Appendix C
Heat Source Model Analytical Framework

CONCEPTUAL MODEL

At any particular instant of time, a defined stream reach is capable of sustaining a particular water column temperature. Stream temperature change that results within a defined reach is explained rather simply. The temperature of a parcel of water traversing a stream/river reach enters the reach with a given temperature. If that temperature is greater than the energy balance is capable of supporting, the temperature will decrease. If that temperature is less than energy balance is capable of supporting, the temperature will increase. Stream temperature change within a defined reach, is induced by the energy balance between the parcel of water and the surrounding environment and transport of the parcel through the reach. The temperature model utilized by ODEQ to estimate stream network thermodynamics and hydrology is Heat Source (Boyd, 1996; Boyd and Kasper, 2003). It was originally developed in 1996 as a Masters Thesis at Oregon State University in the Departments of Bioresource Engineering and Civil Engineering. The general progression of the model is outlined in Figure C- 1.

Figure C- 1: Heat source temperature model flow chart.
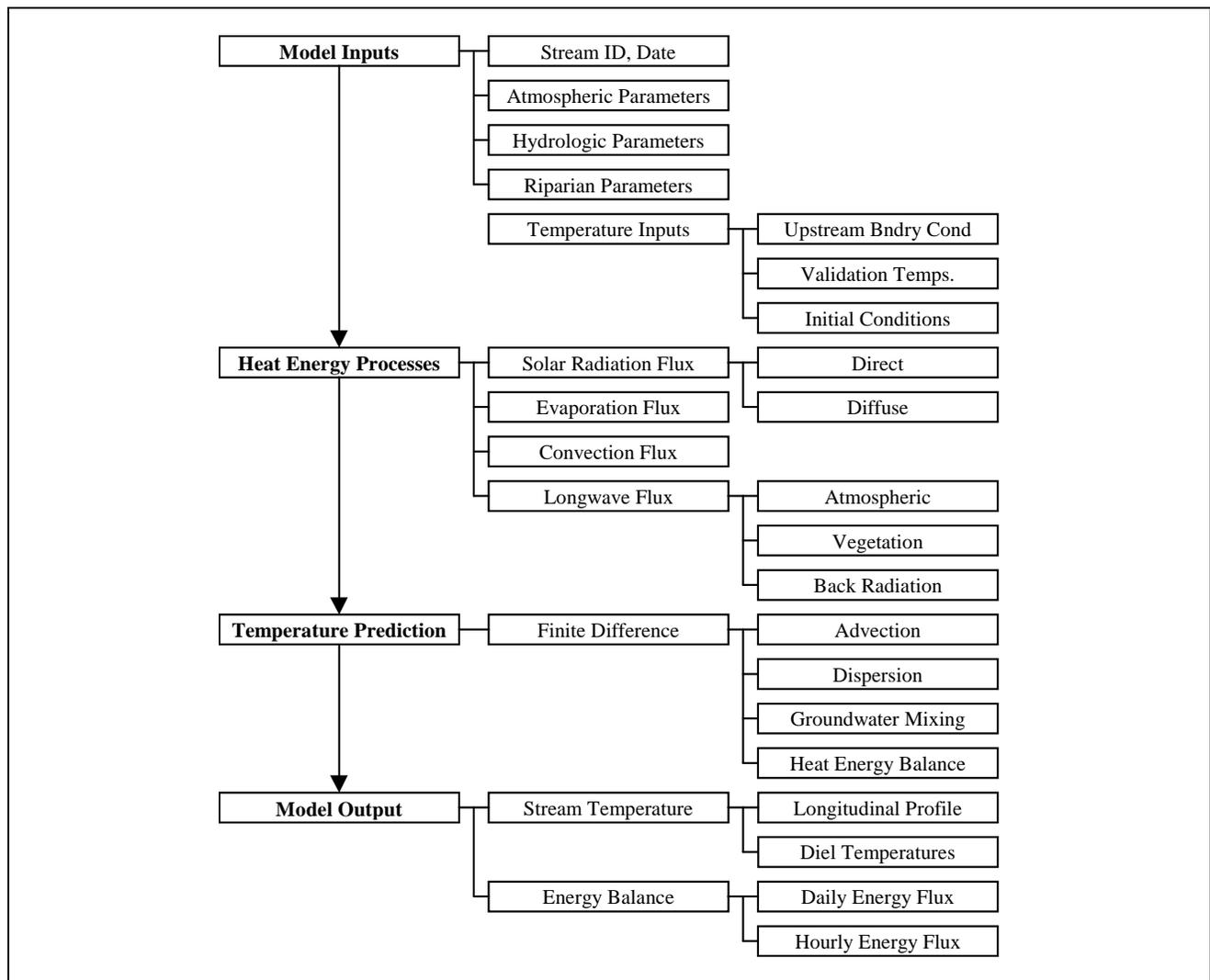
As the water parcel traverses the longitudinal distance of a defined reach, energy processes drive stream temperature change. At any particular instant of time, water that enters the upstream portion of the reach

