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Dynamic processes controlling evolution of rifted basins

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Abstract

The extension of the lithosphere, controlling the development of rifted basins, is driven by a combination of plate-boundary forces, frictional forces exerted on the base of the lithosphere by the convecting asthenosphere and deviatoric tensional stresses developing over upwelling branches of the asthenospheric convection system. Although mantle plumes are not a primary driving force of rifting, they play an important secondary role by weakening the lithosphere and by controlling the level of rift-related volcanic activity. A distinction between "active" and "passive" rifting is only conditionally justified.

The extension of the lithosphere, depending on its rate and magnitude, and the potential temperature of the asthenosphere, can cause by adiabatic decompression partial melting of the lower lithosphere and upper asthenosphere. In rift systems, the level and timing of volcanic activity is highly variable. The lack of volcanic activity implies "passive" rifting. An initial "passive" rifting stage can be followed by a more "active" one during which magmatism plays an increasingly important role. Magmatic destabilization of the Moho may account for the frequently observed discrepancy between upper and lower crustal extension factors. Combined with evidence for thermal thinning of the mantle–lithosphere, this suggest that the volume of the lithosphere is not necessarily preserved during rifting as advocated by conventional stretching models.

The structural style of rifts is controlled by the rheological structure of the lithosphere, the availability of crustal discontinuities that can be tensionally reactivated, the mode (orthogonal or oblique) and amount of extension, and the lithological composition of pre- and syn-rift sediments. Simple-shear extension prevails in rifts that subparallel the structural grain of the basement. Pure-shear extension is typical for rifts cross-cutting the basement grain. Pre-existing crustal and mantle–lithospheric discontinuities contribute to the localization of rift systems.

The duration of the rifting stage of extensional basins is highly variable. Stress field changes can cause abrupt termination of rifting. In major rift systems, progressive strain concentration on the zone of future crustal separation entails abandonment of lateral rifts. Depending on constraints on lateral block movements, crustal separation can be achieved after as little as 9 My and as much as 280 My of rifting activity.

Syn-rift basin subsidence is controlled by isostatic adjustment of the crust to mechanical stretching of the lithosphere, its magmatic inflation and thermal attenuation of the mantle–lithosphere.

Post-rift basin subsidence is governed by thermal reequilibration of the lithosphere–asthenosphere system. Deep-seated thermal anomalies related to syn-rift pull-up of the asthenosphere–lithosphere boundary have decayed after 60 My by about 65% and after 180 My by about 95%. The magnitude of post-rift subsidence is a function of the rift-induced thermal anomaly and crustal density changes, the potential temperature of the asthenosphere and initial water depths. Intraplate

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stresses can have an overprinting effect on post-rift subsidence. Stretching factors derived from post-rift subsidence analyses must be corrected for such effects.

Keywords: rifting; post-rift; lithosphere; crust; rheology; magmatism; stress; strain; extension; pure-shear; simple-shear; subsidence

1. Introduction

Tectonically active rifts, palaeo-rifts and passive-margins form a group of genetically related extensional basins that play an important role in the spectrum of sedimentary basin types (Bally and Snelson, 1980). Extensional basins cover large areas of the globe and contain important mineral deposits and energy resources. A large number of major hydrocarbon provinces are associated with rifts (e.g. North Sea, Syrt and West Siberian basins, Dniepr–Donets and Gulf of Suez grabens) and passive margins (e.g. Campos Basin, Gabon and Angola shelves, Niger and Mississippi deltas; Ziegler, 1996a). On these basins, the Petroleum Industry has acquired large databases that document their structural styles and allow detailed reconstruction of their evolution. Academic geophysical research programs have provided information on the crustal and lithospheric configuration of tectonically active rifts, palaeo-rifts and passive margins. Petrologic and geochemical studies have advanced the understanding of rift-related magmatic processes. Numerical models, based on geophysical and geological data, have contributed at lithospheric and crustal scales towards the understanding of dynamic processes that govern the evolution of rifted basins. This paper summarizes the state of knowledge on dynamic processes that control the evolution of extensional basins.

A natural distinction can be made between tectonically active and inactive rifts and rifts that evolved in continental and oceanic lithosphere. Tectonically active intracontinental (intraplate) rifts, such as the Rhine Graben, the East African Rift, the Baikal Rift and the Shanxi Rift of China, correspond to important earthquake and volcanic hazard zones. The globe-encircling mid-ocean ridge system forms an immense intraoceanic active rift system that encroaches in the Red Sea and the Gulf of California onto continents. Rifts that are tectonically no longer active are referred to as palaeo-rifts, aulacogens, inactive or aborted rifts and failed arms, in the sense that they did not proceed to crustal separation. Conversely, the evolution of successful rifts culminated in the breakup of continents, the opening of new oceanic basins and the development of conjugate pairs of passive margins.

In the past, a genetic distinction was made between "active" and "passive" rifting (Sengör and Burke, 1978; Wilson, 1989; Olsen and Morgan, 1995). "Active" rifts are thought to evolve in response to thermal upwelling of the asthenosphere (Dewey and Burke, 1975; Bott and Kusznir, 1979; Spohn and Schubert, 1982), whereas "passive" rifts develop in response to lithospheric extension driven by far-field stresses (McKenzie, 1978; McKenzie and Bickle, 1988; Khain, 1992). It is, however, questionable whether such a distinction is justified as the study of Phanerozoic rifts revealed that rift-related volcanic activity and doming of rift zones is basically a consequence of lithospheric extension and is not the main driving force of rifting. The fact that rifts can become tectonically inactive at all stages of their evolution, even if they have progressed to the Red Sea stage of limited sea-floor spreading, supports this concept. However, as extrusion of large volumes of rift-related sub-alkaline tholeiites must be related to a thermal anomaly within the upper mantle, a distinction between "active" and "passive" rifting is to a certain degree still valid, though not as "black and white" as originally envisaged.

Rifting activity, resulting in the breakup of continents, is probably governed by forces controlling the movement and interaction of lithospheric plates. These forces include plate boundary stresses, such as slab pull, slab roll-back, ridge push and collisional resistance, and frictional forces exerted by the convecting mantle on the base of the lithosphere (Forsyth and Uyeda, 1975; Bott and Kusznir, 1979; Bott, 1982, 1993; Ziegler, 1993).

On the other hand, deviatoric tensional stresses, inherent to the thickened lithosphere of young orogenic belts, as well as those developing in the lithosphere above upwelling mantle convection cells and mantle plumes (Bott, 1993) do not appear to cause, on their own, the breakup of continents. However, if such stresses interfere constructively with plate boundary and/or mantle drag stresses, the yield strength of the lithosphere may be exceeded, thus inducing rifting (Fig. 1).

It must be understood that mantle drag forces are exerted on the base of a lithospheric plate if its velocity and direction of movement differs from the velocity and direction of the mantle flow. Mantle drag can constructively or destructively interfere with plate boundary forces, and thus can either contribute towards plate motion or resist it. Correspondingly, mantle drag can give rise to the build-up of extensional as well as compressional intraplate stresses (Forsyth and Uyeda, 1975; Bott, 1982, 1993; Sabadini et al., 1992; Artemieva and Mooney, 2002). Although the present lithospheric stress field can be readily explained in terms of plate boundary forces (Cloetingh and Wortel, 1986; Richardson, 1992; Zoback, 1992), mantle drag probably contributed significantly to the Triassic–Early Cretaceous breakup of Pangea, during which Africa remained nearly stationary and straddled an evolving upwelling and radially outflowing mantle convection cell (Pavoni, 1993; Cadec et al., 1995; Ziegler, 1993; Ziegler et al., 2001).

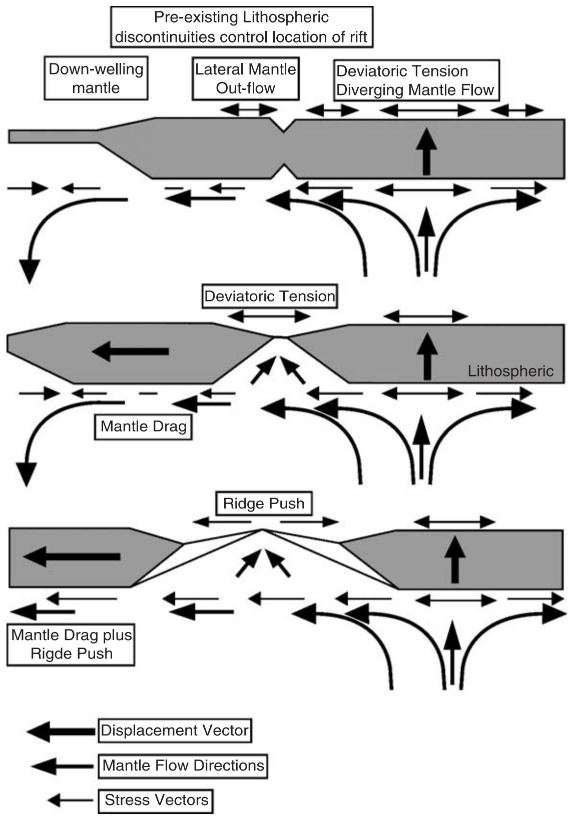

Fig. 1. Diagram illustrating the interaction of shear-traction exerted on the base of the lithosphere by asthenospheric flow, deviatoric tension above upwelling mantle convection cells and ridge push forces (Ziegler et al., 2001).

Mechanical stretching of the lithosphere and thermal attenuation of the mantle–lithosphere is associated with the development of local deviatoric tensional stresses, which play an increasingly important role during advanced rifting stages (Bott, 1992a; Ziegler, 1993). This has led to the development of the concept that many rifts go through an evolutionary cycle starting with an initial "passive" phase which is followed by a more "active" stage during which magmatic processes play an increasingly important role (Wilson, 1989, 1993b; Khain, 1992; Burov and Cloetingh, 1997; Huismans et al., 2001). However, nonvolcanic rifts must be considered as purely "passive" rifts.

2. Megatectonic setting of rifts

Since Archean times, the Earth's history was characterized by several more or less discrete orogenic and rifting cycles during which successive Pangea-type supercontinents were assembled and subsequently disrupted (Salop, 1983; Khain, 1992, Milanovsky, 1992). Through time, rifted basins evolved under a wide variety of megatectonic settings, a summary of which is given below.

2.1. Atlantic-type rifts

Atlantic-type rift systems evolve during the breakup of major continental masses, presumably in conjunction with a reorganisation of the mantle convection system (Ziegler, 1993). During early phases of rifting, large areas around future zones of crustal separation can be affected by tensional stresses, giving rise to the development of complex graben systems. In time, rifting activity concentrates on the zone of future crustal separation, with tectonic activity decreasing and ultimately ceasing in lateral graben systems. As a consequence of progressive lithospheric attenuation and ensuing crustal doming, local deviatoric tensional stresses play an important secondary role in the evolution of such rift systems. Upon crustal separation, the diverging continental margins (peri-continental rifts) and the "unsuccessful" intracontinental branches of the respective rift system become tectonically inactive. However, during subsequent tectonic cycles, such aborted rifts can be tensionally as well as compressionally reactivated (Ziegler et al., 1995, 1998, 2001, 2002). Development of Atlantic-type rifts is subject to great variations mainly in terms of duration of their rifting stage and the level of volcanic activity (Ziegler, 1988, 1990, 1996b).

2.2. Back-arc rifts

Back-arc rifts are thought to evolve in response to a decrease in convergence rates and/or even a temporary divergence of colliding plates, ensuing steepening of the subduction slab and development of a secondary upwelling system in the upper plate mantle wedge above the subducted lower plate lithospheric slab (Uyeda and McCabe, 1983; Tamaki and Honza, 1991; Honza, 1993). Changes in convergence rates between colliding plates are probably an expression of changes in plate interaction. Back-arc rifting can progress to crustal separation and the opening of limited oceanic basins (e.g. Sea of Japan, South China Sea, Black Sea). However, as convergence rates of colliding plates are variable in time, back-arc extensional basins are generally short-lived. Upon a renewed increase in convergence rates, back-arc extensional systems are prone to destruction by back-arc compressional stresses (e.g. Variscan geosyncline, Sunda Arc and East China rift systems, Black Sea domain; Uyeda and McCabe, 1983; Cloetingh et al., 1989; Jolivet et al., 1989; Ziegler, 1990; Letouzey et al., 1991; Nikishin et al., 2001).

2.3. Syn-orogenic rifting and wrenching

Syn-orogenic rift/wrench deformations can be related to indenter effects and ensuing escape tectonics, often involving rotation of intramontane stable blocks (e.g. Pannonian Basin: Royden and Horváth, 1988; Late Carboniferous Variscan fold belt: Ziegler, 1990), as well as to lithospheric overthickening in orogenic belts, resulting in uplift and extension of their axial parts (Peruvian and Bolivian Altiplano: Dalmayrac and Molnar, 1981; Mercier et al., 1992). Furthermore, collisional stresses exerted on a craton may cause far-field tensional or transtensional reactivation of pre-existing fracture systems and thus the development of rifts and pull-apart basins. This model may apply to the Late Carboniferous development of the Norwegian–Greenland Sea rift (Ziegler, 1989a; Ziegler et al., 1995), the Permo-Carboniferous Karoo rifts (Visser and Praekelt, 1998) and the Neogene Baikal rift (Molnar and Tapponnier, 1975). Under special conditions, extensional structures can also develop in forearc basins (e.g. Talara Basin, Peru: Moberly et al., 1982).

2.4. Post-orogenic extension

Extensional disruption of young orogenic belts, involving the development of grabens and pull-apart structures, can be related to their post-orogenic uplift and the development of deviatoric tensional stresses inherent to orogenically overthickened crust (Stockmal et al., 1986; Dewey, 1988, Sanders et al., 1999). The following mechanisms contribute to post-orogenic uplift: (1) locking of the subduction zone due to decay of the regional compressional stress field (Whittaker et al., 1992); (2) roll-back and ultimately detachment of the subducted slab from the lithosphere (Fleitout and Froidevaux, 1982; Bott, 1990, 1993; Andeweg and Cloetingh, 1998); and (3) retrograde metamorphism of the crustal roots, involving in the presence of fluids the transformation of eclogite to less dense granulite (Le Pichon et al., 1997; Bousquet et al., 1997; Straume and Austrheim, 1999). An example is the Permo-Triassic development of the West Siberian Basin (Rudkewich, 1976; Nikishin et al., 2002). Modifications in the convergence direction of colliding continents, causing an important stress reorientation, can give rise to the development of wrench fault systems and related pull-apart basins, controlling the collapse of the an orogen (e.g. Devonian development of the Arctic–North Atlantic Caledonides and Stephanian–Autunian evolution of the Variscan fold belt: Ziegler, 1989a, 1990; Ziegler et al., 2004).

The Basin and Range Province of North America is a special type of post-orogenic rifting. Oligocene and younger collapse of the U.S. Cordillera is thought to be an effect of the North American craton having overridden at about 28 Ma the East Pacific Rise in conjunction with rapid opening of the Atlantic Ocean (Verall, 1989). In the area of the southwestern U.S. Cordillera, regional compression waned during the late Eocene and the orogen began to collapse during late Oligocene with main extension occurring during the Miocene and Pliocene (Keith and Wilt, 1985; Parsons, 1995). By contrast, the Canadian Cordillera remained intact. During the collapse of the U.S. Cordillera, the heavily intruded, at middle and lower levels ductile crust of the Basin and Range Province was subjected to major extension at high strain rates, resulting in uplift of ductilely deformed core complexes by 10–20 km. The area affected by extension, crustal thinning, volcanism and uplift measures 1500 × 1500 km (Wernicke et al., 1987; Coney, 1987; Wernicke, 1990). The Eo-Oligocene magmatism of the Basin and Range Province bears a subduction-related signature, suggestive of an initial phase of back-arc extension, whereas mantle–lithosphere- and asthenosphere-derived magmas play an increasingly important role from Miocene times onwards, presumably due to the opening of asthenospheric windows as the Farralon slab was detached from the lithosphere and sank into the mantle (Keith, 1986; Jones et al., 1992; Parsons, 1995).

2.5. Mantle plumes and hot spots

The notion of "active" rifting is based on the association of some rift systems with major magmatic provinces thought to be relate to mantle plumes, also referred to as hot spots (Dewey and Burke, 1975; Bott and Kusznir, 1979; Spohn and Schubert, 1982; Coffin and Eldholm, 1994). Mantle plumes rise diapirically from the core–mantle boundary through the lower mantle and, upon reaching a density equilibrium, spread out variably at the 670- and 410-km discontinuities. Plume heads spreading out at these discontinuities act as major heat sources and can trigger partial melting of the upper mantle and upwelling of a system of secondary plumes, as evidenced by the tomographic image of the Iceland plume (Bijwaard and Spakman, 1999; Brunet and Yuen, 2000; Nikishin et al., 2002). Partial melts rising through these secondary plumes spread out at the base of the lithosphere, interact with it and ascend into zones of pre-existing lithospheric thinning. Resulting flood basalt provinces can have a radius of 1000–2000 km (Ziegler, 1988, 1990, White, 1992; Wilson, 1989, 1992, 1993a, 1997; Nikishin et al., 2002).

Intraoceanic hot spots, such as Hawaii, St. Helena and Tristan da Cunha, and intracontinental ones, such as the Cenozoic Hoggar and Tibesti volcanic centres (Burke and Whiteman, 1973; Le Bas, 1987; Wilson and Guiraud, 1992, 1998), are not associated with major extensional faulting. Therefore, hot spot activity is unlikely to cause on its own the development of major rifts and the splitting apart of continents. However, mantle plumes impinging on the base of the lithosphere cause its thermal weakening, regional uplift and the development of deviatoric tensional stresses (Wilson, 1989, 1993b; Bott, 1992a). Moreover, the radially out-flowing plume material may enhance mantle drag forces. Although hot spots have contributed in this fashion to the opening of the Central and South Atlantic, the Norwegian–Greenland Sea and parts of the Indian Ocean (White and McKenzie, 1989; O'Connor and Duncan, 1990; Wilson and Guiraud, 1992; Gladczenko et al., 1997), they do not appear to exert a basic control on rifting (Wilson, 1993b, 1997; Nikishin et al., 2002; Ziegler et al., 2001).

In this context, it should be noted that we can distinguish between long-lived and short-lived plumes. Long-lived plumes, such as the Tristan da Cunha, St. Helena, Iceland and Hawaii plumes, can remain active for more than 140 My. On the other hand, short-lived plumes, such as the Siberian, Emeishan and Central Atlantic plumes, can remain active for no more than 1–10 My. Impingement of plumes on zones of crustal extension can variably occur almost at the same time as rifting commences (South Atlantic, Labrador Sea), after 15 My of rifting (Central Atlantic), some 70 My after the onset of rifting (Indian Ocean) or after as much as 280 My of intermittent crustal extension (Norwegian–Greenland Sea). Moreover, plume activity can terminate 10–15 My before crustal separation has been achieved (e.g. Central Atlantic, Karoo; Ziegler et al., 2001; Nikishin et al., 2002).

