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Review Article

The Moho in extensional tectonic settings: Insights from thermo-mechanical models

Sierd Cloetingh [a,*], Evgenii Burov [b,c], Liviu Matenco [a], Fred Beekman [a], François Roure [a,e], Peter A. Ziegler [d,†]

[a] Netherlands Research Centre for Integrated Solid Earth Sciences, Faculty of Geosciences, Utrecht University, Budapestlaan 4, 3584 CD Utrecht, The Netherlands
[b] UPMC Univ Paris 06, ISTEP UMR 7193, Université Pierre et Marie Curie, F-75005 Paris, France
[c] CNRS, ISTEP, UMR 7193, F-75005 Paris, France
[d] Kirchweg 41, 4102 Binningen, Switzerland
[e] IFPEN, 1-4 Avenue de Bois-Préau, 92852 Rueil-Malmaison Cedex, France

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ABSTRACT

The lithospheric memory is key for the interplay of lithospheric stresses and rheological structure of the extending lithosphere and for its later tectonic reactivation. Other important factors are the temporal and spatial migration of extension and the interplay of rifting and surface processes. The mode of extension and the duration of the rifting phase required to lead to continental break-up are to a large extent controlled by the interaction of the extending plate with slab dynamics. The finite strength of the lithosphere has an important effect on the formation of extensional basins. This applies both to the geometry of the basin shape as well as to the record of vertical motions during and after rifting. We demonstrate a strong connection between the bulk rheological properties of Europe's lithosphere and the evolution of some of Europe's main rifts and back-arc systems. The thermo-mechanical structure of the lithosphere has a major impact on continental break-up and associated basin migration processes, with direct relationships between rift duration and extension velocities, thermal evolution, and the role of mantle plumes. Compressional reactivation has important consequences for post-rift inversion, borderland uplift, and denudation, as illustrated by poly-phase deformation of extensional back-arc basins in the Black Sea and the Pannonian Basin region.

Contents

* Corresponding author. Tel.: +31 30 2537314; fax: +31 30 2535030.
E-mail address: sierd.cloetingh@uu.nl (S. Cloetingh).
† Deceased July 19th, 2013.

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1. Introduction

The Moho, originally defined as an acoustic contrast marking the compositional boundary between the crust and the underlying mantle, is also of crucial importance for the mechanical state of the lithosphere. Actually, the effect of the inherited crustal fabric and strength appear to be of paramount influence on the nature of tectonic deformation in the plate interiors. A particularly important parameter for lithospheric strength is the ratio of the thickness-dependent crustal and lithospheric mantle strength (e.g. Burov, 2011) and it is exactly this ratio that is strongly affected by crustal thinning during the rifting process. In addition, changes in thermal regime characteristic for rifting processes leave a strong imprint on the temporal and spatial variations in crustal and lithospheric strength during and after the rifting phases.

The notion that rifting is a very important mechanism for sedimentary basin formation with key implications for the basin fill and thermal regime is of immediate importance for the assessment of the hydrocarbon habitat. This has boosted rifting studies in the context of global exploration for hydrocarbons. Deep seismic profiling, often carried out in close collaboration between academia and industry (e.g. Thybo and Nielsen, 2009; Thybo et al., 2000), has been an essential component of these efforts. As a result, many data are now available to provide constraints on the timing and duration of the rifting phases and overall geometry of rift basins.

At the same time, advances in seismic tomography studies of the mantle structure at convergent plate boundaries, combined with case history analyses, have revealed a much stronger coupling between the subducting lower lithospheric plate and the overriding upper lithospheric plate that affects a much larger area of the latter than originally envisaged. The thermo-mechanical structure of the lithosphere is of fundamental importance for the interaction of both upper mantle and lithospheric processes (Burov and Cloetingh, 2009, 2010; Burov et al., 2007) as well as for the assessment of the horizontal tectonic stress impact on lithosphere deformation (Burov and Cloetingh, 2009; Cloetingh and Burov, 2011).

This paper reviews recent advances in modelling of the initiation and evolution of extensional basins in a lithospheric context. We first summarize key features of the architecture of rifted basins and examine the record of vertical motions during and after rifting in the context of stretching models developed to quantify the development of extensional basins. This will be followed by a discussion on the thermo-mechanical aspects of extensional sedimentary basin development in the context of large-scale lithosphere models for the underlying lithosphere and on tectonic controls on the post-rift evolution of extensional basins.

We highlight the connection between the bulk rheological properties of the lithosphere and the evolution of extensional basins and find that the finite strength of the lithosphere plays an important role in their development. This applies both to the geometry of the basin shape as well as to the record of vertical motions during and after lithospheric extension.

Below we give a brief review of concepts for the rheological structure of continental lithosphere and discuss insight gained from global models of strength distribution, highlighting the impact rifting processes have on the lithospheric strength distribution. This is followed by a discussion of constraints on the duration of rifting based on case studies (e.g. Ziegler and Cloetingh, 2004), the effects of slow versus fast extension on lithospheric strength and crustal configuration (van Wijk and Cloetingh, 2002; Van Wijk et al., 2004), and an analysis of the interaction between surface erosion and lower crustal flow in extensional regimes (see also Burov and Cloetingh, 1997; Burov and Poliakov, 2001). We then analyze the Pannonian basin and the Black Sea back-arc basins, integrating observational constraints on crustal structure and fabric provided by deep seismic profiles with inferences from thermo-mechanical models.

2. Constraints by deep seismic imaging, potential field data and mantle tomography on the architecture of the Moho, volcanic underplating and the occurrence of residual lithospheric slabs in rift basins, passive margins and intracontinental sags

The localization, initiation, development and current architecture of sedimentary basins and orogens directly rely on the thermo-mechanical behaviour of the continental lithosphere, which is in turn depending on the local thermal regime (i.e. lithosphere thickness or depth to the 1300 °C isotherm; Artemieva et al., 2006), and on the vertical and lateral distributions of the various lithologies encountered in the lower crust, crystalline basement and overlying sedimentary rocks (Allemand and Brun, 1991; Bertotti et al., 2000; Burov and Watts, 2006; Cloetingh and Banda, 1992; Davy and Cobbold, 1991; Kusznir and Karner, 1985; Huismans and Beaumont, 2003, 2008; Huismans et al., 2005; Ranalli, 1995; Ranalli and Murphy, 1987).

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At lithospheric scale, the main decoupling horizons operating at plate boundaries but also within lithospheric plates mostly comprise the Moho surface and the ductile lower crust, as well as pre-existing tectonic fabrics still preserved in the brittle crust, such as older low-angle extensional detachments or thrust systems which are likely to be positively or negatively inverted during subsequent compressional (Brooks et al., 1988; Colletta et al., 1997; Cooper and Williams, 1989; Coward, 1983; Gillchrist et al., 1987; Letouzey, 1990; Letouzey et al., 1990; Roure and Colletta, 1996; Van Wees, 1994; Vially et al., 1994) or extensional episodes, respectively (Burg et al., 1994; Dewey, 1988; McClay et al., 1986; Menard and Molnar, 1988; Platt and Vissers, 1989; Seguret et al., 1989; Séranne et al., 1989; van den Driessche and Brun, 1991).

2.1. Architecture of the Moho and lower crust in extensional settings, as revealed by deep seismic reflection profiles

In contrast to Moho depth predictions based on the inversion of potential data, deep seismic reflection profiles provide a direct image of the architecture of the Moho and lower crust beneath many rift basins and passive margins. Furthermore, refraction and tomographic data help also to constrain robust velocity models that can be used to convert time sections into depth sections, and to better control the current attitude of the Moho, as well as the local occurrence of anomalous high velocity/dense bodies that account for granulites in the lower crust, underplated volcanics that have been consolidated in the vicinity of the Moho interface, or even remnants of eclogitized lithospheric slabs that are still attached beneath the Moho.

Despite the fact that the initial stage of the lithosphere prior to the onset of rifting may be difficult to assess, at least the present day attitude of the Moho and lithosphere thicknesses documented by geophysical tools constitute unique data to test/validate thermo-mechanical models of lithosphere/crustal evolution in extensional systems. For instance, in the framework of the French Ecors program deep seismic profiles have been recorded in the eighties in extensional systems, in two different segments of the West European Oligocene rift system across the Rhine (Brun et al., 1991, 1992) and Bresse grabens (Bergerat et al., 1989, 1990), in the Bay of Biscaye across the Albian depocenter of the Parentis Basin (Bois and Gariel, 1994), and in the Gulf of Lions across the northern passive margin of the West Mediterranean Basin (Burrus, 1989; De Voogd et al., 1991; Séranne, 1999).

Also, the rapid development of petroleum exploration in the deep offshore by the industry has significantly promoted the acquisition of crustal-scale geophysical data on various segments of the conjugate passive margins of the Atlantic between Europe and North America (Etheridge et al., 1989; Faleide et al., 2008; Péron-Pinvidic and Manatschal, 2009; and references therein), and between Africa and South America (Castro, 1987; Franke et al., 2012; Moulin et al., 2010; Ueipas-Mohriak et al., 1995; Unternehr et al., 2010, and references therein), as well as across the conjugate passive margins of Australia and Antarctica (Direen et al., 2011; Espurt et al., 2012, and references therein), thus providing a clear picture of the current architecture of mature rifted margin systems with either volcanic-rich or volcanic-poor signatures. Despite the fact that their petroleum potential is more limited, young oceanic basins like the Gulf of California or the Gulf of Aden have also been studied recently by means of crustal-scale geophysical surveys (Leroy et al., 2012, and references therein).

Deep oceanic drilling off the Galicia margin (Boillot et al., 1988; Manatschal and Bernoulli, 1999; and references therein) and field studies in the Alps and the Pyrenees (Lagabrielle and Bodinier, 2008; Manatschal et al., 2001; Manatschal, 2004; Manatschal et al., 2011) have also contributed to control the lithologies and tectonic fabrics of lower crustal and upper mantle rocks which are still currently exposed at the continent–ocean transition in the deep offshore, or are instead preserved as allochthonous tectonic units still preserving the former COT (Continental–Oceanic Transition) in the inner part of these orogens.

Ultimately, COCORP profiles in the Basin and Range province (Allmendinger et al., 1983, 1987; Brown et al., 1986; Hauser et al., 1987) and combined field studies and seismic imagery in the Apennines (Scrocca et al., 2005; and references therein) and Alboran, Algerian basins and Tyrrhenian basins (Déverchère et al., 2013; Medaouri et al., in press; Ranero et al., 2013; Roure et al., 2009) have also contributed to better understand the change operating from compression to extension in over-thickened tectonic wedges and during the opening of back-arc basins.

2.2. Evidence for volcanic underplating from refraction data and inversion of potential field data

Rapid vertical and lateral changes can occur in the lithologies and seismic velocities within the upper and the lower crust, both beneath rift basins and passive margins. For instance, the consolidation of volcanic melts at or near the Moho interface can result in anomalously dense bodies at the base of the crust, likely to induce long-term subsidence, independently to temporary effects of lithospheric cooling and post-rift thermal subsidence. Seemingly, hydration/serpentinisation of denudated mantle at the Continent–Ocean Transition can lead to anomalously slow velocities for the mantle, which should not be misinterpreted as radiogenic lower crust, which is instead always characterized by seismic velocities higher than the brittle upper crust.

Time sections beneath the Parentis basin (Bois and Gariel, 1994; Fig. 1A) and the Dienpr–Donets (Stephenson et al., 2001; Fig. 1B) basin display a flat Moho, which would instead become shallower when the seismic profile is converted into a depth section in the case of the Parentis basin, due to the slow velocity of overlying Mesozoic sediments. In the case of the Dniepr–Donets basin. It is not clear, however, whether the amount of underplating and salt (high velocity rocks) is sufficient to balance the effect of slower velocity clastic and carbonate sediments, and therefore, whether the Moho surface is still ultimately flat, or would instead become uplifted or down-flexed once properly depth converted.

Therefore, it is of prime importance to define a robust velocity model at crustal scale when converting time images into depth images. In addition to refraction data that are frequently recorded along the same regional profiles as deep reflection seismic, the joint inversion of gravimetric and magnetic data is also frequently used to trace the distribution of dense bodies in the vicinity of the Moho and improve our overall understanding of the thermal state of the crust and overlying sedimentary basins, as for instance in the South Atlantic and Norwegian margins (Maystrenko et al., in press; Scheck-Wenderoth and Maystrenko, 2008) or the Asia continent (Stolk et al., 2013).

2.3. Long lasting subsidence of intracratonic sags induced by residual lithospheric slabs

Many passive margins (e.g. the South Atlantic and South Australia margins) are characterized by a long lasting sag episode, following major stretching and associated crustal tilted fault block formation. Although the controls of sag development are still debated, two obvious explanations can be proposed, depending on the assumptions made on the velocity models and the actual architecture of the Moho beneath the sags. The first explanation assumes that either the Moho remains flat beneath the sag basin or the dense anomalies relate to volcanic underplating, as suggested by Maystrenko et al. (in press) and Scheck-Wenderoth and Maystrenko (2008) accounting for a buried load beneath the sag. Alternatively, it is assumed that the Moho is uplifted beneath the sag, with the sedimentary loading in the sag resulting in a lateral flow of the lower crust away from the depocenters.

Last but not least, new tomographic models beneath the Paris Basin have documented a strong velocity anomaly in the upper mantle, which

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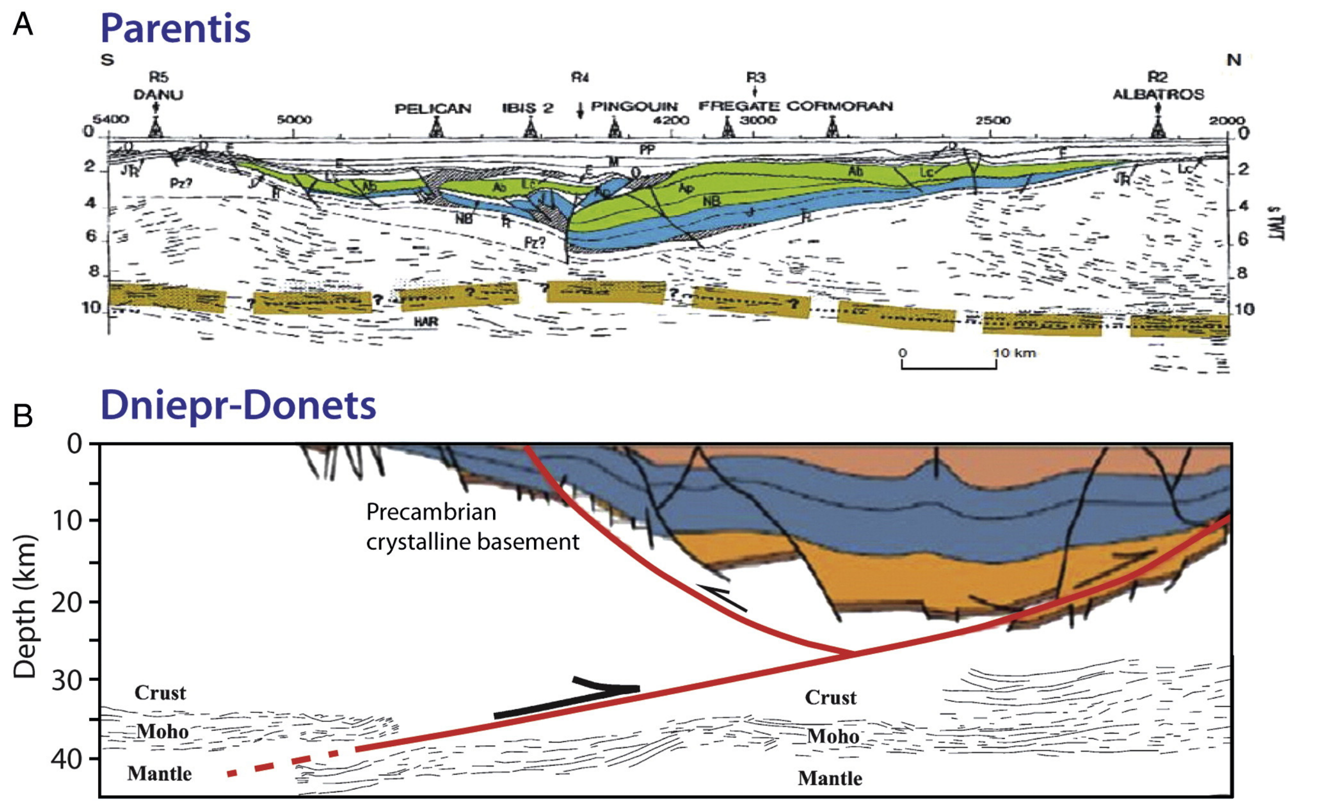

Fig. 1. A) Line drawing and simplified geological interpretation of the Ecors profile crossing the Parentis Basin. Notice the relatively flat Moho on the time section, and the overall asymmetry of the basin. Due to crustal stretching, Albian sediments rest almost directly on top of the lower crust and continental mantle lithosphere (Bois and Gariel, 1994). B) Line drawing and simplified geological interpretation of the DOBRE deep seismic profile crossing the Dniepr–Donets graben. Notice the apparent flat Moho beneath the main Devonian depocenter on the time section. High velocity salt occurs in the sedimentary infill of this basin, whereas a dense high velocity body interpreted as underplated volcanics has been identified. Ultimately, post-Devonian positive inversion of this basin has also been identified, accounting also for reverse faulting (Maystrenko et al., 2003; Stephenson et al., 2001; Stovba and Stephenson, 2003).

is best interpreted as a remnant of the former Hercynian lithospheric slab. The preservation of such a dense sub-crustal load since the Carboniferous is probably the mechanism accounting for the Mesozoic subsidence of the overlying basin, where no Triassic nor Jurassic normal faults have ever been documented, and where Alpine foreland lithospheric buckling can only be advocated to explain the Cenozoic subsidence pattern of the basin (Averbuch and Piromallo, 2012; Fig. 2A; Bourgeois et al., 2007; Fig. 2B).

2.4. Negative inversion of former thrust belts

Negative inversion of Paleozoic thrusts inherited from the Caledonian and Hercynian orogenies has been evidenced in the Celtic Sea by BIRPS, SWAT and WAM deep seismic profiles (Bois et al., 1991; Hobbs and Klemperer, 1991; Lefort et al., 1991; Fig. 3), in association with a lateral flow of the lower crust and development of a Mesozoic sag basin in the British Channel. Surface observations and wells correlations have also provided evidence for the poly-phase reactivation of the Hercynian front (faille du Midi), both negatively during Mesozoic extensional episodes and positively during Paleogene compressional episodes (Averbuch et al., 2004).

3. Rifting and extensional basin formation and its role in the evolution of the continental lithosphere

Rifting and extensional basin formation is a key factor in the geological evolution of the continental lithosphere. During the last decades, substantial progress was made in the understanding of thermo-mechanical processes controlling the evolution of rifts and extensional sedimentary basins and the isostatic response of the lithosphere to surface loads such as sedimentary basins (Watts, 2001). Much of this progress stems from improved insight into the mechanical properties of the lithosphere, from the development of new modelling techniques, and from the evaluation of new, high-quality datasets discussed above from previously inaccessible areas of the globe. Below we focus on tectonic models of processes controlling the evolution of extensional basins.

After the realization that subsidence patterns of Atlantic-type margins, corrected for effects of sediment loading and paleo-bathymetry, displayed the typical time-dependent decay characteristic of ocean-floor cooling (Sleep, 1971), a large number of studies were undertaken aimed at restoring the quantitative subsidence history of rifted basins on the basis of well data and outcropping sedimentary sections. With the introduction of backstripping algorithms (Bond and Kominz, 1984; Steckler and Watts, 1978) in the late 1970s' and early 1980s', a phase of basin analysis commenced that aimed at backward modelling by reconstructing the tectonic basin subsidence from sedimentary sequences. These quantitative subsidence histories provided constraints for the development of conceptually driven forward basin models. For extensional basins this commenced in the late 1970s, with the appreciation of the importance of lithospheric thinning and stretching to basin subsidence. After the initial mathematical formulation of stretching concepts in forward modelling of extensional

Fig. 2. A) Sketch diagram outlining the occurrence of a high velocity dense body in the upper mantle beneath the Paris Basin, as evidenced by mantle tomography, which is interpreted as a relic of the former Hercynian slab (after Averbuch and Piromallo, 2012). B) Top: Patterns of lithospheric folding and orientation of folding axis in the northwest European Alpine foreland. Bottom: cross-section A–A′ (see top figure for location), from the Alpine thrust front to the Paris Basin, outlining the large wave-length Cenozoic buckling of the European foreland lithosphere (after Bourgeois et al., 2007).

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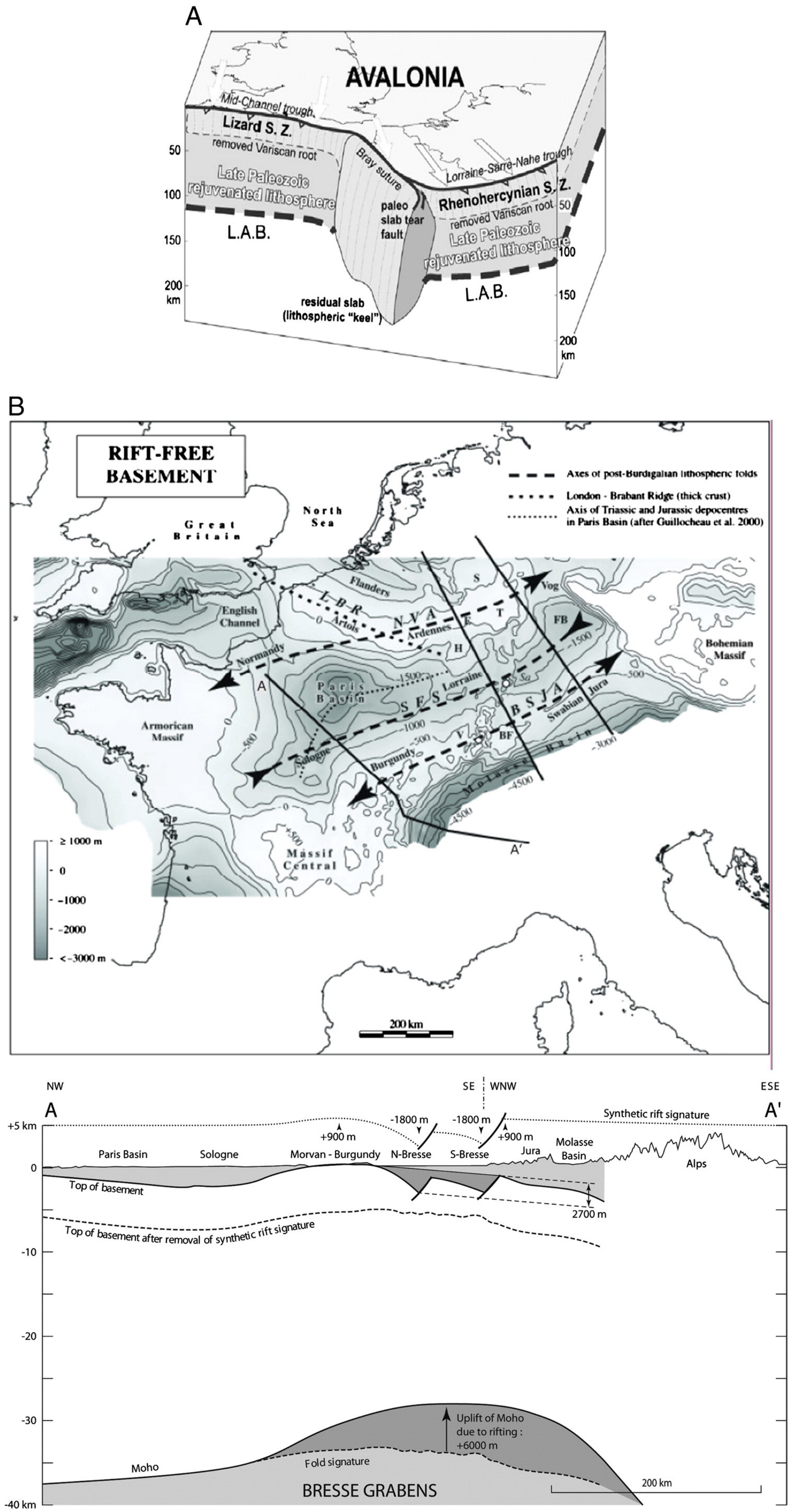

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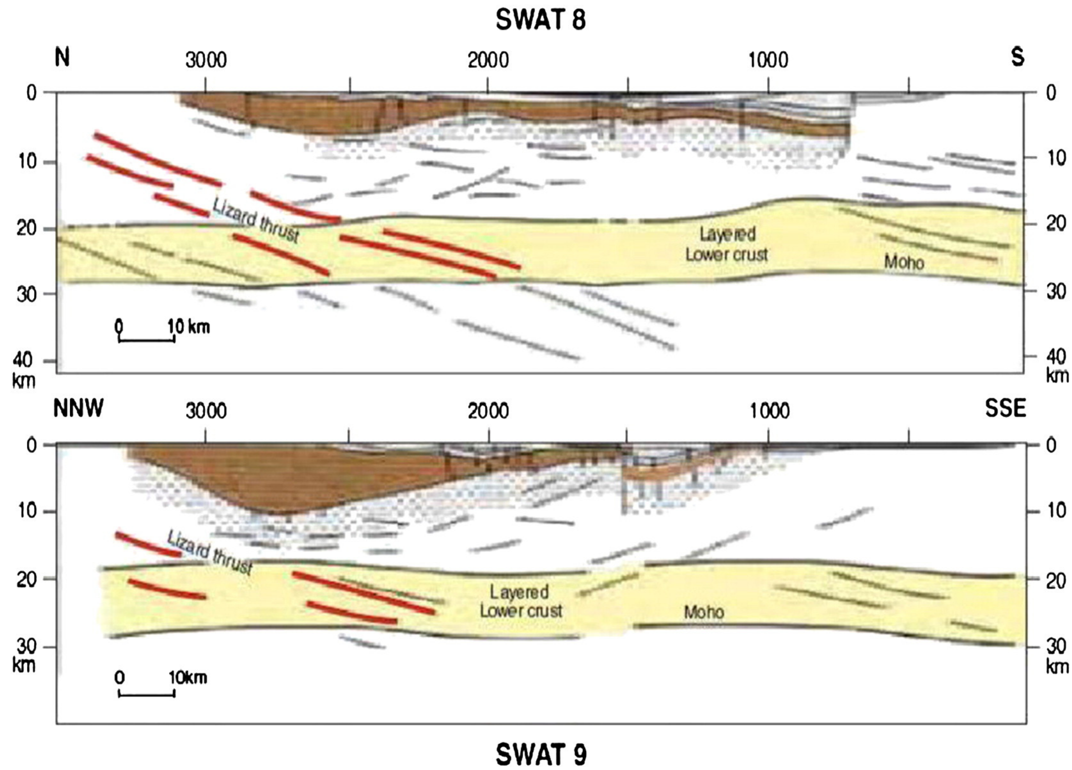

Fig. 3. Line drawing from SWAT and WAM deep seismic profiles across the Celtic Sea and British Channel, outlining the negative inversion of Paleozoic thrusts inherited from the Caledonian and Hercynian orogenies (after Bois et al., 1991; Hobbs and Klemperer, 1991; Lefort et al., 1991). Notice also the lateral flow of the lower crust and coeval development of a Mesozoic sag basin.

basins (McKenzie, 1978), basin fill simulation focused on the interplay between thermal subsidence, sediment loading, and eustatic sea-level changes. To simulate episodic and irregular anomalies commonly observed in otherwise smooth post-rift subsidence curves, changes in sediment supply and eustatic sea-level fluctuations were implemented in an effort to identify their potential tectonic origin (intraplate stress-induced basin subsidence/uplift). In another approach a subsidence curve was used as input, with the basin-modelling package essentially filling the adopted accommodation space (e.g., Lawrence et al., 1990). For the evolution of extensional basins this approach made a clear distinction between their syn-rift and post-rift stages, relating exponentially decreasing post-rift tectonic subsidence rates to a combination of thermal equilibration of the lithosphere–asthenosphere system and to lithospheric flexure (Watts et al., 1982).

Similar simplifications were adopted to describe the syn-rift phase. In the initial version of the stretching model (McKenzie, 1978), lithospheric thinning was described as resulting from instantaneous extension of a semi-infinite half-space with infinite Péclet number. Later more complex tilted fault block models were introduced that decoupled the crust from the lithospheric mantle during the rifting stage (e.g. Brun and Nalpas, 1996; Burov and Cloetingh, 1997; Kaufman and Royden, 1994; Kusznir et al., 1991). These rather theoretical models did, however, not take into account the history of rifted basins and obviously lacked a number of important components of lithosphere mechanics.

A notable feature of most modelling approaches was their emphasis on the basin subsidence record and their very limited capability to handle differential subsidence and uplift patterns in a process-oriented, internally consistent manner (e.g., Kusznir and Ziegler, 1992). Subsequent modelling focused on the quantification and coupling of lithosphere processes and their near-surface expression in response to tectonic controls on basin fill. This demanded a process-oriented approach, linking different spatial and temporal scales in the basin record. In this it was crucial to test and validate model predictions against high-quality data providing information at deeper crustal levels (deep reflection- and refraction-seismics) and the basin fill (reflection-seismics, wells, outcrops); this required a close cooperation between academic and industrial research groups (see Cloetingh and Ziegler, 2007; Cloetingh et al., 2010; Roure, 2008; Roure et al., 2010).

