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Simultaneous inversion of gravity and heat flow data: constraints on thermal regime, rheology and evolution of the Canadian Shield crust☆

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Abstract

Heat flow and gravity are sensitive to crustal composition and thickness. In this paper, we show that heat flow and gravity data, combined with constraints from seismic refraction, can be inverted to obtain the gross composition of the crust. The inversion method is used to determine the crustal structure in different parts of the Canadian Shield. The long wavelength variations in heat flow imply a difference in thermal regime between and within provinces and subprovinces of the Canadian Shield. The crustal heat production determined by inversion is used to construct temperature and rheological profiles for the lithosphere in different parts of the Canadian Shield. Regions enriched in radioelements (the western part of the Abitibi subprovince and the Thompson Belt) had a weak rheology after they stabilized and could not withstand stresses for geologically long times. In contrast, subprovinces that are depleted in radioelements (the eastern part of the Abitibi and the Lynn Lake Belt) had a stronger rheology and might have preserved crustal roots after their formation. © 2002 Published by Elsevier Science Ltd.

1. Introduction

In order to determine the temperature of the continental lithosphere, we need to know the distribution of heat producing radioelements, the thermal conductivity, and the surface heat flow. In general, the surface heat flow is known at sparse sites. Thermal conductivity and heat production can be determined only on available surface or drill core samples. Although it has been known for a very long time that crustal heat production contributes significantly to the continental heat flow

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(e.g. Strutt, 1906), there is still some debate on the respective contributions of crustal heat production and mantle heat flow. Systematic studies have been conducted to determine the average heat production of the main lithologies exposed in the Canadian and other shields (Nicolaysen et al., 1981; Shaw et al., 1986; Fountain et al., 1987; Ashwal et al., 1987; Wedepohl, 1995). These studies have shown that the entire crust contributes to the heat flow, as the heat production of presumably depleted lower crustal samples is $\sim$0.4 μ W m-3. Several recent studies (e.g. Jaupart et al., 1998; Jaupart and Mareschal, 1999) have shown that, in the Canadian Shield, the mantle heat flow is $\approx$10–15 mW m^{-2}, lower than what was previously assumed. In these and similar studies (e.g. Pinet et al., 1991; Guillou et al., 1994), geological and geophysical (in particular gravity) data were used with the surface heat flow to determine the crustal composition and heat production.

If the density of heat flow measurements were sufficient, it would be possible to use heat flow data to determine 2 or 3D crustal models of heat production, in roughly the same manner as crustal density is inferred from gravity. Unfortunately, heat flow measurements are time consuming and it will be impossible to ever obtain the same dense coverage as in gravity studies. Furthermore, heat flow measurements are made in holes of opportunity, drilled mostly for mining exploration purposes, which are extremely unevenly distributed. Finally, because of the geological noise affecting heat flow data, only the long wavelength variations can be determined with some level of confidence. However, there are a number of studies dealing with regional variations in heat flow and the downward continuation of the temperature field (e.g. Cermak and Sass, 1991). Simmons (1967) used direct models to calculate the variations in heat flow due to contrasts in heat generation. Mareschal (1985) developed an inverse method solving the Poisson equation in the frequency domain. This method would permit the calculation of the radioactive sources compatible with 2D-data provided that sufficient data are available to evaluate the Fourier transform. Huestis (1980) applied the Backus–Gilbert theory (Parker, 1994) to determine in 2D the range of crustal temperatures compatible with the heat flow data. The possible distributions of temperature were obtained in a function of constraints on maximal and minimal heat generation through a 35 km thick crust. Beck and Shen (1989) and Shen et al. (1991) also considered the application of inverse methods to infer temperature in the crust. Theoretically, if measurements of heat flow were continuous and covering up different scales, the methods mentioned above would be useful. Unfortunately, both the number of existing boreholes and their depth, limit measurements of heat flow. In the Canadian Shield, that is particularlywell covered, there are less than 200 heat flow sites. Generally, the mean distance between heat flow measurements only allows the determination of long wavelength variations. For this reason, we shall be concerned only with a 1D model.

Numerous seismic studies have shown that the crust is heterogeneous. In the Superior Province of the Canadian Shield, three main lithologies are exposed at the surface: the low grade greenstone assemblages, the granitoids, and the high grade granulitic rocks. With only heat flow data, crustal composition is always under-determined. Gravity data provide an additional constraint because the long wavelength gravity field is dominated by the variations in crustal thickness. The seismic refraction studies have provided a robust estimate of total crustal thickness near some of the heat flow sites. With the combination of gravity and heat flow data and the addition of some seismic data, crustal thickness and composition remain undertermined. Additional assumptions must be introduced to infer a crustal structure. We can rule out some models because they are geophysically implausible. In addition, we shall require the model to be as simple as possible, i.e. the model parameters must vary only when it is required to fit the data.

The objective of this paper is to show how crustal composition can be inferred from heat flow data and how it relates to models of crustal evolution. We present a method for inverting crustal parameters from a combination of heat flow, gravity, and seismic data. We shall apply the inversion to different regions of the the Canadian Shield where heat flow data are available. We will use the results to infer the temperature in the lithosphere, its effect on the rheology, and we shall discuss the implications for the evolution of the crust in the Canadian Shield.

2. Physical model

We are concerned with determining the regional variations in crustal composition from gravity and heat flow data. Gravity data have been low pass filtered. Heat flow data are averaged over areas $1^\circ \times 1^\circ$ ($\approx 110 \times 80$ km). These data are thus representative of the average crustal structure. We may neglect the small scale horizontal variations and consider that the averaged gravity and heat flow are due to a horizontally layered crust (1 dimensional local model). Because we are concerned with variations with wavelengths >200 km, several times the crustal thickness, this approximation is reasonable. For a layered crustal structure, we can consider the variations of either the physical properties or the thickness of the layers. We have used published physical properties (density, heat production) measurements on samples from the main lithologies exposed in the Canadian Shield and assumed that the average physical properties are fixed within each layer. The objective of the inversion will be to determine how the layers' thickness vary between regions.

2.1. Heat flow

The conductive heat flow, $\vec{q}$, is proportional to the gradient of temperature T:

$$\vec{q} = -K\nabla T \tag{1}$$

where the proportionality constant K is the thermal conductivity. At equilibrium, the equation of heat conservation is:

$$\nabla \cdot \vec{q} = A \tag{2}$$

where A is the heat production. If we assume that heat flow varies only vertically,

$$q(z) = q_0 - \int_0^z A(z')\mathrm{d}z' \tag{3}$$

where q_0 is the observed surface heat flow. Note that we use the standard convention that $z>0$ downward but $q>0$ upward. Integrating this equation to the depth of Moho H_M yields:

$$q_\mathrm{M} = q_0 - \int_0^{H_\mathrm{M}} A(z')\mathrm{d}z' \tag{4}$$

where q_M is the heat flow from the mantle. For a crust made up of J layers of constant heat production, we get at the location k:

$$q_0^{(k)} = q_M + \sum_{j=1}^{J} A_j h_j^{(k)} \tag{5}$$

where A_j and $h_j^{(k)}$ are the heat production and the thickness of each layer.

