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# The lithosphere under stress

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*Phil. Trans. R. Soc. Lond. A* 1999 **357**, 901-915
doi: 10.1098/rsta.1999.0357

THE ROYAL SOCIETY

# The lithosphere under stress

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I discuss thermal, rheological and compositional definitions of the Earth's lithosphere and describe the data that can be brought to bear in refining these definitions. Models of the behaviour of the lithosphere are useful in describing such diverse effects as continental basin subsidence, the flexural response to imposed loads such as seamounts or sediment accumulations, the partial melting of mantle beneath rifts, the composition and geochemical signature of the melts, and the depths of seismicity. It is generally the case that oceanic lithosphere is simpler to understand than is continental lithosphere, primarily due to its usually younger age and its consistent mode of formation at mid-ocean ridges.

**Keywords: lithosphere definitions; elastic thickness; rheology; mantle temperature; subsidence; plate model**

## 1. Background

The notion of the lithosphere as a strong outer layer of the Earth has been in circulation since Barrell wrote a series of papers introducing the concept (Barrell 1914*a–c*). His work was based on the observation of significant gravity anomalies over continental crust, from which he inferred that there must exist a strong upper layer (the lithosphere) above a weaker layer which could flow (the asthenosphere). These ideas were enlarged by Daly (1940), and have been broadly accepted by geologists and geophysicists ever since (White 1988). Alternative definitions of the lithosphere based on the assumed thermal structure of the outer layer of the Earth have proven to be particularly powerful in the formulation of simple models of basin subsidence resulting from lithospheric extension, both in the oceanic basins (Parsons & Sclater 1977), and in continental basins (McKenzie 1978).

However, alongside the evident utility of the concept of the lithosphere has been a burgeoning in its use in a wide variety of contexts. Thus the term lithosphere, *sensu lato*, has been used in models to explain the flexural response of the Earth to imposed loads such as seamounts or mountain belts; the production of molten rocks in areas undergoing extension; the formation and geochemical composition of continental flood basalts; the isostatic response of the Earth to explain vertical motions observed in both continental and oceanic settings; the observed heat flow and inferred heat production at depth; and the composition of the depleted mantle beneath Archaean shields.

Though there may be considerable merit in the various models, the lack of a clear definition of the lithosphere leads easily to confusion when the term is used by different authors in different circumstances. Because the properties of the lithosphere depend on the thermal state, the strain rate, the composition and mineralogy, and the stress in the outer part of the Earth, there are often particular dangers of misunderstanding between scientists concerned primarily with the physical properties of the lithosphere (such as the effective elastic thickness of a particular region) and those concerned with the lithospheric geochemical reservoir, which may give distinctive signatures to igneous melts. In part to counteract such problems, Jordan (1975, 1988) introduced the concept of the tectosphere as a chemically distinct layer, and Anderson (1995) suggested using a term 'perisphere' to denote a weak, geochemically enriched layer that may spread across the top of the convecting mantle.

In this paper, I discuss the properties of the lithosphere that are useful in understanding the response of the outer skin of the Earth to extension. The properties of the lithosphere fall into three main categories, which I discuss separately: the thermal structure; the rheological structure and the compositional structure.

## 2. Thermal definitions of the lithosphere

The thermal structure of the lithosphere and underlying asthenosphere in a region undergoing extension exerts an important control on both the subsidence history of the region, and on the generation of partial melt by decompression of the mantle that wells up beneath the extending region. We consider these two aspects next.

### (*a*) *Subsidence*

An example of how a simple thermal understanding of the lithosphere has led to a powerful predictive model of the bathymetric shape of the ocean basins is the lithosphere cooling model of Parsons & Sclater (1977). This model assumes that conductive heat loss through the sea-floor causes the asthenospheric mantle that lies immediately beneath the underlying lithosphere to cool, and thereby to become part of the thickening lithosphere as it moves away from an oceanic spreading centre. It is then straightforward to calculate the cooling history of the lithosphere, and hence the increase of its density with age. With the assumption that local isostasy is maintained, the subsidence history can then be predicted. The base of the lithosphere is thus defined as an isotherm. This conceptually simple model successfully matches the observed subsidence of oceanic crust out to lithospheric ages of *ca.* 70 Ma, beyond which the lithosphere ceases to thicken appreciably, but appears to be maintained at a constant thickness. This model also predicts the heat flow through the surface of the plate, although at young ages the measured heat flow is generally smaller than the theoretically predicted heat flow. This discrepancy arises primarily because significant amounts of heat are lost from young crust by advection through hydrothermal circulation, whereas measurements of heat flow through the sea-floor determine only the conductive heat flow.

