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Role of pre-rift rheology in kinematics of extensional basin formation: constraints from thermomechanical models of Mediterranean and intracratonic basins

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The role of pre-rift rheology on the kinematics of extensional basin formation is examined. Constraints obtained on the effective elastic thickness and level of necking, inferred from forward modelling of a number of Alpine/Mediterranean basins (including the Gulf of Lion margin, the Valencia Trough, the Tyrrhenian Sea and the Pannonian Basin), are interpreted in terms of the pre-rift rheology of the lithosphere underlying these basins. The Gulf of Lion/Tyrrhenian Sea basins and the Pannonian Basin appear to be end-members in terms of the inferred depth levels of necking. The models support the existence of spatial variations in crustal and lithospheric strength, as inferred from previous rheological modelling for other segments of the European lithosphere, and provide constraints on the ratio of crustal and subcrustal strength during extension. The results of these studies are compared with predictions on the kinematics of extension for a number of intracratonic basins, including the Black Sea basins, the Transantarctic Mountains/Ross Sea and Saudi Arabian Red Sea margins, and the Baikal and East African rifts. The kinematics of extension appears to be largely controlled by the (transient) thermal regime of the pre-rift lithosphere and the crustal thickness distribution. These usually result from orogenic processes operating on the lithosphere before extensional basin formation. Predictions are made for the level of external forces required to initiate rifting in intracratonic and Alpine/Mediterranean settings. The models also shed light on the relative parts played by far-field versus near-field stresses and inferred variations in strain rate during the evolution of these basins.

Keywords: basin modelling; rheology; forward modelling; extensional basins

The dynamics of extension form an important component in models developed for the evolution of the lithosphere and a better understanding of the processes controlling rifting and basin formation is required to address some of the questions raised (Cloetingh *et al.*, 1993; 1994a; 1994b; Ziegler, in press). A question that has been addressed by a number of previous studies (e.g. Dewey, 1988; Seranne *et al.*, this issue) is whether extensional basin formation in convergent regimes is a process fundamentally different from 'normal' intracratonic extension. The mechanisms underlying extension in convergent regimes have also been investigated by a large number of studies of orogenic belts of various geological ages (e.g. Chauvet and Seranne, 1994; Seranne and Malavieille, 1994; Tricart *et al.*, 1994). The Mediterranean basins and the Pannonian Basin (Horváth, 1993) form the type localities to study the sedimentary record and crustal structure of extensional basin formation in convergent regimes. Analogues to back-arc basin formation and the association with the presence of subducting slabs have played a dominant part in discussions on the origin of these basins (e.g. Horváth and Berckhemer, 1982; Royden, 1993). Other studies have further addressed evidence for the presence of retreating slabs in the convergence zones of the Mediterranean (Wortel and Spakman, 1992; Royden, 1993). This is of particular importance in view of the possible contribution provided by slab pull and body forces (England and Houseman, 1989) to stress fields in the areas (e.g. Bassi and Sabadini, 1994).

A new database is now available as a result of intensive geophysical and geological studies carried out during the last decade, which have significantly enhanced our current understanding of the basin formation mechanisms operating in these areas. These data allow us to put the question about whether extension in a regime of overall convergence is fundamentally different from intracratonic extension in a quantitative framework. Model studies that use these

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data as constraints enable us to test the assumptions that have been made on the role of pre-rift orogenic events and rheological consequences. The realization that the lithosphere retains finite strength during extension has led to the coupling of pre-rift strength to the process of lithospheric necking (Braun and Beaumont, 1989; Kooi *et al.*, 1992). Still, important questions remain to be solved on the role of detachments (Le Pichon and Chamot-Rooke, 1991) and the dynamic significance of the (kinematic) necking levels (e.g. Weissel and Karner, 1989 versus the Kooi *et al.*, 1992 approach). The same is true for the connections proposed between the bulk strength of the lithosphere and the width of rift zones (Allemand and Brun, 1991; Buck, 1991).

Dynamic models offer a useful framework for a better understanding of the bulk rheology of the lithosphere (e.g. Dunbar and Sawyer, 1989; Buck, 1991; Bassi *et al.*, 1993; Govers and Wortel, 1993). These models have addressed the role of strain rate variations and the rheological properties of the lithosphere. The role of the initial pre-rift lithospheric configuration (Bassi *et al.*, 1993) appears vital in this respect. Questions relating to the relative contributions made by pure shear/simple shear, the role of the crust in the continuum sense or the influence of faults (Boutilier and Keen, 1994) are open to exploration by these models.

In this paper we focus on the relative importance of lithospheric parameters on extensional basin formation in terms of pre-rift rheology, strain rate history and the forces driving extension. To this aim we have selected 12 basins formed in the Africa/Arabia–Eurasia plate collision zone and in intracratonic settings (see *Figure 1* for location) for a closer investigation of basin formation parameters derived from kinematic modelling.

Rheological aspects

Rheology strongly affects the dynamics of basin formation (Buck, 1991; Vilotte *et al.*, 1993; Bassi *et al.*, 1993; Bassi, 1995). The effects of different initial crustal thicknesses and the thermal state of the pre-rift lithosphere determine the effectivity of far-field and near-field stresses and associated strain rates in the process of extensional basin formation. Here we will focus on the role of the pre-rift rheology, constrained by the results of experimental rock mechanics data adopting the yield strength envelope concept (Carter and Tsenn, 1987; Cloetingh and Banda, 1992; Ranalli, 1994; Burov and Diament, 1995; Cloetingh and Burov, in press). These strength profiles provide a

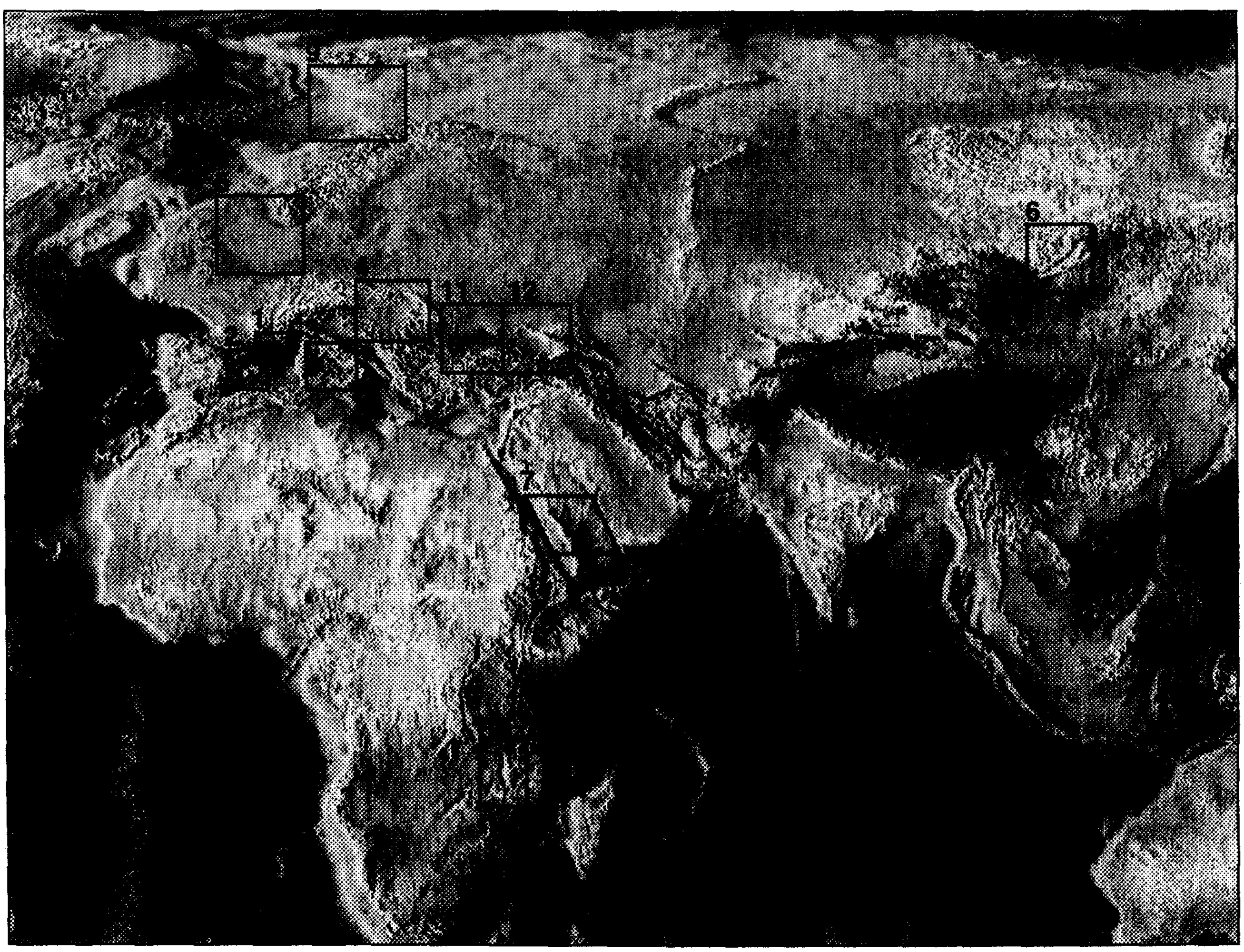

Figure 1 Location of basins discussed in this paper. Different numbers refer to the following basins: 1, Gulf of Lion; 2, Valencia Trough; 3, Southern Tyrrhenian Sea; 4, Pannonian Basin; 5, North Sea Basin; 6, Baikal Rift; 7, Saudi Arabia Red Sea Margin; 8, Transantarctic Mountains; 9, Barents Sea margin; 10, East African Rift; 11, Western Black Sea; and 12, Eastern Black Sea

qualitative picture of the distribution of lithospheric strength with depth and give a measure for the maximum stress levels to be supported by the lithosphere. Composition and temperature are controlling factors on bulk lithospheric strength (Kusznir and Park, 1987), also reflected in indirect observables such as estimates of effective elastic thickness (EET) (Burov and Diament, 1995) and the distribution of intraplate seismicity (e.g. Cloetingh and Banda, 1992). *Figure 2* displays strength profiles constructed for 'normal' average lithosphere, recently thickened Alpine lithosphere and intracratonic 'cool' lithosphere. As illustrated by this figure, important differences can be expected in the starting conditions for rifting in relatively warm, thickened Alpine lithosphere, as opposed to intracratonic lithosphere with a cool thermal regime.

