Introduction

Over the past decade, a large amount of exploration data has been generated at rift and passive margin settings indicating that some of the basic exploration rules for these settings that control present-day exploration efforts are based on flawed assumptions.

For example, the oil industry views ocean basin as an outboard of interesting areas, understanding oceanic environment as one type of basin. Its underlying oceanic crust is understood as cooler than continental crust and as not having sufficiently thick overburden. As a result, source rocks should be immature and petroleum systems should be missing.

However, geologists have discovered a number of hydrocarbon fields on oceanic crust in the Gulf of Mexico and offshore Nigeria, and there is at least one well, far in front of the continental margin, DSDP well 368, which penetrated organic-rich shale, which is not immature. Furthermore, current academic research indicates that there are several types of oceanic crust and each of them has very different thermodynamic history. Some of them are warmer than "classical oceanic crust."

For example, the oil industry routinely uses several standard source rock maturation/hydrocarbon expulsion timing software packages, and this modeling usually utilizes heat conduction algorithms and omits the heat transfer by fluids for the time being. The reason is that the heat conduction is understood as a dominant type of the heat transfer. The heat transfer by fluids is considered insignificant in current exploration settings and, as such, ignored.

However, there are exploration data from areas such as the Canary Islands or the Espírito Santo Basin, where exploring companies noted a significant impact of geothermal fluids on source rock maturation. Other passive margins start to indicate cementation problems with otherwise good reservoir rocks in discharge areas of geothermal fluids. There are also evidences from the rift settings, such as the Rhine Graben, about fluid flow systems reducing and increasing the organic matter maturation level in recharge and discharge areas of the geothermal fluid flow systems. Furthermore, there are present-day geothermal areas in rift and pull-apart settings where the geothermal industry uses significant thermal anomalies for commercial electricity production. The research the geothermal industry has done indicates that some of the thermal anomalies can have life span, which should be long enough to modify maturation history considerably. As documented by heat flow maps of the Salton Sea region, these anomalies can express themselves even on the elevated background heat flow regime.

We also need to appreciate that widely spaced regional 2-D seismic grids can provide a misleading picture of the regional context needed for successful hydrocarbon exploration. Dense 3-D seismic-based programs, in contrast, focus on the prospect level and, consequently, do not provide sufficient regional context about the depositional section under investigation. What is required are exploration models that bridge the regional and prospect level scales to improve understanding of the elements and processes contributing to basin and petroleum systems development, and the role of their controlling factors.

Using the words of Bruce Rosendahl:

> the initial geological problem to be solved is understanding basin development through the transformation of disorganized rifting to organized seafloor spreading. Thus, we must look at the evolution from indecisive continental rift basins and imperfect accommodation zones (Tanganyika region) to pseudo-volcanic lineaments and transfer faults (Turkana region) to gradual, tortuous, and probably sub-aerial spreading (Afar region) to opening (Red Sea region), if located on the successful limb of a triple junction to discern how these environments affect petroleum system elements and processes.
>
> Of course, all of the petroliferous margins are much more complex, more segmented, and more controlled by deep thermodynamics than one tends to envision. This is especially true on continental shelf edges. Perhaps the best way to imagine these exploration terrains is as a bunch of small, fault-connected basins divided by accommodation zones and transfer faults that hoped to evolve into transform faults, and did so rather imperfectly, or not at all.
>
> As a result, one can imagine a series of basins with varying thermal regimes strung along the continental margins, usually oriented at two different oblique directions to the original margin. The kitchens may generate hydrocarbons but they may well end up on the dividing (and overlying) structures. Successful exploration in deep water margins, thus, seems to require a capability of model-driven interpretation of the imaging data, focusing on basin development,

an understanding of controlling factors to predict thermal history, and the implications of both of these factors on the presence or absence of essential petroleum system elements and processes.

Motivated by the aforementioned examples and discussion, this book tries to provide a comprehensive understanding of rifts and passive margins as a whole. The aim is to synthesize existing information devoted to specific aspects of these most important hydrocarbon habitats. This book assembles this information in one volume, in a manner that permits the use of this knowledge to assess the risks of exploring and operating in these settings.

The plan for this book originated with a three-year-long project called "Hydrocarbon exploration models in deep-water rift and passive margin settings," which took place during 2006–8, which summarized various aspects of exploration in rift and passive margin settings provided by large and diverse literature, and which addressed gaps in knowledge.

This synthesis is completed from results of personal, long-term research on rifts and passive margins, numeric validations of various concepts, and extensive tables documenting various factors influencing structural styles, thermal regimes, and petroleum systems, as well as rates of geologic processes. This book should have value for a broad range of readers, spanning from geology students, to exploration geologists, to exploration managers searching for the hydrocarbons in analogous settings. This book's strategy is to develop a comprehensive appreciation of the factors controlling structural architecture, basin development, and fluid flow in rifted margin settings. Synthesis of this knowledge is then used to tackle what is perhaps the most enigmatic aspect of these environments – the factors controlling thermal regimes and temperature peculiarities of rift and passive margin settings. The final phase of this book is the discussion of various aspects of petroleum systems based on the results of the previous structural and thermal phases. Special attention is given to the oceanic basin – the large frontier region, almost untouched by exploration so far.

As a result, this book is divided into three blocks.

The first block accumulates and synthesizes the structural and fluid flow knowledge necessary for fact-based and forward-thinking exploration plans in the rift and passive margin settings. The text begins with a basic description of structural styles, delineation of extension directions, which is necessary for the proper layout of geophysical surveys; includes determination of crust types, which are necessary for definition of different petroleum system candidates; determination of structural segmenting features, which define the sizes and geometry of a potential petroleum system; and determination of structural timing events, which is necessary for specification of source/reservoir/trap/seal temporal relations. This block is also focused on the mechanics of rifting and its transition to drifting and the controlling role of lithospheric composition and its variations on rifting and breakup. The text tries to contribute to the knowledge of why certain passive margin segments are so prolific and why others are (or should be considered) dry. Ultimately, this means a true understanding of the continuum that exists between pure rift systems and pure pull-apart basins, and how the continuum applies to petroleum systems. The structural block concludes with discussion of the role of preexisting anisotropy on rifting and breakup, and occurrences and controlling factors of associated fluid flow systems.

