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A simplified river temperature model and its application to streamflow management

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Abstract

Relations between water temperatures and river discharge or flow depth are developed from a simplified model, using an analytical solution and the concept of surface heat flux and equilibrium temperature. The effects of streamflow and weather conditions on summer river temperatures are analyzed and quantified. The model relationships are compared with 5 summers of field data for river temperature (measured hourly), weather variables (hourly), and discharge (daily) for the central Platte River, Nebraska, USA. The method is applied to the Platte River to derive weather-related flow requirements for controlling summer river temperature maxima through streamflow management. Above a critical discharge, increasing the discharge has little effect on reducing water temperature.

Introduction

River temperature is an important element of the aquatic environment. High temperatures, long periods of elevated temperature, and large diurnal variations in temperature can be lethal to aquatic life (Bovee, 1982; Carreker, 1985; Sibley and Strickland, 1985). Some species of fish can not survive a week of water temperatures exceeding 25°C (Brett, 1956).

River temperatures follows both daily and seasonal cycles, in response to heat inputs and outputs under specific hydraulic conditions (discharge and channel morphometry) and meteorological conditions (air temperature, solar radiation, wind, and humidity). Ambient atmospheric conditions are the main control, but river temperature can be affected by streamside shading. Stream-flow management practices, such as release of cool water from a reservoir, also can alter the water temperature in a river (Morse 1972; Hockey *et al.*, 1982).

The effects of discharge on river temperatures need to be isolated and quantified. A model that simulates river temperature as a function of weather and flow can be used to predict the discharges needed within a river reach to maintain temperatures within desirable limits. Given the variability of weather over a summer season, streamflow requirements need to be weather-dependent. Variable minimum river discharges may make it possible to protect aquatic environments and wildlife habitat, while efficiently utilizing scarce water resources.

Based on field observations during warm months in a 15-km stretch of the Little Deschutes River near Lapine, Oregon, Morse (1972) found an increase in daily mean water temperature by a reduction of flows. Hockey *et al.* (1982) observed that in the Hurunui River, New Zealand, water temperature increased by about 0.1°C for each 1 m^3/s decrease in flow resulting from abstraction for irrigation; they developed a heat budget model for numerical predictions of river temperature. Grant (1977) found that at discharges greater than 10 m^3/s water temperatures decreased with increasing discharge. There were considerable scatters in the statistical relationships determined by Hockey *et al.* (1982) and Grant (1977) because their water temperature measurements were sporadic and involved a broad range of meteorological conditions over months or years.

Although the influence of discharge on water temperature has been recognized, quantitative temperature-flow relationships and the instream flow requirements to manage stream temperatures have not been successfully developed. Water temperatures presented in previous field data analyses represent the combined effects of flow and weather. This made it difficult to successfully develop any definite temperature-flow relationships for a wide range of weather variables and to quantify the effects of instream flow on summer river temperatures. The data used in previous investigations were instantaneous river temperatures measured intermittently or randomly. Continuous field measurements are needed to develop temperature-flow relations that can be used in stream-flow management to control river temperatures. It is necessary to precisely define river temperature variables, such as daily maximum, daily mean, diurnal variation, and time lag. In addition, weather conditions need to be characterized, parameterized and grouped.

In this study, the relationships between water temperatures, meteorological conditions, and discharge are established using a simplified model and an analytical solution. The relationships are validated against continuous observations and measurements of a reach of the Platte River, Nebraska, USA, over five summers, including water temperature, discharge, channel geometry, and weather. The concept of equilibrium temperature and surface heat transfer coefficient are adopted for the analytical solution to the one-dimensional river heat transfer equation. The influence of weather is incorporated into two parameters that represent all the weather variables and meteorological conditions that can influence river temperatures. The effects of flow depth and discharge on stream temperatures, including diurnal variation and daily maximum, are investigated. The weather-related flow requirement curves are derived from the quantitative temperature-flow relationships. Weather-dependent minimum instream flows for maintaining temperatures below a given limit (e.g. 32° or 35°C) are obtained from the flow requirement curves for two locations on the Platte River.

The model and analytical solution

The governing equation for one-dimensional flow and heat transfer in a river is

$$\frac{\partial T}{\partial t} = -\frac{1}{A}\frac{\partial (QT)}{\partial x} + \frac{1}{A}\frac{\partial}{\partial x}\left(EA\frac{\partial T}{\partial x}\right) + \frac{H_f + S_b}{\rho c_p D} \qquad (1)$$

where T = cross-sectional averaged water temperature, t = time, Q = river discharge, A = cross-sectional area, D = mean flow depth in the river, H_f and S_b = net rates of heat exchange across the water surface and sediment bottom, respectively, ρ = water density, c_p = heat capacity of water, and E = longitudinal dispersion coefficient. The exchange of heat through the air-water interface (H_f) is one of the most important factors that govern the temperature of a river (Edinger *et al.*, 1974; Gu and Stefan, 1985; Sinokrot and Stefan, 1993). Stream discharge (Q) influences river water temperatures through advection. Flow depth (D) and river surface area (Ax/D) or the volume of the water body (Ax) are important because of the thermal inertia of water. Water released from a reservoir affects water temperature for only a short distance downstream.

