Tectonophysics, 55 (1979) 361—376

POST-COLLISIONAL TECTONICS OF THE TURKISH—IRANIAN PLATEAU AND A COMPARISON WITH TIBET

A.M.C. ŞENGÖR and W.S.F. KIDD

Department of Geological Sciences, State University of New York at Albany, Albany, N.Y. 12222 (U.S.A.)

(Submitted December 12, 1977; accepted for publication June 19, 1978)

ABSTRACT

Şengör, A.M.C. and Kidd, W.S.F., 1979. Post-collisional tectonics of the Turkish—Iranian Plateau and a comparison with Tibet. Tectonophysics, 55: 361—376.

The Turkish—Iranian Plateau (Fig. 1) is a high region with an average elevation of about 1.5 km. During the late Miocene the last piece of oceanic lithosphere between the Eurasian and Arabian continents was eliminated at the Bitlis/Zagros suture zone. Continued convergence across the collision site resulted in the shortening of the plateau across strike by thickening and by sideways motion of parts of it. Predominantly calc-alkaline volcanism is present on the highest portions of the area, despite the absence of a descending slab of lithosphere. Surface geology and volcanism of the Turkish—Iranian Plateau resemble greatly those of the Tibetan Plateau, and both are underlain by a zone of seismic attenuation. From a comparison of these features and their tectonic setting we argue that the two plateaux are homologous structures, albeit at different stages of their evolution. Both areas appear to be tectonically alive and actively shortening. Available evidence lends little support to the hypothesis of large-scale underthrusting of continental lithosphere and of plastic-rigid indentation where such high plateaux, located directly in front of the "rigid indenter", are considered to be tectonically "dead". Their peculiar features are best explained in terms of shortening and thickening the continental crust whereby its lower levels are partially melted to give rise to calc-alkaline surface volcanism. Minor associated alkaline volcanism may be due to local longitudinal cracking of the crust to provide access to mantle.

INTRODUCTION

If convergence of two colliding continents continues significantly after continental apposition, the displacement must be converted into strain by intracontinental deformation. Argand (1924) suggested that this deformation will be most intense near the collision site (e.g., Himalayas and the Tibetan Plateau) and may be accomplished by thrusting one of the colliding continents (e.g., the Indian subcontinent) under the other, while the overriding continent (e.g., Asia) deforms internally over wide regions to give rise to intracontinental compressional structures (his plis de fond; see his figs. 12 and 13). He explained the high elevation of the Tibetan Plateau, averaging 5 km above sea level, by thrusting a considerable portion (about 600 km) of India under Asia. Argand's model has been nearly directly adopted in some recent attempts to account for the origin of the Tibetan Plateau (e.g., Powell and Conaghan, 1973, 1975); however, McKenzie (1969) has shown that significant underthrusting of continental lithosphere must be prohibited by its buoyancy. If large-scale underthrusting of continents does not occur, there are two other possible mechanisms to take up intracontinental convergence at collision zones. One is "sideways" motion to wedge out pieces of continental lithosphere, as is currently occurring, for example, in the eastern Mediterranean, the Anatolian Plate escaping westwards along the North and East Anatolian Transform Faults from the East Anatolian convergent zone (McKenzie, 1972; Dewey and Şengör, 1979). The other is fairly homogeneous crustal shortening and complementary thickening, which has been suggested to be responsible for the height of the Tibetan Plateau (Dewey and Burke, 1973).

For an understanding of the geometry and mechanics of continental collision, it is necessary to know whether large-scale continental underthrusting occurs, or whether the majority of the displacement across the collision zone is taken up by intracontinental shortening and thickening. As the process of crustal shortening and thickening has also been suggested to be responsible for the formation of wide terranes of basement reactivation such as those of the Variscides, Grenville, and Pan-African (Dewey and Burke, 1973), an understanding of the process is necessary to help understand the tectonic history of the reactivated terranes. The tectonic nature of the Tibetan Plateau and homologous areas may provide a partial test for the plastic-rigid indentation model of continental collision proposed by Molnar and Tapponnier (1975).

Surface geology, coupled with geophysical observations on seismicity and gravity should be able to illuminate the origin and evolution of Tibetan-type high plateaux. Specifically, it should be able to distinguish between wholesale continental underthrusting, and crustal thickening by shortening. It should also show whether such regions are actively shortening or passively transmitting stress. However, it is unfortunately at present not possible to study the detailed surface geology of the best example of such plateaux, namely Tibet. Although much geological work on the Tibetan Plateau is currently being undertaken using the existing literature and Landsat images (for an overview of the ongoing research see Molnar and Burke, 1977), critical details of the geology are very poorly known, or in dispute.

The purpose of this paper is to approach the problem of the origin and evolution of Tibetan-type high plateaux by examining an analogous region, the Turkish—Iranian Plateau (Fig. 1), and compare its post-collisional tectonics with what is known about Tibet in an attempt to shed some light on the nature of the latter.

