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Formation and emplacement ages of the SSZ-type Neotethyan ophiolites in Central Anatolia, Turkey: palaeotectonic implications

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Isolated outcrops of ophiolitic rocks, termed the Central Anatolian Ophiolites, are found as allochthonous bodies in the Central Anatolian Crystalline Complex, that represent the metamorphosed passive northern edge of the Tauride–Anatolide Platform, central Turkey. In terms of pseudostratigraphic relationships of the magmatic units and their chemical designation, the Central Anatolian Ophiolites exhibit a supra-subduction zone (fore-arc) setting within the Vardar–İzmir–Ankara–Erzincan segment of the Neotethys. The epi-ophiolitic sedimentary cover of the Central Anatolian Ophiolites is generally characterized by epiclastic volcanogenic deep-sea sediments and debris flows intercalated with pelagic units. The richest and most significant planktonic foraminiferal association recorded from the lowest pelagic members infer a formation age of early–middle Turonian to early Santonian. K/Ar ages of post-collisional granitoids (81–65 Ma) intruding the basement rocks as well as the Central Anatolian Ophiolites suggest a post-early Santonian to pre-middle Campanian emplacement age. The marked high volume of epiclastic volcanogenic sediments intercalated with the pelagics of the Central Anatolian Ophiolite is suggestive of rifting in a marginal sea adjacent to a volcanic arc. Penecontemporaneous tectonism is reflected in repetitions in the stratigraphy and in debris flows, which result from major slides and mass-gravity reworking of pre-existing units and of arc-derived volcanics and sediments.

Correlating the rock units and formation/obduction ages of the Central Anatolian Ophiolites with further supra-subduction zone type ophiolites in the eastern (Turkey) and western (Greece) parts of the Vardar–İzmir–Ankara–Erzincan segment of Neotethys we conclude that the intraoceanic subduction in the east is definitely younger and the closure history of this segment is more complex than previously suggested.

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1. INTRODUCTION

Sedimentary sequences associated with ophiolites can provide important evidence regarding the age and tectonic setting of the ophiolites and thus help in the interpretation of the geodynamic evolution of oceanic basins. The type of sedimentary sequence overlying the ophiolite assemblages may also allow us to distinguish marginal-sea ophiolites from those generated at mid-ocean ridges. It is commonly believed that all well-developed marginal-basin ophiolites should be conformably overlain by deep-sea sediments that contain a volcaniclastic component and tuffs derived from the adjacent volcanic arc (Karig 1982), whereas mid-ocean ridge ophiolites would typically be overlain only by deep-sea pelagic sediments (Moores 1982). However, this is not the case for the supra-subduction zone (SSZ) type southern branch of the Neotethyan ocean generated within the marginal environment. These ophiolites (e.g. Pindos, Troodos Massif, Antalya, Mersin, Hatay and Semail) are overlain by ferro-manganean sediments and chalks which contain no significant

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volcanogenic component (Pearce *et al*., 1984). However, the ideal SSZ ophiolite would be overlain by pelagic and/or volcanogenic (non-terrigenous) deep-sea sediments.

Unlike the southern Neotethyan ophiolites in the eastern Mediterranean area, the Central Anatolian Ophiolites (CAO) remain incompletely studied and their geological, petrological and geochemical features are too fragmentary to draw any critical palaeogeographical correlations within the Neotethyan oceanic realm. In particular, do SSZ-type CAO that have been derived from the northern branch of Neotethys display a similar epi-ophiolitic sedimentary cover to that which is characteristic of SSZ-type ophiolites found in the southern branch of Neotethys? Although the CAO do not have an exact formation age, whether Jurassic or Cretaceous they clearly have important implications for a regional tectonic interpretation and correlation between the different branches of Neotethys.

The objective of this paper is to present the general characteristics and age of the epi-ophiolitic sedimentary cover of one of the most complete and best exposed CAO bodies–the Sarıkaraman Ophiolite–as a basis for the tectonic environmental evaluation and correlation of the southern and northern branches of the Neotethyan ocean in the Eastern Mediterranean region.

In order to determine the age of the pelagic sedimentary sequence and the relative age of the underlying unit (the Sarıkaraman Ophiolite), planktonic foraminiferal assemblages have been examined in several thin sections prepared from the pelagic carbonate samples collected from the lower and upper pelagic units. Although planktonic foraminiferal zonation is not established due to sporadic occurrences of foraminifers in the samples, the fundamental biostratigraphic schemes such as those of Caron (1985), Wonders (1980), Bolli (1966), Postuma (1971), Van Hinte (1976) and Sigal (1977) were used for verifying the age of the units. The planktonic foraminiferal assemblage age data presented here thus constitute the first reliable age for the Central Anatolian Ophiolites.

2. REGIONAL GEOLOGY

Turkish Neotethyan ophiolites occupy a critical segment in the Alpine–Himalayan orogenic system. They are included together with some further ophiolitic units in the Eastern Mediterranean Ophiolites (Figure 1) (e.g. Pindos, Vourinos, Othris in the west; Troodos, Antalya, Hatay, Baer–Bassit in the south; and Semail in the southeast) and represent the variably disrupted fragments of Neotethyan oceanic branches that were situated between the Eurasia and Gondwanaland (Juteau 1980; Şengör and Yılmaz 1981; Robertson and Dixon 1984). In Anatolia, these branches are constituted by northern and southern segments of the Neotethys; the remnants of both crop out along nearly east–west trending suture belts.

The remnants of the northern branch of Neotethys are made of: (a) a northern belt (the Intra-Pontide Suture) separating the Eurasian derived İstanbul Terrane from the Cimmerian Sakarya Terrane; and (b) a southern belt representing allochthonous units derived from the Vardar–İzmir–Ankara–Erzincan (VIAE) ocean; which separated the Gondwanan Anatolide–Tauride Terrane from the Cimmerian continent. The southern branch of Neotethys, on the other hand, separated the Arabian Platform, that is the main body of Gondwanaland, from the Tauride–Anatolide Platform and is variably called the Peri-Arabic Belt (Ricou 1971) or Southern Neotethyan Ocean (Şengör and Yılmaz 1981) (Figure 1).

These ophiolitic belts exhibit different formation ages; for example, the Intra-Pontide belt consists mainly of different types of Jurassic–Cretaceous ophiolites found as dismembered fragments in mélanges (Göncüoğlu *et al*., 1987), whereas in the VIAE suture and southern belts, ophiolites represent good examples of both unfragmented oceanic crust, as huge thrust-sheets, as well as dismembered allochthonous blocks within mélanges of middle to Late Cretaceous age. The ages of the ophiolitic rocks from the southern belt have undergone considerable investigations, whereas those of the VIAE suture have been studied only in the Greek area, but are relatively less known in Turkey (for references see Robertson and Dixon 1984).