Direct relationships between discharge (Q) and river temperature (T) need to be developed by isolating the effects of discharge on river temperatures from those of meteorological factors. The relationships can be established by analytically solving Equation 1 for T, after some simplification. Various heat exchange processes involved in H_f are collected by two parameters for weather defined in the following section. Two insignificant terms, for which field data are not available, are dropped from the equation – longitudinal dispersion (E) and groundwater and/or heat flux across the water-sediment interface (Grant 1977). Longitudinal dispersion can be neglected for a thermally well-mixed stream, unless pulse injections of heat, or flow, bends and other channel irregularities are to be considered (Thomann and Meuller, 1987; Brocard and Harleman, 1976; Fischer, 1967; Hockey *et al.*, 1982; Edinger *et al.*, 1974). In Minnesota, a neighbor state north of Nebraska, the heat flux from water columns to a lake bottom during June-September and to a riverbed in September were estimated as about 0.5% and 5%, respectively, of the net solar radiation entering these water bodies. These estimates were made from the predictions of temperatures in the lake bottom (Gu and Stefan, 1990) and the riverbed (Hondzo and Stefan, 1994).

Assuming steady state flow, the equation, which describes the temperature of a parcel of water moving down the river, is simplified to

$$\frac{Q}{A}\frac{dT}{dx} = \frac{H_f}{\rho c_p D} \tag{2}$$

The net rate of heat flux across the water surface, a term combining all weather variables, is expressed as

$$H_f = K\,(T_e - T) \tag{3}$$

where T_e = equilibrium river water temperature representing the effects of meteorological conditions (Thomann and Mueller, 1987; Edinger *et al.*, 1974) and K = overall water surface heat exchange coefficient. This coefficient is related to air temperature, wind speed, and relative humidity, and describes the rate at which water temperature responds to these heat exchange processes and approaches the equilibrium temperature. T_e is approximated by (Edinger *et al.*, 1974)

$$T_e = T_d + \frac{R}{K}\ (^\circ C) \tag{4}$$

where R = solar radiation (W/m^2) and T_d = dew point temperature (°C). The surface heat exchange coefficient, K (W/m^2 °C), is computed from air temperature, wind speed, and relative humidity or dew point temperature (Thomann and Mueller, 1987).

After substituting Equation 3, the solution to Equation 2 for a steady state, i.e. no change in environmental variables with time (constant T_e and K), is

$$\frac{T - T_e}{T_e - T_o} = -e^{-\frac{1}{Q^*}} \tag{5}$$

where, $Q^* = \dfrac{Q}{Ax}\dfrac{\rho c_p D}{K}$,

which may be viewed as a non-dimensional time of travel, T = daily mean river temperature, and T_o = temperature at the upstream site, i.e. T at $x = 0$. The equilibrium temperature continually varies in response to changing meteorological conditions in a diurnal cycle and a seasonal cycle. The water temperature is continually being driven toward T_e by the difference between T and T_e. The long-term average water temperature at a large distance x is expected to be equal or very close to the time-average value of the equilibrium temperature where the effect of T_o on T becomes insignificant. The role of Q on T is to affect the distance required to attain T_e.

Daily maximum water temperature is an important indicator of a water body's response to daily weather changes. Aquatic biota can be adversely affected by a brief period of high maximum water temperatures, even though the daily mean may be tolerable. The daily amplitude of the actual water temperature, i.e. difference between maximum and minimum, is only a small proportion of the diurnal amplitude of the equilibrium temperature, due to the attenuation effect of the water column and the lag between T and T_e. Diurnal variation in short-wave solar radiation alone would cause a large daily variation in equilibrium temperature. The large diurnal variation of air temperatures adds to the diurnal amplitude of T_e. The diurnal amplitude of T_e is damped by the surface heat exchange coefficient, which also varies daily as a function of wind and humidity. For a diurnal cycle, the lag between the occurrence of daily maximum T_e and the occurrence of daily maximum T is another important parameter and is related to depth or discharge (Edinger *et al.*, 1974). Generally, maximum T_e occurs near noon due to the influence of solar radiation, while water temperatures reach a maximum in the late afternoon (around 16:00-18:00 hours).

Assuming $\partial(QT)/\partial x = 0$, i.e. $\partial T/\partial x \to 0$ at a large distance x, Equation 1 is simplified to

$$\frac{dT}{dt} = \frac{H_f}{\rho c_p D} \qquad (6)$$

to describe the temperature variation of a parcel of water over time in a stationary a body of water as well as in running waters. Equations 5 and 6 can be linked by considering $x = Qt/A$. Equation 6 can be used to analyze how the daily maximum and the amplitude of water temperature fluctuations vary with discharge or depth and with weather in a diurnal cycle. The equilibrium temperature may be approximately expressed as a single sine function of time in a daily cycle. The solution to Equation 6 gives a sinusoidal variation of water temperature with time (Edinger *et al.*, 1974). The diurnal amplitude of water temperature is then related to daily variation of equilibrium temperature at a large distance (x) by

