Pathways of Human Influence on Water Temperature Dynamics in Stream Channels[1]

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ABSTRACT / While external factors ("drivers") determine the net heat energy and water delivered to the fluvial system, the internal structure of fluvial systems and their components (channel, alluvial aquifer, and floodplain/riparian zone) determines how heat and water are distributed and exchanged amongst or lost from streams. Therefore, channel water temperature is ultimately determined by the interaction between external drivers of stream temperature and the internal structure of the integrated stream system. This paper provides a synoptic discussion of the external drivers of stream temperature, the internal hydrologic processes that insulate and buffer channel water-temperatures, and the mechanisms of human influence on drivers and stream structure which ultimately alter the temperature regime of stream networks. Key management implications include: 1) in-channel water flow is a critical element for re-establishing desirable thermal regimes in streams; 2) in addition to modified riparian vegetation structure, human alteration of groundwater dynamics and channel morphology are critical pathways of human influence on channel-water temperature; 3) the potential success of stream temperature research and monitoring programs will be jeopardized by an inaccurate or incomplete conceptual understanding of complex temporal and spatial patterns of stream temperature response to anthropogenic influences; and 4) watershed assessment, including analyses of land-use history and analysis of historic vs. contemporary structure of the stream channel, riparian zone, and alluvial aquifer, is an important tool in developing effective management prescriptions for meeting water quality targets for in-channel temperature.

Stream temperature directly influences the metabolic rates, physiology, and life history traits of aquatic species and helps to determine rates of important community processes such as nutrient cycling and productivity (Allen 1995). Natural or anthropogenic fluctuations in water temperature can induce a wide array of behavioral and physiological responses in aquatic organisms and more permanent changes in stream temperature can render formerly suitable habitat unusable for native species assemblages (Holtby 1988, Wissmar and others 1994a, Quigley and Arbelbide 1997). Because of the importance of stream temperature in regulating biochemical processes in lotic systems, preventing or mitigating anthropogenic changes in stream temperature has become a common concern for many resource managers (Coutant 1999). Unfortunately, management actions do not always adequately consider the multitude of interacting environmental processes that determine stream temperature regimes nor the wide variety of pathways by which human activities may affect stream temperature. In this paper, we attempt to succinctly describe a number of these important processes and pathways. Our most detailed discussions focus on heat energy exchange and transport *within* stream systems because, in our opinion, these processes provide great promise for successful stream temperature management yet are most often overlooked during the development management plans. Although the discussion and examples in this paper focus on the

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Pacific Norwest, USA, the ecological principles and processes discussed are applicable to lotic systems in general.

Fluvial System Structure

Current understanding of stream ecology indicates that streams are comprised of at least three integrated and interdependent components: the channel, riparian zone, and alluvial aquifer (Stanford and Ward 1988, Ward 1989, Stanford and Ward 1993, Gibert and others 1994, Findlay 1995, Ward 1998a, Ward 1998b). From this perspective, the “edge” of a river is not defined by its channel margin, but rather by the edge of the riparian zone (Gregory and others 1991). Similarly, the “bottom” of a river is not the stream bed, but the bottom of the alluvial aquifer (Ward 1998b). These components are set within the context of the catchment’s surface and groundwater flow network. (Figure 1).

Channel water temperature is ultimately determined by the interactions between external drivers of stream temperature and the internal structure of the integrated stream system. The presence and importance of specific drivers and stream characteristics vary across spatial scales and are determined in part by landscape context. Together, they interact to produce heterogeneity in stream temperature at a variety of spatial and temporal scales.

Although other factors also affect stream temperature, the primary determinants of stream temperature are climatic drivers (such as air temperature and wind speed), stream morphology, groundwater influences, and riparian canopy condition (Sullivan and Adams 1991). Therefore, this paper focuses on the importance of stream morphology, groundwater influences, and riparian canopy conditions as factors that both markedly influence stream temperature and are also substantively altered by various human activities.

The **stream channel** is the portion of a stream system that transports water across the earth’s surface. The channel boundary is approximately the typical annual high water level on each stream bank. Some streams have multiple threads to their channels (Leopold and Wolman 1957, Kellerhalls and others 1976, Mosley 1987). This underscores the fact that a stream channel may be discontinuous in cross section, comprised of the main channel, side channels, and perhaps channels that are active only during the period of annual high flow. Where floodplains are present, the locations of channels change over time (Leopold and others 1964, Naiman and others 1992). Sometimes these changes occur gradually over decades as streams erode the outer banks along stream meanders and deposit sediment along the inner banks. In other instances, streams in flood stage rapidly cut new channels or re-capture previously abandoned channels (Nanson and

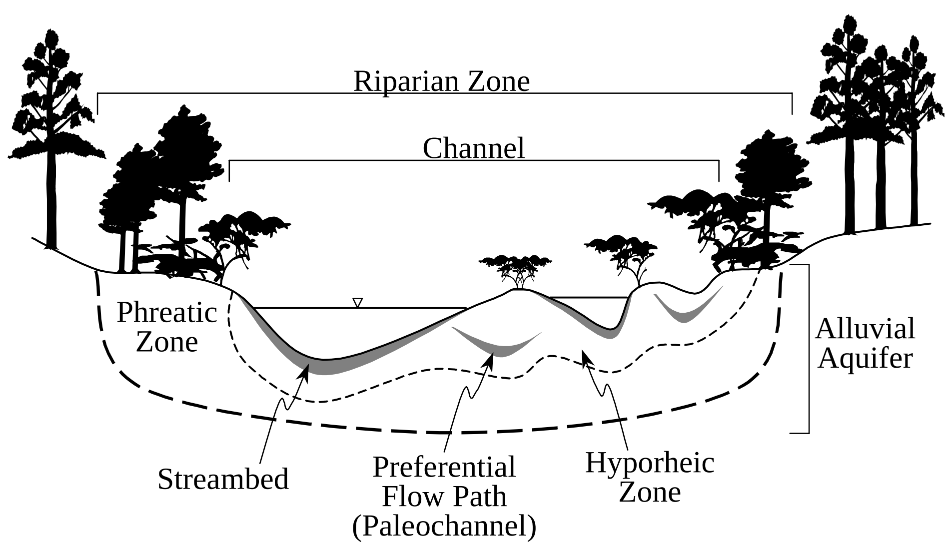

Figure 1. Structural components of a stream system. Not all features exist in all streams.

Knighton 1996). Channel migration processes are important for the creation and maintenance of floodplain complexity. This complexity, in turn, directly influences important in-stream dynamics (e.g., nutrient and carbon cycles, natural floodwater storage, and water temperature buffers) and enhances the variety of available aquatic and terrestrial habitats thereby supporting biological diversity (Sedell and Froggatt 1984, Harvey and Bencala 1993, Creuzé des Chatelliers and others 1994, Abbe and Montgomery 1996).

The **riparian zone** is the area of land influenced by moisture derived directly from the stream. For small streams, this area may only extend a short distance (10^0 to 10^1 m) laterally from the channel margin. However, for larger streams, the riparian zone extends much further (10^1 to 10^3 m), at least to the edge of the active floodplain (Gregory and others 1991). For the great rivers of the world such as the Mississippi and Amazon, the riparian zone sometimes extends even further (10^3 to 10^5 m) (Salo and others 1986). Riparian zones form the transition zone (or *ecotone*) between terrestrial and aquatic systems. Periodic flooding of the riparian zone encourages the exchange of water, nutrients, sediments, and energy between the river channel and the riparian zone. This exchange creates unique habitats, enhances natural productivity, and drives biological processes that contribute to the ecological complexity and integrity of stream systems (Ward 1998a).

