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Geometry and Kinematics of Continental Deformation in Zones of Collision: Examples from Central Europe and Eastern Mediterranean

Ali Mehmet Celâl Sengör
University at Albany, State University of New York

Geometry and Kinematics of Continental Deformation in Zones of Collision: Examples from Central Europe and Eastern Mediterranean

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Ali Mehmet Celâl Şengör

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ABSTRACT

Consideration of world-wide epicenter distribution has shown that deformation in continental lithosphere is not narrowly confined to well-defined plate boundaries but is present in wide, diffuse plate boundary zones. Early studies on the seismicity of the peri-Mediterranean area resulted in the division of the lithosphere in that region into a number of small plates, or microplates. Later studies in central Asia, which integrated seismicity with Quaternary geology, indicated, however, that a continuum approach may be more realistic to describe continental tectonics. This study concentrates on geometry and timing of continental deformation that resulted from continental collision in Central Europe and Eastern Mediterranean. In Central Europe continental collision occurred along the Alps during the Lutetian/Priabonian boundary, Foreland deformation in the form of rifting at high angles to the orogen (the Upper Rhine Graben) and strike-slip faulting at about 45° to 60° to the orogen followed the collision. Rifting was nearly synchronous with the collision; strike-slip faulting happened about 20 m.y. after the collision.

In the Eastern Mediterranean the North Anatolian Transform and the Turkish-Iranian Plateau were the main objects of study.

The North Anatolian transform fault is a morphologically distinct and seismically active strike-slip fault which extends for about 1200 km from Karliova to the Gulf of Saros along the Black Sea mountains of N. Anatolia. It takes up the relative motion between the Black Sea and the Anatolian plates, thereby connecting the E. Anatolian convergent zone with the Hellenic Trench through the complex plate-boundary zone of the Aegean. For most of its length, the transform

has a typical strike-slip fault zone morphology, characterized by a narrow 'rift zone,' offset, captured and dammed streams, sag ponds and other deformed morphological features. The fault zone is a broad region of extensively crushed country rock cut by a number of parallel and/or anostomosing strike-slip faults. The transform has periods of seismic activity the last of which, from 1939 to the present, is characterized by frequent $6 \leq M \leq 7$ earthquakes; these are separated by quiet periods of about 150 years. The crust along the fault zone is thinner than normal. The transform probably originated some time between the Burdigalian and the Pliocene and has an offset of about 85 km. Whether the offset of the fault changes systematically along its strike is not known. The North Anatolian transform fault seems to have originated as a consequence of the Arabia-Anatolia collision during the late (?middle) Miocene, when the Anatolian Plate originated and was wedged out into the oceanic tract of the E. Mediterranean from the converging jaws of Arabia and Eurasia to prevent excessive crustal thickening in E. Anatolia. The westerly motion of Anatolia with respect to Eurasia and Africa caused a great change in the tectonic evolution of the eastern Mediterranean, giving rise to the Aegean extensional regime and to internal deformation of Anatolia.

The Turkish-Iranian Plateau (Fig. 5.1) 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 vulcanism is present on the highest portions of the area, despite the absence of a descending slab of lithosphere. Surface geology and vulcanism 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 vulcanism. Minor associated alkaline vulcanism may be due to local longitudinal cracking of the crust to provide access to mantle.

In conclusion, it appears that although the existing mechanical models of continental collision processes satisfy the first-order properties of collision zones they fail to predict the geological (particularly the temporal) details of these areas. Detailed field-mapping rather than attempting to refine the existing theoretical models seems necessary.

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CHAPTER I

INTRODUCTION

The problem of continental vs. oceanic styles of deformation is a new one in the history of tectonic research, perhaps first brought into focus by a careful plotting of world-wide earthquake epicentres after accurate location of these had become possible through WWSSN. Isacks, et al. (1968) noted that in the oceans seismically active regions, the plate boundaries, were narrow and cleanly defined, whereas in the continental areas they seemed to be wide and diffuse, regardless of the nature of the boundary (extensional, strike-slip, or compressional). Prompted by this fundamental observation, McKenzie (1970, 1972) published two syntheses of the circum-Mediterranean region in terms of plate tectonics, using almost exclusively earthquake data. At the time McKenzie thought that the available data could be explained satisfactorily by subdividing, particularly the eastern Mediterranean area, into a number of "small plates."

In spite of the justified criticism of McKenzie's solution by field geologists, it seemed a viable solution to explain the seismicity of continental areas until Molnar and Tapponnier (1975) suggested that a continuum approach to explain continental deformations may be more realistic.

From an entirely different viewpoint, both Hans Stille (1918) and Leopold Kober (1921) had noted that orogenic deformation (Stille's Alpinotype Gebirgsbildung, 1920) is followed in time by block faulting of generally complicated geometry (Germonotype Gebirgsbildung) and they both ascribed this to a consolidation of rocks by orogeny.

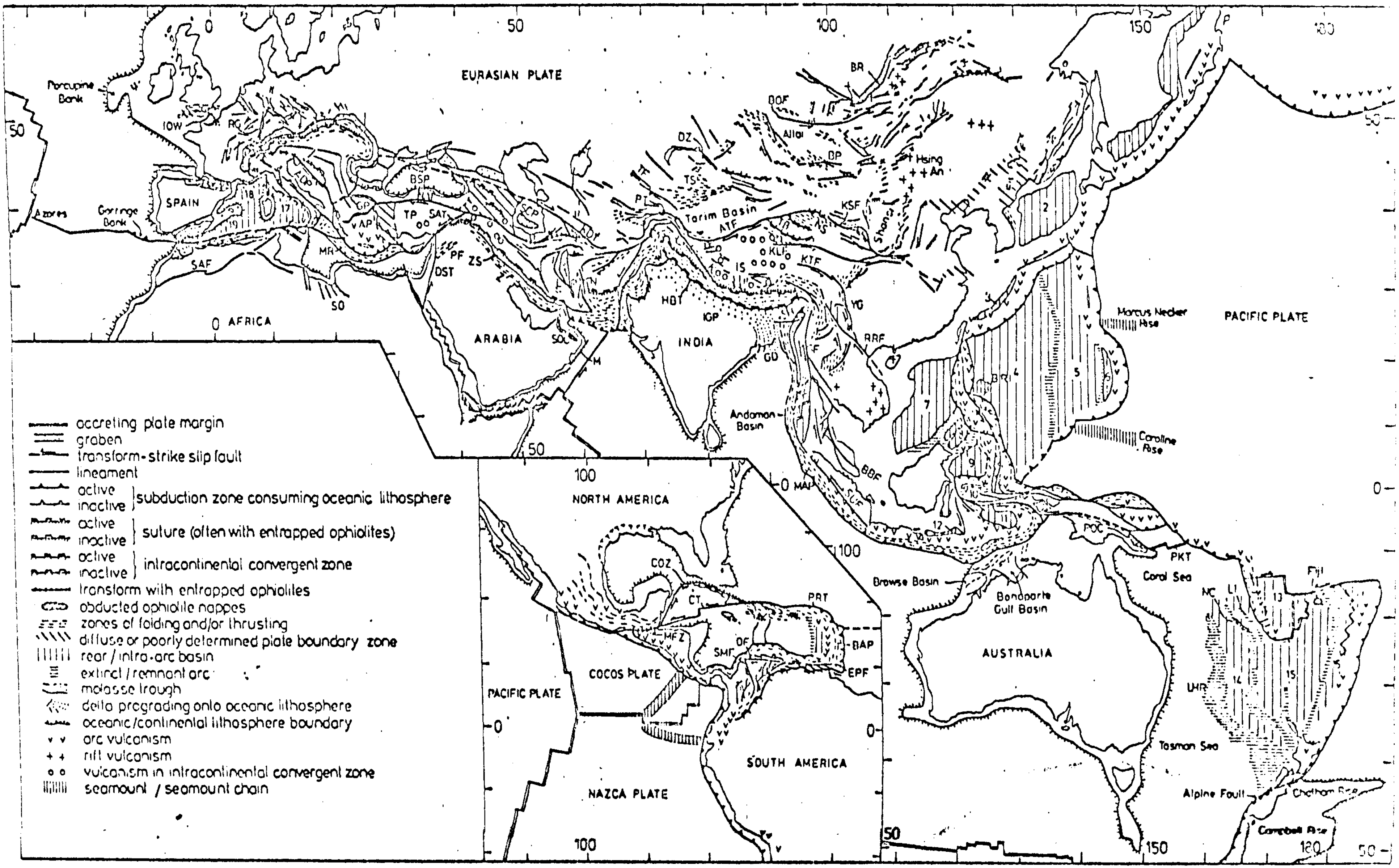
The "plastic" sediments of geosynclines were strongly compressed and intruded during orogeny and converted into "brittle" cratons. Although the observation that alpine-type mountain building is generally followed by german-type mountain building admitted of no argument, the explanation provided by Stille and Kober was far from being satisfactory. More recently Tapponnier and Molnar (1976) indicated, following a suggestion by McKenzie (1972), that at the beginning of continental collision, when regional elevations are generally low, σ_3 will be vertical and thrust- (crustal thickening) dominated deformation will be characteristic, whereas, as the elevations increase as a result of crustal thickening, the lithostatic load will increase as well making σ_2 vertical. At this stage convergence will be accommodated by strike-slip faulting. This generalized scheme is a much more satisfactory explanation of Stille's and Kober's observation than the ill-defined concept of consolidation. Although the basic principle pointed out in Tapponnier and Molnar (1976) is somewhat implicit in Stille's (1918) paper he was extremely vague about the whole thing.

Elegant as they are, the hypotheses put forward by McKenzie (1972) and Molnar and Tapponnier (1975) involve a critical time factor to establish the sequence of events and can only be checked by geological methods. Extremely critical are the temporal relationships between various structures and the evolution of strain-geometries in time. In Central Asia and Tibet we do not as yet have a sufficiently good control on the detailed stratigraphy to be able to date critical events such as the initiation of Altyn Tagh and similar large strike-slip faults or the uplift of the Tibetan Plateau. The purpose of the research reported in this thesis was, therefore, to select regions where continental collision is currently underway, where the widest possible range of collision-related continental deformation styles and associated structures are observed, and, perhaps most important, where there is a good handle on stratigraphy to be able to see detailed temporal relationships among a variety of structures to test the available models for continental deformation styles.

Two regions fulfilled these conditions: the European Alps proper and their extensive foreland in Central Europe and the eastern Mediterranean region. In both areas terminal collision occurred during the Tertiary and continental convergence is currently underway. Both regions have been subjected to detailed geological studies, by academia as well as by industry, during the last century and-a-half and published results of these studies are available. The only serious drawback that has affected this study has been the lack of very detailed, one-to-one correlations between marine and terrestrial successions. This has been less of a problem in the central European case than in the eastern Mediterranean, where correlation problems are just beginning to be addressed.

In Chapter II a very brief account is given of rift-related vulcanism because of the great similarities between mantle-induced "intra-plate" deformation and extensional foreland deformation. Although unique criteria to identify either are rare and/or not readily recognizable, relevant observations are discussed as a useful guide to the arguments presented in the following chapters. Chapter III treats Central Europe and Chapters IV and V are devoted to the post-collisional tectonics of the eastern Mediterranean area (see fig. 1.1). Finally, in Chapter VI the general conclusions of this study are presented with due emphasis on the uncertainties involved.

Figure 1.1 Neotectonics of the Tethyan Orogenic System and its Eurasian fore- and hinterlands. This study deals with the two areas shown in boxes. (From Dewey, 1977, Suture Zone Complexities, Tectonophysics, v. 40).


EURASIAN PLATE
PACIFIC PLATE
AFRICA
ARABIA
INDIA
AUSTRALIA
SPAIN
NORTH AMERICA
SOUTH AMERICA
COCOS PLATE
NAZCA PLATE
PACIFIC PLATE
Porcupine Bank
Azores
Gorringe Bank
Tarim Basin
Andaman Basin
Browse Basin
Bonaparte Gulf Basin
Coral Sea
Tasman Sea
Alpine Fault
Chatham Rise
Campbell Rise
Marcus Necker Rise
Caroline Rise
Altai
Hsing An
accreting plate margin
graben
transform-strike slip fault
lineament
active
inactive
subduction zone consuming oceanic lithosphere
active
inactive
suture (often with entrapped ophiolites)
active
inactive
intracontinental convergent zone
transform with entrapped ophiolites
obducted ophiolite nappes
zones of folding and/or thrusting
diffuse or poorly determined plate boundary zone
rear / intra-arc basin
extinct / remnant arc
molasse trough
delta prograding onto oceanic lithosphere
oceanic/continental lithosphere boundary
arc vulcanism
rift vulcanism
vulcanism in intracontinental convergent zone
seamount / seamount chain

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__________ 1920, Über Art und Alter der variszischen Gebirgsbildungsphasen: Nachr. K. Gesell. Wiss. Göttingen, math.-phys. Klasse, p. 1-7.

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CHAPTER II

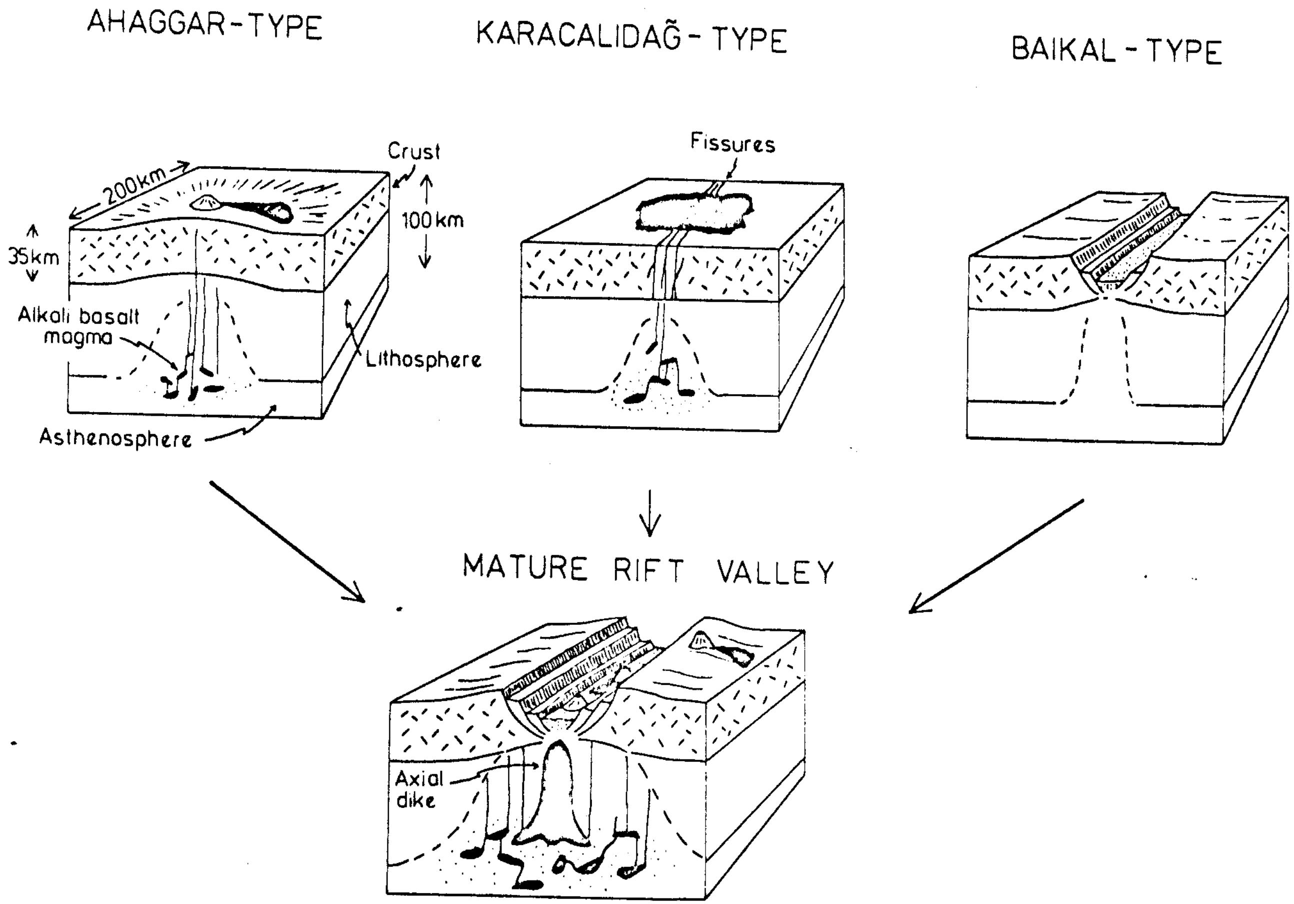
RELATIVE TIMING OF RIFTING AND VOLCANISM ON EARTH AND ITS TECTONIC IMPLICATIONS

Introduction

Rifts, elongate depressions beneath which the entire thickness of the lithosphere has ruptured under extension, occur in diverse tectonic environments that result from the continuous two-dimensional evolution of the multi-plate mosaic of the Earth and also from the interaction between mantle processes and the overlying lithosphere. There is a great variation in the sizes and length-to-width ratios of individual rifts, as well as in their mode of occurrence. Some rifts are seemingly isolated such as that of Lake Baikal (Logatchev and Florensov, 1978), others are aligned along relatively narrow belts such as those of East Africa (Burke and Whiteman, 1973), and yet others are clustered together, in nearly parallel arrangement, in roughly equi-dimensional areas such as those of the Basin and Range Province of the western United States (Atwater, 1970). Although the genetic significance of this variation is far from being clearly understood, it is an empiric conclusion that rifts that occur along 'rift belts' are associated with areas of primary extension, whereas 'rift clusters' are more characteristic of wide, imperfect strike-slip regimes. Isolated rifts occur in both environments.

A feature common to most rifts is basaltic volcanism. In continental rifts, with which we are here concerned, these basalts are mainly alkaline. The relative timing of basaltic volcanism with respect to the associated rifting shows considerable variation. Currently there are two main hypotheses attempting to explain the origin of basaltic magmas beneath rifts: one favors the preliminary cracking of the lithosphere due to differential stresses resulting from two-dimensional plate evolution (e.g., membrane stresses: Turcotte and Oxburgh, 1973; Turcotte, 1974; stresses due to the collision of continents: Molnar and Tapponnier, 1975; Şengör, 1976; stresses resulting from the propagation of existing accreting plate margins into continents: McKenzie and Weiss, 1975) that upsets the T/P balance of the underlying mantle resulting in its partial melting. This results in a volume increase in the partially melted area and may induce a post-rifting uplift of the overlying lithosphere as happened in the Upper Rhine Graben (Şengör, et al., 1978). The other view holds a complicated convection pattern in the mantle responsible for doming and cracking the lithosphere thereby giving rise to extensional fractures and eventually to rifts (Burke and Whiteman, 1973; Burke and Dewey, 1973). In the former view, the expected sequence of events is rifting--(?uplifting)--volcanism, whereas in the latter it is doming-volcanism-rifting (Fig. 1). It is our contention that both hypotheses are compatible with the available data, but are applicable to fundamentally different tectonic environments. The first hypothesis is applicable to regions where rifting is the result of horizontal movements of plates and their interaction in which mantle plays a passive role. The second hypothesis seems to be valid where (?small scale) convection in the mantle (McKenzie and Weiss, 1975) directly affects the overlying lithosphere and induces rifting as a combined result of primary vertical tectonics (uplift) and (?later) horizontal motion (initial spreading). It appears that this latter mode of rifting occurs mainly on plates that are fixed with respect to

Figure 2.1 Idealized and simplified block diagrams showing the evolution of a 'mature rift valley' from three different initial configurations. In Ahaggar-type rift valley evolution mantle is active and induces uplift and subsequent rifting. In this case volcanism is likely to predate rifting. In the Baikal-type mantle is passive and there is no pre-rifting doming or volcanism. Karacalidag-type was added to the diagram to illustrate a complexity: horizontal extension forms fissures through which magma wells up prior to major downfaulting and rift formation. Hence, in the geological record the sequence of events in the formation of a Karacalidag-type rift will look like those of an Ahaggar-type rift, although the mode of rifting was that of a Baikal-type rift. The end products of all rift processes are likely to look very similar.


