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Syn-sedimentary deformation structures in the Early Miocene lacustrine deposits, the basal limestone unit, Bigadiç basin (Balıkesir, Turkey)

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ABSTRACT

In the Western Anatolian region, NE-SW, E-W directional basins were developed which were limited to the extension-related faults beginning in the late Oligocene to early Miocene. The fillings of these basins consist of fluvial – lacustrine deposits containing volcanic and volcaniclastic intercalations. These deposits include intensive local unconformities and soft sediment deformation structures. The filling of the Bigadiç Neogene Basin which is one of these basins, constitute base limestone unit, lower tuff unit, lower borate unit, upper tuff unit and upper borate unit. The base limestone unit composed of claystone, marl, limestone, dolomitic limestone facies was deposited in the deep lacustrine environment. The soft sediment deformation structures were defined in the base limestone unit, which outcroped in the south of Bigadiç. These are: slumps, rock falls, chaotic structures, clastic dykes, synsedimentary faults and breccia limestone. Deformation mechanisms are related essentially to the increase of slopes of layers, liquidization and fluidization. In the study area; regional tectonics, sedimentological data, and deformation structures are evaluated together, it is concluded that these structures are formed by tectonic and seismic (earthquakes related to tectonic origin and syndepositional magmatic activities).

1. Introduction

The extension that began in late Oligocene-early Miocene, the latter periods of the continuing collision following the closure of the northern branch of the Neotethys in the Western Anatolia region, has continued until today (Altunkaynak and Yılmaz, 1998; Westaway, 2006). This extensional tectonism has caused the development of metamorphic core complexes, the fault controlled NE-SW and E-W directional sedimentary basins (Figure 1) and the settlements of magmatic rocks (Savaşçın and Güleç, 1990; Seyitoğlu and Scott, 1994; Seyitoğlu, 1997; Altınkaynak and Yılmaz, 1998; Yılmaz et al., 2001). The extension, which was formed in Neogene in the Western Anatolia region, mainly affected the Menderes massive (Harris et al., 1994; Okay and Satır, 2000; Jolivet et al., 2013). These basins, which contain volcano sedimentary deposits, were developed by detachment faults defined in the Menderes massive. In the period during which the extensional tectonism is affective the NE directional Soma, Bigadiç, Demirci, Gördes and Selendi basins developed (Koçyiğit et al., 1999; Yılmaz et al., 2000; Bozkurt, 2000, 2003; Işık et al., 2003; Bozkurt and Sözbilir, 2004). The fillings of these basins, which unconformably overlie the pre Miocene basement, are represented by the fluvial-lacustrine deposits containing volcanic and volcanoclastic intercalations (Erkül and Tatar Erkül, 2010). These are generally the multi-staged basins (Sözbilir, 2007). The first stage, which represents the Oligocene- early Miocene period, constitutes the opening (formation) period of basins. However; the second stage (20-7 my) is the

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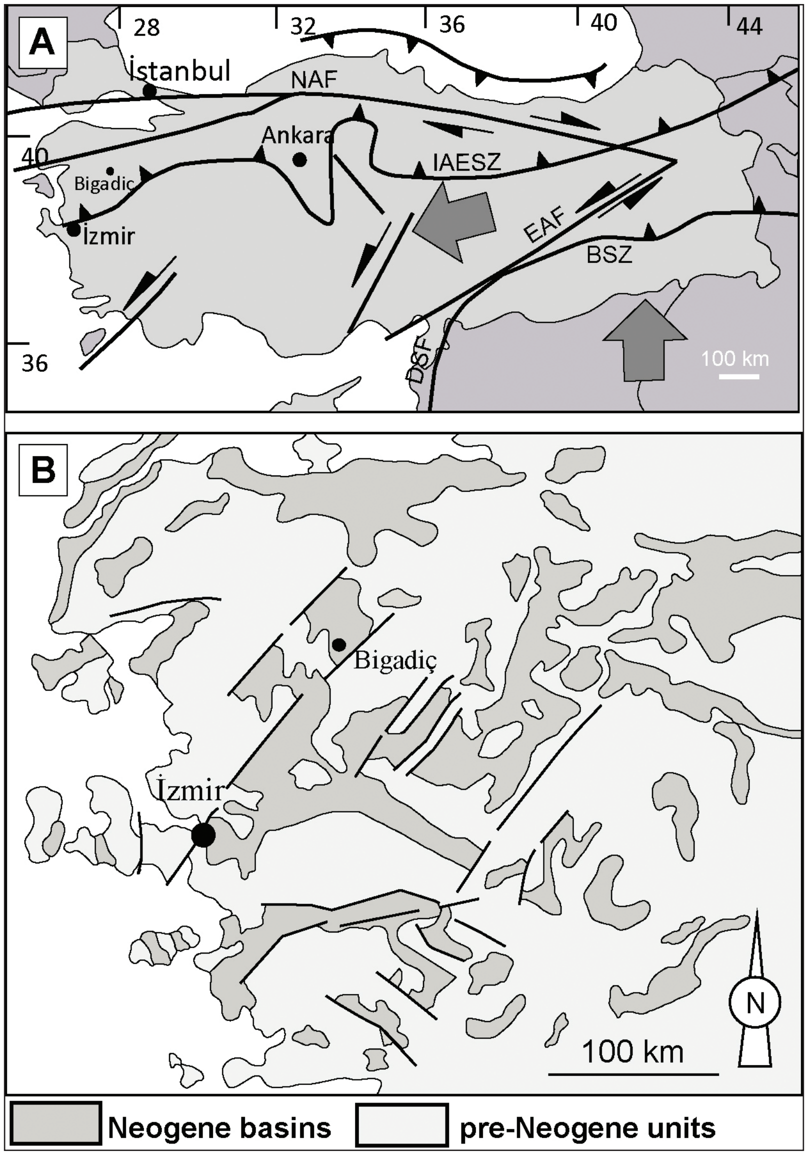

Figure 1- **a)** Main structural characteristics in Turkey; **NAF:** North Anatolian Fault; **IAESZ:** İzmir-Ankara-Erzincan Suture Zone; **BSZ**: Bitlis Suture Zone; **EAF**: East Anatolian Fault; **DSF**: Dead Sea Fault, **b)** Neogene basins in the Western Anatolia (modified from Garcia-Veigas and Helvacı, 2013).

period in which fillings of basins have developed. During the infilling of basins, the normal and slip faults in different angles have also accompanied to the sedimentation. In the last stage; E-W directional, normal and strike-slip faults have developed (Sözbilir, 2007). The Bigadiç Neogene basin, which was opened at the beginning of Late Oligocene-early Miocene, has been filled until the end of Early Miocene. During the sedimentation that controls this basin the tectonic events stated above have been effective and their traces have been observed in it. Besides; the locations and geometries of NE-SW directional slip faults (Figure 2), which cut basin infillings after the sedimentation, show that these are the continuation of faults that control the basin. The deposits here contain local stratigraphic unconformities associated with the tectonism controlling the basin (intra-formational unconformities). The dips of the upper tuff beds and overlying upper borate unit reach up to 80 degrees in some places. The radiometric age data obtained from volcanic rocks varying from basalt to rhyolite show that these basins have been under the effect of volcanism during early-middle Miocene (Erkül and Tatar Erkül, 2010). The unconformity and deformation

