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Controls on low-pressure anatexis

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ABSTRACT Low-pressure anatexis, whereby rocks melt in place after passing through the andalusite stability field, develops under more restricted conditions than does low-pressure metamorphism. Our thermal modelling and review of published work indicate that the following mechanisms, operating alone, may induce anatexis in typical pelitic rocks without inducing wholesale melting in the lower crust: (i) magmatic advection by pervasive flow; (ii) crustal-scale detachment faulting; and (iii) the presence of a high heat-producing layer. Of these, only magmatic advection by pervasive flow and crustal-scale detachment faulting have been shown quantitatively to provide sufficient heat to cause widespread melting. Combinations of the above mechanisms with pluton-scale magmatic advection, shear heating, removal of the lithospheric mantle, or with each other provide additional means of developing suitable high temperatures at shallow crustal levels to generate low-pressure anatexis.

Key words: low-pressure metamorphism; anatexis; thermal modelling.

INTRODUCTION

The thermal structure of the crust plays a major role in numerous geological processes, including deformation and metamorphism, erosion patterns and magnitudes, fluid flow, and magma generation. Accurate description of the thermal structure of the crust is therefore a prerequisite for understanding the rheology and the relationship between surface features and geodynamic processes (e.g. Koons *et al.*, 2002). Stable continental geotherms produce temperatures of 400–500 °C in the range of 15–25 km depth, but many orogenic belts contain rocks that were heated to over 600 °C at these depths. Low-pressure metamorphic belts, which record these high temperatures, are common in orogens throughout the world and therefore must represent transient conditions in Earth's crust that readily develop in active tectonic settings. Because of the widespread presence of these belts, the causes of low-pressure metamorphism have received significant attention (Lux *et al.*, 1986; Wickham & Oxburgh, 1987; Hanson & Barton, 1989; Loosveld & Etheridge, 1990; Sandiford & Powell, 1990, 1991; De Yoreo *et al.*, 1991; Stüwe *et al.*, 1993; Brown, 1998; Escuder Virute, 1999; Bodorkos *et al.*, 2002; Miyazaki, 2004; White, 2005). Many of these studies discuss, if not model, anatexis at lower crustal levels in the context of generating magma that then propagates upwards to provide additional heat to the upper levels of the crust. The causes of anatexis at low pressure, however, have received comparatively little attention. Determining the geodynamic settings for anatexis is particularly relevant, given the recent progress in relating crustal rheology to the presence of melt (Rutter, 1997; Renner *et al.*, 2000; Brown, 2001; Takeda & Obata, 2003; Rosenberg & Handy, 2005).

We do not intend this contribution to be a review of all locations where low-pressure anatexis has occurred nor of the causes of low-pressure metamorphism. Rather, we quantitatively explore the causative agents of conditions suitable for low-pressure anatexis are explored. For the purposes of this paper, low-pressure anatexis is defined as the result of andalusite–sillimanite-type metamorphism (Miyashiro, 1961) where the *P–T* evolution crosses a melt-producing reaction (Fig. 1). Melting commonly occurs not only in the sillimanite field but is also possible in the andalusite field (Cesare *et al.*, 2003). Although we present much of the discussion in the context of pelite melting, the thermal conditions can easily be generalized. The widely cited wet pelite melting curve of Spear *et al.* (1999) is used because it is the lowest temperature melting reaction for common rock types. Recent calculations of the wet melting curve based on thermodynamic data (e.g. White *et al.*, 2001) put the reaction at slightly higher temperatures than do Spear *et al.* (1999), but the exact location of the melting reaction in pressure–temperature space does not affect our results.

Excluded from this analysis is anatexis due to (near-)isothermal decompression. Such decompression allows deeper crustal material to approach the surface, carrying heat with it. If the decompression is fast enough, rocks can enter melting fields at low pressures (Fig. 1). Numerous examples of this type of low-pressure melting exist around the world (Whitney & Teyssier, 2003), including in areas where the decompression is caused by erosion (Zeitler *et al.*, 1993). Melting at low pressure because of isothermal

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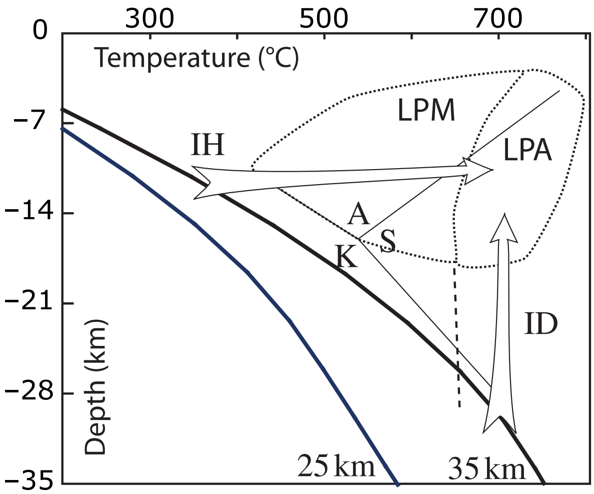

Fig. 1. Equilibrium geotherms and *P–T* paths leading to low-pressure anatexis. Depth–temperature space illustrating typical geotherms for stable continental crust of 25 and 35 km thickness (labelled solid lines) and *P–T* paths that can lead to anatexis at low pressure (arrows: IH, isobaric heating; ID, isothermal decompression). The aluminosilicate phase boundaries (A, andalusite; S, sillimanite; K, kyanite) and the wet pelite melting curve (dashed) after Spear *et al.* (1999) are also shown. LPM, field of low-pressure metamorphism; LPA, field of conditions appropriate for low-pressure anatexis.

decompression is readily explicable by well-known crustal kinematics and has a fundamentally different cause than low-pressure anatexis in rocks that passed through the andalusite field.

NUMERICAL MODEL

We designed a one-dimensional numerical model to evaluate the relationship among deformation, lithospheric mantle removal, sedimentation or erosion, and the thermal structure of the crust. One-dimensional models necessarily simplify the structure of the lithosphere, but they also provide information about fundamental processes that may be obscured by two- or three-dimensional complexities. These models are not appropriate when investigating the effects of magma transport (Hanson & Barton, 1989; De Yoreo *et al*., 1991; Pedersen *et al*., 1998; Ryan & Soper, 2001) or inclined shear zones (Buck *et al*., 1988; Sandiford & Powell, 1991; Escuder Virute, 1999), but they are appropriate for studying homogeneous deformation. The model constructed follows the same principles employed by others (Loosveld & Etheridge, 1990; Sandiford & Hand, 1998; Bodorkos *et al*., 2002), and it is included here to allow direct comparison of different heat-transfer mechanisms.

