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TETHYAN EVOLUTION OF TURKEY: A PLATE TECTONIC APPROACH

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This paper is dedicated to Professor Jan Houghton Brunn on the occasion of his retirement from active teaching and in recognition of his fundamental contributions to our understanding of the tectonics of Turkey over the past decade and-a-half.

ABSTRACT

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The Tethyan evolution of Turkey may be divided into two main phases, namely a Palaeo-Tethyan and a Neo-Tethyan, although they partly overlap in time. The Palaeo-Tethyan evolution was governed by the main south-dipping (present geographic orientation) subduction zone of Palaeo-Tethys beneath northern Turkey during the Permo-Liassic interval. During the Permian the entire present area of Turkey constituted a part of the northern margin of Gondwana-Land. A marginal basin opened above the subduction zone and disrupted this margin during the early Triassic. In this paper it is called the Karakaya marginal sea, which was already closed by earliest Jurassic times because early Jurassic sediments unconformably overlie its deformed lithologies. The present eastern Mediterranean and its easterly continuation into the Bitlis and Zagros oceans began opening mainly during the Carnian—Norian interval. This opening marked the birth of Neo-Tethys behind the Cimmerian continent which, at that time, started to separate from northern Gondwana-Land. During the early Jurassic the Cimmerian continent internally disintegrated behind the Palaeo-Tethyan arc constituting its northern margin and gave birth to the northern branch of Neo-Tethys. The northern branch of Neo-Tethys included the Intra-Pontide, Izmir—Ankara, and the Inner Tauride oceans. With the closure of Palaeo-Tethys during the medial Jurassic only two oceanic areas were left in Turkey: the multi-armed northern and the relatively simpler southern branches of Neo-Tethys. The northern branch separated the Anatolide—Tauride platform with its long appendage, the Bitlis—Pötürge fragment from Eurasia, whereas the southern one separated them from the main body of Gondwana-Land. The Intra-Pontide and the Izmir—Ankara oceans isolated a small Sakarya continent within the northern branch, which may represent an easterly

continuation of the Paikon Ridge of the Vardar Zone in Macedonia. The Anatolide—Tauride platform itself constituted the easterly continuation of the Apulian platform that had remained attached to Africa through Sicily. The Neo-Tethyan oceans reached their maximum size during the early Cretaceous in Turkey and their contraction began during the early late Cretaceous. Both oceans were eliminated mainly by north-dipping subduction, beneath the Eurasian, Sakaryan, and the Anatolide—Tauride margins. Subduction beneath the Eurasian margin formed a marginal basin, the present Black Sea and its westerly prolongation into the Srednogorie province of the Balkanides, during the medial to late Cretaceous. This resulted in the isolation of a Rhodope—Pontide fragment (essentially an island arc) south of the southern margin of Eurasia. Late Cretaceous is also a time of widespread ophiolite obduction in Turkey, when the Bozkir ophiolite nappe was obducted onto the northern margin of the Anatolide—Tauride platform. Two other ophiolite nappes were emplaced onto the Bitlis—Pötürge fragment and onto the northern margin of the Arabian platform respectively. This last event occurred as a result of the collision of the Bitlis—Pötürge fragment with Arabia. Shortly after this collision during the Campanian—Maastrichtian, a subduction zone began consuming the floor of the Inner Tauride ocean just to the north of the Bitlis—Pötürge fragment producing the arc lithologies of the Yüksekova complex. During the Maastrichtian—Middle Eocene interval a marginal basin complex, the Maden and the Çüngüş basins began opening above this subduction zone, disrupting the ophiolite-laden Bitlis—Pötürge fragment. The Anatolide—Tauride platform collided with the Pontide arc system (Rhodope—Pontide fragment plus the Sakarya continent that collided with the former during the latest Cretaceous along the Intra Pontide suture) during the early to late Eocene interval. This collision resulted in the large-scale south-vergent internal imbrication of the platform that produced the far-travelled nappe systems of the Taurides, and buried beneath these, the metamorphic axis of Anatolia, the Anatolides. The Maden basin closed during the early late Eocene by north-dipping subduction, synthetic to the Inner-Tauride subduction zone that had switched from south-dipping subduction beneath the Bitlis—Pötürge fragment to north-dipping subduction beneath the Anatolide—Tauride platform during the later Palaeocene. Finally, the terminal collision of Arabia with Eurasia in eastern Turkey eliminated the Çüngüş basin as well and created the present tectonic regime of Turkey by pushing a considerable piece of it eastwards along the two newly-generated transform faults, namely those of North and East Anatolia. Much of the present eastern Anatolia is underlain by an extensive mélange prism that accumulated during the late Cretaceous—late Eocene interval north and east of the Bitlis—Pötürge fragment.

INTRODUCTION

The analysis of the structure of Turkey has proved exceptionally difficult in the past, because of a large concentration of a number of convergent events throughout its history. The mountain ranges of Turkey constitute the easternmost segment of the Mediterranean Alpides, where Palaeo-Tethyan and Neo-Tethyan deformations have been superimposed on a continental basement (partly Pan African—Baykalian and partly Hercynian ages), whereas a portion of the Neo-Tethyan evolution alone has been recorded by the predominantly late Mesozoic—Cenozoic accretionary complex of eastern Anatolia (see Fig. 1). The Turkish orogen forms a critical link between the Mediterranean and the Asiatic Tethyan* systems, because Tethyan palaeogeographic elements peculiar to each meet and terminate in the Anatolian Peninsula (Asia Minor). Turkey is an ideal location in which to study the relationships of the Palaeo- and the Neo-Tethyan systems, not only because both are well-developed, but also because despite the very intense collisional events that accompanied the closure of Neo-Tethys, pre-Neo-Tethyan-opening palaeogeography of the area can be reconstructed with some confidence as a result of the Neo-Tethyan oceans having opened and closed roughly along the same zones in Turkey. As Tethyan structures cross-cut Hercynian trends in Turkey, the latter can be used as an independent check on the reconstructions. These unusually favourable conditions provide a good data base on which to tackle a variety of stratigraphic—structural problems that have so far eluded explanation.

Because the birth and the further evolution of the Neo-Tethyan oceans in Turkey and their tectonic relationships with older and to some extent contemporaneous structures such as Palaeo-Tethys and the Cimmerian Continent (Şengör, 1979a) can be so clearly observed and relatively easily interpreted in Turkey, the history of this segment of the Mediterranean Alpides adds a hitherto generally unconsidered perspective to the study of the Tethyan evolution of the Mediterranean region as a whole, namely an eastern, "Tethyan" one. So far, most of the plate tectonic syntheses of peri-Mediterranean post-Palaeozoic geology have been constructed in terms of the opening history of the Atlantic Ocean (e.g. Smith, 1971; Dewey et al., 1973; Biju-Duval et al., 1977, 1978; Tapponnier, 1977; Channel et al., 1979; Burchfiel, 1980). Although this has provided a well-constrained kinematic basis for the analysis of the tectonics of the Mediterranean area since the Toarcian, it has done little to clarify earlier Mesozoic events that increase both in complexity and in variety from west to east along the Tethyan chains of the peri-Mediterranean area. The "Tethyan" influence on the Mesozoic evolution of the Mediterranean region has been at least as important as the Atlantic interference and the geology of Turkey provides an excellent data base to discover its nature and to decipher its evolution.

The purpose of this paper is to review the Mesozoic—Cenozoic tectonic evolution of the Turkish orogenic belt in the framework of the theory of plate tectonics. We discuss the implications of our proposed model for the tectonics of Tethys in the general area of the Mediterranean and the Middle East and also for theoretical models of collisional orogeny in general. The

* The term "Alpide" is used here to designate the orogenic belts that descended from Tethys. "Tethys", in general, or "Tethyan domain" refer to the large, triangular, westward-narrowing embayment of Pangaea that separated Laurasia from Gondwana-Land and that include both Palaeo- and Neo-Tethys and their margins. As the Greek suffix "-ide" implies descendence, it would have been perhaps more appropriate to speak of "Tethysides" instead of "Alpides," but we continue to use the latter term for obvious historical reasons. Our usage, however, deviates from the employment of Stille (1924) and Kober (1921) in that it is geographical as opposed to theirs, which was temporal.

model is based on published literature, unpublished data made available to us by numerous colleagues both inside and outside Turkey, and our field observations. All unreferenced data in the following pages represent the results of our own field work. So as not to burden the text, we have stored much of the factual information in Figs. 2—5 and cited sources in their captions. A part of this information plus some other data have also been shown in Figs. 6A—6I with the main references given in the captions.

SUTURES AND MICROCONTINENTAL FRAGMENTS INVOLVED IN THE TETHYAN EVOLUTION OF TURKEY

The mountain belts of Turkey are the result of repeated continental collisions that eventually welded together the Old World segments of the two mega-continents of Laurasia and Gondwana-Land. Since its birth, contemporaneous with the assembly of Pangaea during the late Palaeozoic, the internal geometry of the Tethyan domain has been characterized by a complex array of plate boundary systems composed of a continuously evolving network of ridges, transforms and subduction zones whose record of activity is now found, in various states of preservation, mainly along the sutures of the Alpides, the sites of former Tethyan oceans. The first step in reconstructing the past configurations of continents and oceans is therefore to identify the sutures, which mark the places where oceans disappeared, and the continental fragments once separated by such oceans.

Figure 1 is a tectonic map showing the continental fragments involved in the Tethyan evolution of Turkey and their original geographic affiliations with respect to the Neo-Tethyan palaeogeography. From north to south these are the Rhodope—Pontide Fragment, the Sakarya Continent, the Anatolide—Tauride Platform and its elongated appendage, the Bitlis—Pötürge massifs. The nature of the northern boundary of the Rhodope—Pontide fragment is controversial. In the west, the boundary is marked largely by deformed Tithonian—Berriasian flysch deposits (Hsü et al., 1977; Sandulescu, 1978a, 1978b) and wildflysch-containing mafic volcanic detritus (Burchfiel, 1980) between the Rhodope Massif and the Moesian Platform. Dewey et al. (1973) and Burchfiel (1980) have interpreted the mid-Cretaceous boundary between the Rhodope Massif and the Moesian Platform as a

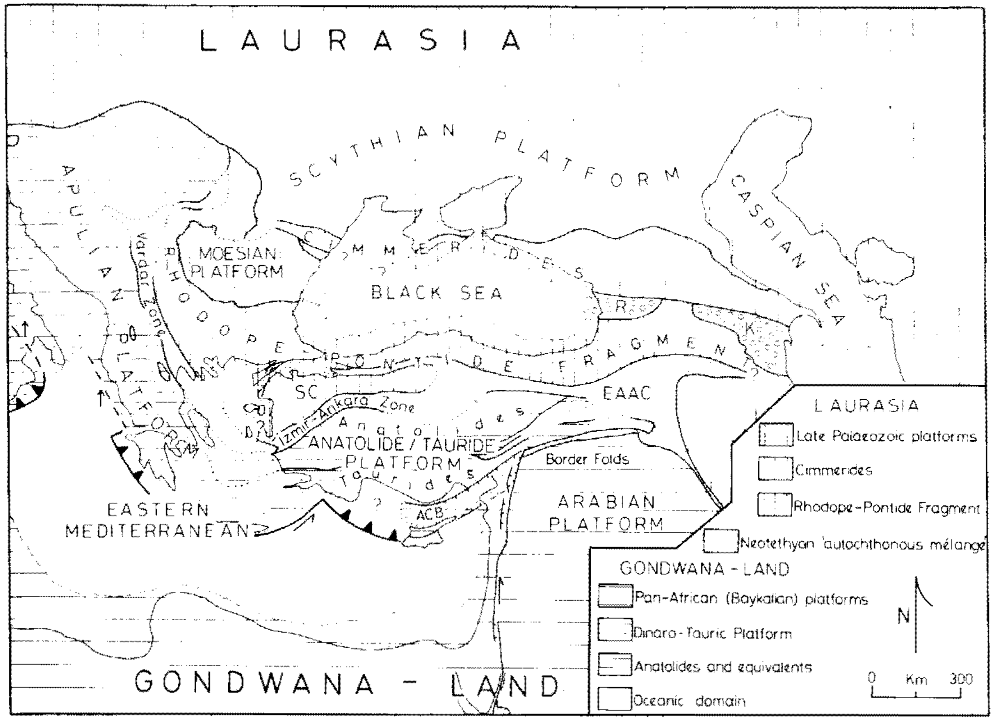

Fig. 1. Tectonic setting of Turkey within the larger framework of the eastern Mediterranean Alpides. The map was drawn from a Neo-Tethyan perspective (i.e., it shows only the sutures that represent Neo-Tethyan oceans) only to avoid too complicated a picture. Although the North Caspian Depression is very likely to have a ?late Devonian oceanic basement (Fedynsky et al., 1972; Garetsky et al., 1972; Burke, 1977; Kidd, in prep.), it has been included in the "late Palaeozoic platforms" because that has been how it has behaved since that time with respect to the Tethyan chains. A northeast striking ophiolitic suture, the "Intra-Pannonian Belt" of Channel et al. (1979), may cut the basement of the Pannonian Basin into two segments and join the Vardar Zone along the latter's northwest extension. The existence of this suture is suspected mostly from bore-hole data, and its geometry is very poorly known. We have left it out in our figure because its existence could also be explained by slicing off pieces of the northernmost extremity of the Vardar Zone along strike-slip faults that shoved a considerable piece of the basement of the Dinaro—Tauric Platform into the Carpathian bend during the Miocene (e.g. Burchfiel, 1980).

Cimmerides represent the orogenic belt(s) that resulted from the elimination of Palaeo—Tethys and her dependencies (marginal basins, etc.) during the late Triassic—middle Jurassic interval. Only those areas in this figure have been included in the Cimmerides, where the outlines of the present structure were determined by Cimmerian deformations. Regions where overprinting by later events has determined the outlines of the present structures have been left out of the Cimmeride domain such as the Rhodope—Pontide Fragment, despite the fact that it contains the main Palaeo—Tethyan suture in this area (Şengör et al., in press). In spite of the current strong deformation in Central Iran (e.g. Şengör and Kidd, 1979), it has been left within the Cimmerides, because there the main orogenic structures are of Cimmerian origin (e.g. Stöcklin, 1977). *SC* is Sakarya Continent (largely equivalent to Brinkmann's (1966) *Mysisch-Galatische Scholle*), *ACB* is Adana—Cilicia Basin, *EAAC* is Eastern Anatolian Accretionary Complex, *R* and *K* are Riou and Khoura depressions respectively. The small circle pattern in them indicates the Oligocene to Present molasse fill of these depressions. Heavy lines with black triangles are subduction zones with triangles on the upper plate. Lines with half-arrows are transform faults (not all of them are shown!) and lines with hachures on them are normal faults with hachures on the down-dropped side. Short, full arrows south of the Adriatic Sea indicate the present relative motion direction of Africa with respect to Europe. The Carpathian—Balkan regions are after Burchfiel (1980), whereas the Cimmeride domains east of the Caspian Sea are modified after Kotanski (1978).

suture along which a major oceanic tract has disappeared. Hsü et al. (1977) have pointed out, however, that the "eugeosynclinal" rocks of late Jurassic—early Cretaceous age of the Balkanides were deposited in an extensional basin, the Nish—Trojan flysch trough (Hsü et al., 1977), which had originated during the Jurassic and was most likely of modest dimensions throughout its life span. The similarity between the Bulgarian and the "Alpine" Triassic facies, the absence of ophiolites and oceanic deep-sea sediments of Triassic to middle Jurassic age in the Nish—Trojan trough and the observation that the main Palaeo-Tethyan suture is probably located within the circum-Rhodope orogenic zone of Kockel et al. (1971) and Kauffmann et al. (1976) (Roeder, 1978; Şengör, 1979; Şengör et al., in press), i.e. south of the Nish—Trojan trough, lead us to follow Hsü et al.'s (1977) interpretation as to the nature of the boundary between the Moesian Platform and the Rhodope Massif, which considers it as a narrow, intracontinental extensional basin prematurely shut before attaining oceanic characters (a "sialic ocean?" see Şengör and Monod, 1980).

In Turkey, the waters of the Black Sea hide the northern boundary of the Rhodope—Pontide Fragment, which reappears in the Caucasus as the flysch zone of the southern slope of the Greater Caucasus (Khain, 1975). Extension here probably started during the early Jurassic as evidenced by the development of the "Lias—Dogger slate—diabase association" (Khain, 1975). Following Cretaceous flysch deposition, the entire belt has been deformed through several episodes the latest of which continues. Widespread olistostrome development in front of the southward-advancing nappes characterizes the Eocene. The basin was probably eliminated synchronously with the Oligocene development of the Khoura and Riou molasse basins. Although the flysch zone in the Caucasus is probably located to the south of the main Palaeo-Tethyan suture (Şengör et al., in press), we do not believe that it ever was a major ocean. Farther east, the northern boundary of the Rhodope-Pontide Fragment passes across the Caspian Sea and is lost beneath the younger deposits of the Turan Platform (Stöcklin, 1977). As we shall see below, palaeontologic data also indicate that the Rhodope—Pontide fragment was never far away from Laurasia, at least not since the Lias. The only exception is probably its western end, where palaeomagnetic data show that the extreme curvature of the southern Carpathian—Balkan mountains along the western periphery of the Moesian Platform is the result of the previously straight (east—west trending) Rhodopian fragment having "wrapped around" the former (Dr. N. Orbay, personal communication, 1979), as has already been pointed out by Burchfiel (1980, figs. 7 and 8) on the basis of geological data.

The southern boundary of the Rhodope—Pontide fragment is marked by an ophiolitic suture zone of largely earliest Tertiary age. In the west this suture is represented by the Vardar Zone (Fig. 1) (Dewey et al., 1973; Bernoulli and Jenkyns, 1974; Biju-Duval et al., 1977, 1978; Channel et al., 1979). Kockel et al. (1971) have traced the internal limit of the Vardar Zone in the Aegean area from Thessaloniki through the peninsulas of Sithonia and

Athos to the island of Samothraki. From Samothraki, the suture strikes into the Gallipoli peninsula (Fig. 1), in the eastern part of which a late Cretaceous ophiolitic mélange crops out from beneath deformed Eocene flysch deposits (Fig. 2) (M.T.A., 1964 and our observations). On the eastern shore of the Sea of Marmara, along strike from the Gallipoli mélange, Akartuna (1968) has mapped what we interpret to be an extensive late Cretaceous ophiolitic mélange with a predominantly pelitic matrix, on the Armutlu Peninsula. This same mélange includes blueschist slivers in the Samanli Mountains north of Geyve, and in the Mudurnu Valley strikes into a relatively more intact, but still internally imbricated ophiolitic complex that comprises, from bottom to top (as restored) pyroxenites, layered gabbros, isotropic gabbros and finally pillow lava—chert intercalations. Following a minor shale horizon a late Cretaceous flysch was deposited on top and ?in front of the moving ophiolite nappe. Today, the ophiolites and the flysch are found tectonically interleaved and overthrust from the north by the so-called "Palaeozoic of Istanbul" (Fig. 2), which here makes up the Hercynian basement of the Rhodope—Pontide Fragment and shows a remarkable similarity to the Kraistides of Bulgaria in its stratigraphic development (Sandulescu, 1978b). Farther to the east-northeast, this ophiolitic zone has been followed through Bolu and Eskipazar up to the Ilgaz Massif (Fig. 2) by Brinkmann (1966) who described the lithologic characteristics of this zone as follows: "... die jungmesozoisch-alttertiären Sedimente [sind] grossenteils als Radiolarit oder Flysch, oft mit stärker Beteiligung basischer Ergussgesteine und Tuffe ausgebildet" (p. 614). Tokay (1973), who mapped the segment between Gerede and Ilgaz, described a late Cretaceous to Palaeocene ophiolitic mélange (the Arkotdag Formation) and ascribed its origin to both sedimentary and tectonic events that took place in a trench associated with a north-dipping subduction zone beneath the Rhodope—Pontide fragment. Between Gallipoli and the Ilgaz Massif, the ophiolitic suture, herein called the Intra-Pontide suture, separates the Rhodope—Pontide fragment from the Sakarya Continent (Fig. 1). Although strong magmatic arc volcanism of late Cretaceous age on the Rhodope—Pontide fragment in this area suggests that a considerable amount of oceanic lithosphere between it and the Sakarya Continent had been removed, the existence on the latter of typically European faunas, including such forms as *Epidoceras*, indicate that this ocean was not very large.