2.2. Gravity

The formulation of the relationship between the average Bouguer gravity anomaly and the vertical density distribution in the crust and mantle is similar to that of the heat flow. For both gravity and heat flow we are using a one dimensional approximation because we are concerned with wavelengths several times the crustal thickness. We shall use a "residual Bouguer anomaly", defined as the difference between the (locally averaged) gravity value and the mean of all the gravity values:

$$g_{\mathrm{res}}^{(k)} = 2\pi G[\rho_{\mathrm{m}}\left(\bar{H} - \sum_{j=1}^{J} h_j^{(k)}\right) + \sum_{j=1}^{J} \rho_j\left(h_j^{(k)} - \bar{h}_j\right)] \tag{6}$$

where the bar indicates the average layer thickness $\bar{h}_j = \left(\sum_{k=1}^{K} h_j^{(k)}\right)/K$. The index k refers to the location ($k = 1,....K$). The average crustal thickness is $\bar{H} = \sum_{j=1}^{J} h_j$. ρ_{m} and ρ_j are the densities of the mantle and j_{th} crustal layer respectively, and G is the gravitational constant. Changes in thicknesses of the layers produce the residual Bouguer anomalies. The first term represents variations of total crustal thickness while the second involves variations of each layer relatively to the mean crust.

If the total crustal thickness at the site k, $H^{(k)}$ is known from seismic studies, we have the additional condition that:

$$H^{(k)} = \sum_{j=1}^{J} h_j^{(k)} \tag{7}$$

The crustal structure remains under-determined when the number of layers is $J > 3$ or when the crustal thickness is not known at all sites. In such cases, we can restrict the range of solutions and/or we can introduce other conditions to select one among all the solutions.

2.3. Additional constraints

High-resolution seismic reflection and refraction profiling studies were conducted in the Canadian Shield by LITHOPROBE (Green et al., 1985; Clowes et al., 1992; Lucas et al., 1993; Lewry et al., 1994; Lacroix and Sawyer, 1995; Nemeth and Hajnal, 1997). The interpretation of these surveys has provided new information about the nature of the crust in the different transects. In the Abitibi subprovince, the seismic velocity models suggest that the crust is made up of three

main lithologies (Calvert and Ludden, 1999; Ludden and Hynes, 2000). In addition, seismic refraction has provided Moho depth which was used whenever available near one the sites.

We have introduced additional constraints to restrict the range to geophysically reasonable solutions. For instance, the total crustal thickness must remain within the range observed in the Shield. When a lithology is observed at the surface at a site, the thickness of the corresponding layer must be >0. The thickness of any layer must be less than the total crustal thickness. We thus have introduced inequality constraints of the type:

$$h_j^{\min} < h_j^{(k)} < h_j^{\max} \tag{8}$$

$$H^{\min} < \sum_{j=1}^{J} h_j^{(k)} < H^{\max} \tag{9}$$

These inequalities restrict the range of the solution but usually are not sufficient to insure a unique solution. Therefore, we shall select among all the solutions the one that minimizes the variation in either the total crustal thickness or the variations in the thickness of all the individual layers. This choice is plausible because seismic investigations show that there is almost norelief on Moho thoughout the Canadian Shield. It also yields the solution that minimizes the stresses due to the crustal heterogeneities.

To minimize the variation in total crustal thickness, we must minimize:

$$\sum_{k=1}^{K}\left(\left(\sum_{j=1}^{J} h_j^{(k)}\right) - \bar{H}\right)^2 \tag{10}$$

where the average crustal thickness $\bar{H}$ is:

$$\bar{H} = \frac{1}{K}\sum_{k=1}^{K}\sum_{j=1}^{J} h_j^{(k)} \tag{11}$$

$h_j^{(k)}$ is the thickness of layer j at site k. Minimizing this functional does not provide additional constraints if the Moho is known everywhere. Alternatively, we can minimize the thickness variations for all the layers. This condition will provide additional constraints even when the Moho's depth is known. In this case, we must minimize:

$$\sum_{j=1}^{J}\sum_{k=1}^{K}\left(h_j^{(k)} - \bar{h}_j\right)^2 \tag{12}$$

Searching for the solution that satisfies a system of linear equations and inequations and minimizes a quadratic functional is the standard problem of quadratic programming. We have used the algorithm developed by Haskell and Hanson (1981) to solve this problem.

The sensitivity of the solutions to errors in the data, and to the selection of the physical properties parameters has been investigated and results are reported by Cheng (1999). In order to verify that the 1D assumption does not produce a large error, we have performed 2D calculations of the heat flow and gravity anomalies. For each of the crustal sections derived by 1D inversion, the differences between the results of 2D direct models and the data were always small (<3 mGal and <2 mW m^{-2} for gravity and heat flow respectively). This is consistent with the assumption that the changes in layer thickness are smooth and produce even smoother changes in the gravity and heat flow fields at the surface. Tests with synthetic data have also been conducted to verify that lateral changes in the average density and crustal heat production of a lithology have a small influence on the solution. Synthetic data were generated with random changes in physical properties (Gaussian distribution around their mean value) and inverted with the uniform physical property approximation. The tests show that a solution reasonably close to the synthetic model will be obtained only if heat production or density constraints remain within 10% of the assumed value (Cheng, 1999).

2.4. Physical properties

Physical properties are assumed constant within each layer. These parameters are determined from all the measurements made on surface and core samples. The main lithologies of the upper and middle crust have been mapped by outcrops in the Canadian Shield: high-grade gneiss and low-grade volcanic–plutonic rocks, probably representing different erosion levels (Hoffman, 1989; Percival and Williams, 1989). Rocks from the lower crust are exposed in crustal sections such as the Pikwitonei–Sachigo and Wawa–Kapuskasing which have been extensively studied. These studies have shown that the upper to middle crust is composed of greenstone belts and tonalitic gneisses extending into the granulite facies of the lower crust (Fountain et al., 1987; Ashwal et al., 1987). All available data of heat production and density measurements, within the Canadian Shield were compiled and values were grouped under different lithological facies (Table 1). Heat generation measurements are made on samples from each of the heat flow sites to assess the local variations in heat production. To estimate the average heat generation of the main lithologies, we have relied on published results of sytematic sampling of the Canadian Shield (Lewis et al., 1984;

Table 1
Heat production (A), number of samples (N_A), density (ρ), and number of samples (N_ρ) for the main lithologies in the Canadian Shield

Lithology	A (μW m^{-3})	N_A	ρ (Mg m^{-3})	N_ρ	References[a]
Gneisses (tonalitic to mafic)	0.3–1.64	114	2.67–2.99	683	1,2,3,4,8,11,12,13
Volcanites	0.09–0.5	54	2.7–3.04	3587	1,2,4,5,8,9,11,12,13
Metasediments	1.1–1.5	7	2.75–2.83	56	1,2,4,8
Gabbro/anorthosite	0.05–0.17	28	2.71–2.99	4003	1,2,4,5,8,11
Granite	2.15–5.22	42	2.63–2.7	26	1,3,6,7,8
Granulite	0.15–0.54	18	2.83–3.2	25	1,2,10

[a] References: (1) Fountain et al., 1987; (2) Ashwal et al., 1987; (3) Jessop and Lewis, 1978; (4) Pinet et al., 1991; (5) Guillou et al., 1994; (6) Drury and Lewis, 1983; (7) Gupta and Grant, 1985; (8) Antonuk and Mareschal, 1993; (9) Fountain et al., 1990; (10) Dion et al., 1992; (11) Telmat et al., 2000; (12) Stettler et al., 1997; (13) Leclair et al., 1997.

