Mineral Res EXpl. Bull., 128. 27-47. 2004

# ACTIVE SLUMPING OFFSHORE AMASRA (SOUTHWEST BLACK SEA) AND ITS RELATION WITH REGIONAL TECTONICS

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ABSTRACT.- The Bartın earthquake of 3rd September, 1968 provided the first evidence for active thrust faulting at the southern margin of the Black Sea. 500 km of 2D seismic reflection profiles acquired in this area show the presence of large scale slumps and landslips due to oversteepening of the shelf sediments, as a result of thrust faulting. The onshore drainage pattern is affected by interaction of the faulting with the geology, and this in turn affects the offshore shelf. Kilometre-scale bathymetric features on the sea-floor indicate a previous mass movement of a relatively coherent block of sediment, which has subsequently been partially buried. The source location for the block is visible on the geological map of the area. Incisional and erosional features in Pliocene sediments far from today's coastline and drainage systems suggest that water level was lower at the time they were formed than at present day.
Key words: The Black Sea, seismicity, seismic reflection study, slumping.

## INTRODUCTION

The seismicity and tectonics of Turkey have long been interpreted in terms of the plate movements in between Africa, Arabia, Eurasia and Anatolia (McKenzie, 1972; Alptekin, 1973; Dewey, 1976; Şengör, 1979; Şengör and Canitez, 1982; Şengör et al., 1985). Most of the large earthquakes occur on two fault systems, the North Anatolian fault (NAF) and the East Anatolian fault (EAF), along which Turkey is moving westward. The focal mechanisms of these events are generally consistent with the relative plate motions in these locations. The focal mechanism of the earthquakes that are away from plate boundaries, however, are difficult to relate with the tectonic movements and this type of earthquakes occur on the Black Sea coast of Turkey.

In Anatolia, three major tectonic provinces have been recognized on the basis of their predominant structural styles and associated strain patterns (Şengör et al., 1980; 1983; 1985) (Fig. 1). They are 1) the East Anatolian high plateau, 2) the Central Anatolian ova regime, 3) the Aegean qraben system. There are three secondary neotectonic regions except for these major tectonic provinces: 1) Black Sea region north of the North Anatolian fault (The North Turkish Province), 2) Thrace, 3) Adana basin - Isparta angle.

The neotectonic regime of the areas lying north of the North Anatolian fault and comprising a major portion of the Pontide paleotectonic unit (Ketin, 1966) has not been studied in detail; largelv because of the neotectonic structures of the region are neither as active, nor as spectacular, nor as abundant as in other parts of Turkey.

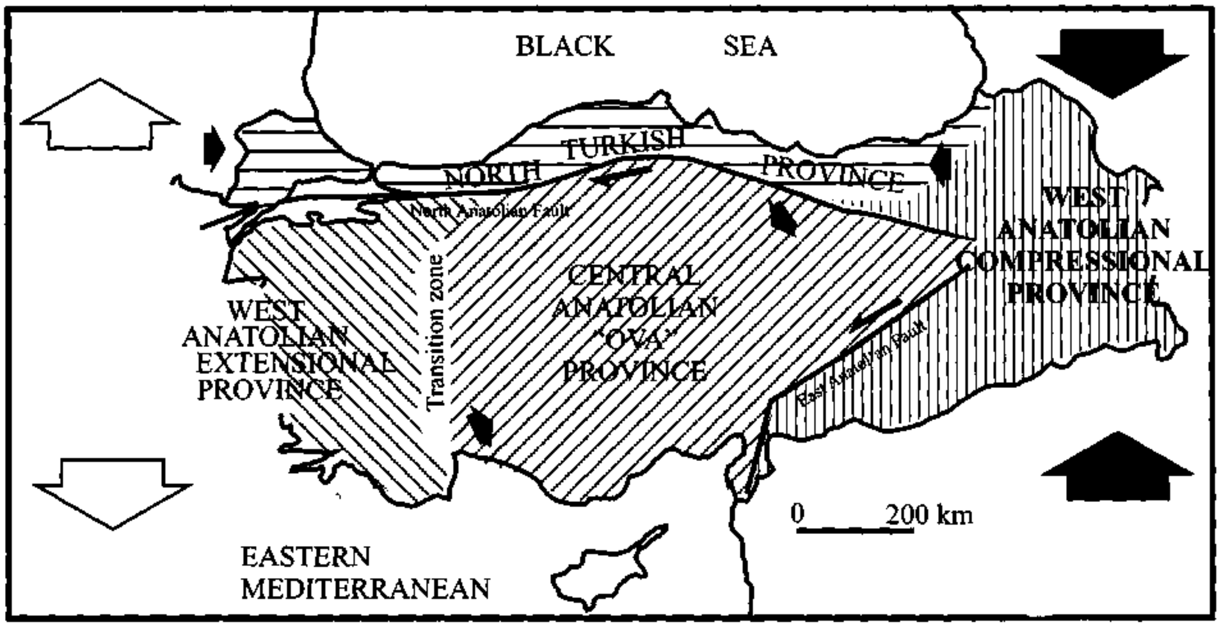


Fig. 1 - Neotectonic provinces of Turkey. Black arrows denote compressional areas while the white arrows denote extensional areas (after Şengör et al., 1985).

Indeed, the areas lying between the Black Sea shore and the North Anatolian fault have long been viewed as inactive. However, a study of the neotectonics of the North Turkish region is of critical importance for the solutions to a number of problems such as the interpretation of the weak regional seismicity and the understanding of numerous, discontinuous, scattered neotectonic features recognizable on land, on aerial photographs and in the field.

In September 1998, seismic reflection data were acquired in the southwestern Black Sea. The cruise was planned to image a possible scarp or fault which might have been associated with repeated slip on the Bartın fault. It is also aimed that, the data acquired will be useful to understand the regional tectonics as well as the new findings on the above-mentioned fault.

## MORPHOLOGY AND SEISMICITY OF THE BLACK SEA

The Black Sea is a large (423 000 $km^3$), deep (2000 m) semi-isolated marine system situated to the north of Turkey, between two Alpine mountain ranges, the Caucasus and the Pontides (Fig. 2). It is thought to be a remnant of the Tethys oceanic system, which existed between Eurasia and Anatolia. It is linked to the Mediterranean by the Strait of Istanbul (the Bosphorus), the Sea of Marmara and the Strait of Çanakkale ( the Dardanelles), and contains water of below normal marine salinity, a result of freshwater input from major rivers such as the Dnieper and the Danube. Extreme changes in the sedimentation regimes of the Black Sea occur as a result of eustatic changes in sea level. The stratigraphic history of the Black Sea is stil a