A stream's *alluvium* (sediments that have been deposited and sorted as the result of hydraulic processes) along with the groundwater contained therein form the **alluvial aquifer** (Creuzé des Chatelliers and others 1994). Generally speaking, the alluvial aquifer includes the sediments that underlie the riparian zone (or floodplain) and the sediments that comprise the streambed. In streams that flow across bedrock, alluvial deposits (and therefore the alluvial aquifer) may be no more extensive than pockets of sediment trapped in depressions in the bedrock. In most large rivers, however, the upper substrate of the floodplain is built entirely from alluvial deposits which can be meters thick. Stream channels actively exchange water back and forth with their alluvial aquifer (Gibert and others 1994). *Hyporheic groundwater* is water that infiltrates into the alluvial aquifer from the stream, travels along localized subsurface flow pathways for relatively short periods of time (perhaps from 10^{-2} to 10^4 days), and re-emerges into the stream channel downstream without leaving the alluvial aquifer. The portion of the alluvial aquifer that contains at least some hyporheic groundwater (White 1993) is referred to as the *hyporheic zone* (Stanford and Ward 1988, Brunke and Gonser 1997). Therefore, there are two types of groundwater that influence streams, hyporheic groundwater and *phreatic groundwater* (water derived from the catchment aquifer). Phreatic groundwater feeding a river enters the alluvial aquifer and eventually mixes with hyporheic groundwater. Depending on subsurface flow dynamics, the groundwater released into the stream channel at a given point may be predominantly phreatic, predominantly hyporheic, or a mixture of both. The hyporheic zone can exert an extremely strong influence on the biological, chemical, and physical processes that occur in a river (Stanford and Ward 1993, Findlay 1995, Brunke and Gonser 1997).

Water Temperature in Stream Channels

Water temperature is not a simple measure of the *amount* of heat energy in a stream reach. Temperature is proportional to heat energy divided by the volume of water:

$$\text{Water Temp} \propto \text{Heat Energy / Water Volume}$$

Conceptually, water temperature can be thought of as a measure of the "concentration" of heat energy in a stream. All water contains heat energy; warmer water simply contains a higher "concentration" of heat energy than does cooler water.

The *heat load* is a measure of the net amount of heat added to a stream channel; any increase or reduction in heat load will affect stream temperature by altering the amount of heat energy in the system. The *flow rate* is a measure of the volume of water flowing in a stream channel. Substituting "heat load" and "flow rate" into the above equation results in:

$$\text{Water Temp} \propto \text{Heat Load / Flow Rate}$$

Therefore, stream temperature is dependent on both heat load and stream flow; any process that influences heat load to the channel *or* stream flow in the channel will influence channel water temperature and can be considered a *driver* of stream temperature. All water contains heat energy. Thus, heat energy is introduced to a stream channel any time water is added to the channel and lost any time water is removed. When cool water is added to a warm stream, the temperature falls not because heat energy is lost, but because the "concentration" of heat energy in the stream is diluted. In spite of the fact that heat energy is lost when water is removed from a stream, the temperature remains unchanged because the "concentration" of heat energy in the stream remains the same. (Note that evaporation is an exception to this rule. The cooling effect of evaporation results from the fact that the water adsorbs additional heat energy as it changes state from a liquid to a vapor. The additional energy removed from the stream alters the ratio of heat energy to water volume in the stream.)

Heat energy is also gained or lost by a stream without adding or removing water. Heat flows between the stream and atmosphere in a variety of ways that does not require the exchange of water (Naiman and others 1992). Heat energy is transferred from the sun to the stream via the process of radiation. Heat in the atmosphere is transported to the stream surface via convection, conduction, and advection and is then transferred into the stream channel via conduction between the air and water. When heat is added to or removed from a stream channel without altering flow, only the heat load is altered. Increasing the heat load while holding flow constant will increase stream temperature while decreasing the heat load will decrease stream temperature. By extension, the same heat load applied to a lesser flow will result in higher water temperatures. This illustrates the importance of flow in determining the stream's ability to resist temperature changes in response to a given heat load.

Drivers of Stream Temperature

Drivers of stream temperature generally operate beyond the boundaries of the stream and help to form the physical setting or context within which the stream flows. Drivers control the rate at which heat and water are delivered to the stream system and therefore have ability to actually cool or warm the water in the stream. Examples of drivers are listed in Table 1.

Table 1. Examples of natural drivers of channel water temperature

Topographic Shade
Upland Vegetation
Precipitation
Air temperature
Wind speed
Solar angle
Cloud cover
Relative humidity
Phreatic groundwater temp. and discharge
Tributary temperature and flow

Climatic drivers (e.g., precipitation, air temperature, etc.) interact with the geographic drivers (i.e., topography, lithology, and upland vegetation) in the basin to determine the rate and means by which water enters the stream. Ultimately, all stream flow derives from precipitation, but precipitation enters the stream via a number of pathways: directly, via surface flow, and via groundwater discharge after infiltrating the catchment aquifer.

Although some streams in arid climates flow only as the result of surface run-off, many streams derive at the bulk of their flow directly from groundwater. Therefore, the temperature of the surrounding upland aquifer is generally the "baseline" temperature from which stream temperature deviates (although streams fed by snowfields and glaciers may be exceptions to this rule). Channel water temperature trends away from "baseline" temperature and toward atmospheric temperatures in a downstream direction (Sullivan and others 1990).

As soon as groundwater enters the stream channel and is exposed to the atmosphere, heat exchange begins and the water begins to equilibrate with atmospheric temperature. In the absence of insulating, and buffering influences, streams will rapidly trend away from groundwater temperature and toward atmospheric temperature. Even in the presence of insulating and buffering influences, streams may naturally reflect a gradual downstream warming trend in temperature (e.g.,

Figure 2), yet these trends in temperature are almost always punctuated by small-scale cooling trends due to changes in local conditions along the stream (Torgersen and others 1999). Groundwater from the catchment aquifer influences channel water temperature when it enters the stream channel; if the water in the channel has warmed or cooled while flowing downstream, lateral groundwater inputs moderate channel water temperature toward groundwater temperature.

Temperature of lateral surface water inputs to the stream network reflect the seasonal climate and is much less consistent over the year than that of groundwater inputs. Like groundwater inputs, however, lateral inputs from tributaries and surface run-off affect water temperature by pulling the channel temperature toward that of the incoming water.

Temperature Dynamics within Fluvial System

Unlike drivers of stream temperature which operate outside the boundaries of the stream, the stream's physical structure (as represented by channel and floodplain morphology, riparian vegetation structure, and the stratigraphy of the alluvial aquifer) exerts internal control over water temperature. While drivers determine heat and water delivery to the stream, the physical structure of a stream determines how well the water in a stream channel *resists* warming or cooling. Additionally, stream structure determines the means and rates of heat and water entry into, flow through, storage within, and release from the stream system and its components. Stream structure is strongly influenced by the physical dynamics occurring within the stream and the interaction with the surrounding landscape (Vannote and others 1980, Beschta and Platts 1986, D'Angelo and others 1997, Hawkins and others 1997).

A wide variety of stream "characteristics" (descriptions or measures of stream structures) affect channel water temperature response to stream temperature drivers (Table 2). Some characteristics enhance insulating processes in streams by reducing the rate of heat or water flux into or out of the channel. Other characteristics influence buffering processes by removing heat/water from the channel when temperatures/flows are high and releasing heat/water to the channel when temperatures/flows are low.

