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3.1 Introduction

It has been more than 30 years since Sleep (1971) first proposed that passive continental margins form by thermal contraction following rifting and continental breakup. According to his model, the lithosphere is heated during rifting which causes thermal expansion and uplift over a broad region. Cooling restores the uplifted lithosphere to its pre-rift level, but sub-aerial erosion thins it. The net result is that the lithosphere subsides below its initial level, thereby forming a marginal depression for continental-derived sediments to infill.

Sleep (1971) assumed that margin evolution could be modelled by the cooling of a plate from an initially high temperature and an erosion rate that was proportional to the regional elevation. He substantiated his model by showing that the depth to horizon tops in wells in the coastal plain of the East Coast, USA, when normalised to a particular depth (in his case the top of the Woodbine) followed an exponential curve with a thermal time constant of about 50 Myr. A similar time constant characterises the subsidence of oceanic lithosphere away from a mid-ocean ridge.

Walcott (1972) showed that although thermal contraction and erosion may indeed combine to produce a marginal depression, sediments act as a load on the surface of the lithosphere and cause additional subsidence. He demonstrated that if sediments prograded into a 5 km deep basin, the lithosphere (which includes the crust) would bend or flex by as much as 8 km, depending on the density of the sediment and, importantly, the flexural rigidity of the underlying lithosphere.

Sleep and Snell (1976) constructed a model that combined thermal contraction, erosion and flexure and showed that it explained many features of the stratigraphic architecture of the coastal plain of the East Coast, USA. One problem with the model, however, was that it was based on a viscoelastic rather than an elastic plate. Flexure studies (e.g., Watts and Cochran, 1974) had earlier

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suggested that the oceanic lithosphere is elastic, rather than viscoelastic, and that if there is any load-induced stress relaxation, it occurs on short time-scales (i.e., <1 Myr) rather than on the long time-scales ($\sim$10–50 Myr) that had been assumed by Sleep and Snell (1976).

Nevertheless, Sleep's model was instrumental in focussing attention on the structural styles of passive margins. In two seminal papers, Montadert et al. (1977) and de Charpal et al. (1978) showed that the Bay of Biscay and Rockall Plateau (eastern North Atlantic) margins were underlain by tilted and rotated 'blocks' that were bounded by listric faults. These authors noted that there was little evidence for sub-aerial erosion preceding rifting. They suggested that the tilted blocks formed in response to extension during rifting and that it was brittle deformation, together with ductile flow, that thinned the crust and caused subsidence.

These observations, together with the lack of evidence for erosion in the deeper parts of the North Sea basin, led McKenzie (1978) to develop an alternative model. He proposed that the subsidence of rift-type basins was caused partly by stretching of the crust during extension (the 'initial subsidence') and partly by thermal cooling of the thinned sub-crustal lithosphere (the 'thermal subsidence'). The McKenzie stretching model, as it is now known, and its modifications for the effects of finite rifting and lateral heat flow (e.g., Cochran, 1983), have subsequently enjoyed much success in explaining the subsidence and uplift history of rift-type basins in both continental interior and passive margin settings.

While stretching is a highly effective way to extend the crust and lithosphere, precisely how it occurs is in dispute. The McKenzie model assumes 'pure-shear' which predicts that the geometry of the thinned crust either side of a rift zone would show a high degree of symmetry. Wernicke (1985), however, suggested an alternative 'simple shear' model based on field observations of highly extended terranes that are now juxtaposed to metamorphic core complexes in the Basin and Range province of the western USA. The main feature of his model is a 'detachment surface' that separates an 'upper plate' consisting of a weakly structured rifted upper continental crust from a 'lower plate' dominated by a highly deformed lower crust. The model predicts spatial variations in the proportion of crust to mantle thinning and, hence, a high degree of asymmetry in the subsidence and uplift history either side of a continental rift zone (e.g., Wernicke, 1985) or newly formed ocean basin (e.g., Lister et al., 1986).

Since about the mid-1980s, there has been a rapid increase in the amount and, importantly, the quality of multichannel seismic (MCS) reflection profile data acquired over the world's ocean basins and their margins. These data have imaged not only the sediments, but also the rift structures that underlie them. The increase has been led by the oil and gas industry as exploration has shifted from the continental shelf to deep-water slope and rise regions. The industry activity has been supplemented by academic groups interested in rifting processes, sedimentary products and fluid flow (e.g., US MARGINS, UK Ocean Margins, France 'Marges', EUROMARGINS) and by government-led groups (e.g., Geoscience Australia) involved in the mapping of the Economic Exclusion Zone.

The new MCS data have shown increasing complexities in passive margins, especially as regards their across-strike and along-strike structure. As a result, a new generation of thermal and mechanical models based on numerical and analogue techniques has been constructed. In this chapter, we will review some of the new observations and the models that have been developed to explain them. We begin, as did Sleep, by considering the subsidence and uplift history as this is still one of the most important data sets against which the predictions of new models need to be tested.

3.2 Subsidence and uplift history

Passive margins are characterised by large thicknesses of sediments (up to ~12 km) that obscure and make it difficult to use structural styles alone to distinguish between the various rifting models. The development of techniques such as backstripping (e.g., Watts and Ryan, 1976), however, has enabled the subsidence and uplift history to be determined directly from stratigraphic data. These data may then be compared to the predictions of different rifting models.

The backstripping of biostratigraphic data from commercial wells shows that the tectonic subsidence (i.e., the subsidence not caused by sediment and water loading) of passive margins decreases with time following rifting (e.g., Watts and Ryan, 1976). The subsidence is exponential in form and bears a striking resemblance to that of oceanic crust away from a mid-ocean ridge.

Unfortunately, commercial wells tend to be drilled on structural 'highs' and so relatively few of these wells penetrate both the syn-rift (i.e., the sediments that form during rifting) and the post-rift (i.e., the sediments deposited after rifting). As a result, there is still much uncertainty about the tectonic subsidence during the syn-rift and post-rift. Some margins (e.g., South China Sea – Lin et al., 2003) appear to have an equal amount of syn-rift and post-rift subsidence. Others, however, exhibit either more syn-rift than post-rift subsidence (e.g., Western Mediterranean – Watts et al., 1993) or less syn-rift than post-rift subsidence (e.g., Labrador, Canada – Royden et al., 1980; northwest Australia – Driscoll and Karner, 1998).

The proportion of syn-rift to post-rift sediments is a useful constraint on rifting models. The instantaneous, uniform extension, McKenzie stretching model, for example, predicts an approximately equal amount of syn-rift and post-rift subsidence, irrespective of the actual amount of crustal heating and thinning. Variations therefore suggest refinements to the model. The large proportion of syn-rift to post-rift subsidence in Western Mediterranean margins, for example, has been explained by a finite rifting model in which the majority of the thinning and, hence, subsidence occurs during the syn-rift rather than the post-rift (e.g., Cochran, 1983). The small proportion of syn-rift to post-rift subsidence in the Labrador, Canada and Northwest Australia margins has been attributed to a greater amount of extension in the mantle than in the crust (Royden et al., 1980). The latter model, however, causes a 'space problem' if the *total* amount of extension is not constrained in depth (e.g., White and McKenzie, 1988). Driscoll and Karner (1998) therefore proposed another model in which extension in the upper mantle and lower crust was partitioned from the upper crust by an intra-crustal detachment.

