Quantifying Ephemeral Streambed Infiltration from Downhole Temperature Measurements Collected Before and After Streamflow

Charles E. Dowman V, Ty P.A. Ferré,* John P. Hoffmann, Dale F. Rucker, and James B. Callegary

ABSTRACT

A constant flux infiltration experiment was conducted to determine the feasibility of using downhole temperature measurements to estimate infiltration flux. Temperatures measured using a downhole thermistor within a 15.4-m-deep borehole compare well with temperatures measured with buried thermocouples in an adjacent borehole to 5 m depth. Numerical forward model simulations were conducted using VS2DI. A numerical sensitivity analysis showed that the temperature profile was most sensitive to the average temperature of the infiltrating water, the infiltration flux, and the specific heat capacity of dry soil. The high sensitivity of these variables allows for a simple sequential optimization to be used to estimate the average temperature of the infiltrating water, the water flux, and the specific heat capacity of dry soil from numerical inversion of temperature measurements. Downhole temperature measurements could be a useful complement to shallow streambed temperature methods, allowing for better quantification of the contribution of streambed infiltration to basin-scale recharge.

Improved ability to monitor the rate of infiltration during flow in an ephemeral stream would provide information that would greatly improve our understanding of how these surface water bodies recharge underlying aquifers. Typically, subsurface water flux is estimated by monitoring changes in water content or water pressure with depth and time. For example, repeat water content profiling can be used to determine the total change in storage during transient flow. Alternatively, measured pressure gradients can be used with knowledge of the soil hydraulic properties to calculate water flux. In addition to these approaches, solute tracer tests can be conducted to infer infiltration rates. Although any of these approaches can be used, each has practical limitations for application beneath ephemeral streams. Soil pressure monitoring using tensiometers or heat-dissipation sensors and water content profiling using time domain reflectometry (TDR) require an individual probe for each monitoring depth. Water content profiling using neutron probes involves licensing restrictions and the need for soil-specific calibrations. All of these physically based methods require either buried instrumentation in the streambed that could be removed by flood waters, or in-stream monitoring, which can present significant safety concerns. Solute tracer tests generally are labor intensive and costly because they require chemical analyses and controlled addition of tracer solutions. As a result, there is a need for an alternative method to determine the infiltration rate beneath an ephemeral stream that does not rely on measurements made during streamflow. Furthermore, it would be advantageous if this method were applicable in deep vadose zones that commonly are associated with arid and semiarid environments and if measurements could be made in existing wells.

The use of heat as a tracer to monitor subsurface water flow has been shown to be a promising alternative to water content or water pressure monitoring and chemical tracer tests (Constantz et al., 2001, 2002, 2003). Suzuki (1960) used measured temperature profiles to estimate infiltration rates below flooded rice fields. Lapham (1989), Silliman and Booth (1993), and Constantz et al. (1994) used measured shallow-sediment temperatures to estimate water infiltration rates below streams. Ronan et al. (1998) used a variably saturated flow and heat transport model, VS2DH (Healy and Ronan, 1996), to estimate infiltration rates below Vicee Canyon in northwestern Nevada. Bartolino and Niswonger (1999) used temperature profiles measured beneath the middle Rio Grande Basin near Albuquerque, NM to evaluate the rate of vertical flux between the Rio Grande and the underlying aquifer system. In addition to these investigations that relied on variations in shallow temperature measurements, deep temperature profiles have been used to estimate fluxes through thick vadose zones and in deep, basin-scale flow systems (e.g., Boyle and Saleem, 1979; Sass et al., 1988; Rousseau et al., 1998; and Reiter, 2001). In some of these studies, measured changes in the geothermal gradient due to the effects of infiltrating water were used to identify areas of active infiltration.

The objective of this study was to determine whether temperature profiles measured within a cased borehole before and after a single flow event could be used to infer the rate of infiltration during the flow event.

