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The control of sea-level changes on sedimentation in the Mut Basin: Late Serravallian-Early Tortonian incised valley-fill

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Research Article

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ABSTRACT

Early-middle Miocene reefal limestones are overlain disconformably by late Serravallian-early Tortonian incised valley-fill deposits of Dağpazarı formation in the Mut Basin. Dağpazarı formation is composed of mudstone, siltstone, sandstone and conglomerates. Facies associations of the formation are: fluvial, lagoon, shoal-water delta, shoreface, beach and barrier island deposits. *Loxoconcha tumida* Brady and *Loxoconcha* sp. in the gray mudstones indicate the freshwater influence and decrease in salinity. *Hemicyprideis* sp. documents brackish water conditions. The shoreface sandstones alternating with the lagoonal mudstones, and oyster-rich beach deposits in different levels of the sequence indicate episodes of marine connections. The age of the Dağpazarı formation is provided by the planktonic foraminifera from the marine mudstones and marls below and above the formation, and corresponds to the MMi8-MMi10 biostratigraphic interval, which spans the late Serravallian-early Tortonian. Late Serravallian eustatic sea-level fall caused to the quick shallowing of the Mut Basin and subaerial exposure of the reefal limestones at the basin margin. Thus, the incised valley, formed upon the reefal limestones of the Mut formation. This incised valley reflects a regional forced regression and unconformity. The Dağpazarı formation was deposited within this incised valley following an early Tortonian relative sea-level rise.

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1. Introduction

The Mut Basin in the Central Taurides consists of Oligo-Miocene terrestrial and marine deposits. In the basin, lacustrine carbonates and fluvial sediments

have been deposited during the Oligocene-early Miocene time. Although the marine sequence of late Burdigalian-Tortonian age in the Mut Basin is mainly composed of reefal and platform carbonates, an unconformity reflecting a late Serravallian forced

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regression can easily be defined because the Miocene sedimentary sequence of the Mut Basin contains well preserved sedimentary records of the relative sea level changes.

Mudstones, sandstones and conglomerates outcropping in vicinity of the Dağpazarı village in north of Mut, located on reefal limestones of the

Mut formation (Figure 1) were studied and mapped as the Köşelerli formation by Gedik et al. (1979). However, these rocks, which reflect terrestrial and transitional environments, were defined as the Sertavul formation by Demir (1997) as they differ from the Köşelerli formation in terms of lithology, depositional environment and age. Sertavul formation reflects back-reef lagoon environment and

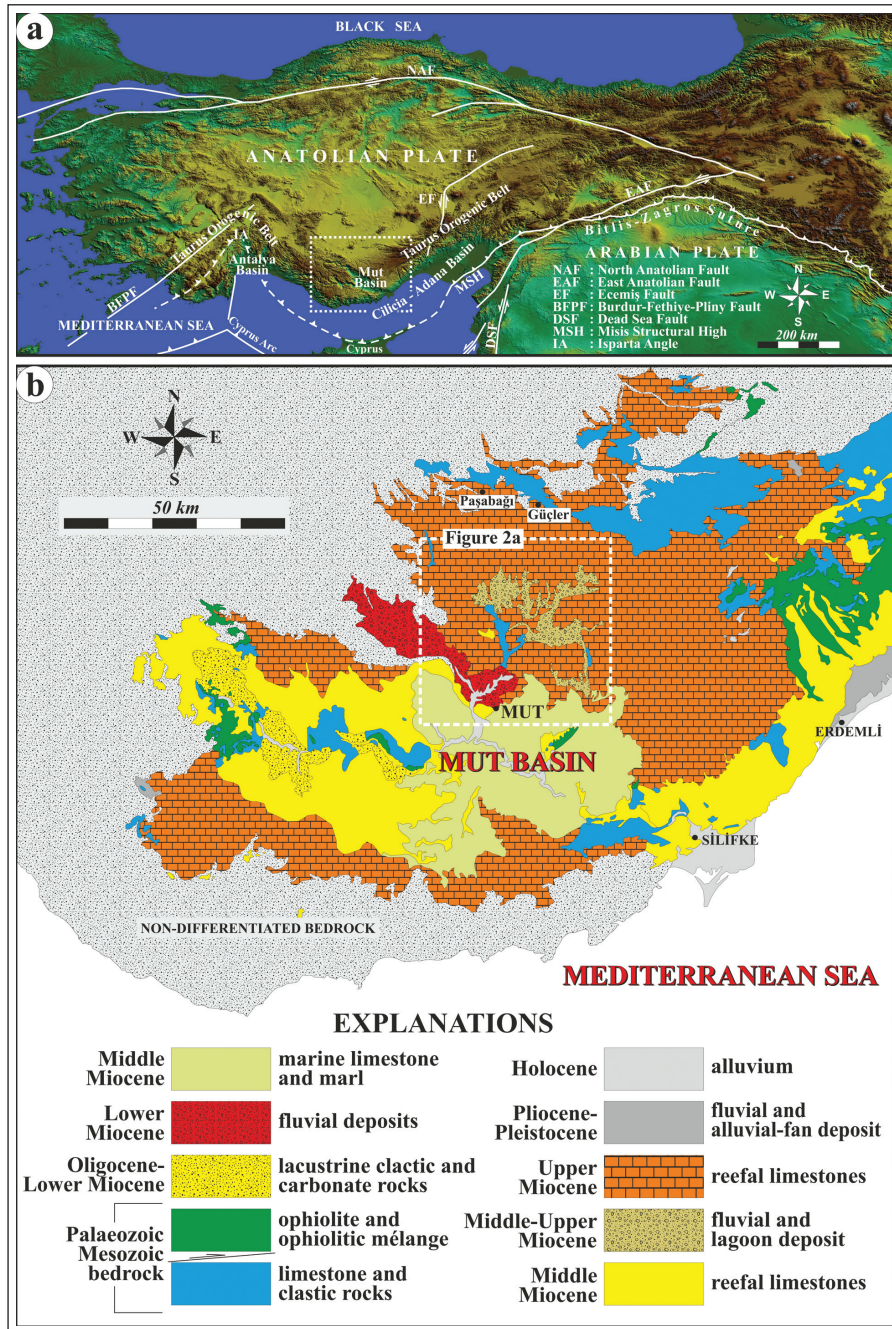


Figure 1- Location map of the study area, (a) location of the Mut Basin and main tectonic lines are seen on the topographical image of Anatolia (SRTM in 90 m resolution, from Jarvis et al., 2008), (b) simplified geological map of the Mut Basin (from Şenel, 2002 and Ulu, 2002).

occasionally fluvial deposits, coastal sands, coals and limestones (Demir, 1997). The same rock assemblage was defined by Atabey et al. (2000) as the Dağpazarı formation and interpreted as products of the back-reef lagoon and alluvial fan environments. According to Atabey et al. (2000), the Dağpazarı formation is a transitional unit with Mut and Köşelerli formations, and was deposited during a regressive phase. Bassant et al. (2005) document the early-middle Miocene sea level changes in the Mut Basin, but they did not present data on regional exhumation related to these changes. Although Cipporalli et al. (2013) did not traced sea level changes in the basin, they argued that there should be effects of sea level changes in the stratigraphic sequence.

The Dağpazarı formation, which is the subject of this study consists of clastic terrestrial and transitional facies, and disconformably overlies the reefal limestones of the Mut formation. This erosional surface forms a negative paleotopographic depositional environment within the Mut formation. The boundary between the Mut and Dağpazarı formations also reflects abrupt facies changes in the stratigraphic sequence, and a basinward erosional shift of the late Serravallian shoreline and the forced regression. Occurrence of proximal facies over distal facies across an erosional surface constitutes the most important defining criteria for incised valleys (Zaitlin et al., 1994; Hampson et al., 1997). The Dağpazarı formation shows the deposition in an incised valley according to its stratigraphic and facies characteristics. This incised valley and valley fill represents the most important records of relative sea level change occurred in the Mut Basin during the late Serravallian-early Tortonian.

Incised valley fills are important deposits in stratigraphic sequences, because these valley fills are depositional complex and represent both the formation of the valley and the deposition of sediments in the formed valley (Boyd et al., 2006). The formation of these deposits is controlled by the interaction between sea level changes, tectonism, climate, sediment discharge into the basin and paleogeomorphology (Posamentier and Vail, 1988). Therefore, the identification of these deposits in stratigraphic sequences is of great importance for interpretation of stratigraphic relationships, paleogeographic reconstruction and understandings of depositional evolution (Archer and Feldman, 1995; Boyd et al., 2006).

The aim of this study is to explain the geological history of the late Serravallian-early Tortonian sea level changes in the Mut Basin and to discuss the relative roles of processes such as eustatic, local tectonic and the amount of sediment transportation that control these changes. For this purpose, the sedimentary facies analysis of an incised valley fill sandwiched between Miocene marine carbonates analyzed and dated with paleontological data.

2. Terminology

The sedimentological terms used in this study were described according to Harms et al. (1975, 1982) and Collinson and Thompson (1982). For the identification of planktonic foraminifer species, Kennett and Srinivasan (1983), Iaccarino (1985) and Bolli and Saunders (1985) were used. In biozone definitions, Sprovieri et al. (2002) and Iaccarino et al. (2007)'s "Mediterranean Planktonic Foraminiferal Biostratigraphy" was taken as the basis and biostratigraphic age was evaluated according to ATNTS2004 (Lourens et al., 2004) scale (Lourens et al., 2004).

The term "regression" denotes seaward displacement of shoreline and this leads to an increase in land areas (Posamentier and Vail, 1988; Posamentier et al., 1992). Regression reflects the interplay between the relative sea-level change and the supply of sediment to the shoreline. This interaction causes normal or forced regression (Posamentier et al., 1992; Posamentier and Morris, 2000). Normal regression refers to the shoreline displacement towards the sea due to high sediment supply during a relative sea-level stillstand or slow rise. The forced regression means the basinward shoreline displacement due to relative sea-level fall. The amount of sediment supply to the shoreline during the forced regression is not significant. The forced regression occurs due to eustatic sea level fall or a tectonic uplift.

An incised valley is a bigger, longer topographic depression than a single river channel excavated by fluvial erosion. Abrupt facies changes occur basinward along the bottom of these valleys (Zaitlin et al., 1994). Incised valleys are filled with sediments transported from rivers and seas (Dalrymple et al., 1992; 1994) and include deposits accumulated under the control of tidal, wave and fluvial processes.

The terminology of sequence stratigraphy was used as suggested by Catuneanu (2006). A relatively

conformable succession of genetically related strata bounded by unconformities are called as sequence (Mitchum, 1977). Each sequence is composed of system tracts and parasequences. A conformable succession of genetically related strata bounded by transgression surfaces form parasequences

In this study, three different system tracts were defined as; the “forced-regressive systems tract”, “lowstand systems tract” and “transgressive systems tract”. System tracts were defined according to the vertical stacking pattern of the sedimentary facies associations and the direction of displacement of the palaeoshoreline (Helland-Hansen and Martinsen 1996). The “forced-regressive systems tract” occurs during a relative sea level fall. The “lowstand systems tract” is a normal regressive coastal progradation deposited during a relative sea level fall and then remains stable. The units deposited during the rise of the relative sea level constitute the “transgressive systems tract”.

