RECONSTRUCTION OF THE HISTORICAL TEMPERATURE TREND FROM MEASUREMENTS IN A MEDIUM-LENGTH BORE HOLE ON THE LOMONOSOVFONNA PLATEAU, SVALBARD

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Abstract

A medium-length ice core was drilled at the ice divide, on the Lomonosovfonna plateau (±1230 m. a.s.l.), in Svalbard in May 1997. As part of this project temperature measurements were performed in the 120 m deep bore hole. At this site the ice thickness based on radar measurements is 126.5 m and the mean annual accumulation rate is 0.38 m w.e. The measurements over the interval from 15 m depth to 120 m depth show a nearly isothermal profile with a mean value of -2.8 ˚C and a standard deviation of 0.2 ˚C.

The measurements reveal a minimum in the temperature at a depth of approximately 70 m and a temperature gradient of 0.011±0.004 ˚C per metre near the bottom. The temperature minimum and relatively low temperature gradient cannot be explained in terms of a steady state climate. Numerical calculations with a simple 1D diffusion-advection model show that the temperature increased at a maximum rate of 0.02-0.025 K/yr over the last hundred years. The total temperature increase being estimated to be 2.0- 3.0 K.

Forcing the model with the observed record at Svalbard airport revealed that in the 19th century the surface temperature was at most 2.5 K lower and that the instrumental observations started during a period with temperatures comparable to the end of the 19th century. The data are of particular interest for historical simulations since often no other temperature data are available in subpolar areas.

1 Introduction

Ice cores contain a wealth of paleoclimatic information (e.g. Langway 1967), but the ice temperature measured in the borehole itself can also reveal information about the past climate. Atmospheric temperature acts as an upper boundary condition for the temperature within the ice sheet. This temperature signal is advected downwards and diffuses over time. At the bottom of an ice sheet the geothermal heat flux serves as a lower boundary condition for the internal temperature distribution. Ice flow might complicate the thermal profile as a result of horizontal and vertical advection, strain heating and friction near the bottom.

Several studies have been carried out to retrieve the temperature history at the surface from ice temperature measurements along ice cores or in deep bore holes in polar areas (e.g. Dahl-Jensen and Johnsen 1986, Cuffey et al. 1995, and Dahl-Jensen et al. 1998). Most of these studies focus on the temperature reconstruction on the glacial-interglacial time scale because the cores come from large ice sheets with a large ice thickness and small accumulation rates. Although Dahl-Jensen (1998) and Cuffey et al. (1994) have also derived climate information over the last century from deep boreholes, the temperature reconstruction from medium-depth boreholes reported by Paterson and Clarke (1978) on Devon Island and Nicholls and Paren (1993) on the Antarctic Peninsula has a greater similarity to the work presented in this study.

Here we consider the temperature profile in a medium-length borehole of 120 m deep drilled at the highest point of the Lomonosovfonna Plateau, Svalbard, Figure 1. Lomonosovfonna is one of the highest ice fields in Svalbard and may therefore be a suitable site for retrieving ice cores. In April 1997 a 120 m deep ice core was drilled at the ice divide at 1230 m asl (78°51'53"N, 17°25'30"E) by a Dutch-Norwegian-British-Swedish-Finnish team. The ice core analysis programme includes measurements of dielectric and electrical properties, ice structures, β-activity, oxygen isotopes, deuterium, major ions and methanesulphonate. The influence of melt on the climatological signals in the ice core are discussed by Pohjola et al. (2001). Interpretation of the chemistry of the upper part of the core covering the period 1920-1997 is presented by Isakkson et al. (2001), while O'Dwyer et al (2000) discuss the possibility of MSA measured in this core as an indicator of ocean climate.

The local ice thickness was measured using a Ramac ground-penetrating radar that generates a monopulse wavelet with a centre frequency of 50 MHz (Malå Geoscience, Sweden). Radar measurements were taken at approximately 2 m intervals on profiles to the west, south, and east of the borehole, with five closest approaches to within 20 m of the borehole, radar travel time corresponding to 119-121 m of solid ice. A correction for the lower density firn was made from detailed density measurements made on the upper 18 m of core, which corresponds roughly to the firn/ice transition depth to give a total ice thickness of 126.5 ± 1.5 m.