South and southeast of the Ilgaz Massif the ophiolitic suture zone continues to delimit the Rhodope—Pontide Fragment to the south (Fig. 1). East of Erzincan (Fig. 2) the mainly volcanic Neogene cover hides many details of the suture and obscures its relations to the Zangezur suture of the Lesser Caucasus (Knipper, 1979) and also to the predominantly oceanic rocks that make up a large portion of the basement of eastern Turkey such as the Yüksekova Complex (Perinçek, 1979) and its equivalents (Figs. 2 and 4). East of Ankara, the Sakarya Continent wedges out and the Ilgaz—Erzincan ophiolitic suture zone juxtaposes the Rhodope—Pontide Fragment and the Anatolide—Tauride Platform directly. The Izmir—Ankara Zone (Brinkmann, 1966, 1972, 1976) forms the boundary between the Sakarya Continent and the Anatolide—Tauride Platform (Fig. 1) and represents the remnants of a Jurassic (?Triassic) to early Palaeocene ocean that closed along a north-dipping subduction zone during the late Palaeocene—early Eocene (Dürr, 1975; Channel et al., 1979; Şengör, 1979b). Because it is largely submerged beneath the waters of the Aegean Sea, the western connection between the Intra-Pontide suture and the Izmir—Ankara zone is obscure. As the internal boundary of the Vardar Zone strikes east-northeast to delimit the Rhodope—Pontide Fragment to the south in the northern Aegean, whereas its external boundary seems to join that of the Izmir—Ankara Zone, it appears that both the Intra-Pontide and the Izmir—Ankara sutures link up with the Vardar Zone in the west and that the Sakarya Continent was an island in the northern branch of Neo-Tethys represented by the Vardar, Intra-Pontide, Izmir—Ankara and the Ilgaz—Erzincan ophiolitic belts. Because the late Mesozoic—Cenozoic structural trends take a sharp bend from a predominantly east—west orientation to an almost north—south orientation at the western end of the Sakarya Continent (e.g. Jacobshagen and Skala, 1977), and because the ophiolites in the island of Lesbos have this general strike, we delimited the Sakarya Continent to the west as shown in Figs. 1 and 6B—E. However, it is entirely possible that it represents a somewhat enlarged eastern continuation of the Paikon "Ridge" of the Vardar Zone (Mercier et al., 1975), with which it shares a common tectonic position, although Eo-Hellenic ophiolite obduction (Jacobshagen et al., 1976) is not known from the Sakarya Continent. It is interesting to note, on the other hand, that its predominantly neritic carbonate sedimentation of early late Jurassic age switches rather abruptly to a pelagic environment (Fig. 5, columns 3 and 4; Fig. 6D) synchronously with the emplacement of the Eo-Hellenic ophiolites onto the Paikon Ridge farther west along strike.

The Anatolide—Tauride Platform is the eastern end of the Dinaro-Tauric Plaform as shown by the remarkable correlations between the Hellenides and the Anatolides and the Taurides (Özgül and Arpat, 1973; Bernoulli and Jenkyns, 1974; Bernoulli et al., 1974; Dürr, 1975; Brunn et al., 1976; Dürr et al., 1978; Channel et al., 1979; Gutnic et al., 1979; Şengör, 1979a, b; Monod, in press). It wedges out in eastern Anatolia in two main digitations (Fig. 1), the Munzur Mountains in the north (Özgül, 1976; Özgül et al., 1978) and the Bitlis—Pötürge crystalline massifs in the south (Yilmaz, 1978) (Fig. 2). These two digitations are separated by an ophiolitic suture zone that extends from north of the Bolkar Mountains in the southwest to Erzincan in the northeast (Fig. 2) that contains mainly early Cretaceous ophiolites, late Cretaceous to middle Eocene flysch and olistostromes and that is unconformably overlain by latest medial Eocene sediments (Demirtaşli et al., 1973). This belt also contains blueschists (Metamorphic Map of Europe, 1/2,500,000. sheet 15) and corresponds to the eastern part of Demirtaşli's (1977) Inner Tauride Belt. To the southeast of this ophiolitic suture, herein called the Inner Tauride suture, are the Malatya—Keban metamorphics (Perinçek, 1979), which are

Fig. 2. Simplified geological map
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ID = Istiranca Daglari, *KaM* =
SD = Sultan Daglari, *KM* = Kirşe
B—HN = Beysehir—Hoyran Napp
Malatya—Keban Metamorphics,
Gölü) and *LV* = Lake Van.

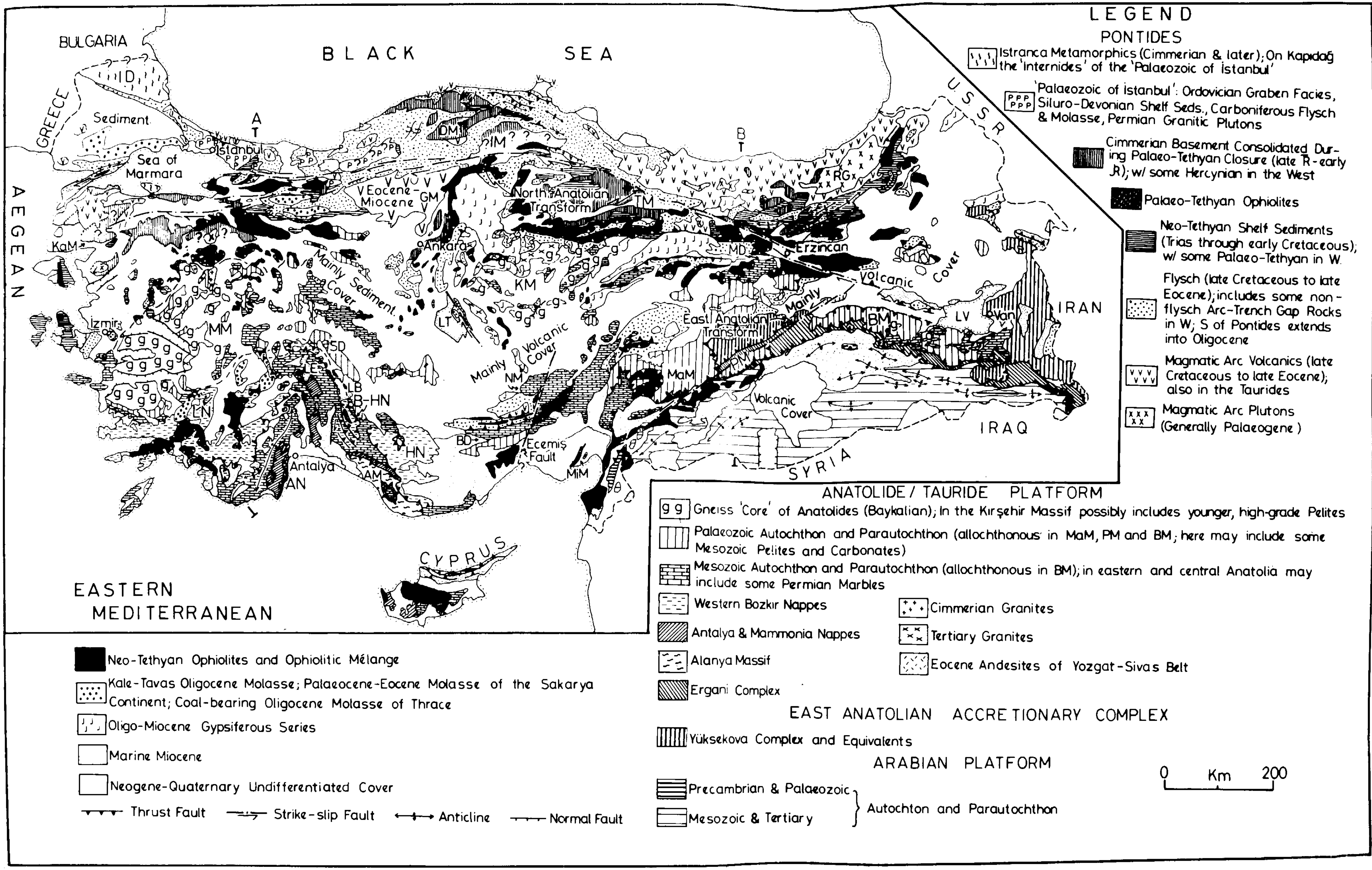
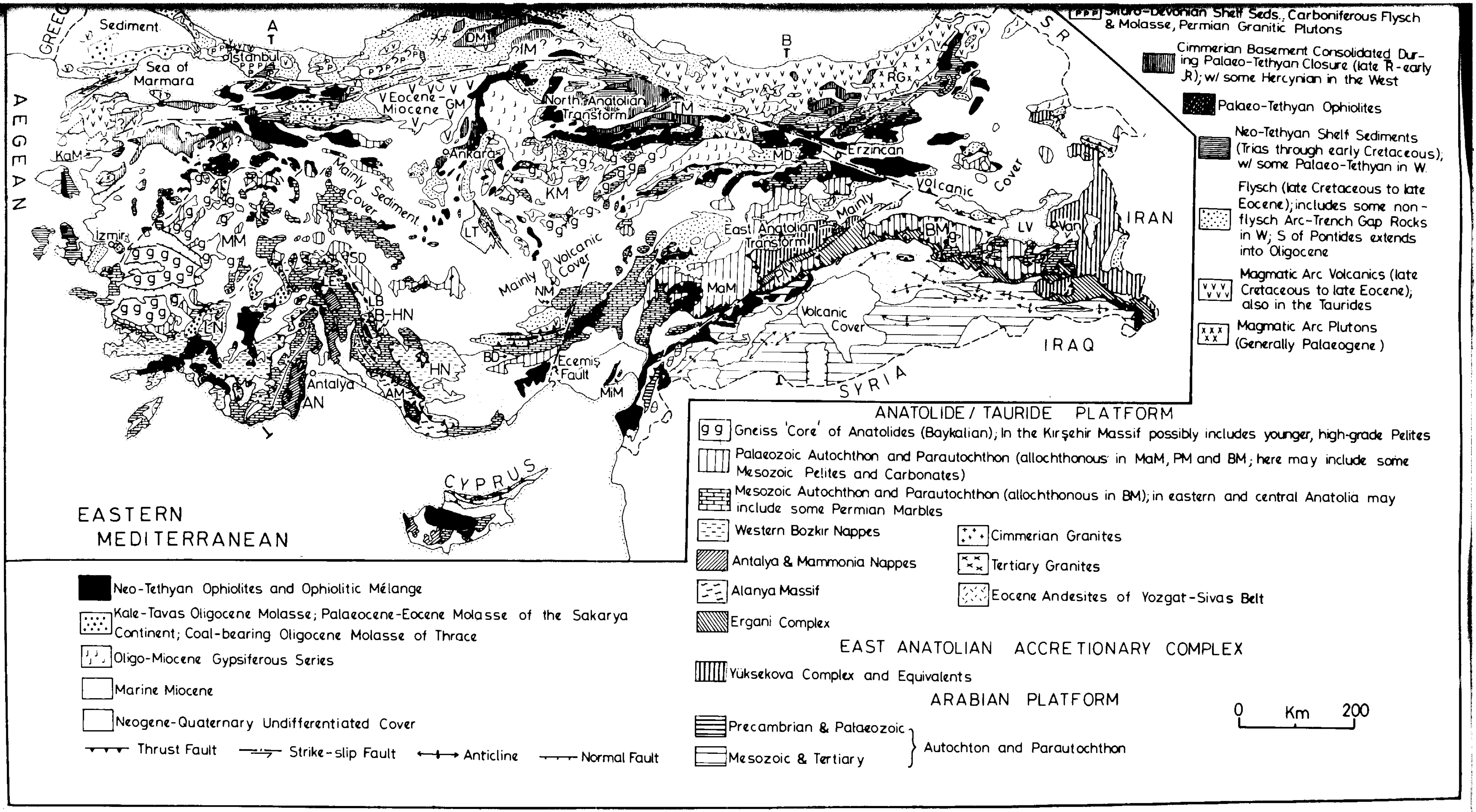


Fig. 2. Simplified geological map of Turkey emphasizing the original palaeo-geographic setting of the various units and de-emphasizing the "Alpine" metamorphism, which, by forming extensive crystalline terrains very much confused the original stratigraphic—structural relationships and for many years misled geological research in Turkey as exemplified by the "Zwischengebirge" interpretation of the Anatolides. In conjunction with our own observations, the map is compiled and somewhat reinterpreted from Anonymous (1979), Bergougnan (1976), Dürr (1975), Geological Map of Cyprus, 1/250,000 (1979), Geological Map of Turkey, 1/500,000 (1961—1964)



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ID = Istiranca Daglari, *KaM* = Kazdag Massif, *DM* = Daday Massif, *IM* = Ilgaz Massif, *TM* = Tokat Massif, *MM* = Menderes Massif, *SD* = Sultan Daglari, *KM* = Kirşehir Massif, *NM* = Nidge Massif, *MD* = Munzur Daglari, *LN* = Lycian Nappes, *AN* = Antalya Nappes, *B—HN* = Beysehir—Hoyran Nappes, *H* = Hadim Nappes, *AM* = Alanya Massif, *BD* = Bolkar Daglari, *Mim* = Misis Mountains, *MaM* = Malatya—Keban Metamorphics, *PM* = Pötürge Massif, *BM* = Bitlis Massif; *LE* = Lake Egridir, *LB* = Lake Beyşehir, *LT* = Salt Lake (Tuz Gölü) and *LV* = Lake Van.

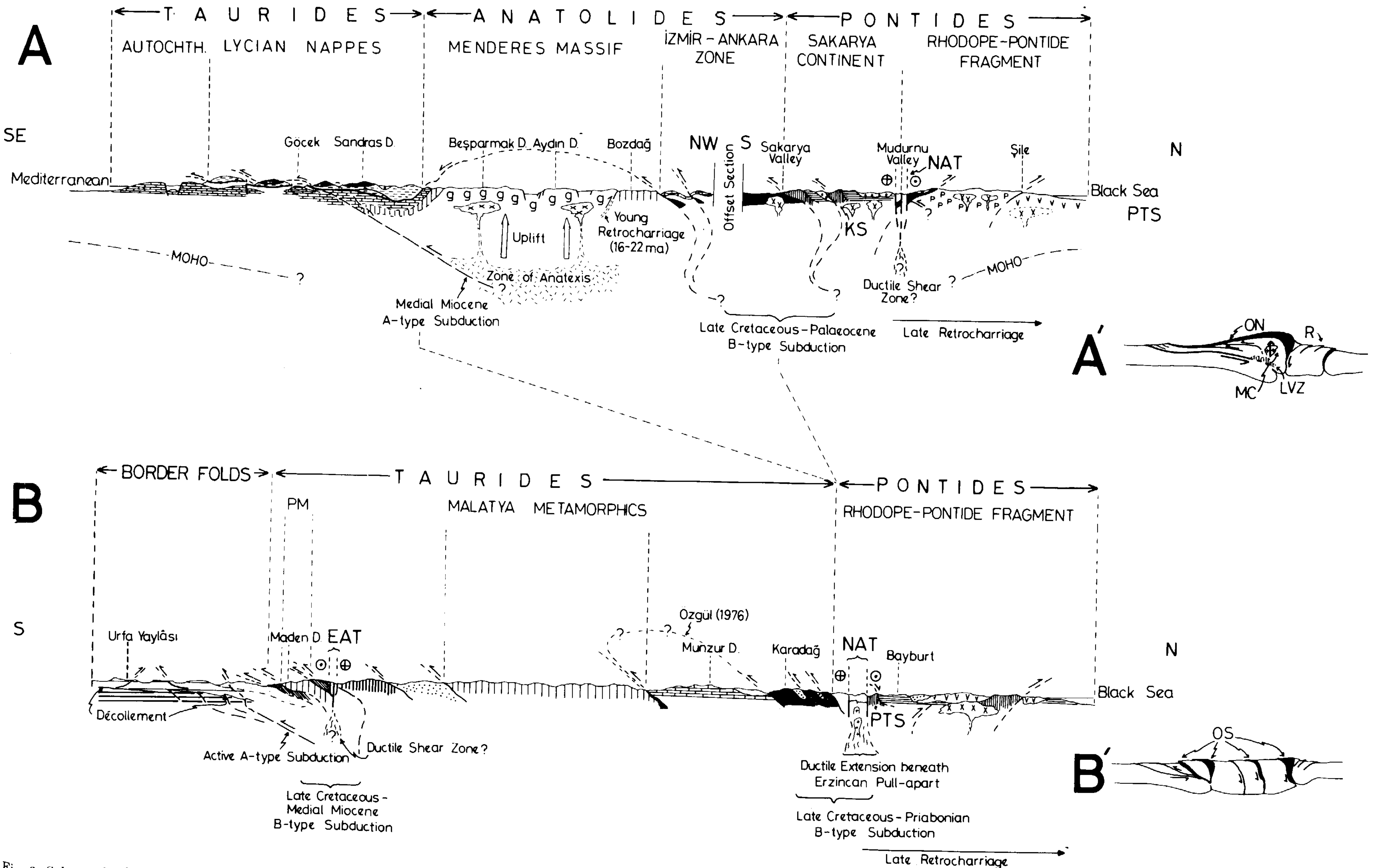
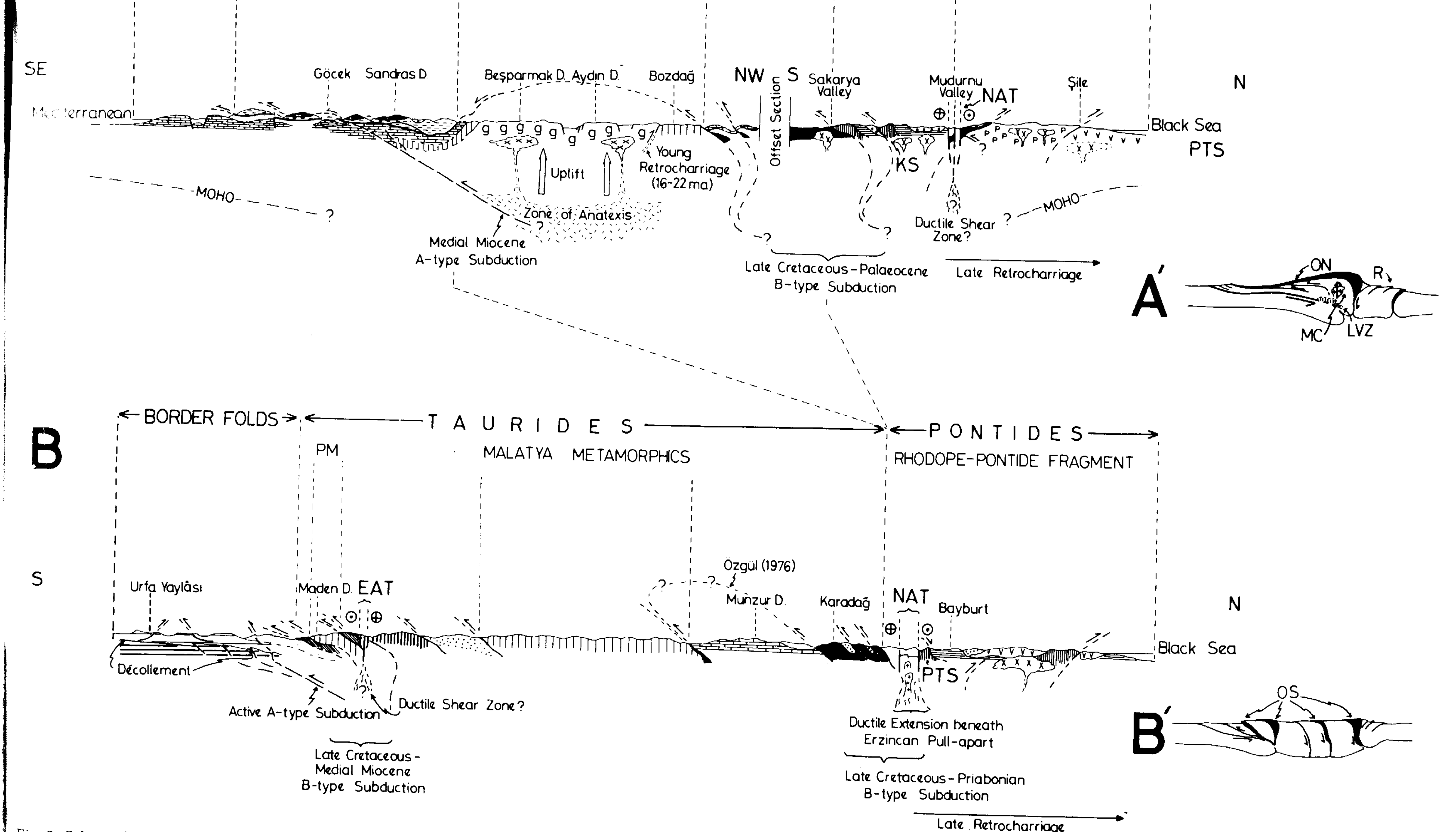