Geometrically, the model consists of continental crust and lithospheric mantle underlain by asthenosphere (Fig. 2). For ease of applying boundary conditions, our model is constructed in two dimensions, but because of the uniformity of the layers, the resulting model is in effect a one-dimensional solution. The upper boundary of the model represents the Earth's surface and has a fixed temperature of 0 °C. The lower boundary of the model is the lithosphere–asthenosphere boundary, assigned a constant temperature of 1400 °C. Parameters include layer thickness, density, crustal heat production, heat capacity, thermal conductivity and asthenosphere temperature (Tables 1 & S1). We began each model run by calculating an equilibrium geotherm for the assigned geometry using Eqs (1) and (2) (modified from Fowler, 1990). The temperature profile of the crust follows the rule:

$$T = -\frac{A}{2k}z^2 - \frac{Q_d + Ad_c}{k}z, \qquad (1)$$

and the lithospheric mantle temperature is

$$T = -\frac{Q_d}{k}z + \frac{Ad_c^2}{2k}, \qquad (2)$$

where T is the temperature, A the heat production (assumed constant in crust, absent in mantle), k the thermal conductivity, z the depth, Q_d the heat flux from the asthenosphere (varies; see Eq. 5), and d_c the crustal thickness. Depth (z) is positive upwards and $T = 0$ °C at $z = 0$.

Constant horizontal velocities are applied to the edge of the block (Fig. 2). Interior horizontal and vertical velocities are calculated to maintain a constant area within each layer, and deformation within each layer is homogeneous. Upon extension or contraction, the model may accept sediment or erode to maintain isostatic balance with the initial geometry. At each time-step, the temperature is calculated at 32 nodes initially spaced evenly throughout the lithosphere. The initial temperature at each node is taken as the value of the stable geotherm at that point, and at each time-step, the temperature changes according to the solution of

$$\frac{\partial T}{\partial t} = \frac{k}{\rho C_p}\frac{d^2T}{dz^2} - \bar{v}\frac{dT}{dz} + \frac{A}{\rho C_p}, \qquad (3)$$

where ρ is the density, C_p the heat capacity, t the time, and $\bar{v}$ the erosion or sedimentation rate (e.g. Stüwe, 2002). The change in temperature with time can be approximated for any given rock particle by

$$\Delta T = \left(\frac{k}{\rho C_p}\frac{d^2T}{dz^2} - \bar{v}\frac{dT}{dz} + \frac{A}{\rho C_p}\right)\Delta t, \qquad (4)$$

for sufficiently small time steps. Additional experiments performed indicate that this approximation is insensitive to time-step size for the range of time steps employed in this study. The spatial derivatives are approximated using central differences except at the boundaries, where forward or reverse differences are used as appropriate.

Some studies fix the heat flux from the mantle (e.g. England & Thompson, 1984), but here instead a constant temperature is prescribed to the base of the lithosphere during each model run. With that constant