3. Regional Geological Setting

The Taurides located in the eastern Mediterranean part of the Alpine-Himalayan mountain belt and the southern part of Turkey are studied by dividing them into three sections as; the West, East and the Central Taurides (Figure 1a). The Western Taurides extend from the Isparta Angle to the west towards the Hellenides. The Central Taurides are between the Isparta Angle and Ecemiş Fault. The Eastern Taurides extend from the Ecemiş Fault to the east into the Zagros Mountains. The orogeny forming Taurides continued until the end of Eocene in the Central Taurides (Şengör, 1987; Clark and Robertson, 2002). However, the compressional tectonics, indicating the latest movement of nappes, has continued until middle Oligocene (Kelling et al., 1987; Andrew and Robertson, 2002). In the Eastern Taurides, the orogenic deformation has continued until the late Miocene and the Misis Structural Uplift has been formed due to the folds and thrusts (Michard et al., 1984; Aktaş and Robertson, 1990; Dilek and Moores, 1990; Yılmaz, 1993; Yılmaz et al., 1993; Robertson, 2000; Sunal and Tüysüz, 2002). At the transition of the Western and Central Taurides, the Lycian Nappes collided with the Isparta Bend (Collins and Robertson, 1998, 2000; Poisson et al., 2003; Sagular and Görmüş, 2006). The Miocene thus saw the last stages of localised compressional deformation, while the Taurides in general had already become

subject to post-orogenic isostatic uplift and crustal extension with the development of orogen-collapse basins (Seyitoglu and Scott, 1991, 1996; Jaffey and Robertson, 2005; Karabıyıkoglu et al., 2005; Bartol et al., 2011; Koç et al., 2012; Cosentino et al., 2012). This orogen-collapse basins occurred in the form of grabens. The continental crust, which has excessively thickened during the Eocene period, caused the formation of the basin under the effect of gravity due to the disappearance of compression (Gautier et al., 1999; Dilek and Whitney, 2000). According to Le Pichon and Angelier (1981) and Gautier and Brun (1994), the extensional tectonic regime was formed by the back-arc spread as a southern retreat process of the northerly subducting Hellenic plate.

4. Dynamic Stratigraphy of the Mut Basin

The Mut Basin (Figure 1), located in Central Taurides, forms one of the molasse basins between the mountains during Neogene period. It is thought that Mut Basin was formed due to the orogenic collapse behind the extensional back-arc of the Cyprus arc in the south (Kempler and Ben-Avraham, 1987; Robertson, 2000; Kelling et al., 2001; Ünlügenç et al., 2001). The basin has then continued to develop under the influence of an extensional regime.

The bedrocks of the basin are Aladağ and Bozkır nappes emplaced in the late Eocene period (Özgül, 1976; Andrew and Robertson, 2002). The nappes are composed of Jurassic-Cretaceous limestones and Late Cretaceous ophiolitic melange (Figure 1b). These units are locally overlain by Eocene shallow marine limestones. Heavily eroded allochthonous bedrocks are overlain by the Oligocene lacustrine carbonates, which are called the Fakırca formation in the study area (Figure 2). These lacustrine units, which have limited lateral extent, form the initial productions of intramountain basins. Fakırca formation consists of thin to medium, planar to undulated bedded, planar parallel stratified and wave-rippled limestone and marl alternation. Pine needles, leaf fossils and *Planorbis* type gastropod fossils commonly observed in limestones and marls indicate the fresh water environment. Coal seams are observed at the bottom of the early-middle Oligocene Fakırca formation sequence (Tanar, 1989; Tanar and Gökçen, 1990).

Following the compressional tectonism that occurred towards the end of the middle Oligocene due to the latest movement of nappes the extensional

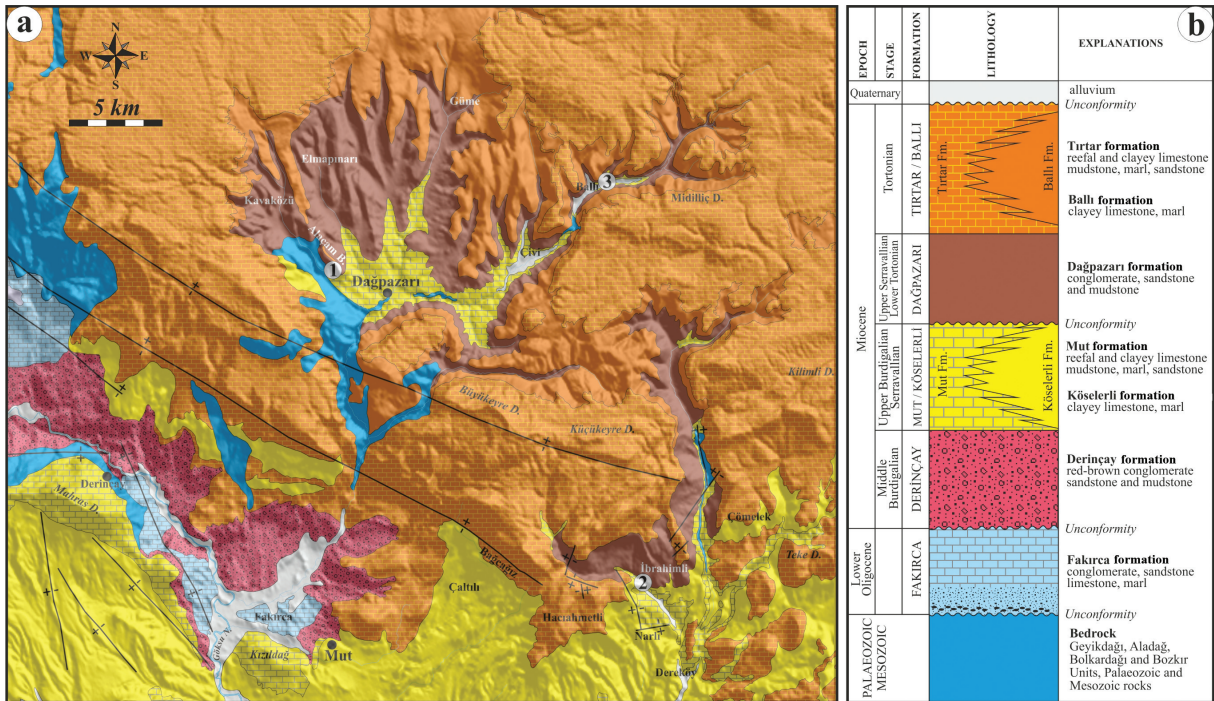


Figure 2- (a) Detailed geological map of the study area in the Mut Basin and the distribution of the Dağpazarı formation. Numbers 1-3 on the map indicate the section localities. 1. Alaçamburnu section, 2. İbrahimli section and 3. Ballı section, (b) Generalized stratigraphical section of the Mut Basin.

tectonic regime in the region has been dominant and the structural development of the basin was established. Due to continuous sedimentation in post Oligocene in the Mut Basin, the fluvial sediments of the Derinçay formation were unconformably deposited on the Fakırca formation in the early-middle Burdigalian period (Ünay et al., 2001) (Figure 2; Ilgar et al., 2016). Derinçay formation consisting of red conglomerate, sandstone and mudstone reflects the channel lag, sigmoidal point bar and flood plain deposits of a meandering river system. The marine transgression, which occurred in the late Burdigalian led to the drowning of Antalya, Mut and Adana basins and the development of the first marine deposition during the Neogene period in these basins. Thus, the reefal limestones, platform carbonates and the marl to thin bedded limestone forming the sediments of the Köselerli formation were deposited in the late Burdigalian–Serravallian (Figure 2). The shallow marine carbonate deposits of the Mut formation (Figure 3a), which forms onlapping structure due to the marine transgression on bedrocks towards the north (Figure 3a), pass into offshore deposits of the Köselerli formation towards south in the basinward direction (Figure 3b).

The relative sea level fall in the late Serravallian period led to the shoaling of the basin, migration of

the reefal limestones of the Mut formation on to the Köselerli formation in basinward direction and to crop out on the basin margin. The sedimentary facies belonging to the Dağpazarı formation, which is the subject of this study, were deposited in an incised valley generated during this forced regression (Figure 2). The early Tortonian transgression caused the re-flooding of the northern parts of the Mut Basin and the deposition of the reefal limestones of the Tirtar formation on basin margins and the marl to thin bedded limestones of the Ballı formation inside the basin (Figure 3c). In the late Tortonian, the marine sedimentation in the Mut Basin terminated with the isostatic uplift of the Taurides and the basin began to expose (Cosentino et al., 2012; Ilgar et al., 2013a, b).

5. Dağpazarı Formation and its Sedimentary Facies

The Dağpazarı formation mainly composed of mudstone, sandstone and conglomerate outcropping in the north of Mut Basin is located between the reefal limestones and carbonate platform deposits of the Mut and Tirtar formations and observed in the form of incised valley fill deposits in the reefal limestones of the Mut formation (Figure 3c). The Dağpazarı formation, which is defined within belt in the north-

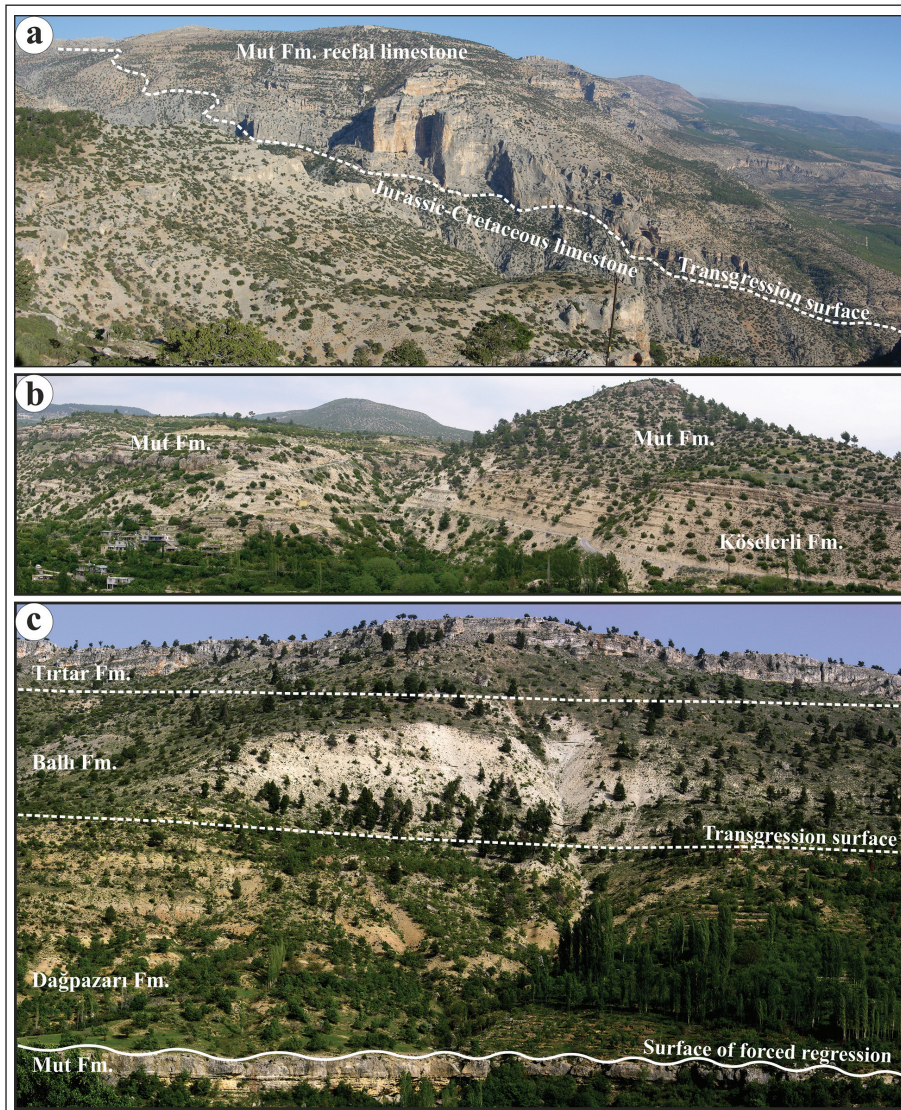


Figure 3- (a) Reefal limestone beds of the Mut formation onlapping on the basement rocks consisting of Jurassic-Cretaceous limestones, (b) Gradational transition of the Mut formation into marl-fine bedded limestones of the Köseleerli formation in the basinward direction. (c) The stratigraphical relations of the Mut, Dağpazarı, Ballı and Tirtar formations around Dağpazarı Village in the north of Mut. The fluvial deposits of the Dağpazarı formation unconformably overlie the Mut formation.

south direction, consists of several branches to the north of the basin. The valley fill is approximately 35 km long, 0,5-2 km wide and 130 m thick.

The Dağpazarı formation unconformably overlies the reefal limestones of the Mut formation or shoreface sandstones and shell-rich fossiliferous pebbly beach deposits which is occasionally observed at the topmost layer of the formation. It is transgressively overlain by the reefal limestone, shoreface sandstone and pebbly beach sands of the Tirtar formation and marl to thin bedded limestones of the Ballı formation (Figure 3c).

