Australian Proterozoic high-temperature, low-pressure metamorphism in the conductive limit

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Abstract: High-temperature, low-pressure (HTLP) metamorphism often reflects transient advection of heat due to magma ascent. However, the origin of HTLP metamorphism in a number of Australian Proterozoic terranes remains contentious either because of the deficiency of magmatic bodies in the terranes, or because the long time delay (>100 Ma) between magmatism and metamorphism precludes heating by existing magmatic bodies. Furthermore, a number of Australian Proterozoic HTLP terranes (such as the Reynolds Range in central Australia) show evidence of an extended history (*c.* 30 Ma) of HTLP mineral growth suggesting metamorphism during a thermal regime dominated by conduction at lithospheric length scales. Australian Proterozoic metamorphic terranes are characterized by both elevated modern-day heat flows (averaging *c.* 85 $mW\,m^{-2}$) and granitic gneisses with anomalously high heat production rates (commonly >5–10 $\mu W\,m^{-3}$). We show that the conditions required for HTLP metamorphism may result from conduction if the crustal heat production responsible for modern-day heat flows is concentrated at mid-crustal levels (15–20 km). Importantly, for low–intermediate mantle heat fluxes (10–20 $mW\,m^{-2}$) and moderate synmetamorphic crustal thicknesses (*c.* 45 km), the conductive geotherms attendant with such HTLP metamorphism do not necessarily lead to significant melting of a refractory lower crust. Importantly, the thermal regimes are very sensitive to the depths at which crustal heat production is localized. The strong dependence of the resulting geotherms on the depth of the heat-producing layer has the important consequence that only minor burial may be required to induce HTLP metamorphism, while only minor erosion (*c.* 5 km) is necessary to terminate the event.

High-temperature, low–intermediate-pressure (HTLP) metamorphism at temperatures in excess of about 600°C and pressures of less than about 5 kbar has been widely interpreted as resulting from transient thermal events associated with the ascent of magmas through the crust (Lux *et al.* 1978; DeYoreo *et al.* 1991; Collins & Vernon 1991; Sandiford *et al.* 1991). In many HTLP terranes the close spatial and temporal association between metamorphism and magma emplacement clearly supports this interpretation (Holdaway *et al.* 1988; Sisson *et al.* 1989; Sisson & Hollister 1988) which remains the governing paradigm for such metamorphism (Barton & Hanson 1989). Metamorphism governed by advective heat transfer should be characterized by dramatic lateral variations of grade and temporal transience (Sandiford *et al.* 1991, 1995*b*). As shown by Sandiford *et al.* (1991), the inverse exponential temperature dependence of crustal strength provides a logical reason for the coupling of deformation and heating in such terranes (e.g. Karlstrom & Williams 1995). In convergent orogens, such a thermomechanical coupling may be expected to be evidenced by 'anticlockwise' pressure–temperature–time paths.

In HTLP terranes that lack compelling field evidence or geochronological data supporting a close genetic association of magmatism and metamorphism, the necessity for magmatic heat advection has been advanced largely on the basis of thermal arguments. For example, Sandiford & Powell (1991) argued '...high-T–low-P metamorphic terranes cannot simply reflect the conductive response to crustal thickening since the resulting Moho temperatures would greatly exceed the crustal solidus.' With regard to this kind of thermal argument, it is important to realize that the uncertainty in the basic properties governing the thermal regime within the mid- to deep continental crust (such as thermal conductivities at elevated temperature and the distribution of heat-producing elements) implies a corresponding uncertainty in the deep crustal thermal structure. For example, available measurements would (conservatively) allow an uncertainty of at least 30% in the thermal conductivity structure of the mid- to deep crust, leading to equivalent uncertainties in deep crustal temperatures (that is, an uncertainty in Moho temperature of around 200°C due to conductivity uncertainties alone).

Recent studies in a number of Australian Proterozoic HTLP terranes (e.g. Mt Isa and northern Arunta Inliers) that contain abundant granitic bodies have shown that the main HTLP metamorphism post-dates the major granite

SANDIFORD, M., & HAND, M. 1998. Australian Proterozoic high-temperature, low-pressure metamorphism in the conductive limit. *In*: TRELOAR, P. J. & O'BRIEN, P. J. (eds) *What Drives Metamorphism and Metamorphic Reactions?* Geological Society, London, Special Publications, **138**, 109–120.

emplacement events by as much as 150 Ma (Jaques *et al.* 1982; Reinhardt 1992; Conners & Page 1995; Williams *et al.* 1996; Hand *et al.* 1995; Vry *et al.* 1996). Furthermore, thermochronological data in the Reynolds Range (Williams *et al.* 1995) indicate that elevated temperatures during the HTLP event were sustained for as much as 30 Ma, with little change in pressure, suggesting that synmetamorphic thermal regimes approached steady state.

A notable feature of many Australian Proterozoic HTLP terranes is the unusually high present-day surface heat production (Fig. 1). For example, in the Mesoproterozoic Mt Painter Inlier in South Australia, heat production in granitic gneisses typically exceeds 10 μW m^{-3} (Fig. 1a). The mean and median heat productions (recalculated at 1500 Ma) of 227 samples (granitic gneisses and metasediments) for which analyses are available are 16 and 8 μW m^{-3}, respectively. In the Reynolds and Anmatjira Ranges in central Australia, 89 analyses of granitic gneisses (which which constitute more than 60% of the terrane) indicate a mean heat production of *c.* 8 μW m^{-3} at the time of metamorphism some 1600 Ma ago (Fig. 1b). Similarly, high heat production values are known from regionally extensive granites in the Mt Isa region, such as the Sybella and Wonga granites (Loosveld 1989). Modern-day heat flows averaging around 85 mW m^{-2}, and locally in excess of 100 mW m^{-2}, through this Australian Proterozoic metamorphic province (Sass & Lachenbruch 1979; Cull 1982; Fig. 2) confirm the regional influence of these high heat production values but also imply that any near-surface, high heat production layer is of only moderate thickness. For example, considering that this province now forms part of a tectonically quiescent craton with a lithosphere estimated to be about 250 km thick (Zielhuis & van der Hilst 1996), we might expect reduced heat flows of around 10–15 mW m^{-2}, with the remaining 70–75 mW m^{-2} of the observed surface heat flow due to crustal sources. For the observed surface heat productions of between 5 and 10 μW m^{-3}