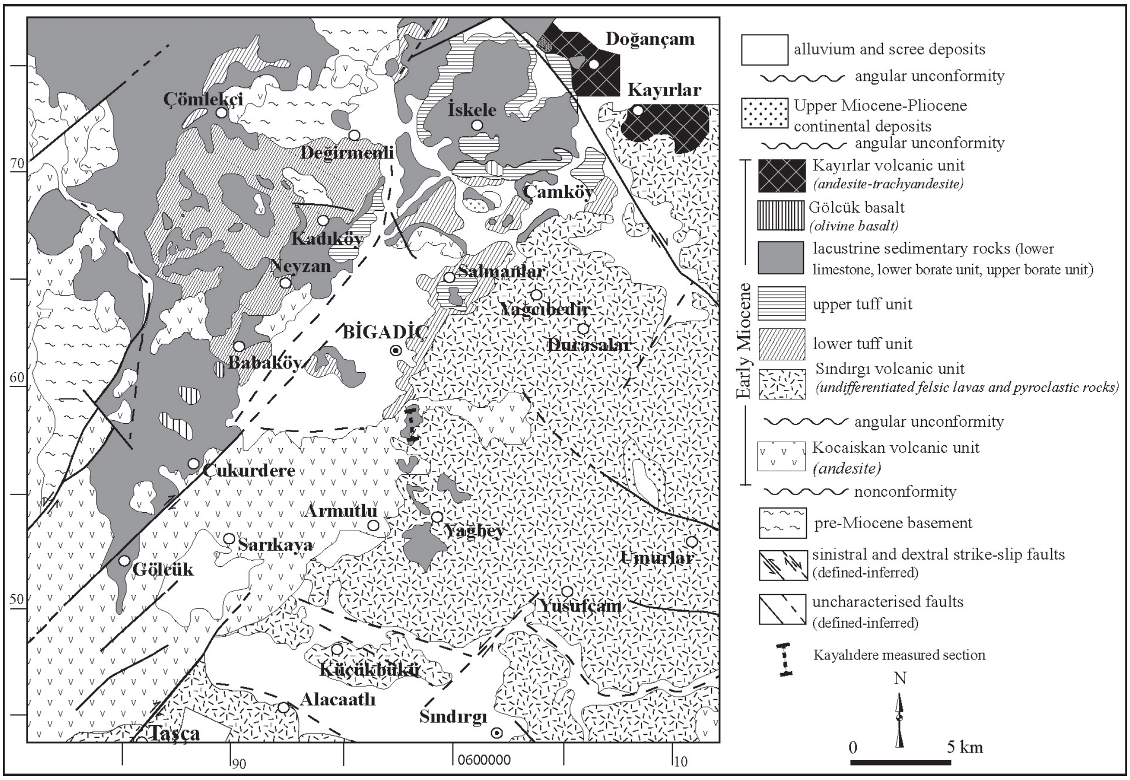

Figure 2- The geological map of the study area and its vicinity (modified from Erkül, 2006).

structures in these deposits can be associated with detachment faults controlling the development of basins. The deformation structures formed by the earthquake induced vibrations that are caused by these faults are defined as seismites (Seilacher, 1969). The seismites have been observed in many environments too (in fluvial, delta, lacustrine, etc.) (Owen, 1995; Gibert et al., 2005; Moretti and Sabato, 2007; Koç-Taşgın and Türkmen, 2009; Mastrogiacomo et al., 2012). The soft sediment deformation structures have been developed in the lacustrine environment in this study. The lacustrine environments are depositional environments, which most clearly reflect the results of seismic and tectonic activity during sedimentation as the deformation structure.

The purpose of this study is to define the soft sediment deformation structures, which were observed within the basal limestone unit located in the Bigadiç volcano sedimentary deposit in the Bigadiç basin, and interpret the formation mechanism.

2. Stratigraphy

The outcropping rock in the vicinity of the study area is the deformed Late Cretaceous- Paleocene flysch zone which is formed by big olistolith and ophiolitic blocks within chaotic sediments (Okay et al., 1996, 2001; Erkül et al., 2005*a*) (Figures 2 and 3). This unit is unconformably overlain by the early Miocene Kocaiskan volcanics, the Bigadiç volcano sedimentary deposit, the late Miocene-Pliocene terrigenous deposits and Quaternary sediments (Erkül et al., 2005*a* and *b*).The Kocaiskan volcanics cover an area of more than 800 km^2 and are the earliest products of the early Miocene volcanic activity in the Bigadiç region. In previous studies, it has been defined as the basal volcanic (Gündoğdu et al., 1989; Helvacı, 1995). The unit is formed by andesitic intrusions, pyroclastic rocks and volcanic origin sedimentary rocks (Erkül et al., 2005*a*).

Sındırgı volcanics, Gölcük basalts, Kayırlar volcanics and Şahinkaya volcanics constitute the volcanic units of the Bigadiç volcano-sedimentary deposit. However; the lacustrine units of this unit is composed by basal limestone, lower tuff, lower borate, upper tuff and upper borate units (Figure 2). Sındırgı volcanics are composed of dacitic and rhyolitic intrusions, massive and autobrecciated lava flows and pyroclastic deposits. Dacitic and rhyolitic rocks cover large areas in the eastern and southern parts of Bigadiç. Gölcük basalts spread out between

