

STRUCTURAL EVOLUTION OF  
THE BABADAG FAULT ZONE IN  
DENIZLI GRABEN,  
SOUTH WESTERN TURKEY

BY

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2002

Submitted to the faculty of the  
Graduate College of the  
Oklahoma State University  
in partial fulfillment of  
the requirements for  
the Degree of  
MASTER OF SCIENCE  
December 2005

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## ACKNOWLEDGEMENTS

Firstly, I express my deepest gratitude to my hero, Osman Ozdemir, whom I have always admired and want to be like. I thank my mother, Merym Meral Ozdemir, and my dearest sister, Nadire Ozdemir for their constant love. My family has been there and always supported me during my life.

I thank my thesis advisor, Dr. Ibrahim Cemen for his great help and support. Without his enthusiasm for geology and guidance, I would not have made it this far. I also thank the rest of my committee, Dr. Elizabeth Catlos for her unwavering support and encouragement but the most her friendship throughout my graduate career and Dr. Surinder Sahai for his constant help and consideration with my project. I also owe my thanks to Dr. Mete Hancer from Pamukkale University, Turkey for helping during my fieldwork.

I would like to thank all the students and my friends of the Geology department. My officemate, Wesley Kruger, Emre Diniz, Cenk Ozerdem and Ali Jaffri have my deepest appreciation for their valuable reviews of this thesis report and for their understanding and support.

I also thank my roommate, Sibel Irmak for her endless friendship, being more than a sister and sharing a full two-year period without any argument. Dr. Salim Hiziroglu shares a deep respect in my heart for guiding me to choose University of British Columbia for doing my PhD and also for his great friendship.

I would like to extend my appreciation to all members of Latin Dancing and Cultural Club who changed my life in the United States in an incredible way.

Finally I extend my deepest gratitude towards Venugopalan Anantharamakumar for his unique friendship and the most for believing in me. He was my refuge during hard times. I am forever grateful to him.

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# **CHAPTER I**

## **INTRODUCTION**

### **1.1 Overview**

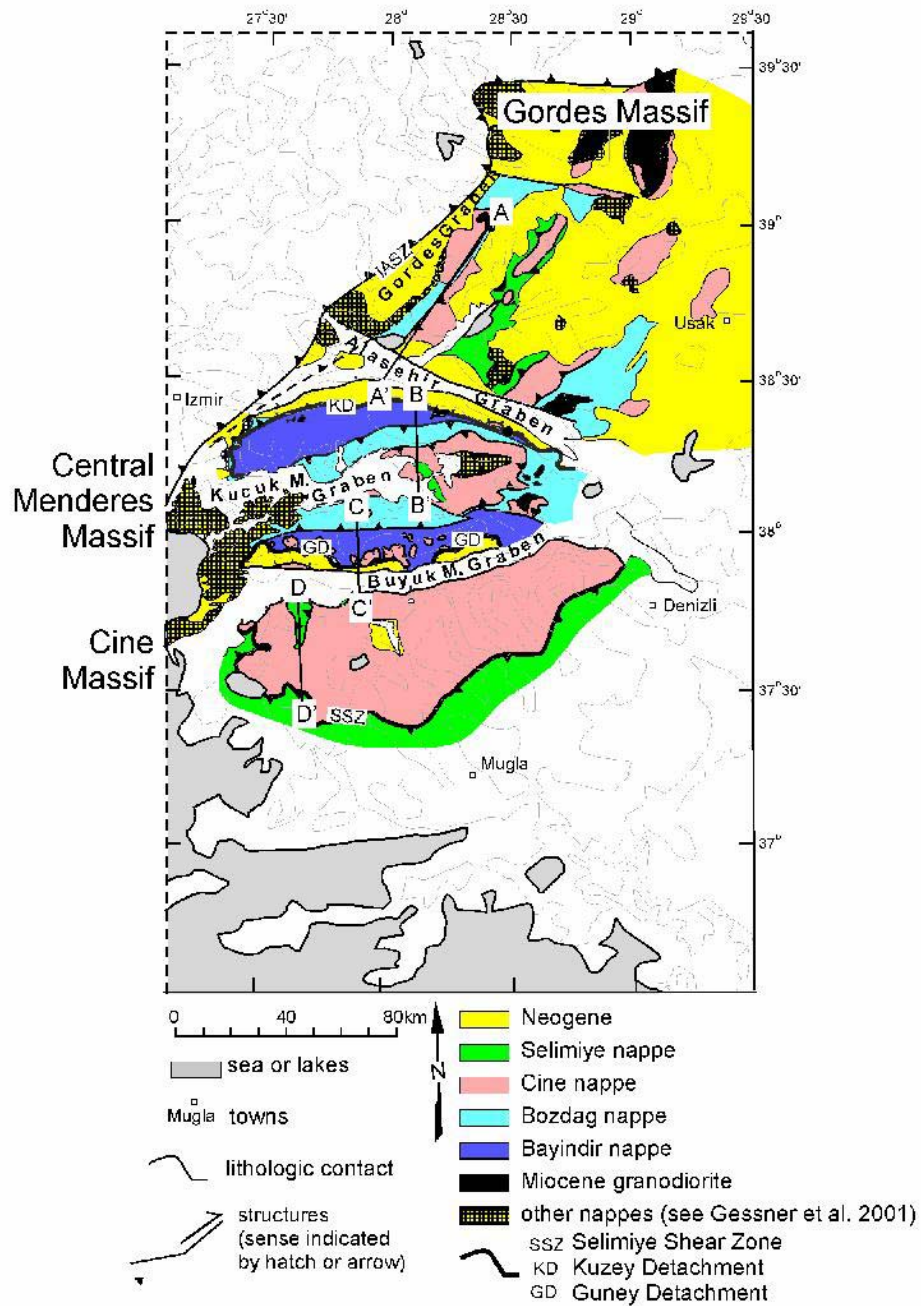
Western Anatolia is part of the Alpine-Himalayan belt. It has experienced post-collisional extension since Late Oligocene (Catlos and Cemen 2005, Gursoy et al., 2003). Western Anatolia is part of the Aegean Extended terrane where several metamorphic core complexes developed during the Cenozoic extension. These metamorphic core complexes are: Rhodope Massif, Kazdag Massif, Cycladic Massif, Crete Massif and Menderes Massif (Figure 1). The largest of these massifs is the Menderes Massif. The massif is divided into northern (Gordes), central and southern (Cine) submassifs by two E-W trending grabens (Hetzl et al., 1995) (Figure 2), the Alasehir, and Buyuk Menderes Graben. The north-dipping Alasehir Detachment surface and south-dipping Buyuk Menderes Detachment surface exhumed in Cenozoic. The Denizli Basin, located in the area where two major grabens of the region join together (Figure 3), is significant for our understanding of extensional tectonics affecting western Turkey.

## 1.2 Objective

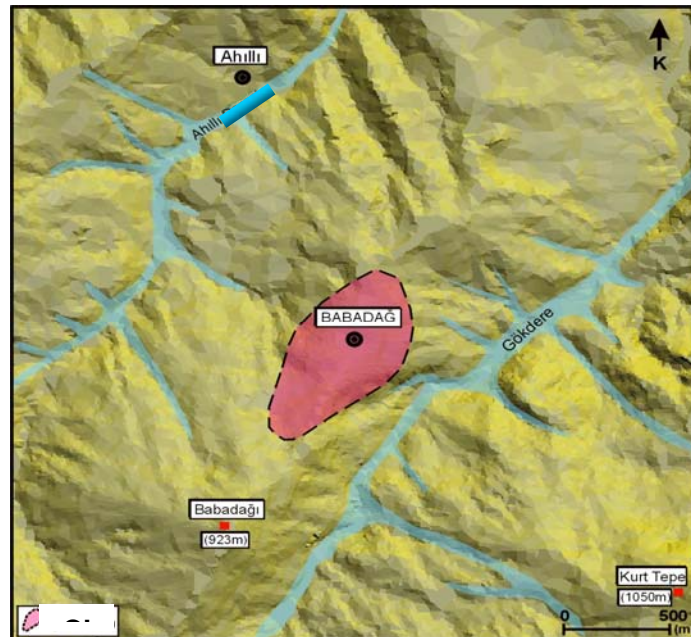
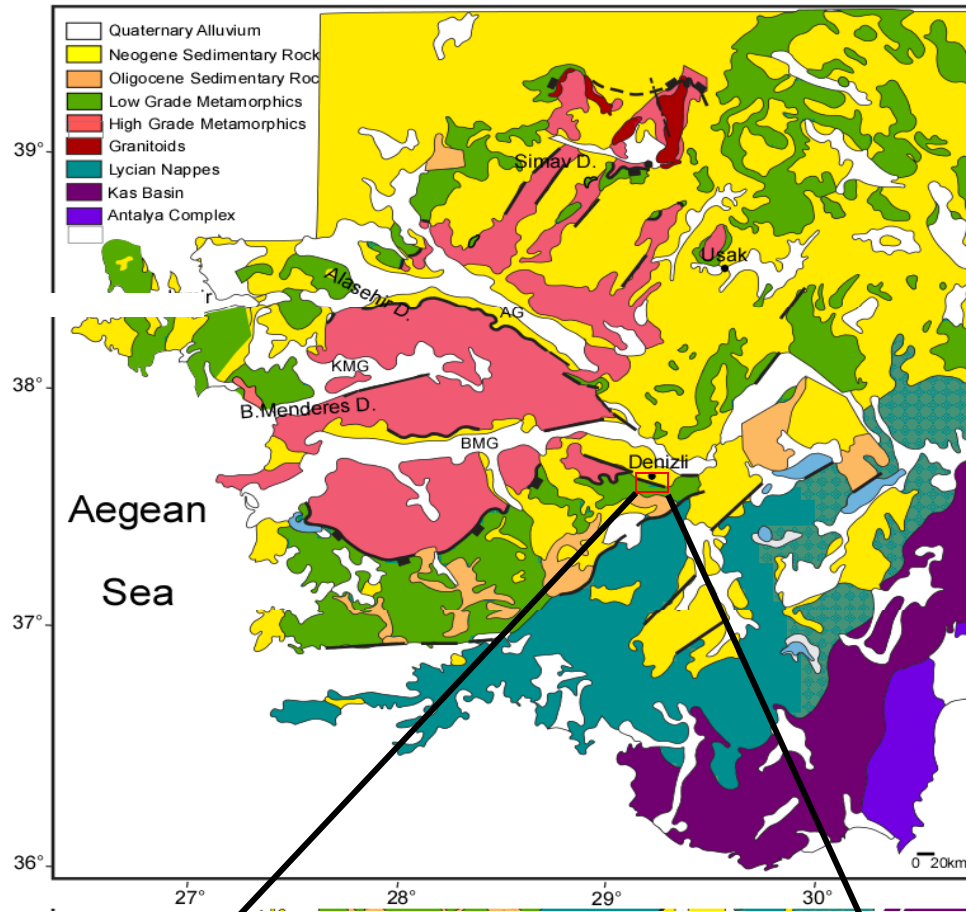
The main goal of this thesis is to present the geometry and structural evolution of the Babadag fault zone, which is located at the southern margin of the Denizli Basin. The fault zone separates metamorphic rocks in its footwall from sedimentary rocks in its hanging wall (Figure 4).



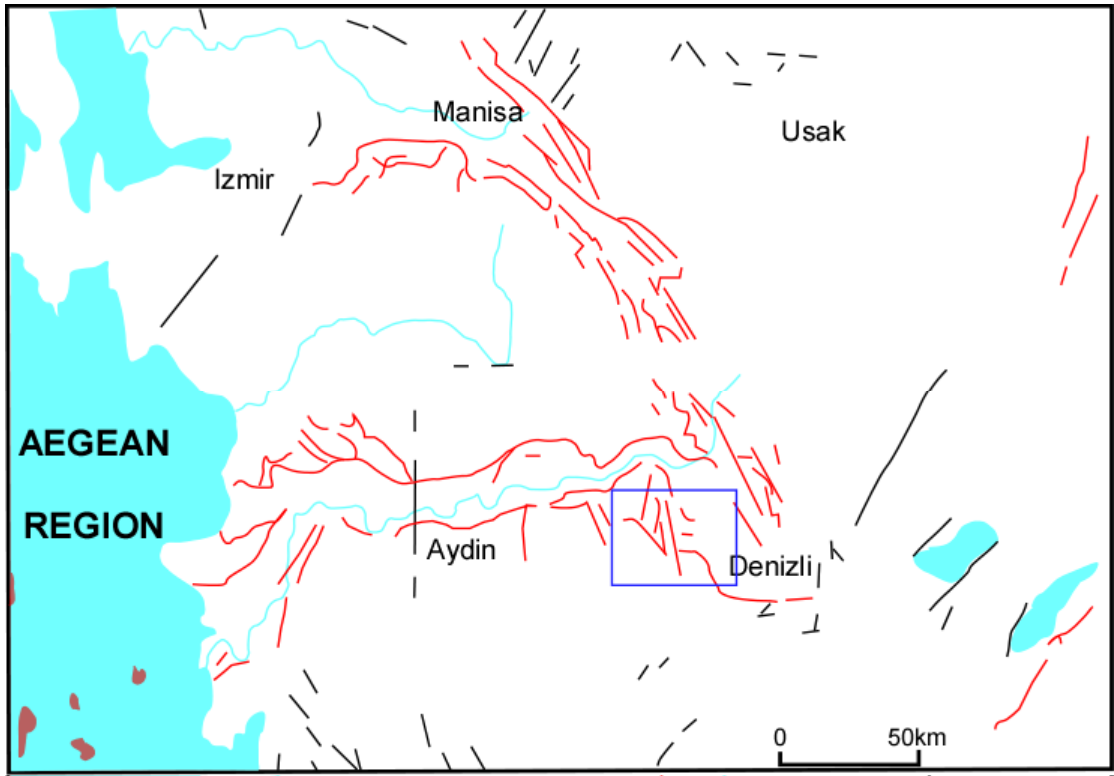
*Figure 1: Generalized map of Western Anatolia showing the metamorphic core complexes (Catlos and Cemen, 2005)*



**Figure 2: Geologic map of the Menderes Massif (Gessner et al., 2001). The Massif is divided into three submassifs by the the Alasehir, and Buyuk Menderes Graben.**



**Figure 3: Denizli and adjacent areas ([after Hetzel et al. (1995a)] b) Topographic map of the study area produced from digital elevation model.**



*Figure 4: Maps show the location of the Babadag Fault (after Hancer, 2003).*



### **1.3 Methodology**

The following field-oriented investigations were conducted:

- 1) The existing geological maps of the study area were field-checked and a 1/10,000 scale geologic map was compiled in summer 2004 (Plate I). Available arial photos of the area were used to identify structural features to provide reconnaissance mapping. These features were then field checked and mapped in detail. The Global Positioning System (GPS) locations, including elevation, latitude and longitude values of all observation points were recorded.
- 2) 45 oriented and random samples of fault rocks were collected and 25 oriented samples were prepared into thin sections (Plate I). Two thin sections were produced from each samples, one parallel to the lineation and the other perpendicular. These samples were prepared to examine microstructures in terms of their ductile and brittle shear sense indicators using regular petrographic microscope.
- 3) Cross-sections were drawn to provide a better understanding of the geometry and structural characteristic of the Babadag fault zone, and its relationship with the sedimentary rocks in its hanging wall.

## 1.4 Geologic Overview

The Menderes Massif is a recognized metamorphic core complex, based on the geological studies that were conducted in the region in the early 1990s (Bozkurt and Park 1993). The massif covers approximately 40,000 km<sup>2</sup> and displays a complex interior structure and variation in its lithology. The core of the massif is represented by high grade gneissic and schistic units where cover formations surrounding the core are represented by low grade rocks such as the greenschist facies schist, marble, phyllite and recrystallized limestone (Sengor et al., 1984).

Earliest geological studies in Western Anatolia were conducted in the second part of the 19<sup>th</sup> century. The first detailed report was written by De Tchihatscheff (1867), who recognized the uplifted Neogene basins in the region and described their fossiliferous brackish-water marl sediments.

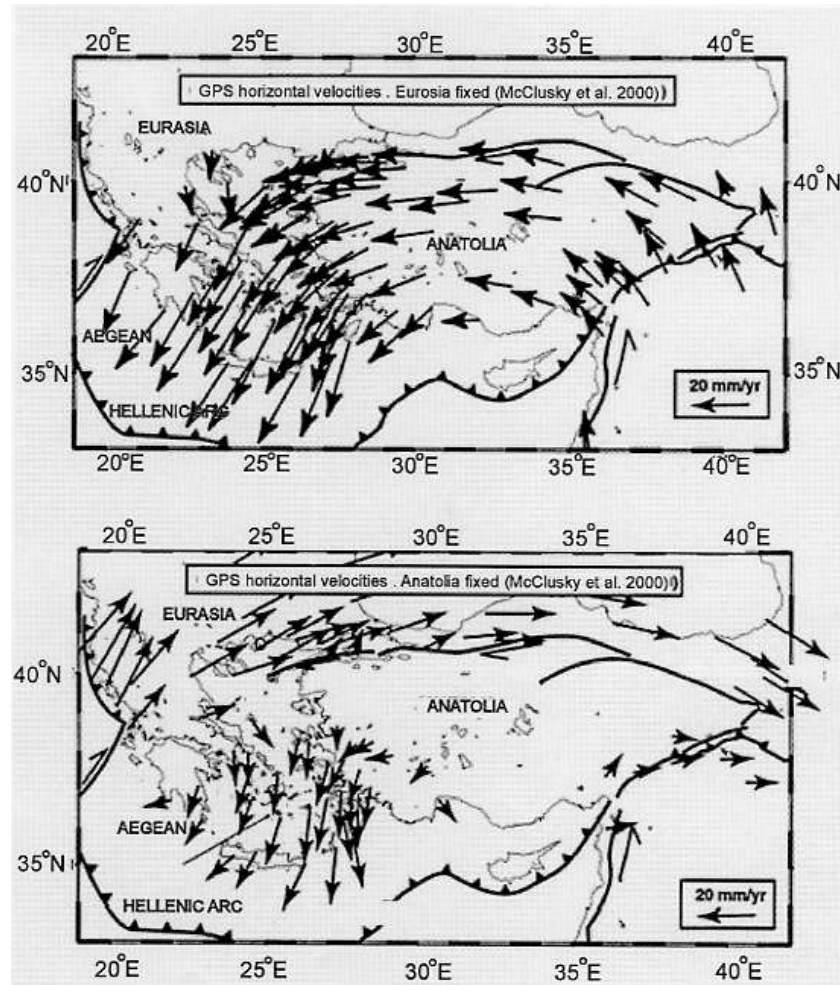
Holzer (1953) examined the effects of Alpine Orogeny on Menderes Massive and mapped Buldan-Cokenez, Dag, Cal, Civril-Karahalli and prepared 1/100000 scale geological map. The pre-Neogene metamorphic rocks are mapped into three units; marble, schist and submarble.

Although the extensional origin of western Anatolia is accepted, the timing and origin remain controversial. Four major models that were proposed for the extensional tectonics in Western Anatolia:

- 1) The Tectonic Escape or Lateral Extrusion Model: Dewey and Sengor (1979), Cemen et al., (1999), Yılmaz et al., (2000). This model suggests that the collision of Arabia and Eurasia across the Bitlis suture zone in southeast Turkey caused the tectonic escape of the Anatolian plate, which moved westward, and in turn caused the extension in western Turkey. As a result of the collision of Arabia and Eurasia during late Serravalian time (c.12 Ma), dextral North Anatolian and sinistral East Anatolian transform faults formed (Dewey and Sengor, 1979, 1982, 1987). Thus, the west-southwestward tectonic escape of the Anatolian plate produced an extensional tectonic regime leading to the development of the horst-graben system in western Turkey. The timing of these tectonic events in western Turkey is suggested to be Late Miocene (Sengor, 1982, 1987; Yılmaz, 1989; Savascin et al., 1990). Cemen et al (1993) and (1999) suggested a modified version of the Tectonic Escape Model. They proposed that lateral extrusion of Central Anatolia due to the collision of Arabia and Eurasia caused the formation of the Cenozoic extension in western Turkey.
- 2) Back-Arc Spreading/Subduction Roll Back Model: McKenzie [(1978) Le Pichon and Angellier (1979), Jackson and McKenzie (1988), Jolivet et al., (1999), Jolivet and Faccenna (2000), Okay and Satir (2000)]. This model suggests that south-southwestward migration of the Hellenic trench system caused the Cenozoic extension in the region. Eastern Mediterranean oceanic crust is rapidly downgoing beneath the Aegean,

and the roll-back motion of this downgoing slab causes severe extension within the upper plate. Commencement of roll-back spreading begun by the Late Serravallian – Early Tortonian (12 – 11Ma) (Meulenkamp et al., 1988).

- 3) Orogenic Collapse Model: (Dewey, 1978; Seyitoglu and Scott, 1991). This model that suggests that the Cenozoic north-south extensional tectonics and related sedimentary basin development in western Turkey started in late Oligocene – Early Miocene and is related to the spreading and thinning of the crust which occurred after the cessation of earlier Alpine shortening. The earlier Alpine compressional regime and related north-south shortening stopped by the end of the Oligocene. The compressional mountain ranges were gravitationally unstable and spread outward under their own weight.
- 4) Geodynamic Model: Doglioni, C. et al.(2002) This model suggests that the Aegean-Western Anatolian rift was opened due to the faster southwestward progress of Greece over Africa, with respect to Cyprus-Anatolia over Africa. Within the realm of this idea, GPS studies provide a detailed view of the present day plate motions in the eastern Mediterranean (McClusky, 2000), and confirmed that the Anatolian block moves westward through the North Anatolian Fault (Figure 5). The westward extrusion of the Anatolian block is defined by counterclockwise rotation (Le Pichon and Angelier, 1979; McClusky et al. 2000).



*Figure 5. GPS velocity field after McClusky et al. (2000) relative to Eurasia (above) and to Anatolia (below).*

The regional guide by Pamir & Erentoz (1974) provides the basis for the present paper which examines relationships between tilting of beds and dips of active normal faults within Denizli basin, and estimates the extension across it. This paper describes field observations that quantify this extension and its partitioning between individual normal faults.

Kocyigit (1984) focuses on the tectonic events which started in the Late Tortonien and the structures formed due to these events. These events were caused by changes in the tectonic regime. He proposed that terrigenous Neogene sediments were formed by the block faulting which was formed in the subsidence in Middle Anatolia and southernwest of Turkey.

Westaway (1993) researched the dips of Neogene sediments in the Denizli basin and noticed that the area shows a various range of dips. He also suggested that some small faults in the Denizli Area have steeper dip, approximately  $80^{\circ}$ , whereas the others have maximum dip of  $40^{\circ}$ .

Menderes Massif was classified into three regions: northern (Gordes), central and southern (Cine) submassifs separated by two E–W trending grabens and suggested the initiation of the Cenezoic extensions to be 20Ma (Early Miocene) (Hetzel, 1995).

The Menderes Massif is a part of the western Anatolia and contains high-grade metamorphic rocks exposed at the surface. Akgun and Sozbilir (1999) carried out a detailed stratigraphic study of the Denizli Basin and found that the basin has a transgressive sequence resting on the pre-oligocene. The basement which is overlain by the alluvial fan deposits of Late Miocene–Pliocene age.

Yilmaz et al.(2000) considered the geology of the grabens and suggested five major stages in the geological evolution of western Anatolia: (1) a pre- graben stage which corresponds to the Late Cretaceous, pre-Miocene time period; (2) an east–west extensional stage in which some of the grabens formed during Early Miocene while others formed during the Late Miocene time; (3) an earlier stage of north – south extension which began during Late Miocene time (4) an interrupting stage of north –

south extension; in this stage continuation of N-S extension ceased at the end of Late Miocene-Early Pliocene(?); 5) a later stage of north – south extension; following the development of the erosional surface, N-S extension was reactivated and E-W grabens began to form.

The study of Cemen, Tekeli, & Seyitoglu (2000) provide data on the discussion of the development of N-trending and E-W trending basins in western Turkey. They divided the fault system into two parts based on the orientation and structural relations. The first fault system is responsible for the graben fill where the second system is responsible for the deposition of the sedimentary units which were developed in the hangingwall of the first system.