At the Alaçamburnu locality, 2,5 km to the northwest of the Dağpazarı village is the type section of the formation (Figures 2a and 4a). The surround of Çivi and Ballı villages (Figure 2a) and the north of the Büyükeyre Mountain (Figure 3c) in the study area allow to observe the contacts with underlying Mut formation and the overlying Ballı and Tirtar formations.

The Dağpazarı formation is mainly composed of dark gray mudstone, light brown-brick red mudstone and siltstone, very fine to fine grained sandstone and medium to coarse grained sandstone and granule to fine

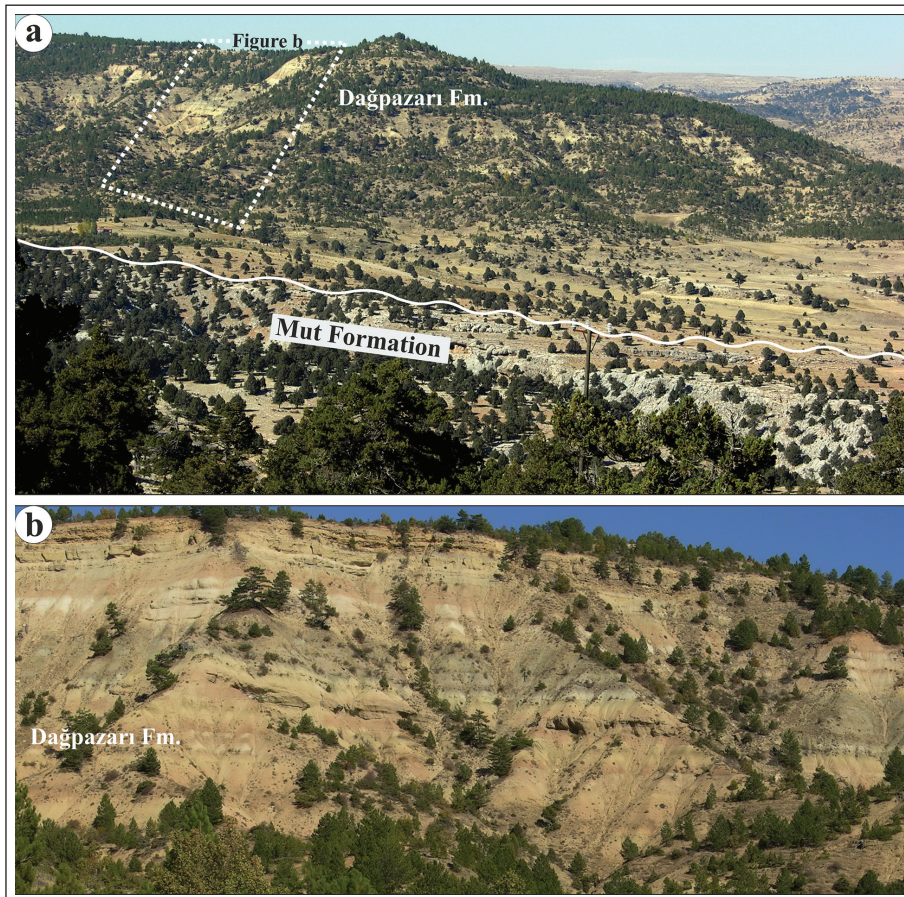


Figure 4- (a) The Alaçamburnu type section where the facies defined in the Dağpazarı formation are best observed. (b) Close-up view of the Alaçamburnu section. The fluvial, shoal water delta, lagoon, shoreface and beach facies associations were defined in this section.

pebble size conglomerate (Figures 4b and 5). In this formation, total of 6 facies associations were defined as fluvial, shoal water delta, lagoon, shoreface, beach and barrier island (Figure 5). These facies associations, which are lateral and vertical transitional, have variable lateral extent and thicknesses and are repeated several times in the succession. Since each facies association reflects the depositional environment and conditions, it is described under the headings of the environments and described below in detail.

5.1. Fluvial Deposits

The fluvial deposits defined in the Dağpazarı formation are mainly composed of pale brown-brick red mudstone, siltstone, sandstone and gray-pale brown sandstone, pebbly sandstone and conglomerate. Gray sandstones and conglomerates consist of bedsets which have maximum thicknesses of 2,5 m and do not have lateral continuity. These bedsets are located on pale brown-brick red mudstones and siltstones with a

sharp and erosional bottom contact (Figure 6a). The amount of erosional relief reaches up to 1 m. The upper surfaces of bedsets are planar or undulatory. Chutes are often observed at the bottom of these deposits. These gray sandstone and conglomerate bedsets show thinning and fining upward units (Figure 6a). These sandstones and conglomerates, which have the main sedimentary structures of planar inclined or sigmoidal beds, are interpreted as meandering river deposits (Figure 6a). Pale brown-brick red, thin bedded mudstones and siltstones alternating with meandering river deposits form floodplain deposits (Figure 6a and b). These facies grades each other within vertical succession and alternates. Sedimentological characteristics of the meandering river and flood plain deposits are mentioned below.

5.1.1. Meandering River Deposits

The meandering river deposits described in the Dağpazarı formation are composed of the channel lag

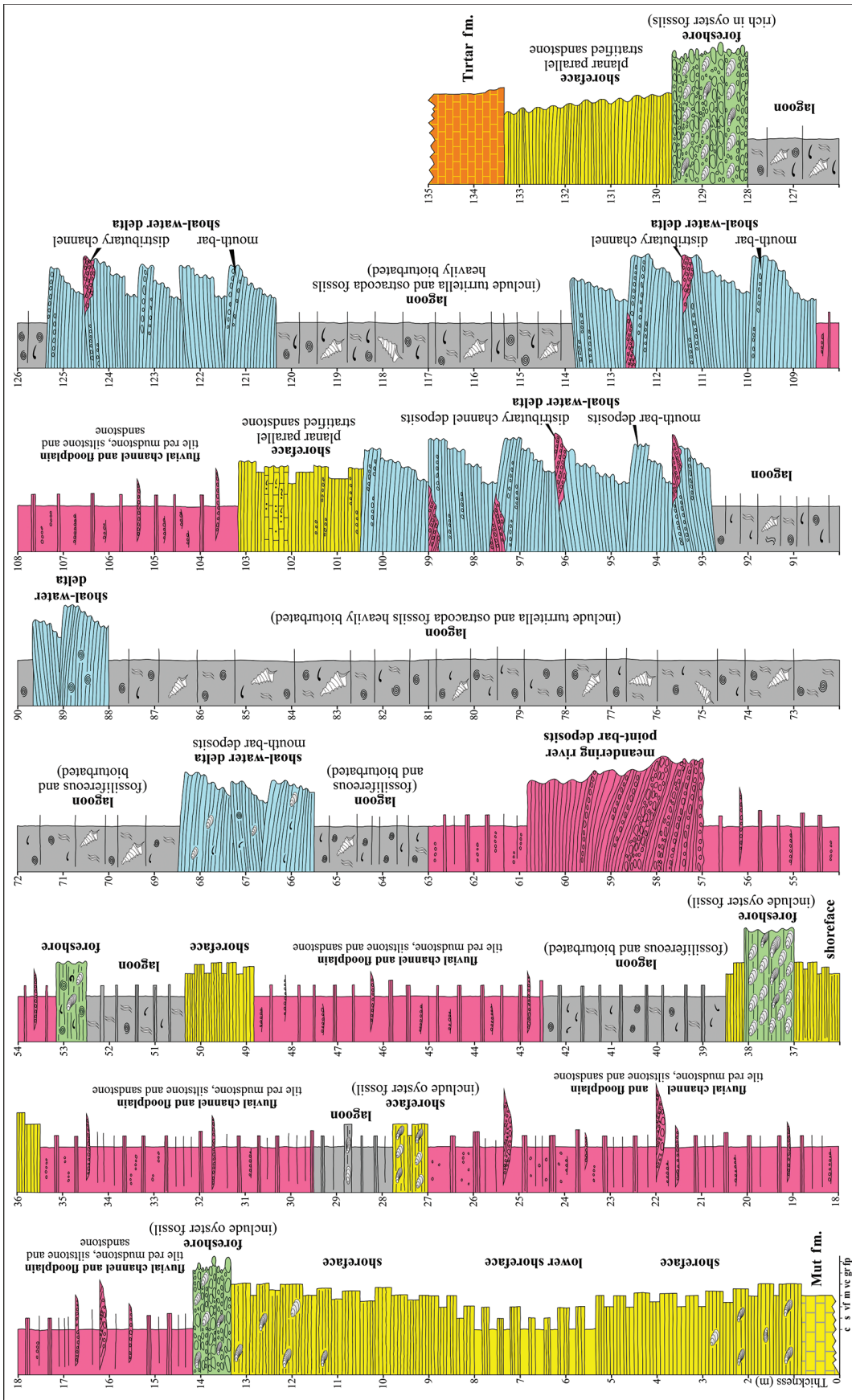


Figure 5- Sedimentological section of the Dağpazarı formation (section locality is seen in Figure 2a) and facies associations defined on this section.

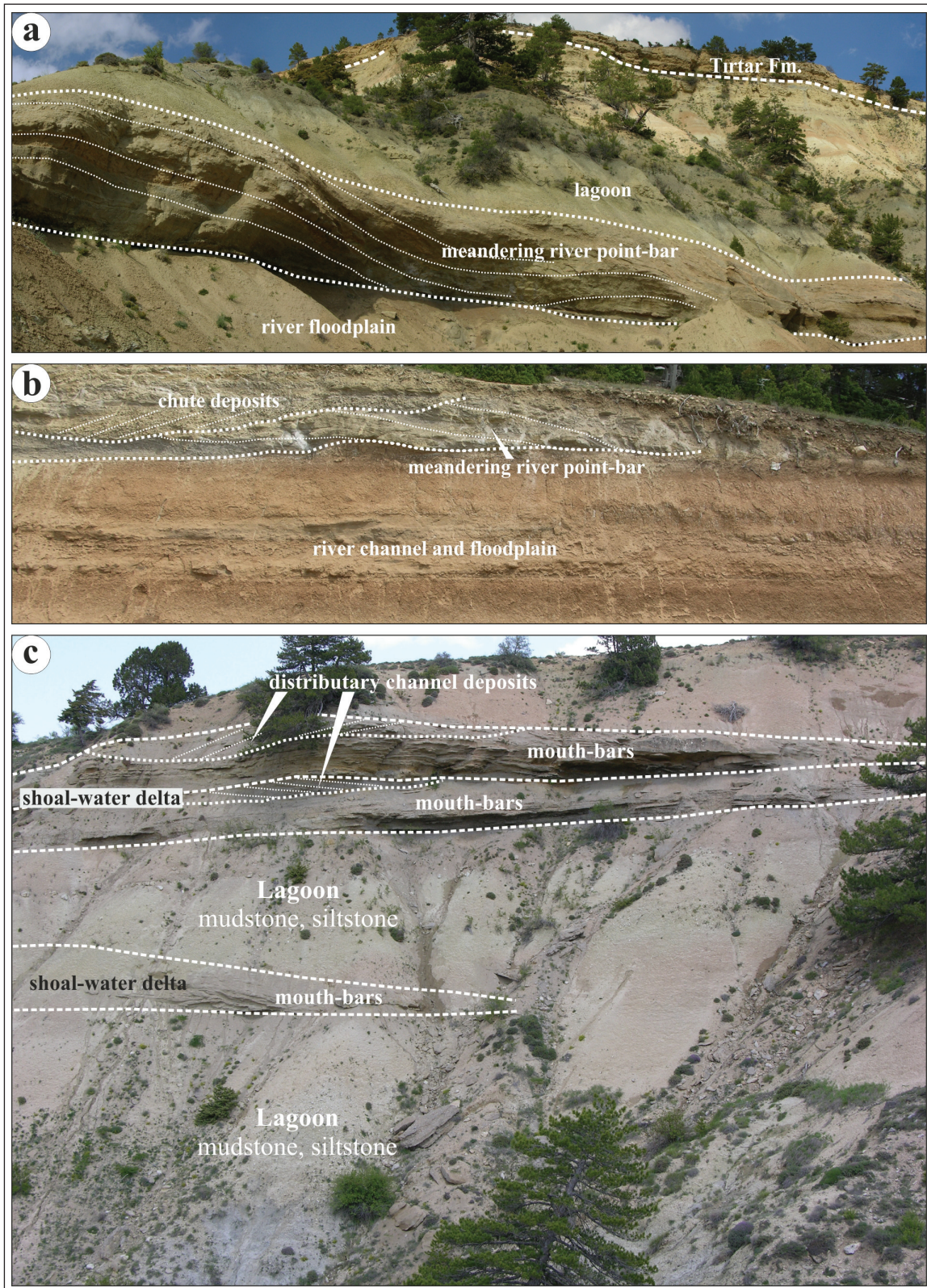


Figure 6- (a) Pale brown-brick red flood plain and meandering river point bar deposits composed of conglomerates and sandstones which are overlain by lagoon deposits. (b) River channel and flood plain deposits and meandering river point bar deposits, and overlying planar cross stratified chute deposits. (c) Shoal water delta deposits alternating with gray lagoonal mudstones. These deposits consists of distributary channel deposits and mouth bars.

deposits stored within the stream channel and point bar and chute sediments on them (Figure 6a, b).

