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Tectonic Inheritance in Continental Rifts and Passive Margins

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Dedicated to Chris Talbot (retired Professor: Uppsala University) for growing our interest in tectonics

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Contents

Abstract

Tectonic inheritance deals with the influence of pre-existing or pre-rift elements on the geometry, genesis and propagation of rift-related faults. Inheritance strongly controls the architecture of continental rifts and passive margins. Experimental results demonstrated the importance of layering and mineralogical anisotropy in extensional deformations. For low-to-intermediate angles of the anisotropy to the maximum compression direction, faults formed within anisotropic rocks parallel to the pre-existing weakness. For high angles, the faults breach the weak planes but follow them in segments. Rocks usually are anisotropic and respond to extension more easily than to compression. Shallow anisotropies at the brittle upper crust are either pervasive or discrete. While foliations and layers define ‘pervasive’ fabrics, widely spaced isolated zones of weakness such as faults and shear zones define the ‘discrete’ ones. Pervasive fabrics govern the overall trend of the rifts in passive margins. The discrete fabrics form oblique to rifts or as transfer zones between propagating rift segments. Rheology of the pre-rift lithosphere controls the architecture of rifts and passive margins predominantly for levels deeper than the upper crust. The parameters controlling the architecture of rifts and passive margins are strength, crustal and lithospheric thicknesses, thermal state and strain rate. The first three factors are soft-linked. For example, the strength of the lithosphere depends on its composition, thickness and temperature (van der Pluijm and Marshak 2004). The thickness of the lithosphere—thicker for mobile belts and thinner for cratons—depends on the thermal age (=age of last tectonothermal event). Lithospheric thickness thus influences its thermal state also. Generally, rifting in thicker lithosphere diminishes rift shoulder topographies, whereas rifting in colder and thinner lithosphere forms $\sim$3–5 km elevated rift shoulders. Warmer lithosphere produces rifts narrower and faster than those within colder lithosphere. In this work, we bring together the concepts of the inheritance of pre-rift shallow (pervasive and discrete fabrics) and deep (lithosphere rheology) elements. Citing examples from intra-continental and rifted passive margins, we show that the process of tectonic inheritance remains active throughout the rifting episode.

Keywords Tectonic inheritance · Pre-existing anisotropies · Pervasive fabrics · Discrete fabrics · Lithosphere rheology · Rifting

Highlights

1. Tectonic inheritance of pre-existing anisotropies in rifting is universal.
2. Pervasive fabrics have a mode of influence different than discrete/isolated fabrics.
3. Lithosphere rheology is important in controlling the geometry, genesis and architecture of rifted basins.

Chapter 1
Introduction

> *...inherited fault zones can clearly influence the pattern of continental deformation at length scales greater than the thickness of the lithosphere.*
>
> Sutherland et al. (2000)

1.1 General Aspects

Rifts are elongate zones of crustal/lithospheric extension. These could extend for thousands of kilometers and are controlled by fold belts within the basement (Gibson et al. 2013). Rifted passive margins/trailing margins (Dennis 1987) depict a range of attitudes, lengths, linkage and segmentation of rift bounding and minor faults (Stuart et al. 2006; Tugend et al. 2013). Rift zone geometries are controlled by litho-layers and extension rates besides geomorphological processes (Ramirez-Arias et al. 2012; Buiter 2014; Huismans et al. 2014). Typical structure of continental rifts are sediment-filled (half/) grabens (Fig. 1.1; Bott 1995; see Morley 1999a for detailed geometries of the East African Rift System). The grabens are bound by segmented, generally listric, normal faults. Adjoining faults may dip oppositely. The area between the two normal faults is called an 'accommodation zone' or a 'transfer zone' (Fig. 1.1; Bosworth 1986; Rosendahl 1987), such as that found in the Gulf of Mexico basin (Palmes 2005), the Suez rift (Moustafa 1996), and the Bay of Biscay-Pyrenees region (Tugend et al. 2014). Review and comparison of passive rifts are available in Melankholina (2008), Melankholina and Suschevskaya (2008), Levell et al. (2010) etc. Role of inheritance in rifting has been described as repetitive throughout geological time in the North American Atlantic by Thomas (2014).

How rifts get segmented, propagate discontinuously and evolve is decided by pre-existing structural features within the basement (e.g. Branco Farnandez et al. 2010). This control is called 'tectonic inheritance' (Fossen 2013). In one way, the entire Wilson cycle is controlled by inheritance (Vetel and Le Gall 2006; Stuart et al. 2006; Adet et al. 2011). Rotation of plates can reactivate prior structures (Geoffroy et al. 2014). Role of heterogeneities in controlling structures in rifts, passive margin

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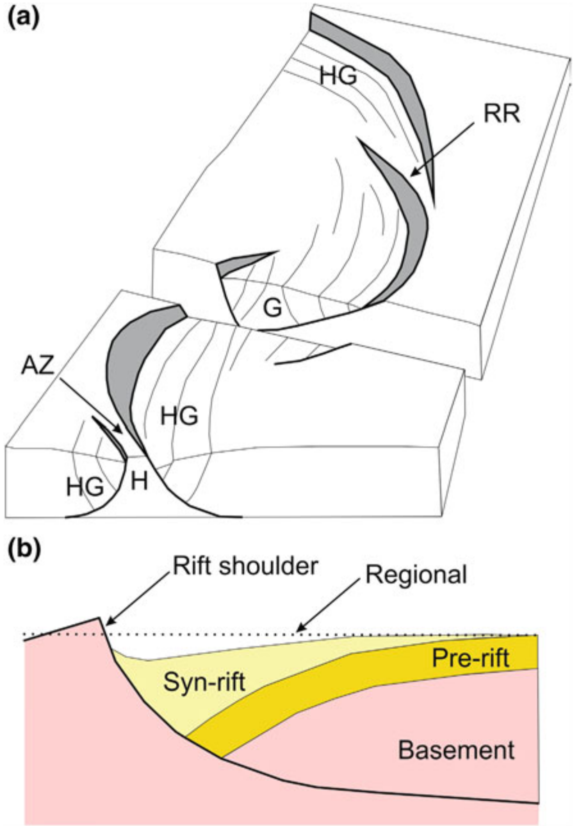

Fig. 1.1 **a** Schematic diagram of common geometries of continental rifts with oppositely dipping faults. The accommodation/transfer zones (AZ) may form horsts (H), grabens (G), or relay ramps (RR) depending on the position and dip of the two faults. *HG* half grabens. **b** Sketch of a rifted half graben showing rift shoulder uplift, demarcated by the rising of the rift shoulder above the "regional elevation"

and hyperextended rift systems contributing to inheritance have been identified in numerous studies (e.g. Cloetingh et al. 1995; references in Morley 1999b, c; Morley et al. 2004; Bellahsen and Daniel 2005; Corti et al. 2007; Aanyu and Koehn 2011; Bellahsen et al. 2013; Cappelletti et al. 2013; Manatschal et al. 2015 etc.). '*Tectonic heritage*' (Missenard et al. 2007), and 'reworking' and 'reactivation' (Ashby 2013) are also used synonymously with inheritance. Older rifts might be *inherited* in later orogeny (Nirrengarten et al. 2014), and we do not discuss those aspects here. Variations in isostasy, depth of strong layer etc. for the entire (mega-) regional passive margin have been noted (Pazzaglia and Gardner 2012). One of the factors for such varied structural styles is the tectonic inheritance of pre-existing anisotropies in the pre-rift lithosphere (Mosar 2003; Leroy et al. 2013). Rifts (and transform/shear margins) can inherit from pre-existing fracture zones (Greenroyd et al. 2008; Tsikalas et al. 2012) or from prior (strike slip) fault zones (Masini et al. in press), foliations, lineaments, bedding planes and shear zones (Cloetingh et al. 1995; Morley 1999b, c). Inherited structures can also act as detachments (Gouiza and Hall 2013), and can control mantle exhumation (Sutra and Manatschal 2012), sedimentation pattern and sub-basin formation/crustal segmentation (Correira et al. 2012;

Soares et al. 2012) within the rift, and may form shortcut structures (Mora et al. 2009). How volcanos distribute when rifting starts (Isola et al. in press), and mode of uplift in orogens (Vernon et al. 2014) can be decided by inheritance.

Transfer zones can run parallel to the pre-existing fabrics (Withjack and Schlische 2005). Such a zone could have been inherited from fractures in older basement rocks (Montenat et al. 1986), or from pre-existing grabens (e.g. Madritsch 2014). For example, complex transfer zones or 'graben shifts' related to conjugate fault sets in Pattani basin in the Gulf of Thailand were attributed to structural inheritance in two stages (Kornsawan and Morley 2002). Thus, an optimally oriented fracture in the basement can influence faults in the superjacent rocks under a stress regime. In a generalized language, therefore, one brittle structure can inherit another brittle structure. Transfer zones could inherit form previous structures (Corti et al. 2002; Heffner 2013).Continental rifts may abort and never form oceans, e.g. Cambay rift (India) or may split continents to form conjugate passive margins or either side, e.g. Iberia-Newfoundland conjugate passive margins (Shelley et al. 2005). The accommodation zones may be normal faulted and appear as horsts or relay ramps (Fig. 1.1a).

Inheritance of previously existing structures take place for strong mechanical anisotropy, less flow of heat, and for favourable angle between existing structure and the applied stress (Edel et al. 2007). As stress direction changes temporally, some pre-existing faults may reactivate after some time (Morley et al. 2007), as possibly happened in the Ethiopian rift (Muluneh et al. 2014). Anisotropy in sediment composition, phase transition of minerals etc. add to anisotropy leading to inheritance for passive margins to develop (Cloetingh and Negerdank 2010). However, when studied in greater detail, the degree of anisotropy created by some of them, foliations in particular, may not be significant as expected initially (e.g. Kocher and Mancktelow 2006; reviewed by Mukherjee 2014a). However, the difference in the degree of influence made by various anisotropies need to be studied in detail. The mechanical anisotropy can be brought within the lithosphere also by either an underlying stack of nappes (produced during orogeny; Mattioni et al. 2006), or by accretion (Tetreault and Buiter 2013). Faults strongly inherited may not obey length-slip linear relation (review in Morley et al. 2007).

Analogue models proved that the geometries of rift systems change significantly by pre-existing fabrics (e.g. Corti 2004; Aanyu and Koehn 2011; Chattopadhyay and Chakra 2013; also see Holdsworth et al. 2013), and lithospheric rheological contrasts (e.g. Cappelletti et al. 2013) along with thermal state and strength profile of the pre-rift lithosphere control the basin dynamics (Fig. A1; Buck 1991; Cloetingh et al. 1995). Experiments demonstrated that anisotropies affect strength of rocks (e.g. Donath 1961; Youash 1969). Fractures/fissures (Nogueira and Marques 2012; Sonnettea et al. 2012; Autin et al. 2013) and even thrust faults (Tavarnelli et al. 2004) within basins can concentrate due to inheritance. Inheritance may alter geometry of rifts and dip of faults (Rocher et al. 2003; Le Pourhiet et al. 2004, 2006; Mattioni et al. 2006; Labails et al. 2009; Heffner 2013). Faults may change trend when they inherit an existing fault in the basement. For example, from the passive margin of Egypt, few ENE trending faults inherited from E-W faults. Rifts that criss-cross each other are inherited products (Cawood et al. 2001). The depth and orientation of heterogeneity in the lithosphere decides the orientation and pattern of inherited faults (Mattioni et al. 2006). However, if the rheological contrast is too high between adjacent lithologies, the interface does not reactivate, such as the Rio Grande rift (Philippon et al. 2013a, b). Inherited faults might join mutually by tear faults (Bollati et al. 2012).

Inheritance is scale-independent (Holdsworth et al. 2013) and localizes shear stresses along anisotropies in the pre-rift lithosphere. Klyuchevskii (2014) considered anisotropy/weak planes and stress change together as 'rifting attractor structures' that fundamentally governs the rifting. The main processes can be clustered at shallow depth where continental lithosphere reactivates and reworks at deeper depth (Holdsworth et al. 2001; Fig. 1.2).

Refining tectonic understanding of passive margins in the context of inheritance is important, which is not well understood till date (Bellingham et al. 2014). Although the concept of passive margin can be traced back to geosynclinal theory in terms of 'miogeosynclines', the idea of inheritance was not incipient in the latter theory (Bond and Kominz 1988). Passive margins are important locales for hydrocarbon exploration (such as Beydoun et al. 1992; Davies et al. 2004), earthquakes and landslides (GEOPRISM, internet reference). These are also sites where fluids flow (Fevre and Stampfli 1992) and modify flow pattern (Holdsworth et al. 2013). Structural trend produced by inheritance can also be the selective site for metamorphic evolution (Beltrando et al. 2014).

Here we focus continental rifts (Fig. 1.3) to cite examples and skip the oceanic types. We discuss parameters like strength, crustal- and lithospheric thicknesses, effective elastic thickness, thermal state and strain rate in this context. We do not use 'inheritance' in the sense of 'remnant' nuclear chemistry in rocks (Stroeven et al.

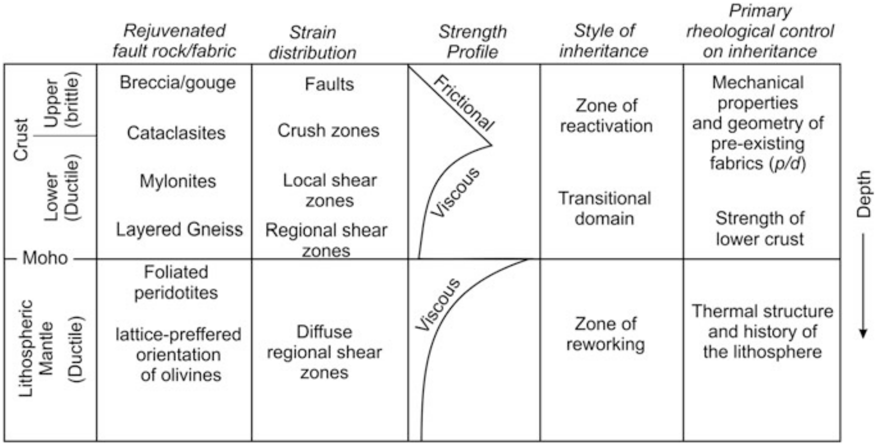

Fig. 1.2 Processes of inheritance and its relation with lithospheric layers, strain distribution, rocks/fabrics, rheological control and strength profile of the continental lithosphere. Notice no sharp boundaries exist for these processes, fabrics and properties. This scheme is built for a three-layered lithosphere comprising of a brittle upper continental crust, a ductile lower crust and lithospheric mantle. *p* pervasive fabrics; *d* discrete fabrics. Modified from Holdsworth et al. (2001)

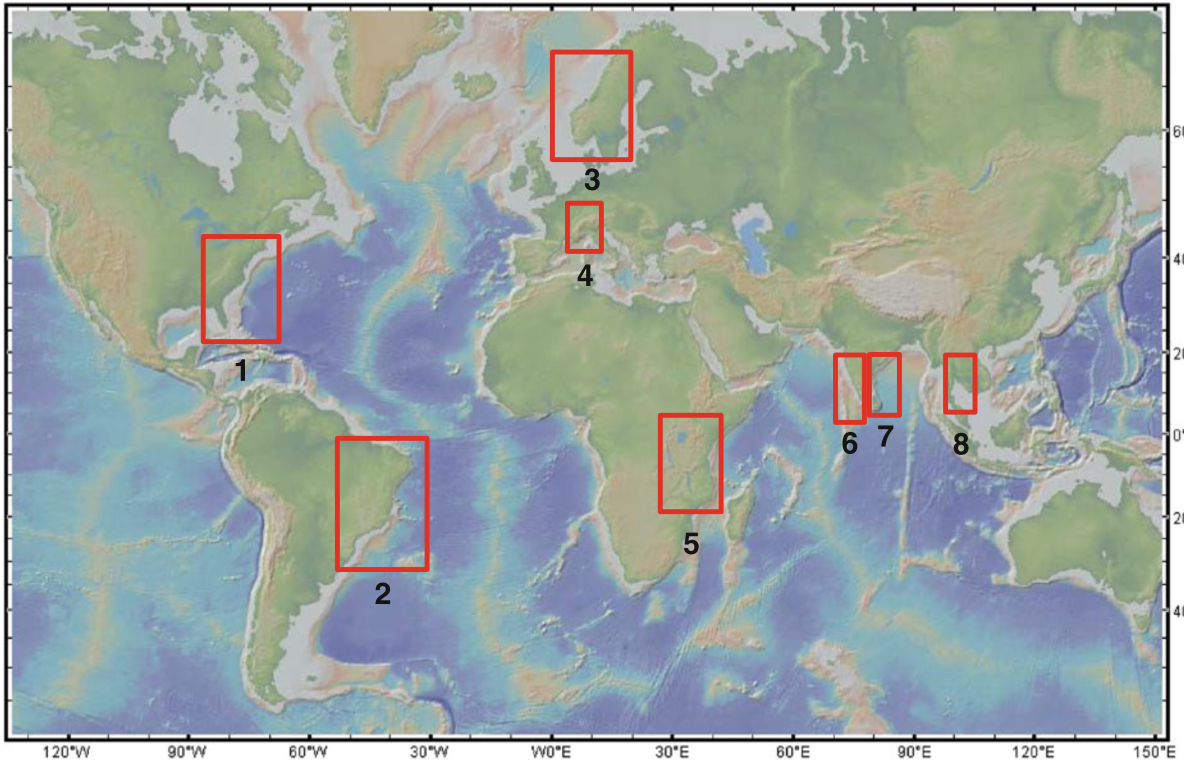

Fig. 1.3 Locations of the rift/passive margins this study reviews. *1* North American passive margin; *2* Brazilian continental rifts and passive margin; *3* Norwegian Sea and eastern North Sea; *4* Cenozoic continental rifts of Europe; *5* East African Rift System; *6* west Indian passive margin; *7* east Indian passive margin; *8* Tertiary rifts of Thailand. Map reproduced from Geomapapp (www.geomapapp.org)

2002 and many other), and avoid inherited topography in basins (Doré et al. 2002) as well. Unlike Mora et al. (2009), Strzerzynski et al. (2010) etc., we do not discuss inheritance specifically in the context of inversion tectonics where normal fault planes act as reverse faults under favourable stress regime and/or guided by sedimentation pattern (Dubois et al. 2002; Copley et al. in press). We selectively review the East African Rift System (EARS; Morley 1999a, c), Tertiary rift basins in Thailand (Morley et al. 2004), Brazilian passive margin (de Matos 1992; Ashby et al. 2010; Ashby 2012; review by Mohriak and Fairstein 2012), and East Indian passive margin (Nemĉok et al. 2007) and West Indian passive margin (Misra et al. 2015b; this study).

Besides their width and geometry (Schoettle-Greene and Pysklywec 2014), passive margins govern subsequent orogeny (as referred in Chabli et al. 2014; Mohn et al. 2014; Schoettle-Greene and Pysklywec 2014) and tectonic heterogeneity along the orogenic trend (Jammes et al. 2014). Sheared margins can be inherited from one- (Woodcock 2012) or more than one previous orogenies (El Harfiet al. 2006). More than one set of structural trends from grabens might be inherited from prior orogeny (Bache et al. 2010)/tectonism (Huetra et al. 2012). The latest orogenic imprints are usually inherited strongly in the subsequent rifts (Harry et al. internet reference).

Passive margins can influence rheological properties of an overlying foreland basin during inversion (Watts 2001). Sedimentation pattern in rifted basins could be controlled by inheritance (Purser and Bosence 1998). Two oblique fabrics in the lithosphere can reactivate and govern the location and the geometry of depo-centers inside the rift (Michon and Sokoutis 2005). Basin evolution (Leeder 1982) is controlled by inheritance, which eventually weakens the passive margins/rift zones (Audet and Bürgmann 2013).

1.2 Gaps in Knowledge

Although tectonic inheritance has been well-studied in compressional settings (e.g. Miller et al. 2001) and a few arc systems (Comas et al. 2014), a concise work on passive margins in relation to inheritance does not exist (Watts 2012). For example, role of inheritance in producing amagmatic/magma-poor and magmatic/magma-rich passive margins is unknown (Bott 1995; Rodgers and Bally 2012; Seiler et al. 2013). Put in another word, what fraction of deformation was controlled by inheritance in magma-rich passive margin is unclear (Olsen 1995; Vetell and Le Gall 2006). Recently, Sutra and Manatschal (2012) pointed out qualitatively link between inheritance and magma poor rifts. The quantification does not exist. Specifically, whether and how the magmatic NE Greenland margin (Helwiget et al. 2012) is linked with Caledonian orogeny is not known. How exactly inherited rifts initiates is unclear (GEOPRISM, internet reference). Notwithstanding some of the reviews on passive margins do mention the role of inheritance (e.g. Le Pichon and Sibuet 1981; Odegard 2005; Levell et al. 2010; Le Pourheit et al. 2013; Reber et al. 2013; Alves et al. 2014), they do not address those issues.

Chapter 2
General Aspects

Foliations, lineations and fractures can develop into inherited structures (Coussement et al. 1994). Basins/margins that underwent inheritance attain mature stage quite faster than those without inheritance (Holdsworth et al. 2013). Rock deformation tests confirmed that the faults/shears formed experimentally in anisotropic rocks disobey the common failure criteria, the Coulomb failure criterion and the Anderson's theory of faulting, for certain angular relations between the anisotropy and stress orientation. Those faults/shears rather follow/inherit anisotropies in such cases (like Donath 1961; Youash 1969; Shea and Kronenberg 1993; reviews by Paterson and Wong 2005). Thus, the geometries of faults that are inherited from prior structures within passive margins could be non-Andersonian (Brun and Autin 2013), e.g. steeply dipping reverse faults and low dipping normal faults are possible. Along with mantle plumes, inheritance work in non-unique ways in shaping individual rifts (Achauer and Masson 2002).

Inheritance/reactivation of faults along more than one direction (Montenat et al. 1986) is quite possible that can separate basins into segments (Branco Farnandez et al. 2010). Secondly, specific set of faults have been identified from rift basins to be product of inheritance (San'kov et al. 1999). Rifts and suture lines can parallel and inherit from basement discontinuity (de Graciansky et al. 2011). Suture zones in the basement can define trends of inherited rifts (Al-Amri 2013). Transfer zones can run parallel to the pre-existing fabrics (Withjack and Schlische 2005). Such a zone could have been inherited from fractures in older basement rocks (Montenat et al. 1986), or from basins/depressions in the pre-rift basement (Madritsch 2014). For example, complex transfer zones or 'graben shifts' related to conjugate fault sets in Pattani basin (Gulf of Thailand) were attributed to structural inheritance in two stages (Kornsawan and Morley 2002). Thus, a fracture in the basement can inherit/influence fault in the superjacent rocks under a stress regime. In a generalized language, therefore, one brittle structure can inherit another brittle structure. Transfer zones could be inherited form previous structures (Corti et al. 2002; Heffner 2013).

However, the rift axis need not always parallel weak inherited crustal lineaments (Beaumont and Ings 2012). Also note that a lineament might get reactivated multiple times (Copley et al. 2014). Extensional direction for rifting may not parallel planes of pre-existing weakness in the basement (Odegard 2005). The passive

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margin then cuts the trend of the mobile belt at high angles e.g. in the equatorial Atlantic conjugate margins of Africa and South America (see Sect. 4.2.3). The propagation direction of basins need not exactly parallel the inherited structure leading to transtension (Odegard 2005; Pereira et al. 2012; Gernigon et al. 2013; Holdsworth et al. 2013).

Inheritance in the context of extensional tectonics has been described by previous workers by a number of terms. Below we use Morley (1999a, b) for pre-existing fabrics in the shallow crustal domains, and Cloetingh et al. (1995) and Holdsworth et al. (2001) for those in the deeper realms in the context of lower crustal strength and mantle anisotropy.