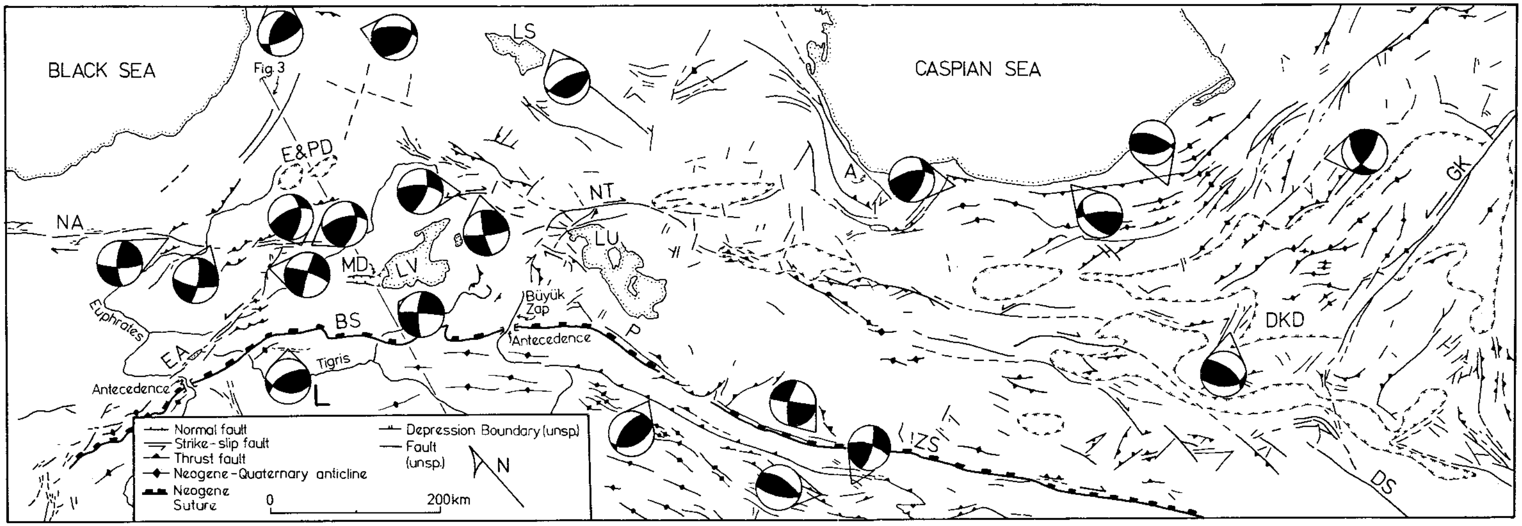

Fig. 1. Structures of the Turkish—Iranian Plateau that have been active during some interval of or throughout the time period late Miocene to present. Some of the structures shown on this map originated during pre-late Miocene times but remained active thereafter; others originated in post-Miocene times; very few may have been inactive throughout the time period considered. (For detailed information on the structures in Iran see, Berberian, 1976a; for Turkey, information is scattered throughout the papers cited as references to this figure.) Bitlis and Zagros Sutures today are marked by active thrust and/or strike-slip faults. Note the two distinct belts of thrusting and strike-slip faulting in eastern Turkey (from the North and East Anatolian Transforms to the east of Lake Van) and northwestern Iran (from the west of Lake Urmiyah to the west of Dasht-i Kavir Depression), merging into the wide diffuse area of deformation in the east. All fault plane solutions except the Lice, 1975 earthquake, are selected from Shirokova, 1962 and McKenzie, 1972 to show the instantaneous strain in the plateau as shown by those earthquakes. The Lice solution (1) was kindly communicated by Prof. M.N. Toksöz.

Key: *NA* = North Anatolian Transform; *EA* = East Anatolian Transform; *E & PD* = Erzurum and Pasinler Depressions; *MD* = Muş Depression; *BS* = Bitlis Suture; *L* = Lice earthquake focal mechanism solution; *NT* = North Tabriz Fault; *A* = Astara Fault; *P* = Piranshar Fault; *ZS* = Zagros Suture; *DKD* = Dasht-i Kavir Depression; *GK* = Grand Kavir Fault; *DS* = Deh-Shir Fault; *LV* = Lake Van; *LS* = Lake Sevan; *LU* = Lake Urmiyah. The map is compiled and somewhat simplified from Arpat and Saroglu (1972, 1975), Arpat et al. (1977), Berberian (1976b), Ketin (1969), Seymen (1975), Seymen and Aydin (1972), and Toksöz et al. (1977).

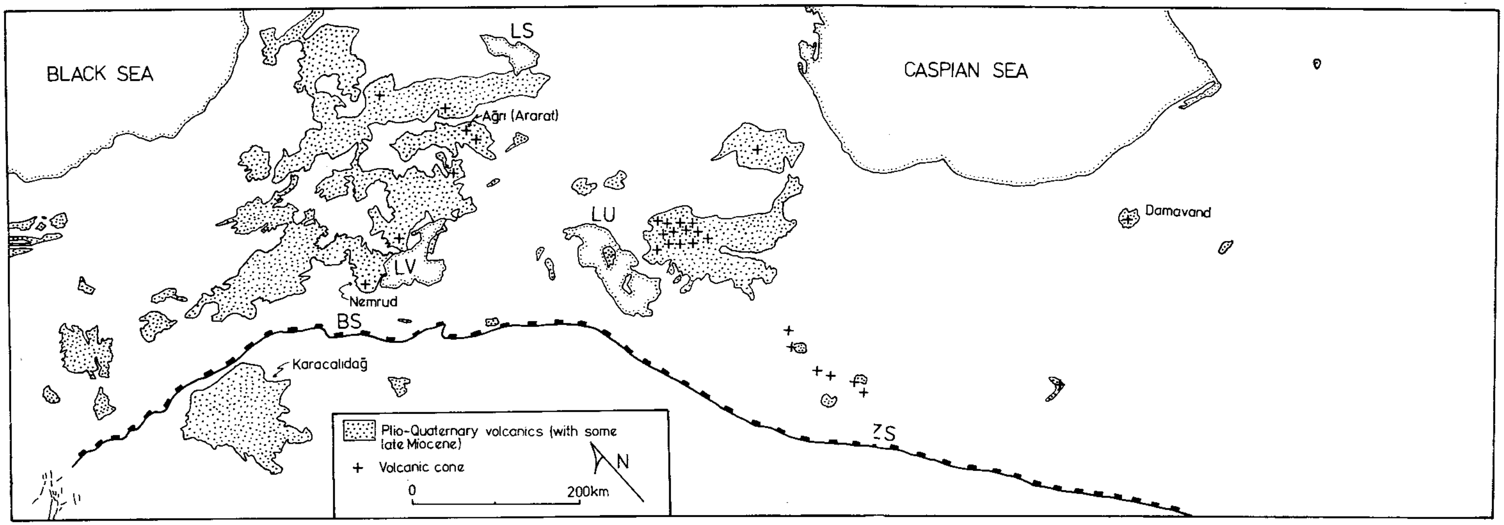

Fig. 2. Distribution of Plio-Quaternary volcanic rocks and cones on the Turkish–Iranian Plateau. The map is compiled from M.T.A. (1962a, 1962b, 1963, 1964, 1966, 1974), Ketin (1961), Altinli (1966b), Berberian (1976b), International Tectonic Map of Europe (1962).

POST-COLLISIONAL TECTONICS AND VULCANICITY OF THE TURKISH—IRANIAN PLATEAU

The pre-collision geology of the Turkish—Iranian Plateau is well enough known to reconstruct the first order tectonic phenomena that finally gave the region its geometry immediately after the collision. Before the Priabonian (late Eocene) two oceans existed in the area, one south of the Pontide—

