

# Lithospheric structure of the Canadian Shield inferred from inversion of surface-wave dispersion with thermodynamic *a priori* constraints

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**Abstract:** We argue for and present a reformulation of the seismic surface-wave inverse problem in terms of a thermal model of the upper mantle and apply the method to estimate lithospheric structure across much of the Canadian Shield. The reformulation is based on a steady-state temperature model, which we show to be justified for the studied region. The inverse problem is cast in terms of three thermal parameters: temperature in the uppermost mantle directly beneath Moho, mantle temperature gradient, and the potential temperature of the sublithospheric convecting mantle. In addition to the steady-state constraint, prior physical information on these model parameters is based on surface heat flow and heat production measurements, the condition that melting temperatures were not reached in the crust in Proterozoic times and other theoretical considerations. We present the results of a Monte Carlo inversion of surface-wave data with this 'thermal parameterization' subject to the physical constraints for upper mantle shear velocity and temperature, from which we also estimate lithospheric thickness and mantle heat flux. The Monte Carlo inversion gives an ensemble of models that fit the data, providing estimates of uncertainties in model parameters. We also estimate the effect of uncertainties in the interconversion between temperature and seismic velocity. Variations in lithospheric temperature and shear velocity are not well correlated with geological province or surface tectonic history. Mantle heat flow and lithospheric thickness are anti-correlated and vary across the studied region, from 11 mW/m<sup>2</sup> and nearly 400 km in the northwest to about 24 mW/m<sup>2</sup> and less than 150 km in the southeast. The relation between lithospheric thickness and mantle heat flow is consistent with a power law relation similar to that proposed by Jaupart *et al.* (1998), who argued that the lithosphere and asthenosphere beneath the Canadian Shield are in thermal equilibrium and heat flux into the deep lithosphere is governed by small-scale sublithospheric convection.

Studies of the structure, composition and evolution of Precambrian continental lithosphere are the foundation of the current understanding of the processes that have shaped the growth and long-term stability of the continents. The thermal structure of Precambrian continental lithosphere has been studied principally with three methods: inversion of surface heat flow measurements (e.g. Nyblade & Pollack 1993; Pollack *et al.* 1993; Jaupart *et al.* 1998; Jaupart & Mareschal 1999; Nyblade 1999; Artemieva & Mooney 2001), geothermobarometry of mantle xenoliths (e.g. a recent review by Smith 1999), and seismic tomography (e.g. Furlong *et al.* 1995; Goes *et al.* 2000; Röhm *et al.* 2000). Each of these

methods has distinct strengths and limitations. Thermobarometry of mantle xenoliths is probably most directly related to deep thermal structure, but high-quality xenoliths are rare and those that are available have recorded the temperature of their formation which may not be representative of the current thermal regime of the lithosphere. In contrast, heat flow measurements directly reflect the recent lithospheric thermal regime, but their ability to resolve the deep structure of the continental lithosphere is limited. Inverting surface heat flow for the mantle geotherm requires strong *a priori* assumptions about the thermal state of the lithosphere and the distribution of heat sources. Seismic data

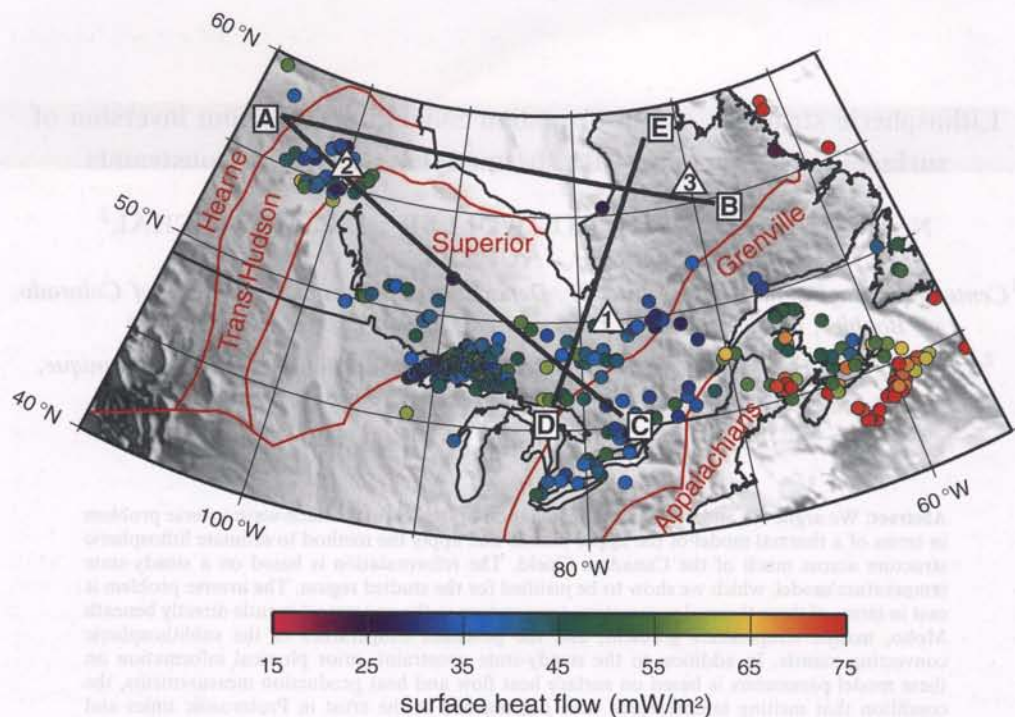


Fig. 1. Reference map of eastern Canada showing the heat flow measurements used in this study as well as the locations of the 1-D (Spatial Points 1, 2, 3) and 2-D profiles (A–B, A–C, D–E) referred to in the study. Red lines, boundaries of principal Precambrian provinces. Adapted from Hoffman (1989).

are directly sensitive to the present deep structure of the lithosphere, but vertical resolution is limited and substantial uncertainties remain in the conversion from seismic velocity to temperature, resulting particularly from ignorance of mantle composition and anelasticity.

The limitations of each of these methods individually lead naturally to exploiting them in combination. For example, Rudnick & Nyblade (1999) describe constraints on the Archean lithosphere that derive from applying xenolith thermobarometry and surface heat flow measurements simultaneously. Shapiro & Ritzwoller (2004) discuss the use of surface heat flux as an *a priori* constraint on inversions of seismic surface-wave dispersion data. The heat flow measurements are used to establish upper and lower temperature bounds in the uppermost mantle directly beneath the Moho discontinuity. The temperature bounds are then converted to bounds on seismic velocity using the method of Goes *et al.* (2000). This approach can improve seismic models beneath continents, particularly beneath cratons and continental platforms, and tighten constraints on mantle temperatures. Shapiro & Ritzwoller also describe an additional thermo-

dynamic constraint that involves replacing *ad hoc* seismic basis functions with a physical model of the thermal state of the upper mantle which is intrinsically a function of temperature. They applied this procedure to the oceanic upper mantle where the thermodynamic model consisted of a shallow conductive layer underlain by a convective mantle. They argued that the constraint produces more plausible models of the oceanic lithosphere and asthenosphere and reduces the uncertainty of the seismic model while negligibly degrading the fit to the seismic data.

This study is an extension of the results of Shapiro & Ritzwoller (2004) in two principal respects. First, Shapiro & Ritzwoller applied heat flow measurements as a prior constraint on seismic inversions only at a few isolated points to test the concept. In the present study, we apply the joint inversion over a wide region of North America, principally in the Canadian Shield, where high-quality heat flow measurements are available and the lithosphere is likely to be in thermal equilibrium. The locations of heat flow measurements used in this study are shown in Figure 1. We refer to these results as deriving from the 'seismic parameterization',

because models are constructed in seismic velocity model space (Shapiro & Ritzwoller 2002). The heat flow constraints are converted to seismic velocities from temperature model space.

Heat flow measurements, however, do not cover the entire Canadian Shield, but are clustered mostly in southern Canada. To apply the heat flow constraint broadly across the Canadian Shield, therefore, would require us either to extrapolate existing measurements to other regions or to apply physical constraints derived from the regions where heat flow measurements exist. Here, we use the latter approach as the second extension of the results of Shapiro & Ritzwoller. Based on inversions in regions where heat flow measurements exist, we argue that the uppermost mantle beneath much of the studied region is likely to be in thermal steady state; i.e. the lithosphere is neither heating nor cooling and the surface heat flow is the sum of the heat entering the base of the lithosphere and the heat production in the crust. We reformulate the inverse problem in terms of a physical model of the thermal state of the upper mantle, in which a lithosphere in thermal steady state overlies a convecting mantle. Models are constructed first in temperature model space and are tested to ensure that they satisfy the steady-state constraint, surface heat flow data (within uncertainties), and bounds on the mantle component of heat flow (discussed later). Temperature profiles that satisfy these constraints are converted back to seismic model space where a seismic crust is introduced and the resulting model is tested to see if it fits the seismic data acceptably. We refer to these results as deriving from the 'thermal parameterization'. Figure 2 presents a schematic outline of the method based on the thermal parameterization. We apply this method to estimate the seismic and temperature structure of much of the Canadian Shield, including the mantle component of heat flux and lithospheric thickness.