Kinematic models (e.g. Weissel and Karner, 1989; Kooi *et al.*, 1992) and their validation through testing with data sets from rifted basins can yield constraints on dynamic basin formation models, providing a robust framework for quantitative analysis of basin stratigraphy and associated vertical motions (Kooi *et al.*, 1992; van Balen and Cloetingh, 1994; van der Beek *et al.*, 1994).

The notion of a finite strength of the lithosphere during extension and its formulation in terms of a level of necking (Braun and Beaumont, 1989; Weissel and Karner, 1989; Kooi *et al.*, 1992) has drawn attention to how basin formation mechanics affect rift shoulder development and basin-fill. The level of necking has been defined as the level which, in the absence of isostatic forces, would remain horizontal during extension. As a result of isostasy, a deep level of necking will result in upward flexural rebound on rifting and pronounced rift shoulder uplift, whereas a shallow level of necking will reduce the development of rift shoulder topography, converging ultimately to the predictions of the flexural loading model of a lithosphere without strength during extension (Braun and Beaumont, 1989; see *Figure 3*). For stratigraphic modelling, both on a basin-wide and a sub-basin scale, the correct assessment of the actual depth of necking is important. This implies the need for a better understanding of the factors controlling the level of necking in connection with our current understanding of lithospheric rheology.

Kooi *et al.* (1992) connected the level of necking with the zone of maximum lithospheric strength, located at the depth of the brittle–ductile transition. In this conceptual framework cratonic lithosphere, characterized by a cool geotherm and a strong subcrustal mantle, would be associated by a deep level of necking centred around a subcrustal level (see *Figure 3*). Alpine orogenic lithosphere would, by its association with thick crust and a high temperature regime, qualify for a low strength lithosphere and a level of necking at shallow depths (see *Figure 3*). In subsequent work (e.g. van der Beek *et al.*, 1995; Spadini *et al.*, 1995) we have realized that these relationships might be more complex, particularly in view of the part played by necking with depths for intra-lithospheric detachment (van der Beek *et al.*, 1995).

Dynamic models indicate that detachment of crust and mantle may result in intracrustal necking levels (Dunbar and Sawyer, 1989; Bassi *et al.*, 1993; Boutilier and Keen, 1994). In addition, questions have arisen about the mantle part of the inferred strength distributions, suggesting that the mantle strengths are possibly overestimated. For example, intraplate seismicity distributions do not show earthquake foci at the levels of the subcrustal mantle lithosphere, where rheological profiles predict considerable strength (Cloetingh and Banda, 1992). Furthermore, depending on the actual crustal configuration, more than one strong layer might be present in the lithosphere, yielding a situation with several stress guides in the lithosphere which is subsequently described by one parameter; the effective level of necking would then coincide with the weak layer in between the strong layers (Spadini *et al.*, 1995).

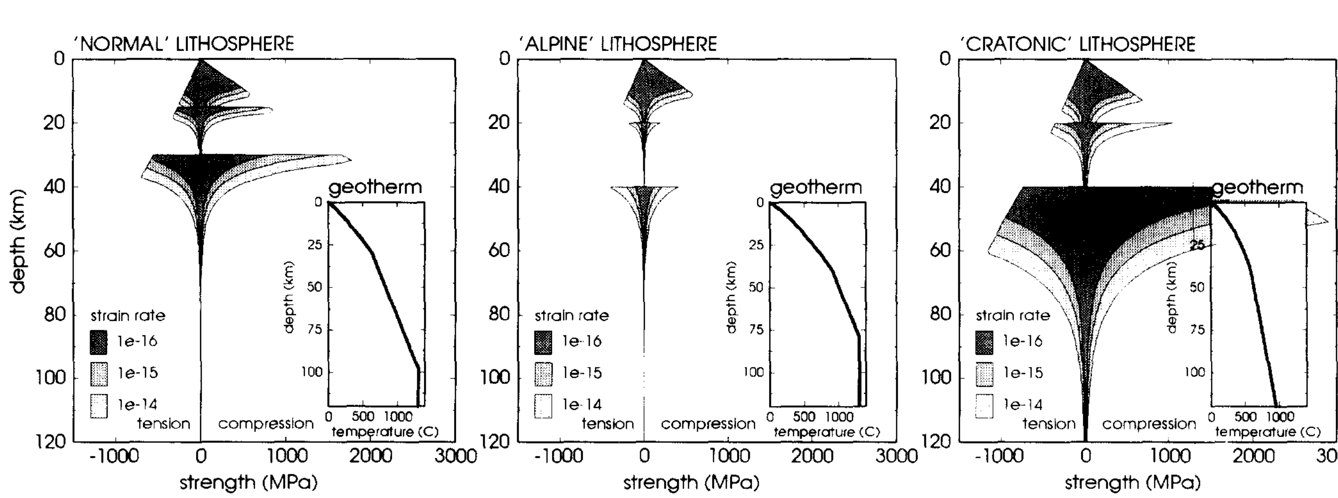

Figure 2 Strength profiles, based on extrapolation of rock mechanics data constructed for idealized petrological models of continental lithosphere, adopting the temperature distributions given in the inset of the figure. The rheology is based on a two-layered crust with an upper crust of quartzite composition and a lower crust made up of diorite, underlain by an olivine upper mantle lithosphere. Different shadings are for different strain rates. Left: strength profile calculated for 'normal' lithosphere with a crustal thickness of 30 km. Centre: strength profile calculated for typical Alpine lithosphere with a crustal thickness of 40 km. Right: strength profile constructed for cratonic lithosphere, adopting a crustal thickness of 40 km and low geothermal gradient

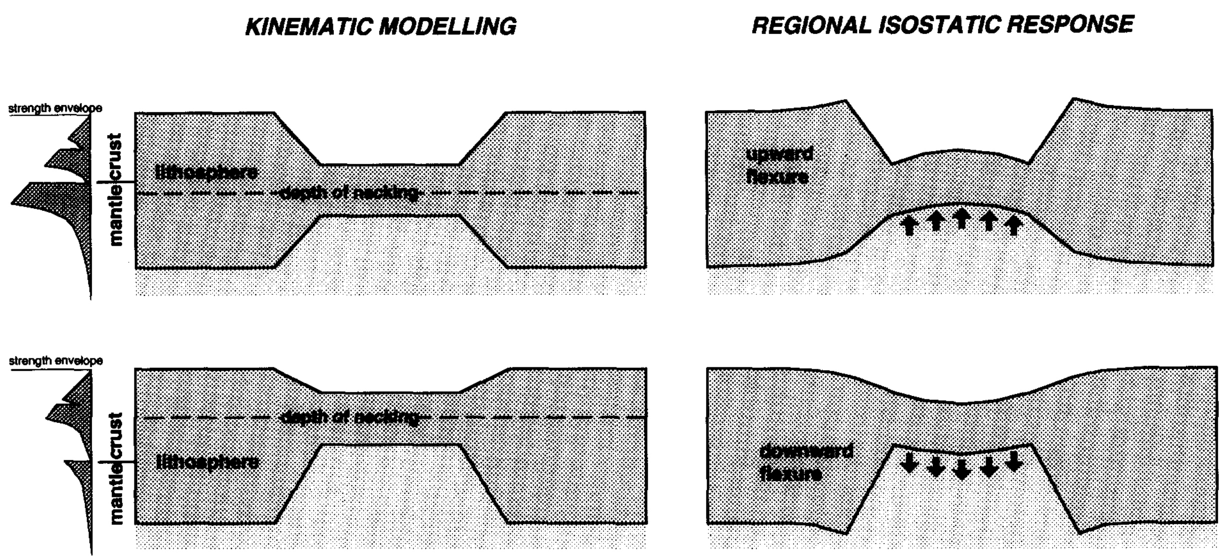

Figure 3 Schematic illustration of the concept of necking of lithosphere with a finite flexural rigidity during extensional basin formation. Panels show the investigated relationships between pre-rift strength distribution, level of necking incorporated into kinematic modelling and the regional isostatic response of the lithosphere to extension. Upper panel, extensional basin formation in cratonic lithosphere with a concentration of lithospheric strength in the subcrustal (mantle) part of the lithosphere, leading to a deep level of necking. An upward state of flexure is induced, resulting in the development of flexurally supported rift shoulders. Lower panel, extensional basin formation in Alpine thickened lithosphere, characterized by a relatively shallow level of necking. A downward state of flexure is induced, resulting in flexurally downwarped rift flanks. Modified after Braun and Beaumont (1989)

Modelling approach

From the preceding discussion it follows that the conceptual advances made in the last few years and their subsequent incorporation into dynamic models and kinematic approaches, for individual basins, have left open a large uncertainty range on pre-rift lithosphere conditions and their effects on pertinent basin formation parameters, such as EET, necking depth, lithospheric strength and strain rate. We now present inferences from forward basin modelling of 12 well constrained basin history data sets carried out during the last two years by the Amsterdam Tectonics Group. The systematic use of the same basin modelling methodology allows an unbiased quantitative comparative analysis of the basin formation parameters. Subsurface structure as imaged by seismic, gravity and borehole data provide the constraints on forward basin modelling and predictions for vertical motions for subsequent comparison with stratigraphic data, back-stripped tectonic subsidence curves and fission track data.

