Review Paper/

Heat as a Ground Water Tracer

by Mary P. Anderson[1]

Abstract

Heat carried by ground water serves as a tracer to identify surface water infiltration, flow through fractures, and flow patterns in ground water basins. Temperature measurements can be analyzed for recharge and discharge rates, the effects of surface warming, interchange with surface water, hydraulic conductivity of streambed sediments, and basin-scale permeability. Temperature data are also used in formal solutions of the inverse problem to estimate ground water flow and hydraulic conductivity. The fundamentals of using heat as a ground water tracer were published in the 1960s, but recent work has significantly expanded the application to a variety of hydrogeological settings. In recent work, temperature is used to delineate flows in the hyporheic zone, estimate submarine ground water discharge and depth to the salt-water interface, and in parameter estimation with coupled ground water and heat-flow models. While short reviews of selected work on heat as a ground water tracer can be found in a number of research papers, there is no critical synthesis of the larger body of work found in the hydrogeological literature. The purpose of this review paper is to fill that void and to show that ground water temperature data and associated analytical tools are currently underused and have not yet realized their full potential.

Introduction

Heat-flow theory has been important in the development of the science of ground water hydrology; yet, the body of literature on heat as a ground water tracer is relatively small. After a flurry of publications in the 1960s, interest waned, although it never died out. A revival began in the late 1980s, with the publication of a methodology for using temperature profiles beneath streams to quantify ground water/stream interaction (Lapham 1989) and a collection of papers on the effects of ground water flow on the thermal regime in basins (Beck et al. 1989). Recent work has considerably expanded the body of literature on heat as a tracer and temperature measurements as surrogates for head measurements, especially in the analysis of temperature profiles to estimate interchange with streams (Stonestrom and Constantz 2003).

Ground water temperature can be measured easily and rapidly by lowering a thermometer down a borehole, although precautions must be taken to ensure that the recorded temperature is representative of water in the aquifer and not influenced by movement of water in the borehole (Mansure and Reiter 1979; Keys and MacCary 1971, p. 106). In recent applications, thermocouples and thermistors are used to obtain a time series of measurements remotely. Waterproof temperature loggers are now inexpensive and widely available, which may help explain the recent revival of interest in subsurface temperatures (Stonestrom and Blasch 2003). Additionally, some investigators have used airborne thermal sensors to detect areas of ground water discharge (Becker 2006).

The purpose of this review paper is to show how pioneering work published in the 1960s on heat as a ground water tracer has been extended and applied to a wide variety of hydrogeological settings. All the research considered herein is concerned with ways of analyzing the movement of heat within ground water systems, including heat produced naturally in the subsurface and heat introduced in seepage from surface water. These studies use temperature, which often can be measured more easily than head, to gain insight into ground water flow. That is, temperature is used as a surrogate for head and to supplement head measurements.

The purpose of this paper is to present a critical synthesis of the body of work on heat as a ground water tracer. After discussion of historical background, the basic theory of heat transport as relevant to ground water is considered, followed by discussion of temperature profiles, heat

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transport at the basin scale, and the use of temperature measurements in formal solutions of the inverse problem. In addition to the literature on temperature measurements in the shallow subsurface, which is the focus of this paper, there are related publications in geothermics or thermal geophysics (e.g., Jessop 1990; Clauser 1999), mining engineering (e.g., Williams et al. 1999), and energy resources including geothermal resources (e.g., Garg and Kassoy 1981), heat pumps (e.g., Andrews 1978; Rafferty 2003; Bose 2003), and subsurface storage of hot water (e.g., Molson et al. 1992). Finally, there is literature on the role of heat in geological processes such as faulting and hydrothermal flow (e.g., Williams and Narasimhan 1989; Wieck et al. 1995), hydrothermal circulation in the oceanic crust (e.g., Yang et al. 1996; Screaton and Ge 1997), and the formation of ore deposits (e.g., Bethke 1986; Deming and Nunn 1991; Garven et al. 1999). Ground water flow and heat transport models are also important in assessing geologic repositories for high-level nuclear waste (e.g., Birkholzer et al. 2004).

Background

Historical Importance

It is well known that Darcy's law has the same form as Fourier's law for heat flow. Soon after the publication of Fourier's (1822) classic monograph, the scientific community recognized that Fourier's work had broader application as a framework for conceptualizing phenomena other than heat flow (Narasimhan 1999). It is likely that Darcy was aware of Fourier's work soon after it was published (Groenevelt 2003) and well before he conducted his famous laboratory experiments in 1856 (Darcy 1856). About the same time as the work of Fourier and Darcy, Humboldt and Arago (see Davis 1999) measured seasonal fluctuations in shallow ground water temperature and speculated that deep circulation of ground water could account for hot springs. Later, Forchheimer (1886) and Slichter (1899) recognized the analogy between ground water flow and heat flow and applied potential theory in

the form of the Laplace equation to ground water problems. Theis (1935), with the help of Lubin (Freeze 1985), turned to an analytical solution of the heat-flow equation to develop his famous model for transient flow to a pumping well, noting that:

> Darcy's law is analogous to the law of the flow of heat by conduction, hydraulic pressure being analogous to temperature, pressure-gradient to thermal gradient, permeability to thermal conductivity, and specific yield to specific heat. Therefore, the mathematical theory of heat-conduction developed by Fourier and subsequent writers is largely applicable to hydraulic theory.

Researchers still turn to analytical solutions in the heat-flow literature (e.g., Carslaw and Jaeger 1959) to find solutions to ground water flow problems (e.g., Townley and Trefry 2000).

Tracing Surface Water Infiltration

Although there was early recognition of ground water temperature as a water quality parameter (Collins 1925), one-time measurements were (and still are) considered adequate for the purpose of assessing quality inasmuch as ground water temperature is relatively stable. Slichter (1905) was among the first of the early workers to recognize the utility of heat as a ground water tracer, noting that relatively high ground water temperatures reflected induced infiltration from a pond in Long Island, New York. Later workers (Rorabaugh 1956; Schneider 1962a, 1962b, 1972; Winslow 1962; Salem et al. 2004) also used heat to trace induced infiltration from streams, as well as thermal pollution from a cooling pond (Andrews and Anderson 1979) and seeps of ground water inflow to lakes (Lee 1985). For example, in Figure 1, a plume of relatively warm water is clearly delineated by isotherms as water is induced to flow out of the Mohawk River near Schenectady, New York, in response to pumping from nearby well fields.

In recent work, streambed temperatures are used to identify losing and gaining reaches in both humid (Silliman and Booth 1993; Conant 2004) and arid regions
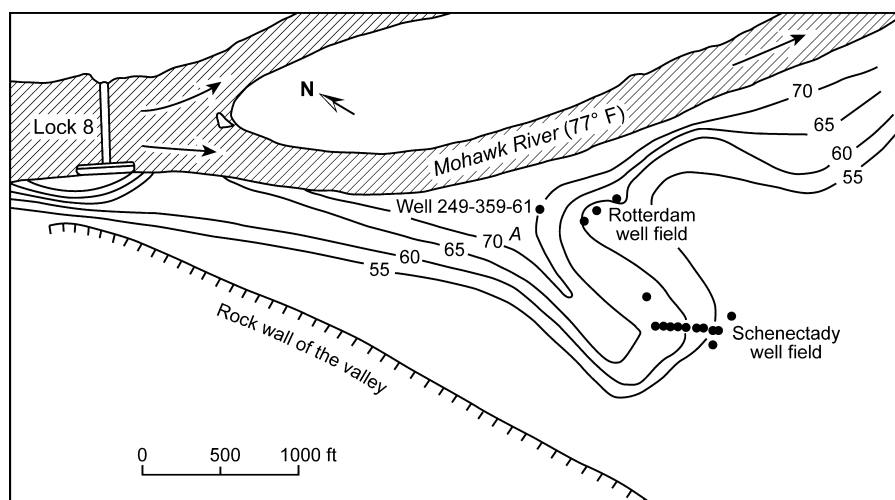
Figure 1. Thermal plume in an aquifer adjacent to a stream; relatively warm surface water is induced to enter the aquifer by pumping (Winslow 1962).

(Constantz et al. 2003a, 2003b). Allander (2003) identified seasonal shifts from losing to gaining conditions in a tributary to Lake Tahoe, California, and Constantz et al. (2001, 2002) used streambed temperatures to characterize stream/aquifer exchange in ephemeral streams in the Western United States. Streambed temperatures were also used to delineate flows in the hyporheic zone (White et al. 1987; Hendricks and White 1991; Evans et al. 1995; Maddock et al. 1995; Dumouchelle 2001; Alexander and Caissie 2003; Storey et al. 2003; Conant 2004).