The channel lag deposits (Miall, 1985; Nemeč and Postma, 1993) are observed on the pale brown-brick red mudstones, siltstones, sandstones of the flood plain deposits or at the bottom of channels. These are isolated beds of pebble conglomerates, which have erosional bottom surface, and are laterally discontinuous. These small-medium pebble conglomerates have clast-supported, cobbly framework filled with medium to very coarse sand, granules. Point bar deposits are located above the channel floor lag deposits (Figures 5 and 6a, b).

Point bar deposits in cross sections are formed by sandstones, pebbly sandstones or conglomerates in planar inclined or sigmoidal geometry (Figure 6a, b). These deposits, which are formed by these planar dipping beds, has 1-2,5 m set heights and 10°-20° dip angles. The dip angles of sandstones with sigmoidal geometry decrease both in upward and downward slopes and tangentially pass to the channel lag (Figure 6a). These layers are planar parallel and planar cross stratified. The current ripples can also be seen in upward slopes of sandstones. The superimposed channel deposits in the sequence are separated from each other by erosional surfaces. These channel deposits also consist of planar inclined deposits at different dip amounts and directions.

At the uppermost part of horizontally inclined sandstones and conglomerates, channel deposits up to 50 cm at thick and 3 m in width take place (Figure 6b). These channels with erosional bottom surfaces were developed diagonally in the dip direction of horizontal dipping layers. Channel deposits consisting of small pebbly conglomerate and coarse grained sandstones are generally planar cross-stratified (Figure 6b).

Horizontally inclined bedded or sigmoidal point bar sediments defined in the meandering river deposits were interpreted as the meander-belt sediments that have been formed by lateral accretion due to the migration of river channels in lateral direction (Jackson, 1976; Nanson, 1980; Brierley, 1991). In upper levels of the point bar deposits planar to cross stratified and current rippled sandstones were formed due to bedload transportation. The channel lag deposits located below the point bar deposits and on the basal erosional surface are associated with laterally migrating channel bottom (Ghinassi et al., 2014). Small channelized deposits on top of point

bars were interpreted as chute channel deposits. Chute deposits represent erosion and filling deposits formed during flood periods of the river (Mc Gowen and Gamer, 1970). The cross bedded sandstones deposited in chutes form chute bars deposited in the downstream side of bars (McGowen and Gamer, 1970; Ghinassi et al., 2014).

5.1.2. Flood Plain Deposits

Flood plain deposits associated with bar and chute deposits of the meandering rivers in stratigraphic succession consist of pale brown-brick red mudstone, gray siltstone and very fine to fine grained sandstones (Figures 5 and 6b). Massive mudstones form the dominant lithology of the succession. Very thin bedded siltstones and sandstones alternate with mudstones. These deposits are planar parallel stratified and are laterally widespread.

This facies association reflects sediments deposited on flood plains of meandering rivers (McGowen and Gamer, 1970). Planar parallel stratified sandstones and siltstones are the products of sedimentations that occurred at upper flow regime during floods (McGowen and Garner, 1970; Miall, 1985). The energy decrease in the non-channelized sheet-floods caused the deposition of mudstone from suspension (Collinson, 1996; Tooth, 1999).

5.2. Lagoon Deposits

Lagoon deposits alternating with fluvial, shoreface, beach and shoal water delta sediments in the succession mainly consist of yellowish-dark gray mudstone, siltstone and very fine to fine grained sandstones (Figure 5 and 6c). Massive mudstones are the dominant lithology of the succession. Mudstones alternate with very thin bedded siltstones and very fine grained sandstones. These deposits are rich in ostracod fossils and occasionally *Ostrea* fossils are observed. Because of dense bioturbation, the lamination or other sedimentary structures have not been generally preserved.

Ostracod species such as; *Loxococoncha tumida* Brady and *Loxococoncha* sp. in gray mudstones in the formation (Figure 7) indicate the fresh water entrances into the depositional environment, *Hemicyprideis* sp. on the other hand reflects brackish water conditions. The shoreface sandstones and oyster-rich beach deposits defined at different levels of the succession indicate the marine connections from time to time. The fossils of *Acanthocythereis hystrix* (Reuss),

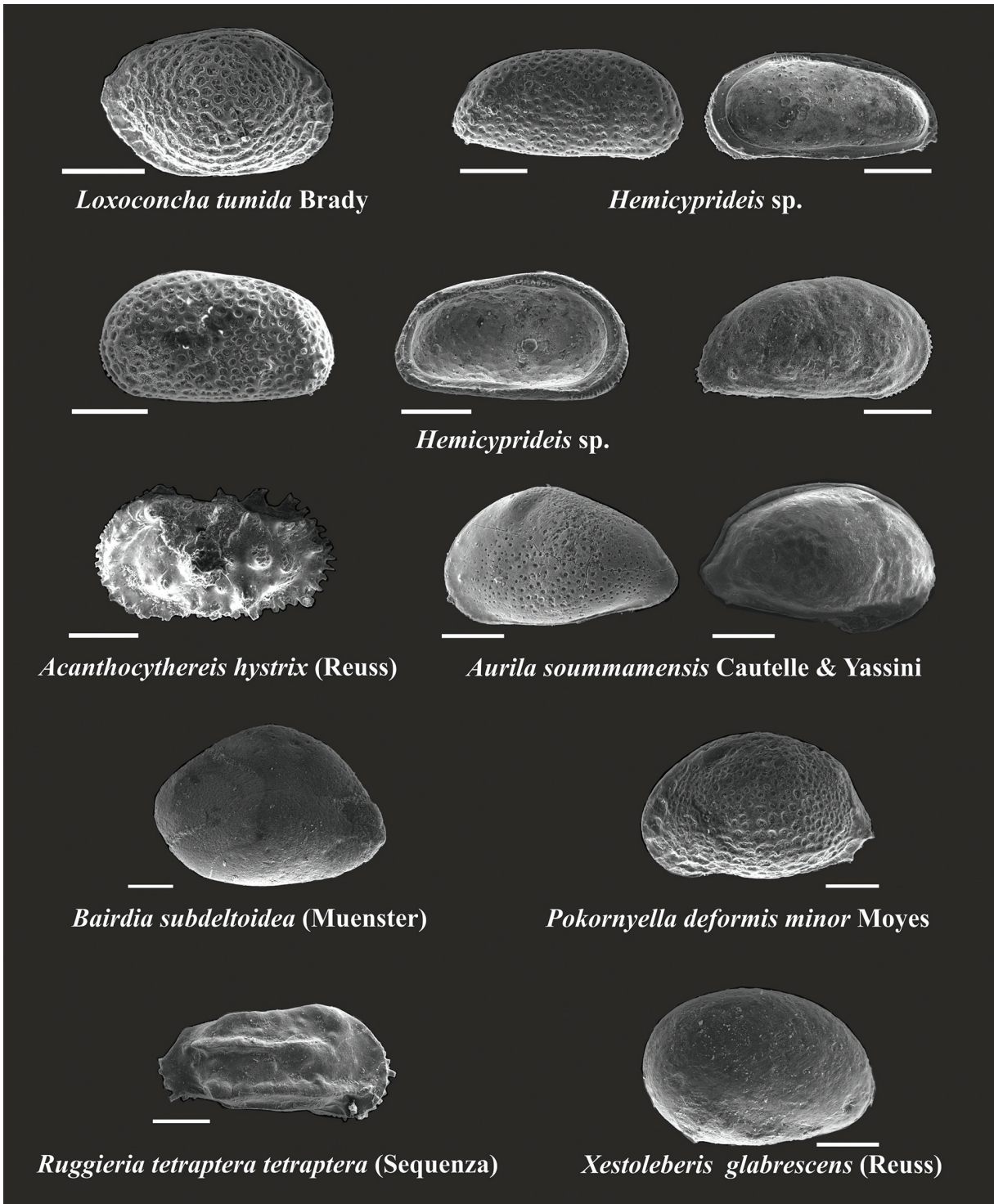


Figure 7- Ostracod species described in lagoonal mudstones of the Dağpazarı formation. Ostracod species such as *Loxoconcha tumida* Brady and *Loxoconcha* sp. reflect freshwater recharge into the depositional environment and the reduction in salinity, and *Hemicyprideis* sp. indicates the brackish water conditions. Fossils of *Acanthocythereis hystrix* (Reuss), *Aurila soummamensis* Cautella and Yassini, *Bairdia subdeltoidea* Muenster, *Paracypris polita* Sars, *Pokornyella deformis minor* (Moyes), *Ruggieria tetraptera tetraptera* (Sequenza), *Xestoleberis glabrescens* (Reuss) indicate that the depositional environment has sometimes marine connections. Bar scale 200 µm.

Aurila soummamensis Cautella and Yassini, *Bairdia subdeltoidea* Münster, *Paracypris polita* Sars, *Pokorniyella deformis minor* (Moyes), *Ruggeria tetraptera tetraptera* (Sequenza), *Xestoleberis glabrescens* (Reuss) also support that the depositional environment has sometimes connection with marine environment (Figure 7).

The lagoon deposits represent the sediments deposited in protected areas against the effects of wave and storm by limiting the open-marine connection due to the development of a barrier island.

5.3. Shoal Water Delta Deposits

The shoal water delta deposits are mainly composed of medium to very coarse grained sandstones (Figures 5 and 6c) with granule and fine pebbles. Deltaic deposits located on lagoonal mudstones with a sharp contact relationship form thickening and coarsening upward successions (Figures 5 and 6c). The shoal water delta deposits have 1,5-2 m thickness and 50-100 m lateral continuity. In these deposits, the beds become thinner and finer both down-dip and laterally, and the delta passes into other deposits (Figure 6c). The sandstone beds are gently sloping ($<10^\circ$) in both directions and display a lenticular package (Figure 6c). Each delta deposit is composed of distributary channel and mouth bars (Figures 5 and 6c). Similar delta deposits form delta packages nearly in 5–8 meters thickness which developed on each other with lateral displacements (Figure 5). The displacement of shoal water deltas in lateral and vertical directions in delta packets is related with the stillstand or increasing relative water level. In case of stable water level, the deltas migrate laterally and when the water level increases the delta packets stack upon one another. These delta packets are overlain by lagoonal mudstones or river flood plain deposits.

5.3.1. Mouth Bar Deposits

These mound-shaped mouth bar deposits are 1,5–2,5 m thick, typically thickening and coarsening upwards (Figure 5 and 6c). These sequences are composed of medium to very coarse grained sandstones, granules and very fine pebbly conglomerates. Among mouth bar successions, which are located in vertical or lateral directions, there are inclined erosional surfaces. Sandstone beds forming the mouth bar deposits have 5-25 cm thick. Sandstones are generally planar parallel stratified and rarely current and wave rippled cross stratified. Within sandstones, occasionally the

granule-fine pebble layers parallel to stratification are observed. Conglomerates defined in mouth bar deposits have a thickness of 5-35 cm and consist of alternation of granule-fine pebbly conglomerate. Conglomerates, which have grain supported textures, are mostly planar parallel stratified. The grain sizes of sandstones and conglomerates forming the mouth bars decrease both laterally and towards basin, and the layer thickness becomes thinner and wedges out (Figure 6c). Thus, these rocks tend to form downlap structures on the beds below.