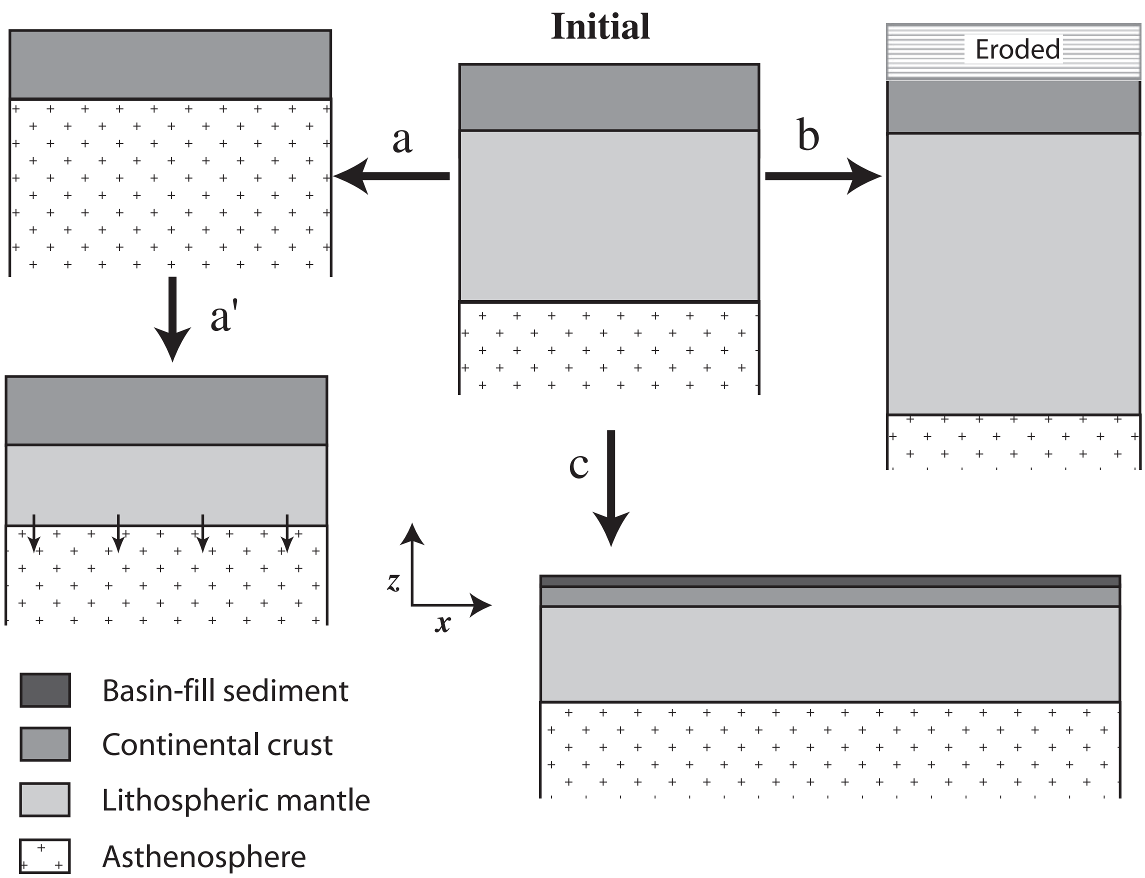

Fig. 2. End-member geometries of numerical model; each starting block is 100 km in the x-direction. (a) Without deforming the crust, the lithospheric mantle is removed and replaced by constant-temperature asthenosphere. Additional parameters include erosion, whereby the crust may erode to maintain isostatic balance with the initial conditions, and when the lithospheric mantle begins to reform by conductive cooling (a′). No mechanism for removal of the lithospheric mantle is implied. (b) The initial two-layer lithosphere contracts horizontally at a prescribed rate and thickens vertically. The extent of erosion is a parameter in the model. (c) The initial lithosphere extends and thins at a prescribed rate. The degree of sedimentation is a model parameter. Variations (b) and (c) may be coupled with removal of the lithospheric mantle.

Table 1. Model parameters.

Parameters and variables	Symbol and units	Units	Value or range
Temperature	T	°C	n.a.
Depth (positive upwards)	z	m	n.a.
Basal heat flow	Q_d	W m^{-2}	n.a.
Heat production	A	W m^{-3}	2×10^{-6} to 4×10^{-6}
Thermal conductivity	k	W m^{-1} °C^{-1}	2.5
Crustal thickness	d_c	m	25,000 to 35,000
Lithospheric mantle thickness	d_m	m	120,000
Crustal density	ρ_c	kg m^{-3}	2800
Lithospheric mantle density	ρ_m	kg m^{-3}	3400
Asthenosphere density	ρ_a	kg m^{-3}	3300
Basin sediment density	ρ_s	kg m^{-3}	2300
Sedimentation or erosion factor	γ	–	0 to 1
Heat capacity	C_p	J kg^{-1} °C^{-1}	1000
Asthenosphere temperature	T_d	°C	1400
Maximum erosion rate	$\bar{v}$	m year^{-1}	0.001
Block edge velocity (extension positive)		m year^{-1}	−0.002 to 0.05

temperature, the heat flux through the lithosphere–asthenosphere boundary varies as the lithosphere thickens or thins. The mantle heat flux, Q_d, used in calculating the initial geotherm is the flux required to achieve the assigned asthenospheric temperature, T_d, at the base of the lithosphere

$$Q_d = -k\frac{(Ad_c^2/2k) - T_d}{d_m + d_c}, \tag{5}$$

where d_m is the initial thickness of the lithospheric mantle.

Sedimentation or erosion is allowed to vary between zero change and that sufficient to maintain isostatic balance. The extent of sedimentation or erosion in the lithospheric column follows the rule

$$d_s = \gamma\frac{(\rho_c - \rho_a)(d_c') + (\rho_m - \rho_a)(d_m')}{(\rho_s - \rho_a)}, \tag{6}$$

where d_c' and d_m' represent the crustal and lithospheric mantle thicknesses, respectively, at a given time, ρ_m is the lithospheric mantle density, ρ_a the asthenospheric density, ρ_c the crustal density, ρ_s the density of sediments, and γ a prescribed constant. The value of γ indicates the fraction of sedimentation or erosion modelled – a value of zero indicates no deposition or erosion, whereas a value of unity produces complete isostatic compensation relative to the original geometry. A maximum erosion rate can also be prescribed.

Heat production was considered to be uniform throughout the crust and absent in the mantle. Different distributions of heat production would affect the temperature profiles (Chamberlain & Sonder, 1990; Sandiford & Hand, 1998; Sandiford & McLaren, 2002; McLaren *et al.*, 2005), but those possibilities are not modelled here. Restriction of heat-producing elements to the upper crust would lower the temperature of the crust at all levels and therefore reduce the likelihood of generating low-pressure anatexis. Further discussion of the potential for concentration of heat-producing elements to drive low-pressure anatexis is given below. Thermal conductivity is assumed constant throughout the lithosphere. In detail, this is not the case (Clauser & Huenges, 1995), but the average conductivity probably does not vary greatly from the values used here and thus the variability does not affect our solutions significantly. For similar reasons, density is assumed constant for each compositional zone. The latent heat of fusion is neglected because we are interested in determining factors that lead to fusion, rather than the thermal evolution following fusion. This omission has implications for the final temperature of melted regions (e.g. the base of the crust), which would be lower due to the buffering effects of the latent heat of fusion (Stüwe, 1995), but does not change the conclusions with regard to low-pressure anatexis. The coefficient of thermal expansion is also neglected. Including thermal expansion affects the total subsidence (McKenzie, 1978), but calculations indicate that the temperature at a given depth is only minimally affected. The model is not sensitive to reasonable variation in lithospheric thickness or asthenosphere temperature. The parameters to which the model is most sensitive (i.e. crustal thickness, strain rate, erosion) are discussed in the following sections.

To simulate the effects of lithospheric delamination, the model varied as follows: without stretching the crust, the temperature of the lithospheric mantle is instantaneously changed to T_d, simulating separation of the lithosphere from the crust. At any time following the detachment event, the lithosphere can be allowed to conductively cool and return to its original thickness.

EVALUATION OF POSSIBLE DRIVING MECHANISMS

Most studies consider that a high thermal gradient in the upper crust, suitable to attain the andalusite field, requires a dominant advection component (e.g. Kühn *et al.*, 2004). Although that is certainly the case in regions of magmatic activity, we wish to explore more broadly the roles that conduction and heat production play, in combination with advection, to produce conditions under which low-pressure anatexis may occur. The specific mechanisms discussed below have been shown to cause low-pressure metamorphism; here we evaluate whether those mechanisms generate anatectic conditions as well. In this paper, we are more concerned with the causes of melt production over a wide area (tens of square kilometres) rather than with the degree of melting. Although the amount of melt generated by any of the mechanisms below depends on a number of factors, including rock composition, water content and time, discussion is restricted to the thermal evolution up to the point of fusion.