The work by Conant (2004) illustrates the type of spatial and temporal detail possible when deploying temperature sensors. Subsurface temperatures in the streambed were measured in a 60-m reach of a stream in Ontario, Canada, at a spacing of 1 m along transects spaced 2 m apart. Stream temperatures were measured every 15 min, and subsurface temperature was measured at 1-h intervals by temperature loggers; the difference between surface water temperature and streambed temperatures ranged from 2°C to 5°C in July when temperature was measured at around 380 points. Five different types of flow within the streambed were identified (short circuit discharge/springs, high discharge, low-moderate discharge, no discharge, and recharge), and a conceptual model of the circulation of water within the streambed was developed giving a first detailed look at the intricacies of flow in the hyporheic zone.

Effects of Temperature on Hydraulic Conductivity

Early workers noted that thermal effects cause changes in hydraulic conductivity, K ($= kg\rho_w/\mu_w$) since density, ρ_w, and viscosity, μ_w, of water are temperature dependent; k is intrinsic permeability and g is acceleration due to gravity. For example, the cone of depression caused by pumping near a river is larger during winter months when viscosity of water is relatively large and hydraulic conductivity is relatively small (Winslow 1962). Consequently, rates of induced infiltration may be comparable in winter and summer even though the gradient between the river and the aquifer is higher in winter (Rorabaugh 1956). Jaynes (1990) observed marked diurnal fluctuation in infiltration from a pond, with maximum infiltration occurring during the day and minimum infiltration at night. Similarly, increases in stream temperature in the afternoon may raise the temperature of the streambed causing an increase in streambed hydraulic conductivity, which induces higher fluxes of water from the stream to the subsurface, resulting in an afternoon decrease in streamflow (Constantz et al. 1994; Constantz and Thomas 1997; Constantz 1998).

Analytical Tools for Solving the Inverse Problem with Temperature

Stallman (1963) was among the first of the early workers to appreciate fully the potential for using temperature measurements to solve the inverse problem for ground water velocity and hydraulic conductivity: "… it appears feasible that temperature measurements can be used for calculating flow velocity and that a combination of head and temperature measurements can be used for calculating aquifer permeability." Pioneering theoretical work in the 1960s (Suzuki 1960; Stallman 1965; Bredehoeft and Papadopulos 1965) showed that temperature measurements could be used in analytical solutions of the one-dimensional form of the heat transport equation to solve for ground water velocity. The extension of this type of analysis to estimate ground water recharge/discharge rates, exchange with surface water, flow through fractures, and effects of surface warming is considered in "Temperature Profiles."

The use of temperature measurements at the basin scale was pioneered by Cartwright (1970) who used the Bredehoeft and Papadopulos (1965) model to calculate ground water discharge in the Illinois Basin. Domenico and Palciauskas (1973) presented a two-dimensional analytical solution of heat transport in a cross section of a ground water basin. Subsequent work using analytical solutions applied to sedimentary basins was pursued mainly by investigators in Japan (Sakura 1993; Inagaki and Taniguchi 1994; Taniguchi 1993, 1994; Dapaah-Siakwan and Kayane 1995). With the advent of numerical models, many investigators turned to numerical solutions of coupled ground water and heat-flow models in two and three dimensions. In 1989, a state-of-the-art monograph on heat transport in ground water basins was published by the American Geophysical Union (Beck et al. 1989). The body of work on basin-scale studies, including studies of thermal and cold springs, are reviewed in "Basin-Scale Studies."

Finally, Stallman's (1963) idea that head and temperature measurements could be used jointly in numerical models to solve the inverse problem for ground water flow and hydraulic conductivity has been realized in recent work reviewed in "Formal Solutions of the Inverse Problem."

Basic Theory

The basic theory of heat flow as applied to ground water problems is reviewed by Deming (2002, chap. 11), Domenico and Schwartz (1998, chap. 9), Ingebritsen and Sanford (1998, chap. 3 and sec. 4.3), deMarsily (1986, sec. 10.3), and Bear (1972, sec. 10.7). Following Domenico and Schwartz (1998), the three-dimensional heat transport equation can be written as:

$$\frac{\kappa_e}{\rho c}\nabla^2 T - \frac{\rho_w c_w}{\rho c}\nabla \cdot (Tq) = \frac{\partial T}{\partial t} \qquad (1)$$

where T is temperature; t is time; ρ_w and c_w are, respectively, density and specific heat of the fluid and ρ and c are, respectively, density and specific heat of the rock-fluid matrix; q is the seepage velocity or specific discharge vector; and κ_e is a term that includes the effective thermal conductivity of the rock-fluid matrix. Equation 1 assumes that ρ_w, c_w, ρ, c, and κ_e are constants. When written with T as the dependent variable, Equation 1 is said to be in the temperature-based formulation; the heat transport problem can also be formulated with enthalpy as the dependent variable (Ingebritsen and Sanford 1998). Rigorous derivations of more general forms of Equation 1 are given by Ingebritsen and Sanford (1998) and Bear (1972).

The analogy between Equation 1 and the advection-dispersion equation, which describes transport of solutes in water, is obvious. The first term in Equation 1 is analogous to the diffusion/dispersion term in the advection-dispersion equation, while the second term, which represents transport of heat by moving ground water, is analogous to the advection term.

The first term in Equation 1 represents transport of heat by conduction (analogous to transport of solutes by molecular diffusion) as well as transport by thermal dispersion (analogous to hydrodynamic dispersion in solute transport theory). The term κ_e includes effects of conduction through the rock-fluid matrix as well as effects of thermal dispersion, which is caused by velocity variation within the pore space. We can show this by writing:

$$\eta = \frac{\kappa_e}{\rho c} = \frac{\kappa_o}{\rho c} + \alpha^* |q| \qquad (2)$$

where $\kappa_o = n\kappa_w + (1-n)\kappa_g$ is the effective thermal conductivity, n is porosity (or moisture content in unsaturated material), κ_w is the thermal conductivity of the fluid, and κ_g is the thermal conductivity of the grains/rock, from which it is obvious that κ_o is strongly dependent on moisture content (figure 3 in Stonestrom and Blasch 2003). (A similar relation can be written for the heat capacity, ρc, where $\rho c = n(\rho_w c_w) + (1-n)\rho_g c_g$.) Thermal conductivity, κ_o or κ_g, is typically reported in units of watts (J/s) per degree celsius (or kelvin) per meter (W/m°C). Values of thermal conductivity of porous materials range from around 0.2 W/m°C for dry sediments to 4.5 W/m°C for sandstone (Table 1). (The thermal conductivity of water is 0.60 W/m°C.) Thus, the range in thermal conductivity is much smaller than the range in hydraulic conductivity.

The term, $\kappa_o/\rho c$, in Equation 2 is analogous to the molecular diffusion coefficient in the advection-dispersion equation for solute transport, while α^* is analogous to solute dispersivity and $|q|$ is the specific discharge vector. In the absence of ground water movement, there is no thermal dispersion and the second term on the right hand side of Equation 2 is zero. In this case, κ_e equals the effective thermal conductivity, κ_o. For this case, η becomes the

thermal diffusivity and is of the order 10^{-2} to 10^{-3} cm²/s (Domenico and Schwartz 1998, p. 196), whereas molecular diffusion is of the order 10^{-6} cm²/s. The larger values for conduction of heat arise partly because heat is transferred through the solid as well as the fluid.

There are conflicting points of view regarding the magnitude of the thermal dispersivity, α^*. Tracer tests of both solute and heat at Bonnaud, Jura, France, gave α^* of the same order of magnitude as solute dispersivity (deMarsily 1986, p. 279–280), leading some workers to set α^* equal to solute dispersivity. For example, in table 2 of Smith and Chapman (1983) regional values of longitudinal and transverse α^* are given, respectively, as 100 and 10 m; in table 1 of Niswonger and Prudic 2003, longitudinal α^* for ground water exchange with streams is given as 0.01 to 1 m. However, Bear (1972), Ingebritsen and Sanford (1998), and Hopmans et al. (2002), among others, concluded that the effects of thermal dispersion are negligible compared to conduction and set α^* to zero.

The second term in Equation 1 represents the transport of heat by flowing ground water, a process known as advection or convection. The terms advection and convection are used interchangeably in the literature and may refer to transport of heat and/or solutes. In this paper, advection is used for transport of solutes, while convection is used for transport of heat. Free convection refers to heat transfer in response to flow driven by temperature-induced density differences, while forced convection refers to heat transfer by flow driven by any other mechanism. Most commonly, forced convection occurs in response to topographically driven ground water flow. Free convection may be enhanced by salinity gradients (thermohaline convection). Mixed convection refers to flows driven by both forced and free convection (Raffensperger and Vlassopoulos 1999).

