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RIPARIAN MICROCLIMATE AND STREAM TEMPERATURE RESPONSE TO FOREST HARVESTING: A REVIEW[1]

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ABSTRACT: Forest harvesting can increase solar radiation in the riparian zone as well as wind speed and exposure to air advected from clearings, typically causing increases in summertime air, soil, and stream temperatures and decreases in relative humidity. Stream temperature increases following forest harvesting are primarily controlled by changes in insolation but also depend on stream hydrology and channel morphology. Stream temperatures recovered to pre-harvest levels within 10 years in many studies but took longer in others. Leaving riparian buffers can decrease the magnitude of stream temperature increases and changes to riparian microclimate, but substantial warming has been observed for streams within both unthinned and partial retention buffers. A range of studies has demonstrated that streams may or may not cool after flowing from clearings into shaded environments, and further research is required in relation to the factors controlling downstream cooling. Further research is also required on riparian microclimate and its responses to harvesting, the influences of surface/subsurface water exchange on stream and bed temperature regimes, biological implications of temperature changes in headwater streams (both on site and downstream), and methods for quantifying shade and its influence on radiation inputs to streams and riparian zones.
(KEY TERMS: stream temperature; forestry; headwater; riparian; microclimate; water quality; watershed management; Pacific Northwest.)

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INTRODUCTION

Riparian microclimate and stream temperature are critical factors in relation to habitat conditions in and near streams and are governed by the interactions of energy and water exchanges within the riparian zone. Riparian microclimate sets the boundary conditions for many of the energy exchanges that influence stream temperature, while stream temperature sets one of the boundary conditions for riparian microclimate. The two topics are therefore closely linked and are covered together in this paper, which focuses on research relevant to two concerns: (1) forest harvesting may change riparian microclimate and have an impact on aquatic and terrestrial habitat; and (2) forest harvesting, particularly with removal of riparian vegetation, may result in stream heating or other changes in water temperature that could have deleterious effects on aquatic organisms.

Despite decades of research on stream temperature response to forest harvesting, there are still vigorous debates in the Pacific Northwest about the thermal impacts of forestry and how to manage them (e.g., Larson and Larson, 1996; Beschta, 1997; Ice *et al.,* 2004; Johnson, 2004). The conventional approach to minimizing the effects of forest harvesting on streams and their riparian zones is to retain a forested buffer strip along the stream. Most jurisdictions in the Pacific Northwest require buffer strips to be left along larger (usually fish bearing) streams (Young, 2000). However, less protection is afforded to smaller, non-fish-bearing streams. For example, in British Columbia, buffer strips are not required along non-fish bearing streams unless they are a designated community water supply, and buffer strips are not mandatory along the fish bearing streams whose

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bankfull width is less than 1.5 m. Thus, small streams are potentially subject to significant changes in riparian microclimate and particularly to increased solar radiation, which is the major factor driving summertime stream warming.

Beschta *et al.,* (1987) presented an excellent review of the physical and biological aspects of stream temperature in a forestry context, but more recent research has expanded the geographic scope of knowledge within the Pacific Northwest (PNW) region, shed new light on governing processes, or made advances in relation to tools for monitoring and prediction. In the interests of completeness, this paper will revisit much of the material reviewed by Beschta *et al.* (1987) in addition to reviewing more recent studies but will focus on physical aspects. It is assumed that the reader has a basic grounding in microclimatological principles and terminology. Readers lacking this background are referred to Oke (1987) for an excellent introductory treatment.

Given that the primary concern is with riparian management around small streams, the review focuses as much as possible on studies in catchments less than 100 ha in area or streams less than 2 to 3 m wide. It also focuses on studies in the Pacific Northwest region, broadly defined to include northern California, Oregon, Washington, British Columbia, and southeastern Alaska. However, studies from outside the PNW region were considered if they provided useful insights that were not available from local studies. Similarly, studies that did not focus specifically on small forest streams were included if the results were relevant to small stream thermal regimes.

RIPARIAN MICROCLIMATE

Characteristics of Forest Microclimates

Microclimate below forest canopies has been studied extensively for decades, though usually without explicit attention to riparian zones (FAO, 1962; Reifsnyder and Lull, 1965; Jarvis *et al.,* 1976; Rauner, 1976; Geiger *et al.,* 1995; McCaughey *et al.,* 1997; Chen *et al.,* 1999). Compared to open environments, the canopy reduces solar radiation, precipitation, and wind speed near ground level and increases longwave radiation received at the surface. These changes in turn influence the thermal and moisture environments under forest canopies.

Solar radiation transmission through forest canopies depends on the heights of the crown and the density and arrangement of foliage elements (Vézina and Petch, 1964; Reifsnyder and Lull, 1965; Federer, 1971; Black *et al.,* 1991). Reductions in solar radiation under forest cover range from more than 90 percent with dense canopies (Young and Mitchell, 1994; Chen *et al.,* 1995; Brosofske *et al.,* 1997; Davies-Colley *et al.,* 2000) to less than 75 percent in open stands (Örlander and Langvall, 1993; Spittlehouse *et al.,* 2004). The forest canopy changes the spectral distribution of light because plant foliage differentially absorbs and reflects the various wavelengths (Federer and Tanner, 1966; Vézina and Boulter, 1966; Atzet and Waring, 1970; Yang *et al.,* 1993). There is a greater reduction in the ultraviolet and photosynthetically active radiation ranges compared to longer solar radiation wavelengths. Longwave radiation to the forest floor increases as the canopy density increases because the forest canopy is usually warmer than the sky being blocked and has a higher emissivity (Reifsnyder and Lull, 1965). Although this increase somewhat offsets the reduction in solar radiation below the forest canopy, daytime net radiation below forest canopies is usually substantially lower than that in the open.

The amount of precipitation intercepted by the canopy and lost by evaporation depends upon tree species and the amount of canopy cover and typically varies from 10 to 30 percent of annual precipitation (Calder, 1990; McCaughey *et al.,* 1997; Pomeroy and Goodison, 1997; Spittlehouse, 1998). The fraction of precipitation intercepted decreases as storm magnitude and intensity increase. Time since the previous storm and weather conditions during the current storm are also important.

Wind speed under forest canopies is usually 10 to 20 percent of that in large openings (Raynor, 1971; Chen *et al.,* 1995; Davies-Colley *et al.,* 2000). Wind speed within forest openings depends on their size, and openings of less than about 0.1 ha will have low wind speeds, similar to those in the forest (Spittlehouse *et al.,* 2004).

Forest canopies tend to reduce the diurnal air temperature range compared to large open areas. Maximum differences (open area minus area under forest canopy) in daytime air temperature at the 1.5 to 2 m height varied from 3°C (Brosofske *et al.,* 1997; Davies-Colley *et al.,* 2000; Spittlehouse *et al.,* 2004) to 6°C or more (Young and Mitchell, 1994; Chen *et al.,* 1995; Cadenasso *et al.,* 1997). At night, air temperatures in forest areas are typically about 1°C higher than in the open (Chen *et al.,* 1995; Spittlehouse *et al.,* 2004), though Brosofske *et al.* (1997) found temperatures about 1°C lower above a stream. Surface and near-surface soil temperatures show the largest differences between forest and open sites, being up to 10 to 15°C lower under forest canopies during the daytime and

1 to 2°C higher at night (Chen *et al.,* 1995; Brosofske *et al.,* 1997; Spittlehouse *et al.,* 2004).

The vapor pressure of the air is mainly a function of the surrounding air mass and will be similar in the open and the forest. Consequently, the relative humidity and vapor pressure deficit will depend on the air temperature. The lower daytime forest air temperature means that relative humidity is typically 5 to 25 percent higher in the forest (Chen *et al.,* 1995; Brosofske *et al.,* 1997; Davies-Colley *et al.,* 2000; Spittlehouse *et al.,* 2004).

Riparian zones typically have elevated water tables and higher soil moisture than adjacent upland areas. Partly due to these hydrologic conditions, riparian forest cover and understory vegetation often differ from those of uplands, which would influence penetration of solar radiation and interception loss of precipitation. Surrounding slopes may also block direct and diffuse solar radiation. In small headwater streams, the riparian zone may be narrow to nonexistent due to topographic constraints imposed by steep side slopes (Richardson *et al.,* 2005). In addition to the effects of distinctive forest cover and higher soil moisture, riparian microclimate may be influenced by the stream channel, which can provide a local source of water vapor and act as a heat sink during the day, producing locally cooler and moister conditions near the stream (Brosofske *et al.,* 1997; Danehy and Kirpes, 2000). Riparian vegetation may also serve as a source of water vapor via transpiration (Danehy and Kirpes, 2000). Danehy and Kirpes (2000) found that enhanced relative humidity was restricted to a narrow zone within 10 m of the stream edge at 12 forested sites in eastern Oregon and Washington, most likely due to the constraining effects of steep local topography. Another topographic influence that is particularly important in mountain regions is the development of drainage winds that flow down valleys and gullies (Oke, 1987), advecting cool air into lower reaches.