Regarding thermo-mechanical controls on continental breakup and associated basin migration processes, we summarize the results of some of our studies on the continental margin of Norway, one of the best documented in the world owing to concerted efforts by academia and industry in the context of hydrocarbon exploration and production. This rifted margin provides close constraints on the timing and magnitude of extension, extension velocities and its thermal evolution, as well as on its post-rift compressional reactivation and partial inversion that had repercussions on the uplift and erosion of its borderland.

As any other passive margin, however, it is difficult in Norway and its Greenland conjugate margin to constrain the initial architecture of the crust and lithosphere prior to the onset of rifting. That would be a pre-requisite to provide consistent input of thermal or analogue modelling. For instance, a prime question is the actual control of Caledonian thrusts (Anderson et al., 1992) in the localization of the rift-related normal faults and the extent of upper Paleozoic series in the offshore domain. Also, the timing of the unroofing of the Paleozoic series and crystalline basement onshore Norway and Sweden is still debated. The other major question being related to the initial extent and thickness of the Jurassic series onshore, prior to the exhumation of their underlying basement as due to the more recent uplift.

Poly-phase deformation of extensional basins is discussed further on in this paper with examples of the Black Sea and Pannonian–Transylvanian back-arc basins. An intriguing feature of these basin systems is the asymmetric relationship between the syn-rift and post-rift subsidence patterns that reflect overall thermo-mechanical asymmetry. This demands an assessment of the rheological controls on basin formation and of implications for the large-scale basin stratigraphy and rift shoulder dynamics. In this context the role of intraplate stresses and the strength evolution of the lithosphere during the post-rift phase are of particular interest for the understanding of the neotectonic reactivation of the Black Sea Basin system. An overview of the interplay

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between extensional and compressional stresses in the Pannonian–Carpathian basin system of Central Eastern Europe is presented. This area is the site of pronounced lithospheric strength contrasts between the Pannonian Basin area, which is probably underlain by the hottest and the weakest lithosphere of continental Europe, the flanking Carpathian arc and the particularly strong East-European Platform lithosphere. Noteworthy features of the Pannonian Basin system are the short duration of the back-arc rifting phases that controlled the collapse of the internal Carpathian–Dinarides system and the subsequent compressional reactivation of this extensional system. This offers a unique opportunity to study basin evolution in the aftermath of continental collision.

4. Extensional tectonics: concepts and global-scale observations

The development of extensional basins and their evolution into intracontinental thermal sag basins, passive continental margins or back-arc basins is controlled by a wide range of parameters, such as their plate tectonic setting and position relative to collisional plate margins, the dynamics of the sublithospheric mantle, the thermo-tectonic age of the lithosphere, the presence of structural inheritances and the origin and persistence of the controlling stress regime.

4.1. Extensional basin systems

Tectonically active rifts, palaeo-rifts, passive margins and back-arc basins form a group of genetically related extensional basins that play an important role in the spectrum of sedimentary basin types (Bally and Snelson, 1980; Ziegler, 1992, 1996b; Ziegler and Cloetingh, 2004; see also Buck, 2007). 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 intracontinental rifts (e.g., North Sea, Sirt and West Siberian basins, Dniepr–Donets and Gulf of Suez grabens) and passive margins (e.g., Campos Basin, Gabon, Angola, Mid-Norway and NW Australian shelves, Niger and Mississippi deltas; Ziegler, 1996a, 1996b). On these basins, the petroleum industry has acquired large databases that document their structural styles and allow for a detailed reconstruction of their evolution. As discussed above, academic geophysical research programs have provided information on the crustal and lithospheric configurations of tectonically active rifts, palaeo-rifts, passive margins and back-arc basins. 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 toward the understanding of dynamic processes that govern the evolution of extensional basins (Ziegler, 1996b; Ziegler and Cloetingh, 2004).

A natural distinction can be made between tectonically active and inactive rifts, and between rifts that evolved in continental and oceanic lithospheres. Tectonically active intracontinental rifts, such as the Rhine Graben, the East African Rift, the Baikal Rift and the Shanxi Rift of China, are commonly associated with seismicity and volcanic activity that pose significant natural hazards. The mid-ocean ridges form an immense intra-oceanic active rift system that encroaches onto continents in the Red Sea and the Gulf of California. 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 progress 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 (Olsen and Morgan, 1995; Sengor and Burke, 1978). 'Active' rifts are thought to evolve in response to thermal upwelling of the asthenosphere (Bott and Kusznir, 1979; Dewey and Burke, 1975), whereas 'passive' rifts develop in response to lithospheric extension driven by far-field stresses (McKenzie, 1978). 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 (e.g., Bay of Biscay-Pyrenean rift), supports this concept. However, as extrusion of large volumes of rift-related subalkaline 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, preceding 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 (Bott, 1993; Forsyth and Uyeda, 1975; Ziegler, 1993; see also Wessel and Müller, 2007).

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.

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 forces are, however, hard to quantify. Mantle drag stresses at the base of the lithosphere cannot be very high (0.05 MPa assuming an asthenosphere viscosity of 5×10^{19} Pa$*$s and strain rate on the order of 10^{-15} s^{-1}). Consequenly, important drag forces ($>10^{11}$ N) may be created only in case of very large-scale mantle–lithosphere interactions, at high deformation rates and when basal stresses are applied on large areas (hundred thousands of km^2) and over large distances (several thousands of km). If these conditions are satisfied, 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 that contribute to the far-field forces driving lithospheric deformation (Artemieva and Mooney, 2002; Bott, 1993; Forsyth and Uyeda, 1975; Warners-Ruckstuhl et al., 2012). Although the present lithospheric stress field can be readily explained in many areas in terms of plate boundary (far-field) 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 radial outflowing mantle convection cell (e.g., Ziegler et al., 2001). However, it appears that with exception of mega-scale events such as formation of spreading centres, the major impact of mantle–lithosphere interactions refers to normal buoyancy-driven forces and thermo-mechanical erosion at the base of the lithosphere rather than to basal shear arriving from the mantle drag. Normal forces create upward or downward bending of the lithosphere that may generate important concentrations of flexural extensional and compressional stresses inside the bending plates (e.g., Burov and Diament, 1995). These stresses, in conjunction with far-field forces, are major factors effecting rift evolution. Mechanical stretching of the lithosphere and thermal attenuation of the lithospheric mantle are associated with the development of local deviatoric tensional stresses, which play an increasingly important role during advanced rifting stages (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 that is followed by a more 'active' stage during which magmatic processes play an increasingly important role (Burov and Cloetingh, 1997; Huismans and Beaumont, 2011; Huismans et al., 2001a; Wilson and Guiraud, 1992).

Atlantic-type rift systems evolve during the breakup of major continental masses, presumably in conjunction with a reorganization 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 and wide graben systems. In time, rifting activity concentrates in the zone of future crustal separation, with tectonic activity decreasing and ultimately ceasing in lateral graben systems. In time and 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 (pericontinental 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, 2002). Development of Atlantic-type rifts is subject to great variations mainly in terms of the duration of their rifting stage and the level of volcanic activity (Ziegler, 1988, 1990).

New deep seismic refraction data have been acquired in the recent years and these data provide constraints for timing of formation and geometry of rift basins. It is important to consider in detail the effect of magmatism on the Moho geometry, subsidence and the evolution of the lithospheric strength; especially for volcanic rifted margins such as the Vøring volcanic passive margin (e.g. data papers by White et al., 2008; Mjelde et al., 2007a, 2007b, 2008; modelling papers by Buck, 2006; Schmeling, 2010; Schmeling and Wallner, 2012; Simon and Podladchikov, 2008). Although continental breakup might be largely a consequence of plume–lithosphere interaction, melting may also develop without invoking a large mantle plume (e.g. van Wijk et al., 2001). In the Northern Atlantic, multi-scale waveform inversion of seismic data has resolved many details of both crustal and upper mantle structures leading to strong evidence for the Iceland Jan Mayen plume system and suggesting a strong impact of mantle dynamics in the North Atlantic region (Rickers et al., 2013). This research demonstrates a close coincidence between the individual plumes and areas of recent pronounced elevation such as the Southern Scandes of Norway.

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). 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 Rheno–Hercynian Basin, Baikal, Sunda Arc, East China, Pannonian and Black Sea rift systems; Cloetingh and Kooi, 1992a, 1992b; Letouzey et al., 1990; Nikishin et al., 2001; Okay et al., 1994; Royden and Horváth, 1988; Thybo and Nielsen, 2009; Uyeda and McCabe, 1983; Ziegler, 1990).

Extensional disruption of young orogenic belts, involving the development of grabens and pull-apart structures, can be related to their postorogenic uplift and the development of deviatoric tensional stresses inherent to orogenically overthickened crust (Dewey, 1988; Sanders et al., 1999; Stockmal et al., 1986). The following mechanisms contribute to postorogenic 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 (Andeweg and Cloetingh, 1998; Bott, 1993); 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; Straume and Austrheim, 1999). Although tensional collapse of an orogen can commence shortly after its consolidation (e.g. the Variscan Orogen, Ziegler, 1990; Ziegler et al., 2004), it may be delayed by as much as 30 My, as in the case of the Appalachian–Mauretanides (Ziegler, 1990).

Dynamic thermo-mechanical models are important to analyse the effect of boundary conditions such as extension velocity and surface erosion on the mode of rifting. Extension velocity is a key parameter, governing the focusing of deformation and breakup. In addition, there are important implications of the lithosphere rheology with respect to extension — the relationship of viscous flow power-law exponent and development of necking (Fletcher, 1974; Schmalholtz, 2008) and Moho geometry (e.g., Tirel et al., 2008). The Influence of melting-related weakening and other strain localization processes can also be considerable. For instance, Huismans and Beaumont (2007) have shown that localization of deformation and rift mode selection during extension may be closely related to dynamic weakening; some recent publications emphasize several physical mechanisms to explain this weakening such as shear heating (Crameri and Kaus, 2010; Kaus and Podladchikov, 2006), damage evolution (Karrech et al., 2011), grain size reduction (Braun et al., 1999), lattice preferred orientation (Tomassi et al., 2009) and some other mechanisms such as fluid-induced methamorphic reactions at different crustal levels (Mohn et al., 2011).

4.1.1. Thermal thinning and stretching of the lithosphere: concepts and models

Mechanical stretching of the lithosphere can be associated with significant magmatic activity and increase in conductive and advective heat flux and consequently an upward displacement of the thermal asthenosphere–lithosphere boundary. Small-scale convection in an evolving asthenospheric diapir may contribute to mechanical thinning of the lithosphere by facilitating lateral ductile mass transfer (Richter and McKenzie, 1978). Progressive thermal and mechanical thinning of the higher-density lithospheric mantle and its replacement by lower-density asthenosphere induces progressive doming of rift zones. In case of slow rifting, these processes may lead to delamination of the lithosphere mantle and hence to accelerated widening of the rift (Burov, 2007). At the same time, deviatoric tensional stresses developing in the lithosphere contribute to its further extension (Bott, 1992). Although rift-related major hot-spot activity is thought to reflect a thermal perturbation of the asthenosphere caused by a deep mantle plume, smaller-scale 'plume' activity may also be the consequence of lithospheric stretching, which is triggered by adiabatic decompression partial melting in areas characterized by an anomalously volatile-rich asthenosphere/lithosphere (White and McKenzie, 1989; Wilson, 1993a).

Depending on the applicability of the 'pure-shear' (McKenzie, 1978) or the 'simple-shear' model (Wernicke, 1985), or a combination thereof (Kusznir et al., 1991), the zone of upper crustal extension, corresponding to the subsiding rift, may coincide with the zone of lithospheric mantle attenuation (pure- and combined-shear) or may be laterally offset from it (i.e. simple-shear). A modification to the pure-shear model is the 'continuous depth-dependent' stretching model which assumes that stretching of the lithospheric mantle affects a broader area than the zone of crustal extension (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 lithospheric mantle. 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 produces footwall uplift of the rift flanks and intrabasin fault blocks (Kusznir and Ziegler, 1992). By the same mechanism, the simple-shear model predicts asymmetric doming of a rift zone or even flexural uplift of an arch located to one side of the zone of upper crustal extension (Wernicke, 1985). A modification to the simple-shear model envisages that massive upper crustal extensional unloading of the lithosphere causes its isostatic uplift and passive flow of the asthenosphere towards the rift axis (Wernicke, 1990).

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 lithospheric mantle 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).

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 (β factor). Although quantification of the extensional strain and of the stretching factor is of basic importance for the understanding of rifting processes, there is often a considerable discrepancy between estimates derived from upper crustal extension by faulting, the volume of the rift zone, the crustal configuration and quantitative subsidence analyses (Hinsken et al., 2011; Ziegler and Cloetingh, 2004). Some of these discrepancies can be explained by interplays between surface processes and subsurface deformation (Burov and Poliakov, 2001) and other factors such as gravitational instabilities in the extending lithosphere mantle (Burov, 2007).

It is also noteworthy that the above hypotheses and concepts are based on the inherent assumption of 2D cylindrical (zero strain in the out-of-plane direction) deformation. It can be understood that in some cases, the out-of-plane outflow or inflow of mantle–asthenosphere material or out-of-plane stress field may significantly affect rift evolution.

4.1.2. Syn-rift subsidence and duration of rifting stage

The balance of two mechanisms controls the syn-rift subsidence of a sedimentary basin. First, elastic/isostatic adjustment of the crust to stretching of the lithosphere and its adjustment to sediment loading causes subsidence of the mechanically thinned crust (Keen and Boutilier, 1990; McKenzie, 1978). Depending on the depth of the lithospheric necking level, this is accompanied by either flexural uplift or down warping of the rift zone (Fig. 4) (Braun and Beaumont, 1989; Kooi et al., 1992). Second, 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 (Turcotte and Emerman, 1983). Thus, the geometry of a rifted basin is a function of the elastic/isostatic response of the lithosphere to its mechanical stretching and the related thermal perturbation (Van der Beek et al., 1994).

The duration of the rifting stage of intracontinental rifts (aborted) and passive margins (successful rifts) is highly variable (Figs. 5 and 6) (Ziegler, 1990; Ziegler and Cloetingh, 2004; 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 buildup rate, magnitude and persistence of the extensional stress field, constraints on lateral movements of the diverging blocks (on-trend coherence, counteracting far-field compressional stresses), and apparently less dependent on the availability of pre-existing crustal discontinuities that can be tensionally reactivated.

Crustal separation was achieved in the Liguro–Provençal Basin after 9 My of crustal extension and in the Gulf of California after about 14 My of rifting, whereas opening of the Norwegian–Greenland Sea was preceded by an intermittent rifting history spanning some 280 My (Ziegler, 1988; Ziegler and Cloetingh, 2004). There appears to be no obvious correlation between the duration of the rifting stage (R) of successful rifts (Fig. 6), which are superimposed on orogenic belts (Liguro–Provencal 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 My; northern South Atlantic R = 29 My; Red Sea R = 29 My; Baffin Bay R = 70 My; Labrador Sea R = 80 My). 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 pre-existing or inherited 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; Ziegler et al., 2001).

4.1.3. 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 density and 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

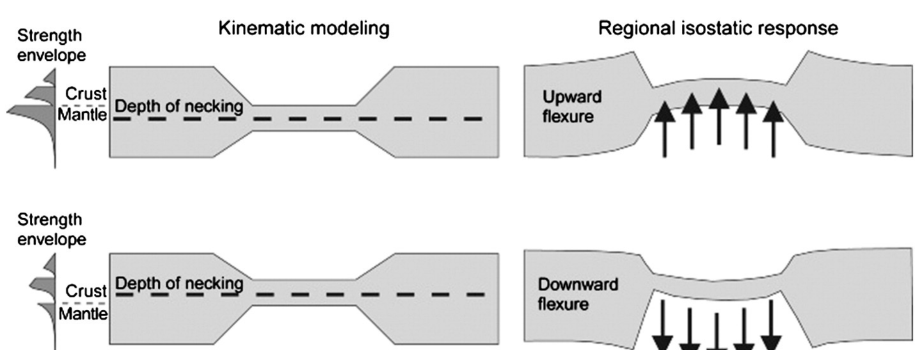

Fig. 4. Concept of lithospheric necking. The level of necking is defined as the level of no vertical motions in the absence of isostatic forces. Left panel: kinematically induced configuration after rifting for different necking depths. Right panel: subsequent flexural isostatic rebound.

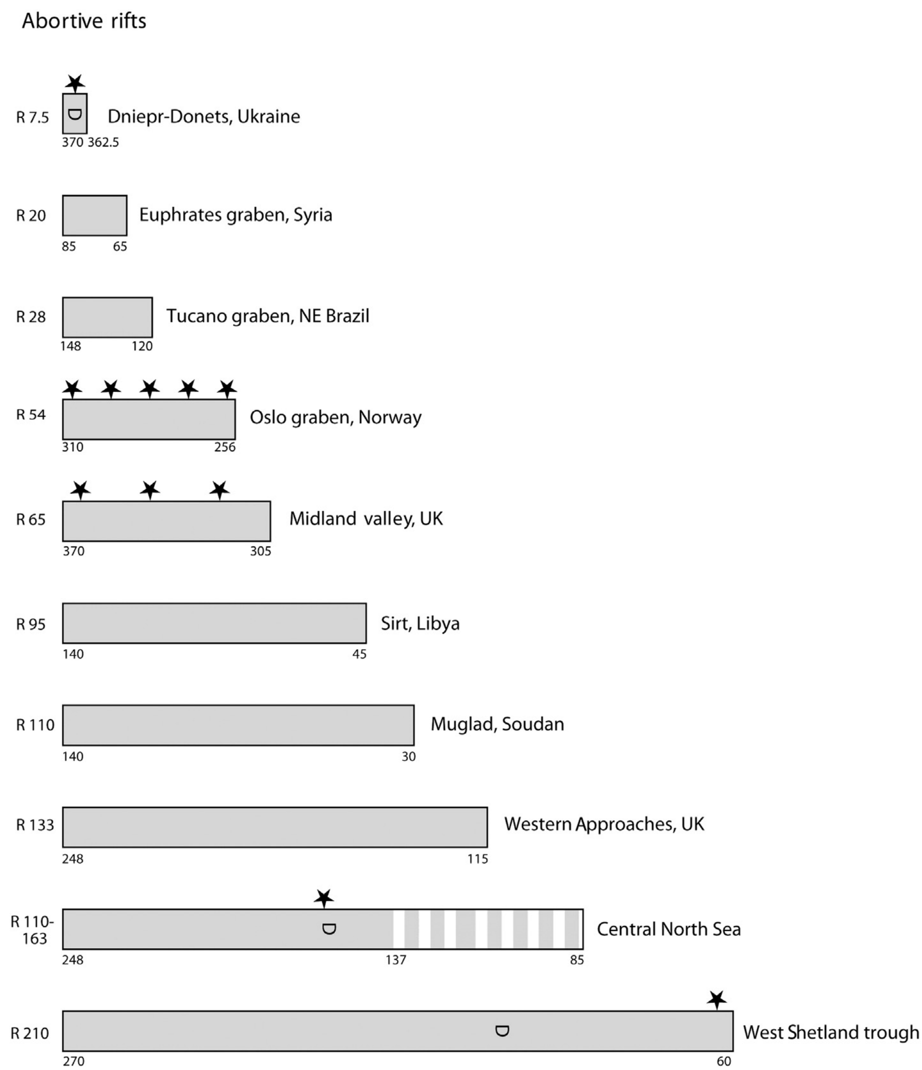

Fig. 5. Duration of rifting stage of 'abortive' rifts (paleo-rifts, failed arms). Horizontal bars in My; numbers below horizontal bars indicate onset and termination of rifting stage in My; numbers besides R left of each bar give the duration of rifting stage in My; stars indicate periods of main volcanic activity; D indicates periods of doming.
From Ziegler and Cloetingh (2004).

of which is thought to be directly related to the lithospheric stretching value (Beaumont et al., 1982; McKenzie, 1978; Royden et al., 1980; Steckler and Watts, 1978; Watts et al., 1982). During the post-rift evolution of a basin, the thermally destabilized continental lithosphere re-equilibrates with the asthenosphere (McKenzie, 1978; Steckler and Watts, 1978; Wilson, 1993b). In this process, in which the temperature regime of the asthenosphere plays an important role (ambient, below, or above ambient), new lithospheric mantle, consisting of solidified asthenospheric material, is accreted to the attenuated old continental lithospheric mantle (Ziegler et al., 1998). On the other hand, lateral heat exchanges and progressive increase of the flexural strength of the lithosphere with cooling may significantly retard it subsidence (compared to the McKenzie model) after the first 10–20 Ma of post-rift subsidence (Burov and Poliakov, 2001). 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, 1992b; Lobkovsky et al., 1996; Van Wees and Cloetingh, 1996).

The shape and dimension of rift-induced asthenosphere–lithosphere boundary anomalies essentially control the geometry of the evolving post-rift thermal-sag basin. Thermal sag basins associated with aborted rifts are broadly saucer-shaped and generally overstep the rift zone, with their axes coinciding 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, 1989) 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 Bear, 1990; Dniepr–Donets Graben, Kusznir et al., 1996, Gulf of Thailand, Hellinger and Sclater, 1983; Sudan rifts, McHargue et al., 1992). Such a geometric relationship is compatible with discontinuous, depth-dependent stretching models that assume that the zone of crustal extension is narrower than the zone of lithospheric mantle attenuation. The degree to which a post-rift basin oversteps the margins of the syn-rift basin is a function of the difference in the width of the zone of crustal extension and the zone of lithospheric mantle attenuation (White and McKenzie, 1988) and the effective elastic thickness (EET) of the lithosphere (Watts et al., 1982).

Conversely, simple-shear dominated rifting gives rise to a lateral offset between the syn- and post-rift basin axes. An example is the Tucano

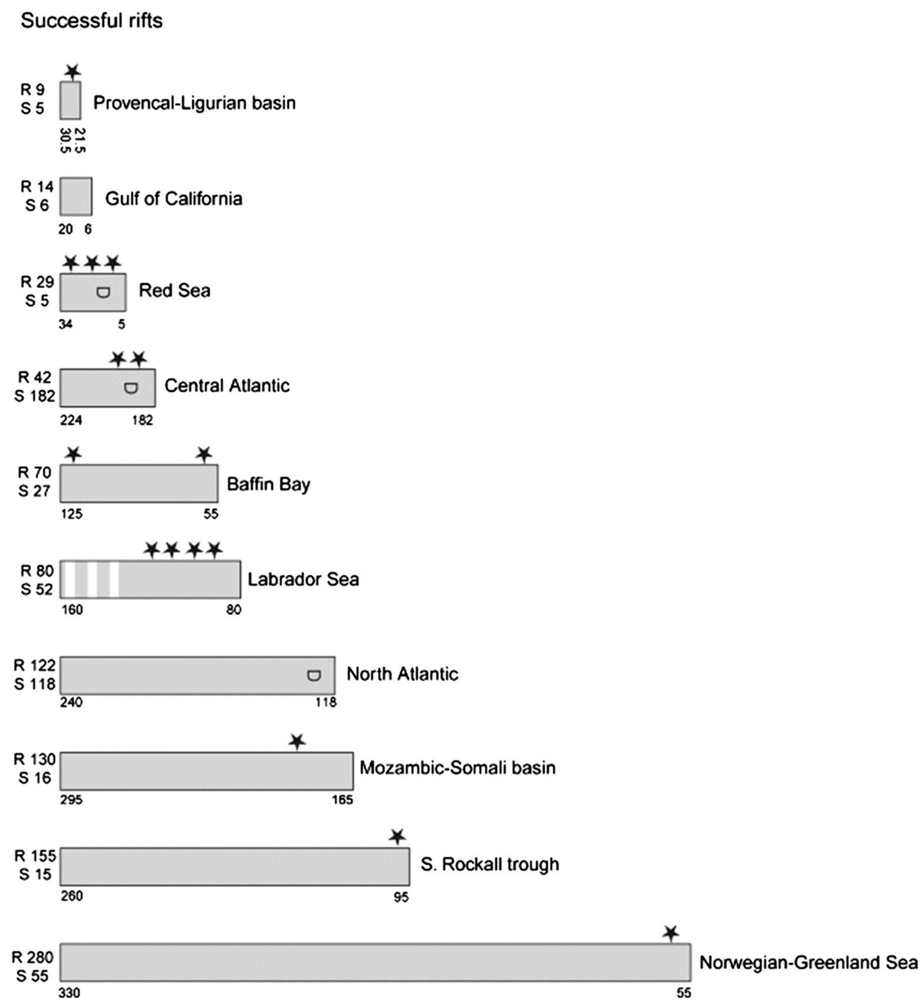

Fig. 6. Duration of rifting stage of 'successful' rifts. Legend same as Fig. 5. Numbers beside letter 'S' indicate duration of seafloor spreading stage in million years. From Ziegler and Cloetingh (2004).

Graben of north-eastern Brazil, which ceased to subside at the end of the 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 and subsequent major post-rift subsidence (Chang et al., 1992; Karner 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 are attributed to differences in their lithospheric configuration at the crustal separation stage. At the end of the rifting stage, a relatively thick lithospheric mantle supports lower plate margins, whereas upper plate margins are underlain by a strongly attenuated lithospheric mantle and partly directly by the asthenosphere. Correspondingly, lower plate margins are associated at the crustal separation stage with smaller thermal anomalies than upper plate margins (Stampfli et al., 2001; Ziegler et al., 1998). These differences in lithospheric configuration of conjugate simple-shear margins have repercussions on their rheological structure, even after full thermal relaxation of the lithosphere, and their compressional reactivation potential (Ziegler et al., 1998).

The magnitude of post-rift tectonic subsidence of aborted rifts and passive margins is a function of the thermal anomaly that was introduced during their rifting stage and the degree to which the lithospheric mantle was thinned. Most intense anomalies develop during crustal separation, particularly when a plume assisted and asthenospheric melts well up close to the surface. The magnitude of thermal anomalies induced by rifting that did not progress to crustal separation depends on the magnitude of crustal stretching (δ-factor) and lithospheric mantle attenuation (β-factor), the thermal regime of the asthenosphere, the volume of melts generated and whether these intruded the lithosphere and influenced the evolution of the Moho. 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 have decayed (mantle-plume model).

The thickness of the post-rift sedimentary column that can accumulate in passive margin basins and in thermal-sag basins above aborted rifts is not only a function of the magnitude of the lithospheric mantle and crustal attenuation factors, but also of the population of upper crustal faults (Ziegler, 1983, 1990). This discrepancy is somewhat reduced when flexural isostasy is assumed (Roberts et al., 1993). However, because during rifting attenuation of the lithosphere is not only achieved by its mechanical stretching 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 cannot be directly related to a mechanical stretching factor. Moreover, intraplate compressional stresses and phase transformations in the lower crust and lithospheric mantle have an overprinting effect on post-rift subsidence and can cause significant departures from a purely thermal subsidence curve (Cloetingh and Kooi, 1992b; Lobkovsky et al., 1996; Van Wees and Cloetingh, 1996).

Furthermore, during rifting the ascent of mantle-derived melts to the base of the crust can cause destabilization of the Moho, magmatic inflation of the crust and its metasomatic reactivation and secondary

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differentiation (Mohr, 1992; Morgan and Ramberg, 1987; Stel et al., 1993; Watts and Fairhead, 1997). For instance, lower crustal velocities of 6.30–7.2 km/s and densities of 3.02×10^3 kg/m^3 characterize the highly attenuated crust of the Devonian Dniepr–Donets rift (Fig. 1B) almost up to the base of its syn-rift sediments, probably owing to its syn-rift permeation by mantle-derived melts (Maystrenko et al., 2003; Yegorova et al., 1999).

Moreover, accumulation of very thick post-rift sedimentary sequences can cause phase transformation of crustal rocks to granulite facies, with the resulting densification of the crust accounting for accelerated basin subsidence. This process can be amplified in the cases of eclogite transformation of basaltic melts that were injected into the lithospheric mantle or that had accumulated at its base (Lobkovsky et al., 1996). 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). 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 of asthenosphere, Cloetingh and Kooi, 1992b), it is uncertain whether physical conditions conducive to such transformations can develop in response to post-rift sedimentary loading of palaeo-rifts and passive margins as, for example, suspected for the East Newfoundland Basin (Cloetingh and Kooi, 1992b) and the West Siberian Basin (Lobkovsky et al., 1996).

An alternative process to account for the development of post-rift sag basins in passive margins is the lateral viscous flow operating in the lower crust away from the main depocenters. Lateral viscous flow in the lower crust is also well evidenced in the core complexes associated with the collapse of orogens (i.e. in the North American Cordillera), and at early stages of back-arc opening, prior to the formation of true oceanic lithosphere (i.e. in the Aegean back arc and Algerian Basin, the later being still characterized by a layered lower crust that could represent the remnants of hyper-extended continental material (Roure et al., 2009).

As rifting processes can take place intermittently over very long periods of time (e.g., Norwegian-Greenland Sea rift, Ziegler, 1988), thermal anomalies introduced during early 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: Nadin and Kusznir, 1995; Ziegler, 1990). Therefore, analyses of the post-rift subsidence of rifts that have evolved in response to multiple rifting phases spread over a long period need to take their entire rifting history into account.