Free convection is thought to occur in areas of high heat flow such as near spreading centers in the ocean but rarely in sedimentary basins where it requires high basal heat flow or relatively thick and permeable layers. The potential for free convection is often investigated using the dimensionless Rayleigh number, which is derived by considering the ratio of buoyant forces to viscous forces. Theoretically, the onset of free convection in an infinitely extensive horizontal layer occurs when the Rayleigh number exceeds $4\pi^2$ (= 39.48) (Lapwood 1948). However, for other configurations, critical numbers ranging from 0 to 27 have been postulated (Blanchard and Sharp 1985; Raffensperger and Vlassopoulos 1999). A number of theoretical studies have investigated the phenomenon (e.g., see Domenico and Schwartz, p. 207–209), but direct field evidence is lacking. Free convection is sometimes implicated by the large mass transfer of solute required in certain types of diagenesis. Blanchard and Sharp (1985) presented graphs (reproduced in Sharp et al. 1988) that show the critical permeability necessary for the onset of free convection for selected layer thickness and geothermal gradients. Following Aziz et al. (1973), they also presented evidence in the form of areal patterns of ground water temperature to support the existence of convection cells in thick sandstone units in the Gulf Coast basin under normal geothermal gradients.

Table 1
Representative Values of Thermal Conductivity of Porous Materials

	Thermal Conductivity (W/m°C)
Dry sediments (e.g., sand, loam, clay)	0.18–0.26[1]
Saturated sediments (e.g., sand, loam, clay)	1.4–2.2[1]
Shale	1–2
Sandstone	2–4.5
Limestone	2–4
Granite	3–4
Basalt	1.5–2.5

[1]Stonestrom and Blasch (2003); values are κ_o for saturated sediments and κ_g for dry sediments; all other values are from Deming (2002) and represent κ_g.

Forced convection, typically the transfer of heat owing to the circulation of ground water driven by recharge and discharge, is a common phenomenon in ground water systems and is the focus of the work considered in this review. The potential for forced convection to perturb the geothermal gradient is quantified by two dimensionless numbers: the Nusselt number (the ratio of the total heat transfer to the heat transfer that would be expected in the absence of convection) and the more commonly used Peclet number (the ratio of convection to conduction). The Peclet number is also used in solute transport studies, where it represents the ratio of advection to dispersion. The thermal Peclet number, Pe, is given by:

$$\text{Pe} = \frac{\rho_w c_w q L}{\kappa_e} \tag{3}$$

where L is a characteristic length.

Early workers focused on analytical solutions of the one-dimensional form of Equation 1. Later, numerical solutions were used to consider more complicated boundary conditions in one-dimensional problems and to study basin-scale transport in two and three dimensions. Generic codes with user's manuals and graphical user interfaces (GUIs) are available to solve coupled ground water flow and heat transport problems (Table 2). Some of these codes were designed to solve complex problems involving geothermal systems (e.g., TOUGH2) or geologic processes (e.g, SHEMAT, BASIN2), while others (e.g., VS2DH, SUTRA, and FEFLOW) were introduced for simulation of

shallow aquifers. The ground water flow equation is coupled to the heat transport equation through the velocity term ($q = vn$ in Equation 1 where v is velocity). Velocity is dependent on hydraulic conductivity, which is partly governed by fluid density and viscosity, both of which vary with temperature. Because changes in viscosity with temperature are greater than changes in density, some codes (e.g., VS2DH) consider the temperature dependence of viscosity but assume a constant density fluid. To incorporate the nonlinearity caused by the dependence of hydraulic conductivity on temperature, the ground water flow and heat transport equations are solved iteratively to allow feedback between the solutions within a time step.

Temperature Profiles

Background

Following Parsons (1970), we consider two zones in the subsurface, the surficial zone and the geothermal zone (Figure 2). In the absence of ground water flow, subsurface temperature in the geothermal zone, which occurs below a depth of around 10 m, normally follows the geothermal gradient (Figure 2), usually represented by an increase of 1°C per 20 to 40 m of depth. Within the geothermal zone, the temperature profile is not subject to seasonal variations and is expected to be approximately linear except when perturbed by ground water flow, although changes in thermal conductivity also curve the thermal profile. Ground water flow perturbs the geothermal gradient by

Table 2
#### Selected Ground Water Flow and Heat Transport Codes

Code	Documentation	Comments	Selected Applications
HST3D[1]	Kipp 1997	three-dimensional, finite differences; ArgusONE GUI	Bravo et al. 2002
SHEMAT	Clauser 2003 (includes CD-ROM with code and GUI)	three-dimensional, finite differences; GUI	Rhine Graben cross section (Clauser 2003, sec. 6.5)
HEATFLOW[2]	Molson and Frind 2000	three-dimensional, finite elements	Molson et al. 1992; Yang et al. 1996
TOUGH2[3]	Pruess et al. 1999; Pruess 2004	three-dimensional, integral finite differences; variably saturated porous and fractured media	Birkholzer et al. 2004
SUTRA[1]	Voss and Provost 2002	three-dimensional, finite elements; variably saturated porous media; ArgusONE GUI	Bundschuh 1993; Screaton and Ge 1997; Burow et al. 2005
FEFLOW	Waterloo Hydrogeologic (http://www.flowpath.com)	three-dimensional, finite elements; variably saturated porous media; GUI	Marechal et al. 1999
BASIN2[4]	Bethke et al. 2002	two-dimensional, finite differences; postprocessor B2plot	Bethke 1985, 1986; Bethke et al. 1988; Corbet and Bethke 1992; Williams et al. 1999
VS2DH[1]	Healy and Ronan 1996	two-dimensional, finite differences; constant density fluid; variably saturated porous media; GUI	Ronan et al. 1998; Constantz 1998; Bartolino and Niswonger 1999; Su et al. 2004

[1]USGS codes (http://water.usgs.gov/software/ground_water.html/; access date: 12/7/04).
[2]http://sciborg.uwaterloo.ca/~molson/; access date 12/7/04.
[3]http://esd.lbl.gov/TOUGH2; access date 12/7/04.
[4]http://www.geology.uiuc.edu/~bethke/hydro_b2.htm; access date 12/7/04.
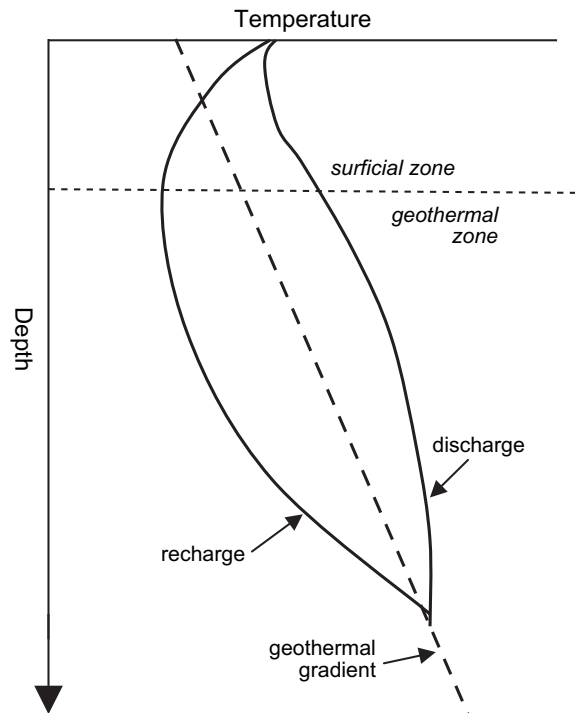
Figure 2. Schematic temperature profiles showing deviations from the geothermal gradient caused by surface warming in the surficial zone and convection in the geothermal zone. Recharge (downward movement of ground water) results in concave upward profiles, whereas discharge (upward movement) results in convex upward profiles. (Modified from Taniguchi et al. 1999a.)

infiltration of relatively cool water in recharge areas and upward flow of relatively warm water in discharge areas, causing concave upward profiles in recharge areas and convex upward profiles in discharge areas (Figures 2 and 3).

Within the surficial zone (Figure 2), temperature is influenced by seasonal heating and cooling of the land surface. Shallow ground water temperature is ~1°C to 2°C higher than the mean annual surface temperature. The amplitude of temperature fluctuations decreases with depth; below ~1.5 m temperatures are not significantly influenced by diurnal fluctuations at the land surface (Silliman and Booth 1993). Temperature profiles in the surficial zone potentially provide information about seasonal recharge/discharge events from precipitation and interchange with surface water.

