Quantifying the effects of stream discharge on summer river temperature

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Abstract Control of summer river temperature is needed for maintaining water temperature standards to protect aquatic biota and wildlife habitats. Given the fact that instream discharge, among meteorological and hydrological factors, may be the only one that can be practically managed, is it feasible to moderate summer river temperature through reservoir and streamflow regulations? An analysis is conducted to quantify the effects of the magnitude of instream flow on summer river temperature with weather as a reference. Relationships between water temperature and river discharge or flow depth are developed using a simplified model and adopting the concept of equilibrium temperature and bulk surface heat exchange coefficient. The relationships are validated against continuous 5-year field measurements at the central Platte River, Nebraska, USA. It was found that the variation of daily maximum water temperature with flow was stronger than that of daily mean. A critical discharge was obtained, which divides dramatic drop and slow variation in river temperature values. The existence of the critical discharge makes it possible to reduce or minimize the occurrence of daily maximum water temperature exceeding a standard at a river reach by increasing discharge to an achievable level. This study advances understanding of impacts of instream flow on summer river temperature and provides information useful in proper planning and design of reservoir operations and streamflow management.

Quantification des effets du débit sur la température des rivières en été

Résumé Il est nécessaire de maîtriser la température des rivières en été afin de respecter les normes de protection de la faune et des habitats aquatiques. Etant donné que le débit, parmi l'ensemble des variables hydrométéorologiques, est sans doute la seule variable à pouvoir être commandée, est-il possible, grâce à des actions sur les cours d'eau et les réservoirs, de maintenir la température des rivières en été à un niveau acceptable? Nous avons entrepris une analyse afin de quantifier les effets du débit sur la température des rivières en été pour différents types de temps. Des relations entre la température de l'eau et le débit ou la hauteur de l'eau ont été établies grâce à un modèle simplifié et à un coefficient global d'échange thermique de surface. Ces relations ont été validées sur cinq années de mesures sur la rivière Central Platte au Nebraska (USA). On a pu montrer que les variations selon le débit de la température maximale journalière étaient beaucoup plus importantes que celles de la température moyenne. On a pu déterminer un débit critique, qui limite les variations de température brusques ou lentes. L'existence de ce débit critique permet, en augmentant convenablement le débit, de réduire ou de minimiser l'occurrence sur un bief de rivière d'une température maximale dépassant une norme. Cette étude a permis de mieux comprendre l'influence du débit sur la température des rivières en été et fournit des informations utiles à la planification et à la conception de la gestion des réservoirs et des cours d'eau.

INTRODUCTION

Among the water quality characteristics of a river, temperature is probably one of the most significant and widely measured variables as most of the physical properties of

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water are functions of temperature. High temperatures with long duration (days) and large diurnal variations in warm months can be lethal to aquatic life (Bovee, 1982; Carreker, 1985; Sibley & Strickland, 1985). Some species of fish can not survive water temperatures exceeding 25°C when exposed for any one-week episode (Brett, 1956). Maintaining water temperature standards during summers is important to the biological integrity of a warm plain river that serves as a habitat for fish and birds (Crawshaw, 1977; Krapa, 1981). Concern has been growing over the persistent exceedance of a water temperature standard of 32°C in the Big Bend reach of the central Platte River downstream of a hydropower dam, Nebraska, USA (Fig. 1). This river reach provides an important habitat for a variety of migratory birds and forage fish that are necessary for the survival of certain endangered bird species (Dinan, 1992; Wilson, 1991; Krapa, 1981).

River water temperature is a resulting variable, which follows a diurnal cycle and a seasonal cycle, due to the net of heat inputs and outputs under specific hydrological (discharge, depth, and volume of groundwater exchange) and meteorological conditions (air temperature, solar radiation, wind and humidity). Water temperature is mainly controlled by the ambient atmospheric conditions. However, of all the factors, discharge may be practically the only one that can be managed to alter the water temperature. Reservoir and dam operations at various discharges can modify physical and chemical conditions and in turn affect the biological and ecological conditions of a downstream river reach. The use of streamflow management to improve river water quality and to moderate high river temperature is an emerging issue. In addressing the river temperature problem it is important to understand fully the impact of the quantity and quality of reservoir water release on the downstream river water quality, especially temperature. Therefore, the effects of discharge on summer river temperature need to be isolated and quantified. The development of quantitative relationships between water temperature and discharge is necessary for engineering practice to control summer river temperature (daily maximum and

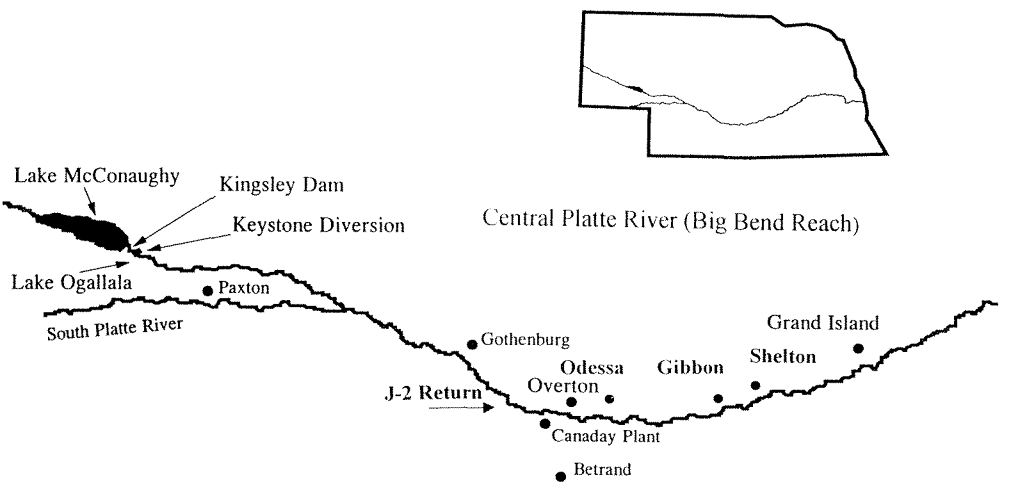

Fig. 1 Study area—the central Platte River, Nebraska, USA.

diurnal variation) through reservoir and streamflow regulations. The relationships can be incorporated into habitat assessment models as they relate to habitat suitability.

