Analysis of Temperature Gradients to Determine Stream Exchanges with Ground Water

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Chapter 4 of

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Chapter 4
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By James E. Constantz,[1] Richard G. Niswonger,[2] and Amy E. Stewart[3]

Introduction

Heat flows continuously between surface water and adjacent ground water, and as a consequence, provides an opportunity to use heat as a natural tracer of water movement between the surface and the underlying sediments. By the early 1900s, researchers intuitively understood that heat is simultaneously transferred during the course of water movement through sediments and other porous bodies (Bouyoucos, 1915). Examination of temperature patterns provided qualitative and quantitative descriptions of an array of ground-water-flow regimes, ranging from those beneath rice paddies to those beneath volcanoes. Quantitative analysis of heat and water flow was introduced via analytical and numerical solutions to the governing partial differential equations. These quantitative analyses often relied on field measurements for parameter identification and accurate predictions of flow rates and directions. Because field measurements of temperature had to be made manually, however, the data were sparse. Early numerical simulation of heat and mass ground-water transport required significant computational resources, which limited modeling to conceptual demonstrations. As a consequence of these challenges, the use of heat as a tracer of ground-water movement was confined to isolated research projects, which could demonstrate only the feasibility of the method rather than progressing toward a routine use of the technique. Recently, both the measurement of temperature and the simulation of heat and water transport have benefited from significant advances in data acquisition and computer resources. This has afforded the opportunity for routine use of heat as a tracer in a variety of hydrological regimes. The measurement of heat flow is particularly well suited for investigations of stream/ground-water exchanges. Dynamic temperature patterns between a stream and the underlying sediments are typical, because of large stream surface area to volume ratios relative to many other surface-water bodies. Heat is a naturally occurring tracer, free from (real or perceived) issues of contamination associated with the use of chemical tracers in stream environments. The use of heat as a tracer relies on the measurement of temperature gradients, and temperature is an extremely robust property to monitor. Temperature data are immediately available as opposed to most chemical tracers, many of which require laboratory analysis. The recent publication of numerous case studies (for example, Su and others, 2004; Burow and others, 2005) greatly extends the temporal range and the spatial scale over which temperature gradients have been analyzed to use heat as a natural tracer of ground-water movement near streams. This chapter reviews early work that addresses heat as a tracer in hydrological investigations of the near-surface environment, that describes recent advances in the field, and that presents selected new results designed to identify the broad application of heat as a tracer to investigate stream/ground-water exchanges. An overview of field techniques for estimating water fluxes between surface water and ground water is provided here; for a comprehensive discussion with numerous case studies, see Stonestrom and Constantz (2003).

Heat Transfer During Stream/Ground-Water Exchanges

When water is present in a stream channel, heat and water transfer because of vapor movement in the streambed sediments generally is negligible relative to heat and water transfer because of liquid water movement. This eliminates the need to address the complex processes of nonisothermal vapor dynamics in porous material when describing heat and water movement below streams. Within the streambed, heat is transferred into and through sediments as a result of three heat-transfer mechanisms. Radiative heat transfer occurs as solar radiation is adsorbed by the streambed surface. This is the dominant mechanism for a dry streambed, but is usually a small component of heat transfer for the streambed beneath a flowing stream. Heat conduction occurs as diffusive molecular transfer of thermal energy between the streambed surface and the underlying sediments. Heat convection and advection often are used interchangeably in hydrology to indicate heat

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transfer resulting from the movement of water (or air). For the present work, it is advantageous to partition their definitions as follows. Heat convection is defined as heat transfer occurring because of the movement of water (or air) above a streambed of dissimilar temperature. Heat advection is defined as the heat transfer that occurs during the movement of water (or air) through the streambed. This alternative definition is useful for the application of heat as a tracer in examining stream/ground water interaction because it aids in delineating between heat transfers as a result of ground-water movement (advection) in contrast to surface-water movement (convection). Thus, heat conduction, convection, and advection all contribute to heat transfer across the stream/streambed boundary, but determination of heat advection is the focus in examining stream/ground-water interaction.

Commonly, all three heat-transfer mechanisms occur simultaneously within stream environments. For example, all three mechanisms occur in a losing stream reach as water infiltrates into the streambed, then percolates through the sediments, potentially recharging the water table. Convective heat transfer occurs between the stream and sediments as stream water flows over the sediments. As a result of this convection transfer, conductive heat exchange occurs between the surface sediments and sediments at depth. Simultaneously, advective heat transport occurs as water infiltrates into the sediment and percolates in a downward (but usually not vertical) direction. The daily and (or) annual temperature extremes are attenuated and delayed with depth in the streambed sediments. The attenuation of temperature extremes is determined by the bulk volumetric heat capacity of the sediments as heat is rapidly exchanged at the pore scale. The delay or lag in temperature extremes is controlled by the rate of downward heat transfer, which is dependent on the thermal conductivity of the sediments and the pore-water velocity through the sediments. The greater the heat transfer, the greater the depth of penetration and the shorter the time lag of temperature extremes. In the vicinity of streams, heat usually is transported more rapidly by moving water than through molecular diffusion and, as a result, higher streambed infiltration rates result in the deeper penetration and shorter lags in temperature extremes (for example, Lapham, 1989; Silliman and others, 1995). For a neutral stream reach (one neither gaining nor losing flow), the streambed-temperature gradients are created by convective heat transfer from the stream to the streambed surface, and heat transport into the sediment is determined by heat conduction alone. (Thus, if the Fourier equation for conductive heat transfer can explain the temperature patterns within a streambed, there is no stream/ground-water exchange.) For gaining stream reaches, as was the case for losing and neutral streams, a temperature gradient is created at the streambed surface because of convective heat transfer. As ground water discharges to the stream, however, the stream-temperature extremes are attenuated at shallow depths because of the heat capacity of the discharging ground water, such that the greater the ground-water discharge the greater the attenuation of temperature extremes and the greater the lag in temperature extremes in the sediments (for example, Silliman and Booth, 1993).

