The Effects of Forest Disturbance on Hydrologic Processes and Watershed Response

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INTRODUCTION

Hydrologic processes that affect the generation of streamflow were described in the previous chapter ("Hydrologic Processes and Watershed Response"). This chapter builds on the understanding of these fundamental processes by describing how changes in forest cover, brought about by logging, insects, disease, fire, or forest regrowth, affect these processes at both the stand and watershed scale. The discussion in this chapter focusses on the water balance, groundwater, water yield, peak and low streamflows, and hydrologic recovery.

STAND- AND HILLSLOPE-SCALE EFFECTS

Surface Processes

Net precipitation

Forest cover significantly influences the amount of precipitation that reaches the ground, as described in Chapter 6 ("Hydrologic Processes and Watershed Response"). The loss of forest cover results in an increase in net precipitation (i.e., gross precipitation minus the amount intercepted and subsequently evaporated or sublimated). The amount of net precipitation depends on stand characteristics (tree species, leaf area, canopy density) and the weather (time since the last rainstorm, existing snow load, air temperature, wind speed, and the size of storm events). Increases in net precipitation may occur immediately after forest disturbance, such as following clearcutting or severe fires, or may occur over a period of years when foliage loss is gradual, as can occur following attack by insects such as the mountain pine beetle.

In British Columbia and similar forested environments, 5–70% more water can accumulate as snow in clearcuts than in the forest, depending on the winter precipitation in a given year and forest cover type (Toews and Gluns 1986; Hudson 2000; Winkler 2001; Pomeroy et al. 2002; Winkler and Moore 2006). At Mayson Lake and Upper Penticton Creek in south-central British Columbia, 47% and 29% higher snow water equivalent (SWE) was measured in a clearcut than in a mature, mixed-species stand, respectively. In the clearcut at Upper Penticton Creek, SWE was only 12% higher in the open than in a mature lodgepole pine stand (Winkler 2001). At these same locations, the largest relative differences in SWE between mature forest and open snow survey sites did not occur in the year of heaviest snowfall, likely because of an upper limit to the interception capacities of these stands and to annual variations in snowfall pattern (few large snowfall events vs. many small events) (Winkler and Moore 2006).

Studies in Colorado indicated that most of the increased snow accumulation in openings occurs during storms, and that the redistribution of intercepted snow by wind between storms is not significant (Wheeler 1987). Comparisons of snow accumulation under forest cover and in clearings, at both the plot and watershed scales, indicated that the increased accumulation in openings was due mainly to a decrease in interception loss rather than preferential deposition by wind patterns (Troendle and Meiman 1984, 1986; Troendle and King 1985; Wheeler 1987). In Alberta, maximum increases in snow accumulation were measured in openings 2–5 tree heights wide (Golding and Swanson 1986).

Losses to evaporation, sublimation, or early melt can be important influences on peak SWE within cutblocks. For example, although more snow accumulated in cutblocks than in the forest on the North Fork of Deadhorse Creek, Colorado, peak SWE averaged over the watershed did not increase significantly following harvesting (Troendle and King 1987). The researchers hypothesized that early ablation in south-facing openings offset the increase in snow accumulation that occurred from reduced interception loss. This effect was also observed in a high-elevation, south-facing clearcut at Upper Penticton Creek, where at the start of the spring melt, less snow remained in the clearcut than in the adjacent forest (R.D. Winkler, unpublished data).

Tree mortality caused by insects or disease can result in reduced interception losses and consequent increases in net precipitation, though not as large as those associated with harvesting. Research into the effects of beetle-killed stands on snow accumulation is under way throughout the interior of British Columbia (Winkler and Boon 2010). Initial results of this research show that averaged over all research sites, maximum SWE is reduced by 24% under mature, green-to-red attacked lodgepole pine relative to the open, and by 13% under grey pine (Table 7.1); however, the wide range in results for both stand conditions highlights the large variability in interception among stands, sites, and years.

Reductions in snow interception and increases in net precipitation are generally proportional to reductions in canopy cover or to percent basal area removal (Harestad and Bunnell 1982; Moore and McCaughey 1997). Maximum snow accumulation, measured as SWE, increased by 16% following 40% basal area removal in lodgepole pine stands in Montana (Woods et al. 2006), and by 21% following 50% removal in similar stands in Colorado (Troendle and King 1987).

Net precipitation may also increase after a forest fire. If only a few small openings are created, the effects on net precipitation may be similar to those following thinning or patch cutting; however, in cases where most of the forest cover is lost, changes in net precipitation can be expected to be similar to those following clearcutting (Neary and Ffolliott 2005). Few studies have quantified the effects of fire on snow accumulation. In a lodgepole pine stand in southern Montana, where fire reduced canopy cover by 90% (from 42 to 4% cover), SWE increased by 9% relative to the adjacent similarly structured unburned forest, the same difference as was observed in the clearcut (Skidmore et al. 1994). Although this increase in SWE was small, the authors suggested

TABLE 7.1 *Percent reduction in maximum snow water equivalent (MSWE) and average ablation rate (AAR) in the forest relative to the open and the number of days difference in timing of snow disappearance in stands affected by mountain pine beetle in the British Columbia Interior (Winkler and Boon 2010, based on data collated in Winkler and Boon 2009)*

		Reduction in MSWE (%)			Reduction in AAR (%)			Difference (days) in snow depletion date (Forest–Open)	
Forest age class (years)	**Attack class**	**No. sites**	**Average**	**Range**	**No. sites**	**Average**	**Range**	**Average**	**Range**
Old (120+)	Green/red	3	22	31 to 6	2	39	42 to 36	7	4 to 10
Mature (40–120)	Green/red	12	26	57 to +9	3	38	48 to 27	2	0 to 3
Intermediate (10–40)	Green/red	12	16	72 to +7	5	22	49 to +7	2	–5 to 9
Old (120+)	Grey	7	11	21 to +9	2	22	29 to 14	9	–1 to 3
Mature (40–120)	Grey	9	16	58 to +28	3	37	57 to 25	3	0 to 6
Intermediate (10–40)	Grey	1	21		1	38		12	12

that larger increases should be expected where stand densities are higher. At Border Lake, south of Keremeos, SWE was 4% higher in a burned stand than in a nearby clearcut (Dobson 2008). At Mayson Lake, near Kamloops, maximum SWE was reduced by 4–11% in a burned stand relative to a clearcut (R.D. Winkler, unpublished data).

Changes in rainfall interception follow a similar pattern to that of snow interception, generally decreasing with increasing loss of forest cover; however, the magnitude of change also depends on the type of forest cover loss. For example, interception in stands where trees are killed by insects and disease may not change until the trees start to lose their needles (Héile et al. 2005; Spittlehouse 2007; Winkler et al. 2008a), whereas disturbance by fire immediately removes much of a stand's rainfall-intercepting capacity (Moore et al. 2008). After harvest, some understorey remains, although interception losses from shrubs, herbs, mosses, slash, and large woody debris tend to be smaller than those from forest canopies (Kelliher et al. 1992). At Mayson Lake, Carlyle-Moses (2007) found no significant difference in rainfall interception loss during the growing season between an unlogged mature pine-spruce-fir stand (20–25 m tall; basal area = 52 m^2/ha) and a portion of the stand where beetle-attacked trees had been removed (remaining basal area = 19 m^2/ha). Similar results have been observed in other forests. For example, Knoche (2005) found that in a 66-year-old Scots pine stand in Germany, rainfall interception losses declined by only 7% (from 38 to 31%) when basal area was reduced by 47% (from 38 to 20 m^2/ha). Thinning of an 11-year-old radiata pine stand in New Zealand reduced rainfall interception by 27% (Whitehead and Kelliher 1991). Both rain and snow interception losses increase as the forest regenerates (see Chapter 6, Table 6.1; see also "Watershed-scale Effects" section below).

Evaporation

Changes in the amount and (or) type of vegetation caused by fires, insects, disease, harvesting, and silvicultural treatments can alter evaporation from the forest (i.e., evaporation or sublimation of intercepted precipitation, transpiration from the vegetation, and evaporation from the soil surface) (Bonan 2008). Understanding how changes in forest cover affect evaporation is critical to comprehend the effects of disturbance on water yield, and on peak and low flows.

Evaporation rates from a wet, bare, exposed soil surface, which often occurs immediately after clearcutting or a fire, may be as high or higher than evaporation rates from a forest (3–4 mm/d) (Novak and Black 1982; Spittlehouse 1989); however, bare soil surfaces usually dry within 1–2 days after a rainfall, and evaporation decreases rapidly (Novak and Black 1982). This results in a dry 0.05–0.10 m surface layer and a moist lower soil profile; consequently, soil moisture stays relatively high throughout the summer compared to the forest (Figure 7.1). Average evaporation rates depend on the frequency of rainfall and on weather conditions. At high-elevation sites in the province's southern interior, bare soil evaporation during intermittent, short, wet and dry periods averaged 1–2 mm/d, similar to that of forests, but rapidly decreased to less than 0.5 mm/d during extended dry periods (Figure 7.2).

Silvicultural practices such as partial retention and thinning remove only part of the forest canopy. If water is not a limiting factor for the site, the relationship between partial vegetation removal and decreases in evaporation is generally considered as linear (Hibbert 1983). When transpiration is water-limited, however, the relationship is not linear. In some situations, where trees are able to use additional water, a threshold level of forest cover must be removed before a change in evaporation loss can be detected (Tang et al. 2003; Bladon et al. 2006; Simonin et al. 2006; Li et al. 2007; Stednick 2008). Canopy openings also allow for increased evaporation from the soil and increased transpiration from understorey vegetation, which compensates somewhat for the reduction in stand-level transpiration caused by the removal of trees (Black 1979; Spittlehouse and Black 1982; Knoche 2005).

Soil moisture levels increase following harvesting (Ziemer 1964; Hart and Lomas 1979; Adams et al. 1991; Elliott et al. 1998; Bhatti et al. 2000). By using a physically based stand water balance model, Spittlehouse (2006) found that summer evaporation from a high-elevation clearcut in the province's southern interior was about 30% less than that from the forest. This lower evaporation rate combined with a reduction in interception losses increased soil moisture in the clearcut and increased the water available for streamflow (Figure 7.3). Similar results were obtained in a clearcut and Douglas-fir stand in coastal British Columbia (Jassal et al. 2009). Liu et al. (2005) found a similar reduction in seasonal and daily evaporation at burned sites in Alaska.

