3

Thermo-mechanical controls on heat production distributions and the long-term evolution of the continents

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3.1 Introduction

The heterogeneous nature of continental geology is testimony to the complex superposition of tectonic processes associated with the growth, differentiation, and reactivation of the continental lithosphere. Despite this heterogeneity, geophysical data sets provide compelling evidence for at least two levels of geochemical ordering, both of which impact on the long-term strength of the continents.The first-order layering is evident in the density contrast across the crust–mantle boundary, or Moho, which typically lies at depths of 30–40 km. The Moho is associated with a dramatic change in mineralogy. Consequently, the depth of the Moho exerts a profound control on the strength of the lithosphere (e.g., Brace & Kholstedt, 1980; Sonder & England, 1986; Molnar, 1989; Ranalli, 1997).

A more subtle level of geochemical ordering is evident in terms of the distribution of *heat producing elements* (HPEs). On geological time scales there are only three elements, and in particular four isotopes, ^{238}U, ^{235}U, ^{232}Th and ^{40}K, that occur in sufficient abundance to contribute to the lithospheric thermal budget; these elements are referred to as the HPEs. The great proportion of the HPEs are contained within felsic igneous rocks such as granites and granodiorites, which have average volumetric heat production of 2.5 μW m^{-3} and 1.5 μW m^{-3}, respectively (Haenel *et al.*, 1988). These values are significantly higher than mafic igneous rocks and most sedimentary rocks, and suggest that the distribution of HPEs is broadly related to the distribution of lithotypes. At the crustal scale, an ordering of the HPEs is evident from an analysis of surface heat flow and heat production data. These data sets show that in most stable continental regions the lithospheric complement of HPEs is largely confined to the upper 10–15 km of the crust and is responsible for about one-half to two-thirds of the characteristic surface heat

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flow of the continents (i.e., about 30–40 mW m^{-2}, e.g., McLennan & Taylor, 1996).

The distribution of HPEs exerts a crucial influence on continental thermal regimes. Because lithospheric rheology is extremely sensitive to the thermal regime (e.g., Brace & Kholstedt, 1980; Ranalli, 1997), the distribution of HPEs must also impact on the strength of the lithosphere. Consequently, processes that redistribute the HPEs should be expected to affect the long-term strength of the lithosphere. Conventional wisdom holds that the absolute abundance of the HPEs in the continents and the concentration of HPEs in the upper half of the continental crust is a primary feature of magmatic-related crustal growth processes. In particular, the lithophile nature of the HPEs implies that they will be progressively differentiated during the partial melting processes attendant with primary crustal growth. However, there is only limited understanding of why such processes should lead to the characteristic abundance and distribution of HPEs inferred from surface heat flow and heat production measurements. As noted by Oxburgh (1980), "... the problem of crustal heat production and its distribution and re-distribution by physical and chemical processes during crustal evolution is of fundamental importance and is at present little understood."

The primary objective of this chapter is to show that minor tectonic reworking of the continental lithosphere may modify the distribution of HPEs. In particular, we focus on the role of isostatic coupling between deformation and surface processes (i.e., erosion and sedimentation). The temperature sensitivity of lithospheric rheology requires that tectonic reworking of the continental lithosphere is sensitive to the thermal state of the lithosphere. By making the link between tectonic reworking and HPE distributions we will illustrate a potentially profound feedback relation that might serve as a long-term control on the distribution of HPEs in the continents. Although the ideas and models developed in this chapter contribute to our understanding of lithospheric evolution in a very general sense, we emphasize that our one-dimensional pure shear models are particularly relevant to the evolution of continental interiors, and the role played by intracontinental, rather than plate-margin deformation. We begin by outlining a new parameterization of crustal HPE distributions that permits a simple graphical scheme for evaluating some important thermal and mechanical consequences of HPE distributions.

3.2 Length scales and heat production distributions

Surface heat flow and heat production measurements imply that the upper 10–15 km of the continental crust is significantly enriched in HPEs relative to

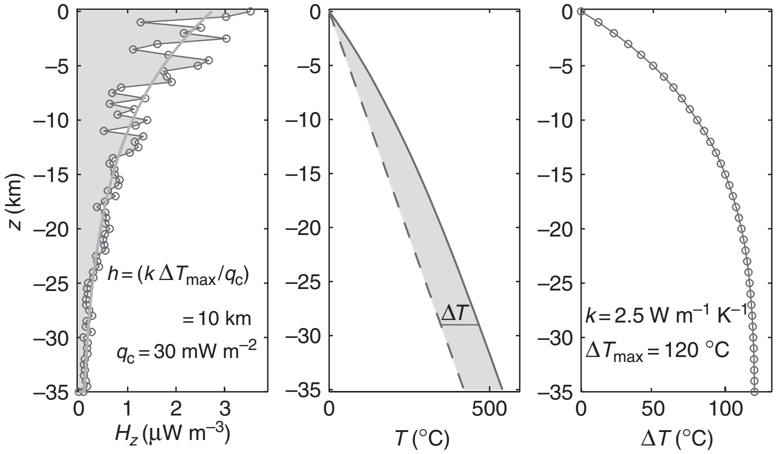

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Figure 3.1. (a) Hypothetical heat production distribution approximating an exponentially decreasing distribution of the form $H(z) = H_{sur}$ (exp(-z/h_r)) with added "noise" designed to simulate the heterogeneity of the continental lithosphere. The model exponential distribution is shown by the solid line. (b) The temperature field produced by such a distribution illustrating the component due to the reduced heat flow (the dashed line), and the component due to heat production in the lithosphere (the shaded area). (c) The temperature deviation, ΔT, is defined as the component of the temperature field due to heat production and corresponds to the shaded region in Fig. 3.1(b). Note that the maximum value, ΔT_{max}, is attained at the base of the heat-producing layer.

the deeper crust and mantle. The general form of the HPE distributions in the crust, and particularly the way HPE concentrations vary with depth, has been the subject of considerable interest. The heterogeneous composition of the continental crust implies that any distribution will be complex and discontinuous (Fig. 3.1a). However, for a number of purposes it is desirable to approximate this distribution with simple analytical models. Since the classic studies of Birch *et al.* (1968), Roy *et al.* (1968) and Lachenbruch (1968), heat-production distributions have been characterized in terms of both a characteristic volumetric heat production value (H, measured in units of W m^{-3}) and a characteristic length-scale (h, measured in kilometers). The most celebrated of these heat-production distributions is the *exponential* model (Lachenbruch, 1968, 1970) where the heat production at any depth z (defined to be negative downwards) is given by

$$H(z) = H_{sur} \exp\left(\frac{z}{h_r}\right). \tag{3.1}$$

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Table 3.1. *Parameters used in text and figures with default values applying to calculations unless otherwise stated.*