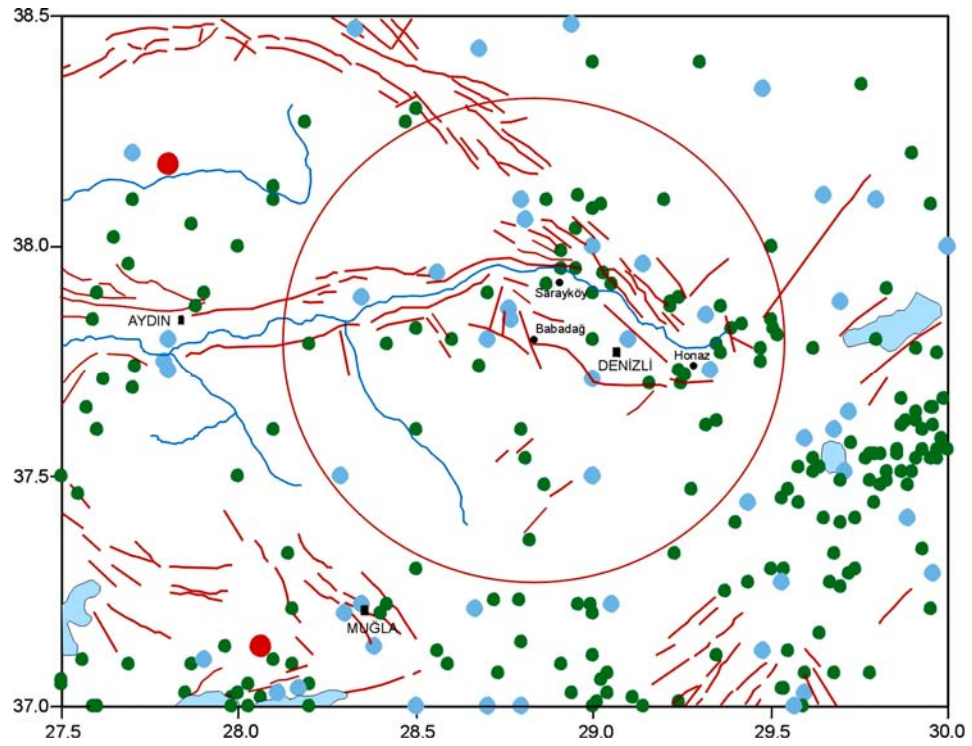
Gessner et al. (2001) studied the area of Central Menderes Massif and pointed out that the kinematic indicators along the Kuzey detachment show top to the north shear sense while they show top to the south shear sense along the Guney detachment. They proposed a bivergent-rolling hinge model for the Central Menderes Massif Core Complex due to the existence of these two oppositely dipping detachments with opposite senses of shear.

## **1.5 Neogene evolution in Denizli Basin**

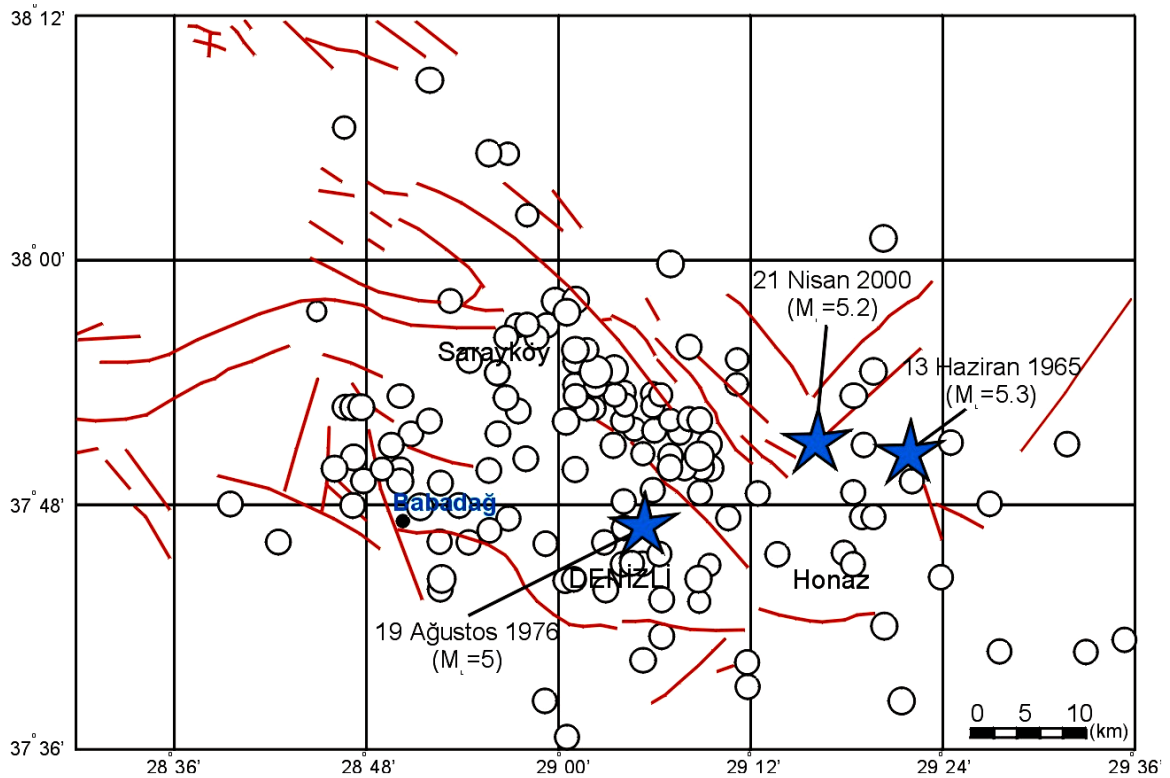
Denizli basin is located between N38°00'-37°36' and E28°36'-29°24', a region that has experienced moderate seismic activity. (Magnitude 4-6). Figures 6a, 6b, and 6c illustrate earth quake history.







*Figure 6b. Map showing the active faults and epicenters of the earthquakes occurred in between 1900 and 2002 in Babadağ and its adjacent areas (YELİZ DAD, modified from the data given by DAD 2003).*



**Figure 6c. Map showing the earthquakes occurred in the first 10 months of 2002 ( $2.8 < M < 5.2$ ) and epicenters of these destructive earthquakes for Denizli and adjacent areas (Ozalaybey et al., 2000).**

Westaway (1993) reviewed his studies on Denizli basin and suggested the Neogene evolution of the basin as below.

The history of the Denizli Basin can be broken down into a series of transgressions and regressions that began in the Early Miocene and continued into the Late Miocene. The earliest highstand deposits in the area are red conglomerates and marls. Extension began in the Middle Miocene and more highstand sediments were deposited in the basin. These include a transition from marine limestones to brackish water limestones and beach deposits. Marine sediments were deposited in an inlet of the sea that followed the hanging wall block of the Büyük Menderes Fault. In the Messinian

Time (7.1 – 5.3 Ma) an uplift episode began. Highstand deposits during this time are characterized by whitish yellow marls. This period of uplift was followed by second episode of extension and the depositional system changed from marine to terrestrial. A fluvial system developed in the latest Miocene.

## **CHAPTER II**

### **LITERATURE REVIEW**

#### **2.1 Introduction**

The main aspect of this thesis is to investigate the shear sense indicators that were found along the footwall rocks of the Babadag fault zone. These shear sense indicators formed during the extensional unroofing of the eastern end of the central Menderes Massif where the Babadag fault zone controls the sedimentation along the southern margin of the Denizli Basin. This chapter a brief summary of the structural features associated with the metamorphic core complexes and shear sense indicators will be discussed.

#### **2.2 Metamorphic Core Complexes**

Metamorphic core complexes (MCC) were first described in the Basin and Range Province in North America (Crittenden et al., 1980). The basic rock units of a MCC are a metamorphic basement and a nonmetamorphic cover. Between these two rock units there is a discontinuity or detachment which consists of mylonitic fabric. These complexes usually bring brittlely deformed sedimentary rocks along low-angle faults onto ductilely deformed rocks.

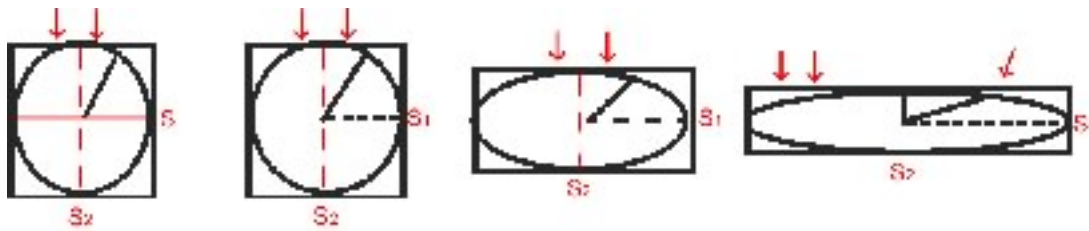
The formation of the MCC is thought to be the overthickening of the continental crust by compressional forces or intrusive igneous processes. The relaxation of the crust

which follows the cessation of the compressional forces produces extensions which then form normal faults and low-angle detachment faults which are important structures for the exhumation of ductilely deformed rocks (Wernike, 1982).

Two main shear models were proposed to define the development of MCC along shear zones:

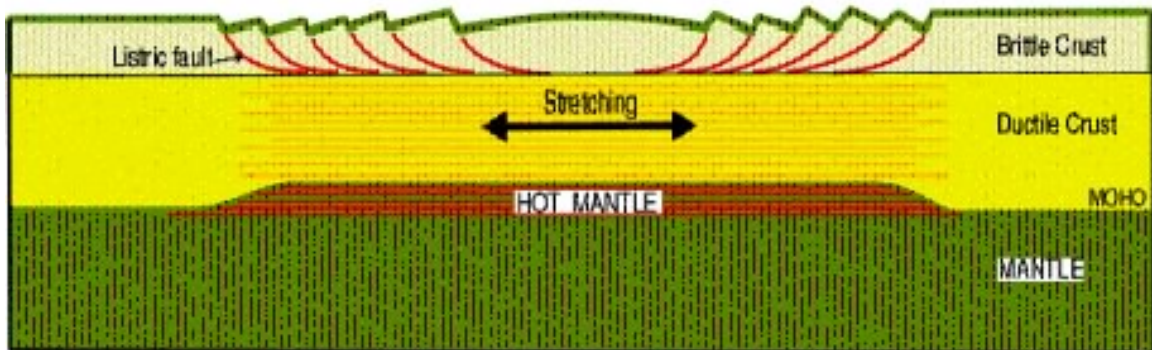
1) Pure Shear

Pure shear is in which the axes of the strain ellipsoid do not rotate and the incremental and finite strain ellipsoids are coaxial (Figure 7).



**Figure 7: The diagram represents strain ellipsoids for pure shear. There is not any rotation but compression and extension can be observed.**

McKenzie's model is one of pure shear in which the homogeneous lithosphere is stretched uniformly to form a symmetrical basin with faulting in the brittle crust and spread away from the center of the uplift toward the sides. Isostatic uplift causes a bowed up structure in the asthenosphere (McKenzie, 1978) (Figure 8).



**Figure8: McKenzie's pure shear model.**

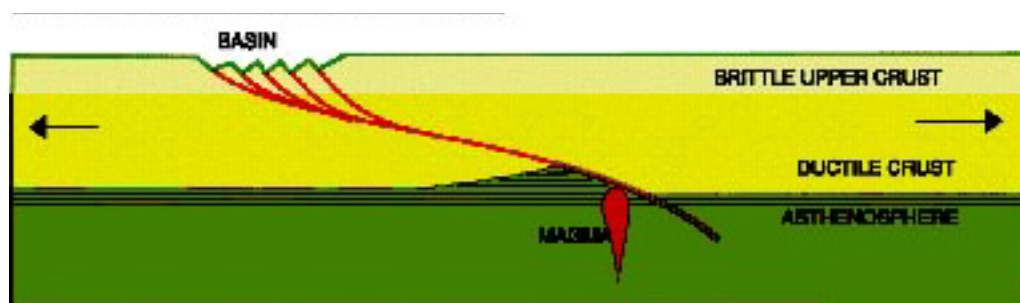
## 2) Simple Shear

Simple shear is a three-dimensional constant-volume strain (Twiss and Moores, 1992; Davis and Reynolds, 1996) (Figure 9). During simple shear, the axes of the strain ellipsoid rotate and are noncoaxial.



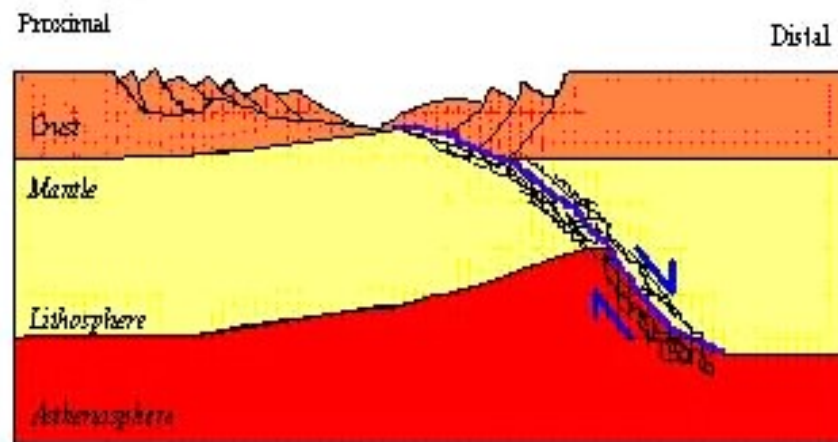
**Figure 9:** This diagram shows that simple shear involves rotation about a point.

There are two simple shear models proposed for extensional regimes. The first suggests deformation is concentrated along detachment surfaces that cross the lithosphere (Wernicke, 1981). This model suggests that when upper level hanging-wall rocks move downward they are thinned by faulting and rotation. Because the deformation on the hangingwall occurs at relatively shallow depths, it is characterized as brittle (Davis and Reynolds, 1996). The deeper level footwall rocks move upward toward the surface, deforming ductilely at depth where the ductile deformation is then overprinted by brittle deformation (Davis and Reynolds, 1996). The partial unroofing of the footwall rocks result in isostatic uplift in a domelike shape and this may modify the early formed normal faults (Davis and Reynolds, 1996) (Figure 10).



**Figure 10:** Wernicke (1981) simple model

The second suggests the 'proximal' area of the upper crust will undergo an initial subsidence but no lithospheric extension occurs (Buck, 1988). Lithospheric behavior at the 'distal' end of the shear zone may undergo an uplift. Buck et al (1988) found that the footwall of the shear zone experiences minor uplift from lateral heat of the asthenosphere that eventually sinks back to normal height after thermal cooling (Figure 11).



*Figure 11: Buck's (1988) simple model.*

In both the simple-shear and the pure-shear models, extensional thinning of crust and mantle lithosphere produces local upwelling of warm asthenosphere to replace thinned lithosphere (David Wilson et al., 2005)

Shear zones can be classified into 3 major groups based on deformation mechanisms and localization behavior varies with temperature and pressure:

- *Brittle shear zones:*

Brittle shear zones represent the shallow parts of the crust with a depth less than 5-10 km of the Earth's surface. The deformation is dominated by brittle mechanisms,

such as fracturing, ranging from a few centimeters to a few meters, and faulting which cause discontinuous margins (Davis and Reynolds, 1996).

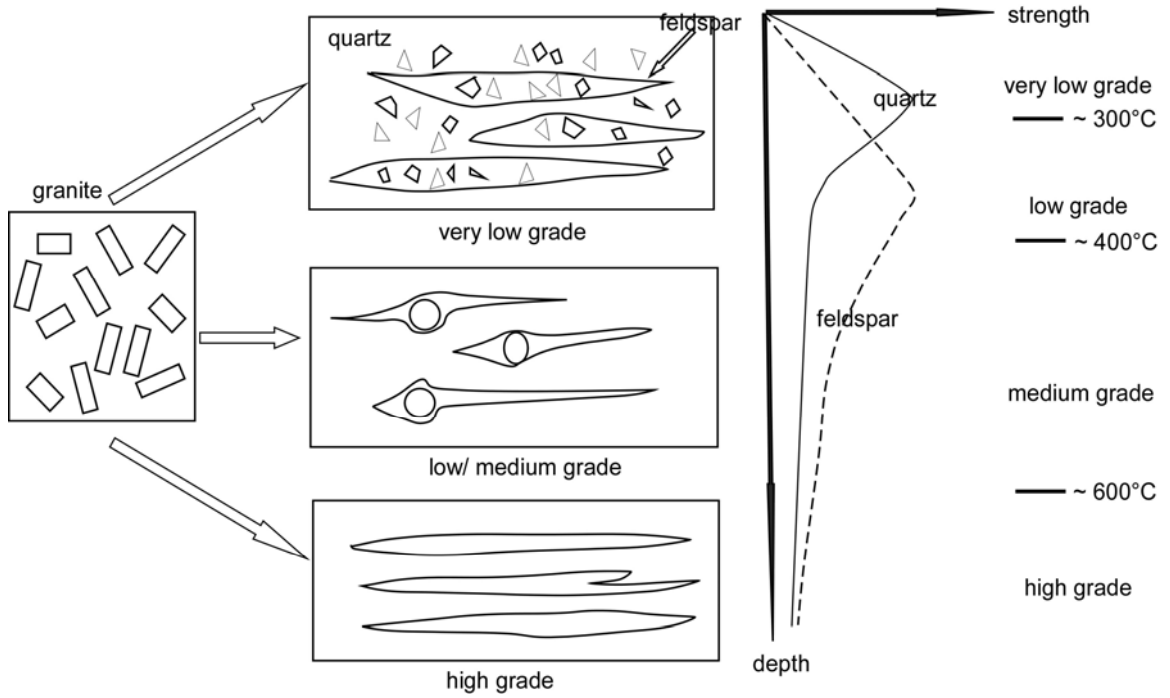
- *Brittle-ductile shear zones:*

Brittle-ductile transition shear zones are zones that contain the product of both brittle and ductile deformation mechanisms. They form when the physical conditions permit brittle and ductile deformation to occur together. This mixed zone indicates that physical conditions are unstable when the deformation is active (Davis and Reynolds, 1996).

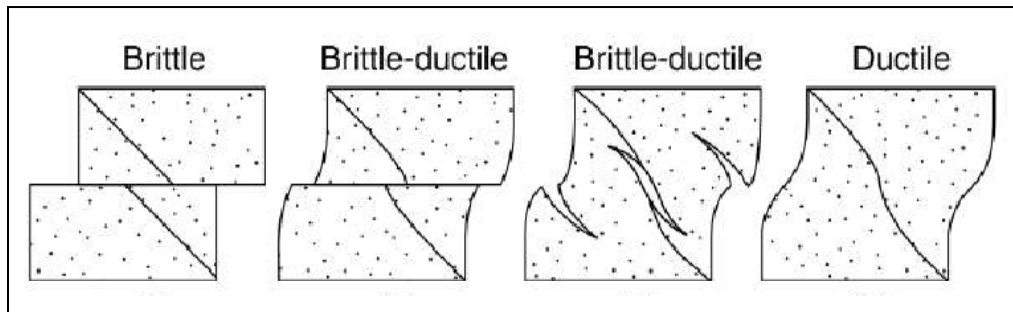
- *Ductile shear zones:*

Ductile shear zones can develop as a result of shearing (simple-shear strain) or "squeezing" (pure-shear strain). They represent the relatively deeper parts of the Earth's crust and are mostly generated in the middle-lower crust. Rocks below 10 km are subjected to relatively high temperatures and pressures which give rise ductile flow rather than by faulting and brittle fracture (Davis and Reynolds, 1996). Ductile shear zone's deformation causes continuous displacements.





**Figure 12: Changes in the deformation behavior of quartz-feldspar grains with depth. At right, a depth-strength graph with brittle (solid line) and ductile (dashed line) segments for quartz and feldspar is shown. (Modified from David and Reynolds, 1996)**



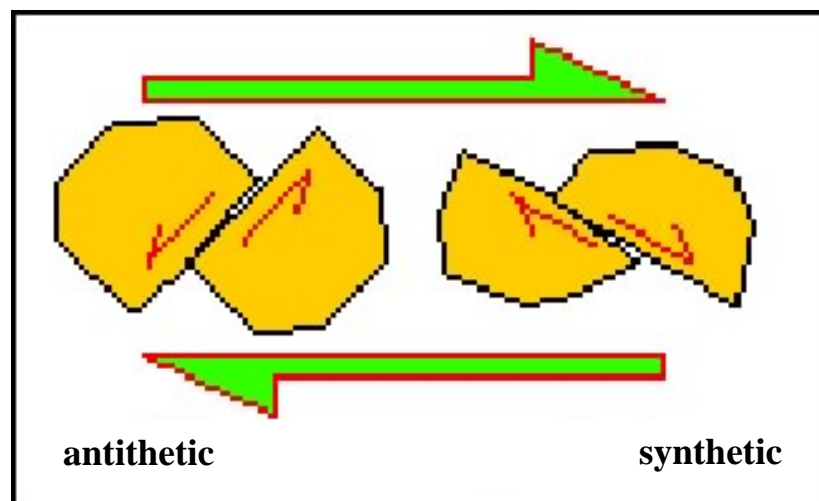
**Figure 13: Different behavior of a rock body subjected to different types of deformation.**

## 2.3 Shear Sense indicators

Metamorphic rocks contain evidence for simple shear deformation which indicates low-angle dip-slip movement. The reason as to why we should define the sense of shear on a simple shear zone is that only a noncoaxial deformation can create a shearing sense on a body while a coaxial deformation causes a thinning or a thickening. Then we can talk about which way  $\sigma_1$  is inclined with the respect to the zone. The following is a short summary of the most commonly observed shear sense indicators.

### 2.3.1 Shear Bands and S-C fabrics

Shear bands are thin zones of very high shear strain within the main shear zone (Davis and Reynolds, 1996). They can be either parallel or oblique to the main shear zone. A shear band is synthetic if it is inclined in the same direction as the sense of shear. It is considered as antithetic if inclined in the opposite direction (Figure 14).

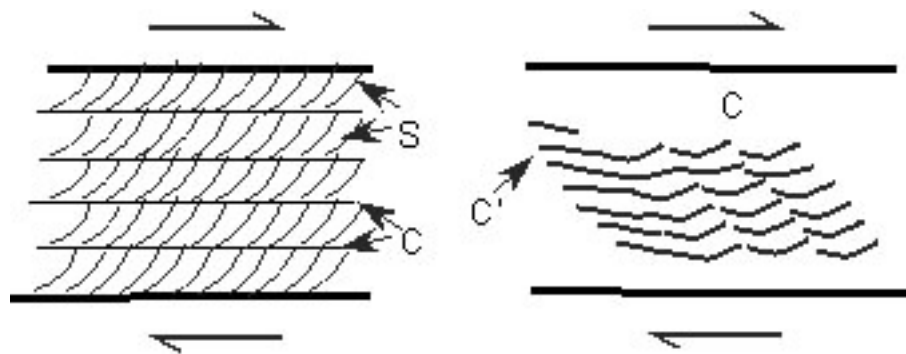


*Figure 14: A shear band is synthetic if it is inclined in the same direction as the main shear zone and antithetic if inclined in the opposite direction.*

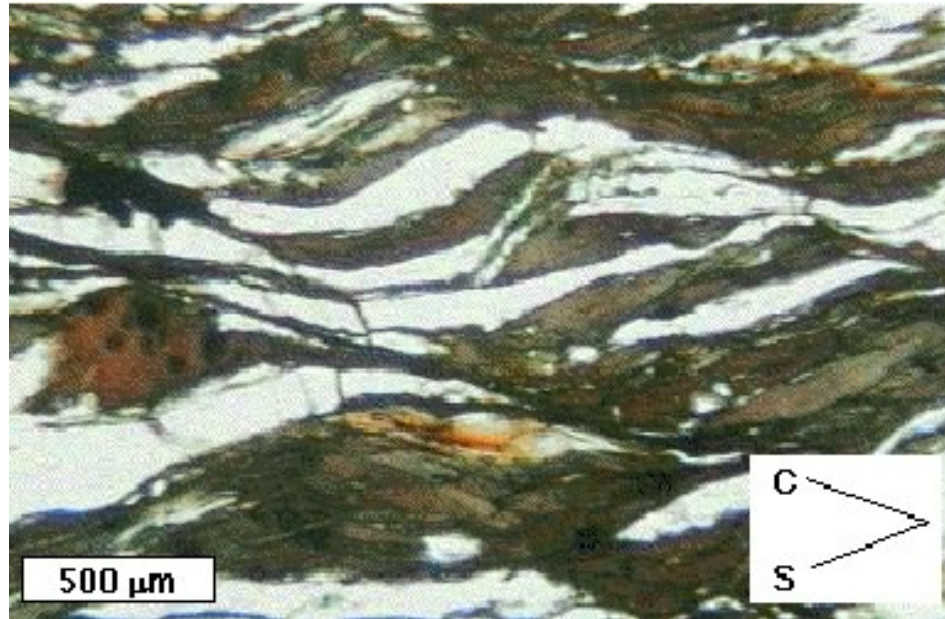
S-C (S refers to schistosity; C refers cisaillement, meaning shear in French) fabrics are among most useful sense-of-shear indicators in ductile shear zones (Hammer and Passchier, 1991). They consist of two sets of planes which are S-surfaces and C-surfaces. S- surfaces are the foliation planes whereas the C-surfaces indicate shear bands within zones of high shear strain. C- surfaces are mostly aligned parallel to the shear zone (Davis and Reynolds, 1996). Berthe et al., 1979 defined two different types of C-surfaces which are C-type and C'-type and stated

“C-type shear band cleavage consists of S planes, transected by planar distinct C-type shear bands or C planes. C-type shear bands are parallel to shear zone boundaries. C'-type shear band is oblique to shear zone boundaries and develops mainly in foliated mylonites where the shear bands fail to continue into more weakly foliated layers (Figure 15).