The second block develops the thermal foundation for successful (and unsuccessful) petroleum system determinations. The text consists of evaluation of critical factors influencing the thermal history of rocks in rift and passive margin settings, including the effects of pre-rift heat flow regime, lithostratigraphy, deposition, structural framework, erosion, deformation, and movement of geothermal fluids.

The third block develops knowledge of source rock distribution, source maturation history, reservoir distribution, seal quality, potential migration scenarios, trap candidates, and preservation issues. The text evaluates the effects of movement of geothermal fluids on diagenetic history of reservoirs and seals, and maturation history of source rocks. Knowledge of reservoir distribution works with prediction tools designed to assess continental margin vertical motion histories and the controlling factors on sediment distribution through breaches of the marginal uplifts.

1 Basic description of structural styles in rift and passive margin settings, including extension directions and key structural elements

In order to place the basins discussed in this book into plate settings, we use the respective categories from the classification of Bally (1982) (Figure 1.1):

1. Basins located on the rigid lithosphere, not associated with formation of megasutures
    - 1.1. basins related to oceanic crust formation
        - 1.1.1. rifts
        - 1.1.2. basins associated with oceanic transform faults
        - 1.1.3. oceanic abyssal plains
        - 1.1.4. passive margins
            - 1.1.4.1. passive margins overlying earlier rift systems
            - 1.1.4.2. passive margins overlying earlier transform systems
            - 1.1.4.3. passive margins overlying earlier back-arc basins
    - 1.2. basins located on pre-Mesozoic continental lithosphere
        - 1.2.1. cratonic basins
        - 1.2.2. basins located on earlier rifts
        - 1.2.3. basins located on earlier back-arc basins
2. Episutural basins located and mostly contained in compressional megasutures
    - 2.1. basins related to B-subduction zone
        - 2.1.1. circum-Pacific back-arc basins
            - 2.1.1.1. back-arc basins floored by oceanic crust
            - 2.1.1.2. back-arc basins floored by continental or transitional crust
    - 2.2. basins related to A-subduction
        - 2.2.1. basins located on continental crust (Pannonian-type basins)
        - 2.2.2. basins located on transitional or oceanic crust (western Mediterranean basins)
    - 2.3. basins related to episutural megashear systems
        - 2.3.1. Great basin–type basin
        - 2.3.2. California-type basin

Successful exploration in these basins requires a detailed knowledge of their architecture, which is the goal of this chapter.

Rift classifications

Passive versus active rifts

Based on the behavior of the asthenosphere, the rifting is driven by passive and active rifting components (Sengör and Burke, 1978; Figure 1.2), which are understood as rather end-members of the natural rift types driven by the mixture of these components (see, e.g., Huismans, 1999; Davis and Kusznir, 2002; Manatschal, 2004; Huismans and Beaumont, 2005; Simon and Podladchikov, 2006).

Passive rifting is characterized by passive asthenospheric upwelling in response to overlying layers separation controlled by regional tectonic extension (Keen, 1985; Ruppel, 1995; Kincaid et al., 1996). In this case, the lithosphere is thinned only in response to extension. The passive asthenospheric upwelling drives many secondary processes such as decompression-driven melting, crustal/lithospheric magma underplating, eruption of continental flood basalts, the onset of secondary convection, and development of large thermal gradients between extended and unextended regions (e.g., McKenzie and Bickle, 1988; Ruppel, 1995; Huismans et al., 2001).

Active rifting is characterized by active asthenospheric upwelling in response to buoyancy instability, which drives the rifting. This upwelling exhibits broad spatial wave lengths in the order of 200–500 km in proportion to asthenospheric thickness (Ruppel,

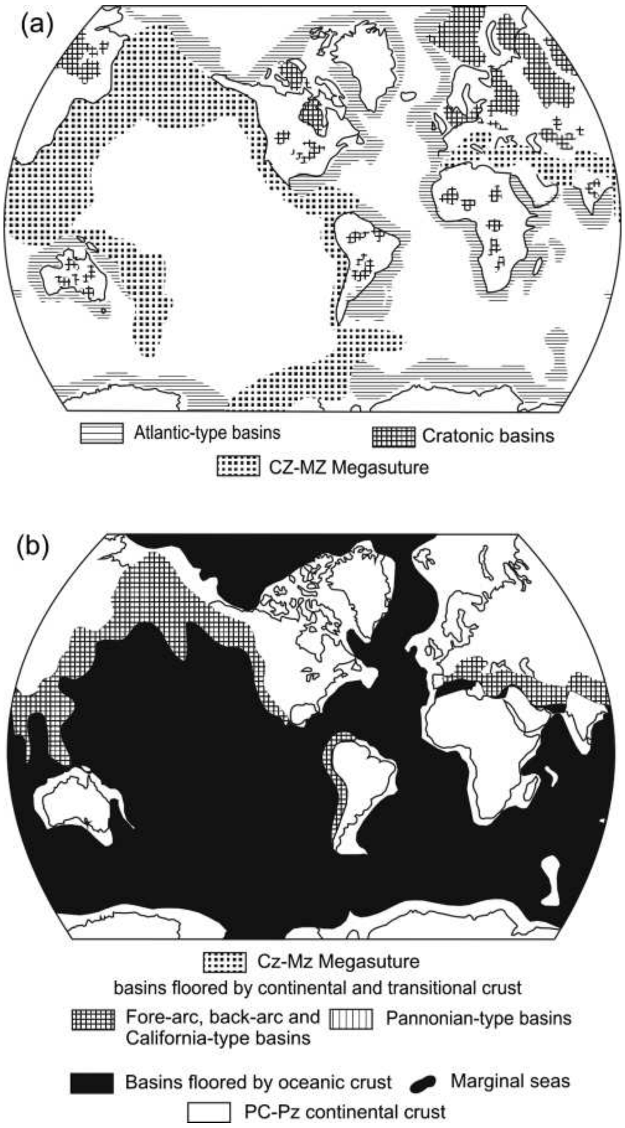


Figure 1.1. Extensional basin classification (Bally, 1982).