Symbol	Description	Units	Default value
z	Depth within the crust. z is defined to be negative downwards	km	
$H(z)$	Distribution of heat sources with depth (z)		
H_{sur}	Surface heat production value	(μ)W m^{-3}	
H_{sed}	Sediment heat production value	(μW m^{-3})	
h_r	Characteristic length scale of the exponential heat source distribution. Originally defined from the linear relationship between surface heat flow and heat production data (e.g., Lachenbruch, 1968)	km	15
H	General formulation for the characteristic length scale of the heat source distribution (this contribution). Defined in text in Eq. (3.7)	km	
K	Thermal conductivity	W m^{-1} K^{-1}	3
q_r	Reduced heat flow which appliesat deep levels within the lithosphere, beneath all significant heat production	(m) W m^{-2}	
q_c	Depth-integrated heat production. q_c is always positive (i.e., heat flowing out of the crust)	(m) W m^{-2}	
T_{qc}	Temperature contribution due to crustal heat sources	°C	
T'_{qc}	Maximum temperature contribution due to crustal heat sources. T'_{qc} is reached at the depth at which heat production becomes negligible	°C	
z_c	Crustal thickness	km	35–40 km
z_s	Amount of subsidence during basin formation	km	
β	Ratio of the thickness of the crust prior to and following deformation; $\beta < 1$ for shortening, $\beta > 1$ for extension	dimensionless	
ρ_c	Density of the crust	kg m^{-3}	2780
ρ_m	Density of the mantle	kg m^{-3}	3300
ρ_{sed}	Density of the basin filling sediments	kg m^{-3}	2400

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lower case h (in italics)

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Table 3.1 lists definitions of all parameters, their units of measurement and default values used in calculations reported here. In Eq. (3.1), the characteristic length-scale h_r is defined as the depth at which heat production falls to 1/e of the characteristic surface value, H_{sur} (i.e., h_r is the e-fold length of the exponential distribution).

The use of such idealized, largely pedagogical models is motivated in part by the desire to obtain simple analytical expressions for steady-state temperature

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distributions. For example, 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 exponential model is given by

$$T_z = -\frac{q_r z}{k} + \frac{H_s h_r^2 (1 - \exp(-z/h_r))}{k} \tag{3.2}$$

Hsur

In Eq. (3.2), q_r is the reduced heat flow that applies at deep levels within the lithosphere, beneath all significant heat production. One interpretation of q_r is that it represents the heat flux provided to the base of the lithosphere by convective processes in the deeper mantle. However, as noted by Jaupart (1983) and Vasseur and Singh (1986), the value of q_r deduced from regression of surface heat-flow–heat-production data will invariably overestimate the non-radiogenic component of surface heat-flow and underestimate the length scale of the crustal heat production. This is because the lateral transfer of heat is significant in regions where there are significant lateral contrasts in heat production. In such cases both the slope and intercept of any fit of surface heat-flow with surface heat production are dependent on the horizontal variation in heat-production parameters, as well as the vertical distribution.

An alternative model for the vertical distribution of heat production is a layer extending to depth h_r with constant heat production, H_{sed}, beneath which heat production is negligible. In this *homogeneous* model the temperature field in the heat-producing layer is

$$T_z = -\frac{q_r z}{k} + \frac{H_{sur} z (h_r - z/2)}{k} \tag{3.3}$$

Note that in Eqs. (3.2) and (3.3), the first term on the right-hand side of the equations represents the component of the temperature field due to the heat flow from beneath the heat-producing parts of the lithosphere (e.g., dashed line in Fig. 3.1(b)). The second term on the right-hand side of the equation represents the contribution due to heat sources in the crust (e.g., the shaded area in Fig. 3.1(b)). We use this second term to define the quantity T_{qc}, the temperature contribution due to crustal heat production. T_{qc} reaches its maximum value (T'_{q_c}) at the depth at which heat production becomes negligible (i.e., at the base of the heat-producing layer; Fig. 3.1(c). The appropriate expressions for T'_{q_c} for the exponential and homogeneous models are, respectively,

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$$T'_{qc} = \frac{H_{sur} h_r^2 (1 - \exp(-z_c/h_r))}{k} \tag{3.4}$$

and

$$T'_{qc} = \frac{H_{\text{sur}} h_{\text{r}}^2}{2k} \tag{3.5}$$

In Eq. (3.4), z_{c} is the depth at which the exponential heat production distribution is terminated, which we approximate as the Moho. Note that provided that $z_{\text{c}} > h_{\text{r}}$, then Eq. (3.4) reduces to

$$T'_{qc} \approx \frac{H_{\text{sur}} h_{\text{r}}^2}{k}. \tag{3.6}$$

One attribute of these two models is that the length scale h_r may be evaluated from the regression of surface-heat production and heat-flow data (with the proviso outlined above). Typically, such regressions yield a value of $\sim$10 km (e.g., Lachenbruch, 1968). An alternative and more general formulation of the characteristic length-scale (h), that makes no explicit assumption about the form of the heat-production distribution, may be made in terms of the heat source distribution

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delete "-" between surface and heat; insert "-" between heat and production

$$h = \frac{1}{q_{\text{c}}} \int_0^{z_{\text{c}}} (H(z))\mathrm{d}z \tag{3.7}$$

where q_{c} is the depth integrated heat production

$$q_{\text{c}} = \int_0^{z_{\text{c}}} H(z)\mathrm{d}z. \tag{3.8}$$

One attribute of this definition is that it leads to a very simple relation between T'_{qc}, q_{c} and h, as follows

$$T'_{qc} = \frac{q_{\text{c}} h}{k}. \tag{3.9}$$

Equation (3.9) implies that very different HPE distributions characterized by similar values of h (cf., Fig. 3.2a–d) and similar integrated abundances of HPEs (i.e., similar q_{c}) will contribute the same temperature deviation at all depths beneath the heat-producing parts of the lithosphere. This parameterization also highlights the fact that the temperature deviation within the heat-producing parts of the lithosphere due to the crustal heat production, $T_{q_{\text{c}}}$, is very sensitive to the depth at which the heat production is located (cf., Fig. 3.2a, b, e, f). We note that for both the homogeneous and exponential heat production models outlined above

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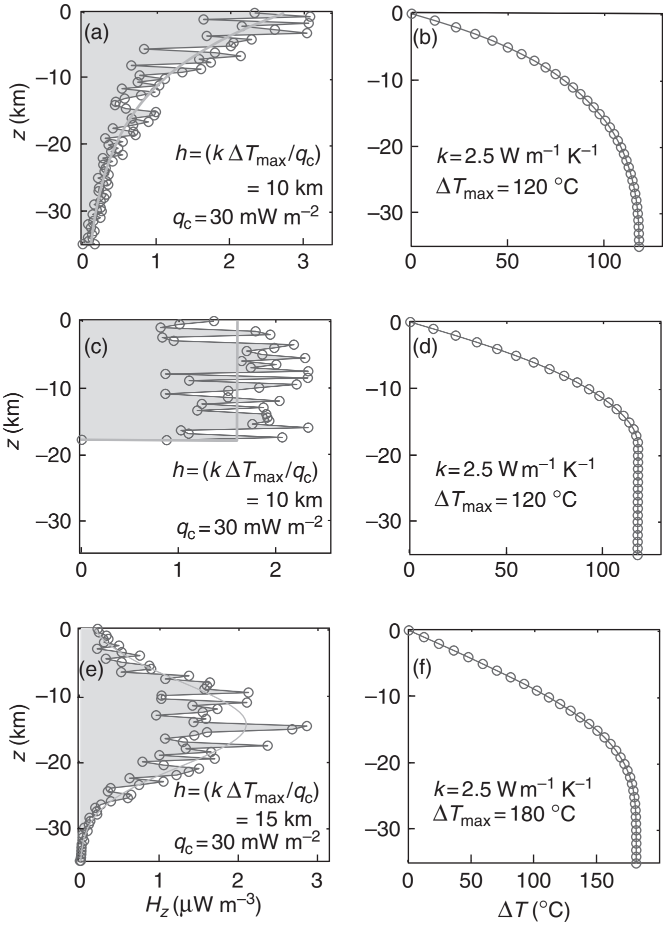

