Earth and Planetary Science Letters, 95 (1989) 85–96
Elsevier Science Publishers B.V., Amsterdam – Printed in The Netherlands

[2]

Secular trends in the thermal evolution of metamorphic terrains

Michael Sandiford

Department of Geology, University of Adelaide, G.P.O. Box 498, Adelaide, S.A. 5001 (Australia)

Received May 17, 1989; revised version accepted July 27, 1989

The thermal evolution of convergent orogens is sensitive to the deformation history of the mantle part of the lithosphere. If the dense mantle lithosphere thickens as a consequence of convergent deformation and remains attached during subsequent denudation of the thickened crust then *high-pressure / low- to intermediate-temperature* glaucophane schist assemblages will form during both thickening and denudation; the actual form of the *P-T* paths for a given deformation depending primarily on the initial thermal state of the lithosphere and, particularly, the ratio of the thickness of the crust and mantle lithosphere. However, if mantle lithosphere is thinned during convergence by processes such as thermal boundary layer detachment then any early formed *high-pressure / low- to intermediate-temperature* glaucophane schist assemblages will be obliterated during subsequent *high-temperature* overprints. Since the stability of the thermal boundary layer during convergent deformations is controlled by the rate of convergence and the convective vigour of the subjacent mantle, any change in the convective vigour of the mantle through geologic time in response to, for example, changing global heat production may therefore be reflected in the thermal evolution of metamorphic terrains formed during convergent deformations. Models of the thermal evolution of deforming lithosphere appropriate to an earth in which the effect of declining global heat production and convective vigour is mediated by the process of thermal boundary layer detachment suggest that the preservation of *high-pressure / low- to intermediate-temperature* glaucophane schist assemblages in convergent orogens has become increasingly probable through geological time; a suggestion in accord with the observed secular trend in the character of metamorphic terrains preserved in the geological record.

1. Introduction

Any unifying theory for the early thermal evolution of the continental lithosphere must attempt to reconcile two distinct and somewhat incongruous sets of observations. The first is the set of observations supporting the notion that the Archaean continental lithosphere was, at least in part, as thick as the modern continental lithosphere [1–4]. The second set relates to the character of the Precambrian metamorphic record which is dominated by *low- to intermediate-pressure / high-temperature* amphibolite and granulite terrains with virtually no *high-pressure / low- to intermediate-temperature* glaucophane schist terrains and which therefore is clearly different from the Phanerozoic metamorphic record [5,6]. The first set of observations points to an essentially time-invariant continental lithospheric thermal structure and consequently it is not clear why there should be any secular trend in the character of continental metamorphic rocks preserved in the geological record.

Regional metamorphism in the continental crust takes place mainly in orogens formed during lithospheric convergence such as continental collisions. In such environments the thermal evolution of the lithosphere is strongly dependent on the way in which the lower part of the lithosphere and, in particular, the lower thermal boundary layer responds to the deformation [7–11]. Houseman et al. [7] have shown that a thermal boundary layer thickened during convergent deformation may detach from the lithosphere and they have suggested that such detachment may effect dramatic thinning of the overlying mantle part of the lithosphere. The stability of a thermal boundary layer to convergent deformation is controlled largely by the vigour of mantle convection in the underlying asthenosphere, with the likelihood of detachment increasing with more vigorous convection as well as faster convergence rates [7]. The vigour of a convective system is related to the rate at which thermal energy must be dissipated and since global heat production has declined through geologic time [12–15] it is probable that mantle convective vigour

0012-821X/89/$03.50

has also declined. While the magnitude of the secular decline in mantle convective vigour is poorly constrained because of the difficulty in modelling convection in the earths interior some simple calculations suggest the possibility of a significant decline. The vigour of mantle convection for an internally heated system is proportional to $\sqrt{R}$, where R is the Rayleigh number for convection, which in turn is inversely proportional to the viscosity [15,16]. A decline in mean mantle potential temperature of $\sim 200\,^\circ C$ since the end of the Archaean is suggested by komatiite petrogenetic constraints [12–14] as well as by the imbalance between heat production and heat loss in the modern earth [14]. This implies a corresponding increase in mantle viscosity, possibly by as much as two orders of magnitude [17,18], raising the possibility that the convective vigour of the Archaean mantle may have been significantly greater than in the modern earth; by as much as an order of magnitude if the vertical dimension of convection has remained constant through this interval. Houseman et al. [7] suggest that such high Rayleigh number convection may effect the very efficient detachment of thermal boundary layers thickened during lithospheric convergent deformations and therefore may significantly influence the thermal evolution of the overlying lithosphere. In this paper I describe models of the thermal evolution of the continental lithosphere for convergent (collisional) deformations involving different thermal boundary layer responses appropriate to an earth in which the convective vigour of the mantle has declined significantly with time. These models highlight the profound importance of the thermal boundary layer in the thermal evolution of crustal metamorphic terrains [7,9] and lend support to the notion that changes in the response of the thermal boundary layer to lithospheric deformations may be responsible for the observed secular trend in the thermal evolution of metamorphic terrains.

2. Thermal boundary layers and lithospheric deformation

Beneath a thermally stabilised lithosphere a lower thermal boundary layer must separate the lithosphere where steady state heat transfer is by conduction (termed the mechanical boundary layer

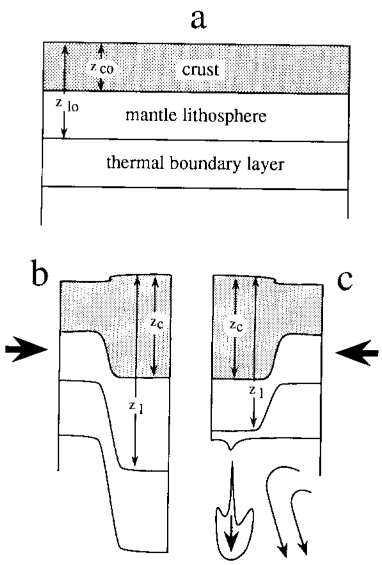

Fig. 1. (a) Schematic illustration of the structure of a thermally stabilised continental lithosphere showing a thermal boundary layer separating the mechanical lithosphere from the asthenosphere. During horizontal shortening deformations the thermal boundary layer may remain attached such that the lithosphere thickens at the same rate as the crust above ($f_l = f_c$), where $f_l = z_l z_{lo}$ and $f_c = z_c/z_{co}$ (b), or it may detach and cause thinning of the mantle part of the overlying lithosphere [7] such that $f_l < f_c$ (c).

or mechanical lithosphere) from the asthenosphere below where conductive heat transfer is insignificant compared to convective heat transfer (Fig. 1). The thickness of the thermal boundary layer, where both conduction and convection contribute to the observed thermal profile, reflects the temperature dependence of mantle viscosity; beneath the modern continental lithosphere the thermal boundary layer is estimated to be ~ 40 km thick with a corresponding temperature difference of ~ 200–$300\,^\circ C$ [1,19,20].

