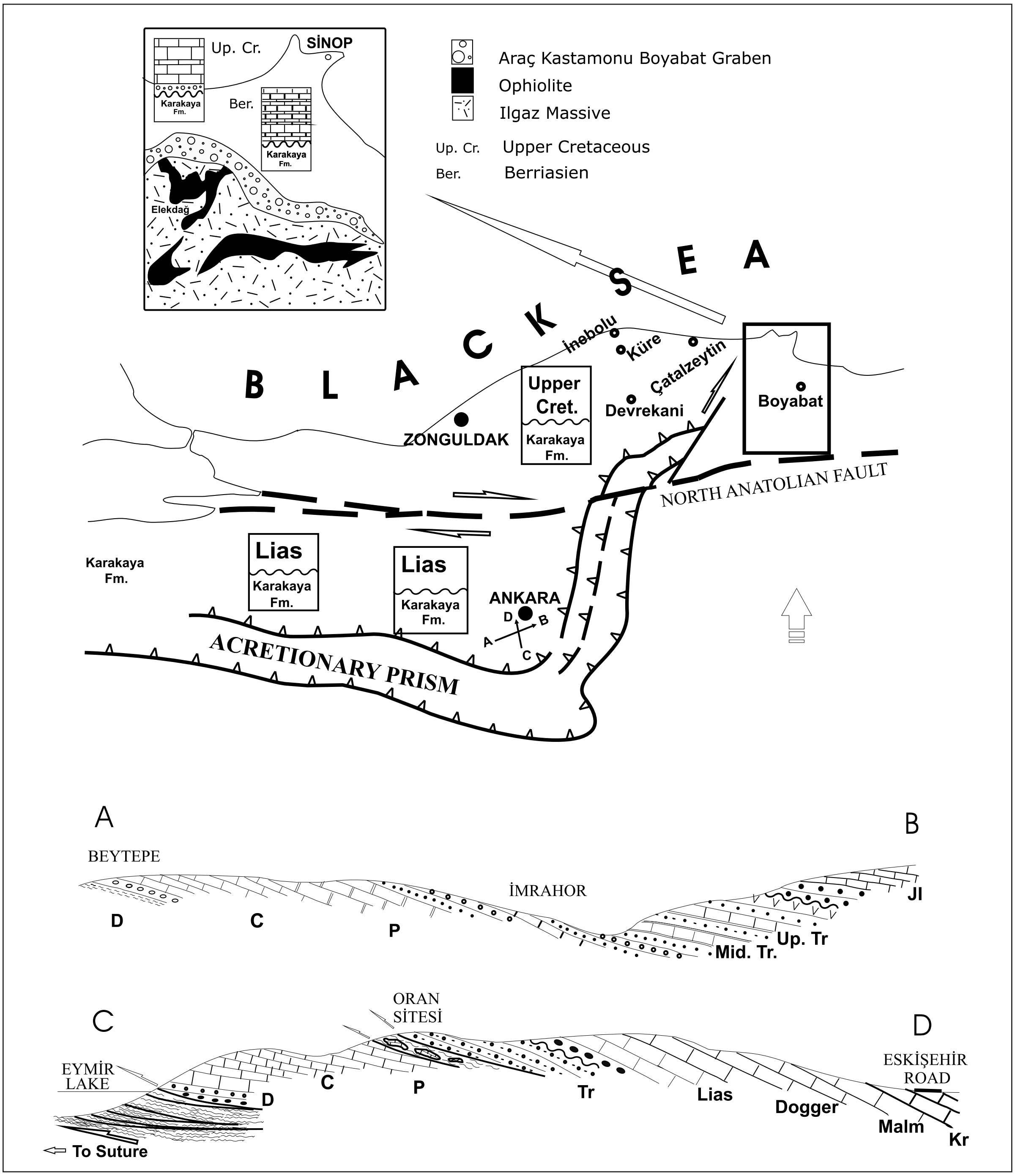

Figure 3- Figure to show various features of the eastern-Central Pontides. Cross-sections AB and CD illustrate the routes along which the type sections of the Karakaya formations ,can be seen. The deformation augments towards the south where a type of melange is encountered representing the shallow segments of shear zones. The type sections are oversimplified to emphasise the fact that rocks and stratigraphic relations are well preserved because of being carried on the back of the uppermost slice.

There is extensive geochemical research, all defending the Triassic-Jurassic (?) magmatism of the Pontides to be of the island arc type (Boztuğ et al., 1985; Kazmin et al., 1986 and Tokel, 1992).

On the northern side of the suture, namely in the Sakarya fragment, there is a post-tectonic sedimentation unconformable on the Cimmerian tectonites. It generally starts in Liassic as a fining upward sequence (Şengün, 1992). The sequence is continuous from the Liassic to the uppermost Cretaceous in the Eskişehir (Bingöl and Neugebauer, 1992) and Ankara regions. On the other hand, a post-tectonic wedge in the range of Portlandian-Lutetian is well established in the Central Pontides. North-facing character of the latter is readily recognised on the Boyabat-Sinop and Devrekani-Çatalzeytin roads. The Upper Jurassic-Lutetian sections are almost identical with their counterparts in the sections starting with the Liassic. This sedimentary wedge is very probably related to the Tethyan onlap in the south while it is definitely a Black Sea sequence north of the positive area. The evidence is indicative of a transgression diminishing the positive area centering the Sakarya fragment. There have been small islands that have not been transgressed until the end of Eocene as indicated by local columnar sections.

The Liassic-Lutetian sediments cover unconformably the Palaeozoic rocks and the Karakaya formation, the latter being restricted to the southern belt juxtaposed to the Tethyan (or the pseudo-Neotethyan) suture. The Karakaya formation, Carboniferous-Triassic in age, bears a genetic relation to this suture, which cannot be of the Liassic Neotethys, but the Carboniferous-Triassic Tethys. The regressive Triassic of the northern Tethyan margin versus the fining upward Mesozoic sequence of the southern is not compatible with a Liassic extensional basin, but a Mesozoic active margin in the north and a passive one in the south.

6- The Mesozoic magmatism of the Western Pontides is of the island arc type and is Upper Triassic in age.

The Upper Triassic basic magmatism of Central Pontides have been differentiated to yield a granitic magma, followed by high silica differentiates. The basic magmatism is displayed very neatly in Central Pontides hosting the granitic. The granitic bodies display extremely wide aphanitic perypheries, implying that they have been very shallow seated and the country rock has been very cold. There are many granite batholithes (comprising basic rocks as the host) in Central Pontides dissecting mutually ophiolites, the Palaeozoic sediments and the Karakaya formation. Boztuğ et al (1985) state that the basic and the granitic magmas of this region belong to the same magmatic suit. The author, agreeing with this statement, suggests furthermore that the granitic magma is the differential product of the basic. The suggestion is based on the basic rocks being the host for the granitic and the striking resemblance of the pleochroism of the hornblend phenocrystals of the diorites and those of the granitic rocks, crucially implying the continuity of crystallisation of hornblend of a specific composition. Further petrochemical investigations should, hopefully, check up the suggestion. This proposal has the crucial implication that the basic and granitic magmatism are of the same age.

On the other hand, the radiometric dating of 165±3 my for the granite (Yılmaz, 1979) cannot be proven wrong for the Central Pontides on the basis of direct evidence. Nevertheless, no granite dyke has been reported to dissect any section of Liassic age. Secondly, the sedimentary wedge covering the Karakaya complex is Upper Jurassic-Lutetian in age in the northern Central Pontides, with full stratigraphic correlation of the section with those of the Ankara and Eskişehir regions (Saner, 1980; Bingöl and Neugebauer, 1992). In other words, the Liassic-Cretaceous sedimentation of these regions are tied to the

Central Pontides, showing very clearly that this wedge covers the magmatism and the Triassic deformation. In conclusion, the granitic magmatism has to be Upper Triassic in age to be compatible with the stratigraphic constraints, and is not post-collisional but of the island arc type in Central Pontides on the basis of extensive geochemical research.

7- The Karakaya enigma

There has been an enigma on the Karakaya formation caused by the similarities between the Triassic sedimentation on Gondwanian and Eurasian margins. The nomenclation (Bingöl, 1968) has been applied to both (Sakarya and Biga).

The Gondwanian Karakaya is Mesozoic in age while the Eurasian is Carboniferous-Late Triassic.

The Gondwanian Triassic (Antalya nappes) grades to deep marine Jurassic-Cretaceous sediments while the Eurasian ends up in the Upper Triassic, being invariably regressive in the Upper Triassic section.

The Gondwanian Permian blocks are covered and underlain by continuous layers that converge on both ends. The strata are turbiditic with blocks of various dimensions. There is no crystallization in Permian limestones. The fauna has Gondwanian affinities. On the other hand, the Permian blocks north of the Tethyan suture are of the broken type and partly crystallised. Block formation of the Eurasian Permo-Carboniferous in the Triassic sandstones is encountered essentially in the shear zones.

The Triassic of the Eurasian Karakaya is restricted to a belt adjacent to north of the Tethyan suture and has never been deposited outside this narrow belt. Thus, it is genetically related to the southward-located Triassic Tethys. On the other hand, the deposition of the Gondwanian sequence is scattered on northern Gondwana and is related to dilatation of the Anatolian microcontinent.

There is a type section/area between Beytepe and İmrahor in the vicinity of Ankara. The northward younging section north of the deformed suture is carried on the back of the uppermost slice (Figure 3, sections AB and CD). It comprises of:

1- A lower Palaeozoic clayey unit exposed in the vicinity of Eymir Lake.

2- Carboniferous-Permian limestones with a conglomeratic base.

3- High-energy clastics of Lower Triassic age gradational to Permian limestones.

4- A continous carbonate sedimentation of Jurassic-Cretaceous age covering these unconformably.

5- Serpentinites juxtaposed to Liassic- Cretaceous carbonates on the Eskişehir road.

Block formation in the Eurasian margin occurs, to the author, mostly in shallow segments of Upper Cretaceous-Paleocene thrusts, which have dragged the inter/back-arc regions towards the Tethyan suture. The deformation diminishes going away from the suture so that undisrupted sequences, with extremely rich faunas, can be seen as in the case of Permo-Carboniferous rocks in the vicinity of Ankara. The Cretaceous deformation appears as rhythmic and southward narrowing shear planes. The deformations are restricted to these shear planes so that the lithons, when fairly distant to the suture, can display the previous, i.e. the Triassic deformations. In other words, the belt with intense Cretaceous deformation grades northward, to the zone with pre-Liassic deformations showing rhythmic and progressive northward diminishing of its Cretaceous imprint.

The Karakaya formation is a marker of the European margin in the Sakarya where it is unconformable on Palaeozoic sequences of European origin. It is of Gondwanian origin in the Biga-Karaburun as will be discussed in the following item.

8- The Sakarya fragment is of European origin while the Biga is of Gondwanian.

Which continental fragment (Figure 1) belongs to which major continent? Is there satisfactory evidence for the origins of the Sakarya and the Biga? These are probably the most significant and appropriate questions that should be raised for the Anatolian segment of the Tethyan belt. There is no direct claim based on paleontologic evidence that the Biga is European but implied to be hypothetically (Okay and Tüysüz, 1999). On the other hand, European origin of Sakarya (Şengün et al., 1990) receives acceptance in the recent years. The present author claims on the basis of sequential evidence that the Sakarya is European while the Biga is of Gondwanian origin.

There is a continuous Triassic-Cretaceous sequence in the Biga peninsula (Koçyiğit and Altıner, 1990) and there are many Mesozoic fining-upward sequences along the Bursa-İzmir zone (Akdeniz, 1985). The D'orsay group (Brunn et al., 1975; Ricou et al., 1974-1986) has defined these as the Antalya nappes, the very typical Gondwanian facies. The Triassic sequence of the Biga, or the Karakaya formation of Bingöl (1968) has also been defined as the Antalya nappes. The Antalya nappes of the Karaburun peninsula, which has been declared as Gondwanian and juxtaposed to the Bursa-İzmir zone (Erdoğan, 1990), are certainly on the continuation of the Biga Peninsula, the Aegean and the Greece. Bursa-İzmir zone comprises other Lower Triassic-Upper Cretaceous sequences (Akdeniz, 1985), substantiating the designation of this zone as intra-Gondwanian.

The Karakaya formation sits, with a sedimentary contact (Şengün et al., 1990), on the Palaeozoic rocks of Eurasian origin in the Daday-Azdavay section of the central Pontides. The Carboniferous of the Beytepe- İmrahor section of the Ankara region sits on the underlying Devonian shales also with a sedimentary contact. Anything European in any part of these sections implies the European character of the whole. The European character of the Sakarya fragment (Şengün et al 1990) is backed up not only sequentially but also paleontologically. The Liassic sedimentation of the Sakarya contains European ammonites (Alkaya, 1990) and the Eurasian origin of the Zonguldak Carboniferous is well established (Kerey, 1982, Toprak, 1984). It has been asserted (personal communication with Prof. E. Ya. Leven of Moscow University) that the Permo-Carboniferous carbonates of the Beytepe (Ankara) section (Figure 3) have a very rich Eurasian fauna.

Nevertheless, further investigations will hopefully complement the sequential evidence and the paleontologic assertion cited, to show that the Sakarya belongs to Europe while the Biga peninsula to Gondwana. A consensus on the origin of the Biga will resolve many of the entailing disputes such as the Intra-Pontide Ocean. Further paleontologic and paleomagnetic works on the Permian limestones of Biga may yield evidence that will hopefully resolve the ongoing debate.

9- The Intra-Pontide Ocean is hypothetical.

Okay et al (1994), Okay and Tüysüz (1999) and Göncüoglu et al. (2000) defend an E-W trending intra-Pontide ocean separating the Sakarya and northern Pontides. The author's objections to existence of this ocean are outlined below.

The sediments, younger than the Liassic, are undeformed on both sides of the suggested Intra-Pontide suture, which is mostly coincident with the NAF.

The belt of Upper Cretaceous volcanics trends perpendicular to the suggested strike-slip faults and shows no offsets in Central Pontides.

An unbroken Pontide belt is unacceptable (Lauer et al., 1981). The abrupt end up of the Intra-Pontide ocean in the middle of the Pontides is not only subject to objection as a concept but also brings in the necessity to explain the eastward ophiolite thrusts onto the Kirsehir massif, which are connected to ophiolites of the Daday.

The Portlandian-Lutetian sedimentary wedge covers the paleo-lineament that causes the offset of the Daday-Devrekani massif with respect to the Ilgaz in northern Central Pontides. On the other hand, the Liassic-Lutetian wedge has started to transgress the northern Pontides in Portlandian as a result of the progressive collapse of the Western Pontides. Therefore, the lineament, by which the Western Pontides has been displaced southwards relative to the Eastern Pontides, must have died out by Portlandian. This interpretation is backed up by the hardly detectable offset between the Elekdag (Figure.3) and the westward continuation (Karadere ophiolites) separated by the Arac-Boyabat graben.

Yılmaz et al (1994) have suggested an Upper Cretaceous closure of the Intra-Pontide Ocean versus the Paleogene collision of Okay and Tuysuz (1999). However, the lack of deformation on the Mesozoic rocks on both sides of the suture seems to be an important drawback for either of these theories. The active margin may be undeformed because of being carried on the back of thrust sheets. However, non-deformation along a passive margin is unacceptable

THE SUTURES OF ANATOLIA

THE NEOTETHYAN SUTURE

The northern branch of the Neotethys (Şengör and Yılmaz, 1981) seems to be annulled (Şengün et al, 1990; Okay et al, 1994), leaving one Neotethyan suture in Anatolia as shown in Figure 1, the southern and the western zones being connected (?) through the Eastern Mediterranean. The western branch possibly connects to the northern Antalya basin of Robertson (1990) and the loop north of Karaburun towards the Vardar zone is theoretical. The loop is theoretically undeletable to maintain the distinction between the Tethyan and İntra-Pontian sutures. In other words, elimination of this loop or acceptance of the Biga as Gondwanian would mean coincidence of the Tethyan and the Intra-Pontian sutures. However, the suture cannot pass through north, in disagreement with Okay and Tüysüz (1999), of the broadly folded Mesozoic sediments of the Gondwanian Karaburun (Erdoğan, 1990), but passes through the southwest of this peninsula (Gökten et al., 2001) where the ophiolites have been ultramylonitised due to imbrication with the continental rocks of the Menderes massif (personal observation with Dr. N. Konak and. Mr. A. Cağlayan). This means that this suture is in between the Karaburun and the Menderes, both of which being of an undoubted Gondwanian origin. The Menderes massif, the domed continuation of the Taurids, has a Palaeozoic-Mesozoic cover (Çaglayan et al, 1980) in the southern segment of the Izmir-Ankara zone. The sections exclusively display the sequential and palaeontological characteristics of the Tauric facies. This evidence backs up the connection of this zone to the Antalya basin. The suture and the Lycien nappes, the latter being presumably related to the compressive field generated in relation to closure of this zone, have probably been subject to post-collisional configureuration by strike-slip faults.

