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PREDICTING THE NATURAL WATER TEMPERATURE PROFILE THROUGHOUT A RIVER BASIN

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ABSTRACT

A one-dimensional mathematical model is presented for predicting natural water temperatures throughout an entire river basin. The model contains an accumulation term, convective term, dispersive term, and source term. Mathematical representations of each term contributing to the heat flux are presented.

This model can assume major importance in the planning and management of a river basin. Natural water temperature and the effects of alteration of this temperature regime have a large effect on the ecology of the entire river basin.

This model is solved numerically utilizing extensive synoptic meteorological and flow rate data obtained for an entire river basin. This solution is compared to experimental data. The results indicate that calculated water temperatures can be obtained which typically have a daily root-mean-square deviation of less than 3°C, a daily amplitude ratio of ± 0.20 of 1.0, and a daily mean within 2°C of observed data. These results can be improved with improved measurement of incoming radiation and average depth. The effects of the average depth, convection, and dispersion are also discussed.

INTRODUCTION

Temperature is a significant water quality parameter. Accurate analysis of the temperature regime in a river or stream is of the utmost importance. Most

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physical properties of water, including dissolved oxygen and dissolved solids content, depend on temperature.

Temperature regimes in river systems can be distinguished as being natural or altered regimes. The natural temperature regime is that which exists or would exist in the absence of man-made alterations. Altered temperature regimes encompass the remaining situations where man has made changes which affect river temperature. Some of the common alterations which impact the temperature regime in a river system are channelization or siltation, impoundment, irrigation, vegetation removal (particularly riparian) and industrial discharges [1].

The difference between the altered temperature and the natural temperature at a point depends on the severity of the alteration. In extreme cases, it may be as great as 30°C. The ecology of a river system can be greatly influenced by these alterations.

The prediction of natural temperature and the effects of alteration of the natural temperature, therefore, can assume major importance in the planning and management of a river basin. A model which predicts natural temperatures can be utilized to determine the natural temperature profiles of rivers and streams already altered and determining the extent of that alteration. Many schemes intended to control the extent of that alteration call for maintaining temperature within specified limits of the natural temperature. Where alterations already exist, a basin model is required to predict natural temperature.

Previous models have been specific in nature and this has been dictated by the small amount of data available. They are specific in that they only can be applied to segments or reaches of a river system. The time-temperature relationship for the inlet to each reach must be known, and then a profile can be predicted at the outlet. The procedure must then be repeated for any subsequent reach. Some models are further restricted in that they only analyze monthly, seasonal, or annual variations.

Kothandaraman has developed a model to predict daily mean temperature of large rivers [2]. He assumes the water temperature can be written in the form of a time dependent Fourier series with three fitted residual terms. He uses four years of mean water temperature data to calculate three series coefficients and four years of mean air temperature data to estimate the three residual coefficients. He uses the model to compute daily mean water temperatures for 1969, one of the four years for which he had data, and gets agreement within (1.0°C). This model is empirical in nature and it cannot predict the resulting water temperature from any alteration to the basin since the coefficients must be determined from existing conditions.

A model for predicting the hourly temperature of small streams has been developed by Brown [3]. The model assumes the time rate of change of temperature is given only by an energy source term. The model neglects convection and dispersion. Calculated results were with (0.5°C) of observed values on two streams. Comparison was limited to one twenty-four hour period.

Raphael [4] and Delay and Seaders [5] propose a general model for rivers and reservoirs analogous to Brown's model with the addition of accounting for tributary inflows. The model also assumes complete mixing, which is rarely encountered in reservoirs. Delay and Seaders calculated a river temperature for one twenty-four hour period and it was within (0.5°C) of the observed data [5].

Morse developed a model for predicting hourly temperatures, which has the same features as Brown's model and also includes a convective term [6]. The model is based on observations along a 14.0 km (8.77 mile) stretch of a tributary of the Columbia River. The calculated results are compared to observed results for a twenty-eight hour period and agree within (0.5°C).

Edinger has developed models for predicting hourly temperatures based on the concept of an equilibrium temperature [7]. This is the temperature the system would ultimately reach if allowed to go to steady-state with constant meteorological conditions. The purpose of this model is to linearize the water temperature dependence of the surface heat exchange terms. The surface heat exchange terms are approximated by the product of a coefficient based on meteorological variables and the difference between the actual and the equilibrium temperature. Therefore, the coefficient and the equilibrium temperature become parameters which must be evaluated at each computational step. The model neglects convective and dispersive effects.

Chen theoretically derives a basin model for hourly temperature prediction which accounts for tributary inflows [8]. The model contains spatial variation of temperature due to convection and dispersion and the heat exchange terms are approximated by heat transfer coefficients and linear temperature differences. No verification of the model using experimental data was attempted.

Jobson and Yotsukura developed a reach model for hourly temperature prediction which contains spatial variation in temperature due to convection and dispersion [9]. They employ a non-linear heat exchange term. Comparison of calculated results with observed data at five different sites for a twenty-four hour period yields results with the maximum deviation being (0.5°C).