is never exactly the temperature that is supported by the defined reach. And, as the water is transferred downstream, heat energy and hydraulic processes that are variable with time and space interact with the water parcel and induce water temperature change. The described modeling scenario is a simplification; however, understanding the basic processes in which stream temperatures change occurs over the course of a defined reach and period of time is essential.

GOVERNING EQUATIONS

HEAT ENERGY PROCESSES

Water temperature change is a function of the total heat energy transfer in a discrete volume and may be described in terms of energy per unit volume. It follows that large volume streams are less responsive to temperature change, and conversely, low flow streams will exhibit greater temperature sensitivity.

Equation 1. Heat Energy per Unit Volume,

$$\Delta T_w = \frac{\Delta Heat \quad Energy}{Volume}$$

Water has a relatively high heat capacity.; one kilocalorie of heat added raises the temperature of a kilogram of water one degree Celsius at standard conditions. Conceptually, water is a heat sink. Heat energy that is gained by the stream is retained and only slowly released back to the surrounding environment, represented by the cooling flux ($\Phi_{cooling}$). Heating periods occur when the net energy flux (Φ_{total}) is positive: ($\Phi_{heating} > \Phi_{cooling}$).

Equation 2. Heat Energy Continuity,

$$\Phi_{total} = \Phi_{heating} - \Phi_{cooling}$$

In general, the net energy flux experienced by all stream/river systems follows two cycles: a seasonal cycle and a diurnal cycle. In the Pacific Northwest, the seasonal net energy cycle experiences a maximum positive flux during summer months (July and August), while the minimum seasonal flux occurs in winter months (December and January). The diurnal net energy cycle experiences a daily maximum flux that occurs at or near the sun's zenith angle, while the daily minimum flux often occurs during the late night or the early morning. It should be noted, however, that meteorological conditions are variable. Cloud cover and precipitation significantly alter the energy relationship between the stream and its environment.

The net heat energy flux (Φ_{total}) consists of several individual thermodynamic energy flux components, namely: solar radiation (Φ_{solar}), long-wave radiation ($\Phi_{longwave}$), conduction ($\Phi_{conduction}$), groundwater exchange ($\Phi_{groundwater}$) and evaporation ($\Phi_{evaporation}$).

Equation 3. Net Heat Energy Continuity,

$$\Phi_{total} = \Phi_{solar} + \Phi_{longwave} + \Phi_{convection} + \Phi_{evaporation} + \Phi_{streambed} + \Phi_{groundwater}$$

Stream temperature is an expression of heat energy per unit volume, which in turn is an indication of the rate of heat exchange between a stream and its environment. The heat transfer processes that control stream temperature include solar radiation, longwave radiation, convection, evaporation and bed conduction (Wunderlich, 1972; Jobson and Keefer, 1979; Beschta and Weatherred, 1984; Sinokrot and Stefan, 1993; Boyd, 1996). With the exception of solar radiation, which only delivers heat energy, these processes are capable of both introducing and removing heat from a stream. Figure C- 2 displays heat energy processes that solely control heat energy transfer to and from a stream.