S-C fabric probably reflects inhomogeneous simple shear (Figure 16) and occurs from the earliest stage of mylonite generation, contrary to C'-type shear band cleavage (Berthe et al., 1979).



**Figure 15: Characteristic geometry of a C-S and C-C' structures in a dextral shear zone (right-lateral). (Berthé et al., 1979).**

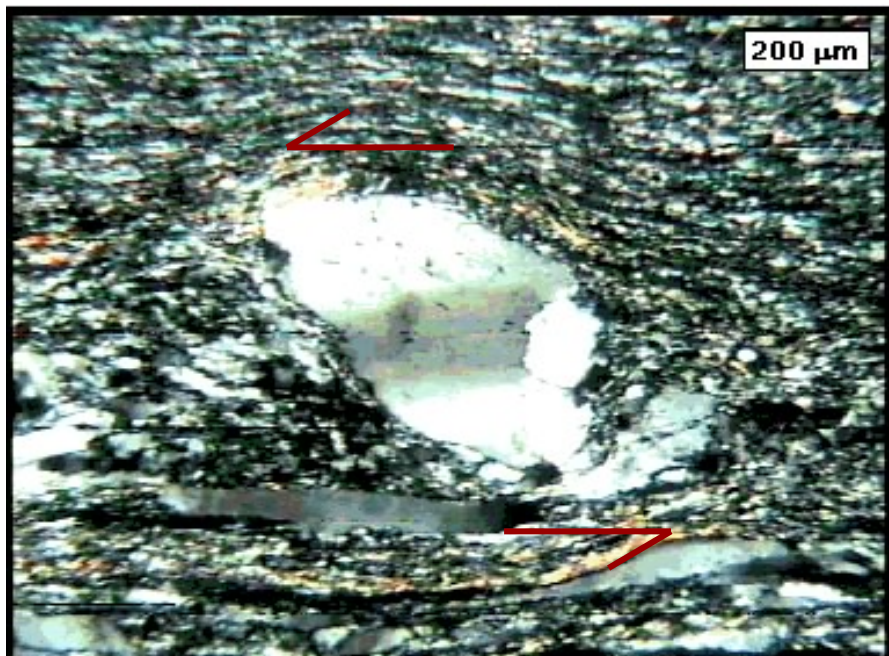


*Figure 16: S-C fabric in a diorite.*

S-C fabrics are considered as reliable shear sense indicators. They develop early in shear zone formation. C' fabrics are often late structures and may develop due to shear zone parallel shortening or stretching.

### 2.3.2 Mantled Porphyroclast / Winged objects

A porphyroclast refers to a rock or mineral fragment surrounded by a finer matrix that has been produced by cataclastic deformation. It may display a halo effect (mantle) of finer materials (Figure 17). The halo extends out into two winged structures that orient parallel to a mylonitic foliation; the wings often show a characteristic curvature. They are believed to be produced by crystalplastic deformation where the crystal lattice shifts to accommodate strain. Shear sense is indicated by the direction that the wings point and their intersection with the porphyroclast.

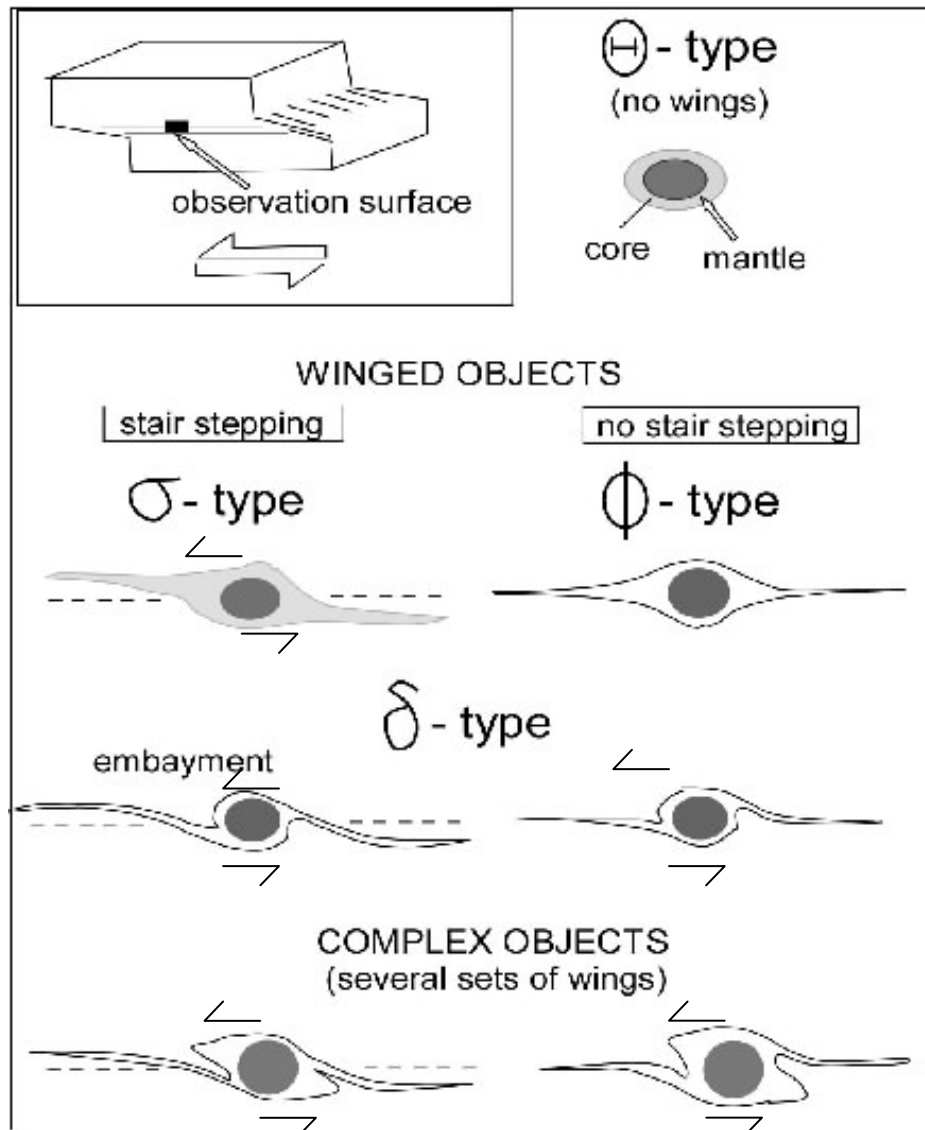


*Figure 17: Porphyroclast of a quartz grain surrounded by quartz-feldspar grains in quartz-feldspar mylonite*

Five main geometries of winged porphyroclast have been identified based on whether the centerlines of the wings are straight or curved, and symmetrical or asymmetrical with respect to an imaginary reference line through the center of the clast

(Trouw and Passchier, 1996) (Figure 18). These five main types are explained by Trouw and Passchier as follows:

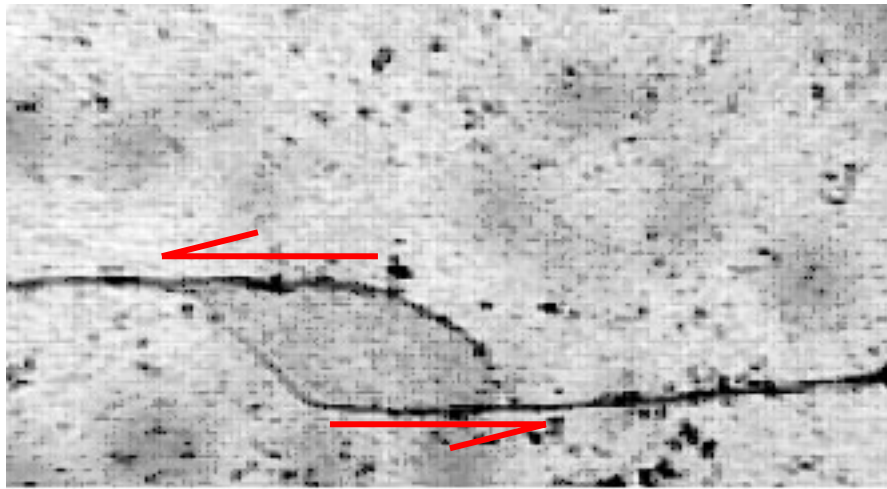
“Theta ( $\theta$ ) objects have round to elliptical mantles, but no real wings. Phi ( $\Phi$ ) objects have straight centerlines and are symmetrical with respect to the porphyroclast. Sigma ( $\sigma$ ) type has wings with curved centerlines and is asymmetric; the wing extends off the top of one side and the bottom of the opposite side, a pattern referred to as ‘stair stepping’. Delta ( $\delta$ ) type has strongly curved wings that are asymmetrical with respect to the porphyroclast. The last type is complex porphyroclast which have several sets of wings.



**Figure 18: Classification of mantled porphyroclasts. Sense of shear is top-to-the-left (Modified from Passchier and Trouw, 1996).**

### 2.3.3 Mica Fish

Mica fish is a single crystal of mica which commonly has a lozenge shaped porphyroblast forming an oblique foliation at an angle to the shear zone boundaries, with thin mica layers running off each end of the lozenge (Figure 19).



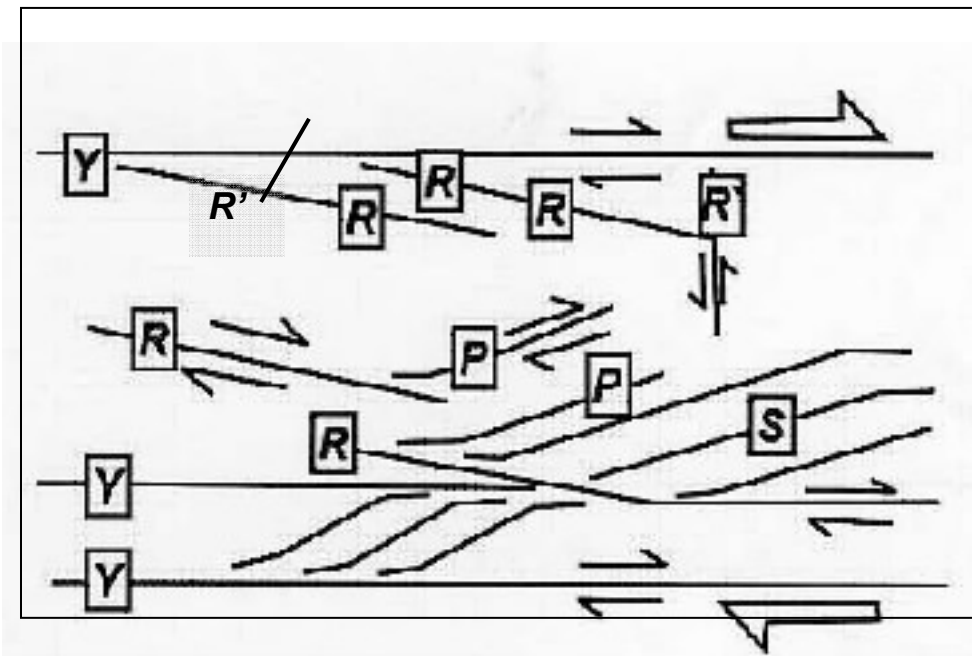
*Figure 19:  $\sigma$ -type of biotite showing sinistral shear of sense*

### 2.3.4 Reidel Shear

Sense of shear can be determined in the field from displacement of markers, slicken fibers or from the shape of striations (Petit, 1987). However, it is difficult to sample brittle rocks such as gouge and incohesive cataclasite compared to sampling ductilely deformed rocks. Although they are difficult to sample and cut, brittlely deformed fault rocks may display decent information on sense of shear in thin sections.

Sets of subsidiary shear fractures with distinct orientation and movement sense are known as Riedel shears (Chester et al., 1985). There are four different types of Riedel shears; R, R', P and Y shears. R refers synthetic (sub-parallel to major fault) and R'

refers to antithetic (opposite to major fault). P shears form at an acute angle to the main line of faulting. Y shears form from the combination of P, R, and R' shears and are parallel to the shear zone boundaries (Davis and Reynolds, 1996) (Figure 20). Riedel shears can be used as indicators of shear sense if suitable displaced markers are available. They are the analogues of ductile shear bands such as S-C shear bands but formed by brittle fracturing and differentiated shapes and orientations.



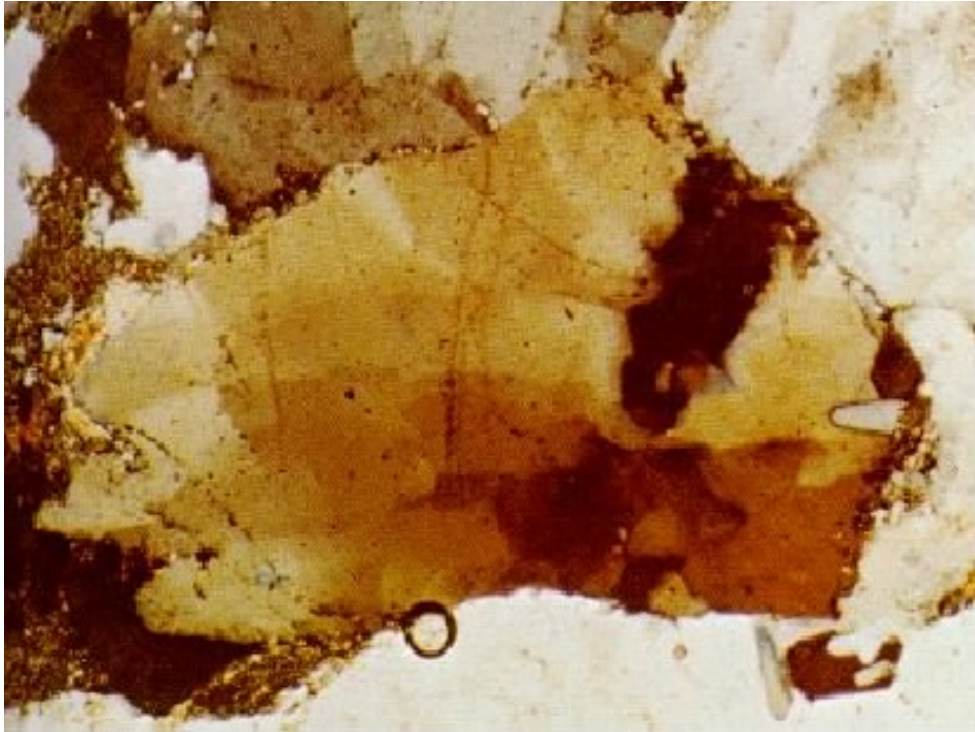
**Figure 20: Schematic diagram showing the characteristic geometry and shear sense of the most common types of Riedel shear**



## 2.4 Other Features

### 2.4.1 Undulose Extinction

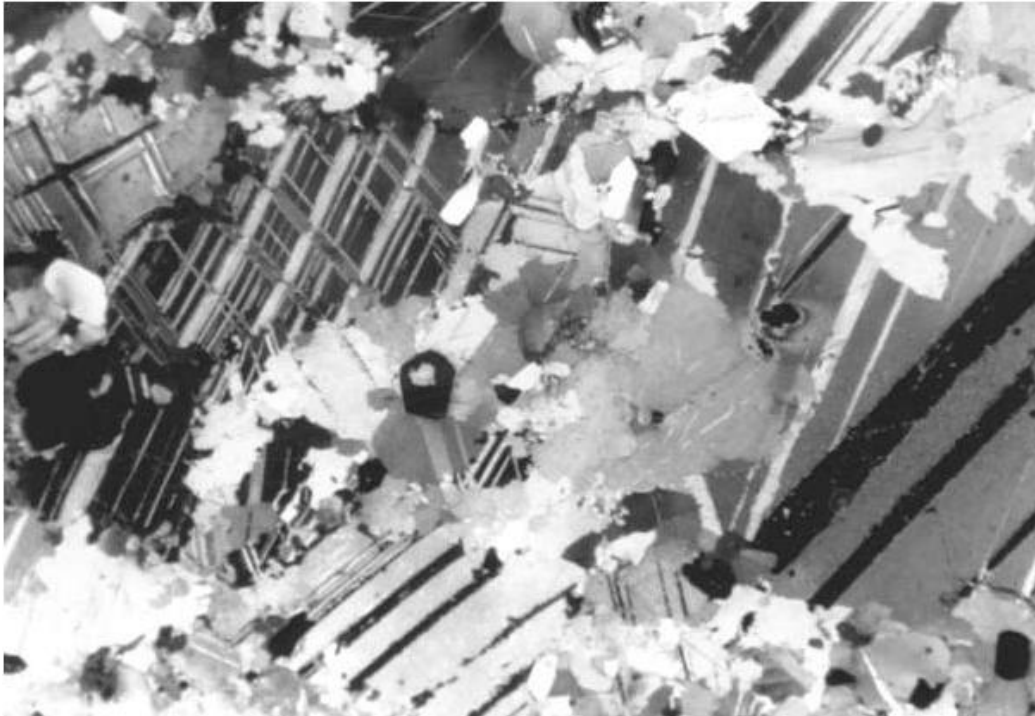
A crystal lattice which contains a large number of similar dislocations can be slightly bent; as a result, the crystal does not extinguish homogeneously as observed with crossed polars, this effect is known as undulose extinction (Passchier and Trouw, 1996) (Figure 21). In general, undulose extinction occurs in the low temperature and pressure condition (Passchier and Trouw, 1996).



*Figure 21: Undulose extinction of a quartz grain in a megacryst.*

### 2.4.2 Twinning

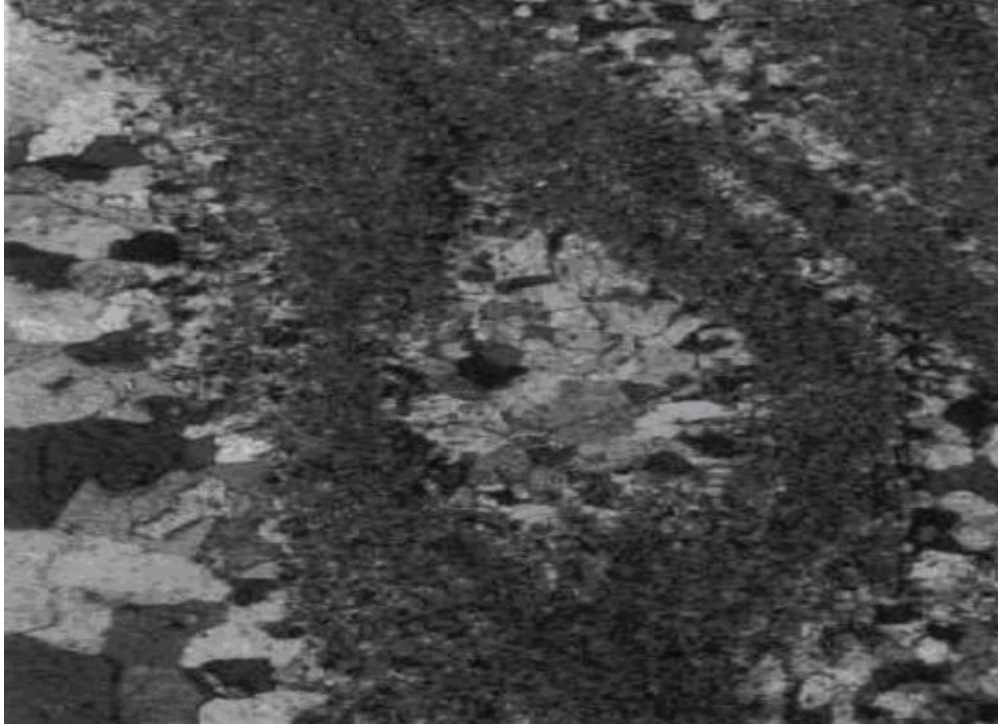
Some minerals can deform by deformation twinning that can accommodate only a limited amount of strain and always operates in specific crystallographic directions (Burkhard, 1993) (Figure 22). In general, twinning occurs in the lower temperature range of deformation (Passchier and Trouw, 1996).



*Figure 22: Calcite twins in marble.*

### 2.4.3 Cataclastic Flow

Cataclastic flow is a brittle process that is achieved by mechanical fragmentation of rocks, and subsequent sliding and rotation of fragments. It occurs along brittle fault zones at non- to low grade metamorphic conditions, and at high strain rates (Trouw and Passchier, 1996) (Figure 23).



*Figure 23: Cataclastic flow in a megabreccha*

## **CHAPTER III**

### **GENERALIZED STRATIGRAPHY OF THE STUDY AREA**

#### **3.0 Introduction**

In the study area, the Babadag fault separates ductilely to brittlely deformed metamorphic rocks of the Menderes Massif in its footwall from the brittlely deformed sedimentary rocks of the Denizli Basin in its hanging wall. The generalized stratigraphic column of these units exposed in Denizli Basin is shown on (Figure 24). These units are discussed below. This chapter will also briefly summarize the contact between the basement metamorphic rocks and its cover series which has been documented since 1990s.



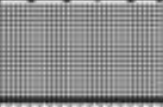







ERA	PERIOD	EPOCH	FORMATION	LITHOLOGY SYMBOLS	LITHOLOGIES
CENOZOIC	QUATERNARY	Holocene			Alluvium
		Pleistocene			Alluvial Fan
	TERTIARY				Unconformity
					Travertine
		Pliocene	Tekkekoy		conglomerate- sandstone- siltstone levels
	E - L Miocene	Cobantas		marine fossiliferous limestones whitish yellow marl	
				Unconformity	
MESOZOIC	Jurassic Cretaceous		Catalcatepe		Massive to thick bedded grey limestones
					<i>tectonic contact</i>
PALEOZOIC			Menderes M. Metamorphics		Schist, metaquartzite, phyllites

Figure 24: Generalized stratigraphic sequence of the study area (Modified from Ozpinar et al., 1999)

### 3.1 The Menderes Massif Metamorphic Rocks

The Menderes Massif comprises metaquartzite, phyllites and various schists such as mica schist, kyanite-chlorite schist (Figure 25). It also contains recrystallized limestones that crop out in different sizes in the study area. These rocks contain well-developed foliation, and are generally greenish in color. Schistic units consist of many different minerals such as muscovite, biotite, garnet and chlorites. Although having subjected various grades of metamorphism, schistose sequences are mostly in the biotite to garnet zone of the greenschist facies. Petrographic studies indicate that the common mineral assemblage in the sequence is quartz+biotite+muscovite+plagioclase+feldspar±chlorite±garnet in the schistic sequences. Petrographic analysis indicates plagioclase grains in microcrystalline texture, calcites in marbles have pressure twinning (Figure 26), quartz with undulose extinction, and quartz-feldspar grains with cataclastic flow.



*Figure 25: A view of Menderes Metamorphic schists towards NW.*



***Figure 26: A plagioclase porphyroblast showing pressure twinning in quartz-mica rich matrix; GPS, 88110; 77193.***

The basement rocks show a gradation from ductile deformation to brittle deformation.

Recrystallized limestones are grey – blackish grey where the fresh surfaces are white – beige in color (Figure 27). They are found as massive blocks of various size including cracks and many fissures which developed in different directions.

No fossils were observed in the schist sequence in the region of Babadag. Aral (2001) pointed out that the schistic unit under cover rocks must be older than Carboniferous where the age of this metamorphic unit is considered as Upper Paleozoic (Caglayan et al., 1980).

The controversy on the contact between core and cover series in the southern Menderes Massif has not yet been resolved. It has been interpreted either as a Pan-African unconformity (Sengor et al., 1984) or as an intrusive contact (Bozkurt et. al., 1999) overprinted by an Eocene top-to-south shear zone during the Late Oligocene (Hetzl and Reischmann 1996). Ring et al., (1999) suggested that the contact is a thrust fault.