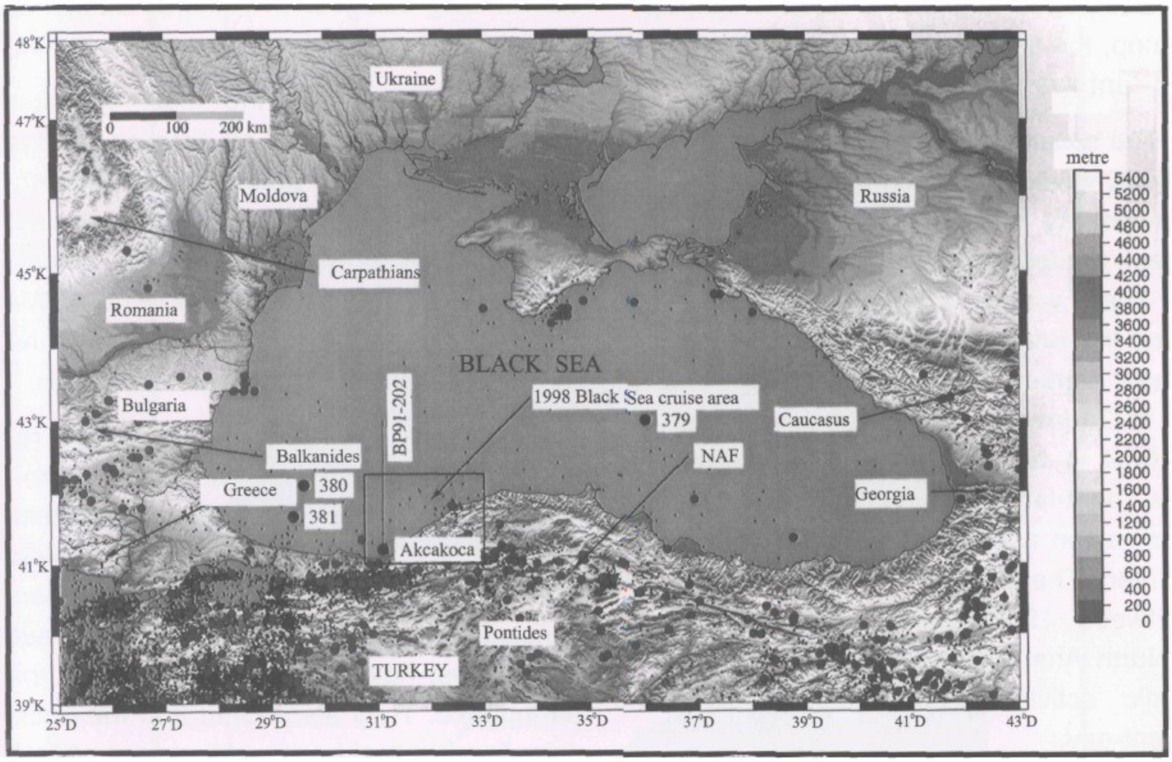


Fig. 2 - Map showing the seismicity in and around the Black Sea in between 1900-2000 (from NEIC catalogues). Small black dots are for epicenters Mw>3.0. Dairk gray circles are for Mw>5.0. DSDP wells (379, 380, 381) and Akgakoca-1 are the major well drilled in the area. KAF denotes the North Anatolian fault. The study area is shown in rectangle.

subject of debate due to the effects of complex extensional and compressional tectonic regimes (Letouzey et al., 1977; Okay et al., 1994; Spadini et al., 1996). Although the Black Sea is an intra-continental sea, it has geological features of a small ocean.

From morphological point of view, there are two types of margins in the Black Sea. There are no wide shelves on the eastern and southern Black Sea, the sea suddenly deepends after a narrow shelf. On the other hand, there are wide shelves on the north and west, following the basin slope, abyssal plain is reached. These features, from only morphological point of view, resemble the Pacific and Atlantic type continental margins (Erinç, 1984).

The development and evolution of the shelves are related to the rivers and their material input. Along the Black Sea coast, the sedimentation rate is not high, this is due to a) the presence of rather small but highly incising rivers, b) high topographic relief, c) lack of estuaries to preserve the deposits, d) the presence of narrow shelves, and e) the presence of canyons along which the sediments were transported to the deeper sections of the basin. There is a rather flat morphology in between İğneada and Ereğli. On Thrace peninsula, there are small rivers carrying the material from

Istranca Massif The most significant river in this region is Sakarya. East of Ereğli, up to Sinop, the shore is steeper and the most significant river in this area is Filyos.

The seismic activity in the Black Sea is relatively weak. In the central part of the sea the seismicity is negligible, however, on the shores moderate earthquakes were recorded. There are two important seismic belts around the Black Sea, these are situated in northern Turkey (the North Anatolian fault) and in Caucasus region. The North Anatolian fault is an east-west trending, highly active, right-lateral strike-slip fault. In Caucasus region active folding and thrusting is observed. The distribution of the epicenters in between 1900-2000 (Fig. 2), shows that the North Anatolian fault has a remarkable seismic activity in historical times and present-day.

There is some tectonic activity which is not located on the major controlling structures in the area, the most notable example being the shallow focus seismicity along the Turkish margin of the Black Sea. Eight moderate sized historical earthquakes have been reported in the area (Soysal et al., 1981). The Bartın earthquake of 3rd September, 1968 (Ms=6.6) provided the first seismological evidence that active thrusting is occurring at the southern margin of the Black Sea.

## THE BARTIN EARTHQUAKE OF 3rd SEPTEMBER, 1968

On 3rd September 1968, at 10h 20min 36 sec GMT, the two small townships of Amasra and Bartın were shaken by an earthquake of magnitude 6.6. The earthquake has caused casualty (official statistics: 24 dead and hudreds of wounded) and heavy damage in these towns and the surrounding villages. In Ankara, Bursa, Istanbul and Samsun the earthquake was felt.

The epicentre of the main shock is located 10 km north of Amasra, in the Black Sea by ISC (International Seismological Center) (Fig. 3a). ISC recorded nine aftershocks following the main shock, magnitudes varying between $4.0<m_b<4.6$ in the area, five of these in the same day and the remaining in the following four months.

Ketin and Abdüsselamoğlu (1969), reported that the epicenter was around Akpınar and Kirlik (Fig. 3b) although no fresh fault breaks were observed in the field following the earthquake. The researchers concluded that the rather younger fault family must have been responsible for the earthquake. They also noted that the. coast near Amasra had been uplifted some 35-40 cm during the earthquake as a proof the movement of these faults, as well as some observed cracks on the alluviums and landslides near the epicentral area. Landers (1969) reported a tsunami in the Amasra Bay just after the earthquake.

A number of isoseismal maps were prepared for the Bartın earthquake (Albers and Kalafatçıoğlu, 1969; Ketin and Abdüsselamoglu, 1969; Ergünay and Tabban, 1983). The maximum intensity on these maps, despite some small variations, is VIII MMS.

McKenzie (1972), Kudo (1983), Şengör et al. (1983) and Jackson and McKenzie (1984) provided fault plane solutions for the earthquake. Except for Şengör et al. (1983), the researchers indicated strike-slip faulting with reverse fault component for the earthquake. Şengör et al. (1983), using the

