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Review Article

Radiogenic heat production, thermal regime and evolution of continental crust

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ABSTRACT

Heat flow and heat production data complement seismic information and provide strong constraints on crustal composition, thickness and evolution. They have helped understand the nature of the Mohorovicic discontinuity and the variations in seismic velocities below the Moho. Notably, heat flow studies have delineated the vertical distribution of heat producing elements throughout the crust and in the upper most mantle lithosphere. Analysis of global data sets on heat flow and crustal thickness demonstrate that there is no correlation between these two variables. This is due to the large spatial variations in crustal composition and heat production that exist within a single geological province. For a given crustal thickness, the Moho temperature varies within a wide range ($\approx$300 K) depending on surface heat flux and crustal heat production. Thus one cannot use generic models based on a “type” crustal column to calculate crustal geotherms. In stable regions, lower crustal temperatures depend on the amount and vertical distribution of heat producing elements in the crust. These temperatures determine the conditions of crustal stability and impose a limit on the maximum thickness of a stabilized crust.

Contents

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1. Introduction

The discovery of the crust mantle boundary and the first estimates of crustal thickness by Mohorovicic in 1909 were not a major surprise to the geophysical community. As a matter of fact, they had been anticipated from heat flow considerations. Following the discovery of radioactivity, the 4th Lord Rayleigh, Baron Strutt (1906) compared the few available estimates of the Earth's heat flow with those of radio-activity and heat production in crustal rocks to conclude that the composition of rocks must change at a depth less than 60 km, otherwise the total heat production would greatly exceed the surface heat flux. That heat flux and heat production data complement seismic data on crustal thickness and composition did not escape Jeffreys (1936). At the time, seismic velocities in the crust were interpreted in terms of granitic and gabbroic layers, each about 20 km thick. Using measurements of heat production in samples of granite and gabbro and following the same line of reasoning as Strutt (1906), Jeffreys argued that a thick granitic upper crustal layer is inconsistent with the heat flux measurements (see also Jeffreys, 1942). More recently, this approach has been pursued by petrologists and geochemists to estimate the ratio of felsic to mafic rocks in the continental crust and its average composition (McLennan and Taylor, 1996; Rudnick and Fountain, 1995; Taylor and McLennan, 1995a,b).

This early focus on the crust was such that it provided the motivation for the first heat flux measurements on the sea floor because it was thought that they would demonstrate the differences in thickness and composition between oceanic and continental crusts. We know now that the high values and large variations of oceanic heat flux have nothing to do with the thickness and composition of the oceanic crust (Bullard, 1954) and that they record the cooling of the oceanic lithosphere. After the effect of hydrothermal circulation has been accounted for, the oceanic heat flux is a simple function of the age of the sea floor (Sclater and Francheteau, 1970; Sclater et al., 1980).

The success of the thermal models of the sea floor led many authors to apply the same approach to continental heat flow. It soon appeared that the heat flow through continents does not depend on a single geological parameter such as age. Continental crust varies in thickness and composition, such that changes of crustal heat production account for a large fraction of heat flow variations in lithosphere that has reached thermal equilibrium. The stability and the evolution of the continental crust, however, are controlled by its thermal structure which we must determine from heat flux and heat production measurements. In this article, we shall focus on the thermal regime of the stable continental crust. We shall briefly review what we have learned about the concentration and distribution of heat producing elements (HPE) in the crust. We shall examine how crustal thickness, composition, and surface heat flux are related and we shall estimate temperatures in the lower crust. Our main objectives are to understand how the continental crust stabilizes and what controls its thickness.

2. Stable continental crust: crustal composition

Continental heat flux varies much within the continents but is generally lower in stable than in active provinces (Fig. 1). Over most of the Precambrian provinces, heat flux is less than 60 mW m^{-2}; over tectonically active regions it is higher than 80 mW m^{-2}. Archean cratons stand out with low heat flow (Jaupart and Mareschal, 1999; Morgan, 1985; Nyblade and Pollack, 1993). Archean cratons are also characterized seismically by a thick lithospheric root with high seismic velocities (Grand et al., 1997). The crustal thickness varies over the continents but, with the exception of active orogenic regions, remains less than 60 km (Fig. 2).

The thermal relaxation time for the continental lithosphere depends on its thickness and on the boundary conditions, specifically whether heat flux or temperature is fixed at the base of the lithosphere (Jaupart and Mareschal, 2007, and references therein). For 250-km thick lithosphere, quasi steady-state conditions are reached after 200–500 My. When the lithosphere has reached thermal steady-state, the surface heat flux Q_0 can be decomposed into several components:

$$Q_0 = Q_c + Q_L + Q_B \tag{1}$$

where Q_c is the total heat production of the crust, Q_L is the total heat production in the lithospheric mantle, and Q_B is the heat flux at the base of the lithosphere. For true steady-state over the whole lithosphere, Q_B must be constant, but it should be seen as a time-averaged value as temporal fluctuations are damped out by diffusion. The heat production of the lithospheric mantle is much smaller than that of the crust but it cannot be neglected in cratons with a thick lithospheric root (Michaut et al., 2007). The heat flux across the Moho, Q_m is such that:

$$Q_m = Q_L + Q_B. \tag{2}$$

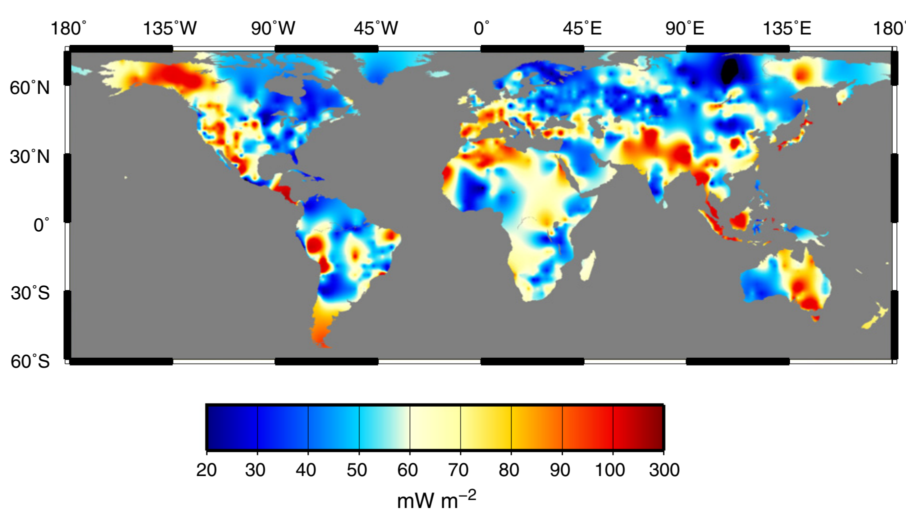

Fig. 1. Continental heat flow variations. In stable continental regions, the variations in heat flux are due to changes in crustal thickness and composition. Data from a compilation by Francis Lucazeau.

