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The role of the Early Tertiary Ulukışla Basin, southern Turkey, in suturing of the Mesozoic Tethys ocean

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Abstract: The Maastrichtian–Late Eocene Ulukışla Basin is representative of the tectonic and sedimentary evolution of prominent Early Tertiary basins in central Anatolia, including the Tuzgölü and Şarkışla basins. The Ulukışla Basin overlies an ophiolitic mélange of Late Cretaceous age between the Bolkar Carbonate Platform to the south and the Niğde–Kırşehir metamorphic massif to the north. The basin stratigraphy records successive phases of transgression, subsidence, volcanism, evaporite deposition, deformation and uplift. Subsidence curves are consistent with an extensional (or transtensional) basin origin terminated by uplift. The Ulukışla Basin includes a thick succession (*c.* 2 km) of Upper Paleocene–Lower Eocene basaltic to andesitic submarine pillow lavas, lava flows, volcaniclastic rocks and intercalated limestones. Whole-rock XRF chemical analysis indicates a within-plate origin, with a marked subduction influence, believed to be inherited rather than contemporaneous. The Ulukışla Basin formed after Late Cretaceous ophiolite and mélange emplacement and ended with Late Eocene emergence, deformation and onset of Oligo-Miocene non-marine deposition. We propose that Late Cretaceous ophiolite and mélange emplacement reflect initial ocean basin closure. This was followed by a long period (*c.* 30 Ma) of microcontinental adjustment, including possible strike-slip, palaeorotation and suture tightening, during which the Early Tertiary Ulukışla Basin developed. Possible driving forces were regional slab pull (of a relict subduction zone) or oblique (diachronous) convergence. Suture tightening was complete by Late Eocene time marked by collisional deformation and regional uplift.

Keywords: Eastern Mediterranean, Neotethys, Late Cretaceous, Early Tertiary, basin analysis.

It is now widely recognized that many orogenic belts, including much of Asia, resulted from the assembly of microcontinents, oceanic arcs and other units (e.g. Şengör & Natl'in 1996). Anatolia is an ideal region for study of continental assembly, as the interaction of microcontinents is assumed in most modern reconstructions (Robertson & Dixon 1984; Şengör *et al*. 1984; Dercourt *et al*. 1986, 1992; Stampfli *et al*. 2001). Previous work on the tectonic history of microcontinents in Anatolia has focused on rifting and ocean basin opening, either from Gondwana to the south (e.g. Robertson 2000; Stampfli *et al*. 2001), or from Eurasia to the north (Ustaömer & Robertson 1997; Yılmaz *et al*. 1997). However, less attention has been paid to the processes of orogenic assembly, potentially including suture tightening, strike-slip and palaeorotation.

Within Anatolia, the Mesozoic Tethys ocean was separated into several oceanic strands separated by microcontinents, of which arguably the largest was the Northern Neotethys (Şengör & Yılmaz 1981). In one interpretation only a single Northern Neotethys existed (Izmir–Ankara–Erzincan branch; Şengör & Yılmaz 1981; Göncüoğlu 1986), whereas in another (Görür *et al*. 1984; Görür & Tüysüz 2001) a discrete Inner Tauride Ocean existed between the Tauride Platform in the south and a microcontinent in the north (centred on the Kırşehir Massif). The Northern Neotethys started to close by dominantly northward subduction after Early Cretaceous time (Şengör & Yılmaz 1981; Yılmaz *et al*. 1997). One interpretation is that subduction continued into Mid-Tertiary time giving rise to peripheral magmatic arcs and related fore-arc basins (Görür *et al*. 1984). The latter are now represented by a series of regionally extensive, variably deformed Lower Tertiary sedimentary basins, including the Tuzgölü, Şarkışla, Haymana–Polatlı and Sivas Basins and the Ulukışla Basin in the south, the main subject of this paper (Fig. 1). In the tectonic model of Görür *et al*. (1984), ocean basin closure was delayed until Late Eocene time when these basins were deformed, uplifted and eroded. Recently, however, some studies, mainly of intrusive igneous rocks associated with the Early Tertiary basins, have suggested that suturing of the Northern Neotethys occurred during Late Cretaceous time (e.g. Boztuğ *et al*. 2001), implying that the Early Tertiary basins are of post-collisional origin. Okay *et al*. (2001) have proposed that continent collision in western Turkey occurred in the Paleocene, although a further important phase of deformation occurred in the Mid-Eocene. On the other hand, arc volcanism in the eastern Pontides continued intermittently until the end of Paleocene or Early Eocene times (Yılmaz *et al*. 1997).

The main objective of this paper is to test the popular model of Paleogene subduction and arc volcanism culminating in Late Eocene continental collision (Görür *et al*. 1984), using evidence mainly from the contemporaneous Ulukışla Basin. Study of the early stages of the central Anatolian basins is complicated by erosion and covering with later, post-collisional sediments (e.g. Tuzgölü Basin). However, the Ulukışla Basin is exceptionally well exposed and can be taken as representative of the Early Tertiary central Anatolian basins. Until now there have been no modern studies of the tectonostratigraphy, depositional history, the nature of associated volcanic rocks, or the deformation history of this important basin. In this paper we present new information on each of the above aspects and use the assembled information to test alternative hypotheses for the timing and processes of collision in central Anatolia. We will conclude in favour of an interpretation in which initial suturing ('soft collision') took place in latest Cretaceous time, then was

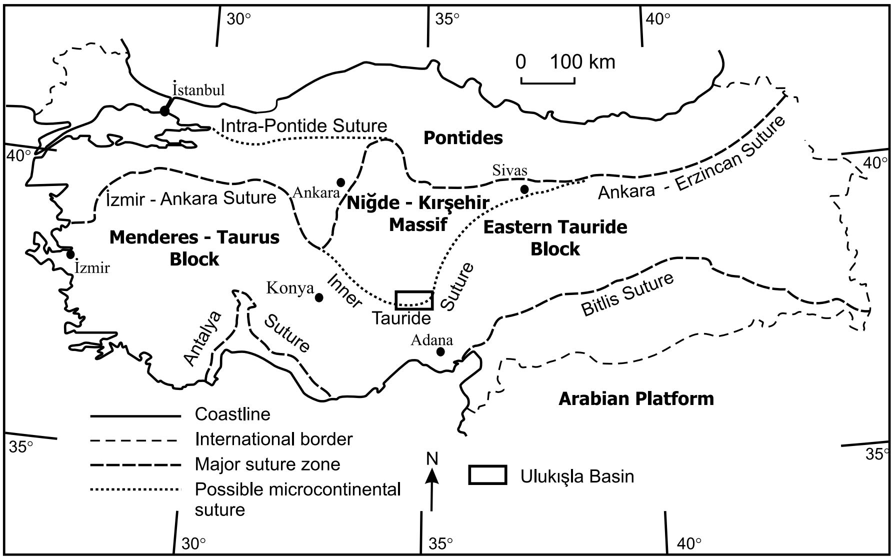


Fig. 1. Neotethyan sutures of Turkey, modified from Okay (2000).

followed by suture tightening, strike-slip, possible block rotation and terminal subduction (over *c*. 30 Ma), until no further convergence could be accommodated by this means. 'Hard collision' involving regional deformation and the initiation of mountain building ensued after Late Eocene time. Further convergence of Africa and Europe was then accommodated by northward subduction of a southern strand of Neotethys further south.

Tectonic setting and lithostratigraphy of the Ulukışla Basin

The Ulukışla Basin (Figs. 2 and 3) contains >5 km of sedimentary and volcanic rocks of latest Cretaceous to early Oligocene age (Blumenthal 1956;Demirtaşlı *et al*. 1975, 1984; Atabey *et al*. 1990; Görür *et al*. 1998; Clark *et al*. 2001). It is bounded to the north by the Niğde–Kırşehir Massif (termed the Central Anatolian Crystalline Complex by Göncüoğlu *et al*. 1991)), and to the south by the Bolkar Carbonate Platform. To the east the basin ends abruptly against the left-lateral Ecemiş Fault zone and associated Oligo-Miocene basinal sediments (Yetiş 1978, 1984; Koçyiğit & Beyhan 1998; Jaffey & Robertson 2001). The boundary of the basin to the west is less clear, as Lower Tertiary basinal sediments dip below continental Neogene to Recent sediments. However, the Ulukışla Basin is genetically linked with the Tuzgölü Basin Complex to the NW (Görür *et al*. 1984; Fig. 2).

During this work a new stratigraphic scheme was devised (Fig. 4) and is detailed below. The initial stratigraphy of the Ulukışla Basin was constructed by Oktay (1973, 1982) and Demirtaşlı *et al*. 1984, utilizing micropalaeontological age data. We have revised the stratigraphy in line with our lithostratigraphic findings and the result of new biostratigraphic dating in collaboration with N. İnan and K. Taslı (Mersin University), to be published elsewhere.

Upper Cretaceous Alihoca Ophiolitic Mélange

The Lower Tertiary Ulukışla Basin is unconformably underlain by Upper Cretaceous ophiolitic rocks, mainly mélange, emplaced onto the Bolkar Carbonate Platform (Fig. 3) in latest Cretaceous

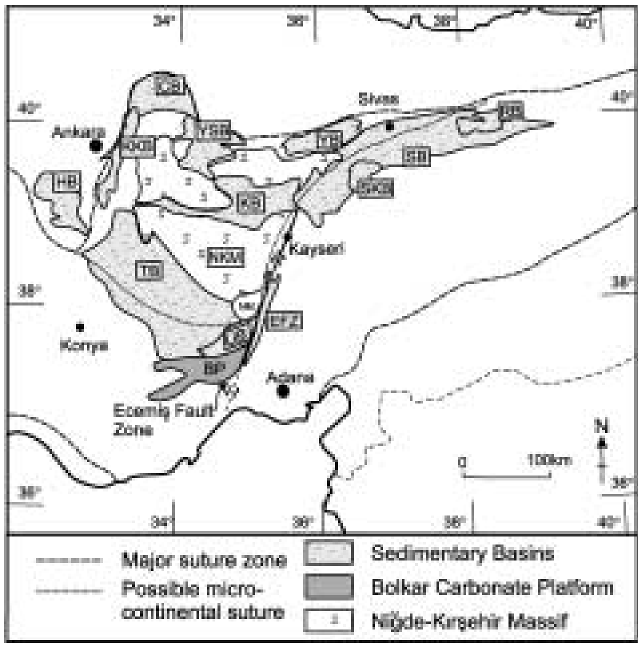


Fig. 2. Major sedimentary basins of central Anatolia, with Bolkar Carbonate Platform (BP) and Niğde–Kırşehir Massif (NKM) also shown. UB, Ulukışla Basin; TB, Tuzgölü Basin; HB, Haymana Basin; KKB, Kırıkkale Basin; CB, Çankiri Basin; YSB, Yozgat–Sorgun Basin; KB, Kizilirmak Basin; YB, Yildizli Basin; RB, Refahiye Basin; SB, Sivas Basin; SKB, Şarkışla Basin; EFZ, Ecemiş Fault Zone. (Adapted from Görür *et al*. 1998.)

time (Demirtaşlı *et al*. 1984; Lytwyn & Casey 1995; Dilek *et al*. 1999). Dykes cutting the metamorphic sole of the ophiolite are dated by the $^{40}Ar/^{39}Ar$ method at 90.8 ± 0.8 Ma (Dilek *et al*. 1999). The Alihoca Ophiolitic Mélange mainly crops out along the southern margin of the Ulukışla Basin (Fig. 3). It comprises ophiolite-derived blocks and clasts, commonly serpentinite and gabbro (several centimetres to 15 m in size) set in a red sheared clay-rich matrix, together with large (*c*. 10 m) blocks of crystal-

Fig. 3. Simplified geological map of the Ulukışla Basin based on this study andDemirtaşlı *et al*. (1984). (See text for locality descriptions and Fig. 14 for cross-sections.)

line neritic limestone (Demirtaşlı *et al*. 1984). Around Maden village (Fig. 3, locality 1) other blocks and clasts include conglomerates, debris-flow deposits, pelagic limestones, red cherts, turbiditic sandstones and blueschists. The pelagic carbonates are dated as 'early Senonian' to Maastrichtian in age (Demirtaşlı *et al*. 1984). We interpret the mélange as an emplaced accretionary prism, related to northward Neotethyan subduction.