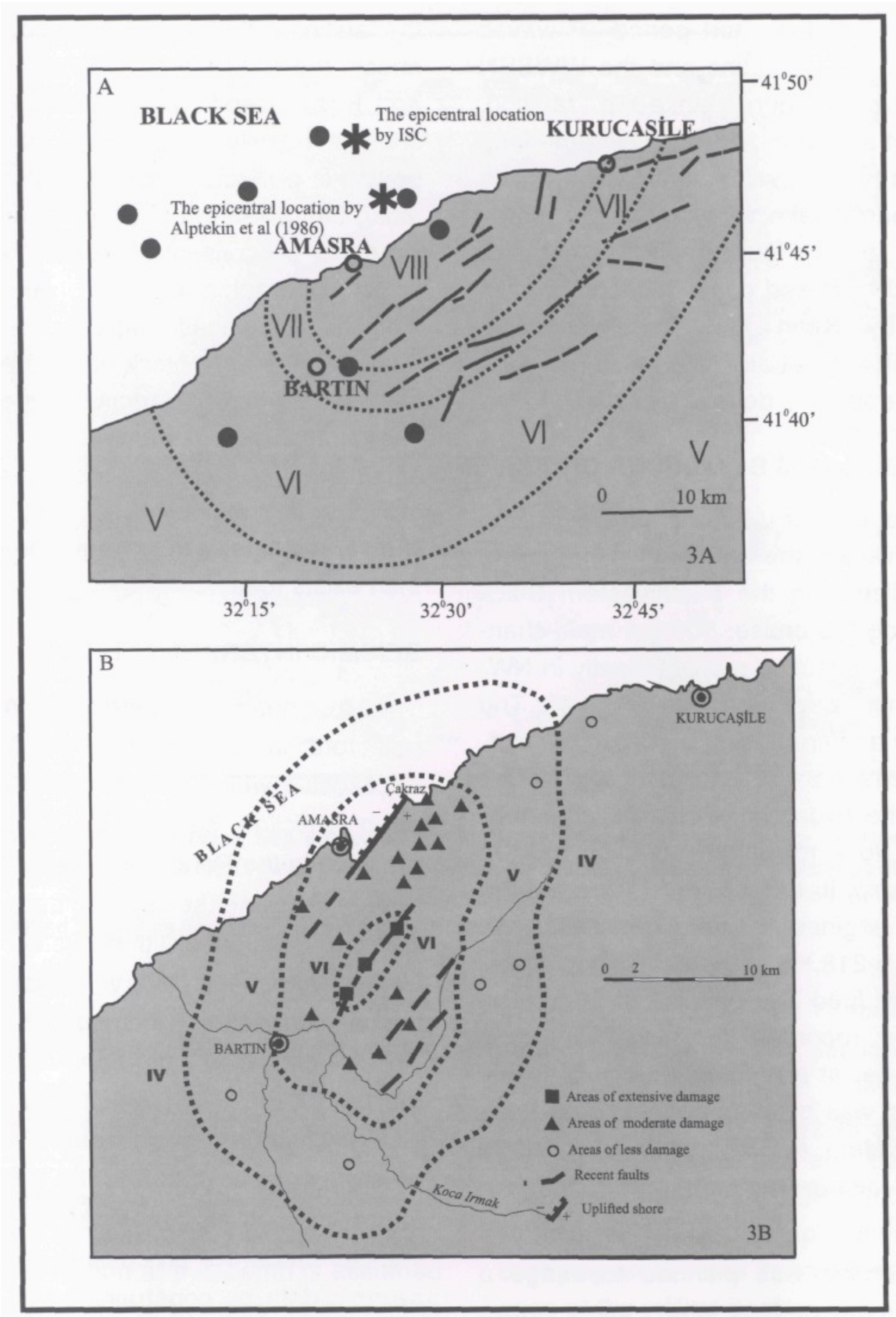


Fig. 3 - A) The epicenters proposed for the Bartın earthquake. Isoseismals are shown by broken lines. The stars denote the epicenters proposed by ISC and Alptekin et al. (1986). Black circles show the epicenters of the aftershocks of Alptekin et al. (1986).

B) Seismotectonic map of the epicentral region of the Bartın earthquake (after Ketin and Abdüsselamoğlu, 1969). Contours are isoseismal lines enclosing areas of varying intensities according to the Mercalli-Sieberg scale (Roman numerals indicate intensities).

first motion of the short period P waves reported in ISC bulletins and the WWSSN data, indicated pure strike-slip faulting. Alptekin et al. (1986), re-examining the long and short period seismograms, proposed that the earthquake was due to the movement of a reverse faulting which is compatible with the raised coast observations in Amasra by Ketin and Abdüsselamoğlu (1969). Alptekin et al. (1986) also has relocated the epicentre defined by ISC.

## SEPTEMBER 1998 BLACK SEA CRUISE

In September 1998, a cruise using MTA Sismik-1 in the area shown in figure 2 was realized in the southwestern Black Sea. During the cruise, 500 km multi-channel seismic reflection profiles mostly in NW-SE direction were acquired (Fig. 4). The data were acquired using a 10-gün, 23 litre, tuned generator-injector compressed air source. The hydrophone had 96 channels, extending to a maximum ofset of 1500 m from the ship. In order to reduce noise from the ship's engines, the data were bandpass filtered at 8-218 Hz at the acquisition stage. Shots were fired at a spacing of 50 m. The data were recorded to a two-way time (TWT) of 5s, at a sample interval of 4ms. Navigation was by Differential Global Positioning System (DGPS), with an onshore VHF radio beacon, allowing a location accuracy of 5 m.

The cruise was planned to image a possible scarp or fault which might have been associated with repeated slip on the Bartın fault. The data show evidence for large scale slumping and landslip due to oversteepening of the continental shelf, as a result of thrust faulting. The onshore drainage pattern is affected by the interaction of the faulting with the geology, and this in turn affects the shelf offshore. Where the Filyos and Koca rivers enter the Black Sea, the shelf is heavily incised and no overstee pening is possible. Several kilometers east, along the coast, the shelf oversteepens and slumping processes can occur. Kilometre-scale topographic features on the sea-floor indicate a previous mass movement of a relatively coherent block of sediment, which has since been partially buried. Other incisional and erosional features in Pliocene sediments far from today's coastline and drainage systems suggest that the water level at the time of their formation was lower than exists today.

## SEISMIC INTERPRETATION

An important constraint on dating reflectors in the seismic profiles is the Akcakoca-1 well. This well drilled on a small ramp anticline caused by Tertiary compression in the Pontides. It penetrated Quaternary-Pliocene sediments above a thick Eocene sequence, thin Paleocene and Upper Cretaceous (UC) volcanics and tuffaceous carbonates (Robinson et al., 1995). The Akgakoca-1 stratigraphy can be tied to the profile Bla98-04, which helps to correlate stratigraphic sequences between the offshore seismic data and the onshore geological units. It also provides a good initial velocity model for processing the reflection seismic data by constraining the depths of strong reflectors. Nonetheless, it lies close to the shore, in a zone affected by Tertiary compression and submarine to subaerial erosion, both of which hinder extrapolation of stratigraphy away from the wells into the Black Sea basin centre.

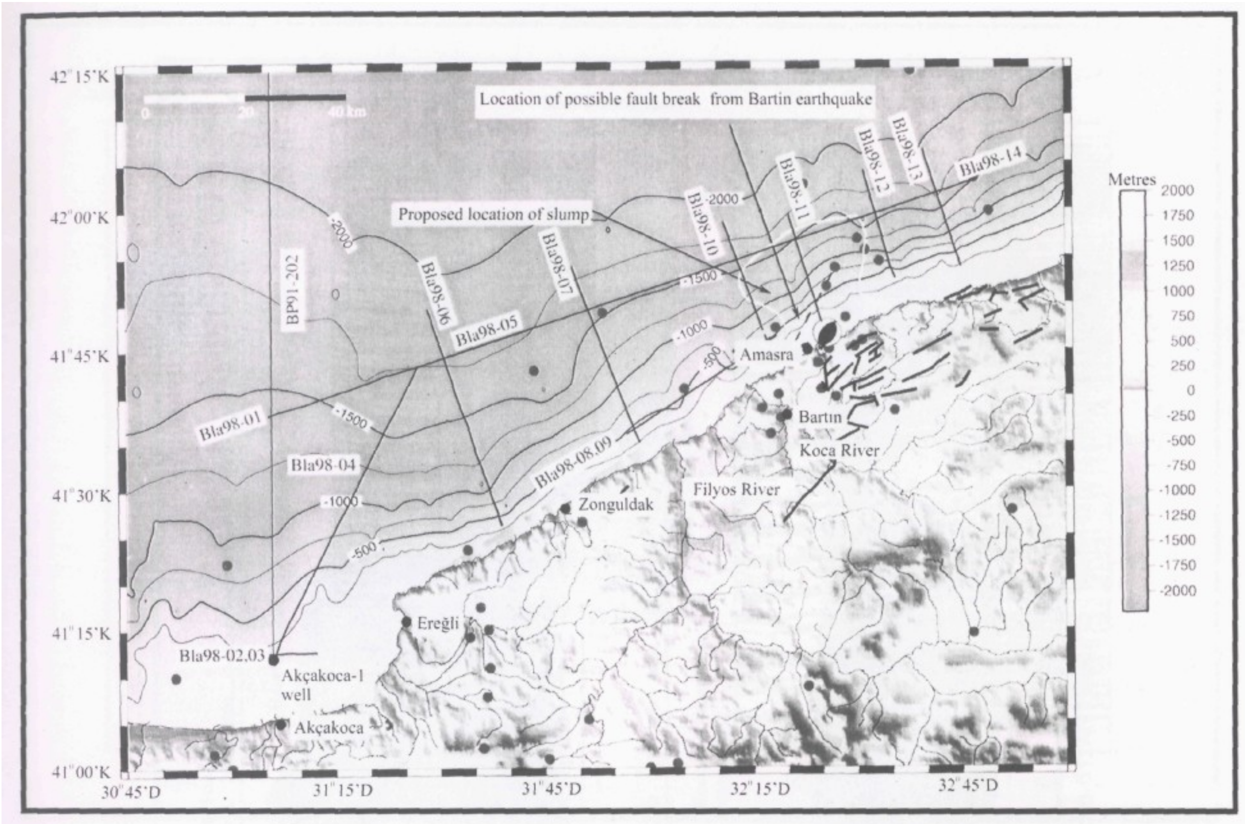


Fig. 4 - Map of the 1998 Black Sea survey area. Thin black lines show the reflection profiles. Earthquake epicenters from the NEIC catalogue are plotted as black dots. The Bartın fault plane solution is plotted, along with the theoretical location of a fault break (dashed line). The location of the proposed slump is marked by a white line. Onshore faulting is taken from Alptekin et al. (1986).