Fig. 3. Schematized geological cross-sections across Turkey, the locations of which are indicated on Fig. 2. The cartoons in A′ and B′ are drawn to emphasize the fundamentally different tectonic styles of the collisional orogen in Turkey along these cross sections (discussion in text). The sections are drawn according to our interpretation of the data reported in the references cited in the caption of Fig. 2, our observations, and Baş (1979), Canitez (1962) and Izdar (personal communication, 1977). Note the tectonic window position of the metamorphic Menderes Massif and its total disappearance eastwards. D stands for mount of mountains. *KS* is the Karakaya



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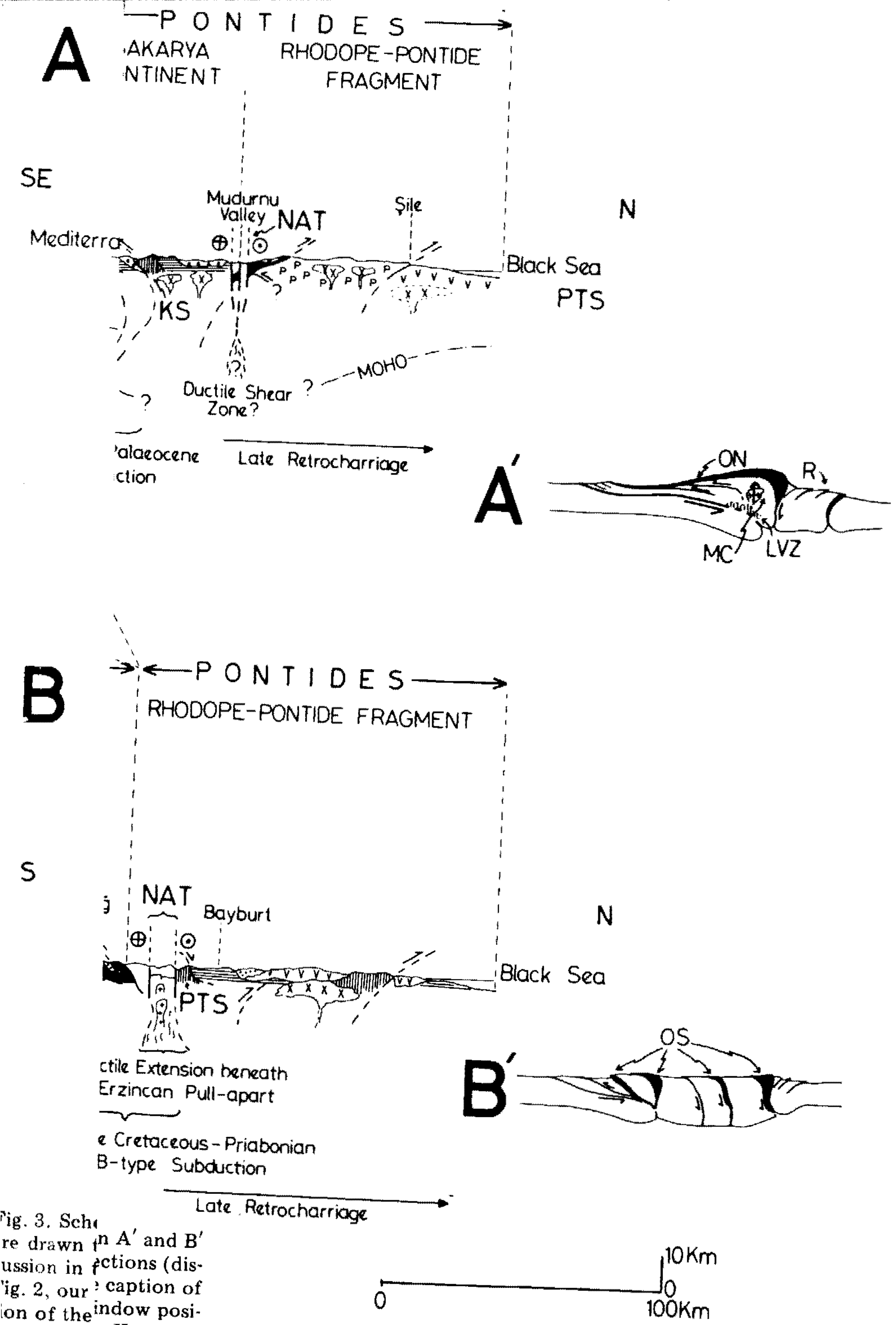

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composed of metamorphosed Permo-Mesozoic carbonates very similar to those that make up the outer envelope of the Bitlis Massif (Yilmaz, 1978). The Malatya—Keban metamorphics are thrust southwards over the oceanic and island-arc lithologies of the Yüksekova Complex, which in turn overthrust the tectonically imbricated lithologies of the Ergani Complex and the Pötürge Massif (Hempton and Savci, in prep.) (Figs. 2 and 3B). Although the detailed relationships in this general area are not yet fully worked out, we believe that the Inner Tauride suture was the original site of much of the Yüksekova oceanic lithologies and the combined Malatya—Keban Pötürge and Bitlis lithologies lay originally to the south of the Inner Tauride ocean. The present distribution of rocks is probably due to the late Eocene and later imbrication and thrust-stacking. Tectonically interleaved within and beneath the Bitlis—Pötürge complex are late Maastrichtian to Oligo-Miocene sediments (the Ergani Complex) that record the distensive development of a deep-sea basin system, the Maden Marginal Sea (Fig. 6E), that disrupted the Bitlis—Pötürge Platform. In the north the Maden Basin sensu stricto achieved its maximum development during the early and medial Eocene, when the widespread pillow lavas, radiolarian cherts and turbidite deposits of the Maden Complex (Perinçek, 1979; Maden Member of Rigo de Righi and Cortesini, 1964; Sason—Baykan Group of Özkaya, 1974; Baykan Complex of Sungurlu, 1974 and Yalçin, 1977), and was tectonized by late Eocene times (Fig. 5, column 16; Fig. 6F). The southern portion, the Çüngüş Basin, however, remained open and closed much later, during the medial Miocene (Figs. 6F—H).

Prior to the opening of these Maden marginal sea complexes (Figs. 6E, F), the southern boundary between the Anatolide—Tauride Platform and Afro-Arabia had been the southern branch of Neo-Tethys (Eastern Mediterranean ocean of Şengör, 1979a). The present eastern Mediterranean represents a remnant of this ocean. In southeastern Turkey and northwestern Syria the large ophiolite nappes of Kizildag and Baer-Bassit (Ricou, 1971; Delaune-Mayere et al., 1977; Yalçin, 1979), Koçali (Sungurlu, 1972; Perinçek, 1979; it would correspond partly to Rigo de Righi and Cortesini's (1964) Kevan nappe), and Cilo (O. Sungurlu, personal communication, 1979, and Y. Yilmaz's own field observations) represent relicts of the closed section of this ocean. These ophiolites were obducted onto the continental margin of the northern Arabian Platform as a result of its collision with the Bitlis—Pötürge Massifs during the late Cretaceous. Almost immediately after this collision event the Maden marginal sea complexes began opening.

Throughout the evolution of Neo-Tethys in the general area of Turkey, the microcontinents constantly changed shape and position as a result of a semi-continuous series of rifting, shearing and collision phenomena. The evolution of the Antalya Nappes and the Alanya Massif (Figs. 2, 6B—E) is a small scale example of the same phenomenon. This makes it difficult to define a number of microcontinents and reconstruct their tectonic evolution through the time period we are interested in, because such microcontinents have not survived that long as individuals — ideally, different microcontinent terminology should have been employed for different times. This becomes particularly clear when we extend back into the early Mesozoic to include the Palaeo-Tethyan elements in reconstructions. In Fig. 1, the Palaeo-Tethyan sutures have not been included to keep the map reasonably simple. They are, however, shown in Figs. 6B and C.

Finally, the eastern Anatolian region in general represents a kind of "suture knot" where the northern and southern Neo-Tethyan, the Zangezur, and the Zagros sutures come together. Until recently, there was very little information available on this part of Turkey. During the last several years, however, considerable amounts of data have accumulated and we have summarized the critical stratigraphic information in our Fig. 4. From this correlation chart, it appears that the basement of eastern Anatolia is largely composed of late Cretaceous and ?older ophiolitic material. Data on pre-late Cretaceous oceanic lithologies are now being gathered (O. Monod, personal communication, 1980) and indicate that the eastern Anatolian oceanic area probably originated sometime during the Jurassic (?older) which is in accordance with the data from the suture zones that converge into it. Ever since Arni (1939) it has been known that the structure of eastern Anatolia consists generally of stacked, steepened, imbricate thrust sheets of coloured mélange (Yüksekova Complex and equivalents in Fig. 2) that includes, other than Mesozoic oceanic rocks, Permian neritic limestone blocks and metamorphic knockers (see Ketin, 1977, for an excellent description of this association). The post-Maastrichtian lithologies of eastern Anatolia comprise an early flysch—olistostrome association of Palaeocene—Eocene age which leads upward to less complete shallow water and finally terrestrial sedimentary sections and locally calc-alkalic volcanics. Intensity of deformation in these sequences decreases with decreasing age. They are often infolded or imbricated into the underlying mélange slices which have generally steep contacts. We interpret the entire basement of eastern Anatolia as an accretionary wedge with at least one ?ensimatic island arc complex (andesitic volcanism shown in Fig. 4, columns 1 and 2; these correspond to the eastern prolongation of the Yozgat—Sivas Belt of Fig. 2, but are not shown there because of the chosen scale) caught up within it. The sediments above the mélange wedge are essentially arc-trench gap and upper-slope basin deposits recording the progressive thickening and progradation of the wedge similar to the Aleutian accretionary prism with its multiple arc-trench gap basins (Dickinson and Seely, 1979, fig. 12) or the largely subaerial Makran accretionary wedge with its large, subaerial forearc basin, the Jaz Murian depression, and the multiple, steep-sided upper-slope basins (Farhoudi and Karig, 1977, figs. 1 and 2). It is interesting to note that Arni (1939) had already pointed out the great similarity between the lithologic associations and tectonic style of eastern Anatolia and those of Makran and included both of these areas into his Iranides. Arni's (1939) cross-sections (plate 4) across his "geschuppte Einheiten mesozoischer und tertiäerer Schichten der Iraniden"

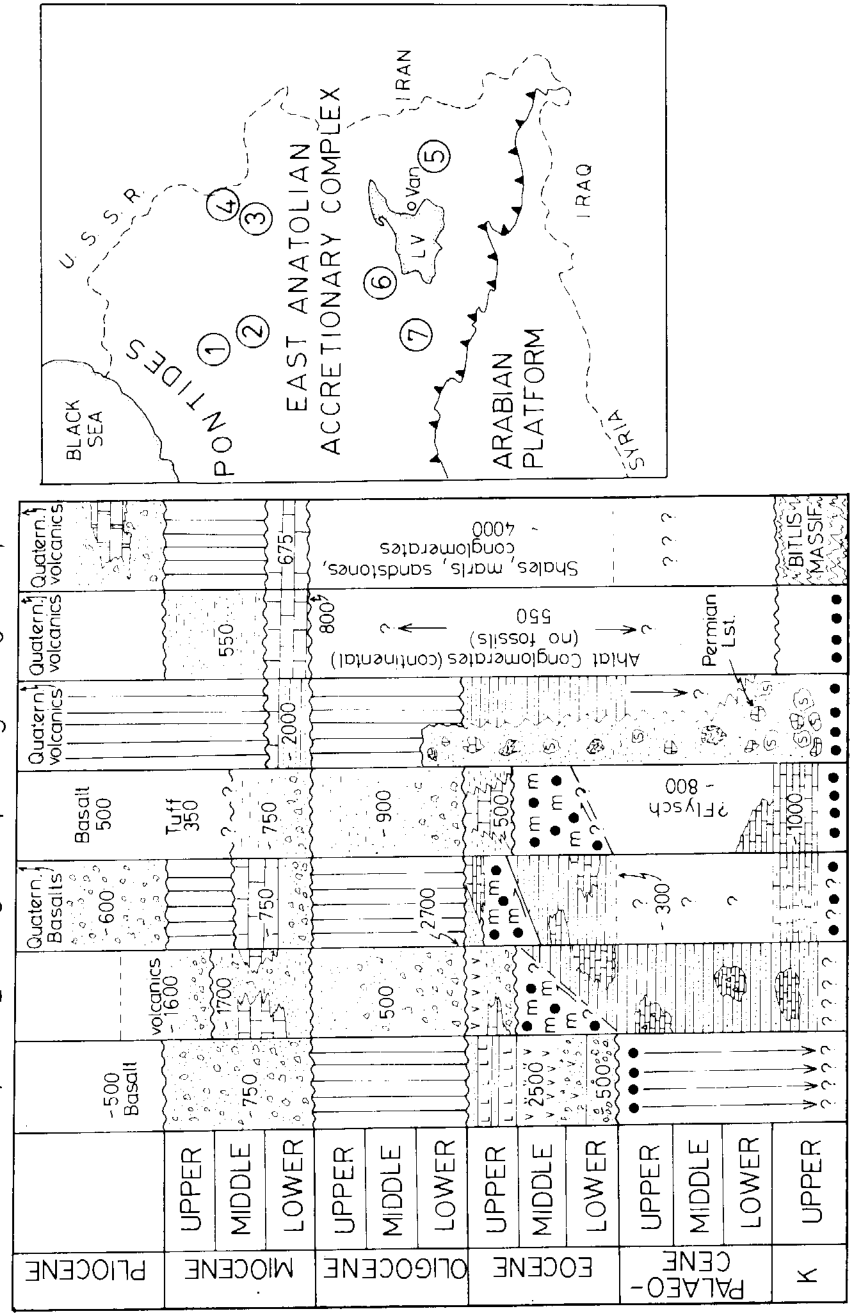
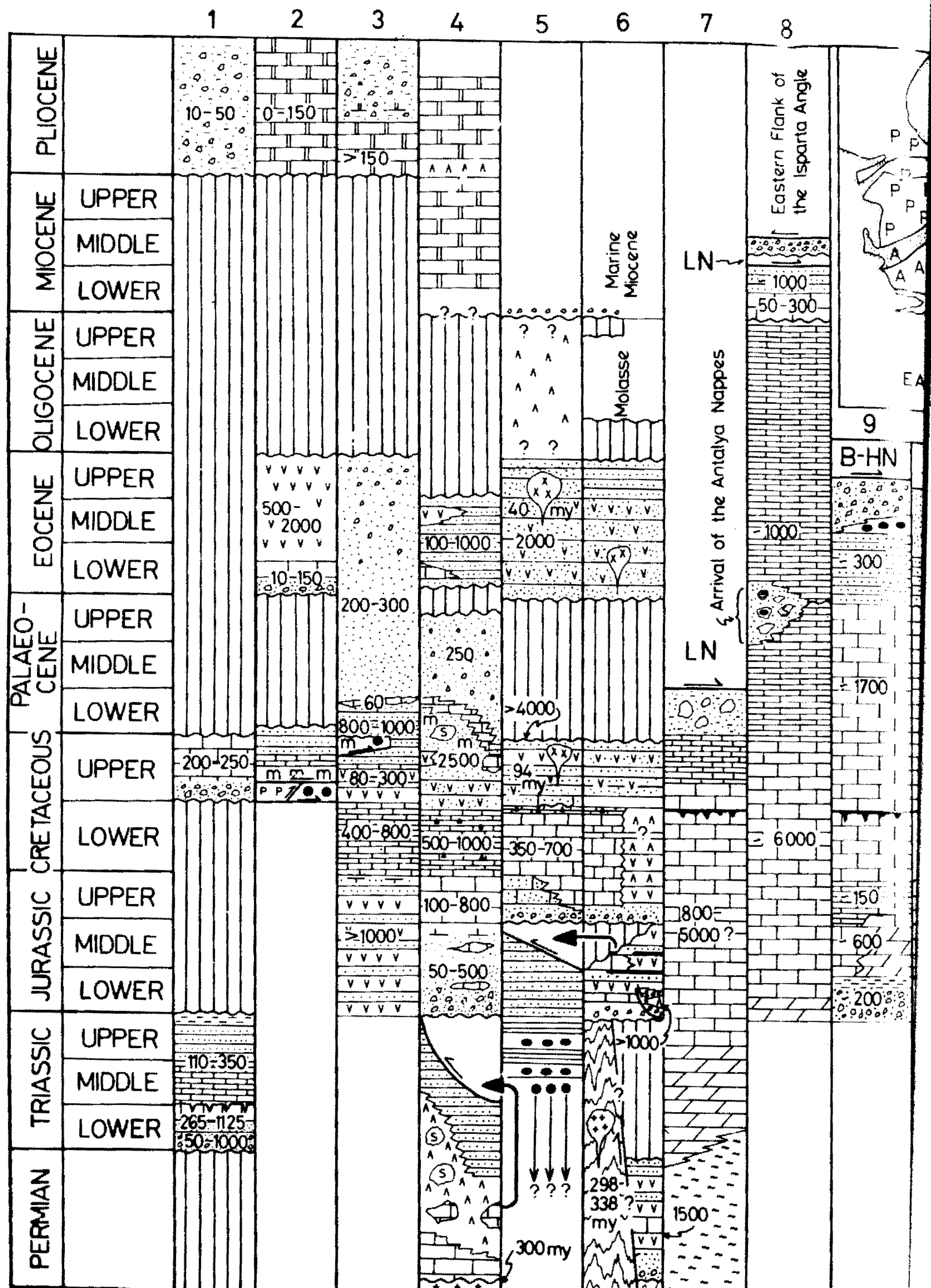
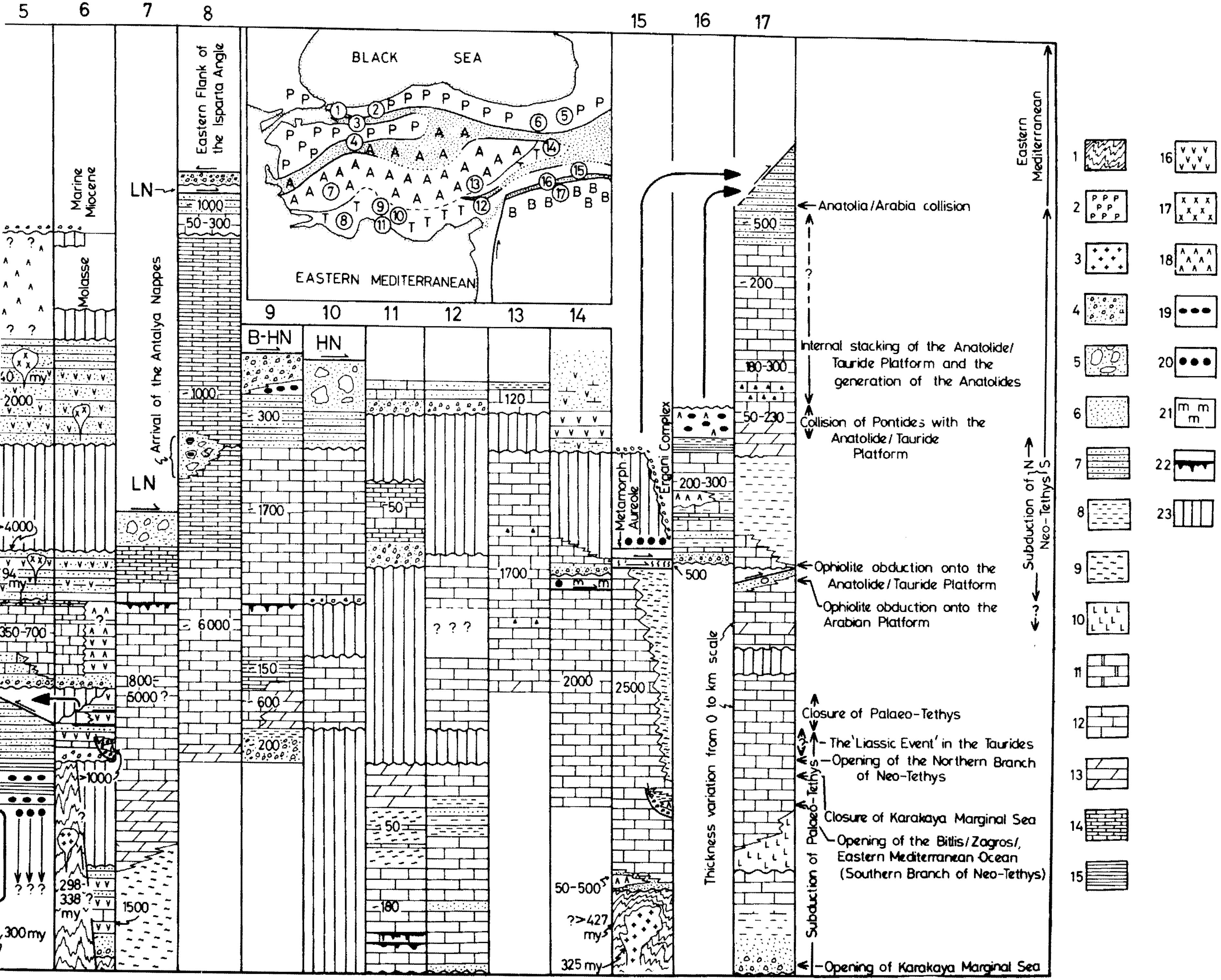