Figure 4 summarizes the flow chart of the modelling strategy of the individual basins, for the analysis of the relation between pre-rift thermo-mechanical properties of the lithosphere and pertinent basin formation parameters. Models obtaining an adequate fit to the basin-fill data and consistent with basement and Moho topography and gravity data generate predictions for EET and the level of necking. The interplay of strength and stress in the extending lithosphere is examined using estimates for necking depths and inferences from rheological modelling for a tectonic scenario adopting a commonly observed sequence of pre-rift collisional thickening and subsequent extensional basin formation. This procedure allows us to quantify the role of pre-extensional transient thermo-mechanical history and to obtain estimates for the level of body forces and regional stresses required to cause extension. *Figure 5* illustrates the individual modelling steps taken in the analysis applied to a synthetic example.

The availability of seismic refraction and reflection studies and gravity data has allowed us to carry out well constrained crustal-scale modelling, in particular for the basins examined here. As an example of the

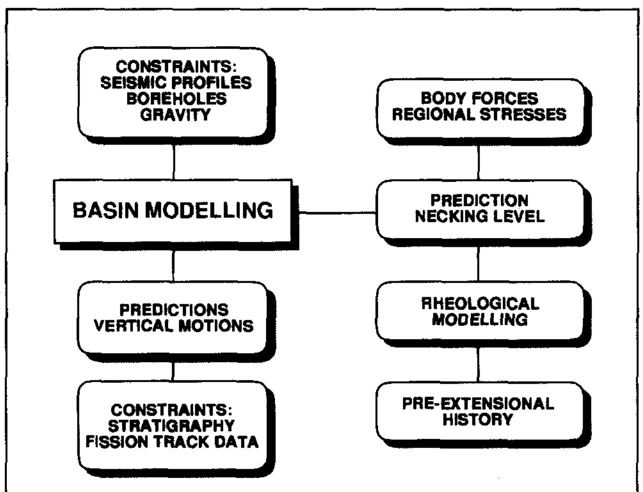

Figure 4 Flow chart summarizing the sequence of steps followed in the modelling procedure applied to the 12 basins examined in this paper

application of our methodology to a real data set, we show in *Figure 6* a modelled crustal cross-section through the southern part of the Tyrrhenian Sea. For this young rifted basin, the available crustal structure and gravity data (*Figure 6*) enable us to discriminate between the effects of different modes of extension. As demonstrated by Spadini *et al.* (1995), models invoking local isostasy or a shallow level of necking generate predictions for basement topography which are incompatible with the data. As shown by *Figure 6*, the basement topography data have the resolving power to discriminate between various assumptions on the depth levels of necking, showing an apparently better fit for a deep level of necking of 25 km. A level of necking can be inferred within a depth range of approximately 5 km on average. Predicted EETs are less accurate (approximately within 10 km). For low EETs the accuracy of necking level predictions becomes less, as the basin is closer to local isostatic equilibrium. *Figure 7a* compares the observed gravity and predictions inferred from the modelling for different levels of necking, showing a best fit for a level of necking of 25 km. The shape of the predicted topography is primarily controlled by the adopted level of necking, whereas variations in EET appear primarily to affect the overall level of the basement topography and Moho shape. The incorporation of these parameters in forward stratigraphic modelling (*Figure 7b*) allows their subsequent testing by back-stripped subsidence data and basin architecture (see Spadini *et al.*, 1995 for a detailed discussion of the Tyrrhenian Sea).

Alpine/Mediterranean basins

The formation of the Alpine/Mediterranean basins occurred in a convergent setting in close association with the evolution of the Alpine mountain belts. As indicated by the interpretation of seismic data, the formatin of the Gulf of Lion was genetically linked to the seaward continuation of the Pyrenean mountain belt (Seranne *et al.*, this issue). In addition to the inferred coupling of various stages in their tectonic evolution, a spatial flexural coupling appears to exist between extensional basin formation in the offshore domains and adjacent onshore foreland fold and thrust belts. Such a coupling is supported by data collected in the northern Tyrrhenian Sea/northern Apennines (Keller *et al.*, 1994), Valencia Trough/southern Pyrenean foreland basin (Zoetemeijer *et al.*, 1990; Verges *et al.*, this issue), the Alborán Sea/Betic System (Cloetingh *et al.*, 1992; Docherty and Banda, 1995) and the Pannonian/Carpathian system (Horváth, 1993). Kinematic modelling studies of the extensional basins in the Gulf of Lion margin (Kooi *et al.*, 1992), the Valencia Trough (Janssen *et al.*, 1993) and the southern part of the Tyrrhenian Sea (Spadini *et al.*, 1995) have provided quantitative estimates for necking depths and other parameters on the pre-rift lithosphere of the western Mediterranean basins. The Pannonian Basin (Horváth, 1993; van Balen and Cloetingh, 1995), has been modelled using constraints from a large data set of wells and seismic profiles acquired through petroleum exploration in this basin.

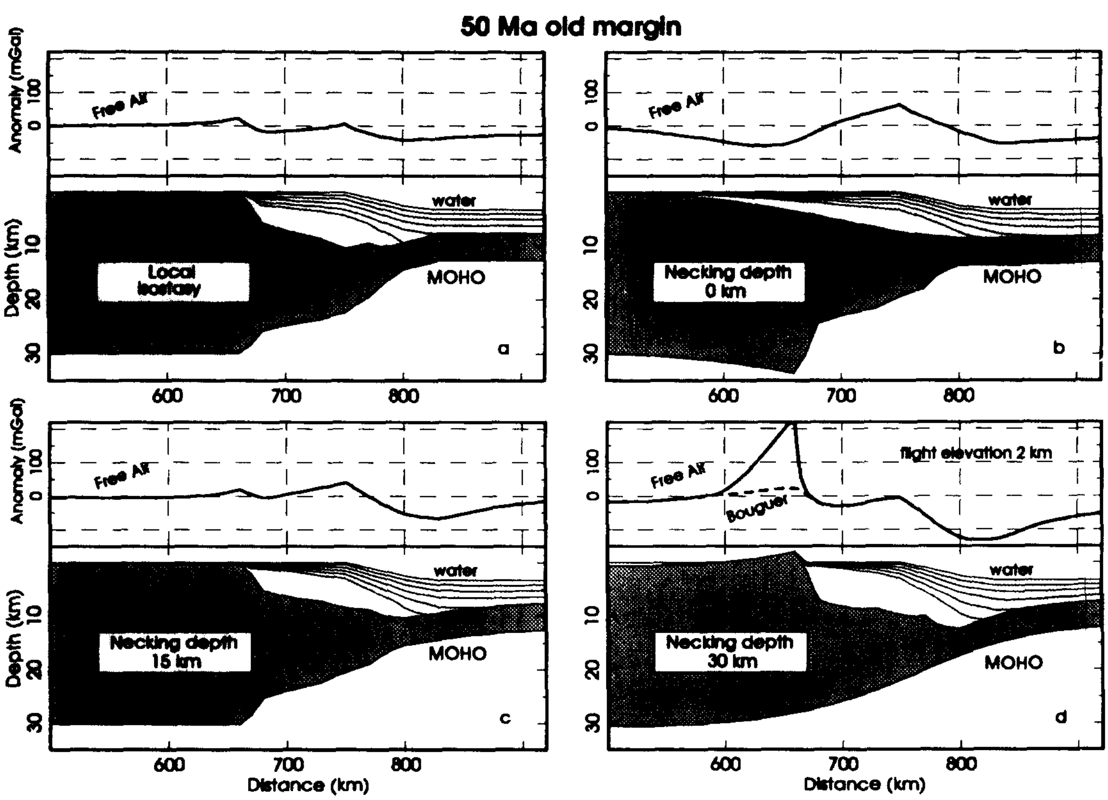

Figure 5 Application of the approach summarized in *Figure 4* to a hypothetical 50 Ma old rifted margin basin. Modelling predictions for crustal structure, gravity and overall basin stratigraphy are displayed for a local isostatic model and models with different levels of necking (0, 15, 30 km), adopting the same stretching model. Different levels of necking strongly affect predictions for Moho and basement topography, rift shoulder development, thickness ratios of syn-rift and post-rift sediments and free air anomalies (after Kooi *et al.*, 1992)

