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High resolution mapping of hyporheic fluxes using streambed temperatures: Recommendations and limitations

Dylan J. Irvine *, Laura K. Lautz

Department of Earth Sciences, Syracuse University, 204 Heroy Geology Laboratory, Syracuse, NY 13244, United States

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S U M M A R Y

Analytical solutions to the 1D heat transport equation can be used to derive point measurements of flux between surface water and groundwater from streambed temperature time series. Recent studies have used empirical relationships between measured flux and point-in-time observations of streambed temperatures to produce detailed plan view maps of flux from instantaneous temperature maps. Here, the accuracy of such flux maps, derived using streambed temperatures as a quantitative proxy, was assessed from synthetic streambed temperature data generated by numerical flow and transport simulations. The use of numerical simulations is advantageous because maps of flux from the temperature proxy method can be compared to known flux maps to quantify error. Empirical flux–temperature relationships are most accurate if developed from data collected when stream temperatures are at a maximum. The true relationship between flux and streambed temperature will generally be non-linear and well approximated as a cubic function, although linear relationships may be applied when data density is low. Intermediate fluxes (±1.0 m/day) returned by the temperature proxy method have errors typically less than ±0.1 m/day. Errors in estimated flux increase for strong upwelling (>1.0 m/day) or downwelling (<−1.0 m/day), although the direction of flux is still accurate.

1. Introduction

Detailed characterization of the spatial and temporal variability of water exchange between streams and groundwater is important for assessing stream health. For example, downwelling of stream water into the subsurface provides dissolved oxygen to organisms living in streambed materials (Boulton et al., 1998), and upwelling of groundwater helps to regulate stream temperatures (Arrigoni et al., 2008). Detailed characterization of spatial patterns of hyporheic flux can be challenging, particularly at the reach scale (Fleckenstein et al., 2010). Hydraulic methods such as seepage meters (Lee, 1977), hydraulic testing at mini-piezometers (e.g. Cardenas and Zlotnik, 2003) or temperature methods (Bredehoeft and Papadopulos, 1965; Hatch et al., 2006; Luce et al., 2013) yield point estimates of flux, but these point estimates can be difficult to integrate spatially. Differential stream gauging and tracer injection techniques can yield integrated measurements of reach-scale net and gross hyporheic flux (e.g. Payn et al., 2009), but only limited information on spatial patterns. None of the aforementioned methods are capable of readily providing detailed spatial maps of flux.

Conant (2004) proposed a method where an empirical relationship between mapped streambed temperatures and fluid flux is used to estimate fluid flux where only a point-in-time temperature is known. This approach generates detailed spatial maps of flux, from only a few direct flux measurements. Through the use of streambed temperature as a quantitative proxy, Conant (2004) was able to identify a range of flow behaviors, including the magnitude of upwelling and downwelling, and the location of preferential flow pathways. To develop the empirical relationship between streambed temperature and flux, Conant (2004) used hydraulic testing at mini-piezometers to determine hydraulic conductivities and hydraulic heads to then calculate flux using Darcy's law. A primary limitation of this approach is that hydraulic based measurements of flux are typically highly uncertain given the large uncertainty in measurements of hydraulic conductivity (e.g. Calver, 2001). Also, point-in-time temperature (T_p) and flux (q) measurements were not always measured at the same location.

An alternative and widely used approach to measure fluid flux in streambeds is 1D analytical heat transport modeling (e.g. Hatch et al., 2006). Although assumptions of 1D heat transport models are rarely met in field settings (e.g. vertical flow, homogeneous streambed materials, sinusoidal stream temperature), heat based flux estimates have been demonstrated to be reliable where flow fields are multi-dimensional (Lautz, 2010; Roshan et al., 2012;

* Corresponding author now at: Monash University, Australia.
E-mail address: dylan.irvine@monash.edu (D.J. Irvine).

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Cuthbert and Mackay, 2013), where streambeds are heterogeneous (Irvine et al., 2015), or where stream temperatures are non-sinusoidal (Lautz, 2010, 2012). Estimates of flux from 1D analytical heat models are ideally suited for shallow sensors (Roshan et al., 2012; Irvine et al., 2015). Uncertainties in flux estimates from temperature time series also occur due to uncertainties in thermal properties. Shanafield et al. (2011) show that these uncertainties can be minimized by reducing sensor spacing, and that errors induced from uncertainties in thermal properties are small for higher fluxes.

The method of using streambed temperature as a quantitative proxy for flux (from herein referred to as '*T*-proxy method') has been investigated in more recent work by Lautz and Ribaudo (2012) and Gordon et al. (2013), building on the initial ideas presented by Conant (2004). In their approach, point estimates of q are calculated from the amplitude ratio between two temperature time series, using equation 6a from Hatch et al. (2006). The use of temperature time series (T_s) to determine q is advantageous because q and T_p (i.e. a T_p selected from T_s) are measured at the same location, and heat based analytical solutions do not require knowledge of hydraulic conductivity, which can be highly uncertain. The empirical relationships between q and T_p (which we denote as rating curves) from Lautz and Ribaudo (2012), and Gordon et al. (2013) are presented in Fig. 1 (where negative values denote downwelling, and positive values denote upwelling). As with the method proposed by Conant (2004), the rating curve is used to estimate q where only T_p is known.

