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Groundwater Temperatures During the Winter in a Forested Basin

D.J. CALKINS
Geological Sciences Branch
U.S. Army Cold Regions Research and Engineering Laboratory
72 Lyme Road
Hanover, New Hampshire 03755-1290 U.S.A.

ABSTRACT

Vertical groundwater temperature profiles were measured in shallow (1–4 m) 5.08-cm-diameter PVC wells at several sites within a small (0.47 km^2), deciduous forested basin. All sites except one indicated that conduction was the predominant means of heat transport, as was evident by the profiles being close to linear with depth. At one site the temperature was uniform with depth, indicating that advecting fluid kept the temperature profile shape from changing significantly during the entire fall or winter season.

By the end of winter and just prior to snowmelt, groundwater temperatures at the 2-m depth ranged between 2 and 4.25°C, and were warmest at the highest elevations in the basin. The spatial variability in groundwater temperatures at the 2-m depth at all sites was nearly 2°C throughout the winter period.

The presence of a snow cover minimizes the infiltration of rainwater and the subsequent advection of heat to the soil. A preliminary analysis of the data has shown that late in winter, the assumption of a steady state heat flux can be made. Vertical upward velocities in saturated soil regions on the order of 0.05 m/day were computed from a one-dimensional steady state heat and mass flow analysis using the temperature profiles.

INTRODUCTION

This study was initiated to obtain information on the temperature regime of groundwater contained in shallow saturated soils in watershed W9 in the Sleepers River Basin. There have been very few studies that have examined the spatial and temporal changes in the vertical temperature regime of shallow groundwater systems. In watershed settings, numerous studies have addressed temperature profiles in unsaturated soils during the winter and snowmelt period, but very limited research has been focused on saturated soils.

From as long ago as the early 1930s, temperature profiles in deep boreholes have been used to estimate geothermal heat flux, and beginning in the early 1960s, calculation of fluid velocities in saturated media has been determined from the temperature profiles. In this paper, vertical velocities in shallow saturated soil media will be determined from temperature profiles measured in observation wells, and these data will be compared with values calculated from slug tests made in the wells.

SITE LOCATION

The field data were collected from a group of shallow wells in the W9 basin (0.47 km^2) of the Sleepers River Watershed in northeastern Vermont between November 1992 and April 1993. This basin is entirely forested and the soils range from glacial tills to lucustrine silts (Engman 1981). The vegetation is mainly deciduous maple and beech with a few thinly scattered conifers. The terrain in this basin is steep, with some slopes in the 15–25% range, with an elevation range of 500–660 m. The mean annual air temperature is approximately 4.5°C.

The holes for the groundwater wells were augered to refusal during 1991 and 1992 by the USGS. More than 40 well sites exist in the basin, but only a dozen were monitored for periodic measurement during the winter period. All wells were 5.08-cm-diameter PVC with a 0.6-m length of screened casing beginning at the bottom of each hole. The wells were sealed above the screened section and the native soil replaced around the pipe. PVC was chosen because its thermal conductivity is similar to that of most soils and for its ease of installation. Also placed in each well was a 1.03-cm-diameter ABS tubing used for withdrawing water for chemical analysis. The effective radius of the well was approximately 2.0 cm.

METHODS

Equipment

The measurements of the water temperature in the wells were made using factory-calibrated thermistors with an accuracy of 0.01°C at the calibration points and 0.02°C between the calibration points. The readout device was a 4.5 digit Fluke DVM (8060A); its accuracy was generally within one ohm at ten thousand ohms. Each thermistor was placed in a small perforated PVC cylinder (0.63 cm o.d., 3.8 cm long with an i.d. of 0.47 cm) to protect it from damage.

A thermistor string was fabricated by taping seven individual thermistors to the outside of a 0.095-cm-diameter, 175-cm-long fiberglass rod (227 gm). The thermistors were spaced 30.5 cm apart beginning at the bottom of the rod, while the seventh was 17.5 cm above the sixth one. The seven pairs of thermistor leads were soldered and waterproofed to a 9-m-long lead cable

On 26 March 1993, fixed thermistor string arrays were installed in seven wells that represented the range of groundwater temperature conditions in the watershed. The thermistors used were 0.1°C interchangeable. The thermistor string was installed inside a 1.6-cm-i.d. PVC tube (o.d. = 2.2 cm). This and another tube of the same o.d. were inserted in the 5.08-cm well so that these wells contained three tubes with an effective radius of 1.25 cm.

