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Tectonic feedback and the ordering of heat producing elements within the continental lithosphere

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Abstract

The distribution of the heat producing elements within the lithosphere provides an important control on continental thermal regimes and the mechanical strength of the lithosphere. Moreover, the strong temperature dependence of lithospheric rheology suggests the possibility of an important feedback between deformation and the distribution of heat producing elements. Simple models for lithospheric rheology are used to illustrate how such feedback might serve as an important control on both the characteristic abundance of, and spatial variation in, the heat production elements in the crust. These models also imply that the organisation of heat producing elements is essential for the long-term tectonic stabilisation of the continental crust. This is particularly relevant to the evolution of cratons in early Earth history, wherein lies the most dramatic evidence for the role played by tectonic processes in achieving a stable ordering of the heat producing elements.

Keywords: tectonic feedback; heat producing elements; heat flow; cratons; intraplate deformation

1. Introduction

In terms of the bulk mechanical properties of the lithosphere a key determinant is the temperature of the uppermost mantle, with the temperature of the Moho providing a useful proxy for lithospheric strength (Fig. 1) [1,2]. In the steady state, the Moho temperature is a function of (1) the thickness of the crust, (2) the distribution of the heat producing elements (HPEs) within the lithosphere, (3) the heat supplied to the base of the lithosphere by convection in the deeper mantle, and (4) the thermal conductivity structure of the lithosphere [3,4]. Surface heat flow–heat production relations imply that HPEs concentrated in the upper 10–15 km of the lithosphere account for 50% or more of the measured surface heat flow of the continents, implying that the continental lithosphere is strongly stratified with respect to the HPEs [5,6]. The significance and origins of this characteristic geochemical stratification of the continental lithosphere, however, have remained poorly understood [7].

Since the distribution of the HPEs impacts on the Moho temperature, it must also impact on the mechanical strength of the lithosphere, suggesting the possibility of a feedback system that may

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modulate the distribution of HPEs in the lithosphere. If this is the case it implies that the geochemical ordering of the continental lithosphere is fundamentally linked to its long-term tectonothermal evolution. This paper investigates the link between tectonic processes and the distribution of HPEs in the continental lithosphere.

2. The distribution of HPEs in the continental crust

Several lines of evidence point to the strong ordering of the HPEs in the continental lithosphere. Most importantly, the heat production rates of typical upper crustal rocks cannot be representative of the whole crust without violating the observed surface heat flows. For example, a crust of typical thickness (~40 km) with uniform heat production of average granite (~2.5 μW m^{-3}; [8]) would contribute ~100 mW m^{-2} to the surface heat flow, well in excess of the average continental surface heat flow of ~65 mW m^{-2} [9]. The view that the lower crust is strongly depleted in HPEs with respect to the upper crust is confirmed by analysis of crustal sections observed in oblique profile (e.g. [10–12]) (Fig. 2), and by analysis of xenoliths derived from the deep crust [13].

It has been observed that many heat flow provinces exhibit a quasi-linear relation between surface heat flow, q_s, and heat production, H_s [5,14,15]. Although the evidence for such a linear relationship has become less compelling as more data have been acquired (e.g. Fig. 3), a general correlation between surface heat flow and near-surface heat production is appealing because it concurs with the observation that a significant fraction of the surface heat flow derives from sources in the upper half of the crust. Moreover, much has been made of this relation because it can be used as a constraint on the vertical distribution of the HPEs [15] (Fig. 3b). For example, the slope of the q_s–H_s regression has been interpreted in terms of the characteristic vertical length scale of surface heat production, h_r, while the intercept has been interpreted as the reduced heat flow, q_r [5]. One interpretation of h_r is that it provides a measure of the thickness of the upper crustal layer that contains the source of the observed variation in surface heat flow. The slopes of such regressions for individual heat flow provinces typically lie in the range 5–12 km [6].

Importantly, as well as containing information about the vertical structure of HPE distributions, surface heat flow–heat production relations also contain information about the horizontal structure of heat production variation [16,17]. Lateral heat transfer, due to horizontal temperature gradients associated with variations in heat production, effectively homogenises the heat flow measured at the surface, thereby reducing the slope of the q_s–H_s regression, yielding an underestimate of the true vertical length scale and an overestimate of the deeper heat flow contribution (Fig. 4). For example, for the classic dataset of Birch et al. [14] lateral heat transfer may easily account for a reduction in the slope of the derived regression line slope, h_r, by ~30%, with the corresponding overestimate of q_r ~50% (Fig. 4b).

In view of the fuzziness of the primary datasets (e.g. Fig. 3b) and the lack of quantitative information about the nature of the horizontal variation of HPEs, the only robust conclusions warranted by the general correlation between surface heat flow and heat production are: (1) that in excess of half the surface heat flow can be attributed to sources in the upper half of the crust, and (2) regression of q_s–H_s data is likely to provide only a lower bound for the characteristic thickness of this upper crustal heat producing region. The remainder of the surface heat flow (for which an upper bound is provided by q_r) can be attributed to sources in the deeper crust, flux from below the mantle lithosphere and, to a minor extent, sources within the mantle lithosphere [18].

Since the classic studies of Birch et al. [14], Roy et al. [15] and Lachenbruch [5], the vertical distribution of HPEs has been characterised in terms of both a characteristic volumetric heat production value (H) and a characteristic length scale (h). The most celebrated HPE distribution model is the *exponential* model [5,19] (Fig. 5) where the heat production at any depth z is given by:

$$H(z) = H_s \exp(-z/h_r) \quad (1)$$

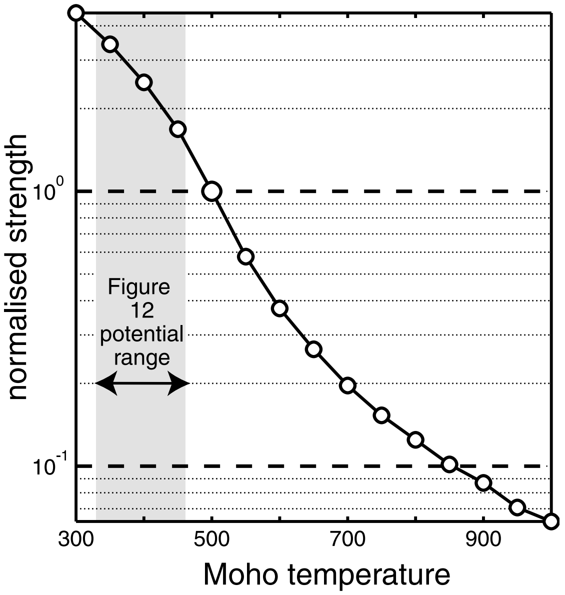

Fig. 1. Illustration of the dependence of the strength of the continental lithosphere on the bulk temperature of the Moho, using a Brace–Goetze rheological model in which the lithosphere is assumed to fail by a combination of frictional sliding and power-law creep. Strength is normalised against the lithospheric strength at $T_{Moho} = 500°C$. A change in T_{Moho} of 100°C leads to a change in bulk strength by a factor of 2–3. Material parameters from [33,34] (Table 2).

where h_r, the characteristic length scale, is defined as the depth at which heat production falls to 1/e of the characteristic surface heat production, H_s. However, in view of the manifest geochemical heterogeneity of the crust, it seems more reasonable to explore the thermal effects of a heat source distribution without regard to the explicit form of the distribution, as is implied by analytical descriptions such as Eq. (1). Thus, we present a general formulation of the characteristic length scale (h), which makes no assumptions about the form of the heat production distribution:

$$h = \frac{1}{q_c}\int_0^{z_c}(H(z)z)\mathrm{d}z \tag{2}$$

where q_c is the depth-integrated heat production, given by:

$$q_c = \int_0^{z_c} H(z)\mathrm{d}z \tag{3}$$

For any lithospheric geotherm we can distinguish between the component of the temperature field due to heat flowing from the deeper mantle, T_{qm}, and the component due to radiogenic sources, T_{qc}. As shown in Fig. 5, T_{qc} attains a maximum value (T'_{qc}) at depths at which heat production becomes negligible (i.e. at the base of the heat-producing layer), where its magnitude is simply:

$$T'_{qc} = \frac{q_c h}{k} \tag{4}$$

where k is the thermal conductivity (strictly, Eq. 4 and the following apply only for temperature-independent conductivity).

