Basin Research (1992) **4**, 169–178

The effective elastic thickness of the lithosphere and the evolution of foreland basins

A. B. Watts
Department of Earth Sciences, University of Oxford, Parks Road, Oxford OX1 3PR, UK

ABSTRACT

The elastic thickness of the continental lithosphere is strongly 'bi-modal'. Foreland basins reflect this bi-modality with narrow, deep basins (e.g. Apennines) and wide, shallow ones (e.g. Ganges). The bi-modal distribution cannot be explained by thermal models which describe the relationship between elastic thickness and plate age in the oceans, suggesting the involvement of factors other than secular cooling. The high values (80–90 km) are consistent with present-day temperature gradients of the cratons and the scatter within cratons may be explained by changes in the radiogenic heat production. The low values (10–20 km), however, are more difficult to explain. Foreland basins develop by flexure in front of thrust/fold loads as they advance over former passive margins and onto the cratons. Recent studies suggest that passive margins are underlain by highly attenuated crust and lithosphere which has a low elastic thickness that remains low for long periods ($> 10^8$ yr) of time. Foreland basins may inherit these low values as they migrate over a passive margin. Stratigraphic modelling suggests that the low elastic thicknesses would have a profound effect on the development of foreland basins predicting as they do the asymmetry, the pattern of onlap and, the transition from the 'underfilled' to 'overfilled' phase. Why stretched crust and lithosphere is so weak on long time-scales is enigmatic. Rifting, however, seems to proceed in such a way that the strong uppermost part of the crust is effectively 'de-coupled' from any support that it might otherwise receive from the strong underlying mantle.

INTRODUCTION

Lithospheric flexure due to sediment loading is one of the primary factors that determines the overall 'architecture' of sedimentary basins. In the case of basins which form by downsagging of the lithosphere due to orogenic loading, flexure controls both their width and depth. Narrow, deep basins such as the Apennine, Ebro and Swiss Molasse are associated with low values of flexural rigidity while wide, shallow basins such as the Appalachian and Ganges yield high values. The role of flexure in the development of rift-type basins, is, however, more controversial. Models which assume that the flexural rigidity of the lithosphere increases with age can account for the dominance of onlap during the post-rift phase of their evolution. Gravity anomalies over the North Sea and Baltimore Canyon Trough basins, however, are small, and require a low value of flexural rigidity in order to explain them. This suggests that rift-type basins originate either on a weak lithosphere that is unable to recover its strength, or on strong lithosphere that is weakened by sediment blanketing, anelastic or other effects. Irrespective of the actual cause, large lateral changes would be expected in the flexural rigidity across a stretched basin. Plate tectonic models suggest that foreland basins develop on the site of former passive margins. The purpose of this paper is (a) to review current knowledge of the elastic thickness (which is determined by the flexural ridigity) of oceanic and continental lithosphere and (b) to examine its implications for better understanding the role of flexure in the development of foreland basins.

LITHOSPHERIC FLEXURE

Oceanic T_e

One of the successes of past oceanic flexure studies has been to demonstrate that the lithosphere responds to long-term ($> 10^6$ years) geological loads as an elastic plate that increases its strength with age (Watts 1978). Volcanic loads that form on or near a mid-oceanic ridge crest, for example, are associated with low values of the elastic plate thickness, T_e, while those that form on the ridge flanks have higher ones. Oceanic flexure studies (e.g. Fig. 1a) show that T_e is given approximately by the depth to the 450°C oceanic isotherm suggesting that as they cool the plates progressively increase their strength with age. In addition, they have demonstrated that because oceanic loads range in age from a few Myr to several tens of Myr, T_e is in effect 'frozen in' *at the time of loading* and does not change significantly with time.

Recent studies (e.g. Fig. 1b) have confirmed the increase in strength of oceanic lithosphere with age but, have shown some complications with the simple model. Calmant & Cazenave (1986), for example, have shown that seamounts and oceanic islands in the French Polynesia region are associated with a significantly *smaller* T_e than predicted. The low values correlate with a broad region of relatively shallow sea-floor, low seismic P-wave velocities and, isotopic anomalies which McNutt & Fischer (1987) termed the Pacific 'superswell'. Calmant & Cazenave (1986) suggest that the oceanic lithosphere is capable of being significantly weakened by regional heating from sources within or below the lithosphere, effectively resetting the thermal age of the plate. Deep sea trenches (McNutt & Menard 1982) and large-offset fracture zones (Wessel & Haxby 1990), on the other hand, reveal T_e values that are significantly *larger* than are predicted. McNutt (1984) questioned whether the trench vales were high, suggesting instead that it was the volcano values that were low because of local heating of the crust and lithosphere during magma ascent. While this explanation may account for some of the differences between the trench and volcano values, it is difficult to invoke at Hawaii which, despite being located near the crest of a broad mid-plate swell, has a T_e (Watts & ten Brink 1989) and heat flow (Von Herzen *et al.* 1989) that is close to what would be expected on the basis of plate age.

An alternative explanation has been advanced by Wessel & Haxby (1990). These authors suggest that the differences between oceanic island/seamount and other T_e values can be explained by the thermal stresses that arise in the oceanic plates due to non-uniform cooling. Thermal stresses generate a bending moment in the plate which will interact with the bending stresses due to flexure. Wessel & Haxby (1990) used a yield stress envelope technique to model the failure of the lithosphere due to the combination of thermal and flexural stresses. They showed that the curvature of the flexed plate beneath seamounts is sufficient to make the plate appear weaker than it actually is whereas at trenches the opposite effect occurs. Thermal stresses therefore produce effects that are of the right sign to reduce the discrepancy between ocean island/seamount and trench values. However, Wessel (1992) has shown recently that they may explain only part of the magnitude of the difference because the effect only seems significant at young ages.

Despite re-heating and bias due to thermal stresses, oceanic T_e values show (e.g. Fig. 2) a generally uniform distribution with most values in the range 10–20 km. This is attributed to the fact that irrespective of load size or type, oceanic T_e is controlled by the *same* mechanism of secular cooling of the lithosphere following its formation at a mid-oceanic ridge crest. Phenomena such as re-heating and thermal stresses do not modify this distribution significantly since they themselves are also controlled, to some extent, by the same plate cooling mechanisms as determine T_e.