THEORY

Heat energy can be transferred via conduction, advection, and radiation. Fourier's Law describes conductive heat flux as the product of the thermal conductivity and the temperature gradient and is similar in form to Darcy's equation for water flux. Radiant-heat flux is the transfer of energy by electromagnetic waves and is governed by the temperature of the heat source. As solar radiation is absorbed at the soil surface, heat energy is transmitted into the shallow subsurface. Radiant

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Abbreviations: TDR, time domain reflectometry.

heat transport is regarded as a negligible component of heat transfer during transient water flow. For most subsurface applications, advection by flowing water and conduction are the primary mechanisms for the transport of heat. An increase in temperature decreases the dynamic viscosity and the density of water. Although these are competing effects, increasing temperature generally leads to an increase in the hydraulic conductivity of the medium. Heat transport is coupled directly with water flow through advection. In addition, under unsaturated conditions, the thermal and hydraulic conductivities are both dependent on the volumetric water content of the medium.

Stallman (1965) initially proposed that the dependence of heat transport on water flux could allow for the use of water temperature measurements to determine water flux. In an investigation of vertical, nonisothermal infiltration through a homogeneous medium, Stallman (1965) used a coupled continuity equation of heat energy and water mass. In developing this continuity equation, one-dimensional water flow through a variably saturated medium can be described by the following form of Richards' equation (Richards, 1931), which describes the time rate of change of water storage due to capillary and gravitational gradients:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[K(\theta,T)\frac{d\Psi}{d\theta}\frac{\partial\theta}{\partial z}\right] + \frac{\partial}{\partial z}[K(\theta,T)] \qquad [1]$$

where θ is the volumetric water content (dimensionless), T is the temperature (°C), $K(\theta,T)$ is the hydraulic conductivity as a function of the volumetric water content and temperature (L t^{-1}), Ψ is the pressure head (L), and t is time.

Advection and conduction of heat through subsurface sediments can be described by the advection–dispersion equation (Kipp, 1987; Nasser and Horton, 1992a, 1992b), which describes the time rate of change of energy storage in the sediments and water due to spatial changes in heat transport via advection, conduction, and dispersion. In one dimension, this can be written as

$$\frac{\partial[\theta C_w + (1-\phi)C_S]}{\partial t} = \frac{\partial}{\partial t}\left\{[k_T(\theta)]\frac{\partial T}{\partial z}\right\} + \frac{\partial}{\partial z}\left(\theta C_w D_h\frac{\partial T}{\partial z}\right) - \frac{\partial}{\partial z}(\theta C_w q T) + q_s C_w T^* \qquad [2]$$

where ϕ is the sediment porosity (dimensionless), D_h is the thermomechanical dispersion tensor (L^2 t^{-1}), q is the water flux (L t^{-1}), q_s is the water flux added as an external or internal source (L t^{-1}), C_s is the volumetric heat capacity of the dry solids (energy per unit volume per change in temperature, E L^{-3} T^{-1}), C_w is the volumetric heat capacity of liquid water (E L^{-3} T^{-1}), $k_T(\theta)$ is the thermal conductivity of the sediment as a function of water content (E/TLt), and T^* is the temperature of the fluid source (°C). These coupled one-dimensional equations are used in this investigation to describe heat transport and transient water flow through a variably saturated medium.

MATERIALS AND METHODS

A constant flux infiltration experiment was conducted in the field to examine the feasibility of using downhole temperature monitoring to quantify the infiltration rate. The subsurface temperature was measured using a thermistor that was lowered within a cased borehole and using thermocouples buried in an adjacent borehole. A numerical model of coupled water flow and heat transport was used to analyze the results.