3. Rifting and magmatism

The intensity and timing of volcanic activity in rifts is highly variable (Fig. 2). Many rifts are totally devoid of volcanic rocks or show only a very low level of volcanism. Examples are the Triassic rifts of Western and Central Europe (Ziegler, 1990) and the Early Cretaceous Central African rifts (Genik, 1992; McHargue et al., 1992; Wilson and Guiraud, 1992, 1998). Such rifts are generally not associated with progressive large-radius crustal doming. Other rifts display a high level of volcanic activity, sometimes shortly after the onset of crustal extension (e.g. Mid-continent rift: Cannon, 1992; Allen et al., 1995; Oslo Graben: Sundvoll et al., 1990; Neumann et al., 1995; Red Sea–East African rift system: Almond, 1986; Mohr, 1992; Braile et al., 1995; South Atlantic rift: Chang et al., 1992; Cenozoic rift system of Europe: Ziegler, 1990; Prodehl et al., 1995). Rifts can also become temporarily volcanic after an initial stage of nonvolcanic subsidence (e.g. North Sea Central Graben: Ziegler, 1990). Volcanic activity is often associated with more or less symmetrical doming of the rift zone. Wrench-induced pull-apart basins and oblique-slip rift zones often display a relatively high level of volcanic activity (e.g. Triassic grabens of North Africa: Manspeizer, 1982; Laville and Petit, 1984; Dead Sea wrench system: Fediuk and Al Fugha, 1999; Garfunkel and Ben-Avraham, 2000). This suggests that major wrench faults transect the entire lithosphere, thus providing conduits for magma migration to the surface (Wilson and Guiraud, 1992).

Major variations in the intensity and timing of volcanic activity are also evident during the rifting stage of passive margins (Fig. 3). For instance, the early rifting stage of the Labrador Sea was accompanied by a high level of volcanic activity, whilst the Arctic–North Atlantic rift remained nonvolcanic during its Late Carboniferous to Cretaceous evolution until the Iceland plume was activated during the latest Cretaceous–Palaeocene, giving rise to the Thulean volcanic surge which immediately preceded and ac-

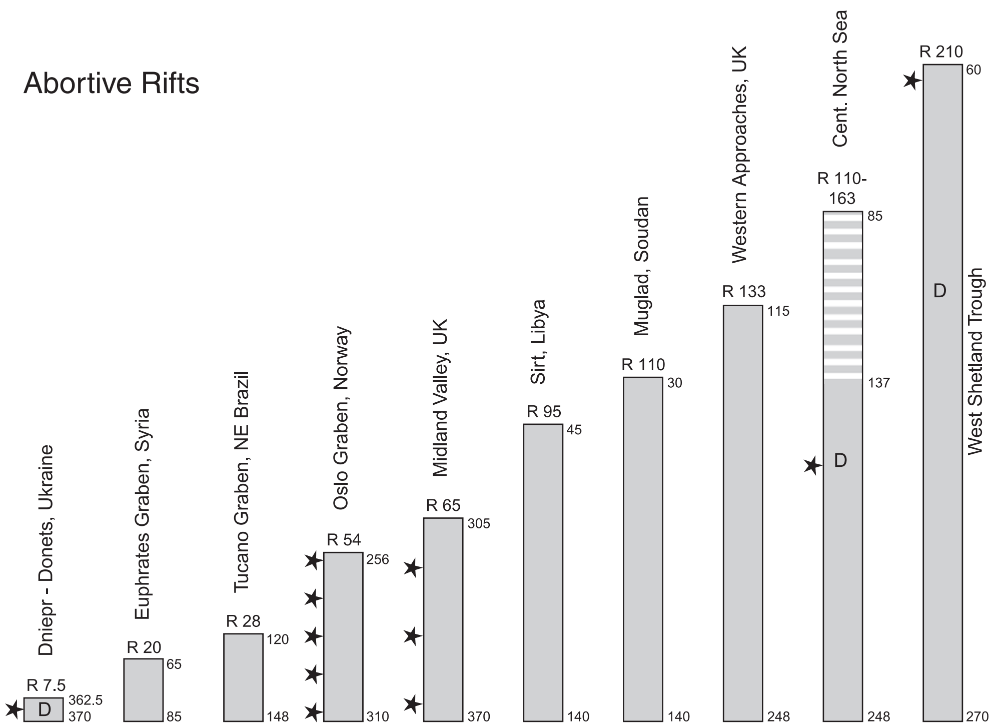

Fig. 2. Duration of rifting stage of "abortive" rifts (palaeo-rifts, failed arms). Vertical columns in My; numbers on side of vertical columns indicate onset and termination of rifting stage in My; numbers under R on top of each column give the duration of rifting stage in My; stars indicate periods of main volcanic activity; D indicates periods of doming.

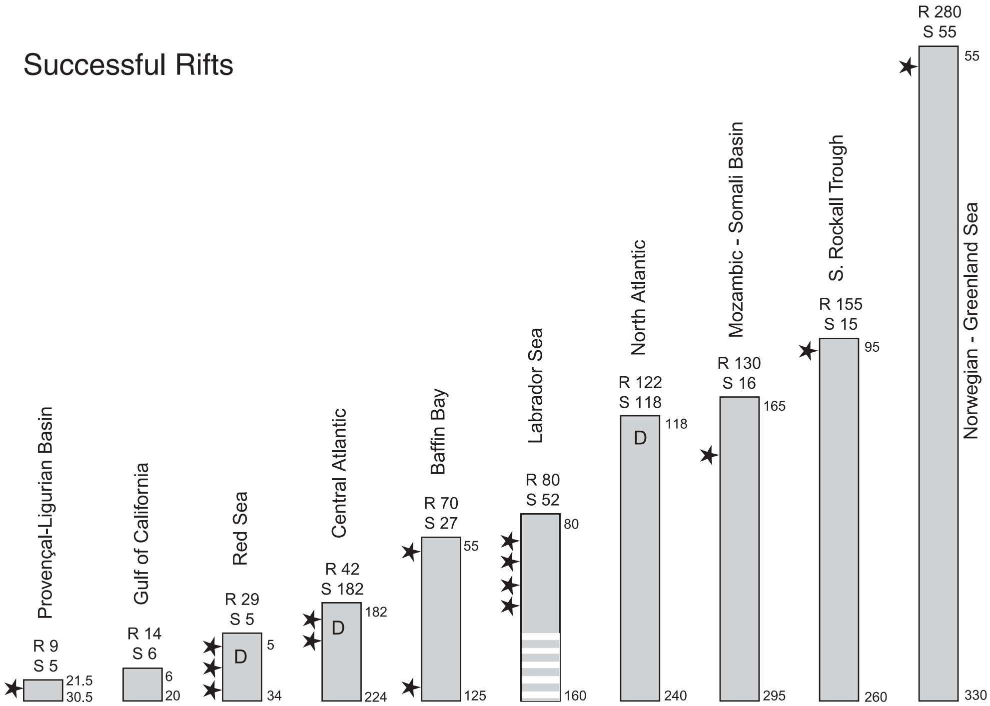

Fig. 3. Duration of rifting stage of "successful" rifts. Legend same as Fig. 1. Numbers beside letter S indicate duration of sea-floor spreading stage in My.

companied crustal separation in the Norwegian–Greenland Sea (Morton and Parson, 1988; White, 1989; Ziegler, 1988, 1990). Similarly, the end Early Jurassic crustal separation in the Central Atlantic was preceded by a short pulse of flood-basalt extrusion (Manspeizer, 1988; Bertrand, 1991; Wilson, 1997; Marzoli et al., 1999), whereas Mid-Cretaceous opening of the southern South Atlantic was preceded and accompanied by the extrusion of the Etendeka and Paraná flood basalts (Wilson, 1992; Gladczenko et al., 1997). However, the development of other passive margins was not accompanied by major syn-rift volcanism (e.g. largely nonvolcanic opening of Canada Basin and North Atlantic: Ziegler, 1988; Sibuet, 1992; development of Australian Northwest Shelf: Williamson et al., 1990; opening of Bay of Biscay and Alpine Tethys: Ziegler et al., 2001).

3.1. Magma source

Volcanic rocks associated with intracontinental rifts display a typically alkaline, mafic–felsic bimodal composition (Martin and Piwinsky, 1972; Burke and Dewey, 1973; Le Bas, 1980; Lameyre et al., 1984; Fitton and Upton, 1987; Wilson, 1989). Mafic melts appear to be generally derived from an incompatible element-enriched mantle source, residing presumably in the mantle–lithosphere, the depleted asthenosphere and/or within mantle plumes. Initial magma generation in intracontinental rifts generally occurs in the 100–200 km depth range, corresponding to the lower parts of the lithosphere and the upper asthenosphere (Wilson, 1989). During the evolution of some rifts, a decrease in alkalinity of the extruded mafic magmas and an increasing contribution of MORB-source melts (depleted mantle) can be recognized, both in time and generally towards the rift axis. This can be attributed to an increasing contribution of melts from the asthenosphere as the lithosphere is progressively thinned (Mohr, 1982; Wendlandt and Morgan, 1982; Morgan, 1983; Lameyre et al., 1984; Baldridge et al., 1991; Wilson, 1993b). Melt contributions from deep, more primitive mantle sources and/or the boundary layer between the upper and lower mantle, typical for mantle plumes (Christensen, 1989; Davies et al., 1989; Loper, 1991), appear to be lacking in many rift-related volcanic suites (Wilson and Downes, 1992), though traces of earlier hot-spot magmatism are sometimes observed (Wilson and Guiraud, 1992). This raises serious doubts about the general applicability of the deep mantle plume-driven "active" rifting model (Spohn and Schubert, 1982; Yuen and Fleitout, 1985) to many examples of intraplate rifting. However, plume-related flood basalt provinces are characterized by a wide range of geochemical and isotopic signatures reflecting mixing of partial melts derived from the upper mantle, the mantle–lithosphere and the original plume material (Smith and Lewis, 1999). At Mid-Ocean ridges, major magma generation occurs at the depth range of 30–40 km, but the onset of partial melting may reach down to depths of 60–80 km beneath normal ridge segments (Wilson, 1989; Latin and Waters, 1991, 1992). Thus, sea-floor spreading axes probably represent zones in which the asthenosphere wells up passively into the space opening between diverging plates (Fig. 1; Anderson et al., 1992; Pavoni, 1993; Ziegler, 1993). The amount of melt generated during rifting in intraoceanic as well as intracontinental domains depends to a large extent on the potential temperature of the asthenosphere with enhanced melt production reflecting above ambient temperatures (Wilson, 1993a,b). Such domains correspond either to areas where deep mantle plumes have impinged on the lithosphere or which are underlain by an upwelling branch of the deep mantle convection systems, possibly activating less vigorous plumes rising from the 670-km discontinuity (e.g. Africa: Cadec et al., 1995; Nikishin et al., 2002).

Surface manifestations of rift-related magmatism represent only a fraction of the total volume of melt generated during rifting (White et al., 1987). As rift-related magmatic processes also affect the mantle–lithosphere and the lower crust, they have a profound effect on the evolution of rift zones during their tectonically active, as well as their post-rifting stage (van Wijk et al., 2001).

Rift-related mechanical stretching of the lithosphere (Figs. 4 and 5) can cause, by adiabatic decompression of the lower lithosphere and upper asthenosphere, their partial melting and the diapiric rise of melts along fractures in the zone of lithospheric thinning. The volume and composition of melts gen-

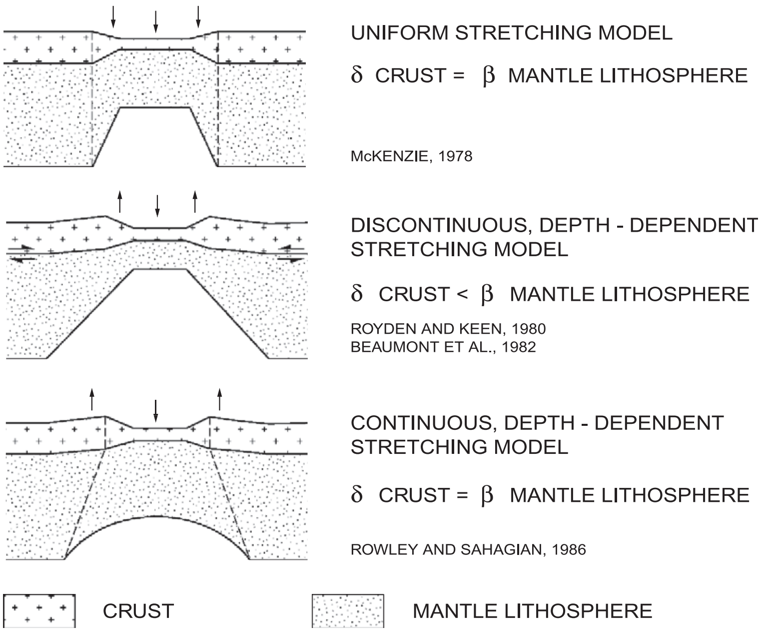

Fig. 4. Lithospheric stretching models.

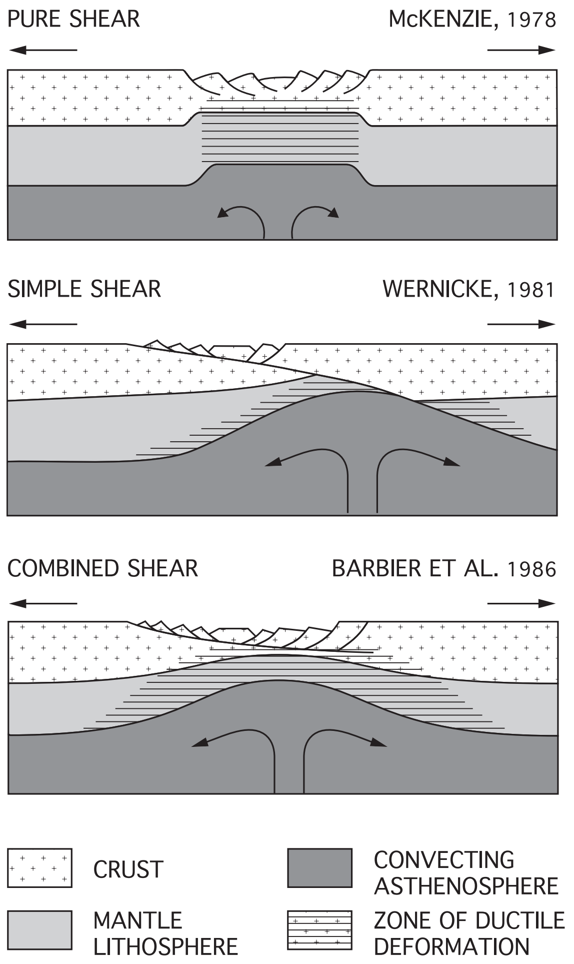

Fig. 5. Lithospheric shear models.

erated is a function of the amount of lithospheric extension, the thermal state of the asthenosphere and lithosphere at the onset of extension, the presence of volatiles and the thickness of the lithosphere (Spohn and Schubert, 1983; Neugebauer, 1983; McKenzie and Bickle, 1988; White and McKenzie, 1989; Wilson, 1989, 1993a; Latin et al., 1990; Gallagher and Hawkesworth, 1992).

Clearly, lithospheric stretching factors play an important role by controlling the degree of adiabatic decompression of the lower lithosphere and the asthenosphere and the upwelling of the latter. Partial melting occurs when the upwelling material crosses the mantle solidus line, the position of which in $P-T$ space is a function of its composition (McKenzie and Bickle, 1988; Wilson, 1993b). In this respect, it must be realized that stress-induced extension of the lithosphere is likely to cause in areas underlain by anomalously hot asthenosphere (upwelling, plume-input) a greater degree of partial melting (e.g. East African rift: Braile et al., 1995; South Atlantic rift: Wilson, 1992; Midcontinent rift: Allen et al., 1995; Dniepr–Donets rift: Wilson and Lyashkevich, 1996; Baikal rift: Keller et al., 1995; Oslo graben: Neumann et al., 1995) than in areas underlain by an asthenosphere characterized by ambient or even below ambient (down welling) temperatures (e.g. Late Palaeozoic and Mesozoic Norwegian–Greenland Sea rift: Ziegler, 1988). Moreover, if very thick (150 km), cold lithosphere is extended, little magmatism can be expected unless a very high degree of extension has occurred. Under such conditions, melts generated are expected to be derived mainly by partial melting from the lower lithosphere (e.g. Hawkesworth et al., 1992). Conversely, stretching of lithosphere having a thickness of some 100 km by a factor of 1.2–1.3 can already result in the generation of significant volumes of melts that are rapidly dominated by an asthenospheric source (Ellam, 1992; Wilson, 1993a). In addition, strain rates appear to play an important role in the volume of melts generated; at low strain rates, conductive and convective heat diffusion probably plays an important role in suppressing partial melting (Pedersen and Ro, 1992). On the other hand, during advanced rifting stages, when strain is concentrated on the zone of future crustal separation, large volumes of melts can be generated, particularly in the presence of an asthenosphere characterized by above ambient temperatures (van Wijk et al., 2001). Finally, in the presence of volatiles, the solidus is significantly lowered and partial melting can start at much smaller stretching factors than under anhydrous conditions (McKenzie and Bickle, 1988; Latin and Waters, 1992; Wilson, 1993a).

3.2. Thermal thinning of lithosphere, doming and flood basalts

Mechanical stretching of the lithosphere, triggering partial melting of its basal parts and the upper asthenosphere, is followed by segregation of melts and their diapiric rise into the lithosphere, an increase in conductive and advective heat flux and consequently an upward displacement of the thermal asthenosphere–lithosphere boundary. Small-scale convection in the evolving asthenospheric diapir may contribute to mechanical thinning of the lithosphere by facilitating lateral ductile mass transfer (Fig. 6; Richter and McKenzie, 1978; Chase, 1979; McKenzie et al., 1980; Morgan and Baker, 1983; Steckler, 1985; Fleitout and Yuen, 1985; Fleitout et al., 1986; Keen, 1987; Moretti and Chénet, 1987; Coleman and McGuire, 1988; McGuire, 1988; Lachenbruch and Morgan, 1990; Mareschal and Gliko, 1991). Progressive thermal and mechanical thinning of the higher density mantle–lithosphere and its replacement by lower-density asthenosphere induces progressive doming of rift zones. At the same time, deviatoric tensional stresses developing in the lithosphere contribute to its further extension (Bott, 1992a).