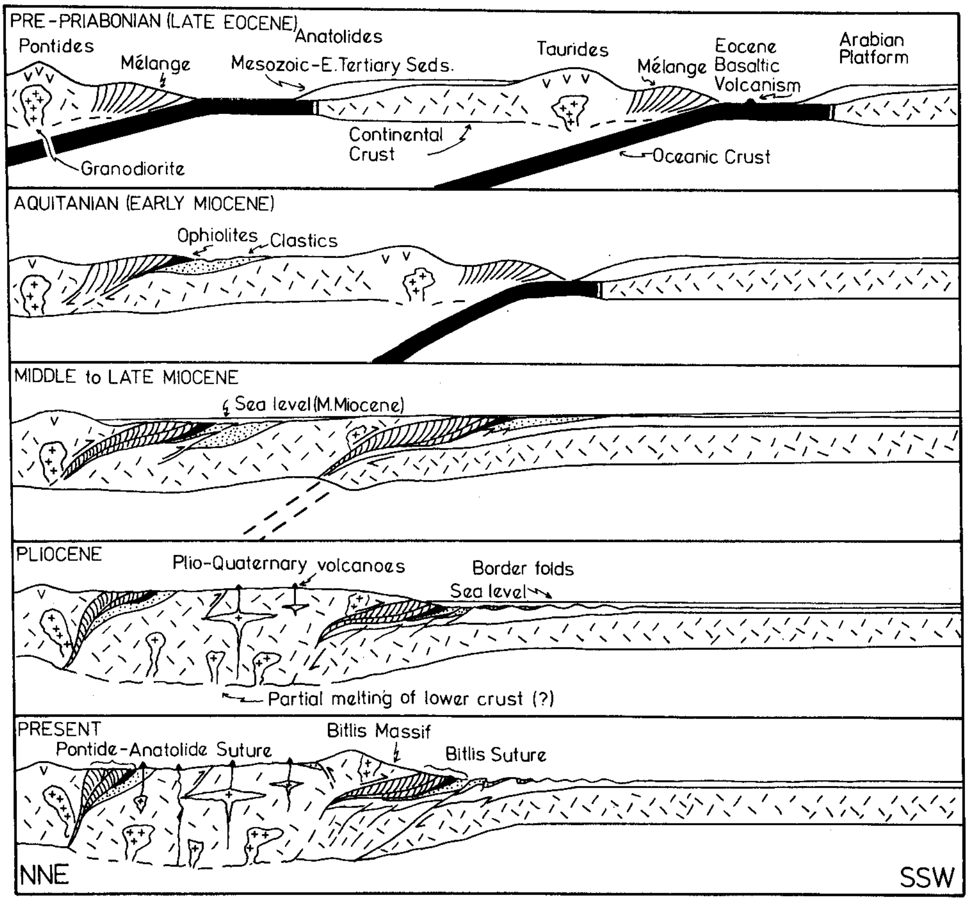

Fig. 3. Schematic sequential diagram to show the successive steps in the evolution of the Turkish—Iranian Plateau from pre-Priabonian to present. Approximate location of the cross-section is shown on Fig. 1. This particular cross-section is selected for illustration merely on the basis of the availability of data to us. The section is based on our interpretation of the data taken from, among others, Adamia et al. (1977), M.T.A. (1962a) and Seymen (1975) for the Pontides and Pontide—Anatolide Suture; Altinli (1966a, b) for the Anatolides; M.T.A. (1962b), Hall (1976) for the Taurides, and M.T.A. (1962b) and Altinli (1966a, b) for the Border Folds. See text for explanation.

Lesser Caucasus—Alborz mountain system and the other south of the Anatolids and the Central Iranian Plateau (Fig. 3). The northern ocean closed during the Priabonian as shown by the deposition of exogeosynclinal sediments and major overthrusting on the Anatolide side (Seymen, 1975); convergence across this suture continued until the Burdigalian (late early Miocene) (Seymen, 1975; Adamia et al., 1977). The southern, Bitlis/Zagros ocean, closed during the late Miocene and convergence across it is still in progress (Dewey et al., 1973; Dewey and Şengör, 1979).

At present, in eastern Turkey more than 70% of the total area, generally represented by moderately flat surfaces, lies between 1.5 and 2.5 km (Tanoğlu, 1947). In western Iran large areas lie also at similar elevations, which, however, decrease to about 500 m farther eastward, near and in the Dasht-i Kavir Depression (Fig. 1). The region as a whole has a plateau character and the only significant peaks in the hypsographic curve of the region result from the Plio-Quaternary volcanic cones, such as Mt. Ağri (5165 m).

A late Oligocene—early Miocene marine transgression coming from the west inundated large areas of the Turkish—Iranian Plateau (Alavi-Naini, 1972 fig. 62; Stöcklin, 1968). This is represented by evaporite (largely gypsum)/ sandstone/limestone lithologies passing upwards into shallow marine marls and reefal carbonates (Oligo-Miocene gypsiferous series and Lower Miocene carbonates in eastern Turkey, Lahn, 1950; M.T.A., 1963, 1964; Altinli, 1966a; Qum Formation in Iran, Geol. Surv. Iran, 1969; Alavi-Naini, 1972). That the area remained under the sea, at least locally, until the Serravallian (late medial Miocene) is shown by the microfossils collected from near Lake Van (Gelati, 1975). A marine regression in late Miocene time is indicated by the lacustrine and fluvial sediments overlying the marine sediments (Pontian in eastern Turkey, Altinli, 1966a; Upper Red Formation in Iran, Geol. Surv. Iran, 1969). As a result of the emergence, a late(st?) Miocene—early Pliocene erosion surface originated that was interrupted by closed drainage basins, probably resembling the present Dasht-i Kavir Depression. These contain playa deposits including salt and gypsum which pass into coarse clastic basin margin facies and become completely conglomeratic near the basin edges (Geol. Surv. Iran, 1969; Erinç, 1953). The late Miocene—early Pliocene erosion surface, onto which the abundant andesitic-dacitic lavas of Pliocene age were erupted (Ketin, 1961; Altinli, 1966a; Innocenti et al., 1976), was considerably uplifted, especially in the western and central sections of the Turkish—Iranian Plateau, towards the end of the Pliocene. This uplift is documented by the deep dissection of the Miocene—Pliocene erosion surface and the infilling of the resulting valleys by Pleistocene lava flows (Erinç, 1953). Tanoğlu (1947) and Erinç (1953) argued that the surprising uniformity of timing of uplift and of the elevations attained by the erosion surface is indicative of a block uplift of the entire region rather than of a progressive wave of uplift. Several depressions in the area such as the Erzurum-Pasinler and Dasht-i Kavir, some of which are fault-bounded (e.g., Muş Depression), may have originated or became isolated during this phase of uplift (see Fig. 1).

In spite of the block character of the uplift in the Turkish—Iranian Plateau in general, the Bitlis suture zone appears to have been uplifted later or more slowly than the rest of the plateau. This is shown by two antecedent rivers originating on the plateau and flowing onto the Arabian Platform after cutting across the Bitlis suture zone in the southeastern Taurus Mountains. Izbirak's (1951) geomorphological studies along the valley of Büyük Zap (Fig. 1) showed that the river cuts straight across the geological structures of the suture irrespective of their orientation and documented several nested alluvial terraces on the sides of its valley showing the progressive uplift of the suture zone. Huntington (1902) showed that the Euphrates (Fig. 1) is also antecedent. Just north of the suture the uplifted and dissected latest Miocene—early Pliocene erosion surface still has a southerly dip (Erinç, 1953).