Procedures

For the portable thermistor array, the field procedure was to carefully lower the thermistor string down the groundwater wells until the uppermost thermistor was submerged just below the water level; if the depth of water was less than 1.7 m, the rod rested on the bottom of the well. The thermistors were allowed to reach equilibrium, which generally occurred within three to five minutes. If the depth of water was greater than 1.70 m, the rod was lowered another 1.50 m and the readings were recorded again. No wells had water depths greater than 3.4 m, so the thermistor string was lowered only once after the initial placement in the well and this allowed overlap for some temperature readings.

The measurement technique inherently induces some error into the actual temperature profile of the water in the well. The time to reach a stable reading is thus a function of how much the fluid was disturbed when the rod was inserted, the residual heating/cooling with the thermistor string, and the temperature decay constant for the thermistor. Thermal instability due to induced fluid motion caused by the insertion of the thermistor string into a well occurred once. Natural convection is also possible at times, but due to the intrusive measurement technique, reliable measurements would not be possible if this was occurring.

The heat storage of the fiberglass rod and lead wires was calculated to be roughly 250 J/°C. The change in water temperature caused by inserting the thermistor string with a temperature difference of 5°C into a 5.08-cm well of the same depth would be less than 0.1°C. To minimize this effect, temperature measurements during the winter period were always made when the air temperature was near or above freezing.

Verification of minimal flow convection in wells

On several occasions data was taken from two sets of nested wells (1 m apart) of different depths to determine if any significant deviation in water temperature had occurred that might be attributed to natural convection of fluid within the well casing. The movement of water by thermal convection in the wells would not represent the true groundwater temperature outside the well. To ensure that natural convection would not be a problem during the summer period, a bundle of tubes was installed inside the 5.08-cm-diameter wells to reduce the effective radius to near 1.25 cm. As a result, data can be collected year-round.

The data for Wells 1 and 2 on January 15 and April 13 are given in Figure 1. The temperature profiles on 15 January were taken with the 7-string thermistor array and the temperature profiles are quite similar and nearly coincident (they differ by 0.2°C near the surface). For the April date during snowmelt, the temperatures in Well 1 were measured with the in-place array, while the temperatures in Well 2 were measured with the portable array. The temperature difference was greatest near the surface, with about a 0.25°C difference at 0.5 m below ground surface; at 2-m depth the profiles had merged. Based on soil temperature measurements at other sites, this temperature difference is within the natural spatial variability.

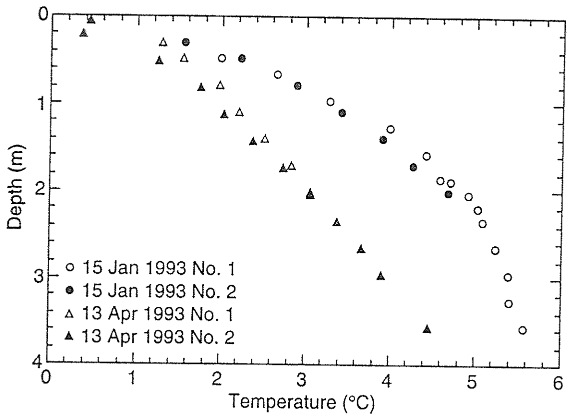

Figure 1. Temperature profiles for Wells 1 and 2, 21 January and 13 April 1993.

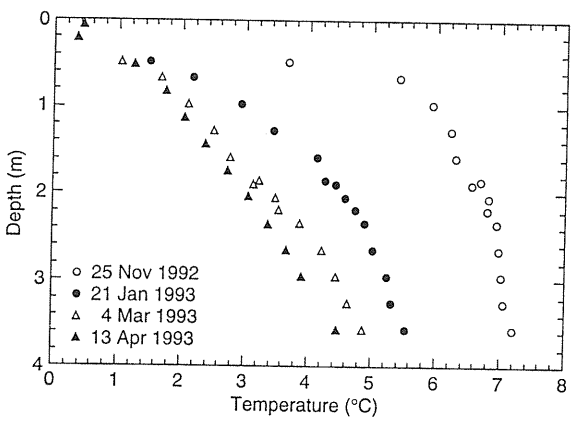

Figure 2. Temperature profiles at Well 1.