A locally intact ophiolitic unit crops out ENE of Alihoca village (Fig. 3, locality 2) beginning with a pegmatitic gabbro containing feldspar and large amphibole crystals, up to 4 cm in diameter. Numerous diabase dykes cut the pegmatite. Up-section, diabase dykes dominate for 100 m laterally and exhibit regular cross-cutting relationships. Above, a coherent unit of basaltic breccias includes subordinate lava flows, overlain by pelagic limestone with a low-angle contact (Çiftehan Unit, Fig. 4). Near the base, dark calcareous shale and thin- to medium-bedded volcaniclastic sandstones are interpreted as turbidites sourced from underlying extrusive ophiolitic rocks. Upwards, the pelagic limestones are in local faulted contact with Lower Tertiary volcanogenic units of the Ulukışla Basin.

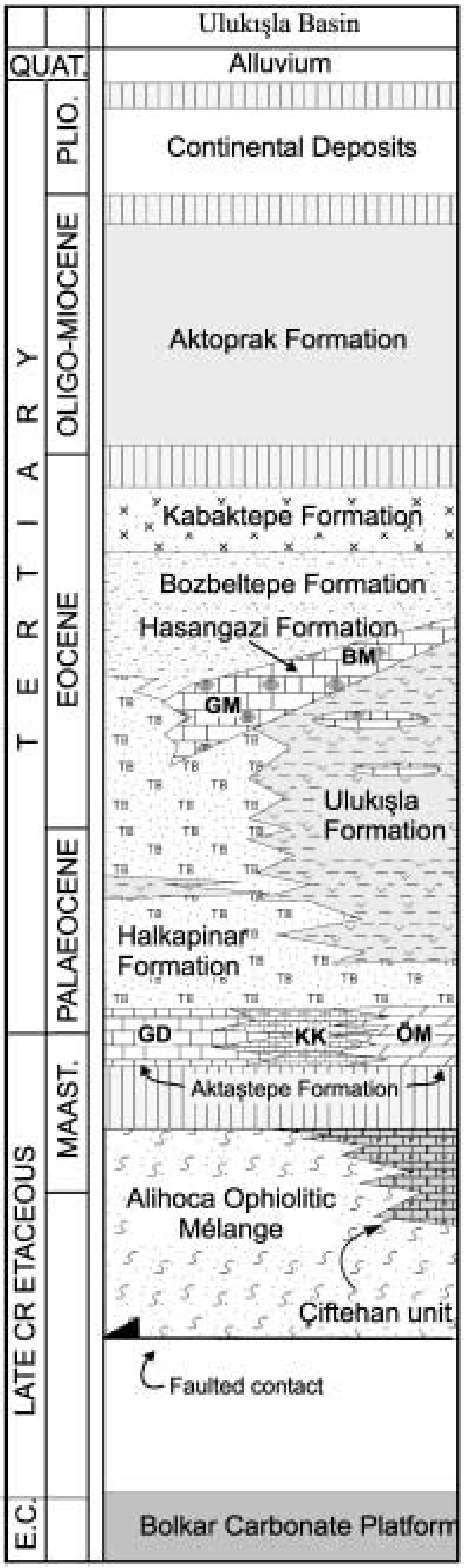
Geochemical studies suggest that the ophiolite was generated by spreading above a subduction zone (Parlak *et al*. 2000) in a Neotethyan oceanic strand located north of the Bolkar Carbonate Platform (Lytwyn & Casey 1995).

Maastrichtian–Lower Paleocene transgressive carbonate units (Aktaştepe Formation)

Ulukışla Basin sedimentation began with several shallow-water limestone units of Maastrichtian to Paleocene age. Although faulted in places, the contact is generally an unconformity above the Late Cretaceous ophiolitic rocks and mélange (Fig. 4).

Marginal terrigenous clastic rocks and carbonates (Kalkankaya Member)

The Kalkankaya Member (Fig. 4), of Late Maastrichtian age (N. İnan & K. Taslı, pers. comm.), was named after a type locality (Kalkankaya Tepe), between the villages of Alihoca and Maden (Demirtaşlı *et al*. 1975; Fig. 3, locality 3). This unit comprises a lower sandstone-dominated part (80 m) and an upper limestone-dominated part (200 m). The base of the lower part is marked by a conglomeratic unit (5 m) with rounded clasts of recrystallized



limestone grading up into alternating sandstones and pebblestones. The pebblestone beds are often channelized (*c.* 10 m wide, 2–3 m deep) and are incised into coarse-grained, thick-bedded sandstones that exhibit planar cross-bedding. The succession fines upward overall, although there are a number of conglomerate and breccia interbeds.

The upper limestone-dominated part of the succession begins with a coarse biomicrite, including rip-up clasts of microbial limestone and local concentrations of bivalve shells. Petrographic study reveals clusters of small grapestone particles (1–3 mm), including benthic foraminifera, spines and calcite crystals encrusted by micrite. Well-bedded micritic limestones with less bioclastic material appear higher up.

The Kalkankaya Member is interpreted as a shallow-marine transgressive unit (Fig. 5). The lower part is interpreted as channel fills associated with the subaqueous toes of small deltas flowing off the Bolkar Carbonate Platform into a marginal marine setting. Comparable deposits occur in the lower part of the Pliocene Nicosia Formation of the Mesaoria Basin, Cyprus (McCallum & Robertson 1995). The overlying unit is interpreted as shallow-marine limestones above storm wave base, including rip-up clasts of microbial limestone and bivalve shell fragments concentrated by currents (i.e. shell lags). The grapestone is indicative of a subtidal environment affected by storm activity (Illing 1954; Reading 1996). The predominance of well-bedded micritic limestones upwards suggests overall deepening.

Off-margin sandstones and limestones (Ömerli Member)

The Ömerli Member (Fig. 4), dated as Late Maastrichtian to Early Palaeoene (N. İnan & K. Taslı, pers. comm.), is exposed in the east of the basin above a north-dipping thrust fault (Fig. 3, locality 4) and begins with a fine-grained breccia containing angular clasts (3–4 cm in diameter) of mainly carbonate rocks derived from the Bolkar Carbonate Platform and the Alihoca Ophiolitic Mélange. Overlying detrital limestones exhibit planar, trough and hummocky cross-bedding, with convolute and planar laminations in many beds. The succession then passes upwards into uniform light blue microfossil-rich massive mudstones. Partial to locally complete turbiditic Bouma sequences (Bouma 1962) are present. However, the presence of hummocky cross-bedding is suggestive of redeposition above the storm wave base (Leeder 1999) in a slope setting, rather than as turbidity current in a deeper-water setting below the storm wave base (i.e. >250 m; Brasier 1980; Sartorio & Venturini 1988).

The Ömerli Member, characterized by a marine slope setting, probably represents a relatively distal facies equivalent of the Kalkankaya Member (Fig. 5).

Near-coastal carbonates (Güneydağı Member)

The Güneydağı Member (Fig. 4), dated as Late Maastrichtian to Early Paleocene (N. İnan & K. Taslı, pers. comm.), unconform-

Fig. 4. Revised stratigraphy of the Ulukışla Basin. (See text for discussion.) Initial palaeontological dating was carried out mainly by Oktay (1973, 1982) and Demirtaşlı *et al*. (1984), with recent revision by N. İnan and K. Taslı of Mersin University. (See Fig. 3 for key.) Previous stratigraphies have been given by Oktay (1973, 1982), Demirtaşlı *et al*. (1984) and (Çevikbaş & Özuntalı 1991). Subdivisions of the Aktaştepe Formation: GD, Güneydağı Member; KK, Kalkankaya Member; ÖM, Ömerli Member; subdivisions of the Hasangazi Formation: GM, Gümüş Member; BM, Basmakçı Member.

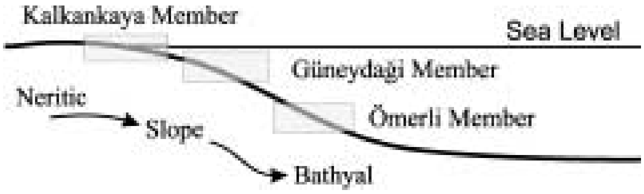


Fig. 5. Inferred Maastrichtian basin morphology.

ably overlies the Alihoca Ophiolitic Mélange in the west of the Ulukışla Basin (Fig. 3, locality 5). It is dominated by medium- to thick-bedded limestones that, although recrystallized, still exhibit abundant microbial carbonate fragments. The base of the limestone is marked by conglomerate with rounded limestone clasts (15–20 cm in diameter) derived from the Bolkar Carbonate Platform.

The Güneydağı Member is interpreted as a shallow-marine near-coastal deposit, as suggested by the abundance of microbial carbonate; the limestone conglomerate at the base of the section is suggestive of a slope setting. Overall deepening from the base to the mid-part of the succession is suggested by an upward decrease in grain size.

Taken together the three members, of Maastrichtian to Early Paleocene age (i.e. Kalkankaya Member, Ömerli Member and Güneydağı Member), record transitional palaeoenvironments following marine transgression in Maastrichtian time. The Kalkankaya Member records a proximal coastal setting, the Güneydağı Member a near-coastal setting, and the Ömerli Member a more off-margin slope setting (Fig. 5).

Middle Paleocene–Lower Eocene turbidites (Halkapınar Formation)

The Halkapınar Formation, dated as Late Maastrichtian to Late Paleocene (N. İnan & K. Taslı, pers. comm.), is a thick (*c.* 1.8 km), variable unit of mudstones, sandstones, carbonates and volcanogenic rocks developed throughout the Ulukışla Basin (Fig. 3).

A complete, representative succession of the Halkapınar Formation is exposed on a hillside south of Gümüş village (Fig. 3, locality 6, and Fig. 6a). The base of the formation is marked by marls and mudstones, passing up into graded sandstones. A notable light blue, calcareous mudstone horizon is rich in well-preserved gastropods. Above, there is a channelized conglomeratic limestone, up to 10 m thick and 30 m wide. The top of this unit is dominantly calcarenite and less clast rich. Up-section, a horizon of megabreccia (up to 30 m thick) is laterally continuous for >3 km, although the thickness and clast size (30 m to <3 m) vary considerably. Clasts are largely shelf-derived grey micritic limestone with rare chert and pink pelagic limestone derived from the Late Cretaceous ophiolite mélange. The megabreccia is interpreted as a rock-fall or debris-flow deposit (Drzewiecki & Simo 2002); such deposits are commonly related to oversteeping (Robertson & Bamakhalif 2001; Drzewiecki & Simo 2002) and are considered to reflect faulting of the margin of the Ulukışla Basin. Redeposited limestone (i.e. bioclastic packstones and wackestones) is an important constituent of the lower part of the successsion.