Intraplate compressional stress, causing deflection of the lithosphere in response to lithospheric folding, can seriously overprint the thermal subsidence of post-rift basins (see for a recent discussion Cloetingh et al., 2008). The example of the Plio-Pleistocene evolution of the North Sea shows that the buildup of regional compressional stresses can cause a sharp acceleration of post-rift subsidence (Cloetingh et al., 1990; Kooi et al., 1991; Van Wees and Cloetingh, 1996). Similar contemporaneous effects are recognized in North Atlantic passive-margin basins (Cloetingh et al., 1990) as well as in the Pannonian Basin (Horváth and Cloetingh, 1996). Also the late Eocene accelerated subsidence of the Black Sea Basin can be attributed to the buildup of a regional compressional stress field (Cloetingh et al., 2003; Robinson et al., 1995).

At present, many post-rift extensional basins are in a state of horizontal compression as documented by stress indicators summarized in the World Stress Map (Heidbach et al., 2010). The magnitude of stress-induced vertical motions of the lithosphere during the post-rift phase, causing accelerated basin subsidence and tilting of basin margins, depends on the ratio of the stress level and the strength of the lithosphere inherited from the syn-rift phase. Moreover, horizontal stresses in the lithosphere strongly affect the development of salt diapirism, accounting for local subsidence anomalies (Cloetingh and Kooi, 1992b), and have a strong impact on the hydrodynamic regime of extensional basins (Van Balen and Cloetingh, 1994) by contributing to the development of overpressure, as seen in parts of the Pannonian Basin (Van Balen et al., 1995). Glacial loading and unloading can further complicate post-rift lithospheric motions, the scope of which requires further analysis (Solheim et al., 1996) (e.g. post-glacial rebound of the Barents Sea shelf).

These considerations indicate that stretching factors derived from the subsidence of post-rift basins must be treated with reservations. Nevertheless, quantitative subsidence analyses, combined with other data, are essential for the understanding of post-rift subsidence processes by giving a measure of the lithospheric anomaly that was introduced during the rifting stage of a basin and by identifying deviations from purely thermal cooling trends.

4.1.4. Post-rift compressional reactivation of extensional basins

Palaeo-stress analyses give evidence for changes in the magnitude and orientation of intraplate stress fields on time scales of a few million years (Bergerat, 1987; Dèzes et al., 2004). 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, 1992b).

In response to the buildup of far-field compressional stresses, rifted basins, characterized by a strongly faulted and thus permanently weakened crust, are prone to reactivation at all stages of their post-rift evolution, resulting in their inversion (Ziegler, 1990; Ziegler et al., 1995, 1998, 2001). Only under special condition can gravitational forces associated with topography around a basin cause its inversion (Bada et al., 2001; Pascal and Cloetingh, 2009).

Rheological considerations indicate that the lithosphere of thermally stabilized rifts, lacking a thick post-rift sedimentary prism, is considerably stronger than the lithosphere of adjacent unstretched areas (Ziegler and Cloetingh, 2004). This contradicts the observation that rift zones and passive margins are preferentially deformed during periods of intraplate compression (Ziegler et al., 2001). However, burial of rifted basins under a thick post-rift sequence contributes by thermal blanketing to weakening of their lithosphere (Stephenson, 1989), thus rendering it prone to tectonic reactivation.

4.1.5. Finite strength of the lithosphere during extensional basin formation

In recent approaches to extensional basin modelling the implementation of finite lithospheric strengths has been an important step forward. Advances in the understanding of lithospheric mechanics (Burov, 2011) demonstrate that early stretching models, which assumed zero lithospheric strength during rifting, are not valid (e.g., Bassi, 1995; Chery et al., 1992). Moreover, the notion of a possible decoupling zone between the strong upper crust and the strong lithospheric upper mantle is also important in the context of extensional basin formation. During the past decades the relative importance of the pure shear (McKenzie, 1978) and simple shear (Wernicke, 1985) mode of extension has been a matter of debate. In the presence of a weak lower crustal layer, decoupling of the mechanically strong upper mantle from the strong upper crust depends on their thermal age and the thickness of their sedimentary cover.

During the last few years, the role played by rheology during extensional basin formation was gradually appreciated due to the availability of improved modelling, increasingly shifting attention away from these end lithospheric members. Finite element models permitted to explore the large-scale implications of a finite lithospheric strength and particularly its sensitivity to the presence of fluids in the crust and lithospheric mantle (Bassi, 1995; Braun and Beaumont, 1989; Dunbar and Sawyer, 1989). These differences are a direct consequence of contrasts in thermo-tectonic age between young continental versus and old cratonic lithosphere affected by extensional basin formation. These dynamic models, which require intensive and expensive computing, are not always suitable for a user-oriented industry environment. However, they provided the background for a more user-friendly class of kinematic models targeted at modelling rift-shoulder uplift and the basin fill. For the development of extensional basins these kinematic models invoke necking of the lithosphere around one of its strong layers (see Braun and Beaumont, 1989; Kooi et al., 1992; Spadini et al., 1995). Fig. 4 illustrates the basic features of these models and their relation to the strength distribution within the lithosphere. In the presence of a strong layer in the subcrustal mantle, the lithospheric necking level is deep, inducing pronounced rift-shoulder uplift. This type of response is to be expected if extension affects cold and correspondingly strong intracontinental lithosphere with an average crustal thickness of 30–40 km; it is commonly observed in intracratonic rifts and rifted margin such as, for example, the Red Sea and the Trans-Antarctic Mountains (Chery et al., 1992; Cloetingh et al., 1995).

For Alpine/Mediterranean basins, which developed on young, orogenically thickened crust and lithosphere, the necking level is generally located at depths of 5 to 10 km within the crust (Fig. 4). An example of such a situation is found in the Pannonian Basin (Horváth and Cloetingh, 1996; Van Balen et al., 1995) where the strength of the lithospheric upper mantle is almost zero. Important exceptions to this general pattern do occur, however, such as in the Southern Tyrrhenian Sea, for which our modelling indicates a deep necking level to fit observational data (Spadini et al., 1995). This is primarily attributed to the fact that the Southern Tyrrhenian Basin developed essentially on Hercynian lithosphere with significant bulk strength of its mantle component.

Similarly, models analyzing the differences between the western and eastern sub-basins of the Black Sea have inferred important variations in lithospheric necking depth and thermal conditions related to the mode and timing of extension, as well as the thermo-tectonic age of the underlying lithosphere (Cloetingh et al., 2003; Robinson et al., 1995; Spadini et al., 1996, 1997). The inter-dependence of the necking depth and pre-rift crustal and lithospheric thicknesses, the effective elastic thickness (EET) of the lithosphere and strain rates was investigated in a comparative study on several basins, which are all superimposed on Hercynian or Alpine destabilized crust (Cloetingh et al., 1995). Results of this study suggest that for Alpine/Mediterranean basins the position of the necking level depends primarily on the pre-rift crustal thickness and strain rate, whereas the key controlling factors in intra-cratonic rifts, which are superimposed on Precambrian crust, appear to be the pre-rift lithospheric thickness and strain rate. These models illustrate the importance of the pre-rift lithosphere rheology for the architecture of the evolving extensional basins and the associated pattern of vertical motions. Moreover, they demonstrate that the better the pre-rift evolution of an area is constrained, the greater the chance to define the proper parameters for successful modelling and the definition of the underlying syn-rift mechanisms.

4.1.6. Rift-shoulder development and architecture of basin fill

As discussed above, the finite strength of the lithosphere has important implications for the crustal structure of extensional basins and the development of accommodation space within them. The development of significant rift-shoulder topography in response to lithospheric extension has drawn attention to the need to constrain the coupled vertical motion of the uplifting rift shoulders and the subsiding basin (Chery et al., 1992; Kusznir and Ziegler, 1992). Whereas the standard approach in basin analysis focused until recently primarily on the subsiding basin, treating sediment supply as an independent parameter, necking models highlight the need for linking sediment supply to the rift-flank uplift and their erosion history. To this end, a two-fold approach was followed. The first research line aimed at constraining the predicted uplift histories by geothermochronology. Modelling of the distribution of fission-track length permits to backstack the eroded sediments from their present position in the basin to their source on the rift shoulder in an effort to obtain a better reconstruction of the rift-shoulder geometries (e.g., Burov and Cloetingh, 1997; Redfield et al., 2005; Rohrman et al., 1995; Van der Beek et al., 1995). This has led to a better understanding of the timing and magnitude of rift-shoulder uplift.

A second research line focused on the development of a model for basin fill simulation, integrating the effects of rift-shoulder erosion through hill-slope transport and river incision with sediment deposition in the basin. These models (e.g., van Balen et al., 1995), predict the progradation of sedimentary wedges into extensional basins and the development of hinterland basins having a distinctly different stratigraphic signature than predicted by standard models, which invoke stretching and post-rift flexure, as commonly applied in the software packages. Testing this model against a number of rifted margins around the world demonstrates that erosion of the rift-shoulder topography, created during extension of a lithosphere with a finite strength, can sometimes eliminate the need to invoke eustatic sea-level changes to explain large-scale stratigraphic features of rifted basins and associated hinterland basins. The tectonically-driven erosion of these rift shoulders is commonly observed on the flanks of extensional basins such as the Catalan Coastal Ranges facing the Neogene Valencia trough of the Western Mediterranean (Gaspar-Escribano et al., 2003), or the large scale exhumation of the onshore Morroccan segment of the Atlantic margin (Gouiza et al., 2010).

Considering the notion that in syn-rift basins different spatial and temporal scales are by their very nature linked, the lack of constraints on the evolution and structural configuration of surrounding areas severely limits modelling of syn-rift sedimentary sequences. Moreover, modelling can assess tectonic controls on syn-rift depositional sequences at a sub-basin scale (see also Nottvedt et al., 1995). The magnitude of footwall uplift of individual fault blocks appears to be a function of the lithospheric necking level and thus cannot be attributed to factors restricted to the sub-basin scale.

5. Rheological stratification of the lithosphere and basin evolution

5.1. Lithosphere strength and deformation mode

The major, directly measurable proxy for the integrated strength of the lithosphere is its EET, (or T_e, see Watts, 2001 and references therein). By comparing observations of flexure in the region of long-term loads such as ice, sediment and volcanoes to the predictions of elastic plate flexure models, T_e of the lithosphere in a wide range of geological settings could be estimated. Flexure studies on oceanic lithosphere suggest that its T_e ranges between 2 and 40 km and depends on plate age and sedimentary load. The T_e of continental lithosphere ranges from 0 to 100 km and shows no clear relationship with the consolidation age of the crust (Watts, 2001). The results of these flexure studies are qualitatively consistent with the results of experimental rock mechanics. The Brace–Goetze failure envelope curves (Brace and Kohlstedt, 1980; Goetze and Evans, 1979), for example, predict that strength increases with depth and then decreases in accordance with the brittle (e.g. Byerlee, 1978) and ductile deformation laws. In oceanic regions, the envelopes are approximately symmetric about the depth of the Brittle–Ductile Transition (BDT) where the brittle–elastic and elastic–ductile layers contribute equally to the strength. Furthermore, the lithospheric strength is also dependent on rock composition and fluid content (Burov, 2011).

The strength of continental lithosphere is controlled by its stratified depth-dependent rheological structure in which the thickness and composition of the crust, the thickness of the lithospheric mantle, the potential temperature of the asthenosphere, and the presence or absence of fluids, as well as 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 (Cloetingh and Burov, 1996; Watts, 2001; see also Burov, 2007, 2011). 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 that 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, 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).

Despite this, Burov and Diament (1995) showed that a rheological model in which a weak lower crust is sandwiched between a strong brittle–elastic upper crust and an elastic–ductile mantle ("jelly sandwich" model, Jackson, 2002) accounts for the wide range of lithospheric T_e values observed due to wide variations in crustal thickness and composition, and geothermal gradients.

It has been also suggested (Jackson, 2002) that the observed maximal depth of earthquake occurrence (T_s) correlates with T_e and hence is also a proxy for the integrated strength of the lithosphere. From observations it is evident that mantle seismicity is rare in continents, suggesting that their lithospheric strength is concentrated in the continental crust (so called "crème-brûlé" rheology model, Burov and Watts, 2006; Handy and Brun, 2003). These concepts were, however, later dismissed (Audet and Bürgmann, 2011; Burov and Watts, 2006; Watts and Burov, 2003) on account of (1) the lack of physical grounds for a $T_e - T_s$ correlation (2) continental T_e values being often larger than T_s, and (3) the rare occurrence of sub-Moho earthquakes in continents. However, the rarity of sub-Moho seismicity still needs to be explained even though some first order explanations have been advanced such as an increase of the brittle strength with depth, domination of grain boundary sliding creep flow in case of lithospheric mantle deformation.

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 (Ziegler and Cloetingh, 2004; see also Burov, 2011). 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, 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. This fabric reflects the presence of deep-reaching lithological inhomogeneities and shear zones (Ziegler and Cloetingh, 2004).

The strength of the continental upper lithospheric mantle depends to a large extent on the thickness of the crust but also on its age and thermal regime (see Jaupart and Mareschal, 2007; Toussaint et al., 2004). Thermally stabilized stretched continental lithosphere with a 20 km thick crust and a lithospheric mantle thickness of 50 km is mechanically stronger than unstretched lithosphere with a 30 km thick crust and a 70 km thick lithospheric mantle. 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, that is, 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 lithospheric mantle is also weak. Extension of this type of lithosphere, involving ductile flow of the lower and the 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 toward the hanging wall (e.g., Basin and Range Province, Bertotti et al., 2000; Buck, 1991; Wernicke, 1990). 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 (Moisio et al., 2000). However, the occurrence of upper mantle reflectors, which generally dip in the same direction as the crustal fabric and probably are related to subducted oceanic and/or continental crustal material, suggests that the continental lithospheric mantle 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 lithospheric mantle. Moreover, even in the face of similar crustal thicknesses, the heat flow of deeply degraded Late Proterozoic 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 lithospheric mantle 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 (Janssen et al., 1995; Ziegler, 1989a, 1989b; 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 (Ziegler et al., 1998). In this respect, it must be realized that, during rifting, progressive mechanical and thermal thinning of the lithospheric mantle 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. Upon decay of the rift-induced thermal anomaly, rift zones are, rheologically, considerably stronger than unstretched lithosphere. 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 lithospheric mantle of rifted basins (Stephenson, 1989). Moreover, as faults may permanently weaken the crust of rifted basins due to strain softening mechanisms activated during their formation (Huismans and Beaumont, 2003), 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 viscoelastic beam (Reston, 1990; ter Voorde et al., 1998). The response of such a system to the buildup 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 (Watts and Burov, 2003; Zeyen et al., 1997). As the structure of continental lithosphere is also areally heterogeneous, its weakest parts start to yield first, once tensional or compressional intraplate stress levels equate their strength.

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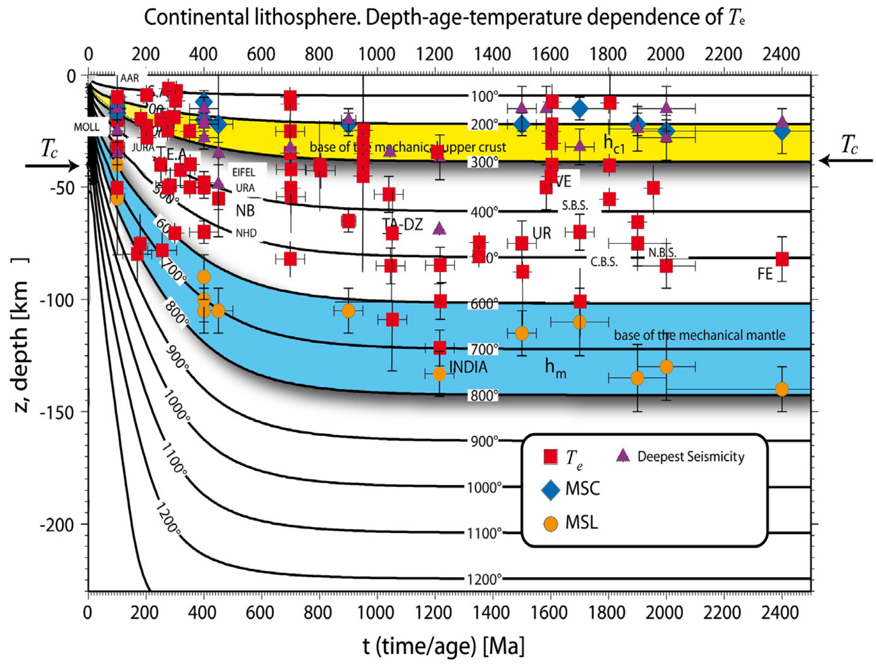

Fig. 7. Compilation of observed and predicted values of effective elastic thickness (EET), depth to bottom of mechanically strong crust (MSC), and depth to bottom of mechanically strong lithospheric mantle (MSL) plotted against the age of the continental lithosphere at the time of loading and comparison with predictions from thermal models of the lithosphere. Labelled contours are isotherms. Isotherms marked by 'solid lines' are for models that account for additional radiogenic heat production in the upper crust. 'Dashed lines' correspond to pure cooling models for continental lithosphere. The equilibrium thermal thickness of the continental lithosphere is 250 km. The yellow band corresponds to depth intervals marking the base of the mechanical crust (MSC) while the blue band marks the strong part of the mantle lithosphere (MSL). 'Squares' correspond to EET estimates, 'circles' indicate MSL estimates, and 'diamonds' correspond to estimates of MSC. 'Bold letters' correspond to directly estimated EET values derived from flexural studies on, for example, foreland basins, 'Thinner letters' indicate indirect rheological estimates derived from extrapolation of rock-mechanics studies. The data set includes (I): Old thermo-mechanical ages (1000–2500 Ma): northernmost (N.B.S.), central (C.B.S.), and southernmost Baltic Shield (S.B.S.); Fennoscandia (FE); Verkhoyansk plate (VE); Urals (UR); Carpathians; Caucasus; (II): Intermediate thermo-mechanical ages (500–1000 Ma): North Baikal (NB); Tarim and Dzungaria (TA-DZ); Variscan of Europe: URA, NHD, EIFEL; and (III): Younger thermo-mechanical ages (0–500 Ma): Alpine belt: JURA, MOLL (Molasse), AAR; southern Alps (SA) and eastern Alps (EA); Ebro Basin; Betic rifted margin; Betic Cordilleras.
Modified from Cloetingh and Burov (1996).

5.2. Mechanical controls on the evolution of rifts: Europe's continental lithosphere

The thermo-mechanical age concept provides the framework for EET estimates of the lithosphere. Fig. 7 was constructed on the basis of a large data set for Eurasian foreland basins (Cloetingh and Burov, 1996) and illustrates the general trend of increasing elastic plate thickness with increasing thermo-mechanical age of the continental lithosphere. Moreover, Fig. 7 summarizes the bulk rheology of the lithosphere based on extrapolations from rock mechanical data, constrained by crustal geophysical data and thermal models. A characteristic feature of these models is the incorporation of a quartz-dominated upper crustal rheology and an olivine-controlled mantle rheology. These models were designed to shed light on the depth to the base of the mechanically strong upper part of the crust (MSC) and the mechanically strong part of the upper mantle lithosphere (MSL). Analysis of Fig. 7 demonstrates that the mechanical properties of the crust are little affected by its age-dependent cooling, whereas the thickness and strength of the lithospheric mantle is very strongly age dependent. Depth- and temperature-dependent rheological models show that the mechanically weak lower crust separates the mechanically strong upper crust and upper lithospheric mantle (e.g., Watts and Burov, 2003). The EET bands for the mechanically strong upper crust and upper lithospheric mantle describe the integrated EET of the continental lithosphere that has a bearing on its response to loads imposed on it. The degree of coupling and/or decoupling between these two mechanically strong layers plays an important role in the structural style compressionally deformed extensional basins.

The bulk-strength of the heterogeneous European continental lithosphere is directly linked to its thermo-tectonic age (Cloetingh and Burov, 1996; Cloetingh et al., 2005b; Perez-Gussinye and Watts, 2005). An understanding of the temporal and spatial strength distribution in the NW European lithosphere may offer quantitative insights into the patterns of its intraplate deformation (basin inversion, up-thrusting of basement blocks), and particularly into the pattern of lithosphere-scale folding, as described by Ziegler et al. (1995, 1998).

Owing to the large amount of high-quality geophysical data acquired during the last 20 years in Europe, its lithospheric configuration is rather well known; though significant uncertainties remain in many areas concerning the seismic and thermal thickness of the lithosphere (Artemieva, 2006, 2011a, 2011b; Artemieva and Mooney, 2001; Babuska and Plomerova, 1992). Nevertheless, available data help to constrain the rheology of the European lithosphere, thus enhancing our understanding of its strength.

Strength envelopes and the effective elastic thickness of the lithosphere have been commonly calculated for a number of locations in Europe (e.g., Cloetingh and Burov, 1996). However, as such calculations were made for scattered points only, or along transects, they provide limited information on lateral strength variations of the lithosphere. Although lithospheric thickness and strength maps have already been constructed for the Pannonian Basin (Lankreijer et al., 1999) and the Baltic Shield (Kaikkonen et al., 2000), such maps were until recently not yet available for all of Europe.

As evaluation and modelling of the response of the lithosphere to vertical and horizontal loads (e.g., Toussaint et al., 2004) requires an understanding of its strength distribution, dedicated efforts were made to map the strength of the European lithosphere by implementing 3-D strength calculations (Cloetingh et al., 2005b; Tesauro et al., 2009).

Strength calculations of the lithosphere depend primarily on its thermal and compositional structure and are particularly sensitive to

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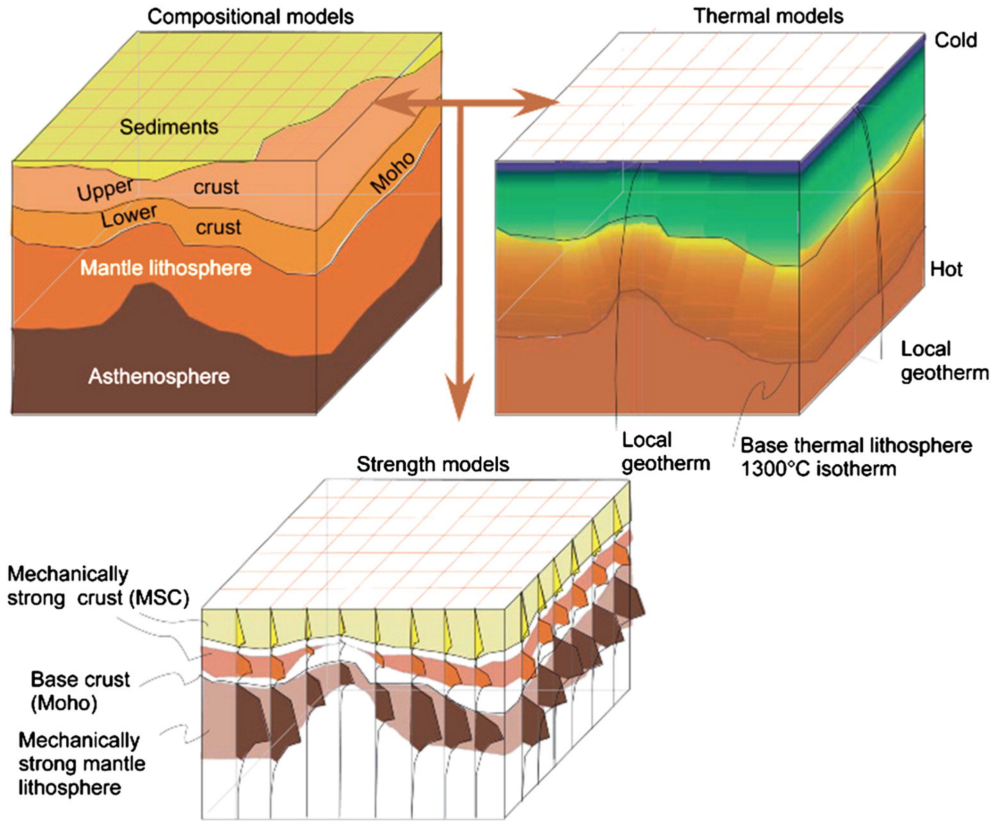

Fig. 8. From crustal thickness (top left) and thermal structure (top right) to lithospheric strength (bottom): conceptual configuration of the thermal structure and composition of the lithosphere, adopted for the calculation of 3-D strength models.
Modified from Cloetingh et al. (2005b).

thermal uncertainties (Burov, 2011; Burov and Diament, 1995; Ranalli, 1995). For this reason, the workflow aimed at the development of a 3-D strength model for Europe was twofold: (1) construction of a 3-D compositional model and (2) calculating a 3-D thermal cube. The final 3-D strength cube was obtained by calculating 1-D strength envelopes for each lattice point (x, y) of a regularized raster covering NW Europe (Fig. 8). For each lattice point, the appropriate input values were obtained from a 3-D compositional and thermal cube. A geological and geophysical geographic database was used as reference for the construction of input models.

For continental realms, a 3-D multilayer compositional model was constructed, consisting of one mantle-lithosphere layer, 2–3 crustal layers and an overlying sedimentary cover layer, whereas for oceanic areas a one-layer model was adopted. For the depth to the different interfaces several regional or European-scale compilations were available that are based on deep seismic reflection and refraction or surface wave dispersion studies (e.g., Artemieva et al., 2006; Blundell et al., 1992; Suhadolc and Panza, 1989). Information on the depth to the base of the crust, was mainly derived from the European Moho map by Tesauro et al. (2008) (Fig. 9). Regional compilation maps of the seismological lithosphere thickness were used as reference base for the thickness of the thermal lithosphere in thermal modelling (Babuska and Plomerova, 1993, 2001; Plomerova et al., 2002).

Fig. 10 shows spatial variations in integrated strength and effective elastic thickness of Europe's lithosphere inferred from forward rheological modelling (from Tesauro et al., 2009). As is evident from Fig. 10, the percentage of total lithospheric strength contributed by the crust is highly variable covering the full spectrum from jelly-sandwich to crème brûlée end-member rheologies (Burov and Watts, 2006).

Europe's lithosphere is characterized by major spatial mechanical strength variations, with a pronounced contrast between the strong Proterozoic lithosphere of the East-European Platform to the east of the Teisseyre–Tornquist line and the relatively weak Phanerozoic lithosphere of Western Europe. A similar strength contrast occurs at the transition from strong Atlantic oceanic lithosphere to the relatively weak continental lithosphere of Western Europe. Within the Alpine foreland, pronounced NE–SE trending weak zones are recognized that coincide with such major geologic structures as the Rhine Graben System and the North Danish–Polish Trough, which are separated by the high-strength North German Basin and the Bohemian Massif. Moreover, a broad zone of weak lithosphere characterizes the Massif Central and surrounding areas. The presence of thickened crust in the area of the Teisseyre–Tornquist suture zone (Fig. 9) gives rise to a pronounced mechanical weakening of the lithosphere, particularly of its mantle part.

Whereas the lithosphere of Fennoscandia is characterized by a relatively high strength, the North Sea rift system corresponds to a zone of weakened lithosphere. Other areas of high lithospheric strength are the Bohemian Massif and the London–Brabant Massif both of which exhibit low seismicity. A pronounced contrast in strength can also be noticed between the strong Adriatic indenter and the weak Pannonian Basin area (see also Fig. 10).

The lateral strength variations of Europe's intraplate lithosphere are primarily caused by variations in the mechanical strength of the lithospheric mantle, whereas variations in crustal strength appear to be more modest. Variations in lithospheric mantle strength are primarily related to variations in the thermal structure of the lithosphere that can be related to thermal perturbations of the sub-lithospheric upper mantle as imaged by seismic tomography (Goes et al., 2000), with lateral variations in crustal thickness playing a secondary role, apart from the Alpine domains which are characterized by deep crustal roots. The high strength of the East-European Platform, the Bohemian and London–

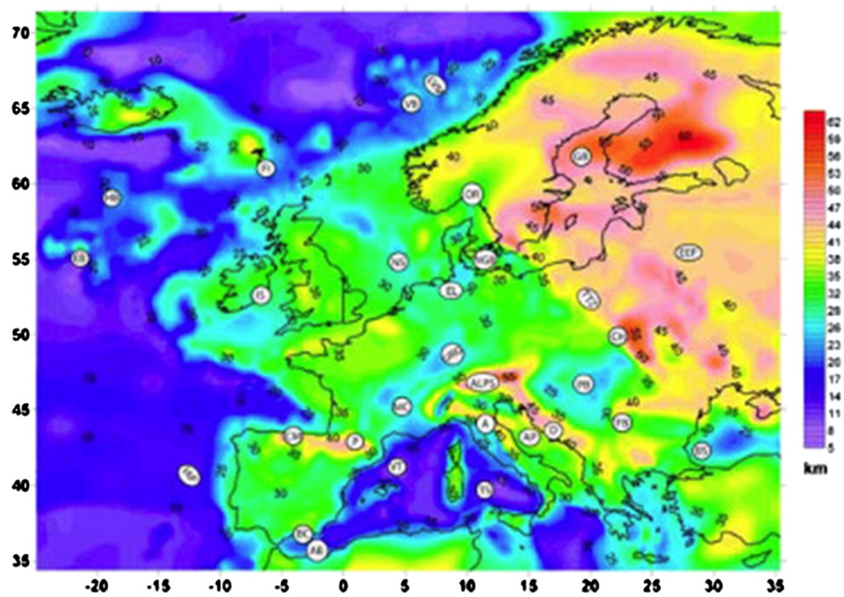

Fig. 9. Depth map of Moho discontinuity (Tesauro et al., 2008). Abbreviations are as follows: A, Apennines; AB, Alboran Basin; AP, Adriatic Promontory; BC, Betic Cordillera; BS, Black Sea; CH, Carpathians; CM, Cantabrian Mountain; D, Dinarides; EB, Edoras Bank; EL, ElbeLineament; EEP, East European Platform; FB, Focsani Basin, FI, Faeroe Islands; GB, Gulf of Bothnia; HB, Hatton Bank; IAP, Iberian Abyssal Plain; IS, Iapetus Suture; LVM, Lofoten–Vesteralen margin; MC, Massif Central; NGB, North German Basin; NS, North Sea; OR, OsloRift; P, Pyrenees; PB, Pannonian Basin; TS, Tyrrhenian Sea; TTZ, Tesseyre–Tornquist zone; URG, Upper Rhine Graben; VB, Vøring Basin; and VT, Valencia Trough.