The ^{137}Cesium activity from the atmospheric thermo-nuclear tests was measured by high resolution γ-spectrometry. The peak level corresponds to the 1963 level, using the integrated density measurements above this level reveals an accumulation rate of 380 kg m^{-2} a^{-1}. Measurements by Gordienko et al (1980), made at ~ 250 m lower and ~ 5 km downstream at the same glacier indicate an accumulation rate of 820 kg m^{-2} a^{-1} over the period 1951-1976. There is no sound explanation for the difference. This shallow ice thickness and accumulation rate imply that only a limited time range is covered by this core. Application of a Nye-time scale (Paterson 1994) indicates that the ice at a depth of 120 metres is approximately 1000 years old. This means that possible temperature changes during the 20th century can be detected. Verification of the estimated depth-age relation could provide better insight into the vertical velocity profile, but is beyond the scope of this paper.

In this paper we will try to reconstruct the temperature history by solving the heat flow equation for an ice divide numerically, neglecting the horizontal advection. After the measurements are presented, we briefly discuss the theory. In section four we discuss time dependent solutions, and a historical simulation. The role of refreezing is considered in the section discussion and conclusions

2 Measurements

Temperature measurements were made in the 120m deep borehole at the end of drilling by taking resistance measurements using a Fluke 8060A meter (characterised by a low current excitation to reduce self-heating) on three calibrated Betatherm Corporation type .3K3A1 thermistors (nominally 300 ohms at 25°C) spaced 2m apart on a 4-core cable. The cable was lowered in 4m steps to overlap the upper and lower thermistors on each measurement cycle. Readings were taken several times at each depth to ensure the resistance had stabilised at the new temperature. Cable resistance was measured and subtracted from the thermistor resistances.

Calibration of the thermistors was carried out in the Cambridge laboratory in a cooled water bath, with salt added to depress the freezing point to –12°C (K. Makinson, BAS, *pers. comm.*). Calibration data is taken at several points in the range of –12°C to +15°C against the water bath temperature measured using an Automatic Systems Laboratories Ltd. F300 Mk2 thermometry bridge and a Tinsley Model 5187SA Primary Standard Platinum Resistance Thermometer (SPRT), with a nominal resistance of 25 ± 0.5 ohms at 0.0000°C. All temperatures use the reference T90 temperature scale and the SPRT is regularly calibrated using the following fixed points:

- Triple point of water (0.01000°C), type 16 cell ±0.001°C
- Triple point of diphenylether (26.863°C), type 16 cell ±0.005°C
- Melting point of Gallium (29.7646°C), type 16 cell ±0.001°C

In addition the thermometer has also been calibrated commercially at National Physics Laboratory against a lower fixed temperature:

- Triple point of Mercury (-38.8344°C).

Combining the calibrations of the thermometry bridge and SPRT, the calibration system has an accuracy of ±0.001°C and resolution of 0.0001°C.

With this level of calibration the thermistors should be accurate to 0.01 ± 0.005 K. However, we found that the upper-most thermistor gave results 0.1 K higher than the other two thermistors, and we assume that this thermistor had been damaged and the data were rejected. Ideally, measurements should be made several days after drilling to allow the borehole temperature to relax after the drilling operation, but this was not possible due to lack of time in the field. One set of measurements over the depth range 0 to 82m were made within hours of the completion of drilling, while a second set were made 16 hours later over the range 68 to 121m. Examination of the overlapping sections suggest the change in borehole temperature over the 16 hours was less than 0.01K, suggesting the measurements were not unduly affected by the drilling operation (which had taken four days to reach the 120m), but there is some noise evident in the lowest readings so we conservatively estimate our accuracy for the results to be no better than ± 0.05 K.