This parameterisation emphasises the sensitivity of crustal thermal regimes not only to the total heat contributed by crustal sources (i.e. q_c) but equally to its depth distribution as represented by the parameter h. An important consequence is that processes that act to change the distribution of the HPEs (i.e. the parameters q_c and/or h) must induce long-term changes in lithospheric thermal regimes, independently of any change in the heat flux from the deeper mantle. Moreover, the sensitivity of thermal regimes to these distribution parameters raises the possibility of potential *feedback* mechanisms if the processes that modify and control the distribution of HPEs are themselves sensitive to the thermal structure of the lithosphere. The remainder of this paper explores the connection between tectonic processes that modify HPE distributions and the thermal and mechanical state of the lithosphere.

3. Tectonic modification of HPE distributions

Changes in the distribution of HPEs within the crust are effected by a range of tectonic processes associated with reworking of the continental lithosphere. These processes include magmatism, deformation, sedimentation and erosion. Because we can quantify the changes in the distribution parameters h and q_c that result from these processes, the h–q_c plane can be used to illustrate the change in the distribution of HPEs and the consequent long-term thermal effects.

Fig. 6a–c shows qualitatively how the major

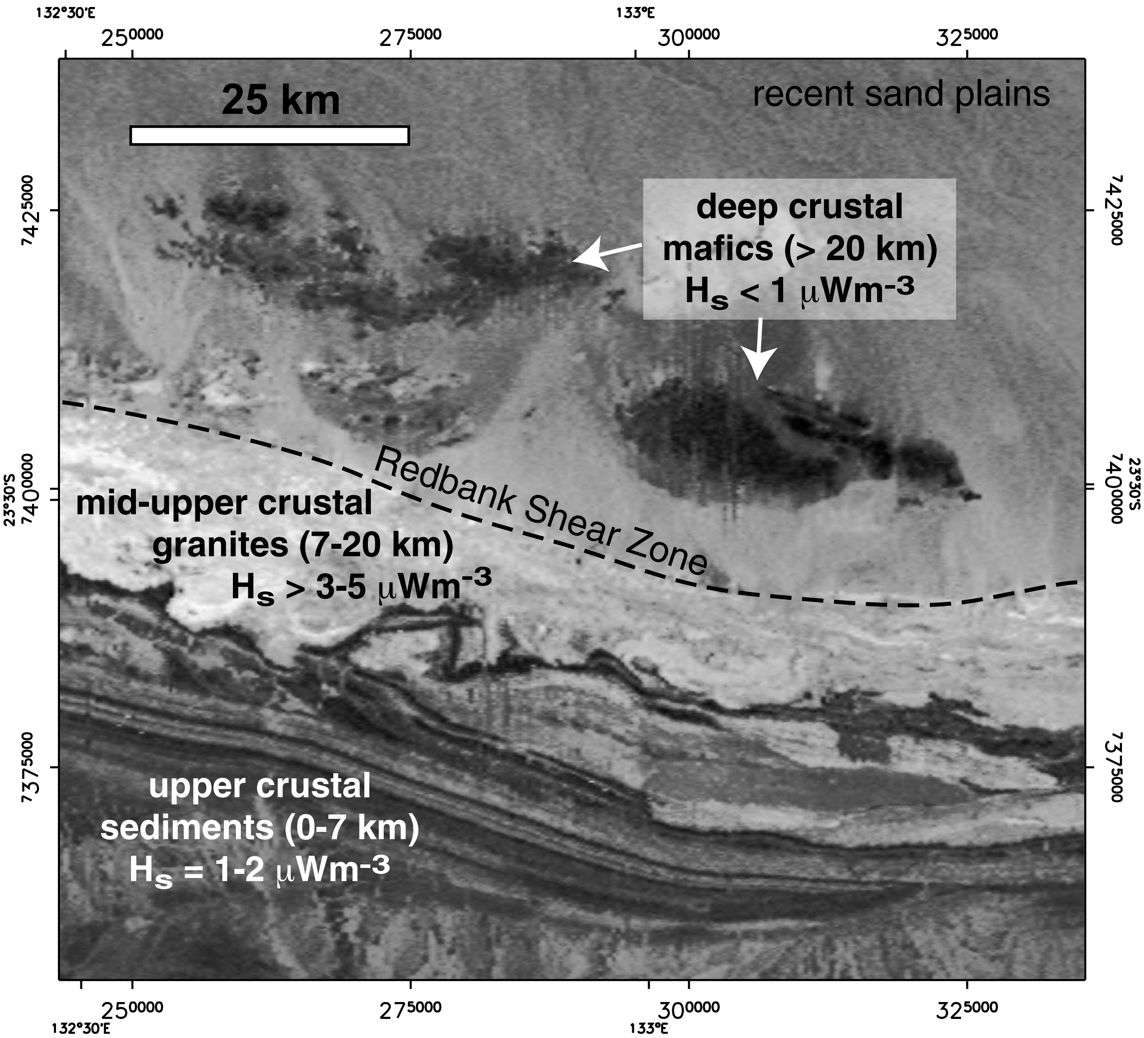

Fig. 2. Image of the distribution of HPEs in an oblique section of crust exposed in central Australia, in the vicinity of Ormiston Gorge [35]. The crustal section comprises a ~7 km thick upper crustal sedimentary package of the Neoproterozoic to Phanerozoic Amadeus Basin in the south; a ~10–15 km thick mid-upper crustal complex of highly radiogenic, Mesoproterozoic granites and metamorphic rocks; and in the north, remnants of a deep crustal, highly depleted, Palaeoproterozoic mafic granulite complex. The various crustal segments were juxtaposed in the Carboniferous across the Redbank Shear Zone, a major crustal penetrating structure associated with one of the largest gravity anomalies known from the continental interiors (see [35]).

processes attendant on crustal reworking are likely to change the HPE distribution using the h–q_c plane. For example, the ascent of granites generated by crustal melting provides an effective mode for moving HPEs upwards in the crust, leading to a reduction in h without reducing q_c. Thus, expressed in terms of relative changes in the heat production parameters, intracrustal magmatism is characterised by $\Delta q_c = 0$ and $\Delta h < 0$. In contrast, deformation can change both the amount and the distribution of crustal heat sources. Crustal shortening (i.e. $\beta < 1$, where β is the ratio of the thickness of the crust prior to and following deformation) increases the total complement of heat sources within the crust, due to structural repetition or layer thickening, and at the same time buries the heat production to deeper crustal levels (i.e. $\Delta q_c > 0$ and $\Delta h > 0$). In con-

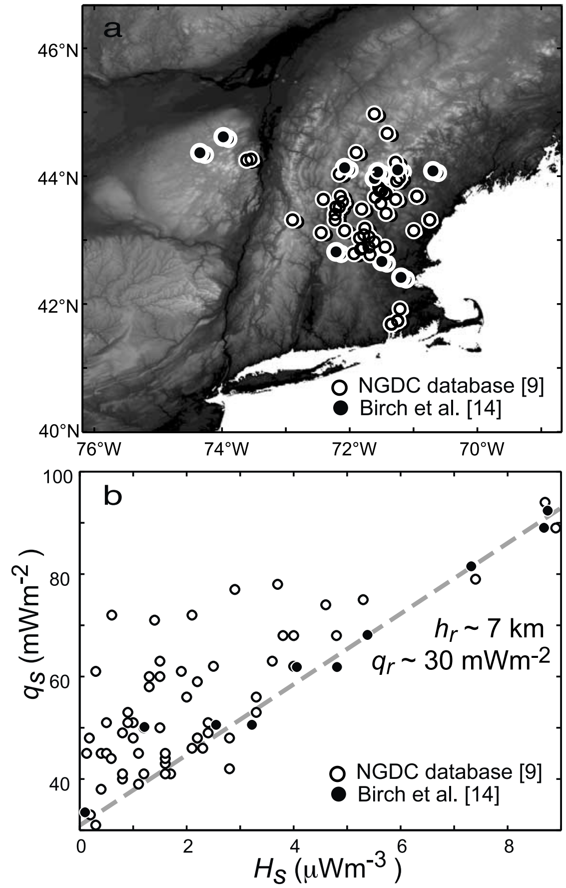

Fig. 3. (a) Distribution of heat flow analyses from the northeastern USA including the Birch et al. [14] data and data from the NGDC database [9]. (b) Surface heat flow (q_s)–heat production (H_s) relations for the data shown in panel a. The regression of the Birch et al. [14] data is shown as the dashed line with y-intercept (q_r) = 30 mW m^{-2} and slope (h_r) = 7 km.