These heat-transfer processes also have important ramifications on stream-temperature patterns. Constantz (1998) examined streamflow and stream-temperature patterns on the Truckee River, California, and its tributaries to demonstrate the use of stream-temperature analysis to determine spatial and temporal patterns of exchange in selected reaches. For example, results in this work showed that stream-temperature patterns could be used to demonstrate that the main-stem Truckee River received significant water from bank storage in response to upstream dam releases, whereas the tributary directly downstream from the dam possessed inadequate bank storage to influence post-release stream-temperature patterns (see Constantz, 1998, fig. 10).

Quantitative Analyses of Heat as a Ground-Water Tracer Near a Stream

Rorabaugh (1954) examined correlations between stream temperature and seepage patterns and proposed the measurement of temperature to quantify heat flow, and thus determine streambed seepage indirectly. He indicated that a ground-water model capable of quantifying heat and water fluxes appeared to be the appropriate tool. A physically based, quantitative analysis of heat and water transport through porous materials was introduced by Philip and deVries (1957). Their analysis resulted in a comprehensive mathematical description of the coupled process of liquid and vapor water transport simultaneous with the transfer of heat in the solid, liquid, and vapor phases of unsaturated porous material. Application of their analysis has demonstrated that the transport of heat and water in the vapor phase often is important in unsaturated soils, and generally dominates in dry environments (for example, Scanlon and Milly, 1994). As the degree of water saturation increases in sediments, heat transport in the vapor phase abruptly declines as the gas phase becomes discontinuous and then vanishes as sediments approach saturation (for example, Stonestrom and Rubin, 1989). As a result, the comprehensive approach developed by Philip and deVries (1957) is unnecessary for analysis of heat and water fluxes in material that is sufficiently saturated to inhibit macroscopic gas flow. Streambed sediments beneath wetted channels are sufficiently saturated to ignore macroscopic vapor transport.

Suzuki (1960) and Stallman (1963, 1965) were able to use a single-phase approach to predict water fluxes through saturated sediments, based on measured ground-water temperatures. Their work formed the basis for examination of flow in environments ranging from deep ground-water systems (Bredehoeft and Popadopulos, 1965) to humid hillslopes (Cartwright, 1974). Stallman (1963) presented a general equation describing the simultaneous flow of heat and fluid in the earth. He indicated that ground-water temperatures could be used to determine the direction and rate of water movement. He also indicated that temperatures in combination with hydraulic gradients could be used to estimate sediment hydraulic conductivity. Stallman's equation for the simultaneous transfer of heat and water through saturated sediments for the one-dimensional case of vertical flow (z direction) is as follows:

$$K_t \frac{\partial^2 T}{\partial z^2} - q\mathrm{C_w} \frac{\partial T}{\partial z} = \mathrm{C_s} \frac{\partial T}{\partial t}, \qquad (1)$$

where

K_t is the thermal conductivity of the bulk streambed sediments in W/(m °C);

T is temperature in degrees Celsius;

q is the liquid water flux through the sediments in meters per second;

$\mathrm{C_w}$ and $\mathrm{C_s}$ are the volumetric heat capacity of water and the bulk sediment in J/(m^3 °C), respectively;

and

t is time in seconds.

The value of q is controlled by the Darcy equation as the product of the hydraulic conductivity, K, and the total head gradient, h. When q is zero, the equation reduces to the Fourier equation for the transfer of heat by conduction, and when q is large, advection dominates the transfer of heat, as well as the change of temperature throughout the porous material.

Thermal parameters can be estimated, given some knowledge of streambed materials. The heat capacity of the sediments can be estimated by the following:

$$\mathrm{C_s} = f_s(\mathrm{c}_s\, \rho_s) + f_w(\mathrm{c}_w\, \rho_w) + f_a(c_a\, \rho_a), \qquad (2)$$

where f_s, f_w, and f_a are the volumetric fractions of the sediment, water, and air, respectively; c_s, c_w, and c_a are specific heats in J/(kg °C) of the sediment water and air, respectively; and ρ_s, ρ_w, and ρ_a are the densities in kg/m^3 of the sediment, water, and air, respectively. The product of the specific heat capacity and the density is the volumetric heat capacity, which is in the range of 0.8×10^6, 4.2×10^6, and 0.001×10^6 J/(m^3 °C) for sediments, water and air, respectively (de Vries, 1963).

An alternative approach to describe simultaneous heat and water transport through sediments has been to use an energy transport approach via the convective-dispersion equation (Kipp, 1987). These coupled heat and water-flow equations are included here as equations 3, 4, and 5.

$$\partial \frac{\left[\theta C_w + (1-\phi) C_s\right] T}{\partial t} = \nabla \cdot Kt(\theta) \nabla T + \nabla \cdot \theta C_w D_h \nabla T - \nabla \cdot \theta C_w T q + Q C_w T, \qquad (3)$$

where

θ is percent volumetric water content;

ϕ is sediment porosity, dimensionless;

D_h is thermomechanical dispersion tensor, in square meters per second;

q is the water flux, in meters per second,

and

Q is rate of fluid source, in seconds.

The left side of the equation represents the change in energy stored in a volume over time. The first term on the right side describes the energy transport by heat conduction. The second term on the right side accounts for thermomechanical dispersion. The third term on the right side represents advective heat transport, and the final term on the right side represents heat sources or sinks to mass movement into or out of the volume. The familiar water-flow equation is as follows:

$$C(\psi, x) \frac{\partial h(x,t)}{\partial t} = \nabla[k(\psi, x) \cdot \nabla h(x,t)], \qquad (4)$$

where

$C(\Psi, x)$ is specific moisture capacity, which is the slope of the water-retention curve;

Ψ is the water pressure, in meters;

h is the total head, in meters;

x is length, in meters;

and

t is time in seconds (Buckingham, 1907; Richards, 1931).

The thermomechanical dispersion tensor is defined as (Healy, 1990):

$$D_h = \alpha_T |v| \delta_{i,j} + \frac{(\alpha_l - \alpha_t) v_i v_j}{|v|}, \qquad (5)$$

where α_l and α_t are longitudinal and transverse dispersivities, respectively, in m; $\delta_{i,j}$ is the Kronecker delta function; ν_i and ν_j are the ith and jth component of the velocity vector, respectively, in meters per second.