A critical question is whether the pattern of tectonic subsidence and uplift deduced from backstripping is similar at *conjugate* passive margins (i.e., the margins that form on opposing sides of a new ocean basin). The 'pure shear' model predicts a symmetric pattern of subsidence and uplift patterns in each margin. The 'simple shear' model, however, predicts an asymmetric, spatially varying, pattern.

While both 'pure shear' (e.g., White, 1989) and 'simple shear' (e.g., Lister et al., 1986) models have been applied to passive margins, there is presently too few well data to be able to distinguish between them. The best constraints on the amount of thinning have come instead from seismic reflection and refraction data. Keen et al. (1989), for example, argued that while the pattern of faulting at the Flemish Cap, Canada and Goban Spur (Southwest Approaches, UK) conjugate margin pair is asymmetric, the amount of crustal thinning shows a high degree of symmetry. Recently, Louden and Chian (1999) using better data have questioned the degree of symmetry at this margin pair. Moreover, these workers have shown that the Labrador–southwest Greenland conjugate margin pair is initially symmetric, but shows progressively more asymmetry as the locus of rifting shifts to one side of the rift system.

Backstripping of restored stratigraphic cross-sections of ancient passive margins that have been preserved in deformed orogenic belts reveals patterns of tectonic subsidence and uplift that resemble those in modern margins. The Canadian and U.S. Rockies (e.g., Armin and Mayer, 1983; Bond and Kominz, 1984), Alps (e.g., Wooler et al., 1992), Betics (e.g., Peper and Cloetingh, 1992) and Himalaya (e.g., Corfield et al., 2005), for example, show examples where there is an equal amount of syn-rift to post-rift subsidence. By backstripping restored stratigraphic sections and comparing them to the predictions of the 'pure shear' and 'simple shear' models, it has been possible to place constraints on the age of rifting, the distribution of stretched crust, and the orientation (e.g., of the proximal and distal facies) of ancient margins.

Not all passive margin backstrip curves reveal the smooth exponential shape expected of thermal and mechanical models. Some show departures due to uncertainties in the parameters used in backstripping (e.g., the compaction depth constant, paleobathymetry and magnitude of sea-level changes in the past). Others, however, show departures due to basin-wide tectonic events and post-depositional sedimentary processes. The tectonic subsidence and uplift of the Tethyan margin in the Betics, for example, show changes in the subsidence rate during the Middle Triassic, Early Jurassic and Callovian–Hauterivian, which Peper and Cloetingh (1992) attributed to multiple rifting events. The Northeast Atlantic margin (Clift et al., 1995) shows excess uplift early on in its evolution that has been attributed to dynamic topography due to plume-related processes. In contrast, the Northwest Australia margin (Muller et al., 2000) shows excess subsidence later in its evolution that has been attributed to dynamic topography due to subduction-related processes. Finally, the eastern seaboard of Canada (Watts and Steckler, 1979) and northern Gulf Coast (Diegel et al., 1995) margins show similar magnitude excesses of uplift and subsidence. The excesses have been interpreted, however, not as tectonic in origin, but as a result of post-depositional processes associated with salt migration.

The departures in the backstrip from simple exponential curves make it difficult to invert subsidence and uplift history data *directly* for the amount of stretching and, hence, the strain rate history of a passive margin. Nevertheless, White (1994) has attempted such inversions on the basis that they are able to discriminate between extensional rifting events from other events, such as those associated with thrust/fold loading, flexure and foreland basin formation. As Newman and White (1997) have demonstrated, the strain rate and its relationship to the amount of stretching, β, is a potentially important constraint on the rheology of extended continental lithosphere.

The Wilson Cycle implies that passive margins ultimately become the sites of orogeny. Although the mechanisms by which this transition takes place are unclear (e.g., Erickson, 1993), there is evidence from backstripping of biostratigraphic data that foreland basins are underlain by stretched crust (e.g., the western deep Gulf of Mexico basin which overlies the western Gulf Coast margin (Feng et al., 1994), the Papuan basin which overlies the Northwest Australia margin (Haddad and Watts, 1999) and the west Taiwan basin which overlies the South China Sea margin (Lin and Watts, 2002)).

Figure 3.1 shows an example of one such backstrip curve from the Colville Trough (north slope, Alaska), a flexural foreland basin that formed as a result of thrust/fold loading in the Brooks Ranges. The figure shows that during the Carboniferous to Jurassic the well site was the location of an exponentially decreasing passive margin type subsidence. The decrease can be explained by a McKenzie stretching model with an initial rifting age of 360 Ma and $\beta = 1.5$, although a finite rifting model with an older rifting age and higher β could explain the subsidence equally well. Irrespective of that, the exponentially decreasing subsidence is interrupted during the Late Jurassic and mid-Cretaceous by an accelerating subsidence that is typical of foreland basins.

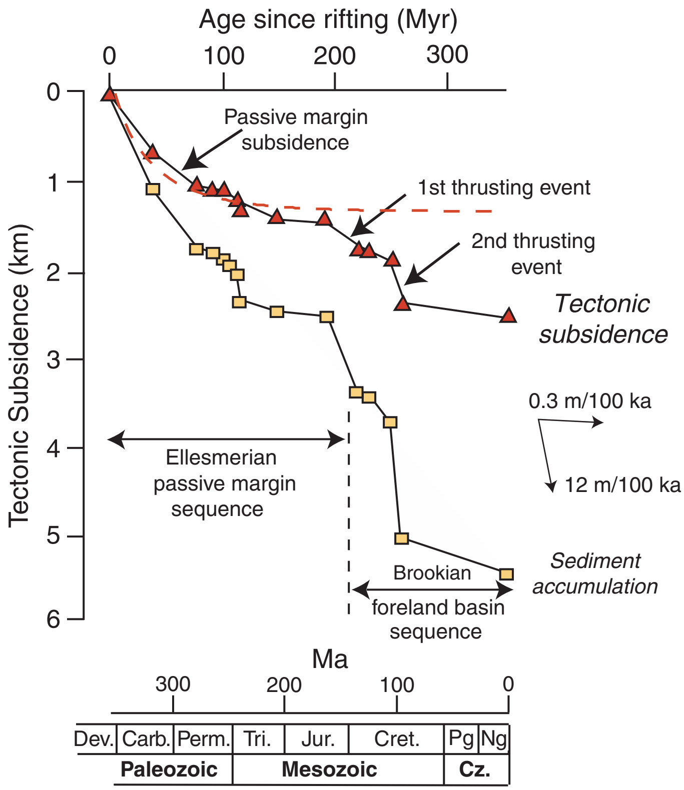

Figure 3.1 Tectonic subsidence and uplift at the Inigok-1 well in the Colville Trough, National Petroleum Reserve, Alaska. Yellow filled squares show the sediment accumulation based on data in Armagnac et al. (1988) and references therein. Red filled triangles show the tectonic subsidence and uplift obtained by progressively backstripping individual sediment layers through time. The dashed line shows the calculated subsidence based on the McKenzie 'pure shear' model with initial rifting at 360 Ma, a stretching factor, β, of 1.5, an initial crust and lithosphere thickness of 31.2 and 125 km respectively, a 0°C crust and sub-crustal mantle density of 2800 and 3330 kg m^{-3} respectively, a coefficient of volume expansion of 3.28×10^{-5} °C^{-1}, an isothermal mantle temperature of 1333°C and a thermal diffusivity of 8.0×10^{-7} m^2 s^{-1}.