The main feature of the Parsons & Sclater (1977) model is the isothermal base of the plate: this is a simple way of describing the additional heat input, which prevents half-space cooling continuing beyond *ca.* 70 Ma. Parsons & McKenzie (1978) proposed that a small-scale convective instability creates a thermal boundary layer between the base of the conductively cooled upper layer and the underlying well-mixed, adiabatic asthenosphere. This small-scale convective instability thus maintains a constant average depth to the base of the conductively cooled section in lithosphere older than *ca.* 70 Ma. A time-averaged equilibrium thermal structure is shown in





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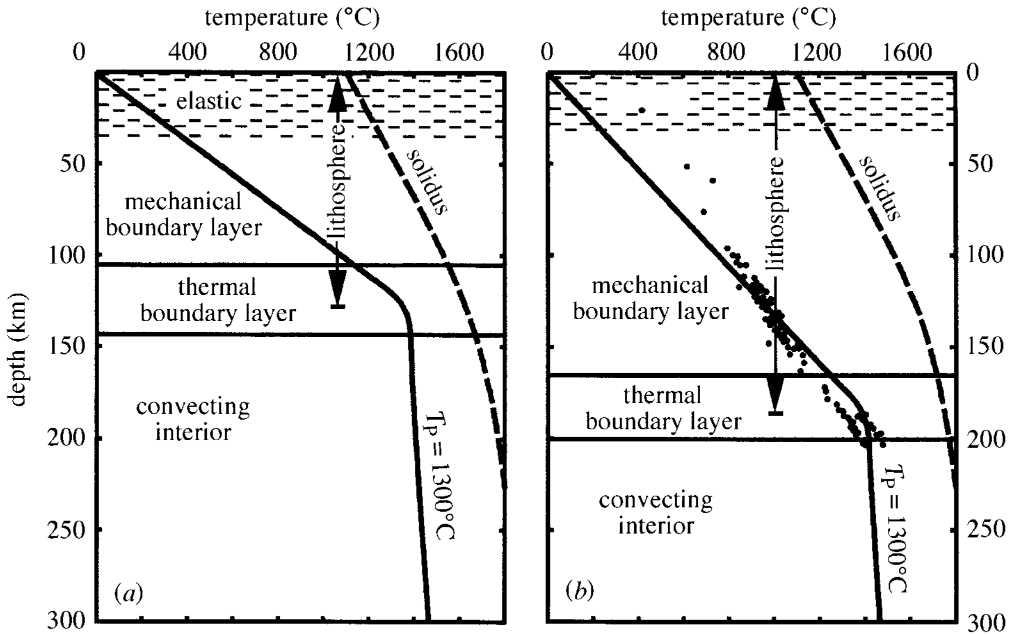


Figure 1. Representative, average temperature–depth curves for (*a*) mature oceanic lithosphere, and (*b*) Archaean shield, to demonstrate some common definitions of the lithosphere based on the thermal structure. The curve in (*a*) is based on Parsons & Sclater's (1977) thermal structure of mature oceanic lithosphere, with a potential temperature, $T_{\mathrm{P}}$, of 1300 °C in the convecting interior, and has an average lithosphere thickness of 125 km, with a mechanical boundary layer 107 km thick and a 33 km thick thermal boundary layer of small-scale convection. The curve in (*b*) for an Archaean shield has a lithosphere thickness of 185 km, with mechanical and thermal boundary layer thicknesses of 165 and 36 km, respectively. Dots are estimates of pressure and temperature from individual garnet peridotite nodules brought to the surface by kimberlite magmas in the Kaapvaal Craton of South Africa (from McKenzie & Fairhead 1997).

figure 1*a* for mature oceanic lithosphere. Using Parsons & McKenzie's (1978) terminology, the mechanical boundary layer maintains a conductive thermal gradient and is mechanically isolated from the underlying well-stirred asthenosphere, while the thermal boundary layer of small-scale convection can be thought of as attached to the overlying plate on short time-scales, but as part of the underlying mantle circulation over long time-scales.

These simple parametrizations of the thermal structure of the oceanic lithosphere do not take account of many factors that in the real world modify the thermal structure: such factors include variable heat production by radioactive decay, particularly in the crust; variations with depth of the thermal conductivity; phase changes; temperature variations in the mantle (such as those caused by mantle plumes); and hydrothermal circulation. Nevertheless, they remain a powerful standard against which anomalies in the sea-floor depth can be measured. Parsons & Sclater (1977) give a mean lithospheric thickness in mature oceanic plates of $125 \pm 10$ km with a basal temperature of $1350 \pm 275$ °C (table 1).

A more recent investigation of ocean subsidence and heat-flow data by Stein & Stein (1992) yields a slightly thinner plate at $95 \pm 15$ km, with a basal temperature of $1450 \pm 250$ °C, which is the same, within uncertainties, as the Parsons & Sclater (1977) basal temperature (table 1). Stein & Stein (1992) comment that the uncertain contribution of radioactive decay might change the basal temperatures by up to 50 °C.

Rifted continental lithosphere exhibits subsidence with a similar time constant to that of oceanic lithosphere (Sleep 1971). This led McKenzie (1978) to propose that lithospheric thinning was a result of stretching, accompanied by a subsequent phase of cooling similar to that which occurs in oceanic lithosphere, to account for the subsidence history of many continental sedimentary basins. It has proven to be a robust model, which can be modified to allow for finite duration stretching, multiple rift episodes, phase changes, lateral heat flow and melt production during rifting. Inversions of subsidence data derived from over 2000 stratigraphic sections in Phanerozoic continental rift basins suggest that the best fit to the lithospheric thickness is $120 \pm 15$ km, with a basal temperature of $1400 \pm 150$ °C (Newman & N. White, this issue). This, again, is indistinguishable from mature oceanic lithosphere (table 1).