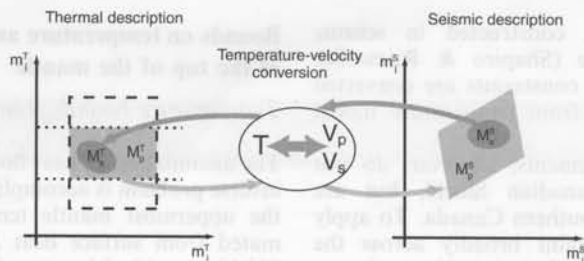
We begin by discussing the temperature bounds applied on the models based both on the seismic and thermal parameterizations. We also discuss uncertainties in the interconversion between temperature and seismic shear velocity. The joint inversion of surface-wave dispersion and heat flow with the seismic parameterization is then introduced and described followed by the inversion based on the thermal parameterization with the steady-state heat flow constraint. Finally, we discuss the results of the inversion with the thermal parameterization, including estimates of lithospheric thickness and the mantle component of heat flow.

## Bounds on temperature and seismic velocities at the top of the mantle

### *Temperature bounds from heat flow*

The assimilation of heat flow data in the seismic inverse problem is accomplished by constraining the uppermost mantle temperatures,  $T_m$  estimated from surface heat flow. The Canadian Shield is an ideal location for the first extended application of this method for two reasons. Firstly, the Canadian heat flow data (Fig. 1) are of exceptional quality because, for example, heat flow has been measured using several deep neighbouring boreholes in many cases (e.g. Jessop *et al.* 1984; Drury 1985; Drury & Taylor 1987; Drury *et al.* 1987; Mareschal *et al.* 1989, 1999, 2000; Pinet *et al.* 1991; Guillou *et al.* 1994; Hart *et al.* 1994; Guillou-Frottier *et al.* 1995, 1996; Rolandone *et al.* 2002) and there is an extensive dataset of crustal heat production measurements (see Jaupart & Mareschal 2004). Secondly, as discussed in detail by Shapiro & Ritzwoller (2004), the joint inversion of seismic data and heat flow is most straightforward for cold lithosphere such as that found beneath Precambrian regimes. This is because uncertainties in the anelastic correction are smallest, which is part of the interconversion between temperature and seismic shear velocity. In addition, the volatile content, which can also affect the conversion to temperature, is believed to be small beneath ancient cratons due to the efficiency of partial melting upon their formation (e.g. Pollack 1986).

Even in the best of cases, however, estimating temperatures at the top of the mantle requires determining the crustal geotherm, which depends on thermal conductivity and on the distribution of crustal heat production. Several methods have been used to determine radioactive heat production in the Canadian crust (e.g. Jaupart *et al.* 1998; Jaupart & Mareschal 1999) and they show that a simple linear relation between surface heat flow and crustal heat production is invalid because, in many terranes, such as the greenstone belts, heat production is lower in the upper crust than in the mid-crust. Lower crustal heat production is believed to be relatively homogeneous ( $\sim 0.4 \mu\text{W}/\text{m}^3$ ). Based on a simultaneous Monte Carlo inversion of heat flow and gravity data across the Abitibi Belt, Guillou *et al.* (1994) proposed that the mantle component of heat flow lies between 7 and  $15 \text{ mW}/\text{m}^2$ . It has been suggested (e.g. Jaupart *et al.* 1998; Jaupart & Mareschal 1999) that such low values for the heat flow from the mantle are characteristic of most of the



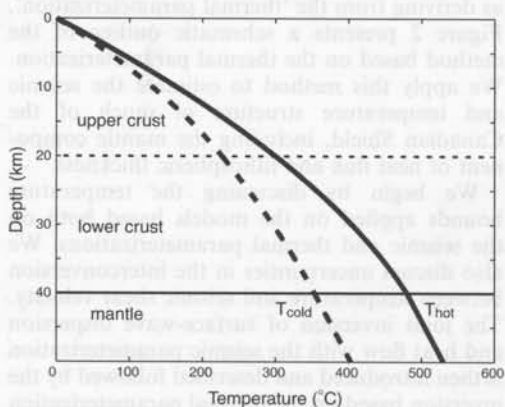
**Fig. 2.** Schematic representation of the Monte Carlo seismic inversion based on a thermal parameterization with *a priori* constraints. The thermal parameterization (left panel) is constrained by the heat flow data (horizontal dotted lines) where they exist, the steady-state constraint on the thermal structure of the mantle (dashed rectangle), and a lower bound on mantle heat flow (not shown). These constraints delimit the range of physically plausible thermal models  $M_p^T$  (light shaded area on left panel). Using a temperature–seismic velocity conversion, this range is converted into a range of physically plausible seismic models  $M_p^S$  (light shaded area on right panel) to which a range of crustal seismic models and radial anisotropy are added. Random sampling within  $M_p^S$  identifies the ensemble of acceptable seismic models  $M_a^S$  (dark shaded area on right panel). Finally, the seismic crust is stripped off and this ensemble is converted back into the ensemble of acceptable temperature models  $M_a^T$  (dark shaded area on left panel).

Canadian Shield. The range was further narrowed down by Rolandone *et al.* (2002) who argued that mantle heat flow cannot be lower than  $11 \text{ mW/m}^2$  for the crust to have stabilized. We will use the assumption that heat flow from the mantle is relatively low and homogeneous across the region of study to compute crustal geotherms by solving the steady-state heat equation with different models of the distribution of heat production in the crust. The major cause for the uncertainties on the estimated temperature in the upper mantle is the limited knowledge of crustal heat production.

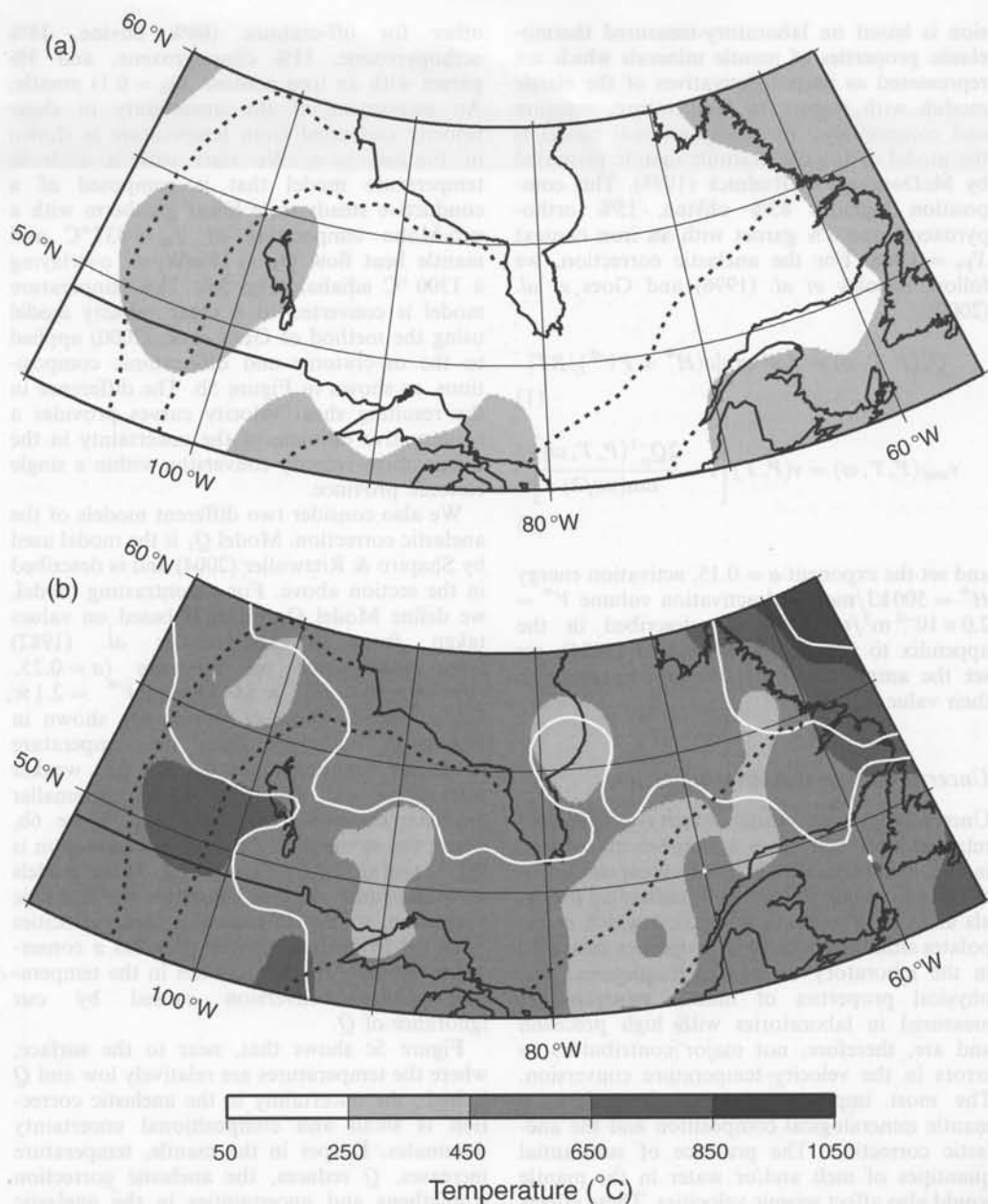
To bound temperatures in the uppermost mantle we consider two end-member models of crustal heat production. The lower bound is set by a two-layer crust that is consistent with a ‘cold’ uppermost mantle with temperature  $T_{\text{cold}}$ . Heat production of  $0.4 \mu\text{W/m}^3$  is used in the 20 km-thick lower crust, and upper crustal heat production is adjusted to match the measured surface heat flow with a constant mantle heat flow of  $15 \text{ mW/m}^2$ . We fix crustal thickness for the Canadian Shield at 40 km here (e.g. Perry *et al.* 2002) and crustal thermal conductivity at  $3.0 \text{ W/m/K}$ . A second, single-layer crustal model that produces a ‘hot’ uppermost mantle with temperature  $T_{\text{hot}}$  was constructed by assuming that heat production is uniform throughout the crust with a value adjusted to the same constant mantle heat flow of  $15 \text{ mW/m}^2$ . The crustal thermal conductivity of  $2.5 \text{ W/m/K}$  is used for the ‘hot’ model. The resulting two crustal geotherms for a surface heat flux of  $45 \text{ mW/m}^2$  are shown in Figure 3.