Sediment thermal conductivity, K_t, varies with texture and degree of saturation; for the typical case of saturated sediment in a general textural class, however, the uncertainty is greatly reduced. For example, the streambed K_t for a sand channel is likely to range only from 1.0 to 2.0 W/(m °C), so that the value of K_t can be estimated as 1.5 W/(m °C) ± 0.5 W/(m °C) (van Duin, 1963).

After the thermal parameters are assigned, q is estimated via an appropriate heat and mass-transport simulation model (discussed in detail below). Generally, hydraulic conductivity cannot be estimated using this procedure. As opposed to K_t, hydraulic conductivity can vary over several orders of magnitude. Even for saturated conditions, the hydraulic conductivity of sand-textured material can vary from values of 10^{-2} down to 10^{-6} m/s (Freeze and Cherry, p. 29, 1979). For a given sand-textured material, as saturation decreased, values of hydraulic conductivity measured in the laboratory ranged from 10^{-5} meters per second down to 10^{-10} (for example, Constantz, 1982). Consequently, hydraulic conductivity is not isolated from q without an accurate measurement of the hydraulic gradient. For many studies, the goal is to develop estimates of q, so that temperature measurements applied to equation 1 or equation 2 have proved useful in determining the rate of water movement through a region of interest. In some studies (as discussed below), hydraulic gradients are determined on the basis of piezometer measurements, so that values of hydraulic conductivity also are estimated from sediment-temperature patterns.

Using reasonable boundary conditions and thermal and hydraulic parameters, a heat- and water-transport simulation code is run to predict temperature patterns in stream sediments. For the present application, predicted temperature patterns are matched to measured data using an inverse-modeling approach. Specifically, hydraulic information, such as stream stage, are determined, temperatures are monitored in the stream and streambed, and predicted temperatures then are compared with measured temperatures by using trial-and-error methods or a parameter-estimation code.

Temperature Instrumentation

Background

Measurement of temperature gradients in the sediments is required to estimate the rate of heat transfer through the streambed. Measurement of temperature over time at two or more depths within the stream/ground-water system is the minimum temperature data needed to estimate heat and water fluxes in the domain bounded by the temperature measurements. When it is desirable to separate water fluxes into hydraulic conductivity and the hydraulic-gradient components, measurements of hydraulic gradients are required in addition to temperature gradients. Accuracy in estimating thermal parameters sometimes is improved through laboratory analysis of sediment samples, especially for variables in equation 2 and equation 4; however, the spatial variability of textures in fluvial environments often diminishes the effectiveness of coring efforts, such that an estimate based on the bulk textural class over the domain of interest may be a more prudent approach. Operationally, measurements of temperature in the stream environment involves logistical issues, which generally do not occur in forest or agriculture settings (for example, Jaynes, 1990). In the stream environment, fluvial processes create installation challenges that often have to be overcome on a site-by-site basis. Some streams are wide and shallow with a mantle of boulders, whereas other streams are deep with steep banks. Furthermore, damage to or loss of temperature equipment, because of high streamflows, is an issue unique to streams. Equipment selection and installation methods are usually site specific, though two common requirements are equipment that is sufficiently durable in high flows and an installation procedure that avoids preferential flow of pore water along the length of the equipment embedded in the streambed. Often, the manner in which temperature is measured may differ for ephemeral channels as compared with perennial channels.

Figure 1 provides a qualitative, pictorial description of the thermal and hydraulic responses to the four possible states of a streambed—a gaining stream, a losing stream, a dry ephemeral channel, and an ephemeral channel with water. The purpose of this figure is to provide graphical depictions of conditions relevant to the installation of monitoring equipment. Within each panel of the figure, a hydrograph is depicted on the right, while a pair of thermographs, representing the diurnal pattern in the stream and streambed temperatures, are depicted on the left. For the case of a gaining stream (fig. 1*A*), the hydraulic gradient is upward as indicated by the positive water pressure in the observation well relative to the stream stage. The stream is shown with a large diurnal variation in water temperature, but the sediment temperature has only a slight diurnal variation. The diurnal variation in the sediment is due to the inflow of ground water to the stream, which is generally of constant temperature on the diurnal time scale. Any variation in sediment temperature is a result of a change in the balance between downward conductive transport of heat and upward advective transport of heat. Thus, for a high inflow of ground water, the sediment temperature will have no diurnal variations, whereas for a slight inflow of ground water, the sediment will have a small diurnal variation in temperature (which will be increasingly damped with depth). Consequently, shallow installation of temperature equipment (in the observation well or directly in the streambed) is desired for a gaining stream reach in order to detect significant temperature variations. For the case of a losing stream (fig. 1*B*), the downward hydraulic gradient transports heat from the stream into the sediments. The combined conductive and advective heat transport can result in large diurnal fluctuations in sediment temperature. Furthermore, because ground water is not flowing into the stream, stream-temperature variations generally are larger than those for gaining streams (Constantz, 1998). Consequently, deeper installation of temperature equipment (in the observation well or directly in the streambed) may be in order for losing streams. For a dry streambed (fig. 1*C*), pore-water pressures are negative relative to atmospheric pressure, and, thus, are not measurable in a observation well. The streambed may have extremely high variations in diurnal temperature because of radiative heat transfer; however, the combined effects of low K_t values and no advective heat transport results in negligible diurnal variations in sediment temperatures below the shallowest (for example, 10 centimeters) streambed depths. For ephemeral stream channels (fig. 1*D*), a dynamic temperature pattern exists at the initiation of streamflow. Again, the observation well remains empty because of negative pore-water pressures until mounding of the water table results in water entry into the well. For the ephemeral case, convective, conductive, and advective heat transport all contribute to the rapid responses in the streambed surface and underlying sediments, as seen in the abrupt response of the streambed thermograph.