trast, crustal extension (i.e. $\beta > 1$) tends to attenuate the pre-existing heat production and move it to shallower levels (i.e. $\Delta q_c < 0$ and $\Delta h < 0$). Erosion reduces q_c ($\Delta q_c < 0$), particularly when the HPEs are already strongly concentrated in the upper crust. The impact of erosion on h is sensitive to the general form of the initial HPE distribution (Fig. 6c). For *exponential* distributions erosion essentially preserves h [5], while for *homogeneous* distributions erosion reduces h in direct proportion to the reduction in q_c. The effects of sediment deposition depend on the heat production character of the sediments (H_s) relative to the pre-existing crust (Fig. 6c). If H_s is much greater than the mean heat production of the pre-existing crust ($\sim q_c/h$) then sedimentation will result in a significant increase in q_c and may reduce h ($\Delta q_c \gg 0$ and $\Delta h < 0$). In contrast, if $H_s \ll q_c/h$ then sedimentation will lead to a significant increase in h for a comparatively minor increase in q_c ($\Delta q_c > 0$ and $\Delta h > 0$). These concepts are

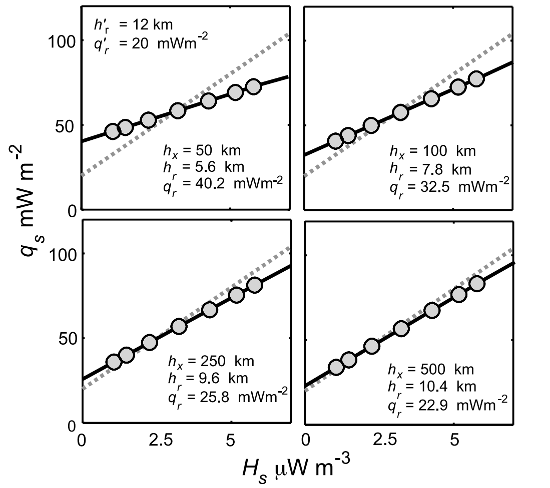

Fig. 4. Illustration of the effect of lateral heat conduction on the surface heat flow (q_s)–surface heat production (H_s) relation for various characteristic horizontal heat production length scales. The variation in HPEs responsible for surface heat flow variation is contained in an upper crustal layer 12 km thick (h'_r). In this layer, H_s varies between 1 and 5 μW m^{-3}, with the horizontal variation following a sinsusoid with a characteristic wavelength h_x (50–500 km). A reduced heat flow, q'_r, of 20 mW m^{-2} is contributed by a mantle heat flow of 15 mW m^{-2} applied at a depth of 150 km, and a lower crustal HPE contribution equivalent of 5 mW m^{-2}. Circles show the apparent surface heat flow as a function of surface heat production, computed by finite element solution of the steady-state heat equation. The best-fit line to the expected observations is shown in the solid lines, together with the values of q_r and h_r derived by regression of the data. For comparison, the dashes show the line $q_s = q'_r + h'_r H_s$ that would apply in the absence of lateral heat flow. For low values of h_x, lateral heat flow produces extreme clockwise rotation of the heat production–heat flow regression line. These simulations show that if the characteristic horizontal length scale for variation in crustal heat production is of the order 100 km (Fig. 4a,b), then the reduced heat flow values derived from natural datasets (such as shown in Fig. 3b) may be as much as a factor of 2 higher than the characteristic deep crustal heat flow, q_r.

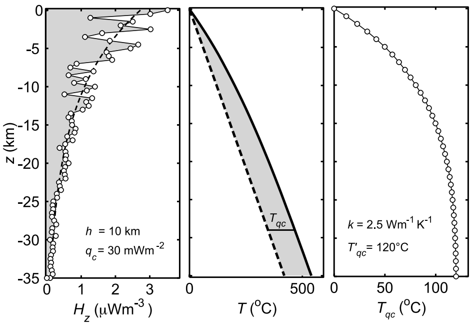

Fig. 5. (a) Hypothetical heat production distribution approximating an exponentially decreasing distribution of the form $H(z) = H_s \exp(-z/h_r)$). Additional 'noise' has been added to simulate the heterogeneity of the continental lithosphere. (b) The temperature field produced by such a distribution illustrating the component due to the non-radiogenic heat flow (the dashed line), and the component due to heat production in the lithosphere (the shaded area). (c) The temperature deviation, T_{qc}, is defined as the additional component of the temperature field due to heat production and corresponds to the shaded region in Fig. 5b, with the maximum value, T'_{qc}, attained at the base of the heat-producing layer.

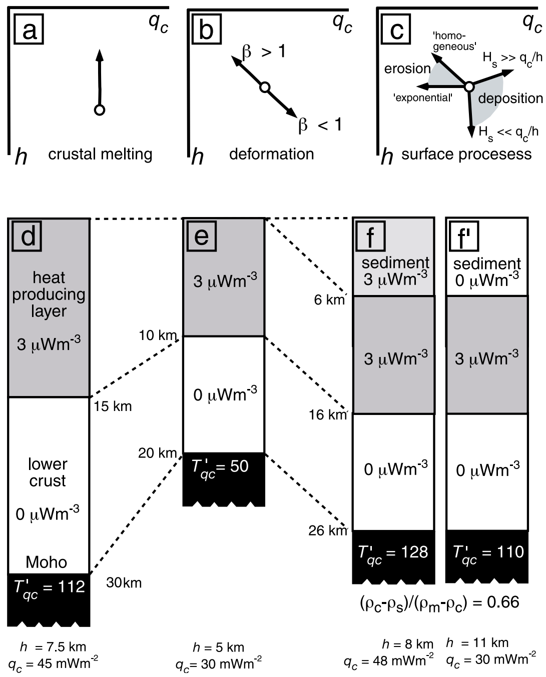

Fig. 6. (a–c) Schematic illustration of the way tectonic processes associated with reworking of the continental lithosphere modify the HPE distribution parameters, h and q_c. The ascent of granites generated by crustal melting provides a mode for redistributing HPEs upwards in the crust without reducing q_c. Crustal deformation will either attenuate and shallow the HPE distribution (for crustal extension, i.e. $\beta > 1$) or concentrate and bury it (for crustal thickening, i.e. $\beta < 1$). Erosion will normally reduce q_c, when HPEs are already concentrated in the upper crust. The impact of erosion on the length scale depends on the general form of the HPE distribution in the crust. The length scale for approximating an exponential distribution (see Appendix) is essentially preserved during erosion [5], while for homogeneous distributions the length scale is reduced in direct proportion to q_c. Likewise, the effect of sedimentation depends on the heat production character of the sediments (H_s) relative to the pre-existing crust. If H_s is much greater than the mean heat production of the pre-existing crust ($\sim q_c/h$) then sedimentation will result in a significant increase in q_c and may reduce h. In contrast, if $H_s \ll q_c/h$ then sedimentation will lead to a significant increase in h and a relatively minor increase in q_c. (d–f) Illustration of the role of deformation coupled with an isostatic response on heat production distributions in the crust. Given an initial lithosphere (d), stretching (e) results in attenuation of pre-existing heat production, while the surface response of basin formation (f) moves it to deeper levels (i.e. an increase in h). The final heat production distribution following restoration of the original surface elevation depends on the heat production character of the basin-filling sediments (with q_c either decreased or increased as shown in panels f and f′).

illustrated in a 1-D crustal section for the case of crustal thinning and subsequent basin formation (Fig. 6d–f).

