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Estimating flow and flux of ground water discharge using water temperature and velocity

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Abstract

The nature of ground water discharge to a stream has important implications for nearby ground water flow, especially with respect to contaminant transport and well-head protection. Measurements of ground water discharge were accomplished in this study using (1) differences between current meter measurements, (2) stream temperature surveys combined with streamflow estimates, and (3) heat transport modeling of measured temperature gradients below the streambed. The first two techniques produced an area-averaged estimate of ground water flow, while the last produced a point estimate of ground water flux. Point measurements differed from area-averaged methods by 1 or 2 orders of magnitude. We hypothesize that discharge to the study creek is spatially heterogeneous, and is dominated by springs and seeps. Thermal gradient measurement did not quantify these local sources of stream inflow. Point measurements of inflow from temperature gradients or seepage meters, therefore, may not represent ground water inflow in some streams. Stream temperature and streamflow surveys were combined using a simple heat-balance to yield a higher-resolution estimate of streamflow than could have practically been obtained with current meters alone. This approach has potential as a cost-effective method of quantifying ground water discharge in streams where stream inflow is highly heterogeneous.

Keywords: Temperature surveys; Temperature methods; Ground water modeling; Hydrology; Streamflow; Rivers and streams

1. Introduction

Ground water discharge to streams strongly influences the character of regional subsurface flow and can be the dominant parameter in the calibration of numerical ground water flow models. Estimates of ground water discharge are particularly important in the development of regional ground water models in which head data may be sparse or unevenly distributed. Methods of quantifying interactions between streams and ground water have yet to be firmly established, however. In general, two approaches have been taken to this problem. In the first, water *flux* through the streambed is measured, e.g. by using seepage meters (Isiorho and Meyer, 1999) or inferring specific discharge through hydraulic or temperature gradients (Constantz, 1998). In the second, water *flow* is measured in the stream,

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e.g. using a current meter or permanent stream gage (Christensen et al., 1998). Each of these approaches have their merits. Flux measurements can quantify the spatial variability of ground water discharge, while flow measurements result in a lumped discharge value for an entire stream or stream reach. Flux measurements do not necessarily result in mass balance, while flow measurements will produce mass balance if surface-water and ground water components of the streamflow are properly separated.

In this article, we compare flux- and flow-based measurements of ground water discharge and discuss the implications for using these data to calibrate numerical models. Our purpose is to highlight the conceptual difficulties underlying the integration of very different estimates of ground water surface-water interactions. The methods of discharge estimation discussed here are (1) differential stream discharge measurements, (2) stream temperature survey, and (3) vertical streambed temperature gradient modeling. The study site is a 40 km long creek that traverses a stratified-drift aquifer in southwestern New York State.

Certainly the most common method of assessing ground water discharge to streams is to separate a stream hydrograph into baseflow and quickflow and then assume that baseflow represents ground water discharge (Winter, 1999). If hydrographs are not available, current meter measurements may be taken under low-flow conditions, and the difference in flow between measurement points along the stream attributed to ground water discharge. Unfortunately, multiple stream gages are rarely installed along a single stream and current meter measurements are time consuming and limited by stream geometry. The limited number of flow measurements along a stream severely constrains the ability to characterize spatially varying inflow to the stream. This leads to a lack of resolution in ground water flow estimation, as there is no a priori reason to assume that specific discharge should be evenly distributed along an entire stream.

Temperature surveys are expected to indicate area of ground water discharge at times of the year when surface- and ground water temperatures are in contrast. If the contribution of heat from ground water is significant in comparison to heat exchange at the stream surface or with the streambed, then stream temperature can be used a rough proxy for relative ground water discharge in the stream. Temperature characterization of ground water discharge is attractive because temperature surveys can be conducted very quickly and cheaply compared with seepage meters or streamflow measurements. They have been the subject of increased interest recently because economical waterproof temperature dataloggers have become widely available. In addition, space-based thermal sensors (e.g. ASTER, MODIS) that have recently come on line hold potential for observing stream surface temperatures. If stream surface temperatures are representative of water column and/or streambed temperatures, it may eventually be possible to remotely quantify ground water discharge to streams.

The primary disadvantage of the thermal survey is that it does not directly quantify water movement. Temperature surveys can locate losing and gaining portions of streams, for example, but an estimate of ground water flux cannot be obtained without additional information (Silliman and Booth, 1993). Vertical temperature profiles can be used to produce quantitative estimates of ground water discharge or recharge. The general process is to measure water temperature at multiple points below the streambed. The temperature gradient is assumed to be a result of the coupled heat transport processes of water advection and heat conduction. Heat transport models are calibrated to temperature history and the water advection parameter (specific discharge) derived from the calibrated model. Heat transport can be modeled in response to seasonal or diurnal variations in surface temperature. For example, Silliman and Booth (1993) surveyed a creek in northern Indiana in which portions of the creek were gaining (ground water discharge) and others were losing (ground-water recharge). During the summer months, ground water temperature was cooler than stream temperature in discharge areas. Ground water temperature and stream temperature were about the same temperature in recharge areas. In a subsequent study, the downward movement of water in recharge areas was quantified by modeling the propagation of diurnal temperature fluctuations from the water column to the deeper sediments (Silliman et al., 1995). The authors found that downward specific discharge rates as low as 0.03 cm/s could be determined from this method.