DESCRIPTION OF THE GOVERNING DIFFERENTIAL EQUATION

Equation 1 is the one-dimensional thermal energy equation which describes the spatial and temporal water temperature variation in response to time dependent meteorological conditions. This equation has been derived elsewhere [9].

$$\frac{\partial T}{\partial t} + V\frac{\partial T}{\partial x} = D\frac{\partial^2 T}{\partial x^2} + \frac{H}{\rho Cpd} \tag{1}$$

where T = temperature, t = time, x = longitudinal space coordinate, D = axial dispersion coefficient, H = net heat flux at the water surface, ρ = density of water, Cp = isobaric heat capacity of water, d = depth of the river or stream, ν = axial stream velocity.

The net heat flux at the water surface H is composed of many terms. The net non-radiative heat flux, H_{NR}, at the water-air interface is written as

$$H_{NR} = -H_E - H_C - H_A. \tag{2}$$

The net radiative heat flux, H_R, is written as

$$H_R = H_I - H_{BR} \tag{3}$$

where H_I = total incoming heat flux which passes through the water surface, H_{BR} is the long-wave radiation emitted by the water surface, H_E is the heat flux due to evaporation, H_C is the heat loss due to conduction, and H_A is the heat flux due to water movement across the air-water interface. The net heat flux H is written as

$$H = H_{NR} + H_R. \tag{4}$$

H_{IT} is the total incoming radiation flux above the river. A portion of this is reflected at the surface and additionally, some portion may be blocked due to shading. The total incoming radiation flux, H_{IT}, consists of directional and diffuse short-wave solar radiation H_{SW} and long-wave atmospheric radiation H_{LW}. (Thus $H_{IT} = H_{SW} + H_{LW}$.) A certain amount of both the short-wave and long-wave components will be reflected at the air-water interface. In the Lake Hefner studies it was found that the empirical relation

$$R = a\,\theta^b \tag{5}$$

was satisfactory to describe the reflectivity R of H_{SW} as a function of the solar attitude above the horizon θ and two constants a and b [10]. For clear skies, the values are a = 1.18 and b = -0.77 for $10^\circ < \theta < 80^\circ$.

The Lake Hefner studies found that the reflectivity of a water surface to long-wave radiation is 0.03 and independent of temperature over the range 0°-30°C [10]. Utilizing Kirchoff's low and the fact that the absorptivity equals one minus the reflectivity, they stated that the absorptivity of a water surface is 0.97. Therefore,

$$H_I = S \times (1.0\text{-}R)\, H_{SW} + 0.97\, H_{LW} \tag{6}$$

where S = the fraction of short-wave radiation not blocked by vegetation or land features.

In deriving Equation 1, the point temperature has been averaged with respect to vertical and transverse coordinates. As such, the model assumes that the net energy flux at the air-water interface is mixed uniformly in the vertical direction. Thus H appears as a source form in the differential equation rather than as a boundary condition.

The heat loss due to evaporation, H_E, is estimated by

$$H_E = (A + BU)(VP_{SR} - VP_{AIR})\rho\lambda \tag{7}$$

which is the Dalton form of the evaporation equation and the most widely used is practice [11]. A and B are constants, U the wind velocity, ρ is the density of water, λ the latent heat of vaporization for water, VP_{SR} the saturated vapor pressure of water at the river temperature and VP_{AIR} the actual vapor pressure of water in air.

The saturated vapor pressure of water was calculated from the empirical relation

$$VP_{SR} = 54.721 - \frac{6788.6}{T} - 5.0016 \ln T \tag{8}$$

where the vapor pressure is in millibars and the temperature T is in K [9].

The actual vapor pressure of water in air VP_{AIR} is calculated by first obtaining the saturated vapor pressure of water at the wet-bulb air temperature using Equation 8 and then using the following psychometric equation [12]:

$$VP_{AIR} = VP_{SWB} - 6.6 \times 10^{-4} \times P \times (T_{DB} - T_{WB}) \times (1.0 + 1.15 \times 10^{-3} \times T_{WB}) \tag{9}$$

where P = atmospheric pressure (mb), T_{DB} = dry-bulb temperature (°C), T_{WB} = wet-bulb temperature (°C), VP_{SWB} = saturated vapor pressure of water at wet-bulb temp. (mb).

The flux describing the conduction of heat between the air and water H_C is calculated from H_E and Bowen's ratio [13] as

$$H_C = \gamma \frac{(T - T_{AIR})}{(VP_{SR} - VP_{AIR})} H_E \tag{10}$$

where

$$\gamma = 0.61 \frac{P}{1013}. \tag{11}$$

Here, the pressure P is in millibars.

The advected energy flux H_A is usually very small and can be neglected except during rainfall.

As noted previously, the long-wave emissivity of a water surface is estimated to be 0.97. Emission of long-wave radiation from the water surface follows the Stefan-Boltzmann relation

$$H_{BR} = 0.97\, \sigma T^4. \tag{12}$$

Because of the "cold-skin" effect at the surface, the outgoing long-wave radiation may be over-estimated by using the bulk-averaged temperature in Equation 12. With the "cold-skin" effect, the actual surface temperature may be as much as 2-3°C lower than the bulk temperature due to high evaporation and conduction losses. Since the error in H_{BR} is on the order of 2-3 per cent, this is considered acceptable.

The boundary conditions in the axial direction are as follows.

1. The temperature of the water entering the headwaters of the river is the groundwater temperature.

$$T(o,t) = Tg. \tag{13}$$

This temperature is constant on a daily basis but can vary slightly on a seasonal basis.

2. Near the mouth of the river, the temperature is not a function of longitudinal position

$$\frac{\partial T(L,t)}{\partial x} = 0. \tag{14}$$

where L = river length.