In the long term, deformation, erosion and sedimentation are linked through the isostatic response of the lithosphere, with the surface processes tending to restore the long-term elevation of the continents to near sea-level. Potentially, this coupling may serve to order the distribution of the HPEs. In the remainder of this section we investigate quantitatively the way in which deformation, erosion and sedimentation can modify various heat source distributions and their consequent impact on crustal thermal regimes. We use two simple models for the distribution of heat sources within the crust, the *exponential* and *homogeneous* models. These models and the assumptions on which they are based, are outlined in the Appendix.

The results of these calculations for the case of a vertical stretch of 10% (thinning or thickening) combined with a surface response that restores the initial surface elevation (sedimentation or erosion), are summarised in Figs. 7 and 8. Fig. 7 shows the pattern of changes in the HPE distribution parameters and Fig. 8 shows the associ-

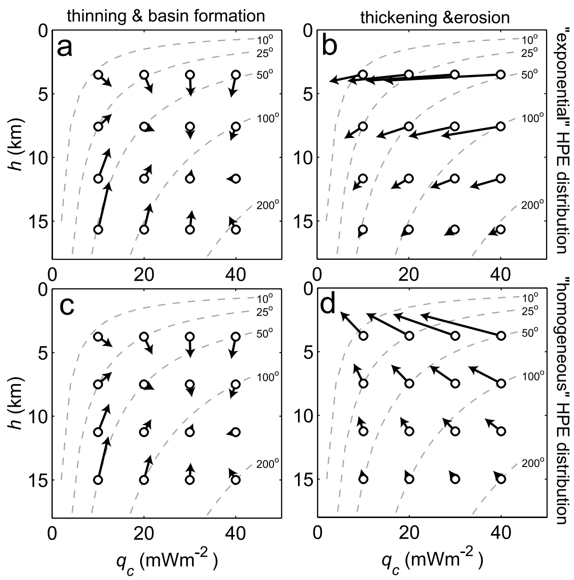

Fig. 7. Impact of small strain events on the HPE distribution parameters, h and q_c, for exponential (a,b) and homogeneous HPE distributions (c,d) (see Appendix for HPE distribution models). The deformation equates to a 10% thinning ($\beta = 1.1$) and a 10% thickening ($\beta = 0.9$), where the β is used in its conventional sense when applied to a 1-D lithospheric column, that is, as the ratio of the initial to deformed crustal thickness. The circles indicate the initial h and q_c values; the arrows show the change in the HPE parameters, due to the combined effect of thinning and basin formation (sedimentation) (a,c), and thickening and erosion (b,d). Diagrams are contoured for T'_{qc}.

ated long-term thermal response T'_{qc}. Figs. 7a,b and 8a,b show the case for an initially exponential HPE distribution and Figs. 7c,d and 8c,d show the case for a homogeneous distribution. As summarised below, these figures illustrate a number of important features concerning the way in which the HPE distribution parameters evolve as a consequence of the isostatic coupling between deformation and surface processes.

The coupling of crustal thickening and erosion provides an effective means for modifying HPE parameters especially when crustal heat production is already concentrated in the upper crust (i.e. initially low h). Although crustal thickening tends to increase q_c, erosion will reduce it. For very low-h configurations the loss of heat production during erosion is much greater than the increase in heat production due to thickening (Fig. 7d), with the linked deformation–surface process resulting in a dramatic decrease in q_c. This can be readily appreciated by the fact that the coupling of crustal thickening and erosion essentially effects the removal of upper crustal material and replaces it with lower crustal material. The consequences for the length scale, h, depend on the initial distribution of heat sources. For the homogeneous distribution, h will be reduced as a consequence of thickening and erosion (Fig. 7d). For the exponential distribution, providing $h \ll z_c$, h will increase by the factor $1/\beta$, because erosion preserves the exponential length scale established by the deformation ([5], Fig. 7b). Fig. 8b,d shows that the long-term changes in the thermal structure of the deep crust resulting from thickening and erosion can be significant. For typical crustal heat production parameters, a 10% thickening fol-

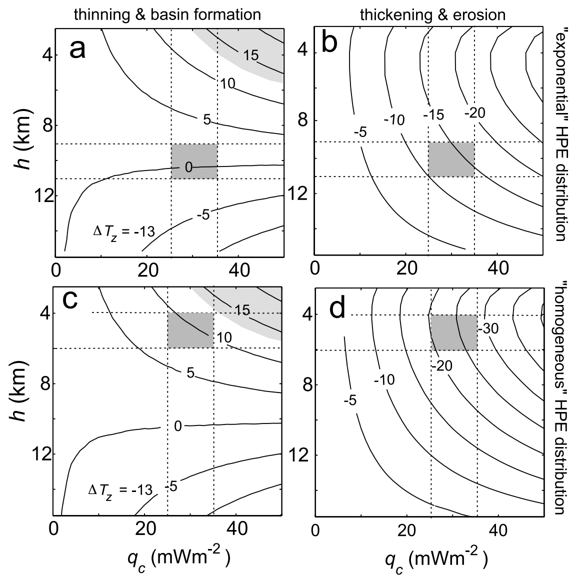

Fig. 8. As for Fig. 7 showing the h–q_c plane contoured for changes in T'_{qc}, i.e. $\Delta T'_{qc}$. The change in Moho temperature is given by $\Delta T'_{qc} - \Delta T_z$, where ΔT_z is the temperature change due to Moho shallowing (see text). In the case of stretching, the depth of the Moho is reduced in the long term because the density of sediments is less than that of the crust. Because of this reduction, the Moho is cooled (by $\sim$13°C for the parameters assumed here; $q_m = 25$ mW m^{-2}). The dark grey box shows the range in h–q_c parameters typical of modern continental crust, derived from compilations in [6,28,29]. Light grey shaded areas show the combination of $h - q_c$ parameters permitting long-term Moho heating.

lowed by erosion will lead to a reduction in deep crustal temperatures by 10–30°C. Because the Moho will be returned to its original depth following erosion, the Moho will experience all of this cooling.

For crustal thinning, the long-term thermal consequences are dependent on whether basin formation occurs and, if so, the heat production parameters of the basin-filling sediments. As outlined in the Appendix, in these models we assume that the basin fill has a heat production of 1.5 μW m^{-3} [20] and a thermal conductivity of 3 W m^{-1} K^{-1}. Crustal thinning without basin formation will result in a reduction in both q_c and h by the factor $1/\beta$ for all heat source models. Where the subsidence resulting from crustal thinning is accompanied by sedimentation sufficient to restore the original surface elevation, the resulting changes in h and q_c may lead to either an increase or decrease in T'_{qc} [21]. As shown by Fig. 8a,c, increases in T'_{qc} are expected when h is less than about 10 km. In contrast, high h configurations tend to lead to long-term reductions in T'_{qc} (up to $\sim$10°C). We note, however, that all of these values are sensitive to both the assumed density and heat production character of the basin filling sediments, with more dense and/or more radiogenic sediments leading to increased values of T'_{qc} for a given stretch [21].