Although thermal gradient measurements are quantitative, they are subject to a number of assumptions concerning unknown boundary conditions and thermal characteristics of the streambed and aquifer material. More importantly, even when the discharge value is accurate it still represents a point-estimate of flux. Combining, comparing, or ‘fusing’ point and flux measurements of ground water discharge can be problematic. One must assume that the ground water discharge through the streambed is uniform, which is usually not the case. Variations in hydraulic conductivity of the streambed results in uneven discharge and flow geometry may lead to greater discharge near the banks than the thalweg. A given point measurement of ground water flux to a stream may not be representative of the stream gain as a whole.

In this study, differential stream discharge was obtained by repeated current meter measurements at specified stations along the stream (Fig. 1). Streambed temperature surveys were performed repeatedly at numerous locations along the entire 40 km length of the stream. Finally, vertical streambed gradient measurements were conducted by installing temperature and pressure probes at multiple depths below the stream. These gradient measurements were modeled using a one-dimensional heat transport equation to obtain estimates of upward or downward water flux through the streambed. The summer in which these studies were conducted was unusually dry; very little rain fell during July and August of 2001. Streamflow was relatively constant over the study period and minor tributaries did not flow. As a result, we could reasonably assume that streamflow was dominated by baseflow and therefore represented a direct measurement of ground water discharge.

2. Description of research site

The characterization of ground water discharge was performed along a 40 km stretch of Ischua Creek between the towns of Machias and Hinsdale, Cattaraugus County, New York State. The 300 km^2 Ischua Creek Watershed is defined here by Ischua Creek discharge at the confluence of Oil Creek and Ischua Creek, which merge to form Olean Creek. The topography of the watershed is characterized by gently rolling hills with wide valleys in the north to steep and narrow valleys in the south, all of which have been shaped by Pleistocene glaciation. The climate of the study area is typical of western New York State with cold, snowy winters and cool, wet summers. The average annual temperature for the region is 7.7 °C (46 °F) (National Weather Service Office, 2003). Temperatures in January range from an average low of −11 °C (12 °F) to an average high of −2.2 °C (28 °F) and a July range of 11.7 °C (53 °F) to 24.4 °C (76 °F) (National Weather Service Office, 2003). Precipitation for the watershed averages 100 cm annually, with about 244 cm of snowfall.

The Ischua Valley floor is underlain by stratified glacial drift, mantled by modern alluvial deposits and bordered by small deltas of tributary streams. The study area has experienced at least two major continental glacial advances during Pleistocene time, which are responsible for most of the surficial deposits throughout the region (Tesmer, 1975). These two advances are represented by the Olean (30,000–40,000 years B.P.) and Kent (19,000 years B.P.) drift sheets. Stagnant ice margins occupying the northern end of the valley resulted in deposition of a heterogeneous group of sediments ranging from permeable sands and gravels to relatively impermeable till. The surficial sediments range from a mixture of gravel, sand, and till in the head waters, to primarily glaciolacustrine silts from in the vicinity of the town of Machias, to well-sorted sand and gravel toward narrow southern end of the Ischua Valley. Aquifers are recharged through a combination of precipitation infiltration, stream losses, and inflow from upland slopes.

Ischua Creek has its headwaters in flat hummocky terrain west of the town of Machias, about 50 km southeast of Buffalo, New York. Channels developing in the wetlands coalesce into a slow moving stream which meanders until the last 7 km of its run, where it straightens and broadens. The creek bed is everywhere rocky, and is covered with organic sediments toward the north. In general, the creek is wide and shallow, with depths ranging approximately from 0.5 to 1.5 m. The creek meanders along some sections and small oxbow lakes can be seen. During baseflow conditions, significant surface turbulence is observed