Direct Versus Indirect Measurements

Water and sediment temperatures can be measured directly by inserting a temperature probe (that is, thermistor wire, thermocouple wire, or platinum resistance thermometer wire) into the medium of interest, or indirectly by inserting the probe to a depth of interest in an observation well. In either case, the selected temperature probe is connected to a data logger. Within

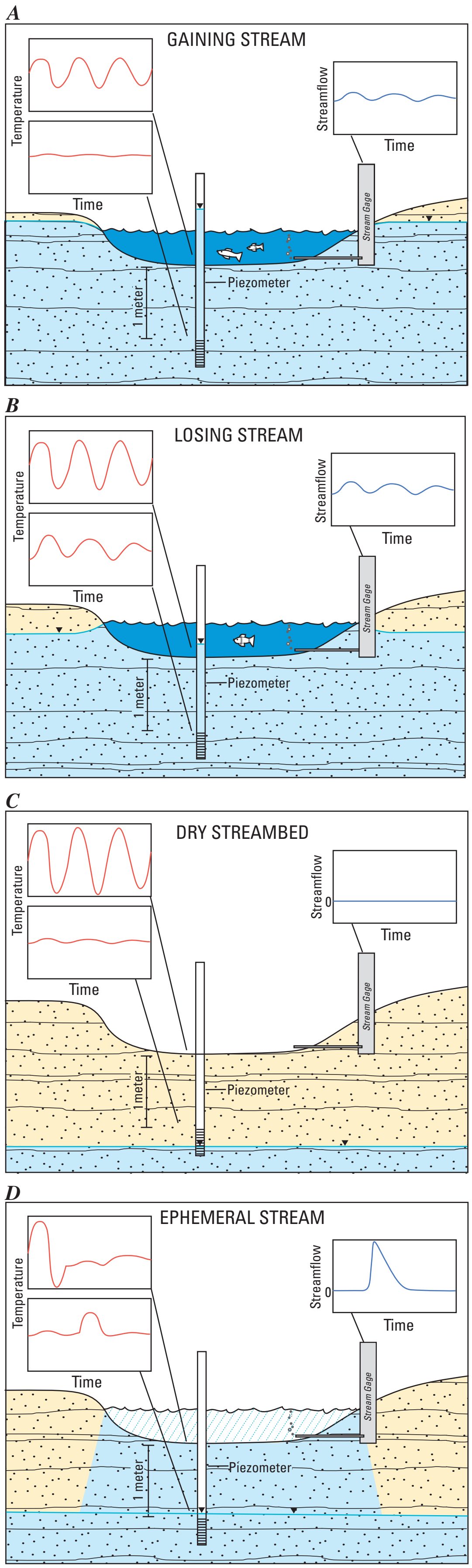


Figure 1. A qualitative description of thermal and hydraulic responses to four possible states of a streambed: *A*, gaining stream, *B*, losing stream, *C*, dry streambed, and *D*, ephemeral stream. Thermographs and hydrographs are displayed in the upper left and right corners, respectively.

observation wells, temperature can be monitored with temperature-logging equipment on a specific schedule such as hourly, daily, or monthly (see Lapham, 1988), or alternatively, temperature can be continuously monitored at fixed locations within the observation well, using either a series of temperature probes at specific depths, or a series of single-channel, submersible microdata loggers tethered at several locations in the observation well (see Bartolino and Niswonger, 1999). There has been some concern about heat conduction down the observation well, and Lapham (1988) suggested that at least the first meter below the streambed may be influenced by the upper boundary. Another concern has been the development of a convection cell in observation wells that would redistribute heat within the well. For a typical shallow observation well with a 0.05-meter-diameter opening, Samuels (1968) calculated that a convection cell 0.5 meter in length could be established at the top of the water column during periods of large upward thermal gradients (for example, during winter). Samuels demonstrated that for large upward thermal gradients, the temperature in the upper water column could be erroneously high by 0.5 °C. These two complications indicate that measurements in the shallow water column within a observation well may less accurately represent the temperature in the surrounding sediments than values obtained deeper in the well-water column. These complications will be addressed in the final section of this chapter.

Observation wells generally are air filled in ephemeral stream channels, resulting in poor estimates of diurnal temperature variations due to the extremely low thermal conductivity of the air cavity. Thus, streambed temperatures need to be taken directly in the sediment rather than in the observation wells for ephemeral channels. Vertical arrays of thermocouples or thermistors (or temperature nests) have been successfully installed directly into flowing ephemeral channels (Thomas and others, 2000). Thermistor or thermocouple wires are inserted down a drill hole, and as the drill stem is withdrawn, saturated sediment collapses into the hole, thus inhibiting preferential flow. Wires then are run horizontally, either bare or in conduit, to a data-acquisition system. For dry ephemeral channels, thermistor or thermocouple wires are installed into the streambed in a similar fashion as for the flowing case; however, backfilling with either native materials or with diatomaceous earth is necessary to inhibit preferential flow next to the wires. Instrumentation also can be installed horizontally from a trench excavated along the bank (see Ronan and others, 1998). Recently, single-channel, submersible microdata loggers have been successfully installed in dry streambeds using a drill rig and backfill (Bailey and others, 2000). For a detailed description of numerous options for using and monitoring of sediment temperature, see Stonestrom and Constantz (2003).

Methods to Analyze Streambed Temperatures

Several researchers have developed simplifying assumptions for specific hydrological conditions that preclude the necessity of using a heat- and ground-water-transport simulation model. A simplistic, first-approximation approach was developed for the case in which pore-water velocities are sufficiently high such that heat transport by conduction is negligible compared with heat transport by advection. This case is typical during flow events in many ephemeral streambeds (see fig. 1*D*), and common in perennial stream channels where a dense clay layer is absent. For those cases in which conduction is small compared to advection, pore-water velocity, v, is approximated by:

$$v = V_T \frac{1}{\theta} \frac{Cs}{Cw} , \quad (6)$$

where

V_T is the vertical velocity of the temperature peak (the "wave" or "front") down into the streambed sediments.