In modern convergent orogens involving continental lithosphere, shortening can be accommodated in the crust by thickening by a factor of up to ~ 2; the limit apparently applying when the buoyancy forces arising from the existence of the isostatically compensated deformed crust balance the driving forces for convergence. The response of the lower mantle part of the mechanical lithosphere and the thermal boundary layer to shortening deformations is less precisely understood, although Houseman et al. [7] have suggested that the thermal boundary layer may detach from the lithosphere during shortening and may effect thinning of the overlying mantle part of the lithosphere. Exactly how this thinning of the lithosphere is accommodated is not well understood (alternative geometries involving decoupling of crustal and mantle lithospheric strain have been suggested by Bird [8]). The study by Houseman et al. [7] does however raise the possibility of a number of qualitatively distinct behaviours in response to thermal boundary layer detachment; the lithosphere may: (1) thicken at a slower rate than the crust, (2) remain at a constant thickness while the crust thickens or (3) thin while the crust thickens.

If we represent the initial (or pre-deformation) and the deformed thickness of the lithosphere and crust as z_{lo}, z_l, z_{co} and z_c, respectively (Fig. 1), then the state of the deformation at any time can be described by the parameters $f_l = z_l/z_{lo}$ and $f_c = z_c/z_{co}$ (note that f is related to the reciprocal of the stretching factor, β, commonly used to describe extensional deformations). Since all possible geometries of lithospheric deformation, including both compressional and extensional deformation, can be described by variations in lithospheric and crustal thickness it is useful to illustrate the deformation path in the $f_l - f_c$ plane (Fig. 2) as well as the thermal, isostatic and mechanical responses to deformation. A useful parameter in describing the response of the lithosphere in this form is the ratio, ψ, of the initial crustal thickness to initial lithospheric thickness ($\psi = z_{co}/z_{lo}$). The qualitatively distinct behaviours described above correspond to the conditions: $1.0 < f_l < f_c$, $f_l = 1.0$, and $f_l < 1.0$, where $f_c \gg 1.0$. In the extreme case Houseman et al. [7] have suggested that the mantle lithosphere may be almost entirely removed with a detaching boundary layer, with the consequent juxtaposition of the lower crust and asthenosphere.

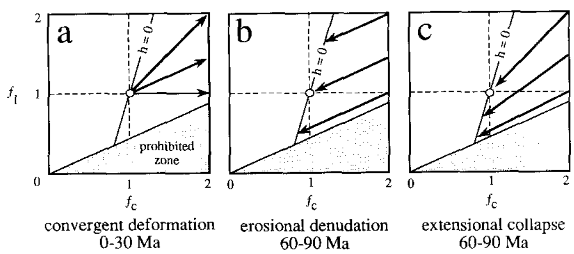

Fig. 2. Any deformation of the continental lithosphere can be represented in $f_l - f_c$ space, where $f_l = z_l/z_{lo}$ and $f_c = z_c/z_{co}$ (see Fig. 1) with most natural deformations described by values of f_l and f_c in the range 0–2. For crustal thickening deformations ($f_c > 1.0$), the possibility of detachment of the lower thermal boundary layer and consequent thinning of the mantle part of the mechanical lithosphere suggests the possibility of a number of different lithospheric scale strain histories: the paths shown here illustrate the deformations modelled in this paper. Fig. 2a shows the crustal thickening paths. Fig. 2b shows the denudation paths for models in which the denudation occurs by erosion. Fig. 2c shows the denudation paths for the models in which denudation occurs by extensional collapse. The line $h = 0$ is the line of constant surface elevation (see Fig. 8). The region in the lower right is prohibited because here the crustal thickness exceeds the lithospheric thickness. The size of the prohibited region depends on ψ, the ratio of the initial crustal thickness to initial lithospheric thickness ($\psi = z_{co}/z_{lo}$), and has an upper bound $f_l = f_c \times \psi$. The deformation paths shown here are simple straight lines in $f_l - f_c$ space, although in reality more complicated paths are likely (see Fig. 7).

3. Thermal models of lithospheric deformation

The different responses of the thermal boundary layer outlined above profoundly affect the amount of heat flowing into the crust during the deformation. If the thermal boundary layer remains attached and the lithosphere thickens at the same rate as the crust ($f_l = f_c$) then the heat flux across the Moho, q_m, will decrease as the mantle part of the lithosphere thickens. Conversely, if thermal boundary layer detachment effects the removal of a significant proportion of the lithosphere q_m may increase dramatically during the deformation. In order to model the thermal effects of such deformations it is therefore necessary to treat q_m as a variable parameter rather than an imposed condition. This can be achieved by considering how the different deformation geometries (corresponding the different thermal boundary layer responses) affect the depth to the critical isotherm at the base of the lithosphere T_{ML} which is here taken to be 1150 °C [10] (strictly, this is the base of the mechanical lithosphere). In the limiting case where the thermal boundary layer remains attached to, and thickens at the same rate as the lithosphere, the depth to T_{ML} will increase at the same rate as the depth to the Moho (i.e., $f_l = f_c$). With thermal boundary layer detachment, the depth to T_{ML} will increase at a slower rate than the depth to the Moho (i.e., $1.0 < f_l < f_c$) or, in extreme cases, will decrease ($f_l \leqslant 1.0$); the actual rate and extent of change in lithospheric thickness depending not only on the shortening rate but also on the rate, extent and timing of thermal boundary layer detachment. It is important to note that thermal boundary layer detachment is not the only way to decouple strains between the crust and mantle part of the lithosphere. Continued subduction of mantle lithosphere beneath a thickening crust provides one alternative mechanism. Moreover, the large rheological contrast across the Moho raises the possibility that mantle response to convergent deformations may in general be considerably different than the response of the overlying crust resulting in considerable disparity between f_l and f_c.

