1. Introduction
The Anatolian block of Turkey, after its tectonic assembling through the collision of the Pontides with the Anatolide-Tauride platform of the East Mediterranean Alpine orogeny, soon became subject to a westward “tectonic escape” (structural strike-slip expulsion) driven by the northward movement of the Arabian plate accommodated by the sinistral East Anatolian and dextral North Anatolian fault zones (Figure 1A; McKenzie, 1972; Dewey and Şengör, 1979; Şengör et al., 1985; Flerit et al., 2004). This postorogenic neotectonic regime is thought to have commenced in the middle Miocene, around 12 Ma BP (Barka and Hancock, 1984; Şengör et al., 1985), rendering Anatolia one of the most seismically active regions of the Eurasian continent – notorious for its numerous and often disastrous earthquakes (Dewey, 1976; Saroğlu et al., 1992).

The Recent fault activity and seismicity in Anatolia are well-documented from the historical record of catastrophic earthquakes and tsunami events (Ambraseys and Finkel, 1995; Altnok and Ersoy, 2000; Ambraseys, 2001; Tan et al., 2008), from surface ruptures and shallow excavations (Ambraseys and Jackson, 1998; Awata et al., 1999; Barka et al., 2000; Witter et al., 2000; Bouchon et al., 2002; Kondo et al., 2005; Özaksoy et al., 2010), from high-resolution shallow seismic-reflection imagery (Parke et al., 1999; Kusçu et al., 2002) and radar interferometry (Wright et al., 2001), and from abrupt local paleogeographical changes (Ryan et al., 1997; Leroy et al., 2002; Özaksoy et al., 2010). Little geological documentation exists of the earlier, pre-Holocene neotectonic activity and its onset in Anatolia (Poisson et al., 2003; Koç Taşgın, 2011; Koç Taşgın et al., 2011), most probably because of the negligible preservation potential of a surface record of older seismotectonic events.

The present study from a middle Miocene shallow-marine siliciclastic succession in the Sinop Peninsula, north-central Turkey (Figure 1B), is the first documentation of a syndepositional seismic fault activity that can be attributed to the early onset phase of neotectonics in
Figure 1. A) Tectonic map of Anatolia, showing the Pontide and Tauride orogenic belts enveloping the Kırşehir and Menderes crystalline massifs (simplified from Leren et al., 2007); note the neotectonic pattern of strike-slip deformation along the East Anatolian (EAF) and North Anatolian (NAF) fault zones driving the westward “tectonic escape” of the Anatolian craton. B) Geological map of the Sinop Peninsula, showing the tectonically-inverted foreland basin of Central Pontides to the south and the associated marginal trough to the North (modified from Gedik and Korkmaz, 1984). The trough’s southern margin is approximated by the Balıfakı thrust, whereas the northern, outer margin corresponds to the belt of Cenomanian volcanics. Note the location of the study area in the NE part of the Sinop Peninsula.
northern Anatolia. The synsedimentary deformation described from the Sinop outcrop section was associated with the subaqueous escarpment of an active, oblique-slip reverse fault and involved both contemporaneous nearshore sediments and their consolidated shallow-marine substrate. Similar features related to underwater or terrestrial fault escarpments may be common in the stratigraphic record of Anatolia and other tectonically active regions, but may have been lumped with postdepositional fault breccias or talus deposits and their significance may have escaped recognition. The present study thus also contributes to the sedimentological criteria for the deciphering of ancient seismotectonic events from the stratigraphic record.

2. Geological setting

The Neogene basin in north-central Turkey formed as a marginal trough at the northern edge of the tectonically inverted, ESE-trending foreland basin of the Central Pontides (Figure 1A). The retroarc foreland basin evolved from a failed Late Cretaceous rift (Leren et al., 2007). The marginal trough, paralleling the folded and thrust-deformed foreland, is presently exposed only in the Sinop Peninsula (Figure 1B). The Neogene siliclastic succession represents the southern reaches of Eastern Paratethys (Görür et al., 2000) and rests unconformably on the preorogenic Late Jurassic–Early Cretaceous carbonate platform and Cenomanian volcanic rocks, whereas to the south, it overlies partly the youngest, middle Eocene deposits of the deformed foreland basin (Figure 1B). The Neogene basin is structurally bounded by Balıfakı thrust to the south. Thrusting of the Paleocene–early Eocene Atbaşı Formation over the middle Eocene Kusuri Formation on the NW part of the Balıfakı thrust (Figure 1B) reflects the post-middle Eocene activity of the fault. This activity continued up to the early Miocene and caused the uplift of the region.