Shaw et al., 1986, 1994; Rudnick and Presner, 1990; Gupta and Grant, 1985; Dion et al., 1992). Although the variability of heat production and density is high within each lithology, the long wavelength heat flow and gravity integrate the density and the heat production over a large volume of rocks and the use of average values for physical properties is appropriate. It is well established that felsic rocks are characterized by high heat production and low density. Mafic compositions are usually linked to very low heat production and higher density. Granulites have the highest density and a low heat production. In others words, because variations in gross crustal composition and thickness affect differently the heat flow and the gravity, these two sets of data are complementary.

3. Inversion of gravity and heat flow data in the Canadian Shield

The inversion method was applied to the Abitibi belt, in the Archean Superior Province, and to different belts of the PaleoProterozoic Trans Hudson Orogen of the Canadian Shield (Fig. 1). These two regions were the focus of seismic, geological and other geophysical studies, including heat flow, that were conducted by LITHOPROBE.

The new heat flow data (Mareschal et al., 1999, 2000) were added to the compilation of Jessop et al. (1984). We have used the published interpretation of the seismic refraction data collected by LITHOPROBE. The gravity data come from the Geophysical Data Centre of the Geological Survey of Canada. The gravity data were first interpolated on a $2.5' \times 2.5'$ grid and low-pass filtered to obtain regional trends of the Bouguer anomaly.

3.1. Abitibi sub-province

The Abitibi Greenstone Belt is the largest, in area, Archean greenstone assemblage of greenschist to lower amphibolites facies terrain (Card, 1990). According to surface lithologies it was divided into three main regions (Fig. 2): the Northern Volcanic Zone, the Central Granite–Gneiss Zone and the Southern Volcanic Zone (Ludden and Hubert, 1986). The widespread tonalitic plutons in the central domain have been interpreted as a tectonic window of the underlying crust (Chown and Mueller, 1992; Sawyer and Benn, 1993; Lacroix and Sawyer, 1995).

Several seismic reflection and refraction surveys were conducted perpendicular to the east–west geological strike of the Superior Province, across the Opatica, Abitibi, and Pontiac subprovinces, into the Grenville Province (Fig. 2). The interpretation of the seismic data suggested that: (1) upper crustal greenstone rocks are allochthonous; (2) the midcrust is made up of metasedimentary and igneous rocks, plus imbricated units of unknown affinity; and (3) the lowermost crust, characterized by high velocity anomalies, probably contains mafic lithologies (Calvert and Ludden, 1999).

Sufficient heat flow data were available along an east–west profile in the central zone (profile A–B in Fig. 2) to obtain a crustal cross section of the Abitibi belt. We have used a three layers model to include the main lithologies identified by the geology and the seismic studies: the greenstone assemblages ($<A> \approx 0.2$ μW m^{-3} and $<\rho> \approx 2.8$ Mg m^{-3}) for the upper crust, the tonalitic gneisses ($<A> \approx 1.1$ μW m^{-3} and $<\rho> \approx 2.75$ Mg m^{-3}) for the midcrust, and high grade and mafic rocks ($<A> \approx 0.4$ μW m^{-3} and $<\rho> \approx 2.9$ Mg m^{-3}) for the lower crust.

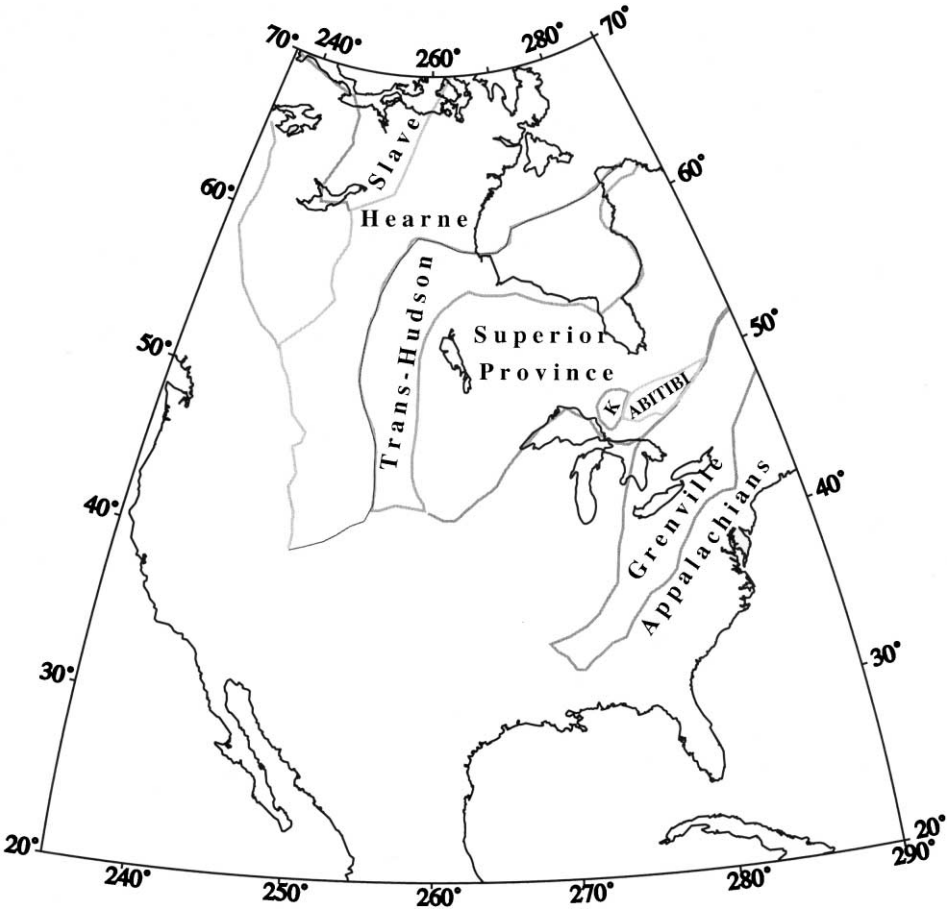

Fig. 1. Map of the Canadian Shield with the main Provinces and the regions referred to in the text. K is the Kapuskasing structural zone. Adapted from Hoffman (1989).

Heat flow and gravity exhibit parallel westward increasing trends (Figs. 3a and b). The mantle heat flow value obtained by inversion ($\approx$13 mW m^{-2}) is consistent with the suggestion that mantle heat flow is low throughout the Canadian Shield (Jaupart et al., 1998). The crustal composition model confirms the trends suggested by previous studies (Pinet et al., 1991; Guillou et al., 1994). In the eastern part of the transect, the low heat flow and gravity require a thick, almost entirely mafic crust. The surface heat flow variations are related to changing crustal composition with an increase in the felsic component toward the west. High heat production, within the mid-crust of the western-central portion of the Abitibi sub-province, indicates a felsic composition suggesting the coexistence of metasedimentary rocks with felsic intrusions. This composition, which predominates southward, is mainly limited between 78o and 81o longitude west. It could represent the composition of imbricated units as observed along seismic reflection and refraction profiles. The crustal thickening near Kapuskasing is consistent with detailed gravity models of the Kapuskasing structure and with some interpretations of the seismic refraction (Atekwana et al., 1994; Bolland and Ellis, 1989).

