Scientific Briefing: Quantifying streambed heat advection due to groundwater-surface water interactions

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Abstract

Stream thermal regimes are controlled by the interaction of external and internal energy fluxes with the water in the channel. Solar radiation is typically the dominant driver of stream water temperature, but streambed heat fluxes can be important in forested headwater streams. Past studies have presented seemingly disparate formulae for quantifying streambed heat advection due to upwelling groundwater. This note details the sources of the differences in these alternative formulations. The equations illustrate the difficulties of attempting to isolate the thermal influence of groundwater-surface water interactions and highlight future research opportunities.

Keywords

Groundwater temperature, stream temperature, thermal regimes, river temperature, gaining streams, energy budget

1. Introduction

The water temperature of streams and rivers influences the growth and distribution of aquatic organisms and thus exerts a major control on the health and complexity of lotic ecosystems (Poole and Berman, 2001; Caissie, 2006; Webb et al., 2008). Consequently, elucidating the fundamental drivers of stream temperature is essential for implementing effective fish management strategies (Huang et al. 2012), informing stream protection or restoration practices (Beechie and Bolton, 1999; Palmer et al., 2009), and predicting future habitat loss due to climate change or other water temperature stressors (e.g., Luce et al., 2014; MacDonald et al., 2014).

Stream thermal regimes are controlled by energy fluxes from above the water surface, within the channel, and across the streambed (Allen and Castillo, 2007). Stream surface heat fluxes have typically received far more attention than streambed heat fluxes (e.g., Maheu et al., 2014), as the former have been shown to typically dominate the latter, at least in rivers or unshaded streams (Evans et al., 1998; Sinokrot and Stefan, 1994). The influence of streambed heat fluxes on stream thermal regimes depends on many factors including the rate and direction of groundwater-surface water interaction, local climate, and the degree of stream shading (Story et al., 2003; Leach and Moore, 2011), and can be important in small headwater streams where shading attenuates the influence of surface heat fluxes and where groundwater tends to dominate stream discharge (Caissie, 2006). Also, highly localized streambed advective heat fluxes at discrete groundwater discharge points can generate in-stream thermal anomalies that are utilized as refugia by cold-water fishes during high temperature events (Kurylyk et al., 2015a).

Several studies have considered the influence of streambed heat fluxes in lower order stream systems (e.g., Hannah et al., 2004; Sridhar et al., 2004; Hebert et al., 2011; Caissie et al., 2014). However, such studies have employed two seemingly disparate approaches for quantifying the streambed advective heat flux. Several studies (e.g., Sridhar et al., 2004; Hebert et al., 2011) have proposed that the streambed advective heat flux in upwelling-dominated streams is proportional to the *difference* between the stream temperature and the groundwater temperature, while other studies (St-Hilaire et al., 2000; Caissie et al., 2014) have suggested that it is proportional to the groundwater temperature and independent of the stream temperature. These two approaches can yield advective heat fluxes that differ in both sign and magnitude, and thus the equations cannot both represent the same physical process.

The objective of this note is to reconcile these differing representations of the streambed advective heat flux by considering the heat and water budgets at the reach and point scales. We also raise some important questions that require further study to improve our ability to quantify and model streambed advective influences on stream temperature.

2. Heat budget formulation for a finite stream segment

For a stream that is well-mixed both vertically and transversely, stream temperature is constant at a given channel cross-section, but varies longitudinally along the stream. The energy balance of a vertically and transversely well-mixed stream segment (e.g., 300 m long, Sinokrot and Stefan, 1993; Moore et al., 2005) with no tributary inputs between an upstream boundary at x_i and a downstream boundary at $x_i + \Delta x$ can be expressed by equating the rate of change of thermal energy in the stream segment (left hand side, Eq. 1) to the sum of the energy fluxes entering/leaving the stream segment (e.g., Moore et al., 2005):

$$\rho C \Delta x \frac{\partial \overline{A(T - T_{dat})}}{\partial t} = J_i - J_{i+1} + \Delta x \overline{w Q_v} + \Delta x \rho C \overline{f_g (T_g - T_{dat})} \qquad (1)$$