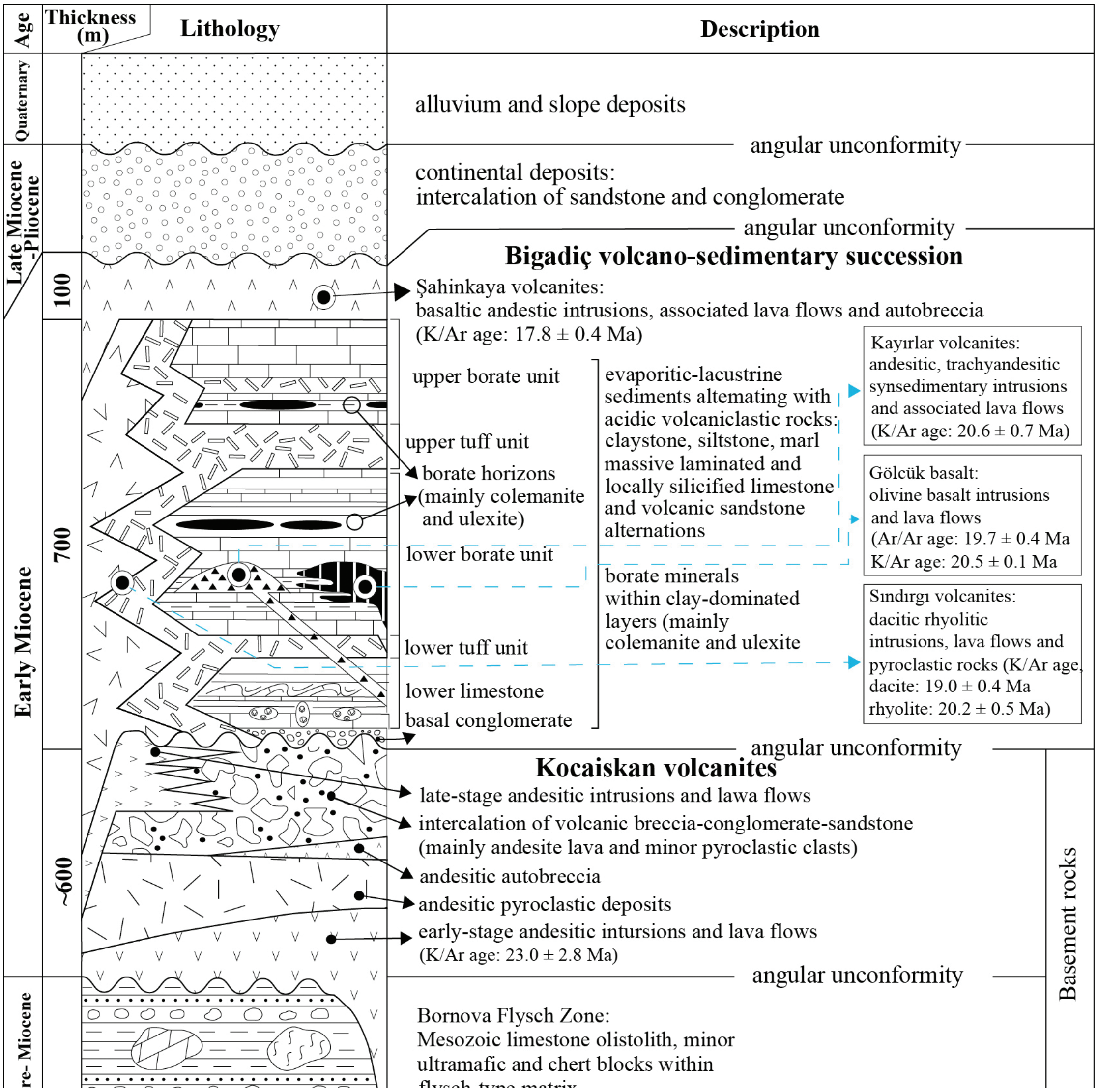

Figure 3- Generalized stratigraphic section of the study area (modified from Erkül, 2005*b*).

Gölcük and Babaköy, and are composed of olivine basalt dykes and volcanic domes in NE direction. It also spreads around Çamköy. Kayırlar volcanics are made up of trachyandesitic dykes, massive and auto brecciated lava flows in Doğançam and Kayırlar regions. Şahinkaya volcanics are basaltic and andesitic dykes, and are composed of lavas intercalating with dykes and auto brecciated pyroclastic rocks (Erkül et al., 2005*a*).

Lacustrine units were divided into 5 units and studied by Helvacı (1995). The same classification was followed in this study too. These units from bottom to top are; the units of basal limestone, lower tuff, lower borate, upper tuff and upper borate. The unit, which is represented by the intercalation of limestone, dolomitic limestone, claystone, marl and tuff, was named as the "Basal Limestone" and constitutes the main topic of this study. The unit unconformably overlies the Kocaiskan volcanics and is also conformably overlain by the lower tuff unit. The lower tuff unit a wide spread unit between Bigadiç and Cagis, is represented by coarse grained, thick layered (25 cm) or thickly bedded gray-white tuffs,which is up to 150 m thick. Economically important lower borate unit in the study area is composed of limestone, cherty limestone, tuffite, claystone and marls. The upper tuff unit is represented by "zeolitic" tuffs in the lower layers and as fine grained tuffs in the upper layers. The upper borate unit is represented by boron-claystone-limestone-tuff intercalation in the lower layers; claystone-limestone-tuff intercalation with organic material in the middle layers and by medium to fine grained laminated sandstone in the uppermost layers. In the same layers of the unit, the slumps and associated syn-sedimentary faults were developed.

The terrestrial deposits unconformably overlie the Miocene volcano-sedimentary deposits. The Upper Miocene-Pliocene red and beige sandstone and conglomerate layers of fluvial origin outcrops around Kocaiskan. These are unconformably overlain by unconsolidated clastic sediments of Quaternary age.

3. Sedimentological Characteristics of the Basal Limestone Unit

The basal limestone unit is represented by the intercalation of limestone, dolomitic limestone, claystone, marl and tuff. The measured thickness of the unit in this study is nearly 200 m (Figure 4). Mention the attitude of beds at proper positions in this paragraph. The deposit begins with dolomitic limestones that have much fractured and jointed structures and passes into banded tuff- cleavaged limestone-marl intercalation in the upper layers, and into claystone-limestone-tuff intercalation in the uppermost layers. The limestones are generally bedded and occasionally massive in character. The bedding thickness of cream to beige colored limestones is approximately 10 cm. They also have 2-3 cm thickness in some places and intercalate with tuffs. The marls are gray to green in color and observed as intercalating with tuff and limestone layers. They occasionally have the characteristics of conchoidal cleavage and consist of volcanic clastics in sand and pebble (1 cm) sizes. The medium to coarse silt size tuffs commonly exhibits lamination while at places they exhibit bedded nature also fractures and cracks in tuffs are filled with calcite.

The organic, laminated facies with carbonate clastics are composed of micritic carbonates, silt size clastic material and the intercalation of organic rich layers. It is considered that these were formed at the bottom sections of a less energetic, cold lake, which is not much saline, and presents seasonal bedding (Donovan, 1980). Similar facies (marl/limestone and mudstone/marl laminites) were interpreted as the perennial lacustrine deposits (Tanner, 2002). In these facies there were not encountered any evidence indicating shore (palustrine) or shallow regions (caliche for palustrine environments, desiccation cracks, wave origin structures for shallow environments, fossil diversity).

Sedimentological data obtained in this study indicate that the unit has sometimes been affected from the volcanism and deposited in deep lake environment. According to its relationship with volcanic units, the age of the unit was given as the Lower Miocene (Erkül et al., 2005*a*). Helvacı and Alaca (1991) detected the age of unit as the Lower Miocene according to its stratigraphic relationship with units at the bottom and top. Even though the washed samples of Basal Limestone unit yielded limited number of shells it was not possible to date and hence Lower Miocene age asigned by previous workers is followed in this study.

4. Soft Sediment Deformation Structures

Soft sediment deformation is a term used for the variation of fabric and layers of recently deposited sediments (Nichols, 2009). It is generally formed in granular sediments of which soft-sediment deformation structures are saturated with water. This strength loss is related with the liquefaction and/or fluidity of the water which develops as a result of the pore water pressure (Allen, 1982; Owen, 1987). In addition; the soft sediment deformation structures were also observed in carbonate rock deposits (Demicco and Hardie, 1994) and defined as seismite by some researchers (Weaver and Jeffcoat, 1978; Pratt, 1998, 2002; Kahle, 2002; Jewell and Ettenshon, 2004; André et al., 2004; McLaughlin and Brett, 2004).