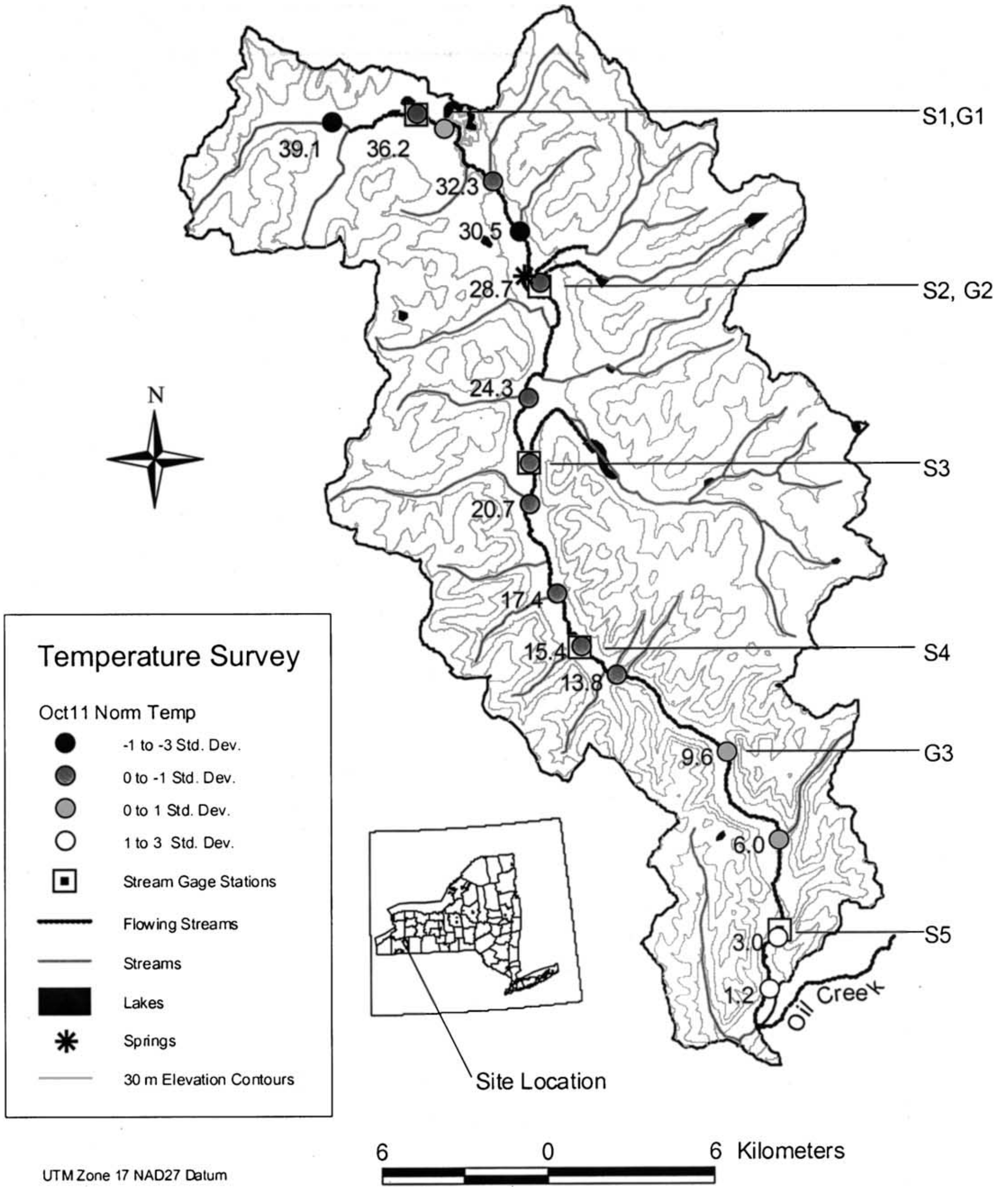

Fig. 1. Map of the Ischua Creek Drainage Basin. Streambed temperatures taken on October 11, 2001 are classified as number of standard deviations above or below the mean for that day. Distances upstream (km) from Oil Creek are shown left of the temperature stations. Locations of stream gage stations (e.g. S1) and temperature gradient measurement stations (e.g. G1) are also shown.

only near temporary obstructions such as beaver dams. Ischua Creek was under baseflow conditions through most of the study period of July–October, 2001. July, 2001 was the driest July in 68 years as recorded in the Buffalo office of the National Weather Service. About one quarter and one-half the normal amount of rainfall fell in July and August, respectively.

3. Description of measurement methods

3.1. Differential streamflow measurements

Streamflow was measured repeatedly at five stations along Ischua Creek at intervals of 5–10 km (see Fig. 1). As permanent gages were not available, streamflow was measured entirely through the use of current meters. A Marsh-McBirney Model 2000 electromagnetic current meter was used to measure streamflow velocity. Ischua Creek is slow moving over most of its course so measurements sites were selected in areas of channel narrowing. At station S1, flow was measured in a culvert. Stream discharge measurements along with their estimated error are given in Table 1. Errors were calculated according to the method described by Herschy (1985).

3.2. Temperature survey measurements

Temperature surveys were conducted by walking sections of the creek with a digital temperature probe (YSI 46 Tele-Thermometer or Oakton Digital 91300). At each measurement station, temperatures were collected at three depths: (a) just above streambed, (b) just below the streambed (2–3 cm), and (c) at the maximum depth that the probe could penetrate. The maximum depth of temperature measurement rarely exceeded 10 cm because of the rocky nature of the streambed. These measurements were repeated at three positions approximately a meter apart, along the stream thalweg. Temperatures reported here are the average of three measurements below the streambed.

Table 1
Selected stream discharge (Q) measurements (see Fig. 1 for station locations)

Streamflow	8/10/2001		8/20/2001		8/24/2001	
Station	Q (m^3/s)	Est err (%)	Q (m^3/s)	Est err (%)	Q (m^3/s)	Est err (%)
S1	0.04	10	0.10	10	0.04	11
S2	0.21	9	0.30	8	0.24	8
S3	0.33	10	0.41	10	0.36	9
S4	0.78	6	0.39	6	0.52	10
S5	0.41	7	0.45	9	0.40	11

Est err is the percent error estimated for measurement of Q.

Because only very shallow streambed measurements could be acquired using the digital thermometer, they are considered to be representative of stream temperature rather than streambed or ground water temperature. Even in reaches of the stream where ground water flowed most strongly to the stream, shallow streambed temperature was dominated by stream temperature. This relationship was confirmed with the temperature profile data collected at the gradient stations, along with one-dimensional heat transport modeling (see below).

3.3. Temperature gradient measurements

Temperature gradients in the streambed were measured at three points along the creek (Fig. 1), by installing nested piezometers in which temperature and head were recorded using Solinst Levelogger Model 3001 dataloggers. Nests consisted of either two or three piezometers, one of which measured stream stage and water temperature (± 0.3 °C) just above the bed. We were limited in our instrumentation by the five available dataloggers but attempted to sample gradients in gaining and losing portions of the creek. Significant head gradients could not be detected over the 0.5–1 m vertical distances at which heads were measured.