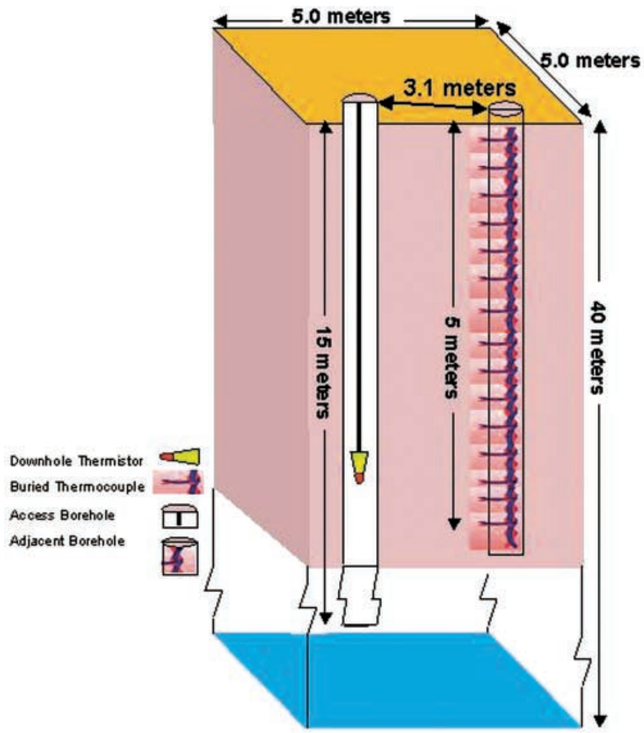

Fig. 1. Illustration of 5 by 5 m infiltration area with an access borehole and an adjacent borehole with buried thermocouples.

Field Experimental Design

From 9 to 12 Nov. 2001, a 71-h constant flux infiltration experiment was conducted at a site on the east bank of the Santa Cruz River in Tucson, AZ. The subsurface material comprises unconsolidated to poorly consolidated interbedded gravel, sand, and clayey sand. The water table was 40 m below ground surface. The borehole used for temperature logging was drilled to the 15.4-m depth within a 5 by 5 m bermed area 16 wk before the beginning of the experiment (Fig. 1). The hole was cased with 5-cm (2-inch) PVC tubing, capped at its base, and the annulus was backfilled with a mix of cuttings and coarse sand. Thermocouples were installed in an adjacent borehole within the bermed area at 0.30-m intervals to a depth of 4.5 m and backfilled with native material. The thermocouples were sampled every 10 s using a Campbell Scientific (Logan, UT) CR10 datalogger.[1] Temperatures were averaged over a 10-min interval to reduce scatter.

During the experiment, water was applied inside the bermed area through porous hoses at a constant rate of 0.34 m d^{-1} for 71 h with minimal ponding. Given that only minimal ponding occurred, this could be considered to represent the minimum flux associated with an ephemeral flow over the site sediments. The applied water was taken from a municipal water source, and the water temperature was measured periodically during the experiment. The average water temperature was 20°C. Hourly air temperature measurements were downloaded from the Tucson meteorological station at the Campus Agricultural Center, which is approximately 1.2 km from the site. No precipitation occurred during the experiment.

Downhole Thermistor Construction and Calibration

For this study, a Fenwal (Fenwal Electronics, Inc., Pawtucket, RI) iso-curve glass-probe thermistor was attached to

[1] Use of trade names is for identification purposes only does not constitute endorsement by the U.S. Geological Survey.

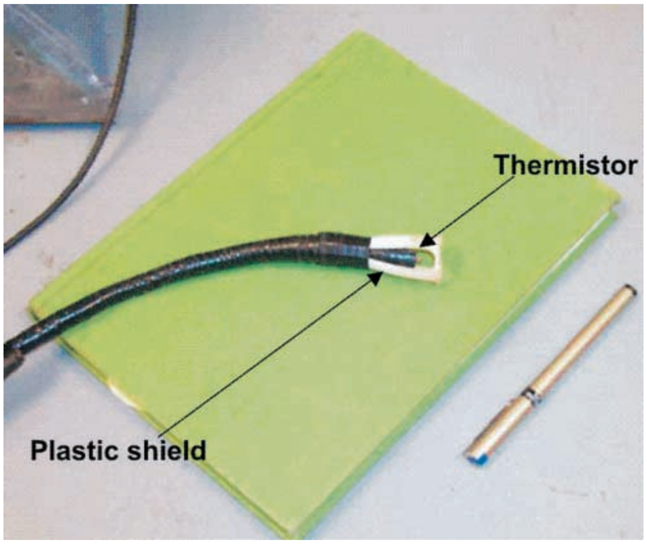

Fig. 2. Hermetically sealed thermistor encapsulated in glass and a protective plastic shield.

a breast reel containing 600 m of insulated four-conductor shielded cable. A winch was used to lower and raise the thermistor in the borehole. Two wires in the four-conductor shielded cable were used to apply a current while the remaining two were used to measure the voltage drop across the thermistor. The hermetically sealed thermistor was encapsulated in glass and protected from impact with the borehole wall by a plastic shield that was 2.5 cm in diameter (Fig. 2). The thermistor was soldered to the cable, and the connection was encased in heat-shrink tubing to prevent moisture from entering.

