Climate Change and Low Flows:
Influences of Groundwater and Glaciers

Final Report for Climate Change Action Fund Project A875

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Executive Summary

Streamflow patterns in BC can be broadly classified as rain-dominated, hybrid (rain and snow), snow-dominated, and glacierized, each having distinct low flow seasons. In this study, we chose to focus on the summer/early autumn low flow period, as that has particular economic and ecological significance.

Variability in stream discharge was measured over a low flow season in Bertrand Creek, Pepin Brook and Fishtrap Creek, which drain the Abbotsford-Sumas aquifer in southwest British Columbia and northwest Washington State. Discharge was measured across a series of profiles, both repeating across a central line (at-a-station measurements) and along consecutive sections from upstream to downstream (successive downstream measurements) to assess the repeatability of the low flow measurements. As the flow in these small streams decreases, the coefficient of variation of the mean discharge tends to increase when the flow is very low (i.e., <0.1 m^3/s), although the trends for Fishtrap are weak and less reliable likely due to the higher flows at that site. Overall, variations in measurements are of sufficient magnitude that subtle changes in streamflow due to, for example, climate change may be difficult to detect.

Unglacierized catchments generally exhibited declining trends in September streamflow through most of British Columbia, but not in August. Glacierized catchments, on the other hand, show the opposite pattern, with dominantly negative trends in August but not September. The decreases in September flows are broadly consistent with a decline in September precipitation over the period studied. For all four regimes, the most important control on August streamflow is August precipitation. August streamflow is also positively but more weakly related to lagged variables including July precipitation and the previous winter's precipitation.

There is clear evidence for the effect of multi-year storage change only for the glacierized catchments. Glacier retreat over the last few decades has apparently reduced glacier area sufficiently so as to reduce meltwater generation and thus streamflow. Snowmelt-dominated catchments showed no tendency to trends or serial correlation in the model residuals, though the runs test suggested a substantial number of stations had non-random residuals. Both rain-dominated and hybrid catchments tended to have positive residuals in August, suggesting either an increasing trend in groundwater storage or a spurious cause such as model mis-specification.

The majority of groundwater well observation records in BC start in the late 1970s or 1980s, providing a total of 20 to 30 years of record. A select few of these wells monitor aquifers in pristine areas that reflect natural variability; the others have been influenced by human activity making them less representative. Climate change detection requires long term records because natural climate variability phenomena, such as the Pacific Decadal Oscillation (PDO), last over multiple decades. Long term fluctuations of precipitation, air temperature, and stream flow can be reflected in groundwater level fluctuations, but the hydrogeology of a given aquifer plays a unique role in the interaction of climate, surface water, and groundwater. By comparing the relationships between groundwater, climate, and surface water within and between groups of well records from the two major hydro-climatic zones in BC, a system for detecting the influence of climate change and variability on groundwater in the absence of long term records is defined, and suitable correlation coefficients are applied to evaluate the strength of these interactions. Different aquifer types are assessed with respect to vulnerability to climate change influences.

Overall, summer groundwater levels appear to have lowered across the province, despite an increase in winter precipitation and recharge during the same time period. Due to the limited availability of long well records near gauged streams, the attribution of whether and how these changes have affected low flows proved difficult. The available groundwater observation wells vary substantially in terms of aquifer properties and the hydraulic connection to rivers and for many, we suspect the natural records may be altered by changes in the abstraction patterns. The few examples where streamflow and groundwater observations are available in the same basin and are nearby furthermore provide some of the exceptions to the trends found. In Abbotsford, even though a field study and a modelling study showed that summer low flow is fed by groundwater, low flows increased while over the same time period groundwater levels decreased. The same was found for Grand Forks, where previous studies have shown that streamflow is fed by groundwater in the summer. However, the recharge mechanisms are different in both examples.

A methodology was developed to simulate the transient effects of glacier retreat on streamflow patterns by coupling a semi-distributed hydrological model (HBV-EC) with a glacier mass balance and glacier scaling model. This study investigated the sensitivity of streamflow to changes in glacier cover for the Bridge River basin in British Columbia, which is an important water source for one of BC Hydro's hydro-electric facilities. The model is driven into the future assuming three different types of climate scenarios: a continuation of the current climate and a moderate warming scenario realized with a weather generator, as well as a set of downscaled

GCM scenarios with greenhouse gas forcing. Considering specifically the effect of changing glacier area in the basin, the modelled glacier mass balance is used to re-scale the glacier every decade using a universal volume-area-scaling relation. The model application shows strong reductions in glacier area and summer streamflow even under the assumption of a continuation of the present climate. The results of this study suggest that climate warming and associated glacier retreat will have significant implications for water resources and aquatic ecology. Glacier-fed rivers are likely to experience a shift from a glacial regime with high flows in mid and late summer, with an associated moderating effect on stream temperature, to a regime that responds to the summer dry period with streamflow recession, low flows and increased temperatures.