Brabant Massifs and the Fennoscandian Shield reflects that they are underlain by old, cold and thick lithosphere, whereas the European Cenozoic Rift System coincides with a major axis of thermally weakened lithosphere. Similarly, weakening of the lithosphere of southern France can be attributed to the presence of tomographically imaged plumes rising up under the Massif Central (Granet et al., 1995; Wilson and Patterson, 2001). Furthermore, the strong Adriatic indenter contrasts with the weak lithosphere of the Mediterranean collision zone.

The major lateral strength variations that characterize the lithosphere of extra-Alpine Phanerozoic Europe are largely related to its Late Cenozoic thermal perturbation as well as to Mesozoic and Cenozoic rift systems and intervening stable blocks, and not so much to the Caledonian and Variscan orogens and their accreted terranes (Dèzes et al., 2004; Ziegler and Dèzes, 2006). These lithospheric strength variations (Fig. 10) are primarily the result of variations in the thermal structure of the lithosphere and, therefore, are compatible with inferred EET variations (see Cloetingh and Burov, 1996; Perez-Gussinye and Watts, 2005).

Crustal seismicity is largely concentrated on the presently still active Alpine plate boundaries, and particularly on the margins of the Adriatic indenter. In the Alpine foreland, seismicity is largely concentrated on zones of low lithospheric strength, such as the European Cenozoic rift system, and areas where preexisting crustal discontinuities are reactivated under the presently prevailing NW–SE directed stress field, such as the South Armorican shear zone and the fracture systems of the Bohemian Massif (Dèzes et al., 2004; Ziegler and Dèzes, 2006), and the rifted margin of Norway (Mosar, 2003).

5.3. Lithospheric folding: an important mode of compressional reactivation of rifts

Folding of the lithosphere, involving its positive as well as negative deflection (see Fig. 11), appears to play a more important role in the large-scale neotectonic deformation of Europe's intraplate domain than hitherto realized (Cloetingh et al., 1999). The large wavelength of vertical motions associated with lithospheric folding necessitates integration of available data from relatively large areas, often going beyond the scope of regional structural and geophysical studies that target specific structural provinces. Recent studies on the North German Basin have revealed the importance of its neotectonic structural reactivation by lithospheric folding (Marotta et al., 2000). Similarly, the Plio-Pleistocene subsidence acceleration of the North Sea has been attributed to folding. Lithospheric folding is a very effective mechanism for the propagation of tectonic deformation from active plate boundaries far into intraplate domains (e.g., Burov et al., 1993; Stephenson and Cloetingh, 1991; Ziegler et al., 1995, 1998, 2002).

At the scale of a microcontinent that was affected by a succession of collisional events, Iberia provides a well-documented natural laboratory for lithospheric folding and the quantification of the interplay between neotectonics and surface processes (Cloetingh et al., 2002; Fernandez-Lozano et al., 2011, 2012). An important factor supporting a lithosphere-folding scenario for Iberia is the compatibility of the thermo-tectonic age of its lithosphere and the wavelength of the observed deformations.

Well-documented examples of continental lithospheric folding also come from other cratonic areas. A prominent example of lithospheric folding occurs in the Western Goby area of Central Asia, involving a lithosphere with a thermo-tectonic age of 400 Ma. In this area, mantle and crustal wavelengths are 360 km and 50 km, respectively, with a shortening rate of 10 mm yr^{-1} and a total amount of shortening of 200–250 km during 10–15 My (Burov and Molnar, 1998; Burov et al., 1993). Quaternary folding of the Variscan lithosphere in the area of the Armorican Massif (Bonnet et al., 2000) resulted in the development of folds with a wavelength of 250 km, pointing to a mantle-lithospheric control on deformation. As the timing and spatial pattern of uplift inferred from river incision studies in Brittany is incompatible with a glacio-eustatic origin, Bonnet et al. (2000) relate the observed vertical motions to deflection of the lithosphere under the present-day NW–SE directed compressional intraplate stress field of NW Europe. Stress-induced uplift of the area appears to control fluvial incision rates and the position of the main drainage divides. This area, which is located at the western margin of the Paris Basin and along the rifted Atlantic margin of France, has been subjected to thermal rejuvenation during Mesozoic extension related to North Atlantic rifting (Robin et al., 2003; Ziegler and Dèzes, 2006) and subsequent compressional intraplate deformation (Ziegler et al., 1995) that affected also the Paris Basin (Bourgeois et al., 2007; Lefort and Agarwal, 2002, see Fig. 2B). Levelling studies in the Bretagne area (Lenotre et al., 1999) also point towards its ongoing

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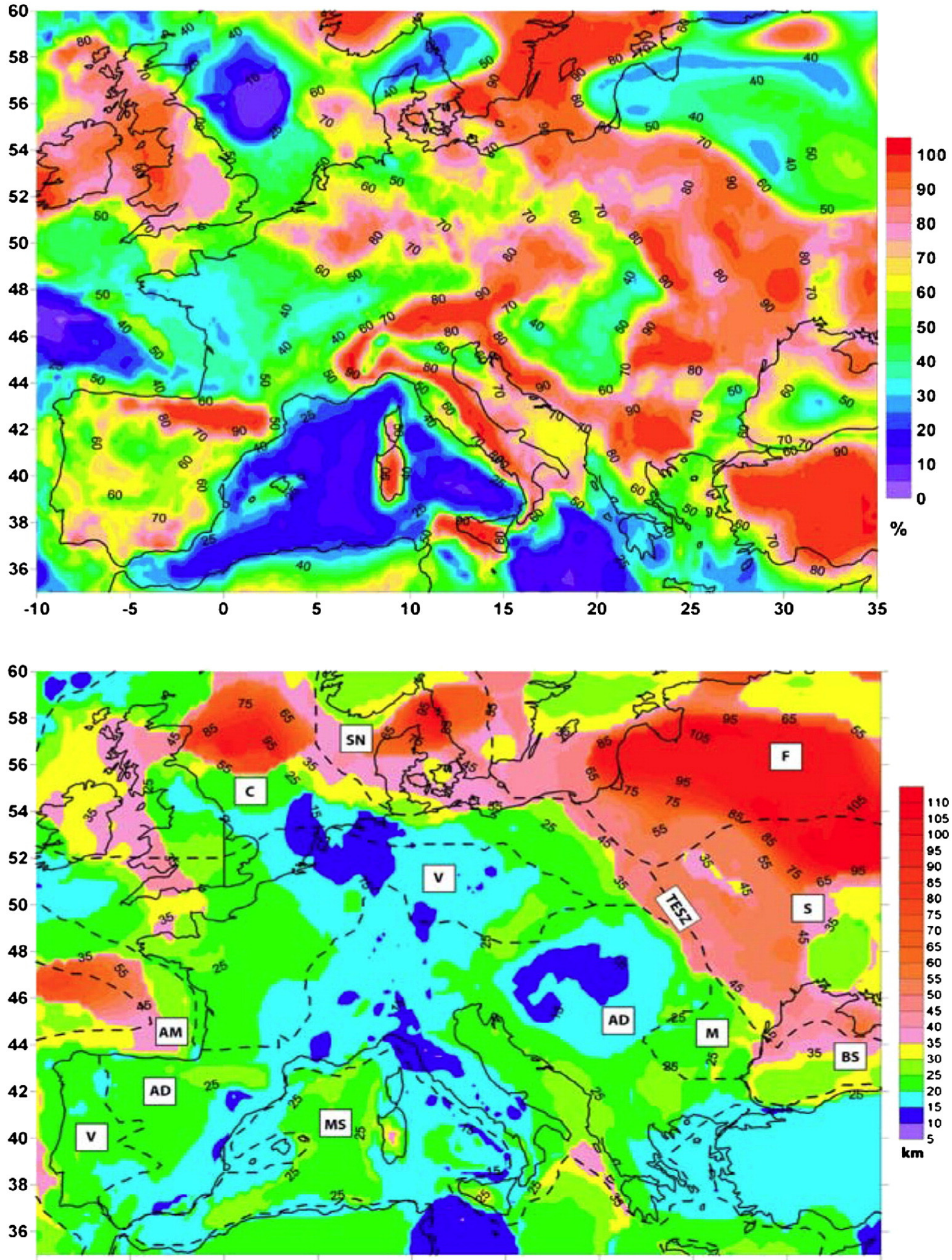

Fig. 10. Spatial variations in integrated strength and effective elastic thickness of Europe's lithosphere inferred from forward rheological modelling (from Tesauro et al., 2009). Top) Percentage of total lithospheric strength due to the crust; Bottom) Effective Elastic thickness (T_e) distribution of the European lithosphere derived from integrated strength of the lithosphere (km). Abbreviations are as follows: F, Fennoscandia; S, Sarmatia, SN, SvecoNorvegian; BS, Black Sea, M, Moesian Platform; C, Caledonides; V, Variscides; AD, Alpine Domain; AM, Atlantic Margin; MS, Mediterranean Sea; and TESZ, Trans European Suture Zone.

deformation. The inferred wavelengths of these neotectonic lithosphere folds are consistent with the general relationship that was established between the wavelength of lithospheric folds and the thermo-tectonic age of the lithosphere on the base of a global inventory of lithospheric folds (Fig. 12) (see also Cloetingh and Burov, 1996, 2011). In a number of other areas of continental lithosphere folding, smaller wavelength crustal folds have also been detected, for example, in Central Asia (Burov et al., 1993; Nikishin et al., 1993).

Thermal thinning of the mantle–lithosphere, often associated with volcanism and doming, enhances lithospheric folding and appears to control the wavelength and amplitude of folds, as seen by the Neogene uplift of the Rhenish Shield and the Vosges–Black Forest arch (Dèzes et al., 2004; Ziegler and Dèzes, 2007). Substantial thermal weakening of the lithospheric mantle is consistent with higher lithosphere growth rates of lithosphere folding in the European foreland as compared to folding in Central Asia (Nikishin et al., 1993), which is characterized by pronounced mantle strength (Cloetingh et al., 1999).

6. Models for continental breakup and rift basins

6.1. Extensional basin migration: observations and thermo-mechanical models

Several examples of extensional basins have been described in which the locus of extension shifted in time toward the zone of future crustal separation (e.g., Bukovics and Ziegler, 1985; Lundin and Doré,

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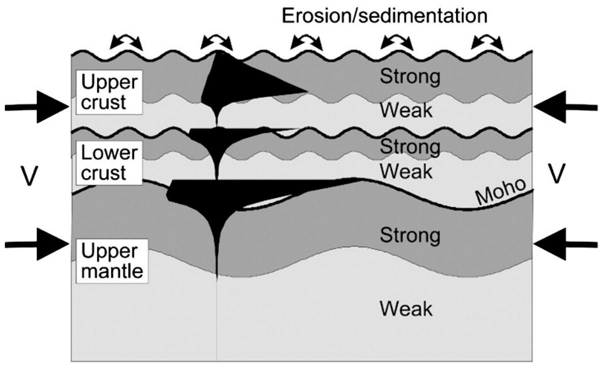

Fig. 11. Schematic diagram illustrating decoupled lithospheric mantle and crustal folding, and consequences of vertical motions and sedimentation at the Earth's surface. V is horizontal shortening velocity; upper crust, lower crust, and mantle layers are defined by corresponding rheologies and physical properties. A typical brittle–ductile strength profile (in black) for decoupled crust and upper mantle–lithosphere, adopting a quartz–diorite–olivine rheology, is shown for reference.

1997; Ziegler, 1988, 1996b). These basins typically consist of several laterally adjacent fault-controlled sub-basins, the main sedimentary fill of which often differs by several tens of millions of years. As such, they reflect lateral changes in their main subsidence phases that can be attributed to a temporal lateral shift of the centres of lithosphere extension. A well-documented example of this phenomenon is the 200 to 500 km wide Mid-Norway passive continental margin. Prior to the final Late Cretaceous–Early Tertiary rifting event that was accompanied by the impingement of the Iceland Plume and culminated in continental breakup, the Mid-Norway Vøring margin was affected by several more or less discrete rifting events (Bukovics and Ziegler, 1985; Skogseid et al., 1992; Ziegler, 1988). These resulted in the subsidence of several depocenters located between the Norwegian coast and the continent–ocean boundary (Fig. 13B). On a basin-wide scale, this passive margin consists of an initial break-away zone of probably Late Palaeozoic age located up to 150 km to the east of the coast, the Late Palaeozoic–Triassic Trøndelag Platform, located adjacent to the coast, the Jurassic–Early Cretaceous Vøring Basin in the central part of the shelf, and the marginal extended zone adjacent to the continent–ocean boundary that was active during the stretching event preceding continental breakup at the Paleocene–Eocene transition (Lundin and Doré, 1997; Mosar et al., 2002; Osmundsen et al., 2002; Reemst and Cloetingh, 2000; Skogseid et al., 2000). The Nordland Ridge, forming the western margin of the Trøndelag Platform, is a classical footwall uplift that is associated with the border fault of the Vøring Basin, whilst the Vøring Marginal High is associated with the continent–ocean boundary. Although there is still considerable uncertainty about the age of the oldest syn-rift sedimentary sequences involved in the different deep-seated fault blocks of the Mid-Norway margin and the width of the area that was affected by the pre-Jurassic extensional phases (Gabrielsen et al., 1999; Mosar et al., 2002), it is generally agreed that in time extension shifted westward, away from the Trøndelag Platform towards the Vøring Basin and ultimately centred on the continental breakup axis (e.g., Bukovics and Ziegler, 1985; Lundin and Doré, 1997; Reemst and Cloetingh, 2000). Other areas in which extensional strain concentrated in time toward the rift axis or the future breakup axis are, for instance, the Viking Graben of the North Sea (Ziegler, 1990), the nonvolcanic Galicia margin and the passively rifted South Alpine Tethys margin (e.g. Bertotti et al., 1997; Manatschal and Bernoulli, 1999).

Several hypotheses have been proposed in an effort to explain this lateral migration of rifting activity by invoking the principle that extensional strain is centred on the weakest part of the lithosphere (Steckler and Ten Brink, 1986). A mechanism for limiting extension at a given location was studied, for example, by England (1983), Houseman and England (1986), and Sonder and England (1989), who found that cooling of the continental lithosphere during stretching may increase its strength, so that deformation shifts to a previously low strain region (Sonder and England, 1989). With this mechanism, other effects such as changes in plate boundary forces are not required to explain basin migration.

Migration of the extension locus has also been attributed to temporally spaced multiple stretching phases, separated by periods during which the lithosphere is not subjected to extension but cools. Under such a scenario, the lithosphere that was weakened during an earlier stretching phase needs sufficient time to regain its strength by cooling and to become indeed stronger than the adjacent unextended area before the onset of the next stretching event. Obviously, this concept

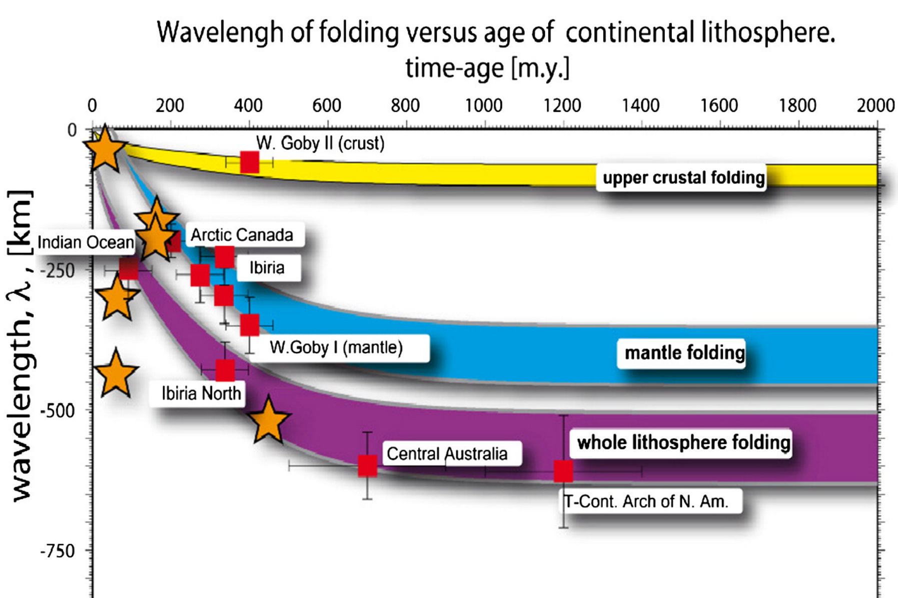

Fig. 12. Theoretically predicted wavelengths as function of thermo-tectonic age (for different lithospheric layers as well as whole-lithosphere folding). Model is compared to the observed folding wavelengths (after Cloetingh et al., 1999). Thermo-tectonic age corresponds to the time elapsed since the last major perturbation of the lithosphere prior to folding.

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requires a long period of tectonic quiescence between successive rifting events. Bertotti et al. (1997) showed in a model for the thermo-mechanical evolution of the rifted South Alpine Tethys margin that its strongly thinned parts could have indeed been stronger than the remainder of the margin. This is compatible with rheological considerations which suggest that stretched and thermally stabilized lithosphere, characterized by a thinned crust and a considerably stronger lithospheric mantle, is much stronger than unextended lithosphere (Bertotti et al., 1997; Ziegler and Cloetingh, 2004; Ziegler et al., 1995). This hypothesis requires sudden time-dependent changes in the magnitude and possibly the orientation of intraplate stress fields, controlling the different stretching and nonstretching phases. Nevertheless, evidence for tensional reactivation of rifts which were abandoned millions of years ago suggests that crustal-scale faults permanently weaken their lithosphere to the degree that under given conditions they are prone to tensional and compressional reactivation (Ziegler and Cloetingh, 2004).

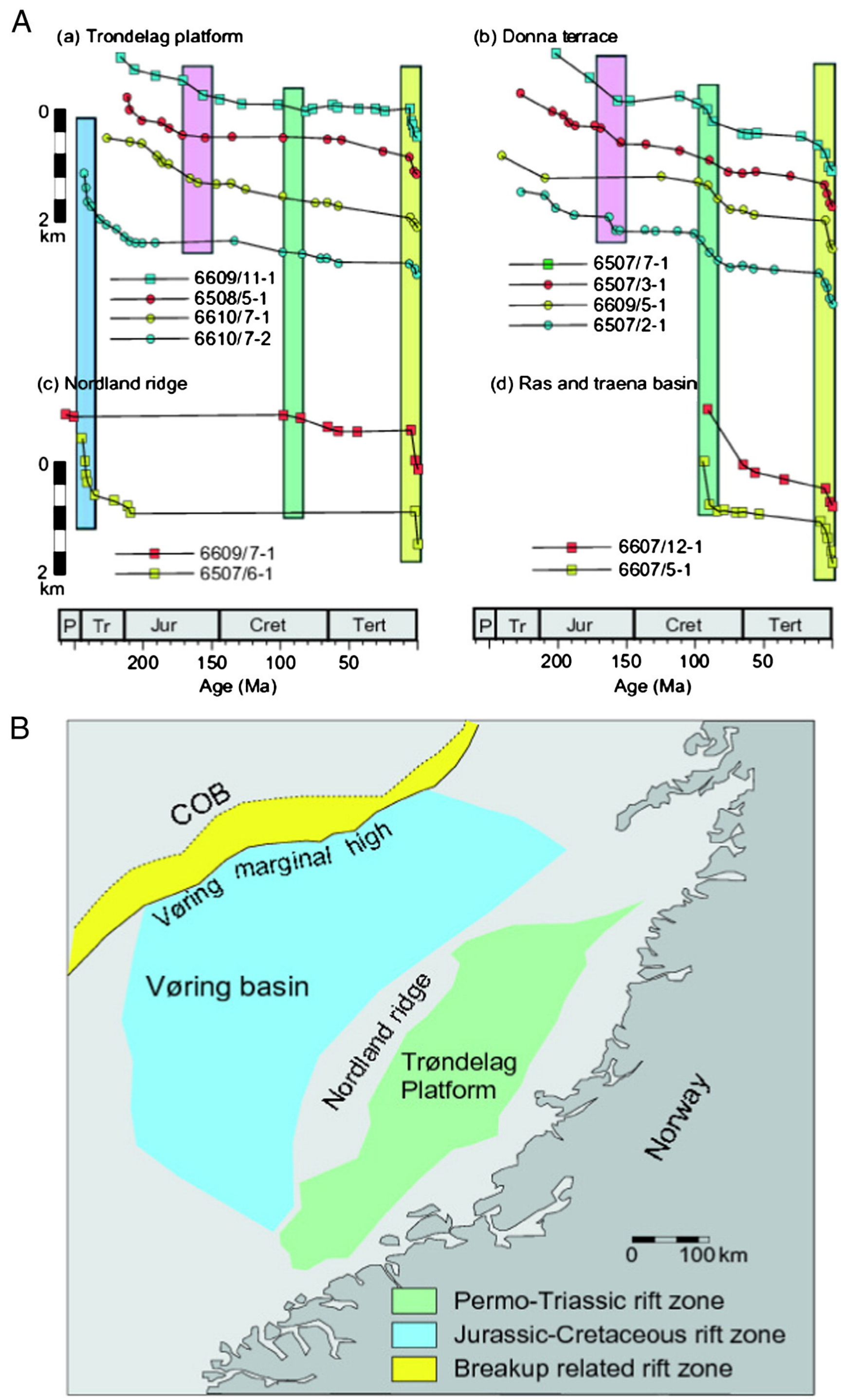

Fig. 13. A) Tectonic subsidence curves for four sub-basins and platforms of the Mid-Norwegian margin. Note the simultaneously occurring subsidence accelerations (coloured bars). The Mesozoic subsidence accelerations reflect pronounced rift-related tectonic phases, whereas the Late Neogene subsidence acceleration (light yellow bar) coincides with a major reorganization of the Northern Atlantic stress field. Modified from Reemst (1995); B) Sketch map of rift zones on the Mid-Norwegian margin showing timing of main rift activity. Modified from Skogseid and Eldholm (1995).

Sawyer and Harry (1991) and Favre and Stampfli (1992) described for the evolution of the Central Atlantic rift a gradual concentration of extensional strain towards the future zone of crustal separation. This phenomenon is attributed to the syn-rift gradual rise of the lithospheric isotherms and a commensurate upward shift of the lithospheric necking level and the intracrustal brittle/ductile deformation boundary. As a result of this, extensional strain concentrates in time on the thermally more intensely weakened part of the lithosphere, generally corresponding to the rift axis, thus causing narrowing of the rift and abandonment of its lateral graben systems (Ziegler and Cloetingh, 2004). In the case of the Central Atlantic, this process was enhanced by the impingement of a short-lived, though major, mantle plume at the Triassic–Jurassic transition (Nikishin et al., 2002; Wilson, 1997).

In the following, we discuss the implications of a viscoelastic plastic finite element model for extension of a lithosphere with an initially symmetric upper-mantle weakness (for details see Van Wijk and Cloetingh, 2002). When large extension rates are applied, deformation becomes focused, causing lithospheric necking and eventually plate separation. Hereafter, this is referred to as 'standard' rifting. A different evolution of deformation localization takes place when the lithosphere is extended at low strain rates. In this case, the necking area may start to migrate laterally, and hence delay or even prevent plate separation.

The modelled domain consists of an upper and a lower crust and a lithospheric mantle (Fig. 14) to which the rheological parameters of granite, diabase and olivine, respectively, have been assigned (Carter and Tsenn, 1987). In order to facilitate deformation localization, the crust was thickened by 2 km in the centre of the domain, forming a linear feature parallel to the future rift axis.

This causes localized weakening of the upper mantle and a corresponding strength reduction of the lithosphere. A lithosphere with this configuration is thought to be typical for orogenic belts after post-orogenic thermal equilibration of their thickened crust (Henk, 1999). The Mid-Norway margin is superimposed on the Caledonian orogen, the crust of which was presumably still thickened prior to the Late Carboniferous onset of rifting (Skogseid et al., 1992).

6.2. Fast rifting and plate separation

As the architecture of rifts that developed in response to variable extension rates ('standard necking cases') does not differ significantly, a representative simulation of (model for) lithosphere stretching at a velocity of about 16 mm yr^{-1} is presented here. At the onset of lithospheric extension deformation is localized at the centre of the domain on which the initial mantle weakness was imposed. Thinning of the crust and mantle lithosphere concentrates here, and mantle material starts to well up (Fig. 14B). Under persisting extension thinning of the crust and mantle lithosphere continues, resulting after 27 My in plate separation. Continental breakup is here defined as occurring when the crust is thinned by a factor 20, though another factor or another definition for plate separation could have been chosen. This definition corresponds to about 40–50% of extension of the model domain at plate separation. Thinning factors for the crust (δ) and lithospheric mantle (β) are shown in Fig. 14C and are defined as the ratio between the initial and the present thickness of the crust or lithospheric mantle, respectively. The base of the lithospheric mantle is the 1300 °C isotherm.

The centre of the model domain, on which the initial mantle weakness was imposed, is the weakest part; this continues to be so until plate separation. Integrated strength values of the continental lithosphere vary between 10^{12} and 10^{13} N/m with the higher values characterizing Precambrian shields (Ranalli, 1995). In our model the strength of the lithosphere falls within this range. During rifting, the strength of the lithosphere decreases with time, owing to its thinning and heating as a consequence of its stretching and nonlinear rheology.

In other cases with higher constant extension rates the localization of deformation is comparable to that discussed above. In all cases, deformation concentrates on one zone in which thinning continues until plate separation is achieved. The duration of the rifting stage preceding plate separation depends on the extension velocity. When the lithosphere is stretched at greater velocities, it takes less time to reach continental breakup. When extension velocities are less than 8 mm yr^{-1}, stretching of the lithosphere does not lead to plate separation. The dependence of rift duration on potential mantle temperature is discussed in Van Wijk et al. (2001). The configuration of the rifted basins shows no clear dependence on the tested stretching velocities (see also Bassi, 1995).

The tendency for the lithosphere to neck (or to focus strain) is weaker with decreasing extension velocities. When stresses exerted on the lithosphere are smaller, the rate of strain localization also slows down, mantle upwelling is slower and syn-rift lateral conductive cooling plays a more important role. When the lithosphere is stretched at high rates, upwelling of mantle material is fast (almost adiabatic) with little or no horizontal heat conduction. The fast rise of hot asthenospheric material further reduces the strength of the lithosphere in the central region, with the consequence that lithospheric deformation and thinning accelerates even further. The result is a short rifting stage preceding plate separation.

These model-derived conclusions on the duration of the rifting stage preceding plate separation need to be viewed in the context of natural examples as summarized in Fig. 5 (Ziegler, 1996b; Ziegler and Cloetingh, 2004; Ziegler et al., 2001).

