Part 1

Classical numerical models of basin formation and evolution with applications to the Central European Basin System (CEBS)

1.1 Kinematic models for basin formation and evolution

1.1.1 Purely thermal models

The first class of models to explain vertical movements in continental regions closely resembled the thermal model which has been successfully used for oceanic lithosphere (e.g. Vogt & Ostenso, 1967). Following this approach, the subsidence of continental shelves could in principle be related to thermal contraction beneath the crust. This conclusion reflects the concept that the tectonic subsidence of continental lithosphere decreases exponentially as a function of time with a time constant very close to that typical of a mid-ocean ridge (Sleep, 1971; Steckler & Watts, 1978). Sleep (1971) proposed a major thermal perturbation as the driving mechanism for subsidence, see Figure 1. Following his model, the thermal anomaly heats the entire lithosphere causing consequent uplifting of the crust by thermal expansion. Subsequent removal of the upper crustal layers by erosion together with the resultant cooling produce subsidence below the original surface level creating a basin.

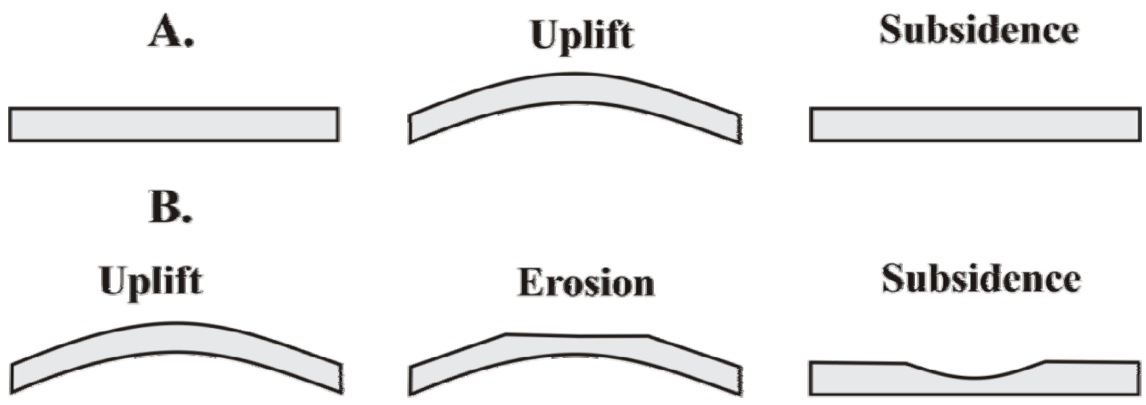

Figure 1. Cartoon illustrating the thermal driven subsidence as proposed by Sleep (1971). Doming due to thermal perturbation causes uplift. Erosion and subsequent subsidence creates a basin.

The model of Sleep (1971) accounts rather well for the time history of subsidence, however, the explanation is inconsistent with the large sediment accumulations frequently observed. Once the temperature of the lithosphere increased, first the surface is elevated and then starts to subside to its original position due to the cooling of the lithosphere, *Case A* of Figure 1. Without erosion this process will not lead to the formation of a sediment flourished basin. The model causes a '*space problem*': given the similarity in density between sediments and upper crustal layers, subsidence should match the amount of material eroded. However, there is no evidence for such a great amount of erosion in deep sedimentary basins. Another problem concerns the mechanism required to explain the heating of the continental lithosphere. An attempt in explaining this '*heating problem*' was suggest by Artemiev & Artyushkov (1971). In order to describe the anomalously high heat flow in the lake Baikal, they proposed an upwelling of hot upper mantle material during the extension phase and a consequent thinning of the crust as driving mechanisms of the observed high heat flow. On the other hand, Haxby et al. (1976) suggested as a cause of the heat flow anomaly a possible intrusion of hot mantle diapirs in the lower portion of the lithosphere which do not produce major deformation at the surface level.

1.1.2 McKenzie's kinematic model

Salveson (1978) then proposed a qualitative model of passive mechanical extension of both crust and sub-crustal lithosphere. He assumed brittle fracture as governing mechanism of failure in the crust, whereas the sub-crustal lithosphere was supposed to fail by ductile necking. The extension-induced formation of a rift sediment-filled basin causes an upwelling of the asthenosphere in order to maintain isostatic equilibrium with subsequent subsidence, erosion of the rift shoulders and continuous sedimentation in the rift basin.
Using the same concept of Salveson (1978), McKenzie (1978a) developed a quantitative model which allows to avoiding both the space and heating problems, i.e. *uniform stretching*. He considered instantaneous and uniform extension of the lithosphere and the crust with passive upwelling of hot asthenosphere to maintain isostatic equilibrium. Mechanical stretching is considered responsible for both heating of the lithosphere and subsidence of the basin. The assumptions related to this model may be summarized as follows:

a. The lithosphere (crust plus mantle) overlies a partially molten asthenosphere.

b. The densities of the crust, lithosphere and the asthenosphere decrease linearly with temperature, whereby the lower lithosphere and the asthenosphere present the same thermal properties. The density of the crust is lower than the densities of the asthenosphere and the mantle-lithosphere.
c. The geotherm is considered to be a linear function from a fixed temperature at the base of the lithosphere (T_a~1300°C) to the surface temperature (T_0).
d. Radiogenic heat production within the crust is neglected.
e. Initial mechanical stretching, β=(original thickness)/(final thickness), is considered instantaneous and volume preserving, i.e. homogeneous within the entire lithosphere=uniform stretching.
f. Lateral temperature gradients are much smaller than vertical gradients, therefore the heat excess is gradually removed only by vertical heat conduction, i.e. no lateral heat flow.

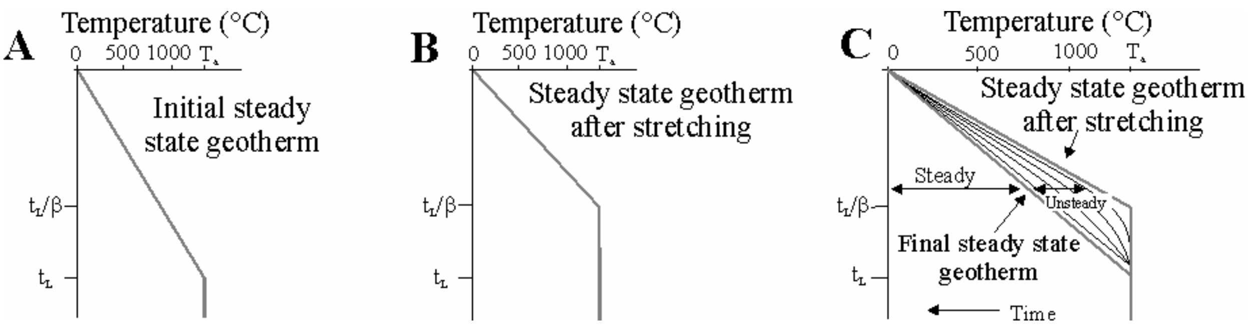

Figure 2. Thermal evolution as a function of time as predicted by McKenzie's (1978a) stretching model.

A: Steady-state geotherm before stretching.

B: Steady-state geotherm after stretching.

C: Final steady-state thermal structure of the lithosphere. After instantaneous increase in heat flow, the temperature decrease exponentially with time. The total temperature is made up of an unsteady and a steady component. Black dashed and dotted curves indicate the transient temperature as a function of time.

Under these assumptions the model predicts the temporal thermal evolution, as shown in Figure 2, and the total amount of subsidence, see Figure 3.
Following McKenzie's (1978a) model, the total amount of subsidence, S_{tot}, consists of two components:

1. *Initial, instantaneous, fault-controlled subsidence* (S_i). Rapid stretching of the lithosphere produces thinning of both the crust and the mantle-lithosphere and

upwelling of hot asthenosphere, perturbing the original structure *case B* of Figure 2. The crust and mantle are thinned while the temperatures at the surface and at the transition to the asthenosphere remain constant. This configuration induces an increase in the thermal gradient. The thinned lithosphere is replaced by less dense asthenosphere causing isostatic readjustment and instantaneous rapid initial subsidence (S_i), *case B* of Figure 3. Following McKenzie (1978a) this initial instantaneous subsidence is given by:

$$(1.1) \qquad S_i = \frac{t_L \left\{ (\rho_m - \rho_c) \frac{t_c}{t_L} \left(1 - \alpha_V \frac{1}{2} T_a \frac{t_c}{t_L} \right) - \frac{\alpha_V T_a \rho_m}{2} \right\} \left(1 - \beta^{-1}\right)}{\rho_m (1 - \alpha_V T_a) - \rho_s}$$

In (1.1) t_L, t_c, and t_m are the initial (un-rifted) thickness of the entire lithosphere, crust and mantle-lithosphere respectively, ρ_c and ρ_m are the (average) density of the crust and of the mantle, β is the stretching factor, ρ_S is the (average) bulk density of sediments or water filling the rifted, and α_V is the volumetric thermal expansion coefficient (for both the crust and the mantle portion).

The sign of the predicted initial subsidence depends on the initial thickness of the crust (t_c) and is independent of the amount of stretching (β). Following McKenzie (1978a) uplift (i.e. $S_i<0$) occurs for pre-stretched initial crustal thickness less than 20 km whereas subsidence (i.e. $S_i>0$) is predicted for crustal thickness initially greater than 20 km.

2. *Long-term, post-rift, thermal subsidence* ($S_T(t)$). The disturbed temperature gradient causes cooling whereby asthenosphere is transformed into mantle. While the crust remains thinned, the entire lithosphere approaches its original thickness, *case C* of Figure 3. This cooling process is non linear with depth, *case C* of Figure 2. The thermal temporal evolution T(z,t) is given by:

$$(1.2) \qquad T(z,t) = Ta \left\{ \left(1 - \frac{z}{t_L}\right) + \frac{2}{\pi^2} \beta \sin\left(\frac{\pi}{\beta}\right) e^{-\frac{t}{\tau}} \sin\left(\frac{\pi z}{t_L}\right) \right\}$$

In (1.2) $\tau = \frac{t_L^2}{k\pi^2}$ with k=thermal diffusivity, is the thermal constant of the lithosphere, z is the vertical spatial direction (i.e. depth) and t is time. The other quantities have the same meaning as in (1.1).

The relaxation of lithospheric isotherms to their pre-stretching position is associated with a second thermal subsidence phase, $S_T(t)$, whose subsidence rate decreases exponentially with time. The thermal subsidence depends only on the amount of stretching (β) and its mathematical formulation is:

$$
(1.3)\quad
\begin{aligned}
S_T(t) &\approx E_0 \frac{\beta}{\pi} \sin\left(\frac{\pi}{\beta}\right)\left(1 - e^{-\frac{t}{\tau}}\right) \\
E_0 &= 4 \frac{t_L \rho_m \alpha_V T_a}{\left(\pi^2 (\rho_m - \rho_S)\right)}
\end{aligned}
$$

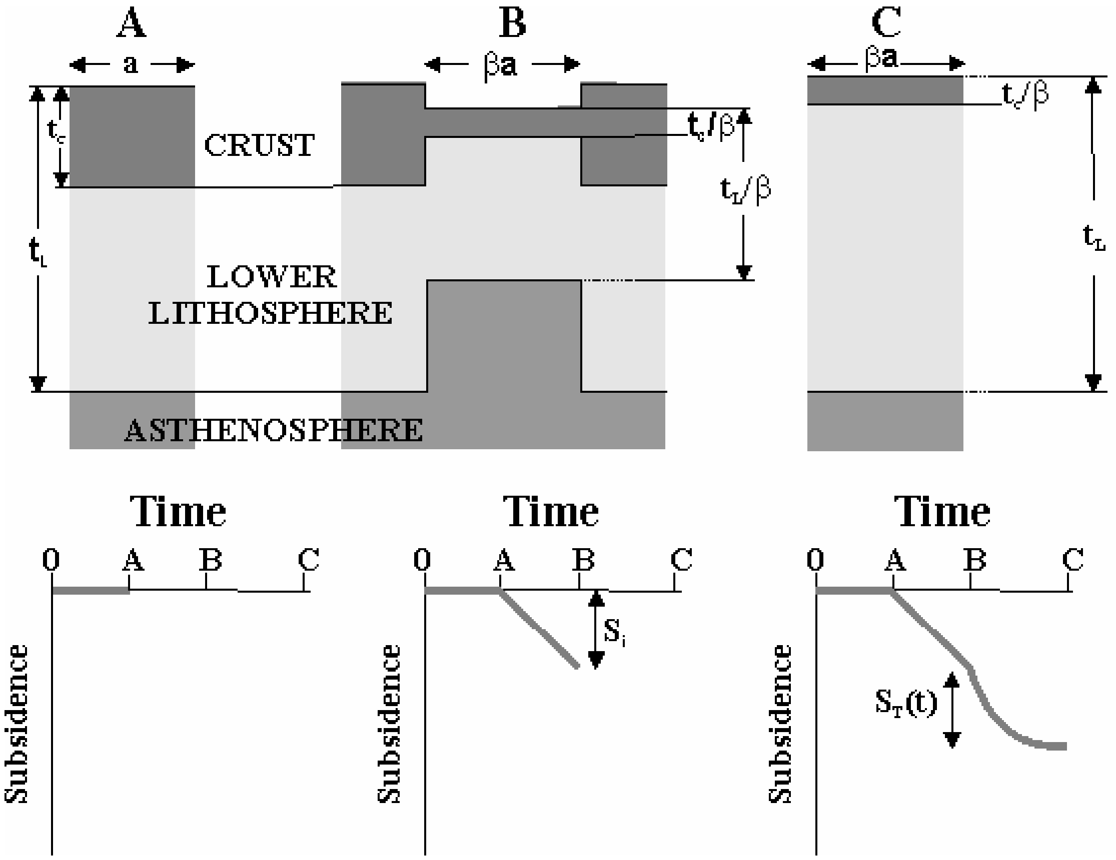

Figure 3. Principal features of the McKenzie's (1978a) subsidence model.