The geomorphological data allow the following inferences: (1) the Turkish—Iranian Plateau, especially its western and central parts, was significantly uplifted, probably as a block, by the beginning of the Pleistocene, the amount of uplift diminishing in the direction of the Dasht-i Kavir Depression; (2) the mountains now on the site of the suture were uplifted only after the uplift of the Turkish—Iranian Plateau, and in places still lie lower than the plateau.

After the collision of the Arabian continent with Eurasia during the late Miocene, convergence between Arabia and Eurasia continued until the present as shown by the Pliocene to Recent folding and thrusting of the Arabian shelf sequence in the Zagros (Ricou et al., 1977), the Border Folds of southeastern Turkey (Ketin, 1966), and the present diffuse seismicity of the entire Turkish—Iranian Plateau (Canitez and Üçer, 1967; McKenzie, 1972; Nawroozi, 1972; Berberian, 1976a, b; M.N. Toksöz, personal communication, 1977). In the Turkish—Iranian Plateau, this convergence is taken up, in the Turkish sector, by wedging out the Anatolian Plate into the oceanic tract of the eastern Mediterranean along the North and East Anatolian Transform Faults (McKenzie, 1972; Dewey and Şengör, 1979) and partly by shortening the continental crust by thrusting. Numerous post-Miocene thrusts have been documented in eastern Turkey (see, for example, Altinli, 1966b, p. 4 and plate II). Several of these occur near the traces of the North and East Anatolian Transform Faults and may be related to strike-slip motion rather than to general crustal shortening across eastern Turkey. However, many others are located well east of the point where the two transforms meet, and therefore appear to be due to general crustal shortening (Fig. 1). Although the majority of these thrusts are south-vergent, a number of them moved to the northeast or to the northwest. The present data indicate that the thrusts in eastern Turkey are crowded along two main lines: one passing near the Erzurum—Pasinler Depressions and Mt. Ağrı, the other following the Bitlis Suture (Fig. 1). Between these two, the easterly continuation of the North Anatolian Fault Zone (Ketin, 1948, 1969) seems to have a major thrust component according to McKenzie's (1972) fault-plane solutions (Fig. 1), although the surface breaks during earthquakes show strike-slip as by far the predominant component (Ketin, 1948, 1969). It is possible that this zone of faults is taking up some of the shortening by thrusting. Thrusting with some minor folding (?) is observed near the Lesser Caucasus (M.T.A., 1974, plate VI). According to reports currently available, folding of post-Miocene age in eastern Turkey appears to be absent or at best very subordinate to thrust and strike-slip tectonics. However, I. Ketin (personal communication, 1978) informs us that the Miocene, and, locally, the Pliocene rocks are folded, in places strongly. Only the youngest lava flows are flat-lying, although cut by faults.

Adamia et al. (1977) believe, as did McCallien (1947), that there are also north—south trending structures, predominantly of extensional nature, in eastern Anatolia extending northwards into the Caucasus, as indicated by the alignment of basaltic volcanic centers. Although D.P. McKenzie (personal communication, 1977) obtained a north—south striking fault-plane solution in northwestern Iran, the only documented north—south oriented extensional structure in eastern Turkey that we know of is the group of north—south fissures of the Pliocene Karacalidag basaltic shield volcano (Fig. 2, south of the Bitlis Suture).

In contrast to eastern Turkey, there is widespread post-Miocene folding in Iran along with major thrust and strike-slip tectonics (Berberian, 1976a, b; A. Gansser, pers. comm., 1978). In northwestern Iran strike-slip and thrust tectonics predominates over folding that may be completely absent here as in eastern Turkey (Geol. Surv. Iran, 1969; Berberian and Arshadi, 1976). Eastward, to the south of the Caspian Sea, folds begin to appear across the full width of the plateau and are especially abundant in and just northwest of the Dasht-i Kavir Depression. Within the Dasht-i Kavir folding is apparently active now, as seen on Landsat images. Earthquake distribution in Iran, as in Turkey, shows that most of the structures shown on Fig. 1 are active and taking up the Arabia—Eurasia convergence. Although between Lake Urmiyah and Dasht-i Kavir the faults are aligned roughly in two discrete belts, farther east the whole width of the plateau is deformed by faults and folds.

The post-Miocene tectonic picture of the Turkish—Iranian Plateau is one of active shortening across the plateau. In Turkey a considerable portion of this convergence is taken up by wedging out the Anatolian Plate along the North and East Anatolian transform faults with throws of 85 ± 5 km and 22 ± 5 km, respectively (Seymen, 1975; Arpat and Şaroglu, 1972; Seymen and Aydin, 1972). East of where these faults meet, active shortening has resulted in some strike-slip faulting (e.g., Toksöz et al., 1977), but thickening on numerous thrusts is predominant. In Iran, thrusts and folds are abundant and they probably greatly predominate over sideways motion. This may be because (1) Iran has less space to "escape" into than does Turkey, and (2) its lower elevation requires less work for the present uplift. Negative Bouguer anomalies (—150 mgal) over a large part of the Turkish—Iranian Plateau

(Özelci, 1973a, b) are most easily, but not uniquely, interpreted in terms of crustal thickening, especially in the light of the geological data.