Despite some slight differences in their petrology, internal stratigraphy and age, most of these ophiolites

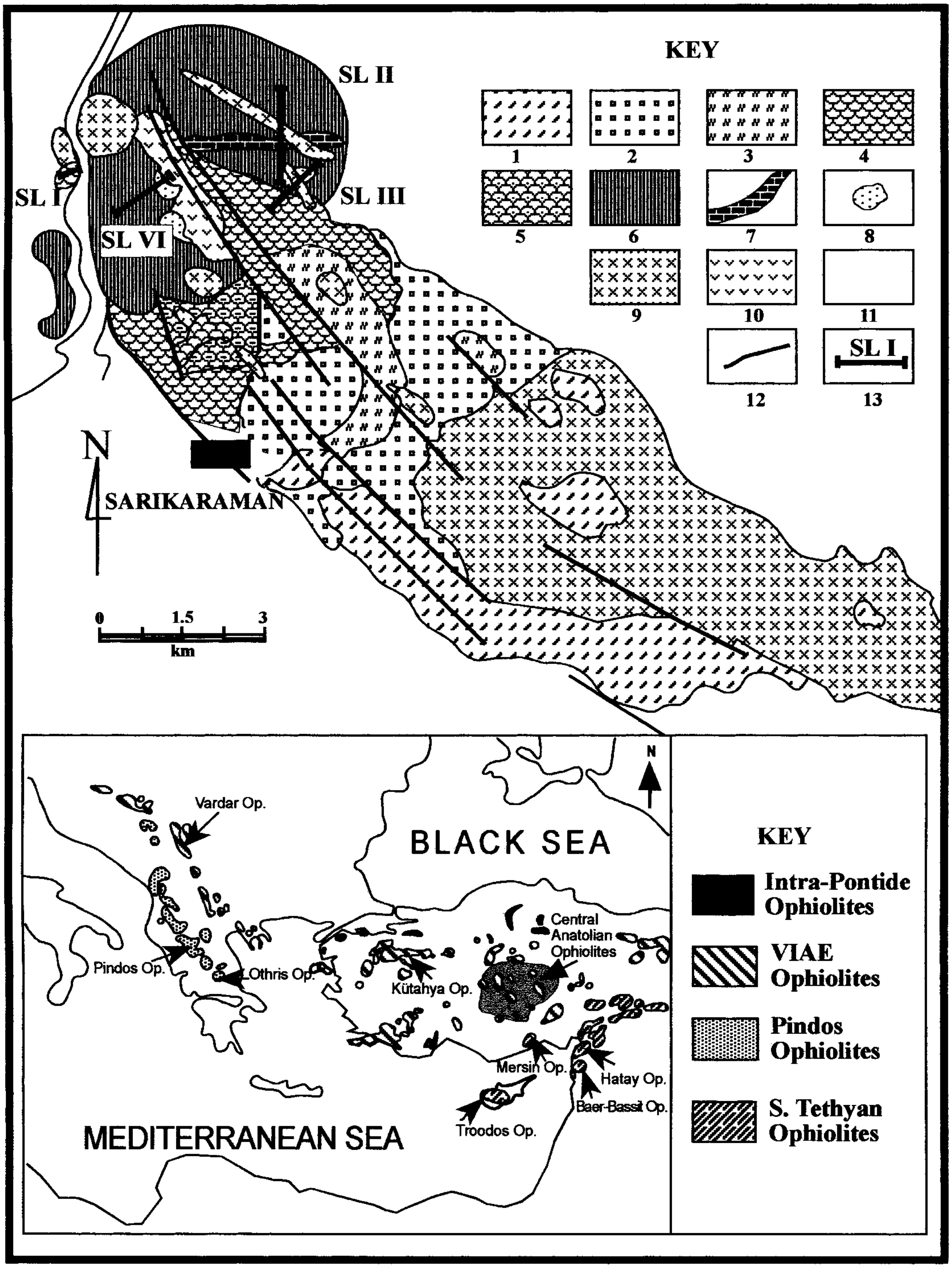


Figure 1. Simplified geological map of the Sarıkaraman Ophiolite, Central Anatolia. Key: 1, isotropic gabbro; 2, plagiogranites; 3, dolerite dyke complex; 4, pillow lava; 5, pillow breccia; 6, epiophiolitic cover; 7, pelagic limestones; 8, felsic volcanic olistoliths; 9, quartz-monzonites; 10, Late Maastrichtian–Early Palaeocene volcaniclastic cover; 11, Neogene cover; 12, faults; 13, location of the measured sections. The inset map shows the distribution and zonation of the Neotethyan Ophiolites in the Eastern Mediterranean and the location of the Central Anatolian Ophiolites.

are characterized by an epi-ophiolitic sedimentary cover dominated mainly by pelagic and/or volcanogenic (non-terrigenous) deep-sea sediments (Pearce *et al*., 1984; Robertson 1994).

3. CENTRAL ANATOLIAN OPHIOLITES

Despite the more or less well-defined Alpine ophiolitic belts, numerous, little known and isolated ophiolitic bodies are exposed in the Central Anatolia Crystalline Complex (CACC), just to the south of the VIAE suture (Figure 1 inset). The CACC is a triangular wedge of metamorphic basement (Göncüoğlu *et al*., 1991) which is considered to represent the northern passive margin of the Mesozoic Tauride–Anatolide Platform, facing the VIAE ocean (Özgül 1976). Those isolated outcrops of ophiolitic rocks within the CACC are termed Central Anatolian Ophiolites (Göncüoğlu *et al*., 1991), and are found as allochthonous bodies in the CACC. It is thought that they were initially derived from the VIAE Ocean and emplaced southwards onto the CACC.

The Central Anatolian Ophiolites are representative of a somewhat dismembered, but partially preserved, ophiolitic sequence consisting of metamorphic tectonites, cumulate and isotropic gabbros, plagiogranites, dolerites of sheeted dyke complexes, pillow lavas and an epi-ophiolitic sedimentary cover (Göncüoğlu *et al*., 1991; Yalınız 1996). In terms of the pseudostratigraphic relationships of the magmatic units and their chemical designation, the Central Anatolian Ophiolites exhibit a supra-subduction zone chemistry rather than one indicating generation at an ocean ridge (Göncüoğlu and Türeli 1993; Yalınız *et al*., 1996, 1999; Floyd *et al*., 1998, 2000; Yalınız and Göncüoğlu 1998). Likewise, the other Eastern Mediterranean ophiolites derived from the southern branch of the Neotethyan ocean (such as, Pindos, Troodos, Antalya, Hatay, Baer–Bassit, Oman) exhibit a similar tectonic setting (Yalınız *et al*., 1996).