The temperature profile contains two striking features. Firstly, one can observe that the profile is nearly isothermal but with a minimum at 70 m depth. Similar reverse temperature gradients are observed for some other ice cores on Svalbard (e.g. Uchida et al. 1996 at Åsgårdfonna and Zagorodnov and Zotikov 1988 at the divide of Grönfjord Eastern and Fritjof glaciers). The observed temperature range between 15 m and 120 m depth is only 0.6 K, which is small. Secondly, one can observe that the gradient near the bottom is about 0.011 K/m in the section from 100-120. This indicates a low value for the geothermal heat flux, because a typical geothermal heat flux of 50 mW/m^2 would yield a gradient of 0.02 K/m under steady state conditions (which is of course not necessarily the case as will be seen later). These two characteristics will be used in a later stage.

Note that a bottom temperature of approximately 270.5 K implies a permafrost layer of 100 to 350 m, depending on the thermal conductivity of the rock. This is within the range observed on Svalbard (J-O. Hagen, pers. communication). However, the permafrost thickness does not influence the ice temperature calculations because the lower boundary condition is prescribed in terms of a temperature gradient, as will be discussed in the next section.

3 Theory

If we consider the thermodynamic equation where heat transfer is balanced by vertical conduction and vertical advection due to ice motion we can write:

$$\frac{\partial T}{\partial t} = k\frac{\partial^2 T}{\partial z^2} - w\frac{\partial T}{\partial z} \qquad 1.1$$

In this equation k is the thermal diffusivity and w the vertical velocity. The vertical co-ordinate ranges from z=h at the surface to z=0 at the bottom and is positive upwards.

Advection in the horizontal direction is neglected as we consider an ice divide. Since the measurements indicate basal temperatures below zero friction at the bottom can be neglected as well. A simple scale analysis, as described by Cuffey et al. (1994), shows that heat production to strain heating can be neglected.

The simplest solution for this equation is the steady state solution with $\partial T/\partial t=0$. We furthermore assume that k is independent of the temperature; this assumption is justified because of the nearly isothermal profile which leads to variations in k smaller than 1% of the mean value. Finally we have to assume that the glacier itself is in balance because a changing ice flow would change the vertical velocity profile and thus the advection and temperature profile.

In order to obtain a steady state solution we impose the following boundary conditions for equation (1.1): $T=T_S$ at $z=h-15$ m (temperature at 15m depth which is a depth considered to be not affected by the seasonal cycle), and dT/dz is constant at z=0 (bottom). Integration of (1.1) (with $\partial T/\partial t=0$) yields:

$$\frac{dT}{dz} = \left[\frac{dT}{dz}\right]_{z=0} \exp\left(\frac{1}{k}\int_0^z w\,dz\right) \qquad 1.2$$

This can be solved analytically for certain functions of z. Three solutions will be considered in this analysis. We know w at the surface from the net accumulation rate and w at the bottom must be zero. The simplest solution is $w= -bz/h$, where b is the net accumulation rate. The second solution $w = -bz^2/h^2$, is more appropriate for an ice divide and has been suggested by Raymond (1983) . The third solution assumes sliding and no deformation, implying that w is everywhere equal to the surface value. The last solution should be considered as a sensitivity experiment to study the influence of the vertical velocity profile on the vertical temperature distribution.

Integration of (1.2) with a linear profile for w yields the Robin solution (Robin 1955) for steady state:

$$T - T_s = \frac{\sqrt{\pi}}{2} l \left[\frac{dT}{dz}\right]_{z=0} \{\mathrm{erf}(z/l)-\mathrm{erf}(h/l)\} \qquad 1.3$$

with $l^2=2kh/b$;b positive and

$$\mathrm{erf}(z) = \frac{2}{\sqrt{\pi}}\int_0^z \exp(-z^2)\,dz' \qquad 1.4$$

4. Time-dependent solutions

In Figure 2a we observed a significant decrease in temperature between 15 m and 70 m below the surface; this cannot be attributed to a steady state solution or uncertainty in the calculations. Therefore, we consider a non-steady state with the following parameter setting H=126.5 m, G= 25 mW/m^2, w_s=-0.38 m/yr, linear profile. For this purpose eq. 1.1 has to be solved numerically. The details of this procedure are explained in Appendix A. The vertical temperature profile is in principle determined by the temperature history, changes in ice thickness and changing geothermal heat flux. We consider cases where variations in ice thickness and geothermal heat flux can be neglected and are decoupled from variations in the temperature history. Given the limited constraints we only consider solutions with a linear trend in temperature. To resolve the trend in the temperature we simply prescribe the temperature at 15 m depth (T_s) linearly in time (t).