For calibration, the thermistor and a YSI (YSI Temperature, Dayton, OH) precision thermometer were submerged in deionized water in an insulated Nalgene container. Paired temperature and thermistor resistance measurements were made for the range of temperatures, from 4 to 36°C, which is expected to encompass the minimum and maximum downhole temperatures encountered during the field study. The resistance and temperature measurements were fitted with the nonlinear relation presented by (Steinhart and Hart, 1968)

$$T^{-1} = A + B\log R + C(\log R)^3 \qquad [3]$$

where T is temperature (K); R is resistance (Ω); and A, B, and C are empirical curve-fitting constants. TABLECURVE (Systat Software, Inc., Richmond, CA) was used to determine the values of the fitting constants. Paired measurements collected during 11 calibrations during a 12-mo period (Fig. 3) show that the temperature measurements were repeatable and that the instrument shows no appreciable drift. The regression correlation coefficient (R^2) was >0.99 for all of the calibrations. The accuracy of the temperature measurements is within 0.1°C for the entire calibration temperature range.

Downhole Temperature Profiling

The temperature profile was measured within the cased borehole at 0.61-m intervals from 0.13 to 15.34 m beneath the ground surface. At each measurement depth, the electrical resistance of a thermistor suspended in the borehole was measured once every minute. When the change in measured resistance was <10.2 Ω (equivalent to a temperature change of approximately 0.05°C) between consecutive readings, the thermistor was considered to be in thermal equilibrium with its surroundings. The manufacturer-reported time constant of the probe in air is 22 s. The average equilibration time for

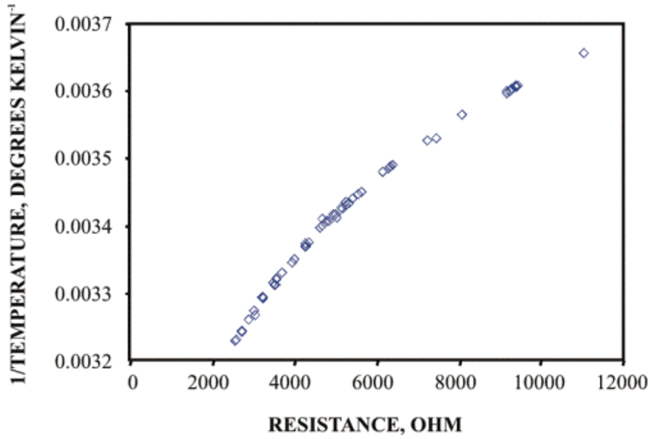

Fig. 3. Paired measurements of the temperature of a water bath and thermistor resistance from 11 calibration experiments performed during a 12-mo period.

each measurement point was 9 min. A minimum of five measurements were taken at each depth. When the thermistor reached equilibrium, the resistance was recorded and the thermistor was lowered to the next measurement depth. To lessen the disturbance of the air column in the cased borehole, the access borehole was plugged at the surface with fabric approximately 5 min before making the first temperature measurement. The temperature profile was logged twice: 13 h before water application and 114.5 h after infiltration began (43.5 h after infiltration ceased).

RESULTS AND DISCUSSION

Because the downhole thermistor was not in direct contact with the soil, there was uncertainty about whether the temperature measured within the borehole was representative of the thermal conditions in the surrounding medium. Direct comparison of the temperature profiles recorded with the downhole thermistor with those measured with buried thermocouples in the adjacent borehole (Fig. 4) showed that the downhole measurements were within 1.25°C of the buried thermocouple measurements for all measurements made below the 1-m depth (horizontal dashed line on Fig. 4). (Note that a second string of thermocouples, buried 1 m away, shows highly repeatable measurements for all depths except 0.6 m, suggesting that the downward temperature deflection at 0.6 m is due to an error in the operation or calibration of the buried thermocouple at this depth.) Shallower measurements made with the downhole and buried instruments do not agree as well, possibly because of differences between radiative and conductive heating of the casing and of the soil. Therefore only those borehole temperatures measured below 1 m depth are used for further analysis.