Insulating processes

Stream characteristics that influence the rate of heat exchange with the atmosphere can be said to insulate the stream. These characteristics include the height, density, and proximity to the channel of riparian vegetation and the width of the stream channel. Riparian vegetation shades the stream, blocking solar radiation from reaching the channel

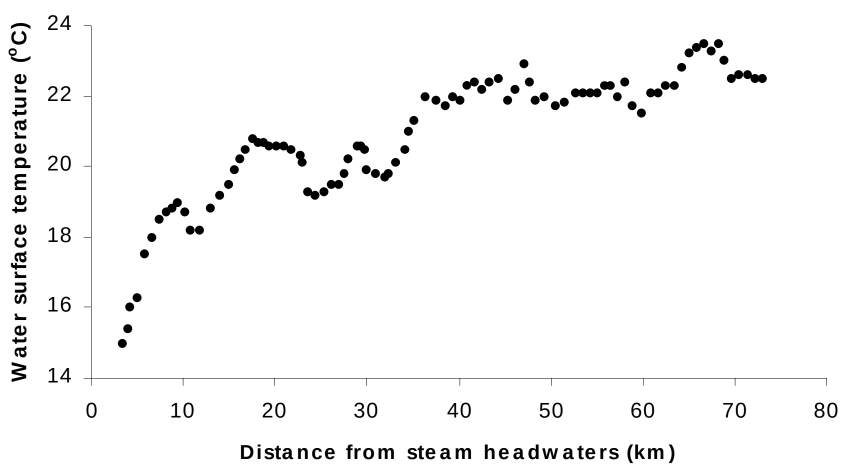

Figure 2. Downstream profile of water surface temperature in the North Fork John Day River, a stream draining relatively pristine, wilderness areas. Data are from Torgersen et al. (1999).

Table 2: Stream structures that influence insulating and buffering characteristics.

Component Characteristic	*Determined by:*	*Ecological influence over:*
Channel		
Channel slope	catchment topography	- Flow rate.
Channel substrate	flow regime, sediment sources, stream power	- Resistance to groundwater flux - Channel roughness and therefore flow rate
Channel width	flow regime, sediment sources, stream power, bank stability	- Surface area for convective heat exchange
Streambed topography	flow regime, sediment sources, stream power, bank stability, large roughness elements (e.g., large woody debris)	- Gradients that drive hyporheic flux
Channel pattern	flow regime, sediment sources, stream power, bank stability, large roughness elements, valley shape	- Gradients that drive hyporheic flux - Potential shade from riparian vegetation
Riparian Zone		
Riparian Vegetation	flow regime, vegetation height, density, growth form, rooting pattern	- Shade to reduce solar radiation - Wind-speed, advective heat transfer, conductive heat transfer - Bank stability
Riparian width	(same as channel pattern)	- Potential for hyporheic flux - Potential for shade
Alluvial Aquifer		
Sediment particle size	(same as channel substrate)	- Potential for hyporheic flux
Sediment particle sorting	(same as channel substrate)	- Diversity of subsurface temperature patterns by determining stratigraphy - Extent of hyporheic flux
Aquifer depth	(same as channel pattern)	- Extent of hyporheic flux

and reducing the heat load to the stream (Hostetler 1991, Naiman and others 1992, Davies and Nelson 1994, Li and others 1994). Vegetation also reduces wind speed across the stream channel thereby trapping air against the water surface. This action reduces conductive heat exchange with the atmosphere by decreasing convection and advection of heat energy to the water surface. (Naiman and others 1992). Width influences channel surface area across which heat is exchanged; a greater surface area allows for more rapid conductive heat transfer. Under the same climatic conditions, narrower, deeper channels will not exchange heat with the atmosphere as rapidly as shallow, wide channels. Similarly, riparian vegetation of a given height will shade a larger percentage of a narrow channel than a wide channel.

Buffering processes

Buffering processes may either heat or cool a stream channel, but buffers differ from drivers in several important ways. First, buffers operate by storing heat already in the stream system rather than by adding or removing heat. For instance, buffers may transfer water and heat between the components of the stream (i.e., from the alluvial aquifer to the stream channel), but water and heat are not added to nor withdrawn from the system. Secondly, buffers operate by integrating variation in flow and temperature over time. If water and heat flux into the stream were constant, buffers would have no effect on channel water temperature.

The two-way exchange of water between the alluvial aquifer and stream channel (*hyporheic flow*) is perhaps the most important stream temperature buffer. The magnitude of hyporheic flow in a stream is determined by the stream channel pattern, the structure of the alluvial aquifer, and the variability in the stream hydrograph (White and others 1987, Creuzé des Chatelliers and others 1994, Henry and others 1994, Evans and others 1995, Hendricks and White 1995, Wondzell and Swanson 1996, Evans and Petts 1997, Morrice and others 1997).

Hyporheic flow occurs at three different spatial and temporal scales. At the finest scale (*streambed scale*), hyporheic flow is driven by alternating pool/riffle sequences in the stream channel (Vaux 1968, White and others 1987). Water enters the stream bed (i.e., the top of the alluvial aquifer) at the downstream end of pools, flows through the streambed sediments, and re-emerges at a downstream riffle (Figure 3). Channels with complex streambed topography have higher rates of streambed hyporheic flow (Harvey and Bencala 1993). Streams with relative little streambed complexity may lack the pool/riffle sequences that drive streambed hyporheic flow. Streambed scale hyporheic flow pathways may be anywhere from 10^{-2} to 10^{1} days in duration.

At an intermediate spatial scale (*meander-bend scale*) hyporheic flow is driven by the development of mid-channel bars and meander bends (Wroblicky and others 1994) and by the presence of side channels, backwaters, and abandoned channels (Stanford and others 1994). Water enters the upstream end of a gravel or sand bar, flows through the underlying alluvium, and re-emerges into the stream at the downstream end. Similarly, hyporheic water follows preferential flow pathways underneath abandoned channels or flood channels and re-emerges in backwaters and side channels or as springbrooks on the floodplain (Stanford and Ward 1992). Stream sinuosity and the presence of geomorphic features such as side channels, flood channels, and backwaters are critical influences on the magnitude of hyporheic flow at the meander-bend scale. Hyporheic flowpath duration at this scale may be anywhere from 10^{0} to 10^{3} days in duration.

Since hyporheic flow pathways at streambed and meander-bend scales are short in duration and are often spatially distinct from the phreatic groundwater flow network, hyporheic flow can buffer channel water temperature. Hyporheic pathways at these scales do not allow hyporheic water temperatures to equilibrate with mean groundwater temperature before re-emerging into the stream. For instance, if a hyporheic flow pathways is four months in duration, the temperature of emerging hyporheic water may be very close to the channel temperature from four months ago (C. Frissell, University of Montana, unpublished data). Since river temperature fluctuates in diel cycles, the most significant

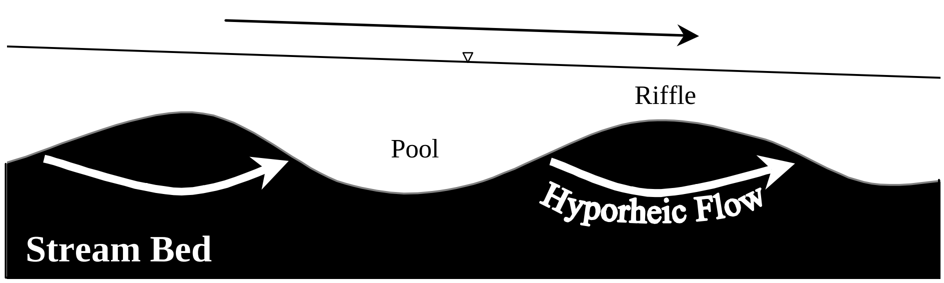

Figure 3. Downstream vertical profile of a stream showing hyporheic flow through the streambed.