1995; Barnouin-Jha et al., 1997). In this case, tensional stresses at the base of the lithosphere are driven by the ascending convecting material (Turcotte and Emerman, 1983); the lithosphere is thermally thinned by heating and adsorption into the asthenosphere, in addition to necking in response to extension, and hence, the volume of asthenosphere rising into the lithosphere exceeds the volume of lithosphere displaced laterally by extension (e.g., Olsen and Morgan, 1995).

Narrow rift, wide rift, and core complex rift modes

The rift styles can be divided into narrow rift, wide rift, and core complex rift modes (Brun and Choukroune, 1983; England, 1983; Buck, 1991; Hopper and Buck, 1996; Brun, 1999 and references therein), reflecting how the lithospheric rheology, lithospheric thickness, existence of mechanic anisotropies, rate of extension, direction of extension, and presence of magma affect their development.

The narrow rift is characterized by the thinning of both crust and lithospheric mantle occupying a narrow zone of up to 100–150 km wide. This style has been attributed to local weakening factors such as thermal lithospheric thinning, local strain weakening affecting the strong layers of the lithospheric multilayer, or local magmatism (England, 1983; Kusznir and Park, 1987; Buck, 1991, 2004). It is characterized by the extension being localized in the weakest and thinnest region. As a consequence, narrow rifts have large lateral gradients in crustal thickness and topography (Corti et al., 2003). The localized lithospheric thinning in narrow rifts is associated with elevated heat flow within rift depressions in comparison to adjacent rift shoulders and cratonic blocks (e.g., Bonatti, 1985; Ruppel, 1995).

Examples of narrow rifts come from the Bajkal Rift (Artemjev and Artyushkov, 1971), the East African Rift system (Rosendahl et al., 1986; Rosendahl, 1987), the Recôncavo-Tucano Rift (Destro et al., 2003a, 2003b), the Gulf of Suez–Red Sea Rift (Bonatti, 1985; Steckler et al., 1988; Patton et al., 1994), and the Rhine Graben (Brun et al., 1992).

The wide rift is typical for its thinning of both crust and lithospheric mantle in a zone that is wider than the lithospheric thickness. Its distributed deformation style is characterized by a large number of separated basins extending across a region more than 1,000 km wide (Brun and Choukroune, 1983). Wide rifts are characterized by high extensional strain, which is not uniformly distributed over the extended region as, for example, in the Basin and Range province (e.g., Bennett et al., 1998; Thatcher et al., 1998). They can contain relatively small lateral gradients in crustal thickness (Shevenell, 2005a) and topography (USGS, 1996). Wide rifts also contain the flat base of the crust, which is attributed to magmatic underplating (e.g., Gans, 1987; MacCready et al., 1997) as contributing to smoothing Moho undulations known from narrow rift terrains (see Brun et al., 1992 and references therein).

This style has been understood as controlled by:

a) a local increase of integrated strength caused by the replacement of the crustal material by stronger material of the lithospheric mantle and accompanying lithospheric cooling driving subsequent extension into weaker unstretched

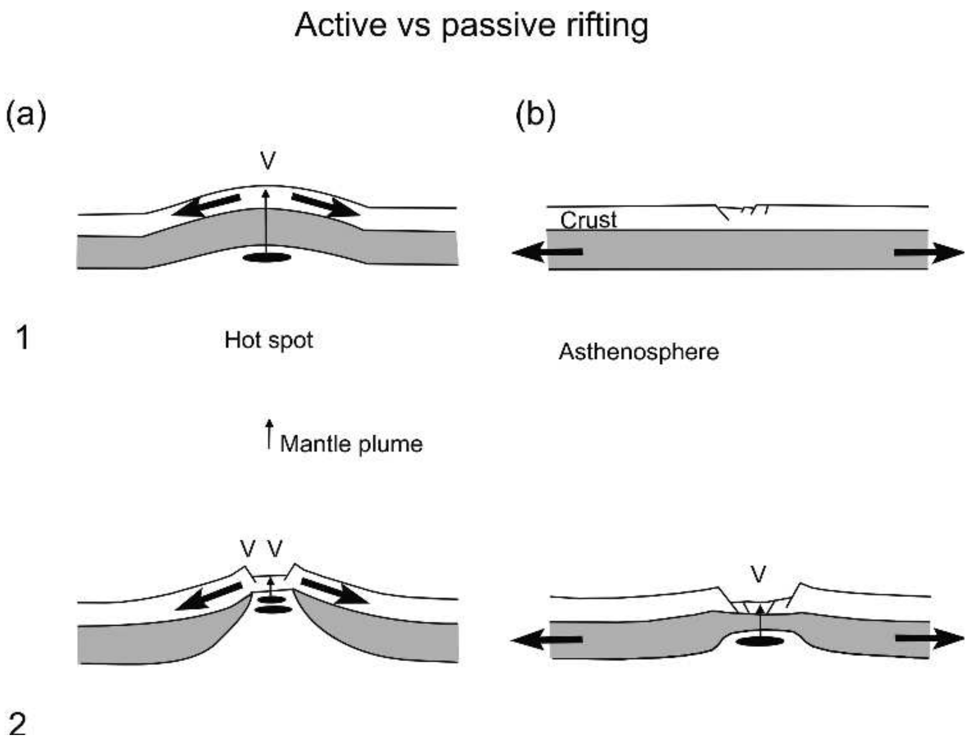


Figure 1.2. Passive versus active rifting (Corti et al., 2003).

lithospheric regions (England, 1983; Houseman and England, 1986);

b) lower crustal flow in response to pressure gradients caused by interaction of regional tectonic, local flexural, and gravity forces, resulting in delocalization of deformation (Buck, 1991; Buck et al., 1999); and

c) the degree of decoupling in lithospheric multilayers containing a viscous lower crustal layer, where the existence of localized versus distributed deformation modes depends on the coupling of the layers and the viscosity of the coupling layer (Huismans et al., 2005; Schueller et al., 2005).

