CHAPTER 19

Climate Change Effects on Watershed Processes in British Columbia

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INTRODUCTION

A changing climate in British Columbia is expected to have many important effects on watershed processes that in turn will affect values such as water quality, water supplies, slope stability, and terrestrial and aquatic habitats. In many parts of British Columbia, the effects of too much or too little water have already been observed and it is possible that an increased probability of droughts, floods, and landslides will result in considerable socio-economic, biological, and (or) physical changes in the future (Spittlehouse and Stewart 2004; Walker and Sydneysmith 2007). The influence of climate change on watershed processes is critically important to understand and to manage for now and in the future, as these functions directly determine human well-being in terms of public health, the economy, communities, and cultures.

In this chapter, we provide a summary of research detailing recent climate changes in British Columbia and possible future climate scenarios. We then discuss how watershed processes may be affected by climate change, and the implications of these changes to hydrology, geomorphology, and aquatic ecology in British Columbia. We conclude with a discussion of requirements for incorporating climate change–affected watershed processes into hydrologic models used at the forest management scale.

This chapter does not provide an overview of the causes of climate change, global climate model projections, downscaling models, or the key issues surrounding them. Further information on these topics can be found in Barrow et al. (editors, 2004), Intergovernmental Panel on Climate Change (2007), Parry et al. (editors, 2007), Parson et al. (2007), Randall et al. (2007), and Solomon et al. (editors, 2007). For material specific to British Columbia, the reader is referred to Rodenhuis et al. (2007), Spittlehouse (2008), and Chapter 3 (“Weather and Climate”).

Historical Trends in Air Temperature and Precipitation

Historical trends[1] in air temperature and precipitation provide important context against which future climate projections may be evaluated. Trend results, however, vary with the time period of analysis (i.e., 30, 50, 100 years), and in particular with the starting point of any trend calculation. Climate variability from atmosphere-ocean oscillations, such as El Niño–Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO), Arctic Oscillation (AO), and Pacific North American Pattern (PNA), can also complicate historical trends, and may amplify responses (Gershunov and Barnett 1998; Storlazzi et al. 2000) or cause changes of the same or greater magnitude than those in historical, long-term trends (Rodenhuis et al. 2007). For example, the 100-year trend analysis conducted over British Columbia is sensitive to the early 1920s drought period that occurred during a warm-PDO phase (Zhang et al. 2000). Further discussion of the influence of sea surface temperatures and large-scale atmospheric circulation patterns on British Columbia's climate can be found in Chapter 3 ("Weather and Climate").

Analyses of historical climate records for British Columbia show a rise in annual air temperatures, with the greatest warming occurring in the winter (Rodenhuis et al. 2007). Across the province, warming has been greater in the north than in the southern and coastal regions (Table 19.1; Figure 19.1). For example, temperature trends from 1971 to 2004 (updated from Rodenhuis et al. 2007) show increased annual mean temperatures and increased winter mean temperatures over British Columbia (Table 19.1). Nighttime temperatures have increased more than daytime temperatures (Vincent and Mekis 2006). This change may be associated with an increase in high clouds[2] occurring at nighttime and a decrease in low–middle cloudiness that might have contributed to the warming of daily minimum and maximum temperatures (Milewska 2008). The changes in temperatures over the past 50 years also have been linked to increased atmospheric water vapour and associated dew point and specific humidity trends during the winter and spring (Vincent et al. 2007).

Changes in daily extreme temperatures have also been observed in Canada. A global study found significant decreases in the number of days with extreme low daily temperatures, while increases in the number of extreme warm days were not significant over the 20th century (Easterling et al. 2000). In Canada, Bonsal et al. (2001) investigated seasonal extremes in southern Canada from 1900 to 1998. These authors found that fewer extreme low-temperature days occurred during winter, spring, and summer, and that the number of extreme hot days did not change from 1900 to 1998 (Bonsal et al. 2001). Some of these changes are related not only to climate change, but also to climate variability, such as ENSO (Bonsal et al. 2001). The warm (cold) phase of ENSO was associated with a significant increase (decrease) in the occurrence of warm (cold) spells and the number of extreme warm (cold) days across most of Canada over the 1950–1998 period (Shabbar and Bonsal 2004).

Trends in annual precipitation across British Columbia for the 100-, 50-, and 30-year periods are variable both spatially and through trend periods, as compared to temperature trends (Table 19.1; Figure 19.2). In general, average annual precipitation has increased (1.4 mm/month per decade) over the past 100 years (Table 19.1), with larger percentage increases occurring in regions with comparatively lower annual precipitation (Rodenhuis et al. 2007). Precipitation indices compiled for Canada over the 20th century illustrate an increase in annual snowfall from 1900 to 1970, followed by a considerable decrease until the early 1980s (Vincent and Mekis 2006). Generally, precipitation over the past 50 years has decreased over the southern portion of the province, most notably in the south coastal region, the Columbia River basin, and in the Peace watershed regions during winter. Conversely, precipitation has increased in spring, particularly in the southern regions (Rodenhuis et al. 2007).

Climate oscillations play a role in the above-mentioned precipitation trends, as presented and discussed in Chapter 3 ("Weather and Climate"). The

1 Paleoclimatic trends in precipitation and temperature are not considered in this chapter.
2 Clouds with a base height of 6–12 km above the Earth's surface, referred to as cirrus, cirrocumulus, or cirrostratus clouds.

TABLE 19.1 *Historical trends in 30-, 50-, and 100-year periods (1971–2004, 1951–2004, and 1901–2004, respectively). Temperatures and precipitation trends calculated from mean daily values as seasonal (winter as December–February and summer as June–August) and annual averages. Values provided for the province as a whole, and for the Coastal, South, North, and Georgia Basin regions (see Figure 19.1).*

Season	Time period (years)	British Columbia	South	North	Coastal	Georgia Basin
Temperature (°C per decade)						
Winter	30	0.77	0.77	0.90	0.60	0.44
	50	0.45	0.38	0.59	0.35	0.22
	100	0.22	0.22	0.25	0.18	0.15
Summer	30	0.33	0.28	0.32	0.40	0.52
	50	0.18	0.21	0.14	0.19	0.30
	100	0.07	0.08	0.07	0.05	0.06
Annual	30	0.41	0.41	0.41	0.42	0.45
	50	0.25	0.25	0.27	0.22	0.22
	100	0.12	0.12	0.13	0.10	0.11
Precipitation (mm/month per decade)						
Winter	30	–4.28	–4.90	–2.47	–6.08	–8.06
	50	–1.90	–2.44	–0.55	–3.06	–5.35
	100	1.77	1.26	1.19	3.39	1.78
Summer	30	1.41	1.83	0.05	3.50	–1.80
	50	1.31	1.28	0.97	2.11	–0.27
	100	1.18	1.37	1.21	0.91	0.93
Annual	30	0.75	1.06	0.07	1.63	–0.42
	50	0.67	0.86	0.41	1.01	–0.43
	100	1.41	1.22	1.06	2.25	1.20

impact of the 1976 positive PDO phase shift has been well documented in British Columbia and the Pacific Northwest (i.e., reduction in snowpack: Moore and McKendry 1996; fisheries effects: Mantua et al. 1997). The recent 30-year trend period (1971–2004) falls almost entirely within this positive phase of the PDO. The positive phase of the PDO in British Columbia has been noted to cause warming throughout western Canada and decreased precipitation in the mountainous and interior regions of the province (Stahl et al. 2006).

Trends in extreme events for the past 50 years indicate that seasonal patterns of precipitation in western Canada are changing. In the Pacific Northwest, recent shifts in the occurrence and magnitude of extreme rainfall intensities have been observed, with storms becoming more frequent and of a greater magnitude for a given frequency. Madsen and Figdor (2007) observed an 18% increase in extreme precipitation events over the 1948–2006 period. Similarly, Rosenberg et al. (2009) observed significant increases in extreme precipitation events in the Puget Sound, with increases up to 37% from the 1956–1980 period to the 1981–2005 period. These increases represented a shift in which the 50-year storm event became an 8.4-year storm event. Stone et al. (2000) found a significant increase in heavy rainfall events during May, June, and July from 1950 to 1995. Zhang et al. (2000) examined the differences between the first and the second half of the century and found an increase in both extreme wet and extreme dry conditions in summer (1950–1998). Although the national trend shows that only the number of days with heavy precipitation increased significantly over the past 50 years, some stations in southern British Columbia show significant increases in two extreme indices: (1) the highest 5-day precipitation, and (2) very wet days (the number of days with precipitation ≥ 95th percentile) (Vincent and Mekis 2006).

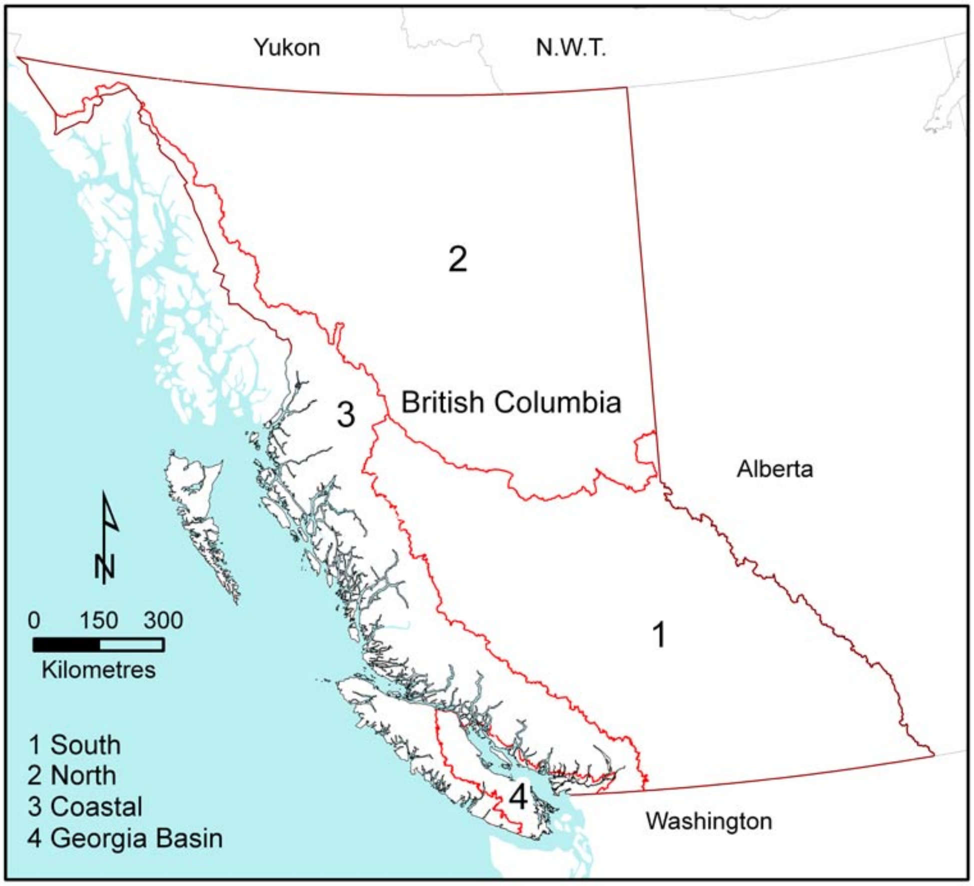


FIGURE 19.1 *Map of British Columbia regions used in Table 19.1.*

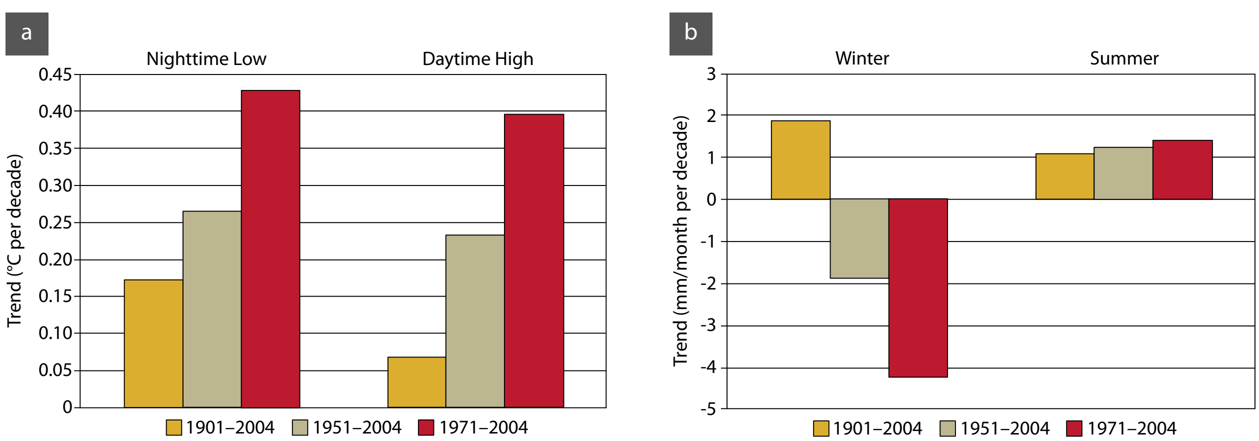


FIGURE 19.2 *Mean of all trends across British Columbia for (a) minimum temperature (nighttime low) and maximum temperature (daytime high), and (b) precipitation based on CANGRID[3] gridded time series of historical climate. (Data from Environment Canada)*

3 CANGRID is a gridded 50 km² product developed by Environment Canada, based on Adjusted Historical Canadian Climate Data (AHCCD).

Historical Trends in Snow, Seasonal Ice Cover, Permafrost, and Glaciers

The interactions between increased temperature and shifts in precipitation form (i.e., snow to rain) in British Columbia are complex and not fully understood. Research in the western United States suggests that the snow-to-rain ratio is changing and less snow is falling during winter at lower elevations on the west coast of the United States (Knowles et al. 2006). Mote et al. (2005) reported a general decline in snowpacks over much of western North America from 1950 to 1997, despite increases in precipitation. Between the mid-1980s and 2008, McCabe and Wolock (2009) reported above-average winter temperatures and below average snow water equivalent (SWE) in the western United States. In British Columbia, the Ministry of Environment reported overall decreasing trends in April 1st SWE from 1956 to 2005 based on data from 73 long-term snow courses (63 decreased, 10 increased). The largest decreases occurred in the mid-Fraser Basin, whereas the Peace, Skeena, and Nechako Basins had no notable change over the 50-year study period, and overall the provincial average SWE decreased 18% (B.C. Ministry of Environment 2007).

Increasing temperatures have also affected the length and date of seasonal lake ice cover. A Canada-wide study showed significantly earlier lake "ice-free" dates for the 1951–2000 period (Duguay et al. 2006). In several British Columbia lakes, the first melt date and ice-free date decreased by 2–8 days per decade from 1945 to 1993, whereas the duration of ice cover decreased by up to 48 days over the 1976–2005 period (B.C. Ministry of Environment 2002; Rodenhuis et al. 2007).

Permafrost in Canada is also changing. In British Columbia, sporadic discontinuous permafrost exists in the northern latitudes, whereas isolated patches of permafrost exist south of Prince Rupert and Fort St. John and at higher elevations to the United States border through the Coast and Rocky Mountain ranges. Recent analysis has shown that permafrost in many regions of North America is warming (Brown et al. 2004). In looking at recent trends from the Canadian Permafrost Thermal Monitoring Network, Smith et al. (2005) reported that while the timing and magnitude of warming varied regionally, warming trends in permafrost were largely consistent with air temperature trends observed since the 1970s. In their analysis, Smith et al. (2005) noted that local conditions importantly influenced the response of the permafrost thermal regime.

Glaciers in British Columbia are also out of equilibrium with the current climate and are adjusting to changes in seasonal precipitation and elevated temperatures, with widespread glacial volume loss and retreat in most regions. For example, the Illecillewaet Glacier in Glacier National Park has receded over 1 km since measurements began in the 1880s (Parks Canada 2005). In general, glaciers have been retreating since the end of the Little Ice Age (mid-19th century), although some glaciers have exhibited periods of stability at the terminus and even advances (Moore et al. 2009). For example, Moore et al. (2009) reported that the terminus of Illecillewaet Glacier remained stationary from 1960 until 1972, and then advanced until 1990. It has subsequently resumed its retreat. This behaviour is consistent with the decadal time scale of glacier terminus response to climate variability (Oerlemans 2001). Schiefer et al. (2007) reported that the recent rate of glacier loss in the Coast Mountains is approximately double that observed for the previous two decades. A compilation of glacier area changes in the period 1985–2005 indicates glacier retreat in all regions of the province (Bolch et al. 2010), with an 11% loss in total glacier area over this period. On Vancouver Island, the central Coast Mountains, and the northern Interior ranges, ice-covered areas have declined by more than 20% over this period (Bolch et al. 2010).

The dominant trend of glacier retreat has influenced streamflow volumes, leading to declines in late-summer streamflow (Moore et al. 2009). However, Moore et al. (2009) also noted that some exceptions exist, particularly in glacier-fed watersheds in northwest British Columbia and the Yukon that experienced increased flows over recent decades (due to ice melt), consistent with the findings of Fleming and Clarke (2003). This is a function of the glacier-covered area in a catchment; runoff per unit area from glaciers is higher as glaciers retreat, but the total glacier contribution to a basin declines with reductions in glacier area. This is discussed further in "Glacier Mass Balance Adjustments and Streamflow Response," below.

Historical Trends in Landslides and Other Geomorphic Processes

In British Columbia, much of the contemporary landscape has been shaped by previous glacial periods (see Chapter 2, "Physiography of British Columbia"). Persistent paraglacial effects exist in British Columbia whereby secondary remobilization of Quaternary sediments has led to a relationship of increasing contemporary sediment yield (sediment yield per unit area) with increasing drainage area (Church and Slaymaker 1989). This suggests that, at a landscape level, British Columbia is still responding to the last Cordilleran glaciation.

Alpine glacial retreat has led to a variety of geomorphic processes in periglacial (proximal to glaciers) and alpine environments (O'Connor and Costa 1993; Evans and Clague 1997; Ryder 1998; Moore et al. 2009). For example, debuttressing of support to lateral slopes caused by glacial retreat has led to deep-seated rock failures in some areas (Holm et al. 2004). Flooding attributed to the failure of moraine-dammed lakes impounded by Little Ice Age deposits has also been observed throughout the Coast Mountains (McKillop and Clague 2007). Glacial lake outburst floods (jökulhaups) have occurred in areas of the province, predominantly along the British Columbia–Alaska border and in the southwest Coast Mountains (Clague and Evans 1997; Geertsema 2000). Though the relationship to climate variability and change in the region is not completely understood, in recently exposed glacial forefield areas, sediment production rates have increased from both primary erosion of exposed slopes and remobilization of stored channel deposits (Orwin and Smart 2004; Schiefer and Gilbert 2007).