AHAGGAR-TYPE
KARACALIDAĞ-TYPE
BAIKAL-TYPE
200km
Crust
100km
35km
Alkali basalt magma
Lithosphere
Asthenosphere
Fissures
MATURE RIFT VALLEY
Axial dike

the underlying mantle, as Africa has been suggested to have been since 25 m.y. ago (Burke and Wilson, 1972). At present, the former mode of rifting is by far the more widespread of the two.

Passive Mantle Hypothesis

The horizontal movement of plates and their interaction give rise to differential stresses within the lithosphere that result in rifting. Turcotte and Oxburgh (1973) and Turcotte (1974) argued that membrane stresses can be generated within a plate moving longitudinally on the surface of the Earth, due to the ellipticity of the Earth, and result in rifting. Burke and Dewey (1974), however, indicated that the elastic stresses thus generated cannot be stored for long periods of time and therefore may not be as important in generating rifts as Turcotte and Oxburgh (1973) suggested. Moreover, the supporting evidence that Oxburgh and Turcotte (1974) used to apply this idea to the East African rifts consisted of incorrect K/Ar data and misinterpretation of palaeomagnetic results.

Molnar and Tapponnier (1975) and Şengör (1976) pointed out that continental collision, especially of irregular margins, also generates extension within continental plates resulting in rifting more or less perpendicular to the direction of convergence; if collision-induced strike-slip faulting also causes rift formation, as is the case for the Baikal rift (Sherman, 1978), such rifts are not likely to have any systematic orientation with respect to the convergence direction. In the foreland of the Alps in Europe, for example, rifting events are well-correlated with the Mesoalpine collision (Trümpy, 1973; see Şengör, 1976, table 1). Here rifting appears to have predated major basaltic volcanicity. In the case of the Upper Rhine Graben, rifting can be shown to have started during medial Eocene with some minor volcanicity of 42 m.y. age along the master faults (Şengör, et al., 1978). Major basaltic volcanic activity began at the northern end of the graben about 25 m.y. ago (Lotze, 1974). This volcanic center, the Vogelsberg, appears to have continued its activity until very recently. In the south, well within the major graben trough, the Kaiserstuhl volcano began its activity about 18 m.y. ago; this however, unlike that of the Vogelsberg, was a short-lived event. In the Lower Rhine Graben there are few volcanics, but the maars of Eifel appear to be related to Plio-Pleistocene NW-trending faults (Greiner and Illies,1977). In Bohemia, both along the Thuringian disturbance and in the grabens between the Bohemian and the Thuringian blocks (Schollen) volcanicity began during the Oligocene and lasted through the Pliocene; rifting along these lines had begun at the beginning of the Tertiary and accelerated during late Eocene (Lotze, 1974). It appears that in Europe it is this kind of intra-plate but inter-Scholle activity that caused basaltic volcanism that is largely of olivine-nepheline bearing alkaline type.

Western Turkey and the Basin and Range Province of the western United States are areas of extensive rift development, characterized by numerous sub-parallel grabens in 'rift clusters'. Both areas are related to large, imperfect strike-slip plate boundary zones (western Turkey: Dewey and Şengör, 1979; western U.S.: Atwater (1970). In western Turkey rifting began during the late Miocene and is currently active. Most rifts here are devoid of volcanics with the exception of the Gediz and Simav grabens (Dewey and Şengör, 1979). On the northern shoulder of the Gediz graben, an area, 50 km long and 20 km wide, is covered with Pliocene to recent alkaline basalts that are nepheline, leucite, and hornblende bearing (kulaite, Washington, 1894; Erinc, 1970). These basalts emanated from fissures that can be shown to be controlled by faults related to the Gediz graben. Zeschke (1954) documented a similar history for the Simav basalts.

In the Basin and Range Province major rifting related to the present regime began about 18 m.y. ago (Noble, 1972) and was closely followed by basaltic/rhyolitic bimodal volcanism (McKee, et al., 1970). The geographic extent and the close temporal association with a strike-slip regime (Hamilton, 1970) of the western Siberian rift system of Triassic age (Logatchev, 1977) suggest that it might be a fossil analog of the Basin and Range-type rift regimes, where generally basaltic volcanism postdates rifting.

Active Mantle Hypothesis

Rifts active on the African Plate can be interpreted as products of relatively simple mantle-lithosphere interaction. The control appears to be the lack of relative motion between the African Plate and underlying convective circulation over the last 25 m.y. (Burke and Wilson, 1972). The interaction beneath the African Plate shows itself in several phenomena: the distinctive basin and swell structure; the occurrence of intraplate volcanism (largely on swell crests); the development of rifts and the evolution (in the Red Sea and in the Gulf of Aden) of rifts into oceans (see Burke, 1977 for a review).

Throughout southern Africa swell crests carry no volcanoes and rifts are mainly not on swells but are reactivated old structures. This indicates that a sequential development from volcanoes on swells to rifts such as Burke and Whiteman (1973) distinguished is not universally recognizable. Further evidence that rift development is complex can be seen in the active rifts of East Africa. Volcanism, there, is very unevenly distributed along the length of the rifts and in some areas (for example, near Addis Ababa) is dominantly tholeiitic, whereas in other areas it is dominantly alkaline (see, for example, Baker, et al., 1972). The implication is that the petrology of the igneous rocks is a very poor indicator of rift style compared with more direct structural/stratigraphic features such as topography, faulting and sediment fill. The general absence of signs of interaction between magma and continental crust in the rift igneous rocks is perhaps the most significant feature and is to be expected in a regime where axial dikes and extension dominate.

Older episodes of rifting induced by similar mantle interaction are hard to identify but the lack of motion between Africa and the spin axis during the break-up of Pangaea has been taken as evidence that African rifting during that time was similarly induced (Burke and Dewey, 1973). The Pangaean-rupturing rifts on either side of the Atlantic formed just before that ocean opened provide further examples of the diversity of styles of volcanism. Some rifts appear to be without volcanic material as are the majority of the rifts of western Turkey; others are associated with extensive pre-rift igneous activity and yet others show igneous activity only when the rifts are well-developed. As in East Africa now, compositional diversity is the rule. In some areas carbonatites and alkaline syenites abound (e.g., Los, Bagbe, Songo), whereas in other tholeiitic basalts are the only igneous rocks.

Conclusions

Rift studies are now at a very exciting stage. Tentative classifications of rifts and rift systems based on global tectonic hypotheses such as that in this paper are being made and a major need is for comprehensive study of all aspects of selected rifts. They are exceedingly complex structures and timing of events in rifts is a difficult task (see Fig.2.1 for a simplified sketch of rifting-volcanism relative timing types and a few of the involved complexities). Some rifts are particularly well-suited for the study of some properties--for example, igneous petrology in the Keweenawan--while others are better adopted for the study of the properties such as subsidence and sediment fill (e.g., the North Sea rifts). A unified approach recognizing that these phenomena are all related to similarly induced processes is likely to prove most rewarding.

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CHAPTER III

RIFTS AT HIGH ANGLES TO OROGENIC BELTS: TESTS FOR THEIR ORIGIN AND THE UPPER RHINE GRABEN AS AN EXAMPLE

Introduction

Although the association between the world's orogenic belts and grabens that strike into them at high angles was recognized early during this century (Weber, 1921, 1923, 1927; Shatski, 1946a, b, 1947, 1955) and later reemphasized (DeSitter, 1956; J.T. Wilson *in* Jacobs, Russel, and Wilson, 1959; Wilson, 1966a), their genetic relationships remained blurred or were denied altogether (Cloos, 1939; see a summary of various objections to Weber's views *in* Weber, 1927) until the development and wide acceptance of the theory of plate tectonics and its geologic corollaries. Burke and Whiteman (1973) showed that, in East Africa, Neogene rifts commonly developed a three-armed pattern that they compared to the pattern of the rifts that had led to the opening of the South Atlantic Ocean establishing an apparently common sequence as follows: doming, development of three- (in some cases more) armed rift systems on the crests of the domes, continental break-up as two of the three-armed rifts link up and develop into an accreting plate margin, while the third arm is left as a graben or a "failed arm." Burke and Dewey (1973) argued, on the basis of their survey of 52 rift systems throughout the globe, that most of the ocean-opening cycles, both young and old, have been initiated by and evolved through the sequence that Burke and Whiteman (1973) described; further studies at oceanic and fold belt margins seem, generally, to have confirmed their views (for example, Curray and Moore, 1974; Burchfiel and Davis, 1975; R.C.L. Wilson, 1975; Rankin, 1976). Burke and Dewey (1973), Dewey and Burke (1974), and Hoffman, Dewey and Burke (1974) suggested that, when an ocean closes, the previously "failed arms" become rifts at high angles to the resulting collisional orogen or, in Shatski's (1946a) terminology, aulacogens[1]. Recently, however, Şengör (1976a, b) proposed that continental collision also produces rift structures very similar to aulacogens in map view and internal geometry, with the important difference that the origin of aulacogens predates (whereas that of collision rifts postdates) ocean closing and the resultant collisional orogeny.

In two recent reviews, Burke (1976, 1977) emphasized the importance of rift structures that strike into mountain belts at high angles, herein called high-angle rifts irrespective of their mode of origin. He pointed out their generally much less severely deformed state when compared with that of the adjacent orogen and indicated that if these rifts predate the orogeny and are related to the initial ocean opening phase they then may preserve the valuable syn-rifting stratigraphic record of the vanished ocean whose early record is generally obliterated by the intense collisional orogeny. If, however, the rifts are collision-induced, their peculiar stratigraphy is unlikely to have any relation to that of the obliterated ocean. It is, therefore, of great interest in tectonics to be able to distinguish between aulacogens and collision grabens.

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[1]In this chapter, I follow Burke's (1977) practice of calling aulacogens only those high-angle rifts striking into fold belts that had originated during and in relation to the ocean opening phase. Rifts that originate during a collision could perhaps be called impactogens.

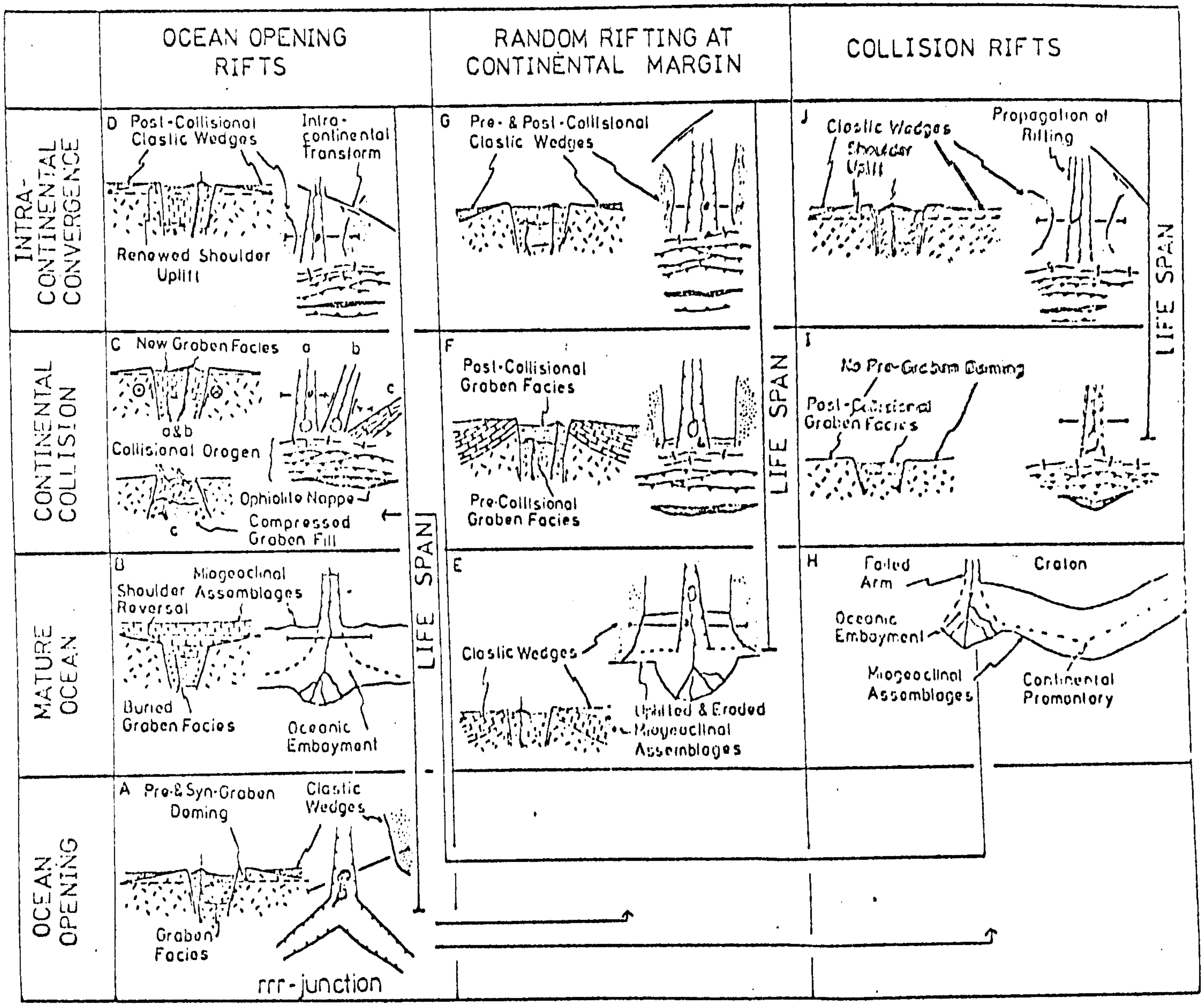
The purpose of this chapter is to outline briefly expected differences, primarily in the stratigraphic and structural evolution, between aulacogens and collision rifts, to point out those differences that may be decisive and those that are likely to be deceptive, and, finally, to describe the evolution of one well-studied example of a high angle rift, the Upper Rhine Graben of Central Europe.

Differences in Geological Evolution Between Aulacogens and Collision Rifts

Burke (1977) suggested that a study of the date of origin of a high-angle rift might reveal its origin, because it would show whether the rift predated or postdated the collisional orogeny. There, are, however, many complications in the evolution of a high-angle rift that warrant a closer examination of the life histories of aulacogens and collision rifts. In order to emphasize the point that aulacogens and collision rifts are not the only types of high-angle rifts, we include the discussion of the life history of one kind of "random high-angle rift" (fig.3.1E-G) and attempt to show how it differs from the other two.

Figure3.1 is a composite diagram that schematically illustrates events during the origin and evolution of an aulacogen, a collision rift and a rift that randomly originates at or near the continental margin after the initial rifting and opening of the associated ocean. In the case of an aulacogen (fig.3.1A-D), rifting is likely to be predated by doming at the future site of formation of an rrr-junction, possibly induced by a mantle plume (Burke and Whiteman, 1973; Burke and Dewey, 1973; Burke and J.T. Wilson, 1976). It may be possible to document and date this doming event by means of peripheral clastic wedges. For

Figure 3.1 A schematic illustration of events during the origin and evolution of an aulacogen (A-D), a random rift at high angle to continental margin (E-G), and a collision rift or impactogen. Cross sections show the expected differences in the stratigraphic evolution of the three kinds of high-angle rifts. For more detailed discussion, see the text.


OCEAN OPENING RIFTS
RANDOM RIFTING AT CONTINENTAL MARGIN
COLLISION RIFTS
INTRA-CONTINENTAL CONVERGENCE
CONTINENTAL COLLISION
MATURE OCEAN
OCEAN OPENING
D Post-Collisional Clastic Wedges
Intra-continental Transform
Renewed Shoulder Uplift
C New Graben Facies
a
b
c
a&b
Collisional Orogen
Ophiolite Nappe
Compressed Graben Fill
B
Miogeoclinal Assemblages
Shoulder Reversal
Buried Graben Facies
Oceanic Embayment
A Pre-& Syn-Graben Doming
Clastic Wedges
Graben Facies
rrr-junction
LIFE SPAN
G Pre- & Post-Collisional Clastic Wedges
F
Post-Collisional Graben Facies
Pre-Collisional Graben Facies
E
Clastic Wedges
Uplifted & Eroded Miogeoclinal Assemblages
LIFE SPAN
J
Clastic Wedges
Shoulder Uplift
Propagation of Rifting
I
No Pre-Graben Doming
Post-Collisional Graben Facies
H
Failed Arm
Craton
Oceanic Embayment
Miogeoclinal Assemblages
Continental Promontory
LIFE SPAN

example, the Miocene clastics of Western Kenya and Uganda were produced by accelerated erosion during arching over the sites of the future western and eastern rifts during the early Neogene and were later preserved beneath the younger volcanics (King, 1970). Similarly, R.C.L. Wilson (1975) presented strong evidence for late Jurassic uplift on the site of the northwestern corner of the Iberian Peninsula, where grabens developed at the end of the Jurassic before the opening of North Atlantic and the Bay of Biscay. A doming event may be accompanied by volcanicity (for example, Kenya, King and Chapman, 1972) and eventually followed by crestal rifting (for example, Kenyan and Ethiopian rifts in East Africa, Baker, Mohr and Williams, 1972). Within the rift, clastics in the form of fanglomerates, arkoses and mudstones, and locally evaporites, the 'graben facies' of Bird and Dewey (1970), may form. The margin fanglomerates are particularly useful in stratigraphic analysis because they record the successive levels of exposure on the graben shoulders, thereby complementing the syngraben doming record of the clastic wedges outside the graben. Fanglomerates are also useful in recording tectonic "pulses" during the evolution of the graben (Steel, 1976).

As two of the arms of the triple rrr-junction evolve into an accreting plate boundary, an ocean will develop, and the newly-formed continental margin will subside following a time-dependent cooling curve (Sclater, Anderson and Bell, 1971; Hays and Pitman, 1973). A miogeoclinal assemblage accumulates on the subsiding margin that will probably be thicker in the failed arm than in the adjacent continental margin and will overlie the graben facies well into the craton (Salop and Sheineman, 1969). Also by this time, the initially uplifted shoulders of the failed arm will reverse and subside (Hoffman, Dewey and Burke, 1974).

As pointed out by Burke and Dewey (1973), failed arms are likely to be located at continental embayments (for example, Gulf of Guinea embayment and the Benue rift, Burke, Dessauvagie and Whiteman, 1971) and localize the continental drainage thereby creating large clastic deposits in front of them in the form of deltas (for example, the Niger Delta as localized by the Benue Trough, Burke, 1972). Such anomalously thick, clastic deposits, perhaps overlying oceanic basement, may be identified in the stratigraphy of the collisional orogen, which, at that locality, may be at a "reentrant" due to the original embayment and thus reveal the pre-collisional age of the associated high-angle rift. (The most familiar example is in the Mackenzie Mountains of Canada, see fig. VIII-2 of Douglas and others, 1970).