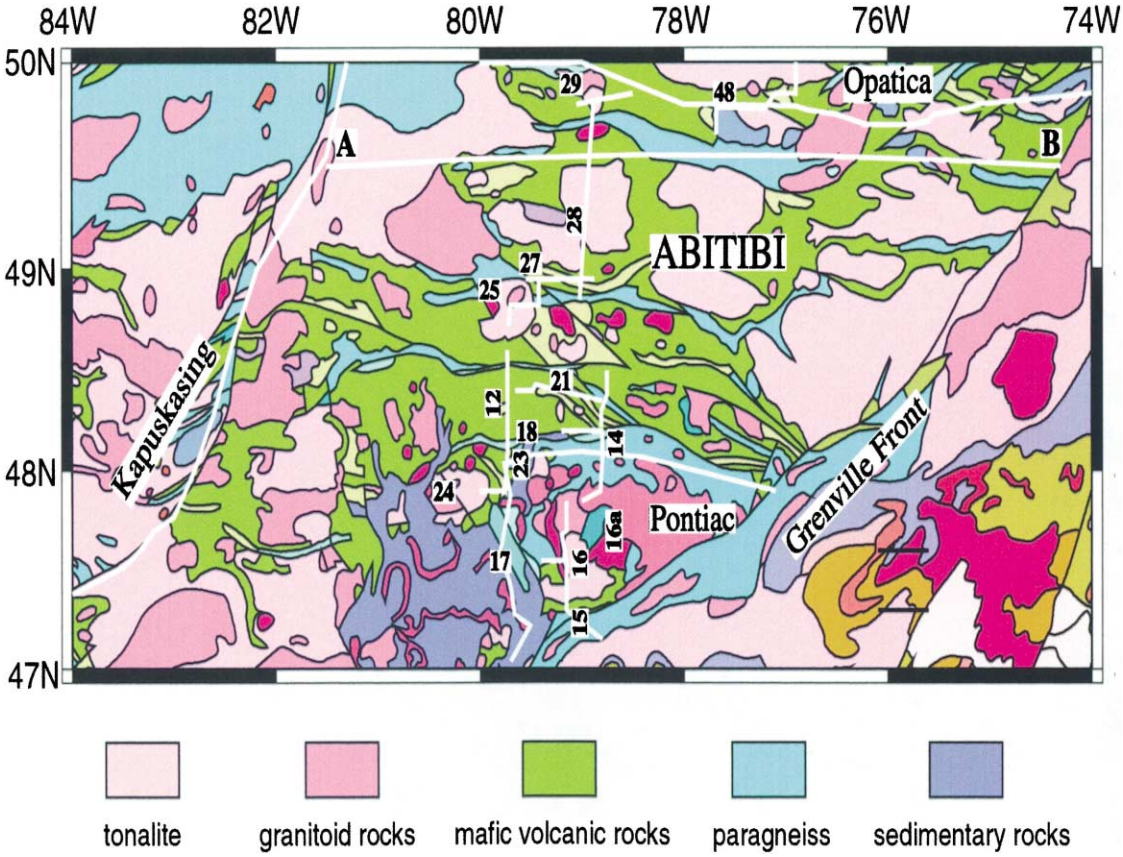

Fig. 2. Geological map of Abitibi sub-province. Adapted from Wheeler et al. (1996). Numbers refer to the seismic reflection and refraction profiles of Lithoprobe. The line A–B shows the inversion profile.

3.2. Trans Hudson Orogen

The Trans Hudson Orogen (THO) resulted from continent–continent collision between the Archean Hearne and Superior cratons (Hoffman, 1989). It has been divided into four principal tectonic domains from east to west (Lewry and Collerson, 1990) (Fig. 4): (1) the Thompson Belt contains interbedded metasedimentary and volcanic rocks on reworked Archean basement; (2) the Reindeer Tectonic Zone is a collage of juvenile intraoceanic rocks; (3) the Wathaman–Chipewyan Batholith is interpreted as an Andean-type continental margin batholith; (4) the Cree Lake Zone is the northwestern hinterland.

The Reindeer tectonic zone consists mostly of arc-derived Proterozoic crust, with small Archean basement windows confirming the existence of a high-grade Archean crust under juvenile volcanic rocks (Lucas et al., 1993). The seismic profiles suggest a broadly symmetric structure with reflections linked to island-arc rocks dipping beneath both bounding Archean cratons (Lucas et al., 1993). Suggestions that the Reindeer zone plunges beneath the Superior (Green et al., 1985) have not been confirmed by the interpretation of the northern seismic profiles (White et al., 1999). Most of the heat flow data come from the Thompson Belt on the former margin of the

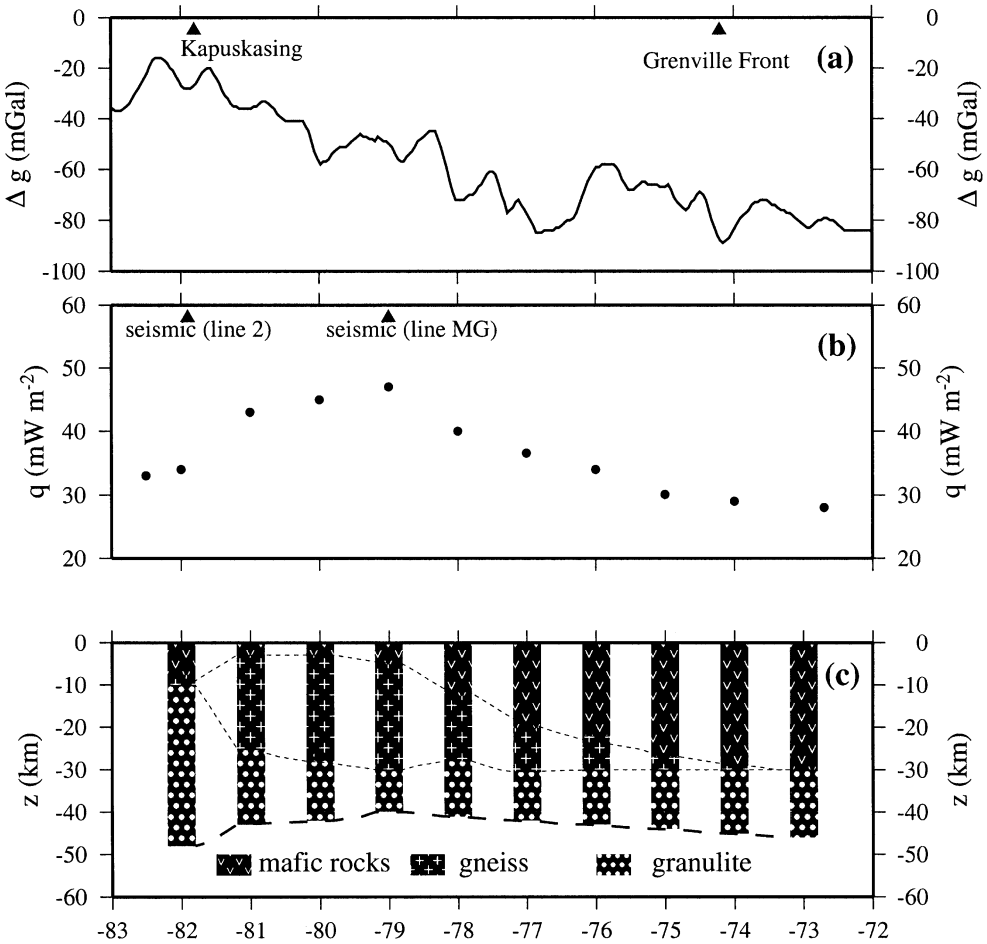

Fig. 3. Gravity, heat flow and crustal cross section across the Abitibi sub-province: (a) filtered Bouguer anomaly, (b) heat flow, and (c) crustal columns derived by inversion.