4. SARIKARAMAN OPHIOLITE

The Sarıkaraman Ophiolite, one of the best exposed portions of the CAO, is representative of a partially dismembered ophiolite body retaining a recognizable magmatic pseudostratigraphy (Figure 1; Table 1). Voluminous ultramafics are not exposed in direct contact with the rest of the ophiolitic slab, the lowest section being composed of isotropic gabbros, which are faulted against a sheeted dyke complex which merges up-section into basalt lavas and breccias. The gabbros are cut by intrusive trondhjemitic plagiogranites which are genetically related to various high-level rhyolitic dykes and sills that traverse the upper volcanic section of the ophiolite (Floyd *et al*., 1998). All units are cut by a late set of isolated dolerite dykes. The ophiolite is overlain by an epi-ophiolitic sedimentary cover. Both the ophiolite and cover sediments have been intruded by bodies of Late Cretaceous quartz-monzonite. These granite rocks have their counterpart in other areas of the CACC and are not related to the ophiolites, being the post-collisional products of the melting of thickened crust (Göncüoğlu and Türeli 1994; Yalınız *et al*., 1996; Yalınız and Göncüoğlu 1998). Finally, the ophiolite and late granites are unconformably overlain by fluvial conglomerates and volcaniclastic sediments of uppermost Maastrichtian–lower Palaeocene age (Table 1; Göncüoğlu *et al*., 1991; Dirik and Göncüoğlu 1996).

4a. Epi-ophiolitic sedimentary cover

The para-autochthonous sedimentary cover of the Sarıkaraman Ophiolite crops out at the northwestern part of the Sarıkaraman area (Figure 1). It has two different primary relationships with the uppermost portion of the basaltic volcanics: (1) passage into red coloured mudstone intercalated with the uppermost portion of the pillow lavas (SL-1 in Figure 2); and (2) contact with pillow breccias situated above the pillow lavas (SL-3 in Figure 2). The sedimentary sequence is generally characterized by epiclastic volcanogenic

Table 1. Summary of field characteristics, petrography, ages, geochemistry and tectonic setting of Sarıkaraman Ophiolite

Field characteristics	Petrography	Age	Geochemistry	Tectonic setting
(7) Volcaniclastic sediments.		Upper Maastrichtian–Lower Palaeocene		
(6) **Terlemez Granitoid**. Cross-cut Sarıkaraman Ophiolite; bearing K-feldspar megacrysts, medium to coarse grained, light coloured, contains ophiolitic enclaves.	Essentially qtz-plag-K-feldspar-hb $\pm$ bio bearing quartz monzonite.	81–65 Ma		Post-COLG
(5) Epiclastic volcanogenic deep-sea sediments intercalated with pelagic units.		Middle Turonian–Early Santonian		Marginal basin
(4) Mainly pillow lavas with subordinate massive and breccias.	Generally vesicular, aphyric and plag-cpx phyric tholeiites.		Sub-alkaline tholeiites; high LILE/HFSE ratio relative to N-MORB-IAT + subduction zone component (especially Th enrichment relative to Ta-Nb).	SSZ
(3) Entirely sub-vertical to vertical dolerite 'dyke in dyke' structure, with asymmetric chilling.	Plag-cpx bearing ophitic to sub-ophitic.		IAT	SSZ
(2) Occur as simple narrow fracture infilling to wide complex zones of netveining in gabbros; dykes and pods in dyke complex and volcanic units; sharp to diffused margin with the host rocks.	Trondhjemite, generally equigranular, consisting largely of plagioclase and quartz. Texture range from hypidiomorphic granular to granopyric.		Low-K; ocean ridge granite	SSZ
(1) Isotropic, coarse to fine grained; frequently amphibolized.	Non-cumulus granular, with primary Ca-plag and relict cpx, with secondary amphiboles.		In bulk, low-Ti in mineral chemistry: Cpx; low-Ti island arc; Plag; high Ca (An 82-94); Amp; high temp., low pressure.	SSZ

deep-sea sediments and debris flow intercalated with pelagic units. The thickness of the epi-ophiolitic sedimentary cover attains approximately 400 m and exhibits three different units: (1) lower volcanogenic unit; (2) pelagic unit; and (3) upper volcanogenic unit (Figure 3). The samples, collected from the lower volcanogenic unit and pelagic unit, yielded poorly preserved (mostly recrystallized forms), moderately

