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Some remarks on the stability of blueschists and related high P–low T assemblages in continental orogens

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ABSTRACT

The thermal evolution of the crust in continental orogens is sensitive to the deformation in the mantle part of the lithosphere, a point obscured in models that apply a boundary condition of constant basal heat flux within the mantle lithosphere. We show that for part of the plausible range of conductivities and heat source distributions within the continental lithosphere the *maximum* or *potential* temperatures in thickened continental crust typical of collision zones do not necessarily exceed the stability field of the high P–low T facies such as blueschists provided the mantle lithosphere is also thickened during the deformation. The appropriate range of parameters includes conductivities towards the upper limit of the plausible range ($k > \sim 2.2$ W m^{-1} K^{-1}) together with low crustal heat production rates with distributions concentrating heat production in the upper crust, and relatively thin initial lithosphere. These observations suggest that at least some of the high P–low T assemblages found within continental orogens, for example in the Dalradian of the Scottish SW Highlands, may reflect the thermal response to whole lithospheric thickening and do not necessarily reflect thermal regimes attendent with subduction or unusually rapid excavation.

1. Introduction

Numerous quantitative thermal models of collisional orogens have been developed in recent years in order to understand the thermal energy budget of regional metamorphic terrains [1–4]. Many such models predict that crust thickened by the collisional process will undergo Barrovian, or higher T, metamorphism. In particular, an important class of models which use a constant heat flow boundary condition at the base of the modelled media show that facies series of lower T and/or higher P than Barrovian may form early in the collisional process but almost inevitably will be overprinted by Barrovian assemblages during the ensuing thermal relaxation of the thickened crust. One consequence of these models is that high P–low T metamorphic facies series, including blueschists, are often (but not always [5,6]) considered diagnostic of subduction-related metamorphism or are inferred to have been preserved due to unusually rapid excavation, and as such are frequently awarded considerable significance in palaeotectonic reconstructions. A subduction related environment is clearly apposite for metamorphism in the many high P–low T terrains formed from the fragmented remnants of accretionary wedges and ophiolite complexes. However, there remains a significant number of regional scale high P–low T terrains which involve rocks of clear continental affinity and where, independently of the thermal arguments, the evidence for subduction related activity during metamorphism is poor or absent; for example the Seward Peninsula Terrain, Alaska [6] and the SW Dalradian (see discussion below).

It is important to realise that the constant heat flow lower boundary condition used in many numerical models of continental metamorphism imposes a unique and possibly somewhat unusual mantle lithosphere deformation (Fig. 1) and thus obscures the important role that mantle lithospheric deformation may play in the thermal evolution of the continental orogens [7]. Modelling strategies which employ different boundary conditions, such as a constant temperature (but not constant depth) lower boundary condition [8–10] show that there is a range of deformation geometries and thermal parameters appropriate to colli-

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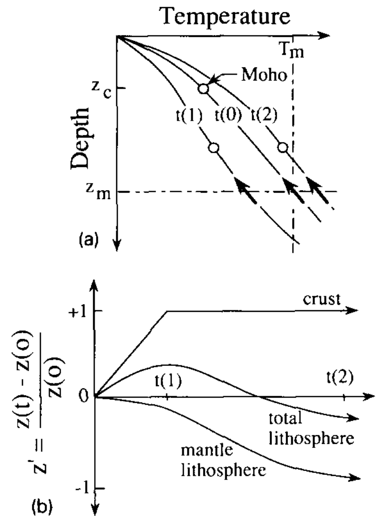

Fig. 1. Schematic illustration of the thermal evolution of a collisional orogen showing how models which impose a constant heat flow lower boundary condition imply a unique vertical strain history for the mantle part of the continental lithosphere. The condition of constant heat flux is imposed at depth z_m initially at temperature T_m appropriate to the base of the lithosphere. Initial crust thickness is z_c. (a) The geotherm prior to thickening at time $t(0)$, immediately after rapid crustal thickening at $t(1)$ and after considerable thermal relaxation of the thickened crust at $t(2)$. (b) Shows schematically how total lithosphere (crust and mantle), crust and mantle lithospheric thickness evolve with time in order to maintain constant heat flow at the depth z_m.