Collins and Robertson (1999) stated that during the extensional collapse of Lycian orogeny some of the thrust faults were reactivated as detachment faults. Similarly, Gessner et. al. (2000) and Ozen and Sozbilir (2003) have interpreted the contact as low angle normal fault (reactivated thrust fault). In Denizli Basin this core sequence is followed by Mesozoic and Cenozoic units with Babadag normal fault.





*Figure 27. Photo shows the weathered recrystallized limestones which are grey – blackish grey where the fresh surfaces of them are white.*

### **3.2 Catalcatepe Formation**

Named by Okay (1989), Catalcatepe Formation consists of massive to thick bedded grey – blackish grey limestones. It also includes cherty limestone which is grey in color (Figure 28). Calcite+quartz+biotite+muscovite, and feldspar are the minerals observed under the petrographic microscope.



*Figure 28: A view of Catalcatepe Limestone looking towards NW.*

Taner (2001) interpreted the Catalcatepe units as deep-marine deposits that were transported as extensional allochthons onto the metamorphic sequence. Ostracods, foraminifera, gastropods, and brachiopods form the fossil assemblage of the Catalcatepe Formation. These taxa along with palynofossils suggest a Jurassic-Cretaceous age for the Catalcatepe Formation (Konak et al., 1987).

### **3.3 Cenozoic Sedimentary Units**

#### **3.3.1 Cobantas Formation**

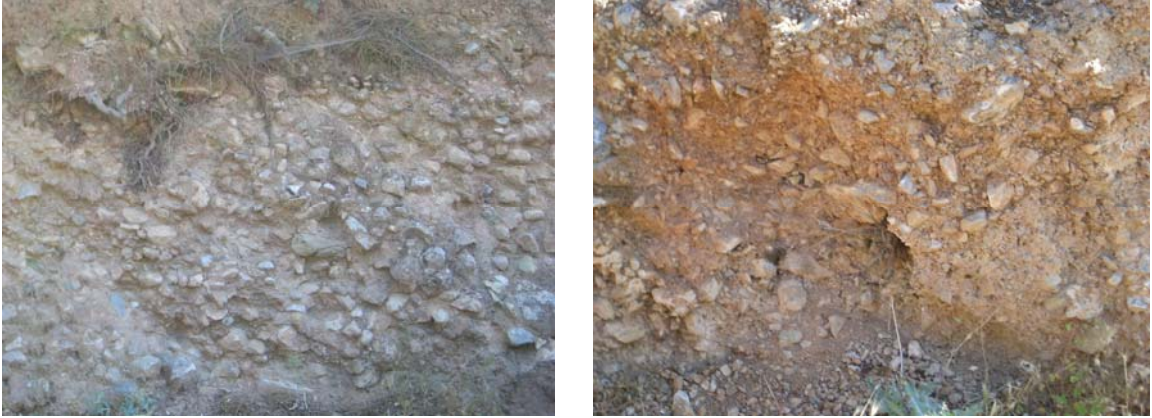
The Cobantas Formation is a sequence in the sense of Vail (1977) as a stratigraphic unit composed of a relatively conformable succession of genetically-related strata, bounded at its top and base by unconformities or their correlative conformity. The lower sequence boundary is Early to Middle Miocene in age., whereas the upper sequence boundary is Late Miocene to Early Pliocene in age (Westaway, 1992; Sun, 1990). This makes the Cobantas Formation a second order sequence. The Cobantas formation probably represents a Highstand Systems Tract. The evidence for this are fossiliferous marine limestones that grade into shallower water white to yellow marls, showing a generally shoaling upwards sequence (Figure 29). These marls contain fossil ostracods and molluscs such as *Didacna* and *Radix Phyrogovat* that suggest a brackish-water environment (Taner 2001). The best exposures of the Cobantas Formation can be studied in the uplifted parts of the basin at elevations between 600 and 700 meters (Westaway, 1993).



*Figure 29: Whitish yellow brackish water marls of the Cobantas formation photographed near the type locality.*

### **3.3.2 Tekkekoy Formation**

The Tekkekoy Formation was first described by Ercan et al (1977), and unconformably overlies the Cobantas Formation. It consists of sandy marl, lacustrine carbonates, conglomerates, and sandstones (Westaway, 1992). The microfossils from the formation are useful for paleoecological analyses and show a transition from brackish water into fresh water. The age of the Tekkekoy is Mid-Pliocene to Pleistocene (Ercan et al., 1977). The most prominent lithologies in the Tekkekoy formation are conglomerates (Figure 30). These include both matrix and clast supported conglomerates that are medium to poor sort. They are highly weathered and show hues of orange, brown and red. The clasts are sub-rounded to rounded clast and include schists and quartzites. The conglomerates are also interbedded with sandstones and these facies are interpreted as representing a fluvial system.



*Figure 30: Matrix and clast cemented conglomerates of the Tekkekoy Formation*

### **3.4 Travertines**

Travertines are the most distinct lithologies present in the Denizli area. They overly older formations and fill topographic lows (Ozkul et al., 2001). These travertines were formed by spring water rich in carbonate minerals that discharged through aquifers truncated by fault zones and extensional fissures (Bozkus et al., 2001). The contact between these Pleistocene travertines and older strata in the Denizli Basin is unconformable.

Holocene Alluvial fan deposits are not discussed in this paper because fieldwork for this study showed no evidence of deformation in these fan deposits.

## **CHAPTER IV**

### **STRUCTURAL GEOLOGY**

#### **4.0 Overview**

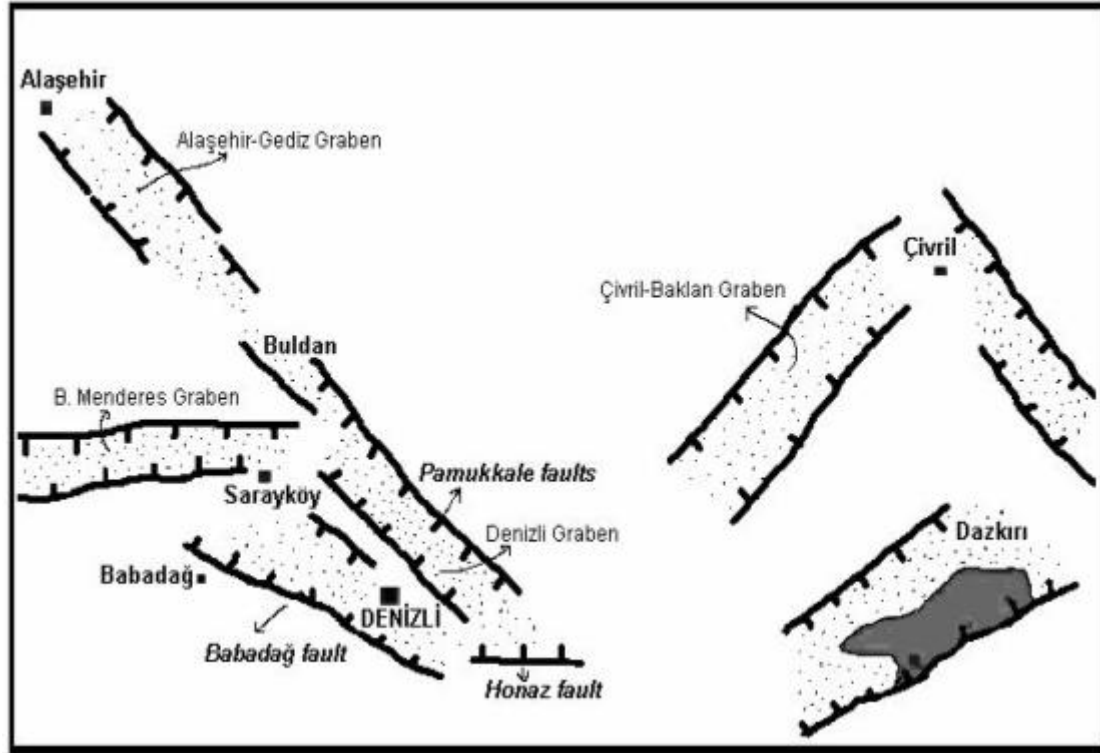
The major structural feature of the study area is the Babadag Fault Zone. In this chapter, the Babadag fault zone, and other structural features will be discussed.

#### **4.1 Babadag Fault Zone**

The Babadag fault zone is located along the eastern part of the Central Menderes Massif and bounds the southern margin of the Denizli Graben (Figure 31). Babadag Fault extends towards the south of the town of Babadag; passes 200m away from Yenikoy and continues towards Denizli (Saroglu, 1992). It separates the Menderes Massif metamorphic rocks in its footwall from the Cenozoic sedimentary rocks in its hanging wall. It is about 50km in length and shows mostly normal separation. It strikes between N60°W and N80°W, and dips about 50° to 80° NE. It has a low angle of dip in the vicinity of the town of Babadag, but, splays into high angle normal faults in the Denizli area.

The map in Figure 31 suggests that it has formed at an acute angle to Buyuk Menderes Graben system. Although the Alasehir-Gediz Graben system and Babadag fault zone have the same kinematic features and are genetically related to N-S extension

in the western Anatolia, the structural relationship between these two systems remains unclear.



*Figure 31: Faults and graben structures of the region (Modified from Taner, 1974, Sun, 1990 et al., 1990)*

The Babadağ fault is well exposed to the west and south of Yenikoy (Plate I). The fault zone contains a thin (10-15cm) fault breccias composed of metamorphic rock fragments. In the Babadağ area, the fault zone separates the metamorphic rocks in its footwall from sedimentary rocks in its hanging wall (Figure 32).



***Figure 32. Field view of Babadag fault showing hanging wall rocks and foot wall rocks (schists).***

The Babadag fault displays many branching characteristics. Two splays occur along the Babadag fault. These splayed faults can be determined as synthetic faults since they dip in the same direction as the Babadag fault.

One of the synthetic faults is exposed NE of the NW end of Babadag Fault that N50°W and dips toward NE. It does not provide a well-exposed fault surface to get an accurate strike and dip measurements.



Another synthetic fault is located at the south of Zeytinkoy. In this locality, the fault strikes N55°W and dips 45°NE (Figure 33). In some areas along the fault, sedimentary units within the hanging wall are brought into contact with metamorphics in the footwall. This fault begins to trend more E-W at its SE end just south of Alazik Hill.

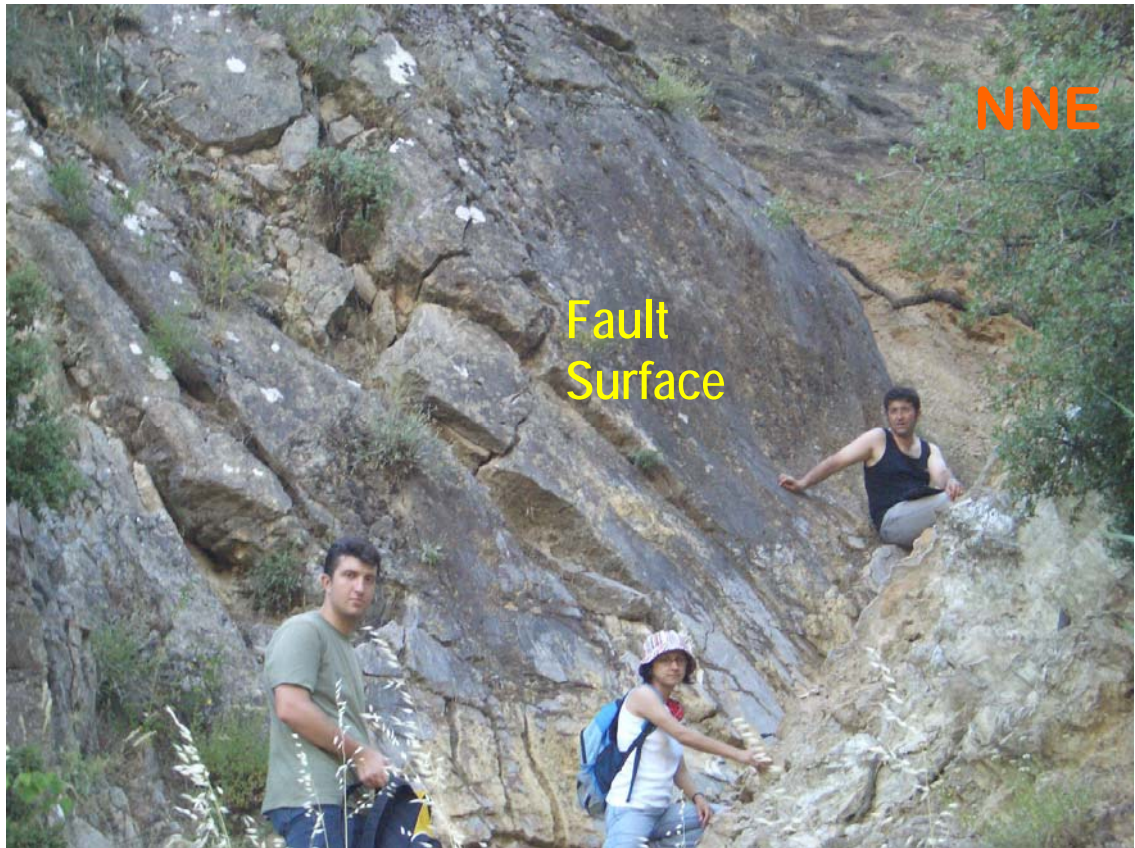


***Figure 33: GPS, 85200; 78600: A view of the footwall block and the footwall (Fw) of the fault. Upper right; the looking direction to the North.***



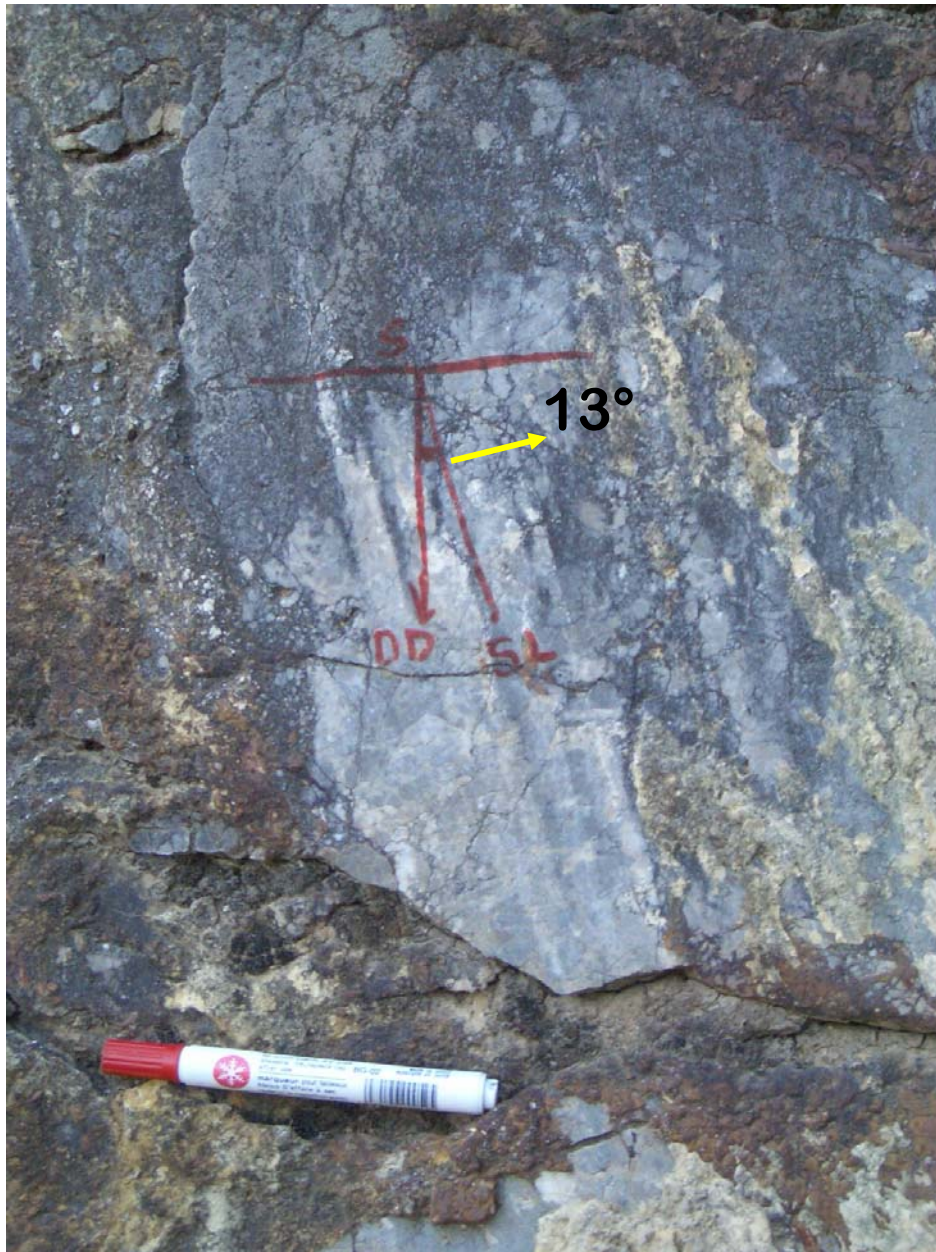
*Figure 34. Looking towards the NNW; a close up view of the small scale fractures formed due to a brittle deformation on the fault surface.*

Northeast from the Toprakbasi Village there is another high angle normal fault which strikes N50°W, 70°NE (Figure 35).



***Figure 35. Looking towards the NNE; a close up view within the fault surface of Jurassic-Cretaceous Limestone.***

At the south of Murtat Hill another splayed fault which strikes  $N60^{\circ}W$  and dips  $80^{\circ}SW$  is exposed. The angle between the dip direction and slicken lines was measured  $13^{\circ}$  indicating mostly dip-slip normal movement along the fault zone (Figure 36). The figure below shows the probable small scale oblique faults that occurred in association with the extensional mechanism of this normal fault (Figure 37).



*Figure 36. A close up view of the fault surface (FS).*



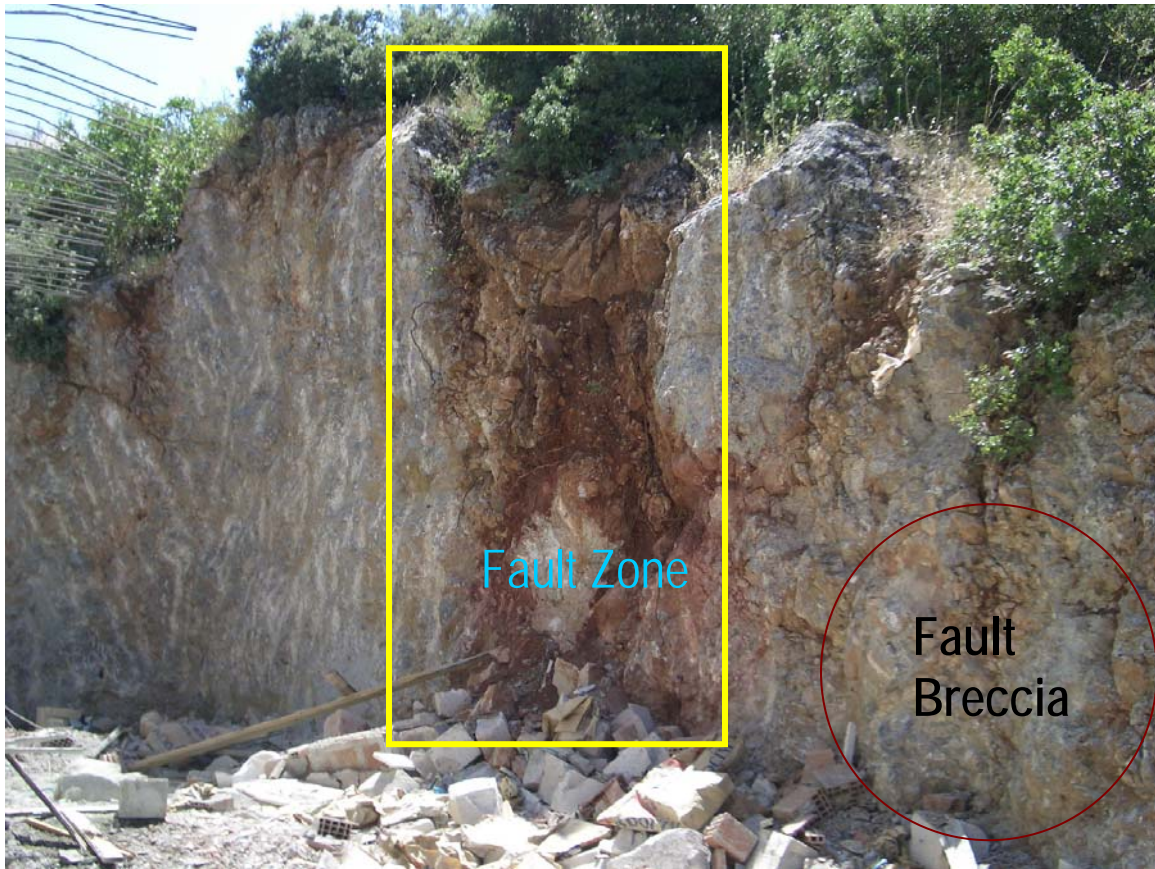
***Figure 37: GPS, 95350; 78338. Looking towards the NNE ; photo shows the probable oblique faults.***

The fault located to the west of Bekirli (GPS: 94276; 76075) strikes N10°E and dips 85° SE.



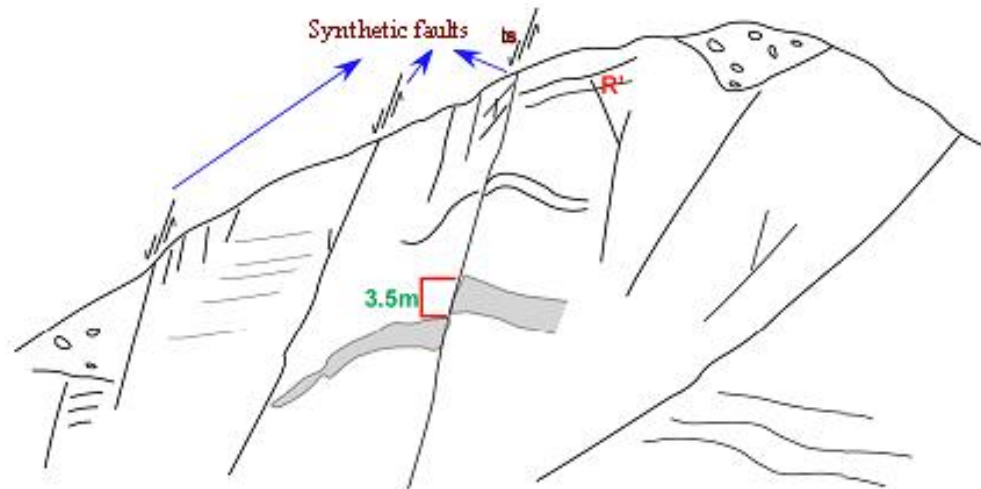
***Figure 38: (GPS: 94276; 76075) Looking towards the WNW; photo shows the fault surface.***

Because the travertine at Alazik Hill does not show a continuous sequence; it can be concluded that it was produced by fluid flow along the fault south of Taskesigi Hill which strikes N35°W and dips 82°NE (Figure 39)



***Figure 39: (GPS: 89648; 77611) Looking towards the N E; an excavation work that revealed the presence of the fault zone.***

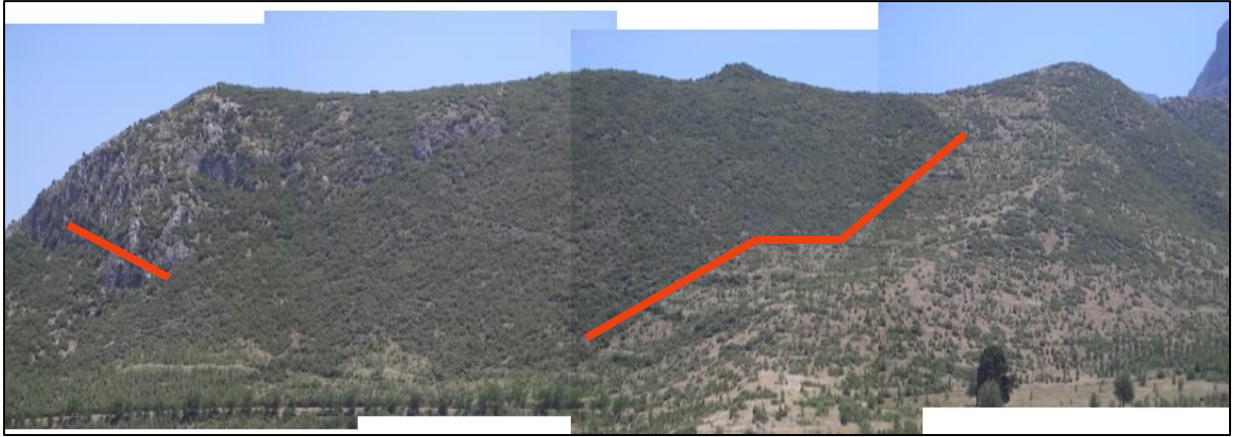
One of the most common structural products of N-S compressional tectonism during the Late Miocene in Western Anatolia is the presence of synthetic faults (Yilmaz et al., 1997a). In the study area, there are many young synthetic faults formed during the opening of the basin by the extensions that affected western Anatolia. One of these young synthetic faults is located to the south of Dereagzi. It strikes N35°W and dips 70°NE with a throw of 3.5m (Figure 40).