The formation of marginal troughs is known to be associated with rift basins tectonically inverted into fold belts (e.g., see the Polish-Danish Anticlinorium in Pozaryski and Brochwicz-Lewiński, 1978; Ziegler, 1990, fig. 71). The early Miocene development of the Sinop marginal trough was due to the uplift and gravitational foundering of the Pontide orogen caused by crustal thickening (see Seyitoğlu and Scott, 1991, 1996; Yılmaz et al., 1997), which reactivated the foreland Balıfakı thrust (Figure 1B) and resulted in crustal downwarping along the orogen margin. The resulting synclinal basin initially hosted a neritic marine environment, but underwent rapid shallowing as the subsidence ceased by the end of early Miocene. Littoral sedimentation prevailed in the middle Miocene, when the abandoned marginal trough eventually became affected by transpressional tectonic deformation (see Barka and Hancock, 1984; Şengör et al., 1985).

3. Basin-fill stratigraphy

Sedimentation in the Sinop marginal trough commenced with the deposition of the shallow-marine Sinop Formation (Figure 1B; Gedik and Korkmaz, 1984), composed of mudstones, sandstones, and minor fine-grained conglomerates. The lower part of the formation consists of dark gray, nonfossiliferous mudstones intercalated with thin sandstone sheets (Figure 2A). These deposits pass upwards into a heterolithic facies composed of fine-grained sandstones with mudstone interlayers and with wavy to lenticular bedding (Figure 2B). Molluscan fauna at the transition of the mudstone to heterolithic facies indicates a Tarkhanian/Tshokrakian (early middle Miocene) age of the deposits. The proportion of sand increases upwards and slightly varies (Figure 2C), until fossiliferous sandstones predominate, intercalated with thin erosional layers of granule and fine-peatle conglomerate. Interbeds of fossiliferous stromatolitic and/or oolitic limestone (Figure 2D) also occur locally in the upper part. Primary sedimentary structures in sandstones include wave-ripple cross-lamination (Figure 2E), hummocky and planar parallel stratification (Figure 2F), and bidirectional planar cross-stratification. A local occurrence of granule-rich fossiliferous sandstones shows lateral accretion cross-stratification (Figure 2G), attributed to an isolated paleochannel.

The sedimentary facies succession of the Sinop Formation indicates a gradual upward shallowing with minor sea-level fluctuations and with the environment evolving from a mud-dominated offshore transition zone into a heterolithic lagoonal tidal flat by the beginning of middle Miocene. Barrier facies are unexposed, but the coarse-grained lateral-accretion unit probably represents a migrating tidal inlet channel extended over the lagoon. The sedimentary environment evolved further into a wave-dominated sandy shoreface, with a gradual encroachment of a seaward-prograding gravelly foreshore zone.

The normal (progradational) marine regression was then interrupted by a strong sea-level fall at the Tshokrakian/Karaganian transition, when a forced regression occurred and the basin was emerged. Incised fluvial valleys formed, more than 120 m deep, and were subsequently filled with marine to deltaic deposits as the sea level began to rise in the Konkian time. The transgressive Gelinçik Formation (Figure 1B) consists of foreshore to shoreface deposits with associated shoal-water and Gilbert-type deltas, recording a gradual rise in sea level followed by highstand. This record of sea-level changes in the basin clearly postdates the orogeny and corresponds to the Paratethyan eustatic history (Harzhauser and Piller, 2004).

The basin was then tectonically uplifted in the late Miocene, with the basin-fill succession presently exposed at up to 130 m above sea level. The Plio-Pleistocene
Figure 2. Sedimentary facies of the Sinop Formation. A) Dark gray offshore-transition mudstone interspersed with thin sandstone sheets. B) Thin lenticular sandstone beds intercalated with mudstones of a lagoonal tidal-flat succession. C) Planar parallel-stratified, fine- to coarse-grained shoreface sandstones with subordinate wave-ripple cross-lamination, intercalated with silty mudstone layers. D) Fossiliferous oolitic limestone interbed in the upper part of the formation. E) Close-up view of wave ripple forms in fine-grained, cross-laminated shoreface sandstone. F) Fossiliferous, granule-rich, planar parallel-stratified foreshore sandstone. G) Fossiliferous, granule-rich sandstone with laterally accreted cross-strata, interpreted as a point bar of tidal inlet channel.
aeolian deposits of the Sarıkum Formation (Figure 1B) unconformably overlie the Sinop and Gelincik formations, smoothing out the post-Miocene paleotopography.