sional orogens for which temperatures within thickened continental crust will not exceed the stability field of high P–low T facies series such as blueschists. While these studies did not explicity evaluate the range of thermal properties and deformation geometries they clearly highlight the profound influence that the mantle lithosphere deformation history plays in the thermal evolution of the overlying crust in continental orogens. Moreover, these studies suggest that the unique tectonic significance often awarded to high P–low T assemblages, where not supported by arguments other than thermal, may need to be re-evaluated. In this paper we investigate the limits to the range of deformation geometries and thermal parameters that allow the preservation of facies series of higher P and/or lower T than Barrovian in crust thickened in continental orogens using potential temperature arguments developed by Sandiford and Powell [9–10]. We use the term "high P–low T facies series" in preference to "blueschist facies series" because of the important control that bulk composition as well as pressure and temperature exerts on the appearance of the blue amphiboles [e.g. 11]; for example the assemblage chlorite-garnet-phengite-quartz is more stable than blue amphibole assemblages at low $a(O_2)$ over much of the "blueschist" P–T range [11]. Thus we consider a stability field for high P–low T facies series that coincides with the stability of blueschist facies assemblages in rocks of appropriate bulk composition as summarised by Thompson and England [12] and Guiraud et al. [11] with a maximum temperature for the formation of blueschist and related assemblages at 10 kbar (i.e. 35 km) of ~ 500°C. Finally, we discuss the significance of the results for the occurrence and preservation of high P–low T assemblages in continental orogens, with particular reference to the Dalradian of the Scottish SW Highlands.

2. Thermal structure of the continental lithosphere

The thermal evolution of a continental orogen depends on: (1) the distribution of heat sources and conductivity; (2) the relative thicknesses of crust and mantle lithosphere which may be related to an initial configuration by the *lithospheric-scale deformation geometry*; (3) the time scale allowed for conductive equilibration; and (4) the role of advective heat transfer mechanisms. For general boundary conditions on finite timescales only numerical solutions can be obtained, and the solutions necessarily reflect the prejudices involved in choosing the sufficient set of boundary conditions. However, here we are only concerned with the potential for preserving high P–low T asssemblages in thickened continental crust and therefore our problem reduces to establishing the range of thermal configurations (i.e., heat source and conductivity distributions) and lithospheric scale deformation geometries which, upon thermal equilibration, will preserve high P–low T facies series within thickened crust (since advective heat transfer by melts is unimportant in high P–low T terrains we ignore this possibility).

The appropriate choice of thermal parameters to model the thermal evolution of continental orogens is not well constrained because of our poor understanding of the distribution of conductivity and heat sources within, and the gross thickness of the continental lithosphere. In the steady state, where crustal thickness tends to ~ 35 km, the surface heat flow, q_s, is typically in the range 55–75 mW m^{-2} and can be used to constrain "acceptable" parameter ranges for heat source distribution, conductivity and lithospheric thickness. The following parameter ranges are considered:

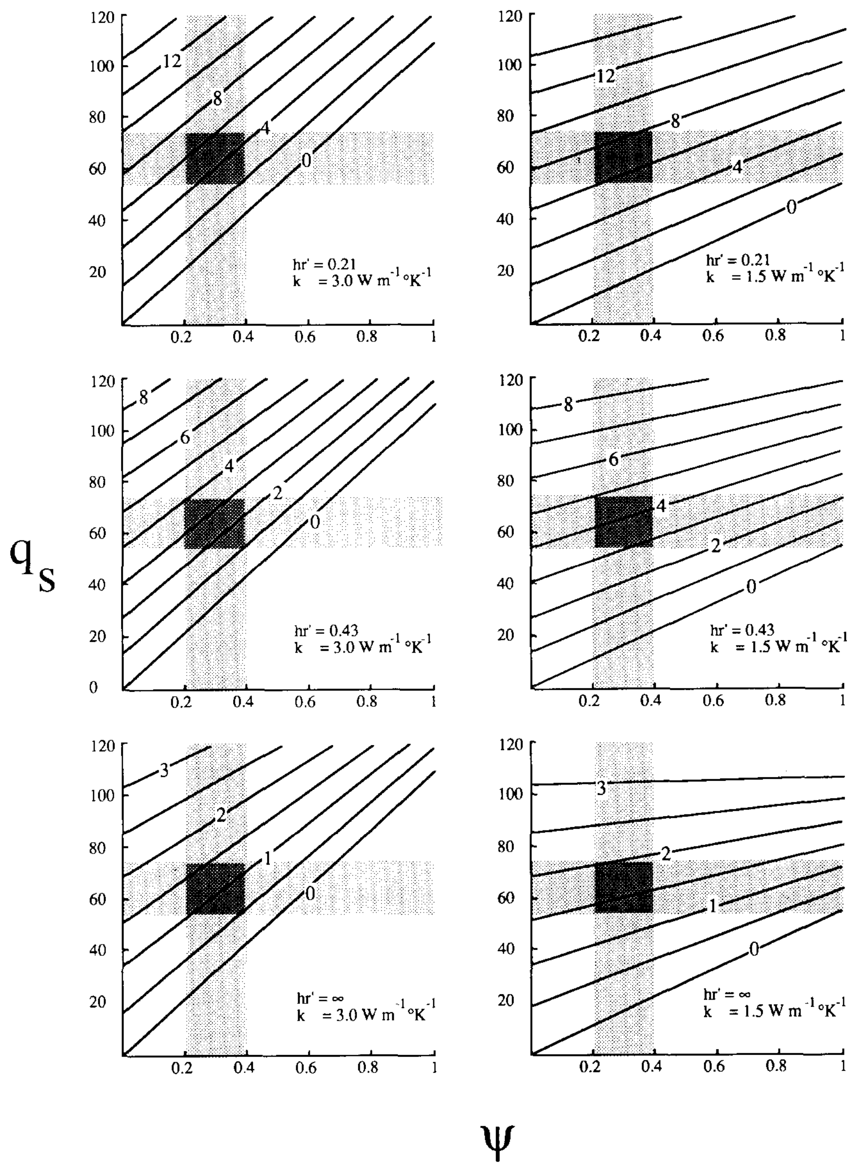

Fig. 2. The heat flow through the lithosphere depends on the thickness and temperature at the base of the lithosphere, and the distribution of conductivity and heat production in the lithosphere. This figure shows the surface heat flow, q_s (mW m^{-2}), as a function of a range of heat production distributions (measured as the surface heat production, H_s, and the length scale for heat source reduction within the crust, h_r'), conductivity, k (which is assumed constant through the lithosphere) and lithospheric thickness (measured as ψ, the ratio of crust to lithospheric thickness). The range of acceptable solutions is shown in the intersection shaded regions which correspond to typical steady state continental heat flows (55–75 mW m^{-2}) and the plausible range of lithospheric thickness (ψ = 0.2–0.4). All diagrams constructed for a crustal thickness of 35 km. Contours are for surface heat production in μW m^{-3}.

(1) Heat source distribution. While it is conceivable that heat production may be greater at some depth within the continental crust than at the surface, it is generally recognised that deep crustal metamorphic rocks have low heat production rates [13]. Thus, we consider only heat source distributions which are either constant or decrease with depth. In time honoured tradition we characterise the heat source distribution as an exponential function of depth [14] with a length scale of heat source reduction, h_r (we also use h_r' as a dimensionless form of the length scale: $h_r' = h_r/z_c$, where z_c is total crustal thickness). Thus the heat production, $H_{(z)}$, at any depth, z, within the crust is related to the surface heat production, H_s, by:

$$H_{(z)} = H_s\, e^{(-z/h_r)}$$

We consider explicitly three heat source distributions: a homogeneous distribution thoughout the crust, i.e., $h_r' = \infty$; a distribution in which heat production is moderately concentrated in the upper crust $h_r' = 0.43$ ($h_r = 15$ km for $z_c = 35$ km); and a distribution in which heat production is strongly concentrated in the upper crust $h_r' = 0.21$ ($h_r = 7.5$ km for $z_c = 35$ km). We consider the mantle lithosphere has negligble heat production.

(2) Conductivity. As discussed by Engand and Thompson [4] the distribution and temperature dependence of conductivity in the continental lithosphere are relatively poorly known, providing a major limitation to thermal models of continental orogens. In accord with common practice [e.g., 4] we consider the likely range of conductivities for the continental lithosphere is 1.5–3 W m^{-1} K^{-1}. However, many of our results, for example those shown in Figs. 3 and 4, can be easily extrapolated to conductivities outside this range. We assume conductivities are temperature and depth independent.

(3) Lithospheric thickness. The thickness of the continental lithosphere is poorly constrained with estimates ranging from as little as 90 km up to as much as 400 km beneath some Archaean cratons (see review by McKenzie [15]). Since most orogenic belts in the continental lithosphere involve collision of continental margins previously attenuated during rifting events, the range of lithospheric thicknesses appropriate to the discussion here is likely to be at the low end of this spectrum. Moreover, there is some uncertainty concerning the significance of the observations relating to very thick continental lithosphere [15]. Therefore, we consider explicitly only lithospheric thicknesses in the range 90–175 km. Following Sandiford [8] and Sandiford and Powell [9] we characterise the lithospheric thickness by the ratio, ψ, of the thicknesses of the crust to the whole lithosphere. In the potential temperature arguments discussed below we consider an initial crustal thickness of 35 km with ψ in the range 0.4–0.2. We consider here a thermal lithosphere in the sense of McKenzie and Bickle [16] extending to the isotherm $T_l = 1280$°C.