Chapter 3
Influence of Pre-existing Anisotropies on Fault Propagation

3.1 Fracture Criteria

Brittle failure of rock follows as cohesion of the material is lost and the rock ruptures along a surface/zone. This plane, called the brittle fracture plane, is the first required structure for subsequent faulting within rocks. Experiments show that two basic types of fractures are generated by brittle failure viz. tension-/Mode-I fractures and shear-fractures/Mode-II fractures. The tension fracture planes or tensile joints develop perpendicular to the minimum principal stress axis (σ_3) and is governed primarily by the Griffith failure criterion. In confined-compression experiments, conjugate shear-fractures develop at an acute angle on both sides of the maximum principal stress axis (σ_1) and is governed by the Coulomb failure criterion (Fig. 3.1; see Fossen (2010); Davis et al. (2012) for details). The Coulomb fracture criterion for a given rock predicts the state of stress at which it is at the verge of failure i.e. a "critically stressed" rock. This is also known as the Mohr-Coulomb failure criterion (Fig. 3.1). It states:

$$\sigma_s = c + \mu \sigma_n \tag{3.1}$$

here

$$\mu = \tan\varphi \tag{3.2}$$

φ: angle of internal friction
where, σ_s: critical shear stress, σ_n: normal stress; μ: frictional constant or coefficient of internal friction, is the cohesion on the plane across which the normal stress, i.e. σ_n = 0. c: also known as the 'cohesive strength' and has its equivalent T, critical tensile strength in the tension domain, i.e. for $\sigma_n < 0$ (Fig. 3.1). Thus, the failure properties of rocks depend on 'c' and 'μ' (as in Eq. 3.1). Additionally, the shear

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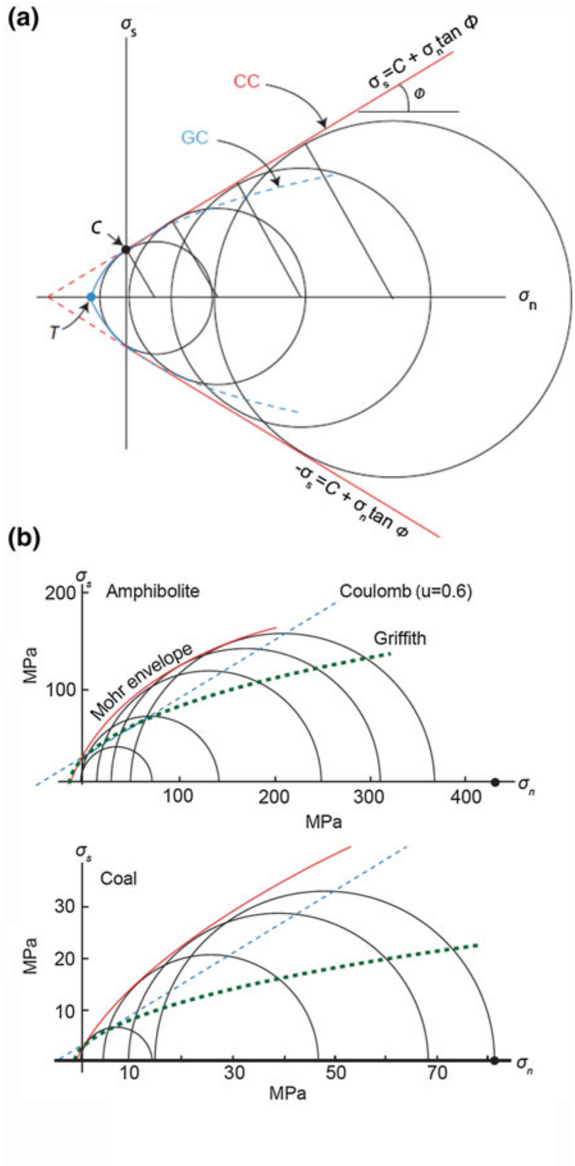


Fig. 3.1 σ_N: normal stress, σ_S: shear stress. **a** The Coulomb fracture criterion occurs as two *straight red lines* in the Mohr space. The *circles* represent examples of critical states of stress. *Blue line* the Griffith criterion. The combination of the two is sometimes used: Griffith criterion (*GC*) in tensile- and Coulomb criterion (*CC*) in compressional regime. C: **c**ohesive strength; T: tensile strength (from Fossen 2010). **b** The Griffith**-** and the Coulomb fracture criteria superimposed on experimental data. The criteria are placed so that they intersect the vertical axis (σ_S) together with the Mohr envelope. Neither of the criteria fit the data accurately. The Griffith criterion works well for tensile stress (*left to origin*), but shows a too low slope in the entire compressional regime. The Coulomb criterion approaches the envelope for high confining pressure: right side of the diagram (from Fossen 2010 and references therein)

stress required to start a shear fracture depends on the normal stress on the newly formed fracture plane (σ_n). With increase of σ_n, larger σ_s is required to break the rock.

The Coulomb (or Mohr-Coulomb) failure criterion for isotropic rocks always considered the maximum and minimum principal stresses (σ_1, σ_3) and ignores the intermediate principal stress (σ_2), which inherently lies on the fracture plane. The attitude of the fracture can be expressed as:

$$\varphi = (45^\circ - \alpha/2) \quad (3.3)$$

where φ: angle of internal friction of the rock; α: angle between the fracture and the maximum principal compressive stress axis σ_1, and ranges in natural unfoliated rocks from 25 to 35° (Paterson and Wong 2005; Zang and Stephansson 2010). Majority of shallow crustal rocks have $\varphi = 30°$ and so $\alpha = 30°$. That denotes dip of a normal fault to be 60°, which is in agreement with Anderson's theory of faulting. However it is to be noted that tan φ cannot be approximated directly as the coefficient of friction in the physical sense (Handin 1969) although the values range between 0.5 and 1.5 generally but tend to show higher values for coefficient of sliding friction (Patterson and Wong 2005 and references therein).

Griffith (1924) described the relation between tectonic joints and pre-existing cracks: the 'Griffith cracks' (see Fossen 2010; Davis et al. 2012). Griffith cracks are (sub-) microscopic, and are modelled as ellipses with apertures much less than their lengths, i.e. of high aspect ratios/ellipticity (Engelder 1987; Pollard and Aydin 1988; Blenkinsop 2000). On extension, stress concentrates at the edges of the cracks and cracks propagate to interconnect under tensile or shear stresses forming either macro-fractures or faults (Fossen 2010). The Griffith failure criterion states:

$$\sigma_s^2 + 4T\sigma_n - 4T^2 = 0 \quad (3.4)$$

where σ_s: critical shear stress, and T: tensile strength. The Griffith criterion holds only for tensile and hybrid stresses ($\sigma_n < 0$; Fig. 3.1). For $\sigma_n > 0$, the Coulomb law of failure suits (Fossen 2010). Griffith (1924) demonstrated that random intra-granular/trans-granular cracks/micro-defects determine the brittle strength of rocks.Thus the experimental tensile strengths of rocks are much less than their theoretical magnitudes (Davies et al. 2012). Note that the depth-wise increase in brittle strength of rocks is linear (Ratheesh-Kumar et al. 2014 and references therein).

In the context of anisotropic rocks i.e. rocks having preferred orientation of weakness planes (e.g. foliated rocks), the resulting strength anisotropy influences the brittle behaviour of the entire rock significantly. So, the Mohr–Coulomb failure criterion was modified by Jaeger and Cook (1967) for anisotropic materials, as follows. The cohesion (c_1) for rock material having anisotropies/discontinuities can also be defined as a continuous variable (Δc), which changes depending on the angle (θ_c) between anisotropy plane and the maximum principal stress axis (σ_1), for different rock samples:

$$c_1 = \Delta c \ \cos 2(\alpha - \theta_c) \quad (3.5)$$

where c: cohesive strength of isotropic rock, and α: angle between plane of shear failure and maximum principal stress axis (σ_1). $0 < \theta_c < 90°$ was assumed. 'c_1' is least when $\alpha = \theta_c$ (Fig. 3.2), and is maximum when the plane of anisotropy/discontinuity is

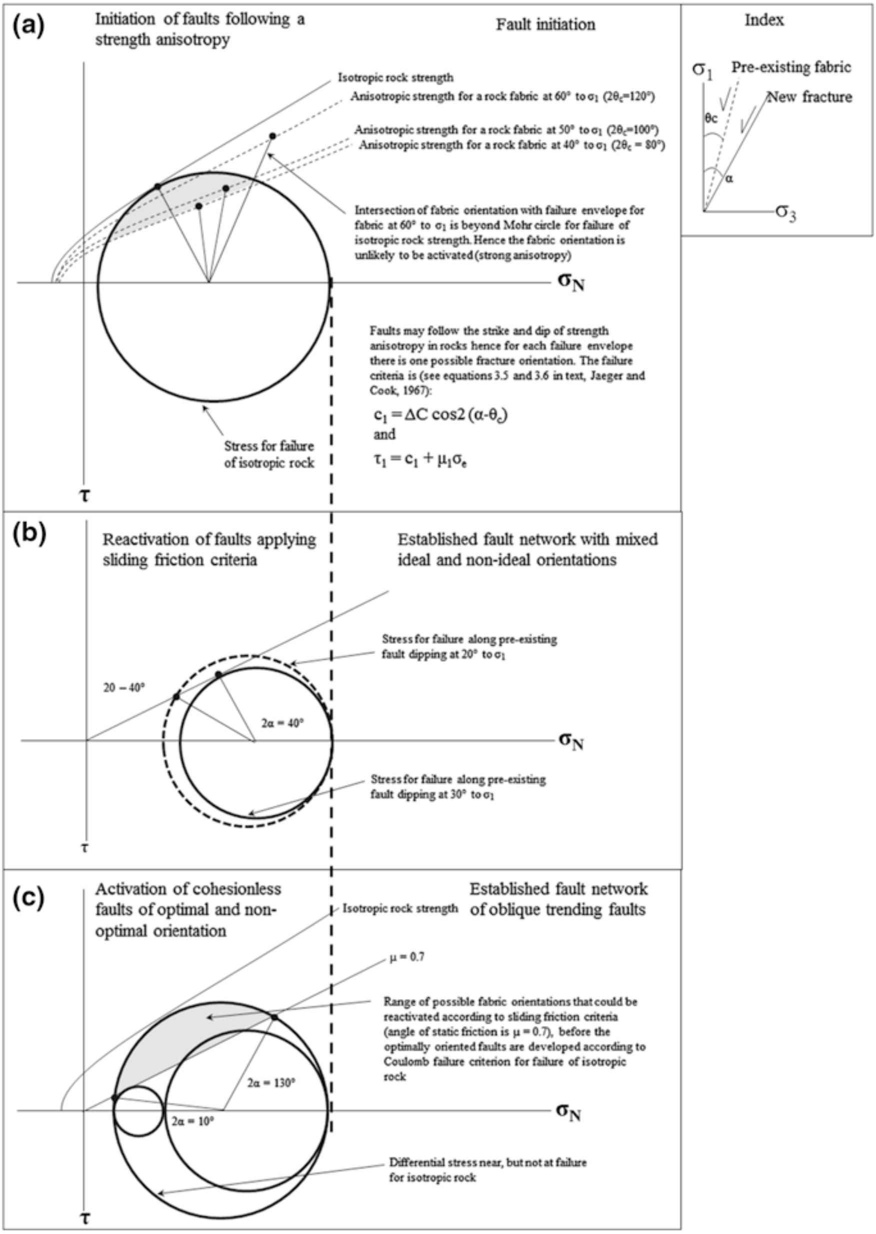


Fig. 3.2 Idealised Mohr circles explaining the relationship between stress state and fracturing/faulting: **a** for initiation of fresh optimally oriented fractures in isotropic rock, **b** activation of oblique fabrics anisotropic rock and **c** activation of incohesive faults of optimal and non-optimal orientations (modified after Morley et al. 2004)

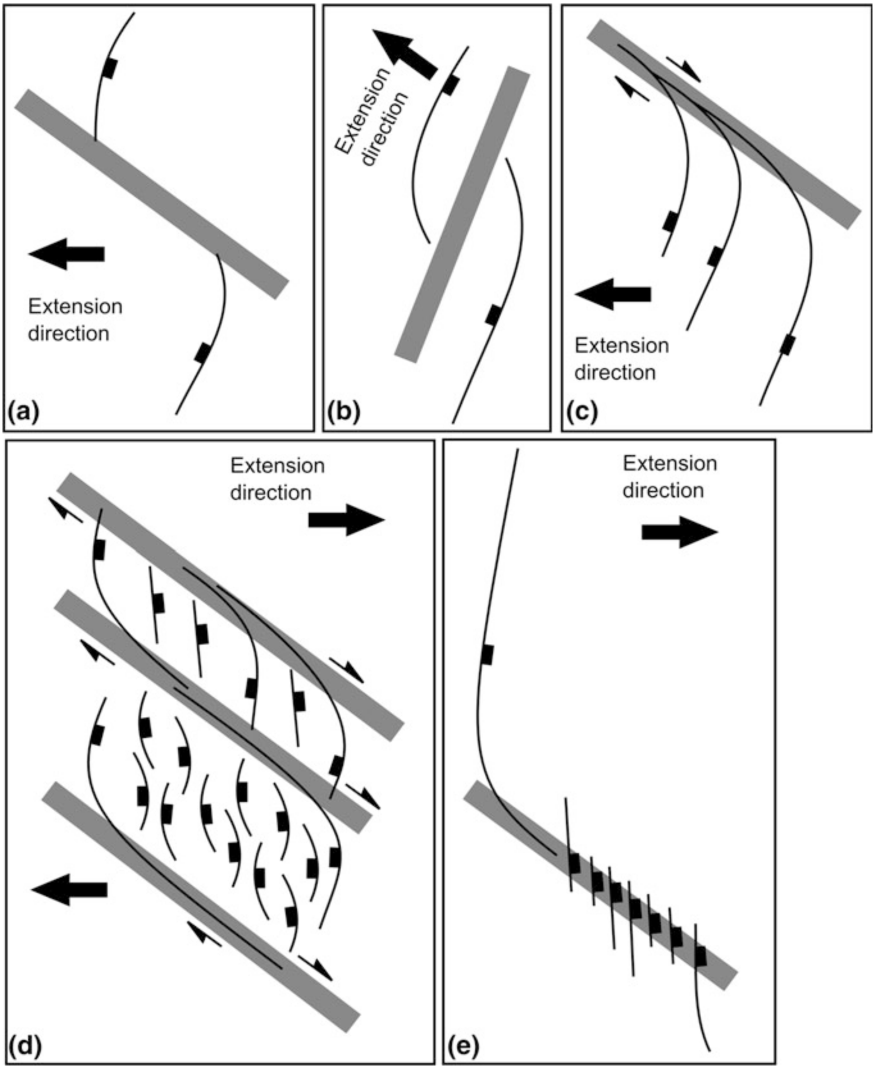


Fig. 3.3 Relationship of oblique "isolated" pre-existing shear zones (*bold grey lines*) with newly formed faults during extension. **a** Newly formed extensional faults abut against the pre-existing shear; **b** the pre-existing shear zone forms a transfer /accomodation zone; **c** the newly formed extension faults tangentially merge into the pre-existing shear zone. This condition forms one end of a horse-tail structure (see Fossen 2010); **d** three pre-existing shear zones with varied spacings. The faults geometries may also depend on the spacing between the shear zones. This situation here is ideal for the formation of localized pull-apart basins; **e** the newly formed fault tangentially joins the pre-existing shear zone and a series of extensional faults are formed at an angle along the shear zone. Note that the extension faults parallel pervasive anisotropies if the angular relationship between the extension direction and the anisotropy trends are favourable (adapted partly from Morley 1999c)

rotated 90° further. Thus, the frictional behaviour of rocks modifies as a function of anisotropy. Its co-efficient of internal friction changes (to say μ_1). Thus, for anisotropic rocks, the modified Coulomb-Mohr failure criterion, is:

$$\tau_1 = c_1 + \mu_1 \sigma_e \tag{3.6}$$

where τ_1: shear strength, and σ_e: effective stress (Jaeger and Cook 1967).

An empirical failure criterion, the Hoek-Brown failure criterion (Hoek and Brown 1980, 1988) has also been used (Ferril and Morris 2003) to analyze brittle failure. This failure criterion equation forms an envelope, which consists of both the tensile and the shear fields, thus avoid using different failure criteria for those fields. Numerous other failure criteria model the strength of both isotropic and anisotropic rocks (see Jaeger and Cook 1976; Lade 1993; Paterson and Wong 2005). Nevertheless, simpler Mohr-Coulomb and Griffith criteria make them most versatile till date. Notice that when a fresh fault is just about to form in an ideal alignment with respect to the principal stress axes, for the case of Andersonian faulting (see Eq. 3.2; Anderson 1951), the prevalent stress state needs to overcome the isotropic cohesive shear strength of the whole rock (Teufel and Clarke 1984).

Strength anisotropy due to discrete or pervasive fabrics in country rock affects the whole rock strength, and attains $c_1 < c$ (Ranalli and Yin 1990). Note, these anisotropies/discontinuities can also prevent fault/fracture propagation if the tensile strength of the intact rock cannot be surmounted. This may occur when the frictional shear/tensile strength of those anisotropies/discontinuities is relatively lower (Teufel and Clarke 1984; Morley 1999). The angular relationship between the regional extension direction and discrete/isolated fabrics also controls fracture/fault propagation (Cliffton et al. 2002; Morley et al. 2004). Those anisotropies/discontinuities can prohibit the fault propagation across their surfaces. As a result the rift bounding faults may not continue through the oblique-trending discrete fabrics. Furthermore, the rift related faults can also orient along a principal oblique trend (Fig. 3.3).

Lab experiments deduced the relation between deformation and pre-existing anisotropies (e.g. Bott 1959; Jaeger 1959; Donath 1961; Youash 1969; Brace and Kohlstedt 1980; Kohlstedt et al. 1995; reviews by Paterson and Wong 2005; Saedi et al. 2014). These studies built the subsequent concepts on the role of pre-existing anisotropies in rifting and on the structures of rifted passive margins. The role of pre-existing fabrics influencing the dynamics of continental rifts has been discussed ever since the development of plate tectonic theory ~1960s (Wilson 1966).

3.2 Rock Deformation Experiments

Planar pre-existing structures/features in basements not only affect the strengths of rocks but also control the attitude of fractures/faults. The possible strength anisotropy is significant: the cohesive strength of intact rock is commonly around 100 bar, whereas that in the fault zone rarely exceeds 10 bar (e.g. Sibson 1977). Sibson

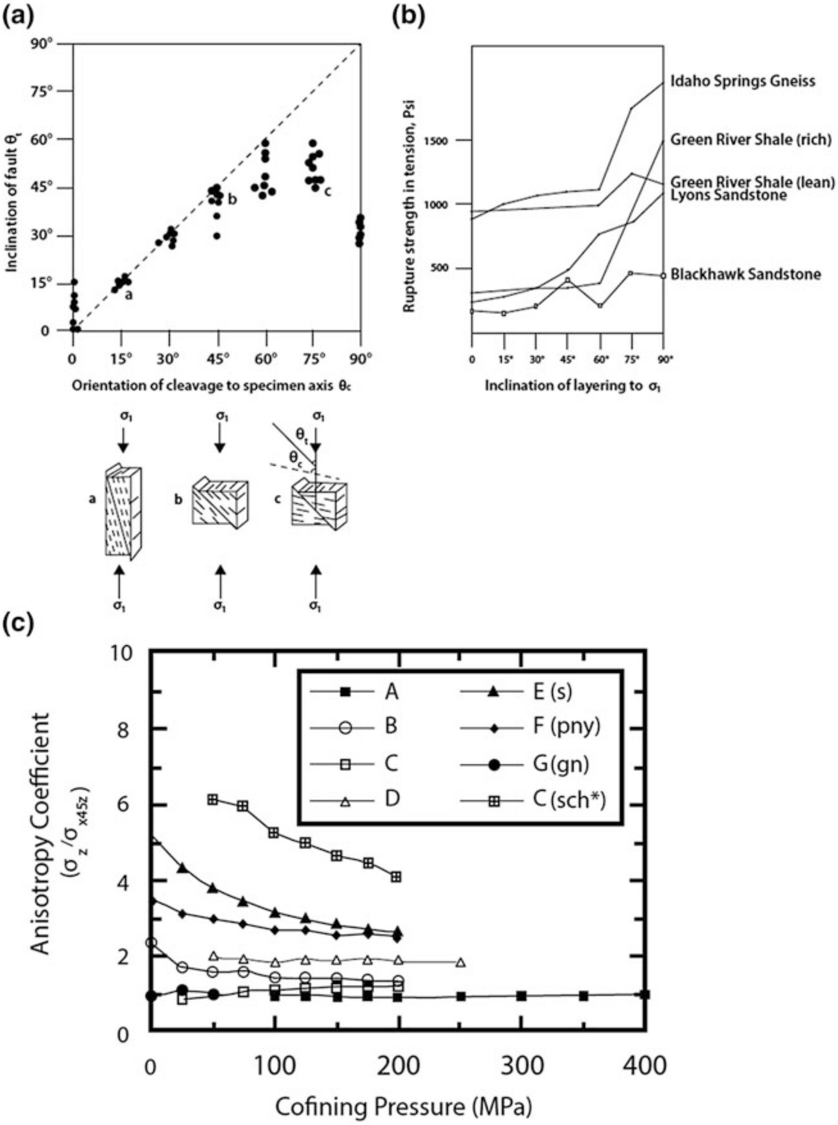


Fig. 3.4 Results of rock deformation experiments: **a** shows the influence of slaty cleavage on the shear fracture orientation, which follows the strength anisotropy, compressive experiments on slate (reproduced from Donath 1961 and Morley 1999c). **b** Variations in tensile strength for different lithologies with relation to layering (reproduced from Youash 1969). **c** Anisotropy coefficients (for degree of anisotropy) of rock types in (a), defined as the ratio of differential stresses ($\sigma_1 - \sigma_3$) measured in the z ($\theta_c = 90°$) and x45z ($\theta_c = 45°$) orientations (σ_z/σ_{x45z}) (reproduced from Shea and Kronenberg 1998)

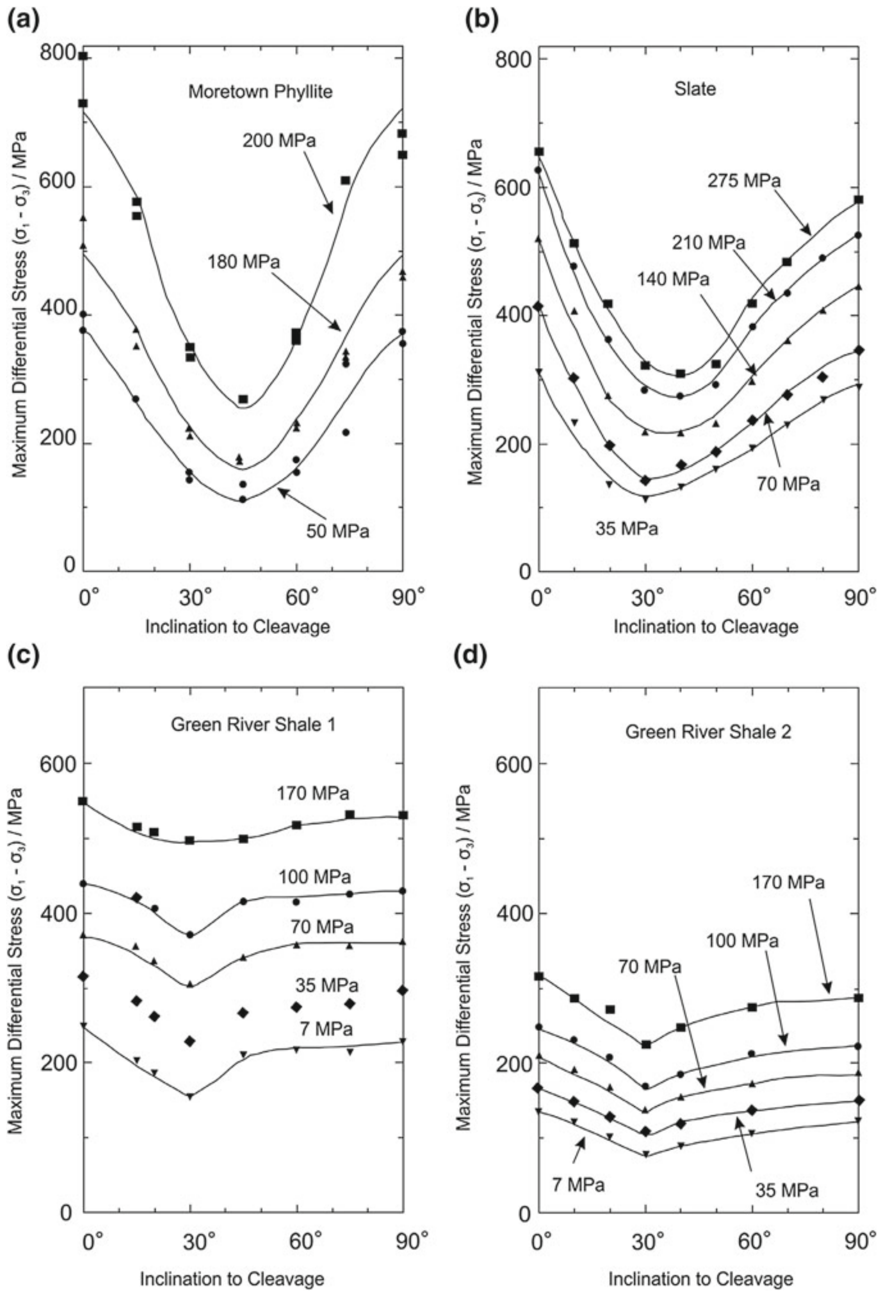


Fig. 3.5 Relationship of differential stress ($\sigma_1 - \sigma_3$) and layering (θ_c) in different lithogies (reproduced from Patterson and Wong 2005), for **a** phyllite **b** slate **c–d** shales

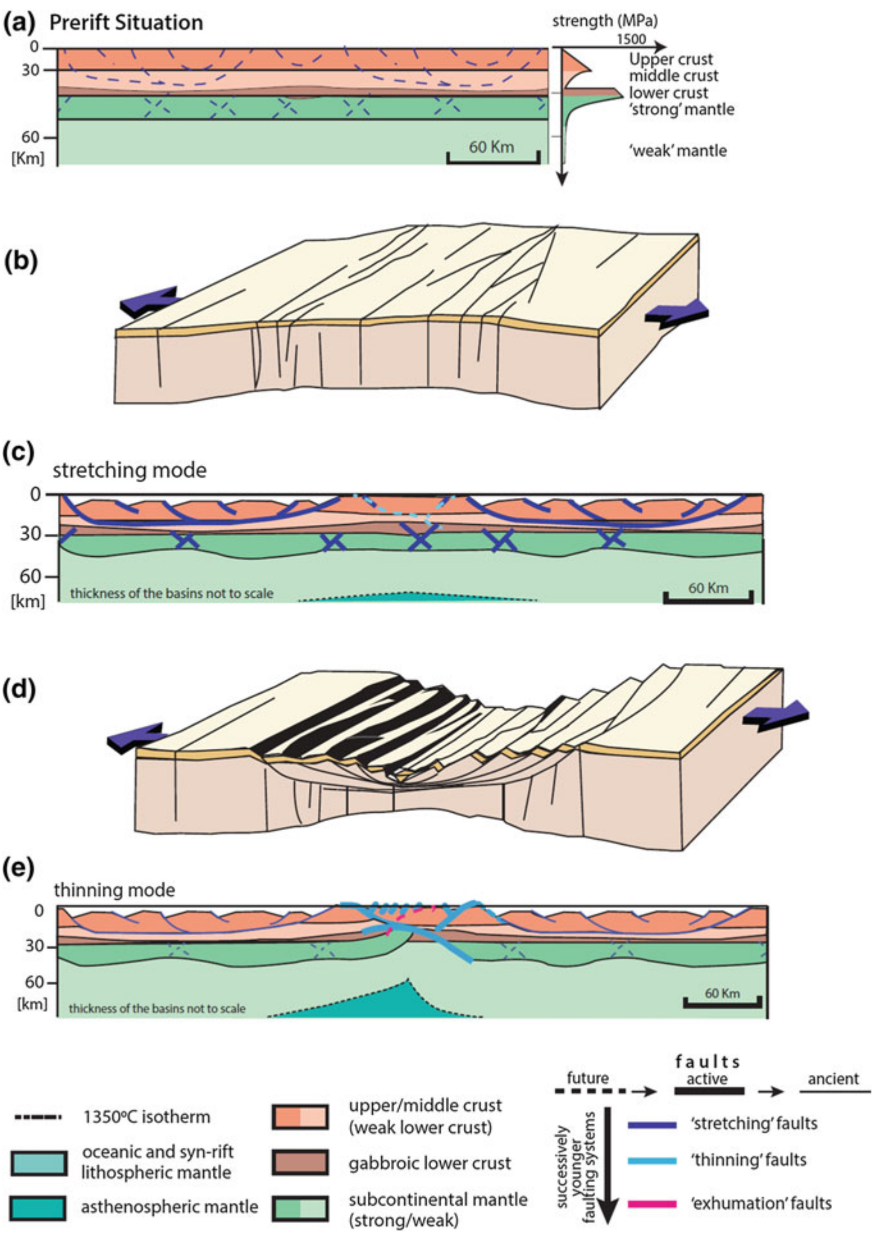
(1977) performed experiments with cylindrical cores of slate. The experiments suggest that when the stress was applied, the fracture orientation strongly correlated to the cleavage orientation (Fig. 3.4a). From lab experiments, Donath (1961) showed that fractures developed along varying orientations depending on the angular relationship between slaty cleavages (θ_c) and the axis of maximum compressive stress (σ_1). Rocks show maximum compressive strength when cleavage planes and maximum compressive stress (σ_1) orient orthogonally ($\theta_c = 90°$). A minimum strength is observed for $\theta_c = 30°$. Fractures develop along cleavages when $15° < \theta_c < 45°$. Fractures disregard the cleavages if $\theta_c > 45°$ (Fig. 3.4a). The 'strike', i.e. the intersection line between the fracture plane and the (horizontal) base of the cylindrical core, parallels the cleavages.