$$\Delta T_w = \frac{\Delta T_e}{\sqrt{1 + (D^*)^2}} \qquad (7)$$

where $\Delta T_e = (T_{e,max} - T_{e,min}) / 2$ and $\Delta T_w = (T_{max} - T_{min}) / 2$, amplitudes of sinusoidal variations of river temperature and its equilibrium, respectively, and

$$D^* = \frac{2\pi\varpi\rho c_p D}{K} \qquad (8)$$

where D^* is a dimensionless flow depth in which ϖ = frequency of sinusoidal variation of equilibrium temperature with time (Brady *et al.*, 1969; Edinger *et al.*, 1974). Using Manning's equation

$$D = \left(\frac{Q}{W}\right)^{3/5} n^{3/5} S_o^{-3/10} \qquad (9)$$

to relate water column depth (D) in a shallow river to discharge (Q), D^* is expressed as

$$D^* = \frac{C_1}{K} \left(\frac{Q}{W}\right)^{\frac{3}{5}} \qquad (10)$$

where W = top (surface) width of the river and $C_1 = 2\pi\varpi\rho c_p n^{3/5} S_o^{-3/10}$ in which n is Manning's roughness coefficient and S_o is the river bottom slope. Equation 7 suggests that a higher flow rate or greater water column depth would result in a lower ratio of diurnal amplitude of water temperature to that of equilibrium temperature. In other words, diurnal variation of water temperature would be smaller as the river discharge increases, for a fixed amplitude of the equilibrium temperature variation. With greater depths, which result from higher discharges, there is more water to heat up, while the heat flux per unit surface area is the same. Thus a greater depth or an increased discharge corresponds to greater thermal inertia. The lag time between T_{max} and $T_{e,max}$ is increased and the maximum water temperature is reduced as depth or discharge increases.

The relationship between daily maximum river temperature and discharge or depth can be obtained from Equations 5 and 7. To solve these equations for T_{max}, the diurnal amplitude of water temperature, $\Delta T_w = (T_{max} - T_{min})/2$, needs to be converted to $(T_{max} - T_{mean})$. In the derivation of Equation 7, a sine function is assumed for equilibrium temperature. However, in a real river, water temperature and its equilibrium may not be a true sine function of time in a diurnal cycle, i.e., $(T_{max} - T_{min})/2 \neq T_{max} - T_{mean}$. An empirical ratio is used to take account of any deviation from the sinusoidal pattern, and is pre-determined by field measurements

$$r_w = \frac{T_{max} - T_{mean}}{\Delta T_w} \tag{11}$$

The ratio r_w is expected to be a constant for a river, depending on river characteristics, and vary with the ratio for measured equilibrium temperature, defined as

$$r_e = \frac{T_{e,max} - T_{e,mean}}{\Delta T_e} \tag{12}$$

The pre-determined ratio r_e represents the deviation from a truly sinusoidal variation of T_e for a daily period and is controlled by weather pattern in the study area. Substituting Equation 7 into Equation 11 and assuming constant K, the analytical solution for daily maximum water temperature is

$$T_{max} = T_{mean} + r_w \frac{\Delta T_e}{\sqrt{1 + (D^*)^2}} \tag{13}$$

where T_{mean} is computed by Eq. 5 and ΔT_e is calculated from $T_{e,max}$, $T_{e,mean}$ and r_e using Equation 12. Equation 13 can be written in a dimensionless form,

$$T^*_{max} = -e^{\frac{1}{Q^*}} + \frac{r_w}{r_e} \frac{T^*_{e,max}}{\sqrt{1 + \alpha(Q^*)^{6/5}}} \tag{14}$$

where $\alpha = C_1^2 x^{6/5} K^{-2/5} (\rho c_p)^{-1/5}$, $T^*_{\max} = (T_{\max} - T_{e,mean})/(T_{e,mean} - T_o)$, and $T^*_{e,\max} = (T_{e,max} - T_{e,mean})/(T_{e,mean} - T_o)$. Equations 13 and 14 indicate that daily maximum river temperature at a river point is inversely related to discharge and depth under constant meteorological conditions. When $T^*_{e,max} = 0$, Equation 14 becomes Equation 5, which is for the daily mean river temperature.

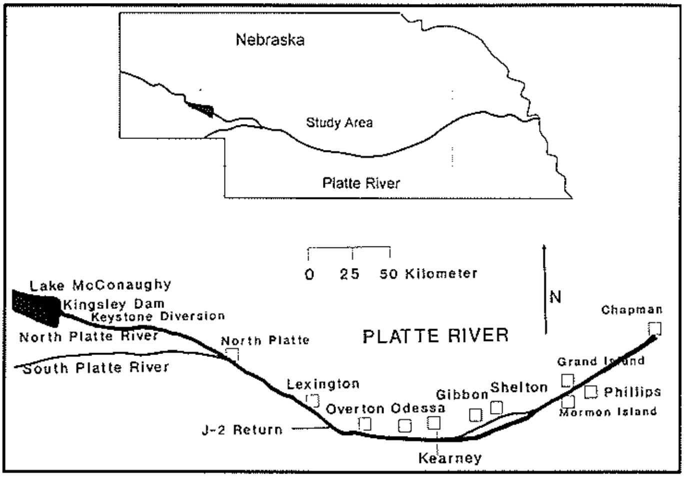
Following a sinusoidal variation similar to that in a diurnal cycle, water temperature in a seasonal cycle is driven toward its equilibrium by the difference between T and T_e during summer warming and winter cooling (assuming no ice is presenting) in response to changes in meteorological factors. Daily mean and maximum water temperatures are usually below their equilibrium values in warm weather and above them on cold days. As indicated by Equations 5 and 13-14, under steady meteorological conditions or fixed weather factors (T_e and K or H_f), higher discharge tends to reduce river temperatures (stop them rising so high above T_o) during summers when $T_e > T_o$ and to elevate them (stop them falling below T_o) during winters when $T_e < T_o$.