As shown in figure 3.1C, failed arms may strike at any angle to the continental margin (a, b, c), but not all of them may survive through the collision to become aulacogens: those that are highly oblique to the colliding margins (c) are likely to undergo compressional deformation with minor strike-slip displacement during the collision: an example of this kind of a compressed, oblique aulacogen is represented in the present Timan Mountains (Kraus, 1972; Siedlecka, 1975). Compression due to shoulder reversal may also occur (see above), but it is unlikely to produce strike-slip displacement along the graben axis. Failed arms that strike at high angles to the collision front may be reactivated as rifts with a strike-slip component. Failed arms that are perpendicular to the collision front would be reactivated as purely extensional features. During the collisional reactivation and further evolution during intra-continental convergence (fig. 3.1C, D), the shoulders of the aulacogen are likely to rise again to provide a post-collisiohal clastic wedge to the periphery of the resulting arch; in the rift trough, a new graben facies will develop above the preserved miogeoclinal assemblages (fig. 3.1D).

In sharp contrast to aulacogens, collision grabens do not necessarily have an associated pre-graben doming, neither do they have any stratigraphic relation to the adjacent mountain belt (with the possible exception of the exogeosynclinal sequences, such as a possible correlation between the Middle Pechelbronn beds of the Upper Rhine Graben and the lowest units of the Lower Marine Molasse of the Alps). Their prerifting basement is composed of the rocks of the orogenic foreland, and the overlying graben facies is clearly post-collisional. Although they begin as purely extensional structures (Şengör and Göçmen, in preparation), a subsequent change in the orientation of the convergence vector across the collision zone may impose a late strike-slip component onto the structure as happened to the Upper Rhine Graben during Miocene and mainly, Pliocene times (Illies, 1974a; see below). Post-rifting doming will give rise to clastic wedges, which, along the correlative fanglomerates of the graben fill, would reveal the age of onset of doming (fig. 3.1H-J).

As seen from the above discussion, the decisive test as to whether or not a high-angle rift is an aulacogen would be the detailed study of the stratigraphy of the graben fill; however, unless the graben is subsequently deformed, its pre-graben basement and the early sedimentary infill are generally not accessible to field examination. Even if the sedimentary record is revealed, either by drilling or by geophysical methods, it is often non-marine and therefore difficult to date. Therefore, the identification, in the stratigraphic record of the adjacent mountain belt opposite a high-angle rift of an exceptionally large delta and a deflection of the orogen toward the rift at this spot may be used as another, perhaps more practical, criterion from the viewpoint of the field geologist, for identifying the graben as an aulacogen. Early, pre-graben doming and early syncollisional strike-slip movement along the rift are additional but risky criteria, as they may have other causes and should be used in conjunction with stronger evidence to identify a high-angle rift as an aulacogen.

Random rifting at or near a continental margin after the initial opening either as a result of a mantle plume or strike-slip tectonics parallel to the margin or as a result of membrane stresses (Turcotte and Oxburgh, 1973; Turcotte, 1974) may further complicate the picture (fig. 3.1E-G). In this case, the pre- (or syn- and/or post-, depending on the mode of origin) rifting doming will result in the erosion of the miogeoclinal strata, and the graben facies will be deposited on a thinner-than-normal miogeoclinal assemblage, as opposed to an aulacogen's thicker-than-normal, and a collision rift's altogether absent miogeoclinal sediment content. Clastic wedges associated with such a rift would be deposited unconformably over the miogeoclinal strata. However, if a failed arm should be subjected to a rejuvenation before the collision, then it may be extremely difficult, if not impossible, to establish its exact date of origin; a line of alkaline intrusives aligned along the graben axis and predating the later rejuvenation may give clues to the age of original rifting.

As a result of the difficulties involved in the study of grabens, with the exception of those that contain economic resources, most rift structures are rather poorly known. Therefore, especially in the case of pre-Phanerozoic high-angle rifts, it may be extremely difficult to deduce the nature of the relation between the graben and the adjacent orogenic belt. In the following sections we describe the evolution of a high-angle rift, the Upper Rhine Graben of Central Europe, which has variously been interpreted as having evolved as a result of doming and key-stone collapse during the Tertiary without relationship to the Alpine Orogeny (Cloos, 1939; Burke and Dewey, 1973); as a collision rift related to the Alpine Orogeny (Illies, 1947a, b; 1975; Şengör, 1976a, b); as a drag feature related to strike-slip faulting (Molnar and Tapponnier, 1975); and as an aulacogen, originated in the Triassic and reactivated by the Alpine Collision (Burke, 1977). The long history of geological investigation and rich economic resources of the Rhine Graben provides an extensive data base that allows us to show it to be collision induced. Recent efforts of German, Swiss, and French geoscientists, under the Upper Mantle and Geodynamics projects, have greatly contributed to our understanding of the evolution and present behavior of the Rhine Graben. Their results are published largely in three recent compendia (Rothé and Sauer, 1967; Illies and Müller, 1970; Illies and Fuchs, 1974), which make up a large portion of our data base.

Permian to Middle Eocene Development of the Upper Rhine Graben Area

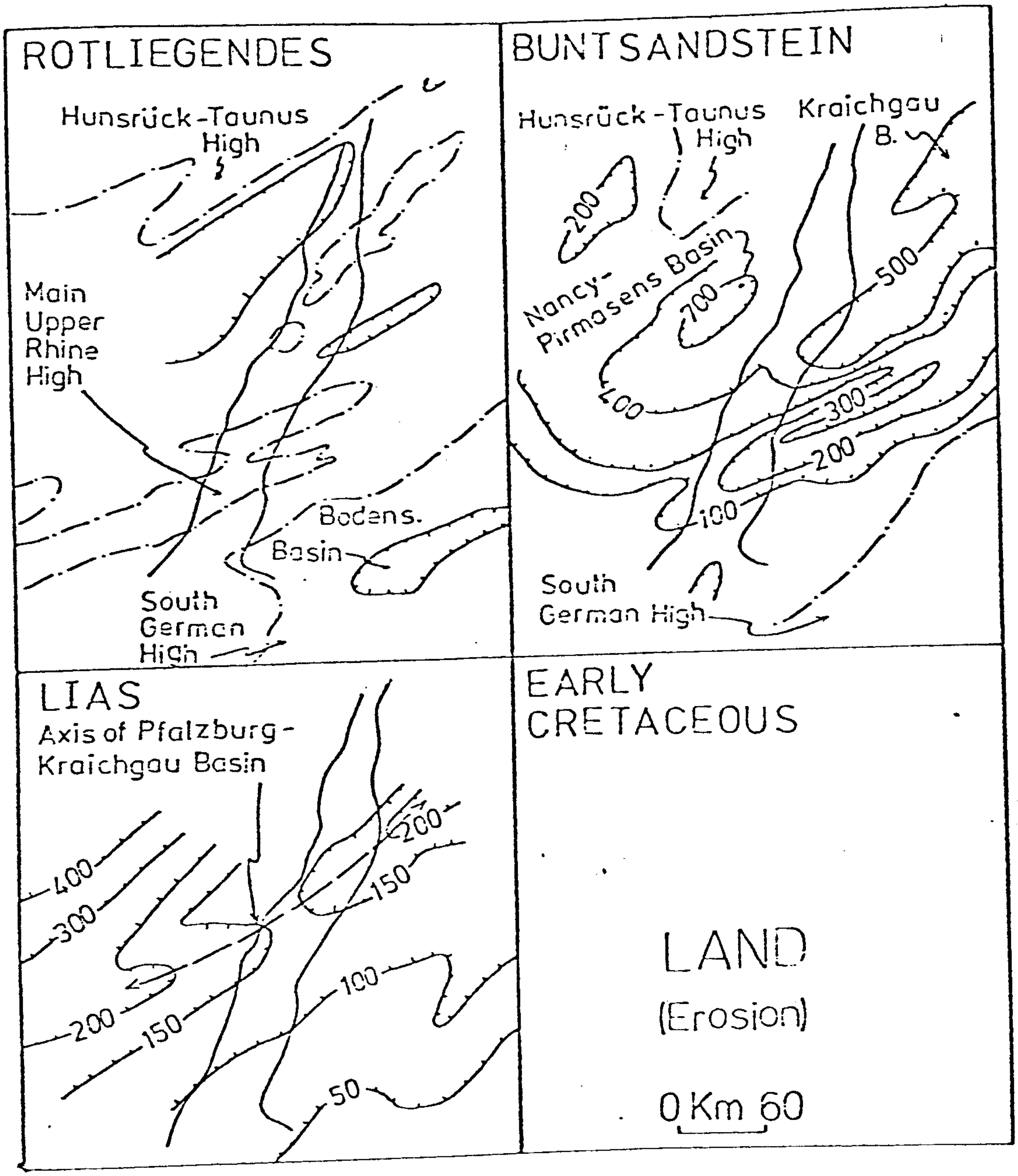
If the Upper Rhine Graben is an aulacogen related to the Alpine Orogen, it should have a rifting history that would date back to the initial rifting in the Alpine area during the Triassic. In this section we describe the stratigraphy and gross basement structure of the Rhine Graben area from Permian to the end of the Mesozoic to show that there was no sign of a rift structure, a "failed arm," on the site of the future Rhine Graben.

Throughout its entire length of about 300 km, from the Jura Mountains to the young volcanic center of Vogelsberg, the Upper Rhine Graben is located mainly in the Rhenoherzynikum and Saxothuringikum zones of the Central European Variscides; the age of deformation in these zones ranges from late Devonian to Stephanian (Lotze, 1974).

The orogenic development of the Central European Variscides was largely completed during the late Pennsylvanian. By Rotliegendes (Early Permian) times, various east-northeast striking basins (for example, Saar-Nahe Basin, Kraichgau Basin) appeared and received largely clastic sediments with occasional intercalated volcanics (Boigk and Schöneich, 1970, 1974; Lotze, 1974; P.A. Ziegler, 1975). It is important to emphasize that all isopachs of the Rotliegendes deposits cross the future site of the Upper Rhine Graben indiscriminately (fig. 3.2). To the south of the Rhine Graben, in the Jura Mountains, continental Permian rests on Upper Carboniferous with angular unconformity.

The number of control points for the construction of the distribution of Zechstein sediments is unfortunately too few to permit an isopach analysis; however, from the existing data (Boigk and Schöneich, 1970, 1974; Lotze, 1974) and from P.A. Ziegler's (1975) paleogeographic map of the Zechstein, the following conclusions may be drawn: the sea of the Permian Basin in northern Germany extended a tongue in the direction of the present Rhine Graben, in which marls, carbonates and evaporites with local salt deposits were laid down. At about this time, the site of the future Hessen Graben appears to be a preferred site of subsidence, although not as large as the future Upper Rhine Graben. Moreover, the

Figure 3.2 The depositional framework of the Upper Rhine Graben area from Rotliegendes (Early Permian to Early Cretaceous). In Rotliegendes time only the 0 m isopach (dashed line with dots between dashes) and the 1000 m isopach (line with hachures) are shown, except around the Bodensee Basin where the hachured line shows the 100 m isopach. Figures are always in meters (data from Boigk and Schöneich, 1970, 1974; Hoffmann, 1967; P.A. Ziegler, 1975).


ROTLIEGENDES
Hunsrück-Taunus
High
Main
Upper
Rhine
High
Bodens.
Basin
South
German
High
BUNTSANDSTEIN
Hunsrück-Taunus
High
Kraichgau
B.
200
Nancy-
Pirmasens Basin
700
500
400
300
200
100
South
German High
LIAS
Axis of Pfalzburg-
Kraichgau Basin
400
300
200
150
200
150
100
50
EARLY
CRETACEOUS
LAND
(Erosion)
0 Km 60

observation that the fold axes of the Variscan structures plunge beneath this depression rather than being truncated by it (Hedemann, 1957; Schenk, 1974) makes it likely that the depression was not necessarily a graben but more of a downbending, like some of the Neogene depressions of Anatolia (Ardel, 1965).

When the Mesozoic Era opened, sedimentary basins in Central Europe were separated from the Tethyan Ocean by the South German--Vindelician High (Lotze, 1974; P.A. Ziegler, 1975). In the Upper Rhine Graben area, northeast and east-northeast directed structural trends, such as the Nancy-Pirmasens Basin and the Kraichgau Basin (fig. 2), generally following the orogenic trend of the underlying Variscan basement, are dominant (Boigk and Schöneich, 1970, 1974). On a large scale isopach map (Boigk and Schöneich, 1970, fig. 2), a narrow (~15 km), north-northeast directed deepening of about 100 m is seen around Basel, related to the Burgundian Basin in the western portion of the future Alpine Tethys (Boigk and Schöneich, 1974).

During Muschelkalk and Keuper times the east-northeast trends of the post-Variscan basins persisted. The deepening, of probably around 75 to 100 m (Boigk and Schöneich, 1970), that extended to Basel during Buntsandstein was extended as far as Strassburg following the trend of the Rhine Graben.

In the Jura Mountains, Buntsandstein rests unconformably on Permian and the entire Triassic, in typical Germanic Facies, has a total thickness of just more than 600 m. This thickness is comparable to that of the other Triassic basins in Central Europe (for example, Franken/Oberpfalz $\cong$845 m; Lotze, 1974) and does not indicate any major graben subsidence.

During the Lias (fig. 3.2), the north-northeast-trending "tongue" of weak depression was completely lost, and all isopachs again cut across the future site of the Upper Rhine Graben indiscriminately (Boigk and Schöneich, 1970; Hoffman, 1967). The Lias is a very important time in the evolution of the adjacent Alpine Ocean, as this is the time of the first major transgression in the Alpine area (Trümpy, 1960). This transgression is most pronounced in the Alpine area proper and rapidly loses its "geosynclinal" expression toward the European platform and also toward the Jura Mountains.

Later in the Jurassic, the area surrounding the future Rhine Graben gradually shallowed (P.A. Ziegler, 1975). Although syn-sedimentary faulting in this area during Oxfordian times has been suspected, it has not been demonstrated, and the arguments for its existence are not convincing (Breyer, 1974). During the early Cretaceous, the surroundings of the Rhine Graben became dry land (fig. 3.2) (Umbgrove, 1947; P.A. Ziegler, 1975). This so-called Rhine Shield (Cloos, 1939) was not confined, however, only to the future Upper Rhine Graben area but encompassed a wide region extending from Holland to the Bohemian Massif (P.A. Ziegler, 1975, fig. 16). Toward the end of the Mesozoic Era, volcanism began to affect various parts of Central Europe, including the Vosges, Schwarzwald, Mainz Basin and the Taunus (Illies, 1947a). A majority of this activity is today preserved in the form of plugs and dikes that give ages in the range of 90 to 100 m.y. (Illies, 1974a). No major uplift, similar in magnitude to those observed in Africa (Burke and Whiteman, 1973), is reported to have accompanied these eruptions, although regional warping, tilting and jostling of rigid to semi-rigid blocks (Schollen), giving rise to local intense deformations (for example, the remarkable Osning overthrust zone, Stille, 1953) have occurred since the Late Jurassic in Central Europe (Lotze, 1953, 1974; Keller, 1976). It is uncertain whether this period of magmatic activity, mainly of olivine-nepheline-bearing basic rocks (Illies, 1974a) can be related to this kind of deformation.

In summary, the Mesozoic Era, when the Alpine Ocean was born and later during Cenomanian began to contract (Dewey and others, 1973; Dietrich, 1976), closed with no indication of any kind of major graben subsidence on the site of the future Rhine Graben. The Mesozoic strata "bear no relation to the graben" (Sittler, 1969, p. 545). They thicken from the south to the north in Europe, probably due to the influence of the Vindelician High. During the early Cretaceous, Central Europe became land, and later in the Cretaceous volcanicity appeared accompanying faster uplift of the northern sector of Central Europe (Sittler, 1969).

Lutetian to Recent Evolution of the Upper Rhine Graben

In this section we present data that indicates that the Upper Rhine Graben originated during the Lutetian as a result of the Mesoalpine collision (Şengör, 1976b). The Tertiary history of the Upper Rhine Graben area is much clearer than the Mesozoic. The initial downfaulting of the present Rhine Graben began in the south and is indicated by the conglomerates of the Siderolitikum of probable Lutetian age (Doebl, 1970). This initial downfaulting was accompanied also by mafic volcanicity along the master faults of the rift, which volcanics yield ages at around 48 m.y. (Illies, written commun.). Drilling and geophysics (gravity and magnetics) show that the graben floor inherited the east-northeast-trending trough and sill structure of the post-Variscan basement, and the initial infill of the graben also shows variable thickness and facies within different sub-basins (Sittler, 1969, figs. 3 and 4). By Priabonian (late Eocene) times, subsidence had accelerated in the south (Sittler, 1969; Illies, 1947b), and the Limnea marls of fresh water origin are up to about 900 m thick to the southwest of Freiburg, 500 m thick near Karlsruhe, and wedge out near Mannheim (Illies, verbal commun., 1977; fig. 3.3B). In early Oligocene times, the Pechelbronn beds, generally of terrestrial origin with the exception of the Middle Pechelbronn strata laid down by a marine incursion, were deposited. The Pechelbronn beds extend into the Hessen Graben, which had become a graben during the Eocene (Illies, 1974b), where they are known as the Melania Clay (Lotze, 1974). During Pechelbronn times along the master faults of the Rhine Graben in the south are the so-called Kustenkonglomerate (coastal conglomerates) that indicate not only the activity of the faults here, but also the post-rifting uplift of the graben shoulders. Early pebbles of this series contain Jurassic material (Illies, 1967), showing that the major uplift of the graben shoulders occurred after faulting (Illies, 1970).