Such structures were encountered in lakes (Sims, 1973; Hempton and Dewey, 1983; Scott and Price, 1988; Karling and Abella, 1992; Alfaro et al., 1997; Jones and Omoto, 2000; Rodriguez-Pascua et al., 2000; Moretti and Sabato, 2007; Koç-Taşgın and Türkmen, 2009), in deltaic environments (Gibert et al., 2005; Owen and Moretti, 2008), in shallow marine and tidal environments (Johnson, 1977; Bhattacharya and Bandyopadhyay, 1998; Molina et al., 1998; Rossetti, 1999; Rossetti et al., 2000; Rossetti and Goes, 2000; Moretti et al., 2001; Spalluto et al., 2007; Mastrogiacomo et al., 2012; Chen and Suk Lee, 2013) and in fan delta deposits (Postma, 1983). Besides; there are experimental studies related with the formation of these structures (Kuenen, 1958; Nichols et al., 1994; Owen, 1996; Moretti et al., 1999).

Within basal limestone unit the soft sediment deformation structures in different types were defined in many layers of the Kayalıdere section (Figure 4). These structures are explained below in detail.

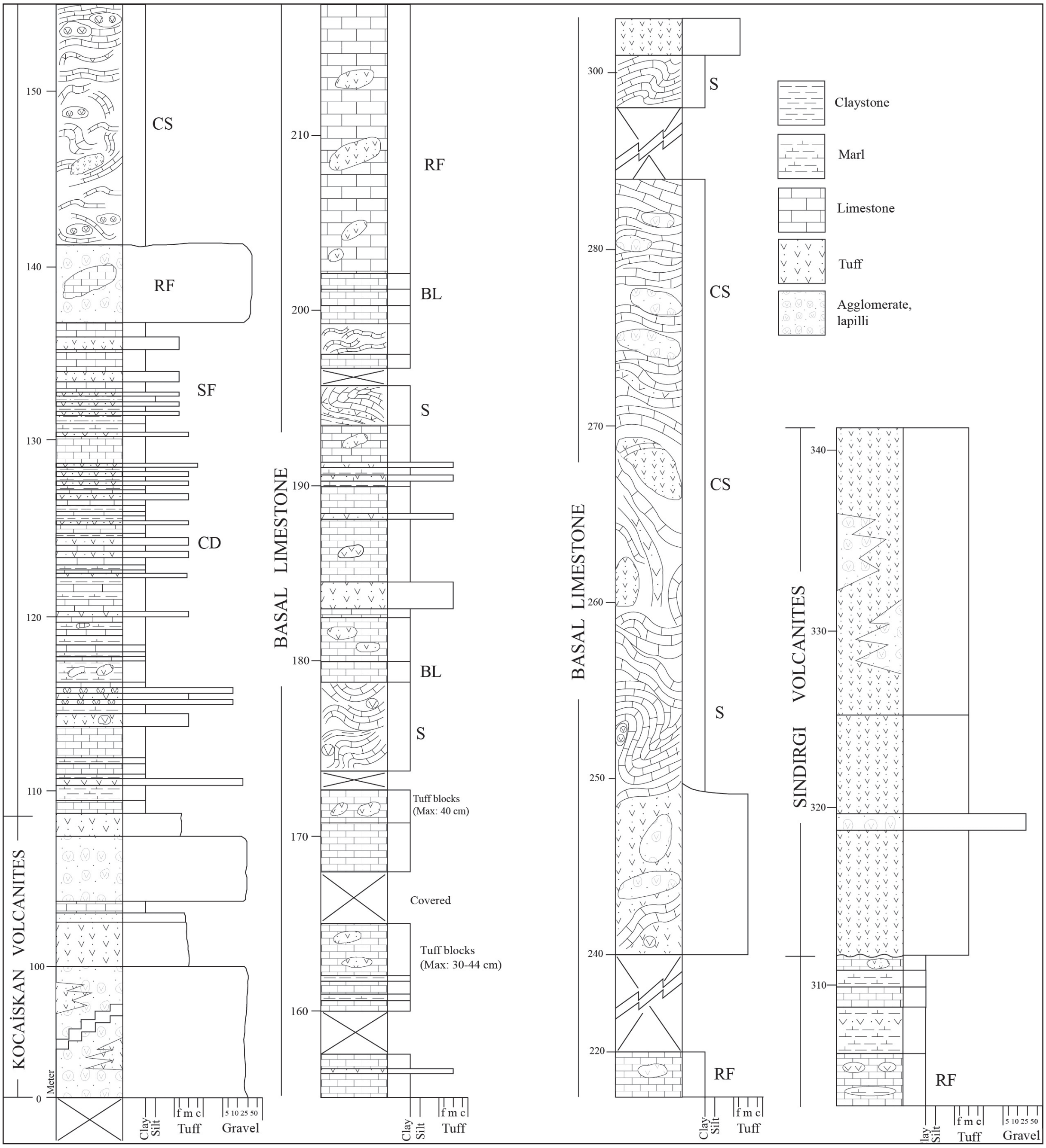

Figure 4- Kayalıdere measured section. Deformation structures are marked on the section. **S:** Slump Features, **CS:** Chaotic Structures, **RF:** Rock Fall, **CDS**: Clastic Dyke and Sills, **SF**: Syn-sedimentary Faults, **BL**: Brecciated Limestone.

4.1. Slump Structures

Slump structures were observed and defined along the road cuts around the Kayalıdere village where Basal Limestone units are well exposed (Figures 4-6). The structures,which are encountered at different levels of the unit, especially have affected bedded limestones. It has sometimes affected the thin bedded limestones and dolomitic limestone layers and sometimes tuff and marl layers, and formed slump structures of different dimensions. The size of small scale slump structures varies in between 20-100 cm. Also, the syn-sedimentary faults were formed towards the end portions of folds related with slump structures (Figure 5D). The slip amounts of these faults, which have the characteristics of inverse fault, are approximately 10 cm. These structures were observed in depth intervals