Thermal perturbations caused by discontinuities in the internal geometry of lithospheric deformation are short lived compared to the thermal evolution of the lithosphere as a whole, because the time constant for the equilibration of thermal perturbation is proportional to the square of the wavelength of the perturbation. Thus the effects of the different thermal boundary layer responses can be modelled by treating the lithosphere as a relatively homogeneous entity with respect to the deformation. Since we are largely concerned with the effects of the thermal boundary layer where, for horizontal shortening deformations, the pressure–temperature trajectories do not intersect the mantle solidus, heat contributions by advection of melts can be ignored. Finally, since the horizontal length scales in collisional orogens are generally large compared to the thickness of the lithosphere the thermal effects of perturbations at the base of the lithosphere can be treated as a one-dimensional problem in which heat transfer occurs only in the vertical direction. These considerations allow the effects of a thermal boundary layer detachment to be modelled by solving the diffusion–advection equation in one dimension:

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} - v \frac{\partial T}{\partial z} + \frac{h}{c_p \rho} \tag{1}$$

where c_p is the heat capacity; κ is diffusivity, v is the vertical velocity, h is the volumetric heat production, and ρ is the density. The models presented here follow from a Crank-Nicolson finite difference approximation to equation (1), assuming temperature-independent thermal diffusivity. In all the models I have assumed an initial crustal thickness of 35 km. Since the thickness of the continental lithosphere is poorly constrained I have constructed models for variable initial lithospheric thicknesses of 75, 100, 125 and 150 km, corresponding to $\psi = 0.47$, 0.35, 0.28 and 0.23, respectively. Corresponding depths to the base of the thermal boundary layer (1250–1300 °C) are ~ 100–200 km and thus span the plausible spectrum of depths to the asthenosphere under the modern continents. The distribution of heat sources is modelled as an exponentially decreasing function of depth, with the length scale, h_r, modified to give a typical continental steady state initial surface heatflow of 65 mW m^{-2} (in all models conductivity = 3 W m^{-1} K^{-1}, surface heat production = 3.0 μW m^{-3}, thermal diffusivity = 1 μm^2 s^{-1}, density of crustal rock = 2850 kg m^{-3}, density of mantle rock = 3300 kg m^{-3}). This corresponds to a Moho temperature of 570 °C for $\psi = 0.47$ ($h_r = 7.6$ km, $q_m = 44$ mW m^{-2}), 490 °C for $\psi = 0.35$ ($h_r = 12.1$ km, $q_m = 30$ mW m^{-2}), 460 °C for $\psi = 0.29$ ($h_r = 14.5$ km, $q_m = 23$ mW m^{-2}) and 440 °C for $\psi = 0.23$ ($h_r = 16.2$ km, $q_m = 18$ mW m^{-2}). Importantly, the constant initial surface heatflow condition results in a strong correlation between lithosphere thickness and total crustal heat production and an inverse correlation between lithosphere thickness and q_m.

The thermal effects of the different thermal boundary layer responses have been compared using an arbitrary 90 Ma orogenic history (Fig. 2). The lithosphere has been subject to crustal thickening (with a limiting crustal thickening factor $f_{c,max} = 2.0$) in the interval 0–30 Ma (corresponding to vertical strain rates of $\sim 2 \times 10^{-15}$ s^{-1}) for three limiting values of $f_{l,max} = 1.0$ (corresponding to extensive thermal boundary layer detachment), $f_{l,max} = 1 + \psi$ (limited thermal boundary layer detachment which maintains a constant thickness of mantle lithosphere) and $f_{l,max} = 2.0$ (no thermal boundary layer detachment or homogeneous thickening, see Fig. 2a). The deformed lithosphere has then been maintained at constant thickness from 30 to 60 Ma (i.e., no deformation or erosion). Finally, the deformed lithosphere has been denuded either by erosion or by extensional collapse in the period of 60–90 Ma so as to restore the initial elevation of the initial reference lithosphere (the deformation paths for the models involving erosional denudation are illustrated in Fig. 2b and for extensional collapse in Fig. 2c). The erosion

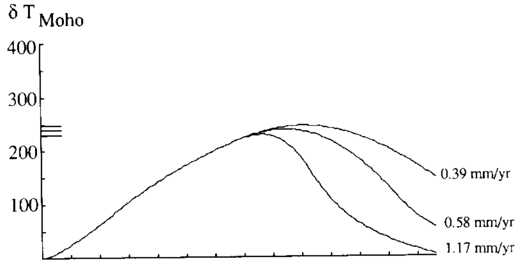

Fig. 3. Change in Moho temperature (δT_{Moho}) plotted as a function of time for the deformation $\psi = 0.35$, $f_{c,\max} = 2.0$, $f_{l,\max} = 1.35$ with erosion rates of 1.17 mm yr^{-1} (corresponding to the removal of all topography in 30 Myr), 0.58 mm yr^{-1} (removal of all topography in ~ 60 Myr) and 0.39 mm yr^{-1} (removal of all topography in 90 Myr). The erosion rate has little effect on the maximum temperature attained at the Moho, which differs by only some 20 ° C in these three models.

rates are towards the fast end of the plausible spectrum [9]. However, slowing the erosion rate by a factor of 2 or 3 makes little difference to the maximum temperatures attained during denudation, and therefore is of little consequence to the type of metamorphic terrains formed, and eventually preserved, within the crust (Fig. 3).

In the models presented here I have assumed that the detachment of the lower thermal boundary layer occurs continuously as it is shortened and that the consequent thinning of the lithosphere is accommodated by a homogeneous strain in the mantle part of the lithosphere. These assumptions lead to straight line deformation paths on the $f_c - f_l$ plane; the possibility of more complex, and probably more realistic, deformation paths is discussed in section 4. The model results are presented here in two forms: as $P - T - t$ paths (Figs. 4 and 5) and as relative change in Moho temperature (δT_{Moho}) as a function of time (Fig. 6) for select deformations ($f_{c,\max} = 2.0$, $f_{l,\max} = 1.0$, $1.0 + \psi$ and 2.0).