The southern Neotethyan suture passes through the immediate north of the Pütürge (Yazgan, 1984) and the Bitlis massifs, and not through the zone known as the Bitlis suture (Hall, 1976) or the southern branch of Neotethys of Şengör and Yılmaz (1981), which is referred to as the Maden-Cungus foredeep in this text. This assertion is based on detailed regional geological mapping of eastern Bitlis (Caglayan et al., 1984) and the Puturge massifs (Yazgan, 1984).

The following information is a summary of the evidence that show Bitlis to be in situ and the uplifted passive margin of the southern Neotethys.

The Mesozoic sequence of the southeastern Bitlis (Cağlayan et al., 1983), exactly the same as that of the border folds, not only shows that Bitlis and the border folds were on the same north-facing Mesozoic platform of Gondwana but also proves that Bitlis is in situ. It also implies that there has been no rifting along the Bitlis suture of Hall (1976) until the end of the Mesozoic.

Southern branch of Neotethys or the Bitlis suture of Hall (1976) is juxtaposed on both sides by undeformed rocks. The Bitlis/Puturge block is represented by a Precambrian basement capped unconformably by incipiently deformed Palaeozoic-Mesozoic (Yılmaz, 1971) and there is no quest for the undeformed nature of the border folds (Şengün, 1990). Can the northern block have a sheared Precambrian basement so that a subduction zone is sealed? No, because there is nowhere any sign of such a deformation. Is a Triassic rifting possible south of Bitlis? There is no sign of changing sedimentologic parameters neither in Permian limestones, nor in the Triassic sedimentation, which is exactly the same as those of the pseudo-suture and the border folds. This means that there has been a Permo-Triassic platform extending from northern Bitlis/ Pütürge to the Arabian platform.

The intensity of alpine deformation, undressed of folding, diminishes towards the south, showing that the suture lies in the north.

Bitlis massif is the uplifted passive margin of the Neotethys on the basis of the island-arc setting in its immediate north (Yazgan and Chessex, 1991) and the southward obducted Gevas ophiolite (Cağlayan et al., 1984).

The overall geophysical evidence shows that there has been a collage between the East Anatolia and the Bitlis/Puturge in the Upper Cretaceous. The ophiolites have been obducted onto the passive margin (Bitlis/Puturge) from the Cenomanian onwards (Yazgan, 1984; Yazgan and Chessex, 1991). There is a consensus that the obducted ophiolites imbricated with the crustal rocks of Bitlis glide gravitationally into this foredeep. This is a crucial support to the dilatational regime implied by the sedimentation between the Upper Cretaceous and Miocene.

The closure of the pockets of unsubducted ocean floor has resulted in formation of foredeeps south of Bitlis and Puturge. This process is reflected in the sedimentation and the magmatic activity. The initiation of the extensional Maden-Cungus trough is indicated by convertion of carbonate sedimentation of the southeastern Bitlis to gradational high-energy clastics of Campanian- Maastrichtian age. The dilatation and sedimentation is continous including the Miocene along this trough.

The rotations generate dilatation causing partial melting in the uppermantle-lower crust. A syncollisional magmatism, the Maden formation, forms and is expelled through diabase dikes along NNE trending paleo-transtensional faults that spread the lavas as flows intercalated in coeval sedimentation in the EW trending troughs extending along both extremities of the Bitlis massif.

The thrusting of Bitlis onto this graben took place in the Late Miocene after the collage between the East Anatolia/ Pontides and completion of the Neotethyan closure, with the aid of the NNE push of the Arabian platform (McKenzie, 1972). Detailed mapping has shown that the Bitlis head-thrust of Altınlı (1963), which borders the massif in the south, is not a single plane but comprises of many unlinked thrusts (Ozkaya, 1982). It is possible to cross from the Eastern Bitlis to the Dodan anticline of the border folds without any thrusting. Thus, the division of the northern Arabian platform into tectonic slices of Bitlis, orogenic flysch and the border folds is a false and unsubstantiated hypothesis, but is currently accepted.

Yazgan (1984) has described the magmatic and stratigraphic relations to show that the Neotethys also lied not in the south but the immediate north of the Pütürge massif.

West of the Puturge massif, the suture has a southward displacement of 105 kilometers (Freund et al., 1970) by the Dead Sea transform so that it passes through the north of well-known pockets of unsubducted ocean floor south of Cyprus and Crete in the Eastern Mediterranean. There has been Neotethyan acretions to the Taurids as in the case of Antalya Complex (Robertson and Woodcock, 1981; Yılmaz, 1984; Poisson 1984 and Özgül, (1984). It is possible that the Northern Antalya basin may be tied to the Ecemis lineament. However, this issue is left out because of the author's inadequate knowledge of this region.

The reason for the misinterpretation (Şengün, 1993) about the Bursa-İzmir segment of the İzmir-Ankara zone of Brinkmann (1972) was the acceptance of the Biga as Gondwanian followed by the consequential false reasoning that this zone could have been a foredeep only. Many have considered ophiolites of the Menderes massif as transported tectonites originating from the north, although the imbrication with the Menderes massif is suggestive of a suture. The SSW trend of the İzmir-Bursa zone also backs up a suture. The author has realized on the basis of the preceding thoughts that this zone is not a Tethyan but an Intra-Gondwanian Neotethyan rift separating the Anatolian microcontinent from Apulia-Greece. It seems that this rift has to connect to the Northern Antalya Basin of Robertson (1990) via the west of the Lycien nappes of southwest Anatolia.

THE TETHYAN SUTURE

A Cretaceous-Paleogene (?) Tethyan suture between Bursa and the coastal areas of the Sea of Marmara substantiates itself by the ophiolite slivers that show the same high pressure/low temperature metamorphism and the same deformational geometry as the eastern Sakarya. The Tethyan suture is theoretically connected to the Vardar zone through the Sea of Marmara also on the basis of the Biga being Gondwanian and the Sakarya of European origin. The Tethyan suture coincides with the northern strand of the Neotethys of Şengör and Yılmaz (1981) between Bursa and Ankara. It is also well marked in the segment bounding the Eastern Pontides and the East Anatolia and is characterised by imbricate southward thrusting (Yılmaz, 1985). The Miocene blocks in emplaced ophiolites juxtaposed to the Tethyan suture shows that the suturing of the Tethys in East Anatolia has not been completed until the Late Miocene.

A wide belt of ophiolites thrusted onto the Kirsehir massif marks the Tethyan (+Neotethyan) suture between Ankara and Ilgaz. This ophiolitic slice is connected to that of the Daday massif via the Arac - Boyabat graben. The suturing has caused uplift of the passive margin, the Kirsehir massif with an intense multistage Alpine deformation with southeastward diminishing grade of metamorphism (Erkan, 1975; Seymen, 1982 and Tolluoğlu, 1987).

A DISCUSSION ON POSSIBLE SCENARIOS OF PRE-LIASSIC GEOLOGIC EVOLUTION

The post-Liassic evolution of Dercourt et. al. (1986) is mostly agreeable except the evolution of the Eastern Mediterranean. Most of the tectonic models accept that closure of a marginal basin is unavoidable to explain the Triassic deformation. The possibilities and selected past hypotheses on the issue will be examined briefly.

I. The Paleotethys dives Southward under Gondwana comprising Southern Sakarya and

the Biga. Northern Neotethys is a Marginal (Back-Arc) Gondwanian Basin (Şengör and Yılmaz, 1981).

The key point has been that there must have been a collage responsible for the Cimmerian orogen in the Sakarya fragment, as there is a Jurassic-Cretaceous ocean floor south of the Cimmerian tectonites in the Sakarya fragment. However, the following geological and geophysical constraints are not consistent with the hyphothesis of Şengör and Yılmaz (1981).

1- The gap between the Taurids and the Pontides is more than 4 000 kms in the Liassic (Westphall, et al, 1986). Thus, a collision by M. Jurassic is almost impossible.

2- The geophysical evidence is suggestive for a uniform crust with a thickness of about 35 kms in Pontides, thinning smoothly towards the Black Sea, justifying the objection of Bergougnan and Fourquin (1982) to the theory of crustal thickening.

3- The Liassic-Dogger piling of the Palaeozoic rocks of the northern Pontides is false also because of the post-Upper Cretaceous age of thrust faults. These have a widening spacing going away from the Tethyan sutuıre. They capture many bodies of serpentinites with Cretaceous caps in the Sakarya fragment and are certainly related to the Tethyan suturing.

4- There is concrete evidence for non-existence of the northern strand of Neotethys of Şengör and Yılmaz (1981).

5- The ascription of the Upper Triassic (Jurassic?) magmatism to crustal thickening is exclusively denied by extensive geochemical research (Boztuğ et al., 1985, Kazmin et al., 1986 and Tokel, 1992), all defending the IA type.

Göncüoğlu et al. (2000) have proposed a similar scenario representing the latest of the defenders of the possibility. The pre-Liassic evolution is the same as that of Şengör and Yılmaz (1981) except that the İzmir-Ankara ocean or the northern strand of Neotethys has opened in the Lower Triassic. The following criticism is additional to the ones enumerated above.

1- The Mesozoic dilatation of the southern margin of the İzmir-Ankara ocean is not compatible with the coeval compressive state of the northern margin in the Triassic. Known constraints back up a passive margin in the south and an active one in the north.

2- The theory has to explain how the Mesozoic fining upward sequences of the Bursa-İzmir zone is replaced on the eastern continuation, the Sakarya, by a weakly deformed Carboniferous-Triassic overlain by the undeformed Liassic-Lutetian sedimentary wedge. Can these sequences belong to the same continental margin? The crucial constraint is the Gondwanian and Eurasian origins respectively for the Biga and Sakarya. Both of these fragments have to be of the same origin for consistency of this model. Both have to be of either Gondwanian origin or Eurasian, as the marginal Karakaya Ocean could have not possibly extended in both of the major mainlands. Therefore this model has to defend the Gondwanian character of the Biga and southern Sakarya for consistency. However, the European character of the Sakarya is well established.

3- The İzmir-Ankara ocean, separating the Menderes-Taurid block from the Sakarya, cannot open in the Triassic, because this means that the Gondwana and Eurasia had formed a single continental mass in the Triassic, unless a very thin continental fragment of Gondwanian origin is separated from the European by an ocean at least 4000 km wide.

II. There has been a Southward diving the European Marginal Basin Responsible for the Triassic deformation (Ustaömer and Robertson, 1992).

The objections to this possibility is summarised as follows.

1. There is no direct evidence showing that a Eurasian sliver has collided with Europe by the Liassic.

2. A marginal European basin can be possible only if it is in the range of Carboniferous-Upper Triassic, coeval with the Karakaya formation, because the Karakaya formation is unconformable on the European Palaeozoic sequences (Şengün et al., 1990). This implies that a southward polarity (Ustaömer and Robertson, 1992) is unavoidable with the chain consequence that the granites of Central Pontides, being located north of any European marginal basin, has to be ascribed to crustal thickening. The theories with a southward polarity are not compatible not only with the incipient deformation of the Palaeozoic sequences underlying the Karakaya formation of northern Pontides, but are in contradiction with the extensive geochemical research (Boztug et al., 1985; Kazmin et al., 1986 and Tokel, 1992), exclusively suggesting an island arc origin for the granitic magmatism.

3. Another objection to a southward diving European basin would be that the Cimmerian deformation intensifies southward showing the cause of deformation is in the south of the accretionary prism.

4. A gradation exists through northward widening tectonic slices in the Western Pontides (Sakarya).

5. The Karakaya formation of the Sakarya fragment is European and is juxtaposed continuously to the northern (the Tethyan) suture, implying that such a marginal basin has to be coincident with the Tethys.

6. Along the Izmir-Bursa zone, there is a continuous fining-upward Mesozoic sequence, which has been defended as the Gondwanian Karakaya, also known as the Antalya nappes. On the other hand, the Sakarya fragment is characterised by the European Karakaya formation of Carboniferous-Triassic age. Thus, these segments are genetically different and are not parts of a single continuous margin.

III. The European (Intra-Pontian) Marginal Basin dives Northward.

A European marginal basin (in a model analogous to that of Adamia et al. (1977), noting that the marginal basin in the Caucasus corresponds to the Black Sea) diving north may be considered possible, only if the Carboniferous-Upper Triassic sedimentation unconformable on European rocks is false or ignored. In that case, the Kure ophiolite needs to have been emplaced on the active margin by rotationary processes of the continental margin so that it can be intruded by the island arc magmatism. The mechanism would not be significantly different from that defended in this paper for both the Eastern Mediterranean and the Tethys. Existence of finite pieces of continental crust that have been deformed into an unrecognisable state by the Triassic/Cretaceous events cannot be proven wrong. However, such a theory has to place the marginal basin in the middle of the accretionary prism, leaving a remarkably narrow strip of southern margin, which has left no fingerprints behind.

IV. There is no Marginal Intra-Pontian Basin.

The Triassic deformation is caused by the marginal ophiolite obduction onto the active European margin without a continent-to-continent collision. This is the possibility preferred, so it will be described and defended below.

GEOLOGIC EVOLUTION

PRECAMBRIAN

The Precambrian basement is exposed in several locations in Turkey. Alanya, Menderes, Kırşehir, Bitlis and Puturge are well-known Gondwanian massifs with Precambrian basements. There are several other exposures in northern Pontides underlying unconformably uniform and incipiently deformed Palaeozoic sequences. The Precambrian basements comprise essentially of amphibolites and micaschists/paragneisses metamorphosed in physical conditions of almandine-amphibolite facies. Granites have intruded these and have been deformed by the Alpine uplift only. Palaeozoic- Mesozoic sediments cover these unconformably in the Central Taurids (Özgül, 1976-1984). The westward narrowing gulf of the Pangea (the Tethys) seems to have initiated with the beginning of the Palaeozoic Era (Figure 4).

PALEOZOIC

There has been immense stratigraphic research since the late 1960's in Central and Western Taurids. Detailed mapping in the following years resulted in new disputes about the structure. However, there is almost full agreement on the Palaeozoic stratigraphy of the Taurids. The continental (coal bearing) to shallow marine Permo-Carboniferous sequence of the Alanya massif and other Lower Palaeozoic sediments of Western Taurids grade to continuous Palaeozoic marine sedimentation northwards, implying a north-facing Tauric platform (Blumenthal, 1951; Özgül, 1984; Demirtaşlı, 1984).

The stratigraphy and sedimentologic parameters of the Palaeozoic fining upward sedimentation in northern Pontides are suggestive of dilatation during the Lower Palaeozoic, converted to compression during the Permo-Carboniferous. There is a continuous Palaeozoic sequence in the Karadere region west of Daday. If a south-facing morphology is acceptable for the Karakaya formation, the continuous Palaeozoic implies that there has been an ocean south of these sequences during the entire Palaeozoic era. The Carboniferous and older rocks of the northern Pontides were subject to southward onlap of the Black Sea, the Central Pontides remaining as a positive area until the Late Jurassic. It was not transgressed completely until the Late Cretaceous or even Eocene.

The stratigraphic, sedimentologic and morphologic evidence for the formation of the Tethys can be complemented with the Atlantic Ocean data for the conclusion that the Tethys had a width of roughly 5000 kms in the vicinity of Anatolia at the end of Permian as suggested by Westphall et. al., (1986).

MESOZOIC-PALEOGENE

TRIASSIC

There have been marked facies changes in the Taurids with initiation of rifting in northern Gondwana. Permian limestone deposition has been converted, along the Neotethyan margins, to high-energy deposition represented essentially by turbiditic sequences with Permian limestone olistolithes. The crustal attenuation initiated basic eruptions during the Carnian-Norian. The cessation of this volcanism presumably marks the onset of ocean-floor spreading in Neotethyan rifts. Facies changes between the rifting margins and the positive areas result in juxtaposition of deep and shallow marine environments.

