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Heat flow variations in a deep borehole near Sept-Iles, Québec, Canada: Paleoclimatic interpretation and implications for regional heat flow estimates

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Abstract. A deep (> 2000m) borehole in the Sept-Iles intrusion, on the north shore of the Saint Lawrence River, in Québec, Canada, was repeatedly logged for temperature. Systematic variations of the temperature gradient with depth are not correlated with the thermal conductivity. We interpreted the temperature profile as follows:

(1) During the last glacial maximum, the temperature at the base of the ice sheet was cold ($\approx -5°$C); (2) When the region was below sea level, between 10 and 5ky B.P., the ground surface temperature was warm ($\approx 6°$C); (3) The average ground surface temperature dropped to $\approx 2°$C at 5ky B.P. when the region rebounded above sea level; (4) The long time averaged ground surface temperature before the last glacial maximum was $\approx 0-1°$C; (5) The reference heat flow ($36-37$mW m^{-2}) is $4-5$mW m^{-2} higher than estimated from the upper 1000m of the heat flow profile.

This interpretation can not be extrapolated to the entire region covered by the Laurentide ice sheet. Except for extremely deep (> 1500m) boreholes, the small uncertainty (< 15%) affecting heat flow estimates can not be eliminated.

Introduction

Pleistocene climatic variations have perturbed the subsurface temperature and may affect the heat flow estimates in regions that were glaciated [e.g. *Birch*, 1948]. Following *Jessop* [1971], a correction has been applied to heat flow estimates from the Canadian Shield. This correction includes the climatic variations since 400,000 years before present (B.P.) and assumes that the ground surface temperature was the same as present during interglacial and -1°C during glacial episodes. A study of a very deep borehole near Flin-Flon, Manitoba, has failed to demonstrate any effect of the last glacial episode. The temperature gradient and conductivity were negatively correlated resulting in a constant heat flow with depth [*Sass et al.,* 1971]. On the other hand, *Nielsen and Beck* [1989] have analyzed several temperature profiles from Ontario and suggested that the ground temperature was -4°C during the last glacial maximum (LGM). The temperature profiles are consistent with the suggestion but the boreholes are too shallow (600m) to constrain temperature during the LGM.

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We have studied the variations in heat flow measured to a depth of 1800m in a borehole drilled in the Sept-Iles intrusion on the north shore of the Saint-Lawrence river, in eastern Québec. The heat flow increases from $\approx$ 28mW m^{-2} in the upper 1000m to $\approx$ 36mW m^{-2} between 1500 and 1800m. We interpret this increase as caused by the warming after the LGM. We shall discuss the implications of this interpretation for continental heat flow estimates.

Heat flow data

The Sept-Iles complex is a layered gabbro intrusion, circular in shape with a radius of $\approx$25km, on the North shore of the Saint-Lawrence River (Figure 1). The thickness of the intrusion has been estimated to $\approx$ 2km from gravity data [*Loncarevic et al.*, 1990]. In 1994, we have logged two exploration holes. The temperature measurements are made with a thermistor at 10m interval. The measurement precision is 2mK; the Kevlar cables experience negligible stretching and we estimate that the total error, including the error on depth, is <5mK. The thermal conductivity of representative core samples is measured with the method of divided bars [*Misener and Beck*, 1960]. Each value, based on five measurements on large samples, represents the bulk conductivity of the rock. The heat flow $Q(z)$ is determined as:

$$Q(z) = k\frac{\partial T}{\partial z} \qquad (1)$$

where T is temperature, z is depth, and k is thermal conductivity. We obtained a heat flow value of 32mW m^{-2} after climatic correction [*Guillou-Frottier et al.*, 1995]. One of the drillholes, > 2000m, contains a record of surface temperature variations dating to at least 20,000y B.P. This site is well suited to infer the ground surface temmperature history (GSTH) because of the homogeneous lithology of the intrusion. In 1994, our equipment did not permit logging below 1000m. We have logged the deep borehole to 1800m with a different probe and cable in 1998. Figure 2 shows the temperature profiles, the thermal conductivity, and the heat flow vs depth obtained in 1994 and 1998. With the exception of one value, the conductity remains within 20% of the mean (1.98W m^{-1} K^{-1}). The two profiles exhibit a similar fine structure although they were obtained with different probes.