PREVIOUS STUDIES AND PRESENT WORK

River temperature is related to flow variables as well as to meteorological parameters. In the past, considerable attempts were made to investigate the response of water temperature to meteorological conditions. Edinger *et al.* (1974) suggested a thermal exchange coefficient and an equilibrium temperature, both of which depend on meteorological variables, for computing the net heat exchange rate at the water surface. The effects of flow variables, including discharge, depth, surface width and area, and upstream flow temperature, on river temperature were also investigated (Morse, 1972; Brown & Krygier, 1970; Grant; 1977; Hockey *et al.*, 1982; Churchill, 1963; Mitchell *et al.*, 1995). Other hydrological factors that influence water temperature may include surface runoff (storm events), snowmelt inputs and groundwater exchange. The interest stems from the need to provide information regarding the river temperature problem in streamflow management and reservoir regulations for aquatic environment protection. Methods used in the previous attempts include field estimates and mathematical calculations by statistical techniques with limited data, simple heat budget models, and empirical formulas. Hockey *et al.* (1982) and Grant (1977) conducted statistical analyses of field data. There were considerable scatters in the statistical relationships because of sporadic and scarce water temperature measurements and involvement of broad ranges of meteorological conditions during a long period of time over several months or years. Water temperature values presented in previous work represent the combined effects of streamflow and weather characteristics. The effect of river discharge on water temperature was not separated from that of weather. This made it difficult to successfully develop any definite temperature–flow relationships under a wide range of weather variables and to quantify the effects of instream flow on summer river temperature. Moreover, the data sets used in previous investigations, which were instantaneous river temperature measured intermittently or randomly, were short. The establishment of any theoretical and quantitative relationships for evaluating the effects of flow on river temperature requires: (a) an analytical solution approach, (b) precisely defined river temperature variables, such as daily maximum, daily mean, diurnal variation and time lag, etc., (c) characterization and parameterization of wide ranges of weather conditions, and (d) sufficient and continuous field measurements.

In this paper, the relationships between water temperature and discharge are established through theoretical and mathematical analyses and validated against 5-summer continuous observations and measurement, including water temperature, discharge, geometry and weather at the central Platte River, Nebraska, USA. Weather parameters are used as a reference to decouple the effects of discharge on water temperature from those of meteorological factors so that the role and significance of discharge can be isolated. The concept of equilibrium temperature and surface heat transfer coefficient are adopted for analytical solutions to the one-dimensional river heat transfer equation. The influence of meteorology on stream temperature is incorporated through, and characterized by, these two parameters which represent all of the weather variables and meteorological conditions that can influence river temperature. The effects of flow depth and discharge on stream temperature, including long-term mean under steady flow and weather conditions, river-reach average, daily mean, daily maximum, and diurnal variation, are investigated by applying the temperature–discharge relationships. The lag between the time of maximum water temperature and the time of maximum equilibrium temperature, to which the diurnal variation and maximum temperature are related, is also examined.

DEVELOPMENT OF RELATIONSHIPS

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

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

where T is the cross-sectional averaged water temperature, t is time, Q is river discharge, A is cross-sectional area, D is mean flow depth in the river, H_f is the net rate of heat exchange across the water surface, S_b is the bottom heat flux, ρ is the water density, c_p is the heat capacity of water, and E is the longitudinal dispersion coefficient. The exchange of heat across the water surface, H_f is one of the most important factors that govern the temperature of a river (Edinger *et al.*, 1974; Brocard & Harleman, 1976; Gu & Stefan, 1985; Sinokrot & Stefan, 1993; Brady *et al.*, 1969). Stream discharge, Q influences river water temperature through advection or dilution. Flow depth, D and river surface area or the water body volume play an important role through thermal inertia and heat capacity of water.

Direct and quantitative relationships between discharge (Q) and river temperature (T) need to be developed in order to isolate the effect of discharge on river temperature from that of a wide range 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 characterizing parameters for weather to be defined in the following section. Two insignificant terms, for which field data are usually not available, are dropped from equation (1), i.e. longitudinal dispersion, E and heat flux across the water–sediment interface. The longitudinal dispersion term can be neglected for a thermally well-mixed stream unless pulse injections of heat and flow, bends and other channel irregularities are to be considered (Brocard & Harleman, 1976; Fischer, 1967). The stream bed acts as an energy sink during mid-day hours over a diurnal period (or summer months in a year cycle) and an energy source at night time (or during winter months). Estimates of bottom heat flux were made from the predictions of temperature in a lake bottom (Gu & Stefan, 1990) and riverbed (Hondzo & Stefan, 1994). The heat flux from water columns to lake bottom during June–September and to riverbed in September were estimated to be about 0.5% and 5%, respectively, of the net solar radiation entering the water bodies in Minnesota, a neighbour state north of Nebraska, USA. Streambed heat flux is further insignificant when a large volume of water is involved in the water body, as predicted by Sinokrot & Stefan (1993) for the Mississippi River, USA, which is about 13 cal cm^{-2} day^{-1} in July and approximately 1% of the shortwave radiation. Therefore, heat flux through the water–sediment interface is considered very small compared to that across the air–water interface (Hockey *et al.*, 1982; Edinger *et al.*, 1974) and is assumed negligible in this study to simplify the problem.

Steady state

Considering steady state flow, the equation for the temperature analysis of flowing water in which a parcel of water is followed along the river in the flow direction is:

$$\frac{Q}{A}\frac{\mathrm{d}T}{\mathrm{d}x} = \frac{H_f}{\rho c_p D} \tag{2}$$

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

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

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

$$T_e = T_d + \frac{R}{K} \tag{4}$$

where R is solar radiation (W m^{-2}) and T_d is dew point temperature (°C). The surface heat exchange coefficient, K (W m^{-2} $°C^{-1}$) is computed from air temperature, wind speed and relative humidity or dew point temperature. Air temperature, instead of both air and water temperature, is used to calculate K so that it is possible to separate the effect of discharge on river temperature from that of meteorological factors. Preliminary analysis of 5-year field data at the central Platte River, to be presented, indicates that the average difference between K values based on air temperature and both air and water temperature is 1.3%.

After substituting equation (3), the solution to equation (2) for a steady state situation, i.e. no change in climatic or environmental variables with time, is obtained as:

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

where $Q^* = Q/C_x$, $T^* = (T - T_e)/(T_e - T_o)$, $C_x = KAx/\rho c_p D$ and T_o is the temperature at the upstream site, i.e. T at $x = 0$. If the normalized discharge, Q^*, is set to zero, the normalized water temperature, T^* is equal to zero, indicating the dominance of weather. When Q^* tends to infinity, T^* approaches −1, showing the dominance of flow.