The replacement of carbonate deposition by high-energy clastics in southern Pontides is suggestive of a Triassic onset of northward subduction of the Tethys. However, the author questions the deepening environment even for the beginning of the Triassic period with the consequent quest on the indispensability of a subduction in the Early Triassic. The Triassic sedimentation of the Karakaya formation is represented by high-energy clastics with rare interbeds of limestones yielding fossils of Lower, Middle and Upper Triassic age. The section in the Ankara region (Figure 3) represents the proximal part of the Eurasian continental slope. The Lower Triassic segment comprises blocks of Permian limestones. The Upper Cretaceous thrusting as

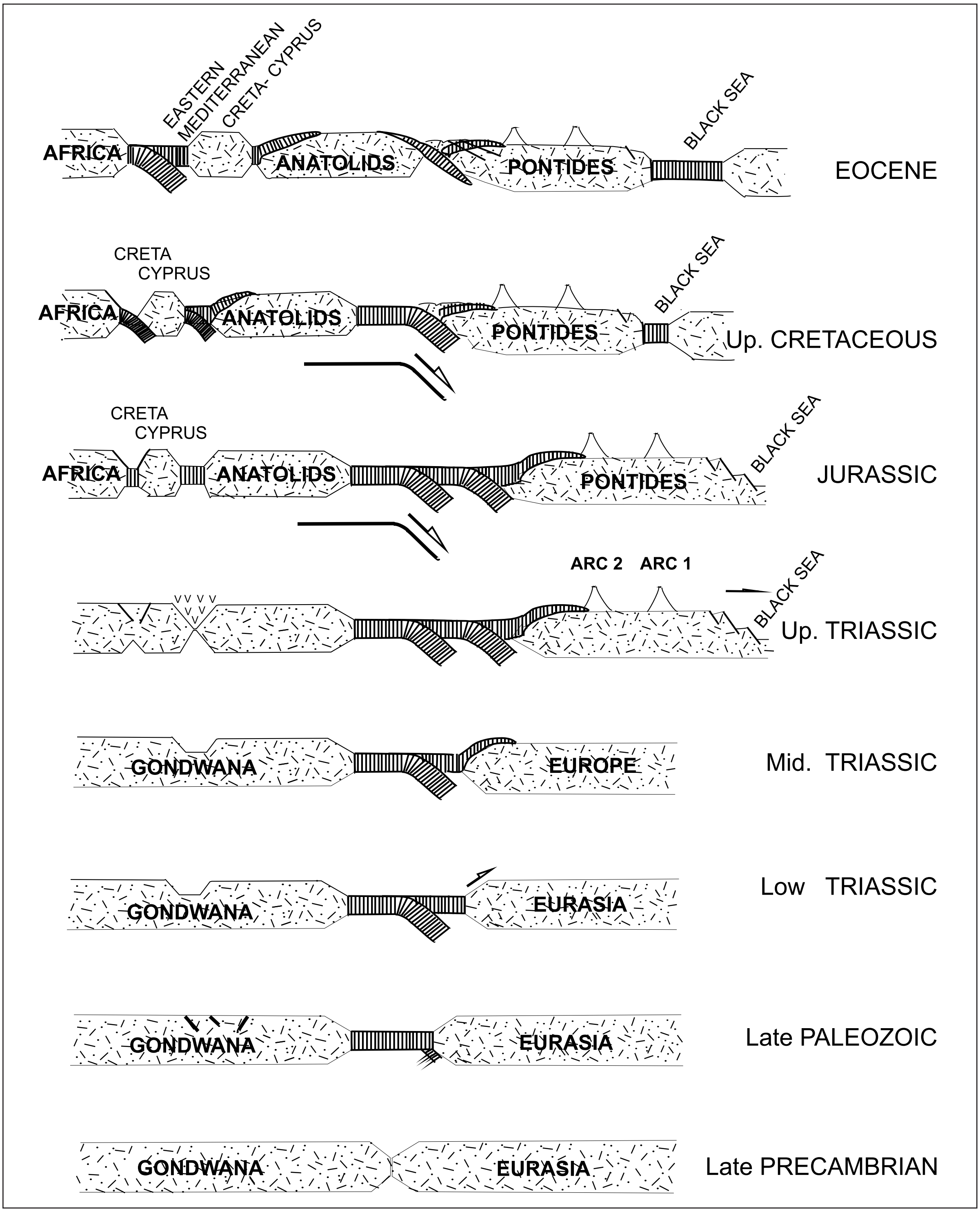

Figure 4- A tentative and unscaled chain of evolutionary cross sections from the Eastern Mediterranean to the Black Sea.

described in the preceeding sections has broken up the Permian limestones resulting in blocks in a Lower Triassic matrix in the Permian-Triassic boundaries. The Triassic sequence of the Kure region sits on the abyssal clays covering the sheeted dykes and is the representative of the marginal ocean floor. It continues as a regressive sequence comprising carbonate interbeds that yield Lower, Middle and Upper Triassic fossils. There are small serpentinite wedges in the medial part of the section with sheared bottoms and sedimentary tops, implying that the emplacement is coeval with the deposition.

JURASSIC

The assesment of the kinematic evidence (Westphall et al., 1986) implies that the western and central Taurids seem to be moving with Africa until the Early Jurassic. Therefore, the suggestion of a Liassic age of detachment is plausible and is compatible with the geophysical constraints (Figure 5).

The Gondwanian sedimentation comprises essentially of deep marine carbonates in subsiding troughs such as Izmir-Bursa zone, Kutahya trough and Karaburun. There are sharp facies changes perpendicular to the axes of these subsiding troughs with gradations to relatively shallower environments and onlaps onto the positive areas. The sedimentologic record is indicative of continuous dilatation during the entire Mesozoic in the Anatolian microcontinent. Rifting has occurred in the Eastern Mediterranean south and north of Cyprus/Crete, the northern strand possibly extending as the Northern Antalya Basin connecting to the İzmir-Bursa zone and the Ecemis (?) lineament in the east.

The Liassic initiation of deposition on the Pontian arc seems to have occurred in areas of earlier collapse along the southern coast of the island arc. The northern part of the Western Pontide block has persisted as a positive area between the Late Carboniferous and the Upper Cretaceous. The Central Pontides have been

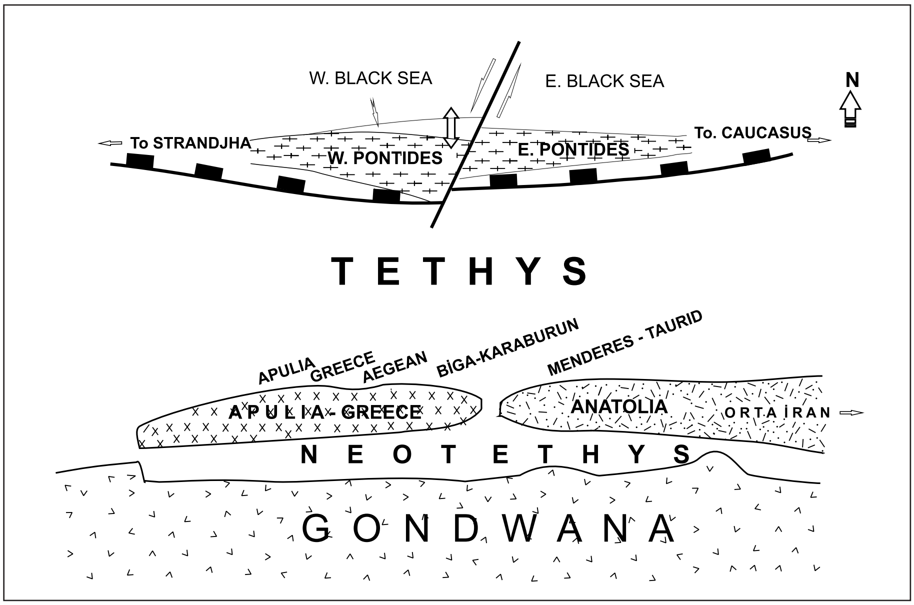

Figure 5- A sketch map to show the relative positions of the main continental and oceanic domains in the Upper Triassic - Liassic.

transgressed in Portlandian-Berriasien in liasion with the progressive collapse. The time of onlap is Albian in the Daday region while it is Upper Cretaceous in areas that are fairly distant from the suture, such as the İstanbul and Zonguldak Palaeozoic. The times of onlap on two sides of the same hill are Berriasien and Campanian in the Elekdag region where the basal columnar section is the same, displaying the same rock sequence. It must be emphasised again, that the Upper Cretaceous transgressions are not related to initiation of rifting of the Western Black Sea, but to onlap of the existing. Thus, the Western Black Sea must have started to rift before the Jurassic, very probably in the Early Triassic to be in consistency with the hypothetical dextral rotation of Western Pontides.

The Liassic deposition from Ankara region to Bursa, in Western Pontides, is suggestive of the subduction to trend parallel to the Tethyan suture. It is not possible to say that there has been coeval subduction in the Ankara-Ilgaz zone during the Liassic, and ocean floor spreading could have been dominant, particularly in the southern segment of this zone, conformably with the possible dextral rotation of Western Pontides.

CRETACEOUS-NEOGENE

The Cretaceous was a period of rapid drifting of the Anatolian microcontinent towards the Pontides. The deposition has continued in the extensional basins of the Anatolian microcontinent and the back-arc basin of northern Pontides. Slicing of the active margin must have continued throughout the period resulting in a HP/LT metamorphism along the suture. Ophiolite obduction onto the passive margins, the Kırşehir and Menderes massifs, must have started towards the end of the Cretaceous period. It is observed that the slicing is imbricate and the lithons widen southward. It is suggested to have progressed towards the south, the earlier slices having been carried on the back of the following. The thrusting has resulted in uplift of the passive margin with consequent gravitational gliding towards the foredeeps that are suggested to form by rotationary processes of the collisional period. There were presumably pockets of unsubducted oceanic crust after the collision (Figure 6), as the continental fragments are not expected to fit like jigsaw puzzles. The closure of these pockets must have been fulfilled with the aid of strike-slip faulting with the consequence of compression and dilatation, the latter being responsible for creation of syn-collisional magmas of essentially Paleogene age. The author disagrees with the theory that crustal thickening may be the cause of partial melting of the upper mantle/lower crust. Because, pressure is hydrostatic in depth and rigid displacements are not possible. Marine sedimentation stops invariably by the end of Lutetian along the Tethyan suture in Western and Central Anatolia.

Suturing along the East Anatolia-Eastern Pontides has not been completed before Late Miocene. The Western Neotethys, or the Bursa Antalya basin, has sutured by the Miocene as indicated by the multistage compressive deformation between Early Eocene- Late Miocene (Gökten et al., 2001), although most of this zone has collided by the Late Eocene. But the imbrication of the suture zone has continued until the Miocene. The sedimentation is continuous including Miocene in the vicinity of the Salt Lake. It seems that there could have been an unsubducted pocket of ocean floor there, which has closed with the aid of NW trending strike-slip faults creating an immense Tertiary magmatism NW of Ankara (Galatya volcanics).

The northward movement of the Arabian plate (McKenzie, 1972) put a brake on rifting of the Maden-Cungus foredeep, continuing so that the Bitlis have been pushed onto this basin following the collision between Eastern Pontides and East Anatolia during the Miocene. The compression has continued to cause uplift and crustal thickening of East Anatolia and formation of new plate margins such as the E-W trending dextral North

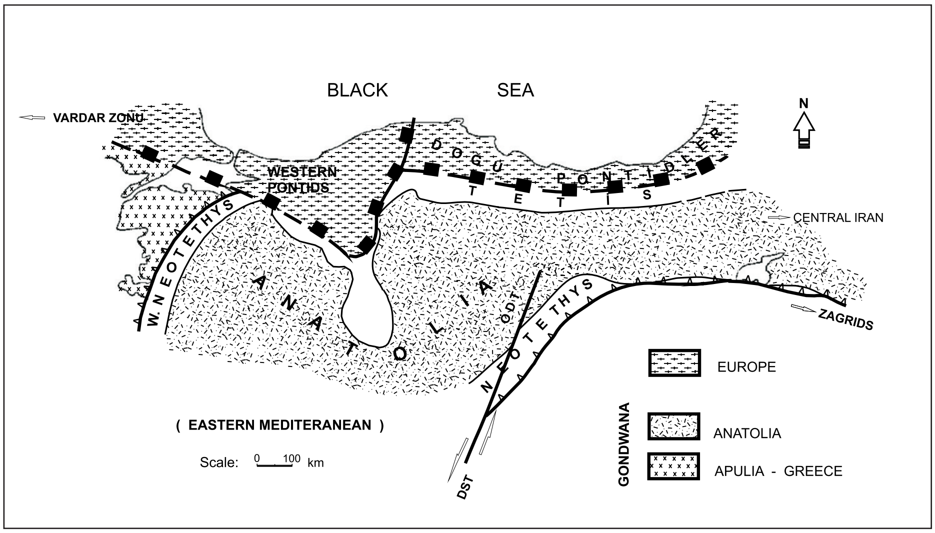

Figure 6- A sketch map showing unsubducted pockets of oceanic crust during the Upper Cretaceous.

Anatolian Fault (NAF) and the NNE-SSW sinistral Ecemis, followed by the NE-SW sinistral East Anatolian (EAF), to push the western part of Turkey onto the ocean floor south of Cyprus and Crete. It seems that the East Anatolian fault and several other sinistral faults en echelon with the Dead Sea transform enable the push of western Anatolia onto the Eastern Mediterranean so that the Ecemiş Fault can be inactive. However, many scientists have questioned this inactivity.

The Anatolian plate escapes west not only because of the northward push of the Arabian plate (McKenzie, 1972) but also the southwestern drag of the Aegean back-arc basin of the Hellenic trench. Otherwise, the movement along the NAF would have died out as the Marmara and the Aegean region had squeezed up. Nevertheless, the North Anatolian tear has occurred because the Pontide plate (The Eurasian) is rigid and stable, implying the southern block to be mobile with respect to a stationary northern block. This is the crucial point on which a deductive process can be started as to locate the area of dilatation on the North Anatolian Fault (NAF) so that a guess can be projected for future earthquakes, which will theoretically migrate eastward. There cannot be a strike-slip fault parallel to the northern coast of the Marmara, for a technical reason, which is the principle that such a strike slip fault has to join a plate margin. Thus, the threat will be from the NAF. The period of time for a new earthquake of the same magnitude along the Marmara segment is not less than 150 my on consideration of the 1-5m displacement in the recent earthquakes and on the assumption of a slip rate of 1.5 (Kasapoğlu, 1984) to 2.5 cm/year.

DISCUSSION AND CONCLUSION

The Tethyan suture is characterised by a HP/LT metamorphism of the active margin and imbrication of ophiolites with the continental crust in the passive margin. A section from the Black Sea to the northern Menderes comprises of very weakly deformed Lower Palaeozoic sediments unconformably capped by the south-facing Karakaya formation. This formation grades into an unrecognizable state towards the suture marked by a fairly wide sliver of ultrabasic rocks obducted onto the passive margin. The passive margin displays ecailling with a widening spacing away from the suture. Concrete field and extensive paleontologic evidence back up the Gondwanian origin of the passive margin, the Taurid-Menderes and the Biga. The Eurasian origin of the Western Pontides is also well established. Origins of these continental fragments locate the Tethyan suture coinciding with that of Brinkmann (1972) between Ankara and Bursa. The basic features of the evolutionary frame may be summarised as:

a) The basic evidence for the evolutionary frame is certainly the geophysical.

b) Sutures are not long distance thrusts, but are rotating systems that are not much longer than 500 kms as in the case of Western Pontides.

c) The Ankara Ilgaz Black Sea line needs further attention, in the sense that separation of Eastern and Western Pontides is far from being thoroughly understood.

This paper comprises of not only substantial evidence but also assertions based on the author's field observations. Nevertheless, the following evidence is independent of the author's perspective.

1- There is a post tectonic sedimentary wedge of Liassic Lutetian age, covering most of the Sa-karya fragment with the implication of dilatation from Liassic onwards (Saner, 1980: Bingöl and Neugebauer, 1992; Şengün, 1992*a*).

2- The Karakaya formation lying adjacent to the Tethyan suture in the Sakarya is European by not only the sequential evidence (Şengün et al, 1990) but also the paleontological (Alkaya, 1990).

3- The sequential and paleontologic evidence (Akdeniz, 1985; Erdoğan. 1990) shows that the Bursa- İzmir segment of the İzmir-Ankara zone (Brinkmann, 1972) is intra-Gondwanian.

4- The overthrust planes of the Sakarya running paralel to the Tethyan suture are of post-Cretaceous age (Şengün et. al.1990).

THE FIELD EVIDENCE WITH REGARD TO SUTURES

Southern branch of Neotethys lied not in the south, but immediate north of the Bitlis/Pütürge massifs (Yazgan, 1984; Çağlayan et al, 1984).

1- There is a continuous Mesozoic sequence exactly the same as the border folds in eastern Bitlis (Çağlayan et. al, 1984).

2- Bitlis suture of Hall (1976) is undeformed on both sides.

