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

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ABSTRACT. Rift valleys striking at high angles to orogenic belts are of two main kinds: those originating at ocean opening (aulacogens) and those originating at collision (impactogens). We here consider differences between the geological evolution of the two kinds of rifts that can be helpful in distinguishing between them. We review the Permian to Recent evolution of the Upper Rhine Graben as an example. There is little evidence to support the idea that there was active rifting in the Rhine Graben in Late Triassic times contemporary with the opening of the Alpine Tethys. Major rifting in the graben began during the Eocene at the same time as the Mesoalpine collision. The subsequent development of the Rhine Graben includes strike-slip and compressional faulting and volcanism. The relationship of the Upper Rhine Graben to the Hessen and Lower Rhine Graben appears more complex than that of a simple triple-rift on a dome as Cloos suggested.

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.

The purpose of this paper 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. 1E-G) and attempt to show how it differs from the other two.

Figure 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. 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

[1] In this paper we 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.

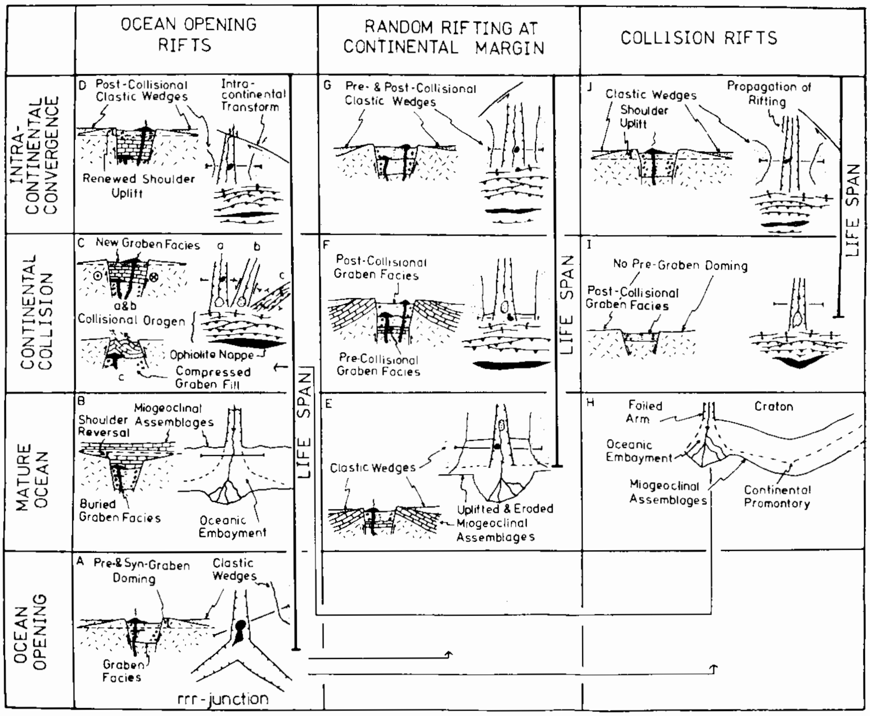

Fig. 1. A schematic illustration of events during the origin and evolution of an aulacogen (A-D), a random rift at high angle to continental margin (E-G), and a collision rift or impactogen. Cross sections show the expected differences in the stratigraphic evolution of the three kinds of high-angle rifts. For more detailed discussion see the text.

by means of peripheral clastic wedges. For 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 lated 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 marginal 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 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. 1C, D), the shoulders of the aulacogen are likely to rise again to provide a post-collisional 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. 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. 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 syn-collisional 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. 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 Upper 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. 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