Most of the Halkapınar Formation consists of medium- to thick-bedded, fine- to medium-grained sandstones. Most sandstones show normal grading, parallel lamination and other sedimentary structures, mainly indicative of the B–E divisions of turbidity current deposits (Bouma 1962). In addition, there are interbeds of pebblestones with intermittent sandstone horizons in which pebblestones fine in and out of sandstone horizons. The coarse pebblestones (*c.* 10 cm clasts) are dominated by limestone clasts. These beds are commonly channelized and always exhibit erosive bases incised into underlying fine-grained turbiditic sandstones.

In the upper part of the Halkapınar Formation, medium-bedded turbiditic sandstones exhibit flute casts, gutter casts, cross-bedding and *Planolites* burrows (Simpson 1975), 2–3 cm in diameter; also *Zoophycos* (Frey & Howard 1970). Where turbidites are well developed, flute casts and cross-bedding indicate palaeocurrents towards the WNW (Fig. 7a).

Near the top of the succession blue–grey silty calcareous mudstones are interbedded with subordinate graded sandstones. The top of the formation is dominated by volcanogenic debris-flow deposits, comprising clasts of feldspar-phyric basalt (up to 30 cm in size) set in a hyaloclastic matrix. The volcanogenic units are locally variable in lateral extent and thickness (up to 10 m thick by 80–100 m wide). In addition, intercalated terrigenous turbidites exhibit considerable lateral variation in bed thickness.

A comparable section through the Halkapınar Formation is seen in a transect from a gorge SE of Çakillar village to Biter Dağ (Fig. 3, locality 8, and Fig. 6b). The succession here fines upwards, interrupted by volcanic material in the mid-part. Coarse conglomerates mark the base then grade upwards into pebblestones interbedded with bioturbated sandstones. Transitionally overlying medium- to fine-grained sandstones are interbedded with mudstones and incised by channelized conglomerate; sedimentary structures include grading, planar cross-bedding and planar lamination. In the mid-part there is a change to interbeds of mudstone, thick-bedded sandstone and conglomerate, together with pillow lava and nummulite-bearing calcarenites. The upper part of the succession is dominated by well-graded gravelstone to mudstone. Planar lamination is common, although trough cross-bedding and convolute lamination are also present. Palaeocurrents are mainly directed northwestwards (Fig. 7b), giving a generally west to northwestwards direction for the formation as a whole (Fig. 7c).

In general, the Halkapınar Formation is interpreted as a 'faulted slope apron' setting (Fig. 8).

Lower to Middle Eocene volcanogenic lithofacies (Ulukışla Formation)

The Ulukışla Formation (Fig. 4) is a thick unit (>2 km) of submarine basaltic to intermediate-composition volcanogenic rocks, typically structureless conglomerates and breccias, subordinate pillow lavas and massive lava flows. The dominant extrusive lithology is dark feldspar-phyric lava (phenocrysts 1–2 mm in diameter) set in a black, green or reddish, altered, finely crystalline groundmass. Other variants include feldspar–pyroxene-phyric lava and, rarely, aphyric lava. Short, well-bedded, fine-grained volcaniclastic intervals are present locally. Considerable, local to regional, variations in lithology and thickness of the volcanogenic units are seen; few are regionally continuous.

Most successions are dominated by thick-bedded, matrix-supported volcanogenic conglomerates and breccias. The breccias contain angular to sub-rounded clasts of basic to intermediate-composition volcanic rock in a silt- to sand-sized volcanogenic matrix, often hyaloclastic. These beds are interpreted as debris-flow deposits indicative of the upper flanks of a

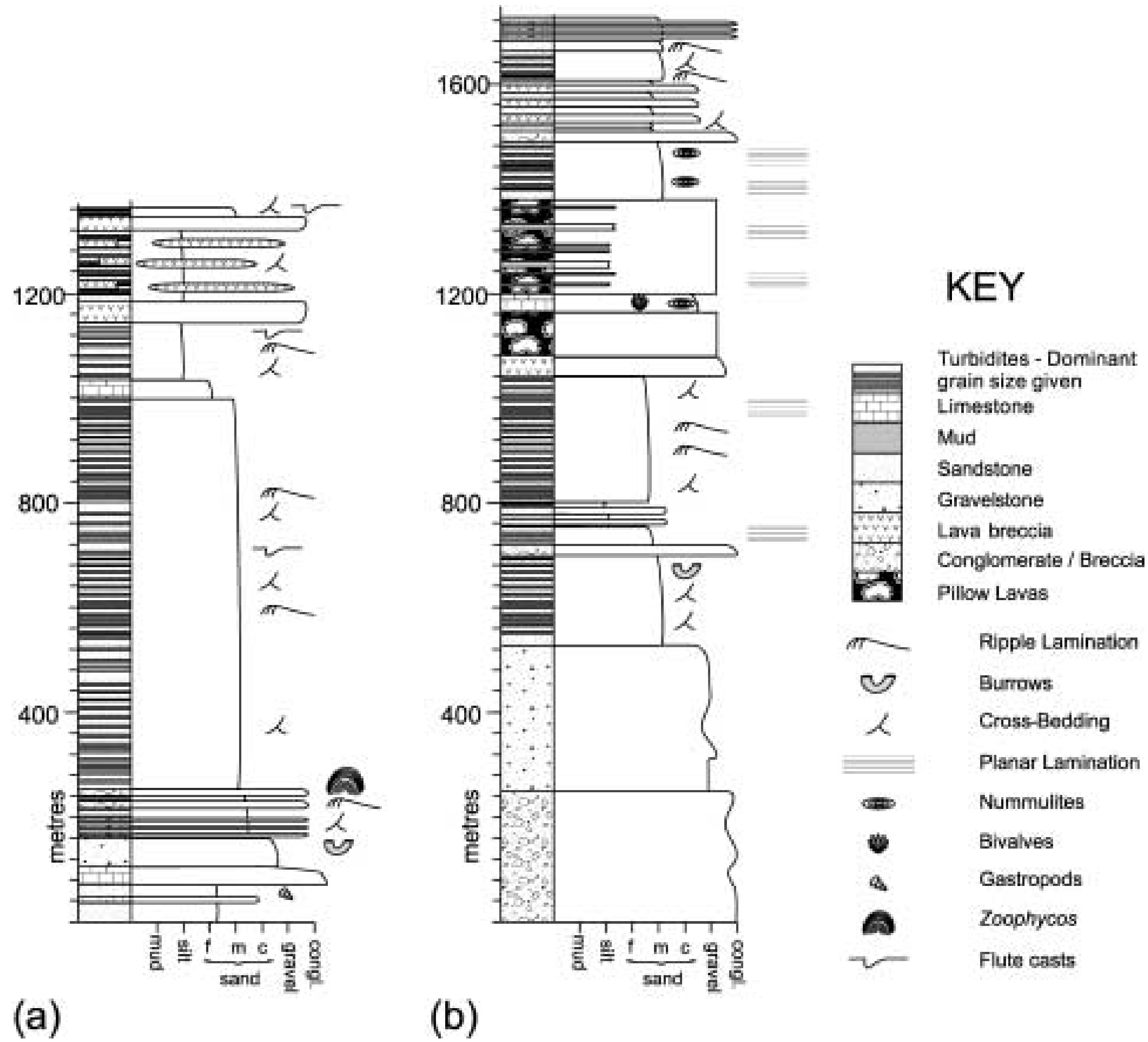


Fig. 6. Representative sedimentary logs of the Halkapınar Formation: (**a**) Gümüş area (Fig. 3, localities 6 and 7); (**b**) Çakillar area (Fig. 3, locality 8).

volcanic edifice (McPhie 1995; Wright 1996). In the lower part of the succession (e.g. near Elmalı, Fig. 3, locality 9) volcanic debris flows are interbedded with fine-grained volcanogenic sediments on a bed-by-bed scale. By contrast, in the higher part of the succession (e.g. Basmakçı area, Fig. 3) volcanic debris-flow deposits (>200 m thick) are structureless and lack such interbedding. The debris flows are, however, locally interbedded with intervals of well-bedded volcaniclastic conglomerates, sandstones and siltstones (e.g. Çifteköy area; Fig. 3, locality 10, and Fig. 9a). Individual beds are well graded and show sedimentary structures indicative of deposition by turbidity currents. There are also intercalations, varying from thin to medium bedded, of off-white, fissile non-calcareous mudstones, interpreted as siliceous airfall tuffs. Discrete sections of pillow-lava and pillow breccia are present thoughout the Ulukışla Formation and are considered to represent the lower flanks of the volcanic edifice (Corcoran 2000).

Volumetrically minor carbonates within the Ulukışla Formation vary from pillow interstices to discrete interbeds of pink or white micritic limestone (wackestone), dated as Late Paleocene to Early Eocene (N. İnan & K. Taslı, pers. comm.). Near the base of the succession (e.g. near Çiftehan, Fig. 3, locality 11) there are local intercalations of thin-bedded pink micritic limestone, up to 20 cm thick. A wackestone in the mid-part of the main exposure of the Ulukışla Formation, near Ünlükaya village (Fig. 3, locality 12), contains large angular lava clasts (<50 cm in diameter) set in a micritic matrix. This bed fines upwards into a calcirudite with very well-rounded volcanic clasts, <5 mm in diameter. Large rip-up clasts (up to 30 cm in size) of carbonate mud are also present, along with fragmented bivalves, nummulites, feldspar grains and fresh biotite. In addition, intercalations of white limestone high in the succession contain bryozoan fragments, echinoderm debris and benthic foraminifera.

During and after volcanism the central part of the volcanic

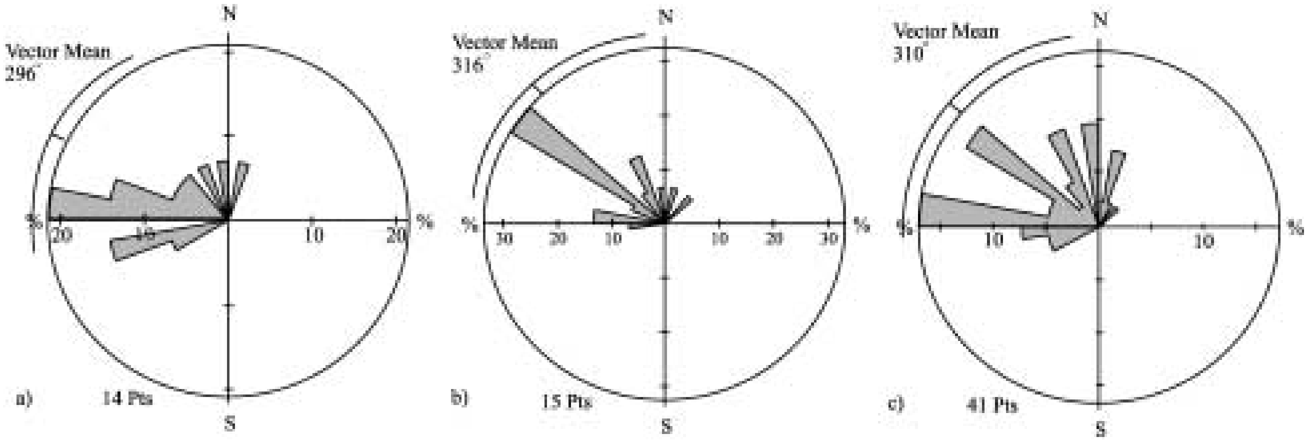


Fig. 7. Palaeocurrent data (azimuthal) from turbiditic sandstones of the Halkapınar Formation, measured from cross-bedding and flute casts: (**a**) Gümüş area (Fig. 3, localities 6 and 7); (**b**) Çakillar area (Fig. 3, localities 5–8); (**c**) Ulukışla Basin as a whole.