Basic Models

Analysis of vertical differences in subsurface temperature was introduced into the hydrogeological literature by Suzuki (1960), who presented an analytical solution of the one-dimensional transient heat-flow equation and used it to estimate infiltration from a rice paddy. Refinements to Suzuki's model by Stallman (1965) firmly established the potential for using temperature profiles in ground water studies of the surficial zone. Stallman (1965) considered one-dimensional ground water flow through an infinitely long column with sinusoidally varying surface temperature and ground water temperature at depth equal to the mean surface temperature. He showed

how Suzuki's solution could be used to analyze the temperature profile and calculate vertical velocity (i.e., rates of recharge and discharge). The seminal work by Suzuki and Stallman initiated continuing contributions on the application of temperature measurements to ground water studies by investigators from Japan and the United States. For example, Cartwright (1974) applied the Stallman model to several sites in Illinois. Taniguchi (1993) developed type curves and applied the method to calculate recharge and discharge at two sites in Japan: Nagaoka Plain (Taniguchi 1993) and Nara Basin (Taniguchi 1994). The Nagaoka profiles (Figure 4) show the seasonal variability in the temperature profile, with measurements taken at 1-m depth intervals every 2 months during 1982 to 1983. These profiles illustrate well the characteristic elongation of the profile in the recharge area caused by downward movement of ground water (Figure 4a) and compression in the discharge area caused by upward movement of ground water (Figure 4b).
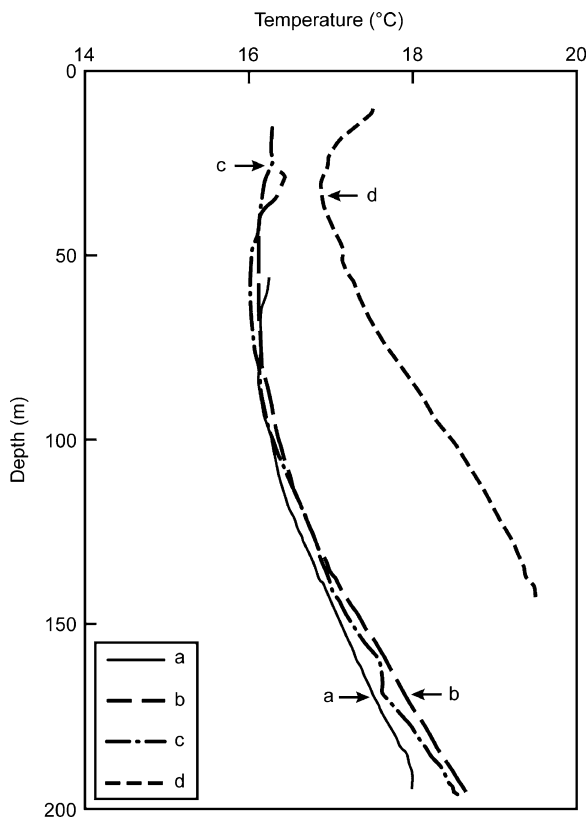
In the same year as the publication of Stallman's paper, Bredehoeft and Papadopulos (1965) introduced a one-dimensional steady-state analytical heat transport model to calculate vertical leakage through a confining bed under constant temperature boundary conditions. Type curves are matched to a temperature profile measured in a well penetrating the confining bed. Vertical ground water velocity is computed from the Peclet number (Equation 3, with L equal to the distance between two measuring points), which is read from a type curve. Although the Bredehoeft and Papadopulos model was developed for analyzing profiles across a confining bed within the geothermal zone, the model is applicable to other one-dimensional problems where temperature is constant at either end of the system. Cartwright (1970), for example, used the model to calculate ground water discharge from the Illinois Basin in southern Illinois, and Taniguchi et al. (2003) used the model to estimate submarine ground water discharge (SGD) in a coastal aquifer. Boyle and Saleem (1979) developed a computer code to do the type curve matching and applied the model to calculate leakage through glacial drift to an underlying aquifer. Other applications of the model include Sorey (1971), Jaynes (1990), Ferguson et al. (2003) and Taniguchi et al. (2003).

Given the uncertainty in parameter values, however, it is possible to fit a temperature profile to a one-dimensional model, which assumes strictly vertical flow, when the profile is influenced by horizontal flow (Reiter 2001; Ferguson et al. 2003), in which case the resulting estimates of velocity will be incorrect. Obviously, care must be taken to ensure that the assumptions implicit in the Bredehoeft and Papadopulos model are appropriate to the field situation. To address concerns over effects of horizontal flow on measured temperature profiles, Lu and Ge (1996) presented an extension of the Bredehoeft and Papadopulos model to include horizontal flow of water and heat by means of a sink/source term in the one-dimensional governing equation. Their method allows for calculation of both vertical and horizontal ground water flux, given vertical and horizontal temperature gradients. They concluded that the error caused by using the

original Bredehoeft and Papadopulos model is negligible when the horizontal heat and fluid flow are <10% of the vertical.

Other researchers considered the problem investigated by Bredehoeft and Papadopulos but used: (1) boundary conditions specified in terms of heat flux (Lachenbruch and Sass 1977, also Deming 2002, sec. 11.4); (2) a plot of the ratio of the change in temperature to change in depth ($\Delta T/\Delta z$) vs. temperature (T) (Mansure and Reiter 1979); and (3) a plot of conducted heat flow ($= \kappa_e \Delta T/\Delta z$) vs. temperature (Reiter et al. 1989). In an application of the Mansure and Reiter (1979) method, Dapaah-Siakwan and Kayane (1995) calculated leakage rates beneath Tokyo, Japan, using temperature profiles in 30 wells (Figure 3). Applications of the Reiter et al. (1989) method were reported by Reiter et al. (1989), Wade and Reiter (1994), and McCord et al. (1992). Reiter (2001) extended the method to estimate both vertical and horizontal ground water fluxes from vertical and horizontal temperature gradients, and Ferguson et al. (2003) used the method to calculate leakage to an aquifer in Manitoba, Canada.

Thus, in the very same year, Stallman (1965) introduced a model for transient analysis of temperature profiles in the surficial zone and Bredehoeft and Papadopulos (1965) presented a model for steady-state analysis in the geothermal zone. These two pioneering papers form the basis for most subsequent work on ground water temperature profiles, in which investigators modified and extended the models to consider other types of boundary conditions and application to a variety of hydrogeological settings.

Interchange with Surface Water

Temperature profiles beneath streams (Figure 5) are similar in nature to profiles measured in ground water basins (compare Figures 4 and 5); profiles are elongated in losing reaches where stream water recharges the aquifer and compressed in gaining reaches (Figure 5a). Moreover, the temperature profile beneath a stream shifts seasonally forming an annual envelope (Figure 5b) similar to the seasonal suite of profiles in recharge/discharge areas in ground water basins (Figure 4). The profiles shown in Figure 5b record temperature to a depth of 15 m beneath the streambed of the Rio Grande River in central New Mexico during the period September 1996 to August 1998.
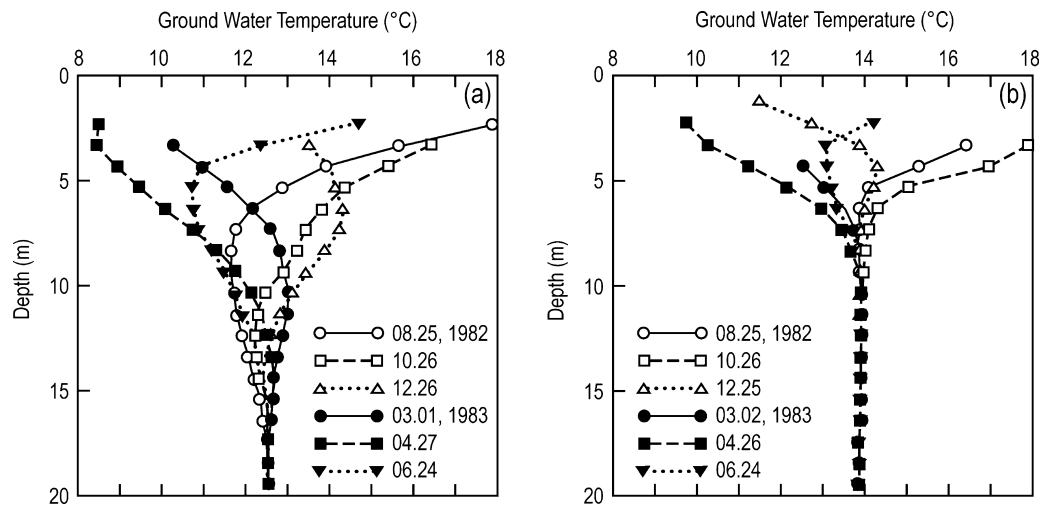
Workers analyzing temperature profiles beneath streambeds use both analytical and numerical solutions of the heat transport equation to calculate seepage through the streambed and hydraulic conductivity of the streambed sediments. For example, Silliman et al. (1995) developed a modified version of Stallman's (1965) analytical solution to calculate fluxes from losing reaches of a stream using the following boundary conditions:

$$T(0, t) = T_w(t) \tag{4a}$$

$$\lim_{z \to \infty} T(z, t) = T_A \tag{4b}$$

where T_w (t) is the temperature at the base of the water column in the stream at time t and T_A is the temperature at large depth beneath the streambed. The upper boundary condition (Equation 4a) is discretized to represent the measured fluctuation of temperature in the stream. The solution is sensitive to velocities greater than ~1 cm/d. The solution by Silliman et al. (1995) was intended for use in losing reaches as larger diurnal fluctuations in streambed temperature are expected beneath losing reaches than beneath gaining reaches (Figure 1 in Constantz and Stonestrom 2003). Nonetheless, Becker et al. (2004) successfully used the solution to calculate ground water fluxes to a gaining stream in New York State.