where ρ is the density of water (assumed constant) (kg m^{-3}); C is the specific heat of water (J kg^{-1} °C^{-1}); A is the cross-sectional area of the channel (m^2); T is water temperature (°C); J is the longitudinal heat flux associated with advection and dispersion (W), where the subscripts i and $i + 1$ refer to values located at x_i and $x_i + \Delta x$, respectively; w is surface width of the channel (m); Δx is the stream segment length (m), Q_v is the net vertical energy exchange associated with radiation, sensible and latent heat across the water surface and the bed heat conduction and friction (W m^{-2}); f_g is the rate of groundwater discharge per unit length of channel (m^{2}s^{-1}); T_g is the groundwater temperature (°C); and T_{dat} is a reference temperature (thermal datum) for computing the thermal energy of the water (°C) (Saur and Anderson, 1956; Lee 1999). The reference temperature is typically set, either explicitly or implicitly, to 0°C, although this is arbitrary from a thermodynamic perspective.

The quantities A, T, J, Q_s, f_g and T_g in Eq. (1) can all vary with location x and in time. For simplicity of notation, the dependence on time will be implicit. Overbars indicate quantities that are averaged along the interval Δx. These are placed over the entire $\overline{A(T-T_{dat})}$ and $\overline{f_g(T-T_{dat})}$ terms to avoid issues with covariance. Figure 1 presents the geometrical configuration and mass balance for the channel cross-section and reach profile under consideration in Eqs. (1) to (6). A rectangular cross-section is presented, but alternative configurations are allowable for the equations presented herein.

Eq. (1) does not include the advective heat transport associated with hyporheic exchange, which is commonly neglected in many physically based stream temperature models. Shallow hyporheic exchanges with the subsurface can significantly attenuate surface water thermal variability on shorter (e.g., sub-daily) time scales (Arrigoni et al., 2008), but typically they do not influence mean temperature on longer time scales. Implications of only considering far-field groundwater discharge are addressed in the discussion.

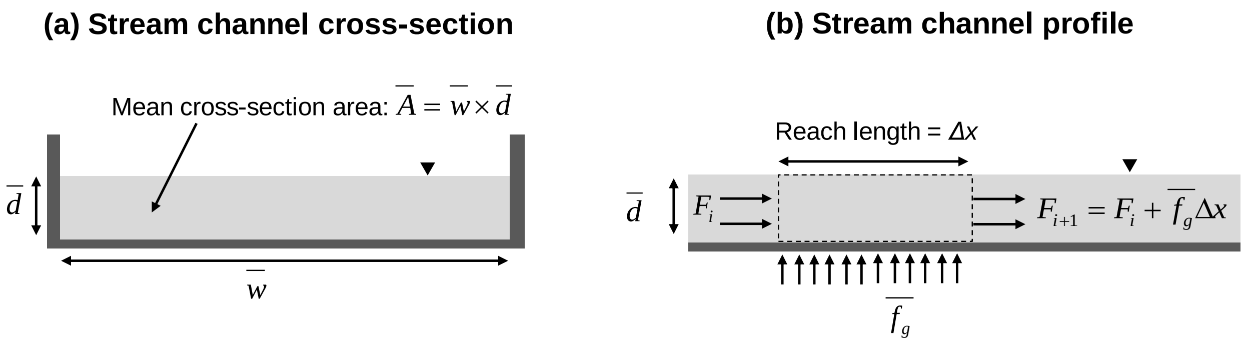

Figure 1: (a) Channel cross-section and (b) channel profile for a gaining stream. The mass balance yields the downstream flow for a particular gaining reach.

The advective component of the longitudinal stream water energy flux at location x_i is given by:

$$J_{adv,i} = \rho C F_i \left(T_i - T_{dat}\right) \tag{2}$$

where F_i and T_i are the stream discharge ($m^3 s^{-1}$) and water temperature (°C), respectively, at location x_i. The dispersive component is given by:

$$J_{disp,i} = -\rho C A_i D_i \frac{\partial T}{\partial x} \tag{3}$$

where A_i and D_i are the stream channel cross-sectional area (m^2) and the longitudinal thermal dispersion coefficient ($m^2\ s^{-1}$), respectively, at location x_i. The longitudinal dispersive flux is typically considered negligible for streams with natural thermal regimes (e.g., Sinokrot and Stefan, 1993; Polehn and Kinsel, 2000; Yearsley, 2009), and will be neglected from this point on for simplicity.