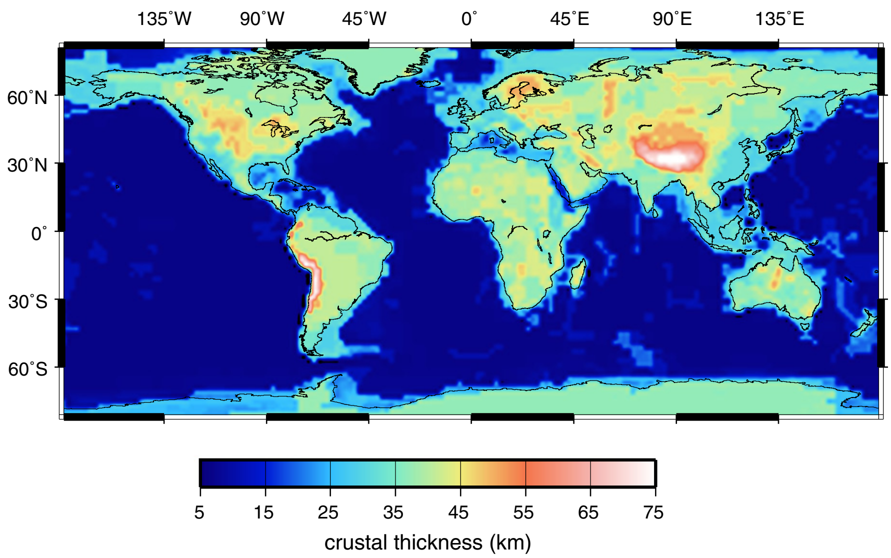

Fig. 2. Crustal thickness variations from the global compilation of seismic crustal thickness data, CRUST2.0 by Mooney et al. (1998).

Heat production and, by way of consequence, heat flux vary spatially by large amounts. The different components Q_c, Q_L and Q_B vary on different horizontal scales. Short wavelength variations of heat flux can only be due to local variations of crustal composition. Long wavelength variations however can be caused by regional geology juxtaposing terranes of different origins and ages, as well as changes of basal heat flux and heat production in the lithospheric mantle. The superposition of different sources of spatial variations is the main cause of our difficulties in thermal calculations. In order to interpret heat flow data, one must work with horizontally averaged quantities. For the crustal component, Q_c, the scale is determined by crustal thickness, z_m; for the other two components, Q_L and Q_B, the scale is the lithospheric thickness (Mareschal and Jaupart, 2004). For horizontally averaged quantities, the total heat production of the crust is:

$$Q_c = \int_0^{z_m} A(z) dz \tag{3}$$

where $A(z)$ is the heat production rate. The heat flux at Moho, Q_m cannot be measured directly and heat production can be measured only for surface samples. For the purposes of calculating temperatures in the lower crust and evaluating the conditions for crustal stability, however, one must determine Q_m and the vertical distribution of the heat producing elements.

2.1. Moho heat flux

In stable continental regions, the variations in surface heat flux often occur with wavelengths shorter than the crustal thickness and cannot be explained by variations of Moho or basal heat flux (Mareschal and Jaupart, 2004; Mareschal et al., 2000). Heat flux measurements are made in holes of opportunity and data are unevenly distributed. With highly variable spatial sampling, the average heat flow field cannot be determined with a precision better than ± 3 mW m^{-2} (Levy et al., 2010), which sets a lower limit on the amplitude of Moho heat flux variations that can be resolved by the data. Within these limits, the Moho heat flux has been found to be in a range of 12–18 mW m^{-2} (see Table 1). This wide range spans 40% of the bulk value. Unfortunately, it is at present impossible to tighten it using heat flux data only, but some progress can be made by combining heat flux and seismic data (Levy and Jaupart, 2011; Levy et al., 2010). In calculations of the amount of crustal heat production and crustal temperatures, the uncertainty on the Moho heat flux is not severely limiting, however.

Table 1
Various estimates of the heat flux at Moho in stable continental regions.

Region	Moho heat flux (mW m^{-2})	References
Norwegian Shield	11[a]	Pinet and Jaupart (1987)
Vredefort (South Africa)	18[a]	Nicolaysen et al. (1981)
Kapuskasing (Canadian Shield)	11–13[a]	Ashwal et al. (1987); Pinet et al. (1991)
Grenville (Canadian Shield)	13[a]	Pinet et al. (1991)
Abitibi (Canadian Shield)	10–14[a]	Guillou et al. (1994)
Siberian craton	10–12[a]	Duchkov (1991)
Dharwar craton (India)	11–19[a]	Roy and Rao (2000)
	14–20[b]	Roy and Mareschal (2011)
Trans-Hudson orogen (Canadian Shield)	11–16[ac]	Rolandone et al. (2002)
Slave province (Canada)	12–24[d]	Russell et al. (2001)
Baltic Shield	11[a]	Kukkonen and Lahtinen (2001)
	7–15[d]	Kukkonen and Peltonen (1999)
Kalahari craton (South Africa)	17–25[d]	Rudnick and Nyblade (1999)

[a] Estimated from surface heat flux and crustal heat production.
[b] Estimated from surface heat flux and crustal heat production, shear wave velocity profiles, and geothermobarometry on mantle xenoliths.
[c] Estimated from condition of no melting in the lower crust at the time of stabilization.
[d] Estimated from geothermobarometry on mantle xenoliths.

Table 2
Low heat flow regions.

Region	Province	Age Gy	<Q> (mW m^{-2})	<A> (μW m^{-3})	Reference
Lynn Lake Belt	THO (Canada)	1.8	22	0.7	Mareschal et al. (2000)
Voisey Bay	Nain Plutonic Suite (Canada)	1.4	22	0.7	Mareschal et al. (2005)
	Baltic Shield	2.5	22-28		Kukkonen and Joehlet (1996)
	Siberian Shield	2.5	21	–	Duchkov (1991)
Niger	West Africa Shield		17–22	–	Chapman and Pollack (1974)
Tagil-Magnitogorsk	Urals	0.4	25	0.3	Kukkonen et al. (1997)

One can obtain constraints on the Moho heat flux directly from heat flux and heat production data. How this can be done has been discussed in many papers (Jaupart and Mareschal, 2007, and references therein) and we shall only recall a few key arguments.

2.1.1. Low heat flow regions

The heat flux cannot be higher at the Moho than at surface. Low values of the surface heat flux provide an absolute upper limit for the Moho heat flux. Values as low as 22 mW m^{-2} have been reported for several regions in different continents (Table 2). The upper bound for Q_m is even lower than 22 mW m^{-2} when one accounts for the minimum possible crustal heat production, ≈4 mW m^{-2}.