There is a distinct change in the temperature profile as a result of infiltration (Fig. 5). The near surface temperature is reduced by as much as 11°C. From 4 to 10 m depth, warmer water has been displaced downward, causing local heating by as much as 2°C. In general, the measured changes in temperature are much larger than the difference between the temperatures measured with downhole and buried probes, supporting the use of either instrument type for tracking water movement.

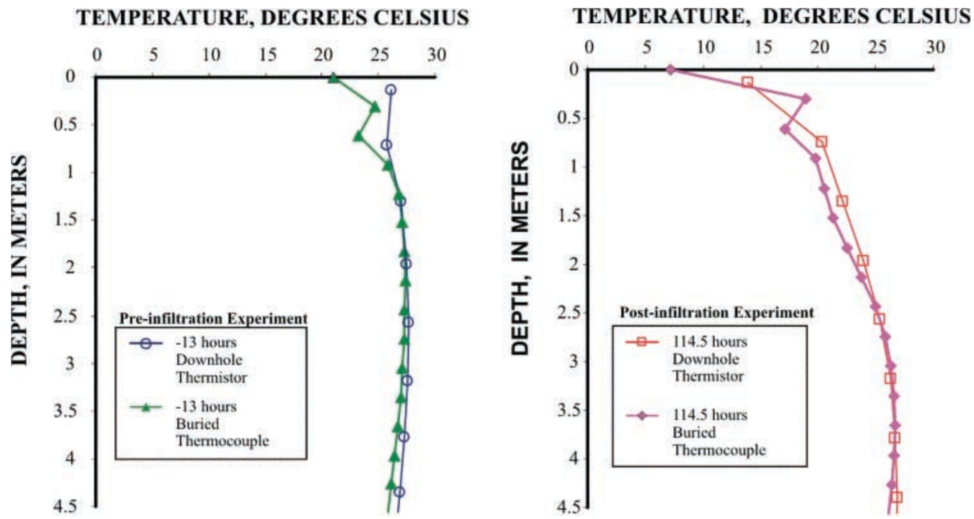

Fig. 4. A comparison of temperature profiles measured with downhole thermistor and buried thermocouples. Profiles are labeled with the elapsed time since the beginning of infiltration.

Coupled Water Flow and Heat Transport Numerical Forward Model

VS2DI (Healy and Ronan, 1996) is an interactive numerical model that iteratively solves the coupled, one- or two-dimensional water flow and heat transport equations, presented above as Eq. [1] and [2]. To represent the constant flux infiltration experiment, a one-dimensional homogenous domain was constructed that was 30 m deep. The bottom water flow and heat transport boundaries were defined as zero flux on the basis of borehole ground penetrating radar and neutron probe measurements made at the site that showed that the wetting front did not reach 15 m depth during the experiment; therefore, the boundary at 30 m depth was considered to have zero water flux. A constant water flux was applied at the upper boundary during infiltration, and zero flux was applied for all other times. The purpose of the investigation was to determine whether the infiltration flux beneath an ephemeral stream could be determined using data that would be readily available at a field site. Diurnal variations in the surface temperature were not included in the model. Average daily air tem-