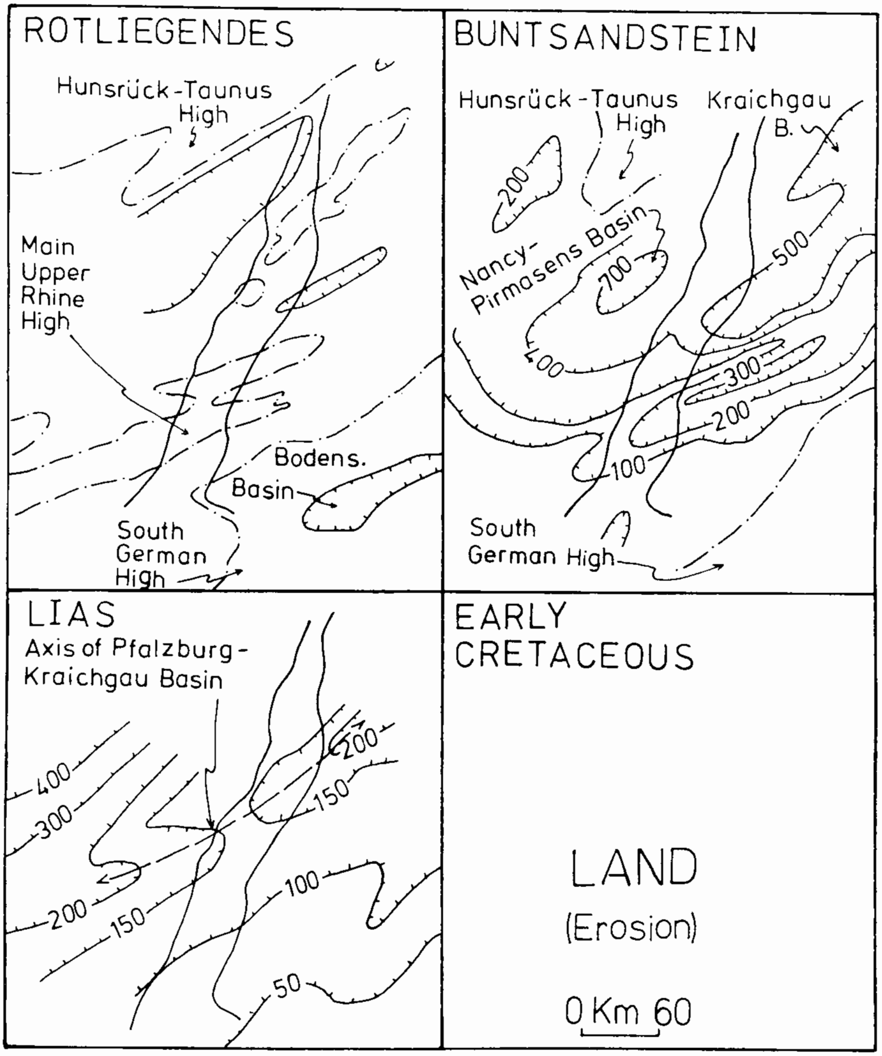

Fig. 2. The depositional framework of the Upper Rhine Graben area from Rotliegendes (Early Permian to Early Cretaceous. In Rotliegendes time only the 0 m isopach (dashed line with dots between dashes) and the 1000 m isopach (line with hachures) are shown, except around the Bodensee Basin where the hachured line shows the 100 m isopach. Figures are always in meters (data from Boigk and Schöneich, 1970, 1974; Hoffmann, 1967; P. A. Ziegler, 1975).

future Hessen Graben appears to be a preferred site of subsidence, although not as large as the future Upper Rhine Graben. Moreover, the 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 ≅845 m; Lotze, 1974) and does not indicate any major graben subsidence.

During the Lias (fig. 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. 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. 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 *Küstenkonglomerate* (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. 3B).

During the medial Oligocene, graben subsidence levelled off and the Grey Beds, a marine marl sequence, were deposited with more or less uniform thickness (Doebl, 1970) (fig. 3B). During the late Oligocene, the subsidence slowed down, and the deposits of this age, the Niederroedern Beds, are disconformably overlain (fig. 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. 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. 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. 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 subvolcanic 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. 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