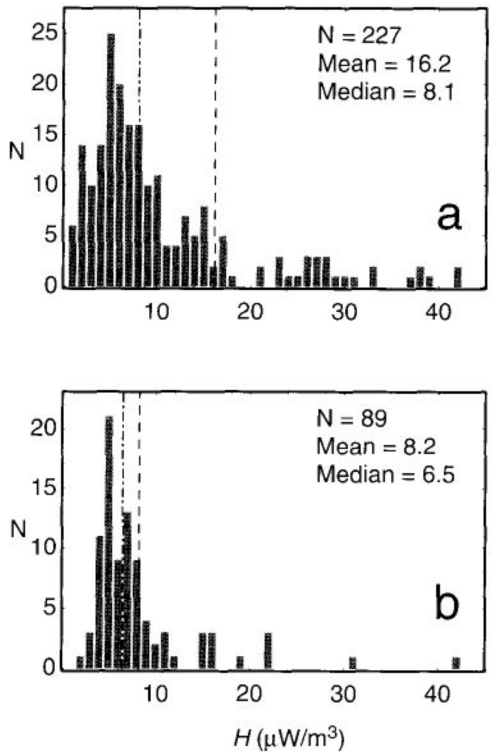

Fig. 1. Heat production values for some Australian metamorphic HTLP terranes. (**a**) Mesoproterozoic Mt Painter Inlier, South Australia. Data were compiled by N. Wall mostly from Minerals and Energy South Australia unpublished reports (227 analyses). (**b**) Anmatjira and Reynolds Ranges (total number of analyses is 89, and includes data from Australian Geological Survey Organisation). For clarity, extreme heat production values (in excess of 45 W m^{-3}) are not shown but have been incorporated in calculation of the mean and median of the distributions.

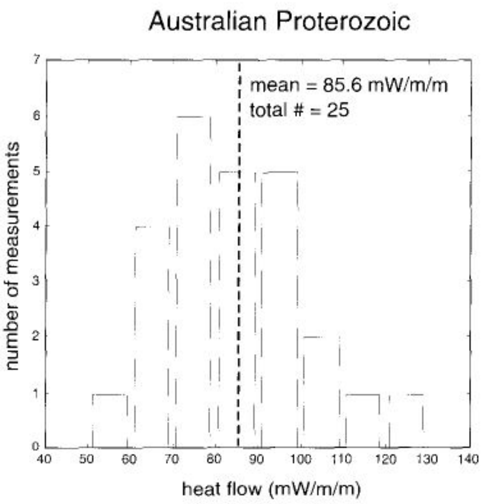

Fig. 2. Heat flow data from Australian Proterozoic terranes, illustrating the generally anomalous nature of this heat flow province (the data are taken from the tabulation of Cull 1982). Two outlying measurements from Warrego (160 mW m^{-2}) and Cattle Creek (48 mW m^{-2}) have been excluded from the analysis. The mean of 25 remaining measurements is 85 mW m^{-2}.

such a contribution would be provided by a layer of total thickness 7–15 km.

The notion that anomalous heat production associated with granites and/or metasediments within the metamorphic pile (as opposed to advective heating) provides a primary control on metamorphic grade in high-grade terranes has been suggested by Chamberlain & Sonder (1990), who were able to correlate variations in the metamorphic grade in New England with current heat flow and surface heat production. Similarly, unconformity-related HTLP metamorphism in the Mt Painter region of the northern Flinders Ranges, South Australia, has been attributed to the anomalous heat production in the basement (Mildren & Sandiford 1995). One obvious question raised by this discussion is *whether it is feasible to generate the conditions associated with HTLP metamorphism without generating significant quantities of melt in the deeper crust*. Motivated by this question, we seek to evaluate the plausibility of HTLP metamorphism in the steady-state thermal limit. Our approach is essentially parametric in that we are primarily concerned with evaluating the thermal parameter space that may allow steady-state HTLP at mid- to upper crustal levels without appreciably melting the deep crust. In the discussion we return to the question of applicability of our calculations to the metamorphic and geochemical character of select Australian HTLP terranes.

A model for heat production distributions in Australian Proterozoic HTLP terranes

The way in which heat production is distributed with depth in the continental crust has been a long-standing source of interest (e.g. Lachenbruch 1968). The upper crust has long been recognized as having significantly greater heat production than the lower crust, both from the analysis of surface heat flow–heat production data (Lachenbruch 1968) and from geochemical studies of the distribution of heat-producing