This implies that the effect of tributary inflows at this point is very small and identical meterological conditions prevail over the final reach. If convective effects are large with respect to dispersive effects, this would not be valid.

The initial condition required is a complete axial temperature profile at the initial time. For this study we chose to approximate the initial axial profile using measured temperatures and linearizing interpolation between measured points. When initial temperature profile data is not available, one can assume that the entire river is initially at the groundwater temperature. When using this condition, one must be careful to utilize only that part of the solution at time t which is upstream of the point which a particle moving from the source at the average stream velocity has reached by time t.

To account for tributary inflows, an energy balance at the intersection of the tributary and main stream yields

$$Q_u T_u + Q_T T_T = (Q_u + Q_T)\, T_{RE} \tag{15}$$

where Q_u = volumetric flowrate of the upstream reach, T_u = water temperature just upstream of the tributary inflow, Q_T = volumetric flowrate of the tributary, T_T = water temperature of the tributary, T_{RE} = water temperature at reach entrance.

The solution of Equation 1 must be accomplished on a reach by reach basis. The predicted temperature at the end of a reach (point of tributary inflow) must be employed in Equation 15 together with the tributary inflow temperature to calculate the upstream boundary condition for the next reach.

METHOD OF SOLUTION

Equation 1 cannot be solved analytically due to the nonlinearity of H and, therefore, an approximate solution scheme must be developed. Numerical dispersion can be accounted for by utilizing a numerical dispersion coefficient D_p reported by Bella and Dobbins [14]. For regular geometry and constant parameters

$$D_p = \frac{V}{2}(\Delta x - V\,\Delta t) \tag{16}$$

Referring to Equation 16, D_p is substracted from the actual dispersion coefficient in the solution to negate the numerical dispersion effect. It is important to note that this holds only for $\Delta x \geqslant V\,\Delta t$.

Using a finite difference technique, with the subscript i referring to spatial position and j referring to time, Equation 1 becomes, after rearrangement

$$T_{i,j+1} = \left[\frac{(D-D_p)\Delta t}{(\Delta x)^2} + V_x \frac{\Delta t}{\Delta x}\right] T_{i-1,j} + \left[1.0 - V_x \frac{\Delta t}{\Delta x} - 2(D-D_p)\frac{\Delta t}{(\Delta x)^2}\right] T_{i,j} + \left[(D-D_p)\frac{\Delta t}{(\Delta x)^2}\right] T_{i+1,j} + \frac{H}{\rho C_p d}. \tag{17}$$

A computational procedure is required to deal with tributary inflow. For any time step j, let i+1 be the first spatial node point downstream of a major tributary inflow. Let $T_{T,j}$ be the temperature of the tributary, Q_T be the volumetric flow rate of the tributary, and Q_i the flow rate of the river at node point i immediately upstream of the tributary. Assuming the mixing process to be rapid and complete, one can write a thermal energy balance between node points i and i+1 and solve for $T_{i+i,j}$

$$T_{i+1,j} = T_{i,j} + \frac{Q_T}{Q_i + Q_T}(T_{T,j} - T_{i,j}). \tag{18}$$

Heat flux terms which contain the river temperature may be evaluated explicitly using the river temperature calculated at that node point for the previous time interval.

The meterological data was collected on an hourly basis. A linear relationship was assumed between hourly points and a weighted average was used such that each meteorological data point at time period j was the average over the time from j-1 to j.

Experimentally observed values of flow rate Q, average velocity V, and dispersion coefficient D were fitted to an equation of the form

$$Q = Ee^{-FJ} \tag{19}$$

where J = Julian date, E and F are constants and $197 < J < 240$.

This form was utilized since the flows tended to diminish slowly with time during this time period and this is better expressed as an exponential decay than a linear one. E and F were determined from a linear regression of ln Q vs. J.

The average reach depth d can be defined by

$$d = \frac{Q\, t_t}{A_s} \tag{20}$$

where A_s is the surface area of the reach and t_t is the time of travel for the reach. This equation works well for a well-defined channel where A_s can be evaluated fairly accurately and Q and t_t are constant throughout the reach. This is usually the case for large rivers.

Where possible, average reach depths were computed using Equation 20. Since this was not possible for the upstream two reaches were time-of-travel data was unavailable, an alternative method for evaluating depth was employed. In addition to problems evaluating d for some reaches, it was not possible to calculate S, the fraction of solar radiation not blocked by vegetation or land features, for any reach. Therefore both d and S were calculated for each reach by fitting values which gave the best prediction of observed river temperature during a three day period.

The values of d and S are calculated for the furthest upstream reach first and then for each reach, in turn, as one moves downstream. This is done since the water temperature in an upstream reach can affect the calculated value in a downstream reach through convective and dispersive effects.

In calculating d and S for each reach, Equation 17 with boundary conditions and initial condition specified previously is solved from the upstream end of the river to a preselected downstream end point.

This preselected endpoint is the point immediately downstream of the downstream tributary boundary of the reach if there is a temperature monitoring station there. This can be used because Equation 18 relates the water temperature downstream of a tributary to the water temperature above that tributary, which is in the reach under consideration. Where this is not possible, the endpoint is selected as the point immediately upstream of the downstream tributary of the reach under consideration.

The water temperature calculated at night is a function of d, but not S since Q_{sw} = o. Therefore, the value of the temperature change during this period is a function of d alone.