It is expected that river temperature, T is driven toward its equilibrium, T_e by weather and toward (upstream) inflow temperature, T_o by discharge. Equation (5) also indicates that the influence of upstream flow conditions (Q and T_o) on the downstream water temperature decreases with distance under constant discharge. To investigate the effects of flow and weather on the averaged water temperature over a river reach, T_l, equation (5) is integrated over the length of a river reach, L and the solution is:

$$T_l^* = Q_l^* \left(e^{-\frac{1}{Q_l^*}} - 1 \right) \tag{6}$$

Equations (5)–(6) exhibit two different exponential functions of stream discharge for river temperature under steady state assumption. If T^* is viewed as the dimensionless water temperature at the end of a river reach of length L, the average over the reach, T_l^* is equal to T^* at a location upstream of the reach's mid-point. Equations (5) and (6) indicate that T^* is expected to decrease more slowly than T_l^* does with discharge ($Q^* = Q_l^*$) because the effect of upstream inflow (discharge and temperature T_o) on downstream water temperature is inversely related to distance.

Diurnal variation

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 the two. The long-term average water temperature is expected to be equal or very close to the time-average value of the equilibrium temperature. Hourly mean water temperature usually cross T_e twice each day as warming and cooling trends alternate in response to the diurnal cycles.

Daily maximum water temperature in a diurnal cycle is an important indicator of a water body's response to weather during a diurnal period. It was noted that an aquatic biota could be impaired by the diurnal variations in river temperature as well as the daily maximum water temperature. The daily amplitude of the actual water temperature, i.e. difference between maximum and minimum, could be only a small proportion of the diurnal excursion of the equilibrium temperature. Diurnal shortwave solar radiation alone would cause a large daily variation in equilibrium temperature. The large diurnal variation of air temperature would add to the diurnal amplitude of T_e. The diurnal amplitude of T_e is damped by the surface heat exchange coefficient that also poses a daily variation, a function of wind and humidity.

Considering $x = Qt/A$, equation (2) can be written in an unsteady form:

$$\frac{dT}{dt} = \frac{H_f}{\rho c_p D} \tag{7}$$

to describe the temperature variation of a parcel of water over time in a stationary water body as well as in running waters. Equation (7) can be employed to analyse the response of daily maximum river temperature and amplitude of variation to discharge or depth and 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 (7) gives a sinusoidal variation of water temperature with time. The diurnal amplitude of water temperature is then related to the daily variation of equilibrium temperature by:

$$\Delta T_w = \frac{\Delta T_e}{\sqrt{1 + (D^*)^2}} \tag{8}$$

where $\Delta T_e = (T_{e,\max} - T_{e,\min})/2$ and $\Delta T_w = (T_{\max} - T_{\min})/2$ are the amplitudes of sinusoidal variations of equilibrium and river temperature, respectively, and $D^* = 2\pi\omega\rho c_p D/K$ is dimensionless flow depth in which ω is the frequency of sinusoidal variation of equilibrium temperature with time. Water column depth, D, in a shallow river can be related to discharge using Manning's equation:

$$Q = \frac{1}{n} W D^{\frac{5}{3}} S_o^{\frac{1}{2}}$$

D^* is expressed as:

$$D^* = \frac{C_l^{\frac{1}{2}}}{K}\left(\frac{Q}{W}\right)^{\frac{3}{5}} \tag{9}$$

where W is the top (surface) width of the river and $C_l = (2\pi\omega\rho c_p)^2 n^{6/5} S_o^{-3/5}$ in which n is the Manning's roughness coefficient and S_o is the river bottom slope. Equation (8) suggests that an increase in water column depth or higher flow rate 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.

Time lag of diurnal peak

For a diurnal cycle, the lag between the occurrence of daily maximum equilibrium temperature and the occurrence of daily maximum water temperature is another important parameter characterizing the response of river water temperature to meteorological conditions. Generally maximum T_e occurs near noon due to the strong influence of solar radiation, and water temperature reaches a maximum in late afternoon (around 16:00–18:00 h). Maximum air temperature occurs after maximum T_e and before the time of maximum water temperature. The time lag between the occurrence of maximum equilibrium temperature and maximum water temperature is also related to depth or discharge. From the phase lag $\theta = \tan^{-1} D^*$ $(0 \leq \theta \leq \pi/2)$, in the sinusoidal variation of water temperature, the lag time, t_l for a diurnal cycle (1 day or 24 h) is determined by:

$$\frac{t_l}{t_d} = \frac{\theta}{2\pi} \tag{10}$$

where t_d is the period of time for a diurnal cycle, equal to 1 day or 24 h. The maximum lag is readily seen to be 6 h because θ has a maximum value of $\pi/2$. With larger depth (which requires 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 a greater thermal inertia. The lag time is increased and the maximum water temperature is reduced as depth or discharge increases.

Daily maximum

The relationship between daily maximum river temperature and discharge or depth can be obtained from equation (5) and equation (8). To solve equations (5) and (8) for $T_{\max}$, the diurnal amplitude of water temperature, ΔT_w needs to be converted to $(T_{\max} - T_{\text{mean}})$ through a ratio determined by field measurements:

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

The ratio r_w is expected to be constant for a river and vary with the ratio for equilibrium temperature defined as:

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

Both r_w and r_e are equal to 1.0 for true sinusoidal variation of water and equilibrium temperature in a daily period. Substituting equation (8) into equation (11) and assuming constant K, the analytical solution for daily maximum water temperature is:

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

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

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

where

$$\alpha = C_1 x^{\frac{6}{5}} K^{-\frac{2}{5}} \left(\rho c_p\right)^{-\frac{1}{5}}$$

$$T^*_{\max} = \frac{T_{\max} - T_e}{T_e - T_o}$$

and

$$T^*_{e,\max} = \frac{T_{e,\max} - T_e}{T_e - T_o}$$

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 them during the summer warming and winter (no ice) cooling processes in response to changes in meteorological factors. Daily mean and maximum water temperature are usually below their equilibriums under warm weather and above it on cold days. Higher discharge tends to reduce river temperature during summers and to elevate them during winters through mixing and dilution under steady meteorological conditions or fixed weather factors (T_e and K or H_f) as indicated by equations (5)–(6) and (13)–(14). Only the summer situation is considered in this study for the development of temperature–flow relationships. The effects of discharge and depth on river temperature are analysed and discussed in detail together with validation of the above derived equations against field data for the central Platte River, Nebraska.