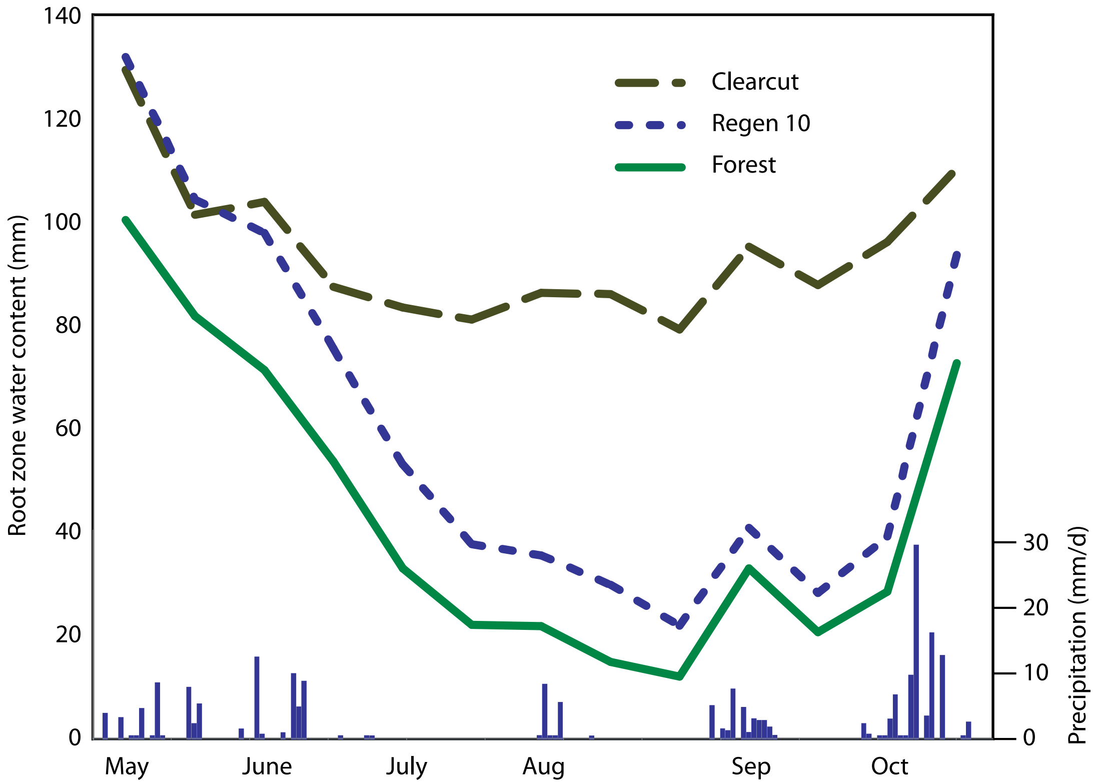

FIGURE 7.1 *Measured root zone soil water storage (mm) and daily precipitation (mm/d) for a lodgepole pine forest (solid line), a clearcut (long dashed line), and a regenerating 10-year-old lodgepole pine stand (short dashed line) at Upper Penticton Creek. Differences in water content between sites in May are the result of differences in water storage capacity (depth and stone content) of the three sites (D.L. Spittlehouse, unpublished data).*

In stands killed by mountain pine beetle, evaporation from the shaded understorey and the forest floor is low, and soils remain moist for most of the summer (Spittlehouse 2007). Numerical modelling indicates that if tree mortality is less than 40%, the water balance (evaporation, soil moisture storage, and drainage) in an attacked stand is similar to an unattacked stand; however, if most of the trees are killed, the stand may have a water balance similar to that of a clearcut (Spittlehouse 2007; Moore et al. 2008).

Changes in vegetation with succession and reforestation may also influence evaporation. Grasses and deciduous plants tend to have a lower canopy resistance to transpiration than do conifers (Kelliher et al. 1993; Bonan 2008) (see Chapter 6, "Hydrologic Processes and Watershed Response"); thus, these plants have a higher transpiration rate than conifers under the same weather conditions. In Australia, Putuhena and Cordery (2000) studied the hydrological effects of clearcutting a watershed covered by a native, dry, sclerophyll eucalypt forest and replacing it with radiata pine. The study results showed that evaporation decreased after clearcutting and then increased with increasing pine age (especially between the 4th and 16th year of growth). The equilibrium evaporation rate reached after approximately 16 years was lower than the rate recorded in the pre-disturbance eucalypt forest. These results are consistent with data for the province's southern interior (Figures 7.2 and 7.3).

Forest practices that alter or remove vegetation affect the amount of evaporation by changing interception losses, transpiration, and direct evaporation from the mineral soil. Decreased evaporative losses result in increased water storage (e.g., soil moisture, groundwater) within watersheds. Higher antecedent soil moisture increases the potential for greater water yield and for more rapid and higher peak flows (Bosch and Hewlett 1982; Iida et al. 2005). The magnitude of such increases varies from small to large depending on available watershed storage capacity relative to the volume of increased water input. In addition, the total area and spatial distribution of disturbance are important to consider because the areas of a watershed contributing to streamflow at any given time will vary in both space and time (i.e., variable source area) (Hewlett and Hibbert 1967).

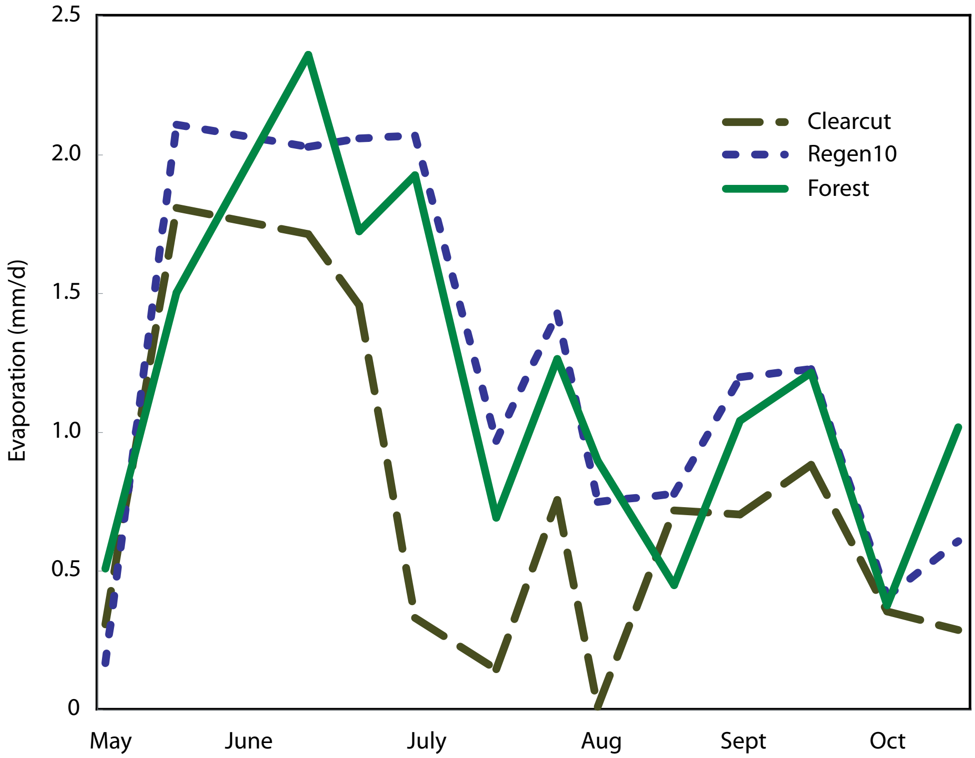

FIGURE 7.2 *Daily evaporation (mm/d) averaged over 10- to 20-day periods for a lodgepole pine forest (solid line), a clearcut (dashed line), and a regenerating 10-year-old lodgepole pine stand (dotted line) at Upper Penticton Creek. Data are based on measurements of water content and precipitation (Figure 7.1), and a water balance model estimate of drainage. Evaporation is the sum of plant transpiration, evaporation of intercepted water, and evaporation from the soil surface (Modified from Spittlehouse 2006).*

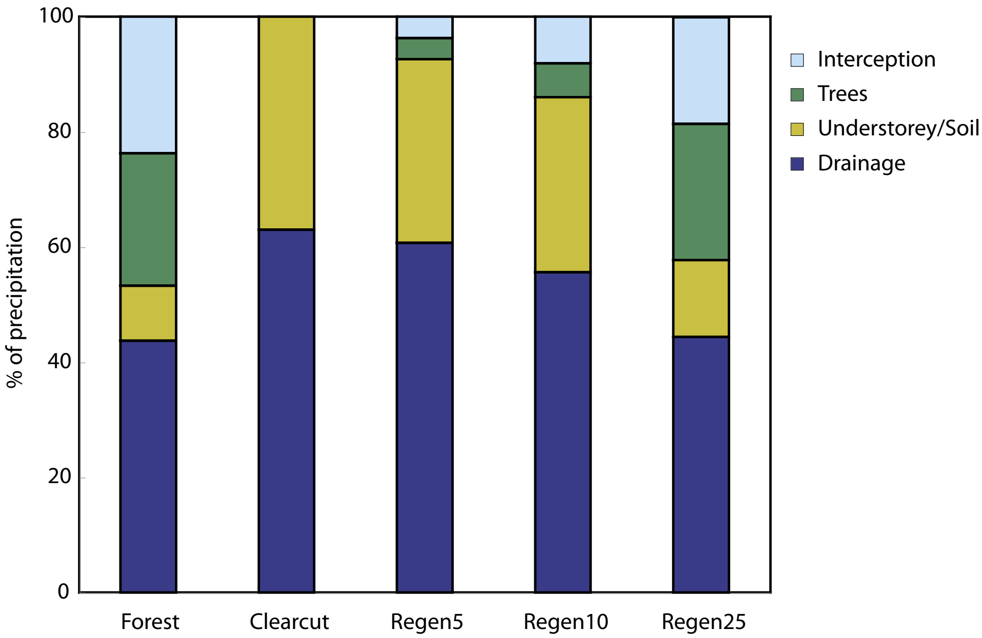

FIGURE 7.3 *Mean annual water balance for an old lodgepole pine forest, clearcut, and regenerating 5-, 10-, and 25-year-old lodgepole pine stands at Upper Penticton Creek from October 2002 to September 2005. Data are from a water balance model calibrated using measured root zone water content, precipitation interception, and tree transpiration data. Shown are interception of precipitation (light blue), tree transpiration (green), evaporation from the soil plus understorey transpiration (yellow), and drainage below the root zone (blue), which are expressed as a percentage of the mean October to September precipitation of 490, 840, and 645 mm for the 3 years (D.L. Spittlehouse, unpublished data).*

Snow ablation

Ablation refers to the loss or disappearance of a snowpack caused by snowmelt (drainage of meltwater at the base of the pack) and evaporation (including sublimation). The separation of ablation into melt and evaporation is necessary to understand snowmelt contributions to spring streamflow. In most situations, melt dominates over evaporation in snowpack ablation as the energy required for evaporation and sublimation is approximately 7.5 and 8.5 times greater than for melt, respectively (see Chapter 6, "Hydrologic Processes and Watershed Response").