Superior craton, and from the Flin-Flon and Lynn Lake Belts in the Reindeer tectonic zone. Because these belts have very different compositions that reflect their different origin, it is not possible to invert all the data together with the same set of physical parameters as we did in the Abitibi. Furthermore, there appear to beshort wavelength variations in heat flow (notably near Thompson and in the Flin-Flon Belt). After filtering out these short wavelength variations, we were able to propose crustal models for the Flin-Flon, Thompson, and Lynn Lake belts (Fig. 5).

3.2.1. Flin-Flon Belt

The Flin-Flon Belt is made up of volcanic rocks of oceanic affinity. The surface heat flow does not vary much within this Belt. One exceptional heat flow anomaly (the West Arm site where the value is 51 mW m^{-2}) is located within a short distance of standard values (42 mW m^{-2}), suggesting that it is due to shallow and local sources (Mareschal et al., 1999). Density measurements on samples from the Flin-Flon Belt show that these juvenile rocks are denser (2.84 Mg m^{-3}) than

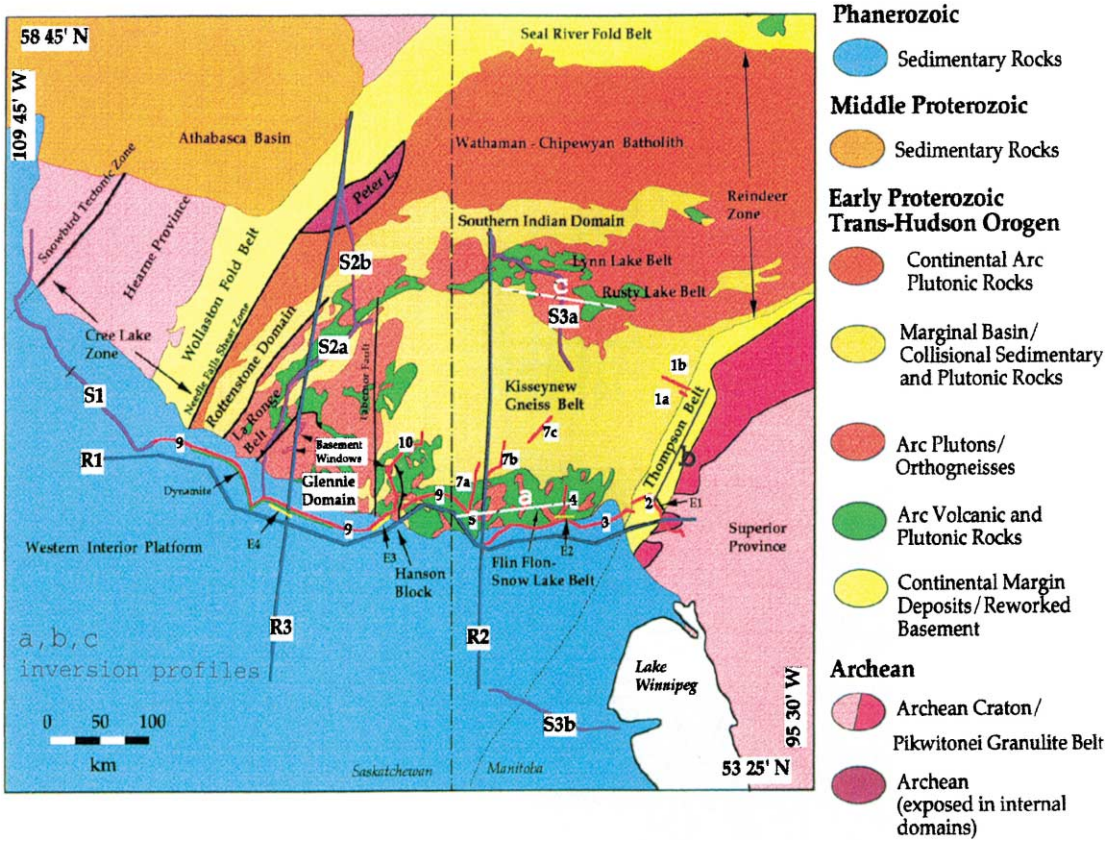

Fig. 4. Geological map of Trans Hudson Orogen. Numbers refer to the seismic reflection lines. The three refraction lines aremarked by R1, R2 and R3. Adapted from White et al. (1999). Three profiles marked a, b and c were inverted independently.

Archean volcanic rocks in the Abitibi sub-province (Leclair et al., 1997), but heat production is similar (Cheng, 1999). The mean density of gneisses is also higher (2.8 Mg m^{-3}) than that of the gneisses in the Abitibi. We have inverted the heat flow and gravity data in the Flin-Flon Belt with a three layer model (Fig. 5a). The juvenile volcanic layer attenuates to the east towards the Thompson Belt. The midcrustal layer with higher heat production (1.1 μW m^{-3}) suggests a composition of felsic gneisses that could be underthrust Archean basement or arc derived gneisses as those exposed within the Hanson Lake region. Physical parameters of the lower crust are similar to those of the granulite facies rocks.

3.2.2. Thompson Belt

This narrow tectonic zone is characterized by average heat flow higher than the Shield (53 vs 42 mW m^{-2}) and a relative Bouguer anomaly low (−50 mGal). Prior interpretations of seismic and electrical resistivity along the southern transect suggested that rocks of the Reindeer Zone dip beneath the Thompson Belt at the middle to lower crustal depth (Green et al., 1985; Bickford et al., 1990; Bleeker et al., 1990). More recent seismic studies along a northern transect do not show the same prominent east dipping reflectivity in the middle to lower crust (White et al., 1999).

The heat flow varies from normal shield values (42 mW m^{-2}) at the southern end of the profile to much higher values (52 mW m^{-2}) near Thompson. To this long wavelength (>100 km) trend, is superposed near Thompson, a short wavelength (<20 km) heat low anomaly, which we have not considered. For this belt where there are no volcanic rocks exposed at the surface, we have used a two layer model with the topmost layer of felsic gneisses with the average heat production and density of outcrop and core samples. The lower crustal layer is assumed to have the physical properties of granulites. Inversion of heat flow and gravity data suggests differences in bulk radioactivity and density between the northern and southern parts of the Thompson Belt (Fig. 5b). The inversion suggests that bulk crustal composition becomes dominantly felsic, possibly because of a higher concentration of Archean metasediments, near the northern end of the profile.