Landslides in British Columbia are often triggered by major storm events (see Chapter 8, "Hillslope Processes," and Chapter 9, "Forest Management Effects on Hillslope Processes"). Septer and Schwab (1995) summarized extreme rainstorm and landslide events in northwest British Columbia over the 1981–1991 period. Guthrie and Brown (2008) estimated the variability in landslide rates over the Holocene and suggested that increases in landslide rates doubled during shifts from drier to wetter periods. Shifts in landslide rates attributed to changes in climatic regimes are thought to be of a similar order of magnitude or smaller when compared to landslide responses to forest management in the 20th century (Campbell and Church 2003; Guthrie and Brown 2008).

In northern British Columbia, shallow slides and debris flows have occurred during infrequent large storms. Egginton et al. (2007) noted that large cyclonic storms and convective thunderstorms have triggered recent landslides in the north, but large slides (greater than 0.5 Mm^3) are more typically preceded by long periods of wetter or warmer climate. Large rock slides appear to have responded to warming trends of the past few decades by destabilizing from snow and ice melt and increasing freeze thaw processes (Egginton 2005; Geertsema et al. 2007). Larger soil slides are more common during long periods (years to decades) of above-average precipitation, likely from soil saturation (Egginton 2005; Geertsema et al. 2007). Prolonged periods of increased precipitation or temperature have increased the vulnerability of slopes to failure in these areas, whereas large or intense storms are often the trigger. All of these conditions are expected to be further enhanced under current climate change scenarios.

Historical Trends in Groundwater Levels

Spatial and temporal variations in groundwater levels are caused by both human and natural factors. Human factors often involve groundwater extraction (e.g., pumping and irrigation) or land use change (urbanization or deforestation). Natural factors may include the effects of tides on coastal aquifers, the influence of seasonal variations in precipitation and recharge, and the effects of longer-duration climatic cycles. Historical time series of groundwater levels (groundwater hydrographs) often illustrate cyclic behaviour ranging from short term (i.e., hours, days) to long term (i.e., years, decades). Long-term monitoring of groundwater levels is therefore necessary to quantify groundwater trends and discern the effects of climatic changes on groundwater hydrology. (A brief introduction to groundwater hydrology is provided in Chapter 6, "Hydrologic Processes and Watershed Response").

Across British Columbia, groundwater levels and quality are monitored in 145 observation wells (as of July 2009).[4] These observation wells are located primarily in developed aquifers to examine the effects of water extraction and development on groundwater availability and quality. Unfortunately, because of a short data record, and location in areas influenced by human activity, many of the wells are likely unsuitable for climate change detection purposes.[5] Furthermore, British Columbia contains a wide range of

4 For details about the observation well network, go to: www.env.gov.bc.ca/wsd/data_searches/obswell/index.html.

5 Moore, R.D., D.M. Allen, and K. Stahl 2007. Climate change and low flows: influences of groundwater and glaciers. Nat. Resour. Can., Can. Climate Action Fund, Ottawa, Ont. Can. Climate Action Fund Proj. No. A875. Unpubl. report.

aquifer types (Wei et al. 2009) with varying physical properties that have a strong control on groundwater response to climatic changes.

Most recently, declining groundwater levels trends have been reported at 35% of the provincial observation wells for the 2000–2005 period (B.C. Ministry of Environment 2007). This represents a percentage increase in the rate of decline compared to groundwater level declines reported for only 14% of wells for the 1995–2000 period. The greater decline was attributed to human activities rather than climate causes, as the majority of monitoring wells showing declines were located in regions with intense urban development and groundwater use (i.e., Vancouver Island, Gulf Islands, and the Okanagan Valley). Unravelling such complexity of causal factors is confounded by climate variability; however, analysis of groundwater hydrographs in combination with climate and streamflow data offers some insight. Fleming and Quilty (2006) investigated groundwater and stream hydrographs for a small area of the lower Fraser Valley (four observation wells) and found that groundwater levels tend to be higher during La Niña years and lower during El Niño years because of the associated variations in precipitation and recharge. These results also indicate that groundwater levels can lag in response to climate variation. Moore et al.[6] examined a larger subsample of the provincial well-monitoring database and correlated groundwater levels with nearby streamflow and precipitation records over a 20–30 year period. Their results indicate that groundwater levels have decreased over the areas examined, whereas winter precipitation and recharge increased over the same time period. The results are highly variable, however, and likely related to differences in aquifer properties, surface water–groundwater interactions, and the effects of water withdrawals.[7]

Historical Trends in Streamflow

Province-wide studies examining historical (undisturbed) streamflow patterns are not generally available. This is likely related to the limited availability of long-term hydrologic records and the large variability in hydrologic regimes occurring across British Columbia. An important component of diagnosing the changes in streamflow is the isolation of natural and human-caused disturbance effects (e.g., forest harvesting) from those effects attributed to climate changes and variability. In many areas in British Columbia, this is a challenge because of the predominance of watershed disturbances.

Several studies have documented streamflow trends for provincial watersheds. Important documented changes include observations of earlier spring peak freshets and prolonged, dry late-summer periods for streams in south-central British Columbia (Leith and Whitfield 1998; Whitfield and Cannon 2000). These changes are attributed to a greater percentage of rain falling versus accumulating as snow, although this hypothesis was only recently verified using standard statistical approaches (P. Whitfield, Environmental Studies, Meteorological Service of Canada, pers. comm., 2007). One recent study using trends in sequential 5-day periods observed that rising air temperatures in December and early January led to decreased snowpack, increased runoff from fall to early winter, and decreased flows from May through August (1976–2006) in the Little Swift River Basin, near Barkerville (Déry et al. 2009).

In a Canada-wide survey, Zhang et al. (2001) documented declining trends in annual mean streamflow for the past 30–50 years (three time periods: 1967–1996, 1957–1996, and 1947–1996); however, these results were variable across seasons, with an increase in mean monthly streamflow across Canada in March and April, and decreases in summer and fall. For many of the variables studied, Zhang et al. (2001) identified southern British Columbia as a significantly affected region. Several important streamflow metrics, including the date of spring high-flow season, annual maximum daily mean streamflow, centroid (date) of annual streamflow, and spring ice break-up, occurred earlier in the season (Zhang et al. 2001).

Advances of 10–30 days in the centre of mass of annual streamflow (i.e., the date by which half of the annual total streamflow runoff has occurred) have been measured in streams in Pacific North America from 1948 to 2002 (Stewart et al. 2005). Other analyses of changes in the date of the centre of volume (a similar metric), gave varying results when computed for the calendar year and hydrologic year (Déry et al. 2009). These varying results illustrate that analyses can be strongly affected by the date metrics used to identify trends in streamflow.

6 Ibid.
7 Ibid.

The magnitude and direction of the changes to streamflow vary across British Columbia depending on the time period of analysis and the hydro-climatic region. For example, Rodenhuis et al. (2007) reported differing trends in mean annual streamflow for British Columbia than those reported by Zhang et al. (2001). Rodenhuis et al. (2007) attributed these differences to the different PDO phases that occurred during their analysis period (1976–2005) compared to the 1967–1996 period used by Zhang et al. (2001).

Analyses conducted for this chapter at several stations (Table 19.2) representative of different regions in British Columbia (updated from Rodenhuis et al. 2007) indicate that the largest amount of change appears to be occurring in coastal watersheds. Regimes are shifting towards increased winter rainfall, and declining snow accumulation, with subsequent changes in the timing and amount of runoff (i.e., weakened snowmelt component). This, coupled with decreased summer precipitation, is shifting the streamflow pattern in coastal watersheds. In other systems throughout British Columbia, increasing temperatures over the past 15 years and changing precipitation patterns have altered the magnitude and timing of snowpack and spring melt. In the Okanagan region, for example, changes in snowpack accumulation are resulting in an earlier spring peak streamflow and leading to declining maximum flows and extended minimum flows in late summer and early fall. The Fraser and Columbia River nival-glacial systems show increased peak flows and lower recessional flows, illustrating changes in the associated watersheds, perhaps away from a glacier-dominated regime towards a snow-dominated regime with an earlier freshet and faster recessional period.

Trend analysis of sequential 5-day average runoff values was conducted at a collection of stations representative of several hydro-climatic regimes in British Columbia (Table 19.2). Two periods were investigated: (1) 1959–2006 (48 years, Figure 19.3a), and (2) 1973–2006 (34 years, Figure 19.3b). Analysis results for the longer record (1959–2006; Figure 19.3a) show that the nival-supported pluvial Chemainus River in British Columbia's coastal region had a varied response over the record, with predominantly increased flow in winter and decreased flow during May. The nival–glacial Adams River had increased flow in spring and decreased flow in all other seasons. The nival/hybrid Similkameen River located in the Okanagan and the nival Swift River in the northwest had increased winter and spring flows and decreased summer flow. In the Peace region, the nival Sikanni Chief had increased flow in December through April and decreases in May through November.

For the more recent years of record (1973–2006; Figure 19.3b) the above-mentioned stations and one additional record for Fry Creek, located in the Columbia River Basin, were analyzed. Fry Creek is a small, nival–glacial system that had decreases in flow in June, July, August, and September over this period, similar to the Adams River in the Interior, another glaciated system. Decreased flow in September was most prominent in the Adams. Decreases in streamflow also occurred for the nival-hybrid Similkameen River and the nival Swift River during the summer months. In these more recent years of record (1973–2006), increased streamflow was observed from November to April and decreased streamflow from June to September across all stations, with the exception of the Sikanni Chief River, which had decreases in October through December, and the Chemainus River, which had decreases in February.

The trend analysis shown here employed techniques developed by Déry et al. (2009). Déry et al. (2009) found that in the pluvial systems of the Yakoun, Zeballos, and San Juan Rivers, positive trends (increasing streamflow) were observed in winter and negative trends (decreasing streamflow) were ob-

TABLE 19.2 *Water Survey of Canada gauging station information for various streamflow regimes in British Columbia*

Name	WSC station no.	Region	Streamflow regime	Basin size (km^2)	Continuous years of record	Period of record
Chemainus River	08HA001	Coastal	nival supported pluvial	355	52	1955–2006
Similkameen River	08NL007	Okanagan	nival/hybrid	1185	62	1945–2006
Adams River	08LD001	Interior	nival–glacial	3080	58	1949–2006
Fry Creek	08NH130	Columbia	nival–glacial	586	34	1973–2006
Sikanni Chief River	10CB001	Peace	nival	2160	47	1960–2006
Swift River	09AE003	Northwest	nival	3320	48	1959–2006

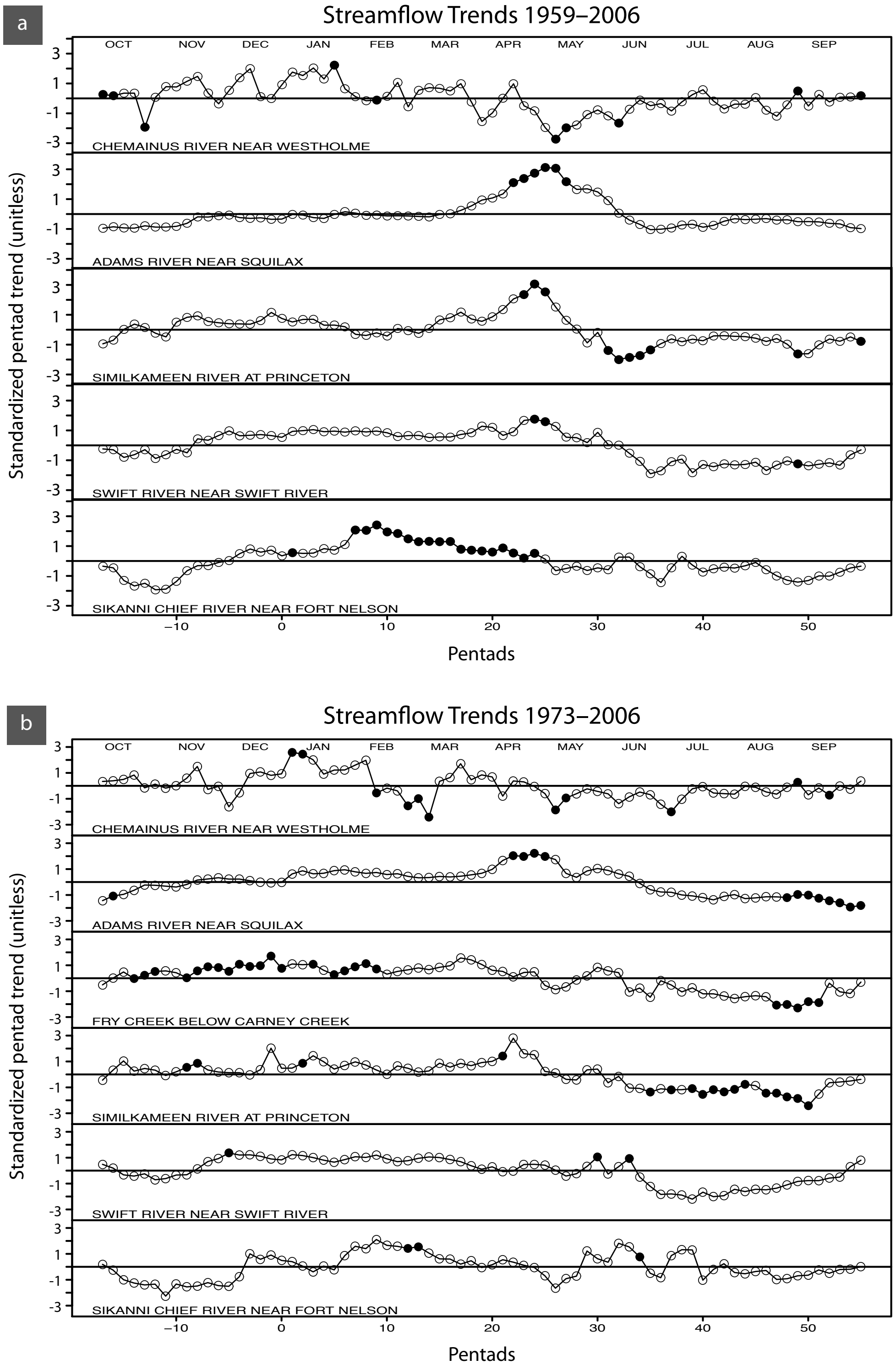


FIGURE 19.3 *(a) Streamflow sequential 5-day average runoff trends for the long-term historical period 1959–2006 for five streams located in different regimes throughout British Columbia; and (b) streamflow sequential 5-day average runoff trends for the recent historical period 1973–2006 for six streams located in different regimes throughout British Columbia (see Table 19.2 for information on gauging stations). Solid circles are significant results, open circles are non-significant results at the 95% confidence interval. Standardized results above zero indicate increased streamflow, whereas results below zero indicate decreased streamflow.*

served in summer from 1972 to 2006. The nival and glacial systems (Dore, Tuya, and Little Swift Rivers) had large positive trends (increasing streamflow) during spring followed by strong negative trends (decreasing streamflow) in summer, which suggests a phase shift towards earlier spring freshets. Surprise Creek, a nival–glacial system, showed a pronounced positive discharge trend throughout the summer, unlike many other similar rivers in western Canada (Déry et al. 2009).

Although trends illustrate how streamflow is changing over long periods, events caused by climate variability may result in short-term shifts in streamflow. (Streamflow variability is discussed in Chapter 3, "Weather and Climate.") The influence of the modes of climate variability (e.g., ENSO and PDO) on streamflow is evident and often confounds identification of historical trends. On the south Coast, some streams that are normally rainfall-dominated have snowmelt runoff in the spring during cool La Niña years (Fleming et al. 2007). This can result in years with two streamflow peaks in watersheds where normally only one would occur (e.g., Figure 19.4a, Chemainus River). During El Niño years, substantially less streamflow may occur from May to August in snowmelt-dominated basins, especially those in the Okanagan Basin (e.g., Figure 19.4b, Similkameen River; Rodenhuis et al. 2007) but may have little effect in the north of the province where ENSO signals are less pronounced (e.g., Figure 19.4c, Swift River). Warm PDO phases, such as the one that occurred from 1977 to 1998, advance the spring or summer freshet, lower peak flows, and cause drier summer periods for many streams in British Columbia (Zhang et al. 2000). Some exceptions occur in northern British Columbia where the opposite response can occur during warm PDO phases (e.g., Figure 19.4d, Sikanni Chief River; Rodenhuis et al. 2007). This is important to note because climate and streamflow responses during different climate oscillations are not necessarily uniform across regions, and often depend on (or are related to) the hydrologic regime, physiography, and climate of the region under consideration.