Piezometers were constructed of $1\frac{1}{4}$ in. (3.2 cm) nominal diameter schedule 80 PVC pipe fitted with a specially machined aluminum drive point. Piezometers were driven into the streambed using a fence-post driver. Cobble-sized stones in the streambed limited penetration depth to less than a meter at most locations. Ports were drilled through the pipe wall to allow head to be measured at a specific depth. The datalogger was hung on a nylon wire inside the pipe at a known depth near the port. Baffles constructed of rolled foam padding were installed above the datalogger in the pipe to prevent temperature-driven convection in the water column.

4. Interpretation of data from individual methods

We will interpret separately the results of the three methods of ground water discharge estimates, although the three methods are not entirely independent. Ground water discharging to a stream is diluted by surface-water flow. Consequently, stream temperature surveys must be combined with streamflow information to infer ground water discharge. Estimation of ground water discharge from temperature surveys, therefore, cannot be decoupled from estimation of ground water discharge using incremental streamflow measurements.

4.1. Incremental streamflow

Table 1 lists the flow rates estimated at current meter stations shown in Fig. 1. Additional measurements were taken earlier in the summer, but some were acquired at stations that were later abandoned due to low flow velocities. As noted previously, these measurements are considered to be representative of baseflow because the summer was unusually dry. A rain gage located at Bower's Hollow, approximately 20 km from the center of the study site, recorded only 32 mm of precipitation from August 1 to August 24, 2001. More than half of this precipitation (18 mm) occurred between August 7 and August 10, so that the streamflow measurement taken on August 10 is more influenced by quickflow than the other two measurements. The total flow in the creek section as measured at S5, however, did not change dramatically over the study period (Table 1). Because the season was so dry, most of the tributaries did not flow during the study period. Only three tributary streams had significant flow (Fig. 1). These tributaries enter Ischua Creek between stations S2 and S3, so potentially could affect baseflow estimates in that reach. The flow in these tributaries was too small to measure, but were estimated to be less than 10% of the incremental streamflow measured for the reach.

Fig. 2 depicts incremental stream gain along reaches of Ischua Creek. Incremental stream gain was calculated by taking the difference between streamflows measured at each successive station. Errors bars were calculated by adding error from the individual flow measurements (Table 1). Note that stream gain varies little in the uppermost two stream reaches (Site 2-1, Site 3-2) during the study period. Ground water discharge in the two downstream reaches varies considerably over time, however. After significant rainfall (August 10, 2003), the reach

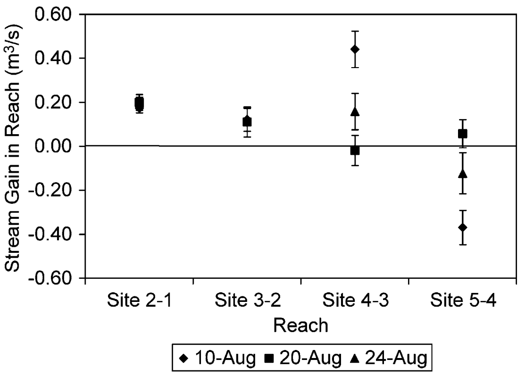

Fig. 2. Ground water discharge measured as difference in flow rate between two current meter stations.

between S3 and S4 is strongly gaining and the reach between S4 and S5 strongly losing. During drier periods (August 20, 2003), these same reaches may switch from losing to gaining and vice versa. Although these measurements clearly show differences in discharge behavior over time and space, the resolution is not sufficient to determine whether the ground water discharge occurs at a point or is distributed along the creek.

4.2. Temperature surveys

Temperature surveys can indicate areas of ground water and stream-water interactions, but correlating temperature changes to ground water discharge measurements requires ancillary data. Consider a simple heat balance on a stream reach (Fig. 3). The heat flow in and out of the reach include advection via stream-water flow, advection via ground water discharge or recharge, conduction through the streambed, and sensible and latent heat exchange with the atmosphere. Based upon the small mean temperature differences measured between the stream water and the deep streambed and the generally upward flow direction of ground water, it seems reasonable to neglect streambed heat conductance. Dividing the stream into reaches defined by temperature survey point results in an incremental balance as depicted in Fig. 3, where Q_i is streamflow rate leaving reach, i, and T is the stream temperature in reach, i. Q_{g_i} and T_g are the inflow of ground water and temperature of ground water, respectively. The temperature of the ground water is assumed to be

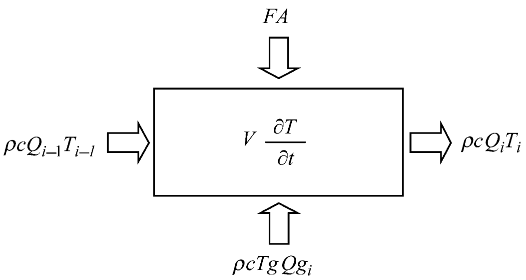

Fig. 3. Heat balance for reach, i, in a stream.

that of the deep ground water, even though the temperature may increase as it rises upward through the streambed. The rationale for this assumption is that warming of the ground water comes from the stream water and, therefore, warming represents a net heat loss or gain from the stream. The density and heat capacity of water are denoted by ρ and c, respectively. The temperature of ground water is assumed to be constant throughout the stream. The surface heat flux, F, represents the sum of all heat fluxes across the stream surface area, A, that are considered to be primarily temperature independent. The amount of latent and sensible heat transfer, for example, is expected to be a function primarily of solar radiation, humidity, and wind speed. The heat flux is assumed to be constant along the stream.