4. Outcrop evidence of synsedimentary deformation

4.1. Description and interpretation of deposits
The evidence of synsedimentary deformation, attributed to seismotectonic activity, has been found in the upper, sandy part of the shallow-marine Sinop Formation exposed in a roadcut section on the eastern side of the Sinop Peninsula (see study area in Figure 1B). The E–W roadcut section is 4.5 m high and 35 m long (Figure 3). An oblique-slip reverse fault formed in, and buried by, sandy shoreface deposits divides the outcrop section into a hanging-wall part to the left and a footwall part to the right. The faultplane dip azimuth and angle are 105/65°. Striations on the fault plane indicate a reverse fault with dextral strike-slip component (Figure 3).

4.1.1. The hanging-wall part
The hanging-wall part of the fault structure in its lower part (Figure 3, lower left) consists of alternating very fine-grained sandstone and subordinate siltstone beds showing planar parallel stratification and minor wave-ripple cross-lamination. The deposits also show open fissures filled with angular fragments of similar rock and a poorly sorted matrix of fine to coarse sand. Directly above, the deposits show intense brecciation into small angular blocks surrounded by a similar, relatively coarse sand matrix. The brecciated sandstone is unconformably overlain by an upper unit of nondeformed, very fine-grained sandstone alternating with siltstone layers (Figure 3, upper left). The sediment is similarly well-sorted, forming tabular layers 1–10 cm thick, with mainly planar parallel stratification and minor wave-ripple cross-lamination (ripple cross-sets of ≤0.5 cm thick).

There is no evidence of unidirectional paleocurrent in the lower and upper unit, and the stratification types indicate a perennial action of waves with near-bottom orbital velocities of mainly ≥0.5 m s⁻¹ (see Allen, 1982, figs. 11–19). For the available grain-size range, the near-bottom oscillatory currents were apparently too strong to allow common formation of wave ripples, whereby plane-bed configuration predominated. Perennial wave action implies deposition above a fair-weather wave base. The fine grain sizes and scarcity of recognizable erosional surfaces suggest deposition in a middle shoreface environment, at the transition of lower to upper shoreface zone (Goldring and Bridges, 1973; Clifton, 1976, 1981; Walker and Plint, 1992; Ainsworth and Crowley, 1994). Wave near-bottom orbital velocities were ≥0.7 m s⁻¹, often >0.9 m s⁻¹ (Allen, 1982, fig. 11–19), and the occurrence of hummocky stratification indicates episodic storm-generated combined flow (Dott and Bourgeois, 1982; Brenchley, 1985; Duke et al., 1991; Brenchley et al., 1993; Cheel and Leckie, 1993). The upper sandstone unit, which onlapped and fully buried the fault escarpment (Figure 3), shows planar parallel stratification and minor wave-ripple cross-lamination and is attributed to deposition in a middle shoreface environment.

4.1.2. Footwall part
The footwall part of the fault structure shows a considerably greater degree of deformation (Figure 3, lower right). The deformation is strongest near the top of the lower sandstone unit, with the deformation zone thinning from 135 cm to less than 50 cm with distance from the fault (Figure 3). The large isolated blocks of stratified sandstone, up to 3.5 m in length, consist of similar shoreface deposits as those in the lower hanging wall, but show steep to vertical internal parallel stratification and have apparently been dislodged from the fault escarpment. The lower block (first block in Figure 3), after its emplacement, was buried by fine- to medium-grained sand, but initially shed some smaller fragments and accumulated massive sand on its lower flanks. The surrounding sandstone shows planar parallel stratification and sporadic wave ripples on bedding planes in its lowest part, 50–75 cm thick, but is increasingly fossiliferous and dominated by hummocky cross-stratification in the upper part, 70–125 cm thick. This lower sandstone unit has an uneven upper boundary overlain by stratified medium-grained sandstone on which the second large block is resting (second block in Figure 3). The sandstone planar parallel strata are deformed both below the block and on its sides, with small convolutions and diapiric folds (Figures 3 and 4). Notably, this middle sandstone unit lacks deformation farther away from the fault and overlies sharply the deformed lower sandstone unit (Figure 3, right-hand part), which means that the second block was emplaced considerably later, after a significant episode of sedimentation. This block and the middle sandstone unit disturbed by its emplacement are overlain unconformably, with slight erosional truncation, by an upper unit of parallel-stratified, very fine-grained sandstone alternating with siltstone layers. As in the underlying units, the seafloor-draping parallel stratification is not perfectly planar, apparently adjusted to the preexisting local relief and also bent slightly by compaction.