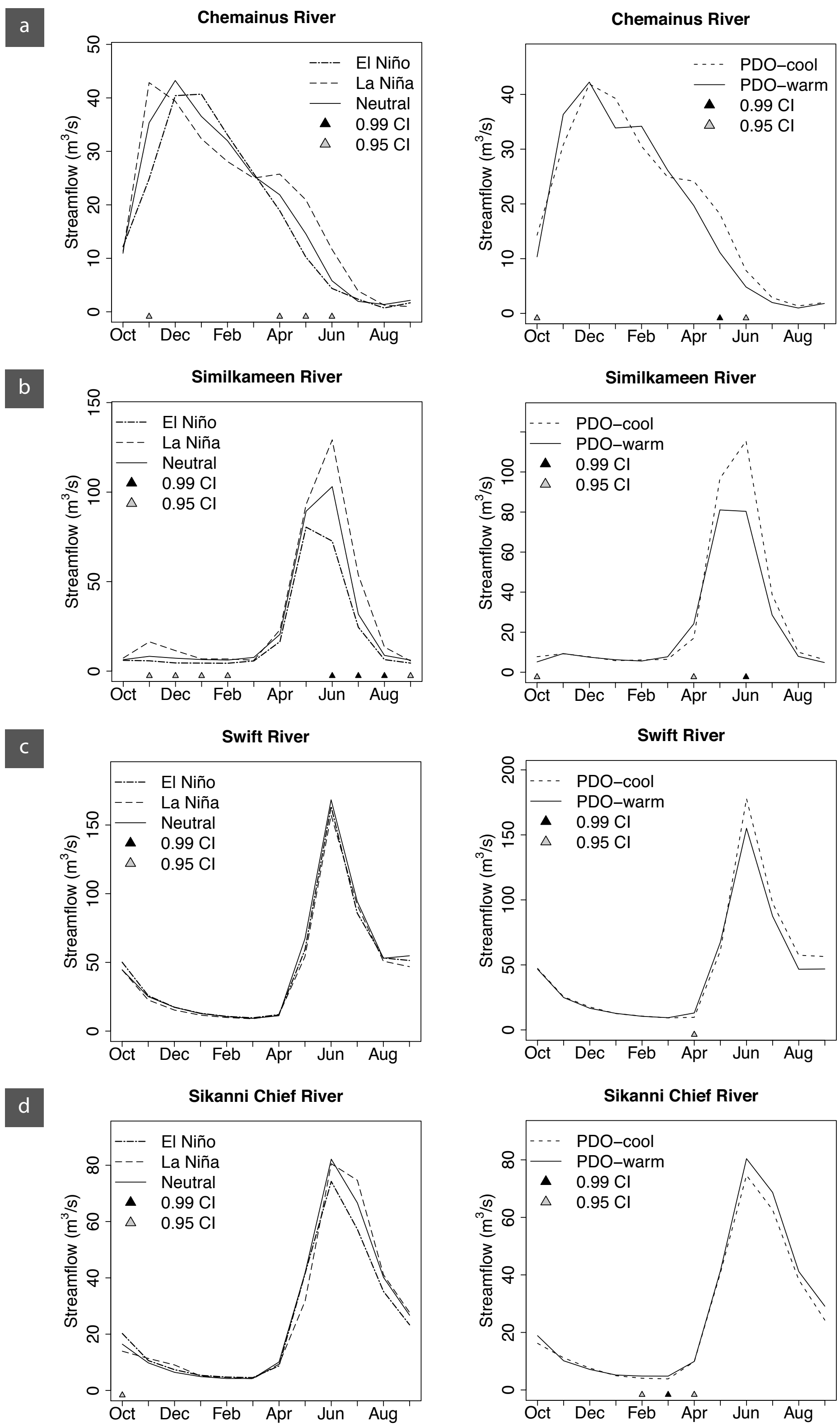


FIGURE 19.4 *Monthly average streamflow occurring during ENSO periods (left-hand plots) and PDO-cool (right-hand plots) for: (a) Chemainus River, (b) Similkameen River, (c) Swift River, and (d) Sikanni Chief River. See Table 19.2 for information on gauging stations. The grey and black triangles indicate significant differences at the 95% (0.95 CI) and 99% (0.95 CI) confidence levels, respectively.*

British Columbia Projections by Emissions Scenario

Projections of future climates are available from numerous global climate models (GCMs) and for a range of greenhouse gas emissions scenarios. These emissions scenarios[8] depend on future population, technology, economic growth, and international trade (Intergovernmental Panel on Climate Change 2007), but do not consider intentional co-operation to prevent climate change. Importantly, each GCM can project different future climates for the same emissions scenario because each models specific processes (e.g., evaporation) differently.

In general, GCM projections agree in the direction and magnitude of temperature changes, but projections of precipitation change are more varied in both direction and magnitude (Barnett et al. 2005; Rodenhuis et al. 2007). Figures 3.24–3.28 in Chapter 3 ("Weather and Climate") illustrate anticipated climate changes for British Columbia, based on simulations by the Canadian Global Climate Model (CGCM2) for the A2 scenario.[9] The Canadian model tends to project warmer and wetter summers compared with the United Kingdom's Hadley Centre model (Spittlehouse 2008). Even under the low emissions scenario (B1), the amount of climate change projected for British Columbia by the end of the century (> 2° C; Figure 19.5, teal line) is comparable to the historical differences between the coldest of the cold years and the warmest of the warm years (troughs and peaks of solid black line); in other words, an entirely different temperature regime is projected for British Columbia than that of the last century.

The University of Victoria's Earth System Climate Model (ESCM; Eby et al. 2009) has also been run for

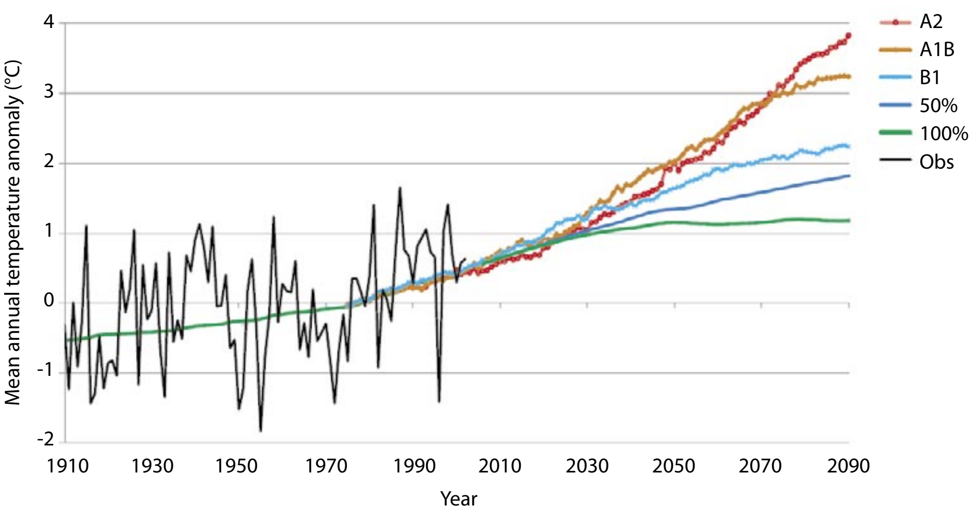


FIGURE 19.5 *Mean annual temperature anomalies for British Columbia using 1961–1990 baseline of the UVic ESCM over the 21st century for emission reduction scenarios compared to median of AR4 GCM projections for the A2, B1, and A1B SRES emissions scenarios. The GCM projections are displayed as 20-year centred means to remove annual and decadal variability. The 50% and 100% lines represent percent reductions by 2050; that is, to half of 2006 greenhouse gas emissions (blue) and to zero net emissions (carbon neutral; green). Sources: Environment Canada (historical data – CANGRID), Lawrence Livermore National Laboratory (GCM projections), and UVic Climate Modelling Lab (ESCM projections).*

8 SRES refers to the *Special Report on Emissions Scenarios*, published by the Intergovernmental Panel on Climate Change (IPCC) in 2000. For summary information about these scenarios, see Intergovernmental Panel on Climate Change (2007).

9 The A2 scenario is one of the highest emissions scenarios of the SRES group. In contrast, the B1 emission scenario represents roughly half of the emissions of A2.

several emissions that incorporate emissions reductions (Figure 19.5). It is reasonable to expect that the ESCM projections would be similar to those of an ensemble of several GCMs for similar emissions, as the ESCM follows the median of these GCMs for trajectories leading to similar greenhouse gas concentrations (not shown). These lower emissions are important because they represent the concentrations now the subject of serious policy debate. Importantly, Figure 19.5 shows that considerable climate change is projected for the province by the end of the century even under large emission-reduction scenarios, thus requiring adaptation.

2050s Projections for British Columbia

An ensemble of 30 projections from 15 GCMs was used to compute a range of projections for the 2050s (2041–2070) climate of British Columbia (Rodenhuis et al. 2007). Based on these results, the provincial annual average temperature is projected to warm by 1.7° C compared with the recent 1961–1990 period (Table 19.3; Figure 19.6). Uncertainty is represented by the range 1.2–2.5° C (the 10th–90th percentile of projections). The 2050s annual precipitation is projected to increase by 6%, with a range of 3–11%. The seasonal temperature projections were relatively uniform, but seasonal precipitation projections varied from 2% drier to 15% wetter for winter and 9% drier to 2% wetter for summer. Further information on model projections can be found in Rodenhuis et al. (2007). All models and emissions scenarios project an increase in winter and summer temperatures with the greatest increases for the higher emissions scenarios.

Regional 2050s Projections

At a regional scale, the same ensemble of 30 projections described above in "2050s Projections for British Columbia" (above) show that projected warming will be greater in the Interior than on the Coast (Table 19.3). Changes in precipitation will vary spatially as well as temporally. Southern and central British Columbia are expected to become drier in the summer, whereas northern British Columbia will likely become wetter (Table 19.3; see also Chapter 3, Figures 3.27 and 3.28). Overall, wetter winters are expected across British Columbia (Rodenhuis et al. 2007).

TABLE 19.3 *Changes in seasonal[a] and annual air temperature and precipitation by the 2050s for regions in British Columbia for the ensemble of 30 GCM projections described above in "2050s Projections for British Columbia" (updated from Rodenhuis et al. 2007)*

	Air temperature change (°C)				
Region	**Winter**	**Spring**	**Summer**	**Fall**	**Annual**
Columbia Basin	1.8	1.5	2.4	1.8	1.9
Fraser Plateau	1.9	1.6	2.0	1.8	1.8
North Coast	1.5	1.3	1.4	1.5	1.4
Peace Basin	2.4	1.7	1.8	1.8	1.9
Northwest	2.0	1.6	1.8	1.7	1.8
Okanagan	2.0	1.8	2.6	2.0	2.1
South Coast	1.5	1.3	1.7	1.6	1.5
British Columbia	*1.9*	*1.6*	*1.8*	*1.7*	*1.7*
	Precipitation change (%)				
Columbia Basin	7	9	–8	8	4
Fraser Plateau	8	10	–4	11	7
North Coast	6	7	–8	9	6
Peace Basin	9	9	3	10	7
Northwest	10	9	4	8	8
Okanagan	5	12	–8	8	5
South Coast	6	7	–13	9	6
British Columbia	*7*	*8*	*–3*	*9*	*6*

a Winter = December–February; Summer = June–August; Spring = March–May; Fall = September–November.

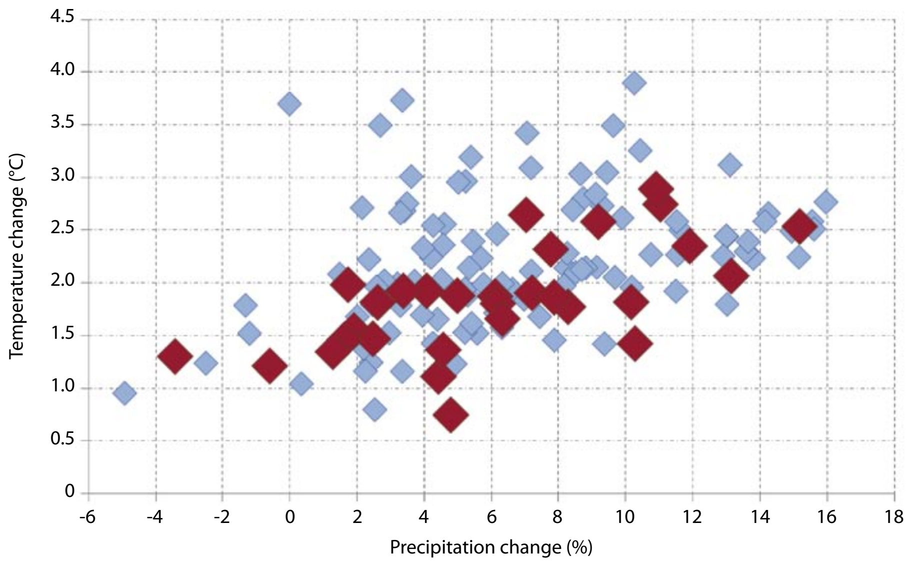


FIGURE 19.6 *Range of 2050s annual temperature and precipitation averaged over British Columbia from 140 GCM projections. The larger/darker diamonds represent the 30 projections in the ensemble described in the section "2050s Projections for British Columbia."*

Although we primarily present mean changes in climate in this chapter, future changes in the variability or the extremes of temperature and precipitation are anticipated to have many important effects on watershed processes. Changes in warm temperature extremes generally follow changes in the mean summer temperature (Kharin et al. 2007). This suggests that extreme maximum temperatures would be higher than at present and cold extremes would warm at a faster rate, particularly in areas that experience a retreat of snow with warming. An increase in the intensity and maximum amount of precipitation is also expected (Kharin et al. 2007); however, changes in extreme events may not be proportional to mean changes, and the changes may not be equal in both directions. For example, increasingly frequent extreme maximum temperatures are anticipated; however, the frequency of extreme cold temperatures is anticipated to decline in the future (Tebaldi et al. 2006; Kharin et al. 2007).

WATERSHED PROCESSES AFFECTED BY A CHANGING CLIMATE

It is expected that the effects of a changing climate on watershed processes will vary across British Columbia, depending on which specific watershed processes and responses are sensitive to change. In this section, we discuss the potential effects of a changing climate on watershed processes and outputs. Specifically, the following changes in watershed processes may be expected in British Columbia.

- Increased atmospheric evaporative demand
- Altered vegetation composition affecting evaporation and interception processes
- Decreased snow accumulation and accelerated melt
- Accelerated melting of permafrost, lake ice, and river ice
- Glacier mass balance adjustments
- Altered timing and magnitude of streamflow (peak flows, low flows)
- Altered groundwater storage or recharge
- Changes in frequency and magnitude of hillslope and geomorphic processes
- Changes in water quality, including increased stream or lake temperatures and altered chemical water quality

Atmospheric Evaporative Demand

Evaporative demand is a function of air and surface temperature, solar radiation, humidity, and wind speed (see Chapter 3, "Weather and Climate," and Chapter 6, "Hydrologic Processes and Watershed Response"). The climate scenarios previously described could increase the atmosphere's ability to evaporate water (Huntington 2008). This will occur if the saturation vapour pressure of the air (a function of air temperature) increases more rapidly than the actual vapour pressure (i.e., the vapour pressure deficit increases). It will also increase if net radiation and wind speed increase. An increase in evaporative demand would significantly affect water resources through evaporative losses from water bodies, vegetation, and soils, and through subsequent changes in water demands. Increased evaporative demand will also affect vegetation survival and growth through changes in water availability and fire risk. For example, Spittlehouse (2008) estimated the magnitude of change in evaporative demand (calculated following methods in Allen et al. 1998) for the Campbell River, Cranbrook, and Fort St. John areas using current weather station data and climate change model output for the B1 and A2 scenarios from the Canadian General Circulation Model (CGCM3). Evaporative demand, which is calculated for months when the air temperature is above 0° C, increased at all locations because of an increase in the length of time that air temperature remained above 0° C and an increase in the vapour pressure deficit (drier air). By the 2080s, evaporative demand increased by about 8% under the B1 scenario and by 15–20% under the A2 scenario (Spittlehouse 2008).

Estimates of evaporative demand and precipitation can be combined to give indicators of plant water stress and to predict water demand for agricultural irrigation and domestic use. A climatic moisture deficit occurs when the monthly precipitation is less than the evaporative demand for the month; conversely, if precipitation is greater than the evaporative demand, a moisture surplus occurs. By the 2080s under the B1 scenario, Spittlehouse (2008) reported that the deficit at Campbell River increased by 20%,[10] at Fort St. John by 25%, and at Cranbrook by 30%. For the A2 scenario, Campbell River and Fort St. John increased by 30%, whereas Cranbrook increased by 60%. The larger increase at Cranbrook reflects the decrease in summer rainfall and an initially relatively low average deficit for the 1961–1990 reference period. A moisture surplus did not occur during the summer at any of the locations for the climate change scenarios examined (Spittlehouse 2008).

Vegetation Composition Affecting Evaporation and Interception

Terrestrial vegetation influences water balance through the interception of rain and snow and the removal of water from the root zone as a result of plant transpiration and evaporation from the soil surface. As vegetation composition responds to climate change, so too will the amounts of water intercepted, evaporated, and transpired, thus altering snow accumulation and melt processes (see "Snow Accumulation and Melt," below), water balance, groundwater recharge, and ultimately streamflow and mass wasting processes. Increases in the length of the snow-free season and changes in atmospheric evaporative demand are likely to increase plant transpiration, assuming soil water is available. For example, Spittlehouse (2003) estimated that transpiration from a coastal Douglas-fir forest could rise by 6% with an increase of 2° C and by 10% with an increase of 4° C. The projected changes in climate are sufficient to affect forest productivity and species composition (Barber et al. 2000; Hamann and Wang 2006; Campbell et al. 2009). Changes may also occur in age-class distribution and in the form of vegetation (e.g., forest die-off, alpine encroachment, grassland expansion) (Breshears et al. 2005; Hebda 2007). Thus, changes to the amount of plant biomass on a site and the physiological characteristics of the new vegetation will have an important effect on water balance in the future.

Snow Accumulation and Melt

By the 2050s (2041–2070) increased air temperatures will lead to a continued decrease in snow accumulation (Rodenhuis et al. 2007; Casola et al. 2009), earlier melt (Mote et al. 2003), and less water stor-

10 Deficits are calculated for the months with an air temperature greater than 0°C. For Cranbrook, this is March to November, for Fort St. John, April to October, and for Campbell River, January to December. Formula are not appropriate for snow cover situations. Deficits occur only if monthly evaporation is greater than monthly precipitation. Cranbrook and Fort St. John have deficits in all months, whereas Campbell River has deficits only from May through September.

age for either spring freshet (Stewart et al. 2004) or groundwater storage. Changes in air temperature and wind may also affect snow adhesion in the snowpack and the subsequent amount of snow drift or scour that occurs in an area. "Wetter" snowpacks may be more resistant to redistribution by wind, which could be important for avalanche forecasting and management. The influence of vegetation on snow accumulation and melt processes (see Chapter 6, "Hydrologic Processes and Watershed Response") will also be an important factor to consider as the composition of vegetation on the landscape changes. To simplify the discussion, we next consider the implications of increased temperature on snow processes.