Comparison with field measurements

The study area used to validate and apply the relationships is a 125 km section of the central Platte River downstream of Lake McConaughy and the Kingsley Dam, extending from Overton to Grand Island, Nebraska (Fig. 1). The sand-bed river has a bottom slope of 0.001-0.0017. The Manning's roughness coefficient (n) ranges from 0.02 to 0.05 with a mean of 0.0355 (Dinan, 1992). The discharge in this reach ranged from 0.1 to 180 m^3/s during the summers of 1989-93, with a mean flow of 20 m^3/s, at which the reach-averaged surface width was 200 m, flow depth was 0.26 m, and cross-sectional mean velocity was 0.38 m/s.

Field data have been collected along the reach for the summer periods from June to August since 1988. Available data include river water temperatures, streamflow discharges, and climatic parameters (air temperature, solar radiation, humidity, and wind speed). River temperatures were measured hourly at Overton, Odessa, Gibbon, Shelton, Mormon Island, and Philips. Weather variables were measured hourly at stations located at Lexington, Gibbon, and Grand Island. Daily discharges were available at gauging stations near Overton, Odessa, Kearney and Kearney Canal, and Grand Island. The hourly river temperatures and meteorological data were analyzed to obtain daily maximum, mean, and minimum values. Daily mean T_e and K, daily maximum T_e, and corresponding K at the time

Figure 1 – Study area – the central Platte River, Nebraska, USA.



Daily mean T_e and K, daily maximum T_e, and corresponding K at the time of maximum T_e were obtained from hourly values of T_e and K calculated from meteorological observations. River temperatures measured at Overton were used as T_o for the analyses of temperature-discharge relations at Shelton, 66 km downstream, and Phillips, 120 km downstream.

Based on the preliminary analysis of 1989-93 river temperature data, the ratio, r_w, had a mean of 1.12 at Overton and Odessa, 1.13 at Gibbon, 1.10 at Shelton, and 1.13 at Grand Island and Phillips: i.e. the ratio did not vary much from station to station. The values represent small deviations from a truly sinusoidal variation in water temperature (for which $r_w = 1.0$). The ratio r_e, which was independent of flow, was found to be 1.23 for the study area and deviates more from the exact sine function than r_w does. The reach-wide averages of K, T_e, and river temperature T over the five summers were 29.6 W/m² °C, 24.5°C, and 23.5°C, respectively. These values were computed from the weather observations and water temperatures during 1989-93 summers. Meteorological conditions vary over a summer from June to August. The range of the daily means for K was 17-76 W/m² °C, for T_e was 11-36°C, and for T was 13-30°C for the 1989-93 summers. The calculated daily maxima for $T_{e,max}$ ranged from 15° to 56°C and the measured T_{max} ranged from 14.7 to 38°C, with five-summer means of $T_{e,max} = 42$ and $T_{max} = 29.4$°C. The average amplitudes of sinusoidal variation of river temperature (ΔT_w) and its equilibrium (ΔT_e) were estimated to be 6°C and 15°C for the five summers.

Sinusoidal variations of equilibrium and water temperatures are exhibited by the central Platte River. Figure 2 shows the 5-summer averages of diurnal variations measured at the Shelton station; they are described reasonably well by a sine function of time, as assumed in the derivation of Equation 7. The amplitude of diurnal water temperature variation is a proportion of that of equilibrium temperature. On a typical day during 1989-1993 summers at Shelton on the Platte River, T_{max} occurred in late afternoon (16:00-18:00), while the equilibrium temperature peaked at about 13:00. Water temperatures usually cross T_e twice each day as the water warms and cools in response to the diurnal cycles.

The ratio of diurnal amplitude of measured river temperature to that of equilibrium temperatures at Shelton is plotted against the normalized flow depth in Figure 3, together with the theoretical curve (Eq. 7). Flow depth is calculated from the discharge data using Manning's Equation. The general trend of the data seems to follow Equation 7. The standard errors of prediction is 0.166. The mean value of the measured ratios is 0.366. The observed ΔT_w ranged from 0.8° to 9°C (mean 5°C). The observed ΔT_e ranged from 2.5° to 26°C with an average value of 13.5°C. The field measurements and calculated values show some agreement, although the theory overpredicts the ratio, and more scatter of the data is seen at low flows or small depths (D^*). This over-prediction and scatter are probably caused by the omission of some heat components for which data are not available and some uncertainties in field measurements of river temperatures. Under low flow conditions, the contribution of low-temperature groundwater may not be negligible. In a shallow and clear body of water, about 5% of solar radiation reaches the bottom and is absorbed by bottom sediments (Hondzo and Stefan, 1994). Conductive heat flux across the water-sediment interface causes the water column to lose heat to the bottom during daytime, resulting in a lower daily maximum river temperature, and to gain heat from the sediments during nighttime, leading to a higher minimum temperature. In addition, the top width of the river was used as the wetted perimeter in the computation of flow depths, as no cross-sectional profiles were available. This simplification is valid for rivers with wide, shallow channels. Water temperatures were measured in the main channel of the river, while the computed values represent the temperatures averaged over an entire cross section. According to Equation 7, shallow waters have higher temperatures than deeper flows in the main channel. Therefore, lateral variation in river temperatures may also contribute to the discrepancy and scatter in Figure 3. More accurate prediction may be obtained by computing values for the main channel