Figure 2 shows the range of values of H_s capable of generating normal continental surface heat flows (55–75 mW m^{-2}) for the parameter range discussed above. The dark shaded region in Fig. 2 gives the limits on the "acceptable" combinations of thermal configurations for the continental lithosphere prior to incorporation within a continental orogen. These "acceptable" parameter ranges provide the basis for our evaluation of the stability of high P–low T assemblages in continental orogens presented below.

3. Potential temperatures

A useful way to characterise the potential heating in an orogen which has been deformed from some initial configuration in which the thermal properties are assumed, is with reference to the equilibrium geotherm for the deformed lithosphere (e.g. Sandiford and Powell [9,10]). In the absence of advective heat transfer processes due to migration of melts or the rapid upward advection of rock with respect to the earth's surface through erosion or extension the temperature field produced by the orogenic processes must lie between the initial equilibrium condition and the deformed equilibrium or *potential* condition [9,10]. For any heat source distribution and crustal thickness the potential temperature at any depth is dependent only on the lithospheric thickness and the ratio of surface heat production to conductivity, which we term $\zeta = H_s/k$ (K m^{-2}), see Appendix 1. In the calculations presented below we consider deformations which are homogeneous on the scale of the crust, such that the dimensionless length scale for the depth dependent heat source distribution, h_r', remains invariant through the deformation.

Figure 3 shows the potential temperatures at half crust depths (i.e., $T^{p}_{p=0.5}$, where p is the depth at which potential temperatures are calculated expressed as a proportion of the total crustal thickness) for the three heat source distributions modelled here appropriate to a deformed crust with a thickness of 70 km (i.e, a crust which has been thickened by the factor $f_c = 2.0$). The lithospheric thickness is expressed as the ratio, f_l, of deformed lithospheric thickness to initial lithospheric thickness. The shading in Fig. 3 shows the range of deformation geometries (i.e, range of f_l at $f_c = 2$), and thermal configurations, expressed as ζ, which, in the absence of advective heat transfer mechanisms allow temperatures at mid crustal levels to remain below 500°C, and thus allow the preservation of high P–low T assemblages. Clearly not all the solutions indicated by the shaded regions in Fig. 3 represent plausible thermal configurations for the continental lithosphere, for example, the lower bound of $\zeta = 0$ is unacceptable since it is appropriate to zero crustal heat production. The important question then is do any of the parameter ranges appropriate to the shaded areas in Fig. 3 also fall within the acceptable parameter range defined by the shaded regions in Fig. 2?

Figure 4 show f_l–k space contoured for heat production distributions which give $T^{p}_{p=0.5}$ =