The characteristics of the lower sandstone unit, ~2.5 m thick, differ from those of the hanging-wall sandstone and suggest deposition in an upper shoreface environment (Clifton, 1976, 1981; Bourgeois, 1980; Walker and Plint, 1992; Ainsworth and Crowley, 1994). Wave near-bottom orbital velocities were ≥0.7 m s⁻¹, often >0.9 m s⁻¹ (Allen, 1982, fig. 11–19), and the occurrence of hummocky stratification indicates episodic storm-generated combined flow (Dott and Bourgeois, 1982; Brenchley, 1985; Duke et al., 1991; Brenchley et al., 1993; Cheel and Leckie, 1993). The upper sandstone unit, which onlapped and fully buried the fault escarpment (Figure 3), shows planar parallel stratification and minor wave-ripple cross-lamination and is attributed to deposition in a middle shoreface environment.

4.2. Deformation features

4.2.1. Neptunian dykes
Structures interpreted to be neptunian dykes occur in the mid-shoreface sandstone of the fault hanging wall.
Figure 3. The studied outcrop section and overlay drawing, showing the deformation features and vertical facies changes (log). Note the inset frame indicating the close-up detail shown in Figure 4.
On the basis of their extent and sediment infill, 2 types of dykes can be distinguished. One type is represented by the deep vertical dyke that extends from the base of the brecciated top part of the hanging wall to a depth of at least 3 m (Figure 3, left) and is probably longer in strike-parallel lateral extent. This fissure is filled with angular fragments of the fine-grained host sandstone, mainly 15–45 cm in size, and an interstitial matrix of fine- to coarse-grained sand. The medium to coarse sand component of the matrix resembles texturally the upper shoreface sandstone of the fault footwall, rather than the host hanging-wall rock. The other dyke type comprises relatively small fissures that form short wedges inclined at a high angle to the large dyke (Figure 3, left). These dykes are 7–45 cm in width and 50–150 cm in down-dip extent. The fissure-fill is mainly fine to coarse sand, with minor parabreccias (sensu Spalletta and Vai, 1984) composed of host sandstone angular fragments of ≤10 cm in size.

Neither the large nor the smaller dykes show any offset of the host rock (Figure 3, left), which means that they represent sediment-filled open fissures rather than small faults. The angular geometry of dykes implies host-rock brittle dilation, attributable to relaxation of the fault scarp. The dyke parabreccia component was formed by a brittle disintegration of the fissure sidewalls. The sandstone fragments show no evidence of weathering or dissolution, and their sharp edges suggest that the sand was well-compacted and perhaps slightly cemented prior to its rupture (see Cozzi, 2000).

The sedimentary infill of open fractures and cavities in seabed, referred to as neptunian dykes, consists of sediment derived from the sidewalls and from above (Winterer and Sarti, 1994; Cozzi, 2000; Črne et al., 2007). The origin of neptunian dykes is widely attributed to seismic activity (Playford, 1980; Winterer and Bosellini, 1981; Lehner, 1991; Montenat et al., 1991; Cozzi, 2000), although similar features have also been reported from the surfaces of submarine slides (Winterer et al., 1991; Winterer and Sarti, 1994). The formation of open fissures is related to extension in cohesive or consolidated sediment (Moretti
and Sabato, 2007), caused by such factors as a surficial stretching associated with the strain relaxation of fault scarps, slide extension gashes, diapir rising, or overloading (Montenat et al., 2007). Only this first factor is relevant in the present case, as the others can be precluded.

4.2.2. Convolute stratification
Convolute stratification occurs in the fine- to medium-grained sandstones of the fault footwall (Figures 3 and 4). The convolutions consist of disorderly anticlines and synclines, 10–35 cm in amplitude and 5–25 cm in width. Fold axial planes vary from vertical to subhorizontal and lack preferential orientation. The fold size decreases as the intensity of folding increases upwards, and the higher convolutions are commonly disharmonic with respect to the lower ones. The convolutions show a gradual transition to the underlying nondeformed parallel stratification, whereas their upward transition to a diffuse plumose structure above is relatively abrupt, although the contact itself is highly irregular and digitated (Figure 4). The gradational lower contact and the chaotic orientation of fold axial planes indicate an in situ deformation rather than a slump. The intensity of convolutions and the thickness of convoluted zone decrease away from the fault (Figure 3).