Projected declines in snow are most notable on the central and north coast of British Columbia and at high-elevation sites along the south coast (Rodenhuis et al. 2007). Watersheds that may be the most sensitive to change are those occupying the boundary between rainfall and snow deposition in the winter (mixed regimes). For example, recent work in the Fraser River Basin (Figure 19.7) illustrates the 2050s changes in SWE projected by six different GCM emissions scenarios as a percentage difference from the 1961–1990 historical baseline period. These six sce-

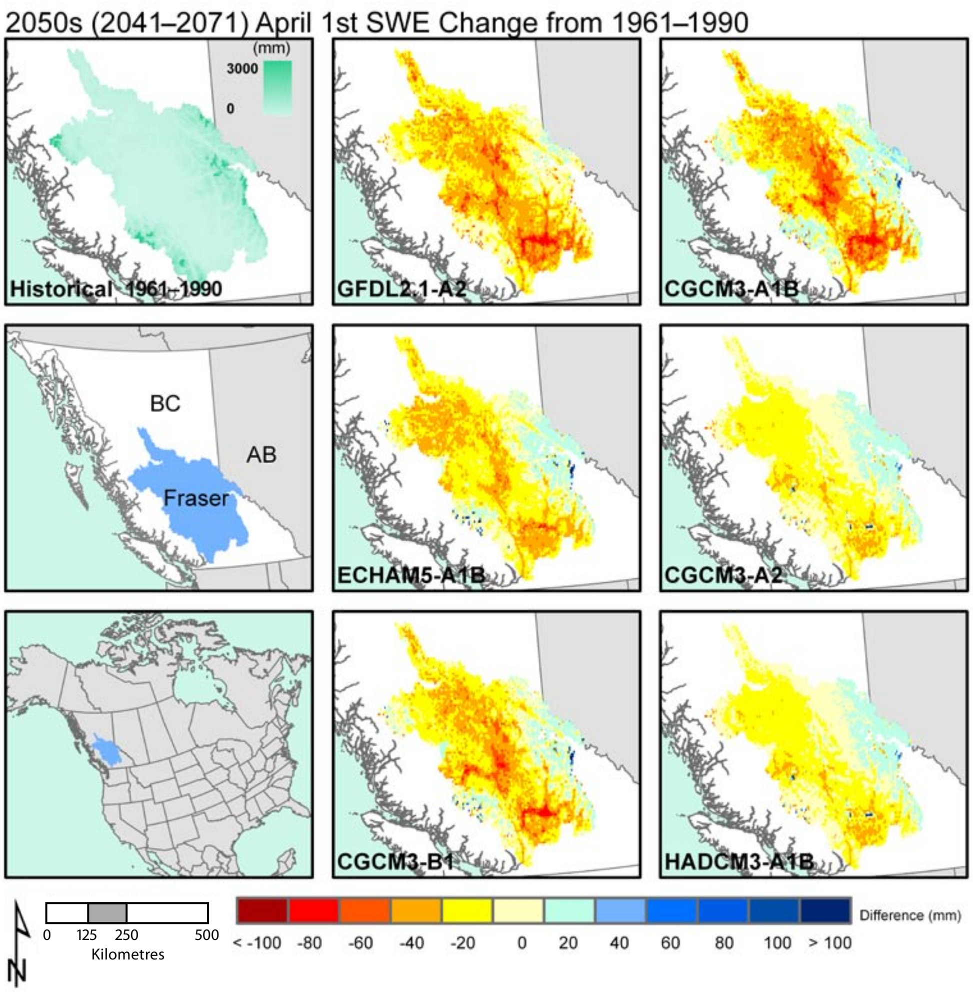


FIGURE 19.7 *Six GCM emissions scenarios projecting April 1st snow water equivalent (SWE) change to the 2050s in the Fraser River, British Columbia. Historical (1961–1990 April 1st) average SWE (mm) is illustrated in the top left-hand panel. The six scenarios are shown as 2050s (2041–2070) anomalies (mm) from the 1961–1990 baseline period on April 1st. For more information see "Case study: Fraser River Basin climate change projections," page 719.*

narios include: Geophysical Fluid Dynamics Laboratory, version 2.1 (GFDL2.1-A2); Canadian Centre for Climate Modelling and Analysis, Canadian Global Climate Model, version 3 (CGCM3-B1, -A1B, -A2); Max Planck Institute for Meteorology, European Centre Hamburg Model, version 5 (ECHAM5-A1B); and Hadley / United Kingdom Meteorological Office, Hadley Centre Coupled Model, version 3 (HADCM3-A1B).

Snowpack is projected to decline in the central plateau region of the basin and increase in the upper reaches of the Rocky Mountains and at high elevations in the Coast Mountain ranges (Figure 19.7). Some variation is evident in the spatial distribution of change but, on average, models project a 28% decline in SWE by the 2050s across the Fraser River Basin. Projected increases in precipitation will only slightly offset the changes resulting from increased temperature alone (Table 19.3); however, if a large portion of winter precipitation shifts to rain, the amount and timing of discharge will significantly change (see discussion in "Streamflow: Peaks, Lows, Timing," below). For nival regimes, especially in southern British Columbia, the warming trend may result in an earlier freshet, leading to lower flows in late summer and early autumn (Loukas et al. 2002; Merritt et al. 2006). Hydrologic scenarios for snowmelt-dominated basins in the Okanagan are projected to change in this way (Merritt et al. 2006); however, the degree of change projected depends on the GCM used. Simulations performed by this chapter's authors found that, on average, snow disappeared 16 days earlier under 2° C of warming and 37 days earlier under 4° C of warming in the Okanagan Plateau region (Figure 19.8). The length of the snow season was reduced, on average, by 25 and 60 days under 2° C and 4° C of warming, respectively. Changes in seasonal snow accumulation and melt will result in changes to the streamflow regime, which has important implications for water supply, hydroelectric power, and fish and aquatic habitat.

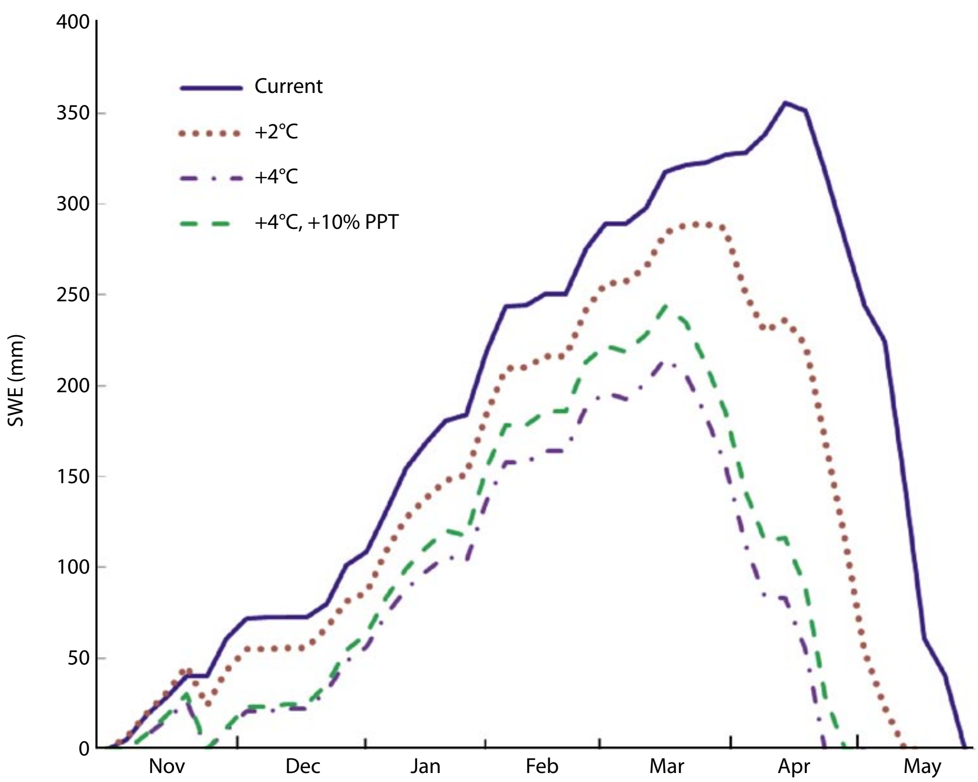


FIGURE 19.8 *Simulated winter snow water equivalent (SWE) in the mature lodgepole pine forest at the Upper Penticton Creek Experimental Watershed (1600 m elevation) under typical winter temperature and precipitation (2005–2006) conditions (solid blue line) and three climate change scenarios: (1) 2°C warming with no precipitation change (dotted red line); (2) 4°C warming with no change in precipitation (dashed and dotted purple line); and (3) 4°C warming with a 10% increase in precipitation (dashed green line).*

Less snow also has major implications for winter recreation and associated tourism activities (i.e. ski hills, Scott et al. 2006).

Permafrost, Lake Ice, and River Ice

Ice-related hydrologic features will be affected by rising temperatures. Projections of milder winter temperatures indicate that river and lake ice could occur later and disappear earlier than normal. These hydrologic changes will have implications for forest harvest scheduling (e.g., operable ground, seasonal water tables, timing), and transportation (e.g., ice bridges). In northern British Columbia, discontinuous permafrost is also expected to respond to temperature and precipitation changes. As with glaciers, all permafrost that exists today is not necessarily in equilibrium with the present climate. Unlike glaciers, however, adjustment to present climate lags on a longer time scale because of the insulating effects of the ground.

In the discontinuous permafrost region where ground temperatures are within 1–2° C of melting, permafrost will likely disappear as a result of ground thermal changes associated with global warming (Geological Survey of Canada 2006). In areas where the ice content is high, thawing permafrost can lead to increased thaw settlement and thermokarst activity, whereas reduced soil strength related to melting will lead to ground instability, increasing the incidence of slope failures (Smith and Burgess 2004). The integrity of engineered structures such as bridge footings, building foundations, roads, railways, and pipelines will also be affected (Woo et al. 2007).

Thawing permafrost may also affect aquatic ecosystems through changes to the storage and release of soil water (i.e., increasing the storage capacity) caused by the melting of ice. Under climate warming, the permanent thawing of permafrost may also add another source component to the hydrologic cycle. The overall thermal response of permafrost to increased temperatures will depend on the characteristics of the permafrost and surface buffer factors (e.g., snow, vegetation, and organic ground cover), which can attenuate temperature changes (Smith and Burgess 2004). The continuation of warming trends will likely increase the prevalence of thaw-related landslides in British Columbia and cause changes in soil water balances affecting the storage and release of water. About 50% of the Canadian permafrost region could ultimately disappear or become thinner in response to future climate warming (Smith et al. 2005).

Glacier Mass Balance Adjustments and Streamflow Response

Over the last few decades, the province's glaciers have dominantly had a negative mass balance (i.e., ablation of snow and ice exceeds the accumulation of snow) and continue to lose mass (Moore et al. 2009; Bolch et al. 2010; see also "Historical Trends in Snow, Seasonal Ice Cover, and Glaciers," above). Given future climate scenarios, glaciers will ultimately retreat under sustained conditions of negative net balance, although a lag is often associated with glacier dynamics (e.g., Arendt et al. 2002). Glacier retreat will continue until the glacier loses enough of its lower-elevation ablation zone that total ablation matches total accumulation. In some cases, climate warming can result in ablation exceeding accumulation over all elevations on a glacier, in which case the glacier would ultimately disappear. This will likely occur for glaciers in Montana's Glacier National Park and the North Cascades in Washington (Hall and Fagre 2003; Pelto 2006).

Projections based on future climate scenarios indicate that a negative net balance will continue over at least the next few decades. Hall and Fagre (2003) modelled glacier response to climate change in Montana's Glacier National Park under two climate scenarios. In the first scenario, with a doubling of CO_2 and a summer mean temperature increase of 3.3° C, all glaciers disappeared by 2030. In the second scenario, with a linear increase in temperature over time and a 0.47° C increase in summer mean temperature, glaciers remained until 2277. Stahl et al. (2008) used three scenarios to model the response of the Bridge Glacier in the southern Coast Mountains: one was a continuation of current climatic conditions until 2150, and two others were based on the A2 and B1 emissions scenarios developed by the Intergovernmental Panel on Climate Change (IPCC 2007) and simulated by the CGCM3. Even with no further climate warming, the Bridge Glacier is sufficiently out of equilibrium with current climatic conditions that it is projected to lose approximately 20% of its current area, reaching a new equilibrium by about 2100. Under the two warming scenarios investigated, glacier net balance remained negative and the glacier continued to retreat over the next century, with a projected loss of over 30% of its current area by the end of this century.

If glaciers are initially in equilibrium with current climatic conditions (i.e., snow accumulation balances ablation of snow and ice), then the onset of climatic warming will produce an initial increase in glacial melt and runoff contributions to streamflow (Hock et al. 2005; Moore et al. 2009). This increase in melt results from a lengthening of the melt season (i.e., advanced onset of melt in spring, delayed onset of accumulation in autumn) and an increase in melt intensity, particularly as firn (snow deposited in previous years) melts away, exposing the less reflective glacier ice to solar radiation. Eventually, however, the loss of glacier area will reduce total meltwater generation, resulting in a decrease in glacier runoff contributions to streamflow. Although this pattern of response to climate warming is generally accepted, the time scale over which the response shifts from increasing to decreasing discharge is not known.

Negative trends have been documented for summer streamflow in glacier-fed catchments in British Columbia, with the exception of the northwest, where streamflow has been increasing in glacier-fed catchments (Fleming and Clarke 2003; Stahl and Moore 2006). Similar negative trends have been documented for the late summer to early autumn "transition to base flow" period for glacier-fed headwaters draining the eastern slopes of the Canadian Rocky Mountains (Demuth and Pietroniro 2001; Comeau et al. 2009). Thus, it appears that the initial phase of streamflow increases associated with accelerated glacier melt has already passed for most of the province, whereas the northwest is still experiencing augmented streamflow. Stahl et al. (2008) found that future glacier retreat produced continuing declines in summer flows for Bridge River, particularly for July to September.

In addition to changes in streamflow, future glacier retreat will influence a range of aquatic habitat characteristics, including stream temperature, suspended sediment concentrations, and stream water chemistry. It is possible that the number and rates of geomorphic events or processes associated with glacial retreat will also increase in the future (see "Changes in Geomorphic Processes," below). Reduced glacier cover will also affect tourism and outdoor recreation activities in much of the province. Physical considerations and empirical evidence consistently indicate that summer stream temperatures should increase as a result of glacier retreat; however, the magnitude of this change is difficult to predict. Changes in other aspects of aquatic habitat will depend on a range of site-specific factors, and generalizations, even about the direction of change, cannot be made with confidence (Moore et al. 2009).

Altered Groundwater Storage and Recharge

The most direct interaction between climate and groundwater is through the process of recharge, which occurs when water from the ground surface (i.e., precipitation inputs, surface water bodies) has percolated to the water table. Recharge is the net result of energy and moisture transfer that occurs at the land surface, and is controlled by climate, vegetation, topography, soil characteristics and physical characteristics of the aquifer (i.e., geology). Thus, the recharge process will exhibit different degrees of sensitivity to the state of climate in a given region. Decreased recharge and persistent declines in groundwater storage can lead to a reduction in water supplies, degradation of water sources for groundwater-dependent ecosystems, land subsidence, increased conflicts between water users, and saltwater intrusion in coastal areas (Rivera et al. 2004). Groundwater discharge to streams is a critical factor governing low flows across much of British Columbia, with steeper mountain areas typically having little groundwater storage capacity (periodically leading to drying streams in the summer), and larger valleys with deeper alluvial sediments providing greater groundwater contributions through the low-flow period (Burn et al. 2008).

Changes in recharge fluxes will be influenced by several of the previously discussed processes, including increased atmospheric evaporative demand, changes in vegetation composition, snow accumulation and melt, and streamflow. Changes in the amount, timing, and form of precipitation (snow vs. rain) will all affect the rate and timing of groundwater recharge. Changes to streamflow will also affect groundwater recharge in locations where surface water is the main recharge source (i.e., alluvial valley-bottom aquifers recharged by streamflow during periods of high flow and discharge to streams during low flow periods). Depending on aquifer size and depth, any changes in groundwater hydrology caused by change in climate will likely occur more slowly than surface water changes.

The physical characteristics of aquifers have a strong influence on how groundwater systems respond to climatic changes. For example, shallow aquifers with highly permeable sediments (e.g., fractured bedrock or unconsolidated coarse sediments) are more responsive to climatic changes than deeper bedrock aquifers (Rivera et al. 2004). Deeper aquifers have a greater ability to buffer short-term perturbations; however, these aquifers will also preserve the signature of longer-term trends in climate change.

Relatively little research has been directed toward the effects of climate change on groundwater in British Columbia (Allen 2009) or elsewhere in the world (Dragoni and Sukhija [editors] 2008; GRAPHIC Team 2009). Provincial research consists of case studies of the Grand Forks Aquifer (Allen et al. 2003; Scibek and Allen 2006a, 2006b; Scibek et al. 2007), the Abbotsford-Sumas Aquifer (Scibek and Allen 2006a), the Okanagan Valley (Toews and Allen 2009; Toews et al. 2009), and the Gulf Islands (Appiah Adjei 2006). From these studies, the authors concluded that groundwater resources in the southern Interior are potentially the most sensitive to climate change in British Columbia. This is because of the strong influence of snow accumulation and melt on recharge and the potential changes in the magnitude and timing of nival processes under future climates (Allen 2009). Nevertheless, the estimated changes in groundwater storage and recharge are within the uncertainty range of the groundwater and GCM models (Allen 2009), which makes it difficult to identify or predict actual climate change influences (GRAPHIC Team 2009).

For the Grand Forks Aquifer, research combining GCM projections and groundwater models indicates that peak runoff in the Kettle River would occur earlier, and that the shift in peak streamflow would be accompanied by an earlier annual peak in groundwater levels. Away from the floodplain, groundwater recharge is predicted to increase in spring and summer months, and decrease in winter months (Scibek et al. 2007).

Research in the Okanagan Valley shows that direct (vertical) recharge along the valley bottom is driven largely by regional precipitation (e.g., frontal precipitation) rather than localized precipitation (e.g., convective storms) (Toews et al. 2009). When combined with future climate scenarios, peak recharge is expected to occur earlier in the year when evaporative demand is lower. The net effect is a minor increase in annual recharge for predicted future climate scenarios (Toews and Allen 2009), which could possibly buffer higher water demand in hotter and drier summer months in this region.

The potential effects of climate change on groundwater levels and recharge in the Lower Mainland and coastal regions generate fewer concerns. A modelling study of the Abottsford-Sumas Aquifer indicates only small absolute decreases in water levels, which are generally limited to upland areas (Scibek and Allen 2006a). However, lower groundwater levels will result in decreased base flow during low flow periods, which may have a negative influence on fish habitat. For aquifers on the Gulf Islands and in other coastal locations, concerns about decreasing recharge and declining water levels are related to the potential for saltwater intrusion (Rivera et al. 2004), a problem that may be compounded by rising sea levels (Allen 2009). In combination, a decrease in groundwater recharge and increase in sea level will cause the interface between seawater and fresh groundwater to move further inland, potentially increasing aquifer salinity to a point where its water is not fit for human consumption or use in irrigation.