### (*b*) *Melt generation by mantle decompression*

An independent measure of the temperature of the asthenosphere, at the base of the plate, is provided by the products of partial melting of the mantle that occurs when the mantle rises and decompresses beneath thinning lithosphere. Again it is oceanic spreading centres that provide the best constraints on the asthenospheric temperature from this process, because the complete separation of the lithospheric plates at such locations means that it is possible to model the path of the upwelling mantle. Provided the full sea-floor spreading rate is faster than about 20 mm $a^{-1}$, which is true for considerable lengths of the mid-ocean ridge system (including, for example the East Pacific Rise and most of the Mid-Atlantic Ridge), then the mantle will well up isentropically, without significant loss of heat by conduction. As the asthenospheric mantle rises toward the surface, it eventually crosses the solidus (figure 1*a*) and starts to melt. The melt then rises buoyantly toward the surface, where eventually it freezes to form the oceanic crust. Provided the melt is extracted efficiently from the source mantle rock, as indeed appears to be the case, and provided no parcel of mantle rock passes through the melting zone more than once, then the thickness of the igneous crust will be a good measure of the total amount of melt generated, and should be independent of the spreading rate.

Compilation of oceanic crustal thicknesses in areas away from perturbing factors such as mantle plumes and fracture zones shows that the thickness is remarkably uniform around the world, although fast-spreading ridges (full rate greater than 40 mm $a^{-1}$) tend to exhibit slightly thinner crusts than do slow-spreading ridges (full rate 20–40 mm $a^{-1}$) (White *et al.* 1992). However, slow-spreading ridges are characterized by numerous fracture zones, which typically have thinner crusts than normal. If the crustal thickness is averaged along spreading centres, including both normal spreading segments and fracture zones, then the observed average crustal thickness is close to 6.5 km along both fast- and slow-spreading ridges (Bown & White 1994). This crustal thickness can then be compared with theoretical models of melt generation, such as those derived by McKenzie & Bickle (1988), to deduce the mantle temperature. At very slow spreading ridges (full rate less than 20 mm $a^{-1}$),

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Table 1. *Mature thermal lithosphere, remote from mantle plumes*
($\alpha$ is the coefficient of thermal expansion, assumed constant throughout the lithosphere; $L$ is the latent heat of fusion; $\Delta S$ is the entropy change on melting.)

| constraints | thickness | asthenosphere temperature | other parameters | authors |
|---|---|---|---|---|
| oceanic subsidence and heat flow | $125 \pm 10$ km | $1350 \pm 275$ °C | $\alpha = (3.2 \pm 1.1) \times 10^{-5}\ \mathrm{K}^{-1}$ | Parsons & Sclater 1977 |
| oceanic subsidence and heat flow | $95 \pm 15$ km | $1450 \pm 250$ °C | $\alpha = (3.1 \pm 0.8) \times 10^{-5}\ \mathrm{K}^{-1}$ | Stein & Stein 1992 |
| continental subsidence | $120 \pm 15$ km | $1400 \pm 150$ °C | $\alpha = (3.3 \pm 0.3) \times 10^{-5}\ \mathrm{K}^{-1}$ | Newman & N. White (this issue) |
| melt generation for oceanic crust | — | 1320–1355 °C[a] | $L = 3.34 \times 10^{-5}\ \mathrm{J\ kg}^{-1}$ | Foucher *et al.* 1982 |
| melt generation for oceanic crust | — | $1340 \pm 20$ °C[a,b] | $\Delta S = 250\ \mathrm{J\ kg}^{-1}\ \mathrm{K}^{-1}$ | McKenzie & Bickle 1988 |
| melt generation for oceanic crust | — | $1360 \pm 20$ °C[a,c] | $\Delta S = 400\ \mathrm{J\ kg}^{-1}\ \mathrm{K}^{-1}$ | Bown & White 1994 |

[a]Temperature at a depth of 100 km, assuming an adiabatic gradient in the lithosphere of 0.6 °C $\mathrm{km}^{-1}$.
[b]This calculation ignores the mantle compaction that occurs as melt is extracted, and allows the mantle to decompress to the surface.
[c]This calculation includes the mantle compaction that occurs as melt is extracted, and allows the mantle to decompress to the base of the igneous crust.