Figure C- 2: Heat Energy Processes
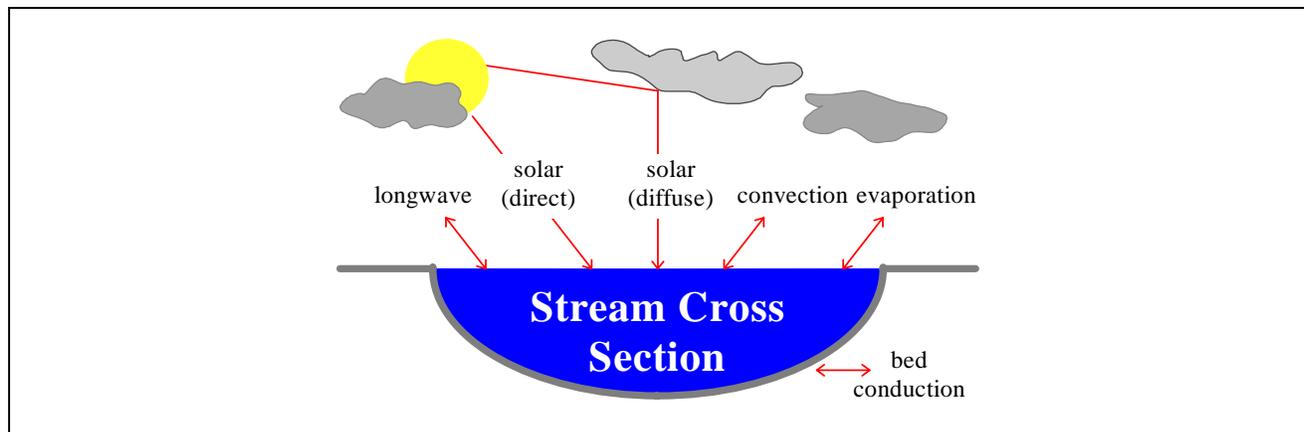
When a stream surface is exposed to midday solar radiation, large quantities of heat will be delivered to the stream system (Brown 1969, Beschta et al. 1987). Some of the incoming solar radiation will reflect off the stream surface, depending on the elevation of the sun. All solar radiation outside the visible spectrum (0.36μ to 0.76μ) is absorbed in the first meter below the stream surface and only visible light penetrates to greater depths (Wunderlich, 1972). Sellers (1965) reported that 50% of solar energy passing through the stream surface is absorbed in the first 10 cm of the water column. Removal of riparian vegetation, and the shade it provides, contributes to elevated stream temperatures (Rishel et al., 1982; Brown, 1983; Beschta et al., 1987). The principal source of heat energy delivered to the water column is solar energy striking the stream surface directly (Brown 1970). Exposure to direct solar radiation will often cause a dramatic increase in stream temperatures. The ability of riparian vegetation to shade the stream throughout the day depends on vegetation height, width, density, position relative to the stream, and stream aspect.

Water Temperature ⇑

Rise above natural conditions as a result of increased

Solar Radiation ⇑
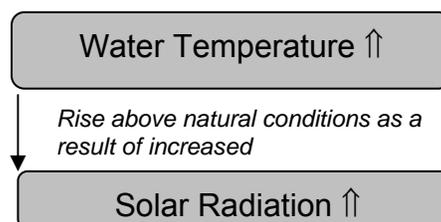
Both the atmosphere and vegetation along stream banks emit longwave radiation that can heat the stream surface. Water is nearly opaque to longwave radiation and complete absorption of all wavelengths greater than 1.2μ occurs in the first 5 cm below the surface (Wunderlich, 1972). Longwave radiation has a cooling influence when emitted from the stream surface. The net transfer of heat via longwave radiation usually balances so that the amount of heat entering is similar to the rate of heat leaving the stream (Beschta and Weatherred, 1984; Boyd, 1996).

Evaporation occurs in response to internal energy of the stream (molecular motion) that randomly expels water molecules into the overlying air mass. Evaporation is the most effective method of dissipating heat from water (Parker and Krenkel, 1969). As stream temperatures increase, so does the rate of evaporation. Air movement (wind) and low vapor pressures increase the rate of evaporation and accelerate stream cooling (Harbeck and Meyers, 1970).