The sandstones forming the mouth bars were originated by stream frictional effluent (Wright, 1977). Thus, the planar parallel stratified sandstones and current ripples were formed. The wave ripples observed on sandstones are the products of wave activity developed after deposition. Similar mouth bar deposits developed in shoal water deltas were also defined by Ilgar and Nemeč (2005), Leszczyński and Nemeč (2014) and Ilgar (2015). Most of the mouth bar deposits are erosionally overlain by distributary channel deposits. In some of them, granule-fine pebbled, planar parallel stratified, well sorted beach deposits take place (Bluck, 1967, 1999). Beach deposits indicate that mouth bar and distributary channel deposits were reworked by wave activities (Ilgar, 2015).

5.3.2. Distributary Channel Deposits

Channel deposits, which are located in axial sections of the mouth bar bodies and have concave erosional bottom surfaces, form the distributary channel deposits of the shoal water deltas (Figure 5 and 6c). Distributary channel deposits consist of grain supported, medium sorted, semi rounded-rounded granule and fine pebble conglomerates. Distributary channel deposits, which are as solitary channel and form a fining upward sequence, have a thickness of 30-80 cm and maximum width of 5 m (Figure 6c). Medium to coarse pebble conglomeratic layers interpreted as the channel bottom lag deposits are observed at the erosional bottom surfaces of distributary channel deposits (Miall, 1985; Nemeč ve Postma, 1993). Alternation of planar cross stratified granule-fine pebble conglomerate takes place on the channel bottom lag deposits. Planar cross stratified conglomerates are 25-60 cm high and reflect the transverse or mid channel bar deposits of the braided rivers (Miall, 1985).

5.4. Shoreface Deposits

Shoreface deposits consist of yellow-gray sandstones and very few amounts of siltstone and mudstones (Figures 5 and 8a). Well sorted, medium to coarse grained sandstones forming the shoreface facies are 15-50 cm thick and are laterally continuous. Sandstones, which have a sharp bottom contact, have horizontal lower boundary and planar to undulatory

upper boundary. Planar parallel stratification or wave ripples in less amounts form main sedimentary structures of the sandstones (Figure 5 and 8a). Sandstones are rich in *Ostrea* (Figure 8b) and shells are observed in occasion (Figure 8c). The first shoreface deposits located over the Mut formation are quite rich in broken coral fragments. Some levels are formed by fully intricate, well developed coral fragments (Figure 8d).

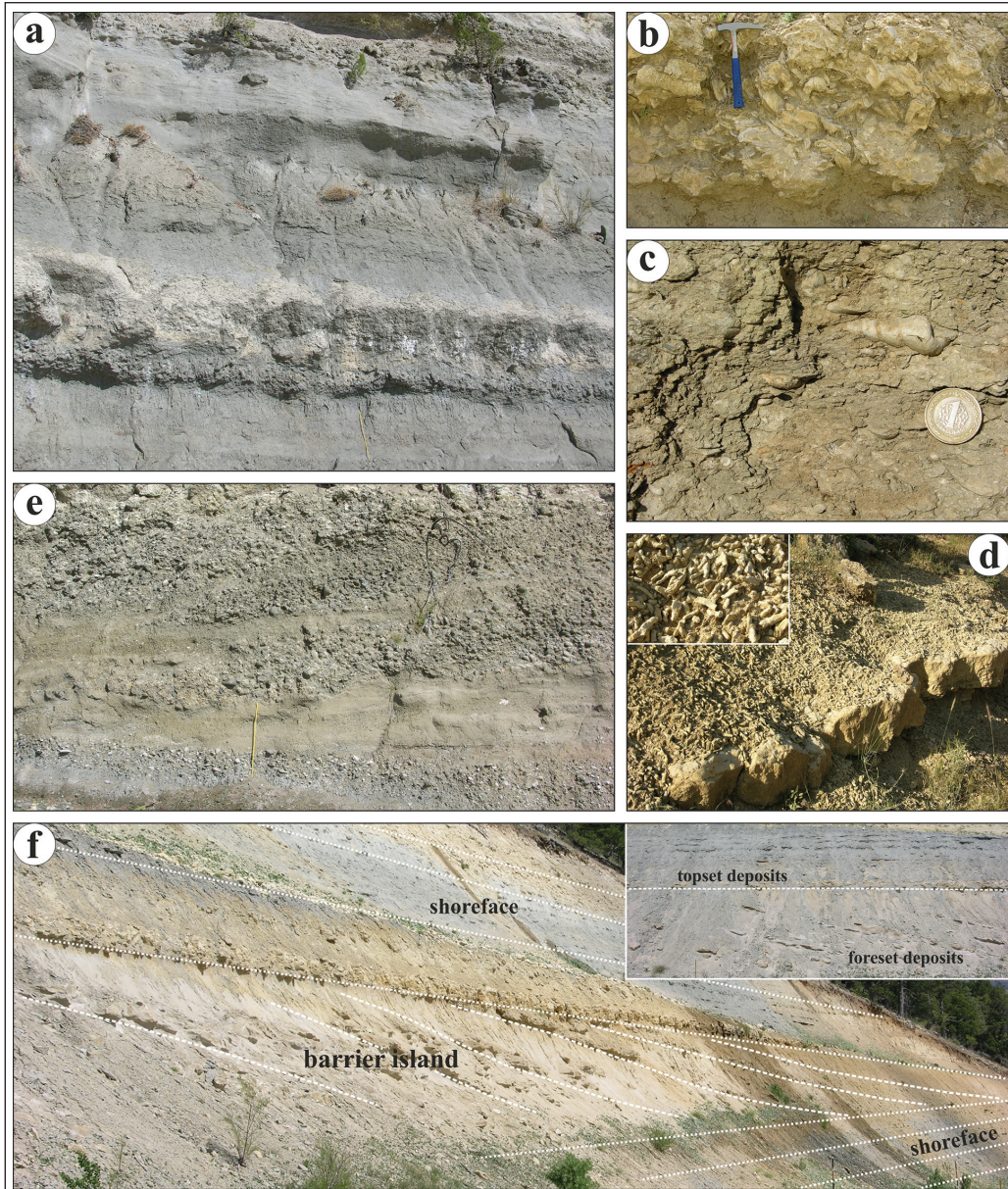


Figure 8- (a) Gray colored, planar parallel stratified shoreface sandstones defined in the Dağpazarı formation. (b) Sandstones are rich in *ostrea* fossils, (c) they also contain mollusc fossils in places. (d) Shoreface sandstones defined just on top of the Mut formation is quite rich in fragmented reef fossils. (e) Beach conglomerates which are laterally and vertically transitional with shoreface sandstones. (f) Barrier island deposits downlapping basinward defined between planar parallel stratified shoreface sandstones.

In intermediate levels of this shoreface deposit (Figure 5 between 5,5-8 m) very fine grained sandstone, siltstone and mudstone alternations take place. Thinly stratified (1-20 cm) sandstones and siltstones are laterally continuous. The upper boundaries of sandstones, which have a sharp and horizontal bottom contact on mudstones, are planar or undulatory. Well sorted sandstone and siltstone layers are normal graded, planar parallel stratified and wave rippled. Gray mudstones alternating with sandstones and siltstones are laterally continuous. Mudstones have massive appearance due to bioturbations. Shoreface sandstones are both laterally and vertically transitional with beach conglomerates.

Planar parallel stratified and wave rippled, medium to coarse grained, amalgamated sandstones indicate the deposition in the upper shoreface environment on the normal wave base (Walker and Plint, 1992; Ainsworth and Crowley, 1994). Mud sedimentation and preservation is difficult in this environment. The alternation of thin bedded sandstone, siltstone and mudstone defined in the shoreface succession indicate the sedimentation transported at certain time intervals such as the stormy periods in stable conditions. While the muds were deposited in normal conditions, the silt and other sand size grains were transported into the environment due to stormy conditions (Hunter and Clifton, 1982; Walker, 1984; Walker and Plint, 1992). This depositional environment indicates the lower shoreface environment at the normal wave base boundary or the offshore transition environment just below that environment (Ainsworth and Crowley, 1994).

5.5. Beach Deposits

Beach deposits, which are defined in the Dağpazarı formation and generally alternated with shoreface sandstones, consist of sandstones and conglomerates. These sandstones and conglomerates are rich in ostrea and these fossils are observed as mounds in occasion (Figure 5). Granular conglomerates and 5-20 cm thick, coarse to very coarse grained interbedded sandstones are planar to parallel stratified (Figure 8e).

Conglomerates are small-medium pebble sized, mostly well sorted, well rounded, spherical pebbles (Figure 8e). The intergranular voids in grain-supported conglomerates are filled with coarse sand or granule. The parallel or low angle stratifications are observed due to grain size difference in

conglomerates. Sandstone and conglomerate beds are inclined basinward at a very low angle (Figure 8e). Conglomerates have gradational or sharp contact relationship with the sandstone. The bed boundaries with sharp contacts are mostly erosional (Figure 8e). There are observed coarsening upward sequences in places where the sandstones gradually pass into conglomerates.

The characteristics features of this facies association reflect the deposition that had occurred in the beach environment (Bluck, 1967, 1999; Clifton, 1973; Massari and Parea, 1988; Postma and Nemeč, 1990). Well-sorted and well-rounded conglomerates and the stratifications developed due to grain size differences indicate that pebbles were washed and reworked by the waves and storms during the deposition (Clifton, 1973; Postma and Nemeč, 1990). Coarsening upward sequences in beach deposits indicate prograding beaches (Maejima, 1983; Nemeč and Steel, 1984).

5.6. Barrier Island Deposits

The barrier island deposits are located between the horizontally bedded shoreface sandstones and mainly consist of sandstones, granule and very fine pebble size conglomerates in few amounts. Foresets inclined towards the basin and the overlying topsets form the barrier island deposits (Figure 8f). Foreset deposits formed by inclined beds have 3-6 m thickness. Medium to coarse sandstones show thickness of 20-50 cm. Their lower boundary surfaces are planar whereas the upper contact is either planar or undulating. Sandstones are mostly planar parallel and in addition planar cross stratified in few amounts. Foreset deposits have the slope angle of 20-50° and the layers tangentially pass into shoreface deposits below. The boundaries of these sandstones with topset deposits are mostly erosional. Small scale erosional surfaces are also observed in foreset deposits.

Foreset deposits are overlain by nearly 1 m thick topset deposit and generally consist of coarse grained sandstones and granular conglomerates. Topset deposits are not laterally continuous in transverse sections and thin out into shoreface sandstones (Figure 8f). Sandstone and conglomerate layers have the thickness of 10-30 cm and planar parallel or planar cross stratified.

Deposits forming the barrier island pass landward into lagoonal mudstones and to shoreface sandstones in basinward direction (Figure 8f).

Barrier island, lagoon and estuaries are the characteristic depositional systems of transgressive shores (Swift et al., 1991). Sediments transported by waves and longshore currents form the barrier island ridges by being deposited in the form of sand bars parallel to the shore. The deposition occurs by the upward growing of beaches and shoreface bars due to the sea level rise (Swift, 1975). Erosional surfaces defined in foreset deposits reflect abrasions that occasionally developed due to storm events. The cross stratifications defined in topset deposits are mega ripples formed by strong longshore currents (Nielsen et al., 1988). Barrier islands are not stable in places and migrate landward as the sea level rises (Swift et al., 1991).

6. Biostratigraphic Dating

In order to determine the age of the Dağpazarı formation, total of 19 samples collected from the İbrahimli Stratigraphic Section at the lower boundary and from the Ballı Stratigraphical Section at the upper boundary of the formation are analyzed based on the planktonic foraminiferal biostratigraphy (Figure 2a). Planktonic foraminifera of the İbrahimli section are recorded in marine mudstones of the lower boundary of the Dağpazarı formation that can be conformably correlated in the basin. Planktonic foraminiferal assemblages are also determined in the marl samples from the lowermost part of the Ballı formation which overlies the Dağpazarı formation. Thus, the Dağpazarı formation is dated based on the age data obtained from the upper and lower boundaries of the formation.