VALIDATION AND DISCUSSION

Study area and field data

The study area for validation and application of the relationships extends from Overton to Grand Island, Nebraska, a 125 km section of the central Platte River, downstream of Lake McConaughy and associated Kingsley Dam (Fig. 1). The reach is a wide, small gradient, and shallow sand-bed river with areas of relatively permanent and semi-permanent trees and woody vegetation. Some sections include large forested islands. The upstream sections of the reach are more heavily forested. The average channel width is about 200 m. The river width increases with distance in the downstream direction. The width increases by 20% from Overton to Mormon Island (about 78% of the total length of the reach) and by 42% over the much shorter lower reach (22%) from Mormon Island to Grand Island. The river has a braided reach between Wood River and Grand Island. The bottom slope of the river varies from 0.001 in the lower reach to 0.0017 in the mid-reach with an average value of 0.0013 in the upper reach. The Manning's roughness ranges from 0.02 to 0.05 with a mean of 0.0355 (Dinan, 1992) and it decreases with distance downstream from Overton.

River discharge during summer in the central Platte River is affected by operations at several major hydropower and irrigation projects or storage and diversion structures and by irrigation return flow. About 70% of average annual flow is diverted before reaching the study area. The discharge in this reach ranged from 0.1 to 180 $m^3 s^{-1}$ in 1989–1993 summers with a mean flow of 20 $m^3 s^{-1}$ at which the reach-averaged surface width was 200 m, flow depth was 0.26 m, and cross-sectional mean velocity was 0.38 $m^3 s^{-1}$.

Field data have been collected continuously along the Big Bend reach of the central Platte River for the summers from June to August since 1988. Available data include river water temperature, stream discharge, and climatic parameters: air temperature, solar radiation, humidity and wind speed. Water temperature was measured hourly at Overton, Odessa, Gibbon, Shelton, Mormon Island and Grand Island (Philips). Hourly meteorological data from the nearest weather station were used for each thermograph for which river temperature analysis was performed. Weather stations are located at Lexington, Gibbon, and Grand Island. Daily flow data are from gauging stations at Overton, Odessa, Kearney and Kearney Canal and Grand Island (Fig. 1). The summers of 1989–1993 (460 days) were chosen as the study periods because some records are incomplete for other years.

The meteorological data were analysed to obtain daily mean weather parameters (solar radiation, air temperature, wind speed and humidity) and daily maximum values of solar radiation and air temperature. The values of T_e and K were calculated from hourly values of the meteorological parameters. Daily mean water surface heat exchange coefficient, daily mean equilibrium water temperature, daily maximum equilibrium temperature and corresponding heat exchange coefficient at the time of maximum T_e were obtained from hourly values of T_e and K. Daily discharges at the nearest upstream US Geological Survey gauging station were used for water temperature stations where no flow data were available. Discharge in most streams fluctuates during a warm season. Its daily variation is expected to be significant when a large storm comes. However, daily mean discharges were used to simplify the analysis.

Based on the preliminary analysis of 1989–1993 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. 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 has a larger deviation from the exact sine function than r_w does.

The average heat exchange coefficient, K, is 29.6 W m^{-2} °C^{-1} based on air temperature, and 30.0 W m^{-2} °C^{-1} using both air and water temperature. The equilibrium temperature, T_e, is 24.5°C. The average river temperature, T, is 23.5°C. These values were computed from the observed weather variables and water temperature during the 1989–1993 summers. The average amplitudes of diurnal (sinusoidal) variations of river temperature and their equilibria were obtained as ΔT_w = 6°C and ΔT_e = 15°C for the five summers. Broad variations of meteorological conditions exist over a summer from June to August. The ranges of the daily means (K, T_e and T) were 17–76 W m^{-2} °C^{-1}, 11–36°C, and 13–30°C respectively in the 1989–1993 summers. The daily maxima were $T_{e,\max}$ = 15–56°C and $T_{\max}$ = 14.7–38°C with five-summer means of 42 and 29.4°C, respectively.

Time-average water temperature

Equation (5) computes water temperature at various river points for given discharges and longitudinal distances. Equation (6) displays the relationship between discharge and river temperature averaged over a river reach. Because both equations were obtained with the assumption of steady state conditions (flow, weather, and upstream flow temperature), the periods of measurements during which discharges were relatively constant or steady need to be selected for validation of the analytical solutions (equations (5) and (6)). All durations chosen for analysis should be equal to or longer than the travelling time of flow from the upstream boundary to the point considered. Based on preliminary computations of river velocity using Manning's equation and available width–discharge relationships, it was estimated that it took approximately 3 days for a water parcel to travel from Overton to Grand Island under an average flow rate of 20 $m^3\ s^{-1}$ over the 1989–1993 summers with an average channel bottom slope of 0.0013 and a typical Manning's roughness coefficient of 0.03 for the central reach of Platte River. In total twelve periods, ranging from 4 to 71 days with an average of 16 days, were chosen after viewing the flow history of the river in the summers of 1989–1993. Measured water temperature (T_o and T), discharges (Q), and calculated meteorological parameters (K and T_e) were averaged over each of the selected periods to obtain mean values so that equations (5) and (6) could be applied to the analysis of water temperature variations with discharge and distance.

Figure 2 shows measured and computed river temperature at Shelton averaged over each of the periods in a dimensionless form. It is seen that the observed variation of normalized temperature (T^*) with discharge (Q^*) followed the exponential decay described by equation (5). For the average over the entire reach of central Platte River (Overton to Phillips) during the 1989–1993 summers, theoretical relationship between river temperature and discharge (equation (6)) is also validated against field measurements (Fig. 3). It is shown that the dimensionless river temperature for the reach presented in Fig. 3 is lower than that plotted in Fig. 2 for Shelton under constant flow and weather conditions. This is because, for an exponential increase of water temperature with distance, the water temperature averaged over the reach is equal to that at or near Odessa which is upstream of Shelton. It is shown in Figs 2 and 3 that the reductions in time- and length-average river temperature become less significant when discharge is very large ($Q^* > 4$ and $Q_1^* > 2$). As

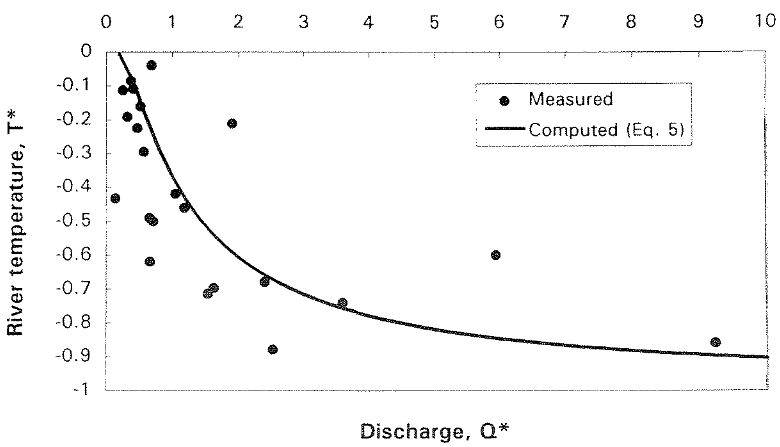

Fig. 2 Time-average river temperature *vs* discharge at Shelton.