Northern branch of Neotethys has never existed. The Neotethyan suture coincides mostly with that of the Tethys (Palaeotethys). The Bursa-İzmir zone, presumably extending to northern Antalya basin and emplacing ophiolites onto southwestern Anatolia (Lycien nappes), is hereby proposed as an intra-Gondwanian ocean. It has started to rift not in Liassic but in Early Triassic.

Concrete evidence is presented showing that the Intra -Pontide Ocean is also a pseudo-suture.

1- There are flat-lying Mesozoic sequences on both sides of this suture.

2- The controlling strike-slip faulting suggested by Okay et. al, (1994) is unsubstantiated.

3- The deformation coincides with that of the North Anatolian fault zone

OUTLINE OF THE GEOLOGIC EVOLUTION OF THE ANATOLIAN SEGMENT OF TETHYS/NEOTETHYS

1- Initiation of the formation of the Tethys at the end of Precambrian.

2- Initiation of rifting in the Anatolian segment of northern Gondwana and onset of northward Tethyan subduction in Early Triassic.

3- Obduction of marginal ophiolites onto the Pontian active margin as the consequence of the dextral rotation of Western Pontides during Lower and Middle Triassic.

4- Recess of the subduction zone and initiation of the Pontian arc in Early Upper Triassic.

5- Detachment of the Anatolian microcontinent from Africa in Upper Triassic-Liassic.

6- Northward drift of the Anatolian microcontinent during Jurassic and Cretaceous.

7- Ophiolite obduction onto the passive margin and the incipient collision of the Apulia Greece with the Strandjha in the Upper Cretaceous.

8- Incipient collision of western Anatolia with Western Pontides in the uppermost Cretaceous-Paleogene.

9- Onset of rotations to close unsubducted pockets of ocean floor, coeval formation of foredeeps and formation of syn-collisonal magmas in Paleogene.

10- Closure of the Salt Lake pocket, collision of the Anatolian microplate with the Aegean and collision of East Anatolia with the Eastern Pontides in Miocene (?).

11- Formation of the plate boundaries, the NAF and the EAF in Late Miocene.

The scenario presented in this paper will hopefully progress in future through questioning of the various other aspects of the evolutionary history of Anatolia.

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PRELIMINARY APPROACH TO GENESIS OF BERYL GROUP MINERALS IN KAYMAZ, NW SİVRİHİSAR, ESKİŞEHİR

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ABSTRACT.- Kaymaz (Sivrihisar) is located at 80 km east of the Eskişehir. The study area consists of metamorphites, ophiolites, phonolites and pegmatites. Beryl crystals are detected in the sediments which cut the metamorphic units. Optical microscopic determinations show that beryl crystals have 35 µm diameter, green and pale greenish blue colours, hexagonal features, basal cleavages, n_o = 1.584 and n_e = 1.584 refractive indexes, uniaxial (-) and euhedral hexagonal prismatic characters which was supported by scanning images (SEM). Be content of phonolites between 9-31 ppm, of pegmatites between 4-17 ppm and of metamorphic units as 1 ppm were determined. Elements associated with beryllium in phonolites with F (260-440 ppm), Ba (1088-3106 ppm), La (300 ppm), Y (16-19 ppm) content, pegmatite with F (300 ppm), W (10 ppm), Sn (5 ppm) contents and presence of beryl crystals in the sediments cutting the metamorphic units which have tectonic relation with phonolite and pegmatite, seem to imply that formation of the beryl minerals have close releationship with these units.

Key words: Beryl, phonolite, pegmatite, metamorphite, Sivrihisar

INTRODUCTION

The study area, situated in 80 km east of Eskişehir province, northwest of Sivrihisar, is shown on the 1/25.000 scaled Eskişehir İ-26 c4 map sheet (Figure 1).

Regional geology was explored by Romieux (1942), Weingart (1954), Kulaksız (1981), and Gözler et al. (1996). However, there is not any known study related with mineralogy, geochemistry and origin of emerald and accociated minerals originated from Sivrihisar locality famous for gemstones market and recorded at historical times.

If world beryllium occurrences are taken into considerations in general, they seem to form mostly in schists, nepheline syenites, phonolites and pegmatites within ophiolite belts. Based on these informations and carrying the whole parameters, Sivrihisar region seems to be a suitable locality to be considered carefully for this research.

MATERIAL AND METHOD

29 clastic and 29 rock samples were collected along creek valleys cutting metamorphics, ophiolites and pegmatites, complying with the principles of geochemical prospecting. Transparent and opaque minerals collected from clastic samples by using stereomicroscope are classified as groups based on their color and crystal forms. Of transparent minerals, colored minerals considered to be beryl were separated, and refractive indexes of these minerals were determined under polarized light with using 0.002 mm spaced immersion oils, and their optical properties were also determined. In addition, these beryl crystals were analyzed by electron microscope (SEM) (Zeiss Supra 50 VP). Rock samples were petrographically analyzed and their mineral constituents and textures were determined. Trace element analyses of 12 samples were performed for the presence of beryl to be determined with ICP-AES method at ACME laboratory (Canada).

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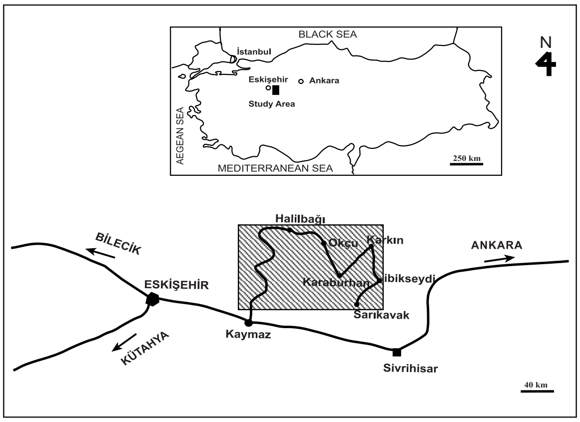

Figure 1- Location map.

GEOLOGICAL SETTING

The study area consists of metamorphics, Karabayır Metaophiolites, Karakaya Granodiorite, Höyüklü formation and Sarıkaya formation (Kulaksız 1981). Metamorphics of Upper Cretaceous aged units forming the basement in the study area include an intercalation of metaquartzite, metapelite, marble, calcschist, metabasite, serpentine schist and metacalcirudite (Okay, 1984; Figure 2). Tectonically overlying Karabayır Metaophiolites contain of metagabbro, metahornblendite, metapyroxenite, metaharzburgite, metaserpentine and metaperidotites. There are also phonolite domes intruding serpentinite along the crack formed by a fault line which have effects up to penetrating depths. K-Ar age of phonolites is given as Middle Miocene (Özgenç, 1982). Karakaya Granodiorite of Upper Cretaceous intruding ophiolites as a pluton bears a pegmatitic vein with quartz, feldspar, tourmaline and pyrite. Höyüklü Formation of Miocene is composed of volcanogenic sandstone, greywacke, tuff, agglomerate and lava flows. Overlying all these units, Sarıkaya Formation of Pliocene begins with volcanic tuff and conglomerate at the bottom and lasts with lacustrine limestone, sandstone, claystone and marl units. This formation overlays locally metaophiolites with angular unconformity (Kulaksız, 1977). The youngest deposits in the study area are Quaternary alluviums (Figure 3).

Emerald and aquamarine were detected by optical analyses on creek sands collected from the valley of Karakız Creek cutting metamorphics to research beryl mineralization. Phonolites were selected as second area within the ophiolites. Pegmatitic vein within granodiorite is considered as third area.

PETROGRAPHIC STUDIES

a- Rock Petrography

Petrographic studies present that metamorphics seem to include garnet-glaucophane schist, epidote-chlorite schist, garnet-lawsonite-glaucophane schist, epidotite, chlorite-lawsonite schist and marbles, Karabayır Metaophiolites contain diabase, serpentinite, ophicalcite and phonolites, and Karakaya Granodiorite seems to bear a pegmatitic vein as well.

In thin sections of lepidoporphyroblastic-textured garnet glaucophane schist traces of chloritization are present on the edges of subhedral garnets with varying sizes between 0.25 and 0.75 mm (Plate I - Figure 1). Prismatic glaucophane crystals with varying sizes between 0.125 mm and 0.5 mm have significant strong pleochroism. With using immersion oils, refractory index of glaucophane is determined as n_x= 1.642, n_y = 1.656 ve n_z = 1.657 and it implies crossite of glaucophane series. As well as garnet and glaucophane, there are muscovite, epidote, chlorite and quartz.

In lepidoblastic - textured epidote - chlorite schist, there are epidote as elongate crystals in the direction of b axis, clinozoisites with bluish interference colors, and chlorite with no aligned groups of crystals.

In nematoporphyroblastic-textured garnet-lawsonite-glaucophane schist, glaucophane min-

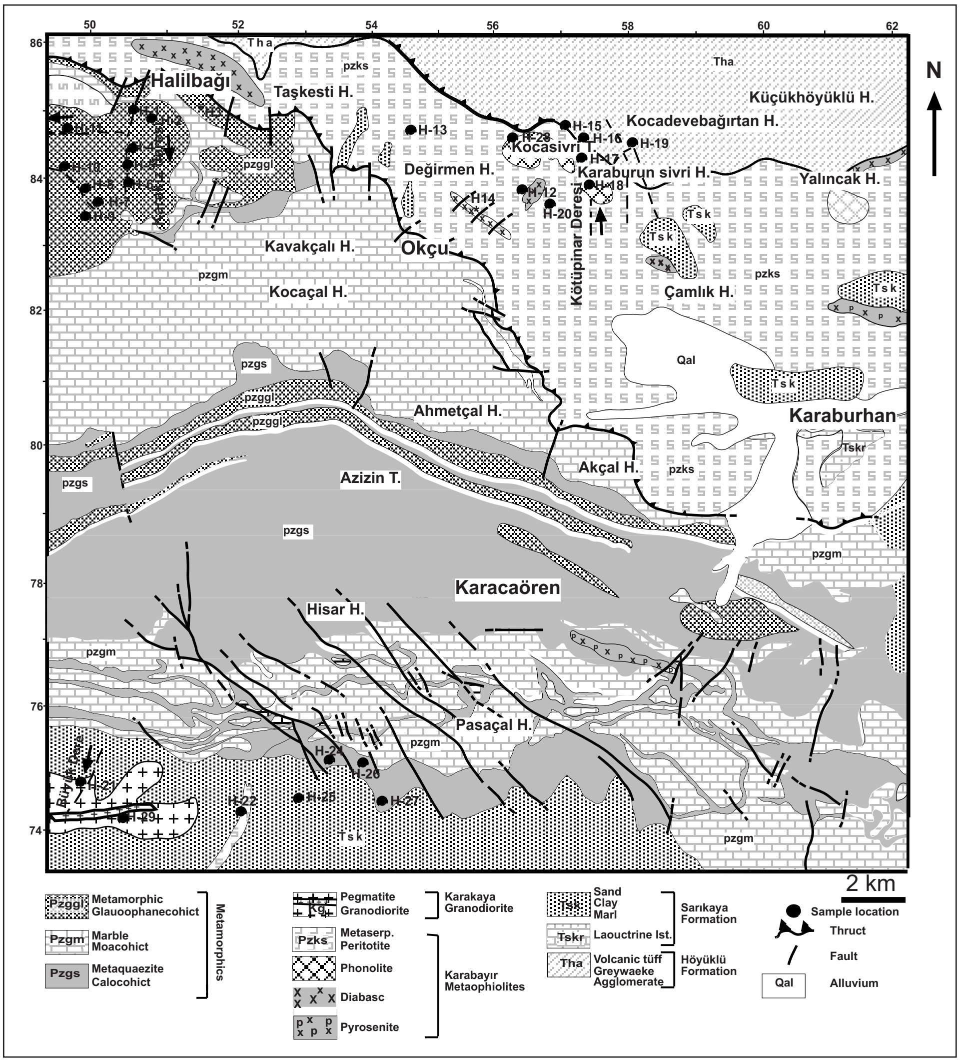

Figure 2- Geological map of the study area (modified from Kulaksız, 1981).

erals have sizes between 0.2 and 0.3 mm. Lawsonite minerals varying sizes between 0.08 and 0.22 mm are short prismatic crystals. There are inclusions in garnet minerals. Epidote minerals are between 0.2 and 0.65 mm, in size.

In nematoblastic-textured epidotite, euhedral albit crystals are found together with epidotes. The size of the calcite mineral inclusions seem to be between 0.15 and 0.3 mm. İn size (Plate I - Figure 2).

In nematoblastic-textured chlorite lawsonite schist, foliated crystals of chlorites surround lawsonite and epidote minerals in bunches. Lawsonite minerals are short prismatic crystals and show parallel extinction. Epidote minerals are 0.13 mm in size and augite minerals are between 0.15 and 0.3 mm in size (Plate I - Figure 3).

In marble sample enclosed in metamorphics, there are very few quartz, sericite, plagioclase and orthoclase besides calcite. Twin glidings and

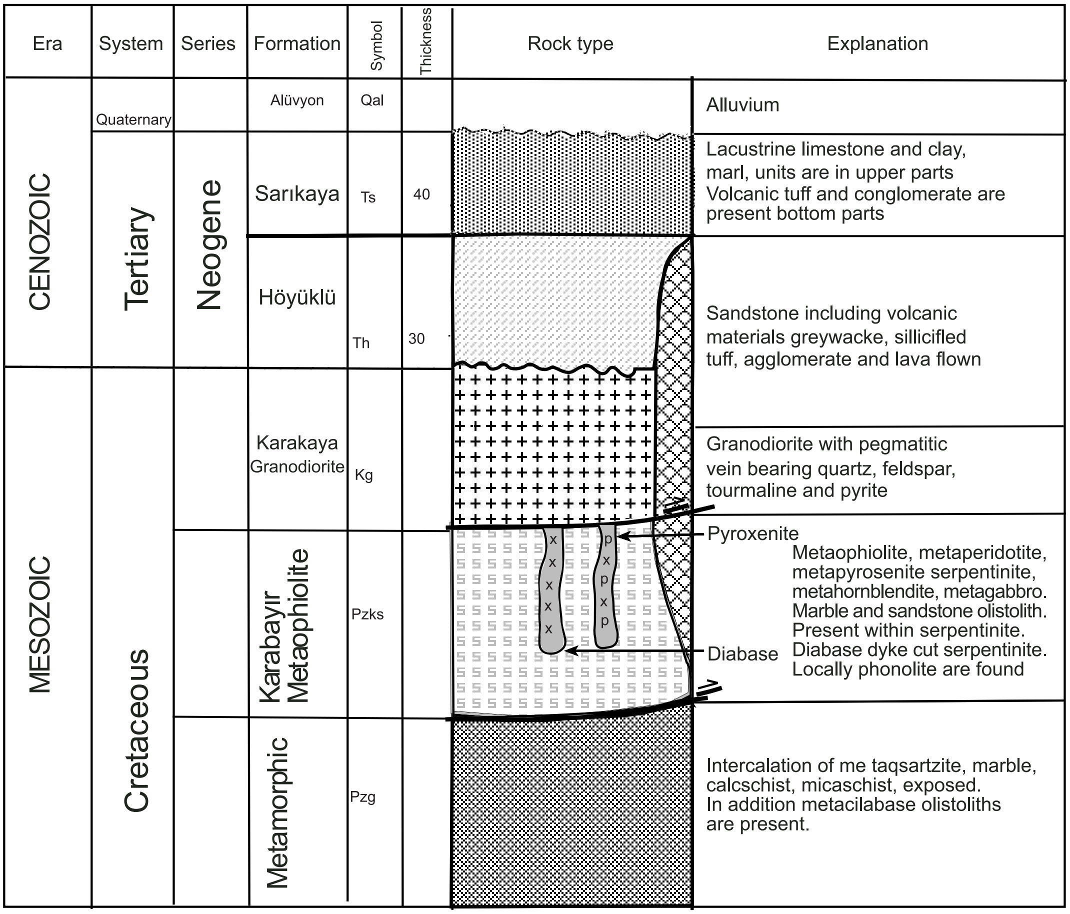

Figure 3- Generalized vertical section of the study area (modified from Kulaksız, 1981; Okay, 1984)

bendings on these twin lamellas are observed in calcite crystals. They show symmetrical extinctions according to the traces of cleavages. Quartz minerals are of anhedral rounded grains and sericite minerals are flakelike assemblages. Few plagioclase and orthoclases of Carlsbad twinning are observed.