Examples of wide rifts come from the Basin and Range province of the western United States (e.g., Hamilton, 1987), the Pannonian Basin system (e.g., Tari et al., 1992), the Aegean region (Jackson, 1994), and the Tyrrhenian Sea (Faccenna et al., 1996).

The core complex rift mode is characterized by the localized extension of the upper crust accompanied by the broadly distributed extension of the lower crust. This mode is attributed to a rapid lower or even middle crustal flow removing the crustal thickness variations that would otherwise control the wide rift style (Buck, 1991; Martínez-Martínez et al., 2004). In core complex structures, high-grade metamorphic rocks originating in the middle-lower crust are exposed at the surface, exhumed by shallow-angle normal faults, uplift, and erosion. They are typically separated by a detachment carrying low-grade rocks. The crust of the core complex areas has a similar thickness to the crust of surrounding less-extended areas (Corti et al., 2003). A close association of core complexes with wide rifts led Brun (1999) to describe them as local anomalies within wide rifts, developed above heterogeneities in underlying ductile lower crust, which are weak enough to localize stretching. Consequent local enhanced brittle crustal thinning becomes compensated by the upwarping and exhumation of the deep crustal levels. As Martínez-Martínez and colleagues (2004) documented, regions that underwent orogenic thickening and that are characterized by the brittle-to-ductile crustal thickness ratio of 1:3 are especially prone to development of core complexes (e.g., Coney, 1980; Crittenden et al., 1980; Lister and Davis, 1989). However, core complexes have also been described in different extensional settings such as the East African Rift system (e.g., Talbot and Ghebreab, 1997; Morley 1999; Ghebreab and Talbot, 2000), the Tertiary rift basins of the Thai Peninsula (Morley et al., 2001 and references therein), or ultra-slow and slow seafloor-spreading ridges (e.g., Cann et al., 1997; Tucholke et al., 1998; Escartín et al., 2003).

Cambridge University Press
978-1-107-02583-7 — Rifts and Passive Margins
Michal Nemčok
Excerpt
More Information

Examples of core complexes come from the Basin and Range province (Wernicke et al., 1988; Hurlow et al., 1991; Wdowinski and Axen, 1992), the Hellenic Arc (Dinter and Royden, 1993, Fassoulas et al., 1994), and the Pannonian Basin system (Tari et al., 1992).

Rift system components

The largest architectural component of the rift terrains is a rift system. An example comes from the East African Rift system (Figure 1.3a; Rosendahl, 1987). The rift systems are generally agreed as evolving in a more-or-less continuous manner without an interruption of a different tectonic process.

The rift system can be divided into rift branches (Figure 1.3b; Rosendahl, 1987) such as the Western and Eastern rift branches of the East African Rift system. They coincide with marked differences in character, whether it is the character of accompanying magmatism, detailed development timing, or opening mechanisms.

Rift branches are formed by rift zones (Figure 1.3c; Rosendahl, 1987). The average rift zone width and length in the northern South Atlantic region are 40 and 500 km, respectively. The rift zone boundaries are represented by offsets, bends, and any other major trend changes in the rift branch. A natural example of a rift zone is the Tanganyika Rift Zone of East Africa. The zones can contain several discrete basins separated by accommodation zones (Bosworth, 1985; Rosendahl et al., 1986), transfer zones (Morley et al., 1990), or interference accommodation zones (Versfelt and Rosendahl, 1989).

Each rift zone propagates until it becomes arrested because of the lack of critical stress available for further propagation. Depending on the reason for the arrest of neighbor rift zones, there are various geometries of the accommodation zones between them. As the rifting evolves, each such distinct offset of the rift branch becomes a potential nucleus for the future transform. Natural examples include transforms such as the dextral Rukwa transform connecting the adjacent Tanganyika and Malawi Rift Zones in East Africa (Chorowicz and Mukonki, 1980; Rosendahl, 1987); the sinistral Sergipe-Alagoas transform linked to the South Gabon and Recôncavo-Tucano-Jatobá Rift Zones in West Africa–Brazil (Rosendahl et al., 2005); the dextral transform connecting the adjacent Cauvery and Krishna-Godavari Rift Zones in India; and the sinistral Gettysburg-Tarfaya transform in the Central Atlantic region (Nemčok et al., 2005b).

Discrete basins of the rift zone are called rift units (Rosendahl, 1987). The average width and length in the northern South Atlantic region are 40 and 120 km, respectively. The ideal rift unit can be envisaged as an isolated half-graben bounded by a spoon-shaped normal fault (Figure 1.4; Rosendahl, 1987). The fault is characterized by the dip-slip and oblique displacements at its center and ends, respectively. A supporting systematic study of slip-vector normal faults comes from normal fault terrains in the Gulfs of Corinth and Evia, Greece, and normal fault terrains in the Central and Southern Apennines, focusing on basin-bounding faults (Roberts, 2007). The common characteristics of these faults are complex slip distributions, involving