A: Initial conditions. A thermally equilibrated continental lithosphere of total thickness t_L consisting of a crust (thickness=t_c) and a lithospheric mantle (thickness=t_m=(t_L-t_c)) overlies a partially molten asthenosphere.

B: Uniform instantaneous stretching (β). At the time t=0, uniform instantaneous mechanical extension of the lithosphere by a factor β occurs causing vertical thinning of both the crust (thickness=t_c/β) and the mantle lithosphere (thickness=t_m/β=(t_L-t_c)/β). Since the temperature of the material remains unchanged during the extension, isostatic compensation causes upwelling of hot asthenosphere. The resulting gradual decaying of the thermal perturbation produces an initial instantaneous subsidence (S_i).

C: Post-rift evolution. The cooling of the lithosphere following rifting causes a second phase of relative slower time-dependent thermal subsidence (S_T(t)).

Since the initial and final thermal states are known, the temperature distribution and the surface heat flow can be obtained as a function of time. Once the thermal structure is known, the subsidence history of the extended region is easily estimated.
The quantitative aspects of the extensional concept were first applied by McKenzie (1978b) to the Aegean Sea.
Sclater & Christie (1980) then developed an extensional model to account for the Mid-Cretaceous subsidence of the Central Graben area in the North Sea domain. In their work they showed that a stretching factor β of about 2 in the centre and of about 1.25 across the flanks may explain the post-tectonic subsidence across the basin. However, the pure shear model adopted was not able to explain the entire subsidence history of the basin.
Le Pichon & Sibuet (1981) applied the stretching model of McKenzie to the formation of passive continental margins, focusing on the Armorica and Galicia continental margins of the northeast Atlantic. They found that the simple stretching model was capable to explain the relationship between initial subsidence, thermal subsidence and continental crustal thinning providing a reasonable first approximation to the actual physical process of formation of the margin. Moreover, the simplicity of their approach enables to reconstruct the edge of the continent prior to break-up as well as to account for the dynamics of the transition from continental stretching to oceanic accretion. Several other groups have applied these concepts with generally successful results to other basins and shelves (e.g. Royden & Keen, 1980; Wood & Barton, 1983; Brunet & Le Pichon, 1982).

1.1.3 Limitations of McKenzie’s model and corresponding implementations

The simple one-dimensional uniform extension model developed by McKenzie (1978a) provides a quantitative explanation for the general aspects of continental basin evolution. However, many authors have demonstrated discrepancies occurring between the crustal extension β and the initial and thermal subsidence as predicted by the McKenzie’s model with regard to geological observations. As an example, subsidence analysis of well data from Nova Scotia by Royden & Keen (1980) showed that McKenzie’s model predictions may lead to unrealistic overestimates of the initial subsidence. Additionally, crustal extension is commonly much less than the model predicts. Time-dependent models for lithospheric deformation have demonstrated that a purely mechanical source of extension may not be able to generate the heat flow and uplift history as actually observed in a large number of rift zones. Scatler & Celerier (1987) pointed out that the post-rift thermal subsidence exceeds the syn-rift subsidence too much. The assumption of instantaneous stretching of the lithosphere followed by thermal subsidence during the re-equilibration stage of the lithosphere provides a simple initial condition for the thermal subsidence of the basin. However, it is reasonable for modelling a slab of infinite horizontal extension undergoing a stretching event for less than 20 Ma. Jarvis & McKenzie (1980) demonstrated that for a finite time of rifting the relative amount of subsidence during extension, i.e. syn-rift subsidence, increases. Prolonged periods of rifting may cause significant amounts of lateral heat loss thus enhancing the amount of subsidence during the stretching phase. The longer the stretching event, the more important lateral heat loss should become. Additionally, Cochran (1983) suggested that supplementary heat loss away from the basin centre may facilitate increased early basin subsidence and parallel thermal uplift of the basin flanks. Steckler (1981) showed that also for basins with a width less than about 100 km, the lateral heat loss plays a crucial role. Again, cooling and subsidence in this case are more rapid than those predicted by the McKenzie's model.

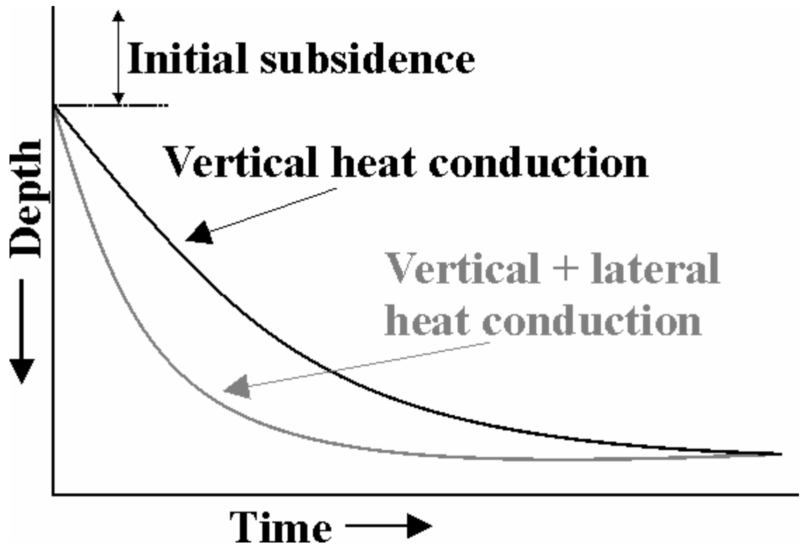

Figure 4. Effects of lateral heat conduction on subsidence history. Lateral heat conduction (light grey curve) provides a faster subsidence than the one predicted by McKenzie's (1978a) simple stretching model (black curve).

To account for lateral heat conduction effects, Steckler (1981) expanded the one-dimensional vertical heat flow model in two dimensions by including conduction also in one horizontal direction, see Figure 4. Cochran (1983) applied this implementation and developed an analytical technique to examine an arbitrary rifting history in both time and space. The adopted technique allowed him to calculate the thermal structure of the lithosphere throughout the rifting event and to trace the subsidence history and surface heat flow of the developing basin. To simulate a continuum finite rifting process, he subdivided the extension in a large number of short (about 0.5 Ma), discrete events during which the lithosphere was alternately allowed to rift and to cool. He investigated the effects of an extended rifting event for the simple case of a sediment free basin, i.e. water filled. The obtained results showed several deviations between the instantaneous case and the finite rifting case in terms of subsidence history, distribution of sediments and thermal history for extension time as short as 5 Ma. The main effect of considering a finite-duration extension event is a heat loss during rifting which leads to an increase in the syn-rift subsidence at the expense of the post-rift. Specifically, a decrease of more than 24% for the post-rift subsidence associated with a rifting event of 20 Ma and of 10-15% for a shorter 10 Ma rifting event were calculated.

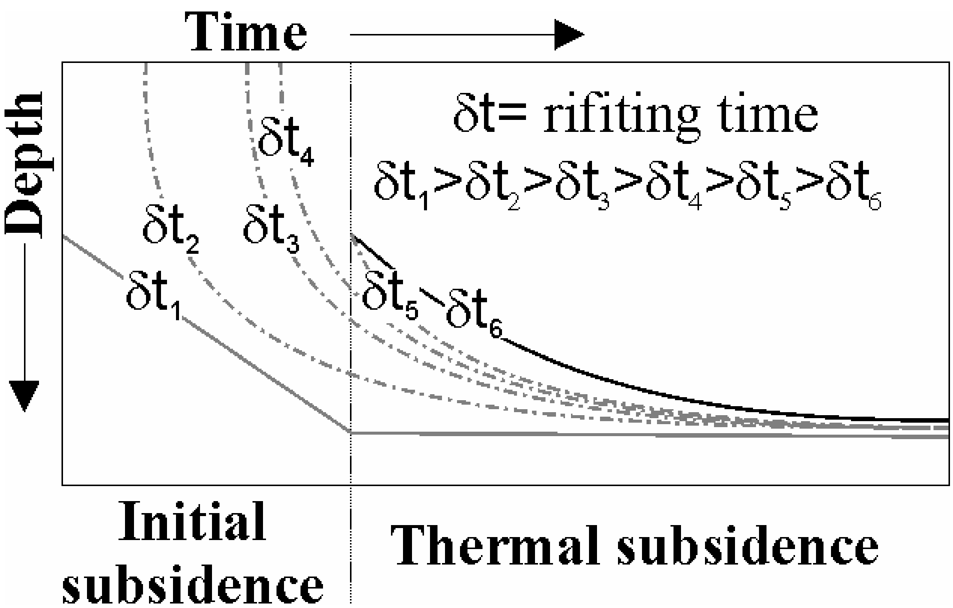

Figure 5. Effects of finite rifting rates on the thermal (post-rift) subsidence. The continuous black curve represents the end-member subsidence curve for the case of instantaneous (δt_6=0) uniform stretching as developed by McKenzie (1978a). The continuous grey curve is the (opposite) end-member subsidence curve for the case of infinitely slow rifting rate ($\delta t_1 \sim \infty$). In this situation the mantle lithosphere will remain cool and almost no thermal (post-rift) subsidence will occur after the rifting has ceased. The dashed and dotted grey curves (δt_i, i=2,…,5) are subsidence curves for different finite rates of rifting. As a consequence of considering a finite duration of extension, an increase in the syn-rift (initial) subsidence occurs at the expense of the post-rift (thermal) subsidence. This feature is more important the longer the rifting event is considered and it is reflected by a progressively flattening of the post-rift subsidence curves.

This aspect is reflected by a flattening of the post-rift subsidence curve which may lead to underestimate the real amount of extension in comparison with theoretical curves as derived from the one-dimensional rifting model. As a result a more rapid subsidence in the early history of the basin evolution occurs followed by a slower thermal subsidence than predicted by McKenzie's model, see Figure 5.

A further consequence of considering lateral heat flow is an increase in total heat flow and subsequent thermal expansion which causes significant uplift across the basin flanks. However, when applied to a sediment filled basin, the model of Cochran (1983) gives unrealistically large ratios of syn-to-post-rift sediments: for a 20 Ma rifting time, about 77% of the sedimentary column is represented by syn-rift sediments.

The rapid subsidence and sediment accumulation characterizing pull-apart basins can in principle be approximated by the model of lithospheric extension as developed by McKenzie (1978a). However, the small size of pull-apart basins implies very large lateral temperature gradients, so that lateral heat conduction through the basin walls becomes crucial. Following the studies of Steckler (1981) and Cochran (1983), the critical width for significant lateral heat loss ranges from 100 km up to 250 km. Within narrow basins the syn-rift subsidence is greater than predicted by the uniform instantaneous stretching model due to the high amount of cooling during extension. As an example, Pitman & Andrews (1985) adopted a model close to the one used by Cochran (1983) to study the subsidence and thermal history of small pull-apart basins associated with the San Andreas transform system. Syn-rift heat loss was simulated by repeatedly stretching and cooling phases of the lithosphere during small time increments (about 0.05 Ma) with reference to a two-dimensional model. The model was tested with three hypothetical basins of different initial widths (respectively 10 km, 20 km, and 30 km) stretched up to a maximum of 1.6 times their original width under a constant stretching rate of 3 cm/yr. The results pointed to a strong control exerted by the initial basin width on the magnitude of the cooling time constant. They found that the subsidence during extension was more rapid for the basin associated with the least initial width. Unfortunately, following the model of Pitman & Andrews (1985) due to lateral heat loss, the subsidence curves as well as the isotherms soon tend to approach equilibrium. Consequently, sediments deposited after the initial stage of rifting appear unlikely to reach temperatures sufficient enough for maturation.

In summary, McKenzie's (1978a) model provides an attractive tool to quantify one-dimensional subsidence by simple parameters like the stretching factor even if only well data are available. However, it remains a first order approximation with strong limits even for simple basins. The limits are related to the strictly one-dimensional nature of this approach which neglects lateral variations observed in the real world as well as to the simplifying assumptions concerning the temperature distribution in the lithosphere and the rheology of the earth.

1.1.4 Non-uniform stretching models: discontinuous and continuous stretching with depth

The simple geometrical uniform, depth-independent stretching model does not take into account the layered rheological stratification of the lithosphere. Since rheological properties vary with temperature and pressure, the lithosphere is expected to extend more realistically in a non-uniformly manner with depth. The distribution of lithospheric extension with depth may be discontinuous or continuous as represented in Figure 6.