Youash (1969) demonstrated that rocks show only 25–75 % of tensile strength when loaded at 0°–60° to pre-existing anisotropies than those loaded at 90° (Fig. 3.4b). Compressive tests on anisotropic dolomitic limestone (McGill and Raney 1970) indicate that faults parallel the pre-existing anisotropy when the angle between the maximum stress direction and the laminae ranges 13–51°.

Shea and Kronenberg (1993) performed compressional deformation tests on schists and gneisses for varying (15–75 %) mica content at $\theta_c = 45°$ and 90° to test the influence of mica in rock anisotropy. They defined degree of anisotropy (Fig. 3.4c) in terms of '*anisotropy coefficient*', which is the ratio of the compressive strength of the rock for $\theta_c = 90°$ and 45°. The same lithology shows varying degrees of anisotropy at different confining pressures. Additionally, they concluded that anisotropic strength decreases with increase in mica content in foliations. It was shown that for θ_c within 30°–45°, the strength attains minimum magnitude and is independent to lithology (Fig. 3.5). A number of studies (Jaeger 1959; Handin and Stearns 1964; Lane and Heck 1964; references in Sirieys 1966; Okusa 1971) introduced anisotropy by a process resembling saw-cut in an isotropic rock. The rocks showed shearing along the induced anisotropy over a large range of θ_c values. Only for θ_c within 0°–90°, the rock shears across the anisotropy. In case of insignificant shear stress along the anisotropy plane due to the specific orientation of the plane, the rock does not fracture.

3.3 Bearing on Rift Systems

At the shallow crustal levels, pre-existing fabrics can control the geometry and location of rifts by diverting or preventing fracture propagation (e.g. Lezzar et al. 2002). Continental rifts evolve through multiple phases (Fig. 3.6): from a distributed deformation to a localised one (Manatschal et al. 2007). In the earliest pre-riftphase, fracture systems develop, sometimes as deep as to detach on the ductile lower crust (Fig. 3.6a, b). During this stage, the pre-existing anisotropies of the basement and/or the pre-rift sedimentary rocks control the fractures/faults trends depending upon the strength, geometry, spacing and orientation of those heterogeneities. Rift zone geometries are also controlled by litho-layers and extension rates besides


(a) Prerift Situation
strength (MPa)
1500
Upper crust
middle crust
lower crust
'strong' mantle
'weak' mantle
0
30
60
[Km]
60 Km
(b)
(c)
stretching mode
0
30
60
[km]
thickness of the basins not to scale
60 Km
(d)
(e)
thinning mode
0
30
60
[km]
thickness of the basins not to scale
60 Km
1350ºC isotherm
oceanic and syn-rift lithospheric mantle
asthenospheric mantle
upper/middle crust (weak lower crust)
gabbroic lower crust
subcontinental mantle (strong/weak)
faults
future
active
ancient
successively younger faulting systems
'stretching' faults
'thinning' faults
'exhumation' faults

◀ **Fig. 3.6** Schematic diagram to illustrate evolution of a continental rift system for a five layer lithosphere (refer to the strength profile). **a** Pre-rift situation where lithosphere is just subjected to extension and will form faults in the future stretching phase, along the *dotted lines*. **b** In the shallow crust (>15 km) the extension is accommodated by numerous tectonic joints, normal to it; pre-rift sediments in *pale yellow*. **c** Stretching mode, when extension is maximum, faults (*deep blue lines*) form in the brittle (or strong) upper crust, detaching on the weak middle crust. The strong lower crust-upper mantle also gets faulted in response to the extension. Rift-basins bound by stretching faults characterise the deformation in the uppermost crust. **d** The tectonic joints formed in the previous phase are inherited in this phase in the upper crust to form the stretching faults. The steep joints may be disregarded in preference to low-dipping normal faults, although the strikes maintain. **e** Thinning mode, where the lithosphere actively thins across a necking zone, which comprises of thinning faults (*light blue lines*). Note, extension is lesser compared to stretching mode (reproduced from Manatschal et al. 2007 and Fossen 2010)

geomorphological processes (Ramirez-Arias et al. 2012; Buiter 2014; Huismans et al. 2014). Spacing of faults inherited from basement faults tend to be spaced maintaining some relation (Montési and Zuber 2003b). Also, not all regularly spaced faults are inherited (Montési and Zuber 2003b). Relationship between the anisotropies in the pre-rift basement and the trends of the faults have been noted in many studies (Morley 1999b, c; Corti 2004; Morley et al. 2004; Aanyu and Koehn 2011; Chattopadhyay and Chakra 2013). Peacock and Sanderson (1992) reported that faults run along the anisotropies. When those anisotropies are aligned, possibly 45°–90° to the regional extension direction (σ_3), normal faults develop (Morley 1999c). Some non-optimally/obliquely aligned anisotropies are disregarded and some, like those having the least strength, later develop into sheared segments or oblique rifts (e.g. Morley et al. 2004). In the next important stage of stretching, crust and the lithospheric mantle extend laterally (Manatachal 2004; Lavier and Manatschal 2006; Manatschal et al. 2007; Whitmarsh and Manatschal 2012) (Fig. 3.6c, d). The fractures/faults, which started earlier, active at forming large fault blocks.The steep fractures/faults may be disregarded to follow the “modified Mohr-Coulomb failure criterion” (Eq. 3.7) but the strikes/trends are usually maintained (Fig. 3.2b; Morley 1999c). During this stage, individual faults segment strongly and soft link into relay structures/accommodation or transfer zones. Further extension links these faults by breaking down relay structures and hardlink transfer faults (Versfelt and Rosendahl 1989; McClay et al. 2002). These result in rhombic basin geometries. If the stress regime changes during rifting, non-optimally oriented fractures/faults of the early extensional stage may reactivate, based on change in coefficient of internal friction (μ) and cohesion (c_1) (Morley et al. 2004).

In an array of faults with different trends but similar μ and c_1, i.e. those closest to the mechanically ideal Andersonian orientations for failure in response to stress, reactivate favourably (Fig. 3.2). The lithosphere thins subsequently by deep mantle reaching faults (Fig. 3.6e) and the basin subsides rapidly (also see Fowler 2005). The thinning faults (e.g. Peron-Pinvidic et al. 2013) also follow the same regional trends as set by the previous rifting phases. The deformation progresses from distributed in the pre-rift to early rift into a localised one in the thinning stage, and the weakest faults evolve as thinning faults.

On the other hand, in case of oblique fabrics, the rift faults in all the phases show oblique slip of varying magnitudes (Versfelt and Rosendahl 1989; Morley et al. 2004). The basin geometries evolve by interfering pre-existing oblique fabrics with orthogonal extension commonly resemble those evolved through purely strike-slip extension (Clifton et al. 2000).

Chapter 4
Pre-existing Fabrics

4.1 General Discussion

Continental rifts are not random and generally tend to follow the mobile belts (Fig. 4.1), diverging around the cratons (Tommasi and Vauchez 2001). Mobile belts are fossilized orogens, which possess numerous weak brittle and ductile planes viz. faults, shear zones, foliations etc. Holdsworth et al. (2013) categorized inheritance to be 'lithosphere-scale control' when thickness of crust, thermal age etc. are important; and 'grain-scale control' when depth, stress, composition of rocks, size of grains, intensity of fabrics developed, and presence and role of fluids play importance. However, a better accepted classification of factors that give inheritance is: (i) discrete (large, widely spaced anisotropies), and (ii) pervasive (small, closely spaced and present in the entire rock volume) as proposed by Morley (1999c). Whether or not orogeny is active, olivine in the lithospheric mantle attains a lattice preferred orientation, which effectively localises strain during extension (Tommasi and Vauchez 2001 and references therein). Repeated reactivation of mobile belts makes them weaker than the cratons (Cloetingh et al. 1995). Thus rifting facilitates more in mobile belts than in cratons (Cloetingh et al. 1995; Rey 2001; Corti et al. 2007). Also, mobile belts form positive topographies and after compressive stress stops, they collapse gravitationally and initiate rifts (Rey 2001) e.g. in Basin and Range Province, North America.

4.2 Pervasive Fabrics

Pervasive fabrics- slaty cleavages, close-spaced joints or beddings, laminations, flow layers of pre-rift sedimentary or volcanic rocks, and foliations such as schistosity and gneissosity- may persist throughout the rock volume and contrast strength within the rock body. Due to primary- (and secondary) ductile shear planes (Passchier and Trouw 2005), mylonites are prone to brittle reactivation (Holdsworth et al. 2001;

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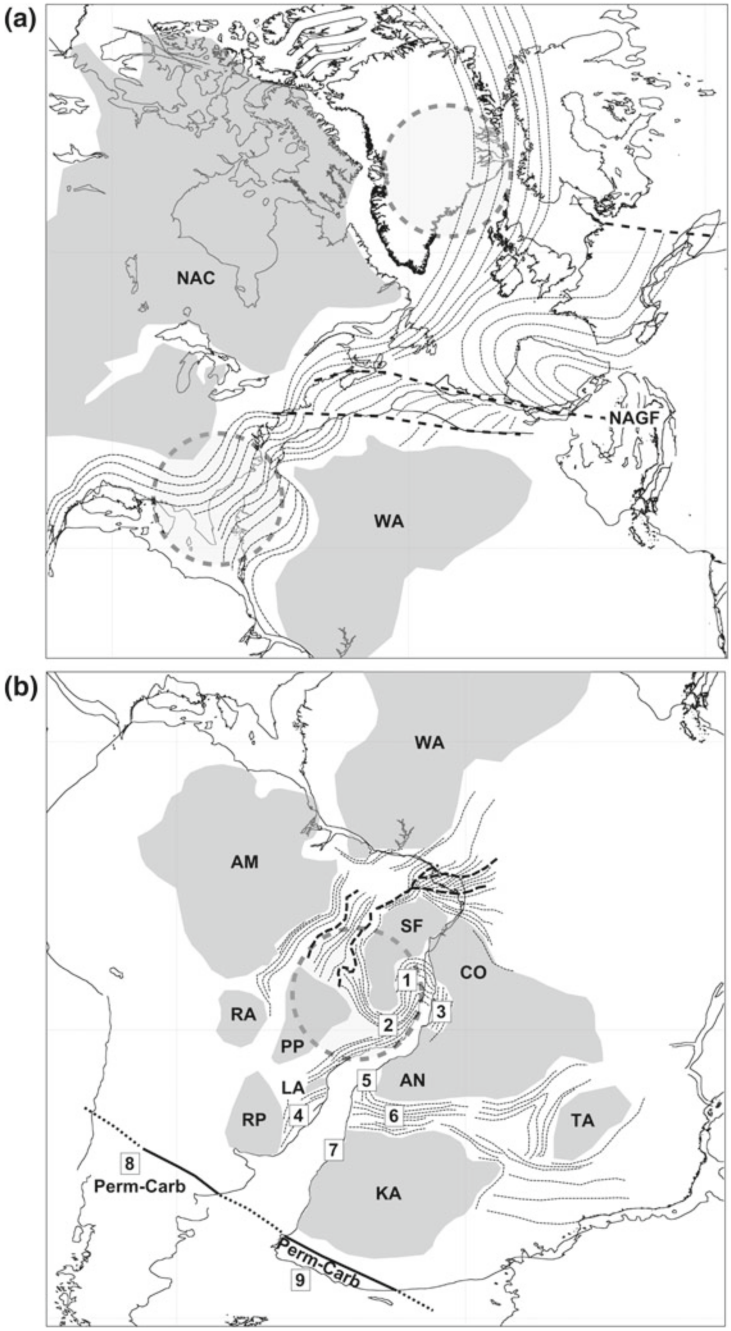
DOI 10.1007/978-3-319-20576-2_4


(a)
NAC
NAGF
WA
(b)
WA
AM
SF
CO
1
2
3
RA
PP
LA
RP
4
5
AN
6
TA
7
KA
8
Perm-Carb
Perm-Carb
9

◀ **Fig. 4.1** Schematic plate reconstruction diagram at 200 Ma before present showing **a** the Caledonian-Hercynian mobile belts in north Atlantic conjugate margins and **b** the Pan-African mobile belts in South Atlantic conjugate margins. Note the parallelism maintained between the passive margins and the mobile belts. Grey circles with bold broken outlines demarcate the approximate position of mantle plumes; *NAGF* Newfoundland-Azores-Gilbraltar fault zone (modified after Tommasi and Vauchez 2001). Cratons are shown in grey. *NAC* North American Craton (those stable for the last ~1 Ga shown here), *AM* Amazonia, *WA* West Africa, *SF* São Francisco, *CO* Congo, *PP* Paranapanema, *AN* Angola, *RA* Rio Alba, *RP* Rio de la Plata, *KA* Kalahari, *TA* Tanzania. The orogens are shown with dashed lines indicating structural trends. *1* Araçuaí, *2* Ribeira, *3* West Congo, *4* Dom Feliciano, *5* Kaoko, *6* Damara, *7* Gariep, *8* Sierra de la Ventana, *9* Cape Fold Belt, Perm-Carb: Late Carboniferous-Permian mobile belts (modified after Almeida et al. 2013 and references therein)

Mukherjee and Koyi 2010a, b; Mukherjee 2013a, b, c; Mukherjee 2015a; Mukherjee and Biswas 2015; Mukherjee et al. 2015; Mulchrone and Mukherjee 2015a, b). These fabrics usually follow the regional trends of foliations within the metamorphic basement. Pervasive fabrics affect large rock volumes even at the micro-scale (Morley et al. 2004) or possibly still beyond. Rocks having pervasive fabrics show minimum strengths throughout the entire volume (Sect. 3.2). As a result, trends of newly formed faults get optimally aligned (Sect. 3.3). Experiments established relations amongst fault trends, anisotropies and strength of anisotropic rocks (Sect. 3.2; see Figs. 3.3, 3.4, 3.6). Pervasive fabrics control the general trend of the rift faults. Therefore, most rift systems and passive margins parallel pervasive fabrics e.g. in the East African Rift System, the N and S Atlantic (Morley 1999b; Fig. 4.1). Detail research work at Atlantic margin is on progress on the role of inheritance in hyper extended rifts (Chenin et al. 2014; Mohn et al. 2014a). Pre-existing pervasive fabrics in pre-rift rocks also affect overlying lithology during extensional deformation by forming new fractures that parallel the pre-existing pervasive fabrics (Cortés et al. 2003). In such cases, transfer faults may locally parallel the anisotropy (Chattopadhyay and Chakra 2013). Pre-existing pervasive fabrics and offset angles between rift zones strongly influence structure of transfer zones between propagating rifts (Corti 2004).

Field observations by Peacock and Sanderson (1992) studying effects of anisotropy in sedimentary layers gave results similar to those from the lab (e.g. Donath 1961; Youash 1969; Shea and Kronenberg 1993) and theoretical analyses (e.g. Ranalli and Yin 1990). Analogue models showed that most of the rift faults, not just the bounding faults, initially follow the pervasive fabrics and may breach the trend later to form faults at high-angles to the fabric (Aanyu and Koehn 2011). The relationship of pervasive fabrics, extensional faults and stress state is shown in Fig. 4.2. Mathematical models by Dyksterhuis et al. (2007) revealed the followings: (i) initial weakness within the lithosphere governs rifting significantly; (ii) a single focused weak zone creates symmetric and narrow rifts; and (iii) diffuse weak zones create wider rifts; (iv) prior weak faults create asymmetric rifts. Manifestation of shearing on passive margins that inherited structures from bottom is under study (Salvi et al. 2013).

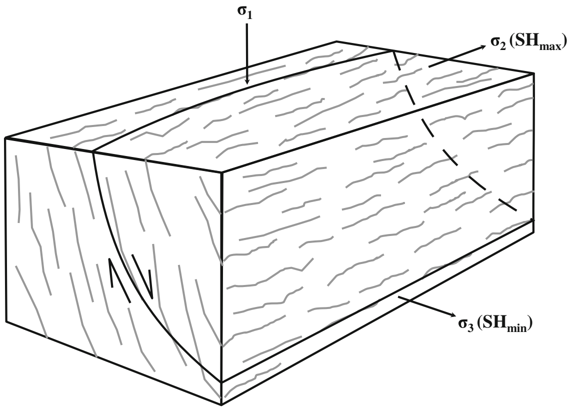


Fig. 4.2 Schematic block diagram depict relation of pervasive fabrics, stress state and extensional faults. σ_1: maximum compressive stress axis; σ_3: minimum compressive stress axis; SH_{min}: minimum horizontal compression; SH_{max}: maximum horizontal compression. See Sect. 3.2 for details, modified from Morley 2010

Model lithosphere inside stiffer material reactivates quicker than that within the plastic material (Chenin and Beaumont 2013). When the crust is attached with a flexible mantle, offset basins develop solely within weak zones, and such basins develop only when the flexible crust is not attached/coupled with the mantle (Chenin and Beaumont 2013).

The above relations and their tectonic manifestations are observed globally.

4.2.1 East African Rift System

The East African Rift System (EARS) (Fig. 4.3), considered also by some as magma-rich/magmatic rift (Moores and Twiss 1995; Allen and Allen 2013), is one of the most well-studied region for the influence of pre-existing anisotropies on the rift structure (reviews in Morley 1999c; Ebinger 2012), especially at its western branch (Klerkx et al. 1998). A $\sim$ E − W extension has been deciphered from the northern and central parts of this rift system (Saria et al. 2014). The Tanganyika craton, bifurcating the rift into the Eastern- and Western- branches, comprises of ortho- and para-gneisses with some basic to ultrabasic rocks (Anhaeusser et al. 1969) of $\sim$3000–3600 Ma age (Spooner et al. 1970; McConnell 1972; Kroner 1977). The mobile belts: Ruwenzori, Kibaran, Ubendian, Usagaran and Mozambique (Fig. 4.3) show three major deformation phases (Tack et al. 2010). Those are: (i) The Eburnean deformation (2100–1800 Ma) at Ubendian mobile belt, which formed the 140° trending ductile pervasive fabric south to the Tanganyika craton, (ii) the Kibaran orogeny at $\sim$1400–900 Ma (Tack et al. 2010) that shaped N/NNE trending ductile fabric north to the Lake Tanganyika; and (iii) the Pan-African orogeny at $\sim$600 Ma that deformed the mobile belts in the region.

Interestingly, the prior Pan-African orogeny controlled the geometries of the later grabens in the EARS (Michon and Sokoutis 2005). The rift trends follow those of the mobile belts. The Western branch is convex towards W, and foliations trend dominantly NE-SW in N and NW-SE in S (Fig. 4.3). The western branch swerves N-S south of the Ubendian belt. The eastern branch is convex towards E and turns NE-SW north of the Lake Turkana. The Eastern Branch is older, starting rifting possibly in Eocene and abandoned around Mio-Pliocene. The presently activeWestern branch started rifting in late Miocene. Both these branches consist of large asymmetric half grabens with <6–7 km of fluvio-lacustrine/lacustrine/ fluvio-deltaic sediments with volcanics/volcaniclastics (Morley 1999a). The main extensional faults (e.g. Lokichar-, Lothagam-, Lagh Bogal-, Lupa faults) parallel the foliation trends of the Precambrian basement (Morley 1999a, c). The Lake Kivu-Lake Edward region is however unique because NE, NW, E-W trending pervasive foliation meet here. Nevertheless, the Kivu and Edward rift faults follow the overall convex W-wards trend of the Ubendian-Kibaran mobile belt (Fig. 4.3). The eastern- and western branches (Fig. 4.3) warp in response to the Precambrian trend and paralleling the Proterozoic mobile belts, which swerve around the Tanzanian craton (Fig. 4.3; Rosendahl 1987; Versfelt and Rosendahl 1989; Corti et al. 2007). This swerving geometry governed the kinematics and the geometry of rifts within the East Africa Rift System (Corti et al. 2006).

The Tanganyika craton apparently acted as a resistant core forcing the Eocene to Miocene EARS initiation to follow surrounding Proterozoic mobile belts. The Rukwa rift (Fig. 4.3) formed an oblique rifted segment in response the NW-SE trend of the Precambrian pervasive fabrics and the E-W to NW-SE regional stress directions (Morley et al. 1992; Ring 1994; Morley 2010). Oblique/broken transform rifted margins are in general linked to inheritance of oblique fabrics (Manatschal et al. 2013). The other segments of the EARS are predominantly orthogonal rifts. In the youngest (possibly Pleistocene or younger), incipient branch (Kinabo et al. 2008; Bufford et al. 2012) of the EARS, the Okavango Rift zone in NW Botswana, basement fabrics and pre-existing faults control the NE-SW trend of the rift related normal faults (Modisi et al. 2000; Alvarez Naranjo and Hogan 2013). The limbs of meta-sedimentary tight folds and foliation in meta-rhyolites define its basement fabrics. This example shows that the pervasive fabrics exert a strong influence on defining the overall geometry of the rift system in the sense that nucleation of the initial rift faults parallel the strike of the pervasive anisotropies. In the EARS, complex rift geometries are seen because rift segments propagate and the pervasive fabrics breach (Aanyu and Koehn 2011). Analogue modelling suggested that such interactions were either shear (mostly simple/non-coaxial shear) or extensional, depending on the angular relation between the transfer faults and the extension direction (σ_3).