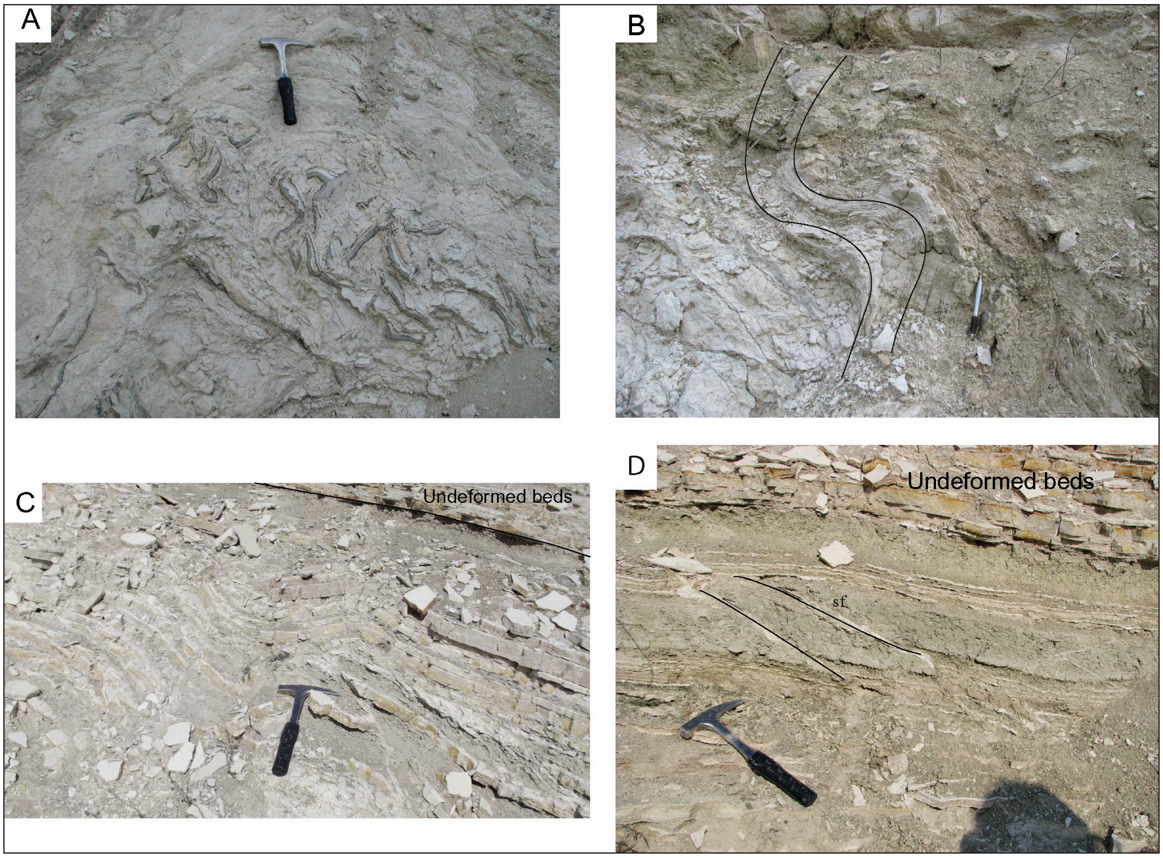

Figure 5- Small scale slump structures. These structures were developed in; **a)** dolomitic limestones **b)** bedded limestones and **c)** marl-limestone intercalation. **d)** syn-sedimentary inverse faults associated with slump structures.

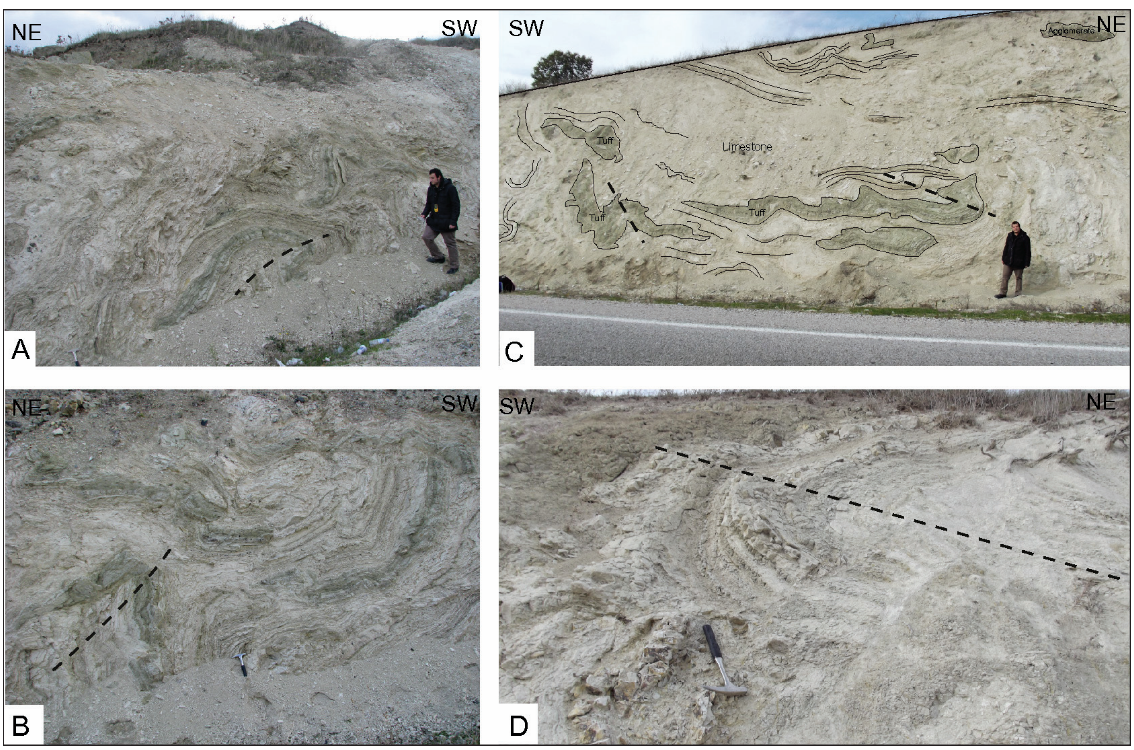

Figure 6- Large scale slump structures; **a)** and **b)** claystone, marl, limestone and tuff layers, **c)** tuff layers associated with limestones, **d)** slump structures observed in bedded limestones.

of 142-155 m, 174-179 m, 193-198.5 m and 240-301 m in the Kayalıdere measured section (Figure 4). The slump structures detected in limestones between of 142-147 meter are associated with chaotic sediments (tuff blocks and volcanic rock clastics). The layers are observed as bended and folded in different directions. The sediments here moved generally in the direction of SW (220°).

Interpretation: Slump structures develop due to the downward slip of the mass of sediment from the slope. These structures are characterized by inverse faults, isolated or continuous folds in the ends and by extensional structures over the head (Martinsen, 1994; Spalluto et al., 2007; Owen et al., 2011; Alsop and Marco, 2011, 2013). Slump structures are formed by steepening of slopes due to excess loading (due to rapid sedimentation), deposition (accumulation) and by earthquakes (Allen, 1982; Mills, 1983; Keefer, 1984; Owen, 1987; Van Loon and Brodzikowski, 1987; Moretti, 1996; Shanmugam, 2017). They may also occur due to the sloping of layers related with tectonic activities (faulting etc.) (Maltman, 1994*a*, 1994*b*; Siegenthaler et al., 1987; Mastrogiacomo et al., 2012; Perucca et al., 2014).