The evolution of a passive margin into a foreland basin has important implications for the petroleum system. In Eastern Venezuela, for example, the subsidence caused by thrust/fold loading has taken the otherwise immature Guyana passive margin sequence into the petroleum window (Summa et al., 2003).

3.3 Thermal and mechanical structure

The main result of backstripping (e.g., Fig. 3.1) has been to show that the accumulation of sediments at passive margins is the result of two main factors: cooling following heating of the lithosphere at the time of rifting and loading due to the sediment flux. Other factors (e.g., sea-level, compaction, salt tectonics, magmatism) contribute, but it is generally agreed that these are of secondary importance. We now consider some observational constraints on the thermal and mechanical structure of passive margins.

Heat flow

An important constraint on the thermal structure of a passive margin is the present day heat flow. Of particular significance is the heat flow away from a margin, and such regions are indicative of the pre-rift 'background' heat flow of the continental crust. A comparison between the heat flow at the margin and surrounding regions is indicative of whether or not a margin is still experiencing thermal subsidence.

Unlike other marine geophysical data, there are still relatively few heat flow measurements at passive margins. Della Vedova and Herzen (1987) and Ruppel et al. (1995) have shown that the heat flow over the old (rifting ~180 Ma) East Coast, USA, margin is in the range 30–49 m Wm^{-2}. Louden et al. (1991) have shown that heat flows across the intermediate age (rifting ~100 Ma) Goban Spur and Galicia Bank margins average 50–55 and 30–35 m Wm^{-2}, respectively. Finally, Nissen et al. (1995) have shown that the heat flow over the young (rifting ~56–36 Ma) South China Sea margin averages 75–79 m Wm^{-2}. These estimates (which include, where applicable, corrections for the effects of sediment blanketing) suggest that young margins have a high overall basement heat flow while old margins have low heat flow.

Whether a margin is presently 'hot' or 'cold', however, depends on the 'background' heat flow. The old East Coast, USA, margin, for example, appears to have a similar heat flow to the adjacent crust. The intermediate age Goban Spur has higher heat flow than adjacent oceanic crust while Galicia Bank has a lower heat flow. Interestingly, *both* Goban Spur and Galicia Bank margins have a lower heat flow than the flanking continental crust. This is probably because of the relatively high radioactive heat content of the Variscan terranes that extend from Brittany and southwest England, across the Southwestern Approaches to the English Channel and into the Iberian peninsula. Finally, the young South China Sea margin heat flow is similar to that of adjacent oceanic crust, but higher than that of flanking continental crust.

Various models have been proposed to explain heat flow data at passive margins. Thermal models predict that during rifting heat flow should increase in

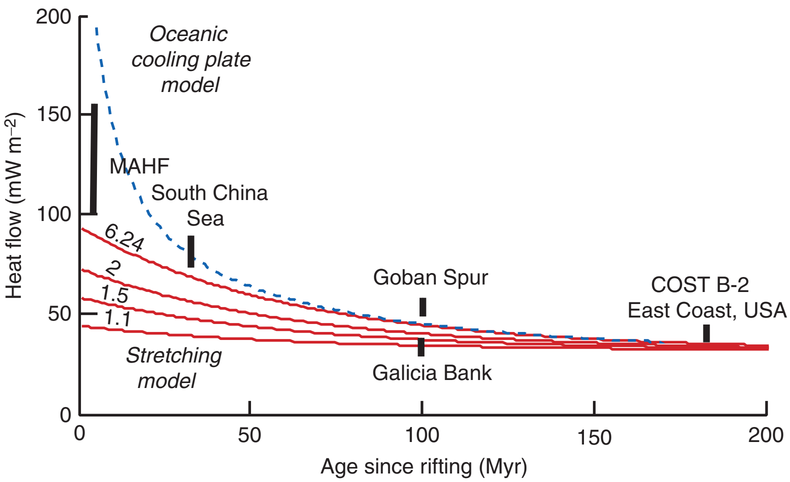

Figure 3.2 Comparison of the observed average surface heat flow over stretched continental crust at the East Coast, USA, South China Sea and Bay of Biscay margin to predictions based on a stretching model. The observed data (thick bars) is based on Nissen et al. (1995), Louden et al. (1991), Della Vedova and Von Herzen (1987) and Ruppel et al. (1995). MAHF, mean axial heat flow at the ridge crest is based on Sclater et al. (1980). The calculated solid lines are based on the McKenzie stretching model with $\beta = 1.1$, 1.5, 2 and 6.24, assuming no radiogenic heat production from the crust. Other parameters are as assumed in the subsidence calculations in Fig. 3.1. The calculated dashed lines are based on the Parsons and Sclater (1977) cooling oceanic plate model.

an oceanward direction from relatively low values over unstretched crust to relatively high values over the stretched crust. Once seafloor spreading begins, heat flow over stretched continental crust may increase, in part, due to lateral heat flow from the relatively hot oceanic crust into the relatively cold stretched crust. The increase in heat flow across a margin may persist for a few tens of Myr, but model calculations suggest that it will probably have decayed by about 50 Myr following rifting.

Probably the most detailed comparison that has been carried out to date of observed and calculated heat flow is that by Nissen et al. (1995) at the South China Sea margin. These authors used seismic refraction data to determine the amount of thinning along transects of the South China Sea margin. By using rift models (e.g., Fig. 3.2) to compute the heat flow from the amount of crustal thinning and comparing it to observed values, Nissen et al. (1995) were able to constrain parameters such as the initial lithospheric thickness and radiogenic heat production.

The South China Sea is a relatively young margin that underwent rifting during the early Paleogene (~65–35 Ma) and seafloor spreading during the late Paleogene (~35–17 Ma). The 'pure-shear' model predicts that despite oceanic crust generation ending at ~17 Ma, there should be an increase in present day heat flow across the margin. Observations, however, do not show an increase; heat flow being high over both stretched crust (Fig. 3.2) and adjacent oceanic crust. Nissen et al. (1995) suggested that this was due to either a thinner initial lithosphere (50–60 km instead of a 'standard' thickness of 125 km) or high radiogenic heat production. This would require, however, too high a heat flow at the margin prior to rifting. Their preferred model therefore was a combined one of a 'standard' initial lithosphere thickness and a moderate radiogenic heat production.