Figure 2 – Diurnal variations of water temperature and equilibrium temperature at Shelton.

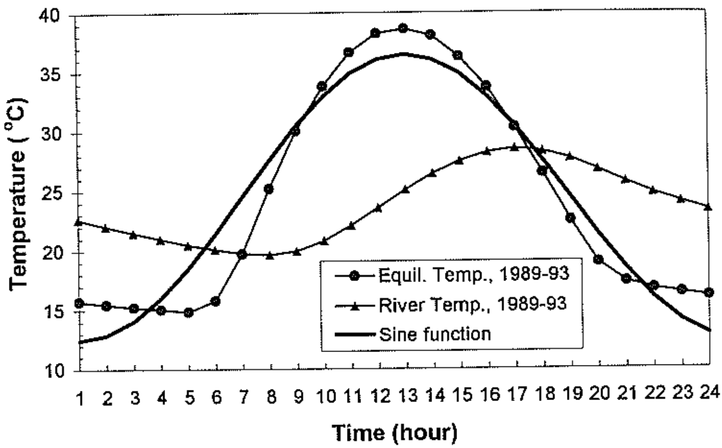
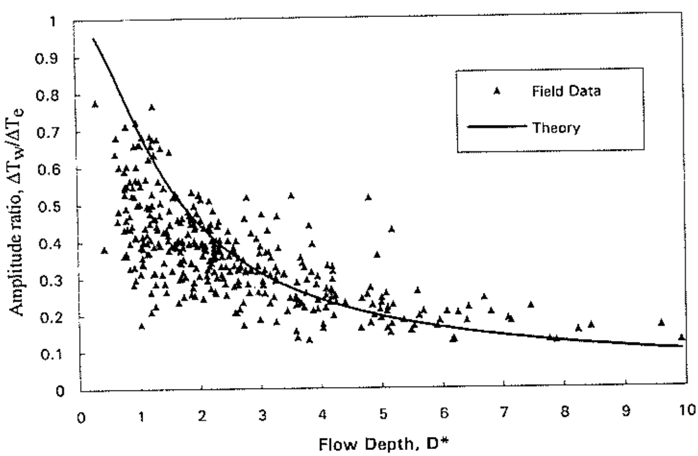


Figure 3 – Ratio of amplitude of diurnal river temperature variation to that of equilibrium temperature versus flow depth at Shelton for 1989-93 summers (June-August).



and the shallow overflow areas separately. Moreover, the effects of shading may become more significant during low flows.

Application of Equation 7 to data for the central Platte River at Shelton indicates a strong relationship between the diurnal water temperature fluctuations and flow depth/discharge. The theoretical line (Eq. 7) corresponds to the average condition, $K = 30$ W/m^2 °C. Figure 3 clearly shows a decrease in the ratio $\Delta T_w/\Delta T_e$ with increasing flow depth or discharge (D^*). A large volume of water takes longer to be heated or cooled than a small one with equivalent surface area. The thermal inertia of water is increased by high discharges, reducing water temperature rises during the hot daytime and falls during the cool nighttime. Daily maximum and minimum river temperatures are increasingly influenced by weather (mainly air temperature and solar radiation) as discharge, and therefore thermal mass, decreases. It is shown in Figure 3 that $\Delta T_w/\Delta T_e$ decreases from 1 to 0.2 as D^* varies from 0 to 5, while it is reduced from 0.2 to 0.1 by an increase in D^* from 5 to 10. The attenuation of $\Delta T_w/\Delta T_e$ by flow in the range $D^* = 0\text{-}5$ is approximately 8 times of that by flow in the range of $D^* = 5\text{-}10$.

Daily maximum river temperatures (T_{max}) measured at Shelton and Phillips during 1989-93 summers and computed using Equation 13 and their variation with discharge, are presented in Figures 4 and 5. Daily maximum river temperatures under high flows are expected to be significantly lower than those under low flows, because of the greater thermal inertia of the larger volume of water and greater corresponding depth (Fig. 3). To separate the impact of discharge (Q) on T_{max} from that of weather, the T_{max}-Q relationships were developed for days with the same or similar weather conditions. The relative daily maximum net heat exchange between atmosphere and river water columns ($H_{f,max}$) was used as the sole parameter that collects all weather factors, it reflects the main heat transfer processes. $H_{f,max}$ were calculated from hourly weather data and assume a constant river temperature, i.e., a long-term average (24°C at Shelton and 24.7°C at Phillips) to exclude the effect of water temperature on $H_{f,max}$ (Eq. 3). The relative values of $H_{f,max}$, which vary from -250 to 1200 W/m^2, served to characterize the wide range of meteorological conditions (cold to hot). Since the values of $H_{f,max}$ are used only for sorting the weather conditions, the absolute value of $H_{f,max}$ is not important. The role of the constant water temperature in $H_{f,max}$ calculations is similar to that of mean sea level in river stage measurements.