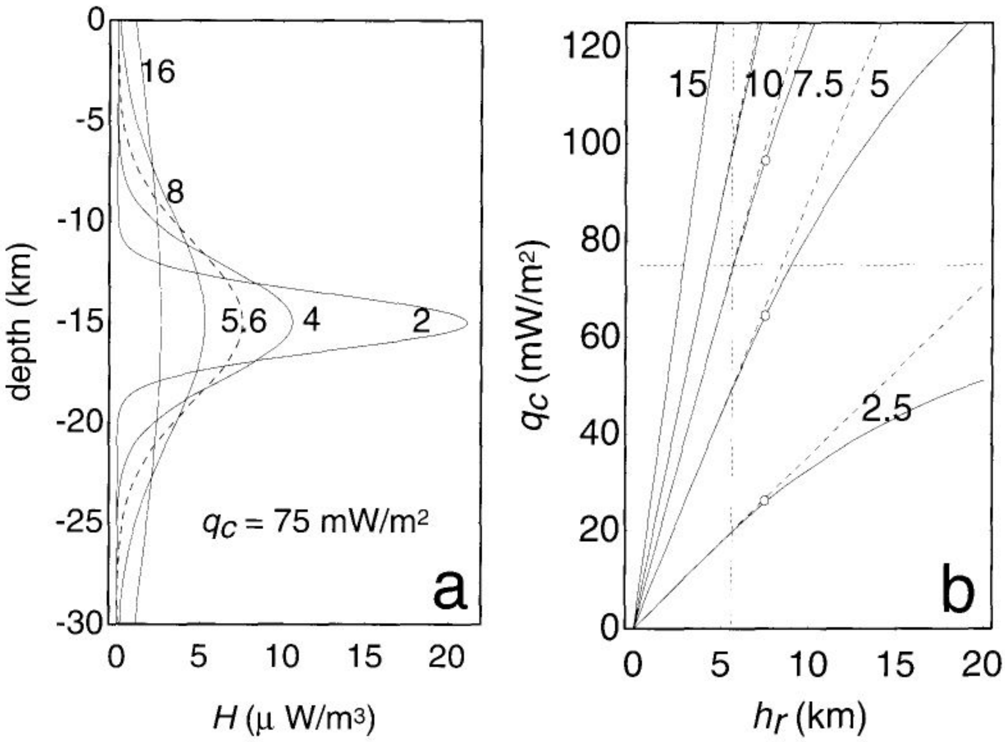

Fig. 3. (**a**) Heat production distribution specified by equation (2) for a range of values of h_r (2 km, 4 km, 5.6 km (dashed), 8 km and 16 km), with H_i set so the total crustal heat production $q_c = 75\,\mathrm{mW\,m^{-2}}$.
(**b**) Total crustal heat production values for a range in H_i (2.5 μW m^{-3}, 5 μW m^{-3}, 7.5 μW m^{-3}, 10 μW m^{-3} and 16 μW m^{-3}) and h_r. Note that providing h_r is less than about 7.5 km ($< z_i/2$) then q_c is linear in H_i and h_r (equation (3) reduces to $q_c \approx H_i h_r \pi^{0.5}$, as shown by the dashed lines).

elements. Such studies have led to the notion that heat production ($H(z)$) varies approximately with an inverse exponential dependence on depth (z), with a form:

$$H(z) = H_i \exp\left(\frac{-z}{h_r}\right) \quad (1)$$

with a characteristic length scale (h_r) of about 10 km. At the time of metamorphism of many Australian Proterozoic HTLP terranes (when current surface exposures were at 15–20 km depth) this relation would imply implausibly high surface heat production rates (>30 μW/m^{-3}), and thus is unlikely to be appropriate. Rather, the high heat production at the current surface exposures is likely to have been anomalous not only in terms of the deeper crust but also the shallower crust during metamorphism. The implication of an anomalous heat production at the crustal levels appropriate to these HTLP terranes suggests that the attendant thermal regimes may be approximated by a heat production distribution that is concentrated at a discrete horizon within the crust (we leave the question of the origin of such heat production distributions until the discussion). In order to investigate the thermal consequences of heat production distributions that are concentrated at a discrete level within the crust we propose a variation of equation (1):

$$H(z) = H_i \exp\left(\frac{-(z-z_i)^2}{h_r^2}\right) \quad (2)$$

In this distribution $H(z)$ varies exponentially with depth but rather than reaching its maximum value H_i at the surface, the maximum heat production occurs at depth z_i (Fig. 3a). As for equation (1), the parameter h_r provides a measure of the spread of the heat production distribution (Fig. 3a), with the heat production falling to $H_i\,e^{-1}$ at depths $z_i \pm h_r$.

In the steady state, the integrated crustal heat production (q_c) represents the crustal contribution to the surface heat flow:

$$q_c = \int_0^{z_c} H(z)\,\mathrm{d}z$$

For the heat production distribution given by equation (2), the integrated crustal heat production is given by:

$$q_c = \frac{H_i h_r \sqrt{\pi}}{2}\left(\mathrm{Erf}\left(\frac{z_c - z_i}{h_r}\right) + \mathrm{Erf}\left(\frac{z_i}{h_r}\right)\right) \quad (3)$$

and the steady-state temperature distribution in the crust of thickness z_c subject to a basal heat flux q_m with a depth-independent thermal conductivity k is:

$$\begin{aligned} T(z) = &-\frac{q_m z}{k} + \frac{H_i h_r^2}{2k} \\ &\times\left(\exp\left(-\frac{z_i^2}{h_r^2}\right) - \exp\left(-\frac{(z_i-z)^2}{h_r^2}\right)\right) \\ &+\frac{h_r H_i \sqrt{\pi}}{2k}\left(z\,\mathrm{Erf}\left(\frac{z_c - z_i}{h_r}\right)\right. \\ &+(z_i - z)\,\mathrm{Erf}\left(\frac{z-z_i}{h_r}\right) \\ &\left.+z_i\,\mathrm{Erf}\left(\frac{z_i}{h_r}\right)\right) \end{aligned} \quad (4)$$

In the following sections we use equation (4) to evaluate the steady-state limit to HTLP metamorphism. One useful attribute of the distribution specified by equation (2) is that it allows evaluation of the effect of variably concentrating a given total heat production (as highlighted by Fig. 3a). Providing $z_i > 2h_r$ then the integrated crustal heat production is linear in H_i and h_r (see Fig. 3b) and consequently the effect of lumping up (or spreading out) the heat production for a given total crustal heat production is obtained by varying h_r as $1/H_i$.

Parameter ranges and solutions to equation (4)

We are interested in seeking solutions to equation (4) that allow HTLP metamorphism without violating (our) imposed Moho melting condition (see below). Such solutions must comply with known constraints on the independent parameters and boundary conditions that govern the behaviour of equation (4), notably thermal conductivity (k) and mantle heat flow (q_m). We also require a definition of the *Moho melting criterion* and realistic bounds on the magnitude of q_c and z_c during metamorphism. In the following discussion we consider explicitly thermal conductivity in the range 1.5–3.5 W m^{-1} K^{-1} and mantle (or reduced) heat flows in the range 10–40 mW/m^2.

The thermal regimes required for large-scale melting of the lower crust are likely to depend on the lithological makeup of the deep crust which in turn will depend greatly on its prior tectonothermal history. Fertile crust may undergo significant melt extraction at temperatures as low as 700–750°C. However, in the HTLP terranes we are concerned with here, metamorphism typically post-dates (by up to

150 Ma) very significant granite melt extraction events, that are likely to have produced a refractory lower crust (as well as enriching the mid-crustal levels in heat production; see Discussion). Such a lower crust is probably able to withstand very high temperatures (up to 1000°C) without significant melting. An example of a lower crustal fragment that withstood temperatures of at least 1000°C without significant melting is provided by the Napier Complex in Enderby Land, Antarctica (Harley 1991; Sandiford 1985). In the following discussion we set, somewhat arbitrarily, an upper limit of 1000°C on the maximum sustainable temperatures $T(Moho)$ in the deep crust without significant melting.