Ophitic-textured diabase dyke within Karabayır metaophiolites contain of oligoclase, diopsite, actinolite and zeolite minerals. Diopsites performing uralitization were converted to partial or completely acicular actinolites. Zeolite mineral, defined as natrolite, are found as acicular-fibrous aggregates. In serpentinite, there are antigorite bearing olivine remnants, and fibrous chrysotiles. In the porphyritic-textured phonolite sample (H18), there are euhedral phenocrysts of sanidine varying between 0.2 and 0.7 mm size, hornblende with 0.015 mm. size and leucite minerals. Its groundmass consists of volcanic glass and rock fragments (Plate I - Figure 4). Phonolite sample numbered H28 showing typically trachytic texture contains nepheline and apatite minerals (Plate I - Figure 5). Sieve-textured ophicalcite include twin lamellar calcite, fibrous chrysotile and subparallel aligned antigorite crystals.

Pegmatite sample enclosed in Karakaya granodiorite consists of orthoclase, biotite and quartz minerals. Some of orthoclase minerals show Carlsbad twinnings, some have a combination of Carlsbad and Baveno twinnings (Plate I - Figure 6). Chloritization of biotites ranging sizes from 1 to 4.75 mm are seen from the edges. Quartz grains are between 0.25 and 0.5 mm in size.

b- Beryl Mineralogy

Beryl, with displaying dark green color is defined as emerald, and displaying bluish green to blue coloration, called as aquamarine, and is beryllium aluminosilicate in reality, carrying a formula as $Be_3Al_2Si_6O_{18}$.

25 green and light-green colored beryl grains owning the largest size as 35 µm, numbered as H1 from Karakız Creek cutting metamorphics, were defined as emerald and aquamarine based on the analyses (Plate II - Figure 1).

Views of green crystals (grains) on a stereomicroscope imply that they are crystallized in hexagonal systems (Plate I - Figure 2).

With using immersion oils having 0.002 different values, refractory indexes of the grains were found as n_o=1.568, n_e=1.584 using (Plate II Figure 3).

Beryl minerals, in optical studies, determined as uniaxial (-) and existed as hexagonal prismatic euhedral crystals were defined as emerald when green, and as aquamarine when light bluish green varieties at polarized light in immersion oils. (Plate II - Figure 4). Basal cleavage (0001) is significant (Plate II - Figure 5).

Beryl crystals determined on stereomicroscope were analyzed by SEM (Plate III - Figs 1 and 2) as well. Beryl crystals are euhedral hezagonal prismatic in general. 0001 planar sections of these crystals are six cornered and locally circular. Prismatic beryl crystals showing circular and elipsoidal 0001 planar sections were probably derived from partial abrasion of hexagonal prismatic crystals resulting from probable transportation within detrital sediments.

GEOCHEMICAL ANALYSES

12 samples were taken so as to determine if beryl is existed by performing chemical analyses, in the probable emerald promising areas. Of these samples, 6 samples were taken from Karakız Creek cutting metamorphics (H1-H6), one sample was taken from Kötüpınar Creek cutting phonolites (H18) and one sample was taken from Büyük Creek clastics cutting pegmatites (H29). Rock samples were analyzed as metamorphics (H1), phonolites (H18 and H28), and pegmatites (H29). To compare sediments and rock samples and if there is possibility of any rock types to bear beryllium mineralizations, both rock and sediment samples were analyzed.

Samples taken from Karakız creek (H1- H6) bear 1 ppm Be, creek sands taken from Büyük Creek (H21) cutting pegmatites contain 4 ppm Be (Chart 1). The highest Be value were determined as 9 ppm on creek sands taken from Kötüpınar Creek (H18) cutting phonolites. Related with these sediments, Be content of Karaburunsivri phonolite sample (H18) is 31 ppm, of Kocasivri phonolite (H28) is 19 ppm, of pegmatitic rock (H29) is 17 ppm and metamorphics (H1) bear less than 1 ppm Be. These data show us beryl has probably been derived and enriched from alkaline phonolitic and pegmatitic rocks.

Useful indicators to determine presence of beryl are F, Li, Rb, Cs, Sn, W, and accociated elements are Ba, Sr, B, Sc, Y, and other rare earth elements are U, Th, Nb, Ta, P, Ti, Mo and Mn (Boyle, 1974). Because of Be and Li, Rb, Cs, Sr, Ba, B, Sc, Y, Ti, Th, P, V, Nb, Ta, Cr, W, U, Mn and F are lithophile elements, they are interrelated (Akyol et al., 1985). Of accompanying elements with beryllium, various cations of small radius and often greater valence (U 10-22, Th 4-142, Mo 2-14, W 4-17, Nb 3-182, Sn 2-5 ppm) are called as incompatible elements. Owing to ionic radius and valences, the element Exchange with major ions in silicate minerals seem to be

Table 1- Results of geochemical analyses of the samples taken from the study area

Sample No Type	H1 Sediment	H2 Sediment	H3 Sediment	H4 Sediment	H5 Sediment	H6 Sediment	H21 Sediment	H18 Sediment	H1 Schist	H28 Phonolite	H29 Pegmatite	H18 Phonolite
Mo(ppm)	<2	<2	<2	<2	<2	<2	<2	14	<2	<2	2	<2
Cu(ppm)	34	38	44	56	65	31	14	35	10	3	4	31
Pb(ppm)	11	21	14	11	6	11	58	32	8	189	134	18
Zn(ppm)	51	50	62	78	81	42	32	94	33	308	22	84
Ag(ppm)	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
Ni(ppm)	306	192	188	327	247	111	95	107	681	17	5	30
Co(ppm)	31	26	33	40	47	16	9	27	50	<2	<2	27
Mn(ppm)	1590	2596	1595	1718	2663	1014	373	2080	644	1474	294	1033
As(ppm)	<5	<5	<5	<5	<5	<5	25	34	<5	16	14	9
Au(ppm)	<4	<4	<4	<4	<4	<4	<4	<4	<4	4	<4	<4
Th(ppm)	<2	5	4	4	4	5	8	65	<2	142	39	5
Sr(ppm)	137	150	125	143	196	232	131	290	53	976	37	467
Cd(ppm)	<0.4	0.6	0.7	0.5	<0.4	0.5	0.5	0.6	0.5	<0.4	<0.4	<0.4
Sb(ppm)	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5
Bi(ppm)	7	7	9	8	<5	<5	<5	<5	<5	<5	<5	<5
V(ppm)	103	111	114	148	204	86	21	203	18	99	4	196
La(ppm)	20	17	27	29	33	18	18	165	<2	64	19	21
Cr(ppm)	523	521	278	490	423	219	325	336	619	2	<2	80
Ba(ppm)	96	109	194	210	172	121	275	5008	36	3106	71	1088
W(ppm)	<4	<4	<4	<4	<4	<4	4	8	<4	<4	10	17
Zr(ppm)	7	12	17	24	17	10	21	72	<2	801	35	67
Sn(ppm)	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	5	<2
Y(ppm)	19	22	19	22	30	14	6	32	2	19	4	16
Nb(ppm)	5	7	11	10	11	4	9	22	3	182	25	9
Be(ppm)	1	1	1	1	1	1	4	9	<1	19	17	31
Sc(ppm)	13	16	13	20	27	10	2	14	3	<1	<1	25
F(ppm)	-	-	-	-	-	-	-	-	10	260	300	440
U(ppm)	<10	<10	<10	<10	<10	<10	<10	<10	14	22	<10	15
Ca(%)	6.13	7.69	7.61	3.36	3.77	7.83	3.21	1.22	6.43	1.16	0.27	3.77
Mg(%)	3.34	2.35	2.24	3.88	3.69	1.51	0.61	1.35	9.6	0.13	0.07	2.38
Ti(%)	0.29	0.31	0.43	0.41	0.58	0.21	0.10	0.47	0.01	0.16	0.02	0.53
Al(%)	3.54	3.81	4.25	5.32	5.81	3.48	4.40	7.01	7.35	10.4	6.67	8.26
Na(%)	0.62	0.62	0.62	0.80	1.08	0.53	0.91	1.67	0.72	5.86	2.77	3.35
K(%)	0.28	0.44	0.60	0.74	0.64	0.45	2.83	2.81	0.04	4.55	3.64	1.76
P(%)	0.058	0.054	0.059	0.072	0.094	0.040	0.018	0.078	0.004	0.019	0.002	0.065
Fe(%)	4.17	4.95	4.18	5.88	7.26	3.44	1.34	5.17	3.16	2.59	0.73	5.63

difficult (Krauskopf, 1979). On phonolite samples (H18,H28) values of U as 15-22, of Th as 5-142, of Nb as 9-182, of Mn as 1033-1474, of Sr as 467-976, of La as 21- 64, of Ba as 1088-3106, of Al as 8.26-10.4, of Y as 16-19, and of F, a significant indicator for determining beryl, as 440-260 ppm were determined. On pegmatitic rock sample (H29) higher values of W as 10, of Sn as 5, of F as 300 ppm than other rock samples support that beryllium occurrences on the veins and pyrometasomatic beds own a geochemical affinity between beryllium and W, Sn and F in particular. (Warner et al., 1959).

On clastics from Karakız Creek cutting metamorphics, the values of Cu, Ni, Co, Mn, Ca, Cr, Mg, Sc, V, and Fe; on a garnet-glaucophane rock sample taken from the surroundings of same creek values of Ni and Cr associated with mafic volcanic and ultramafic rocks; on garnet, glaucophane, epidote and chlorite minerals values of Co, Fe, Ca, and Mg are high. It is noteworthy that the rock sample bear less Cu, Mn, Sc and V than clastic sample does.

On clastics from the creek cutting pegmatite Pb and K got concentrated, and pegmatite sample (H29) enriched in W, Sn and F, and depleted in Sr and Ba. This explains that pegmatite veins bear wolfram, tin and fluorine beds. On sediment sample taken from Kötüpınar Creek cutting Karaburunsivri phonolite, the values of Al, Na, Mo, Zn, As, Th, Sr, La, Ba, K, W, Zr, Y, and Nb seem to be considerably high. On phonolite samples (H18, H28), LFSE elements (Pb, Mn, Na, Th, Sr, Ba, U, K), HFSE (Y, Nb, Zn, Ti, Zr), LREE (La), transition elements (V, Sc), lithophile elements (W, F) and other elements (As, Cu, Fe, Al) got enriched.

Beryl may include alkaline ions like Na and K, and its total alkaline content may rise up to 5-7 %. Beryllium replaces Na, on the contrary, doesn't replace K. Alkaline elements remain within hexagonal channels in the lattice structure of a beryl (Çelik ve Karakaya, 1998), so high-beryllium containing samples bear alkaline elements of considerably enriched values (0.04 - 5.86 %) like Na and K. In fact, Kocasivri and Karaburunsivri phonolitic rock samples enriched in Na, K and Al, and depleted in Ca and Mg show that these rocks carry alkaline character.

It is indicated that there is a slight positive correlation between beryllium and Nb, Mo, Sn, F, Ba, Sr, U, Th and La elements; a negative correlation between beryllium and Sc, P, Ti, Mn, Cr and V, and no correlation between W and Y (Figure 4).

Results of chemical analyses show that a positive correlation between Be and Al may exist except for H1 sample, and Be element enters the lattice structure of Al-bearing minerals (sanidine, leucite, nepheline, hornblende, biotite, orthoclase). Because ionic radii and ionic valences of Be and Al are similar, these elements are isomorphic.

DISCUSSION

It can be referred that beryl minerals were formed due to phonolites and pegmatites in the northwest of Kaymaz (Sivrihisar). Beryllium element's indicating an anomaly on phonolitic and pegmatitic units, and the determination of beryl crystals in sediments cutting metamorphic rocks related with these units seem to exhibit significant geochemical relationship between Be and the rocks. According to Marshall et al. (2003), emerald occurrences are closely related with Cr (+/- V) and Be-bearing solutions in general. The authors state that Cr and V could be derived from local mafic and ultramafic rocks during hydrothermal alteration. At the sample of Kocasivri phonolite (H28), having a high value of V and a negative association, and a higher value of Cr in metamorphic rocks than that of other rocks indicate that chromium transportation is probably resulted from thrusting. Optically determined beryl minerals might have been formed due to an association with schist minerals following the

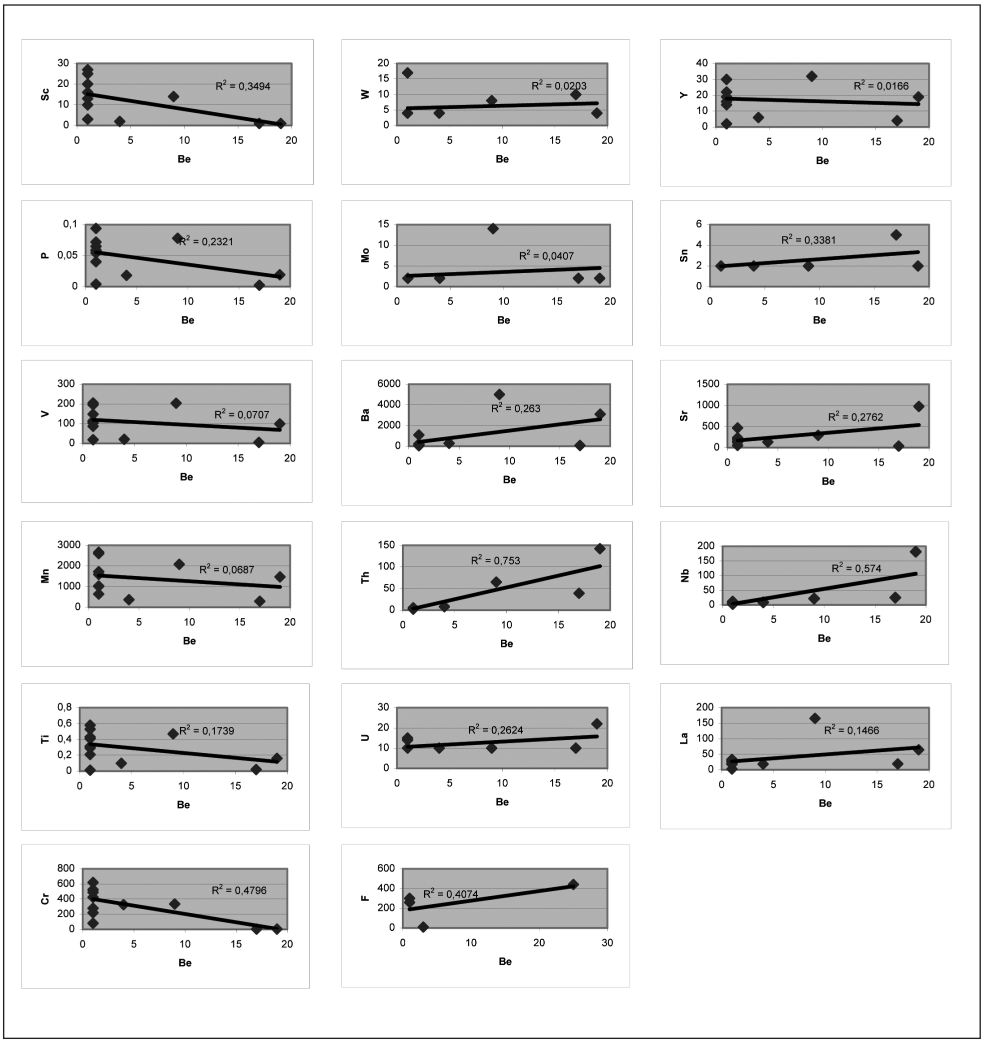

Figure 4- Correlation charts of elements associated with Be.

metamorphism of original rock during a hydrothermal process. Depending with upthrust, alkaline magma derived solutions were filled the fissures and joints of the metamorphic rocks, as result, beryl got deposited as trace amounts.

A positive correlation between Be and Na, and also K in phonolitic and pegmatitic rocks shows that beryllium could be present in a crystal lattice structure of some Na and K-bearing minerals like in sanidine, nepheline, hornblende, orthoclase and apatites. Because of trace amounts of beryllium element, beryl minerals can not be observed in thin sections of phonolite and pegmatite rocks and are mostly determined by chemical analysis methods.