In the models appropriate to no detachment (i.e., where the lithosphere and crust thicken at the same rate, $f_l = f_c$), crustal metamorphic terrains undergo only minor heating ($\delta T_{\text{Moho}} < 60$ ° C for $\psi = 0.47$, $\delta T_{\text{Moho}} < 180$ ° C for $\psi = 0.23$) as they are buried and then decompressed on a timescale of 90 Ma (Figs. 4–6). With $f_{l,\max} = f_{c,\max} = 2.0$, erosional denudation sufficient to reduce the topography of the deformed lithosphere to that of the reference lithosphere (Fig. 2b) results in the eventual exposure of terrains which have been buried to a maximum depth of 20–30 km, and which have experienced maximum temperatures of ~ 350–400 ° C (Fig. 5) while decompressing through the depth range 25–30 km (corresponding to ~ 8–9.5 kbar). For the thermal conditions assumed here the metamorphic terrains eventually exposed at the end of the 90 Ma orogenic cycles terminated by erosion will exhibit glaucophane schist assemblages for initial lithospheres with $\psi > 0.3$ (Fig. 5). The amount of erosion in models with $\psi < 0.3$ is not sufficient to excavate glaucophane schist terrains, however much of the crust preserved following erosion will have glaucophane schist assemblages. Moho temperatures in all cases are less than ~ 650 ° C and since this is too low for any significant crustal melting implies that the results of the model calculations presented here will not be seriously invalidated by the magmatic redistribution of heat within the lithosphere.

In contrast, in models appropriate to significant detachment ($f_l \ll f_c$) crustal metamorphic terrains undergo significant heating on orogenic timescales. For deformation at constant lithosphere thickness ($f_{l,\max} = 1.0$), Moho temperatures increase by between 90% ($\delta T_{\text{Moho}} = 510$ ° C, $\psi = 0.47$) and 68% ($\delta T_{\text{Moho}} = 300$ ° C, $\psi = 0.23$). Pressure temperature trajectories for the lower half of the crust pass through the *high-pressure / low- to intermediate-temperature* glaucophane–schist stability field during tectonic burial but are heated out of this field following burial. For deformation at constant mantle lithosphere thickness ($f_{l,\max} = 1 + \psi$) maximum Moho temperatures exceed the wet granite solidus (~ 650 ° C). Assuming that temperatures of at least 750–800 ° C are required for the generation of significant quantities of granitic melt, then the deformation required for granite genesis is strongly dependent on ψ. For deformations which produce such high lower crustal temperatures the model conditions will clearly be violated by magmatic redistribution of heat, and realistic thermal regimes, particularly in intermediate to shallow crustal levels are likely to be considerably hotter than predicted by these models giving rise to the possibility of *high-T / low-P metamorphism* in the upper part of the crust. Independently of the amount of heat advected by magmatic processes, the metamorphic terrains eventually preserved and exposed at the end of the orogenic cycles terminated by erosion and involving significant thermal boundary layer detachment will exhibit *high-temperature* assemblages. The model shown in Figs. 4c and 5c corresponds to a deformation which almost entirely removes the mantle part of the lithosphere (only 5