The formation of convolute stratification is widely attributed to a hydroplastic deformation of nonconsolidated sediment in a state of moderate or partial liquefaction (Lowe, 1975; Allen, 1986; Jones and Omoto, 2000; Rossetti and Goes, 2000; McLaughlin and Brett, 2004; Neuwerth et al., 2006; Montenat et al., 2007). Despite the apparent geometrical complexity of convolution structures, the deformation is most commonly driven by unidirectional stress, whether due to a seismic shaking or to gravity (Collinson, 1994).

4.2.3. Plumose structure
A diffuse and mainly amorphous plumose structure occurs in the footwall of the fault, at the transition from the strongly convoluted fine/medium-grained sandstone to the overlying coarse-grained massive sandstone (Figures 3 and 4). As in Matsuda’s (2000) description of similar features, the structure shows an interweaving of sediment from the upper and lower layers and consists of concave-upwards diffuse flow bands between the downwards-sagged pockets of the upper, heavier sand. The resulting interdigitation of sediment layers reaches up to 40 cm in amplitude (Figure 4), and the complexity of the plumose structure increases towards the fault (Figure 3).

The formation of plumose structures requires the applied shear stress to overcome the frictional shear strength of the deposit, which normally happens due to an increase of pore-water pressure that leads to liquefaction or fluidization (Elliot, 1965; Matsuda, 2000). This pattern of sediment transposition may be regarded as a combination of dish structures formed by dewatering and load structures caused by unstable sediment-density gradient (see Dżułyński and Walton, 1965; Collinson et al., 2006), resulting in a varied direction of transposition.

4.2.4. Injection dykes
At least one major injection dyke and a couple of minor ones, possibly its branches, are recognizable in the fault’s hanging wall (Figure 4). In contrast to neptunian dykes, these features extend obliquely upwards from the fault plane, wedge out blindly, and have apparently drained a mixture of fine- to coarse-grained liquefied sand from the zone of convolution and plumose deformation in the fault’s footwall. The main dyke is 5–25 cm wide and its dip-parallel extent from the fault plane is ~90 cm.

The emplacement of injection dykes is generally attributed to sediment fluidization combined with hydraulic fracturing (Montenat et al., 1991). These features are common in synsedimentary deformation horizons interpreted as seismites (e.g., Hesse and Reading, 1978; Seilacher, 1984).

4.2.5. Diapiric features
Small diapirs are associated with the edges of the upper large sandstone block (second block in Figures 3 and 4), which was apparently dislodged from the fault escarpment. These features are tight, shallowly protrusive folds of the underlying medium-grained sand, no more than 10–15 cm wide, with a penetration height of 20–30 cm. Their internal stratification is preserved, but tightly folded in accordance with the diapir shape. The deformation decreases towards the diapir root zone, where the folded strata pass into undisturbed ones.

The diapirs are synsedimentary deformation features that, in the present case, may have been caused by the impact shock of the fallen sand block or by its subsequent shaking.

5. Discussion
5.1. Trigger mechanism of synsedimentary deformation
The synsedimentary deformation structures owe their origin to the formation of open fissures in well-compacted, consolidated shoreface sand exposed in the fault hanging wall and to the process of sand liquidization (i.e., liquefaction or fluidization; sensu Allen, 1982) in the footwall. Liquidization increases pore-water pressure, breaks down the grain-supported sand framework, and causes a dramatic decrease of the sand shear strength, which effectively renders the sand mobile (Allen and Banks, 1972; Youd, 1973; Allen, 1982; Owen, 1987). This process is the main agent for the development of hydroplastic deformation structures in water-saturated cohesionless sediments (Lowe, 1975; Allen, 1986; Collinson, 1994; Rossetti, 1999; Jones and Omoto, 2000; McLaughlin and Brett, 2004; Neuwerth et al., 2006; Montenat et al., 2007; Moretti and Sabato, 2007). The triggering factor may be an
overloading or differential loading, gravitational slumping, seismic shaking, groundwater movement, cyclic and/or impulsive effect of storm wave loading, shear stress of an overriding current, or possibly an abrupt change in relative sea level or ground-water level (Jones, 1972; Lowe, 1975; Allen, 1982; Glennie and Buller, 1983; Mills, 1983; Owen, 1987; Molina et al., 1998; Moretti, 2000).