An initial value of d (d_1) is chosen (based on an estimate from Equation 20 where possible). The solution is then propagated for the three-day period. The actual value of the change in water temperature during the night AT is recorded as is the calculated value CT. The value of $\frac{AT}{CT}$ is then averaged over the three-day period.

$$\frac{AT}{CT} = \frac{(\text{maximum observed T} - \text{minimum observed T})\ \text{night}}{(\text{maximum calculated T} - \text{minimum calculated T})\ \text{night}} \tag{21}$$

$$m = \frac{\sum_{\ell=1}^{3} \left(\frac{AT}{CT}\right)_\ell}{3}. \tag{22}$$

The value of d used is

$$d = \frac{d_1}{m}. \tag{23}$$

Using the above value of d in Equation 17, the value of S is then calculated in the following manner. S is varied by 0.01 and for each S, the following equation is calculated for each day of the three-day period

$$\emptyset = \frac{\sum_{\ell=1}^{24} (TO-TC)^2}{24} \tag{24}$$

where $\emptyset$ = root-mean-square error in calculating T for that day, TO = hourly observed water temperature, and TC = hourly calculated water temperature at a given point in the interval under consideration. The three daily values of $\emptyset$ are then averaged. This procedure is repeated for different values of S until a minimum value of the average $\emptyset$ is calculated. The value of S which corresponds to this minimum value of $\emptyset$ is chosen.

The data utilized in this study were obtained in the Mattole River basin in northern California for a period July 16, 1975 to August 31, 1975. Figure 1 shows the major features of the river basin as well as the location of water temperature and meteorological monitoring stations within the basin. All data collected in this study is available elsewhere [15].

DISCUSSION OF RESULTS

The results of model calculations will be analyzed by comparison with the observed water temperature. The three quantities utilized for this analysis are the daily root-mean-square residual as calculated by Equation 24, the daily mean temperature defined as

$$T_{M_i} = \frac{\sum_{j=1}^{24} T_{i,j}}{24} \tag{25}$$

and the daily amplitude ratio defined as

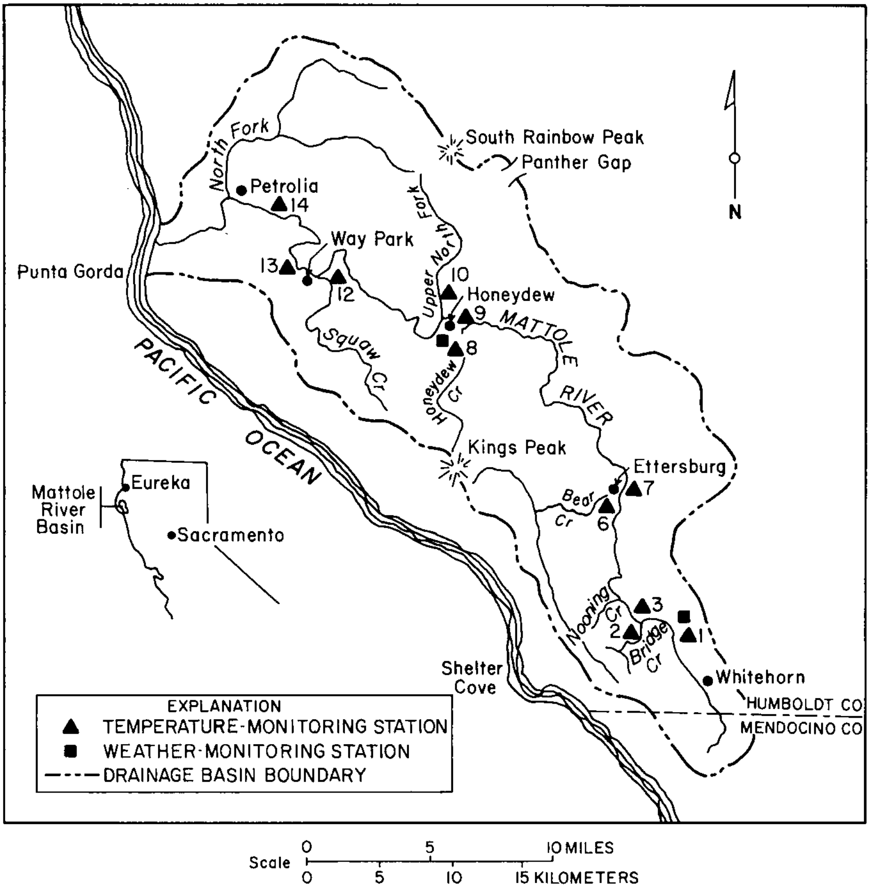

Figure 1. Mattole River basin.

$$AR = \frac{\text{maximum calculated T} = \text{minimum calculated T}}{\text{maximum observed T} = \text{minimum observed T}} . \qquad (26)$$

It is desirable for a model to have a small root-mean-square error. The magnitude of this value can be affected by timing errors in the temperature monitoring system which would offset the observed temperature curve. Timing errors were generally small but some periods may have errors large enough to significantly contribute to the RMS error.