Magmatic advection

In numerous sites worldwide, magmatism has heated host rocks through the andalusite and into the sillimanite fields, locally causing anatexis. Magma-related heating is isobaric or nearly so, because of the relatively rapid rate of conduction relative to regional deformation. In several orogens, including the Sierra Nevada (Barton & Hanson, 1989) and Appalachians (Solar & Brown, 1999), voluminous magmatism accompanied low-pressure metamorphism of regional extent. For areas where plutons make up more than 50% of the exposure, magmatically advected heat can explain regional low-pressure metamorphism into the sillimanite field (Barton & Hanson, 1989; Hanson & Barton, 1989) because the heat brought into the crust serves both to raise the background temperature of the surrounding area and cause localized prograde metamorphism (De Yoreo *et al.*, 1991; Stüwe *et al.*, 1993). Nevertheless, even for such high plutonic volume, the host rocks between the plutons do not warm enough to melt or, with some geometries, even to enter the sillimanite field from the andalusite field (see figs 7 & 11 of Hanson & Barton, 1989). So, although heat transferred to the middle and upper crust by rising magma can produce low-pressure metamorphic conditions and localized low-pressure anatexis, advection of magma as plutons does not appear to be sufficient to generate widespread low-pressure anatexis.

As an alternative to bulk magmatic flow, Miyazaki (2004) calculated that pervasive flow of melt (cf. Brown & Solar, 1999; Weinberg, 1999; Leitch & Weinberg, 2002) could raise the temperature of a region to conditions suitable for low-pressure metamorphism and low-pressure anatexis. Thermal modelling indicates that temperatures of 600–800 °C at 15 km depth are attainable (Miyazaki, 2004). Recent studies by White (2005) support these calculations. He quantitatively showed that magma advected from the lower crust, if apportioned evenly across the sink region, can carry sufficient heat to melt pelitic host rocks.

Heat production

Independent of advection, unusually high radiogenic heat production in the crust can generate low-pressure metamorphism (Chamberlain & Sonder, 1990; Sandiford & Hand, 1998; Sandiford *et al.*, 1998). If, for example, a 25-km-thick crust within a 120-km-thick lithosphere were to produce heat to the order of 4 μW m^{-3}, the geotherm would pass through the andalusite field, and the base of the crust would be ~780 °C (Fig. 3). Although high heat production throughout the crust can cause melting in the lower crust, conditions for low-pressure anatexis are not acheived. Thickening of a high heat-producing crust may produce *P–T* paths that pass from the andalusite field across the pelite melting reaction; but in doing so, the base of the crust would be heated to well over 1000 °C (cf. Sandiford & Hand, 1998), even considering the buffering effects of the latent heat of fusion.

Recognizing the unrealistic nature of the whole crust having high heat production and the attendant geological complications, Sandiford *et al.* (1998) showed that rapid burial of a layer of high heat-producing sediments can have a first-order effect on the temperature profile of the crust. For example, crust at 15 km depth may be heated to ~600 °C with a layer of high heat production and a thick (250 km) lithosphere. Sandiford *et al.* (1998) did not address low-pressure anatexis directly, but their model results did not generate conditions appropriate for melting. Nevertheless, based on this work, Goscombe & Hand (2000) hypothesized that anatexis in the Greater Himalayan Sequence of Eastern Nepal may be due to high heat production. They suggested that heat from the melt itself was not sufficient to generate the anatexis, but the long-term cumulative effect of the U- and K-rich leucogranites may have been more substantial. The suggestion that a high heat-producing layer can cause anatexis has not yet been demonstrated quantitatively.

Another form of internal heat production is shear heating. Although not cited as the sole cause for regional low-pressure metamorphism, several studies have quantitatively demonstrated that shear heating can raise the temperature of the deforming and adjacent rocks (e.g. an additional ~100 °C (Stüwe,

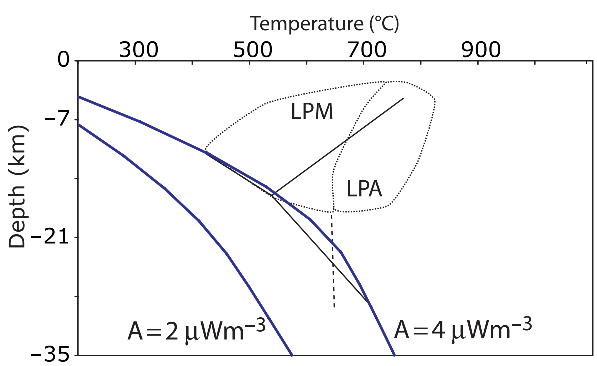

Fig. 3. Effect of high heat production on stable geotherm. Heat production is uniformly distributed and crust is 25-km-thick crust. Different distributions of heat-producing elements would yield different temperature distributions (labelled solid lines), but would not, on their own, produce low-pressure anatexis. Symbols and fields as in Fig. 1.

1998) or produce on the order of 1 μW m^{-3} (Burg & Gerya, 2005). Such heat production does not exert primary influence over the crustal thermal structure, but it may be enough to generate melting in rocks heated by another mechanism.

Lithospheric mantle removal

During or following crustal thinning or thickening, the lithospheric mantle may become unstable. The primary mechanisms proposed for replacement of part or all of the lithospheric mantle with asthenosphere are convective thinning (Houseman *et al.*, 1981; Molnar *et al.*, 1998) and delamination between the crust and the mantle (Bird, 1979; Schott & Schmeling, 1998). Herein, we do not consider the causes or likelihood of such an event, but rather are concerned with the consequences. In this section, crust–mantle delamination operating alone is discussed, and in subsequent sections delamination accompanying crustal thinning and thickening is discussed.

The initial consequence of the asthenosphere rising to or near the base of the crust is a dramatic heating of the lower crust. Bodorkos *et al.* (2002) investigated the thermal consequences of complete lithospheric mantle removal without surface erosion, and concluded that conditions for low-pressure metamorphism (550–600 °C at 17 km) may develop within a 35-km-thick crust if the asthenosphere remains at the base of the crust. Our modelling produced similar results. The thermal anomaly propagates upwards to the middle and upper crust, generating low-pressure metamorphism but not anatectic conditions at pressures less than that of the aluminosilicate triple point in 35-km-thick crust (Fig. 4). The shorter length scale in 25-km-thick crust allows anatectic conditions to develop at 10 km and deeper (Fig. 5). In both cases, the base of the crust must remain hotter than 1000 °C for several million years.