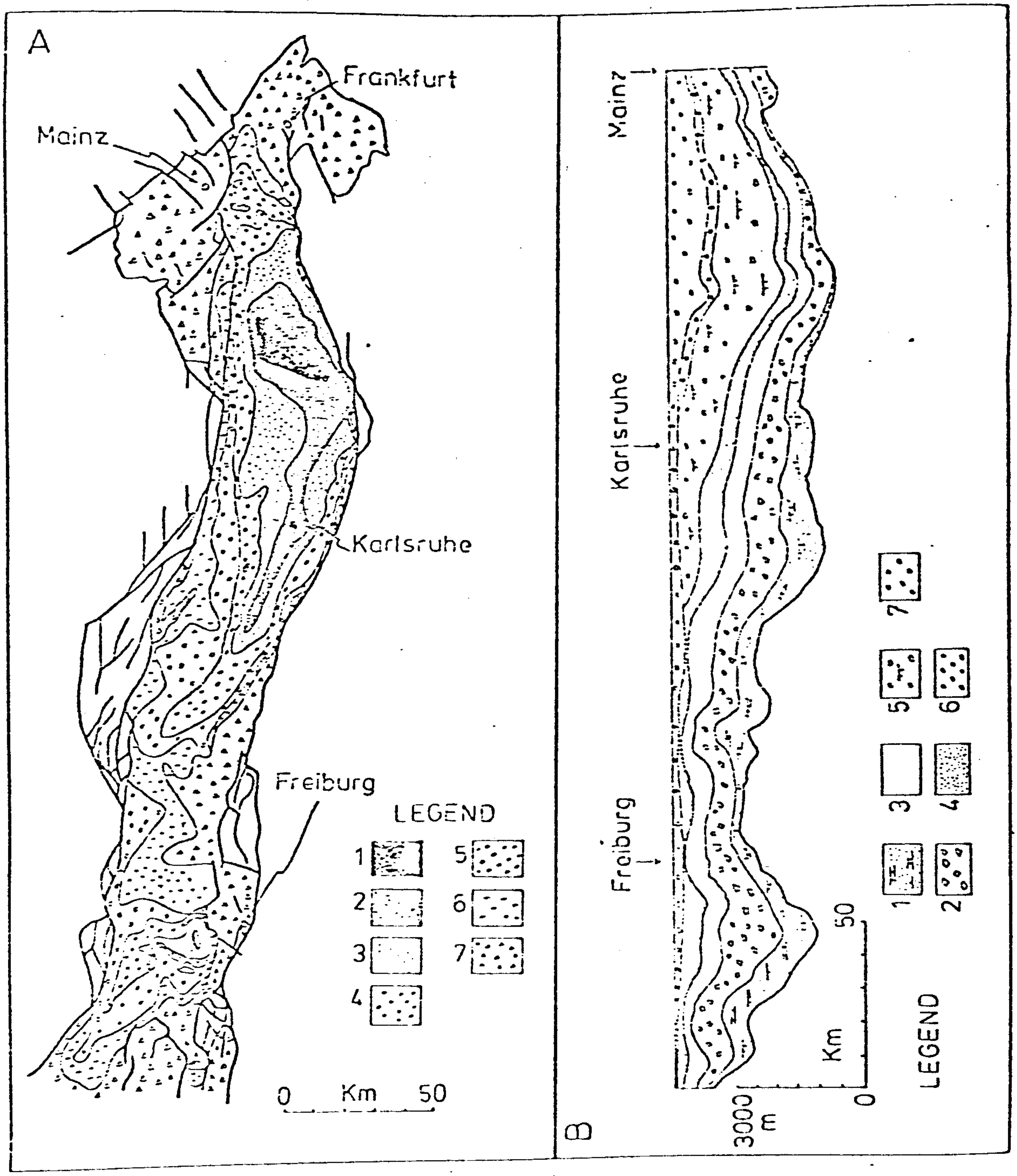
The area of most rapid subsidence of the graben floor was still in the south during Pechelbronn times, although rifting in general had already reached the Hessen Graben as indicated by the continuity of the Pechelbronn beds and the Melania Clay to the north. The thickness of the Pechelbronn beds reaches 1600 m to the southwest of Freiburg and about 900 m near Karlsruhe (Illies, 1974b; fig. 3.3B).

During the medial Oligocene, graben subsidence levelled off and the Grey Beds, a marine marl sequence, were deposited with more or

Figure 3.3 (A) Distribution of the thicknesses of the Tertiary sediments in the Upper Rhine Graben trough. Heavy lines are faults. Key: 1, > 3000 m; 2, 3000 to 2500 m; 3, 2500 to 2000 m; 4, 2000 to 1500 m; 5, 1500 to 1000 m; 6, 1000 to 500 m; 7, < 500 m.

(B) Generalized stratigraphic section along the Upper Rhine Graben trough. Key: 1, Lymnea marls (inclusive of Siderolitikum here); 2, Pechelbronn beds; 3, Gray beds; 4, Niederrödern beds; 5, Aquitanian deposits; 6, Upper Miocene; 7, Plio-Pleistocene.

Both (A) and (B) are simplified after Illies, 1974b.


A
Frankfurt
Mainz
Karlsruhe
Freiburg
LEGEND
1
2
3
4
5
6
7
0 Km 50
B
Mainz
Karlsruhe
Freiburg
3000 m
0
Km
50
LEGEND
1
2
3
4
5
6
7

less uniform thickness (Doebl, 1970) (fig. 3.3B). During the late Oligocene, the subsidence slowed down, and the deposits of this age, the Niederroedern Beds, are disconformably overlain (fig. 3.3B).

With the onset of the Miocene, two important changes occurred in the evolution of the Rhine-Graben: first, the center of subsidence shifted to the north, where today the thickest accumulations of Tertiary sediments are found (Illies, 1974a, b; Sittler, 1969; fig.3.3A), and later in Miocene times (18 m.y.) volcanicity began within the rift (Illies, 1975).

The Aquitanian section in the Rhine Graben consists of ~1500 m of clastics and carbonates (fig. 3.3B). However, after close of the Oligocene, the subsidence axis in the Upper Rhine Graben shifted from a north-northeast-trend to almost northwest (Illies, 1974a, b). This is also reflected by the thickness of the late Tertiary/(Quaternary fill of the graben (fig. 3.3A). Likewise, the Aquitanian section was deposited in a "secondary" northwest-trending rift within the main trough of the Upper Rhine Graben (Illies, 1974a).

The Kaiserstuhl volcanic center began its activity about 18 m.y. ago (Illies, 1974a, b, 1975), about 27 m.y. after the Rhine Graben had originated and in the south subsided to more than 2.5 km. The sub-volcanic breccias of the Kaiserstuhl (Baranyi, 1974) and the close correspondence of volcanism and graben suggest that faulting controlled the volcanism.

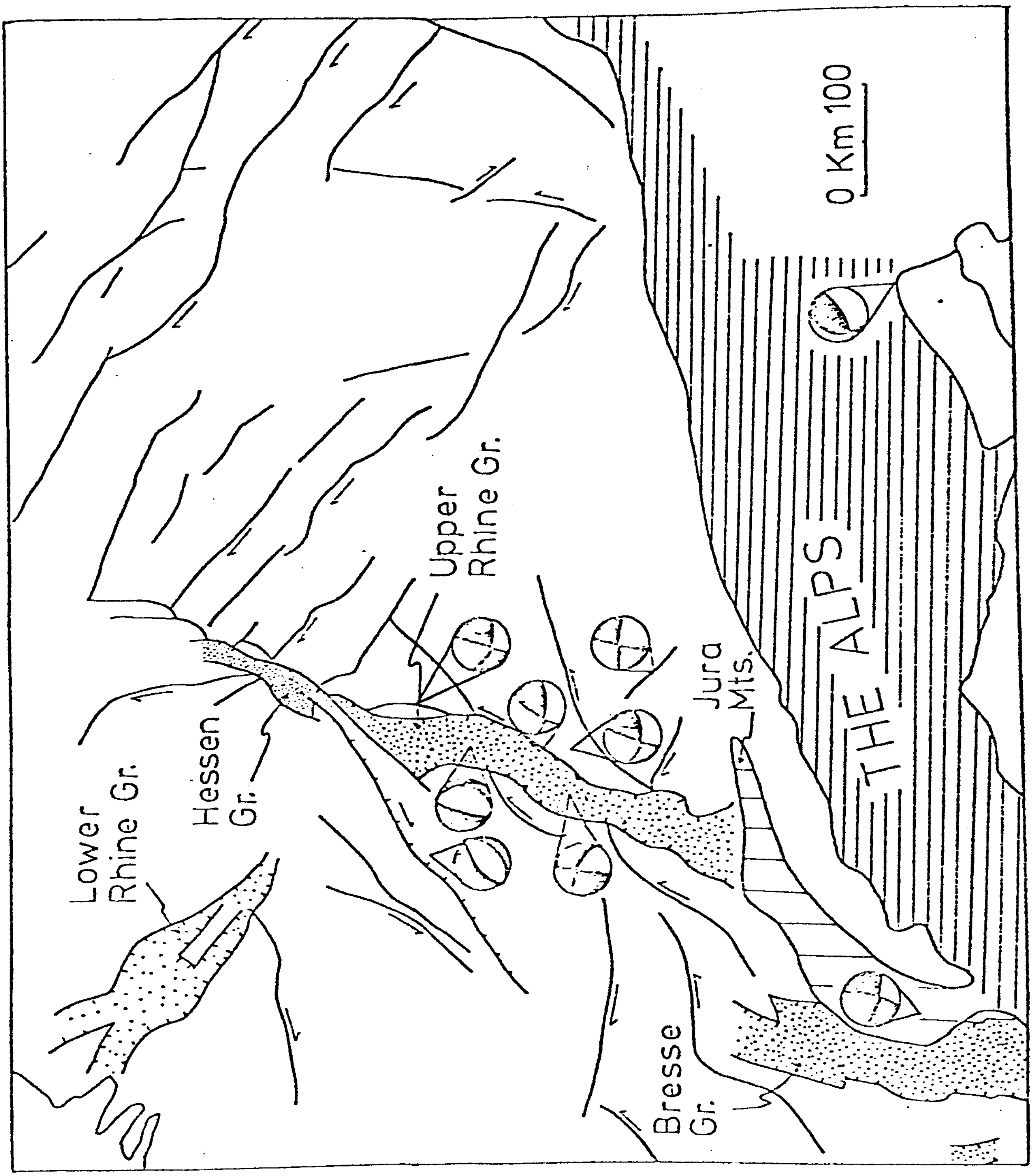
Toward the end of the Tertiary, Rhine Graben tectonism began to change from pure extension normal to the graben axis to sinistral strike-slip (Illies, 1972, 1974a, b; 1975; Illies and Greiner, 1976). Today the graben as a whole is a broad strike-slip zone with associated second-order extensional and compressional features (Ahorner, 1970; Ahorner and Schneider, 1974; fig. 4). The central portion of the graben is no longer a rift but a ramp valley as shown by the high-angle thrust faults near Baden-Baden (Illies and Greiner, 1976, fig. 3) and by seismic first motion studies (Ahorner and Schneider, 1974) (fig. 3.4).

Detailed knowledge of the structure and stratigraphy of the Upper Rhine Graben trough and its basement enabled Illies (1967) to reconstruct palinspastically the pre-rifting geometry and thereby determine the amount of extension since the initial rifting. After doming was taken out, he found a remaining gap of 4.8 km, indicating extra extension not accountable by doming. If the reversed syn- and antithetic normal faults of the graben are listric, then this gap may be even greater.

Discussion

The above review of the late Paleozoic to Recent geological evolution of the Upper Rhine Graben area of Central Europe shows that (1) the miogeoclinal assemblages of the Alpine Ocean do not extend into the present graben; (2) during late Palaeozoic and Mesozoic times, the Rhine Graben area did not have a "precursor" of the present graben; (3) before the graben formed during Lutetian time, there had been no pre-graben doming in the area that could have led to a "key stone-drop" to form the Rhine Graben. From this, it is evident that the Upper Graben is not, as suspected by Burke (1977), an aulacogen. Cloos (1939) and Burke and Dewey (1973) suggested that it formed as a result of doming around the Vogelsberg volcano; the Hessen and the Lower Rhine Grabens were also ascribed to the same doming, and the three rifts were

Figure 3.4 Active tectonics of the Alpine Foreland in Central Europe. Compiled from Ahorner, 1970; Ahorner and Schneider, 1974; Illies, 1974b; Illies and Greiner, 1976; Müller, in press; Pavoni and Peterschmitt, 1974. In the stereographic projections of the fault plane mechanisms white quadrants are dilational, black quadrants compressional.


0 Km 100
THE ALPS
Upper Rhine Gr.
Jura Mts.
Hessen Gr.
Lower Rhine Gr.
Bresse Gr.

suggested to form the Frankfurt triple junction (Burke and Dewey, 1973). However, the data, not only from the Upper Rhine Graben but also from the Lower Rhine Graben and Hessen Graben, do not support the doming-rifting-triple junction formation hypothesis. The northwestern part of the Lower Rhine Graben was active during the Maastrichtian (Teichmüller, 1974), and possibly was related originally to the North Sea rift system (Whiteman and others, 1975; P.A. Ziegler, 1975; W.H. Ziegler, 1975). Renewed rifting during Pliocene and Recent times extends from the Rheinische Schiefergebirge to the North Sea and is associated with Quaternary maar explosions and eruptions in Eifel area (Lotze, 1974; Greiner and Illies, in press). This new period of northwest-trending normal faulting coincides in time with sinistral shearing along the north-northeast striking Upper Rhine Graben. It may, therefore, be a second-order extensional feature related to the Upper Rhine Graben shear zone. The Hessen Graben has been largely inactive during late Tertiary times. Cloos' (1939) interpretation of the Upper and Lower Rhine and the Hessen Grabens does not explain the temporal evolution of the three rifts and especially the late Tertiary strike-slip movement of the Upper Rhine Graben. Moreover, the Upper Rhine Graben began its subsidence in the south and propagated northward; the doming-rifting hypothesis of Cloos demands the opposite.

Molnar and Tapponnier's (1975) suggestion that the Upper Rhine Graben originated as a "drag" structure suffers from the lack of a suitable wrench fault at the southern end of the graben. The zone of left-lateral shear connecting the southern end of the Upper Rhine Graben and the northern end of the Fosse Bressan (Contini and Theobald, 1974) appears to be younger than the initial graben formation and was perhaps formed as an intra-continental transform fault between the two rifts (Illies, 1974b).

Illies (1975) and Şengör (1976b) correlated the rifting events in the Upper Rhine Graben with the Alpine collision and later intra-continental convergence in the Western Alps, concluding that the Upper Rhine Graben is the <u>result</u> of the Alpine orogeny. Illies (1974a, 1975), however, believes that subduction in the Alpine Realm first gave rise to a subcrustal swelling of the mantle beneath the southern Upper Rhine Graben, which caused or greatly facilitated rifting. However, this view is difficult to reconcile with the southerly dip of the Alpine subduction zone (Dewey and others, 1973; Trümpy, 1975; Hawkesworth, Waters, and Bickle, 1975; Dietrich, 1976) notwithstanding Oxburgh's (1972) flake model. Even if the system involved a three-plate geometry with two subduction zones of opposing polarity, as suggested by Roeder (1976), the conspicuous absence of a subduction related arc and the paucity of andesitic volcanism make it unlikely that the magnitude of northerly and/or southerly subduction was great enough to have given rise to a mantle swell almost 200 km away from the area of the Alpine suture in the Ivrea Zone. The late Miocene collision along the Bitlis suture (Hall, 1976) in southeastern Turkey gave rise without any indication of a crustal uplifting or preliminary dike injection to north-south trending rifting and fissure formation, along which the Plio-Pleistocene alkaline basalts of the Karacalidag volcanic province erupted (Şengör and Göçmen, in preparation). We therefore follow Şengör's (1976a, b) interpretation that holds the Alpine collision directly responsible for the Rhine Graben rifting.

In conclusion, we emphasize that high-angle rifts may have multiple causes of origin both genetically related to the evolution of the associated mountain belt and also independent of it. Because oceans are likely to open and close several times along roughly the same lines (J.T. Wilson, 1966b), high-angle rifts may be rejuvenated several times during their life history with all the complexities of their pre- and post-collisional evolution, as schematically shown in figure 3.1. Such multiple events may obscure the origin of a high-angle rift or rift system beyond any reasonable hope of unraveling their history. Geologists working in such high-angle rift areas should be aware of the bewildering complexities of their evolution and hesitate to draw hasty conclusions as to their origin and tectonic significance; for instance, if a collisional orogeny is immediately followed by ocean opening along the same lines it may be especially difficult to decide whether the rift belongs to the collision event and is therefore a collision rift, or to the opening cycle and is therefore a failed arm.

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CHAPTER IV

THE NORTH ANATOLIAN TRANSFORM FAULT: ITS AGE, OFFSET AND TECTONIC SIGNIFICANCE IN THE GENERAL NEOTECTONICS OF THE EASTERN MEDITERRANEAN

The North Anatolian transform fault (Fig. 4.1) is one of many large strike-slip faults striking at low angles to the general trend of the Alpine-Himalaya System of Eurasia that originated late in the orogenic history of the segments they cut (e.g. Insubric Line of the Western Alps (Gansser, 1968; Trümpy, 1973, p. 240, but especially Laubscher, 1971); Pustertal Line of the Eastern Alps (Burchfiel, in press); Kraistide-Vardar Lineament of the Balkan-Hellenide Chain (Laubscher, 1971; Boncev, 1974); Zagros Fault of SE Iran (Berberian, 1976); Karakorum Fault of the W. Himalaya-Karakorum (Molnar and Tapponnier, 1975). Generally, these faults formed shortly after the continental collision that resulted in the paroxysmal deformation of the orogenic segments in which they are located and they appear to be closely related to the given collision (e.g. disintegration of colliding promontories and change in direction of relative motion along the suture after collision). Some, particularly the older ones, such as the Insubric Line, are now completely inactive or characterized only by dip-slip motion. Those that are still active are moving in extremely complicated areas of widespread deformation in which it is not always easy to delineate plate boundaries.

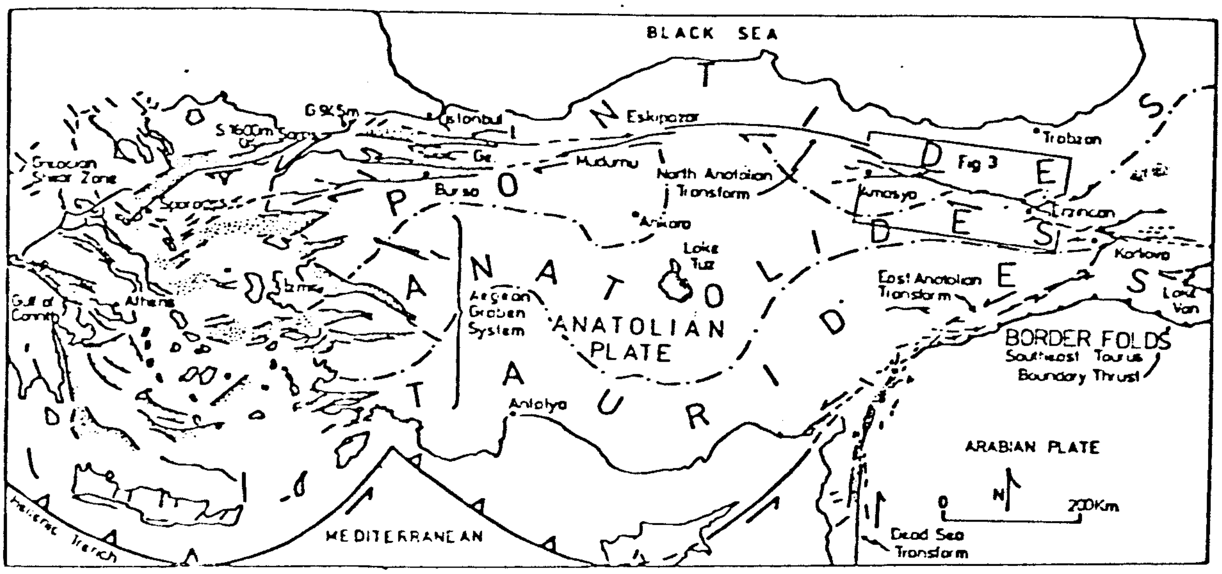
Definition and Short Historical Review

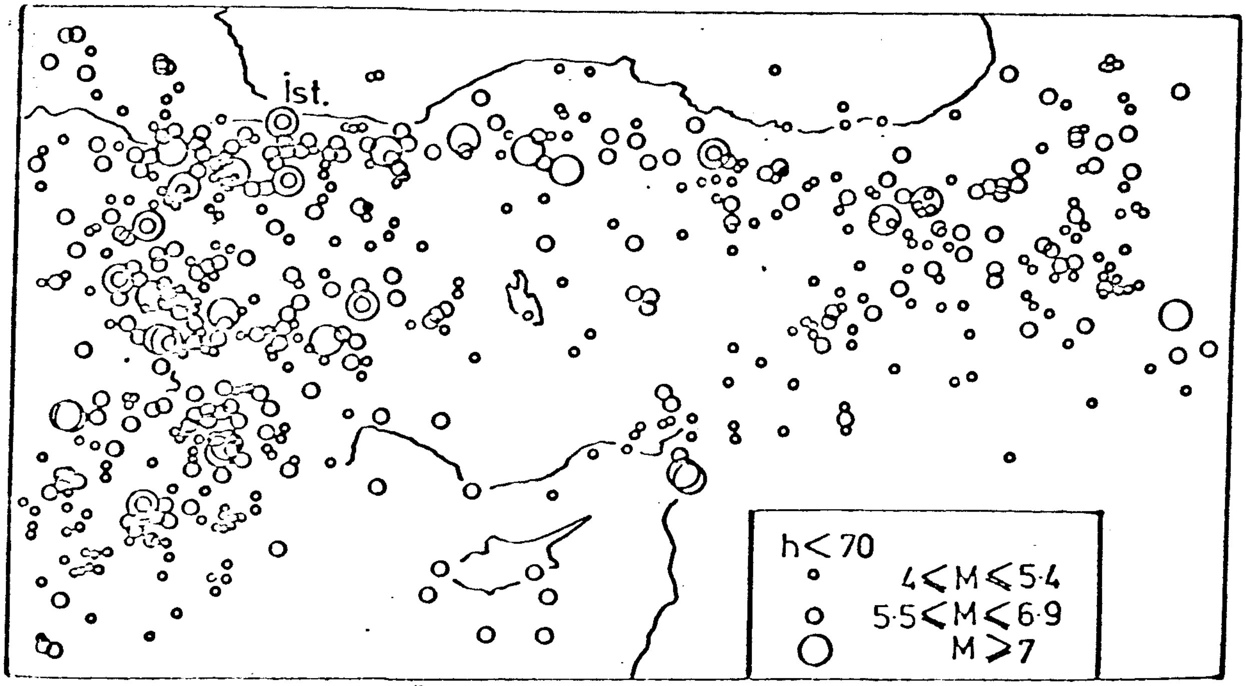
The morphologically distinct and seismically active North Anatolian transform fault is defined as that right-lateral fault zone taking up the relative motion between the Black Sea and the Anatolian plates and running sub-parallel to the Black Sea coast of Anatolia from Karliova in the E. to the Gulf of Saros in the W., thereby connecting the E. Anatolian convergent zone (McKenzie, 1972; Şengör, 1977; Şengör and Kidd, in press) with the Hellenic Trench through the complex plate boundary zone of the Aegean (Dewey and Şengör, 1979) (Figs. 4.1 and 4.2). This definition follows that of Allen (1969) and McKenzie (1972) and is therefore not exactly equivalent to the North Anatolian earthquake fault (Kuzey Anadolu Deprem Fayi) of Ketin (1957) and the North Anatolian strike-slip fault (Nordanatolische Horizontalverschiebung) of Pavoni (1961) and Ketin (1969, 1976), who continue it as a unified structure into Iran.