The Turkish—Iranian Plateau, especially its western and central, most elevated sections, is also the locus of intense late Tertiary—Quaternary volcanism (Fig. 2) and contains volcanoes that have erupted during historic times (e.g., Mt. Nemrud eruption, 1441 A.D., Erinç, 1953). Although parts of the region were characterized by Cretaceous to Miocene calc-alkaline volcanism prior to the Pliocene to Recent volcanic phase (Ketin, 1961; Altinli, 1966a), this was probably related to a subduction zone consuming Bitlis—Zagros ocean floor and dipping north-northeast under the future Turkish—Iranian Plateau as shown by the progressive increase in K_2O/SiO_2 ratios in the associated igneous rocks from south to north across the plateau (Adamia et al., 1977). This phase of volcanic activity came to an end in Turkey about 6 m.y. ago (Innocenti et al., 1976) and in Iran sometime during the late Miocene (Jung et al., 1976). Volcanic activity recommenced during the Pliocene and is still active (Ketin, 1961; Gansser, 1966). There is no evidence of a descending lithospheric slab beneath the plateau that can be connected with the active volcanism; the age of collision (approx. 10 m.y. ago) and the average convergence rate over this time period (about 4.5 cm/yr, McKenzie, 1972) indicate that the slab must long be past the 100—150 km depth where the majority of the calc-alkaline melts are generated. Pliocene to Recent volcanism is very extensive in the highest parts of the plateau (Fig. 2) and includes both calc-alkaline and alkaline associations, although the former predominates over the latter. The calc-alkaline association is represented by andesites, dacites, and rhyolites with some ignimbrites, whereas basalts and very limited phonolites and trachytes represent the alkaline association. Some volcanoes (e.g., Nemrud, Özpeker, 1973) appear to have erupted both alkaline and calc-alkaline rocks. Lambert et al. (1974) have reported a $^{87}Sr/^{86}Sr$ ratio of 0.7050 ± 0.0005 from the calc-alkaline lavas of Mt. Ağrı. If the Devonian and Permian that outcrop very near the volcano represent the total basement beneath it, then this ratio precludes the continental crust as the source material for the melt. But in a terrane as complex as eastern Turkey, that contains a huge amount of accretionary mélange material, this may not be the case. The observation that the Plio-Quaternary volcanics are almost entirely confined to the highest parts of the plateau and that there is a zone of seismic attenuation beneath the plateau (Toksöz and Bird, 1977) supports the view that a great portion of the calc-alkaline volcanics here may be the products of the partial melting of the lower levels of the thickened continental crust. The less abundant alkaline rocks are probably the result of the local longitudinal cracking of the crust, under north—south shortening, to provide access to the mantle.

In summary, all the available tectonic and volcanic evidence from the Turkish—Iranian Plateau indicates active shortening and partial melting of the lower levels of the thickened crust. Some small amount of lateral extension is indicated by the alkaline volcanics and minor north—south trending normal faulting.

OUTLINE GEOLOGY OF THE TIBETAN PLATEAU

The main sources of information on the geology of the Tibetan Plateau are Backstrom and Johannsen (1907), Hennig (1915), Norin (1946), Chang Ta (1959), Chang and Zdeng (1973) and the Geological Map of China (1976). The pre-Mesozoic geology of the plateau is separable into two unequal and contrasting parts (Fig. 4). The northern border, now occupied by these eastern and western Kun Lun ranges and the Altyn Tagh, consists dominantly of medium to high-grade regionally metamorphosed rocks, whose deformation and metamorphism is pre-Devonian (Norin, 1956; Geological Map of China, 1976), and, in places, Precambrian (Geological Map of China, 1976). From this region all the way south to the Indus Suture, the rocks in the undissected part of the plateau (up to about 92°E) consist mainly of low-grade slates and greywackes, for the most part of Carboniferous—Permian age (Hennig, 1915; Norin, 1946; Geological Map of China, 1976). Subordinate areas of island-arc-type volcanics and plutonics of Carboniferous and Permian age are reported within this terrane east of 92°E (Chang and Zdeng, 1973; Geological Map of China, 1976), rare pre-Mesozoic granitic rocks found west of 92°E may be correlative. This terrane is interpreted (Kidd, et al., in prep.) as a huge accretionary prism of sediments, mostly derived from the extensive area of late Paleozoic orogenesis in Central Asia, and containing minor subduction related plutonic and volcanic rocks (the Makran area is a present day analog; see Farhoudi and Karig, 1977). Granodiorites in the southernmost part of this terrane (Kailas and Kyu-Chu "granites") may be of late Cretaceous age and represent the eroded roots of an Andean-type arc constructed during the subduction that removed the ocean between India and Asia. Alternatively, some of these granodiorites may post-date the collision. No rocks of continental shield type are known south of Kun Lun, and therefore Tibet is apparently not a fragment of a pre-Mesozoic continent.

Over large areas of the Tibetan Plateau, as far as 92°E, the deformed pre-Mesozoic rocks are unconformably overlain by a sequence of Jurassic and Cretaceous sedimentary rocks. Clastic rocks, mainly red sandstones at the base, pass up into rudist-bearing limestones. In the northern part of the plateau the rocks are mainly red sandstones and locally contain gypsum. These Mesozoic sediments are found over the whole plateau south to near the Indus Suture, including near Lhasa, where they are strongly folded (Hayden, 1907). Dips recorded by Hedin (Backstrom and Johannsen, 1907; Hennig, 1915) and analysis of Landsat images (Kidd et al., in prep.) show that the Mesozoic sequence is buckle folded, in places strongly, over the whole width of the plateau with east—west trending axial surfaces. Folds of this type are usually accompanied by thrust faults (e.g., in the Jura Mountains), although it is difficult to identify these positively from the imagery. The amount of shortening represented by these folds (Kidd et al., in prep.) is probably not less than 20%; in addition, compressive deformation that results in buckle

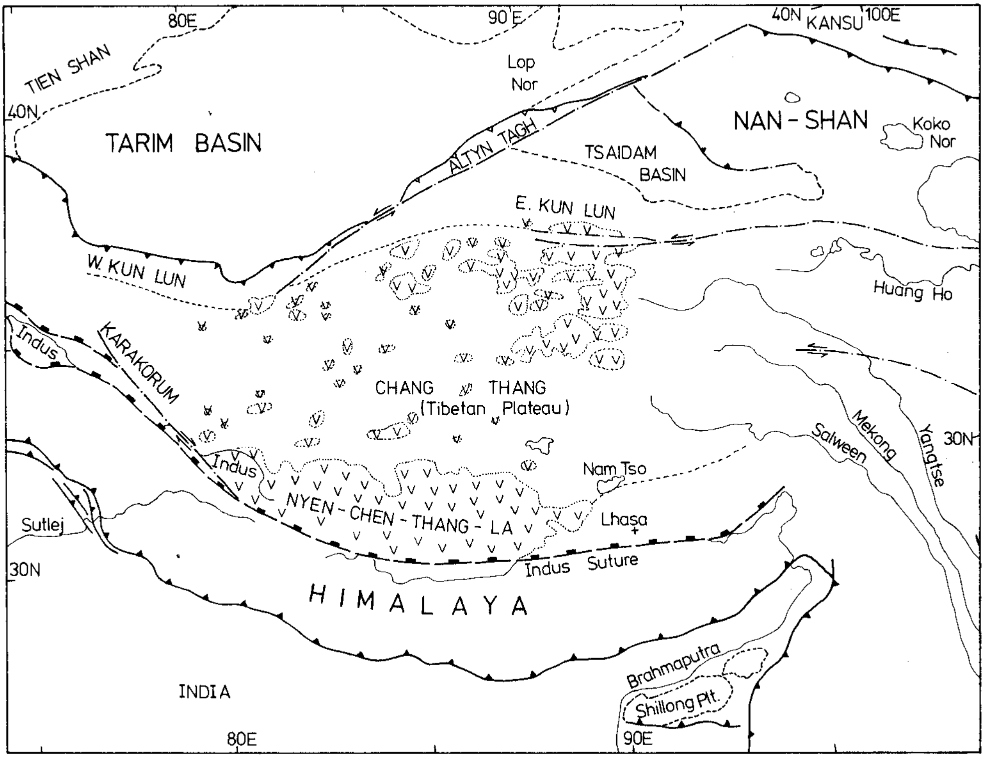

Fig. 4. Tectonic sketch map of the Tibetan Plateau and surrounding areas. Lines with black triangles: active thrust faults; lines with open triangles: inactive thrust boundary. v = Neogene and younger volcanic rocks.

