Temperature in a glacier: Part II Why surface temperature history is accurately preserved.

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Background

Geophysicists use T-d curves to obtain past climate, or more accurately, histories of ground surface temperature. As others have suggested (1), and I have attempted to show elsewhere (2,3) past climate is difficult to obtain from these data for time periods beyond a century or two into the past. The reason for this is that the means of heat transfer in solid rock, thermal diffusion, or in porous or fractured rock, combined thermal diffusion and convection, are irreversible processes, which lead to a rapid dissipation of the information needed to construct temperature history. In a glacier, however, a different means of heat transfer is available that is reversible, and which approximates wave propagation. This explains why surface temperatures are available over tens of thousands of years in glacier boreholes (4,5).

Heat Transfer in a Glacier

Heat transfer in a glacier differs from that in rock or soil in one important way. Thermal convection in a glacier involves flow of the entire material medium; whereas, in rock what convection occurs (on a short time scale at least) is confined to fluid moving through pores and fractures. Unless the material is isothermal, convection in porous rock necessarily involves mixing fluids at one temperature with rock at another. This mixing is an irreversible process, which dissipates temperature information. In a glacier the entire medium participates in convection, which means that heat transfer doesn't involve mixing(6).

A simple model of flow in a glacier is as follows. As snow accumulates upon the glacier, its weight bears upon the snow below, which compresses the glacier vertically. At the same time the glacier moves laterally. At the bedrock surface the vertical strain stops completely and the resulting flow is entirely lateral. An extremely simplified model of this motion is the complex velocity potential $f\,(w) = \frac{1}{2}\,\varepsilon w^2$; where ε is the vertical strain rate of the glacier and w is the complex variable $x+jz$. At any location the vertical velocity of material is $-\varepsilon z$ where z is the height above the bedrock. Using this flow model, the 1-D equation for heat transfer is

1. $\alpha\partial^2 T/\partial z^2 + \varepsilon z \partial T/\partial z = \partial T/\partial t$

Use the changes of variable, $z = ly$ where l is the thickness of the glacier, and $\tau=\varepsilon t$, to transform the differential equation into dimensionless form(7).

2. $\delta\partial^2T/\partial y^2+y\partial T/\partial y =\partial T/\partial\tau$

The variable y ranges from 0 at the bedrock surface to 1 at the glacier surface. The single parameter $\delta=(\alpha/\varepsilon l^2)$ is of the order of 0.04 for the Greenland Ice Sheet. Because δ is so small, we may consider neglecting the first term in the differential equation as long as y is not too near 0. The result is

3. $y\partial T/\partial y -\partial T/\partial\tau=0$

The solution to which is $T=F(\mathrm{Log}(y)+\tau)$. At the surface of the glacier, y=1, which shows that $F(\tau)$ is simply the scaled history of surface temperature at the site. Because no smoothing of the function $F(\tau)$ is involved, material is incorporated and conveyed to depth conserving its temperature. Of course, there is a little diffusion of heat near the ground surface and within the flowing ice, so that past temperature cannot be recovered perfectly. In particular, temperature variations with less than a century duration are not preserved well. Near the base of the ice sheet, the vertical strain is insignificant, and a diffusive boundary layer develops which eventually ends the possibility of recovering long past temperature.

In the previous paper, Part I, I showed the idealized steady solution as being approximately isothermal to a depth of 2/3 of the thickness of the ice sheet. Thus, the diffusive boundary layer occupies, at most, the bottom 1/3 of the glacier. The time required to reach this depth is of the order of $1/\varepsilon$. In the case of the Greenland Ice Sheet this is about 10,000 years. Beyond 10,000 years, approximately, diffusion becomes important, and past temperatures are recovered with much diminished resolution.

References and Notes

1/ MacAyeal, D. 1995. Challenging an ice-core paleothermometer. Science 270, 444-445. MacAyeal cites R. L. Parker's Geophysical Inverse Theory as saying that borehole paleothermometry is "…worthless for all practical purposes." My reading of Parker's statement is not that he condemns all such efforts, but only that one which he specifically addresses. Nevertheless, the method is practically worthless for many purposes.

2/ Kilty, K. T. 1997&1998, Past climate and borehole temperatures. The paper is available on my web site in PDF form. It got me into such an ugly battle with a reviewer at Geophysics that I withdrew it from consideration. Nature declined to consider it.

3/ Kilty, K. T. 1998, Technical comment concerning a paper by Pollack et al in Science. The editor appeared interested in the comment and then decided not to publish it. The entire comment is available on my web site in HTML format.

4/ Dahl-Jensen, D., Mosegaard, K., Gunderstrup, N., Clow, G.D., Johnson, S. J., Hansen, A.W., Balling, N. 1998. Past Temperatures Directly from the Greenland Ice Sheet. Science, **282**, 268-271.

5/ Cuffey, Kurt M., Clow, Gary D., Alley, Richard B., Minze, Stuiver, Waddington, Edwin D., and Saltus, Richard W. 1995. Large Arctic Temperature Change at the Wisconsin-Holocene Glacial Transition. Science. **270**. 455-458.

6/ Actually a sufficiently large groundwater flow could accomplish the same thing, but the magnitude of the flow that is required is not realistic over the necessary depth range.

7/ There is an alternate scale length available in this problem; that being $\sqrt{(\alpha/\varepsilon)}$. However, this does not change the behavior of the diffferential equation.