Figure 3.2. Illustration of various heat-production models and their associated thermal effects, used in this chapter. Each model is based on an analytical expression with associated geological noise. The analytic expressions are respectively: (a and b) the exponential model; (c and d) the homogeneous model, and (e and f) show that, for a given complement of HPEs, an increase in the depth of burial (as represented by the parameter h) results in greater perturbation of lower crustal temperatures.

$$q_c = h_r H_{sur} \tag{3.10}$$

allowing h to be expressed in terms of h_r. Consequently, for the exponential model

$$h = h_r(1 - \exp(-z_c/h_r)) \tag{3.11}$$

that reduces to

$$h = h_r \text{ for } z_c \gg h_r \tag{3.12}$$

and, for the homogeneous model,

$$h = h_r/2 \tag{3.13}$$

The analysis presented above emphasizes the sensitivity of crustal thermal regimes not only to the total heat contributed by crustal sources (i.e., q_c) but also to its depth distribution as represented by the parameter h. An important consequence of this is that processes that act to change the distribution of the HPEs (i.e., q_c and/or h) are capable of inducing significant long-term changes in deep crustal thermal regimes, independently of any change in the heat flux from the deeper mantle. Moreover, the sensitivity of crustal thermal regimes to these distribution parameters raises the possibility of potentially profound feedback relations if the processes that modify and control the distribution of HPEs are themselves sensitive to the thermal structure of the lithosphere. In the remainder of this chapter we explore the connection between tectonic processes that modify HPE distributions and the thermal and mechanical state of the lithosphere.

3.3 Tectonic modification of HPE distributions

Tectonic processes associated with reworking of the continental lithosphere (e.g., magmatism, deformation, metamorphism, sedimentation and erosion) effect changes in the distribution of HPEs in the crust, and thus must impact on its long-term thermal and mechanical state. Because the long-term changes in the thermal structure of the lithosphere may be directly related to changes in the distribution parameters, the h–q_c plane may be used to illustrate the thermal consequences of such tectonic processes. Figure 3.3 shows qualitatively how various processes attendant with crustal reworking are likely to change the HPE distribution parameters. For example, the generation of granites by crustal melting provides a mode for redistributing HPEs upwards in the crust with a corresponding reduction in h without reducing q_c (Fig. 3.3a). Expressed in terms of relative changes in the heat-production parameters, infracrustal magmatism is characterized by $\Delta q_c = 0$ and $\Delta h < 0$.

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Deformation results in changes in both the amount and the depth of crustal heat sources. Crustal shortening $\beta < 1$ 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, and at the same time buries it to deeper crustal levels (i.e., $\Delta q_c > 0$ and $\Delta\ h > 0$). In contrast, crustal extension (where $\beta > 1$) tends to attenuate the pre-existing heat production and move it to shallower levels (i.e., $\beta q_c < 0$ and $\Delta h > 0$). Erosion reduces

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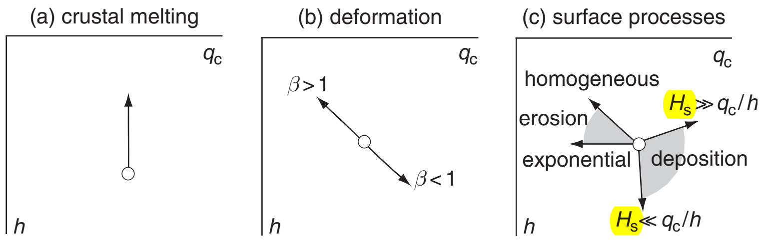

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Figure 3.3. Schematic illustration of the way tectonic processes associated with reworking of continental lithosphere modify the HPE distribution parameters, h and q_c. ~~The generation of granites by crustal melting provides a mode for redistributing HPEs upwards in the crust without reducing q_c (3a). Crustal deformation (3b) 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 (3c). The length-scale for *exponential* distributions is essentially preserved during erosion, 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 (3c). 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.~~

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q_c ($\Delta q_c < 0$), particularly so 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 (Fig. 3.3c). The length-scale for *exponential* distributions is essentially preserved during erosion, while for *homogeneous* distributions the length scale is reduced in direct proportion to the reduction in q_c. The effect of sedimentation depends on the heat production character of the sediments (H_{sed}) relative to the pre-existing crust (Fig. 3.3c). If H_{sed} 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 $H_{sed} \ll q_c/h$, then sedimentation will lead to a significant increase in h for a comparatively minor increase in q_c ($\Delta q_c > 0$ and $\Delta h > 0$).

3.4 Thermal modeling

In the long term, deformation, erosion, and sedimentation are linked through the isostatic response of the lithosphere, with the surface processes tending to sub-figure labels, a, b, c, etc should be more prominent and in brackets (a), (b) ...

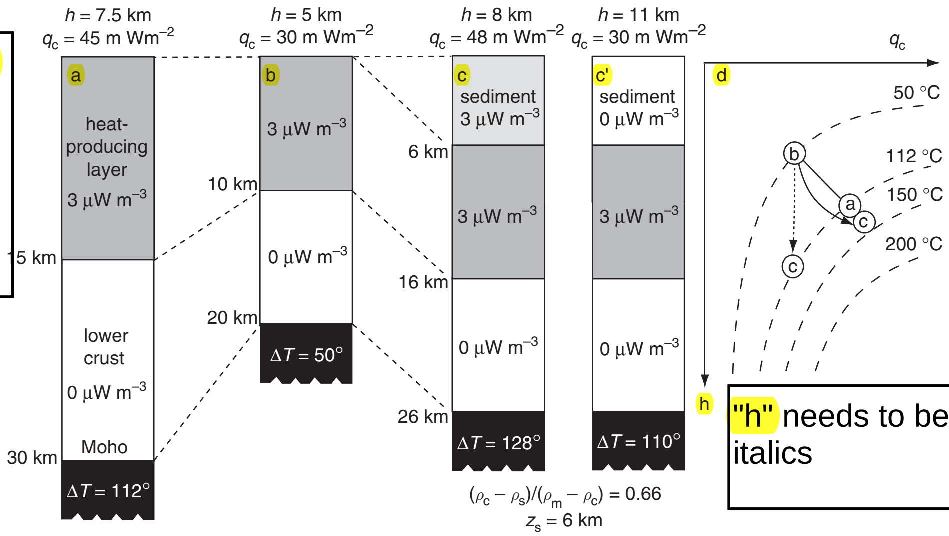

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Figure 3.4. Illustration of the role of deformation coupled with an isostatic response on heat-production distributions in the crust. Given an initial lithosphere (a), stretching (b) results in attenuation of pre-existing heat production, while the surface response of basin formation (c) 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-fillings sediments (with q_c either decreased of increased). The changes in *h* and q_c may be mapped on to the h–q_c plane (d) that may also be used to illustrate the steady-state contribution of crustal radiogenic sources to thermal regimes at deep crustal levels (i.e., ΔT_{max}).