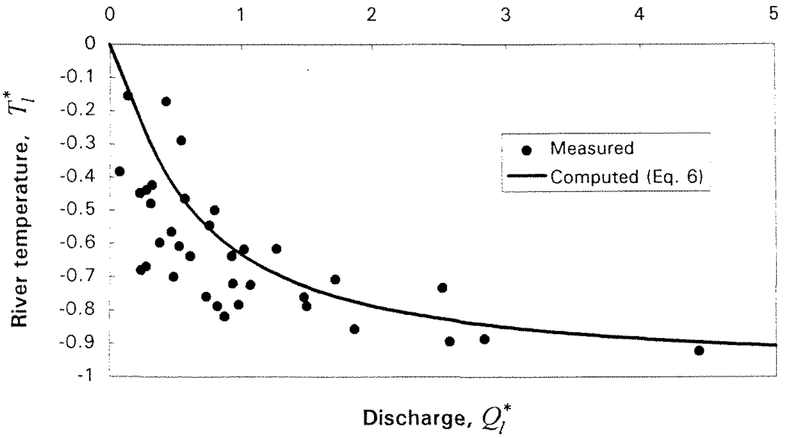

Fig. 3 Variation of river temperature with discharge averaged over the Big Bend reach (from Overton to Phillips).

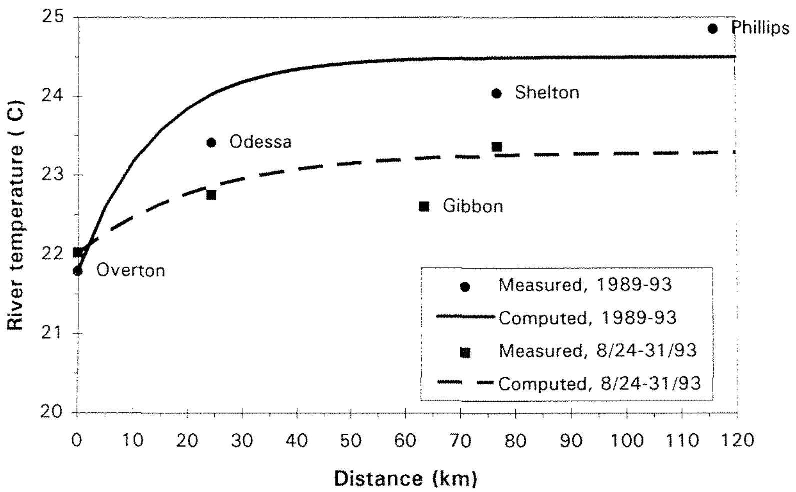

Fig. 4 Variation of time-average water temperature with river distance.

indicated by equations (5) and (6), discharge plays a lesser role in moderating water temperature downstream at a longer distance. The scatter of data points in Figs 2 and 3 may be attributed to factors such as shading, lateral variation, groundwater exchange and intermittent runoff during storm events which are not taken into account in the analytical solution method.

Longitudinal distributions of time-average river temperature in the river reach from Overton to Phillips computed by equation (5) and measured during two periods, 24–31 August 1993 and all five summers of 1989–1993, are plotted in Fig. 4. River temperature measurements at Overton were used as the upstream inflow temperature for the computational domain from Overton to Shelton (T_o = 21.79 and 22.02°C for 1989–1993 summers and 24–31 August 1993, respectively). Weather conditions at

Lexington and Gibbon were averaged for the reach. The values are T_e = 23.29 and 24.5°C, K = 26.39 and 29.6 W m^{-2} °C^{-1}, and Q = 30 and 20 m^3 s^{-1}, for 24–31 August 1993 and the entire 1989–1993 (summer) period, respectively. Surface width ($W = A/D$) was computed using empirical width–discharge relationships presented by Dinan (1992). The measured and computed river temperature profiles along the river reach exhibit an exponential variation of summer water temperature with downstream distance (x) under constant T_o (flow temperature at Overton), T_e, K, and discharge as described by equation (5). Comparison of field data and analytical solutions for variation of water temperature with river discharge illustrates that, when applied over a long-term mean, equation (5) agrees reasonably well with the measurements. The accuracy of prediction may be improved by using variable slope, roughness and cross-sectional geometry for different sections of the reach and applying equation (5) section by section. The results in Fig. 4 indicate that water temperature increases as one moves downstream to Grand Island where it approaches its equilibrium.

Diurnal amplitude

As expressed by equation (8), flow depth moderates the diurnal amplitude of water temperature variation, i.e. the difference between daily maximum and minimum. The ratio of diurnal amplitude of measured river temperature to that of equilibrium at Grand Island is plotted against the normalized flow depth in Fig. 5, together with the theoretical curve (equation (8)). Flow depth is calculated from discharge data using the Manning's equation. The general trend of the data seems to follow equation (8). The standard error of prediction is 0.166. The mean value of the measured ratios is 0.366.

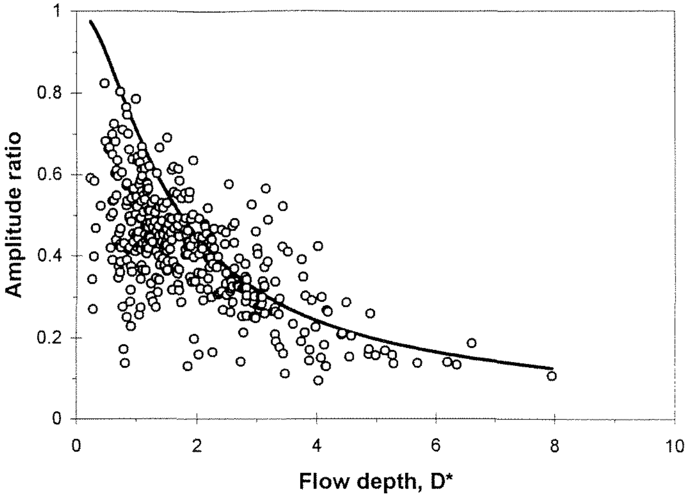

Fig. 5 Measured (circles) and predicted (line) ratio of amplitude of diurnal river temperature variation of that of equilibrium temperature at Grand Island for 1989–1993 summers.