As shown in Figures 4b, 4c, 5b, and 5c higher, $H_{f,max}$ resulted in higher T_{max}. Reducing the $H_{f,max}$ range from 500 W/m^2 (Figs. 4a and 5a) to

Figure 4 – Measured and computed daily maximum river temperatures and their variations with discharge at Shelton (1989-93) for three $H_{f,max}$ ranges.

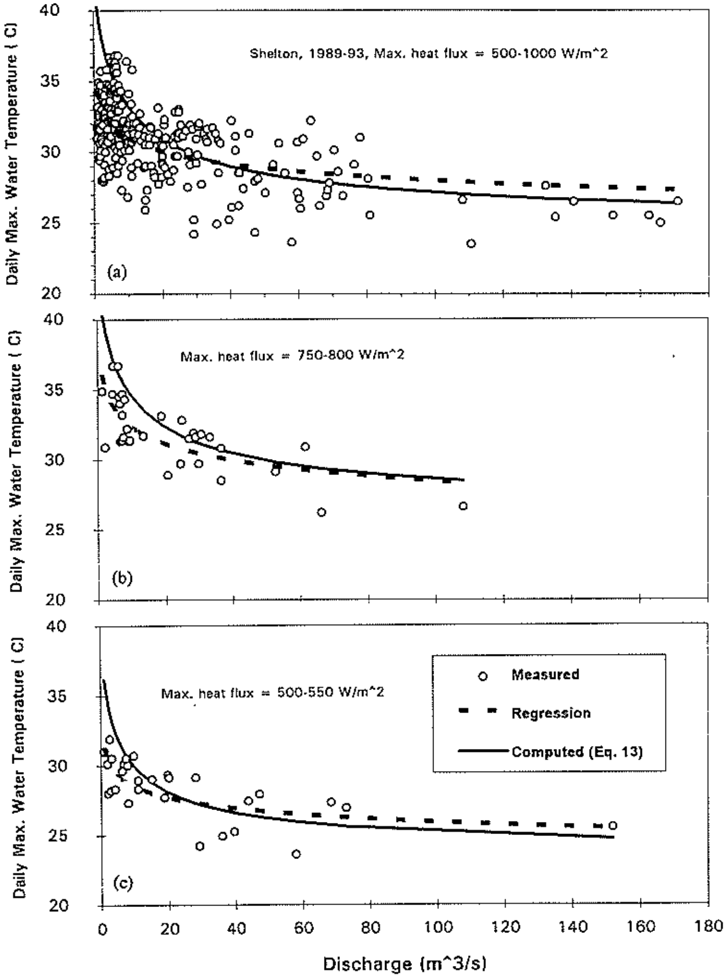
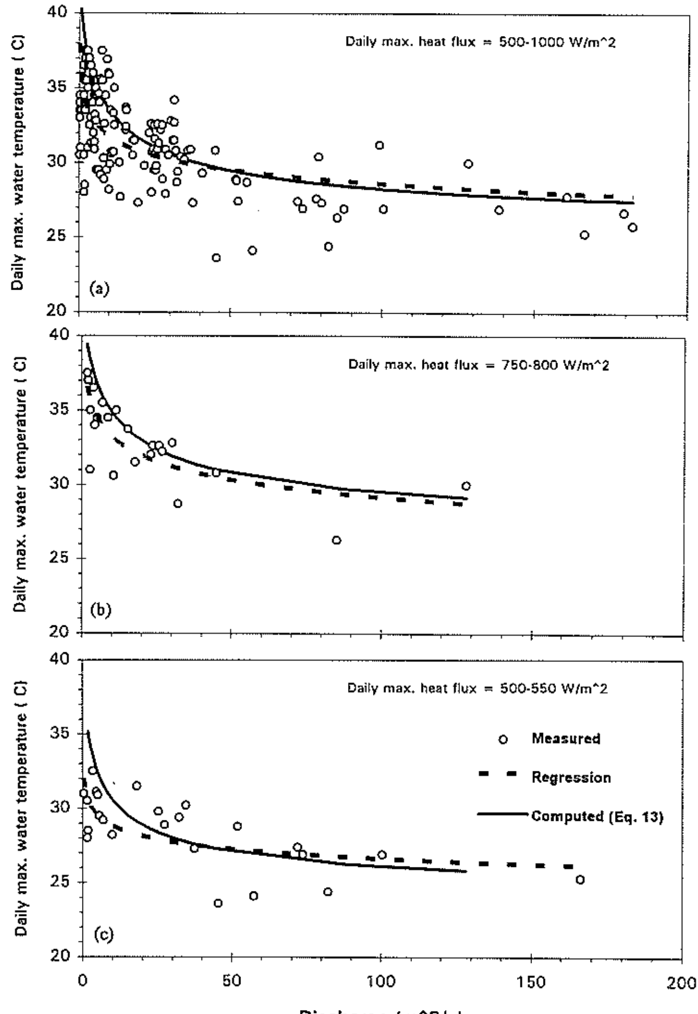


Figure 5 – Measured and computed daily maximum river temperatures and their variations with discharge at Phillips (1991-93) for three $H_{f,max}$ ranges.