This simplification has been shown to work well in a laboratory column (Taniguchi and Sharma, 1990) and in artificial recharge basin studies (Cartwright, 1974). The flux, q, can be determined from the product of v and θ. The value of θ can be approximated by the porosity of the streambed sediments, although Constantz and others (1988) determined that a value of 0.9θ may be more typical for the initial stages of ponded infiltration. Constantz and Thomas (1996, 1997) have successfully applied this simplification at Tijeras Arroyo, New Mexico, by monitoring temperatures between the surface and a depth of about 3 meters during ephemeral streamflow events. Stewart (2003) examined the error in using this simplistic approach compared to a complete description of conductive and convective heat transport. Stewart reported that the use of equation 6 could overestimate water fluxes by 30 percent for cases in which heat conduction is a significant component of the total heat flux within streambeds.

Silliman and others (Silliman and Booth, 1993; Silliman and others, 1995) used time-series analysis of stream and sediment temperature patterns in Indiana to identify losing reaches. In a similar fashion to Suzuki (1960) working in rice fields, Silliman and others used a one-dimensional solution to equation 1, with an assumed sinusoidal temperature pattern for upper boundary condition. Silliman and Booth (1995) examined the range of the Peclet number (a measure of advective to conductive transport) for which a solution should be applicable (see Silliman and Booth, 1995, p. 106, for the specific values for Peclet parameters that they chose for a streambed environment). They concluded that for Peclet numbers of less than 2×10^{-4}, which represent a flux of 8×10^{-8} meters per second, the advective component of the solution is negligible. Thus, this approach may not be useful for the very low water fluxes typical of streambed environments with extensive clay-textured streambeds and (or) very low hydraulic gradients.

Ronan and others (1998) used the heat- and water-transport simulation code VS2DH (Healy and Ronan, 1996) to model the ground-water-flow pattern below Vicee Canyon, Nevada. The ephemeral stream channel within Vicee Canyon meanders over an alluvial fan on the east side of the Carson Range of western Nevada. Along the fan, temperature was monitored in the stream channel and streambed using a 3-meter by 3-meter grid of 24 thermocouples at three locations. The two-dimensional simulation code was used in an inverse modeling approach to match simulated temperature against measured temperature to estimate heat and water fluxes into or out of the streambed vertically and horizontally. After calibration of the model during one season, simulation results were able to predict streamflow loss and streambed infiltration based only on temperature data. Their results used values for dispersivity of about 0.01 meter, indicating that thermal dispersion does not appear to be significant in this type of environment for this length scale (3 meters). Incorporating two-dimensional temperature patterns as input into the model was useful in demonstrating the asymmetrical pattern of substream ground-water flow as a result of down-canyon ground-water flow as the stream meandered across the fan.

The use of heat as a tracer to examine stream/ground-water exchanges has not been limited to shallow investigations. Deeper monitoring of substream temperatures has been done by Lapham (1989) and Bartolino and Niswonger (1999) to estimate annual patterns of stream/ground-water exchanges, where temperatures were periodically logged in observation wells as deep as 50 meters below the streambed. Long-term temperature monitoring provides a series of temperature profiles that can be useful in characterizing streambed fluxes. Figure 2 shows hypothetical streambed-temperature profiles for a losing stream (downward water flux) compared with a gaining stream (upward water flux) over either a year or a day. The temperature profiles for the annual (or daily) extremes

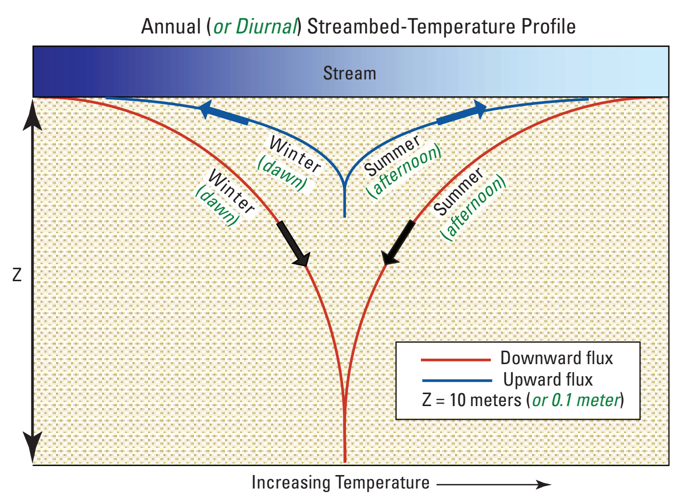
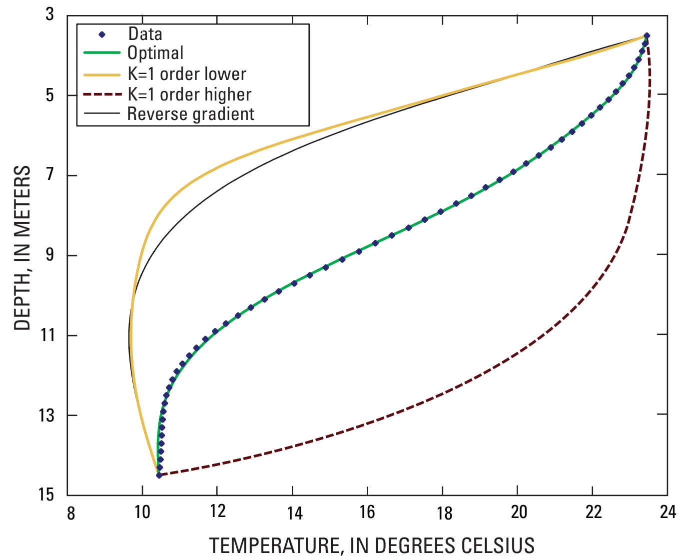


Figure 2. The streambed temperature profiles (temperature envelopes) for a losing stream (downward water flux) compared to a gaining stream (upward water flux) for the annual example and the diurnal example.