4.2. Chaotic Structures

Chaotic structures, which are observed in a couple of layers within the Basal Limestone unit (especially in the upper layers), are seen in the mixed form of slump structures and rock falls (Figures 4-7). The slump structures observed here have mostly moved in the direction of SW (210°-220°) and occasionally in the direction of NE (40°-45°). The planes of fold axes are horizontal, sub-horizontal and vertical. Slump structures influenced bedded limestones and tuff layers intercalating with them, and formed chaotic folds. The limestones consist of agglomerate and tuff blocks with sizes even reaching 3 m. The long axes of these blocks are both horizontal and vertical. These chaotic structures are either bounded by calcarous cement or tuffaceous material. It is also seen that these structures are occasionally associated with normal faults.

Interpretation: The complex or chaotic soft sediment deformation structures may occur in layers which have been affected by a couple of deformation phase. These deformation phases should have been repeated in short intervals. The deformation phase or phases, which follow the complexities formed by the main deformation phase (e.g. such as the aftershocks following an earthquake), could make the succession more complex (Mazumder et al., 2016). In the formation of chaotic sediments here, the faults controlling the basin and syn-tectonic activities such as volcanic and seismic activites associated with these faults should have been effective (Basilone et al., 2014).

4.3. Rock Falls

These generally occurs in pebble, fragment (mention size) and large blocks (mention dimension) of varying lithologies that recurs at different levels within the Basal limestone unit (Figures 4, 8 and 9). It is seen that these rock fragments are sometimes related with slumps and sometimes with chaotic sediments. There are also observed floating rock blocks within limestone. Within marl and limestone,

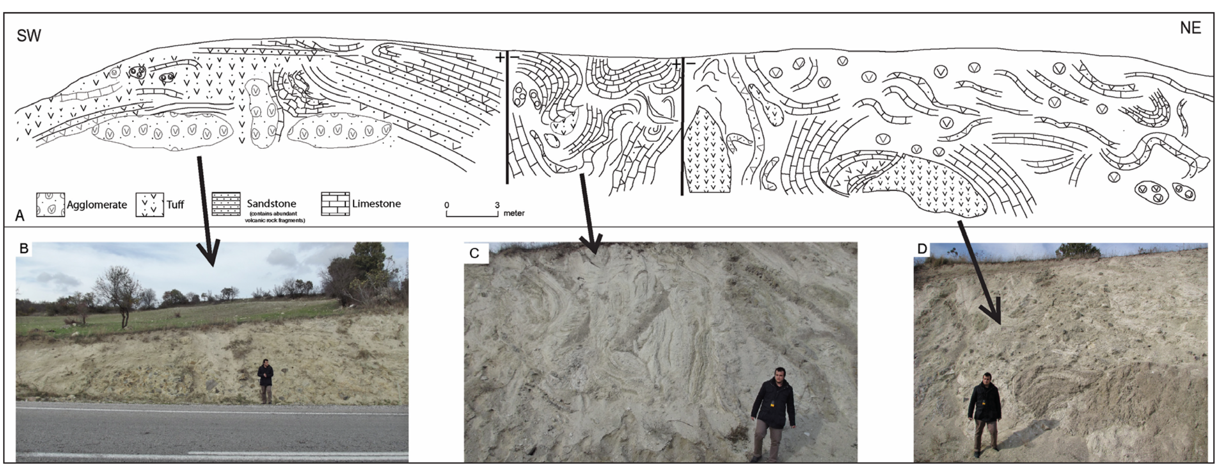

Figure 7- Chaotic structures, **a)** geological cross section showing chaotic structures, **b)** agglomerate blocks and slump structures, **c)** slump structures affecting the bedded limestones and tuff clastics, **d)** tuff block and slump structures.

the fragments and blocks of tuff and volcanic rock fragments (10-15 cm) take place. The size of tuff fragments vary between 20 and 100 cm. The size of agglomerate blocks observed in tuff reaches 6 m (Figure 7). Commonly the limestone beds below such agglomeratic blocks are observed to be deformed / buckled and sunken down/downwarped, similar to blocks (Figure 8a, b).

Interpretation: Rock falls are the most frequently seen mass movements associated with earthquakes. These movements occur on slopes at angles more than 40° (Keefer, 1984). They accumulate as colluvial or in tens of meters ahead the bottom section of steep slopes (Keefer, 1999). According to Montenant et al. (2007), the rock falls are gravity associated events originating from earthquakes. These rock fall deposits that reaches couple of meters in thickness could have been developed because of seismic event related with block faulting and extension effective in basin (Bozkurt and Sözbilir, 2004; Sözbilir, 2007; Erkül and Tatar Erkül, 2010).

Figure 8- **a)** and **b)** agglomerate blocks observed within limestones. Limestone layers under the block are deformed.

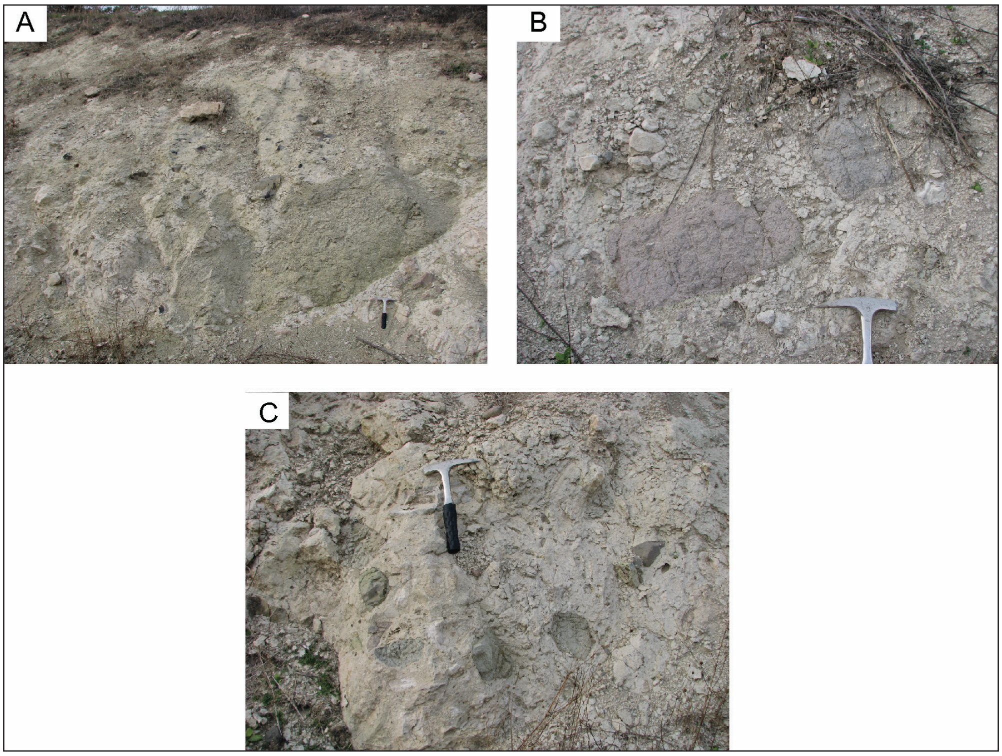

Figure 9- Rock falls in small scale, **a)** and **b)** tuff clastics, **c)** volcanic rock fragments.