**Figure 40. (GPS: 84726; 77551) Young synthetic faults striking N35°W; dips 70°NE  
Throw is 3.5m.**

There are other synthetic normal faults existing in the study area that are parallel to the Babadag Fault Zone. These faults are observed extensively in the Neogene units. They are noted as listric faults and strike NW-SE.





*Figure 41. Looking towards the NE; fault is passing north of Alazik Hill strikes NW-SE.*

## **4.2 Strain Partitioning Along the Babadag Fault**

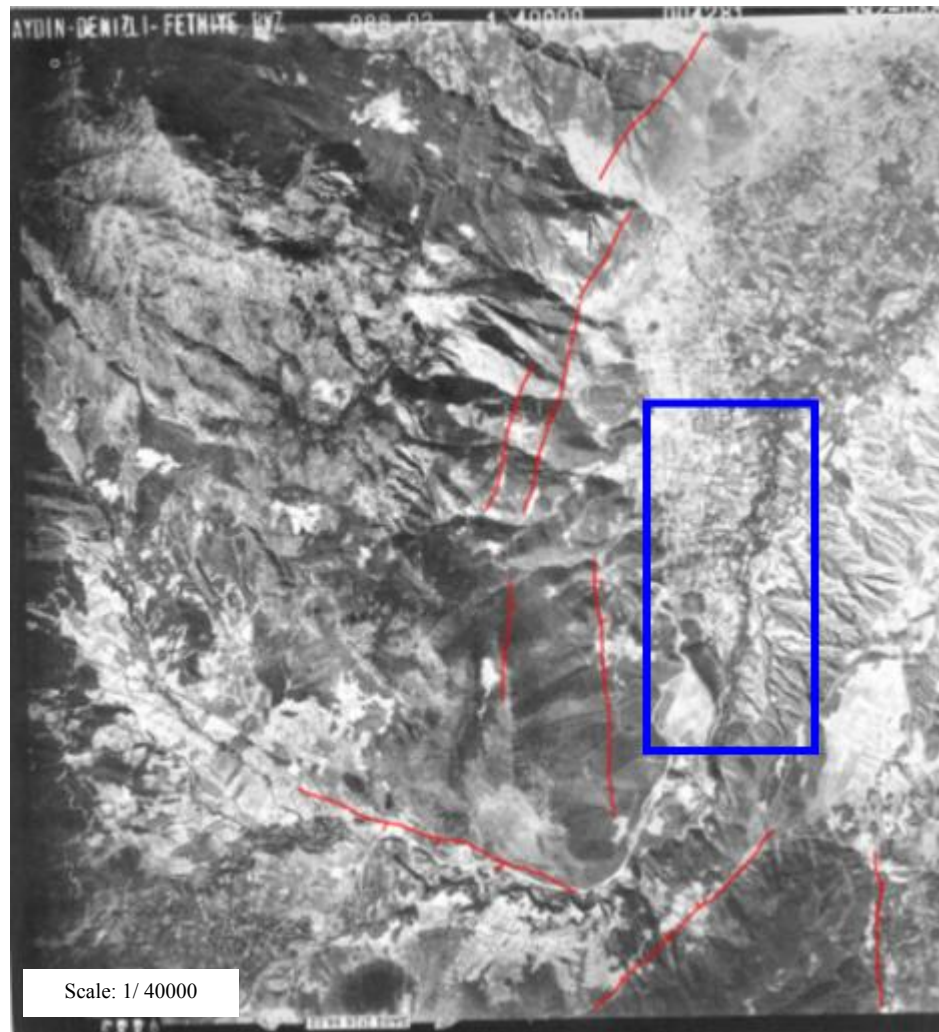
Strain partitioning is the concentration of deformation into specific parts of a rock mass by different behavior or mechanisms (Hatcher, 1995). The normal strain along the Babadag fault zone partitions by splaying along strike into smaller faults. The fault zone contains two major step-overs in the Denizli area. They are synthetic faults since they dip in the same direction as the Babadag fault. The high angle normal fault splays of the Babadag fault zone displace the quaternary rock deposits including the famous Denizli travertines which are concentrated in the areas where fault splays overlap each other.

## **4.3 Hanging wall rocks of Babadag Fault Zone**

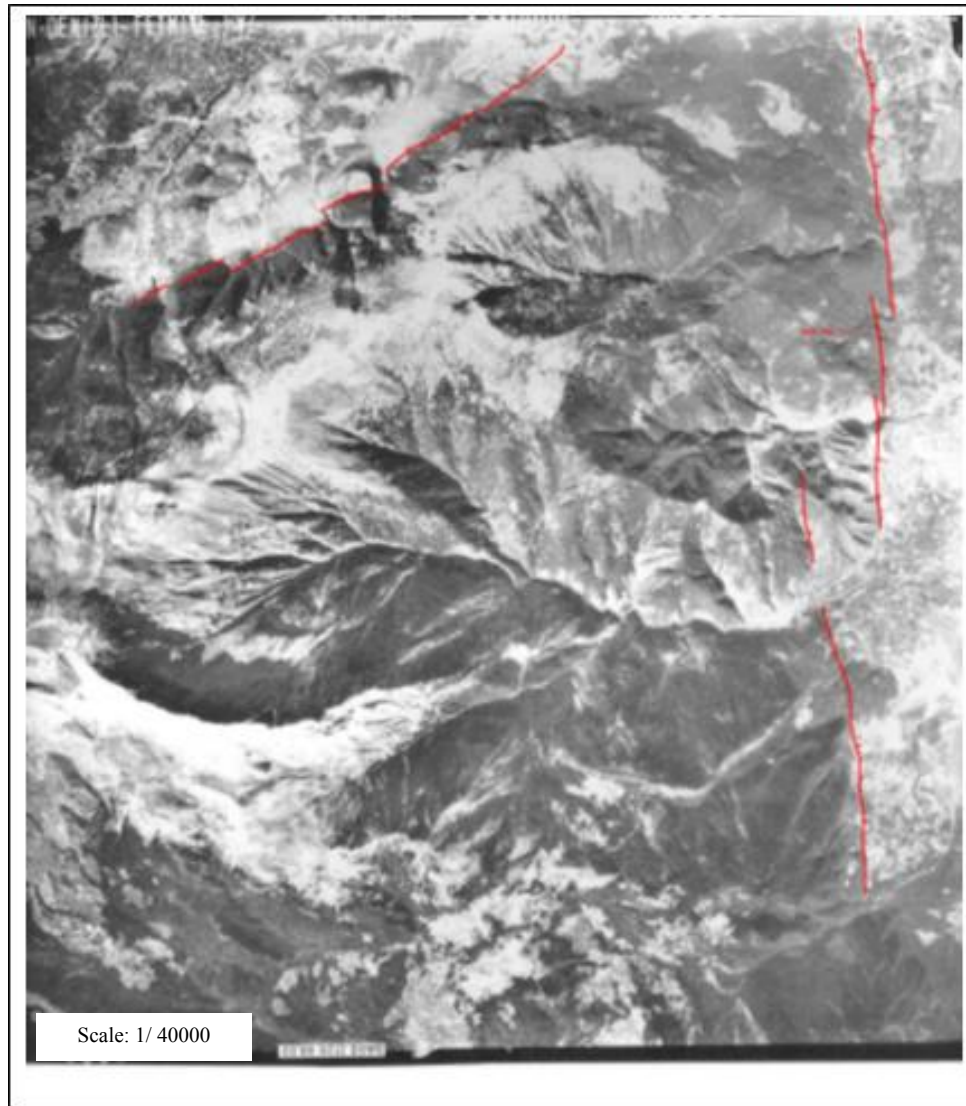
The orientation of strata close to the stream was measured N33E/19°SE and N35E/14°SE in north of Babadag (Cevik, 2002). No change in strike and dip was observed when these two data are compared each to other. However, the dip of bedding increases at some locations close to the stream. In the western part of Babadag, the dominant strata orientation is N37E/17°SE. From west toward east, the dip of bedding is decreases. Field data from the Babadag region does not show any folded structure such as anticline and syncline.

#### 4.4 Interpretation of the Aerial Photos

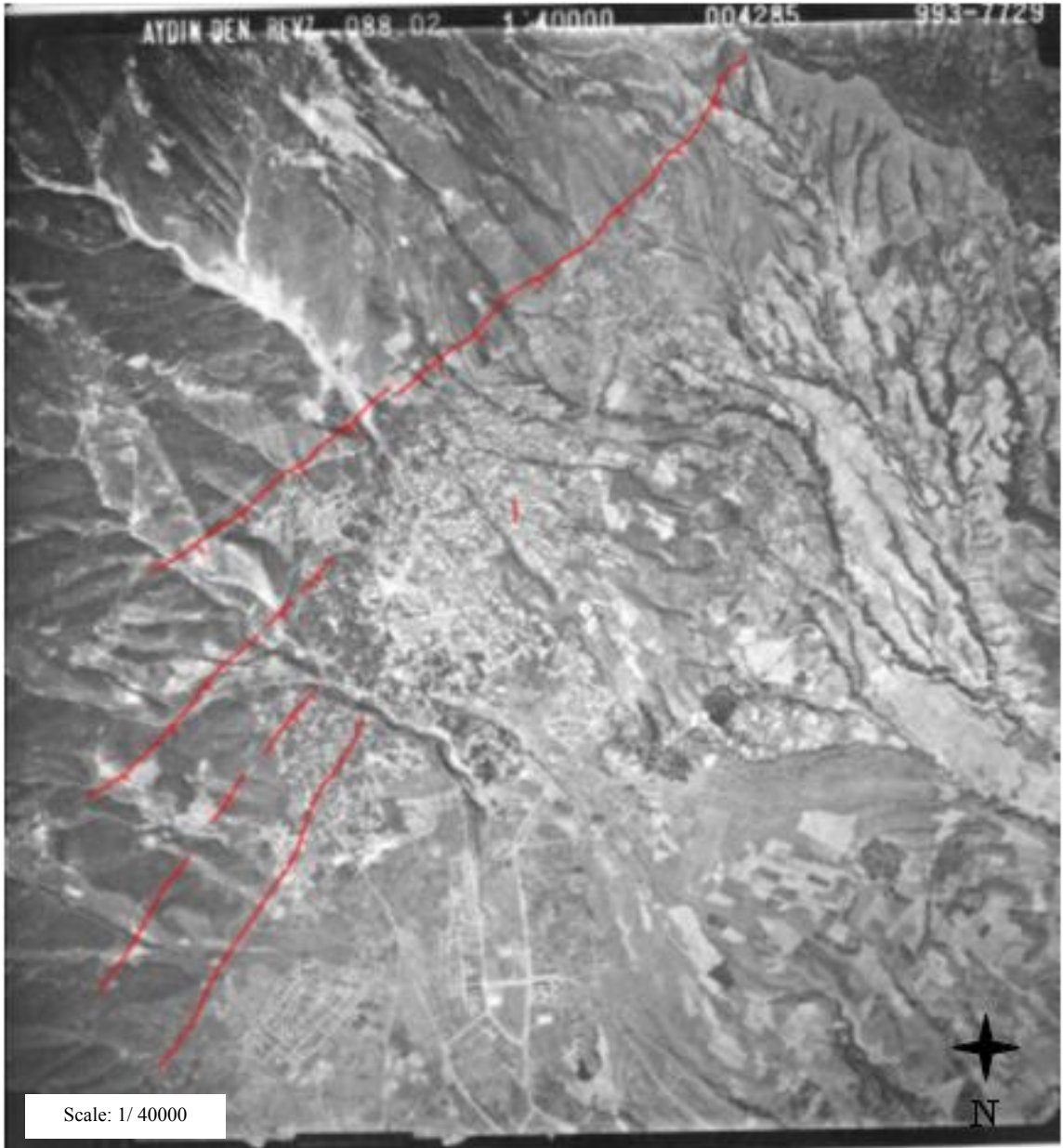
Examination of available aerial photos of the study area revealed presence of triangular facets along the Babadag fault zone. Former planner fault scarps may be dissected by erosion to yield triangular facets, physiographic signatures that indicate locations of fault contacts (Figure 42).



*Figure 42: Triangular facets along the front of Babadag fault.*



*Figure 43: Aerial photo showing the splays of the Babadag fault zone to the south of the Denizli Basin.*



*Figure 44: Aerial photo of Denizli region. Probable normal faults are indicated by red lines.*

## **CHAPTER V**

### **KINEMATIC ANALYSIS ALONG THE BABADAG FAULT ZONE**

#### **5.0 Introduction**

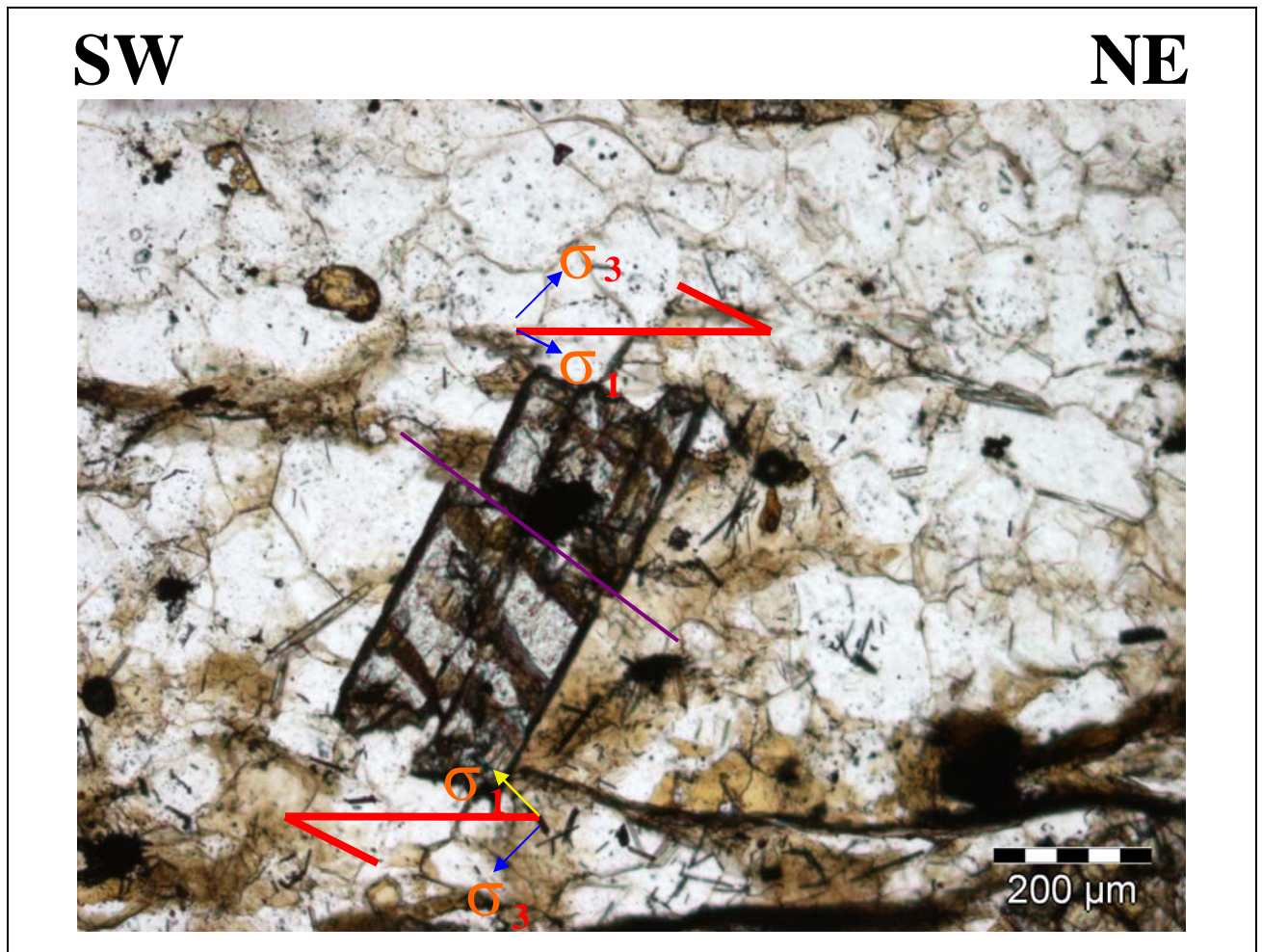
The footwall rocks of the Babadag fault zone contain well developed shear sense indicators. In order to examine the kinematic history of the Babadag fault zone, 25 oriented metamorphic rock samples were collected along the Babadag fault zone for determining shear sense. The samples were cut and analyzed for shear-sense determinations under regular petrographic microscope. The thin-sections were prepared parallel to stretching in order to get the best geometry of the kinematic indicators. Some samples contain excellent shear-sense indicators such as asymmetric porphyroclasts, mica fish, S-C and S-C' fabrics and microfaults.

The samples 04 - 11; 04 -04; 04 -07; 04 -14 contain mostly brittle-shear-sense indicators. The samples 04 -11; 04 - 06; 04 -07; 04 -08; 04 -10; 04 -20; 04 -24 contains mostly ductile-shear-sense indicators. The samples 04 -11, 04-07 show both brittle-ductile shear sense indicators (Plate I).

In the following section, the shear sense indicators and other microstructural features will be described. They will then be interpreted in terms of kinematic history of the Babadag fault zone.

## 5.1 Brittle Shear Sense Indicators

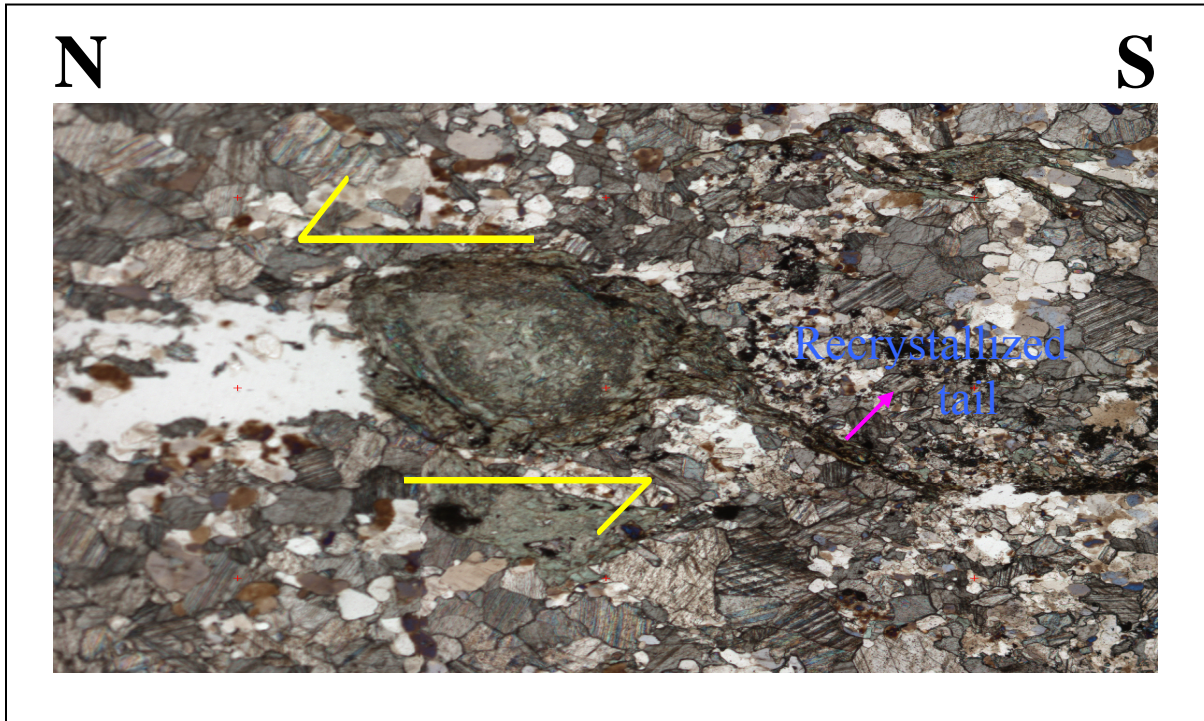
*Sample 04-11: GPS, 93132; 77823.* Synthetic microfault transecting kyanite porphyroclast in chloritoid schist. The tension fracture has formed normal to  $\sigma_3$  direction and indicates tensile fracturing. The fracturing does not continue into the mantle of quartz surrounding the kyanite (Figure 45).



*Figure 45 (Sample 04-11) Photomicrograph depicting synthetic microfault found in kyanite crystal inferred sense of shear is top to the north.*

## 5.2 Ductile Shear Sense Indicators

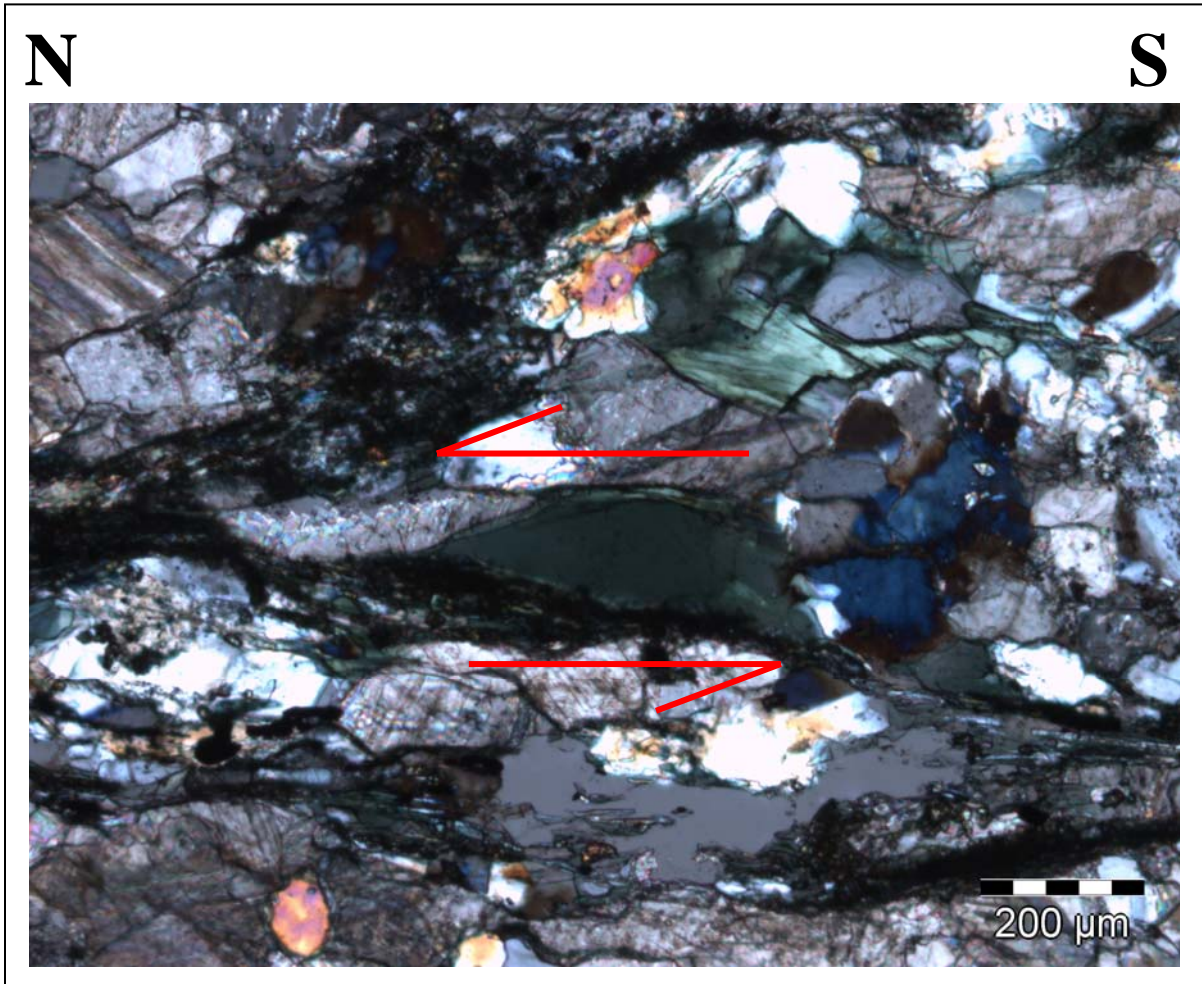
**04-06: GPS**, 87118; 77726. Sigmoid shape ( $\sigma$ -type) altered biotite in recrystallized limestone, showing top to the north ductile shearing. The assemblage is biotite+calcite+plagioclase+quartz (Figure 46).