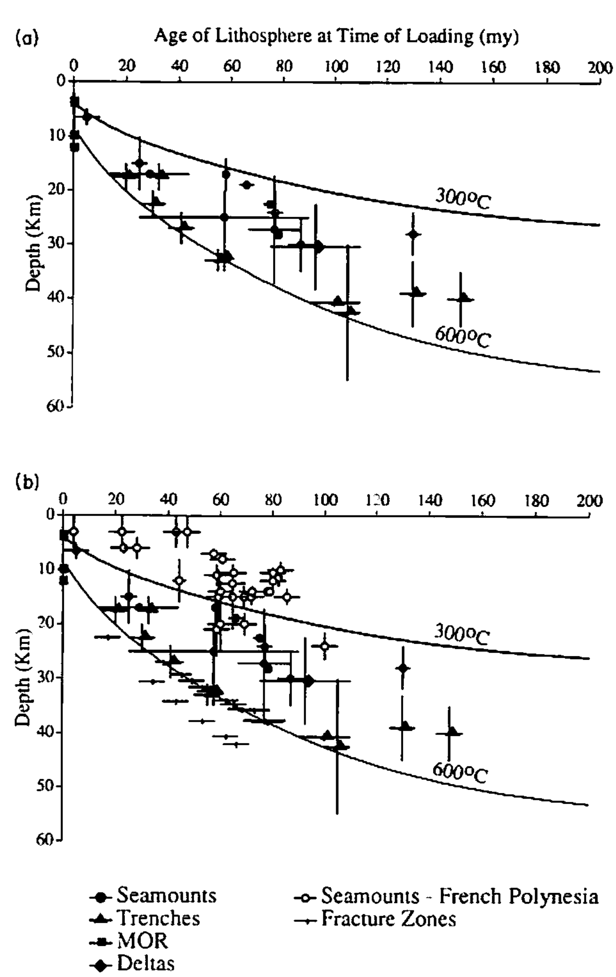

Fig. 1 Compilation of T_e against age of the oceanic lithosphere at the time of loading. (a) All values *excluding* fracture zones and seamounts/oceanic islands in the French Polynesia region of the central Pacific. (b) All values. Data sources are as given in Smith *et al.* (1989) with additional points from Kuo & Parmentier (1986), Watts *et al.* (1988), Wessel & Haxby (1990) and Calmant & Cazenave (1986). The solid lines show the depth to the 300°C and 600°C isotherms based on a cooling plate model (Parsons & Sclater 1977) with equilibrium thermal thickness of 125 km.

Continental T_e

Although Karner *et al.* (1983) have proposed that secular cooling may also control continental T_e, the case is not as clear as it is for the oceans. One reason is that surface loads are not as well preserved in the continents because of the modifying effects of denudation, tectonic erosion, rifting and orogenisis. These processes may either remove or add loads and, in some cases, impose new loads *within* the crust in such a way so as not to give a surface expression. This has led to some doubt about the significance of continental

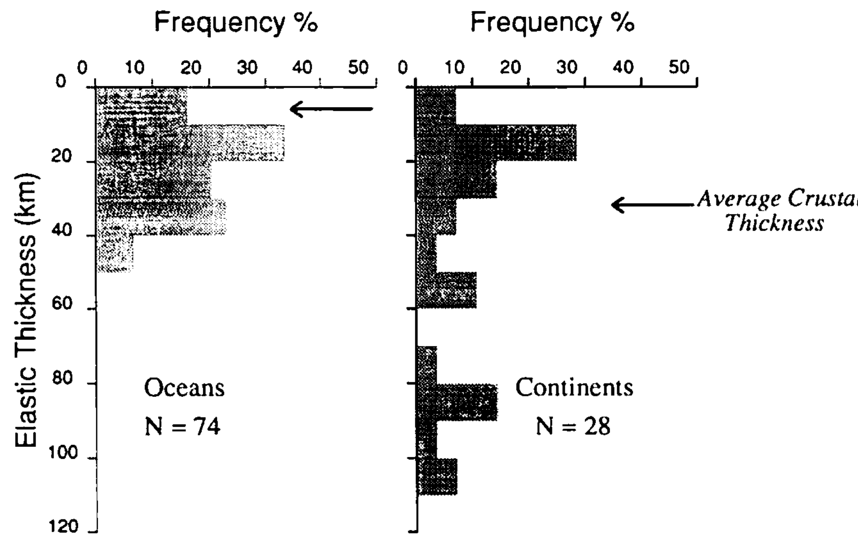

Fig. 2 Frequency plot of oceanic and continental T_e estimates. The oceanic values are based on the data sources in Fig. 1b and the continental values on values from foreland basin and late-glacial rebound studies (Fig. 4). Both loads are considered to be old enough (> 10,000 years) to 'sample' the long-term mechanical properties of the lithosphere. The arrows indicate the average thickness of the oceanic and continental crust based on regional compilations of seismic refraction data.

T_e estimates, especially those (e.g. Dorman & Lewis 1970; Lewis & Dormon 1970; McNutt & Parker 1978) based on spectral studies of gravity anomaly and topography data which do not take into account the effects of buried loading.

If the discussion is limited to foreland basins and late glacial lakes where the data are sufficiently good (e.g. raised beaches, Bouguer anomaly 'highs') to define the load geometry then, continental T_e values (Fig. 2) show a number of differences with the oceans. First, they appear to have a 'bi-modal' distribution with a primary peak at 10–20 km and a secondary one at 80–90 km. Secondly, they display a wide range of values (5–110 km). These observations are difficult , as pointed out by McNutt *et al.* (1988), to explain in terms of a *single* secular cooling model. For example, Fig. 3 shows that an oceanic cooling type model (equilibrium thermal thickness of the lithosphere = 125 km) can account for some continental T_e values but, a single set of thermal parameters is unable to explain both the low values at young 'apparent' ages *and* the high ones at old ages. This suggests that factors other than secular cooling play a role in controlling continental T_e.

One possible factor is composition such that, unlike the oceans, there is a relationship between continental T_e and the *present day* temperature gradient. Pinet *et al.* (in press), for example, used heat flow together with other geophysical data to estimate the gradient in the Grenville and Superior province of the North American craton. They showed that the sub-crustal gradients are low enough that the depth to the 450°C isotherm would be about 70 km which is in general agreement (Fig. 4) with estimates of T_e in the region. Thus, cratons may appear relatively strong. This is consistent with yield strength envelope studies that are based (e.g. Vink *et al.* 1984; Steckler & ten Brink 1986) on a combination of Byerlee's linear frictional law and, quartz and olivine flow laws for the crust and mantle respectively. Quartz deforms by ductile flow at lower temperature than Olivine which has led to the suggestion (e.g. Vink *et al.* 1984) that continents are weaker than oceans. However, this conclusion depends on the temperature gradient in the oceans and continents being the same (Steckler & ten Brink 1986). As Fig. 5 shows, the temperature gradients determined by Pinet *et al.* (in press) are low enough that they would lead to more involvement of the Olivine layer in the support of surface loads than would be the case for high gradients. In fact, the secondary peak in the bimodal distribution of continental T_e agrees well (Fig. 5) with the depth depends not only on the strain rate but also on the size of the load.

Other factors that are related to composition such as lateral variations in radiogenic heat production may help explain some of the scatter that is observed within cratonic regions and between the cratons and the adjacent stabilized orogenic belts. Pinet *et al.* (in press) have argued, for example, that the higher heat flow that is observed over the Appalachians than the North American craton (Fig. 4) can be accounted for almost entirely by an increase in the thickness of the heat-producing crustal layers, especially a lower tonalitic layer. The temperature that is implied for the crust beneath the Appalachians is about 100°C higher than beneath the craton suggesting a T_e which is about 25 km lower. This variation in T_e is in agreement in both sign and magnitude with those estimated by Bechtel *et al.* (1990) based on spectral studies (Fig. 4), thereby providing further support to the suggestion that composition, rather than secular cooling, may play a major role in determining continental T_e.