In Finland, Kuusisto (1986) found that snow evaporation losses are generally low (less than 0.6 mm/d). At Mayson Lake, B.C., latent heat flux densities measured under forest cover and in a clearcut averaged less than 2 W/m^2 (Adams et al. 1998). These fluxes indicate negligible (less than 0.5 mm/d) evaporation losses from, or condensation on, the snowpack during seasonal snowmelt. Bernier and Swanson (1992) found that daily snow evaporation rates were low in both openings and in a lodgepole pine forest in Alberta, and ranged from an average of 1.1 to 2.5 mm/d, respectively. Bernier (1990) studied snow evaporation rates in the open and in an artificial stand of 2.5 m tall lodgepole pine trees in Alberta. At a stand density of 1650 stem per hectare, the evaporation rate was equal to that in the open. At 2500 stems per hectare, the rate decreased to one-third of that in the open. The consensus from available research is that vapour losses are minor relative to the rates of snowmelt, and that measured ablation provides a good approximation of meltwater draining from a snowpack. The relative importance of sublimation is greater for thinner snowpacks, as it becomes a larger proportion of the water balance.

Forest cover affects snow ablation by changing the surface energy balance. The amount of shortwave radiation reaching the snow surface is reduced, longwave radiation is increased, and the exchange of latent and sensible heat is attenuated through the reduction of wind speed, near-surface temperature, and humidity gradients relative to the open. Removal of the forest canopy exposes the snow surface to greater incident solar radiation and higher wind speeds, which can increase sensible and latent heat inputs (Berris and Harr 1987; Adams et al. 1998). Snow ablation rates are typically 30% to more than 100% higher in the open than in the forest (e.g., Toews and Gluns 1986; Berris and Harr 1987; Spittlehouse and Winkler 2004; Winkler et al. 2005).

Snow ablation rates reported in the literature are usually average values over a 1- to 4-week period. The rates vary from 4 to 25 mm/d in open areas and 3 to 17 mm/d under forest canopies similar to those found in south-central British Columbia (Winkler 2001). Metcalfe and Buttle (1998) found that in boreal stands, snowmelt rates increased exponentially with increasing canopy gap fraction (the fraction of sky visible to a sensor under the canopy). Hardy and Hansen-Bristow (1990) found that average snowmelt rates in a mature stand were 0.6 and 0.8 times those in 4 and 14 m tall juvenile stands, respectively. At Mayson Lake average melt season ablation rates measured over a 3-year period were 4 mm/d in a spruce-fir stand and 10 mm/d in a clearcut (Winkler et al. 2005). At long-term snow research sites on the Thompson-Okanagan Plateau, snow ablation averaged 15% lower and snow persisted up to 8 days longer in mature lodgepole pine stands than in the open. Snow ablation rates in mixed-species stands varied by up to 70% of those in the open (Winkler 2007).

Maximum snowmelt rates indicate how quickly snow cover can generate meltwater under extreme conditions. These rates are more difficult to measure and are less frequently reported than seasonal melt rates, but often exceed seasonal average rates by two to three times. Kattelmann et al. (1998) summarized snowmelt rates reported in the literature and suggested that under typical mountain conditions, maximum rates in the open are generally less than 50 mm/d. The authors concluded that the following conditions are most conducive to rapid snowmelt in the absence of rain: long periods of daylight around the time of the summer solstice; clear days with cloudy nights; warm humid air combined with high wind conditions; low snow albedo; discontinuous and shallow snowpacks so that light can penetrate to the ground surface; and local sources of longwave radiation (e.g., burned stems, logs, or rocks). In a clearcut at Mayson Lake, the maximum snowmelt rate measured as lysimeter outflow was 29 mm/d, which was more than three times the average of 8 mm/d measured as SWE loss over the melt season (Winkler et al. 2005). Haupt (1969) reported a maximum daily lysimeter outflow rate of 52 mm during clear weather at a high-elevation, south-facing site in Idaho. At Murphy Creek near Castlegar, B.C., Nassey reported[1] a maximum lysimeter outflow of 23 mm/d compared to the average SWE loss of 12 mm/d over the melt period.

Changes in snow ablation rates also occur following natural disturbance. After a forest fire in southern Montana reduced lodgepole pine canopy cover by 90%, average snow ablation rates increased by 57%, the same increase as was observed following clearcutting (Skidmore et al. 1994). In lodgepole pine stands near Fraser Lake, Boon (2009) found that average ablation rates were 14–17% higher in dead stands than in live stands, whereas differences between a live forest and open areas varied from 15 to 34%, depending on year. The average snow ablation rate in a dead stand near Vanderhoof was similar to that in a green stand and about one-half the rate of that in the open (Boon 2007). Initial results from other studies in British Columbia, summarized in Table 7.1, show that average ablation rates are 27–48% lower in green-to-red attacked pine stands, except at Border Lake where ablation rates were slightly higher in the forest than in the clearcut, and 14–29% lower in grey stands compared to rates in the open. Snow disappeared 3–10 days later in the green-to-red attacked stands and 0–6 days later in the grey stands than in the open. In a comparison of maximum ablation rates, Teti (unpublished data) found reductions of 17–48% in green-to-red attacked stands and rates 3% higher to 47% lower in grey stands.

Hillslope Runoff Generation

Because undisturbed forest soils in the Pacific Northwest, including British Columbia, generally have sufficiently high infiltration capacities, infiltration-excess overland flow (also known as *Hortonian* overland flow) is generally not an important process for streamflow generation (Cheng 1988; Wondzell and King 2003). Therefore, runoff generation in undisturbed forest catchments tends to be dominated by subsurface flow. In montane catchments in

1 Nassey, J.M. 1994. Measurement and modelling of snowmelt on a clear-cut site in the West Kootenays, British Columbia. Directed Stud. Rep., Simon Fraser Univ., Vancouver, B.C.

coastal British Columbia, which are typically dominated by shallow soils overlying bedrock or relatively impermeable glacial till, downslope flow occurs through the development of a transient saturated zone overlying the till or bedrock, with rapid flow facilitated by root channels and other preferred pathways (Hetherington 1982; Hutchinson and Moore 2000; Anderson 2008).

Forest harvesting can influence subsurface hydrologic response through decreased interception and evaporation of water, alteration of soil physical properties that control infiltration and transmission of water, and the construction of roads that reroute water. Decreased interception loss can lead to an increase in the amount of water that infiltrates the soil, which can result in higher water-table levels during storms (Dhakal and Sidle 2004). Forest canopy removal reduces interception losses. This increases rainfall intensity at the soil surface, which may cause more rapid subsurface flow and larger peak flows (Keim et al. 2006). Logging can also affect subsurface stormflow by compacting surface soils (i.e., reducing macropore space), which slows the transmission of water through the soil (deVries and Chow 1978).

Forest harvesting commonly leads to higher soil moisture content because removal of forest cover leads to a reduction in interception and evaporation of water. This trend has been observed in several different ecosystems (e.g., Megahan 1983; Adams et al. 1991; Keppeler et al. 1994; Troendle and Ruess 1997; Elliott et al. 1998; Hetherington 1998). Increases in subsurface stormflow after harvest are typically related to higher antecedent soil moisture content, resulting in a smaller soil moisture deficit that must be satisfied prior to initiation of lateral flow (Keppeler et al. 1994). This effect is most pronounced in summer and early autumn when soil moisture deficits are high (e.g., Ziemer 1981).

Very few published studies focus on changes in hillslope runoff generation following harvesting. At Carnation Creek on Vancouver Island, Hetherington (1998) noted greater post-harvest hillslope flow inputs to the valley bottom, but did not provide details. In Idaho, Megahan (1983) measured a large, post-disturbance (harvest and wildfire) increase in both subsurface stormflow volume (+ 96%) and flow rate (+ 27%) during the snowmelt period. The increase in volume and rate was attributed to a post-disturbance increase in both SWE and snowmelt rates. Similarly, by using a combination of hydrometric and geochemical methods, Monteith et al. (2006a, 2006b) found that a higher proportion of snowmelt-generated streamflow originated from surface and near-surface soil horizons in harvested than in unharvested watersheds on the boreal shield in Ontario. At the Fraser Experimental Forest in Colorado, Troendle and Ruess (1997) found that annual plot outflows increased from 15% of annual precipitation under mature forest cover to 60% in a clearcut plot. Keppeler and Brown (1998) measured a 400% increase in peak pipeflow after harvesting in the Caspar Creek watershed in northern California.

At many sites, particularly in British Columbia's interior, harvesting is conducted with skidders that can compact soil surfaces and cause overland flow. This can lead to an increase in the flashiness of streamflow response and the magnitude of surface erosion, and a decrease in the chemical interactions of water with the subsurface environment. The significance of this soil compaction and resulting overland flow depends on the degree of compaction and how much of the watershed area is disturbed (Putz et al. 2003), as well as whether the skid trails direct water to the natural drainage network. In south coastal British Columbia, Cheng et al. (1975) found that infiltration rates remained sufficiently high, even following compaction, and that infiltration-excess overland flow was not observed. Tracked machinery, such as hoe-forwarders and feller-bunchers, can also cause soil compaction, particularly if used when soil moisture levels are high (Greacen and Sands 1980). Excavated trails and constructed haul roads typically have compacted surfaces with lower permeability than forest floors, and can generate overland flow even in moderate rainstorms (Luce and Cundy 1994). The significance of this overland flow at the watershed scale depends on the proportion of the watershed that is covered by logging roads and landings, and on the connectivity of these surfaces to streams via ditches, road surfaces, and culverts.

Harvesting in riparian zones can have a significant effect on riparian zone hydrology. Changes in transpiration and water table drawdown caused by harvesting can decrease the fluctuations in discharge through the day and lead to an increase in low flows (Dunford and Fletcher 1947). These effects should lessen over time as riparian forests regrow. Nevertheless, changes in species composition during forest succession can affect low flows. For example, in the Oregon Cascades, a change in riparian vegetation from conifers to deciduous species following clearcut logging resulted in increased transpiration by streamside vegetation and reduced dry weather streamflow (Hicks et al. 1991).

Groundwater

The effects of forest harvesting on groundwater are generally driven by the same processes that result in greater hillslope flow and water yield (Smerdon et al. 2009a, 2009b). Very few published studies have focussed on the direct link between forest management activities (both harvesting and road construction) and groundwater (i.e., subsurface water in the saturated zone). Some studies have reported on the effects of tree removal (i.e., clearcutting, partial retention, or selective cutting) and have variously observed changes in the position of the water table, estimated changes in catchment water yield, or changes in streamflow. Most often, reductions in evaporation and interception lead to an increase in groundwater recharge, which results in elevated water tables. Changes in water-table elevations after harvesting are important because groundwater is the source of most base flow in streams, which has many economic and ecological values (Douglas 2008).

This section draws from two recent publications that summarize the available literature on forest management effects on groundwater hydrology, specifically water-table position, groundwater recharge, and the effects of road construction on groundwater flow (Smerdon et al. 2009a, 2009b). These authors also developed a hydrogeological classification for British Columbia and used it to place the potential effects of forest management on groundwater within a landscape context.