2.1.2. Crustal sections

In many places of the world, the lower crust has been transported along a ramp and thrust over the upper crust, resulting in the exposure of large parts of the crustal column (see for instance Percival et al., 1992). The rebound following meteoritic impacts has also exposed the entire crust in some large impact structures, such as the Vredefort structure in the Kaapvaal craton, South Africa. After measuring the heat production of samples from different crustal levels and estimating the crustal composition, it is possible to calculate the crustal heat production and Moho heat flux (Ashwal et al., 1987; Nicolaysen et al., 1981). Values of Moho heat flux estimated from different crustal sections range between 13 mW m^{-2} for the Kapuskasing structural zone, Superior Province of the Canadian Shield, and 18 mW m^{-2} for the Vredefort.

2.1.3. Sampling different crustal levels

Crustal slices that have been brought up from different crustal levels are now juxtaposed at the surface in old orogenic belts, such as the Grenville Province in the Canadian Shield. The average crustal heat production can be obtained from systematic sampling of the surface rocks. In such regions, the Moho heat flux is obtained by subtracting the average heat production from the average surface flux. For the Grenville province, the Moho heat flux has been estimated at 13 mW m^{-2}, a value consistent with the other estimates (Mareschal et al., 2000).

2.1.4. Xenoliths

One can also estimate the Moho heat flux using (P, T) determinations on mantle xenoliths brought to the surface by kimberlite eruptions. Unfortunately, this method can only be used in a few areas, which does not allow comparison with the heat flow data on a continental scale. Another potential pitfall is that some kimberlite eruptions are older than 1 Ga (e.g. the kimberlites in the Dharwar craton in southern India) and provide a record of past conditions. In areas where it has been possible to compare the xenolith and heat flux based methods, they were found to be in good agreement with each other (Michaut et al., 2007).

2.2. Crustal heat production

2.2.1. Average crustal heat production

The Moho heat flux has been determined in regions where the entire crustal column can be sampled and crustal heat production determined. As discussed above, heat production varies on a very small scale, within a seemingly homogeneous pluton for instance, while heat flux integrates the entire crustal column and varies on much larger scale (Jaupart and Mareschal, 2012). In stable regions, Moho heat flux does not appear to vary much on the scale of a geological province (Mareschal and Jaupart, 2004). Therefore, in stable provinces, the crustal heat production can be calculated by subtracting the Moho heat flux from the surface heat flux averaged over a sufficiently wide area (relative to crustal thickness). Results from heat flow studies yield an average heat production of 0.77 ± 0.08 μW m^{-3} for the Precambrian crust and 1.08 ± 0.13 μW m^{-3} for the Phanerozoic with a range of 0.79–0.95 μW m^{-3} for the entire continental crust (Jaupart and Mareschal, 2012). The latter range is consistent with the bulk continental crust estimate of 0.93 μW m^{-3} derived from geochemical models (Rudnick and Fountain, 1995; Rudnick and Gao, 2003).

2.2.2. Vertical distribution of the heat producing elements

The concentration of heat producing elements (HPE) in the crust varies on many different scales. In most provinces, upper crustal rocks have a heat production higher than the bulk crust, otherwise crustal heat production would exceed the surface heat flux. Lower crustal rocks tend to be depleted in radio-elements compared to the upper crust, but variations in heat production with depth are not monotonic and cannot be described by a simple function. This has been demonstrated by sampling in exposed crustal sections (Brady et al., 2006; Ketcham, 1996) as well as by measurements in deep scientific drill holes (Arshavskaya et al., 1987; Clauser et al., 1997; He et al., 2008). Measurements on granulite facies rock samples from many locations worldwide yield consistently low heat production values (0.2–0.5 μW m^{-3}), that are assumed to be representative for deep crustal levels (Ashwal et al., 1987; Fountain et al., 1987).

Measurements in the small number of available deep boreholes demonstrate that the vertical distribution of heat production depends on the local history of crustal accretion and deformation, and cannot be used in calculations of the total amount of heat producing

Table 3
Average heat flux, crustal heat production, and differentiation index for different provinces and subprovinces of the Canadian Shield. <Q> average surface heat flux, N_Q number of sites, <A> average surface heat production, N_A number of values, z_m crustal thickness, DI differentiation index.

Province	Age Ga	<Q> ± σ_Q mW m^{-2}	N_Q	<A> ± σ_A μW m^{-3}	N_A	z_m km	DI
Slave	3.1–2.9	51 ± 6	5	2.3	[a]	36	2.3 ± 0.5
Superior	2.9–2.6	40 ± 10	79	0.72 ± 0.6	62	40	1.2 ± 0.1
Abitibi	2.7	38 ± 7	30	0.4 ± 0.3	24	40	0.7 ± 0.1
Trans Hudson Orogen	2.1–1.8	42 ± 11	49	0.7 ± 0.5		40	1.1 ± 0.2
Thompson Belt	2.1–1.9	53 ± 6	10	1.1 ± 0.3	10	40	1 ± 0.1
Flin Flon Snow Lake Belt	1.9–1.8	42 ± 5	15	0.3 ± 0.2	14	40	0.7 ± 0.1
Lynn Lake Belt	1.8	32 ± 7	9	0.7 ± 1	20	36	2.3 ± 0.1
Grenville	1.3–1.1	41 ± 11	–	0.8	[a]	40	1.3 ± 0.2
Appalachians	0.4	57 ± 13	79	2.6 ± 1.9	50	40	2.5 ± 0.2

[a] Area weighted average.

elements in the crust. As explained above, in principle, one needs to determine horizontal averages of heat production over the whole crustal thickness, but this is not feasible. In practice, one can show that crustal geotherms are not sensitive to the exact form of the vertical distribution of heat production and that they depend mostly on the thickness of enriched upper crustal rocks. To get around this problem, one can use the average values of both surface heat flow and heat production together with an estimate of the Moho heat flux to characterize the amplitude of crustal stratification (Perry et al., 2006a). The differentiation index (DI) is defined as the ratio of the averaged heat production measured on surface samples, $<A>$, to the vertically averaged crustal heat production, $(Q_0\text{-}Q_m)/z_m$

$$DI = \frac{<A> \times z_m}{Q_0 - Q_m}. \quad (4)$$

Estimates of DI in North-America show that it increases with the average crustal heat production (Perry et al., 2006a) (Table 3). Because of crustal differentiation and the resulting enrichment of the upper crust in HPE, it is expected that DI should always be >1. This is not always observed because the uppermost crustal layer may have been emplaced on or transported over a more radioactive basement (e.g. Abitibi belt, Flin-Flon Snow Lake belt in the Trans Hudson orogen, Canada, or the Kola peninsula, Baltic Shield).

