GROUNDWATER HEAT FLUX INTO A SMALL HEADWATER STREAM DURING THE WINTER

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ABSTRACT

Water temperatures during the winter period in small headwater streams can remain above the freezing point for a major portion of the time, even if a snow or ice cover exists over the channel. Two factors are primarily responsible: the air gap that develops between the bottom of the ice/snow cover and the water surface and the thermal energy from groundwater inflow.

This paper examines a portion of the field data collected during the 1992-93 winter season. The ice cover was generally less than 5 cm thick, it formed early in the season and the remainder of the cover developed over it as a seasonal snowpack and it reached a thickness of 80 cm at the end of the season. The snow/ice cover was suspended above the water surface and supported by the stream banks, in-stream boulders and woody debris. The data analyzed included longitudinal stream temperature profiles under different winter conditions and selected the flow and temperature data at two weirs. The thermal energy flux from a groundwater source in one stream was calculated to be in the order of 40-60 W/m^2, significantly higher than possible values of frictional heating or streambed conduction. The groundwater flow in the same reach was calculated to be in the order of 0.0002 m^3/s for flows in

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January and March. This estimate of the groundwater energy flux is the first reported in the literature beneath ice covered streams and additional work is warranted to refine the field procedures and analysis.

INTRODUCTION

This study was initiated to obtain information on the spatial and temporal distribution of water temperatures beneath ice/snow-covered small streams in two adjacent basins of the Sleepers River Watershed in Danville, VT, (N44°-30', W72°-05') throughout the winter period. Longitudinal water temperature profiles were measured along the streams from their headwater position to the installed weirs to determine if groundwater inflow zones could be identified by their thermal signatures.

The two streams are located in the headwater basin, W-9 of the Sleepers River Watershed. The sub-basins are labeled W-9A and W-9B, and they have drainage areas of 20 and 18 hectares, respectively. The stream length measured in W-9A is 375 m, with another 50 m of generally dry channel except during snowmelt or rain events. Stream W-9B has a length of nearly 300 m, where it splits into two smaller tributaries of additional lengths of 40 and 150 m. The data presented in this report were collected between December 1992 and April 1993 and come from the W-9A watershed as groundwater flow was identified over a long reach of channel. The stream width at the weirs is roughly 1 m, and at the upper reaches of the streams the average width is close to 0.5 m. The water depths at low flows in the winter are in the 5- to 10-cm range. The stream slopes are steep, ranging from 0.08 to 0.23 m/m.

Both basins are entirely forested, and the soils are glacial tills and lucustrine silts (Engman, 1981). The vegetation is mainly deciduous forest, and the terrain is steep with some slopes between 15 and 25%. The elevation ranges from 600 to 700 m. The mean annual air temperature is near 5.0°C.

The streamflow was measured by the USGS at small, plywood, 90° weirs fitted with PVC plates in the notch to maintain accuracy and nape. The weirs are located about 10 m upstream from the confluence on each stream from sub-basins W9A and B. A 10.2-cm. stilling basin housed a float assembly connected to a potentiometer that measured the stage, and a Campbell Scientific Datalogger, CR-10, recorded data every 5 minutes. The accuracy of the stage readings is 0.15 cm, and the volumetric measurements confirmed the theoretical stage–discharge curve.

The water temperature at the weirs was measured using a Campbell Scientific, model 107B, thermistor probe. The probe was placed 5 m upstream of the weir in a shaded section. The temperature data were taken every 5 minutes, and the accuracy of the readings is 0.1°C. The calibration of the water temperature sensor was checked on a monthly basis using a precision thermistor.

The longitudinal stream temperatures were measured manually using a thermistor with an accuracy of 0.01°C at roughly 20- to 30-m intervals; it generally took about 20 minutes to make the measurements along the stream. Complete meteorological energy budget data are available at a site located 130 m upstream of the weir in sub-basin W9B, which is only 100 m from stream A.

BACKGROUND

Water temperature measurements are recorded at some USGS gaging sites throughout the winter period, but in the larger rivers with an ice cover and without thermal sources, the water temperatures are very close to the freezing point because of the thermal instability provided by the ice cover to grow or melt. When the streams become narrow and/or have large substrate material, suspended ice covers can develop, with air gaps between the water and the ice underside. As a result, water temperatures may stay above 0°C for extended periods due to the insulating effects of both the snow/ice cover and the air gaps, as well as through the input of thermal energy from groundwater, streambed conduction and frictional heating. Heat convected by the groundwater should be the most dominant, although no information is known to this author. Ashton (1986) stated the same.