The well data are presented in fig. 5, along with the relevant time section of the seismic profile through the borehole, a depth migrated section and a plot of the interval velocities used to stack the data. The thin Paleocene horizon is not visible in the section at the location of the well, and the main reflection at this depth is assumed to be from the UC carbonates. Further north along the section, where a unit appears between the base of the Eocene and the top of the Cretaceous, it is interpreted as a thickening Paleocene unit.

Line Bla98-04 ties the Akçakoca-1 well to the main regional seismic profile (Bla98-01) (Figs. 4 and 6). The Cretaceous basement is visible in the section, and has undergone extensional faulting with evidence of syn- and post-rift deposition in the overlying Paleocene and Eocene units. The section between CDP 8200 and 9400 exhibits a classic onlap pattern with the overlying sediments terminating along the edge of the infilled basin.

The deepest identifiable reflection in this dataset is interpreted as the top of the Upper Cretaceous. This geological unit is also widely observed outcropping onshore, and is discussed later. At the southern end of the seismic section the UC reflection dips

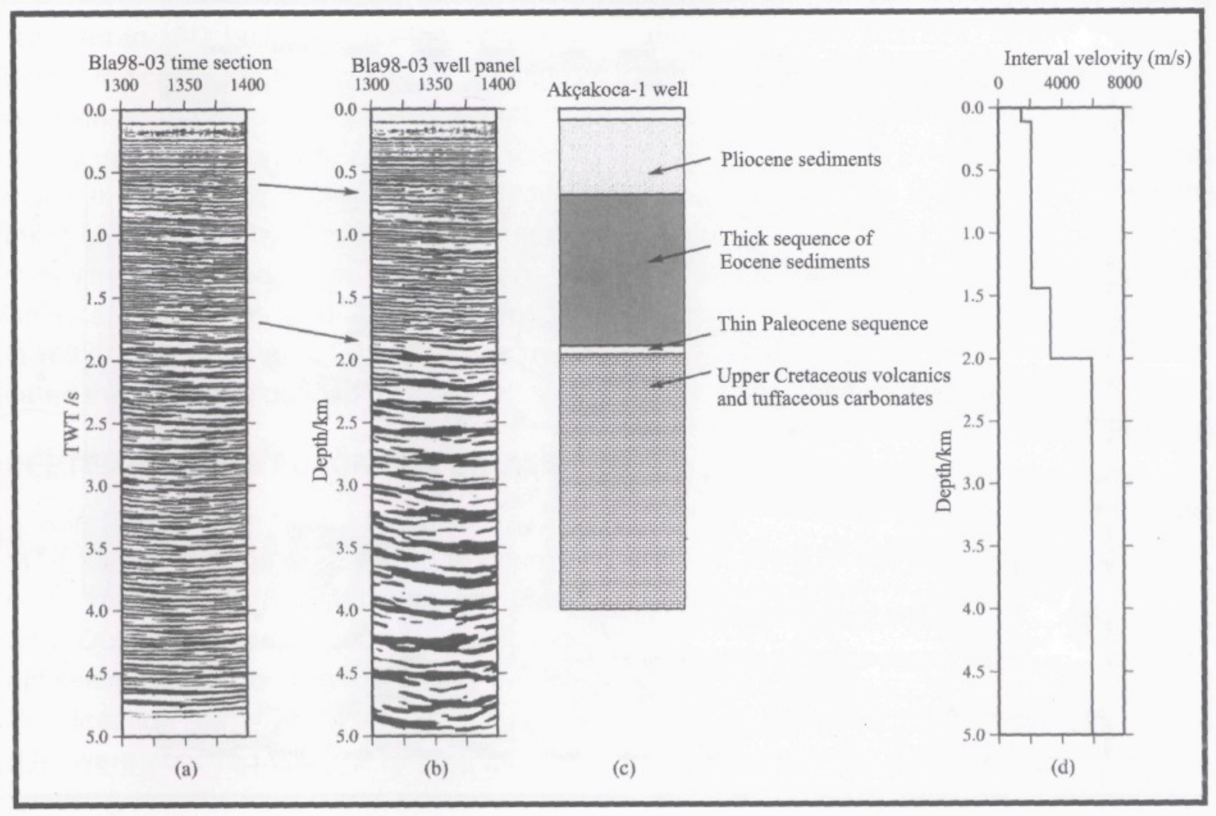


Fig. 5 - The Akgakoca-1 well. Shown here are a panel of: (a) Time-migrated reflection seismic profile data through the Akgakoca-1 well, (b) Depth conversion, (c) A sketch of the stratigraphy in the well, and (d) The interval velocities used to stack and depth convert the data.

down to the north, away from the coast. The Akgakoca-1 well was drilled into a small anticline in a compressional area in the shelf, and the northern limb of this anticline is observed from CDP 1200-2400 in the section. Further north, the UC horizon becomes flat lying, and enters an extensional regime. The northernmost end of the line shows a buried normal fault block which is also visible on the regional survey of the area (Robinson et al., 1995).

The sediments overlying the UC basement have had a complicated history. The well enables identification of the base of the Eocene and Pliocene sequences, neither of which is conformable with their respective underlying units. A thickening unit is observed further north between the UC and the base of the Eocene, and is interpreted as the Palaeocene basin fill at the northern end of the line. The top of the Eocene terminates against the UC at CDP 9200, in a characteristic onlap style, so that when the intersection with the regional line (Bla98-01) is reached, there remains only Pliocene or post- Pliocene deposits, unconformably overlying the Cretaceous basement. There are no Oligocene or Miocene horizons at the well site. The DSDP wells (Fig. 2) further offshore show that the region was subaerial during the late Miocene (Stoffers et al., 1978). It seems likely there-