$$T_s = T_0 + \alpha(t - t_0) \qquad 1.5$$

At time t_0 the temperature at the z=15 m depth equals T_0 and the vertical profile is in steady state. In other words one can say that if $t < t_0$ then $T_s=T_0$. Note that we introduce three parameters to be optimised (T_0, α, t). The reason for choosing this simple approach is the uniform temperature gradient between 15 m and 70 m, which suggests a simple form of temperature forcing. In practice only a few attempts are needed to find the best solution for the rate of increase (α) and time (t). This is due to the fact that the thermal inertia leads to a temperature minimum which travels downwards with a wave velocity that depends on the perturbation.

Figure 3 shows an example of the development of a temperature minimum and the downward migration of this minimum. This shows that there are three parameters, initial temperature, forcing rate and time, which are coupled, leading to unique solutions for the temperature distribution. Changing only one parameter will change the match between observations and calculations. The best solution is found after 56 years for a temperature increase of 0.03 K/yr, see Fig. 5. Increasing or decreasing the rate of the temperature change and simultaneously changing T_0 and t does not improve the match between observed and calculated temperature, as one can see in Figure 4. Increasing the rate of temperature increase at the surface leads to an overestimation of the surface temperature and decreasing the rate of increase leads to an overestimate of the temperature minimum. This means that we have at least one satisfactory temperature history which explains the upper half of the temperature profile.

So far we have considered a solution obtained from an initial equilibrium state at t_0. The limitation of this solution is that we have to use rather a low value for the geothermal heat flux in order to get a reasonable match between the observed temperature gradient and the calculated one near the bottom. Physically there is of course no reason to assume that the initial state is in equilibrium. The observed low gradient near the bottom actually points to a non-steady state, even near the bottom of the core.

Figure 5 illustrates the evolution of the gradient between 100 and 120 m in a 126.5 m deep bore hole in the case of two different perturbations. The geothermal heat flux is 37.5 mW/m^2 in these calculations. This means that at t_0 the gradient is 0.02 K/m in this depth interval under steady state conditions. The gradient observed was 0.011 K/m. If the boundary condition at the surface is changed the temperature profile has to adjust in time with this forcing. Diffusion and advection will slowly change the temperature from the top to the bottom. As a direct result the gradient near the bottom will decrease, see Fig 6. One can observe that soon after the perturbation at the surface has terminated the temperature gradient near the bottom increases again. The initial steady state gradient will be reached after about 500 years. The typical response time (1 - 1/e^2), before the temperature gradient is close to the equilibrium value again is about 228 yrs (which is independent of the magnitude of the temperature increase).

The results presented in Figure 5 show that an observed gradient of 0.01 K/m can be explained by a perturbation at the surface, even if the geothermal heat flux is 37.5 mW/m^2, instead of the lower value of 25 mW/m^2 which we used in the previous experiment (Figure 4) to explain the gradient in the upper half of the profile. Figure 6 shows how the gradient near the bottom reduces for five different rates of increase in the 15 m temperature. One can observe that, in the case of a larger perturbation, the gradient is reduced faster. With a slow increase of 1 K/100 yrs it takes a very long time for the gradient to reduce to 0.01 K/m. A perturbation of 4 K/100 yrs leads to a 50% reduction in ±100 years.

In summary the results obtained so far indicate that the increase in temperature in the upper half of the profile can be explained by a temperature increase, but one needs an extremely low value for the geothermal heat flux (figures 4 and 5). On the other hand, the calculations in Figure 5 and 7 suggest that it is possible to explain the low observed gradient near the bottom, even without an extremely low value for the geothermal heat flux.