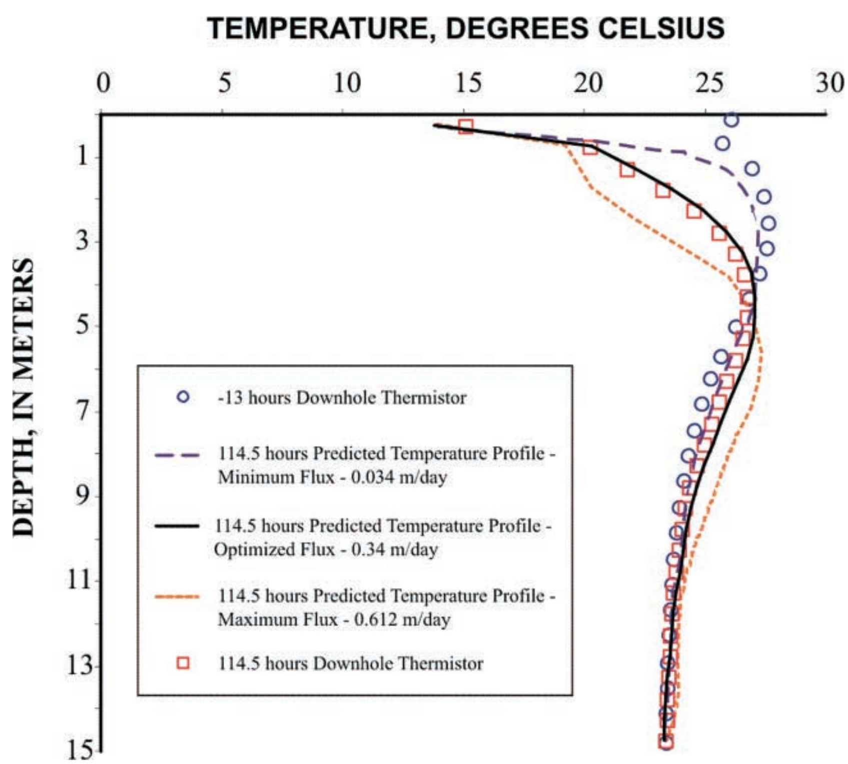

Fig. 5. Comparison of the predicted temperature profile for a constant infiltration rate of 0.34 m d^{-1} with temperatures measured using a downhole thermistor during the constant flux infiltration experiment. Profiles are labeled with the elapsed time since the beginning of infiltration. The initial temperature profile measured before infiltration is shown for comparison.

Table 1. Hydraulic, transport, and thermal properties and initial and boundary conditions used in the base case numerical simulation of the infiltration experiment. The range of realistic values is included, where appropriate. Those values listed as NA were not varied. The optimized property value, based on a sensitivity analysis, is shown for each parameter.

Model variables	Units	Initial Value	Optimized	Maximum	Minimum	Source
		Hydraulic properties				
Saturated hydraulic conductivity	$m\ d^{-1}$	5	3.5	35	0.5	Hoffmann et al., 2002; Hillel, 1998; Stephens, 1996; Jury et al., 1991
Porosity	$m^3\ m^{-3}$	0.375	0.41	0.46	0.31	Lappala et al., 1987; Prill et al., 1965
Specific storage	$1\ m^{-1}$	0	NA	NA	NA	Prill et al., 1965
Residual water content	$m^3\ m^{-3}$	0.1	NA	NA	NA	Lappala et al., 1987; Prill et al., 1965
van Genuchten α	$1\ m^{-1}$	4.31	2.59	9.05	2.59	Stephens, 1996; Prill et al., 1965
van Genuchten n	–	3.1	3.1	5.2	2.2	Stephens, 1996; Prill et al., 1965
		Thermal properties				
Heat capacity of dry soils (E + 06)	$J\ m^{-3}\ °C^{-1}$	2.5	2.5	4.23	0.986	Hoffmann et al., 2002; Hillel, 1998; Jury et al., 1991
Heat capacity of water (E + 06)	$J\ m^{-3}\ °C^{-1}$	4.18	NA	NA	NA	Hillel, 1998
Thermal conductivity at full saturation	$W\ m^{-1}\ °C^{-1}$	1.79	3.76	17.9	0.18	Hoffmann et al., 2002; Hillel, 1998; Jury et al., 1991
Thermal conductivity at residual saturation	$W\ m^{-1}\ °C^{-1}$	1.64	2.09	14.9	0.19	Hoffmann et al., 2002; Hillel, 1998; Jury et al., 1991
Longitudinal dispersivity	m	0.01	NA	NA	NA	VS2DI default
Transverse dispersivity	m	0.01	NA	NA	NA	VS2DI default
		Boundary and initial conditions				
Flux	$m\ d^{-1}$	0.34	0.34	0.612	0.034	Measured
Temperature of applied water	°C	20	20	30	8	Measured
Initial water content	$m^3\ m^{-3}$	0.22	0.22	0.31	0.13	Measured

peratures of 26°C before infiltration and 14°C after infiltration were assigned as the constant-temperature upper-boundary condition. A constant temperature of 20°C was used to represent the average applied water temperature during infiltration. A constant initial water content was assigned throughout the subsurface. The downhole temperature profile measured before infiltration was used to define the initial temperature profile. The domain was discretized into 81 equally spaced vertical increments, with a cell height of 50 cm. Twenty-six observation points were placed in the shallowest 15 m of the domain to coincide with depths at which the borehole temperature was measured.