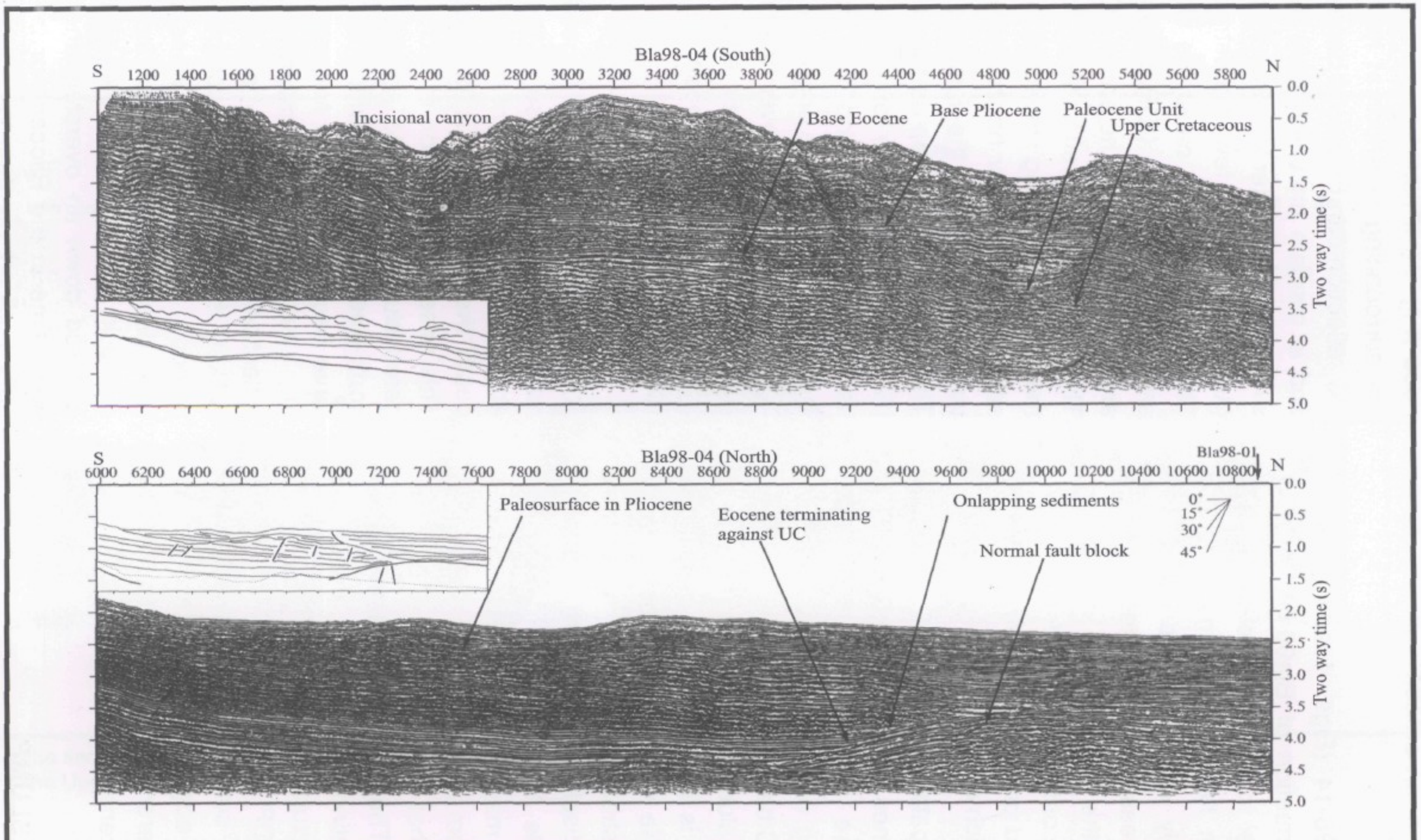


Fig. 6 - Line Bla98-04, tying the Akçakoca-1 well to the regional line Bla98-01. An incised canyon is marked between CDP 1400-3000.

fore that any Oligocene-Early Miocene deposits have been eroded from the top of the anticline.

Line Bla98-14 (Figs. 4 and 7) is an easterly continuation of Bla98-01, approximately parallel to the coastline. It shows a transition from flat lying sediments in the west to highly disturbed sediments and slumps in the east. The UC reflector can be traced along the entire length of this line, giving good control on the depth of the basement, and providing a useful tie to the other lines which cross this one.

The palaeo-surface previously identified on lines Bla98-04 and Bla98-01 is continued here from CDP 19800-14800. The underlying unit has a similar reflective style to the UC basement in the area. Given the nature of slumps and mass movement of sediment, it is likely that some units in the reflection profile will not be in their correct place in the stratigraphic column, as they have been displaced from their original location. While the material forming the paleo-surface may be displaced Cretaceous carbonates from a landslide, we know from the well tie that it is resting on Pliocene sediments. The relatively undeformed basement continues to the east, underlying different structural features. The section between CDP 15000-10000 shows examples of small kilometre-scale listric faulting at the edges of incised canyons, as well as coherent slumped and rotated blocks of Pliocene sediment.

## A PLIOCENE SLUMP

The most noticeable bathymetric feature visible on profile Bla98-14 is the large hill, some 19 km wide, between CDP 8000 and 5000 (Fig. 7). It rises almost 750 m above the surrounding sea-floor at its peak. The UC reflection can be traced beneath the hill and is overlain unconformably by deposits similar in character to the Pliocene seguences visible a few kilometres laterally. These deposits first onlap the sides of the small basin, between CDP 6800 and 6000, and then drape over the entire UC reflector. The large hill has a reflective style similar to that of the underlying limestone, though it is sitting on a particular horizon which can be traced laterally into the Pliocene deposits. The deposits also onlap up the flanks of the hill, on both sides. The top of the hill exhibits some canyoning, and it is thought that the canyons are due to ongoing incision, as no obvious sediment cover has been deposited since this unit was emplaced. The flanks, however, have been buried under some 300 m of sediment.

This feature, between CDP 8000 and 5000, is interpreted as a largely coherent, massive slump. The base of the slump is above the last coherent reflector underneath the pile, which is of Pliocene age. This raises the question of technically triggered slumping, due to seismic activity in north-western Turkey, which is thought to have begun in the late Miocene (Görür et al., 1995). From its similarity with dated reflection packages elsewhere in the data, it is likely that the slump is composed of Cretaceous limestone. This limestone would have decoupled from the underlying Jurassic rock, somewhere further to the south. and slid down an oversteepened shelf, at some time in the Pliocene or later, ending up emplaced on then recent sedimentary deposits. It has since been partly buried by up to 300 m of younger sediments.

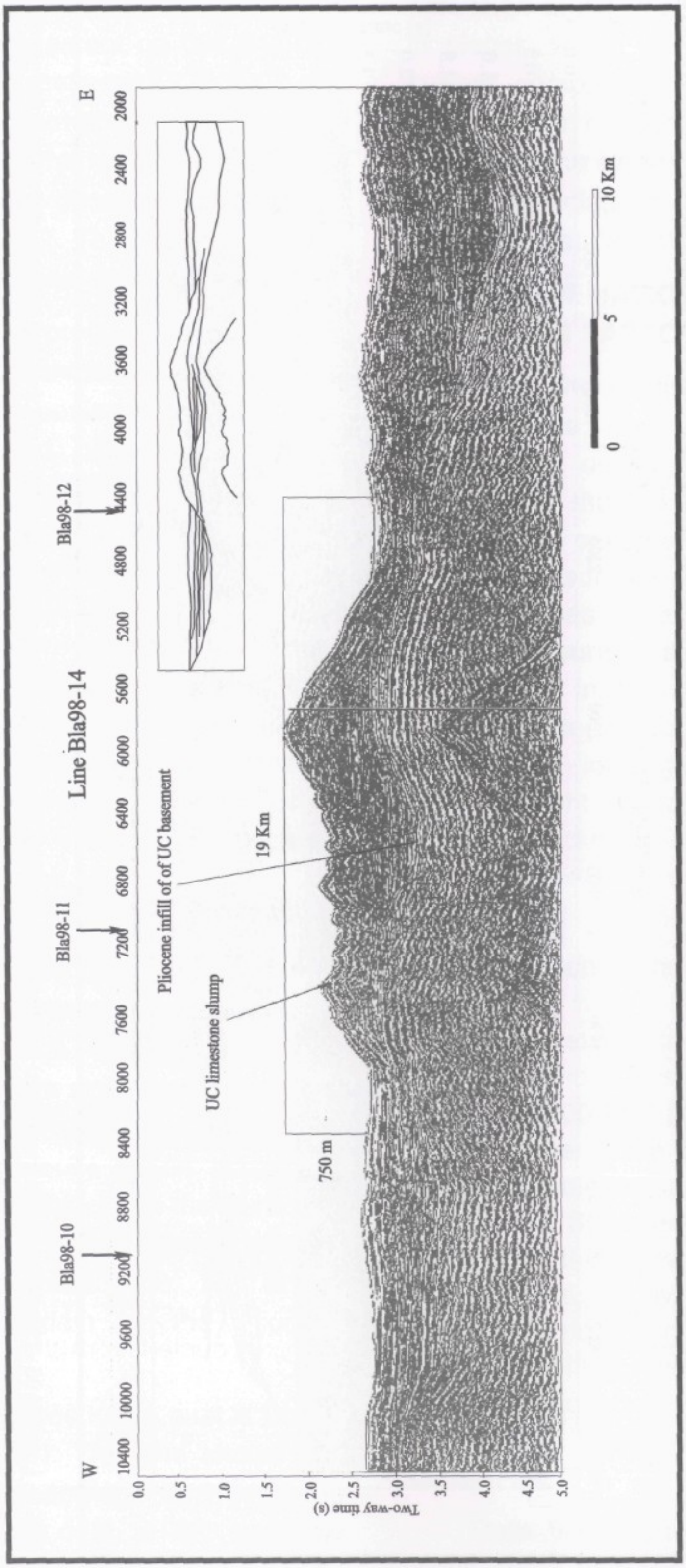


Fig. 7 - The section of Line Bla98-14 relevant to the slumping. The dotted lines on the seismic profile mark the boundaries of the slump. The Upper Cretaceous (UC) basement reflector is visible across the entire length of this line.