A more rigorous treatment of the time-dependent solutions is to optimise the parameters such that the root mean square error between observations and measurements along the entire bore hole is minimised. For this purpose the 5-dimensional search space determined by:

$$\left(\begin{array}{cr} 37.5 \leq G \leq 62.5 & mW/m^2 \\ 0 \leq \alpha \leq 0.05 & K/yr \\ 0 \leq t \leq 300 & yrs \\ -12 \leq T_0 \leq 0 & ^\circ C \\ w_s = -0.25, -0.38, -0.75, \text{ linear or quadratic} & m/yr \end{array}\right)$$

has been screened for optimal solutions. Figure 7 shows a typical result for G=37.5 mW/m^2, α=0.02 K/yr, w_S=-0.38 m/yr and a quadratic profile. It is obvious that there is a narrow time band with optimal solutions for a given setting of G,α,W,T_0. A minimum root mean square error of 0.022 K is found for the above-mentioned parameter setting 102 years after the start of the perturbation. Nevertheless, small changes of T_0 still yield reasonable results, but at different times. However, this hardly affects the optimal value for α, which is of primary interest. Differences between solutions are not always statistically significant so it does not make sense to define one single best solution. Results are, for example, not very sensitive to the magnitude of the vertical profile or the shape of the vertical profile, but for the parameter of key interest, α, sensitivity is higher.

The best solutions in the search space defined above are presented in Figure 8a. It turns out that the geothermal heat flux should be around 37.5 mW/m^2 and the rate of temperature increase 0.025 K/year. It should be noted that slightly better profiles can be found if a smaller value is used for the geothermal heat flux. However, this seems to contradict with the sparse data on the geothermal heat flux. Liestøl (1977) presented temperature profiles in Svalbard which indicated a geothermal heat flux of 40 mW/m^2. On the other hand these measurements are only indicative since they were made in other parts of the island.

The statistically best parabolic fit to the measurements yields a root mean square difference of 0.02 K, whereas the optimised model yields 0.022 K. The small remaining difference between measurements and calculations can be attributed to noise in the observations or a more complicated temperature increase in time. One cannot expect a statistically better result from the advection diffusion model used.

Larger values for the geothermal heat flux result in poorer results, as can be observed in Figure 8b for G=50 mW/m^2. All profiles are too steep near the bottom irrespective of the assumptions for the vertical profile, and also show a temperature minimum that is both deeper and colder than observed. The experiments presented in the figures 8 and 9 show the best statistical solution within the range defined.

Experiments with larger values for the vertical velocity at the surface also showed that the best results are obtained for a geothermal heat flux of 37.5 mW/m^2.

Figure 8c shows the optimal solutions for somewhat higher values of the temperature trend. Results are somewhat poorer than presented in figure 8a, but hardly significant.

An independent test of the validity of these experiments can be obtained by a historical experiment. For this purpose we ran the model with the observations of the longest temperature record from Svalbard airport (Figure 9). This record covers the period 1912-1996. Before this period we used a constant temperature. The constant was optimised in order to find the best modelled temperature profile at the end of the run in 1996. The best profile was again defined as the profile with the smallest root mean square error between model and observations. The temperature in the 19th century was found to be 2.4 K colder than the average over the period 1912-1996. This experiment showed firstly that the estimated temperature increase from minimising the root mean square error over the entire profile as presented in the figure 7 and 9 was not in contradiction with the meteorological observations. Secondly, it showed that the temperature difference between the 20th century and the 19th century is about 2.4K and that the record of observations started (1912-1920) in a relatively cold period, comparable to the conditions at the end of the 19th century.