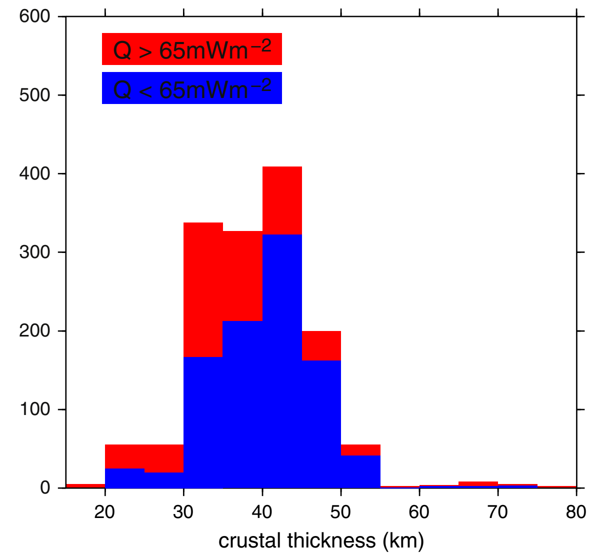

Fig. 4. Histogram of crustal thickness averaged on 2°×2°cells in continents, for all cells where heat flow data are available, divided between cells where the heat flux is higher and lower than the continental average (65 mW m^{-2}).

3. Heat flow and crustal thickness

The heat flux data base (now accessible at http://www.heatflow.und.edu/) has recently been updated by Francis Lucazeau and Derrick Hasterok. It contains more than 35,000 values on land including new data in Canada and India which provide very detailed sampling of Precambrian provinces. Also included are bottom hole temperature measurements in oil exploration wells from sedimentary basins and continental margins. However, the glacier covered areas of Greenland and Antarctica remain practically un-sampled. In high latitude regions where measurements are impractical because of the permafrost, sampling remains very poor. There are several biases in the data sets because of the uneven geographical distribution of measurements, but also because many measurements were made for geothermal exploration in high heat flux areas. Consequently, the mean of all continental heat flux measurements is biased toward a high value (>80 mW m^{-2}). Different methods have been used to remove such bias, either by area weighting the heat flux data (Jaupart and Mareschal,

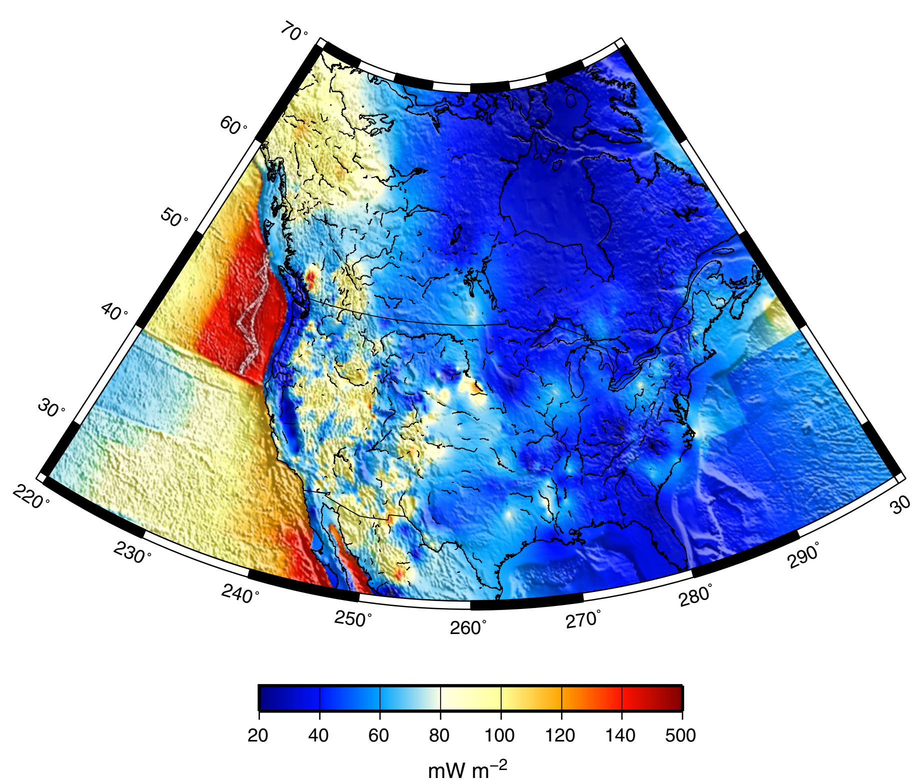

Fig. 3. Heat flow map of North America. The data are from the compilation by Blackwell and Richards (2004) with some addition of new data in Canada. Note the dichotomy between the stable eastern and the active western parts of the continent.

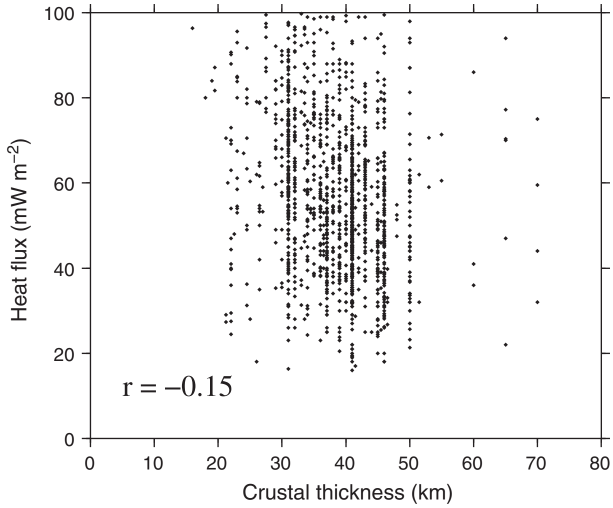

Fig. 5. Heat flux vs crustal thickness (both averaged over 2°×2°cells) show no positive correlation.

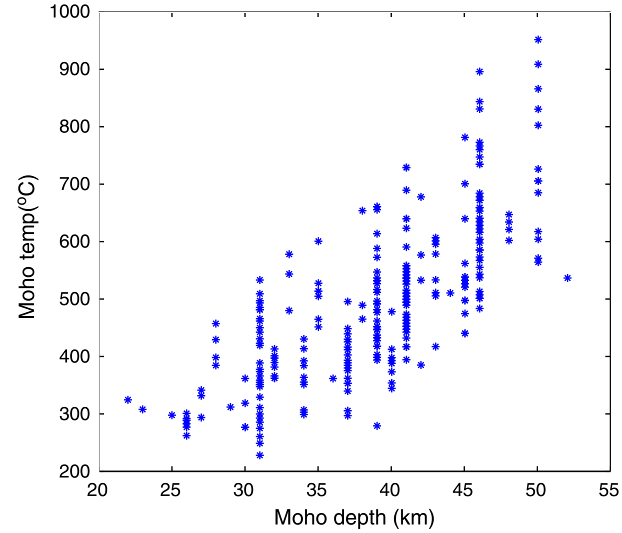

Fig. 6. Calculated Moho temperatures as a function of crustal thicknesses as explained in the text.