Convection transfers heat between the stream and the air via molecular and turbulent conduction (Beschta and Weatherred, 1984). Heat is transferred in the direction of warmer to cooler. Air can have a warming influence on the stream when the stream is cooler. The opposite is also true. The amount of convective heat transfer between the stream and air is low (Parker and Krenkel, 1969; Brown, 1983). Nevertheless, this should not be interpreted to mean that air temperatures do not affect stream temperature.

Depending on streambed composition, shallow streams (less than 20 cm) may allow solar radiation to warm the streambed (Brown, 1969). Large cobble (> 25 cm diameter) dominated streambeds in shallow streams may store and conduct heat as long as the bed is warmer than the stream. Bed conduction may cause maximum stream temperatures to occur later in the day, possibly into the evening hours.

Conduction may also occur between groundwater, tributary, or point source inputs and the stream. The rate of conduction between two volumes of water depends upon 1) their relative volumes and 2) the temperatures of each water volume.

The instantaneous heat transfer rate experienced by the stream is the summation of the individual processes:

$$\Phi_{Total} = \Phi_{Solar} + \Phi_{Longwave} + \Phi_{Evaporation} + \Phi_{Convection} + \Phi_{Conduction}.$$

Solar Radiation (Φ_{Solar}) is a function of the solar angle, solar azimuth, atmosphere, topography, location and riparian vegetation. Simulation is based on methodologies developed by Ibqal (1983) and Beschta and Weatherred (1984). *Longwave Radiation* ($\Phi_{Longwave}$) is derived by the Stefan-Boltzmann Law and is a function of the emissivity of the body, the Stefan-Boltzmann constant and the temperature of the body (Wunderlich, 1972). *Evaporation* ($\Phi_{Evaporation}$) relies on a Dalton-type equation that utilizes an exchange coefficient, the latent heat of vaporization, wind speed, saturation vapor pressure and vapor pressure (Wunderlich, 1972). *Convection* ($\Phi_{Convection}$) is a function of the Bowen Ratio and terms include atmospheric pressure, and water and air temperatures. *Bed Conduction* ($\Phi_{Conduction}$) simulates the theoretical relationship ($\Phi_{Conduction} = K \cdot dT_b / dz$), where calculations are a function of thermal conductivity of the bed (K) and the temperature gradient of the bed (dT_b/dz) (Sinokrot and Stefan, 1993). Bed conduction is solved with empirical equations developed by Beschta and Weatherred (1984).

The ultimate source of heat energy is solar radiation both diffuse and direct. Secondary sources of heat energy include long-wave radiation, from the atmosphere and streamside vegetation, streambed conduction and in some cases, groundwater exchange at the water-stream bed interface. Several processes dissipate heat energy at the air-water interface, namely: evaporation, convection and back radiation. Heat energy is acquired by the stream system when the flux of heat energy entering the stream is greater than the flux of heat energy leaving. The net energy flux provides the rate at which energy is gained or lost per unit area and is represented as the instantaneous summation of all heat energy components.

NON-UNIFORM HEAT ENERGY TRANSFER EQUATION

The rate change in stream temperature is driven by the heat energy flux (Φ_i). A defined volume of water will attain a predictable rate change in temperature, providing an accurate prediction of the heat energy flux. The rate change in stream temperature (T) is calculated as shown in **Equation 4**.

Equation 4. Rate Change in Temperature Caused by Heat Energy Thermodynamics,

$$\frac{\partial T}{\partial t} = \left(\frac{Ax_i \cdot \Phi_i}{\rho \cdot c_p \cdot V_i} \right),$$

Which reduces to,

$$\frac{\partial T}{\partial t} = \left(\frac{\Phi_i}{\rho \cdot c_p \cdot D_i} \right).$$

Where,

Ax_i:	cross-sectional wetted area (m^2)
C_p:	specific heat of water (cal kg^{-1} · ^{o}C^{-1})
D_i:	average stream depth (m)
t:	time (s)
T:	Temperature (^{o}C)
V_i:	volume (m^3)
Φ_i:	total heat energy flux (cal m-^{2} · s^{-1})
ρ:	density of water (kg/m^3)

Advection (U_x) redistributes heat energy in the positive longitudinal direction. No heat energy is lost or gained by the system during advection, and instead, heat energy is transferred downstream as a function of flow velocity. In the case where flow is uniform, the rate change in temperature due to advection is expressed in the first order partial differential equation below.