When the river width is large, the major heat loss is through the ice surface, and the minor heat gains such as groundwater, bed conduction, frictional heating and solar radiation penetration are usually ignored when evaluating energy budgets for a stream. These minor fluxes can become important for small streams and thin ice covers.

Assessing energy budgets in the winter for small streams in the midlatitudes has received almost no attention. Water temperature measurements taken during the winter period and beneath the ice/snow cover along a stream can easily identify sources of groundwater inflow. Other clues may be a bare streambank void of a snow pack, most likely due to the melting of the snow from the near-surface groundwater, identifying a local inflow to the stream.

Small streams exhibit the same ice processes at freeze-up as some of the larger rivers, but the magnitude and duration of these processes are at significantly shorter time and spatial scales because of the groundwater thermal regimes, shallow depths, narrow widths and steep slopes. Freeze-up proceeded with anchor ice on the channel substrate and woody debris that created major blockages of the channel cross section. These blockages create temporary and localized backwater reaches of 1-2 m in length, which reduce the flow velocity and allow the formation of a surface ice cover. With the formation of the ice cover comes a reduction in the surface heat loss through the cover, and within hours or a day or so, the thermal energy within the streamflow begins melting the anchor ice deposits, which enlarges the flow area and makes the water level drop. The narrow stream width, along with the ever-present substrate of boulders and woody debris, provides structural support to keep the ice suspended when the water level drops, creating an air gap.

ENERGY AND FLOW BALANCES IN SMALL STREAMS

A small stream with an air gap and no open water leads could be considered a semi-closed system, where there is no direct connection to the atmosphere except for the conduction and convection of heat through the snow pack and ice cover. The microclimate in the air gap is not known, and only visual observations have been made. The underside of the cover is not smooth, and it includes hoarfrost ice, indicating evaporation and subsequent condensation off the water surface. It is also possible for multi-layered ice sheets to be present, with two air gaps.

The energy budget for a stream with a suspended ice/snow cover with an air gap is

$$\Phi_i = \Phi_{sw} + \Phi_{se} + \Phi_t + \Phi_b + \Phi_f + \Phi_{gw} \tag{1}$$

where Φ_i is the heat available to warm the flow, Φ_{sw} is the short-wave radiation component that passes through the snow and ice covers to the stream or bed, Φ_{se} is the sensible heat and evaporative fluxes between the air in the gap and the water (which may be positive or negative), Φ_t is the tributary heat flux, Φ_b is the bed and geothermal heat fluxes, Φ_f is the frictional heat component and Φ_{gw} is the groundwater heat flux.

Ashton (1986) reported typical values of 2–4 W/m^2 for the bed heat conduction when geothermal sources do not exist. Frictional heating is also a small term, and in these two steep mountain streams, the values would be in the range of 1–4 W/m^2. The penetration of solar radiation through a snow cover is minimal in this setting because of shading by the trees, the low sun angle and the depth of the snow over the cover. An analysis presented in Prowse and Gridley (1993) of

Geiger's work showed that with an extinction coefficient of 14, a 30-cm snow cover would attenuate the radiative flux to 1.5% of its original value; at noon a value of 3 W/m^2 might reach the stream during the winter at 45° N latitude. The remaining unknown that supplies heat to the small streams is groundwater inflow.

The groundwater thermal flux is a term that is extremely difficult to calculate, much less measure, because it is hard to determine the locations and temperatures of groundwater inflow. Yet, in small streams during the winter, the groundwater flow is often the dominant water source. No data have been presented that are known to this author that give values for the thermal fluxes from groundwater sources entering a stream.

The thermal energy passing a given point for a flowing fluid is the product of the streamflow (Q), its temperature (T_s) relative to a 0°C base (T_m), and its volumetric heat capacity (ρC_p): $\Phi_{se} = \rho C_p Q(T_s - T_m)$. The discharge and temperature data gathered at the weirs allow the computation of the stream energy flux during the winter period.