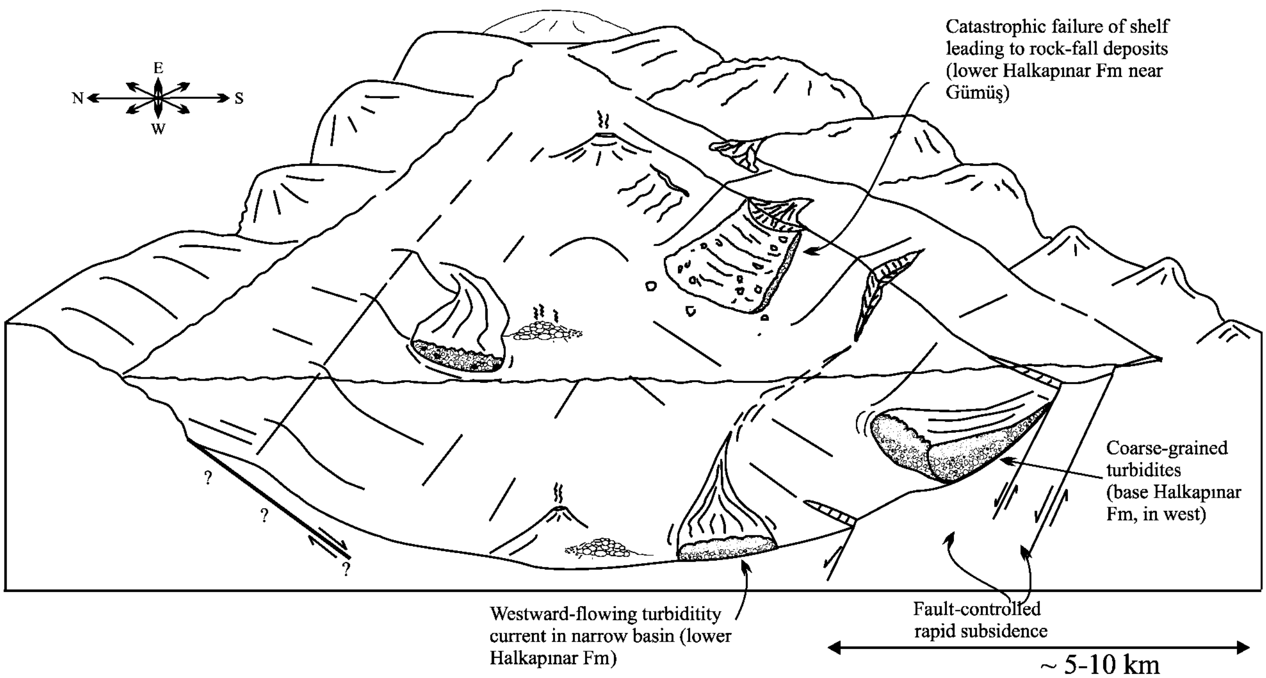


Fig. 8. Basin reconstruction for the Halkapınar Formation.

pile was cut by monzonite and syenite intrusions up to 1 km wide (Boztuğ *et al*. 2001), with well-developed contact metamorphic aureoles.

Geochemistry

To shed light on the volcanic affinity and possible tectonic setting of eruption, little-altered basalts, mostly pillow lavas and massive flows, were analysed for major and trace elements by XRF, using the method specified by Fitton *et al*. (1998). Most of the samples were collected from near the base (near Ardıçlı, Fig. 3, locality 18) and the top of the Ulukışla Formation (near Çaykavak, Fig. 3, locality 19). Figure 10a and b shows relative enrichment of the large ion lithophile elements (LILE) Sr to Ba, and light rare earth element (LREE) enrichment of La to Nd, with a lesser enrichment of Nb. Nb is, however, more enriched than the high field strength elements (HFSE) Ti and Y, whose values are similar to those of mid-ocean ridge basalt (MORB). Enrichment of Nb (most incompatible HFSE) relative to Ti and Y (less incompatible HFSE) is indicative of a mantle source enriched by intra-plate processes involving low degrees of partial melting (Pearce *et al*. 1990; Keskin *et al*. 1998). However, the lesser enrichment of Nb compared with the LILE and LREE is indicative of the existence of a subduction component (Pearce *et al*. 1990). We therefore infer that the Ulukışla Formation was erupted in an intra-plate tectonic setting (e.g. a rift) influenced by subduction (either coeval or inherited). The trace-element geochemistry of the plutonic intrusive rocks is similar to that of the volcanic rocks (Fig. 10c), and therefore suggests that the intrusive and extrusive rocks are cogenetic.

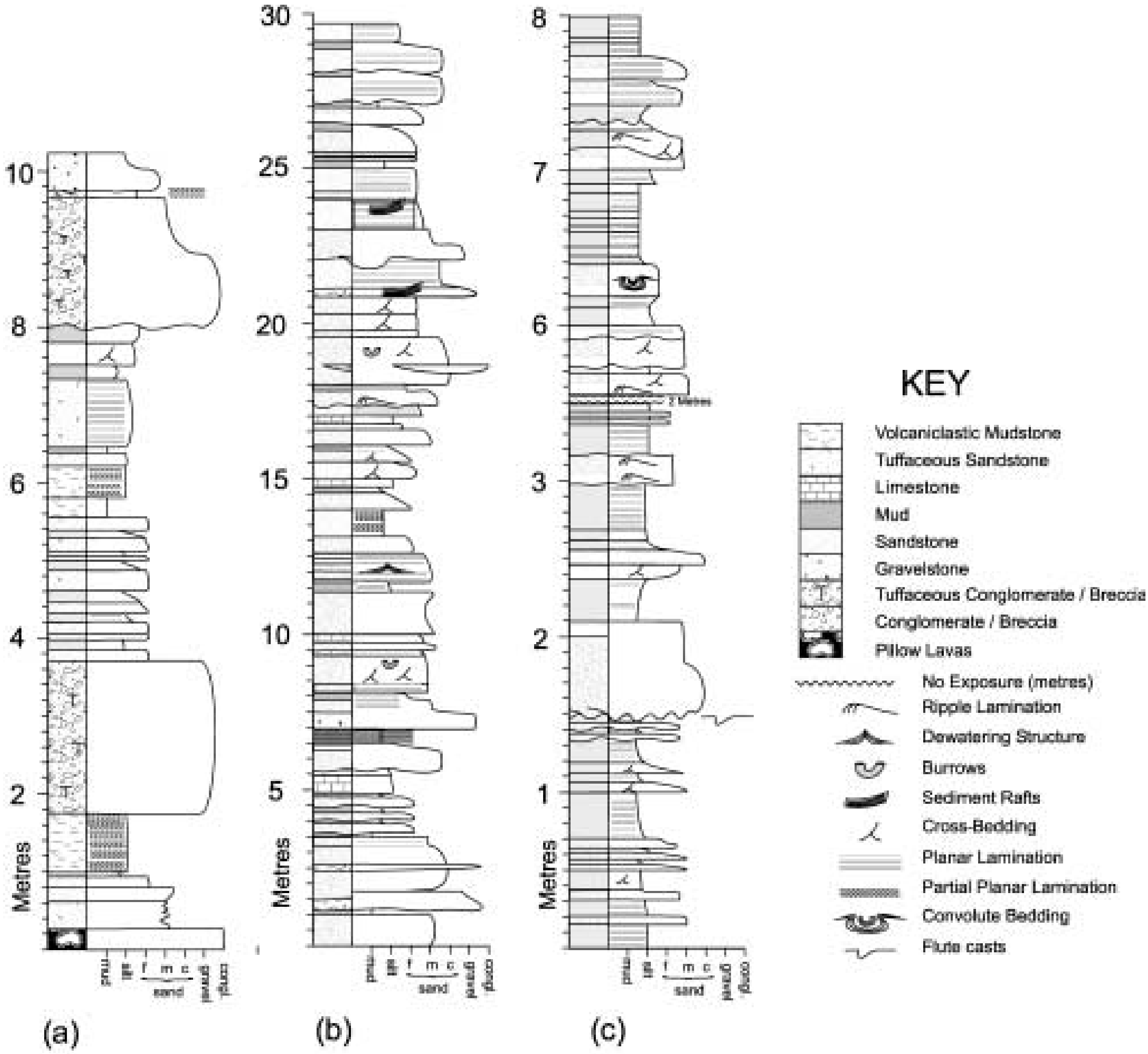


Fig. 9. Sedimentary logs: (**a**) volcaniclastic sediments of the Ulukışla Formation, Çifteköy area (Fig. 3, locality 10); (**b**) Upper Lutetian turbidites of the lower part of the Bozbeltepe Formation from the Yeniköy area (Fig. 3, locality 15); (**c**) Upper Eocene turbidites of the upper part of the Bozbeltepe Formation from Güney village area (Fig. 3, locality 16). The greater proportion of mud and silt to sand in the upper part of the Bozbeltepe Formation should be noted.

The Ulukışla Formation, thus, records extensive volcanism, with associated plutonic intrusion, during Late Paleocene–Early Eocene time. The magmatism generated high-temperature hydrothermal activity, as shown by the occurrence of disseminated sulphide and epidote (Çevikbaş & Özuntalı 1991). The predominance of sÇub-aqueous volcanogenic debris-flow deposits from the base to the top of the succession (*c.* 2.5 km thick) implies continuing subsidence to maintain the slope conditions necessary for debris-flow deposits and volcanogenic turbidites to form. The absence of ophiolite-derived sediment, in contrast to the underlying or laterally equivalent Halkapınar Formation, suggests that the volcanic pile was by then isolated from continental input. A variable sea-floor topography existed, with copious pillow lava extrusion in the west (Çaykavak), rather than volcanogenic debris-flow deposits as seen elsewhere. The interbedded pink limestones near the base of the succession contain planktonic foraminifera suggesting relatively deep-water (bathyal) deposition, similar to the underlying or laterally equivalent Halkapınar Formation. The occurrence higher in the succession of nummulites and neritic fossils (e.g. bivalves, echinoids) indicates overall shallowing upwards. The presence of locally interbedded airfall tuffs implies subaerial (or at least shallow-water) eruption.

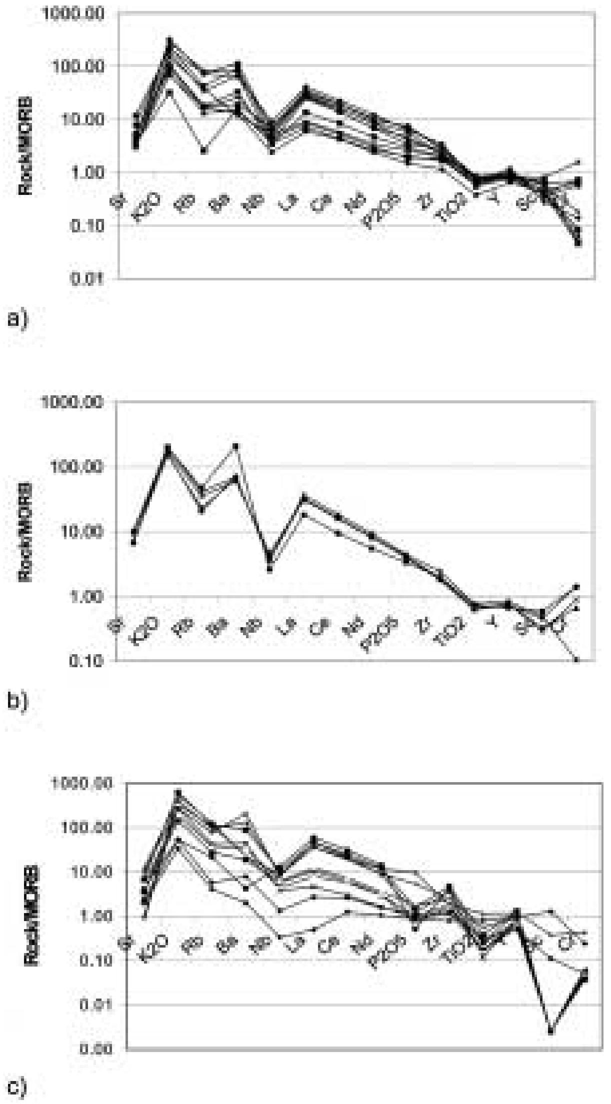


Fig. 10. MORB-normalized 'spider diagrams' for aphyric lava flows and pillow lavas from the Ulukışla Formation: (**a**) Ardiçli road section (Fig. 3, locality 18); (**b**) Çaykavak Pass (Fig. 3, locality 19); (**c**) from plutonic rocks intruded into the volcanic rocks. Data are shown in Table 1.