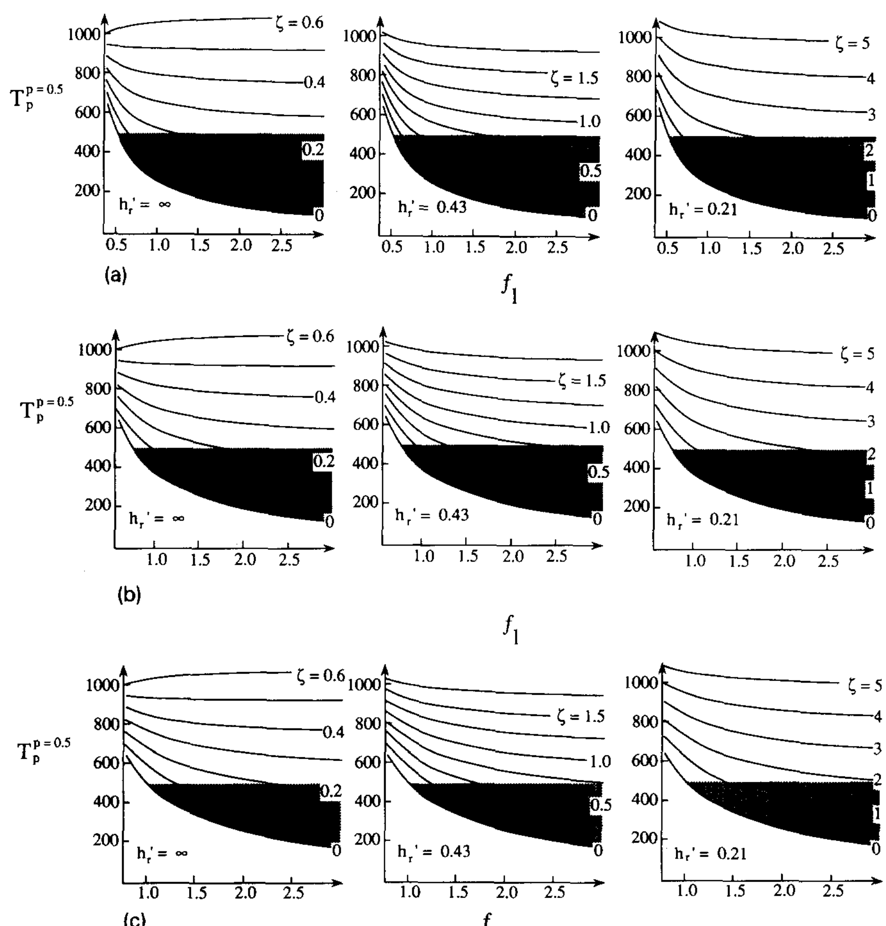

Fig. 3. Potential temperatures at 35 km depth in a 70 km thick crust (i.e., $T^{p}_{p=0.5}$ for $f_c = 2.0$) illustrated as a function of the lithospheric vertical strain, f_l, the heat source depth exponent, h_r', and the initial ratio crust and lithosphere thicknesses, ψ, and contoured for variable ζ (expressed as the ratio of surface heat production to conductivity). The stippled region represents potential temperatures which remain within the high P–low T "blueschist" stability field for depths of ca. 35 km. (a) $\psi = 0.2$; (b) $\psi = 0.3$; and (c) $\psi = 0.4$.

500°C and initial surface heat flows in the range 45–85 mW m^{-2}. Assuming that the acceptable range of surface heat flow prior to deformation is 55–75 mW m^{-2}, then Fig. 4 indicates that there is indeed an acceptable thermal parameter range for which potential temperatures do not exceed the stability field of high P–low T facies series (shown by the shading): this range includes conductivities towards the high end of the plausible spectrum ($k \gtrsim 2.2$ W m^{-1}K^{-1}), large f_l (or thickened mantle lithosphere), heat production distributions strongly concentrated in the upper crust, and high initial values of ψ.

The role of mantle lithosphere deformation (i.e., f_l) in the potential thermal structure of the overlying crust is readily appreciated; increasing the thickness of the mantle lithosphere decreases the thermal gradient at the Moho and hence decreases the heat flowing through the base of the crust. For whole lithospheric thickening this must, to some extent, counter the heating resulting from increased heat production within the thickened crust.