Edge Effects and the Microclimate of Riparian Buffers

The magnitude of harvesting related changes in riparian microclimate will depend on the width of riparian buffers and how far edge effects extend into the buffer. Studies by Chen *et al.* (1993a,b, 1995) in an old-growth Douglas fir forest in Washington state (tree heights 50 to 65 m) are commonly cited in relation to edge effects and required buffer widths. Their results are consistent with those of Ledwith (1996), Brosofske *et al.* (1997), and Hagan and Whitman (2000), as well as with a range of other studies including Raynor (1971) (10.5 m tall red and white pine, closed canopy, New York state), Öerlander and Langvall (1993) (22 to 25 m tall Norway spruce and Scots pine stands of varying density, Sweden), Young and Mitchell (1994) (mixed podocarp-broadleaf forest in New Zealand), Cadenasso *et al.* (1997) (60+-year-old oak, birch, beech, and maple forest in New York state), Davies-Colley *et al.* (2000) (mature, 20 m tall native broadleaved rainforest in New Zealand), and Spittlehouse *et al.* (2004) (25 to 30 m tall Engelmann spruce-subalpine fir forest with a 40 percent canopy cover in British Columbia). All of these studies show that much of the change in microclimate takes place within about one tree height (15 to 60 m) of the edge. Solar radiation, wind speed, and soil temperature adjust to interior forest conditions more rapidly than do air temperature and relative humidity. Nighttime edge temperatures are similar to interior forest conditions. Daytime relative humidity decreases from interior to edge in response to the increased air temperature.

Edge orientation can be important, particularly for a south-facing edge (in the northern hemisphere), where solar radiation can penetrate some distance into the forest for much of the day. Dignan and Bren (2003) found that light penetration diminished rapidly within 10 to 30 m of the buffer edge for a riparian mountain ash forest in Australia, but that light penetration at 10 m was significantly greater for buffers that faced the equator than for other orientations. Wind blowing directly into the edge penetrates farther into the forest than from other directions (Raynor, 1971; Davies-Colley *et al.,* 2000).

Few studies appear to have examined microclimatic conditions within riparian buffers. In a study in northern California, above stream air temperatures measured in the early afternoon decreased with increasing buffer width, at decreases of about 1.6°C per 10 m for buffer widths up to 30 m and 0.2°C per 10 m for buffer widths from 30 m to 150 m (Ledwith, 1996). Above stream temperatures in the 150 m wide buffer treatments were about 6°C lower than at the no-buffer sites. In the same study, relative humidity was 10 to 15 percent higher than at a clear-cut site for 30 m wide buffers and increased another 5 to 10 percent as buffer widths increased to 150 m. At a study conducted at a first-order stream in Maine (Hagan and Whitman, 2000) where a 23 m wide buffer had been left on each side, air temperature 10 m from the stream in the buffer exhibited local differences from the reference sites of up to about 2°C. Differences up to about 4°C were observed within about 10 m from the buffer edge.

Only one study, covering 15 small streams in western Washington, appears to have examined changes in riparian microclimate using both pre-harvest and post-harvest data (Brosofske *et al.,* 1997). Prior to harvest, gradients from the stream into upland areas existed for all variables except solar radiation and wind speed. After harvest, conditions at the edges of riparian buffers tended to approximate those in the interior of the clear-cut. Solar radiation increased substantially within the buffers relative to pre-harvest conditions. Soil surface temperatures were higher after harvest. For buffers less than about 45 m wide (about one tree height), the pre-harvest gradient from riparian zone to upland was interrupted, which could influence habitat conditions for riparian fauna.

THERMAL PROCESSES AND HEADWATER STREAM TEMPERATURE

An understanding of thermal processes is required as a basis for understanding stream temperature dynamics, in particular for interpreting and generalizing from experimental studies of forestry influences. As a parcel of water flows through a stream reach, its temperature will change as a function of energy and water exchanges across the water surface and the streambed and banks (Figure 1) as described by the following equation (modified from Polehn and Kinsel, 2000).

$$\frac{dT_w}{dx} = \frac{\Sigma Q}{\rho C_p v D} + \frac{F_{gw}}{F}\left(T_{gw} - T_w\right) + \frac{F_{hyp}}{F}\left(T_{hyp} - T_w\right) \quad (1)$$

where dT_w/dx is the rate of change in the temperature (˚C) of the water parcel with distance, x(m), as it flows downstream; ΣQ is the net heat exchange by radiation, turbulent exchange, and conduction across the water surface and bed (W/m^2); F is the streamflow (m^3/s); F_{gw} is the ground water inflow rate (m^3/s/m); F_{hyp} is the hyporheic exchange rate (m^3/s/m); T_{gw} and T_{hyp} are the ground water and hyporheic water temperatures, respectively (˚C); ρ is the water density (kg/m^3); C_p is the specific heat of water (J/kg/˚C); v is the local mean velocity (m/s); and D is the local mean depth (m). Equation (1) assumes steady state flow and ignores longitudinal dispersion. It also ignores the heat input of precipitation, which is typically much less than 1 percent of the total energy input to a stream (Webb and Zhang, 1997; Evans *et al.,* 1998). Similarly, frictional heating is neglected because it can be shown to be important relative to other energy exchanges only for steep streams with relatively high flows, under low radiation conditions. This section provides an overview of the dominant processes represented in Equation (1), followed by a discussion of spatial and temporal dynamics of stream temperature regimes.

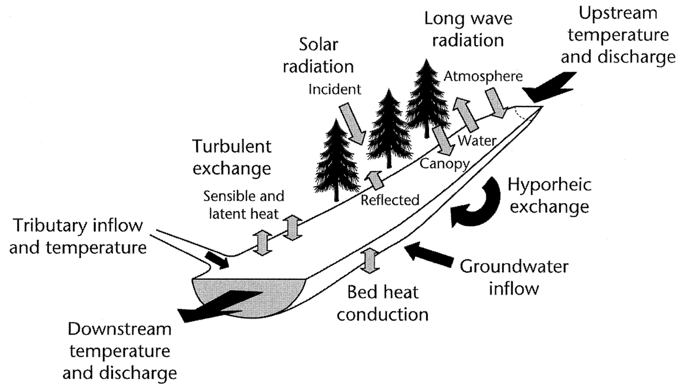

Figure 1. Factors Controlling Stream Temperature. Energy fluxes associated with water exchanges are shown as black arrows.

Radiative Exchanges

Radiation inputs to a stream surface include incoming solar radiation (direct and diffuse) and longwave radiation emitted by the atmosphere, forest canopy, and topography. Canopy cover along the sun's path will reduce the direct component of solar radiation, some of which will be scattered and transmitted through the canopy as diffuse radiation. Transmission of diffuse solar radiation will depend on both the spatial pattern of diffuse radiance from the sky dome and its interactions with the spatial arrangement of canopy elements. The details of solar radiation transmission through canopies are complex. It is often represented by simplified models based on extinction coefficients (e.g., Black *et al.,* 1991; Sridhar *et al.,* 2004) or the spatial distribution of canopy gaps (e.g., Dignan and Bren, 2003). Channel morphology can also influence incident solar radiation at a stream surface. Narrow, incised channels can be effectively shaded by streambanks (Pluhowski, 1972; Webb and Zhang, 1997). Wide channels tend to be less shaded because they have a canopy gap overhead, which will be particularly important for streams oriented north-south.

For solar elevation angles greater than 30 degrees, less than 10 percent of incoming solar radiation will be reflected from the water surface (Oke, 1987). Most incoming solar radiation thus enters the water column, where absorption can occur within the water column and at the bed (Evans *et al.,* 1998). The net effect is that roughly 90 to 95 percent of incident solar radiation is absorbed in the water column or at the bed and thus potentially available for stream heating, except at low solar elevation angles (Evans *et al.,* 1998; Johnson, 2004).