Effect of forest harvesting on water-table position
Forest harvesting generally leads to a rise in the elevation of the water table towards the ground surface. This is primarily attributed to the reduced interception and evapotranspiration that results from the loss of forest cover. The effect of changes in the interception of rain and snow by the canopy has been relatively well documented; however, changes in plant transpiration before and after harvest are complex and depend on vegetation type. The net effect is that wetter soil more readily conducts water to the saturated zone, which in turn may increase the elevation of the water table (Table 7.2).

The degree of water-table rise depends on specific watershed characteristics, including bedrock geology, surficial geology, soil type, and landform topography. At Carnation Creek on west Vancouver Island, late summer water-level rises on the floodplain of 30–50 cm persisted for 10 years after harvesting (Hetherington 1998), and contributed to the process of triggering preferential flow in humid, steep watersheds (e.g., Beckers and Alila 2004). Rex and Dubé (2006) noted elevated water-table levels in lowland areas (e.g., toe slopes, wetlands) after harvesting in the Vanderhoof Forest District (Table 7.2). On the boreal plains, the water table in low-relief watersheds with dry climates increased by 26 cm following harvesting (Evans et al. 2000).

An increase in water-table elevation may have different effects on forest management operations (e.g., trafficability) depending on soil characteristics and slope gradient. Reduced surface-water drainage density and understorey vegetation, and increased area of poorly drained soils, promote wetter ground conditions (Dubé and Rex 2008). In the Vanderhoof Forest District, summer logging on wet ground has been stopped in favour of logging when the ground is frozen and generally more stable (Rex and Dubé 2006). In steep mountainous watersheds, a rise in the water table may lead to a greater risk of slope failure (Sidle and Ochiai 2006). Each of these scenarios imposes constraints on logging operations, which illustrates the potential feedback between forest harvesting and trafficability.

A rise in the water table and wetter ground can also affect the selection of appropriate silvicultural systems (Pothier et al. 2003) and post-harvest species selection. Forest regeneration may be affected by higher water tables because some tree species do not tolerate saturated conditions in the rooting zone, which may decrease productivity or result in regeneration failure (e.g., Landhäusser et al. 2003). Some species, however, promote higher evapotranspiration rates, thereby helping to lower the water table following harvesting. These "nurse crops" could contribute to water-balance recovery in recently harvested and replanted sites (Landhäusser et al. 2003). Restoration of a vegetation cover that has similar evapotranspiration characteristics to the original species is an important step in minimizing the long-term effects of harvesting on groundwater systems and in maintaining forest productivity.

Effect of forest harvesting on groundwater recharge
Water entering the groundwater flow system (i.e., recharge) is difficult to quantify (deVries and Simmers 2002); however, inferences are possible from changes in water-table position, watershed water yield, and

TABLE 7.2 *Effects of forest harvesting on water-table position (from Smerdon et al. 2009b)*

Source	Study site	Location	Annual precipitation (mm)	Forest management practice[a]	Change in water table
Hetherington (1998)	Carnation Creek	Vancouver Island, British Columbia	2100–4800	CC	30–50 cm rise that persisted for 10 years following harvest
Fannin et al. (2000)	Carnation Creek	Vancouver Island, British Columbia	2100–4800	CC	50–150 cm rise (approx.) following individual storm events. Large spatial variability because of soil conditions, but all water-table response was rapid. An upper limit to pressure-head increase was observed, above which preferential flow pathways activated.
Rex and Dubé (2006)	Vanderhoof Forest District	Central British Columbia	496	CC + MPB	10 cm (approx.) higher water table in toe-slope of cut area compared to MPB-kill area; 30 cm (approx.) higher water table in upland of cut area compared to MPB-kill area
Evans et al. (2000)	TROLS	Central Alberta	468	PC	26 cm higher in cut area compared to uncut area
Dubé et al. (1995)	Beaurivage Forest	St. Lawrence lowlands, Quebec	957	CC	7–52 cm rise, depending on soil texture
Pothier et al. (2003)	Villroy	St. Lawrence lowlands, Quebec	510	PC + CC	Up to 22 cm rise in cut areas. Water-table rise increased linearly with percentage of cut area in the first year following harvest. Five years after harvest, water tables remained elevated but were less dependent on the percentage of area cut.
Megahan (1983)	Pine Creek	Central Idaho	890	CC	90 cm rise in water table, decreasing to approximately 40 cm after 2 years
Bliss and Comerford (2002)	-	Gainesville, Florida	1150	CC	21–49 cm rise after 900 days. Larger seasonal fluctuations observed for 4 years following harvest.
Peck and Williamson (1987)	Collie River Basin	Western Australia	820–1120	CC + PC	100–400 cm rise following wet season. Water table increased by 260 cm/yr in clearcut areas and 90 cm/yr in partially cleared areas.

a CC: clearcut; PC: partial cut; MPB: mountain pine beetle

base flow (in some cases). Similar to water-table rises following harvesting, wetter soil conditions may lead to increases in groundwater recharge rates because of the wetter antecedent conditions and the lower available storage in the unsaturated zone. Appreciable amounts of groundwater recharge (compared to runoff) are not expected in watersheds in steep terrain (e.g., Hudson and Anderson 2006). Although no published studies directly measured increased rates of groundwater recharge following harvesting, a few have shown increases in watershed yield, which may be a result of higher groundwater recharge (Table 7.3). For example, in the northeastern United States, a harvested headwater watershed supplied an

TABLE 7.3 *Effects of forest harvesting on groundwater recharge (from Smerdon et al. 2009b)*

Source	Study site	Location	Annual precipitation (mm)	Forest management practice[a]	Inferred change in groundwater recharge
Bates (2000)	Fernow Experimental Forest	West Virginia	1470	PC	Harvested watershed had greater low flows (base flow) to headwater streams caused by higher soil moisture in the years following harvest. Minor amount of (event) stormflow noticed, compared to deeper subsurface flow.
Bent (2001)	Cadwell Creek	Massachusetts	1174	PC	Groundwater recharge increased by 68 mm/yr for six seasons following harvest
Cornish (1993)	Karuah	Australia	1450–1750	PC	Yield increased 150–250 mm/yr following harvesting, depending on percentage of area cut. Increased recharge and overall water yield remained higher for 3 years following harvesting.
Bren (1997)	Cropper Creek	Southeast Australia	660	CC	Increase in amplitude of diurnal streamflow fluctuations attributed to increased subsurface flow, after removal of vegetation
Cook et al. (1989)	Western Murray Basin	Australia	340	CC	Recharge increased by 20 mm/yr very gradually following harvest (approx. 200 years based on simulation modelling)

a CC: clearcut; PC: partial cut

increased base flow for a few years that was apparently not generated from shallow stormflow (Bates 2000). In another case, recharge increased by 68 mm/yr for 6 years following harvesting (Bent 2001). Pike and Scherer (2003) summarized similar results for snowmelt-dominated hydrologic regimes. Results of stand water-balance models for the Upper Penticton Creek Experimental Watershed in the Okanagan Basin showed an increase in water drainage from the soil rooting zone after harvesting and natural disturbance (e.g., mountain pine beetle) compared to undisturbed mature forest stands (Spittlehouse 2007, Figure 7.3). Similarly, simulations of fire effects on pine stands at Mayson Lake indicated that greater root zone drainage is a result of high antecedent soil moisture conditions (Moore et al. 2008).

The implications of changes in groundwater recharge vary according to watershed characteristics (e.g., bedrock geology, surficial geology, soil type, and topography). Further study of the relationship between groundwater recharge and streamflow in forest management areas is needed to quantify the hydrologic mechanisms that control flow regimes. The magnitude of increase in water yield, and potentially the increase in groundwater recharge, may be linearly related to the percentage of area cut (partial harvesting or clearcutting), especially in the first few years following harvesting (e.g., Stednick 1996). Groundwater flow systems adjust to increased water input, and effects may be short- or long-term depending on the scale of the flow system. Recharge that is part of a local-scale system may discharge relatively quickly to nearby headwaters (e.g., Bren 1997), in which case the effects of harvesting could be detectable and (or) ecologically significant. Conversely, the effects of relatively short-term forest disturbances (i.e., decades) may not be detectable in larger-scale flow regimes (e.g., regional-scale flow systems), which tend to respond over long time scales.

Forestry roads and groundwater flow

The effects of logging roads on hillslope hydrology and watershed response are a major focus of concern and debate (Luce and Wemple 2001). Roads that cut into a hillside may intersect shallow groundwater. This creates a seepage face along the road cut (Megahan and Clayton 1983) and more rapidly redirects surface water flow in ditches and culverts into the stream network, which potentially increases peak flows (Jones and Grant 1996; Wemple and Jones 2003). Forestry roads that intercept and re-direct shallow groundwater may also reduce groundwater flow to downslope environments (e.g., springs and seepage areas).

The effect of flow interception by roads depends on how much subsurface flow is intercepted and how much is conveyed directly to the stream network. The proportion of intercepted subsurface flow is a function of many factors, including soil depth, permeability of the bedrock underlying the soil, depth of the road cut/ditch surface, permeability of the roadbed material, and location of the road on the slope. These factors combine with specific watershed characteristics (e.g., bedrock geology, surficial geology, soil type, and topography) that vary across the province. On the south coast of British Columbia, Hutchinson and Moore (2000) found that most of the net rainfall input to a hillslope segment was intercepted at a road cut that was more than 1 m deep with soils averaging 1 m deep and underlain by a compacted basal till. At Carnation Creek on Vancouver Island, road construction at different locations either increased, decreased, or resulted in no change to downslope water-table elevations (Hetherington 1998). The differences in downslope effects were related to road location and construction practices. The effect of road interception on groundwater flow may have significant effects on aquatic ecosystems that rely on specific rates and timing of groundwater discharge or duration of base flow. Subsurface flow intercepted at a road cut during snowmelt in the Idaho Batholith was about 30–40% of the total upslope water input, suggesting that over one-half of the meltwater flowed downslope below the road cut, through the bedrock (Megahan 1972).

Runoff from a road network can flow into stream channels via two pathways: (1) roadside ditches that drain directly to streams, and (2) roadside ditches that drain to culverts that feed water into incised gullies (Wemple et al. 1996). In other cases, ditch flow may be diverted back onto the slope below a road by a cross-drain or culvert, where it will re-infiltrate and flow downslope as subsurface flow. In this case, the road may not increase the rate of transmission of water to the stream channel, but rather redistribute subsurface flow laterally across the slope. For steep valley-side streams without well-defined watershed boundaries, this redistribution could transfer water from one small stream to the next. The redistribution of flow from its initial path into gullies and onto slopes can increase the potential for slope failures (see Chapter 9, "Forest Management Effects on Hillslope Processes").