FORELAND BASINS

It is generally agreed that one of the best measures of continental T_e is the geometry of foreland basins that develop in front of advancing thrust and fold loads (e.g.

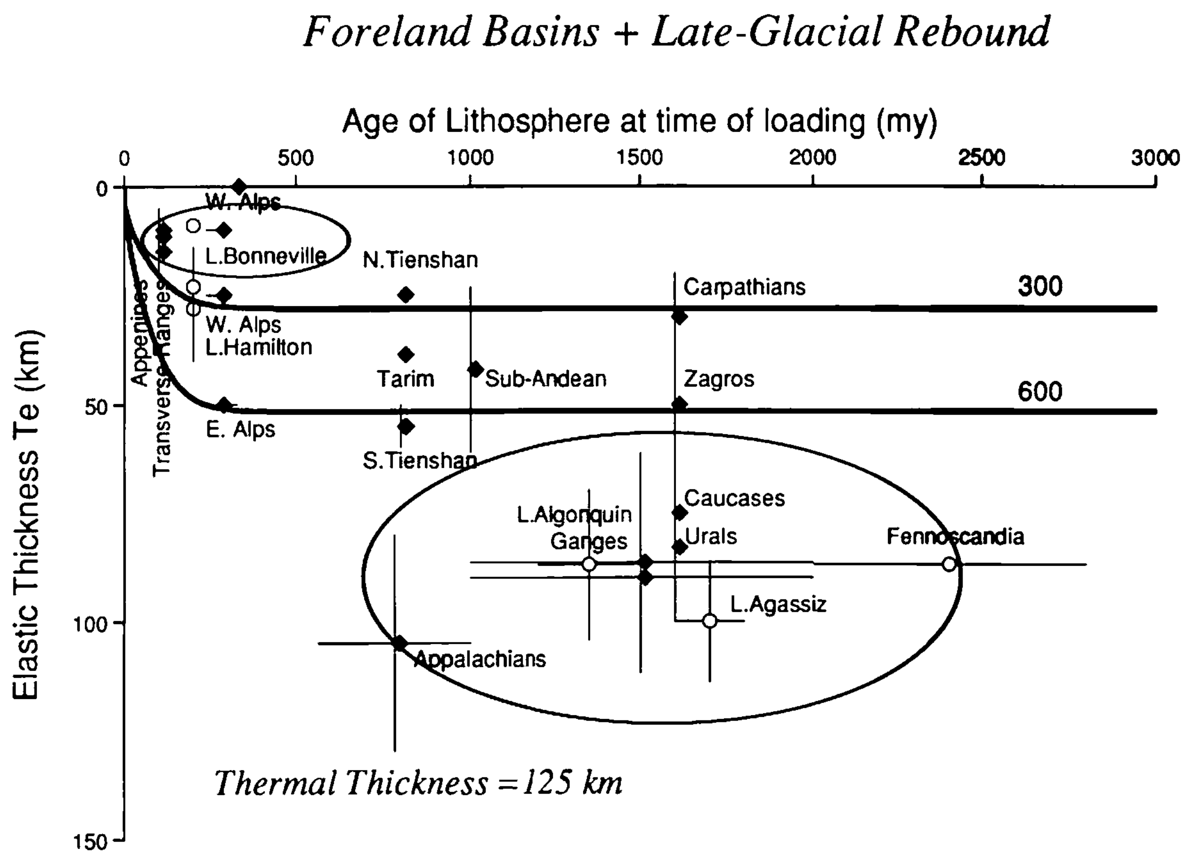

Fig. 3 Compilation of T_e against age of the continental lithosphere at the time of loading. Data sources are: (a) *Foreland basins*: Karner & Watts (1983) – W/E Alps, Appalachians, Ganges; Lyon-Caen & Molnar (1983) – Ganges, Tarim, Sub-Andean; Royden & Karner (1984) – Carpathians, Apennines; Snyder & Barazangi, (1986) – Zagros; Sheffels & McNutt (1986) – Transverse Ranges; Royden (1987) – Apennines; Rupple & McNutt (1987) – Caucasas; Kruse & McNutt (1988) – Urals; Burov *et al.* (1990) – Pamir, N/S Tien Shan; – 'Ref' W. Alps. (b) *Late glacial rebound*: Walcott (1970) – Lake Algonquin, Lake Agassiz; McConnell (1968) – Fennoscandia; Fulton & Walcott (1975) – Lake Hamilton; Crittenden (1970) – Lake Bonneville. The heavy lines indicate the depth to the 300°C and 600°C isotherm based on a cooling plate model with equilibrium thermal thickness of 125 km. The ellipses show 'groupings' of the data that are either too low to be explained by the cooling model or too high.

Beaumont 1981; Jordan 1981; Karner & Watts 1983; Stockmal *et al.* 1986; Sinclair *et al.* 1991). Simple models (Menke 1981) show that irrespective of load shape, T_e controls the width and, to a lesser extent, the depth of a foreland basin. In general, foreland basin geometry reflects the bi-modality of the continental T_e values with narrow, deep basins such as the Appennines (Royden & Karner 1984) and Ebro (Zoetemeijer *et al.* 1990) indicating low T_e and wide, shallow ones (> 250 km) such as Ganges (Karner & Watts 1982) and Appalachian (Turcotte & Schubert 1982) a high T_e. According to Kominz & Bond (1986), T_e in foreland basins is an inherited feature which is not modified significantly by orogenic loading.

Evolution from passive continental margin to foreland basin

Plate tectonic models (e.g. Dewey 1982; Houseknecht 1986; Tankard 1986) suggest that foreland basins develop by flexure in front of thrust/fold loads as they advance across former passive margins. The palinspastic restoration of mountain belts (Pfiffner 1986), the analysis of back-stripped subsidence curves (Homewood *et al.* 1986) and deep seismic reflection profile data (Cook *et al.* 1981) suggest a close association between foreland basins and passive margins. Watts (1981) suggested that during orogeny the hinge zone (Watts & Steckler 1979) may act as a ramp which serves to 'telescope' individual thrust sheets over and onto the coastal plain of the underlying passive margin. Stockmal *et al.* (1986) were the first to develop a numerical model for thrust loading of the stretched crust of a former passive margin. They assumed that the strength of stretched crust increases with age such that if thrusting developed on thermally young lithosphere it would inherit a relatively low T_e while if it loaded thermally older lithosphere it would develop on stronger lithosphere. They found, however, that the T_e of the stretched crust only slightly affects foreland basin geometry. This was because they used a model in which T_e is determined by the depth to the 750°C isotherm which implies that the strength of stretched lithosphere is greater