This paired data, along with other data and the details of the analysis in Calkins (1993), demonstrated that natural convection within the well does not appear to have any major influence on the temperature profile and that the well temperatures represent the surrounding groundwater temperatures.

Field data

Data from two wells (1 and 8) that represent the spatial and temporal variability in the temperature profiles for four dates during the winter season are selected for presentation and analysis. The saturated soil temperature profile shapes for Well 1 (Figure 2) during the winter period is typical for the majority of the sites in the basin, but Well 8 (Figure 3) displays a nearly uniform temperature profile with depth from 0.5 m to 2.7 m below the ground surface.

The temperature profile for Well 1 shows a gradual cooling from fall through the winter period; by the end of winter the temperature decrease is almost linear with depth. A linear temperature profile means conduction is the dominant process of heat transport. Because the wells were shallow, the depth at which

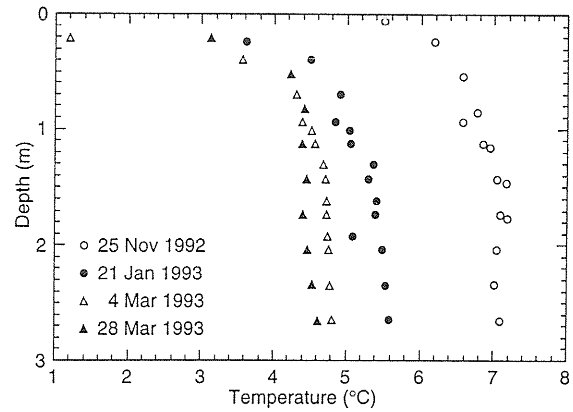

Figure 3. Temperature profiles at Well 8.

the surface heat fluxes do not influence the temperature regime could not be measured at these sites. At another site in the Sleepers River Watershed about 10 kilometers to the east, the depth where a constant temperature of 7.4°C exists appears to be roughly eight meters below the ground surface.

The temperature profile shape for Well 8 is distinctly different from all of the other wells, showing an almost uniform temperature distribution with depth except near the surface. This uniform temperature profile existed in October and was maintained for the entire winter period. This temperature profile shape is indicative of a pronounced upward vertical movement of water from the local groundwater system in this area.

The siting of Well 8 was made without any prior knowledge of the subsurface thermal regimes, but subsequent soil temperature probing in the immediate area does indicate a significantly warmer region than other areas probed in the basin.

To gain a better appreciation of the spatial distribution of the groundwater temperatures in this small basin, Figures 4 and 5 present the composite data from all the wells sampled for 21 January 1992 and 28–30 March 1993, respectively. Time and space do not offer a complete assessment of the temperature profiles for each site, but the following observations can be made. Groundwater temperatures at the 2-m depth in November ranged from roughly 5–7°C, and by the end of March the range was 2 to 4.25°C at this depth.

Another interesting feature of the data is that the groundwater temperature in the wells at the higher elevations (7, 8, 9, 11 and 15) were all warmer than the temperature in the wells at the lower elevations at the end of the winter season, with the exception of Well 27. It is not clear why this occurred; possibly the higher elevation wells have deeper unsaturated surface regions.

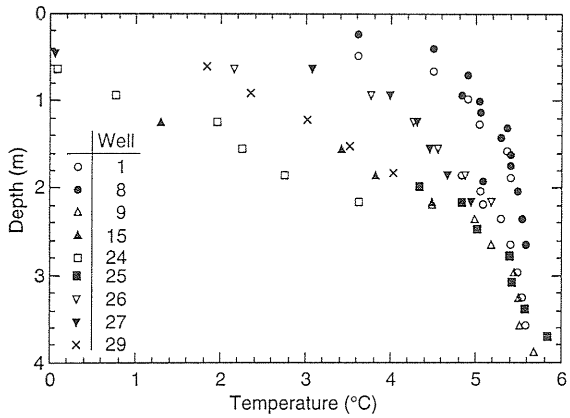

Figure 4. Temperature profiles on 21 January 1993.