4.4. Clastic Dykes and Sills

The clastic dykes observed in the study area occur between clayey limestone and fine to medium grained tuff. The deformation in question continues tens of meters laterally (Figure 10). The fine grained tuffs located in lower layers deformed limestones by intruding into them. The vertical length of dykes reaches 30 cm occasionally and that at places dike intrude through both limestone and medium grained tuff and intrudes in to the coarse grained tuff. Dyke formation began in the lower layer in the form of a very thin fracture and reached 7 cm thickness in maximum in the upper layers. The sills, which are the product of tuffs, show lateral continuity within limestones as connected with dykes.

Interpretation: The best indicator of liquefaction and fluidization as the soft sediment deformation structure is the water escape structures. For example; the dish and column structures, sand volcanoes, clastic dyke and sills (Mills, 1983). Dykes are generally formed by the upward transportation of sediments with pore water (Lowe, 1975, Owen et al., 2011; Mazumder et al., 2016; Onorato et al., 2016). The water escape structures are formed by the liquefaction and fluidization of the water in sands restricted by low permeable layers (Owen, 1987; Moretti and Sabato, 2007). Such clastic dykes could also be formed as a result of the upward movement of liquefied sediment under the pressure of upper layers (Daley, 1971; Rossetti, 1999; Montenant et al., 2007). Dykes and sills observed in the study area should have developed as being associated with the upward and lateral movement of tuffs as a result of liquefaction and fluidization (Rodriguez-Pascua et al., 2000).

4.5. Syn-sedimentary Faults

The syn-sedimentary fault, which is observed in the Basal Limestone unit especially affected tuff-limestone-marl facies, (Figure 11) and are generally in the characteristics of steeply inclined normal fault. Normal faults, which affect the deep lacustrine deposits, have formed horst and grabens in occasions. Over the layers of horst portions the breakdowns and detachments were developed. The net slip amount of the faults vary between 30 cm to 1 m.

Interpretation: The brittle deformation is associated with cohesive behavior of the sediment. When the pore water pressure in sediments increases and it is not strong enough to liquefy the sediment pressure then the brittle deformation occurs (Owen, 1987; Vanneste

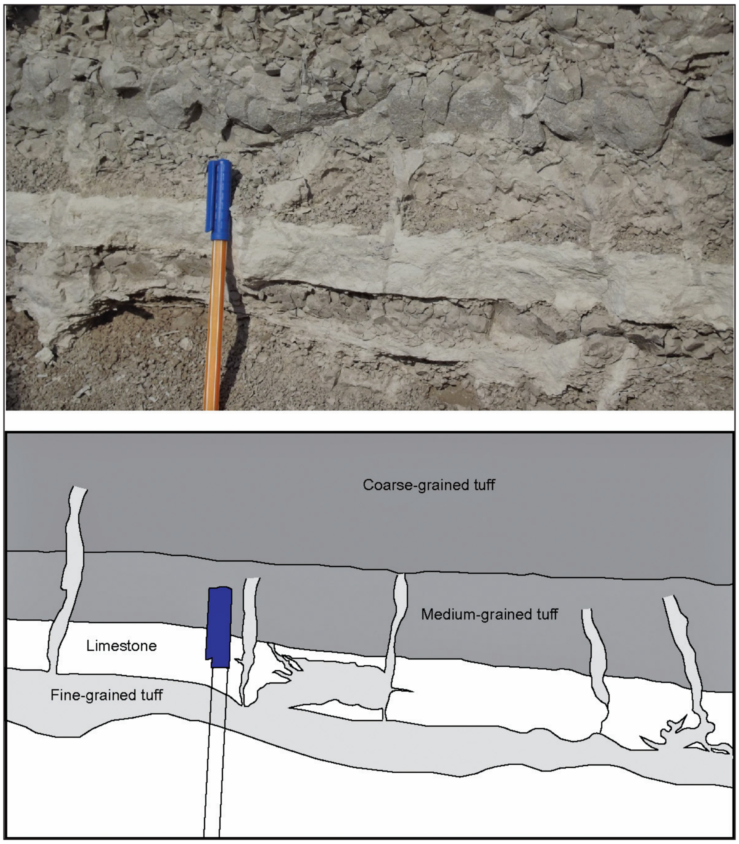

Figure 10- Clastic dyke and sills. Dyke and sills formed by the fine grained tuffs affected clayey limestones and medium to coarse grained tuffs.

Figure 11- Syn-sedimentary normal faults affecting the claystone, marl, limestone and tuff layers.

et al. 1999). Rosetti and Goes (2000) emphasize that these structures are associated with unconsolidated or poorly consolidated sediments. The structures investigated in the study area have developed after the partial consolidation of sediment.

4.6. Brecciated Limestones

The brecciation is observed in bedded limestones and partly in massive limestones (Figure 12). It has also affected limestone blocks which are observed in the form of rock fall. In brecciated layers, occasionally the angular limestone pebbles with sizes reaching 15 cm are observed. There were also observed brecciated limestone fragments within tuffs. These are generally grain supported.

Interpretation: The liquefactions, which are formed by the increasing pressure of entrapped water in pores of the early calcite cemented carbonate sediments cause brecciation (Clukey et al., 1985). Breccias defined in this study area associated with the liquefaction and some of the breccias observed in limestone blocks within tuffs should have developed during the transportation.

5. Triggering Mechanism

In order to detect the triggering mechanism, it is necessary to discuss all triggering mechanisms in the light of paleo-environmental analyses.

The presence of steeply inclined slopes is important for the formation of slump structures. The facies analyses carried out in the succession during the study indicate that the depositional environment is flat or sub-flat. Though it is considered that steep slopes are the main factors for the formation of slump structures, these may also occur at low angle slopes (even at degree of 1°) (Shepard, 1955; Field et al., 1982; Mills, 1983). It is stated that these structures, which occur on flat areas, are generally associated with paleoseismic activities (Bhattacharya and Bandyopadhyay, 1998; Rossetti and Santos, 2003; Spalluto et al., 2007; Garcia-Tortosa et al., 2011). Slump structures may occur due to excess load (related to the rapid sedimentation) (Allen, 1982). The entrapped waters among grains cause the increase in pore water pressure in next periods and the grains to become weak during rapid sedimentation. There was not observed any sudden coarse grained facies entrance in the study area. In poorly consolidated sediments, the most probable reason for the formation of slump structures is the increase in slope angles (steepening). Slump structures are formed when the bedded layers are inclined enough to exceed the stability limit. The slope increase in layers develops due to the deposition and tectonic movements. At the same time; the erosions, which are formed by water flows or turbiditic currents, may cause the increase in the slope angle (Mills, 1983). The facies characteristics and environmental data of the study area indicate that the deposition and current activities are not effective in the development of structures here. Slump structures here should have been formed as a result of increase in slope angle with the effect of seismic activities. The tremors, which occurred as a result of seismic shocks and/or volcanic activities, might have caused the decrease in the cohesion of sediments in inclined layers and sliding.