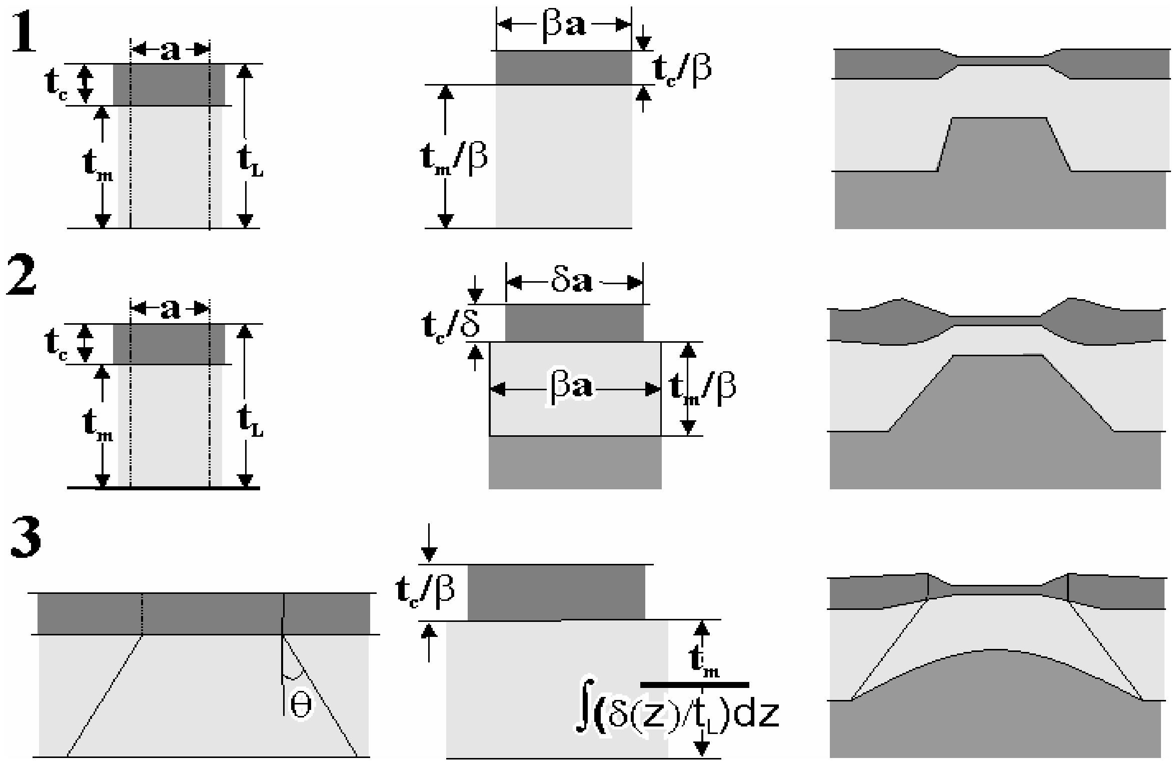

Figure 6. Differences between (1) uniform, (2) discontinuous with depth, and (3) continuous with depth stretching models. (1) Uniform stretching model: the crust (initial thickness=t_c) and the mantle lithosphere (initial thickness=t_m) are stretched by an identical amount (β). (2) Discontinuous stretching with depth: the crust is stretched by a different amount (δ) than the mantle lithosphere (β, with β>δ). The difference in the amount of crustal and sub-crustal extension requires a decoupling between the two layers. Both crustal and sub-crustal extensions are considered independent but uniform throughout their respective layers. (3) Continuous stretching with depth: in the crust the stretching is the same as in the above described situation (2), whereas in the sub-crustal layer the stretching is a continuous function of depth.

In the first case the upper and lower portion of the lithosphere are decoupled at a certain depth that may or may not correspond to the crust-mantle boundary. The portion of the lithosphere above this horizon extends by a certain factor δ, with regard to the underlying lower lithosphere, which extends by a factor β. The upper and lower lithospheric extensions are considered independent but uniform throughout their respectively layer thickness. Clearly, when δ=β uniform extension occurs. The main differences with respect to the uniform stretching may be summarized in the following points:

- The thermal gradient after extension is no longer linear with depth, but '*two-legged*'.
- Discontinuous stretching raises the lithosphere-asthenosphere boundary to a more shallow level than in the uniform case thus solving the 'heating problem'. Thinning the sub-crustal lithosphere more than the upper crust may lead to an increased heat input during extension.
- A minor amount of crustal thinning for the same total subsidence is required.

As in the uniform case, the total subsidence is made up by two contributions, an initial, fault-controlled, subsidence and a following thermal subsidence. The initial subsidence depends on the amount of the initial crustal thickness (t_c), the decoupling depth and the relative magnitudes of δ and β whereas the thermal subsidence is hardly affected. Considering the situation of instantaneous stretching, i.e. no heat loss during the rifting process, the thermal subsidence reflects the amount of sub-crustal thinning. This provides a simple way to estimate the stretching factors of the crust and mantle-lithosphere directly from the amount of syn- and post-rift subsidence.

Royden & Keen (1980) applied this model to the Nova Scotia and Labrador continental margins. In their formulation the decoupling horizon was located at the base of the crust and instantaneous extension was assumed. A general agreement was found between the theoretical subsidence predicted by their model and the subsidence calculated from deep well data. Moreover, the results explained the uplift typically experienced by many basin margins during early rifting phases and showed that the thermal subsidence can account for the long-term tectonic subsidence observed.

A more refined model approach was adopted by Van Wees et al. (2000) to explain the Late Permian-Early Jurassic evolution of the Southern Permian Basin, the southern part of the CEBS. Detailed thickness and facies analyses based on wells and reflection-seismic data indicating only minor faults contradict the simple pure shear model used previously (e.g Dadléz et al., 1995). In order to establish a quantitative framework for the evolution of the Southern Permian Basin, Van Wees et al. (2000) proposed a model allowing for multiple rifting phases as well as differential extension of the crust and the mantle-lithosphere as illustrated in Figure 7. The obtained results indicated that the basin wide Late Permian-Triassic subsidence required an active mechanism, although there is almost no evidence of synchronous active faulting. In this regard, a relative important component of Late Permian and Triassic tectonic subsidence can be explained by thermal relaxation of a major lithospheric thermo-mechanical attenuation during Early Permian and by consequent delayed infilling of the topographic depressions developed during Late Permian. The results of Van Wees et al. (2000) suggested a thermal-induced thinning of the mantle rather than a mechanical driven crustal extension as responsible mechanism for the Early Permian stretching event. Following their hypothesis, thinning of the crust below the Southern Permian Basin may be partly attribute to its mechanical extension and partly to magmatic destabilization of the crust-mantle boundary followed by reactivation of the lower crust.

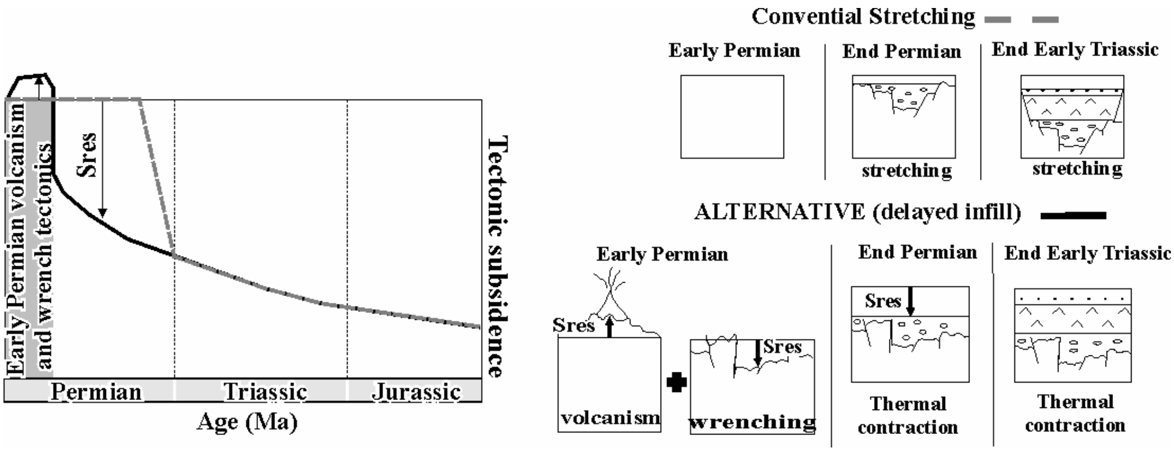

Figure 7. Classical stretching model („Conventional stretching“) and alternative (“delayed infill”) model as proposed by Van Wees et al. (2000) for the Permo-Triassic evolution of the Southern Permian Basin. **Sres** depicts the evolution of the relative magnitude of accommodation space as a function of time for both models (grey dotted line refers to the conventional stretching model, and continuous black line to the alternative delayed infill model).

Although discontinuous non-uniform stretching models have been successful in explaining some first order features in long-term subsidence patterns of basins, they rely on a number of requirements:

- The focal depths of earthquakes in old cratons suggest that the upper part of the lithosphere has relatively high strength and shows active seismic activity. In contrast, the underlying lower part is almost completely a-seismic. This difference may contribute to the ductile deformation mechanisms within the lower portions of the lithosphere which may lead to a rheological de-coupling at mid-crustal level. This feature may induce different amounts of extension within the two layers (e.g. Sibson, 1983; Ranalli, 1995). However, the existence of an intra-lithospheric discontinuity is not universally proven.

- The mechanism by which the lithospheric mantle may stretch differently is an 'a-doc' requirement which creates another space problem within the mantle. To overcome this limitation, a combination of extension and magmatic intrusion during rifting has been proposed (e.g. Royden & Keen, 1980). However, it leads to a complex formulation hiding the natural simplicity of the model.

All the above requirements may be removed by considering a non-uniform but continuous stretching with depth. This formulation requires the stretching to be a continuous function of depth in the mantle-lithosphere with a decreasing rate with depth as the extension is diffused over a wider area. The amount of stretching depends on the depth beneath the crust and on the angle θ between the vertical and the boundary of the stretched region. Greater values of θ increase the amount of the initial subsidence, while at the same time reducing the amount of the post-rift thermal subsidence. There are mainly two important implications of stretching the mantle over a wider area than the crust:

1. A point located at the rift shoulder experiences an initial uplift followed by a subsequent phase of subsidence. Sleep (1971) has demonstrated that in absence of erosion it will approximately return to its initial elevation, while if erosion occurs it will sink to a deeper level than the initial one.
2. Predicted stratigraphic onlap at basin margins causes the so-called '*steer's head*' geometry during the following post-rift subsidence phase (e.g. White & McKenzie, 1988).

1.1.5 Simple shear model of Wernicke

In general, pure shear models cannot account for the asymmetry and/or uplift of the flanks as often observed in basins. To overcome these limitations, a different method for lithospheric extension has been proposed by Wernicke (1981, 1985). Based on studies of Basin and Range tectonics, he suggested that lithospheric extension may be accomplished by displacement on a large-scale, gently dipping shear zone cutting throughout the lithosphere, Figure 8. This shear zone transfers extension from the upper crust to a region of the lower crust and lithospheric mantle elsewhere. Consequently, a physical separation of the zone of fault-controlled extension from the zone of upwelling asthenosphere occurs.

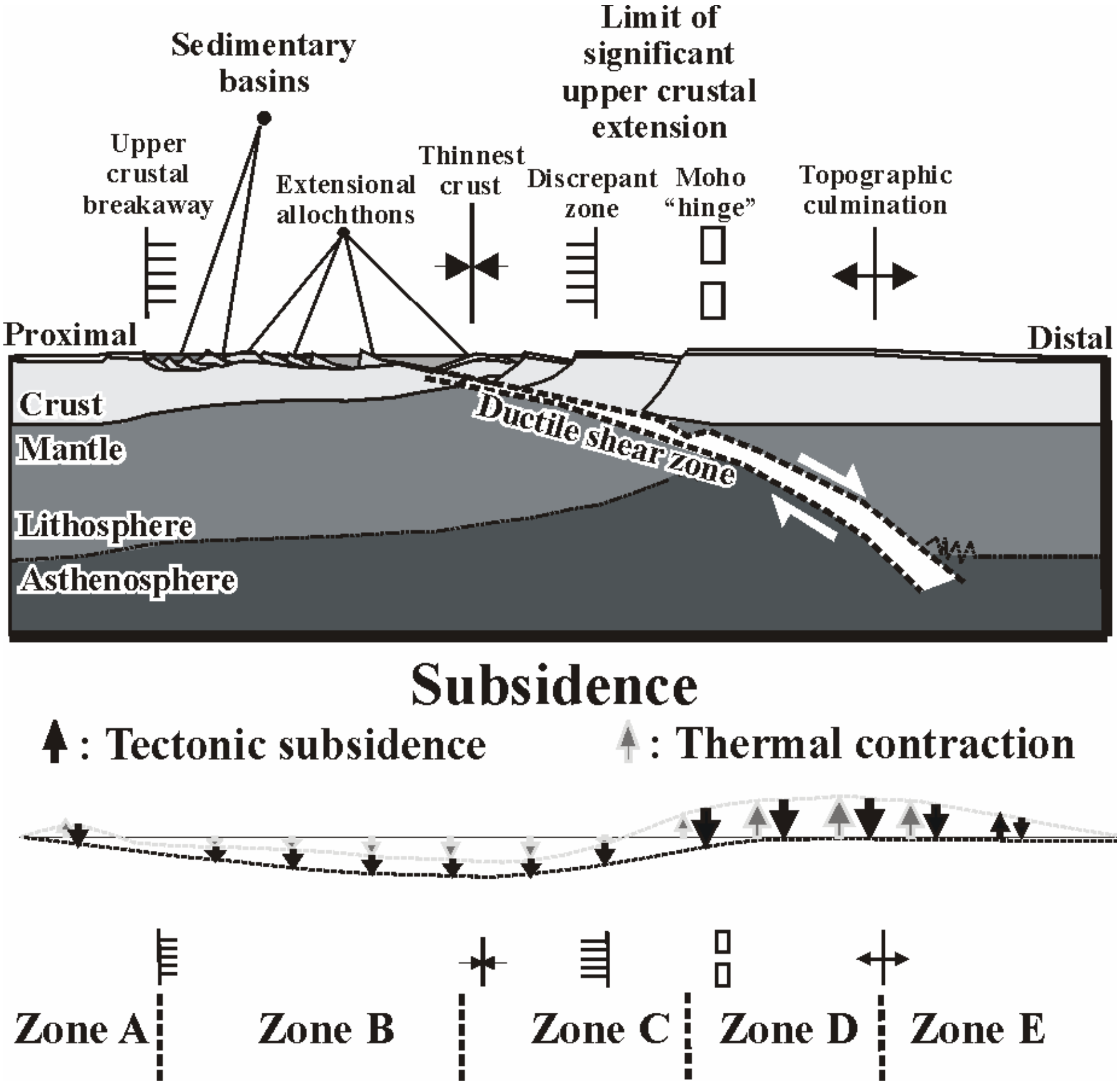

Figure 8. Normal simple shear model of the entire lithosphere on a low-angle ductile shear zone, after Wernicke (1985).