Incoming longwave radiation will be a weighted sum of the emitted radiation from the atmosphere, surrounding terrain, and the canopy, with the weights being their respective view factors (Rutherford *et al.,* 1997). The water surface, canopy, and terrain have high emissivities (typically ≥ 0.95) (Oke, 1987), while the atmospheric emissivity is normally lower, except under overcast conditions. Outgoing longwave radiation includes that emitted by the water surface plus a small fraction (typically 3 to 8 percent) of the incoming longwave radiation that is reflected (Oke, 1987).

Peak daytime net radiation over a stream within a clear-cut can be more than five times greater than that under a forest canopy during summer (Brown, 1969), primarily due to the increase in incident solar radiation. Longwave radiation losses at night may be reduced slightly under forest canopy (Brown, 1969). It has been suggested that longwave radiation losses during autumn and winter may increase following removal (harvest) of forest canopy, leading to more rapid seasonal cooling (e.g., Macdonald *et al.,* 2003b), but this does not appear to have been investigated.

Sensible and Latent Heat Exchanges

Transfers of sensible and latent heat occur by conduction or diffusion and turbulent exchange in the overlying air. Sensible heat exchange depends on the temperature difference between the water surface and overlying air and on the wind speed. Where the stream is warmer than the air, heat transfer away from the stream would be promoted by the unstable temperature stratification, which enhances turbulence. Where the stream is cooler, heat transfer from the air to the stream would be dampened by the stable air temperature stratification (Oke, 1987). Evaporation and associated energy loss occur where the vapor pressure at the water surface (equal to the "saturation" value for the water temperature) exceeds the vapor pressure in the overlying air (a function of the air temperature and relative humidity); condensation and associated energy gain occur where the vapor pressure of the air exceeds the vapor pressure at the water surface. Latent heat exchange also depends on atmospheric stability over the stream.

Most field and modeling studies have used empirical "wind functions" to compute sensible and latent heat fluxes over small streams (e.g., Brown, 1969; Rutherford *et al.,* 1997; Webb and Zhang, 1997; Evans *et al.,* 1998; Johnson, 2004; Moore *et al.,* 2005). There can be great uncertainty in fluxes computed from wind functions, particularly because mean wind speeds under canopies may be less than the stall speed of typical anemometers (Story *et al.,* 2003).

Under intact forest cover, lack of ventilation appears to limit the absolute magnitude of sensible and latent heat exchanges over small streams (Brown, 1969; Webb and Zhang, 1997; Story *et al.,* 2003). Even at open sites such as clear-cuts, sensible and latent heat fluxes over small streams may be limited by bank sheltering, particularly for narrow, incised channels (Gulliver and Stefan, 1986). Brown (1969) and Moore *et al.* (2005) estimated the sensible and latent heat exchanges to be an order of magnitude lower than net radiation on sunny days in recent clear-cuts at coastal sites. Johnson (2004) computed higher values for latent heat flux at a stream in a recovering clear-cut in the Oregon Cascades, though it was still an order of magnitude lower than incident solar radiation.

Bed Heat Exchanges and Thermal Regime of the Streambed

Radiative energy absorbed at the streambed may be transferred to the water column by conduction and turbulent exchange and into the bed sediments directly by conduction and indirectly by advection (in locations where water infiltrates the bed). Given that turbulent exchange is more effective at transferring heat than conduction and that the flowing portions of streams are fully turbulent, much of the energy absorbed at the bed is transferred into the water column, and the temperature at the surface of the bed will generally be close to the temperature of the water column (Sinokrot and Stefan, 1993), except perhaps in pools with upwelling ground water or hyporheic exchange flow.

Bed heat conduction depends on the temperature gradients within the bed and its thermal conductivity and will normally act as a cooling influence on summer days and a warming influence at night, thus tending to reduce diurnal temperature range (Brown, 1985; Moore *et al.,* 2005). For streams within clear-cuts on sunny days, it has been estimated to be approximately 10 percent of net radiation in a step-pool stream (Moore *et al.,* 2005) and up to 25 percent in a bedrock channel (Brown, 1969). Bed heat conduction should depend on stream-subsurface interactions: stream reaches with upwelling ground water tend to have stronger daytime bed temperature gradients than those without and thus should have higher heat loss by conduction (Silliman and Booth, 1993; Story *et al.,* 2003).

Temperatures within the streambed are significant in their own right, since they may influence conditions for post-spawning egg development and fry emergence, as well as conditions for benthic invertebrates. Ringler and Hall (1975) observed summer bed temperature gradients in three catchments in the Oregon Coast Range. Gradients in an unlogged catchment were negligible. Differences of 2°C between the bed surface and 50 cm depth were observed in the streambed of a catchment subject to 25 percent patchcut with riparian buffers, while bed temperatures in artificial redds in a fully clear-cut catchment reached 21˚C with diurnal variations of up to 7˚C at 25 cm depth and vertical changes of about 8˚C over 50 cm. Bed temperatures varied greatly among locations within the clear-cut, likely due to variations in surface water exchange across the bed (Ringler and Hall, 1975). Consistent with this inference, Moore *et al.* (2005) found that bed temperatures in a step pool unit within a clear-cut followed stream temperature more closely in areas of downwelling flow into the bed than in areas of upwelling flow. Given the documented influence of subsurface hydrology on bed temperatures in a range of stream sizes and types and the potential interactions between stream temperature and stream subsurface exchanges (e.g., Shepherd *et al.,* 1986; White *et al.,* 1987; Silliman and Booth, 1993; Constantz, 1998; Curry *et al.,* 2002; Malcolm *et al.,* 2002; Alexander and Caissie, 2003; Moore *et al.,* 2005), the degree to which post-logging bed temperatures reflect changes in surface temperature likely depends on the local hydrologic environment.

Ground Water Inflow

Ground water is typically cooler than stream water in summer during daytime and warmer during winter and thus acts to moderate seasonal and diurnal stream temperature variations (Webb and Zhang, 1999; Bogan *et al.,* 2003). Forest harvesting can increase soil moisture and ground water levels due to decreased interception losses and transpiration (Hetherington, 1987; Adams *et al.,* 1991). Increases in ground water levels following forest harvesting could act to promote cooling or at least ameliorate warming. Alternatively, several authors have speculated that warming of shallow ground water in clear-cuts could result in heat advection to a stream, exacerbating the effects of increased solar radiation or decreasing the effectiveness of riparian buffers (e.g., Hewlett and Fortson, 1982; Hartman and Scrivener, 1990; Brosofske *et al.,* 1997; Bourque and Pomeroy, 2001), and this process has been incorporated into a catchment scale model of hydrology and water quality (St.-Hilaire *et al.,* 2000). Although there is ongoing research on the thermal response of ground water to forest harvesting (Alexander *et al.,* 2003), no published research appears to have examined ground water discharge and temperature both before and after harvest as a direct test of the ground water warming hypothesis.

Hyporheic Exchange

Hyporheic exchange is a two-way transfer of water between a stream and the saturated sediments in the bed and riparian zone. It often occurs where a stream meanders or where there are marked changes in stream gradient. For example, stream water typically flows into the bed at the top of a riffle and re-emerges at the bottom of the riffle (Harvey and Bencala, 1993). If the temperature of hyporheic water discharging into a stream differs from stream temperature, then hyporheic exchange can influence stream temperature dynamics (Equation 1). Several studies have shown that hyporheic exchange creates local thermal heterogeneity in larger streams (e.g., Bilby, 1984; Malard *et al.,* 2002), and recent studies suggest that it can be important in relation to both local and reach scale temperature patterns in headwater streams (Johnson, 2004; Moore *et al.,* 2005). However, there are significant methodological challenges associated with quantifying rates of hyporheic exchange and its influence on stream temperature (Kasahara and Wondzell, 2003; Story *et al.,* 2003; Moore *et al.,* 2005).

Tributary Inflow

Effects of tributary inflow depend on the temperature difference between inflow and stream temperatures and on the relative contribution to discharge, according to a simple mixing equation.

$$T_m = f_i T_i + (1 - f_i)T_s = T_s + f_i(T_i - T_s) \quad (2)$$

where T_i is the inflow temperature (˚C); T_s is temperature at the upstream end of the reach (˚C); T_m is the temperature of the stream inflow mixture (˚C); and f_i is the ratio of inflow rate to streamflow at the downstream end of the reach. Equation (2) assumes complete mixing and may not be valid in the immediate vicinity and some distance downstream of the tributary mouth, where lateral mixing of the tributary flow with the main stream may be incomplete.

Longitudinal Dispersion and Effects of Pools

Longitudinal dispersion results from the variation in velocity through the cross-section of a stream. It would act to "smooth" temperature waves as they propagate downstream, potentially causing a progressive decrease in the diurnal temperature maximum as clearing heated water flows downstream through forested reaches. It is often assumed to be negligible in modeling studies of both small and large streams (e.g., Sinokrot and Stefan, 1993; Rutherford *et al.,* 1997; Polehn and Kinsel, 2000), but no published studies appear to have evaluated its influence in small streams.