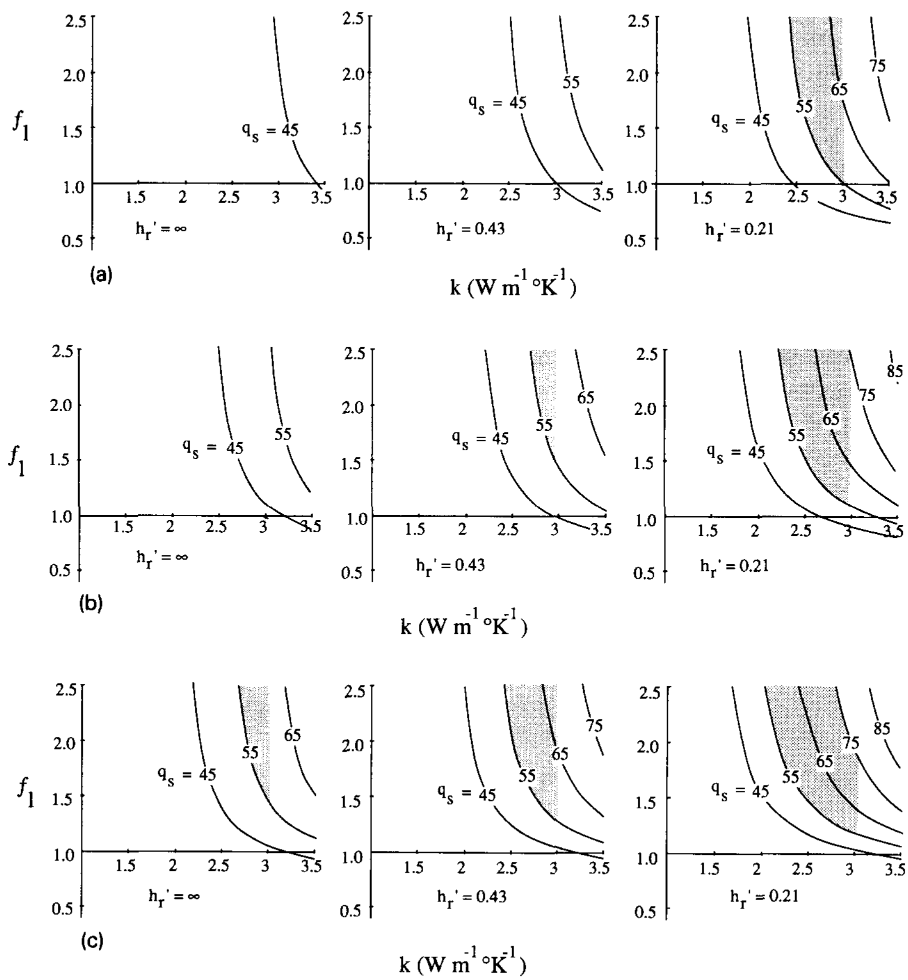

Fig. 4. The influence of conductivity, k, on the potential thermal structure illustrated for surface heat productions giving initial (i.e., prior to deformation) surface heat flows (q_s) of 45–85 mW m^{-2} and which yield $T^P_{p=0.5} = 500$°C as a function of the lithospheric scale vertical strain, f_l, the ratio of thickness of crust and lithosphere, ψ, and the heat source depth exponent, h_r', for double thickness crust, i.e., $fc = 2.0$. (a) $\psi = 0.2$; $\psi = 0.3$; and (c) $\psi = 0.4$. The shaded region shows the range in k–f_l space in which potential temperatures do not exceed the stability of the high P–low T facies series assuming initial surface heat flows fall between 55 and 75 mW m^{-2} and conductivities do not exceed 3 W m^{-1} K^{-1}. This range is favoured by conductivities near the upper acceptable bound ($k > 2.2$ W m^{-1} K^{-1}), together with heat production distributions concentrated in the upper crust, high values of ψ, and lithospheric thickening factors in excess of 1.0–1.5.

The implication of Figs. 3–4 is that mantle lithosphere thickening may provide a very effective "refrigeration" of the overlying crust, provided conductivities are high. The reason large ψ is favoured is more obscure. In the initial, or pre-deformational, configuration the ratio of the heat flow through the mantle lithosphere to the crustal heat production for a prescribed surface heat flow is proportional to ψ. Consequently the effect on the temperature field in the lithosphere of a given deformation of the mantle lithosphere is proportionaly larger for larger ψ.

4. High *P*–low *T* metamorphism in continental orogens: discussion

Having identified the "acceptable" thermal parameter ranges for the continental lithosphere, and the thermal configurations and deformation geometries allowing potential temperatures not exceeding the stability field of high *P*–low *T* facies series at the mid levels of thickened crust, we now discuss in further detail factors most likely to lead to the preservation of high *P*–low *T* terrains in continental orogens, and give an example of a metamorphic terrain in which some of these processes may have occurred. It is important to reiterate that in the potential temperature arguments presented so far we have only considered thermally equilibrated lithospheres. In reality, the lithosphere deforms at rates fast enough to cause significant heat advection, such that geotherms defined by potential temperatures are unlikely to be attained in natural orogens. Most importantly, rapid exhumation of rock buried to deep levels may prevent heating beyond blueschist stability fields even though potential temperatures at any instant exceed the stability field of the high *P*–low *T* facies series. As a result, high *P*–low *T* terrains are often considered diagnostic of rapid exhumation rates (e.g. > 1 km Ma^{-1} [17]) having taken place, with intermediate to slow exhumation rates generally believed to allow sufficient thermal relaxation that Barrovian stability fields will be intersected. However, our potential temperature arguments imply that if the lithospheric thermal configuration allows potential temperatures to remain within high *P*–low *T* stability fields, even slow exhumation rates will favour the survival of regional high *P*–low *T* facies series. Of course potential temperatures during exhumation will only remain low if the thickened mantle lithosphere remains intact, and if exhumation is driven by mantle lithospheric thinning [e.g. 7–9] then heating during exhumation may be dramatic. Finally, the potential temperature arguments developed above imply that it not necessary to invoke specific tectonic settings where heat flow is suppressed for the development of high *P*–low *T* assemblages such as subduction zones. While subduction zones certainly provide an ideal setting for high *P*–low *T* metamorphism, we stress that such specific tectonic assignments on the basis of thermal considerations alone are unwarranted.