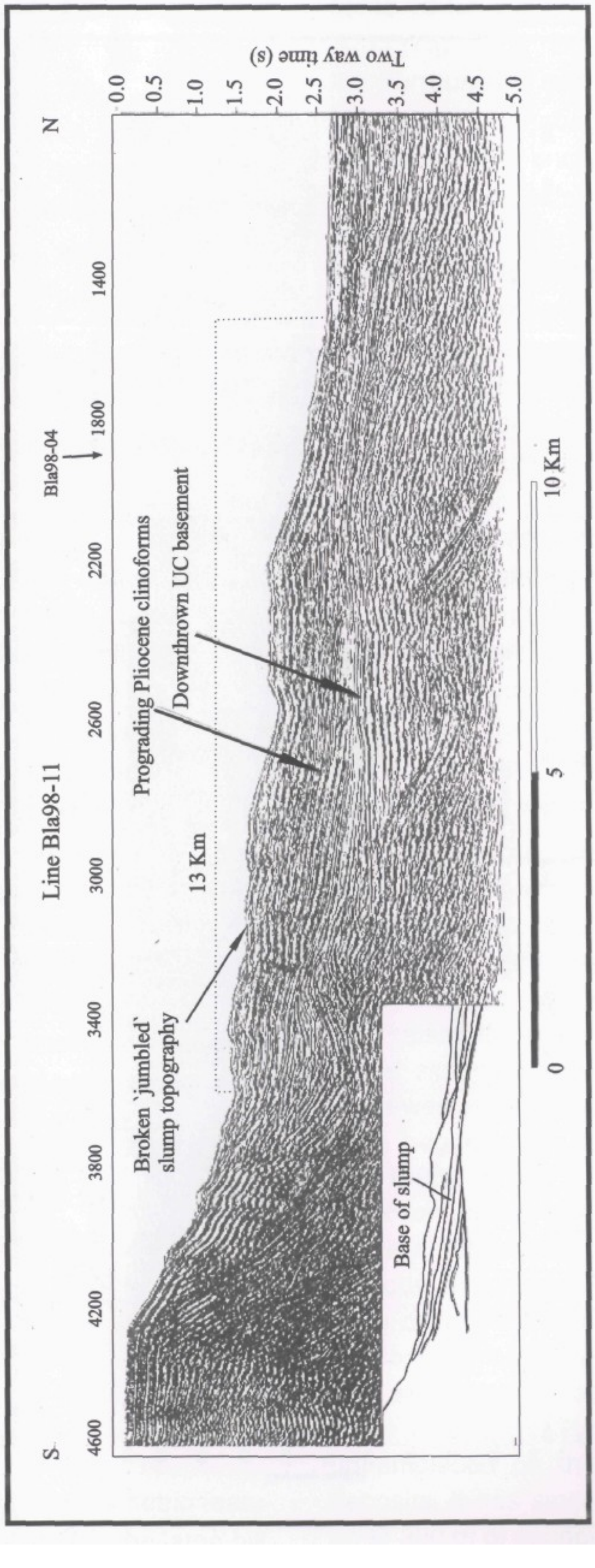


Fig. 8- Line Bla98 -11, shot over the theoretical location of the surface break of the bartın earthquake. Although the fault plane is not visible. the slumping and the over are cut by the profile.

Bla98-11 (Figs. 4 and 8) intersects Bla98-14 at CDP 7100, on the proposed slump. The UC basement on this line dips down to the south, between CDP 3200 and 2200. The unconformable overlying sediments dip down to the north. Both of these units fit with the observed stratigraphy interpreted in figure 7. The basement shows evidence of extensional faulting.

This reflection profile also exhibits all the classic characteristics of a slump caused by oversteepening of the shelf sediments. The jumbled, blocky nature of the topography starts at CDP 3800 and extends out to CDP 1800, giving a good indication of the thickness and offshore extent of the feature. Again, the toe of the slump extends under the sea-floor, where it has been partially buried by more recent sediments. Underneath the slump structure, Pliocene sediments are observed, terminating against the basement. The Pliocene deposits form a series of dipping, prograding clinoforms which downlap onto the basement.

Two other perpendicular lines can be used to delineate the lateral extent of the slump. Bla98-10 (Figs. 4 and 9) lies 12 km to the west, but has a significantly shallower slope. Between CDP 2000 and 3400 is the highly disturbed, blocky region associated with the slump. The 'blocks' in the slump are about 1 km in size. The basement reflector is also less deformed here, but is still unconformably overlain by the Pliocene deposits.

Sixteen kilometres to the east is Bla98-12 (Figs. 4 and 10). This line shows evidence of oversteepening, and has the steepest slope in the data, at -38°, probably related to the seismicity in the area. The UC basement has been downthrown by extension at the basin margin, but is folded in places, which may be the result of the current compressional tectonics. There is no evidence of slump related topography. The toe of the slope is at CDP 3000, and as before, is onlapped from the north by the basin filling sediments.

## SLUMP KINEMATICS AND THE AFFECTING FACTORS

Using the constraints provided by the seismic data, and the bathymetry, it is possible to define the perimeter of the slump, and thus to estimate its area and volume. The slump area is marked in Fig. 9. It covers a surface area of 400 km$^2$. The slump reaches a maximum height of 750 m above the surrounding sea-floor, from the data visible in the sections, although appears to extend some 250 m beneath the current sea-floor, giving it a total topographic height of about 1 km. Estimating an average thickness of 500 m across the whole feature gives a total slump volume of 200 km$^3$.

The volume, though significant, is low compared with the largest reported submarine slide, the Agulhas slide off southern Africa. This has an estimated volume of 20000 km$^3$ (Dingle, 1977). Slumps on the flanks of the Hawaiian islands have recorded volumes that exceed 5000 km$^3$, as does the Storegga slide off the continental slope of the Norway (Hampton and Lee, 1996). The area covered by the Black Sea slump is not large, particularly when compared within other slumps such as the 1080 km$^2$ coverage of a slump from the Malaspina Glacier, in Alaska (Carlson, 1978).