"of" should be "or"

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 this section we illustrate quantitatively the way in which deformation, erosion, and sedimentation modify various heat-source distributions (including the *exponential* and *homogeneous* models described in the previous section). The assumptions used in the development of these models are listed here.

First, we link deformation and surface processes through an isostatic response that tends, in the long term, to restore the surface elevation to some constant value (Fig. 3.4). Thus, when deformation results in crustal thickening, the isostatic response of surface uplift induces erosion, whereas crustal thinning deformation is coupled with subsidence and sedimentation (i.e., basin formation). By long term we imply a sufficient time to dissipate all thermal transients accruing from the act of change itself (i.e., a period greater than the thermal time constant of the lithosphere). The treatment of the lithosphere as 1-D vertical column requires application of a local isostatic balance appropriate to the first-order dynamics of continental lithosphere for processes that operate on spatial scales greater than the thickness of the lithosphere (i.e., greater than several hundred kilometers).

Second, in order to preserve surface elevation, we make several assumptions about the densities of the various components of the crust. 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. Further, we also assume that the basin-filling sediments are significantly less dense than the crust. We express this density in terms of the density of the basin filling sediments (ρ_{sed}), crust (ρ_{c}) and mantle (ρ_{m}), as follows

$$\rho' = \frac{\rho_{\mathrm{c}} - \rho_{\mathrm{sed}}}{\rho_{\mathrm{m}} - \rho_{\mathrm{c}}} \approx 0.66. \tag{3.14}$$

Consequently, the crust shows significant long-term thinning following extension. In terms of the changes in the compositional structure, the subsidence scales as

zsub not defined anywhere, change to "zs" (subscript s)

$$z_{\mathrm{sub}} = \frac{z_{\mathrm{c}}(1 - 1/\beta)}{1 - \rho'}. \tag{3.15}$$

Third, in our models, 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 important components 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 intra-continental (or intra-plate) deformation associated with reworking of continental interiors. In such settings, deformation rarely results in the exposure of deep crustal rocks. As discussed ~~later,~~ we believe our analysis is particularly relevant to the role played by intracontinental deformation.

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Lastly, in modeling the effects of sedimentation, we assume a typical upper crustal heat production of 1.5 μW m^{-3} for the basin fill (e.g., McLennan & Taylor, 1996). The results of this modeling are summarized in Figs. 3.5 and 3.6. The initial HPE distributions used in the calculations correspond to the *exponential* (Fig. 3.5) and *homogeneous* models (Fig. 3.6) described in the previous section.

Figures 3.5a, b and 3.6a, b show the effect of a 10% ~~crustal stretching~~ (i.e., $\beta = 0.9$ for crustal thickening and $\beta = 1.1$ for crustal thinning). Figures 3.5c, d

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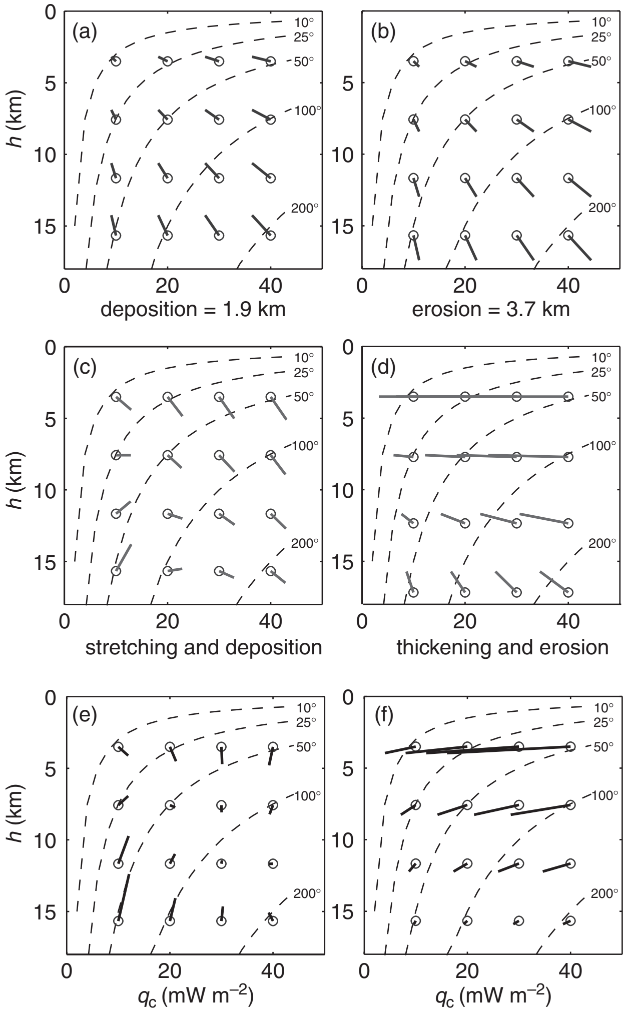

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Figure 3.5. (a), (b) illustrate the effects of a 10% attenuation ($\beta = 1.1$) and a 10% thickening ($\beta = 0.9$ on the distribution parameters h and q_c for ~~an~~ initial *exponential* distribution (see Fig. 3.2a); ~~the parameter β is used in its conventional sense when applied to a 1-D lithospheric column (i.e., as a measure of the deformation given by the ratio of the initial to deformed crustal thickness);~~ (c), (d) show, respectively, the role of deposition and erosion sufficient to restore the original value surface elevation, and (e), (f) show, respectively, the combined effect of stretching and basin formation, and thickening and erosion. Each figure is contoured for ΔT_{max} (see Fig. 3.1).