In regions where the density of heat flow measurements is high, we use these bounds on

uppermost mantle temperatures:  $T_{\text{min}} = T_{\text{cold}}$  and  $T_{\text{max}} = T_{\text{hot}}$ . In regions away from heat flow measurements, however, we widen the temperature range by increasing the upper bound on temperature to  $T_{\text{max}} = T_{\text{hot}} + (T_{\text{hot}} - T_{\text{cold}})$ , but retain  $T_{\text{min}} = T_{\text{cold}}$  because this lower bound is already very low. These bounds are varied spatially in a smooth way. Figure 4 displays the spatial variation of these temperature bounds in the uppermost mantle. The temperature limits  $T_{\text{min}}$  and  $T_{\text{max}}$  are sufficiently different to account for uncertainties in the crustal thermal parameters but still provide useful constraints on the seismic inver-



**Fig. 3.** End-member crustal models that define  $T_{\text{cold}}$  and  $T_{\text{hot}}$  at the top of the mantle for a surface heat flux of  $45 \text{ mW/m}^2$ . For both models, the same values are assumed for the mantle heat flow and thermal conductivity.



**Fig. 4.** Uppermost mantle (sub-Moho) temperature bounds inferred from the heat flow data. (a) Lower bound  $T_{\min}$ . (b) Upper bound  $T_{\max}$ . The white contour in (b) encircles the points within 200 km of a heat flow measurement. Temperature bounds are tightest near heat flow measurements. Dashed lines, boundaries of principal Precambrian provinces. Adapted from Hoffman (1989).

sion, as demonstrated by the results below. As a final step, we interpolate the temperature bounds,  $T_{\min}$  and  $T_{\max}$  onto the same  $2^\circ \times 2^\circ$  geographical grid on which the surface-wave dispersion maps are defined.

#### *Interconversion between temperature and seismic velocity*

We convert temperature to shear velocity using the method of Goes *et al.* (2000). This conver-

sion is based on laboratory-measured thermoelastic properties of mantle minerals which are represented as partial derivatives of the elastic moduli with respect to temperature, pressure and composition. The compositional model is the model of the old cratonic mantle proposed by McDonough & Rudnick (1998). This composition includes 83% olivine, 15% orthopyroxene, and 2% garnet with an iron content  $X_{\text{Fe}} = 0.086$ . For the anelastic correction, we follow Sobolev *et al.* (1996) and Goes *et al.* (2000):

$$Q_{\mu}(P, T, \omega) = A\omega^a \exp[a(H^* + PV^*)/RT] \quad (1)$$

$$v_{\text{anel}}(P, T, \omega) = v(P, T) \left[ 1 - \frac{2Q_{\mu}^{-1}(P, T, \omega)}{\tan(\pi a/2)} \right] \quad (2)$$

and set the exponent  $a = 0.15$ , activation energy  $H^* = 500$  kJ/mol, and activation volume  $V^* = 2.0 \times 10^{-5}$  m<sup>3</sup>/mol but, as described in the appendix to Shapiro & Ritzwoller (2004), we set the amplitude  $A = 0.049$  in contrast with their value of 0.148.

### Uncertainties in the interconversion

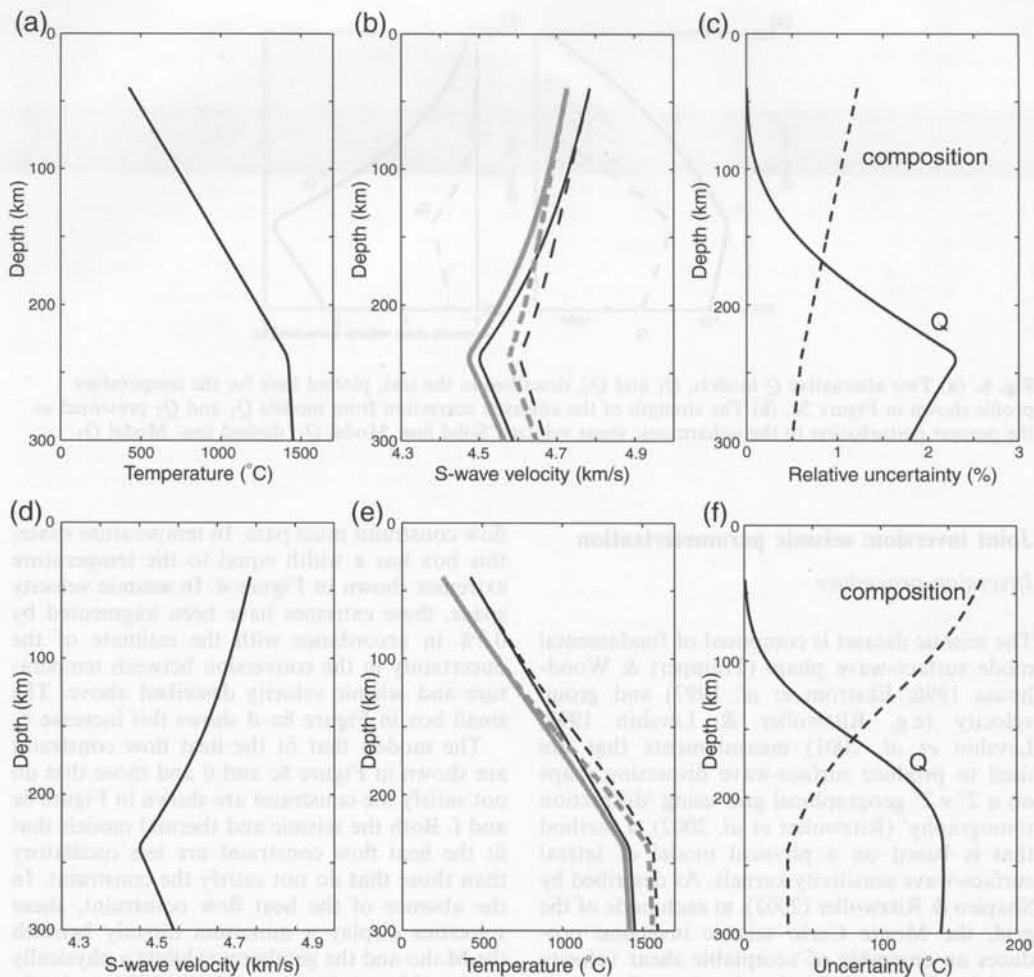
Uncertainties in the seismic velocity–temperature relationship result from a number of sources, including uncertainties in mantle composition, in the thermoelastic properties of individual minerals and in the anelastic correction which extrapolates anharmonic mineral properties measured in the laboratory to seismic frequencies. The physical properties of mantle minerals are measured in laboratories with high precision and are, therefore, not major contributors to errors in the velocity–temperature conversion. The most important uncertainties relate to mantle mineralogical composition and the anelastic correction. The presence of substantial quantities of melt and/or water in the mantle would also affect seismic velocities. These effects are expected to be negligible beneath old continental lithosphere, which is believed to have been largely desiccated during multiple episodes of melting during cratonic formation and is too cold for substantial quantities of melt currently to reside in the uppermost mantle.

To bound the effect of uncertainties in mantle mineralogical composition, we consider a pair of mantle compositional models proposed by McDonough & Rudnick (1998), one for on-cratonic (see previous section) and the

other for off-cratonic (68% olivine, 18% orthopyroxene, 11% clinopyroxene, and 3% garnet with an iron content  $X_{\text{Fe}} = 0.1$ ) mantle. An assessment of the uncertainty in shear velocity converted from temperature is shown in Figures 5a–c. We start with a cratonic temperature model that is composed of a conductive steady-state linear geotherm with a sub-Moho temperature of  $T_m = 437$  °C and mantle heat flow  $Q_M = 15$  mW/m<sup>2</sup> overlaying a 1300 °C adiabat (Fig. 5a). This temperature model is converted to a shear velocity model using the method of Goes *et al.* (2000) applied to the on-cratonic and off-cratonic compositions, as shown in Figure 5b. The difference in the resulting shear velocity curves provides a conservative estimate of the uncertainty in the temperature–velocity conversion within a single tectonic province.

We also consider two different models of the anelastic correction. Model  $Q_1$  is the model used by Shapiro & Ritzwoller (2004) and is described in the section above. For a contrasting model, we define Model  $Q_2$  which is based on values taken from Berckhemer *et al.* (1982) from experiments on forsterite ( $a = 0.25$ ,  $A = 2.0 \times 10^{-4}$ ,  $H^* = 584$  kJ/mol,  $V^* = 2.1 \times 10^{-5}$  m<sup>3</sup>/mol). These  $Q$  models are shown in Figure 6a computed from the temperature model in Figure 5a. Model  $Q_2$  has weaker attenuation and, therefore, will have a smaller anelastic correction, as shown in Figure 6b, where the strength of the anelastic correction is  $2Q^{-1}/\tan(\pi a/2)$ , by Equation 2. These models represent fairly extreme values for the anelastic correction, so the difference in shear velocities obtained from these models provides a conservative estimate of uncertainties in the temperature–velocity conversion caused by our ignorance of  $Q$ .