Taniguchi et al. (2003) applied both the Stallman (1965) and the Bredehoeft and Papadopulos (1965) models to temperature profiles measured in a coastal area in Australia to estimate the fresh water component of SGD. The ground water flux calculated from the temperature measurements was less than half the SGD measured in seepage meters, which included recirculated sea water as well as fresh water discharge from inland areas. This work shows the potential for using temperature measurements in studies aimed at assessing the magnitude and significance of SGD (Burnett et al. 2003). In another study of a coastal region, Taniguchi (2000) used temperature profiles to determine the depth to the salt-water interface by calculating the deviation of the measured temperature from the geothermal gradient.

Figure 3. Temperature profiles affected by convection and surface warming, Tokyo Basin, Japan. Profiles a to c are concave upward, indicating recharge, and profile d is convex upward below 50-m depth, indicating discharge. (Modified from Taniguchi et al. 1999a and based on work by Dapaah-Siakwan and Kayane 1995.)

Figure 4. Seasonal variation in temperature profiles, Nagaoka area, Japan, showing (a) elongated profiles in the recharge area and (b) compressed profiles in the discharge area (Taniguchi 1993).

Numerical solutions allow more flexibility in defining boundary conditions and variability in hydrogeological parameters. In an early application of a numerical model to a ground water/stream system, Wankiewicz (1984) analyzed temperatures beneath streams in the Northwest Territories, Canada, in an area of permafrost. In an important extension of Stallman's method, Lapham (1989) used an explicit finite-difference approximation to the transient, one-dimensional form of Equation 1 considered by Stallman (1965) to provide flexibility in representing boundary conditions. In Lapham's method, nonsinusoidal temperature variations in the stream may be simulated and ground water temperature at depth may be set at ambient conditions, whereas Stallman's model prescribes sinusoidal variations in surface temperature and specifies ambient ground water temperature at depth equal to the mean surface temperature. Simulated annual or daily temperature profiles are matched to the envelope of measured profiles (e.g., Figure 5b) to estimate ground water flux. The method is sensitive to ground water velocities in the range of 0.3 to 30 cm/d when the yearly fluctuation in stream temperature is around 25°C. By substituting ground water flux and measurements of hydraulic gradient into Darcy's law, the effective hydraulic conductivity of the streambed sediments is calculated. Lapham (1989) used the model to estimate vertical ground water flux and effective hydraulic conductivity of streambed sediments in aquifer-stream systems in the Northeastern United States. Other applications of the method include calculating ground water flux to a lake (Krabbenhoft and Babiarz 1992) and a wetland (Hunt et al. 1996) in Wisconsin and fluxes to gaining and losing reaches of streams in Ohio (Fryar et al. 2000).

Similarly, Constantz and coworkers at the USGS (Stonestrom and Constantz 2003; Constantz et al. 2002; Bartolino and Niswonger 1999; Constantz 1998) used the ground water flow and heat transport code VS2DH (Table 2) to estimate vertical seepage and hydraulic conductivity of streambed sediments. (See Niswonger and Prudic 2003, for a summary of the procedure.) These studies rely on a one-dimensional analysis assuming strictly

vertical flow beneath the streambed. Two-dimensional applications of VS2DH to aquifer-stream systems were reported by Ronan et al. (1998) and Su et al. (2004). Bravo et al. (2002) and Burow et al. (in press) used temperature measurements in transient heat-flow simulations with HST3D and SUTRA, respectively (Table 2). Bravo et al. (2002) used a three-dimensional model to estimate hydraulic conductivity and flux across the water table in a wetland; Burow et al. (2005) performed two-dimensional simulations to estimate hydraulic conductivities and used the model to calculate dissolved organic carbon (DOC) flux to a wetland.
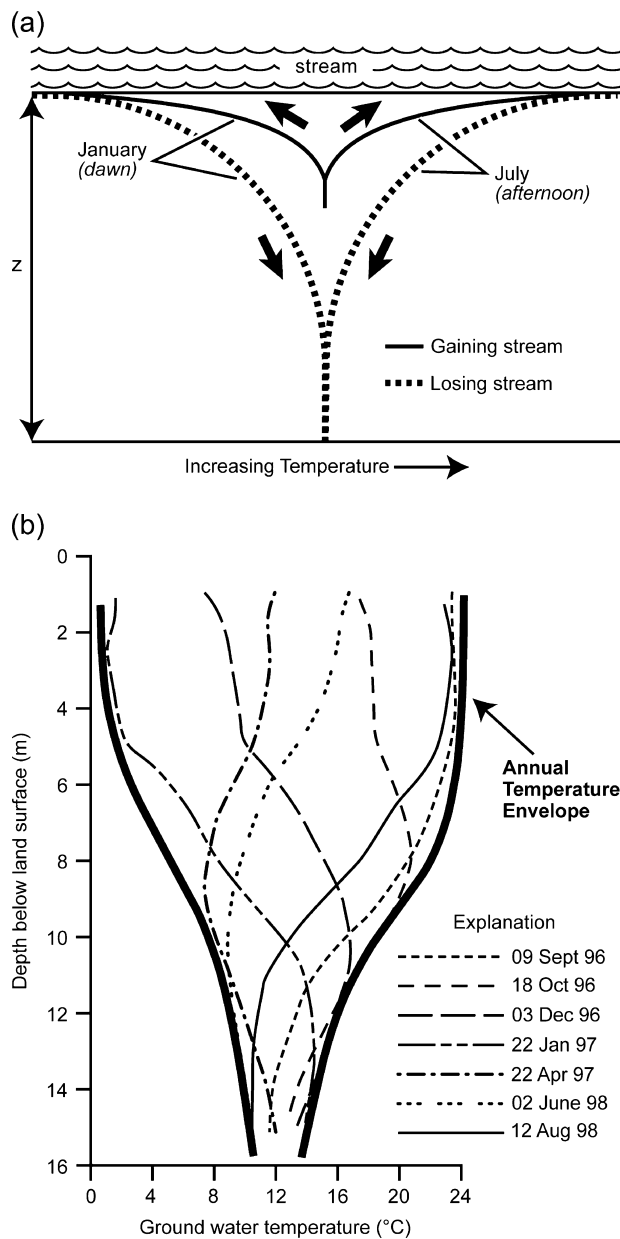
In summary, the Stallman (1965) model has been extended and generalized for application to aquifer-stream systems, most powerfully applied with the help of numerical solutions. The Stallman (1965) model has also proved useful to estimate fluxes to lakes and wetlands; both the Stallman and the Bredehoeft and Papadopulos models have been used in a coastal setting to estimate SGD.

Fractures

Temperature profiles measured in wells that intercept fractures may exhibit anomalies caused by movement of relatively cool or hot water in or out of the fracture. The inflection is sharp when the profile is influenced by isolated fractures that intersect the borehole in contrast to a large continuous fracture plane. In an early study, Trainer (1968) traced laterally continuous bedding plane fractures in a carbonate aquifer for several hundred meters by correlating inflections in temperature profiles. Drury (1989) used temperature logs to identify a fracture zone in granite. Other workers used changes in temperature to estimate fracture locations (Silliman and Robinson 1989) and to study the infiltration of surface water contaminated by sewage into a fractured limestone aquifer (Malard and Chapuis 1995).