A seismic trigger is the most likely factor in association with a fault rupture, especially if alternative nonseismic causes can be precluded (Moretti, 2000). The deformed shoreface sandstones in the footwall part of the fault structure (Figure 3, right) were deposited by waves combined with storm-generated currents, and hence the origin of the convolute stratification and plumeose structure cannot be attributed to a rapid emplacement of sediment or an unstable, reversed sediment density gradient. The gradual transition of these deformation features to the underlying nondeformed horizontal parallel stratification, the disorderly orientation of convolutions, and the lack of lateral displacement indicate an in situ deformation and preclude both gravitational slumping and basal shear stress of an overriding current (see Collinson, 1994). There is no facies evidence of an abrupt major change in relative sea level and no sign of sediment disturbance by bioturbation, and also rapid groundwater movements are unlikely to have occurred in a littoral environment.

Substrate deformation can be triggered by the impact of a collapsing sand block, but the emplacement of the second large block in the outcrop section (Figure 3) apparently caused only shallow deformation directly beneath the block and at its edges. The block’s impact cannot explain the deeper zone of more extensive deformation, covered by little deformed to nondeformed sand (Figure 3). The thickness of the deformation zone increases towards the fault, which suggests a genetic link with the latter. Therefore, the laterally extensive deformation of seafloor sediment and the collapses of sand blocks from the fault escarpment are thought to have been triggered by seismic shocks.

It is worth noting that the deformation structures described here are similar to those reproduced experimentally by Kuenen (1958) and Sims (1975) in laboratory simulations of seismic shaking. Earthquake-produced ground fissures and liquefaction phenomena, from distinct horizons of load structures and convolutions to sand dykes, have long been recognized as seismites (Seilacher, 1969, 1984) and used to identify ancient earthquakes (e.g., Talwani and Cox, 1985; Allen, 1986; Anand and Jain, 1987; Obermeier, 1996; Vanneste et al., 1999; Matsuda, 2000; Rodriguez-Pascua et al., 2000; Obermeier et al., 2002; Singh and Jain, 2007). Apart from the similarity to experimental structures, Matsuda (2000) considered the following aspects of synsedimentary deformation to be diagnostic of seismic events: 1) the vertical narrowness and considerable lateral extent (synchronicity) of deformation zones; 2) a systematic lateral decrease in the deformation zone thickness and intensity; and 3) the evidence of sediment liquefaction or fluidization with minor horizontal transposition of sediment, implying in situ deformation. Rossetti and Gónes (2000) emphasized the diagnostic significant of deformation horizons separated by nondeformed sediment zones. All these criteria are met in the present case.

Obermeier et al. (2002) argued that if strong seismic shaking occurs in a sedimentary basin, liquefaction features should inevitably be formed. Earthquake-induced stresses cause a back-and-forth shear strain, which increases pore-water pressure, results in a loss of sediment shear strength, and causes deformation (Youd, 1973; Allen, 1977; Obermeier, 1996).

Many studies have focused on earthquake-induced soft-sediment deformation in an attempt to make inferences about the magnitude, epicenter location, and frequency of past earthquakes (Sims, 1975; Talwani and Cox, 1985; Allen, 1986; Anand and Jain, 1987; Obermeier, 1996; Vanneste et al., 1999; Matsuda, 2000; Rodriguez-Pascua et al., 2000; Obermeier et al., 2002; Singh and Jain, 2007). It has been argued that, for ground liquefaction to be triggered, the earthquake magnitude should be at least Mw 5, because weaker earthquakes lack sufficient strength and duration (Audemard and De Santis, 1991). Marco and Agnon (1995) suggested that seismic ground-level fault ruptures develop with earthquake magnitudes of Mw ≥ 5.5.

5.2. Chronology of seismotectonic events

Three distinct episodes of syndepositional seismotectonic activity can be recognized in the outcrop section on the basis of the stratigraphic succession of deformation features separated by renewed sediment deposition (Figure 3). They are reconstructed hypothetically in Figure 5 and are discussed below.