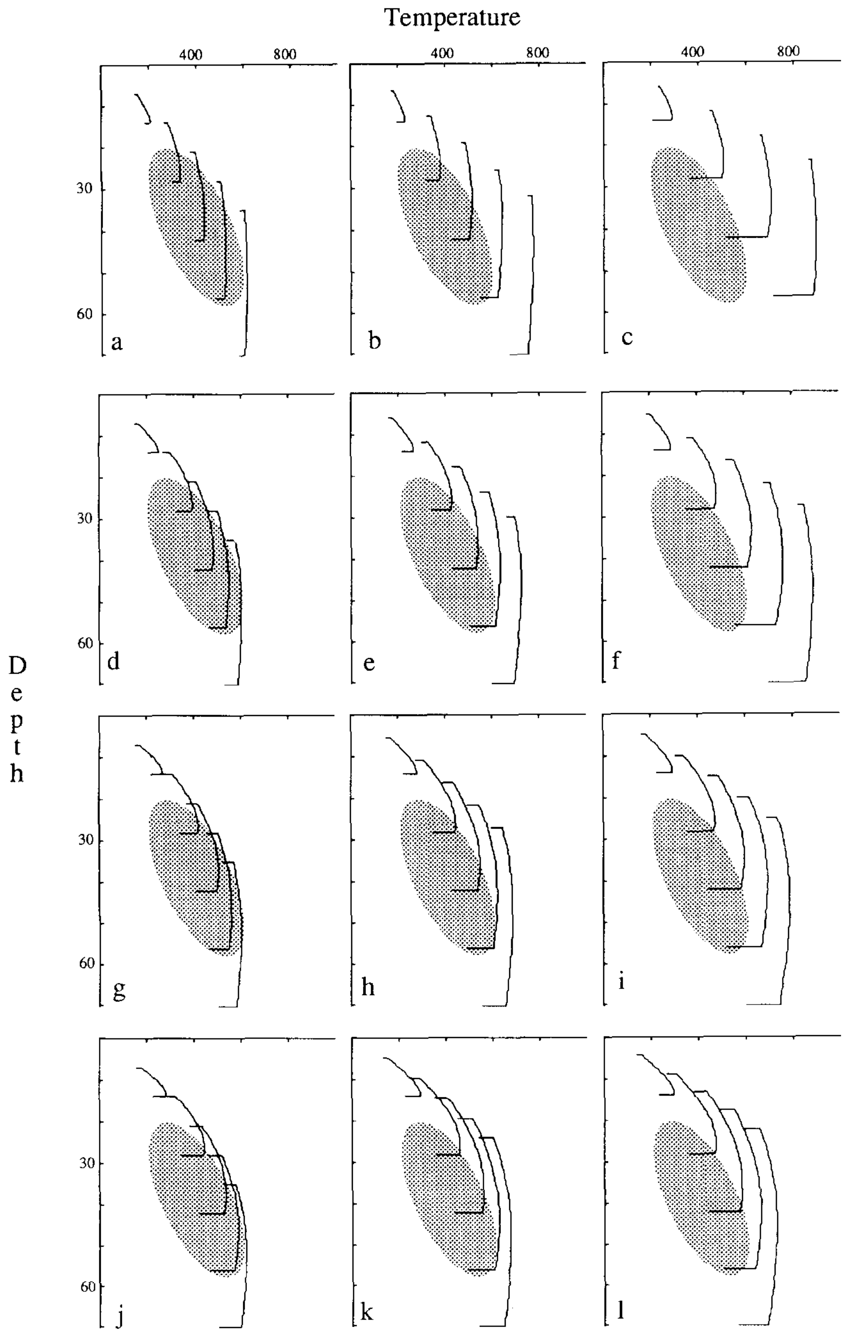

Fig. 4. Pressure–temperature–time paths for the crustal part of the lithosphere for orogenic cycles involving extensional collapse of thickened crust. Individual paths are shown for rocks spaced at 7 km depth intervals in the initial lithosphere. For clarity, only the post-thickening portions of the paths are shown. Stiple shows region in which glaucophane schists assemblages are stable [9]. Fig. 4a–c correspond to $\psi = 0.47$; Fig. 4d–f to $\psi = 0.35$; Fig. 4g, h to $\psi = 0.28$; Fig. 4i–l to $\psi = 0.23$. Deformation parameters (see text for details) are $f_{c,max} = 2.0$ and $f_{l,max} = 2.0$ in a, d, g and j, $f_{l,max} = 1 + \psi$ in b, e, h and k; $f_{l,max} = 1.0$ in c, f, i and l.

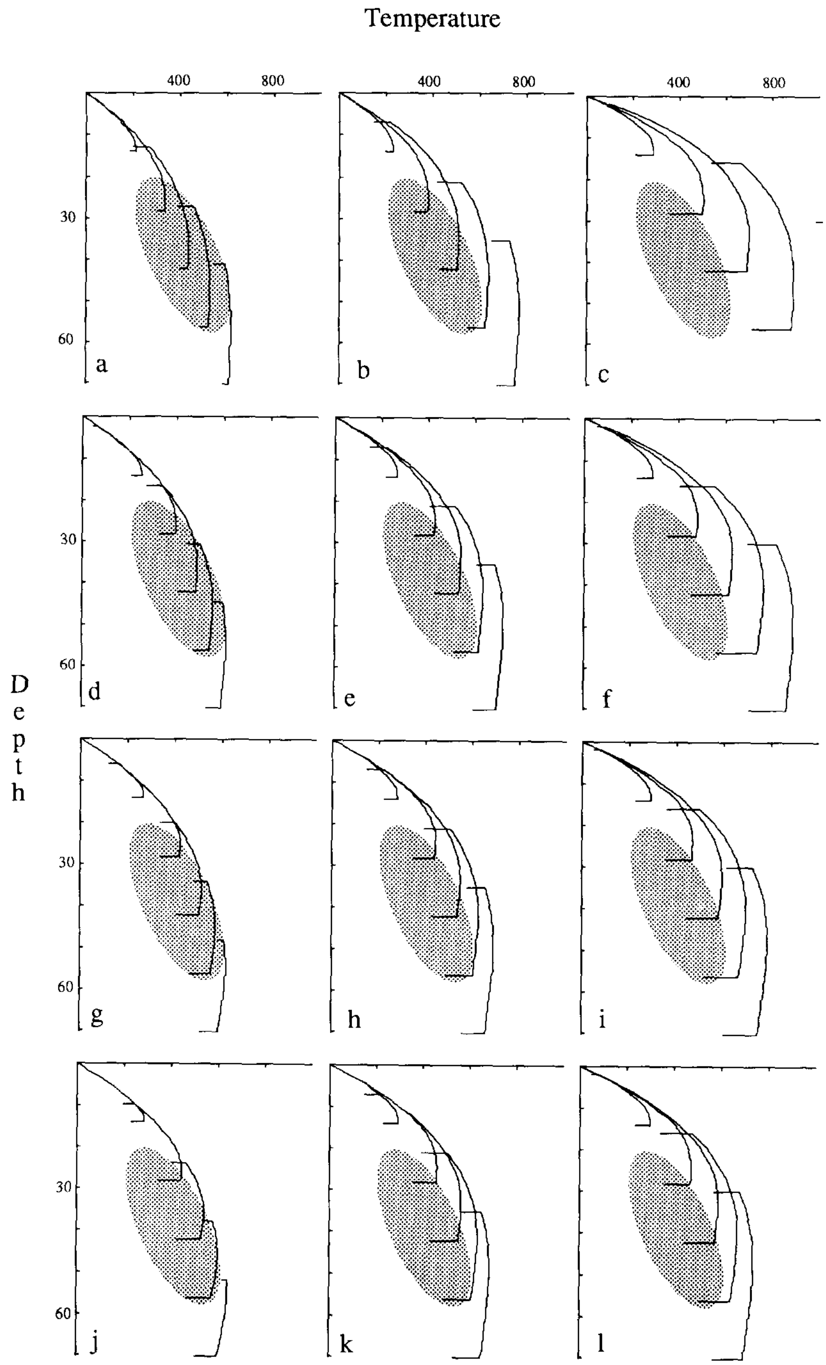

Fig. 5. Pressure–temperature–time paths for the crustal part of the lithosphere for orogenic cycles involving erosional denudation of thickened crust. Individual paths are shown for rocks spaced at 7 km depth intervals in the initial lithosphere. For clarity, only the post thickening portions of the paths are shown. Stiple shows region in which glaucophane schists assemblages are stable [9]. Fig. 5a–c correspond to $\psi = 0.47$; Fig. 5d–f to $\psi = 0.35$; Fig. 5g, h to $\psi = 0.28$; Fig. 5i–l to $\psi = 0.23$. Deformation parameters (see text for details) are $f_{c,max} = 2.0$ and $f_{l,max} = 2.0$ in a, d, g and j, $f_{l,max} = 1 + \psi$ in b, e, h and k; $f_{l,max} = 1.0$ in c, f, i and l.

