Electronic supplement

S1. Brief description of the OTIS model

The One-Dimensional Transport with Inflow and Storage (OTIS) model (Runkel, 1998) was used to simulate the tracer breakthrough curves and thereby estimate the transient storage parameters: cross sectional area of the storage zone (A_s) and stream-storage exchange coefficient (α). These metrics are used to describe areas with slow flow compared to the main stream channel. This includes pools, eddies and the area beneath the streambed known as the hyporheic zone. OTIS solves the following equations, for solute in the stream channel and the storage zone, using a finite-difference method (Bencala, 1983):

$$\frac{\partial C}{\partial t}=\frac{-Q}{A}\frac{\partial C}{\partial x}+\frac{1}{A}\frac{\partial}{\partial x}\left(AD\frac{\partial C}{\partial x}\right)+\frac{Q_L}{A}\left(C_L-C\right)+\alpha\left(C_s-C\right) \quad \text{(S1-1)}$$

$$\frac{\partial C_s}{\partial t}=-\alpha\frac{A}{A_s}\left(C_s-C\right) \quad \text{(S1-2)}$$

where:

C is the solute concentration in the stream (mg L^{-1})
Q is the volumetric flow rate (m^3 s^{-1})
D is the dispersion coefficient (m^2 s^{-1})
A is the cross sectional area of the channel (m^2)
Q_L is the lateral volumetric inflow rate (m^3 s^{-1} m^{-1})
C_L is the solute concentration in the lateral inflow (mg L^{-1})
C_s is the solute concentration in the storage zone (mg L^{-1})
A_s is the cross-sectional area of the storage zone (m^2)
α is the stream-storage exchange coefficient (s^{-1})
t is time (s) and x is distance downstream (m)

Furthermore, a package known as OTIS-P allows for automated parameter estimation using observed concentration data. Full model documentation can be found in Runkel (1998).

The observed tracer breakthrough curves for both downstream locations were used, meaning that our model is divided into two sub-reaches. Discharge at the start of the main reach was obtained from GS1. Discharge at the end of the two sub-reaches was calculated using:

$$Q = Q_i C_i / C_p \quad \text{(S1-3)}$$

where Q is discharge (m^3 s^{-1}), Q_i is the injection pump rate (m^3 s^{-1}), C_i is the injectate concentration (g L^{-1}) and C_p is the plateau stream concentration (g L^{-1}) (Lautz & Siegel, 2007).

S2. Brief description of the HFULX model

HFLUX is a one-dimensional transient model that uses the finite difference method to calculate stream temperatures by solving the mass and energy balance equations (Glose et al., 2017):

$$\frac{\partial(AT_w)}{\partial t}+\frac{\partial(QT_w)}{\partial x}=Q_L T_L+\frac{Wq_t}{\rho_w c_w} \quad \text{(S2-1)}$$

Q is the volumetric flow rate (m^3 s^{-1})
A is the cross sectional area of the channel (m^2)
T_w is the stream temperature (°C)
q_t is the total energy flux to the stream per surface area (W m^{-2})

W is the width of the stream (m)
Q_L is the lateral volumetric inflow rate (m^3 s^{-1} m^{-1})
T_L is the temperature of the inflow (°C)
t is time (s) and x is distance downstream (m)
ρ_w is the density of water (kg m^{-3})
c_w is the specific heat capacity of water (J kg^{-1} °C^{-1})

The open-source code is written in MATLAB (MathWorks, 2020b), which allows for easy modification and selection of different methods to compute the relevant energy fluxes. Here we describe the simulation strategy and model modifications. A full list of model parameters can be found in Table 1.

Net Shortwave radiation (K*, W m^{-2}) was calculated by:

$$K^* = (1 - a_s)(1 - SF)\,K\downarrow \qquad \text{(S2-2)}$$

where SF is the shading factor (0 to 1) determined by calibration, $K\downarrow$ is measured shortwave radiation (W m^{-2}) and a_s is the albedo of the water surface assumed to be 0.05 (Magnusson, Jonas, & Kirchner, 2012).

Net longwave radiation (L*, W m^{-2}) was calculated by:

$$L^* = L_{atm} + L_{veg} + L\uparrow \qquad \text{(S2-3)}$$

where L_{atm} is atmospheric radiation (W m^{-2}) absorbed by water, L_{veg} is radiation from surrounding vegetation (W m^{-2}) absorbed by water, and $L\uparrow$ is radiation emitted by the stream surface (W m^{-2}). In most stream temperature studies atmospheric radiation is not measured but rather calculated from air temperature and cloud cover. To circumvent this, we modified the code to use measured values:

$$L_{atm} = 0.96\,L_{ws}\,VTS\,/\,VTS_{ws} \qquad \text{(S2-4)}$$

$$L_{veg} = 0.96\,(1 - VTS)\,0.98\,\sigma_{sb}\,(T_{air} + 273.15)^4 \qquad \text{(S2-5)}$$