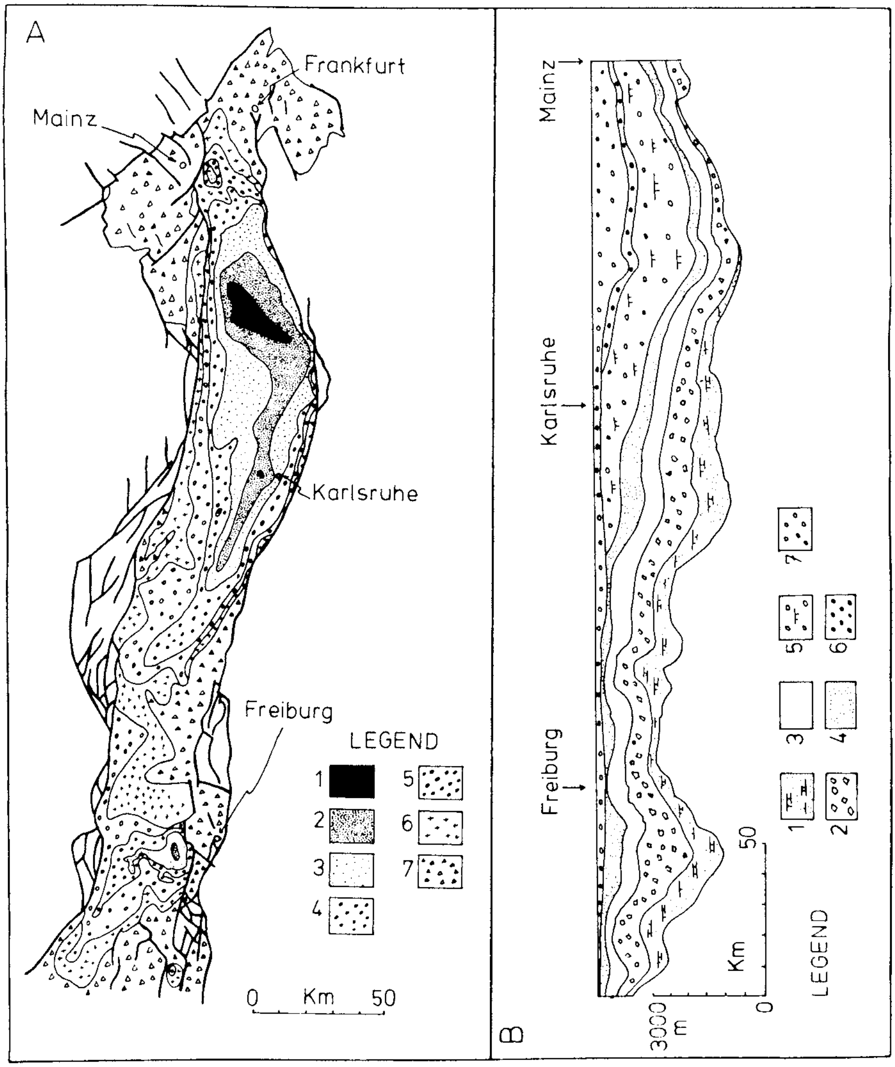

Fig. 3. (A) Distribution of the thicknesses of the Tertiary sediments in the Upper Rhine Graben trough. Heavy lines are faults. Key: 1, > 3000 m; 2, 3000 to 2500 m; 3, 2500 to 2000 m; 4, 2000 to 1500 m; 5, 1500 to 1000 m; 6, 1000 to 500 m; 7, <500 m.

(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.

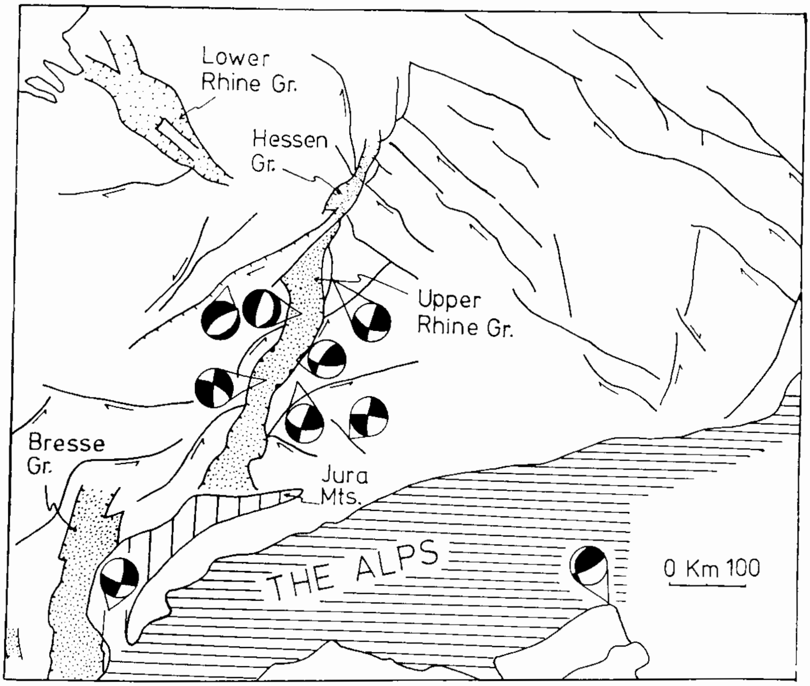

Fig. 4. Active tectonics of the Alpine Foreland in Central Europe. Compiled from Ahorner, 1970; Ahorner and Schneider, 1974; Illies, 1974b; Illies and Greiner, 1976; Müller, in press; Pavoni and Peterschmitt, 1974. In the stereographic projections of the fault plane mechanisms white quadrants are dilational, black quadrants compressional.

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 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 *result* 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 Karacalidağ 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 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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