**Figure 46 (Sample 04-06):** Photomicrograph depicting sigmoid shape biotite, inferred sense of shear is top to the north.

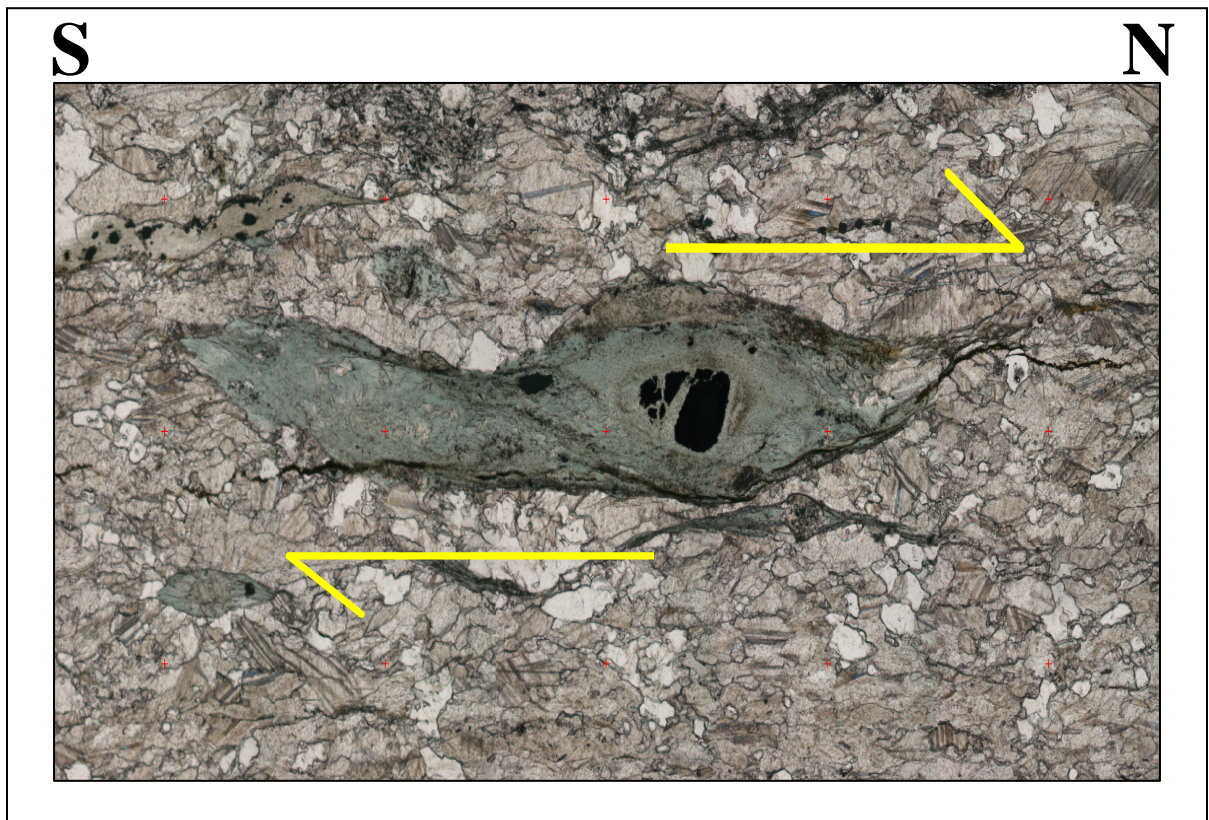


**04-07: GPS**, 88110; 77193. Altered biotite in recrystallized limestone with a  $\sigma$ -type porphyroblast. The assemblage is calcite+chlorite+muscovite+plagioclase+quartz. Tail step up to the left, indicating top to the north shear sense. Sharply curved tails in quartz-rich areas represent strain shadows (Figure 47).



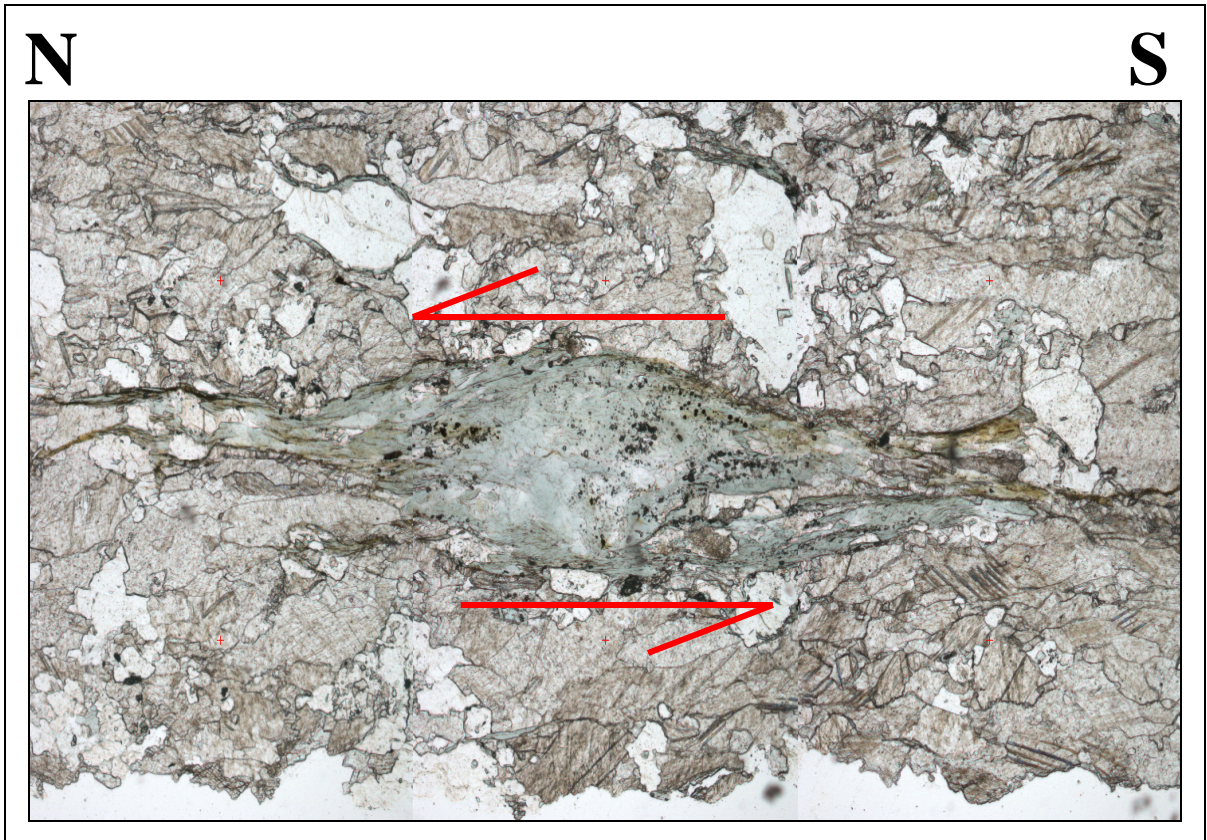
**Figure 47 (Sample 04-07): Altered Biotite in recrystallized limestone with a  $\sigma$ -type porphyroblast showing top to the north shear of sense.**

**04-20: GPS,** 90522; 77418. Complex mantled porphyroblast of chlorite in chloritoid schist enveloped by a schistose fabric comprising chlorite, plagioclase, muscovite, quartz, calcite and biotite. The asymmetry of the chlorite porphyroblast tilted sub-parallel to S surfaces can be used as an indicator of sense of shear. The structure indicates a top to the north (right) sense of shear (Figure 48).



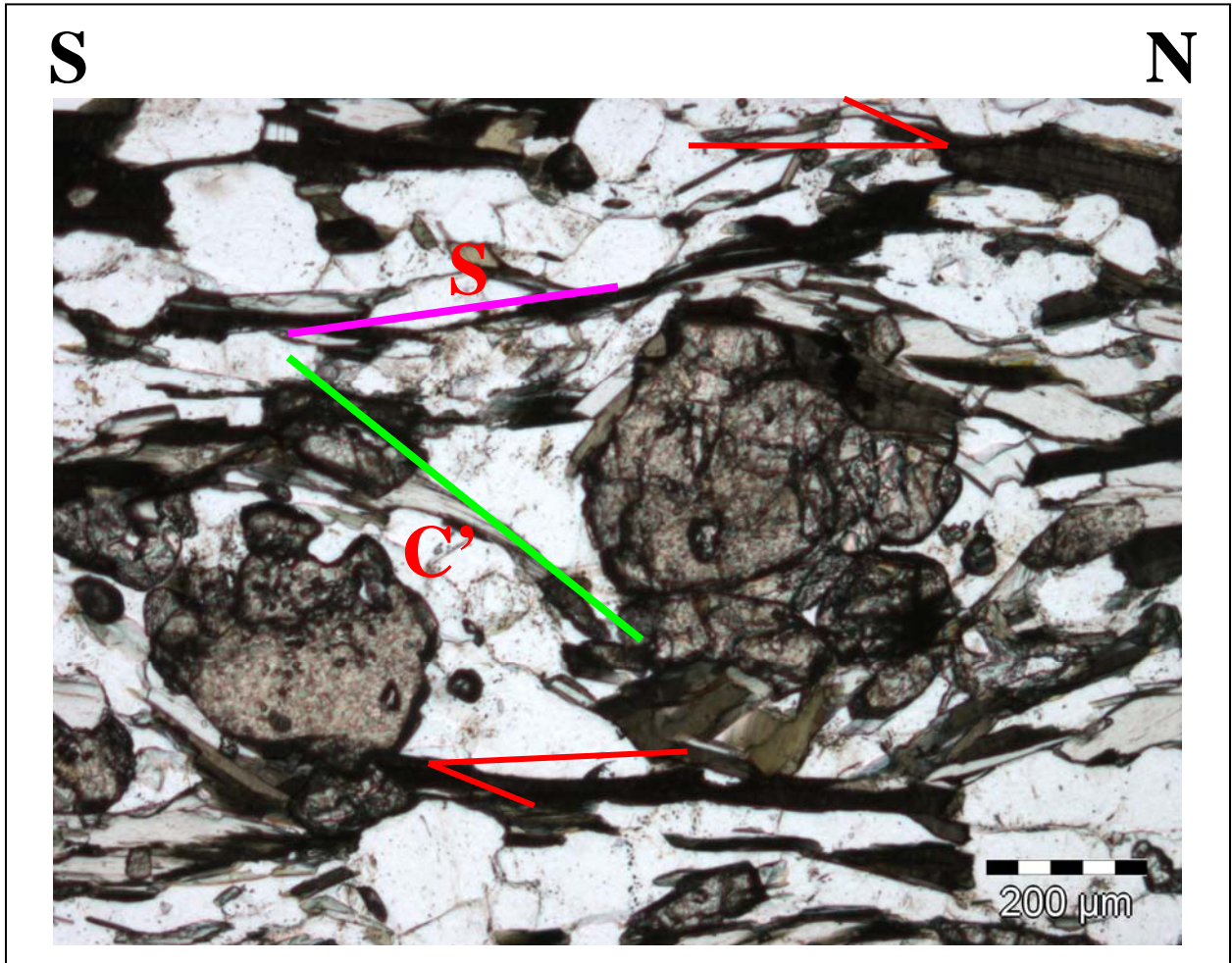
**Figure 48 (Sample 04-20): Chlorite porphyroblast in chloritoid schist showing top to the north shear of sense.**

**04-20: GPS, 90522; 77418.** Complex mantled porphyroblast of chlorite in chloritoid Schist. The structure indicates a top to the right sense of shear (Figure 49).



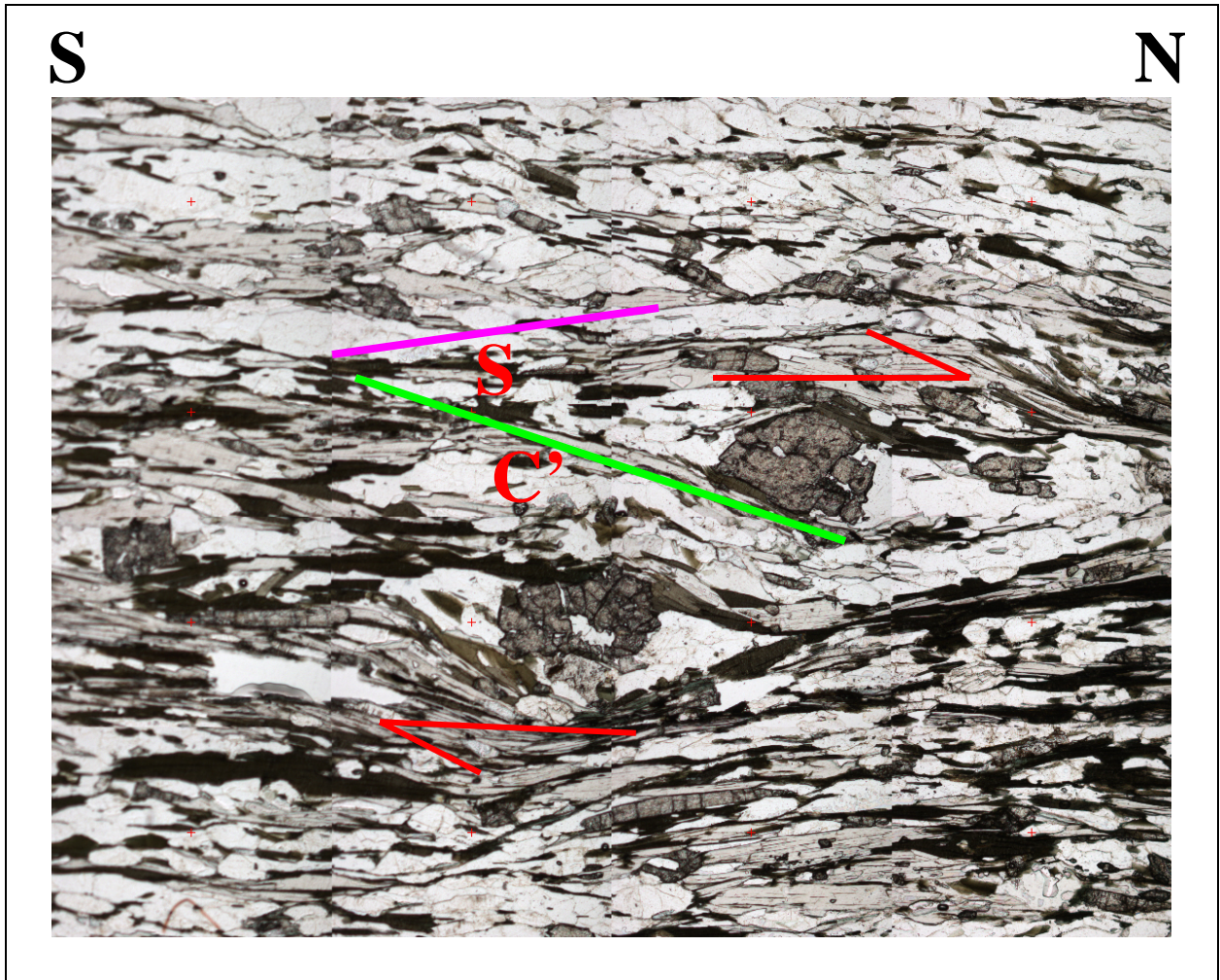
***Figure 49 (Sample 04-20): Chlorite porphyroblast in Chloritoid schist showing top to the north shear of sense.***

**04-10: GPS:** 93118; 77433 Garnet porphyroblast showing C'-type shear bands in a quartz + feldspar + biotite + garnet mylonite, indicating dextral shear sense (Figure 50).



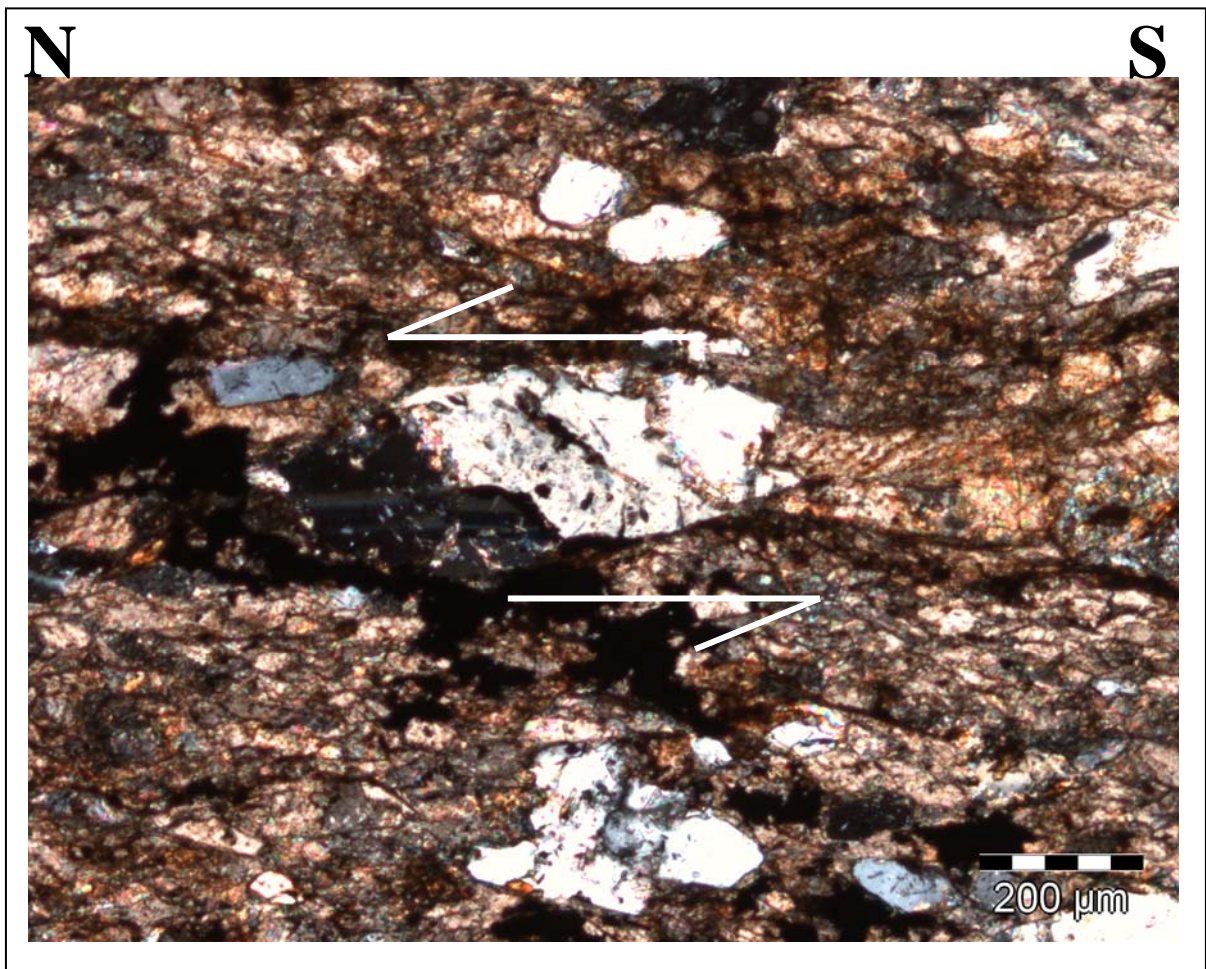
**Figure 50 (Sample 04-10): C'-type shear bands quartz-feldspar garnet mylonite, indicating dextral sense.**

**04-10: GPS:** 93118; 77433. Garnet porphyroblast showing C'-type shear bands in a quartz+feldspar+biotite+garnet mylonite, indicating top to the north dextral shear sense (Figure 51).



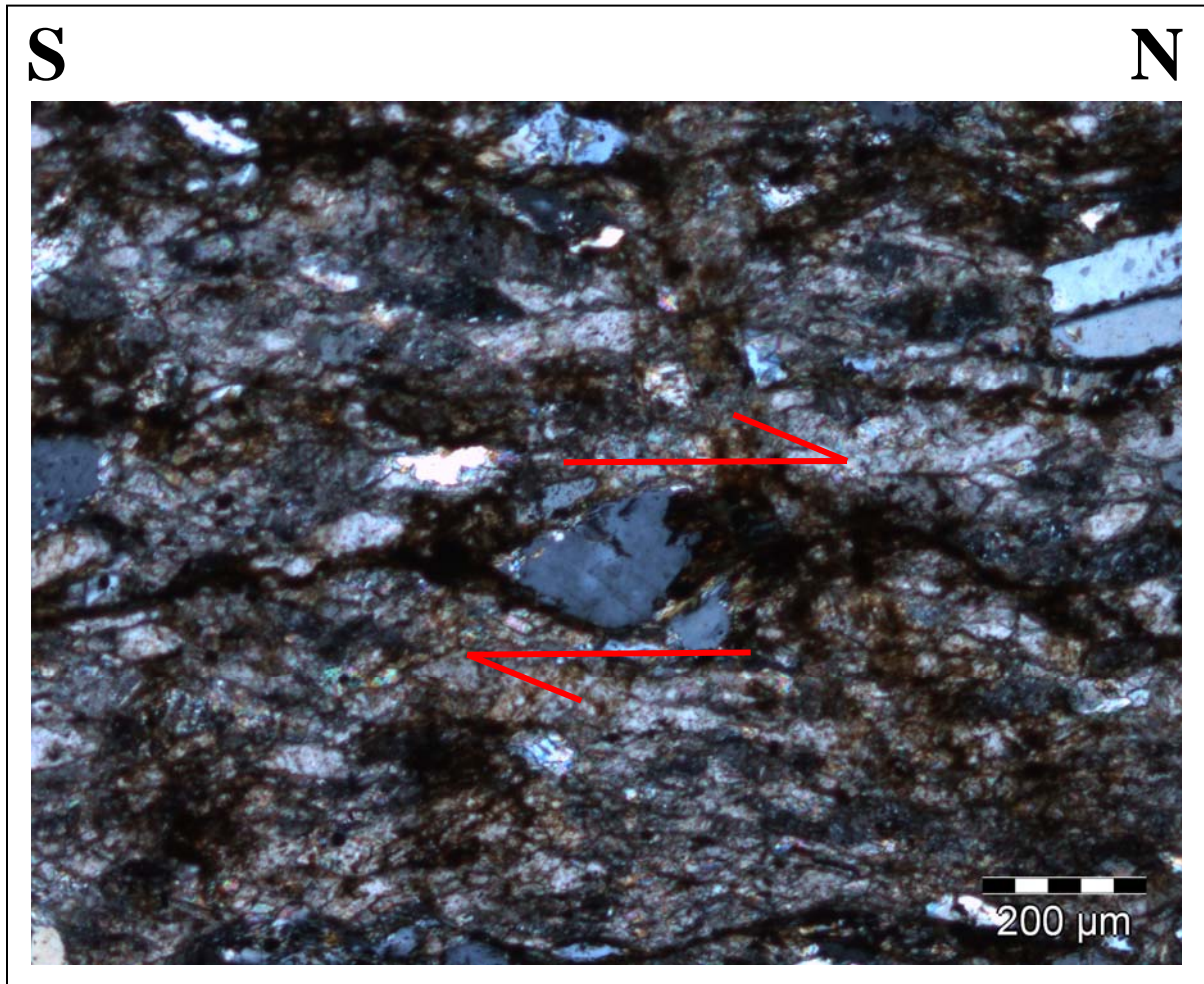
*Figure 51 (Sample 04-10): C'-type shear bands quartz-feldspar garnet mylonite, indicating dextral sense.*

**04-07: GPS, 88110; 77193.** Recrystallized limestone with an  $\sigma$ -type porphyroblast of plagioclase grain showing top to the north shear of sense (Figure 52).