Acknowledgements

Financial support from the Government of Canada's Climate Change Impacts and Adaptation Program via Grant A875 is gratefully acknowledged. BC Ministry of Environment (BCMoE) provided the funding for digitizing the water level records described here. Vanessa Sadler and Mandy Mo digitized the hundreds of monthly charts involved. Carl Lee and Mike Wei from BCMoE were helpful with dealing with and advising on a wide range of technical issues and problems associated with paper records of water levels. Terrance Dong performed quality assurance of the digitized record and critical reconstructions. Bill Hodge, retired from BCMoE (formerly BCMELP) performed critical reviews of problematic records and his experience and advice was crucial in helping us identify and eliminate records unsuitable for analysis. Mike Toews and Alex Cannon developed code in R to present and analyse the groundwater level hydrographs. Throughout our work, the continued personal interest of Jenny Fraser from the Climate Change section of BC MoE has been appreciated. For the groundwater components of the study a NSERC Undergraduate Scholarship to Mary Ann Berg and a NSERC Discovery Grant held by Diana Allen supported the Abbotsford Case Study. For the glacier components of the study a Natural Sciences and Engineering Research Council (NSERC) Scholarship to JM Shea and a NSERC Discovery Grant held by RD Moore supported research that provided important background material for this project. BC Hydro kindly provided quality controlled meteorologic data and Mike Demuth supplied the mass balance data for Bridge Glacier. Markus Weiler helped with the implementation of the glacier scaling algorithm into IDL.

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1. Introduction and objectives

1.1 Background

"Low flow" is used in a variety of ways among scientists, aquatic resources managers and the general public. A general usage refers to the portions of an annual hydrograph when flows are typically lower than during other parts of the year. For example, streams in areas with Mediterranean climates typically have low flows during late summer and early autumn, when relatively low rainfall coincides with high evapotranspiration. The term "low flow" can also be used more specifically, for example, to refer to the lowest flow in a given year, averaged over some defined period, typically 7 days. In this project, we adopted the former general definition.

During low flow periods, streamflow may be inadequate to meet needs for economic uses such as domestic consumption, hydroelectric generation, irrigation and effluent dilution, as well as ecological functions such as in-stream habitat. There is growing concern that climate change may result in more extreme low flows, associated with increases in air temperature and evapotranspiration, lower snow accumulation and earlier snowmelt, and possibly more extreme summer drought. Low flows occur during periods of minimal rainfall or snowmelt inputs and are sustained by release of catchment storage in the forms of groundwater, glaciers, lakes and wetlands. Therefore, to understand the potential impacts of climate change on the magnitude of low flows, more knowledge is required on how the climate changes will influence the magnitude and timing of storage releases. Of the main forms of catchment storage, our understanding is weakest for groundwater and glaciers, particularly in British Columbia. The following section provides a brief review of current knowledge, to help set the context for the objectives of this research project, which will be presented in section 1.3.

1.2. Previous work

1.2.1 Climate change and groundwater

Groundwater is the dominant source of baseflow in many catchments and is, by nature, a difficult entity to quantify and evaluate. Consequently, studies of the impact of climate variability and change on groundwater are limited, both within BC and elsewhere around the world. Elsewhere in Canada, Piggott et al. (2001) separated stream flow data for a network of 174 watersheds into surface runoff and groundwater discharge components and analyzed these in a time series context. The results of that analysis indicated that as much as 75 percent of melting snow and rain is available for runoff and groundwater recharge during the winter, while as little as 10 percent is available during the summer. The portion of this excess precipitation that recharges groundwater is a function of physiography and varies from 10 to 80 percent when calculated by watershed. The rate at which this recharge subsequently discharges to form base flow and, therefore, the persistence of this flow, is also a function of physiography. These results were used in combination with two climate change scenarios to determine the potential impacts of this change on groundwater conditions. The two scenarios resulted in a significant, and more consistent, impact on the annual distribution of base flow, with increased flows during the winter due to a reduction in snow accumulation and decreased flows during the spring and early summer due to the corresponding decrease in snow melting. These results clearly indicate a potential for significant impacts on groundwater conditions and, therefore, water supplies, in-stream conditions, and aquatic habitat. To date, there have been no targeted studies in British Columbia that specifically address the influence of climate change on low flows and groundwater conditions.