The presence of pools can also potentially influence stream temperatures. Being locally deeper zones, pools would tend to change temperature more slowly than the shallower, flowing portions of the stream. However, Brown (1972) observed that there was incomplete mixing in many pools in pool riffle streams in Oregon such that the effective width and depth of flowing water through pools were much smaller than the pool dimensions. Thermal influences of pools do not appear to have been examined in smaller, steeper step pool streams.

Equilibrium Temperature and Adjustment to Changes in Thermal Environment

For a given set of boundary conditions (e.g., solar radiation, air temperature, humidity, wind speed), there will be an "equilibrium" water temperature that will produce a net energy exchange of zero and thus no further change in temperature as water flows downstream (i.e., $dT_w/dx = 0$; Edinger *et al.,* 1968). For stream water being warmed as it flows through a clear-cut, the equilibrium temperature represents the maximum possible temperature the parcel could achieve within the reach at a given time, assuming that boundary conditions remain constant in time and space. However, equilibrium temperature may not be achieved because the boundary conditions may change in time or space before the water parcel can adjust fully to the thermal environment. The concept applies most simply to streams or time scales for which the energy exchanges across the air/water interface dominate the energy budget (Edinger *et al.,* 1968). Stream temperatures influenced by substantial ground water inputs will be consistently less than equilibrium temperature computed from atmospheric conditions during summer and higher in winter (Bogan *et al.,* 2003). Equilibrium temperatures for unshaded reaches are higher than those under shade during summer afternoons (Bartholow, 2000; Bogan *et al.,* 2003).

The rate at which a parcel of water adjusts to a change in the thermal environment depends on stream depth because for deeper streams, heat would be added to or drawn from a greater volume of water. Shallow streams should thus adjust relatively quickly to a change in thermal environment. In addition, flow velocity influences the length of time the parcel of water is exposed to energy exchanges across the water surface and the bed and thus the extent to which the parcel can adjust fully to its thermal environment within a given reach (Figure 2). Given that the depth and velocity of a stream tend to increase with discharge, the sensitivity of stream temperature to a given set of energy inputs should increase as discharge decreases (Brown, 1985; Beschta *et al.,* 1987; Moore *et al.,* 2005).

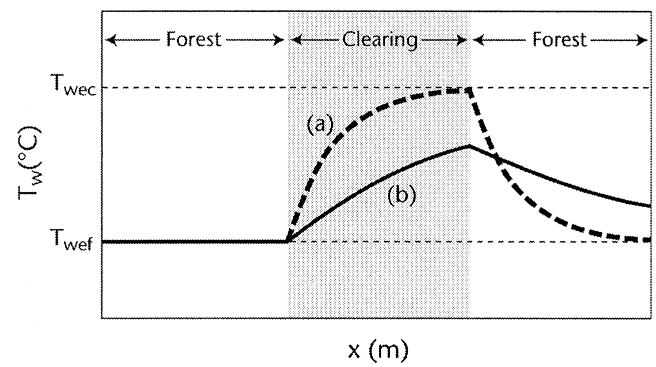

Figure 2. Schematic Temperature Patterns Along a Stream Flowing From Intact Forest, Through a Clear-Cut, and Back Under Intact Forest for (a) Shallow, Low Velocity, and (b) Deep, High Velocity Conditions (T_{wef} = equilibrium temperature in forest; T_{wec} = equilibrium temperature in clearing).

Thermal Trends and Heterogeneity Within Stream Networks

Small forest streams tend to be colder and exhibit less diurnal variability than larger downstream reaches, up to about fourth or fifth order (Vannote and Sweeney, 1980; Holtby and Newcombe, 1982; Macdonald *et al.,* 2003a). Small streams will be more heavily shaded by riparian vegetation and near stream terrain, will have a higher ratio of ground water inflow in a reach to the total downstream flow, and are located at higher elevations and thus experience a generally cooler thermal environment. However, local deviations from a dominant downstream warming trend may occur as a result of ground water inflow, hyporheic exchange, or thermal contrasts between isolated pools and the flowing portion of a stream. In addition, lakes, ponds, and wetlands can produce elevated water temperatures at their outlets, resulting in downstream cooling below them over distances of hundreds of meters, even through cut blocks (Mellina *et al.,* 2002).

Thermal heterogeneity at a range of spatial scales has been well documented in intermediate and large streams (i.e., third order and larger; Bilby, 1984; Arscott *et al.,* 2001; Malard *et al.,* 2001; Ebersole *et al.,* 2003), where it is an important aspect of stream habitat (Neilsen *et al.,* 1994; Ebersole *et al.,* 2003). Thermal heterogeneity in small streams has apparently received less attention, though Story *et al.* (2003) and Moore *et al.* (2005) observed substantial temperature variations in small streams for reaches within a clear-cut and downstream of forest clearings, both along the reach and within channel units.

Stratification of pools can be an ecologically important source of thermal heterogeneity, although its occurrence is variable. Brown (1972) found that only one pool in an intermediate-sized stream with a pool-riffle morphology exhibited significant vertical stratification, with a temperature decrease of 6.5˚C over 1.2 m depth. Nielsen *et al.* (1994) observed more prevalent thermal stratification in pools in three larger rivers in northern California and noted their significance as thermal refugia for steelhead. No published studies appear to have examined stratification of pools in smaller, steeper streams.

STREAM TEMPERATURE RESPONSE TO FOREST MANAGEMENT

The effects of forest management on stream temperature have been estimated using a variety of study designs. The most rigorous approach is the BACI (before-after/control-impact) design, which involves monitoring both before and after treatment and includes untreated control sites (e.g., Harris, 1977). A variation is to use a regression of stream temperature on weather data in place of a calibration with a control catchment (e.g., Holtby and Newcombe, 1982; Curry *et al.,* 2002). Some studies used synoptic surveys of streams that had been subjected to a range of treatments (e.g., Rashin and Graber, 1992; Mellina *et al.,* 2002), while others monitored downstream temperature changes in clear-cuts (Brownlee *et al.,* 1988). This review focuses primarily on studies employing a BACI design, which are summarized in Table 1.

Influences of Forest Harvesting Without Riparian Buffers

Almost all study streams in rain-dominated catchments experienced post-harvest increases in summer temperatures, with increases in summer maximum temperatures ranging up to 13˚C (Table 1). The strong response at Needle Branch may reflect the harsh treatment: clear-cutting to the streambank, slash burning, and removal of wood from the stream. The difference in response between Needle Branch and H.J. Andrews (HJA) Watershed 1, which was subjected to similar treatment, may reflect the differences in aspects (i.e., south for Needle Branch versus northwest for HJA Watershed 1), but other factors also could have influenced the responses. At HJA Watershed 3, where streamside harvesting influenced only part of the stream length, a debris torrent removed riparian vegetation and scoured the channel to bedrock, ultimately leading to similar temperature increases as observed in HJA Watershed 1. At HJA Watersheds 1 and 3, the timing of summer maximum temperatures shifted from August for predisturbance conditions into late June and early July after disturbance, probably because inputs of solar radiation came to dominate other factors such as seasonal variations in discharge (Johnson and Jones, 2000).

In contrast to the results summarized in Table 1, Jackson *et al.* (2001) found that daily maximum temperature for four of seven study streams within clear-cuts in the Washington Coast Range either did not change significantly or decreased following harvesting, likely due to the large volumes of slash that covered the streams and provided shade. However, the post-harvest summer was substantially cooler than the pre-harvest summer, possibly confounding the results.

Effects on summer minimum daily temperatures do not appear to be as marked as those on maximum temperatures, with both small increases and decreases (on the order of 1 to 2˚C) having been reported (e.g., Feller, 1981; Johnson and Jones, 2000). Summer daily temperature ranges after logging have increased up to about 7 to 8˚C, compared to pre-logging ranges of about 1 to 3˚C (Feller, 1981; Johnson and Jones, 2000). Carnation Creek and one of its tributaries experienced smaller increases in diurnal temperature range than found in other studies, but the reason is not obvious from available information (Holtby and Newcombe, 1982).