A lower limit on total crustal heat production q_c relevant to the Australian Proterozoic terranes is provided by the measured heat flows (averaging $85\,\mathrm{mW\,m^{-2}}$) which, together with evidence for a 250 km thick lithosphere (Zielhuis & van der Hilst 1996), suggests that presently the integrated crustal heat production contributes, on average, at least $70\,\mathrm{mW\,m^{-2}}$ to the surface heat flow. At the time of metamorphism, total crustal heat productions are likely to have been considerably higher, due to the contribution from the upper 10–20 km that has subsequently been removed from the metamorphic pile, and the secular decline in radiogenic heat production (resulting in a decline in heat production of about 20% over the last 1500–1600 Ma for the analyses shown in Fig. 1). We consider the likely range for synmetamorphic q_c to be $75\text{–}100\,\mathrm{mW\,m^{-2}}$. Note that this is a factor of 2–3 higher than is typically used in modelling metamorphic conditions in the crust (e.g. England & Thomson 1984).

As discussed by Sandiford & Powell (1991), total crustal thicknesses (z_c) during metamorphism of HTLP terranes are unlikely to exceed about 45–50 km, because greater crustal thicknesses are likely to have induced deeper levels of denudation. Note that estimating z_c simply by summing the total denudation and current crustal thicknesses is compounded by significant post-metamorphic deformation in many of the Australian Proterozoic terranes relevant to our discussion.

Main results

A summary of the principal results is provided in Figs 4–9. Figures 4a, 5a, 6a and 7a show geotherms for a range of k, q_m, z_i and h_r values. Note that, in these figures, the heat production parameters have been set such that total crustal heat production $q_c = 75\,\mathrm{mW\,m^{-2}}$. In Figs 4–6 this has been achieved by using a fixed parameter set ($H_i = 7.5\,\mu\mathrm{W\,m^{-3}}$ and $h_r = 5.6$ km) while Fig. 7 shows the effect of varying H_i as $1/h_r$ for a constant $q_c = 75\,\mathrm{mW\,m^{-2}}$. Figures 4b, 5b, 6b

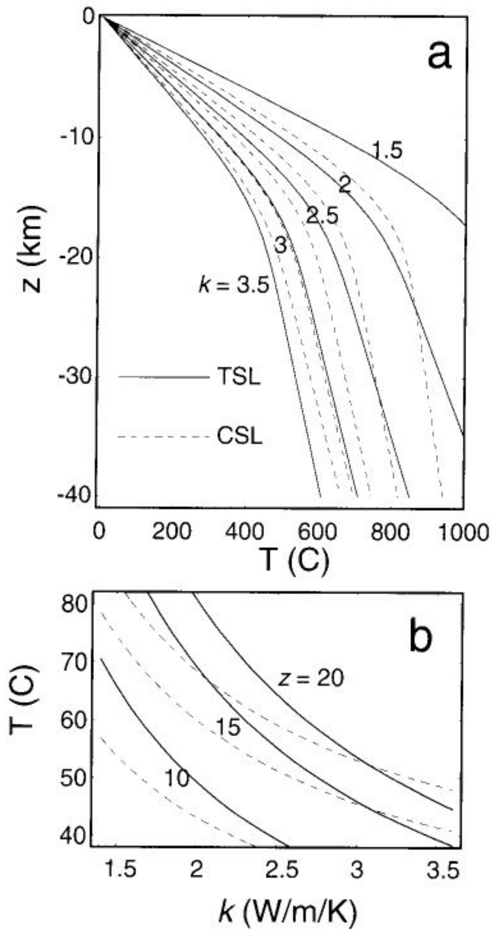

Fig. 4. (**a**) Geotherms for a range of thermal conductivities with total crustal heat production $q_c = 75\,\mathrm{mW\,m^{-2}}$. In this figure (and Figs 5–7) solid lines represent solutions to equation (4) and are appropriate to a thermally stabilized lithosphere (TSL) while dashed lines are appropriate to a chemically stabilized lithosphere (CSL). (**b**) Variation of temperatures T_z at depths of 10, 15 and 20 km with thermal conductivity. Note that variations of thermal conductivity within the experimentally plausible range ($1.5\text{–}3\,\mathrm{W\,m^{-1}\,K^{-1}}$) have a dramatic effect on the geotherm, allowing a range in temperatures at 15 km in excess of 300°C for both the TSL and CSL (note that not all solutions shown are viable since some lead to unrealistically high Moho temperatures). Default parameter ranges for Figs 4–7 are $z_c = 40$ km, $z_i = 15$ km, $h_r = 5.6$ km, $H_i = 7.5\,\mu\mathrm{W\,m^{-3}}$, $q_m = 25\,\mathrm{mW\,m^{-2}}$ and $k = 2.25\,\mathrm{W\,m^{-1}K^{-1}}$.

and 7b show the temperatures attained at depths of 10, 15 and 20 km as a function of the variable parameters. Default parameter ranges for Figs 4–7 are: $z_c = 40$ km, $z_i = 15$ km, $h_r = 5.6$ km, $H_i = 7.5\,\mu\mathrm{W\,m^{-3}}$, $q_m = 25\,\mathrm{mW\,m^{-2}}$ and $k = 2.25\,\mathrm{W\,m^{-1}\,K^{-1}}$.

Figures 4–7 indicate that the thermal structure of the crust is very sensitive to the thermal conductivity (Fig. 4), and to the depth of the heat-producing layer z_i (Fig. 6). In comparison, the variations in the reduced heat flow q_m and the spread of the heat production distribution h_r (for a given total crustal heat production) provide only second-order controls. The important results illustrated in these figures are that lowering the conductivity and increasing the depth of the heat-producing layer both effect a dramatic rise in mid-crustal temperatures. More details relating to the individual diagrams can be found in the figure captions.