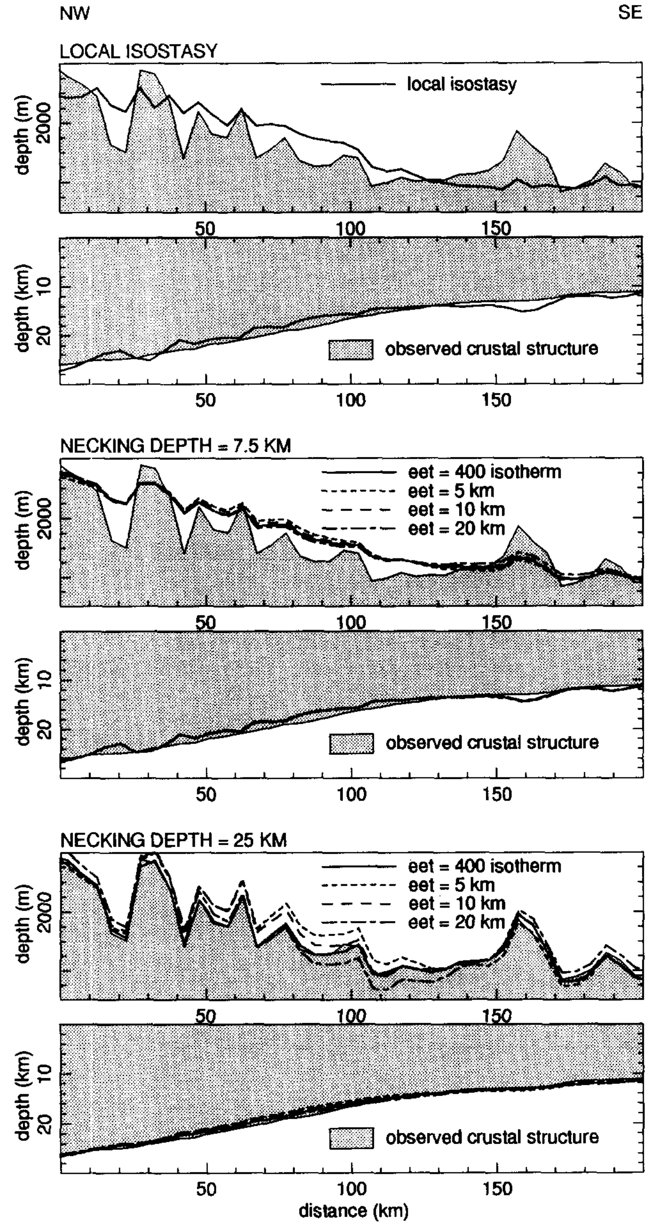

Figure 6 Illustration of application of modelling procedure for the simulation of basin history using real data. Constraints on the level of necking are provided by a case study of the Sardinian margin of the southern Tyrrhenian Sea (after Spadini *et al.*, 1995). Basin topography allows testing of various levels of necking, supporting a deep level of necking in the lithosphere. Comparison of model predictions and crustal structure data for models invoking local isostasy (upper panel), a necking depth of 7.5 km (centre) and a necking depth of 25 km (lower panel)

Table 1 gives a compilation of necking levels, strain rates, β estimates, thermo-mechanical age, pre-rift crustal and lithospheric thicknesses at onset of rifting, EET estimates and rifting duration for the Gulf of Lion, Valencia Trough, Tyrrhenian Sea and Pannonian Basin. As a group, these four basins are characterized by rapid extension operating on pre-rift lithosphere with a young thermo-tectonic age (10–20 Ma). The estimates for pre-rift crustal thicknesses vary between 30 and 40 km, with the Gulf of Lion, Valencia Trough and Tyrrhenian Sea (30–35 km) in the lower values of this range and the Pannonian Basin (40 km) in the higher crustal thickness range.

Important differences and similarities are found in the mechanical properties of the pre-rift lithosphere inferred from the kinematic modelling studies. We note that most extensional basins in the Alpine/Mediterranean region are associated with intermediate levels of necking (*Table 1*). A shallow level of necking (5 km), however, is found for the Pannonian Basin (van Balen *et al.*, unpublished data), whereas the Tyrrhenian Sea (25 km) and the Gulf of Lion (25–35 km) represent the other end-members with a deep level of necking. Intermediate–deep levels of necking are encountered in the Valencia Trough (17–33 km). Two other main differences exist between the Pannonian Basin and the Gulf of Lion and Tyrrhenian Sea. In the Tyrrhenian Sea and the Gulf of Lion, fast extension (rates of $8.3–6.7 \times 10^{-15}$ s^{-1}) are coupled with a low ratio of subcrustal/crustal thinning (see also Spadini *et al.*, 1995), whereas a medium fast extension (rates of 1.6×10^{-15} s^{-1}) and high ratio of subcrustal/crustal thinning is characteristic of the Pannonian Basin. Secondly, the style of crustal and lithospheric thinning also seems to be reflected in the observed differences in overall basin shape. Whereas extension in the Pannonian Basin, after being initiated along a system of narrow, small graben, has resulted in a subsequent wide basin, the Tyrrhenian Sea extension has created a series of relatively narrow rifted basins without a subsequent change in shape. The observed differences in basin shape could reflect a possible shallowing of the necking level during the rifting phase of the Pannonian Basin. The duration of rifting is relatively short for all basins (16 Ma for the Valencia Trough and 6–8 Ma for the other basins) and variations in this parameter appear to have limited influence on the inferred estimates for necking depths.

Table 1 Estimates for level of necking (*Z*-neck), effective elastic thickness (EET), strain rate ($\dot{\varepsilon}$), thermal age at the onset of extension (Ma), duration of extension (Ma) and pre-rift lithospheric and crustal thicknesses (km) for the Alpine/Mediterranean basins considered here. β and strain rate estimates are averaged through the entire rifting period. EET estimates are best-fit pre-rift values. Numbers in parentheses refer to data sources: (1) Kooi (1991); Van Wees (1994); (2) Janssen *et al.* (1993); (3) Spadini *et al.* (1995); and (4) Van Balen *et al.* (in press)

Basin	*Z*-neck (km)	$\dot{\varepsilon}$ (10^{-15} s^{-1})	β	Thermal age (Ma)	Pre-rift crustal thickness (km)	Pre-rift lithospheric thickness (km)	EET (km)	Rifting duration (Ma)
Gulf of Lion[1]	25–35	6.7	2.48	20	30–35	80–100	20–25	7
Valencia Trough[2]	17–33	2.9	2.46	10	30–35	100–120	9–11	16
Tyrrhenian[3]	23–27	8.3	2.57	20	30–35	80–90	20–25	6
Pannonian[4]	4–6	1.58	1.3	20	33–43	120–140	6–9	6

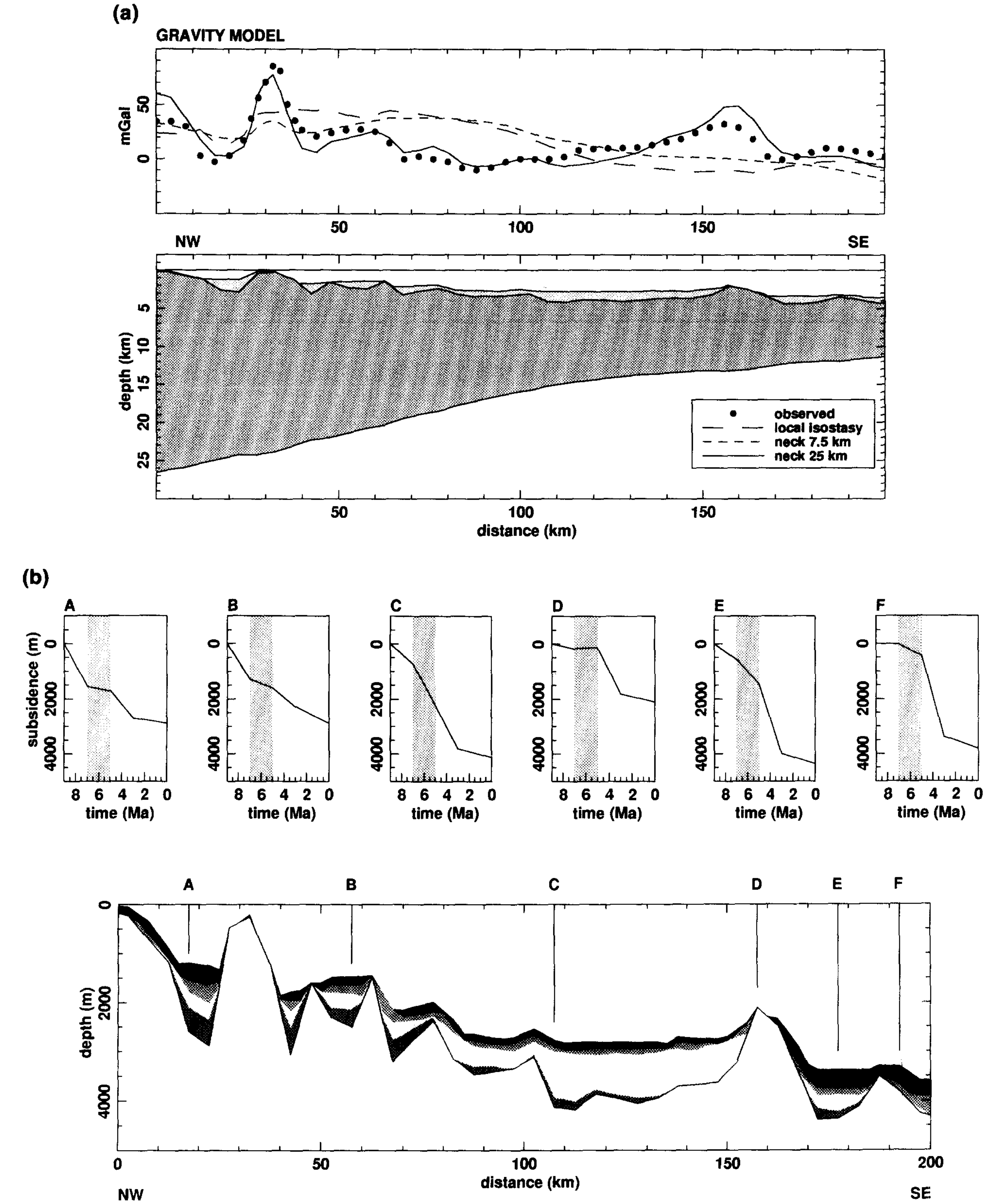

Figure 7 (a) Comparison between observed gravity and modelling predictions for the three levels of necking specified in *Figure 6*. A necking level of 25 km fits the gravity data. (b) After testing with geophysical data, the modelling package is used to simulate basin stratigraphy and the calculation of subsidence histories for different positions in the basin, allowing a comparison with backstripped subsidence curves