Zone A: region of no extension in both the crust and the mantle lithosphere ($\beta_{crust}=\beta_{mantle}=1$).

Zone B: region of thin-skinned fault-controlled tectonic ($\beta_{crust}>\beta_{mantle}=1$).

Zone C: region where the shear zone enters the mantle lithosphere (β_{crust} or $\beta_{mantle}>1$).

Zone D: '*discrepant zone*' ($\beta_{mantle}>\beta_{crust}=1$).

Zone E: boundary for the discrepant zone, (β_{crust} or $\beta_{mantle}=1$).

Following Wernicke (1985), there are three main domains within an extensional shear zone:

1. A zone where the upper crust has thinned and abundant faults occur above the detachment zone (thin skinned tectonics).
2. A '*discrepant zone*' where the lower crust has thinned while there is negligible thinning in the upper crust.
3. A zone where the shear zone extends through the mantle-lithosphere.

These zones then have a different subsidence history: tectonic subsidence beneath the thin-skinned domain is counteracted by tectonic uplift in the region overlying the thinned lower crust and lithospheric mantle, the 'discrepant zone' of Wernicke (1985). The discrepant zone will then suffer thermal subsidence caused by cooling of the asthenosphere. This thermal subsidence will restore the crust to its initial level if no erosion has occurred, or, if sub-aerial erosion has taken place, it will lead to the formation of a shallow basin. The basement of thermal subsidence should be un-faulted. Beneath the zone of thin-skinned tectonics no detectable thermal subsidence should be observed. As discussed by Wernicke (1985), a thin-skinned extensional zone should (a) show tectonic subsidence but no thermal subsidence beneath the extended upper crust, and (b) show a 'discrepant zone' where the lower crust and the mantle-lithosphere have thinned. This 'discrepant zone' should subside to its initial crustal level or below, in case of erosion, and hence should be the site for a shallow basin which does not present extensional faults, i.e. a sag-type basin.

Wernicke (1985) applied this model to the Basin and Range province. Buck et al. (1988) proposed a similar extensional model for the Viking Graben in the northern North Sea. The obtained results explained the geometry of the crustal-scale faults but were not able to account for the presence of the thermal subsidence beneath the North Sea. Specifically, the Wernicke shear zone model cannot explain basins which experience a thermal subsidence spatially superimposed on a fault-controlled subsidence.

1.1.6 Asymmetrical stretching of the crust

In order to overcome the limitations associated with the simple model of Wernicke (1981, 1985), Coward (1986) proposed an alternative taking into account the evidence for crustal scale simple shear as well as the subsidence history. The model involves a combination of widespread upper crustal fault-controlled extension above a more localized region of concentrated sub-crustal extension, see Figure 9.

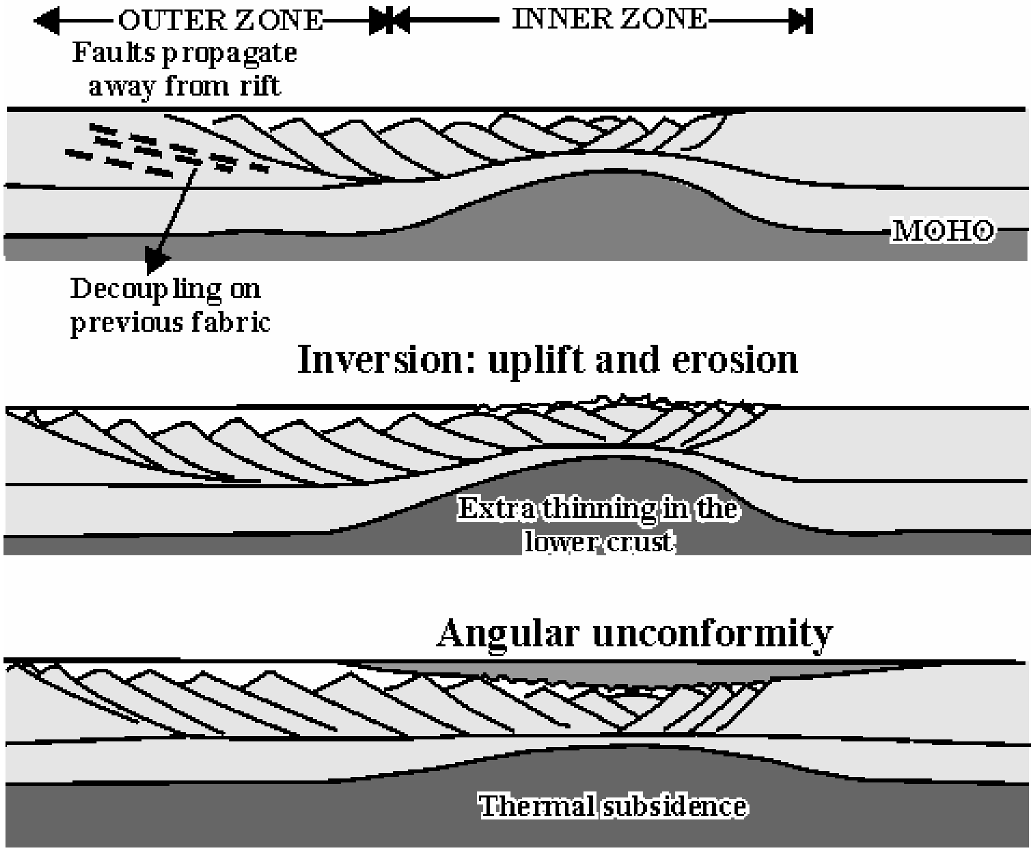

Figure 9. Model for heterogeneous thinning of the lithosphere after Coward (1986).

The invoked combination may originate either during a single period of stretching or as a result of a two-phase process. The latter would comprise a first stage of roughly uniform extension throughout the lithosphere followed by a second phase with extension confined to a relatively narrow domain in the sub-crustal portion of the lithosphere. Localized extension of the lower crustal/mantle-lithosphere is favoured by strain softening which may be caused by increased geothermal gradient, fabric development and reduction in grain size. These processes induce a non-uniform deformation in the sub-crustal portion of the lithosphere leading to localized deformation in zones of more intense stretching or simple shear. On the contrary, in the upper crust, planar brittle faults tend to lock after a certain amount of extension so that deformation is transferred to new steep faults. According to the model of Coward (1986), the new generation of faults may spread outwards essentially along one direction away from the initial rift zone widening the zone of extended upper crust. Additionally, the presence of some earlier gently dipping anisotropy(ies) or compositional layering(s) within the crust may enhance the expansion of the upper crustal stretching domain. Within this model, patterns of uplift and subsidence vary across the basin. A typical sedimentary basin could have an *outer zone* where only the upper crust is stretched and an *inner zone* where the upper crust is stretched by a factor β smaller than the lower crust/upper mantle (β+β'). The additional stretching factor β' in the lower portion of the lithosphere is required to balance the extension in the upper crust at the basin margin. The extension and fault development in the upper portion of the lithosphere is not necessarily symmetrical across the basin.

The model predictions may be summarized as follows:

- The zone where only the upper crust has thinned, i.e. the outer zone, undergoes an initial tectonic subsidence related to thinning of the upper crust. The amount of this tectonic subsidence is less than in the case when thinning has occurred throughout the entire crust. No thermal subsidence will then occur within this portion of the lithosphere.
- In the inner zone, where the whole lithosphere has thinned, both initial tectonic subsidence and later thermal subsidence occur. Stretching of the whole lithosphere of a factor β will cause subsidence, however the additional factor β' in the lower crust and lithospheric mantle causes early uplift.
- With extreme density variations in the lower crust a net uplift can occur. This will raise the stretched upper crust above sea level causing unconformities.

These processes may explain the Late Jurassic-Early Cretaceous 'Cimmerian' unconformity observed in the North Sea domain. Following Coward (1986), the development of Jurassic unconformities in the North Sea may be due to slight uplift caused by heterogeneous stretching during regional extension rather than being the product of different phases of regional 'Cimmerian' compression as suggest for example by Barton & Wood (1984).

1.2 Flexural models for basin formation and evolution

1.2.1 Lithospheric flexure due to sediment loading

In conventional stretching models the vertical loads which act on the lithosphere are usually subdivided into three main components (e.g. Kooi et al., 1992):

1. Loads due to crustal thinning.
2. Changing loads due to the gradually decaying thermal anomaly.
3. Loads due to the presence of sediment or water infill.

The first two members of the above classification are directly attributed to the stretching process, whereas the third contribution is preferentially regarded as an indirect consequence of extension. All the models presented above invoke local (Airy) isostasy to compensate these loads. Local isostasy assumes that the lithosphere is unable to support vertical stresses so that any vertical force is compensated only by lithospheric buoyancy. Consequently, the lithosphere behaves like a solid with zero threshold shear stress under any impose vertical load. At the same time, it presents enough lateral strength to prevent deformation induced by horizontal stress gradients caused by variations of lithostatic pressure at given compensation depths within different balancing columns. Local isostasy is irreconcilable with any realistic and self-consistent rheological representation of the lithosphere (e.g. Fernandez & Ranalli, 1997). The weak point of this assumption is that the lateral strength of the lithosphere is not taken into account. The strength of the lithosphere is able to redistribute the support of a load by a lateral, i.e. horizontal, transmission of stresses. Consequently, the lithosphere will respond to an applied load by flexure. The simplest model for regional isostatic compensation is that of a thin elastic beam. Many flexural studies have modelled the response of the lithosphere to long-term geological surface loads adopting either an elastic or a visco-elastic (Maxwell) plate model for the lithosphere overlying an inviscid substratum (e.g. Walcott, 1970; Hanks, 1971; Watts & Cochran, 1974; Watts et al., 1975; Banks et al., 1977; McKenzie & Bowin, 1976).

Considering a thin elastic plate, the parameter needed is the flexural rigidity (D). For an elastic plate of thickness=h to which a vertical load q(x) is applied, the flexural rigidity D is defined by:

(1.4) $$q(x) = -D\frac{d^4\omega(x)}{dx^4}$$

In (1.4) $\omega(x)$ is the distance from the neutral surface as a function of horizontal distance (i.e. displacement of the elastic plate).

An alternative common way to quantify the rigidity of the lithosphere is by mean of its effective elastic thickness, T_e. Explicitly, the effective elastic thickness (T_e) is related to the flexural rigidity (D) by means of the lithospheric elastic parameters, i.e. Young’s modulus (E) and Poisson’s ratio (υ), by:

(1.5) $$D = \frac{ET_e^{\,3}}{12\left(1-\upsilon^2\right)}$$

The parameter determines the amplitude and the wavelength of lithospheric flexure. The end-member case of zero T_e corresponds to local isostasy. The effective elastic thickness is defined by an isotherm (~450°C) for oceanic lithosphere. However, this empirical relation is debatable for continental lithosphere. For the case of continental lithosphere T_e values depend on several factors like the thermal structure of the plate, crustal and mantle thickness, and on the way these elements interact during the process of lithospheric deformation (e.g. Cloetingh et al., 1995; Van der Beek et al., 1995). Moreover, the same value of T_e is not constant through time. During rifting the isotherms move downwards as a result of the cooling of the lithosphere. This is kept with an increase of the effective elastic thickness with age, which leads to a widening of the basin.

On the other hand, whenever adopting a visco-elastic formulation, the amplitude and wavelength of lithosheric flexure is an exponential function of the Maxwell relaxation time constant (τ), i.e. $\omega(x) \propto e^{-\frac{t}{\tau}}$.

One of the most important sources of vertical load acting on the lithosphere is represented by its sedimentary infill. The sediments that have been accumulated in sedimentary basins during geological times act as a load on the lithosphere which responds by flexure. Watts et al. (1982) were the first to apply regional compensation to sediment loading. In their study, they referred both to elastic and visco-elastic mechanical flexural models of the lithosphere under similar sediment loading histories. They found that the dominant mechanisms affecting basin subsidence are thermal contraction, thinning of the plate at the time of basin formation and sedimentary loading. While thermal contraction controls space available for sedimentation, sedimentary loading has a strong impact on the stratigraphic evolution of the basin. A comparison between the results obtained by considering a purely elastic and a visco-elastic analogue for the lithosphere, led Watts et al. (1982) to assert that an elastic model comprising an increase with time of the basement rigidity was more suitable to explain the overall shapes of many sedimentary basins rather than a visco-elastic model. In this regard, the main tecto-stratigraphic features predicted are: (1) increased in basin widths during their evolution as observed in several geological settings (e.g. West Siberia, Gippsland, North Sea during the Permian, costal plains of Atlantic-type continental margins), and (2) a progressive onlap of younger sediments onto basement at the basin edges. After sedimentary flexural-controlled deposition, in the case of erosion of the basin and/or its edges the elastic plate model were capable to explain basic features of basins with the youngest sediments restricted only to the basin centre (e.g. Michigan Basin or Paris Basin).