folding is often accompanied by about 10% bulk shortening and thickening before folding occurs. Although Norin (1946) reported, in the western part of the plateau just north of the Karakorum, some areas of flat-lying late Cretaceous rocks unconformably overlying folded Jurassic and early Cretaceous sediments, Landsat images and Norin's map (1946) suggest that the Upper Cretaceous rocks in this area are mostly folded and are only locally flat-lying; such regions of flat-lying strata may in fact represent wide, flat-bottomed synclines, characteristic of regions where surficial buckle folding is the dominant tectonic style, as, for example, in the northwestern Jura Mountains. In the Tsaidam Basin, northeast of Tibet (Fig. 4), the youngest playa sediments appear to be presently folding and are overthrust by the Permian rocks on the northeast border of the basin. The axis of maximum shortening, as judged from the strike of the fold axial planes, is northeast as opposed to the north—south shortening axis given by the east—west axial planes over Tibet. This may be due to a relatively recent reorganization of the active deformation as it spreads away from the relatively stable Tarim Block. In terms of style of folding, overall morphology and tectonic setting, the Tsaidam Basin has an astonishing similarity to the Dasht-i Kavir Depression of Iran (an observation made independently by A. Gansser, pers. comm., 1978).

Young volcanic rocks of andesitic and more silicic calc-alkaline compositions are widely developed over the Tibetan Plateau as far as 92°E (Backstrom and Johanssen, 1907; Hennig, 1915; Norin, 1947; Burke et al., 1974, Kidd, 1975). Those in the northern and central parts of the plateau clearly postdate the folding of the Mesozoic strata, because they are seen on Landsat images to drape the folded Mesozoic strata in many places (Kidd et al., in prep.). In the southern part of the plateau a 200 km wide belt of volcanics adjoins the northern side of the Indus Suture, and stretches from the Indus in the west to near Lhasa in the east. It contains a few obviously young volcanic features on the Landsat imagery, but the youth of most of these abundant volcanics is not morphologically obvious. Specimens collected by Hedin from this terrane are mostly ignimbrites and subordinate related igneous rocks. While this belt could be the remains of an Andean-type magmatic arc, the reports of abundant hot and boiling springs in this terrane indicate the widespread presence of magma at no great depth. For this reason, and because draping relations to folded Cretaceous sediments can be seen on the Landsat images in the northern part of this terrane, it is thought (Burke et al., 1974; Kidd et al., in prep.) that the bulk of these volcanics are young, that is, Neogene and younger, although a small portion could be of early Tertiary or late Cretaceous age. The uplift of Tibet, although poorly dated, clearly predated the ongoing uplift of the Himalayas; this is shown by the antecedent Indus and Brahmaputra rivers.

DISCUSSION AND CONCLUSIONS

The great resemblance between the Turkish—Iranian Plateau and Tibet with respect to overall morphology and tectonics was first emphasized by Von Zahn (1906) and the foregoing descriptions show that the geological resemblances between the Tibetan and the Turkish—Iranian Plateau are readily apparent. Both plateau areas adjoin a suture where continental apposition has occurred and collision is in progress. The start of the collision being older, perhaps 30—40 m.y. ago (Dewey and Burke, 1973; Molnar and Tapponnier, 1975), and convergence being at a faster rate (about 5.0 cm/yr., Molnar et al., 1977 vs. 4.5 cm/yr, McKenzie, 1972) are perhaps the reasons why Tibet is higher and more extensive than the Turkish—Iranian Plateau. In both areas the plateaux were uplifted before the suture zone. We take this and the available seismic evidence to indicate that large-scale continental underthrusting, as suggested by Argand (1924), is not the cause of uplift. Shortening and resultant thickening as proposed by Dewey and Burke (1973) and favored by Le Fort (1975) for the origin of Tibet also appears to be the model most consistent with the presently available data from the Turkish—Iranian Plateau, as well as from Tibet. Therefore the lower elevation of especially the Iranian segment of the Turkish—Iranian Plateau may be also due to the existence of cratonic nuclei within the Central Iranian Plateau (Stöcklin, 1974) that resist deformation by shortening better than the relatively weaker accretionary prism material that appears to make up a large portion of the Tibetan basement.

Both plateau areas exhibit folding of covering sedimentary rocks in at least part of their areas. In the Turkish—Iranian Plateau this folding, accompanied by extensive thrusting, started nearly synchronously with the collision. The time of folding in Tibet is less well-constrained, but the huge area affected by folding, particularly across strike, is remarkable and it seems to us, considering the active folding in the Tsaidam Basin, unlikely to have happened before the collision. Thrusting in Tibet does not seem to be as widespread as it is in the Turkish—Iranian Plateau, but we believe this to be an artifact of recognizing thrusts on Landsat imagery and not the result of the actual absence of the process. Both plateaux show minor normal faulting at high angles and strike-slip faulting at low angles to the suture.

Volcanism on the Tibetan Plateau greatly resembles that found in eastern Turkey and western Iran in composition, wide extent, occurrence on high ground, and, at least for a large proportion of the Tibetan volcanics, in its post-collisional age. On the Turkish—Iranian Plateau alkaline rocks, in the form of alkaline basalts, are present; their apparent absence from Tibet may merely reflect inadequate sampling.

We believe the Tibetan and the Turkish—Iranian Plateaux to be homologous structures. The geomorphological and structural data lend little support to the concept of large-scale continental underthrusting to form such high plateaux. The post-Miocene tectonics and volcanicity of the Turkish—Iranian Plateau and the present folding in the Tsaidam Basin and the volcanism over large areas of the Tibetan Plateau indicate that these plateaux are tectonically "alive" and active shortening is taking place. The view that these regions represent tectonically "dead" areas (e.g., Molnar and Tapponnier, 1975) does not seem justified in the face of the available geological data. However, its should be borne in mind that our knowledge of Tibetan-type high plateaux is still exceedingly limited and that no hypothesis for the origin and evolution of such regions can be considered wholly satisfactory until it also accounts for the existence of the Altiplano of the Andes, which has very similar properties to the Tibetan and Turkish—Iranian Plateaux, except for the absence of continental collision (Audebaud et al., 1973).