Flood basalts, extruded prior to crustal separation, which often cover wide areas around rift zones, must be related to a hotter than ambient sublithospheric mantle (mantle plumes, upwelling asthenospheric convection cells). Flood basalts generally bear the geochemical signature of the enriched lithosphere (mantle and crust), though a contribution from enriched plumes cannot be excluded. As such, they testify to massive thermal thinning of the lithosphere, involving metasomatism of the mantle–lithosphere. This is in keeping with the frequently observed large-scale thermal doming of major rift zones preceding crustal separation (e.g. Early Jurassic Central Atlantic border lands: Favre and Stampfli, 1992; Wilson, 1997). However, as the observed large volumes of basalts cannot be generated entirely from the continental lithosphere, major contributions from plume material- and asthenosphere-derived partial melts is likely. By mixing of these mantle-derived partial melts with lithosphere-derived ones, flood basalts acquire the ‘‘fingerprint’’ of the lithosphere in terms of their trace element and isotope geochemistry. (Wilson, 1989, 1993b, 1997; Ellam, 1991; Gallagher and Hawkesworth, 1992; Menzies, 1992; Bertrand, 1991; Sheth, 1999; Smith and Lewis, 1999; Marzoli et al., 1999; McHone, 2000). Thermal thinning of the lithosphere, combined with its mechanical stretching, results in greater attenuation factor of the mantle–lithosphere (β) than at crustal levels (δ), to the point where β is considerably larger than δ. This has led to the development of the ‘‘discontinuous, depth-dependent’’ lithosphere stretching model (Fig. 4; Royden and Keen, 1980; Beaumont et al., 1982a,b; Hellinger and Sclater, 1983; Steckler, 1985; Keen, 1987).

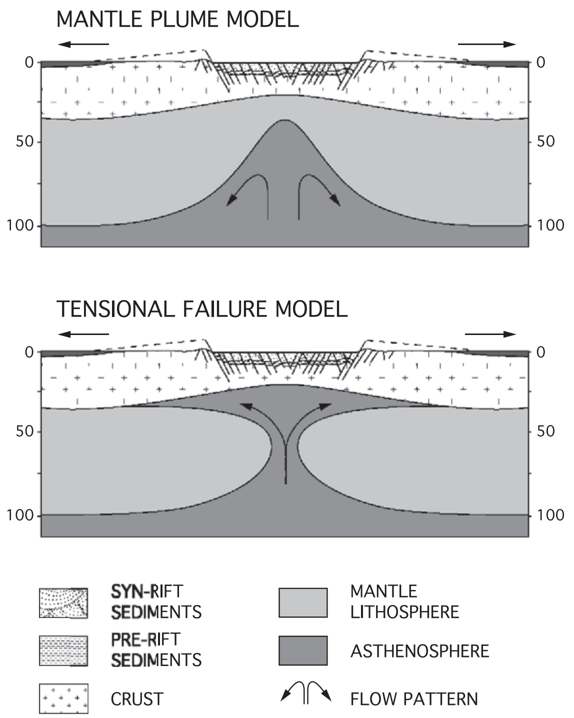

Fig. 6. Rift models.

Major differences in the level of pre-separation volcanic activity is probably not only related to the thermal state of the asthenosphere (e.g. presence or absence of deep mantle plumes) and the availability of volatiles in the mantle–lithosphere, but also to plate interaction. If the motion of the diverging continental blocks is not restrained, crustal separation can be effected after a relatively short rifting stage that is accompanied by only minor surface volcanic activity and asthenospheric melts advecting into the space opening between the diverging plates are accreted to the new plate margins as oceanic lithosphere (e.g. Miocene opening of Liguro-Provençal Basin: Roca, 2001; Early Cretaceous opening of North Atlantic: Ziegler, 1988). Conversely, if divergence of the continental blocks is impeded, either by their continued on-trend coherence across major shear zones against which the respective rifts terminate (e.g. Early Jurassic Central Atlantic and Early Cretaceous South Atlantic rifts), or by compressional far-field stresses (e.g. Palaeocene northern North Atlantic and Norwegian–Greenland Sea rift, Plio-Pleistocene Red Sea rift), massive, plume-related thermal thinning of the mantle–lithosphere may occur that is accompanied by the extrusion of voluminous plateau basalts covering wide areas around the actual rift zone (e.g. Jurassic Karoo basalts, Early Cretaceous Paraná-Etendeka basalts, Palaeocene Iceland hot spot, Cenozoic plateau basalts of Ethiopia; White and McKenzie, 1989; Menzies, 1992; Wilson, 1992, 1997). Moreover, magmas that have ascended to the base of the crust underplate it and may migrate laterally over considerable distances due to horizontal pressure gradients developing in the lower crust in response to upper crustal extension (Harry and Sawyer, 1992). Once constraints on plate divergence are removed, either by the activation of major shear zones at the termination of the respective rift system (e.g. Central Atlantic: Azores–Gibraltar–Maghrebian and Caribbean fracture zones: Ziegler, 1988; South Atlantic: Equatorial fracture zone, Fairhead and Binks, 1991; Nürnberger and Müller, 1991), or by relaxation of the far-field compressional stress system (end Palaeocene northern North Atlantic–Norwegian–Greenland Sea: Ziegler, 1988), crustal separation occurs, melts are accreted as oceanic crust at the new plate boundaries, and volcanism in the adjacent continental borderlands terminates (Ziegler, 1990, 1992).

Although major hot spot activity is thought to be related to a thermal perturbation within the asthenosphere caused by a deep mantle plume, smaller-scale "plume" activity may also be the consequence of lithospheric stretching triggering by adiabatic decompression partial melting in areas characterized by an anomalously volatile-rich asthenosphere/lithosphere (Wilson, 1989, 1993a,b; White and McKenzie, 1989; White, 1992). In this context, it is noteworthy that extension-induced development of a partially molten asthenospheric diapir, which gradually rises into the lithosphere, causes by itself further decompression of the underlying asthenosphere and consequently more extensive partial melting and melt segregation at progressively deeper levels. Thus, the evolving diapir may not only grow upwards but also downwards. Similarly, acceleration of plate divergence and the ensuing increase in seafloor spreading rates probably causes at spreading axes partial melting and melt segregation at progressively deeper asthenospheric levels reaching down to 80–100 km, as imaged seismic tomography (Anderson et al., 1992).

Melts, which intrude the lithosphere and pond at the crust/mantle boundary, provide a further mechanism for thermal doming of rift zones (Fig. 6). Emplacement of such asthenoliths, consisting of a mixture of indigenous subcrustal mantle material and melts extracted from deeper lithospheric and upper asthenospheric levels may cause temporary doming of a rift zone and a reversal in its subsidence pattern (e.g. Mid-Jurassic Central North Sea arch: Ziegler, 1990; Underhill and Partington, 1993; Neogene Baikal arch: Kiselev and Popov, 1992; Suvorov et al., 2002).

3.3. Moho discontinuity and its destabilization

In continental lithosphere, the geophysically defined Moho corresponds to the P-wave velocity (Vp) break-over from ≤ 7.8 to 8.0–8.2 km/s that is generally considered to mark the boundary between lower crustal granulite-facies rocks and the olivine-dominated mantle–lithosphere. However, the continental Moho is not always a sharp discontinuity but often a complex and variable transition zone that generally ranges in thickness between < 1 and 5 km but can expand to 10 km. The normal-incidence reflection–seismically defined Moho is characterized by very variable reflectivity and reflection patterns and does not always coincide with the Moho defined by wide-angle reflections, perhaps due to crustal anisotropy or the gradual nature of the crust–mantle boundary (Bott, 1982; Calcagnile et al., 1982; Jones et al., 1996; Hammer and Clowes, 1997; Sapin and Hirn, 1997). Moreover, it is questionable whether the geophysical and petrological crust–mantle boundaries always coincide (Ryan and Dewey, 1997). This is compatible with the fact that: (1) the orogenic fabric of the crust can be sharply truncated by the Moho; (2) upper mantle reflection bundles which dip in the same direction as the orogenic fabric of the crust must be related to subducted and eclogitized crustal material; (3) the Vp break-over associated with the Moho can be explained in terms of the granulite (Vp = 7.0–7.8) to eclogite (Vp = 8.0–8.4) transformation (Ziegler, 1996b; Le Pichon et al., 1997; Ziegler et al., 1998). Therefore, it is suspected that the Moho discontinuity is not a permanently stable boundary and that it can be destabilized by tectono-magmatic processes and subsequent phases of thermal reequilibration of the lithosphere.

Regarding the composition of the lower crust, it should be noted that reflection–seismic data show that, for instance, the crustal orogenic fabric of the Variscan fold belt of Europe reaches down to the Moho by which it is truncated (Fig. 7; Cazes et al., 1985; Behr and Heinrichs, 1987; DEKORP, 1988, 1994; Vollbrecht et al., 1989; Meissner and Bortfeld, 1990). Xenolith studies indicate that the lower crust of the eastern Rhenish Shield consists of felsic granulites and that mafic granulites occur only in the crust–mantle transition zone (Mengel, 1990; Mengel et al., 1991). Felsic lower crustal xenoliths have also been reported from other parts of the Variscan fold belt (Downes and Leyreloup, 1986). Similarly, exposures in the Norwegian Caledonides show that the lower crust is heterogeneous in composition and consist of mafic as well as felsic granulite- and eclogite-facies rocks (Austrheim, 1987; Austrheim and Mørk, 1988). Further examples of monoclinal crustal reflections extending from upper and middle crustal levels to the base of the crust where they are truncated by the Moho, come from the Baltic Sea (Balling, 1992; Abramovitz et al., 1997), the Grenville belt in Canada (Green et al., 1988), the southern Canadian Cordillera (Cook et al., 1992) and the central Australian Arunta block (Goleby et al., 1989).

This shows that the lower crust is not necessarily and exclusively composed of high-pressure mafic rocks (Taylor and McLennan, 1985; Meissner, 1986; Shaw et al., 1986). In the presence of a felsic lower crust, mantle-derived mafic melts, which ponded at its base, can induce its metasomatic reactivation and, thus, destabilization of the Moho discontinuity and its upwards displacement. This process, together with erosional unroofing of the crust, apparently played an important role in Western and Central Europe during the Permo-Carboniferous phase of wrench-faulting, rifting and magmatism during which the Variscan orogenic roots were destroyed and the Moho reequilibrated at depths of 28–35 km (Ziegler, 1990; Mengel et al., 1991; van Wees et al., 2000; Prijac et al., 2000; Ziegler et al., 2004).

Mantle-derived mafic melts, which ascended during rifting to the crust/mantle boundary (Müller, 1978), interact with the felsic lower crust and can produce by its direct melting, or a combination of assimilation and fractional crystallisation, acidic magmas. The resulting low-density granitic to granodioritic–tonalitic magmas (Cortesogno et al., 1998;

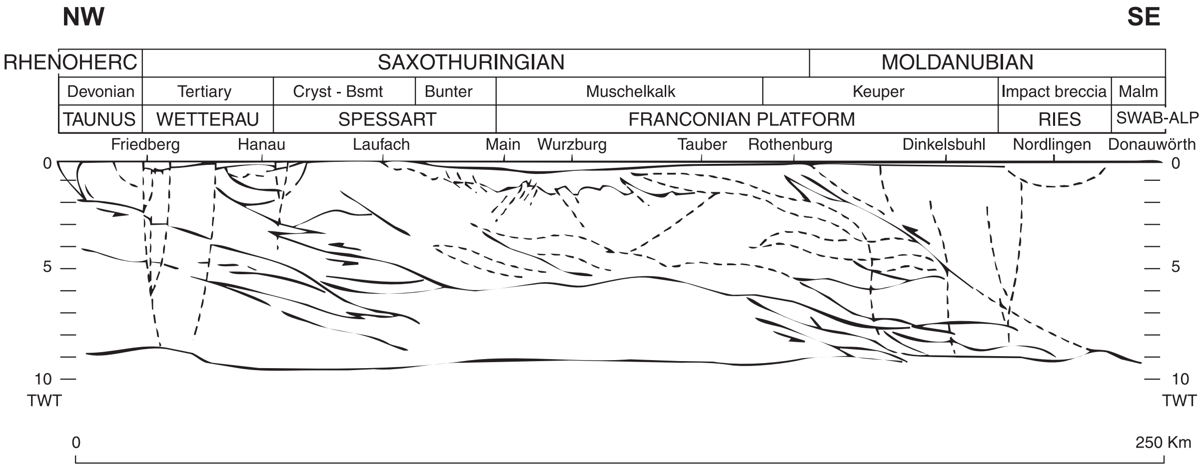

Fig. 7. Crustal fabric of the Variscan Orogen as imaged by the DEKORP 2-S reflection–seismic line, South Germany (modified after Behr and Heinrichs, 1987).

Breitkreuz and Kennedy, 1999) migrate upwards along fracture zones, adding to the mass of the middle and upper crust, thus decreasing its density. Complementary mafic and ultramafic cumulates will cool at the base of the remnant crust into the granulite stability field and may enter at depths of 30–35 km and Moho temperatures of around 600 °C the eclogite stability field (Griffin et al., 1990). This secondary crustal differentiation process involves a destabilization of the Moho discontinuity and, depending on the original composition of the lower crust (felsic or mafic), its upwards or downwards displacement (Müller, 1978; Meissner, 1986; Mohr, 1992). The occurrence of silicic extrusive rocks and anorogenic A-type granites in rift zones, having a geochemical and isotopic signature that indicates that the respective magmas evolved by fractional crystallisation of mantle-derived basaltic melts and partial fusion of the continental crust, speaks in favour of this secondary crustal differentiation concept (e.g. Oslo Graben: Neumann et al., 1995; Paraná-Etendeka province: Kirstein et al., 2001). However, in rift zones, silicic magmas can also be extracted from mantle-derived melts, stored in high-level magma chambers, solely by crystal fractionation (e.g. Iceland: Loiselle and Wones, 1979; Coleman and McGuire, 1988; Wilson, 1989; Coleman et al., 1992; Neumann et al., 1990; Mohr, 1992; Gamble et al., 1992).

Injection of mantle-derived melts into the lower crust is probably associated with the development of lower crustal laminations that are imaged by reflection–seismic data beneath palaeo-rifts (e.g. Precambrian Midcontinent rift: Hinze et al., 1992; Late Precambrian–Cambrian Reelfoot rift: Nelson and Jie, 1991; Devonian Prypiat–Diepr–Donets rift: Stephenson et al., 2001; Permian Oslo rift: Ro and Faleide, 1992; Mesozoic North Sea rift: Beach et al., 1987), active rifts (e.g. Rhine Graben: Fuchs et al., 1987) and wrench-induced basins (e.g. Paris Basin: Cazes et al., 1985). These laminations consist of densely packed, subhorizontal high-amplitude intra-lower crustal reflectors and diffraction sources that indicate the presence of high-density and velocity contrasts. Such impedance contrasts may be associated with mantle-derived mafic sills, metamorphic layering (granulite/eclogite) or extension-induced ductile shear zones or a combination thereof (Wever and Meissner, 1986; Matthews, 1986; Moretti and Pinet, 1987; Austrheim and Mørk, 1988; Warner, 1990; Blundell, 1990; Bois, 1992; Meissner and Rabbel, 1999).

Generally, such lower crustal laminations disappear away from rift zones and are absent beneath flanking stable cratonic blocks (Fig. 8); this suggests that their origin is rift-related. If these lower crustal laminations are indeed related to the injection of mantle-derived magmas into the lower crust (Müller, 1978; Matthews, 1986; Meissner, 1986; Bois et al., 1988; Warner, 1990), it would be plausible that this process could ultimately lead to a basification of a felsic lower crust.

The occurrence of crust–mantle transition zones, characterized by Vp in the 7.0–7.8 km/s range, is probably related to the permeation of the lower crust by mantle-derived melts and hydrous fluids (Fig. 9; Müller, 1978; e.g. Newfoundland shelf: Keen and Barrett, 1981; Mid-Norway shelf: Planke et al., 1991; Mjelde et al., 1998; western margin of Rockall-Hatton Bank: White, 1989, 1992; Morgan and Barton, 1990). This process has also been referred to as "magmatic underplating" (Keen and de Voogt, 1988) or "magmatic inflation" (Thompson and McCarthy, 1990), implying the addition of mantle-derived material to the lower crust, resulting in its thickening (Wilson, 1993a,b). In the case of the Oslo and Midcontinent palaeo-rifts (Ro and Faleide, 1992; Hinze et al., 1992) and the tectonically active East African rift (Mohr, 1992), magmatic underplating has apparently contributed substantially towards thickening of the strongly attenuated crust (see also Morgan and Ramberg, 1987). During the rifting stage, addition of basaltic material to the lower crust may therefore cause a downward displacement of the Moho (Gans, 1987; Mohr, 1992). During the post-rift stage, the underplated material probably cools directly into granulite-facies stability field, resulting in an increase of the lower crustal density and velocity, and by continuous cooling and burial its lower parts may enter the eclogite facies stability field, causing an upward displacement of the geophysically defined Moho (Austrheim, 1987, 1990; Austrheim and Mørk, 1988; Griffin et al., 1990; Henry et al., 1997). In this respect, the availability of hydrous fluids percolating up from the asthenosphere plays an important role in the granulite–eclogite transformation ratio, the resulting potential upward displacement of the geophysically defined Moho and rapid basin subsidence

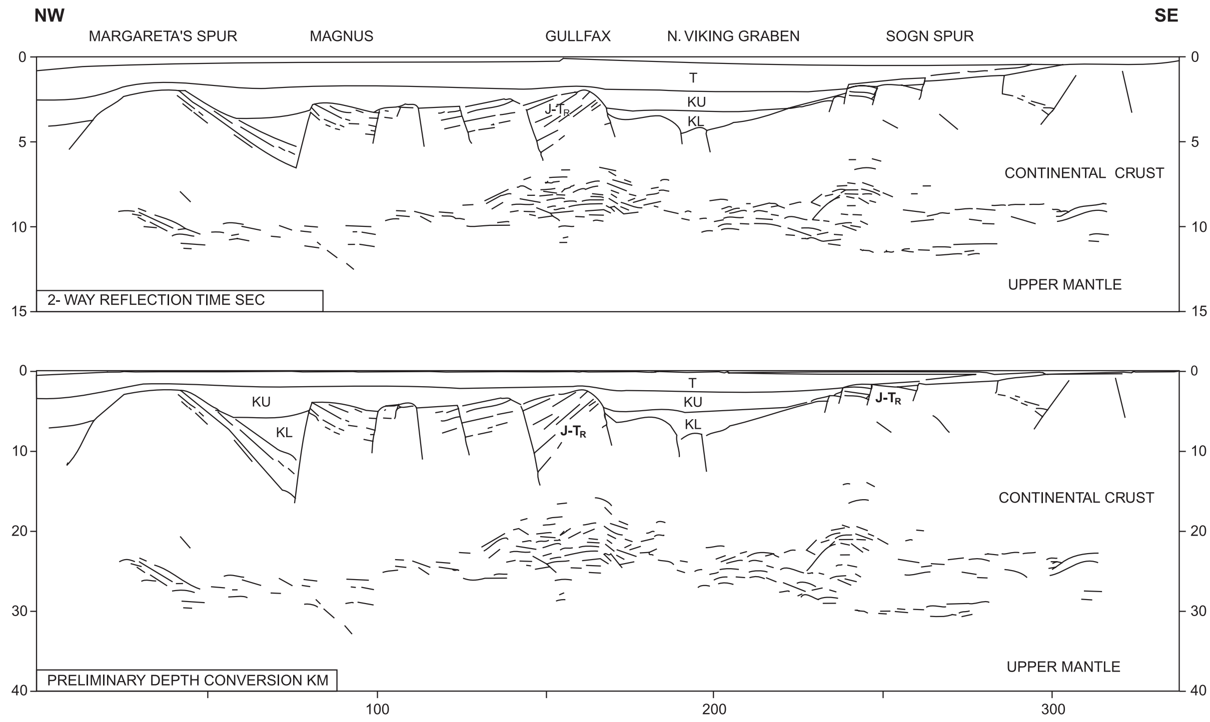

Fig. 8. Lower crustal laminations in Viking Graben, North Sea (Ziegler, 1988).