In their study, Lautz and Ribaudo (2012) generated their rating curve (Fig. 1a) from a linear fit, which was applied to six q–T_p pairs. Gordon et al. (2013) followed the same procedure as Lautz and Ribaudo (2012), using 13 q–T_p pairs to generate their rating curve (Fig. 1b), which was derived using a cubic fit of their data points.

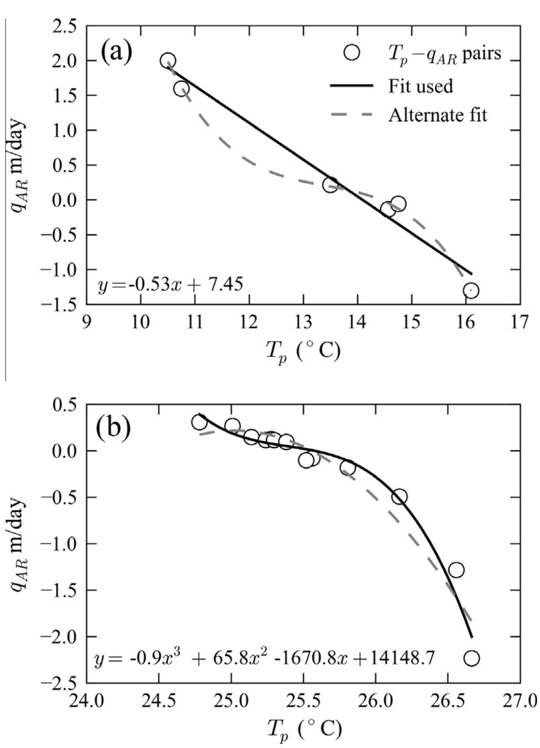

Fig. 1. Examples of rating curves of fluid flux (q) against point-in-time measurements of streambed temperature (T_p). Shown in (a) is rating curve from Lautz and Ribaudo (2012), and (b), Gordon et al. (2013). For both figures, the white circles denote the q–T_p pairs used to generate the rating curve, solid black line denotes the line of best fit (i.e. the rating curve) used in each publication, and gray dashed line represents a plausible alternative best fit.

Alternative fitting procedures are presented in Fig. 1, where a plausible cubic fit could also be applied to the data presented by Lautz and Ribaudo (2012). A reasonable quadratic fit could be applied to the Gordon et al. (2013) data.

It is difficult to determine the actual shape of the q–T_p relationship using field data. For example, the cubic fit applied in Fig. 1a could be a plausible rating curve in addition to the linear fit. The quadratic fit applied in Fig. 1b could also have been a reasonable alternative fitting procedure. A numerical model is an ideal tool to investigate the use of streambed temperatures as a quantitative proxy, and to determine the optimal fitted relationship between q and T_p because streambed temperatures can be generated by the model, and predicted fluxes can be compared against the known flux from the model.

The aims of this paper are to: (1) determine the best time of day to map temperature to produce the most accurate flux maps, (2) determine the optimal fit and shape of the rating curve, (3) provide guidance on the number of points required for rating curve development, and (4) present how the *T*-proxy method performs for a range of flow conditions and depths. These questions are explored through the use of synthetic streambed temperature time series generated using numerical simulations of flow and transport.

2. Methods

2.1. Heat model

Point estimates of flux were determined using the equation for amplitude ratio (A_r) from Hatch et al. (2006):

$$q=\frac{C}{C_w}\left(\frac{2\kappa_e}{\Delta z}\ln A_r+\sqrt{\frac{\alpha+v_{th}^2}{2}}\right), \tag{1}$$

$$\alpha=\sqrt{v_{th}^4+(8\pi\kappa_e/P)^2}, \tag{2}$$

where q is the vertical Darcy flux (m/s), C is the volumetric heat capacity of the saturated sediment (J/m^3/°C), C_w is the volumetric heat capacity of water (J/m^3/°C), κ_e is thermal diffusivity (m^2/s), A_r is the amplitude ratio of the temperature signals (calculated as $A_r = A_d/A_s$ where subscript s and d denote shallow and deep sensors respectively) (–), Δz is the spacing of the sensor pair (m), v_{th} is the velocity of the thermal front (m/s), α is calculated from Eq. (2), and P is the period of the temperature signal (i.e. 86 400 s). The effect of thermal dispersivity (β, m) was excluded from the analysis, and hence κ_e is a constant. The relevant parameters used in the analytical and numerical modeling in this study are shown in Table 1.

Hatch et al. (2006) also presented a method where q can be calculated from the phase shift ($\Delta\phi$, i.e. time lag) between the signals observed at two sensors. The A_r method is preferable because the $\Delta\phi$ method cannot determine flow direction (Hatch et al., 2006). The A_r method has been shown to be less prone to error compared to $\Delta\phi$ in both multi-dimensional flow fields (Lautz, 2010), and in heterogeneous streambeds (Irvine et al., 2015). The drawback of

Table 1
Parameters used in analytical and numerical modeling.

Model parameter (symbol)	Value	Unit
Period of stream temperature signal (P)	86 400	s
Sensor spacing (Δz)	0.1	m
Porosity (θ)	0.3	–
Volumetric heat capacity of saturated sediment (C)	2.83×10^6	J/m^3/°C
Volumetric heat capacity of water (C_w)	4.19×10^6	J/m^3/°C
Thermal diffusivity (κ_e)	4.92×10^{-7}	m^2/s
Thermal dispersivity (β)	0.0	m