50 W/m^2 (Figs. 4b, 4c, 5b and 5c) leads to less scatter of the data points for the measured T_{max}. Although predictions can be made for individual days using Equation 13, a temperature-discharge relationship can be more clearly seen by plotting T_{max} at variable discharges under a constant weather conditions. T_o, T_e, and K were averaged over the days selected to obtain the predicted T_{max}-Q relationships. River temperatures observed at Overton were taken as T_o. Overton is 66 km upstream of Shelton and 120 km upstream of Phillips; this satisfies the assumption of a large distance x for Equations 6-7 and 13. Each of the prediction curves and the regression curves (using a power function) for the field data in Figures 4 and 5 represents an average relationship over a given range of weather conditions ($H_{f,max}$). An overall agreement between predictions and measurements is shown in Figures 4 and 5, though Equation 13 over-predicted T_{max} at low flows ($Q < 20$ m^3/s). The over-prediction may be caused by the neglection of some heat sources, the simplification of channel cross sections, and the effect of shading. T_{max} decreases with increasing discharge, as indicated by a sharp drop of T_{max} when $Q < 30$ m^3/s and a slow reduction when $Q > 30$ m^3/s. T_{max} at Phillips is slightly higher than at Shelton.

Applications to streamflow management

Many states in the USA use 32°C as their upper limit for river temperature for habitat protection. The 35°C level, however, is close to the "critical thermal maximum" for many fish species found in the Platte River basin (Dinan, 1992). For a specified river temperature standard (32° or 35°C), the instream flow requirements under different weather conditions (cold, warm, hot) can be derived from the relationships between daily maximum river temperature and flow. Information on local weather and on river characteristics, including bottom roughness, slope, length, and cross-sectional geometry, are required in the analytical solution to determine quantitative relationships. The weather-dependent requirement curves are then used to determine minimum instream flows (Q_{min}), expressed as a function of characterizing weather parameter. The critical discharge (Q_c) divides flows that have a significant effects on temperature from flows that have little effect. Q_c can be used for evaluating the most effective minimum flow requirements for streamflow management. If $Q_{min} > Q_c$, Q_c should be used as the Q_{min} for economic reasons, because an increase in Q ($>Q_c$) does not effectively reduce T_{max}.

For given weather conditions and river characteristics, the relationships between T_{max} and Q are obtained with selected ranges of $H_{f,max}$ (Eq. 13 and

Figs. 4 and 5) or $T_{e,max}$ (Eq. 14) as a reference parameter. Using the relationships, one can determine the minimum river discharges required to keep water temperatures below limits (32 or 35°C) under different ranges of weather conditions, by setting $T_{max} \leq 32°$ or 35°C in Equations 13 or 14. When Equation 13 is used to predict Q_{min}, the calculated $H_{f,max}$ must be sorted and classified into groups, and $T_{e,max}$, T_e, K, T_o in Equation 13 need to be set to constants equal to the means in each $H_{f,max}$ range. For the Platte River at Shelton, the minimum flow required for maintaining the 32°C river temperature limit under a given weather regime ($H_{f,max}$) was determined by letting $T_{max} = 32°C$ in Equation 13 and solving the equation for Q_{min}. Similarly, different Q_{min} for the limit of 35°C could be obtained for various weather conditions, i.e. $H_{f,max}$ values. Figure 6 presents the curve for average in-stream flow requirements for the central Platte River at Shelton based on the 5 years of data. Since $T_{e,max}$ is directly involved in Equation 14, it is simpler to use $T_{e,max}$ as the weather index. $T_{e,max}$ varies from 15° to 60°C with an average value of 40.6°C for the central Platte River area during the 1989-93 summers. Figure 7 presents the simulated T_{max}-Q curves for selected ranges of $T_{e,max}$ at Phillips. Flow requirements at the river site for the two river temperature limits, 32° and 35°C, were developed for different weather regimes (Fig. 8).

By simulating temperature variations due to flow changes, instream flow requirements can be determined for various weather conditions to control summer river temperatures and protect the aquatic habitat. The T_{max}-Q relationships and the Q_{min}-$H_{f,max}$ or Q_{min}-$T_{e,max}$ curves, developed from the simplified analytical solution, can aid in long- and short-term planning for streamflow management and reservoir or dam operations. Daily real-time control or adjustment of river discharge to maintain given water temperature limits can be computed for various operational scenarios using the curves with short-term weather forecasts as an input. Different minimum flow requirements can be established for different meteorological conditions. For a specific day, the minimum discharge required for maintaining the standard of 32° or 35°C can be determined from the curves in Figure 6 or Figure 8 with $H_{f,max}$ or $T_{e,max}$ calculated from the weather forecast.