This simple heat balance leads to the equation

$$\rho c Q_i T_i = \rho c Q_{i-1} T_{i-1} + \rho c Qg_i T_g + FA, \qquad (1)$$

where under baseflow conditions

$$Q_i = Q_{i-1} + Qg_i. \qquad (2)$$

Substituting Eq. (2) into Eq. (1) results in the following expression for Q_i :

$$Q_i = \frac{\rho c Q_{i-1}(T_{i-1} - T_g) + FA}{\rho c (T_i - T_g)}. \qquad (3)$$

Thus, stream temperature is a function of the ground water discharge rate, the difference in stream-water and ground water temperature, streamflow, and additional heat gains and losses through the stream surface. To predict ground water discharge, one must model the coupled heat and mass balance in the stream. The level of sophistication of the stream heat and mass balance model depends upon the complexity of the system and the accuracy and level of detail required for the estimate of ground water discharge. Our view is that because most ground water models are calibrated to a single flow measurement along a stream, even a simple stream heat and mass balance model such as (3) will provide useful information to a ground water model. This is particularly true when few measurements of head are available to constrain variations in head gradients and hydraulic conductivity in surficial aquifer systems.

Eq. (3) was fit to current meter flow measurements by assuming that the temperature of the ground water was everywhere constant, and varying the value of the surface heat flux, F, until the best fit was achieved (measured as the sum of squared errors between measured and predicted streamflow). The ground water temperature was assumed to be equal to the average annual air temperature ($T_g = 8$ °C). Modeled flow rate was forced to be equal to the measured flow rate at Site 1. This heat dilution at reach $I = 1$ is necessary to initiate the series in Eq. (3). Fig. 4 represents a comparison of modeled streamflow based upon temperatures collected on October 11, to streamflow measured on August 24, where the fitted value of the heat flux was approximately 35 W/m^2. The modeled streamflow was generally able to predict gaining and losing sections of the stream.

It is important to note that streamflow and stream temperatures were not measured on the same day. The model is only reasonable if the stream is under similar baseflow conditions on both days. As an illustration, a model in which baseflow conditions were perceived to

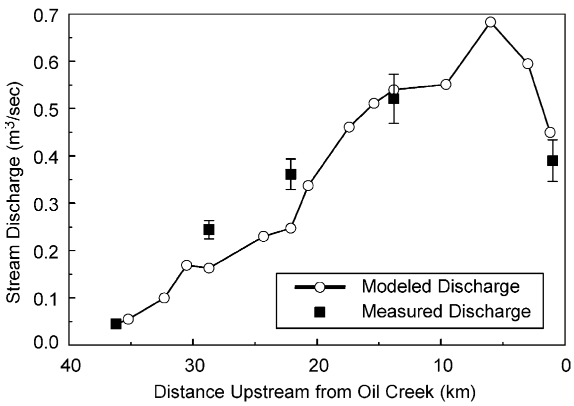

Fig. 4. Modeled discharge based upon streambed temperatures measured October 11, 2001, compared to measured discharge collected August 24, 2001. A value of $F = 35$ W/m^2 provided the best fit between measured and modeled (Eq. (3)) streamflow.

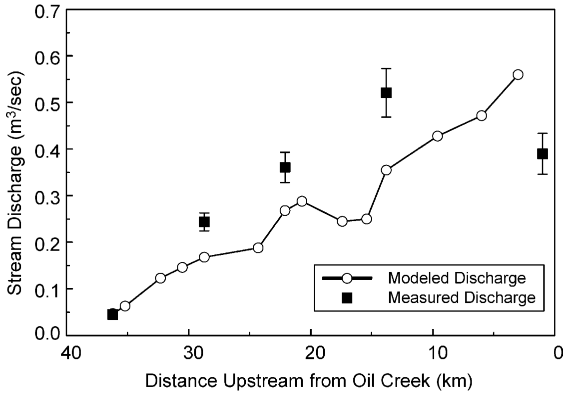

Fig. 5. Modeled discharge based upon streambed temperatures measured October 11, 2001, compared to measured discharge collected September 6, 2001. A value of $F = 53\ \mathrm{W/m^2}$ provided the best fit between measured and modeled (Eq. (3)) streamflow.

be similar during temperature and streamflow measurements (Fig. 4) can be compared to the model in which baseflow conditions were perceived to be different (Fig. 5). Fig. 5 compares stream discharge modeled from a temperature survey collected on September 6 ($F = 53\ \mathrm{W/m^2}$) and actual stream discharge measured on August 24. Even though measurements were taken only 2 weeks apart and streamflow was generally dominated by baseflow on both dates, minor rainfall events on August 26 (8 mm) and August 30 (9 mm) increased the quickflow component of streamflow on September 6. Note that the September 6 temperature survey predicts generally gaining stream 0–15 km upstream of Oil Creek but the flow survey on August 24 shows a generally losing stream over that reach. Obviously, the best approach to reducing these discrepancies is to conduct flow and temperature surveys on the same day.

4.3. Temperature gradient

The propagation of diurnal heat fluxes from a stream into the streambed can be used to estimate ground water discharge. Silliman et al. (1995) used this method, for example, to estimate ground water discharge to gaining portions of a creek in Indiana. The one-dimensional heat transport equation is (Stallman, 1965)

$$\frac{K_e}{\rho' c'}\frac{\partial^2 T}{\partial x} - \frac{n\rho_\mathrm{w} c_\mathrm{w}}{\rho' c'} v_x \frac{\partial T}{\partial x} = \frac{\partial T}{\partial t}, \tag{4}$$

where T is temperature, t is time, x is depth below the streambed, v_x is the steady-state average linear velocity of the water in the sediments (positive in the downward direction), K_e is the thermal conductivity of the saturated sediments, n is the porosity of the sediments, ρ' and c' are the density and heat capacity of the saturated sediment, respectively, and ρ_w and c_w are the density and heat capacity of the water, respectively. We may solve Eq. (4) subject to the initial and boundary conditions

$$T(x,0) = 0, \tag{5a}$$

$$T(0,t) = \Delta T_\mathrm{w}, \tag{5b}$$

$$T(x \to \infty, t) = 0, \tag{5c}$$

where ΔT_w is the perturbation of water temperature at the streambed.