Fewer studies have examined stream temperature response to forest harvesting in snowmelt-dominated regimes, and no published studies employed a BACI design to estimate effects of no-buffer harvesting in these environments. Brownlee *et al.* (1988) measured downstream increases in summertime mean daily temperature of 1 to 3˚C in three small streams flowing through clear-cuts in the central interior of British Columbia (BC), with increases in daily maximum temperatures of 4.5 to 9˚C on the warmest days. Assuming that downstream temperature changes in these reaches were modest under pre-logging conditions, these upstream/downstream comparisons

TABLE 1. Summary of Experiments Documenting Stream Temperature Changes after Forest Harvesting.

Study Location	Latitude (°N)	Treatment Catchment	Harvesting Type[1]	Riparian Buffer	Aspect	Temperature Variable	Observed Value After Treatment (°C)	Change Due to Treatment (observed-predicted) (°C)	Recovery to Pre-Treatment Conditions	Reference
RAIN DOMINATED										
Oregon Coast Range (Alsea Watershed)	45	Needle Branch Creek (71 ha)	CC (100%)	no buffer	S	Mean of monthly max. T (Apr.-Oct.)	17.5	5.5	~70% recovery in 7 years	Harris, 1977
						Maximum summer T	26	11.6	~70% recovery in 7 years	
Oregon Coast Range (Alsea Watershed)	45	Deer Creek (304 ha)	PC (25%)	30 m	S	Mean of monthly max. T (Apr.-Oct.)	14.5	2.0	No obvious recovery over 7 years	Harris, 1977
British Columbia, Southern Coast Mountains	49	A[2] (59 ha)	CC (20%)	no buffer	SSW	Maximum difference between observed and predicted for daily max. T		5.0	No obvious recovery in 4 years	Moore *et al.*, 2005
British Columbia, Southern Coast Mountains	49	A[2] (23.1 ha)	CC (61%)	no buffer	S	Maximum recorded T	21.8	3.9[3]	Apparently full recovery after 6-7 years	Feller, 1981
British Columbia, Southern Coast Mountains	49	B (68 ha)	CC (19%) followed by slash burn	no buffer	S	Maximum recorded T	20.3	1.8[3]	No apparent recovery after 7 years	Feller, 1981
Oregon Cascades (H.J. Andrews)	45	WS1 (96 ha)	CC (100%)	no buffer	WNW	Maximum summer T	23.9	~7[4]	Apparently full recovery in 15 years	Johnson and Jones, 2000
						Summer mean weekly max. T	not given	5.4 to 6.4 (first 4 years after logging)[4]		
						Summer mean weekly min. T	not given	1.8 to 2.0 (higher)[4]		
Oregon Cascades (H.J. Andrews)	45	WS3 (101 ha)	PC (25%)	Riparian vegetation removed by debris flow after logging	NW	Maximum summer T	23.9	~7[4]	Apparently full recovery in 15 years	Johnson and Jones, 2000

TABLE 1. Summary of Experiments Documenting Stream Temperature Changes after Forest Harvesting (cont'd.).

Study Location	Latitude (°N)	Treatment Catchment	Harvesting Type[1]	Riparian Buffer	Aspect	Temperature Variable	Observed Value After Treatment (°C)	Change Due to Treatment (observed-predicted) (°C)	Recovery to Pre-Treatment Conditions	Reference
RAIN DOMINATED (cont'd.)										
Oregon Cascades (H.J. Andrews) (cont'd.)						Summer mean weekly max. T	not given	3.5 to 5.2 (first 3 years after disturbance)[4]		
						Summer mean weekly min. T	not given	-0.1 to 1.0 (first 3 years after disturbance)[4]		
Oregon Cascades (Bull Run)	45	FC1 (59 ha)	PC (25%) and burned	"sparse strips" left on south banks	W	Maximum summer T	15	3.0	Effect on max. T decreased to < 1°C within 6 years	Harr and Fredriksen, 1988
Oregon Cascades (Bull Run)	45	FC3 (71 ha)	PC (25%)	"sparse strips" left on south banks	SW	Maximum summer T	16	2.5	Effect on max. T decreased to < 1°C within 6 years	Harr and Fredriksen, 1988
Vancouver Island, British Columbia (Carnation Creek)	49	J tributary (24 ha)	CC (100%)	no buffer		Summer (JJA) diel T range	2.3	1.6[5] (after logging)	Only one post-treatment year	Holtby and Newcombe, 1982
							3.2	2.5[5] (after logging and burning)	Only one post-treatment year	
Vancouver Island, British Columbia (Carnation Creek)	49	H tributary (12 ha)	CC (100%)	no buffer		Summer (JJA) diel T range	1.8	1.4[5]	Only one post-treatment year	Holtby and Newcombe, 1982
INTERIOR										
Central Interior of BC (Stuart-Takla FFIP)	55	B5 (42.5 ha)	CC (38%)	10-30 m, all trees > 30 cm dbh harvested	NW	Weekly T_{mean} (max. change)	not given	2.5	No apparent recovery over 5 years	Macdonald *et al.*, 2003b
Central Interior of BC (Stuart-Takla FFIP)	55	B5 (150 ha)	CC (40%)	10-30 m, all trees > 15-20 cm dbh harvested	NW	Weekly T_{mean} (max. change)	not given	3.0	No apparent recovery over 5 years	Macdonald *et al.*, 2003b

TABLE 1. Summary of Experiments Documenting Stream Temperature Changes after Forest Harvesting (cont'd.).

Study Location	Latitude (°N)	Treatment Catchment	Harvesting Type[1]	Riparian Buffer	Aspect	Temperature Variable	Observed Value After Treatment (°C)	Change Due to Treatment (observed-predicted) (°C)	Recovery to Pre-Treatment Conditions	Reference
INTERIOR (cont'd.)										
Central Interior of BC (Stuart-Takla FFIP)	55	B2 (18 ha)	CC (89%)	20 m high retention buffer on lower 60% of stream length within cut block	W	Weekly T_{mean} (max. change)	not given	3.8	No apparent recovery over 5 years	Macdonald *et al.*, 2003
Central Interior of BC (Stuart-Takla FFIP)	55	B1 (313 ha)	CC (6%)	30 m high retention	W	Weekly T_{mean} (max. change)	not given	0.5	No apparent recovery over 5 years	Macdonald *et al.*, 2003b
Central Interior of BC (Stuart-Takla FFIP)	55	G5 (25 ha)	CC (90%)	20 m low retention	NE	Weekly T_{mean} (max. change)	not given	At least 5.4 (missing data)	No apparent recovery over 5 years	Macdonald *et al.*, 2003b
Central Interior of BC (Stuart-Takla FFIP)	55	118-48 (410 ha)	CC (13%)	30 m, all commercial trees harvested	SW	Mean T_{max} in Aug.		0.3	No apparent recovery over 3 years	Mellina *et al.*, 2002
						Mean T_{min} in Aug.		-0.2	No apparent recovery over 3 years	
						T_{max} in Aug.	20.1	2.2	Insufficient info.	
Central Interior of BC (Stuart-Takla FFIP)	55	118-16 (310 ha)	CC (9%)	30 m, all commercial trees harvested	SE	Mean T_{max} in Aug.		0.3	No apparent recovery over 3 years	Mellina *et al.*, 2002
						Mean T_{min} in Aug.		-1.1	No apparent recovery over 3 years	
						T_{max} in Aug.	20.1	5.1	Insufficient info.	

[1]CC = clear-cut, PC = patch cut and number in brackets is % of catchment area treated.
[2]Different creeks with same name.
[3]Computed as difference in maximum observed temperatures between treatment and control streams after logging, compared to difference before logging.
[4]Computed by authors as difference between treatment and control streams due to lack of pre-logging regression.
[5]Computed as difference pre-logging and post-logging for the treatment stream due to lack of calibration with control.

provide an estimate of the effect of clear-cut logging. Winkler *et al.* (2003) inferred similar effect sizes by comparing summer water temperatures for small, high-elevation streams in the southern interior of BC, one in a clear-cut and one in undisturbed forest.

Winter temperatures have received less attention. Feller (1981) found short lived, modest increases in winter temperatures following logging and decreases following logging and slash burning, though there was no clear explanation for these divergent patterns. Post-harvest temperature differences between clear-cut Needle Branch and Flynn Creek (the control) were positive during winter, though smaller than summer differences (Brown and Krygier, 1970). In rain dominated catchments, smaller effects would be expected in winter than in summer, based on the lower energy inputs and higher discharges. In small snowmelt fed catchments, particularly at high elevation or northern sites, ice formation and snow cover within the channel should reduce temperatures to near 0˚C regardless of canopy cover (e.g., Mellina *et al.,* 2002; Macdonald *et al.,* 2003b), except possibly in ground water discharge areas.