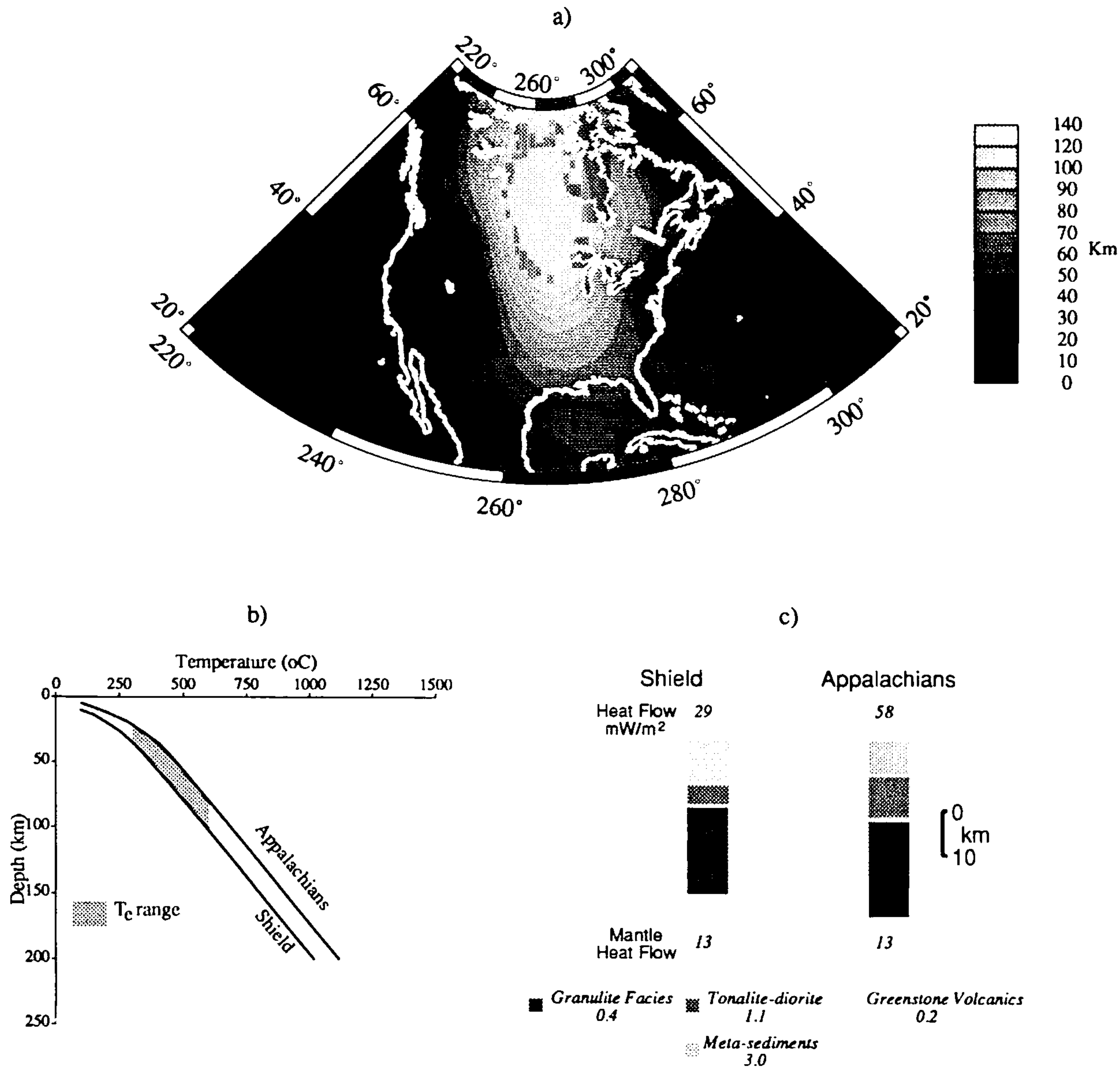

Fig. 4 Summary of principal results of T_e and heat flow studies in North America (Bechtel *et al.* (1990); Pinet *et al.* (in press)). (a) T_e map of North America based on data sources in (Bechtel *et al.* 1990) with additional values by Watts (1988). In oceanic regions it is assumed that T_e is given by the thermal structure of the plate. (b) Temperature gradient for the Appalachian and Grenville/Superior province (i.e. shield areas) according to Pinet *et al.* (in press). The gradients have been calculated from the measured heat flow and by estimating the relative contributions of the heat-producing layers in the crust. (c) Crustal structure assumed by Pinet *et al.* (in press) in calculating the temperature gradients in (b). The shaded region in (b) indicates the expected range values of T_e assuming it to be controlled by the depth to the 300–600°C isotherm.

than oceanic lithosphere of the same thermal age. A high T_e is therefore acquired relatively quickly following rifting in the Stockmal *et al.* (1986) model and as a result the T_e of young and old passive margins do not differ significantly.

Passive margin T_e

Several studies (e.g. Fowler & McKenzie 1989; Watts 1988) have argued, however, that stretched lithosphere is of significantly *lower* strength than oceanic lithosphere. Beaumont *et al.* (1982) and Karner & Watts (1982) showed that the best fit to observed free-air gravity anomaly data at the Nova Scotia, South Africa and Coral Sea passive margins is a model in which T_e follows a lower, rather than higher, controlling isotherm (250°C, 150°C and 150°C, respectively) than the oceanic lithosphere (450°C). A lower controlling isotherm might be interpreted in terms of the average response of the margin with the earliest sediments loading a weak lithosphere and later sediments a stronger one. Fowler & McKenzie (1989) have argued, however, from the sediment 'starved' Rockall and Exmouth Plateau margins that T_e of stretched crust is fundamentally low