Assessing the effects of climate change on groundwater is a difficult task because highly detailed subsurface information is required to develop quantitative models (Allen 2009). The aquifer classification system for the Canadian Cordillera (Wei et al. 2009) could be used as a starting point to identify the aquifer types with the greatest potential to be affected by climate change. Several successful attempts have been made to quantify areal or regional changes in groundwater storage using the Gravity Recovery and Climate Experiment (GRACE). Essentially, GRACE satellites record changes in Earth's gravity field that are then related to changes in terrestrial water storage (Rodell and Famiglietti 2002). This technique has been applied on the Canadian Prairies (Yirdaw et al. 2008) and the Mackenzie River basin (Yirdaw et al. 2009). Although these applications of GRACE data are encouraging, further research and testing is required to determine whether the method is appropriate for the complex physiography and geology of British Columbia and whether the impact of climate variation on groundwater can be determined at a resolution useful for water management.

Streamflow: Peaks, Lows, and Timing

Streamflow regimes are controlled primarily by seasonal patterns of temperature and precipitation, as well as watershed characteristics such as glacier cover, lake cover, and geology. In British Columbia, the four main hydrologic regimes are: (1) rain-dominated, (2) snowmelt-dominated, (3) mixed/hybrid, and (4) glacier-augmented (see Chapter 4, "Regional Hydrology"). The relative importance of climatic changes, therefore, will vary by region and depend on the current sensitivity of the hydrologic regime to regional temperature and precipitation changes. Also, groundwater storage and release strongly control streamflow (particularly low flows) in some watersheds (see discussion above). Variations in underlying geology that influence whether snowmelt goes into groundwater reserves or directly into runoff can determine the magnitude and timing of late-summer streamflow, and thus affect the overall response to climate change (Thompson 2007; Tague and Grant 2009).

The hydrologic effects of climate change will have an important influence on all types of watersheds, not just those with cold-season precipitation storage as snowpack. The response of rain-dominated regimes will likely follow predicted changes in precipitation (Loukas et al. 2002). For example, increased magnitude and more numerous storm events will result in increasingly frequent and larger storm-driven streamflow (including peaks) in the winter. Projected warmer and drier summers also raise concerns about a possible increase in the number and magnitude of low flow days.

Projected warming will result in less snow stored over winter (Figures 19.7 and 19.8) and more winter precipitation falling as rain. In these situations, hybrid/mixed regimes might transition to rain-dominated regimes through the weakening or elimination of the snowmelt component (Whitfield et al. 2002). Similarly, snowmelt-dominated watersheds might exhibit characteristics of hybrid regimes and glacier-augmented systems might shift to a more snowmelt-dominated pattern in the timing and magnitude of annual peak flows and low flows. For example, in the southern Columbia Mountains at Redfish Creek, an increase is evident in the incidence of fall to early-winter peak streamflow events, which up to 10 years ago were relatively rare in the hydrometric record (P. Jordan, Research Geomorphologist, B.C. Ministry of Forests and Range, pers. comm., Dec. 2007). With projected elevated temperatures, the snow accumulation season will shorten (Figure 19.8) and an earlier start to the spring freshet in snowmelt-dominated systems will likely occur, which may lengthen the period of late-summer and early-autumn low flows (Loukas et al. 2002; Merritt et al. 2006). Where snow is the primary source of a watershed's summer streamflow, loss of winter snowpack may reduce the late-summer drainage network, transforming once perennial streams into intermittent streams (Thompson 2007). Conversely, where groundwater is the primary source of a watershed's summer streamflow, flows will still continue but with volume reductions in response to changes in the seasonal snowpack accumulation that recharges groundwater (Thompson 2007).

In glacier-augmented systems, peak flows would decrease and occur earlier in the year, similar to snowmelt-dominated regimes. In the long term, the reduction or elimination of the glacial meltwater component in summer to early fall would increase the frequency and duration of low flow days in these systems.

In hybrid regime watersheds on the Coast, some snowpacks above 1000–1200 m can be up to 4–5 m deep (e.g., Russell Creek), especially in north-facing open bowls or subalpine forests (B. Floyd, Research Hydrologist, B.C. Ministry of Forests and Range, pers. comm., 2007). Normally, snowpacks in these hybrid regimes are deep enough to store a significant amount of rain, thus dampening the response of watersheds to large midwinter rain events. If these snowpacks no longer form or are very shallow, and increases in temperature and wind speeds occur, large midwinter snowfall events will become large rain or melt events, and thereby increase the frequency of high flows occurring throughout the winter in these watersheds. Subsequently, spring peak flow volumes will decrease and occur earlier because less precipitation is stored as snow during the winter, and winter flows will increase because precipitation will fall as rain instead of snow.

For all streamflow regimes, a complex relationship will likely develop between rain-on-snow events and changes in regional air temperature and precipitation patterns. This is because the magnitude of rain-on-snow floods fluctuates depending on the duration and magnitude of precipitation, the extent and water equivalent of the antecedent snowpack, and the variations in freezing levels (McCabe et al. 2007). Climatic changes will influence all of these factors. For example, McCabe et al.'s (2007) modelling study showed that as temperatures increase, rain-on-snow events decrease in frequency primarily at low-elevation sites. Higher elevations are likely less sensitive to changes in temperature as these sites remain at or below freezing levels in spite of any temperature increase that would affect snow accumulation (McCabe et al. 2007).

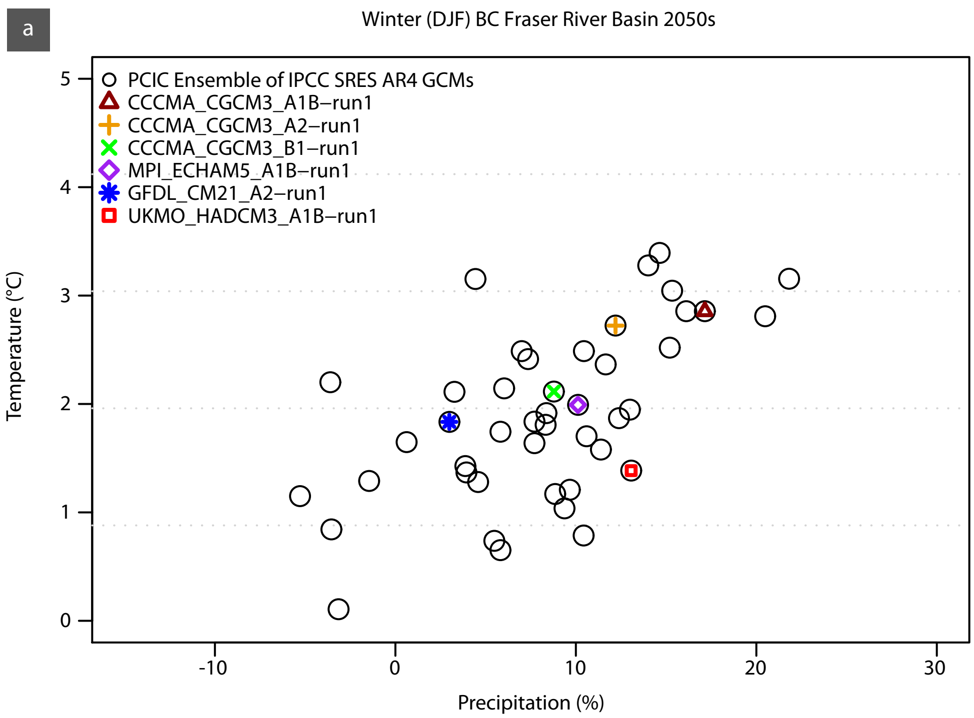
Case study: Fraser River Basin climate change projections

The Fraser is one of British Columbia's largest rivers, and one of the most productive salmon rivers in the world. Approximately two-thirds of B.C.'s population resides in the basin, and 80% of the provincial economy is generated within the basin. Because of its importance to the residents of British Columbia, concerns have been raised over the effect of future climatic changes in the basin. To address some of these concerns, the Fraser River Basin was modelled using the semi-distributed, Variable Infiltration Capacity (VIC) hydrologic model (Liang et al. 1994, 1996; Schnorbus et al. 2009). The aim of this computer modelling was to examine the impacts of climate change on basin hydrology for the 2050s (i.e., the period of 2041–2070 from the baseline period of 1961–1990). Although the hydrologic impacts of climate change in British Columbia have been examined (e.g., Slaymaker 1990; Brugman et al. 1997; Whitfield and Taylor 1998; Loukas et al. 2002; Whitfield et al. 2002; Merritt et al. 2006; Toth et al. 2006), only three studies have specifically presented results for the Fraser River Basin (i.e., Moore 1991; Coulson 1997; Morrison et al. 2002).

Creating projections of future streamflow and snowpack conditions across the entire Fraser River Basin is valuable for several reasons. For example, the distributed hydrologic model produces simulations at various spatial and temporal scales. To estimate projected streamflow responses for specific watersheds, the model "forces" future simulations with temperature and precipitation downscaled from gridded GCMs. By examining a suite of GCMs and emissions scenarios, it is possible to analyze a range of potential futures for the Fraser River Basin. This approach to modelling provides practitioners with valuable information to support planning initiatives and to develop suitable adaptation plans for the Fraser River Basin.

The work we present here applies the bias-corrected spatial downscaling technique to estimate future temperature and precipitation change for six GCM emissions scenarios selected from the IPCC's fourth assessment report database (see Intergovernmental Panel on Climate Change 2007). This particular subset of the GCMs performs well across North America when compared to historical data (Plummer et al. 2006; Salathé et al. 2007; Gleckler et al. 2008). The six scenarios described in "Snow Accumulation and Melt," (above) were chosen to represent a wide range in future conditions, from warm-wet to cool-dry climates occurring in the Fraser River watershed. Downscaling measures included bias-correction of the monthly "coarse"-resolution GCM emissions scenarios model output to match the spatial and temporal resolution of the VIC hydrologic model (based on methods described by Wood et al. 2002; Widmann et al. 2003; Salathé 2005; see also "Downscaling for Watershed Modelling," below). These downscaled forcings were used to run the VIC model out to the year 2100 across the entire 225 000 km^2 Fraser Basin above Hope, at a grid-scale resolution of approximately 32 km^2 (as described in Schnorbus et al. 2009). This updates previous work by employing the latest GCMs as well as a statistical downscaling technique that produces a bias-corrected transient simulation on a monthly basis for the entire distribution, gridded to the scale of the hydrologic model resolution (32 km^2). For the historical baseline period, the model performance was 0.89 for the calibration period (1985–1990) and 0.82 for the validation period (1991–1995) based on Nash-Sutcliff model efficiency (Nash and Sutcliffe 1970).

Future projected changes for the Fraser River Basin by the 2050s include an increase in median annual precipitation of 5% and potential increase in median annual air temperature of 2° C. Figure 19.9 presents winter and summer scatterplots for all IPCC SRES AR4 GCMs along with the selected GCM and emissions scenarios for precipitation against temperature as projected anomalies from the historical baseline. Figure 19.10 shows the projected increase in annual mean air temperature by the 2050s, an increase that is observed across all six scenarios. Across most models, the warming is greatest in the Thompson-Nicola region in the southeastern part of the basin. The median summer (June–August) air temperature is projected to increase by 3–4° C. Although the southern portion of the basin will warm faster than the north in the summer, the strongest winter warming is projected for the northern region of the basin (e.g., in the Stuart River watershed above Fort St. James, 2.6° C). Figure 19.11 illustrates the projected increase in annual median precipitation across the basin, although some scenarios show a small decrease in the southern part of the basin, including the Chilcotin Plateau and the Thompson-Nicola region. Most model projections illustrate an increasing gradient of precipitation to the northeast of the basin, with the least amount of precipitation change projected for the southwestern portion of the Fraser along the Coast Mountain ranges (8%). Summer precipitation change is projected to decrease in most scenarios, although the ECHAM5-A1B scenario projects a wetter summer in the northern watersheds near Quesnel.



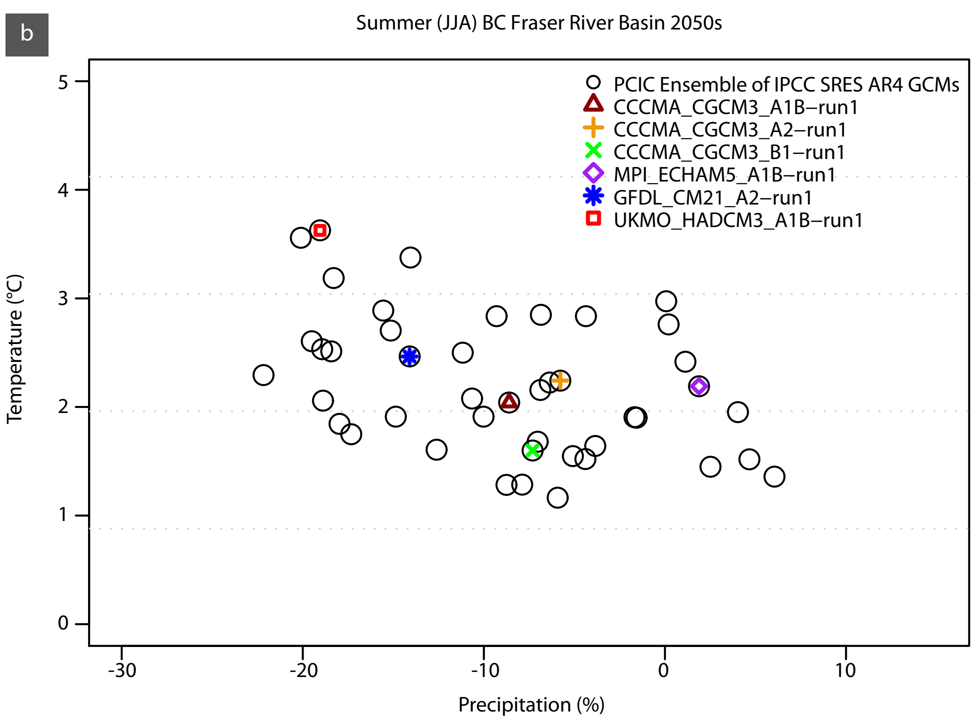


FIGURE 19.9 *Scatterplots of (a) winter and (b) summer precipitation versus air temperature projections for the Fraser River Basin provided by the six GCM emissions scenarios. The modelling centre is identified in the legend followed by the GCM name/version: CCCMA – Canadian Climate Centre for Modelling and Analysis; MPI – Max Planck Institute; GFDL – Geophysical Fluid Dynamics Laboratory; UKMO – United Kingdom Meteorological Office.*

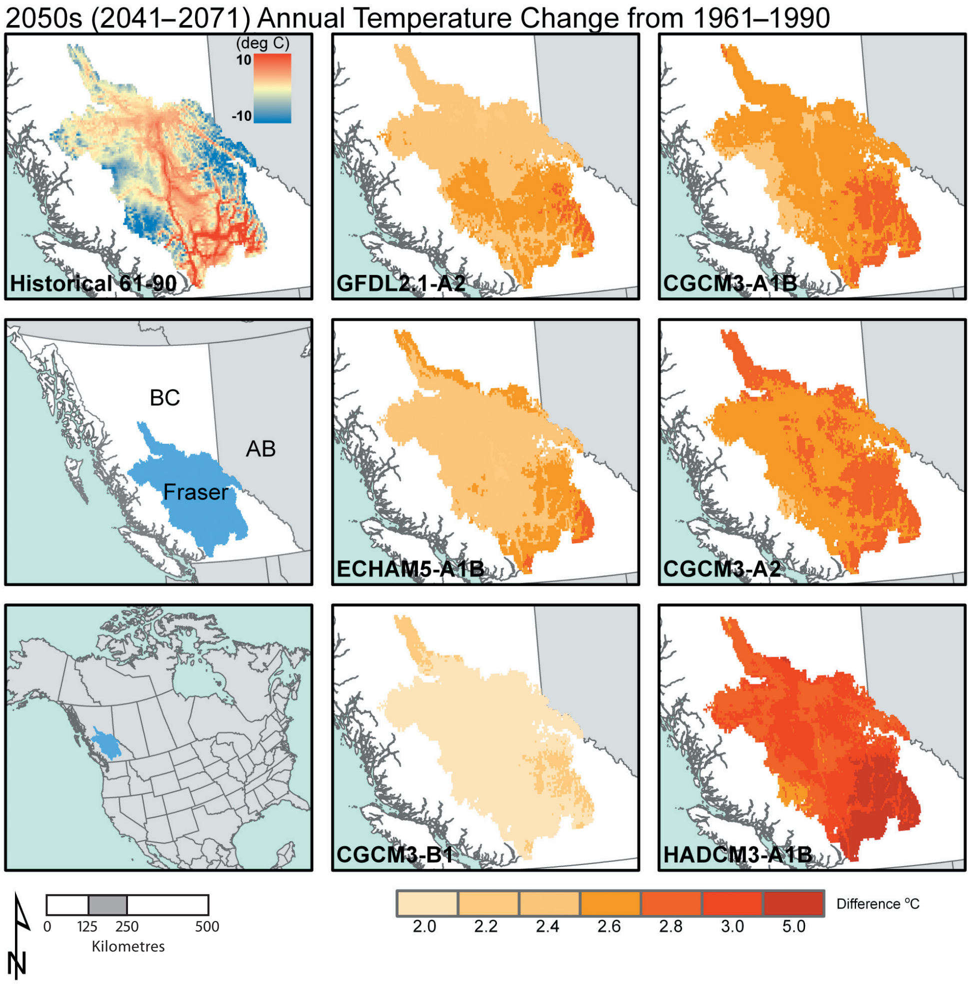


FIGURE 19.10 *Six GCM emissions scenarios projecting annual air temperature changes to the 2050s in the Fraser River Basin. Historical temperatures are illustrated in the top left-hand panel. The six scenarios are shown as degree Celsius anomalies from the 1961–1990 baseline period.*

For the most part, these projections agree with findings from previous work but with some differences. Morrison et al. (2002) projected an increase in mean annual air temperature of 1.5° C with the CGCM1 doubled CO_2 simulation, which is similar but lower than the CGCM3-B1 emissions scenario projection we present here for the Fraser Basin. Moore (1991) based his analysis on GCM projections provided in Slaymaker (1990), which selected "boundary" model results, estimating that temperatures would increase by 2.4–6° C in the winter and 0.6–4.2° C in summer. Moore's (1991) precipitation projections ranged from no change at all to increases of 15% and 20%, which is similar to the wet ECHAM5-A1B and CGCM3-A1B GCM emissions scenarios presented here. Coulson's (1997) projection for a 9% increase in precipitation for Prince George is also within the range represented by the six scenarios analyzed here. Morrison et al. (2002) projected approximately a 15 % increase (decrease) in winter (summer) precipitation at Kamloops, which is similar to the winter projections provided here; however, the summer decrease may be too extreme, based on the downscaled projections analyzed in this study. Differences in the projections may be partly attributable to the GCM versions applied in previous studies (e.g., CGCM3 vs. CGCM1), and the use of new transient emissions scenarios (SRES vs. doubled CO_2) applied in this case study.