In more gently sloped terrain, the potential for road cuts to intersect groundwater flow systems is typically lower than in steep terrain, except near groundwater discharge areas (streams, wetlands) where the water table is shallow. Under such conditions, the physical attributes of the road may be more important than those of the surrounding landscape. For example, compacted road surfaces can limit infiltration and pre-existing lateral flow (if present). Whether this effect is significant depends on the area of the watershed that is covered by compacted surfaces (Putz et al. 2003), the location of roads, and the type and quality of road construction.

Forest management effects on regional groundwater resources

Groundwater flow systems at the regional scale tend to buffer short-term variability in climate and land use changes (including forest management activities), but also tend to integrate long-term changes, which makes deleterious impacts more difficult to reverse. As a result, widespread changes in upland recharge areas caused by forest harvesting could go unnoticed for decades in adjacent valley-bottom aquifers. The effects may also be masked or magnified by climate variation and change. Widespread forest clearing in Western Australia throughout the 1950s and 1960s provided an example of the time involved for changes to propagate through groundwater regimes. Low-rainfall areas (850 mm/yr) and high-rainfall areas (1120 mm/yr) were clearcut, and groundwater response was monitored over the succeeding decades (Hookey 1987). Water budget studies and simulation modelling for the flow systems in Western Australia have shown that groundwater equilibrium for the basin re-establishes 25–30 years after cutting (Hookey 1987).

To date, no studies on the potential large-scale effects of forest disturbance on regional groundwater resources have been conducted in British Columbia. Therefore, studies that focus on the dynamics of complete groundwater flow systems and the mechanisms that govern groundwater recharge and discharge in forest management areas are warranted, particularly in the context of the current mountain pine beetle infestation and associated salvage harvesting in central British Columbia. Rex and Dubé (2006) found that low-relief watersheds with fine-textured soils and dead pine stands have wet soils and a raised water table. Cutblocks may have higher water-table elevation than beetle-kill areas, and the difference between beetle-kill and cutblock areas could be more pronounced at upper slope locations (Rex and Dubé 2006). Such findings suggest that the effects of forest disturbance on regional groundwater resources may not manifest for decades. Clearly, further research on the effects of forest disturbance on regional groundwater resources in British Columbia is needed, especially as the use of groundwater resources and urban interface forest management increases in the province.

WATERSHED-SCALE EFFECTS

Although many of the hydrologic effects of forest operations are reasonably well understood at the site or stand scale, it is more difficult to make quantitative predictions at the watershed scale (Committee on Hydrologic Impacts of Forest Management 2008; also see Chapter 16, "Detecting and Predicting Changes in Watersheds"). Three streamflow variables of primary interest are: (1) water yield, (2) peak flows, and (3) low flows (see Chapter 6, "Hydrologic Processes and Watershed Response). Understanding the effects of forest disturbance and regrowth on total annual and seasonal water yield, peak flows, and low flows is important to sustain water supplies, protect aquatic habitat, design infrastructure, and mitigate risks to lives and property.

Detecting the effects of disturbance on streamflow variables at the watershed scale is difficult because of the natural variability in driving factors, such as climate, geology, topography, and forest cover within and between watersheds. The most statistically rigorous method of quantifying the effects of forest operations on streamflow involves paired-watershed experiments, which include pre- and post-harvest data and at least one untreated control (Hewlett 1982). Alternative approaches to paired-watershed experiments include: retrospective studies that use existing operational streamflow data; pre- and post- treatment studies without a control; chronosequence analysis in which trends in streamflow in a suite of watersheds with different management histories are correlated with changes in land use and (or) climatic variables; and computer simulation models. Most recently, Alila et al. (2009) suggested pairing peak-flow events based on frequency rather than chronology to evaluate changes after logging. Table 7.4 summarizes information about watershed experiments within British Columbia that quantify disturbance effects on streamflow. Figure 7.4 shows the locations of these watersheds. Table 7.5 summarizes similar information for watershed experiments outside British Columbia.

TABLE 7.4 *Watershed experiments that quantify forest cover effects on streamflow in (a) coastal, (b) northern interior, and (c) southern interior regions of British Columbia*

Forest region /watershed (Reference)	Status[a]	Leading tree species[b]	Sub-basins	Area (km^2)	Elevation range (m)	Hydrologic regime[c]	Duration[d]	Average annual precipitation (mm/yr)	Average annual water yield (mm/yr)	Design[e]	Treatment[df] (years): total of area cut (%)	Changes[g] measured (%) or under investigation (√): Annual yield	Changes[g] measured (%) or under investigation (√): Peak flow
a) Watershed experiments in coastal British Columbia													
Carnation Creek													
(Hetherington 1982, 1998; Hartmann and Scrivener 1990)	O	Hw, Cw, Ba, Fd	B	10.0	8–884	R	1970–p	2100-4800	2544	PP	CC (1976–1981): 41 CC (1987–1995): 25; (66% total)		
	O		C	1.5	46–700	R	1970–p				Control: 0%	√	√
	O		E	2.6	150–884	R	1970–p				Control:	√	√
											CC (1987–1993): 38	√	√
	O		H	0.1	152–305	R	1970–p				CC (1977–1978): 90	14	20
	C		J	0.2	30–300	R	1970–1996				CC (1977–1978): 94	√	√
Flume Creek													
(Roberts)	C	Hw, Fd	F4	< 1	505–850	ROS	1995–2005	1640	1416	BACI	VR (1998): 44		80
(Hudson 2001)			F5	< 1	505–850				1405		SC (1998): 32		85
			F6	< 1	395–560				1486		Control: 0		
Jamieson Creek													
(Golding 1987)	C	Fd, Hw, Cw	Jamieson	3	305–1310	ROS	1972–1984	3525	2995	BACI	CC (1978, 1982–1984): 19		13.5
			Elbow	1	275–1065				1525		Control: 0		
Russell Creek													
(Anderson 2008; Floyd and Weiler 2008)	O	Hw, Fd	Stephanie	32	275–1700	ROS	1991–p	2395	1993	M, ISS	CC (1978–p): 35		√
b) Watershed experiments in northern interior British Columbia													
Baker Creek													
(Forest Practices Board 2007)	O	Pl		1570	470–1530	SM	1963–2005, 2006–p	400	80	M, ISS	CC (2005*): 34; MPB (2005*): 53; CC (< 1970–p)	31*	61*
Bowron River													
(Lin and Wei 2008)	C	Se, Bl		3590	602–2447	SM	1954–1996	1149 (upper)	594	PP	CC (1966–1996): 25	NS	NS

c) Watershed experiments in southern interior British Columbia													
Camp Creek (Cheng 1989; Moore and Scott 2005)	C	Pl	Camp	34	1070–1920	SM	1970–2000	600	140	CI	CC (1976–1978): 30	NS	30 (April)
			Greata	41	880–1620						Control: 0		
Cotton Creek (Jost et al. 2007)	O	Pl, Se, Bl		17	1100–2100	SM	2004–p	650	284	ISS	CC (1975–p): 34	√	√
Fishtrap Creek (Moore et al. 2008)	O	Pl, Se, Bl, Fd		135	370–1620	SM	2004–p	471	180	ISS	B (2003): 75; CC(?–p)	√	√
Palmer Creek (Cheng and Bondar 1984)	C	Fd, Pl, Se, Bl	Palmer	18	960–1950	SM	1967–1977	750	350	BACI	B(1973): 50	24(Apr–Aug), 37(Aug–Nov)	50
			Upper Salmon	143							Control: 0		
Redfish Creek (Whitaker et al. 2002)	O	Fd, Lw, Pl, Se, Bl		26	700–2370	SM	1992–p	1582	1018	M, ISS	CC (*): 22max CC (1979p)		22*
Upper Penticton Creek (Winkler et al. 2008b)	O	Pl	240	5	1620–1920	SM	1984–p	700	325	BACI	Control: 0		
		Pl	241	5	1600–2015				326		CC (1995): 10 CC (1998): 18 CC (2002): 28 CC (2006): 47	NS, NS, √ √	NS, NS, √ √
		Se, Bl, Pl	Upper Dennis	4	1780–2065				389		CC (1995): 10 CC (1998): 21 CC (2000): 52	NS, NS, √	NS, NS, √

a O: ongoing; C: complete
b Ba: amabilis fir; Bl: subalpine fir; Cw: western redcedar; Fd: Douglas-fir; Hw: western hemlock; Lw: western larch; Pl: lodgepole pine; Py: ponderosa pine; Se: Engelmann spruce
c R: rain; ROS: rain on snow; SM: snowmelt
d p: present
e BACI: before after control intervention (paired watersheds all with pre- and post-disturbance data); CI: control intervention (paired watersheds with no pre-disturbance data); PP: pre- and post- (single watershed time series); ISS: intensive study site for detailed modelling; M: modelling
f CC: clearcut; B: burned; VR: variable retention (% of canopy removed); SC: shelterwood (% of canopy removed); PC: partial cut (% basal area removed); MPB: attacked by mountain pine beetle (% of watershed area attacked)
g * modelled; NS: non-significant or non-detectable