It was recognized very early that a fundamental tectonic boundary was present in N. Anatolia and its location was estimated to be roughly parallel with the later discovered North Anatolian fault zone. This, Nowack's (1928) Paphlagonische Narbe, Salomon-Calvi's (1936a, 1940) Fortsetzung der Tonale-Linie, and Pamir's (1950) Cicatrice Nord-Anatolienne, was believed to be an orogenic feature, the vertex of the Alpine Orogen in Anatolia. Salomon-Calvi (1963a, 1940) viewed it in the context of Wegener's theory of continental drift and, following Argand's (1924) general scheme, considered it to be the suture zone (his Synaphie) between the collided Gondwana and Eurasia elements. Later, however, a series of disastrous earthquakes along this zone, beginning with the Tercan quake on 21st November 1939 and especially the Erzincan catastrophe on 28th December 1939, resulted in a number of detailed observations (e.g. Pamir and Ketin, 1940, 1941; Parejas, et al., 1942; Pamir and Akyol, 1943; Blumenthal, 1945a, b). In his

Figure 4.1 Map of Anatolia and surrounding regions showing the active plate boundaries or plate boundary zones and the tectonic subdivisions. Heavy lines with half arrows are strike-slip faults; lines with black triangles are thrust faults; lines with hachures are normal faults; lines with open triangles are subduction zones with triangles on the upper plate; simple solid lines are unspecified faults; dotted regions are depressions; broken lines with dots in between specify the boundaries of the tectonic divisions. G is Ganosdag; Ge is Gemlik Graben; I is izmit/Sapanca Graben; S is the island of Samothraki. Figures give elevations above sea level. Both Ganosdag and Samothraki may be regions of active over-thrusting in restraining bends along the North Anatolian transform fault. The map is compiled and simplified after Dewey and Şengör (1979), Şengör and Dewey (in press) and Şengör and Kidd (in press).

Figure 4.2 Map of the epicenters of shallow earthquakes in Anatolia and surrounding regions between 11 AD and 1964 (After Ergin, et al., 1967).


BLACK SEA
Istanbul
Eskipazar
Mudurnu
North Anatolian Transform
Bursa
Ankara
Lake Tuz
Aegean Graben System
ANATOLIAN PLATE
Antalya
MEDITERRANEAN
Hellenic Trench
Gulf of Corinth
Athens
Izmir
Sporades
Grecian Shear Zone
Fig 3
Amasya
Erzincan
Trabzon
Karlova
Lake Van
East Anatolian Transform
BORDER FOLDS
Southeast Taurus Boundary Thrust
ARABIAN PLATE
0
N
200Km
Dead Sea Transform


İst.
h< 70
4<M⩽5.4
5.5<M<6.9
M>7

remarkable synthesis that interpreted the collective fieldwork on these earthquakes carried out between 1939 and 1948, Ketin (1948) concluded that the earthquakes were the manifestations of an active right-lateral strike-slip fault zone that extended along the entire length of the Black Sea mountains of N. Turkey and showed that this structure had nothing to do with the main orogenic structure of the country (op. cit. fig. 1). Instead, it was interpreted as a young feature along which an Anatolian block, comprising central and W. Anatolia S. of the fault, was drifting westwards with respect to the Black Sea basin (see, however, Pamir, 1950). He further speculated that, if his hypothesis were true, there had to be a complementary left-lateral strike-slip fault bounding the Anatolian Block to the S. and cited the earthquake of Kozan in S.E. Turkey as supporting evidence.

Ketin's hypothesis found strong support from later studies on the continued earthquake activity along the North Anatolian fault (see, for example, the summary in Ketin and Roesli, 1953) and by 1953 Ketin and Roesli were able to compare it with the San Andreas Fault of California. Following Ketin's (1957) new synthesis, Pavoni (1961) pointed out the importance of the North Anatolian fault in the tectonic context of the Mediterranean and made an unsuccessful attempt to estimate the age and offset of the fault.

From 1961 to the present, numerous detailed investigations of the fault zone have been undertaken and syntheses have been formulated, especially in the light of the theory of plate tectonics. In several of the recent plate-tectonic interpretations of the E. Mediterranean the North Anatolian transform plays an important role (e.g. McKenzie, 1972; Tapponnier, 1977; Dewey and Şengör, 1979; Şengör and Dewey, in press). Moreover, this fault and analogous strike-slip faults in collision environments are the key elements of the recent models of continental collision processes (e.g. Molnar and Tapponnier, 1975; Şengör, 1976; Dewey, 1977). Thus it is of great importance to know the detailed tectonic evolution of the North Anatolian Transform, specifically its age and cumulative offset, to be able to test and improve such models. In this paper, I therefore seek to summarize the current state of knowledge on the fault, to review some of the recent data pertaining to its age and offset and, in the light of these, to discuss its tectonic evolution and implications.

General Characteristics of the North Anatolian Transform Fault

Tectonic setting. Figure 4.1 shows the North Anatolian Fault within the tectonic framework of Turkey. The Anatolian Peninsula is a composite orogen made up of the following tectonic subdivisions defined by Ketin (1966a); from north to south.

1) Pontides: Bounded to the south by an ophiolitic suture zone, which extends from Izmir, through Ankara and Erzincan, to the ophiolitic suture of the Lesser Caucasus (Fig. 4.1) and which represents the remnants of the northern branch of the Mesozoic Neo-Tethyan Ocean that separated the Pontides from the Anatolide/Tauride Platform (see below) (Fourquin, 1975; Bergougnan, 1975, 1976; Seymen, 1975) and to the north by the Black Sea (Letouzey, et al., 1977), the Pontides acted as a magmatic arc from ?early Cretaceous to late Eocene, constructed partly on Hercynian and partly on Eo-Cimmerian basement. Especially from late Cretaceous to late Eocene (locally into early Oligocene) a widespread calc-alkaline magmatism characterized this region (Tokay,

1973; Seymen, 1975; Dr. Yücel Yilmaz, pers. comm., 1978), concurrent with the accumulation of a large mélange wedge to the south (Tokay, 1973; Gansser, 1974; Seymen, 1975). The Black Sea opened as a marginal basin behind the Pontide arc in several phases from late Cretaceous to Eocene (Letouzey, et al., 1977; Adamia, et al., 1977). The main, predominantly south-vergent deformation of the Pontides occurred during the Palaeocene/Eocene.

2) Anatolides and Taurides: These two tectonic subdivisions (Fig. 4.1) are defined on the basis of structural style and metamorphism; both stem from a single palaeogeographic domain, the Anatolide/Tauride Platform (Şengör and White, 1978; Şengör, in press) that constitutes the along-strike continuation of the Apulian Platform to the west (Burchfiel, in press). This platform seems to wedge out to the east, somewhere in the present eastern Anatolia, so that either no or a very narrow continental connection between the Anatolide/Tauride Platform and Central Iran existed until about the Eocene. The platform is characterized, in a general way, by a neritic environment in which Triassic to Neogene carbonates and subordinate clastics accumulated, although in the internal parts, the future Anatolides, sedimentation was interrupted during the Maestrichtian by the rather "quiet" arrival of large ophiolite nappes expelled from the ocean that separated the Anatolide/Tauride Platform from the Pontide arc (Ricou, et al., 1975).

Collision of the Pontide island arc and the Anatolide/Tauride Platform occurred during the late Palaeocene/early Eocene in the west and somewhat later in the east; shortly thereafter large-scale imbrication of the Anatolide/Tauride Platform began. This imbrication resulted in the large composite nappe systems of the Taurides (Özgül, 1976) and symmetric, southerly-migrating flysch/molasse troughs, characteristic of the Eocene/Oligocene tectonics of Anatolia (Şengör, in press). Another consequence of the internal imbrication of the platform was that its internal parts became deeply buried under the composite nappe systems. High T/P metamorphism, even local anatexis, occurred during burial, which, following a strong Neogene uplift perhaps similar to that seen today in the Guntschu and Gurla Mandhata highs of the Himalaya (Gansser, 1977), formed the central crystalline axis of Anatolia after the overlying nappes have been eroded off (Dürr, 1975; Şengör and White, 1978; Şengör, in press).

3) The Border Folds: During the medial Miocene, another ocean separating the already welded Pontide/Anatolide/Tauride realms (Eurasia) from the Arabian Platform closed as a result of the collision of the Arabian Platform with Eurasia (Dewey, et al., 1973; Hall, 1976; Şengör, et al., in press). The Bitlis Suture and the Border Folds in southeastern Turkey (Fig. 4.1) are the manifestations of this collision (Yalçin, in press), which is still in progress.

As seen in figure 4.1 the North Anatolian Fault is largely located within the Pontide unit, although it cuts the Pontide/Anatolide boundary three times. According to Bergougnan (1976) in the easternmost intersection the fault passes from the Pontides directly into the Taurides, because the crystalline Anatolides no longer intervene between the Pontide and the Tauride units. Ataman, et al. (1975) have argued that the Pontide/Anatolide suture was a major factor in the localization of the fault. However, the detailed studies of Fourquin (1975) in the western Pontides, of Bergougnan (1975, 1976) and Seymen

(1975) in the eastern Pontides, and a general comparison of the maps showing the course of the North Anatolian Fault and those showing the Pontide/Anatolide boundary (e.g., figure 4.1) indicate that the suture seems to have exercised almost no control on the nucleation of the fault.

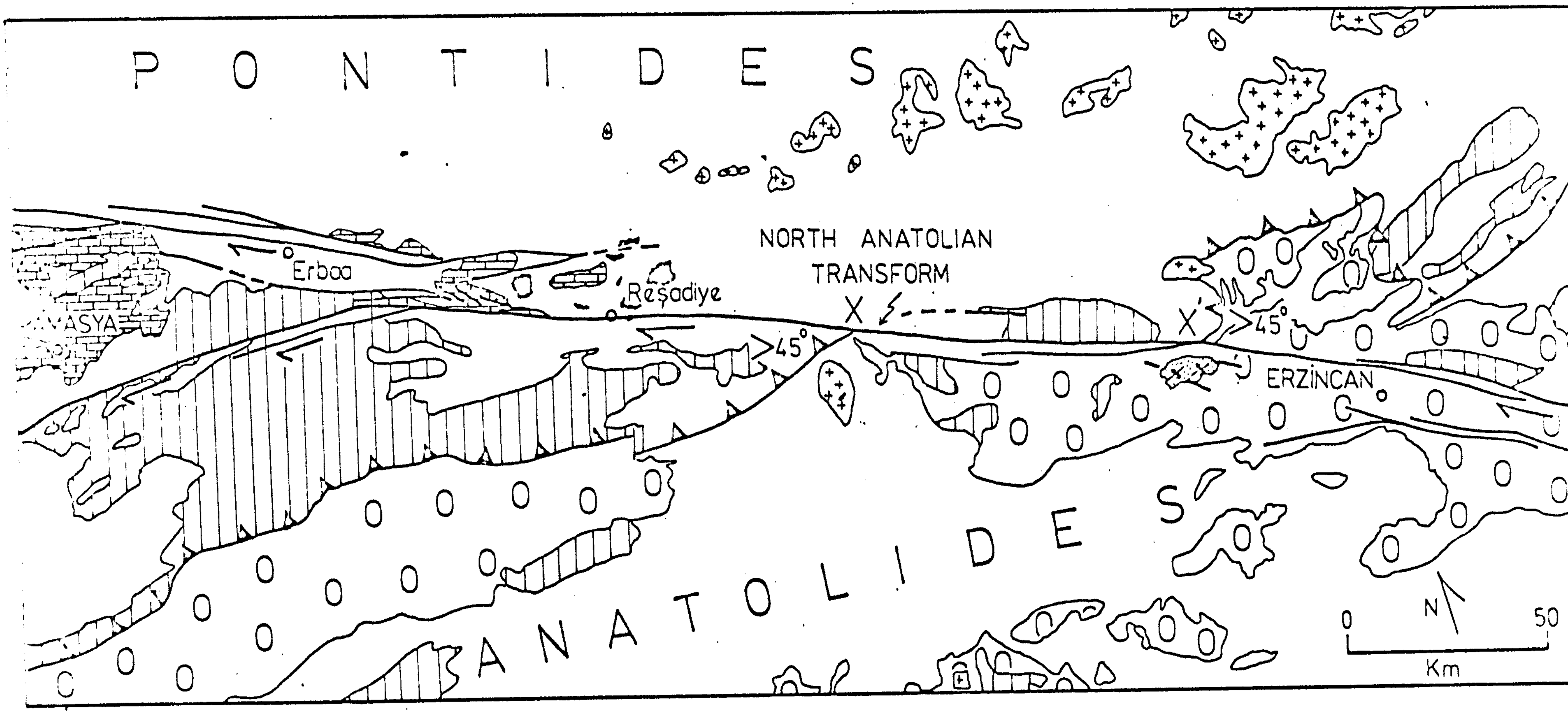
Morphology of the Transform

Like most of the large, active strike-slip faults of the circum-Pacific region (Allen, 1965) or Central Asia (Molnar and Tapponnier, 1975), the North Anatolian transform has an extremely well developed surface expression for most of its length of about 1200 km, defined by a sharp 'rift morphology' delineating a broad fault zone composed of numerous sub-parallel and/or anastomosing faults, offset, captured and/or dammed streams, sag ponds, island-like hills within major valleys following the course of the fault zone, and other deformed geomorphological features. Earthquakes, resulting from the activity of the fault sometimes cause sizable landslides, and lakes, formed through damming by such landslides, are not uncommon along the course of the fault (e.g. Pamir and Ketin, 1941).

Morphologically, the transform can be followed as a fairly continuous, strike-slip fault zone from Karliova to about Mudurnu (Fig. 4.1). E. of the point where it joins the East Anatolian transform, about 10 km E of Karliova, it is lost in the block-fault and thrust terrain of E. Anatolia, also characterized by intense Pliocene to Recent, largely calc-alkaline vulcanism (Allen, 1969, Şengör and Kidd, in press). Although several earthquakes E. of Karliova (e.g. Varto

(Ambraseys and Zatopek, 1968) and Caldiran (Arpat, et al., 1977; Toksöz, et al., 1977) produced right-lateral surface breaks, they lack the continuity and uniformity of the North Anatolian transform breaks and resemble the irregular and discontinuous strike-slip faults of NW Iran (Berberian, 1976). Moreover, many of the earthquakes E. of Karliova seem to have thrust components in contrast to the pure strike-slip earthquakes along the transform (McKenzie, 1972; see next section). The Karliova region itself is characterized by a complex fault pattern and morphology (Allen, 1969; Arpat and Şaroğlu, 1972; Seymen and Aydin, 1972), which may be due to the peculiarities of the continental triple junction here (see the last section). From Karliova to Erzincan (Fig. 4.1) the fault zone is continuous. Near Erzincan, the course of the fault is interrupted and it jumps for about 10 km to the N. The two traces of the fault zone, here, are connected by the extensional Erzincan Plain, a typical pull-apart basin (Crowell, 1974, fig. 10), characterized by young sediments and small basaltic volcanoes (Ketin, 1976). From Erzincan to Reşadiye, the trace of the fault zone is again continuous (fig. 4.3). The characteristic morphological features of this segment are elongate sag ponds, springs that are sometimes associated with travertines, fault scarps cutting the alluvium in the valley floor (related to the 1939 Erzincan quake) and deformed stream valleys (Seymen, 1975). To the W. of Erzincan, recent anticlines within the Pliocene sediments have excellent morphological expression (Fig. 4.3, Tatar, 1974). Between Reşadiye and Erbaa, the continuity of the fault is again lost and, whereas the trace coming from Erzincan turns into an E.-W. orientation S. of Amasya, a new trace begins to the N. of Reşadiye (Fig. 4.3). Between the S. and N. branches there

Figure 4.3 Simplified tectonic map of the segment of the North Anatolian transform between Amasya and Erzincan. Vertical ruling is Palaeozoic basement; 'O' denote ophiolitic mélange; carbonate pattern is Jurassic-Cretaceous rocks; dotted region W. of Erzincan is Pliocene sediments; crosses are Tertiary granodiorites and granites; black is recent basaltic volcanics. Line with black triangles is the Pontide-Anatolide suture. Note the flattening of the suture zone to the N. of Erzincan. Short lines with double arrows across them within the Pliocene sediments are Recent anticlines. Figures give the dip of the suture near the transform trace. X-X' is the apparent offset of the suture along the transform. Simplified after Seymen (1975) with data added from Tatar (1975).