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.

Comparison of field measurements and mathematical computations shows a reasonable agreement although the theory overpredicts the ratio and more scatter of the data is seen at low flows or small depths. The overprediction and scatter are probably caused by the omission of some heat components for which data are not available and some uncertainties in conditions under which field measurements of river temperature were taken. Under low flow conditions, the contribution of groundwater with low temperature may not be negligible. In a shallow and clear water body, a portion of solar radiation, approximately 1–5% (Sinokrot & Stefan 1993; Hondzo & Stefan 1994), may reach, and be absorbed in, bottom sediments. 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 gains heat from sediments during night time, 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 due to unavailability of cross-sectional profiles. This simplification is valid for wide, shallow channels, which is the case if the river has broad floodplains under large discharges. Water temperature was measured at the main channel of the river, while the computed temperature values represent the laterally-averaged values over an entire cross section. According to equation (8), shallow waters have higher temperatures than deeper flows in the main channel. Therefore, lateral variation in river temperature may also contribute to the discrepancy and scatter in Fig. 5. A more accurate prediction may be obtained by means of a stream tube concept, i.e. making computations for the main channel and the shallow overflow areas separately. Moreover, the effects of shading conditions may become more significant during low flows due to increased foliage in summer (warmer) months.

Application of equation (8) to the central Platte River at Shelton indicates a strong relationship between the diurnal water temperature fluctuations and flow depth–discharge. The theoretical line (equations (8) and (9)) corresponds to the average condition, K = 30 W m^{-2} °C^{-1}. Figure 5 clearly shows a decay in diurnal river temperature amplitude with flow depth or discharge. It takes a longer time for a large volume of water to be heated or cooled than a small one with equivalent surface area. The thermal inertia of water is increased by high discharges, leading to fewer water temperature rises during hot daytime and falls during cool night time. The higher spread of the maximum and minimum temperature around the mean water temperature occurs at lower discharges because of the reduced depth, and lower thermal inertia. It is, therefore, expected that discharge has a stronger impact on daily maximum river temperature (T_{max}) than on daily mean (T_{mean}), i.e. T_{max} drops faster than T_{mean} when flowrate increases. This is because daily maximum and minimum temperatures are driven toward the daily mean by elevated flow while daily mean water temperature is driven toward its equilibrium by weather. As indicated by equations (5) and (6), river temperature is more influenced by weather (mainly air temperature and solar radiation) under low flows than under high flows. It is shown in Fig. 5 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 with D^* in the range 0–5 is approximately eight times that by flow in the range of D^* = 5–10.

Daily peak and time lag

Presented in Fig. 6 are measured and calculated dimensionless daily maximum water temperature values and their variations with the normalized discharge at Shelton during 1989–1993 summers under the weather conditions represented by $T^*_{e,\max}$ = 3–12 (mean 6.8). Given the fixed range of the reference parameter $T^*_{e,\max}$, water temperature values computed using equation (14) follows the trend of observed values and the band of data points agree well with that defined by the theoretical lines (equation (14)). The mean values of r_w/r_e and K were used in the computation. The significant moderation effect of high discharge on daily maximum river temperature is shown in Fig. 6. The theoretical curve for $T^*_{e,\max} = 0$ corresponds to the relationship described by equation (5), i.e. T^*_{mean} *vs* Q^* curve (Fig. 2). Figure 6 indicates that $T^*_{\max}$ drops faster with increasing Q^* than T^*_{mean} does. It is seen that the daily peak of river temperature is attenuated by high discharges and the amplitude of water temperature is narrower under a high discharge than under a low flow. Although daily mean water temperature may not change much, approaching its equilibrium, daily maximum river temperature under a high flow is expected to be significantly lower than that under a low flow. This reduction is because of greater thermal inertia resulting from elevated discharge and a greater corresponding depth.

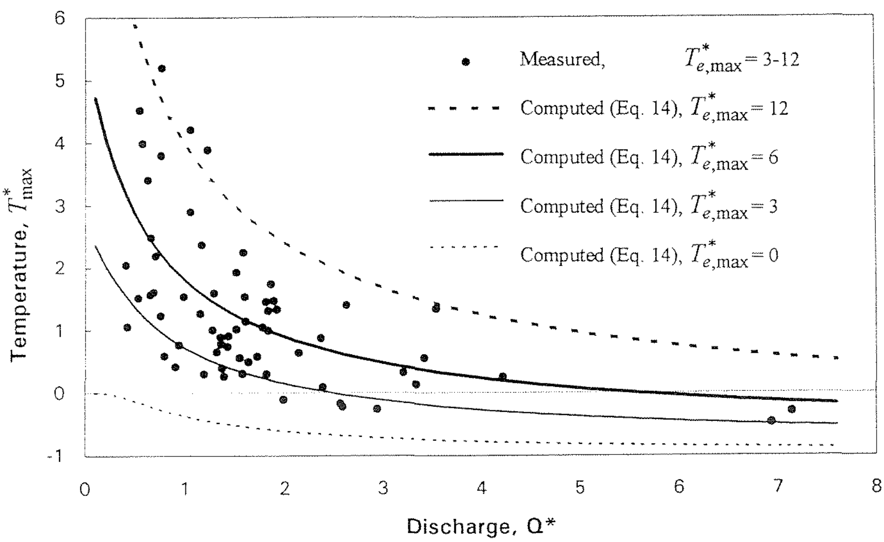

Fig. 6 Variations of daily maximum water temperature with discharge for different daily maximum equilibrium temperature at Shelton.