In summary, heat flow data provide useful constraints on the thermal structure of passive margins. Unfortunately, there are only a small number of heat flow transects of margins. In most margins, other observations (e.g., seismic refraction, gravity anomaly, subsidence history) are used to constrain the amount of thinning and thermal models are used to *predict* the basement heat flow. While such models have been useful in calculating the temperature structure of the sediments and thermal maturity, they are limited. Heat flow observations are crucial, therefore, not only to constrain thermal models, but also to understand the contribution of the various processes not included in these models such as radioactive heat production, deep flow of hot fluids and salt diapirism.

Elastic thickness

An important constraint on the mechanical structure of a passive margin is the elastic thickness, T_e, which is a proxy for the long-term (i.e., $>10^5$ a) strength of the lithosphere. Simple models (e.g., Watts et al., 1982) show that T_e exerts a strong control on the overall stratigraphic 'architecture' of a passive margin, explaining the existence, for example, of a coastal plain at old margins. Furthermore, there is evidence that flexure may account for some of the stratigraphic patterns that are observed in passive margins, such as onlap and offlap (e.g., Watts, 1989).

There have been a number of estimates of T_e at passive margins. Some are based on reconstructions of the footwall uplift and hanging-wall geometry of syn-rift basins (e.g., Clift et al., 2002). The majority, however, are based on gravity anomaly data (e.g., Cochran, 1973) which are sensitive not only to the local rift geometry, but also to the deformation that is caused by all the sediment and other (e.g., magmatic underplate) loads that have been applied to a margin during its evolution.

One problem in using the gravity anomaly is the requirement that all the loads acting on the crust (and lithosphere) be specified. This may not be a problem with magmatic underplating, the geometry of which is usually constrained by seismic refraction data, but it is a problem with sediments where it is difficult to determine the pre-load configuration of the margin. Holt and Stern (1991), for example, used the paleobathymetry derived from biostratigraphic data to define the base of the sediment load and present day bathymetry to derive the top of the load.

Watts (1988) suggested a method, dubbed Process-Oriented Gravity Modelling (POGM), that does not require *a priori* estimates of the paleobathymetry. The method regards passive margins as simple mechanical systems (e.g., Fig. 3.3) in which the crust thinned during rifting is subsequently subject to sediment (i.e., the non-volcanic type passive margin), volcanic (i.e., volcanic margin) or some combination of these loads. Flexural backstripping (i.e., backstripping that explicitly takes into account the strength of the lithosphere) of the sediment

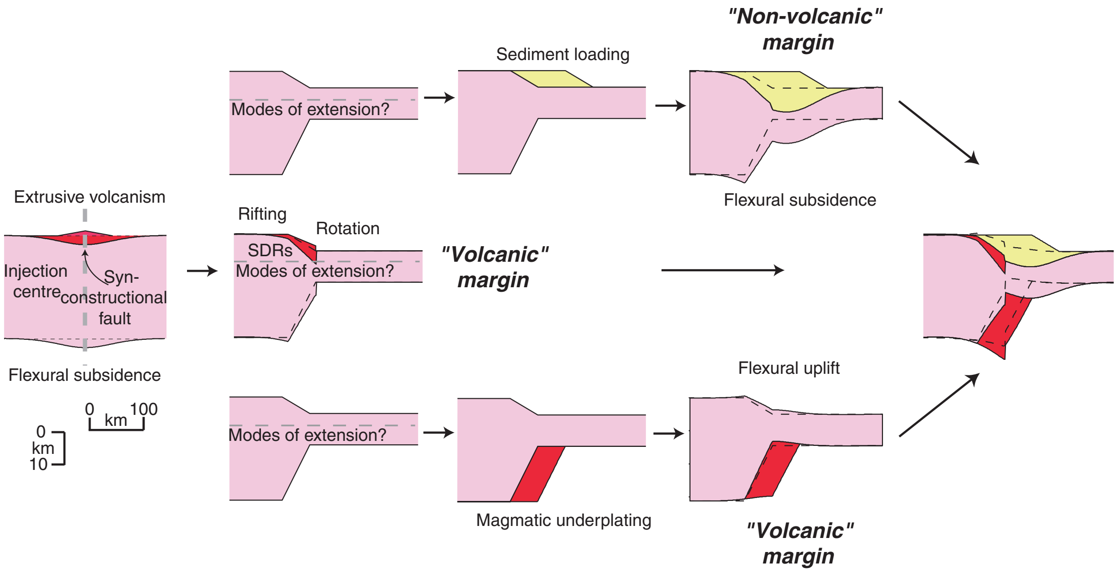

Figure 3.3 Simple models for the mechanical evolution of passive margins. The models view that passive margins form by a combination of rifting, which thins the crust and sediment loading, sub-aerial volcanic loading and magmatic underplating which deforms it. Seaward dipping reflectors (SDRs) refer to the material that initially infilled the flexural moats that flanked the main region of extrusive volcanism and has since subsided (and rotated) following rifting. The pre-rift crust is assumed to have a zero elevation thickness of 31.2 km.

reveals the Total Tectonic Subsidence (TTS) at a margin and, hence, the geometry of the rifted crust for different assumed values of T_e. By calculating the 'sum' gravity anomaly due to the combined effects of the restored rifted crust (the 'rifting anomaly') and sediments (the 'sedimentation anomaly') and comparing it to observations, it is possible to constrain T_e at the margin. The effects of magmatic underplating and sub-aerial volcanism can be included in POGM by computing their flexural loading effect on both the TTS and the gravity anomaly.

Figure 3.4 shows a compilation of T_e estimates over passive margins based on POGM, as well as other forward models. The figure shows a wide range of values and there is no simple relationship between T_e and age since the initiation of rifting. However, when margins are considered that have been loaded at some stage during their evolution by a large discreet load, there is evidence that T_e may depend on the age. The South China Sea, for example, is a young margin that was loaded by thrust/fold loads in Taiwan during the Pliocene and has a low T_e while Northeast Brazil is an old margin that was loaded by the Amazon Cone deep-sea fan system during the late Miocene and has a high T_e. The increase appears to follow the depth to the 450°C isotherm and is in accord with the earlier suggestions of Karner and Watts (1982) based on spectral studies and of