Intracratonic extensional basins

For eight intracratonic basins we have compiled information on the kinematic parameters for subsequent comparison with the four Alpine/Mediterranean basins discussed here. The eight intracratonic basins comprise the Saudi Arabian Red Sea margin, Transantarctic Mountains/Ross Sea basins (van der Beek *et al.*, 1995), Tanganyika–Malawi rifts (East Africa) (Ebinger *et al.*, 1991; 1993; van der Beek *et al.*, 1995), Baikal rift (Déverchère *et al.*, 1993; van der Beek, 1995), the Barents Sea margin (Reemst *et al.*,

Table 2 Estimates for level of necking (*Z*-neck), strain rate ($\dot{\varepsilon}$), effective elastic thickness (EET), stretching factor (β), thermal age at the onset of extension (Ma–Ga) and pre-rift lithospheric and crustal thicknesses (km) for the intracratonic rifts/rifted margins considered here. Numbers in parentheses refer to data sources: (5) Van Wees (1994); (6) van der Beek (1995); (7) van der Beek *et al.* (1994); (8) van der Beek *et al.* (1994); (9) Reemst *et al.* (1994); (10) Ebinger *et al.* (1991); van der Beek (1995); (11) Spadini *et al.* (unpublished data); and (12) Spadini *et al.* (unpublished data)

Basin	*Z*-neck (km)	$\dot{\varepsilon}$ ($10^{-15}\,s^{-1}$)	β	Thermal age (Ma)	Pre-rift crustal thickness (km)	Pre-rift lithospheric thickness (km)	EET (km)	Rifting duration (Ma)
North Sea(5)	15–20	0.2	1.5	300	32–35	110–130	20–35	60
Baikal Rift(6)	20–25	0.38	1.3	400	40–45	150–200	30–50	25
Saudi Arabia Red Sea(7)	10–15	22	3.5	600	35–40	150–200	40–60	5
Transantarctic(8)	30–35	2.1	5	600	40–45	180–200	40–60	60
Barents Sea(9)	15–20	0.2	2	400	30–35	110–130	25–35	11
East Africa(10)	30–40	5	1.5	600	35–40	150–200	30–50	8
Western Black Sea(11)	22–28	2.2	3	400	35–40	180–200	50–60	30
Eastern Black Sea(12)	10–15	4.9	2.23	150	35–40	80–90	20–30	8

1994), the North Sea Basin (Kooi *et al.*, 1992) and the eastern and western parts of the Black Sea (Spadini *et al.*, unpublished data; Robinson *et al.*, this issue) (*Table 2*). Two age groups can be distinguished in the basins listed in *Table 2*. The first group of basins, with thermo-tectonic ages of 300–400 Ma at the onset of rifting, has developed on Variscan/Caledonian crust (North Sea, Barents Sea margin), overlying 'normal' lithosphere for time intervals long enough after Caledonian orogeny. The second group of basins, with thermo-tectonic ages of approximately 600 Ma, has developed within Precambrian cratons (e.g. the East African rift, Transantarctic Mountains) overlying old, cold and strong lithosphere. Modelling of these rifts and basins is constrained by crustal structure data from refraction seismics and gravity. For the Barents Sea and the North Sea, additional seismic reflection data on the basin architecture have been utilized. In the case of rifted margins, as shown by van der Beek *et al.* (1994; 1995) and van Balen *et al.* (unpublished data), it is important to take onshore rift shoulder erosion into account when explaining the observed topography and basin architecture. Practically the only record of rift margin erosion in these settings is provided by apatite fission track thermo-chronology of the exposed basement rocks (Dumitru *et al.*, 1991; van der Beek *et al.*, 1994; 1995).

Kinematic modelling, incorporating margin erosion estimates constrained by fission track thermochronology, has been applied to the Transantarctic Mountains and the Saudi Arabian Red Sea margin (van der Beek *et al.*, 1994). Both margins show significant rift-related tectonic uplift of the order of 3 km for the Saudi Arabian Red Sea margin to >5 km for the Transantarctic Mountains. Local isostasy models predict tectonic uplift patterns which are not compatible with the observed topography, amounts of erosion and inferred morpho-tectonic evolution, nor with the crustal structure and gravity anomalies observed. Models that incorporate flexural support and medium-deep levels of necking (15 km for the Saudi Arabian Red Sea margin, 30 km for the Transantarctic Mountains) do successfully predict these features.

Table 2 provides a compilation of necking levels, strain rates, β estimates, thermo-mechanical age at onset of rifting, pre-rift crustal and lithospheric thicknesses and rifting durations for the eight intracratonic basins and rifted margins mentioned earlier. As a whole, compared with the Mediterranean/Alpine group of basins, the inferred average level of necking appears to be deeper, whereas the thermal age is considerably larger. The levels of necking inferred from kinematic modelling studies fall into two groups: medium (Saudi Arabian Red Sea margin, 10–15 km; Eastern Black Sea, 15 km; Barents Sea, 17 km; North Sea, 18 km) and deep levels of necking (Baikal Rift, 20 km; Western Black Sea, 25 km; Transantarctic Mountains, 30 km; Tanganyika–Malawi Rift system, 30–40 km). A separation can be observed in terms of low (North Sea, Baikal rift, Barents Sea) and high strain rates (Transantarctic Mountains, Eastern and Western Black Sea, East African Rift and in particular the Saudi Arabian Red Sea margin), corresponding to low and high extension factors, respectively. The Red Sea opening is anomalous in being accompanied by major magmatism (e.g. Menzies *et al.*, 1992). An inverse correlation can be observed between strain rate and necking level. A correlation is absent between the thermo-tectonic age grouping with a 150 Ma (Eastern Black Sea), 300–400 Ma (North Sea, Baikal rift, Barents Sea and Western Black Sea) and a 600 Ma (Transantarctic Mountains, East Africa) subset and levels of necking. Similarly, the pre-rift crustal thickness appears to vary more or less randomly between 30 and 40 km without an obvious empirical control on the differences observed between the individual basins.

How can we explain the inferred differences and similarities in necking level and rifting styles?

Important differences occur between the main parameters used to characterize the kinematics of extension inferred from these 12 case histories. These differences appear in the first place between the group of four Alpine/Mediterranean basins as a whole and their counterparts in the group of eight intracratonic rifts and margins, but also between individual cases within these groups. For a full understanding of the observed differences and similarities, a consideration of all the different factors involved, including the interplay of extension velocities and initial rheology, would be required. At this stage we limit ourselves to some remarks addressing aspects which relate to the possible link between the kinematics of extension and the collisional orogenic processes operating on the lithosphere before extension. We restrict ourselves to tectonic scenarios to be expected for a situation encountered in the Mediterranean/Alpine and cratonic case histories discussed earlier. *Figure 8* shows correlation diagrams constructed to examine more closely the presence or absence of correlations between factors controlling the extensional basin formation. Eleven of the 12 basins plotted in *Figure 8*, with the exception of the Saudia Arabia Red Sea margin, show the existence of correlations between necking level, pre-rift crustal and lithospheric thickness and strain rate. Inspection of these diagrams reveals a particularly strong inverse correlation between pre-rift crustal structure and necking depths in the Alpine/Mediterranean basins: with an increasing thickness of the pre-rift crust, necking levels are shallowing. A less pronounced similar correlation can also be observed between lithospheric thickness and necking depth. We infer from these findings that the crustal control on necking depth prevails over the lithospheric control on this extensional basin formation parameter, suggesting rheological decoupling in the pre-rift lithosphere. In contrast, and although characterized by a much larger scatter in the inferred parameter values, the case histories for cratonic extension discussed here indicate a trend of a deepening of the level of necking with increasing lithospheric thickness, which may suggest that the level of necking is primarily controlled by lithospheric thermal structure in these cases. The diagrams also show an absence of a correlation between pre-rift crustal thickness and necking depth in the intracratonic basins. The highest lithospheric thickness values probably reflect low mantle heat flow and a thermal structure which would suppress intra-lithosphere decoupling during extensional basin formation.