Table 2. *Effective elastic thickness beneath rifts with* $\beta > 1.3$

| location | elastic thickness (km) | authors |
|---|---|---|
| from forward modelling techniques | | |
| Central Graben, North Sea | $\leqslant 5$ | Barton & Wood 1984 |
| Rockall Continental Margin | $\leqslant 5$ | Fowler & McKenzie 1989 |
| Exmouth Plateau Margin, Australia | 5 | Fowler & McKenzie 1989 |
| Michigan Basin, USA | 4.6 | Nunn & Sleep 1984 |
| Lake Tanganyika Rift, East Africa | $\sim 3$ | Kusznir *et al*. 1995 |
| Turkana Rift, East Africa | 3 | Kusznir *et al*. 1995 |
| Turkana Rift, East Africa | 3.5 | Hendrie *et al*. 1995 |
| North Viking Graben, North Sea | 1.5–2.0 | Kusznir *et al*. 1995 |
| Basin and Range, USA | 1–2 | Buck 1988 |
| Basin and Range, USA | 1.5 | Kusznir *et al*. 1995 |
| Recôncavo Sub-basin, Brazil | 5 | Magnavita *et al*. 1994 |
| North Tucano Basin, Brazil | 5 | Magnavita *et al*. 1994 |
| Metamorphic core complexes | $< 5$ | Buck 1988 |
| Dobe Basin, Afar | 5 | Hayward & Ebinger 1996 |
| Corinth, Aegean | 8 | King 1998 |
| Gabon Margin, West Africa | 0–10 | Watts & Stewart 1998 |
| Baltimore Canyon Trough, USA | $\sim 5$ | Watts & Marr 1995 |
| from coherence techniques | | |
| Basin and Range, USA | $\sim 5$ | Betchel *et al*. 1990 |
| Basin and Range, USA | 4.6–16 | Lowry & Smith 1995 |
| Dobe, Afar | 5.6 | Hayward & Ebinger 1996 |
| Addo-Do, Afar | 6 | Hayward & Ebinger 1996 |
| Corinth, Aegean | 8–10 | King 1998 |

the mantle welling up beneath the spreading centre loses heat by conduction, resulting in the generation of less melt and thus of thinner oceanic crust (Bown & White 1994).

McKenzie & Bickle (1988) used published data on the pressure, temperature and composition of melts formed from samples of presumed mantle material in laboratory experiments, together with thermodynamic arguments, to calculate the volume and composition of the melt that would be generated by isentropic mantle decompression beneath an oceanic spreading centre. Their results show that the volume of melt is extremely sensitive to the mantle temperature, with an increase in mantle temperature of as little as 100 °C causing twice as much melt to be generated beneath an oceanic spreading centre.

Using McKenzie & Bickle's (1988) original parametrization, the average 6.5 km of igneous crust found in the ocean basins requires a mantle temperature at 100 km depth of $1340 \pm 20$ °C (table 1). The original parametrization did not include the effects of mantle compaction that occurs as melt is extracted, which tends to increase the amount of melt generated from mantle of a given temperature (White *et al*. 1992). McKenzie & Bickle (1988) also used a rather lower value than is currently accepted for the entropy change on melting: use of a higher value causes a decrease in the calculated amount of melt generated from mantle at a given temperature. In the original calculations they also allowed the mantle to decompress to the surface, and thus to continue generating melt: restricting the shallowest depth to which the mantle can well up to be the base of the oceanic crust at the spreading axis also causes a decrease in the amount of melt generated from mantle at a given temperature. The first change works in the opposite direction to the second two changes, and the revised estimate of mantle temperature is little altered when they are included, at $1360 \pm 20\,^{\circ}\mathrm{C}$ (Bown & White 1994). Using a simple analytical expression for the degree of partial melting as a function of temperature and pressure, Foucher *et al*. (1982) had earlier reported that similar mantle temperatures of 1320–1355 °C (recalculated from their values, using an adiabatic gradient of $0.6\,^{\circ}\mathrm{C\,km^{-1}}$) were required for the generation of normal thickness oceanic crust (table 1).

As can be seen from table 1, the estimates of mantle temperature derived from mantle-melting models are indistinguishable from those derived from plate models of the oceanic heat flow and subsidence history. The absolute temperature estimates from mantle melting are subject to a number of uncertainties, including details of whether fractional or batch melting is assumed, the melt extraction process, the value of the entropy change on melting, the original composition of the mantle, and the extent to which volatiles in the mantle play a part, since volatiles lower the solidus temperature considerably. Nevertheless, these effects probably only have a relatively minor effect on the inferred mantle temperature, and the agreement is striking with temperatures determined from plate models.

A second way in which melt products can be used to infer temperatures in the mantle is by inverting the rare earth element (REE) concentrations in oceanic volcanic rocks to calculate the depths at which partial melting occurred, and the total amount of mantle melting that has taken place (McKenzie & O'Nions 1991). For oceanic crust the melt 'thicknesses' inferred from REE inversions are in agreement, within the uncertainties, with the seismically measured crustal thicknesses (White *et al*. 1992). The inference of mantle temperature from the melt distribution with depth carries more uncertainty, because it depends on comparing the melt distribution deduced from REE concentrations with that derived from a theoretical forward calculation of mantle decompression melting with depth. Mantle temperatures of *ca.* 1350 °C beneath normal oceanic crust are inferred by such a comparison of melt distributions derived from REE inversions with forward melting models. So, again, this is consistent with the mantle temperatures at the base of the lithosphere derived from the plate models and from the oceanic crustal thickness.