Analytical solutions for two-dimensional transient analysis of fractured systems (Bodvarsson 1969; Ziagos and Blackwell 1986) are difficult to apply to field problems owing to difficulties in acquiring the necessary input data (Ingebritsen and Sanford 1998, p. 90–92). Ge (1998) presented a one-dimensional, steady-state analytical

(a)

stream

January
(dawn)

July
(afternoon)

z

—— Gaining stream
•••• Losing stream

Increasing Temperature ⟶

(b)

Depth below land surface (m)

Annual
Temperature
Envelope

Explanation

- - - - - 09 Sept 96
– – – – 18 Oct 96
— — — 03 Dec 96
— - — - 22 Jan 97
—·—·— 22 Apr 97
···· ···· 02 June 98
———— 12 Aug 98

Ground water temperature (°C)

Figure 5. Temperature profiles beneath streams. (a) Conceptual model showing profiles in gaining (upward flux) and losing (downward flux) stream reaches (modified from Constantz and Stonestrom 2003). (b) Ensemble of profiles showing seasonal variability in temperature beneath the Rio Grande River, central New Mexico; the envelope around the profiles defines the annual variation. Temperature was measured at 15-cm intervals using a probe equipped with a thermistor. (Modified from Bartolino 2003 and Bartolino and Niswonger 1999.)

solution to simulate the temperature signal caused by an isolated fracture or fractured zone. Normalized temperature profiles are matched to a set of type curves from which values of flow velocity in the fracture zone are calculated. In an application of the method to two temperature profiles from 1800-m-deep boreholes in volcanic tuffs and dolomite in southern Nevada, the calculated velocity was three to four orders of magnitude larger than the average fluid velocity in the vicinity of the borehole, indicating higher velocities in the fracture zone (Ge 1998). Fault zones also focus ground water and heat (e.g., Bodner et al. 1985; Painter et al. 2003).

Effects of Surface Warming

Changes in surface temperature affect subsurface temperatures in the surficial zone, while deeper portions of the profile reflect the influence of past climate. In thermal geophysics, it is standard practice to extract ground surface temperature (GST) histories from temperature profiles prior to analysis of the geothermal gradient for calculation of heat flow (e.g., Kohl 1998), but the surface thermal effects themselves are of interest and can be analyzed to reconstruct past climates (e.g., Lachenbruch and Marshall 1986; Pollack and Huang 2000). A thermal wave with a period of 1000 years can be detected to a depth of 400 m (Figure 3 in Pollack and Huang 2000). Paleoclimatic temperature signals are persistent and, although altered, are rarely "washed out" by ground water flow (Kohl 1998).

To analyze temperature profiles affected by surface warming, Taniguchi et al. (1999a) turned to Carslaw and Jaeger (1959) for a one-dimensional transient solution for heat flow in an infinitely long column subject to a linearly increasing surface temperature such as might be caused by urbanization or global warming. The initial condition and the boundary condition at the surface are, respectively:

$$T(z, 0) = T_0 + \lambda_I z \qquad (5a)$$

$$T(0, t) = T_0 + \lambda_t t \qquad (5b)$$

where z is depth from the land surface; T_0 is the surface temperature at $t = 0$; λ_I is the initial geothermal gradient; and λ_t is the rate of linear increase in surface temperature. The initial condition (Equation 5a) requires that λ_I is linear and therefore ground water velocity everywhere in the profile is initially assumed equal to zero.

The depth of the minimum temperature observed in the profile, which is related to the magnitude of the downward ground water flux, is plotted on a type curve and a value of dimensionless specific discharge, U ($= q_z c_w \rho_w / c \rho$), is read from the plot, from which q_z is calculated. In an application of the method, Taniguchi et al. (1999a) assumed surface temperatures in Tokyo had increased 0.025°C/year over the past 100 years and calculated vertical ground water flux by matching the solution to temperature profiles measured by Dapaah-Siakwan and Kayane (1995) (Figure 3). The depth of minimum temperature occurred 30 to 80 m below surface. Ferguson and Woodbury (2005) expanded upon the work of Taniguchi et al. (1999a) by using a numerical model to input measured GST at the upper boundary in lieu of Equation 5b. Other researchers in Japan investigated the effects of surface warming on temperature profiles qualitatively (Sakura et al. 2000); Salem et al. (2004) concluded that recent temperature profiles measured in the Nagaoka area are affected by surface warming to a depth of 60 m.

Related work investigated the effects of land use on subsurface temperature (Pluhowski and Kantrowitz 1963; Heath 1964; Taniguchi et al. 1997, 1999b). Studies of surface warming effects on ground water to date are limited but will be useful in future studies aimed at addressing concerns over the potential effects of global warming on water resources (Sophocleous 2004).

Basin-Scale Studies

Heat transfer is relevant to many topics in Earth Sciences including basin-scale processes (Person et al. 1996). Those approaching the problem from the field of geothermics regard perturbations in the thermal profile as anomalies, which might be explained by ground water flow; anomalously low thermal gradients may be due to ground water recharge, whereas anomalously high thermal gradients may be caused by ground water discharge. Following Lachenbruch and Sass (1977), Deming (2002, p. 344–350) showed that the effect of ground water flow on the thermal regime depends on the magnitude of the seepage velocity, q, the depth of circulation, and the term $\rho_w c_w / \kappa_e$.

Hydrogeologists, on the other hand, are interested in analyzing the anomalies caused by ground water flow in order to delineate ground water recharge and discharge areas, to estimate recharge/discharge rates independently of head data, and to constrain estimates of hydraulic conductivity. According to van der Kamp (1982, p. 32), "Detailed field studies frequently encounter thermal anomalies due to convection, to the extent that in areas of active ground water flow such anomalies may be expected to be the rule rather than the exception."

Schneider (1964) and Cartwright (1970) were among the first to use temperature profiles in the context of regional ground water systems. Domenico and Palciauskas (1973) presented a two-dimensional analytical solution of the heat transport equation to analyze perturbations in the thermal regime caused by convection in a ground water basin. They considered a cross section through a basin of length L_B and thickness z_o (Figure 6a,b) with constant hydraulic conductivity. The head at the water table is given by:

$$h(x, z_o) = z_o + \frac{\Delta H}{2}\left[1 - \cos\left(\pi x / L_B\right)\right] \qquad (6)$$

where ΔH is the change in elevation of the water table across the basin (Figure 6b). The bottom of the cross section is assumed impermeable and the sides are ground water divides. The solution of the flow problem is incorporated into the solution of the two-dimensional steady-state heat transport equation with constant temperature along the upper boundary of the cross section, specified heat flux at the lower boundary, and zero flux at the side boundaries. The Domenico and Palciauskas model was applied to ground water basins in Japan by Sakura (1993) and Inagaki and Taniguchi (1994).

Of course, when the entire aquifer is in the surficial zone, the expected thermal patterns and temperature profiles in the geothermal zone (Figures 2, 3, and 6c) will never develop. For example, Parsons (1970), who was among the first to use a numerical model to demonstrate the effect of ground water flow on the thermal regime, found that a hypothetical example of a 160-m-thick aquifer showed the classic deflections in isotherms, later predicted by Domenico and Palciauskas (1973), but temperatures in a 60-m-thick glacial complex were dominated by heat transfer from the surface.
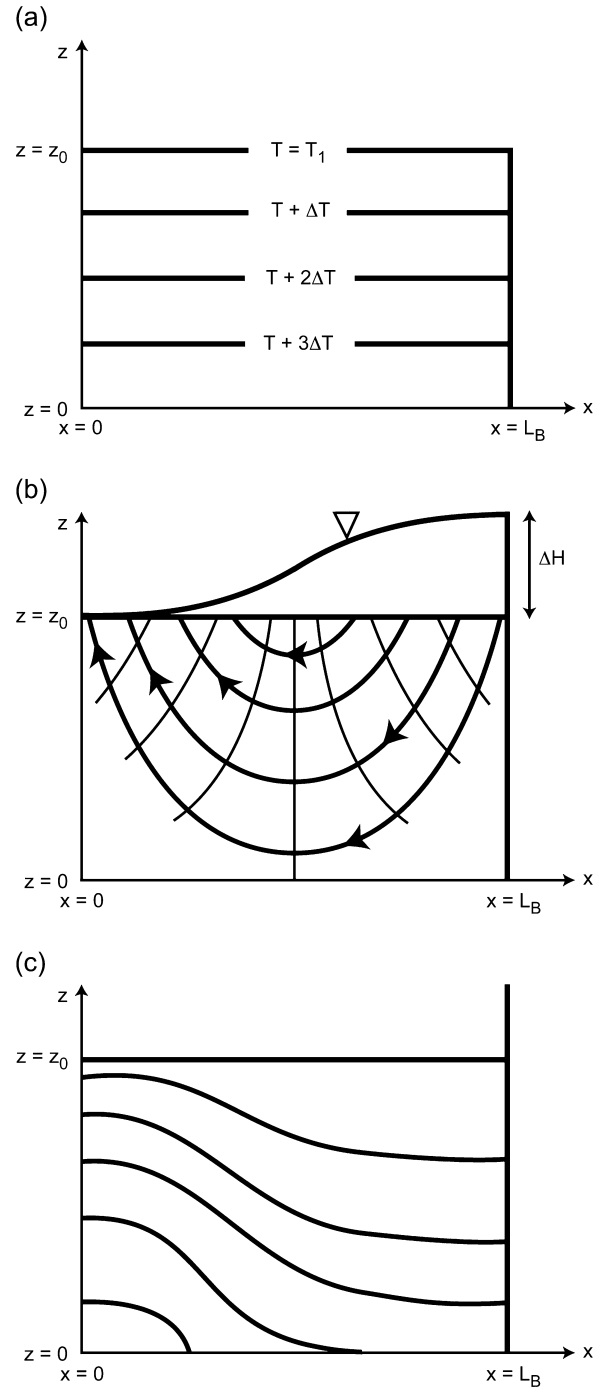
Figure 6. Basin-scale heat transport. (a) Horizontal isotherms in a cross section of a basin unaffected by convection; (b) Water table, equipotential lines, and ground water flow lines; (c) Isotherms perturbed by ground water flow for the flow system depicted in (b). (Modified from Domenico and Palciauskas 1973.)