Beryllium containing solutions are products of alkaline volcanism together with F in phonolites. Be composing between 9 and 31 ppm, and F between 260 and 440 ppm, and also Be between 4 and 17 ppm, and F as 300 ppm in pegmatites show the presence of a positive correlation between Be and F, and a probable relation as in origin, between them. Where beryllium's having hihger values, F gets high values, fluorine play a reactive role separating F from a solution phase, and beryllium was facilitated to get separate in pneumatolytic and hydrotermal solutions. Be-bearing minerals form due to conditional changes of the solutions like pH, pressure and temperature or due to decaying of complexes like $[BeF_4]^{2-}$, $[BeF_2]^{-}$, $[BeF_2]°$, $[Be(CO_3)F]^{-}$ ve $[Be(CO_3)_2]^{2-}$ when they interact with surrounding rocks (Gökçe, 2000). Either beryllium enters a lattice structure of silicate minerals or enriches in residual solutions during magma crystallization. Which of these happened is dependent upon the fluorine content of magma, hydrostatic pressure, alkalinity of surrounding rocks and other factors. Therefore, Be either is deposited in granitic pegmatites or is carried with fluorine and formed in hydrothermal occurrences and in greisens. Higher F value on Kızılcaören fluoride barite ore formed at the contact of phonolites in the study area shows that they are derived from the same volcanism. Hence, beryl occurrences are associated with phonolites and pegmatites in origin. It is also reported by Preinfalk and Morteani (2002) that similar occurrences derived from anatectic pegmatites in Belmont and Capoeriana (Minas Gerais, Brazil) reacted with Cr-rich ultrabasic rocks in metasomatism, Beryl occurrences in the study area might have been formed as a result of metasomatic reaction between pegmatites and Cr-rich ophiolitic rocks

CONCLUSIONS

1- The presence of beryl as emerald is firstly identified in sediments of Kaymaz locality (NW of Sivrihisar- Eskişehir) in Turkey by mineralogical and geochemical studies.

2- Optical properties of beryl minerals are hexagonal prismatic euhedral crystals, green-colored varieties called emerald, bluish green varieties called aquamarine, basal cleavage, determined light refractory indexes n_o= 1.568, n_e= 1.584, 1st row interference colors and uniaxial (-). On the other hand, beryl crystals on SEM are identified as euhedral hexagonal prismatic forms.

3- For the identification and origin of beryllium, Beryllium shows positive correlations with F, Ba, Sr, U, Th, La, Nb, Sn and, W elements, seem to be considerably important data.

4- Beryl minerals associated with beryllium in geochemistry as volcanogenic hydrothermal products related to phonolites and pegmatites.

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PLATES

PLATE I

Figure 1- Garnet glaucophane schist (H1)
Garnet (gr) indicating chloritization at its environs, and glaucophane (gl).
Single nicole, Obx Ok= 5x10

Figure 2- Epidotite (H7)
Epidote (ep) and calcite
Double nicole, Obx Ok= 10x10

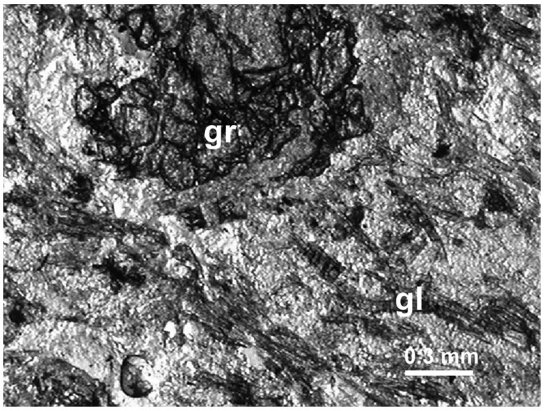
Figure 3- Chlorite lawsonite schist (H9)
Chlorite (kl) with foliated crystals, and short prismatic crystals of lawsonite (lv), Double nicole, Obx Ok= 5x10

Figure 4- Karaburunsivri phonolite (H18)
Hornblende (hnb), sanidine (sa) and leucite (lö)
Double nicole, Obx Ok= 4x10

Figure 5- Kocasivri phonolite (H28)
Hornblende (hnb) and sanidine (sa) minerals
Single nicole, Obx Ok= 4x10

Figure 6- Pegmatite (H29)
Baveno and Carsbad twins on orthoclase are seen.
Double nicole, Obx Ok= 5x10

PLATE - I



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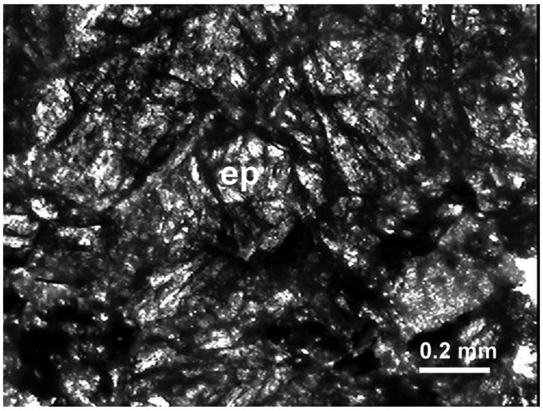

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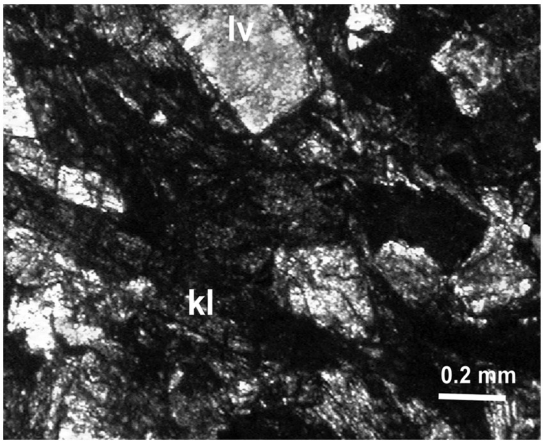

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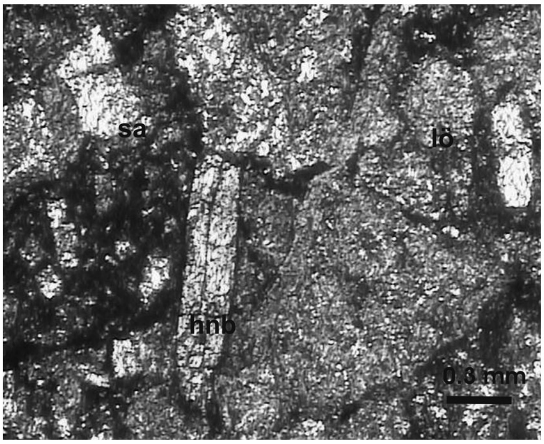

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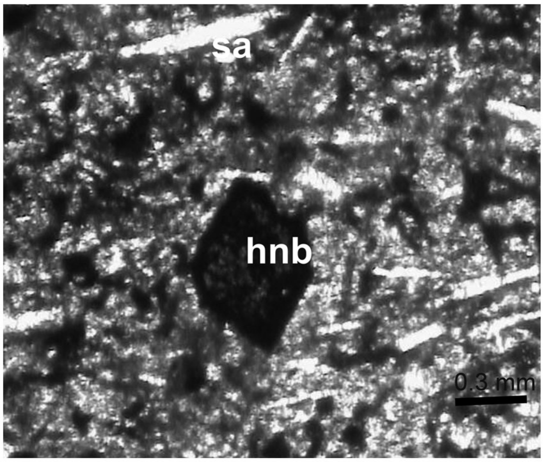

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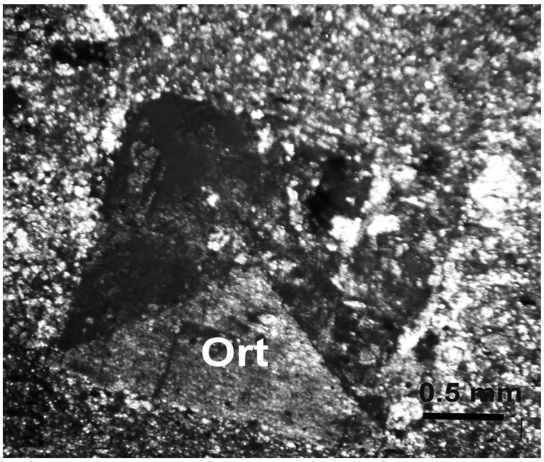

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PLATE II

Figure 1- Stereomicroscobic view of the crystals of aquamarine (A) at the left, emerald (Z) at the right, enlargement = 35X

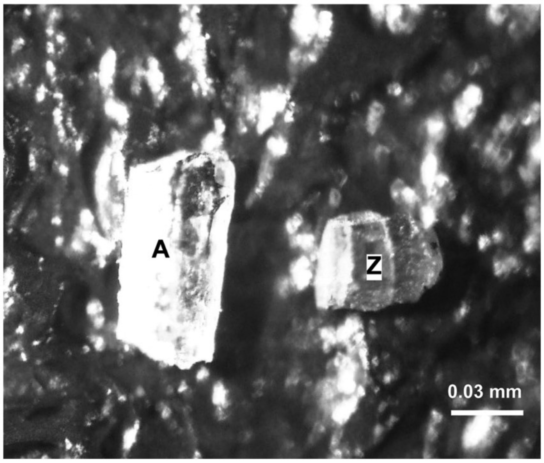
Figure 2- Emerald photo taken by stereomicroskobic view, enlargement = 35x

Figure 3- View of a beryl grain in a 1.60-immersion fluid

Figure 4- Emerald at single nicole, Obx Ok= 5x8

Figure 5- Basal cleavage (0001) on aquamarine is seen. Single nicole, Obx Ok= 5x8

PLATE - II



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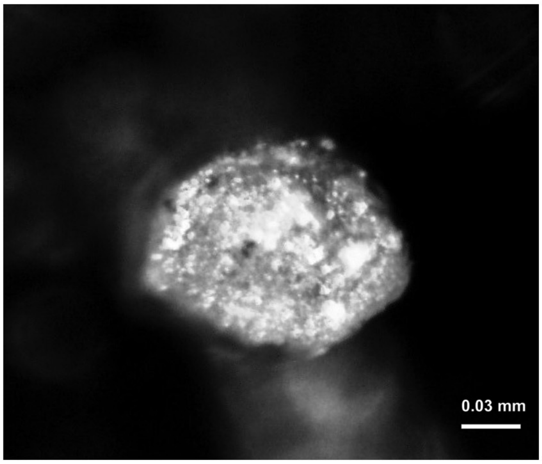

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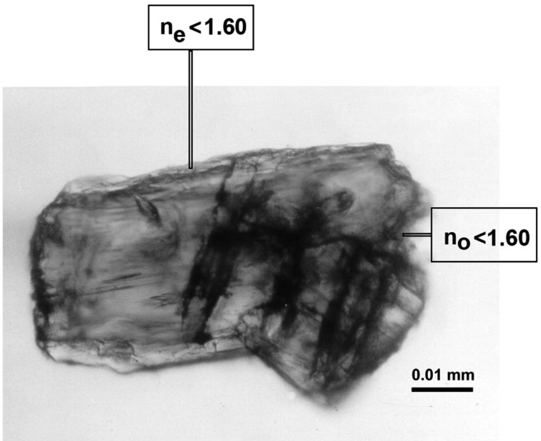

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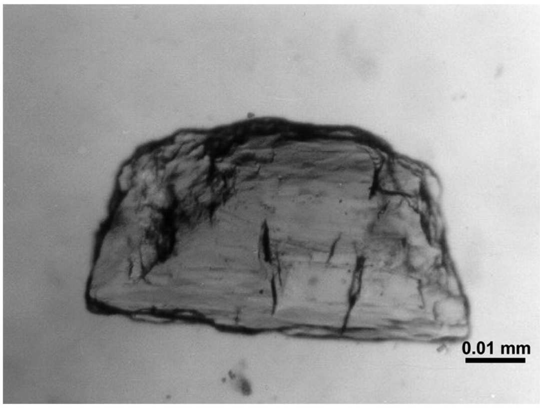

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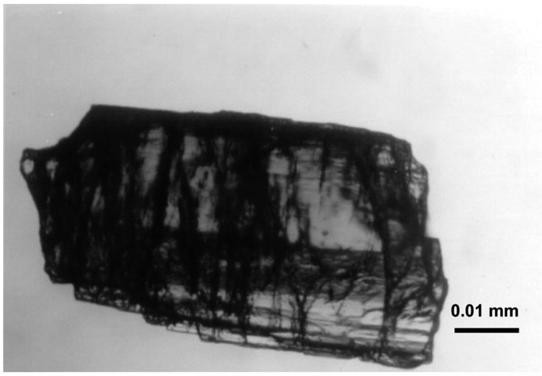

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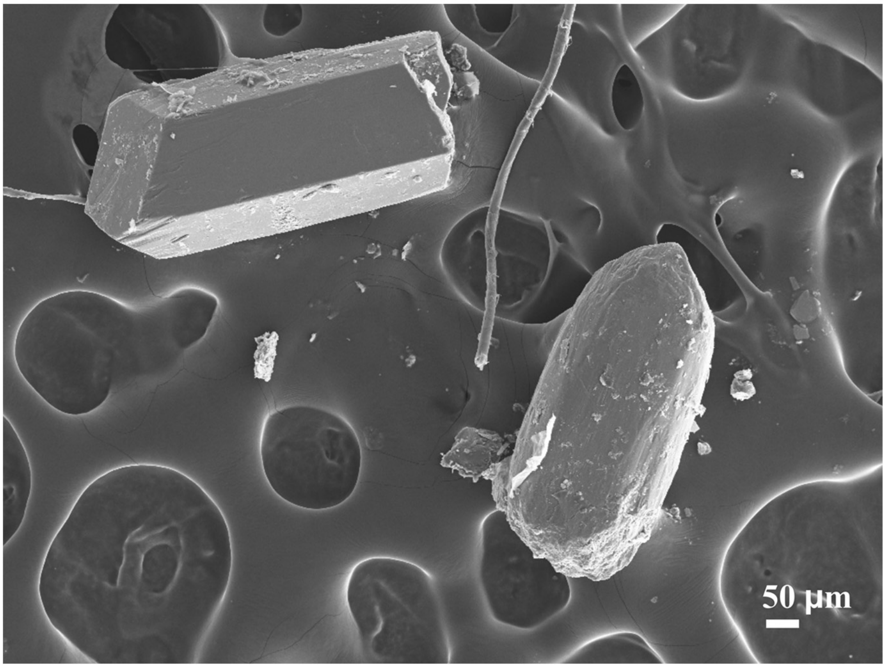
PLATE III

Figure 1- General view of beryl crystals

Figure 2- Surficial view of a hexagonal prismatic beryl crystal (0001)

PLATE - III



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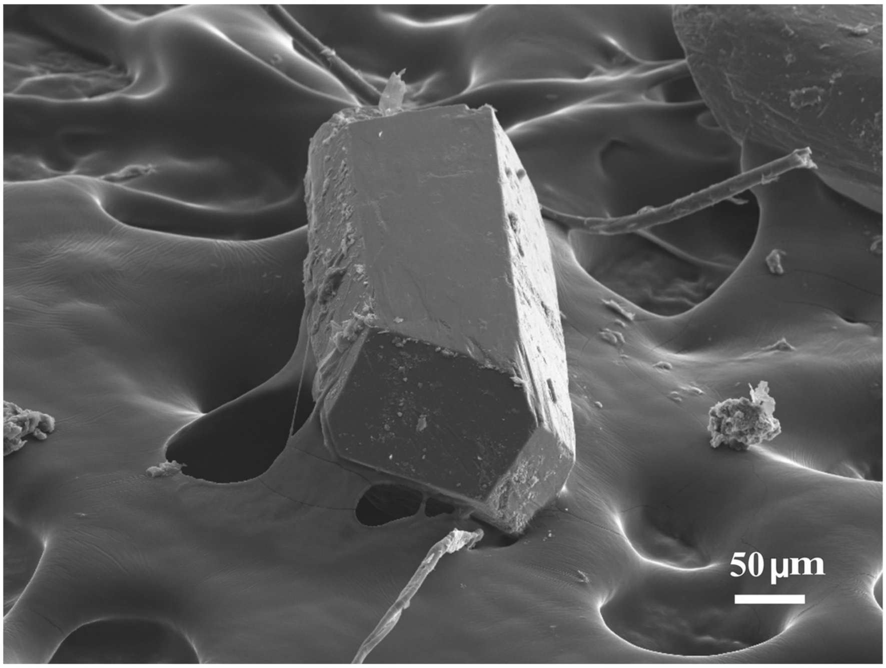

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Mineral Res. Exp., Bull., 133, 41-49, 2006

THERMAL REACTION OF ANTIGORITE: A XRD, DTA-TG WORK

Gökçe GÜRTEKİN* and Mustafa ALBAYRAK*

ABSTRACT.- It is very important to determine the stability limits of serpentine minerals for investigation of geologic processes taking place in the oceanic lithosphere which consists of serpentinized peridotites. As a result of temperature increase in the subducting oceanic lithosphere, thermal decomposition of serpentine minerals and new mineral formations have a great effect on geologic processes related to subducting plate. In this study, serpentine minerals (antigorite) collected from the Konya-Çayırbağı region were dehydrated under constant atmospheric pressure, then mineralogical changes were determined by using X-Ray diffraction and DTA-TG analyses and finally the stability limits of antigorite was determined. Dehydration reactions on antigorite started at approximately 100-150°C, dehydroxilation reactions started at approximately 550°C and beyond this temperature forsterite started to crystallize from antigorite+brucite. Association of antigorite+forsterite continued until 650°C at which enstatite started to be formed. During all the reactions, which were carried out at the atmospheric pressure, talc was not formed but some H_2O and amorphous silica were released. Dehydroxilation reaction on antigorite occurred between 550°C and 690°C and antigorite was stable until 650°C-690°C.