Beneath continental areas, large volumes of igneous rock may be generated by decompression melting of the mantle, particularly where the continental lithosphere overlies a mantle plume (White & McKenzie 1989). However, it is more complicated to unravel the thermal structure of the lithosphere from continental melts than from oceanic spreading-centre melts because the initial thickness of the unrifted continental lithosphere means that some melts may be sourced in enriched lithospheric mantle (McKenzie 1989), and some in the underlying asthenospheric mantle, and the melts may also pick up a geochemical and isotopic signature from the lithosphere as they rise through it toward the surface. Nevertheless, REE inversions from flood basalts (White & McKenzie 1995) are consistent with the thermal structure of continental lithosphere being similar to that shown in figure 1.





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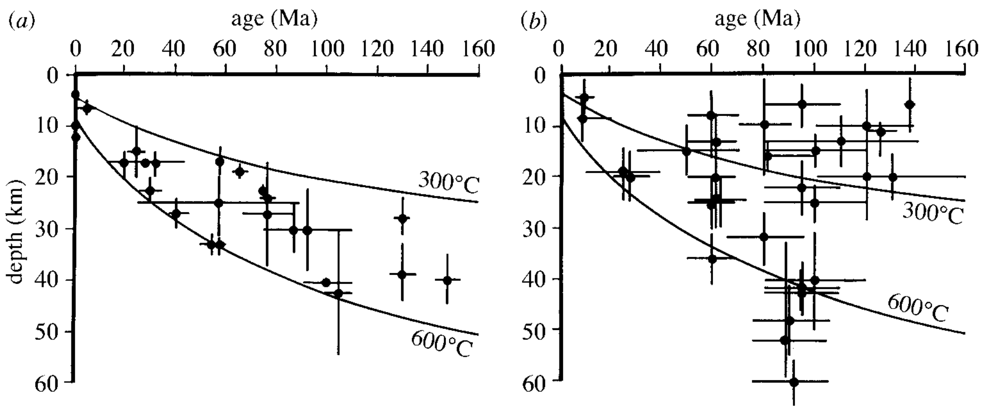


Figure 2. (*a*) Compilation of $T_e$ against age of the lithosphere at the time of loading, for oceanic lithosphere, excluding fracture zones and seamounts/oceanic islands in the French Polynesia region of the central Pacific. Redrawn from figure 1*a* in Watts (1992). (*b*) Focal depths versus lithospheric age for earthquakes in oceanic interiors and on the seaward side of oceanic trenches. Data from Chen & Molnar (1983). Curves show depths to the 300 °C and 600 °C isotherms based on Parsons & Sclater's (1977) cooling-plate model.

## 3. Rheological definitions of the lithosphere

Only the cold, uppermost part of the Earth within, at most, a few tens of kilometres of the surface is capable of supporting long-term stresses over geological time-scales. Such stresses may be imposed locally by, for example, the emplacement of a volcanic seamount on pre-existing oceanic crust. The normal way of parametrizing the elastic response of the lithosphere is to calculate an 'effective elastic thickness', $T_e$: this is the thickness of a hypothetical layer with uniform properties, which can be used to represent the undoubtedly more complex response of the real Earth. The elastic thickness may be defined as the thickness of a perfectly elastic plate which has the same overall flexural strength as has the lithosphere of the real Earth on time-scales of millions of years. It follows, therefore, that the base of this hypothetical elastic plate does not necessarily correspond to any real geological horizon, and indeed that the top of it may not lie at the surface if there are weak materials such as unconsolidated sediments in the uppermost crust.

As with the use of simple conductive models of the development of the thermal lithosphere to explain the bathymetry of the ocean basins, the concept of effective elastic thickness was first applied to oceanic lithosphere and it has proven to be a simple and durable concept. Away from active plate boundaries, the effective elastic thickness beneath the oceans increases with age, with a thickness that approximates to the depth of the 450–600 °C isotherms (figure 2*a*) (Watts 1978, 1992). This is well within the mantle, and the simple relationship between elastic thickness and lithospheric age is attributable to the relatively uniform and consistent composition, and hence physical properties, of the suboceanic mantle, at least in contrast to the complex structure and diverse composition of the crust.

However, in continental lithosphere the concept of effective elastic thickness has had a more chequered history. It has often been estimated using the coherence between the two-dimensional Fourier transforms of the topography and Bouguer gravity anomalies (Forsyth 1985). The results of such methods suggest that the elastic thicknesses of Archaean shields may be as large as 130 km. Since the temperatures at such depths are 800–1000 °C, about double those of the temperature at the base of the effective elastic plate in oceanic lithosphere, these results are difficult to understand. Recently, McKenzie & Fairhead (1997) have shown that the effective elastic thicknesses calculated using Forsyth's (1985) method are often upper bounds.