Data from current slumps and landslides show that the ratio of their run-out

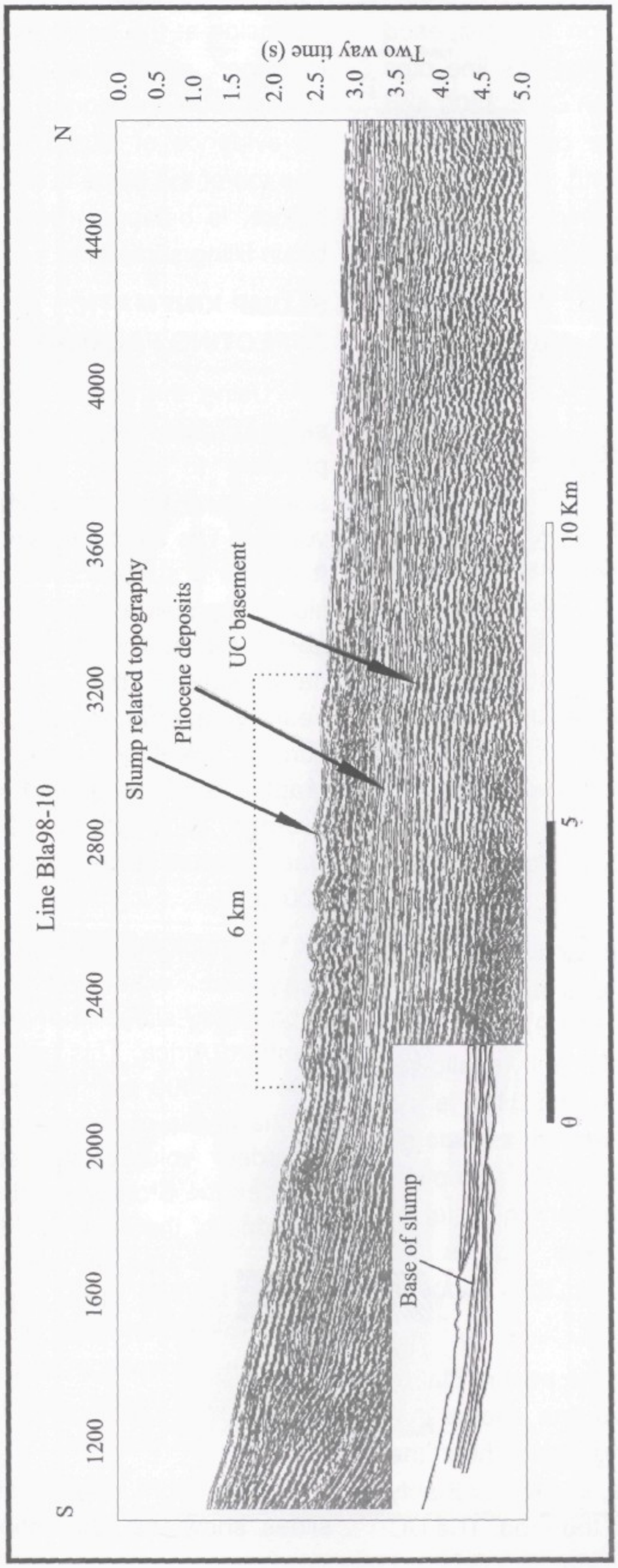


Fig. 9- Line Bla98 -10, shot 12 km west of Line Bla98-11. The characteristic topography related to slumping is visible on the line

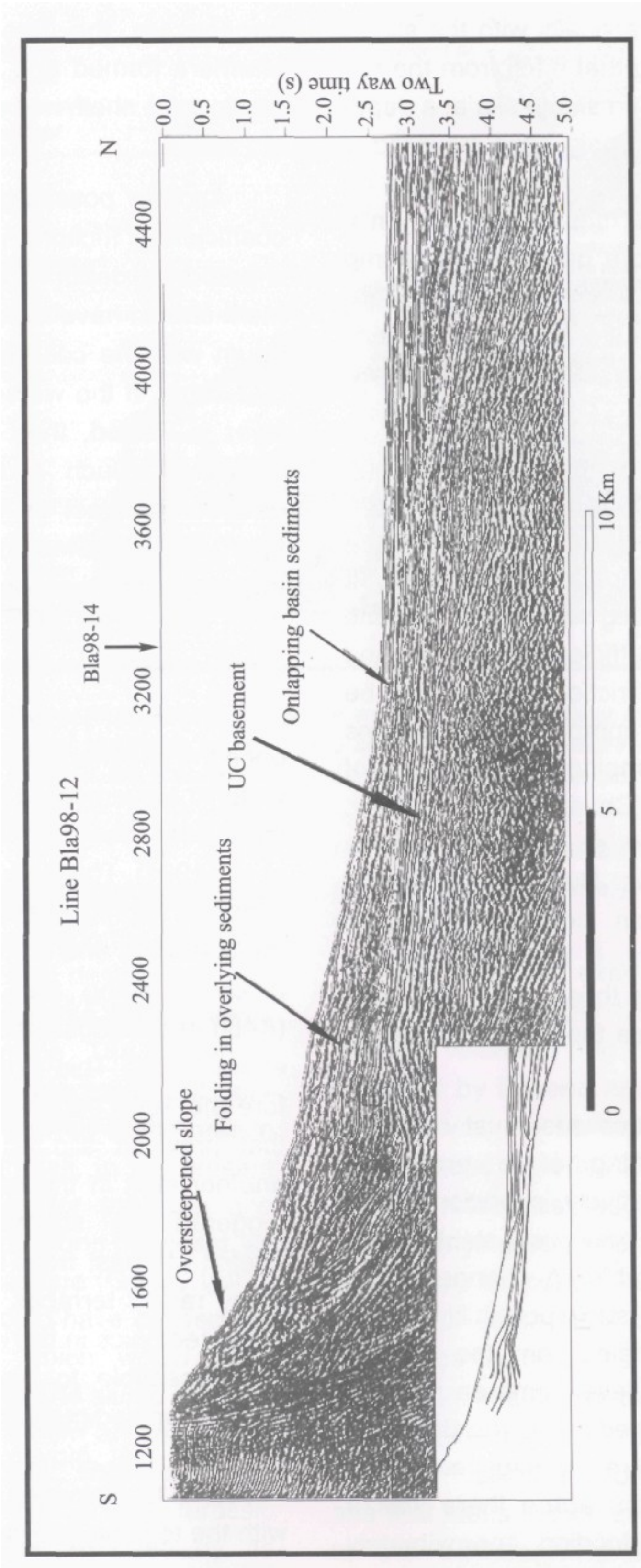


Fig. 10- Line Bla98 -12, shot 16 km east of Line Bla98-11. No slumping has occurred at this part of the coast, though the shelf is severely oversteepened.

length (L) to drop height (H) varies considerably and systematically with the slump mass (M). Assuming that it fell from the top of the shelf (the Bartın slump sits at a depth of 1500 m), then it could have travelled a distance between 20-60 km depending on the volume of the mass (250±50 $km^3$). Based on the onshore geology, this slump does not seem to have moved much more than 20 km, it appears to have had less than average mobility, based on observations of other slumps.

The above mentioned model based on a mass sliding down a slope is, however, too simplistic to explain all the phenomena associated with sumps and landslides. It does indicate that long runouts correspond to small effective coefficients of friction, and suggests that a low friction model might be a more appropriate approach. The reasons for initiation of slumping is a subject of ongoing research (Chuang and Greeley, 2000; Chamberlain et al., 2001). The force of gravity alone is not likely to be the sole cause of failure on continental margin slopes (Ross, 1971). If it were, then the large slumps such as those observed would not be possible, since the sediments would fail much earlier.

Some other processes must occur to enable oversteepening of a previously stable slope. Alternatively, conditions must change such that a previously stable slope then becomes unstable. A change in the water depth is one such possibility. It is known that depending on the drastic change in the water level some parameters (i.e. density of the sediment, resistance to failure, pore pressure of sediment) may change. Most authors agree there was a substantial marine flooding approximately 8000 years ago, when a connection to the Aegean via the Bosphorus and Sea of Marmara formed and over 100 000 $km^3$ of continental shelf was submerged (Ryan et al., 1997).

Another possibility is a change in the coefficient of friction. A layer of sediments at a particular depth in a submarine environment should have a pore pressure in equilibrium with the confining water pressure at that depth. If the water level drops, and the layer is sealed, then it will become overpressured. Such a layer could become unstable and act as a low-friction horizon upon which slip could occur. Some previously observed submarine slumps have been associated with rapid changes in sea level (Hampton and Lee, 1996).

The most likely trigger for this slump, once the sediments had become unstable is tectonic movement in the area. Such effects have been observed elsewhere by Hazlett et al. (1991). The Bartın earthquake of 1968 occurred directly between this slump and the coastline, and an uplift of 30-40 cm was recorded along the coast near Amasra (Ketin and Abdüsselamoğlu, 1969; Alptekin et al., 1986). This slump occurred well before the 1968 earthquake, but it is possible that previous slip on the same fault, or another fault in the same area could have triggered the submarine landslide. The stretch of coast from Ereğli to Amasra has many raised terraces, probably due to the thrust tectonics in the area, and is known to be susceptible to landslides. The sparse post-siump sedimentary drape on the slump suggests that movement occurred in the Pliocene to holocene, which is consistent with the tectonic activity in the area today.