Figure 3. The simulated temperature profile below the Rio Grande in central New Mexico for an optimized value of *K* (hydraulic conductivity), compared to a value of *K* one order of magnitude greater or less than the optimal value, and the temperature profile with an optimal value of *K*, but with an upward value of *H* (heat) to simulate a gaining stream, based on measured results in Bartolino and Niswonger (1999).



forms a "temperature envelope" for a particular site, within which all other temperature profiles reside. On an annual scale, the January and July temperature profiles typically form an envelope in which other monthly temperature profiles reside. When ground water is discharging to the stream, the annual envelope is collapsed toward the streambed surface, and when the stream is rapidly losing water to the sediments, the envelope extends to great depths. This is true on a daily time scale as well, with the dawn and afternoon temperature profiles forming the daily envelope, in which all other hourly temperature profiles reside. A salient difference between the annual and daily temperature envelopes is the depth scale. See Lapham (1989) for a series of annual and daily example temperature envelopes from streams in the eastern United States.

Figure 3 depicts the effect of changing values of hydraulic conductivity on temperature profiles, based on results for the Rio Grande at Albuquerque, New Mexico (Bartolino and Niswonger, 1999). The figure compares the optimal fit value for hydraulic conductivity (6.7×10^{-6} meters per second) with an order of magnitude increase in hydraulic conductivity, an order of magnitude decrease in hydraulic conductivity, and the optimal hydraulic conductivity with a reversed hydraulic gradient. The large sensitivity of streambed temperature to different hydraulic conditions is clearly apparent. To a lesser extent, simulated flux estimates also are sensitive to uncertainty in thermal parameters. Niswonger and Rupp (2000) used Monte Carlo analysis to examine the relative importance of errors in estimating temperature, K_t, and C_s to the resulting simulated water fluxes for Trout Creek, Nevada. When isolating thermal properties, they determined that for the expected mean and standard deviation in thermal parameters, resulting VS2DH simulated water fluxes were most sensitive to uncertainties in sediment temperature and least sensitive to uncertainties in K_t. In general, predicted fluxes were highly sensitive to variations in hydraulic properties and slightly sensitive to variations in thermal properties for the range of properties reported for sediments.

Example Sites

Study results from two sites are summarized below to provide example applications of the use of heat as a natural tracer of ground-water movement near streams. These sites were chosen because: (1) they are characterized by distinctly different seasonal streamflow patterns; and (2) at the first site, a direct temperature monitoring technique was used, whereas at the second site, an indirect monitoring technique was used. Both sites were losing stream reaches during the study period; however, the techniques described work equally as well on gaining stream reaches. For an example of direct and indirect temperature measurements used in gaining reaches to estimate upward water fluxes, see Silliman and Booth (1993) for direct temperature measurements made in sediments, and Lapham (1988) for indirect temperature measurements made in observation wells.

Bear Canyon, New Mexico

Bear Canyon is on the eastern edge of Albuquerque, New Mexico. The small ephemeral stream within the canyon is a representative example of more than 100 similar streams that drain from the western flanks of the Sandia and Manzano Mountains into the Middle Rio Grande Basin. The flows in these ephemeral streams have the common characteristics of being bedrock-controlled in their upper gaining reaches, and alluvium-controlled in their lower losing reaches. The streamflow and stream-loss patterns of these stream channels are poorly documented, but their cumulative streambed infiltration might contribute significantly to potential recharge to the basin.

The use of streambed-temperature data was included in a suite of monitoring methods and field-reconnaissance procedures intended to estimate streamflow loss and potential recharge along a reach of Bear Canyon. This reach extends from the exposed bedrock at the mountain-front downslope in a westward direction for about 3 kilometers, at which point the stream channel has been modified as a result of urbanization. The stream is perennial east of the bedrock exposure at the mountain front, and flows rarely extend more than 1 kilometer from the mountain front, though summer monsoons occasionally induce streamflow to the confluence with the Rio Grande, approximately 20 kilometers to the west of the mountain front.

Two temperature-monitoring methods were used within the stream channel of Bear Canyon from 1996 through 1999. Streambed surface temperatures were monitored at sites between the mountain front and the modified reach of the channel, 3 kilometers to the west of the mountain front. Surface-temperature patterns were analyzed as part of the characterization of the spatial and temporal pattern of streamflow in Bear Canyon. Procedures and results for the surface-temperature measurements are described in detail in Constantz and others (2001). Vertical temperature patterns were monitored by using a series of thermocouple wires installed at depths between the streambed surface and about 3 meters below the channel to create a temperature nest in a fashion similar to that described in Thomas and others (2000). Temperature nests were installed at two locations in the middle reach of the Bear Canyon study site, where ephemeral streamflows were expected to be present for extended periods. After backfilling installation holes, the completed temperature nests monitored temperature at 0.40, 0.60, 1.10, 2.10, and 3.10 meters below the streambed surface. Temperatures were monitored at 15-minute intervals via a data logger in an enclosure near the stream channel until September 1999.

Seasonal snowmelt resulted in a gradual progression of the downstream limit of flow down-channel over several months in the spring, followed by a retreat up-channel in early summer. Flashy, summer monsoon streamflow occurred in some, but not all years. Details of the late stages of spring streamflow at one temperature nest in the channel are shown in figure 4. As expected, the greatest diurnal temperature variations are those at a depth of 0.40 meter, and the smallest diurnal temperature variations are those at a depth of 3.10 meters. The abrupt retreat of streamflow upstream in Bear Canyon also is clearly detectable. As streamflow retreated up-channel from this site on June 5, 1999, the abrupt transition from advection-dominated heat transport to conduction-dominated heat transport is quite distinct. Reduced magnitudes in diurnal variations in streambed temperature result from the loss of advective heat transport with the cessation of streamflow into the streambed.