The formation of deformation structures such as chaotic sediments and rock falls are associated

Figure 12- Brecciated limestones.

with seismic and tectonic activities (Keefer, 1999; Montenant et al., 2007). The normal faults, which affected the basin during sedimentation, caused topographic reliefs in the basin. These rises caused rock falls in block and fragment sizes belonging to basal volcanics.

The liquefaction of buried layers begins with seismic activity (Clague et al., 1992) and the groundwater movements can cause these layers to be fluent (Guhman and Pederson, 1992). However; the hydraulic tension, which develops depending on the instant periods of the groundwater, widely causes the formation of local structures in young sediments. The dykes defined in this study show continuity in tens of meters. The rapid sedimentation may cause the formation of sand dykes (Parize and Fries, 2003). Facies overlying the dykes in this study show that these are not related with the rapid sedimentation. The sand dykes may also be formed by big storm waves (Martel and Gibling, 1993). However; the probability of big storm movements to be effective is weak in relatively deep lacustrine environments. The periodical tensions that are formed by seismic waves cause the pore water pressure to increase and the liquefaction (Owen and Moretti, 2011). The mechanism, which initiates the formation of sand dykes here, may be associated with seismic shocks (Mills, 1983; Audemard and De Santis, 1991; Obermeier et al., 1993; Obermeier, 1996; Rodríguez- Pascua et al., 2000) and/or tremors caused by the volcanic activities. The tremors, which are related to volcanic activities and frequently control the basin, should have initiated the liquefaction and fluidization event (Samaila et al., 2006; Tian et al., 2014; Zhou et al., 2017).

The sedimentological characteristics of deposits, their abundances and relationships with other deformation structures, which were formed by small scale normal faults show that these are developed based on the seismic activities (Vanneste et al., 1999). The normal faults known in the study area indicate that the region is controlled by an extensional tectonic movement. Syn-sedimentary faults with normal character are compatible with the regional tectonism. In other words; the syn sedimentary faults in the study area should have developed as associated with seismic movements due to the extensional tectonic activity in the region.

The breccias observed in limestones deposited in marine environments were developed by big storm movements (Seguret et al., 2001; Chen and Lee, 2013). The limestones deposited in the lacustrine environment have weak probability to get influenced from big storm movements. The observation of brecciation in footwall blocks in places indicates that these are associated with both transportation and seismic activity.

The liquefaction can be initiated depending on several factors. These factors affect the deposition environment both externally (allogenically) and internally (autogenically). The allogenic factors are tectonic movements and earthquakes. The factor affecting the depositional environment internally are autogenic in character, and these are; the wave motions, strikes due to the breaking of waves, stormy pressure vibrations in strong water flows, shear tension due to tsunami and tidal movements, rapid sediment rise, glacial melting in badly drained sediments or the groundwater movements (Owen and Moretti, 2011). There was not detected any evidence supporting autogenic factors that could initiate the formation of soft sediment deformation structures in the Basal Limestone unit. In other words; it seems quite difficult to associate these deformation structures, which developed in the lacustrine environment, with the triggering mechanism such as the shear tension related to wave motions, tsunami. So; in this case, the allogenic factors (tectonic movements, volcanism and earthquakes) should have been effective in the formation of deformation structures observed in the study area.

It is known that the faults associated with extension, which began in late Oligocene-early Miocene in the Western Anatolia region, are very effective during the formation of NE-SW and E-W directional basins and the deposition of volcano-sedimentary deposit. The sedimentation in the Bigadiç Neogene basin was controlled by tectonism and volcanism (Helvacı and Alaca, 1984, 1991). There are many and significantly large faults in the region. The step faulting system constitutes one part of these faults (Gündoğdu, 1982, 1984; Yılmaz et al., 1982; Baysal et al., 1985, 1986). During the sedimentation in the Bigadiç Neogene Basin, the NE-SW directional oblique slip, normal faults, strike slip faults and anticlines/synclines have developed (Erkül et al., 2005*a*). When the locations of faults and the characteristics of the basin fill are studied, it is seen that these faults are the basic structures controlling the development of the basin and one part of these continue their functions as syn sedimentary faults (Figure 13) (Baysal et al., 1986; Erkül et al., 2005*a*).It is seen that the sedimentation

Figure 13- Sinsedimanter normal faults observed in Kocaiskan volcanics.

in the basin, which developed in trans-tensional zone, is intensively accompanied by volcanism in addition to the faulting. During the sedimentation, the volcanism and dykes 100 m in width 2 km in length developed in the region together with intrusions (Erkül et al., 2005*a*). Accordingly; the earthquakes, which were formed as a result of magmatic activities synchronously with tectonics and deposition, should have been effective in addition to the tectonism, which is the main mechanism triggering the development of deformation structures here.

Seismic shocks may cause liquefaction and/or fluidization in unconsolidated sediments (Seilacher, 1969; Lowe, 1975; Sims, 1975). The tendency of seismic activities to form in the basin, which is restricted by fault, is higher (Mastalerz and Wojewoda, 1993; Bhattacharya and Bandyopadhyay, 1998; Koç-Taşgın and Türkmen, 2009; Koç-Taşgın, 2011; Koç-Taşgın et al., 2011). For the formation of liquefaction, the magnitude of the smallest earthquake should be greater than 5 (Audemard and De Santis, 1991). So; the earthquakes with magnitudes greater than 5 should have been effective during deposition in the region.

6. Results

In this study, the morphological characteristics of soft sediment deformation structures observed in the Early Miocene basal limestone unit around Bigadiç were established and formation mechanism was interpreted. In the lake, where the basal limestone was deposited, it was seen that both the tectonism and volcanism accompanied the sedimentation. Generally; tuffs and agglomerate levels in fewer amounts developed as being associated with the volcanism. Tectonic activities effective in the basin and earthquakes associated with tectonic and magmatic activities caused the formation of deformation structures.

The deformation structures restricted with undeformed layers from lower and upper layers and show lateral continuity in tens of meters (clastic dykes) indicate that these were developed in response to seismic activities. The structures defined in the study area show resemblance to seismic and tectonic origin deformation structures, which were defined by Seilacher, (1969); Moretti et al. (1999); Rodríguez-Pascua et al. (2000); Rossetti and Góes, (2000); Moretti and Sabato, (2007); Mastrogiacomo et al.

(2012) and experimentally approved by Kuenen, (1958) and Owen, (1996) in previous studies. It was determined that other factors (the shear tension due to wave motions, tsunami and tidal movements, rapid sedimentation, and groundwater movements), which could form deformation, were not effective in the study area.

In and around the study area, the soft sediment deformation structures were intensely observed in the Early Miocene lower and upper borate unit (Günen and Varol, 2004; Koç-Taşgın and Türkmen, 2014). This situation indicates that tectonic, seismic and associated magmatic activities in the region (Erkül et al., 2005*a* and *b*) have continued during periods when these sediments had been deposited.

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