The planktonic foraminiferal fauna in both sections are similar in terms of species diversity whereas they are different each other in the number of individual (abundance) and degree of preservation (fossilization). The İbrahimli section is characterized by abundant and well-preserved assemblages (Figure 9, 1-20). On the other hand, it is observed that the dominant fossil group is benthic foraminifers and planktonic foraminifera is low abundant and moderately preserved in the Ballı section (Figure 9, 21-25).

The planktonic foraminiferal fauna is represented by 25 species in the 11 marl samples from the İbrahimli section. *Paragloborotalia siakensis* (LeRoy), *Pg. partimlabiata* (Ruggieri and Sprovieri) and *Neogloboquadrina acostaensis* (Blow) are identified in all samples, whereas *Paragloborotalia mayeri* (Cushman and Ellisor) is recorded in a few samples

(İBR 3, 4, 8) and *Globigerinoides subquadratus* Brönnimann is only in one sample (İBR 1) (Figure 9). While, *Globigerinoides trilobus* (Reuss), *Gs. quadrilobatus* (d'Orbigny), *Gs. sacculifer* (Brady), *Globigerina bulloides* d'Orbigny, *G. falconensis* Blow, *Orbulina universa* d'Orbigny, *O. suturalis* Brönnimann, *Globigerinella obesa* (Bolli), *D. baroemoenensis* (LeRoy) are represented by a large number of specimens, *G. praebulloides* Blow, *G. occlusa* Blow and Banner, *G. ciperoensis* Bolli, *Gs. bulloideus* Crescenti, *Dentoglobigerina altispira* (Cushman and Jarvis), *Globoturborotalita decoraperta* (Takayanagi and Saito), *Glt. woodi* (Jenkins), *Globigerinita glutinata* (Egger), *Tenuitellinata angustiumbilitata* (Bolli), *Globorotalia scitula* (Brady) and *Globigerinella siphonifera* (d'Orbigny) are observed in fewer specimens (Figure 9).

The presence of *Neogloboquadrina acostaensis* (Blow) and *Paragloborotalia siakensis* (LeRoy), which are the characteristic species of MMi8 - MMi9 zonal interval, shows that the İbrahimli section is late Serravallian - early Tortonian in age (Figure 10; Sprovieri et al., 2002; Lirer et al., 2002; Hilgen et al., 2005; Lirer et al., 2005; Iaccarino et al., 2007). The other marker planktonic foraminiferal species, *Pg. partimlabiata* (Ruggieri and Sprovieri), *Pg. mayeri* (Cushman and Ellisor) and *Globigerinoides subquadratus* Brönnimann also support this age interval (Figure 10; Hilgen et al., 2003; Aguilar et al., 2004; Hilgen et al., 2005).

The 8 mudstone samples of the Ballı section don't include *Paragloborotalia siakensis* (LeRoy) which is the biomarker of the MMi9 Zone and is present in all samples of the İbrahimli section. Two characteristic species of the Serravallian, *Pg. mayeri* (Cushman and Ellisor) and *Pg. partimlabiata* (Ruggieri and Sprovieri) are not also recorded in the samples. In addition, *Neogloboquadrina acostaensis* (Blow), one of the marker species of the late Serravallian MMi8 Zone and *Globigerinoides subquadratus* (Brönnimann) are found only in one each sample (BAL 6 and BAL 1), similar to those in the İbrahimli section (Figure 9). The Ballı section includes *Globigerinella pseudobesa* (Salvatorini) (BAL 1, 2, 7), *Globoturborotalita nepenthes* (Todd) (BAL 1) and *Neogloboquadrina atlantica atlantica* (Berggren) (BAL 1, 2, 3) (Figure 9) in contrast to the İbrahimli section in which these species are not determined. The other species identified in the assemblage are *Globigerinoides trilobus* (Reuss), *Gs. quadrilobatus* (d'Orbigny), *Gs.*

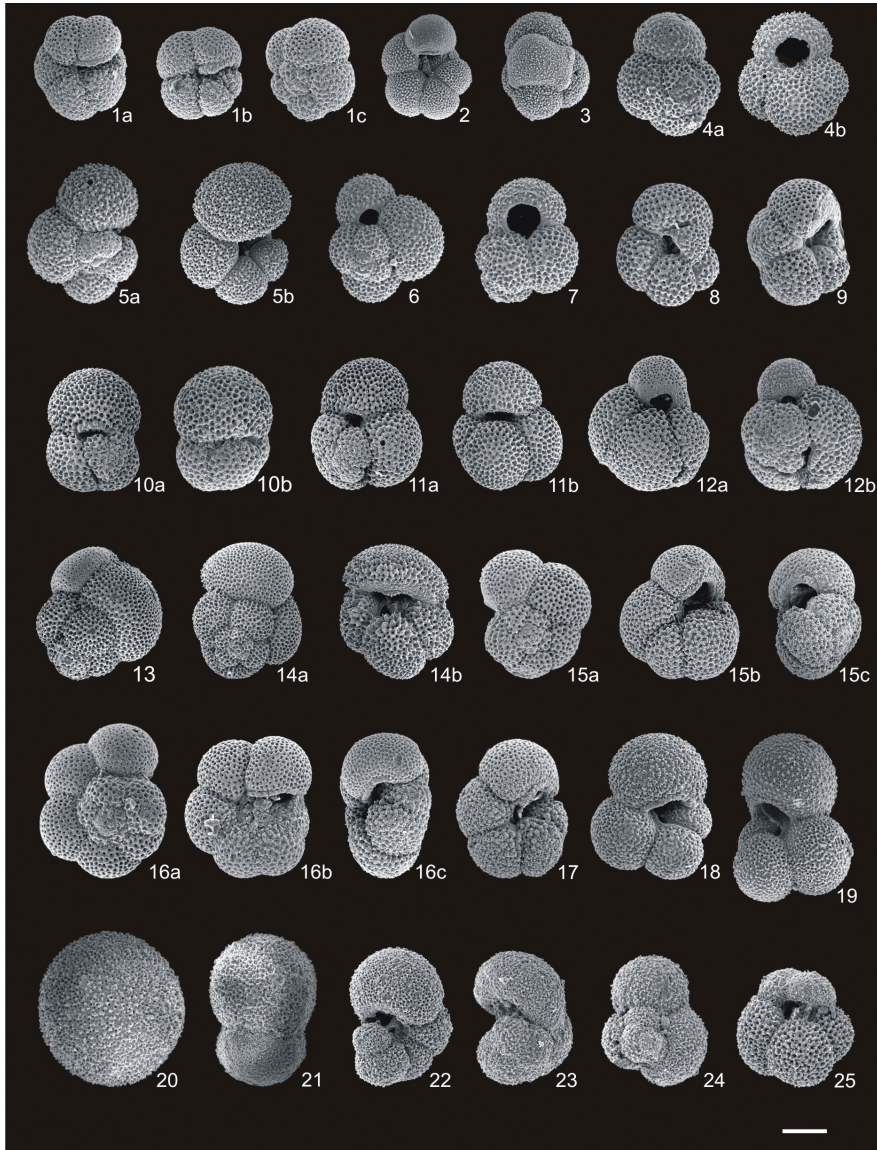


Figure 9- SEM views of the planktonic foraminifera described in the İbrahimli and Ballı sections (Scale: 1-9, 15-19, 24-25: 100 μm ; 12b, 20, 21: 110 μm ; 23: 120 μm ; 10-12a, 13b-14: 125 μm). 1a-c. *Neogloboquadrina acostaensis* (Blow), (a) umbilical view, (b) umbilical view, (c) spiral view, İBR.4; 2. *Tenuitellinata angustiumblicata* (Bolli), umbilical view, İBR.10; 3. *Globigerinita glutinata* (Egger) umbilical view, İBR.10; 4a, b. *Globoturborotalita woodi* (Jenkins), (a) spiral view, (b) umbilical view, İBR.4; 5a, b. *Globigerinella obesa* (Bolli), (a) spiral view, (b) umbilical view, İBR.10; 6. *Globigerinoides bulloideus* Crescenti, spiral view, İBR.5; 7. *Globigerinoides subquadratus* Brönnimann, umbilical view, İBR.1; 8. *Catapsydrax parvulus* Bolli, Loeblich and Tappan, umbilical view, İBR.10; 9. *Neogloboquadrina atlantica praeatlantica* Foresi, Iaccarino and Salvatorini, umbilical view, İBR.4; 10a, b. *Globigerinoides trilobus* (Reuss), (a) spiral view, (b) umbilical view, İBR.11; 11a, b. *Globigerinoides quadrilobatus* (d'Orbigny), (a) spiral view, (b) umbilical view, İBR.3; 12a, b. *Globigerinoides sacculifer* (Brady), (a) umbilical view, (b) spiral view, İBR.3; 13. *Dentoglobigerina altispira* (Cushman and Jarvis), spiral view, İBR.2; 14a, b. *Dentoglobigerina baroemoenensis* (LeRoy), (a) umbilical view, (b) spiral view, İBR.6; 15a-c. *Paragloborotalia partimlabiata* (Ruggieri and Sprovieri), (a) spiral view, (b) umbilical view, (c) side view, İBR.4; 16a-c. *Paragloborotalia siakensis* (LeRoy), (a) spiral view, (b) umbilical view, (c) side view, İBR.4; 17. *Paragloborotalia mayeri* (Cushman and Ellis), umbilical view, İBR.4; 18. *Globigerina occlusa* Blow and Banner, umbilical view, İBR.10; 19. *Globigerina praebulloides* Blow, umbilical view, İBR.10; 20. *Orbulina universa* d'Orbigny, İBR.10; 21. *Orbulina bilobata* (d'Orbigny), BAL.6; 22. *Globigerinella siphonifera* (d'Orbigny), oblique view, BAL.2; 23. *Globigerinella pseudobesa* (Salvatorini), oblique view, BAL.1; 24. *Globigerinoides ruber* (d'Orbigny), spiral view, BAL.1; 25. *Neogloboquadrina atlantica atlantica* (Berggren), umbilical view, BAL.3.

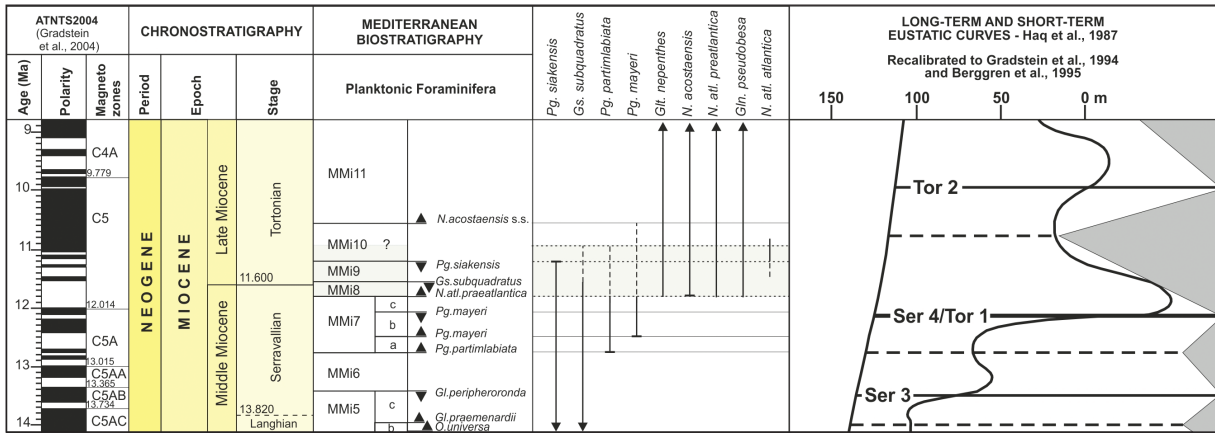


Figure 10- Mediterranean planktonic foraminifer stratigraphy and ATNTS2004 magnetic chronostratigraphy. The stratigraphical distribution of characteristic species described in samples of the Dağpazarı formation and eustatic sea level curves (Hardenbol et al., 1998).

sacculifer (Brady), *Orbulina universa* d'Orbigny, *O. suturalis* Brönnimann, *O. bilobata* (d'Orbigny), *Tenuitellinata angustiumbilitata* (Bolli), *Globigerina bulloides* d'Orbigny, *G. cf. ciproensis* Bolli, *G. cf. praebulloides* Blow, *Globigerinella obesa* (Bolli), *Gln. siphonifera* (d'Orbigny), *Dentoglobigerina altispira* (Cushman and Jarvis), *Globorotalia scitula* (Brady), *Neogloboquadrina continua* (Blow), *Globoturborotalita decoraperta* (Takayanagi and Saito), *Neogloboquadrina atlantica praeatlantica* (Foresi, Iaccarino and Salvatorini) and *Globigerinina glutinata* (Egger) Thus, 23 planktonic foraminiferal species are determined in the Ballı section.