(Austrheim and Mørk, 1988; Artyushkov et al., 1991; Artyushkov, 1992; Lobkovsky et al., 1996).

In many areas where crustal underplating has apparently taken place, the amount of upper crustal extension by faulting is generally significantly smaller than the amount of lower crustal attenuation, assuming material balance of the crust during extension and neglecting the addition of mantle-derived material to the lower crust (e.g. East Newfoundland Basin: Keen et al., 1987; Provençal Basin: Le Douaran et al., 1984; Bessis and Burrus, 1986; Parentis Basin: Moretti and Pinet, 1987). This suggests that physicochemical processes associated with crustal underplating entail also incorporation of crustal material into the mantle and thus an upward displacement of the geophysically defined Moho. Therefore, the "underplating" concept, implying thickening of the crust, is in so far unfortunate as the invoked magmatic processes can apparently also contribute towards thinning of the lower crust.

A further mechanism that may cause displacement of the Moho is small-scale convection in a mantle diapir that has risen to the base of the crust, inducing lateral ductile mass transfer of (mafic?) lower crustal material away from the axial rift zone to its flanks. This may occur during advanced rifting stages preceding crustal separation (Fleitout and Yuen, 1985; Fleitout et al., 1986; Moretti and Pinet, 1987). This process could possibly be responsible, for instance, for thickening of the lower crust under the Arabian flank of the Red Sea (Voggenreiter et al., 1988).

In most rifts, the amount of upper crustal thinning by extensional faulting is considerably smaller than the amount of extension derived from the crustal thickness (Table 1; Pinet et al., 1987, Ziegler, 1988, 1990, Ziegler and van Hoorn, 1989; Bois et al., 1990; Bois, 1992; Artyushkov et al., 1991; Artyushkov, 1992). Assuming that the crust had a uniform thickness prior to its extension, this suggests that the geophysically defined Moho discontinuity can indeed become seriously destabilized during rifting processes and that the volume of the crust is not necessarily preserved during its extension, as advocated by conventional lithospheric stretching models (McKenzie,

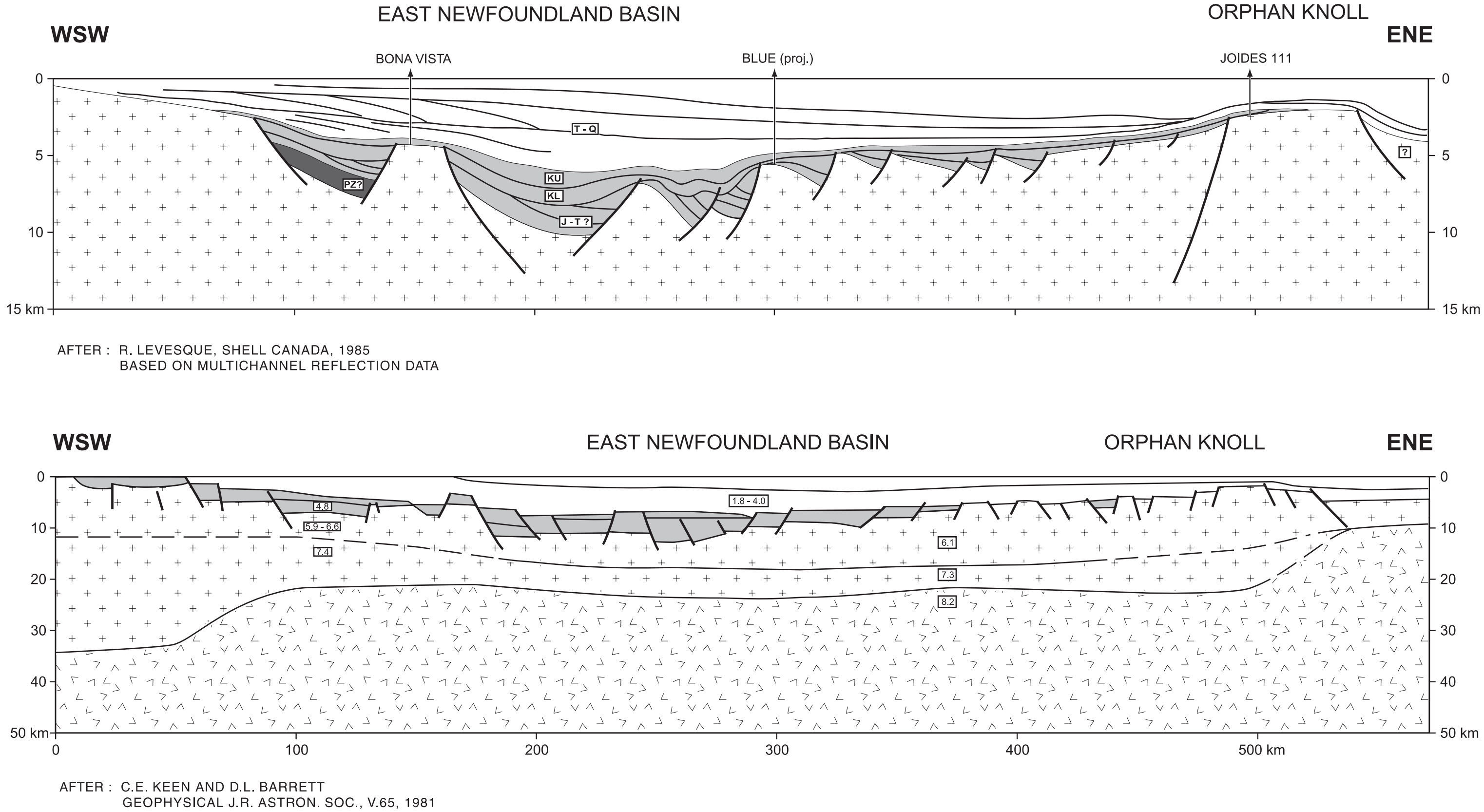

Fig. 9. East Newfoundland Basin regional structural cross section and crustal profile (after B. Levesques, Shell Canada and Keen and Barrett, 1981).

Table 1
Examples of discrepancies between extension values derived from reflection–seismic mapping of upper crustal faulting and from the crustal configuration, assuming preservation of crustal volume during rifting

Rift system	Estimates of crustal extension based on	
	Major faulting (km)	Crustal configuration (km)
Upper Rhine Graben	5–7	17
Oslo–Skagerrak Graben	10–20	40–60[a]
Baikal Rift	3–10	15–20[a]
North Sea Central Graben	25–35	100–105
North Sea Viking Graben	30–40	100–130
Termit Basin, Chad	± 40	± 80
North Biscay Margin	± 50	± 150
East Newfoundland Basin	± 100	± 280

[a] Excluding possible underplating effect.

1978; Wernicke, 1981. This raises questions on the validity of stretching factors derived from the crustal configuration of a rift, particularly as during syn-rift magmatic destabilization of the Moho crustal material can apparently be incorporated into the mantle–lithosphere and mantle material can be added to the crust.

4. Syn-rift subsidence

Syn-rift subsidence of a sedimentary basin is controlled by the balance of two mechanisms. Firstly, elastic/isostatic adjustment of the crust to stretching of the lithosphere and its adjustment to sediment loading causes subsidence of the mechanically thinned crust (Figs. 4 and 5; McKenzie, 1978; Keen and Boutilier, 1990). Depending on the depth of the lithospheric necking level, this is accompanied by either flexural uplift or downwarping of the rift zone (Fig. 12; Kooi, 1991; Kooi et al., 1992). Secondly, uplift of a rift zone is caused by upwelling of the asthenosphere into the space created by mechanical stretching of the lithosphere, thermal upward displacement of the asthenosphere–lithosphere boundary, thermal expansion of the lithosphere and intrusion of melts at the base of the crust (Fig. 6; Turcotte and Emermann, 1983). Thus, the geometry of a rifted basin is a function of the elastic/isostatic response of the lithosphere to its mechanical stretching and related thermal perturbation (van der Beek et al., 1994).

4.1. Stretching models

Depending on the applicability of the "pure-shear" (McKenzie, 1978) or the "simple-shear" model (Wernicke, 1981, 1985), or a combination thereof (Barbier et al., 1986; Kusznir and Egan, 1989; Kusznir et al., 1987, 1991), the zone of upper crustal extension, corresponding to the subsiding rift, may coincide with the zone of mantle–lithospheric attenuation (pure- and combined-shear) or may be laterally offset from it (simple-shear, Fig. 5). Under conditions of pure-shear lithospheric extension, magmatic activity should be centred on the rift axis where in time also MORB-type magmas can be extruded after a high degree of extension has been achieved. By contrast, under simple-shear conditions, magmatic activity is asymmetrically distributed with respect to the rift axis and MORB-type extrusives may occur on one of the rift flanks (e.g. Basin and Range Province: Jones et al., 1992; Red Sea: Favre and Stampfli, 1992; Ethiopian Rift: Kazmin, 1991).

A modification to the pure-shear model is the "continuous depth-dependent" stretching model which assumes that stretching of the mantle–lithosphere affects a broader area than the zone of crustal extension (Fig. 4; Rowley and Sahagian, 1986). In both models, it is assumed that the asthenosphere wells up passively into the space created by mechanical attenuation of the mantle–lithosphere. In depth-dependent stretching models, this commonly gives rise to flexural uplift of the rift shoulders, and in the flexural cantilever model, which assumes ductile deformation of the lower crust, this causes footwall uplift of the rift flanks and intrabasinal fault blocks (Kusznir and Egan, 1989; Kusznir et al., 1991; Kusznir and Ziegler, 1992). By the same mechanism, the simple-shear model predicts asymmetrical doming of a rift zone or even flexural uplift of an arch located to one side of the zone of upper crustal extension (Wernicke, 1981, 1985; Wernicke and Tilke, 1989). A modification to the simple-shear model envisages that massive upper crustal extensional unloading of the lithosphere causes its isostatic uplift and passive inflow of the asthenosphere (Etheridge et al., 1989; Wernicke, 1990; Karner et al., 1992).

4.2. Thermal uplift of rift zones

Models of purely mechanical, "passive" stretching of the lithosphere (Figs. 4 and 5) do not take into account the possibility of thermal upward displacement of the asthenosphere–lithosphere boundary. Conversely, "active" rifting models assume that progressive thermal thinning of the mantle–lithosphere is caused by a mantle plume-related temperature perturbation of the lithosphere–asthenosphere boundary only (Fleitout et al., 1986; Mareschal and Gliko, 1991). However, as discussed above, stress-induced extension of the lithosphere, causing upwelling of the asthenosphere, and by adiabatic decompression partial melting of the basal parts of the lithosphere and upper asthenosphere, is accompanied by thermal destabilization of the asthenosphere–lithosphere boundary, which contributes to its progressive upward displacement and thus to doming of a rift zone. Moreover, intrusion of melts at intralithospheric levels, such as the crust–mantle boundary, can also cause uplift of a rift zone (Fig. 6; mantle plume and tensional failure models; Turcotte, 1981; Turcotte and Emermann, 1983; Olsen and Morgan, 1995).

The notion of plume-related domal uplift of future rift zones, preceding and accompanying early phases of crustal extension (Burke and Whiteman, 1973; Dewey and Burke, 1974) is not compatible with the stratigraphic record of rifts in which thick pre-rift sediments are overlain in depositional continuity by syn-rift deposits or are separated from the latter by only a minor hiatus. For instance, in the North Sea Central Graben, in which crustal extension began during the Early Triassic, sedimentation was continuous during Permian and Triassic times (Ziegler, 1990). Similarly, the stratigraphic record of the Tucano and Sergipe–Alagoas basins (Karner et al., 1992) and the Pripyat–Dniepr–Donets graben (Stephenson et al., 2001) shows no break in sedimentation at the pre- to syn-rift transition. In the Gulf of Suez, a minor hiatus separates late Eocene platform carbonates and early Oligocene partly marine sands from late Oligocene earliest syn-rift sediments (Moretti and Chénet, 1987; Bosworth and McClay, 2001). In the Rhine Graben, down-faulted, thick Triassic and Jurassic pre-rift sediments are separated by a major hiatus from late Eocene and younger syn-rift deposits; however, this hiatus is related to a phase of intraplate compressional deformation, which preceded the onset of graben subsidence (Ziegler, 1990; Schumacher, 2002). In the Rio Grande Rift, several-kilometer-thick Palaeozoic to Eocene pre-rift sediments are overlain with a minor hiatus by syn-rift Oligocene volcanic and Neogene clastic rocks (Russell and Snelson, 1990; Baldridge et al., 1995). In rifts evolving on young orogenic belts, pre-rift sediments are often involved in deeply truncated compressional structures; a distinction between uplift caused by compressional deformation and early rift doming is not feasible (e.g. Basin and Range Province: Jones et al., 1992; Pannonian Basin: Horvàth, 1993). Also, for rifts that evolved on ancient cratons lacking an extensive sedimentary cover, it is difficult to determine whether early rifting stages were accompanied by regional doming of the rift zone. However, in the East African rift system, early phases of plateau basalt extrusion were apparently associated with crustal downwarping (Mohr, 1992; Morley et al., 1992). Similarly, the early rifting stages of the Midcontinent and Oslo rifts were characterized by regional crustal downwarping (Cannon, 1992; Sundvoll et al., 1992; Sundvoll and Larsen, 1994). This is indicative for a shallow lithospheric necking level (Fig. 12; Kooi et al., 1992). In very deep rifted basins, there is often insufficient information to determine whether syn-rift deposits are underlain by pre-rift sediments (e.g. South Atlantic rifts, Teisserenc and Villemin, 1990); this inhibits assessment of their possible early syn-rift doming.

On the other hand, massive thermal doming of rift zones can commence 15–60 My after the beginning of crustal extensions (e.g. Gulf of Suez: 15–20 My; East African rift: 20–25 My; Baikal rift: 20–30 My; Rhenish Shield: 25–30 My; Central Atlantic rift: 30–40 My; North Sea: 60 My; Ziegler, 1992). Uplift of rift flanks can exceed 2 km whereas grabens may be elevated by nearly as much above sea level (e.g. surface of Lake Baikal +450 m, Lake Tanganyika +773 m, Lake Khubsugul +1650 m; see Logatchev and Zorin, 1992). However, some rifts, such as the Mesozoic Polish Trough and the Cretaceous Sudan rifts, show throughout their evolution neither evidence for regional doming nor for significant flank uplift (Ziegler, 1990; Kutek, 2001; McHargue et al., 1992).

The subsidence pattern of a rifted basin can be reversed if the rate of thermal uplift exceeds the rate of isostatic subsidence caused by lithospheric extension. This can result in uplift of the rifted basin above the erosional base level and truncation of its sedimentary fill (e.g. Mid-Jurassic Central North Sea Rift dome: Ziegler, 1990; Underhill and Partington, 1993; Mio-Pliocene uplift of Rhenish Shield straddling the triple junction of the Rhine, Roer and Hessian grabens: Ziegler, 1994; Neogene uplift of Sayan–Baikal dome: Logatchev and Zorin, 1992). This implies that a slowdown or even a reversal in the subsidence pattern of a rift does not necessarily imply that crustal stretching has decreased or terminated altogether (e.g. Gulf of Suez: Moretti and Chénet, 1987; Evans, 1988; Richardson and Arthur, 1988). In tectonically silled rift basins, sedimentation may continue despite their uplift above sea level, as shown by Lake Baikal, the East African rift lakes and the occurrence of a thick Bajocian–Bathonian continental clastic series in the North Sea Central Graben (Ziegler, 1990).

Most rift domes straddle more or less symmetrically the zone of upper crustal extension with volcanic activity showing no marked concentration on one of the rift flanks (e.g. Rhine Graben: Prodehl et al., 1992, 1995; Jurassic Central North Sea: Ziegler, 1990; East African rifts: Rosendahl, 1987; Braile et al., 1995). Such rifts conform essentially to a modified pure-shear model. However, a strongly asymmetric relationship between the rift and the domed area can be taken as an indication for simple-shear deformation at least at crustal levels (e.g. Lake Baikal: Logatchev and Zorin, 1992; Keller et al., 1995). Doming of the Red Sea area is markedly asymmetric and volcanic activity is concentrated on the Arabian Shield; this suggests that the rising mantle diapir is laterally offset from the zone of crustal extension and that simple-shear deformation of the lithosphere plays an important role in the observed asymmetry. In such a model, the Arabian flank of the Red Sea corresponds to the upper plate and the African side to the lower plate (Almond, 1986; Voggenreiter et al., 1988; Coleman and McGuire, 1988; Favre and Stampfli, 1992).

Development of a "separation unconformity" on passive margins during their final rifting stage preceding crustal separation is a function of thermal uplift exceeding extension-induced isostatic subsidence, resulting in subaereal exposure of large parts of the rift zone. Although such regional erosional unconformities are typical for a number of passive margins (e.g. Northwest Shelf of Australia: Williamson et al., 1990; Butcher, 1990; Grand Banks: Tankard and Welsink, 1989), they are not evident on other margins. For instance, on the Mid-Norway Shelf, the absence of a regional Palaeocene separation unconformity indicates that despite the impingement of the Iceland plume, thermal uplift of this basin was not large enough to overcome water depths that had been established in it during the Late Cretaceous (Bukovics and Ziegler, 1985; Roberts et al., 1997; Swiecicki et al., 1998). Local unconformities over rift shoulders and intrabasinal fault blocks can be attributed to footwall uplift in response to extensional unloading of the lithosphere, a phenomenon which can be enhanced by thermal uplift of the rift zone and gradual strain concentration to the axial rift zone (Kusznir et al., 1991; Kusznir and Ziegler, 1992; Ziegler, 1988).

5. Structural style of rifts

The structural style of rifts, as defined at upper crustal and syn-rift sedimentary levels, is influenced by the thickness and thermal state of the crust and mantle–lithosphere at the onset of rifting, by the amount of crustal extension and the width over which it is distributed, the mode of crustal extension (orthogonal or oblique, simple- or pure-shear) and the lithological composition of the pre- and syn-rift sediments (Cloetingh et al., 1995; Ziegler, 1996b).