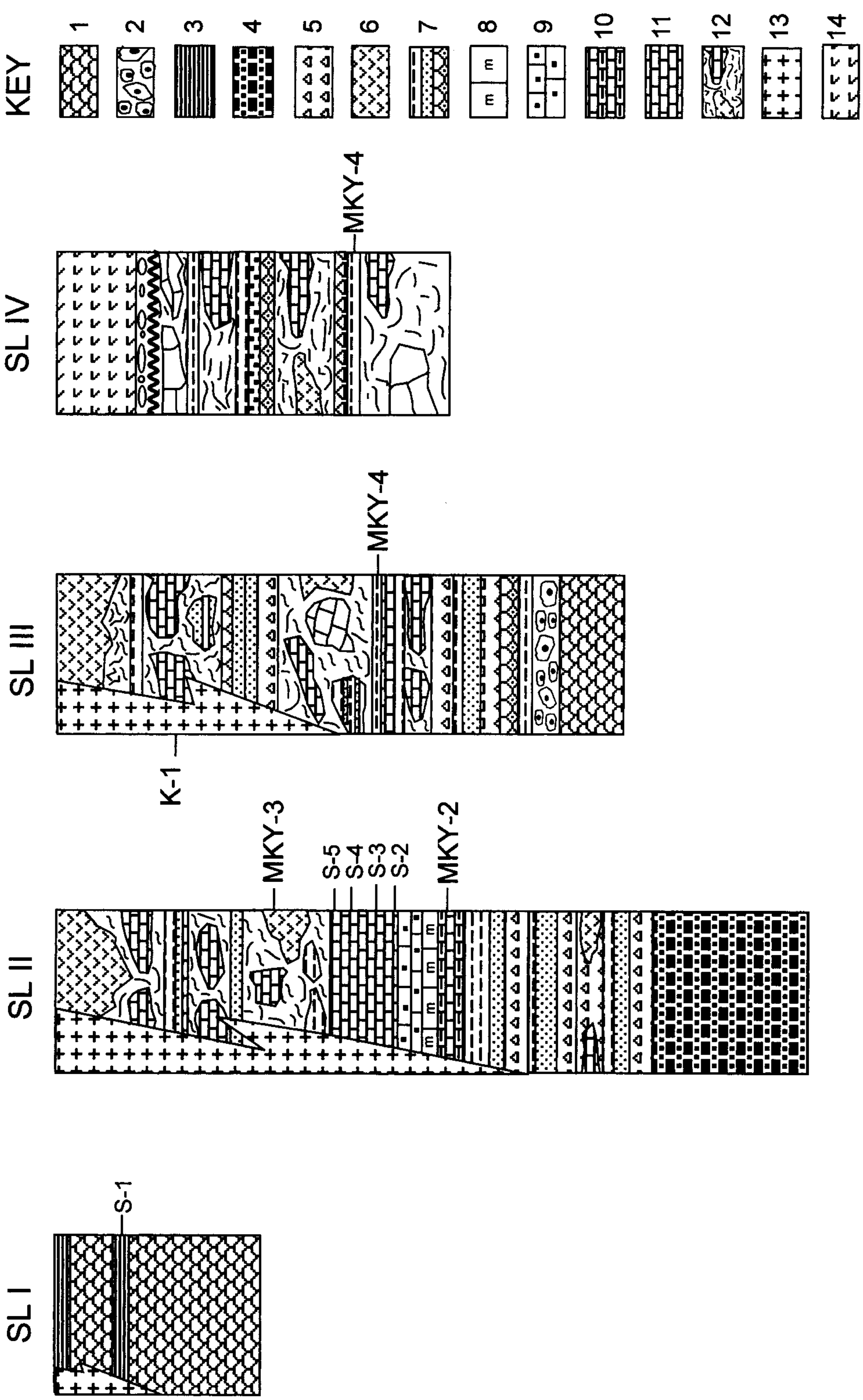


Figure 2. Measured columnar section of the epi-ophiolitic cover of the Sarkaraman Ophiolite, Central Anatolia. Key: 1, pillow lava; 2, pillow breccia; 3, interlava red mudstone; 4, turbiditic siltstone and sandstone; 5, volcanogenic breccia; 6, rhyolitic tuffs and flows; 7, volcanogenic sandstone; 8, manganiferous cherts; 9, radiolarian rich cherty limestone; 10, cherty micritic limestone; 11, dark pelagic limestone; 12, volcanogenic debris flow deposits; 13, post-collisional granitoids; 14, volcaniclastic cover sediments.

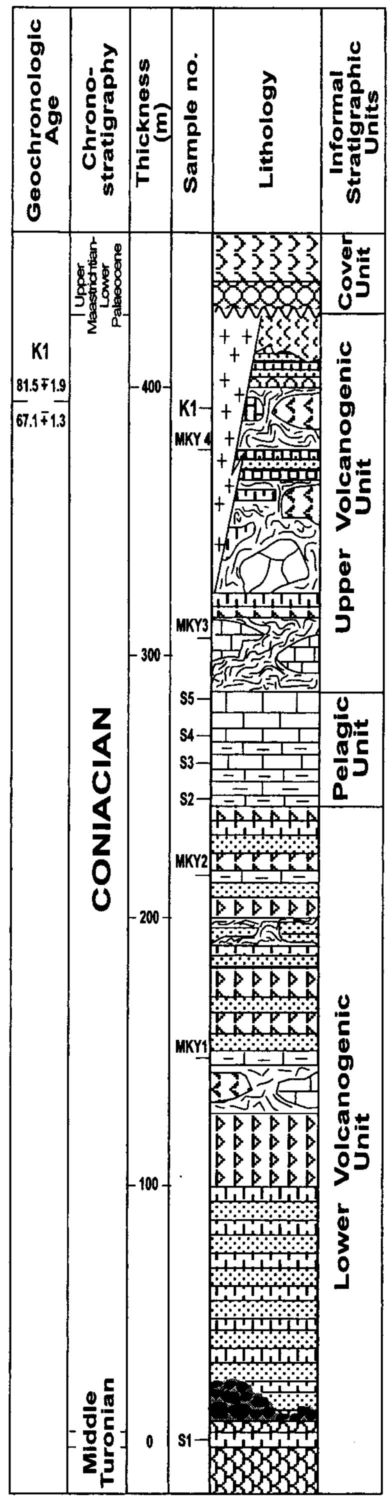


Figure 3. Generalized columnar section of the epi-ophiolitic sediments in the Sarıkaraman area (for explanations see Figure 2).

diversified planktonic foraminiferal assemblages. The diagnostic and critical forms for age determination of the epi-ophiolitic sequence of the Sarıkaraman Ophiolite are as follows.

The lower volcanogenic unit represents the lowermost part of the epi-ophiolitic sequence with a total thickness of approximately 240 m. The base of this unit is marked by red-coloured mudstones intercalated with the uppermost pillow lavas which are overlain by 100 m thick green-coloured structureless volcanogenic turbiditic sandstone, siltstone and shale alternations.