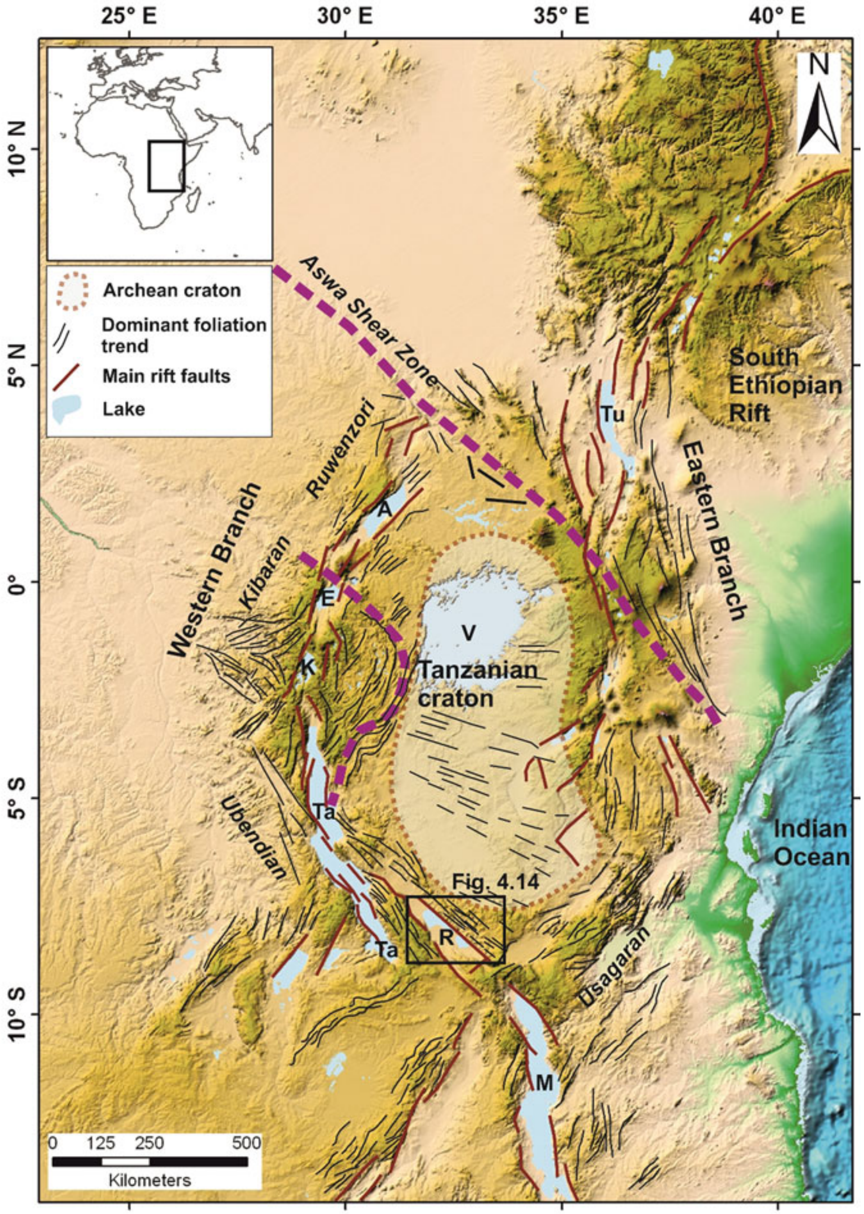


Fig. 4.3 Main structural features of the eastern and western branches of the East African Rift (after Morley 1999a; Corti et al. 2007 and references therein). Rift zones parallel mobile belts and swerve around the strong Tanzanian craton. Lakes: *A* Albert; *E* Edward; *K* Kivu; *M* Malawi; *Ta* Tanganyika; *Tu* Turkana; *V* Victoria. *IO* Indian Ocean

4.2.2 *Thailand Tertiary Rift System*

The Thailand Tertiary rift zone covers the Northern Intramontane-, Central plains- and Gulf of Thailand provinces in Myanmar (Burma), Thailand, the Lao People's Democratic Republic, Cambodia and Vietnam in SE Asia (see Barber et al. 2004). This rifted zone is a part of the Thai-Malay mobile belt, which extends from N Thailand to Sumatra (Figs. 4.4 and 4.5; Polachan et al. 1991). This mobile belt formed when the Shan Thai- and Indochina craton collided in Late Triassic (Figs. 4.4 and 4.5; Atlas of Mineral Resources of the ESCAP Region: Mineral Resources of Thailand 2002; Ridd et al. 2011). Both these cratons resemble the broad stratigraphy in terms of Precambrian metamorphic basement overlain by Palaeozoic marine meta-sediments: quartzite, phyllite and schist, and meta-volcanics: meta-tuff and some ultrabasic rocks (review in Ridd et al. 2011). The Thailand Tertiary (Middle –Late Miocene) rift province consists of pull-apart basins formed on N-S trending oblique-slip faults and NW and NE trending conjugate strike-slip zones (Figs. 4.4 and 4.5; Polachan et al. 1991). These onshore and offshore basins formed in the Late Oligocene Period in response to India-Asia collision ~55 Ma back (Yin 2006) and the resultant strike-slip displacement due to escape tectonics between the Shan-Thai and Indochina continental blocks (Figs. 4.4 and 4.5) (Polachan et al. 1991; Leloup et al. 2001). However Morley et al. (2000, 2001, 2004) and Rhodes et al. (2002) favoured an E-W extension in the region due to strike-slip faults, which followed pre-existing weaknesses in the pre-rift lithology combined with some strike slip movement during basin inversion in Late Tertiary. NE-SW striking bedding planes with dip ~40–70° in intensely folded pre-rift Triassic sediments strongly influenced the trends of the rift faults. The faults tend to parallel beddings (Morley et al. 2004). The bedding controlled NE-SW faults formed relatively earlier than Tertiary faults and prevented rift orthogonal N-S faults to propagate (Morley et al. 2004).

4.2.3 *South Atlantic Passive Margins*

The S Atlantic passive margins formed in Jurassic (~180 Ma) along mobile belts and avoided cratons (Figs. 4.1b and 4.6). This separation formed the conjugate SE Brazil and West Africa passive margins and opened the Atlantic Ocean. The São Francisco- and Congo-Angolan cratons were part of the same lithospheric unit prior to Jurassic. The dominant lithologies of these Archaean (3600–2000 Ma) cratons are tonalite-tronjhemite-granodiorite associations with calc-alkaline plutons, greenstone belts and Paleo-proterozoic metasedimentary successions (Shelley et al. 2005). The Ribeira-, Araçuaí- and Dom Felicano belts are a part of the Mantiqueira Orogenic System (Figs. 4.1b and 4.6) with the West Congolian-, Kaoko- and Gariep belts as the African counterparts (Shelley et al. 2005). The Ribeira- and Araçuaí mobile belts separate the São Francisco- and Congo cratons; and the

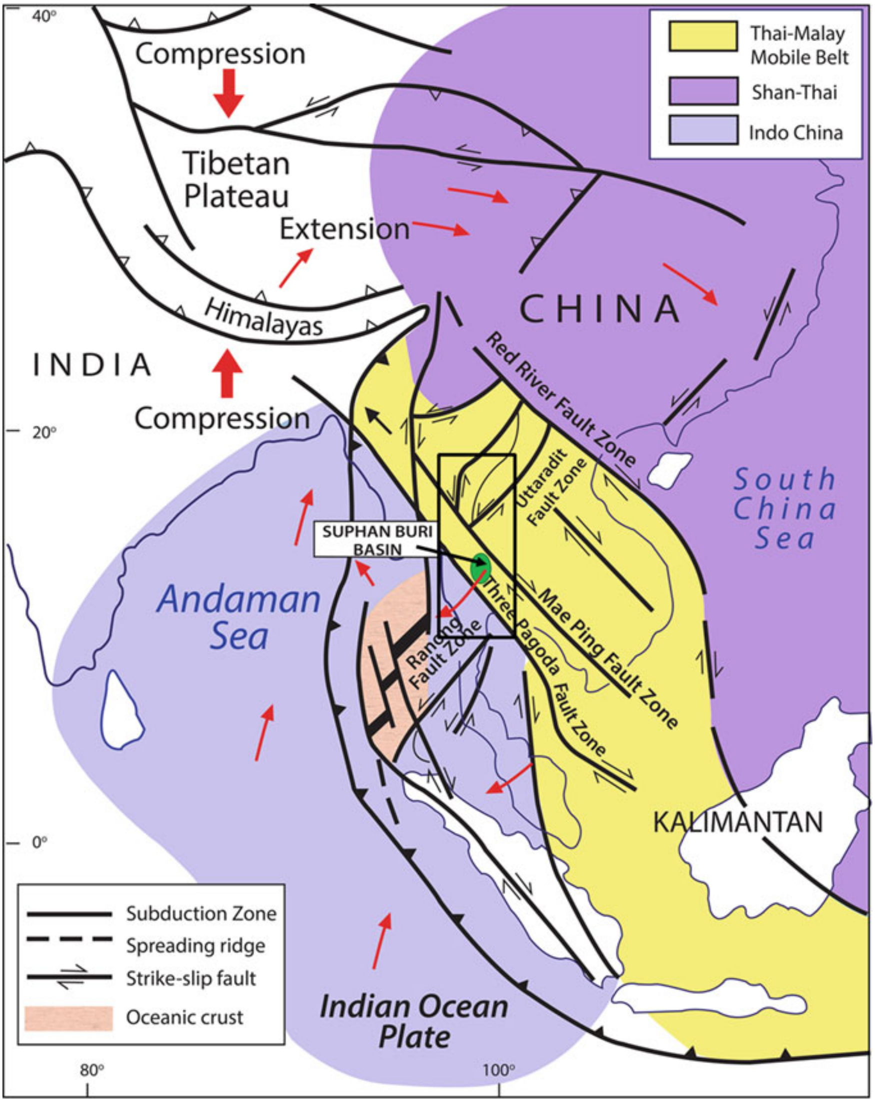


Fig. 4.4 Regional tectonic elements of SE Asia (after Polachan et al. 1991). The Indochina block was extruded towards the SE due to escape tectonics because of the India-Eurasia collision. The Thai-Malay and Shan-Thai blocks accommodated this deformation by strike-slip movements of blocks. This is a result of the N-S compression and E-W extension. Black rectangle marks the area shown in Fig. 4.5

Ribeira- and Dom Feliciano belts are sandwiched between the Angolan- and Rio de Plata cratons with the Luis Alves craton trapped in between (Brito Neves and Cordani 1991; Trompette 1997; Cordani et al. 2003; Fuck 2007; Fuck et al. 2008) (Figs. 4.1b and 4.6). The Luis Alves craton (Fig. 4.1b) was a rather small

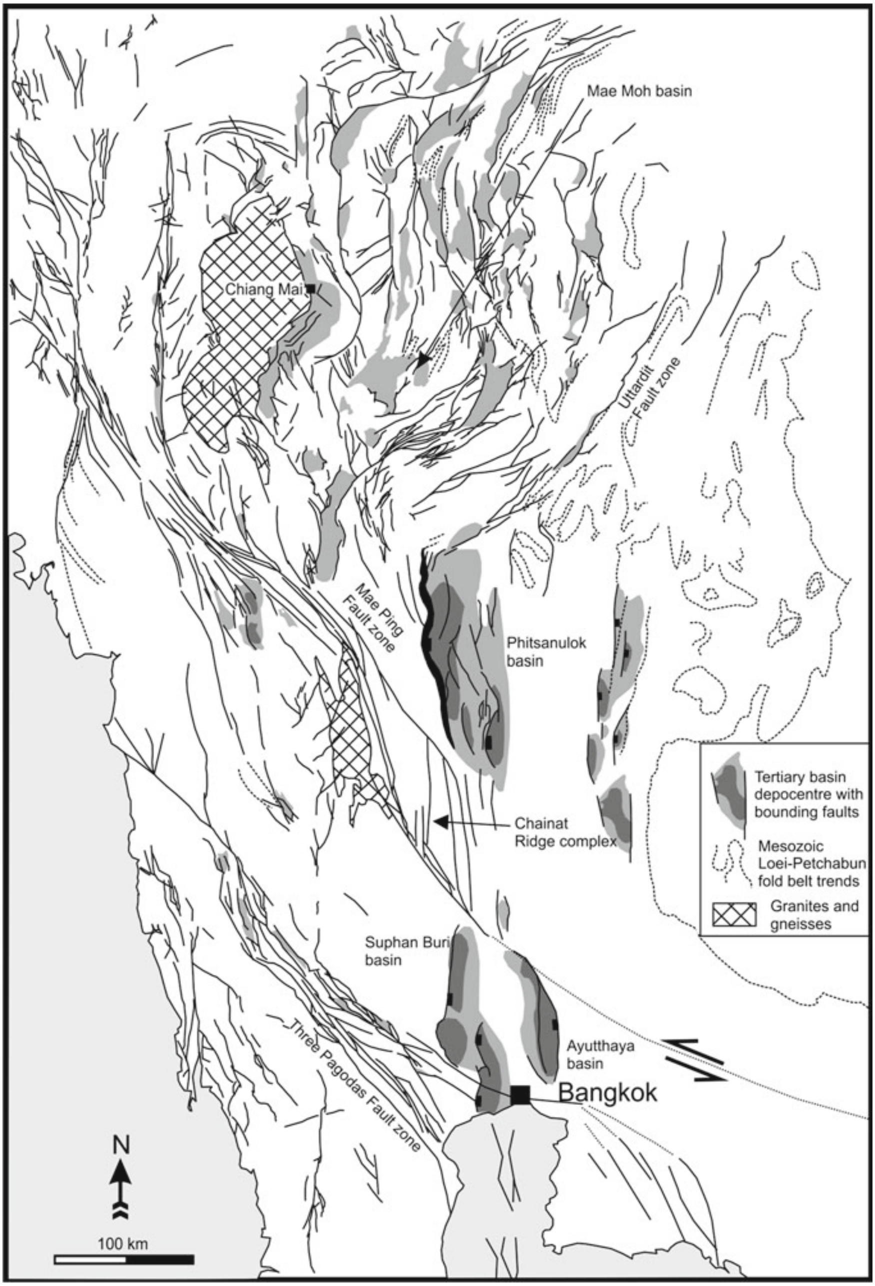


Fig. 4.5 Locations of the rift basins of Thailand showing the major faults and rift basins. Note the parallelism of Tertiary rift faults with the Mesozoic Loei-Petchabun fold trends, especially N of the Mae Moh basin and around the Phitsanulok basin. Modified after Morley et al. (2004)

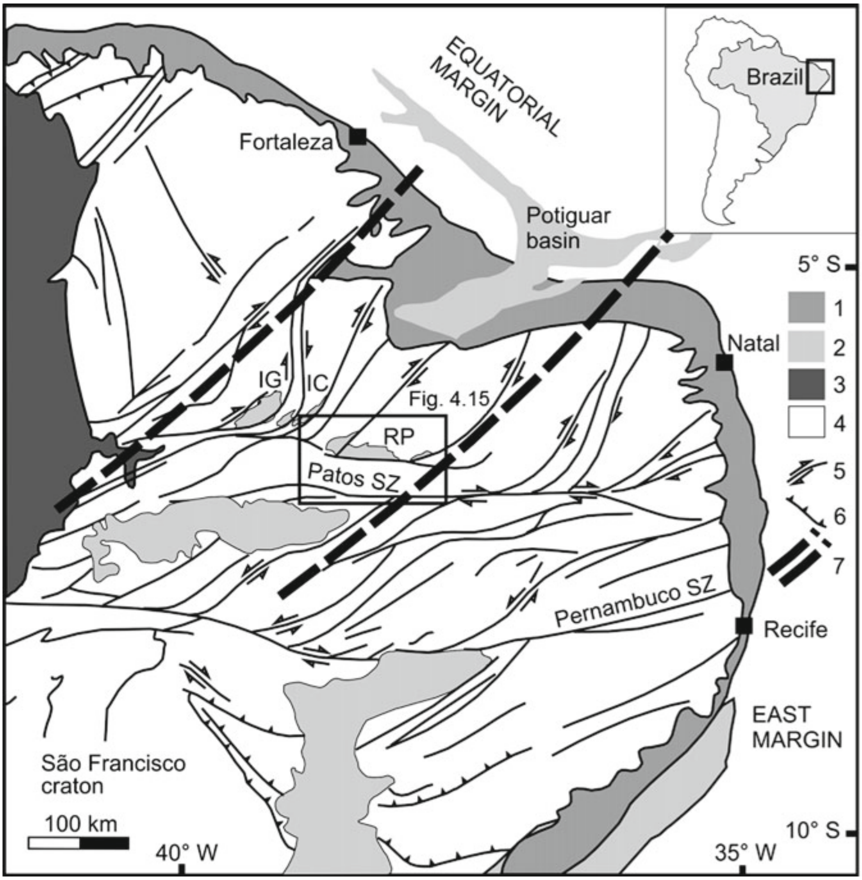


Fig. 4.6 Simplified tectonic map of NE Brazil showing the major shear zones and rift basins Rectangle shows the location of Fig. 4.15. *1* Post-rift sedimentary cover, *2* Cretaceous basins, *3* Palaeozoic basins, *4* Precambrian crystalline basement, *5* strike-slip ductile shear zones, *6* thrust faults, *7* Cariri–Potiguar trend; *IC* Icó, *IG* Iguatu *RP* Rio de Peixe (reproduced from Kirkpatrick et al. 2013)

continental block, which was marginally affected during the Pan-African- or the Brasiliano orogeny (Fuck 2007; Fuck et al. 2008). The ∼NE-SW Ribeira- and Dom Feliciano belts are almost coast parallel and continue underneath the thinned continental crust beneath the offshore Santos Basin (Schmitt et al. 2008). These belts are characterised by Archaean to Paleo-proterozoic basement with Meso- to Neo-proterozoic and Eo-Cambrian meta-sediments with pre- and post-collisional granites. This collision, a part of the Brasiliano/Pan-African orogeny, was dextral transpressional and formed NW verging fold-thrust belts and nappes and NW trending shear zones (Ebert and Hasui 1998; Trouw et al. 2000). The Ribeira belt continues northwards as the ∼N-S Araçuaí belt, which also parallels the coastline of the offshore Campos basin. The Araçuaí belt consists of Palaeoproterozoic basement, amphibolite or granulite facies Neoproterozoic passive margin to foreland strata with Brasiliano/PanAfrican granites (625–530 Ma), deformed by an array of ~NNW striking dextral transpressional zones (Cruz and Alkmim 2006 and references therein).

In the SE Brazil and West Africa conjugate passive margins (Fig. 4.1b), the trends of the rifts follow strongly the Precambrian basement fabric, while the dips of the former vary considerably (Ashby et al. 2010; see also Fig. 8 of Almeida et al. 2013). Rifts swerve around cratons and follow the trends of mobile belts and resemble the EARS (see Ashby 2012; Almeida et al. 2013). Notably the passive margin strongly parallels the pervasive fabric of the Mantiqueira Orogenic System and bend around the São Francisco- and Congo-Angolan cratons (Fig. 4.1b). In the Portiguar basin (Fig. 4.6), NE Brazil, the influence of the pervasive fabrics is presumably scale dependant. In other words, at regional scale, the rift faults follow the basement fabrics. However, at scales of hundreds of meters, the rift faults seem to cross-cut the pervasive foliation (Kirkpatrick et al. 2013). Further south, the Dom Feliciano and Kaoko mobile belts control the ~NE trend of the passive margin (Fig. 4.1b). However on its conjugate counterpart of the passive margin i.e. the Damarapaleo-suture (630–510 Ma) between the Congo and Kalahari cratons in the African plate occurs at a high-angle to the west African passive margin and also to the mobile belts of the Mantiqueira Orogenic System in S America. This means that this fabric possibly did not influence the genesis of this passive margin. Such trends oblique to the overall rift trend are also seen in the EARS e.g. the pervasive fabric in the EARS trends ~E-W west of Lake Kivu in the Western Branch (Fig. 4.3).

4.2.4 East and West Indian Passive Margins

The ~2000 km Late Jurassic-Early Cretaceous east Indian passive margin (Rao 2001) is ~N-S in the Cauvery and Palar-Pennar basins and NE-SW in the Krishna-Godavari and Mahanadi basins along the E coast of India formed due to rifting of India from Antarctica (Bastia and Radhakrishna 2012; Murthy et al. 2012; Roy et al. 2015; Fig. 4.7). Two mobile belts- the Meso-proterozoic Eastern Ghats Mobile Belt (EGMB) and the Paleo-proterozoic Southern Granulite Terrain (SGT) involved in the separation of India and Antarctica (e.g. review by Chetty and Santosh 2013). The NE trending EGMB is characterised by a suite of lithologies, where charnockites, khondalites, enderbites, leptynites, basic granulites, quartzites, calc-alkaline gneisses, marble, meta-volcanics are dominant (Valdiya 2010 review). The deformation is manifest by mainly NE (e.g. Sileru Shear Zone), with minor N-S trending (e.g. Nagavalli Shear Zone), ductile shear zones due to NW verging thrust sheets with ~NE-SW foliations (Chetty and Murthy 1993, 1994, 1998; Chetty et al. 2003b; Chetty 2010). The SGT is present in the southernmost part of peninsular India and resemble EGMB in lithology. The deformation in this mobile belt is represented by ~E-W crustal scale shear zones (e.g. Palghat-Cauvery Shear

Zone, Moyar Shear Zone) with minor NE /NW (e.g. Bhavani Shear Zone, Achankovil Shear Zone) trending ones (e.g. Chetty and Santosh 2013). The mobile belts are also exposed in Sri Lanka and Antarctica (Napier Complex) (Collins et al. 2014). The Cuddapah Basin is a Meso- to Neo-proterozoic sedimentary basin, which formed due to sedimentation in the foreland of the EGMB (Naqvi and Rogers 1987; review by Valdiya 2010; Chetty 2011). The characteristic crescent shaped basin (Figs. 4.7 and 4.8) occurs at the boundary of the Krishna-Godavari and the Cauvery-Palar Basins along the Indian east coast. These mobile belts and meta-sedimentary lithologies are accompanied by three cratons in the peninsular India viz. Dharwar-, Bastar- and Singhbhum cratons. The Archaean (~3300–2550 Ma) Dharwar craton is characterised by granites and gneisses, meta-volcanics and thick meta-sedimentary sequences and traversed by numerous ~NNW-SSE intra-cratonic shear zones and foliations (e.g. Chadwick et al. 2003). The Dharwar craton covers most of the Peninsular India and lies below the Cretaceous-Paleocene Deccan volcanic province (e.g. Chenet et al. 2007; Misra et al. 2014) in the N. The 3500–2500 Ma Bastar craton comprises of granites, gneisses and meta-volcanics and meta-sediments (e.g. Roy and Prasad 2003). The ~3600–2800 Ma Singhbhum

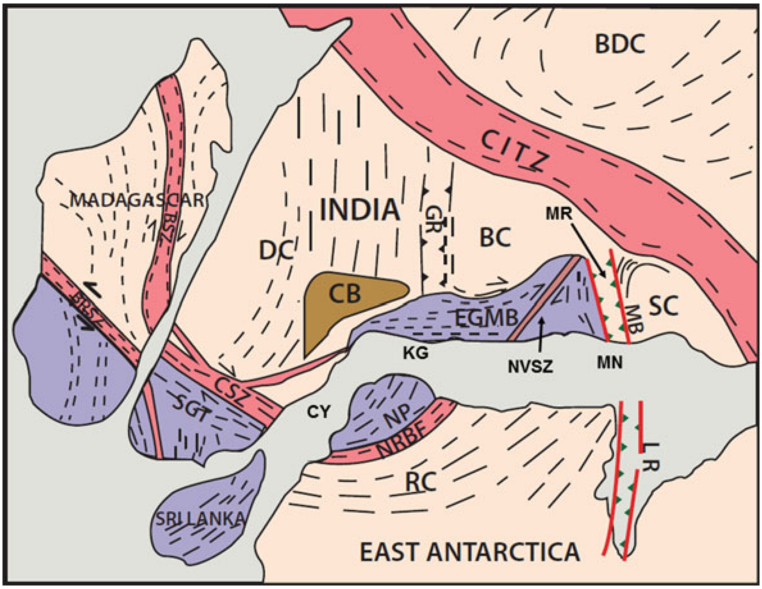
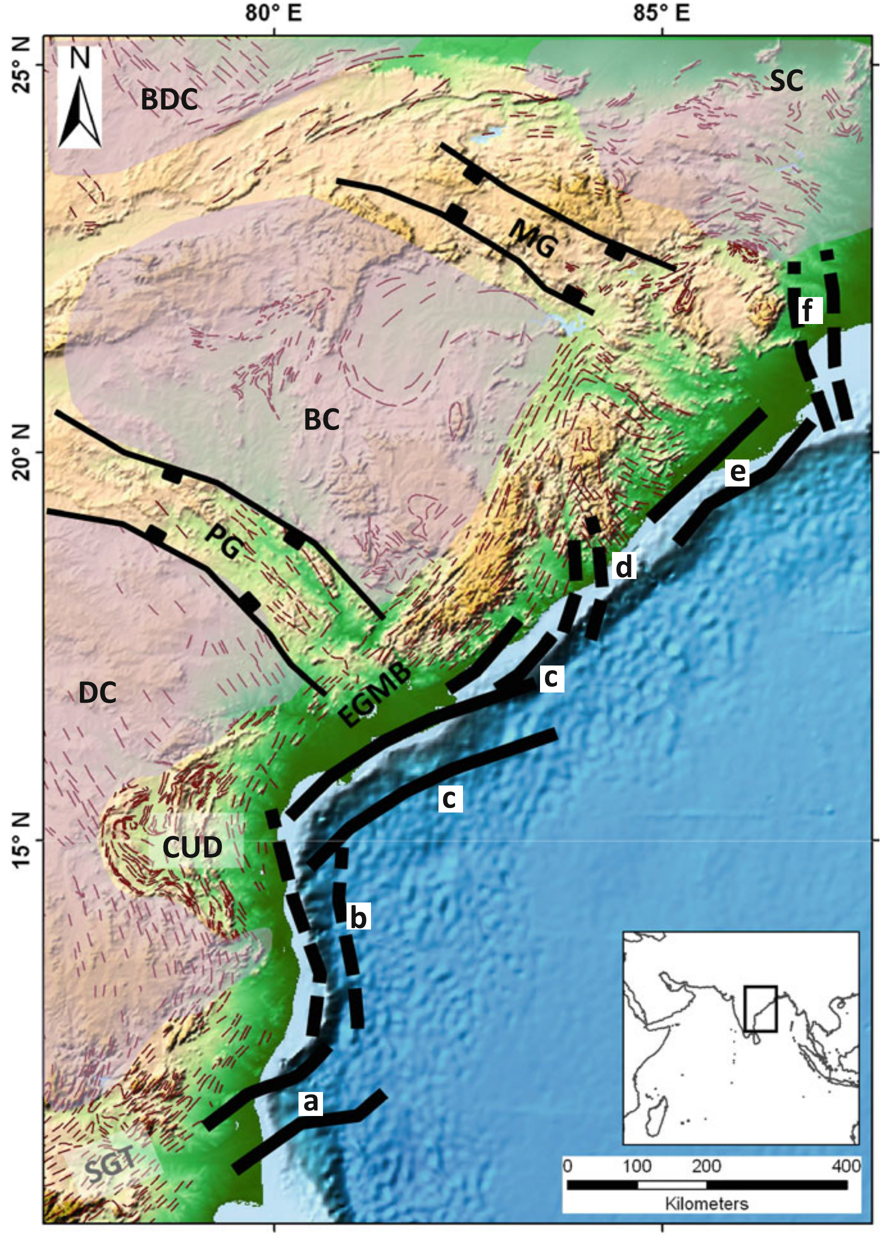


Fig. 4.7 Reconstruction of Madagascar, India, Sri Lanka and Antarctica showing the major shear zones, mobile belts and cratons. *MN* Mahanadi basin; *KG* Krishna Godavari basin; *CY* Cauvery basin; *NVSZ* Nagavali–Vamsadhara Shear Zones; *NRBF* Napier Rayner Boundary Fault; *CSZ* Cauvery Shear Zone; *BSZ* Betsimisaraka Suture Zone; *BRSZ* Bongolava–Ranotsara Shear Zone; *BDC* Bundhelkhand Craton; *BC* Bastar Craton; *SC* Singhbhum Craton; *DC* Dharwar Craton; *CUD* Cuddapah Basin; *GR* Godavari Rift; *MR* Mahanadi Rift; *LR* Lambert Rift; *RC* Rayner complex; *EGMB* Eastern Ghats Mobile Belt; *SGT* Southern Granulite Terrane; *CITZ* Central Indian Tectonic Zone (reproduced from Chetty and Santosh 2013)


80° E
85° E
25° N
20° N
15° N
N
BDC
SC
MG
BC
PG
DC
EGMB
CUD
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b
c
c
d
e
f
0
100
200
400
Kilometers

◀ **Fig. 4.8** Map of the major tectonic segments of the Indian east coast and the structural features of the adjacent onshore (modified fromNemčok et al. 2013 and references therein). There are six major segments: **a** the NE–SW-trending Cauvery rift zone; **b** the NNW–SSE-trending dextral Coromondal transfer zone; **c** the NE–SW- to ENE–WSW trending rift units of the Krishna–Godavari rift zone; **d** the NNE–SSW-trending North Vizag transfer zone between the Krishna–Godavari and Mahanadi rift zones; **e** the NE–SW-trending Mahanadi rift zone; and **f** the NNW– SSE-trending dextral Konark transfer zone. Tectonic lines (denoting predominantly foliation) in *grey lines* and cratons in *grey shaded areas*; *BDC* Bundhelkhand Craton; *BC* Bastar Craton; *SC* Singhbhum Craton; *DC* Dharwar Craton; *CUD* Cuddapah Basin; *EGMB* Eastern Ghats Mobile Belt; *SGT* Southern Granulite Terrane

craton includes granites, tonalitic gneisses, meta-volcano-sedimentary complexes and minor charnockites-khondalites (Saha et al. 2004).