Figure 5c shows that, near to the surface, where the temperatures are relatively low and  $Q$  is high, the uncertainty in the anelastic correction is small and compositional uncertainty dominates. Deeper in the mantle, temperature increases,  $Q$  reduces, the anelastic correction strengthens and uncertainties in the anelastic correction become appreciably more important. Overall, the estimated uncertainty grows from about  $\pm 0.5\%$  of the seismic velocity in the uppermost mantle near the Moho to about  $\pm 1\%$  in the asthenosphere. A similar assessment of the uncertainty in temperature converted from shear velocity is shown in Figures 5d–f. Again, uncertainty in composition dominates at shallow depths and uncertainty in  $Q$  is more important deeper in the upper mantle. The resulting uncertainty in temperature converted



**Fig. 5.** Assessment of uncertainties in the interconversion between temperature and shear velocity. (a) Input mantle temperature profile. (b) Shear velocity profiles converted from temperature with variable compositional and  $Q$  models. Blackline, on-cratonic composition; grey line, off-cratonic composition; solid line, mantle model  $Q_1$ ; dashed line, mantle model  $Q_2$ . (c) Relative uncertainties in the shear velocity produced by uncertainties in composition (dashed line) and  $Q$  (solid line). (d) Input mantle shear velocity profile. (e) Temperature profiles converted from shear velocity with variable compositional and  $Q$  models, as in (b). (f) Relative uncertainties in temperature produced by uncertainties in composition (dashed line) and  $Q$  (solid line).

from shear velocity is about  $\pm 100^\circ\text{C}$  at all depths.

It is important to account for these uncertainties in the inversions both with the seismic parameterization and the thermal parameterization presented below, because both involve interconversion between temperature and shear velocity. With the seismic parameterization, we convert the temperature bounds to bounds in seismic velocities explicitly. To account for the uncertainty of this conversion, we increase the

range of seismic velocities by  $\pm 0.5\%$  of the seismic velocity. The width of the bounds on seismic velocity, therefore, approximately doubles. With the thermal parameterization, trial models are constructed in temperature space so we introduce these uncertainties in the conversion to shear velocity by increasing the bounds on the temperature model parameters. We describe how we introduce these uncertainties into the range of allowed thermal parameters in following sections.

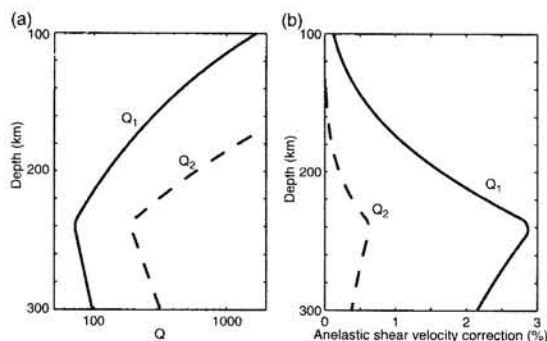


Fig. 6. (a) Two alternative  $Q$  models,  $Q_1$  and  $Q_2$ , described in the text, plotted here for the temperature profile shown in Figure 5a. (b) The strength of the anelastic correction from models  $Q_1$  and  $Q_2$  presented as the percent perturbation to the anharmonic shear velocity. Solid line, Model  $Q_1$ ; dashed line, Model  $Q_2$ .

## Joint inversion: seismic parameterization

### *Inversion procedure*

The seismic dataset is composed of fundamental mode surface-wave phase (Trampert & Woodhouse 1996; Ekström *et al.* 1997) and group velocity (e.g. Ritzwoller & Levshin 1998; Levshin *et al.* 2001) measurements that are used to produce surface-wave dispersion maps on a  $2^\circ \times 2^\circ$  geographical grid using 'diffraction tomography' (Ritzwoller *et al.* 2002), a method that is based on a physical model of lateral surface-wave sensitivity kernels. As described by Shapiro & Ritzwoller (2002), at each node of the grid, the Monte Carlo seismic inversion produces an ensemble of acceptable shear velocity models that satisfy the local surface-wave dispersion information, as illustrated in Figure 7 for Spatial Point 1 (whose location is indicated on Fig. 1). The model is radially anisotropic in the mantle ( $V_{sv} \neq V_{sh}$ ) to a depth of about 200 km, on average. The model is constructed to a depth of 400 km. We summarize the ensemble of acceptable seismic models with the 'Median Model', which is the centre of the corridor defined by the ensemble, and assign an uncertainty at each depth equal to the half-width of the corridor. When converting to temperature, we need the effective isotropic velocity in the upper mantle, which we define as  $V_s = (V_{sv} + V_{sh})/2$ .

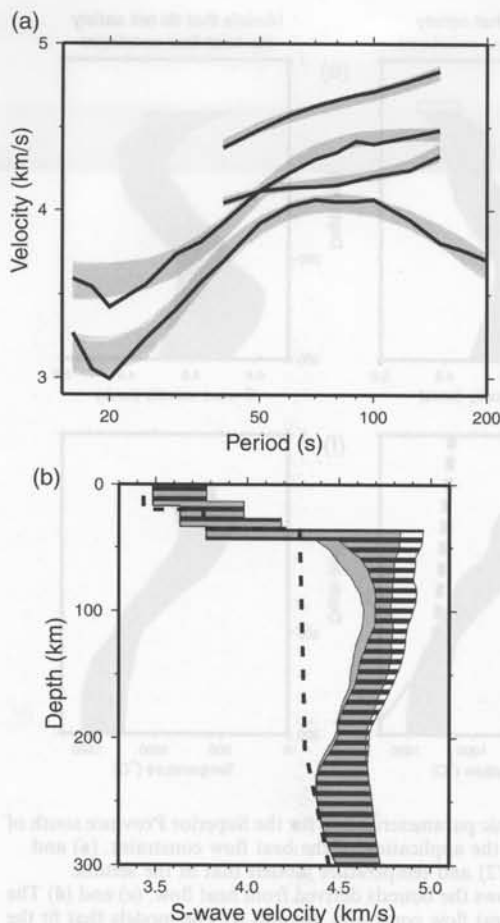
Figure 8 illustrates how assimilating heat flow information into the surface-wave inversion affects the inversion at Point 1, which is located south of Hudson Bay within the Superior Province (see Fig. 1). The heat flow constraint is shown in Figure 8a and b as a small box through which all models that satisfy the heat

flow constraint must pass. In temperature space, this box has a width equal to the temperature extremes shown in Figure 4. In seismic velocity space, these extremes have been augmented by 0.5% in accordance with the estimate of the uncertainty in the conversion between temperature and seismic velocity described above. The small box in Figure 8a–d shows this increase.

The models that fit the heat flow constraint are shown in Figure 8c and d and those that do not satisfy the constraint are shown in Figure 8e and f. Both the seismic and thermal models that fit the heat flow constraint are less oscillatory than those that do not satisfy the constraint. In the absence of the heat flow constraint, shear velocities display a minimum directly beneath the Moho and the geotherm exhibits a physically implausible minimum at about 100 km depth. The introduction of the heat flow constraint eliminates most of the models with this non-physical behaviour, systematically favouring models with very high seismic velocities in the uppermost mantle directly beneath the Moho. This is consistent with the results of LITHOPROBE refraction studies which have shown very fast  $P_n$  ( $\approx 8.2$  km/s) velocity beneath most of the Canadian Shield (e.g. Perry *et al.* 2002; Viejo & Clowes 2003). In converting from seismic velocity to temperature in constructing Figure 8b, d and f we have not included the uncertainty in the conversion from shear velocity to temperature. This uncertainty would widen the ensemble of temperatures somewhat.

Similar behaviour can be seen at Points 2 and 3, located in northern Manitoba within the Trans-Hudson Orogen and in the Ungava Peninsula within the Superior Province, respectively (Fig. 1). The results for these two sites are presented in Figures 9 and 10. In the Ungava





**Fig. 7.** Monte Carlo inversion with the seismic parameterization, but without thermodynamic constraints. (a) Surface-wave dispersion curves (black lines) at Spatial Point 1 (whose location is indicated on Fig. 1) and the range of curves (grey lines) predicted from the constitutive seismic models in (b). (b) The ensemble of radially anisotropic models that acceptably fit the dispersion curves found in (a). Solid grey corridor,  $V_{sv}$ ; cross-hatched corridor,  $V_{sh}$ ; dotted line,  $V_s$  from the 1-D model ak135 (Kennett *et al.* 1995).

Peninsula (Point 3), which is remote from heat flow measurements, the heat flow bounds on the seismic model are weaker and the seismic and temperature profiles displayed in Figure 10a and b are substantially more oscillatory and, hence, more physically questionable than those shown in Figures 8 and 9.

#### Characteristics of the upper mantle geotherm

As shown in Figure 8d, the temperature profiles that satisfy the heat flow constraint near heat

flow measurements are consistent with a nearly linear shallow mantle geotherm. At Point 1, the temperature gradient  $dT/dz \approx 5.5$  K/km, which translates to a mantle heat flux  $Q_M = kdT/dz \approx 16.5$  mW/m<sup>2</sup> for thermal conductivity  $k = 3$  W/m/K. The linearity of the mantle geotherm is consistent with a 'steady-state' thermal regime with no mantle heat sources. The shallow geotherm in Figure 8d displays a knee at about 200 km, below which the geotherm has a different nearly linear temperature gradient. This gradient,  $\approx 0.5$  K/km, is similar to the mantle adiabatic gradient. The shallower temperature gradient defines a thermal boundary layer whose thickness we will refer to as the lithospheric thickness. The definition of 'lithospheric thickness' is somewhat arbitrary but, for simplicity, we will define it as the depth where the shallow linear gradient intersects the mantle adiabat. For Point 1 (Fig. 8d), therefore, lithospheric thickness is estimated to be about 200 km. The main sources of errors in this estimate are due to uncertainties in the shallow geothermal gradient and in the potential temperature of the convecting mantle (i.e. the horizontal position of the mantle adiabat).