Bodorkos *et al.* (2002) did not incorporate erosion caused by isostatic uplift; our calculations indicate that erosion can have a significant effect on the *P–T* paths of rocks originating in the middle to upper crust. Although we considered mantle delamination to occur instantaneously, a maximum erosion rate of 1 mm year^{-1} was set. The simplification of instantaneous delamination is justified because it represents the maximum possible heating effect on the crust; slower delamination would produce only cooler conditions in the crust at any given time. With the erosion constraint, exhumation in the upper crust dominates, producing nearly isothermal decompression (Fig. 5). The warming effects of the shallow asthenosphere are limited to the base of the crust if uplift and erosion accompanies removal of most or all of the lithospheric mantle. Incorporating erosion into the model for 25-km-thick crust allows the middle crust to develop conditions appropriate for low-pressure anatexis, but only over an unrealistically hot lower crust (Figs 4 & 5).

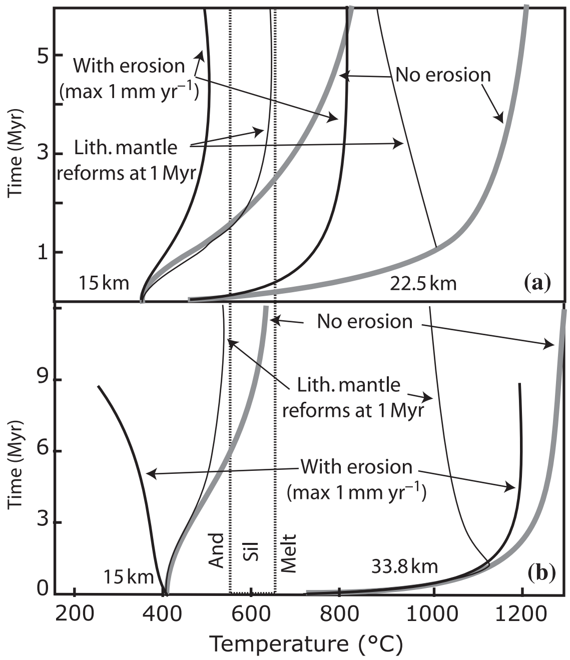

Fig. 4. Time-dependent effects of removing the lithospheric mantle. Temperature-time paths for two depths within (a) 25 km and (b) 35 km thick crust. Labels indicate which paths represent results with and without erosion and with and without reformation of the lithospheric mantle. Vertical lines mark the sillimanite stability field at 15 km depth. See Table S1 for model parameters.

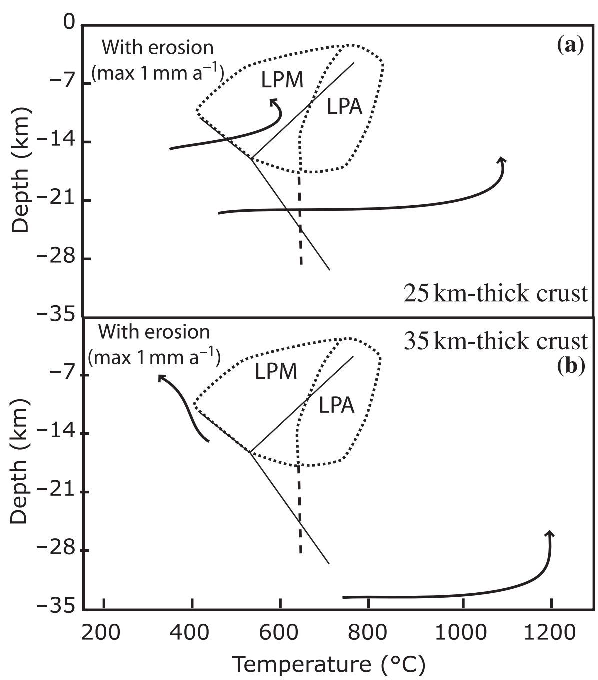

Fig. 5. Effects of erosion accompanying removal of the lithospheric mantle. Temperature-depth paths for rocks starting at 15 km depth and near the base of the crust for crustal thicknesses of (a) 25 km and (b) 35 km. Fields as in Fig. 1. See Table S1 for model parameters.

The above discussion is predicated on asthenosphere remaining near the base of the crust throughout the experiment. If cooling and reformation of the lithospheric mantle commences immediately following its removal, conditions suitable for low-pressure anatexis do not develop. For both 25- and 35-km-thick crusts, the asthenospheric heat source decays too rapidly to infuse sufficient heat into the crust to generate melt, even though low-pressure metamorphism occurs (Fig. 4).

Crustal thickening

Convergent orogens are well-known sites of metamorphism and crustal melting (e.g. England & Thompson, 1986; Jamieson *et al.*, 1998). Homogeneous thickening of the lithosphere results in downward bowing of isotherms in the crust, and heating of individual rocks (Fig. 6, curves E and N; England & Thompson, 1984). The trajectories of crustal *P–T* paths depend on the relative rates of thickening and conduction: faster thickening, and hence higher vertical velocities, produces steeper *P–T* paths; slower thickening and slower vertical velocities produce shallower paths. Thompson (1989) proposed that slow thickening of thinned lithosphere may have produced the conditions for low-pressure metamorphism in the Slave Province, Canada. Because of the nearly isobaric heating required for *P–T* paths that pass from the andalusite field to the muscovite breakdown reaction, crustal thickening must occur slowly if low-pressure anatexis is to develop. After quantitatively exploring this relationship, Loosveld (1989) and Sandiford & Powell (1991) concluded that homogeneous lithospheric thickening alone is insufficient to generate rock trajectories suitable for even low-pressure metamorphism.