Solution of (4) subject to (5a)–(5c) yields

$$T(x,t) = \frac{\Delta T_\mathrm{w}}{2}\left[\mathrm{erfc}\left(\frac{x - Zt}{2\sqrt{Dt}}\right) + \exp\left(\frac{Zx}{D}\right)\mathrm{erfc}\left(\frac{x + Zt}{2\sqrt{Dt}}\right)\right], \tag{6a}$$

where

$$Z = n\beta v_x, \tag{6b}$$

$$\beta = \frac{\rho_\mathrm{w} c_\mathrm{w}}{\rho' c'}, \tag{6c}$$

$$D = \frac{K_e}{\rho' c'}, \tag{6d}$$

where erfc is the complementary error function, and D is known as the thermal diffusivity. Under experimental conditions, however, temperature at position x is not known at all time so (6a)–(6d) must be written for incremental time steps that correspond to the sampling rate of temperature by the dataloggers. This may be accomplished by the superposition of (6a)–(6d), where the initial temperature perturbation, ΔTwi, corresponds the incremental change in stream water temperature over each time interval, $t_i - t_{i-1}$:

$$\Delta T_\mathrm{wi} = \Delta T_\mathrm{w}(t_i) - \Delta T_\mathrm{w}(t_{i-1}). \tag{7}$$

As we are interested in actual temperature, rather than temperature perturbation, a reference temperature, T_0, must be specified. The incremental heat transport equation is then

$$T_i(\tau) = T_0 + \sum T_i(x, \tau), \tag{8}$$

where τ is the time elapsed since the initial temperature reading at t_0, i.e. $\tau = t_i - t_0$. Initial condition (5a) implies that the initial sediment temperature distribution, T_0, is constant with depth.

The 'semi-infinite' solution has been interpreted by Silliman et al. (1995) to imply that the Eqs. (6a)–(6d) can be used only to predict downward ground water movement, i.e. temperature below losing streams. This argument stems from condition (5b) which specifies that a perturbation in the temperature profile occurs at the streambed. It is important to note, however, that the same conditions can be applied via a probabilistic 'first-passage-time' construct, in which the equation calculates the probability density function of arrival times of a 'particle' introduced at $x = 0$, arriving at the point x, at time, t (Becker and Charbeneau, 2000). Under this interpretation, there is no reason why the equation cannot be used for upward or downward heat advection and transport. The solution does imply continuity of heat transfer between the streambed and the overlying water column, and requires that the temperature at depth is equal to the initial temperature distribution ((5a) and (5c)).

The solution is somewhat sensitive to the initial temperature, T_0. After a period of time, the influence of the choice initial temperature is reduced, but is never completely removed. Following the approach of Silliman et al. (1995), we chose to ignore the first 100 h of model results to avoid influence of this initial temperature distribution. The initial distribution was chosen through trial and error. The final choice of T_0 minimized the difference between modeled and measured temperatures over the entire modeled history. The model was insensitive to changes in initial ground water temperature within 0.5 °C.

Eqs. (6a)–(6d) were calculated in the spreadsheet MathCad, and fitted to the time-series temperature data from the three temperature gradient stations. Fits were first attempted manually, and then finalized using an automated Levenberg-Marquardt error minimization routine available in the spreadsheet. No reasonable fit could be found to the station G1. The assumption of a constant deep ground water

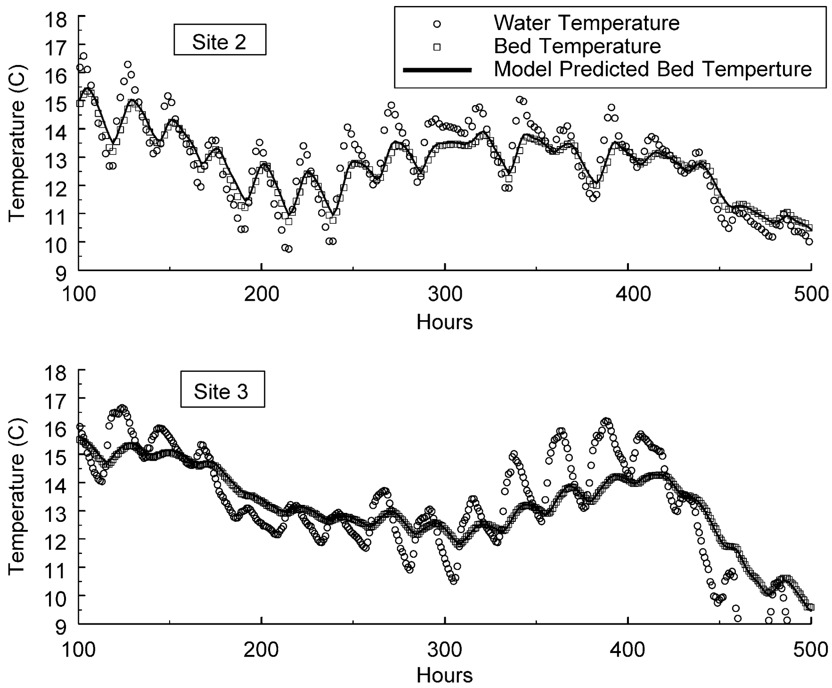

Fig. 6. Comparison of temperature 14 cm below the streambed to temperature modeled using a one-dimensional heat transport model and measured stream water temperature.