The E Indian passive margin also follows closely the trend of the Precambrian foliations of the EGMB and SGT. The major rift-fault trends also parallel the foliations (Fig. 4.8). The passive margin follows the N-S trend in the Cauvery and Palar-Pennar basins, bends following the Cuddapah basin trend (Figs. 4.7 and 4.8) and follows the NE-SW trend of the Eastern Ghats mobile belt in the Krishna-Godavari and Mahanadi basin in Indian east coast. The east India passive margin avoids obviously the major cratons viz. Dharwar-, Singhbhum- and Bastar cratons (Fig. 4.8).

The W Indian passive margin also follows the NNE foliation trend of the Western Dharwar gneisses (Fig. 4.9). But, this margin did not develop along any mobile belt and separated the ~3300–2500 Ma Antongil-Masora cratons of Madagascar (Fig. 4.10) in west from the Dharwar craton of India (Veeraswamy and Raval 2004; Schofield et al. 2010). The Paleo-proterozoic to Neo-proterozoic Bemarivo mobile belt lies immediately N of the Antongil craton with the Andaparaty thrust in between (Schofield et al. 2010; Key et al. 2011) (Fig. 4.10). Note that there is a conspicuous ~NNW trending crustal-scale lithological boundary between the Dharwar craton and the Closepet Granite (Fig. 4.9), which also parallels the passive margin. Whether such deep litho-contacts influence the trend of the rifts requires more study. The western edge of the Dharwar craton weakened by repeated Neo-proterozoic to Late Paleozoic reworking of the adjacent mobile belts (Key et al. 2011) and strain localized and fragmented it (Veeraswamy and Raval 2004). The northern part of the western continental margin of India formed as a result of India-Seychelles rifting (Misra et al. 2015a) also shows inheritance of the Dharwar trend in the trend of brittle shears and dykes (Misra et al. 2015b).

4.3 Discrete Fabrics

Discrete/isolated fabrics are planar to curvi-planar elements e.g. ductile shear zones (Mukherjee 2012; Mukherjee and Biswas 2014) or fault planes that lead to anisotropy in terms of strength and material properties with respect to the

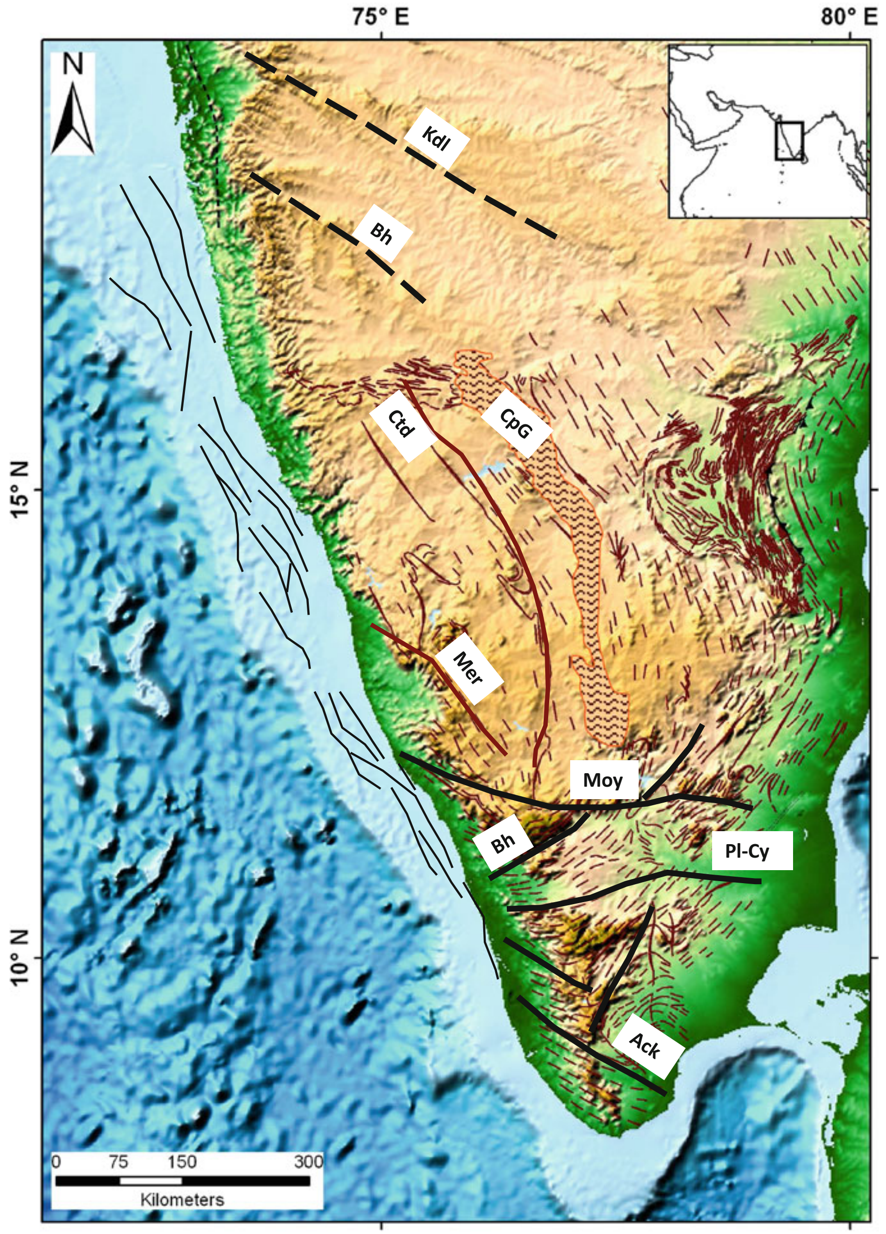

Fig. 4.9 Major tectonic elements of the western continental margin of India, showing the on-land pervasive fabrics as thin brown lines, predominantly foliation planes in Precambrian rocks. *Thin black lines* faults in the Kerala-Konkan shelf. *Kdl* Kurduwadi lineament; *Bh* Bhima Shear; *Ctd* Chitradurga Shear Zone; *Mer* Mercara Shear Zone; *Moy* Moyar Shear Zone; *Bh* Bhavani Shear Zone; *Pl-Cy* Palghat-Cauvery Shear Zone; *Ack* Achankovili Shear Zone; *CpG* Closepet Granite

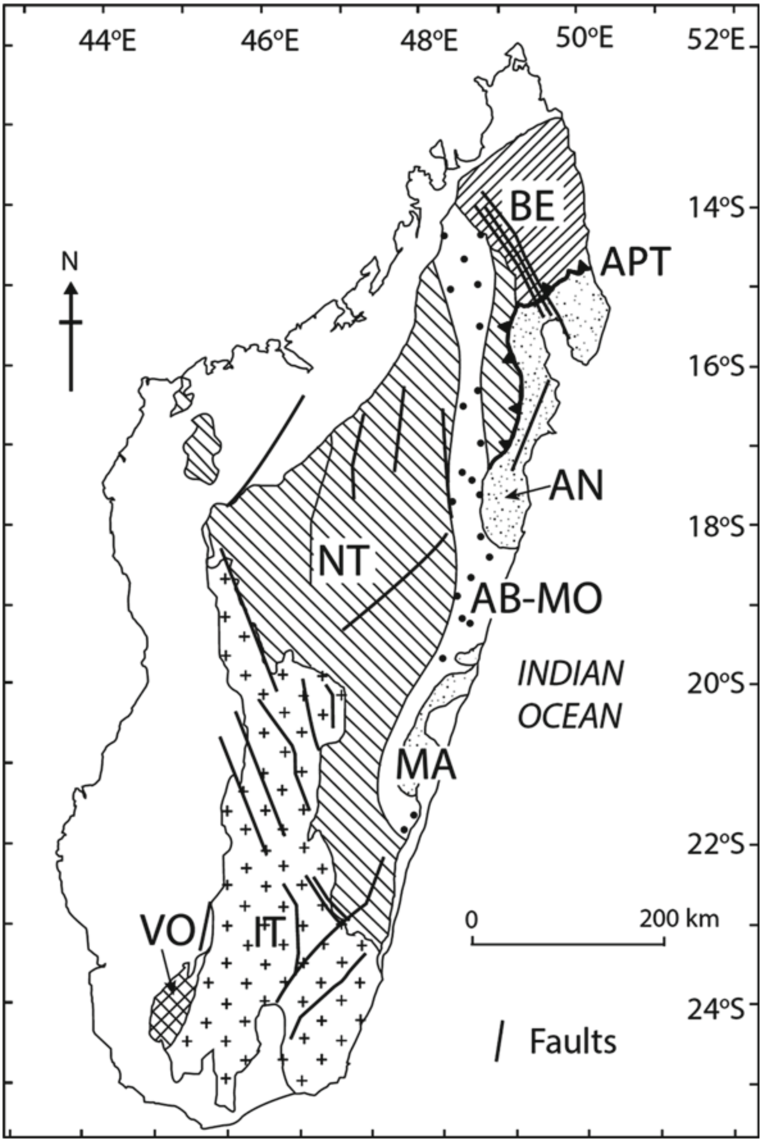

Fig. 4.10 Map showing the major Precambrian tectonostratigraphic units of Madagascar, (reproduced from Schofield et al. 2010 and references therein). *APT* = Andaparaty Thrust; *AB–MO* = Anaboriana–Manampotsy Belt; *AN* = Antongil Craton; *BE* = Bemarivo Belt; *MA* = Masora Craton; *NT* = Antananarivo Craton; *IT* = southern mobile belts including the Itremo Group; *VO* = Vohibory Unit

surrounding rocks (Morley 1999b). Deformation event(s) develop discrete/isolated fabrics such as brittle/ductile fault/shear zones, joints etc. These fabrics act as weak planes in rocks.

Notice that "fabric" refers here a crustal-scale geological feature and thrusts have been described as 'discrete fabric'. The second point to note is that the meaning of 'discrete fabric' differs from 'discrete foliation' (in the sense of Twiss and Moores 2007). 'Discrete' in both cases means isolated/wide-spaced at the scale of observation.

In contrast to pervasive fabrics, discrete fabrics influence only at large-scale: they control transfer zone geometries, prevent propagation of rift-bounding faults and segment the rift systems depending upon their anisotropic strength. Discrete fabrics do not govern the overall trend of all the rift faults and rather develop faults of unique and similar trend. Whether or not discrete fabrics govern the dip of rift faults is not yet worked out. Depending on their density and attitude, discrete fabrics segment and localise rift zones (Delvaux and Barth 2010). Discrete fabrics weaker than the rock volume reactivate during extension. Some of the discontinuities affect not only the attitude of faults under extension, but also their lengths (Versfelt and Rosendahl 1989; Morley 1999b; Morley et al. 2004). A fault may continue partly or entirely along the pre-existing fabric depending upon the strength of the fault along strike (Morley et al. 2004).

In rifting, basin-bounding normal faults develop with their strikes paralleling preferential pre-existing discontinuities. The basin formation is accentuated by selective extensional reactivation of the youngest and weakest fabrics. Here, the 'weakest' fabric is not the one with weakest isotropic strength, but are those which are most favourable to shear failure depending on the angular relationship between the fabric and the extension direction (Fig. 3.6) (Donath 1961; Swanson 1986; Morley 1999c).

When weakest fabrics reactivate, strain focuses at weak planes and rift-bounding faults propagate laterally. These fabrics have near-zero cohesion and almost no friction for stress needed to propagate rift-bounding faults (Versfelt and Rosendahl 1989; Nemčok et al. 2007). The dip of pre-existing discrete fabric also has bearing in reactivation. In analogue models and natural prototypes, fabrics reactivate as normal faults if they dip at > 40° (Faccena et al. 1995).

Discrete anisotropies in the basement rocks influence the fracture pattern in younger overlying cover showing parallel relationship (e.g. Cortés et al. 2003). This might be because the discrete fabrics reactivate and do not produce new faults depending on the following factors (i) The deviation in the attitude of the pre-existing fabric from the ideal attitude of fresh faults due to the prevalent stress field (Fig. 3.6) (Morley 1999b; Morley et al. 2004; Morley 2010). (ii) The difference in strength between the pre-existing fabric and the rock devoid of such fabrics (McGill and Raney 1970; Ranali and Yin 1990; Ranalli 2000). The strength depends on coefficient of friction, cohesion, depth, pore fluid pressure (Sect. 3.1). Figure 3.2 present failure of intact rock and weaker rock associated with pre-existing shear zones in the Mohr space. For pre-existing weak zones, the discontinuities with favourable orientations reactivate (McGill and Raney 1970; Ranalli and Yin 1990)

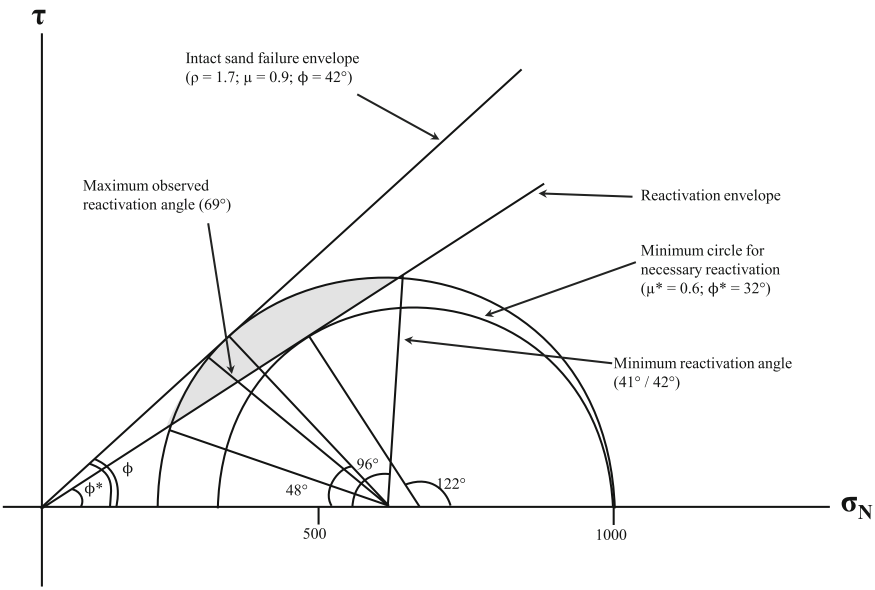

Fig. 4.11 An example case (for sandstone) for reactivation of anisotropies with a favourable angular relationship with the stress state, represented by the Mohr circle; ρ = density of rock; μ = co-efficient of sliding friction; φ = angle of internal friction; μ^* = modified μ due to anisotropy; φ^* = modified φ due to anisotropy; reproduced from Paterson and Wong (2005)

(Fig. 4.11). (iii) The sense of slip on reactivation of the fabric (Morley et al. 2004). The rift faults, which were controlled initially by pervasive fabrics, interact with the new set of fractures developed by interference with discrete fabrics (Fig. 4.12). Such interactions may reorient rift faults under the same extensional regime (Morley et al. 2004). These faults may be oblique to < 80° to the overall regional extension direction. There are methods (stated below) to decipher whether faults/fractures formed or reactivated. Field-based paleostress analyses (e.g. Misra et al. 2014) may be best when relative chronology of stress markers like superposed slickensides, abutting tectonic joints etc. are available (e.g. Nickelsen 2009). Such studies can specify attitude of newly formed fractures/faults based on paleostress tensor (Bosworth and Strecker 1997; Morley 1999; Delvaux and Barth 2010; Žalohar and Vrabec 2007). An analytical technique is used, which differentiates strength between the intact rock and that of the pre-existing fabrics (Ranalli and Yin 1990).

Pre-existing discrete anisotropies also control geometry and slip-sense of faults in transfer zones (Acocella et al. 1999). Those faults tend to be strike-slip when the angle between the extension direction and the axis of the transfer zone is < 50° (Fig. 4.13; Acocella et al. 1999). When the angle (γ) between the normal of the extension direction and the anisotropy is < 90°, narrow transfer zones develop. For γ > 90°, wider transfer zones sub-parallel to the extension direction form (Acocella et al. 1999).

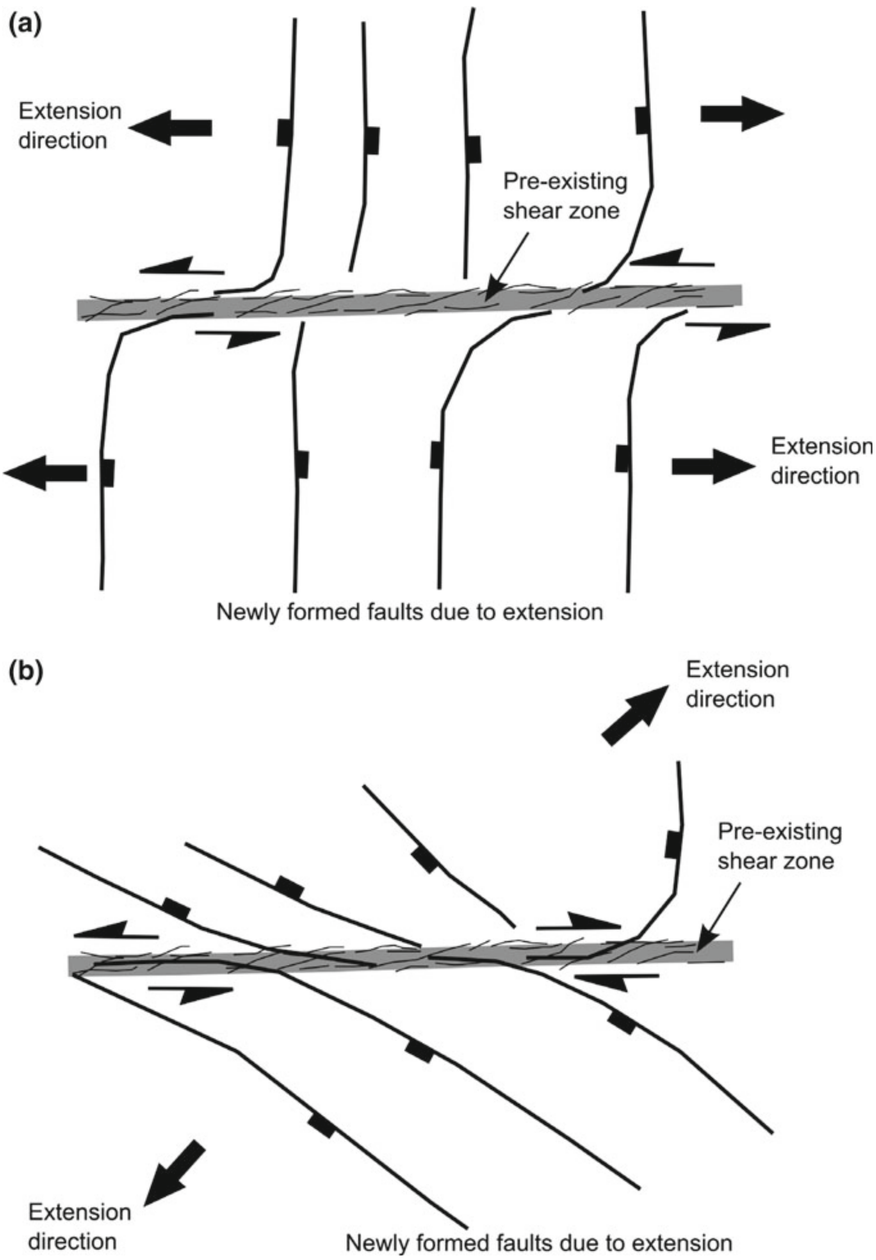

Fig. 4.12 Sketch depicting interaction of newly formed faults and oblique pervasive anisotropies (pre-existing shear zone, represented by *thick grey line*). **a** Active fabric, where the shear sense along the shear zone during later reactivation remains constant throughout the shear zone. The extensional faults mostly abut against or seldom merge tangentially into the pre-existing shear zone. **b** Passive fabric, where the sense of shear along the shear zone is not constant during later reactivation. The extensional faults mostly merge into the pre-existing shear zone tangentially

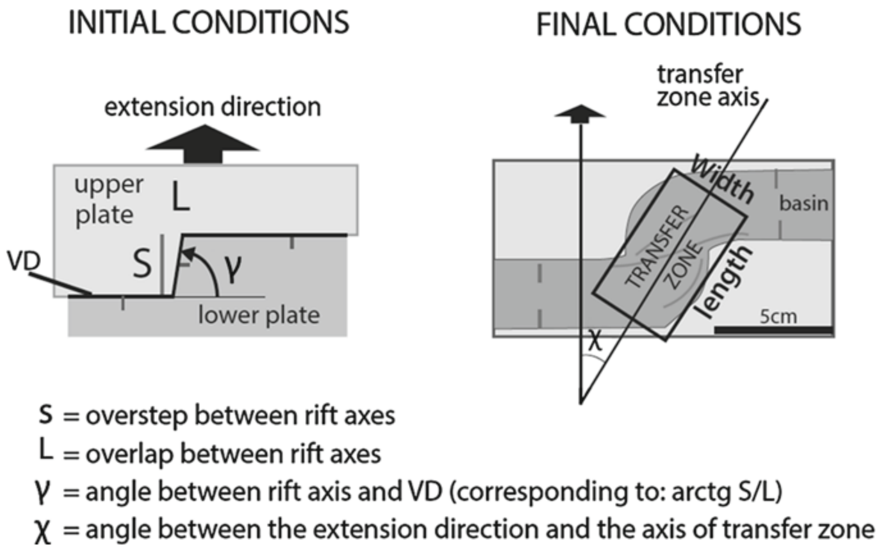

Fig. 4.13 Schematic diagram showing the different parameters for setups relating to experiments to study transfer zone geometries. **a** Imposed (initial) conditions and **b** observed (final) conditions; *VD* Velocity Discontinuity (reproduced from Acocella et al. 1999)

Pre-existing discrete fabrics control dominantly (i) geometry of primary rift faults and their temporal evolution, (ii) the geometry of transfer faults, (iii) the secondary faulting associated with primary rift faults, and (iv) and the location of depocenters/deepest parts of basins controlled by the rift faults (Dooley and McClay 1997; Bellahsen and Daniel 2005). Based on their response to extensional stresses, discrete fabrics can either be active or passive (Morley 1999c). The active discrete fabrics constitute dominant regional strike- or oblique-slip fault zones cutting across all other fabrics. Such fault planes sub-parallel the regional extension direction (Fig. 4.12a). The amount of slip and shear sense remain constant along the fault length (Morley 1999c). The passive discrete fabrics are unrelated to the extension direction. Segments of such fabrics may parallel some other faults with none or weak linkage (soft linked) (Fig. 4.12b). Such faults do not have large slips and may show opposing oblique-slips across the fault plane (Fig. 4.12b) (Morley 1999b). The fault patterns generated by discrete passive oblique fabrics vary significantly. They can prevent fault propagation, so that rift parallel faults either turn or terminate at the oblique fabric; or short rift-parallel faults may parallel an oblique fabric (Fig. 4.12b). Active- and passive discrete fabrics thus do not link directly to the composition, geometry etc. of the fabrics that define them. Rather they play role in segmentation of the margin or limiting extents of individual grabens.