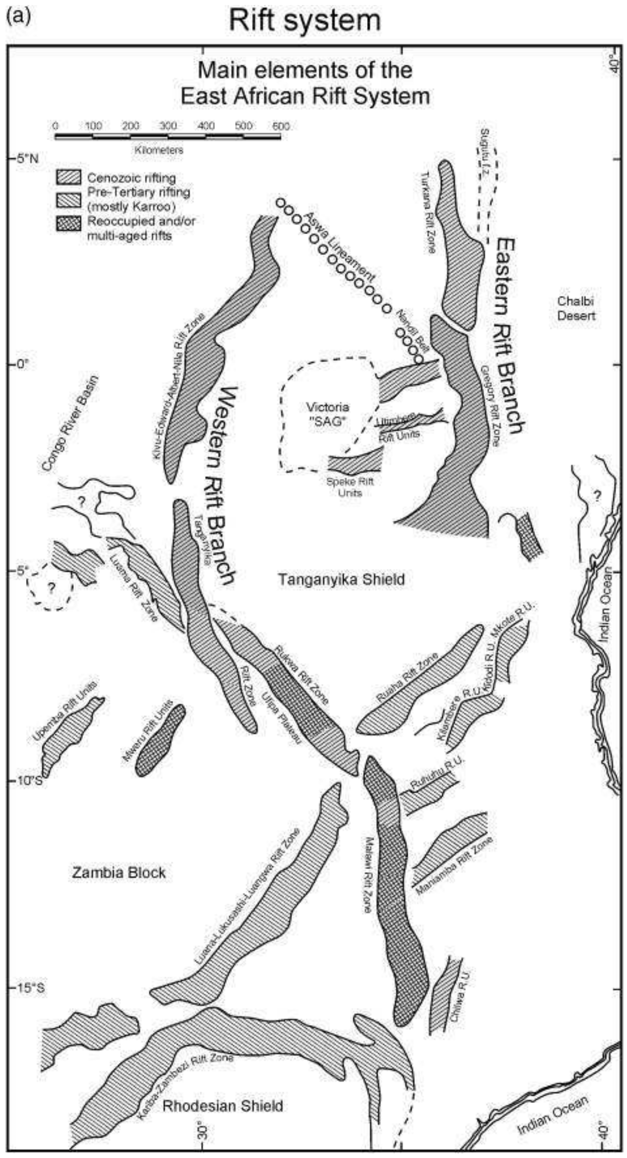
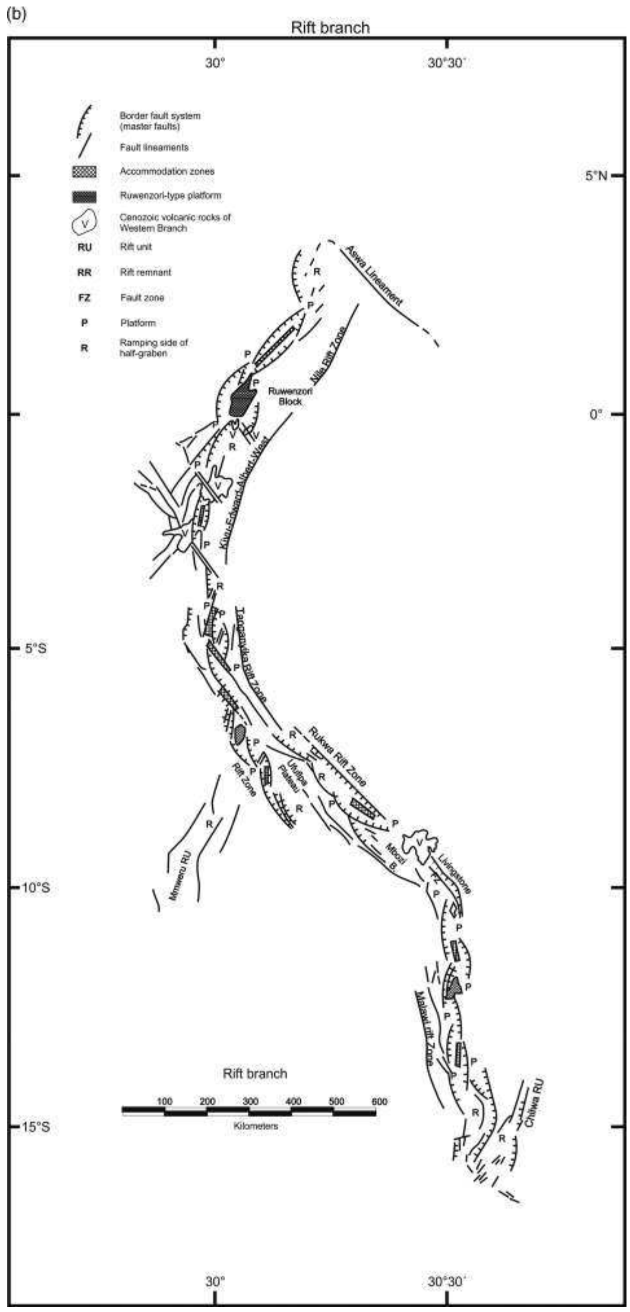


Figure 1.3. (a) Rift system example (Rosendahl, 1987). (b) Rift branch example (Rosendahl, 1987). (c) Rift zone example (Rosendahl, 1987).

Cambridge University Press
978-1-107-02583-7 — Rifts and Passive Margins
Michal Nemčok
Excerpt
More Information

converging patterns of slip toward the hanging walls (Figure 1.5). All faults can be characterized by dip-slip vectors dominating in central portions, where the overall displacement is largest, and oblique slip vectors typical for lateral propagation tips, where the displacement converges to zero. Extensive fieldwork documents that these converging slip patterns are the result of single-phase tectonic events, which is in accordance with the results of earthquake research in the Gulf of Corinth (e.g., Jackson et al., 1982). Striae and corrugation populations in various studies of faults indicate that the total displacement is a result of repeated earthquakes, each of them having a converging slip pattern (Roberts and Ganas, 2000). This observation implies