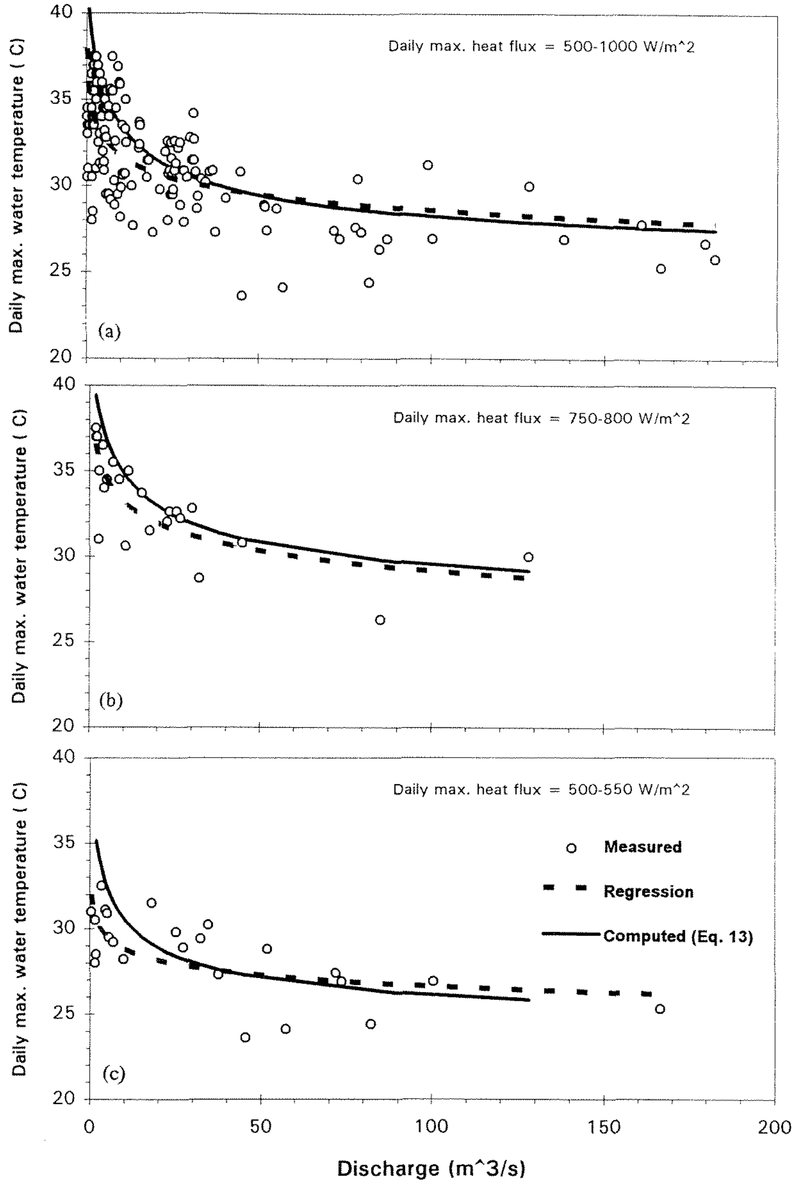

Fig. 7 Measured and computed daily maximum river temperature values and their variations with discharge at Grand Island (1991–1993) for three $H_{f,\max}$ ranges.

Daily maximum river temperature, $T_{\max}$ measured at Grand Island during the 1991–1993 summers and computed using equation (13) and the variation with discharge are presented in Fig. 7 in dimensional forms. To separate the impact of discharge, Q on $T_{\max}$ from that of weather, each of the $T_{\max}$ *vs* Q relationships was developed for those days having the same or similar weather conditions. The daily maximum heat flux from atmosphere to river water columns ($H_{f,\max}$) was used as the sole reference parameter which collects all weather factors and reflects main heat transfer processes. Values of $H_{f,\max}$ were calculated from hourly weather data and a constant river temperature, i.e. a long-term average (24°C at Shelton and 24.7°C at Grand Island) to exclude the effect of water temperature. The relative values of $H_{f,\max}$, which varied from -250 to 1200 W m^{-2}, served to characterize and sort the wide range of meteorological conditions (hot to cold). As displayed in Fig. 7(b) and (c), higher $H_{f,\max}$ resulted in higher $T_{\max}$. Reducing the $H_{f,\max}$ range from 500 W m^{-2} (Fig. 7(a)) to 50 W m^{-2} (Fig. 7(b) and (c)) lead to less scatter of the data points for the measured $T_{\max}$. Although predictions can be made for individual days by equation (13), a temperature-discharge relationship can be more clearly seen by plotting $T_{\max}$ at variable discharges under a constant weather condition. The values of T_o, T_e, and K were averaged over days selected to obtain the predicted $T_{\max}$ *vs* Q relationships. Each of the regression curves (using power function) for the field data and the prediction curves in Fig. 7 represents an average relationship over a given range of weather conditions ($H_{f,\max}$). An overall agreement between predictions and measurements is exhibited in Fig. 7, though equation (13) overpredicted $T_{\max}$ at low flows ($Q < 20$ m^3 s^{-1}). The value of $T_{\max}$ decreases with discharge as indicated by a sharp drop of $T_{\max}$ when $Q < 30$ m^3 s^{-1} and a slow reduction when $Q > 30$ m^3 s^{-1}. Slightly higher $T_{\max}$ values are also seen at Phillips than at Shelton. The critical discharge (Q_c) can be used to evaluate the minimum flow ($Q_{\min}$) for maintaining a given temperature standard in decision-making on selection of streamflow management with respect to feasibility, effectiveness, and achievability. If $Q_{\min} > Q_c$, Q_c should be used as the $Q_{\min}$ from economic consideration because increase in Q ($>Q_c$) does not effectively reduce $T_{\max}$.

Sinusoidal variations of equilibrium and water temperature are exhibited by

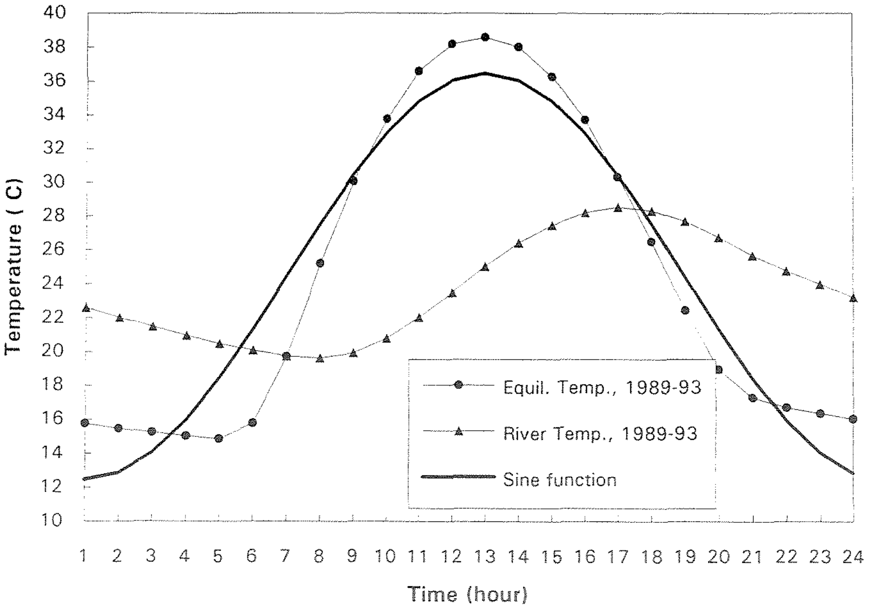

Fig. 8 Diurnal variations of water temperature and equilibrium temperature at Shelton.