2007), or by averaging the data by geologic type (Davies and Davies, 2010; Pollack et al., 1993). Within the error margin, both methods yield identical values of 64 mW m^{-2} for the mean continental heat flux.

When the heat flux data are averaged over 2°×2° cells, they can be compared to the global crustal thickness data set CRUST2.0 that is provided with a similar resolution (Chulick et al., 2002; Mooney et al., 1998). Both data sets have many serious shortcomings. The heat flow data set is a compilation of all reported measurements. These data were obtained in variable conditions (e.g. the depth range of the temperature measurements); various corrections are applied, but not in a systematic way; many of the data are affected by noise (mostly groundwater convection), and some are simply erroneous. Data from stable regions, particularly from Shield areas, are in general of much higher quality than data from active regions because they are less affected by groundwater flow or topography driven convection. The major difficulty with the seismic "model" CRUST2.0 is that a crustal column is defined for each cell: for many cells where data are not available or have not been included, the column definition is based on geological type. In addition, some data, now obsolete, have not been replaced by the more recent ones. For example, in Canada, comparison between the measured crustal thickness data from LITHOPROBE and the former version of the crustal model, CRUST5.1, yields a root mean square difference of 5.5 km between the two sets of crustal thicknesses (Perry et al., 2002).

Nevertheless, one can extract useful information from these large data sets. Comparing heat flux with crustal thickness and elevation suggests a dichotomy between tectonically stable regions, with little variations in elevation and crustal thickness, where the surface heat flux is relatively low (<65 mW m^{-2}) and active regions with high heat flux (>65 mW m^{-2}), high elevation, and where the crust is often anomalously thin (<30 km) or thick (>55 km). For the heat flow field, this dichotomy is well demonstrated by the contrast between the eastern and western halves of the North American continent (Fig. 3). This shows that surface heat flow cannot be reliably estimated from estimates of crustal thickness and average heat production. We return to this important point below.

On a global scale, there is a difference in the distribution of crustal thicknesses between regions with lower and higher than average heat flux (Fig. 4). In this comparison of heat flux with crustal thickness, we have excluded all the cells with an average heat flux >120 mW m^{-2} because we suspect that some of the very high heat flux values in the data base are either erroneous, and/or are affected by transient perturbations at shallow depth and thus not representative of the crustal heat flux. In tectonically active regions where heat flux is higher than 65 mW m^{-2}, the average crustal thickness 36±10 km is not significantly different from that in regions with lower than average heat flux, but the histogram shows differences in the distribution of crustal thickness with relatively more very thin (<30 km) or very thick (>60 km) crust than in stable regions. The sampling of heat flow is poor in high elevation regions with a thick crust. The sampling is sufficient in regions with a thin crust which are definitely characterized by higher than average heat flow and a transient thermal regime.

3.1. Stable provinces

If variations in crustal composition in stable continental regions were random, there should be a trend of increasing heat flux with crustal thickness. Regardless how the heat flux data are selected (for example, by eliminating cells where the average heat flux is too

Table 4
Preserved thick crustal roots.

Location	Age (Ga)	Thickness (km)	Heat flux (mW m^{-2})	Reference
Kapuskasing	1.8	48–50	33	Pinet et al. (1991)
Eastern Grenville Front	1.1	50–55	32	Mareschal et al. (2000)
Lynn Lake Belt (THO)	1.8	48–50	32	Mareschal et al. (2005)
Siberia	Archean	48–55	22	Cherepanova et al. (2010)
Baltic Shield	Archean	50–58	25	Luosto et al. (1990); Kukkonen and Joehlet (1996)
Central Australia	Proterozoic	60	?	Clitheroe et al. (2000)

Table 5
High heat flow regions in stable continental Provinces.

Province	Age (Ga)	Heat flux ($mW\ m^{-2}$)	Surface heat production ($\mu W\ m^{-3}$)	Crustal thickness (km)	DI	References
Canada						
Wopmay orogen	2.1	90 ± 15	4.8	32	2.	Lewis et al. (2003)
Australia						
SAHFA [a]	1.6	92 ± 8	6	35	3	Neumann et al. (2000)
Eastern Gawler craton	1.6	72 ± 24	5	35	3.2	Neumann et al. (2000)
Central Shield (Australia)	1.8	78 ± 19	3.6 ± 1.9	40	2.4	Neumann et al. (2000)
Northern Indian Shield						
Aravalli	1.5–2.4	68 ± 13	–	–	–	Roy and Rao (2000)
Bastar craton	3.0–3.5	55 ± 7	–	–	–	Gupta et al. (1993)
Singbum craton	3.0–3.4	61 ± 2	–	–	–	Rao and Rao (1974)
Chottagnapur gneiss complex	2.5	59	5	–		Kumar et al. (2009)
South Africa						
Lesotho	Archean	61 ± 16				Jones (1992)
Namaqa	Prot	61 ± 11	2.3	43	2	Jones (1987)

[a] The South Australia heat flow anomaly (SAHFA) belongs to the Gawler craton.

high to represent steady state thermal regime), there is absolutely no correlation between heat flux and crustal thickness (Fig. 5). If the heat flux and the total crustal heat production do not increase with crustal thickness, the concentration in HPEs must decrease as the thickness of the crust increases. There is even a slightly negative correlation between heat flux and crustal thickness, but it is not significant because of the large spread of the data sets. The decrease in the concentration in HPE may be due to the removal by erosion of the enriched upper section of a thick differentiated crust. It may also be due to the relaxation of a hot crustal root when the crust is rich in HPE and Moho temperatures are high.

Crust thicker than 50 km is unusual in Precambrian Shields, but all the crustal roots in old provinces are associated with low heat flux (Table 4). It is expected that, over a time scale of 1 Gy, a crustal root should relax unless the lower crust is strong. Crustal strength depends on composition and temperature: mafic rocks are stronger and depleted in HPE relative to felsic rocks. Thus, crust with large amounts of mafic rocks is strong (Ranalli, 1995) and has lower Moho temperatures than a more felsic one. This accounts for the preservation of thick crust in low heat flow regions: cold mafic lower crust can withstand the horizontal stresses induced by the root (e.g., Mareschal et al., 2000; Perry et al., 2006c).