ANALYSIS OF DATA

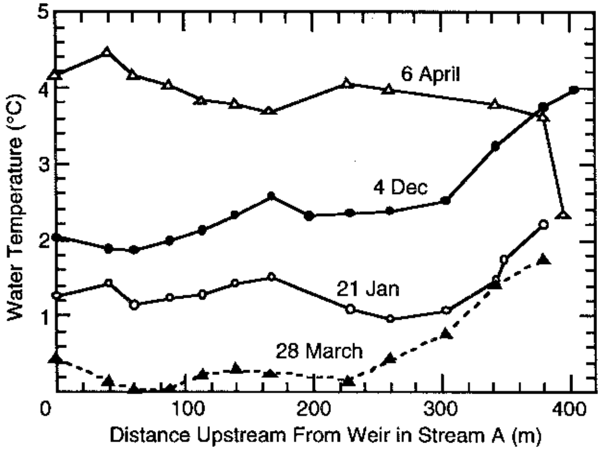
The ice cover thicknesses on both small streams ranged from 3 to 5 cm for the entire winter and the cover formed in early January. These very small thicknesses indicate that the water did not maintain contact with the ice cover for very long, and the remainder of the cover over the stream consisted of an increasing snow depth as the winter persisted. In the upper 50 m of stream A the cover was almost entirely snow, the stream widths were less than 0.5 m, and the substrate was exposed, which permitted the snow cover to bridge. The snow cover plays a significant role in insulating the stream from a rapid loss of energy, as well as cutting off solar radiation penetration.

Field Measurements

The actual measurements of the longitudinal water temperatures in the small streams started in the upper reaches of the basins and proceeded downstream toward each weir. To measure the water temperature, small holes were poked through the cover, and the thermistor was inserted into the stream.

Longitudinal profiles at four different dates are given for stream A in Figure 1. On December 4, 1992, with the air temperature below freezing, the stream had no ice cover, and the temperatures were generally above 2°C; there were two reaches

Figure 1. Water temperatures along stream A in sub-basin W-9A.



of cooling and warming. By January 15, 1993, the stream was fully ice and snow covered, and for the set of measurements on January 21, 1993, these warming and cooling trends were very distinct and could even be identified at the start of snowmelt on March 28, 1993, when the channel was still 75–80% covered. The stream was fully free of snow and ice by April 6, and the cooling and warming reaches had reversed; the significance is not understood at this time. However, measurements made during the summer low flows again confirmed the presence of these groundwater inflow reaches.

By January 21, 1993, stream A was totally covered by 20 cm of snow over a thin ice cover of 3–5 cm, with no open leads. On this date the temperatures were generally above 1°C over the entire reach. The shape of the longitudinal temperature profiles is somewhat similar for the open-water situation in December as for the ice-covered condition in January and March. Two zones where the flow is warming can be found immediately upstream of the weir (0–40 m) and then in the reach roughly between 170–260 m upstream of the wier. The first reach just upstream of the wier that shows a warming trend is actually due to a tributary inflow.

Snowmelt was just underway on March 28, 1993, and the low stream temperatures in the lower 250 m of the channel were not the sole result of incoming snowmelt water lowering the water temperature, but also of the water level rising in the channel, making contact with the ice/snow cover and melting it. The temperature profile in the upper reaches of stream A were influenced by solar radiation penetrating into open water leads within the ice/snow cover.

By April 6, 1993, stream A was entirely open, as the entire snow/ice cover had melted or collapsed into the channel. The impact of the solar radiation on the water temperature is seen by the temperatures near 4°C over the entire length. Even with the strong solar input to the channel, a cooling zone between 180 and 220 m upstream reflects the location where the groundwater inflow was warming the flow in midwinter.

One can infer a region of groundwater inflow quite convincingly from the stream temperature data for January 21, 1993. The stream is totally snow/ice covered, with an air gap, and there are no open-water reaches anywhere along the channel. Between 170 and 260 m upstream of the weir, the water temperature rose nearly 0.55°C, and from 60 to 170 m the water cooled by 0.37°C. The rise in water temperature means groundwater inflow, while the drop in the stream temperature suggests minimal or no inflow along that reach. There are no visible surface tributary inflows in this reach either.