Fig. 4. Correlation chart for seven localities (shown in the map on the right) in the East Anatolian Accretionary Complex. Symbols are the same as in Fig. 5. Data for the columns 1—4 were kindly supplied by O. Sungurlu (written communication, 1980). Column 5 is from Ketin (1977) and Kurtman et al. (1978). Columns 6 and 7 are from Kurtman et al. (1978). Not all of the sedimentary thicknesses are well-controlled and few may actually represent tectonically thickened (most often by thrust-stacking) sections. Note that *all* sections except one (7) have late Cretaceous or older ophiolites for a basement and column 7 may not be in place on top of the continental metamorphics.

have an amazing similarity to those of the rotated portions of modern large accretionary complexes (compare them with the sections in Karig, 1974; Mascle et al., 1977; Farhoudi and Karig, 1977; Dickinson and Seely, 1979).

Fig. 5. Stratigraphic correlation chart showing the typical successions encountered in the autochthonous and parautochthonous regions of Turkey plus the allochthons of south-eastern Turkey. On the right is a summary of major tectonic events in Turkey plotted against the evidence from those areas represented by the columns. The inset shows the approximate locations of the stratigraphic columns within the context of Ketin's (1966) tectonic subdivisions of Turkey (*P* = Pontides, *A* = Anatolides, *T* = Taurides, *B* = Border folds), although they have been constructed so as to reflect the overall traits of the sub-division in which they are located. The references for individual columns are as follows:
1 — Özdemir et al. (1973), Brinkmann (1976), Abdüsselamoglu (1977) and our observations.
2 — Abdüsselamoglu (1959) and unpublished observations by Drs. A.M. Gözübol and Y. Yilmaz.
3 — Altinli (1973a), Saner (1978) and unpublished observations by Drs. A.M. Gözübol and Y. Yilmaz.
4 — Altinli (1973a, b), Bingöl et al. (1973), Çogulu et al., (1965), Erk (1942) and Saner (1978).
5 and 6 — Agrali et al. (1966), Baykal (1952), Bergougnan (1976), Brinkmann (1976), Erguvanli (1950), Gattinger (1956), Gedikoglu (1978), Ketin (1951), M.T.A. (1977) Nebert (1961, 1963), Özsayar (1973) and Şengör et al. (in press). Heavy black lines in the Middle Jurassic section represent coal measures.
7 — Boray (in Özgül, 1976), Dürr (1975), Gutnic et al. (1979).
8 — Gutnic et al. (1979).
9 — Monod (1979).
10 — Özgül (1976).
11 — Argyriadis (1974) and Özgül (1976).
12 — Özgül (1976).
13 — Unpublished Turkish Petroleum Co. section by A. Aziz, M. Meshur and H.S. Serdar (O. Sungurlu, written communication, 1979).
14 — Bergougnan (1975) and Özgül et al. (1978).
15 — Boray (1975), Yilmaz (1978) and Y. Yilmaz's observations. The patternless slice and the slice with horizontal wavy lines beneath the ophiolites represent metamorphic aureole and the cataclastic rocks respectively.
16 — Perinçek (1979) and personal communications by Dr. D. Perinçek and O. Sungurlu (1979).
17 — Rigo de Righi and Cortesini (1964), Sungurlu (1974) and written communication by O. Sungurlu (1979).
Key to legend: *1* = deformed and metamorphosed stratigraphic basement, *2* = the "Palaeozoic of Istanbul," *3* = late Palaeozoic—early Mesozoic granitic intrusives, *4* = conglomerates and breccias of various kinds, *5* = olistostromes, *6* = sandstones of unspecified sedimentologic environment, *7* = flysch, *8* = shales, *9* = pelites and semi-pelites of the Alanya and Bitlis Massifs, *10* = evaporites, *11* = lagoonal and limnic carbonates, *12* = neritic limestones, *13* = dolomites, *14* = pelagic limestone, *15* = radiolarites, *16* = magmatic arc volcanism, *17* = magmatic arc plutonism, *18* = rift volcanism (mixed rift and arc volcanism symbols indicate bimodal, Tibetan volcanism in column 6), *19* = pillowed basalts, *20* = ophiolites, *21* = mélange, in part ophiolitic, *22* = bauxites, *23* = no record. *S* represents serpentinite in the columns; small, open triangles, chert nodules or cherty beds. *LN* = Lycian Nappes, *B—HN* = Beyşehir—Hoyran Nappes, *HN* = Hadim Nappes. Figures give approximate thicknesses.





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BLACK SEA
EASTERN MEDITERRANEAN
Eastern Flank of the Isparta Angle
Marine Miocene
Molasse
Arrival of the Antalya Nappes
LN
B-HN
HN
Metamorph. Aureole
Ergani Complex
Thickness variation from 0 to km scale
Anatolia/Arabia collision
Internal stacking of the Anatolide/Tauride Platform and the generation of the Anatolides
Collision of Pontides with the Anatolide/Tauride Platform
Ophiolite obduction onto the Anatolide/Tauride Platform
Ophiolite obduction onto the Arabian Platform
Closure of Palaeo-Tethys
The 'Liassic Event' in the Taurides
Opening of the Northern Branch of Neo-Tethys
Closure of Karakaya Marginal Sea
Opening of the Bitlis/Zagros/Eastern Mediterranean Ocean (Southern Branch of Neo-Tethys)
Opening of Karakaya Marginal Sea
Eastern Mediterranean
Subduction of N Neo-Tethys S
Subduction of Palaeo-Tethys
40 my
2000
4000
94 my
350-700
300 my
1000
50-300
6000
1800
5000 ?
1500
>1000
298-338 my
300
1700
150
600
200
50
180
120
2000
50-500
?>427 my
325 my
2500
500
200-300
180-300
50-230

In this paper we call the accretionary prism or prism system of eastern Anatolia the East Anatolian Accretionary Complex following Şengör et al. (in press) (Figs. 1 and 4). As we shall see below, the East anatolian Accretionary Complex exercised a very profound influence on the neotectonics of Turkey (Şengör and Kidd, 1979; Şengör, 1980).

As to the positions of the various continental fragments at different times, little can be said for the time being. Existing palaeobiogeographical and palaeomagnetic data on Turkey in general allow us only to make the very crudest estimate concerning the sizes of oceans and locations of continental fragments. The Rhodope—Pontide Fragment and the Sakarya Continent share Eurasian faunas since at least the Lias and contrast sharply from the Lias to the early Tertiary with the Anatolide—Tauride Platform and Arabia, which display southern Tethyan faunas during this time interval (e.g. Bassoullet et al., 1975; Bergougnan, 1975; Fourquin, 1975; Enay, 1976; Gutnic et al., 1979). This probably indicates that the ocean today represented by the Izmir—Ankara—Ilgaz—Erzincan ophiolitic suture was fairly wide. That the Sakarya Continent was never very far away from Eurasia is also supported by preliminary palaeomagnetic data on its palaeolatitudes during the Jurassic (I. Evans, personal communication, 1980). However, it must have been at least a few hundred kilometres away from the Rhodope—Pontide Fragment, because the elimination of the intervening ocean by north-dipping subduction beneath the Rhodope—Pontide Fragment during the late Cretaceous created a well-developed island-arc on top of it.

Neither the faunas (Bassoullet et al., 1975) nor, it appears, palaeomagnetics (Zijderveld and Van der Voo, 1973; Channel et al., 1979) are aware of the existence of the southern branch of Neo-Tethys. This is understandable, because it terminated westwards around Sicily, not far away from Turkey and, as the palaeomagnetic data do not show its existence, its angular opening must not have exceeded the inherent 10° error in the data, which amounts to about a 700-km north—south opening between Arabia and eastern Anatolia. After the late Cretaceous collision between the Bitlis—Pötürge Massifs and the Arabian Platform, faunal exchange between the latter and the Anatolide—Tauride Platform must have been facilitated.

Palaeomagnetic evidence for the successive positions of the Cimmerian Continent in Turkey since the late Palaeozoic is extremely scanty, and palaeobiogeographic data are non-existent. Zijderveld and Van der Voo (1973) concluded that during the Permian the entire present Turkish area belonged to Africa, a conclusion entirely consistent with the geological evidence.

In the following paragraphs we outline the tectonic evolution of Turkey in terms of nine selected time intervals. These intervals have been selected to show the major changes during the tectonic evolution of the country and do not necessarily reflect data resolution. With the available data from Turkey one could draw much more detailed palaeogeographic and palaeotectonic maps for much shorter time intervals. Such an attempt would be far beyond the scope of this paper, whose purpose is simply to present the outlines of the geological development in terms of plate tectonics.

In the palaeotectonic maps the present shapes of the microcontinental blocks have been dotted, following Burchfiel's practice (1980), for reference purposes. Very large amounts of internal strain characterize all of the microcontinental blocks shown in our Fig. 6 and that is why we avoid the term "plate" for such blocks. Large strains generally begin, however, only after continental collision has brought together two continental pieces along a suture zone. As long as a continental block is located in a plate with only oceanic boundaries, little distortion occurs in it. In our palaeotectonic maps we did not indicate the positions of ridges. There are very few data available to do this and unless good transform fault strike control exists (as in Fig. 6A, for example) to constrain spreading directions, they are nearly impossible to use for making unique reconstructions of ridge geometries.

The original shapes of the microcontinents shown in the palaeotectonic maps of Fig. 6 are poorly constrained. We assumed, largely for graphic reasons, an overconservative 40% total across-strike shortening in Turkey since the Permian (excluding the width of subducted oceans) and extended the respective pieces in a north—south direction to restore back that amount of area. The real amount of shortening in Turkey since the Permian, when the amount due to oceanic subduction is taken out, is probably closer to about 60% or more. The amount of post-Miocene east—west shortening that has affected Turkey (Şengör, 1979b, 1980) has not been taken into account at all and left unrestored. As long as oceanic lithosphere is present between the continents the distance by which they are apart has been treated, for practical purposes, as unknown and the oceans in our Fig. 6 are drawn only symbolically. Only in the case of the Neo-Tethyan oceans have we roughly attempted to show relative size.

PERMO-TRIASSIC EVENTS (Fig. 6A)

During the Permian, the entire area of present Turkey constituted a part of the northern margin of Gondwana-Land facing Palaeo-Tethys. Şengör et al. (in prep.) recently summarized the available data on the nature of this margin. In the eastern part of the eastern Pontides *, in the Gümüşhane—Bayburt area a thick marine sequence (approximately 1.5 km) composed of red arkoses, orthoquartzites, and fossiliferous dark limestones with interlayered hornblende—biotite andesites, tuffs and other mainly silicic lavas was deposited during the Permo-?Carboniferous over a metamorphic basement. This basement consists of polyphase-deformed quartz-feldspathic schists, phyllites, and slates (Yilmaz, 1974a). Post-kinematic, low-melting composition granites, such as the Gümüşhane granite pluton, intruded this basement. Whole rock-Pb ages on this pluton range from 298—338 m.y. Şengör et al. (in press) interpreted the Permian andesitic volcanism as products of a south-dipping subduction zone that was consuming Palaeo-Tethyan ocean floor. Whether the intrusion of the Gümüşhane pluton is also related to this or to some older "Hercyanian" regime is as yet uncertain. If the age of the flysch-sequences overlying Palaeo-Tethyan ophiolites in the Küre, Cangal Dagi and Daday regions (Şengör et al., in press) extends down into the Permian as hypothesized by Ketin (in M.T.A., 1962), they may be taken as evidence for Permian flysch deposition on Palaeo-Tethyan oceanic crust. In this region there is documented Triassic flysch. Associated with the tectonized ophiolites are blueschists and eclogites that are of pre-Dogger age. In the rest of the eastern Pontides, Triassic rocks are very sparse, but because of the apparent continuity between the Permian and Jurassic tectonic events, Şengör et al. (in press) assumed that there had been no significant change in the tectonic regime of the eastern Pontides during the Triassic.

* For Turkey's tectonic subdivisions as defined by Ketin (1966), Pontides, Anatolides, Taurides and the Border Folds, see the inset in Fig. 3.

In the northern part of the western Pontides, the Permian saw the end of the late Palaeozoic Hercynian deformations. The coastal regions of the Black Sea were the site of continental red bed deposition (Brinkmann, 1976), and were probably located at some distance to the south of the active Palaeo-Tethyan margin. With the exception of the Orhanlar Greywacke, a late Carboniferous—early Permian marine molasse related to the waning Hercynian deformations and located to the south of the Sea of Marmara (Brinkmann, 1971, 1976), the "porphyroids and the Verrucano of Sandikli" (Parejas, 1943a; Gutnic et al., 1979), and the predominantly clastic Lower Permian section of southeastern Turkey (Rigo de Righi and Cortesini, 1964), the rest of the country was the site of neritic carbonate deposition indicating quiet platform conditions.

During the Triassic this platform ruptured to produce extensional basins in two areas at two different times. During the early Triassic a rifting event began along a line that extends from the Biga Peninsula north of Bursa through Bilecik and Ankara to the Tokat Massif (Bingöl, 1976). It may extend farther east in the direction of Erzincan (O. Tekeli, personal communication), but the present data there are equivocal.

On the Biga Peninsula Bingöl et al. (1973) and Bingöl (1976) described a series of spilitic basalts, mudstones and radiolarites that interfinger with sandstones, siltstones and conglomerates, collectively termed the Karakaya Formation. The blocks within the conglomerates contain pieces of the Permian neritic carbonates. The Karakaya Formation is overlain by medial Triassic sediments (Bingöl et al., 1973). Bingöl (1976) interpreted the early Triassic Karakaya Formation as indicative of rifting of the previously extensive Permian carbonate platform and subsidence of the rift floor. He assumed that the early Triassic rifting here never produced real oceanic crust, because no evidence for the existence of a well-developed ophiolite suite had been found in the Karakaya Formation. Tekeli (in prep.) however, has later pointed out the existence of nearly all the members of the ophiolitic association here and thought the Karakaya basin to have been underlain by real

Fig. 6. Palaeotectonic maps depicting the tectonic evolution of Turkey and some neighbouring regions from the Permian to the present. Maps show oceans (widths are *not* to scale and shown only symbolically; see text; horizontally-ruled = Palaeo—Tethys and dependencies; vertically ruled = Neo-Tethyan oceans), continents (white), predominant lithologic and lithofacies types (widely spaced carbonate pattern = neritic carbonates, closely spaced carbonate pattern = pelagic carbonates, J = evaporites of central and eastern Anatolia, c = coal; open arrows are very generalized sedimentary dispersal directions for clastics (no pattern = Atlantic-type continental margin turbidites, unless otherwise specified, F = flysch) solid black circles are blueschist facies metamorphics, half-filled circle is eclogites, whereas g represents greenschist, an amphibolite facies metamorphics, B = tholeiitic basalt, AB = alkalic basalt, B = unspecified basalt, Tr = trachyte, v = arc volcanics, ++ = arc plutonics, $\wedge$ = Tibetan-type volcanics, s = untectonized ophiolites. Thin lines with hachures on them are Atlantic-type continental margins (normal faults in Fig. 6I), heavier lines with black triangles on them are subduction zones (triangles on the upper plate) and lines with half-arrows are transform faults. Lines with open triangles show major intracontinental thrust faults (triangles on the upper plate), lines with upside-down triangles portray rétrocharriage (triangles on the lower plate), lines with double-headed arrows across them are folds, and the normal fault + conglomerate symbol indicates rifting. Ladder-patterned solid lines are sutures. The main references to the data used for individual time frames are as follows (discussion in text):

A — (Permo-Triassic). Bingöl et al. (1973), Bingöl (1976), Brinkman (1976), Çogulu (1975), Çogulu et al. (1965), Dürr (1975), Eren (1979), Erk (1942), Erol (1953), Gürpinar (1976), Gutnic et al. (1979), Ketin (1951), Lisenbee (1971), Marcoux (in press), M.T.A. (1964), Özgül (1976), Özkocak (1969), Perinçek (1979), Rigo de Righi and Cortesini (1964), Robertson and Woodcock (1979), Saner (1978), Sungurlu (1974), Van der Kaaden (1959, 1966), Yilmaz (1972, 1973, 1974a, b).

B — (Early Jurassic). Adamia et al. (1977), Argyriadis (1974), Baykal (1952), Bergougnan (1975), Fourquin (1975), Gedikoglu (1978), Gutnic et al (1979), Kauffman et al. (1976), Ketin (1951), Monod (1979), M.T.A. (1964), Nebert (1963), Özgül (1976), Rigo de Righi and Cortesini (1964) Şengör et al. (in prep.), Seymen (1975), Wedding (1963), Yilmaz (1972).