Influences of Harvesting With Riparian Buffers

Studies in rain dominated catchments suggest that buffers may reduce but not entirely protect against increases in summer stream temperature. In the Oregon Coast Range, the mean of the summer monthly maximum temperatures increased by only 2˚C at buffered Deer Creek, compared to the 5.5˚C increase observed at unbuffered Needle Branch (Harris, 1977; Table 1). However, this comparison is confounded by the fact that the Deer Creek watershed was 25 percent patch-cut, with only a portion of the stream network adjacent to cut blocks, compared to the 100 percent cutting at Needle Branch. Post-logging increases in maximum summer stream temperature of up to 3˚C were observed at the two Fox Creek streams in the Oregon Cascades, where sparse or partial-retention buffers were left (Harr and Fredriksen, 1988). In the Washington Coast Range, post-harvest changes in daily maximum temperature ranged from -0.5˚C to 2.6˚C for three streams with unthinned buffers (15 to 21 m wide), while streams with buffers of nonmerchantable species warmed by 2.8 to 4.9˚C (Jackson *et al.,* 2001).

Two studies in snowmelt dominated subboreal catchments examined stream temperature response to harvesting with partial retention buffers, both conducted as part of the Stuart-Takla Fish-Forestry Interaction Project in the central interior of BC (Mellina *et al.,* 2002; Macdonald *et al.,* 2003b). Macdonald *et al.* (2003b) reported maximum changes in mean weekly temperatures that ranged from less than 1˚C to more than 5˚C for a set of streams subject to a range of forestry treatments (Table 1). Greater warming was observed for the low retention buffers and a patch retention treatment than for the high retention buffers. The protective effect of the buffers was compromised by significant blowdown, which reduced riparian canopy density from about 35 percent to 10 percent at one high retention buffer and from about 15 percent to less than 5 percent at one low retention buffer. Mellina *et al.* (2002) documented temperature responses to clear-cut logging with riparian buffers for two lake headed streams. Both streams cooled in the downstream direction both before and after logging. Mean August temperatures at the downstream ends of the cut blocks were slightly warmer (less than 1˚C) after logging, although the maximum daily temperature in August increased by more than 5˚C at one stream. The dominant downstream cooling observed both before and after harvest was attributed to the combination of warm source temperatures associated with the lakes and the strong cooling effect of ground water inflow through the clear-cut, as well as the residual shade provided by the partially logged riparian buffer.

Thermal Recovery Through Time

Post-harvest summer stream temperatures should decrease through time as riparian vegetation and shade levels recover. Summers (unpublished, cited in Beschta *et al.,* 1987) found that shade levels at sites that had been clear-cut and burned recovered more rapidly in wetter forest types and at lower elevations. Shade recovery to old-growth levels occurred within about 10 years in the Coast Range western hemlock zone and about 20 years in the Cascade Mountain western hemlock zone. Shade recovery was only 50 percent complete after about 20 years in the higher-elevation Pacific silver fir zone in the Cascades. Shade recovery depends not only on vegetation growth but also stream width: narrow streams should recover more rapidly.

In experimental studies, temperature recovery occurred within 5 to 10 years or was at least under way for several rain dominated streams (Brown and Krygier, 1970; Harris, 1977; Feller, 1981; Harr and Fredriksen, 1988). However, recovery took longer in other cases or was not detectable in the post-harvest period in some cases. Johnson and Jones (2000) found that summer stream temperatures recovered after about 15 years for streams that had their channels and riparian zones disturbed by debris flows in the Oregon Cascades, while Feller (1981) found no evidence of recovery seven years after harvest for a catchment subject to logging and slash burning. In the subboreal environment of B.C., Mellina *et al.* (2002) found no evidence of recovery within the first three years, while Macdonald *et al.* (2003b) found no evidence for recovery of summer temperatures within the first five years following harvesting with partial-retention buffers. Because the streams studied by Macdonald *et al.* (2003b) were well shaded by shrubby vegetation both before and after harvest (E. MacIsaac, Fisheries and Oceans Canada, November 29, 2004, personal communication), it appears that shading by low vegetation may not be as effective at maintaining low stream temperatures as that from trees. In addition, blowdown within the buffers may have contributed to the apparent lack of recovery reported by Macdonald *et al.* (2003b).

Comparison With Studies Outside the Pacific Northwest

Studies of the effects of forestry on stream temperature have been conducted at locations outside the PNW, including Great Britain (Stott and Marks, 2000), eastern and southern United States (e.g., Swift and Messer, 1971; Hewlett and Fortson, 1982; Rishel *et al.,* 1982; Lynch *et al.,* 1984), Quebec (Prevost *et al.,* 1999), and New Zealand (Rowe and Taylor, 1994). Consistent with results from the PNW, these studies have found that streams subject to canopy removal become warmer in the summer and exhibit greater diurnal fluctuations. However, differences in environmental conditions (climate, hydrology, vegetation), forestry treatments, and reported temperature metrics limit the comparability of quantitative results.

Effects of Forest Roads

Forest roads and their rights-of-way would have a similar influence to cut blocks in terms of enhanced solar radiation inputs. Brown *et al.* (1971) observed downstream warming of up to 7°C in a 46 m reach of Deep Cut Creek in Oregon, which was completely cleared of vegetation during road construction. In the central interior of B.C., streams warmed over 2°C across a 50 m right-of-way, 1.4°C across a 30 m right-of-way, and about 0.4°C across a 20 m right-of-way (Herunter *et al.,* 2003). Another possible effect of forest roads is the interception of ground water and its conveyance to a stream via ditches, where it is exposed to solar radiation, effectively replacing the cooling effect of ground water inflow with inflow of warm ditch water. This process has been observed in the central interior of B.C. (D. Maloney, B.C. Ministry of Forests, Northern Interior Region, October 3, 2000, personal communication) and may be most important in low relief terrain, where high water tables could maintain ditch flow during periods of warm weather.

Downstream and Cumulative Effects

The potential for cumulative effects associated with warming of headwater streams is a significant management concern. Beschta and Taylor (1988) demonstrated that forest harvesting between 1955 and 1984 in the 325 km^2 Salmon Creek watershed produced substantial increases in summer water temperature at the mouth of the watershed. Given that current forest practices in the Pacific Northwest require or recommend buffers around all but the smallest streams and require more careful treatment of unstable terrain, cumulative effects resulting from current practices may be of lower magnitude than those found by Beschta and Taylor (1988). At smaller scales, downstream transmission of clearing heated water would increase the spatial extent of thermal impacts and possibly reduce the habitat value of localized cool water areas that form where headwater streams flow into larger, warmer streams, which tend to be cooler and have higher dissolved oxygen concentrations than other types of cool water areas (Bilby, 1984).

Some authors have argued that downstream cooling is unlikely to occur except in association with cooler ground water or tributary inflow (e.g., Beschta *et al.,* 1987), while others have contended that streams can recover their natural thermal regimes within relatively short distances downstream of forest openings (e.g., Zwieniecki and Newton, 1999). Streams can cool in the downstream direction by dissipation of heat out of the water column or via dilution by cool inflows. Dissipation to the atmosphere (and thus out of the stream-riparian system) can occur via sensible and latent heat exchange and longwave radiation from the water surface. Heat loss via evaporation (latent heat) can be a particularly effective dissipation mechanism at higher water temperatures for larger streams (Benner and Beschta, 2000; Mohseni *et al.,* 2002). However, the effectiveness of evaporation may be reduced in small forest streams by negative feedback caused by accumulation of water vapor above the stream due to poor ventilation. Dissipation of heat from the water column into the bed can occur via conduction and hyporheic exchange (assuming the bed and hyporheic zone are cooler than stream water), but reciprocally, these mechanisms would add that heat to the bed and hyporheic zone (Poole *et al.,* 2001). Therefore, cooling of the water column may occur at the expense of warming the streambed and riparian zone, which can influence rates of growth and development of benthic invertebrates and influence salmonid incubation (Vannote and Sweeney, 1980; Crisp, 1990; Malcolm *et al.,* 2002).