It is also desirable to have the model produce an amplitude ratio very close to 1.0. An underestimation of the average depth will cause an overestimation of the calculated amplitude and an amplitude ratio greater than 1.0. An overestimation of the average depth will have the opposite effect. Overestimating

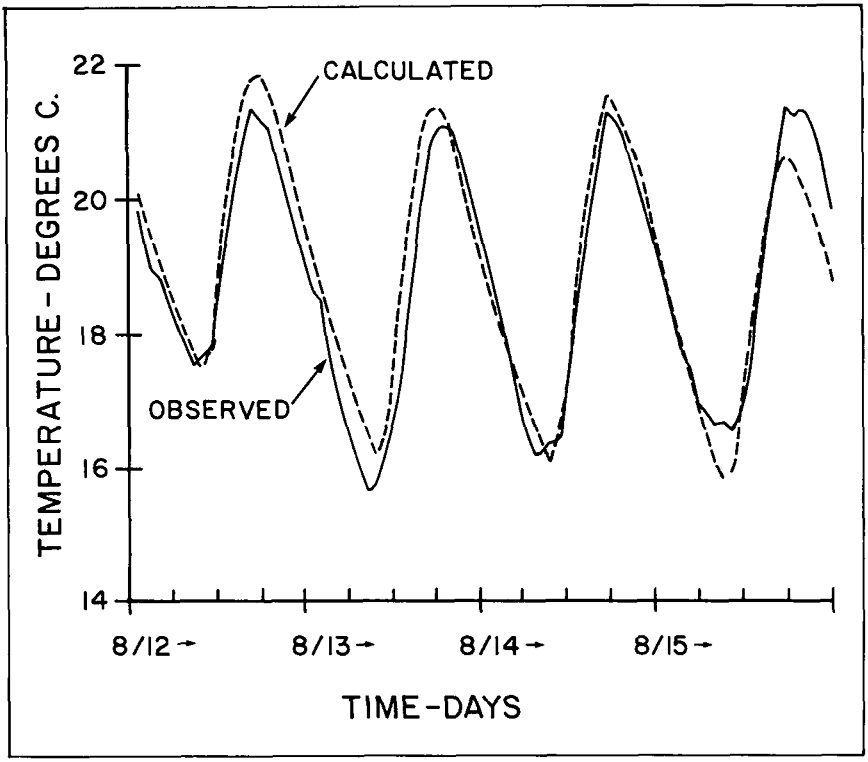

Figure 2. Temperature variation vs. time, Monitoring Station 1.

the total incoming radiation will cause the amplitude ratio to be greater than 1.0 and vice-versa. Also, small errors in the observed temperature will affect the amplitude ratio. If the observed amplitude is overestimated by 0.5°C, it will cause the amplitude ratio to be underestimated by 12 per cent for a 4.0°C observed amplitude and 6 per cent for an 8.0°C observed amplitude.

The model should have the calculated daily mean temperature reproduce the observed mean temperature. Overestimation of the source term will cause the calculated mean to be larger than the observed mean. This can be due to overestimating the total incoming radiation, overestimating S, and to a lesser degree, underestimating the average depth.

It is necessary to use all three parameters to evaluate the performance of the model. Figure 2 represents a good overall fit. Figure 3 represents an overestimation of the mean by 5°-6°C while the amplitude ratio is ± 0.1 of 1.0. Figure 4 represents an overestimation of the amplitude ratio by approximately 0.60 even though the RMS value was less-than 2°C. Figure 5 represents an RMS value of approximately 1.5°C which is considered good, but the amplitude ratio is as high as 1.43 and the mean is overestimated by as much as 1.0°C.

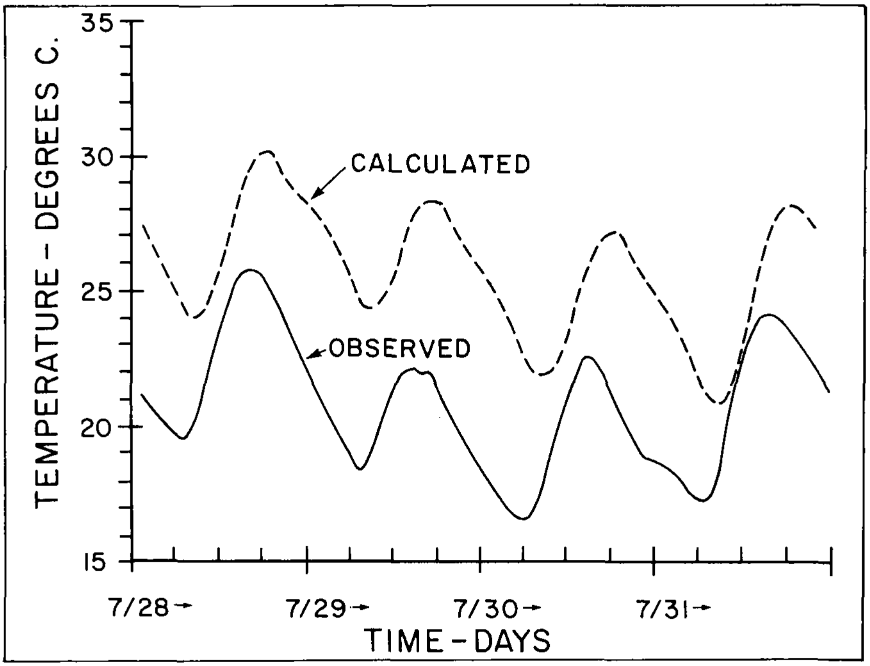

Figure 3. Temperature variation vs. time, Monitoring Station 14.