buffering effect of streambed scale hyporheic flow occurs when water from the alluvial aquifer re-enters the channel at a time of day opposite that of its entry into the aquifer. Similarly, meander-bend scale hyporheic flow will be most effective as a temperature buffer if water infiltrates and re-emerges at opposite times of the year. Thus, hyporheic exchange results in a horizontal and vertical mosaic of groundwater temperature across the alluvial aquifer, the pattern of which is determined by alluvial aquifer structure, stream channel morphology, and variations in water flow and temperature (White and others 1987, Stanford and others 1994, Evans and others 1995, Evans and Petts 1997). Because of intra- and inter-day variations in stream temperature, streambed and meander-bend flow pathways of virtually any duration have the potential to buffer stream temperature.

At the coarsest scale (*floodplain scale*) water tends to enter the alluvial aquifer at the upstream end of floodplains, flow laterally through the alluvial aquifer, and re-emerge at the lower end of the floodplain (Stanford and Ward 1993). Valley morphology and sediment characteristics are the primary drivers of hyporheic flow at this scale. Where river morphology alternates between reaches confined by bedrock and those with well-developed floodplains, floodplain-scale hyporheic flow is apt to be common. Hyporheic flow duration at the floodplain scale may perhaps be on the order of 10^2 to 10^5 days. At this scale, however, the distinction between hyporheic flow and catchment aquifer recharge from the stream is blurred. Floodplain scale hyporheic flow arguably may better conceptualized as more “classic” aquifer recharge/discharge dynamics depending on the duration and magnitude of the flow dynamics.

The flow path duration of floodplain scale hyporheic flow is likely long enough to allow temperature to equilibrate with the mean subsurface temperature. Therefore, floodplain scale hyporheic flow can be viewed either as a buffer or a cooling effect depending on the time-scale in question. In either case, floodplain scale hyporheic flow affects stream water temperature by extracting water of varying temperature from the channel and returning that water to the channel at a relatively constant temperature approximating mean annual air temperature.

The hydrograph of the stream also plays an important role in driving hyporheic exchange of water. Although hyporheic exchange (both recharge and discharge of the alluvial aquifer) occurs year-round, the *net* recharge to the alluvial aquifer varies seasonally depending on the flow regime in the channel (Creuzé des Chatelliers and others 1994, Hendricks and White 1995, Morrice and others 1997, Wroblicky and others 1998). Positive net recharge generally occurs during high-flow periods; negative net recharge occurs during periods of low flow. In streams where flood spates occur during winter and spring months, the highest aquifer recharge period occurs while the stream channel is coldest. In these systems, hyporheic exchange and floodplain storage of floodwaters may be an especially effective buffer against stream channel warming because the aquifer is recharged predominantly with cold water. This cold water is discharged to the stream during baseflow periods when the highest stream temperatures are apt to occur.

Variation in Temperature Dynamics

Over time, humans have substantively altered the structure of stream systems and the physical context through which streams flow. It is sometimes difficult to imagine the historic structure of streams based on an examination of their current state. A conceptual understanding of the processes and structures that influence stream temperature in unaltered systems can provide a framework from which to understand the breadth of human activities that may substantively influence stream temperature. The following discussion attempts to provide a brief synopsis of stream and catchment dynamics that influence stream temperature and a discussion of how those dynamics are influenced by the natural diversity in stream structure.

The physical structure of stream channels, riparian zones, and alluvial aquifers changes along the continuum from headwaters to river mouth (Vannote and others 1980, Creuzé des Chatelliers and others 1994). For a summary of the ecological implications of structural changes from low-order (headwater streams) to mid-order to high-order (mainstem rivers) streams, see Naiman et al. (1992). As stream structure changes, the processes that drive and mediate stream temperature vary in their relative importance. Generally speaking, as streams become larger, insulating processes become less effective and buffering processes (which are driven by stream morphology) become more important.

Low-order Streams

While notable exceptions exist (e.g. alpine meadow streams), headwater streams, as a rule, have smaller, steeper, narrower channels and narrower riparian areas. These small channels generally carry small amounts of water and therefore (in the absence of processes that cool, insulate, or buffer the stream) experience wide temperature swings as they exchange even relatively small amounts of heat with the atmosphere. Substrate particle sizes in the alluvial aquifer of low-order streams are generally coarse suggesting that there is little resistance to the flux of water between the stream bed and stream channel, subsurface flow rates are high (D'Angelo and others 1993) and subsurface residence times are short. However, the alluvial aquifer may be poorly developed. Limited aquifer size combined with the low porosity of coarse alluvium results in limited potential for water storage in the alluvial aquifer.

Small channels, on the other hand, are easily shaded by topography and riparian vegetation, which provides substantial resistance to the exchange of heat with the atmosphere. Small streams derive a large percentage of their water from lateral groundwater inputs except during snowmelt periods and heavy precipitation events. These lateral groundwater inputs can provide substantial thermal stability during periods of low flow.

Since most headwater streams generally lack significant alluvial aquifers, hyporheic flow occurs predominantly at the streambed scale. In forested streams, individual pieces of large wood (trunks of from fallen trees) lodge in the channel and trap sediments that would otherwise be washed downstream (Beschta and Platts 1986, Montgomery and Buffington 1993, Nakamura and Swanson 1993). Additionally, large wood creates turbulent flow that contributes substantially to variation in streambed topography – a critical driver of streambed-scale hyporheic flow. Therefore, large wood may play an important, albeit indirect role in buffering small streams against temperature changes by increasing the storage capacity of the alluvial aquifer and by contributing to streambed complexity that drives streambed-scale hyporheic flow.

Mid-order Streams

Moderate gradients and somewhat wider channels characterize mid-order streams. Morphology often alternates between reaches closely confined in their valleys and unconfined reaches that occupy montane flood plains. Substrate particle size is medium to coarse, allowing for substantive hyporheic exchange within and across the streambed, though streambed resistance may be higher than in low-order streams (D'Angelo and others 1993). Alluvial aquifers can be somewhat to very extensive in floodplain reaches. The high porosity of sand/gravel alluvium allows for substantive water storage and transport in these alluvial aquifers, but, relative to headwater streams, finer grained sediments suggest slower (though still rapid) subsurface flow rates and short to moderate residence times.

Because mid-order channels carry more water, their capacity to absorb heat without substantive changes in temperature is higher than low-order streams. However, the somewhat wider channels are less easily shaded by riparian vegetation and have more surface area to exchange heat with the atmosphere. In floodplain reaches, riparian vegetation likely becomes a less effective insulator as the channel widens, the littoral zone widens pushing vegetation away from the low-flow water surface, and topographic shading is reduced as the sides of the valley retreat from the stream. Still, in confined reaches where channels are narrower, riparian vegetation and topographic shade may be important insulators against atmospheric heat exchange while hyporheic buffering capacity is likely reduced. Flow from small tributaries is often the predominant source of lateral water inflow; therefore, the riparian condition of tributaries may play a major role in determining channel temperature in mid-order streams.