Reported downstream temperature changes below forest clearings are highly variable, with some streams cooling but others continuing to warm (e.g., McGurk, 1989; Caldwell *et al.,* 1991; Zwieniecki and Newton, 1999; Story *et al.,* 2003). The maximum cooling reported in the literature was almost 7˚C over a distance of about 120 m (Greene, 1950). The magnitude of downstream cooling may be positively related in some cases to the maximum upstream temperature. Keith *et al.* (1998) found that greater cooling occurred on sunny days, when maximum stream temperatures were greater than 20˚C, than on cloudy days, when maximum stream temperatures were only approximately 13˚C. Storey and Cowley (1997) observed downstream cooling of 1 to 2˚C for two streams in New Zealand where upstream temperatures were 20˚C or greater. In a third stream, which had a narrow margin of forest in the riparian zone upstream of the study reach, upstream temperatures were lower, approximately 17˚C, and no downstream cooling was observed. However, a high upstream temperature does not ensure that downstream cooling will occur, as illustrated by Brown *et al.* (1971), who observed no significant cooling despite an upstream temperature of 29˚C. These studies all employed only post-treatment data, so that even where cooling was observed, there is no basis to assess whether the stream temperature had recovered to pre-logging levels.

Of the studies reviewed, only three attempted to quantify the processes governing downstream temperature changes under shade (Brown *et al.,* 1971; Story *et al.,* 2003; Johnson, 2004). For one clear July day, Brown *et al.* (1971) found that the latent and conductive heat fluxes were the only cooling (negative) terms because ground water inflow was negligible, and these were offset by the warming influences of net radiation and sensible heat, even though the forest canopy substantially reduced inputs of solar radiation. This estimated net input of heat is consistent with the observed lack of significant downstream cooling. Story *et al.* (2003) found that radiative and turbulent energy exchanges at heavily shaded sites on two streams represented a net input of heat during most afternoons and therefore could not explain the observed cooling of up to more than 4˚C over distances of less than 150 m. Instead, downstream decreases in daily maximum temperatures were caused by energy exchanges between the streams and their subsurface environments via ground water inflow, hyporheic exchange, and heat conduction. In contrast, Johnson (2004) demonstrated that downstream cooling could occur in an artificially shaded stream with no ground water inflow or hyporheic exchange. Clearly, more research is required to clarify the mechanisms responsible for downstream cooling and how they respond to local conditions.

Three factors may mitigate against cumulative effects of stream warming. First, although cooling by dilution of streamwater with colder inflow water cannot reduce downstream temperatures to pre-harvest levels, dilution may be great enough, especially at larger spatial scales, to render the changes ecologically insignificant, as long as the total discharge of clearing-heated streams is not a substantial fraction of the total discharge (Equation 2). Second, the effects of energy inputs will not be linearly additive throughout a stream network. This is a consequence of the relation between energy exchange (particularly energy losses via evaporation and longwave radiation) and stream temperature: increased temperatures in one reach due to reduction of riparian shade may reduce the propensity for the stream to warm in downstream reaches, even in the absence of dilution by ground water or tributary inflow. Finally, where streams flow into lakes, ponds, or wetlands, the resetting of stream temperatures may minimize the possibility for cumulative effects below the lentic environment (Ward and Stanford, 1983).

An important aspect of cumulative effects is the indirect impacts of forest harvesting. For example, removing riparian vegetation not only reduces shade but can result in a stream becoming wider and shallower due to bank erosion, which can produce a greater temperature response to the additional heat inputs. Aggradation caused by logging related mass movements and subsequent sediment loading can similarly cause stream widening and promote warming (Beschta and Taylor, 1988). In addition, debris flows that remove vegetation and scour channel beds to bedrock can lead to marked warming in headwater tributaries (Johnson and Jones, 2000).

MONITORING AND PREDICTING STREAM TEMPERATURE AND ITS CAUSAL FACTORS

Successful management of forestry operations for maintenance of stream temperature regimes requires accurate, cost effective tools for monitoring stream temperature and its causal factors and for predicting the effects of different harvesting options.

Monitoring Stream Temperature

Most recent studies have employed submersible temperature loggers to monitor temperature. These are relatively inexpensive and sufficiently accurate (typically within 0.2˚C) for forestry related applications. They also provide sufficient temporal resolution to allow calculation of temperature metrics at a range of time scales, such as maximum daily temperature and accumulated seasonal degree days. Multiple loggers should be used within and downstream of clearings to avoid sampling problems resulting from small scale spatial variability (Story *et al.,* 2003; Moore *et al.,* 2005).

Forward looking infrared radiometry from helicopters has been used for investigating stream temperature patterns in medium to large streams (Torgerson *et al.,* 1999, 2001). However, its application to headwater streams is limited by the sensor resolution relative to typical channel widths for small streams and the fact that low vegetation overhanging the channel may obscure the water surface. However, the technology may be invaluable in identifying cool water areas at tributary mouths and their significance as thermal refugia.

Measuring Shade

Given the importance of solar radiation in causing stream warming following forest harvesting, reliable and practical methods for measuring shade are required for use as indicators of the effectiveness of riparian buffers in protecting against stream temperature changes and for use in predictive models of stream temperature. Many models use canopy and terrain angles, either field measured with a clinometer or estimated from the geometry of the riparian canopy and stream, to determine whether direct solar radiation is blocked. Where blockage by vegetation occurs, the direct radiation reaching the stream is reduced according to estimates of the transmissivity or shade density of the riparian canopy (e.g., Beschta and Weatherred, 1984; Rutherford *et al.,* 1997; Sridhar *et al.,* 2004).

Ocular estimates of canopy cover using instruments such as a spherical densiometer are often used as indices or as model input (e.g., Sullivan *et al.,* 1990; Mellina *et al.,* 2002). Although ocular instruments are generally inexpensive and easy to use in the field, they are prone to operator error due to subjective interpretation. In addition, measurements such as spherical density may not provide a good index of solar radiation blockage except in a uniform canopy. Brazier and Brown (1973) developed an instrument for measuring angular canopy density (ACD), which is the canopy density in the portion of the sky through which the sun passes during the time of maximum potential stream heating, typically July or August, depending on location and hydrologic regime. Teti (2001) described an alternative, robust instrument for measuring ACD based on a convex mirror. Another instrument, the Solar Pathfinder™, focuses on the portion of the canopy responsible for blocking direct solar radiation throughout the day.

Hemispherical photography offers an alternative that is less prone to operator error than ocular methods and allows computation of a range of parameters that are strongly related to solar radiation exposure (Ringold *et al.,* 2003), but it requires off-site analysis. Digital cameras that can be used with fish-eye lenses are steadily decreasing in price, and functional software packages are available both commercially and by free distribution (Frazer *et al.,* 1999).

Shade can also be characterized by comparing radiation or light levels measured above the stream to those at an open site. For example, Webb and Zhang (1997) used a hand-held photographic light meter, following Bartholow (1989), while Davies-Colley and Payne (1998) used a leaf area index canopy analyzer.

Although studies have compared canopy density parameters estimated by different methods (e.g., Englund *et al.,* 2000; Ringold *et al.,* 2003), few studies appear to have assessed which approach provides the best measure of shade for stream temperature assessment. Brazier and Brown (1973) estimated the amount of "heat blockage" caused by the canopy cover in riparian buffers by comparing observed water temperatures to temperatures estimated for a situation of no canopy shade. The good relation between estimated heat blockage and measured ACD confirmed the relevance of ACD as an indicator of buffer effectiveness for temperature control. Rutherford *et al.* (1997) found substantial sampling variability in their shade estimates for a small stream in New Zealand. Using the average field measured shade value in the physically based model STREAMLINE resulted in overestimates of stream temperature. Moore *et al.* (2005) used the spatial distribution of canopy gaps derived from hemispherical canopy photographs, in conjunction with measurements of total and direct solar radiation at an open site, to model the temporal variation of solar irradiance at a stream surface for a clear sky day. Their inability to close a reach scale energy budget may have resulted from sampling bias associated with the canopy photographs but could also have arisen from errors in estimates of the other energy exchanges. Further work is needed to verify predicted solar radiation based on shade measurements, ideally using solar radiation measurements to avoid confounding factors involved in stream heat budgets.

These efforts will be particularly important for application in complex shade environments such as partial-retention riparian buffers or variable retention harvesting units.

In addition to the quantitative measurement of shade, there are questions about shade "quality" in terms of minimizing energy inputs to a stream. For example, Hewlett and Fortson (1982) presented evidence that shade from low, brushy vegetation was less effective than taller trees at moderating water temperatures for a stream in the Georgia Piedmont. Similarly, Macdonald *et al.* (2003b) observed significant temperature increases in central BC despite cover by low vegetation. If these effects are real, it may be that overhanging low vegetation transmits more solar radiation than a coniferous canopy that obstructs the same fraction of sky view, or that it promotes net energy inputs to a stream by influencing longwave radiation and sensible and/or latent heat.