Middle Eocene post-volcanic limestones (Hasangazi Formation)

The Ulukışla Formation is depositionally overlain by two contrasting lithofacies of relatively shallow-water carbonate, of Late Paleocene to Early Eocene age (N. İnan & K. Taslı, pers. comm.).

Shallow-water carbonate build-ups (Basmakçı Member)

The first shallow-water carbonate facies (Basmakçı Limestone of Demirtaşlı *et al*. 1984) crops out south of Basmakçı village as several folded and thrust outcrops (Fig. 3, locality 13). However, locally the limestone is seen to overlie the Ulukışla Formation depositionally (e.g. in a quarry near Basmakçı village). A locally exposed basal conglomerate consists of sub-rounded, poorly sorted lava clasts set in a recrystallized carbonate matrix. The basal limestone there is coarse bioclastic calcarenite with clasts of coral, locally bound by algal mats, generally set in a sparry cement. Microbial limestone is present as oncoliths; also coral is locally observed. North of Basmakçı, thin (5–6 m), lenticular limestone is mostly fragmental with gastropods, coral fragments and large foraminifera, including nummulites and miliolinids. The carbonates are interpreted as an accumulation of biogenic carbonate after volcanism ended. Small carbonate build-ups developed on local highs, flanked by slope talus and carbonates rich in benthic foraminifera. Mainly slope and inter-reef facies are preserved. Detailed facies correlations are precluded by strong deformation and limited outcrop.

Open-shelf deposits (Gümüş Member)

The second limestone facies crops out in the southern part of the Ulukışla Basin in the core of a large syncline (Fig. 3). A conformable contact with the underlying Ulukışla Formation (e.g. near Hasangazi village, Fig. 4, locality 14) is indicated by clear depositional intercalation of volcanogenic material and carbonate. The typical limestone is a light grey, micrite-supported, bioclastic calcarenite containing nummulites (1 mm–3 cm), bivalves (3 mm–2 cm in diameter), microbial carbonate and echinoid fragments. Glauconite is a common constituent. The density of nummulites varies from sparse in mud-rich layers to dense in bioclastic layers and dominant in certain packstone intervals. The bioclastic layers are very poorly sorted and contain volcanogenic grains, up to 5 mm in size. In some beds the nummulites are oriented parallel to bedding. The most southerly outcrops (near Gümüş village) lack preserved nummulites and comprise relatively coarse calcarenite.

The second facies records open-marine, shallow-water (shelf-derived) deposition after volcanism ended, with the presence of glauconite suggestive of a transgressive setting (Jeans *et al*. 2000). The variable concentration of nummulites reflects differential current reworking that concentrated these relatively dense bioclasts and winnowed away fine-grained sediment (Aigner 1985). The southerly facies, lacking nummulites, reflects a more clastic-dominated setting near an inferred Bolkar landmass to the south.

Mid–Late Eocene: second phase turbidites (Bozbeltepe Formation)

The Early Eocene shallow-water carbonates of the Ulukışla Formation are depositionally overlain by a second, mainly turbiditic unit (Bozbeltepe Formation; post-volcanic 'flysch' deposition of Demirtaşlı *et al*. 1984), assigned to an Early–Late Eocene age (N. İnan & K. Taslı, pers. comm.). Where the shallow-water limestones are not present the unit directly overlies the Ulukışla Formation.

The Bozbeltepe Formation is marked by conglomerates and gravelstones at the base of the sequence (e.g. at Yeniköy village; Fig. 3, locality 15, and Fig. 9b). Well-rounded clasts include lava, volcaniclastic sandstones and bioclastic limestones with nummulites, bryozoa, microbial carbonate and bivalve material. Most beds are clearly graded, commonly with cross-bedded intervals towards the top. Near the base these sediments interfinger with the volcanogenic Ulukışla Formation. Slightly higher exposures are very thin-bedded, graded sandstones and siltstones with a high proportion of interbedded mudstone (SE of Güney village; Fig. 3, locality 16, and Fig. 9c). Flute casts, convolute lamination and other structures indicate deposition by turbidity currents. Palaeocurrent analysis reveals flow mainly away from

Table 1. *MORB-normalized data used in construction of spidergrams in Fig. 10*

Sample	50AR	51AR	52AR	53AR	54AR	55AR	59AR	60AR	63AR	80AR	81AR	82AR	107AR	MC52	MC53	MC54		
Sr	4.29	4.70	3.14	5.04	2.97	5.56	5.49	8.44	3.30	3.46	3.54	3.10	9.01	8.01	11.95	4.44		
K_2O	327.00	91.78	345.20	183.22	150.75	107.09	285.13	190.77	110.93	69.87	32.47	71.61	193.83	233.52	214.70	289.20		
Rb	72.81	18.09	71.74	34.23	38.93	18.27	61.89	52.81	16.26	17.28	2.59	13.60	44.55	43.41	74.86	83.40		
Ba	65.20	33.57	79.50	68.32	11.93	19.46	84.45	21.66	26.06	13.00	15.96	14.87	144.64	91.96	113.39	79.45		
Nb	9.17	3.54	8.80	3.81	5.60	7.11	5.49	6.78	5.22	2.41	3.50	4.37	5.35	4.86	6.52	7.10		
La	41.36	13.63	30.41	24.89	30.48	26.58	39.39	20.18	9.28	6.00	7.16	8.48	31.63	28.02	32.32	36.31		
Ce	22.95	8.44	15.18	13.81	16.50	14.11	20.36	11.21	5.83	4.32	4.37	5.32	16.28	16.00	19.17	19.52		
Nd	12.40	4.69	6.47	7.00	8.47	6.90	9.41	4.91	3.46	2.40	2.62	3.03	7.39	8.84	9.97	10.62		
P_2O_5	6.40	2.96	2.52	4.66	4.14	3.44	4.27	0.96	2.60	1.49	2.05	1.89	3.73	5.89	7.49	7.18		
Zr	3.46	2.07	3.81	2.21	2.35	2.87	2.88	1.37	1.97	1.15	1.69	1.83	2.79	2.18	2.92	2.79		
TiO_2	0.66	0.75	0.45	0.82	0.72	0.59	0.60	0.29	0.62	0.38	0.64	0.56	0.67	0.80	0.85	0.81		
Y	1.18	0.79	0.93	0.84	0.87	0.95	0.88	1.02	1.04	0.65	0.76	0.78	0.95	0.85	0.98	0.98		
Sc	0.28	0.39		0.47	0.81	0.37			0.43	0.42	0.73	0.38		0.63	0.45	0.56		
Cr	0.14	0.74	0.04	0.64	1.59	0.57	0.04	0.10	0.04	0.09	0.05	0.06	0.03	0.70	0.08	0.18		
Sample	1 CA	2 CA	3 CA	4 CA	MC1	MC2	MC3	39 IN	71 IN	72 IN	73 IN	78 IN	115 IN	116 IN	117 IN	127 IN	146 IN	156 IN
Sr	6.73	7.80	7.13	9.28	6.79	10.07	1.99	4.39	3.47	5.41	3.38	0.93	2.31	10.12	12.15	0.80	1.99	6.78
K_2O	163.93	156.73	154.54	183.59	207.51	200.92	383.61	53.40	145.14	192.52	269.50	547.62	52.93	276.42	406.24	331.79	33.53	627.40
Rb	20.81	28.53	23.40	35.09	41.51	47.72	60.31	5.69	27.95	38.04	43.33	126.18	21.58	77.85	95.43	108.28	4.06	116.46
Ba	71.74	45.66	63.56	61.00	69.43	210.48	75.75	7.82	18.71	33.86	44.69	22.78	4.24	205.97	130.23	24.52	2.01	89.54
Nb	3.65	2.39	4.57	4.46	3.68	2.61	3.58	1.33	5.65	6.21	3.80	8.48	13.10	8.30	8.81	4.34	0.34	10.63
La	32.18	21.85	31.32	30.97	36.53	18.33	20.32	2.71	10.52	12.27	4.63	45.59	59.83	36.17	36.24	7.97	0.50	45.90
Ce	16.54	12.08	16.32	15.97	18.52	9.52	11.48	2.58	6.19	7.28	2.94	22.03	31.09	20.75	18.93	5.69	1.23	25.40
Nd	8.17	6.57	8.34	8.09	9.41	5.51	6.54	1.57	3.01	3.50	1.63	8.63	14.23	11.63	9.01	2.96	1.04	12.20
P_2O_5	4.00	3.27	4.17	4.17	4.45	3.40	2.70	1.11	1.00	1.84	0.91	1.18	0.52	9.57	5.38	0.69	0.85	1.53
Zr	1.83	1.53	1.94	1.92	2.46	1.98	1.80	1.22	1.82	2.27	1.15	3.67	4.25	1.83	2.44	1.65	0.79	4.67
TiO_2	0.63	0.07	0.65	0.66	0.77	0.74	0.78	0.32	0.17	0.24	0.18	0.12	0.25	1.14	0.57	0.16	0.79	0.38
Y	0.73	0.73	0.78	0.78	0.84	0.67	0.81	0.50	0.78	0.81	0.54	0.79	1.26	1.19	0.92	0.67	0.97	1.32
Sc	0.46	0.35	0.33	0.29	0.50	0.59	0.48	0.11						0.38			1.27	
Cr	1.42	1.32	0.67	0.89	0.11	1.41	0.96	0.05	0.05	0.05	0.05	0.04	0.04	0.43	0.06	0.04	0.24	0.04

MORB data from Pearce (1982) and Saunders & Tarney (1984): Sr 121 ppm, K_2O 0.2 wt%, Rb 2 ppm, Ba 20 ppm, Nb 4.6 ppm, Ce 11 ppm, P_2O_5 0.12 wt%, Zr 90 ppm, TiO_2 1.4 wt%, Ni 90 ppm, Y 33 ppm, Sc 40.6 ppm, Cr 351 ppm.

the centre of the volcanic outcrop (Ulukışla Formation) (Fig. 11). In addition, in the south the Bozbeltepe Formation contains ophiolitic clasts and grains not seen in the units to the north.