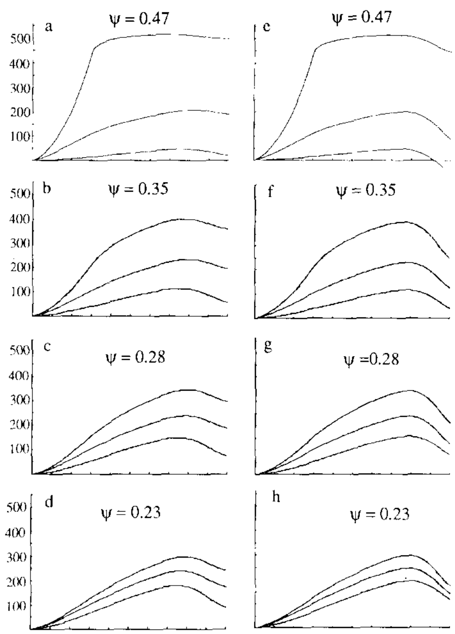

Fig. 6. Change in Moho temperature (δT_{Moho}) plotted as a function of time for the deformations shown in Fig. 2. Fig. 6a–c show results of models where denudation is achieved by extensional collapse while Fig. 6d–f show results of models with erosional denudation. In each case three curves are shown representing, respectively, deformations with $f_{\mathrm{l,max}} = 2.0$ (lowermost curve), $f_{\mathrm{l,max}} = 1 + \psi$, and $f_{\mathrm{l,max}} = 1.0$ (uppermost curve).

km separates the Moho from the base of the lithosphere at the end of convergent deformation). The consequent heating of the crust is extreme allowing the possibility of almost substantial fusion of parts of the lower crust (Fig. 6a and e).

Figs. 4–6 clearly show that the response of the lower lithosphere has a more profound effect on the crustal thermal regime in the large-ψ (or thin initial lithosphere) models ($\psi = 0.35$–0.47) than in the small-ψ (or thick initial lithosphere) models ($\psi = 0.23$–0.28). This is primarily because of the way in which the initial thermal regime of the lithosphere has been configured; in the large-ψ models the heatflow from the mantle provides a much greater contribution to the lithospheric thermal budget than in the small-ψ models.

4. Discussion

The results of the numerical models presented here reflect, as much as anything, the way in which convergent strain history in the lithosphere has been configured, and especially how the strain has been distributed in the lower part of the lithosphere. While such models must be sensitive to the assumed thermal parameters and to the rates of deformation [9], the models presented here illustrate that the behaviour of the lower lithosphere during convergent deformation provides a profound control on the thermal evolution of the crust for a wide range of initial thermal states. In line with the suggestion of Houseman et al. [7] I have allowed the possibility that the lithosphere may be thinned as a direct consequence of thermal boundary layer detachment. For want of better understanding, I have modelled the thinning of the lithosphere induced by detachment as a homogeneously distributed strain in the mantle part of the lithosphere. This is unlikely to be realistic (see below) and other deformation geometries will influence the precise thermal history of the lithosphere. Moreover I have assumed that the detachment of the thermal boundary layer occurs throughout convergence. The numerical experiments of Houseman et al. [7] show that this is unlikely; detachment probably occurs after some critical thickening of the thermal boundary layer and, importantly, once initiated may occur very rapidly (possibly as an effectively instantaneous event relative to the deformation history of the overlying lithosphere). However, Houseman et al. [7] have shown that the critical thickness for the onset of detachment, and the extent of detachment once initiated, is clearly related, and sensitive, to the vigour of convection in the subjacent mantle. Thus the models presented here, while not quantitatively appropriate no real collisional orogenic processes, are believed to provide a qualitative insight into the thermal responses of the lithosphere in a system where the effects of declining mantle convective vigour on collisional processes are mediated by the response of the thermal boundary layer.

In accordance with Houseman et al. [7] the models presented here indicate that the response of the thermal boundary layer exerts a profound control on the thermal evolution of crustal metamorphic terrains formed in zones of lithospheric shortening. If the lower thermal boundary layer remains attached to the lithosphere during the shortening then *high-pressure / low- to inter-*

mediate-temperature glaucophane schist terrains will be preserved during the denudation of the thickened crust and, ultimately, if denudation occurs by erosion exposed at the surface of the earth. However, if the lower thermal boundary layer is detached to any significant extent from the lithosphere then early formed glaucophane schist assemblages will be obliterated by higher-temperature metamorphic overprints during the denudation of the orogen. As such these models provide an insight into a mechanism which may explain the observed secular trend in the thermal evolution of metamorphic terrains, and more importantly, reconcile such trends with the apparently incongruous evidence for a time-invariant steady state continental thermal structure. The models emphasize that the thermal evolution of lithosphere in orogenic zones is not only dependent on the thermal state of the typical continental lithosphere (the initial lithospheres used in the models represent very different thermal configurations) but also on the nature of the interaction between the lithosphere and the underlying asthenosphere. In the case of the thermally stabilised lithosphere discussed here, this interaction will be mediated by the response of the thermal boundary layer which is particularly sensitive to the convective vigour of the subjacent mantle [7].

The deformation paths modelled here are straight lines in $f_l - f_c$ space (e.g., Fig. 2a) as a result of the way in which the convergent strain history of the lithosphere has been configured. The numerical experiments of Houseman et al. [7] suggest that more realistic deformation histories resulting from thermal boundary layer detachment will produce curved paths in $f_l - f_c$ space (Fig. 7). This is because some finite thickening of the thermal boundary layer is apparently necessary before detachment can be initiated. Curved paths such as shown in Fig. 7 allow considerable insight into the potential mechanical evolution of convergent orogens since such paths not only have implications for the amount of heat flowing into the crust but also for the isostatically compensated surface elevation of the lithosphere (Fig. 8). Since isostatically supported elevation contrasts produce significant horizontal deviatoric stresses in the lithosphere [21,22] such curved paths must have important consequences for the evolution of the force balance in convergent orogens. In models considered here the isostatically supported topography caused by a variation in $f_{l,max}$ of 1 is between 1.3 km ($\psi = 0.47$) and 2.6 km ($\psi = 0.23$) for the cases of $f_{l,max} = 2.0$ and $f_{l,max} = 1.0$ (Fig. 8). The gravitational potential energy increase induced by the deformation can be represented by the vertically integrated horizontal buoyancy force (F_b) per unit

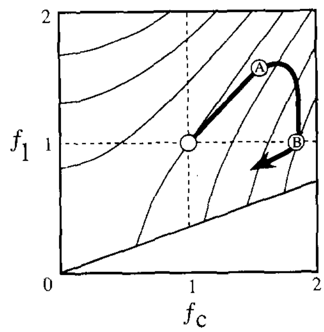

Fig. 7. Hypothetical deformation path in which the onset of thermal boundary layer detachment after some critical thickening (point A) leads to the thinning of the overlying lithosphere. The consequent thermal weakening of the lithosphere and the change in force balance induced by lithospheric thinning leads eventually (point B) to the extensional failure of the whole lithosphere (see text for discussion). Light lines represent force balance contours illustrated in Fig. 9 for $\psi = 0.35$.