In general, when comparing a basin with Airy isostasy with a basin with flexural isostasy with the same basement subsidence, different stretching factors can be computed. Specifically, neglecting the effects of lateral strength may lead to a significant underestimate of the total amount of extension across the entire basin.

1.2.2 Finite strength of the lithosphere: the depth of necking

Until recently, most flexural basin models assumed the basin fill as the only load inducing flexure. Examples of this type of '*flexure sediment loading models*' comprises: Beamount et al. (1982), Cloetingh et al. (1985), Kooi & Cloetingh (1989a,b). Following this approach, thermal contraction and crustal thinning do not contribute to flexure and are only locally compensated. Stephenson et al. (1987) were the first to invoke regional compensation for the forces arising from both sediment loading and thermal contraction. In their study, they applied this approach only to the situation of a non-flexed lithosphere at the end of rifting. This condition requires a zero flexural rigidity for the lithosphere during the rifting process. However, as pointed out by Braun & Beaumont (1989a,b), a finite strength of the lithosphere may strongly affect model predictions. The presence of a finite strength of the lithosphere during the rift phase implies a further internal source of vertical loading apart from sediment loading. The incorporation of this internal vertical load requires a detailed knowledge of the necking process during rifting. Predictions from dynamical models of lithosphere extension have demonstrated that variations in the necking depth level affect the flexural state of extensional basins. Following Braun & Beamount (1989a) the depth of necking can be defined as "the level of non vertical motions in the absence of gravity or buoyancy forces". At present, the depth of necking is difficult to constrain. Theoretically, it can be estimated from predictions derived from dynamical studies of rifting incorporating a realistic rheological configuration of the lithosphere (e.g. Braun & Beamount, 1989a,b; Kooi et al., 1992). The depth of necking is likely to be controlled by the vertical position of maximum strength. For oceanic lithosphere, it coincides with the brittle-to-ductile transition at lithospheric mantle level. Unfortunately, continental lithosphere may present several brittle-to-ductile transitions at different depth levels due to its complex rheological layering. This might result in independent necking levels within the lithosphere decoupled by domains of relatively low yield strength. Moreover, the rheological configuration of continental lithosphere is usually very heterogeneous and temperature dependent. The depths of the brittle-to-ductile transition at intra-crustal level tend to shift upwards during rifting. Accordingly, also the associated crustal necking level will move upwards thus favouring detachments at shallow levels in the crust.

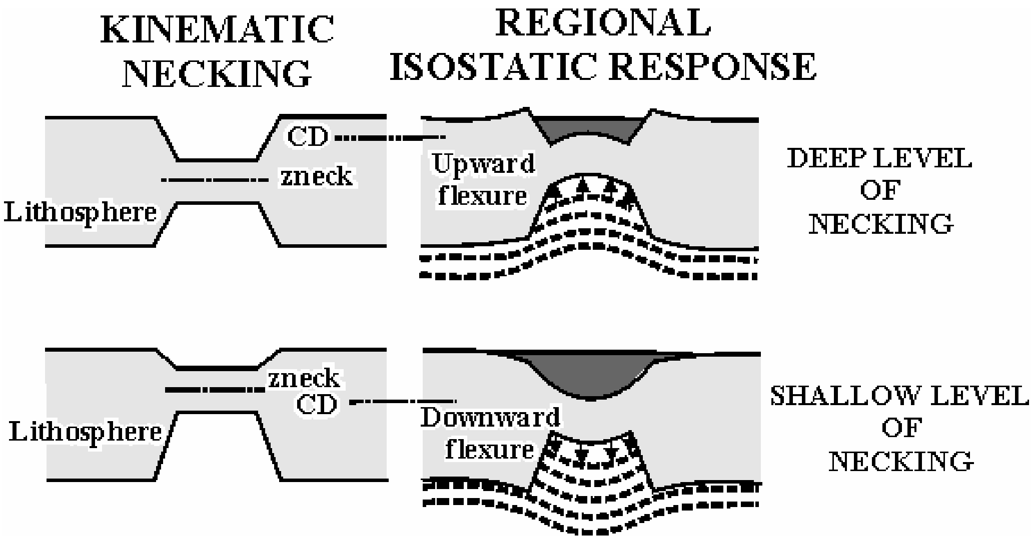

Figure 10. Main features of necking of the lithosphere, modified after Braun & Beaumont (1989). Levels of necking (z_{neck}) creating a surface depression deeper than compensated depth (**CD**=locally compensated basin depth) produce an upward load acting on the lithosphere and a corresponding upward flexural state, upper figure. On the contrary, shallow necking levels result in a downward load acting on the lithosphere and a consequently downward state of flexure, lower figure.

In general, necking levels creating a surface depression deeper than the depth of isostatic compensation result in an upward state of flexure and consequent flexurally supported rift shoulders. On the contrary, a down-ward flexural state may be regarded as the consequence of a shallower surface depression than the depth of isostatic compensation. This will result in down-warped basin flanks and in the production of a buried hinge zone, see Figure 10. In case the depth of necking prevents lithospheric (un)loading ('neutral level of necking') no flexural deflection occurs, i.e. local isostatic equilibrium dominates (e.g. Fernandez & Ranalli, 1997).
To investigate the effects of the depth of necking in the evolution of extensional basins, Kooi et al. (1992) developed a numerical study on the state of lithospheric flexure concentrating on the area of the Gulf de Lions rifted margin in SE France. Their results suggested that necking depths shallower than 15 km produce a downward flexural state associated with down-warping of the rift shoulders. In contrast, 'deep levels of necking' (>20km) result in an upward state of flexure with associated flexurally supported rift flanks. For intermediate values (15-20 km) a transition from upward to downward states of flexure occurs. Deep necking levels (shallow necking levels) are associated with overdeep (underdeep) basins relative to models invoking local isostasy. Moreover, in the absence of significant thermal anomalies and/or underplating beneath the rift shoulders, deep levels of necking provide a reasonable mechanical explanation for long-standing rift flank uplifts.

1.2.3 The role of intra-plate stresses: uplift and basin formation in compression

Although intra-plate stresses can play an important role in the post rift evolution of the stratigraphic sequence, early studies of the flexural response of the lithosphere have often neglected their contribution. Assuming zero vertical loads on the lithosphere, these studies showed that for reasonable levels of compressional forces, the induced vertical displacements of the lithosphere can be neglected. This feature, combined with the lack of folding of the whole lithosphere, led to underestimate the role of intra-plate stresses on continental deformation.

In principle, the first-order effects of variations in intra-plate stresses on basin formation and evolution can be estimated within all the above described models for continental deformation. As an example, most of the stretching models discussed above can also be applied in compressional regimes causing crustal shortening, uplift or inversion. In case of McKenzie's (1978,a) pure shear model the only requirement is that $\beta<1.0$ and that lithosphere is transformed into asthenosphere during subsequent heating of the root. The same of course holds for the other simple non-uniform stretching models.

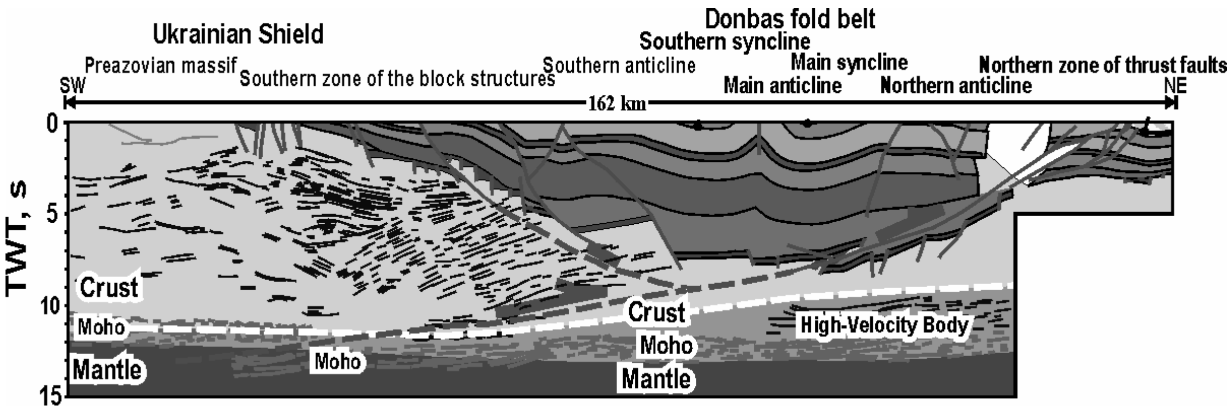

Figure 11. Interpreted structure of the Donbas area along the deep reflection line DOBREflection, Maystrenko et al. (2003a,b). The dash white line indicates the top of the high velocity lower crust modified after DOBREFRACTION'99 Working Group (2003).

The simple shear model of Wernicke has an exceptional compressional counterpart in the inverted southern part of the Dniepr-Donets paleorift (Ukraine), the Donbas (Maystrenko et al. 2003a), see Figure 11. The basin originated in the Devonian and was induced by a typical wrenching event as well documented by a sequence of mostly small scale basement-faults at the base of the sedimentary sequence. A thermal subsidence phase followed, as expected, with some additional complications in the Permian. Probably at the end of the Cretaceous the

Donbas-part of the basin finally was inverted. Due to the interpretation of the deep reflection seismic line DOBREflection by Maystrenko et al. (2003a,b), the inversion caused failure at the Moho-level of the old Archaean and cold crust, continuing into a lystric shear zone which finally ends at the former eastern end of the basin or rift. In addition a backthrust developed, generating a mega horst or "pop up structure". Here Wernicke's model is put from the head to the feet: it is not the initial stretching event causing the lystric detachment but rather the inversion with a detachment surface below the Moho, an aspect which requires rheological considerations as described in the next sub-chapter.

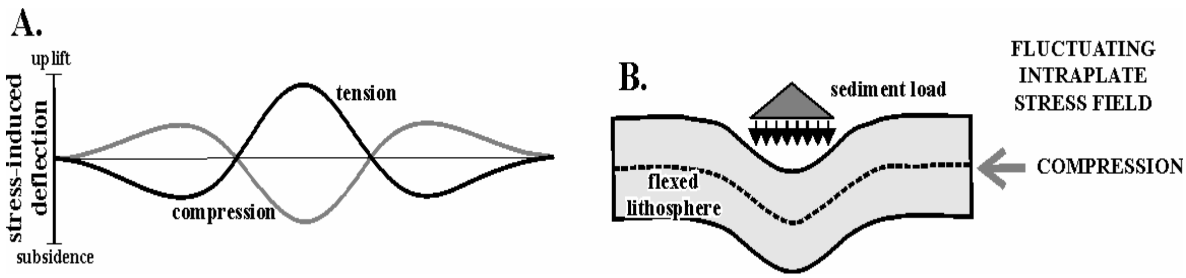

Figure 12. Effects of intra-plate stresses on the development of basin stratigraphy, modified after Cloetingh & Kooi (1992).

A: Stress-induced differential deflection for tension (black curve) and compression (grey curve).

B: Flexural deflection of an elastic lithosphere due to the loading effect of the sedimentary cover superimposed on a compressional stress regime. Superimposed compressional stresses cause relative uplift of the basin flanks and induce higher amounts of subsidence at the basin centre with respect to those predicted by conventional 'sediment loading models'.

Studies of the tectonic stress field within the lithospheric plates have shown a causal relationship between processes affecting the plate boundaries and intra-plate deformation (e.g. Cloetingh & Kooi, 1992). Intra-plate stresses can modulate the long-term thermal subsidence of post-rift sedimentary basins (e.g. Cloetingh et al., 1985) and can induce differential vertical motions of sign and magnitudes which are variable along the basin profile as shown in Figure 12. Cloetingh et al. (1985) have demonstrated that for conventional sediment loading models, e.g. Watts et al. (1982), a superimposed compressional stress state causes relative uplift of the basin flanks, subsidence at the centre of the basin and seaward migration of the shoreline thus giving rise to an unconformity and to the development of an offlapping sequence basinward. In contrast, increasing tensional stresses induce widening of the basin, subsidence of the flanks and landward migration of the shoreline together with the development of an onlapping sedimentary sequence. Stress-induced vertical motions can at the same time influence sedimentation rates. For example, flanks uplift related to an increase of intra-plate compression can enhance sedimentation rates and so modify the infilling pattern, leading to the formation of unconformities (e.g. Galloway, 1989).

Following these concepts, a mechanism of compressive-induced flexure of the continental lithosphere has been suggested to explain geological structures for tectonic regional settings where classical models for basin formation have previously failed. An example of this type of structural setting is represented by the North East German Basin (NEGB).

The present-day crustal structure below the NEGB is characterized by a normal Variscan crust of almost constant 32 km of thickness along its margins which thins under the basin centre to values of about 22 km, see Figure 13 for the seismic cross section. The observed change in thickness of the crystalline crust can be attributed to the change in sedimentary thickness across the basin. However, the observed thinned crust beneath the basin centre lies over an almost perfectly flat Moho which presents a distinct flexural-bulge shape along the southern present-day basin margin. The bulge is located beneath a fault system, THF of Figure 13, which is interpreted as a thrust zone within the crust and may be related to the Late Cretaceous-Early Tertiary inversion phase. In addition, modelling of wide angle seismic data (e.g. Bayer et al., 1999; Beilecke & Rabbel, 1999) and gravimetric studies (Scheck-Wenderoth et al., 1999) have constrained high seismic P-wave velocities and densities in the lowermost crust between Grimmen and the Elbe Line. A similar feature has been proposed also further to the west on the base of refraction lines crossing the North German Basin (e.g. Thybo, 1990; Rabbel et al., 1995).