$$L\uparrow = -0.96\,\sigma_{sb}\,(T_w + 273.15)^4 \qquad \text{(S2-6)}$$

where VTS is the view to sky coefficient (0 to 1) determined through calibration, VTS_{ws} is the calibrated view to sky coefficient at the weather station, T_{air} is the air temperature (°C), T_w is the water surface temperature (°C), σ_{sb} is the Stefan-Boltzmann constant (5.67×10^{-8} W m^{-2} K^{-4}) and L_{ws} is the measured incoming longwave radiation at the weather station (W m^{-2}). In equation S2-5 the emissivity of surrounding vegetation is assumed to be 0.98 (Oke, 1987, as cited in Pomeroy et al., 2009) and the vegetation surface temperature is set to be equal to the air temperature. In equation S2-6 emissivity of the water surface is set to 0.96 (Glose et al., 2017) and it is assumed that water surface temperature is equal to the bulk water temperature because flow is turbulent.

The latent heat flux of water vapour (q_e, W m^{-2}) was calculated using a mass transfer approach:

$$q_e = -\rho_w\,\Delta H_{vap}\,E \qquad \text{(S2-7)}$$

where ΔH_{vap} is the latent heat of vaporization (J kg^{-2}) which is a function of water temperature, and E (m s^{-1}) is evaporation rate estimated by:

$$E = (1.505{\times}10^{-8} + 1.6{\times}10^{-8}\,u)\,(e_s^{\,w} - e_a) \qquad \text{(S2-8)}$$

where u is the wind speed (m s^{-1}) at 2 m height, $e_s^{\,w}$ is the saturation vapour pressure of the evaporating surface (kPa) and e_a is the actual vapour pressure (kPa) at 2 m height. The default model wind function coefficients, reported by Dunne and Leopold (1978), are used for a 2 m height above the water surface. In reality, the weather station is located in a clearing on the hillslope (Figure 4) with measurements at 3.45 m height, below the height of the trees. Wind speed data were used without height correction. This is discussed in Section 5.2.

The sensible heat flux (q_h, W m^{-2}) was calculated using the Bowen ratio method:

$q_h = B_r\, q_e$ (S2-9)

where B_r is the Bowen ratio estimated by:

$B_r = 6.1\times10^{-4}\, P_a\, (T_w - T_{air}) / (e_s^{\,w} - e_a^{\,w})$ (S2-10)

where P_a is average atmospheric pressure estimated from elevation (kPa).

Streambed conduction (q_b, W m^{-2}) was calculated using:

$q_b = (P / W) \times [-k_{sed}\, (T_w - T_{sed}) / d]$ (S2-11)

where P is the wetted perimeter (m), W is the stream width (m), k_{sed} is the thermal conductivity of the sediment (W m^{-1} °C^{-1}), T_{sed} is the temperature of the sediment (°C) measured at P1 and d is the depth of measurement (m). In HFLUX the value of k_{sed} is based on the choice of sediment type: clay, sand, gravel, or cobbles. Gravel was chosen for this study so k_{sed} = 1.4 W m^{-1} °C^{-1}. The role of hyporheic exchange in the channel is addressed in Section 4.5.

Temperatures were computed using the Crank-Nicolson scheme for every minute and every meter along the study reach. However, the upstream boundary was set at temperature sensor T2 rather than T1 since complete mixing had not yet occurred below the spring system (Figure 5).

HFLUX uses a triangular channel shape and the Manning's equation to adjust stream dimensions with changing discharge:

$n_r / S^{1/2} = (A / P)^{2/3}\, (A / Q)$ (S2-12)

where n_r is the Manning roughness coefficient (m$^{-1/3}$ s) and S is the bottom slope of channel. The left-hand side of the equation is determined with measured discharge and channel dimensions. The model assumes that $n / S^{1/2}$ remains constant and adjusts A and P (wetted perimeter) as discharge changes. The channel shape was modified to a rectangle to reduce the flashiness of the modeled temperatures and better match observation data. The width of the stream was an order of magnitude greater than the depth, so we assume that wetted perimeter equals width. This means that

$n_r / S^{1/2} = (W\, d_{ch}^{\,5/3}) / Q$ (S2-13)

where d_{ch} is average channel depth (m). As a result, streambed conduction remains constant with changing discharge.

Furthermore, HFLUX does not include surface water inflows. Tributary inflow was therefore lumped with groundwater discharge as was performed in another study (Somers et al., 2016). The Western tributary was treated as a distinct water source with its own temperature (T12) and discharge (GS3) values. It contributed to the main channel over a sub-reach surrounding the inflow location. The model was modified to allow for changing inflow temperatures through time to include T12 measurements.