Equation 5. Rate Change in Temperature Caused by Advection,

$$\frac{\partial T}{\partial t} = -U_x \cdot \frac{\partial T}{\partial x}$$

Dispersion processes occur in both the upstream and downstream direction along the longitudinal axis. Heat energy contained in the system is conserved throughout dispersion, and similar to advection, heat energy is simply moved throughout the system. The rate change in temperature due to dispersion is expressed in the second order partial differential equation below.

Equation 6. Rate Change in Temperature Caused by Dispersion,

$$\frac{\partial T}{\partial t} = D_L \cdot \frac{\partial^2 T}{\partial x^2}$$

The dispersion coefficient (D_L) may be calculated by stream dimensions, roughness and flow. In streams that exhibit high flow velocities and low longitudinal temperature gradients, it may be assumed that the system is advection dominated and the dispersion coefficient may be set to zero (Sinokrot and Stefan 1993). In the event that dispersion effects are considered significant, the appropriate value for the dispersion coefficient can be estimated with a practical approach developed and employed in the QUAL 2e model (Brown and Barnwell 1987). An advantage to this approach is that each parameter is easily measured, or in the case of Manning's coefficient (n) and the dispersion constant (K_d), estimated.

Equation 7. Physical Dispersion Coefficient,

$$D_L = C \cdot K_d \cdot n \cdot U_x \cdot D^{\frac{5}{6}}$$

Where,

C:	Unit conversion
	C = 3.82 for English units
	C = 1.00 for Metric units
D:	Average stream depth (m)
D_L:	Dispersion coefficient (m^2/s)
K_d:	Dispersion constant
n:	Manning's coefficient
U_x:	Average flow velocity (m/s)

The simultaneous non-uniform one-dimensional transfer of heat energy is the summation of the rate change in temperature due to heat energy thermodynamics, advection and dispersion.   Given that the stream is subject to steady flow conditions and is well mixed, transverse temperature gradients are negligible (Sinokrot and Stefan 1993). An assumption of non-uniform flow implies that cross-sectional area and flow velocity vary with respect to longitudinal position. The following second ordered parabolic partial differential equation describes the rate change in temperature for non-uniform flow.

Equation 8. Non-Uniform One-dimensional Heat Energy Transfer,

$$\frac{\partial T}{\partial t} = -U_x \cdot \frac{\partial T}{\partial x} + D_L \cdot \frac{\partial^2 T}{\partial x^2} + \frac{\Phi}{c_p \cdot \rho \cdot D_i}$$

$$\text{Steady Flow: } \frac{\partial U_x}{\partial t} = 0$$

$$\text{Non-Uniform Flow: } \frac{\partial U_x}{\partial x} \neq 0$$

The solution to the *one-dimensional heat energy transfer equation* is essentially the summation of thermodynamic heat energy exchange between the stream system and the surrounding environment and physical processes that redistribute heat energy within the stream system. It is important to note that all heat energy introduced into the stream is conserved, with the net heat energy value reflected as stream temperature magnitude. Further, heat energy is transient within the stream system, due to longitudinal transfer of heat energy (i.e., advection and dispersion). The net heat energy flux (Φ) is calculated at every distance step and time step based on physical and empirical formulations developed for each significant energy component. The dispersion coefficient (D_L) is assumed to equal zero.

BOUNDARY CONDITIONS AND INITIAL VALUES

The temperatures at the upstream boundary (i_o) for all time steps (t_o ,t_1,..., t_{M-1}, t_M) are supplied by the upstream temperature inputs. The downstream boundary temperature at longitudinal position i_{n+1} is assumed to equal that of i_n with respect to time t. Initial values of the temperatures at each distance node (i_o ,i_1,..., i_{N-1}, i_N) occurring at the starting time (t_o) can be input by the model user or assumed to equal the boundary condition at time t_o.