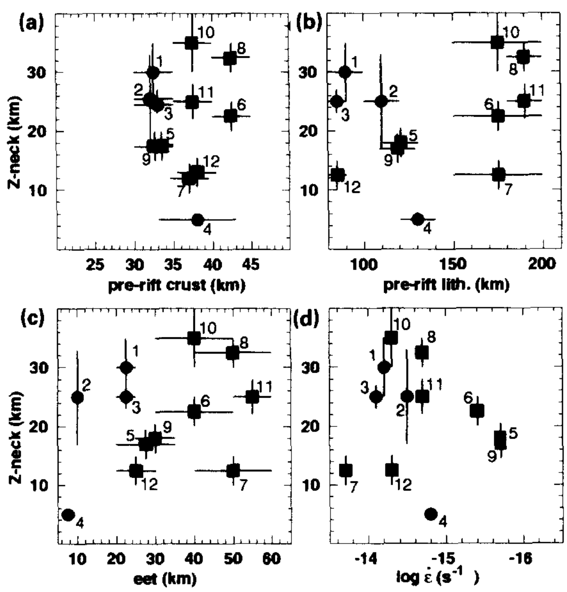

Figure 8 Correlation diagrams for the graphic representation of relationships between parameters displayed in *Tables 1* and *2*. (a) Necking depth and pre-rift crustal thickness; (b) necking depth and pre-rift lithosphere thickness; (c) EET and necking depth; and (d) necking depth and strain rate. Closed dots and squares indicate data from Alpine/Mediterranean basins and intracratonic rifts, respectively. Numbers in diagrams refer to basins listed in *Figure 1*. Note the anomalous position of the Saudi Arabia Red Sea margin (7) in these diagrams

Both the Alpine/Mediterranean and intracratonic case histories demonstrate a positive correlation between an increase in EET values and an increase in the level of necking, indicating a link between these indicators for bulk crustal and lithospheric strength. A striking correlation can be observed between strain rates and necking depths. A decrease in strain rate corresponds to a lower value of necking depth, reflecting the well known reduction of lithospheric strength with decreasing strain rate. This observation provides confidence in the reliability and robustness of the estimates for necking depths obtained from the kinematic modelling studies. The consistent anomalous behaviour of the relation between strain rate and necking depth for the Saudi Arabia Red Sea margin, together with the deviations in the other rifting parameters for this margin, suggest a systematic contribution by other tectonic processes to its basin formation. Volcanism and magmatism may have significantly contributed to the extremely rapid basin extension, leading in this case to ocean basin formation in such a short time. In general, major differences in pre-rift rheology are the result of a cool versus a warm pre-rift lithosphere in the 12 basins modelled. Whereas cratonic rifting probably took place under relatively stable thermal conditions, the Alpine/Mediterranean setting can be characterized by a transient thermo-tectonic regime. As observed from a comparison of the Alpine/Mediterranean basins and the cratonic basins, rifting durations tend to be longer for the intracratonic basins, probably reflecting a larger integrated strength for this class of basins (see *Figure 2*).

A comparison of the inferred parameters for necking depth and other parameters for the Eastern and Western Black Sea basins derived from the modelling studies by Spadini *et al.* (unpublished data) and Robinson *et al.* (this issue) illustrates a number of these points. Their model results support the existence of pronounced differences in the lithospheric properties of the Western and Eastern Black Sea before rifting. The Western Black Sea was formed in middle Cretaceous times on cold, 200 km thick lithosphere. In contrast, in the Eastern Black Sea Palaeocene rifting occurred on a warm and relative thin lithosphere of approximately 80 km, as a consequence of preceding Early Jurassic extension. The extension in the Western Black Sea is characterized by a deep level of necking (25 km), whereas a relatively shallow level of necking (15 km) is predicted for the Eastern Black Sea. The inferred differences in necking depths therefore appear to reflect the presence of a warm, weak rejuvenated lithosphere in the Eastern Black Sea and a strong intracratonic type of lithosphere with significant strength in the upper mantle in the Western Black Sea. The differences in inferred necking levels are also compatible with observed differences in gravity and total tectonic subsidence (Spadini *et al.*, unpublished data).

Differences between the magnitudes of the driving forces for lithospheric extension can also been inferred. Whereas cratonic rifts seem to have been predominantly affected by far-field stresses, in some instances possibly in combination with plume activity, the basins formed in a regime of general convergence have mostly been affected by near-field stresses resulting from slab retreat (e.g. Bassi and Sabadini, 1994).

It appears that observed differences in the duration of rifting phases and strain rates are largely the consequences of pre-rift lithosphere evolution. For low strain rates, encountered in the Pannonian Basin, strength profiles constructed for a thick crust and high heat flow indicate the presence of a shallow level of necking (see *Figure 9d*) and would lead to the development of a wide basin (Buck, 1991). Extension in a tectonic setting with a normal crustal thickness and a low initial heat flow can manifest itself by a deep level of necking (see *Figure 9f*) and the development of a narrow basin (Buck, 1991). The first situation appears to be encountered in the Pannonian Basin. For the other Mediterranean basins discussed here, pre-rift rheology appears to be more important than strain rate, possibly explaining the presence of observed variations in these basins of intermediate width.

A number of interesting interrelationships can be observed between rift width, strain rate and necking levels. Initial rheology, which depends largely on (transient) heat flow and crustal thickness, seems to exert the major control on these factors. Dynamic modelling (Bassi, 1995) has demonstrated that a cold rheology with a major contribution of brittle strength will promote the development of rapid and narrow necking, whereas a warm creep-dominated rheology will lead to wider rifts. Her models also indicate a minor control by the rate of extension when rifting is initiated in a low yield strength regime, or when it reaches the yield strength during extension. These results confirm earlier findings by Buck (1991) which suggest that pre-rift rheology will be more important than strain rate variation in controlling the mode of extension. The lithospheric temperature structure is generally transient, in contrast with the steady-state situation often assumed as the initial conditions in modelling. This implies the need for understanding the factors controlling thermal age and, therefore, the nature of the underlying tectonic processes. The thermal age concept (e.g. Burov and Diament, 1995) further implies the need to involve additional boundary conditions if the resulting strength is less than that of the surrounding lithosphere.