A base case model was formed using estimates of 15 variables (Table 1). These include hydraulic, transport, and thermal properties, as well as boundary and initial conditions. The estimated initial volumetric water content was based on the average water content measured with borehole ground penetrating radar during previous field work. The specific heat capacity of dry soils, thermal conductivity of water sediment at full saturation, and thermal conductivity of the sediment at residual water content were based on measurements made on core samples of similar sediments collected within the Tucson Basin (Hoffmann et al., 2002). The base case saturated hydraulic conductivity was based on previously reported values for well-sorted sand (Jury et al., 1991; Hillel, 1998; Stephens, 1996). The porosity, specific storage, residual water content, longitudinal dispersivity, transverse dispersivity, and the van Genuchten parameters (α and n) were default values defined using a database within VS2DI for a medium sand (Lappala et al., 1987; Prill et al., 1965).

Determining the Infiltration Flux by Sequential Optimization

For an upper boundary condition representing constant flux at a rate equal to the measured infiltration of 0.34 m d^{-1}, the temperature profile predicted by the base case numerical model agrees well with the temperatures measured with the downhole thermistor (Fig. 5). However, given that coupled heat transport and water flow depends on many independent properties, the good agreement between the predicted and measured temperatures does not ensure that all of the hydraulic and thermal properties used in the model are correct or that the model can be used to determine the infiltration flux.

To examine the utility of the method for determining infiltration flux (q_s in Eq. [2]), a sensitivity analysis was conducted using VS2DI to determine the relative sensitivity of the resulting temperature profile to the values of the hydraulic and thermal properties, and of the boundary and initial conditions. Specifically, the following 10 model variables were varied independently from the base case values: saturated hydraulic conductivity, porosity, van Genuchten parameters α and n, initial volumetric water content, specific heat capacity of dry soils, thermal conductivity at full saturation, thermal conductivity at residual moisture content, infiltration flux, and average temperature of the infiltrating water. While holding all other variable values the same, each variable was increased and decreased over a range that extended from 0.1 to 10 times the base case value, to the extent that the variable values remained within the reported range of reasonable values for each property. The ranges of values used are listed in Table 1. The RMSE between the measured temperature profile and the modeled profile was computed for each deviation from the base case model.

The results of the sensitivity analysis (Fig. 6) are presented to compare directly the sensitivity of the model to each property. The error is not sensitive to the van Genuchten parameters α and n, the initial volumetric water content, the thermal conductivity at full saturation or at residual water content, or the saturated hydraulic conductivity. For clarity of presentation, only some of