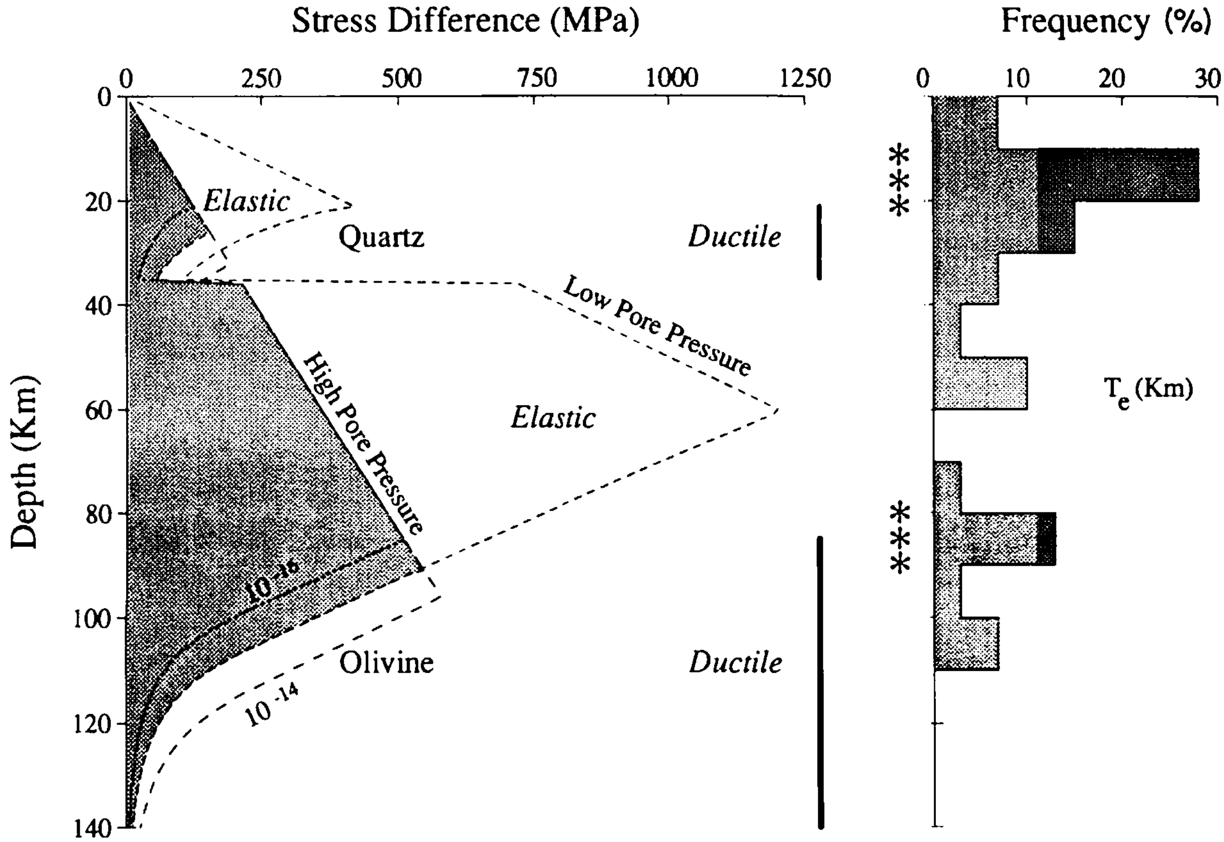

Fig. 5 Yield strength envelope for the continents based on data from experimental rock mechanics compared to the continental T_e estimates. The envelope shows the limits of extensional stress that the crust and mantle can support assuming: Byerlee's linear frictional law for low (0.0) and high (0.7) ratios of pore to lithostatic pressure from Brace & Kohlstedt (1980), the flow laws for wet and dry quartz from Koch *et al.* (1990) with activation energy, Q, for wet and dry quartz of 160.7 kJ mol^{-1} and 151.7 kJ mol^{-1}, respectively, the Dorn and Power laws for olivine from Goetze & Evans (1979) with $Q = 441$ kJ mol^{-1}, the temperature gradient for shields shown in Fig. 4b which is based on Pinet *et al.* (in press) and a crustal thickness of 35 km. The flow laws have been calculated for assumed strain rates of 10^{-14}, 10^{-15} and 10^{16} s^{-1} and values for the universal gas constant, R, of 8.32 J °K^{-1} mol^{-1}. The shaded region shows the envelope for a strain rate of 10^{-15} s^{-1} and a high (0.7) pore to lithostatic ratio. Solid vertical lines show the ductile region and the asterisks indicate the peak values in the bi-modal distribution of continental T_e estimates.

and of the order of 5 km. Their results are in accord with studies of intra-cratonic rifts such as the North Sea (Barton & Wood 1984) and the Michigan basin (Nunn & Sleep 1984). Watts (1988) carried out an integrated study of seismic and gravity anomaly data at the East Coast, US margin and showed that the amplitude and wavelength of the free-air 'edge effect' anomaly could not be explained unless the stretched crust was weak as far landward as the hinge zone and as far seaward as the ocean–continent boundary. When the results from this study are combined with the results of spectral studies landward of the hinge zone and oceanic flexure studies seaward of the stretched crust (e.g. Fig. 4), then they suggest large *lateral* variations in T_e are possible across a passive margin.

The existence of weak stretched crust is not inconsistent with the observation that rift-type basins are characterized by progressive onlap of strata onto the basement during their post-rift development. Flank uplift, for example, may restrict the early distribution of sediments in a basin due either to a relatively high initial T_e (Weissel & Karner 1989), thermal effects (White & McKenzie 1988) or some form of small-scale convection in the mantle (Steckler 1988). As the uplifts gradually disappear due either to erosion, thermal subsidence or, convective decay then sediment is gradually deposited over a broader area.

Stratigraphic models

If passive margins are characterised by large lateral changes in T_e, as is suggested here, then the stratigraphy of foreland basins which develop in front of advancing thrust/fold loads should reflect these changes as they migrate across them. Figure 6 shows, for example, a simple model for the

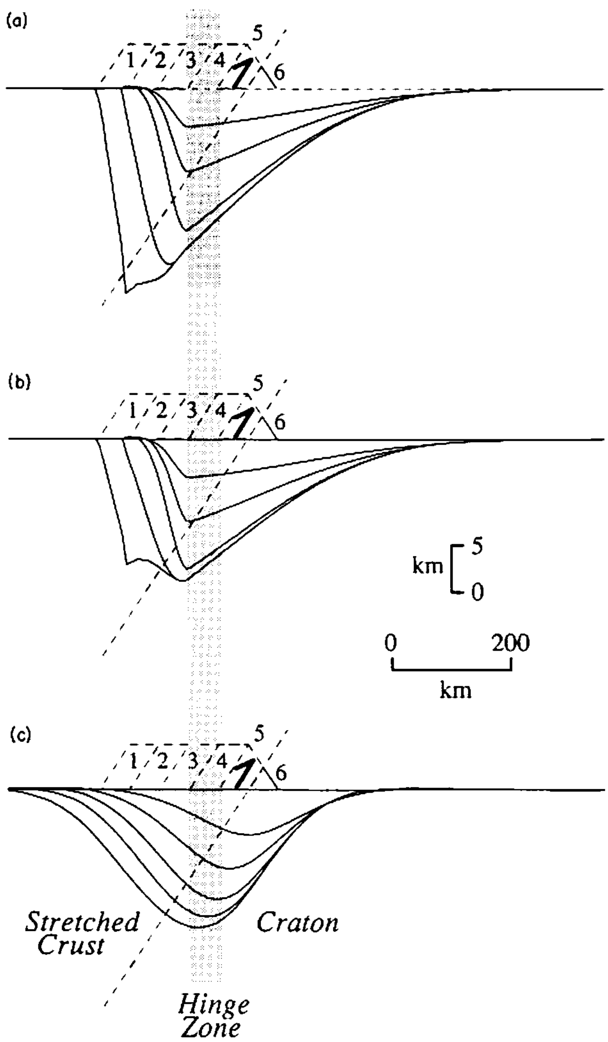

Fig. 6 Plot of the stratigraphy for a simple model of thrust/fold loads that migrate across the surface of the lithosphere into 6 'sub-loads' of density 2700 kg m^{-3}. (a) Constant load that displaces air on a variable T_e plate. The T_e has been chosen to represent a passive margin with a relatively low value ($T_e = 5$ km) over stretched crust and a higher one ($T_e = 75$ km) over the craton. T_e is assumed to change linearly across a 75 km wide hinge zone which separates stretched crust from an unstretched craton. (b) Same as in (a) except that the first two loads are assumed to displace sea-water (density = 1030 kg m^{-3}) while the last three loads displace air (density = 0 kg m^{-3}). (c) Same as in (b) except that T_e is assumed to be constant ($T_e = 75$ km) across the plate. The heavy arrow indicates the direction of thrusting and the shaded region shows the location of the passive margin hinge zone that separates stretched crust from the craton.