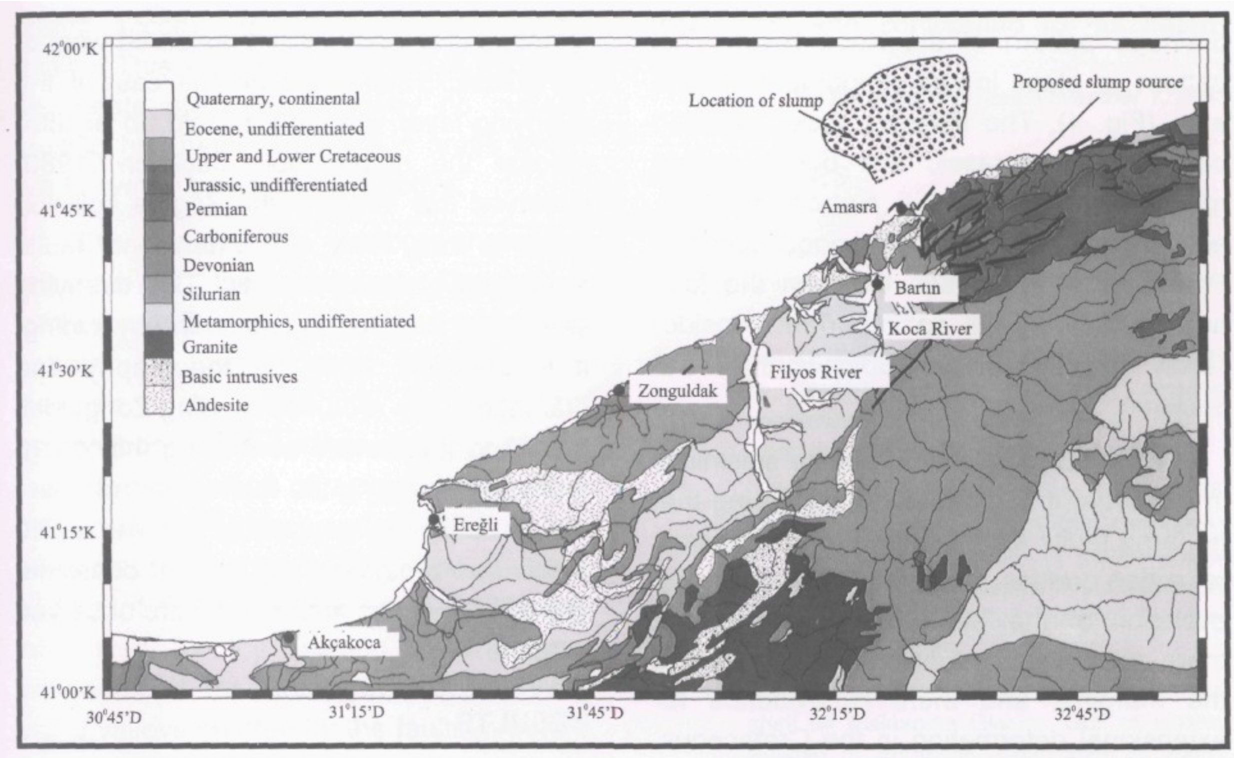


Fig. 11-1/500.000 scale geological map of the Zonguldak area (Tokay, 1964). Most of the area is covered by Cretaceous and younger sediments except for the area east of Bartın where underlying Permian and Jurassic rocks are exposed: this is assumed to be the source of the offshore slump.

## DISCUSSIONS

The reflection profile evidence suggests that at some time during the Pliocene, a large volume of material destabilised and slumped to a position some 20 km away from the present coastline. The trigger for the event was probably tectonic, but consideration must also be given to the reason for the oversteepened shelf. In the present day, there is evidence for thrusting and folding in both directions along the coast, as well as uplift of the coastline (Görür, 1997). Folding and faulting could have caused the oversteepened shelf, which would have subsequently collapsed. The slump material is probably Cretaceous from an analysis of the reflection seismic data. This means that in the source area, the underlying Jurassic rocks would have been exposed by its removal. Examination of the onshore geology shows a likely source location for the slump (Fig. 11).

The dominant exposure in the Zonguldak area is of Upper Cretaceous carbonates, and volcanics (Fig. 11). This is overlain in places by Eocene and Quaternary deposits, as well as having been intruded by later igneous material. To the east of Amasra the Cretaceous has been removed, and the underlying Jurassic and Permian rocks have been exposed. This exposure covers an area of some 1000 $km^2$, and is affected by surface faulting. The faults have a trend which matches that of the instrumental fault plane solutions, and are parallel to the current coastline. The different lithology means that faults in this area are able to break to the surface, and are controlling the drainage pattern of the Koca River. This pattern is clear in the topography of the area (Fig. 4). The Permian rocks exposed beneath the Jurassic are being eroded much faster than is the carbonate cover, and show up the faults in good contrast. This type of interaction between the fault and the lithology is an important consideration in active areas (Goldsworthy and Jackson, 2000).

With the high level of diffuse seismicity in this area, it is at first sight surprising that no large faults are observed on the seismic reflection profiles. It seems likely, given the evolution of the western Black Sea, that there should be extensional faults around the margins, and there is evidence for extensional deformation in the Cretaceous. The later thrusting associated with the closure of the Tethys Ocean, and continuing today, could be accommodated on reactivated normal faults. If this is the case, then since most of the faults on the southern margin of the Black Sea are north dipping, and related to previous extension, we might not see any large fault breaks on the reflection profiles as they would be too close to the shore to be covered by the acquisition. Another possibility is that whichever way the fault plane is dipping, it may not have broken along a plane to the sea-floor. On land, when thrusts reach the surface, they diverge from the fault plane dip at depth, and break closer to the vertical. The nose of the hanging wall often breaks off and masks the true location of the fault, causing a blind thrust (Dunne and Ferrill, 1988).

Given that the seismicity is distributed throughout this region, it is likely that faults exist all along the coast, even though no surface breaks have been observed west of Zonguldak. This would be the case if the overlying layer was able to fold, to accommodate the deformation. Jackson (1980) observed this effect in the Zagros collision zone, in Iran. Here, old extensional faults are reactivated as thrusts. The. overlying sediments take up the shortening by folding. It is apparent from the topography and drainage in the area surrounding Zonguldak that blind thrusts are controlling the topography, and causing the folding. In this case, the thrusting is responsible for the uplift, which has produced the present coastline, and this is why no major faults are observed offshore.

## RESULTS

A large recent slump, of estimated area 400 $km^2$ and volume 200 $km^3$, has moved material approximately 20 km offshore, east of Amasra. The slump is interpreted as being of Cretaceous material, and this analysis is consistent with an examination of the onshore geology. The aspect ratio and run-out distance of the slump are comparable with other features observed elsewhere, though the slump was less mobile and retained its internal structure better than have most other slumps: we suggest that this is due to its composition of competent Cretaceous limestones.

The pattern of onshore faulting, drainage and geology shows that the dominant recent tectonic deformation in the area is thrust faulting. These thrusts are responsible for oversteepening of the shelf, which then collapses, most likely because of a tectonic trigger. It is likely that some of the thrusting is being taken up on reactivated extensional faults, with the overlying sedimerits accommodating the shortening by folding. This explains the folding of Pliocene sediments offshore, and why no surface faulting has been observed in areas which are known to be seismically active.

The present day geology shows an absence of the dominant Cretaceous formation to the east of Bartın. The faulting pattern in the exposed Jurassic and Permian rocks highlights the important contribution which surface lithology makes towards recognition of active faulting. It is only in this area that surface faulting has been mapped on the ground. The drainage pattern of the rivers has been significantly influenced by the faulting. The eastern tributaries of the Koca River are running along valleys created by the faulting, before turning north and draining into the Black Sea. This means that for much of the coast to the east of Bartın, there is very little sedimentary input. When the major rivers enter the Black Sea, incision and canyoning are visible in the sea-floor relief of the reflection profiles, and no oversteepening has occurred.

## ACKNOWLEDGEMENTS

We thank the captain, officers, crew and scientific party of the R/V MTA Sismik-1, for their assistance during the Black Sea cruise in September 1998 which was repeated many times due to very bad weather conditions prevailed. This research was supported by General Directorate of Mineral Research and Exploration (MTA) of Turkey, TÜBİTAK (in context of project number YDABÇAG 198Y085), Royal Society and British Council in funding this research. The first author thanks Esen Arpat and Aral Okay who reviewed the manuscript and contributed for its improvement.

*Manuscript received November 1, 2003*

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