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"distribution" should be "distributions of varying h and qc". [Lower case h and qc in italics]

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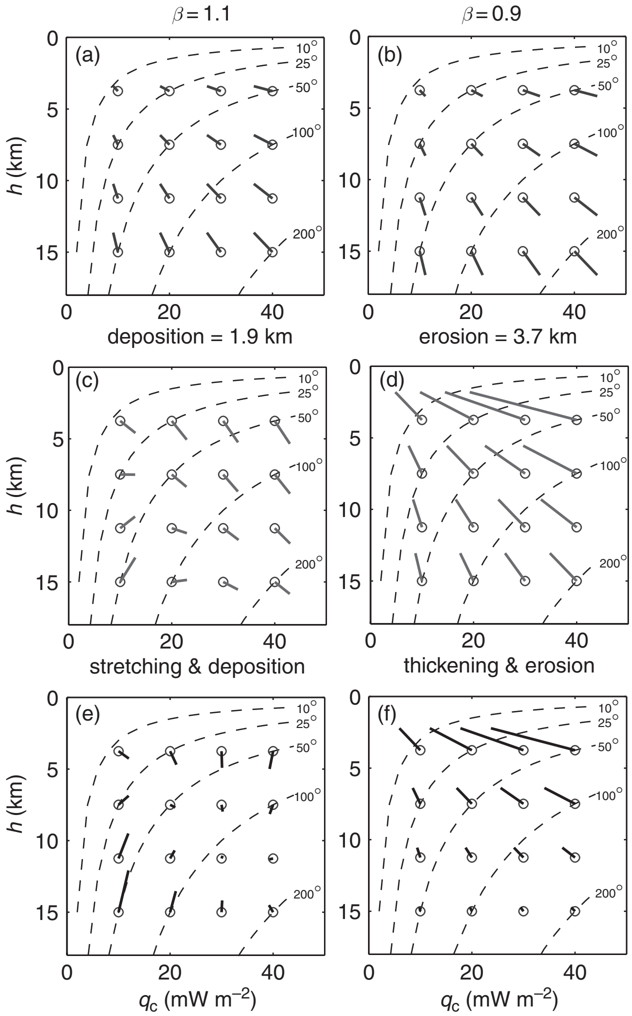

Figure 3.6. As for Fig. 3.5, but with an initial *homogeneous* HPE distribution (i.e., Fig. 3.2c).

and 3.6c, d show the effects of an associated surface response (erosion or sedimentation) that, as noted above, restores the original surface elevation assuming local isostatic balance. Figures 3.5e, f and 3.6e, f show the combined effects of the deformation and the linked surface response. Figures 3.5 and 3.6 illustrate a number of important features concerning the way in which heat-production parameters evolve as a consequence of deformation and the surface processes, which are summarized below.

The coupling of crustal thickening (i.e., $\beta < 1$) and erosion provides an effective means for modifying HPE parameters, particularly where crustal insert "(e.g., 3.5f and 3.6f)" after differentiated

heat-production is already significantly differentiated. Whereas crustal thickening tends to increase q_c, erosion will reduce it. If the existing heat production is already strongly concentrated near the surface (i.e., small h) the loss of heat production during erosion is much greater than the increase in heat production due to thickening (see Fig. 3.4b), resulting in a dramatic decrease in q_c. This reflects the fact that the coupling of crustal thickening and erosion essentially removes 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 distributions approximating the homogeneous model, h will be reduced as a consequence of combined homogeneous thickening and erosion (Fig. 3.6).

For a distribution which is approximately exponential, then providing $h \ll z_c$, h will increase by the factor $1/\beta$, because erosion preserves the exponential length scale established by the deformation (e.g., Lachenbruch, 1968). Note that 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. Figures 3.7c, d show that the long-term changes in the thermal structure of the deep crust resulting from thickening and erosion may be significant. For typical crustal heat production parameters, a 10% thickening followed 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, it will be expected to experience all of this cooling.

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For crustal thinning deformations ($\beta > 1$) the long-term thermal consequences are critically dependant on whether basin formation occurs and, if so, the heat production parameters of the basin-filling sediments. As noted above, in calculating Figs. 3.4 and 3.5 we have assumed that the basin fill has heat production of 1.5 $\mu W\,m^{-3}$ (e.g., McLennan & Taylor, 1996). 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} (e.g., Sandiford, 1999). As shown by Figs. 3.7a, b, increases in T'_{qc} are expected when h is less than about 10 km. For initial configurations characterized by high h, the combination of stretching and basin filling may lead to long-term reductions in T'_{qc} of up to ~10 °C for an initial 10% stretch. However, note that 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

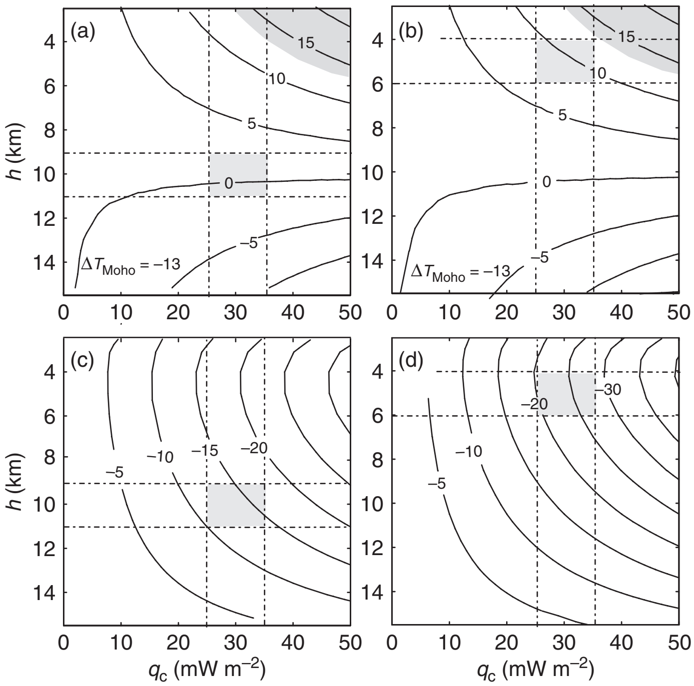

Figure 3.7. The h–q_c plane contoured for changes in ΔT_{max} following a 10% stretch and an associated surface response that restores the initial surface elevation. (a) and (b) show the effects of a 10% stretch with basin formation for an initial *exponential* and *homogeneous* HPE distribution, respectively. (c) and (d) show the effects of a 10% thickening followed by erosion, for an initial *exponential* and *homogeneous* HPE distribution, respectively. 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 ~13 °C for the parameters assumed here). The change in Moho temperature is given by $T'_{qc} - \Delta T_{Moho}$. The gray box shows the relatively narrow range in h–q_c parameters that characterize stable modern continental crust. This data is summarized from a compilation by Taylor and McLennan (1985) (modified from Vitorello and Pollack, 1980, and Morgan, 1984).

delete "assumed" replace with "used"

of T'_{qc} for a given stretch (see more detailed discussion in Sandiford, 1999). Because the Moho will not be returned to its original depths in the long term for typical basin-fill densities, the estimates for T'_{qc} as shown in Figs. 3.7a, b cannot be directly equated with long-term changes in the Moho temperature. The amount of additional cooling of the Moho due to its long-term shallowing depends on the thermal gradient at the Moho level. For a thermal gradient of 10 °C km^{-1}, this long-term shallowing of the Moho due to a stretch of 10% leads to Moho cooling (ΔT_{Moho}) of ~13 °C, with the change in Moho temperature given by T'_{qc}–ΔT_{Moho}. For the parameters used in our calculations, only extremely high q_c, low h initial configurations lead to long-term Moho heating following crustal stretching (i.e., the shaded region in the upper right of Fig. 3.7a, b).