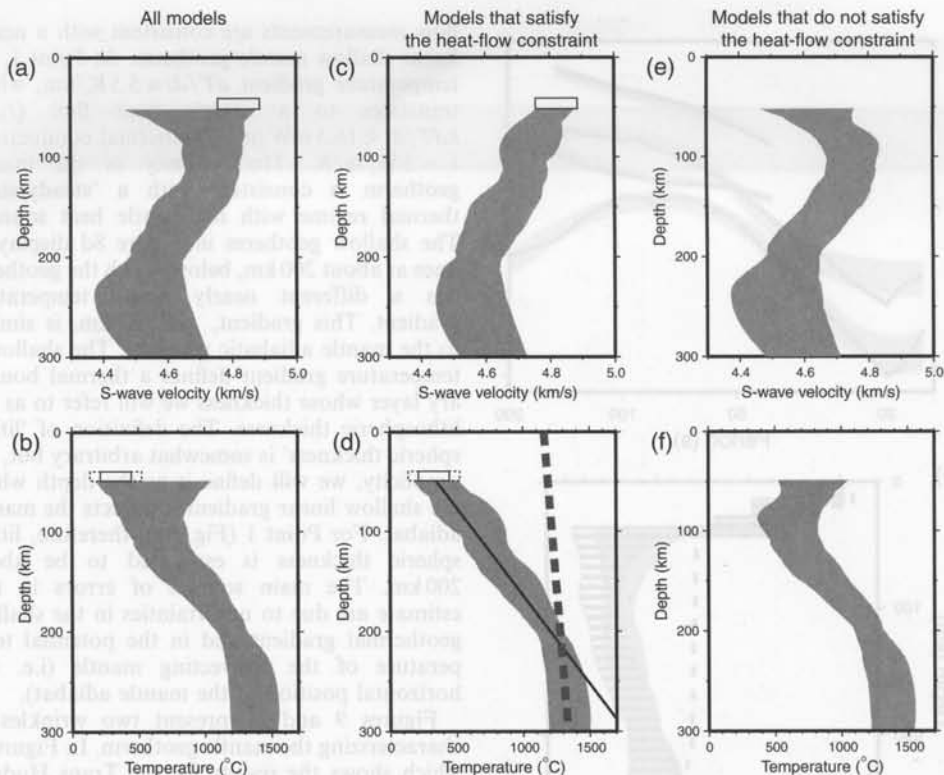
Figures 9 and 10 present two wrinkles in characterizing the mantle geotherm. In Figure 9, which shows the results for the Trans Hudson Orogen (Point 2), the lithosphere is so thick that the transition to the mantle adiabat is not observable. Thus, lithospheric thickness cannot be directly constrained more than by a lower bound of about 300 km. Mantle heat flux, however, is fairly well constrained to about 11 mW/m<sup>2</sup> and the temperature gradient is consistent with thermal steady state. At Point 3, in the Ungava Peninsula away from heat flow measurements (Fig. 10), the vertical oscillations in the temperature profile make it difficult to estimate either the mantle heat flux or the lithospheric thickness, or to test the steady-state hypothesis.

The problems exemplified by Figures 9 and 10 motivate us to apply further constraints in the inversion that are based on tightening physically reasonable bounds on allowed temperature models. These constraints are designed to allow us to obtain better estimates of mantle heat flux and lithospheric thickness.

### Joint inversion: thermal parameterization

#### Inversion procedure

Figures 9 and 10 motivate us to introduce a parameterization based on a physical model in thermodynamic steady state. The linearity of the

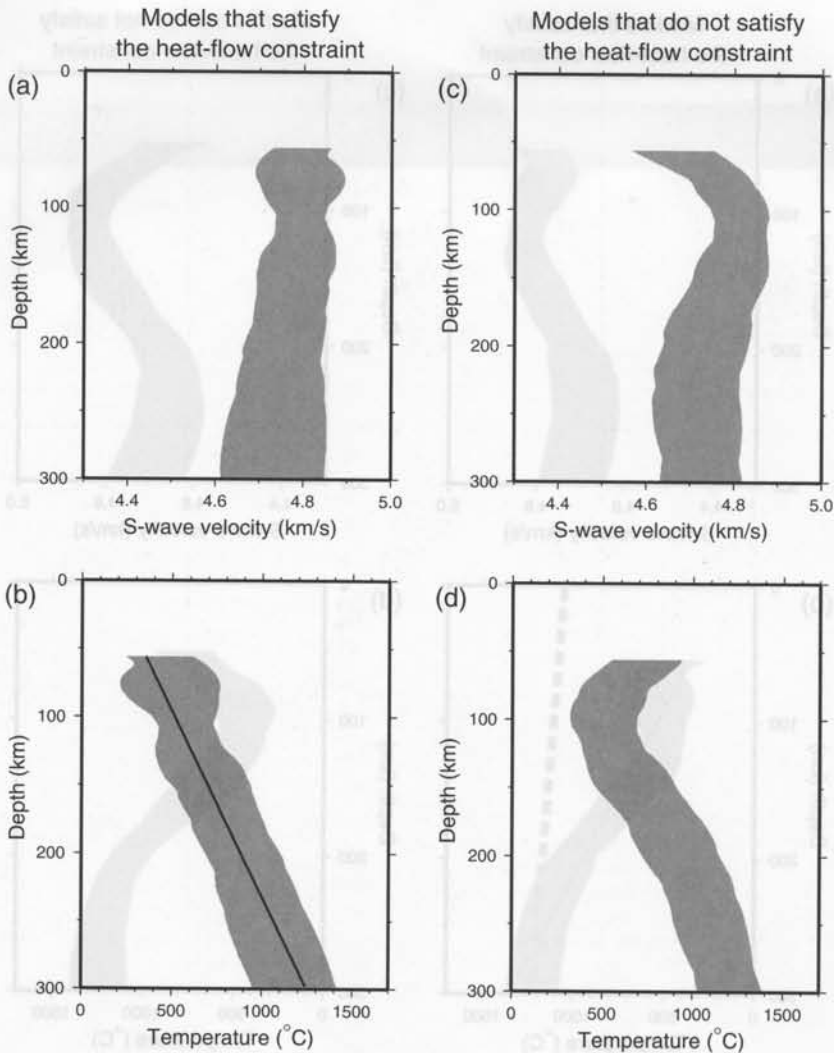


**Fig. 8.** Results of the Monte Carlo inversion with the seismic parameterization for the Superior Province south of Hudson Bay (Point 1 on Fig. 1), illustrating the effect of the application of the heat flow constraint. (a) and (b) The ensemble of acceptable seismic ( $V_s = (V_{sv} - V_{sh})/2$ ) and temperature models that fit the seismic dispersion curves acceptably. The small box at the top shows the bounds derived from heat flow. (c) and (d) The models that fit both the local dispersion curves and the heat flow constraint. (e) and (f) The models that fit the local dispersion curves but not the heat flow constraint. In (d), the best-fitting linear geotherm (solid line,  $Q_M = 16.5 \text{ mW/m}^2$ ) is shown as the solid line. The thick dashed line indicates the adiabat, whose horizontal offset is determined from the deep part of the temperature profile and whose vertical gradient is  $0.5 \text{ }^\circ\text{C/km}$ .

shallow mantle geotherm, which is consistent with the steady-state hypothesis, is a general feature of the seismic model near heat flow measurements. The physical model we have adopted is schematized in Figure 11. The thermal parameterization consists of a linear gradient in the shallow mantle over a deeper adiabatic gradient set equal to  $0.5 \text{ K/km}$ . These two gradients meet in a narrow transition region to eliminate a kink in the temperature profile. The Monte Carlo procedure randomly generates three numbers: the mantle temperature directly beneath Moho ( $T_m$ ), the shallow mantle temperature gradient ( $dT/dz$ ), and the potential temperature ( $T_p$ ). Uppermost mantle temperature and the shallow gradient define the lithospheric geotherm. The potential temperature (i.e. the upward continuation of the adiabatic

temperature profile to the surface) sets temperatures in the asthenosphere. Lithospheric thickness,  $L$ , is defined by the intersection between the lithospheric geotherm and the adiabat. Other parameters could also be varied within some bounds, for example, the adiabatic gradient, but doing so does not significantly increase the range of temperature profiles retained. As with the seismic parameterization, the inversion is performed at each point on a  $2^\circ \times 2^\circ$  grid across the region of study.