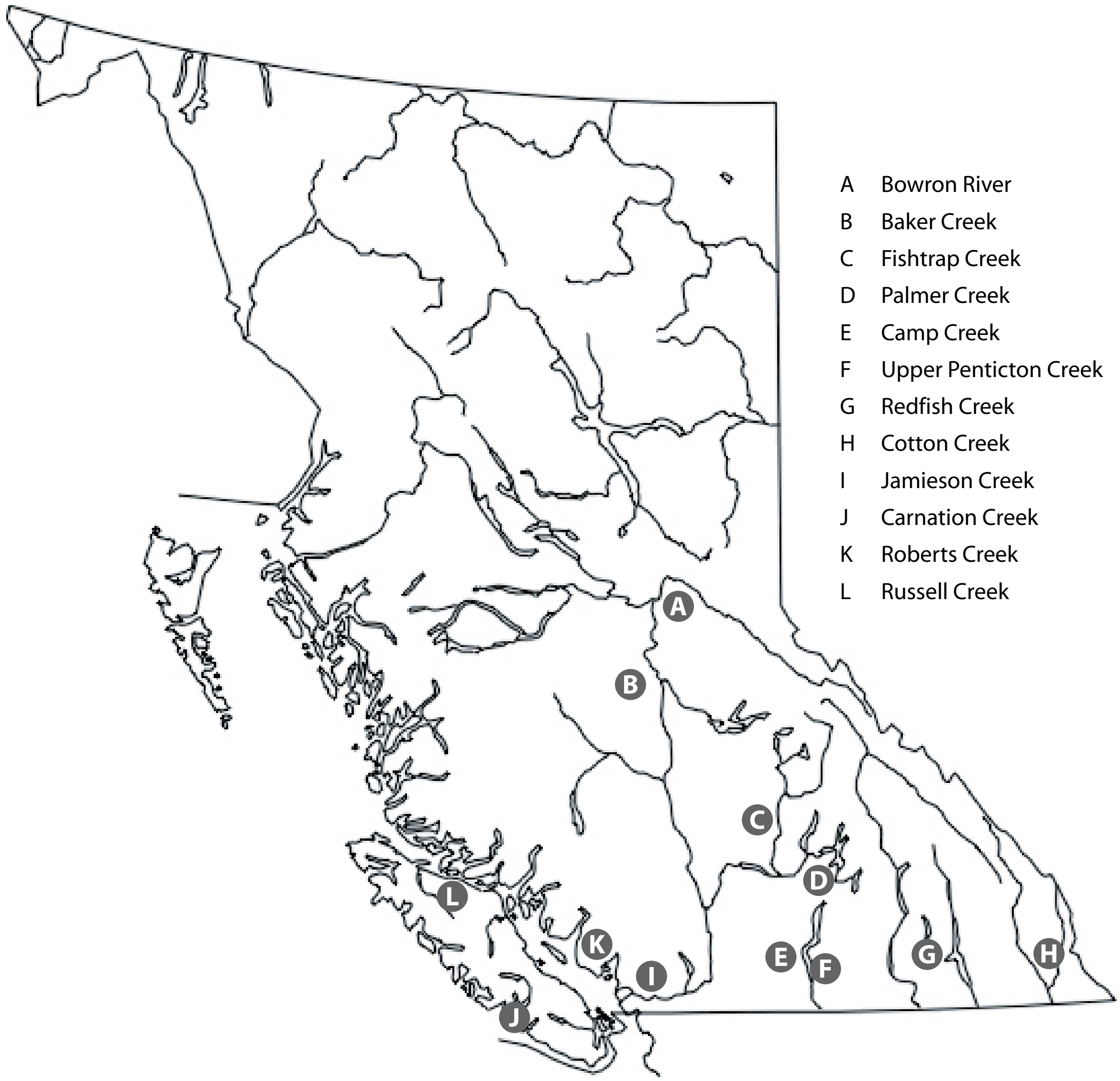

FIGURE 7.4 *Locations of watershed experiments in British Columbia.*

TABLE 7.5 *Watershed experiments that quantify forest cover effects on streamflow in hydrologic regimes similar to those in British Columbia*

Study site	Hydrologic regime[a]	References
Cabin Creek (Alberta)	SM	Swanson et al. 1986
Alsea Watershed Study (Oregon Coast Range)	R	Harr et al. 1975; Harris 1977; Harr 1983; Stednick 2008
Hinkle Creek (Oregon Coast Range)	R, ROS	n/a
H.J. Andrews (Oregon Cascades)	R, ROS	Rothacher 1973; Harr 1983; Harr 1986; Hicks et al. 1991; Jones and Grant 1996; Thomas and Megahan 1998; Beschta et al. 2000; Jones 2000
Fox Creek (Oregon Cascades)	R	Harr 1980, 1983
Coyote Creek (Oregon Cascades)	R	Harr et al. 1979; Harr 1983
Mica Creek Experimental Watershed (Idaho)	SM	Hubbart et al. 2007
Horse Creek (Idaho)	SM	King and Tennyson 1984
Fraser Experimental Forest (Colorado)	SM	Alexander et al. 1985; Troendle and King 1985
Deadhorse Creek (Colorado)	SM	Troendle and King 1987
Wagon Wheel Gap (Colorado)	SM	Bates and Henry 1928; Van Haveren 1988

a R: rain; ROS: rain on snow ; SM: snowmelt

Annual Water Yield

Studies in temperate forest environments around the world have demonstrated that annual water yield generally increases following removal of forest cover (Bosch and Hewlett 1982; Trimble et al. 1987; Stednick 1996). Moore and Wondzell (2005) reviewed results of paired-watershed studies in the Pacific Northwest and found that, in rain-dominated watersheds, water yield increased by up to 6 mm for each percentage of the watershed area clearcut or patch cut, whereas selective cutting increased yields by up to 3 mm for each percentage of basal area removed. Most of the increased yield occurred in the wet autumn–winter period (Harr 1983; Keppeler and Ziemer 1990). Water yield decreased slightly following patch cutting in two watersheds in the northern Oregon Cascades (Fox Creek 1 and 3), likely because of a decrease in fog drip (Harr 1982).

In snow-dominated watersheds, post-logging water yields increased from 0.25 mm to more than 3 mm per percentage of watershed area harvested (Van Haveren 1988; King 1989; Stednick 1996). At the Fraser Experimental Forest, Colorado, water yield from Fool Creek was 40% higher than the long-term expected flow and 45% higher for the first 5 years post-logging, even though the SWE averaged over the watershed increased by only 9% (Troendle and King 1985). This disproportionate increase in water yield relative to snowpack likely resulted from the combined effect of increased snow accumulation and decreased evaporation (and thus reduced soil moisture deficits) within the harvested areas. At the Horse Creek watersheds in Idaho, annual yields increased by about 3.6 mm per percentage of watershed cleared (King 1989). Small increases in yield were reported at Cabin Creek, Alberta (Swanson et al. 1986) and Deadhorse Creek, Colorado (Troendle and King 1987), but these increases were not statistically significant.

Few studies have been conducted on the effects of roads on water yield because harvesting usually occurs either at the same time as, or immediately following, road building. At the Horse Creek watersheds in Idaho, water yield changes for road-only treatments were not statistically significant (King and Tennyson 1984); however, the short study period (i.e., small sample size) likely limited the ability of the analysis to detect a change.

Peak Flow

The effect of forest disturbance on peak flows has been a source of controversy for decades (Food and Agriculture Organization 2005; Committee on Hydrologic Impacts of Forest Management 2008; Grant et al. 2008). Data from numerous studies (notably those from the H.J. Andrews Experimental Forest in the Oregon Cascades) have been analyzed repeatedly using different data-processing approaches and statistical analyses to assess the magnitude of peak flow increases following forest harvesting (Jones and Grant 1996; Thomas and Megahan 1998; Beschta et al. 2000; Jones 2000; Grant et al. 2008; Alila et al. 2009). The term "peak flow" has been used in various ways to refer to the single largest flow event recorded in a year, the largest flow over a 7-day period, and the total volume of water delivered over the high-flow season and other time periods depending on the flows of interest. Peak flow events are described in terms of magnitude, frequency of occurrence, and variability. Often, the data will limit the type of analysis that can be undertaken. The most common limitation is a short data record, which may not include extreme events. The type of analysis can further limit interpretation of the data. For example, the average of a series of high flow events does not provide insights about flow extremes or the frequency at which events of various magnitudes occur.

The effects of disturbance on peak flows can differ between rainfall, rain-on-snow, and spring snowmelt events. Reported changes in mean peak flows following logging in rain-dominated watersheds range from increases of 20–30% (Harr 1983) to a decrease of 22% (Cheng et al. 1975). Cheng et al. (1975) also reported a several-hour delay in peak flow following harvesting in south coastal British Columbia. The authors suggested that extensive skid road development and soil disturbance (affecting 50% of the watershed area) had reduced the efficiency of macropore flow, which resulted in slower delivery of water to the stream channel (Cheng et al. 1975).

Lewis et al. (2001) analyzed peak flow responses to forest harvesting in 10 watersheds in the North Fork of Caspar Creek, northern California. In this study, roads occupied less than 7% of the area in each watershed. In most watersheds, maximum peak flow increased by less than 100%, but in some, maximum flows increased by up to 300%. The size of the treatment effect increased with the proportion of watershed harvested and decreased with an index of antecedent wetness; it therefore tended to be greatest in early autumn. The mean percentage increase in peak flows decreased with storm size. For a storm event with a 2-year return period, the average increase in peak flows was 27% in the watersheds that were 100% clearcut.

Peak flow responses to forest harvesting during rain-on-snow events can be highly variable, depending on the magnitude of rainfall, wind speed, temperature, water equivalent of the remaining snowpack, and the extent of snow cover (Kattelmann 1987; McCabe et al. 2007). In some studies, peak flows during rain-on-snow events increased following harvesting because of increased melt rates and reduced rainfall interception (Harr 1986). At temperatures close to zero, Beaudry and Golding (1983) found that snow intercepted by forest canopies melted, whereas snow on the ground remained frozen and resulted in increased runoff from the forest relative to the open. Hudson (2001) found highly variable peak flow responses in two watersheds in south coastal British Columbia that could be related to the role of transient snow cover and the percentage of canopy removal. Jones (2000) re-analyzed peak flow data for the H.J. Andrews Experimental Forest (Oregon Cascades) using analysis of variance, and found that winter rain-on-snow peak flows increased by 25–31% in four of five watersheds.

Increased peak flows have been reported in snow-dominated watersheds where more than 20% of the watershed was harvested; however, there was no direct correlation between the extent of harvesting and peak flow change. For example, in Colorado, mean peak flows increased by 20% to almost 90% in research watersheds where 20–40% of the watershed had been harvested (Troendle and King 1985, 1987; King 1989). Van Haveren (1988) found that 100% clearcutting produced a 50% increase in mean peak flow. Alila et al. (2009) re-analyzed the 48-year streamflow record for Fool Creek, Colorado, by pairing peak flow events according to estimated frequency rather than chronology. This re-analysis allowed an assessment of logging effects on the magnitude of peak flow associated with a given return period. They found that peak flows increased after logging across the full range of event frequencies and that the frequency of peak flow events larger than the mean also increased. Because of the non-linear relationship between peak flow magnitude and frequency, a small change in peak flow translated into a large change in the frequency of the event (i.e., a given peak flow was equalled or exceeded more frequently following forest harvesting). Results of forest harvest modelling for the 25-km^2 Redfish watershed near Nelson suggest that removing 100% of the forest cover in 60% of the watershed (without the simulation of roads) would cause a 7-day annual maximum discharge event (i.e., an event that would normally be expected once every 30 years) to occur every 8 years (Schnorbus and Alila 2004). These modelling results also suggest that an event previously expected to occur once every 8 years could be expected every 3 years (Schnorbus and Alila 2004). The magnitude of the snowmelt-generated peak in this study increased by 15%. This is a small increase compared to other studies, and was attributed to desynchronization of the snowmelt in the upper-elevation alpine zone from that under forest cover.

Peak flows can also be affected by forest cover losses caused by fire, insects, and disease. The effects of fire on peak flow vary with burn severity (spatial extent and intensity). High-intensity fires can alter infiltration and overland flow processes, which can result in increases in peak flow that may be as large as one or two orders of magnitude during intense rainstorms (Scott 1993; Neary et al. 2005). A summary of the effects of fire on peak flow indicated that increases in peak flow ranged from negligible in a snow-dominated fir forest, to 1.4 times in a coastal Douglas-fir forest, to 20–2000 times in ponderosa pine forests (Committee on Hydrologic Impacts of Forest Management 2008). Moore et al. (2008) found that after a fire burned throughout the Fishtrap Creek watershed near Kamloops, flows early in the freshet season increased and high flows were sustained longer than before the fire. In an Engelmann spruce beetle epidemic in Colorado, defoliation of up to 80% of the trees in 30% of the area in a north-facing and a west-facing watershed resulted in 4% and 27% increases in maximum annual instantaneous flow, respectively (Bethlahmy 1975). These increases were expected to persist for 25 or more years after the infestation.