6.3. Thermo-mechanical evolution and tectonic subsidence during slow extension

The lithosphere reacts differently to low and high extension rates (Burov, 2007; Huismans and Beaumont, 2003; Van Wijk and Cloetingh, 2002). Results of a representative model simulating an extension rate of 6 mm/yr over a period of 100 My are here presented to illustrate the thermal evolution of the lithosphere (Figs. 14 and 15). The total strength of the lithosphere, obtained by integrating the stress field over the thickness of the lithosphere (Ranalli, 1995), is also shown in Fig. 16. During the first 30 My after the onset of stretching, deformation localizes in the centre of the model domain where the lithosphere was pre-weakened (Figs. 14–16). The top of the crust subsided, a sedimentary basin developed and mantle material welled up. Subsequently, as lithospheric stretching proceeds, temperatures begin to decrease (see time slices corresponding to 45 My, 50 and 60 My, Fig. 15), in contrast to what happened in the standard necking case shown in Fig. 14. Development of the mantle upwelling zone ceases and the lithosphere cools in the centre of the domain. Cooling of the central zone continues while after 70 My increasing temperatures are evident on both sides of the previously extended central zone. As these new upwelling zones develop further (Fig. 15, 110 My panel) two new basins develop adjacent to the initial stage basin. Temperatures in the lithosphere are now lower below the initial basin as compared to the surrounding adjacent lithosphere; a 'cold spot' is present in an area that underwent initial extension (see also corresponding lithosphere strength evolution in Fig. 16). This thermal structure is reflected by the surface heat flow, which is the surface heat flow values are lower in the first stage basin than in the adjacent areas at 110 My along cross-section normal to the rift axis. Burov (2007) suggested that slow extension may be associated with the development of gravitational instabilities in the lower lithospheric mantle along the margins of the area of mantle lithosphere thinning, resulting in its sinking into the asthenosphere, and, hence, in "gravitational" thinning of the lithosphere by mantle lithosphere delamination (Fig. 17). These instabilities develop because the mantle lithosphere is colder and thus denser than the upwelling hot asthenosphere (1330 °C) while at each moment of time the degree of assimilation of the mantle lithosphere by the asthenosphere and its density contrast with the later is limited by the ratio of the advection to diffusion rate (Peclet number). If the extension rate is low enough, the characteristic time scale of growth of basal mantle–lithosphere gravitational

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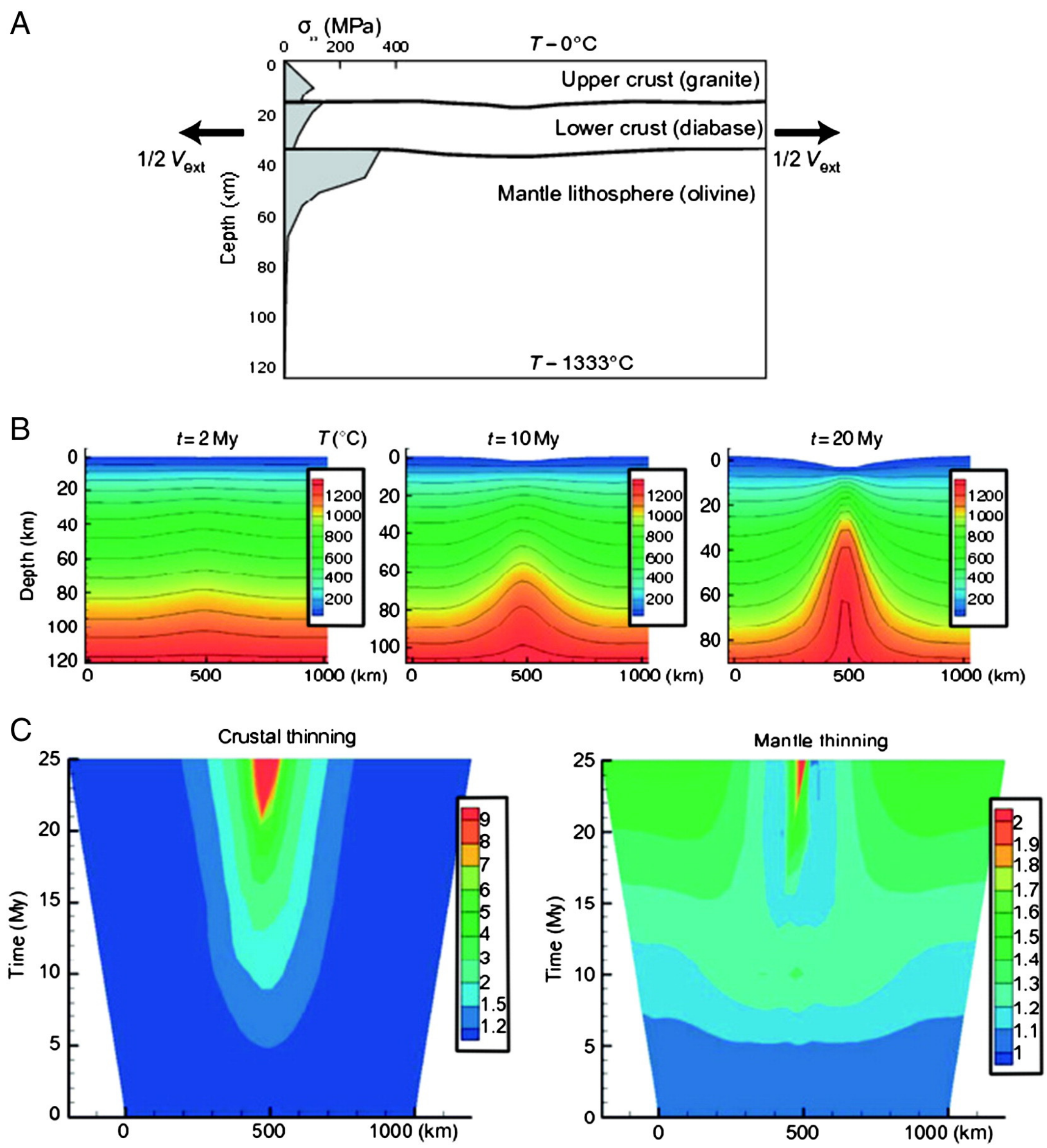

Fig. 14. Modelled thermal evolution of the lithosphere and evolution of thinning factors for a stretching rate of $V_{ext} = 16$ mm/yr. Modified from Van Wijk and Cloetingh (2002). A) Initial model configuration; B) Thermal evolution at 2, 10, and 20 My after the onset of stretching; Note the changing horizontal and vertical scales in the panels indicating the changing sizes of the model domain upon stretching; C) Evolution of thinning factors of crust (δ) and mantle lithosphere (β) for $V_{ext} = 16$ mm/yr. Breakup occurs after 27 My. Width of the model domain (horizontal axis) versus time (vertical axis). The width of the model domain increases as the lithosphere is extended.
Modified from Van Wijk and Cloetingh (2002).

instabilities becomes comparable or even smaller than the characteristic time scale of the syn-rift phase. These instabilities may eventually grow even at higher rates than the rate of other processes.

Hence the Rayleigh –Taylor instabilities may interplay with rifting processes, resulting in asymmetric rifting and/or in increased amount of thinning, specifically in the mantle part of the lithosphere. Mantle lithosphere delamination may also eventually lead to additional partial melting resulting from the ascent of the replacing asthenospheric material to sub-Moho depths (<50 km). It can be further suggested that such delamination-caused partial melting can be generated even under non-volcanic margins. One of the other consequences of the mantle lithosphere instabilities is the development of series of periodic changes in subsidence associated with delamination of dense mantle. To characterize the relative role of the gravitational instabilities during rifting, Burov (2007) introduced a special parameter, the “rift Deborah number”:

$$De_r = 13.04\ \mu_{eff} u_x/(\Delta\rho g d\ \ L)$$

where μ_{eff} is effective viscosity. Assuming a ductile flow law:

$$De_r = 13.04\mu_0\, exp(H/nRT)u_x/((\Delta\rho_c + \alpha\rho_0\Delta T)gd\ \ L)$$

with $\mu_0 = e_{II}^{d(1-n)/n}(A^*)^{-1/n}$, $e^d_{II} = (Inv_{II}(e_{ij}))^{1/2}$ is the effective strain rate and $A^* = 1/2A \cdot 3^{(n+1)/2}$ is the material constant, H is the activation enthalpy, R is the gas constant, and n is the power law exponent. L is the final width of the rift and u_x is the extension rate, $\Delta\rho_c + \alpha\rho_0\Delta T = \Delta\rho$, $\Delta\rho_c$ is the compositional density contrast, α is the coefficient of thermal expansion, ρ_0 is the density at reference temperature and ΔT is the temperature change in respect to the reference temperature. Burov (2007) has shown that gravitational instabilities play an important role for $De_r < 10$, specifically for $De_r < 1.5$. It is also noteworthy that Moho depth and geometry are largely controlled by extension and the evolution of mantle-lithospheric gravitational instabilities. In particular, slow extension may result in subsidence of the Moho during the syn-rift phase. Fig. 18 shows a critical case at an extension rate of 10 mm/yr corresponding to $De_r \sim 1$, below which, inclusively, rifting

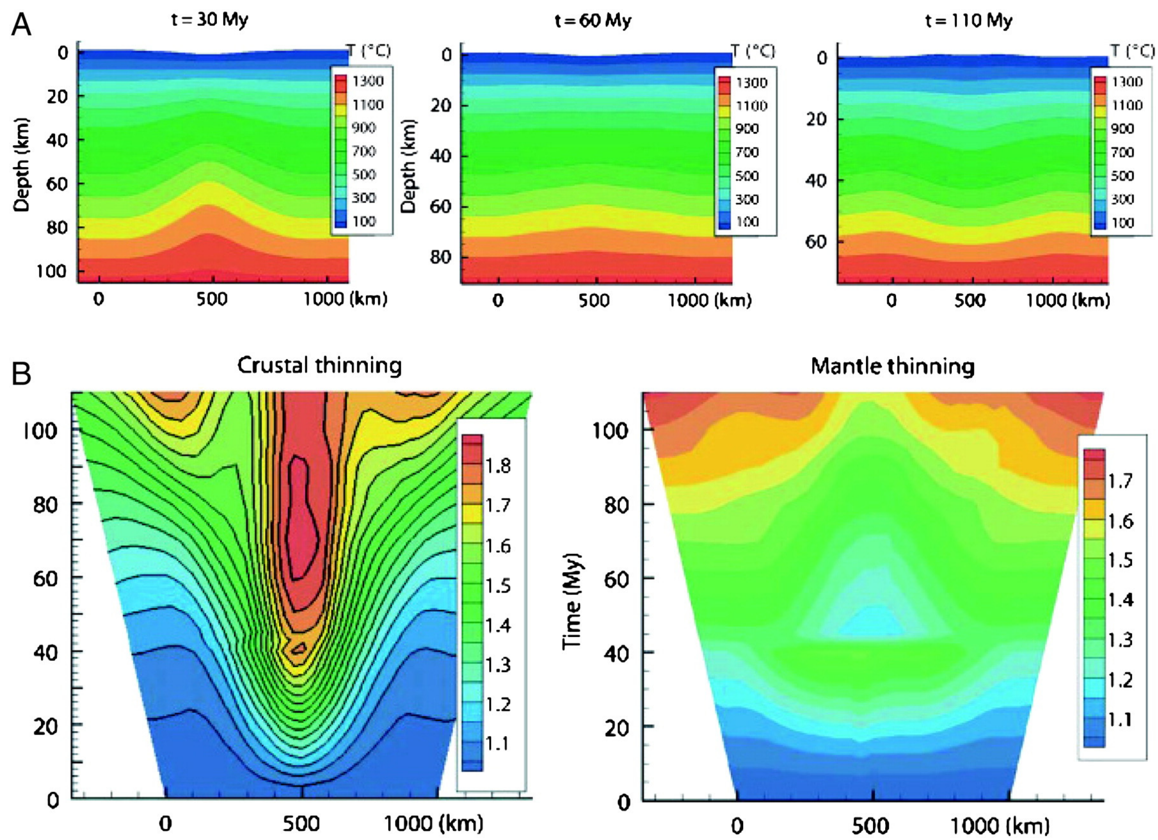

Fig. 15. A) Thermal evolution of the lithosphere for a migrating rift at an extension rate of $V_{ext} = 6$ mm/yr, times in My after the onset of stretching.; B) Evolution of thinning factors for the crust (δ) and mantle (β) at an extension rate of $V_{ext} = 6$ mm/yr. No continental breakup.
Modified from Van Wijk and Cloetingh (2002).

can be treated as "slow" in terms of the impact of gravitation), so that Moho remains almost flat or goes down instead of rising up. Hence, in the presence of a deep lithospheric necking level, and under low extension rates, the Moho may remain stable or even subside in response to lower mantle-lithosphere gravitational instabilities. This does not necessarily mean that upon detachment of significant volumes of lithospheric mantle, the lithosphere rebounds and the graben flanks can be uplifted, even in the absence of further extension. The detachment may occur in pace with thermal re-equilibration of the lithosphere so that the density contrast between the latter and the underlying mantle can be quite small at the moment when the detachment finally takes place. The models also show that the detachment may laterally migrate in the direction of the

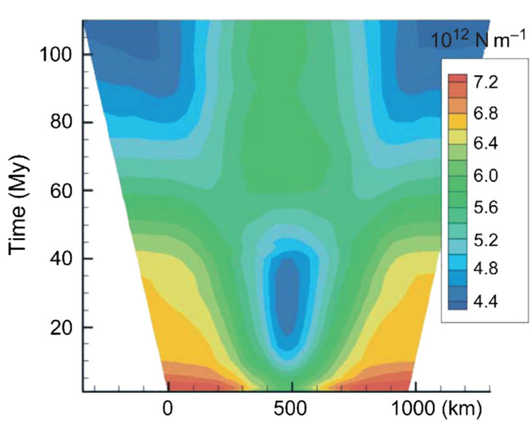

Fig. 16. Lithosphere strength evolution for a migrating rift, $V_{ext} = 6$ mm/yr.
Modified from Van Wijk and Cloetingh (2002).

extension at rates exceeding the extension rate. Hence, for relatively high extension rates it may occur at some distance from the rift flanks.

It can be questioned if the impact of the gravitational instabilities will be smaller in case of differential stretching of the lithosphere with mantle β-factors stronger than the crustal ones. β-Factors, however, are not connected to a particular mechanism of lithospheric thinning: the gravitational instabilities that erode mantle lithosphere in addition to the effect of pure or simple shear mechanism are by themselves a cause of differential stretching.

Thinning factors for the crust and mantle lithosphere corresponding to the experiments shown in Figs. 15–16 vary at a large extent. Thinning of the crust starts, as expected, in the central weakness zone of the domain where a single basin forms, with a maximum thinning factor of 1.85 for the crust. Crustal thinning in the central basin continues until about 65 My after the onset of stretching, at which time the locus of thinning shifts towards both sides of the initial basin. The zones of second-stage maximum thinning are located at a distance of about 500 km from the centre of the initial rift. Although stretching of the lithosphere continues after 65 My, extensional strain is no longer localized in the initial rift basin but is centred on the two flanking new rifted basins. The thinning factor of the mantle lithosphere reflects this behaviour. During the first 45 My of stretching upwelling mantle material is primarily localized in the central zone rather than in its surroundings. Thereafter, however, temperatures decrease rapidly in the central zone (see also the associated lithosphere strength variations in Fig. 16 as new upwelling zones develop on its flanks with decreasing activity in the area of the first basin and increasing upwelling beneath the new basins.

After the onset of stretching (Fig. 16) the central part of the model domain progressively weakens, reaching its minimum strength by 30 My. Thereafter its strength increases, but remains lower than the rest of the domain. From 55 My onward, the smallest values of

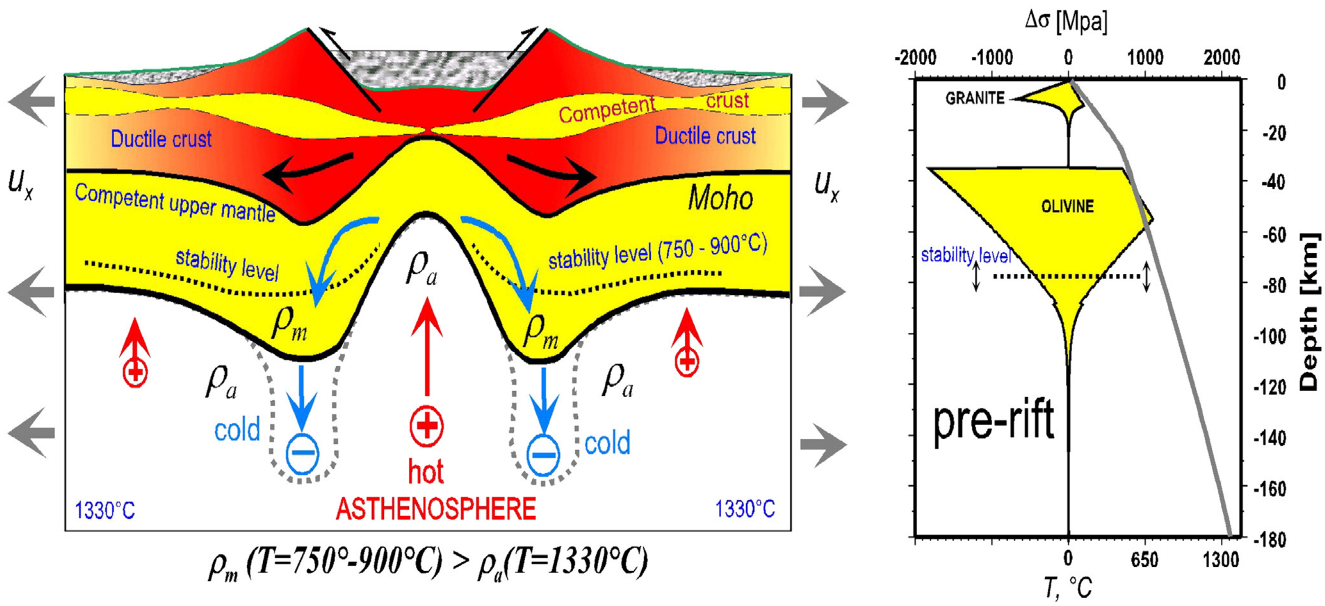

Fig. 17. Simplified cartoon showing interaction between the asthenosphere and mantle lithosphere during slow rifting (Burov, 2007). Left: The mantle lithosphere below the rift flanks becomes gravitationally unstable due to the negative density contrast with the hot asthenosphere and the loss of mechanical resistance resulting from lateral heat transfer from the asthenosphere below the rifted zone. The asthenosphere is positively unstable because of its positive density contrast with the embedding mantle lithosphere. Since the viscosity of the mantle lithosphere is exponentially dependent on temperature, there is a relatively net "stability level" that separates the regions of high viscosity from thermally weakened regions of low viscosity. These regions are subject to development of rapid RT instabilities. Right: simplified yield-stress pre-rift envelope of the continental lithosphere and typical pre-rift temperature profile.

lithospheric strength occur on both sides of the central basin beneath the second-stage rifts. By comparing the strength of the lithosphere with its thermal structure (Fig. 16), the strong dependence of strength on the temperature is evident. From this modelling study it is concluded that lateral rift migration may occur under conditions of long-term low-strain-rate lithospheric extension. Similar results have been also obtained in the earlier study by Burov and Poliakov (2001) who have demonstrated a potentially strong impact of lithospheric strengthening due to mechanical coupling of rheological layers occurring at some degree of thinning, specifically when the rate of cooling of the lithosphere is compatible or higher than the advection rate. Even if we are unaware of present-day examples of two younger rifts flanking an older rift, the exposed mechanism could have played an important role in past rifting tectonics. Its viability can be demonstrated, for example, on the case of the East-African rift, which is subdivided by strong Tanzanian craton into two simultaneously localized rifts (Western and Eastern branches, e.g., Corti et al., 2007).

6.4. Interplay of lower crustal flow and surface erosion in rifts

Surface processes play a very important role during both syn-rift and post-rift evolution (Burov and Cloetingh, 1997; Burov and Poliakov, 2001), not only in terms of their effect of thermal blanketing causing temperatures to rise in the sedimentary cover, but also due to redistribution of surface loads (Fig. 19). Surface processes modify topography and sedimentary rates comparable with the rate of the tectonic uplift/subsidence (a few 0.1 mm/yr), leading to the erosion of thousands meters of topography from uplifted rift flanks and deposition of equivalent amount of sedimentary infill. The associated dynamic loading and unloading exerted on the supporting crust and lithosphere are significant, suggesting possible feedbacks between the surface and tectonic processes. In particular, an increase of sedimentary loading leads to localised yielding and deflection of the supporting lithosphere. At the same time, erosional unloading of rift shoulders leads to their flexural rebound and strengthening. These processes, resulting in acceleration of the subsidence of the rift "neck" (strongest layer of the thinned lithosphere) and uplift of the rift shoulders, create pressure gradients sufficient to drive ductile flow in the low-viscosity lower crust (Fig. 19, Burov and Cloetingh, 1997). Ductile flow occurs until the ductile channel thickness is greater than a few km (Burov and Cloetingh, 1997). This flow is also largely gravity driven, due to lateral Moho gradients associated with crustal thinning. Hence, the flow rate first progressively increases as the crust thins, and then decreases when there is no more ductile crustal material below the rift zone. This flow, directed outward from the centre of the basin may facilitate uplift of the rift shoulders (Fig. 20, Burov and Poliakov, 2001). It may even drive some post-rift "extension" (Burov and Poliakov, 2001). In the limiting case of slow erosion and sedimentation rates, gravitational stresses can reverse the flow, resulting in retardation of basin subsidence rate, homogenisation of the crustal thickness, accelerated collapse of the shoulders and in some post-rift "compression". These effects may significantly change predictions of basin evolution inferred from the conventional stretching models. It is noteworthy, however, that it is rather difficult to refer to direct observations allowing for testing crustal flow models, as well as any models of subsurface dynamics, against natural cases. Model-driven hypothesis can be verified indirectly by showing that simpler models fail to explain vital observations such as poly-phase subsidence, post-rift doming and flank uplift.

6.5. Breakup processes: timing and mantle plumes

Continental breakup is a consequence of plume–lithosphere interactions when a low density, high temperature and low viscosity mantle plume, separated from large-scale convective motions, rises up from the core–mantle boundary to the base of the lithosphere (e.g., Sleep, 2006) applying normal and basal shear stresses and producing magmatism associated with decompressional partial melting and thermo-mechanical erosion that eventually lead to thinning and rupture of the overlying plate. This upwelling, if continuously fed from a deeper source area, forms a "superplume", i.e. a persisting diapir of very large scale coming from the D″ boundary (e.g. Condie et al., 2000; Romanowicz and Gung, 2002). There may be also smaller scale

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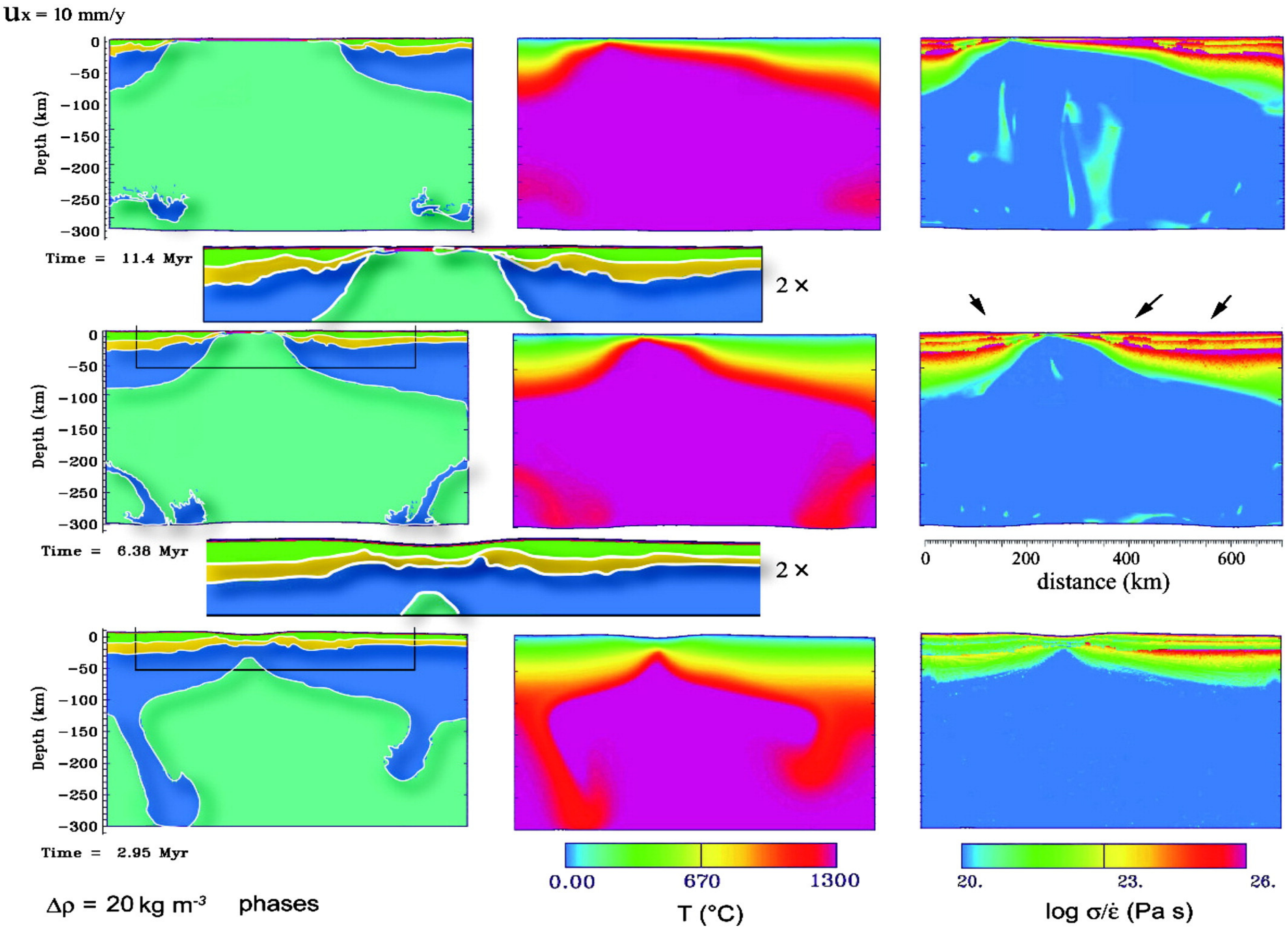

Fig. 18. Results of the experiments (Burov, 2007) for extension rate (10 mm/yr, Der 1) assuming compositional density contrast between mantle and asthenosphere, $\Delta\rho c = 20$ kg m-3. As suggested by the corresponding value of Der (1), 10 mm/yr is a rate below which the rifting can be treated as "slow" in terms of the possibility of the development of gravitational instabilities. Left: evolution of the material field (colour code: upper crust — salad-green; middle-lower crust — yellow; mantle lithosphere — blue; asthenosphere — marine-green; syn-rift sediment or otherwise reworked material — purple). Centre: temperature field. Left: logarithm of stress to strain ratio (Pa s), which is equivalent to the effective viscosity for the ductile domains. Note: 1) removal of a large part of the mantle lithosphere by RT instabilities leads to change of style of rifting; 2) Strong asymmetric rifting at developed stages of extension. 3) Breakup and oceanization of the lithosphere that commences at 9 Myr and is preceded by coupling of the crustal and mechanical layers within the lithosphere. 4) Exhumation of small amounts of lower crust and mantle. Arrows show the position of the interior basins. Inserts correspond to 2× blow-ups of the framed areas. Purple (reworked) material most often directly overlies its source material, i.e. purple above green layer means material reworked from the upper crust. Purple above blue means either reworked mantle or sediment derived from the upper and lower crust. Note: 1) removal of a large part of the mantle lithosphere by RT instabilities leading to change of style of rifting; 2) strong asymmetric rifting at developed stages of extension (at the oceanization stage); and 3) exhumation of continental and oceanic mantle and punctual exhumation of lower crustal material at ocean–continent limits. In this experiment, continental break up and oceanization commence at 6 Myr. Arrows show the position of the interior basins.

short-lived diapirs originating from different depths in the upper mantle (Courtillot et al., 2003; Montelli et al., 2004; Ritter, 2005). The resulting thermo-mechanical consequences for surface geodynamics and geology appear to be largely dependent on not only plume characteristics, but also on the thermo-rheological and density structure of the lithosphere (Burov and Cloetingh, 2010; Burov and Guillou-Frottier, 2005; Burov et al., 2007). Both the thermo-tectonic age and rheological stratification of the lithosphere have a strong impact on the effect on plume impingement (Burov and Guillou-Frottier, 2005). As a result, the presence of low-viscosity ductile lower crust may result in essential damping of the long-wavelength dynamic topography and appearance of short-wavelength tectonic scale features, associated with tensional and compressional instabilities in the crustal layers. The assumption of a horizontally uniform lithosphere at the site of future break-up is also not realistic, in view of the abundant evidence from the geological record that incipient rifts and rifted margins are usually localized at suture zones separating stronger lithosphere. Examples include the Caledonides suture of Laurentia–Greenland and Baltica, localizing Late Carboniferous and Permo-Triassic rifting and subsequent continental break-up around 65 Ma in the Arctic–Northern Atlantic and the rift systems created at the edges of the African cratonic lithosphere (Corti, 2005; Janssen et al., 1995). It should be noted that some examples do not always fit these assumptions. For instance, the Labrador Sea opened orthogonally to the basement grain, whereas the North Sea rift cuts obliquely across the Caledonian basement grain (Ziegler, 1988). Plume impingement may also occur at the onset of the post-rift phase or even later. In this case the rifted margin lithosphere has been thermally reset to very young thermo-mechanical ages for the thinned continental lithosphere and the adjacent oceanic lithosphere.

The interplay between extension and magmatism during continental breakup is still debated and recent numerical modelling studies suggest that the volumes of melts extruded at volcanic margins may also be generated by 'standard' thermal conditions, provided high extension rates can be implied (Huismans and Beaumont, 2011; Van Wijk et al., 2001).