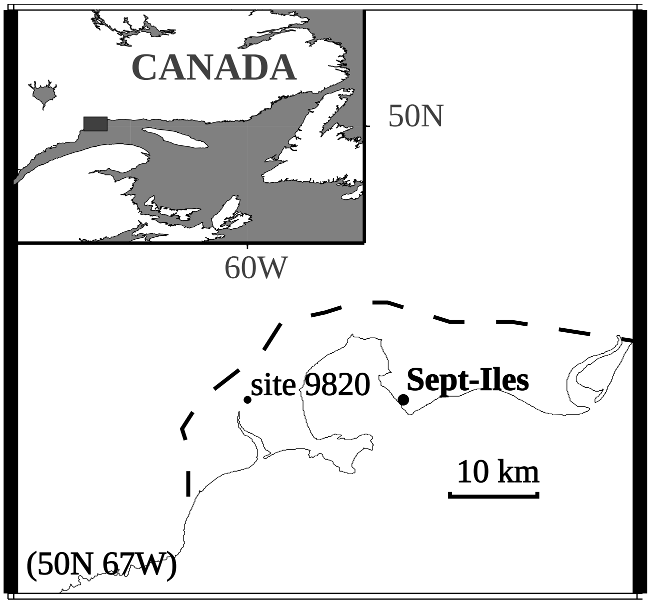

Figure 1. Map of the region of Sept-Iles, with the location of the drillhole, and the limit of the intrusion (discontinuous line). The map of eastern Canada is inset.

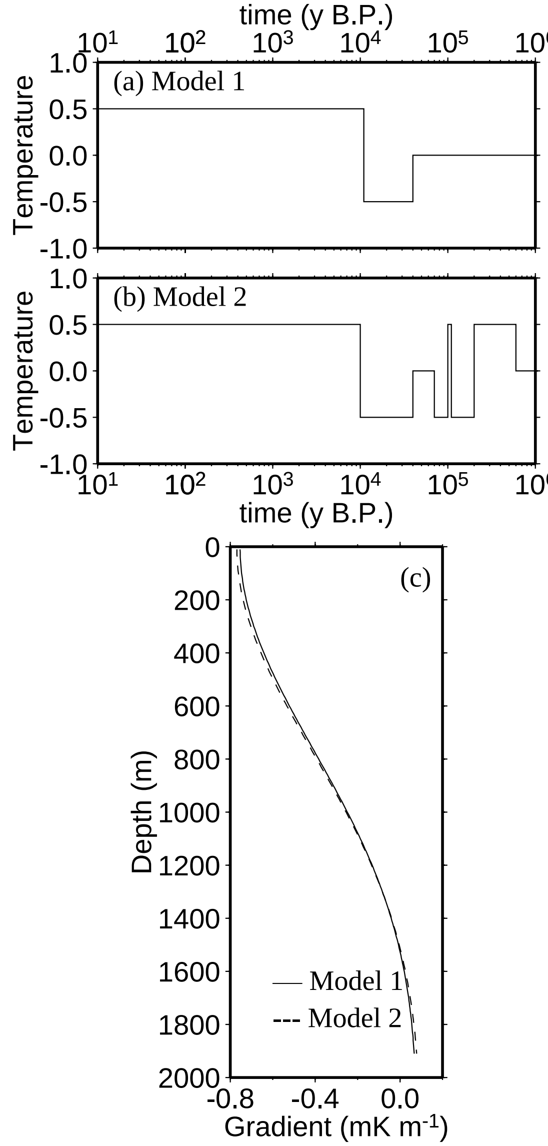

Figure 3. Two models of GSTH starting at 50ky B.P. (a) and starting at 500ky B.P. (b). The calculated profiles (c) are undistinguishable.

Interpretation

Direct models

Two features must be noted on the heat flow profile (Figure 2): (1) The marked decrease of the heat flow toward the surface (i.e. < 200m) is characteristic of recent climatic warming; (2) The increase below 1000m could be due to warming following the deglaciation.

The temperature T at depth z in a conductive half space experiencing horizontally uniform variations in surface temperature is:

$$T(z) = T_{ref} + q_{ref} \int_0^z \frac{dz'}{k(z')} + T_t(z) \qquad (2)$$

where T_{ref} and q_{ref} are the reference surface temperature and heat flow, and $T_t(z)$ is the perturbation caused by the variations in ground surface temperature. If the ground surface temperature is approximated by its average value ΔT_k