C — (Middle Jurassic). Adamia et al. (1977), Agrali et al. (1966) Gutnic et al. (1979), Khain (1975), M.T.A., 1962 and Nebert (1961), Özgül (1976), Rigo de Righi and Cortesini (1964), Seymen (1975), Şengör et al. (in prep.) and O. Yilmaz (1978). (See p. 213.).

D — (Late Jurassic—early Cretaceous) Adamia et al. (1977), Akin (1979), Altinli (1973a, b), Batman (1977), Baykal (1952), Brinkmann (1972, 1976), Gutnic et al. (1979), Ketin (1951), M.T.A. (1962), Özgül (1976), Wedding (1963) and Zankl (1961). (See p. 215.).

E — (Late Cretaceous—Palaeocene). Adamia et al. (1977), Al Maleh (1976), Bergougnan (1975), Bingöl (1976, 1978), Brinkmann (1976), Brunn et al. (1971), Çogulu (1967), Çalapkulu (1978), Demirtasli et al. (1973), Dürr (1975), Fourquin (1975), Gedikoglu (1978), Gutnic et al. (1979), Hall (1976), Lapierre (1975), Letouzey et al. (1977), M.T.A. (1962), Özgül (1976) Özgül et al. (1978), Perinçek (1979), Rigo de Righi and Cortesini (1964), Saner (1977, 1978), Seymen (1975), Sungurlu (1974), Taner (1977), Tekeli (1978), Tokel (1977), Yilmaz (1972, 1978) and the unpublished data both furnished by Drs. N. Görür, D. Perinçek and Messrs E. Arpat, M. Hempton, N. Özgül, G. Savci and O. Sungurlu and collected by ourselves. (See p. 217.).

F — (Early—middle Eocene). Abdüsselamoglu (1959), Ataman (1972), Bergougnan (1975), Cogulu (1975), Demirtasli et al. (1973), Dürr (1975), Gedikoglu (1978), Gutnic et al. (1979), Izdar (1975), Kalafatçioglu and Uysal (1964), Norman (1973), Özgül (1976), Rigo de Righi and Cortesini (1964), Saner (1977), Seymen (1975), Sungurlu (1974), Tokay (1973), Tokel (1977), Vachette et al. (1968). (See p. 222.).

G — (Late Eocene and early Miocene). Bingöl (1976), Dürr (1975) Gutnic et al. (1979), Lüttig and Steffens (1976), Ösgül (1976), Sungurlu (1974) and unpublished data by Dr.

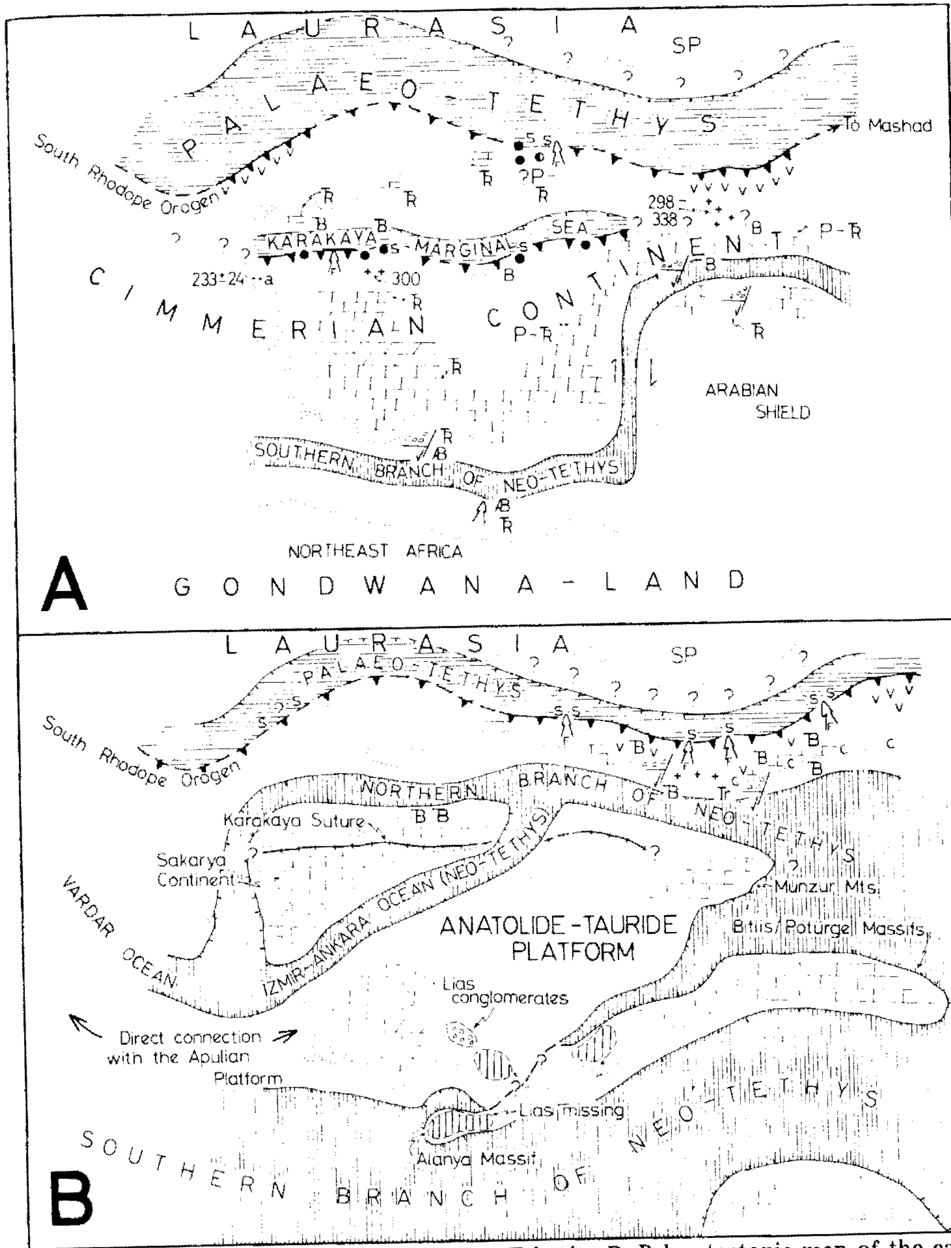

Fig. 6A. Palaeotectonic map of the Permo-Triassic. B. Palaeotectonic map of the early Jurassic.

D. Perinçek, Messrs. M. Hempton, O. Sungurlu and ourselves. (See p. 224.).

H — (Middle Miocene—Pliocene). Lüttig and Steffens (1976), Şengör (1978, 1979b, 1980), Şengör and Dewey (in press), Şengör and Kidd (1979) (See p. 226.).

I — (Pliocene—present). Same as Fig. 6 H (See p. 228.).

oceanic lithosphere. The Upper Triassic in the Karakaya basin is in flysch facies and indicates syntectonic deposition. The basin closed before the beginning of the early Jurassic and the Lias lies unconformably on top of the north-vergent structures of the deformed Karakaya basin with a basal conglomerate (Radelli, 1970; Fourquin, 1975; Bingöl, 1976). The deformation of the Karakaya basin produced a typical ophiolitic mélange involving all the lithologies of the Karakaya Formation, metamorphosed under a variety of conditions producing blueschist, greenschist and high-grade Barrovian assemblages (Tekeli, in prep.). Because of its location behind the Palaeo-Tethyan trench and arc and its relatively short life span, we regard the Karakaya basin as a marginal sea that opened above the Palaeo-Tethyan subduction zone. We have been unable to trace the Karakaya suture either to the west of the Sakarya Continent or to the east of the Tokat Massif. This may be either because of lack of data or the original termination of the basin, perhaps like the present Black Sea.

Throughout southern and southeastern Turkey, as in many other places along the margins of the eastern Mediterranean, there is evidence for a Ladinian—Norian rifting event. In the Antalya Nappes (Delaune-Mayere et al., 1977; Marcoux, in press) the early Triassic has a neritic carbonate environment similar to the Permian platform regime. In the late Anisian, development of breccias with pelagic matrix and Daonellide-containing pelagic limestones with red, manganiferous radiolarites of Ladinian age mark the disintegration and subsidence of the platform. Pietre Verdi (mainly intermediate-silicic tuffs) volcanism at the Anisian—Ladinian boundary (a very common characteristic of the entire Dinaro-Tauric Platform) is followed by a strong, predominantly alkalic basaltic volcanism of late Carnian—early Norian age in the area, which produced abundant pillow-lavas. This abundant late Triassic mafic volcanism is also present in the Mamonia Complex of Cyprus (Lapierre, 1975; Lapierre and Rocci, 1976). As equivalents of the Antalya Nappes are present beneath the allochthonous body of the Alanya Massif (Gutnic et al., 1979), the rifting of the latter from the Anatolide—Tauride Platform must have begun at this time as well, to open at least a part of the basin in which the lithologies of the Antalya Nappes were deposited (*Bassin Pamphylien* of Dumont et al., 1972). In southeastern Turkey evidence of Carnian—Norian rifting and subsidence of the carbonate platform is recorded in a number of places within the lowest tectonic slice of the Bitlis Massif. To the south of Bigra Dagi, for example, submarine basaltic flows interfinger with Megalodont-bearing marbles and pass upwards into red siltstones and cherty limestones. Local development of polygenic breccias here may be indicative of active faulting. (D. Perinçek and O. Sungurlu, personal communication, 1979). Within the Triassic successions of the autochthon of the Border Folds region, i.e. the northern part of the Arabian Platform, Sungurlu (1974) has shown that sedimentation was controlled by active normal faulting.

Elsewhere around the eastern Mediterranean there is other, albeit indirect

evidence for the Triassic rifting of Turkey away from Gondwana-Land. Friedman et al. (1971), Goldberg and Friedman (1974), and Bein and Gvirtzman (1977), have shown that during the late early Jurassic a continental shelf—slope—rise triplet existed along the Levant coast of Israel and they have inferred that the same geometry must have existed also during the early early Jurassic in the area. G.M. Friedman (personal communication, 1978) believes this to be a result of a late Triassic rifting event here. Farther south, in the Sinai area Ginzburg and Gvirtzman (in press) have shown the existence of a similar picture. This and other complementary evidence from the western part of the eastern Mediterranean (W.B.F. Ryan, personal communication, 1979) show that the opening of the eastern Mediterranean probably began during the Carnian—Norian interval with some preliminary events during the latest Anisian—Ladinian (Pietre Verdi time).

Although Dewey et al. (1973) had already inferred the Triassic opening of the eastern Mediterranean ocean, which they had incorrectly connected with the Vardar Ocean, some later workers denied its oceanic nature. The existing equivocal geophysical data have been interpreted by some authors to indicate a continental-type basement beneath the present eastern Mediterranean (Morelli, 1973; Morelli et al., 1975; Lort, 1977; Woodside, 1977). On the other hand, the Orsay group of geologists have primarily developed the concept that *all* the ophiolites in the eastern Mediterranean have their roots in the Izmir—Ankara—Ilgaz—Erzincan—Zagros ophiolitic suture and that they have been thrust over the Anatolide—Tauride Platform during the late Senonian (Ricou et al., 1974, 1975, 1979; Delaune-Mayere et al., 1977; Gutnic et al., 1979). This interpretation considers the eastern Mediterranean as an intra-cratonic rift or a foredeep (Brunn, 1979) that never had any oceanic substratum. Other workers following this view have argued that the Neo-Tethyan evolution of the eastern Mediterranean ranges has been the result of the closure of a single Neo-Tethyan ocean to the north of the Dinaro-Tauric Platform and that the latter has always been an integral part of Afro-Arabia (e.g. Channel and Horvath, 1976; Laubscher and Bernoulli, 1977; Channel et al., 1979). This argument was originally developed primarily to explain the great similarity in the timing of ophiolite emplacement in diverse parts of the eastern Mediterranean orogen (L.-E. Ricou, personal communication, 1978). As a result of recent field work on the ophiolites of southeastern Turkey and northwestern Syria, however, the age of emplacement of these ophiolites became somewhat older (Campanian) than the mainly late Senonian nappes of the Anatolide—Tauride Platform (Al Maleh, 1976). However, the most decisive evidence against a supra-Anatolide—Tauride origin for the Cypriot and southeastern Turkish ophiolites comes from the central part of the Anatolide—Tauride Platform itself. Near Sariz, the geologists of the Turkish Petroleum Company recently discovered a complete stratigraphic sequence from Upper Jurassic to Middle Eocene (Fig. 5, column 13) with no evidence of ophiolite nappes passing over the platform during the late Cretaceous. This automatically puts the roots of the eastern Mediterranean ophiolites south of the Anatolide—Tauride Platform and shows that even if the present eastern Mediterranean does have a "continental" substratum (see Şengör and Monod, 1980), it must have had an oceanic floor at least during the Cretaceous. This resolves the controversy over the origin of the eastern Mediterranean. It is an ocean that began opening during the late Triassic and partially closed, along the Bitlis Suture (Şengör et al., 1979) during the middle Miocene (see below).

The Permo-Triassic geologic record of Turkey records the subduction of the Palaeo-Tethyan ocean floor along a south-dipping subduction zone beneath Turkey. This subduction gave rise to the opening of the Karakaya marginal sea during the early Triassic, which closed shortly thereafter during the late Triassic. The eastern Mediterranean began opening during the Carnian—Norian interval, an event that marks the opening of Neo-Tethys in this area; towards the east the opening continued through the Zagros Ocean (Stöcklin, 1974, 1977) all the way into the Himalayas, separating a northern strip, the Cimmerian Continent (Şengör, 1979a), from Gondwana-Land as Neo-Tethys opened in its wake.

EARLY JURASSIC EVENTS (Fig. 6B)

As far as the Palaeo-Tethyan palaeogeography is concerned, the early Jurassic events represent a continuation of the Permian regime. However, at this time the ocean floor of Palaeo-Tethys was receiving more abundant flysch sediments as they are today seen above the preserved Palaeo-Tethyan ophiolites all the way from the Artvin area near the Russian border (Şengör et al., in prep.) through the Kelkit exposures, to the Küre-Daday region (Ketin, in M.T.A., 1962b; Şengör et al., in press). Because the Palaeo-Tethys terminally closed during the middle Jurassic (Fig. 6C), the increased clastic influx in the early Jurassic may reflect partly the presence of an approaching northern continent, most likely the Scythian Platform. Although in the Lesser Caucasus strong early Jurassic calc-alkalic volcanism was active (Adamia et al., 1977), in the eastern Pontides it assumed a subordinate position with respect to an increasing tholeiitic volcanism (Yilmaz, 1972). Concurrent with this intense tholeiitic activity, rifting began near the volcanic axis of the north-facing Palaeo-Tethyan arc. This could perhaps be interpreted as a result of the subduction of progressively older oceanic lithosphere. While the Laurasian continental margin moved closer to the magmatic arc, the latter became an extensional arc (Dewey, 1980), along the volcanic axis of which marginal basins began to open. Within the initial rift trough very thick conglomerate deposits (>1 km) had intercalated tholeiitic basalts and trachytes (Bergougnan, 1976).

During the Sinemurian a marine transgression began from the south (Akin, 1979) onto the Palaeo-Tethyan arc (Ketin, 1951), which had become an island arc due to the early Jurassic rifting, indicating the beginning establishment of a south-facing Atlantic-type continental margin here. This early Jurassic rifting south of the eastern Pontides is well documented all the way from the east of the Ilgaz Massif to the northeast of Erzincan (Seymen, 1975; Bergougnan, 1976; Tokel, 1977). This is also the time when faunal differentiation between the Pontides and the rest of Turkey began (Bassuolet et al., 1975; Enay, 1976), which also indicates that the Palaeo-Tethys had been reduced to a size that permitted the Laurasian faunas to reach the Pontides. In fact, evidence from the Mashad suture in northern Iran indicates that there the ocean had already closed by the time of the deposition of the Shemshak Formation (?Rhaeto-Liassic) (Stöcklin, 1974, 1977; Majidi, 1978). The equivalent of the Shemshak in the Pontides is the coal-bearing Kelkit Formation of Bergougnan (1976), where it occupies a back-arc setting, along the northern margin of the opening northern branch of Neo-Tethys.

Along the Intra-Pontide suture and the Izmir—Ankara zone the evidence for the Liassic rifting event is less abundant than in the eastern Pontides. South of the Mudurnu area the early Jurassic basaltic volcanics of tholeiitic affinity are overlain by typical shelf deposits of medial to late Jurassic age (Altinli, 1973a) and present a picture very similar to that observed in the eastern Pontides. Fourquin (1975) points out that the Izmir—Ankara ocean also began opening during the Lias (or earlier) as this is the time of faunal differentiation as it was in the eastern Pontides. This widespread early Jurassic rifting event is recorded also in the Vardar Zone and in the rest of the peri-Dinaro-Tauric Platform oceans (Channel et al., 1979) with the exception of its southeastern margin that had already rifted during the Triassic. The opening of this new arm of Neo-Tethys, herein called the northern branch, greatly reduced the width of the Cimmerian Continent in this area and isolated the Dinaro-Tauric Platform, except in Sicily where there has never been an oceanic gap between Africa and the Dinaro-Tauric Platform, as shown by the continuity of the facies belts from Africa to Italy (e.g. Channel and Horvath, 1976, Caire, 1977; Channel et al., 1979).

Unfortunately, we have little evidence on the Anatolide—Tauride Platform rifting away from the Cimmerian Continent during the early Jurassic, because its northern margin has been heavily tectonized during the later collision and the evidenced has been destroyed.

The similarity between the basement of the Bitlis—Pötürge massifs and the eastern Pontides and their latest Palaeozoic sedimentary covers (compare the lowest sections in Fig. 5, columns 4 and 15) lead us to believe that the continental piece that was rifted from the eastern Pontides east of the Munzur Mountains was the Bitlis Massif. The fact that its late Triassic—early Jurassic cover sediments consist mainly of neritic carbonates may be because the Palaeo-Tethyan arc at this time had become an extensional arc with mainly mafic volcanic constructions of low relief that did not supply much clastic material to surrounding areas. As the Bitlis Massif was rotating away from the Pontides it ripped the Pötürge and Malatya—Keban massifs and the Bolkar Mountains away from the rest of the Anatolide—Tauride Platform. The timing of this rifting event that opened the Inner-Tauride ocean is very poorly constrained but sparse evidence of alkalic magmatism at the northeastern end of the rift zone indicates a Jurassic opening (I. Seymen, personal communication., 1980).

Early Jurassic is a time of emergence in many parts of the Central Taurus and has witnessed the deposition of Liassic polygenic conglomerates (Çayir and Üzümdere Formations: Gutnic et al., 1979; Monod, 1979) now found in the autochthon of the eastern flank of the Gulf of Antalya. Argyriadis (1974) has pointed out, (following a joint excursion with N.Özgül) the simularities between the Alanya Massif and the Bolkar Mountains with respect to their stratigraphic content and geological evolution. Both areas particularly display the Liassic emergence as do the autochthonous series just north of the Alanya Massif (Fig. 5, columns 10—12). This suggests the possibility that the Inner-Tauride ocean may have connected with the Pamphylian basin that had already separated the Alanya Massif from the Anatolide—Tauride Platform during the late Triassic. This rifting event may explain the rather abrupt early Jurassic emergence in this localized area and the Çayir and Üzümdere conglomerates can be viewed as correlative deposits generated by the erosion of the uplifted rift shoulders. This hypothesis (N. Özgül, personal communication, 1978) implies that during the Lias the combined Alanya—Bolkar—Malatya—Keban—Pötürge—Bitlis continental fragment began to be separated from the main body of the Anatolide—Tauride Platform and supports Özgül's (1976) grouping of these units under a single heading, his Alanya Unit. It is unfortunately almost impossible to directly test this hypothesis as the critical region between the Bolkar Mountains and the Alanya Massif are buried beneath the Neogene fill of the central Anatolian ovas and the Miocene marine limestones of south—central Turkey (Fig. 2).