Inserting Eq. (2) into (1) and dividing by $\rho C\Delta x$ yields:

$$\frac{\partial \overline{A(T-T_{dat})}}{\partial t}=\frac{F_i(T_i-T_{dat})-F_{i+1}(T_{i+1}-T_{dat})}{\Delta x}+\frac{\overline{wQ_v}}{\rho C}+\overline{f_g(T_g-T_{dat})} \tag{4}$$

Note that the streambed heat advection component in Eq. (4) (last term) implies that the heat advection is independent of the surface water temperature as suggested by St-Hilaire et al. (2000), Caissie et al. (2014), and others. Because the thermal energy and fluxes are measured with respect to the chosen thermal datum, the associated streambed advective heat flux when $T_g = T_{dat}$ is, by definition, zero. Thus, in this formulation, the magnitude of the streambed heat advection depends on the selection of the datum, whereas other vertical fluxes (e.g., shortwave radiation and streambed heat conduction) are independent of T_{dat}.

Moore et al. (2005) modified Eq. (4) for a finite stream reach with steady-state flow and obtained (using the same notation as above):

$$\frac{\partial \overline{T}}{\partial t}=\frac{F_i(T_i-T_{i+1})}{\Delta x\overline{wd}}+\frac{\overline{Q_v}}{\rho C\overline{d}}+\frac{\overline{f_g(T_g-T_{i+1})}}{\overline{wd}} \tag{5}$$

where $\overline{d}$ is the mean hydraulic depth (cross-sectional area divided by surface width) in the reach (m) (note there is an error in Eq. (11) of Moore et al. (2005): T_{us} should have been T_{ds}). Eq. (5) is algebraically obtained from Eq. (4) by assuming the cross-sectional area is temporally constant ($\overline{A}=\overline{wd}$) given the steady-state assumption and by noting that the difference between F_{i+1} and F_i is equal to $\overline{f_g}\,\Delta x$ via continuity (see Fig. 1b). Eq. (5) is independent of the choice of the thermal datum as this term cancels out for temperature differences. Note that Eq. (5) is only valid for a gaining reach.

The last term in Eq. (5) implies that streambed heat advection depends on both the groundwater temperature and the surface water temperature (e.g., Sridhar et al., 2004; Hebert et al., 2011). This term was intended by Moore et al. (2005) to represent the apparent sensible heat flux due to groundwater inflow, and is similar in form to the corresponding term in the point-scale energy balance derived later. However, Eq. (5) can alternatively be rearranged by recalling that $F_i = F_{i+1} - \overline{f_g}\Delta x$. In this format, the third term on the right-hand side includes the difference between T_g and the stream water temperature at the upstream end of the reach (T_i), with an associated change to the first term on the right-hand side:

$$\frac{\partial \overline{T}}{\partial t} = \frac{F_{i+1}\left(T_i - T_{i+1}\right)}{\Delta x \overline{wd}} + \frac{\overline{Q_v}}{\rho C \overline{d}} + \frac{\overline{f_g (T_g - T_i)}}{\overline{wd}}. \tag{6}$$

Thus, at the reach scale (i.e. for a stream segment Δx), it is difficult to isolate the thermal influence of groundwater discharge because the mass flux due to groundwater inflow also affects the longitudinal advection term. The upwelling mass flux contributes more thermal mass to the stream segment, but the sensible effect of this flux is difficult to assess without considering other parameters (F_{i+1}, T_{i+1}) in addition to the last term in Eq. (6).