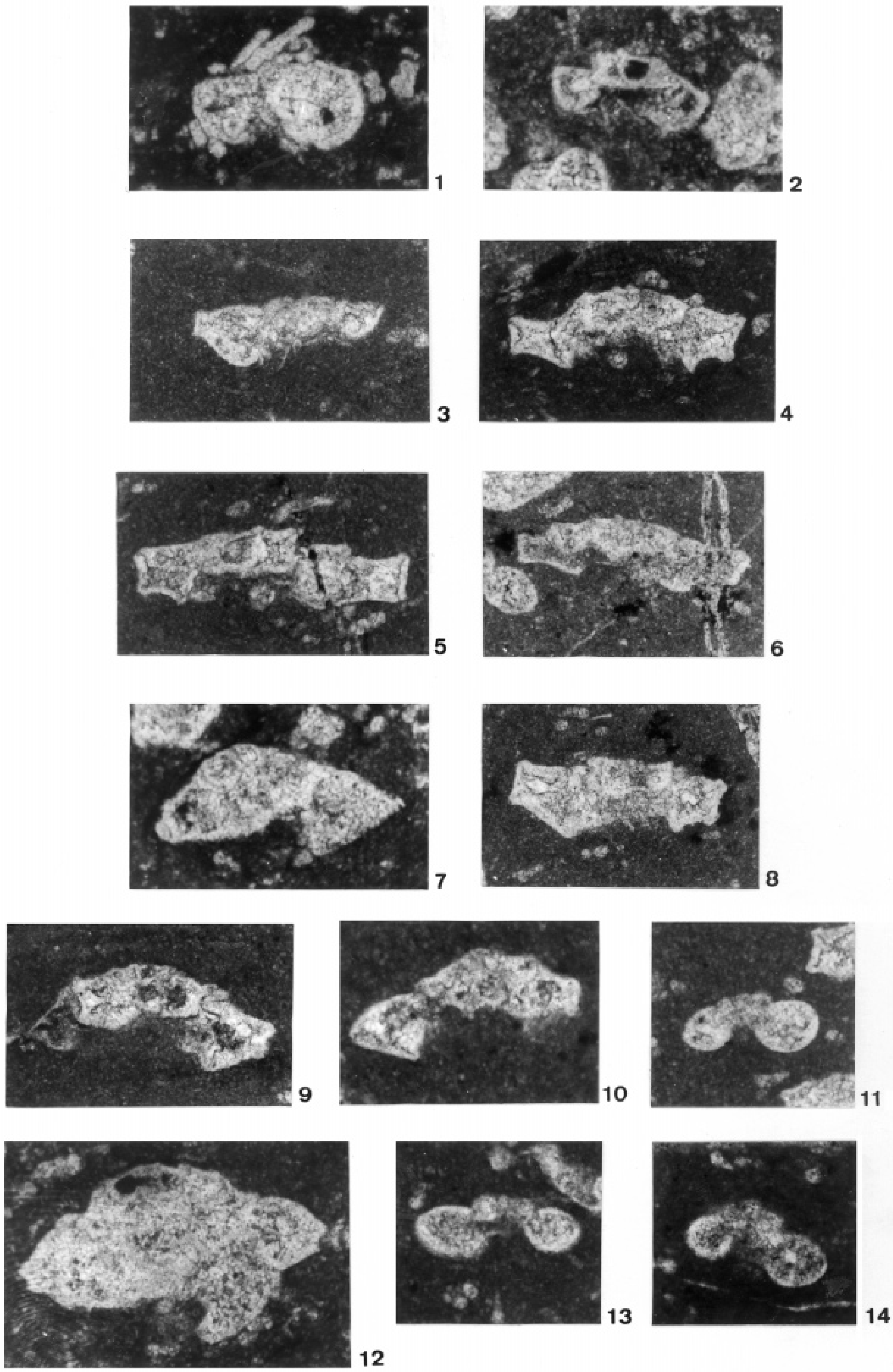
The richest and most significant planktonic foraminiferal association was recorded from a sample taken from the red-coloured mudstone intercalated with the uppermost pillow lavas: *Helvetoglobotruncana helvetica* (Bolli) (Figure 4.2), *H. praehelvetica* (Trujillo) (Figure 4.1), *Dicarinella algeriana* (Caron) (Figure 4.12), *Marginotruncana sigali* (Reichel) (Figure 4.7), *M. schneegansi* (Sigal) (Figure 4.10), *Globigerinelloides* sp., *Hedbergella* sp., *Heterohelix* sp., *Sigalia* sp.

According to Caron (1985), the stratigraphic distributions of *Dicarinella algeriana* (Caron) and *Helvetoglobotruncana praehelvetica* (Trujillo), *Marginotruncana sigali* (Reichel), and *Marginotruncana schneegansi* (Sigal) are late Cenomanian–middle Turonian, middle Turonian–early Santonian and middle Turonian–late Santonian, respectively. However, the stratigraphic range of *Helvetoglobotruncana helvetica* (Bolli) is restricted to middle Turonian (Caron 1985; Sliter 1989). The Turonian marker, *Helvetoglobotruncana helvetica* has been encountered sporadically in sample S-1. Sigal (1977) firstly proposed the *Helvetoglobotruncana helvetica* Zone as a total range zone. Although Postuma (1971) used this form as the marker species of Turonian and Bolli (1966) as that of late Turonian, Van Hinte (1976), Sigal (1977), Wonders (1980), Caron (1985), Sliter (1989) and Robaszynski *et al.* (1990) recorded *Helvetoglobotruncana helvetica* as the marker species of middle Turonian. In the most recent paper on the biostratigraphy of the planktonic foraminifers (Robaszynski and Caron 1995), the stratigraphic ranges of the zones and some marker fossils have been calibrated with ammonites and/or with data from palaeomagnetic reversals. Hence, the stratigraphic distribution of *Helvetoglobotruncana helvetica* has been slightly shifted from middle Turonian, to early to lowest part of the middle Turonian. In addition to this marker fossil, the last occurrence of *Dicarinella algerina* (Caron) and the first occurrences of *Marginotruncana sigali* and *Marginotruncana schneegansi* only overlap in the lowermost part of the Middle Turonian age. After this revision in the planktonic foraminiferal zonal scheme (Robaszynski and Caron 1995), the planktonic foraminiferal assemblages including *Helvetoglobotruncana helvetica* are attributed to an early middle Turonian age for this sample.

The continuous deposition of turbiditic units is interrupted by about 140 m of poorly bedded volcanogenic breccia. Breccias are unstratified and composed of mainly angular to sub-angular clasts of volcanic (basalt, rhyolite and tuffs) and sedimentary (clastic and pelagic) rocks. Clast sizes range from <1 mm to blocks with dimension up to 5 m. These breccias are very poorly sorted and clast supported, with a variable volcanogenic sandy or silty clayey matrix. Poor sorting, the presence of a fine-grained matrix, lack of stratification and a chaotic disorganized fabric all suggest that the coarsest-grained sediments were deposited as debris flows. The massive volcanogenic breccias are frequently interbedded with turbiditic–epiclastic medium- to thin-bedded sandstones, siltstone and shale. The epiclastic rocks are structureless or crudely graded with angular to sub-angular clasts of igneous fragments (basalt–rhyolite) and very poor sortings.