The record of *Neogloboquadrina acostaensis* Blow, *N. atlantica praeatlantica* (Foresi, Iaccarino and Salvatorini), *Globigerinella pseudobesa* (Salvatorini) and *Globoturborotalita nepenthes* (Todd) which are firstly occurred in the MMi8 Zone indicates that the lowest age for the Ballı section is late Serravallian. On the other hand, the absence of *Paragloborotalia siakensis* (LeRoy), zonal marker of the MMi9 Zone, in any samples from the section, but the occurrence of *Neogloboquadrina atlantica atlantica* (Berggren) in some samples (BAL 1, 2, 3) reveals that the Ballı section can be correlated with the MMi10 Zone (Figure 10; Sprovieri et al., 2002; Lirer et al., 2002; Hilgen et al., 2005; Lirer et al., 2005; Iaccarino et al., 2007).

On the basis of the planktonic foraminiferal biostratigraphy, the valley fill deposits of the Dağpazarı formation correspond mainly to the MMi8 – MMi9 zonal interval (*Neogloboquadrina atlantica praeatlantica* - *Paragloborotalia siakensis* zones) and partly to the MMi10 Zone (*Globigerinoides obliquus*

Zone) which span the 11,78-10,90 million years interval.

7. Sequence Stratigraphy and Paleogeographic Evolution

The marine transgression, which began in late Burdigalian in the Mut Basin, has continued until late Serravallian and the reefal and platform carbonates of the Mut formation in the basin have been deposited in this period. The late Serravallian basin boundary of the Mut Basin is represented by the reefal limestones of the Mut formation (Figure 1b) around Paşabağı and Güçler villages in the north. Shoreface, beach, fluvial, lagoon and shoal water delta deposits of the Dağpazarı formation disconformably overlie the reefal limestones which represent a highstand systems tract (Figures 11 and 12). These sediments, which were deposited in an incised valley in late Serravallian-early Tortonian time (Figure 12), are located 20 km to the south of the late Serravallian basin boundary of the Mut Basin in basin direction. The sedimentological characteristics of Mut and Dağpazarı formations reflect an abrupt facies change, an erosional displacement of the late Serravallian shoreline towards basin and, therefore, a forced regression in the stratigraphic sequence (Zaitlin et al., 1994; Hampson et al., 1997; Plint ve Nummedal, 2000). The forced regression and the incised valley indicate a relative sea level fall in the basin in late Serravallian. The decrease in relative sea level caused the partial exposure of the Mut Basin and the formation of valleys on the reefal limestones (Figure 12). Three different system tracts were defined as the “forced regressive systems tract”, “lowstand systems tract” and “transgressive systems tract” which

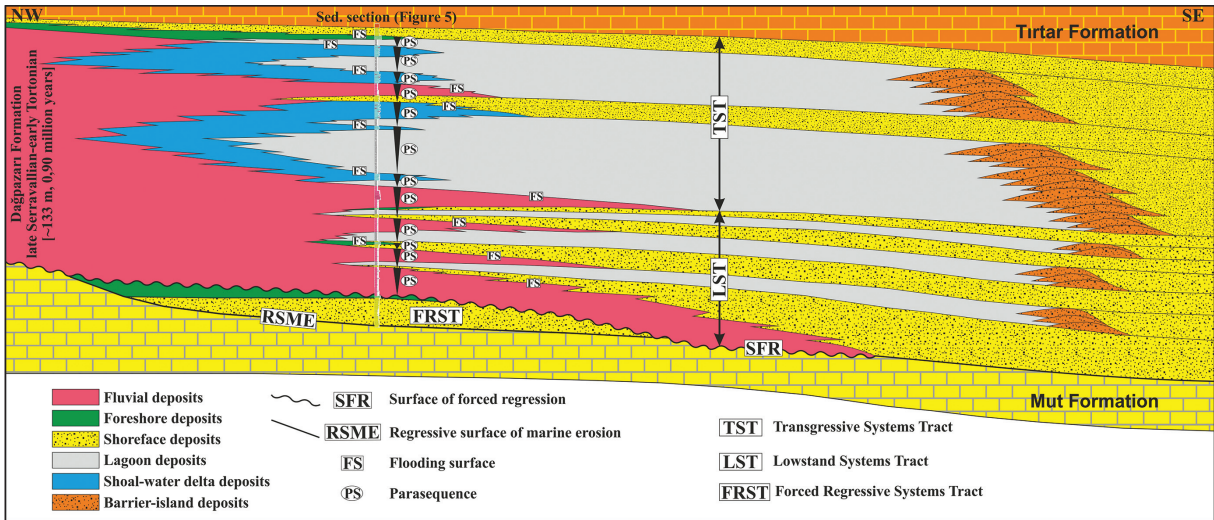


Figure 11- Sequence stratigraphy interpretation of the incised valley fill forming the Dağpazarı formation. This model was constructed based on the section shown in detail in figure 5.

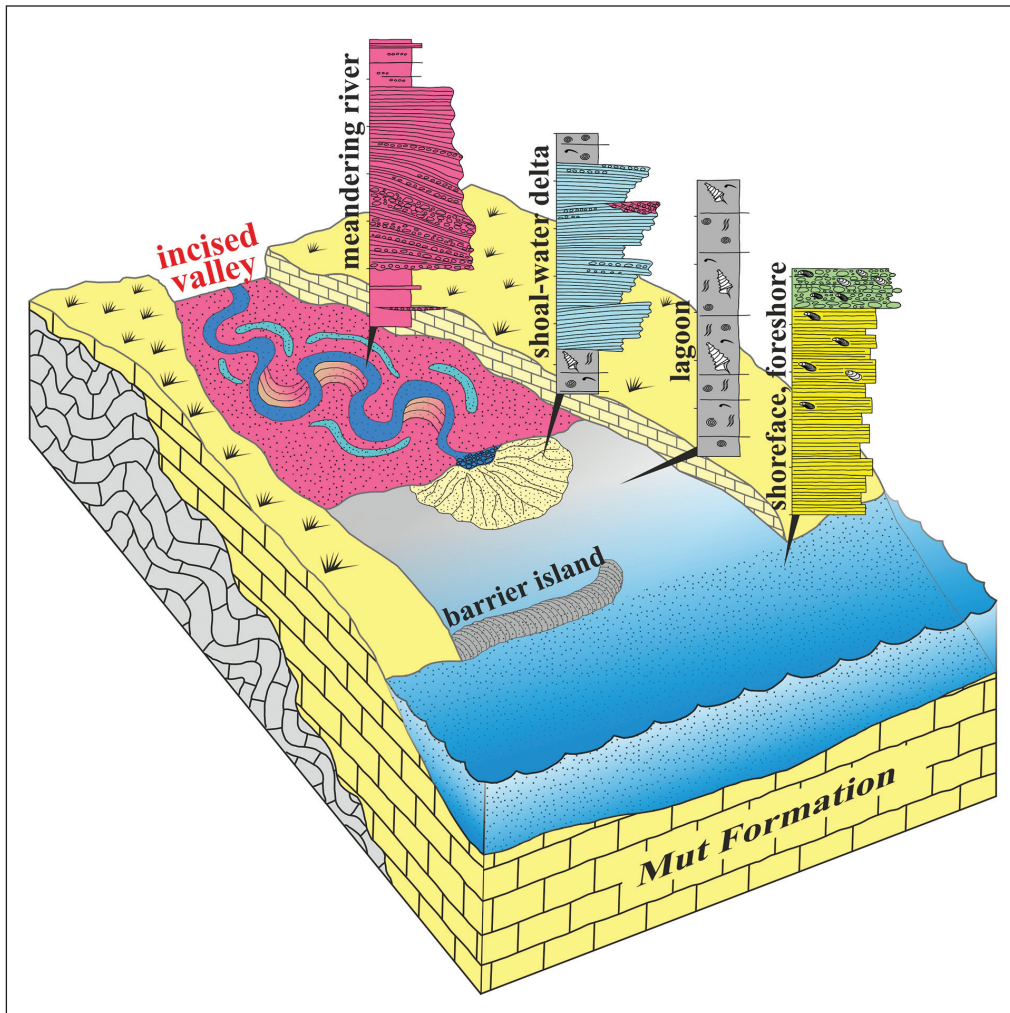


Figure 12- Schematic model of the incised valley where the facies associations forming the Dağpazarı formation deposited (not to scale).

formed during the relative sea level fall, lowstand and the increase of the sea level in the basin. System tracts reflect the paleogeographic changes through different types of deposition and boundary relationships.

7.1. Forced Regressive Systems Tract

While large erosional areas have been developed outside the basin due to the relative sea level decrease and the shoaling have occurred in the depositional area on the basin margin. Sediments derived from new erosional areas were transported into the basin and deposited on carbonate platforms with a sharp contact and formed shoreface and beach deposits. The presence of branched corals within shoreface sandstones in the form of fragmented components (figure 8d) indicates the exposure of reefal limestones and the transportation into the basin.

The sharp boundary between carbonate platforms and the overlying shoreface sandstones forms the “marine erosional surface of forced regression” (Plint, 1988) (Figure 11). The beginning of decrease in the relative sea level caused the lowering of the wave base in the shoreface environment and formation of marine erosional surface. Shoreface sandstones and beach pebbles were deposited on this surface.

In different sections of the basin, beach and shoreface deposits and the underlying platform carbonates were eroded by stream processes and fluvial sediments of the Dağpazarı formation were unconformably deposited (Figures 3c and 11). The deposition of fluvial facies indicates that the relative sea level fall continues. The unconformity surface is accepted as the sequence boundary (Mitchum, 1977) and units deposited on this boundary constitute the “lowstand systems tract” (Posamentier and Vail, 1988; Van Wagoner et al., 1988; Helland-Hansen and Gjølberg, 1994). The deposits, which were defined between the “highstand systems tract”, “lowstand systems tract” and deposited during relative sea level fall were named as the “falling stage systems tract” by Nummedal et al. (1992) and the “forced regressive systems tract” by Hunt and Tucker (1992). The shoreface and beach deposits, which overlie the highstand carbonates of the Mut formation with a sharp contact and disconformably underlie the fluvial deposits, are interpreted as the “forced regressive systems tract” (Figure 11).

7.2. Lowstand Systems Tract

The fluvial sediments of the Dağpazarı formation were deposited on an unconformity surface restricting the upper surface of the forced regressive system tract and forming the sequence boundary in the Mut Basin (Figure 5 and 11). The fluvial deposits vertically alternate with shoreface and lagoon deposits in the lower parts of the Dağpazarı succession (Figures 5 and 11). This alternation indicates the relative sea level rise after the first fluvial deposition in the incised valley. Thus, parasequences restricted by marine flooding surfaces in the succession were formed. Each parasequence that begins with shoreface deposits passed upward into lagoon, beach and fluvial mud plains and reflects normal regressive shore progradations. Total of 5 parasequences with thickness varying between 2,5 and 14 m were defined in the lower part of the sequence. These parasequences form an aggradational deposition and were interpreted as the “lowstand systems tract” (Posamentier and Vail, 1988; Van Wagoner et al., 1988; Helland-Hansen and Gjølberg, 1994) (Figure 11). High sediment supply from the basin margin prevented the relative sea level rise to flood the incised valley. Thus, the development of a rapid transgression in the basin was prevented and the sediments of the lowstand systems tract were deposited although the relative sea level rise had begun.