4.3.1 East African Rift System (EARS)

In the EARS, the effect of pre-existing discrete fabrics is evident from oblique/sheared segments of the rift system, such as the NW trending Rukwa rift of the Tanganyika-Rukwa-Malawi (TRM) segment. The normal and oblique slip faults in the Rukwa rift parallel the pervasive fabrics in the Pre-Cambrian basement (Fig. 4.3). The basin bounding Lupa fault (Figs. 4.3 and 4.14) formed by Tertiary reactivation of Permo-Triassic Karoo rift faults. The Karoo rift faults and thus the Lupa fault also follow the Precambrian pervasive fabric (Morley 1999b). N to Lake Victoria, the NW trending ~300 km Aswa Shear Zone (=Aswa Dislocation, Aswa mylonite belt) of Neoproterozoic age is the most conspicuous oblique element in the EARS (Chorowicz 1989; Grantham et al. 2003; Schlüter 2006). The Aswa Shear Zone terminates the northernmost part of the western branch of the EARS.

The offset zone of the central Kenya rift (eastern branch of the EARS) also coincides with the Aswa Shear Zone (Fig. 4.3). Both the oblique trends viz. the Rukwa rift and the Aswa shear zone are sub-parallel, dextral and with minor dip-slip components. They seemed to be discrete fabrics of active nature (Chorowicz and Munkonki 1980). However, later studies (Grimaud et al. 1994; Coussement 1995) revealed the fabric to be passive. Analogue models for the EARS explained the role of discrete anisotropies in the mechanism of the individual

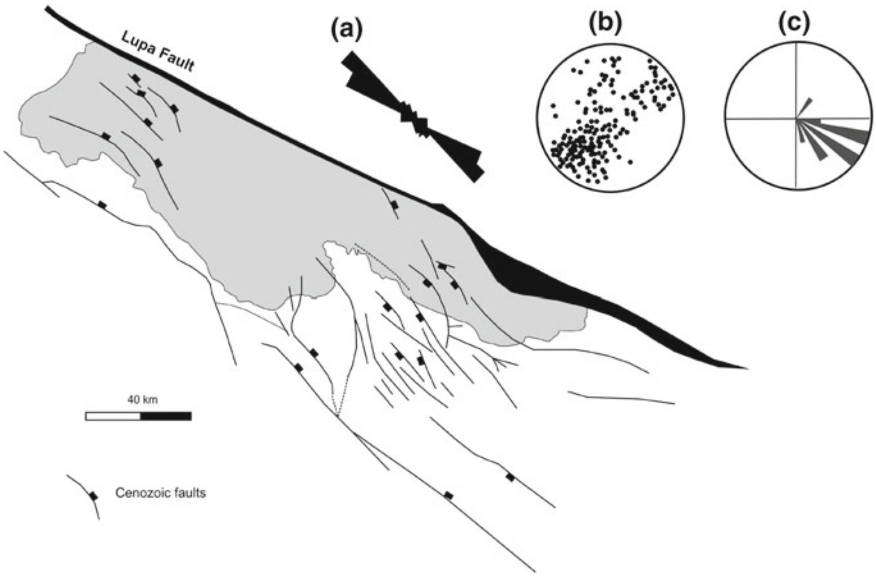

Fig. 4.14 Structural map (in two-way-time) on top of basement at the Rukwa basin. See Fig. 4.3 for location (reproduced from Morley 2010 and compilations therein). **a** Rose diagram of newly formed faults; **b** Stereonet of poles of Pre-cambrian foliation surrounding the Rukwa basin; **c** Rose diagram of SH_{max} orientations calculated from earth quake focal mechanisms around the Rukwa basin

rifted grabens (Corti et al. 2007; Aanyu and Koehn 2011). The rift oblique grabens such as Lake Rukwa (Figs. 4.3 and 4.14) develop lesser subsidence than rift perpendicular grabens like the Lake Tanganyika (Fig. 3.6).

The pre-existing E/ESE trending Sekaka shear zone, possibly of Neoproterozoic age, caused termination of the Okavango rift at south (Modisi et al. 2000; Kinabo et al. 2007, 2008). Therefore, pre-existing discrete fabrics prevented along-strike rift-fault propagation. Such shear zones usually evolve into accommodation (transfer) zones in rift evolution.

4.3.2 The Brazilian Rifts

Discrete fabrics affected the large-scale rift at the NE Brazilian passive margin. Here, the pre-existing Neoproterozoic fault zones constitute the pre-rift structural grain of the ~620–580 Ma (van Schmus et al. 2008) Pan-African (locally, Brasiliano) orogeny. NE trending fold belts and E/NE striking fault zones characterize this structural grain (Corsini et al. 1991; de Matos 1992; de Castro et al. 2007; van Schmus et al. 2008). Out of these, the most prominent is the E trending tens of km wide Neoproterozoic Patos- and Pernambuco- Shear Zones that prevented the along axis propagation of the N trending Brazilian rifts. The Pernambuco Shear Zone separates the southern Jatobá-Recôncavo-Tucano (JRT) rift system from the central Araripe rift basin, which is separated from the northern Potiguar basin by the Patos Shear Zone (de Matos 1992; Destro et al. 2003) (Figs. 4.6 and 4.15). Thus, the Patos- and the Pernambuco fault zones acts as a mega transfer zone between the southern and the northern rifts. The NW-SE extension during Cretaceous induced inversion and formed normal faults in the pre-existing transpressional NE-SW segment of the Patos Shear Zone (see Fig. 13 of de Matos 1992). The Cretaceous rift basins: Icó, Iguatu and Rio do Peixe basins in Brazil developed interesting geometries. The basin bounding faults reactivated the Neoproterozoic shear zones from isolated half grabens by transtension (Fig. 4.15). The Rio do Peixe basin has three sub-basins named Brejo das Freiras, Souza and Pombal, all of which also developed their main rift faults along the Neoproterozoic shear zones (Fig. 4.15). The larger Brejo das Freiras and Souza sub-basins are separated by the Santa Helena high, which formed since the NE trending Portalegre Shear Zone interacted with the E trending Malta Shear Zone (de Castro et al. 2007). This shows how pre-existing discrete fabrics control the geometry and extent of rift basins.

4.3.3 Tertiary Rifts of Thailand

In the Tertiary rifts of Thailand, the ENE-WSW Uttardit fault zone follows a mega-scale discrete fabric: the Nan-Uttardit suture. The Uttardit fault zone parallels Precambrian foliation trends of the Indo-Sinian orogeny near the Nan-Uttardit

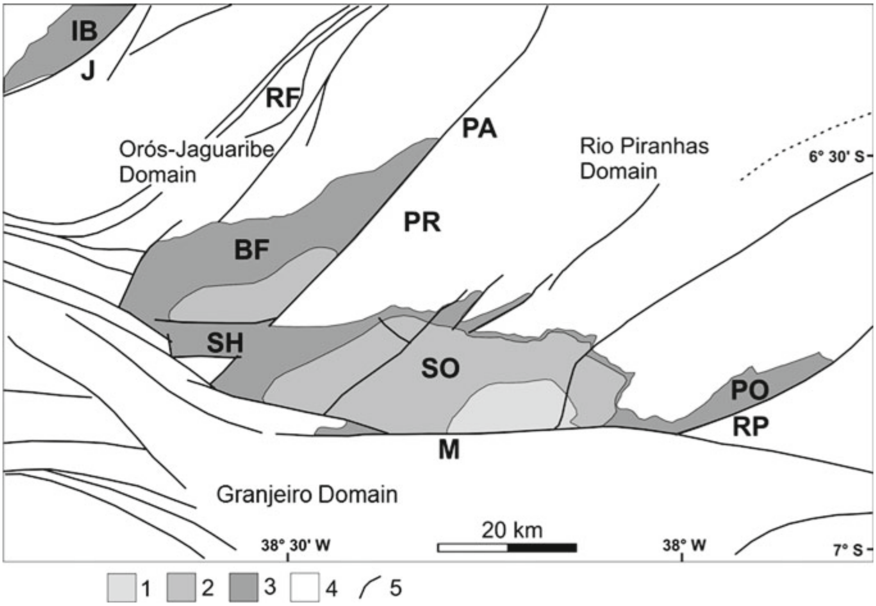

Fig. 4.15 Map of the Rio de Piexe basin showing the Geology of the Rio do Peixe Basin (modified from de Castro et al. 2007), located at the intersection of three structural domains: Orós–Jaguaribe, Rio Piranhas, and Granjeiro. Early Cretaceous basins: Icó (IB) and Rio de Peixe (PR, sub-basins: BF, Brejo das Freiras; SO, Souza; PO, Pombal). *1* Rio Piranhas Formation, *2* Sousa Formation, *3* Antenor Navarro Formation; *4* pre-Brasiliano basement; *5* faults (J, Jaguaribe; RF, Rodolfo Fernandes; PA, Portalegre; M, Malta; RP, Rio Piranhas); SH, Santa Helena High

suture zone. The Nan-Uttardit suture zone is oblique (40–50°) to the regional extension direction during late Paleogene–recent (Fig. 4.4; Fig. 9 of Morley et al. 2004). Due to this obliquity, the Uttardit fault zone slipped obliquely and retained half-graben like geometries of orthogonal rift basins (Bal et al. 1992; Morley et al. 2004; Morley 2007). The sinistral strike-slip component of this fault zone manifests maximum thickness of sediments of Uttardit basin against a transtensional bend in the Uttardit fault zone. This indicates an oblique-extensional nature of the Uttardit fault zone (see Fig. 10 of Morley et al. 2004; Fig. 4.5). Other basins e.g. the Chiang Mai-, Phrae- and Fang basins show dissimilar geometries depending on the angular relationship between the pre-existing anisotropy and the regional stress directions (Fig. 4.16). This shows the profound influence of the discrete fabrics on the geometry of the rift basins.

4.3.4 North Atlantic Passive Margin

The North Atlantic opening is a complex rift history starting from Paleozoic and completing in early Cenozoic by breaking eastern Greenland from western Europe

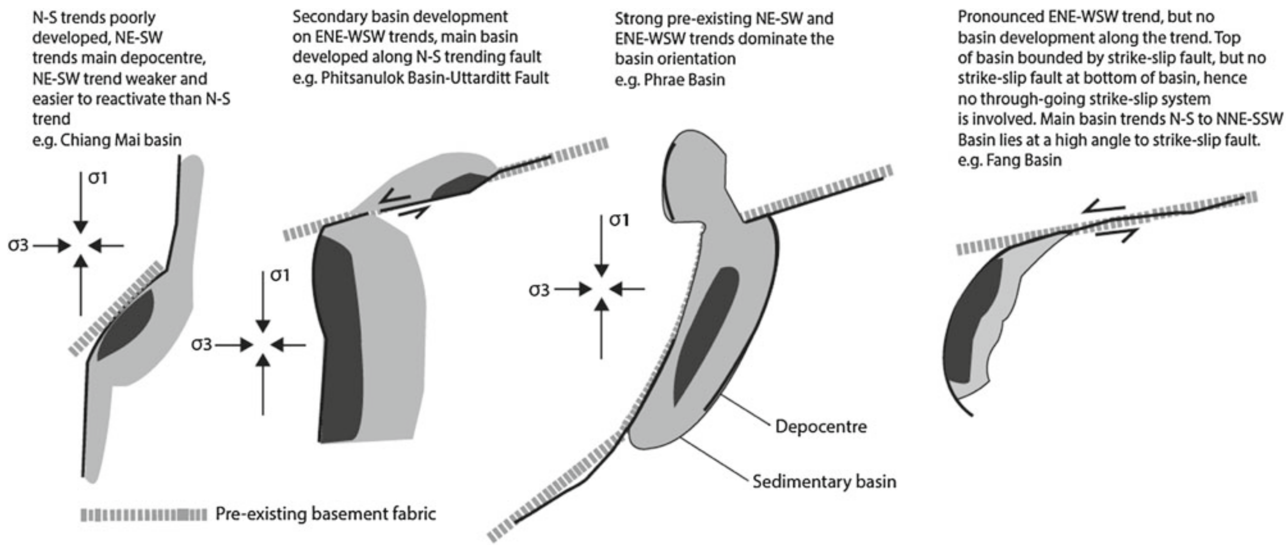

Fig. 4.16 Oblique-extensional rift basin geometries by interaction of the rift faults with pre-existing discrete oblique anisotropies (reproduced from Morley 2007)

(Tsikalas et al. 2012 review; Khani et al. 2015). The west Norwegian margin is a classic example of syn-orogenic to post-orogenic extension related to gravitational collapse of Scandinavian Caledonian orogen (e.g. Andersen 1998; Fossen 2000). It obviously follows that rifting continued along the orogenic belt. The extensional tectonics reactivated the same faults as normal faults NW to N trending deep crustal shear zones e.g. Hardangerfjord Shear Zone and the Bergen Arc Shear Zone (discrete anisotropies). Such shear zones were compressional (indicated by reverse faults) during the Caledonian orogeny (Osmundsen and Andersen 1994; Fossen 2000). Extensional faults also parallel ~NE trending shears and litho-contacts between Precambrian and Paleozoic in the Møre-Trøndelag Fault Zone/Complex (MTFZ/C; Fig. 4.17; Doré et al. 1997; Lidmar-Bergstrom and Naslund 2002; Nasuti et al. 2012). The MTFZ accommodated strike- to oblique-slip deformation in Jurassic and thereafter in Cretaceous-Cenozoic the extension was precisely orthogonal. This is because after Cretaceous, the MTFZ was perpendicularto the regional extension direction, following the angular relationship between the Jan Mayen Fracture Zone and the MTFZ (Fig. 4.17) (Doré et al. 1997; reviews by Nasuti et al. 2012). During Jurassic, the MTFZ probably acted as mega-scale transfer zone and inhibited along-strike fault propagation between the North Sea-Viking Graben and the Møre-Vøring basin (Doré et al. 1997).

In the Paleozoic-Mesozoic North Sea rift (Fig. 4.18), pre-existing N-S and NW trending shear zones of the Early Paleozoic Caledonian orogeny reactivated and numerous new rift faults also formed depending on strain localization (Whipp et al. 2014). The North Sea rifted in two stages: an early E-W Permo-Triassic (possibly as early as Devonian) and a subsequent N-S Late Jurassic extension (Whipp et al. 2014; Khani et al. 2015). The N-S earlier Permo-Triassic rift faults reactivated during Jurassic and also formed new E-W trending fault systems (Fossen 2013). Recent seismic data interpretation (by Khani et al. 2015) shows that physical linkage occurs between the basement reflections and overlying Permo-Triassic faults along strike, but only locally. Nevertheless, the basement structures have played an important role in controlling the locations and trends of major rift related faults in the overlying sediments. Interestingly, heterogeneities of a part of the North Sea crust deciphered from geophysical studies could be due either to inheritance or serpentinization (Fichler et al. 2011).

4.3.5 Eastern North American Rift System

The eastern North America preserves two tectonic stages of the Wilson cycle: Iapetus Ocean opened at ~530 Ma, contractional structures of the Rodinia assembly at ~1350–1000 Ma, and late Triassic opening of the Atlantic paralleled those of the Pangea assembly of Ordovician-Permian Periods (reviews by Thomas 2006; Withjack et al. 2012). Analyses of pre-existing pervasive-and discrete fabrics and their relation with individual rifted grabens do not exist possibly because geometry of the buried basins remained indeterminate (reviews by Withjack et al. 2012).

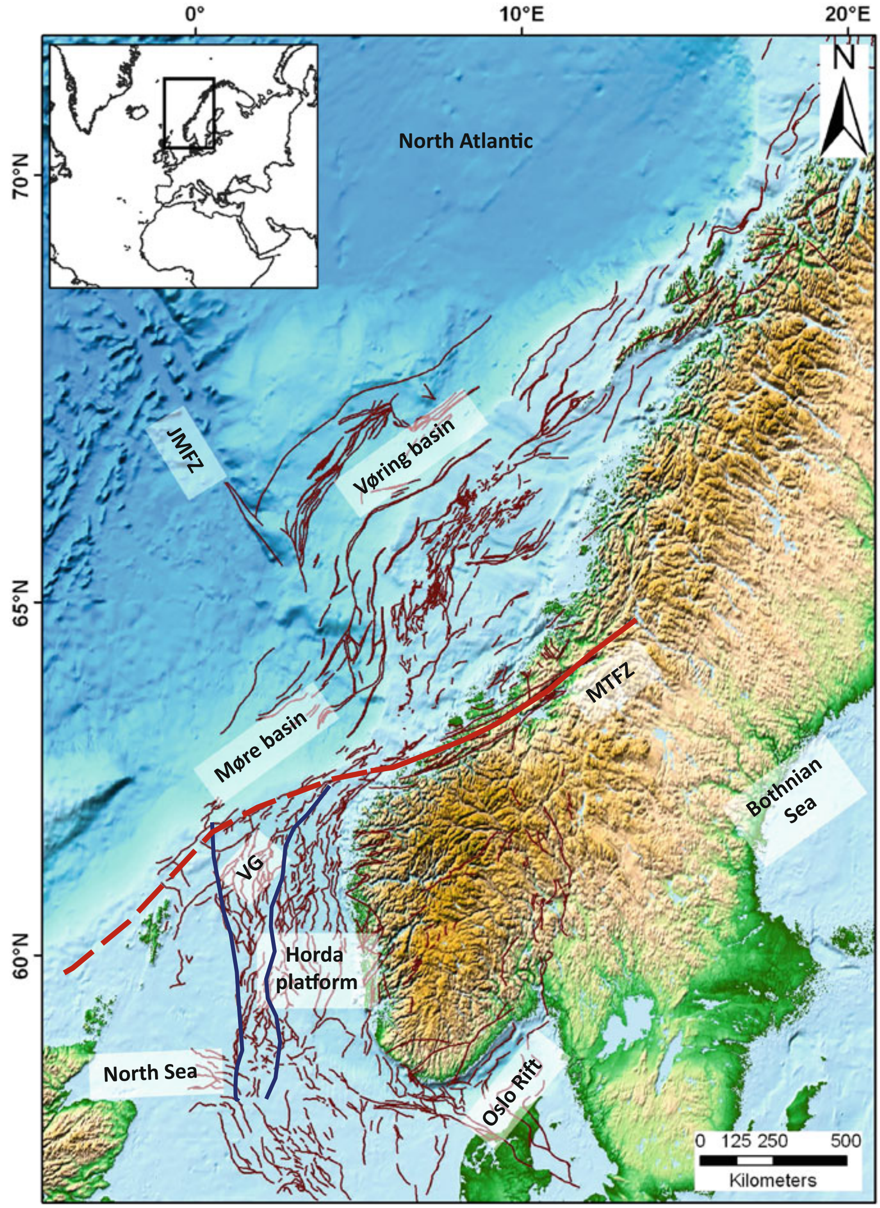

Fig. 4.17 Tectonic elements map of the Norwegian passive margin. *MTFZ* Møre-Trøndelag Fault Zone; *JMFZ* Jan Mayen Fracture Zone; *VG* Viking Graben. Brown lines denote rift faults (after Norwegian Petroleum Directorate)

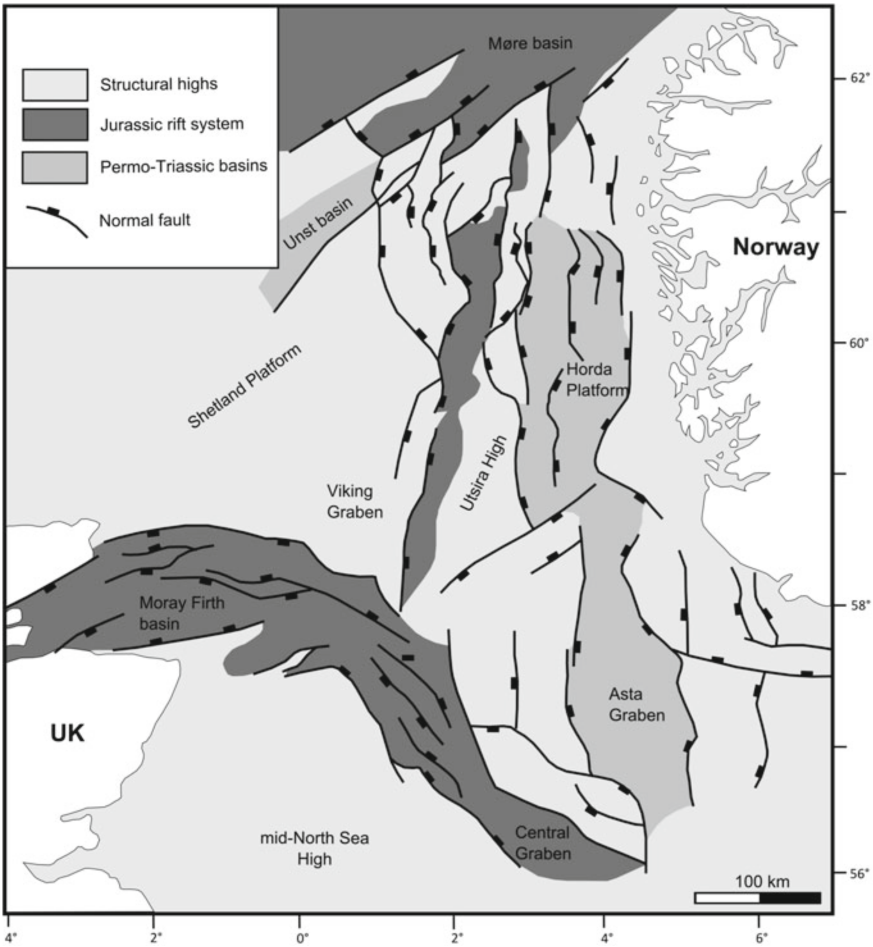

Fig. 4.18 Structural elements map of the North Sea showing the major rift grabens with the rift bounding faults (reproduced from Whipp et al. 2013)

However, a discrete fabric in the form of a dextral bend in the Grenville front (Fig. 4.19a) was inherited as the Iapetan Alabama-Oklahoma transform (Fig. 4.19b) (Thomas 2010). The WNW trending Bahamas fracture zone subsequently inherited the Alabama-Oklahoma transform (Thomas 1988). During Mesozoic rifting, high-angle reverse faults and strike-slip faults of Paleozoic and older age reactivated as discrete fabrics during rifting between eastern North America and western North Africa (Fig. 4.20; Swanson 1986; Huerta and Harry 2012; Withjack et al. 2012). Geophysical studies also indicated a strongly anisotropic crust and upper mantle due to orogeny that later inherited in the rifting events (e.g. Vauchez and Baruol 1996; Pollitz and Mooney 2014).