The sequence is capped by approximately 55 m of thin- to medium-bedded deep-sea pelagics. These units consist mainly of very well-stratified green-coloured micrites to grey-pink or brown-coloured cherty limestone, dark red-coloured radiolarian and manganiferous chert. They are capped by a thick, uniformly

Figure 4. (1) *Helvetoglobotruncana praehelvetica* (Trujillo), axial section, Sample S-1, x40. (2) *Helvetoglobotruncana helvetica* (Bolli), axial section, Sample S-1, x40. (3) *Dicarinella primitiva* (Dalbiez), axial section, Sample S-5, x40. (4) *Marginotruncana coronata* (Bolli), axial section, Sample S-2, x65. (5) *Marginotruncana coronata* (Bolli), axial section, Sample S-5, x70. (6) *Marginotruncana coronata* (Bolli), axial section, Sample S-3, x65. (7) *Marginotruncana sigali* (Reichel), axial section, Sample S-1, x75. (8) *Marginotruncana pseudolinneiana* Pessagno, axial section, Sample S-3, x65. (9) *Marginotruncana sinuosa* Porthault, axial section, Sample S-2, x50. (10) *Marginotruncana schneegansi* (Signal), axial section, Sample S-4, x50. (11) *Whiteinella archaeocretacea* Pessagno, axial section, Sample S-3, x70. (12) *Dicarinella algerina* (Caron), axial section, Sample S-1, x70. (13) *Whiteinella* sp., axial section, Sample S-5, x70. (14) *Whiteinella archaeocretacea* Pessagno, axial section, Sample S-2.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bedded pinkish cherty limestone. The following planktonic foraminifers have been recorded from the pelagic units in ascending stratigraphic order: *Marginotruncana coronata* (Bolli) (Figure 4–6), *M. pseudolinneiana* Pessagno (Figure 4.8), *M. sinuosa* Porthault (Figure 4.9), *Dicarinella primitiva* (Dalbiez) (Figure 4.3), *Whiteinella archaeocretacea* Pessagno (Figure 4.11, 14), *Globigerinelloides* sp., *Hedbergella* sp., *Whiteinella* sp., (Figure 4.13), Radiolaria.

The stratigraphic ranges of *Marginotruncana coronata* (Bolli), *Marginotruncana pseudolinneiana* Pessagno, *Marginotruncana sinuosa* Porthault and *Whiteinella archaeocretacea* Pessagno are middle Turonian–early Campanian, middle Turonian–Santonian, late Turonian–late Santonian and Turonian–Coniacian, respectively (Caron, 1985). *Dicarinella primitiva* (Dalbiez) is a typical Coniacian–early Santonian marker (Caron 1985). The first occurrence of this species, however, has been recorded in the late middle Turonian (Robaszynski *et al*., 1990). Wonders (1980) used this form as the marker of Coniacian which corresponds to the lower part of *Dicarinella concavata* zone of Sigal (1977) and *Marginotruncana schneegansi* zone of Bolli (1966) and Postuma (1971). Caron (1985) used *Dicarinella primitiva* as the nominal species of the interval zone of early Coniacian. Therefore, the presence of index form *Dicarinella primitiva* (Dalbiez) together with *Whiteinella archaeocretacea* Pessagno in sample S-5 infers an age of late middle Turonian–Coniacian for the uppermost level of the pelagic unit (Robaszynski and Caron 1995).

The well-bedded pelagic sequence is sharply overlain by massive cohesive debris-flow deposits, approximately 150 m thick, comprising randomly oriented clasts of volcanogenic clastics and pelagics. These debris-flow units are frequently interbedded with fine- to coarse-grained epiclastic and turbiditic units characterized by a thinning- and fining-upward sequence (mudstones, sandstones and conglomerates). They are composed largely of ophiolite-derived clasts (pillow lavas and breccias) and probably formed by erosion of submarine fault scarps on the ocean floor. The matrix of debris-flow units consists mainly of silty to sandy shale with a cleavage parallel to the bedding of the interbedded epiclastic and turbiditic units. The clasts range in size from a few millimetres to several metres and represent more or less disrupted remnants of the originally stratified underlying sequence. The continuous deposition of debris flows and the intermittent deposition of epiclastic and turbiditic units are periodically interrupted by massive influxes of volcanogenic units (mainly rhyolitic volcanics–tuffs). The uppermost part of this sequence is characterized by such slides of huge blocks of silicified rhyolitic tuffs within a marly matrix derived from an adjacent arc by debris flow. This unit is unfossiliferous and unconformably overlain by terrigenous conglomerates and volcaniclastic sediments of latest Maastrichtian–Early Paleocene age (Table 1; Göncüoğlu *et al*., 1991; Yalınız 1996; Dirik and Göncüoğlu 1996).

4b. Age and palaeo-environmental significance

Moores (1982) has suggested that the type of sedimentary sequences overlying ophiolite assemblages may allow marginal-basin-spreading ophiolites to be distinguished from those generated at mid-ocean ridges. He indicated that a back-arc-basin ophiolite might typically be overlain by volcaniclastic sequences, whereas a mid-ocean-ridge ophiolite would typically be overlain by deep-sea pelagic sediments. However, volcaniclastic sequences above ophiolites indicating a back-arc origin are unclear because many fore-arc ophiolites are also characterized by similar volcaniclastic sedimentation prior to obduction (e.g. Gealey 1980; Casey and Dewey 1984). It is difficult to distinguish back-arc from fore-arc volcaniclastic sequences (Casey *et al*., 1985). Recent deep-ocean drilling shows that fore-arc and arc volcanism is characterized by marginal-sea-type sedimentation were acidic volcanism is also known in the fore-arcs of the Bonin, Mariana and Tonga arcs (references in Leggett 1982; Robertson 1994).

However, the well-known and most complete example of inferred SSZ ophiolites of the southern branch of Neotethys in the Eastern Mediterranean (e.g. the Pindos, Antalya, Troodos, Hatay, Baer–Bassit, Semail Nappe) is overlain by pelagic sediments which contain no significant volcanogenic component (Pearce *et al*., 1984). The ideal SSZ ophiolites would be overlain by pelagic and/or volcanogenic (non-terrigeneous) deep-sea sediments (Robertson, 1994). These ophiolites are formed above the subducted slab during the initial stage of subduction prior to the development of any volcanic arc when subducted slabs of oceanic crust roll-back and slab pull leading to a short period of crustal extension in a supra-subduction zone setting (i.e. pre-arc or fore-arc spreading) (Pearce *et al*., 1984). This result is not surprising as none of these complexes is related in space or time to subaerial arc volcanoes as exemplified by Moores *et al*. (1984) for the southern branch of the Neotethyan ocean. The simplest explanation is that the spreading events that formed these complexes took place immediately after subduction and that the subduction event was too short-lived for any subaerial arc to develop and hence for any source of volcanogenic sediments to be available in these marginal basins. Ophiolites that were associated with longer lived subduction events are, however, overlain by sediments or volcanogenic sedimentary sequences of arc derivation (Pearce *et al*., 1984).