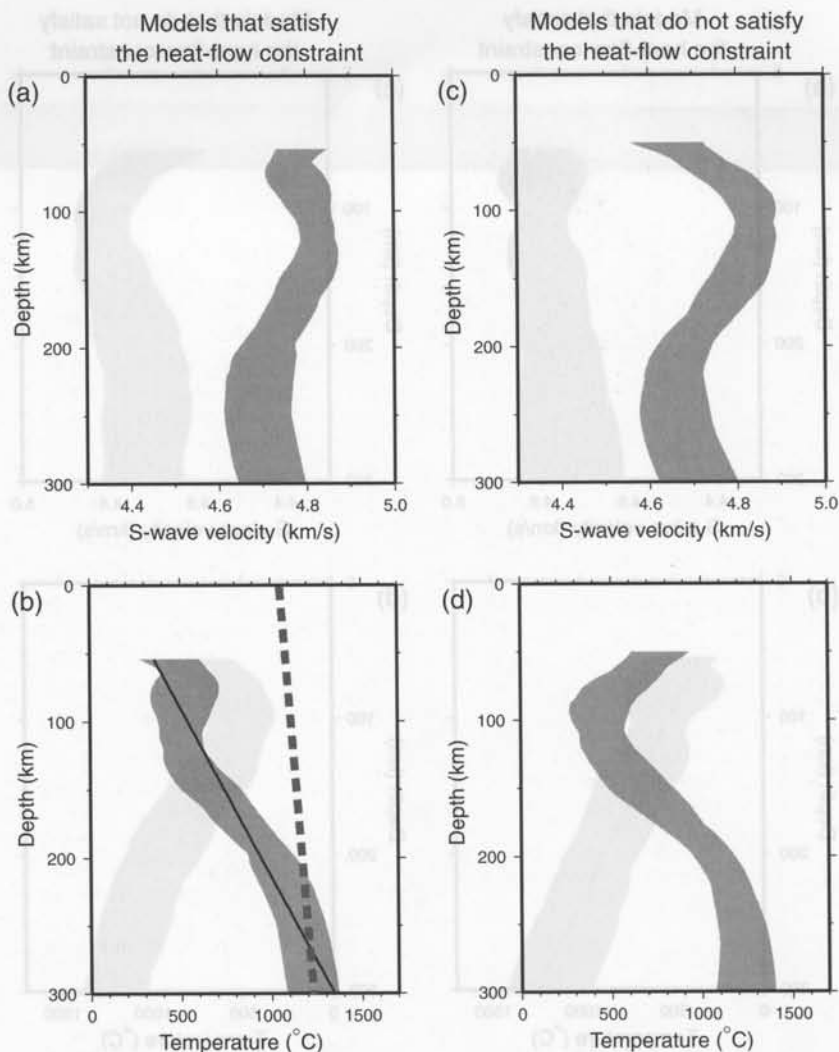
Each of the three parameters is subjected to a constraints. Firstly, the uppermost mantle temperature is within the same temperature bounds as for the seismic inversion,  $T_{\max} \leq T_m \leq T_{\min}$ , where  $T_{\max}$  and  $T_{\min}$  are shown in Figure 4. This constraint remains tightest near the heat flow measurements. In accordance with Figure 5, to



**Fig. 9.** Results of the joint inversion with the seismic parameterization at the point in the Trans-Hudson Orogen (Point 2 in Fig. 1). Here (a) and (b) are the models that fit both the local dispersion curves and the heat flow constraint and (c) and (d) are the models that do not fit the heat flow constraint. The mantle adiabat cannot be discerned, as the knee in the temperature profile appears to be deeper than the extent of the model (i.e. 300 km). As in Figure 8, in (b) the best-fitting linear geotherm ( $Q_M = 11 \text{ mW/m}^2$ ) is shown.

account for uncertainty in the conversion to shear velocity, we increase these bounds by  $\pm 0.5\%$  in the seismic velocity, as was also done in Figure 9a–d. Secondly, following Rolandone *et al.* (2002), lithospheric heat flux  $Q_M = kdT/dz$  is constrained to be larger than  $11 \text{ mW/m}^2$ . It is also constrained to be less than surface heat flux,  $Q_s$ . Thus,  $3.67 \text{ K/km} \leq T/dz \leq Q_s/k$ , where thermal conductivity  $k = 3.0 \text{ W/m/K}$ . Finally, the range of

allowed potential temperatures is somewhat difficult to quantify. McKenzie & Bickle (1988) have proposed a mean upper mantle potential temperature of  $1280 \text{ }^\circ\text{C}$  and Jaupart *et al.* (1998) argue that uncertainties in this value are at least  $\pm 50 \text{ }^\circ\text{C}$ . The mean value beneath continents may be somewhat lower than the value proposed by McKenzie & Bickle. To be conservative, we extend the range somewhat and apply the following intrinsic bounds on potential tempera-



**Fig. 10.** Results of the joint inversion with the seismic parameterization at the point in the Superior Province in the Ungava Peninsula of New Quebec (Point 3 in Fig. 1). In contrast to Spatial Points 1 and 2, this point is remote from heat flow measurements, which results in a weak heat flow constraint so that the oscillations in the seismic and temperature profiles have not been eliminated. As in Figures 8 and 9, in (b) the best-fitting linear geotherm (solid line,  $Q_M = 12 \text{ mW/m}^2$ ) is shown with the mantle adiabat, but there are large uncertainties due to irregularities in the temperature profile with depth.

ture:  $1100^\circ\text{C} \leq T_p \leq 1300^\circ\text{C}$ . To account for uncertainty in the temperature to shear velocity conversion, we expand these bounds by  $\pm 100^\circ\text{C}$  to  $1000^\circ\text{C} \leq T_p \leq 1400^\circ\text{C}$ .

One of the principal advantages of the thermal parameterization is the possibility to apply physically based constraints on the model parameters. Although the bounds on potential temperature are poorly known, the lower bound

of  $11 \text{ mW/m}^2$  on mantle heat flow strongly constrains the seismic model.

After a trial model is constructed in temperature space, it is converted to shear velocity using the method of Goes *et al.* (2000). Trial seismic crustal structures are introduced as well as mantle radial anisotropy similar to the generation of these features of the model in the seismic parameterization. At each grid node, some

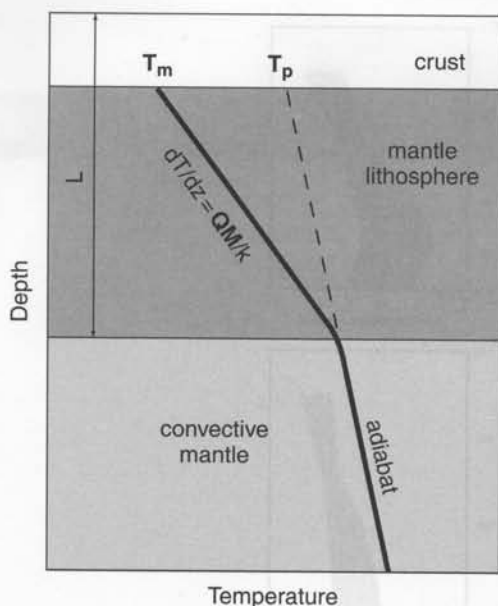


Fig. 11. Schematic representation of the thermal model which is defined by three parameters. The steady-state lithospheric geotherm is defined by temperature directly beneath the Moho ( $T_m$ ) and the linear gradient ( $dT/dz$ ) which is simply related to mantle heat flow ( $Q_M$ ). Below the lithosphere the temperature gradient in the convecting mantle is adiabatic with potential temperature  $T_p$  and an adiabatic gradient of 0.5 K/km. These two gradients join smoothly through a narrow transition region in order to eliminate a non-physical kink in the temperature profile.

temperature profiles will be rejected entirely, but some will be found to fit the seismic data acceptably for an appropriate subset of seismic crustal models and models of radial anisotropy. These profiles define an ensemble of acceptable profiles in temperature space. They are also combined with the crustal and radial anisotropic models to define an ensemble of acceptable models in seismic space.

Results for Points 1 and 3 (Fig. 1) are shown in Figure 12. At Point 1, the results are very similar to those obtained with the seismic parameterization (Fig. 8b, d). The estimated average mantle component of heat flow is  $15 \text{ mW/m}^2$  and its standard deviation is  $2.0 \text{ mW/m}^2$ . Average lithospheric thickness is 246 km with a standard deviation of 33 km. At Point 3, the thermal parameterization yields an ensemble of models that fit the surface-wave data but do not display the physically question-

able vertical oscillations that appear in Figure 10. The estimated average mantle heat flow and average lithospheric thickness are  $12.2 \text{ mW/m}^2$  and 294 km with corresponding standard deviations of  $1.4 \text{ mW/m}^2$  and 46 km.

### Seismic and thermal models

The middle of the ensemble of acceptable models in the seismic and temperature model spaces are the Median Models. Slices of the Median Models of shear velocity and temperature are shown in Figures 13 and 14. Although shear velocity and temperature are fairly homogeneous at 80 km depth, at greater depths the variability is greater, reflecting variations in lithospheric thickness across the study area, as discussed below. The lithosphere is warmer and thinner to the southeast, in the Appalachians, and becomes thicker toward the north and, especially, towards the northwest, as can be seen in Figure 14b and f. Vertical oscillations that plague the seismic parameterization are absent from the model, as the cross-sections in Figure 14 attest.

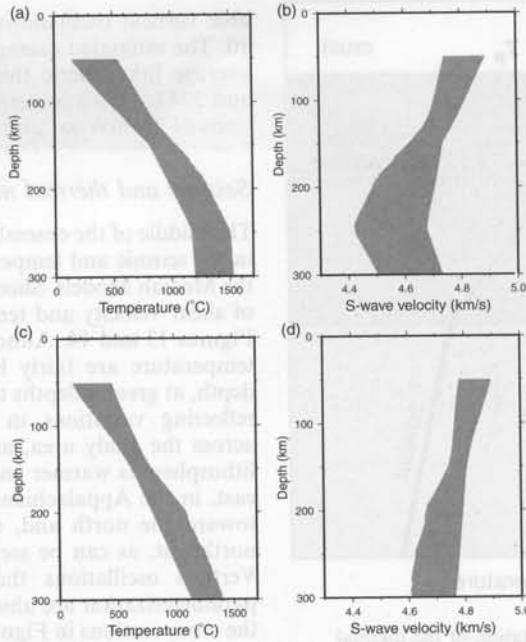
### Mantle heat flux and lithospheric thickness

At each point, we construct an ensemble of mantle heat flow ( $Q_M$ ) and lithospheric thickness ( $L$ ) estimates derived from the ensemble of acceptable temperature profiles ( $Q_M = kdT/dz$ ,  $L =$  depth where the lithospheric geotherm intersects the mantle adiabat). The average of the ensemble of acceptable mantle heat flow and lithospheric thickness estimates is shown in Figure 15.