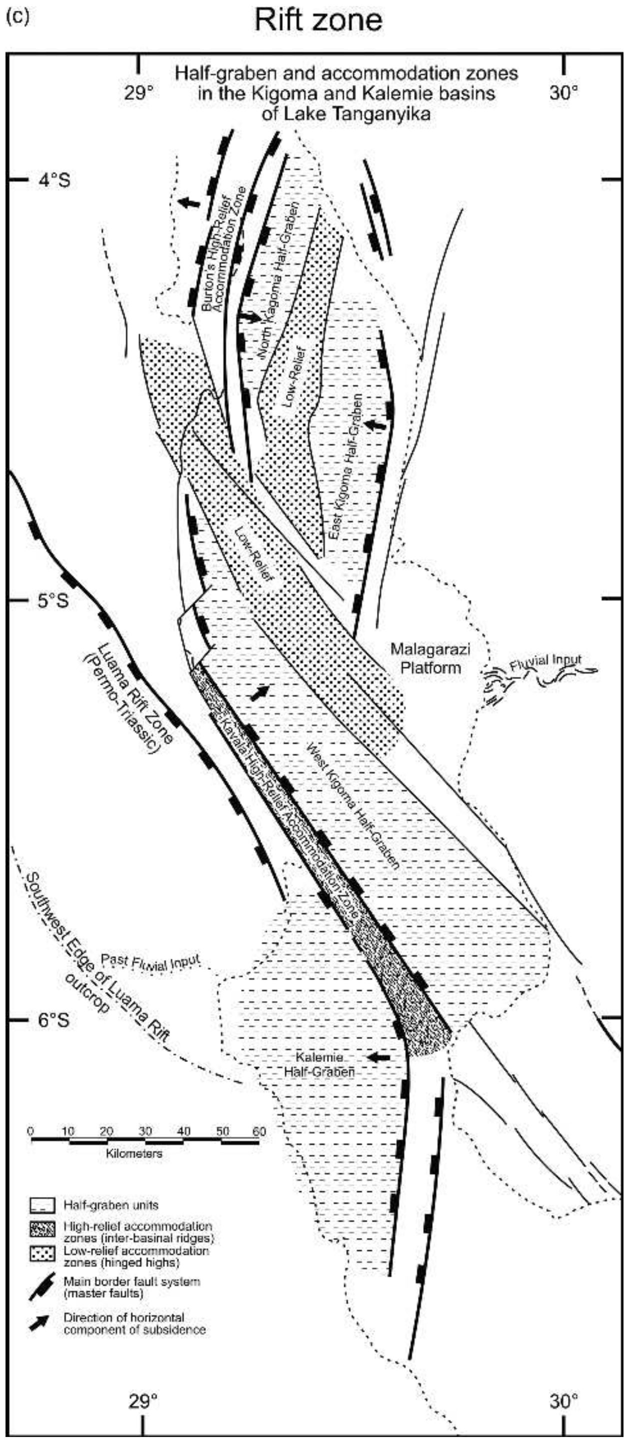


Figure 1.3 (*cont.*)

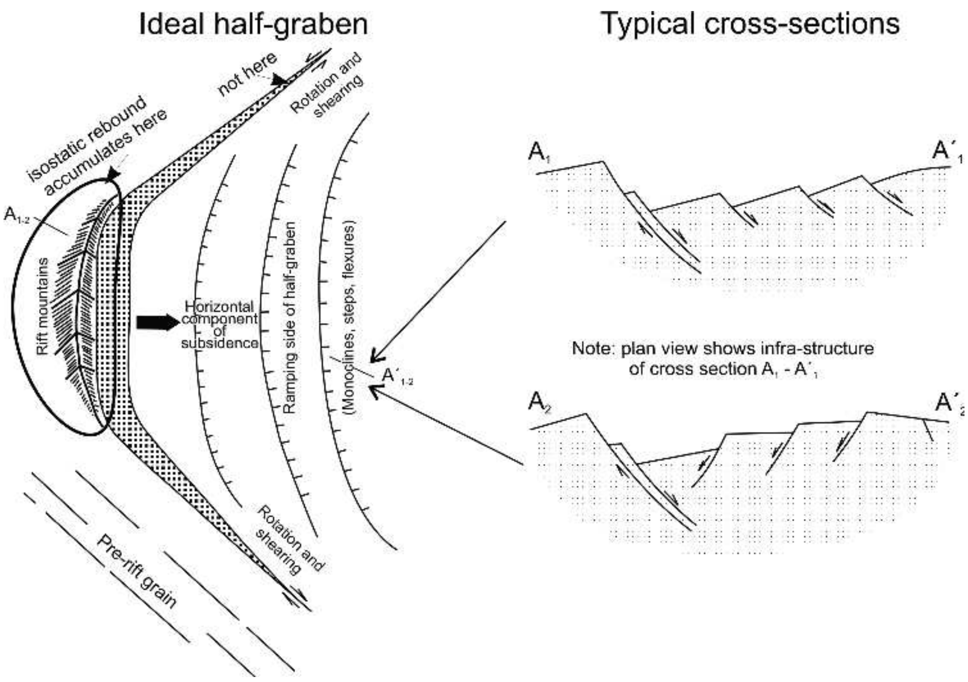


Figure 1.4. Sketch of the ideal rift unit (Rosendahl, 1987).

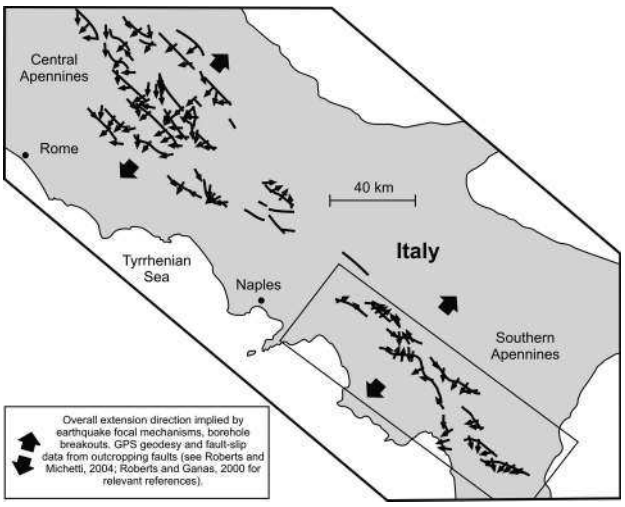


Figure 1.5. Examples of normal faults with converging patterns of slip toward the hanging walls (Roberts, 2007).

spatial and temporal stress field complexity, which causes slip in the direction of shear stress resolved in studies of spoon-shaped faults.