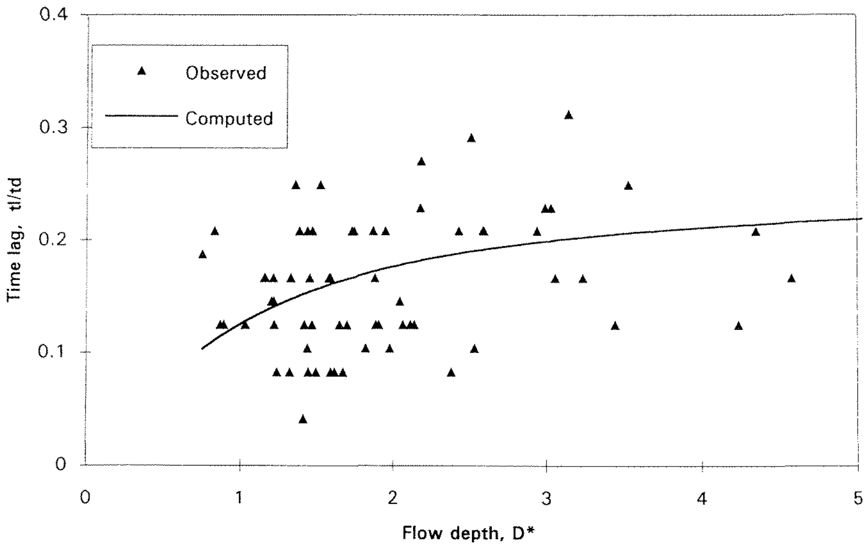

Fig. 9 Time lag between the occurrence of daily maximum equilibrium temperature and water temperature *vs* discharge at Shelton during summer 1992.

measurements at the central Platte River. Figure 8 shows the 5-summer averages of diurnal variations measured at the Shelton station, which are described and followed reasonably well by a sine function of time as assumed in the derivation of equation (8). It is also seen that the amplitude of diurnal water temperature variation proportional of that of equilibrium temperature. On a typical day during the 1989–1993 summers at Shelton on the Platte River, the daily minimum water temperature occurred in the early morning (7:00–9:00), the daily peak appeared in the late afternoon (16:00–18:00), and the equilibrium temperature peaked at about 13:00. Displayed in Fig. 9 are the observed and calculated time lags between the occurrence of daily maximum water temperature and daily maximum equilibrium temperature at Shelton for summer 1992. It can be seen that a larger discharge, through a deeper flow depth, resulted in a longer time lag and in turn a lower daily peak water temperature. The increase in time lag by elevated flow becomes less significant when the discharge is very large. Based on the analysis of the 1992 data, the average occurrence time of daily maximum temperature at Shelton was about 13:00 for equilibrium, 16:00 for air, and 17:00 for water. The average lag was approximately 3 h between daily maximum equilibrium and air temperature and 4 h between equilibrium and river temperature.

CONCLUSIONS

The effects of instream flow discharge and depth on summer river temperature were successfully separated from those of meteorological conditions and quantified by using a simplified model and treating weather as a reference. The theoretical relationships between water temperature and river discharge/depth developed in this study with an analytical solution method were validated against 5-summer field data at the central Platte River for daily maximum, daily mean, diurnal amplitude, time lag, steady state river temperature, variation with distance, and average over a river reach. The reasonably good agreement between computational results and field measurements indicated the potential usefulness of the method for quantitatively describing the temperature–discharge relationships.

It was shown by the analytical solutions and field observations that river water temperature was significantly affected by discharge or flow depth and that an increase in discharge caused a reduction in summer river temperature under given weather conditions. The impact of discharge on water temperature occurred through the associated increase in depth and then in the thermal inertia of river water column, which lead to a longer lag time between the occurrence of the daily maximum equilibrium temperature and river temperature. Instream discharge primarily influenced the amplitude of diurnal water temperature variation and lowered the daily maximum river temperature. The daily maximum and minimum water temperature values were driven towards the daily mean or zero amplitude by increased discharge or depth, whereas the role of weather factors was to drive water temperature towards its equilibrium. The influence of upstream reservoir temperature was limited to a short distance on the river downstream. It was evident that the daily peak temperature was reduced and the daily minimum temperature was increased by elevated discharges.

It is concluded that the influence of instream discharge on summer river temperature became less significant when discharge or depth was very large, i.e. $Q^* > 4$, $Q_l^* > 2$, and $D^* > 5$. A critical discharge for a specific river was obtained, e.g. $Q = 30$ m^3 s^{-1} for the Platte River, below which river temperature dropped dramatically with an increase in discharge or depth, whereas above which temperature varied slowly. Temperature reduction due to elevated flow asymptotically approached a constant. The critical discharge can be used as a criterion for instream flow requirement to protect aquatic biota. The relationships between river discharge and water temperature imply that there may be an opportunity to moderate high river temperature with streamflow management by choosing appropriate flowrate of water release from a reservoir. It is suggested that implementing summer river temperature control through streamflow management is feasible. The existence of the critical discharge makes it possible to reduce or control the occurrence of daily maximum water temperature exceeding a standard at a river reach by increasing discharge to an achievable level.

The validated relationships can be used to evaluate, based on historical flow data and weather conditions, the effects of instream flow on river temperature and their significance. They can also be employed to predict possible consequences of future river and reservoir management scenarios as the impacts of various discharge levels on river temperature and in turn on aquatic environment and wildlife are considered. The predictions of temperature variations due to discharge changes can provide information useful for the control of summer river temperature through streamflow regulations. The method and results can serve as an instrument to be used as a needed environmental aid in the decision making process required for proper planning and design of reservoir operations and river management, e.g. reservoir water release and streamflow diversion.

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