Volcanic margins such as the Mid-Norway margin are a representative part of the Norwegian–Greenland Sea rift along which crustal separation between NW Europe and Greenland was achieved in earliest Eocene times (Mosar et al., 2002; Torsvik et al., 2001). The Norwegian–Greenland Sea rift, which had remained intermittently active for some

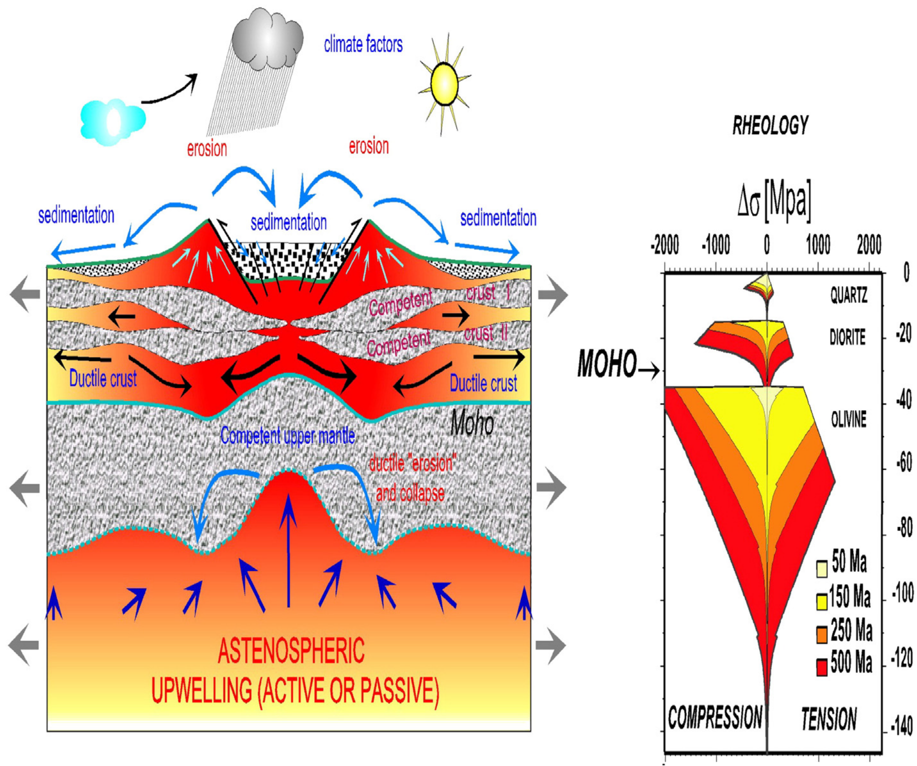

Fig. 19. Proposed scenario of basin subsidence: sediments derived from erosion on slopes of rift shoulders result in increase of loading in the basin. Strong parts of the crust and of mantle lithosphere bend and undergo flexural weakening at the inflexion points (Burov and Cloetingh, 1997). As result, the equivalent elastic thickness, EET, drops beneath the basin and rift shoulders (bottom) and becomes lower than the EET immediately after extension. Lower crustal material flows from the centre of basin towards the shoulders facilitating their uplift.

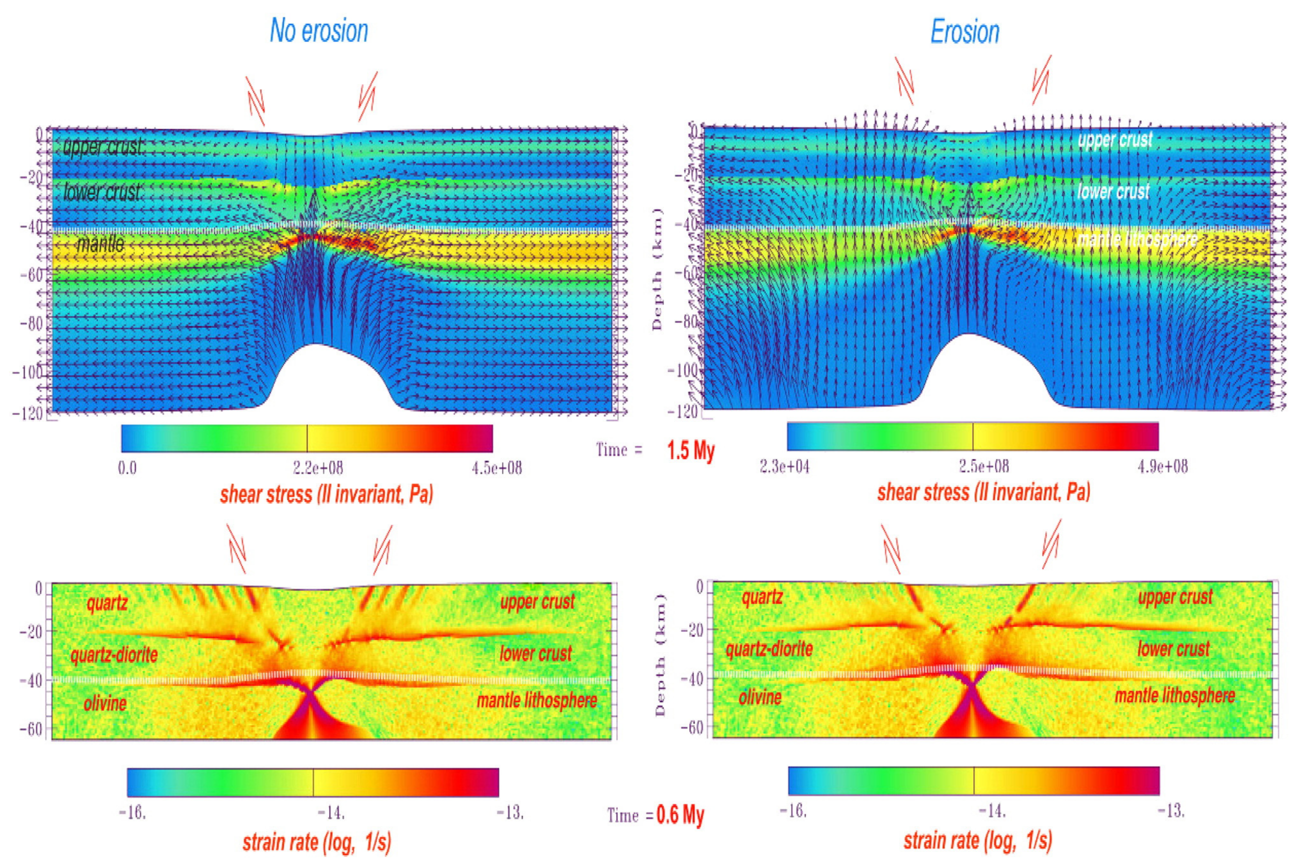

Fig. 20. Numerical experiments showing impact of surface processes on syn-rift evolution (modified from Burov and Poliakov, 2001). Shown are two cases, with and without erosion. Thermo-tectonic age of the lithosphere in this experiment is 400 Ma, the rheological envelope includes 3 layers: 2 layer crust (quartz-dominated upper crust and quartz–diorite lower crust) and dry olivine mantle. Note that erosion accelerates rift flank uplift and allows localization of deformation on major border faults.

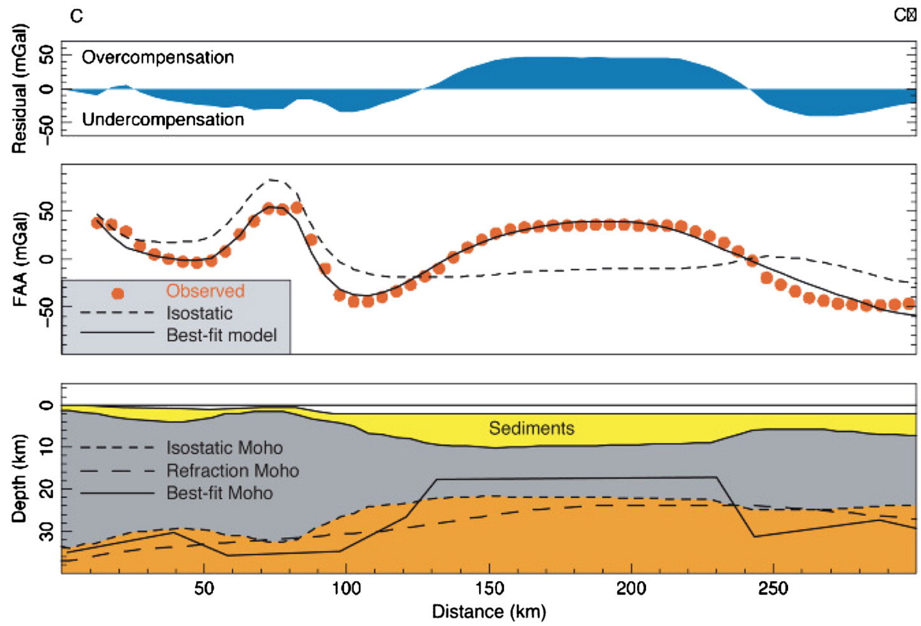

Fig. 21. Results of gravity modelling for the Eastern Black Sea, demonstrating isostatic flexural overcompensation in the centre of the basin. For location see Fig. 24, line C–C′. Modified from Cloetingh et al. (2003).

280 Ma from the Late Carboniferous until the end of the Palaeocene, forms part of the Arctic–North Atlantic rift (Larsen et al., 1999; Skogseid et al., 2000; Ziegler, 1988). Tomographic data image a mantle plume beneath Iceland rising up from near the core–mantle boundary (Bijwaard et al., 1998; Rickers et al., 2013). It should be noted that a temporal relationship between plume activity and the duration of rifting is lacking (e.g. Ziegler et al., 2001).

6.6. The evolution of the Moho geometry during rifting

There are only few data that allow to constrain the evolution of the Moho boundary during the syn-rift phase (e.g., Figs. 21, 22). According to conceptual models (Fig. 4), the Moho of a rift zone is uplifted in case of a deep necking level (strong lithospheric mantle) or downwarped in case of a shallow necking level (weak lithospheric mantle). It is noteworthy that this is independent from plume-driven active or regional stress-driven passive rifting (Ziegler, 1992, 1994) but is related to the strength of the lithosphere that is subjected to extension. As mentioned in previous sections, in nature rifting might never occur in a purely active or passive mode. At least, the presence or absence of surface expression of syn-or pre-rift volcanic activity below the rifted continental crust cannot be directly associated with a particular mechanism of rifting (e.g., Huismans and Beaumont, 2011). Extension that started in a passive mode may result in convective instability in the underlying

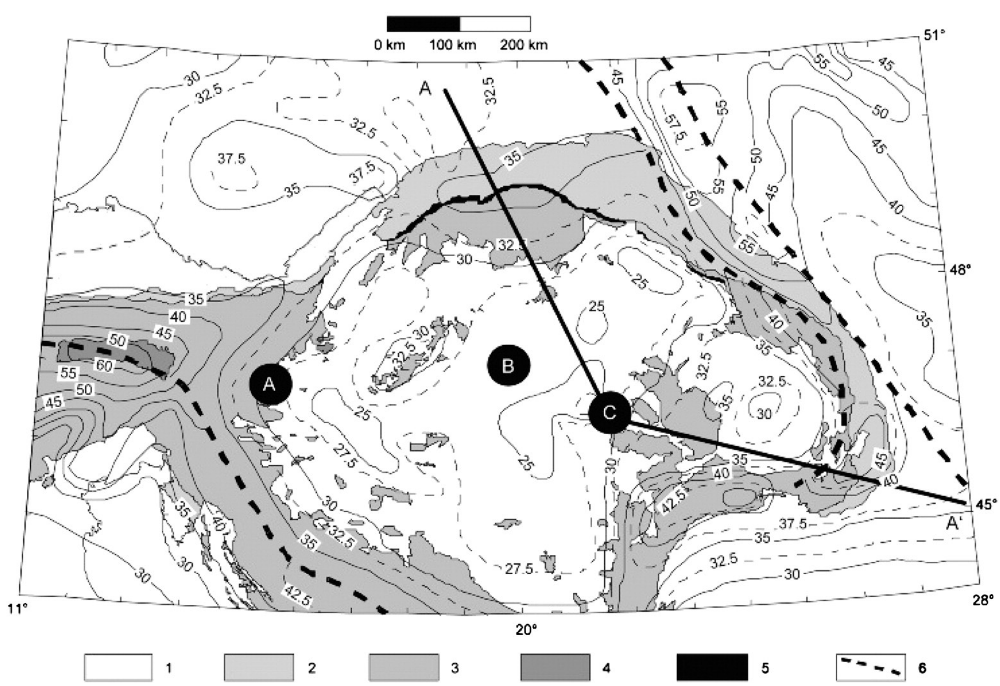

Fig. 22. Crustal thickness in the Pannonian Basin and the surrounding mountains. Values are given in km. 1: foreland (molasse) foredeep; 2: flysch nappes; 3: pre-Tertiary units on the surface; 4: Penninic windows; 5: Pieniny Klippen Belt; and 6: trend of abrupt change in crustal thickness.
Modified from Horváth et al. (2006).

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mantle and lead to further development of rifting in an active mode. Inversely, initially active rifting may affect far-field force balance leading to modification of the regional tectonic force field. Decompressional melting may be produced in all of these scenarios. But, since the mechanisms of melt ascent to the surface depend on many additional conditions, it cannot be taken for granted that melts will be unambiguously present at surface or at shallow depths. The rheological stratification of the lithosphere implies multiple levels of necking (Burov and Poliakov, 2001; 2003) with not a single dominant level during the entire duration of rifting. At different stages of rifting different rheological layers may "work" as temporal dominant levels of necking. As shown by (Burov and Poliakov, 2001; Burov, 2007, Fig. 23), lithospheric necking is initially concentrated in the deepest rheologically competent layer, for example, the lithospheric mantle. This happens because pure shear thinning at lower levels provokes passive and, later, active upwelling of hot asthenosphere below the necking area, which dramatically accelerates thinning of the bottom layer of remnant lithospheric mantle or crust. If the lithosphere initially has a strong mantle layer, the starting level of necking is deep and the Moho may be downwarped (i.e., have a concave-down shape, not necessarily with downward displacement into the mantle) during the initial phase of rifting. As soon as the bottom layer is ruptured (according to models, Burov and Poliakov, 2001), 30–50 km of extension might be sufficient), the level of necking switches to the shallower rheological layer (eventually lower crust), and the Moho boundary is warped upward. There may be about 3 or even 4 levels of consecutive necking during the syn-rift phase. Consequently, 3 to 4 different phases of syn-rift evolution (vertical acceleration and deccelaration of subsidence and of rift opening) should be reflected in the syn-rift subsidence data. In some cases, initially decoupled mechanical layers can "weld" with each other as the crust becomes thinner resulting in a single layer of necking and stronger resistance to extension. This eventual structural increase of the integrated strength during rifting may result in deceleration of subsidence and in the formation of new rifting zones at the borders of the extended area.

In theory, model predictions should be verified by observational facts before the model can be presented as a viable mechanism. In practice, the geological observations are sometimes scarce and generally of insufficient quality to provide a robust measure of validity of the models. Physical laws, however, make a part of reliable observations. Hence, even though it can be a problem connecting model predictions to natural cases, the fact that state-of-the-art numerical tectonic models are physically consistent makes their predictions probably more relevant to nature than intuitive conceptual or semi-analytical considerations. Numerical models dismiss, for example, the common intuitive assumption that faulting results in significant weakening of the crust. Instead, the models demonstrate that normal faulting reduces the initial bulk strength by no more than 40% (assuming Mohr–Coulomb behavior of brittle rocks at geological scales). Hence, without additional strength softening, faulting is not an efficient strength-reduction mechanism. The other point that can be questioned refers to the intuitively accepted impact of melting and dyking on crustal strength. Models show, however, that the effect of melting is likely highly context-dependent: the mechanical properties of melts upon their cooling are not markedly different from those of the lower crust or mantle. In case of slow rifting melts cool as soon as they ascent to the surface and form rather strong flat layers in the crust. Hence, depending on the conditions, melting may lead to either weakening or to strengthening of crustal levels.

Although the asthenosphere will passively well up into the space created by initial extension this will also involve its decompressional partial melting. These partial melts will segregate from the asthenosphere and ascent into the lithosphere. Partial melting may increase with progressive extension, but only at relatively fast rates. Development of an actively rising asthenospheric diapir in response to lithospheric extension appears to occur at seafloor spreading axes but there is little evidence that this occurred during continental rifting, giving rise to the extrusion of MORB-source melts as seen in the highly volcanic East African rift (Ziegler, 1992). Note that the Norwegian–Greenland Sea rift was totally non-volcanic for about 275 My before the Iceland plume impinged on it. The North Atlantic rift was totally non-volcanic during its entire rifting history of 122My (Figs. 5 and 6). The timing of extension phases, the strain accomplished during them and the subsidence associated with them is generally derived from industry-type seismic lines.

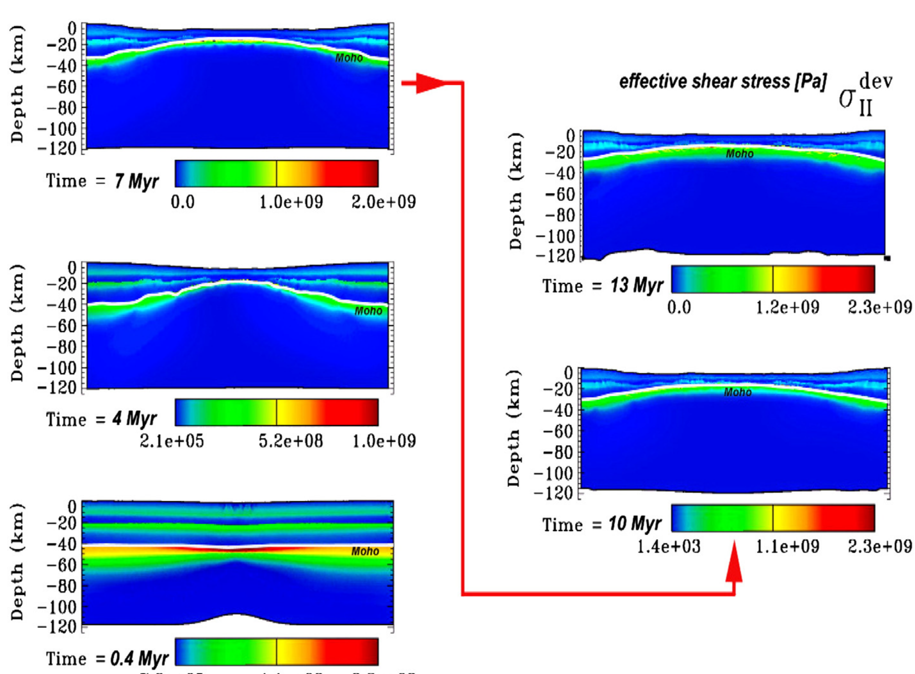

Fig. 23. Time evolution of the model from the experiments showed in Fig. 20. As can be seen, tensional stresses are first concentrated in the mantle part, and at this initial stage the Moho goes down. Once the mantle layer is thinned, the deformation is migrating to the upper layers, and the Moho goes up.

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6.7. Post-rift inversion, intraplate stresses, borderland uplift, and denudation

Mechanisms controlling the development of inversion structures remain enigmatic as theoretical models predict the inversion of passive margins in response to the buildup of compressional ridge-push forces only a few tens of million years after crustal separation has been achieved. Most of the shortening on the Mid-Norway margin was accommodated along preexisting major fault zones (Gabrielsen et al., 1999; Pascal and Gabrielsen, 2001). Compressional structures, such as long-wavelength arches and domes, strongly modified the architecture of the deep Cretaceous basins and controlled sedimentation patterns during Cenozoic (Bukovics and Ziegler, 1985; Vågnes et al., 1998).

From the Oligocene onward, the near-shore parts of the Mid-Norway margin were uplifted and deeply truncated (Doré and Jensen, 1996). Mechanisms controlling the observed broad uplift of the inner shelf and the adjacent on-shore areas, as evident along the Norwegian coast also remain enigmatic. However, the long-wavelength of the uplifted area is suggestive for mantle processes (Rohrman and Van der Beek, 1996; Rohrman et al., 2002). Olesen et al. (2002) interpreted the long-wavelength component of the gravity field in terms of both Moho topography and large-scale intrabasement density contrasts.

The seaward facing part of uplifted land areas was affected by strong glacial erosion, which in turn enhanced their uplift and increased sedimentation and subsidence rates in the flanking basins (Cloetingh et al., 2005b). Fission track geothermal chronology analyses along major on-shore lineaments of southern Norway show that preexisting major normal faults dating back to the Late Palaeozoic and Mesozoic, played a significant role in the Cenozoic uplift pattern (Cloetingh et al., 2005b; Hendriks and Andriessen, 2002; Redfield et al., 2005). Under the presently prevailing northwest-directed compressional stress field the Mid-Norwegian margin and its adjacent highlands are seismically active (Grünthal, 1999) with some faults showing evidence for recent movement (Mörner, 2004). Uplift of the South Norwegian highland continues, whilst the North Sea Basin experiences a phase of accelerated subsidence that began during the Pliocene and is attributed to stress-induced deflection of the lithosphere (Van Wees and Cloetingh, 1996).

Southwestern Norway was uplifted by as much as to 2 km during Neogene times (Rohrman et al., 1995), as evidenced by the progradation of clastic wedges into the North Sea Basin (Jordt et al., 1995). However, the uplift patterns and timing of southwestern Norway and the northern Scandes differ (Hendriks and Andriessen, 2002). By Late Tertiary times, cold climatic conditions prevailed.

Intraplate stresses can significantly influence the post-rift evolution of extensional basins, particularly when transmitted from orogens into adjacent back-arc basins, such as the Black Sea or the Pannonian Basin (e.g., Bada et al., 2007; Cloetingh et al., 2003; Grenerczy et al., 2005). Field studies, kinematic indicators and numerical modelling of present-day and palaeo-stress fields in selected areas (e.g., Bada et al., 1998, 2001; Gölke and Coblentz, 1996) yielded new constraints on the causes and the expression of intraplate stress fields in the lithosphere. Ziegler et al. (2002, 2006) discussed the role of mechanical controls on collision related compressional intraplate deformation. Temporal and spatial variations in the level and magnitude of these stresses have a strong impact on the record of vertical motions in sedimentary basins (Cloetingh and Kooi, 1992b; Cloetingh et al., 1990; Zoback et al., 1993). Stresses propagating from the margins of plates spreading axes into their interior had not only a strong effect on their stratigraphic record, but can also contribute to observed late-stage subsidence accelerations (Cloetingh and Burov, 2011; Cloetingh et al., 1999), similar to what is recognized in the Pannonian Basin and the North Sea Basin (Horváth and Cloetingh, 1996; Van Wees and Cloetingh, 1996). Over the last few years, increasing attention has been directed to this topic, advancing our understanding of the relationship between changes in plate motions, plate interaction, and the evolution of rifted basins (Doré et al., 1997; Janssen et al., 1995; Johnson et al., 2008) and foreland areas (Ziegler et al., 1998, 2001).

A continuous spectrum of stress-induced vertical motions can occur in the sedimentary record, varying from subtle faulting effects to basin inversion (Brun and Nalpas, 1996; Ter Voorde and Cloetingh, 1996; Ter Voorde et al., 2007; Ziegler et al., 1998) to the enhancement of flexural effects and to lithospheric folding induced by high levels of stress approaching lithospheric strengths (Bonnet et al., 1998; Burov et al., 1993; Cloetingh and Burov, 2011; Cloetingh et al., 1999; Stephenson and Cloetingh, 1991).

Numerical models have been developed for simulating the interplay of faulting and folding during intraplate compressional deformation (Beekman et al., 1996; Cloetingh et al., 1999; Gerbault et al., 1998). Models have also been developed to investigate the effects of faulting on stress-induced intraplate deformation in rifted margin settings (Van Balen et al., 1998).

In general, compressional intraplate tectonics appears to have a strong impact on the post-rift evolution of extensional basins (e.g. transpressional fault reactivation causing uplift of anticlinal structures, phases of accelerated basin-wide subsidence in response to stress-induced deflection of the lithosphere (see for a recent overview, Cloetingh et al., 2008).

7. Back-arc basin formation and evolution

Back-arc extensional basins evolve in response to a decrease in the convergence rate or even a temporary divergence of colliding plates, causing steepening and roll-back of the subducting lower plate lithospheric slab and the development of a secondary upwelling system in the upper plate mantle wedge above the subducting slab (e.g. Doglioni et al., 2007; Honza, 1993; Tamaki and Honza, 1991; Uyeda and McCabe, 1983). Back-arc basins can be developed over oceanic as well as continental lithosphere. Continental back-arc rifting can progress to the opening of limited oceanic basins (e.g. Sea of Japan, Black Sea, South China Sea). However, as convergence rates between colliding plates vary in time, back-arc extensional basins are generally short-lived. Upon a renewed increase in convergence rates and swallowing of the subducting lower plate lithospheric slab, back-arc extensional basins are subjected to compressional stresses that can lead to their inversion and closure. Typical examples may include the Variscan geosyncline Sunda arc and East China or the Black Sea (Uyeda and McCabe; Cloetingh et al., 1989; Hall et al., 2011; Jolivet et al., 1994; Munteanu et al., 2011; Nikishin et al., 2011; Ziegler, 1990). The sedimentary geometry and architecture of back-arc basins is controlled by such parameters as the age of subducted lithosphere, the subduction direction, the type of underlying lithosphere (oceanic versus continental) or the uplift of the orogenic arc (e.g., Dewey, 1981; Doglioni et al., 2007; Mathisen and Vondra, 1983; Uyeda and Kanamori, 1979). These parameters control the large variety of back-arc basins of various ages presently overlying different types of crust, such as the Caribbean, Banda-Sunda back-arcs or the Black Sea Basin (e.g., Hall et al., 2011; Meschede and Frisch, 1998; Munteanu et al., 2011; Spakman and Hall, 2010).

An observation that is common to many extensional back-arc basins floored by continental crust is the discrepancy between the observed amounts of crustal and lithospheric mantle stretching, as derived from the magnitude the upper crustal faulting and crustal/lithospheric thicknesses. This type of back-arc basin is often characterized by a much thicker post-rift basin fill than predicted by syn-rift stretching as seen in the Pannonian Basin (Horváth et al., 2006; Lenkey, 1999). Such contrasting features occur in the hinterland of subduction driven orogens (sensu Doglioni et al., 2007; Royden and Burchfiel, 1989), as for example the case of subduction driving the highly arcuate Mediterranean orogens (e.g., the Pannonian–Transylvanian system, the Aegean Sea or

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the Western Mediterranean, Faccenna et al., 2004; Jolivet and Brun, 2010; Krézsek and Bally, 2006; Matenco and Radivojević, 2012; van Hinsbergen and Schmid, 2012) or SE Asia subduction zones (e.g., the Malay and Pattani basins, Hutchison and Tan, 2009; Morley and Westaway, 2006).

In the following, two large scale extensional back-arc basins that underwent significant amounts of inversion post-dating extensional episodes are discussed, the Black Sea and the Pannonian–Transylvania Basin system of Central Europe.

7.1. Black Sea

The Black Sea is superimposed on the Euxinus orogenic system that developed along the southern margin of the East-European craton during the closure of the Rheic and Paleo-Thethys oceans. The Euxinus orogen was repeatedly subjected to back-arc extension and compression during the Triassic to Early Cretaceous final closure of the Paleo-Tethys and the opening of the Neo-Tethys oceans (Fig. 24, e.g., Nikishin et al., 2011; Okay et al., 1996; Robinson et al., 1996; Saintot et al., 2006; Sengor, 1987; Stephenson et al., 2004). The Western Black Sea opened in Late Cretaceous times (e.g., Finetti et al., 1988; Görür, 1988; Munteanu et al., 2011; Nikishin et al., 2001, 2011), and is commonly interpreted as an extensional back-arc basin related to the N-ward subduction of the Neotethys (e.g., Letouzey et al., 1977; Nikishin et al., 2011; Okay et al., 1994). Although a coeval opening of all domains of the Black Sea during Late Cretaceous times has been inferred (e.g., Nikishin et al., 2003; Zonenshain and Le Pichon, 1986), most studies agree that the Eastern Black Sea has opened later, during Paleocene–Eocene times (e.g., Banks, 1997; Kazmin et al., 2000; Robinson et al., 1996] by the rotation of the Shatsky Ridge away from the Mid Black Sea High (Fig. 24) (see also Okay et al., 1994], creating thinned continental to oceanic crust (Edwards et al., 2009; Shillington et al., 2008). The Early Cretaceous opening of the Western Black Sea took place in successive deformation phases (Munteanu et al., 2011) creating possibly oceanic crust, followed by overall sea-floor spreading and subsidence during Late Cretaceous and Cenozoic times (Finetti et al., 1988; Görür, 1988; Nikishin et al., 2011; Starostenko et al., 2004). In the Pontides the Early Cretaceous back-arc opening was followed by a main Late Cretaceous extensional episode (Görür, 1988; Yilmaz et al., 1997) that can be followed laterally also onshore in the Balkanides (e.g., Georgiev et al., 2001). The Eocene opening of the Eastern Black Sea (e.g., Görür, 1988; Meredith and Egan, 2002) has induced renewed extension in the western basin (e.g., Dinu et al., 2005; Munteanu et al., 2011; Robinson et al., 1996; Tari and Nemcok, 2009). The passive margin evolution is interrupted by the late (middle) Eocene collision, recorded by the major tectonic units of the Pontides and Taurides and smaller continental fragments accreted in between (Okay et al., 1994; Sengor and Yilmaz, 1981). The major contraction taking place at the southern margin of the Black Sea led to the onset of inversion recorded in the extensional basins and to the formation of other foreland and thrust-sheet top basins (e.g., Dinu et al., 2005; Finetti et al., 1988; Nikishin et al., 2003; Robinson et al., 1996; Sunal and Tuysuz, 2002). In the western Black Sea basin, the contraction gradually migrated in time N-wards until the Odessa shelf (Munteanu et al., 2011), where the shortening continued during Miocene–Pliocene times (Afanasenkov, 2007; Khriachtchevskaia et al., 2009). The eastern Black Sea was inverted with an opposite polarity, the Crimean and Caucasus orogens being thrusted S-wards over the Black Sea domain starting with Oligocene times (e.g., Stovba et al., 2009). The change in polarity is accommodated by the transcurrent movements recorded by the Odessa–West Crimean fault system and along the Mid-Black Sea High (Munteanu et al., 2011).