PONTIDES
NORTH ANATOLIAN
TRANSFORM
X
X'
45°
45°
Erbaa
Reşadiye
AMASYA
ERZİNCAN
ANATOLIDES
N
0
50
Km

is a third branch (somewhat ill-defined and not shown on Fig. 4.3) that appears to be a secondary extensional feature within a broad pull-apart basin similar to the Erzincan Plain. Seymen (1975) has mapped this region in detail and shown that this extensional feature is also the locus of Recent basaltic vulcanism. He also showed that on both sides of the Kelkit Valley, which here follows the main southern branch of the North Anatolian transform fault, the ridges bounding the auxiliary stream valleys are bent in a clockwise fashion, indicating retardation drag here.

From Amasya to about E. of Eskipazar (Fig. 4.1) the trace of the fault zone is again continuous with a superb rift morphology formed locally by 2 or 3 subparallel fault families (Tokay, 1973). The drainage net is badly disrupted and numerous streams are cut and offset, especially between Eskipazar and Mudurnu, where the fault zone cuts across the topographic gradient (Erinç, et al., 1961; Erinç, 1973).

From Eskipazar westwards, as seen on Fig. 4.1, the trace of the transform forks (Ketin and Roesli, 1953). This forking is probably the initial expression of the distinct division of the North Anatolian transform into a northern and a southern strand farther W., in and around the Sea of Marmara (Dewey and Şengör, 1979). The depression of Çaga is located within this and is thus tectonically controlled (Erinç, et al., 1961). The formation of fault-bounded basins that occur where two splays of a strike-slip fault diverge, called fault-wedge basins by Crowell (1974, fig. 11), has been discussed by him at some length, especially in relation to the San Andreas transform.

As one follows the two main strands of the North Anatolian transform W. of Mudurnu, the clear strike-slip fault zone trace is lost and the strands are delineated by a series of E.-W. and WSW striking grabens and closed depressions, all of which appear to result from further westward splaying of the two main branches of the transform (Fig. 4.1). The northern strand of the transform runs into Izmit-Lake Sapanca Graben (Fig.4.1, I) and reappears on the W. coast of the Sea of MarmaraS. of Ganosdag (Fig. 4.1, G). Between Ganosdag and Saros, the northern strand is characterized by a continuous strike-slip trace that broke during the Mürefte quake on 9th August 1912 (Ambraseys, 1970; Allen, 1975). However, before the northern strand comes ashore S. of Ganosdag, the highest point in eastern Thrace S. of the Istiranca Ranges, a restraining bend occur just E. of it. The existence of a bathymetric depression deeper than 1000 m just E. of the restraining bend and the overall geometry of the fault here led Şengör and Dewey (in press) to suggest that where the restraining bend is located, Ganosdag may be overthrusting the floor of the Sea of Marmara, accounting for the anomalous elevation (945 m, Fig. 4.1) and abrupt relief of the former and the depression of the latter. Thus, the high region of Ganosdag may be an analogous feature, albeit on a smaller scale, to the Transverse Ranges in California. Except for the short segment between Ganosdag and Saros, both the N. and S. strands of the North Anatolian transform are characterized by horst and graben morphology, in and around the Sea of Marmara. Especially in the E. and SE of the Sea of Marmara the morphology is sharp and impressive. However, in the southern as well as in the northern branches both recent surface breaks during earthquakes (Ketin, 1966b, Allen, 1975) and fault plane solutions (McKenzie, 1972) show predominantly right-lateral strike-slip movement with subordinate, NW-striking, secondary extensional faulting (e.g. Canitez and Toksöz, 1971).

Generally, the North Anatolian Transform forms a wide belt of numerous, sometimes parallel, sometimes anastomosing strike-slip faults. Canitez (1962) has shown, on the basis of seismic and gravity observations, that the crust beneath the fault zone is thinner than normal. Within the fault zone the local lithologies often appear extensively crushed and mixed; the low resistance of these fault rocks to subaerial erosion seems to be largely responsible for the 'rift morphology' along the trace of the transform. This rift morphology extends from Karliova to Mudurnu with only two minor interruptions by the pull-apart basins of Erzincan and Reşadiye and finally merges with the horst and graben regime of W. Anatolia of Mudurnu.

Seismicity of the Transform

One of the better known aspects of the North Anatolian Transform is its seismicity. The existence of this seismic zone was already known during the last century (e.g., Dalyell, 1862) and numerous large earthquakes have occurred along it throughout historical times (Ambraseys, 1970). Figure 4.2 shows the epicentres of shallow earthquakes (h<70 km) that occurred between 11 AD and 1964 and the North Anatolian transform appears as a distinct seismic belt on this map.

The first serious seismic studies on the fault began after the 1939 Erzincan earthquake. The fault is characterized by periods or 'bursts' of seismic activity separated by quiet periods of about 150 years (Ambraseys, 1970). The present cycle of activity began with the 1939 quake and progressed generally from E. to W. (Ketin, 1948). This was followed by a burst of seismic activity in the western Anatolian portion of the Aegean Graben System (Dewey, 1976). The mode of activity of the North Anatolian transform, at least for this episode of its seismicity, appears to resemble what Scholz (1977) called 'San Jacinto-type behaviour' characterized by frequent shocks of magnitude between 6 and 7 or equal to these values. He suggested that this kind of behaviour occurs where the fault strikes close to the regional slip vector, resulting in low normal stresses across the fault plane(s). This is approximately the case for the North Anatolian transform, at least E. of the Sea of Marmara, if one uses McKenzie's (1972) pole for the Anatolia-Black Sea rotation, which he located at 18°48'N, 35°E. However, Scholz's (1977) conclusion that earthquakes of magnitude 8 that are separated by long periods of quiescence occur in those segments of large strike-slip faults where a component of convergence is present is contradicted by the only M = 8 earthquake along the North Anatolian transform, which occurred after a long period of tranquility near Erzincan (December 28, 1939) where a component of divergence is present. In the middle portions of the North Anatolian transform earthquakes are infrequent and aseismic slip may be important (Ambraseys, 1970).

Fault plane solutions of the major shocks along the North Anatolian transform have been presented mainly by Canitez and Üçer (1967), McKenzie (1972) and Dewey (1976). Between the E. end of the Sea of Marmara and Karliova, the fault plane solutions give consistently pure right-lateral strike-slip; because all the major earthquakes produced surface breaks, there is no nodal plane ambiguity for any of these solutions (McKenzie, 1972). From Karliova eastwards, the fault-plane solutions give thrust and strike-slip components which are consistent with the overall convergent tectonics of E. Anatolia

(McKenzie, 1972; Şengör, 1977; Şengör and Kidd, in press). The right-lateral strike-slip component here is due to the orientation of the faults with respect to the Arabia-Eurasia convergence vector. At the W. end of the fault, within and around the Sea of Marmara, there are strike-slip and normal fault-plane solutions (McKenzie, 1972, 1977) as discussed above.

Recently, Canitez and Toksöz, (in press) studied the stress release during earthquakes along the E. portion of the North Anatolian transform fault to identify the points of stress concentration. They infer, by means of the locations of the points of high stress, the potential earthquake locations along the fault.

The Age and Offset of the North Anatolian Transform Fault

Because of the paucity of field evidence, attempts to estimate the age and throw of the fault since Ketin's (1948) initial discovery of it have been speculative. In his 1948 paper Ketin remarked that the feature is young, i.e. post-orogenic, but did not propose any specific time of initiation. Pavoni (1961) thought that the fault could be as old as early Tertiary and, based on incorrect information, estimated its offset to be of the order of 350-400 km. Despite Ketin's (1969) judicious warnings, some workers followed Pavoni's opinion, either because of experience of large strike-slip faults gained elsewhere (e.g. Allen, 1969) or extrapolation of possibly incorrect slip rates (e.g. McKenzie, 1972). During the last 20 years, however, considerable amounts of field data have accumulated to show that neither the age nor the offset along the fault is as great as Pavoni (1961) initially believed. This information, much of which appeared in Turkish, may be summarized in three major groups, namely the geomorphological data, geological data and large-scale tectonic correlations with surrounding related structures, and is reviewed briefly below.

Geomorphological Data

Although not able to provide direct evidence for the cumulative offset of the fault, the geomorphological data gathered in and around the North Anatolian transform put certain constraints on the age of the structure. Erinç (1973) pointed out that the original drainage net around the fault zone had been established during the late (?) Miocene. The activity of the transform has since largely modified this network and a new drainage system, in places following the trace of the crushed zone for considerable distances, has been formed. Ketin (1976) pointed out that within the rift zone of the fault no sediments older than middle Miocene have been found, indicating that at least the morphological expression of the fault did not exist prior to this time. Near Mudurnu, within the early Pliocene sediments Abdüsselamoglu (1959) found what he identified as landslide deposits resulting from large earthquakes. He correlates these postulated earthquakes with the activity of the North Anatolian transform and argues that the fault must have begun to move before the Pliocene. Just W. of Erzincan Tatar (1975) mapped inactive branches of the North Anatolian transform now covered by Pliocene sediments and concluded that the fault originated in pre-Pliocene times. Therefore, the geomorphological data constrain the age of the fault between late early Miocene and Pliocene.

Geological Data

The greatest single contribution to our present knowledge of the age and offset of the North Anatolian transform was made by Seymen's (1975) detailed mapping around Reşadiye (Fig. 4.3). Here he showed that the boundary between the Pontide and Anatolide regimes is cut and offset by the transform. In this area the major overthrusting of the Pontides onto the Anatolides took place during the Burdigalian, and Seymen (1975) interpreted this as the manifestation of the terminal suturing between the two tectonic provinces, as it terminated the marine conditions near the suture and resulted in large-scale basement-nappe formation. The suture is also characterized by extensive ophiolitic mélange units (Fig. 4.3). The suture zone has a dip of more than 45° near the fault on both sides, but, especially N. of Erzincan, rapidly flattens as suggested by the mapped pattern. Seymen argued that, at least in the segment between Amasya and Erzincan (Fig. 4.3), the fault must be of post-Burdigalian age because it cuts and displaces a suture of this age. The lower limit for the age of the fault in this area is provided by Tatar's (1975) observation that now inactive splays of the transform here are covered by Pliocene sediments. Therefore, the age is bracketed between the Burdigalian and Pliocene, at least in the area mapped by Seymen, and this bracket is in excellent agreement with the independent geomorphological data.

The apparent offset of the Pontide-Anatolide suture between Amasya and Erzincan (Fig.4.3, X-X') is about 85 ±5 km. Bergougnan's (1976) mapping revealed a similar offset. Seymen (1975) argues that, since the dip of the offset suture near the fault zone is sufficiently steep and the vertical motions along the fault zone are negligible compared with the horizontal component of movement, one can therefore take the apparent offset to be a very close approximation of the real offset; however, further data are needed to determine the offset exactly. Several other lines of indirect evidence seem to converge on an offset of the order of 80-100 km. Starting from different assumptions, Canitez (1973) and Arpat and Şaroglu (1975) estimated the average rate of motion along the fault to be about 1-2 cm/year. If we extrapolate this rate to the time period between the Burdigalian and Pliocene, we get an amount of offset comparable in order of magnitude to that argued for by Seymen (op.cit.). Various other authors, such as Tokay (1973) and Tatar (1975), provided several constraints on the possible minimum and/or maximum amounts of throw along the North Anatolian Transform which bracket the cumulative offset between 50 and 100 km.

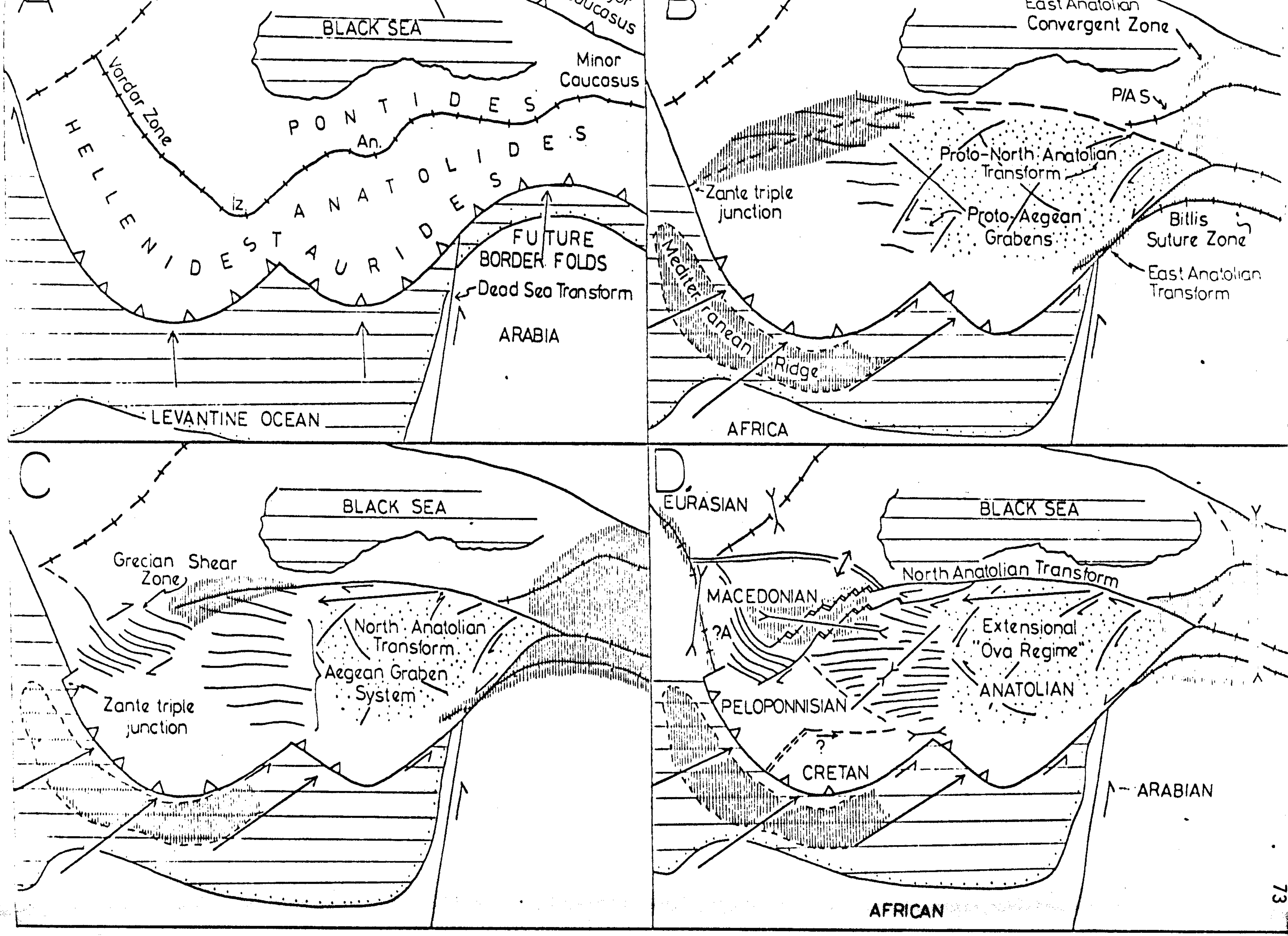
Large-scale Tectonic Correlations

Large-scale tectonics of Anatolia and surrounding regions do not support the idea of a large (more than a couple of 100 km) offset along the North Anatolian transform fault. Especially, the Anatolian tectonic units are traceable into the Hellenides of Greece (e.g. Bernoulli, et al., 1974; Monod, in press) and there is neither enough deformation in nor enough room between Anatolia and Greece to accommodate an offset more than 300 km along the North Anatolian transform. On the other hand, a smaller offset on the order of 80-100 km would be compatible with the known geology and post-Oligocene tectonics of the Aegean area as pointed out by Dewey and Şengör (1979). An even smaller offset of 22 km along the East Anatolian transform (Arpat and Şaroglu, 1972) substantiates this view.

Discussion and Conclusions: Evolution of the North Anatolian Transform Fault and Its Tectonic Significance

It is becoming widely recognized that large strike-slip faults striking at low angles to collision-type orogenic belts (Dewey and Bird, 1970) are common elements of collision environments, both within the orogenic belt and in the forelands (McKenzie, 1972; Pavoni, 1961a and b; Dewey and Burke, 1973; Molnar and Tapponnier, 1975; Şengör, 1976; Dewey, 1977). This appears to be largely the result of the high buoyancy of continental rocks that prohibit their extensive underthrusting and subduction and also of their lower shear strength compared with the stronger oceanic rocks that facilitates their failure (McKenzie, 1969). If convergence between two continents continues after continental collision, the displacement must be converted into intracontinental strain either by shortening and resultant thickening of continental crust or by 'wedging out' pieces of it into areas where there is still easily subductable oceanic material. In places where no ocean is left to be consumed between converging continental blocks, the first mechanism seems to dominate. This is probably going on today in places such as Tibet (Dewey and Burke, 1973), Iran, and E. Anatolia E. of the Karliova junction (Şengör and Kidd, in press). In areas where there are pieces of remnant oceans left between collided promontories, pieces of these promontories are sliced off and pushed into the oceanic tracts as this process is probably easier than thickening the continent against gravity (McKenzie, 1972). McKenzie (1972) and Dewey and Şengör (1979) argued that the North and East Anatolian transform faults owe their origin to such a mechanism. The temporal and spatial evolution of the North Anatolian transform fault and related structures in the E. Mediterranean are shown schematically in Fig. 4.4. During the early middle Miocene (Fig. 4.4a), the Pontide-Anatolide suture was completed, with perhaps some convergence still going on in the E. section, and the Bitlis Ocean was about to be obliterated by the northward-marching Arabia along the Dead Sea transform (Freund, 1965) with respect to Anatolia. During the late Miocene, the Bitlis Ocean closed (Dewey, et al., 1973; Hall, 1976; Dewey and Şengör, 1979) and Arabia-Eurasia convergence began to be taken up by the entire E. Anatolian convergent zone (Şengör and Kidd, in press) as shown in Fig. 4.4b. Within the resolution of the available stratigraphical data, there is no way to decide whether the North Anatolian transform or the Bitlis suture originated earlier. However, the data are perfectly compatible with the model proposed by McKenzie (1972) and Dewey and Şengör (1979) and, as it appears to be the most likely explanation for the generation of the North and East Anatolian transforms I will adopt it here. According to this model, the generation of the two faults must have followed the Anatolia-Arabia collision very closely. Dewey and Şengör (1979) further argued that the sharp SW bend in the trend of the transform W. of Saros (Fig. 4.1) put the fault in a locking geometry in the N. Aegean. This resulted in roughly E.-W. compression over much of the N. and central Aegean, which, especially in W. Anatolia and central Aegean, resulted in N.-S. extension. During the late Miocene, the extensional regime of the Aegean had already set in and Dewey and Şengör (1979) correlated this extension with the activity of the North Anatolian transform.

Figure 4.4 Tectonic evolution of the eastern Mediterranean since the early medial Miocene (A), through late Miocene (B) and late Pliocene (C) to present (d) (modified from Şengör, in press). Lines with a ladder-pattern are sutures and/or zones of intracontinental high strain; lines with open triangles are subduction zones with triangles on the upper plate; lines with half arrows are transform faults. Vertically ruled regions are zones of intraplate compression. Widely spaced horizontal ruling indicates oceanic and/or stretched, thinned and diked, subductable continental regions. Stippled area is the extensional ova province of Central Anatolia. Full arrows show the direction of relative motion across plate boundaries; length of arrows are somewhat proportional to rate of relative movement.