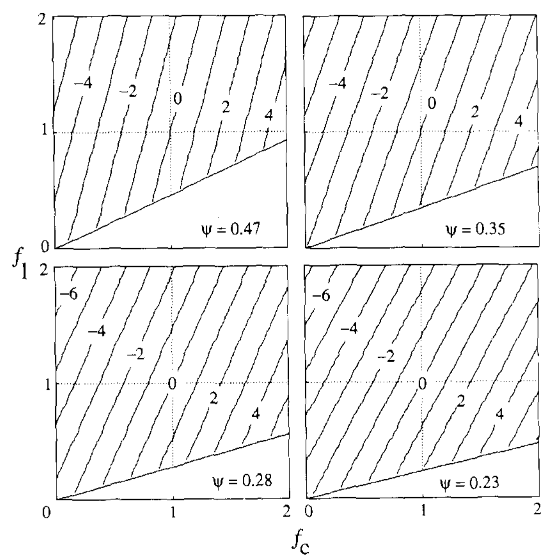

Fig. 8. The $f_c - f_l$ plane contoured for isostatically compensated elevation (see Appendix 1). Elevation contours are 1 km and represent the elevation contrast between the deformed and initial or reference lithosphere (i.e., $f_c = 1.0$, $f_l = 1.0$). In the case of negative elevation differences, no sedimentary or water filling is assumed.

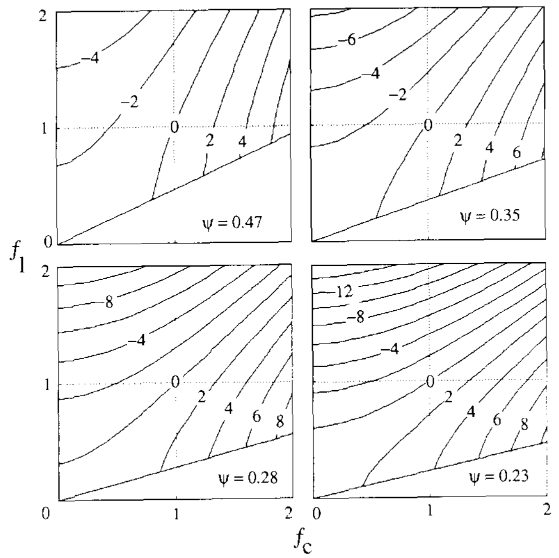

Fig. 9. The $f_c - f_l$ plane contoured for the horizontal buoyancy force (F_b) per unit length of orogen (in 10^{12} N m^{-1}) arising from the isostatically supported topographic gradients between the deformed lithosphere and surrounding undeformed lithosphere (see Appendix 1). Contour intervals are 2×10^{12} N m^{-1}.

length of orogen arising from elevation contrasts between the deformed lithosphere (assuming isostatic compensation) and surrounding undeformed lithosphere:

$$F_b = \int_s^{z_a} (\sigma_{zz})_d \, dz - \int_s^{z_a} (\sigma_{zz})_{ud} \, dz \qquad (2)$$

where $(\sigma_{zz})_d$ and $(\sigma_{zz})_{ud}$ are the vertical stresses in the deformed and undeformed lithosphere respectively, s is the surface of the lithosphere and z_a is the depth at which the density structure is equalised beneath both deformed and adjacent undeformed lithosphere. The density structure assumed here is outlined in Appendix 1, with the results shown in Fig. 9. The difference in F_b in orogens with double thickness crust ($f_{1,max} = 2.0$) varies considerably depending on f_l as well as ψ; the difference between $f_l = 2.0$ and $f_l = 1.0$ is 2×10^{12} N m^{-1} for $\psi = 0.47$–12×10^{12} N m^{-1} for $\psi = 0.23$ (Fig. 9). For a given f_c, F_b increases with decreasing f_l. Importantly, for $\psi < 0.29$ homogeneous lithospheric thickening, $f_l = f_c$, results in a vertically integrated tensional force in the adjacent undeformed lithosphere and potentially provides a self-localising mechanism resembling the development of subduction in zones of convergence in the oceanic lithosphere. In orogens with $f_l \ll f_c$, these forces may be of the same magnitude as, or even greater than, the driving forces for convergence (F_{dc}), in which case they must result in a vertically integrated extensional force in the orogen [11,22]. A process of thermal boundary layer detachment leading to curved paths such as shown in Fig. 7 may therefore lead to an evolutionary sequence involving the transition from a self-localising orogen ($F_b < 0 < F_{dc}$), to a spreading orogen ($0 < F_b = F_{dc}$) which builds outwards through time, eventually to a collapsing orogen ($0 < F_{dc} < F_b$).

A vertically integrated extensional force is a necessary, but not sufficient, condition for the extensional collapse of a collisional orogen [11]. Whether or not collapse occurs will depend on the strength of the lithosphere, which is dependent on its thermal state. Since detachment of the thermal boundary layer both increases the potential energy of, and the heat supply to, the overlying lithosphere, it may provide the necessary trigger for the termination of convergent deformation and the onset of collapse [11]. We may expect therefore that more efficient detachment of thickened thermal boundary layers beneath convergent orogens in early earth history due to more vigorous asthenospheric convection would have resulted not only to the obliteration of early formed *high-pressure / low- to intermediate-temperature* metamorphic assemblages but also in the development of metamorphic terrains which show evidence of extensional collapse. Elsewhere I have argued that the high-temperature deformation history of a number of Precambrian gneiss terrains may be best interpreted as the consequence of extensional deformation rather than thickening deformations [23,24]. It also interesting to note that two recent studies of the finite lifespan of Early Proterozoic orogens have produced compelling evidence for very short-lived orogenic histories involving high-temperature metamorphism [25,26]. The results of these studies suggest that termination of the mountain-building process was effected more quickly than in modern orogens at a stage where the crustal thickening was probably considerably less than 2. Assuming that the forces driving convergent orogens have not changed significantly through time, these observations are consistent with a model in which more vigorous mantle convection has influenced continental tectonics through the more efficient detachment of the thickened thermal boundary layer.