5.1. Lithosphere strength and deformation mode

The strength of continental lithosphere is controlled by its depth-dependent rheological structure in which the thickness and composition of the crust, the thickness of the mantle–lithosphere, the potential temperature of the asthenosphere, the presence or absence of fluids and strain rates play a dominant role. By contrast, the strength of oceanic lithosphere depends on its thermal regime, which controls its essentially age-dependent thickness (Panza, 1980; Kusznir and Park, 1987; Buck, 1991; Stephenson and Cloetingh, 1991; Cloetingh and Banda, 1992; Burov et al., 1993; Cloetingh and Burov, 1996).

Fig. 10 gives synthetic strength envelopes for three different types of continental lithosphere and for

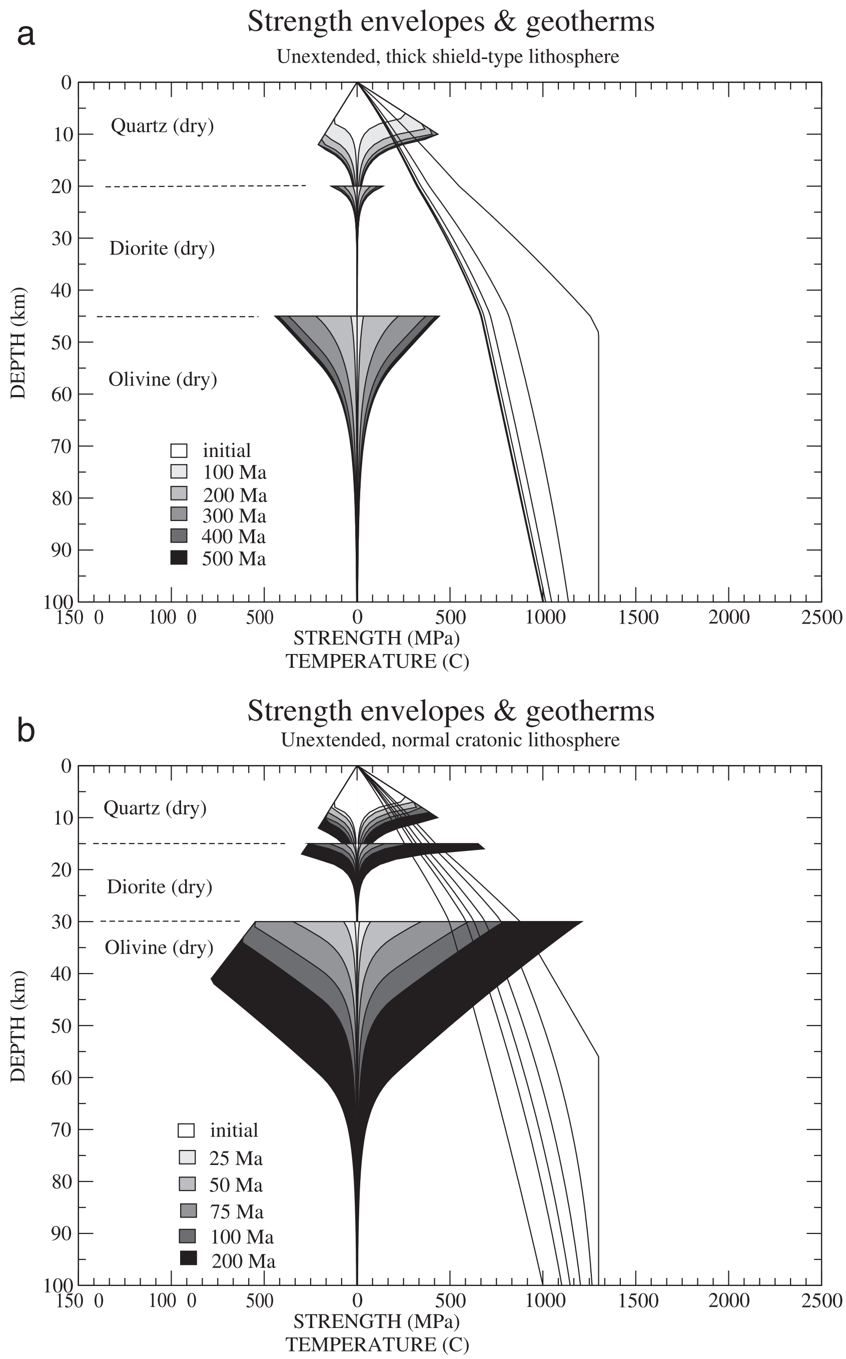

Fig. 10. Depth-dependent rheological models for various lithosphere types and a range of geothermal gradients, assuming a dry quartz/diorite/olivine mineralogy for continental lithosphere (Ziegler, 1996b; Ziegler et al., 2001). (a) Unextended, thick-shield-type lithosphere with a crustal thickness of 45 km and a mantle–lithosphere thickness of 155 km. (b) Unextended, "normal" cratonic lithosphere with a crustal thickness of 30 km and a mantle–lithosphere thickness of 70 km. (c) Unextended, young orogenic lithosphere with a crustal thickness of 60 km and a mantle–lithosphere thickness of 140 km. (d) Extended, cratonic lithosphere with a crustal thickness of 20 km and a mantle–lithosphere thickness of 50 km. (e) Oceanic lithosphere.

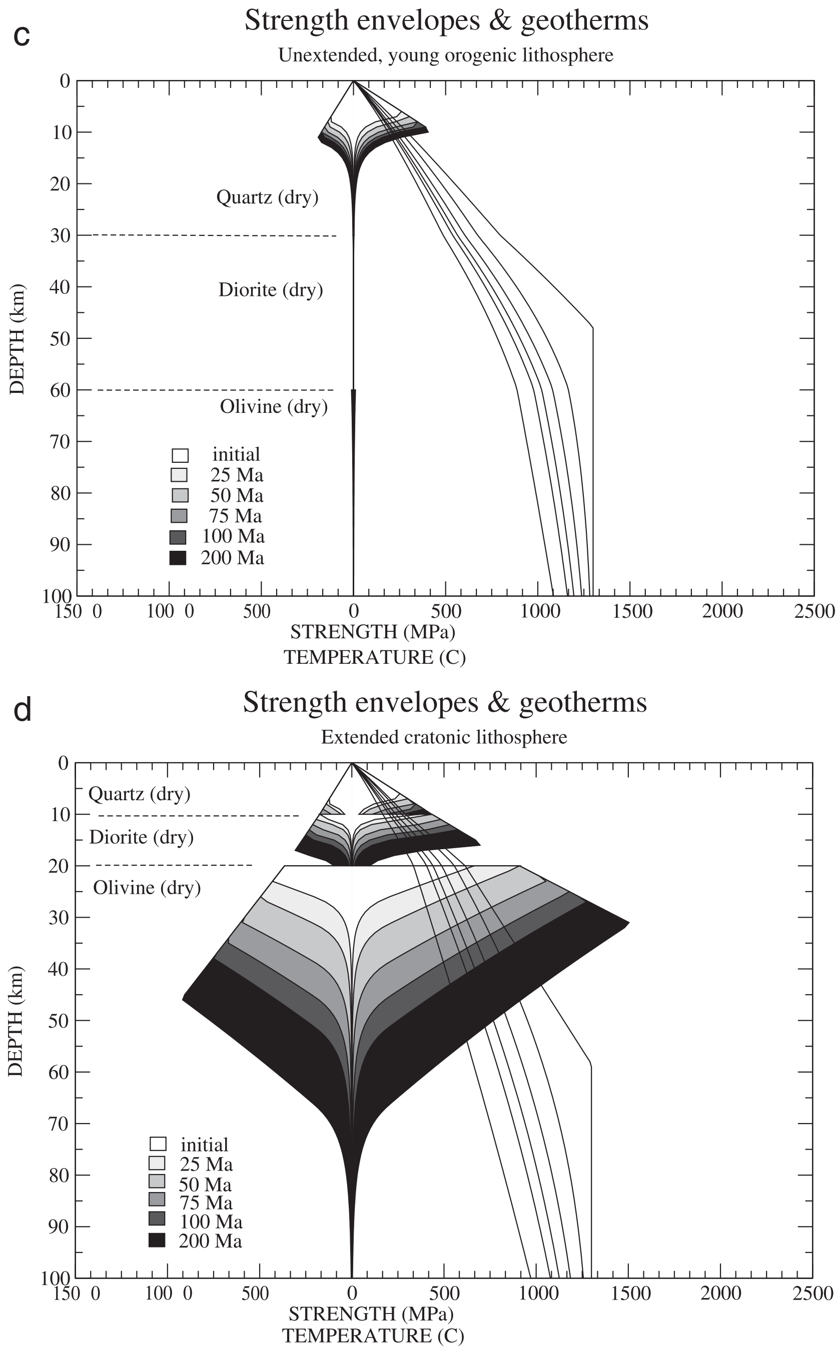

Fig. 10 (*continued*).

oceanic lithosphere at a range of geothermal gradients. These theoretical rheological models indicate that thermally stabilized continental lithosphere consists of the mechanically strong upper crust, which is separated by a weak lower crustal layer from the strong upper part of the mantle–lithosphere, which

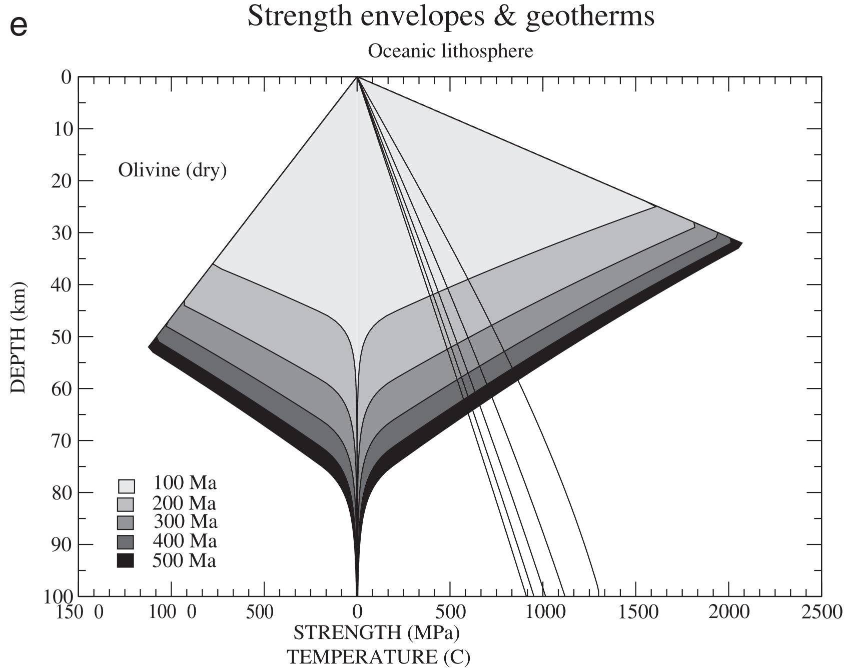

Fig. 10 (*continued*).

in turn overlies the weak lower mantle–lithosphere. By contrast, oceanic lithosphere has a more homogeneous composition and is characterized by a much simpler rheological structure. Rheologically speaking, thermally stabilized oceanic lithosphere is considerably stronger than all types of continental lithosphere. However, the strength of oceanic lithosphere can be seriously weakened by transform faults and by the thermal blanketing effect of thick sedimentary prisms prograding onto it (e.g. Gulf of Mexico, Niger Delta, Bengal Fan; Ziegler et al., 1998).

The strength of continental crust depends largely on its composition, thermal regime and the presence of fluids, and also on the availability of pre-existing crustal discontinuities. Deep-reaching crustal discontinuities, such as thrust- and wrench-faults, cause significant weakening of the otherwise mechanically strong upper parts of the crust. As such discontinuities are apparently characterized by a reduced frictional angle, particularly in the presence of fluids (van Wees, 1994), they are prone to reactivation at stress levels that are well below those required for the development of new faults. Deep reflection–seismic profiles show that the crust of Late Proterozoic and Palaeozoic orogenic belts is generally characterized by a monoclinal fabric that extends from upper crustal levels down to the Moho at which it either soles out or by which it is truncated (Fig. 7; e.g. Cazes and Torreilles, 1988; Green et al., 1988; Meissner and Bortfeld, 1990; Klemperer and Hobbs, 1991; Abramovitz et al., 1997). This fabric reflects the presence of deep-reaching lithological inhomogeneities and shear zones.

The strength of the continental upper mantle–lithosphere depends to a large extent on the thickness of the crust but also on its age and thermal regime. Thermally stabilized stretched continental lithosphere with a 20-km-thick crust and a mantle–lithosphere thickness of 50 km is mechanically stronger than unstretched lithosphere with a 30-km-thick crust and a 70-km-thick mantle–lithosphere (compare Fig. 10b and d). Extension of stabilized continental crustal segments precludes ductile flow of the lower crust and faults will be steep to listric and propagate towards the hanging wall, i.e. towards the basin centre (Bertotti et al., 2000). Under these conditions, the lower crust will deform by distributing ductile shear in the brittle–ductile transition domain. This is compatible with the occurrence of earthquakes within the lower crust and even close to the Moho (e.g. southern Rhine Graben: Bonjer, 1997; East African rifts: Shudofsky et al., 1987).

On the other hand, in young orogenic belts, which are characterized by crustal thicknesses of up to 60 km and an elevated heat flow, the mechanically strong part of the crust is thin and the mantle–lithosphere is also weak (Fig. 10c). Extension of this type of lithosphere, involving ductile flow of the lower and middle crust along pressure gradients away from areas lacking upper crustal extension into zones of major upper crustal extensional unroofing, can cause crustal thinning and thickening, respectively. This deformation mode gives rise to the development of core complexes with faults propagating towards the hanging wall (e.g. Basin and Range Province: Wernicke, 1990; Buck, 1991; Jones et al., 1992; Parsons, 1995; Bertotti et al., 2000). However, crustal flow will cease after major crustal thinning has been achieved, mainly due to extensional decompression of the lower crust (Bertotti et al., 2000).

Generally, the upper mantle of thermally stabilized, old cratonic lithosphere is considerably stronger than the strong part of its upper crust (Fig. 10a; Moisio et al., 2000). However, the occurrence of upper mantle reflectors, which generally dip in the same direction as the crustal fabric and are probably related to subducted oceanic and/or continental crustal material, suggests that the continental mantle–lithosphere is not necessarily homogenous but can contain lithological discontinuities that enhance its mechanical anisotropy (Vauchez et al., 1998; Ziegler et al., 1998). Such discontinuities, consisting of eclogitized crustal material, can potentially weaken the strong upper part of the mantle–lithosphere. Moreover, even in the face of similar crustal thicknesses, the heat flow of deeply degraded Late Precambrian and Phanerozoic orogenic belts is still elevated as compared to adjacent old cratons (e.g. Pan African belts of Africa and Arabia; Janssen, 1996). This is probably due to the younger age of their mantle–lithosphere and possibly also to a higher radiogenic heat generation potential of their crust. These factors contribute to weakening of former mobile zones to the end that they present rheologically weak zones within a craton, as evidenced by their preferential reactivation during the breakup of Pangea (Ziegler, 1989b; Janssen et al., 1995; Ryan and Dewey, 1997; Ziegler et al., 2001).

From a rheological point of view, the thermally destabilized lithosphere of tectonically active rifts, as well as of rifts and passive margins that have undergone only a relatively short post-rift evolution (e.g. 25 Ma), is considerably weaker than that of thermally stabilized rifts and of unstretched lithosphere (Figs. 10d and 11; Ziegler et al., 1998). In this respect, it must be realized that during rifting, progressive mechanical and thermal thinning of the mantle–lithosphere and its substitution by the upwelling asthenosphere is accompanied by a rise in geotherms causing progressive weakening of the extended lithosphere. In addition, its permeation by fluids causes its further weakening (Fig. 11). Upon decay of the rift-induced thermal anomaly, rift zones are rheologically speaking considerably stronger than unstretched lithosphere (Fig. 10). However, accumulation of thick syn- and post-rift sedimentary sequences can cause by thermal blanketing a weakening of the strong parts of the upper crust and mantle–lithosphere of rifted basins (Stephenson, 1989). Moreover, as faults permanently weaken the crust of rifted basins, they are prone to tensional as well as compressional reactivation (Ziegler et al., 1995, 1998, 2001, 2002).

In view of its rheological structure, the continental lithosphere can be regarded under certain conditions as a two-layered visco-elastic beam (Fig. 10; Reston, 1990; ter Voorde et al., 1998). The response of such a system to the build-up of extensional and compressional stresses depends on the thickness, strength and spacing of the two competent layers, on stress magnitudes and strain rates and the thermal regime (Zeyen et al., 1997). As the structure of continental lithosphere is also areally heterogeneous, its weakest parts start to yield first once intraplate stress levels equate their strength.

Thus, depending on their lithospheric configuration, large areas around future zones of crustal separation can be tensionally deformed during rifting phases. In this respect, spatial variations in the rheological structure of the lithosphere play an important role. Although such variations are primarily controlled by the thickness of the crust and mantle–lithosphere and the prevailing thermal regime, the presence of crustal and mantle–lithospheric discontinuities can significantly reduce the strength of the lithosphere. In this, the orientation of such discontinuities with respect to the prevailing stress field plays an important

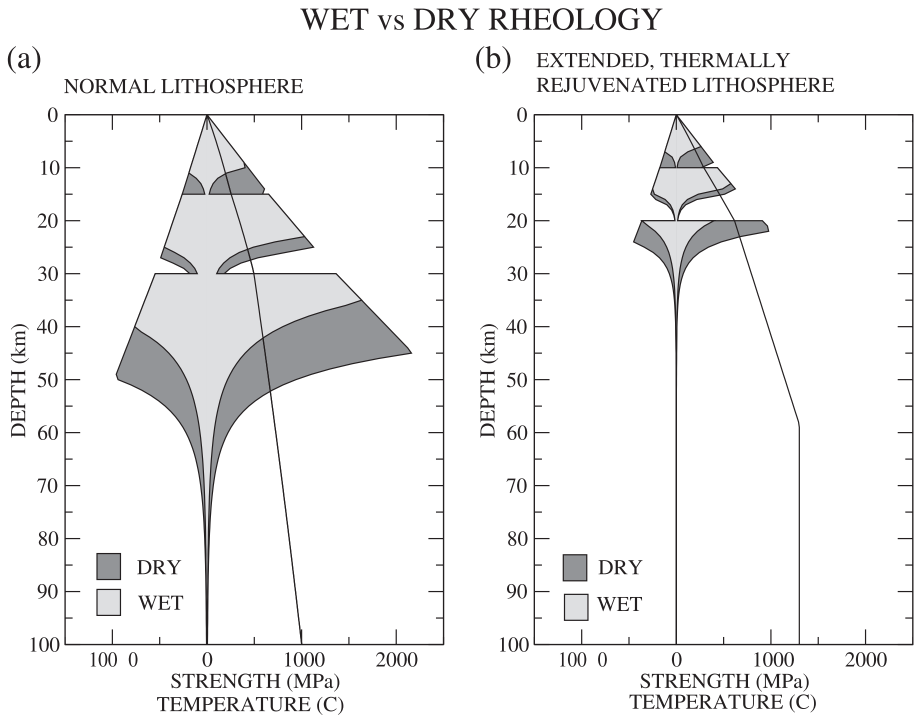

Fig. 11. Depth-dependent rheological models for dry and wet, unextended "normal" cratonic lithosphere and stretched, thermally attenuated lithosphere, assuming a quartz/diorite/olivine mineralogy (Ziegler et al., 2001). (a) Unextended, cratonic lithosphere with a crustal thickness of 30 km and a mantle–lithosphere thickness of 70 km. (b) Extended, thermally destabilized cratonic lithosphere with a crustal thickness of 20 km and a mantle–lithosphere thickness of 38 km.

role in terms of their reactivation potential (Ziegler et al., 1995; Brun and Nalpas, 1996).