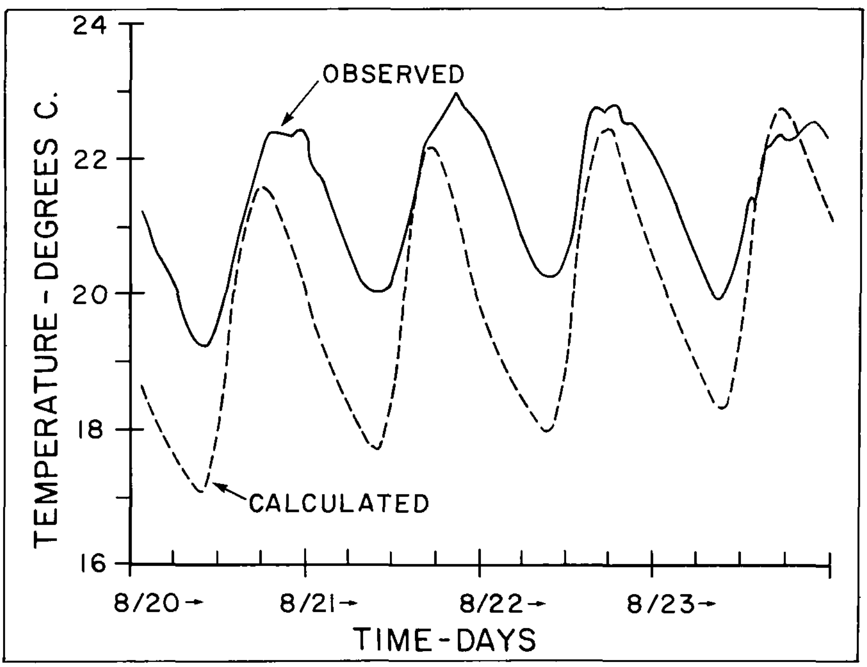

Figure 4. Temperature variation vs. time, Monitoring Station 7.

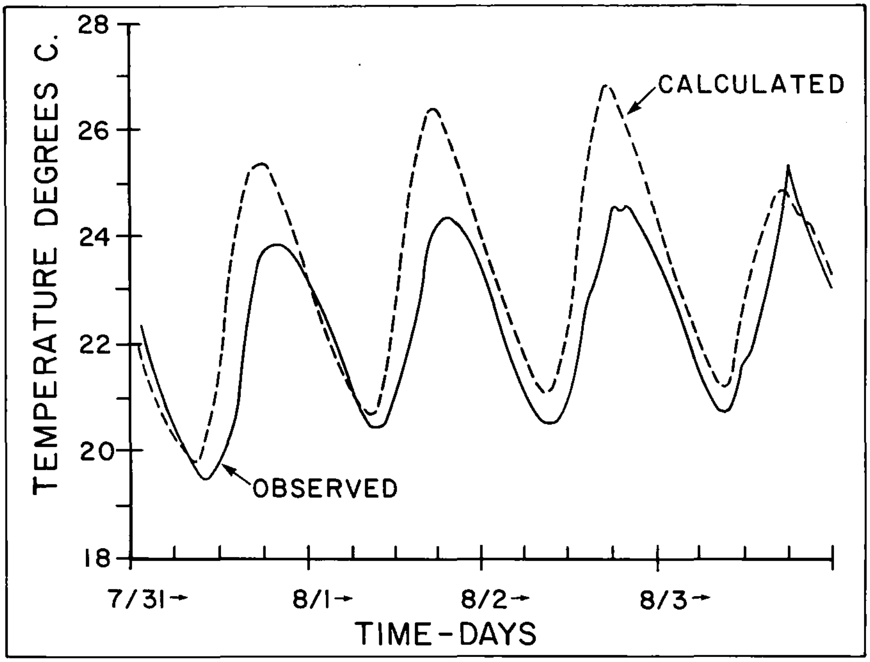

Figure 5. Temperature variation vs. time, Monitoring Station 7.

DEPTH, CONVECTION AND DISPERSION

Table 1 tabulates the three factors for each station for a one week period along the main channel on a daily basis for each model listed below plus the actual daily mean temperature.

Models

Model 1 – fitted average depth, convection and dispersion–In this model, the average depth was fitted using Equation 23. Both this quantity and S were fitted using data from the three-day period August 6, 1975 to August 8, 1975.

Model 2 – variable average depth, convection and dispersion included–In this model, the average depth is a variable calculated from Equation 20. The values of S used were the same as those used for Model 1 since the values obtained for Model 1 and 2 were approximately equal.

Model 3 – fitted depth, convection included and dispersion neglected–This model is included to ascertain the significance of including dispersion in the solution.

Table 1. Daily Values of Root-Mean-Square Deviation (Ø), Amplitude Ratio (AR), and Mean Temperature (T_M)

Date	*Location*	*Model*	*Ø*	*AR*	T_M
8/6	1				18.9
		1	1.24	1.20	20.0
		2	1.77	0.78	20.6
		3	1.24	1.20	20.0
		4	1.24	1.20	20.0
8/7	1				18.3
		1	0.71	0.84	18.7
		2	1.56	0.55	19.5
		3	0.71	0.84	18.7
		4	0.71	0.84	18.7
8/8	1				19.0
		1	0.35	1.02	18.9
		2	0.73	0.68	19.3
		3	0.35	1.04	18.9
		4	0.35	1.02	18.9
8/9	1				19.4
		1	0.33	1.05	19.2
		2	0.68	0.71	19.4
		3	0.33	1.06	19.2
		4	0.33	1.05	19.2
8/10	1				19.6
		1	0.45	0.92	19.4
		2	0.74	0.63	19.5
		3	0.45	0.93	19.4
		4	0.45	0.92	19.4
8/11	1				19.8
		1	0.38	1.07	19.6
		2	0.59	0.74	19.6
		3	0.38	1.06	19.6
		4	0.38	1.07	19.6
8/12	1				19.4
		1	0.41	1.15	19.7
		2	0.46	0.79	19.8
		3	0.41	1.13	19.7
		4	0.42	1.15	19.7
8/6	3				18.8
		1	1.25	1.36	19.5
		2	1.32	1.01	19.9
		3	1.25	1.36	19.5
		4	1.26	1.36	19.5

Table 1. (Cont'd.)