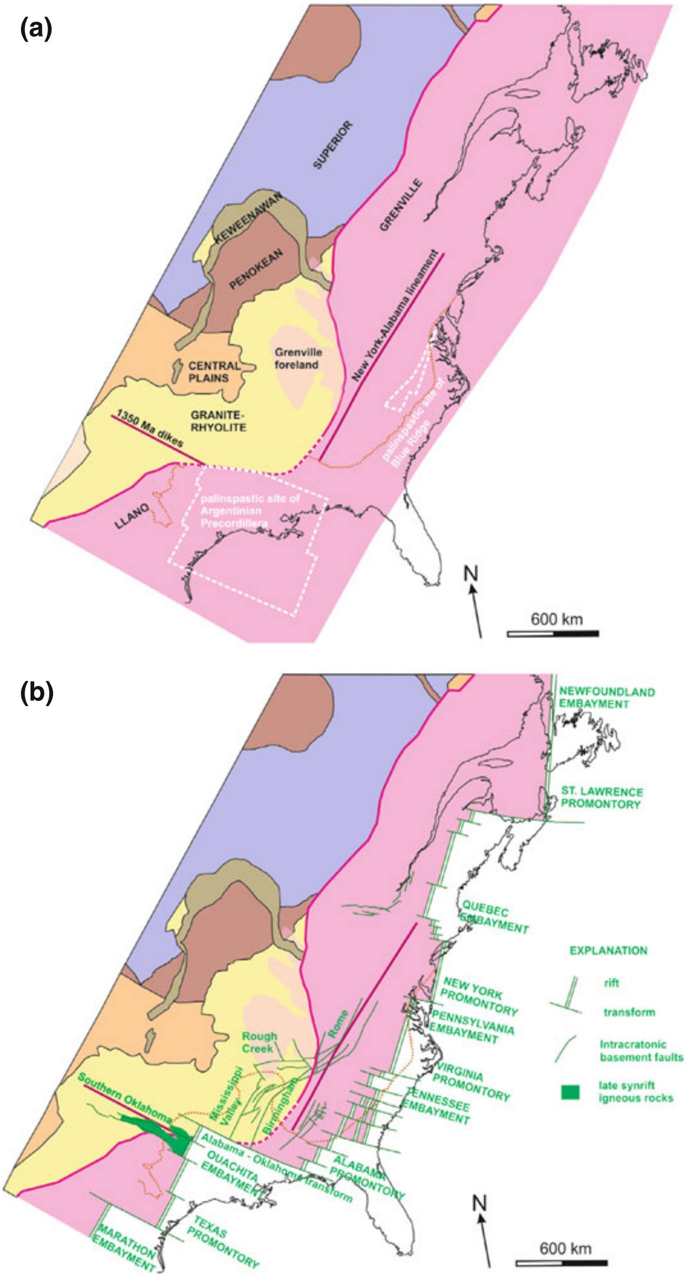

Fig. 4.19 Schematic illustrations of **a** The Grenville orogeny during the assembly of Rodinia; **b** opening of the Iapetus Ocean during breakup of Rodinia (reproduced from Thomas 2006)

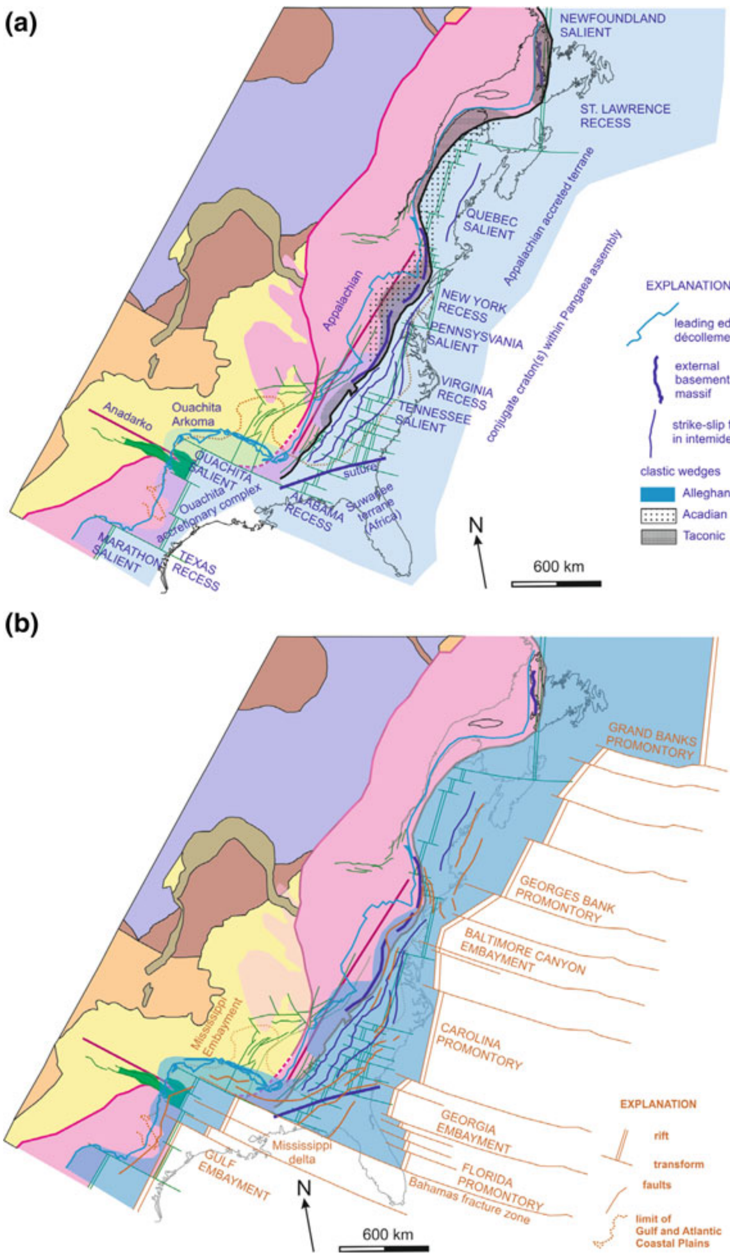

Fig. 4.20 Schematic illustrations of **a** assembly of Pangaea due to the Appalachian-Ouachita orogeny; **b** formation of the Atlantic Ocean following the breakup of Pangaea. (reproduced from Thomas 2006)

4.3.6 Rhine Graben

Pre-existing NE trending discrete anisotropies influenced the Upper Rhine Graben (W Europe) selectively in its northern part up to Late Oligocene (Eden et al. 2007). Regional stress reorientation from Oligocene onwards reactivated existing faults in Upper Rhine Graben (Homuth et al. 2014). The anisotropies reactivated as sinistral strike-slip faults and dip-slip to oblique-slip faults to form the Rhine Graben by NW-SE extension during Mid Oligocene. In its S part, the Upper Rhine Graben transferred strain towards W along the Rhine-Bresse Transfer Zone to the Bresse Graben (Fig. 4.21) (Frisch et al. 2011; Jolivet et al. 2013).

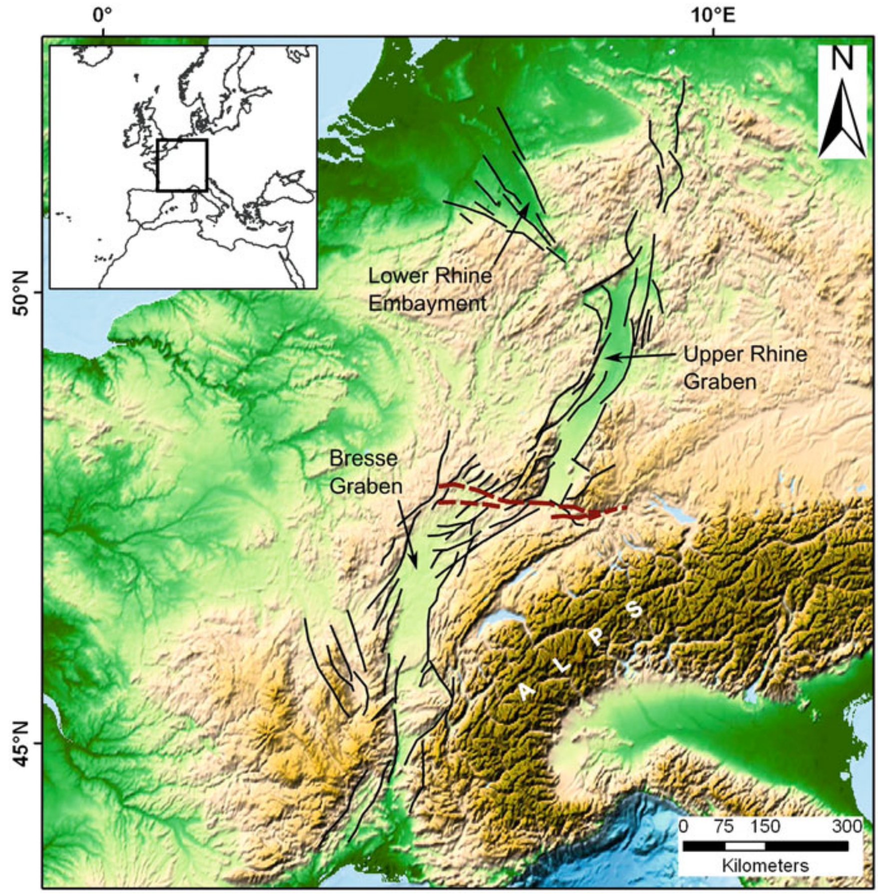

Fig. 4.21 Fault map of the Central European rift system showing the relationship of the Upper and Lower Rhine Grabens and the Bresse Graben. The presence of an accommodation zone between the Upper Rhine and Bresse grabens helps the transfer of deformation towards the later (adapted from Frisch et al. 2011)

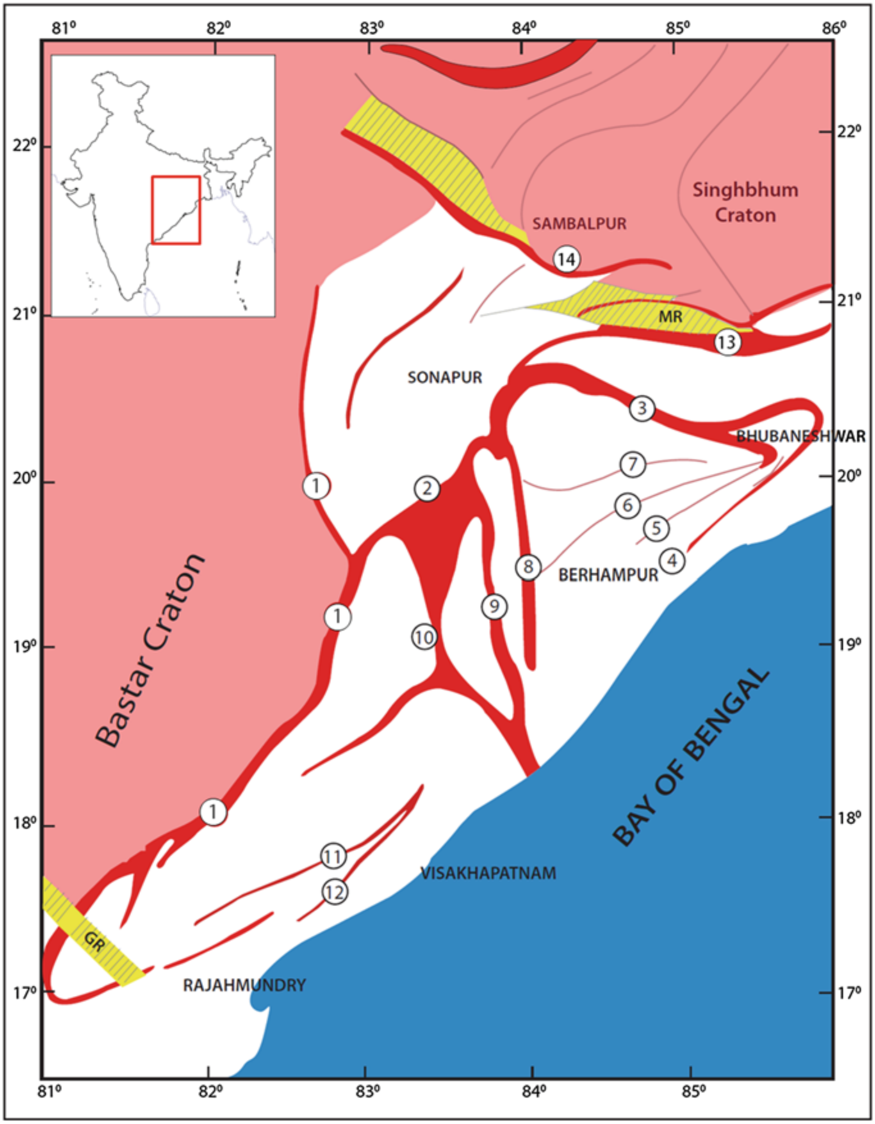

Fig. 4.22 Map of the Eastern Ghats Mobile Belt showing the major shear zones, *1* Sileru Shear Zone; *2* Koraput–Sonapur Shear Zone; *3* Mahanadi Shear Zone; *4* Chilka Lake Shear Zone; *5* Digapahandi Shear Zone; *6* Aska Taptapani Shear Zone; *7* Banjanagar Shear Zone; *8* Balgura Tel Shear Zone; *9* Vamsadhara Shear Zone; *10* Nagavali Shear Zone; *11* Narsipatnam Shear Zone; *12* Tuni–Eleshwaram Shear Zone; *13* Angul-Dhenkanal Shear Zone; and *14* Northern Boundary Shear Zone; *MR*—Mahanadi Rift, *GR*—Godavari Rift (reproduced from Chetty and Santosh 2013)

4.3.7 East and West Indian Passive Margins

The eastern continental margin of India formed during Late Jurassic to Early Cretaceous by rifting and breakup of India and Antarctica (Rao 2001). This margin formed along the Mesoproterozoic Eastern Ghats Mobile Belt (EGMB) in India and the Napier complex in Antarctica (Chetty and Santosh 2013). In the mobile belt, majority of the shear zones trend ~NE and parallel the overall rift trend, apart from the NNW trending Nagavalli- and Vamsadhara Shear Zones and NW trending Mahanadi- and the Northern Boundary Shear zones (Fig. 4.22). These transverse shear zones probably precluded the along strike propagation of the rift faults (Fig. 4.8). The Nagavalli- and the Vamshadhara shear zone acted during Neo-Proterozoic as large transfer zones in the form of the North Vizag Transfer Zone (Fig. 4.8) between the Krishna-Godavari- and the Mahanadi rift zones (Nemčok et al. 2013). The Proterozoic Mahanadi- and the Northern Boundary Shear Zones reactivated during Permo-Triassic rifting in the Indian subcontinent as the Mahanadi Rift (Fig. 4.22). The ~NW-SE trend of these shear zones mismatch with the NNW trend of the Konark transfer zone (Fig. 4.8) and the reason remained indeterminate.

The west Indian passive margin is unique since it did not break along any mobile belt but separated the Archean craton into the Western Dharwar Craton, in India, and the Antongil and the Masora cratons in E Madagascar (Schofield et al. 2010) (Fig. 4.10). The Anaboriana–Manampotsy (AM) mobile belt (Fig. 4.10) immediate W to the Antongil-Masora cratons was avoided i.e. the passive margin did not form along it. The continental breakup between India and Madagascar at ~90 Ma reactivated several shear zones e.g. the NNW trending Chitradurga Shear Zone (Sharma 2009) (Fig. 4.9) of the western Dharwar craton and the continents separated along one of them (Raval and Veeraswamy 2003). Interestingly, a segment of the West Indian passive margin also parallel the NNW trend of the Closepet granite (Fig. 4.9). Repeated tectonism along the Anaboriana–Manampotsy mobile belt probably weakened the Western Dharwar-Antongil-Masora craton. Possibly, the cratonic boundary became weaker than the mobile belt itself and thus broke the craton instead of along the mobile belt (Raval and Veeraswamy 2003). The ~E to NW trending shear zones of the Southern Granulite Terrain e.g. the Palghat-Cauvery- Shear Zone, Moyar-, Bhawani-, Achankovil- shear zones (Santosh and Sajeev 2006), cumulatively active during Neo-Proterozoic (Santosh and Sajeev 2006), possibly have little influence on the passive margin architecture. Neither the LateCretaceous-Early Paleocene off-shore faults nor the Western Ghats Escarpment deviate significantly from the NNW trend (Fig. 4.9).

Chapter 5
Role of Lithosphere Rheology on Rift Architecture

5.1 General Discussion

Extensional geodynamics is controlled strongly by rheology- most important being strength and pore fluid pressure (Buck 1991; Vilotte et al. 1993; Bassi et al. 1993; Bassi 1995; Cloetingh et al. 1995; Ranalli 2000; Corti et al. 2003 and references therein; van Avendonk et al. 2009), and thickness of the pre-rift lithosphere (Cloetingh et al. 1995; Hirth and Kohlstedt 1995; van Avendonk et al. 2009). The other obvious parameters are extensional stress rates, duration and change in direction of extension. Rheology considerably varies spatially and temporally and segments rifts and passive margins (Ranalli 2000).

5.2 Lithospheric Strength

Lithospheric strength inversely relates to its thickness and temperature (Lynch and Morgan 1987) and also to its composition (Kusznir and Park 1987). Mobile belts, which are fossil orogens, have a warmer and thicker ($> \sim 80$ km) lithosphere than "Normal" crust. They have much warmer geotherm than Acrhean cratons (Fig. 5.1), The strength profiles in the "normal", mobile belt and cratonic lithospheres are characteristic (Fig. 5.1). The mobile belt lithospheres or Alpine lithosphere (e.g. Cloetingh et al. 1995) usually have very low crustal tensile- and compressive strength in the crust and upper mantle. 'Normal' and cratonic lithosphere are equally strong.

Reorientation of stress axes can uplift rift shoulders (Favre and Stampfli 1992). The cratonic lithospheric mantle shows large strength to depths of >80 km. Generally, rifting along mobile belts, e.g. South Atlantic passive margins, causes low topography rift shoulders. Whereas, high topography rift shoulders develop when cratons break e.g. the Saudi Arabian Red Sea margin or the Trans-Atlantic

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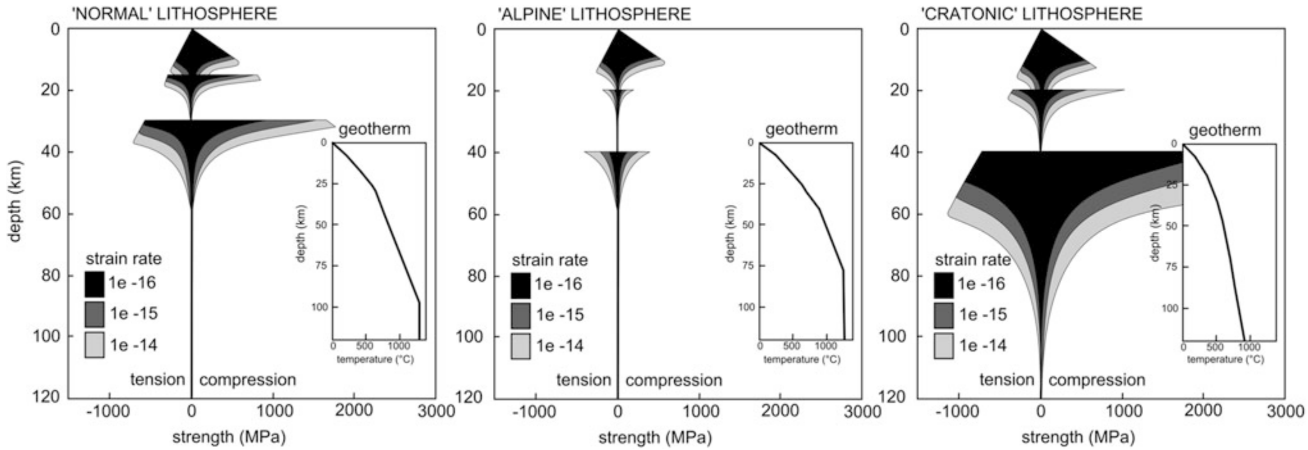


Fig. 5.1 Lithospheric strength profiles of **a** 30 km thick normal crust, **b** 40 km thick Alpine (or *mobile belt*) crust with; and **c** cratonic with 40 km thick crust, assuming the geotherms inset within each. These profiles further assume a two layer crust with a quartzitic upper continental crust and a dioritic lower continental crust. Also, an olivine-rich strong lithospheric mantle is considered (reproduced from Cloetingh et al. 1995)

Mountains (Cloetingh et al. 1995; Ollier and Palin 2000). The high topography rift shoulders of western India and Madagascar passive margins also represent such a case (Ratheesh Kumar et al. 2014). More precisely, the rift shoulder topography depends on the necking depth/level of necking for a layer (Braun and Beaumont 1989), which in other words, is the depth of a strong zone (Fig. 5.1; Watts 2012). This depth, same as the 'intrinsic' necking depth of Braun and Beaumont (1989), is defined by a layer of strong olivine-rick mantle beneath the Moho, which would remain straight without isostasy (Braun and Beaumont 1989). The necking depth, in turn, is related to faulting in the brittle- and thinning in the ductile layer (Mohn et al. 2012). Shallow depth/level of necking can uplift significant proportion of mantle, and vice versa (Watts 2012). The rift shoulder uplifts when the intrinsic necking depth exceeds the actual isostatic cratonic necking depth (Fig. 5.2b). Thickness of crust, and the difference in density between mantle and crust decide the force for shoulder uplift (Buck 2007). The topography diminishes in the reverse case such as in Alpine lithosphere (Fig. 5.2d) (Braun and Beaumont 1989; Weissel and Karner 1989; Kooi et al. 1992; Cloetingh et al. 1995). However, places such as South Africa previously thought to be elevated by inheritance may not be so (Japsen et al. 2006). With the same stretching model i.e. equal stretch factors: β (McKenzie 1978), δ (Royden and Keen 1980), or α (Turcotte and Schubert 2002) values, different degrees of necking strongly affect Moho depth predictions, rift shoulder development, basement topography, thickness ratios of syn-rift and post-rift sediments and free air gravity anomalies (Kooi et al. 1992; Cloetingh et al. 1995; Figs. 5.2 and 5.3).

Strength profiles show that the strongest layer lies near the top of the lower crust for Alpine lithosphere i.e. at the brittle ductile transition zone at ~8–15 km depth

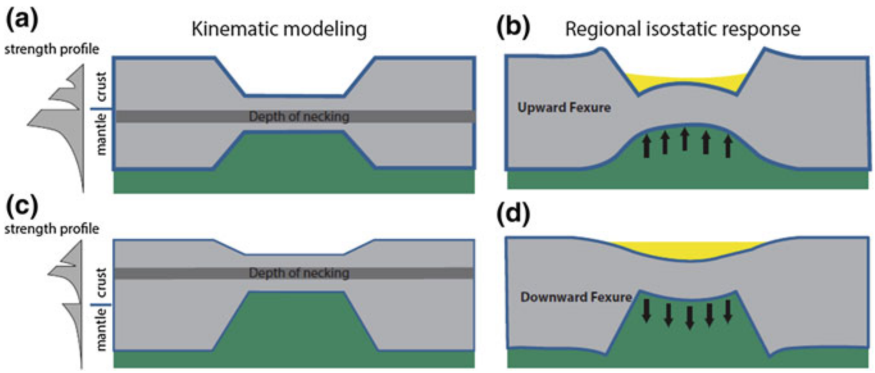


Fig. 5.2 Schematic diagrams illustrate the concept of necking depth/level. **a** and **b**: Cratonic lithosphere; **c** and **d**: mobile belt/orogenic lithosphere; **a** and **c**: intrinsic necking levels devoid of isostasy; **b** and **d**: state of the lithosphere under regional isostasy. *Green* asthenospheric mantle; *Yellow* sediments; *Grey* lithosphere. Note the strength profile for the strongest level in the lithosphere. This strong layer marks the level of necking; modified from Buck (1991) and Cloetingh et al. (1995)

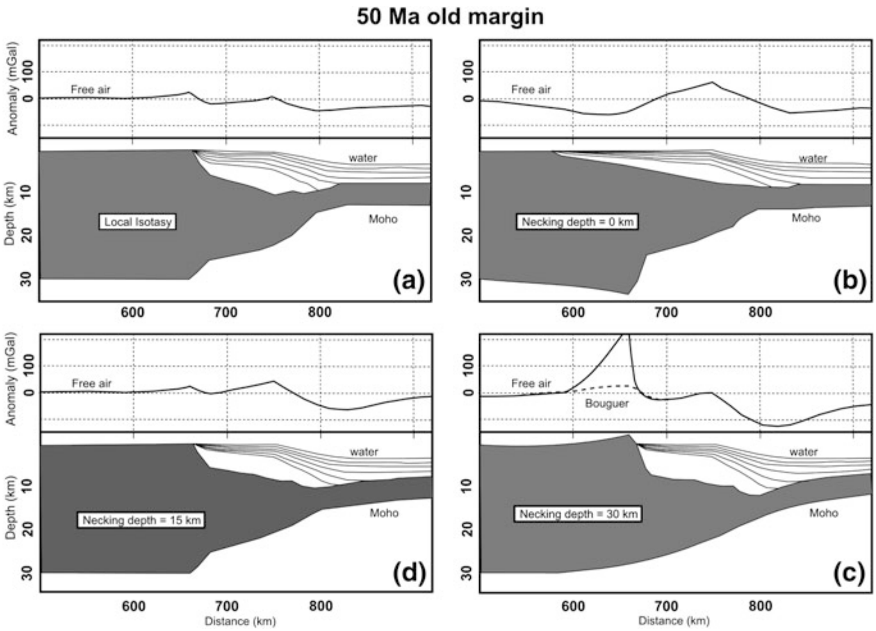


Fig. 5.3 Modelling predictions (for a 50 Ma old margin) for crustal structure, gravity and general basin stratigraphy illustrated for **a** a local isostatic model and **b**–**d** models with different levels of necking (0, 15, 30 km), implementing the same stretching model. Different levels of necking predict differently Moho depth and basement topography, rift shoulder development, thickness ratios of syn-rift and post-rift sediments and free air anomalies (reproduced from Cloetingh et al. 1995)

and a temperature of ~250–300 °C (Kooi et al. 1992), which is the zone of maximum lithospheric strength (Fig. 5.1). Following this concept, stable cratons with a cold geotherm (Fig. 5.1) and strong sub-crustal mantle pinches near the uppermost mantle. On the other hand, mobile belts with thicker orogenic lithosphere and higher geotherm (Fig. 5.1) would have low strength lithosphere as represented in the strength profiles in Fig. 5.1 and necking depth near the brittle-ductile transition zone (Cloetingh et al. 1995; Fig. 5.2).

Subsequent studies revealed that though the simple concept above is true, considering the strongest layer in the lithosphere as the level of necking is not that straightforward (e.g. van der Beek et al. 1995; Spadini et al. 1995). For example, the lithosphere may consist of two or more strong layers where the brittle-ductile transition zone and mantle are equally strong. Thus, stress can localize at many levels within the lithosphere. Again, crustal composition controls the thickness and sometimes the occurrence of ductile layer in the lower crust. For a felsic composition, a thicker ductile zone persists than that for intermediate or basic composition (Ranalli 2000; Afonso and Ranalli 2004). In such a condition, the 'effective level of necking' will be demarcated by the weakest layer enveloped between the strongest

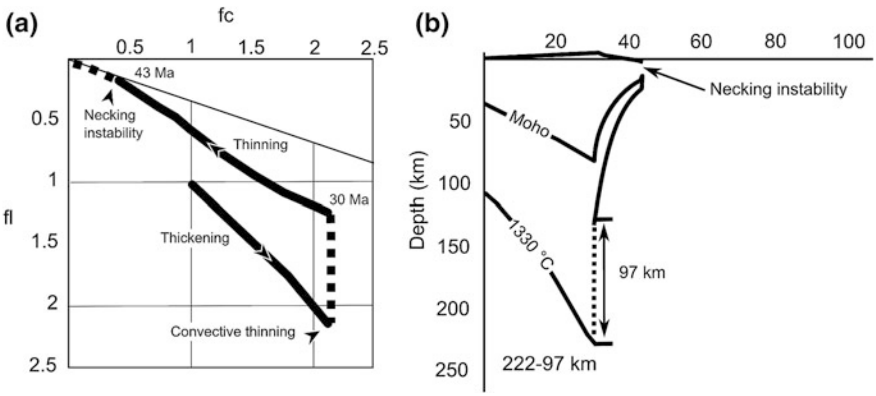


Fig. 5.4 Modeling results show evolution of the lithosphere following homogeneous orogenic thickening (reproduced partly from Rey 2001). Convergence continues for 30 Ma, immediately followed by convective thinning of the mantle. The relative thickening and thinning are represented in the left hand side fc-fl plots; fc: ratio of the final thickness of the crust to that of the reference crust; fl: ratio of the final thickness of the lithosphere to that of the reference lithosphere. Convective thinning is shown for mantle removal of 97 km. Here, necking instability arises and possibly develops a spreading centre

layers (e.g. Spadini et al. 1995; Handy and Burn 2004). Gravitational instabilities may occur in the lithosphere when a large volume (>43 %) of continental mantle of the thick lithosphere in mobile belts drags into the convective asthenospheric mantle (Rey 2001). The thickened crust then collapses gravitationally (Fig. 5.4) and may rift continents e.g. the Basin and Range province, N America (Rey 2001; Montési and Zuber 2003a). Lithospheric layers of varying rheology deform and spread at different rates during orogenic collapse (Fig. 5.4).