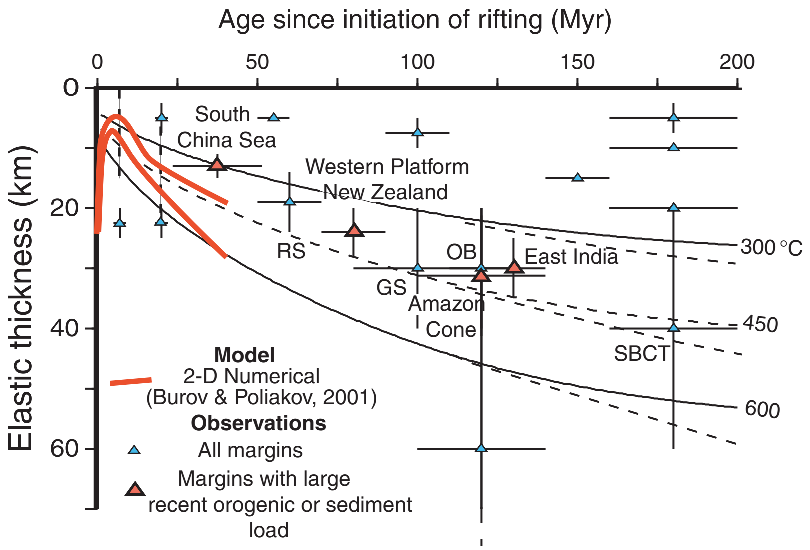

Figure 3.4 Plot of elastic thickness, T_e versus age since the initiation of rifting at passive margins. The T_e estimates are based on Table 6.3c in Watts (2001) with additional estimates from Cochran (1973), Lin and Watts (2002) and Krishna et al. (2000). The ages are also based on Watts (2001) except that ages for the Goban Spur, Jeanne d'arc, Nova Scotia and Grand Banks margins have been amended to 100, 180, 180 and 180 Ma respectively. RS, Ross Sea, GS, Goban Spur, OB, Orphan Bank (Canada), SBCT, southern Baltimore Canyon Trough (East Coast, USA). The solid and dashed black lines show the depth to the 300, 450 and 600°C isotherms based on the cooling oceanic plate model. The thick red lines show the predicted relationship between T_e and age since the initiation of rifting on the basis of the numerical models of Burov and Poliakov (2001).

Burov and Poliakov (2001) based on numerical models that stretched continental lithosphere may increase its strength with age as it cools following rifting.

There are, as Fig. 3.5 shows, a number of T_e estimates that are less than the depth to the 450°C isotherm. They suggest that if the sub-crustal (stretched) continental mantle gains strength following rifting, then there are processes that might act to weaken it. Possible processes include fluidisation of the sub-crustal mantle (e.g., Pérez-Gussinyé and Reston, 2001), thermal blanketing (e.g., Lavier and Steckler, 1997) and load-induced increases in curvature and, hence, yielding (e.g., Wyer, 2003).

Data from experimental rock mechanics suggest that the strength of the lithosphere is limited by brittle deformation and ductile flow (e.g., Kirby and Kronenberg, 1987). Moreover, they suggest that it might be possible to construct a Yield Strength Envelope (YSE) that is applicable to the lithosphere. Although the YSE has a number of limitations (Rutter and Brodie, 1991), it has proved a useful way to quantitatively evaluate the strength of the lithosphere for different strain rates, compositions and geothermal gradients and, importantly, tectonic stresses (including the stresses generated by flexure).

Unfortunately, the YSE does not explicitly define the strength of stretched continental lithosphere. This is because heating at the time of rifting increases the geothermal gradient and, hence, may reduce the strength of the lithosphere while crustal thinning replaces weak crust by strong mantle and, hence, strengthens it. There is therefore a 'competition' between heating that weakens the lithosphere and crustal thinning that strengthens it.

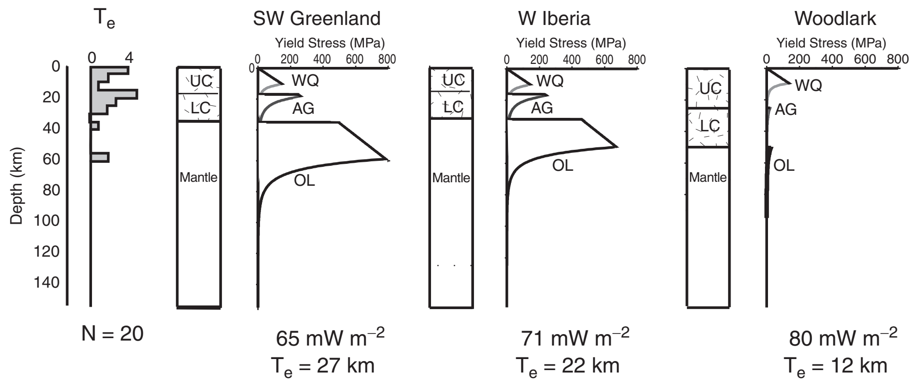

Figure 3.5 Comparison of the observed distribution of T_e at passive margins to calculations of the YSE at selected margins. T_e based on Table 6.3 in Watts (2001). Heat flow and YSE for the SW Greenland, W. Iberia and Woodlark passive margins based on Pérez-Gussinyé et al. (2001). The numbers below each YSE show the heat flow and the T_e estimated from the YSE using the method described by Burov and Diament (1995).

Irrespective of the amount of heating and thinning, YSE considerations suggest that the strength of stretched continental lithosphere will depend not only on the rheological properties of the initial, pre-rift lithosphere, but on its thermal state (e.g., Buck, 1991). Figure 3.5 shows, for example, the YSE for the southwest Greenland, west Iberia and Woodlark margins according to Pérez-Gussinyé et al. (2001). The Greenland and Iberia margins have relatively thin crust and are strong ($22 < T_e < 27$ km) compared to the Woodlark margin which has relatively thick crust and is weak ($T_e \sim 12$ km). A thicker initial crust is not the only reason, however, that the Woodlark margin appears weaker. The background heat flow, according to Pérez-Gussinyé et al. (2001), is high (Fig. 3.5) which implies a higher geothermal gradient and, hence, a lower T_e than for the Greenland and Iberia margins.

3.4 Models

These considerations of heat flow and elastic thickness data suggest that the thermal and mechanical structure of passive margins should be combined in some way. The thermal structure determines heat flow in the cooling basement and, hence, the amount of tectonic subsidence and uplift whereas the mechanical structure determines how it responds to sediment and other loads imposed on it during and following rifting.

Early attempts to combine the thermal and mechanical structure were based on relatively simple kinematic models. Watts and Thorne (1984), for example, used the uniform and depth-dependent extension models to estimate the thermal structure of the cooling basement and a Finite Difference model to compute the flexure. They linked the thermal and mechanical structure by assuming that T_e is given by the depth to a particular isotherm (450°C). The stratigraphy was calculated by assuming that the sediment load could be determined from the *difference* between an old sediment surface and a new surface, the depth to which could be estimated from the paleobathymetry (assumed fixed through time) and the relative height of sea level.