In summary, the early Jurassic was a time of continued disintegration of the Cimmerian Continent in Turkey that gave birth to the Anatolide—Tauride Platform and possibly another independent continental fragment, that of Alanya—Bolkar Mountains—Malatya—Keban—Pötürge—Bitlis. At this time the southern branch of the Neo-Tethys presumably continued its growth and the northern branch originated as a Palaeo-Tethyan marginal basin and extended all the way into the present western Mediterranean area.

MEDIAL JURASSIC EVENTS (Fig. 6C)

The most important change that occurred during the middle Jurassic in Turkey (and in a large portion of the rest of the Tethyan domain between the Balkans and China) was the terminal closure of Palaeo-Tethys which resulted in the collision of the Cimmerian Continent with the Scythian Platform (Şengör et al., in press). In the eastern Pontides, the northern areas of the magmatic arc emerged and sedimentation was interrupted (Ketin, 1951; Bergougnan, 1976) while in the southern portions it continued, albeit in a regressive facies, without any serious interruption (Agrali et al., 1966). Flysch deposition on the Palaeo-Tethyan oceanic crust ceased during the late

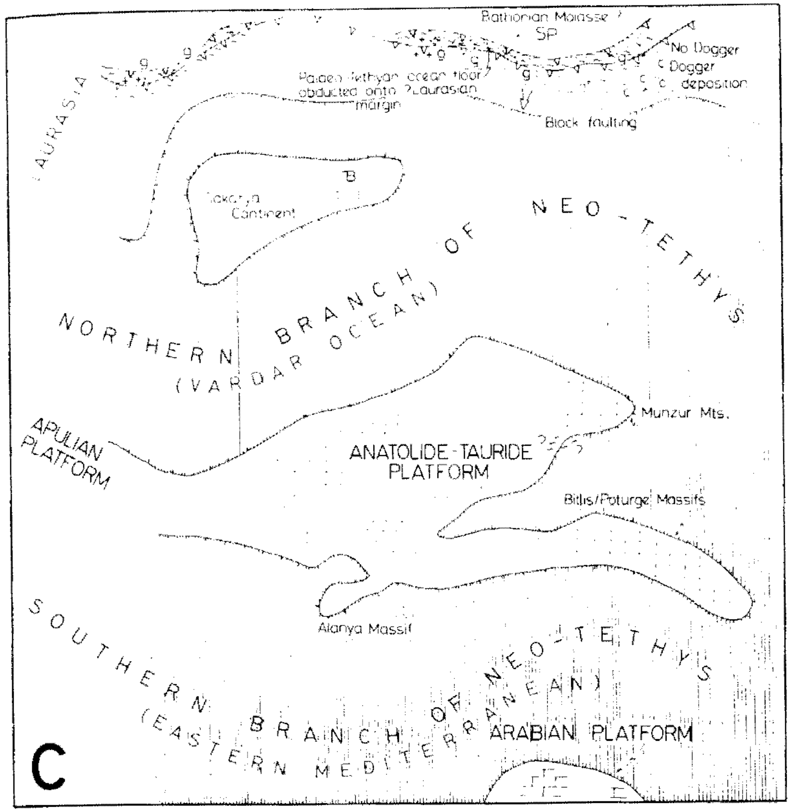

Fig. 6C. Palaeotectonic map of the middle Jurassic. For legend see p. 206.

Lias—early Dogger and very shortly thereafter both the sediments and their ophiolitic substratum were overridden from the south by the continental basement of the eastern Pontides. This event has been interpreted in terms of the collision between the Scythian Platform and the eastern Pontide basement (a part of the Cimmerian Continent) (Şengör et al., in press). The collision gave rise to a north-vergent two-phase penetrative deformation, the second one being accompanied by a greenschist metamorphism in the oceanic lithologies. The Cimmerian Continent itself was not penetratively deformed, except for a wide zone of intense cataclasis and mylonite generation along its basal thrust (Şengör et al., in press). The thrust contacts were later intruded by a variety of plutons that range in composition from granodiorite to tonalite and give isotopic ages around 165 m.y. (Yilmaz, 1979). This event also seems to have reset the K/Ar clocks of the Gümüşhane pluton, which indicate a thermal event around 165 m.y. ago (Çogulu, 1975). Intensive north-vergent deformation, granite intrusion and greenschist metamorphism of probable

Dogger age has obliterated Palaeo-Tethys along the Circum-Rhodope orogen (Kockel et al., 1971; Kaufmann et al., 1976; Roeder, 1978) as well, where Roeder (1978) has identified a north-vergent ophiolite obduction of this age. With the terminal closure of Palaeo-Tethys all oceanic and most marine conditions terminated in the peri-Black Sea regions until the opening of the Black Sea itself during the Cretaceous (Brinkmann, 1974).

In the rest of Turkey carbonate deposition and quiet platform—shelf development dominated the scene. Only to the south of the Munzur Mountains an as yet poorly understood and poorly dated deformational event provided some variety for the otherwise rather monotonous course of events. This deformation is expressed in an angular uncomformity where Upper Cretaceous covers folded Jurassic rocks (I. Seymen, personal communication, 1980). Although the Neo-Tethyan oceans presumably continued to expand, we have no way of knowing how large they were during the middle Jurassic.

LATE JURASSIC—EARLY CRETACEOUS EVENTS (Fig. 6D)

The late Jurassic and the early Cretaceous events in Turkey are intimately connected in time, thus we treat them together.

Following the closure of Palaeo-Tethys during the middle Jurassic, continued convergence led to crustal thickening and Tibetan-type volcanism (i.e., volcanism caused by the partial melting of the lower crust and its fracturing perpendicular to shortening due to collision: see Dewey and Burke, 1973) throughout the eastern part of the eastern Pontides and the entire Caucasus area. In the northern part of the Lesser Caucasus the calc-alkalic volcanism terminated during the late Jurassic and was replaced by localized alkali basalt—trachyte lavas intercalated with salt- and gypsum-bearing terrestrial deposits (collision related rifting? see Şengör et al., in press), which were coeval with ongoing calc-alkalic volcanism in the south (Adamia et al., 1977). A similar setting is shared by the Neocomian bimodal volcanism of the northern Gümüşhane area (Zankl, 1961). The abundant Bajocian—Bathonian granitic plutons that are found both in the Main Ranges of the Greater Caucasus and in the basement of the Trans-Caucasus depression (Khain, 1975) may be the earliest manifestation of the "Tibetization" process in the area.

Throughout the late Jurassic—early Cretaceous carbonate shelf deposition and the growth of the continental rise, mainly by turbidites shooting between reefs (e.g. Seymen, 1975), continued in the eastern Pontides. The area of the Rhodope—Pontide Fragment between western Thrace and Zonguldak appears to have been emergent during this time period, probably because of the Palaeo-Tethyan deformations. Upper Cretaceous sits with angular unconformity and a basal conglomerate on the deformed Triassic lithologies of the Kocaeli (Bithnian) peninsula (Abdüsselamoglu, 1977). In the eastern part of the eastern Pontides an as yet enigmatic "sill event" produced extensive tholeiitic basaltic sills (Lower Basic Series: see Akin, 1979)

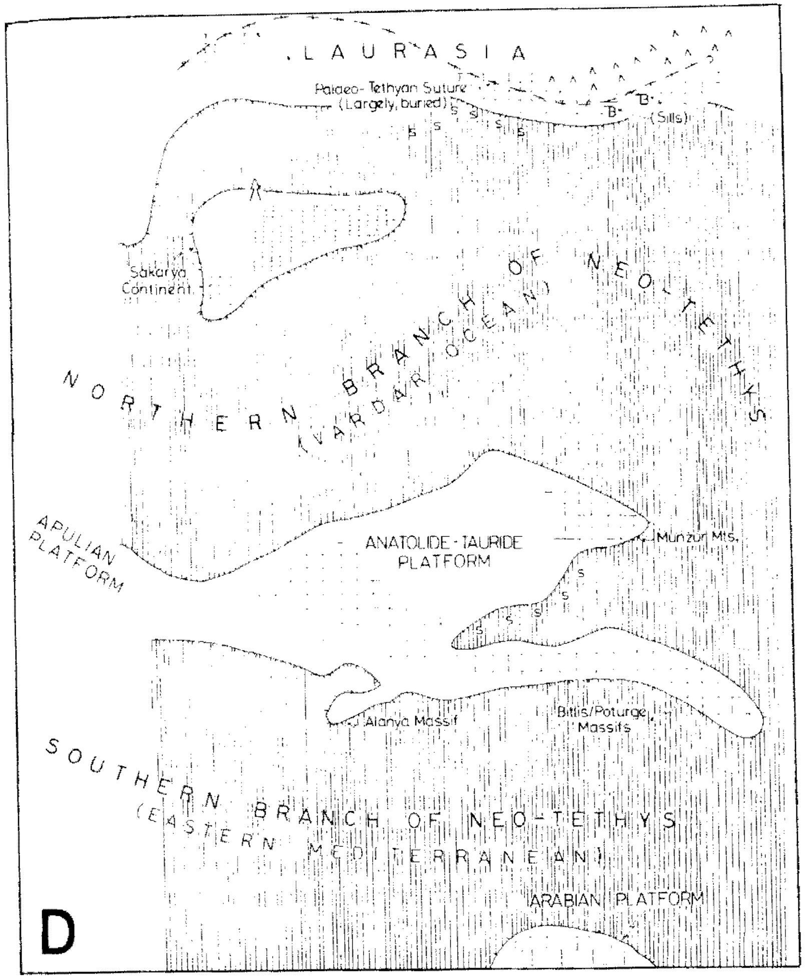

Fig. 6D. Palaeotectonic map of the late Jurassic—early Cretaceous. For legend see p. 206.

that invaded Albian—Aptian rocks. Whether this isolated event marks an initial attempt at starting a subduction zone south of the eastern Pontides is unclear.

The Sakarya Continent was the site of neritic carbonate deposition (Bilecik Limestone: see Altinli, 1973a; Saner, 1977) during the late Jurassic. The conditions changed rather rapidly to a pelagic environment during the early Cretaceous and deeper water cherty carbonates (Soğukçam Limestone, Altinli, 1973a) were laid down. The submarine relief must have increased during this time as evidenced by extensive slump deposits in this pelagic milieu. The reason why the Sakarya Continent suddenly foundered is unclear, but if it can be considered that the westerly extension of the Paikon "Ridge", the isostatic loading of the latter by the Eo-Hellenic ophiolite nappes during the late Jurassic—early Cretaceous obduction event (Mercier et al., 1975; Jacobshagen et al., 1976) may have extended its effects westward onto the Sakarya Continent. Whether this is the case or not, we note here the remarkable temporal correlation between the two events.

On the Anatolide—Tauride Platform also, at least two pelagic domains made their appearance during the late Jurassic—early Cretaceous. Although neritic carbonate sedimentation continued, on most of the platform, pelagic—sediment-filled elongated troughs disrupted its continuity. One such basin may have separated the Menderes Massif from the Bey Daglari autochthonous domain and may represent an easterly continuation of the Ionian trough of the Hellenides (Gutnic et al., 1979). The number, geometry and evolution of these pelagic "troughs" on the Anatolide—Tauride Platform are as yet not clear, however, (see fig. 26 in Gutnic et al., 1979), and that is why we have not included them in our Fig. 6D.

Late Jurassic—early Cretaceous is also the time in which most of the preserved Neo-Tethyan ophiolites of Turkey were produced. This is the same picture as in the Alps and the eastern European Tethyan ranges (Dewey et al., 1973) and similar to the situation there, is accompanied by the pelagization of parts of the previously neritic bordering carbonate platforms. The coincidence of the appearance of ophiolites *in the preserved record* (not necessarily their actual intial appearance) and the rather sudden foundering of parts of the bordering carbonate platforms constitutes one of the current enigmas of tectonics. It is difficult to ascribe this foundering to lithospheric stretching alone because it occurs long after the continental rifting process has ceased and synchronously with the first *direct evidence* of spreading (not necessarily *first spreading*) in the adjacent ocean. Most of the stretching of the continental crust generally occurs during the initial stages of rifting and there is little extension in Atlantic-type continental margins after plate accretion begins (e.g. Montadert et al., 1979). Initial rifting of the Neo-Tethyan oceans in Turkey occurred during the late Triassic—earliest Jurassic, whereas the appearance of the first preserved ophiolites and widespread foundering of previously neritic carbonate platforms were delayed until the late Jurassic and/or early Cretaceous. A thermal explanation is also difficult to devise, because the age-dependent cooling and subsidence curve of Atlantic-type continental margins is close to exponential and most of the subsidence occurs shortly after rifting takes place (Hays and Pitman, 1973; Pitman, 1978). No model as yet satisfactorily explains this delayed rapid subsidence, a situation very similar to the problems encountered in modelling rift-related intra-continental "epeirogenic" basins such as the North Sea (see Burke, in press).

LATE CRETACEOUS—PALAEOCENE EVENTS (Fig. 6E)

During the earliest late Cretaceous clear evidence of subduction activity appears all along the Pontides. According to the Africa—Europe relative mo-

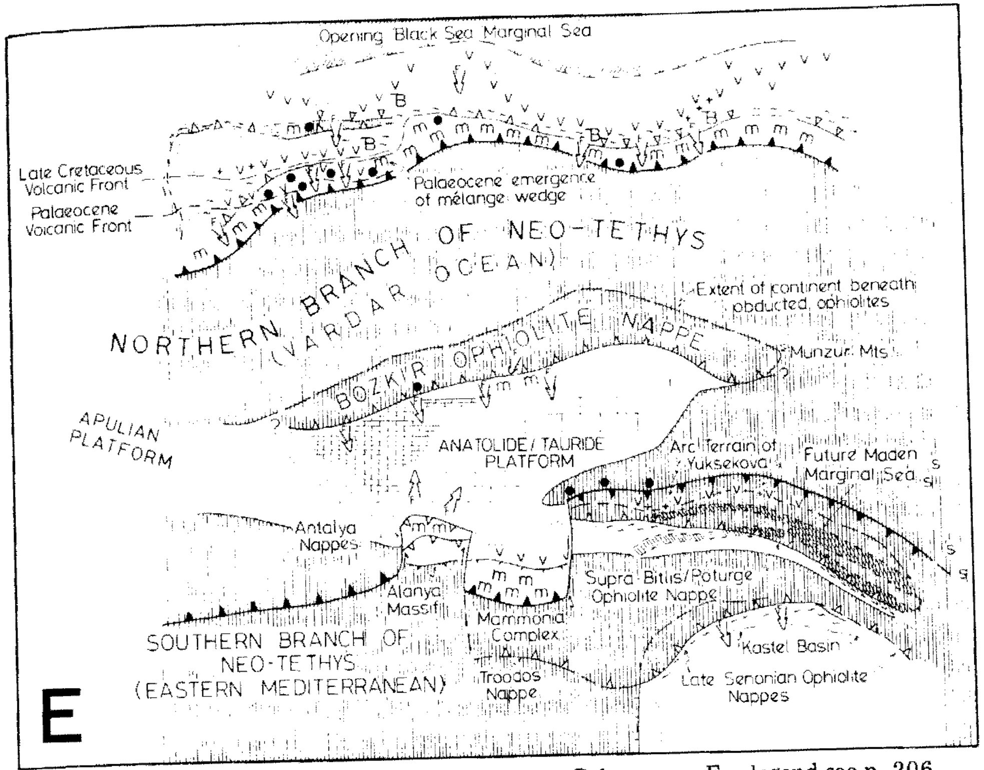

Fig. 6E. Palaeotectonic map of the late Cretaceous—Palaeocene. For legend see p. 206.

tion paths of Pitman and Talwani (1972) and Dewey et al. (1973) there should have been a considerable amount of oblique north—south convergence between Europe and Africa between 148 m.y. ago (middle Kimmeridgian) and 80 m.y. ago (Santonian). Until the Cenomanian—Turonian (90—100 m.y. ago), however, we see almost no evidence of convergence anywhere in Turkey with the possible exception of the localized pre-late Cretaceous folding event south of the Munzur Mountains and the enigmatic early Cretaceous sills (?initial subduction) of the eastern Pontides. According to the Africa—Europe relative motion path of Biju-Duval et al. (1977), on the other hand, the main Africa—Europe convergence began during the latest Cretaceous (76—68 m.y. ago), about 20 m.y. later than the local geological data indicate.

Following a brief episode of overall subsidence and pelagization during the early/late Cretaceous transition (Fig. 5, columns 5 and 6), a strong magmatism involving tholeiitic basalts, rhyolites and extensive tuff deposits began on the eastern Pontides. These volcanics locally sit on the underlying pelagic limestones with a slight unconformity. Later in the Cretaceous, during the Campanian—Maastrichtian, extensive volcanogenic flysch deposits began to accumulate, particularly in the southern parts of the Pontides. The initial intrusions of the Rize granitic pluton began 94 m.y. ago (Gedikoglu, 1978).

The late Cretaceous volcanic axis on the Rhodope—Pontide Fragment in Turkey swung northwestwards to the west of Samsun and closely followed the present Black Sea coast of the western Pontides. In a few places, such as to the north of Istanbul there are onshore outcrops of late Cretaceous arc-related volcanic rocks. By and large, the presently exposed western portion of the Rhodope—Pontide Fragment was in a fore-arc position during the late Cretaceous, in which thick flysch sequences of Turonian—Cenomanian (Cenomanian wildflysch) age (Ketin, 1955) were deposited. Later during the Maastrichtian and the ?Palaeocene the flysch basin shallowed considerably and sedimentation switched to biogenic limestone deposition (Ketin, 1955).

As the magmatic arc was developing on top of the Rhodope—Pontide Fragment a mélange wedge began growing in front of it. It is predominantly of late Cretaceous age, but in places appears to extend into the Palaeocene (Tokay, 1973). Blueschists are seen to be associated with this mélange locally along the strike of the Pontides.

During the late Cretaceous the Black Sea began opening behind the Rhodope—Pontide island arc (Letouzey et al., 1977). This opening event was probably connected with that of the Srednogorie Province of the Balkanides, where pelagic sediments composed of limestones, radiolarites and "couches rouges", turbidites, and submarine volcanics of late Cretaceous age were deposited. Hsü et al. (1977) pointed out the great similarity between the setting and the contents of the Srednogorie province and modern marginal basins and argued that it opened as a marginal basin behind the Rhodope—Pontide Fragment. During the Palaeocene—Eocene the Srednogorie portion of the marginal basin closed and the Emine Flysch was deposited in a foredeep in front of north-vergent thrusts (Hsü et al., 1977), while the Black Sea portion remained open and subsided further, perhaps as a result of the continued cooling of its oceanic substratum (Letouzey et al., 1977). Because of its short lifespan we did not include the Srednogorie portion of the greater Black Sea marginal basin in our reconstruction in Fig. 6E.

On the Sakarya Continent, intensive calc-alkalic volcanism began during the Turonian with the volcanics of the Vezirhan Formation (Altinli, 1973) (Fig. 5, column 4). In the northern part of the continent this volcanism is evidenced by numerous tuff intercalations in the red pelagic limestones of the same formation (Altinli, 1973a). To the south of the Sakarya Continent, an ophiolitic mélange wedge developed between the Turonian and the Campanian and was overlain during the late Maastrichtian—Palaeocene by shallow water sediments in a regressive facies (Saner, 1978). The Intra-Pontide ocean closed sometime between the Palaeocene and the Lutetian, because plant-bearing Lutetian sediments unconformably cover its suture northwest of Ankara (Tokay, 1973).