3. Heat and water budget formulation for a point along a stream

Taking the limit of Eq. (4) as $\Delta x \to 0$ and applying the product rule for derivatives yields:

$$A\frac{\partial T}{\partial t} + \left(T - T_{dat}\right)\frac{\partial A}{\partial t} = -\left(T - T_{dat}\right)\frac{\partial F}{\partial x} - F\frac{\partial T}{\partial x} + \frac{wQ_v}{\rho C} + f_g\left(T_g - T_{dat}\right) \tag{7}$$

where A, T, F, w, Q_v, f_g, and T_g are evaluated at location x and time t. By continuity:

$$\frac{\partial F}{\partial x} + \frac{\partial A}{\partial t} = f_g \tag{8}$$

Combining (7) and (8), dividing through by *A*, and rearranging terms yields:

$$\frac{\partial T}{\partial t} = -v\frac{\partial T}{\partial x} + \frac{Q_v}{\rho C d} + \frac{f_g\left(T_g - T\right)}{wd} \tag{9}$$

where *v* is the cross-sectional average flow velocity in the channel (*F*/*A*) at location *x* and *d* is the hydraulic depth in the channel cross-section at location *x* (m). Note that T_{dat} cancels out as in the case of the reach formulation.

The third term on the right-hand side of Eq. (9) can be expressed in energy flow units as:

$$Q_{gw} = \frac{\rho C f_g\left(T_g - T\right)}{w} \tag{10}$$

Eq. (10) is commonly termed the heat flux associated with groundwater discharge. However, as previously noted, thermal energy storage and fluxes, like potential energy, should be expressed relative to a constant reference thermal state (Saur and Anderson, 1956; Lee 1999). Eq. (10) expresses the sensible heat flux associated with groundwater discharge relative to local stream temperature, which varies in both time and space. Hence, we advocate referring to this term as an *apparent* sensible heat flux.

Furthermore, Eq. (10) only represents the *local* (i.e., at a point within the stream), apparent sensible flux because the stream discharge *F*, which is retained in Eq. (7), is also a function of the upstream groundwater inflow, as are the channel area *A* and width *w*.

4. Challenges in specifying groundwater discharge temperature, T_g

In many studies, upwelling groundwater temperature has been assumed to be equal to mean annual air temperature plus a potential thermal offset up to 3 °C (e.g., Ficklin et al., 2012; McDonald et al., 2014), which is a reasonable approach for deeper groundwater temperature in a stable climate. However, in some catchments, discharge to the stream is dominated by shallow subsurface flow, especially during rain and snowmelt events. These shallow flow paths, which induce lateral heat advection to the stream, can be characterized by relatively dynamic temperatures that differ substantially from that of deeper groundwater (Leach and Moore, 2014).

Even for deeper regional groundwater discharging across the streambed, the temperature T_g assigned when applying Eqs. (1), (5), or (10) should be the temperature of groundwater as it enters the surface water column, which will, in general, be modified by bed heat conduction and possibly mixing with hyporheic water. The standard practice of computing streambed heat conduction while calculating streambed heat advection by assigning groundwater discharge a temperature based on deeper groundwater runs the risk of "double counting" the thermal effect of groundwater, given that streambed heat conduction is typically enhanced in areas of strong groundwater discharge (e.g., Story et al., 2003; Briggs et al., 2014; Caissie et al., 2014).

As an illustrative example, Figure 2 presents the mean annual air temperature (MAAT) and mean daily subsurface water temperatures at 30 cm, 150 cm, and 300 cm below the streambed of an upwelling-dominated zone within a small third order stream (Catamaran Brook) in New Brunswick, Canada (46° 52.7’ N, 66° 06.6’ W). See Caissie et al. (2014) for site description. Mean annual groundwater temperature (MAGT) at some depth below the streambed is typically higher than MAAT in seasonally snow-covered areas due to the insulating effect of winter snowpack (Kurylyk et al., 2013). The shaded area in Figure 2 denotes the likely range of MAGT at this site. During the summer, the temperature of the upwelling water at a depth of 30 cm is up to 8°C higher than the expected MAGT range at this site, and the deviation from MAGT would likely be greater at even shallower (e.g. 5 cm) depths. The upwelling temperature at this site may be influenced by hyporheic exchange, and the difference between upwelling groundwater temperature and deeper groundwater temperature may be less pronounced under very strong upwelling conditions. However, these data illustrate the limitations of assuming that the upwelling water temperature can be represented by the mean annual conditions deeper within the aquifer.