For typical basin fill densities, the Moho will not be returned to its original depth in the long term. Consequently, the estimates for T'_{qc} as shown in Fig. 8a,c will not equate with long-term changes in the Moho temperature. The amount of Moho shallowing is dependent on the density contrast between the basin-fill and the pre-existing crust, while the cooling of the Moho resulting from this long-term shallowing depends on the prevailing thermal gradient. For a deep crustal thermal gradient of $\sim$10°C/km, the long-term shallowing of the Moho following a 10% thinning leads to cooling, ΔT_z, of -13°C. The actual long-term change in Moho temperatures is a function of both the change due to the Moho shallowing and any effect of the redistribution of the sources, thus, $\Delta T_{Moho}=\Delta T'_{qc}+\Delta T_z$. For the parameters used in our calculations, the characteristic response to crustal thinning and basin filling is Moho cooling; only extremely high q_c, low h initial configurations lead to significant long-term Moho heating (i.e. the shaded region in the upper right of Fig. 8a,c).

The coupling of crustal thickening and erosion is ineffective in changing the heat production parameters when the crust is largely undifferentiated (i.e. large h), because the heat production character of the lower crust is essentially the same as the upper crust (Fig. 8b,d).

One context for understanding the potential significance of these calculations is provided by the widespread, low level tectonic activity indicated by the intraplate seismicity and neotectonic structures within modern continental interiors such as Australia. This activity suggests that the long-term evolution of continental interiors is punctuated by many small magnitude deformation events (< 10% stretches), associated with minor reworking and reactivation. Such events involve both crustal shortening and extension, and are typically associated with small amplitude changes in surface elevation (of less than about 1–2 km). These processes, in turn, induce a surface process response that tends, in the long term, to keep the crustal thickness at the continental average of $\sim$35 km. The calculations summarised in Figs. 7 and 8 suggest that the repeated imposition of such deformation can significantly redistribute the HPEs. For individual events involving stretches of $\sim$10% the redistribution of HPEs will affect the steady-state temperature field in the deep crust and upper mantle by up to $\sim$30°C (e.g. Fig. 8d). One implication of these results is that the repeated 'processing' of continental crust by the type of mild events that typify continental interiors is likely to effect significant long-term thermal structuring of the lithosphere and therefore should impact on its long-term mechanical behaviour.

4. Mechanical consequences of the redistribution of HPEs

The strong temperature dependence of lithospheric rheology [22,23] suggests that continental regions characterised by elevated thermal regimes

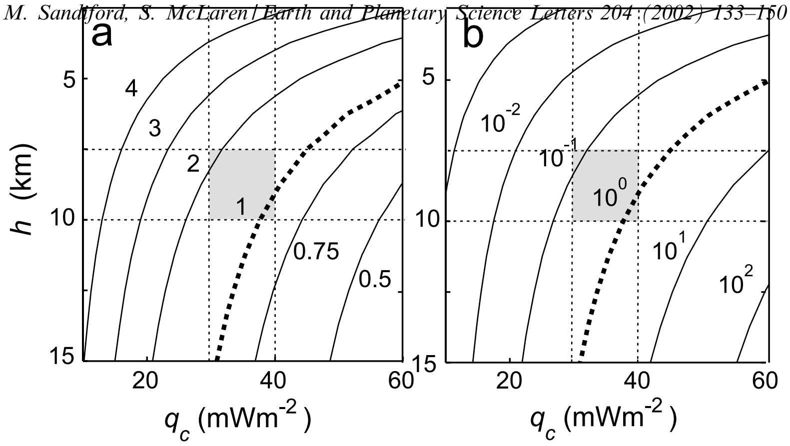

Fig. 9. (a) h–q_c plane contoured for integrated lithospheric strength. (b) h–q_c plane contoured for rate of deformation subject to an imposed tectonic load. The assumed rheology is that of the Brace–Goetze lithosphere, in which deformation is governed by a combination of frictional sliding and temperature-dependent creep (rheological parameters from [33,34]; Table 2). The strength parameters are shown normalised against a configuration characterised by $q_c = 45$ mW m^{-2} and $h = 7$ km. The shaded area illustrates that, all other factors being equal, a 25% reduction in both h and q_c will result in an increase in strength by a factor of 2 or a decrease in strain rate of an order of magnitude in response to an imposed load. Note that the calculations are sensitive to a large number of assumed parameters, including thermal conductivity (here $k = 3$ W m^{-1} K^{-1}) and mantle heat flux (here $q_m = 30$ mW m^{-2}), as well as the material parameters constraining creep and frictional sliding of crustal and mantle rocks [33,34].

associated with high q_c and/or h values will generally be weaker than regions characterised with low q_c and h values, and therefore more susceptible to deformation in response to a given tectonic load. In this section we attempt to quantify how the variations in q_c and h documented in the previous section might affect the strength of the lithosphere.

In all our calculations we use a Brace–Goetze model for lithospheric rheology that combines both temperature-independent, frictional sliding and temperature-dependent creep mechanisms (e.g. [22]; Table 2). The application of the *strength-envelope* approach implicit in this model has become widespread in geodynamic problems. Nevertheless, it is important to note that any calculations of lithospheric strength are subject to very large uncertainties. Basic uncertainty stems from a number of sources [24], including: (1) the extrapolation of rheological flow laws from laboratory conditions and time scales to the geological realm, (2) an imprecise understanding of the compositional and mineralogical structure of the lithosphere, which results in uncertainties in knowing what particular flow laws to apply to which part of the lithosphere, and (3) uncertainties in the absolute thermal structure of the lithosphere due to imprecise knowledge of its thermal property structure, particularly thermal conductivity. These problems make any calculation of absolute strength subject to very large uncertainties, and in the following section we emphasise the role of temperature changes accompanying the various tectonic processes to *relative* changes in lithospheric strength. Although the absolute magnitude of these relative changes will vary with the composition and thermal structure of the reference frame, the relative changes are likely to be far more robust than any absolute measure of strength, because many of the uncertainties described above will cancel.

Recognising that the h–q_c plane can be contoured for thermal structure (e.g. Fig. 8) also allows it to be contoured for lithospheric strength (Fig. 9). Fig. 9a shows variations in the strength of a Brace–Goetze lithosphere forced to deform at a specified strain rate, while Fig. 9b shows variations in the effective strain rate for the lithosphere

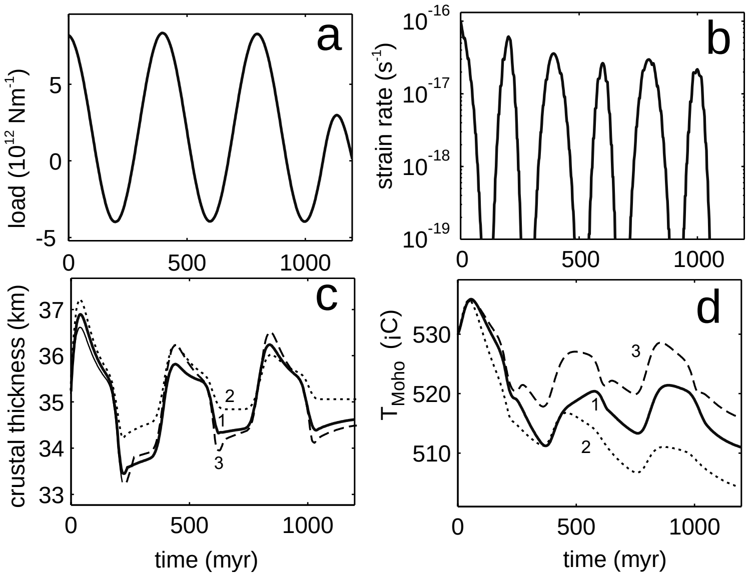

Fig. 10. Results of numerical simulations designed to explore the long-term changes in the HPE parameters caused by extremely low-level tectonic deformation in response to time-varying tectonic loads that may be appropriate to the evolution of continental interiors. Here the simulation runs for ∼1200 million years. (a) This shows the variation in tectonic load; (b) shows the variation in strain rate. During periods of high tectonic load these low-strain rate deformations lead to small (1–2 km) changes in crustal thickness, as shown in panel c. In the models, the associated isostatic response induces a surface process that restores the initial crustal thickness over a characteristic time scale. Moho temperatures (d) evolve as a function of changes in crustal thickness as well as the long-term changes in the HPE parameters due to deformation and the associated surface processes (see Fig. 11). The three paths (1–3) shown in panels c and d correspond to different initial HPE configurations, as shown by the open circles in Fig. 11c.

when subject to an imposed tectonic load. For the Brace–Goetze lithosphere, varying either h or q_c by a factor of 2 leads to a variation in effective strength by a factor of 2–3, and effective strain rate by one to two orders of magnitude. The sensitivity of effective strain rates to the heat production parameters highlights the inherent temperature sensitivity of the Brace–Goetze lithosphere and raises the possibility of feedback between deformation processes and HPE distributions.