Date	*Location*	*Model*	*Ø*	*AR*	T_M
8/7	3				18.2
		1	0.92	1.10	18.4
		2	0.99	0.84	18.9
		3	0.92	1.10	18.5
		4	0.92	1.10	18.4
8/8	3				18.8
		1	0.78	1.06	18.7
		2	0.66	0.83	19.0
		3	0.77	1.06	18.7
		4	0.78	1.06	18.7
8/9	3				19.3
		1	0.94	0.99	19.1
		2	0.82	0.79	19.2
		3	0.94	0.98	19.1
		4	0.94	0.99	19.1
8/10	3				19.5
		1	0.71	0.97	19.3
		2	0.62	0.77	19.4
		3	0.70	0.96	19.4
		4	0.71	0.97	19.3
8/11	3				19.7
		1	0.78	1.02	19.5
		2	0.69	0.82	19.5
		3	0.77	1.02	19.5
		4	0.78	1.02	19.5
8/12	3				19.6
		1	0.74	0.83	19.6
		2	0.69	0.67	19.6
		3	0.74	0.84	19.6
		4	0.74	0.83	19.6
8/6	7				21.2
		1	0.89	2.01	21.3
		2	0.78	1.84	21.3
		3	0.89	2.00	21.3
		4	1.18	2.04	21.9
8/7	7				20.1
		1	0.72	1.34	20.5
		2	0.73	1.25	20.5
		3	0.73	1.37	20.5
		4	1.21	1.45	21.1

Table 1. (Cont'd.)

Date	*Location*	*Model*	*Ø*	*AR*	T_M
8/8	7				20.8
		1	0.68	1.29	20.5
		2	0.62	1.18	20.5
		3	0.68	1.30	20.5
		4	0.81	1.34	21.3
8/9	7				21.6
		1	1.12	1.19	20.5
		2	1.14	1.09	20.5
		3	1.12	1.21	20.5
		4	0.54	1.15	21.1
8/10	7				21.8
		1	1.01	1.37	20.9
		2	1.01	1.26	20.9
		3	1.00	1.36	20.9
		4	0.62	1.39	21.5
8/11	7	NO DATA			
8/12	7	NO DATA			
8/6	9				20.6
		1	1.88	0.94	22.0
		2	1.85	0.94	22.0
		3	1.90	0.93	22.1
		4	1.88	0.95	22.1
8/7	9				20.8
		1	1.56	0.77	21.2
		2	1.56	0.77	21.2
		3	1.58	0.78	21.2
		4	1.49	0.80	21.2
8/8	9				21.8
		1	1.42	0.82	21.9
		2	1.42	0.80	21.9
		3	1.42	0.82	21.9
		4	1.36	0.83	21.9
8/9	9				21.9
		1	1.50	0.74	22.2
		2	1.51	0.73	22.2
		3	1.50	0.75	22.2
		4	1.42	0.76	22.1
8/10	9				22.0
		1	1.31	0.74	22.2
		2	1.33	0.72	22.2

Table 1. (Cont'd.)

Date	*Location*	*Model*	*Ø*	*AR*	T_M
		3	1.32	0.74	22.2
		4	1.21	0.78	22.2
8/11	9				22.1
		1	1.44	0.74	22.4
		2	1.47	0.70	22.4
		3	1.46	0.73	22.4
		4	1.37	0.76	22.4
8/12	9				21.2
		1	1.48	0.71	21.9
		2	1.51	0.68	21.9
		3	1.48	0.72	21.9
		4	1.41	0.74	21.9
8/6	13				21.6
		1	0.86	1.03	22.4
		2	1.52	1.91	21.8
		3	0.89	1.07	22.4
		4	0.74	1.05	22.0
8/7	13				21.3
		1	0.50	0.81	21.4
		2	1.21	1.49	21.0
		3	0.46	0.82	21.4
		4	0.92	0.84	20.8
8/8	13				22.2
		1	0.45	0.91	21.9
		2	1.07	1.50	22.3
		3	0.40	0.92	22.0
		4	1.15	0.91	21.3
8/9	13				22.1
		1	0.37	0.87	22.3
		2	1.03	1.49	22.4
		3	0.36	0.88	22.4
		4	0.84	0.88	21.6
8/10	13				22.4
		1	0.52	0.84	22.4
		2	1.11	1.42	22.3
		3	0.49	0.86	22.4
		4	1.16	0.83	21.7
8/11	13				22.4
		1	0.48	0.91	22.5
		2	1.13	1.52	22.6
		3	0.47	0.93	22.6
		4	0.93	0.90	21.9

Table 1. (Cont'd.)