The reach where the water temperatures rises passes partially through a much steeper section of the stream (s = 0.23), and it is flanked on both sides by steep banks some 3–5 m high, while the remainder of the channel has banks roughly 0.5 m high.

Tables 1 and 2 reflect selected flow and temperature data for both streams. Beginning in December, the flows in stream A recede much faster than stream B, and the temperatures in stream A are consistently higher than in stream B at the weirs. During the ice-cover period, the energy fluxes in stream B are greater than stream A, even though the stream temperatures are lower in stream B; this is primarily due to the higher sustained flow. On January 21 the flow in both streams was similar, and by March 4 they had decreased such that stream B had three times the flow of stream A. However, the interesting feature is that the water temperatures in each stream on these two dates are the same.

Table 1. Flow and temperature data for stream A.

Date	*Discharge (m^3/s)*	*Temp. (°C)*	*Energy flux (W)*
Dec. 4, 1992	0.0034	2.08	29,700
Jan. 15, 1993	0.0015	1	6,300
Jan. 21, 1993	0.0011	1.21	5,500
March 4, 1993	0.0002	1.22	1020
March 28, 1993	0.0053	0.43	9,570
April 6, 1993	0.0069	4.2	122,000

Note: Preliminary flow data

Table 2. Flow and temperature data for stream B.

Date	*Discharge* *(m^3/s)*	*Temp.* *(°C)*	*Energy flux* *(W)*
Dec. 4, 1992	0.0026	1.62	17,700
Jan. 15, 1993	0.0011	0.53	2,450
Jan. 21, 1993	0.0010	0.86	3,610
March 4, 1993	0.0005	0.85	1,790
March 28, 1993	0.0033	0.16	2,220
April 6, 1993	0.0046	2.86	55,300

Note: Preliminary flow data

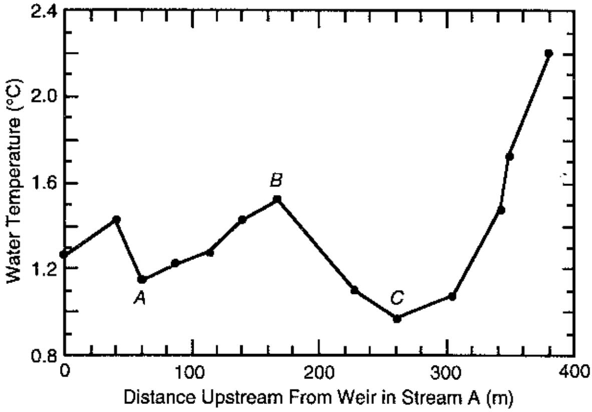
If we assume the groundwater inflow temperatures have not changed, we can make the following claim. To have nearly the same water temperature but at significantly different flows requires 1) the surface area must have been reduced by this factor over the entire channel reach, or 2) the water was emerging from lower in the basin and it had a shorter travel path, or some combination of both. It was noted on 4 March during measurements of the groundwater temperatures that the upper part of stream A did not have any measurable surface flow; no measurements in stream B were undertaken.

Groundwater Inflow Estimates

The relative input of groundwater to the total flow in stream A along one reach may be estimated using the longitudinal temperature data of January 21, 1993. The atmospheric air temperature was –5°C. The increase in temperature from 0.97° to 1.52°C between 260 and 170 m indicates a continuous increase in the energy flux of the stream. Since the air temperatures are below freezing, the sources of positive energy are coming from bed conduction, frictional heating or groundwater. The air temperature in the air gap should have been relatively stable during the time it took to take the water temperature measurements (20 minutes). Cooling of the stream water took place from 170 to 60 m.

The channel between 60 and 270 m upstream of weir A will be considered a control volume for further analysis. The channel characteristics of depth, width, slope and bed materials are similar. The upstream boundary begins at 270 m, called station C; an intermediate section at 160 m is labeled station B; and the downstream boundary at 60 m is station A (Fig. 2). We will assume that groundwater flow only enters the control volume between stations B and C, based on the water temperature measurements. The flow continuity equations can be written as

Figure 2. Water temperatures along stream A on January 21, 1993.



$$Q_b = Q_c + Q_{gw} \tag{2}$$

$$Q_a = Q_b \tag{3}$$

where Q is the flow discharge; the subscript gw refers to the groundwater inflow between sections B and C. Flow losses from the stream through the substrate are considered minimal.