To model the observed structures beneath the NE German Basin, a pure shear model was proposed by Bachmann & Grosse (1989) and by Brink et al. (1990). However, the stretching and thinning of the crust predicts isostatic uplift of the Moho below the basin centre where the crust has thinned and the sediments reached their maximum thickness in contrast to the observed flatness of the Moho. Moreover, there is no evidence supporting the presence of a principal shear zone extending from the crust through the sub-crustal lithosphere as proposed by Brink et al. (1990).

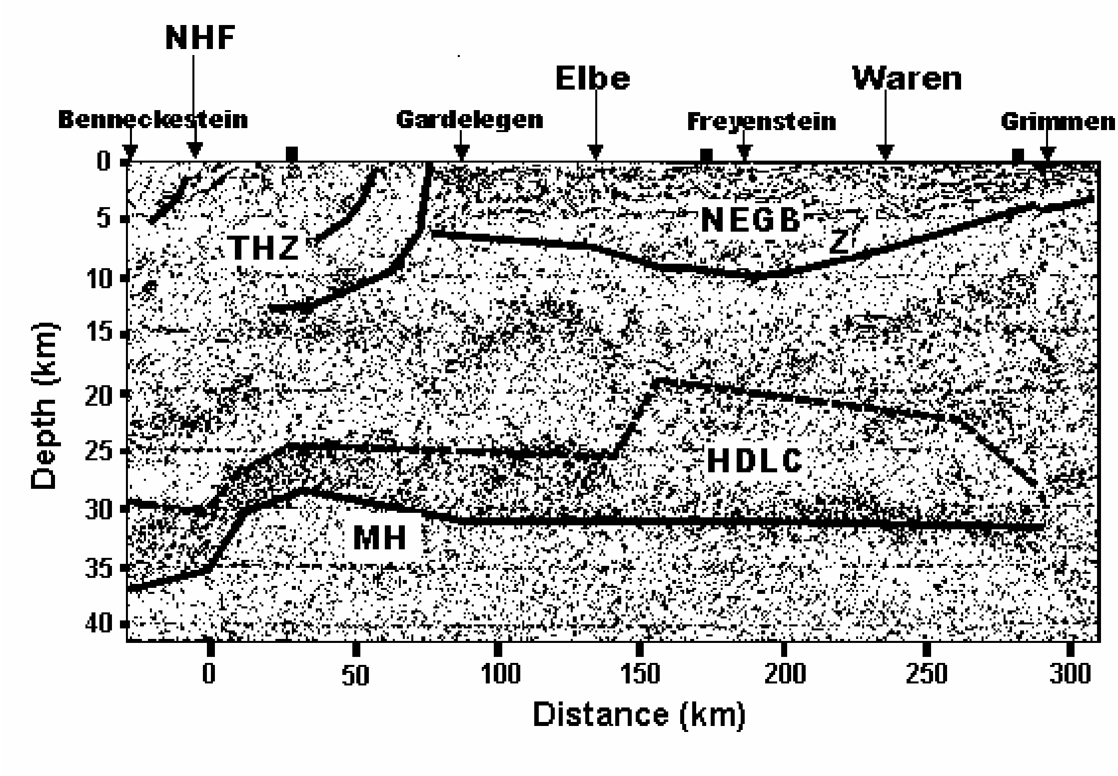

Figure 13. Deep Seismic Reflection Profile BASIN9601, illustrated by a line drawing of main reflections superimposed onto the depth converted seismic section, modified after Marotta et al. (2000). Main crustal entities are illustrated. Abbreviations: **NHF**=Northern Harz Boundary Fault; **THZ**=Thrust Zone; **MH**=Moho; **HDLC**=High Density Lower Crust; **NEGB**=North East German Basin; **Z**=base Zechstein reflector.

Marotta et al. (2000) proposed an alternative. In their study, they developed a simple 2-D thin elastic plate flexural model for the area of the NEGB. The basic assumption was that the deformation induced by Late Cretaceous-Early Tertiary inversion phase might have overprinted previous deformation patterns, comprising a thinned crust and a shallow Moho after the initial basin formation. Based on this assumption, the observed Moho updoming in the southern part of the basin was regarded as a consequence of flexural buckling of the previously thinned lithosphere, induced by the Alpine compressive stress regime. To test this hypothesis, the lithosphere was approximated as a loaded elastic beam by adopting the simple flexural model as often used for the oceanic lithosphere. The effects produced by the compressive forces were modelled as equal to a bending moment applied to one end of the beam plus a vertical load at the free end of the plate so that to represent the solution in terms of the so called *universal flexural profile* as shown in Figure 14, (see also Turcotte & Schubert, 2002). The modelling results supported the hypothesis that the present-day Moho topography may be explained as a consequence of flexural buckling of a previously thinned lithosphere under an active compressive stress field which caused contemporaneous subsidence of the NEGB during the Upper Cretaceous-Early Tertiary inversion event. In fact, for a bulge of about 33 km and height of 3 km the predicted and the observed Moho were found in very good agreement, Figure 15.

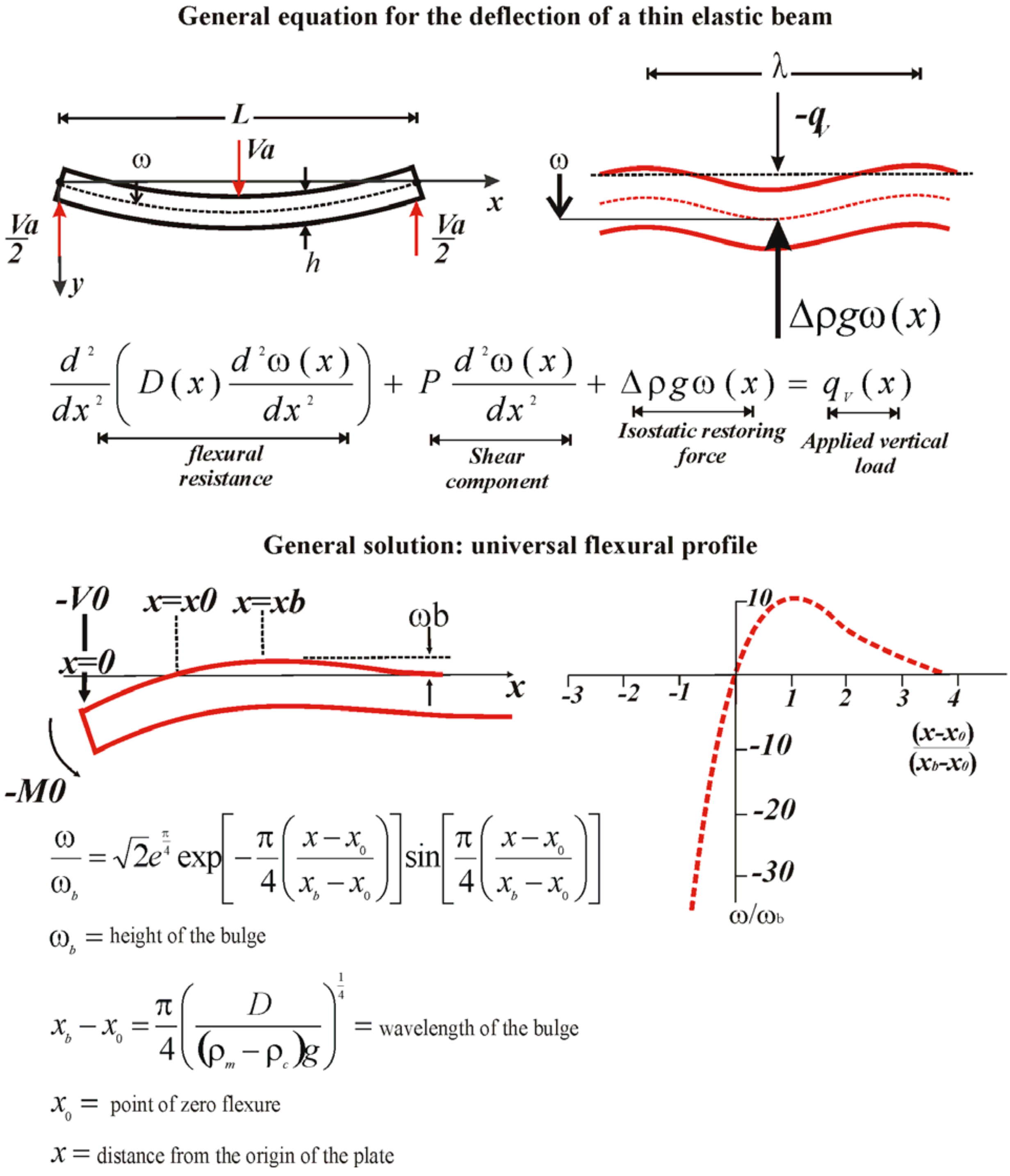

Figure 14. Two-dimensional flexure of a thin elastic beam (upper panel), and universal flexural profile general solution (lower panel).

Also the match between the predicted flexural profile and the present-day topography was satisfactorily by considering the eroded amount of sediments originally located above the actual southern margin of the basin.

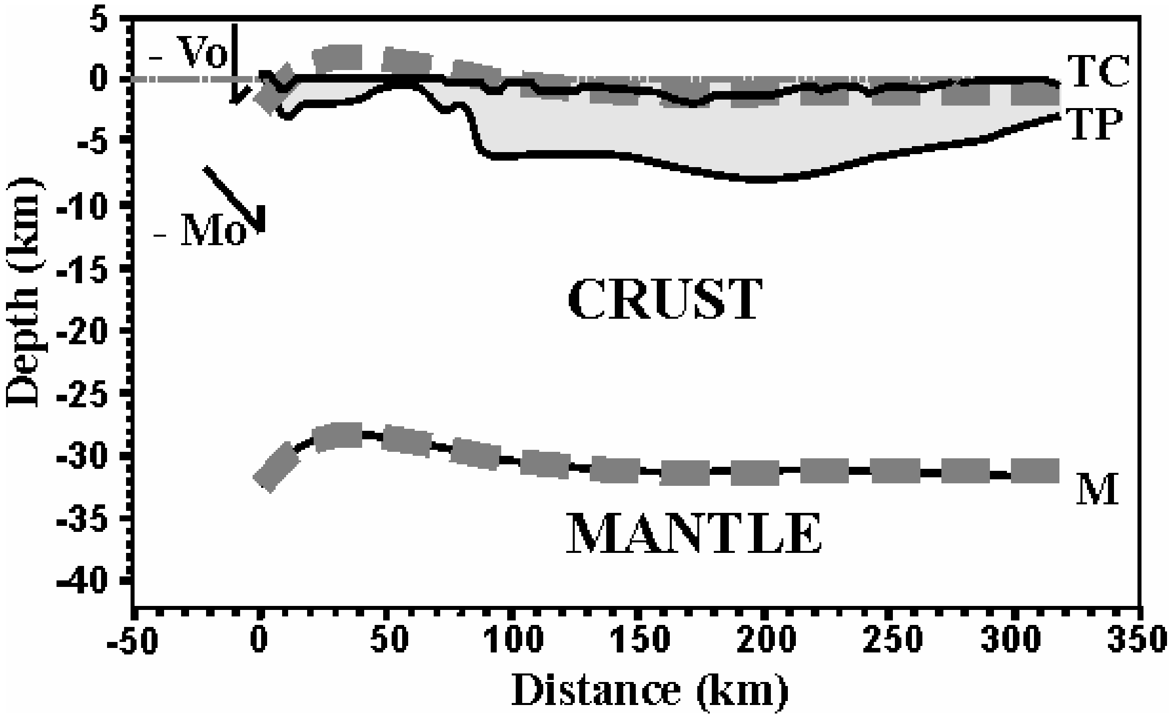

Figure 15. Predictions (thick grey dashed lines) from the flexural study made by Marotta et al. (2000) compared to the interpreted Moho (**M**) and the Top Cretaceous reflector (**TC**). Abbreviations: **TP**=Top Palaeozoic reflector; $\mathbf{V_0}$=vertical line load; $\mathbf{M_0}$=bending moment. The post-Zechstein fill is also shown (light grey colours).

Although flexural models have some limitations concerning the continental crust due to its complex rheology, they well explain the structure of subducting oceanic plates and in parts the evolution of forearc and molasses basins (e.g. Marotta et al., 2000).

Several other studies, focusing on a large number of basins in the northern Atlantic and in the Mediterranean region favour compressional stresses as driving mechanisms for the observed rapid vertical motions (e.g. Cloetingh & Kooi, 1992). As an example, Van Wees & Cloetingh (1996) applied a three dimensional flexure model incorporating lateral variations in flexural rigidity and necking depth to quantify the effects of intra-plate stresses on Quaternary accelerated subsidence and uplift in the North Sea Basin and adjacent areas. To explain the short-wavelength Quaternary depocentre in the North Sea, marked by sediment thickness up to 1000 m, they consider tectonic processes as main driving mechanism. Their results showed that a relative increase of compressive intra-plate forces allows predict the accelerated subsidence values up to 700 m, thus successfully explaining the observed Quaternary isopach values. Similarly, Horvath & Cloetingh (1996) were able to explain the overall Quaternary uplift and subsidence in the Pannonian Basin considering an increase in magnitude of horizontal compressional intra-plate stress in their two dimensional flexural model.