The rift unit footwall undergoes the largest uplift next to the fault with dip-slip displacement. Natural examples come from the Western branch of the East African Rift system (Rosendahl, 1987). The rift unit hanging wall undergoes the largest subsidence next to the fault with dip-slip displacement. Natural examples come from the Strathspey-Brent-Statfjord half-graben, North Sea (McLeod et al., 2002). The ideal spoon-shaped fault case does not occur very frequently in nature because of the influence of preexisting anisotropies and rift unit linkages on the geometry of the boundary faults. The spacing of the boundary faults varies between 2 and 15 km (Angelier and Colletta, 1983; Rosendahl, 1987). The spacing of the internal synthetic and antithetic faults is usually denser.

The rift units are composed of rift blocks (Rosendahl, 1987). This smallest unit of rift architecture has the average width and length of 10 and 40–100 in the East African Rift system (Rosendahl, 1987). The rift blocks can form structural highs or lows, but they can be frequently tilted and rotated, changing their topographic expression along their trend, as can be seen at the North Gabon passive margin.

Linkages of rift architecture elements

Half-grabens forming the rift zone can be linked in various ways, which can be grouped into linkages of half-grabens with opposing and similar polarities (Figure 1.6; Rosendahl, 1987). Both categories can be further subdivided into overlapping and non-overlapping ones.

If the opposing overlapping half-grabens develop roughly coevally, the space problem in the linkage area results in the development of a positive structure known

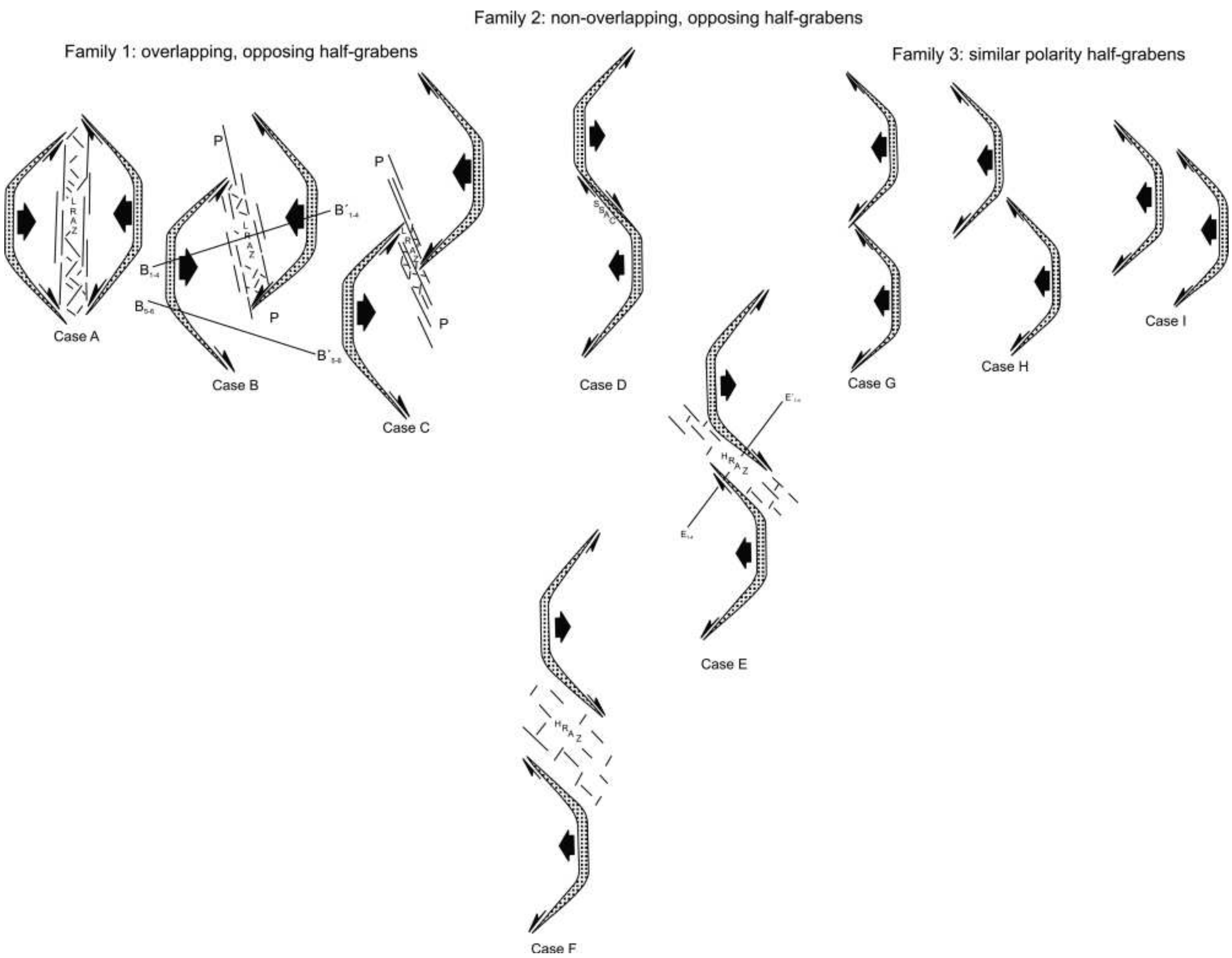


Figure 1.6. Half-graben linkages grouped into families (Rosendahl, 1987). Opposing geometries create either low-relief or high-relief accommodation zones (LRAZ or HRAZ), depending on the extent of overlap. Strike-slip accommodation zones (SSAC) may be the intermediate case.

as a low-relief accommodation zone (Figure 1.6, cases A–C; Rosendahl, 1987; McClay et al., 2002). The subsidence of these hinged highs pales in comparison with that at depocenters near graben-bounding faults (Rosendahl, 1987). Seismic data from Lakes Tanganyika and Malawi (Rosendahl, 1987) and outcrop data from western Tasmania (Noll and Hall, 2006); the Gulf of Corinth, Greece (Jackson et al., 1982; Roberts, 2007); and the Apennines, Italy (Roberts, 2007) indicate that even in the case of roughly coeval grabens, neighbor faults move individually and not exactly simultaneously. Being used in the breakup of the upper brittle crust, the low-relief accommodation zones are also widely recognized on passive margins. Examples come from the Sergipe-Alagoas and Barreirinhas Basins of Brazil (de Azevedo, 1991; Mohriak et al., 1998) and the offshore regions of Nova Scotia and Morocco (Nemčok et al., 2005b).