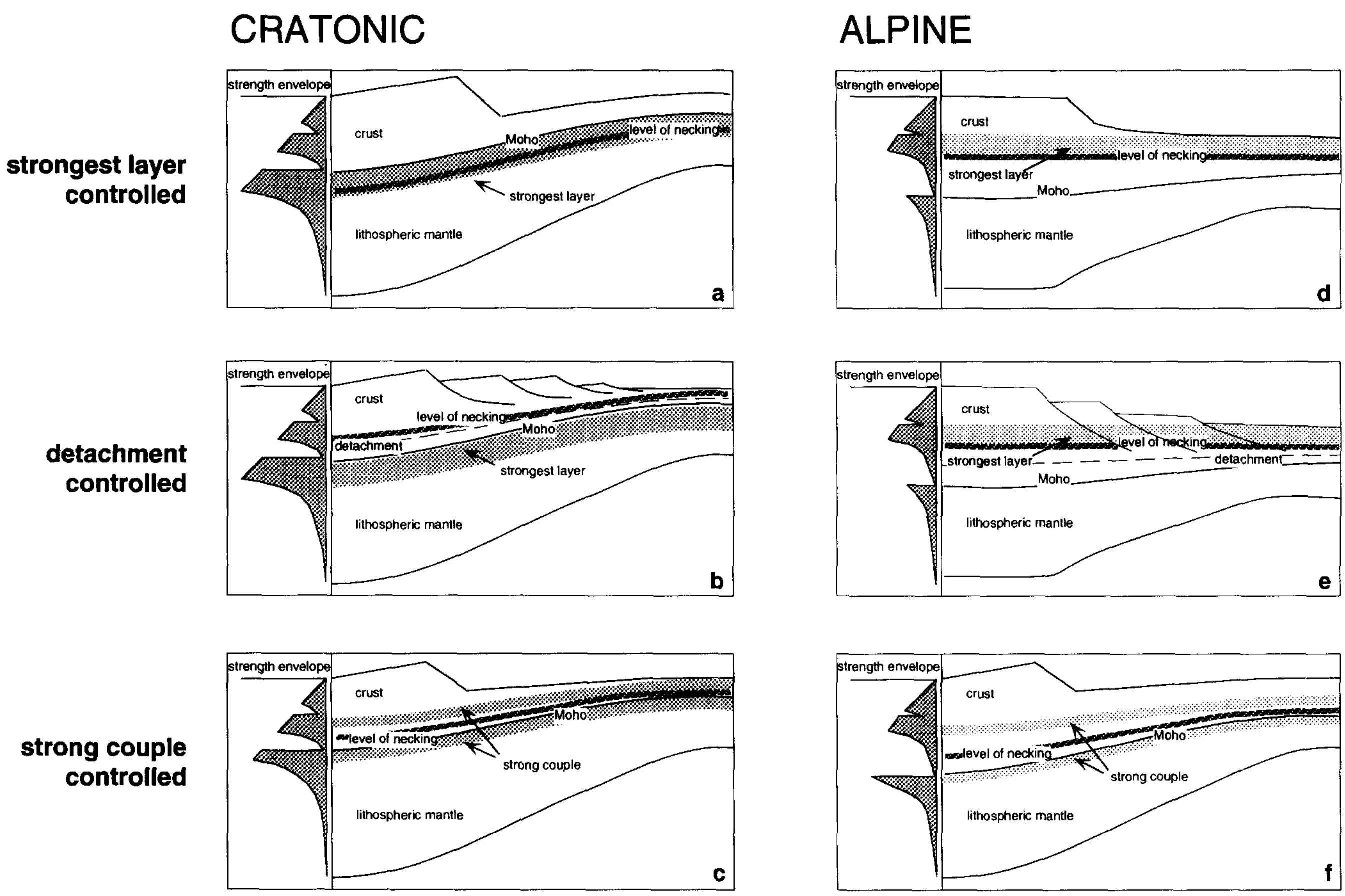

Figure 9 Three different interpretations of the mechanical significance of the levels of necking inferred from kinematic models of extensional basin formation. Differences between the various interpretations differ in terms of the relationships between necking depths and detachment levels in the lithosphere and the presence or absence of strong layers in the pre-rift lithosphere involved in extensional basin formation in cratonic (left) and Alpine (right) tectonic settings. (a, d) The level of necking corresponds to the zone of maximum strength in the lithosphere; for typical cratonic (strong) lithosphere this layer is at depths just below the crust-mantle boundary (Kooi *et al.*, 1992). (b, e) The level of necking represents a level in the crust where faults sole out, roughly corresponding to the level of intracrustal detachment zones (van der Beek *et al.*, 1995). (c, f) The level of necking represents the effective equivalent of the response of a multilayered rheology with a couple of strong layers in the crust and upper mantle. As a result, the level of necking is found at the depth level in between these two layers, coinciding with the weakest segment of the lithosphere at lower crustal levels (Spadini *et al.*, 1995). Models invoking Alpine pre-rift lithosphere predict a level of necking deeper than anticipated for cratonic lithosphere. Note also the differences in the depth spacing predicted by the models

Lithospheric stress, strength and basin formation: the importance of strength evolution during pre-rift thickening and extension

Until now we have discussed the parameters characterizing the rheological state of the extending lithosphere and the associated deformation signature. The initiation of extensional basin formation depends on the interplay of forces operating on the continental lithosphere and the spatial distribution of lithospheric strength. In intracratonic settings, a large distance away from plate boundaries, far-field lithospheric stresses should be relatively constant (Zoback, 1992) and, therefore, lithospheric deformation is expected to be primarily controlled by the initial distribution and subsequent evolution of lithospheric strength, possibly amplified by local sources of stresses. At plate margins, close to the main sources of intraplate stress, large horizontal and vertical fluctuations occur. For intraplate settings and adopting constant intraplate stresses, the localization of extensional basin deformation is in agreement with a predicted reduction of extensional strength of the basin centre relative to its margins (e.g. Buck, 1991; Bassi, 1995; Van Wees and Stephenson, in press). These models assume an initial steady lithospheric configuration. Here we examine the evolution of lithospheric strength in regimes characterized by pre-rift thickening to analyse the effect of orogeny on the evolution and distribution of lithospheric strength. For this purpose, we have carried out a set of simple one-dimensional model calculations (*Figure 10*) to obtain a qualitative picture of the temporal evolution of bulk strength of the lithosphere during pre-rift thickening and subsequent thermal relaxation.

In the models a pre-orogenic lithospheric configuration with an upper crust of quartzite rheology, a lower crust of diorite and an olivine mantle is adopted. The initial crustal thickness is taken as 35 km. For the initial geotherm we adopt a spectrum of surface heat flux values between 50 and 80 mW m^{-2}, representing thermal settings ranging from old cratonic to young thermally rejuvenated continental lithosphere. We assume that heat production of the upper crust accounts for 40% of the surface heat flow, whereas heat production in the lower crust is 0.5×10^{-6} W m^{-3}. For pre-rift thickening of the lithosphere a pure shear (homogeneous) thickening by a factor of 2 is adopted, taking place in a timespan of 30 Ma. In the solution of the thermal evolution a finite difference technique is used (Van Wees *et al.*, 1992). Temperatures are kept fixed to T = 0°C at the surface. At the lithosphere–asthenosphere boundary (T = 1333°C) a constant heat flow is adopted, corresponding to the heat flow at the base of the steady-state solution of the geotherm. As a result of tectonic movements the lithosphere–asthenosphere boundary migrates for the basal heat flow boundary condition.

Model results for the integrated strength evolution (*Figure 11*), during pre-rift homogeneous thickening and subsequent thermal relaxation, are marked by a pronounced increase of lithospheric strength during thickening as a result of syn-tectonic cooling, followed by a progressive decrease during thermal relaxation in the post-orogenic (extension) phase.

Post-orogenic extension occurs favourably in the thickened lithosphere if the integrated strength becomes less than the initial pre-orogenic strength. Consequently, for different initial heat flux values in *Figure 11*, we can distinguish temporal domains in the post-orogenic evolution where extension in thickened lithosphere is possible and where it is highly unlikely. The extent of these temporal domains can be compared with documented time laps observed between orogenic thickening and the onset of post-orogenic basin formation in Alpine and cratonic settings (see *Figure 11*). In the case of documental Alpine post-orogenic extensional settings of the Tyrrhenian Basin and Gulf of Lion (*Table 2*), the inferred presence of a thin pre-rift lithosphere indicates relatively high initial surface heat flow (70–80 mW m^{-2}). These surface heat flow values correspond to relatively low pre-orogenic integrated strength and short predicted time laps between orogeny and the possibility for onset of post-orogenic extension. These time laps are in remarkable agreement with observed thermal age values given in *Table 1*. For the Valencia Trough and Pannonian Basin, marked by considerably thicker initial lithosphere, predicted relaxation times of about 40 Ma are higher than the observed thermal ages. These probably indicate longer durations of thickening or may reflect thermal perturbations in the subcrustal lithosphere.

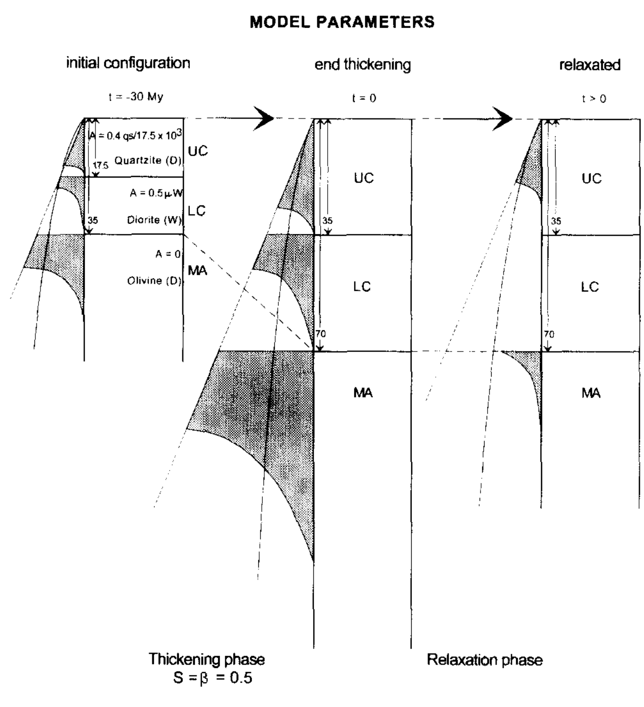

Figure 10 Modelling used for the calculation of the strength evolution schematically illustrated for a sequence of lithospheric thickening followed by relaxation and basin extension. Initial temperatures for the lithosphere are taken equal to those of a steady-state geotherm (Pollack and Chapman, 1977; Chapman, 1986). A three-layer lithosphere is adopted (upper crust, lower crust, mantle) with a crustal thickness of 35 km. Upper crustal thickness is equal to lower crustal thickness. We assume that heat production of the upper crust accounts for 40% of the surface heat flow, whereas the heat production in the lower crust is characterized by a heat production of 0.5×10^{-6} W m^{-3}. For thickening of the lithosphere a pure shear thickening by a factor of 2 is adopted, taking place in a timespan of 30 Ma

In the case of cratonic lithosphere, the inferred presence of a thick pre-rift lithosphere (*Table 2*) indicates relatively low initial surface heat flow (50–60 mW m^{-2}). For these geotherms, the initial strength is high and the model results indicate that only after hundreds of millions of years (to more than 1.5 Ga) of post-orogenic thermal relaxation, the strength has decayed to its initial pre-collisional value, allowing extension to occur in the previously thickened areas. This implies that the mechanical conditions for extension are not optimal during a long time span after the completion of collisional events, which appears to agree well with inferred thermo-tectonic ages in *Table 2*. In this case, the role of external forces probably has to be fairly prominent to overcome the intrinsically high levels of lithospheric strength to be exceeded before extension can be initiated. In this respect, the control by pre-existing discontinuities seems to be extremely important (Daly *et al.*, 1989; Van Wees and Stephenson, in press).

A number of features are not included in the model. For example, we refrain from a full two-dimensional modelling effort (Van Wees *et al.*, 1992). Such an approach would result in slightly faster thermal relaxation times in the modelling predictions, but would not affect the essential characteristics of the model results. We also ignore the effects of erosion and sedimentation. Similarly, the mode of orogenic thickening is largely simplified in our modelling approach. On the pertinent time-scales, thermal and associated mechanical relaxation is primarily controlled by the thickness of the lithosphere (*Figure 11*) and not by the mode of orogenic thickening. This result confirms earlier findings by Sonder *et al.* (1987) on the existence of an inverse dependence of the time lap between thickening and extension on initial Moho temperature. Orogenic stacking, instead of homogenous thickening, would slightly alter the strength predictions during thickening and the early stage of thermal relaxation (England and Thompson, 1986). This will not affect the essential outcome of the modelling presented here, as we address the first-order characteristics of bulk strength evolution during thickening and post-orogenic thermal relaxation.