As has already been stated above, Fig. 4 show that continental orogens most likely to produce and preserve high *P*–low *T* facies series will be characterised by high conductivities and thickened mantle lithosphere, in combination with low heat production rates or high initial ψ. The values of thermal parameters that allow potential temperatures to remain within high *P*–low *T* facies conditions tend toward the limits of the "acceptable" parameter ranges that give normal heat flows in undeformed continental lithosphere. However, the "allowable" range is large enough to suggest that high *P*–low *T* metamorphism can occur as a simple consequence of the thermal relaxation of thickened continental crust, and does not require some additional factor such as rapid exhumation provided that the mantle lithosphere is also thickened during the deformation.

5. High *P*–low *T* metamorphism in continental orogens: an example from the SW Highlands, Scotland

The Scottish Dalradian is well known for being the type area for medium and low *P* (Barrovian and Buchan) regional metamorphism. Importantly, the Dalradian shows little direct evidence for active subduction during its formation and subsequent deformation of the terrain; most importantly the late Precambrian TayVallich Volcanics [18] demonstrably lie within the Dalradian sedimentary sequence implying deposition in a continental extensional basin. A number of distinct provinces have been recognised in the Dalradian on geochemical, geophysical and stratigraphic grounds, separated by lineaments along which abrupt contrasts in tectonometamorphic histories occur [19]. One of these provinces is the South West Highlands of Scotland in which high-pressure greenschist to epidote-amphibolite facies pelitic and mafic assemblages grew early in the regional deformation history [20]. The assemblages in this terrain are superficially like those of the lower grade parts of the classic Barrovian zones, eg. garnet-chlorite-phengite. However, geothermobarometric studies indicate equilibration conditions close to blueschist facies $P-T$ conditions [21,22], and the relatively detailed available constraints on its $P-T$ history [21,22] summarised below make it appropriate for discussion here. Moreover, crossite-barroisite-bearing possible equivalents to the South West Highlands assemblages have been recognised in Mayo, North West Ireland [23].

The metamorphic zonation in the South West Highlands comprises both biotite and garnet zones based on index minerals in mafic [24] and pelitic [21,22,25] assemblages and the $P-T$ conditions of metamorphism are between 410 and 530°C and 8–10 kbar [21,22]. The highest grade assemblages occur in the deepest exposed structural levels and are often pervasively overprinted by albite porphyroblastic assemblages during an episode of later deformation [22,26]. However, the lower grade assemblages are well-preserved and free from overprinting metamorphism. Geochronological constraints suggest that the earliest phases of deformation in the Dalradian terrain, hence the high-pressure metamorphism in the South West Highlands, occurred before 590 ± 2 Ma ago [27]. Elsewhere in the Scottish Dalradian, mineral cooling ages suggest that cooling from the metamorphic climax associated with Barrow's Zones probably started between 490 and 510 Ma ago [28,29], though age constraints on the time of attainment of the metamorphic peak are currently lacking.

In very marked contrast with the rest of the Scottish Dalradian, significant Barrovian-style metamorphism substantially post-dating the earliest phases of deformation did not take place in large areas of the South West Highlands. Some factor unique to this region has therefore allowed the survival of high P–low T assemblages, not as localised remnants, but as the dominant regional isograd sequence in this part of the terrain, through an orogenic history lasting for over 100 Ma. Graham [30] suggested that the crust in the South West Highlands is characterised by a high abundance of rock with relatively low radiogenic heat production rates. This suggestion was made because, in the data set of thermal parameters of Richardson and Powell [31], lithologies that are dominant in the South West Highlands are strongly biased towards low heat production rates. In particular, up to 30% of the surface exposure in the South West Highlands crust is made up of syn-depositional mafic rock [32,33], while a large proportion of the rest of the succession is made of quartz-rich metasediments. We stress that currently, heat production rates and heat flow measurements are lacking for the South West Highlands, though the contrasts in crustal composition between the South West Highlands and the rest of the Dalradian correlate strikingly with the contrasts in the metamorphic histories between these two parts in this terrain.

In addition to the possible low heat production rates in the Dalradian, geophysical evidence [34,35] suggests that the South West Highlands Dalradian is underlain at depth by large thicknesses of orthogneissic rock forming a wedge tapering to the southeast. In contrast, Graham [30] points out that the Dalradian to the northeast of the South West Highlands is underlain by older metasedimentary rock, the "Older Moine" metasediments which, according to Richardson and Powell [31], have similar heat production rates to those of most Dalradian metasediments in their data set. It seems likely that these heat production rates are substantially larger than those in the orthogneiss underlying the South West Highlands, though, again, heat production rate measurements are lacking from this gneiss.