Acknowledgements

Many of the ideas and solutions presented here result from work undertaken while I was supported by a CSIRO postdoctoral fellowship (1986–1988) through the Division of Geomechanics; and I extend my appreciation to those concerned with the CSIRO fellowship scheme. Roger Powell is thanked for illuminating conversations on, and an enthusiastic response to $f_l - f_c$ diagrams and Mike Bickle for a constructive review.

Appendix 1

Calculation of the change in elevation and potential energy follows closely the development of Turcotte [22] using the density distribution with depth, $\rho(z)$:

in the crust:

$$\rho(z) = \rho_c + \rho_c \alpha (T_l - T_s)\left(1 - \frac{z}{f_l z_l}\right) \qquad \text{(A-1)}$$

in the mantle lithosphere:

$$\rho(z) = \rho_l + \rho_l \alpha (T_l - T_s)\left(1 - \frac{z}{f_l z_l}\right) \qquad \text{(A-2)}$$

here ρ_l and ρ_c are the density of mantle and the crust at the temperature, T_l, at the base of the lithosphere at depth z_l, T_s is the temperature at the surface of the lithosphere, and α is the volumetric coefficient of thermal expansion. In solving equation (2) I have assumed the same thermal expansion for crust and mantle, and that the geotherm is linear (i.e., no heat production), with no conductive equilibration during deformation; these assumptions introduce negligible error as the α terms are second order. Calculation of the isostatically-supported elevation change involves equating the vertical stress at a common depth immediately beneath the lithosphere, for the undeformed and the deformed lithosphere respectively. The buoyancy force per unit orogen length, F_b, is given by the difference between the double integrals of density, both with respect to z, from the Earth's surface down to a common depth beneath the lithosphere, for the deformed and undeformed lithosphere respectively. The full equations are given in Sandiford and Powell [27].

References

1 F.M. Richter, Models for the Archaean thermal regime, Earth Planet. Sci. Lett. 73, 350–360, 1985.
2 P.C. England and M.J. Bickle, Continental thermal and tectonic regimes in the Archaean, J. Geol. 92, 353–367, 1984.
3 D. McKenzie, E. Nesbit and J.G. Sclater, Sedimentary basin development in the Archaean, Earth Planet. Sci. Lett. 48, 35–41, 1980.
4 S.H. Richardson, J.J. Gurney, A.J. Erlank and J.W. Harris, Origin of diamonds in old enriched mantle, Nature 310, 198–202, 1984.
5 J.V. Watson, Precambrian thermal regimes, Philos. Trans. R. Soc. London 288, 431–440, 1976.
6 J.A. Grambling, 1981, Pressures and temperatures in Precambrian metamorphic rocks, Earth Planet. Sci. Lett. 53, 63–68, 1981.
7 G.A. Houseman, D.P. McKenzie and P. Molnar, Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts, J. Geophys. Res. 86, 6115–6132, 1981.
8 P. Bird, Continental delamination and the Colorado plateau, J. Geophys. Res. 84, 7561–7571, 1979.
9 P.C. England and A.B. Thompson, Pressure–temperature–time paths of regional metamorphism, 1. Heat transfer during the evolution of regions of thickened crust, J. Petrol. 25, 894–928, 1984.
10 L.J. Sonder, P.C. England, B.P. Wernicke and R.L. Christiansen, A physical model for Cenozoic extension of western North America, in: Continental Extensional Tectonics, M.P. Coward, J.F. Dewey and P.L. Hancock, eds., Geol. Soc. London, Spec. Publ. 28, 1987.
11 P.C. England and G.A. Houseman, The mechanics of the Tibetan Plateau, Philos. Trans. R. Soc. London, Ser. A326, 301–320, 1988.
12 M.J. Bickle, Heat loss from the earth: a constraint on Archaean tectonics from the relation between geothermal gradients and the rate of plate production, Earth Planet. Sci. Lett. 40, 301–315, 1978.
13 M.J. Bickle, Implications of melting for stabilisation of the lithosphere and heat loss in the Archaean, Earth Planet. Sci. Lett. 30, 314–324, 1976.
14 U.R. Christensen, Thermal evolution models for the earth, J. Geophys. Res. 90, 2995–3007, 1985.
15 S.F. Daly, Convection with decaying heat sources: constant viscosity, Geophys. J. R. Astron. Soc. 61, 519, 1980.
16 D.P. McKenzie and F.J. Richter, Parameterized thermal convection in a layered region and the thermal history of the earth, J. Geophys. Res. 86, 6678, 1981.
17 D.L. Turcotte and G. Schubert, Geodynamics, Applications of Continuum Physics to Geological Problems, John Wiley and Sons, New York, N.Y., 1982.
18 J. Weertman, The creep strength of the earths mantle, Rev. Geophys. Space Phys. 8, 145–168, 1970.
19 B. Parsons and D. McKenzie, Mantle convection and the

thermal structure of the plates, J. Geophys. Res. 83, 4485–4496, 1978.

20 D. McKenzie and M.J. Bickle, The volume and composition of melt generated by extension of the continental lithosphere, J. Petrol. 29, 625–679, 1988.

21 E.V. Artyushkov, 1973, Stresses in the lithosphere caused by crustal thickness inhomogeneities, J. Geophys. 78, 7675–7705, 1973.

22 D.L. Turcotte, Mechanisms of crustal deformation, J. Geol. Soc. London 140, 701–724, 1983.

23 M. Sandiford and R. Powell, Deep crustal metamorphism during continental extension; modern and ancient examples, Earth Planet. Sci. Lett. 79, 151–158, 1986.

24 M. Sandiford, Horizontal structures in granulite terrains; a record of mountain building or mountain collapse, Geology 17, 449–452, 1989.

25 P.F. Hoffman and S.A. Bowring, Short-lived 1.9 Ga continental margin and its destruction, Wompay Orogen, northwest Canada, Geology 12, 68–72, 1984.

26 J.A. Cooper, G.E. Mortimer and P.R. James, Rate of Arunta Inlier evolution at the eastern margin of the Entia Dome, Central Australia, Precambrian Res. 40/41, 217–231, 1988.

27 M. Sandiford and R. Powell, Some thermal and isostatic consequences of the vertical strain geometry in convergent orogens (in preparation).