3.2.3. Lynn Lake Belt

Some of the lowest heat flow values (<25 mW m^{-2}) in the Canadian Shield have been reported in the Lynn Lake Belt where juvenile volcanic rocks are exposed. The Lynn Lake Belt has the lowest average heat flow in the THO (32 mW m^{-2}) although the average heat production of core samples (0.7 μW m^{-3}) is not exceptionally low and is higher than that of the Flin-Flon Belt. Interpretation along seismic refraction profiles R2 shows that there is some crustal thickening at the boundary between the Kisseynew Basin and the Lynn Lake Belt (Németh and Hajnal, 1997). We have used a two-layer model (Fig. 5c), with the physical parameters of the topmost layer determined by those of surface samples (i.e. heat production 0.75 μW m^{-3}) and granulitic for the lower crust. The low heat flow implies that there is no felsic intermediate layer as beneath the Flin-Flon Belt. The upper crustal layer is thin and covers a depleted lower crust.

4. Rheology of the Canadian Shield

Changes in crustal radioactivity affect the temperature regime. The more radioactive the crust or the thicker the crust, the higher the temperature is at Moho. For an average shield crust, 40 km thick with mean heat production 0.7–0.8 μW m^{-3} and heat flow of 41 mW m^{-2}, it is ≈400 °C. Higher heat production or a thicker crust result in higher temperature in the lower crust and at Moho. From the heat flow and distribution of radioelements, we have calculated temperature profiles for different regions in the Canadian Shield (Fig. 6). The calculations indicate differences in the present mantle temperature below the different belts. Higher temperatures are found beneath the western Abitibi and the Thompson Belt, lower temperatures are beneath the Lynn Lake Belt and the eastern Abitibi. One should note that the Thompson Belt is narrow and that, because of lateral variations, the temperature in the lowermost crust and mantle may be lower than suggested by this calculation. Because crustal heat production was higher in the past, the surface heat flow and lithospheric temperature were also higher (e.g. West and Mareschal, 1979). We have tried to estimate the temperature after the crust became stable (i.e. at 2.5 Ga for the Superior Province, and at 1.8 Ga for the Trans Hudson Orogen and for the Kapuskasing structure where the last tectonic event was the uplift ca 1.8 Ga). We calculated backwards the radioactive decay equations for the relevant radioisotopes to obtain past crustal heat production. We have assumed that the mantle heat flow was the same as at present. The estimated profiles in Figs. 6c and d give the temperature in the crust after it returned to thermal equilibrium following the

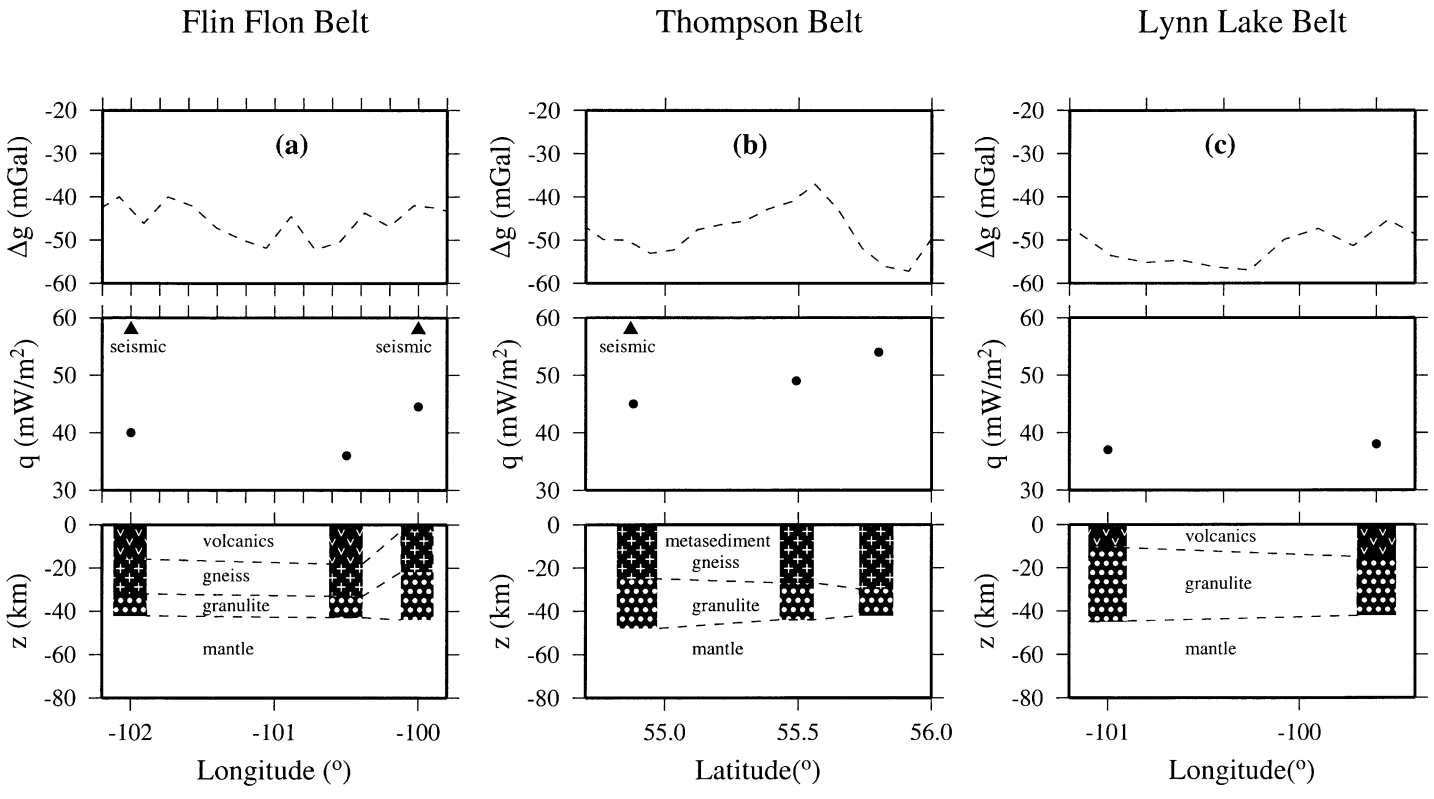

Fig. 5. Crustal columns obtained by inversion for the different belts in the Trans Hudson Orogen: (a) Flin-Flon Belt, (b) Thompson Belt (c) Lynn Lake Belt.