A problem with the representation in Figs 4–7 is that it is not obvious how mid-crustal and

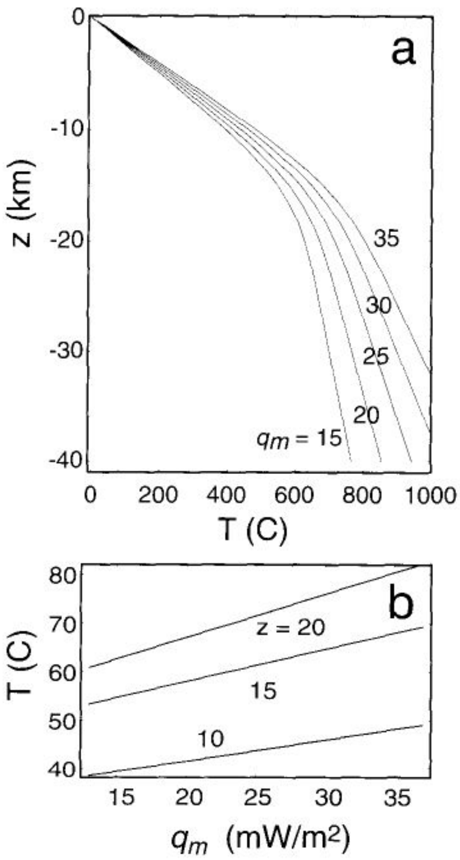

Fig. 5. As for Fig. 4 but with variable reduced heat flow q_m. Varying q_m within the defined range (20–30 $\mathrm{mW\,m^{-2}}$) results in a 70°C range in temperature at 15 km depth (**b**) and thus in comparison with the effect of conductivity (Fig. 4b) and depth of the radiogenic layer (Fig. 7b), provides only a second-order control.

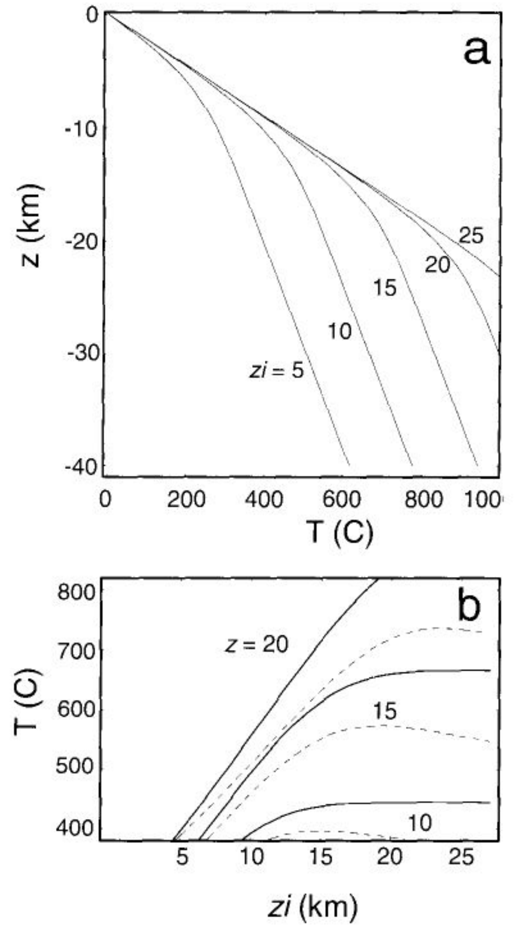

Fig. 6. As for Fig. 4 but with variable depth (z_i) of the radiogenic layer. The figure shows that the parameter z_i has a first-order effect on the temperatures at mid-crustal levels (of the same order as variations in conductivity). The effect is somewhat diminished in the CSL model (dashed lines in Fig. 4b), but remains profound. Note that once the heat-producing layer is significantly below the depth of the observation, then temperatures cease to change with further increases in z_i (and for the CSL model even decline a little with further burial of the heat-producing layer). This figure has profound implications when we consider that changes in z_i will reflect the processes of burial and excavation essential to orogenesis. For example, the burial of radiogenic granitic basement beneath a sedimentary succession provides an excellent model for the heat production distribution specified by equation (2).

Moho temperatures correlate for a given thermal parameter range. Furthermore, these diagrams do not show the effects of varying the total crustal heat production q_c with the other independent parameters. Figure 8a is an attempt to show these effects by plotting solutions to equations (3) and (4) in a parameter space defined by the Moho temperature $T(Moho)$ that results from a particular thermal configuration and either k (Figs 8a and 8b) or z_i (Figs 8c and 8d). Note that we have specifically chosen k and z_i as the independently variable parameters because, as shown in Figs 4–7, they provide the first-order controls on the thermal structure of the middle crust. In Fig. 8, shaded areas indicate regions where parameter ranges produce the approximate conditions of HTLP metamorphism without violating our *Moho melting criterion* or requiring excessive total crustal heat production.

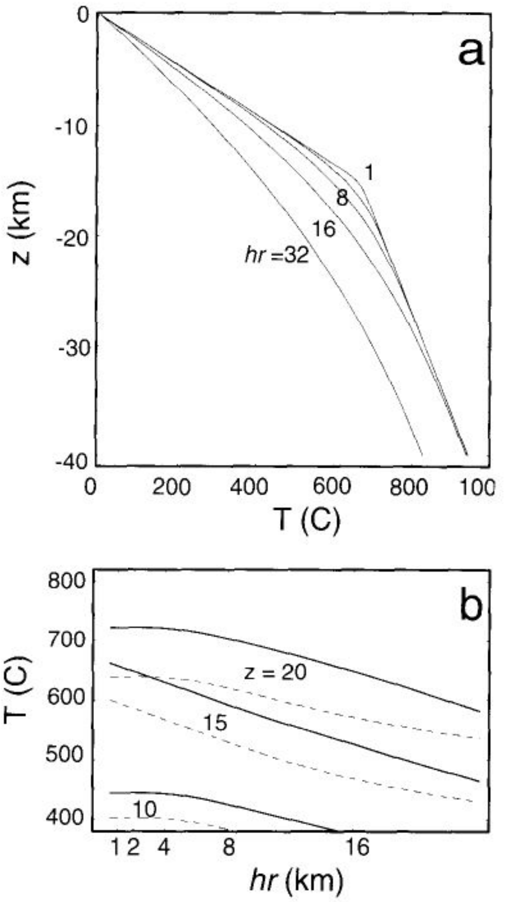

Fig. 7. As for Fig. 4 but with variable heat production parameters h_r and H_i. Note that the total heat production H_i is varied inversely with h_r such that the product $H_i h_r \pi^{0.5}$ remains fixed at $75\,\mathrm{mW\,m^{-2}}$. For $h_r < 7.5$ km ($0.5z_i$) the total crustal heat production q_c is given by $H_i h_r \pi^{0.5}$ (Fig. 3b). (**b**) shows that variations in h_r for a given q_c provide only a small (second-order) variation in temperatures $T(z)$ at depths near z_i and very limited variation in temperatures at any significant depths above or below z_i.