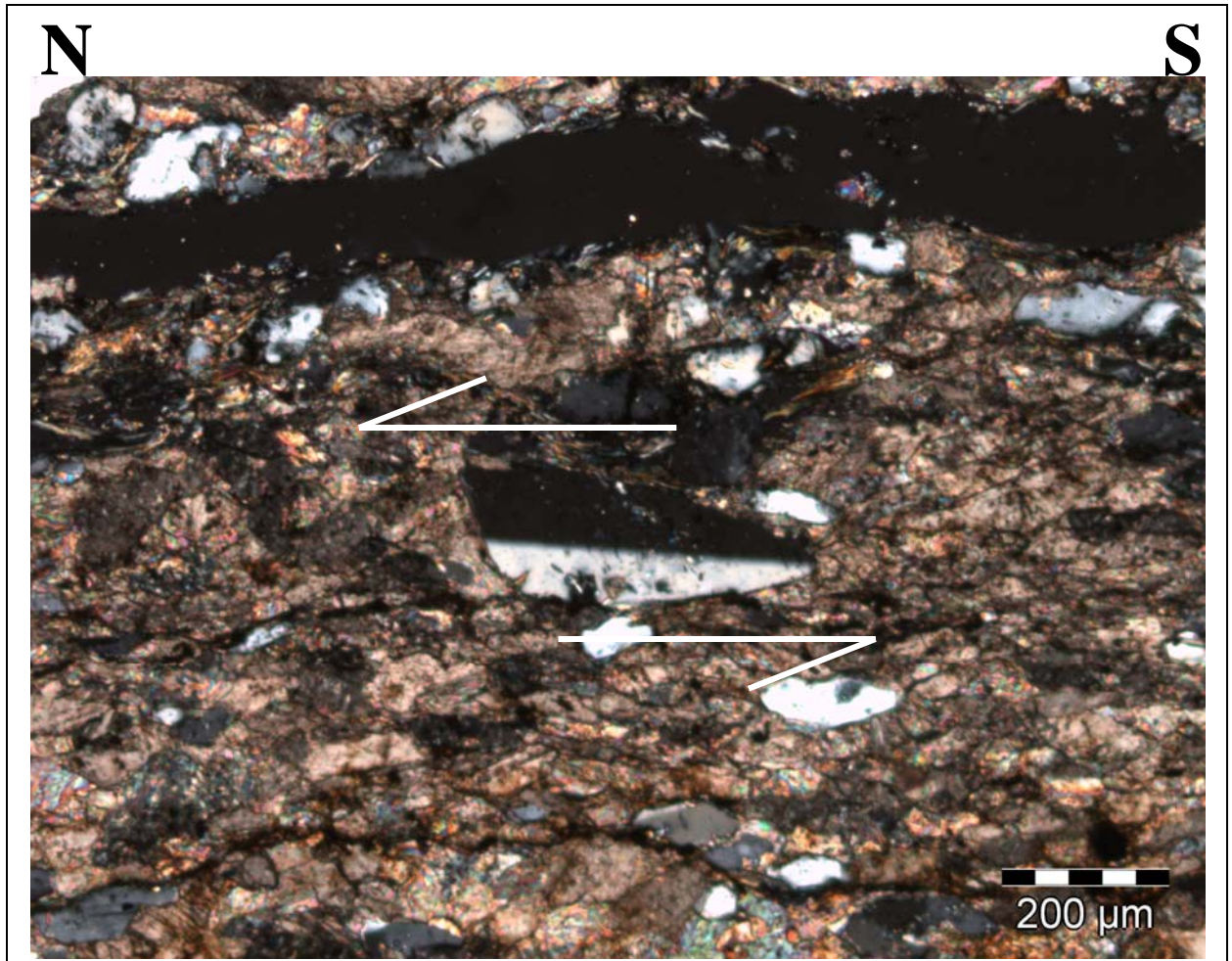
**Figure 52 (Sample 04-07): Sigmoid shape ( $\sigma$ -type) porphyroblast of plagioclase showing a left-lateral shear sense indicates a top to the north sense of shear.**

**04-08: GPS, 87689; 77113.** Recrystallized limestone with an  $\sigma$ -type porphyroblast of plagioclase grain showing a dextral shear sense. The structure indicates a top to the north shear of sense (Figure 53).



**Figure 53 (Sample 04-08): Sigmoid shape ( $\sigma$ -type) porphyroblast of plagioclase showing a right-lateral shear sense indicates a top to the north shear of sense.**

**04-07: GPS**, 88110; 77193. Porphyroblast of plagioclase in a quartz-mica rich matrix, showing a left-lateral shear sense. The structure indicates a top to the north sense of shear (Figure 54).



**Figure 54 (Sample 04-07):** Porphyroblast of plagioclase showing a twinning structure indicates top to the north shear of sense.

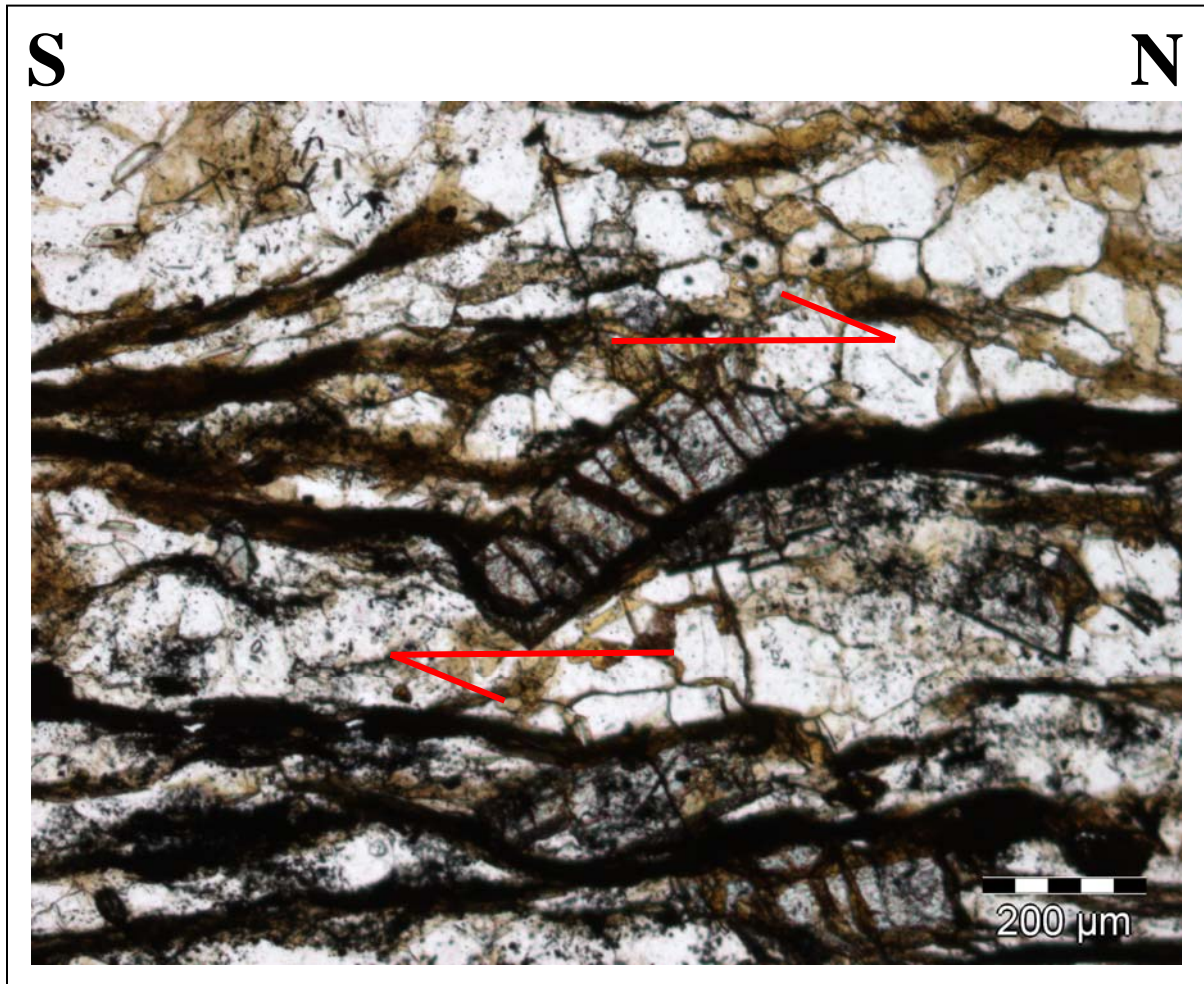


**04-24: GPS, 94927; 78501.** Recrystallized limestone with  $\sigma$ -type porphyroblast of K-feldspar surrounded by a matrix of recrystallized quartz in mylonitic schist. The structure indicates top to the north dextral sense of shear, resulting in internal deformation of grains, rotation recrystallization, and the development of elongated subgrains (Figure 55).

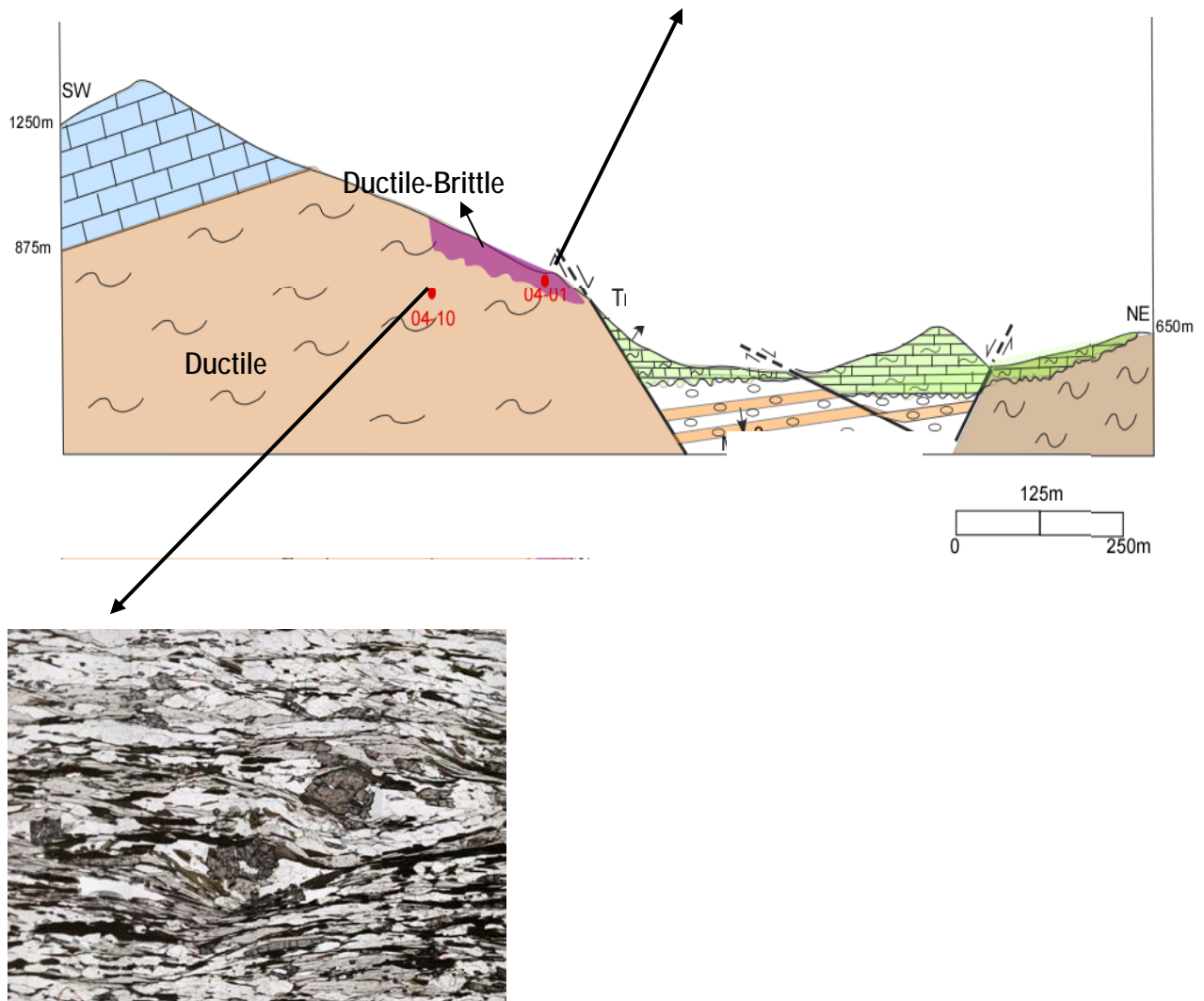
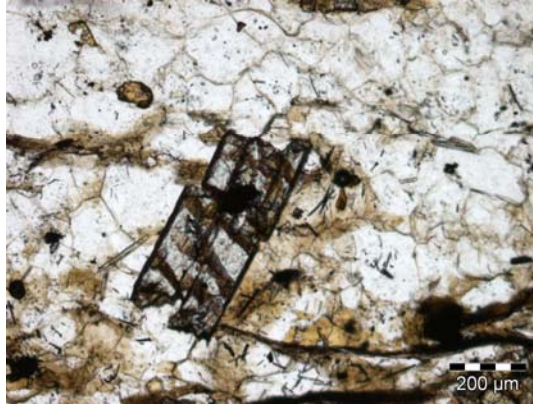


***Figure 55 (Sample 04-24):  $\sigma$ -type porphyroblast of K-feldspar surrounded by a matrix of recrystallized quartz in mylonitic schist.***

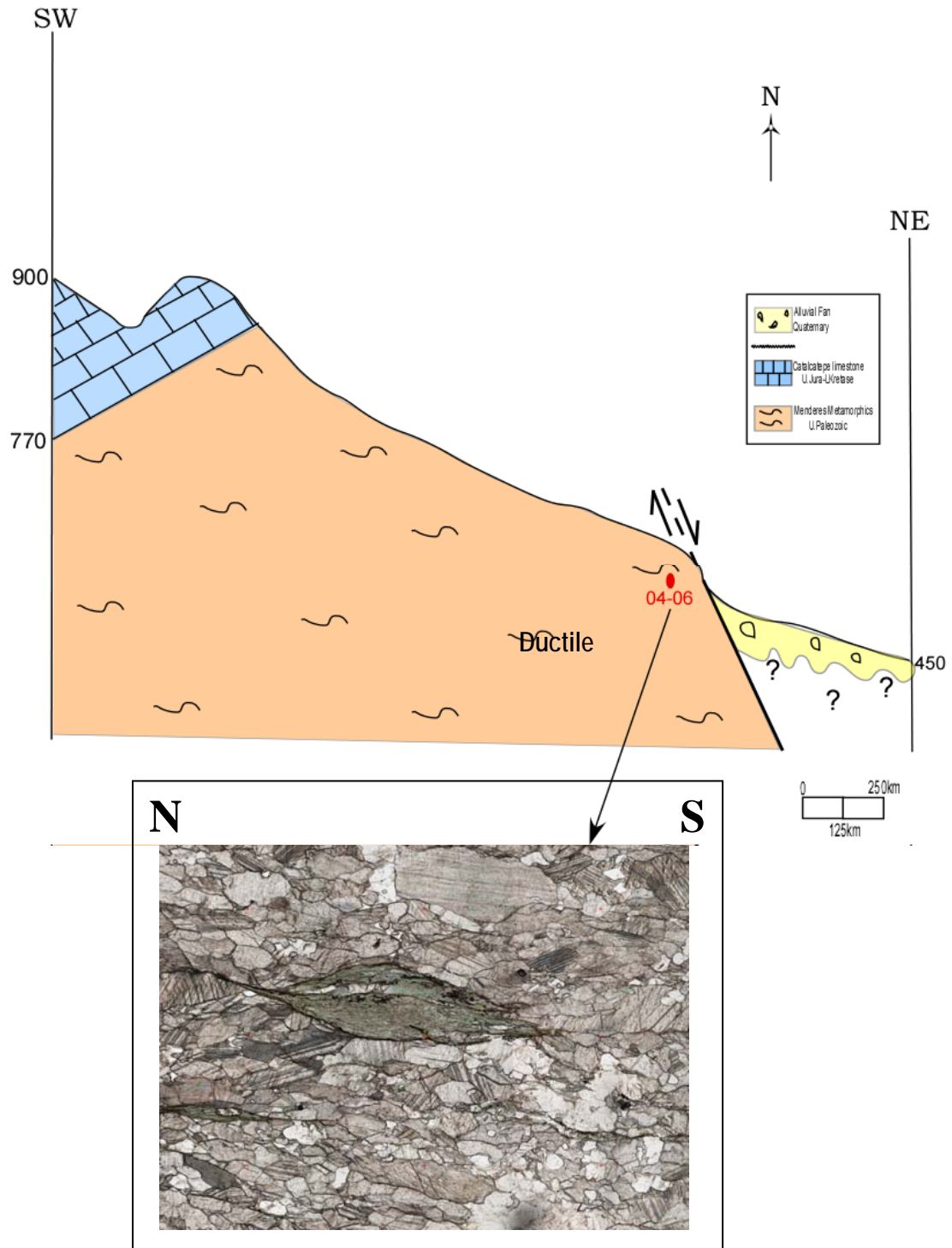
**04-11: GPS, 93132; 77823.** Brittlely deformed kyanite biotite schist with asymmetric porphyroblast feature. Sense of shear is top to the north and right lateral (Figure 56).



***Figure 56 (Sample 04-11): Asymmetric porphyroblast in kyanite biotite schist showing top to the north shear of sense.***



**Figure 57: Cross section with the photomicrograph of synthetic microfault in kyanite crystal and C'-type shear bands quartz-feldspar garnet mylonite indicating dextral shear of sense.**



**Figure 58: Cross section with the photomicrograph of ductilely deformed biotite clast.**

### **5.3 Interpretations of the microtectonic studies**

Microstructural analysis of the footwall fault rocks along the splays of the Babadag fault zone to the southwest of the Denizli basin reveals the presence of ductile shear-sense indicators such as  $\sigma$ -type porphyroclasts and C'-type bands. These all indicate top to the north simple shear. The footwall rocks also contain top to the north brittle shear sense indicators such as synthetic tension fractures, and Riedel fractures overprinting the ductile shear sense indicators. These observations suggest that the prominent movement along the Babadag fault zone is top to the north.

The timing of this top to the north normal movement, however, is unknown. The Babadag fault zone contains Upper Miocene and Plio-Quaternary sedimentary rocks of the Denizli Basin in its hanging wall. The ductile top to north movement may have been formed during the extension in the Menderes Massif, which is earlier than Miocene, perhaps as early as Late Oligocene (Catlos and Cemen, 2004; Cemen et al. (in press).

## CHAPTER VI

### CONCLUSION

The following conclusions were reached regarding the nature of extension within the Babadag fault zone.

- 1) The Babadag fault zone that controls the southern margin of the Denizli Basin is a steeply dipping normal fault. It contains two major step-overs as displacement along the fault partitions westward.
- 2) Extensional elements exhumed in the Babadag fault zone have shown that extension was ductile deformation during the early stage of deformation.
- 3) In some areas the Babadag fault is defined by its topographic expression and interpretation of arial photographs.
- 4) Deformation features in the Buyuk Menderes shear zone range from ductile-brittle.
- 5) In some samples, biotite shows retrogression to chlorite. In sample 04-01 they are found in association with sphene.
- 6) Microstructural information of mica fish-type porphyroclasts, complex mantled porphyroclasts, garnet porphyroclasts showing C'-type shear bands, syntectonic spiral S- garnet,  $\sigma$ -type chlorite porphyroclasts,

found in the fault rocks along the Babadag fault zone, indicate top to the north movement of the upper plate.

- 7) The fault rocks consisted of brittle shear sense indicators such as undoluse extinction, twinning, cataclastic flow and synthetic microfault showing top to the north movement of the upper plate are overprinting the ductile shear sense indicators. These observations suggest that the prominent movement along the Babadag fault zone and its high-angle splays is top to the north.
- 8) The timing of the top to the north normal movement, however, is in question. The fault zone contains Upper Miocene and Plio-Quaternary sedimentary rocks of the Denizli Basin in its hanging wall. Therefore, it may have been formed in Late Miocene.
- 9) The potential also exists to date monazite in rock thin sections from the fault rocks with an ion microprobe to determine the exact age of Babadag fault zone and its splays controlling the southwestern margin of the Denizli Basin.

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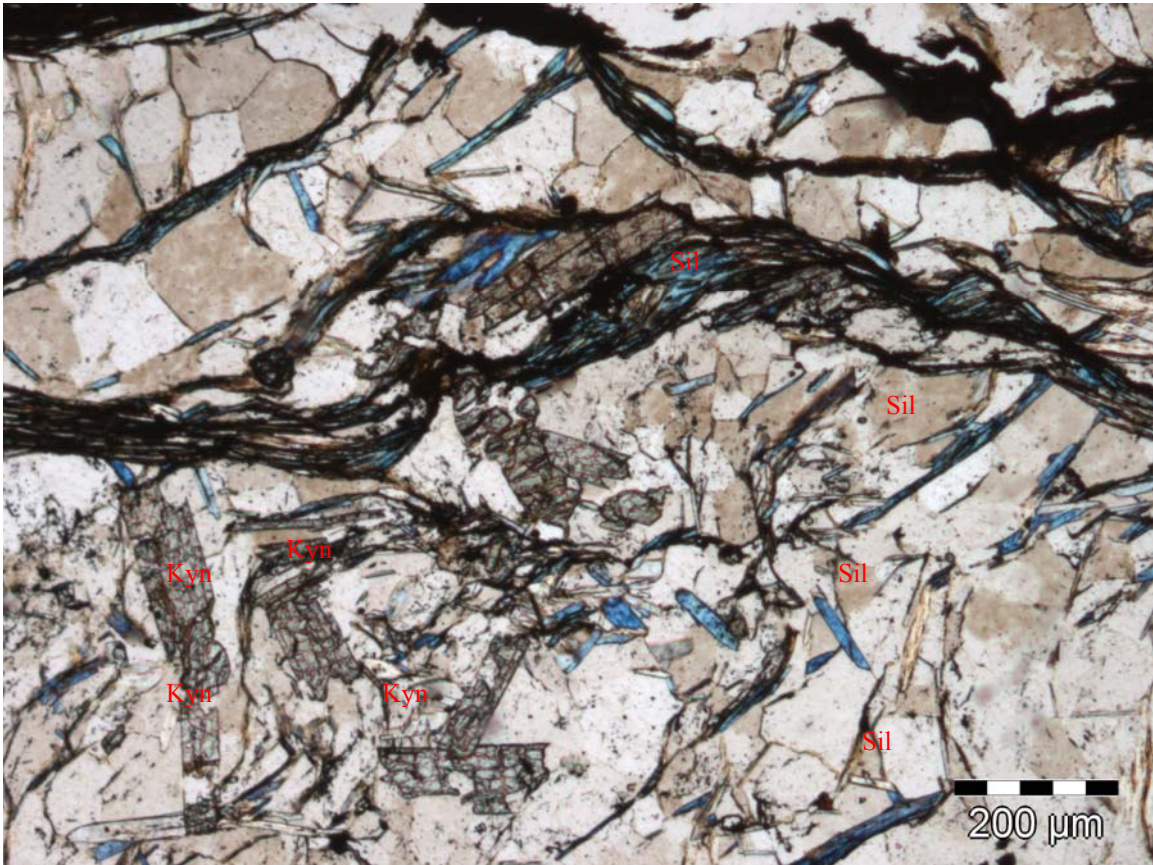
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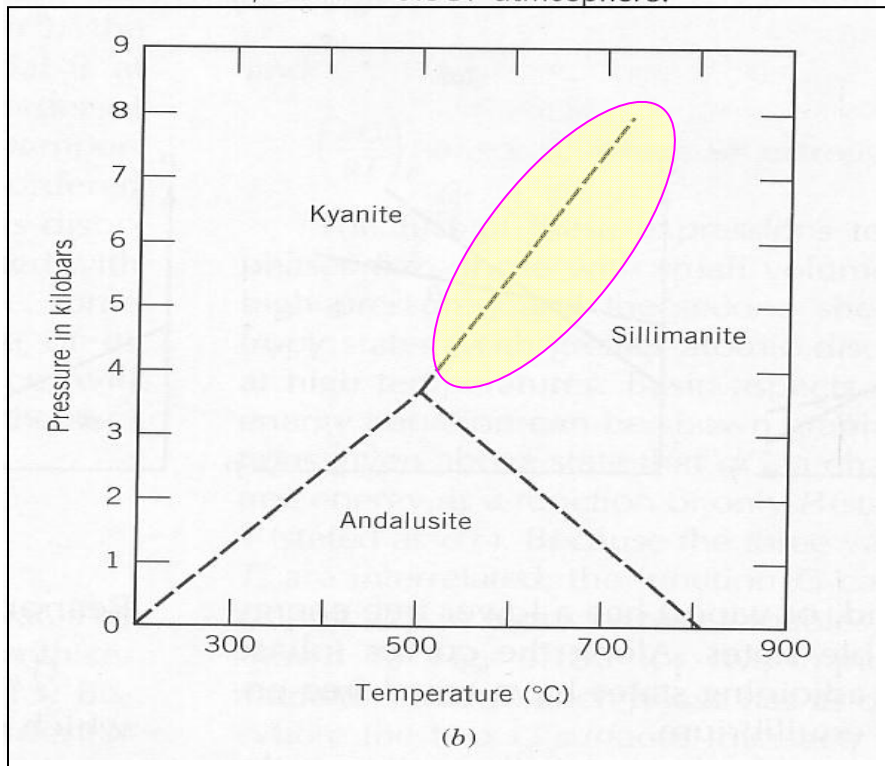