If the opposing non-overlapping half-grabens are active at roughly the same time, there is no space problem to be solved in the linkage area. This area tends to be “left behind” as a relatively unsubsided structure called a high-relief accommodation zone (Figure 1.6, cases D–F; Reynolds, 1984; Burgess et al., 1988; Rosendahl, 1987). In cross-sections, the high-relief accommodation zones are represented by prominent structural highs (Rosendahl, 1987). All high-relief accommodation zones in the Western branch of the East African Rift system are apparently located along preexisting anisotropies. Switches in half-graben polarities in the Gulf of Suez–Red Sea Rift system occur where rift-bounding faults impinge on or are influenced by preexisting basement structures such as the Precambrian shear zones.

Being used in the breakup of the upper brittle crust during continental breakup (Rosendahl, 1987), the high-relief accommodation zones are also widely recognized on passive margins. Examples come from offshore Benin, Ghana Ridge, Guinea Plateau; offshore Côte d'Ivoire; the Acaraú, Mundaú, Para-Maranhão, and Piauí Basins of the Brazilian Equatorial Atlantic margin (Nemčok et al., 2012a); and offshore Nova Scotia (Nemčok et al., 2005b).

Accommodation zones between opposing graben-bounding faults contain en echelon horst-graben faulting (Moustafa, 2002) or linking transfer faults (e.g., Bally, 1982; Rosendahl, 1987).

If the half-grabens of the same polarity are active at roughly the same time, there are space problems to solve in the linkage area. A large overlap area is affected by tilt, which is driven by uplift of the footwall in relationship to the first bounding fault and subsidence of the hanging wall in relationship to the second bounding fault. Areas characterized by a small overlap develop either direct linkages of the neighbor boundary faults (Noll and Hall, 2006) or relay ramps in the overlap between them (Figure 1.7; McLeod et al., 2002).

Figure 1.7. Small-scale relay ramp, Arches National Park, Utah.

Detailed fracture/fault patterns in accommodation zones between graben-bounding faults of the same polarity are (Moustafa, 2002):

a) relay ramp
b) linking transfer fault
c) en echelon step faulting
d) tilted fault blocks
e) zigzag fault array.

A well-constrained example of the spatial and temporal variation of activity of normal faults with the same polarity and located above the same detachment fault comes from the Corinth-Patras Rift, Greece (Sorel, 2000). The active low-angle, north-dipping detachment fault is indicated by micro-earthquakes occurring on a 12°–20° plane dipping to the north. The timing of the faults above detachment was derived from ages of sediments in respective half-grabens, which are the oldest in the south and become younger northward in jump-like fashion (Sorel et al., 1997; Sorel, 2000; Figure 1.8). The whole described extensional history took place during the last 1 Ma (Sorel, 2000). During this time, the total distance traveled by the rolling hinge of the system was about 30 km.

The northerly shift of the normal fault activity above the detachment fault was accompanied with rotation of the detachment fault into shallower dip geometry in the soon-to-be non-active portion of the system, which expanded from south to north (Sorel, 2000). The rotation caused the higher-friction geometry, which eventually led to stalling of the detachment fault, which progressively expanded to the north.

The lack of deeper seismicity north of the Gulf of Corinth along the detachment fault probably indicates its ductile character at this depth. Earthquakes occur only along the southern portion of the detachment fault, where the lower dip angle results in higher friction

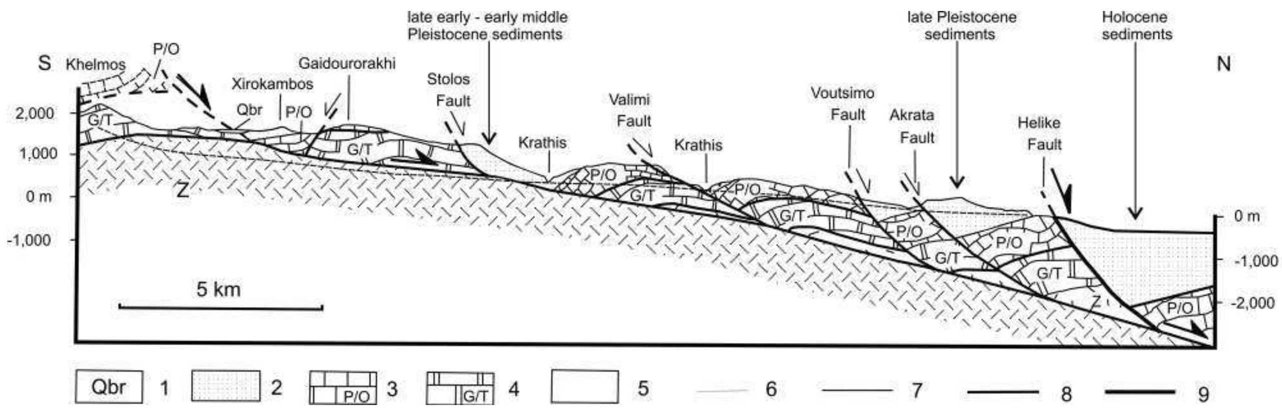


Figure 1.8. Extensional basin system in the Gulf of Patras (Sorel, 2000).