Indeed, the location of most of the major Mesozoic and Cenozoic rift systems appears to be controlled by the distribution of the Pan African, Caledonian, Hercynian and Alpine orogens, the lithosphere of which is characterized by deep-reaching crustal shear zones and lithological discontinuities, and possible upper mantle reflectors related to subducted crustal material (Panza, 1980; Vauchez et al., 1998; Ziegler et al., 1998, 2001). Only few rifts cross-cut the orogenic fabric of the crust, as seen for instance in the Labrador Sea, the area between Rockall–Hatton Bank and Greenland, the North Sea, the Rhine Graben (Ziegler, 1989b, 1990) and the Devonian Dniepr–Donets rift (Shchipansky and Bogdanova, 1996; Stephenson et al., 2001).

The geometry of rift systems that are superimposed on former mobile belts and follow their strike is generally strongly influenced by the pre-existing basement fabric ((Braun and Beaumont, 1989a,b; Dunbar and Sawyer, 1989; Govers and Wortel, 1993). During lithospheric extension, crustal weakness zones can be reactivated at stress levels that are considerably below the bulk yield strength of homogeneous crust, provided their orientation is such that they can be reactivated under the prevailing stress field. Such weakness zones present areas of preferential strain concentration, even if they are located at considerable distances (up to several 100 km) to one or the other or both sides of the zone of mantle–lithosphere stretching. Linkage of such crustal weakness zones and decoupling of the upper crust from the mantle–lithosphere at the level of the ductile lower crust can give rise to large-scale "simple-shear" extensional deformation (Wernicke, 1981; Lister et al., 1986; Sawyer and Harry, 1991; Harry and Sawyer, 1992). Under such conditions, pre-existing crustal and mantle–lithospheric discontinuities probably determine the location and polarity of a rift system (e.g. South Atlantic: Chang et al., 1992; Karner et al., 1992; Maurin and Guiraud, 1993; Central

Atlantic and Red Sea: Favre and Stampfli, 1992; Piqué and Laville, 1995; see also Ziegler, 1996b; Ziegler et al., 1998; Gulf of Thailand: Watcharanantakul and Morley, 2000).

However, in areas where an evolving rift cross-cuts the orogenic fabric of the crust, pre-existing crustal and mantle–lithospheric discontinuities cannot be reactivated under the prevailing stress field. Correspondingly, new faults will develop and "pure-shear" lithospheric deformation is likely to prevail (McKenzie, 1978; Govers and Wortel, 1993; Ziegler, 1996b). This is exemplified by the geometry of e.g. the North Sea rift (Ziegler, 1990), the Rhine Graben (Prodehl et al., 1995) and the Dniepr–Donets rift (Shchipansky and Bogdanova, 1996; Ilchenko, 1996; Stephenson et al., 2001).

On the other hand, oceanic lithosphere behaves as a single-layer beam that is thinner than the competent parts of thick cratonic continental lithosphere. However, in view of the high strength of mature oceanic lithosphere, its extensional deformation requires considerably higher stress levels than the deformation of continental lithosphere (Fig. 10d; Cloetingh et al., 1989). This suggests that extensional stresses transmitted through mature oceanic lithosphere can be large enough to cause failure of the continental lithosphere forming part of the same plate, without at the same time causing deformation of the oceanic lithosphere (e.g. Permian separation of Cimmerian terranes from Africa–Arabia: Ziegler et al., 1998, 2001; Stampfli et al., 2001).

These considerations indicate that the mode of extension of continental lithosphere is controlled by the thickness and viscosity of the crust, the yield strength of the mantle–lithosphere and extensional strain rates, with viscosity and yield strength being controlled by the temperature structure and composition of the lithosphere (Buck, 1991). Under conditions of pure-shear extension, the rheological properties of the lithosphere control its tensional necking depth (Braun and Beaumont, 1989a,b). Necking of cratonic lithosphere occurring at depths of 15–20 km and even as deep as the crust/mantle boundary causes upward flexure of the rift zone. Necking at crustal levels shallower than 15 km causes downward flexing of the rift zone and thus the absence of shoulder uplifts (Fig. 12; Kooi, 1991; Kooi et al., 1992). The latter is observed, for instance, during the evolution of Triassic rifts in Europe which developed on a lithosphere that was thermally destabilized by Permo-Carboniferous wrench-induced magmatism (Ziegler, 1990).

During rifting, gradually rising lithospheric isotherms entail an upward shift of the lithospheric necking level and the intracrustal brittle/ductile deformation boundary. Model experiments indicate that the width of a rift zone depends on the thickness of the crust and the depth at which the brittle/ductile transition zone is located at the onset of lithospheric extension (Fig. 13; Allemand and Brun, 1991; Buck, 1991; Brun, 1999). As during rifting, this interface rises in response to rising isotherms, crustal strain concentrates in time on

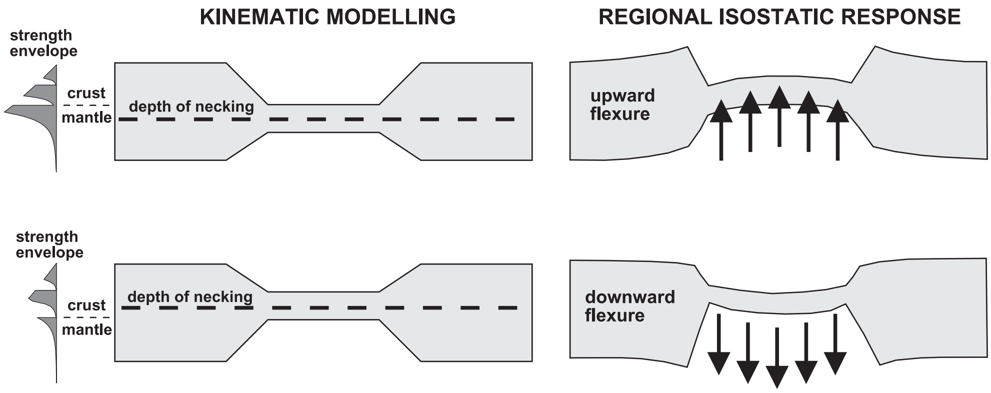

Fig. 12. Concept of lithospheric necking (after Kooi, 1991; modified after Braun and Beaumont, 1989a). The level of necking is defined as the level of no vertical motions in the absence of isostatic forces. Left panel kinematically induced configuration after rifting for different necking depths. Right panel, subsequent flexural rebound (van Wees and Cloetingh, 1996).

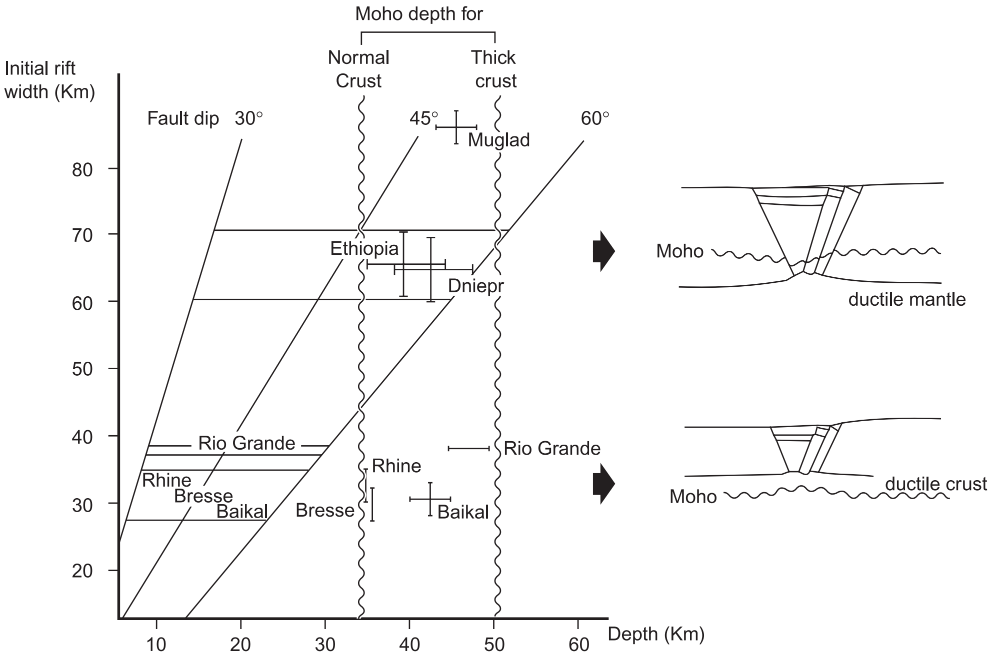

Fig. 13. Initial rift width versus depth range of brittle–ductile transition zone (modified after Allemand and Brun, 1991).

the thermally more intensely weakened parts of a rift, generally corresponding to the rift axis, and thus causes narrowing of the rift (Sawyer and Harry, 1991). This phenomenon is observed during the evolution of the North Sea, the Norwegian–Greenland Sea (Ziegler, 1988, 1990; Reemst and Cloetingh, 2000) and the Central Atlantic rift systems (Favre and Stampfli, 1992). However, at slow strain rates, and particularly during intermittent rifting activity, crustal thinning, involving uplift of the crust/mantle boundary, entails strengthening of the upper mantle–lithosphere with conductive heat loss preventing its thermal weakening (compare Fig. 10b and d) (Buck, 1991). This process may contribute towards the abandonment of lateral rift systems and the shift of rifting activity to previously less extended areas (van Wijk and Cloetingh, 2002). Nevertheless, evidence for tensional reactivation of rifts, which had been abandoned millions of years ago, suggests that crustal-scale faults permanently weaken their lithosphere to the degree that rifts are is prone to tensional and compressional reactivation (Ziegler et al., 1995, 2001, 2002) (e.g. Devonian reactivation of Early Riphean Pachelma and Abdulino rifts and Ordovician Pechora–Kolva rift system on East-European craton: Nikishin et al., 1996; Fokin et al., 2001; Early Cretaceous reactivation of Permo-Triassic grabens on Namibia margin: Clemson et al., 1997; Cenozoic reactivation of Permo-Triassic Rukwa graben: Delvaux, 2001).

Particularly in the presence of pre-existing crustal fractures and mantle–lithospheric discontinuities, which can be tensionally reactivated, simple-shear deformation of the entire lithosphere can play an important role during early rifting stages. However, during advanced rifting stages and in the face of a progressive thermal weakening of the lithosphere, simple shear will generally be confined to upper crustal levels with pure shear dominating the deformation of the lower crust and mantle–lithosphere (e.g. Sudan rifts: McHargue et al., 1992; Sergipe-Alagoas rift: Karner et al., 1992). Yet, at all stages of rift evolution, strain rates are likely to play an important role in the mode of lithosphere deformation.

Tensional reactivation of thrust faults plays an important role in the extension of orogenically destabilized lithosphere (e.g. Basin and Range Province: Wernicke et al., 1987; Pannonian Basin: Tari et al., 1992; Triassic half-grabens of the Appalachians: Costain and Ç oruh, 1989; Sawyer and Harry, 1991). Similarly, ancient suture zones can be tensionally reactivated (e.g. Baikal Rift: Logatchev and Zorin, 1992; Delvaux et al., 1997). Tensional reactivation of pre-existing shear zones has played an important role in the development of many rift systems and in the geometry of their fault patterns (e.g. East African rift system: Rosendahl, 1987; Rosendahl et al., 1992; Delvaux, 2001; South Atlantic rifts: Chang et al., 1992; Cenozoic rift system of Europe: Ziegler, 1990, 1994; Norwegian–Greenland Sea: Doré and Gage, 1987; Osmundsen et al., 2002). However, in rifts which cross-cut the structural grain of the basement more or less orthogonally, new faults develop (e.g. Labrador Sea: Ziegler, 1989b, 1990; Dniepr–Donets rifts: Stephenson et al., 2001). The fault system of the North Sea rift, which obliquely cross-cuts the Caledonides, consists of a combination of reactivated Caledonian and Devonian crustal discontinuities and new Mesozoic faults (Doré and Gage, 1987; Ziegler, 1990; Bartholomew et al., 1993).

During initial rifting stages, reactivation of pre-existing crustal discontinuities can lead to the subsidence of isolated grabens and half-grabens that are linked by shear zones. With increasing strain, such grabens propagate towards each other, coalesce and evolve into a more or less continuous rift system (e.g. East African rift: Nelson et al., 1992; Cenozoic rift system of Western and Central Europe: Ziegler, 1994). Propagation of established rift systems into previously unextended domains can entail reactivation of pre-existing crustal discontinuities (e.g. Triassic southward propagation of Norwegian–Greenland Sea rift into the Central Atlantic domain: Ziegler, 1988), or the development of new fault systems cross-cutting the basement grain (e.g. Late Jurassic–Early Cretaceous northward propagation of North Atlantic rift into Labrador Sea–Baffin Bay: Ziegler, 1988, 1989b).

At the scale of the lithosphere, the strength of the mechanically strong upper part of the mantle–lithosphere, which depends on its thermal state and the thickness of the crust, plays an important role in the localization of rift zones. Moreover, lateral thickness heterogeneities of the lithosphere appear to play an important role in the localization of rifts (e.g. Oslo Graben: Pascal et al., 2002). At the scale of the crust, its composition, the thickness of its mechanically strong upper part and the availability of pre-existing crustal discontinuities, which can be tensionally reactivated, play a dominant role in the width and deformation mode of an evolving rift.

5.2. Extensional strain

A major factor controlling the structural style of a rift zone is the magnitude of the crustal extensional strain that was achieved across it and the distance over which it is distributed (δ factor). Although quantification of the extensional strain and of the stretching factor δ is of basic importance for the understanding of rifting processes, there is considerable controversy about the reliability of estimates derived from upper crustal extension by faulting.

The amount of upper crustal extension across rift zones is usually derived from two-dimensional multichannel reflection–seismic data, by mapping fault-offsets at the base of the syn-rift sediments or within pre-rift sediments. Only in rare cases is partial coverage by three-dimensional seismic data available, providing a much higher resolution of fault systems. In this respect, the resolving power of unmigrated and migrated reflection–seismic data must be kept in mind. Moreover, with increasing depth, the resolution of small faults decreases due to the progressive absorption of high frequencies and the enlargement of the Fresnel zone of the seismic wavefront. Therefore, extension values derived from two-dimensional reflection–seismic data are minimum values. However, it is unlikely that in terms of basin-wide extension, offsets on small faults can account for a doubling or even a quadrupling of the cumulate heave on major faults (Barton and Wood, 1984; Shorey and Sclater, 1988; Ziegler and van Hoorn, 1989). Although clay and sand model experiments suggest that extension by major faulting may account for only 30–50% of the total extensional strain (Kautz and Sclater, 1988), statistical analyses on fault populations in rifted basins indicate that heaves on large-scale faults account for at least 60–70% of the total strain (Walsh et al., 1991; Yielding et al., 1992). Only in rare cases is there evidence for a second generation of normal faults cross-cutting an earlier extensional fault system (Stiros, 1991), significantly disturbing this empirical relationship. On the other hand, the dip-slope angle of fault blocks, and their synthetic or antithetic arrangement, can be generally readily mapped on reflection–

seismic data and provides an indirect measure of the amount of extension along the controlling planar or listric normal faults and the depth at which these sole out (McClay, 1989; White, 1990; Dula, 1991; McClay and Scott, 1991; Artyushkov, 1992). However, a potential major source of error in defining the amount of upper crustal extension is erosion of the crustal parts of intrabasinal footwall uplifts and particularly of rift shoulders (Kusznir and Egan, 1989; Kusznir et al., 1991; Kusznir and Ziegler, 1992).

Even if these considerations are taken into account, there remains in many rifted basins a significant discrepancy between upper crustal extension by faulting and the amount of extension inferred from their crustal configuration (see Table 1).

5.3. Fault geometries

The geometry of the fault systems of a rift zone reflects whether it evolved in response to orthogonal or oblique-slip divergence of its flanking stable blocks (Fig. 14; Tron and Brun, 1991; Rosendahl et al., 1992; Ben-Avraham, 1992). Classical examples of orthogonal crustal extension are the interior Sudan grabens (McHargue et al., 1992), and of oblique crustal stretching the western branch of the East African rift system (Rosendahl et al., 1992). However, it must be kept in mind that during the evolution of a rift, the stress regime governing its development may change. In time, orthogonal extension can give way to transtensional faulting and even to transpressional deformation (e.g. North Sea rift: Ziegler, 1990; Rhine Graben: Illies, 1975, 1981; Sunda Shelf: Letouzey et al., 1991; Central African Doba and Doseo grabens: Genik, 1992; East African rift: Delvaux, 2001), and vice versa (e.g. Gulf of Suez: Ott d'Estevou et al., 1989; Bosworth and McClay, 2001), causing a modification of the structural style (Keep and McClay, 1997).

Most rifts consist of a system of half-graben shaped basins, the polarity of which often changes along trend across accommodation (transfer) zones (Gibbs, 1984; Etheridge et al., 1985; Reynolds and Rosendahl, 1984; Rosendahl et al., 1992). The latter are characterized by complex fault geometries involving local positive and negative flower-structures and folding (Rosendahl, 1987; Morley et al., 1990). In plan view, master faults of half grabens often display a curvilinear geometry, indicative of their listric configuration. The sole-out level of such faults is thought to correspond to the crustal brittle–ductile transition zone below which deformation of the crust is dominated by ductile shear (Gibbs, 1987, 1989; Kusznir and Egan, 1989). Progressive strain gives rise to the rotational subsidence of individual fault blocks. This accounts for the accumulation of wedge-shaped sedimentary units, which expand towards the footwall fault and thin towards the leading edge of the hanging-wall blocks (Nøttvedt et al., 1995; ter Voorde et al., 1997). As long as the lithosphere retains a certain amount of strength during rifting, the subsidence and uplift pattern of such fault blocks conforms closely to the flexural cantilever model (Kusznir et al., 1991; Kusznir and Ziegler, 1992). The level of (partial) decoupling between the upper crust and the mantle–lithosphere depends on the rheological stratification of the extended lithosphere (ter Voorde et al., 1998, 2000), which can be readily derived from the geometry of the syn-rift sedimentary series (ter Voorde and Cloetingh, 1996; ter Voorde et al., 1997). An extreme form of listric faulting, controlling the development of core complexes, can occur during extension of orogenically thickened and thermally destabilized crust at high strain rates and stretching factors, involving ductile flow of the middle and lower crust (e.g. Basin and Range Province, δ up to 1.6–1.77: Wernicke and Tilke, 1989; Wernicke, 1990; Parsons, 1995; Tyrrhenian Sea: Bertotti et al., 2000; Aegean Sea, δ $\pm$ 1.9: Gautier et al., 1999).