The Bozbeltepe Formation records topographic differentiation around the Ulukışla Basin. The core area, dominated by the Ulukışla Formation volcanic rocks, emerged subaerially and was eroded, shedding large volumes of volcaniclastic sediment outwards into open-marine peripheral basins (estimated as 100–200 m deep) dominated by sand and silt turbidites. High-density volcaniclastic turbidity current deposits locally dominate the base of the succession and indicate steep slope gradients. The turbidites locally interfinger with mainly volcanic breccias near the base of the formation. Supply of terrigenous (non-volcanogenic) sediment in the south, mainly detrital carbonate and ophiolitic material, reflects input from the Bolkar Carbonate Platform and the Alihoca Ophiolitic Mélange. A possible reason for basin differentiation is that the region as a whole subsided following the end of volcanism (i.e. by thermal subsidence). However, the core area had undergone high-level intrusions by plutons and remained buoyant, above sea level, creating a source of detrital volcaniclastic sediment.

Upper Eocene evaporites (Kabaktepe Formation)

Turbiditic siltstones and mudstones of the Bozbeltepe Formation pass transitionally upwards into mudstone interbedded with redeposited selenitic gypsum of inferred Mid- to Late Eocene age, marking the base of the Kabaktepe Formation (Fig. 4). The evaporites are thickest (200–300 m) and best exposed on the northern limb of the east–west-oriented syncline mentioned above (Kabak Tepe Hill; Fig. 3, locality 17). Much of the evaporite is recrystallized massive, structureless sugary selenitic gypsum; however, gypsum preserves primary depositional features on the northern limb of the syncline. Higher levels of the succession include steeply dipping, mud-supported selenite conglomerate–breccia, interpreted as gypsum debris flows. There are also thick-bedded graded selenite interbeds, interpreted as high-density gypsum turbidites. This interval includes disorganized horizons, up to 3 m thick, interpreted as slumps. The upper parts of the succession comprise more regularly bedded (less steeply inclined) intervals of well-formed selenitic gypsum (gypsum swallowtails) and massive granular evaporite.

The evaporites record hypersaline deposition within the Ulukışla Basin during its latest stages of regression. Selenite mostly formed in shallow marginal settings (e.g. Schreiber *et al*. 1976) and was redeposited into a small remnant basin with a depocentre near the present synclinal axis. The shallowing was probably tectonically driven and may have triggered the input of selenitic debris flows, turbidites and slumps from marginal areas, as for example seen in the Messinian evaporites of Cyprus (Robertson *et al*. 1995). Eventually, basin shallowing allowed *in situ* precipitation of selenite in shallow lagoons and ponds near the basin depocentre. Further emergence then terminated deposition in the Ulukışla Basin and created a regional low-angle unconformity. Subsequently, Oligocene sediments record non-marine fluvial conditions and exhumation (Jaffey & Robertson 2001).

Subsidence history

As a further guide to the geohistory of the Latest Cretaceous to Late Eocene Ulukışla Basin a subsidence curve was constructed using data collected around the south–central part of the basin (Fig. 12). The curve shows an initial pulse of relatively gentle subsidence at the start of the basin history followed by a period of very rapid subsidence, which coincides with deposition of turbidites and volcanic rocks in the basin. Following volcanism, the curve displays a period of gentle subsidence during the regressive phase of the basin. Compared with the standard subsidence curves of well-known basin types (Emery & Myers 1996; Turner 1996) the trend shows marked deviations from a

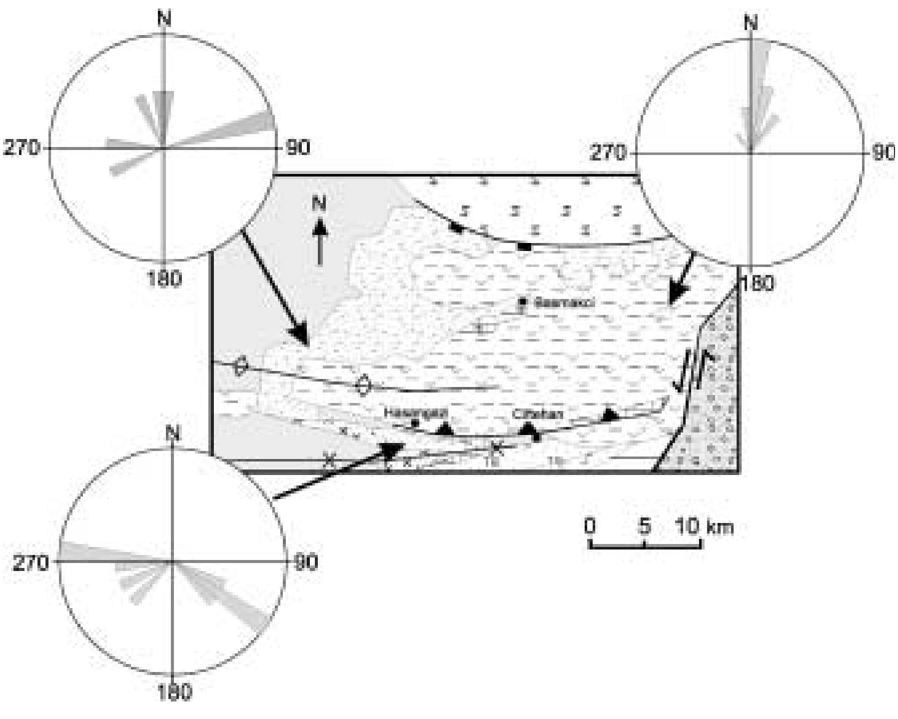


Fig. 11. Palaeocurrents from the Middle to Upper Eocene Bozbeltepe Formation. Readings from cross-bedding, except for locality 16, measured from cross-bedding and flute casts.

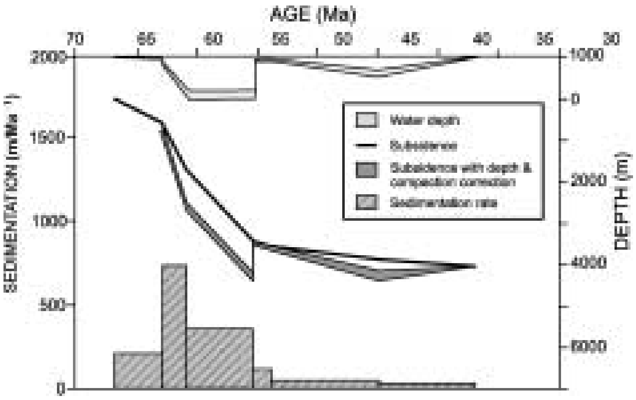


Fig. 12. Subsidence curve plotted from composite log of the Ulukışla Basin succession, data from the Gümüş area. The lack of any major unconformities in the Maastrichtian to Oligocene succession allows construction of the curve. Both the water-depth estimates and the dating of the rocks are possible sources of error. Total decompaction subsidence was calculated using the approach of Steckler & Watts (1978). Each layer of rock was removed to reconstruct the total subsidence history of the basin and remaining layers were decompacted using standard porosity–depth relationships determined by Sclater & Christie (1980). The 'zero' level for depth equates to local sea level at the beginning of the basin history.

foreland basin setting. The curve is more similar to an extensional basin, or a strike-slip pull-apart setting, exhibiting an early rapid subsidence pulse and a latter more gradual phase that could tentatively be attributed to thermal subsidence.

Structure

The structural history of the Ulukışla Basin is summarized here to allow integration into the regional tectonic picture. Four main stages in basin development and deformation are recognized, which all relate to the progressive demise of the Northern Neotethys.

The first stage relates to inferred subduction and emplacement of the Alihoca Ophiolitic Mélange in latest Cretaceous time. During tectonic emplacement onto the submerged Bolkar Carbonate Platform the Alihoca ophiolite was dismembered into thrust sheets, large blocks and clasts, and reworked as debris-flow deposits. The Late Cretaceous pelagic cover of the ophiolitic extrusive rocks locally survived intact (i.e. Çiftehan Unit, Fig. 4).

The emplacement of the Alihoca Ophiolitic Mélange reflects the regional ophiolitic emplacement event throughout the Taurides in latest Cretaceous time (Tekeli 1984; Dilek *et al*. 1999; Okay *et al*. 2001; Andrew & Robertson 2002). However, the mélange shows little evidence of intense folding, high-temperature metamorphism, uplift or deep erosion and is, thus, unlikely to reflect regional continental collision.

The second stage of deformation involved inferred extension to create the Ulukışla Basin. Widespread rapid subsidence, fault-controlled sedimentation and within-plate basalt type volcanism are indicative of extension during Late Paleocene–Early Eocene time; the subsidence curves do not support other mechanisms of basin formation (e.g. compression-related foreland basin).

The northern margin of the Ulukışla Basin is marked by a relatively low-angle (*c*. 30°) south-dipping extensional fault contact with the metamorphic Niğde Massif (Toprak and Göncüoğlu 1993; Whitney & Dilek 1997; Dilek & Whitney 2000;Göncüoğlu *et al*. 2000; Dirik *et al*. 2001; Gautier *et al*. 2001). The timing of activity on this fault is disputed. Dilek & Whitney (2000) reported extensive movement in the Miocene to accommodate exhumation of the Niğde Massif. However, Gautier *et al*. (2001) inferred fault movement in the Late Eocene based on the presence of nummulite-rich limestones overlying the faulted metamorphic contact (Yetiş 1984; Toprak and Göncüoğlu 1993). We observed that sandstones of the Bozbeltepe Formation in the north of the Ulukışla Basin contain quartz, implying that the Niğde Massif was exposed to erosion at least by the Early Eocene, no other sources of quartz being available at this time. Eocene nummulitic limestones overlie the Niğde Massif near Çamardi (25 km NE of Basmakçı) and are correlated with the nummulitic facies overlying the Ulukışla Formation volcanic rocks (Gümüş Member) in the south of the basin. This, in turn, implies that the Niğde Massif was exhumed before Ulukışla volcanism ended and was transgressed. It is thus possible that the fault along the north of the basin accommodated much of the extension that resulted in Maastrichtian to Paleocene subsidence of the Ulukışla Basin. This would not preclude additional extension and further exhumation of the Niğde Massif in Miocene times, as inferred by Whitney & Dilek (1997).

The third stage of deformation of the Ulukışla Basin relates to regional convergence that gave rise to the main structural features preserved in the basin. The basin experienced east–west folding and reverse faulting (Figs. 3 and 13a and b) soon after deposition ended. A strike-swing from east–west to ENE–WSW is displayed moving from west to east along the basin. Folding is generally upright, symmetrical and relatively open. A gentle plunge to the WSW (10–20°) is commonly seen, although this reverses to ENE in the north of the basin (Fig. 14a–d and h). Large map-scale fold orientation is mimicked at an outcrop scale (Fig. 14e–g), suggesting that folding of the basin resulted from one phase of deformation, with compression acting in a NNW–SSE orientation. Thrust faults are common (Fig. 14j) with fault-plane dips clustering around 40° to the south and north, although varying from 16° to 84°; no preferred vergence is apparent. Large-scale reverse faulting is also seen at the base of the Ulukışla Basin in the south, where marbles of the Bolkar Carbonate Platform were thrust northwards over the Alihoca Ophiolitic Mélange (thrust plane 035/39°S); local folding is observed beneath the thrust plane. Intense north-vergent deformation of the adjacent Bolkar Carbonate Platform is also shown by large-scale folding in marbles (Jaffey 2001; Jaffey & Robertson 2001).

We interpret the pervasive Late Eocene deformation as the result of regional suture tightening. Associated uplift terminated marine deposition and resulted in a low-angle (<10°) unconformity at the top of the evaporites (Kabaktepe Formation) in the south. In the north of the area the Oligo-Miocene Aktoprak Formation (Demirtaşlı *et al*. 1984; Jaffey 2001) overlies the Ulukışla Formation above this regional unconformity. Initial compression probably also caused the shallowing of the basin before deposition of the evaporites. Syndepositional deformation may also have triggered debris flows, turbidites and large-scale slumping of the gypsum deposits. Compressional deformation may have inverted pre-existing normal faults related to subsidence of the Ulukışla Basin, thus accounting for reverse-fault dips up to 84°.