Basin-wide annual runoff projections for the 2050s range from −20% in some watersheds to +35% by the CGCM3-A1B scenario (Figure 19.12). The drier scenarios (i.e., GFDL2.1-A2 and HADCM3-A1B) project

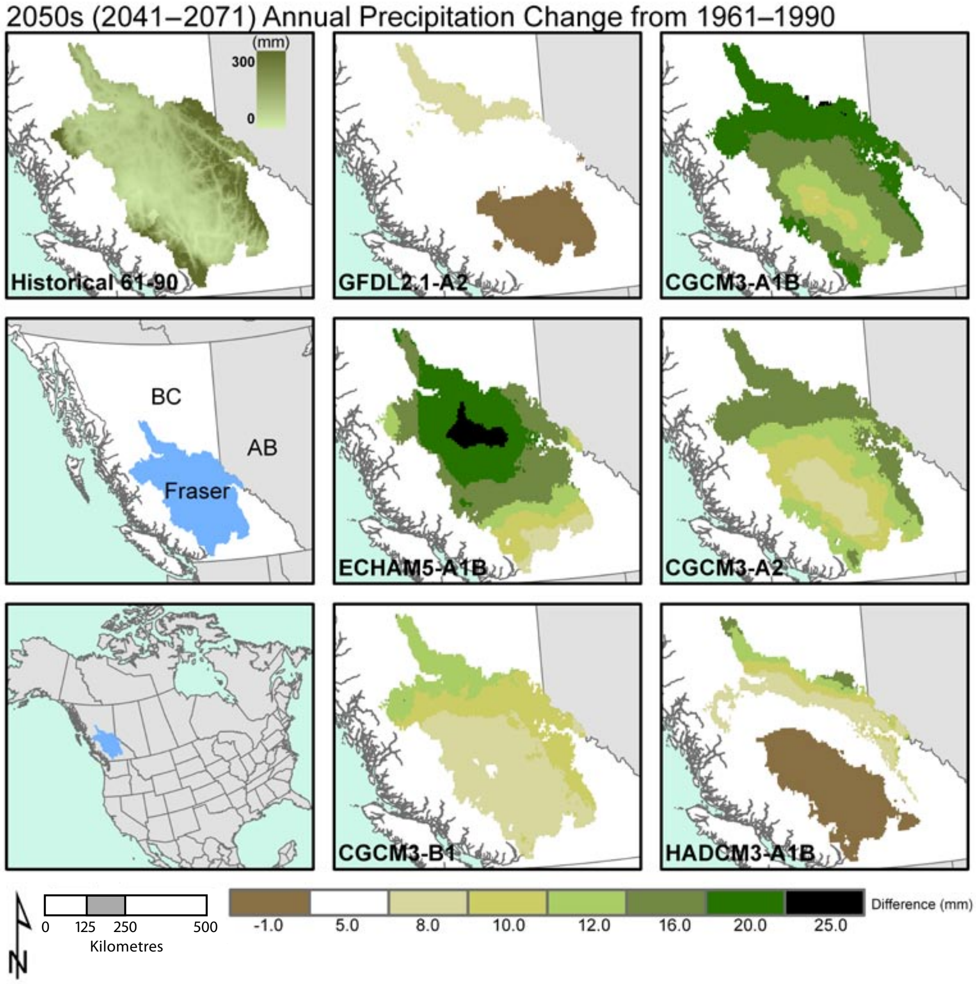


FIGURE 19.11 *Six GCM emissions scenarios projecting annual average precipitation changes to the 2050s in the Fraser River Basin. Historical precipitation (mm) is illustrated in the top left-hand panel. The six scenarios are shown as percentage differences from the 1961–1990 baseline period.*

a decline in annual runoff through the southern plateau regions of the Basin, in the Cariboo-Chilcotin region (including West Road River and Baker Creek), and into the lower reaches of the Thompson River watershed. Notably, most of the scenarios project a positive runoff condition for the future in the northern reaches of the watershed above Prince George, which reflects the projected 16% increases in runoff estimated by Colson (1997). A 14% increase in runoff is projected for the Fraser River at Hope, although projections range from almost no change in runoff (1%, GFDL2.1-A1B) to a larger increase (23%, CGCM-A1B), with a standard deviation between scenarios of 8%. On a seasonal basis, flows for the Fraser at Hope are projected to increase by approximately 1500 m^3/s in the spring, and decrease by approximately 1400 m^3/s in the summer on average (Figures 19.13 and 19.14).

Most scenarios project winter runoff increases, but some scenarios (i.e., GFDL2.1-A2, HADCM3-A1B) project a drier winter for the headwater basins of the Cariboo-Chilcotin region (Figure 19.15). Decreases in runoff for these models correspond to moderate increases or slight decreases in annual precipitation and declines in fall soil moisture (results not shown). The median winter 2050 projection for the Fraser Basin illustrates large increases in runoff (100% or greater) for mid-elevation reaches along the Rocky Mountain headwater regions, as opposed to the Coast Mountains, where slight decreases are projected by the 2050s. Increases in runoff correspond to a 25% increase in the winter precipitation for

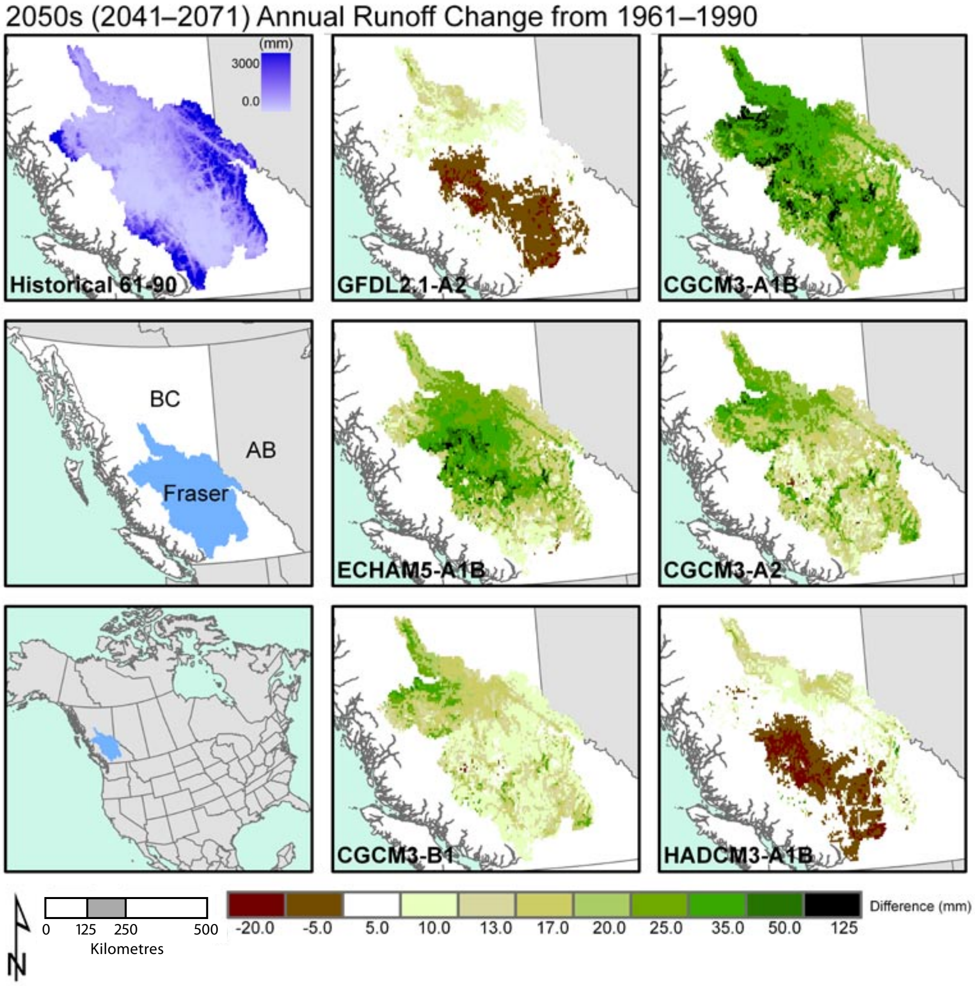


FIGURE 19.12 *Six GCM emissions scenarios projecting annual average runoff changes to the 2050s for the Fraser River Basin. Historical runoff (mm) is illustrated in the top left-hand panel. The six scenarios are shown as percentage differences from the 1961–1990 baseline period.*

the Rocky Mountains. The median summer runoff projection (Figure 19.14) is drier for most areas of the basin, especially in the watersheds of the Quesnel, McGregor, Salmon, and South Thompson Rivers. Projections for the lower reaches of the Thompson River watershed appear to have virtually no change in runoff for almost every scenario (Figure 19.14).

These runoff projections corroborate with Morrison et al.'s (2002) results with some important deviations. The Morrison et al. study (2002) indicated that the change in peak flow is projected to decline into the future, whereas the modelled flows presented here are projected to increase. This important difference may be caused by the higher precipitation amounts (particularly in the spring) projected by the more recent, transient SRES emissions scenarios. Additionally, the Morrison et al. study may have underestimated precipitation distributions across the high-elevation regions of the Fraser Basin, whereas the gridded downscaling approach of the VIC model allows for a more accurate analysis of high-elevation regions and shows these areas as receiving increased amounts of precipitation (still falling as snow) by the 2050s. This can be seen in the April 1st SWE maps where high-elevation sites in the Rocky and Coast Mountain ranges experience slight increases in snowpack (see Figure 19.7). However, the different tools and modelling approaches used in each study prevents a definitive explanation of why these model projections are so divergent.

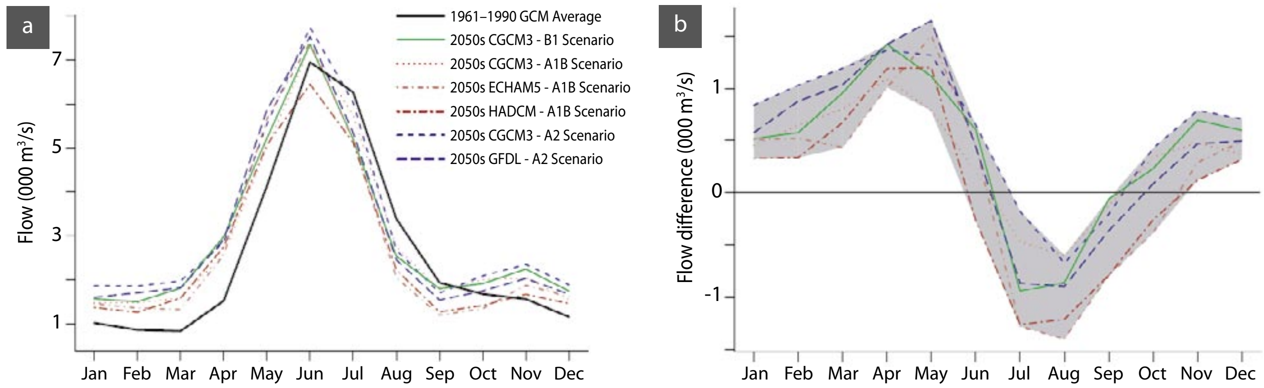


FIGURE 19.13 *Fraser River streamflow: (a) future projections of Fraser River streamflow at Hope, with the 1961–1990 baseline period (average of all GCMs) depicted by a black line; (b) differences in streamflow from the baseline period with the range in GCM emissions scenarios shown in grey to illustrate the variation across the scenarios.*

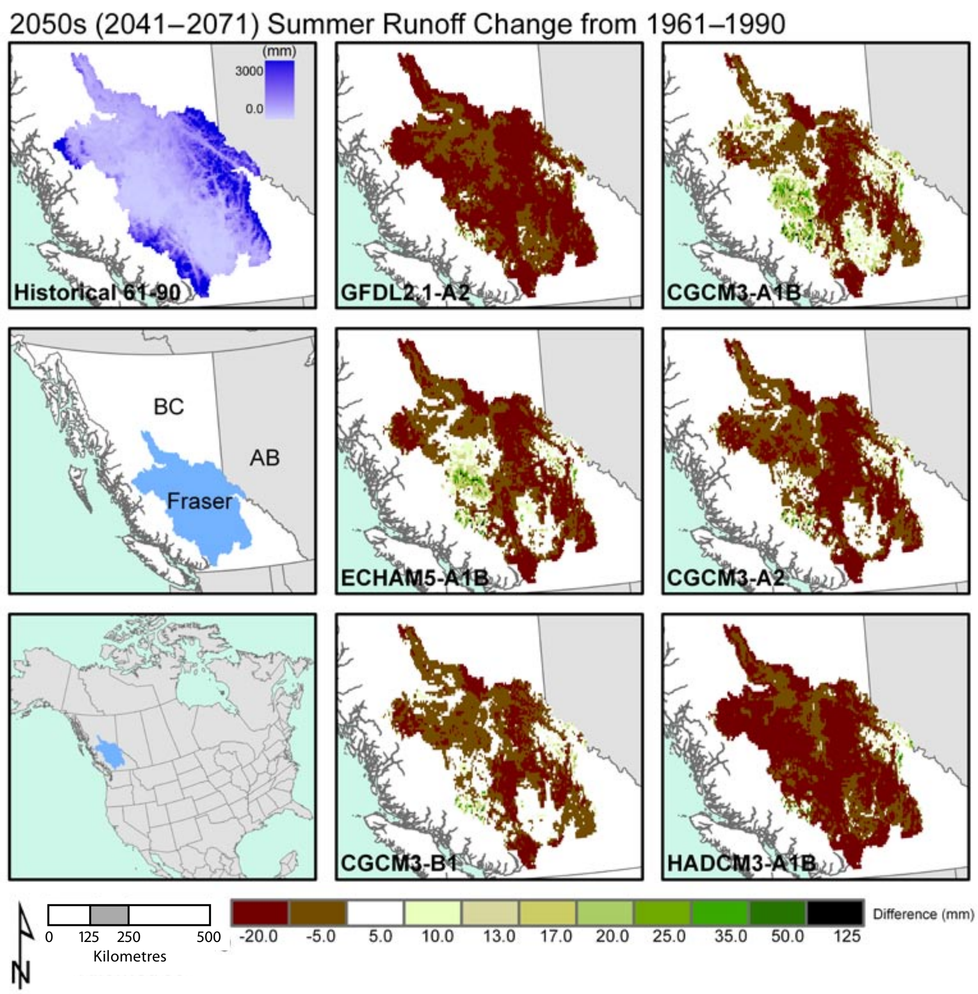


FIGURE 19.14 *Six GCM emissions scenarios projecting summer (JJA) runoff changes to the 2050s for the Fraser River Basin. Historical runoff (mm) is illustrated in the top left-hand panel. The six scenarios are shown as percentage differences from the 1961–1990 baseline period.*

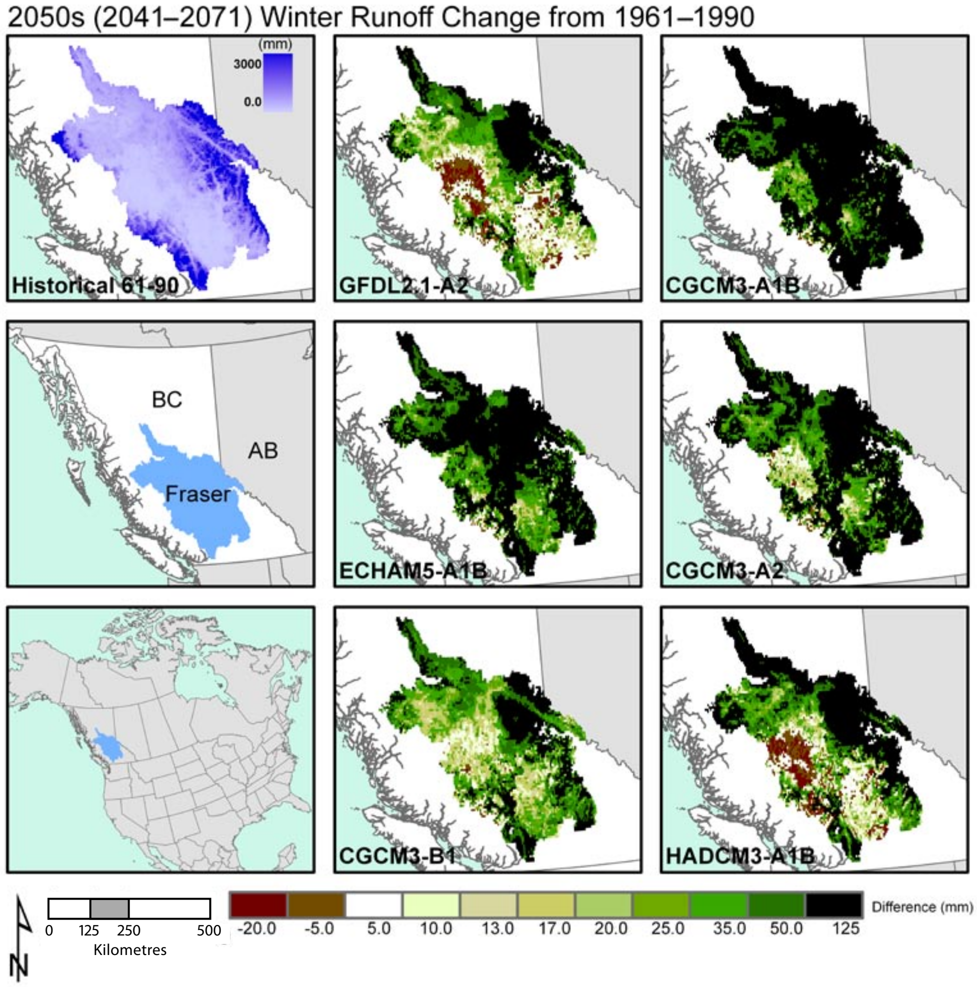


FIGURE 19.15 *Six GCM emissions scenarios projecting winter (DJF) runoff changes to the 2050s for the Fraser River Basin. Historical runoff (mm) is illustrated in the top left-hand panel. The six scenarios are shown as percentage differences from the 1961–1990 baseline period.*

Changes in Geomorphic Processes

Landslides in British Columbia are driven by climate, topography, geology, and vegetation. Landslide response to climatic changes will vary depending on the type of landslide and the initiation process (Geertsema et al. 2007; and see Chapter 8, "Hillslope Processes," and Chapter 9, "Forest Management Effects on Hillslope Processes"). Future changes in geomorphic processes will be driven primarily through changes in precipitation and temperature regimes. Recent trends, as detailed above in "Historical Trends in Landslides and Other Geomorphic Processes," are expected to continue.