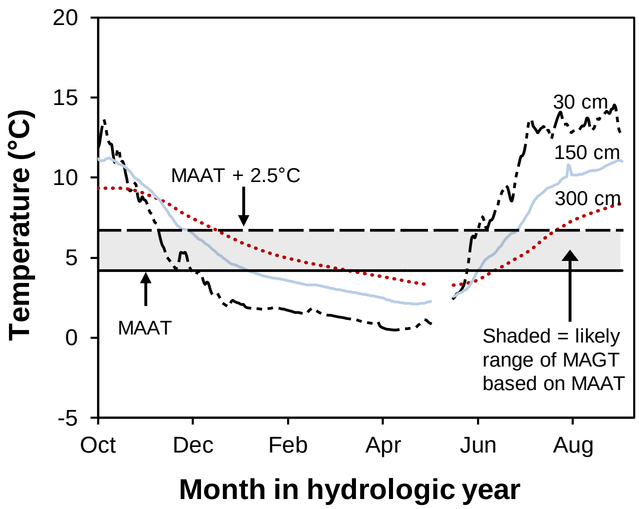

Figure 2: Mean annual air temperature (MAAT), expected range of mean annual groundwater temperature (MAGT) and mean daily shallow groundwater temperature measured at depths of 30 cm, 150 cm and 300 cm below the surface of the streambed for the 2002-2003 hydrologic year. Data were collected from an upwelling location in Catamaran Brook, New Brunswick, Canada (data source: D. Caissie, unpublished data).

The interactions between downward heat diffusion, upward heat advection, and the temperature of upwelling water can be investigated using simple analytical solutions (e.g., Stallman 1965; Luce et al., 2013; Briggs et al., 2014), or numerical models. Hence, stream temperature modellers could apply these methods to estimate upwelling temperature from deeper groundwater temperature, surface water temperature, and estimated upwelling rates. Alternatively, very shallow streambed temperature records could be employed to estimate upwelling temperature conditions (Caissie et al., 2014). Further research should address the limitations and/or suitability of these methods.

Finally, modeling studies projecting the impacts of future climate change on stream temperatures influenced by streambed heat advection should account for the influence of gradually warming groundwater (Kurylyk et al., 2015b). Multi-decadal groundwater warming rates should not be presumed to be identical to climate warming rates as complex interactions between the lower atmosphere, land surface, vadose zone, and aquifer can lag and dampen subsurface warming in comparison to atmospheric warming (Kurylyk et al., 2013). These noted challenges associated with specifying groundwater discharge temperature are particularly relevant for forested headwater streams where the thermal influence of groundwater inflow may be considerable.

5. Conclusion and summary

When working at a reach scale, it is difficult to isolate the thermal effects of the advective heat flux due to groundwater discharge because the associated groundwater mass flux also contributes to the stream longitudinal heat flux divergence. Hence, two distinct formulations may be obtained (Eqs. 5 and 6). The least ambiguous approach is to assess groundwater thermal influences in the context of Eq. (1), in which the advective heat flux associated with groundwater is proportional to the difference between groundwater temperature and a thermal datum. However, the advective flux calculated with this method cannot be directly used to infer the sign or magnitude of an associated change in stream temperature (i.e. a sensible influence). For example, if the upwelling groundwater temperature is less than the stream water temperature but greater than the thermal datum (e.g. 0^{o}C), this approach would produce a positive energy flux to the stream, but the resultant stream temperature change would typically be negative. This is the expected situation during summer, which has been the seasonal focus of most previous stream temperature research.

When working at a point scale, the local thermal effect of groundwater discharge can be expressed in terms of an apparent sensible heat flux (Eq. 10), which is proportional to the difference between groundwater temperature and local stream temperature. Thus, at the local scale, a positive advective flux calculated with this approach will result in an increase in the local stream water temperature.

Assigning an appropriate temperature for calculation of the advective flux due to upwelling groundwater is also challenging. In many cases, discharge to the stream occurs by flow paths that have thermal signatures that are more dynamic than those of deeper groundwater. Even where groundwater discharge originates from a deeper regional aquifer with a relatively stable temperature, the temperature of water discharging across the streambed will have been influenced by bed heat conduction and mixing with hyporheic water. Appropriate representation of groundwater discharge temperature is a key issue for application of stream temperature models in reaches with strong groundwater upwelling, and deserves further research attention.

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