We have estimated the temperatures at the crust mantle boundary for all the cells where the surface heat flux is less than 100 $mW\ m^{-2}$. In steady state, temperature must always increase with depth but the rate of increase depends on the thermal conductivity $\lambda(T)$, the surface heat flux, Q_0 and the crustal heat production, $A(z)$. The temperature at depth z is obtained by solving the 1-D heat equation:

$$\lambda(T)\frac{dT}{dz} = Q_0 - \int_0^z A\left(z'\right)dz'. \quad (5)$$

For these calculations, we have used the following experimental relationship to include the effect of temperature on lattice thermal conductivity (Durham et al., 1987):

$$\lambda(T) = 2.264 - \frac{618.2}{T} + 3.0\left(\frac{355.6}{T} - 0.3205\right) \quad (6)$$

where T is absolute temperature, and the parameters are such that $\lambda = 3\ W\ m^{-1}\ K^{-1}$ for $T = 20$ °C.

There is a general trend of increasing Moho temperature with crustal thickness (Fig. 6), which simply reflects that temperature always increases with depth. The trend is well-defined but, for a given thickness of the crust, the range of Moho temperatures is wide (>300 K). The Moho temperature can be calculated as the sum of two components, corresponding to crustal heat production and Moho heat flux, respectively. The former is proportional to z_m^2 and is the largest one in most cases. Thus, if the average heat production was about the same in all provinces, one would expect surface heat flux to be proportional to crustal thickness and a roughly quadratic trend of Moho temperature with crustal thickness. This is clearly not the case as shown by Figs. 5 and 6. Because of the wide Moho temperature range (±300 K), crustal thickness cannot be used to predict Moho temperature, as would be the case if the crust mantle boundary was a phase change boundary in thermodynamic equilibrium (Kennedy, 1959; Lovering, 1958; Mareschal et al., 1982).

3.2. Very high heat flow regions

Values of the heat flux higher than the continental average are not exclusive of the active regions. High (>60 $mW\ m^{-2}$) and very high (>75 $mW\ m^{-2}$) heat flow anomalies are found in Precambrian provinces of different continents, most notably in the Proterozoic (Table 5). These are always associated with very high concentrations of HPEs in surface rocks and a high value of the differentiation index. The extremely high surface heat flux values ($\mu(Q) = 90\ mW\ m^{-2}$) observed in the South Australian craton are associated with very high heat production in surface rocks ($\mu(A) = 6\ \mu W\ m^{-3}$), reaching up to 62 $\mu W\ m^{-3}$ in the Yerila Granite (Neumann et al., 2000). The 2.1 Ga Wopmay Orogen in the Canadian Shield also exhibits very high heat flux and surface heat production values (Lewis et al., 2003). For both the Wopmay and the South Australian heat flow anomalies, the excessive heat flux is entirely accounted for by very high heat production in the uppermost crust. With high values of the differentiation index, the thickness of the enriched layer does not exceed 10 km.

3.3. Mantle P_n velocity

The velocity of P and S waves in the upper mantle depends on composition and temperature (Cammarano et al., 2003). As expected, high heat flow regions usually have low P_n velocities, and low heat flow regions have higher P_n velocity. The correlation is far from perfect because Moho temperature is not proportional to surface heat flux but also depends on crustal thickness and composition. Where enough data are available, reliable thermal crustal models can be established and the Moho temperature can be calculated. One finds that P_n velocities are correlated with Moho temperatures, with values of $\partial VP_n/\partial T$ that are on the order of $-6.0 \times 10^{-4} \pm$

10% km s^{-1} K^{-1}, within the range of temperature derivatives obtained in laboratory studies of ultramafic rocks (Perry et al., 2006b). Systematic studies of the relationship between P_n velocity and Moho temperature have not been conducted because the global data sets used in this study are inadequate, lacking the necessary resolution and precision. Regional studies should be conducted where sufficient good quality heat flux measurements are available and variations in P_n velocities are precisely determined.

3.4. Active provinces

Recently-active provinces, regions that are not in thermal steady state, are usually characterized by high heat flux and thick crust in regions of convergence or by thin crust in regions of extension. There are few heat flow data from tectonically active mountain belts, but zones of extension are very well sampled. In these zones, crustal thinning, and possibly magmatic intrusions, result in elevated surface heat flux (Lachenbruch and Sass, 1978). In this case it is the dynamics of crustal thinning that determines the surface heat flux, but the style and amount of extension are controlled by crustal heat production and lithosphere thickness (which both determine the steady-state heat flux) (Buck, 1991).

4. Controls on crustal thickness

Continents (including their submerged margins) have a mean elevation of a few hundred meters against a mean depth of 3800 m for the oceans. The average thickness of the continental crust is between 35 and 40 km. This, however, is not very well determined for different reasons. One is that the Moho is not always well recognized. A second reason is that the sampling of seismic crustal studies is almost as poor as that of heat flow. Seismic surveys have often targeted regions where the crust is anomalously thick or thin, and large regions have been left out. Global compilations are thus incomplete, and regions that are not sampled are assigned a typical crustal column supposedly characteristic of the geological type. Despite the inadequacy of the present compilations, the point can be made that very thick continental crust is exceptional and crust thicker than 60 km accounts for less than 10% surface area of the continents and is found only in tectonically active regions (Mooney et al., 2004).

The thermal regime regulates crustal thickness in two ways. The temperature at the base of the crust increases with crustal thickness. If crustal heat production were constant and uniform throughout the crust, temperature would increase nearly quadratically with crustal thickness, and temperature would be high enough for melting to occur near the base of a thick crust. Variations in crustal thickness result in gravitational potential energy differences and induce large stress (Artyushkov, 1973, 1974; Fleitout and Froidevaux, 1983). A local increase in crustal thickness increases the stress and lowers the strength of the lithosphere. Whether the lithosphere is strong enough to sustain large stress depends on its thermal regime. As long as compressional tectonic stresses counter balance the difference in gravitational potential energy, the thickened crust can survive (England, 1987), but it will collapse when the compressional stress has been relaxed (Gaudemer et al., 1988).

4.1. Temperatures in the crust

The temperature at the base of the crust increases with crustal thickness but there is no simple relationship between the two variables due to several factors. One is that the average crustal heat production is not constant and varies horizontally, even in a single geological province. Another factor is the vertical distribution of heat production, which changes from province to province. One can see from Eq. (5) that, with the same surface heat flux, the shallower the heat producing elements, the lower the crustal temperatures will be. Finally, a third factor is the effect of temperature on thermal conductivity which results in higher Moho temperature than in calculations with uniform conductivity. One can still see that, for a given average crustal composition, the Moho temperature will always exceed melting temperature if the crust is too thick.