The darker shaded areas in Fig. 8 indicate that thermal conductivities in the range 2.0–2.5 $\mathrm{W\,m^{-1}\,K^{-1}}$ combined with an anomalous heat-producing layer buried to depths of around 15–20 km may allow steady-state conditions appropriate to HTLP metamorphism without excessively high Moho temperatures (i.e. $T(Moho) < 1000°C$), or requiring excessive total crustal heat production (i.e. $q_c < 75\,\mathrm{mW\,m^{-2}}$). For higher total heat productions (up to $100\,\mathrm{mW\,m^{-2}}$) the solution space pertinent to low HTLP metamorphism spans the much larger range in thermal conductivities 2.0–3.0 $\mathrm{W\,m^{-1}\,K^{-1}}$. At $q_c = 75\,\mathrm{mW\,m^{-2}}$ and $T(Moho) = 1000°C$, the maximum attainable temperatures at 15 km $T(15\,km) = 620°C$, and $T(20\,km) = 715°C$; for $q_c = 100\,\mathrm{mW\,m^{-2}}$ the corresponding temperature maxima are $T(15\,km) = 660°C$, and $T(20\,km) = 755°C$.

The role of conductivity highlighted by Fig. 8, which suggests that intermediate to high thermal conductivities favour the prospect of HTLP metamorphism, may seem at odds with Fig. 4, in which low thermal conductivities clearly result in steeper geotherms. This apparent contradiction reflects the role of conductivity in mediating the geothermal gradient not only in the middle to upper crust but also in the deep crust. Since decreasing k results in steeper lower crustal geotherms (for a given mantle heat flux) it will reduce the maximum sustainable mid-crustal temperature for a given Moho temperature. Conversely, a higher k will, by reducing lower crustal thermal gradients, allow higher mid-crustal temperatures for a given Moho temperature. For very high k, steep geotherms in the upper crust can only be maintained by unrealistically high heat production levels. Thus the apparent contradiction between Figs 4 and 8 simply reflects the important balance resulting from the requirement to minimize lower crustal temperatures whilst maximizing mid-crustal temperatures.

Alternative boundary conditions

In deriving equation (4) we have assumed a thermally stabilized lithosphere (TSL) effecting a constant heat flow through the mantle part of