In northern British Columbia, shallow slides and debris flows happen during infrequent large storms; large rock slides appear to respond to warming and may be triggered during convective storms; and larger soil slides are more common during periods of increasing precipitation (Egginton et al. 2007; Geertsema et al. 2007; and "Historical Trends in Landslides and Other Geomorphic Processes," above). Long-term increases in temperature and precipitation may be preconditioning slopes to fail, whereas intense or large-scale storms may also be triggers of such failures (Egginton 2005). Both scenarios are expected to increase with future climatic changes.

In coastal British Columbia, debris slide and debris flow initiation typically occurs during high-intensity precipitation events, often augmented with additional input from snowmelt, which occurs during fall or winter frontal storm systems. Predictions of the influence of projected climate changes to precipitation have typically focussed on average precipitation and long-duration conditions rather than extreme or short-duration events. As such, regional predictions of changes in precipitation intensity–duration relationships remain a significant knowledge gap in British Columbia, particularly for durations shorter than 24 hours. Landslide response to climate change in these areas will largely follow the projected peak flow response in rain-dominated and hybrid streams. In the Georgia Basin, for example, relationships between annual precipitation and short-duration precipitation intensity were examined by Miles as a predictive approach to estimating changes in storm frequency.[11] For 24-hour rainfall events sufficiently large enough to initiate slope failures, the reported 10% increase in annual precipitation over 80 years could lead to a decrease in storm return periods from 10.4 to 6.3 years.[12] Similarly, Jakob and Lambert (2009) correlated GCM modelling with antecedent precipitation and short-duration rainfall observations to evaluate projected changes in landslide initiation in southwest British Columbia. They estimated that a 6–10% increase in antecedent and short-duration precipitation amounts by the 2071–2100 period could lead to expected increases in landslide initiation of 28%.

Ongoing glacial recession will continue to promote periglacial processes in recently deglaciated areas. This includes increased geomorphic hazards such as outburst flooding, rock debuttressing, slope failures on over-steepened slopes, changes to sediment production, and suspended sediment fluxes (Moore et al. 2009).

Snow avalanche activity will likely also be affected through various processes that are forecast to change; however, the overall implications are likely complex and variable. Increased storm intensities during the winter may lead to increased avalanche activity. Countering this process will be warmer-than-present winter temperatures which, in general, will result in lower temperature gradients within snowpacks, and therefore increased slope stability. This may have a more pronounced effect for Interior ranges and northern British Columbia, which currently have very cold winters and typically strong snowpack temperature gradients. In some areas, the winter snow line may migrate high enough so that lower-elevation areas do not exceed threshold snow depths sufficient to initiate avalanches. This upward migration of the snow line, and encroachment of vegetation into avalanche paths, may lead to a corresponding upslope shift in avalanche runout zones. This process is most likely to be pronounced in coastal British Columbia, and particularly at or near the current tree line.

Changes in the timing and amounts of streamflow and cumulative watershed conditions will likely influence stream channel morphology and riparian function. Increased frequency of channel-forming peak flows is most likely in rain-dominated and hybrid systems. This could lead to channel instabilities, particularly in alluvial stream channels (e.g., Millar 2005). Changes to the return period of flood events also will have implications for engineering design criteria. In mountainous headwater stream systems, hillslope processes are coupled to stream channel processes such that changes in sediment delivery will affect sediment transport, channel morphology, and aquatic ecology (Benda et al. 2005). Similarly, changes in channel stability (i.e., bank erosion), windthrow, or landslides will likely affect supply and function of large woody debris (LWD) in streams (Hassan et al. 2005).

Potential changes in disturbance patterns at the watershed or landscape scales can also influence cumulative watershed effects. With warmer and drier summers projected for parts of British Columbia, fire seasons in these areas are expected to become longer with increased total area burned in each fire season (Flannigan et al. 2002, 2005). Wildfires can lead to widespread and severe surface erosion, debris flows, and flooding within watersheds (Curran et al. 2006). Severe impacts on stream channel morphology have been observed in response to changes in peak flow regime or increased sediment supply (Wondzell and King 2003), or related to loss of bank strength (Eaton et al. 2010). With increased fire activity, increases in erosion and flood processes can also be expected.

Widespread forest disturbances, such as insect infestations or disease, can also affect watershed processes. For example, changes in forest canopy structure in stands affected by the mountain pine beetle have resulted in changes to site-level hydrology. Across larger areas, this could lead to increased flood frequency–magnitude relationships (Hélie et al. 2005; Uunila et al. 2006). Increased frequency of flood events can influence channel morphology.

11 Miles, M. 2001. Effects of climate change on the frequency of slope instabilities in the Georgia Basin, B.C.: Phase 1. Natural Resources Canada, Canadian Climate Action Fund, Ottawa, Ont. Can. Climate Action Fund Proj. No. A160. Unpubl. report.

12 Ibid.

For example, Grainger and Bates (2010) examined increased flood risk attributed to the mountain pine beetle infestation and subsequent salvage harvesting in Chase Creek, and found that flood frequency increased by approximately 2.5 times. These changes resulted in significant channel changes and increased risks to private property and public infrastructure. Widespread tree mortality within riparian zones can affect the delivery of LWD to streams, riparian function, and instream dynamics of LWD (Everest and Reeves 2007). Riparian response to widespread forest disturbance, however, can be complex. For example, in beetle-affected stands near Vanderhoof, Rex et al. (2009) found that the dominance of unaffected spruce in the riparian areas of pine forests allowed for the maintenance of riparian function despite widespread pine mortality. Climate change is expected to affect ecological disturbance processes such as disease and insect outbreaks (Campbell et al. 2009), and therefore may also affect related riparian processes.

While fluvial geomorphic processes and disturbances are important for the renewal and diversity of fish habitat, altered rates and magnitudes of watershed processes above normal levels will have other implications for stream ecology and fish populations. Disturbances directly connected to stream channels, such as landslides and debris flows, can reduce the quantity and quality of fish habitats for several years or decades, and consequently the local abundance of salmon populations in affected stream reaches (Hartman and Scrivener 1990; Tschaplinski et al. 2004). Additionally, related processes such as local streambed scour can isolate the main stream channel from important seasonal fish habitats and refuges located in the floodplain, thus potentially reducing salmon survival and annual smolt production (Hartman and Scrivener 1990; Tschaplinski et al. 2004).

Changes in Water Quality

A considerable amount of research has focussed on the potential effects of climate change on water supplies; however, relatively little is known about the related effects on chemical water quality. Recent IPCC publications provided only cursory details on the effects of climate change on water quality (Kundzewicz et al. 2007; Bates et al. 2008). Limited predictions in this area may be partly related to the challenge of separating the potential effects of climate change on water quality from those of land and water use on surface and ground waters. Nevertheless, interest in this topic is growing (Whitehead et al. 2009).

The effects of climate change on chemical water quality are likely complex and will vary with the physical, geographical, and biological characteristics of each watershed. Changes in climatic conditions have the potential to either mitigate or worsen existing water quality issues, especially when combined with the effects of natural resource use (Dale 1997). The most important factors that influence the effects of climate change on water quality are increases in atmospheric and water temperatures and changes in the timing and amount of streamflow.

Changes in stream or lake temperatures and effects on fish

Climate change has the potential for both direct and indirect effects on stream temperature. Most directly, the energy exchanges that govern stream temperature may change. Solar radiation, generally the dominant driver of daily maximum temperatures, depends on the Sun's position in the sky and the transmissivity of the atmosphere (a function of humidity, cloud cover, dust content, and other factors), and is therefore not directly related to air temperature; however, incident solar radiation will be influenced by any changes in cloudiness that accompany climate change. Incident longwave radiation, which acts to suppress nighttime cooling, increases with increasing air temperature and also with increasing cloud cover, and thus should be influenced by climate warming. Groundwater 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 1997; Bogan et al. 2003). Deep groundwater temperatures tend to be within about 3° C of mean annual air temperature (Todd 1980). It is reasonable, therefore, to assume that climate-induced groundwater warming will influence stream temperature regimes, particularly during base-flow periods when groundwater is a dominant contributor to streamflow and especially when energy inputs at the stream surface are relatively minor (e.g., at night).

Projected hydrologic changes in some areas may produce lower streamflow in late summer, and also less groundwater discharge. Both of these influences could promote higher late-summer water temperatures. Similarly, reductions in late-summer streamflow associated with glacier retreat are expected to result in higher stream temperatures (Moore et al. 2009).

Less directly, climate change may result in changes to vegetation and (or) land use patterns, which could influence stream shading and possibly channel morphology. For example, it is generally accepted that the area burned by wildfires will increase in some areas under a future warming climate (Flannigan et al. 2005). Debris flows often increase in frequency following wildfire, and can generate increased stream temperatures by producing wider channels (which reduces shading) and removing substrate. This decreases the potential for hyporheic exchange, which can moderate stream temperatures (Johnson 2004). Where wildfires burn through riparian zones, the reduction in canopy shade can produce higher stream temperatures caused by increased insolation (Leach and Moore 2010) and also cause channel widening due to loss of bank strength (Eaton et al. 2010). Dunham et al. (2007), working in the Boise River basin in Idaho, found that streams in undisturbed catchments were cooler than streams subject to riparian wildfire, which in turn were cooler than streams that experienced channel disturbance in addition to riparian wildfire.

Other indirect influences may occur through human-induced changes in drainage patterns to address changing patterns of water availability and scarcity. For example, withdrawals of water for irrigation or other uses typically cause increased stream temperatures (e.g., Hockey et al. 1982), whereas the effects of impoundments are more complex, depending on the depth of the reservoir and the depth from which downstream flow releases originate (e.g., Webb and Walling 1997).

Most attempts to evaluate potential stream temperature responses to climate change have used the statistical relationship between stream temperature and air temperature to assess sensitivity related to the projected changes in air temperature derived from GCM output (e.g., Eaton and Scheller 1996; Mohseni et al. 1999; Morrill et al. 2005). Morrison et al. (2002) conducted a more comprehensive assessment for the Fraser River. They used a conceptual model of catchment hydrology (the University of British Columbia Watershed Model), in conjunction with projections of future temperature and precipitation, to generate scenarios for streamflow for sub-basins of the Fraser River. They then used these climate and streamflow projections, together with a model of energy exchanges and water flow in the Fraser River stream network, to simulate stream temperatures. The scenarios suggest an increase in the spatial and temporal frequency of temperatures exceeding 20° C, particularly below the confluence with the Thompson River.

Stream and lake temperatures are projected to increase with climate change, which will result in several specific concerns for aquatic and fish species including salmon (Levy 1992; Mote et al. 2003). Increased water temperatures could affect metabolic rates and increase biological activity and decomposition. In aquatic systems with sufficient nutrient and oxygen supplies, an increase in biological productivity can increase nutrient cycling and possibly accelerate eutrophication (Murdoch et al. 2000). However, it is likely that in aquatic systems currently stressed by high biological oxygen demand any subsequent increase in water temperatures could decrease biological productivity as a result of a decline in the oxygen-holding capacity of the water.

The vulnerability of fish to climate change will partly depend on how much the water body warms and the sensitivity of individual fish species to temperature and habitat changes. Temperature-related risks for fish include both acute (short-term) and chronic (also termed "sublethal" or "cumulative") effects.[13] The vulnerability of fish may depend on local-scale watershed management strategies, which have the potential to exacerbate or mitigate the effects of climate change. For example, research on the Little Campbell River (a tributary entering Boundary Bay, about 35 km south of Vancouver) concluded that watershed remediation or degradation can greatly affect the ultimate impacts of climatic change on chronic thermal risks to fish.[14]

Responses to increased water temperatures will generally be defined by fish species or specific stocks, and how these changes will affect the various life stages (from egg to spawning adult). Nelitz et al. (2007) provided a useful species and life-stage-

13 Fleming, S.W. and E.J. Quilty. 2006. A novel approach: reconnaissance analysis of the Little Campbell River watershed. Report prepared for Environmental Environ. Qual. Sect., Lower Mainland Reg., B.C. Ministry of Environment. Aquatic Informatics Inc., Vancouver., B.C. Unpubl. report. www.env.gov.bc.ca/epd/regions/lower_mainland/water_quality/reports/ltl-campbell-riv/pdf/ltl-camp-riv-analysis.pdf (Accessed May 2010).

14 Ibid.

specific summary of potential biological vulnerabilities to climate-induced changes in water flows and temperatures. Increased temperatures in temperature-sensitive systems may result in increased frequencies of disease, increased energy expenditures, altered growth, thermal barriers to both adult and juvenile migration, delayed spawning, reduced spawner survival, altered egg and juvenile development, changes in biological productivity and other rearing conditions, and altered species distribution.

Changes in baseline conditions of aquatic ecosystems could also influence the outcomes of competition between species with differential temperature tolerances, as well as affect the necessary habitat requirements and survivability of sensitive species (Schindler 2001). Watersheds with warm water temperatures or low flows that currently affect salmonid survival are centred in the southwest, southern Interior, and central Interior of British Columbia (Nelitz et al. 2007). Under a changing climate, it is projected these areas will be further stressed. Salmonids show species-specific thermal optima and tolerances (Selong et al. 2001; Bear et al. 2007), and even small (1–2° C) differences in these conditions may result in marked differences in species distribution (Fausch et al. 1994). Distribution changes may be the direct result of the effects of water temperature on fish physiology, or (indirectly) a consequence of displacement of temperature-sensitive species such as bull trout (*Salvelinus confluentus*) by competing species such as rainbow trout (*Oncorhynchus mykiss*). Therefore, shifts in population distributions may be unavoidable and likely will result in the loss of salmonids in some areas where habitat conditions are currently close to tolerable limits (Nelitz et al. 2007). The effects of increased water temperatures are likely compounded wherever hydrologic regime changes reduce seasonal flows. For example, the limits of fish distribution in headwater areas are further altered by changes in the abundance and distribution of perennial, intermittent, and ephemeral watercourses.

Alternatively, in regions or specific water bodies where temperatures are below thermal optima for fish or temperature sensitivity is not a concern, increased water temperatures may promote fish growth and survival. Even minor temperature increments can change egg hatch dates and increase seasonal growth and instream survival in juvenile salmon. At Carnation Creek, minor changes in stream temperatures in the fall and winter due to forest harvesting profoundly affected salmonid populations, accelerating egg and alevin development rates, emergence timing, seasonal growth, and the timing of seaward migration (Tschaplinski et al. 2004).

The combination of increased temperatures and decreased late-summer base flows (low flows) could increase the stress for fish and other aquatic biota in the future. Low flows can cause a reduction in habitat availability, food production, and water quality, and can heighten the effects of ice on smaller streams during the winter time (Bradford and Heinonen 2008).

Changes in chemical water quality processes

Water quality changes related to temperature effects on terrestrial ecosystems are also possible. For example, increases in air temperature can increase soil productivity and rates of biogeochemical cycling, which may influence the chemical composition of runoff from terrestrial ecosystems. Soil microbes play an important role in influencing nitrogen retention and release to surface waters in forested watersheds (Fenn et al. 1998). Specifically, nitrification rates in soils are generally temperature-dependent; thus, nitrate concentrations in stream water are highly correlated with average annual air temperature (Murdoch et al. 1998) and future projected temperature changes.

Another important climate change factor that may change the rates of nutrient cycling in watersheds is the projected shifts in tree species composition related to temperature changes. This is because different tree species have different nutrient cycling regimes. Similarly, increases in other climate-related disturbances, such as wildfire or forest pest infestations, have the potential to increase nutrient cycling and leaching of mobile nutrients (e.g., nitrate) to surface waters (Eshleman et al. 1988). The effects of these disturbances are discussed in Chapter 12, "Water Quality and Forest Management."

One of the most direct effects of a changing climate on water quality is linked to changes in the timing and volume of streamflow. For example, as streamflows decline, the capacity of freshwaters to dilute chemical loadings will be reduced (Schindler 2001). Where the greatest temperature increases are projected during the summer and declines in surface water volumes are likely (i.e., the Columbia Basin and the Okanagan), water quality deterioration is possible as biologically conservative nutrients and contaminants could become more concentrated.

Where precipitation is expected to decline (i.e., southern and central British Columbia), deteriorating water quality will become a greater issue than in regions experiencing only an increase in air temperature. The key issue in these regions will be the decreased dilution capacity (higher pollutant concentrations) related to altered flows. Declines in surface water flows result in longer resident times for chemicals entering lakes (Whitehead et al. 2009). This is of greatest importance for biologically reactive chemicals for which longer resident times can result in increased biological reaction and increased potential for eutrophication (Schindler 2001).

Some of the effects we have described here may be mitigated in regions such as the Peace Basin and northwest British Columbia, where increases in summer precipitation and an overall wetter climate are predicted. For example, increased flows may potentially result in increased dilution of some nutrient contaminants, offsetting the effects of temperature increases and the associated evaporative demand. In some instances, greater dilution of pollutants may actually result in a positive effect on water quality. Similarly, an increase in the dilution capacity of streams may occur during the spring freshet in regions with predicted increases in winter precipitation. However, a counterbalancing effect may become evident on any water quality improvements because of an increase in stream power and non-point source pollutant loadings to watercourses. Higher runoff can lead to an increase in erosion and sediment transport in aquatic systems and reduced residence times, resulting in a decrease in chemical and biological transformations. This is of greatest concern for nutrients and chemicals that tend to adsorb to suspended solids, such as phosphorus and heavy metals. Higher concentrations of phosphorus, along with warmer temperatures, can promote algal blooms that reduce water quality (Schindler et al. 2008).