Linkages between low flows, groundwater and climate fluctuations have been documented in some regions of BC. For example, within the Grand Forks aquifer in south-central BC, the aquifer is strongly connected to the aquifer (Allen et al., 2004a). Seasonal fluctuations of the water table within the floodplain are damped compared to river stage, and also exhibit phase shifts of at least 15 days, and possibly up to as much as 30 days. During the rising and peak stages of the stream hydrograph during snowmelt, streamwater recharges the floodplain aquifer. Some time after freshet, the drop in river stage induces groundwater discharge toward the stream, which can help sustain late-summer flow. Numerical groundwater flow modeling, conducted as part of CCAF project (A369) (Allen et al., 2004b), successfully captured the temporal connection between the Kettle River and the aquifer in Grand Forks and simulated the response of the groundwater system to predicted climate change. The key effect of climate change was to advance the timing of snowmelt, and thus the timing of groundwater recharge, ultimately resulting in a decrease in late-summer groundwater discharge.

1.2.2 Climate change and glaciers

Mountain glaciers are an important source of freshwater to the rivers they feed. They tend to moderate interannual variability in streamflow and, in particular, maintain higher flow during extended periods of summer drought (e.g., Fountain and Tangborn, 1985; Fleming and Clarke, 2005). Glacier runoff during warm, dry weather also appears to regulate summer stream temperatures (Brown et al., 2005; Moore, 2006) and thus maintains high-quality habitat for cold-water species such as salmonids. In sum, glacier cover can play an important role in the aquatic ecology of downstream reaches (Fleming, 2005).

Mountain glaciers dominantly have been receding since the end of the Little Ice Age in the Canadian Cordillera (Osborn and Luckman, 1988), and worldwide (IPCC, 2001a) and continued global warming is predicted to result in continued retreat of mountain glaciers worldwide (IPCC, 2001b; Barnett, 2005). Changes in glacier mass balance and glacier coverage can influence streamflow over a range of time scales. In the short term, a shift to negative mass balance, associated with decreased winter accumulation and/or increased summer temperature, should increase glacier runoff due to the earlier disappearance of high-albedo snow and the exposure of lower-albedo firn and/or ice, as well as the effects of increased energy inputs (Singh and Kumar, 1997). However, in the longer term, glacier recession may decrease glacier area sufficiently to reduce meltwater volumes, generating concerns about the future sustainability of summer flows (Barnett et al., 2005).

Several studies have examined longer-term streamflow variations in glacier-fed catchments. Braun and Escher-Vetter (1996) found that annual streamflow below the Vernagt Ferner in Austria increased following a shift from roughly neutral to strongly negative mass balance. In Switzerland, most basins with more than 10% glacier cover have tended to exhibit increasing summer streamflow, while basins with less than 10% glacier cover have exhibited negative trends (Birsan et al., 2005). Fleming and Clarke (2003) found similar contrasting trends in glacierized and unglacierized catchments in the western subarctic of Canada. However, further south in western Canada, decreasing streamflow trends were found in the Rocky Mountains (Demuth and Pietroniro, 2003). In the Coast Mountains, Moore and Demuth (2001) found no clear trend in raw August mean streamflow for Place Creek below Place Glacier during a period of negative mass balance and retreat. However, after accounting statistically for the effects of winter snow accumulation and August air temperature, they found a significant negative trend, which they attributed in part to the decrease in ice area available for melt. Other mountainous regions in the world show similar patterns (Collins, 2006).

The potential effects of future changes in climate regimes and glacier cover on hydrology has received considerable attention using simulation models. The response of glaciers to climate change has been modelled for individual glaciers (e.g. Radic and Hock, 2006) as well as for large regions (e.g. van de Wal and Wild, 2001). Most studies used GCM climate output to drive a glacier mass balance model, and many applied a volume-area scaling relation (Bahr, et al. 1997) to determine the areal evolution, in lieu of explicit modelling of glacier dynamics.