Although Fig. 9 illustrates the way in which variations in HPE parameters may alter the strength of the lithosphere, it is important to realise that other factors also influence lithospheric strength. In particular, long-term changes in the depth of the Moho, as may be expected following basin formation (as outlined in Section 3), will also contribute to variations in lithospheric strength. In the case of basin formation, the cooling associated with long-term Moho shallowing tends to increase the strength of the lithosphere, while the changes in HPE parameters usually tend to decrease its strength (e.g. Fig. 8a,c). Depending on the initial HPE distribution parameters, basin formation may either result in mild long-term weakening or strengthening of the lithosphere [21].

In order to explore how the distribution of HPEs impacts on the mechanical response of the lithosphere, and the way in which this response might in turn change the HPE distribution (motivated by the possibility of a feedback mechanism), we have carried out a set of numerical simulations. These simulations are designed to evaluate the response of a lithosphere with temperature sensitive rheology to long-term, time-dependent loading (Figs. 10 and 11). We have chosen a load history that varies from compressional to extensional on time scales of around 300 million years (Fig. 10a). This load history is also scaled to provide a mean state that is mildly compressional (as is suggested by the potential energy differences

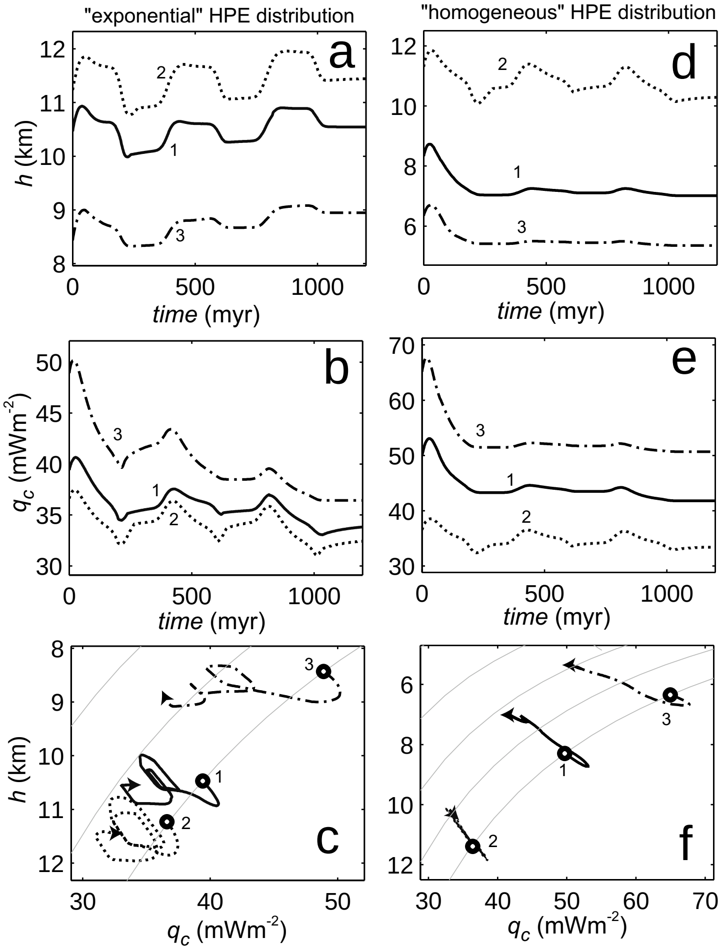

Fig. 11. The changes in HPE distribution parameters accruing for the models shown in Fig. 10. (a,c) The case for an initially exponential HPE distribution; (d,f) for an initially homogeneous HPE distribution. Panels a and d show the changes in h with time; panels b and e show the changes in q_c with time. The initial HPE parameters for each of the models (1–3) are indicated by the open circles in panels c and f. These configurations give identical initial Moho temperatures (see Fig. 10d) and highlight the fact that the evolving HPE distribution can lead to significant differences in the long-term evolution of the system. Contours in panels c and f show intervals of T'_{qc} of 20°C and imply long-term cooling of the deep crust of 10–40°C as a function of the ordering of the HPE parameters (see text for further discussion).

between mid-ocean ridges and normal continental crust [25]), and such that the deformation rates during the maximum loading intervals are sufficiently low ($<10^{-16}$ s^{-1}, Fig. 10b), particularly when compared to plate margin deformation rates ($\sim 10^{-14}$ s^{-1}) that stretch increments during any discrete 'episode' are only small ($<\sim 5\%$). The long-term behaviour of the model is mediated by the isostatic response to the ongoing deformation. This elevates or depresses the surface inducing a restorative surface process (sedimentation or erosion).

In accord with our earlier discussion, these simulations show that the response of the lithosphere is sensitive to its thermal state, and that HPE distributions are significantly modified by repeated small-scale tectonic reworking of the lithosphere (Fig. 11). The simulations also show that initial configurations characterised by high q_c (> 35 mW m^{-2}) experience significant reductions in q_c during repeated tectonic reworking, even though the scale of the reworking is very minor (Fig. 11c,f). For an initially exponential distribution, the erosion-dominated reduction in q_c (Fig. 11b) tends to largely preserve the initial value of h (Fig. 11a) in accord with the results in Fig. 7. In contrast, initially homogeneous distributions can show significant reduction in h, as well as reduction in q_c, as a result of tectonic working (Fig. 11d,e). Reductions in q_c of up to 20% equate to long-term Moho cooling of ~20–40°C (Fig. 10d) and result in an increase in crustal strength (Fig. 9).

Although these models are clearly very sensitive to the input parameters, they do highlight the potential role that tectonic reworking may play in modulating HPE distributions in the continental crust. Crust initially configured with elevated q_c and/or h will, through the isostatic coupling of deformation and surface processes, tend to both redistribute, and shed, HPEs until it ceases to respond to the normal fluctuation in the tectonic load. In terms of the HPE distribution parameters, this is expressed as a reduction in q_c and, depending on the form of the initial distribution, a reduction in h.

5. Lateral heat transfer and the natural variation in heat production

In addition to dramatic vertical differentiation, the continental lithosphere also shows significant lateral heterogeneity in HPE distributions (Figs. 2 and 3). As shown in Fig. 4, and pointed out by Jaupart [16], such lateral variations in the distribution of HPEs may significantly impact upon the relationship between surface heat flow and measured upper crustal heat production. However, despite recognition of the importance of lateral heat flow, comparatively little is known about the natural length scales for the horizontal variation in HPEs within the continents. This section explores the possibility that the amplitudes and length scales of horizontal variation in crustal heat production are limited by the thermo-mechanical response of the lithosphere.

The continental heat flow record shows significant variability [9]. Much of the observed variation can be attributed to transients associated with active tectonism and/or magmatism. However, even in areas where there is no evidence of such young activity there is substantial variation in surface heat flow, from as low as about 35 mW m^{-2} to at least 100 mW m^{-2}. Much of this variation can be attributed to the crustal complement of HPEs. For example, in South Australia, heat flow in Proterozoic terranes varies from ~50 mW m^{-2} to 90 mW m^{-2} and correlates with a variation in the median heat production rates of granitic rocks between 2.5 and 6 μW m^{-3} [26]. Variations in surface heat flow of this magnitude would, if attributable to HPE in the upper half of the crust, equate to variations in Moho temperature of the order of 100°C (i.e. ~$h\Delta q_c/k$), provided that the characteristic horizontal length scale of this variation was sufficiently large to prevent significant lateral heat flow. As implicit in Fig. 1, such large variations in Moho temperature imply variations in mechanical strength of the lithosphere by a factor of 2–3 and would be expected to lead to the selective exploitation of the thermally weakened zones (i.e. high q_c and/or h) by intraplate tectonism. As discussed in the previous section, the notion of tectonic feedback as proposed here suggests a long-term tendency for the heat production parameters to converge towards a region of h–q_c space that imparts sufficient strength to the lithosphere to withstand the natural fluctuation in intraplate stress levels associated with evolving plate configurations.