Modified Thermal Peclet Number

The relative importance of convection over conduction can be quantified using the thermal Peclet number, Pe (Equation 3). van der Kamp and Bachu (1989) argued that the characteristic length in Equation 3 should be taken in the main direction of heat flux; for basins, the appropriate length is the vertical dimension z_o (Figure 6). In order to include the geometry of the basin, the Peclet

number in Equation 3 is multiplied by the aspect ratio, $A = z_o/L_B$, giving the modified Peclet number, Pe*:

$$Pe^* = \left(\frac{\rho_w c_w}{\kappa_e}\right)(q_x z_o A) \qquad (7)$$

where q_x is the horizontal ground water flux through the basin. van der Kamp (1984) pointed out that Equation 7 can be obtained by taking the ratio of the amount of heat convected horizontally (approximately equal to $\rho_w c_w q_x z_o \Delta T$, where ΔT is the change in temperature between the top and bottom of the system) to the amount of heat transferred vertically by conduction (approximately equal to $\kappa_e L_B \Delta T/z_o$). When the modified Peclet number or basin Peclet number (Equation 7) is $>10^{-1}$, it is expected that ground water flow will affect the thermal regime (van der Kamp and Bachu 1989); thermal springs occur when the basin Peclet number is ~1 (van der Kamp and Bachu 1989). When it is >5, it is expected that the system will be dominated by ground water flow and essentially isothermal (Woodbury and Smith 1988).

It is clear from Equation 7 that the importance of convection increases with flow through the basin ($q_x z_o$) and aspect ratio ($A = z_o/L_B$). In other words, for basins of length L_B, perturbation of the thermal regime is more likely where there is deep circulation of ground water (i.e., large z_o) and/or high velocities. Basin Peclet numbers and aspect ratios for five ground water basins (Table 3) show that for the systems considered, the modified Peclet number ranges from 7×10^{-6} for a system dominated by conduction up to 4.2 for a system with strong ground water flow.

Examples

In addition to the work of Domenico and Palciauskas (1973), numerous modeling studies of hypothetical ground water basins have demonstrated theoretically that ground water flow causes perturbations in the thermal regime in two dimensions (Parsons 1970; Smith and Chapman 1983; Garven and Freeze 1984; Woodbury and Smith 1988; Forster and Smith 1989) and three dimensions (Woodbury and Smith 1985). These studies show that the nature and magnitude of the perturbations caused by ground water flow are affected by anisotropy and heterogeneity in hydraulic conductivity, recharge rate, and configuration of the water table.

Yet, the thermal regime in some basins (e.g., the Paris Basin and the Michigan Basin) appears to be unaffected by convection (Jessop 1989). Moreover, while "modeling exercises using reasonable parameter values demonstrate the feasibility of basin-scale heat transfer by ground water, applications to real world basins are often ambiguous, because direct measurement of ground water velocities (or permeabilities) on the appropriate scale is impossible, and any spatial variations in heat flow are often subject to alternative explanations" (Ingebritsen and Sanford 1998, p. 121). An example of such ambiguity is found in the Western Canada sedimentary basin, where there is a general correspondence between areas of low thermal gradients in the recharge area and high thermal gradients and upward flow of ground water in the discharge area. Some researchers (Hitchon 1984; Majorowicz et al. 1984) suggested that ground water flow affects the thermal regime in this area, but others (Bachu 1985, 1988) argued that permeabilities are too small to allow for significant thermal effects by convection and suggested that the thermal regime is controlled by conditions in the underlying crystalline basement, an argument refuted by Majorowicz (1989). A regional-scale ground water model (Garven 1989) showed that while deep circulation of ground water during the late Tertiary period may have perturbed the thermal regime, the current flow system is likely dominated by shallow local flow systems. Corbet and Bethke (1992) also showed that current rates of ground water flow appear to be too slow to transport significant heat. Their two-dimensional coupled ground water and heat-flow model reproduced observed temperature gradients by using spatially variable values of thermal conductivity of a shale layer. The low Peclet number in the Beaverhill Lake aquifer (Table 3), which is located in this basin, also suggests a system dominated by conduction. In another controversy over the importance of convection (Forster and Smith 1989; Ingebritsen et al. 1992; Blackwell and Priest 1996), Ingebritsen et al. (1996) made convincing arguments in support of convection in the Cascade Range, Oregon, United States.

Some argue that low permeabilities measured in the field are often representative of the core scale rather than larger basin-scale permeabilities; therefore, the presence of low-permeability material does not necessarily preclude thermal perturbation by convection. For example, McCord et al. (1992) hypothesized that regional permeability in the northern San Juan basin of Colorado-New Mexico is controlled by fracture zones so that ground water flow through fractures could be responsible for the observed thermal anomalies.

Table 3
Modified Thermal Peclet Numbers (Pe*) and Aspect Ratios (A = z_o/L_B) for Five Basins

Ground Water Basin	Pe*	A	References
Beaverhill Lake aquifer, Alberta, Canada	7×10^{-6}	5×10^{-5}	van der Kamp and Bachu 1989; Bachu 1985
Nara Basin, Japan	0.27	0.011	Inagaki and Taniguchi 1994
Yonezawa Basin, Japan	0.44	0.024	Sakura 1993
Bath-Bristol Basin, England	1.8	0.029	van der Kamp and Bachu 1989; Andrews et al. 1982
Eastern Snake River Plain, Idaho, USA	4.2	0.016	van der Kamp and Bachu 1989; Brott et al. 1981

In fact, evidence for convection is found in many basins. In the North Slope Basin, Alaska, thermal gradients in the foothills are lower than in the coastal plain, suggesting regional ground water flow transports heat by convection from the foothills to the coast (Deming et al. 1992). Using this conceptual model, heat-flow measurements were reproduced in a two-dimensional profile model of ground water and heat flow (Deming 1993). A two-dimensional model across the Rio Grande Rift, New Mexico, showed that ground water flow caused an increase in temperature by as much as 40°C in the discharge area around Socorro Springs (Mailloux et al. 1999) as postulated from temperature measurements (Barroll and Reiter 1990). Other examples of thermal disturbance by convection were reported by Kilty and Chapman (1980) and for the Rhine Graben (Person and Garven 1989; Clauser 1989, 2003); the Bohemian Cretaceous Basin, Czechoslovakia (Cermak and Jetel 1985); the Uinta Basin, Utah (Chapman et al. 1984; Willet and Chapman 1989); the Eastern Snake River plain, Idaho (Brott et al. 1981); the Denver, Kennedy, and Williston Basins of the Great Plains of the United States (Gosnold 1985, 1990); the Gulf Coast basin of the United States (Bodner et al. 1985); and sites in Hungary (Alfoldi et al. 1985) and Japan (Sakura 1993; Inagaki and Taniguchi 1994; Dapaah-Siakwan and Kayane 1995; Uchida 1998; Salem et al. 2004).

In mountainous regions where ground water flows in low-permeability, fractured bedrock with a thin covering of surficial material (Forster and Smith 1989) or in mountainous karst regions (Alfoldi et al. 1985), ground water flow systems and temperature profiles differ from the typical conceptual model for sedimentary basins (Figures 6b and 6c) and are dependent on site-specific geology including location of faults and fracture zones, topography, and depth to water table. Results of numerical modeling studies (Forster and Smith 1989) show that when the water table is deep, ground water recharge rather than permeability determines whether ground water flow perturbs the thermal regime. Ingebritsen et al. (1992, 1996) presented convincing evidence for perturbation of the thermal regime by convection in the Oregon Cascade Range, United States. In other studies in mountainous regions, Yuhara et al. (1989) simulated convection to hot springs in the Japan Alps, and Marechal et al. (1999) simulated the effects of construction of a tunnel on heads and temperatures in the French and Italian alps. Thomas and Paillet (1996) made deductions about regional ground water flow from the temperature profile in a single deep borehole in Hawaii.