Precipitation-runoff models have been used to estimate the effect of climate change on streamflow in glacier-fed basins. However many of these studies did not adjust the glacier cover to account for glacier response to the imposed climatic changes (e.g., Moore, 1992; Singh and Bengtsson, 2005). Hagg et al. (2006) examined the effect of an assumed reduction in glacier area in central Asia, but did not model the transient streamflow response associated with changing glacier area. Horton et al. (2006) updated the glacier area for simulation of future conditions assuming a constant accumulation-area ratio. Only the study by Rees and Collins (2006) appears to have considered the transient response associated with glacier retreat. They assumed a simplified glacier geometry and removed elevation bands as ice thickness depleted.

1.3. Objectives of research

The specific objectives of the research are:

1. To evaluate relations between groundwater fluctuations and past climatic variations using available climate data and data from the recording well network in BC;
2. To examine the links between low flow magnitudes, groundwater and climate fluctuations for selected catchments in BC;
3. To characterize the interactions between climate variations and glaciers, and their influence on streamflow, particularly during the late summer/early autumn “transition to winter” period in mountain catchments;

4. To explore the effect of future climate change scenarios on groundwater and glaciers, and the subsequent changes in streamflow, by applying statistical and conceptual streamflow models.

5. To transfer the scientific and technical knowledge both before the project is initiated and upon completion of the project from/to water users (e.g., BC Hydro), water purveyors, fisheries researchers, and local authorities.

1.4 Outline of the report

The remainder of this report comprises 4 sections, as outlined below.

- Section 2 provides an overview of low flows in British Columbia. It reviews existing knowledge on the spatial and temporal variability of low flows, and also provides analyses of current trends in important climate variables and summer streamflow. It also presents the results of statistical analyses aimed at identifying the sensitivity of summer low flows to hydroclimatic variables and catchment characteristics, including the timing and magnitude of the snowmelt freshet, post-freshet rainfall, and glacier cover.

- Section 3 covers the linkages between climate, groundwater and late-summer streamflow. It builds on the trend analyses presented in section 2 by examining trends in groundwater levels and attempts to link these to trends in later-summer streamflow.

- Section 4 uses computer simulation to examine how glacier response to future climate scenarios may influence summer streamflow.

- Section 5 provides a discussion of the main findings, with suggestions for future research.

The report also includes four appendices, comprising details on processing groundwater level data, a description of the Abbotsford aquifer case study, the method used for downscaling General Circulation Model output, and a listing of publications produced and outreach activities conducted as part of the project.

2. Low flows in British Columbia: measurability, patterns and controls

2.1 Introduction

This section provides an overview of low flows in British Columbia in relation to the dominant hydroclimatic regimes. It also reviews the socio-economic and environmental impacts of winter versus summer low flows, and summarizes what is currently known about the influences of climate variability, climate change and land use on late-summer flows. The measurability of low flows is then examined based on field measurements made at a site in the lower mainland of BC. Empirical analyses are then presented to examine trends in climatic and streamflow variables, and the influence of climatic variability on streamflow. These analyses provide the context for more detailed analysis of climate-groundwater-streamflow and climate-glacier-streamflow linkages in sections 3 and 4.

2.2 The regional context

2.2.1 Description of the study region

The Pacific and western mountain region of Canada (also called the "Canadian Cordillera") stretches from the Pacific Ocean in the west to the Rocky Mountains in the east and encompasses the Province of British Columbia, the southern part of the Yukon Territory and the southwestern part of Alberta. The region is characterized by several mountain chains with elevations of over 3000 m.a.s.l., vast plateau areas, and deep valleys. Geology and geomorphology are highly variable. The mountains expose volcanic and sedimentary bedrock. Pleistocene glacial, fluvioglacial and glaciolacustrine deposits can be found almost anywhere around in the region and considerable Holocene alluvial deposits exist in the valleys of most large rivers.

The complex relief of the Canadian Cordillera creates strong gradients and differences in temperatures and in precipitation amounts. During the winter months frontal weather systems formed in the Pacific Ocean bring moisture into the region. When storms that travel with the prevailing westerly winds aloft hit the north-south oriented mountain chains, the air is forced to rise and as it cools and condenses, high amounts of precipitation are dropped onto the western slopes of the mountain ranges. Conversely, the eastern slopes and the valleys in the lee of the mountains are dry. Except at low elevations along the coast, winter precipitation falls as snow. Summers are relatively dry throughout the region and considerable climatic moisture deficits develop, particularly in the southern Interior where summer temperatures are highest.