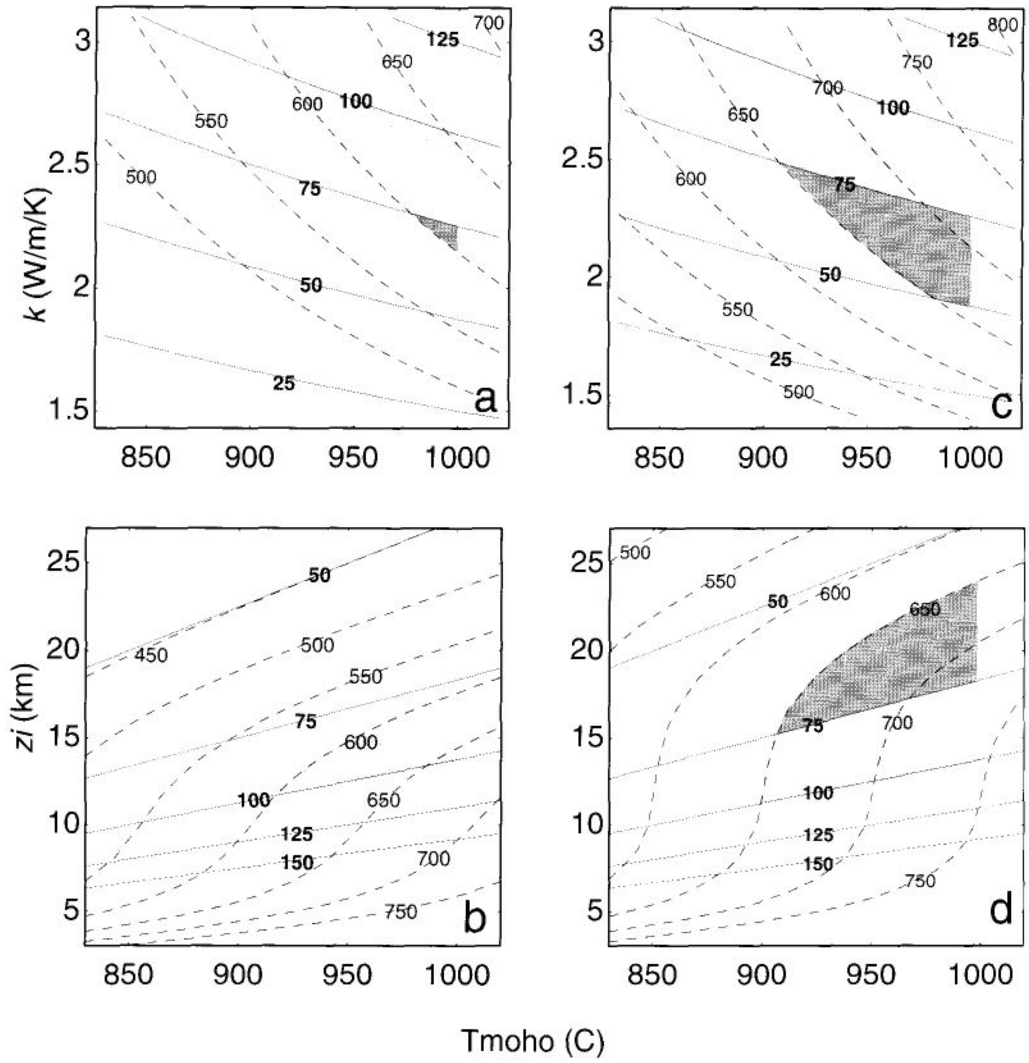

Fig. 8. Solutions to equations (2) and (4) plotted in a parameter space defined by the Moho temperature $T(Moho)$ for $z_c = 45$ km and either k (**a** and **c**) or z_i (**b** and **d**). Solid contours indicate the total crustal heat production needed to achieve the limiting Moho temperature, while dashed contours indicate the resulting temperatures $T(z)$ at depths of 15 km (**a** and **b**) and 20 km (**c** and **d**). Shaded areas indicate regions where parameter ranges produce conditions acceptable for HTLP metamorphism without violating our Moho melting criterion or requiring excessive total crustal heat production (heavy shade $q_c < 75\,\mathrm{mW\,m^{-2}}$, light shade $75\,\mathrm{mW\,m^{-2}} < q_c < 100\,\mathrm{mW\,m^{-2}}$ and $T(z = 15\,\mathrm{km}) > 600°\mathrm{C}$ or $T(z = 20\,\mathrm{km}) > 650°\mathrm{C}$). Note that we have used a fixed value for h_r (=5.6 km) and thus contours for total crustal heat production can be equated with contours for H_i ($q_c = 50\,\mathrm{mW\,m^{-2}}$ equates to $H_i \approx 5\,\mu\mathrm{W\,m^{-3}}$, $q_c = 100\,\mathrm{mW\,m^{-2}}$ equates to $H_i \approx 10\,\mu\mathrm{W\,m^{-3}}$). In **a** and **c**, $z_i = 15$ km, while in **b** and **d**, $k = 2.25\,\mathrm{W\,m^{-1}\,K^{-1}}$.

the lithosphere. An alternative model that may be equally applicable is that of a chemically stabilized lithosphere (CSL) in which the basal boundary condition is specified as a constant temperature condition at the base of the lithosphere, z_l. The important difference is that in the CSL model the thermal properties of the crust will modulate the effective heat flow from the mantle, whereas for the TSL model the heat flow through the mantle lithosphere is independent of the overlying crust. There is a good deal of uncertainty in just how thick the continental

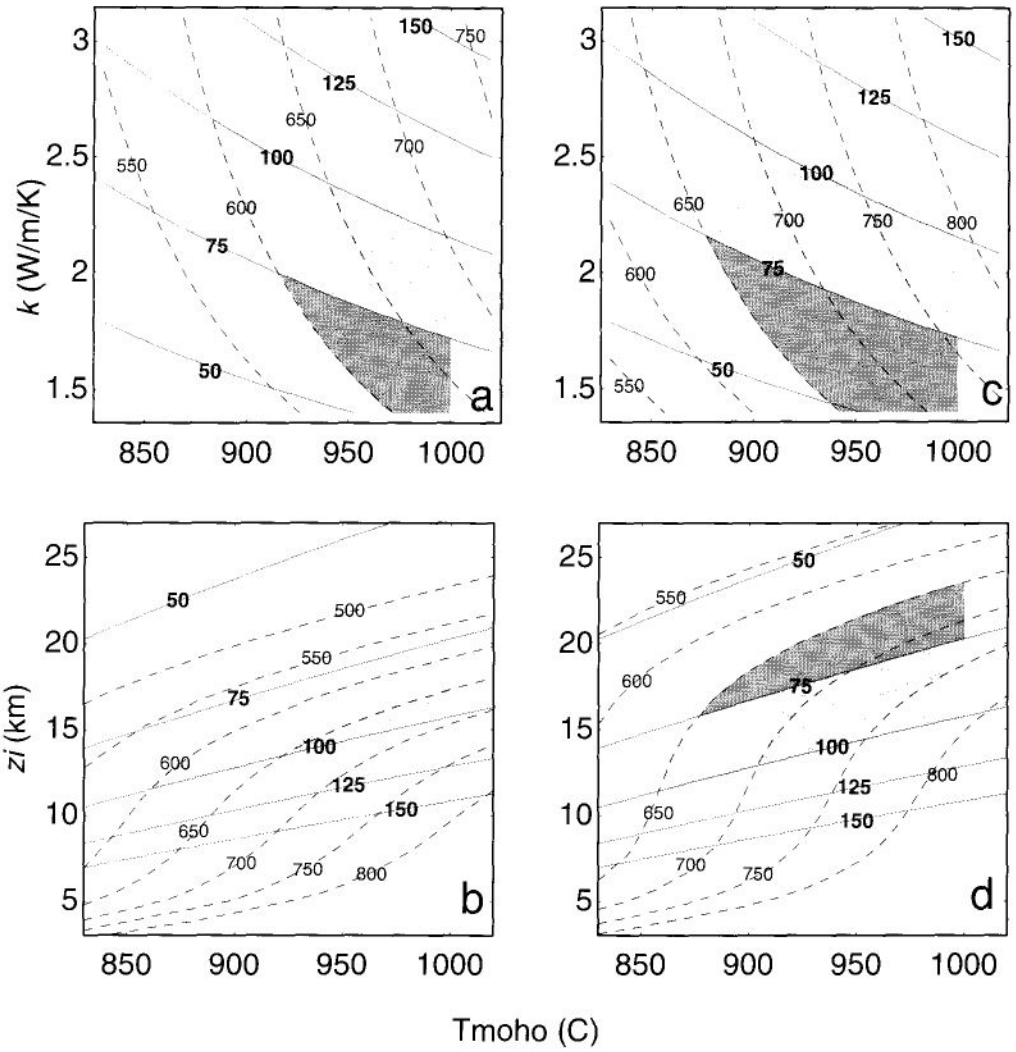

Fig. 9. As for Fig. 8 but with $T(z)$ contours (dashed) based on the CSL model with total lithospheric thickness z_l set at 100 km. The assumed temperature at the base of the lithosphere $T(z_l)$ is 1300°C.

lithosphere may be, and thus specifying the boundary condition for the CSL model is subject to the same kind of uncertainty as specifying q_m for the TSL model.