The Pontides (Rhodope—Pontide Fragment and Sakarya Continent) are one of the few places in the world, where unequivocal evidence exists to show that subduction began at the ocean—continent contact zone and the associated arc has been constructed directly above the sedimentary prism of

the previous Atlantic-type continental margin (Yilmaz, in prep.; see Dewey, 1969a). On the Sakarya Continent the magmatic front marched southwards into the area of the mélange prism during the Palaeocene and migrated back to the north for a brief time during the early Eocene following the closure of the Intra-Pontide ocean.

Throughout the Anatolide—Tauride Platform and the northern portion of the Arabian Platform, the late Cretaceous is a time of extensive, preterminal ophiolite obduction (Ricou, 1971; Ricou et al., 1975). Previously, all these events were thought to have been coeval, but new data indicate that obduction onto the Arabian Platform may have begun somewhat earlier than onto the Anatolide—Tauride Platform. Particularly the western portion of the Anatolide—Tauride Platform began subsiding *en masse* during the Maastrichtian as indicated by the invasion of nearly all previously neritic domains by pelagic sediments (Delaune-Mayere et al., 1977; Gutnic et al., 1979, particularly fig. 26). During the Campanian—Maastrichtian the Bozkir Ophiolite Nappe (named after Özgül's (1976) Bozkir Unit; the "Peridotite Nappe" of the Lycian Nappes: Bernoulli et al., 1974) began climbing onto the Anatolide—Tauride Platform. This is evidenced by the deposition of Maastrichtian flysch and ophiolitic olistostromes on top of the northern continental shelf of the Anatolide—Tauride Platform, which was actively subsiding under the weight of the advancing ophiolite sheet(s) (e.g. the Köycegiz region, De Graciansky, 1972; the Haymana Basin, N. Görür, personal communication, 1980). Sparse blueschist facies metamorphism beneath the advancing thrust sheets, such as those of the Emirdag—Kütakya—Eskişehir—Balikesir belt, where such metamorphics are found in an allochthonous position above the Anatolide—Tauride Platform (Bingöl, 1978) are probably related to the ophiolite emplacement. The obduction of the Bozkir Nappe occurred all along the northern margin of the Anatolide—Tauride Platform during the Campanian—Maastrichtian (Bergougnan, 1975; Dürr, 1975; Ricou et al., 1975; Özgül, 1976; Özgül et al., 1978) and marks the beginning of the Neo-Tethyan deformations here. The Maastrichtian events have not carried the ophiolitic allochthons very far onto the platform, as shown by the continuous platform sedimentation along its central axis from the Menderes Massif—Bey Daglari area in the west to the Sariz region in the east (Fig. 5, columns 7—10, 13), contrary to the interpretations of the Orsay Group (Ricou et al., 1975; Delaune-Mayere et al., 1977; Gutnic et al., 1979), who would like to extend the Bozkir Nappe to Cyprus and southeastern Turkey in the south.

Ophiolite obduction onto the Bitlis—Pötürge continental fragment also occurred at about the same time. On top of the Upper Cretaceous section of the outer envelope, the ophiolite nappe rests above a zone of intense cataclasis (approximately 500 m thick) and a thin ?slice of (0—30 m) metamorphic aureole containing albite—actinolite—epidote—chlorite assemblages (Fig. 5, column 15) ?Palaeocene—Eocene molasse unconformably covers the ophiolites in the western part of the Bitlis Massif (Yilmax, 1978; in prep.). Yilmaz (in prep.) ascribes much of the metamorphism of the Malatya—

Keban, Pötürge and Bitlis massifs to this obduction event. The metamorphism of these massifs is of burial type and may have occurred as a result of the descent of the Bitlis—Pötürge continent into progressively hotter regions as it was tucked beneath the overriding oceanic lithosphere, similar to the situation in the western Fleur de Lys in Newfoundland (Williams, 1977). The ophiolite emplacement in the Aladag region (Tekeli, 1978), along strike from the Bitlis—Pötürge area to the west-southwest, is probably a part of the same event.

Immediately after or synchronously with this obduction event, a south-dipping subduction zone beneath the ophiolite-laden Bitlis—Pötürge continent began consuming the floor of the Inner Tauride Ocean. The evidence for this subduction zone comes largely from the Yüksekova arc lithologies that include pyroclastics, pillow basalts, diorites of various compositions and possibly some andesites (Hempton and Savci, in press). Near Gözeli, to the west of Lake Hazar, the plutonic equivalents of the Yüksekova arc magmatism include large granodiorite plutons that are unconformably overlain by Palaeocene lithologies. In the Bolkardag region to the south of the western extremity of the Inner Tauride ocean, arc magmatism and associated ore mineralization lasted until the late Palaeocene—?early Eocene (Çalapkulu, 1978).

The Alanya Massif overrode the small part of the Pamphylian basin between itself and the Anatolide—Tauride Platform and terminated its sedimentation during the late Cretaceous. N. Özgül (personal communication, 1978) indicates that there might be evidence for an incipient arc magmatism of late Cretaceous—Palaeocene age in the Alanya Massif that could be responsible for its metamorphism, but this remains to be demonstrated.

The southernmost of the late Cretaceous obduction events in Turkey and the eastern part of the eastern Mediterranean was the Senonian ophiolite emplacement that extended all the way from Cyprus to the Turkish—Iraqi border.

In Cyprus the Mammonia Nappes (Lapierre, 1975) probably originated as a subduction-related mélange complex in front of the Kyrenia magmatic arc (Baroz, 1979) and was later thrust over, together with the Troodos complex (Gass, 1968; Biju-Duval et al., 1976) on top of African oceanic or stretched and thinned continental lithosphere. Our Fig. 6E shows the penultimate imbrication stage to emphasize the original north—south thrusting of the Mammonia Nappes as originally proposed by Lapierre (1975) as opposed to the south—north thrusting suggested by Robertson and Woodcock (1979), evidence for which seems flimsier than for southward thrusting. Along the northern periphery of the Arabian shield, obduction may have begun as early as earlier Campanian (Al Maleh, 1976) in the western part, but the bulk of the ophiolites seem to have been emplaced during the late Campanian—early Maastrichtian (Sungurlu, 1974) which is also the age bracket of the Kastel flysch that developed in front of the advancing allochthons. The obduction onto the Arabian Platform may have been a result of the collision between the latter and the Bitlis—Pötürge continent as in several places the massifs and the supposed correlatives of the ophiolite nappes (the Guleman ophiolite, Perinçek, 1979) have a common latest Maastrichtian sedimentary cover. However, the correlation between the Senonian nappes of southeastern Anatolia and the Guleman ophiolite is not yet well established. We adopt the collision hypothesis, because it seems to be the best one currently available.

The opening of the Maden marginal sea complexes probably began during the latest Maastrichtian, i.e. shortly after the south-dipping subduction had been initiated north of the now combined Bitlis—Pötürge and Arabian continent.

The Simaki "Flysch" (Perinçek, 1979), a late Maastrichtian—Palaeocene sequence that begins with coarse clastics on top of the Bitlis—Pötürge metamorphics and Guleman ophiolites and rapidly changes to progressively deepening shallow water limestone lithologies and some turbidites probably mark the initial rifting and beginning subsidence of the Maden marginal sea complexes (Fig. 5, column 16); there is a small amount of volcanics in the Middle Palaeocene section that may have been related to the rifting event. The Campanian—Maastrichtian ophiolites in the east of Lake Van (O. Sungurlu, personal communication, 1979) may represent continued spreading activity in the Neo-Tethyan ocean floor here and imply a complex plate boundary geometry.

The late Cretaceous was a time of revolution in the Neo-Tethyan tectonic development of the Alpides in general and this has been so in Turkey as well. It marked the beginning of the convergent regime here "at all fronts" and was particularly characterized by the emplacement of spectacular ophiolite nappes of large dimensions. These nappes moved onto extensive carbonate platforms that began subsiding *en masse* synchronously with the onset of obduction to become sites of either pelagic or flysch—olistostrome deposition depending on their distance to the advancing ophiolite sheets. Along the southern borders of the Rhodope—Pontide Fragment and the Sakarya Continent ophiolitic mélange was overthrust northwards onto the arc terrains (Bergougnan, 1975; Fourquin, 1975; mostly our observations). The magnitude of this overthrusting is small, however, and nowhere does it involve large, coherent slabs of oceanic lithosphere as is the case on the passive margins to the south. We interpret this thrusting as a rétrocharriage phenomenon at the contact between the arc massif and the accretionary mélange wedge. As the mélange wedge grows oceanward, the wedge-shaped thrust packages rotate to steepen and eventually to overturn (Karig, 1974). This overturning, combined with the increased gravitational potential of the thickened mélange wedge begins driving thrust sheets at the back of the wedge backwards onto the arc massif (Hamilton, 1979). Weber (1978, fig. 8) has recently described a very beautiful fossil example of this phenomenon from the Hohensteiner Sattel from the Rhenohercynian zone near Giessen, W. Germany. This "simultaneous" ophiolite thrusting onto the Pontides and the Anatolide—Tauride Platform during the late Cretaceous and the existence of shallow-water and even terrestrial deposits on thickened mélange wedges during the Palaeocene has been incorrectly interpreted in the past to indicate a terminal closure of the northern branch of Neo-Tethys in Turkey. As we shall see in the next section, however, that closure actually occurred during the latest Palaeocene—early Eocene times.

EARLY TO MIDDLE EOCENE EVENTS (Fig. 6F)

During the ?latest Palaeocene—early Eocene the Anatolide—Tauride Platform collided with the Pontides. Immediately following this collision large-scale internal deformation of the Anatolide—Tauride Platform began that was synchronous with the extensive rétrocharriage development in the Pontides.

Arc magmatism continued throughout the Pontides during the early and medial Eocene as evidenced by extensive calc-alkalic andesitic lavas, pyroclastics and volcanogenic flysch (Fig. 5, columns 4—6), while the deformed Pontide—Anatolide suture zone was overlain by shallow water and terrestrial, largely medial Eocene deposits along the western sector of the suture (Tokay, 1973).

The first evidence of the beginning of the internal imbrication of the Anatolide—Tauride Platform is provided by Paréjas' (1943b) "phase anatolienne" which corresponds to the burial of the central Anatolian crystalline massifs (Menderes and Kirşehir), the Anatolides of Ketin (1966), under the advancing Bozkir Nappes (Lycian, Beyşehir Hoyran and Hadim Nappes plus

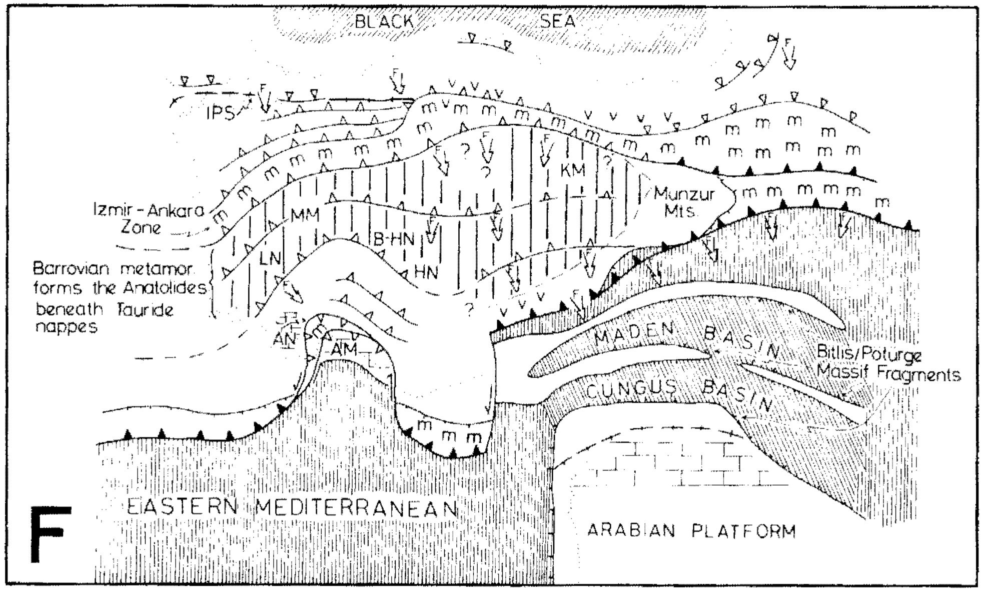

Fig. 6F. Palaeotectonic map of the early—middle Eocene. For legend see p. 206.

the thrust sheets that once lay atop the Kirşehir Massif; Özgül, 1976) (Gutnic et al., 1979). Ypresian and Lutetian sediments everywhere lie unconformably on the structures of this phase, which also corresponds to the beginning metamorphism that eventually formed the Anatolides.

There has been a great diversity of opinion as to the age of the Anatolides. Earlier during this century workers such as Kober (1921) interpreted them as pre-Alpine median massifs ("Zwischengebirge") that had influenced the Alpine development of the Turkish orogen, a view adopted, among others, by Leuchs (1943) and more recently by Brinkmann (1976). Another extreme is represented by Ketin's view (1959, 1966), who considered the late Cretaceous—early Tertiary deformations as the *first* orogenic event that had ever affected this area. Some other workers have taken more compromising views (e.g. Izdar, 1975). Recent work on the Anatolides, particularly on the Menderes Massif (Boray et al., 1973; Dürr, 1975; Dürr et al., 1978; Gutnic et al., 1979), has shown that the massifs have a late Precambrian (Pan African—Baykalian) continental basement on which Palaeozoic, Mesozoic and early Tertiary (Lower Eocene inclusive) sediments were deposited *without any major break*. The same story holds true for the Bitlis and Pötürge massifs until the late Cretaceous ophiolite event. The Anatolides metamorphism as a result of their burial beneath the Boxkir Nappes during the Anatolian Phase (Dürr, 1975; Şengör, 1979b). R. Akkök (personal communication, 1979) has shown that the Menderes Massif was probably buried beneath some 6 km of overburden.

The events in the Kirşehir Massif may have happened somewhat earlier than in the case of the Menderes Massif. Both in its northern (Ketin, 1956) and southern extremities (the Nidge Massif, Göncüoglu, 1977) the metamorphics of the massif are covered by unmetamorphosed early and medial Eocene flysch deposits. Late Eocene andesitic volcanics also sit on the metamorphics of the Kirşehir Massif at various places.

While the main body of the Anatolide—Tauride Platform was being internally imbricated and the initial burial metamorphism of the Anatolides began, the Maden and Çüngüş basins in southeastern Anatolia were about to reach their maximum development. In both of them deep-sea sediments (pelagic limestones and radiolarian cherts) were being deposited accompanied by a strong mafic volcanism mainly in the form of pillow lavas and some turbiditic activity (Perinçek, 1979 and our observations). In the meantime, the Yüksekova arc had ceased its activity during the early Palaeocene and the anti-clockwise rotation of the Bitlis—Pötürge Massif fragments was being accommodated by a north-dipping subduction zone beneath the eastern half of the Anatolide—Tauride Platform. The earliest evidence for this northward subduction comes from the Palaeocene calc-alkalic volcanics of the Ulukişla Basin (Çalapkulu, 1978). The blueschist metamorphics of the Bolkardag area (Fig. 6E) are associated with the late Cretaceous ophiolites and have formed sometime between the Campanian and the Lutetian (Çalapkulu, 1978). The Palaeocene—early Eocene times also witnessed a strong flysch and wildflysch deposition in the Ulukişla area, the western-most portion of the Inner Tauride Ocean, synchronously with the development of the East Anatolian Accretionary Complex to its maximum size. During the same time the neritic Midyat Limestones were being deposited on the southern shelf of the Çüngüş Basin, indicating the return of the quiet shelf conditions to the northern margin of the Arabian Platform (Sungurlu, 1974; Perinçek, 1979).

The Alanya Massif with its cushion of Antalya Nappes was emplaced into the future Isparta Angle (Courbure d'Isparta, Blumenthal, 1963; Monod, 1976) during the late Palaeocene—early Eocene (Brunn et al., 1971; Delaune-Mayere et al., 1977).

LATE EOCENE TO EARLY MIOCENE EVENTS (Fig. 6G)

During the late Eocene to early Miocene interval the general north—south tightening of the Turkish orogen continued while the Anatolides were uplifted and unroofed. Late Eocene in the east (Kirşehir) and Oligocene in the west (Menderes) cover most of the area of the crystalline massifs. However, as the tightening continued and the Bozkir Nappes gradually marched into their final resting places more and more continental material was being fed beneath the massifs, which helped their uplift and gave rise to deep crustal melting under the over-thickened sectors. This anatectic partial melting caused the later widespread silicic volcanism throughout the western Anatolia and granitic plutonism in the Aegean islands (Izdar, 1975; Bingöl, 1976; Dürr et al., 1978). During the late Eocene—Oligocene the Beyşehir—Hoyran and the Hadim composite nappe systems reached their final destinations. This was accompanied by the southerly-imbrication of the autochthonous series

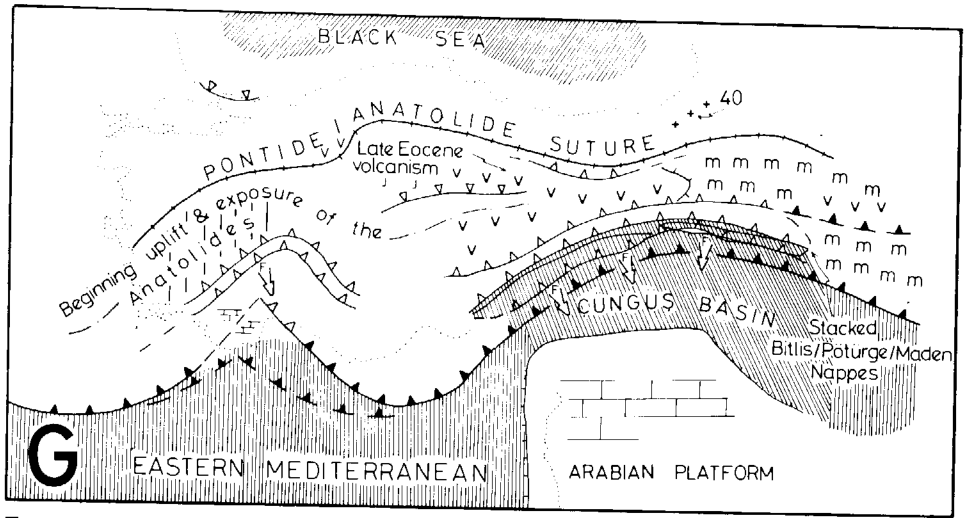

Fig. 6G. Palaeotectonic map of the late Eocene and early Miocene. For legend see p. 206.

of the eastern flank of the Isparta Angle (Monod, 1979). Before the deposition of the Oligocene molasse in the Kale—Tavas Basin (Fig. 2; Becker-Platen, 1970), the Lycian Nappes also advanced onto the autochthonous Bey Daglari region during the Kale Phase of Poisson (in Gutnic et al., 1979). The Oligocene Kale—Tavas molasse seals the thrust contacts between the Lycian nappes and the unroofed Menderes Massif that has a tectonic window position in western Turkey (Dürr, 1975; Gutnic et al., 1979; Şengör, 1979b).

In east-central and southeastern Anatolia the late Eocene saw the closure of the Inner Tauride ocean and the Maden basin (Çalapkulu, 1978; Perinçek, 1979). This event internally imbricated and interleaved the Yüksekova, Maden and the Bitlis—Pötürge lithologies and the Eocene flysch fill of the Inner Tauride ocean. The separation of the Malatya—Keban metamorphics into a separate slice from the Bitlis—Pötürge metamorphics was probably a manifestation of this collision event which also deformed the western half of the East Anatolian Accretionary Complex and essentially terminated its growth, perhaps with the exception of the area to the southeast of the present Lake Van. Immediately following this collision, the Çüngüş basin began receiving large amounts of flysch sediments and olistostromes with blocks of mainly Maden and Bitlis—Pötürge lithologies. Sandstone blocks in the Çüngüş olistostromes may well represent eroded pieces of reworked Inner Tauride flysch.