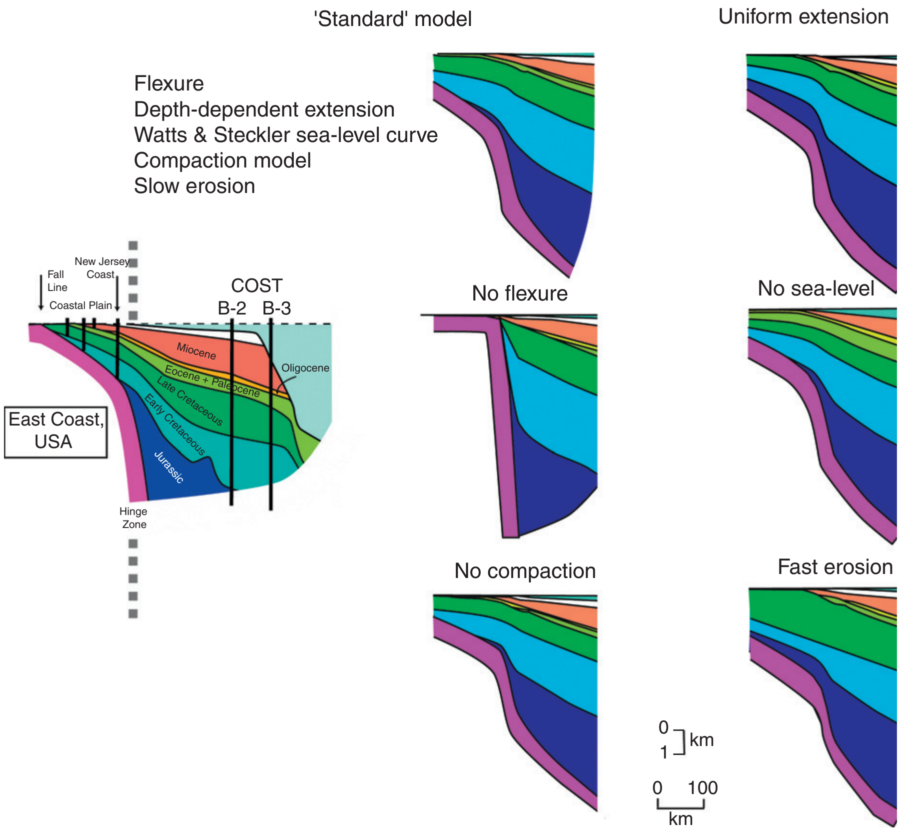

Figure 3.6 Comparison of the observed and calculated stratigraphy at the East Coast, USA, passive margin (modified from Watts and Thorne, 1984). The left panel shows the observed stratigraphy based on well and seismic data along a transect of the margin that intersects the coast near Long Beach, New Jersey. The middle and right panels show the calculated stratigraphy based on a model in which the thermal structure is defined by either a uniform or depth-dependent extension model and a T_e that is given by the depth to the 450° isotherm. The top section in the middle panel shows the 'standard model'. The other sections show the sensitivity of the 'standard model' to the different types of extension model, flexure, sea-level changes, erosion rate and compaction.

Figure 3.6 compares the observed stratigraphy at the East Coast, USA, margin to the calculated stratigraphy based on a combined thermal and mechanical model. The figure shows that the models explain well the overall stratigraphic 'architecture' of the margin. However, their real significance is in determining the role of the various input parameters that control the stratigraphic 'architecture' of the margin. The early history of the margin is determined, for example, mainly by the mode of extension (i.e., whether it is uniform or varies with depth) and flexure. Together, these tectonic controls determine whether or not Jurassic sediments underlie the coastal plain and the extent to which they onlap the basement. The later history, in contrast, is determined less by tectonics, and more by other factors, such as changes in sea level.

Similar models have subsequently been constructed by Lawrence et al. (1990), Kendall et al. (1991) and Reynolds et al. (1991). The main difference between them and earlier models was the inclusion of a variable sediment flux. Reynolds et al. (1991) assumed, for example, a constant continental shelf water depth

(0 km) and slope angle (2.2°) and showed that sediment flux determined the position of the Depositional Shelf Break (DSB) separating the shelf and slope. By tracking the DSB during evolution of the margin they were able to determine the role of sediment flux compared to that of other factors such as tectonics and sea level in determining the nature of the control on the development of the stratigraphic record.

Later models, especially those developed as part of the STRATAFORM programme (Steckler et al., 1996), have incoporated increasingly sophisticated ways of modelling the sediment dynamics. Steckler (1999), for example, included a shoreface which moved independently of the DSB. He showed that, unlike the DSB, the shoreface was more sensitive to first-order sea-level changes.

While the models discussed thus far have provided important insights into passive margin evolution, they are limited to a kinematic description of their thermal and mechanical structure. The structure in the Reynolds et al. (1991) model, for example, is based on a uniform elastic plate that ignores the effect of a vertically layered rheology. Weissel and Karner (1989) and Kusznir et al. (1991) constructed a thermal and mechanical model for rift-type basins that combined brittle deformation in the crust and ductile flow in the sub-crustal mantle. However, their models did not incorporate the YSE, which suggests (e.g., Fig. 3.5) one or more strength maxima in the pre-rift lithosphere.

The effect of a strength 'maxima' on the response of continental lithosphere to extension has been investigated by Kooi et al. (1992). They pointed out that the region of a strength 'maxima' would be difficult to deform and so acts to vertically partition the strain that results from rifting. The depth to a strong zone, dubbed by Kooi et al. (1992) as the depth of necking, determines the magnitude of a basin that overlies it and the uplift of the mantle that underlies it. Shallow depths predict a shallow basin and a large amount of mantle uplift whereas greater depths predict a deep basin and a small amount of mantle uplift. The isostatic response to such a crust/mantle geometry depends on the T_e during rifting. If $T_e = 0$, then the depth of the basin for a particular mantle uplift will be as predicted by an Airy model, irrespective of the depth of necking. If $T_e > 0$, however, then shallow depths of necking predict a broad flexural downwarp at a basin margin while deep depths predict flexurally supported rift flank uplifts.

Although the depth of necking model has been applied with some success (e.g., Kooi et al., 1992 – Gulf of Lyon margin; Keen and Dehler, 1997 – Canada and southwest Greenland margins; Watts and Stewart, 1998 – Gabon margin), Govers and Wortel (1999) have criticised it as well as other kinematic models. They preferred a Finite Element Method (FEM) to solve the coupled heat flow and mechanical equilibrium equations and showed that while T_e during rifting may indeed be finite, the depth of necking is not constant and varies both spatially and temporally. They used a brittle strength failure criterion to represent

Byerlee's law and an effective viscosity to simulate power law creep to show that the depth of necking at a rift axis only approaches a constant depth after a certain transient period (up to a few Myr). Moreover, they argued that the depth of necking does not actually coincide with the strength maxima, but rather occurs above it. Thus, in their view, the depth of necking parameter cannot be used in the same way as, for example, T_e in inferring information about rifted continental lithosphere.

Dynamical models of the type used by Govers and Wortel (1999) have, in fact, been applied to passive margins since the late 1980s (e.g., Bassi et al., 1993; Braun and Beamount, 1987; Dunbar and Sawyer, 1989; Harry and Sawyer, 1992). These models track temperatures using an FEM 'mesh' that includes the thermal lithosphere and at least part of the hot upwelling asthenosphere. The mechanical structure is approximated by experimentally constrained empirical laws for brittle and ductile deformation (e.g., Goetze and Evans, 1979) and isostasy is incorporated through a combination of gravitational body forces and forces on the base of the model that simulate buoyancy. Finally, forces on the side of the model simulate the in-plane extensional stresses.