Keywords: XRD, DTA-TG, Antigorite, Dehydration, Dehydroxilation

INTRODUCTION

Although several types of serpentine minerals are found in the nature, the most common ones are lizardite, chrysotile and antigorite. These minerals have the general formula of $Mg_3Si_2O_5(OH)_4$ and are described as trioctahedric (1:1) layered silicates with MgO, SiO_2 and H_2O contents between 85-95%.

Serpentine minerals are formed as a result of hydration of olivine, pyroxene and other Mg-rich silicate minerals under suitable conditions, the process so called serpentinization. Lizardite and chrysotile are formed at lower conditions of greenschist facies while antigorite mostly occurs at greenschist-amphibolite facies (Evans, 1977).

In studies conducted on serpentine minerals, minerals are generally described by petrographical methods and X-ray analyses and their formation conditions are determined and thus the rock formation processes are investigated (Gürtekin, 2001). Hydrated oceanic crust material is composed of more than 90% of serpentine minerals. During the subduction process, thermal reactions on serpentine minerals cause water release which can play an important role in subduction-related volcanic processes. Completely serpentinized peridotites contain significant amount of water (13%) which released as a result of subduction process this water can be transported to great depth which is represented by high temperature-high pressure conditions at which serpentine minerals can be stable and arc magmatism starts. Therefore, thermal reactions of serpentines and related mineralogical changes are very important for identification of some geologic processes.

The aim of this study is to describe serpentine minerals on the basis of petrographic, X-ray and DTA-TG methods, to investigate the mineralogical changes occurring during the thermal reactions and to determine the stability of antigorite under constant pressure. Samples used in the

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study were taken from the Çayırbağı ophiolite in the Çayırbağı-Hatip region in Konya (Figure 1) and serpantinites in the Hatip ophiolitic complex (Özcan et al., 1990).

THERMAL REACTION OF SERPENTINE MINERALS

Loosing of the absorbed water (H_2O) related to the increasing of heat in a mineral is described as dehydration and loss of water in its crystal structure is descbried as dehydroxilation. Process of thermal reactions in serpentine (dehydration and dehydroxilation) and forming of olivine+ pyroxene+talc+chlorite, that minerals include a few or no water is described a deserpentinization (O'Hanley, 1996). Because, lizardite is firstly formed from olivine during serpentinization, in deserpentinization olivine mostly formed from antigorite and lizardite in deserpentinization.

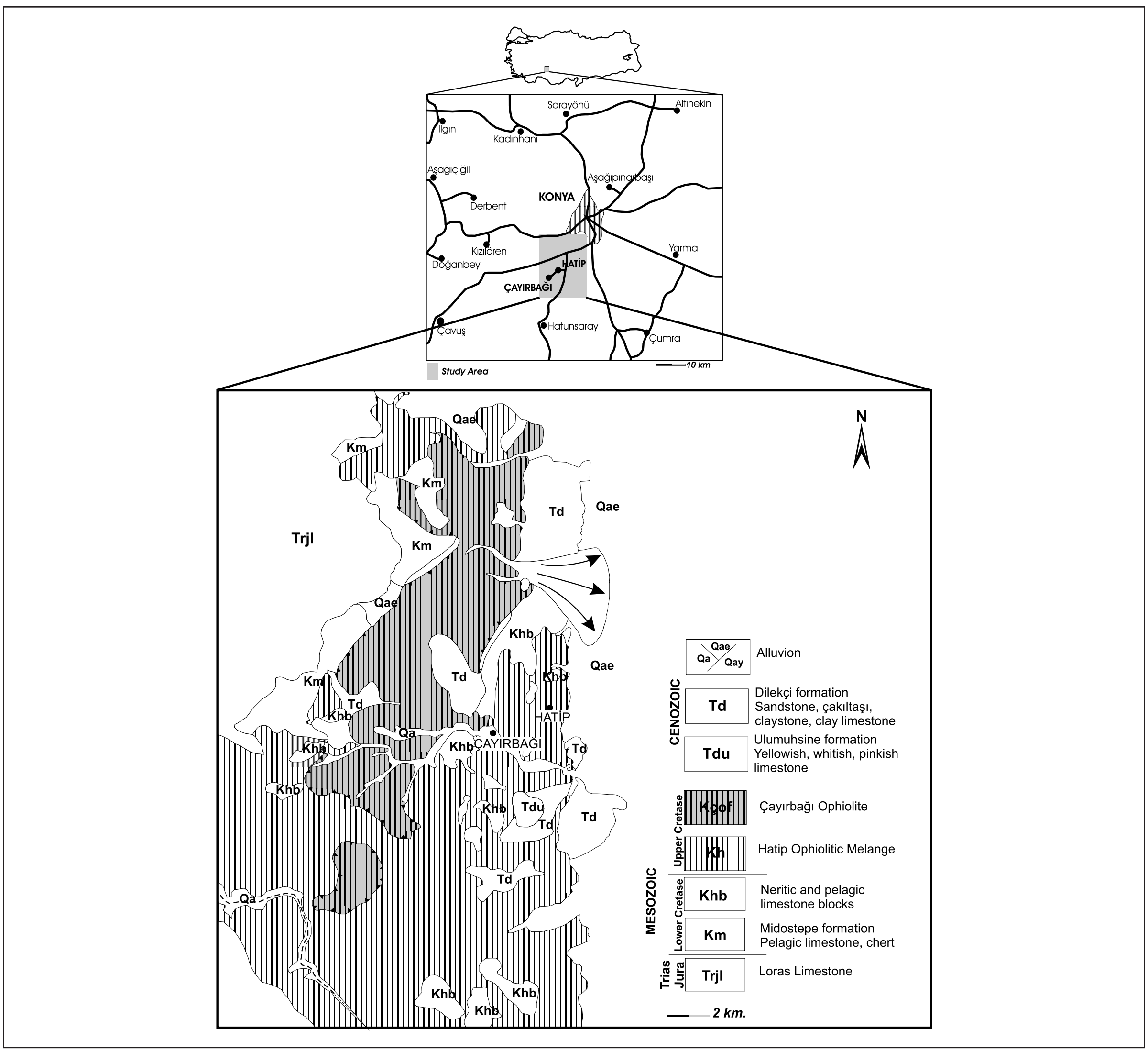

Figure 1- Geological and location maps of the study area (Revised fron Özcan et al., 1986)

In general, thermal reactions in antigorite during recessing serpentinization are in the form of antigorite + brucite ↔ olivine + H_2O orantigorite ↔ olivine ± talc + H_2O and new minerals are formed at temperatures about 500-690°C (Evans, 1977). Considering the stability boundaries of serpentine minerals (Figure 2), lizardite and chrysotile are found at lower conditions of green-schist facies while antigorite is stable over a wide range from lower green-schist to amphibolite facies.

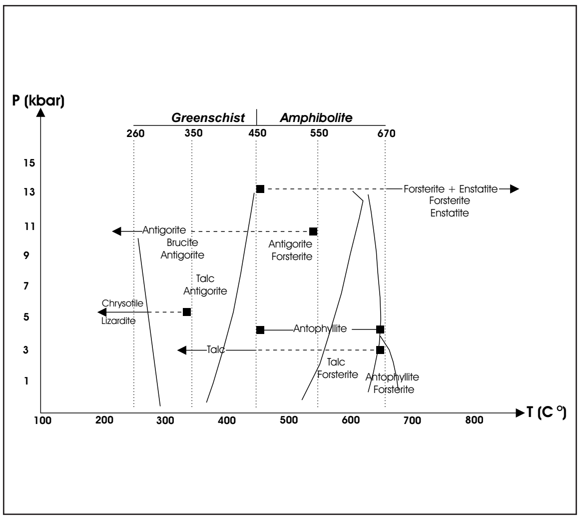

Figure 2- Stability limits of serpentine minerals (Buc-ker and Frey, 1994)

MATERIAL AND METHOD

This work was conducted as two stages: X-ray analyses and DTA-TG analyses. However, prior to X-ray and DTA-TG analyses, petrographical studies were carried out to determine textural characteristics, mineral paragenesis and source rock types of serpentinites. In petrographical studies, mineral paragenesis and textural properties of rocks were described and classifications of Wicks and Whittaker (1977) which were later developed by O'Hanley (1991, 1996) were used.

X-ray analyses of 5 samples from the study area were made with Rigaku Geigerflex brand X-ray diffractometer at the MTA General Directorate. Samples were thermally treated at constant pressure at the laboratory and the resulting mineralogical changes were determined with the X-ray analyses. Since the main peaks have very close values, secondary peaks are very important for evaluation of serpentine minerals and therefore, analyses were conducted at 2θ =0-70º. Following the routine analyses performed at ambient conditions, powdered samples were first mixed with pure water and taken into suspension then they were dried on glass lamellas. Before the X-ray analyses, dried samples were left in furnace for about one hour at temperatures of 200°C, 400°C, 500°C, 550°C, 600°C, 650°C, 700°C, 800°C, 900°C, 1000°C and 1100°C. In order to prevent melting of glass lamellas, porcelain containers were used for heating processes at 900°C, 1000°C and 1100°C.

DTA-TG analyses, which give information on phase transformations in parallel to temperature increase and the amount of mass loss, were performed with Rigaku Thermoflex TG 8110 brand device at the MTA General Directorate. During the analyses which were conducted at temperatures between 18.7°C and 1100°C, temperature increasement were selected as 20°C/minute and 100 mg sample was used for the analyses.

Petrography

In petrographical studies, source rock of serpentines was found as harzburgite. In completely serpentinized rocks with pseudomorphic texture, serpentine±talc±magnetite mineral paragenesis is observed but olivine (forsterite) and orthopyroxene (enstatite) minerals are absent and serpentinized (bastitized) orthopyroxene relicts showing plastic deformation signs were also detected. According to classification of Wicks and Whittaker (1977) and O'Hanley (1991, 1996) on the basis of textural features under microscope, antigorite±lizardite and vein-type chrysotile were found in samples of pseudomorphic texture.

X-Ray Analyses (XRD)

In order to determine mineralogical changes as a result of thermal reactions, samples that were heated at certain temperature were subjected to X-ray analyses. For easily determination of mineralogical changes, block diagrams prepared using the Jade program that show the results of analyses at various temperatures were presented in figures 3, 4, 5, 6 and 7.

Results of routine analyses are conformable with those of petrographic studies and antigorite was the main mineral observed. However, due to its lower abundance (2-3%), chrysotile could not be detected in X-ray analyses. There was no mineralogical change in analyses of all samples conducted at 200°C, 400°C and 500°C while only a little decrease was observed in the antigorite abundance with temperature increase. It was observed that the antigorite abundance was continued to decrease at 550°C and olivine (forsterite) started to crystallize. At 600°C, antigorite abundance was significantly decreased and forsterite continued to crystallize and its abundance was increased. At 650°C, antigorite was mostly disappeared and forsterite continued to crystallize and enstatite was started to appear. In analyses performed at temperatures higher than 700°C, forsterite and enstatite continued to crystallize and their abundances were increased. Considering the analyses conducted at 550°C, 600°C and 650°C, a significant amorphousization was detected in samples 11S and 29A (Figure 8).

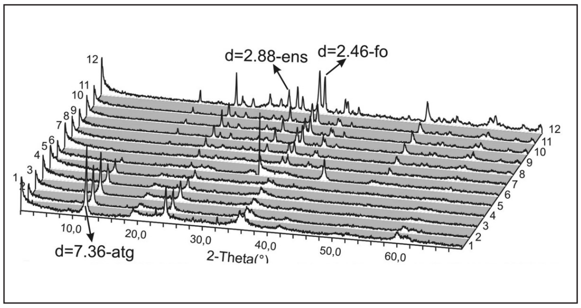

Figure 3- XRD block diagram of sample 1. The num bers 1: Dried, 2:200°C, 3:400 °C, 4:500°C, 5:550°C, 6:600°C, 7:650°C, 8:700°C, 9:800°C, 10:900°C, 11:1000°C, 12:1100°C show the XRD analyses at that temperatures. (atg: antigorite, fo: forsterite, ens:enstatite)

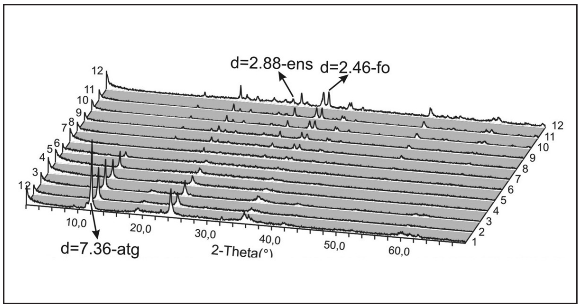

Figure 4- XRD block diagram of sample 6. The numbers 1: Dried, 2:200°C, 3:400 °C, 4:500°C, 5:550°C, 6:600°C, 7:650°C, 8:700°C, 9:800°C, 10:900°C, 11:1000°C, 12:1100°C show the XRD analyses at that temperatures. (atg: antigorite, fo: forsterite, ens:enstatite)

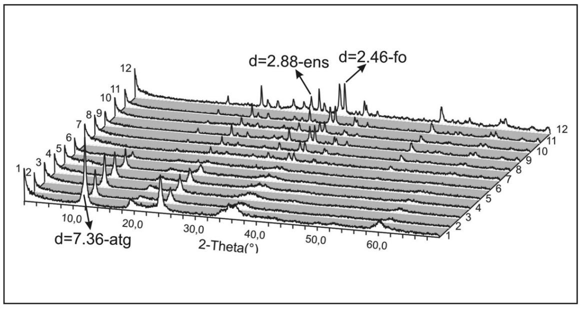

Figure 5- XRD block diagram of sample 11B. The numbers 1: Dried, 2:200°C, 3:400 °C, 4:500°C, 5:550°C, 6:600°C, 7:650°C, 8:700°C, 9:800°C, 10:900°C, 11:1000°C, 12:1100°C show the XRD analyses at that temperatures. (atg: antigorite, fo: forsterite, ens:enstatite)

DTA-TG Analyses

According to results of DTA analyses, samples 1 and 29A yielded endothermic and exothermic peaks at 695°C and 810°C, respectively. Samples 6, 11B and 11S also showed similar features with three endothermic peaks at 210°C, 380°C and 690°C and, one exothermic peak at 810°C (Figures 9, 10, 11, 12 and 13).

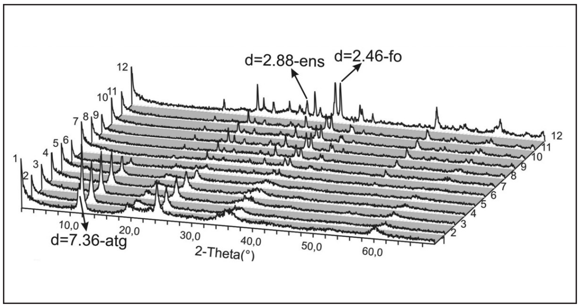

Figure 6- XRD block diagram of sample 11S. The numbers 1: Dried, 2:200°C, 3:400 °C, 4:500°C, 5:550°C, 6:600°C, 7:650°C, 8:700°C, 9:800°C, 10:900°C, 11:1000°C, 12:1100°C show the XRD analyses at that temperatures. (atg: antigorite, fo: forsterite, ens:enstatite)

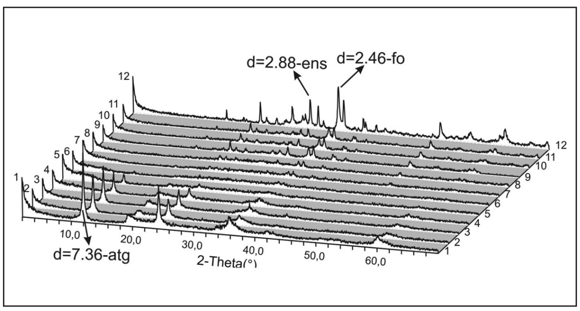

Figure 7- XRD block diagram of sample 29A. The numbers 1: Dried, 2:200°C, 3:400 °C, 4:500°C, 5:550°C, 6:600°C, 7:650°C, 8:700°C, 9:800°C, 10:900°C, 11:1000°C, 12:1100°C show the XRD analyses at that temperatures. (atg: antigorite, fo: forsterite, ens:enstatite)

Thermal reactions in antigorite are developed in parallel to temperature increase and at a temperature of 100°C, free water in samples and absorbed water on the surface are lost due to dehydration. Examination of DTA and TG curves reveals that there is such a change in all samples at 100°C. Endothermic peaks at 210°C and 380°C for samples 6, 11B and 11S are attributed to clay minerals and brucite within the serpentine minerals that could not be detected in X-ray analyses due to their lower abundance. According to McKenzie (1970), montmorillonite group clay minerals give two endothermic peaks at 125-260°C and 625-750°C temperature range while brucite gives only one endothermic peak at 356-455°C temperature range. Examination of DTA diagrams yields endothermic peaks of clay minerals at about 210°C and of brucite at 380°C. The endothermic peak expected for the clay minerals at temperatures between 625°C and 750°C coincides with that of antigorite.