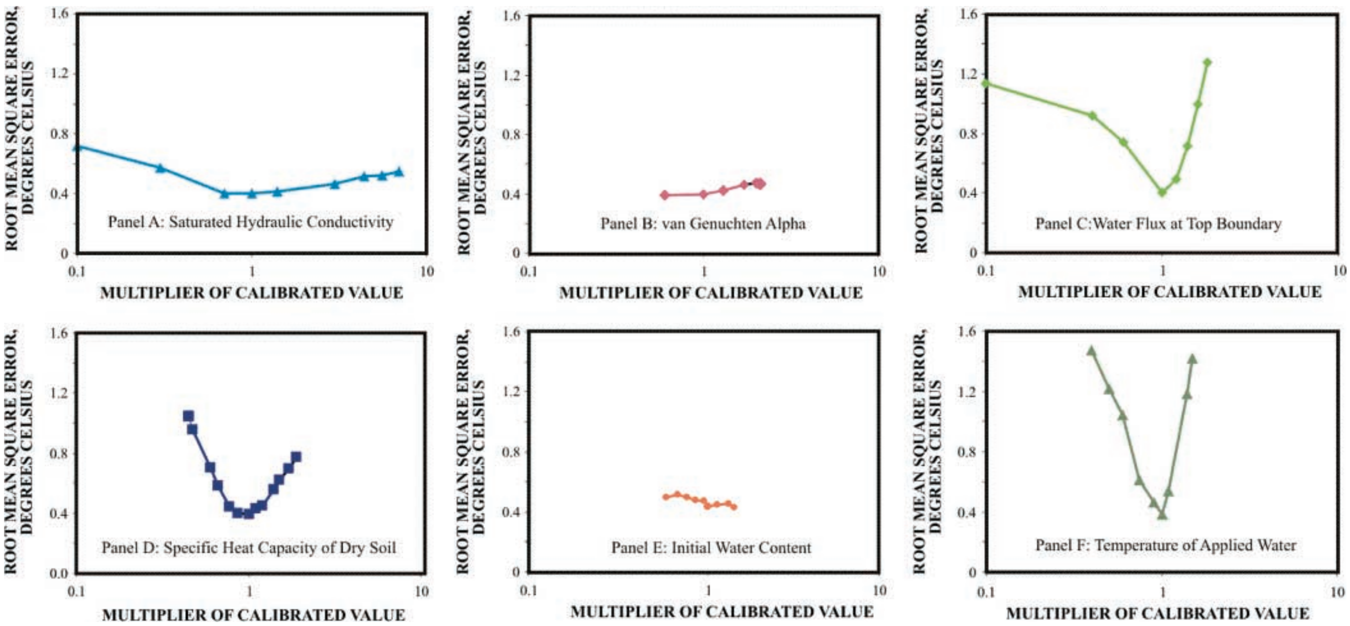

Fig. 6. Sensitivity of the RMSE of the modeled temperature profile 114.5 h after the beginning of infiltration to select hydraulic and thermal properties and boundary and initial conditions. The remaining parameters (van Genuchten's *n*, initial water content, thermal conductivity at full saturation, and residual water content) showed very low sensitivity and are not displayed here.

these results are shown. Furthermore, reasonable values of porosity were restricted to a narrow range, limiting their impacts on the resulting temperature profile. In contrast, the model is highly sensitive to the specific heat capacity of the dry soils (Panel C), the water flux (Panel D), and the average temperature of the infiltrating water (Panel F). The minimum errors are found for a flux equal to the known applied flux of 0.34 m d^{-1} and for an average temperature of the applied water equal to the measured value of 20°C. This result suggests that the flux and the temperature of the water can be defined accurately through a sequential optimization such as that used here. The high sensitivity of the model to the specific heat capacity of the dry soil suggests that this variable also can be defined using the method presented here, although there is no directly measured value available for comparison with the optimized value. Finally, the low sensitivity of the error to the other model variables suggests that a reasonable estimation of the infiltration flux can be determined even if values of these soil properties and initial conditions are known only within an order of magnitude.

The use of a constant surface temperature during water application is a simplification of the actual, time varying surface temperature caused by direct solar heating. To ensure that this simplification does not affect the accuracy of the infiltration flux estimate adversely, the model was run using temperatures measured with the shallowest buried thermocouples to define the surface temperature boundary condition. This analysis showed that the use of these temperatures did not affect the predicted temperature profiles below the 1-m depth (results not shown).

CONCLUSIONS

Vertical temperature profiles measured in a PVC-cased borehole were shown to be useful for inferring the infiltration rate during a constant flux infiltration experiment. A sensitivity analysis performed using a coupled water flow and heat transport numerical model showed that the temperature profile measured after infiltration ceased was most sensitive to the average temperature of the infiltrating water, the water flux, and the specific heat capacity of dry soil. The sensitivity to these properties was high enough to allow for a simple, sequential optimization approach to accurately determine the average temperature of the infiltrating water, the water flux, and the specific heat capacity of dry soil by numerical inversion of temperature measurements. The predicted temperature profile is less sensitive to other hydraulic and thermal properties and the initial water content, suggesting that the infiltration flux can be determined accurately even if these soil properties and initial conditions are not well defined. Because of the long equilibration times necessary to achieve thermal stability at each measurement point, each profile required approximately 4 h of continuous measurement. As a result, this method is not recommended for monitoring short-duration events. Rather, the results suggest that deep temperature profiles collected in existing, PVC-cased boreholes before and after an ephemeral flow event could be used to infer the rate of infiltration during streamflow. These measurements could be a useful complement to shallow streambed temperature methods developed by Constantz et al. (2001), allowing for better quantification of the contribution of streambed infiltration to basin-scale recharge.

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