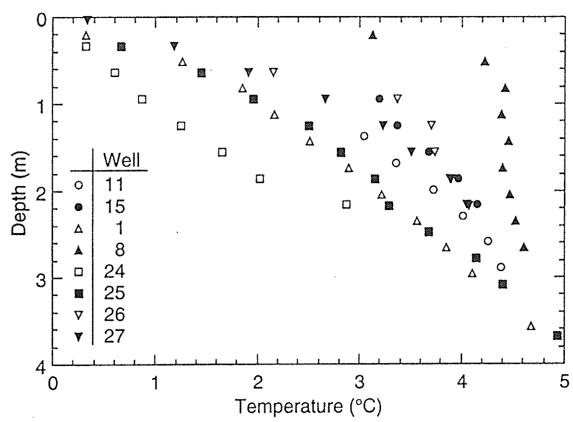

Figure 5. Temperature profiles, 28–30 March 1993.

The groundwater temperatures at Well 24 cooled at a faster rate than the other sites. This site is located in a hardwood wetland at the base of a steep, well-drained hillslope. A small perennial stream flows through the wetland area. The profile shape is opposite all of the others, which suggests vertical downward advection of heat by fluid flow.

Vertical groundwater flow

The governing differential equation describing a one-dimensional vertical simultaneous nonsteady heat and mass flow in a homogeneous, isotropic and completely saturated porous media can be found in many texts. If we assume that the vertical transport of heat and fluid is steady state and fluid flow is uniform with depth, the equation is

$$\frac{\partial^2 T}{\partial z^2} - \frac{\rho_o c_o V_z}{k}\frac{\partial T}{\partial z} = 0 \qquad (1)$$

where T = temperature at any position z;
z = vertical position;
c_o = specific heat of the fluid;
ρ_o = density of the fluid;
k = thermal conductivity of the fluid/soil matrix; and
V_z = fluid velocity in vertical direction.

The assumption of a steady, uniform vertical upward flow occurring during the winter period in the presence of a persistent snow cover is the basic hypothesis that is being tested. By midwinter the downward infiltration of rain or early fall snowmelt has completely passed through the unsaturated zone and will not affect the saturated soil medium. Soil moisture data collected from TDR measurements by the USGS in this small watershed have shown that the response in the volumetric soil moisture content to rainfall events are relatively rapid; after a week or so, the impact of the event is hard to see. It is therefore assumed that after 2–3 weeks of a snow cover without rain or snowmelt, temperature profiles should not be under the influence of a downward percolation.

Bredehoeft and Papadopulos (1965) have given a solution to equation (1) such that the temperature distribution with depth is

$$\frac{T_z - T_o}{T_L - T_o} = \frac{\left[\exp\left(\frac{Bz}{L}\right) - 1\right]}{\exp(B) - 1} \qquad (2)$$

where T_o = upper boundary temperature;
T_L = lower boundary value;
L = total depth of profile between T_o to T_L;
B = $c_o \rho_o V_z L/K$ = a dimensionless Peclet parameter.

The Peclet number can be positive or negative depending on the direction of groundwater flow. High positive or negative values of the Peclet number indicate convection-dominated flow. Bredehoeft and Papadopulos' analysis also indicated that the lowest value of the Peclet number was 0.5 or –0.5 for detecting groundwater flow convection.