The long-term evolution of many continental interiors is punctuated by the accumulation of small to very small magnitude deformation events, associated with minor reworking and reactivation. Such events involve both crustal shortening and extension, and are typically accompanied by an associated surface response that tends, in the long term, to keep the crustal thickness at the continental average of ~35 km. Figures 3.5–3.7 show that such small deformation events may effect significant redistribution of the HPEs, raising the possibility that tectonic reworking of continental interiors plays an important role in the geochemical ordering of the continental lithosphere.

3.5 Mechanical consequences of the redistribution of HPEs

In the previous section we have shown that significant long-term thermal consequences may accrue from the redistribution of crustal heat sources that accompanies deformation when coupled with associated surface processes, even for relatively minor deformation events that are likely to typify continental interiors. The strong temperature-dependence of lithospheric rheology suggests that deformation is likely to be localized in continental regions characterized by elevated thermal regimes, such as would be associated with high q_c and/or h values. If all other factors, including composition and strain rate, are equal, then continental crust with low q_c and h values would be expected to be significantly stronger than crust with higher q_c and h values. Therefore, such crust would be less likely to experience deformation in response to an applied tectonic load. In this section we attempt to quantify how variations in q_c and h affect the strength of the lithosphere. We use a lithospheric rheology in which the lithosphere is assumed to deform by a combination of frictional sliding and ductile creep (e.g., Brace & Kholstedt, 1980; Table 3.2). Throughout the remainder of this section we refer to this model as the *Brace–Goetze* rheology.

The application of the *strength-envelope* approach implicit in this model has become widespread in addressing 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 (e.g., Paterson, 1987), including: (1) the extrapolation of rheological flow laws from laboratory conditions and time scales to the geological realm; (2) an

Table 3.2. *Rheological parameters used in thermo-mechanical models*

	Model rock type	Flow parameter, A ($\mathrm{MPa}^{-n}\,\mathrm{s}^{-1}$)	Power index, n	Activation energy, Q ($\mathrm{kJ\ mol}^{-1}$)	Reference
Crust	Adirondack granulite	7.93×10^{-3}	3.1	243	Wilks & Carter (1990)
Upper mantle	Aheim Dunite	398.11	4.5	535	Chopra & Paterson (1981)

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 emphasize the role of temperature changes accompanying the various tectonic processes to relative changes in lithospheric strength. Whereas 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 alluded to above will cancel.

"relative" should be in italics

Recognizing that the h–q_c plane may be contoured for thermal structure (e.g., Figs. 3.5–3.7) also allows it to be contoured for lithospheric strength (Fig. 3.8; ~~Table 3.2).~~ Figure 3.8a shows variations in the strength of a *Brace–Goetze* lithosphere forced to deform at a specified strain rate, whereas Fig. 3.8b shows variations in the effective strain rate for the lithosphere when subject to an imposed tectonic load (the rheological parameters assumed in the calculations are summarized in Table 3.2). 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 rates by 1–2 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, to be discussed in the next section, of a feedback relation between deformation processes and HPE distributions in the crust. One way to view Fig. 3.8b is that it expresses the likelihood that a lithosphere will experience significant deformation when subject to an imposed tectonic load.

delete "Table 3.2"

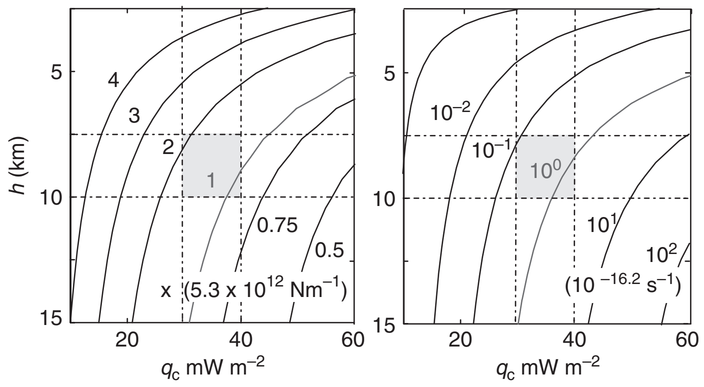

Figure 3.8. (a) h–q_c plane contoured for integrated lithospheric strength. (b) h–q_c plane contoured for rate of deformation subject to an imposed tectonic load. The assumed rheology is that of the *Brace–Goetze* lithosphere in which deformation is governed by a combination of frictional sliding and temperature-dependent creep. The strength parameters are shown normalized against a configuration characterized by $q_c = 45$ mW m^{-2} and $h = 7$ km. The shaded area illustrates that, all other factors being equal, a 25% h and q_c will result in an increase in strength by a factor of 2 or a decrease in strain rate in response to an imposed load by an order of magnitude. Note that the calculations are sensitive to a large number of assumed parameters including thermal conductivity ($k = 3$ W m^{-1} K^{-1}) and mantle heat flux ($q_m = 30$ mW m^{-2}), as well as the material parameters constraining creep and frictional sliding of crustal and mantle rocks.

after "25%", insert "reduction in"

Whereas Fig. 3.8 illustrates the way in which variations in HPE parameters may alter the strength of the lithosphere, it is important to realize 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 (see earlier discussion), will also contribute to variations in lithospheric strength. In the case of basin formation, the cooling associated with shallowing of the Moho tends to increase the strength of the lithosphere, while the changes in HPE parameters usually tend to decrease its strength (e.g., Fig. 3.7a, b). Depending on the initial HPE distribution parameters, basin formation may either result in mild long-term weakening or strengthening of the lithosphere (a detailed discussion is given in Sandiford, 1999).

Figure 3.8 suggests that the mechanical response of the lithosphere to imposed tectonic loads is extremely sensitive to the distribution of HPEs. In order to explore how the distribution of HPEs impacts on the way the lithosphere responds, and the way in which this response might change HPE distributions, we have carried out a set of numerical simulations in which a series of model 1-D *Brace–Goetze* lithospheric columns, differing only in terms of their HPE parameters, are subject to an identical loading history (Figs. 3.9 and 3.10). The load history, as shown in Figs. ~~3.9d~~ and ~~3.10d,~~ is scaled so as to provide a mildly compressional stress regime, as is suggested by the potential energy differences between mid-ocean ridges and normal continental crust (e.g., Coblentz *et al.*, 1994), with a hierarchy of fluctuations of up to $\sim 10^{13}$ $\mathrm{N m^{-1}}$ over a range of timescales up to $\sim$500 Ma. Deformation rates are relatively insignificant ($\varepsilon_{zz} < 10^{-16}\ \mathrm{s^{-1}}$, Figs. ~~3.9d~~ and ~~3.10d)~~ especially when compared to modern plate-margin deformation rates, with the result that strain increments during any given episode are only small (individual stretch increments are < 15%, Figs. 3.9e and 3.10e). The isostatic response to changes in crustal thickness (Figs. 3.9d and 3.10d) either elevates or depresses the surface (e.g., Figs. 3.9a and 3.10a), thereby inducing a surface response that tends to restore the surface to sea level (e.g., Figs. 3.9f and 3.10f) on a characteristic time scale. The three simulations shown in Fig. 3.9 are characterized by an initially *exponential* heat source distribution with HPE parameters, $q_c = 37$, 42 and 27 $\mathrm{mW\,m^{-2}}$ and $h = 13.5$, 11.8, and 11.8 km, respectively (Fig. 3.9b), whereas Fig. 3.10 is characterized by an initially *homogeneous* distribution.