Date	*Location*	*Model*	*Ø*	*AR*	T_M
8/12	13				21.7
		1	0.62	0.86	22.1
		2	0.84	1.45	21.7
		3	0.62	0.90	22.1
		4	0.78	0.83	21.5
8/6	14				20.7
		1	1.62	0.90	22.2
		2	1.42	1.72	21.2
		3	1.65	0.89	22.2
		4	1.38	0.91	22.0
8/7	14				20.1
		1	1.40	0.84	21.1
		2	1.36	1.57	20.4
		3	1.43	0.82	21.1
		4	1.06	0.84	20.8
8/8	14				21.2
		1	1.09	0.81	21.7
		2	1.25	1.46	21.7
		3	1.12	0.79	21.7
		4	0.80	0.83	21.3
8/9	14				21.1
		1	1.21	0.89	22.0
		2	1.56	1.60	21.8
		3	1.21	0.86	22.0
		4	0.83	0.91	21.6
8/10	14				20.9
		1	1.59	0.77	22.1
		2	1.58	1.45	21.8
		3	1.60	0.75	22.1
		4	1.15	0.82	21.7
8/11	14				20.6
		1	1.90	1.06	22.3
		2	2.22	1.97	22.1
		3	1.90	1.02	22.3
		4	1.51	1.12	21.9
8/12	14				19.0
		1	2.12	0.87	21.9
		2	1.77	1.70	21.2
		3	2.13	0.86	21.9
		4	1.74	0.94	21.5

Model 4–fitted depth, convection and dispersion neglected–The purpose of this model is to determine the importance of the convective term by comparing this model with Model 1 and 3.

By analyzing the information in Table 1, the first result which becomes apparent is that Model 1 and 3 are virtually identical for all stations. This indicates that axial dispersion can be neglected without any loss in accuracy.

Model 4 can then be compared with Model 1 to determine the validity of neglecting the convective term. The results generally show that Model 4 is inferior to Model 1. The convective effect becomes more dominant as you proceed downstream, so the results for the two models deviate as you move downstream. This result indicates that the convective term should not be neglected.

Comparison of Model 1 and 2 would demonstrate if Equation 23 can be used to empirically determine the average depth if data is not available. Model 1 compares very favorably with Model 2 in the prediction of daily mean temperature and gives a better estimation of the amplitude ratio.

SUMMARY AND CONCLUSIONS

Using an initial temperature distribution, and a groundwater temperature in the headwaters region, it is possible to obtain good temperature estimates throughout an entire river basin using the one dimensions model presented here. It must be recognized that such model results will not be as accurate as those obtained using similar models on individual short reaches where RMS errors of less than 1°C can be expected. RMS errors of 2-3°C appear to be more representative of the results when modeling an entire basin.

Tributary mixing need only be incorporated for major tributaries. As a rule of thumb, a tributary with a discharge less than 5 per cent of the discharge of the mainstem at the point of confluence can be ignored.

The meteorological data required is wet-bulb and dry-bulb air temperature, wind velocity, incoming short-wave radiation, and atmospheric long-wave radiation. Very special care should be taken to obtain radiation measurements as accurately as possible since they are the major contributors to the net heat exchange term. This is especially difficult during periods with varying amounts of cloud cover. An integrator greatly simplifies the analysis of data during cloudy periods.

Estimation of the average depth by Equation 20 is crucial to obtaining accurate results since this quantity affects the entire source term. This requires obtaining accurate flow rate, time of travel, and surface area data repeatedly throughout the period one wishes to investigate. Equation 20 is superior to Equation 23 since it allows the average depth to vary continuously with time.

If this is not possible, then a fitted average depth should be estimated on a monthly basis. This will prove satisfactory during stable flow conditions but will be highly inaccurate during rain storm activity since the depth changes rapidly during this period.

The fraction of short-wave radiation actually passing through the water surface S will vary on a seasonal basis. Therefore, S should be estimated on a monthly basis to obtain an estimation of the seasonal variation.

Dispersion has been shown to be insignificant during this study, but one should not neglect it *a priori.* Since time of travel data is required for measuring the average depth, one can easily obtain dispersion coefficients from this data and then calculate a Peclet Number to ascertain if dispersion can be neglected.

Convective effects cannot in general be neglected. Again, accurate estimation of the time of travel for calculating the average depth will also insure accurate velocity measurements.

The results for Model 2, which contains just one fitted parameter S, show that one can obtain solutions where the root-mean-square deviation will be less than 3°C, the amplitude ratio will be ± 0.20 of 1.0, and the calculated mean will be within 2°C of observed data. This can be improved by improved estimation of the average depth.

The results for Model 1, which contains two fitted parameters d and S, will yield comparable results during steady flow conditions but will produce larger errors during rain storm activity since the actual average depth will very considerably.

Model 2 is superior to Model 1 in two respects. There is one less parameter to evaluate and Model 1 allows the average depth to vary continuously with time. which Model 2 does not do. Model 1 has the benefit of giving results comparable to Model 2 during steady flow conditions. This is especially useful if there is insufficient data to calculate the average depth by Equation 20.

In some cases temperature data may not be available for a major tributary. By estimating d and S and approximating or neglecting convection and dispersion (e.g., utilizing Model 4), one can solve Equation 17 using available meteorological data and obtain an estimate of the water temperature of the tributary. If velocity data is available, it can be incorporated to obtain a better fit.

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