In the fourth stage, post-collisional extension is represented by normal faulting that localized the Oligo-Miocene Aktoprak sub-basin (Figs. 3 and 13b). As noted above, this could be coeval with further extensional unroofing of the Niğde metamorphic

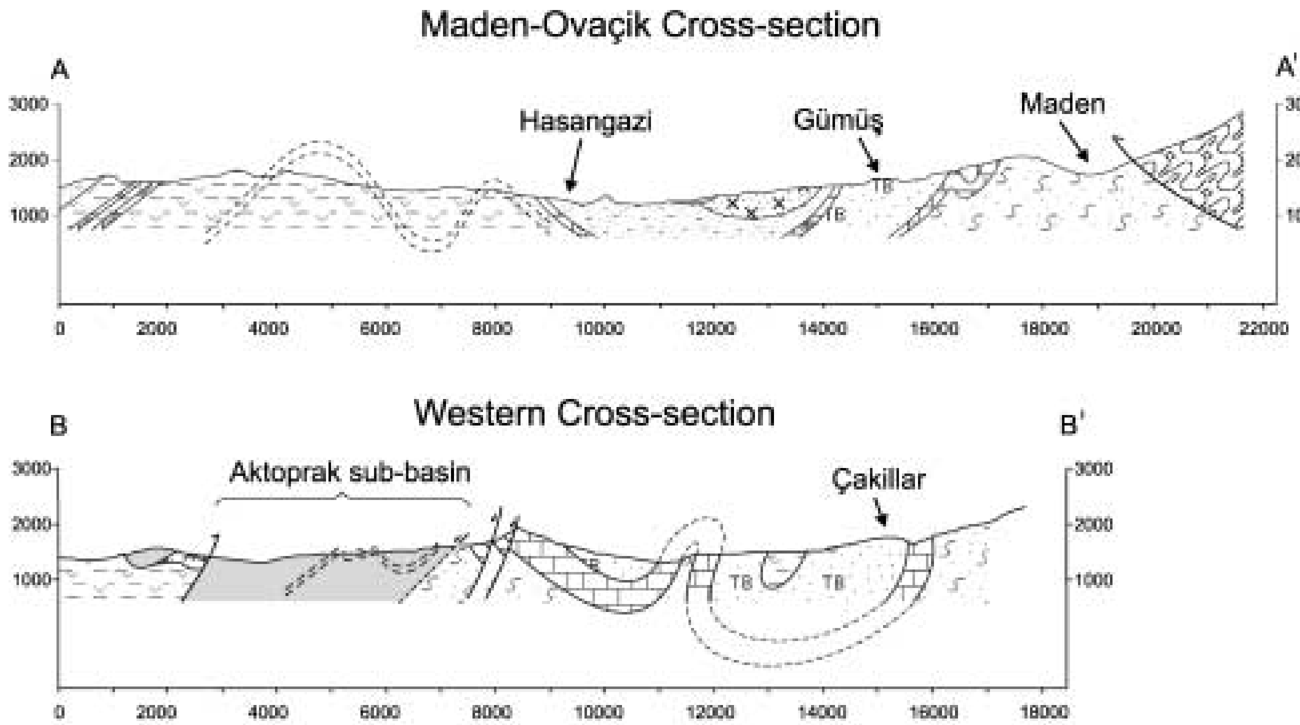


Fig. 13. Structural cross-sections of the Ulukışla Basin (see Fig. 4 for location of cross-sections and key to symbols).

massif to the north. Subsequently, the area was compressed in the Late Miocene, resulting in open folding of the Aktoprak Formation (Jaffey 2001).

Regional tectonic implications

In this study we attempted to establish the tectonic setting in which the Ulukışla Basin formed. One set of published alternatives assume that an Inner Tauride Ocean existed between the Bolkar Carbonate Platform and the Niğde–Kırşehir microcontinent to the north. In one such model this Inner Tauride Ocean was subducted northwards in Late Cretaceous–Early Tertiary time, giving rise to the Ulukışla Basin as a fore-arc basin constructed on the margin of Kırşehir microcontinent (Görür *et al*. 1984; Fig. 15a). The Ulukışla volcanic rocks could thereby relate to contemporaneous subduction. However, the geohistory of the Ulukışla Basin, as established during this study, questions this interpretation for several reasons. First, although internally deformed, the Ulukışla Basin exhibits an unconformable contact with the southerly Bolkar Carbonate Platform, rather than an Early Tertiary, subduction-related thrust, as in Fig. 15a. Second, the Ulukışla Formation volcanic rocks exhibit a within-plate type geochemistry. In another such model, Oktay (1973, 1982) suggested that subduction was instead southwards, with the Ulukışla Formation volcanic rocks forming a subduction-related arc in a remnant oceanic basin, considered as the Ulukışla Basin (Fig. 15b). This interpretation is also questionable in view of the lack of arc-type volcanism, the lack of a contemporaneous accretionary prism preserved in the basin lithostratigraphy and the suggestion of a coeval southward-slipping extensional fault contact with the Niğde Massif rather than a northward subduction-related overthrust, as implied in this model (Fig. 15b). Also, the basal unconformity, lithostratigraphy and subsidence curves all indicate extension at the start of the Ulukışla Basin history, rather than inheritance of a pre-existing ocean basin setting.

The second set of tectonic alternatives assume that only a single northerly Neotethys existed (Göncüoğlu 1986). In this scenario the Niğde–Kırşehir Massif existed as a promontory of the Mesozoic Bolkar (Tauride) continent to the south, rather than as a microcontinent, and no suture existed beneath the Ulukışla Basin (Özgül 1976; Göncüoğlu 1986; Fig. 15c). The Late Cretaceous–Early Tertiary basins bordering the southern margin of the Niğde–Kırşehir Massif could then be considered to have formed as a result of post-collisional extension following closure of the Northern Neothethys along the Ankara–Erzincan suture zone (Çemen *et al*. 1999 for the Tuzgölü Basin; Dirik *et al*. 1999 for the Sivas Basin). The Ulukışla Basin might relate to the break-off and rotation of such a continental promontory. If no suture existed, all the units (i.e. ophiolitic mélange) were thrust from the Ankara–Erzincan suture bounding the southern margin of Eurasia, implying >350 km thrust, which may be unlikely. Also, the most obvious explanation of the subduction component (e.g. negative Nb anomaly) in the Ulukışla basalts is that it was inherited from some earlier subduction in the area, possibly related to formation of the Alihoca Ophiolitic Mélange, itself interpreted as a dismembered supra-subduction zone type ophiolite (Parlak *et al*. 2000). A widespread view (Lytwyn & Casey 1995; Polat & Casey 1995; Dilek *et al*. 1999) is that the Pozantı–Karsantı Ophiolite (and with it the Alihoca Ophiolite) was rooted in a local (Inner Tauride) oceanic basin adjacent to the northern margin of the Tauride continent.

The evidence presented here shows that an ocean basin is unlikely to have existed adjacent to the Ulukışla Basin after latest Cretaceous time. How typical is this of the other central Anatolian Early Tertiary basins? The large Tuzgölü Basin, considered as a possible extension of the Ulukışla Basin (Görür

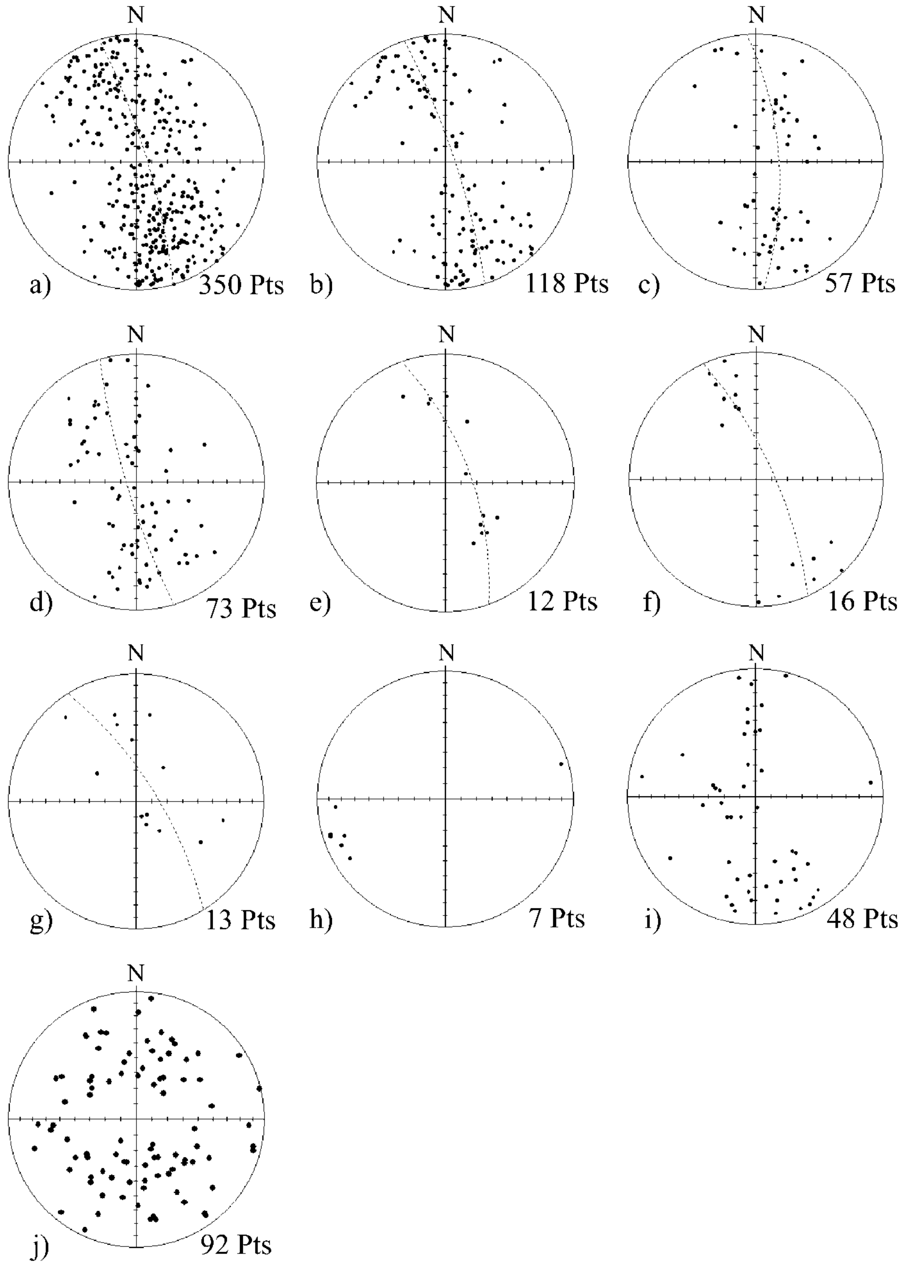


Fig. 14. Equal-area stereographic plots of poles to bedding, poles to profile planes and poles to fault planes for the Ulukışla Basin: (**a**) poles to bedding for the Ulukışla Basin; (**b**) poles to bedding from area around Maden and Hasangazi; (**c**) poles to bedding from Çakillar area; (**d**) poles to bedding from Basmakçı to locality 16; (**e**) fold at Bozbeltepe; (**f**) fold at Kalkankaya; (**g**) fold from east of locality 19 (Fig. 3); (**h**) poles to profile planes from above stereographic projections; (**i**) poles to fold axial planes; (**j**) poles to planes of reverse faults.