5 Discussion and Conclusions

The temperature observations certainly cannot be explained by a steady state solution because the difference in temperature between surface and bottom is so small and because of the fact that we observe a temperature minimum. For this reason we performed time dependent experiments. One of the uncertainties in this kind of calculation is due to the lack of a vertical velocity profile. However, by varying the vertical velocity between 0.25 and 0.75 m/yr a series of results are obtained which can be compared with the measured data. These solutions indicate that the rate of increase in temperature (α), which is of primary interest, is 0.02-0.025 K/yr, but that the elapsed time (t) for the best solution varies. The elapsed time of the solutions in Figure 8a is 112-151 yr and from 80 -107 yr for the results in Figure 8c. This implies that the range of temperature increase (αt) varies from 2.2 to 3.0 K for the results in Figure 8a and from 2.0-2.7 K for those in Figure 8c. This paper therefore shows that ice core temperatures from medium-length ice cores at an ice divide can be used for the reconstruction of the temperature over the last 100 years, despite the fact that the vertical velocity profile is not known precisely.

As indicated in the previous section a statistically better solution can be found for extremely low values of the geothermal heat flux. Figure 10 shows a contour diagram of the root mean squared error as a function of the geothermal heat flux and the temperature increase. The figure shows that the absolute minimum is found for a scaled geothermal heat flux of 50% (=25mW/m^2). The black dots in the figure indicate the position of three optimal solutions which are presented in the Figures 8a,8b and 8c. Since the results presented in Figure 8a are better, especially near the bottom, than those in Figure 8c, one can conclude from Figure 10 that increasing the geothermal heat flux will not improve the results. Since we consider the temperature increase as the main parameter and the geothermal heat flux as an important source of uncertainty we can conclude from Figure 10 that the uncertainty in the temperature increase does not depend very much on the uncertainty in the geothermal heat flux. The line indicates the dependence of the temperature increase on the geothermal heat flux. A vertical line would indicate that the predicted temperature increase is independent from the geothermal heat flux. For a realistic range of 37.5 mW/m^2 to 50 mW/m^2 the temperature increase ranges from 0.02 K/yr to 0.025 K/yr. Higher values of the geothermal heat flux yield bad solutions and lower values of the geothermal heat flux are considered to be physically unrealistic. In fact the low values for the geothermal heat flux might be explained by the fact that we neglect the heat capacity of the underlying bedrock which tends to surpress the heating near the bottom. Including this effect might improve the solutions near the bottom, but is not influencing the main result of the calculated temperature increase.

The historical experiment described at the end of section 4 showed that the temperature in the 19th century was estimated to be 2.4 K colder than the mean temperature over the period 1912-1996 and that the temperature trend derived from the borehole temperatures does not contradict the meteorological observations at Svalbard airport. Air temperature records for Svalbard over the period 1912-1920 are also 2.4K lower than the mean of the entire record, indicating that temperatures at the start of the record were comparable to temperatures at the end of the 19th century.

One might wonder whether no other temperature history can lead to a snap-shot which matches the observations as well as the solution proposed here. Here, we considered first-order solutions with a simple linear trend in time. As no non-linearities are involved one might expect to find higher frequencies with the same trend which explain the observations as well. Rapid variations in time will be diffused and will certainly lead to identical results. This however, is of less climatological importance since it is the trend that counts. No evidence could be found for a step wise change in the forcing at the surface.

One might also argue that changes in temperature are coupled to changes in accumulation rate and therefore influence the results obtained here by increasing the vertical advection. However, this is probably not the case as increased accumulation will also increase the ice thickness which compensates the effect of increased accumulation. To be able to understand fully the non-steady state ice thickness and vertical velocity effect one would need a thermodynamic ice flow model. However, this is not feasible for this site due to a lack of constraints.

The largest source of uncertainty is the role of refreezing. At the warmest point in the profile at 15 m, the ice temperature is probably higher than the mean annual temperature (unknown at this site) due mainly to refreezing. However, structural analyses on the ice core indicates that water percolates less than a meter. Calculations with the vertical advection and diffusion model show that nearly all heat from the refreezing in summer escapes by diffusion back to the atmosphere. Only deep percolation, to about 10 meters over the short one month summer period can lead to a substantial increase in the 15 m temperature. Our reconstructed temperature increase should be taken as a temperature increase at 15 m below the surface and not a surface temperature or atmospheric temperature. The increase in 15 m temperature could imply either an equivalent air temperature increase or an increase in the refreezing rate. The likelihood is that both refreezing and air temperature have changed because if the air temperature rises at this site the melting and refreezing rate will increase as well. However, it is difficult to estimate the importance of changes in refreezing since the yearly mean temperature of this site is unknown, and our calculated trend is therefore an upper limit for the increase in air temperature.