McKenzie & Fairhead's (1997) revised estimates, using as input the free air gravity anomalies and topography as a load whose geometry is known, suggest that the effective elastic thicknesses, away from dynamically supported areas, and with the exception of the Himalayan foredeep, are rarely greater than about 25 km, and are usually less. These estimated values of $T_e$ are often comparable to the thickness of the seismogenic layer under continents. The base of the seismogenic layer under continents is estimated by Chen & Molnar (1983) as being marked by a temperature of $350 \pm 100$ °C, while in the Archaean and Proterozoic crust of east Africa, Foster & Jackson (1998) report a seismogenic layer thickness of *ca.* 35 km, and suggest that at the base of this layer the temperatures are *ca.* 325–475 °C. These results are consistent with what we understand of the strengths of oceanic and continental lithosphere. The strength of a given material is probably characterized most simply by its homologous temperature, which is the ratio of the actual temperature of the material to its melting temperature. In oceanic lithosphere, the base of the elastic layer lies in the mantle at a depth corresponding to a temperature of 450–600 °C (Watts 1992), which represents a homologous temperature of about 0.5. In continental lithosphere the elastic layer lies entirely within the crust, but since the melting temperature of continental crust is lower than that of the mantle, at the same homologous temperature of 0.5, the actual temperature of the continental crust at the base of the elastic layer will be somewhat lower.

The suboceanic lithospheric mantle is probably largely dry, containing little water or other volatile elements. If any volatiles had originally been present in the mantle, they would probably have been stripped out of it as the mantle passed through the melting region beneath the oceanic spreading axes. Since crustal rocks have a lower melting temperature than do mantle rocks, it is consistent that the base of the seismogenic layer in continental crust represents a lower temperature than it does in the suboceanic mantle. However, the apparent absence of earthquakes in subcontinental mantle, apart from the special case of underthrust regions at convergent plate margins, suggests that the subcontinental lithospheric mantle may actually be weaker than the lower continental crust. A possible explanation would be that the subcontinental mantle contains volatiles from metasomatic melts bled into it over a long period from the underlying circulating asthenospheric mantle, which would lower its melting temperature and hence its strength at a given temperature. By contrast, the lower continental crust may be free of water or other volatiles, and thus stronger than the underlying mantle (D. McKenzie, personal communication).

In continental rift basins that have undergone extension of at least 30% (i.e. $\beta$ factors of 1.3 and higher), the effective elastic thickness, $T_e$, is usually less than 6 km (figure 3). Although small, it is significantly not zero. In table 2 and figure 3, I show a compilation of reported small elastic thicknesses from a variety of tectonic settings: they range from continental rifts that have undergone relatively small amounts of extension, such as the Dobe Basin in Afar, northeast Africa ($T_e = 5.6$ km, $\beta < 1.5$), the Central Graben of the North Sea ($T_e \leqslant 5$ km, $\beta < 1.6$), and the Corinth Basin of the Aegean ($T_e =$ 8–10 km, $\beta \cong 1.4$) through highly stretched intracontinental





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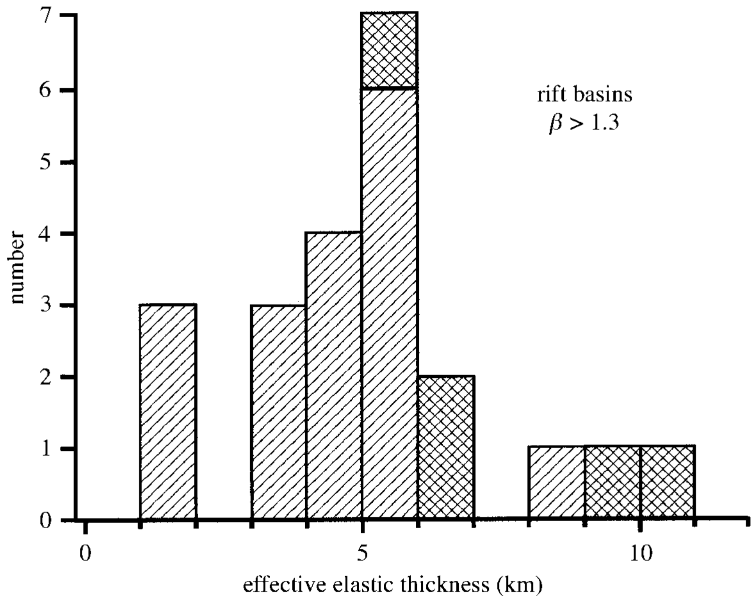


Figure 3. Histogram of effective elastic thickness determinations from rifts with more than 30% extension (see table 2 for sources of measurements). Diagonal shading is from forward modelling of subsidence, fault-block and gravity anomalies; cross-hatched shading is from coherence techniques.

settings such as the Basin and Range Province of the western USA ($T_e \approx 4$–16 km) to continental margins where stretching led to complete separation and the formation of a new oceanic basin, such as the Rockall Continental Margin ($T_e \leqslant 5$ km), the Exmouth Plateau Margin ($T_e \leqslant 5$ km), and the Baltimore Canyon Trough off the eastern USA ($T_e \sim 5$ km).