As a result of the diverse physiographic and climatic characteristics, all large river basins encompass many different biogeoclimatic zones (Krajina, 1969), and even small basins commonly span a large range of elevations and include a diverse landscape of exposed rock, old and young forest, and possibly alpine meadows with small wetlands and lakes. Almost half of the currently gauged rivers in British Columbia have a mountain glacier in their watershed.

2.2.2 Low flow processes and timing

Climatic influences on low flow timing and influences of basin storage properties on streamflow recession and magnitude are modified by elevation and basin hypsography. Hydrological regimes in the Canadian Cordillera are predominantly nival with winter low flows (Figure 2.1). An exception is low-elevation pluvial regimes along the coast, where summer is the primary low flow season (Figure 2.1 – Carnation Creek). Beginning and end of the winter low flow season depends on when temperatures drop below/rise above freezing. As winter temperatures generally decrease with elevation, with distance from the coast, and with latitude, the timing of winter low flows is controlled by the location as well as the elevation of the basin.

During summer, after a prolonged period with a climatic moisture deficit, streamflow in nival regimes is sustained either by late-lying snow or glacier melt or by the release of stored water from groundwater, lakes and wetlands. A secondary low flow period then typically develops first in basins where snowmelt occurred early in spring and/or over a short period of time, i.e. basins at lower elevations or with small elevation differences (e.g. Figure 2.1 – Hidden Creek). In glacierized basins, increased ice melt contributions during the warm season maintain high streamflow levels beyond the snowmelt period (e.g. Figure 2.1 – Nass River). Groundwater flow in alpine areas has only recently emerged as a topic of research and its significance appears to vary with location. Mountain soils and aquifers tend to be shallow with limited storage capacity, and during a dry summer it is not unusual for small streams in the southern part of the Cordillera to become dry in August or September. In a field study at Lake O'Hara in the Rocky Mountains, however, Hood et al. (2006) found that groundwater contributions to the lake are considerable and remain stable during late summer. Extensive alluvial aquifers in the larger valleys across the region can provide late summer base flow to rivers after being recharged from the river during the spring freshet (Scibek et al., 2007).