replace with "3.9g" and "3.10g"

replace with "3.9h" and "3.10h"

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 modified significantly by repeated small-scale tectonic reworking of the lithosphere. The simulations show that initial configurations characterized by high q_c (> 35 $\mathrm{mW\,m^{-2}}$) experience significant reductions in q_c during repeated tectonic reworking, even though the scale of the reworking is very minor (note that the amplitude in variations in crustal surface elevation is generally less than 500 m!). For an initial exponential distribution, the erosion-dominated reduction in q_c tends largely to preserve the initial value of h, in accord with the results in Fig. 3.5. Figure 3.10b shows that initial homogenous distributions may show significant h reduction, as well as q_c reduction, on reworking. The reductions in q_c of up to 20% equate to long-term Moho cooling of $\sim$20–40 °C. In contrast, the initial configurations with low q_c and low h (point 3 in Fig. 3.9b) show no significant response to the imposed loads by virtue of their inherent mechanical strength, a minor reduction in q_c is induced as a consequence of the long-term erosional denudation of the initial topography.

after "500m!" still inside brackets, insert "e.g., Fig. 3.9a"

While these models are clearly very sensitive to the input parameters, they do highlight a potentially profound role that tectonic reworking may play in modulating the HPE distributions in the continental crust. Crust initially configured with elevated q_c and/or h will, through the isostatic coupling of the difference in the three lines is not clear, very small dashed line should be made less like the solid line. Same applies to Fig. 3.10

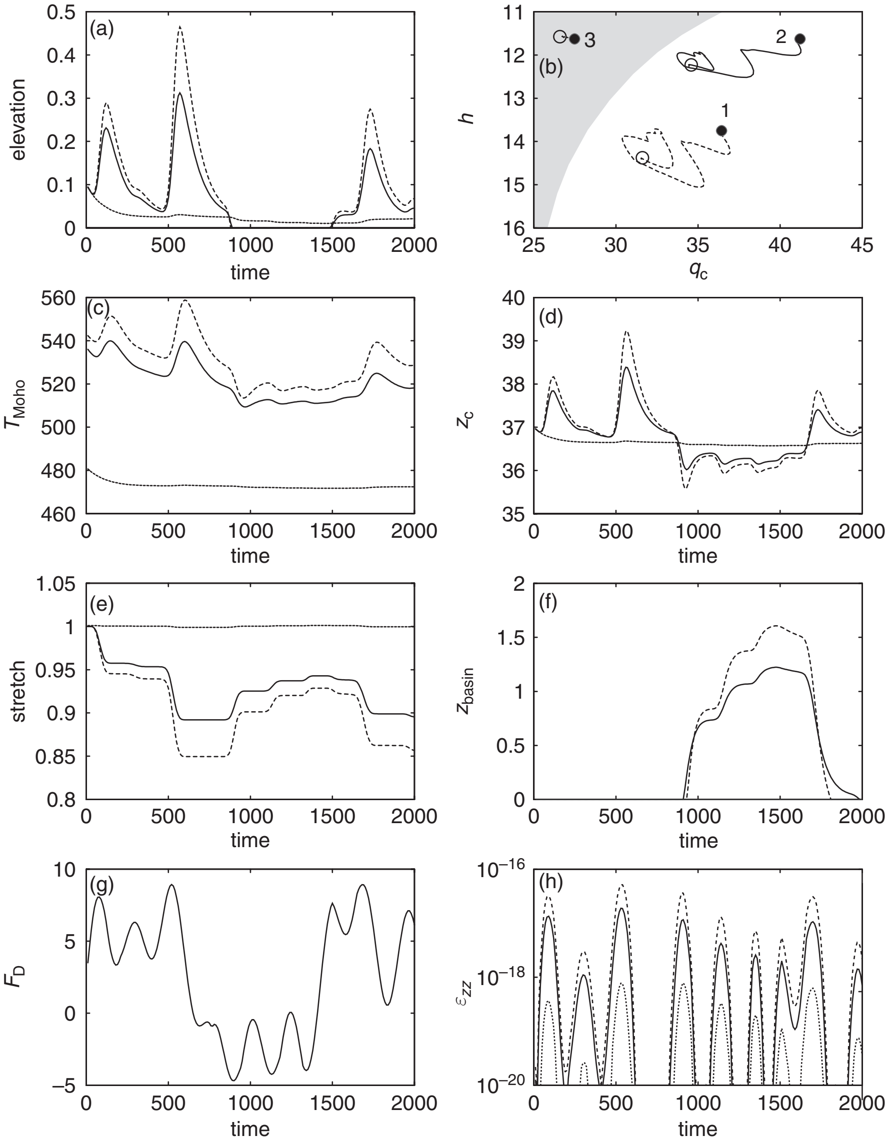

Figure 3.9. Numerical simulations in which 1-D *Brace–Goetze* lithospheric columns are subject to a tectonic load (g). The response of three columns is shown, these columns being identical in all respects apart from the initial HPE distribution parameters (b), which translates to variations in the initial thermal regime (c). The initial HPE distribution is exponential. The tectonic response is measured in terms of vertical strain rate (*h*), as well as incremental stretching history (e). The interaction of deformation and surface response leads to long-term changes in surface elevation (a) and crustal thickness (d). Accumulation of sediments in basins, and its subsequent removal during basin inversion, is shown in (f). The change in HPE distribution parameters is shown in (b). Time is measured in Ma, with the thermal response of the lithosphere (c) calculated using a finite element algorithm. Closed circles show initial HPE parameters while open circles show final HPE parameters. The shaded region represents parameter range that shows stable (or cratonic) behavior throughout the imposed loading history. See text for discussion.