A more sophisticated model was proposed by Grünthal & Stromeyer (1992, 1994). In their study, they applied a 2-D, steady-state, elastic finite-element approach to model the observed direction of the maximum compressive horizontal stress (S_{Hmax}) for the brittle crustal domain within the entire western portion of the Eurasian plate. In order to model the present-day tectonic setting as well as the related style of deformation observed in Europe, Grünthal & Stromeyer (1994) coupled local variations of lithospheric stiffness with specific dynamic boundary conditions. The plate tectonic scheme adopted to simulate the main forces acting at the boundaries of the Eurasian platform comprised: (1) ridge-push forces acting along the Mid-Atlantic; (2) continental collisional forces arising from the northward-directed motions of Africa relative to Europe; and (3) push forces acting across the Anatolian and Aegean microplates. Lateral stiffness contrasts were introduced in terms of variations in rigidity values, i.e. elastic Young's modulus parameter, imposed to different lithospheric blocks as well as in terms of local extensional and shear zones. The lateral heterogeneities in the elastic properties were considered by Grünthal & Stromeyer (1994) to reproduce small scale features as the Adriatic Promontory, the Bohemian Massif, and the Pannonian Basin. To constrain the modelling results a step-wise trial-and-error approach was adopted taking as test criterion the direction of the maximum horizontal stress component (S_{Hmax}) based on stress data interpolation. The fit between the generalized trajectories and the derived S_{Hmax} direction was found remarkably. The calculated horizontal crustal stress orientation was able to resume the observed broad-scale pattern, imaging a uniform and consistent NW-SE direction of maximum compression in the western domain of the study area and a gradual bending toward NE-SW approaching the eastern part of the Eurasian plate. Moreover, the introduction of small-scale features, i.e. sub-domains with different elastic stiffness, allowed resembling also more regional features like the fan-like crustal stress pattern along the south-eastern margin of the European plate.

A similar study was proposed by Gölke & Coblentz (1996). They carried out a 2-D elastic finite-element analysis to evaluate the control exerted by the interaction between the principal sources of tectonic stresses on both the magnitude and the orientation of the first order regional stress field in Western Europe. In addition to the ridge push forces within the Mid-Atlantic and the collisional boundary forces acting along the southern border of the Eurasian plate, they included internal sources of tectonic stresses caused by lateral variations associated with continental margins and elevated continental topography. The quality of the model predictions, i.e. the predicted stress component minus the reference state was tested through a direct comparison with the stress indicator from the World Stress Map. The obtained results imaged the major features of the observed regional stress field (dominant NW-SE compression), leading to the conclusion that the nature of the intra-plate stress field within the European platform may be related to the tectonic forces acting on the plate. Moreover, the numerical results showed that forces due to lateral density contrasts associated with topographic features may induce local changes in the magnitude of the predicted regional stress, whereas they do not alter significantly its orientation.

All the previous described models were based on several not well constrained assumptions which limited their quantitative predictability only to first order features. In this regard, the main simplification was to adopt an elastic plate model. Plate interiors may retain an elastic 'core' under very low deviatoric stresses acting on relative short time scales. However, as the amount of deviatoric stresses increases, frictional or brittle failure mechanisms will be set up within the crust. Consequently, the elastic approximation prevents a quantification of the role played by the different driving mechanisms which are active during continental deformation. In this context, elasticity tends to overestimate the influence of boundary conditions and/or internal loads on the stress field and deformation pattern. At the same time, elastic models are not able to correctly account for more local features related to the internal thermal and mechanical (rheological and compositional) structure of the lithosphere. The incorporation of weak and/or stiff areas in terms of lateral variations of not-well constrained and unrealistic elastic parameters cannot help in understanding the nature, whether it is thermal, mechanical, dynamical or compositional, of the driving mechanisms of the observed local anomalies. In addition the assumption of an elastic rheology is incompatible with the occurrence of intra-plate seismicity (e.g. Cloetingh et al., 2006) and may also lead to large modelling errors (e.g. Bird, 1999).

1.3 Rheological models for basin formation and evolution

1.3.1 The role of rheology for the modes of continental deformation: limitations of a kinematic approach to continental deformation

Continental deformation is mainly determined by the rheological structure of the lithospheric plate. The rheology of the lithosphere is a function of several controlling factors: (1) its composition and structure, (2) pressure, (3) temperature, (4) principal deformation mechanism at given P-T conditions, and (5) state of stress. Variations with depth of these parameters may lead to a "sandwich-like", i.e. stratified, structure with an alternation of rheologically weak layers and relative stronger layers that can be further complicated by horizontal variations in rheology. While the pressure dependence can be neglected to the first order in the lithosphere, the temperature dependence together with composition and deformation rates are fundamental factors in determining the rheological behaviour of continental lithosphere. As derived from experiment, it is generally assumed that the deformation regime for any type of rock may be broadly subdivided into two domains: a brittle (frictional) domain within the colder and/or upper portions of the lithosphere, and a ductile domain at deeper levels and/or higher temperatures (higher than about one-half of the solidus temperature of the relevant material), see also Part 2. The distinction between brittle and ductile domains is not an easily task to achieve. As a first order approximation, the distribution of crustal and upper mantle seismicity outside subduction zones can be taken as an indication of the depth extent of brittle layers. The first-order predictions of rheological studies may be confirmed by geophysical observations. However, the distribution of deformation within the lithosphere is highly heterogeneous and this may induce large uncertainties and errors in constructing rheological profiles.

As an example of such uncertainties, the geological evolution of several continents as inferred from deep seismic studies has revealed the long-lasting heritage role of deep seated and/or pre-existing structures. In addition, the rheological structure of continental lithosphere is strongly controlled by variations in the local thermal state. Unfortunately, the thermal history and thermal structure of the continents are difficult to constrain. Moreover, the thermo-mechanical properties of the continental crust are controlled by lithologies significantly different from those present in the mantle-lithosphere. The main difference is that crustal materials generally show lower temperatures of creep activation with regards to mantle rocks. This difference may result in heterogeneous modes of deformation which may finally lead to mechanical crust-mantle decoupling. Unlike the lithospheric mantle, the continental crust is usually very variable in its lithological and mineralogical composition so that segments of continental crust of the same thickness may present different mechanical strength. In this regard, many studies (e.g. Roth & Fleckenstein, 2001; Marotta et al., 2002; Maystrenko et al. 2006) have demonstrated that the growth of salt structures as represented by salt pillows or diapirs within the sedimentary sequence strongly influences basin evolution. In fact, local salt movements may affect and even determine the stress pattern in the overburden (e.g. Bayer et al., 1999; Marotta et al., 2000; Röckel & Lempp, 2003; Lempp & Lerche, 2006).

The basic features of the described kinematic models for basin formation and evolution is that the deformation is imposed by prescribing a velocity field without incorporating any constitutive equations. Contrastingly, due to the high variability of deformation modes that can be prescribed by prior imposed velocity fields, such models have been able to describe the main features of basin evolution. As an example, the models have been very often used in passive rifting settings to account for a large variety of observations like rates of subsidence and/or uplift of the basement, differential stretching, lithospheric detachments (e.g. McKenzie, 1978a,b; Royden & Keen, 1980; Buck et al., 1988; Cloetingh et al., 1995). The major limitation related to a kinematic approach is the impossibility to establish a casual relationship between the imposed deformation pattern and the mechanical behaviour of the lithosphere. Many studies in the recent past have added more or less realistic rheological constraints within their model formulations in order to control and explain the imposed mode of deformation.

Cloetingh et al. (1995) tried to correlate the concept of the level of necking with the rheological structure of the lithosphere. Following their study, the level of necking was related to the top of the strong structural layer within the lithosphere. Unfortunately, the structural complexity of continental lithosphere strongly limits such an approach: in the presence of two different strong layers the geometrical relationship between depth of necking and rheology is no longer valid.

In a similar way, Buck (1991) developed a model to study the evolution through time of a passive rifting setting for a rheological layered lithosphere combining strength envelopes with constant strain rates and gravitational buoyancy forces. The results predicted the development of narrow rifts for an initially cold lithosphere (heat flow values $\leq$ 75 nWm^{-2}) almost independently of the applied strain rate. On the contrary, a hot un-rifted lithosphere combined with high strain rates led to wide rifts. Hot lithosphere before rifting and low strain rate values resulted in complex mixed deformation modes.

Another example is given by Marotta et al. (2000). To constrain the results obtained from the flexural model for the North East German Basin, as discussed above, they performed an independent rheological study along the seismic reflection profile shown in Figure 13. To model the rheological structure of the lithosphere under the basin, an elasto-plastic rheology at the time of loading was adopted within two different rheological models. In the first case they considered the crust composed of two structural layers (sediments and one single crystalline crustal body), while in the second model they further subdivided the crust in two different sub-layers of varying thicknesses in accordance with the seismic reflection profile of Figure 13. Brittle failure was modelled assuming plastic rock behaviour, i.e. Byerlee's law, and ductile behaviour was implemented through the use of a steady-state power law creep equation. For both the rheological modelling analogues, the calculations of rheological profiles and total elastic thickness of the lithosphere were in agreement with the results obtained by the simple elastic model. The elastic thickness predicted by the rheological study was almost identical with the one derived from the flexural elastic model confirming the presence of a weak lithosphere along the southern margin of the basin and a probable decoupling between the lower and the upper lithosphere at the Moho level as already suggested by the elastic model.

The use of rheological constraints in kinematic models has led to some improvements in understanding basin formation and evolution. However, in order to account for continental deformation dynamic modelling approaches are necessary. Only dynamical models, explicitly invoking constitutive relations can realistically relate dynamic quantities, i.e. stresses, with kinematic quantities, strain and/or strain rates.

1.3.2 Dynamic models for basin formation and evolution

Although intra-plate stresses have been shown to be very important in controlling the subsidence record as well as the stratigraphic architecture of extensional basins, most models for basins formation and evolution, e.g. simple and/or pure shear models, have usually related lithospheric strain patterns to unspecified and/or unrealistic stress fields. In a similar way, models for basins developed under compressional tectonic settings have been mainly based on lithospheric flexure profiles without addressing the role of the active compressional tectonic stress field. Additionally, results from different studies, e.g. Cloetingh et al. (2006), have demonstrated that lateral strength variations of intra-plate lithosphere are primarily caused by variations in the mechanical strength of the lithospheric mantle which are related to variations in the thermal structure of the lithosphere (e.g. Goes et al., 2000a,b). Following all these aspects, models for continental deformation should try to address both the control exerted by lithospheric stresses and the role played by variations of thermal mechanical properties of the lithosphere. To address these aspects, analytical as well as laboratory-scale models have often been used in the past (e.g. Sales, 1968; Sonder et al., 1986). However, due to the high non-

linearity of deformation mechanisms these approaches have given very few exhaustive results. To overcome these restrictions, attention focused at dynamic modelling techniques.
First attempts to simulate anelastic deformation within the lithosphere restricted the deformation to plane horizontal strains in a vertical cross section perpendicular to the strikes of linear structures, i.e. the plane strain approach. A measure of qualitative success has been obtained in modelling convergent tectonic scenarios (e.g. Tapponier & Molnar, 1977). The major limitation of their method is the restriction to processes that can be approximated as purely two dimensional. An alternative approach to model continental deformation is the "plane stress" model, now widely referred to as the thin-sheet or thin-plate model (e.g. Bird & Piper, 1980; England & McKenzie, 1982, 1983). In this technique, deviatoric stresses are assumed to vanish beneath the lithosphere, whereas within the lithosphere the differential equations for the equilibrium of stresses are vertically integrated following the isostatic approximation, i.e. lithostatic vertical stress condition. The equations of conservation of momentum and mass are numerically solved, within the finite difference and/or finite element techniques, in order to predict (horizontal) deformation velocities, stresses, and therefore strain and strain rates patterns. Once the dynamic quantity, i.e. stress, is obtained it is possible to derive the kinematic quantity, i.e. strain. The quality of numerical models then can be tested and quantified by comparing the obtained results with observations. Many data sets exist differing in spatial and temporal resolution as well as coverage. A control data set, thereby, should at least fulfil the following criteria:

- It should cover the entire modelled area.
- Its spatial resolution should reflect the spatial resolution of the model.
- The observed data should be predictable by the numerical model.
- The observed data should be valid when extrapolated to longer time spans.

The thin-sheet approach has been widely used by several research groups which have applied it to a variety of tectonic scenarios. A non-exhaustive list comprises: Vilotte et al. (1982, 1984, 1985), England & McKenzie (1983), England & Houseman (1985, 1989), Houseman & England (1986a,b), Sonder & England (1989), Bird (1989). Although an accurate and more realistic representation of lithospheric strength should involve a fully three dimensional integral model, the thin-sheet approach can be regarded as a good and reasonable approximation to continental deformation. At the same time, it requires less computational costs and elaboration times with respect to three dimensional integral models. Previous thin-sheet models accounted only for mechanical deformation (e.g. England & McKenzie, 1983). Further implementations included temperature-dependent creep parameters as well as the possibility to address more complex processes like kinematic detachments within the crust and the mantle-lithosphere (Bird, 1989; Sonder & England, 1989). In the meantime, several numerical studies also started to address lithospheric deformation as a function of the internal structure and composition of the lithosphere. Within this context, many groups of modellers have tried to represent lithospheric strength by use of a single vertical averaged power law rheology (e.g. Vilotte et al., 1985; England & McKenzie, 1983; England & Houseman, 1985; Houseman & England, 1986a,b). They have found that for a typical two-layered lithosphere, crust and mantle, the vertical averaged stress-to-strain rate relationship can be approximated over a wide range of strain rate values (10^{-13}-10^{-17} s^{-1}) by a power law rheology which accounts for more than one deformation mechanism. Following this approach, only two parameters are necessary to describe the mechanical behaviour of continental lithosphere: the stress exponent and the stress coefficient. The simplicity of this approach is limited due to the strong dependence of these two parameters on the geothermal field in the uppermost mantle as well as on the state of stress within the upper crust (e.g. Houseman & England, 1986b; Sonder & England, 1986).