stratigraphy of the foreland in which it is assumed that loads move at a constant rate and that T_e of the underlying margin resembles that at the present day East Coast, US margin. Because loading occurs initially on a weak lithosphere there is a large amplitude, short wavelength flexure. As the load moves onto strong lithosphere, however, the basin shape shallows and broadens. The *total* basin shape (Fig. 6a) is strongly asymmetrical. Asymmetry is a characteristic feature of many foreland basins, especially the Appalachian (Tankard 1990) and the Swiss Molasse basins (Pfiffner 1990). The basement profile in Fig. 6a has a distinct curvature which results from a superposition of flexural effects from the weak and strong region. The profile would be difficult to account for in terms of a single T_e which may explain some of the difficulties that have been encountered in the past (e.g. Coakley & Watts 1991; McNutt *et al.* 1988) in modelling the shape of the basement beneath forelands.

The geometry of a foreland basin that develops on a passive margin also depends (Stockmal *et al.* 1986) on the load configuration. For example, Fig. 6b shows the case of a lateral change in T_e where the initial loads displace water while later loads displace air. The asymmetry is reduced and the basement profile would better approximate a single value of T_e. Both models show, however, a strong pattern of onlap in the deeper sections of the basin and a less well developed one in the upper sections. This is in contrast to the uniform T_e model (Fig. 6c), which shows a more limited pattern of onlap.

Although Fig. 6 assumes that sediments infill to a constant (horizontal) pre-deformation surface, it is likely that lateral variations in T_e would influence the actual environments of deposition. Initial loading on weak lithosphere, for example, would result in a large flexural depression beneath the load, an increase in 'accomodation space' above it, and a tendency for sedimentation to be outpaced by tectonics. In the terminology of Allen *et al.* (1991) the basin would then be said to be 'underfilled'. However, as the depocentre moves onto high T_e lithosphere there will be less flexure and an increased likelihood that sediments will 'overfill' the basin. This may help explain the general observation in foreland basins of a transition from the flysch (underfilled) to the molasse (overfilled) stage although, as pointed out by Stockmal *et al.* (1990), the inherited bathymetry of the passive margin may also be a factor that determines the onset of this transition. Other factors such as changes in the rate of advance of the load, vertical uplift of the wedge, erosion rate and, changes in sediment flux may also (e.g. Allen *et al.* 1991) contribute.

DISCUSSION

The thermal and mechanical properties of passive margins may help to explain the bi-modal distribution of continental T_e values (Fig. 2), most of which come from modelling basement geometry beneath foreland basins. In particular, basins that develop on stretched crust inherit low T_e values and therefore will be narrow and deep while basins that develop more on the high T_e cratonic interiors will be wide and shallow. The contrast in T_e is large enough that it should easily be possible to distinguish between basins which develop on stretched and unstretched crust.

Foreland basins develop in response to a migrating load and so there will be cases where an individual basin develops initially on weak stretched crust and then gradually moves onto more rigid lithospheres as thrusting encroaches on and crosses the passive margin hinge zone. The progression from weak to strong lithosphere will, as shown in Fig. 6, give a characteristic flexural shape to the foreland basin. The bi-modal distribution suggests, however, that few basins actually show evidence for progressive flexure. Rather, the evidence points to *either* narrow, deep *or* wide, shallow basins suggesting other factors may be involved. For example, basins may appear narrow and deep because erosion has removed the wide shallow part (e.g. Anadarko Basin). Alternatively, basins appear wide and shallow because the narrow, deeper, part is too deeply buried to be visible in seismic, well or other data (e.g. Ganges Basin).

The stratigraphic features predicted by the models in Fig. 6 may not, of course, be preserved in ancient foreland basins because new loads are likely to modify, by thrusting and folding, the flexural depressions and uplifts of previous ones. Fig. 6 schematically shows the surface of the last 'thrust' which if it cuts the infill and flexure of the previous loads could significantly 'distort' the geometry of the basin. However, irrespective of such distortions, the asymmetry, distinct curvature, relative onlap patterns, and facies transition should still be seen in the regions immediately flanking the thrust/fold loads.

The conclusion that T_e exerts a strong influence on foreland basin stratigraphy differs from that of Stockmal *et al.* (1986) who suggested that load size was more important. The main difference between this model and that of Stockmal *et al.* (1986) is in the assumptions that are made concerning the spatial and temporal variations in T_e of the crust and lithosphere that underlies a foreland basin. Here, a model is assumed, based on studies of present-day passive margins, in which stretched crust is fundamentally weak and is unable to recover its strength by cooling following rifting. Large lateral changes in T_e are therefore implied in foreland basins, *irrespective* of the age of the underlying passive margin. Stockmal *et al.* (1986), on the other hand, assumed that stretched crust acquires a high T_e relatively rapidly after rifting such that large lateral differences may exist for young margins but not for old ones.

The 'Wilson cycle' (e.g. Dewey 1982) suggests that a considerable amount of time (up to a few hundred Myr) may elapse before a passive margin becomes the site of foreland basin development. It is believed (e.g. Watts 1988) that large lateral changes in T_e persist at passive margins and that they will continue to influence foreland basin stratigraphy, irrespective of the amount of time that elapses between rifting and orogeny. In this sense, T_e of stretched crust is 'frozen in' and does not change significantly with time. Unfortunately, there is not much evidence for such behaviour in the continents. However, the argument that T_e is low at the East Coast, US margin, which received a significant proportion of its sediment 100–180 Myr after rifting, together with the ability of the shield areas to support gravity anomalies for long periods of time, are strong arguments that T_e of continents, like that of the oceans, does not change significantly as the load ages. In other words, the lithosphere appears to behave elastically and not viscoelastically, as proposed by Beaumont (1981). What is not known is whether in response to a new load there is any increase in strength of the continental lithosphere, in the meantime, due to secular cooling, as is known to be the case for the oceans

Irrespective of the ability of the continental lithosphere to acquire strength by cooling, if T_e is frozen in then we should be able to use the T_e values inferred from flexure studies to infer, in the absence of other information, the tectonic setting of ancient foreland basins. For example, the low T_e determined by Grotzinger & Royden (1990) for the Kilohigok Basin may reflect its formation on stretched continental crust. Another possibility is that the basin formed by progressive flexure and that the upper wide part has subsequently been removed by erosion. In neither case, however, it is necessary to invoke that Archean lithosphere as a whole is fundamentally hot and weak as proposed by Grotzinger & Royden (1990).