Channel pattern and morphology begins to play a key role in buffering channel water temperature on montane floodplains. Sinuosity and the presence or absence of gravel-bars, backwaters, and multiple channels determines the potential for hyporheic flow at the meander-bend scale (Stanford and Ward 1993). Multiple channels also allow for more effective riparian shade (Sedell and Froggatt 1984) since the width of each channel is less than the width of a single channel conducting the same amount of water.

Large wood continues to play an important role in determining stream morphology. Aggregates of large wood act as roughness elements that redirect flow, causing avulsions and creating pools, bars, and side channels (Abbe and Montgomery 1996, Nanson and Knighton 1996). Single pieces of large wood are often mobile and therefore might not store sediments from year to year. However, hydraulic forces in the proximity of large wood continue to contribute to streambed complexity and streambed-scale hyporheic flow.

High-order Streams

Low gradients and wide channels are typical of high-order streams. Although most are single channels today, many high order streams once had complex assemblages of active and seasonally active channels, meander-bends, and oxbow lakes (Sedell and Froggatt 1984). Substrate particle size is typically fine to very fine, reducing the rate of flux into the streambed and alluvial aquifer. Alluvial aquifers are large and well to extremely well developed; combined with the moderate porosity of the sediments, this results in a large potential for water storage in the alluvial aquifer. High-order channels move large amounts of water and therefore can absorb and release relatively large amounts of heat energy without the substantive temperature swings observed in smaller channels. Riparian vegetation and topography generally provide little to no insulation for a wide, single channel with a well-developed littoral zone. The sheer volume of water delivered from upstream may overwhelm temperature effects of lateral inflow from phreatic groundwater sources and tributaries.

The catchment aquifer may influence channel water temperature as much by removing water from the alluvial aquifer as by supplying water to it. Where high-order stream systems lose water to the catchment aquifer, hyporheic exchange may be reduced; water entering the alluvial aquifer from the stream channel is apt to be drawn out of the bottom of the alluvial aquifer rather than being returned to the stream channel. This has the effect of both reducing the amount of water in the stream channel as well as damping a potentially important temperature buffer within the stream system.

Hyporheic flow at the meander-bend and floodplain scales likely provides buffering against temperature changes in the stream and result from the stream's channel pattern and morphology. Meander-bends, side channels and other features such as oxbow lakes enhance floodplain scale hyporheic flow. Variable hydrographs likely play an important roll in alluvial aquifer discharge and recharge. The fine-grained substrate associated with high-order systems has higher resistance to groundwater flow thereby increasing the duration of hyporheic flow paths resulting in discharges from the hyporheic zone being a more constant temperature over the course of the year. Substantial networks of side-channels and mid-channel bar formation allow for the inter-digitation of channels with riparian vegetation, providing a much greater opportunity for shading and other interactions between the channel and riparian zone (Sedell and Froggatt 1984). In short, the complexity of channel patterns across the floodplain creates a diversity of surface and subsurface flow pathways through which water moves downstream. These differential flow rates, when combined with seasonal variation in temperature and river stage, allow for stratification, storage, insulation, and remixing of waters with different temperature within and across the floodplain. The resulting mosaic of water temperatures across the floodplain surface and within the floodplain sediments ultimately buffer the main channel against temperature extremes so long as the physical connections between the floodplain and the stream channel are operational (Ward and Stanford 1995).

Pathways of Human Influence

Several key conclusions can be drawn by understanding how drivers, physical stream characteristics, and resulting insulating and buffering processes influence channel temperature:

1) Human activities that alter the ecological drivers of stream temperature can affect water temperature in stream channels by changing the timing or magnitude of: a) the amount of heat energy delivered to the channel (heat

Table 3. Mechanism and influences of human influence on channel water temperature.

Process / Implication	**Influence and Mechanism**
Reduced phreatic groundwater discharge results in reduced assimilative capacity	- Removal of *upland vegetation* decreases infiltration of groundwater on hillslopes and reduces baseflow in streams. - Pumping *wells* for irrigation or municipal water sources can reduce baseflow in nearby streams and rivers.
Reduced stream and tributary flow during low-flow periods reduces assimilative capacity	- *Water withdrawals* reduce baseflow and draw down the watertable in the alluvial aquifer. - *Dams* alter the flow regime of a river. - Removal of *upland vegetation* results in flashy stream flow. - *Dikes and levies* confine flows that would otherwise interact with the floodplain and recharge the alluvial aquifer.
Simplified alluvial system structure reduces assimilative capacity by reducing hyporheic flow.	- *Dams* reduce peak flows that preventing rejuvenation of alluvial aquifer structure. - Removal of *upland vegetation* increases fine sediment load which clogs gravels and reduces hyporheic exchange. - *Dikes and levies* confine peak flows which eliminates floodplain inundation and rejuvenation of alluvial aquifer structure; channelization severs subsurface flow pathways. - *Riparian management* may remove large woody debris (and its sources) that contributes to streambed complexity.
Simplified channel morphology reduces hyporheic flow and assimilative capacity; wider, consolidated channels are less easily shaded and have greater surface area leading to increased heat load	- Removal of *upland vegetation* increases peak stream power and/or increases sediment volumes altering the interaction between water and sediment regimes and changing channel morphology. - *Dams* remove peak flows that maintain channel morphology - *Dikes and levies* confine flood flows that maintain channel morphology and decrease subsurface floodwater storage and, therefore, reduce groundwater discharge during baseflow periods. - *Riparian management* may remove large woody debris (and its sources) that contributed to streambed complexity.
Reduced riparian vegetation reduces shade and increases heat load.	- *Riparian management* may reduce shade to the channel and may reduce the amount of air trapped by the vegetation, increasing convective and advective heat transfer from the atmosphere to the riparian zone and stream surface.

load), or b) amount of water delivered to the channel (flow regime).

2) The dominant mechanism controlling water temperature differs among stream systems with different structural characteristics (e.g., low-, vs. mid-, vs. high-order; constrained vs. unconstrained; forested vs. non-forested). Therefore, streams with different structural characteristics will differ in their sensitivity to specific human activities that alter ecological drivers and/or stream system structure.

3) The physical structure of streams influences how the water temperature in a stream channel will respond to a given heat load and flow regime. Changing the physical structure of a stream system has the potential to influence both the heat load to the channel *and* the streams ability to withstand a given heat load without substantive increase in channel water temperature (i.e., the stream's "assimilative capacity" for heat).

Dams, water withdrawals, channel engineering (e.g., straightening, bank hardening, diking, etc), and the removal of vegetation (upland or riparian) alter the drivers of stream temperature, the structure of stream systems, or both. Therefore, they are all potential mechanisms by which human activities can influence stream temperature. Table

3 provides a summary of many of these impacts along with their operative mechanisms; Figure 4 is a schematic representation of the web of pathways by with temperature may be increased during low flow periods.

Dams

Dams directly effect temperature downstream depending upon their specific mechanism of water release (top- or bottom-release). When considering stream temperature alone, dams can be operated to provide "desirable" stream temperature regimes directly downstream (e.g. through selective withdrawal of water from varying reservoir depths) (Stanford and Hauer 1992). However, from a broader perspective, other ecologically deleterious impacts from flow regulation (Ward and Stanford 1995) including effects on temperature insulating and buffering processes may not be so easily addressed.

Especially in the western United States, dams often store spring and summer flows for use in irrigation, recreation, and to generate hydropower during periods of peak electrical demand. In basins where water rights are overallocated, there is a tendency for dams to be operated such that summertime flows below dams are severely restricted. Large reductions in flow (sometimes to the point of river stagnation) affect water temperature by reducing or virtually eliminating the stream's assimilative capacity for heat.