The methods by which these small values of $T_e$ shown in table 2 have been calculated also vary greatly: they include matching gravity anomalies along profiles; using coherence methods with gravity and topography data; modelling the subsidence history recorded in sediments accumulated over rifted crust; and modelling the geometry of normal faults and associated tilted fault blocks. So there seems to be little doubt that these small, effective elastic thicknesses are robust measurements and are representative of many rifted basins. Rifts that have undergone relatively little extension (i.e. $\beta \leqslant 1.3$) generally exhibit rather larger effective elastic thicknesses than those with $\beta > 1.3$. For example, in the East African Rift system, many different basins, which have been stretched by amounts ranging from less than 10% up to 30%, have effective elastic thicknesses of typically 20–35 km (see table 1 of Ebinger *et al.*, this issue; and McKenzie & Fairhead 1997). A few intracratonic basins, such as the Eyasi Basin in East Africa and the Baikal Rift in the former Soviet Union, may have $T_e$ of up to 40 km.

Only a few measurements of $T_e$ obtained using coherence techniques are shown in figure 3 and table 2 because the area needed to obtain a reliable determination usually far exceeds the size of the rift. The resultant $T_e$ may be biased to that of the unrifted lithosphere if much of the area used in the coherence estimation lies outside a rift. I have included only coherence-based results from the Basin and

Range, Afar and the Aegean, where the rifted areas are sufficiently large that it is possible to use data windows entirely within the rifts. Even so, these measurements of $T_e$ based on coherence techniques tend to be larger than those derived by forward modelling. The reason may be that forward modelling focuses on loads generated over short wavelengths, either on individual faults or in local areas. The $T_e$ from forward modelling is thus sensitive to high amplitude, short-wavelength loads that of necessity require low values of $T_e$, because otherwise very large bending stresses are produced which cannot be supported by real crustal materials (Hendrie *et al*. 1994).

It is worth noting that some modellers who have adopted necking depth models have reported much larger $T_e$ values than those shown in table 2. However, I have not included those results here because such large values of $T_e$ require large bending stresses that are much greater than the stresses at which crustal materials fail.

Another interesting observation is that effective elastic thicknesses appear to stay low on many rifted continental margins over geological time-scales of the order of 100 Ma. At first sight this is surprising because it might be expected that cooling following rifting would lead to increased crustal strength and thus to increased $T_e$. However, in the case of continental margins, these are often the sites of thick accumulations of sediment eroded off the adjacent continent. The effect of rapidly depositing thick sediments on extended crust is to insulate the underlying crust, consequently keeping it hotter than it would otherwise be. The result of this may be to keep the effective elastic thickness low for a considerable period of time.

On a whole-lithosphere scale, Newman & N. White (this issue) suggest that the total amount of extension in a continental rift is controlled by the viscosity of the mantle part of the lithosphere. A consequence of this may be that the lithosphere becomes stronger after a large rift event. This effect is enhanced over a period of 50 Ma or so, when sufficient time has elapsed for the thermal structure of the lithosphere to have substantially re-equilibrated. The main reason for increased overall strength is that the proportion of continental crust to mantle in the lithosphere decreases as the lithosphere is extended: therefore, the mantle under the rift is nearer the surface, and therefore is cooler and stronger. The result of this is that in areas that have undergone two phases of extension, large first stretching events are commonly followed by relatively small second events and vice versa, as Newman & N. White (this issue) show. On a larger scale, this effect may explain the overall migration of continental rift basins on the northwest European Continental Margin. The North Sea Basin east of Britain was created largely in the Jurassic, while renewed extension in the Cretaceous operated largely to open basins west of Britain and Norway, with the final phase of extension in the Early Tertiary jumping still further westward, to open the North Atlantic Oceanic Basin west of the Edoras Bank–Hatton Bank–Faroe Islands–Vøring escarpment continental blocks.

## 4. Composition of the lithosphere

The growth and behaviour of oceanic lithosphere through time can be modelled well by a simple thermal-cooling model, as has been discussed above. In broad terms the oceanic lithosphere has a simple and uniform composition, partly because it is generated by a straightforward mechanism from asthenospheric mantle at oceanic spreading centres with subsequent cooling, and partly because it is all young in geological terms, with none of the present lithosphere in the major oceanic basins being older than *ca.* 200 Ma. Thus there has been relatively little time for the oceanic lithosphere to be modified greatly by the addition of metasomatic melts, although some limited enrichment may occur (e.g. Class & Goldstein 1997). There are also likely to be some variations in the oceanic lithospheric composition due to variations in the geochemical composition of the asthenospheric mantle from which it was formed, and due to variations in the temperature of the asthenospheric mantle caused by the presence of hot mantle plumes or mantle 'cold spots'.