Deviations from listric extensional fault geometries can be generally related to non-orthogonal extension (Tron and Brun, 1991; Ben-Avraham, 1992; Naylor et al., 1996) or to the tensional reactivation of pre-existing crustal discontinuities. However, the occurrence of planar faults in rift zones, penetrating the entire seismogenic zone, has been advocated by Jackson and McKenzie (1983), Jackson and White (1989), Kusznir et al. (1991) and Westway (1991), and is documented by earthquake data from the Basin and Range Province (Stein and Barrientos, 1985), the Aegean (Jackson, 1987) and the Gulf of Suez rift (Stampfli, personal communication). Mixed planar and listric fault geometries are typical for many rifts (e.g. Sudan rifts: McHargue et al., 1992).

Under conditions of the simple-shear extension, the lithospheric configuration and structural style of conjugate margins can differ considerably at the end of the rifting stage (Fig. 5). At lower plate margins, the

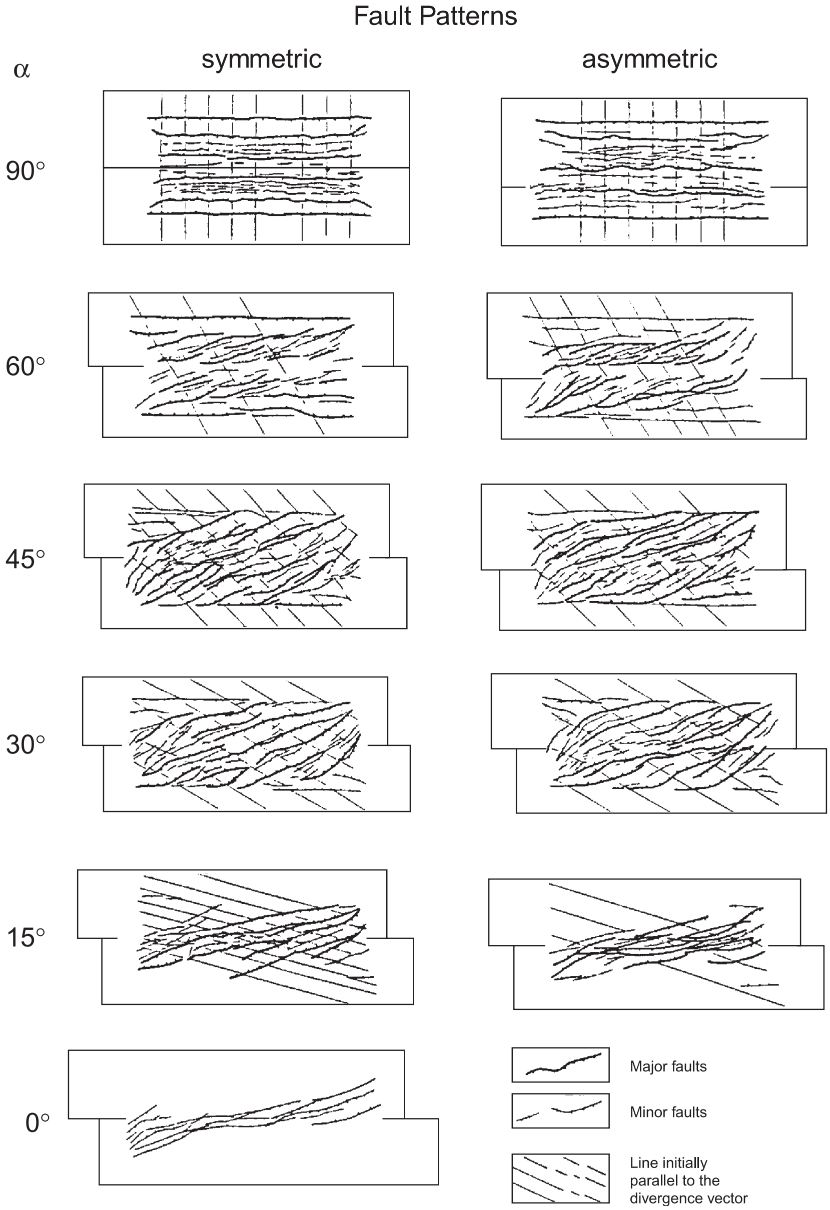

Fig. 14. Model experiments on orthogonal and oblique extensional fault patterns. α, stretching angle (Tron and Brun, 1991).

crust can be highly extended and dominated by elongated rotational fault blocks, whereas the mantle–lithosphere may be little attenuated; in their distal parts, denuded mantle–lithosphere can be in sheared contact with rotated upper crustal fault blocks and syn-rift sediments. By contrast, at upper plate margins, the crust may be less extended and involved in a flexural role-over structure, whereas the mantle–lithosphere can be strongly attenuated with asthenospheric material ascending close to the base of the crust (Wernicke, 1985; Wernicke and Tilke, 1989; Boillot et al., 1989; Kusznir et al., 1991; Lister et al., 1991; Piqué and Laville, 1995; Brun and Beslier, 1996). This contributes towards the frequently observed asymmetry of conjugate passive margins (Favre and Stampfli, 1992; Sibuet, 1992) and to differences in their post-rift subsidence pattern (Ziegler et al., 1998, 2001; Stampfli et al., 2001).

At shallow levels, the structural style of a rift is strongly influenced by the lithological composition of the down-faulted pre-rift and syn-rift sediments. Halites and shales can give rise to the development of multilevel extensional detachment faulting (Fig. 15; Jarrige et al., 1990; Jarrige, 1992) and can act also as sole-out levels for secondary fault systems developing in response to gravitational instability of accumulated sediments (e.g. slumping along fault scarps; Nøttvedt et al., 1995). Ramping of extensional faults at sedimentary and intrabasement levels can give rise to complex deformation patterns in hanging wall blocks, involving folding and the development of secondary tensional fault systems (McClay and Scott, 1991; e.g. Jeanne d'Arc Basin on Newfoundland shelf: Tankard et al., 1989; Gulf of Thailand: Watcharanantakul and Morley, 2000). Moreover, displacements along deep-seated faults can dissipate and attenuate upwards in clays and thick salt layers, giving rise to the development of detachment surfaces and local forced folds (flexures) over the leading edges of fault blocks (Withjack et al., 1989, 1990; Naylor et al., 1996). Faulting and tilting of the basin floor, causing differential sediment loading of thick salt layers, commonly triggers their halokinetic deformation, which in turn overprints extensional fault geometries at post-salt levels (Geil, 1991; Zaleski and Julien, 1992; e.g. pre-rift salt: Permian in North Sea: Hospers et al., 1988, Ziegler, 1990; syn-rift salt: Triassic–Jurassic Grand Banks area: Balkwill and Legall, 1989; Devonian Dniepr–Donets: Stephenson et al., 2001; early post-rift salt: Late Aptian, Gabon and Angola shelf: Teisserenc and Villemin, 1990; Duval et al., 1992; Campos basin: Guardado et al., 1990).

6. Duration of rifting stage

The duration of the rifting stage of intracontinental rifts (aborted) and passive margins (successful rifts) is highly variable (Figs. 2 and 3; Ziegler, 1990; Ziegler et al., 2001). Overall, it is observed that in time, rifting activity concentrates on the zone of future crustal separation with lateral rift systems becoming inactive. However, as not all rift systems progress to crustal separation, the duration of their rifting stage is obviously a function of the persistence of the controlling stress field. On the other hand, the time required to achieve crustal separation is a function of the strength (bulk rheology) of the lithosphere, the build-up rate, magnitude and persistence of the extensional stress field, constraints on lateral movements of the diverging blocks (on-trend coherence, counter-acting far-field compressional stresses), and apparently not so much of the availability of pre-existing crustal discontinuities that can be tensionally reactivated.

6.1. Role of crustal discontinuities

Crustal separation was achieved in the Liguro-Provençal Basin after 9 My of crustal extension (Roca,

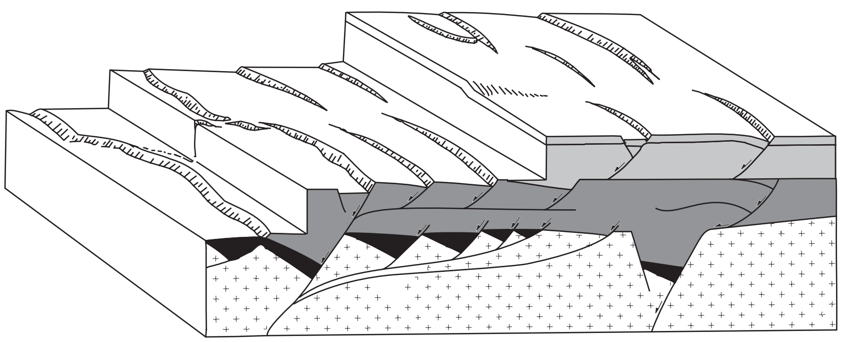

Fig. 15. Model of ramp-flat extensional faulting, employing pre-existing basement (cross pattern) discontinuities and detachment horizons within the sedimentary sequence (stippled) (Jarrige, 1992).

2001) and in the Gulf of California after about 14 My of rifting (Dokka and Merriam, 1982; Coney, 1987), whereas opening of the Norwegian Greenland Sea was preceded by an intermittent rifting history spanning some 280 My (Ziegler, 1988). There appears to be no obvious correlation between the duration of the rifting stage (R) of successful rifts, which are superimposed on orogenic belts (Liguro-Provençal Basin, Pyrenees $R = 9$ My; Gulf of California, Cordillera $R = 14$ My; Canada Basin, Inuitian fold belt $R = 35$ My; Central Atlantic, Appalachians $R = 42$ My; Norwegian Greenland Sea, Caledonides $R = 280$ My) and those which developed within stabilized cratonic lithosphere (southern South Atlantic $R = 13$; northern South Atlantic $R = 29$ My; Red Sea $R = 29$ My; Baffin Bay $R = 70$ My; Labrador Sea $R = 80$ My; Fig. 2). This suggests that the availability of crustal discontinuities, which regardless of their age (young orogenic belts, old Precambrian shields) can be tensionally reactivated, does not play a major role in the time required to achieve crustal separation. However, by weakening the crust, such discontinuities play a role in the localization and distribution of crustal strain. Moreover, by weakening the lithosphere, they contribute to the preferential tensional reactivation of young as well as old orogenic belts (Janssen et al., 1995; Ryan and Dewey, 1997; Ziegler et al., 2001).

6.2. Stress regime and strain concentration

Modifications in the drift-pattern of lithospheric plates, and consequently their interaction, can give rise to changes in the stress regime governing the subsidence of the different branches of a rift system. This can result in a transition from orthogonal to oblique extension or even to transpressional wrench deformation, causing partial inversion of rifts, and vice versa. For example, Senonian changes in the drift pattern of Africa were associated with a reorientation of the stress regime governing the evolution of the Cretaceous–Paleogene Central African rift systems, resulting in inversion of, e.g. the Benue, Doba and Doseo troughs, whereas the related Niger, Chad and Sudan grabens continued to subside (Castro, 1987; Genik, 1992; McHargue et al., 1992; Guiraud et al., 1992; Guiraud and Bosworth, 1997). A further example is provided by the latest Jurassic–earliest Cretaceous rifting pulse of the North Sea that was accompanied by the local development of transpressional structures indicative for oblique extension (Ziegler, 1990) rather than for gravity-driven crustal shortening as advocated by Knott (2001). In the North Sea, the change from Triassic–Middle Jurassic orthogonal to Late Jurassic–Early Cretaceous oblique extension reflects a reorientation of the stress field, which resulted from Mid-Jurassic crustal separation and the onset of sea-floor spreading in the Alpine Tethys (Ziegler, 1988, 1990; Stampfli et al., 2001; Ziegler et al., 2001).

Furthermore, during the evolution of a mega-rift system, tensional strain will in time concentrate on the zone of future crustal separation in response to rising isotherms and increasing strain rates (Kusznir and Park, 1987; Sawyer and Harry, 1991; Ellam, 1991). This can be accompanied by a gradual reduction of tectonic activity in marginal graben systems, as seen in the Central Atlantic rift system (Favre and Stampfli, 1992), the area of the southern Barents Sea and on the Mid-Norway Shelf (Ziegler, 1988; Lundin and Doré, 1997; Swiecicki et al., 1998; Reemst and Cloetingh, 2000). In the North Sea rift, strain concentration to its axial rift system is evident during the Late Jurassic–earliest Cretaceous phase of accelerated crustal stretching. During the Cretaceous, crustal extension rates decreased rapidly in the North Sea as extensional strain concentrated on the Arctic–North Atlantic rift system (Ziegler, 1990). On the other hand, the grabens of northeastern Brazil remained tectonically active until shortly before crustal separation was achieved in the northern South Atlantic (Castro, 1987; Chang et al., 1992; Karner et al., 1992).

We conclude that the duration of the rifting stage of a rift system depends on the persistence of the regional stress regime governing its evolution. Stress regime changes can be a function of extensional strain concentration to the future zone of crustal separation, of crustal separation occurring in genetically related rift systems (e.g. North Sea and Norwegian–Greenland Sea rift), or of fundamental changes in plate motions and interaction (e.g. Oslo Graben, Pyrenean rift, West and Central African rift system; Ziegler, 1992; Ziegler et al., 2001). The time required to achieve crustal separation is largely a function of constraints on lateral movements of the diverging blocks and, thus, of plate interaction.

7. Post-rift subsidence

Similar to the subsidence of oceanic lithosphere, the post-rift subsidence of extensional basins is mainly governed by thermal relaxation and contraction of the lithosphere, resulting in a gradual increase of its flexural strength, and by its isostatic response to sedimentary loading. Theoretical considerations indicate that subsidence of post-rift basins follows an asymptotic curve, reflecting the progressive decay of the rift-induced thermal anomaly, the magnitude of which is thought to be directly related to the lithospheric stretching value (Sleep, 1973, 1976; McKenzie, 1978; Sclater and Tapscott, 1979; Jarvis and McKenzie, 1980; Royden et al., 1980; Le Pichon and Sibuet, 1981; Beaumont et al., 1982a,b; Steckler and Watts, 1982; Watts et al., 1982; Watts and Thorne, 1984; Morgan and Ramberg, 1987; White and McKenzie, 1988; Thorne and Watts, 1989; Bott, 1992b). During the post-rift evolution of a basin, the thermally destabilized continental lithosphere reequilibrates with the asthenosphere (McKenzie, 1978; Steckler and Watts, 1982; Wilson, 1993a). In this process, in which the temperature regime of the asthenosphere plays an important role (ambient, below or above ambient), new mantle–lithosphere consisting of solidified asthenospheric material is accreted to the attenuated old continental mantle–lithosphere (Ziegler et al., 1998). In addition, densification of the continental lithosphere involves crystallization of melts that accumulated at its base or were injected into it, subsequent thermal contraction of the solidified rocks and under certain conditions their phase transformation to eclogite facies. The resulting negative buoyancy effect is the primary cause of post-rift subsidence. However, in a number of basins, significant departures from the theoretical thermal subsidence curve are observed. These can be explained as effects of compressional intraplate stresses and related phase transformations (Cloetingh and Kooi, 1992; Stel et al., 1993; van Wees and Cloetingh, 1996; Lobkovsky et al., 1996).

7.1. Shape and magnitude of thermal anomalies

The shape and dimensions of the rift-induced asthenosphere–lithosphere boundary anomaly controls to a large extent the geometry of the evolving post-rift thermal sag basin (Fig. 5). Thermal-sag basins associated with aborted rifts are broadly saucer-shaped and generally overstep the rift zone; their axes are thought to coincide with the zone of maximum lithospheric attenuation.

Pure-shear dominated rifting gives rise to the classical "steer's head" configuration of the syn- and post-rift basins (White and McKenzie, 1988) in which both basin axes roughly coincide, with the post-rift basin broadly overstepping the rift flanks. This geometric relationship between syn- and post-rift basins is frequently observed (e.g. North Sea Rift: Ziegler, 1990; West Siberian Basin: Artyushkov and Baer, 1990; Dniepr–Donets Graben: Stephenson et al., 2001; Gulf of Thailand: Hellinger and Sclater, 1983; Watcharanantakul and Morley, 2000; Sudan rifts: McHargue et al., 1992). Such a geometric relationship is compatible with discontinuous, depth-dependent stretching models, which assume that the zone of crustal extension is narrower than the zone of mantle–lithosphere attenuation (Fig. 4). The degree to which a post-rift basin oversteps the margins of the syn-rift basin is a function of the difference in width between the zone of upper crustal extension and the zone of mantle–lithosphere attenuation (White and McKenzie, 1988).

Conversely, simple-shear dominated rifting gives rise to a lateral offset between the syn- and post-rift basin axes. An example of this is given by the Tucano Graben of northeastern Brazil, which ceased to subside at the end of its rifting stage, whereas the coastal Jacuipe–Sergipe–Alagoas Basin, to which the former was structurally linked, was the site of crustal and mantle–lithospheric thinning culminating in crustal separation (Chang et al., 1992). Moreover, the simple-shear model can explain the frequently observed asymmetry of conjugate passive margins and differences in their post-rift subsidence pattern (e.g. Central Atlantic and Red Sea: Favre and Stampfli, 1992). Discrepancies in the post-rift subsidence of conjugate margins is attributed to differences in their lithospheric configuration at the crustal separation stage. At the end of the rifting stage, lower plate margins are supported by a relatively thick mantle–lithosphere whereas upper plate margins are underlain by strongly attenuated mantle–lithosphere and partly directly by the asthenosphere. Correspondingly, lower plate margins are associated with smaller thermal anomalies than upper plate margins (Ziegler et al., 1998; Stampfli et al., 2001).

The magnitude of post-rift tectonic subsidence of aborted rifts and passive margins, which can be attributed to cooling and contraction of the lithosphere and its return to thermal equilibrium with the asthenosphere, is a function of the magnitude of the thermal anomaly that was introduced during their rifting stage. Most intense thermal anomalies develop during crustal separation, particularly when plume assisted, when hot asthenospheric melts well up close to the surface. Thermal anomalies induced by rifting that did not progress to crustal separation are generally smaller. Their magnitude depends on the magnitude of crustal stretching (δ-factor) and mantle–lithospheric attenuation (β-factor), the thermal regime of the asthenosphere, the volume of melts generated and/or whether these intruded the lithosphere and destabilized the Moho (Fig. 6).

