Environmental Modelling & Software 82 (2016) 295–307

Contents lists available at ScienceDirect

Environmental Modelling & Software

journal homepage: www.elsevier.com/locate/envsoft

A Reduced Parameter Stream Temperature Model (RPSTM) for basin-wide simulations

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ARTICLE INFO

Article history:
Received 23 February 2016
Accepted 13 April 2016
Available online 20 May 2016

Keywords:
Stream/water temperature modeling
Integrated modeling
Energy/heat balance
Climate change
Hydrology
Muskegon river

ABSTRACT

Water temperature is a crucial variable that shapes biological communities and controls rates of ecosystem processes in rivers. Fully parameterized heat balance models have been used to provide accurate estimates, but high parameterization costs make them difficult to apply at basin-wide scales. As parts of a collaborative modeling project to address future impacts of climate and land-use management on the Muskegon River, we developed a Reduced Parameter Stream Temperature Model (RPSTM), a mechanistic, spatially explicit but easier to parameterize model. Here we describe and test RPSTM's applicability by conducting a series of daily water temperature simulations (1985–2005). RPSTM performed well along the river network. The predictions were most sensitive to air temperature, depth, and solar radiation, but relatively insensitive to rates of surface runoff. This modeling approach is easily integrated into complex multi-modeling systems to evaluate effects of long-term changes in watershed hydrology, climate, and land management across river networks.

1. Introduction

Temperature in streams and rivers is an important water quality attribute that mediates many physical and biological processes in rivers (Caissie et al., 2005; Caissie, 2006; Wehrly et al., 2006). Given the controlling influence of stream temperature on aquatic ecosystems, models to predict changes in stream temperature have played an important role in understanding potential impacts of both climate and landuse change (Borman and Larson, 2003; Bartholow et al., 2005; Brown et al., 2006; Cadbury et al., 2008). These models fall into two general categories: empirical methods and physically-based heat balance models.

Empirical relationships are based on time-series and/or spatially varied observations of stream temperature and both local and catchment properties (Reckhow and Chapra, 1983). Water temperature in natural streams varies temporally following two primary patterns— a cyclic diurnal variation nested in a larger cyclic seasonal pattern (Sinokrot and Stefan, 1993). These are mainly responses to temporal variation in solar radiation and air temperature (Sinokrot and Stefan, 1994; Leopold, 2003; Allan, 2004). Spatially there are longitudinal (upstream-downstream) variations in the mode, amplitude and timing of temporal variations, associated with emerging thermal equilibria, and heat transport and loading patterns across the network. While they are often used to provide summary descriptions for a specific locale (LeBlanc et al., 1997; Baker et al., 2005), empirical approaches provide no basis for anticipating effects of future changes in catchment routing or climate. However, large changes in climate, landuse, and human consumption are already occurring, and they seem an inescapable aspect of our near future. This fact makes finding robust methods for predicting future river water temperatures more challenging but also more essential.

Recently, physically-based heat balance models have been receiving renewed interest (Bartholow et al., 2005; Chapra et al., 2006; Pelletier et al., 2006; Fujihara et al., 2008). These models can accommodate anticipated changes in the energy balance driven by changing environmental conditions (TVA, 1972; Crittenden, 1978; Sinokrot and Stefan, 1993; LeBlanc et al., 1997). Because they preserve individual heat flux terms, tracking the impacts of relatively complex environmental scenarios is possible (Bartholow, 2000a; Borman and Larson, 2003). Therefore, physically-based heat balance models can be reasonably used to predict in hypothetical scenarios, such as the impacts of dam removal (Bushaw-Newton

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http://dx.doi.org/10.1016/j.envsoft.2016.04.015
1364-8152/

et al., 2002; Bartholow et al., 2005), in-stream flow manipulations (Sinokrot et al., 1995; Sinokrot and Gulliver, 2000; Zorn et al., 2008), catchment land use and anticipated climate change (Stefan and Sinokrot, 1993; Gooseff et al., 2005; Wiley et al., 2010). Unfortunately, the flexibility of these models often comes at a cost of substantial difficulties in parameter estimation. Most widely used temperature models depend on time averaged solutions of full energy budgets, and therefore, require detailed local parameterization to capture major heat fluxes into and out of streams (TVA, 1972; LeBlanc et al., 1997; Bartholow, 2000a; Borman and Larson, 2003; Chapra et al., 2006). These typically include: radiation, convection, conduction, evaporation, and advection (Kreith, 1973; Sinokrot and Stefan, 1993; Lanini et al., 2004; Caissie, 2006).