Of course, this last statement can only be relevant to domains sufficiently large such that the lithosphere is able to accommodate significant strain when subject to an in-plane stress. This implies that the minimum horizontal dimension relevant to heat production reorganisation is likely to be at least the order of the thickness of the lithosphere (i.e. 100–200 km). Consequently, variations in the HPE parameters at smaller spatial scales will be largely irrelevant to the mechanical response of the lithosphere, allowing short-wavelength, near-surface HPE enrichments to be preserved. Moreover, for wavelengths of less than

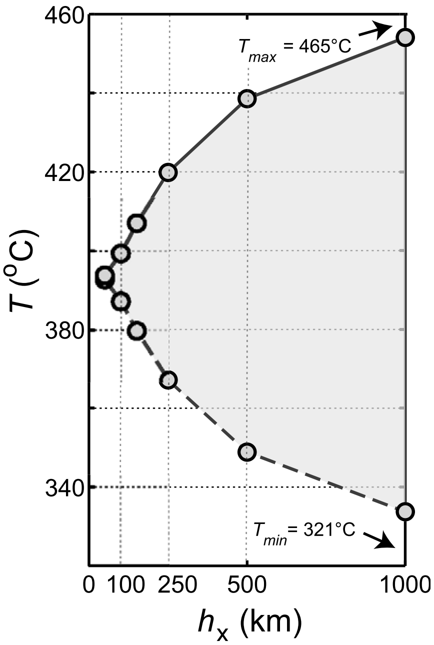

Fig. 12. Illustration of the variation in Moho temperature as a function of different characteristic wavelengths (h_x) of horizontal variation in q_c. See also Fig. 4. The amplitude in q_c variation is 48 mW m^{-2}, (from 17 to 65 mW m^{-2}), yielding theoretical minimum and maximum Moho temperatures of 320 and 465°C for the modelled parameters ($k = 2.5$ W m^{-1} K^{-1}). The upper (solid) curve represents the Moho temperature beneath the most enriched parts of the crust, while the lower (dashed) curve represents the Moho temperature beneath the most depleted parts of the crust. The amplitude in the Moho temperature variation is given by the difference between the two curves at the appropriate value of h_x. For h_x less than 100 km the amplitude in Moho temperature variation is extremely attenuated (< 30°C). Because the mechanical strength of the lithosphere is sensitive to Moho temperature (see text for discussion), the horizontal length scale for heat production variation therefore impacts on the interplay between HPE distributions and the long-term mechanical response of the lithosphere.

about 100 km, the effects of lateral thermal conduction substantially attenuate the variations in temperature at Moho depths (Fig. 12), limiting further any mechanical instability. In view of the hypothesis that tectonic feedback is important in organising the distribution of HPEs, these considerations suggest that the natural length scale for the horizontal variation of heat production parameters may be less than ~100 km.

6. HPE redistribution during craton formation

The correlation between the distribution of HPEs and the strength of the continental lithosphere implies that the differentiation of HPEs is essential for the long-term stability of the continental lithosphere. Our understanding of the stabilisation of continental lithosphere is intimately intertwined with the notion of craton development. While craton formation is undoubtedly a complex process, most probably involving long-term changes in crust–mantle interaction, the analysis of tectonic reworking presented in this paper suggests that crustal-scale differentiation of HPEs is a necessary and important part of this process. Although cratons are manifest by long-term tectonic stability, cratonisation is usually preceded by a lengthy history involving not only crustal growth but also extensive crustal reworking spanning many hundreds of millions of years. In a thermal and mechanical sense such reworking can be viewed in terms of the way it redistributes the HPEs [27].

Such a view may be consistent with the observation that Archaean terranes are characterised not only by lower surface heat flow (~40 mW m^{-2}), but also by lower values of h than most younger geological provinces. The compilation of Taylor and McLennan [6] (see also [28,29]) suggests that the mean (and standard deviation) length scale for heat production in Archaean terranes is 6.9 ± 1.7 km, compared to 10.1 ± 3.6 km in Proterozoic terranes. In part, the low surface heat flow in Archaean cratons is due to the low present-day abundance of HPEs, but also to the presence of very thick mantle lithosphere, which has the effect of diminishing the mantle contribution. For example, McLennan and Taylor [20] have suggested that the mantle heat flow beneath Archaean cratons is typically ~14 mW m^{-2}, implying that the HPEs contribute ~25 mW m^{-2}. Given that at 3 Ga heat production was more than twice modern day rates, the characteristic Archaean q_c could conceivably have been as high as 50 mW m^{-2} (i.e. some 50–66% greater than typical of the modern continents).

One way to mechanically stabilise lithosphere with such high values of crustal radiogenic heat production is to concentrate the HPEs at shallow levels in the crust (i.e. low h). We note with interest that in many Archaean cratons the crustal complement of HPEs is carried in granites that form the cores of large 'diapiric' domes [30,31]. Indeed the origin of the characteristic 'dome and basin' geometry of these Archaean granite–greenstone terranes has been the subject of much discussion and we speculate that such geometries may reflect a distinctive 'high-q_c' mode of stabilising continental crust by redistributing the HPEs into the shallow crust [31].

7. Concluding remarks

The origin of the characteristic HPE distribution inferred from the analysis of surface heat flow and heat production data is central to the differentiation of the continental crust, not least because the distribution of HPEs provides a fundamental control on the thermal and mechanical state of the lithosphere. The available data imply that HPEs are concentrated in the upper 10–15 km and contribute on average $\sim$30–40 mW m^{-2} to the observed surface heat flow. An inevitable consequence of the lithophile nature of the HPEs is that primary crustal growth processes dominated by magmatism will lead to values of h that are significantly less than the characteristic crustal thickness. Nevertheless, no compelling reason has emerged as to why these primary processes would naturally lead to such a narrow range of characteristic length scale or characteristic abundance of the HPEs.

However, the gross chemical and structural architecture of the crust is not only a function of primary crustal accretion, but also reflects complex tectonic reworking in zones of continental deformation, both at plate boundaries (e.g. collisional orogens) and in intraplate settings. In both cases the continental crust is subject to temporal fluctuations in tectonic loads due to changing plate configurations (e.g. [32]) as well as interactions with the convective mantle. Hot, weak continental lithosphere will respond to relatively small loads by deforming. Changes in the load regime will lead to further changes in the deformation with the isostatic response modulating the nature and efficiency of the associated surface processes. In contrast, cold strong continental lithosphere will be able to withstand much greater tectonic loads without appreciable deformation. The simple parameterisation of crustal HPE distributions defined here provides a framework for understanding how crustal-scale deformation associated with tectonic loading may effect long-term changes in thermal regimes. Our models show that deformation, when linked with a surface process through an isostatic response, is capable of effecting significant change in heat production parameters. Thus, such tectonic reworking provides an additional mechanism for the ordering of HPEs over long time scales, and particularly in intraplate settings. However, the greatest response is seen in crust that is already strongly differentiated in the HPEs, concurring with the notion that primary differentiation is due largely to magmatic processes attendant on primary crustal growth.