ACKNOWLEDGMENTS

We thank A. Gansser for encouragement and discussions, M. Nafi Toksöz, and D. McKenzie for unpublished information. We also thank D. McKenzie for helpful criticism.

REFERENCES

Adamia, S.A., Lordkipanidze, M.B. and Zakariadze, G.S., 1977. Evolution of an active continental margin as exemplified by the alpine history of the Caucasus. Tectonophysics, 40: 183—199.

Alavi-Naini, M., 1972. Étude géologique de la Région de Djam. Geol. Surv. Iran, Rep. 23: 288 pp.

Altinli, E., 1966a. Doğu ve güneydoğu Anadolu'nun jeolojisi. Kisim I. M.T.A. Dergisi, 66: 35—74.

Altinli, E., 1966b. Doğu ve güneydoğu Anadolu'nun jeolojisi. Kisim II. M.T.A. Dergisi, 67: 1—24.

Argand, E., 1924. La tectonique de l'Asie. C.R. Congr. Géol. Intern., 13e, 1: 171—372.

Arpat, E. and Şaroğlu, F., 1972. The East Anatolian fault system: thoughts on its development. Bull. Miner. Res. Explor. Inst. Turk., 78: 33—39.

Arpat, E. and Şaroğlu, F., 1975. Türkiye'deki bazi önemli genç, tektonik olaylar. Türk. Jeol. Kurumu, Bül. 18: 91—101.

Arpat, E., Şaroğlu, F. and Iz, H., 1977. 1976 Çaldiran Depremi. Yeryuvari ve Insan, 2; 29—41.

Audebaud, E., Capdevila, R., Dalmayrac, B., Debelmas, J., Laubacher, G., Lefevre, C. Marocco, R., Martinez, C., Mattauer, M., Megard, F., Paredes, J. and Tomasi, P., 1973. Les traits géologiques essentiels des Andes Centrales (Pérou—Bolivie). Geogr. Phys. Geol. Dyn., 15: 73—114.

Backstrom, H. and Johanssen, H., 1907. Geology. In: S. Hedin, (Editor), Scientific Results of a Journey in Central Asia 1899—1902. Vol. 6, pt. 2, Stockholm.

Berberian, M., 1976a. Contribution to the Seismotectonics of Iran (Part II). Geol. Surv. Iran, Rep., 39: 516 pp.

Berberian, M., 1976b. Seismotectonic map of Iran, scale 1/2,500,000. In: M. Berberian (Editor), Contribution ot the Seismotectonics of Iran (Part II). Geol. Surv. Iran, Rep., 39: 516 pp.

Berberian, M. and Arshadi, S., 1976. On the evidence of the youngest activity of the North Tabriz Fault and the seismicity of Tabriz City. In: M. Berberian (Editor), Contribution to the Seismotectonics of Iran (Part II). Geol. Surv. Iran, Rep., 39: 397—418.

Burke, K., Dewey, J.F. and Kidd, W.S.F., 1974. The Tibetan Plateau: its significance for tectonics and petrology. Geol. Soc. Am., Abstr. with Progr., 6: 1027—1028.

Canitez, N. and Üçer, S.B., 1967. Computer determinations for the fault-plane solutions in and near Anatolia. Tectonophysics, 4: 235—244.

Chang, Chen-fa and Zdeng, Shi-lang, 1973. Tectonic features of the Mt. Jolmo Lungma region in Southern Tibet, China. Sc. Geol. Sinica, 1973, 1: 1—12.

Chang, Ta, 1959. The Geology of China, Peking (Translated In: U.S.J.P.R.S. item 19, 209, 1963, 623 pp.).

Dewey, J.F. and Burke, K.C.A., 1973. Tibetan, Variscan and Precambrian basement reactivation: products of continental collision. J. Geol., 81: 683—692.

Dewey, J.F. and Şengör, A.M.C., 1979. Aegean and surrounding regions: complex multiplate and continuum tectonics in a convergent zone. Geol. Soc. Am. Bull. 90: 84—92.

Dewey, J.F., Pitman, W.C. III, Ryan, W.B.F. and Bonnin, J., 1973. Plate tectonics and the evolution of the Alpine System. Geol. Soc. Am. Bull., 84: 3137—3180.

Erinç, S., 1953. Doğu Anadolu Coğrafyasi. Istanbul Üniv. Yay., No. 572, 124 pp.

Farhoudi, G. and Karig, D.E., 1977. Makran of Iran and Pakistan as an active arc system. Geology, 5: 664—668.

Gansser, A., 1966. Catalogue of the active volcanoes of the world, including solfatara fields. Part 17, Iran. Int. Volc. Assoc., Naples, 20 pp.

Gelati, R., 1975. Miocene marine sequence from Lake Van, eastern Turkey. Riv. Ital. Paleontol. Stratigr., 81: 477—490.