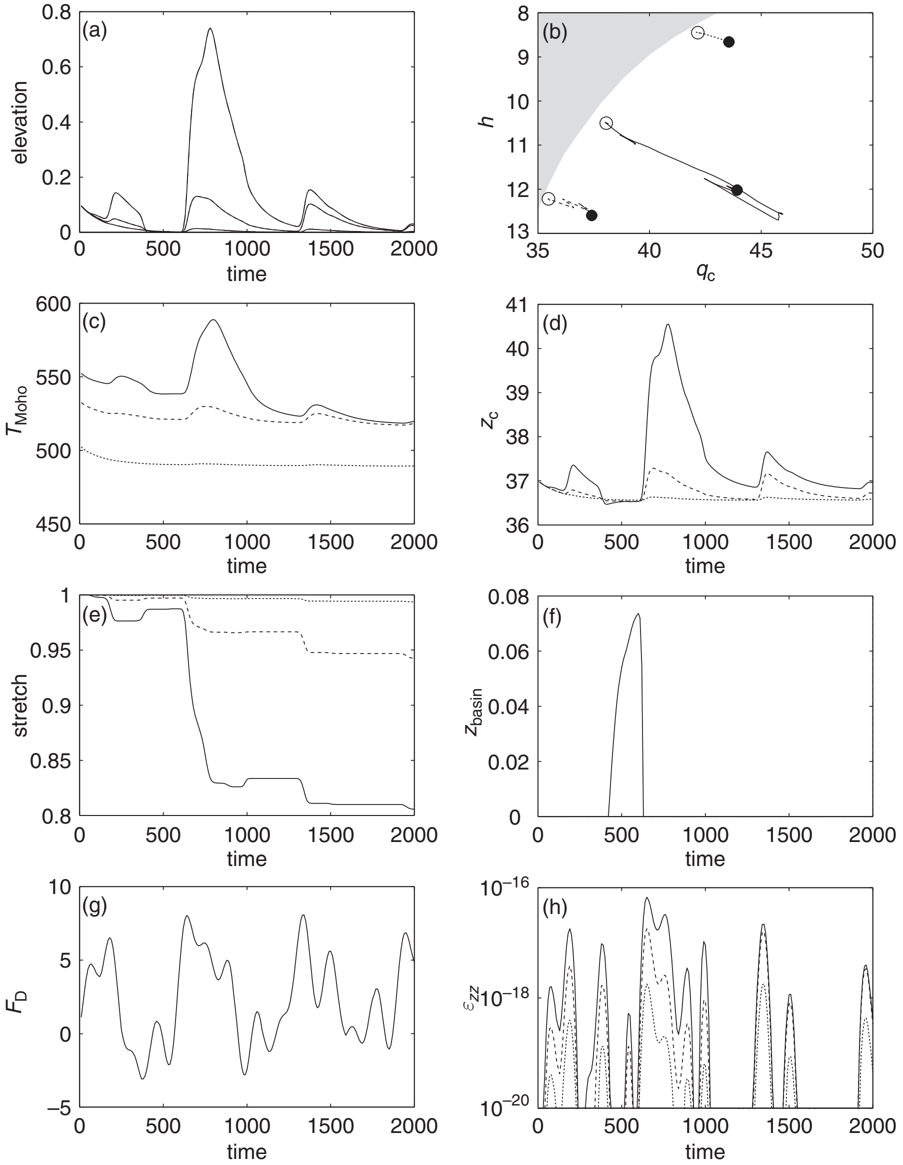

Figure 3.10. As for Fig. 3.9, but with an initial HPE distribution approximating the homogeneous model (i.e., Fig. 3.2c).

deformation and surface processes, tend to remove HPEs from the crust until it ceases to respond to the normal fluctuation in intraplate stress levels. 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.

3.6 Discussion

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 in the 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 ~30–40 mW m^{-2} to the observed surface-heat flow. An inevitable consequence of the lithophile nature of the HPEs is that primary magma-dominated, crustal growth processes will lead to length scales for the heat source distribution that are significantly less than the characteristic crustal thickness. Nevertheless, no compelling reason has emerged as to why such processes would naturally lead to a characteristic length scale or to a characteristic abundance of HPEs that contributes typically between one-half and two-thirds of the surface-heat flow (i.e., q_c = 30–40 mW m^{-2}). It is important to realize that 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. One of our main purposes in this chapter has been to show that, in intraplate settings in particular, HPEs are redistributed as a direct consequence of the deformation as well as through related processes such as erosion and sedimentation that follow as the isostatic consequence of the deformation. Such tectonic reworking provides an additional mechanism for ordering HPEs in the continental lithosphere.

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The continents are subject to temporal fluctuations in tectonic loads due to changing plate configurations (e.g., Sandiford & Coblentz, 1994) as well as interactions with the convective mantle beneath. Hot, weak continental lithosphere will respond to relatively small loads by deforming. Under any prevailing load regime, the deformation may be limited through the accumulation of potential energy in the deformed lithosphere. 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. In the previous sections we have introduced a simple parameterization of crustal HPE distributions that provides a framework for understanding how crustal-scale deformation associated with tectonic loading may effect long-term changes in thermal regimes through the redistribution of HPEs. This framework shows that deformation, when linked with a surface process through an isostatic response, is capable of effecting significant change in heat production parameters. The most significant response to the coupled deformation-surface response regime outlined here, when measured as a function of strain increments, occurs for crust already strongly differentiated. As noted above, this concurs with the notion that the primary differentiation of the HPEs within the crust is due to magmatic processes attendant with primary crustal growth.

The new insight provided by our analysis comes with the recognition that the likelihood of deformation affecting the continental lithosphere is dependant on the thermal regime and therefore on the abundance and distribution of HPEs. This is best exemplified by Fig. 3.8b, 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 may 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–q_c 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 providing 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 stabilized.

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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 stabilization of continental lithosphere is intimately intertwined with the notion of craton development. Whereas 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 chapter suggests that crustal-scale differentiation of HPEs is a necessary precursor to cratonization. Whereas cratons are manifest by long-term tectonic stability, cratonization is usually presaged by a lengthy "early" history involving not only crustal growth but also extensive crustal reworking during episodic tectonism spanning many hundreds of millions of years. In a thermal and mechanical sense, such reworking may be viewed in terms of the way it redistributes the HPEs (e.g., McLaren & Sandiford, 2001). Indeed, such a view may be consistent with the observation that ~~that~~ Archean cratons are

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characterized not only by lower surface heat flow (~40 mW m^{-2}), but somewhat lower characteristic length scales than younger geological provinces.

The data compilation in Taylor and McLennan (1985), largely based on Vitorello and Pollack (1980) and Morgan (1984), suggests that the mean (and standard deviation) length scale for heat production in Archean 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 Archean 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 (1996) have suggested that the mantle heat flow beneath Archean cratons is typically ~14 mW m^{-2}, implying that the HPEs contribute ~25 mW m^{-2}. ~~Given that at 3 Ga~~ heat production was twice modern-day rates, the characteristic Archean 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 stabilize 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 Archean cratons the crustal HPE complement is carried in granites that form the cores of large "diapiric" domes (e.g., Choukroune *et al.*, 1997). Indeed, the origin of the characteristic "dome and basin" geometry of these Archean 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 stabilizing continental crust by redistributing the HPEs into the shallow crust.

delete "Given that at 3 Ga" and replace with "Due to secular decay"

before "was" insert "at 3 Ga"

before "the characteristic" insert "and so"

3.7 Acknowledgements

The ideas presented in this chapter have benefited from discussions with Martin Hand. We are grateful to Becky Jamieson, Chris Beaumont and Keith Klepeis for their helpful reviews of the manuscript.

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