The nondimensional depth-temperature profiles for Wells 1 and 8 are plotted in Figures 6 and 7. For Well 1 in Figure 6, the nondimensional temperature profiles change with time, indicating that a steady state condition of heat transport was not attained until March. The profiles in March and April were very similar; therefore the use of equation 2 to solve for the advecting velocity term is now valid after March. The best fit B value is close to –0.5, yielding an upward velocity of roughly 0.003 m/day using a thermal conductivity of 1 W/m-°C and a heat capacity of 4.2×106 J/m^3-°C for the fluid. This value is in the same order of magnitude as that calculated by

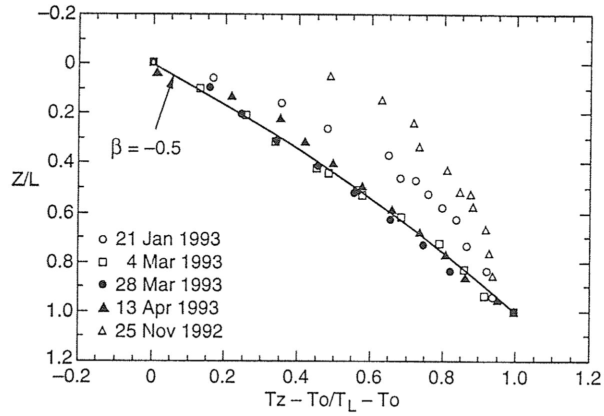

Figure 6. Nondimensional profiles at Well 1.

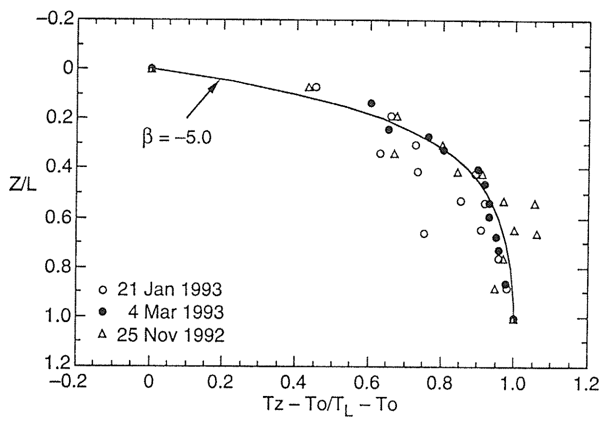

Figure 7. Nondimensional profiles at Well 8.

Thomas (1992) using mini-slug tests for Well 6, which is nearby (50 m) and in the same overburden material.

The profiles in Well 8 (Figure 7) show similar patterns for measurements taken in December, January and March. The scatter in the data is likely due to the thermistors not reaching an equilibrium temperature or the result of convection induced by inserting the temperature probe. The similarity in profiles over the winter season indicates that steady state heat and fluid flow are reasonable assumptions for this site. A curve with a $B = -5.0$ seems to fit the data reasonably well and this yields a vertical velocity of roughly 0.04 m/day. This value is an order of magnitude greater than that determined for Well 1, but closer to the value of 0.1 m/day for Well 9 (70 m away) as determined from slug tests (Thomas 1992).

CONCLUSIONS

Initial observations have been presented for groundwater temperature distributions in a small forested watershed. This year's observations indicate that the warmest groundwater temperatures in the saturated soil regimes were at the highest elevations, which comes as somewhat of a surprise; continued monitoring will assess this trend next winter.

Well 24, located in a wetland at the bottom of the steep slope of a hardwood stand, cooled to the lowest temperatures of all the sites. Well 8, one of the highest wells in the watershed, had the warmest groundwater temperatures throughout the winter. Subsequent soil probing to determine temperatures in the near-surface zone around Well 8 confirmed the presence of a warmer-than-normal region. Well 8 is situated just below a very steep, 50-m-high rock outcrop; it is quite possible that groundwater flow is emerging due to the abrupt change in slope and a more highly permeable soil horizon. The temperature in this well was also the coldest during the summer and early fall of all the sites in this watershed. Although both wells are at the base of steep slopes, their thermal regimes are significantly different.

The computation of the vertical velocities for Wells 1 and 8 from the temperature profiles using a steady state one-dimensional heat and mass flow analysis are remarkably close to the values computed from mini-slug tests in nearby wells. When the nondimensional temperature profiles stabilize with time, it can be assumed that the transport of heat in the profile is steady state, but comprises both conduction and vertical fluid advection of heat. From the initial data, it appears that vertical groundwater velocities can be resolved to only 0.003 m/day.

The period when a winter snow cover persists allows the computation of a vertical velocity component without the interference of summer rain infiltration effects.

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