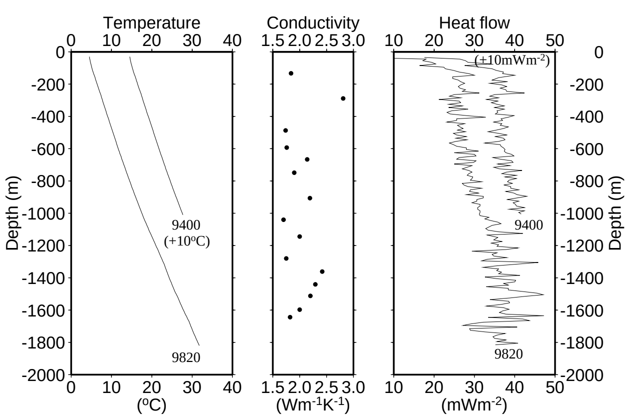

Figure 2. The two temperature profiles obtained in 1994 and 1998, the measured thermal conductivity values, and the heat flow profiles.

during K time intervals (t_k, t_{k+1}) [*Carslaw & Jaeger*, 1959, p.63]:

$$T_t(z) = \sum_{k=0}^{K} \Delta T_k (\mathrm{erfc}\frac{\mathrm{z}}{2\sqrt{\kappa \mathrm{t_{k+1}}}} - \mathrm{erfc}\frac{\mathrm{z}}{2\sqrt{\kappa \mathrm{t_k}}}) \qquad (3)$$

where κ is the thermal diffusivity.

Direct models show that the temperature profile is not affected by the GSTH before 50ky B.P.. The gradient obtained from a detailed GSTH to 500ky B.P. (model 1) and that obtained for a constant ground temperature before 50ky B.P. (model 2) are practically undistinguishable (Figure 3), if we select correctly T_{ref}, i.e. the "time-averaged" temperature of the ground surface before the start of the GSTH [*Harris & Chapman*, 1995].

Monte-Carlo Inversion

In order to determine the range of GSTH, reference surface temperature, and heat flow compatible with the data,

Table 1. Parameterizations used in the Monte Carlo inversions.

parameterization	a	b	c
parameters	22	21	20
trials ($\times 10^6$)	100	200	200
success	80	143	173
	limits of time intervals		
	100	100	100
	200	200	200
	300	300	300
	400	400	400
	500	500	500
	1000	750	750
	2000	1000	1000
	3000	1500	1500
	4000	2000	2000
	5000	3000	3000
	6000	4000	4000
	7000	6000	6000
	8000	8000	8000
	9000	10000	10000
	10000	15000	15000
	15000	20000	20000
	20000	30000	35000
	30000	40000	
	40000	50000	
	50000		

we have used a Monte Carlo inversion [*Press et al.*, 1992, p. 684]. We generated random model parameters, and compared the corresponding tempera ture profile with the data.

The set of parameters is retained when the distance (L_2 norm) between the data and the calculated temperature profile is such that the mean difference between the two profiles is <0.025K. The parameters to be determined are the reference heat flow and temperature and mean ground surface temperatures during K time intervals (Table 1), which are chosen accordingly to the resolution of the data [e.g. *Beltrami & Mareschal*, 1995]. The surface temperatures vary between -15°C and +15°C. The thermal diffusivity is assumed $0.8 \times 10^{-6}\text{m}^2\ \text{s}^{-1}$ for an average thermal conductivity of 2W $\text{m}^{-1}\ \text{K}^{-1}$. Model a and b start at 50ky B.P. and differ by the selection of time intervals. Model c was introduced because the marine $\delta^{18}O$ shows that the LGM did not start until 35ky B.P. [*Imbrie et al.*, 1984].

Results are presented in Figure 4 for different parameterizations. All the GSTH show a cold episode before 10ky B.P., i.e. the retreat of the Laurentide Ice Sheet dated by ^{14}C dates on moraines [*Hillaire-Marcel*, 1979]. A relatively warm episode is seen between 10 and 6ky B.P., with temperature ≈ 2K higher than present. This episode explains why the heat flow increases mostly between 800 and 1200m (Fig. 2). The region was below shallow waters in the Gulf of Saint-Lawrence after the glacial retreat and rebounded above sea level at 6ky B.P. [*Hillaire-Marcel*, 1979]. Palynological studies in the Gulf indicate that the surface water temperature was warm (15°C) in the summer with a mean annual temperature in the range of 7-9°C [*de Vernal et al.*, 1993]. Finally, the results confirm the 1-2K warming during the past 200 years recorded by many temperature profiles from eastern Canada [*Beltrami & Mareschal*, 1992]. Because we could not make temperature measurements above 30m, this recent warming is poorly constrained.