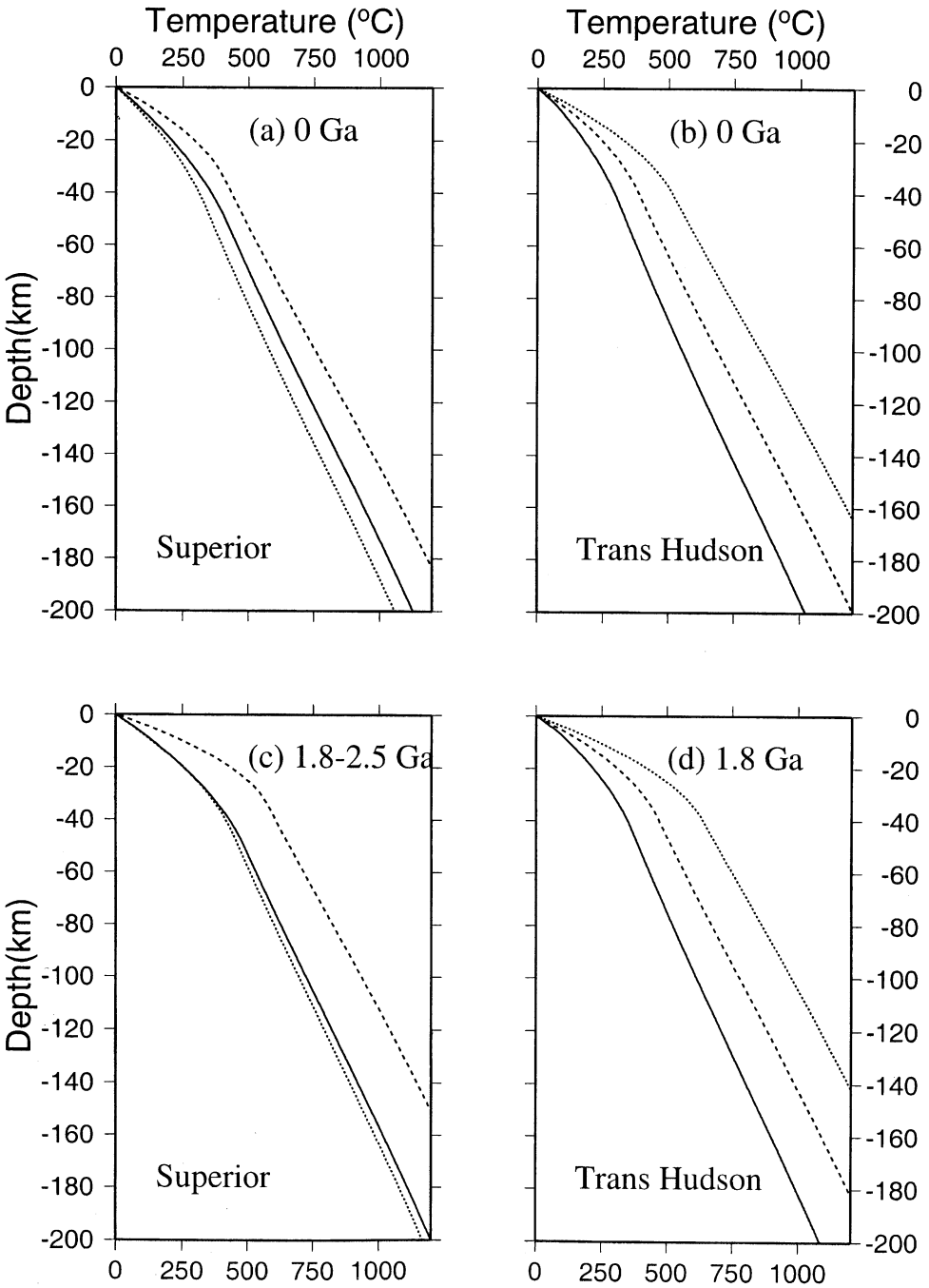

Fig. 6. Temperature profiles in selected regions of the Canadian Shield: (a) present temperature profiles for regions of the Superior Province (Kapuskasing structural zone (full line), western Abitibi (dashed line), eastern Abitibi (dotted line); (b) present temperature profiles for the Flin-Flon (dashed line), Thompson (dotted line) and Lynn Lake (full line) Belts of the Trans Hudson Orogen; (c) steady state temperature profiles after crustal stabilization (1.8 Ga for the KSZ and 2.5 Ga for the rest of the Abitibi; (d) steady state temperature profiles for the TransHudson Orogen after crustal stabilization at 1.8 Ga.

last tectonic event. Because of possibly higher mantle heat flow and transients from the last tectonic event, temperature immediately after stabilization might have been higher than equilibrium.

Crustal rocks are deformed either by brittle fracturing or by ductile creep. The stress required for brittle fracturing depends only on the effective pressure (ratio of pore fluid pressure to overburden pressure). The fracture is usually of the shear Coulomb type (Byerlee, 1978), and the frictional shear fracture criterion is (Sibson, 1974):

Table 2
Geothermal and rheological parameters for different regions of the Canadian Shield: surface heat flow $<Q>$, Tectonic age, layer thickness h, present heat production $<A_0>$, heat production after crustal stabilization $<A_T>$, thermal conductivity K, rheological parameters C, exponent n and activation enthalpy ΔH. For brittle failure, we have used Eq. (13) with $\alpha = 0.75$ and a hydrostatic pore fluid pressure ratio 0.33

	$<Q>$ (mW m^{-2})	Age (Ga)	h (km)	$<A_0>$ (μW m^{-3})	$<A_T>$ (μW m^{-3})	K (W m^{-1} K^{-1})	$\log_{10}$ C (MPa^{-n} s^{-1})	n	ΔH (kj mol^{-1})
Thompson	54	1.8	25	1.2	1.8	3.0	−6.0	2.8	150[a]
			23	0.4	0.6	2.5	−2.5	3.2	270[b]
Flin-Flon	42	1.8	15	0.3	0.4	3.0	−3.0	3.0	230[c]
			15	1.2	1.8	3.0	−6.0	2.8	150[a]
			12	0.4	0.6	2.5	−2.5	3.2	270[b]
Lynn Lake	32	1.8	5	0.7	1.1	3.0	−3.0	3.0	230[c]
			45	0.4	0.6	2.5	−2.5	3.2	270[b]
East Abitibi	31	2.5	15	0.2	0.4	3.0	−3.0	3.0	230[c]
			7	1.1	2.2	3.0	−6.0	2.8	150[a]
			20	0.4	0.8	2.5	−2.5	3.2	270[b]
West Abitibi	42	2.5	4	0.2	0.4	3.0	−3.0	3.0	230[c]
		2.5	20	1.1	2.2	3.0	−6.0	2.8	150[a]
		2.5	17	0.4	0.8	2.5	−2.5	3.2	270[b]
Kapuskasing	33	1.8	52	0.4	0.6	2.5	−2.5	3.2	270[b]

The rheology of the mantle is that of dunite (Chopra and Paterson, 1981).
[a] Rheology of Westerly granite (reported in Carter and Tsenn, 1987).
[b] Rheology of Pikwitonei granulites (Wilks and Carter, 1990).
[c] Rheology of metagabbro (Wilks and Carter, 1990).