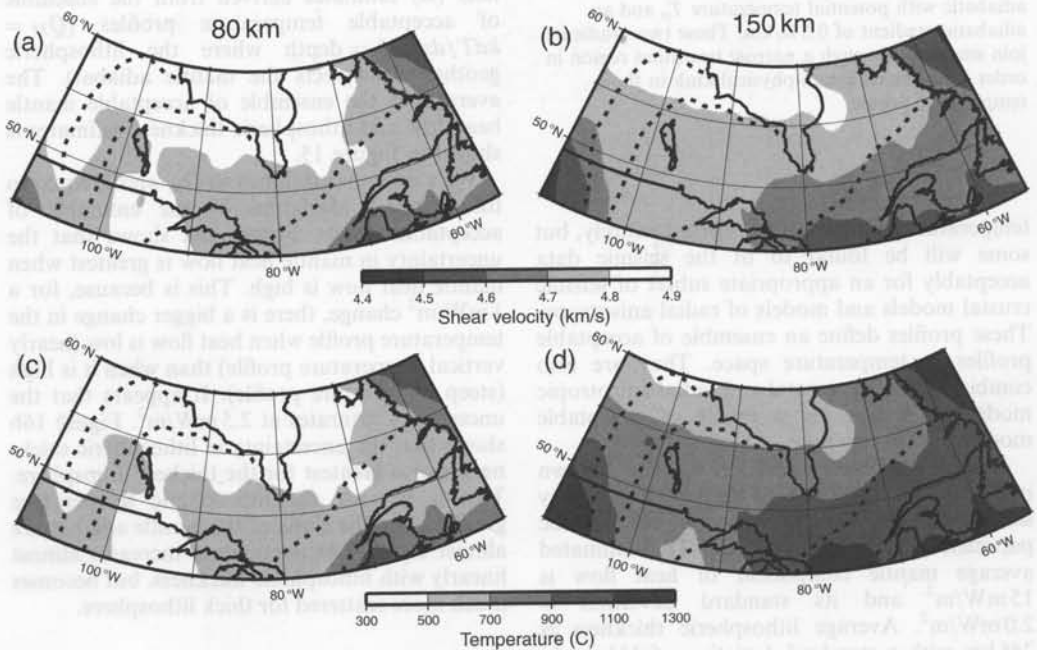
We assign uncertainties to  $Q_M$  and  $L$  equal to the standard deviation of the ensemble of acceptable values. Figure 16a shows that the uncertainty in mantle heat flow is greatest when mantle heat flow is high. This is because, for a  $1 \text{ mW/m}^2$  change, there is a bigger change in the temperature profile when heat flow is low (nearly vertical temperature profile) than when it is high (steep temperature profile). It appears that the uncertainty saturates at  $2.5 \text{ mW/m}^2$ . Figure 16b shows that the uncertainty in lithospheric thickness is also greatest for the thickest lithosphere. This is because the lithospheric temperature gradient and the slope of the mantle adiabat are almost equal. The uncertainty increases almost linearly with lithospheric thickness, but becomes much more scattered for thick lithosphere.

### Discussion

The seismic velocities and temperatures displayed in Figures 13 and 14 demonstrate



**Fig. 12.** Example results of the Monte Carlo inversion with the thermal parameterization at Spatial Points 1 and 3. (a) Ensemble of acceptable temperature models at Point 1. (b) Ensemble of acceptable seismic models at Point 1. (c) Ensemble of acceptable temperature models at Point 3. (d) Ensemble of acceptable seismic models at Point 3.



**Fig. 13.** (a) and (b) Horizontal slices of the seismic  $V_s = (V_{sh} + V_{sv})/2$  model and (c) and (d) the temperature model at depths of (a, c) 80 km and (b, d) 150 km. Dashed lines, boundaries of principal Precambrian provinces. Adapted from Hoffman (1989).

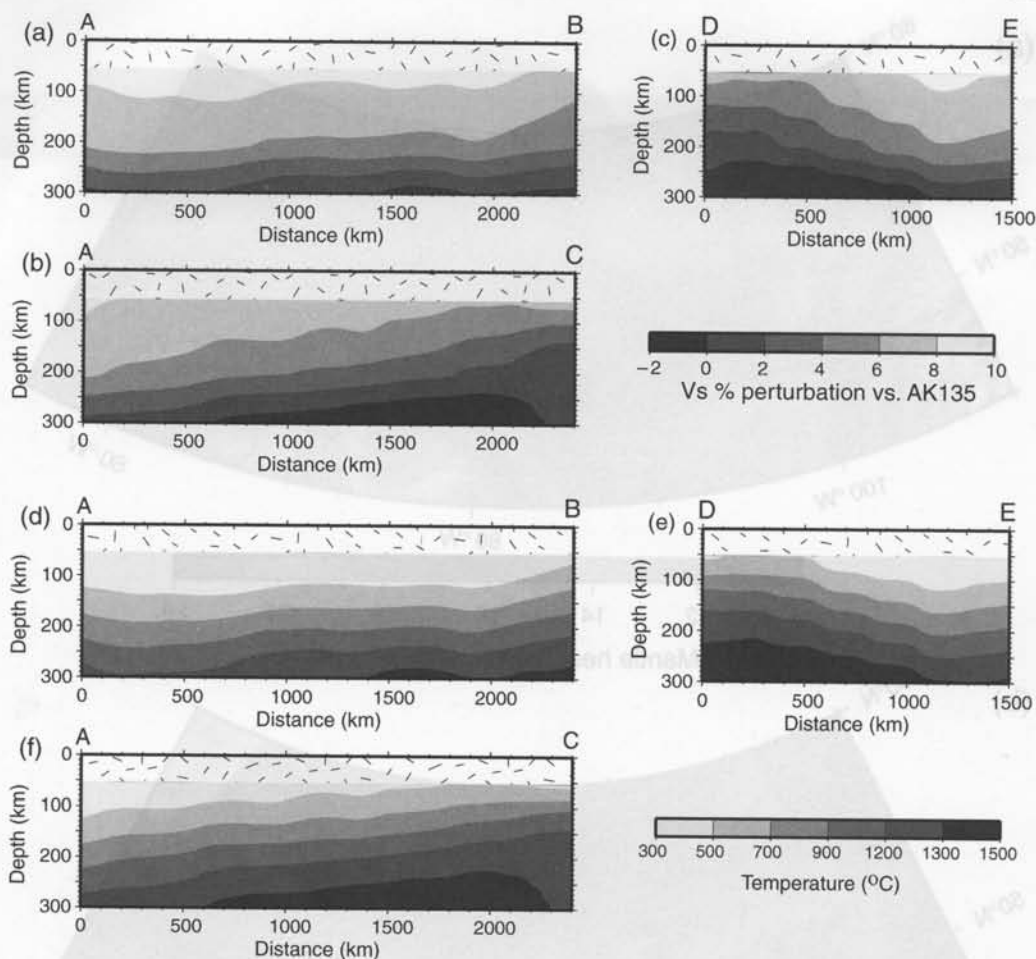


Fig. 14. Vertical slices of the mantle seismic and temperature models along the three profiles shown in Figure 1. (a)–(c) Isotropic shear velocity ( $V_s = (V_{sh} + V_{sv})/2$ ) presented as percent perturbation relative to the global 1-D model ak135. (d)–(f) Temperature in  $^{\circ}\text{C}$ .

considerable variability across the study area. This information is probably best summarized by mantle heat flow and by the lithospheric thickness (Figure 15). The mantle heat flow appears to increase and lithospheric thickness to decrease beneath the Appalachians to the southeast. Within the Canadian Shield, mantle heat flow from seismic inversions ranges between 11 and 18  $\text{mW}/\text{m}^2$ , with apparently higher values in the Grenville Province. Within most of the Superior Province, mantle heat flow ranges between 11 and 15  $\text{mW}/\text{m}^2$ , with small amplitude, short wavelength ( $< 1000$  km) spatial variations.