Although low-pressure metamorphism cannot occur solely because of homogeneous lithospheric thickening, it may accompany crustal thickening if the lithospheric mantle thins substantially or is removed (Loosveld, 1989; Sandiford & Powell, 1991). Thermal modelling by Loosveld (1989) and Loosveld & Etheridge (1990) suggested that peak temperatures during slow crustal thickening and removal of the lithospheric mantle may reach 550 °C at the depth of the aluminosilicate triple point. Subsequently, Sandiford & Powell (1991) predicted temperatures up to 700 °C at ~20 km depth during thickening of 35-km-thick crust if no lithospheric mantle is present; they predicted cooler temperatures with slower lithospheric mantle thinning. Our calculations support the above conclusions: without accounting for erosion, rocks at all levels of thickening crust over a thinned mantle lithosphere will warm (Fig. 6, curve NM). Incorporating even minimal erosion (maximum of 1 mm year^{-1}) at a shortening rate of -0.002 m year^{-1}, the loss of the dense lithospheric slab overcomes burial caused by shortening, so rocks are exhumed while

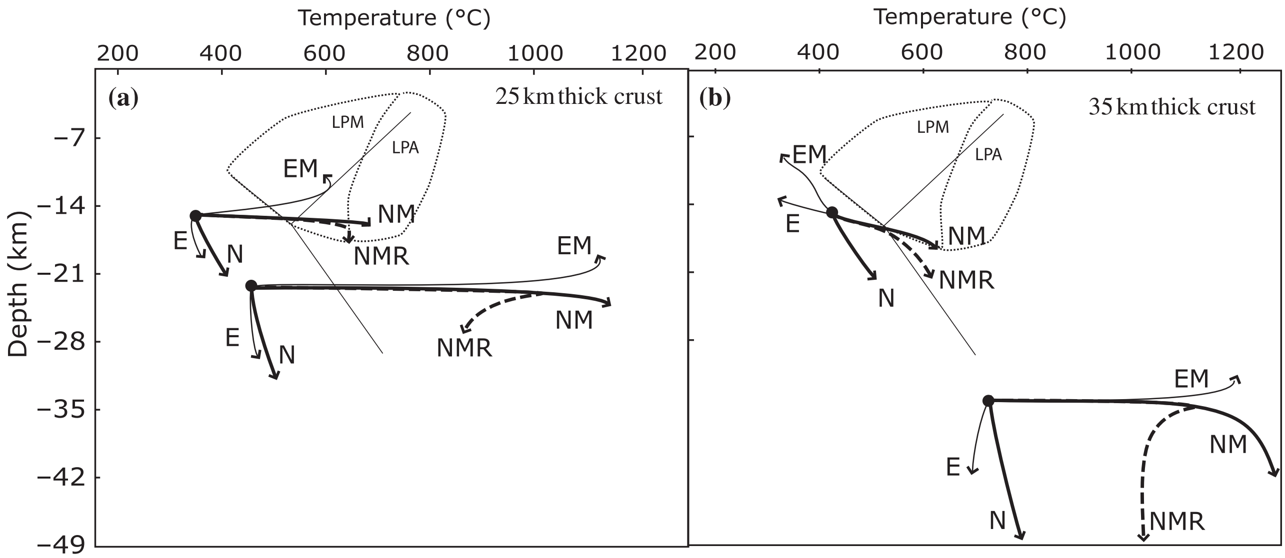

Fig. 6. Effects of crustal thickening. Temperature-depth paths resulting from model runs for crust initially (a) 25 km and (b) 35 km thick. Varied parameters are listed by letter: N, no erosion; E (thin lines), erosion; M, removal of lithospheric mantle; R (dashed lines), reformation of the lithospheric mantle begins 1 Myr after removal. In (a), paths are shown for rocks starting at 15 and 22.5 km depth; in (b), paths shown for rocks starting at 15 km, and 33.8 km depth. Other symbols as in Fig. 1. See Table S1 for complete parameters.

warming (Fig. 6, curve EM). Low-pressure metamorphism conditions may develop in a 35-km-thick crust. With 25-km-thick crust, some mid-crustal rocks may melt after passing through the andalusite field, but the lower crust is heated to over 1100 °C (Fig. 6, curve NM).

In search of causes of low-pressure metamorphism that do not require removal of the lithospheric mantle, Huerta *et al.* (1998, 1999) proposed that much low-pressure metamorphism in collisional orogens may be a product of accretion of heat-producing material. Based on a lithosphere-scale thermal model, Huerta *et al.* (1998) calculated that temperatures >700 °C could occur at 20 km depth in the crust if the heat-producing wedge was >50 km thick and heat production was $>2.5\ \mu W\ m^{-3}$. As in the cases of lithospheric delamination described above, the thermal model of Huerta *et al.* (1998) produces low-pressure metamorphism, but not low-pressure anatexis. In fact, *P–T* paths based on this model do not even transect the andalusite field (Huerta *et al.*, 1999).

Crustal thinning

Because it reduces the length scale of conductive heat transfer between the asthenosphere and the surface, lithospheric thinning has long been considered a driving force for low-pressure metamorphism and possibly low-pressure anatexis. Wickham & Oxburgh (1985, 1987) concluded that shallow asthenosphere was a major component driving metamorphism in the Pyrenees. Other studies in Scotland (Ryan & Soper, 2001), Norway (Pedersen *et al.*, 1998), Antarctica (Smith, 1997), New Zealand (Bibby *et al.*, 1995), and Spain (Cesare & Gomez-Pugnaire, 2001), ascribed the observed thermal structure in large part to high heat flow from the mantle brought on by lithospheric thinning.

Homogeneous thinning of the lithosphere produces cooling paths in the crust if deposition does not occur (Fig. 7, curve N). If the basin produced by subsidence accompanying extension (cf. McKenzie, 1978) fills with sediment, most of the crust cools, but the uppermost crust warms – it is in effect insulated by the sediment. Reasonable variations in any of the parameters, including crustal thickness, lithopheric thickness and strain rate, do not result in heating even into the andalusite field. As such, low-pressure metamorphism can develop during homogeneous crustal thinning only if the lithosphere as a whole behaves inhomogeneously, with the lithospheric mantle thinning much more than the crust (Loosveld, 1989; De Yoreo *et al.*, 1991; Sandiford & Powell, 1991). Conditions for low-pressure anatexis are even more restrictive.