In many practical applications, however, we do not have sufficient data at the required scales to estimate parameter values needed in a spatially explicit manner. As a result, comprehensive parameterization schemes make many widely used temperature models hard to apply to large river networks, multiple locales, or regional-scale systems (Edinger et al., 1974; Crittenden, 1978). In addition, most existing heat balance models use steady state solutions to simplify their calculations. These models assume that flows are constant for the entire simulation period and/or the boundary conditions being simulated are homogeneous and constant (Bartholow, 2000a; Borman and Larson, 2003; Bartholow et al., 2005). Yet, the temperature dynamics of streams and rivers is highly sensitive to changing diurnal environments and to changing rates of water accrual and discharge (Sinokrot and Stefan, 1993; Caissie, 2006). This makes it difficult to capture the temperature dynamics using steady state models. Furthermore, in the context of basin-wide and/or regional-scale ecological assessments, relationships between temperatures and flows can be quite different geographically, requiring very large data collection efforts simply to correctly parameterize a large-scale model.

To address the aforementioned problems in the context of dynamic simulation, we developed a new Reduced Parameter Stream Temperature Model (RPSTM). In this paper, we show the derivation of the simplified RPSTM heat balance equation from the full energy balance with the intent of capturing correct parametric influences without the need for detailed site-specific data. Following that, we provide a brief example of its implementation, examine input sensitivity, and discuss sources of prediction error of the RPSTM approach.

2. Theoretical derivation

2.1. Heat flux in river channel systems

Temperature is a measure of the amount of energy a system contains. Heat flux (dq/dt) in the full energy budget includes three standard mechanisms of heat transport: radiation, convection, and conduction.

In a stream system, energy flux related to radiative processes include solar radiation (i.e., shortwave radiation), long wave radiation, and back radiation. Consequently, physically-based in-stream temperature models treat the heat exchange processes as a combination of these major thermal processes (Bartholow, 2000a; Borman and Larson, 2003; Caissie, 2006; Chapra et al., 2006):

$$\frac{\mathrm{d}q}{\mathrm{d}t} = (SR + LR - BR) + C_v + C_d \tag{1}$$

where dq/dt is heat flux transferring through a unit surface over a specified unit of time (J/m^2h), SR is heat flux from solar radiation at the water surface (J/m^2h), LR is heat flux from longwave radiation (J/m^2h), BR is heat flux of back radiation from the water (J/m^2h), C_v is heat flux of convection (J/m^2h), C_d is heat flux of conduction (J/m^2h).

Shortwave solar radiation is typically the largest thermal input and strongly affects in-stream water temperature (TVA, 1972; LeBlanc et al., 1997), but its quantity is highly variable both within and between days. Latitude, longitude, and attenuation rate, also affect the quantity of solar radiation reaching a stream. While latitude and longitude are easily incorporated into a model, attenuation of solar radiation by local variations in atmospheric transmission, cloud cover, reflection, and canopy shade, is extremely difficult to specify other than by local measurement.

Longwave radiation is the radiation emitted from nearby objects including the surrounding atmosphere and ground itself. It falls within the infrared portion of the spectrum (Adam and Sullivan, 1990). The downward flux of longwave radiation from the atmosphere to a stream can be calculated using the Stefan-Boltzmann Law:

$$LR = \sigma(T_a + 273)^4 \varepsilon_a (1 - R_L) \tag{2}$$

where T_a is air temperature ($^\circ$C), σ is the Stefan-Boltzmann constant (5.67×10^{-8} W/m^2 K^4), ε_a is emissivity of the atmosphere, and R_L is the reflection coefficient, which is typically assumed to equal 0.03, and is negligible.

A stream also radiates back to the atmosphere, the ground near the stream, and the riparian vegetation. The amount of this back radiation from the water surface can also be approximated by the Stefan-Boltzmann law:

$$BR = \varepsilon\sigma(T_w + 273)^4 \tag{3}$$

where T_w is stream surface temperature ($^\circ$C), and ε is emissivity of water.

Furthermore, convection, including atmospheric convection and evaporation, is also important to the overall energy budget of a stream. Convection occurs mostly across the air-water interface when air and water temperature differs. As a consequence, the rate of the convective heat flux (C_v) can be computed as (Bowen, 1926; Kreith, 1973):

$$C_v = h_c(T_a - T_w) \tag{4}$$

where C_v is heat flux of convection (J/m^2h), T_w is stream surface temperature ($^\circ$C), T_a is air temperature ($^\circ$C), and h_c is a heat transfer coefficient (J/m^2h °C). The heat transfer coefficient (h_c) can be calculated as:

$$\frac{h_c}{k_e} = 1.5 \times 10^6 \tag{5}$$

where k_e is the evaporative coefficient for evaporation, and can be estimated as:

$$k_e = 1.74 \times 10^{-6} \times (1 + 0.72V_a) \tag{6}$$

where V_a is wind velocity (m/s). The wind speed dependency is a function of the elevation above the water surface where the wind is measured.

2.2. Stream reach temperature

Empirically, streambed and stream water temperatures typically follow variations in air temperature, but are lagged with increasing depth (Acornley, 1999; Brown et al., 2006). For example, water temperature in a stream responds to the atmospheric conditions in the time constants on the order of about 40 hours for every meter of