The energy balance between A and B is

$$\Phi_a = \Phi_b - \Phi_{exc} \tag{4}$$

and the energy balance between B and C is

$$\Phi_b = \Phi_c + \Phi_{gw} - \Phi_{exc} \tag{5}$$

The energy fluxes are at the three sections are

$$\Phi_a = \rho C_p Q_a T_a \tag{6}$$

$$\Phi_b = \rho C_p Q_b T_b \tag{7}$$

$$\Phi_c = \rho C_p Q_c T_c \tag{8}$$

and the groundwater heat flux is

$$\Phi_{gw} = \rho C_p Q_{gw} T_{gw} \tag{9}$$

with the remaining heat fluxes lumped into the term Φ_{exc}, which includes the fluxes associated with bed conduction, frictional heating and water surface/air exchanges in the air gap.

As discussed earlier, the temperature data in reach A-B suggest little to no groundwater inflow; thus the term Φ_{exc} is evaluated between A and B, and this rate of heat loss is assumed to be the same for the B-C reach. Thus

$$\Phi_{exc} = \rho C_p Q_a (T_b - T_a) L_{bc} \bullet W_{bc} / (L_{ab} \bullet W_{ab}) \tag{10}$$

where L is the distance between B-C and A-B, and W is the stream width for the respective reaches. The stream widths are similar for both reaches. Using equations 3 through 10, a single equation results with unknowns of groundwater flow, Q_{gw}, and temperature, T_{gw}:

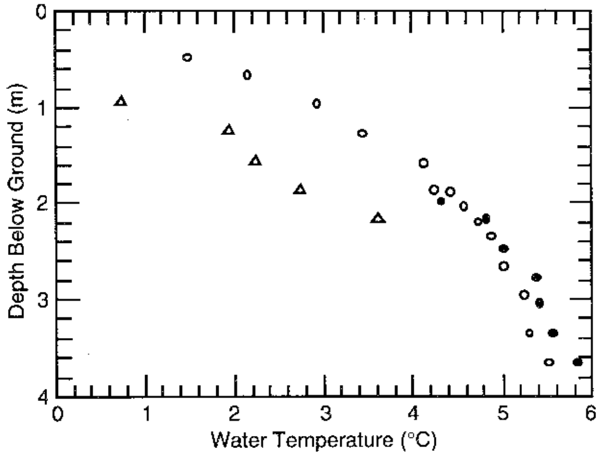
$$Q_{gw} = Q_b[T_b - T_c - (T_a - T_b)(L_{bc}/L_{ab})]/(T_{gw} - T_c) \tag{11}$$

The solution of the above equation can only be obtained by assuming values for one variable and solving for the other. Since the range in the temperature of the groundwater is known from water temperature measurements in surrounding wells and the surface temperatures at the head of the channels, the problem can be bounded. Figure 3 shows the temperature distribution from three wells located within 25 m of stream A; the temperature range is from 0.8 to 6°C. The stream temperatures at the uppermost point in these two streams were 2°C for stream A and almost 4°C for stream B. The streamflow used in the calculation below was taken from measurements at wier A, but it has been reduced by 20% to reflect flow from a small tributary that enters just upstream of weir A. Additional weirs were installed in the summer of 1993, and the flow data indicated that this factor is reasonable.

The following values were used in equation 10 for the data set of January 21, 1993: $T_a = 1.15$°C, $T_b = 1.52$°C, $T_c = 0.97$°C, $Q_b = 0.00088$ m^3/s, $L_{ab} = 110$ m and $L_{bc} = 90$ m.

Table 3 shows the results of evaluating the above equation for groundwater temperatures between 2° and 8°C. Based on visual observations of the low flows

Figure 3. Groundwater temperatures in three wells adjacent to stream A.



in the summer of 1993 along stream A, the groundwater flow component probably could not be greater than 25–35% of the flow for this reach. This means that we could further bound the problem by eliminating groundwater temperatures above 4°C. Groundwater temperatures at 8°C would be an absolute maximum.

Table 3. Solution to Equation 11.