7.3. Transgressive Systems Tract

The continuous rise in the relative sea level flooded the incised valley and caused the valley to turn into a bay. The fresh waters discharging into the bay by rivers and the barrier island development in the bay mouth diluted the normal water salinity and caused the formation of brackish water conditions in the lagoon. Mainly lagoonal mudstones and shoal water delta deposits and subordinate shoreface and fluvial sediments were deposited in this environment (Figures 5 and 11). The shoal water deltas developed on lagoonal mudstones reflect the normal regressive shore progradations and parasequences. In the upper section of succession (between 63-129 meters) seven parasequences were defined with thicknesses varying between 4-21 meters (Figure 11). The development of these parasequences within incised valley reflect the interaction between the marine transgression and sediment transportation into the basin. In periods when the sediment supply into the basin are high and the

relative sea level rise are low, the normal regressive deltaic shore progradations on basin margin occurred. The increase in the relative sea level rise and the decrease in the sediment supply caused an increase in the accommodation space and formation of lagoonal mudstones. These parasequences form backstepping bodies in which marine processes are dominant. The deposition which reflects the deepening in the basin and the retreat of the shoreline were interpreted as the “transgressive systems tract” (Figure 11).

Thick fluvial and delta deposits alternating with lagoon and shoreface deposits within the Dağpazarı incised valley were interpreted as an “overfilled incised valley” as a whole (Breda et al., 2009). The continuing transgression in the basin caused the drowning of the incised valley and flooding of very large areas of the Mut Basin. The retreat of the shoreline due to transgression allowed re-establishment of carbonate deposition, which has continued until late Tortonian.

8. Discussion

Incised valley fills are significant deposit types in stratigraphic successions as they present the formation and the deposition of the valley (Boyd et al., 2009). The formation of these sediments is controlled by the relative sea level changes, tectonism, climate, sediment supply into the basin and paleogeomorphology (Posamentier and Vail, 1988). Accordingly the definition of these sediments in stratigraphical successions has a great importance in interpreting the stratigraphical relationships, paleogeographical developments and evolutions of deposition (Archer and Feldman, 1995; Boyd et al., 2006).

The Mut Basin forms one of the molasse basins, which was opened due to post-orogenic collapse in Neogene (Kempfer and Ben-Avraham, 1987; Robertson, 2000; Kelling et al., 2001). There is no any compressional tectonic deformation in Neogene in the basin, normal faults associated with post-orogenic tectonic extension in basin margin occur (Figure 13) (İlgar et al., 2013; 2016). In addition, there is no any

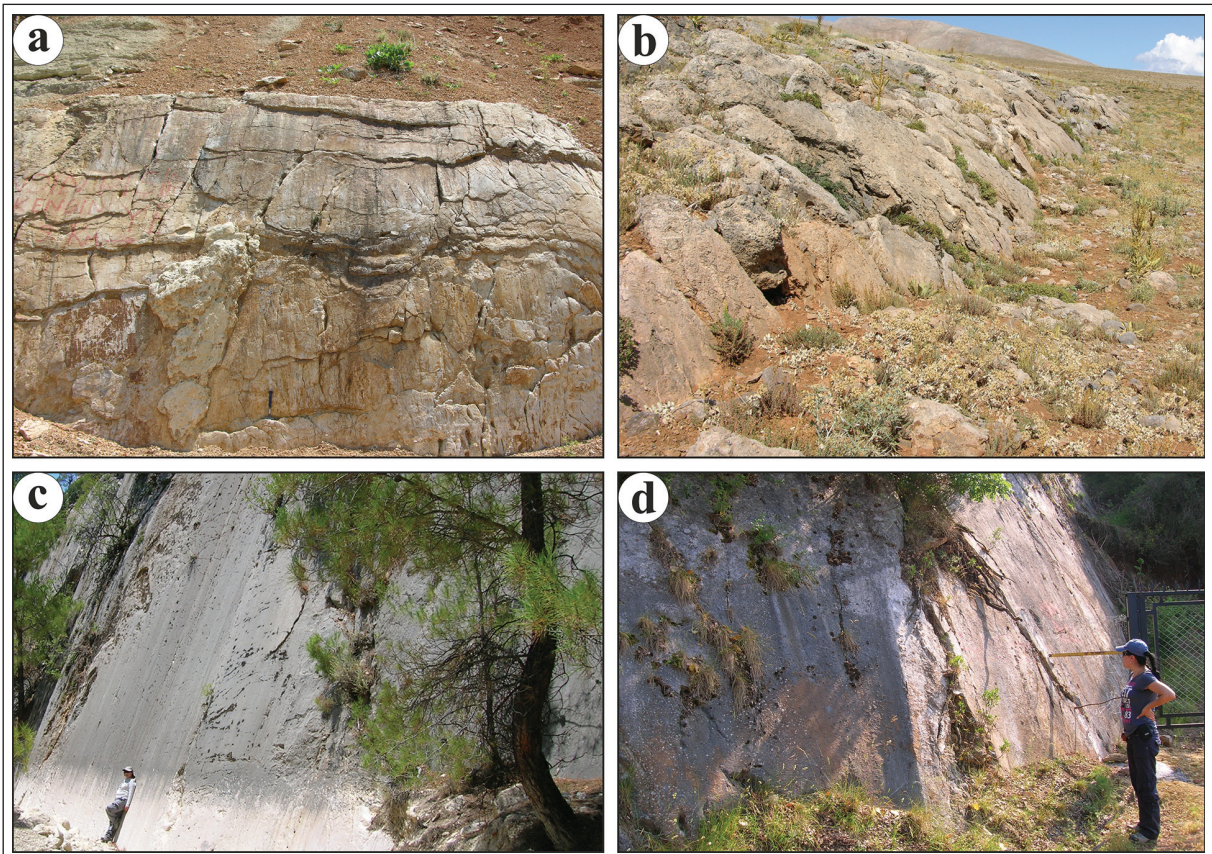


Figure 13- Normal fault planes cutting the bedrocks and Miocene limestones in different parts of the Mut Basin.

angular unconformity between the Mut formation and the overlying Dağpazarı formation. All these stratigraphical and structural characteristics of the Mut Basin show that the Miocene sedimentation in the basin has occurred in a post-orogenic quiet tectonic period. The isostatic rise of the Tauride began in late Tortonian and the Mut basin began to expose due to this rise which was seen in regional scale (Cosentino et al., 2012; Ilgar et al., 2013a). Local tectonic data of the Mut Basin and isostatic data of Tauride show that the exposure of the Mut Basin in late Serravallian and the formation of the incised valley did not develop depending on tectonism.

The planktonic foraminiferal biostratigraphy show that the youngest deposits below the forced regressional surface are located in the MMi7 biozone (12,77-11,78 million year interval) and the incised valley fill in MMi8 - MMi10 biozones (11,78-10,90 million years interval) (Ilgar et al., 2013b). The age of the forced regression surface, which causes the formation of the incised valley in the basin and can be correlated in regional scale, is conformable with the late Serravallian eustatic sea level fall (Ser4/Tor1; Figure 10). Therefore, the formation of the Dağpazarı incised valley can be associated with the late Serravallian eustatic sea level fall (eg. Rouchy and Saint Martin, 1992; Larsen, 2003; Roveri and Manzi, 2006; Ilgar et al., 2013a, b) which are also defined in Miocene basin around the Mediterranean.

During the migration of the shoreline, the sediments of the “forced regressive systems tract” can be either redeposited or completely eroded and shifted basinward. This situation depends on the amount of sediment supply and the drop velocity and amount in the relative sea level fall (Plint, 1988; Helland-Hansen and Gjølberg, 1994). The formation of the “forced regressive systems tract” during the relative sea level fall were interpreted in a such way that the sediment discharged into the Mut Basin was high. The fact that the exposure did not occur rapidly during the fall of the relative sea level probably indicates the contribute of subsidence in the basin. The deposition of thin-bedded sandstones and mudstones, which were interpreted as the lower shoreface facies during the deposition of coastal sandstones (Figure 5, 5.5-8 m), indicates the tectonic deepening of the basin.

Ser4/Tor1 eustatic sea level fall controls the exposure and the formation of the incised valley in the Mut Basin. In addition, it is also considered that the eustatic sea level rise (Figure 10) is one of the

factors controlling the infilling of the incised valley. The beginning of the eustatic sea level rise has also caused the base level rise and terminated the erosion that caused the valley formation. Thus, thick fluvial deposits began to deposit within the incised valley.

In addition, the basin subsidence and the sediment supply into the basin also controlled the facies development in the incised valley. Eustatic sea level rise and the subsidence of the basin caused the deepening of the incised valley and the increase of the depositional area. The marine transgression in these periods flooded the incised valley and allowed the deposition of shoreface sandstones in the valley.

In times when relative sea level increased slowly, the formation of barrier-island and lagoonal basin conditions in protective areas behind barrier-island have developed. Thus, the mudstone deposition occurred in the lagoon during periods when it was protected from the effects of open sea processes such as waves and tides and was not intensely affected by fluvial processes. The shoal water deltas were formed within lagoon during high sediment supply periods. Most probably, the regional climatic fluctuations and changes in the amount of river sediment, have controlled the intensity of the stream activity.

The shoreface sandstones deposited on lagoon sediments indicate the times when the open sea connection was re-established through passing over barrier-islands due to the eustatic sea level rise and the subsidence in the basin.

The “lowstand systems tract” and “transgressive systems tract” forming the incised valley fill is consisted by 12 parasequences. These parasequences indicate small sea level rises occurred during the infilling of the incised valley. The facies analysis and sequence stratigraphy of the Dağpazarı formation in the Mut Basin, which is a subsidence area due to the post-orogenic tectonic extension, showed that the interaction between the eustatic sea level rise in small amounts, extensional tectonic regime and the variable sediment supply rate controlled the deposition.

9. Conclusion

The sedimentary facies analysis of the middle-late Miocene clastic sequence of the Mut Basin was carried out and then the sequence stratigraphic evolution was interpreted by detailed sedimentological study. The relative sea level changes that control the

development of sedimentary facies were dated by means of the planktonic foraminiferal biostratigraphy. Thus, the sedimentological and biostratigraphical data and the sequence stratigraphic interpretations were used in the basin and the middle and late Miocene sedimentological (paleogeographical) evolution of Mut basin and the processes controlling this development were discussed.

The reefal limestones and platform carbonates of the Mut formation were deposited during the late Burdigalian-late Serravallian in the Mut Basin. The shoreface, beach, fluvial, lagoon and shoal water delta deposits forming the Dağpazarı formation unconformably overlay the reefal limestones of the Mut formation and reflect a valley fill sequence deposited within incised valley. This sequence is located about 20 km to the south of the basin margin of the late Serravallian highstand reefal carbonates. The facies characteristics of the Mut and Dağpazarı formations reflect abrupt facies changes in the stratigraphic sequence, basinward erosional shift of the late Serravallian shoreline and the forced regression.

There is no any compressional tectonic deformation in Neogene Mut Basin which opened depending on post-orogenic tectonic extension. The Miocene sedimentation occurred relatively in a quiet tectonic period. The local tectonic data of the Mut Basin and the regional isostatic data of Tauride indicate that the late Serravallian exposure and incised valley formation in the basin did not develop due to tectonic.

The planktonic foraminiferal biostratigraphy shows that the incised valley fill was formed between the biozones of MMi8–MMi10 (11.78-10.90 million years interval) in the late Serravallian-early Tortonian. The age of the forced regression, which caused the formation of the incised valley in the basin, is conformable with the late Serravallian eustatic sea level fall (Ser4/Tor1). Therefore, the formation of the incised valley of Dağpazarı was associated with the late Serravallian eustatic sea level fall. It is thought that the post Ser4/Tor1 eustatic sea level rise in the Mut Basin is one of the factors controlling the valley filling. In addition, the basin subsidence in the extensional tectonic regime and varying sediment supply into the basin also controlled the facies development within incised valley.

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