The streambed-temperature profiles generated at temperature nests during annual spring streamflow were used as input in a fitting procedure that compared measured temperatures to simulated temperature using VS2DH in order to estimate streambed infiltration rates. A commercially available optimization program was used to determine the streambed-sediment hydraulic conductivity from the best fit between the simulated streambed temperatures and measured sediment temperatures. Figure 5 shows sediment temperatures during June 1997 at four depths below the streambed at a vertical temperature site approximately 275 meters west of the mountain front. The figure also shows the simulated best fit at 0.60 meter using an optimization program. The measured streambed temperatures were applied as the upper thermal boundary condition, and measured hydraulic gradients and stream stage were used for hydraulic-boundary conditions. Saturated conditions existed below the stream channel as determined by measuring water levels with piezometers set in the streambed. An optimized seepage rate of 0.75 meter per day resulted in the fit for a depth of 0.6 meter as shown in figure 5 (fits for 0.4 and 1.1 meter depth were comparable but not shown in the figure for clarity of individual thermographs). Optimized simulations for the duration of spring streamflow for this site indicated an average vertical streambed seepage rate of 0.77 meter per day. The consistency in the estimated seepage rate over the duration of the spring season indicates that neither the hydraulic conditions nor the streambed sediments in Bear Canyon were transient. This magnitude of streambed infiltration persisted until retreat of streamflow up-canyon, at which time water fluxes rapidly declined during drainage of the streambed.

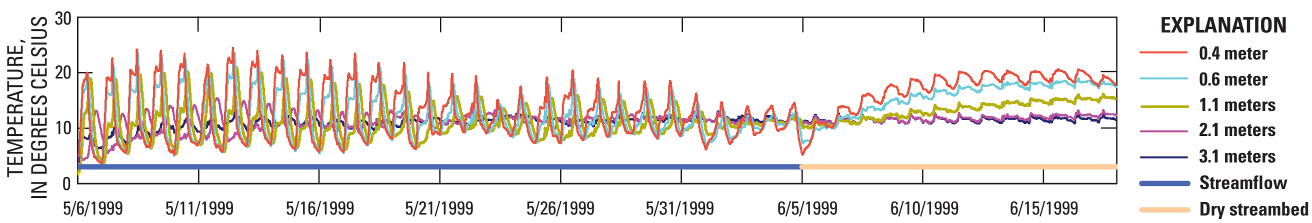


Figure 4. The streambed-temperature patterns resulting from late stages of continuous spring streamflows progressing to abrupt cessation of streamflow, approximately 275 meters west of the mountain front in Bear Canyon, New Mexico, during June 1999.

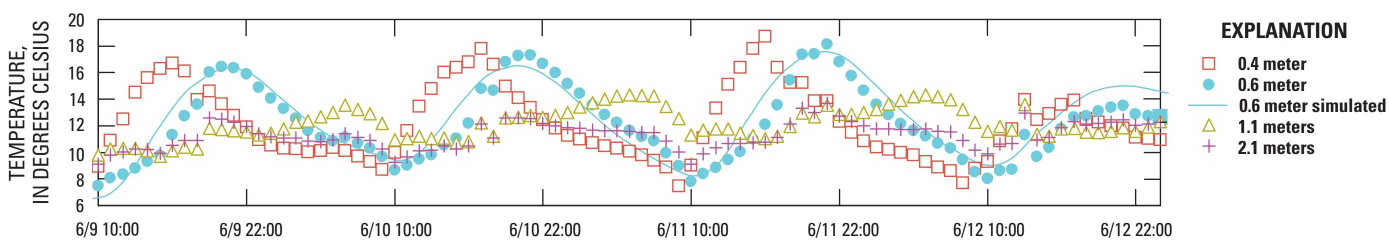


Figure 5. The streambed-temperature patterns compared to optimized (simulated) sediment temperatures at a depth of 0.6 meter, approximately 275 meters west of the mountain front in Bear Canyon, New Mexico, during June 1997.

Santa Clara River, California

The Santa Clara River is in southern California, flowing from the San Gabriel Mountains approximately 200 kilometers to the Pacific Ocean. In the upper reaches, the gradient is steep, and the stream generally flows over bedrock with a steady gain of ground water. In the middle reaches, the stream flows over a wide sandy channel, resulting in large diurnal stream-temperature fluctuations, as well as substantial potential for stream/ground-water interaction. A 17-kilometer study section was defined in the middle reaches of the river, and a variety of hydrological properties were monitored using a range of surface- and ground-water instrumentation. As part of this larger study, an observation well was installed in the deepest section of a losing reach (referred to as SCR5) in October 1999. The observation well was approximately 4 meters in length with a 0.08-meter internal diameter. The observation well was driven approximately 2.5 meters into the streambed, at which time the drive point was driven from the bottom of the observation well. Temperature between the streambed and the bottom of each observation well was monitored by tethering one single-channel, submersible temperature microdata logger outside the observation well to monitor stream temperature and by tethering three microdata loggers inside the observation well at about 0.6, 1.2, and 2.4 meters below the streambed. The VS2DH simulation code was used to compare one-dimensional simulated temperatures with measured temperatures with a best-fit trial and error match, in order to estimate streambed-percolation rates. Temperature at depth for SCR5 during October 1999 varied during periods in which the stream did not flow and when the stream did flow at this observation-well site. Simulation results matched measured data very well during streamflow, but resulted in a poor match during the intermediate no-flow period at all depths monitored (fig. 6). The poor match during this period is expected because of water drainage from inside the observation well. Thus, microdata loggers suspended in the air-filled interior of the observation well were thermally isolated from the adjacent sediment during the no-flow period. Once streamflow returned to this location in the stream, the microdata loggers again were submerged and able to effectively monitor sediment temperatures. Consequently, the simulated results probably more correctly matched the sediment temperature during the no-flow period than did the microdata loggers. Direct burial of temperature equipment in the streambed sediments would have avoided the difficulty in monitoring temperature during this period without streamflow. The best-fit simulated temperatures shown in figure 6 resulted in a streambed infiltration rate of 1.8 meters per day, and based on the measured hydraulic gradient in the observation well of 0.41 meter per meter, the derived hydraulic conductivity was 5.1×10^{-5} meters per second.