The exhumation history of the South West Highlands is constrained by Rb-Sr and K-Ar ages on phengites and biotites [22]. These ages range between 500 and 440 Ma, the Rb-Sr phengite ages being between 460 and 500 Ma, the K-Ar phengite ages between 465 and 475 Ma, and the Rb-Sr and K-Ar biotite agesbeing about 440 Ma. Based on closure temperature estimates for these isotope systems [36], these ages suggest cooling and exhumation of the terrain between 500 and 440 Ma, the average cooling rate being ca. 2–4°C Ma^{-1}. This cooling rate does not account for the probable 80–90 Ma time gap between regional metamorphism and the oldest Rb-Sr phengite age. It is likely that the Rb-Sr phengite ages were reset to some extent by later deformation (possibly in part the deformation associated with the albite-bearing overprinting assemblages to the southeast of the area); these data therefore point to a protracted cooling history which, on the available data, is as long as the cooling history of the classic Dalradian Barrovian sequences [28].

The mineral isotope data imply that the preservation of the high P–low T assemblages in the SW Highlands cannot be attributed to an unusually rapid excavation history in comparison with the better known Barrovian sequences, and therefore the observed regional variations in the Dalradian thermal regime must be explained otherwise. It seems likely that regional variation in heat production may account for much of the variation in metamorphic heating, however our calculations suggest that the role of mantle lithospheric deformation is also likely to be important; the implication being that the SW Highlands high P–low T assemblages have been preserved through more extensive, or longer lived, mantle lithospheric thickening.

Acknowledgements

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Appendix 1

The basic equations for potential temperature (T^{p}) used in Figs. 3 and 4, parameterized in terms of f_c, f_l' and $\zeta(=H_s/k)$ derive from the steady state 1-dimensional heat equation:

$$\frac{\mathrm{d}\left[k_{(z)}\dfrac{\mathrm{d}T^{\mathrm{p}}}{\mathrm{d}z}\right]}{\mathrm{d}z} + H_{(z)} = 0 \qquad \text{(A1)}$$

with

$$H_{(z)} = H_s\, \mathrm{e}^{(-z/h_r)} \quad \text{for } 0 < z < z_c$$
$$H_{(z)} = 0 \quad \text{for } z_c < z < z_l$$
$$k_{(z)} = k \quad \text{for } 0 < z < z_l$$

Assuming the following boundary conditions

$$\frac{\mathrm{d}T^{\mathrm{p}}}{\mathrm{d}z} = \frac{T_l - T_c}{z_l - z_c} \quad \text{for } z = z_c$$
$$T^{\mathrm{p}} = T_s \quad \text{for } z = 0$$

yields the following expression for T^{p}:

$$T^{\mathrm{p}} = T_s + \zeta h_r^2 - \zeta h_r\left[h_r\, \mathrm{e}^{(-z/h_r)} + z\, \mathrm{e}^{(-z_c/h_r)}\right] + z\left(\frac{T_l - T_c}{z_l - z_c}\right) \qquad \text{(A2)}$$

where T_c and T_l are the temperatures at the depths of the Moho, z_c, and base of the lithosphere, z_l, respectively. Finally, the appropriate parameterizations are derived by noting that:

$T^{\mathrm{p}} = T_l$ for $z = z_l$

solving (A2) for T_c and substituting:

$$z_c = z_{c0}\, f_c$$
$$z_l = f_l(z_{c0}/\psi)$$
$$z = p z_{c0}\, f_c$$

where z_{c0} is the thickness of the crust prior to deformation, ψ the ratio of inital crust to lithosphere thickness, and p is the depth at which potential temperatures are calculated expressed as a ratio to total crustal thickness ($p = z/z_c$).

In Fig. 4 the crustal heat production has been parameterised in terms of the surface heat flow in the initial or pre-deformation lithosphere, q_{si}:

$$H_s = \frac{\mathrm{e}^{1/h_r'}(T_l k - q_{si}\, z_{l0})}{z_{c0}\, h_r'\left[h_r'\, z_{c0}(\mathrm{e}^{1/h_r'} - 1) - z_{c0} + z_{l0}(1 - \mathrm{e}^{1/h_r'})\right]} \qquad \text{(A3)}$$

where $h_r' = h_r/z_{c0}$ and z_{l0} is the initial thickness of the lithosphere ($z_{l0} = z_{c0}/\psi$).

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