In general, salvage harvesting is expected to have a greater hydrologic impact than mortality related to the mountain pine beetle (Redding et al. 2008). Modelling of four scenarios involving beetle-related tree mortality and subsequent salvage harvesting at Baker Creek in British Columbia's central interior suggested that with extreme disturbance, peak flows would occur more than 2 weeks earlier than normal, and average peak flows would increase by 60–90% (Forest Practices Board 2007). These results indicate the potential for a major shift in flood frequency to occur in the watershed. With extensive salvage harvesting (80% of the watershed area), the pre-beetle disturbance flood with a return period of 20 years may increase to a return period of 3 years. This shift would have major implications on the design of infrastructure on the floodplain. It also highlights the need to plan the extent of clearcut salvage harvesting in infested watersheds, and the need to designate reserve areas and carefully design stream crossings (Forest Practice Board 2007).

An important issue in understanding the effects of forestry operations and natural disturbance on snowmelt-driven peak flows is synchronization of snowmelt. In some situations, partial logging of a watershed could reduce peak flows through desynchronization of snowmelt in openings and under forest cover. For example, clearcutting 50% of a low-relief peatland watershed in north-central Minnesota resulted in two relatively low-magnitude peak flows instead of a single, higher peak flow (Verry et al. 1983). The first peak resulted from snowmelt in the cut area; the second occurred several days later in response to melting of the forest snowpack. After the remaining upland area was clearcut, snowmelt was uniform over most of the watershed, and peak flows increased compared to flows that would have occurred in the absence of clearcutting. At Fishtrap Creek, harvesting also appears to have desynchronized melt runoff over the watershed, resulting in lower peak flow magnitude but longer duration of high flow periods (Moore et al. 2008); however, this inference is based on a relatively short post-disturbance record and requires further study. Logging of north-facing slopes or high-elevation areas could advance the timing of melt from those areas and synchronize it with melt on forested south-facing or lower-elevation slopes (Toews and Gluns 1986).

Logging roads did not appear to have a measurable effect on peak flows in most studies that included a road-only treatment (Harr et al. 1975; Ziemer 1981; King and Tennyson 1984; Jones 2000). In these studies, however, the road-only condition was typically short (1–2 years), so the analyses would have low power (i.e., low ability to detect an effect). One exception was the Deer Creek 3 watershed (Alsea Watershed Study, Oregon), where peak flow increases averaged 18% following road construction that affected 12% of the area (Harr et al. 1975).

A particular point of controversy is the hydrologic and geomorphic significance of peak flow changes. Some studies have concluded that changes in peak flows are largest for small events that have "no hydraulic consequence" (e.g., less than a 2-year return period) (Ziemer 1981). Several studies have shown that the mean effect of harvesting decreases with increasing event magnitude (as indexed by the return period for the control watershed peak flow response) (Beschta et al. 2000; Moore and Wondzell 2005; Grant et al. 2008). However, this type of analysis does not provide a basis for interpreting the effect of forest harvesting on flood frequencies, or for assessing the change in peak flows of a given return period or the change in recurrence interval for a specified magnitude of peak flow. For example, the change between return-period events is not linear, so a 5% change in a 25-year flood can result in a peak flow that approximates a 100-year event.

One difficulty in assessing the effect of forest harvesting on larger peak flow events is that few large storm events have been sampled in experimental studies. This is due, in part, to the relatively short periods of pre- and post-treatment monitoring and the rarity of high-magnitude events. The ability to draw generalizations about the effect of forest harvesting on peak flows is further confounded by the wide range of observed response, which is linked to the nature of the events generating peak flows (e.g., rain vs. rain-on-snow vs. spring melt), watershed characteristics, details of the road and harvesting systems, and the time since harvesting. Separating the effects of road building from forest harvesting is particularly difficult because, in most studies, road building and harvesting occur either simultaneously or in close succession.

Even if forest harvesting does not have an effect on major channel-disturbing events, changes in small to medium events can potentially affect channel morphology by changing the frequency of events that can move substantial amounts of sediment. A more detailed discussion of channel morphology is provided in Chapter 10 ("Channel Geomorphology: Fluvial Forms, Processes, and Forest Management Effects").

Development of guidelines for managing the risks associated with forestry-related peak flow increases is hampered by uncertainties about how logging changes peak flows and is compounded by uncertainties regarding a channel's sensitivity to those changes. Grant et al. (2008) generated some tentative guidelines for managing these risks in coastal catchments in Oregon. For snowmelt-dominated catchments, the current knowledge base suggests that risks are low when up to about 20% of the catchment is clearcut. Risks increase, but become highly uncertain, as the area clearcut approaches and exceeds 30%. Further research that combines field studies and modelling with long-term monitoring is needed to quantify the hazards associated with forestry-related peak flow increases in British Columbia watersheds of varying sizes, biophysical characteristics, and different hydrologic regimes.

Low Flow

Low flows are minimum flows that are sustained by groundwater during the dry season or prolonged periods without rain. They may occur in summer or during the time of year when precipitation falls as snow. Reductions in low flows can have a significant effect on aquatic ecology and on water supplies for irrigation, domestic use, and industry. Various metrics have been used to describe low flows in experimental studies. These include the number of low flow days each year (the number of days that daily discharge was less than some arbitrary threshold, which varies among studies), water yield during August (typically a period of low flows), and the total flow for months when the discharge from the control watershed was below an arbitrary threshold. It is difficult, therefore, to make quantitative comparisons across studies without re-analyzing the original data.

In a review of 28 studies on low flows conducted in the United States, Austin (1999) found that 16 cited a statistically significant increase in streamflow, 10 showed no change, and two studies noted less streamflow following logging. Pike and Scherer (2003) reviewed eight forest harvesting studies in snow-dominated environments and found that four reported an increase in low flow volumes, whereas four showed no significant change.

In coastal rain-dominated watersheds in northern California, the Oregon Cascades and Coast Range, and Vancouver Island, more summer streamflow was noted in the first few years after harvesting (Harris 1977; Harr et al. 1982; Hetherington 1982; Keppeler and Ziemer 1990; Hicks et al. 1991). An exception occurred in the Bull Run watershed in the northern Oregon Cascades, where the number of days with low flows increased significantly following patch cutting of 25% in Fox Creek 1 and 3 watersheds.

The decrease in streamflow during the low flow period was attributed to a reduction in fog drip (Harr 1982).

Greater streamflow during the low flow period has been shown to diminish over time with the establishment of vegetation after logging (i.e., because of increased evaporative losses), and in some cases, can be reduced below pre-harvest levels. For example, in the H.J. Andrews Watershed 1 (HJA-1) in Oregon, August flows were higher than expected in the first 8 years after logging, but then dropped below expected flows during the next 18 years (Hicks et al. 1991). This response was attributed to the post-harvest colonization of the watershed's relatively wide valley floor by hardwoods, including red alder, which has lower stomatal resistance and thus a higher transpiration rate per unit of leaf area (which was also higher) than conifers (Moore et al. 2004). The high transpiration rate from the riparian zone reduced streamflow. At the South Fork of Caspar Creek in northern California, summer flow volume increased and the number of low flow days decreased after harvesting, but these effects persisted for only about 5–10 years (Keppeler and Ziemer 1990). In fact, there was some indication that summer flows 10 years after harvest decreased to levels less than expected for pre-harvest conditions. Similar, though not statistically significant, trends were recorded at Needle Branch in the Alsea watershed study (Harris 1977).

Low flow responses to forest harvesting in snow-dominated watersheds are not well documented, as the lowest annual flows often occur in mid- to late winter when it is difficult to measure flows accurately because of ice formation in channels and on weirs. At Wagon Wheel Gap, Colorado, 100% clearcutting increased base flows by 17%, but had no statistically significant effect on 30-day low flows (Van Haveren 1988). At Cabin Creek, Alberta, clearcutting 20% of the watershed area increased flows in August, September, and October by 10–15%, but these changes were not statistically significant (Swanson et al. 1986).

Roads may influence low flows in small, upland watersheds by diverting subsurface flow laterally across hillslopes, particularly where streams are weakly incised. The net effect could be an increase in flows in some streams at the expense of others; however, this effect is not well documented by field measurements.

HYDROLOGIC RECOVERY

Hydrologic recovery, or the return to the pre-disturbance hydrologic regime, involves complex processes that are difficult to measure or estimate at stand and watershed scales. Recovery is driven primarily by the return of the canopy (i.e., leaf surface area) to pre-disturbance levels. As the forest canopy re-establishes, increases in water yield and peak flows following harvesting decrease with time. The timing and rate of recovery depends on site- and season-specific hydrologic processes, the species that occupy the site, stocking density, site productivity, the rate of forest regrowth, and the cumulative effects of varying levels of recovery across the watershed. Where streamflow is dominated by radiation snowmelt that produces a single annual peak flow event, the reduction in snow accumulation and melt caused by forest regrowth can be the largest single influence on hydrologic recovery. In coastal watersheds, where multiple streamflow peaks occur as a result of rain or a combination of rain, rain-on-snow, and snowmelt events, rainfall interception and snow interception influence hydrologic recovery. In Canada, most hydrologic recovery research has focussed on stand-scale processes, primarily interception, which represents the recovery of the canopy or leaf area. At the watershed scale, cumulative recovery at the watershed outlet or other point of interest along the stream channel (such as a community water system intake) depends on the integrated effect of changes in all components of the water balance, both surface and subsurface, over time and space. Watershed-scale recovery has been studied at several long-term research sites in the United States. In British Columbia, information at the watershed scale will become available as harvested areas in long-term watershed experiments regrow and as areas are monitored for mountain pine beetle recovery. Most long-term field experiments are limited in area, so the results are extrapolated to larger watersheds through hydrologic modelling.