In summary, our analysis has shown that the likelihood of deformation affecting the continental lithosphere is dependent on the thermal regime and consequently on both the abundance and distribution of HPEs. This result is best summarised in Fig. 9b, which shows the way in which the HPE distribution influences the rate of deformation of a Brace–Goetze lithosphere in response to an imposed tectonic load. This diagram, which can be considered as representing the probability of deformation in response to an imposed tectonic load, shows that for a given complement of HPEs (i.e. constant q_c), relatively undifferentiated crust (i.e. high h) is very much weaker than differentiated crust (i.e. low h). An important consequence of the inherent weakness of high h–q_c configurations is their susceptibility to tectonic reworking. Even though high h distributions are expected to show only relatively modest changes in h–q_c parameters per unit increment of deformation, this susceptibility implies that tectonic loading will result in very much greater strains and thus may, in the long term, effect significant redistribution of HPEs, provided there is some primary differentiation. Indeed, only once reworking has effectively removed much of the heat production from the crust, or moved it to shallow levels, will such lithosphere be mechanically stabilised.

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Appendix. Model assumptions

The models designed to illustrate the effect of deformation, deposition and erosion on the HPE parameters h and q_c rely on the following assumptions:

1. Deformation and surface processes are linked through an isostatic response that tends, in the long term, to restore the surface elevation to some constant value (Fig. 6). Thus, when deformation results in crustal thickening the isostatic response of surface uplift induces erosion, while crustal thinning deformation is coupled with subsidence and sedimentation (i.e. basin formation). By 'long term' we imply a time sufficient to dissipate the thermal transients accruing from the act of change itself (i.e. time intervals greater than the characteristic conductive response time of the lithosphere, >50 million years). We treat the lithosphere as a 1-D column. This assumption requires application of a local isostatic balance, appropriate to first-order continental lithospheric dynamics, for processes that operate on horizontal scales greater than the thickness of the lithosphere (i.e. greater than several hundred kilometres).
2. In order to preserve surface elevation, several assumptions are made about the densities of the various components of the crust. First, we assume that there is no significant density difference between the upper and lower crust. The important implication of this assumption, for all the models presented here, is that following crustal thickening, the amount of erosion required to restore the original surface elevation will also, to a first approximation, restore the original crustal thickness. Second, we assume that the basin filling sediments are significantly less dense than the crust. We express this density in terms of a ratio of the differences in the density between the crust (ρ_c) and the basin-fill (ρ_s) and the mantle (ρ_m) and the crust:

$$\rho' = \frac{\rho_c - \rho_s}{\rho_m - \rho_c} \approx 0.66 \tag{A1}$$

For $\rho' = 0.66$, the crust shows significant long-term thinning following extension, equivalent to 2/3 the thickness of the basin. In terms of the changes in the compositional structure, the subsidence scales as:

$$z_s = \frac{z_c(1-1/\beta)}{1-\rho'} \tag{A2}$$

3. Deformation is assumed to operate homogeneously in the crustal column. Rather different changes to the HPE distribution may result when large-scale discontinuities in the deformation pattern are allowed, such as crustal-scale thrust faulting. Clearly, such discontinuities are an important component of deformation in the continental crust when large strains accumulate. They are less likely to be important when finite strains are small, as is typical of intracontinental (or intraplate) deformation.
4. In modelling the effects of sedimentation we assume a typical upper crustal heat production of 1.5 μW m^{-3} for the basin fill (e.g. [20]).

A summary of the parameters used in the text and figures is given in Table 1. The rheological model used in the Brace–Goetze lithosphere is described in Table 2.

Definition of h and q_c for various analytic heat source distributions

It is commonly assumed that the heat production within the crust can be described by some

Table 1
Parameters used in text and figures

Symbol	Description (unit)	Default value
k	Thermal conductivity (W m^{-1} K^{-1})	3 W m^{-1} K^{-1}
q_s	Measured surface heat flow (mW m^{-2})	
H_s	Characteristic surface heat production value (μW m^{-3})	
h_r	Estimate for vertical length scale of heat production derived from the slope of the regression of surface heat flow (q_s)–heat production (H_s) data (km)	
q_r	Estimate for the reduced heat flow (mantle and lower crustal contribution) derived from the intercept of the regression of surface heat flow (q_s)–heat production (H_s) data (mW m^{-2})	
q_m	Mantle heat flow (mW m^{-2})	25 mW m^{-2}
h	Characteristic length scale of HPE distribution formulated with no assumptions about the form of the HPE distribution (km)	
q_c	Depth-integrated heat production (mW m^{-2})	
z_c	Thickness of crust (km)	35 km
h_x	Wavelength of horizontal variation in the characteristic upper crustal heat production (km)	
T_{qc}	Component of the lithospheric geotherm due to radiogenic sources (mW m^{-2})	
T'_{qc}	Maximum value of T_{qc}, occurring at the depth at which heat production becomes negligible (assumed here to be the Moho)	
T_{qm}	Component of the lithospheric geotherm due to the heat flowing from the deeper mantle (mW m^{-2})	
ΔT_z	Temperature effect due to changes in Moho depth (i.e. Moho shallowing due to lithospheric extension)	
ΔT_{Moho}	long-term change in Moho temperatures; equal to the change in temperature due to changes in Moho depth, as well as any change due to the redistribution of crustal heat sources	

quasi-analytic function of depth. Although the use of such analytic functions clearly defies our geological experience (e.g. Fig. 2), there is some pedagogical value in admitting simple analytical expressions for the steady-state temperature field. Assuming that there is no lateral variation in heat production and that the thermal conductivity is constant throughout the crust (and independent of temperature), the steady-state temperature field for the heat production distribution that shows inverse exponential dependence on depth (the exponential model) is given by:

$$T_z = -\frac{q_m z}{k} + \frac{H_s h_r^2 (1-\exp(-z/h_r))}{k} \quad \text{(A3)}$$

where q_m is the reduced heat flow that applies at deep levels within the lithosphere, beneath all significant heat production.

Another model for the vertical distribution of heat production comprises a layer extending to depth h_r with constant heat production, H_s, beneath which heat production is negligible. In this homogeneous model the temperature field in the heat-producing layer is:

$$T_z = -\frac{q_m z}{k} + \frac{H_s z(h_r - z/2)}{k} \quad \text{(A4)}$$

The first term on the right hand side of Eqs. A3 and A4 represents the component of the temper-

Table 2
Rheological parameters used in the mechanical models

	Model rock type	Flow parameter, A (MPa^{-n} s^{-1})	Power index, n	Activation energy, Q (kJmol^{-1})	Reference
Crust	Adirondack granulite	7.93×10^{-3}	3.1	243	[34]
Upper mantle	Aheim dunite	398.11	4.5	535	[35]

ature field due to the heat flow from beneath the heat producing parts of the lithosphere (as indicated by dashed line in Fig. 5b). The second term on the right hand side of the equations represents the contribution due to heat sources in the crust (e.g. the shaded area in Fig. 5b). This second term can be used to define the quantity T_{qc}, the temperature contribution due to crustal heat production. T_{qc} reaches its maximum value, T'_{qc}, at the depth at which heat production becomes negligible (i.e. at the base of the heat-producing layer; Fig. 5c). The appropriate expressions for T'_{qc} for the exponential and homogeneous models are, respectively:

$$T'_{qc} = \frac{H_s h_r^2 (1-\exp(-z_c/h_r))}{k} \tag{A5}$$

and

$$T'_{qc} = \frac{H_s h_r^2}{2k} \tag{A6}$$

In Eq. A5, z_c is the depth at which the exponential HPE distribution is terminated, which we approximate as the Moho. Note that, provided that $z_c > h_r$, Eq. A5 reduces to:

$$T'_{qc} \approx \frac{H_s h_r^2}{k} \tag{A7}$$

With regard to the homogeneous and exponential HPE distribution models (Eqs. A3 and A4) outlined above:

$$q_c = h_r H_s \text{ for } z_c \gg h_r \tag{A8}$$

Consequently, h can be expressed in terms of h_r as:

$$h = h_r(1-\exp(-z_c/h_r)) \tag{A9}$$

For $z_c \gg h_r$ Eq. A9 reduces to:

$$h = h_r \tag{A10}$$

For the homogeneous HPE distribution model:

$$h = h_r/2 \tag{A11}$$

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