Crustal differentiation effectively lowers the temperature at the base of the crust, allowing stabilization of a thicker crust. If temperatures are too high, partial melting occurs in the lower crust (Michaut et al., 2009; Perry et al., 2006b; Sandiford and McLaren, 2002). Assuming that partial melting in crustal rocks starts when temperature reaches 850 °C, we have calculated the depth where steady state temperature exceeds 850 °C for different concentrations of the HPEs, different values of DI, and a value of 15 mW m^{-2} for the Moho heat flux (Fig. 7).

These calculations show that even for low average crustal heat production (0.5 μW m^{-3}), it is impossible to avoid melting in a crust thicker than 60 km unless it is differentiated. With heat production higher than average (1 μW m^{-3}), a thick crust cannot be stabilized even when it is very differentiated. With a relatively high average crustal heat production today (≈1.0 μW m^{-3}), and twice that much when it stabilized ca 3.0 Ga, the Slave Province in Canada must have been very differentiated and its crustal thickness could

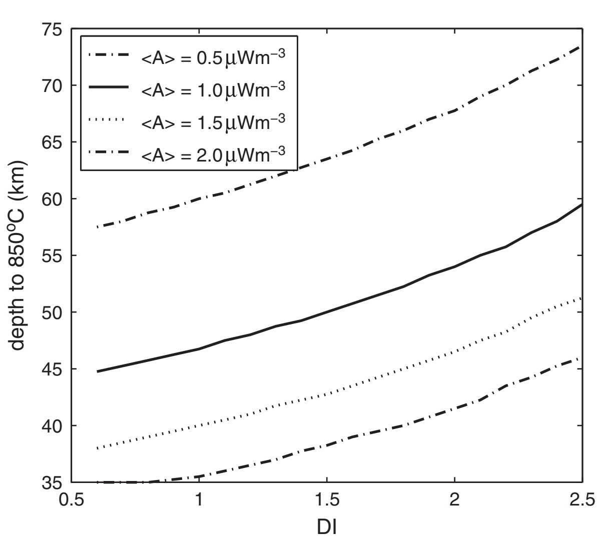

Fig. 7. Depth where the steady state temperature in the crust reaches 850 °C as a function of differentiation index (DI) and average crustal heat production. For these calculations, heat flux at the Moho is assumed to be 15 mW m^{-2} and the thermal conductivity is given by Eq. (6).

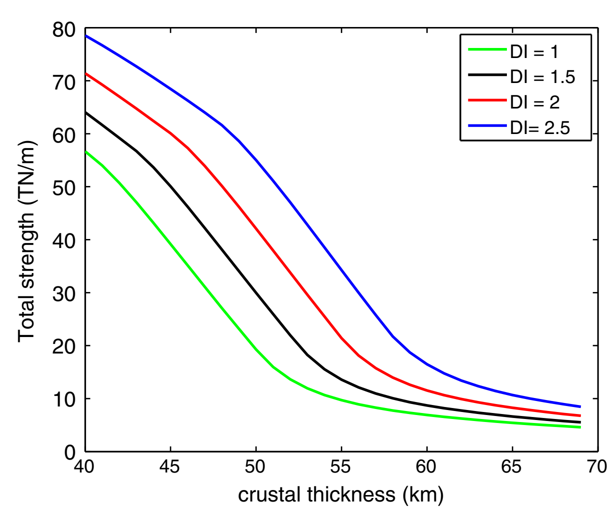

Fig. 8. Strength of the lithosphere as a function of crustal thickness for different values of the differentiation index. For all calculations the average crustal heat production is 0.8 μW m^{-3} and the layer enriched in radioactive elements is assumed to be 15 km thick.

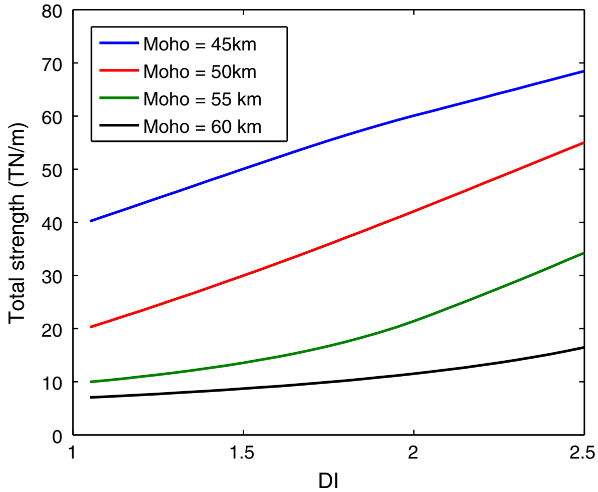

Fig. 9. Strength of the lithosphere as a function of differentiation index for different crustal thicknesses. For all calculations the average crustal heat production is 0.8 μW m^{-3} and the layer enriched in radioactive elements is assumed 15 km thick.

not have exceeded 45–50 km when it reached thermal steady state. Present crustal thickness in the Slave craton (36 km) is less than the average Canadian Shield.

4.2. Strength of the lithosphere

Crustal thickening by thrusting and creep deformation requires the resultant stress (difference between the compressional tectonic stress and the tensile stress due to crustal thickening) to be sufficiently large to overcome the total strength of the lithosphere. After relaxation of the tectonic stress, the tensile stress will dominate. A very crude estimate of this stress can be obtained by considering that elevation differences are proportional to crustal thickness differences. The effect of lithospheric thickening can be included but it requires further assumptions on the mechanism of thickening (e.g. Mareschal, 1994; Zhou and Sandiford, 1992). For a density contrast between mantle and crust $\Delta\rho/\rho \approx 1/8$, the stress throughout most of the crust increases by ≈3.75 MPa/km of crustal thickening, i.e. 75 MPa for 20 km crustal thickening. In terms of the gravitational potential energy, this amounts to an increase ≈5 TJ m^{-2} for 20 km crustal thickening. Whether the crust can withstand such stress depends on its strength, and ultimately on its thermal structure.

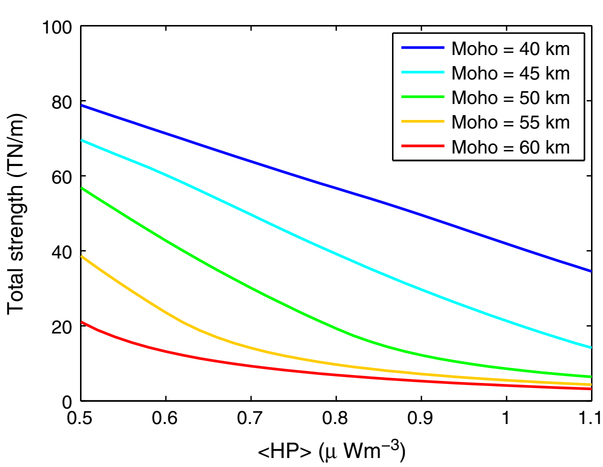

Fig. 10. Strength of the lithosphere as a function of average crustal heat production for different crustal thicknesses. For all calculations, DI = 1.