Temperatures logged beneath the streambed at 0.3 meter varied in comparison to temperature logged inside the piezometer at the same depth as a function of presence or absence of flowing water in the stream (fig. 7). The agreement between the streambed and piezometer temperatures is excellent when streamflow is present and, as expected, agreement is poor during the no-flow period because of drainage of the piezometer and resultant thermal isolation of the microdata logger. The period of excellent agreement between the measured temperatures probably is a result of the strong advective transport of heat at SCR5, such that conduction of heat down the piezometer is small relative to the total transport of heat. Further research is needed to determine the depth of influence of piezometer heat conduction for stream sites where heat conduction is the dominant heat-transport process with the streambed. Realistically, in environments where conduction is the dominant mechanism of heat transport, piezometer design needs to incorporate features that inhibit heat transport vertically along the piezometer while still allowing heat transport horizontally from the sediments to the piezometer. Though the results depicted in figure 7 show good agreement between temperatures, the flux range for good agreement warrants further examination.

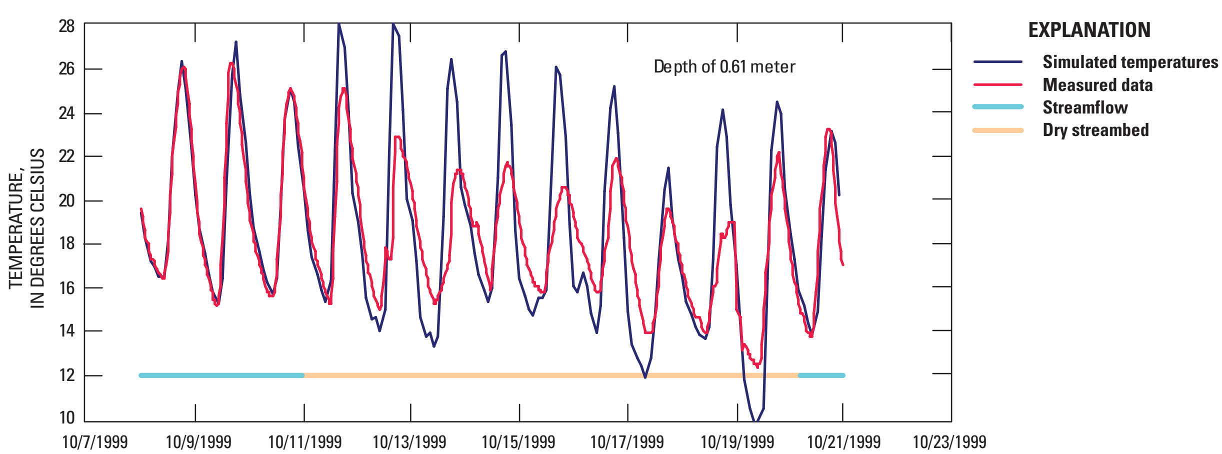


Figure 6. The measured sediment temperatures compared with simulated streambed temperatures at a depth of 0.6 meter below the streambed surface at observation site SCR5 on the Santa Clara River, California, during October 1999. (Note that streamflow ceased and resumed at the site in the middle of the period of record, as indicated by the horizontal lines at the bottom of the graph.)

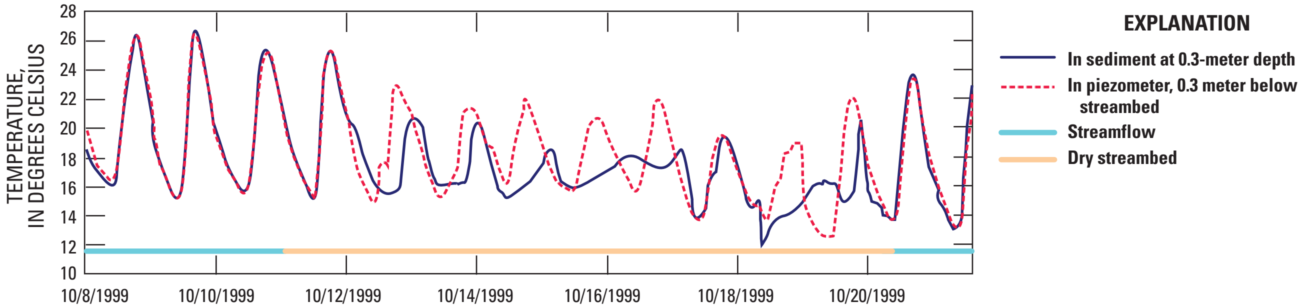


Figure 7. A comparison of streambed temperatures measured directly in the sediments and temperatures measured inside the observation well for observation site SCR5 at a depth of 0.3 meter below the streambed surface on the Santa Clara River, California, during October 1999. (Note that streamflow ceased and resumed at the site in the middle of the period of record.)

Summary

In summary, the measurement and analysis of temperature gradients in streambed sediments provide qualitative patterns and quantitative estimates of rates and direction of water movement through sediments. Both the temporal range, and spatial scale, over which temperature gradients have been analyzed to use heat as a natural tracer of ground-water movement near streams has been greatly extended by numerous recent case studies. Currently, research is ongoing in the areas such as thermal and hydraulic parameter optimization and time-series analysis of temperature gradients to expand the use of heat as a natural tracer in more complex, highly heterogeneous environments.

Recent improvements in acquisition of sediment temperatures and in simulation modeling of heat- and ground-water transport are leading to widespread implementation of methods in which heat is used to trace ground-water fluxes near streams. This chapter provides a brief historical review of the use of heat as a tracer of shallow ground-water movement, and details current theory used to estimate stream/ground-water exchanges. Techniques for installation and monitoring of temperature and stage equipment are discussed in detail for a range of hydrological environments. These techniques are divided into either direct temperature measurements in streams and sediments or indirect measurements in observation wells. Methods of analysis of acquired temperature measurements include analytical solution, heat- and water-transport simulation models, and simple heat-pulse arrival-time procedures. Temperature and derived-flux results are presented for field sites in Bear Canyon, New Mexico, and the Santa Clara River, California. Direct monitoring of temperatures in the sediments below Bear Canyon resulted in estimates of streambed infiltration of 0.75 meter per day, whereas indirect monitoring of sediment temperature using observation wells installed in the Santa Clara River, resulted in streambed-infiltration rates of 1.8 meters per day. The accuracy of measurements within piezometers was confirmed by comparing sediment temperatures acquired directly in sediments with temperatures acquired in a piezometer at the same depth.

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