Antigorite peaks in all samples are observed as endothermic at 690°C and exothermic at 810°C. As shown in DTA curves, dehydroxilation was occurred at about 550°C to 690°C. As determined by the X-ray analyses, reactions associated with dehydroxilation resulted in formation of forsterite from antigorite ($Mg_3Si_2O_5(OH)_4$) at temperatures above 550°C and enstatite ($MgSiO_3$) at temperatures above 650°C. In XRD analyses, antigorite was lastly found at 650°C and considering the results of both DTA and XRD analyses reveal that antigorite can be stable at temperatures mostly 650°C-690°C. At temperatures higher than 690°C which defines the upper stability limit of antigorite, forsterite and enstatite continued to crystallize and this new mineral formation was recorded on DTA diagrams as two exothermic peaks at around 810°C. As a result, all these mineralogical changes determined with the XRD studies are found to be conformable with DTA analyses.

According to results of TG analyses, 11% mass decrease was found in all samples (Figures 9, 10, 11, 12 and 13).

RESULTS AND DISCUSSION

Water plays an important role in formation of various geological processes. Serpentinized oceanic lithosphere is believed to be the source of water that is effective in development of subduction-related calc-alkaline arc volcanism (Ulmer and Trommsdorf, 1995). In this respect, stabilities of serpentine minerals are very important for determination of mode of occurrence and depth of these processes. Therefore, thermal reactions of serpentine minerals and mineralogi-

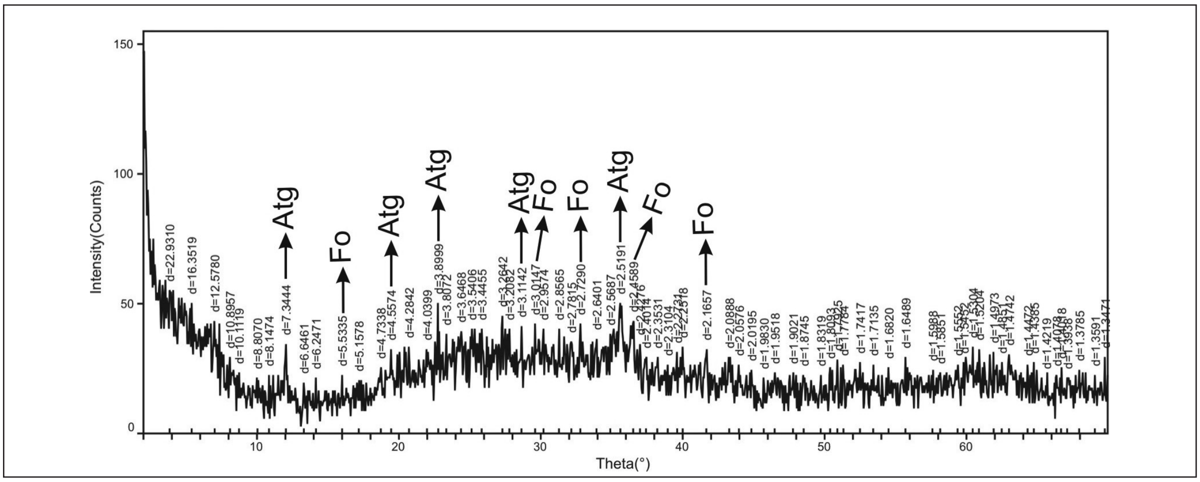

Figure 8- The amorphousization at 600°C (Sample 29A, Atg: antigorite, Fo: Forsterite)

cal changes occurring during these events are primarily important.

A number of experimental works were conducted for determination of stabilities of serpentine minerals (Wunder and Schreyer, 1997) (Figure 14). In this study, serpentine minerals were subjected to thermal treatment to investigate the upper stability limits of these minerals and resulting new mineral paragenesis. In this study antigorite was used which is more stable in comparison to other serpentine minerals and, it was determined that different mineral paragenesis were obtained under various pressure and temperature conditions. Wunder and Schreyer (1997) compiled results of previous works (Figure 14) and determined the stability limits of antigorite and following reactions were stated to be taken place during dehydroxilation of antigorite:

Antigorite ↔ talc + forsterite + H_2O
(15 kbar, 640 °C) (1)

Antigorite ↔ forsterite + clinoenstatite + H_2O
(77 kbar, 680 °C) (2)

Antigorite ↔ clinoenstatite + brucite + H_20
(>77 kbar, >680 °C) (3)

According to Ulmer and Trommsdorf (1995), forsterite is formed by the reaction between antigorite and brucite (if available) at low temperatures (antigorite + brucite ↔ forsterite + H_2O) and about 3.5% H_2O is released. With further increase of pressure and temperature, brucite is completely removed and as shown in reactions 4 and 5, different minerals are formed. Diagram showing the stability limits of antigorite is shown in figure 15.

Antigorite ↔ forsterite + talc + H_2O
(low pressure) (4)

Antigorite ↔ forsterite + enstatite + H_2O
(high pressure) (5)

In this study, antigorite minerals were subjected to thermal treatment under constant pressure (laboratory conditions). Mineralogical changes occurred during this process were determined with XRD and DTA-TG analyses and stability limits of antigorite were determined under constant atmosphere pressure.

All the samples are mostly composed of antigorite. However, on the basis of DTA analyses results, little amount of montmorillonite and brucite were also found in samples 6,11B and 11S.

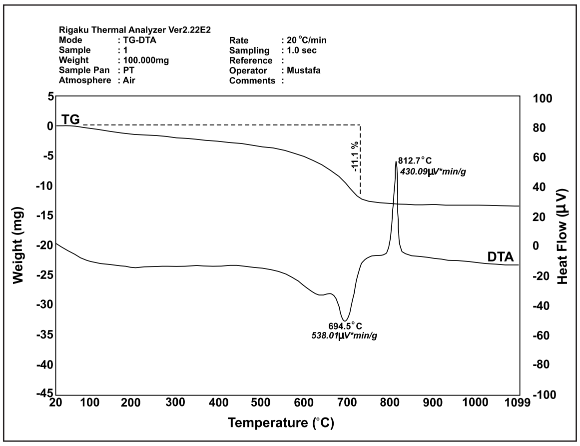

Figure 9- DTA-TG curves of sample 1

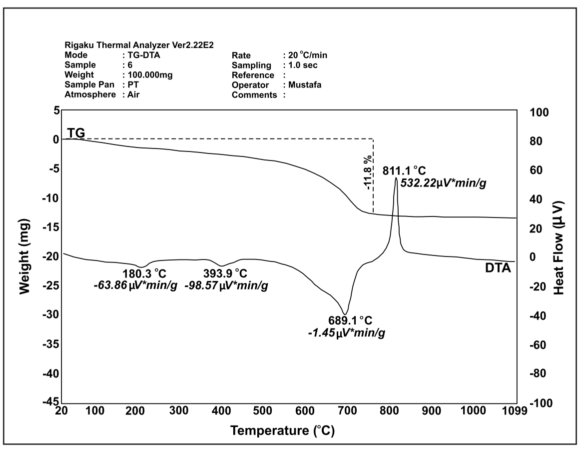

Figure 10- DTA-TG curves of sample 6

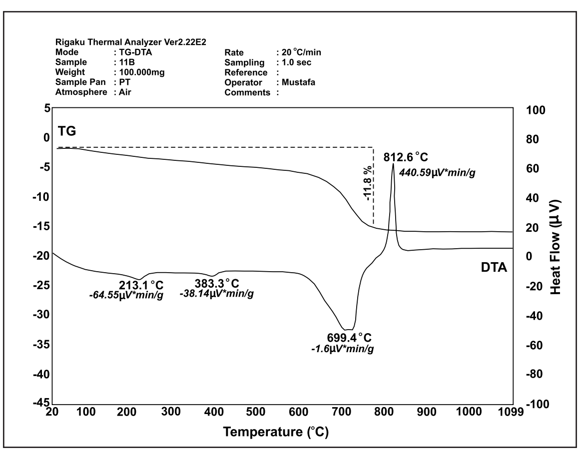

Figure 11- DTA-TG curves of sample 11B

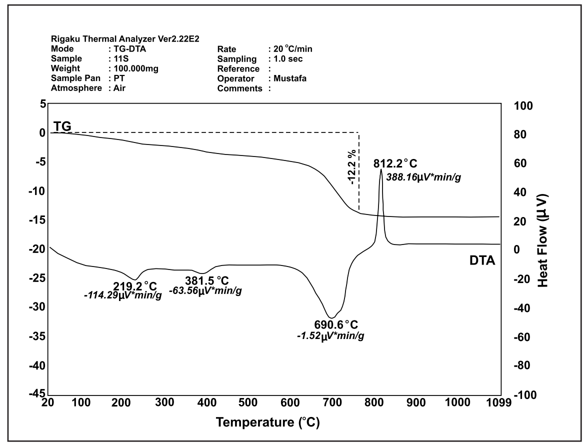

Figure 12- DTA-TG curves of sample 11S

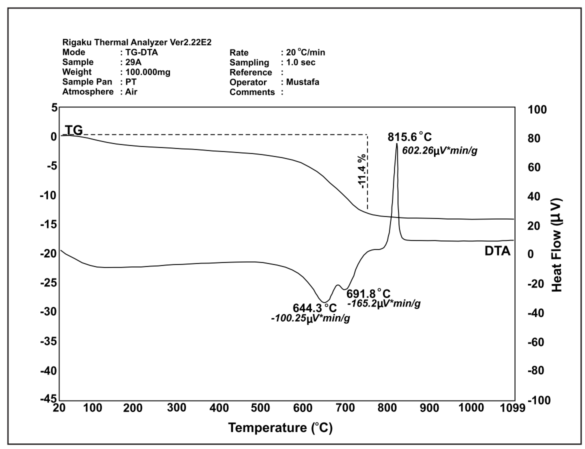

Figure 13- DTA-TG curves of sample 29A

Considering the results of XRD analyses, antigorite can be stable until a temperature range of 650°C-690°C but its abundance decreases depending on reactions associated with temperature increase. Increase in antigorite abundance was determined from peak intensities on XRD charts. In analyses at temperatures above 550°C, the amount of antigorite was further decreased and forsterite started to be formed. Antigorit+forsterite association at temperatures above 550°C was continued until 650-690°C and antigorite was removed at higher temperatures. Results of DTA analyses which were conformable with X-ray determinations indicated that

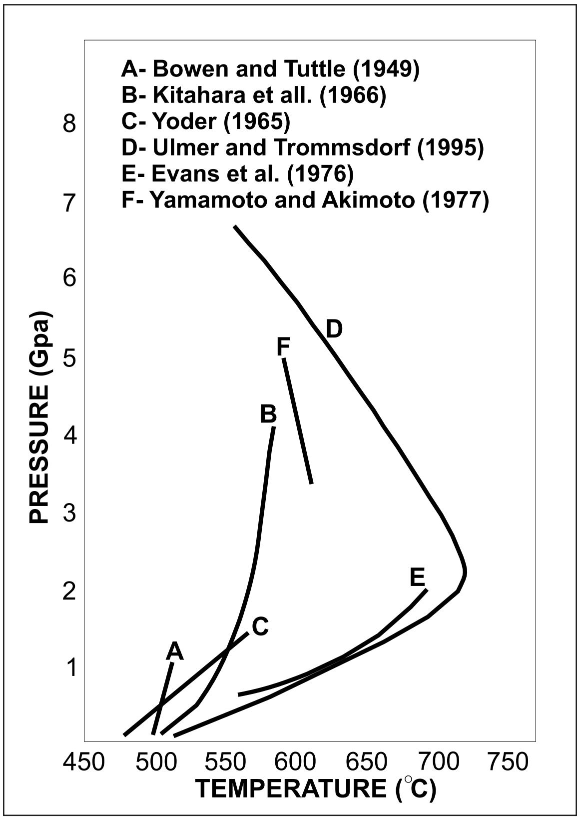

Figure 14- Previous experimental study of stability limit of serpentine minerals (Wunder ve Schreyer, 1997)

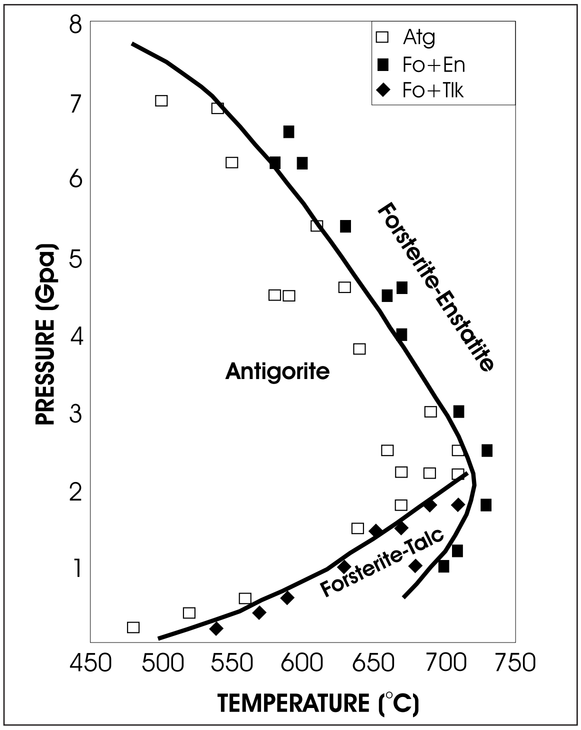

Figure 15- Stability limit of antigorite (Ulmer ve Trommsdorf, 1995)

antigorite dehydroxilation reactions were started at 550°C and continued until 690°C.

X-ray analyses conducted at temperatures above 650°C reveal that in addition to forsterite, enstatite also occurs and abundance of both minerals increases with temperature increase. According to results of DTA analyses , endothermic peak shown at 810°C following the dehydroxilation at 690°C correspond to formation of forsterite and enstatite. However, XRD results imply that enstatite forms in a later stage than forsterite (after 650°C).

In samples 6, 11B and 11S, antigorite is accompanied by little amount of brucite and, as a result of their reaction forsterite is formed and some amount of water is released. In brucite-free samples (e.g., 1 and 29A), forsterite is formed from antigorite and similarly some amount of water is released. Examination of XRD analyses conducted at 550°C, 600°C and 650°C reveal the presence of amorphousization that is observed in all samples but particularly significant in samples nos. 1 and 29A. The amorphousization is thought to be derived from silica that is released during the thermal reactions. Considering this, following reactions were first occurred during thermal reaction of antigorite:

$$\text{Antigorite + brucite} \leftrightarrow \text{forsterite} + H_2O + SiO_2 \quad (550°C\text{-}650°C) \ (6)$$

$$\text{Antigorite} \leftrightarrow \text{forsterite} + H_2O + SiO_2 \quad (550°C\text{-}650°C) \ (7)$$

According to results of XRD analyses, during thermal reaction of antigorite enstatite is formed following the forsterite (650°C) and following reactions took place during the formation of enstatite:

Antigorite ↔ forsterite + enstatite + H_2O (650°C-690°C) (8)

Forsterite + SiO_2 ↔ enstatite (>650°C) (9)

As a result of dehydroxilation reactions occurring between 550-690°C it was found that first forsterite forms antigorite. The amorphousization that was determined by XRD analyses and observed in between dehydration and crystallization reactions may indicate that in addition to water some amount of silica was also released during these reactions. At temperature above 650°C forsterite is accompanied by enstatite.

Considering all these data, antigorite was found to be stable up to temperatures about 650 to 690°C. During dehydroxilation, antigorite-forsterite association was observed at temperatures above 550°C and enstatite starts to crystallize above 650°C. Forsterite+enstatite association is dominant by 690°C.

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