A
BLACK SEA
Minor Caucasus
Vardar Zone
PONTIDES
An.
HELLENIDES
ANATOLIDES
Iz.
TAURIDES
FUTURE BORDER FOLDS
Dead Sea Transform
ARABIA
LEVANTINE OCEAN
B
East Anatolian Convergent Zone
P/A S
Proto-North Anatolian Transform
Zante triple junction
Proto-Aegean Grabens
Bitlis Suture Zone
East Anatolian Transform
Mediterranean Ridge
AFRICA
C
BLACK SEA
Grecian Shear Zone
North Anatolian Transform
Aegean Graben System
Zante triple junction
D
EURASIAN
BLACK SEA
North Anatolian Transform
MACEDONIAN
?A
Extensional "Ova Regime"
ANATOLIAN
PELOPONNISIAN
?
CRETAN
ARABIAN
AFRICAN

In Greece, the transform can no longer be found as a strike-slip fault zone. Dewey and Şengör (1979) argued that the predominantly NW-striking, normal faulting there may be the expression of a right-lateral shear zone, called here the Grecian Shear Zone (Figs. 4.1 and 4.4). Along this shear zone the westward moving Anatolian plate is suggested to rip away Macedonia and Albania from the rest of the Balkans (Dewey and Şengör, 1979).

The westerly motion of Anatolia with respect to Eurasia and Africa caused a change of direction of relative motion across the Hellenic Trench (see Fig. 4.4a and b). The previously northward subduction turned to northeastward subduction along the NW-striking segments of the trench S. of Anatolia and to left-lateral strike-slip faulting in the NE-striking segments.

By early Pliocene times, the E. Mediterranean plate geometry looked something like that shown in Fig. 4.4c. The extensional ova* regime (Fig. 4.4d), characterized by large, somewhat equant, generally fault-bounded depressions (ovas) filled with Neogene terrestrial sediments and locally young, basaltic volcanics, had already set in during the late Miocene and was well developed during the late Pliocene. The existence of this peculiar internal deformation of the Anatolian plate and its implications to the nature of the Karliova triple junction specifically, and to intracontinental tectonics generally, pose interesting problems and the rest of this chapter will be devoted to a short discussion of these problems.

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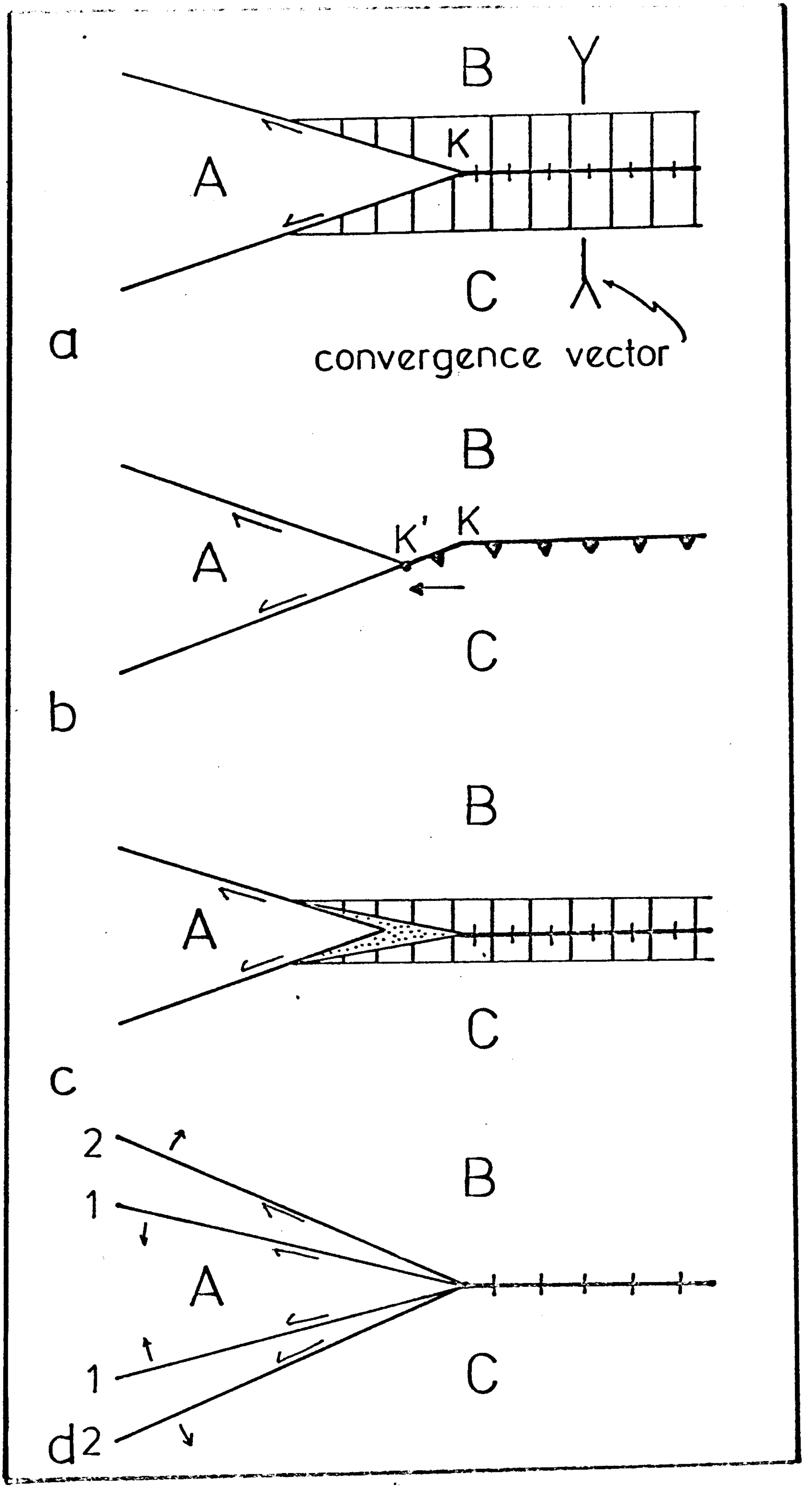
*An ova (Turkish) is a roughly equant or elongated Neogene-Quaternary depression.

The generation of the North and East Anatolian transform faults gave rise to a peculiar 'triple junction' near Karliova (Fig. 4.5). In principle it is an FFT-type triple junction (McKenzie and Morgan, 1969) and had it been wholly contained in easily-subductable oceanic lithosphere or if at least one of the plates B or C were oceanic, the boundary between them would have acted as a trench and from a time t_0 represented by Fig. 4.5a to a time t_1, represented by Fig. 4.5b, the triple junction K would have simply migrated to K'. However, the fact that the triple junction is contained entirely within continental lithosphere greatly complicates the picture, due to the characteristics of continental lithosphere discussed above. As an illustration of the resulting complexity, we can allow the vertically ruled portion of the plates B and C in Fig. 4.5a to shorten 50% by pure vertical plane strain. The resulting geometry, shown in Fig. 4.5c, has a gap (dotted) near the triple junction. As such 'holes' through the lithosphere cannot be maintained in nature, either the material in the vertically ruled area will extrude by plastic flow or by some kind of complex faulting to close the gap, or the gap will be filled by igneous rocks and will be the locus of intense volcanic activity. The complex fault pattern around Karliova (Seymen and Aydin, 1972) may, in fact, be due to such a complication. Where complex intra-continental strike-slip faults are generated in convergent environments, as in Iran, Afghanistan and central Asia, such triple-junction holes may be responsible for basin formation and basaltic vulcanicity. However, in E. Anatolia the pattern of deformation is known to be more complex than that suggested by Figs. 4.5a and c (McKenzie, 1972;

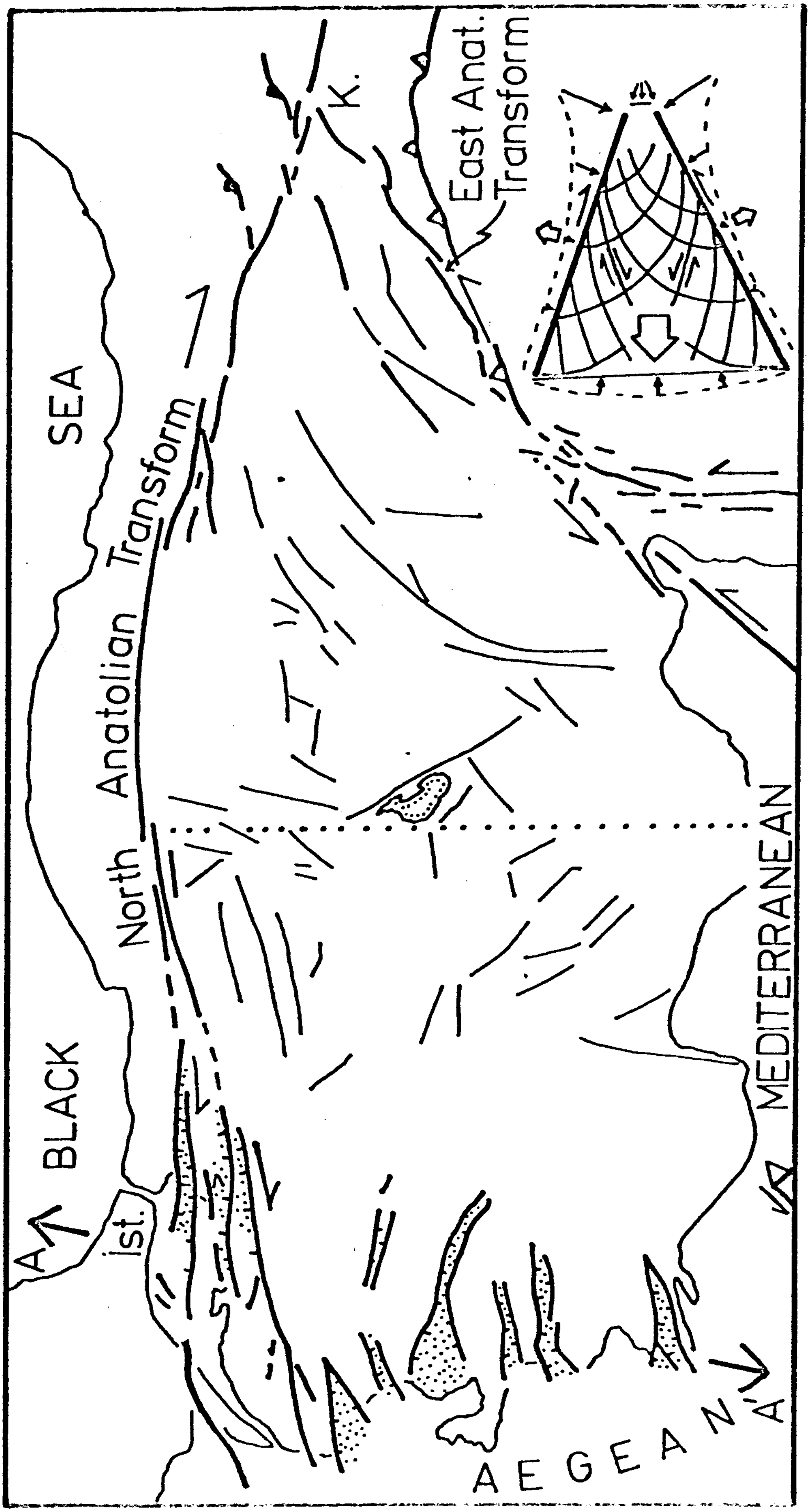
Figure 4.5 Conceptual diagrams of the Karliova triple junction (K). In reality, the boundary between plates B and C, which represent Eurasia and Arabia, respectively, is not perpendicular to the convergence vector. The boundary strikes E.-SE and this induces a right-lateral strike-slip component onto it. Thus, the right-lateral faults E. of Karliova are not the continuations of the North Anatolian transform fault, as assumed by Ketin (1969, 1976), but elements of the east Anatolian convergent zone. Further explanation in text.


B
Y
K
A
C
a
convergence vector
B
K' K
A
C
b
B
A
C
c
2
1
B
A
1
C
d2

Şengör and Kidd, in press), and a description of continuum tectonics, across the entire width of the E. Anatolian Plateau with local high strain zones, appears to be geologically more reasonable than any rigid-plate tectonics description. Therefore, instead of letting the triple junction migrate while the rigid plate A is wedged out from the jaws of the converging plates B and C, we can hold the triple junction <u>in situ</u> and allow the boundary transforms of plate A converge (Fig. 4.5d (1)) while plate A is being internally deformed and extruded westwards. In the case of the Anatolian 'plate' this seems an attractive solution, because there is a widespread Neogene-Quaternary deformation and vulcanicity within it, sometimes accompanied by earthquakes (Ketin, 1948). However, the observations that W. Anatolia and the entire Aegean are characterized by numerous E.-W. striking grabens (Fig. 4.1) which are today under roughly N.-S extension and that there is no tectonic boundary between this extensional region, in which the total post-middle Miocene extension probably exceeds 30% (Şengör, 1978), and the rest of the Anatolian plate (Fig. 4.6), make the solution suggested above an unlikely one. In fact, the Neogene-Quaternary deformation within Anatolia is characterized by the ova regime. The formation of the Anatolian ovas in the Neogene has been one of the long-standing puzzles of Anatolian tectonics (for good summaries of this subject see: Salomon-Calvi, 1936b; Ardel, 1965; Ketin, 1970; Stille, 1919). Their direct continuation into the Aegean extensional regime (Ketin, 1970; Dewey and Şengör, 1979) and the existence of basaltic vulcanism (Sungur, 1970) suggest an extensional origin for the Anatolian ovas. If this is the case, then the geometry shown in Fig. 4.5d (2) which represents the widening of the angle between the two bounding transforms of plate A, or a combination of Figs. 4.5d (2) and 4.5c, may describe the actual situation much more accurately than the geometry shown in Fig. 4.5d (1). In this instance, it is clear geometrically that some factor other than the convergence between plates B and C must be interfering to force the two transform faults to diverge in time. It is equally clear from Fig. 4.1 that this interfering factor is the extensional regime of the Aegean.

The Aegean Graben System (Figs. 4.1 and 4.4) appears to have started during the late Miocene (Berckhemer, 1977; Dewey and Şengör, 1979 and the references cited in the later paper) and this correlates well with the origin of the North and East Anatolian transforms. Once the N.-S. extension was set up in the W. end of the Anatolian plate, it appears to have propagated rapidly eastwards with ever diminishing intensity until it reached a zero value near Karliova. Thus, the Anatolian plate has been, at least since the early Pliocene, escaping from a N-S convergent to a N-S extensional environment. The motion of the Anatolian 'plate' with respect to its northern and southern neighbours and the movement of the bounding transform faults with respect to the bisector of their dihedral angle resembles greatly the situation encountered in plastic extrusion in a modified Prandtl Cell shown in Fig. 4.6. The faults that bound the Central Anatolian ovas are shown for comparison with the slip lines in the Prandtl Cell and the resemblance is striking. Similar models of plastic behaviour have been proposed to explain certain large-scale tectonic patterns in Asia (Molnar and Tapponnier, 1975), in the Mojave Desert of California (Cummings, 1976), and in the Caribbean (Burke, et al., 1978). Such comparisons and their implications should be approached very carefully,

Figure 4.6 Map showing a simplified version of the plate boundaries and plate boundary zones in Anatolia and the faults bounding Anatolian ovas. The latter are modified from Pinar-Erdem and Ilhan (1977). The symbols are the same as in Fig. 1.AA' show the Aegean extension direction after Dewey (1976). The geometric figure in the lower right corner is the modified Prandtl Cell (after Cummings, 1976, fig. 2A) with which the Anatolian internal fault system is compared. The dotted line in the middle of the map depicts the possible western boundary of the Prandtl Cell analogy in Central Anatolia.


BLACK SEA
North Anatolian Transform
East Anat. Transform
K.
Ist.
A
A'
MEDITERRANEAN
AEGEAN

though, for the necessary conditions for the theoretical and experimental models are never fulfilled in nature and, in this case, the models apply only to two-dimensional and instantaneously produced patterns; the geological structures with which they are compared are three-dimensional objects evolving over millions of years. However, the resemblance exists, the vertical motions involved are small compared with the horizontal motions, and there is also the temporal correlation between the generation of the ovas, the beginning of the Aegean extensional regime, and the generation of the North and East Anatolian transforms. I, therefore suggest that there is a causal relationship between the North and East Anatolian transform faults and the extensional ova regime of Central Anatolia. This provides a possible and testable solution to the geometrical as well as temporal aspects of the long-standing ova problem in Anatolia. Similar pairs of large-scale, strike-slip fault exist elsewhere in continental lithosphere, e.g. in Asia (Molnar and Tapponnier, 1975) in Europe (Şengör, 1976), in California (Cummings, 1976); in the Mojave Desert of California their possible role in controlling the internal deformation of pieces of continental lithosphere bounded by them was noted by Cummings (1976). Molnar and Tapponnier (1975), McKenzie (1977), and Dewey and Şengör (1979) remarked that plate tectonics is not particularly useful in describing continental deformation. Plate boundary zones or systems in continental lithosphere are wide regions of extremely complicated deformation patterns, but they still separate torsionally rigid pieces of lithosphere. The question is whether continuum or small-plate tectonics characterize such zones. The detection of patterns and temporal relationships, such as those described above and also by Molnar and Tapponnier (1975) and Cummings (1976), may eventually lead to a useful understanding of the rheology and behaviour of continental lithosphere. However, as remarked above, the pattern comparison should be made with great caution and only where the detailed geological picture is known. It is clear from the foregoing description that this is as yet not the case for Anatolia. A lot more detailed field evidence will have to be obtained, especially to discover whether or not there is any systematic change in the amount of offset along the North and East Anatolian transform faults, and to determine (a) the precise geometry of the Karliova triple junction and the fault system that formed the Anatolian ovas, (b) the ages of the ovas and (c) the amount of subsidence of the basements of ovas. Micro-earthquake surveys within Anatolia may provide further information as to whether or not the deformation in Anatolia, which appears from Quaternary geology and few earthquakes to be active, is generally free of earthquakes and is thus perhaps proceeding non-elastically.

In conclusion, it appears that since the late Miocene the North Anatolian transform fault has been a dominating factor in the tectonic evolution of Anatolia. Its effects are by no means confined to its course but are spread over the entire surface of Anatolia. The appreciation of this may shed some light on the future neotectonic and geomorphological studies of this very interesting portion of the E. Mediterranean region as well as on the nature of intracontinental deformation in analogous areas.

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The contents of this chapter form, in part, the substance of a paper entitled "The North Anatolian Transform Fault: its age, offset and tectonic significance" that is in press in Jour. Geol. Soc. London.

CHAPTER V

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

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 acorss 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 whole-sale 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-Indian Plateau (Fig. 5.1), 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.