As discussed earlier, the Pannonian Basin and the Valencia Trough are both characterized by post-orogenic extension relatively rapidly after orogeny compared with the predicted relaxation times. An example of this is probably also found in the cratonic lithospheric setting of the Scandinavian Caledonides, where a 10 Ma time lag has occurred between the timing of the onset of collapse and the timing of the onset of extensional basin formation (e.g. Chauvet and Seranne, 1994). This notion seems to be incompatible with the findings from our modelling for old cratonic settings displayed in *Figure 11*. However, in these settings, delamination of subcrustal lithosphere can lead to an abrupt and drastic reduction of integrated strength. This process would seem to affect primarily cratonic lithosphere due to more favourable conditions for delamination in old lithosphere combined with a large relative contribution of subcrustal strength to total lithosphere strength. In Alpine settings conditions for mantle delamination are probably less favourable.

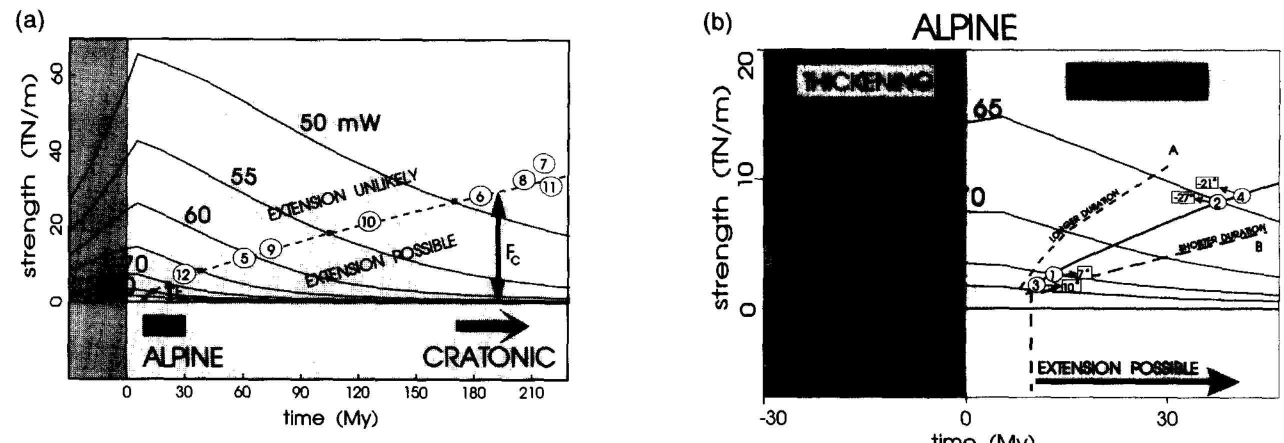

Figure 11 Evolution of integrated lithospheric strength for the sequence of collision (with a thickening factor of 2) and subsequent extension displayed in *Figure 10*. Different curves correspond to different values of initial surface heat flow. The thick broken line connects the points where the strength after extension reaches the strength levels characteristic for the lithosphere before the onset of the collision and thickening of the lithosphere. This line separates the domains where extension in thickened lithosphere is possible and where it is highly unlikely. Vertical distance indicated by arrows indicates the maximum level of external force F required to exceed the strength of the lithosphere in an extensional regime in Alpine/Mediterranean (F_A) and intracratonic settings (F_C). (a) Model range for ages spanning Alpine to cratonic tectonic settings. Numbers refer to intracratonic basins listed in *Figure 1* and *Table 2*. Comparison with documented thermal ages indicates that the observed onset of extension in intracratonic basins falls well within the domain where extension is predicted to be possible. (b) Alpine lithosphere with high initial surface heat flow (70–80 mW m^{-2}). The thick solid line corresponds to a segment of the broken curve displayed in *Figure 11a*. Broken line A shows the effect of a collision of a duration longer than 30 Ma. Broken line B shows the effect of a collisional phase shorter than 30 Ma. Numbers refer to the four Alpine/Mediterranean basins listed in *Figure 1* and *Table 1*. Numbers annotated with asterix display the ofset (in Ma) between the timing of the documented post-orogenic extension and minimum time co-ordinates predicted. Comparison with documented thermal ages indicates that the observed onset of extension in the Gulf of Lion and Tyrrhenian Sea basins falls well within the domain where extension is predicted to be possible; the relatively large positive ofsets for these two basins even allow for considerable shorter duration of pre-extensional thickening. In contrast, negative ofsets for the Valencia Trough and Pannonian Basin most likely require a longer duration or the thickening processes possibly in combination with a thermal perturbation in the subcrustal lithosphere

As a result of this process, through the primary control of crustal strength on the mechanical lithosphere, a more modest effect on total lithospheric strength distribution is expected. Questions remain on the nature of the material replacing the delaminated subcrustal lithosphere, and whether in Mediterranean settings this material will have a comparable high temperature signature as the original configuration. In contrast, low initial heat flow values, characteristic of cratonic settings, could promote slab detachment as a result of gravitational instability. Slab detachment processes could therefore strongly affect the rheological strength evolution by removing the strong subcrustal lithosphere. In the calculated strength evolution displayed in *Figure 11* this effect is not incorporated. Incorporation of this effect will particularly affect the strength predictions made for old intracratonic type of lithosphere, whereas strength profiles for Alpine-type lithosphere with a very modest contribution by mantle strength (Okaya *et al.*, unpublished data) will be hardly affected by this process.

In the case of Alpine lithosphere, the duration of the collisional phase appears to be fairly important for the strength evolution. *Figure 11b* shows the evolution of integrated lithospheric strength in the case of Alpine lithosphere with high initial surface heat flow (70–80 mW m^{-2}), showing the effects of different durations of the collision phase or, equivalently, different thickening factors. The figure also shows the duration of orogenic stacking characteristic for the Eastern Alps (Genser *et al.*, in press) and the southern Apennines (Patacca *et al.*, 1990), followed by extensional basin formation in the Pannonian Basin and the Tyrrhenian Sea, respectively. Inspection of the figure shows that a longer duration of the orogenic phase (as in the Austro-Alpine case) causes lithospheric strength to drop below intracratonic values much quicker than a shorter duration of collision.

The levels of external forces required to induce extension appear to be different in the Alpine/Mediterranean and cratonic settings. In the Alpine case a very low level of external tectonic force is required, well within the realm of force levels associated with subduction and slab retreat, characteristic of this tectonic setting (Wortel and Spakman, 1992). It appears that thickened lithosphere in this setting, after a few tens of millions of years in the relaxation phase on collision, is extremely vulnerable to even a very modest trigger by an external force such as slab pull. In contrast, the intracratonic situation appears, even after hundreds of millions of years, still to be associated with an intrinsically high level of lithospheric force required to initiate extension, whereas in these settings the association with subduction and retreating slabs is not evident, suggesting a primary control by far-field stresses.

Implications: the need for better constraints on pre-rift collisional processes

The models discussed here illustrate the need to obtain a better understanding of the pre-rift mechanical structure to improve current models for extensional basin formation. Lithospheric thickening in collisional regimes has important rheological consequences, which can be tested using presently available two-dimensional modelling techniques (e.g. van Wees *et al.*, 1992; Wijbrans *et al.*, 1993). Of particular importance in this respect is the availability of additional constraints from P–T–t data through metamorphic geology and thermochronology. The incorporation of these data sets in thermo-mechanical modelling in Alpine/Mediterranean settings, characterized by rapid successions of extension and collision, promises to be a fruitful approach to modelling the rheological evolution of continental lithosphere (van Wees *et al.*, 1992; Wijbrans *et al.*, 1993; Genser *et al.*, in press).

It therefore appears that a steady-state approach to the characterization of pre-rift thermal evolution has to be abandoned in favour of a transient thermal modelling approach. The feedback to rheology and orogenic stacking also puts constraints on the mechanical control versus the parts played by the processes of subduction/roll-back in extensional basin formation. This emphasizes the need for better constraints on pre-rift rheology from forward modelling of extensional basins in collisional settings and parallel studies in intracratonic rift settings. Full incorporation of P–T–t paths and orogenic stacking of pre-rift lithosphere is in principle possible, as demonstrated by a number of recent modelling studies on well exposed mountain chains affected by late orogenic extension, including the Betics (van Wees *et al.*, 1992), Eastern Alps (Genser *et al.*, in press) and metamorphic core complexes in the Aegean (Wijbrans *et al.*, 1993). These studies provide support for the presence of intrinsically weak lithosphere in the cores of most Alpine/Mediterranean mountain belts (see also Okaya *et al.*, unpublished data), providing optimum conditions for late-stage extension.

Differences in the mode of extension between intracratonic rifts and margins and Alpine/Mediterranean basins appear to be primarily controlled by differences in (transient) thermal regime during the pre-rift phase and the duration of pre-rift lithospheric thickening, affecting the level of external stress required to result in extension. The latter are directly coupled to the presence of near-field versus far-field regional stresses.

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