## Appendix A

*Sample 04-11 GPS*, 93132; 77823. Co-existing of kyanite and sillimanite indicates a certain temperature and pressure condition (500°C-750°C and 4-8kbar)



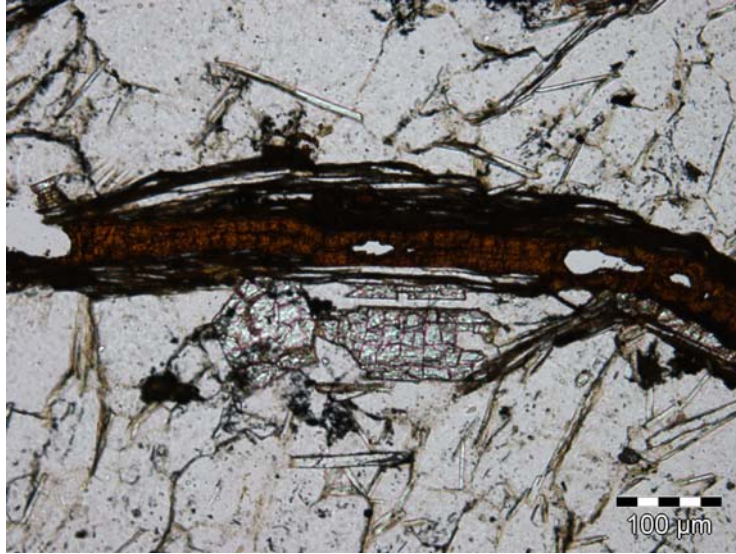
*Figure59: Sample 04-11: Co-existing of kyanite and sillimanite.*



**Figure 60: P-T conditions of kyanite and sillimanite in a stable state (Klein, Mineral Science, 1993)**

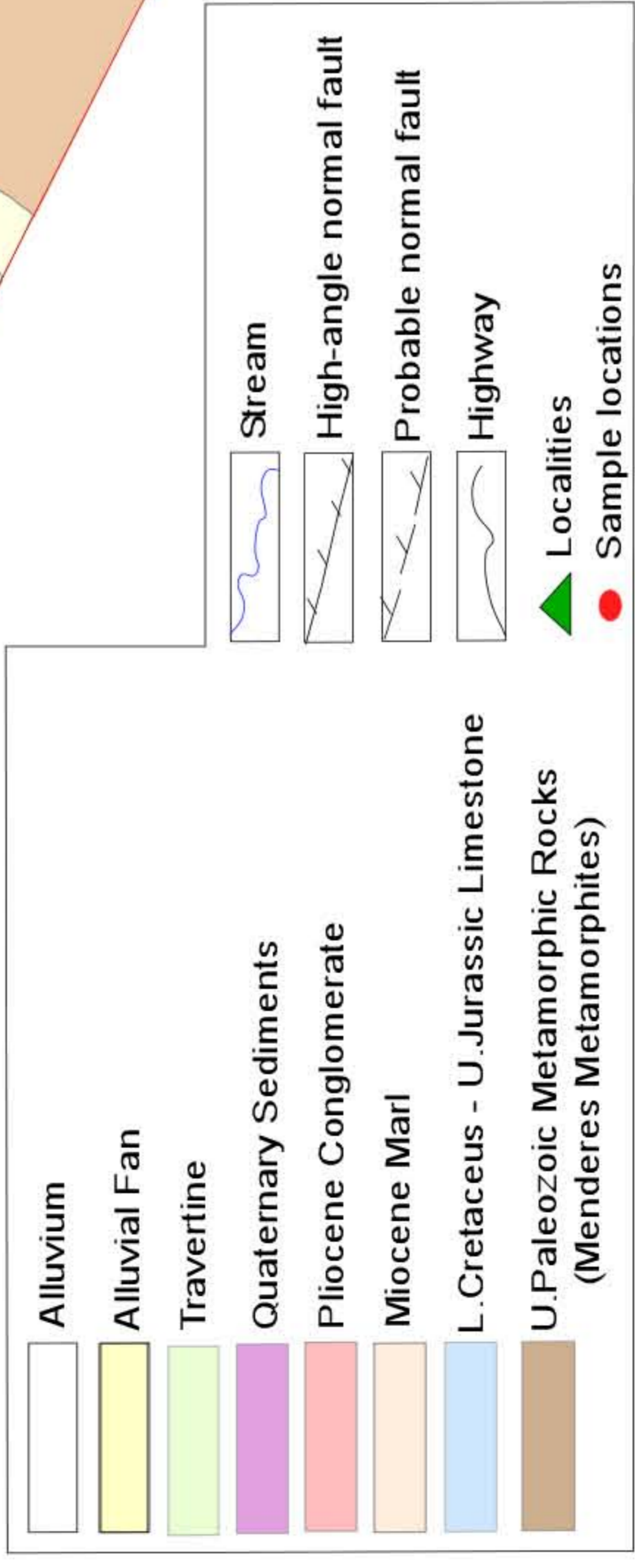
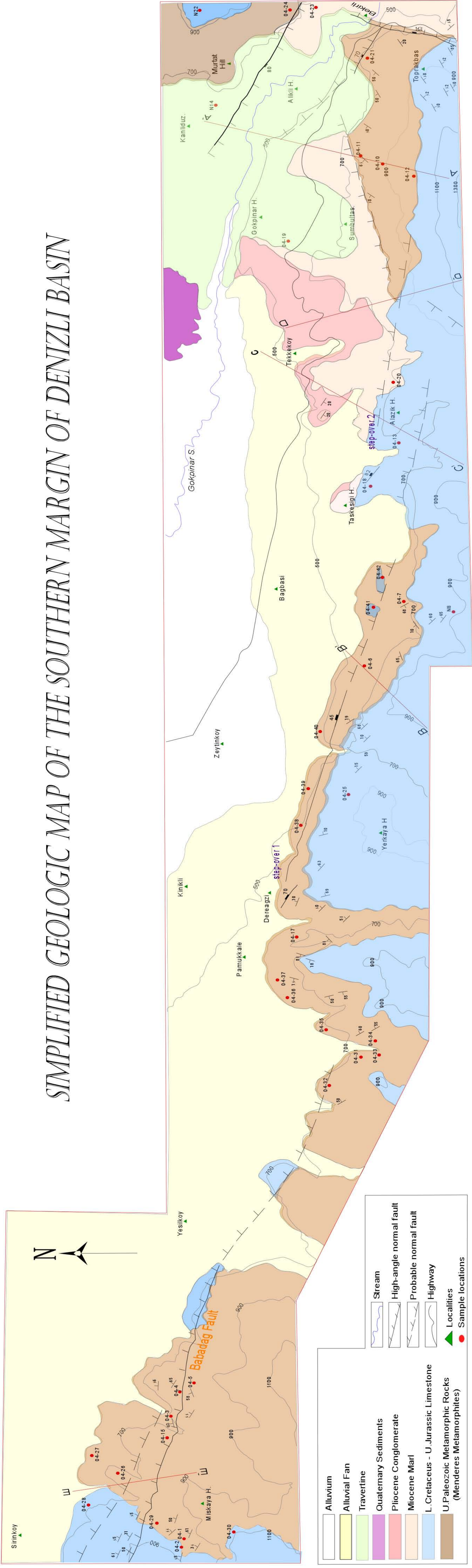


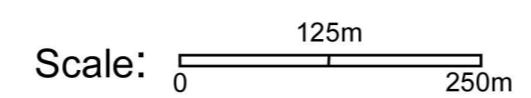
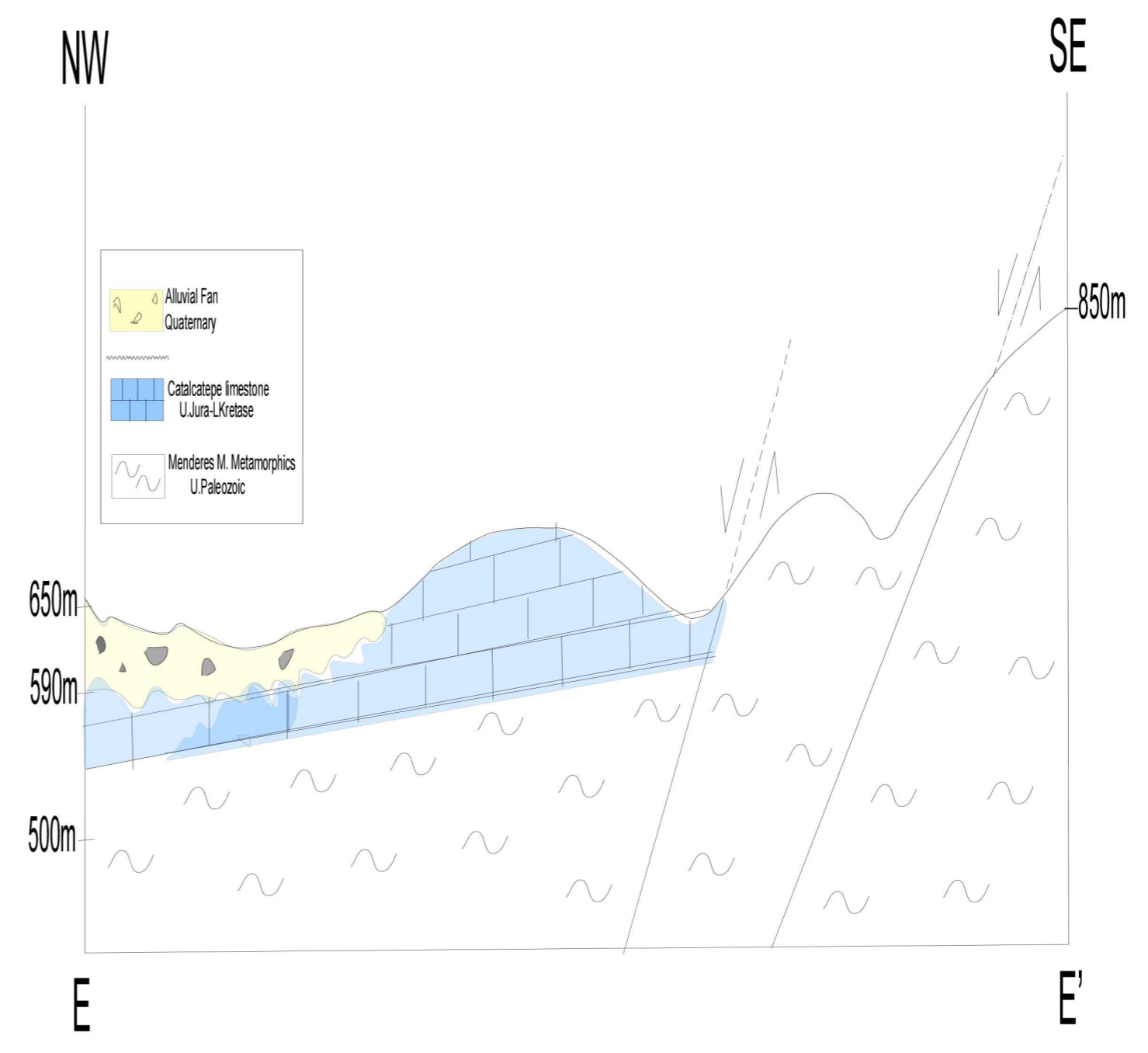
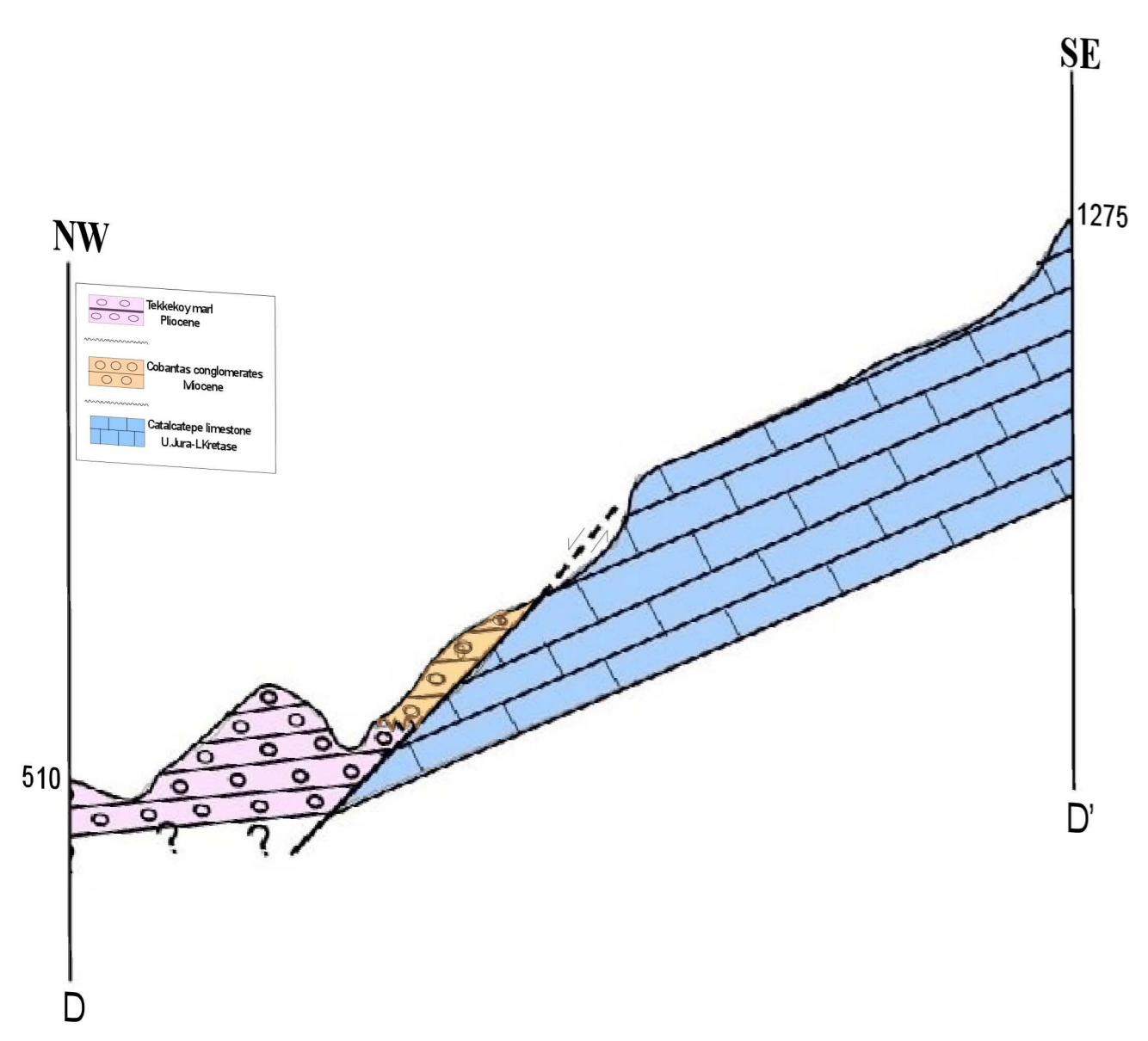
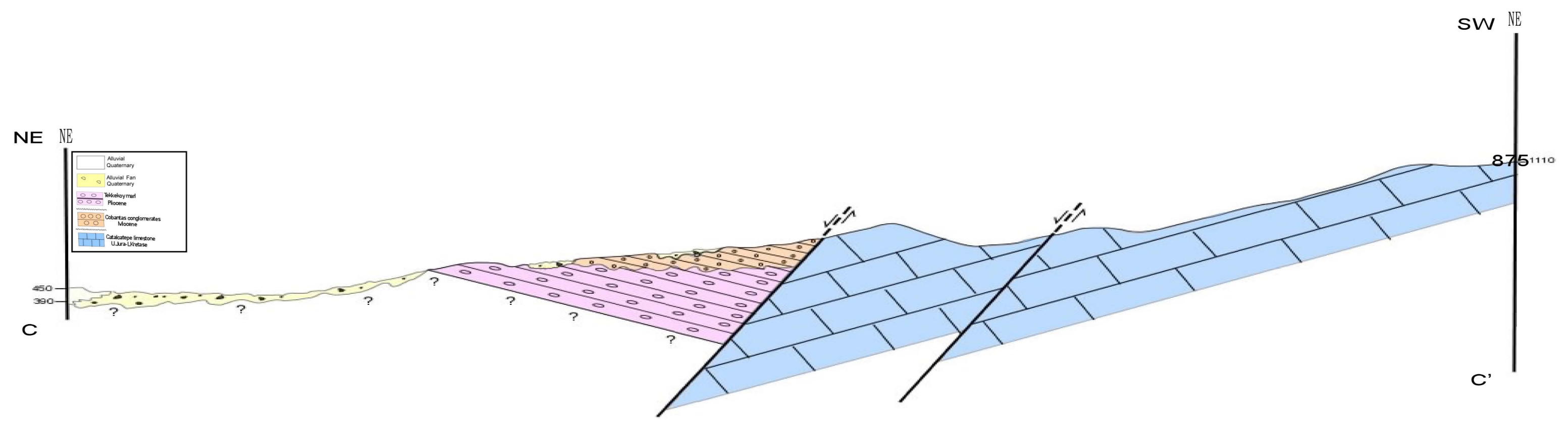
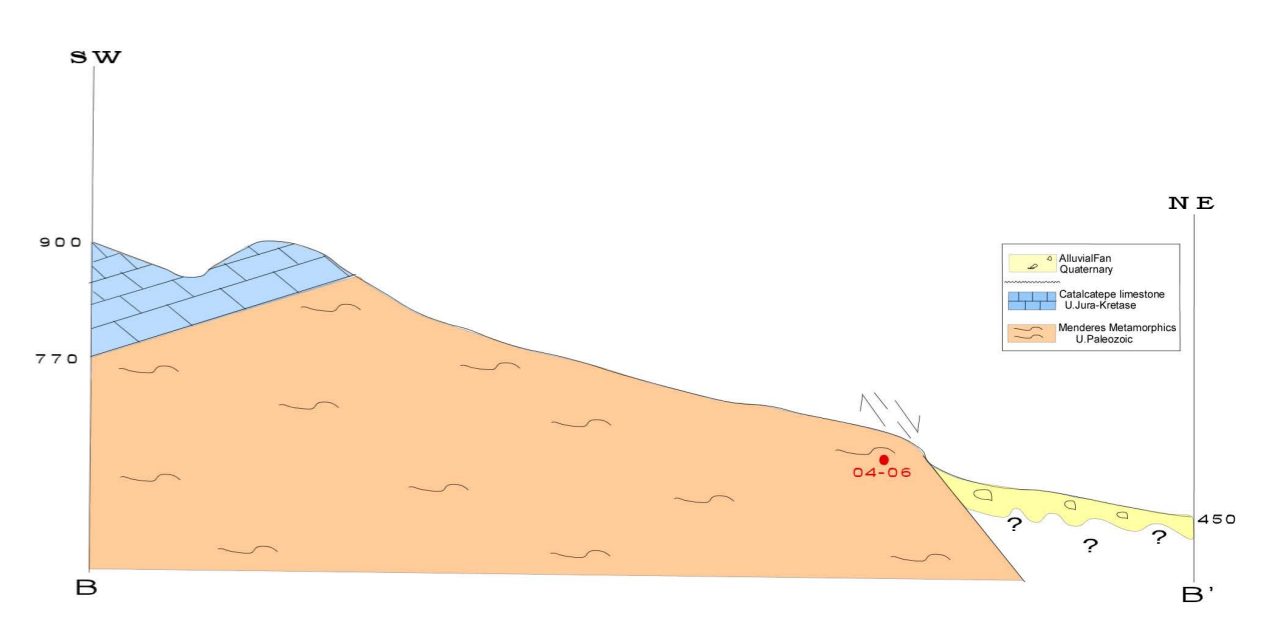
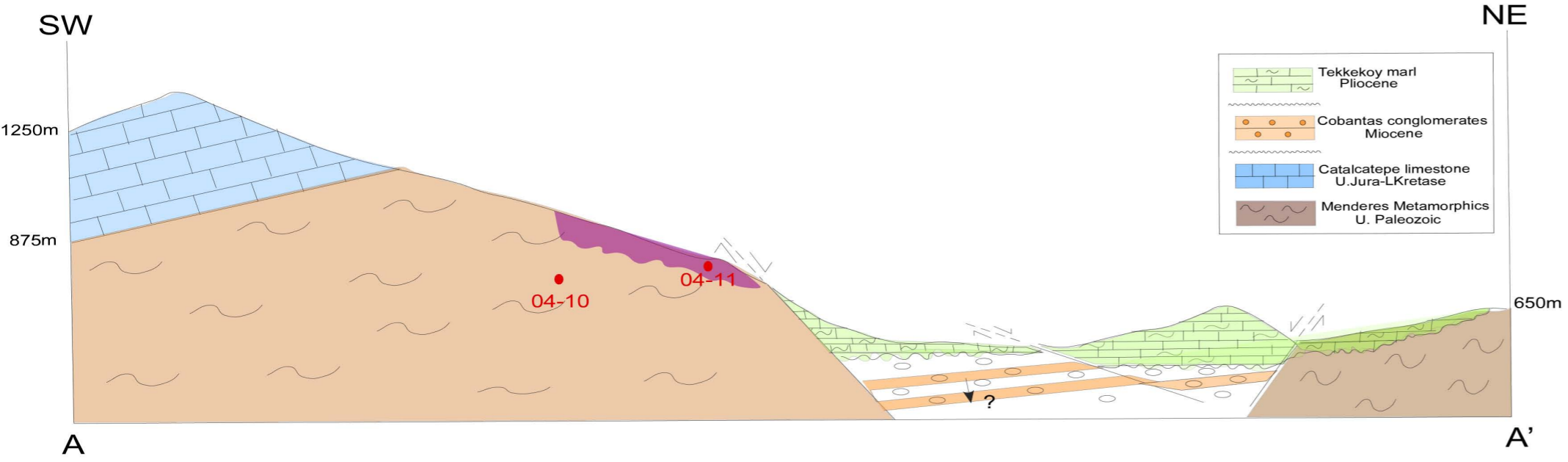
**Figure 61: Sample (04-03) GPS, 79201; 79922. Sphenes in kyanite biotite schist indicates that sphenes were produced because at high temperatures and pressure.**



***Figure 62: Sample 04-05: GPS, 79198; 79801. Rutil in kyanite biotite schist indicates high pressure and temperature conditions.***

# SIMPLIFIED GEOLOGIC MAP OF THE SOUTHERN MARGIN OF DENIZLİ BASIN





# VITA

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Location: Stillwater, Oklahoma

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IN, DENIZLI GRABEN, SOUTH WESTERN TURKEY

Pages in Study: 88

Candidate for the Degree of Master of Science

Major Field: Geology

Scope and Method of Study: The main goal of this thesis is to present the geometry and structural evolution of the Babadag fault zone, which is located at the southern margin of the Denizli Basin and separates metamorphic rocks in its footwall from sedimentary rocks in its hanging wall. The study also investigates the shear sense indicators that were found along the footwall rocks of the fault zone.

Findings and Conclusions: The Babadag fault zone that controls the southern margin of the Denizli Basin is a steeply dipping normal fault. It contains two major step-over as displacement along the fault partitions westward. Deformation features in the Buyuk Menderes shear zone range from ductile-brittle and the early stage of the deformation during the extension was ductile. The fault rocks consisted ductile shear sense microstructures and brittle shear sense indicators, found in the fault rocks along the Babadag fault zone, indicate top to the north movement of the upper plate. These observations suggest to us that the prominent movement along the Babadag fault zone and its high-angle splays is top to the north.

Dr .Ibrahim Cemen

ADVISER'S APPROVAL: \_\_\_\_\_