Most dynamical models incorporate some sort of pre-rift weak zone (e.g., a thicker than normal crust or a thermal anomaly) that serves to focus the initial response of the lithosphere to extension. Despite this limitation, these models have provided useful constraints on the evolution of passive margin, especially the various factors that control the width of the transitional crust that separates unstretched continental and oceanic crust.

Bassi et al. (1993) showed, for example, that the width of stretched continental crust in passive margins depends strongly on the initial thermal conditions. Narrow margins (i.e., margins with widths of the stretched crust of up to 200 km), such as Flemish Cap, Canada, could be produced using models of a thick, cold, lithosphere, even after relatively small amounts of extension. In these models, the rapid localisation of strain is due mainly to the strong, plastic, part of the mantle necking almost to the point of rupture. The actual pattern of necking depends, however, on rheology, with a 'dry' rheology leading to a more rapid change in crust and mantle thinning than a 'wet' rheology. Wide margins (i.e., widths of up to 400 km), such as the Orphan Basin, Canada, could be produced using models of thin, hot lithosphere. In these models, there is a delocalisation of strain because of cooling at the rift axis that inhibits thinning and causes necking to take place along the edges of the deformed lithosphere.

During the past 8 years there has been a rapid development in both numerical (e.g., Bassi et al., 1993; Buck et al., 1999; Burov and Poliakov, 2001; Hopper and Buck, 1996; ter Voorde et al., 1998) and analogue (e.g., Brun and Besslier, 1996) models of rift basin evolution. Of particular interest have been models that determine the role of crust and mantle coupling, lower crustal flow and crustal buoyancy in controlling the distribution of stretched crust, structural styles and the degree of along-strike segmentation at rifted margins.

Buck et al. (1999), for example, considered the forces involved in extension and the relative role of processes that serve to localise rifting and those that de-localise it. They argue that narrow rifts do not need a preexisting weakness and that necking, magmatic addition and cohesion loss on faults are all ways that rifting may be localised in narrow zones. Wide rifts, on the other hand, are produced by de-localising effects such as viscous flow, crustal buoyancy and flexure.

The role of buoyancy forces is a potentially important one given their association with the abrupt changes in the thickness and, hence, density of the crust and mantle that occur at passive margins. As pointed out by Newman and White (1999), lateral density variations generate buoyancy forces that may either oppose or enhance extension. These authors showed that a relationship exists between the amount of stretching and strain rate in continental rift basins. This suggests that extension ceases because the lithospheric mantle cools and strengthens following a rifting event, rather than because the extensional stresses have been removed.

Davis and Kusznir (2002) showed that despite their higher stretching factors (up to $\sim$3.5) and strain rates (up to $\sim 10^{-14}\ \mathrm{s}^{-1}$) a similar relationship exists between the amount of stretching and strain rate at passive margins as at continental rift basins. They took into account crustal and thermally driven buoyancy (dubbed by them ‘rift push’). As crustal buoyancy forces oppose extension and ‘rift push’ forces enhance it, they argued that thermally driven buoyancy forces may dominate over crustal buoyancy forces.

While the models discussed thus far take into account the brittle and ductile deformation of rocks, they are mostly based on a single- rather than a multi-layer rheology. As pointed out by a number of workers (e.g., Burov and Diament, 1995) continents, unlike the oceans, comprise of more than one strong competent layer separated by weak layers. Multi-layer rheology lithosphere responds in fundamentally different ways to tectonic stresses than single-layer lithosphere.

ter Voorde et al. (1998) showed that a multi-layer rheology implies a different degree of coupling between the various competent layers. They showed that during rifting, the upper crust, lower crust and mantle could be either ‘fully coupled’, ‘partly coupled’, or ‘fully decoupled’, depending on the temperature structure in the lithosphere, especially the lower crust. The degree of coupling determines the thickness of the layer that supports flexural loading or unloading: fully decoupled indicates that the load is supported by the strength of the upper crust and that compensation takes place, at least in part, in the weak lower crust whereas fully coupled indicates that the lower crust is sufficiently strong that the load is supported by *both* the upper crust and the mantle and is compensated at greater depths.

ter Voorde et al. (1998), while discussing passive margins, suggested that the Bay of Biscay was an example of partially coupled margin, as is evidenced by the relatively short-wavelength (wavelength, λ, $\sim$30–40 km) of its syn-rift grabens and their associated hanging-walls and footwall uplifts, and the relatively long-wavelength ($\lambda > 80$ km) undulations of the underlying Moho.

A multi-layer rheology also implies, as Burov and Poliakov (2001) have pointed out, certain feedbacks, which involve a lower crust that either accelerates or retards subsidence during the syn-rift and post-rift. For example, while crustal thinning due to sub-aerial erosion increases subsidence, it decreases the distance between strong, competent, elastic 'cores' of the upper crust, lower crust, and mantle. This would cause T_e to increase which would arrest the subsidence. Other feedbacks could occur during the post-rift. For example, yielding during sediment loading would increase subsidence whereas the replacement of weak crust by strong mantle would decrease it. If such effects balance, then the subsidence history of a margin may be in accord with predictions of simple kinematic models such as 'pure shear' and 'simple shear'. Otherwise, they would disagree, which might help explain the observations at some margins of syn-rift sequences that greatly exceed the thickness of the post-rift (e.g., Grand Banks, Canada) and of post-rift sequences that greatly exceed the thickness of the syn-rift (e.g., East Coast, USA).

While the introduction of multi-layer rheologies and their associated feedbacks significantly complicates numerical models, advancements in computing may make it possible to take them into account in the future. An example of one such model is illustrated in Fig. 3.7. The main difference between this and previous models is that *both* the Brittle–Ductile Transition (BDT) and the Crust Mantle Boundary (CMB) can be tracked during rifting. Therefore, the pressure-temperature history of the crust (and lithosphere) can be determined.

One problem with the current generation of numerical models is that it has proved difficult to simulate faulting. Although several workers have traced faults through strain localisation (e.g., Behn et al., 2002), it is difficult to use numerical models to resolve between different extensional models (e.g., 'pure shear' and 'simple shear'). Laboratory experiments on brittle and ductile materials are unable to incorporate the temperature dependence of rock rheology though they have been useful to examine the structural styles that develop during extension, including core complexes.