Table 3
Strength of the lithosphere in the Canadian Shield, at present and after crustal stabilization

	Age (Ga)	Present strength (10^{12} N m^{-1})	Initial strength (10^{12} N m^{-1})
East Abitibi	2.6	71	49
West Abitibi	2.6	67	24
Kapuskasing	1.8	70	52
Flin-Flon	1.8	60	43
Thompson	1.8	32	15
Lynn Lake	1.8	95	78

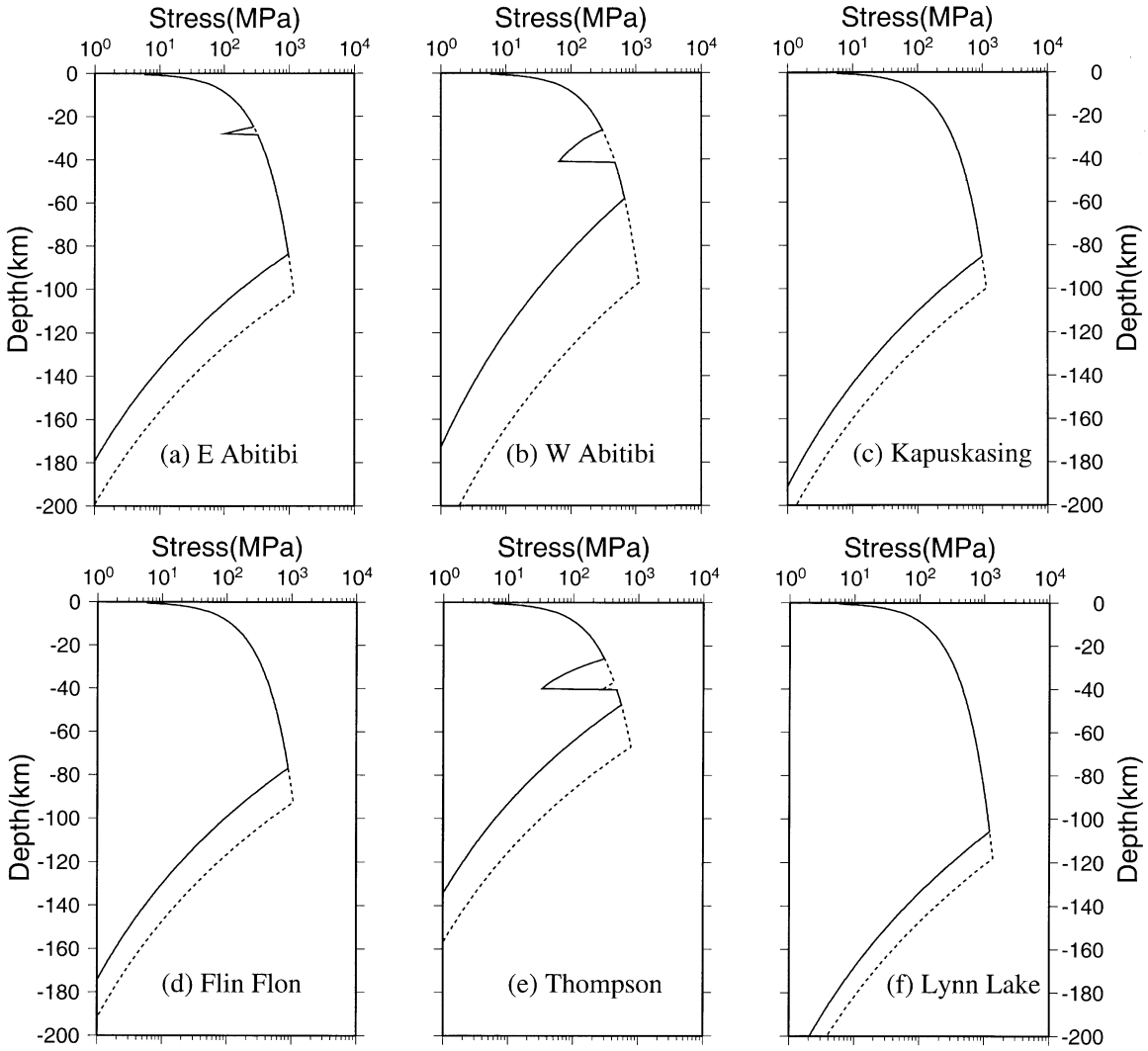

Fig. 7. Rheological profiles in selected regions of the Canadian Shield at present (dashed line) and after crustal stabilization (continuous line): (a) eastern Abitibi, (b) western Abitibi, (c) Kapuskasing structure, (d) Flin-Flon Belt, (e) Thompson Belt, (f) Lynn Lake Belt. Rheological parameters are given in Table 2.

$$(\sigma_1 - \sigma_3) \geqslant \alpha \rho g z (1 - \lambda) \tag{13}$$

where σ_1 and σ_3 are the maximum and minimum principal stress, g the acceleration of gravity, z the depth, ρ the average density of rocks above z, λ is the pore fluid factor (ratio of pore fluid to lithostatic pressure), and α is a numerical coefficient which, for a friction coefficient of 0.75, is equal to 3, 1.2 and 0.75 for thrust, strike-slip and normal faulting respectively.

For ductile creep, the rate of deformation $\dot{\varepsilon}$ depends (non linearly) on the stress σ, and the temperature T. The creep law can be written as:

$$\sigma = \left(\frac{\dot{\varepsilon}}{C}\right)^{\frac{1}{n}} \exp(\Delta H/nRT) \tag{14}$$

where ΔH is the activation enthalpy, and R is the gas constant. Changes of composition in the crust as well as in the mantle produce a rheologically layered lithosphere. The profiles of strength represent the stress needed to maintain a strain rate of 10^{-15} s^{-1} by ductile creep orbrittle fracture, whichever is less. For brittle failure, we have assumed $\alpha = 0.75$ and hydrostatic pore fluid pressure ($\lambda = 0.33$). The lithospheric temperature profiles (Fig. 6) have been used to determine the lithospheric strength in extension at present and at thermal steady state after crustal stabilization (Fig. 7). The thermal and rheological parameters used for these calculations are given in Table 2.

To characterize the total strength of the lithosphere, we have integrated vertically the strength of the lithosphere. This integrated strength can be viewed as the extra potential energy that the lithosphere can accumulate without an extensional collapse. Table 3 shows that presently there is a factor of 3 difference between the weakest lithosphere (Thompson) and the strongest (Lynn-Lake). Such a marked rheological contrast due to differences in thermal regime within the trans Hudson Orogen has also been suggested by studies of the elastic lithospheric thickness which show the juxtaposition in central Canada of strong and weak regions (Wang and Mareschal, 1999). The crust seems to be brittle throughout the Canadian Shield today and even for the Thompson Belt there is no weak lower crustal layer.

Rheological contrasts were even larger when the crust became stable and the western Abitibi (Fig. 7a) and the region of Thompson (Fig. 7b) had a weak and ductile lower crust at 1.8 Ga. In the Superior province, before the uplift, the Kapuskasing region was comparable to the western Abitibi sub-province where a weak lower crust was present between 2.5 and 2.0 Ga. A rheologically stratified crust favoured the formation of the midcrustal decollement during the uplift of the Kapuskasing structure (Fig. 7a). The Kapuskasing structure appears to have been strong after thermal equilibration at 1.8 Ga. The lithosphere in the eastern Abitibi (Fig. 7a) was strong and without a weak lower crust from the time of crustal stabilization at 2.5 Ga.

5. Conclusions

The study has shown the usefulness of combining heat flow and gravity data to infer crustal composition.

There are strong contrasts in thermal regime within the Canadian Shield with some regions with low (28–32 mW m^{-2}) and regions with high (>50 mW m^{-2}) heat flow.

The contrast in thermal regime implies differences in lithospheric strength. After crustal stabilization, a weak and ductile lower crust was present in regions with high heat production and heat flow such as the western Abitibi and the Thompson Belt.

The two regions (Lynn Lake and eastern Abitibi) where heat flow is low and the lithosphere is now strongest did not have a ductile lower crust after they stabilized. In these regions, where a crustal root has been preserved, the lithosphere could sustain the increased potential energy of crustal thickening.

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