The Sarıkaraman Ophiolite is directly overlain by up to 500 m thickness of epi-ophiolitic sediments, which range from early middle Turonian to early Santonian in age. The sediments are characterized by repetition of slow and rapid deposition of epiclastics, turbiditic volcanogenic sediments and debris flows intercalated with pelagic units. The pelagic units, which are typically found in the uppermost part of the ophiolitic sequence, provide strong evidence for a deep-water origin for the lavas. However, the volcanogenic debris-flow sedimentary rocks interbedded with pelagics are suggestive of deposition in an active submarine rift adjacent to a penecontemporaneous active volcanic arc. Penecontemporaneous tectonism with rifting is reflected in repetitions in the stratigraphy and in debris flows which result from major slides and mass-gravity reworking of the pre-existing deep-sea sediments (mainly pelagics) and of arc-derived volcanics and sediments. In addition, the absence of andesitic detritus suggests that deposition was either before commencement of calc-alkaline arc volcanism, or that the site was topographically isolated from the arc. The former is more likely and these sediments possibly represent the initial stages of island arc evolution. This also supports the assertion proposed by Yalınız *et al*. (1996) that the Central Anatolian Ophiolites were generated in a supra-subduction zone setting.

Unlike the ophiolites of the southern branch of Neotethys characterized by the presence of pelagic sediments which contain no significant volcanogenic component, the marked high volume of volcanogenic component intercalated with the pelagic rocks of the Sarıkaraman Ophiolite are suggestive of adjacent active arc volcanism in a marginal basin setting.

5. DISCUSSION

Although much remains to be clarified in the evolution of Neotethyan basins, the consensus is that the Neotethyan ocean was initiated by Permo-Triassic rifting of the passive margin of Gondwanaland followed by sea-floor spreading (Dewey 1976). This opening marked the birth of Neotethys behind the Cimmerian continent which, at that time, began to separate from northern Gondwanaland. During the Middle Jurassic only three oceanic basins were left in Turkey (Figure 5): the Intrapontide Ocean separating the Eurasian continent from Gondwana-originated continental microplates (e.g. Sakarya Microplate, Tauride–Anatolide Platform); the northern and presumably multi-armed northern branch (Vardar and İzmir–Ankara–Erzincan oceans) separating the Sakarya Microplate from typically Gondwanan Apulian–Tauride–Anatolide Platform; and the southern branch between the Tauride–Anatolide Platform and the main body of Gondwanaland (Göncüoğlu *et al*. 2000).

As the North Atlantic started to open in early Late Jurassic, the western Neotethyan oceanic basins came under compression. The subduction of the wedge-shaped Neotethyan oceans was first initiated within the Pindos and Vardar marginal ocean basins in the Greek area in early Late Jurassic and SSZ-type ophiolites were generated above a westward subduction (Robertson *et al*., 1990). The ophiolites include Pindos and Vourinos in Greece, the Mirdita Zone ophiolites of Albania and those of the Southwest Belt in Yugoslavia. Further east, there is no evidence of compressional tectonics during Late Jurassic–Early Cretaceous; instead

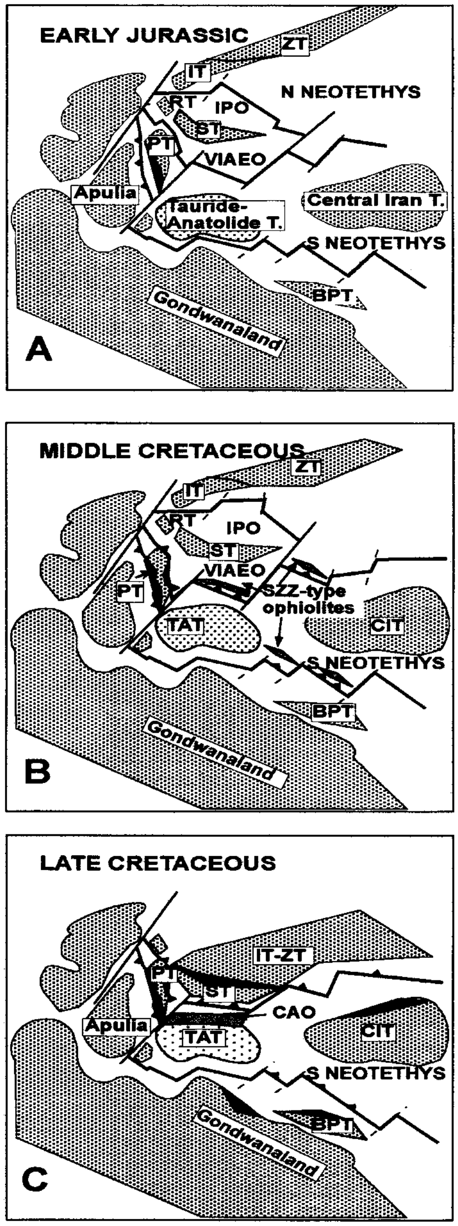

Figure 5. Reconstructions of the Eastern Mediterranean Tethys. (**A**) Early Jurassic; (**B**) middle Cretaceous; (**C**) Late Cretaceous. BPT, Bitlis–Pötürge Terrane; CIT, Central Iran Terrane, IPO, Intrapontide Ocean; IT, İstanbul Terrane; PT, Pelagonian Terrane; RT, Rodope Terrane; ST, Sakarya Terrane; TAT, Tauride–Anatolide Terrane; VIAE, Vardar–İzmir–Ankara–Erzincan Ocean; ZT, Zonguldak Terrane. See text for discussion.

crustal extension and volcanism were renewed, as seen in SW Cyprus (Mamonia Complex), N Syria (Baer–Bassit) and SW Turkey (Antalya Complex) (Robertson *et al*., 1990).

In Early to middle Cretaceous time, the relative convergence between Afro-Arabia and Eurasia changed to more north–south orientation resulting in a regional convergence further east in the multi-armed northern branch and southern branch of the Neotethys. The small Neotethyan SSZ-type ophiolites in the southern branch of Neotethys that extends from Cyprus (Troodos) through Turkey (Mersin–Antalya–Hatay), Syria (Baer–Bassit) and Iran (Zagros) to Oman (Semail) were created in Early to middle Cretaceous due to intra-oceanic subduction. During this convergent regime, away from the passive margin of the Anatolide–Tauride Platform, a northward subduction was also initiated within the multi-armed northern branch (VIAE ocean) in Early to middle Cretaceous. CAO (of which the Sarıkaraman Ophiolite is one example) were formed above this intra-oceanic subduction as fore-arc ophiolites by the partial melting of already depleted MORB-type VIAE oceanic lithosphere during early middle Turonian–early Santonian (Yalınız *et al.*, 1996, 1999).