Brun and Besslier (1996), for example, have used sand and silicone putty to represent the brittle and ductile layers of the lithosphere and a low-viscosity syrup to represent the underlying asthenosphere. They showed that by applying displacements at a constant rate along the edges of the model, necking of a multi-layer rheology lithosphere is nearly symmetric (i.e., 'pure shear'), but that asymmetrical structures develop internally, for example, because of faulting. As displacements increase and the brittle and ductile layers thin, lower crust

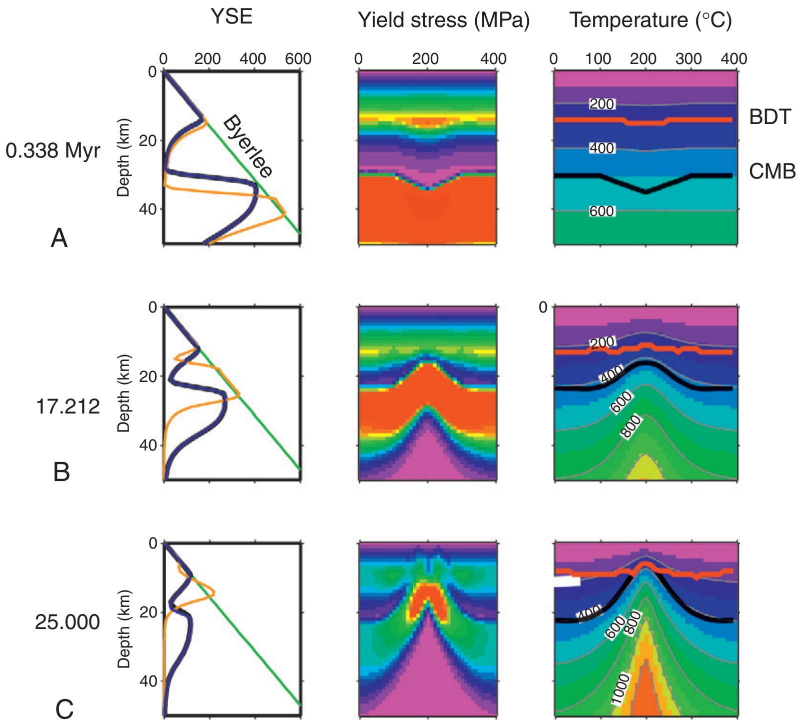

Figure 3.7 Numerical model for the response of the continental lithosphere to extension (Pérez-Gussinyé, pers. comm.). The model is based on a spatially and temporally varying multi-layer rheology and takes into account the heat consumed during rifting by melting. YSE, Yield Strength Envelope. Blue line = YSE at 100 and 300 km. Orange line = YSE at rift centre. Green line = Byerlee's law. BDT = Brittle–Ductile Transition. CMB = Crust/Mantle Boundary. A – 0.3 Myr after initial rifting. Extension is assumed to nucleate in a site of a previous weakness, simulated, in this case, by thicker than normal crust. B – 17.2 Myr. The CMB rises as extension proceeds. There is little change, however, in the depth to the BDT. C – 25.0 Myr. The CMB intersects the BDT such that the strong, cold, brittle upper crust now overlies weak, hot, upwelling mantle. Pérez-Gussinyé (pers. comm.) has used such models to study the onset of serpentinisation since it is likely to occur once faults which act as conduits for fluids can pass entirely through the crust and into the mantle.

and mantle may become juxtaposed. During extreme stretching, boudinage in the high strength sub-Moho layer allows ductile mantle to come into contact with ductile lower crust, leading to the exhumation of ductile material.

The occurrence of exhumed mantle rocks in the west Iberia margin has been interpreted as indicative of 'simple shear' with a detachment fault that cuts either the entire lithosphere (Boillot et al., 1988) or crust (Whitmarsh et al., 2001). However, multi-layer analogue models do not show any evidence of such a large-scale fault (Brun, 1999). Moreover, seismic reflection profile data over passive margins show little evidence of the prominent reflectors expected of detachment faults.

A notable exception is the 'S-type' reflector observed on the north Biscay and Galicia Bank margins (de Charpal et al., 1978). Previous studies suggested that the reflector corresponds to the BDT, a detachment fault that penetrates the entire lithosphere, an intra-crustal detachment fault, a detachment between the upper crust and the serpentinised mantle. The most probable interpretation, however, (e.g., Louden and Chian, 1999; Reston et al., 1996) is that it marks the transition between a low velocity, highly faulted, upper crust and high velocity serpentinised peridotite.

One puzzling observation is that the width of the exhumed mantle at the west Iberia margin varies along-strike, being 10–20 km wide west of Galicia Bank (Pickup et al., 1996) and >100 km wide beneath the southern Iberia Abyssal Plain (Sibuet et al., 1995). Analogue modelling shows that wide regions of exhumed mantle can be produced if a low viscosity heterogeneity is placed below the lowest brittle-ductile interface in the central part of a developing rift (Brun, 1999). Such viscosity heterogeneities could arise from a zone of partial melting.

Callot and Geoffrey (2002) introduced heterogeneities (made of silicone putty) in their analogue models in order to simulate melting zones that locally weaken the extending lithosphere. They showed that such 'soft spots' within the brittle layer act to localise the strain and may control the along-strike magmatic segmentation that is observed in some passive margins (e.g., southwest Greenland – Geoffroy, 2001; and East Coast, USA – Behn and Lin, 2000).

The dynamical models discussed thus far have mainly been concerned with how the lithosphere responds to extension *during* rifting. While this is of fundamental importance to understanding the mechanisms of continental break-up, it sheds little light on the thermal and mechanical processes that occur during the post-rift when the margin is deformed by sediment and other types of loads.

An exception is the study of Kjeldstad et al. (2003) who used a numerical model to simulate the progradation of sediment loads across the Vøring Plateau margin, Norway, during the Pliocene–Pleistocene. They showed that the Helland Hansen Arch, a prominent anticlinal ridge in the outer part of the Vøring basin, could be explained, at least in part, by differential loading that pushes sediment downwards and laterally towards the front of the prograding wedge. Their conclusion, however, appears to be in conflict with recent analogue models for the arch. Leroy et al. (2004), for example, suggest that the Fles Fault Complex (which is located just to the north of the Helland Hansen Arch) is the consequence of incipient shortening, for example, because of 'ridge push' acting on preexisting lines of weakness.

3.5 Conclusions

We draw the following conclusions from this review:

1. The progressive backstripping of sediments though geological time reveals that the main factors controlling the subsidence of passive margins are sediment loading and thermal contraction following heating, thinning and cooling at the time of rifting.
2. Kinematic models suggest that tectonics (in the form of thermal contraction and uplift and flexure) are a major control on the early stratigraphic evolution of passive margins. Later stages, however, appear to be dominated more by changes in sediment flux and/or sea level than by tectonics.
3. The heat flow at young margins is generally higher than at surrounding cratonic areas whereas at old margins it is either similar or less. This suggests that the sub-crustal mantle at passive margins progressively cools with time.
4. The elastic thickness at young margins is generally lower than that at old margins, although there are a number of low values. This suggests that while the sub-crustal mantle at passive margins may increase in strength as it cools, other processes such as fluids, increases in loading, curvature and yielding and sediment blanketing act to weaken it.
5. Numerical and analogue models have provided new constraints on passive margin evolution, most notably on the factors that control their width, structural style and segmentation.
6. While preexisting zones of weakness act to localise strain in rifts, tectonic stresses appear to be large enough to cause stable lithosphere to neck almost to the point of rupture.
7. Necking is modulated by processes such as magmatic intrusion, which localise strain and processes such as viscous flow, crustal buoyancy and flexure, which delocalise it.
8. The net result is that while most conjugate passive margin pairs show a high degree of symmetry in their large-scale crust and mantle structure, they also show asymmetry, especially in their faulting styles, stratigraphic architecture and thermal and mechanical properties.

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