As discussed above, removal of the lithospheric mantle on its own can generate low-pressure metamorphism, but low-pressure anatexis appears unattainable without also inducing wholesale melting of the lower crust. Here we describe quantitative model results from combining lithospheric mantle removal and crustal thinning. With no erosion, an original crustal thickness of 25 km, lithospheric thickness of 120 km, and an extension rate of 0.05 m year^{-1} (distributed uniformly over 100 km), rocks in the thinning crust rise from 15 to 6 km while heating from ~350 to 775 °C over the course of 3.5 Myr (Fig. 7). Conditions such as these can induce low-pressure anatexis, but this anatexis would be accompanied by extensive melting in

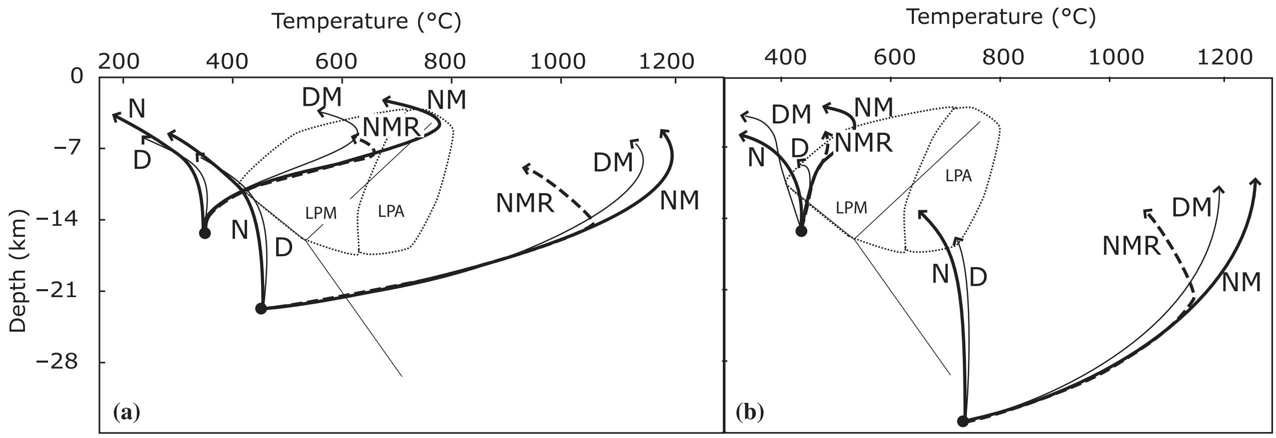

Fig. 7. Effects of crustal thinning. Temperature–depth paths resulting from crustal thinning model runs for crust initially (a) 25 km and (b) 35 km thick. Varied parameters are listed by letter: N, no deposition or erosion; D (thin lines), deposition or erosion; M, removal of lithospheric mantle; R (dashed lines), reformation of the lithospheric mantle begins 1 Myr after removal. In (a), paths are shown for rocks starting at 15 and 22.5 km depth; in (b), paths shown for rocks starting at 15 km and 33.8 km depth. Other symbols as in Fig. 1. See Table S1 for complete parameters.

the lower crust (which warms to nearly 1200 °C, neglecting fusion buffering). If the crust begins 35 km thick, rocks starting at 15 km depth may heat to ~530 °C (Fig. 7), enough to cause low-pressure metamorphism but not low-pressure anatexis. The lower temperature in thicker crust is a result of the longer length scale for conduction. Slower strain rates produce less decompression over time, but can allow more time for conduction before the crust becomes unreasonably thin. For example, extending 25-km-thick crust at a rate of 0.01 m year^{-1} produces maximum temperatures of nearly 810 °C, but again the lowermost crust should melt extensively. Without incorporating erosion, other combinations of stretching rates and initial conditions produce fundamentally the same results as above.

Including erosion results in somewhat different *P–T* paths. Rocks that began at 15 km depth within 25-km-thick crust extending 0.05 m year^{-1} over replaced lithospheric mantle and eroding up to 1 mm year^{-1} may warm to 700 °C, roughly 75 °C less than without erosion (Fig. 7). More significantly, the lower part of the crust achieves only a slightly lower maximum temperature, 1170 °C (compared with 1185 °C without erosion). Other experiments run, varying extension rate, crustal thickness and erosion rate, yielded similar results. Incorporating erosion only serves to cool the upper crust, reducing the solution space for low-pressure anatexis. Moreover, erosion does little to alleviate the problem of extensively overstepping melt-inducing temperatures at the crust–mantle boundary.

Another kinematic scenario for inhomogeneous lithospheric thinning involves inhomogeneous thinning in the crust itself, in the form of a dipping shear zone or detachment fault (Wernicke, 1985; Lister *et al.*, 1986). Extensional activity along this shear zone drops cooler upper crustal rocks onto warmer lower crustal rocks or mantle; the upper plate rocks then heat nearly isobarically. For this detachment mechanism to produce low-pressure anatexis, footwall temperatures must be sufficiently hot to warm the down-dropped hangingwall into the melting field. This would require either thickened crust, as modelled by Zen (1995), or a juxtaposition of middle to upper crustal rocks against the upper mantle. Without thickened crust, temperatures at the base of the crust would likely not exceed 700–750 °C. Escuder Virute (1999) modelled the detachment geometry in two dimensions, with extension occurring at ~25 km depth within a 70-km-thick crust, and concluded that for a shear zone dipping 45° and instantaneous extension, the upper plate could readily develop temperatures > 500 °C. He showed that temperatures much greater than 600 °C, however, were not possible in the upper plate. Zen (1995) predicted low-pressure anatexis with a one-dimensional model, with the detachment located between 22 and 32 km depth within a 60-km-thick crust. The calculations of Zen (1995) employed higher basal heat flow and greater extension than did Escuder Virute (1999), so in Zen's model the upper crust of the hangingwall was juxtaposed against warmer footwall crust. Zen's (1995) model achieves a basal heat flux of ≥60 mW m^{-2}, and even with that high basal heat flow, anatexis does not occur at depths shallower than 14 km. Anatexis at 14 km requires heat production of 2–3 μW m^{-3} throughout the crust, a basal heat flow up to 75 mW m^{-2}, and detachment level of 22 km depth. Four of the five models run by Zen (1995) that produce anatexis at 14 km depth require more than 30 Myr for rocks to undergo prograde metamorphism from the andalusite field to the sillimanite field and continue on to cross the muscovite dehydration melting reaction. Although the solution space is limited, Zen's (1995) model does quantitatively predict that low-pressure anatexis is possible along a crustal-scale detachment fault. A shallower level of detachment or detachment faulting in normal thickness crust could

Table 2. Summary of heat-transfer mechanisms.