In this context, England (1983) applied a thin-sheet viscous model for continental lithosphere in which extension occurred homogeneously. The main feature then was that the force required to deform the lithosphere at a given strain rate has an inverse proportional relation with the geothermal gradient and depends exponentially on the temperature of the strength controlling layer, the mantle portion of the lithosphere in his study. More precisely, he demonstrated that in the early rifting phase the force required to deform the lithosphere decreases until strain reaches a critical level whose value depends on the rheological structure of the lithosphere, identified by the "*Peclet number*", and on the thickness of the un-rifted crust. After having reached this critical value, the strength of the lithosphere starts to increase rapidly preventing further deformation. Although the model of England (1983) is in agreement with constraints derived from the continental margin of the North Atlantic, it is based on several assumptions with regard to the mechanical behaviour of the lithosphere which prevent its application in complex basinal areas. England's (1983) model does not consider the role of crustal rocks or changes in the thermal gradient within the lithosphere, and therefore tends to overestimate lithospheric strengthening.

A different attempt to model the rheological behaviour of the lithosphere has been the introduction of a Newtonian (i.e. linear) viscosity, e.g. Richardson & Cox (1984). However, there is no laboratory evidence for linear viscosity for earth materials to support this assumption.

Others authors in the past have used a more complex combination of plastic and viscous rheologies (e.g. Villotte et al., 1982; Bird, 1999). Such an approach reflects the concept that friction on faults and plastic failure in the upper portions of the lithosphere may control the vertical averaged rheology of the entire continental lithosphere. The main inconvenience is to determine a single vertical averaged rheology on the base of the distribution of deformation mechanisms with depth which requires a very accurate knowledge of the thermo-mechanical structure of the lithosphere. Rheological stratification of continental lithosphere and variations in the local geothermal field may account for the observed heterogeneous style of continental deformation (e.g. Behn et al., 2002). Following these principles many authors in the past have tried to address the occurrence of weak and/or stiff domains in terms of lateral variations of the thermal regime (e.g. Molnar & Tapponier, 1981; England & Houseman, 1989; Tommasi et al., 1995). However, the lack of a full thermo-mechanical formulation has strongly limited the capability of understanding the retro-effects between deformation and temperature variations. More recent versions have been implemented where lateral variations of lithospheric stiffness have been coupled with specific boundary conditions in order to model the observed style of deformation in both collisional and extensional tectonic settings (e.g. England & Houseman, 1985; Vilotte et al., 1985; Lynch & Morgan, 1990; Tommasi et al., 1995; Marotta et al., 2001; Jimenez-Munt et al., 2001).

In this context, Marotta et al. (2004) applied a spherical viscous thin-sheet model to investigate the pattern of crustal deformation within Central Europe as a function of tectonic loading forces and changes in the internal structure of the lithosphere. They modelled the main tectonic forces acting across the boundaries of the European plate (i.e. Mid-Atlantic Ridge push forces, collisional forces related to the collision between Africa and Europe). Lateral variations in lithospheric stiffness were imposed assigning different viscosity values to the domains in which the entire study area was subdivided, see Figure 16. To quantify the sensitivity of the modelling results with regard to changes of the imposed boundary loads as well as to lateral variations in lithospheric stiffness, the predictions derived from the tectonic model were compared first to the deformation pattern computed for the same region using a spherically symmetric, self-gravitating, visco-elastic (Maxwell) Earth model of Glacial Isostatic Adjustment (GIA) and finally to observed baseline rates (i.e. length changes) relative to three different fixed sites within the permanent ITRF200 and BIFROST GPS networks. The reference sites (Potsdam in Germany, Vaas in Finland and Onsala in Sweden) were chosen at locations with different latitudes in order to resemble the different levels of deformation associated with tectonics and GIA.

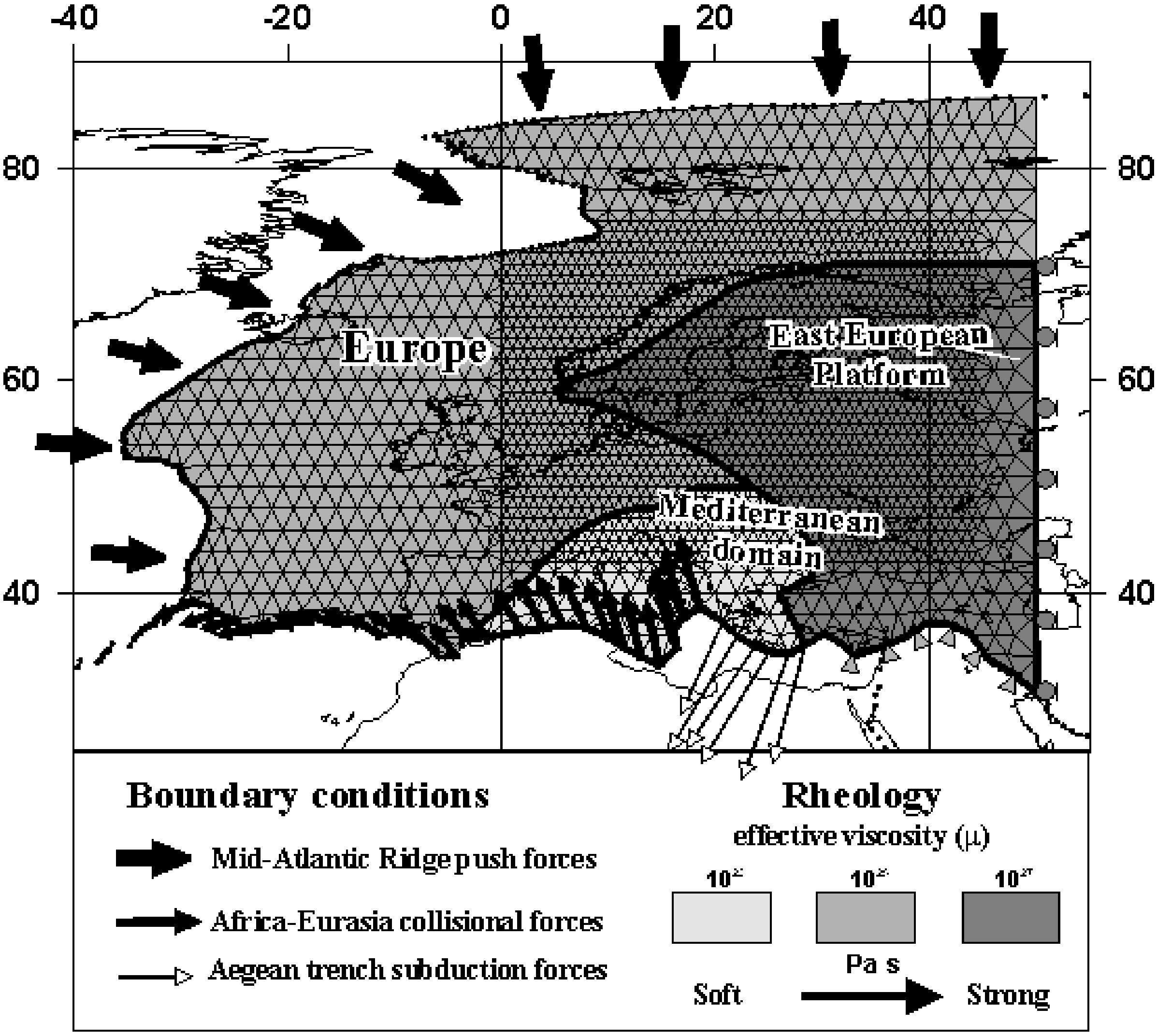

Figure 16. Finite element grid and net boundary conditions adopted for the tectonic predictions in the modelling study of Marotta et al. (2004). Black thick arrows along the western and northern plate boundaries represent push forces from the Middle and North Atlantic. The counter-clockwise rotation of the African plate with respect the Eurasian plate is reflected by the thin black arrows along the south-western boundary. Thin white arrows along the south-eastern border represent the horizontal components of the Aegean trench subduction forces (McClusky et al., 2000). The southern border between the model domain and the Arabian region is held fixed (light grey triangles), whereas the eastern boundary is assumed shear stress free (grey circles). The grid finally distinguished three iso-viscous domains: (1) East European Platform, (2) European platform, and (3) Mediterranean domain. Different colours indicate different values of lithospheric strength in accordance with the colour legend at the bottom of the figure.

The model demonstrated that tectonic deformation associated with boundary stress forces (i.e. Middle Atlantic Ridge push forces and African to Europe collision-related forces), GIA effects, and lateral variations in the strength of the lithosphere must be taken simultaneously into account to reconcile the broad style of intra-plate deformation in Europe. In fact, with reference to the ITRF2000 velocity solution database, the best fit with observations was achieved combining both, the tectonic predictions comprising a weakened lithosphere in the Mediterranean sub-domain, as shown in Figure 16, together with the standard GIA predictions. Moreover, the adopted multi-step approach enabled Marotta et al. (2004) to discriminate the geometric impact that each of the above described signals has on the European region. In this context, they found that GIA dominates the deformation at northern latitudes (north of 58°-60° N latitude), Africa-Eurasia collision boundary forces has the strongest impact on baseline rates within the southern part of Europe (toward the Alpine Front and the Mediterranean region), and that lateral variations in lithospheric strength strongly moderate deformation patterns associated with tectonic forces.

The above described study is a regional analysis concerning the broad scale stress and/or strain field. It covers a very wide region extending from the Mid-Atlantic Ridge to about 40° E and from the natural plate boundary between the African and the Eurasian plate to about 90° N. The great extension of the study area allows predicting possible changes in the stress field and in lithospheric deformation style within a minimum wavelength of a few hundreds of kilometres. Consequently, when the modelling results are compared with available datasets (e.g. the World Stress Map database and/or GPS detected crustal velocities) a satisfactory agreement is achieved only for first order features. Many other large-scale models have been carried out in the past (e.g. Grünthal & Stromeyer, 1994; Richardson & Coblentz, 1994; Gölke & Coblentz, 1996). The main reason was the lack of detailed information concerning the structure of the lithosphere. This lack of knowledge has led to adopt unrealistic simple rheologies thus refraining from a full examination of the dynamical aspects of continental deformation preventing at the same time basin-wide models.

However, during the last decade substantial progress has been made in terms of availability of new and high-quality data sets both from long time investigated areas and from new unexplored areas. These new information have shed new insights on the subsidence and uplift basins history, the temporal and spatial evolution of the stress and strain fields, the thermal history and structural development for several basin systems. Based on this progress, new dynamic models have been developed which successfully incorporate realistic constitutive equations in the lithosphere. This allows to simulate more complicated rheologies including structural and/or thermal in-homogeneities of earth’s materials as well as to test their role on basin development under different tectonic loading boundary conditions.

1.4 Summary

Part 1 provides a short overview of different modelling approaches to continental deformation using the Central European Basin System as a guiding example where most of the known models have been applied.

The chapter starts with a description of the “classical” kinematic models developed during the 1980^{th} (e.g. McKenzie’s uniform stretching and its further implementations). It is shown how these models may account for some simple and/or young basins. Unfortunately, they can account only for special aspects of the historical development in complex sedimentary basins. The reason is that these models rely on extreme simplifications concerning the lithospheric structure, temperature distribution with depth, and tectonic scenarios. In this sense, the major limits of the McKenzie’s model are related to its strictly one-dimensional approach as well as to the simplifying assumptions concerning the temperature distribution in the lithosphere and the rheology of the earth. Consequently, the McKenzie’s model remains a first-order approximation with strong limits even for simple basins. In a similar way, the simple shear model of Wernicke has provided just a reasonable concept lacking however hard data to support it. The use of buckling models with regards to the continental crust, originally developed for the oceanic crust, are limited due to the complex rheology of the continental crust which can support elastic behaviour only to a minor degree. Nevertheless, though simple, all these models have helped to understand specific phases in the evolution of even complex basins.

Further on in the chapter, the more advanced models like the thin sheet or thin plate models are described. These models provide insight concerning the interaction between far field boundary stresses or strains and local variations in material parameters. However, there are still unresolved geological problems concerning the evolution and role of major geological discontinuities like faults and/or variations in crustal composition. The existence of sometimes more than one decoupling horizon within the lithosphere, with salt being a dominating one, provides a critical point to integrate over the entire lithosphere in order to reduce the problem dimensions to fit computational capabilities. Another major problem is the generally poor knowledge of the deeper crust especially in basins characterized by a thick sedimentary cover. This lack of knowledge has often led to unrealistic simple rheologies preventing a full examination of the dynamical aspects of continental deformation. However, during the last decade substantial progress has been made in terms of new and high-quality geophysical data. This new information has provided insights in the subsidence and uplift

history, the temporal and spatial evolution of the stress and strain fields, the thermal history and structural development. Based on this progress and the development of computer power, new numerical models have been developed which successfully incorporate realistic constitutive equations and boundary conditions. These implementations allow to simulate more complex rheologies including structural and/or thermal heterogeneities as well as to test their role on basin development under different tectonic loading and boundary conditions as described in the final part of the chapter.
Concluding, all the numerical models presented in Part 1 may be regarded as useful tools to understand the processes characterizing the development of complex basins. However, their results should be considered within the constraints given by the assumptions adopted in their theoretical formulations and the limits related to uncertainties concerning details of geological data.