Either synchronously with or just before the terminal closure of the Inner-Tauride and the Maden basins a flare-up of andesitic volcanism occurred along a wide belt from Yozgat to Kars (Yozgat—Sivas volcanics of Fig. 2 are the western representatives of this event). We associate this volcanism with the subduction of the Inner Tauride ocean beneath the region of volcanism. The Palaeocene volcanism of the Ulukişla area (Çalapkulu, 1978) is here considered as the initial appearance of the same event, which prematurely terminated as a result of the terminal closure of the Inner Tauride ocean.

Following this closure the Africa—Eurasia convergence began to be taken up along a sinuous, uniformly north-dipping subduction zone beneath southern Turkey. This subduction zone was consuming very young ?oceanic lithosphere of the Çüngüş basin along its eastern extent whereas south of central and western Anatolia much older (?middle Mesozoic) lithosphere was being subducted. This geometry might well be responsible for the Eocene initiation of the left-lateral Ecemiş Fault (Fig. 2) that accentuated the sinuosity of the subduction zone by acting as a trench—trench transform separating the extensional (western) segment of the arc from the compressional (eastern) one — we have not indicated this geometry in Fig. 6G, as the details to confirm or reject this hypothesis are not available to us.

Within the Anatolides, continued uplift under ongoing north—south shortening resulted in some rétrocharriage development that is evidenced in the Menderes Massif by ductile, south-dipping shear zones of Miocene age (E. Izdar, personal communication, 1977), whereas in the Kirşehir Massif, near Kaman, cold basement slices overthrust Eocene and some Neogene sediments with a northerly vergence. (I. Seymen and F. Oktay, personal communication, 1977).

The Rize Pluton in the eastern Pontides completed its intrusion during the late Eocene (Çogulu, 1975).

EARLY TO MIDDLE MIOCENE EVENTS (Fig. 6H)

During the early to middle Miocene interval the Çüngüş basin terminally closed and the Arabia—Eurasia collision started along the Bitlis Suture while the Lice molasse (Sungurlu, 1974; Perinçek, 1979) was being deposited in the foredeep that formed on the Arabian Platform as a result of this collision. Very shortly thereafter terminal collision also occurred along the Zagros suture, and Arabia became weled to Eurasia along the Bitlis-Zagros suture zone (Dewey et al., 1973; Şengör, 1979b; Şengör and Kidd, 1979, Şengör et al., 1979). This collision had very profound effects on the overall tectonics of Turkey and has essentially governed its post-Seravallian evolution. As the north—south shortening across eastern Turkey continued between the converging jaws of Eurasia and Arabia the relatively soft and irresistant East Anatolian Accretionary Complex took up much of the initial post-collision convergence by shortening and thickening. However, the rapidly rising elevations made it eventually more economic to wedge out of the way a considerable piece of Turkey, roughly coincident, particularly in the east, with the original outlines of the Anatolide/Tauride platform, onto the easily subductable eastern Mediterranean floor (McKenzie, 1972; Şengör, 1979b; 1980; Şengör and Kidd, 1979). Thus, the North and the East Anatolian transform faults, and with them the Anatolian Plate, originated (Şengör, 1979b).

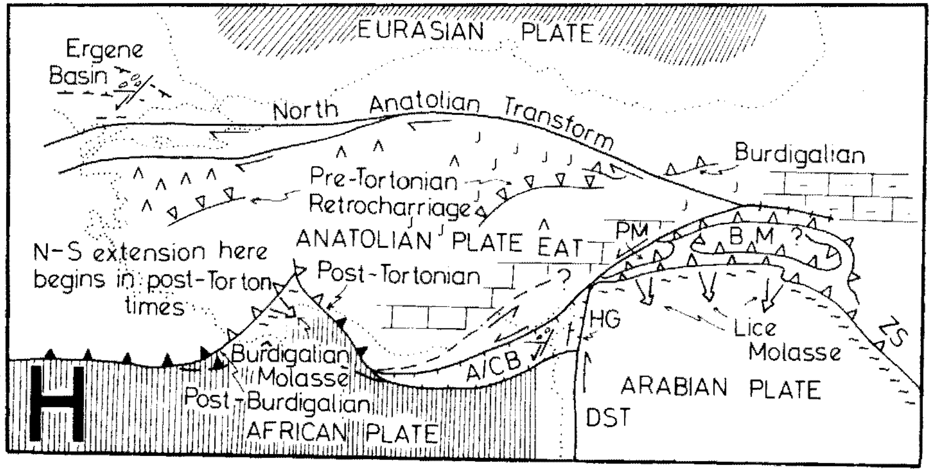

Fig. 6H. Palaeotectonic map of the middle Miocene—Pliocene. For legend see p. 206.

Early Miocene was also the time of the final emplacement of the Lycian Nappes, which overrode a Burdigalian molasse. The associated thrust contacts are sealed by Tortonian conglomerates (Altinli, 1945; Delaune-Mayere et al., 1977). As the fill of the Kale—Tavas basin seals the Oligocene thrust contacts at the back of the Lycian Nappes, the Miocene thrusts must pass below this seal and probably represent a new crustal imbrication of the western part of the Anatolide—Tauride Platform with a minimum thrust transport of 100 km (Gutnic et al., 1979).

The Adana—Cilicia basin, an intracontinental basin that formed as a result of the incompatibility problems arising in a continental FFF triple junction near Maraş (Şengör et al., in press) also formed during this time. The East Anatolian and the Miocene Dead Sea transform faults meet at this triple junction, and because not all of the left-lateral offset can be accommodated by shortening across the Bitlis suture zone, the dagger-shaped Adana—Cilicia basin has to extend its area in a roughly east—west direction.

During the late middle Miocene (Tortonian) the eastern flank of the Isparta angle overrode the western flank at the northern tip of the structure (Delaune-Mayere et al., 1977). Şengör and Dewey (in press) interpreted this tightening of the Isparta Angle as a result of the westward movement of the Anatolian plate. The early Pliocene sediments unconformably overlie this thrust (the Aksu thrust). Since that time the consuming and transform plate boundaries in the eastern Mediterranean appear to have assumed their present geometry.

The Aegean extensional regime also began at the same time due to a relief of the east—west shortening caused by the abrupt south-westerly bend in the North Anatolian transform fault west of the Sea of Marmara by north—south extension (Dewey and Şengör, 1979; Şengör and Dewey, in press).

PLIOCENE TO PRESENT (Fig. 6I)

Continued convergence and thickening in the Turkish—Iranian high plateau resulted in the widespread Plio-Quaternary Tibetan-type volcanism on the plateau (Şengör and Kidd, 1979), while the Border Folds formed on the Arabian Platform as a foreland fold-thrust belt associated with the Miocene collision along the Bitlis—Zagros suture (Ketin, 1966). This collision also gave rise to two impactogens on the Arabian Platform, namely the Akcakale Graben (O. Sungurlu, personal communication, 1979) and the Karacalidag shield volcano (Şengör et al., 1978). While the Aegean extensional regime continued its activity, the strike-slip dominated ova regime became established in central Anatolia (Şengör, 1979b, 1980). The active tectonics of Turkey represents the continuation of the Plio-Quaternary regime and has been described in detail in the recent papers by Şengör (1978, 1979b, 1980), Dewey and Şengör (1979) and Şengör and Kidd (1979). A detailed discussion of the post-Tortonian tectonics of Turkey is therefore omitted in this paper.

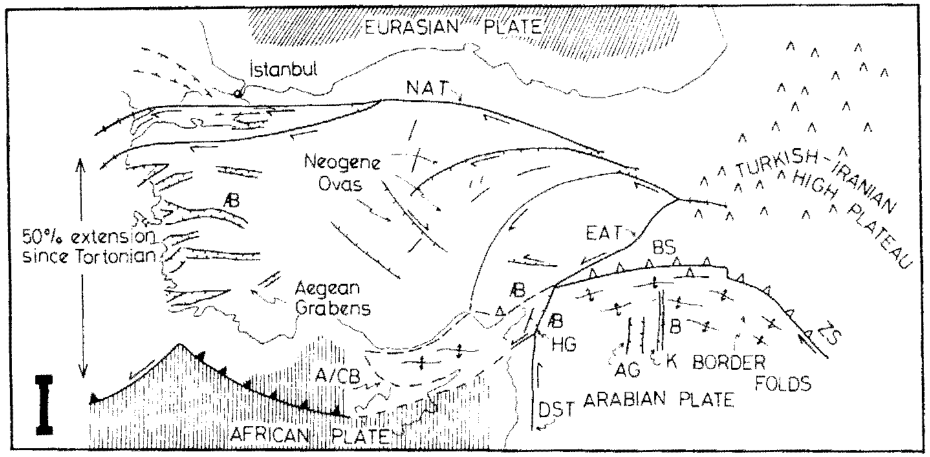

Fig. 61. Palaeotectonic map of the Pliocene—Present. For legend see p. 206.

DISCUSSION AND CONCLUSIONS

We have traced the outlines of the plate tectonic evolution of Turkey and its immediately neighbouring areas since the Permian. This evolution has been, in the simplest terms, the result of the destruction of Palaeo-Tethys and the birth and the subsequent demise of Neo-Tethys. In detail, however, the sequence of events has been much more complicated, characterized by a number of mini-oceans that have repeatedly opened and closed throughout the time interval considered. What is particularly instructive is that although the very broadest outlines of Tethyan geology can be deduced from the evolution of the world plate mosaic, its details are nearly impossible to deduce from the motions of the bounding mega-continents of Laurasia and Gondwana-Land. For example, when head-on convergence between Africa and Europe was going on during the middle Eocene (44 m.y. ago; Biju-Duval et al., 1977) the Maden and the Çüngüş basins were experiencing the zenith of their extensional history. Similarly, as the two continents were receding with respect to each other during the early and middle Jurassic, the Pontides were experiencing intense subduction and later collision tectonics. Throughout the Tethyan history plate boundaries were constantly shifting their position and often changing character, particularly during the late Cretaceous—late Eocene interval (Figs. 6E—G), when subduction zones were constantly "jumping around" in Turkey to continually accommodate the ongoing convergence. Some of these jumps, for example that which occurred during the late Eocene and moved the subduction activity from the northern margin of the Inner-Tauride ocean to the northern margin of the Çüngüş Basin resulted from continental collisions, whereas others, such as the switch from the late Cretaceous—Palaeocene south-dipping subduction along the southern margin of the Inner Tauride ocean, to Palaeocene—late Eocene north-dipping subduction along its northern margin with coeval "quitting" of the former, occurred for less obvious reasons. Modern examples of such "quitting" subduction zones are also present, the best known example being perhaps the subduction zone of the Palawan Trench in the Phillipines that quit during the Miocene (Hamilton, 1979), and are probably the consequence of local plate boundary reorganization. As a result of this rapid change in plate boundary position and geometry, the life histories of the platelets bounded by them become extremely complicated. In many places the geologic record indicates decidedly episodic deformation although the overall displacement between plates was more or less continuous as a result of such abandonment and re-establishment of plate boundaries, particularly of subduction zones. Dewey (1975) has shown the amazing level of complexity in tectonic evolution that can be generated as a result of the continuous evolution of only three plates. In regions where more plates interact, the continuous shifting of the instantaneous rotation poles could create bewildering complexities, the record of which may be completely obliterated by subduction or by deformation during a subsequent continental collision. During the Permian to present tectonic evolution of Turkey we have observed a series of events that rifted, combined and re-rifted, along different lines, the microcontinents within the Tethyan domain. Some of these "microcontinents" were of enormous size and extended well-beyond the boundaries of Turkey, such as the Cimmerian Continent, which extends from northern Greece and Bulgaria to the east of Tibet (Şengör, 1979a). Others were much smaller such as the Alanya Massif and the Bitlis—Pötürge Continent.

Another interesting observation is the "triggering" effects that plate boundaries seem to have on each other. We have seen that the Neo-Tethys in Turkey began opening, at least partly, as a series of back-arc basins over the Palaeo-Tethyan subduction zone. Similarly Maden and Çüngüş basins are marginal seas related to the Yüksekova subduction zone, whereas the Black Sea is a marginal basin triggered by the Cretaceous subduction zone that dipped beneath the Rhodope—Pontide Fragment. Such reciprocal effects of plate boundaries on each other are very widespread particularly in the history of the Tethyan belt (Şengör, in press) and, as the prediction of the tectonic style of any arc system depends on a knowledge of the age of the underthrusting oceanic lithosphere (Molnar and Atwater, 1978) and of the relative motion of the overriding plate with respect to an inert asthenosphere reference frame (Wu, 1978; Dewey, in press), it is almost impossible to foresee where and when they should occur due to the inherent uncertainties of field data and the self-destructive character of plate tectonics.

In Turkey, there is also an important change in tectonic style of collision along strike. The profiles in Figs. 3A and B were drawn to show this difference and Figs. 3A′ and 3B′ to emphasize it schematically.

In western Turkey the structure is dominated, south of the Izmir—Ankara suture, by far-travelled composite nappe systems that become progressively younger in the direction of transport (suture progradation of Roeder, 1979) and tectonic windows into rising metamorphic complexes (the Menderes Massif) that represent the parautochthon (inner part of the Anatolide—Tauride Platform) on which the nappes (Bozkir Unit) rest. The rise of the Menderes Massif began about 15 m.y. after its burial and continued until the extensional regime of western Anatolia interfered in post-Tortonian times. Synchronously with the uplift of the Menderes Massif, south-directed thrust—progradation in the Lycian Taurus and north-directed rétrocharriage in the western Pontides continued. This geometry and evolution seems to be the result of the continuous imbrication of the Anatolide—Tauride Platform following the terminal closure of the Izmir—Ankara ocean. During the Anatolian phase, a part of the platform gets buried beneath allochthons and becomes metamorphosed. The hot, ductile core of the metamorphic complex begins rising in a convergent environment and starts rotating the suture zone into a vertical position. As the convergence continues the suture zone overturns and the hinterland that was originally the overriding plate begins inserting itself "into" the side of the ductile metamorphic uplift as a wedge. This is expressed at the surface as late rétrocharriage (Fig. 3A′). This wedge initiates a separation between the upper and the lower continental crust which propagates to the surface in the south as listric thrusts that involve only the upper part of the crust and eventually root into the hot metamorphic core complex, the active examples of which may be expressed as low-velocity zones in the cores of orogens (e.g. beneath the Central Alps: Müller et al., 1976; Hsü, 1979). While the upper crust is being piled up in the form of successive thrust sheets (the late Eocene, "pre-Kale—Tavas—Oligocene" and the post-Burdigalian thrusting episodes in the Lycian Taurus), the lower crust can be removed by subduction (Molnar and Gray, 1979).

In eastern Turkey, along the profile B, the style is dominated by steeply to moderately dipping thrusts and numerous ophiolitic sutures. Along this profile there is no equivalent of the Menderes-type metamorphic uplifts. A glance at the geological evolution of this part of Turkey reveals the essence of this style. Here, much of the convergence until the Miocene was accommodated by the subductive removal of the floors of the numerous mini-oceans that separated a number of microcontinents. Intra-continental imbrication, the essence of the flat nappe structures and the Menderes-type metamorphic uplifts, has been absent in eastern Anatolia until the Miocene. Continuous suturing of microcontinents has created there a series of moderately to steeply dipping suture zones with abundant ophiolites. The limiting case of the latter tectonic style is to squash a mélange wedge between two continents (see Dewey, 1969b for an excellent example of such a situation), which has been the case for the East Anatolian Accretionary Complex still farther east.

Another important aspect of the East Anatolian Accretionary Complex is that it represents a "continental hole" created as a result of the collision of four continental objects that did not have perfectly matching margins and

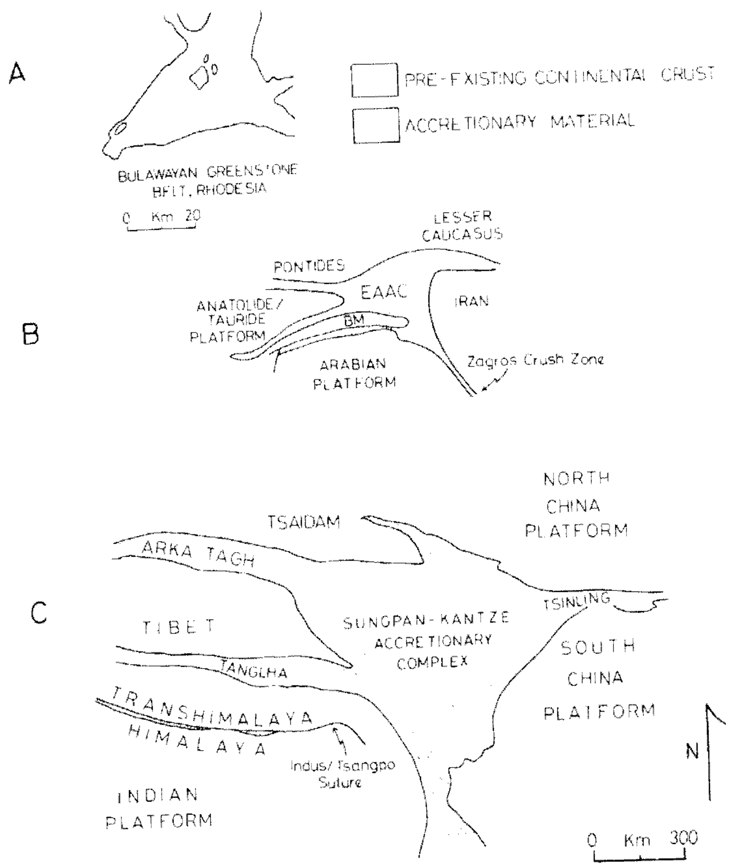

Fig. 7. Sketch maps showing three "deer's head" shaped, accretionary material-filled continental holes that resulted from the collision of more than two continental objects. B and C were drawn to the same scale. A is drawn after Burke et al. (1976) and C is from the geological map of China (1976). Discussion in text.

filled with oceanic island arcs and mélange complexes. The final collision of it with Arabia further deformed it internally and "consolidated" it by magmatic intrusions (see Şengör and Kidd, 1979). Figure 7 shows a comparison of the East Anatolian Accretionary Complex with two other possible analogues, the Precambrian greenstone belt of Bulawaya and the Sungpan—Kantze system of China. Burke et al. (1976) pointed out that the Bulawayan greenstone belt probably originated as a result of the collision of small island arcs today represented by the granodioritic terrain that surrounds the triangular greenstone belt. Continental crust has grown, they argued, by the coalescence of such arc systems by repeated collisions during the Archaean. Similarly, the Sungpan—Kantze system is the "continental hole" that resulted from the imperfect closure of Palaeo-Tethys among four continental objects (Şengör, in press). Its fill and overall geometry have an amazing resemblance to the East Anatolian Accretionary Complex, the difference being that the former is about three times as large as the latter and about 150 m.y. older. Such "deer's head"-shaped accretionary complexes, caught up in suture knots where continents cannot fill holes that result from irregular margin collision, may be more widespread than previously recognized and may contribute substantially to the areal growth of continents as has obviously been the case in Turkey and China.

Although extremely complicated and less well-exposed than many other orogenic belts in the world, Turkey offers perhaps a unique place for the study of Tethyan geology in particular and collision processes in general. It not only has a representative selection of nearly all of the main Tethyan palaeogeographic elements, but has an amazingly rich assortment of collision-related structures. After we obtain a reasonably clear idea of what the outlines of Tethyan evolution were, the next important step will be to try to fit the Hercynian deformations of Turkey into their proper place within the overall geometry of the Hercynian orogen in Euro-Africa.

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