Mechanism	Can cause low-pressure anatexis?	Notes	Reference
Plutonism	Locally	Widespread low-pressure metamorphism only if plutons constitute >50% of the exposure	Hanson & Barton (1989), De Yoreo *et al.* (1991)
Pervasive melt migration	Yes	Need sufficient flux; both injected and anatectic migmatitic material exposed	Miyazaki (2004), White (2005)
Fluid advection	No		
High heat production	Possibly*	High surface heat flow, high heat-producing rocks exposed	Sandiford *et al.* (1998), Goscombe & Hand (2000)
Removal of lithospheric mantle alone	Yes	Crust must be less than 35-km-thick Should also induce significant mantle and lower crustal melting	Bodorkos *et al.* (2002)
Crustal thickening	No		Loosveld (1989), Loosveld & Etheridge (1990), Sandiford & Powell (1991)
Crustal thickening and removal of lithospheric mantle	Yes	Should also induce significant mantle and lower crustal melting	Loosveld (1989), Loosveld & Etheridge (1990), Sandiford & Powell (1991)
Crustal thinning (homogeneous)	No		
Crustal thinning (homogeneous) and removal of lithospheric mantle	Yes	Should also induce significant mantle and lower crustal melting	This paper
Crustal thinning along major dipping shear zone	Yes	Successfully modelled for thick crust	Zen (1995); Escuder Virute (1999)

*Qualitative extrapolation of the thermal models done to date suggest this possibility, but quantitative determination has yet to be performed.

produce warmer temperatures at shallower crustal levels and may more readily produce conditions for low-pressure anatexis.

DISCUSSION

Solution space

For low-pressure anatexis to occur under geologically reasonable conditions, two primary criteria must be met: (i) during heating, the middle to upper crust must have low dP/dt; and (ii) while the middle to upper crust heats to melting conditions, the lower crust must stay cool enough not to melt extensively. Qualitatively, the first criterion exists because the geotherm in a tectonically stable region (i.e. before orogenesis and consequent perturbation) lies outside the andalusite field. For a rock to heat through the andalusite field and cross-melting reactions in P–T space, the slope must be sufficiently shallow to allow the rock to remain at a pressure lower than that of the aluminosilicate triple point ($\sim$4.5 kbar) as it heats past 550 °C (triple point of Pattison, 1992).

Conditions suitable for generating low-pressure metamorphism may develop by each of several proposed general mechanisms: (i) magmatic advection (Lux *et al.*, 1986; Hanson & Barton, 1989; De Yoreo *et al.*, 1991; Miyazaki, 2004; White, 2005); (ii) heat production (Chamberlain & Sonder, 1990; Sandiford *et al.*, 1998); (iii) removal of the lithospheric mantle (Loosveld, 1989; Loosveld & Etheridge, 1990; Sandiford & Powell, 1991; Bodorkos *et al.*, 2002); and (iv) crustal extension along detachment faults (Zen, 1995; Escuder Virute, 1999). In these studies, little or no discussion was devoted to whether low-pressure anatexis could develop as well. After evaluating each of those mechanisms, we conclude that four scenarios could generate low-pressure anatexis (Table 2): (i) pervasive migration of melt; (ii) presence of a high-heat producing layer; (iii) crustal thickening or thinning accompanied by removal of the lithospheric mantle; and (iv) crustal extension along a major detachment fault. Of these, crustal thickening or thinning accompanied by removal of the lithospheric mantle may not be geologically reasonable due to the high temperatures attained in the lower crust. Moreover, replacement of the lithospheric mantle by upwelling asthenosphere should induce mantle melting at levels shallower than 40 km (McKenzie & Bickle, 1988). The mantle melts could propagate upward and preheat the crust (Bodorkos *et al.*, 2002). So although mathematically possible, crustal thickening or thinning accompanied by removal of the lithospheric mantle probably cannot produce low-pressure anatexis on its own.

Each of the three remaining mechanisms we consider as viable sole sources of low-pressure anatexis should leave different evidence in the rock record. Evidence for a high heat-producing layer could include high modern heat flow and/or exposed high heat-producing rocks (Goscombe & Hand, 2000). If low-pressure anatexis was generated by detachment faulting, the anatectic rocks should lie within a few kilometres of a detachment zone. Moreover, if low-pressure anatexis was caused by pervasive flow (Miyazaki, 2004), both anatectic and injected magmas should be present.

Multiple mechanisms

Although only a few single processes are capable of inducing low-pressure anatexis, combinations of the many mechanisms shown to drive low-pressure metamorphism may cause melting. Thermal modelling during case studies in Scotland (Ryan & Soper, 2001), the Pyrenees (Wickham & Oxburgh, 1987), and Australia (Bodorkos *et al.*, 2002) leads to the broad conclusion that a combination of advected heat from magma and high basal heat flow, such as is generated by thinning or removal of the lithospheric mantle, is necessary and sufficient to cause low-pressure anatexis although neither mechanism can on its own cause anatexis. Extension or lithospheric removal may precede (Wickham & Oxburgh, 1987; Ryan & Soper, 2001) or post-date (Bodorkos *et al*., 2002) magmatism, but whichever mechanism operates first, it effectively serves to pre-heat the crust and raise the thermal profile well above the position of the equilibrium geotherm. A secondary heat source is then sufficient to warm the crust nearly isobarically to generate low-pressure anatexis. For example, if crustal extension along a detachment fault were sufficient to generate widespread low-pressure metamorphism (Zen, 1995; Escuder Virute, 1999), shear heating or magmatic emplacement could induce low-pressure anatexis. Similarly, if Late Neoproterozoic plutons had intruded the Mt. Painter province of the Flinders Ranges, Australia, the crust that was already at ~500 °C at 3 kbar depth from the extremely high heat production (Sandiford *et al*., 1998) could have melted in place. Many combinations of heat transfer mechanisms could occur in many different geometries to generate low-pressure anatexis, but each mechanism involved should leave evidence of its operation.

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SUPPLEMENTARY MATERIAL

The following supplementary material is available for this article online:

Table S1. Varied parameters for model runs discussed in text.

This material is available as part of the online article from http://www.blackwell-synergy.com