Flow regulation also reduces the magnitude of hyporheic flow. For hyporheic flow to act as a temperature buffer, differential storage of heat and water over time must occur. Differential heat and

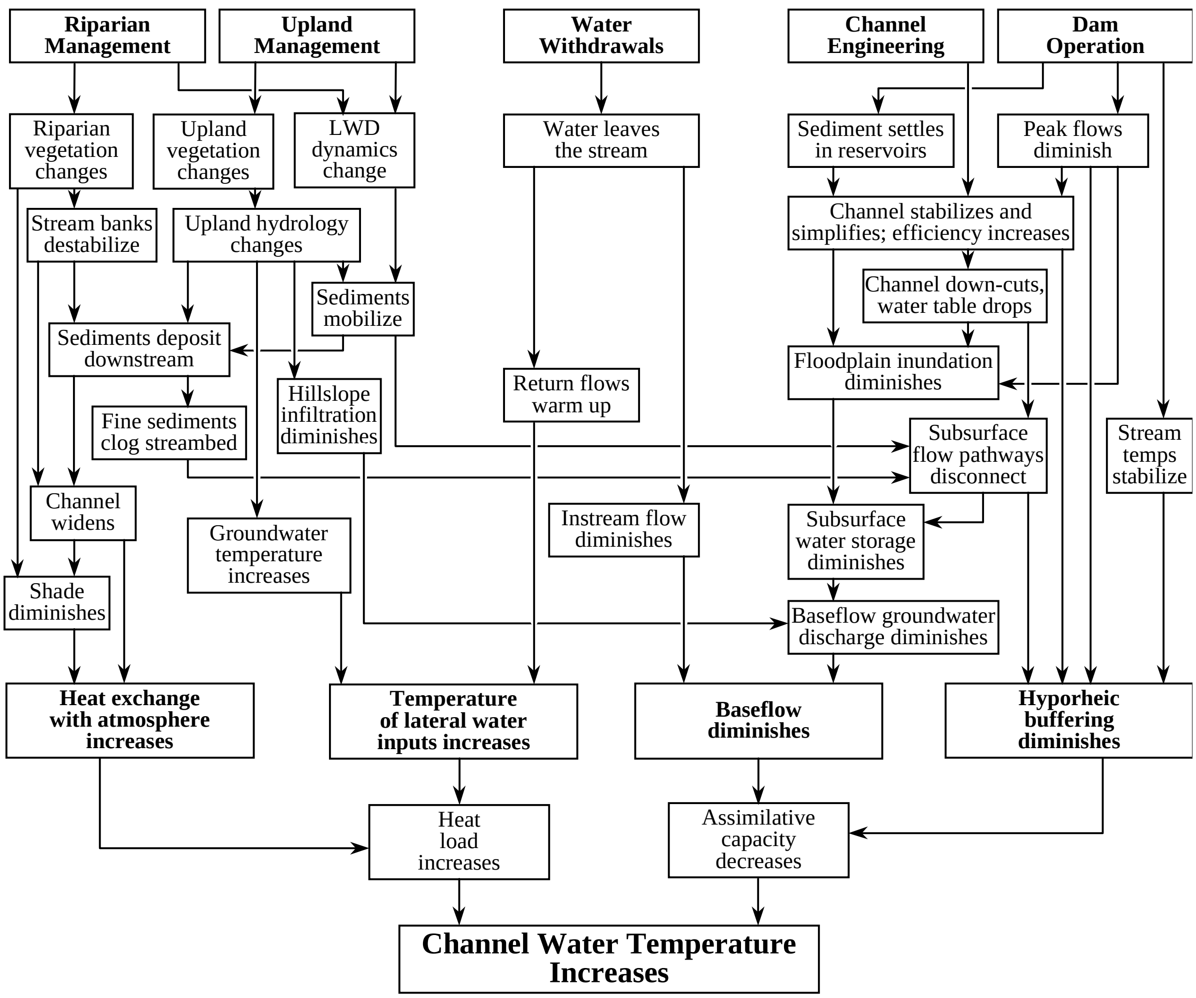

Figure 4. Pathways of human-caused warming of water stream channels.

water storage is driven by *variation* in stream temperature and flow. Since flow regulation dampens variation in both flow and temperature, the potential for hyporheic exchange to act as a temperature buffer is reduced by flow regulation (Ward and Stanford 1995). Dams also affect hyporheic flow by altering the downstream morphology of the channel and geomorphology of the alluvial aquifer. The downstream flux of sediment along the river continuum is disrupted, resulting in downcutting, bed armoring, and, when combined with reduced peak flows, channel stabilization. (Simons 1979, Church 1995). The lack of channel migration and avulsion disrupts fluvial processes critical to creating and maintaining heterogeneous channel patterns (Ward and Stanford 1995, Stanford and others 1996) and alluvial aquifer structure (Creuzé des Chatelliers and others 1994) that drive hyporheic flow at the streambed and meander-bend scales.

To maximize the reservoir storage capacity while minimizing the physical size of the dam, dams are often built at constrictions in rivers just below large alluvial floodplains. Therefore, dams tend to inundate alluvial river segments where hyporheic buffering is most prevalent, thereby reducing the stream's assimilative capacity for heat.

Water Withdrawals

Water withdrawals from reduce instream flow and therefore also reduce the assimilative capacity of streams (Dauble 1994). Although some of this water is eventually returned to the stream, the fraction is typically low. Solley et al. (1993) estimated that only approximately one-third of the water withdrawn in the Pacific Northwest was returned to lakes and streams (as cited in National Research Council 1996). Additionally, water returned to the river after withdrawal is often at a markedly different temperature than it was when withdrawn, thereby affecting the heat load to the stream. The water withdrawals are typically used for industry, municipal water supplies, and agriculture. Regulations may sometimes require that the temperature of industrial and municipal effluent be restored before discharging to the stream, but the fate of water withdrawn for agriculture is less certain. Water from agricultural withdrawals that is not transpired or evaporated will eventually return to the stream. After application, this water sometimes percolates into the phreatic flow network and returns to the stream as groundwater discharge. Although there is the theoretical potential for irrigation to moderate stream temperature by increasing phreatic groundwater inputs to the stream, in practice the impact of the initial reduction in stream flow is not overcome by returning a small fraction of that water through phreatic flow pathways.

Drain tiles are commonly installed in agricultural fields to remove excess water from the soil after irrigation. Water flowing out of these drain tiles usually enters a network of artificial ditches, which deliver the water back to the stream. The temperature of these returns can differ substantially from stream temperatures, further exacerbating the temperature affects of agricultural withdrawals (Dauble 1994, National Research Council 1996).

Major withdrawals from wells penetrating the phreatic groundwater network feeding a stream may reduce flows in a stream channel (Glennon 1995, Wilber and others 1996, Bouwer and Maddock 1997). Additionally, withdrawals via wells can result in the loss of hyporheic water to the larger phreatic groundwater system (Hibbs and Sharp 1992). Therefore, a substantial influence on water temperature may *precede* marked reductions in in-channel flows due to changes in the groundwater flow within the alluvial aquifer and changes in net water exchange between the hyporheic zone and phreatic groundwater system (Long and Nestler 1996). In this case, the buffering capacity of the hyporheic flow network may be substantially reduced because hyporheic water would not be returned to the stream channel to moderate channel-water temperature.