Because of horizontal diffusion of heat, such variations in mantle heat flow are damped at the surface and cannot be resolved by the heat flow data. Pinet *et al.* (1991) had concluded from their analysis of heat flow and heat production data that the mantle heat flow is the same beneath the Grenville Province and the Superior Province. Within the Canadian Shield, the variations are not always well correlated with geological provinces. The lowest mantle heat flow values are found beneath the Archean Rae Province and the Palaeo-Proterozoic Trans Hudson Orogen, in northern Manitoba and Saskatchewan, where

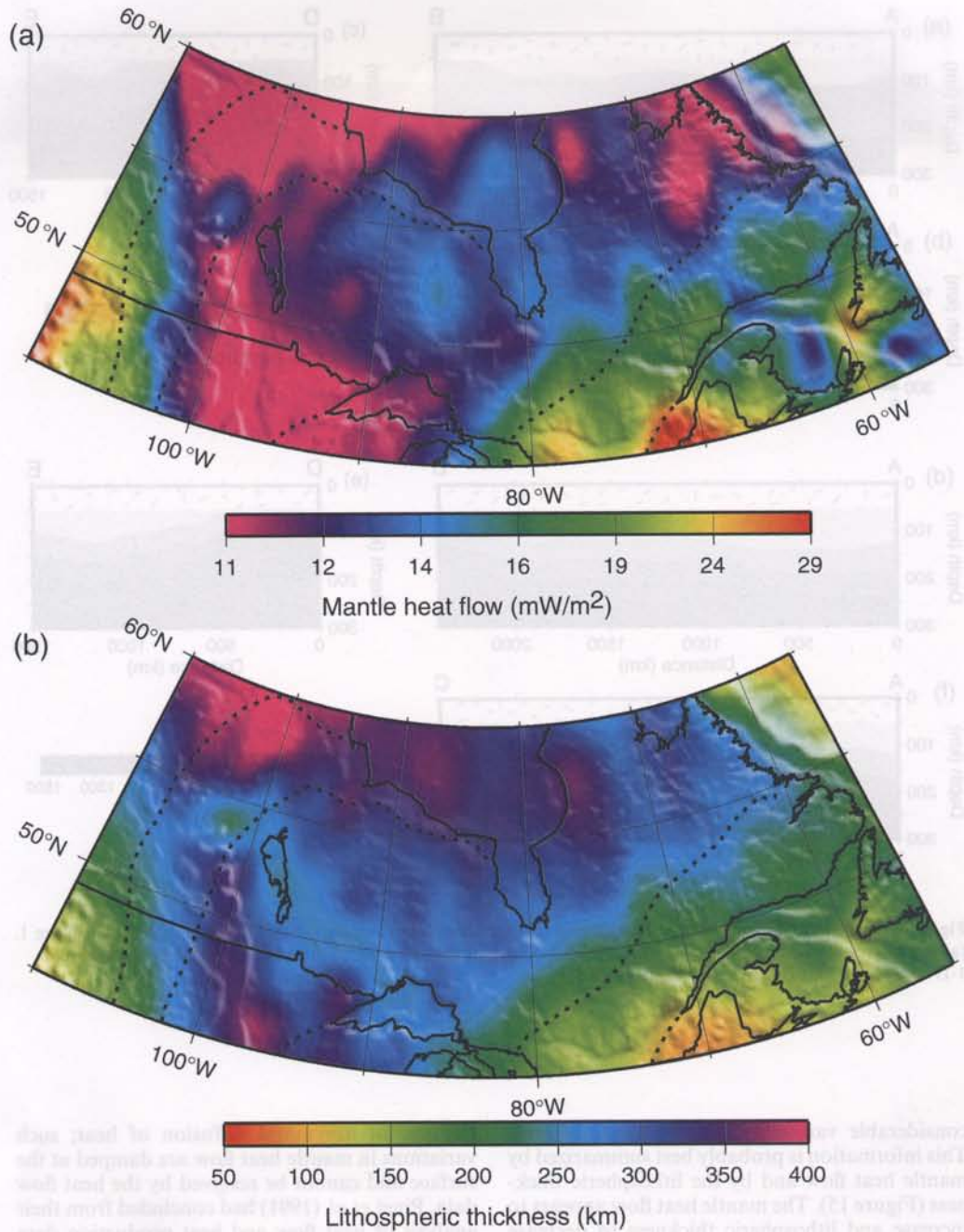


Fig. 15. (a) The estimated mantle component of heat flow,  $Q_M$ . (b) The estimated lithospheric thickness,  $L$ . Dashed lines, boundaries of principal Precambrian provinces. Adapted from Hoffman (1989).

juvenile crust is thrust over the Archean Sask craton. Although the thermal regime of the Canadian Shield does not simply reflect its

surface geology and is not simply related to the last tectonomagmatic event, it reveals the deeper structure of the lithosphere.



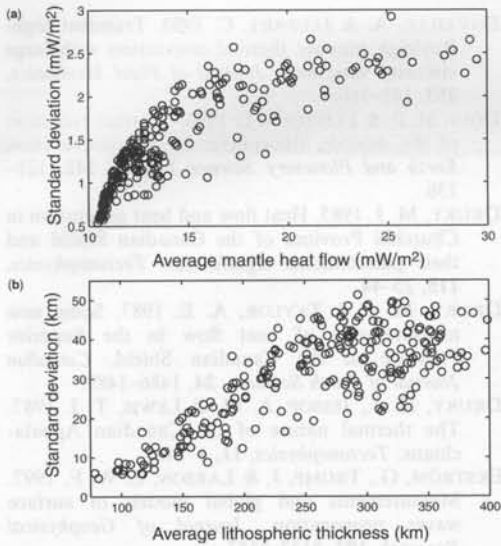


Fig. 16. (a) Estimated standard deviation of mantle heat flow plotted v. mantle heat flow. (b) Estimated standard deviation of lithospheric thickness plotted v. lithospheric thickness. Values are taken at the model nodes on a  $2^\circ \times 2^\circ$  grid across the region of study.

Patterns of lithospheric variability do emerge, however. As Figure 17 shows, lithospheric thickness and mantle heat flow are anti-correlated. A scaling law between mantle temperature and heat flow from the convecting mantle was used by Jaupart *et al.* (1998) to determine the lithospheric thickness. The analysis assumes that the heat flow at the base of the lithosphere is supplied by small-scale convection (e.g. Davaille & Jaupart 1993), and results in an approximately power law relation between mantle heat flow and lithospheric thickness. Results of our seismic inversion shown in Figure 17 are also well approximated by a power law curve (Fig. 17) that is relatively close to the shape of the  $Q_M$  to  $L$  relationship given by Jaupart *et al.* (1998). However, Jaupart *et al.* pointed out that changes in lithospheric thickness do not require changes in mantle heat flow and that Moho temperatures (determined also by crustal heat production) also control lithospheric thickness. This is consistent with Figure 17 showing much more variability in lithospheric thickness than in mantle heat flow.

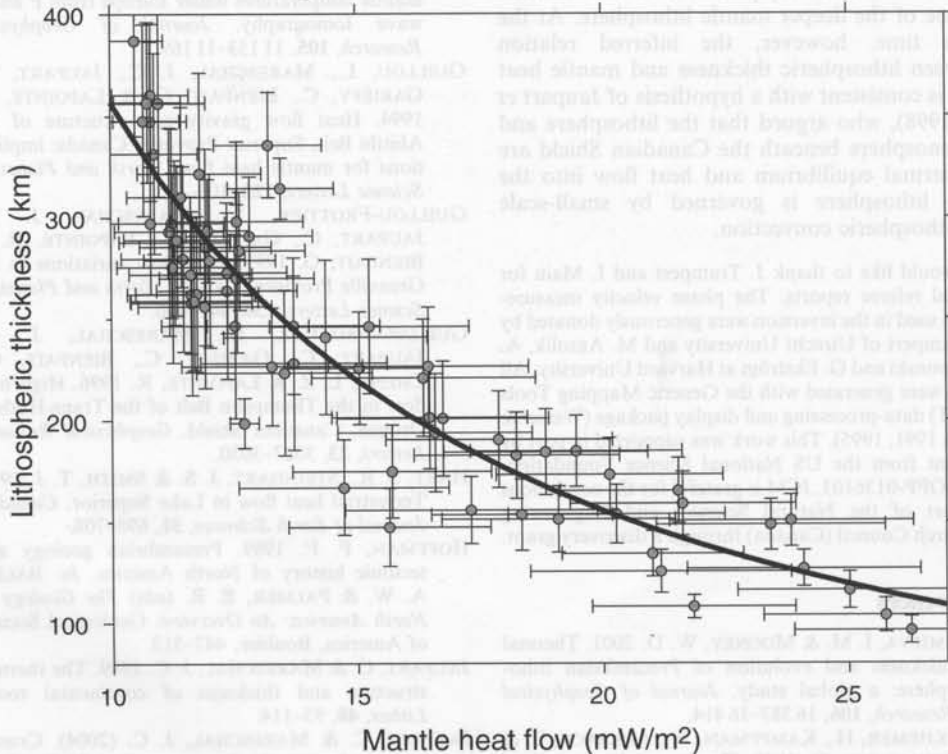


Fig. 17. Lithospheric thickness ( $L$ ) v. mantle heat flow ( $Q_M$ ) taken from model nodes near the heat flow measurements shown in Figure 1. One standard deviation error bars in both  $L$  and  $Q_M$  are shown. The solid line is a power law curve that fits the data well ( $L = 5660Q_M^{-1.2}$ ).

## Conclusions

The primary conclusion of this work is that seismic surface-wave data and surface heat flow observations can be reconciled over broad continental areas; i.e. both types of observations can be fit simultaneously with a simple steady-state thermal model of the upper mantle. This has motivated the reformulation of the seismic surface-wave inverse problem in terms of a thermal model described by three parameters: temperature in the uppermost mantle directly beneath Moho, the mantle temperature gradient, and the potential temperature of the sublithospheric convecting mantle from which we also estimate lithospheric thickness and mantle heat flux. In addition to the steady-state thermal constraint, prior physical information based on surface heat flow measurements is applied.

The results of a Monte Carlo inversion of the surface-wave data with this 'thermal parameterization' across the Canadian Shield demonstrate that lithospheric temperature and shear velocity are not well correlated with surface tectonic history, which implies that the tectonic regime of the crust is not simply related to the thermal regime of the deeper mantle lithosphere. At the same time, however, the inferred relation between lithospheric thickness and mantle heat flow is consistent with a hypothesis of Jaupart *et al.* (1998), who argued that the lithosphere and asthenosphere beneath the Canadian Shield are in thermal equilibrium and heat flow into the deep lithosphere is governed by small-scale sublithospheric convection.

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