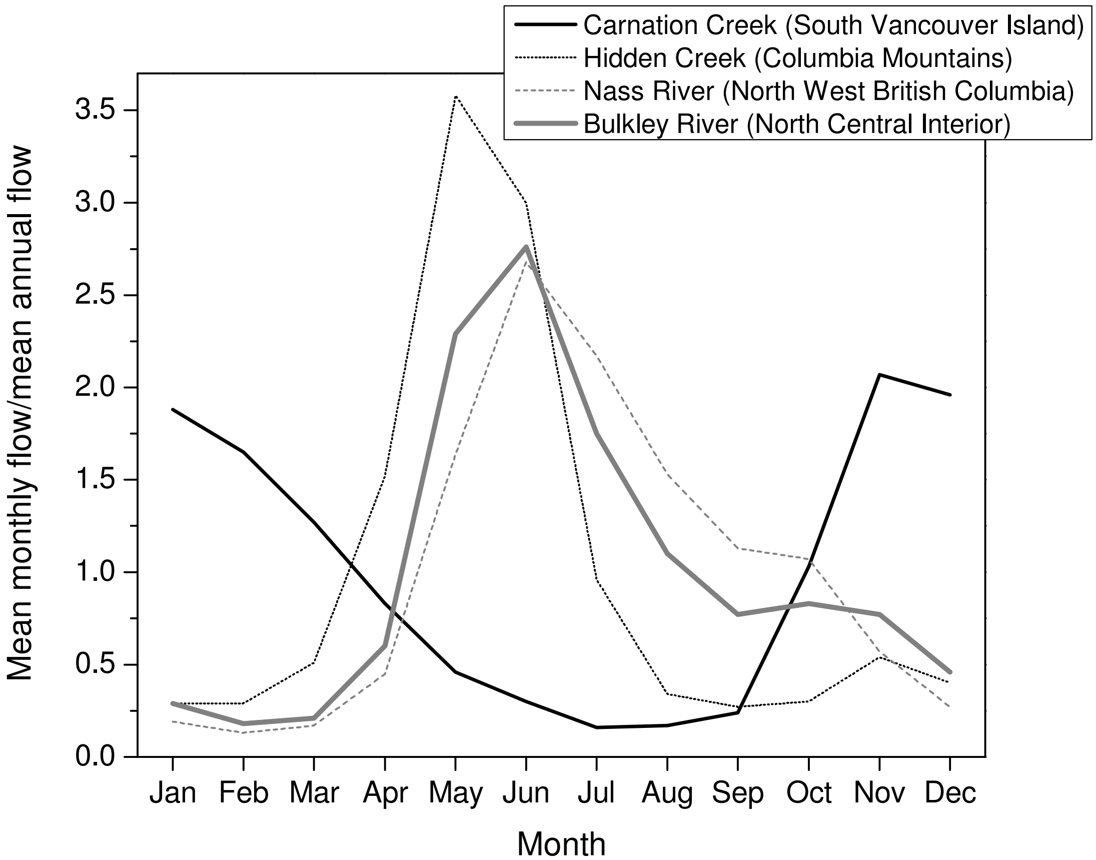


Figure 2.1 Hydrological regimes in the Canadian Cordillera

Low flow regionalisation in mountainous environments is difficult due to competing impacts of climate, elevation, geology, existence of stores such as lakes/wetlands/glaciers, all of which can vary strongly over short distances. Figure 2.2 shows the specific discharge of the mean 7-day minimum flow during the summer half year across BC. Summer low flows vary strongly across the region and low and high streamflow may occur in neighbouring basins. Generally, however, summer low flow per basin area is lowest in the southern plateau areas and valleys and highest in glacierized high-mountain areas.

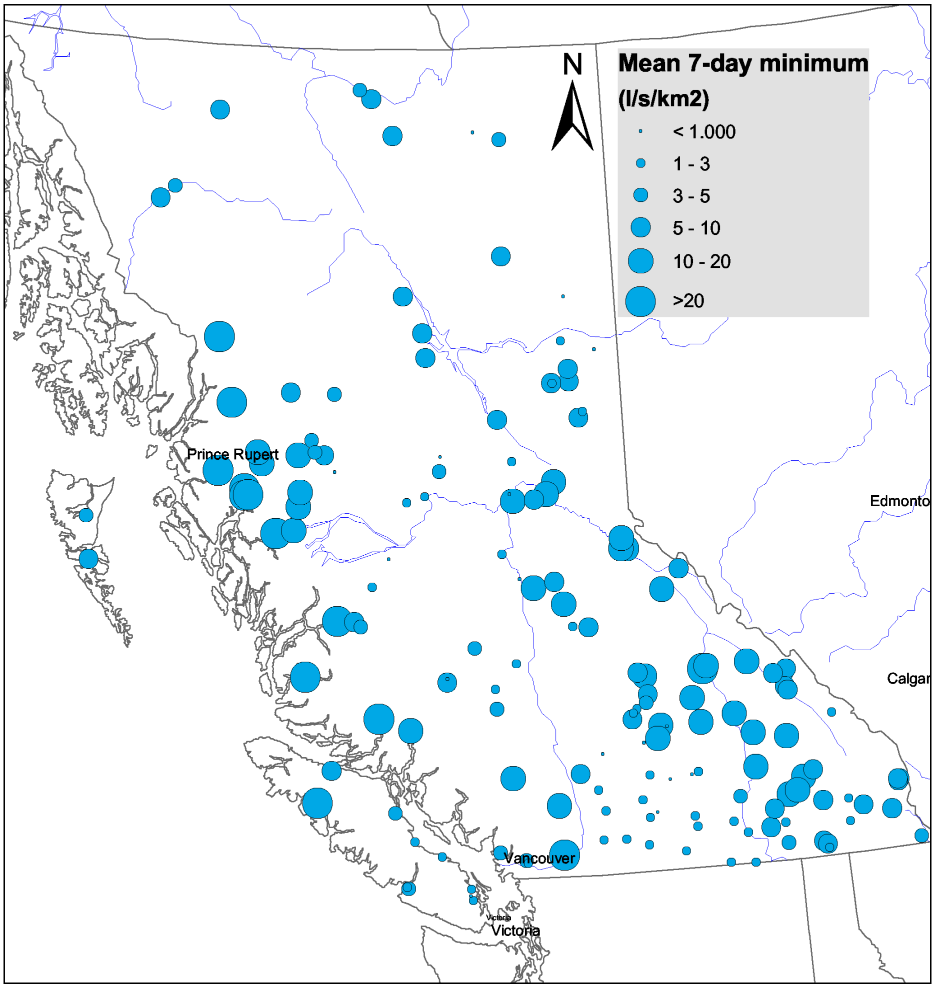


Figure 2.2 Specific discharge of the mean 7-day minimum flow from April to October in BC.