Discussion

Our report has implications for the adjustment made to continental heat flow measurements to remove the effect of the glaciations. Such adjustments are made to the heat flow data from Canada according to a model proposed by *Jessop* [1971], similar to that shown in Figure 3a. This model assumes that the temperature at the base of the glacier was -1°C during glacial episodes. In Greenland, the temperatures measured at the bedrock are -8.58 and -13.22°C for

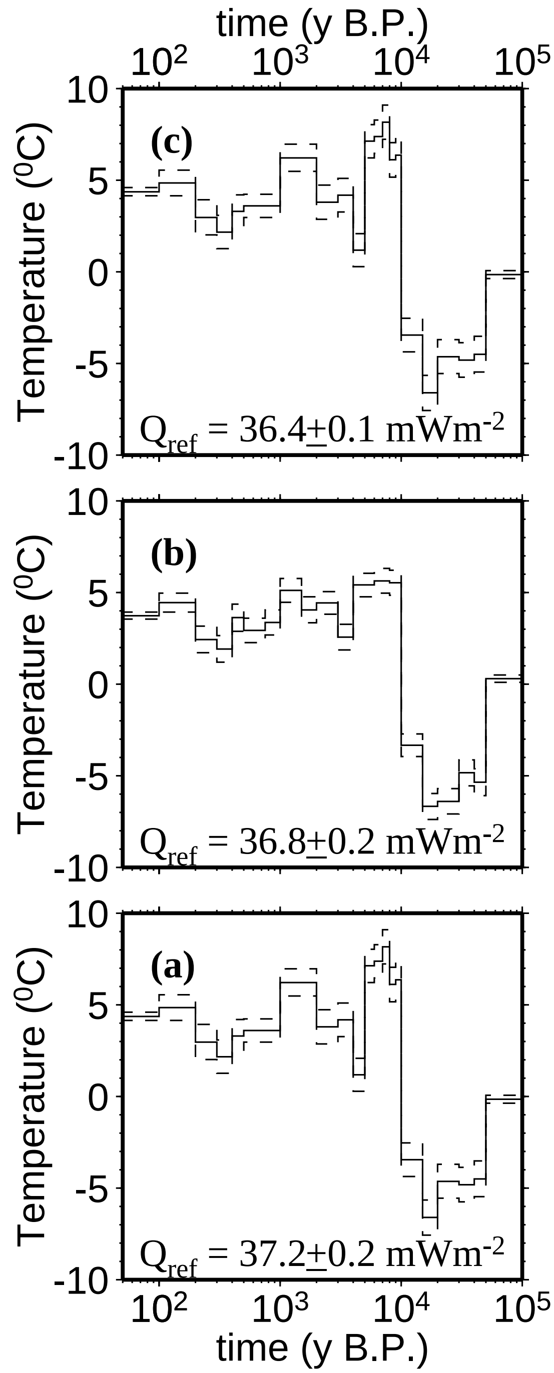

Figure 4. GSTH compatible with the heat flow profile: average value of the parameters (solid line) and standard deviation (dotted lines). The parameterization for (a), (b), and (c) is given in Table 1.

the GRIP and the Dye 3 drillhole respectively [*Dahl-Jensen et al.*, 1998]. This suggests that the adjustment to the heat flow might be underestimated, as already suggested by *Beck* [1977]. The heat flow values at Sept-Iles obtained from the shallower 1994 log were 28 and 32mW m^{-2} before and after climatic adjustment [*Guillou-Frottier et al.*, 1995]. Following our interpretation, the reference heat flow, 36mW m^{-2}, is the average heat flow beneath 1200m. The temperature conditions at the base of the ice sheet during the LGM can not be extrapolated to the entire Canadian Shield from the measurement at Sept-Iles. Like in Greenland today, the temperature during the LGM, was not uniform beneath the ice sheet. This could explain why the temperature profile in Flin-Flon does not show the effect of the LGM [*Sass et al.*, 1971].

Except for extremely deep boreholes, some uncertainty will affect the value of the deep heat flow because of the climatic effects. This uncertainty concerns not only regions that have been glaciated but the entire continents. In order to keep the heat flow data comparable, the procedure to correct for climate the heat flow values from the Canadian Shield should not be modified.

Sclater et al. [1981] observed that, without correction, average and reduced heat flow values from northern hemisphere shields are 6mW m^{-2} lower than those from the southern hemisphere. After correction, a difference of 3mW m^{-2} remains [I. Artemieva & W. Mooney, manuscript in preparation]. This difference is within the uncertainties of the climate corrections.

Heat flow data have been used to infer changes in crustal composition and structure [e.g. *Pinet et al.*, 1991]. It is unlikely that the climatic effect varies on this scale. A uniform error on the heat flow will affect the mantle heat flow estimate but not that of the crustal contribution.

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