Groundwater temperature (°C)	*Streamflow entering (m^3/s)*	*Streamflow leaving (m^3/s)*	*Groundwater flow (m^3/s)*	*% of total*
2	0.00015	0.00088	0.00073	83
3	0.00051	0.00088	0.00037	42
4	0.00063	0.00088	0.00025	28
5	0.00069	0.00088	0.00019	22
6	0.00073	0.00088	0.00015	17
7	0.00076	0.00088	0.00012	14
8	0.00077	0.00088	0.00011	13

If we assume that 6°C is the best estimate of the groundwater influx temperature based on the temperature data from the deeper groundwater wells, the corresponding flow is 0.00015 m^3/s, or 17% of the total flow. This yields an energy flux of 42 W/m^2 over the 90-m-long reach.

To check the validity of the above values for the groundwater inflow, the March 28, 1993, water temperature data set is used, even though the streamflow has increased by a factor of three. This was the start of snowmelt, and the groundwater flow contribution should still be a minor term when compared to the shallow surface runoff. The values used were $T_a = 0.03°C$, $T_b = 0.31°C$, $T_c = 0.16°C$, $Q_b = 0.0026$ m^3/s, $L_{ab} = 80$ m and $L_{bc} = 88$ m, where the flow discharge at the weir was again decreased by 20% to be consistent. The lengths of the contributing reaches changed slightly due to the increased surface flows overwhelming the groundwater inflow thermal flux.

If we assume that 6°C temperature is still a valid figure for the groundwater temperature in March, the groundwater flow is calculated to be 0.00021 m^3/s. This yields an energy flux of 57 W/m^2. This groundwater inflow value is surprisingly close to the value determined in January. An inspection of the groundwater temperatures in the deep wells for March 28 reveals that the temperature at the bottom of the deep wells has dropped to 5°C at the 4-m depth. Using 5°C as the groundwater temperature, it yields a flow of 0.00025 m^3/s and a flux of 60 W/m^2. At this time all that can be claimed without more accurate streamflow data and ground temperature probing is that it appears the values of groundwater flow are in the proper order of magnitude and the groundwater temperature that is influencing the surface flow is probably decreasing as the winter season progresses. Groundwater temperatures in the 5–6°C range seem reasonable, close to the average annual air temperature at this location; another site some 10 km away has a nonfluctuating groundwater temperature of 7.4°C at a depth of 8 m.

The groundwater source appears to occur when the channel transitions from one plateau to another via a steep, incised channel between 140 and 220 m. The depth of the channel walls is on the order of 3–5 m, and this depth is consistent with the groundwater temperatures in surrounding wells that yielded values of 5–6°C. Additional field work is warranted to further investigate this area.

These values of groundwater heat flux (43–60 W/m^2) are more than an order of magnitude higher than typical frictional or bed conduction fluxes. In an earlier paper by Calkins and Brockett (1988), we had ignored the energy flux associated with groundwater inflow as a contributor in melting anchor ice and assisting in the development of an air gap; this was an oversight in light of these new calculations, but groundwater flux values cannot be applied uniformly over the channel surface area. Each contributing reach must be identified.

SUMMARY AND CONCLUSIONS

The winter period appears to be an excellent time to estimate groundwater contributions (flow and temperature) to the surface flows, as the ice/snow cover provides a cover of insulation from the usually dominant solar radiation term. When an ice/snow cover becomes suspended and an air gap forms, minor thermal sources can be quantified. Sources of groundwater could be identified from longitudinal stream temperature profiles taken beneath the ice cover, and a preliminary estimate of the groundwater flow rate was determined from the limited stream flow and temperature data.

Estimated groundwater temperatures near 6°C in January indicated the groundwater inflow in a 90-m reach was roughly 0.00015 m^3/s. An estimated temperature of 5°C for a data set in March yielded a computed groundwater flow of 0.00025 m^3/s, but the streamflow in March was three times the January flow. Groundwater inflow heat flux values are thus in the 40–60 W/m^2 range, which indicates that it is not an insignificant term when considering energy budgets of small streams in the winter. However, these energy fluxes are only present in zones of groundwater discharge and can't be applied indiscriminately to the entire channel reach. Additional instrumentation has been installed in stream A to refine the streamflow data and temperature profiles in the upcoming winter.

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