5.2.1. Episode 1

The first episode of syndepositional deformation involved the formation of an oblique-slip reverse fault (Figure 5A). The reverse fault lifted up well-compacted and possibly weakly cemented mid-shoreface deposits of the hanging wall above the seafloor level and contemporaneous upper-shoreface deposits of the footwall. Brittle deformation occurred along the fault plane and a subaqueous topographic escarpment was formed, with soft sediment stripped by waves from the hanging wall. Stress relaxation in the uplifted hanging wall caused its brittle disintegration, whereby penetrative open fissures developed and were filled in with wave-washed sand, forming neptunian dykes, while also the first large sandstone block fell off from the fault escarpment (Figure 5B). The sandy matrix of the
Figure 5. Cartoon reconstruction of the time series of mid-Miocene seismotectonic and sedimentation events recognized in the outcrop section (Figure 3). A) A reverse fault rupture forms at seafloor in medium- to coarse-grained sand of an upper shoreface environment. B) The fault escarpment reaches ~2 m in amplitude, with erosion and fracturing of the hanging wall and a large sandstone block dislodged from the fault scarp. C) The fallen block is buried by upper-shoreface sand, with little or no accumulation on the hanging wall. D) The second episode of seismotectonic activity, with partial liquefaction and strong deformation of footwall sediment. E) An erosional covering of the deformed sediment by upper-shoreface sand. F) The third episode of seismotectonic activity, with another large sandstone block dislodged from the fault scarp. G) The seafloor rapidly subsides and fine-grained mid-shoreface sand gradually buries the fault escarpment; the fault becomes extinct.
neptunian dyke parabreccias is texturally similar to coeval upper-shoreface sand, coarser than the host sandstone, which supports the notion of a contemporaneous derivation by wave action.

The development of neptunian dykes signifies extensional strain and is typically associated with normal faults (Montenat et al., 1991; Cozzi, 2000). The hanging wall in the present case was likely fractured due to its internal relaxation stresses (Montenat et al., 2007). Shock-induced dilation would likely contribute to the fracturing, as the formation of ground rupture may suggest an earthquake magnitude of $M_w \geq 5$ (Marco and Agnon, 1995). The amplitude of vertical seafloor displacement by the fault at this first stage is uncertain, estimated as at least ~2 m on account of the exposure of consolidated substrate in the hanging wall and the size of the first block detached from the escarpment.

5.2.2. Episode 2
The first stage of deformation was followed by continued sand accumulation in upper-shoreface regime in the accommodation space provided by the fault footwall, whereas little or no deposition occurred on the wave-swept elevated hanging wall. Sand draped and buried the first collapsed block (Figure 5C). The action of fair-weather and storm waves accumulated over 2 m of stratified sand before a second episode of seismotectonic deformation occurred, convoluting the strata and resulting in a plumose liquefaction structure with associated injection dykes (Figure 5D). It is uncertain if the hanging wall accumulated sand, but its further fracturing probably occurred. The hydroplastically deformed sediment was then slightly truncated by waves, probably smoothing the deformation relief, and was overlain by parallel-stratified sand (Figure 5E).

The thickness of the hydroplastic deformation zone decreases rapidly with distance from the fault (Figure 3), which suggests a ground-roll shock wave propagating away from the fault. The fault movement may have been incremental, with multiple shocks, since the multilobate geometry of convolutions (Figure 3) suggests that the strata were liquefied at least twice (Bhattacharya and Bandyopadhyay, 1998). Allen (1982) and Tuttle and Seeber (1991) reported on the occurrence of superimposed or cross-cutting liquefaction structures as indicative of multiple, closely-successive seismic events. One such superposition reportedly common in seismites (Matsuda, 2000) is the outward development of convolutions into a plumose structure with homogenized sand crown, indicative of stepwise liquefaction. Accordingly, this deformation episode is inferred to have involved at least 2 successive seismic events that followed shortly each other. The deformation included the injection of sand dykes into the hanging wall (Figure 4), which implies some accompanying fluidization.

The bulk amplitude of vertical displacement associated with this second seismotectonic event probably did not exceed ~1.5 m (see further inferences below). Although sediment liquidization might suggest an earthquake magnitude of $M_w \geq 5$ (Audemard and De Santis, 1991; Marco and Agnon, 1995), the actual magnitude could be lower, as the ground deformation seems to have been limited to the fault's close neighborhood.