Geologic Map of China, 1 : 4 m. Peking, 1976.
Geological Survey of Iran, 1969. Explanatory text of the Zanjan quadrangle map, scale 1/250,000. Geol. Surv. Iran, Tehran, 61 pp.
Hall, R., 1976. Ophiolite emplacement and the evolution of the Taurus Suture Zone, southeastern Turkey. Geol. Soc. Am. Bull., 87: 1078—1088.
Hayden, H., 1907. Geology of the provinces of Tsang and Ü in central Tibet. Mem. Geol. Surv. India, 36 (2).
Hennig, A., 1915. Zur Petrographie und Geologie von Südwest-Tibet, Vol. 5. In: S. Hedin (Editor), Southern Tibet. Stockholm, 220 pp.
Huntington, E., 1902. The valley of the upper Euphrates River and its people. Bull. Am. Geogr. Soc., 34: 301—310, 384—393.
Innocenti, F., Mazzuoli, R., Pasquare', G., Radicati di Brozolo, F. and Villari, L., 1976. Evolution of the volcanism in the area of interaction between the Arabian, Anatolian, and Iranian plates (Lake Van, eastern Turkey). J. Volcan. Geoth. Res., 1: 103—112.
International Tectonic Map of Europe, 1962. Akad. Nauk, Moscow.
Izbirak, R., 1951. Cilo Daği ve Hakkâri ile Van Gölü Çevresinde Coğrâfi Araştirmalar. Ankara Üniv., Dil ve Tarih-Coğrafya Fak. Yay., No. 67, 149 pp.
Jung, D., Kürsten, M.O.C. and Tarkian, M., 1976. Post-Mesozoic volcanism in Iran and its relation to the subduction of the Afro-Arabian under the Eurasian plate. In: A. Pilger and A. Rösler (Editors), Afar between Continental and Oceanic Rifting. Schweizerbart, Stuttgart, pp. 175—190.
Ketin, I., 1948. Über die tektonisch-mechanischen Folgerungen aus den grossen anatolischen Erdbeben des letzten Dezenniums. Geol. Rundsch., 36: 77—83.
Ketin, I., 1961. Über die magmatischen Erscheinungen in der Türkei. Türk. Jeol. Kurumu Büll., 7: 16—33.
Ketin, I., 1966. Tectonic units of Anatolia. Bull. Miner. Res. Explor. Inst. Turk., 66: 23—34.
Ketin, I., 1969. Über die nordanatolische Horizontalverschiebung. Bull. Miner. Res. Explor. Inst. Turk., 72: 1—28.
Kidd, W.S.F., 1975. Widespread late Neogene and Quaternary calc-alkaline vulcanism on the Tibetan Plateau. EOS, Trans. Am. Geophys. Union, 56: 453.
Kidd, W.S.F., Burke, K.C. and Bewey, J.F., in preparation. Geology of the Tibetian Plateau.
Lahn, E., 1950. La formation gypsifère en Anatolie (Asie mineure). Bull. Soc. Géol. Fr., 20: 451—457.
Lambert, R. StJ., Holland, J.G. and Owen, P.F., 1974. Chemical petrology of a suite of calc-alkaline lavas from Mount Ararat, Turkey. J. Geol., 82: 419—438.
Le Fort, P., 1975. Himalayas: the collided range. Present knowledge of the continental arc. Am. J. Sci., 275-A: 1—44.
McCallien, W.J., 1947. Scotland and Turkey — a comparison. Geol. Soc. Turk. Bull., 1: 91—109.
McKenzie, D.P., 1969. Speculations on the consequences and causes of plate motions. Geophys. J. R. Astron. Soc., 18: 1—32.
McKenzie, D.P., 1972. Active tectonics of the Mediterranean Region. Geophys. J. R. Astron. Soc., 30: 109—185.
Molnar, P. and Burke, K., 1977. Erik Norin Penrose Conference on Tibet. Geology, 5: 461—463.
Molnar, P. and Tapponnier, P., 1975. Cenozoic tectonics of Asia: effects of a continental collision. Science, 189: 419—426.
Molnar, P., Chen, W.P., Fitch, T.J., Tapponnier, P., Warsi, W.E.K. and Wu, F.T., 1977. Structure and tectonics of the Himalaya: a brief summary of relevant geophysical observations. In: Ećologie et Géologie de l'Himalaya. Colloq. Int. C.N.R.S., Paris, pp. 269—294.

M.T.A., 1962a. Explanatory text of the geological map of Turkey, scale 1/500,000: Trabzon sheet. M.T.A., Ankara, 75 pp.
M.T.A., 1962b. Explanatory text of the geological map of Turkey, scale 1/500,000: Diyarbakir sheet. M.T.A., Ankara, 69 pp.
M.T.A., 1963. Explanatory text of the geological map of Turkey, scale 1/500,000: Erzurum sheet. M.T.A., Ankara, 131 pp.
M.T.A., 1964. Explanatory text of the geological map of Turkey, scale 1/500,000: Van sheet. M.T.A., Ankara, 90 pp.
M.T.A., 1966. Explanatory text of the geological map of Turkey, scale 1/500,000: Sivas sheet. M.T.A., Ankara, 116 pp.
M.T.A., 1974. Explanatory text of the geological map of Turkey, scale 1/500,000: Kars sheet. M.T.A., Ankara, 57 pp.
Nawroozi, A.A., 1972. Focal mechanism of earthquakes in Persia, Turkey, west Pakistan, and Afghanistan and plate tectonics of the Middle East. Bull. Seismol. Soc. Am., 62: 823—850.
Norin, E., 1946. Geological explorations in Western Tibet. Publ. 29 of the Sino-Swedish expedition, Stockholm, 205 pp.
Özelçi. F., 1973a. Gravity anomalies of the eastern Mediterranean. Bull. Miner. Res. Explor. Inst. Turk., 80: 54—92.
Özelçi, F., 1973b. Zagros tektonik zonu ve bu zonun Anadolu'da uzanimi ile ilgili jeofizik gözlenimler. Türkiye Birinci Jeofizik Bilimsel ve Teknik Kongre Tebliğleri, pp. 241—249.
Özpeker, I., 1973. Nemrut Yanardağinin Petrojenezi. Thesis. I.T.Ü. Maden Fakültesi, Istanbul, 68 pp.
Powell, C.McA. and Conaghan, P.J., 1973. Plate tectonics and the Himalayas. Earth Planet. Sci. Lett., 20: 1—12.
Powell, C.McA. and Conaghan, P.J., 1975. Tectonic models of the Tibetan Plateau. Geology, 3: 727—731.
Ricou, L.-E., Braud, J. and Brunn, J.H., 1977. Le Zagros. Mém. Soc. Géol. Fr., 8: 33—52.
Seymen, I., 1975. Kelkit vadisi kesiminde Kuzey Anadolu Fay Zonunun tektonik özelliği. I.T.Ü. Made Fak. Yay., Istanbul, 192 pp.
Seymen, I. and Aydin, A., 1972. Bingöl deprem fayi ve bunun Kuzey Anadolu Fay Zonu ile ilişkisi. M.T.A. Dergisi, 79: 1—8.
Shirokova, E.I., 1962. Stresses effective in earthquake foci in the Caucasus and adjacent districts. Izv. Akad. Nauk S.S.S.R., Ser. Geofiz., 10: 809.
Stöcklin, J., 1968. Structural history and tectonics of Iran: a review. Bull. Am. Assoc. Pet. Geol., 52: 1229—1258.
Stöcklin, J., 1974. Possible ancient continental margins in Iran. In: C.A. Burk and C.L. Drake (Editors), The Geology of the Continental Margins. Springer, Berlin, pp. 873—887.
Tanoğlu, A., 1947. Zones d'altitude de la Turquie: détermination et interprétation. Türk Coğrafya Dergisi, 3: 37—55.
Toksöz, M.N. and Bird, P., 1977. Formation and evolution of marginal basins and continental plateaus. In: M. Talwani and W.C. Pitman, III, (Editors), Island Arcs, Deep Sea Trenches, and Back-Arc Basins. A.G.U., Maurice Ewing Ser., 1: 379—394.
Toksöz, M.N., Arpat, E. and Şaroğlu, F., 1977. East Anatolian earthquake of 24 November 1976— Field Observations. Nature, 270: 423—425.
Von Zahn, G.W., 1906. Die Stellung Armeniens im Gebirgsbau von Vorderasien. Ernst Siegfried Mittler und Sohn, Berlin, 89 pp.