Figures 4a, 4b, 6b, 7b and 9 show the main results for the CSL model. Note that, for high Moho temperatures, a lower boundary condition of fixed temperature at fixed depth appropriate to the CSL effectively limits the heat flow into the lower crust and thus is equivalent to reducing q_m in the TSL model. For the CSL, this effective heat flow parameter is inversely coupled to the thermal state of the overlying lithosphere (and thus reduces with decreasing conductivity k, as shown by the dashed lines in Fig. 4a, or increasing the depth of the radiogenic layer z_i). Consequently, the thermal effects of changing k and z_i are somewhat diminished for the CSL, compared to the TSL. An important implication of this is that the effective range of parameters allowing HTLP metamorphism is significantly larger for the CSL than the TSL, as shown by the comparative sizes of the shaded regions in Figs 8 and 9.

Discussion

As illustrated in Figs 4–9, solutions to equation (4) clearly indicate that burial of an anomalously radiogenic layer *c.* 5 km thick, to depths appropriate to HTLP metamorphism, allows the possibility of generating HTLP metamorphism in conductive thermal regimes for a plausible range in thermal parameters. Importantly, at Moho depths of 40–50 km the temperatures may remain lower than 1000°C and thus could preclude widespread melting of a refractory lower crust during the HTLP metamorphism in the middle crust. Of course, for more fertile lower crust, such thermal regimes engender the possibility of extensive melting, with any consequent advection of the melt serving to boost the already elevated geotherms in the middle crust (e.g. Chamberlain & Sonder 1990). We believe that the calculations summarized in the previous section cast significant doubt on the view that HTLP metamorphism cannot result from conductive heat transfer because it would result in widespread melting of the lower crust. We note that, while our analysis clearly does not demonstrate a pre-eminent role for conduction in any particular HTLP terrain, it does demonstrate a fallacy in this conventional thermal argument. While we are greatly encouraged (and indeed were initially motivated) by the existing datasets that point to very high heat production levels (5–10 $\mu W\,m^{-3}$) in a number of Australian Proterozoic terranes (Fig. 1), we must stress that the quality of the dataset pertinent to the thermal energy budget of these terranes is still very poor. Any thorough analysis of the thermal energy budget of specific HTLP terranes will require much more detailed evaluation of heat production parameters, and we urge workers interested in assessing the thermal energy budgets of metamorphic terranes to undertake such evaluations.

A principal finding of our analysis concerns the influence of the depth of an anomalous heat-producing layer z_i on the thermal structure of the crust. As shown in Fig. 6, a change in depth of the radiogenic layer by *c.* 5 km may induce a change in $T(z = 15\,\text{km})$ of *c.* 150°C. This finding has a number of potentially important ramifications. For example, it implies that only minor tectonic burial may be required to induce such HTLP metamorphism, while only minor erosion (*c.* 5 km) is necessary to terminate the event. Of great relevance here is the fact that the resulting Moho temperature is also strongly coupled to the depth of the heat-producing layer particularly for the thermally stabilized lithosphere (Figs 8 and 9). In view of the very strong dependence of lithosphere strength on Moho temperature (e.g. England 1987; Zhou & Sandiford 1992), changes in the depth of burial of a radiogenic layer may have dramatic implications for the mechanical response of the lithosphere. We note that many of the Australian Mesoproterozoic provinces of relevance to this discussion, show extensive reactivation during the Phanerozoic, where they have been deformed in 'thick-skinned' fashion along with a thick cover succession. The implication is that the burial of these high-heat-production Mesoproterozoic provinces beneath Neoproterozoic and Phanerozoic sedimentary basins effected a local thermal weakening of the lithosphere (Sandiford *et al.* 1995*a*).

A second implication of the sensitivity of conductive temperatures to the depth of an anomalous heat-producing layer is the relationship between the rates of cooling and the shapes of retrograde *P–T* paths. The notion that HTLP events result from advective thermal processes (Lux *et al.* 1978; DeYoreo *et al.* 1991; Collins & Vernon 1991; Sandiford *et al.* 1991) provides a framework for understanding *P–T* paths that show isobaric cooling (England & Thomson 1984; Sandiford *et al.* 1991), which should consequently reflect very fast cooling rates. In contrast, genuine isobaric cooling paths would not be possible in the case where the primary control on the thermal regime is an anomalously enriched heat-producing layer (at least, for cooling rates that exceed the time constants for secular decay in heat production). However, Figs 6 and 8 show that relatively small reductions in z_i could induce significant temperature reductions in the middle crust. In many systems, the limiting retrograde *P–T* trajectories implied by Fig. 8 of about 100°C kbar^{-1}, may be petrologically indistinguishable from truly isobaric paths.

As we have noted in the Introduction, Australian Proterozoic HTLP terranes typically form in crustal provinces that have experienced very significant differentiation events associated with voluminous granite magmatism up to 150 Ma prior to metamorphism. This early differentiation presumably has had the twofold effect of developing a very refractory lower crust depleted in heat-producing elements and a complementary middle to upper crust enriched in heat-producing elements, thus providing an ideal setting for the style of HTLP metamorphism advocated here. Of course, in order to generate the heat production distributions appropriate to HTLP metamorphism, it is not enough to differentiate the crust. Equally important is the requirement for the total crustal heat production to be anomalous (c. 75–100 mW m^{-2}), and we suggest that the absolute heat production levels in the Australian Proterozoic provinces are likely to be intimately related to the style of granite magmatism that is characterized by very high temperature, strongly fractionated granites, with a strong mantle signature. Of course, while such high-heat-production granite provinces abound in the Proterozoic record, they are clearly not unique to the era. The Cornubian batholith, which is associated with total crustal heat productions of around 90–100 mW m^{-2}, provides a younger example. The analysis presented here suggests that anywhere such enriched granites (or other rocks types) occur there is the potential for developing HTLP metamorphism as a con-= ductive response to the burial of the granitic layer.

We thank N. Wall who collated the data shown in Fig. 1a as part of her BSc Honours research programme, and A. Bingemer whose research around the Mount Painter Inlier has helped clarify many of our views on how geology works in this astonishing region. I. Cartwright, J. Clemens and S. Harley are thanked for their reviews of an earlier version of this manuscript.

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