In contrast, the continental lithosphere may be extremely old, reaching an age of several gigayears in the oldest known areas, and thus may accumulate a long and complex history. This is reviewed more fully by Hawkesworth *et al.* (this issue) and Anderson (1995). Continental lithosphere is likely to have suffered multiple episodes of decompression melting as it responded to tectonic deformation, and to have been enriched by the accumulation of metasomatic melts throughout its long history. In addition to thermal stabilization, similar to what occurs through simple cooling in the oceanic lithosphere, the continental lithosphere is likely to be stabilized by the accretion of mantle whose density has been decreased by the extraction of partial melt. This may explain the greater thickness of Archaean lithosphere in areas such as the Kaapvaal Craton in South Africa that have suffered little deformation in the last 2.5 Ga compared to oceanic lithosphere (figure 1*b*).

Our knowledge of the composition of Archaean lithosphere is only indirect, coming from three main sources. The first is from nodules and inclusions brought to the surface rapidly by kimberlite pipes and from detrital remains, such as diamonds, that have been carried to the surface from deep in the lithosphere. The second is from the geochemistry and isotopic content of melts erupted at the surface that have been sourced either in the lithosphere itself, or which have travelled from a deeper source in the underlying asthenospheric mantle, but have picked up a geochemical signal from the lithosphere during their ascent to the surface. The third is from comparison of the measured seismic velocities of the lithosphere with models of the velocities to be expected from appropriate combinations of minerals at the relevant pressures and temperatures (e.g. Qiu *et al.* 1996).

Although the pressure–temperature estimates from diamonds and from peridotite nodules are likely to be improved in the future by better experimental data, they appear to be consistent with simple thermal models of Archaean lithosphere that indicate a lithospheric thickness of 170–200 km (e.g. see pressure–temperature values from nodules shown as dots on figure 1*b* (from McKenzie & Fairhead 1997)). Surface wave seismic studies generally suggest that the Archaean lithosphere is less than *ca.* 200 km thick (e.g. Grand & Helmberger 1984; Anderson 1995; Qiu *et al.* 1996; Priestley 1999). There is still some debate as to whether there is a recognizable root beneath cratons that extends another 100 km or more deeper than 200 km, although the only evidence available is from earthquake seismology. Two main types of seismic data are available: horizontally travelling surface waves and almost vertically travelling waves returned from deep interfaces, such as the ScS phase that has returned from the Earth's core. There are resolution difficulties associated with each type of seismic data. Surface waves have poor resolution at depths below 200 km, and a long horizontal path is required to determine the structure: at the very low frequencies required for the depth penetration, the cratonic areas are relatively small so there is little horizontal resolution. The ScS and similar phases have travelled long vertical paths and so average out the velocity structure along the entire path:

it is difficult to localize any anomalies to one depth region. Thus measurements of anisotropy from ScS phases have been ascribed to depths of 150–400 km (Gaherty & Jordan 1995; Vinnik *et al*. 1995), and have been used to argue that the mantle is deforming in this depth range due to lateral movement of an overlying more rigid plate. This interpretation would suggest that there is a deformable root beneath the cratonic lithosphere of the upper 150–200 km. But at present such arguments are subject to further debate and there is no independent compositional evidence that can be brought to bear on them.

In any case, the main concerns of this paper are with rifted regions and it is clear that wherever continental lithosphere has undergone extension, its compositional and seismic characteristics are consistent with simple thermal models of the lithosphere. If there are any deeper roots beneath continental lithosphere, and their reality is still the subject of vigorous debate, they are confined to Archaean regions that have not undergone deformation for at least 2 Ga.

## 5. Concluding remarks

Models of the lithospheric structure provide an important means of interpreting major geological processes, particularly in areas that have undergone extension and rifting. They provide powerful conceptual frameworks for understanding such diverse topics as the sedimentary sequences and subsidence history of intracontinental basins and continental margins, the structure and tectonics of faulted rift basins, and the magmatism that is often associated with rifting. The geological manifestations that we can observe and sample at the Earth's surface are invariably controlled primarily by deeper, lithospheric-scale processes. Since we cannot sample deep into the lithosphere, our understanding of it is dependent largely on indirect methods such as the interpretation of surface waves and deep penetrating body waves from earthquakes. Limited direct samples of presumed deep origin from nodules preserved in kimberlites provide some compositional control on the deep lithosphere.

The other methods of inferring lithospheric structure, such as flexural response to loads, basin subsidence and heat-flow observations, are all model-dependent. However, the degree of consistency of parameters derived from different types of data, such as the mantle temperature in the lithospheric plate model inferred from ocean subsidence, from continental sedimentary basin subsidence and from the generation of partial melt beneath sea-floor spreading centres (table 1), gives confidence that the model is reflecting reality. It can then be used as a basis for comparison between different areas, and as a constraint in developing further models of the structure and composition of the Earth.

As a general rule, the oceanic lithosphere is simpler and less heterogeneous than its continental counterpart, presumably in large part due to its relative youthfulness, to its common mode of generation at mid-ocean ridges, and to the limited extent of tectonic and magmatic modification subsequent to its initial formation.

I am grateful to D. McKenzie for assistance with figure 1, and to C. J. Hawkesworth, J. Jackson, T. Jordan, N. Kusznir, D. McKenzie, A. B. Watts and N. White for discussions on the nature of the lithosphere. Department of Earth Sciences, Cambridge, contribution number 5382.

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