Figure 6 – Minimum discharges (Q_{min}) required for the river temperature limit of 32°C at Shelton under various weather conditions ($H_{f,max}$).

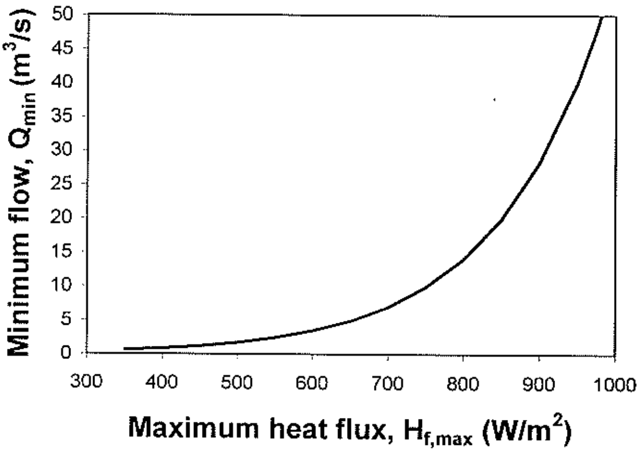


Figure 7 – Predicted variations of T_{max} with Q in the Platte River at Phillips for different $T_{e,max}$.

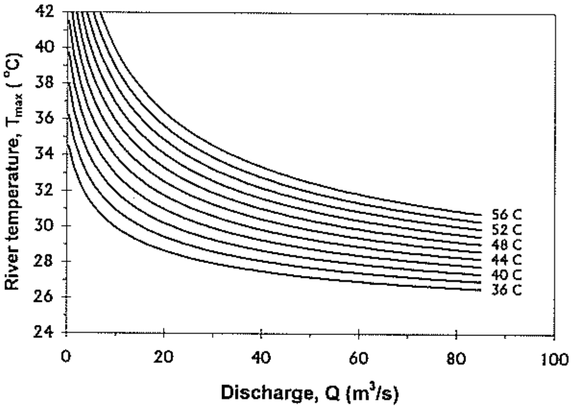
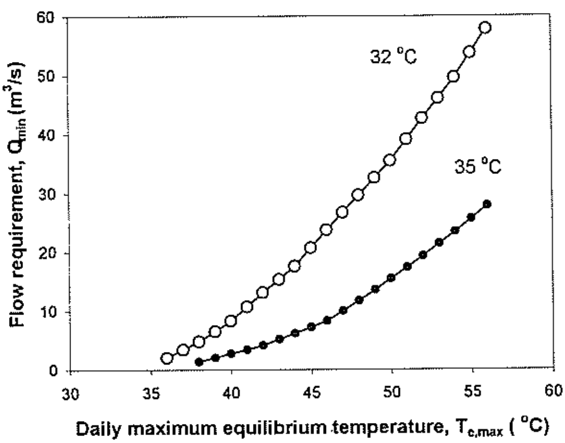


Figure 8 – Simulated weather-dependent flow requirement curves corresponding to river temperature limits of 32° and 35°C for the central Platte River at Phillips.



Conclusions

A simplified model, which simulates river temperature as a function of discharge or depth and weather, was used to quantify the effects of instream flow discharge or depth on summer river temperatures. Quantitative relationships between water temperatures and river flow were developed using an analytical solution and validated against 5 summers of field data for the central Platte River for daily maximum temperatures and their diurnal amplitude. The analytical results and field measurements agreed reasonably well. The method was used to derive minimum flow requirements for maintaining temperature standards under different weather conditions. As temperature is strongly related to flow, occurrences of maximum temperatures exceeding 32° or 35°C can be reduced by maintaining a minimum discharge in the central Platte River.

The analytical solutions and field observations showed that daily maximum river temperatures were significantly affected by streamflow under given weather conditions, as temperatures were delayed from reaching their equilibrium and attenuated by greater flow depth. Increasing discharge affects water temperatures primarily by reducing the amplitude of daily temperature variation, and thus lowering daily maximum water temperature. In a river where water temperatures can vary by up to 20°C over a day, reducing the amplitude of water temperature variation benefits the aquatic biota. It was found that maximum daily water temperatures in the Platte River were highly influenced by river flows below a critical discharge in the range of 20-40 m^3/s and a dimensionless flow depth (D^*) of 5. The occurrence of daily maximum river temperature greater than 32° or 35°C in the central Platte River could be effectively and significantly reduced by increasing stream-flow to 20-40 m^3/s. The reduction in river water temperature by increasing river discharge, however, became insignificant at very large flows ($Q_{min} > Q_c$).

The validated relationships were used to evaluate, based on historical flow data and weather conditions, the effects of instream flow on river temperatures. They were also used to calculate the weather-dependent minimum discharges needed to meet temperature standards in the Platte River to improve habitat conditions. The simplified model and analytical solutions provide a tool for predicting the possible consequences of future river and reservoir management scenarios on river temperatures and in turn on aquatic environment and wildlife. The predictions of temperature variations due to discharge changes provide information useful for the control of summer river temperature through streamflow regulations, and for the planning of reservoir water releases and streamflow diversions.

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