In the context of the endemic Eastern Parathethys, the Neogene evolution of the western Black Sea basin is of special interest (Rögl, 1999; Senes, 1973). While fine clastic to pelagic Lower–Middle Miocene sediments are generally thin throughout the Western Black Sea, higher subsidence rates are recorded during Latest Miocene–Quaternary times (Cloetingh et al., 2003; Munteanu et al., 2012; Nikishin et al., 2003). The sea level drop of the Messinian Salinity Crisis in the Mediterranean (e.g., Krijgsman et al., 1999) is recorded in the Black Sea by large scale shelf erosion and massive progradation of clastics during the lower Pliocene transgressive and high-stand, the estimated sea-level drop probably exceeding 1 km (e.g., Bartol et al., 2012; Gillet et al., 2007; Hsu and Giovanoli, 1979; Munteanu et al., 2012). Rapid sea level changes are also inferred for the Pliocene–Quaternary evolution of the Western Black Sea (Lericolais et al., 2013; Winguth et al., 2000), when sea level low stands triggered the transport of important volumes of sediments toward the deeper parts of the basin. Consequently, thick successions of mass-transport and turbidite deposits are observed along a number of deep-sea fans in front of modern rivers discharging into the Black Sea (e.g., Popescu and Radulian, 2001).

Below we address the relationship between the pre-rift finite strength of the lithosphere and the geometry of extensional basins in the Black Sea area and discuss the effects of differences in pre-rift rheology on the Mesozoic–Cenozoic stratigraphy of the western and

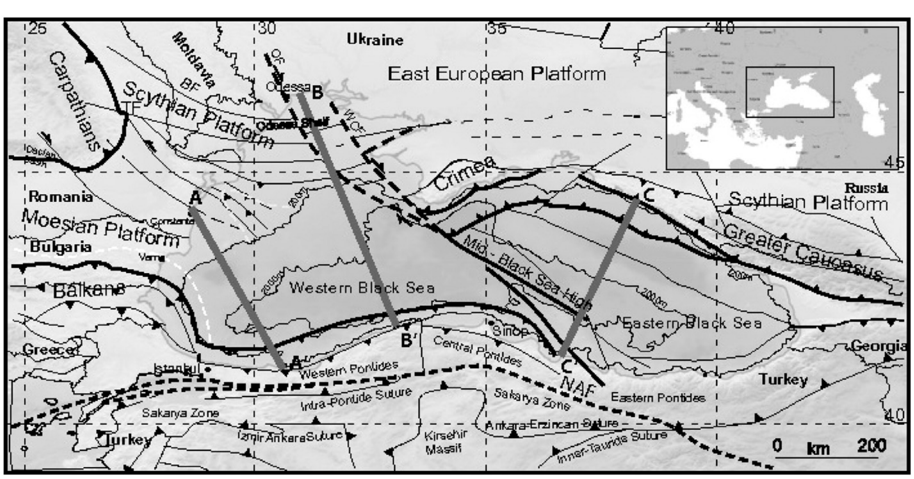

Fig. 24. Tectonic map of the Black Sea and adjacent areas (simplified and modified from Munteanu, 2012) with the location of profiles in Fig. 25. Continuous lines are faults separating tectonic units in and around the Black Sea. dashed lines are zones of diffuse and large scale deformation. NAF — North Anatolian Fault; WCF — West Crimea Fault; OF — Odessa Fault; BF — Bistrita Fault; TF — Trotus Fault;.

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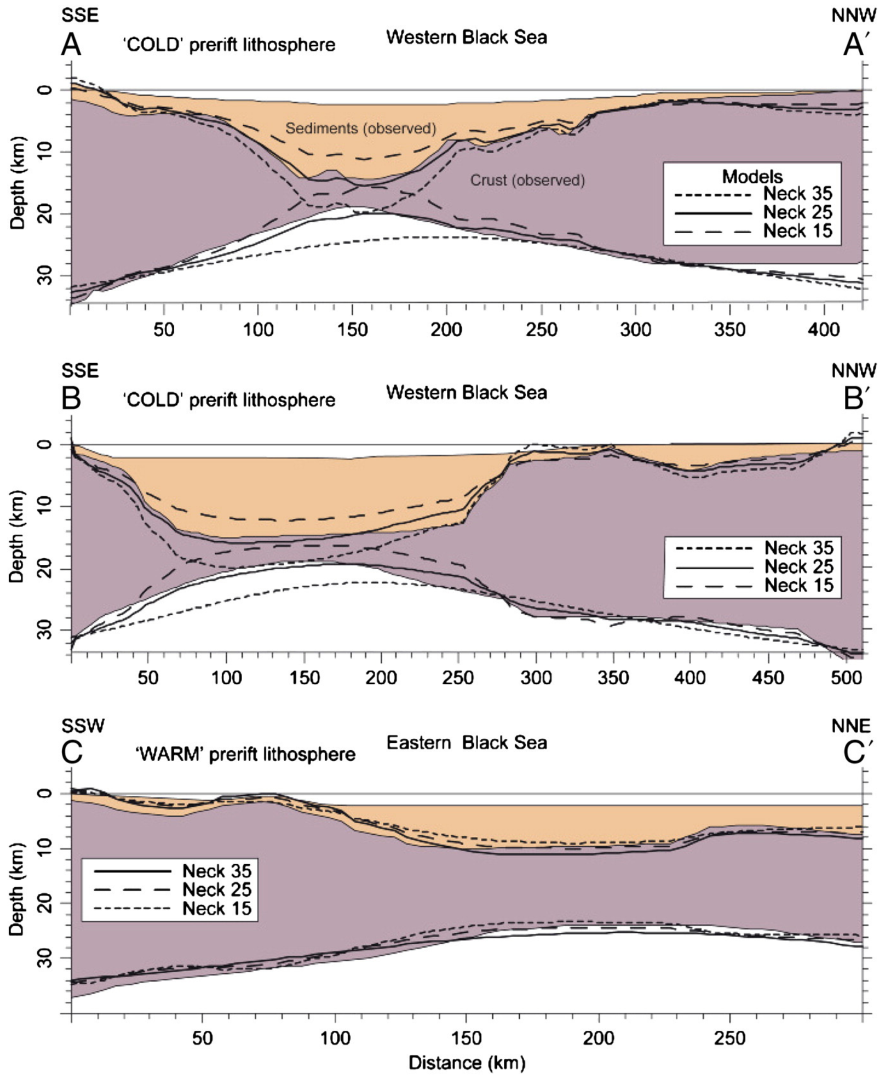

Fig. 25. Crustal scale models for basin evolution for the Western and Eastern Black Sea. For location of cross sections see Fig. 24. A comparison of predicted and observed Moho depths provides constraints on levels of necking and thermal regime of the pre-rift lithosphere. The models support the presence of cold pre-rift lithosphere compatible with a deep necking level of 25 km in the Western Black Sea. For the Eastern Black Sea, the models suggest the presence of a warm pre-rift lithosphere with a necking level of 15 km. Modified from Cloetingh et al. (2003).

eastern Black Sea Basin. These findings raise important questions on post-rift tectonics and on intraplate stress transmission into the Black Sea Basin from its margins.

7.1.1. Rheology and back-arc rift basin formation

Inferred differences in the mode of basin formation between the western and eastern Black Sea (Figs. 24–27), basins can be largely explained in terms of paleo-rheologies. The pre-rift lithospheric strength of the western Black Sea (Fig. 25) appears to be primarily controlled by the combined mechanical response of a strong upper crust and strong upper mantle (Spadini et al., 1996). The shallow necking level in the eastern Black Sea is compatible with a pre-rift strength controlled by a strong upper crust decoupled from the weak hot underlying mantle (Fig. 25). These contrasts point to important differences in the thermo-tectonic age of the lithosphere of the two sub-basins (Cloetingh and Burov, 1996). The inferred lateral variations between the western and eastern Black Sea suggest thermal stabilization of the western Black Sea prior to rifting whilst the lithosphere of the eastern Black Sea was apparently already thinned and thermally destabilized by the time of Eocene rifting. The inferred differences in necking level and in the timing of rifting between the western and eastern Black Sea suggest an earlier and more pronounced development of rift shoulders in the western Black Sea Basin as compared to the eastern Black Sea (Robinson et al., 1995).

7.1.2. Strength evolution and neotectonic reactivation at the basin margins during the post-rift phase

Automated backstripping analyses and comparison of results with forward models of lithospheric stretching (Van Wees et al., 1998) provide estimates of the integrated lithospheric strength at various syn- and post-rift stages.

Fig. 26 shows a comparison of observed and forward modelled tectonic subsidence curves for the centre of the Western Black Sea Basin. Automated backstripping yields a stretching factor of 6. Modelling, however, fails to predict the pronounced Late Neogene subsidence acceleration, documented by the stratigraphic record that may be attributed to late stage compression. Because post-rift cooling of the lithosphere leads in time to a significant increase in its integrated strength, its early post-rift deformation is favoured. Present-day lithospheric strength profiles calculated for the centre and margin of western Black Sea show a pronounced difference. On the contrary, the onset of the basin inversion during late Middle Eocene times at short times after the extension ceased earlier during Middle Eocene means that the lithosphere was weak when compression was activated. The

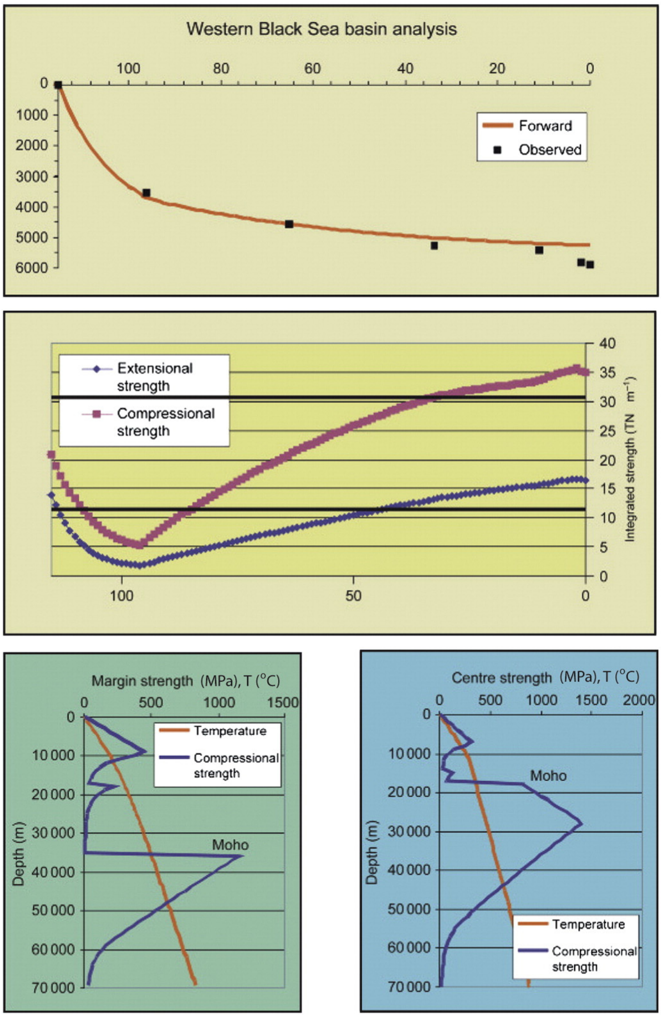

Fig. 26. Upper panel — Comparison of observed and forward modelled tectonic subsidence for the Western Black Sea centre. Automated backstripping yields an estimated stretching factor of 6 (top panel). Post-rift cooling leads in time to a significant increase in the predicted integrated strength for both compressional and extensional regimes (middle panel; 1 TN/m $= 10^{12}$ N/m); Middle panel — Present-day lithospheric compressional strength profiles calculated for the centre and margin of western Black Sea show with depth a pronounced difference (bottom panels). Temperature profiles (in C) and Moho depth are given for reference. Note the important role of the actual Moho position in terms of mechanical decoupling of the upper crust and mantle parts of the Black Sea lithosphere. Modified from Cloetingh et al. (2003). Horizontal lines indicate corresponding strength of the lithosphere at the time of the reversal of the stress field from the initial Cretaceous–Middle Eocene extension (lower horizontal line) to late Middle Eocene compression (upper horizontal line) in the Black Sea Basin. The latter marks the onset of the inversion in the basin.

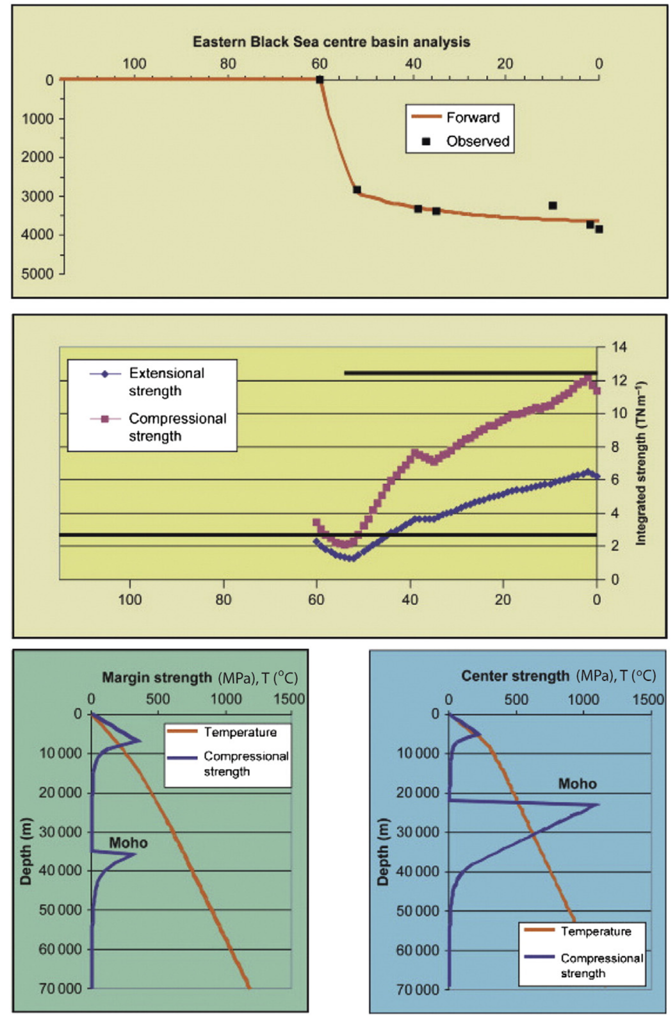

Fig. 27. Comparison of observed and forward modelled tectonic subsidence for the Eastern Black Sea centre. Automated backstripping yields an estimated stretching factor of 2.3 (upper panel). Post-rift cooling leads in time to a significant increase in the predicted integrated strength for both compressional and extensional regimes (middle panel 1 TN/m $= 10^{12}$ N/m). Present-day lithospheric strength profiles calculated for the centre and margin of the Eastern Black Sea show with depth a pronounced difference (bottom panels).
Modified from Cloetingh et al. (2003).

presence of relatively strong lithosphere in the basin centre in the very late stages of Pliocene inversion and weaker lithosphere at the basin margins favours deformation of the latter during late-stage compression.

In Fig. 27 the observed and forward modelled tectonic subsidence curves for the centre of eastern Black Sea Basin are compared, adopting for modelling a stretching factor of 2.3 that is compatible with the subsidence data and consistent with geophysical constraints. During the first 10 My of post-rift evolution, integrated strengths are low but subsequently increase rapidly owing to cooling of the lithosphere. This would correspond to the initial moments of post-Middle Eocene inversion. A short time after the ultimate cessation of rifting (~10 My) in Middle Eocene times, the basing was still underlain by very weak and extended lithosphere. This is the moment when the entire Black Sea basin was inverted due to the collision recorded in the Pontides domain (e.g., Munteanu et al., 2011). It should be noted, however, that with a cooling-related progressive increase of the integrated lithospheric strength, with time increasingly higher stress levels are required to cause large-scale deformation (Figs. 26 and 27). This is probably why the amplitude of contractional deformation decreases in time in the western Black Sea basin (Munteanu et al., 2012). Important in this context is also that, due to substantial crustal thinning, a relatively strong upper mantle layer is present in the central part of the basin at relatively shallow depths that may contrast substantially with an inherited weaker lower crust and, therefore, favour the transmission of contractional stresses over large distances (Munteanu, 2012).

Based on the present thermo-mechanical configuration (Figs. 26 and 27) with relatively strong lithosphere in the basin centre and relatively weak lithosphere at the basin margins, we predict that the bulk of shortening will be taken up in the centre of the basin during the initial phase of shortening, while the late-stage shortening induced by orogenic activity in the surrounding areas will be taken up along the basin margins, with only minor deformation occurring in the relatively stiff by now central parts of the basin. The relative difference in rheological strength of the marginal and central parts of the basin is more pronounced in the eastern than in the western Black Sea. These predictions are in agreement with data mapping deformation in the Black Sea (e.g., Dondurur et al., 2013; Munteanu et al., 2012).

Basement and surface heat flow in the eastern and western Black Sea show markedly different patterns in timing of the rift-related heat flow maximum. The predicted present-day heat flow is considerably lower in the western than in the eastern sub-basin. This is attributed to the presence of more heat producing crustal material in the eastern than in the western Black Sea that is partly floored by oceanic crust. In heat flow modelling studies the effects of sedimentary blanketing were taken into account (Van Wees and Beekman, 2000). Heat flow values vary between 30 mW m^{-2} in the centre of the basins up to 70 mW m^{-2} at the Crimean and Caucasus margins (Nikishin et al., 2003). With its sedimentary thickness of up to 15 km, thermal blanketing has a pronounced effect on the heat flow in the western Black Sea. As a result, its present-day integrated strength is not that much higher than its initial strength. By contrast, the integrated strength of the eastern Black Sea is much higher than its initial strength as the blanketing effect of its up to 12 km thick sedimentary fill is less pronounced and as water depths are greater.

Fig. 12 show a comparison of theoretical predictions for lithosphere folding of rheologically coupled and decoupled lithosphere, as a function of its thermo-mechanical age with estimates of folding wavelengths documented in continental lithosphere for various representative areas of the globe (see Cloetingh et al., 1999). The western Black Sea centre is marked by a thermo-mechanical age of around 100 My with rheological modelling indicating mechanical decoupling of the crust and lithospheric mantle (see Figs. 26 and 27). These models imply an EET of at least 40 km (Burov and Diament, 1995) and folding wavelengths of 100–200 km for the mantle and 50–100 km for the upper crust (Cloetingh et al., 1999). For the eastern Black Sea, a probably significantly younger thermo-mechanical lithospheric age of 55 My implies an EET of no more than 25 km, and indicates mantle folding at wavelengths of 100–150 km and a crustal folding wavelength similar to the western Black Sea.

The large dimension of the pre-existing rift basin causes during the late-stages of inversion a pronounced increase in the wavelength of the stress-induced down-warp (Cloetingh et al., 1999; Munteanu et al., 2011). This effect has been observed in the North Sea Basin (Van Wees and Cloetingh, 1996) and the Pannonian Basin (Horváth and Cloetingh, 1996), both of which are characterized by large sediment loads and a wide rift basin. This provides an alternative to previous explanations for recent differential motions in the northern Black Sea Basin (Smolyaninova et al., 1996) that were attributed to convective mantle flow.

7.2. Modes of basin (de)formation, lithospheric strength, and vertical motions in continental back-arc rifts

The Pannonian–Transylvanian Basin system is surrounded by the Carpathian and Dinaridic orogens (Fig. 28) and has been the focus of major research efforts integrating a wide range of geophysical and geological data thus representing a key area for quantitative basin studies and the study of the dynamics of basin - orogens interaction (see Cloetingh et al., 2006; Horváth et al., 2006; Schmid et al., 2008 for recent reviews). A vast geophysical and geological database has been established during the last decades as a result of major international research collaborations in this area, largely carried out in the framework of European programs such as the EU Integrated Basin Studies project (Cloetingh et al., 1995; Durand et al., 1999), the EUROPROBE- PANCARDI (Decker et al., 1998) programs, the Peri-Tethys program (Brunet and Cloetingh, 2003; Ziegler and Horvath, 1996), and more recently the ESF EUROCORES TOPO-EUROPE initiative and its dedicated sub-projects, such as the SourceSink and Topo-Alps projects (Cloetingh et al., 2007; Matenco and Andriessen, 2013).

An important asset of this natural laboratory is the existence of high-quality constraints on both the evolution of Neogene basins such as the Pannonian or Transylvanian basins, but also on the mechanics of surrounding orogens (see Cloetingh et al., 2006; Horváth et al., 2006; Matenco et al., 2010; Royden, 1988). A focal area has been the Vrancea seismogenic area of the SE Carpathians that displays a high-velocity teleseismic intermediate-mantle anomaly commonly related to the rapid roll-back evolution of a slab inherited from the evolution of the Alpine Tethys (e.g., Cloetingh et al., 2004; Ismail-Zadeh et al., 2012; Martin and Wenzel, 2006 and references therein). This is associated with significant recent tectonic activity and associated natural hazards (e.g., Cloetingh et al., 2005a; Matenco et al., 2007; Merten et al., 2010; Necea et al., 2005).

In the Pannonian–Transylvanian Basin system, extensive industrial reflection-seismic coverage and well data acquired in the context of petroleum exploration and surface studies permitted to construct a high-resolution kinematic, structural, sedimentological and stratigraphic framework (e.g., Bada et al., 2007; Horváth et al., 2006; Krézsek and Bally, 2006, 2010; Matenco and Radivojević, 2012; Sacchi et al., 1999; Sztano et al., 2013; Tari et al., 1999; Vasiliev et al., 2010) that set the

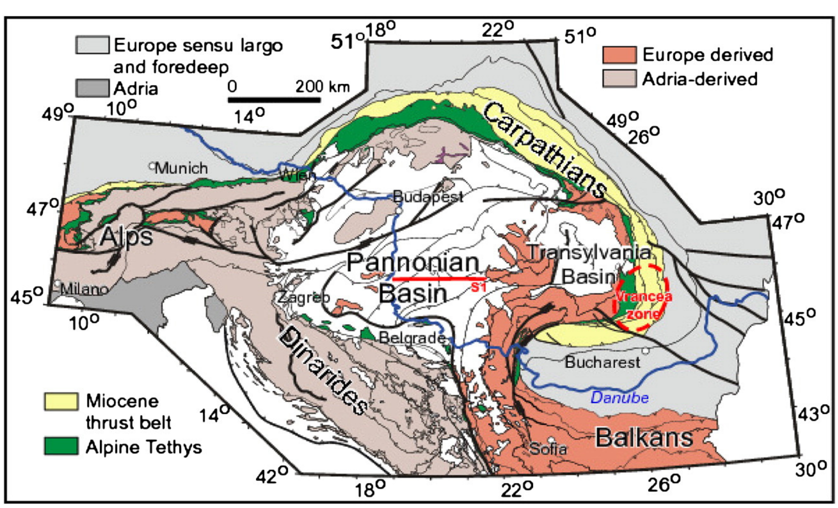

Fig. 28. Tectonic map of the Alps–Carpathians–Dinaridic system (modified after Matenco and Radivojević, 2012; Schmid et al., 2008) with the extent of the large Pannonian and Transylvanian back-arc basins and the location of other Miocene basins superposed over the Dinarides and Carpathians structures. 1 is the location of the cross section presented in Fig. 31. Red dashed oval is the location of the epicentres of the Vrancea seismogenic zone in the SE Carpathians.

stage for numerical and analogue modelling (e.g., Bada et al., 1999; Dombrádi et al., 2010; Jarosinski et al., 2011).

The Pannonian–Carpathian system, therefore, permits to test models for the development of sedimentary basins and their subsequent deformation, and for on-going continental collision. The lithosphere of the Pannonian Basin is a particularly sensitive recorder of changes in lithospheric stress induced by near-field intraplate and far-field plate boundary processes (Bada et al., 2001).

7.2.1. Neogene development and evolution of the Pannonian Basin

The Pannonian Basin of Central Europe (Fig. 28) is a classical extensional back-arc basin formed during Miocene times in response to the rapid roll-back of a slab attached to the European continent, which is still underlain by highly thinned continental lithosphere (e.g., Balla, 1986; Cloetingh et al., 2006; Horváth, 1993; Horváth et al., 2006; Royden, 1988). The roll-back is partly responsible for the creation of the highly arcuate geometry of the Carpathian Mountains (e.g., Matenco et al., 2010), a process that is common with many other Mediterranean orogens (e.g., Faccenna et al., 2004; Jolivet and Faccenna, 2000). The Pannonian basin is spatially juxtaposed over an orogenic area that formed in Cretaceous–Paleogene times in response to the subduction and closure of a number of oceanic domains which were kinematically linked with the evolution of the larger Alpine Tethys and Neotethys oceans (e.g., Csontos and Vörös, 2004; Schmid et al., 2008). Evolutionary models of the Pannonian Basin assume the onset of extension at ~20 Ma, subsequently followed by a peak tectonic activity along normal faults during Middle Miocene times, which was subsequently followed by a post-rift, thermal sag phase starting in late Miocene times (e.g., Tari et al., 1999 and references therein). A latest Miocene–Quaternary contractional event has subsequently overprinted the basin during the translation and counter-clockwise rotation of the Adriatic indenter (Bada et al., 2007; Fodor et al., 2005; Horváth, 1995; Horváth and Cloetingh, 1996; Pinter et al., 2005).

7.2.2. Dynamic models of basin formation

Several models have been proposed to explain the dynamics of Neogene rifting in the Pannonian Basin. An active versus passive mode of rifting has been a matter of continued debate (see Bada and Horváth, 2001) resulting in the proposal of various dynamic models that take into account such prominent features of the Pannonian–Carpathian system as thinned and hot versus thickened and colder lithosphere in its central and peripheral sectors, respectively. A number of models have proposed the presence of a mantle diapir beneath the intra-Carpathians area (e.g., Stegena, 1967). Other models proposed rifting as the driving mechanism for the Pannonian Basin (Horváth et al., 1975; Stegena et al., 1975), attributed to thermal thinning of the Pannonian lithosphere in response to upwelling of the mantle above the subducted European and Adriatic lithospheric slabs that dip beneath the Pannonian domain.

At a different scale and with a different mechanical behaviour, the plume model has been also recently inferred by Burov and Cloetingh (2010) to be responsible for the formation of the asthenospheric upwelling beneath the Pannonian basin and the initiation of continental lithosphere subduction in neighbouring Carpathians and Dinarides orogens.

Other models stress the back-arc position of the Pannonian domain with respect to the Carpathian arc and postulate that gravity-driven passive roll-back of the subducting European lithospheric slab is the driving force for the tensional subsidence of the Pannonian Basin (e.g., Csontos, 1995; Csontos et al., 1992; Horváth, 1993; Linzer, 1996; Royden and Karner, 1984; Royden and Horváth, 1988). In a modification of this model, eastward mantle flow is thought to control roll-back of the subducted slab and related retreat of its hinge line.

Based on thermo-mechanical finite element modelling, Huismans et al. (2001b) were able to simulate temporal changes in rifting style in the Pannonian Basin, suggesting a two-phase evolutionary scheme for the system. According to their model, the initial early Middle Miocene basin subsidence was driven mainly by passive rifting in response to roll-back of the subducted Carpathian slab, involving gravitational collapse of the thickened pre-rift Pannonian lithosphere. This triggered small-scale convective upwelling of the asthenosphere that has favoured the late-stage thermal subsidence (Horváth, 1993) as well as the compressional inversion of the Pannonian domain during Late Miocene–Pliocene times (e.g., Bada et al., 1999; Fodor et al., 2005).

One other category of models involves lithospheric mantle instabilities for the formation of the asthenospheric upraise beneath the Pannonian basin and the high-velocity anomalies observed by teleseismic tomography in neighbouring orogens (e.g., Dando et al., 2011; Gemmer and Houseman, 2007; Houseman and Gemmer, 2007; Ren et al., 2012). These models assume a convective mantle removal beneath the Pannonian basin, correlated with high heat flow and interpreted as upwelling asthenosphere correlated with a fast anomaly extending outwards as far as the Carpathians, the Dinarides and the Eastern Alps. This higher velocity mantle material is interpreted as being produced by a mantle downwelling, whose detachment from the lithosphere above may have triggered the extension of the Pannonian Basin.