Although we may therefore better understand the architecture of foreland basins, several questions remain concerning the long-term mechanical properties of the continental plates. The most important of these concern why, in contrast to the oceans, the continental lithosphere is capable of exhibiting both great strength *and* weakness. The high values appear to be associated with cratons and presumably reflect the low temperature gradients that have been inferred for these regions. For example, a T_e of 80 km would imply an equilibrium thermal thickness of about 250 km which although twice that of oceanic lithosphere is not in marked disagreement with the seismic thickness of the lithosphere as revealed in P–SV data (Lerner-Lam & Jordan 1987). Why stretched crust is so weak is still unexplained. It most probably reflects fundamental differences between the way that the lithospere responds to 'side-driven' loads of rifting and the surface loads of flexure. Continental T_e estimates suggest that stretched crust responds to sediment loads, as well as subsequent thrust/fold loads, as a thin elastic sheet. A comparison of these estimates to continental yield strength envelopes (e.g. Fig. 5) suggest that the stresses associated with loading are concentrated in an upper quartz-rich layer. Thus, the strong upper crustal layer has, in effect, been 'de-coupled' from any support it might have otherwise have received from the underlying strong olivine-rich mantle layer. The heating associated with continental rifting and the formation of new ocean basins is an appealing way to reduce the strength of the mantle and may explain why stretched crust is associated with T_e values that resemble those from volcanic loads that form on or near mid-ocean ridges. Compositional effects such as increases in quartz content and, possibly, radiogenic heat production may further help isolate the strong upper part of the crust from the mantle. However, once stretched crust cools, it should regain its long-term strength. Thermal effects cannot therefore explain why stretched crust once it is weakened apparently remains so. It is speculated that during rifting other, chemical processes, such as the introduction of water, play a much greater role in permanently weakening the mantle than prevously thought.

ACKNOWLEDGMENTS

I thank Phil Allen for helpful discussions and Jeff Nunn, Marcia McNutt, Phil Allen and Roxby Hartley for critically reading of the manuscript.

REFERENCES

ALLEN, P. A., CRAMPTON, S. L. & SINCLAIR, H. D. (1991) The inception and early evolution of the North Alpine Foreland Basin, Switzerland. *Basin Res.* 3, 143–163.

BARTON, P. J. & WOOD, R. J. (1984) Tectonic evolution of the North Sea Basin: Crustal stretching and subsidence. *Geophys. J.R. astr. Soc.* 79, 987–1022.

BEAUMONT, C. (1981) Foreland basins. *Geophys. J.R. astro. Soc.* 65, 291–329.

BECHTEL, T. D., FORSYTH, D. W., SHARPTON, V. L. & GRIEVE, R. A. F. (1990). Variations in effective elastic thickness of the North American lithosphere. *Nature* 343, 636–638.

BRACE, W. F. & KOHLSTEDT, D. L. (1980) Limits on lithospheric stress imposed by laboratory experiments. *J. geophys. Res.* 85, 6248–6252.

BUROV, E. V., KOGAN, M. G., LYON-CAEN, H. & MOLNAR, P. (1990) Gravity anomalies, the deep structure, and dynamic processes beneath the Tien Shan. *Earth Planet. Sci. Letts.* 96, 367–383.

CALMANT, S. & CAZENAVE, A. (1986) The effective elastic lithosphere under the Cook-Austral and Society islands. *Earth Planet. Sci. Letts.* 77, 187–202.

COAKLEY, B. & WATTS, A. B. (1991) Tectonic Controls on the development of unconformities: The North Slope, Alaska. *Tectonics* 10, 101–130.

COOK, F. A., BROWN, L. D., KAUFMAN, S., OLIVER, J. E. & PETERSEN, T. A. (1983) COCORP seismic profiling of the Appalachian orogen beneath the coastal plain of Georgia. *Bull. geol. Soc. Am.* 92, 738–748.

CRITTENDEN, M. D. (1970) Discussion. *Can. J. Earth Sci.* 7, 727–729.

DEWEY, J. F. (1982) Plate tectonics and the evolution of the British Isles. *J. geol. Soc. Lond.* 139, 371–412.

DORMAN, L. M. & LEWIS, B. T. R. (1970) Experimental isostasy 1: Theory of determination of the Earth's response to a concentrated load. *J. geophys. Res.* 175, 3357–3365.

FOWLER, S. & MCKENZIE, D. P. (1989). Gravity Studies of the Rockall and Exmouth Plateaux using SEASAT altimetry. *Basin Res.* 2, 27–34.

FULTON, R. J. & WALCOTT, R. I. (1975) Title? *Mem. geol. Soc. Am.* 142, 163–173.

GOETZE, C. & EVANS, B. (1979) Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. *Geophys. J. R. astr. Soc.* 59, 463–478.

GROTZINGER, J. & ROYDEN, L. (1990) Elastic strength of the Slave craton at 1.9 Gyr and implications for the thermal evolution of the continents. *Nature* 347, 64–66.

HOMEWOOD, P., ALLEN, P. A. & WILLIAMS, G. D. (1986) Dynamics of the Molasse Basin of western Switzerland. In: *Foreland Basins* (Ed. by P. A. Allen and P. Homewood). *Spec. Publs Int. Ass. Sedim.* 8, 199–218.

HOUSEKNECHT, D. W. (1986) Evolution from passive margin to foreland basin: the Atoka formation of the Arkoma basin, south-central USA. In: *Foreland Basins* (Ed. by P. A. Allen and P. Homewood). *Spec. Publs Int. Ass. Sedim.* 8, 327–346.

JORDAN, T. E. (1981) Thrust loads and foreland basin evolution, Cretaceous, western United States. *Bull. Am. Ass. Petrol. Geol.* 65, 2506–2520.

KARNER, G. D., STECKLER, M. S. & THORNE, J. (1983) Long-term mechanical properties of the continental lithosphere. *Nature* 304, 250–253.

KARNER, G. D. & WATTS, A. B. (1982) On isostasy at Atlantic-type continental margins. *J. geophys. Res.* 87, 2923–2948.

KARNER, G. D. & WATTS, A. B. (1983) Gravity anomalies and flexure of the lithosphere at mountain ranges. *J. geophys. Res.* 88, 10, 449–10, 477.

KOCH, P. S., CHRISTIE, J. M. & GEORGE, R. P. (1980). Flow law of 'wet' quartzite in the a-field. *EOS, Trans. Am. Geophys. Un.* 61, 376.

KOMINZ, M. A. & BOND, G. C. (1986) Geophysical modelling of the geothermal history of foreland basins. *Nature* 320, 252–256.

KRUSE, S. & MCNUTT, M. K. (1988) Compensation of Paleozoic orogens: A comparison of the Urals to the Appalachians. *Tectonophys.* 154, 1–17.

KUO, B.-Y. & PARMENTIER, E. M. (1986) Flexure and thickening of the lithosphere at the East Pacific Rise. *Geophys. Res. Lett.* 13, 681–684.