And finally, if we consider the solutions presented in the Figures 8a and 8c as the best solutions, we can calculate the temperature evolution near the bottom in the future, under the assumption that we neglect the heat generation from melting of the permafrost layer. We speculate that roughly 90 years from now temperatures near the bottom will be at pressure melting point. This means that the increasing temperatures near the surface as reconstructed will probably lead to changes in this ice field in near future.

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Appendix A: Numerical details

Equation 1.1 is solved on a equidistant grid with 5.575 m grid point distance, which equals 5% of the ice thickness below 15 m depth. The finite difference equation describing the change in temperature is solved implicitly. The initial temperature profile is uniform with depth and equal to the surface temperature. For given boundary conditions the solver iterates as long as temperature changes at one of the levels by more than 10^{-5} K. The numerical solution for steady state with a linear velocity profile is compared with the analytical solution presented by eqs. 1.3 and 1.4. Temperatures differ less than 10^{-3} K at any level.

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Figure Captions

Figure 1: The Lomonosovfonna drilling site at 1230 m. a.s.l.

Figure 2: (a) Measurements of the temperature in the bore hole; (b) contains the same data as (a), but on a stretched horizontal scale.

Figure 3: Temperature evolution in time for α=0.03 K/yr. Shown are the initial profile and temperature profiles after 25 and 50 years. The thick line represents the observations. Note that the initial profile is determined by the observations near the bottom. Parameters: H=126.5 m, G= 25 mW/m^2, w_s=-0.38 m/yr, linear profile.

Figure 4: Temperature profiles for different values of α (K/yr). Shown are the snap shots in time which resulted in the best match between observation and calculation for a particular value of α. Parameters: H=126.5 , G= 25 mW/m^2, w_s=-0.38 m/yr, linear profile.

Figure 5: The evolution of the temperature gradient near the bottom of the core in the case of an increase of 0.02 and 0.04 K/yr over a period of a hundred years (indicated by the vertical line). The horizontal line indicates the observed gradient in this section. Parameters: H= 126.5 m, G= 37.5 mW/m^2, w_s=-0.38 m/yr, linear profile.

Figure 6: The evolution of the temperature gradient in the interval between a depth of 100 and 120 metres for an increase in temperature of 0.01 to 0.05 K/yr. Parameters: H=126.5 m, G=37.5 mW/m^2, w_s=-0.38 m/yr, linear profile.

Figure 7: Typical example of the root mean square error between the observations and the calculations for the variables T_0 and t. The other parameters are constant:H=126.5 m, G=37.5 mW/m^2, α=0.025K/yr, w_s= -0.38 m/yr and a quadratic profile.

Figure 8: Optimal solutions for different vertical velocity profiles. The geothermal heat flux is 37.5 mW/m^2 for (a and c) and 50 mW/m^2 for (b). H=126.5 m, α=0.025 K/yr for (a and b) and α=0.02 K/yr for (c). The year mentioned in the legend is the elapsed time (t) for the optimal solution (see eq. 1.5). The left column of three

figures shows the absolute temperatures and the right column the difference between observed and modelled temperatures.

Figure 9: Mean annual air temperature at Svalbard airport (Hansen-Bauer et al. 1990) updated with some improvements by Hansen-Bauer, DMNI. Mean temperature over the period 1912-1996 is -6.3˚C.

Figure 10: Root mean squared error between the observations and the calculations for the variables temperature increase and scaled geothermal heat flux. The geothermal heat flux is scaled by 50 mW/m^2. H=126.5 m, w_s=-0.38 m/yr and a quadratic profile. The black dots are solutions which are presented in the figures 8a and 8b and 8c as a function of depth. The line indicates the local minima as a function of the geothermal heat flux. The shaded area is considered to be physically unrealistic.