et al. 1984, 1998), was extensively studied because of its hydrocarbon potential and the occurrence of salt domes (e.g. Yüksel 1970; Norman & Rad 1971; Sirel & Gündüz 1976; Uygun 1981*a, b*; Fig. 16a). Only the eastern margin is well exposed, albeit deformed and cut by the recently reactivated Tuzgölü Fault (Çemen *et al*. 1999). To the NE of the study area the Şarkışla Basin exhibits a similar stratigraphy to the Ulukışla Basin (Gökten 1986; Gökten & Floyd 1987; Fig. 16b). Further NE again, the NE–SW-trending Sivas Basin unconformably overlies emplaced ophiolitic rocks and includes large detached blocks of ophiolitic rocks within overlying basinal clastic sediments (Poisson *et al*. 1996; Görür *et al*. 1998; Dirik *et al*. 1999; Fig. 16c). Currently, the Oligo-Miocene part of the Sivas Basin is attributed to development of a (post-collisional) foreland basin generated by overthrusting of Pontide units (Temiz *et al*. 1993; Gürsoy *et al*. 1997; Görür *et al*. 1998). No record of Early Tertiary ophiolites or accreted oceanic sediments of this age has been confirmed from mélanges adjacent to any of these Early Tertiary basins. We therefore infer that the Northern Neotethys, at least in the vicinity of the Ulukışla Basin and the basins mentioned above, was closed by latest Cretaceous time. Assuming such a syncollisional setting, the formation of the Ulukışla Basin might relate to relative displacement of the Tauride continent to the south and the Niğde–Kırşehir microcontinent to the north.

Irrespective of whether a single or two Northern Neotethys ocean basins existed, we believe that the Ulukışla Basin (and perhaps also other basins along the southern margin of the

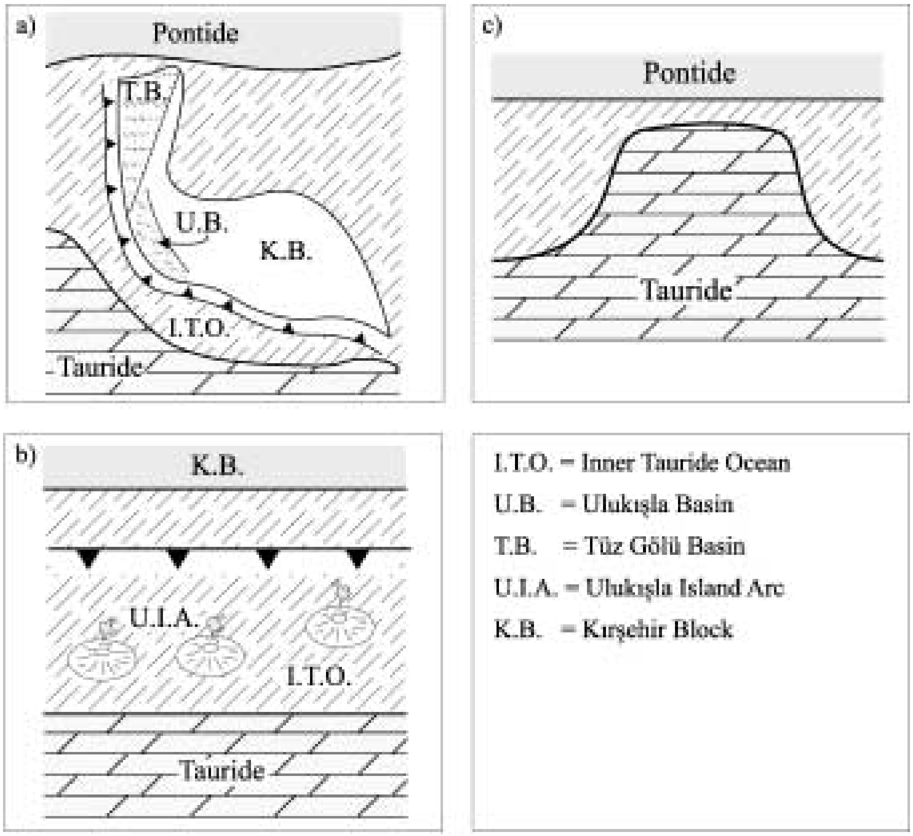


Fig. 15. Existing models for the Ulukışla Basin and the Inner Tauride Ocean: (**a**) fore-arc basin associated with Tuzgölü Basin (Görür *et al*. 1984); (**b**) island-arc model (Oktay 1982); (**c**) Tauride promontory without Inner Tauride Ocean (Göncüoğlu 1986;Göncüoğlu *et al*. 1992).

Niğde–Kırşehir Massif) formed as a result of readjustment of the suture zone after initial collision. Adjustments occurred during the period between initial 'soft collision' in the latest Cretaceous (emplacing the Tauride ophiolites) and final suture tightening, as evidenced by extensive compressional deformation of the Ulukışla Basin, around the end of the Eocene. Possible scenarios are as follows.

(1) The basin formed as a strike-slip pull-apart basin as the Bolkar Carbonate Platform and Niğde–Kırşehir Massif slid past each as microplates following initial latest Cretaceous collision.

(2) The setting was similar to that in (1), but with rotation of one or several microplates following initial collision. Sanver & Ponat (1980) reported anti-clockwise rotation of the Niğde–Kırşehir Massif in the Eocene, although this needs to be tested using modern palaeomagnetic techniques.

(3) If subduction of an Inner Tauride Ocean left a remnant oceanic slab suspended beneath the suture zone, slab-pull forces might have caused subsidence leading to formation of the Ulukışla Basin.

(4) Following closure of a single or multiple Northern Neotethyan basins subduction could have 'stepped-back', initiating closure of the southern Neotethys (Aktaş & Robertson 1990; Yılmaz *et al*. 1993; Robertson 2000). Subduction could then have exerted extensional forces further north in the vicinity of the Ulukışla Basin. Such subduction might also have introduced the observed subduction component in the Ulukışla Formation volcanic rocks.

In summary, the above alternatives are not mutually exclusive and all relate to the progressive closure and suturing of Neotethys in this region. Similar considerations may apply to other syncollision basins in other orogenic belts.

Conclusions

The latest Cretaceous (Maastrichtian)–Late Eocene Ulukışla Basin documents extensional (or transtensional) basin formation following initial closure of what was probably a local strand of the Northern Neotethys ocean (Inner Tauride Ocean). Early Tertiary subsidence resulted in a transition from neritic, mainly carbonate, deposition to basinal turbiditic accumulation with extensive within-plate type, basic–intermediate volcanism. A superimposed, geochemically recognized, subduction signature (i.e. relative Nb depletion) was probably inherited from preceding Late Cretaceous subduction in the area. Inferred basin extension correlates with one stage in the unroofing of the northerly basin margin (Niğde Massif), above a low-angle detachment fault. Following cessation of volcanism, further turbidites accumulated, followed by shallowing, localized evaporite deposition, regional folding and faulting in Late Eocene time. The geohistory of the Ulukışla Basin, and similarities with other central Anatolian basins, suggests that Neotethys in this area was initially closed, by 'soft collision', in latest Cretaceous time. The Ulukışla Basin then formed during a 30 Ma period when the suture zone was reactivated to accommodate continuing regional convergence, culminating in Late Eocene 'hard collision'.

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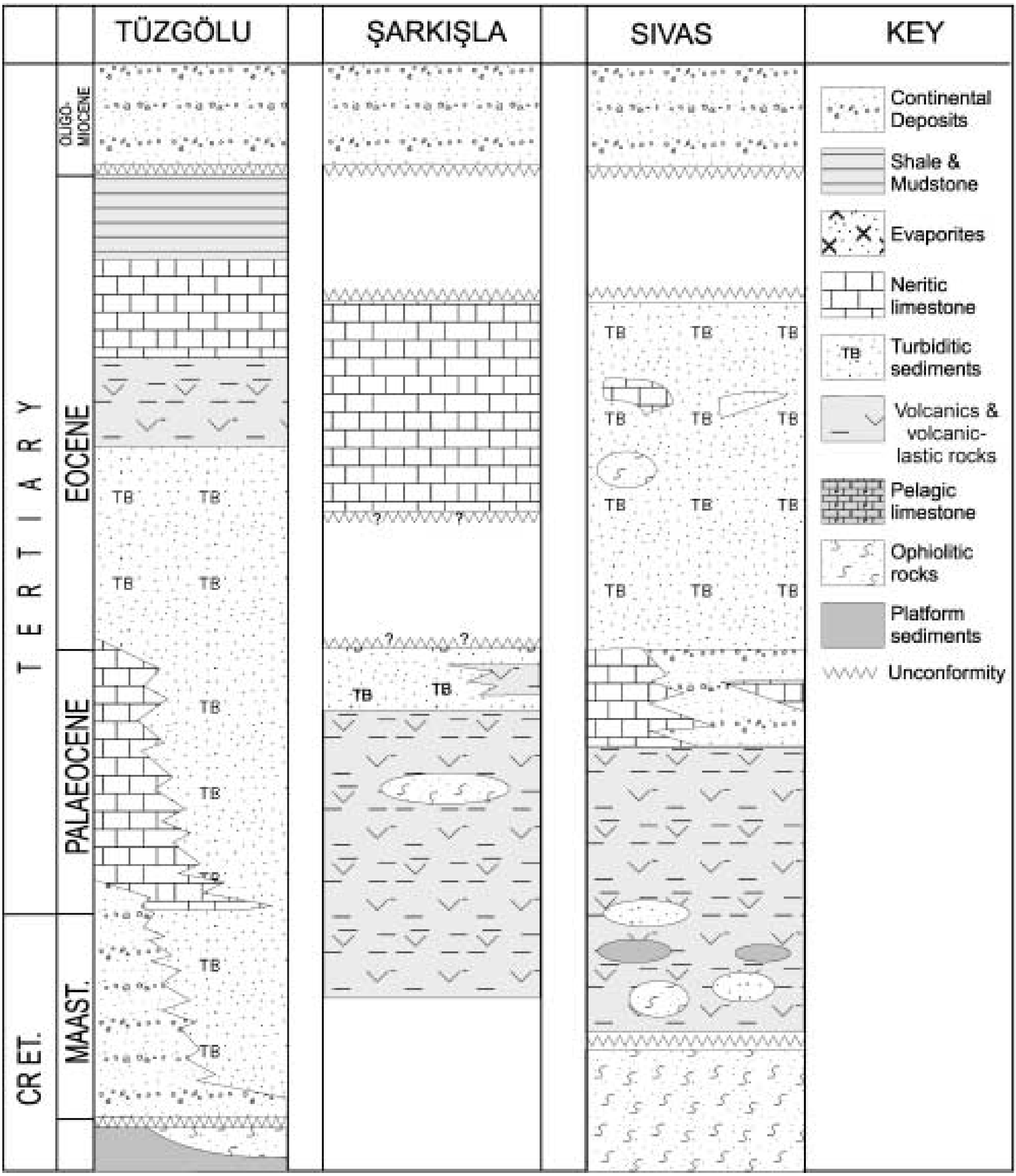

Fig. 16. Chronostratigraphic diagram comparing geohistory of the Tuzgölü, Şarkışla and Sivas basins. Tuzgölü Basin stratigraphy from Çemen *et al*. (1999), Şarkışla Basin stratigraphy from Gökten (1986) and Sivas basin from Dirik *et al*. (1999).

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