MODELLING REQUIREMENTS FOR CLIMATE CHANGE APPLICATIONS AT THE FOREST MANAGEMENT SCALE

Because of the uncertainty associated with the prediction of local climate change using climate models, natural resource managers must consider the effects of drier, wetter, more variable, less variable, or simply warmer conditions depending on the interactions of several site-specific environmental factors. Given the uncertainty of future climate projections at a regional level, as well as the incremental effects of various land uses on watershed processes, watershed-scale hydrologic models possess the potential to address short- and long-term forest management questions. These analyses may include problems such as an assessment of possible future growing conditions, the permanence of wetlands and small streams, or the potential changes to flooding, low flows, and other disturbances as a result of a changing climate. Yet, as a recent review of hydrologic models points out, numerous challenges are likely related to the inherent limitations of these models and the data inadequacies that exist across British Columbia (Beckers et al. 2009c).

In this section we highlight the specific qualities required in a hydrologic model *for climate change applications at the forest management scale*, and discuss several of the suggested improvements for climate change or forest management applications. Much of this information is summarized from Beckers et al. (2009a, 2009b), who provided a detailed review of several currently available hydrologic models and the suitability of these models for applications related to climate change. For a general discussion of weaknesses and limitations of using numerical models and other methods for detecting and predicting changes in watersheds, the reader is directed to Beckers et al. (2009a, 2009b, 2009c), and Chapter 16, "Detecting and Predicting Changes in Watersheds."

The suitability of any model depends on several components, such as available data and resources, and the ultimate end use of the modelled results. The presence of these components in selected watershed models will enable the simulation and investigation in a climate change context. Table 19.4 summarizes these critical components.

TABLE 19.4 *Climate change hydrologic model components (adapted from Beckers et al. 2009c)*

Modelled output	Required model component
Atmospheric evaporative demand	Solar radiation, humidity, and wind speed
Evaporation and precipitation interception	Leaf area index Stomatal resistance Forest growth (productivity) Forest survival (mortality) Temporal input control
Snow accumulation and melt	Physical or analytical snowmelt equations Rain-on-snow simulation
Permafrost, river ice, and lake ice	Frozen soil influence on water movement River and lake ice model component
Glacier mass balance adjustments	Glacier accumulation or melt model Glacier geometric response
Streamflow	Groundwater Lakes Wetlands Water consumption (water supply systems)
Stream and lake temperatures	Water temperature model component
Frequency and magnitude of disturbances	Channel routing (floods) Multiple vegetation layers (wildfires, pests) Vegetation albedo, radiation transmissivity (wildfires, pests) Soil albedo (wildfires) Hydrophobicity (wildfires) Landslide simulation

Downscaling for Watershed Modelling

Projected changes to climate are available at scales of greater than 10 000 km^2, whereas watersheds of interest generally range from 5–500 km^2 in size. Linking large-scale global climate model projections to hydrologic models requires downscaling of climatic data. The downscaling method will depend on the hydrologic model used and the nature of the question to which the model is applied. Statistical methods are most common, as these are computationally less intensive than dynamical methods. These methods range from the bias correction spatial downscaling techniques designed for use with gridded models to draw on monthly GCM data (Wood et al. 2002; Widmann et al. 2003; Salathé 2005), to more sophisticated applications such as hybrid methods that use daily information from GCMs and draw on the strengths of statistical tools and stochastic weather generators. Dynamic downscaling results from regional climate models (RCMs) are being produced at higher resolution over British Columbia (approximately 15 km^2), and multiple RCMs have been compared over North America via the North America Regional Climate Change Assessment Program. The Pacific Climate Impacts Consortium is developing methods that apply statistical downscaling to dynamically downscaled projections to provide higher temporal and spatially resolved information similar to approaches applied outside of Canada (e.g., Bürger 2002).

Global Climate Model Selection for Watershed Modelling

Modelling future changes requires a clear rationale for GCM selection. The GCMs selected will dictate the range and median of future projected changes (Pierce et al. 2009). For example, the range in projections for the 2050s depends more on the choice of models than on emissions scenarios (Rodenhuis et al. 2007). To reduce computational time, remove outliers, and ease interpretation of results, Hamlet et al.[15] have selected a subset of the GCMs; however, a clear consensus on how to evaluate model performance and select outliers does not currently exist. Overland and Wang (2007) identified and eliminated outliers by comparing historical GCMs to observational data, whereas Manning et al. (2009) weighted less biased models more greatly to create a probabilistic ensemble. A carefully selected subset will likely represent the range of possible wet-dry and warm-cool futures to present an adequate characterization of the related uncertainty. In British Columbia, knowledge of GCM model selection is currently expanding. The Pacific Climate Impacts Consortium expects to publish foundation papers and a guidance report on this topic in 2010. For more information, go to: http://pacificclimate.org/.

Modelling Atmospheric Evaporative Demand

Increases in atmospheric evaporative demand may lead to greater evaporative losses from water bodies and changing water demands of vegetation. Incorporating weather variables into calculations of reference evapotranspiration is therefore critical. Subsequently, physically based approaches to calculating evapotranspiration should provide the greatest level of confidence in results. Because empirical methods are based on historical data, physically based equations are better suited for predicting possible shifts in hydrologic responses outside historical data ranges. Many of the models reviewed by Beckers et al. (2009c) employ the Penman-Monteith equation recommended by the Food and Agricultural Organization of the United Nations and the American Society of Civil Engineers to determine reference evapotranspiration (Allen et al. 2005).

Although the theoretical understanding of suitable equations to calculate reference evaporation is advanced, the main challenge in anticipating future increases in evaporative demand arises from a lack of understanding regarding possible changes in temperature, solar radiation, humidity, and wind speed. Projections of future climate change have focussed primarily on analyzing and downscaling mean temperature and precipitation outputs from GCMs. Relatively little research has been done to extract and analyze the remaining variables, or to find adequate methods for downscaling modelled output into formats suitable for use in hydrologic models, to points (representative of meteorological stations), or to high-resolution grids. Thus, we need to develop improved methods of downscaling solar radiation, humidity, and wind speed from GCMs to drive hydrologic models.

Modelling Future Evaporation and Precipitation Interception

To apply hydrologic models for planning purposes, we must consider the issues surrounding forest growth and mortality. When conducting long-term model simulations, it may be important to determine whether the model input is easily adapted to represent gradual or abrupt changes in vegetation disturbance. The ability to vary vegetation properties over time within a single model simulation (i.e., the ability to change properties without having to re-start the model) is referred to as "temporal input control" (Table 19.4).

The amount and type of vegetation and its physiological characteristics have an important effect on site water balance. The interaction between vegetation and the atmosphere (i.e., evapotranspiration, precipitation interception) is determined by vegetation surface area (Monteith and Unsworth 1990; Shuttleworth 1993), typically represented as leaf area index (LAI) in most hydrologic models. Leaf area index is also a primary reference parameter for plant growth. Thus, within a climate change context, explicit representation of vegetation (i.e., LAI) is a critical model parameter to describe forest characteristics, and potential effects of episodic or long-term changes.

Stomatal resistance (or its inverse, stomatal conductance) is another crucial parameter (see Table 19.4) used to calculate the vegetation transpiration rate from humidity (vapour pressure) gradients (Monteith and Unsworth 1990). Stomatal resistances vary between plant species and are an important physiological model parameter. Hydrologic models need to simulate the closing of stomata (i.e., an increase in stomatal resistance) when atmospheric water demand exceeds water availability (i.e., to describe plant response to atmospheric and soil drying). Therefore, inclusion of multi-layered vegetation and associated vegetation parameters can be an important quality for a hydrologic model to pos-

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sess. To improve the ability of hydrologic models to simulate the hydrologic effects of altered vegetation composition, suggested model improvements include adapting watershed models to include forest growth and mortality, linking to existing forest growth and mortality models, and (or) adding temporal input control to some models.

Modelling Future Snow Accumulation and Accelerated Melt

For long-term simulations of climate change, a key challenge is the ability of a hydrologic model to spatially simulate both snow accumulation and snowmelt processes. Over a single model run, these models must also be able to initially represent predominantly nival conditions that then become hybrid (mixed) conditions or even pluvial (Beckers et al. 2009c). Additionally, changes in the form of precipitation (rain or snow) in the late fall or early spring may become increasingly important factors to simulate. As such, the ability of hydrologic models to accurately model mixed regimes (i.e., rain-on-snow energy transfer) can be crucial. Snowpack accumulation and melt is also an important factor for other water balance components, as these processes relate to albedo and snow-covered versus bare ground. Where models do not accurately model the spatial extent of snow, errors can occur in estimating snowmelt contributions to streamflow or in predicting the onset or rate of evapotranspiration. Model testing approaches (e.g., Jost et al. 2009) that incorporate SWE data measurements from a range of elevations and aspects hold promise in helping to validate model output in mountainous, data-sparse watersheds. Models with physically based or analytical (temperature-radiation) snowmelt routines are better suited than empirical models to predict the potential for accelerated melt under a changing climate (Table 19.4) for the same reasons mentioned previously.

Modelling Soil Freezing, Permafrost, Lake Ice, and River Ice

River and lake ice formation and break-up processes are often the focus of specialized kinematic models (e.g., Beltaos 2007) that are not typically incorporated into watershed-scale hydrologic models used in forest management applications. Soil temperatures, however, are more widely accounted for in watershed models, typically to calculate the ground heat flux component of the snowpack energy balance (e.g., Wigmosta et al. 1994). Only the Cold Regions Hydrological Model (Pomeroy et al. 2007) has the ability to assess frozen soil conditions (via soil temperatures) and associated effects on water movement among the models reviewed by Beckers et al. (2009c). The following general modelling improvements are therefore suggested.

- Increase the ability of hydrologic models to simulate the effects of permafrost thaw on hydrological processes applicable to the northern portions of British Columbia, Alberta, and other areas where permafrost occurs. Frozen soil conditions may also be important to model in non-permafrost areas (e.g., effects on infiltration).
- Improve our understanding of how climate change will alter the three-way interaction between streamflow generation, water temperatures, and river and lake ice formation and break-up.
- Develop tools that allow resource managers to assess the importance of these interactions (and how they may change in the future) for forest management.

Modelling Glacier Mass Balance

For some watersheds, the ability to simulate changes in glacial melt contributions to streamflow may be critically important. Glacial processes are represented in some models that simulate the increased melt rates related to climate change (Beckers et al. 2009c); however, for long-term simulations, it is also necessary to calculate glacier mass balance and to adjust glacier area and volume (i.e., to simulate glacial retreat) . Two important components are the capacity to: (1) track glacier mass balance, and (2) account for glacier geometric response to mass balance. This latter function was built into a version of HBV-EC (Stahl et al. 2008) by drawing on the concept of volume-area scaling. The Western Canadian Cryospheric Network is currently working on a model suite that will project glacier response using a physically based glacier dynamics model, which will then be used in parallel with a hydrologic model to generate scenarios. Alternatively, stand-alone models of glacier mass balance can be used to estimate future glacier volume, which will become an input to hydrologic models with glacier processes.

Modelling Future Stream Temperatures

Models to predict stream temperatures fall into two general classes (Sridhar et al. 2004): (1) empirical relationships based on observations of stream temperature and stream properties (such as discharge, channel geometry, and streamside vegetation characteristics); and (2) models that represent the energy balance of the stream. Recently, the use of physically based models to predict stream temperature has become feasible by interfacing with GIS methods. Although numerous models have been developed to predict stream temperature (Webb et al. 2008), none of the hydrologic models reviewed by Beckers et al. (2009a, 2009c) possessed this capability inherently. At a larger scale and as mentioned above in "Changes in stream and lake temperatures and effects on fish," Morrison et al. (2002) conducted a comprehensive assessment using the University of British Columbia Watershed Model, in conjunction with projections of future temperature and precipitation, to generate streamflow scenarios for sub-basins of the Fraser River. Other temperature models are used operationally in British Columbia, such as the FJQHW97 river temperature model. The federal Department of Fisheries and Oceans has used this model for the Fraser River during the salmon migration period and it has played an important role in aiding decisions to open or close commercial fisheries (Foreman et al. 2001). This model was also used for climate change analysis (Foreman et al. 2001).

To improve future stream temperature simulations, existing watershed models could be adapted to spatially simulate stream temperatures or couple to existing aquatic (e.g., salmonid) habitat simulation models. However, where surface water–groundwater interactions are strong controls on stream temperature, fully coupled models that include subsurface processes at a relevant scale would be necessary.

Modelling the Future Frequency or Magnitude of Forest Disturbances

Watershed modelling can be used to assess the suitability of current infrastructure (e.g., stream crossings) under potential future climate conditions, and (or) to determine the suitability of engineering design criteria using scenarios. In some rain-dominated regimes, the ability of watershed models to examine such questions may depend on the accurate simulation of preferential runoff mechanisms (e.g., Carnation Creek on Vancouver Island; Beckers and Alila 2004). In snow or mixed regimes, accurate simulation of melt rates is important for predicting peak flows (e.g., Redfish Creek in southeast British Columbia; Schnorbus and Alila 2004).

Other disturbances that are projected to increase include wildfire, forest pests (insects), windthrow, breakage of trees, and landslides. Of these disturbances, the modelling of landslides provides a clear synergy with watershed simulation (Table 19.4). Landslide modelling has been the focus of specialized physically based models, such as the distributed Shallow Landslide Analysis Model (dSLAM; Wu and Sidle 1995) and the Integrated Dynamic Slope Stability Model (IDSSM; Dhakal and Sidle 2003), and has been incorporated in the Distributed Hydrology Soil Vegetation Model (DHSVM; Doten et al. 2006).

In contrast, specialized windthrow models (e.g., Lanquaye and Mitchell 2005) currently offer minimal synergies with watershed modelling. This lack of synergy also holds true for predicting the occurrence of pests. It is critically important, however, for hydrologic models to incorporate (as inputs) the changes in physical watershed characteristics that may occur as a result of these disturbances. For example, an important aspect related to tree mortality is the change in canopy albedo and solar radiation transmissivity (Table 19.4), which in turn affects the radiation energy balance of affected stands.

Forest fires also cause vegetation changes that, depending on fire behaviour, may include either removal of the understorey without canopy disruption or full combustion of the overstorey, resulting in standing dead timber. These complex changes can be represented in a straightforward fashion only with models that allow for multiple (stratified) vegetation layers (Table 19.4). Fires can also cause changes in soil properties that affect the hydrologic response, including altered soil albedo, and (under certain conditions) the formation of hydrophobic conditions, which limit soil infiltration (Agee 1993). Although soil hydrophobicity is known to decline over time, the overall process is poorly understood (DeBano 2000) and, as such, the ability to simulate these conditions is challenging. For example, although it is possible to alter soil physical properties in existing hydrologic models, representing the potential effect of soil hydrophobicity on infiltration is problematic because no models allow temporary changes to soil properties within a single model run to account for a reduction in hydrophobicity over time (Beckers et al. 2009c).

The current understanding of climate change influences on average meteorological conditions is much further developed than that of understanding potential changes in the frequency and magnitude of extreme events (Rodenhuis et al. 2007). An improved understanding of extreme events (temperature, precipitation, and wind) under a changing climate is needed to advance hydrologic modelling. An increased ability to use models to investigate potential forest disturbances such as landslides, fire hazards, pests (insects), and windthrow is also needed. The outputs from these models could then be used to parameterize hydrologic models for forest management purposes.

Modelling Future Streamflow

Most currently available watershed models will calculate changes in streamflow, infiltration, soil moisture conditions and shallow subsurface runoff, and the subsequent discharge of water to the stream channels without applying any modifications to the model. Nonetheless, specific questions regarding the interaction of forest management and climate change may create difficulties for existing models in certain settings. For example, changes in groundwater recharge rates associated with climate change (e.g., Scibek and Allen 2006a, 2006b) may have consequences for base flow contributions to low flows. The capability to account for the anticipated increased competition between human use and instream needs may be another important feature in selecting a model (Table 19.4).

Improvements in simulating altered peak and low flows in a changing climate are often contingent on advances in the previously discussed topic areas (evapotranspiration, snow accumulation and melt, permafrost and river and lake ice processes, glacier mass balance adjustments, etc.). Furthermore, if a model was developed and calibrated to simulate snowmelt-dominated watershed conditions and is subsequently used to assess the consequences of a regime shift to mixed or rainfall-dominated regimes, its accuracy in predicting future streamflow conditions may be reduced. Additional model improvements include processes related to groundwater, wetland and lakes, and other factors such as human water consumption (water competition) that affect streamflow. This capability is currently limited in those models reviewed by Beckers et al. (2009a, 2009c).

The watershed models reviewed by Beckers et al. (2009c) had varying capabilities for examining climate change questions; however, incremental enhancements to existing models (rather than the development of new models) will help guide forest management decisions. For instance, to apply the complex, physically based models better suited to addressing climate change questions, further efforts are required to enhance and organize data resources. Examples include producing spatially coherent vegetation data sets with up-to-date LAI and stomatal resistance information, and incorporating weather variables such as solar radiation, humidity, and wind speed into climate change projections. A fundamental barrier to considering climate change in a forest management context is the uncertainty in possible future climates (and emissions), with current projections offering a wide range of possible outcome scenarios.

SUMMARY

British Columbia's climate has changed over the last 100 years and will continue to experience change with the future looking warmer and wetter. Transformation of local air temperature and precipitation regimes will drive changes in groundwater and the magnitude and timing of both low and high streamflows in any given watershed. Many areas will see accelerated snowmelt and increased water levels in the winter. Projected warming coupled with altered streamflows will likely increase stream temperatures affecting water quality and, consequently, fish in many areas. Glaciers and permafrost will to continue to melt, and landslide regimes will ultimately respond to all of these drivers. The associated effects will have many important implications for the fisheries, agriculture, forestry, recreation, hydroelectric power, and water resource sectors. As this chapter has illustrated, the effects at a local scale will be complex and vary in importance according to the sensitivity of local watersheds conditions to climatic changes.

Currently, practical management responses to climate change are not well formalized, as the focus of the past few years has largely been on projecting and understanding what the future might hold. As a next step, the development of effective climate change management responses will likely involve local-level strategies that result in both short- and long-term benefits to ecosystems and society beyond climate change applications. The selection of such a suite of approaches may be the best chance to ensure the effective stewardship of watershed resources and associated values in the future.

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