5.2.3. Episode 3
Sediment accumulation resumed on the footwall and the deformed substrate was unconformably covered with a wave-worked, planar parallel-stratified upper-shoreface sand (Figure 5E). This depositional unit is 0.70 m thick near the fault and thins to 0.25 m away from it (Figure 3), which reflects mainly the footwall postdeformational topography. Another episode of the fault activity then occurred and the second large sand block collapsed from the fault scarp (Figure 5F). The block's impact deformed the substrate directly beneath it, forming small shallow convolutions and diapirc features. On the account of the fallen block's size, the hanging wall is estimated to have been elevated by an additional ~2 m and subsequently eroded to a depth of ~4 m before eventually becoming draped with sediment. A minor tsunami generated by the basin's northern-margin fault (Figure 1B) may have possibly contributed to the rapid erosion, as the basin concurrently underwent an abrupt subsidence and relative sea-level rise (see facies change in the log in Figure 3).

The rise in relative sea level instigated sediment deposition, also in the hanging wall, whereby the fine-grained mid-shoreface sand draped and eventually buried the fault escarpment (Figure 5G). The fault became extinct, as the sand cover lacks deformation (Figure 3). The total vertical displacement on the fault is estimated at ~5.5 m, since the footwall accommodated ~2 m of shoreface sand between the fall of the first and the second block, whereas the size of the latter block (Figure 3) indicates dislodgement from a scarp of at least 3.5 m high.

5.3. Evaluation of the possible cause of the reverse fault
The formation of the Karasu anticline (Figure 1B) in relation with the mid-Miocene activity of the Balkıfä reverse fault and the structural data of Yıldırım et al. (2011) from Neogene marine deposits show NNE–SSW shortening in the Sinop Peninsula. However, the dip azimuth of the studied reverse-fault (105°/65°) suggesting NW–SE compressional deformation is not compatible with this stress distribution of the region. The structural, morphometric, and plate kinematic analyses recently carried out in the region by Yıldırım et al. (2011) document the fundamental influence of the North Anatolian Fault (NAF) on the evolution of the Central Pontides (CP) and show that the transpressional deformation that is related to the NAF is distributed in the CP and extends to the
abyssal plain of the Black Sea. The NAF is the main driving mechanism for wedge tectonics and uplift in the CP and the onset of the dextral crustal shearing along the NAF controlled the tectonic inversion and NW–SE contractual deformation of CP Neogene basins (Andrieux et al., 1995; Yıldırım et al., 2011). Barka (1992) also pointed out that a wide dextral shear zone developed before the initiation of the NAF in the region, which caused NE–SW directed compressional structures including thrusts, reverse faults, and folds. Thus, the oblique-slip reverse fault may be attributed to the mid-Miocene activity of the dextral shear zone in the region, preceding full-scale development of the North Anatolian Fault Zone (see Barka and Hancock, 1984; Şengör et al., 1985; Andrieux et al., 1995).

The shallow-marine, neritic to littoral Miocene deposits in the Sinop Peninsula were deposited in a marginal trough developed at the edge of a tectonically inverted foreland basin of the Central Pontides. The development of the marginal trough and its sedimentation history indicate a transition from postorogenic rebound tectonics to the regional neotectonic regime in northern Anatolia. The middle Miocene shoreface deposits of the Sinop Formation show a well-preserved sedimentary record of syndepositional deformation associated with the escarpment of an oblique-slip dextral reverse fault and attributed to seismotectonic events in the basin. The events are thought to represent the early phase of neotectonics in the region. The deformation features include neptunian dykes, convolute stratification, plumose transposition structure, injection dykes, diapirs, and large sandstone blocks dislodged from the fault escarpment. The stratigraphic order of deformation structures deciphered from the outcrop section indicates 3 main episodes of seismotectonic activity separated by renewed sediment accumulation. The first episode formed a surface rupture, about a third of the total amplitude of vertical fault displacement (estimated at ~5.5 m), and dislodged a sandstone block from the fault scarp. The second episode is inferred to have involved at least 2 consecutive earthquakes, causing liquidization of footwall sediment. The last episode of fault activity detached another large sandstone block from the escarpment and was followed directly by abrupt basin subsidence, possibly accompanied by a minor tsunami. This study contributes to sedimentological research concerned with the deciphering of paleoseismic events from the stratigraphic record. Subaqueous features similar to those described in the present paper could potentially be common in association with ancient fault escarpments in tectonically active basins, but may have escaped recognition by being lumped with fault breccias or talus deposits.

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