In such situations, the pre-existing faults of favouring trends (see Sect. 4.2) reactivate. Such faults are either listric (Fig. 4.2) or discontinuous/segmented. Extrusive doming of the low viscosity middle layer can also form metamorphic core complexes (Kaisa et al. 2013).

Forward modelling revealed pre-rift lithospheric parameters viz. effective elastic thickness (T_e; detail in Watts 1992; Burov and Diament 1995), level of necking, and strain rate (Cloetigh et al. 1995; Pourhiet et al. 2015) control the necking depth and consequently the rift geometry. Figure 5.5 summarises the correlations found from their modelling. The thermal state of the lithosphere governs its pre-rift rheology. The plots (Fig. 5.5) show possible correlation of necking depth with pre-rift crustal/lithospheric thickness and strain rate. The depth of necking is shallow for higher depth (60–80 km) of pre-rift crust but such a correlation is weaker for lithosphere (Cloetingh et al. 1995). Effective elastic thickness (T_e) (see Appendix) does not correlate with the level of necking. T_e can represent rigidity (Ratheesh-Kumar et al. 2014 and references therein).

Note the Saudi Arabian Red Sea margin is an outlier in the plot of effective elastic thickness versus necking depth and it also has a significantly large

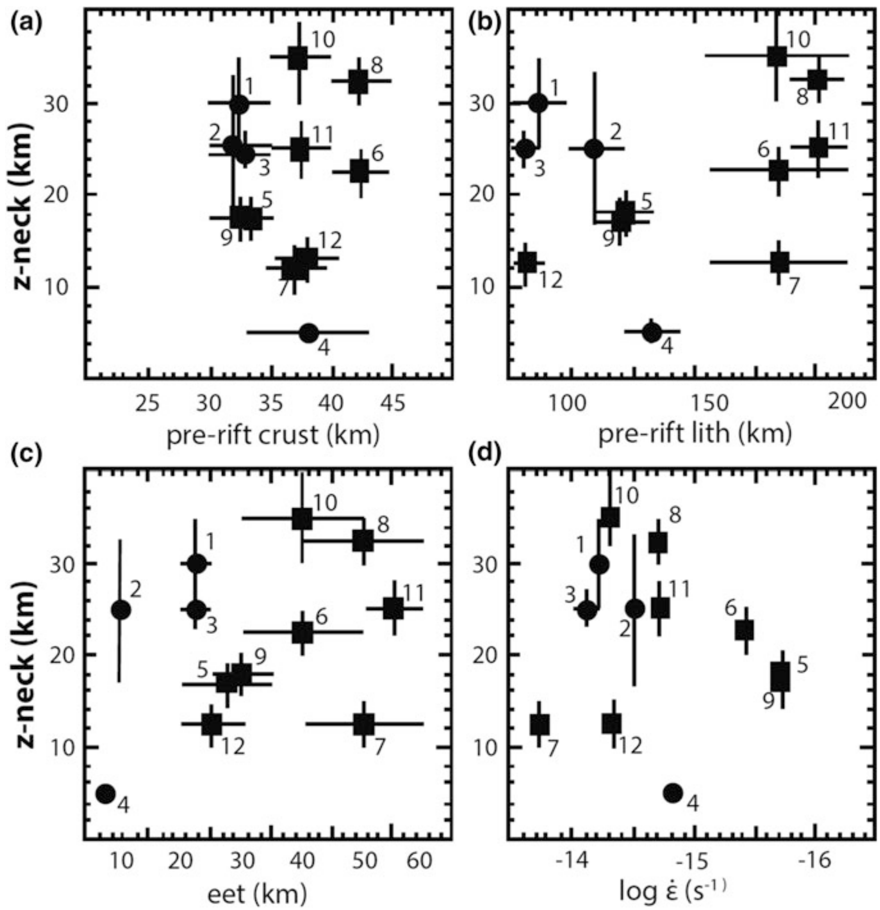


Fig. 5.5 Correlation illustrations for the relationships between different parameters **a** necking depth and pre-rift crustal thickness; **b** necking depth and pre-rift lithosphere thickness; **c** EET and necking depth; and **d** necking depth and strain rate. *Closed dots* and *squares* indicate data from Alpine/Mediterranean basins and intracratonic rifts, respectively. *Numbers in diagrams 1* Gulf of Lion *2* Valencia Trough *3* Southern Tyrrhenian Sea *4* Pannonian Basin *5* North Sea Basin *6* Baikal Rift *7* Saudi Arabia Red Sea Margin *8* Transantarctic Mountains *9* Barents Sea margin *10* East African Rift *11* Western Black Sea *12* Eastern Black Sea (Cloetingh et al. 1995). Note the anomalous position of the Saudi Arabia Red Sea margin (*7*) in these diagrams; $\dot{\varepsilon}$: strain rate (reproduced from Cloetingh et al. 1995)

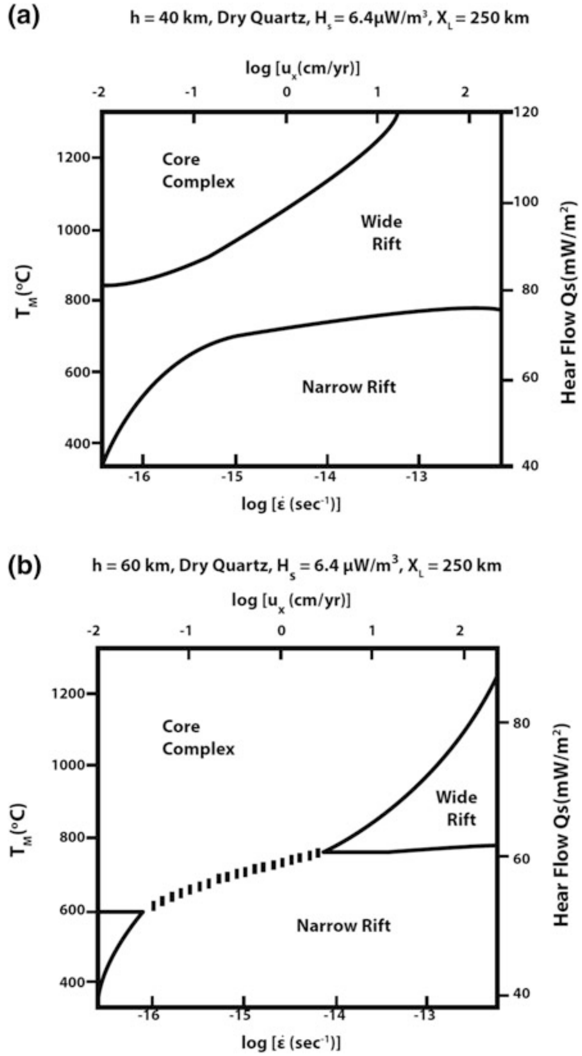
rift-related ~3 km uplift. The Trans-antarctic also shows >5 km rift shoulder uplift (Cloetingh et al. 1995). The anomalous nature of the Saudi Arabian Red Sea margin is a product of other tectonic process like volcanism leading possibly a very fast extension: 5 Ma from rift initiation until breakup (Cloetingh et al. 1995). Prior deformations can either strengthen or weaken the lithosphere. Change in strength may fracture the lithosphere locally (Autin et al. 2013). Most orogens are weaker than adjacent crust due to repeated reactivations (Miller et al. 2001). But, some orogens may be harder than adjacent crust and might not rift e.g. the Urals did not rift during Jurassic to early Paleocene Gondwana breakup (Krabbendam 2001).

5.3 Temperature and Strain Rate

A narrow-, wide- and core complex mode of rifting happens in progressively hot and weak lithosphere (e.g. Buck 1991). Based on numerical modelling, Buck (1991) concluded that width of rifts depends on the thermal state of the lithosphere. He considered wide rifts form in hot lithosphere of 80 mWm^{-2} of surface heat flow and narrow rifts in cold lithosphere of 60 mWm^{-2} heat flow (Fig. 5.6). Wide rift may also form if the weak crust decouples from mantle (Hopper and Buck 1996). For high crustal thickness and very high Moho temperatures, core complexes may form as in the Basin and Range province (Corti et al. 2003 and references therein). Bassi's (1995) dynamic model demonstrated that cold brittle lithosphere thins rapidly under extension to develop narrow rifts. In contrast, warm ductile lithosphere develops wider (>100 km) rifts. Models indicate that pre-rift rheology has a greater control on the geometry of rifted margins than the strain rate (Buck 1991; Bassi 1995). The strength difference between the lower crust and the upper mantle at Moho is best developed in regions with moderate geotherm (i.e. for 700–900 °C at Moho) and decreases with any change in Moho temperature (Ranalli, 2000). A 'normal' geothermal gradient of 22–41 °C km^{-1} and a heat flow of 24–77 mWm^{-2} were reported from south China offshore passive margin region (Liao et al. 2010). The amount is somehow lower than 55–102 mWm^{-2}, which is required normally through mantle plume to initiate rifting (Gliko et al. 1978). Boreholes drilled for hydrocarbon exploration can provide information on temperature and geothermal gradient (as in Liao et al. 2014). Strain rate influences greatly the heat diffusion and strain localisation during extension (Corti et al. 2003 and references therein). Lower the strain rates, wider the rifts. At lower strain rates, deformed crust cools substantially faster and strain localisation front may move laterally towards the position of future breakup widening the rifts. In contrast, for higher strain rates (Fig. 5.1), strain concentrates at specific locations faster and develops narrow rifts (Kuznir and Park 1987).

Combined thermal state of the lithosphere and strain rates manifest differences in rift architecture. Numerical geodynamic models by van Avendonk et al. (2009) demonstrated slow rifting in a cold lithosphere leads to narrow rifts, and sparse melt generates in the end stages of rifting. Fast and hot rifting (Fig. 4.7 of van Avendonk et al. 2009) widens rifts and its late phase characterises profuse melting. Strain localizes slowly and several faults- some parallel to rift axis- activate coevally. In both the fast and slow cases, strain localises along crustal-scale (brittle at top and ductile at bottom) shear zones, which evolve into the main rift faults. However, less stretch is required for complete breakup. Thermal state of the crust varies widely for similar crustal types e.g. the heat production of the

Fig. 5.6 Mode boundaries in Moho temperature (T_M)—strain rate space for dry quartz crustal rheology with crustal heat sources: **a** 40 km thick crust **b** 60 km thick crust; reproduced from Buck (1991)



Australian Proterozoic crust is—twice that of global averages (Taylor and McLennan 1985). The mechanical strength of the lithosphere and thus rift architecture is strongly influenced by the unique thermal state of the crust (e.g. Sandiford and Hand 1998).

Chapter 6
Lessons from Analogue Models

Outcrop analogues of lower crust and mantle in the Alps (e.g. Manatschal 2004; Manatschal and Müntener 2009; Mohn et al. 2011) and Pyrenees (e.g. James et al. 2009; Leleu et al. 2009; James et al. 2010) and deep regional seismic sections reveal lithospheric processes. Nevertheless very few models (e.g. Bellahsen and Daniel 2005; Corti et al. 2007; Chattopadhyay and Chakra 2013) addressed the influence of pervasive fabrics since weak fabrics are difficult to generate on model materials (Morley 1999a). Recently Curren and Bird (2014) tested the influence of pre-existing close-spaced fractures in strike slip faulting. The fractures accommodate larger strains. The width of the Principal Deformation Zone (PDZ) is controlled by the width of the zone of the pre-existing fractures. These experiments form lesser number of long shears than in an isotropic case and the lateral extent of the shears are bound by the pre-existing fractures. The fractures/joints/anisotropies perpendicular to the extension direction undergo dilation or normal faulting. Other orientations experience oblique to strike-slip movements. The base of the sand in those analogue models is usually a rigid plate of acetate, plexiglass or rubber (e.g. Tron and Brun 1991; Aanyu and Koehn 2011; Cappelletti et al. 2013). Unlike these models, the crustal fabrics- pervasive and discrete- lie in the upper crust and not at the base of the crust, and the 'similarity factor' in analogue model has been questioned (e.g. Morley 1999a). In contrary, the discrete fabrics are modelled accurately (e.g. Dooley and McClay 1997; Acocella et al. 1999; reviews by Corti et al. 2003). Moreover, there is a sharp contrast in strength between the modelling sand and the base plate, and also between the rubber sheet and the edge of the base plate, which seems to control the results (Morley 1999a).

Geometry of the fault population for oblique rifts, the obliqueness of extension (Ω) in particular, has been studied in analogue models (e.g. Tron and Brun 1991), but without considering pervasive fabrics. When the angle between the fracture and the maximum principal compressive stress axis (Ω) is 30–90°, faults not affected by pervasive fabrics will tend to be orthogonal to the regional extension direction (Morley 1999a from references therein; Fig. 6.1). For hyper-oblique extensions, where Ω = 0–30°, faults not affected by pervasive fabrics run along the primary

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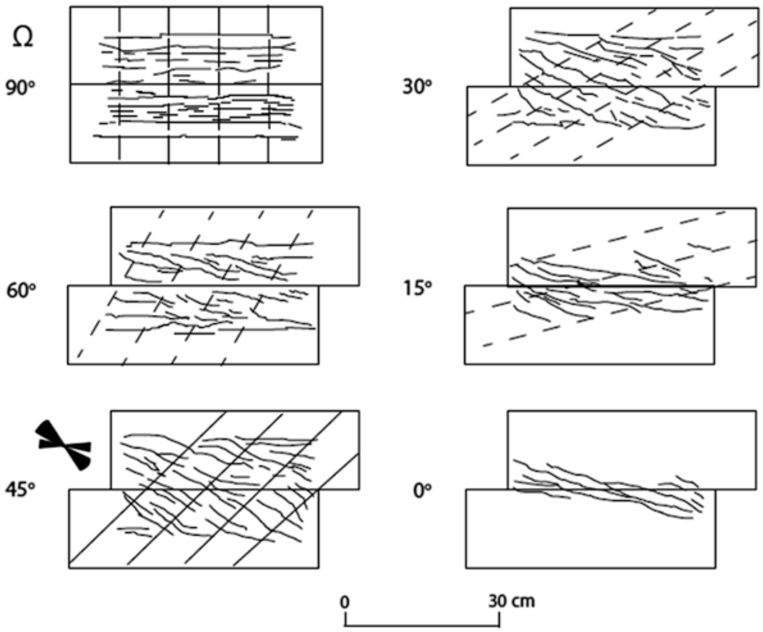

Fig. 6.1 Analogue experiments for oblique rifting (from Tron and Brun 1991). Dotted lines: extension direction, **Ω**: angle between the cut in the base plate and extension direction.

strike-slip orientation. Thus, analogue models showed that the degree of oblique opening as the major control on fault geometry and evolution for oblique extension. Other important factors are the number, spacing, relative strength, attitude and type (pervasive/ discrete) of pre-rift fabrics.

Chapter 7
Summary

We review tectonic inheritance/tectonic heritage/reworking/reactivation in continental rifts and passive margins. Evolution of rifts by segmentation, genesis of transfer zones, changes in geometry and attitude of faults, and concentration of fractures and faults are controlled by inheritance. Depth and orientation of the lineament dictates the inherited faults. Variation in isostatic state and depth of strong layer, criss-crossing geometry etc. are the result. Inheritance acts uniquely but sometimes non-homogeneously on individual passive margins and can reactivate faults in several directions. Existing fractures/fissures and fault zones, metamorphic and sedimentary foliations, shear zones, lineations etc. in the basement can get inherited provided an optimum rheologic contrast exist between the basement and the overlying rocks under low heat flow and an optimum range of angle exists between lineament and the stress direction. Rocks bearing these anisotropies behave under deformation differently from isotropic rocks. In short, two major factors determine whether new trends will develop or pre-existing fabrics will be reactivated—(a) variations in trend between pre-existing fabrics and ideal trend of new fractures with respect to principal stress directions, (b) difference in mechanical strength between pre-existing fabric and surrounding undisturbed country rock volume (e.g. Ranalli and Yin 1990).

The fabric pattern in the cover or overlying bulk rock, inherited from the basement fabric pattern, i.e. deep seated fault trends depends on—(a) kinematic nature of deep seated faults—normal, reverse or strike slip and (b) rheological property of the overlying sequence. If the basement has strike slip faults, Riedel shears (R, R′) and tensile fractures (T) fractures can be formed in the overlying sequence. In case of normal and reverse faults in basement, trends are replicated in the overlying sequence upon reactivation. Rheological property (single/multi layer) of overlying sequence influences the width of the deformation zone within it, overlying the basement fabric.

Inherited planes control fluid flow and metamorphism. A passive margin/rift with inherited structures makes subsequent orogeny heterogeneous. Pattern and site of sedimentation in the overlying foreland basin might be controlled by passive margins with inherited structures. How inheritance controls the magma content in rifts, and what is the link between specific rifts with orogeny is not well understood. Inheritance acts uniquely on individual passive margins. Analogue models suggest

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that inherited structures may parallel pre-existing lineament. However, the rift axis may not follow inherited lineament in the basement, nor is it guaranteed that rift related extension direction will parallel the lineament. Also, the rift basin may not propagate along the lineament. Inherited faults may or may not be regularly spaced.

The Coulomb-Navier's fracture theory holds true for isotropic rocks. Instead, modified Coulomb's failure criterion works for anisotropic rocks and are used more frequently for such cases. Rift parallel faults may terminate against oblique discrete fabrics and assume oblique trends. When the angle (θ_c) between the anisotropic plane and the maximum principal stress ranges 15–45°, fractures parallel the anisotropy. For $\theta_c > 45°$, inheritance seems to be absent. Faults parallel pre-existing anisotropy when the angle between the later and maximum stress direction ranges 13–51°. Degree of anisotropy in the rock depends also on confining pressure. Normal faults form when regional extension direction makes 45–90° with the anisotropy. Reactivation occurs because pre-existing fault zones are generally weaker than the surrounding bulk rocks. Weakness results from (*i*) decrease in cohesion between faulted zone and surrounding rocks, (*ii*) decrease in friction coefficient and (*iii*) high pore pressure within the fault zone. All of these reduce shear strength along pre-existing weak zones and reactivate the faults oriented to the effective principal stress directions. Also, due to recurrent reactivation and thickening, olivine crystals in the lithospheric mantle can develop a lattice-preferred-orientation, which concentrate strain and positions rift zones.

Inheritance can be by structures that are either discrete: large and wide-spaced anisotropies, or pervasive: small and close-spaced anisotropies throughout the rock volume. Rifts and passive margins usually parallel pervasive fabrics. Faults may follow previous anisotropy initially, but later develop an obliquity. Spread of rifts could be decided by spread of anisotropies.

Interaction between normal faults and pre-existing crustal anisotropies such as thrust planes affects the extensional processes. It is generally accepted the pre-existing thrust system influences the geometry of the newly formed normal faults —its location, dip and trend. The interaction has been summarised in three basic situations (*i*) normal faults can crosscut the pre-existing thrust planes without any inheritance, (*ii*) normal fault can branch out from thrust faults at a depth on weak fault zone showing little inheritance and (*iii*) pre-existing thrust system can be fully or partially reactivated resulting ramp and flat geometries by inheritance. Regional rift systems almost always form along mobile belts that are presently inactive earlier orogens. That is because the mobile belts are weaker due to their repeated deformation viz. collision and extension. Graben geometries in the East African Rift System (EARS) inherited from Pan-African orogeny. In this rift system, the four major extensional faults follow the trend of the basement. The first order shape of the rift was pervasive fabric controlled. An E-W extension in the Thailand Tertiary rift zone followed the weaknesses developed before rifting. Attitude of pre-rift bedding planes influenced this rift. Bedding controlled faults terminated propagation of a set of rift related faults. Rifts in SE Brazil and West Africa conjugate passive margins parallel the basement fabric. Like EARS, these rifts follow mobile belts. The rift in the Potiguar basin parallels basement fabric but only regionally. The E Indian passive margin and the major rift faults parallel foliationsof the Eastern Ghat Mobile Belt and the Southern Granulite Terrain. The W Indian passive margin parallels foliation of the Western Dharwar gneisses. The contact between the Dharwar craton and the Closepet Granite also follows the passive margin.

Discrete fabrics do not control the trend of all the rift faults. These fabrics can segment rifts, preclude fault propagation, reactivate if they are weaker than the rock volume, control secondary faulting and location of depocenter, and their dips can decide the nature of faults. The width of the transfer zone depends on the angle between the extension direction and the anisotropy. Discrete fabrics affected oblique/ sheared segments and grabens of the East African Rift System.

Pre-existing fault zones controlled the later Brazilian rifts. These rifts were stopped propagating by two bounding shear zones. A suture zone acted as a discrete fabric for the rifts in Thailand. Basins related to these rifts vary geometrically because of the discrete fabric. Deep crustal shear zones in the Norwegian margin reactivated shallower reverse faults as normal faults. Reverse faults in the west Norwegian margin reactivated as normal faults. Faults reactivated in North Sea and also formed a new fault set. A bend and a fracture zone in the eastern North America inherited as transform structures. Reoriented stress directions reactivated faults in the Upper Rhine Graben. Transverse faults in the east Indian continental margin stopped propagation of rift faults. The Mahanadi rift is possibly bound to the N by a reactivated shear zone. The E-W trending older shear zones in south India probably had no role in West Indian passive margin. The topography of rift shoulder depends on the level of necking. Faults within multiple lithology reactivates more easily.

When deformation occurs at continental scale and affects the entire lithosphere, it is unlikely to have a crustal origin of structural inheritance. Rheological model suggests that it is the mantle rheology that influences and controls the mechanical behaviour of lithosphere. Mechanical anisotropy of lithospheric mantle may develop from—(*i*) plastic anisotropy between single olivine crystal and aggregates, observed in experiments and numerical simulations, (*ii*) strong crystallographic orientation of olivine and some other minerals in mantle xenoliths and (*iii*) seismic anisotropy (shear wave splitting, P-residuals etc.) and electrical conductivity anisotropy. Mantle tectonic fabrics developed during orogeny are probably retained for longer time periods if lithosphere is not strongly modified by deformation or mantle upwelling etc. Deformation can make the lithospheric part weaker. Warmer and ductile lithosphere develops wider rifts under lower strain rate. Mantle plume can localise rift for a heat flow of 55–102 m Wm^{-2}.

Tectonic inheritance in rift has significant impact in hydrocarbon exploration (Mukherjee 2015b). Inheritance has bearing on trap definitions, sediment fairways, localising basins, fracture delineations, fracture predictions etc. Inherited faults show departure from usual trends and understanding the nature of inheritance will allow a better analysis of the trends. In areas of unfit data, knowledge of tectonic inheritance permits better appreciation of the structures. For naturally fractured reservoirs, acquaintance with tectonic inheritance helps delineate sweet spots with higher concentration of fractures and also gives an idea on the trends of the fractures.

Appendix

1. **Effective Elastic Thickness (T_e)**: T_e (elastic) = $(M_{elastic}\ 12(1 - \nu^2)/EK)^{1/3}$ A1 (Watts and Burov 2000)
   Here $M_{elastic}$: bending moment; ν: Poissons's ratio; E: Young's modulus; K: curvature. $M_{elastic}$ and K together indicate the total flexure of the lithosphere. T_e, the effective elastic thickness, is considered to be the depth of a specific isotherm: most commonly, 450–600 °C.
   Low geothermal gradient explains high values of T_e of 70–90 km for cratons. Low values of T_e up to ~40 km usually denote either a young oceanic- or a thinned (rifted) continental crust for higher geothermal gradients (Ratheesh-Kumar et al. 2014).
2. **Lithospheric Strength Profiles**: Lithospheric strength varies with depth and is controlled by the temperature distribution within the lithosphere and the mineralogy (such as Burov and Diament 1995; Burov et al. 1998) (Fig. A.1).
3. **Thermal Age**: It is the last thermal event e.g. orogeny, metamorphism etc. the lithosphere underwent (Rudnick et al. 1998).

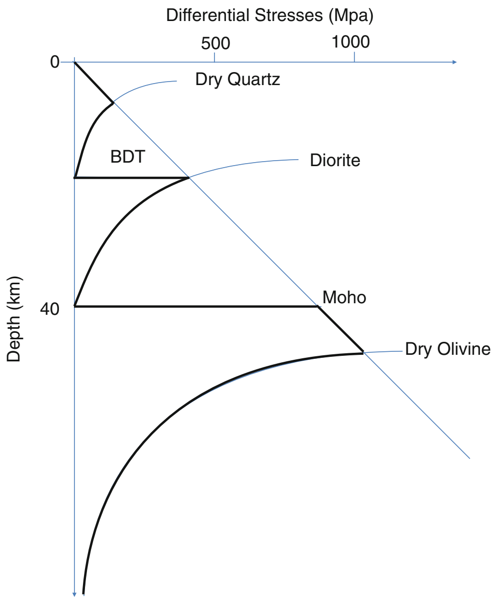

Fig. A.1 Schematic illustration of lithospheric strength profile and the various lithologies

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