Post-Collisional Tectonics and Vulcanicity of the Turkish-Iranian Plateau

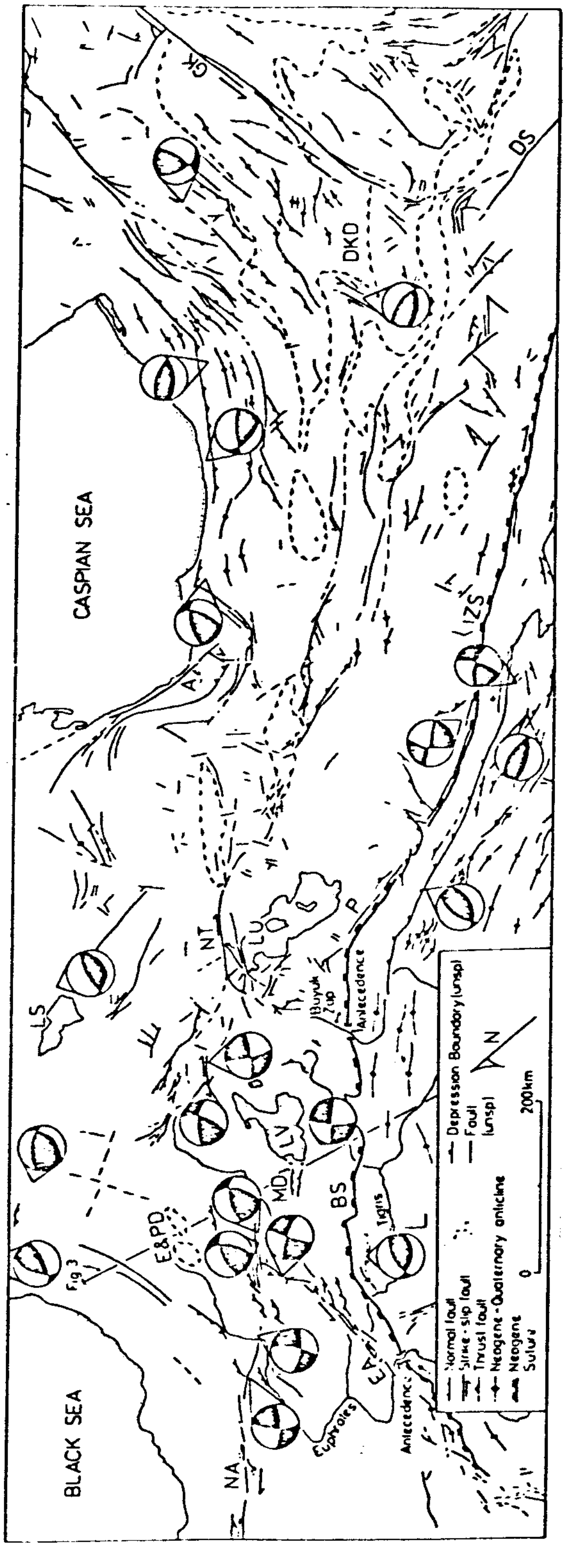
The pre-collision geology of the Turkish-Iranian Plateau is well enough known to reconstruct the first order tectonic phenomena that finally gave the region its geometry immediately after the collision. Before the Priabonian (late Eocene) two oceans existed in the area, one south of the Pontide-Lesser Caucasus-Alborz mountain system and the other south of the Anatolids and the Central Iranian Plateau (Fig. 5.3). 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 Sengör, 1979).

At present, in eastern Turkey more than 70% of the total area, generally represented by moderately flat surfaces, lies between 1.5

Figure 5.1 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.

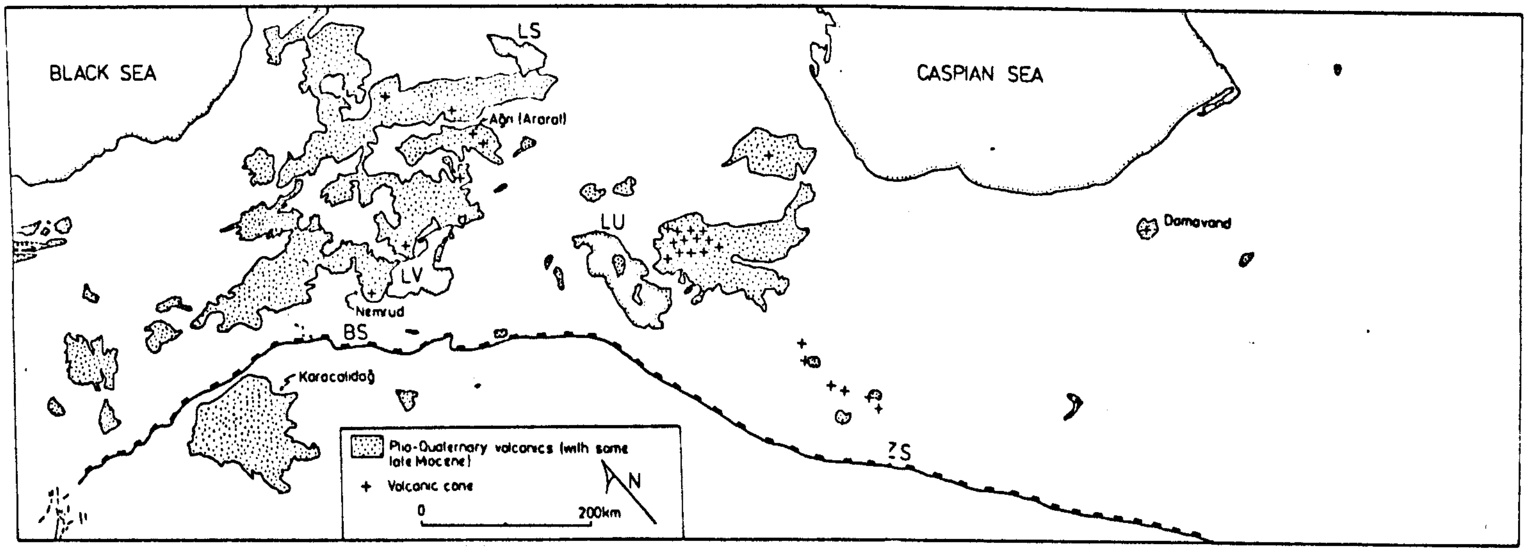
Key: NA = North Anatolian Transform; EA = East Anatolian Transform; E & PD = Erzurum and Pasinler Depressions; MD = Mus Depression; BS = Bitlis Suture; L = Lice earthquake focal mechanism solution; NT = North Tabriz Fault; A = Astara Fault; P = Piranshar Fault; ZS = Zagros Suture; DKD = Dasht-i Kavir 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. (in press).

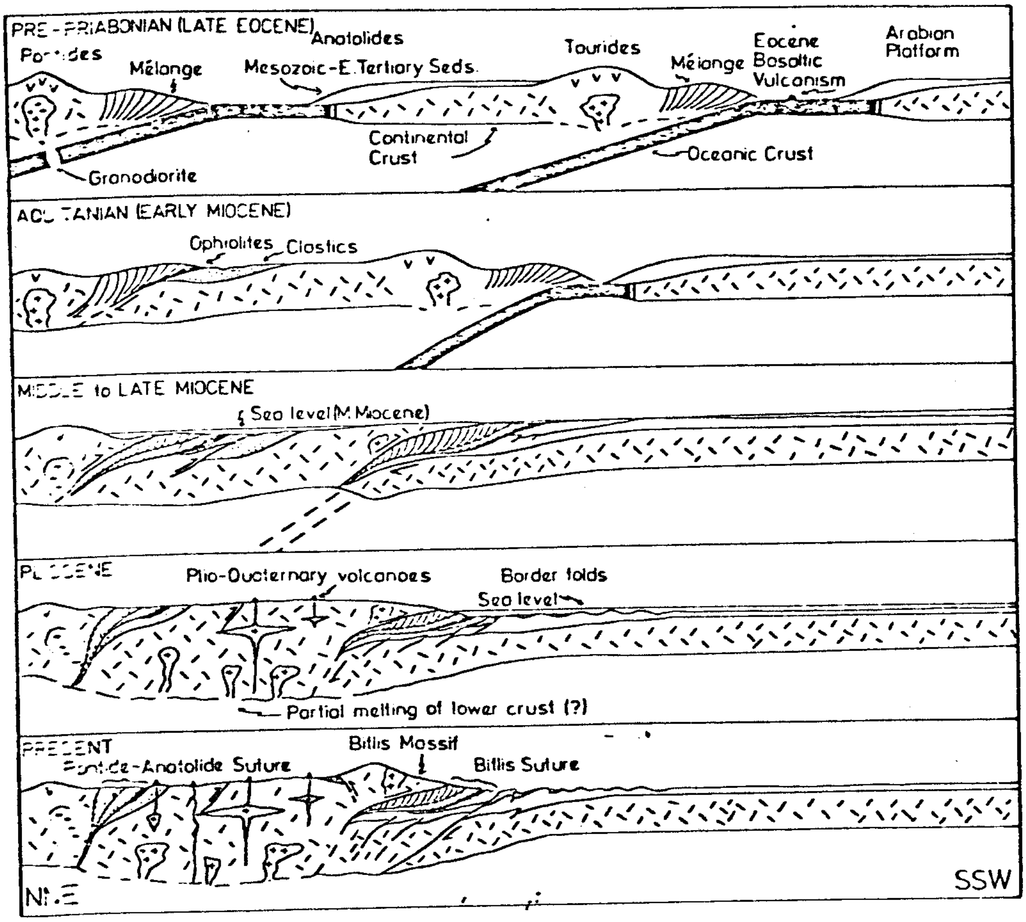


BLACK SEA
CASPIAN SEA
NA
Euphrates
Antecedence
EA
BS
MD
LV
Tigris
E&PD
Fig 3
LS
NT
LU
Buyuk Zap
Antecedence
P
ZS
A
DKD
DS
GK
Normal fault
Strike-slip fault
Thrust fault
Neogene-Quaternary anticline
Neogene Suture
Depression Boundary (unspl)
Fault (unspl)
0
200km
N

Figure 5.2 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), International Tectonic Map of Europe (1962).


BLACK SEA
CASPIAN SEA
LS
Ağrı (Ararat)
LU
LV
Nemrud
BS
Karacalıdağ
Damavand
ZS
Plio-Quaternary volcanics (with some late Miocene)
+ Volcanic cone
0
200km
N

<u>Figure 5.3</u> 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. 5.1. 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, <u>et al</u>.(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.


PRE-PRIABONIAN (LATE EOCENE)
Pontides
Mélange
Anatolides
Mesozoic-E. Tertiary Seds.
Taurides
Mélange
Eocene Basaltic Vulcanism
Arabian Platform
Continental Crust
Oceanic Crust
Granodiorite
AQUITANIAN (EARLY MIOCENE)
Ophiolites
Clastics
MIDDLE to LATE MIOCENE
Sea level (M. Miocene)
PLIOCENE
Plio-Quaternary volcanoes
Border folds
Sea level
Partial melting of lower crust (?)
PRESENT
Pontide-Anatolide Suture
Bitlis Massif
Bitlis Suture
NNE
SSW

and 2.5 km (Tanoglu, 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. 5.1). 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. Agri (5165 m).

A late Oligocene-early Miocene marine transgression coming from the west inundated large areas of the Turkish-Iranian Plateau (Alavi-Naini, 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; Altini, 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). Tanoglu (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. 5.1).

In spite of the block character of the uplift in the Turkish-Iranian Plateau in general, the Bitlis suture zone zppears 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. 5.1) 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. 5.1) is also antecedent. Just north of the suture and 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. 5.1). Although the majority of these thrusts are south-vergent, a number of them moved to the north-east 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. Agri, the other following the Bitlis Suture (Fig. 5.1). Between these two, the easterly continuation of the North Anatolian Fault Zone (Ketin, 1948, 1969) seem to have a major thrust component according to McKenzie's (1972) fault-plane solutions (Fig. 5.1), 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, Ketin (personal communication, 1977) 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. 5.2, 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.). 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. 5.1 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. 5.2) 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 vulcanism 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 recommended during the Pliocene and is still active (Ketin, 1961; Ganser, 1966). There is no evidence of a descending lithospheric slab beneath the plateau that can be connected with the active vulcanism; 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 vulcanism is very extensive in the highest parts of the plateau (Fig. 5.2) 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. Agri. 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. 5.4). 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

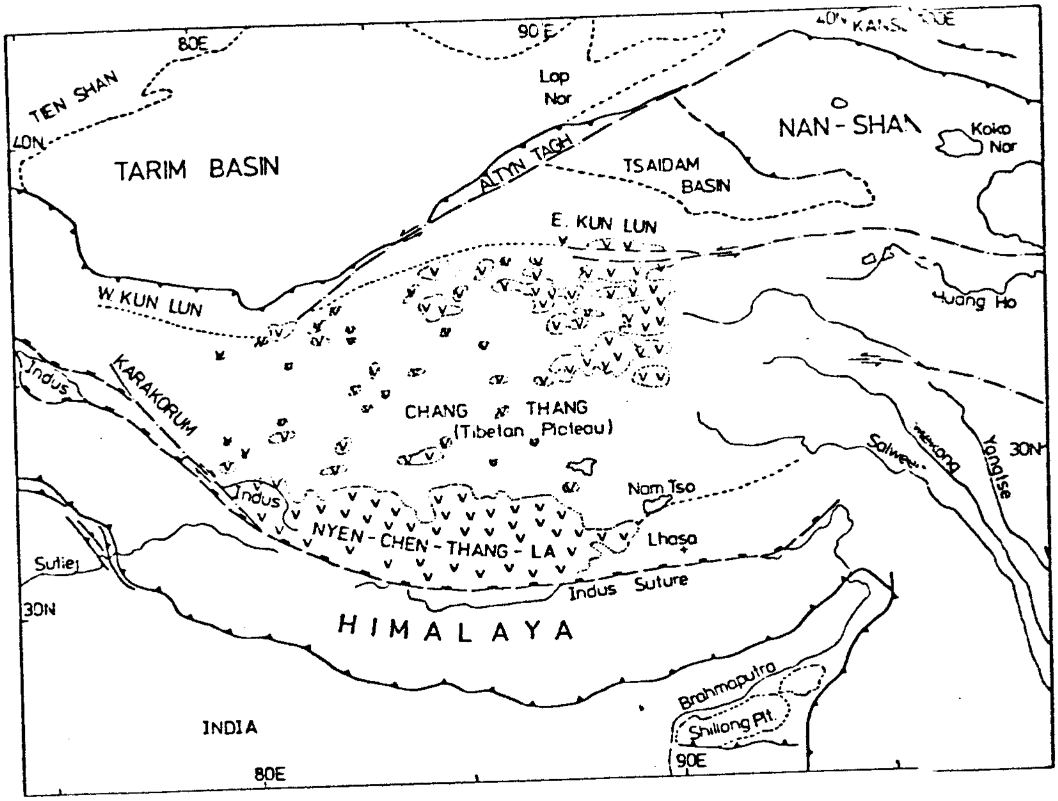
Andean-type arc constructed during the subduction that removed the ocean between India and Asia. Alternatively, some of these granodiorites may post-date the collision. No rocks of continental shield type are known south of Kun Lun, and therefore Tibet is apparently not a fragment of a pre-Mesozoic continent.

Over large areas of the Tibetan Plateau, as far as 92°E, the deformed pre-Mesozoic rocks are unconformably overlian 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 folding is often accompanied by about 10% 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 (1947) 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. 5.4), 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 Johannsen, 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

Figure 5.4 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.


80E
90 E
TIEN SHAN
40N
Lop
Nor
TARIM BASIN
ALTYN TAGH
TSAIDAM
BASIN
NAN - SHAN
Koko
Nor
E. KUN LUN
W. KUN LUN
Huang Ho
Indus
KARAKORUM
CHANG THANG
(Tibetan Plateau)
Salween
Mekong
Yangtse
30N
Indus
Nam Tso
NYEN - CHEN - THANG - LA
Lhasa
Sutlej
Indus Suture
30N
HIMALAYA
Brahmaputra
Shillong Plt.
INDIA
80E
90E

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 Brahmapatra 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 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 vulcanicity of the Turkish-Iranian Plateau and the present folding in the Tsaidam Basin and the vulcanism 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, it 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).

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The contents of this chapter are in press in Tectonophysics, in co-authorship with W.S.F. Kidd.

CHAPTER VI

CONCLUSIONS

Specific conclusions of individual studies reported in the preceding chapters are indicated in those chapters, so they are not repeated here. Instead, I outline here the overall conclusions of the study and point out the uncertainties that still remain.

The first and the obvious result of this study is that local geological data confirm the initial picture gained from a study of the present world-wide epicenter distribution. That is, once pieces of continental lithosphere are brought into contact across a plate boundary of any kind, the zone that takes up the displacement across that boundary very rapidly widens. The best example of this cited in this study is the east Anatolian/Iranian region (Chapter V). Deformation there had been largely confined to the environs of the future suture zone until collision occurred. After the medial Miocene closure along the Bitlis/Zagros suture, deformation rapidly (instantaneously within the present resolution of the stratigraphic data) expanded to the north and northeast to encompass regions including the Greater Caucasus and the Kopet Dag.

The change from thrust- (crustal thickening) dominated compressional deformation to strike-slip dominated shortening is not as sharp as would be expected from the theory as applied by Tapponnier and Molnar (see Chapter I). In Central Europe, for example, the beginning of strike-slip faulting on the foreland (early to mid-Miocene) clearly post-dated the Mesoalpine post-paroxysmal uplift of the Alpine edifice during the medial and late Oligocene. However, crustal thickening in the Alps continued as shown by the early (?medial) Miocene thrusting of the Helvetic complex and the Verschluckung of almost the whole of the Tavetsch Massif (Trümpy, 1975), and later, during the Pontian, the folding of the Jura (Laubscher, 1973). This somewhat episodic shortening history may be explained by erosion. When elevation (therefore the lithostatic load) increases to a certain value where σ_2 becomes vertical thrust-dominated deformation stops until erosion reduces the lithostatic load and restores σ_3 to a vertical position. Under the assumption of constant convergence rate, then, the periods of "orogenic interruption" should correspond to increased activity on foreland structures. To my knowledge this is not documented. In fact, perhaps the opposite is true. I have shown in 1976 (Şengör, 1976, Table 1) that periods of quiescence in the Alpine edifice, accompanied by uplift, correlated with similar periods of slackening of foreland deformation. How this apparent contradiction can be solved is not now clear and the effects of lithospheric strain rate and geometries on rates and geometries of plate motions should be considered.

In the case of the eastern Mediterranean the temporal difference between the initiation of strike-slip regime and the onset of rapid uplift of the Turkish-Iranian plateau seems so small that the presently available data are unable to tell which system originated first. However, there is no doubt that in the Turkish-Iranian Plateau thrust- and strike-slip tectonics are going on now as shown by the current fault-plane solutions. Although the current hypotheses (Chapters IV and V) seem to make a good story in the context of the available data they are by no means tested. To check them rigorously we need, first of all, a very detailed time-table. To get that, in turn, two things that are interrelated should be accomplished.

1) Detailed geological mapping to tie down the precise geometries of the structures in question and erection of very detailed time-stratigraphic tables for such localities.

2) One-to-one correlation of Neogene marine and terrestrial successions.

Another question that this study has answered in the negative is whether continental tectonics is predictable from what we know of major plate motions. For example, that the Anatolian Plate (Chapters IV and V) should "escape" into the oceanic eastern Mediterranean area was predictable from the relative motions of Arabia and Eurasia and the general behaviour of the continental lithosphere, but there was no way to tell that the North Anatolian Transform would bend southwards in Greece and cause the widespread intra-continental deformation within Anatolia. The only way to understand such unforeseen complications is to know the geological history of the structures involved. Perhaps one day we shall understand the processes of deformation well enough to be able to make field geology truly predictable. As our knowledge stands now this is not the case. I would argue that local structural fabrics - or a whole set of signatures that typify a tectonic environment - are very poor indicators of large-scale plate motions.

Uncertainties involved in this study lie, therefore, not in the conceptual models of what continental deformation should be like, but in the field data that would show what it is. Models are very useful because they direct observations, but as far as I can see, the immediate need now in continental tectonics is <u>not</u> in any kind of physical modelling of what should be there or in laboratory experiments but in very detailed field mapping of critical areas suggested by the existing models with emphasis on stratigraphy (time) and structural geology (space and geometry).

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