It follows that the post-separation evolution of passive continental margins probably reflects the decay of the largest thermal anomalies, whereas the subsidence of aborted rifts is governed by the decay of smaller anomalies. On the other hand, the post-rift development of volcanic rifts is likely to reflect the decay of larger thermal anomalies than those of nonvolcanic rifts. After 60 My about 65%, and after 180 My about 95% of a deep-seated thermal anomaly associated with a major pull-up of the asthenosphere–lithosphere boundary will have decayed (mantle–plume model). Thermal anomalies related to intralithospheric intrusions (Fig. 6; asthenoliths; tensional-failure model) have apparently a faster decay rate. For instance, the Mid-Jurassic North Sea rift dome had subsided below the sea-level 20–30 My after its maximum uplift, that is well before crustal extension had terminated (Ziegler, 1990; Underhill and Partington, 1993).

7.2. Can stretching factors be derived from quantitative subsidence analyses?

The thickness of the post-rift sedimentary column that can accumulate in passive-margin basins and in thermal sag basins developing above aborted rifts is not only a function of the magnitude of the mantle–lithospheric attenuation factor β, but also of the crustal attenuation factor δ, the crustal density and the water depth at the end of their rifting stage, as well as of the density of the infilling post-rift sediments (carbonates, evaporites, clastic rocks). Moreover, it must be kept in mind that during the post-rift cooling process, the flexural rigidity of the lithosphere increases gradually, resulting in distribution of the sediment load over progressively wider areas.

Quantitative analyses of the post-rift subsidence of extensional basins, assuming the absence of intraplate stresses, are thought to give a measure of the thermal contraction of the lithosphere and, conversely, of the magnitude of the thermal anomaly that was introduced during their rifting stage. Such analyses, assuming Airy isostasy, yield often considerably larger δ-factors than indicated by the populations of upper crustal faults (Ziegler, 1983, 1990; Watcharanantakul and Morley, 2000). This discrepancy is somewhat reduced when flexural isostasy is assumed (Roberts et al., 1993). However, as during rifting, attenuation of the lithosphere is not only achieved by its mechanical extension but also by convective and thermal upward displacement of the asthenosphere–lithosphere boundary, the magnitude of a thermal anomaly derived from the post-rift subsidence of a basin can probably not be directly related to a mechanical stretching factor. Moreover, intraplate compressional stresses and phase transformations in the lower crust and mantle–lithosphere can have an overprinting effect on post-rift subsidence and can cause significant departures from a purely thermal subsidence curve (Cloetingh and Kooi, 1992; van Wees and Cloetingh, 1996; Lobkovsky et al., 1996).

Furthermore, crustal extension can be associated with syn-rift density changes of the crust in response to its magmatic underplating, destabilization of the Moho, injection of mantle-derived melts into the crust, its metasomatic reactivation and secondary differentiation (Morgan and Ramberg, 1987; Mohr, 1992; Stel et al., 1993; Watts and Fairhead, 1997). For instance, the highly attenuated crust of the Devonian Dniepr–Donets rift is characterized by lower crustal velocities (6.8–7.2 km/s) and densities (3.02×10^{-3} kg m^{-3}) almost up to the base of the syn-rift sediments, probably due to its syn-rift permeation by mantle-derived melts (Fig. 16; Yegorova et al., 1999; Stephenson et al., 2001).

As rifting processes can take place intermittently over very long periods of time (e.g. Norwegian–

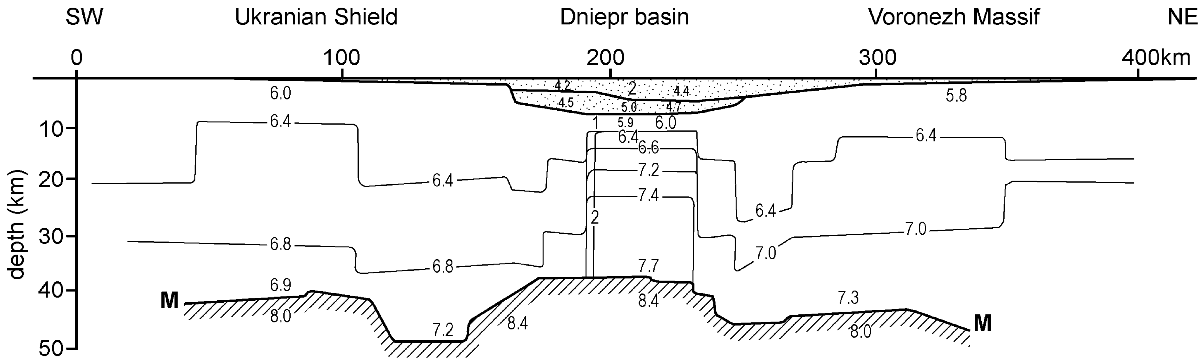

Fig. 16. Crustal configuration of the Dniepr and Donets basins, showing P-wave velocity distribution. Stippled interval: sediments; barbed line marked with M: Moho discontinuity (Stephenson et al., 2001).

Greenland Sea rift: Ziegler, 1988), thermal anomalies introduced during early lithospheric stretching phases start to decay during subsequent periods of decreased extension rates. Thus, the thermal anomaly associated with a rift may not be at its maximum when crustal stretching terminates and the rift becomes inactive. Similarly, late rifting pulses and/or regional magmatic events may interrupt and even reverse lithospheric cooling processes (e.g. Palaeocene thermal uplift of northern parts of Viking Graben due to Iceland plume impingement: Ziegler, 1990; Nadin and Kusznir, 1995). Therefore, analyses of the post-rift subsidence of rifts that have evolved in response to repeated rifting phases or a long period of more or less continuous crustal extension must take into account their entire rifting history.

A major compounding effect is related to intraplate stresses, which induce deflection of the lithospheric, thus overprinting the thermal subsidence of a basin. The example of the Plio-Pleistocene evolution of the North Sea shows that the build-up of regional compressional stresses can cause a sharp acceleration of post-rift subsidence (Fig. 17; Cloetingh, 1988; Cloetingh et al., 1990; Kooi and Cloetingh, 1989, Kooi et al., 1991; van Wees and Cloetingh, 1996). Contemporaneous similar effects are recognized in North Atlantic passive margin basins (Cloetingh et al., 1990). Also, the late Eocene accelerated subsidence of the Black Sea Basin can be attributed to the build-up of a regional compressional stress field (Robinson et al., 1995; Cloetingh et al., 2003). Similarly, the sharp subsidence acceleration of the West Siberian Basin at the Jurassic–Cretaceous transition (Nikishin et al., 2002) may be related to the build-up of intraplate compressional stresses during the evolution of the Verkhoyansk–South Taimyr orogen (Zonen-

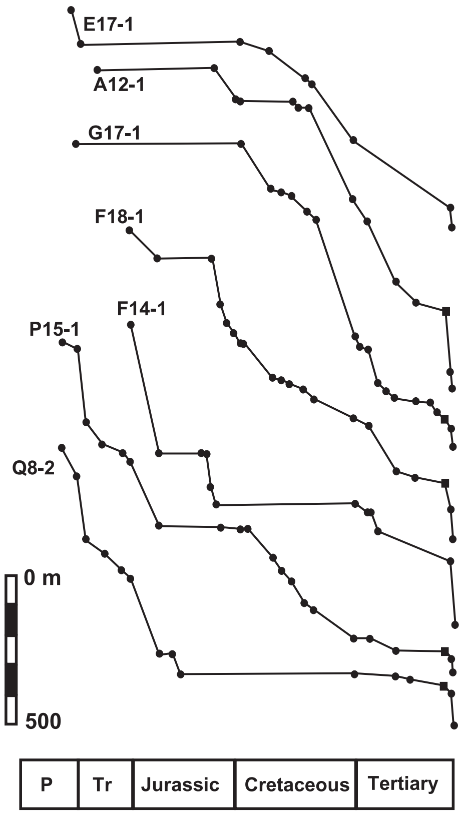

Fig. 17. Tectonic subsidence curves of southern North Sea, showing accelerated subsidence during Plio-Pleistocene. Post-rift stage starts during the Cretaceous (Kooi et al., 1991).

shain et al., 1993). The present-day World Stress Map demonstrates the existence of large-scale, consistently oriented stress patterns in the lithosphere (Zoback, 1992). Palaeo-stress analyses give evidence for changes in magnitudes and orientations of stress fields on time scales of a few My (Letouzey, 1986; Philip, 1987; Bergerat, 1987). Thus, in an attempt to understand the evolution of a post-rift basin, the effects of tectonic stresses on subsidence must be separated from those related to thermal relaxation of the lithosphere (Cloetingh and Kooi, 1992).

Post-rift cooling and contraction of the lithosphere and its gradual return to thermal equilibrium with the asthenosphere can be accompanied by the accumulation of very thick sedimentary sequences. Upon sufficient burial, phase transformation of crustal rocks to granulite-facies, resulting in densification of the crust, could also account for rapid basin subsidence. This process could be amplified in the cases of eclogite transformation of basaltic melts that were injected into the mantle–lithosphere or that had accumulated at its base (Lobkovsky et al., 1996). Granulite to eclogite transformation at the base of the crust, possibly involving an upward displacement of the geophysically defined Moho discontinuity, has been advocated by Artyushkov and Baer (1990) to explain the subsidence acceleration of the West Siberian Basin at the Jurassic/Cretaceous transition (Artyushkov, 1992). However, it is uncertain whether large-scale eclogite formation can indeed occur at crustal thicknesses of 30–40 km (Carswell, 1990; Griffin et al., 1990). The transformation of mafic and felsic rocks to granulite- and eclogite-facies is known to be controlled to a large extent by pressure and temperature (Ringwood and Green, 1966; Ito and Kennedy, 1971) as well as by the presence of hydrous fluids, CO_2 and nitrogen (Fowler and Nisbet, 1985; Austrheim, 1990; Austrheim and Mørk, 1988; Pearson and O’Reilley, 1991; Touret, 1992; Vapnik, 1993, Andersen, 1993; Austrheim et al., 1997). Although phase transformations, entailing an upward displacement of the geophysical Moho, are thought to occur under certain conditions at the base of very thick Proterozoic cratons (increased confining pressure due to horizontal intraplate stresses and/or ice load, possible cooling and local downwelling of asthenosphere, Cloetingh and Kooi, 1992), it is still uncertain whether physical conditions conducive to such transformations can develop commonly in palaeo-rifts and passive margins. Nevertheless, it is suspected that sharp deviations from theoretical exponential thermal decay curves, defined by quantitative subsidence analyses, may not be exclusively related to deflection of the lithosphere in response to the build-up of intraplate compressional stresses, but that under the load of thick post-rift sediments conditions for transformation of crustal rocks and/or underplated mantle-derived basalts to granulite- and ultimately to eclogite-facies may be a contributing factor (e.g. East Newfoundland Basin: Cloetingh and Kooi, 1992; West Siberian Basin: Lobkovsky et al., 1996).

A special problem is presented by grabens which are located marginal to main rift systems and which are characterized by the lack of post-rift or post-crustal separation subsidence, or which were even uplifted after their rifting stage. Examples are the deeply truncated Permo-Triassic half-grabens on the Hebrides Shelf (Cheadle et al., 1987; Duindam and van Hoorn, 1987), the Triassic–Early Jurassic grabens of the Appalachian Piedmont (Manspeizer, 1988), the Reconcavo, Tucano and Jatoba grabens of northeastern Brazil (Chang et al., 1992; Karner et al., 1992) and the Late Palaeozoic–Mesozoic Jameson Land Basin of central East Greenland (Ziegler, 1988). This phenomenon could be explained either in terms of the simple-shear model, involving upper-crustal extensional unloading and long-wavelength flexural isostatic rebound of the lithosphere as a consequence of the zone of upper-crustal extension being offset from the zone of lower crustal and mantle–lithosphere thinning (e.g. Tucano graben: Ussami et al., 1986; Karner et al., 1992), by ductile flow of lower crustal material away from the axial rift zone to peripheral areas (e.g. Arabian flank of Red Sea: Moretti and Pinet, 1987; Basin and Range: Jones et al., 1992), or by magmatic underplating of the rift-flanks (Rockall-Hatton Bank: White, 1992). In the case of East Greenland, a post-rift uplift by 1.5–2 km has been determined (Dam and Christiansen, 1990; Christiansen et al., 1992; Larsen and Marcussen, 1992) that may be related to plume-support of the lithosphere (Rohrman, 1995; Rohrman et al., 1995; Bijwaard and Spakman, 1999). As these processes are not mutually exclusive, their combined effects may have to be taken into consideration to explain the lack of post-rift subsidence of some grabens. Erosion of rift-flanks and particularly intraplate compressional stresses can play an important role in the post-rift uplift of marginal grabens (Cloetingh and Kooi, 1992). Moreover, inversion of rifted basins during their post-rift stage is essentially controlled by the build-up of far-field compressional stresses (Ziegler, 1990; Ziegler et al., 1995, 1998) and probably only under special conditions by gravitational forces associated with the topography created around the basin (Bada et al., 2001).

A particularly baffling case is the Oslo-Skagerrak Graben, which lacks a thick syn-rift sedimentary package, although voluminous basalts are present within the rift zone, and which shows little evidence for the development of a post-rift sedimentary basin (Ro et al., 1990; Kinck et al., 1991; Ro and Faleide, 1992; Neumann et al., 1995). The configuration of the Oslo Rift precludes that this can be attributed to simple-shear deformation nor to major magmatic underplating of the rift zone. Triassic post-rift sediments are only present in the Sagerrak Graben and are locally preserved in the northern parts of the Oslo Graben. Results of fission-track analyses indicate that some 50 My after the end of its rifting stage the Oslo Graben area was uplifted by 3–4 km during the Late Triassic and Jurassic and subjected to erosion, resulting in the destruction of its post-rift basin and some of its syn-rift fill (Rohrman et al., 1994). Dynamic processes governing this post-rift uplift are still not fully understood but could be related to regional uplift of the western parts of the Fennoscandian Shield in conjunction with the evolution of the Arctic–North Atlantic rift system (Ziegler, 1988, 1990).

These considerations cast doubt on the validity of stretching factors that are derived from the subsidence of post-rift basins. Firstly, stress-induced effects must be removed from back-stripped subsidence curves before the thermal anomaly present at the end of the rifting stage can be quantified. Secondly, it is suspected that the magnitude of a thermal anomaly derived from a post-rift subsidence curve cannot be directly related to a crustal stretching factor since the rift-induced disturbance of the lithosphere depends on the thermal state of the asthenosphere, the presence of volatiles and probably also on extension rates. High strain rates provide presumably for larger thermal anomalies than low extension rates (syn-rift conductive and convective heat diffusion). Moreover, thermal and convective thinning of the mantle–lithosphere amplifies the extension-induced lithospheric anomaly. Density decrease of the crust due to partial melting and secondary differentiation and/or density increase of the crust due to injection of mantle-derived melts are likely to influence its post-rift subsidence. In addition, magmatic destabilization of the Moho discontinuity during the rifting stage can cause thinning of the crust that is only indirectly related to its mechanical stretching. Finally, phase transformations of underplated basalts may play an important role during post-rift basin subsidence. Despite this, quantitative subsidence analyses, combined with other data, contribute substantially to the understanding of post-rift subsidence processes by giving a measure of the thermal anomaly that was introduced during the rifting stage of a basin and by identifying deviations from purely thermal subsidence patterns.

8. Conclusions

The principal driving mechanism of plate movements and lithospheric extension, governing the evolution of rifts and the breakup of continents, is a combination of frictional forces exerted by the convecting asthenosphere on the base of the lithosphere, deviatoric tensional stresses developing over mantle upwellings and far-field stresses related to plate boundary processes (Bott, 1993; Ziegler, 1993; Ziegler et al., 2001). Plate interaction plays a major role in the evolution of rifts. Geochemical criteria indicate that deep-mantle plumes, rising from the $\pm$670 km discontinuity or the core/mantle boundary, play a subordinate role in the evolution of most rift systems; however, they account for high syn-rift volcanic activity. Moreover, by causing weakening of the lithosphere, hot spots may contribute to the localization of a rift system as seen for instance in the South Atlantic rift system (Wilson and Guiraud, 1992), the evolution of the Cenozoic East African Rift (Braile et al., 1995) and the Mid-Proterozoic Midcontinent Rift (Cannon and Hinze, 1992; Allen et al., 1995).

Stretching of the lithosphere can cause by adiabatic decompression partial melting of the lower lithosphere and upper asthenosphere. The thermal state of the sublithospheric mantle and the availability of volatiles play an important role in the volume of melts generated (McKenzie and Bickle, 1988; Wilson, 1993a,b). Upwelling of the asthenosphere into the space provided by lithospheric stretching can cause thermal and convective thinning of the mantle–lithosphere. Segregated melts which ascended to the crust/mantle boundary interact with the lower crust and can cause its metasomatic reactivation and partial substitution, resulting in destabilization of the Moho discontinuity. Magmatic processes may explain the frequently observed discrepancy between crustal stretching factors derived from upper crustal faulting and crustal thicknesses. This requires a fundamental modification of conventional lithospheric stretching models that assume syn-rift volume conservation.

At the onset of rifting, the rheological structure of the lithosphere controls the location, structural style and width of the evolving rift system, as well as the necking depth of the lithosphere (Cloetingh et al., 1995). During rifting, uplift of the crust–mantle boundary in response to crustal extension causes initially a strength increase of the mantle–lithosphere; however, thermal attenuation of the mantle–lithosphere accompanied by rising isotherms results in progressive weakening of the lithosphere, upward displacement of the intracrustal brittle/ductile deformation transition zone and strain concentration on the rift axis. Progressive doming of rift zones is attributed to gradual thermal and mechanical substitution of the mantle–lithosphere by the asthenosphere; temporary doming can be caused by intrusion of melts at the crust/mantle boundary and their subsequent rapid cooling and contraction. Both processes give rise to the development of deviatoric tensional stresses that contribute towards extension of the lithosphere.

The duration of the rifting stage of passive margins and palaeo-rifts is highly variable. Interaction of lithospheric plates controls to a large extent the duration of the rifting stage and whether a rift succeeds in crustal separation or aborts at an earlier stage. Even after crustal separation, forces acting on a plate, controlling its drift pattern and interaction with other plates, can cause tensional and/or compressional reactivation of such crustal discontinuities as rifts (West and Central African rift system: Wilson and Guiraud, 1992).

The post-rift evolution of extensional basins is governed by cooling and contraction of lithosphere and its re-equilibration with the asthenosphere, involving phase transformations at mantle–lithospheric levels and, under special conditions perhaps at the base of the crust (Lobkovsky et al., 1996), and by compressional intraplate stresses (van Wees and Cloetingh, 1996). In view of this, and as syn-rift magmatic processes can cause a serious destabilization of the lithosphere–asthenosphere and the crust–mantle boundaries, stretching values, derived from the crustal configuration of a rift and from its post-rift subsidence, cannot be taken at face value.

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