LERNER-LAM, A. L. & JORDAN, T. H. (1987) How thick are the continents?. *J. geophys. Res.* 92, 14,007–14,026.

LEWIS, B. T. R. & DORMAN, L. M. (1970) Experimental Isostasy 2: An isostatic model for the U.S.A. derived from gravity and topographic data. *J. geophys. Res.* 75, 3367–3386.

LYON-CAEN, H. & MOLNAR, P. (1983) Constraints on the structure of the Himalaya from an analysis of gravity anomalies and a flexural model of the lithosphere. *J. geophys. Res.* 88, 8171–8191.

MCCONNELL, R. K. (1968) Viscosity of the mantle from relaxation time spectra of isostatic adjustment. *J. geophys. Res.* 73, 7089–7105.

MCNUTT, M. & PARKER, R. L. (1978) Isostasy in Australia and the evolution of the compensation mechanism. *Science* 199, 773–775.

MCNUTT, M. K. (1984) Lithospheric flexure and thermal anomalies. *J. geophys. Res.* 89, 11,180–11,194.

MCNUTT, M. K., DIAMENT, M. & KOGAN, M. G. (1988) Variations of elastic plate thickness at continental thrust belts. *J. geophys. Res.* 93, 8825–8838.

MCNUTT, M. K. & MENARD, H. W. (1982) Constraints on yield strength in the oceanic lithosphere derived from observations of flexure. *Geophys. J.* 71, 363–394.

MENKE, W. (1981) The effect of load shape on the deflection of thin elastic plates. *Geophys. J.R. astr. Soc.* 65, 571–577.

NUNN, J. A. & SLEEP, N. H. (1984) Thermal contraction and flexure of intracratonal basins: A three-dimensional study of the Michigan Basin. *Geophys. J. R. astr. Soc.* 76, 587–635.

PARSONS, B. E. & SCLATER, J. G. (1977) An analysis of the variation of ocean floor bathymetry and heat flow with age. *J. geophys. Res.* 82, 803–827.

PFIFFNER, O. A. (1986) Evolution of the north Alpine foreland basin in the Central Alps. In: *Foreland Basins* (Ed. by P. A.

Allen and P. Homewood). *Spec. Publs Int. Ass. Sedim.* 8, 219–228.

Pinet, B., Jaupart, C., Mareschal, J.-C., Gariepy, C., Bienfait, G. & Lapointe, R. (in press) Heat flow and structure of the lithosphere in the Eastern Canadian Shield.

Royden, L. & Karner, G. D. (1984) Flexure of the lithosphere beneath the Apennine and Carpathian foredeep basins. *Nature* 309, 142–144.

Sheffels, B. & McNutt, M. (1986) Role of subsurface loads and regional compensation in the isostatic balance of the Transverse Ranges, California: Evidence for intracontinental subduction. *J. geophys. Res.* 91, 6419–6431.

Sinclair, H. D., Coakley, B. C., Allen, P. A. & Watts, A. B. (1991) Simulation of foreland basin stratigraphy using a diffusion model of mountain belt uplift and erosion: an example from the central Alps, Switzerland. *Tectonics* 10, 599–620.

Smith, W. H. F., Staudigel, H., Watts, A. B. & Pringle, M. S. (1989) The Magellan Seamounts: Early Cretaceous record of the South Pacific isotopic and thermal anomaly. *J. geophys. Res.* 94, 10,501–10,523.

Snyder, D. B. & Barazangi, M. (1986) Deep crustal structure and flexure of the Arabian plate beneath the Zagros collisional mountain belt as inferred from gravity observations. *Tectonics* 5, 361–373.

Steckler, M. S. (1985) Uplift and extension in the Gulf of Suez: Indications of induced mantle convection. *Nature* 317, 135–139.

Steckler, M. S. & Ten Brink, U.S. (1986). Lithospheric strength variations as a control on new plate boundaries: examples from the Northern Red Sea region. *Earth Planet. Sci. Letts.* 79, 120–132.

Stockmal, G. S., Beaumont, C. & Boutilier, R. (1986) Geodynamic models of convergent margin tectonics: Transition from rifted margin to overthrust belt and consequences for foreland basin development. *Bull. Am. Ass. Petrol. Geol.* 70, 181–190.

Tankard, A. J. (1986) On the depositional response to thrusting and lithospheric flexure: Examples from the Appalachian and Rocky Mountain basins. In: *Foreland Basins* (Ed. by P. A. Allen and P. Homewood). *Spec. Publs Int. Ass. Sedim.* 8, 369–392.

Von Herzen, R. P., Cordery, M. J., Detrick, R. S. & Fang, C. (1989) Heat flow and the thermal origin of hot spot swells: The Hawaiian swell revisited. *J. geophys. Res.* 94, 13,783–13,799.

Vink, G. C., Morgan, W. J. & Wu-Ling, Z. (1984) Preferential rifting of continents: A source of displaced terranes. *J. geophys. Res.* 89, 10,072–10,076.

Walcott, R. I. (1970) Isostatic response to loading of the crust in Canada. *Can. J. Earth Sci.* 7, 716–727.

Watts, A. B. (1978) An analysis of isostasy in the world's oceans: 1. Hawaiian-Emperor Seamount Chain. *J. geophys. Res.* 83, 5989–6004.

Watts, A. B. (1981) The U.S. Atlantic continental margin: Subsidence history, crustal structure and thermal evolution. *Am. Ass. Petrol. Geol., Education Course* 19, 75.

Watts, A. B. (1988) Gravity anomalies, crustal structure and flexure of the lithosphere at the Baltimore Canyon Trough. *Earth Planet. Sci. Letts.* 89, 221–238.

Watts, A. B., Duncan, R. A. & Larson, R. L. (1988) The origin of the Louisville Ridge and its relationship to the Eltanin Fracture Zone system. *J. geophys Res.* 93, 3051–3077.

Watts, A. B. & ten Brink, U.S. (1989) Crustal structure, flexure and subsidence history of the Hawaiian Islands. *J. geophys. Res.* 94, 10,473–10,500.

Weissel, J. K. & Karner, G. D. (1989) Flexural uplift of rift flanks due to mechanical unloading of the lithosphere during extension. *J. geophys. Res.* 94, 13,919–13,950.

Wessel, P. (1992) Thermal stresses and the bimodal distribution of elastic thickness estimates of oceanic lithosphere. *J. geophys. Res.* 97, 14,177–14,194.

Wessel, P. & Haxby, W. F. (1990) Thermal stresses, differential subsidence, and flexure at oceanic fracture zones. *J. geophys. Res.* 95, 375–391.

White, N. & McKenzie, D. P. (1988) Formation of the 'Steer's Head' Geometry of sedimentary basins by differential stretching of the crust and mantle. *Geology* 16, 250–253.

Zoetemeijer, R., Desegaulx, P., Gloetingh, S., Roure, F. & Moretti, L. (1990) Lithospheric dynamics and tectonicstratigraphic evolution of the Ebro Basin. *J. geophys. Res.* 95, 2701–2711.