The sub-ophiolitic amphibolites of the VIAE belt from western Central Anatolia (Önen and Hall 1993) yielded closely spaced K/Ar and Ar/Ar ages that range between 93 and 90 Ma, which clearly indicates a middle Cretaceous initial decoupling of oceanic crust. The upper level gabbros and dykes of SSZ-type oceanic crust in the same area, on the other hand, yielded isochron ages of 85 Ma, which obviously suggest that the formation of these ophiolites post-date the initial intra-oceanic subduction.

This suggestion would not rule out that the sea-floor spreading in the VIAE may have already started in the early Mesozoic. New fossil data from blocks of sediments associated with pillow lavas within the mélanges along the VIAE suture indicate a wide chronological range from late Norian to Turonian (Göncüoğlu *et al*. 2000; Rojay *et al*., 1995; Bragin and Tekin 1996). The geochemical data on the associated volcanics are very limited and indicate a very wide series of palaeotectonic settings ranging from MORB to OIB (Floyd 1993) including WPB and arc-related calk-alkaline types (Yalınız *et al*. 2000). However, there is no adequate control on the ages of these volcanics and thus on the palinspastic reconstruction of the oceanic crust generation. Nevertheless, these data imply that besides the early Late Cretaceous SSZ-type oceanic crust that is mainly encountered as dismembered allochthonous bodies on the Tauride–Anatolide passive margin, remnants of an older and MORB-type oceanic crust representing the main body of Neotethys has existed since late Norian. We assume that this main oceanic basin was consumed by northward subduction beneath the Sakarya Microcontinent and some parts were accreted to form the mélange complexes along the VIAE suture (e.g. Ankara Mélange) during latest Cretaceous.

Obduction of the ophiolitic nappes occurred all along the northern margin of the Tauride–Anatolide Platform during Campanian–Maastrichtian times (Ricou *et al*., 1975; Özgül 1976; Göncüoğlu *et al*. 2000) and marks the beginning of the Neotethyan deformation. In central Anatolia (Göncüoğlu *et al*., 1991) obduction of the ophiolitic nappes onto the northern margin of the Tauride–Anatolide platform was realized in two successive events. Initial obduction of the older MORB-type oceanic assemblages onto the Tauride–Anatolide margin resulted in the deposition of an ophiolite-bearing olistostrome within a flexural foreland basin formed in front of the ophiolitic nappes. The obduction of these ophiolitic nappes caused deep burial and metamorphism of the platform-margin units together with the ophiolitic olistostromes (metamorphic olistostromes around the Niğde region; Göncüoğlu 1986). The second event was the obduction of SSZ-type ophiolites and the subsequent collision of the ensimatic arc. Structural data indicate obduction towards the SSW. The generation of the Central Anatolian Granitoids (CAG) in early Late Cretaceous time is related to this terminal collision and subsequent post-collisional uplifting (Göncüoğlu *et al*., 1991).

The SSZ-type Sarıkaraman Ophiolite with related early middle Turonian–early Santonian sediments is intruded by the post-collision type Terlemez Quartz-Monzonite. The K/Ar mineral and whole-rock data from the Terlemez Quartz-Monzonite ($81{\cdot}5 \pm 1{\cdot}9$ to $67{\cdot}1 \pm 1{\cdot}3$ Ma) indicate a late Campanian–early Maastrichtian intrusion age (Table 1; Yalınız and Göncüoğlu 1998; Yalınız *et al*., 1999). The oldest sediments disconformably overlying the CAO and the CAG are represented by continental to shallow-marine deposits related to post-collisional extension of latest Maastrichtian–Early Palaeocene age (Table 1; Göncüoğlu *et al*., 1991). The radiometric age of the Terlemez Quartz-Monzonite and the palaeontological data from the cover units are in good accordance with the palaeontological data obtained from the epi-ophiolitic sediments of the Sarıkaraman Ophiolite.

A further regional implication of the new palaeontological data is the demonstration that SSZ-type oceanic crust formation in the Turkish part of the VIAE ocean is definitely younger than in the Carpathian–Balkan region, which was suspected in previous studies (e.g. Robertson and Dixon 1984).

There are several models to explain the difference both in formation and emplacement ages of the SSZ-type ophiolites in the eastern and western parts of the VIAE ocean, e.g. half-ridge/transform fault model of Smith and Spry (1984) or progressive asymmetrical spreading/ridge collapse model of Robertson and Dixon (1984). The new age data in this study indicate that Robertson and Dixon's (1984) model cannot be applied to the western and central Anatolian parts of the VIAE branch of Neotethys. However, the available data on the geochemical characteristics of the oceanic crusts, age of epi-ophiolitic sediments, and sub-ophiolitic soles are still too fragmentary and imprecise for a reliable geodynamic reconstruction of the Eastern Mediterranean area.

6. CONCLUSIONS

(1) Volcanogenic debris flows and epiclastics interbedded with pelagics of the epi-ophiolitic cover of the Sarıkaraman Ophiolite confirm the presence of an adjacent active arc during the formation of SSZ-type Central Anatolian Ophiolites.
(2) It is suggested that the oceanic crust of the CAO, of which the Sarıkaraman Ophiolite is a detailed example, was formed above a northward-dipping intra-oceanic subduction in a supra-subduction zone setting within the Vardar–İzmir–Ankara–Erzincan branch of Neotethys in the early middle Turonian–early Santonian.
(3) It is also suggested that the CAO was emplaced onto the CACC between late Santonian and late Campanian and intruded during late Campanian–early Maastrichtian by COLG-type CAG, that had been generated from CACC basement due to crustal thickening. During latest Maastrichtian–Early Palaeocene all these CAO, CAG and Central Anatolian metamorphics were disconformably overlain by sediments related to post-collisional extension.
(4) Although the Central Anatolian Ophiolites display a similar SSZ-type tectonic setting to the other SSZ-type ophiolites of the Late Cretaceous southern branch of Neotethys, epi-ophiolitic sedimentary cover is lithologically different and is characterized by the presence of a significant island arc volcanogenic component.
(5) SSZ-type oceanic crust formation in the Turkish part of the VIAE ocean is definitely younger than in the Carpathian–Balkan region.

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