Channel engineering

Straightening, diking, dredging, snagging (removal of large wood), and rip-rapping of channels are all undertaken in an effort to prevent lateral movement of stream channels and increase channel efficiency. These activities focus the erosive energy of streams toward the middle of the channel, encouraging downcutting (National Research Council 1996), and ultimately decreasing the interaction of stream channels with their floodplain in all but extreme flood events. This loss of ecological connectivity between the channel and floodplain can occur through one or all of the following mechanisms. First, since engineered channels carry water more efficiently, both the amount of time floodwaters spend on the floodplain and the surface area inundated is reduced during average annual high-flow events. This action reduces the opportunity for floodwaters to penetrate the alluvial aquifer (Steiger and others 1998) and, in turn, decreases baseflow by reducing groundwater discharge during the low-flow season. Second, engineered channels typically lack heterogeneity in channel pattern and streambed topography (Jurajda 1995), thereby reducing hyporheic flow (See *Physical Structure of Streams*, above). Third, removal of large wood from the channel eliminates major structural elements responsible for creating channel pattern heterogeneity (Sedell and Froggatt 1984, Abbe and Montgomery 1996, Piegay and Gurnell 1997). Fourth, when downcutting occurs, the stream bed is lowered; stream water no longer reaches the floodplain surface and existing subsurface preferential flow pathways can be disconnected from the stream channel (Wyzga 1993). In a manner similar to flow regulation below dams, channel modifications sever linkages between the channel and floodplain thereby reducing groundwater buffering of stream flow and temperature (Ward 1998a) as well as eliminating interactions between the channel and riparian zone that would insulate the stream from exchange of heat with the atmosphere.

Upland vegetation

Whether the catchment of a stream is urban, forested, rangeland, or agriculture, disturbance of upland vegetation associated with human activities has the tendency to increase sediment delivery, warm lateral water inputs, alter the relative amount of surface runoff (and therefore, peak flows), and alter upland water infiltration and groundwater recharge. (Naiman and others 1992, National Research Council 1996). When considering stream channel temperature, perhaps the most pervasive and best studied effect of upland land use is arguably the change in channel morphology (usually widening and shallowing of channels) in response to increased sediment load (Dose and Roper 1994, Knapp and Matthews 1996, Richards and others 1996, Sidle and Sharma 1996). Wider channels have greater surface area and are not as easily shaded by riparian vegetation, thereby facilitating the exchange of heat with the atmosphere. Increasing sediment load can also clog coarse streambed gravels with fine sediments (Megahan and others 1992) thereby decreasing streambed conductivity and reducing the exchange of groundwater and surface water across the streambed (Schälchli 1992). Depending on basin characteristics and the nature of the land use, upland land-use may augment (Harr and others 1982, Ziemer and Keppeler 1990) or reduce (Harr 1980, Burt and Swank 1992) baseflows thereby altering the assimilative capacity of the stream and the stream's erosive power. When stream power is altered, the historic channel morphology is likely to be disrupted, altering the physical structure of the stream and therefore the dynamics of heating, cooling, and temperature buffering. Where shallow groundwater systems are important sources of stream water, removal of vegetation in the catchment can alter upland groundwater temperatures, increasing the temperature of water delivered to the stream (Hewlett and Fortson 1982).

Riparian Vegetation

Removal or alteration of riparian vegetation can have important implications for stream temperature (Beschta and Taylor 1988, Hostetler 1991, Naiman 1992, National Research Council 1996). The primary mechanism by which riparian vegetation controls temperature is through insulation (i.e., shading the stream and trapping air next to the stream surface). However, riparian vegetation removal can also destabilize streambanks, thereby facilitating erosion, increasing sediment loads, and ultimately changing the physical structure of the stream (Li and others 1994). These actions may alter the rate of heat exchange with the atmosphere and restrict hyporheic flow by reducing streambed permeability. Loss of riparian vegetation may have major consequences for in-channel processes for forested streams since riparian vegetation is the primary source of large wood to the channel. The size of large wood (Ralph and others 1994, Hauer and others 1999) and rate of large wood recruitment determine its influence on the channel; therefore current land-use practices such as the selective removal of standing riparian vegetation may have important ramifications for channel morphology (and therefore channel temperature) over time.

Stream Temperature Response to Anthropogenic Influence

Without an understanding of expected patterns of response, we are more apt to attempt to study and monitor stream temperatures in the wrong way, at the wrong location, or at the wrong time. Given a more comprehensive understanding of stream temperature dynamics, we can begin to describe the expected response of stream temperatures to anthropogenic influence. Clearly, it is possible for anthropogenic actions to change the average daily temperature of a stream at any particular sampling location. However, different measures (for instance, variation in temperature) may be more sensitive to anthropogenic influences and therefore may occur long before a measurable change in average stream temperature. Here we discuss three expected patterns of stream temperature change that may be ecologically significant, but could easily fail to be captured by monitoring experiments not designed specifically to detect them: 1) increased amplitude in diel temperature swings; 2) loss of spatial temperature variability at the habitat-unit and stream segment scales (sensu Frissell and others 1986); and 3) variable response in stream temperature along the downstream profile.

As is the case with almost any buffers, any reduction in buffer efficiency results in larger swings in cyclical response patterns within the buffered system. Temperature is no exception. Anthropogenic reductions in the efficiency of stream temperature buffers will likely result in higher maximum and lower minimum daily and seasonal temperatures in the stream. Monitoring methods that do not provide a means of capturing daily maxima and minima (such as continuous data recording across days and/or seasons) will be insufficient to document this expected change.

Spatial variability in temperature within stream reaches may provide localized “refugia” against stream temperature extremes for fishes and other organisms (Gibson 1966, Kaya and others 1977, Berman and Quinn 1991). Localized temperature variation is driven by habitat heterogeneity (Cavallo 1997) and the associated changes in the relative influence of stream temperature drivers across small (10^0 to 10^2 m) spatial scales. Simplification of localized habitat structure (dredging, diking, bank hardening, etc) will reduce localized habitat and therefore temperature variability. Loss of small-scale refugia will affect an organism’s ability to avoid undesirable temperatures associated with diel temperature fluctuations, potentially rendering good habitat marginal, and marginal habitat unusable. Similarly, changes in variability along the downstream profile are likely to affect the spatial variability and distribution of organisms along the stream (Theurer and others 1985, Roper and others 1994, Torgersen and others 1999). If interruption of buffering processes results in a reduction in thermal stability in stream segments that act as refugia, habitat variability and quality is apt to be reduced. Monitoring programs that do not first document and then monitor existing thermal variability at multiple scales will not be able to document changes in spatial temperature patterns over time. Given our growing understanding of the importance of thermal heterogeneity across multiple spatial scales, it seems clear that monitoring programs may be inadequate it they cannot capture expected changes in the thermal variability of streams.