SPATIAL AND TEMPORAL SCALE

The lengths of the defined reaches are 100 feet. The temperature model is designed to analyze and predict stream temperature for one day and is primarily concerned with daily prediction of the diurnal energy flux and resulting temperatures in July or August. Prediction time steps are limited by stability considerations for the finite difference solution method.

INPUT PARAMETERS

Data collected during this TMDL effort has allowed the development of temperature simulation methodology that is both spatially continuous and spans a full days length (diurnal). Detailed spatial data sets have been developed for the following parameters:

- ✓ River and Tributary Digital Mapping at 1:5,000 scale
- ✓ Riparian Vegetation Species, Size and Density Digital Mapping at 1:5,000 scale
- ✓ West, East and South Topographic Shade Angles calculations at 1:5,000 scale
- ✓ Stream Elevation and Gradient at 1:5,000 scale,
- ✓ Hydrology Developed from Field Data - Spatially Continuous Flow, Wetted Width, Velocity and Depth Profiles.

All input data are longitudinally referenced in the model allowing spatial and/or continuous inputs to apply to certain zones or specific river segments.

SPATIAL INPUT PARAMETERS

Longitudinal Distance (meters): Defines the modeled reaches for which spatial input parameters reference. Model reaches are 100 feet each, are derived from DOQ 1:5000 river layer digitized from

Digital Orthophoto Quarter Quads (DOQs) and geo-referenced color aerial photographs, and are measured in the downstream direction.

Elevation (meters): Sampled for each model reach either from Digital Raster Graphic (DRG) or Digital Elevation Model (DEM).

Gradient (%): Is the difference between the upstream and downstream elevations divided by the reach length.

Bedrock (%): The percent of streambed material that has a diameter of 25 cm or greater. Values are derived from stream survey data or assumed where data are limited.

Aspect (decimal degrees from North): Calculated for each reach break and represents the direction of stream flow.

Flow Volume (cubic meters per second): Measured by ODEQ with standard USGS protocols with interpolation between flow measurement sites, while taking into account known water withdrawals and inputs.

Flow Velocity (meters per second): Derived from Manning's equation and Leopold power functions calibrated to measured flow velocity data.

Wetted Width (meters): Derived from Manning's equation and Leopold power functions calibrated to measured wetted width data.

Average Depth (meters): Derived from Manning's equation and Leopold power functions calibrated to measured average depth data. Calculated based on assuming rectangular channel.

Near-Stream Disturbance Zone Width (meters): Commonly referred to as bankfull width or channel width. Based upon ODEQ field measurements and aerial photograph interpretation.

Channel Incision (meters): Depth of the active channel below riparian terrace or floodplain. Based on ODEQ field measurements.

Riparian Height (meters): Determined from the professional expertise of foresters with ODF and the USFS and from ODEQ field observations.

Canopy Density (%): Determined from the professional expertise of foresters with ODF, the USFS, and from ODEQ field observations and aerial photograph interpretation.

Riparian Overhang (meters): Distance of riparian vegetation intrusion over the bankfull channel width. Based on ODEQ field observations.

Topographic Shade Angle (decimal degrees): The angle made between the stream surface and the highest topographic features to the west, east and south as calculated from DEM at each stream reach.

CONTINUOUS INPUT PARAMETERS

Wind Speed (meters per second): Hourly values measured by NOAA, municipal airports, or the Oregon Climate Service.

Relative Humidity (%): Hourly values measured by NOAA, municipal airports, or the Oregon Climate Service.

Air Temperature (^{o}C): Hourly values measured by NOAA, municipal airports, or the Oregon Climate Service.

Stream Temperature (^{o}C): Hourly values measured by ODEQ and other agencies as stated in each subbasin TMDL.

Tributary Temperature (^{o}C): Hourly values measured by ODEQ, and other agencies as stated in each subbasin TMDL.

Tributary/Flow Volume (cubic meters per second): Measured flow volumes for all major tributaries by ODEQ and other agencies as stated in each subbasin TMDL.

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