temperature (i.e. (5c)) seemed to be inappropriate at this site. Although the effect of diurnal temperature changes was observed at depth, the temperature followed surface-water temperature with considerable lag. We believe that ground water in this area is recharged very locally from a nearby moraine, which leads to fluctuations in 'deep' ground water temperatures over the time span of weeks. Unfortunately, this was the only site outfitted with temperature probes at three depths, instead of two. Due to the problems with temperature gradient station, G1, only the model fits of data from stations G2 and G3 are discussed here.

Temperature gradient data could be adequately represented (Fig. 6) where surface-water temperatures at sites G2 and G3 were used as the temperature perturbation (Eq. (4)). Although water flux (specific discharge) was small, there was a clear difference in the fit of the model to data when flow was downward, upward, or neutral. Confidence in the model fits was improved by the use of a large number of hourly sampling points (684 at site G2 and 414 at site G3). The parameter β was assumed to be 2, and porosity 0.3 after Silliman et al. (1995). Best fit of data from site G2 produced an estimate of water flux (specific discharge) of -0.05 cm/h (upward) and thermal diffusivity of 45 cm^2/h. Best fit of data from site G3 produced an estimate of water flux (specific discharge) of -0.03 cm/h (upward) and thermal diffusivity of 14 cm^2/h. The sensitivity of the model to advection at G2 was poor, however, because the best fits were achieved with a relatively high thermal diffusivity. This value (45 cm^2/h) is within the range of published values but is higher than most measured thermal diffusion rates.

5. Comparison of ground water discharge estimates

Incremental streamflows, temperature gradients, and temperature surveys represent fundamentally different indicators of ground water discharge to streams. Incremental streamflow measurement is the only method that is mass-conservative and, therefore, may be considered the most reliable approach. Streamflow differences are only practical over long stream reaches, however, and therefore, represent an integrated result. Temperature surveys also produce integrated measurements, but the distance of integration can be practically much smaller, due to the ease and rapidity of acquiring temperature measurements. The major drawback of temperature surveys is that, because ground water discharge is diluted by surface flow, temperature survey values do not produce discharge flux unless they are combined with a streamflow information. Temperature gradient measurements coupled with a heat and water flow transport model produce point measurements of ground water discharge flux. Because discharge measurements at a point may not be representative of a stream reach, however, vertical gradient discharge estimates do not ensure water mass balance along a stream reach.

Fig. 7 displays a comparison of the three methods. Line plots represent 'integrated' measurements (differential streamflow and temperature surveys) and the symbol plot represents the point measurement method (temperature gradient). All values are given in water flux (specific discharge). Flow rates obtained from the differential streamflow and temperature survey results were converted to flux by dividing by the estimated area of the creek bed. The area of the creek bed was obtained using a GIS program (ArcView, ESRI). Lengths were calculated along a digital line graph (DLG) representation of the creek which did not always track the frequently meandering Ischua Creek. Stream widths measured at temperature station position. A linear regression of widths versus

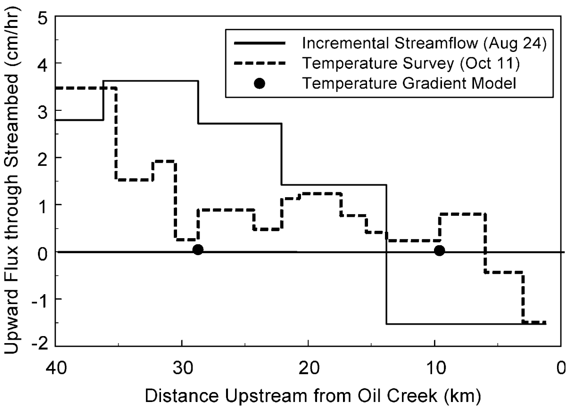

Fig. 7. Flux of ground water to Ischua Creek as measured by differential streamflow, temperature survey, and streambed temperature gradient. Note that the first two methods give integrated results, while streambed temperature gradient gives a point result.

stream distance was used to estimate stream widths at all points along the stream. The regression was performed to minimize bias of the mass balance due to choice of stream temperature locations.

Fig. 7 suggests that there is considerably more variation in the interaction of ground water and surface-water than is discerned by streamflow measurements alone. Although the differential streamflow and temperature survey models predicted the same trend in streamflow gain and loss, the temperature survey model simulated a highly variable stream–aquifer interaction over the scale less than about a kilometer. The greater variability in the temperature survey approach may result from the greater sampling density, or may be an artifact of the model assumptions. The heat balance (Eq. (3)) assumes that conductive exchange with the streambed is insignificant and that the water–atmosphere transfer is evenly distributed. Local variations in shading and wind speed may have resulted in a heterogeneous heat exchange which would be expressed as heterogeneous baseflow in the model. Monitoring of surface energy fluxes along the stream would be required to discriminate between ground water/surface-water and surface-water/atmosphere heat exchange.