Streams temperature may respond differently to anthropogenic impacts in different parts of the stream. For instance, where streams naturally trend towards average air temperature along their downstream profile (Sullivan and Adams 1991), stream temperatures may be dominated by groundwater (or snow melt) temperature in the stream’s headwaters and by mean daily air temperature near the stream’s mouth (Figure 5A). Therefore, alteration of processes determining heat transfer rates may not drastically affect stream temperatures at the top or perhaps even the bottom of the stream. Rather, the most dramatic (and perhaps most measurable) change may occur in the middle reaches where the stream’s temperature regime transitions from being dominated by groundwater temperature to being dominated by air temperature (Figure 5B). This could drastically reduce the length of stream that contains usable habitat if the temperature change occurs in a critical range for stream biota (Figure 5C) even

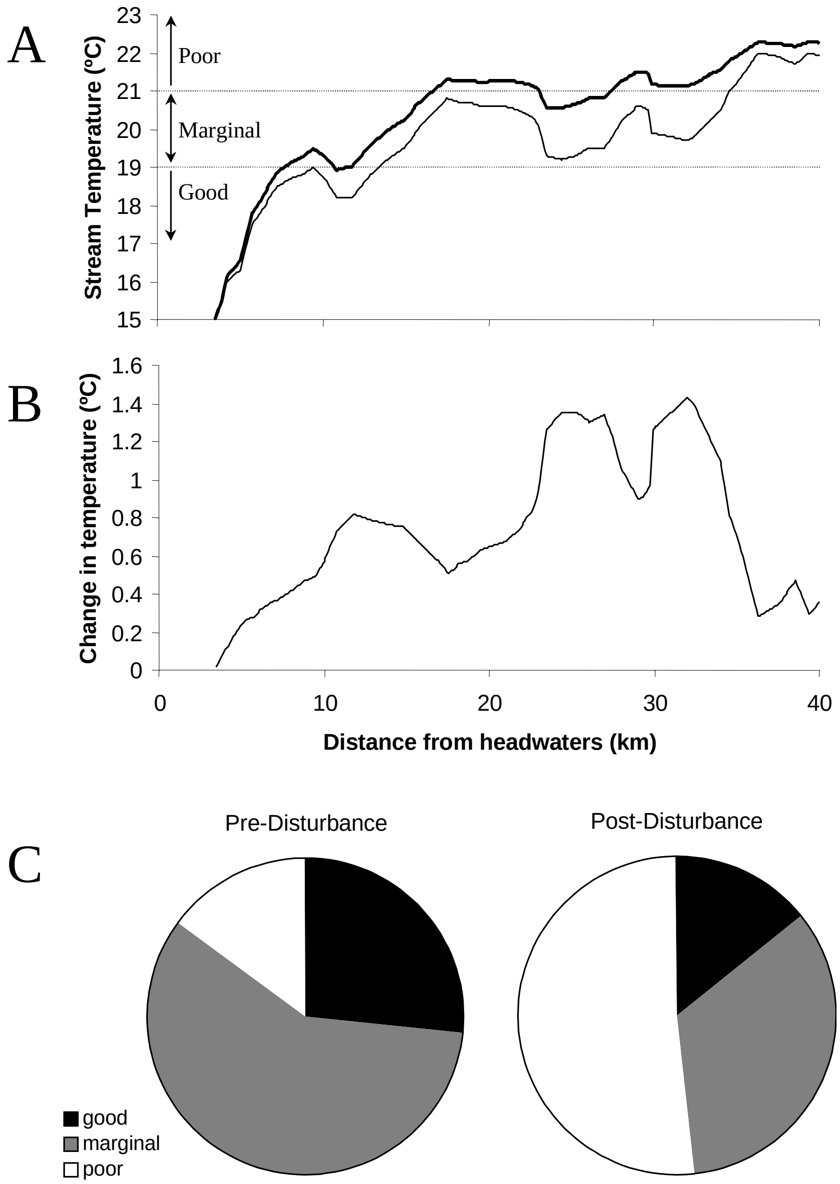

Figure 5. Quantitative depiction of results from a conceptual model of stream warming in response to a hypothetical disturbance resulting in a cumulative 0.25% increase per km of stream in the rate at which water equilibrates with mean air temperature (22.5 ºC). A) Pre-disturbance (thin line – see Figure 2) and post-disturbance (thick line) stream temperature. Zones demarcated by dashed lines show associated habitat quality of a hypothetical species of concern. B) Increase in stream temperature resulting from the hypothetical disturbance. C) Change in thermal quality of habitat after disturbance.

though the change in temperature at the mouth of the stream is limited. In essence, loss of insulating and buffering processes can reduce the distance which groundwater temperature dominance extends downstream thereby decreasing thermal stability in the stream.

Similarly, stream temperature drivers may have different relative influences in different stream segments (see *Variation in Temperature Dynamics*, above). Anthropogenic influences that affect a particular driver or stream structure cannot be expected to influence stream temperature where the driver or structure is not influencing stream temperature. Results from experiments designed to test for stream temperature response to anthropogenic influence will provide conflicting results if investigators do not account for expected variable response to anthropogenic influence. Further, monitoring programs that do not take measurements in areas prone to temperature change will not detect ongoing degradation as soon as might otherwise be documented.

Management of Stream Water Temperature

A holistic understanding of the pathways of human influence on water temperature in stream channels underscores the need for an integrated approach to managing and restoring channel water temperature. To be effective, management programs designed to prevent degradation of water temperature or restore previously degraded systems should consider the breadth of practices occurring in the basin to determine which are the most influential on water temperature. Restoration of historic channel structures, channel-forming processes, sediment delivery, and flow regimes (Stanford and others 1996, Poff and others 1997) may be critical to the re-establishment of historic temperature regimes in large rivers.

Not all of the pathways illustrated in Figure 4 are likely to operate in any one catchment. Determining which human activities have been or may be most influential on water temperature is important for designing an effective management strategy. To accomplish this, watershed analysis is a powerful assessment tool (Montgomery and others 1995). The analysis should include an assessment of historic stream structures and processes, thereby providing a referent for assessing the present-day influences on stream temperature (Kondolf and Larson 1995). It should also attempt to document, in a spatially explicit manner, the historic channel morphology, riparian structure, and extent of the alluvial aquifer along the stream network. An assessment of management history and ongoing activities within the basin (Wissmar and others 1994b) is useful for interpreting identified changes in stream structure and for making strong inferences regarding causal linkages between management activities and degradation of water temperature. Additionally, an analysis of the present day channel morphology, riparian structure, and extent of alluvial aquifer is helpful in prioritizing stream segments for restoration and in the design of effective management prescriptions.

Conclusions

Since stream temperature is a measure of the amount of heat energy per unit volume of water, changing either the amount of heat energy entering the stream or the amount of water flowing in the channel has the potential to alter stream temperature. Further, since a diversity of physical processes in the stream channel, riparian zone, and alluvial aquifer influence the temperature of water in stream systems, degradation of stream temperature can result from modification of external drivers as well as modification of the internal structure of the integrated stream system.

The primary processes determining stream temperature are strongly influenced by the stream's structure. To be effective, management prescriptions designed to restore or protect water temperature dynamics must be aligned with the dominant processes that influence (or historically influenced) channel-water temperatures in a given stream system. For instance, restoration of riparian vegetation will likely not be sufficient to meet temperature standards in streams where degraded channel morphology is the largest cause of undesirable stream temperatures. Instead, processes affecting channel morphology should be addressed. Recovery and protection of stream temperature dynamics would be accomplished best by identifying the dominant historic external drivers and internal structural modifiers of water temperature in a spatially and temporally explicit manner across a basin. This information should then be used to develop spatially explicit management prescriptions relevant to identified human influences.

To be successful, monitoring and research programs need to account for the functional dynamics of stream temperature. Poorly designed programs that do not account for spatial and temporal variability in stream temperature, in the relative influence of various drivers and/or structure, and in the expected response of stream temperature to anthropogenic influences will not ultimately provide reliable answers to relevant scientific or management questions. In short, if we are to better manage and protect valuable aquatic resources from degradation due to altered stream temperatures, scientific and management issues must be set in the context of a more holistic understanding of the functional basis for the expression of stream temperature regimes across space and time.

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