7.2.3. Stretching models and subsidence analysis

Sclater et al. (1980) were the first to apply the stretching model of McKenzie (1978) to the intra- Carpathian basins. They found that the development of peripheral basins could be fairly well simulated by the uniform pure shear extension concept with a stretching factor of about 2. For the more centrally located basins, however, their considerable thermal subsidence and high heat flow suggested unrealistically high stretching factors up to 5. Thus, they postulated differential extension of the Pannonian lithosphere with moderate crustal stretching (δ factor) and larger stretching of the lithospheric mantle (β factor). Building on this and using a wealth of well data, Royden et al. (1983) introduced the non-uniform stretching concept according to which the magnitude of lithospheric thinning is depth-dependent. This concept accounts for a combination of uniform mechanical extension of the lithosphere and thermal attenuation of the lithospheric mantle (Ziegler, 1992, 1996b; Ziegler and Cloetingh, 2004). This is compatible with the subsidence pattern and thermal history of major parts of the Pannonian Basin that suggest a greater attenuation of the lithospheric mantle as compared to the finite extension of the crust. Horváth et al. (1988) further improved this concept by considering radioactive heat generation in the crust, and the thermal blanketing effect of basin-scale sedimentation. By reconstructing the subsidence and thermal history, and by calculating the thermal maturation of organic matter in the central region of the Pannonian Basin (Great Hungarian Plain), a major step forward was made in the field of hydrocarbon prospecting by means of basin analysis techniques.

These studies highlighted difficulties met in explaining basin subsidence and crustal thinning in terms of uniform extension, and point toward the applicability of anomalous sub-crustal mantle thinning. This issue was central to subsequent investigations involving quantitative subsidence analyses (backstripping) of an extended set of Pannonian Basin wells and cross sections and their forward modelling (Juhász et al., 1999; Lankreijer et al., 1995; Lenkey, 1999; Sachsenhofer et al., 1997). Kinematic modelling, incorporating the concept of necking depth and finite strength of the lithosphere during and after rifting (Van Balen et al., 1999), as well as dynamic modelling studies (Huismans et al., 2001b), suggested that the transition from passive to active rifting was controlled by the onset of sub-crustal flow and small-scale convection in the asthenosphere. In order to quantify basin-scale lithospheric deformation, Lenkey (1999) carried out forward modelling applying the concept of non-uniform lithospheric stretching and taking into account the effects of lateral heat flow,

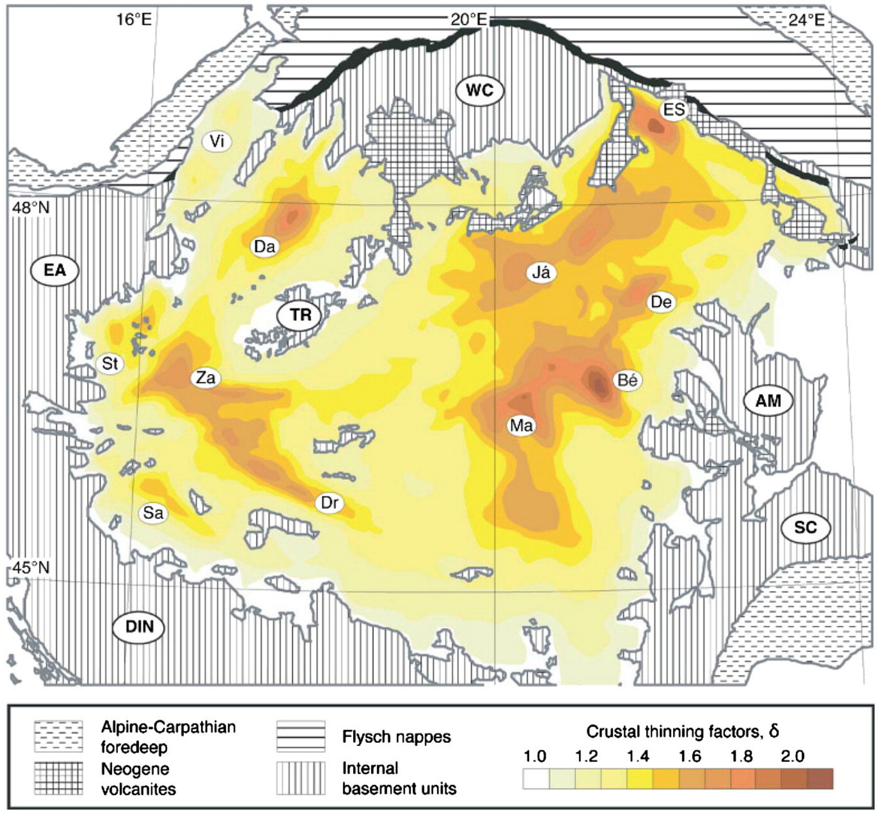

Fig. 29. Crustal thinning factors calculated by forward modelling for the Pannonian Basin applying the non-uniform stretching concept and taking the effects of lateral heat flow and flexure of the lithosphere into account (after Lenkey, 1999). Note the pronounced lateral variations in crustal extension controlling the development of deep sub-basins separated by areas of limited deformation. AM: Apuseni Mts.; DIN: Dinarides; EA: Eastern Alps; TR: Transdanubian Range; SC, WC: Southern and Western Carpathians, respectively. Local depressions of the Pannonian Basin system: Be: Bekes; Da: Danube; De: Derecske; Dr: Drava; ES: East Slovakian; Ja: Jaszsag; Ma: Mako; Sa: Sava; St: Styrian; Vi: Vienna; Za: Zala. Modified from Lenkey (1999).

flexure, and necking of the lithosphere. Calculated crustal thinning factors (δ) indicate large lateral variation of crustal extension in the Pannonian Basin (Fig. 29). This is consistent with the areal pattern of the depth to the pre-Neogene basement (Horváth et al., 2006). The indicated range of crustal thinning factors of 10–100% crustal extension is in good agreement with the pre-rift palinspastic reconstruction of the Pannonian Basin, and the amount of cumulative shortening in the Carpathian orogen (e.g., Fodor et al., 1999; Roure et al., 1993).

As a major outcome of basin analysis studies, Royden et al. (1983) provided a two-stage subdivision for the evolution of the Pannonian Basin with a syn-rift (tectonic) phase during Early to Middle Miocene times, and a post-rift thermal subsidence phase during the Late Miocene–Pliocene. Further development of the stratigraphic database, however, demonstrated the need to refine this scenario. According to Tari et al. (1999), the regional Middle Badenian unconformity, marking the termination of the syn-rift stage, is followed by a post-rift phase that is characterized by only minor tectonic activity. Nevertheless, the subsidence history of the Pannonian Basin can be subdivided into three main phases that are reflected in the subsidence curves of selected sub-basins (Fig. 30). The initial syn-rift phase is characterized by rapid tectonic subsidence, starting synchronously at about 20 Ma in the entire Pannonian Basin. This phase of pronounced crustal extension is recorded everywhere in the basin system and was mostly limited to relatively narrow, fault bounded grabens or sub-basins. During the subsequent post-rift phase much broader areas began to subside, reflecting general down warping of the lithosphere in response to its thermal subsidence. This is particularly evident in the central parts of the Pannonian Basin, suggesting that in this area syn-rift thermal attenuation of the lithospheric mantle played a greater role than in the marginal areas (e.g., Royden and Dovenyi, 1988; Sclater et al., 1980). The third and final phase of basin evolution is characterized by the gradual structural inversion of the Pannonian Basin system during the Late Pliocene–Quaternary. During these times intraplate compressional stresses gradually built up and caused basin-scale buckling of the Pannonian lithosphere that was associated with late- stage subsidence anomalies and differential vertical motions (Horváth and Cloetingh, 1996). As evident from subsidence curves (Fig. 30), accelerated late- stage subsidence characterized the central depressions of the Little and Great Hungarian Plains (Fig. 30b-c), whereas the peripheral Styrian and East Slovakian sub-basins were uplifted by a few hundred meters after mid-Miocene times (Fig. 30d–e) and the Zala Basin during the Pliocene–Quaternary (Fig. 30f). The importance of tectonic stresses, both during the rifting (extension) and subsequent inversion phase (compression), is highlighted by this late-stage tectonic reactivation, as well as by other episodic inversion events in the Pannonian Basin (Fodor et al., 1999; Horváth, 1995).

A more recent analysis of the subsidence evolution of the Pannonian basin in the sector connecting the Carpathians with the Dinarides has observed significant discrepancies with the above mentioned models by analysing the development of syn- and post-rift sedimentation (Matenco and Radivojević, 2012). This study has documented large-scale erosional unconformities between the classically defined syn- and post-rift phases that migrate in time from the Dinarides to the Carpathians, while the post-rift deposition for the Early and Middle Miocene rifting is largely missing (Fig. 31). Therefore, these authors propose that the low-angle Early–Middle Miocene normal faults responsible for the basin formation are connected to a lower crustal detachment that was prolonged eastwards beneath the Apuseni Mountains and affected the mantle lithosphere below the Transylvanian Basin. The coupling between the mechanical

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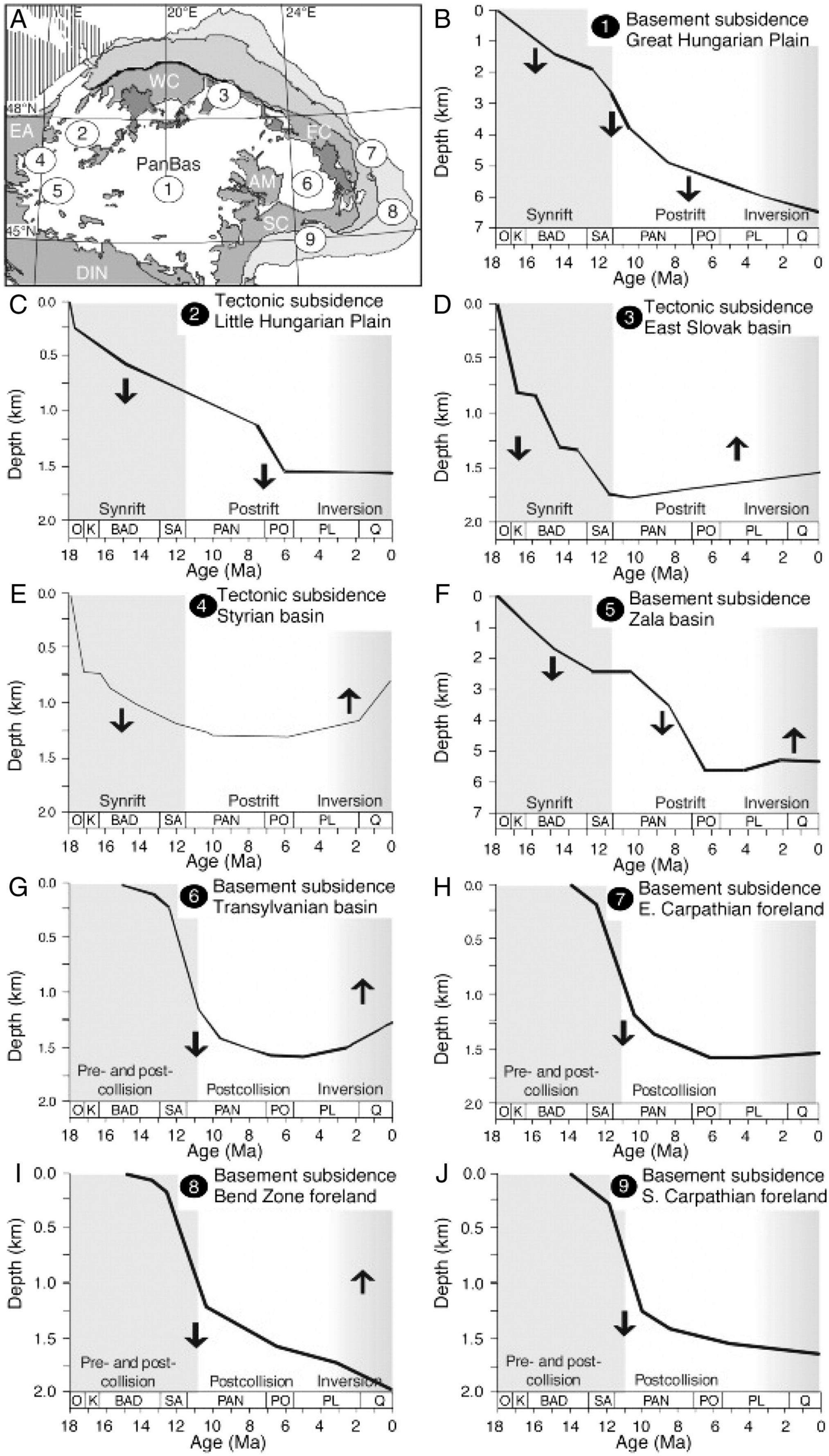

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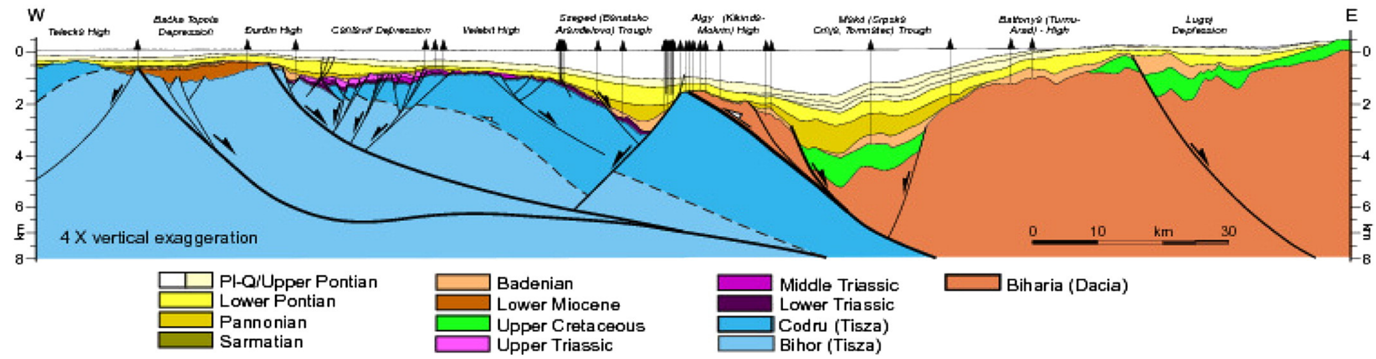

Fig. 31. Depth-converted regional transects over the Pannonian Basin of Serbia and Romania. After Matenco and Radivojević (2012).

extension of the Pannonian Basin and its thermal effects recorded in the Transylvanian Basin should have taken place in such a way that the intervening Apuseni Mountains did not underwent any significant Miocene vertical movements, whose exhumation and large scale dome structure is mainly the effect of Cretaceous–Paleogene deformations (Merten et al., 2011). In contrast to earlier periods of extension, massive thermal sag sedimentation is observed in direct relationship with and post-dating the Pannonian–Lower Pontian detachments and/or normal faults (such as the Makó depression, Fig. 31). The main depocenters of the thermal sag sedimentation overlie these asymmetrical extensional structures. Furthermore, this thermal sag sedimentation has a far larger areal extent than the one of the half-grabens, covering an area that corresponds to the asthenospheric upraise presently observed beneath the Pannonian Basin (Horváth et al., 2006).

This analysis demonstrated the close relationship that existed between the timing and nature of stress changes in extensional basins and structural episodes in the surrounding thrust belts, pointing to mechanical coupling between the orogen and its back-arc basin. In parallel, significant efforts have been devoted to reconstruct the spatial and temporal variations in thrusting along the Carpathian orogen (Matenco and Bertotti, 2000; Matenco et al., 2007, 2010; Roure et al., 1993; Schmid et al., 1998) and its relationship to foredeep depocenters (Cloetingh et al., 2005a; Matenco et al., 2003; Meulenkamp et al., 1996; Necea et al., 2005; Tarapoanca, 2004), changes in foreland basin geometry and lateral variations in the flexural rigidity of the foreland lithosphere.

7.2.4. Lithospheric strength in the Pannonian–Carpathian system

The Pannonian–Carpathian system shows remarkable variations in crustal thickness (Fig. 22) and thermo-mechanical properties of the lithosphere. Lithospheric rigidity varies in space and time, giving rise to important differences in the tectonic behaviour of different parts of the system. As rheology controls the response of the lithosphere to stresses, and thus the formation and deformation of basins and orogens, the characterization of rheological properties and their temporal changes has been a major challenge to constrain and quantify tectonic models and scenarios. This is particularly valid for the Pannonian– Carpathian region where tectonic units with a different history and rheological properties are in close contact.

By conversion of strength predictions to EET values at a regional scale, Lankreijer (1998) mapped the EET distribution for the entire Pannonian– Carpathian system (Fig. 32). Calculated EET values are largely consistent with the spatial variation of lithospheric strength of the system. Lower values are characteristic for the weak central part of the Pannonian Basin (5–10 km), whereas EET increases toward the Dinarides and Alps (15–30 km) and particularly toward the Bohemian Massif and Moesian Platform (25–40 km). This trend is in good agreement with EET estimates obtained from flexural studies and forward modelling of extensional basin formation. Systematic differences, however, can occur and may be the consequence of significant horizontal intraplate stresses (e.g., Cloetingh and Burov, 1996) or of mechanical decoupling of the upper crust and uppermost mantle that can lead to a considerable reduction of EET values.

The range of calculated EET values reflects the distinct mechanical characteristics and response of the different domains forming part of the Pannonian–Carpathian system to the present-day stress field. These characteristics can be mainly attributed to the memory of the lithosphere. In this respect it must be kept in mind that the tectonic and thermal evolution of these domains differed considerably during the Cretaceous–Neogene Alpine development of both the outer and intra-Carpathians units and the Neogene extension of the Pannonian Basin, resulting in a wide spectrum of lithospheric strengths. These, in turn, exert a strong control on the complex present-day pattern of on-going tectonic activity.

7.2.5. Deformation of the Pannonian–Carpathian system

The present-day deformation pattern and related topography development in the Pannonian–Carpathian system is characterized by pronounced spatial and temporal variations in the stress and strain fields (Fig. 33) (see, e.g., Cloetingh et al., 2006). Horváth and Cloetingh (1996) established the importance of Late Pliocene and Quaternary compressional deformation of the Pannonian Basin that explains its anomalous uplift and subsidence, as well as intraplate seismicity. Based on the case study of the Pannonian–Carpathian system, these authors established a novel conceptual model for the structural reactivation of back-arc basins within orogens. At present, the Pannonian Basin has reached an advanced evolutionary stage as compared to other Mediterranean back-arc basins (e.g., Bache et al., 2010) in so far as it has been partially inverted during the last few million years. Inversion of the Pannonian Basin can be related to temporal changes in the regional stress field, from one of tension that controlled its Miocene extensional subsidence, to one of Pliocene–Quaternary compression resulting in deformation, contraction, and flexure of the lithosphere associated with differential vertical motions.

Model calculations are in good agreement with the overall topography of the system (Fig. 33). Several flat-lying, low-elevation areas

Fig. 30. Subsidence curves for selected sub-basins of the Pannonian Basin and the Carpathian foreland. Note that after a rapid phase of general subsidence throughout the entire Pannonian Basin, the sub-basins show distinct subsidence histories from Middle Miocene times onward. Arrows indicate generalized vertical movements. O, K: Ottnangian and Karpatian, respectively (Early Miocene); BAD, SA: Badenian and Sarmatian, respectively (Middle Miocene); PAN, PO: Pannonian and Pontian, respectively (Late Miocene); PL: Pliocene; Q: Quaternary. AM: Apuseni Mts.; DIN: Dinarides; EA: Eastern Alps; PanBas: Pannonian Basin; SC, WC: Southern and Western Carpathians foreland basins, respectively. Modified from Cloetingh et al. (2006).

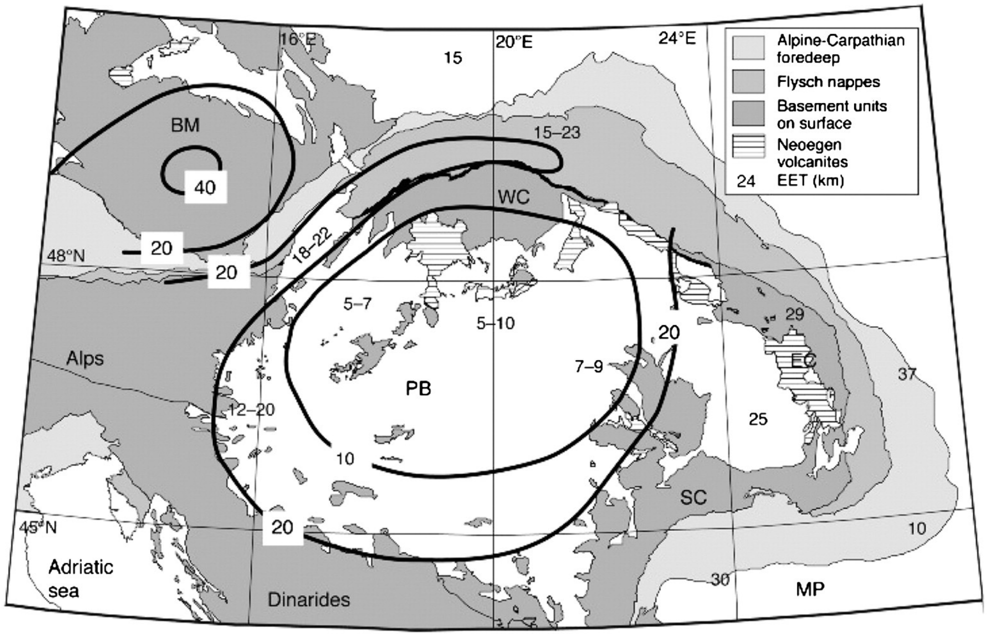

Fig. 32. Effective elastic thickness (EET — in km) of the lithosphere in and around the Pannonian Basin predicted from rheological calculations. BM: Bohemian Massif; MP: Moesian Platform; PB: Pannonian basin; EC, SC, WC: Eastern, Southern and Western Carpathians, respectively.
Modified from Lankreijer (1998).

(e.g., Great Hungarian Plain, Sava and Drava troughs) subsided continuously since the Early Miocene beginning of basin development and contain 300–1000 m thick Quaternary alluvial sequences. By contrast, the periphery of this basin system, as well as the Transdanubian Range, the Transylvanian Basin and the adjacent Carpathians underwent late-stage uplift.

Therefore, the spatial distribution of uplifting and subsiding areas within the Pannonian Basin can be interpreted as resulting from the build-up of intraplate compressional stresses, causing large-scale positive and negative deflection of the lithosphere at various scales. This includes basin-scale positive reactivation of Miocene normal faults, and large-scale folding of the system leading to differential uplift

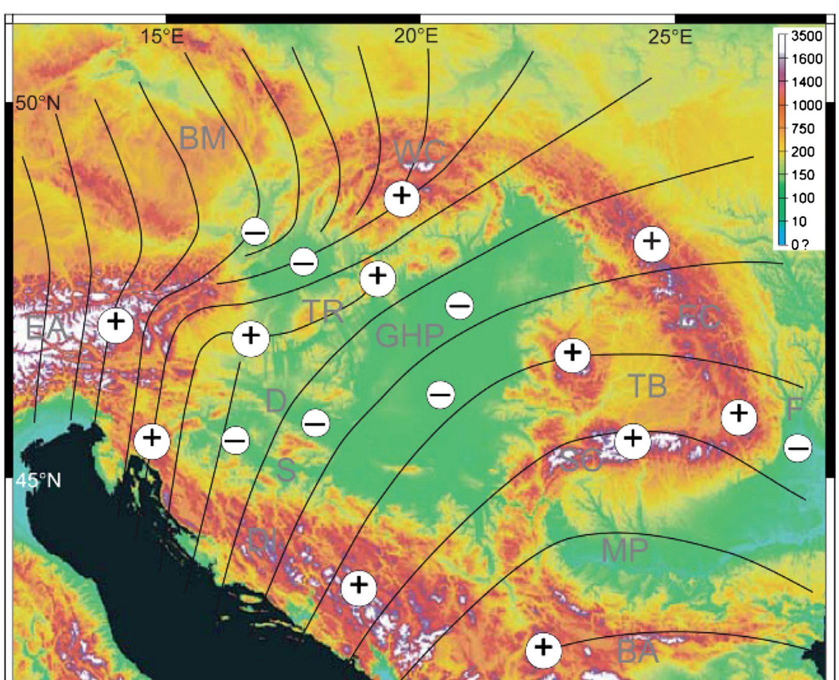

Fig. 33. Topography of the Pannonian–Carpathian system showing present-day maximum horizontal stress (SHmax) trajectories (after Bada et al., 2001). '+' and '-' symbols denote areas of Quaternary uplift and subsidence, respectively. BA: Balkanides; BM: Bohemian Massif; D: Drava Trough; DI: Dinarides; EA: Eastern Alps; EC: Eastern Carpathians; F: Foc ani Depression; MP: Moesian Platform; PB: Pannonian Basin; S: Sava Trough; SC: Southern Carpathians; TB: Transylvanian Basin; TR: Transdanubian Range; WC: Western Carpathians.
Modified from Cloetingh et al. (2006).

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and subsidence of anticlinal and synclinal segments of the orogen were uplifted and considerably eroded from Late Miocene–Pliocene times onward (see Fig. 33). Quantitative subsidence analyses confirm that late-stage compressional stresses caused accelerated subsidence of the central parts of the Pannonian Basin (Van Balen et al. (1999) whilst the Styrian Basin (Sachsenhofer et al., 1997), the Vienna and East Slovak Basins (Lankreijer et al., 1995), and the Transylvanian Basin (Ciulavu et al., 2002) were uplifted by several hundred meters starting in Late Mio–Pliocene times (Fig. 33). The mode and degree of coupling of the Carpathians and Dinarides with their foreland controls the Pliocene to Quaternary deformation patterns in their hinterland in the Pannonian - Transylvanian basins (Ciulavu et al., 2002; Dombrádi et al., 2007; Jarosinski et al., 2011). These findings can be related to the results of seismic tomography that highlight upwelling of hot mantle material under the Pannonian Basin and progressive detachment of the subducted lithospheric slab that is still ongoing in the Vrancea area (Wenzel et al., 2002; Wortel and Spakman, 2000).

In summary, results of forward basin modelling show that an increase in the level of compressional tectonic stress during Pliocene–Quaternary times can explain the first-order features of the observed pattern of accelerated subsidence in the centre of the Pannonian Basin and uplift of basins in peripheral areas. Therefore, both observations and modelling results lead to the conclusion that compressional stresses can cause considerable differential vertical motions in the Pannonian–Carpathian back-arc basin–orogen system (Dombrádi et al., 2007; Horváth et al., 2006; Jarosinski et al., 2011).

8. Conclusions

Rifts and extensional basin systems are sensitive recorders of dynamic processes controlling the deformation of the lithosphere and its interaction with the deep mantle and surface processes. The thermo-mechanical structure of the lithosphere exerts a prime control on its response to basin-forming mechanisms, such as rifting and its deflection under vertical loads, as well as its compressional deformation. Poly-phase deformation is a common feature of some of Europe's best-documented rifts and extensional sedimentary basin systems. The relatively weak lithosphere of intraplate Europe renders many of its rifted basins prone to reactivation. Tectonic processes operating during basin formation and during the subsequent deformation of basins can generate significant differential topography in sedimentary basin systems. Numerical modelling provides a novel approach to assess the feedback mechanisms between deep mantle, lithospheric, and surface processes. Compressional reactivation of extensional basins during their post-rift phase appears to be a common feature of Europe's intraplate rifts and passive margins, reflecting temporal and spatial changes in the orientation and magnitude of the intraplate stress regime. In the intraplate domain of continental Europe, thermally perturbed by Cenozoic upper mantle plume activity, lithospheric folding plays an important role and strongly affects the pattern of vertical motions, both in terms of basin subsidence and the uplift of broad arches.

Rift shoulder topography is directly linked to the thermo-mechanical properties and rheological stratification of the underlying lithosphere. The temporal evolution of the strength of continents and the spatial variations in stress and strength at continental margins and rifts govern the mechanics of basin development and destruction in time and space. Structural discontinuities and pre-existing weakness zones are prone to reactivation in response to the build-up of extensional stresses and thus play an important role in the localization of rifts. Their reactivation in response to the build-up of compressional stresses propagating from plate boundaries into continental platform areas controls the inversion of extensional basins and the thrusting of basement blocks and contributes to the localization of lithospheric folding. Systematic differences occur in the timing of the transition from rifting to intraplate compression and the development of contractional belts and lithospheric folds.

Recent deformation, involving tectonic reactivation, has strongly affected the structure and fill of many sedimentary basins. The long-lasting memory of the lithosphere appears to play a far more important role in basin reactivation than hitherto assumed. A better understanding of the 3D linkage between basin formation and basin deformation is, therefore, an essential step in research aiming at linking lithospheric forcing and upper-mantle dynamics to crustal uplift and erosion and their effects on the dynamics of sedimentary systems. Structural analysis of the basin architecture, including paleo-stress assessment, provides important constraints on the transient nature of intraplate stress fields and their effects on the evolution of sedimentary basins.

These results and inferences demonstrate the key role of the Moho during the formation and evolution of extensional basins. Originally defined as an acoustic contrast and subsequently demonstrated as a compositional boundary, the Moho has proven to be a highly dynamic feature of crucial importance to understand spatial and temporal variations in the mechanical state of the lithosphere. Of particular importance is the ratio of crust and lithospheric mantle in extensional regimes during the formation of rifts and their subsequent tectonic reactivation or inversion.

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