Table 6
Creep parameters for lithospheric materials used in calculating the strength of the lithosphere (Carter and Tsenn, 1989; Ranalli, 1995).

	A (MPa^{-n}s^{-1})	n	H (kJ mol^{-1})
Upper crust (dry granite)	$1.\times10^{-7}$	3.2	144
Lower crust (mafic granulites)	1.4×10^{4}	4.2	445
Mantle (dry dunite)	$3.\times10^{4}$	3.6	535

Over a time scale of the order of 100 My, the thickened crust reaches thermal steady-state with higher temperatures than initially (Jaupart and Mareschal, 2010). The strength of the thickened crust is controlled by temperature, i.e. essentially by the crustal heat production and its vertical distribution (Kusznir and Park, 1984; Mareschal, 1994; Zhou and Sandiford, 1992). In order to illustrate how crustal heat production controls the strength, we have calculated the integrated strength of the lithosphere assuming a "standard" rheology for the crust and mantle (see Appendix A). We have determined how the strength varies with crustal thickness, heat production, and differentiation. The point of such calculations is not to make accurate predictions but to illustrate how these parameters affect the strength of the lithosphere. Calculations show that, for the average crustal heat production of 0.8 μW m^{-3}, the value of the lithospheric strength (in TN m^{-1}, or TJ m^{-2}) decreases to a level comparable to the difference in potential energy regardless of DI when crustal thickness exceeds 60 km (Fig. 8). For a crust thinner than 55 km, the differentiation increases the lithospheric strength and might become sufficient to sustain potential energy differences (Fig. 9). Regardless of the value of DI, the lithosphere is always weak for a 60 km thick crust when the average heat crustal heat production is 0.8 μW m^{-3}. The only way a thickened crust can be maintained is when the average crustal heat production is low, i.e. <0.5 μW m^{-3} (Fig. 10). This is consistent with the observation that several crustal roots coincide with low heat flux regions (Table 4). Considering how the strength of the lithosphere depends on crustal thickness and average heat production, it is interesting to note that a thick crust is always weak and that it requires a low average heat production, i.e. <0.5 μW m^{-3}, for lithospheric strength to exceed 20 TJ m^{-2}. Vertical differentiation and redistribution of the heat producing elements can follow partial melting in the lower crust (e.g. Perry et al., 2006a), but can also be the result of tectonic redistribution as proposed by Sandiford and McLaren (2002).

4.3. Discussion

The point of the calculations above is not to predict exactly when a thick continental crust will collapse, but to understand what are the controlling factors that permit crustal stability. The calculations show that the concentration of HPEs and their vertical distribution in the crust are the parameters that determine whether partial melting of the lower crust can be avoided and whether the lithosphere is strong enough to maintain thickened continental crust. Other important effects that have not been included in this study (e.g. lateral heterogeneities in the concentration in HPEs, the presence of volatiles) can only make the crust and lithosphere even more difficult to stabilize. The calculations above show that, for the average concentration of HPEs in the crust, the temperature in the lower crust exceeds 850 °C and the total strength of the lithosphere drops by one order of magnitude when crustal thickness increases from 40 to 60 km. Crustal differentiation with relative enrichment of the upper crust in radio-elements leads to lower temperatures at the base of the crust. But crustal differentiation alone is not sufficient to maintain the strength of the lithosphere when the crust is thick and has a concentration in HPEs close to the continental crust average (0.8 μW m^{-3}). Unless the average concentration in HPEs is very low (<0.5 μW m^{-3}), the strength of the lithosphere will be comparable to the excess potential energy in regions where the crust reaches 60 km.

5. Conclusions

Averages of continental heat flux and crustal thickness show no positive correlation, which implies that average crustal heat production decreases with crustal thickness. In stable regions, crust thicker than 55 km is extremely rare. In the few examples where the crust is thicker than 55 km, the surface heat flux is very low (<30 mW m^{-2}). Simple calculations show that lower crustal temperatures in a thickened crust would be near melting unless the crust is very differentiated with the HPEs concentrated in the uppermost crust. Regardless of the crustal differentiation, the strength of the lithosphere decreases when the crust is thickened. The strength of the lithosphere drops below the level where it can sustain the increase in gravitational potential energy due to crustal thickening. Survival of a thick crustal root is possible only if the concentration in HPEs and the surface heat flux are very low.

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Appendix A. Rheology and strength of the lithosphere

We have followed the usual procedure to calculate the strength of the lithosphere. Crustal rocks usually deform by power law creep (Ranalli, 1995):

$$\sigma = \frac{\dot{\epsilon}^{1/n}}{A^{1/n}} exp((E + PV^*)/nRT) \tag{A.1}$$

where $\dot{\epsilon}$ is the strain rate, σ the deviatoric stress, A and n are constant characteristics of the material, E is the activation energy, V^* the activation volume, R the gas constant, P the pressure, and T the thermodynamic temperature. We compared the strength with rheological parameters for dry or wet mantle and a wet crust (Table 6). In particular, the rheology of lower crustal samples from the Superior Province, including the Pikwitonei and Kapuskasing granulites, has been studied in the laboratory (Wilks and Carter, 1990). At low temperatures, very large stress is required to maintain steady-state creep, and deformation occurs by frictional sliding on randomly oriented fractures, leading to a linear increase in deviatoric stress with depth known as Byerlee's Law (Brace and Kohlstedt, 1980; Byerlee, 1978). The shear stress τ to overcome friction is proportional to the stress normal to the plane of fracture:

$$|\tau| = f\sigma_n \tag{A.2}$$

where f is the coefficient of friction, and σ_n the effective normal stress (i.e. lithostatic less the fluid pore pressure, usually assumed to be hydrostatic). The coefficient of friction was determined to be 0.85 for σ_n<200 MPa. For horizontal tension, where the maximum principal stress is horizontal and the minimum is vertical, and the dip of normal faults is ≈60°, the deviatoric stress is:

$$\delta\sigma = 0.75\sigma_n \quad \sigma_n < 200\text{MPa} \tag{A.3}$$

$$\delta\sigma = 0.6\sigma_n + 150 \quad \sigma_n > 200\text{MPa}. \tag{A.4}$$

A lower deviatoric stress is needed in extension than in compression. The minimum stress needed to maintain a given deformation rate (typically 10^{-15} s^{-1}) either by frictional sliding or steady state creep defines the yield strength envelope. Vertical integration of this minimum stress gives the total strength of the lithosphere.

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