At the stand scale, changes in snow accumulation and melt caused by forest regrowth were investigated in coastal British Columbia by Hudson (2000) and in the southern interior of the province by Winkler

(2001) and Winkler et al. (2005). Hudson (2000) used tree height as a predictor variable and space-for-time substitution to develop snow recovery curves for the coast, whereas Winkler (2001) and Winkler et al. (2005) compared snow accumulation and melt in clearcuts and in young and mature stands to infer recovery rates in the southern interior. Hudson (2000) found that maximum SWE recovery was 50% in 4 m tall mixed subalpine fir, western hemlock, and cedar stands with 20% crown closure, and 75% in 8 m tall stands with 50% crown closure. The author suggested that complete recovery would occur when trees were 20 m tall and crown closure exceeded 95%. Hudson and Horel (2007) subsequently incorporated changes in rainfall interception in young stands into their estimates of recovery, which also considered differences in elevation, precipitation regimes, and storm types. The authors suggested that for a stand at 550 m elevation, recovery during a 35-mm rain-on-snow event was approximately 65% in 10 m tall stands and 90% in 20 m tall stands. Using a mature mixed pine, spruce, and subalpine fir stand to represent the "undisturbed" hydrologic condition, Winkler (2001) and Winkler et al. (2005) found that maximum SWE in a 4.5 m tall lodgepole pine stand with 28% crown closure was 43% recovered and that average ablation rates over the melt season were 29% recovered. Snow accumulation recovery (SAR) was calculated using measured SWE as follows, and ablation recovery was measured by replacing SWE with ablation rate.

$$\text{SAR}\ (\%) = \left\{1 - \left[\frac{\text{SWE}_{\text{open}} - \text{SWE}_{\text{regen}}}{\text{SWE}_{\text{open}} - \text{SWE}_{\text{mature}}}\right]\right\} \times 100 \qquad (1)$$

The recovery of snow ablation rates is generally slower than that of snow accumulation rates. In the boreal forest of northern Ontario, maximum SWE recovered to 80% of that in the mature forest within approximately 15 years after clearcutting, when the regenerated stand was about 7 m tall and had a canopy density of 40% (Buttle et al. 2005). Average ablation rate recovery was not expected to reach 50% until trees were 16 m tall or the same height as the unlogged stands. Melt rates in very young regenerating stands exceeded those in clearcuts, possibly as a result of increased longwave radiation and sensible and latent heat fluxes to the snow surface as young trees and other vegetation began to extend above the snowpack (Buttle et al. 2005). Talbot and Plamondon (2002) found that ablation recovery in a clearcut balsam fir stand in southern Quebec was 50% at 15 years after clearcutting, when young trees were 4 m tall, stand crown closure was 50%, light interception was 55%, and basal area was 16 m^2/ha. In Montana, a lodgepole pine stand of 10–14 m tall trees with a basal area of 17 m^2/ha achieved a 41% and 48% recovery of maximum SWE and ablation, respectively, compared to a mature stand of predominantly subalpine fir (Hardy and Hansen-Bristow 1990).

In general, results of snow surveys in pine stands in the southern interior of British Columbia suggest that a 6% reduction in maximum SWE can be expected for approximately every 10% increase in crown closure (Winkler and Roach 2005). Moore and McCaughey (1997) found a 6.4% decrease in maximum SWE per 10% increase in canopy density. The authors also noted that the relationship between canopy density and SWE was stronger for spruce-fir than pine stands, which suggests that other forest structure variables (e.g., crown shape) also influence SWE.

Regenerating stands also slowly increase rainfall interception losses at a site. At Upper Penticton Creek, rainfall interception losses in a 25-year-old, well-stocked, 4 m tall lodgepole pine stand were similar to those in the adjacent mature forest. In coastal ecosystems, however, where mature canopy biomass is high, interception loss in 20- to 25-year-old juvenile stands was about one-half that of a mature stand (Spittlehouse 1998a, 1998b; see also Table 6.1 in Chapter 6, "Hydrologic Processes and Watershed Response").

Within 5 years of disturbance, the root systems of young trees, grasses, and shrubs can effectively withdraw moisture from most of the soil profile. Consequently, during the summer, soil water content and evaporation on these sites is similar to that of a mature forest (Adams et al. 1991; Elliott et al. 1998; Spittlehouse 2002). Similar results were found in boreal forest sites in Saskatchewan (Amiro et al. 2006) and black spruce stands in Alaska (Chambers and Chapin 2003). Further regrowth increases interception losses and increases transpiration to levels where water losses can exceed those of some types of mature forests (Vertessy et al. 1993). In the Oregon Cascades, summer soil moisture content in a regenerated clearcut was 20 mm less than in a forested site 5 years after harvest (Adams et al. 1991). The declining surplus and shift to a slight deficit was attributed to the rapid increase in plant growth, primarily by fireweed, vine maple, and snowbrush.

Figure 7.3 shows model estimates of the mean annual water balance over 3 years in a mature lodgepole pine stand, a recent clearcut, and juvenile lodgepole pine stands after 5, 10, and 25 years of regeneration within the Upper Penticton Creek Experimental Watershed in south-central British Columbia. For this forest type and region, the water balances for a well-stocked 25-year-old stand and the mature uncut forest are similar, whereas those for 5- and 10-year-old stands resemble a clearcut. Figure 7.3 also illustrates an interesting pattern in the water balance throughout the various stages of recovery. The sum of evaporation from the soil and transpiration from understorey plants and overstorey trees remains relatively constant under all forest scenarios, and the decrease in drainage from clearcut to mature stand conditions appears to be a consequence of increased interception loss with increasing stand age (i.e., increasing canopy cover). As noted earlier in this chapter (and in Chapter 6, Table 6.1), the recovery of rainfall interception in coastal forests is slower than in interior forests because of the additional time needed to develop the high canopy biomass characteristic of coastal old-growth forests. Jassal et al. (2009) found that annual evaporation from a 15-year-old and a 60-year-old stand of coastal Douglas-fir stand was similar, which suggests that the higher transpiration rates of younger trees and brush somewhat compensated for lower rainfall interception.

Estimates of post-treatment streamflow recovery rates vary among studies and depend partly on the method of analysis. Thomas and Megahan (1998) found that in the western Cascades of Oregon, treatment effects on peak flows decreased through time, persisting for more than 20 years on the clearcut HJA-1 watershed, but for only 10 years on the patch cut and roaded HJA-3 watershed. Jones and Grant (1996) suggested that the slower recovery of HJA-1 compared to HJA-3 was likely caused by the slower regeneration of conifers. Jones (2000) reported recovery times of at least 10 years in all cases where a significant treatment effect was evident and at least 30 years in two cases (HJA-1 and HJA-3). Hydrologic recovery for annual water yield in Needle Branch, a clearcut watershed in the Alsea watershed, took 31 years to return to pre-treatment water yields (Stednick 2008). At the Coyote Creek watersheds in the southern Oregon Cascades, effects of clearcutting appeared to decline over the 5-year, post-harvest period (Harr et al. 1979). An exponential model fitted to the data suggested that the effect would decrease by about 60% over the first 10 years and by 95% after 30 years. For practical applications, recovery curves can be developed that relate the percent recovery toward pre-logging conditions to some index of stand development, typically canopy height or basal area. These are surrogates for leaf area, which is more difficult to measure and not routinely available.

In British Columbia, estimates of recovery are used to calculate equivalent clearcut area (ECA), which is a commonly used index of the extent of forest disturbance and regrowth in a watershed. Equivalent clearcut area was originally used in provincial watershed assessment procedures as an indicator of potential hydrologic change due to forest harvesting (B.C. Ministry of Forests 1999). In these procedures, data from government and industrial forest inventory databases (e.g., the date of harvest, extent, and distribution of logging and roads in a watershed, and forest regrowth) were used to infer the potential for past and new developments to affect peak flows, channels, aquatic habitat, water supplies, and communities. It was assumed that the greater the proportion of the total watershed area disturbed, the greater the potential for hydrologic change. Additional weight was given to roads and logging near stream channels as well as harvesting in zones that contributed snowmelt at the time of peak flow.

The potential for harvesting-related changes in streamflow, particularly snowmelt-dominated spring peak flows, was one of the key hydrologic changes of concern. Information on quantitative relationships between forest cover removal or regrowth and streamflow at the watershed scale in British Columbia is not widely available; therefore, changes in streamflow generation processes at the stand scale were assumed to result in changes at the watershed scale if the area affected was of sufficient size. In completely forested watersheds, the potential for hydrologic change was assumed to be large if much of the forest was removed. In a watershed with limited distribution of forest cover—for example, where a large portion of the watershed was alpine—even if most of the forest was altered, little change in streamflow would be expected since flows would originate mainly in the non-forested area. It was further assumed that forest regrowth, or recovery, would mitigate the effects of past development. Deactivation of roads was also considered.

The ECA of a clearcut is derived by reducing the total area cut by recovery, which is estimated from relationships between snow accumulation and melt and crown closure (Winkler and Roach 2005) or tree height (Hudson and Horel 2007). The cumulative ECAs for all openings are summed to provide an ECA for the entire watershed. High ECAs would trigger a detailed field assessment of watershed condition. Lewis and Huggard (2010) combined snow and stand information from the coast and the interior, and from research in similar forest types in Montana, to derive a single relationship between ECA and tree height.

Several sources of uncertainty complicate the application of recovery curves and simple indices of hydrologic change, such as ECA. The changing hydrologic function of a growing forest stand depends on the tree and understorey species present, tree spacing, climatic characteristics, and site topography. Consequently, estimates of recovery at one site may not apply to another. Standard forest inventory metrics, such as tree height or stocking, may not best represent the influence of forests on hydrologic processes, including interception, evaporation, soil moisture, and water yield. These processes are generally better correlated with leaf area, canopy density, or basal area (Pomeroy et al. 2002; Talbot and Plamondon 2002; Teti 2003). The complex interactions between forests and hydrologic processes are highly dependent on the weather. For example, in the southern interior of British Columbia, attributes of the forest canopy account for some of the variability in hydrologic processes, such as snow accumulation and melt, among stands, but year-to-year differences in snowfall patterns account for the largest proportion of the variability (Winkler et al. 2005; Winkler and Moore 2006). Therefore, estimates of recovery derived from measurements made over a relatively small number of seasons may not adequately represent the range of weather conditions expected within a given watershed, and at best, represent average conditions. The greatest uncertainty lies in the application of stand-scale recovery estimates and ECA indices to the evaluation of hydrologic change at the watershed scale. Linkages between stands, hillslopes, and entire watersheds are complex and vary with the weather and the watershed; consequently, these linkages have not been quantified. A 50% recovery in snow accumulation or melt may not represent a 50% recovery in peak flow magnitudes. The ECA index provides only an indication of potential hydrologic change based on the extent of disturbance. It should not be used as a substitute for professional analyses and field assessments.

SUMMARY

The effects of forest disturbance on hydrologic processes and the generation of streamflow are highly variable and are influenced not only by the disturbance, but also by the weather and the biophysical characteristics, of the watershed. This chapter has described how forest harvesting and natural disturbance-related changes in fundamental hydrologic processes, including precipitation, interception, evaporation, and soil moisture, can influence hillslope flow, groundwater, and streamflow. The details of past and ongoing research in British Columbia and elsewhere can be found in the references cited. New initiatives are building on this foundation by linking stand-scale processes to watershed response; by conducting modelling over greater temporal and spatial scales, with scenarios of extensive harvesting and changing climates; and by conducting field research to quantify the effects of natural disturbance on hydrology and hydrologic recovery.

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