Other studies focus on the occurrence of hot springs in regional flow systems influenced by thermal effects such as the famous springs at Bath, England (Andrews et al. 1982; Atkinson and Davison 2002); hot springs in Japan (Yuhara et al. 1989), Hungary (Alfoldi et al. 1985), Switzerland (Bodmer and Rybach 1985), Spain (Sanz and Yelamos 1998; Sanchez-Navarro et al. 2004), and in the Valley and Ridge Province of the United States (Hobba et al. 1979). In a typical conceptual model, ground water flows deep into the basin where it is heated and then flows back to the surface to discharge in hot springs. Faults may act as preferential flow paths to focus flow to springs (Forster and Smith 1989; Mailloux et al.

1999). Also, inferences about flow systems can be made using temperature of springs, including cold springs (Bundschuh 1993; James et al. 2000).

While models of ground water flow and heat flow in individual basins lend insight into site-specific processes, the power of numerical models lies in using temperature data to solve the inverse problem.

Formal Solutions of the Inverse Problem

All the aforementioned studies relate in some way to solving the inverse problem by using temperature data to estimate ground water flux and hydraulic conductivity. For example, in an early qualitative solution of the inverse problem, temporal variations in temperature profiles in a well near the Mohawk River, New York, were used to make inferences about the location of zones of high hydraulic conductivity (Winslow 1962). More recently, Macfarlane et al. (2002) injected heated water and then used temperature logs to characterize heterogeneities in the Dakota Aquifer, Kansas. As aforementioned in "Temperature Profiles," Stallman (1965) and Bredehoeft and Papadopulos (1965) used temperature data to solve for ground water flux in the one-dimensional form of the heat transport equation. Following Stallman (1963), there is general recognition that temperature data can be used to constrain estimates of hydraulic conductivity (e.g., Lapham 1989; Niswonger and Prudic 2003) including vertical hydraulic conductivity of streambed sediments (Constantz et al. 2003a, 2003b, Conlon et al. 2003; Prudic et al. 2003; Su et al. 2004) and basin-scale permeability (Forster and Smith 1989; McCord et al. 1992; Deming 1993). In these studies, observed temperature data are fitted to simulated temperature profiles by trial and error or are used in trial-and-error calibration of coupled ground water flow and heat-flow models.

Instead of trial-and-error fitting, some investigators have built on Stallman's (1963) suggestion that heads and temperatures might be used jointly to estimate ground water velocity and hydraulic conductivity, by attempting formal inversion of a coupled ground water flow and heat transport model (Woodbury and Smith 1988; Wang et al. 1989; Bravo et al. 2002). Requiring that simulated temperatures match observed temperatures provides an additional constraint on calibrating a flow model because the velocities required to calibrate the heat-flow model must also calibrate the ground water flow model. Use of temperature data in calibration can be quite powerful provided that thermal properties of the flow system can be specified a priori so as not to introduce additional calibration parameters. Fortunately, thermal properties including thermal conductivity vary over a much smaller range than hydraulic conductivity (Figure 2 in Constantz and Stonestrom 2003).

Bravo et al. (2002) used heads and hourly temperature data in HST3D (Table 2) to calibrate a coupled ground water flow and heat transport model of a ground water/wetland system in Wisconsin using the parameter estimation code PEST (Doherty 2004). Time series of head and temperature were collected in more than 30 wells for a 6-year period using different sampling frequencies. Data were analyzed to determine the response

time of the system to fluctuations in head and temperature. Based on this analysis, a steady-state ground water flow model was assumed, but the heat transport model considered transient fluctuations in temperature during a 1-d simulation period in October 1996 when shallow subsurface temperatures varied in the range of 9.4°C to 14.6°C. (Annually, shallow subsurface temperature varied around 20°C.) A 120-m-thick section of aquifer was simulated with temperature specified at the top boundary and heat flux specified at the bottom boundary. Spatially constant thermal properties were assumed. Temperature measurements taken every 4 h at four depths below the surface were used in the calibration. Calibrations that failed to converge using only head data, did converge when constrained by temperature data. The parameter estimation procedure yielded hydraulic conductivity (230 to 340 cm/d) and flux to the wetland (0.66 to 0.95 cm/d). The Bravo et al. (2002) study is significant because it reports a successful use of temperature time series data in parameter estimation in a field setting characterized by low heat flow (a basal heat flow of 42mW/m^2, compared to the continental average of 60 mW/m^2). In addition to specifying the thermal properties of the system, success of the method requires high-frequency time series of both head and temperature. Moreover, the fluctuation in annual and daily temperature must be sufficiently strong to perturb the surficial thermal regime and ground water velocities must be above a minimum threshold (thought to be around 0.3 cm/d; Lapham 1989) to perturb the surface temperature signal.

Summary and Conclusions

Although heat-flow theory has been influential in the development of the theory of ground water flow, interest in using temperature measurements themselves in ground water investigations has been sporadic. Early papers on this topic, published in the 1960s and 1970s, called attention to the potential for using heat as a tracer and temperature as a surrogate for head measurements to estimate ground water fluxes. However, workers understandably continued to focus on measuring and simulating head directly. With the recent availability of improved temperature sensors and relatively inexpensive data loggers, which enable remote and continuous measurements, as well as improved numerical codes for simulating coupled ground water flow and heat transport (Table 2), there is renewed interest in using heat as a ground water tracer.

Workers have continued to explore themes investigated by early researchers: (1) using heat as a tracer to detect infiltration of surface water and fractures and (2) solving the inverse problem by using temperature to estimate ground water velocity and hydraulic conductivity. It is perhaps not surprising that temperature measurements are useful when there is a marked contrast in surface water and ground water temperatures and in basin settings where there is high heat flow and associated thermal springs. Suzuki (1960) and Stallman (1965), however, postulated that ground water velocity could be estimated from seasonal fluctuations of temperature at the land surface. Using a version of Stallman's model, Lapham (1989) verified that monthly and yearly variations in subsurface temperature

profiles beneath streams can be used to estimate ground water velocities in the range of 0.3 to 30 cm/d when the yearly fluctuation in stream temperature is around 25°C. Building on ideas of Stallman (1963), Bravo et al. (2002) successfully used measurements of surficial temperature in parameter estimation with a coupled ground water and heat-flow model of a wetland system in Wisconsin where the basal heat influx is relatively low.

Capitalizing on the availability of improved temperature sensors and numerical codes, investigators are just starting to explore the full potential for using temperature measurements in a wide variety of hydrogeological settings. The utility of temperature measurements in estimating fluxes in ground water-stream systems is now well established (Stonestrom and Constantz 2003). Temperature measurements have also proved useful in estimating ground water flux in wetland settings (e.g., Hunt et al. 1996; Bravo et al. 2002; Burow et al. 2005), in lakes (Lee 1985; Krabbenhoft and Babiarz 1992), and in coastal aquifers (Taniguchi 2000) including estimation of SGD (Taniguchi et al. 2003). Building on early work by Bredehoeft and Papadopulos (1965) and Stallman (1965), temperature profiles have been used with success in basins to estimate ground water recharge and discharge rates (e.g., Cartwright 1970; Taniguchi 1993, 1994; Ferguson et al. 2003). Analysis of surface warming effects on temperature profiles (Taniguchi et al. 1999a; Ferguson and Woodbury 2005) has potential application in studies of the effects of climate change on ground water resources. The work of Conant (2004) shows the possibility of using temperature to help delineate the details of flow in the hyporheic zone, which is a recent focus of interest of hydrogeologists and ecologists (Hayashi and Rosenberry 2002). Additionally, many investigators have used coupled ground water and heat-flow models to analyze heat transport in ground water basins.

One of the most powerful uses of temperature data is in formal solutions of the inverse problem. It is generally recognized that head data alone are not sufficient to calibrate a ground water flow model, while estimates of ground water flux and/or information on the movement of solute and/or heat help constrain the calibration. Joint inversion of ground water and heat-flow models, as recently demonstrated by Bravo et al. (2002) for a wetland system in Wisconsin, holds promise for general application to sites where there is sufficient fluctuation in surficial temperatures and velocities are sufficiently high to perturb the temperature signal. While additional studies of joint inversion of head and temperature data in a variety of hydrogeological settings are needed to determine the general applicability and limitations of the method, the body of work on heat as a ground water tracer leaves no doubt that analysis of temperature measurements ought to be a standard tool in the ground water hydrologist's toolbox.

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