The point measurements of ground water flux did not correspond well with the other measurements of ground water flow. At site G2, for example, modeling of the temperature profile estimated an upward flux of 0.05 cm/h, while the differential streamflow method estimated an upward flux of 3 cm/h (average of adjacent stream reaches), and the temperature survey model estimated an upward flux of 0.8 cm/h (average of adjacent stream reaches). At site G3, modeling of the temperature profile estimated an upward flux of 0.03 cm/h, while the differential streamflow method estimated a downward flux of 1.5 cm/h (average of adjacent stream reaches), and the temperature survey model estimated an upward flux of 0.7 cm/h (average of adjacent stream reaches).

Heat transport modeling of the temperature profile in both cases resulted in a much smaller flux than the differential streamflow or temperature survey methods. Although the parameters n and β were taken from the literature and could be in error, these parameters are expected to range within an order-of-magnitude and could not have produced such a large discrepancy between the gradient and other estimates of flux. It is conceivable that the gradient method produced a smaller flux because the temperature stations happened to be installed in areas of transition between gaining and losing portions of the stream, but that would be highly fortuitous. It is unlikely that meter-scale variability in the streambed could account for such large discrepancies between the temperature gradient and other measurement methods, because thermal diffusion would tend to smooth the temperatures laterally. We hypothesize, therefore, that bulk of ground water discharge to Ischua Creek is not diffuse, but occurs through buried springs or similar small-scale discharge features. This hypothesis is supported by the highly heterogeneous nature of the surficial geology in the area and the spatial variability of the stream temperatures. Neither incremental discharge or temperature gradients alone, could have indicated the spatial distribution of ground water discharge to the creek.

Conversion between integrated measurements (incremental discharge and temperature survey) and point measurements (thermal gradient) requires understanding of the local distribution of discharge flux. In streams dominated by spring recharge, significant correlation between integrated and point measurements is not expected. Even in those streams dominated by diffuse ground water discharge, discharge flux is not expected to be evenly distributed. Ground water discharge may vary across the stream section (shore versus thalweg), at meanders, and with the heterogeneity of the streambed hydraulic conductivity. Even if discharge is assumed perfectly even across the streambed, conversion between point and integrated discharge estimates requires measurements of the streambed area. This can be a painstaking process at the watershed scale.

Silliman et al. (1995) found reasonable agreement between the ground water discharge estimates from temperature gradient modeling and Darcy's Law estimates based upon nearby piezometers. Although Silliman et al. (1995) discuss the many limitations and possible pitfalls of this method, they concluded that the approach was appropriate, at minimum, as an order-of-magnitude estimate of diffuse flux through streambeds. We found, by contrast, that estimates of ground water flux using temperature gradient modeling differed from other estimates by over an order of magnitude. Because the one-dimensional heat-transport model provided an excellent representation of the diurnal temperature behavior in the streambed, we do not think this difference was due to errors in the temperature gradient measurement or modeling. We conclude, therefore, that discharge in Ischua Creek is fundamentally different than the creek studied by Silliman et al. (1995), in that it is dominated by point rather than diffuse ground water discharge. It appears that temperature gradient methods are much more useful when it can be determined that ground water discharge to a creek is primarily diffuse.

6. Conclusions

Three methods of determining ground water/surface-water exchange to stream were investigated for this article. Differences between streamflow measurements at adjacent stream reaches were used to determine net exchange along stream reaches. Temperature surveys were combined with streamflow measurements to determine net exchange between temperature measurement locations. Finally, the measured temperature gradient below the streambed was modeled with a one-dimensional heat transport model to derive water exchange rates. These methods represent very different approaches to obtaining the same information, and may therefore be considered complimentary. The vastly different nature of these approaches, however, makes the comparison or 'fusion' of these data sources inherently difficult. The most problematic difference with the comparison of these methods is that streamflow and temperature survey methods integrate water fluxes along a finite distance of the stream, while the temperature gradient method estimates flux at a single point. The impact of heterogeneities in hydraulic conductivity and hydraulic gradient along the stream have a very different influence on the results of these methods. In addition, to compare the estimates of ground water and surface-water interactions, fluxes must be converted to flow rates or vice versa, requiring assumptions concerning stream geometry and spatial distribution of fluxes. Similar problems would have been encountered had we used seepage meters rather than temperature gradients to measure discharge at a point in the streambed.

The field example demonstrates that comparisons between flux and flow measurements are not easily made, unless discharge to a stream is primarily diffuse. If a stream receives most of its baseflow via springs and other local discharge phenomenon, then flux measurements in the streambed (e.g. temperature gradients, seepage meters) cannot be expected to be a reasonable estimate of ground water discharge to a stream. In our study, estimates of ground water flux to the stream via the temperature gradient method was 1–2 orders-of-magnitude smaller than inflow determined through incremental current meter surveys. Stream temperature surveys showed strong variability along the course of the creek, which may have been due to local increases in ground water discharge. Thus, stream temperature surveys are potentially useful for determining the nature of ground water discharge to streams.

We combined stream temperature and streamflow surveys to yield a higher-resolution estimate of streamflow than could have been obtained practically with current meters. This approach, although subject to error, provides spatial information that could be important to ground water flow modeling studies. For example, if ground water discharge occurs primarily through local discharge phenomenon, then the near-stream ground water flow would be expected to be heterogeneous as well. This has important implications for contaminant transport modeling and well-head protection in near stream-environments. Given the low cost and ease of collecting stream temperature data, it seems worthwhile to collect stream temperature data whenever near-stream ground water studies are performed. If temperature appears to be highly variable, it may be worthwhile to quantify discharge variability with current meters, or a combination of current meter and temperature data in the manner suggested here.

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