Predicting the Influences of Forest Harvesting on Stream Temperature

Empirical models for predicting stream temperature response to forest harvesting in the PNW include Mitchell's (1999) regression model for predicting the mean monthly stream temperature following complete removal of the riparian canopy, a "temperature screen" for predicting stream temperature as a function of elevation and percent stream shade in Washington (Sullivan *et al.,* 1990) and a multiple regression model that predicts downstream temperature changes as a function of upstream temperature and canopy cover in the central interior of B.C. (Mellina *et al.,* 2002). Although empirical models have the virtues of simplicity and low requirements for input data, they usually involve significant uncertainties, especially when applied to situations different from those represented in the calibration data (e.g., different locations, weather conditions).

Physically based models incorporating energy balance concepts have been developed for application to individual stream reaches, including the seminal model introduced by Brown (1969, 1985), TEMP-84 (Beschta and Weatherred, 1984), TEMPEST (Adams and Sullivan, 1989), Heat Source (Boyd, 1996), and STREAMLINE (Rutherford *et al.,* 1997). Models to simulate stream temperatures at the stream network or catchment scale include SNTEMP (Mattax and Quigley, 1989; Bartholow, 1991, 2000) and a model based on the HSPF (Hydrological Simulation Program – FORTRAN) model developed by the U.S. Environmental Protection Agency and the U.S. Geological Survey (Chen *et al.,* 1998a,b). Other models have been developed, but the ones mentioned are broadly representative of the range of complexity.

Sullivan *et al.* (1990) tested the ability of four reach scale models (Brown's model, TEMP-86, TEMPEST, and SSTEMP) and three catchment scale models (QUAL2E, SNTEMP, and MODEL-Y) to predict forestry related temperature increases in Washington. The catchment scale models required more input data than would be available for operational applications and did not provide accurate temperature predictions. TEMP-86 provided accurate predictions for mean, minimum, and maximum temperatures but required upstream temperatures as input to achieve the high level of performance. TEMPEST was less sensitive to specification of input temperatures, making it more suitable as an operational tool (Sullivan *et al.,* 1990).

Sridhar *et al.* (2004) addressed the problem of unknown upstream temperatures by using a reach length of 1,800 m above the prediction point. For this reach length, the effect of the upstream boundary condition on modeled downstream temperatures became negligible for low flow conditions. However, this approach would not necessarily be appropriate for the headmost streams in the channel network, where the reach of interest may extend only a few hundred meters or less downstream from the channel head. In such cases, an estimate of ground water temperature may be appropriate as an upstream boundary condition.

As mentioned previously, Rutherford *et al.* (1997) found that their model predictions were biased when the mean field measured values for shade were used as input. Although they were able to match the daily maximum and minimum temperatures by increasing the shade values to the maximum observed values, the timing of the diurnal temperature wave was incorrect, suggesting that some process was not properly represented. They hypothesized that flow through gravels (i.e., hyporheic exchange) could have been one of the causes. The significance of hyporheic exchange on reach scale temperature patterns should be investigated further.

DISCUSSION AND CONCLUSIONS

Summary of Forest Harvesting Effects on Microclimate and Stream Temperature

Forest harvesting can increase solar radiation in the riparian zone as well as wind speed and exposure to air advected from clearings, typically causing increases in summertime air, soil, and stream temperatures and decreases in relative humidity. Riparian buffers can help minimize these changes. Edge effects penetrating into a buffer generally decline rapidly within about one tree height into the forest under most circumstances. Solar radiation, soil temperature, and wind speed appear to adjust to forest conditions more rapidly than air temperature and relative humidity.

Clear-cut harvesting can produce significant daytime increases in stream temperature during summer, driven primarily by the increased solar radiation associated with decreased canopy cover but also influenced by channel morphology and stream hydrology. Winter temperature changes have not been as well documented but appear to be smaller in magnitude and sometimes opposite in direction in rain-dominated catchments. Although retention of riparian vegetation can help protect against temperature changes, substantial warming has been observed in streams with both unthinned and partial retention buffers. Road rights-of-way can also produce significant warming. Changes to bed temperature regimes have not been well studied but can be similar to changes in surface water in areas with downwelling flow.

Although the experimental results are qualitatively consistent, it is difficult to make quantitative comparisons of experimental results because the studies have expressed temperature changes using incommensurable temperature metrics. For the studies where similar metrics were available (e.g., maximum summer temperature), treatment effects exhibited substantial variability, even where the treatments appeared to be comparable (e.g., HJA Watershed 1 and Needle Branch). Thus, on their own, experimental results cannot easily be extrapolated to other situations. Application of heat budget models may help to diagnose the reasons for variations in response in experimental studies and provide a tool for confident extrapolation to new situations.

Increased stream temperatures associated with forest harvesting appear to decline to pre-logging levels within five to ten years in many cases, though thermal recovery can take longer in others. There is mixed evidence for the efficacy of low, shrubby vegetation in promoting recovery.

Temperature increases in headwater streams are unlikely to produce substantial changes in the temperatures of larger streams into which they flow, unless the total inflow of clear-cut heated tributaries constitutes a significant proportion of the total flow in the receiving stream. Clearing heated streams may or may not cool when they flow into shaded areas. Where downstream cooling does not occur rapidly, the spatial extent of thermal impacts is effectively extended to lower reaches, which may be fish bearing. In addition, warming of headwater streams could reduce the local cooling effect where they flow into larger streams, thus diminishing the value of those cool water areas as thermal refugia.

Biological Consequences and Implications for Forest Practices

It is difficult to estimate the biological consequences of harvesting related changes in riparian microclimate and stream temperature based on the existing results. In terms of terrestrial ecology in riparian zones, there is incomplete knowledge regarding the numbers of species that are unique to small streams and their riparian zones, as well as their population dynamics, sensitivity to microclimatic changes, and ability to recolonize disturbed habitat (Richardson *et al.,* 2005). The ecological effects of stream temperature changes in small, nonfish bearing streams are also unclear. While it is generally acknowledged that changes in thermal regime can influence macroinvertebrates (Vannote and Sweeney, 1980; Ward and Stanford, 1992), the metrics typically presented for stream temperature changes (e.g., maximum summer temperature) may not be the most biologically significant for streams that remain at sublethal temperatures. Given the emerging appreciation for the role of small streams in providing organic matter to downstream fish bearing reaches (e.g., Wipfli and Gregovich, 2002), a better understanding is required of how changes in the physical conditions in small streams and their interactions with chemical and biological processes influence their downstream exports.

Based on the available studies, a one-tree-height buffer on each side of a stream should be reasonably effective in reducing harvesting impacts on both riparian microclimate and stream temperature. Narrower buffers would provide at least partial protection, but their effectiveness may be compromised by wind throw, and they could still incur costs by complicating access and yarding operations. Alternative approaches to protecting riparian values may be possible that avoid at least some of the problems associated with buffers. For example, in B.C., many companies retain green tree patches within a cut block to provide future wildlife habitat. If these were positioned where they could shade the stream, they could provide at least some of the function of a riparian buffer but perhaps with lower wind throw risk and with less impact on ease of access and yarding.

Issues for Future Research

Riparian microclimates appear to have been relatively little studied, both in general and specifically in relation to the effects of different forest practices. Further research needs to address these knowledge gaps.

Shade is the dominant control on forestry related stream warming, and although algorithms exist for estimating it based on riparian vegetation height and channel geometry, there is a need to refine methods for measuring it in the field and for modeling it. Ground-based hemispherical photographs offer great potential for developing both static indices of shade as well as a tool for modeling the temporal variation of solar transmission as a function of the spatial distribution of canopy gaps. Further research should focus on the application of hemispherical photography, including an assessment of sampling variability and bias. In addition, the effects of low deciduous vegetation on the heat budget of small streams should be examined to help understand and predict trajectories of thermal recovery in time.

Further research should address the thermal implications of surface/subsurface hydrologic interactions. Studies should focus on both the local scale and reach scale effects of heat exchange associated with hyporheic flow paths, particularly those associated with step pool features, which are common in steep headwater streams. Bed temperature patterns in small streams and their relation to stream temperature should be researched, especially in relation to the effects on benthic invertebrates and other nonfish species. The hypothesis that warming of shallow ground water in clear-cuts can contribute to stream warming should be addressed, ideally by a combination of experimental and process/modeling studies.

The physical basis for temperature changes downstream of clearings needs to be clarified. In particular, it may be useful to determine whether diagnostic site factors exist that can predict reaches where cooling will occur. Such information could assist in the identification of "thermal recovery reaches" to limit the downstream propagation of stream warming. It could also help to identify areas within a cut block where shade from a retention patch would have the greatest influence.

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