In summary, the following are the dominant factors influencing the magnitude and timing of low flows in British Columbia:

- Sub-freezing temperatures in winter (except near coast, and depending on elevation)
- Climatic moisture deficit in summer
- Snowpack and snowmelt (amount, timing and duration of melt, depending on elevation)
- Limited storage in shallow soils and aquifers and at high elevation

- Augmentation of summer flow by glacier melt (and groundwater) during warm/dry summer season

2.2.3 Environmental and socio-economic impacts of low flows

While dilution of effluents from industry, mining, etc., can be a problem during both winter and summer low flows, environmental and socio-economic impacts of reduced flows and degraded water quality tend to be greatest during the warm and dry season. The salmon fishery is an important industry in British Columbia and an economic and cultural pillar for many First Nations. Low water levels and high stream temperatures, however, threaten the habitat for cold-water species such as salmonids. In dammed rivers, reservoir operation has to be adjusted during low flows to provide certain in-stream flow targets, mainly for fish but also for other aquatic species. Meeting these targets can compromise hydropower production during summer. In the wetter parts of the Cordillera, for example along the coast, municipal water supplies as well as irrigation facilities rely on inflows during the dry season. In years with exceptionally low flows during summer, communities may need to impose water restrictions. Given the broad range of impacts associated with summer low flows, we focused in this study on late-summer streamflow and how it is influenced by climate, groundwater and glaciers.

2.2.4 Influences of climate variability, climate change and land-use change on summer low flows

Climate warming has caused concerns over various seasonal streamflow characteristics in the Canadian Cordillera. Trend analyses have shown that while the warmer winters of the last decades have increased the number of mid-winter melt events and have therefore directly increased winter low flows in hybrid (pluvial-nival) regimes, summer low flows have mainly decreased (e.g., Whitfield, 2001, Burn and Hag Elnur, 2002). This decrease in summer low flows was interpreted as the result of an extended summer recession period due to lower snowpacks and earlier snowmelt.

Streamflow in the southern portion of the Cordillera also fluctuates with the Pacific Decadal Oscillation (PDO) and the El Niño Southern Oscillation (ENSO). Wang et al. (2006) showed that (relative) low flows across the south and in the northern interior of British Columbia were lower and more frequent during the PDO warm phase and during El Niño events. For these reasons, the predicted continuation of climate warming as well as a potential increase in El Niño/positive PDO events in the future, can be expected to increase the occurrence of severe summer low flows. A modelling experiment in the Georgia Basin by Whitfield et al. (2003) demonstrated an extension of the summer low flow period in pluvial systems and a shift of the primary low flow time from winter to summer in hybrid (nival/pluvial) systems in response to future climate scenarios. Combined with an increasing water demand due to a rapidly growing population this is a cause of concern in particular for the southern part of the Canadian Cordillera.

Besides snow and direct summer climate, long term storage is important, in particular groundwater and glaciers. In some areas, however, groundwater is increasingly exploited to meet rising irrigation demands, and glaciers have been receding as a result of global warming.

The major land use change in the Cordillera is forest harvesting. Historically, clearcutting was the dominant forest practice, but variable retention methods are increasingly popular (Beese et al., 2003). Moore and Wondzell (2005) reviewed the effects of forest harvesting on streamflow. Most studies have found that late-summer flows increase for the first 5 to 10 years following harvesting, due to the reduction in summer transpiration. However, there is some evidence that regrowth of vegetation can increase transpiration, resulting in more extreme low flows, particularly if deciduous species become established in the riparian zone (Hicks et al., 1991). However, there is a limited base of empirical research to draw upon, particularly for British Columbia conditions. It is therefore difficult to separate the impact of forest management on streamflow from that of climate and storage changes.

2.3 Measurability of low flows

2.3.1 Introduction

Low flows are difficult to measure accurately due to equipment limitations and the nature of the flow. In order to understand the relationship between the surface water and groundwater, it is imperative to obtain accurate measurements of stream discharge and to relate these to the regional hydrogeology. It is also important to assess the range and magnitude of uncertainties that arise during low flow conditions, and how these may change with flow conditions and time. Without accurate measurements of discharge at low flows, it will become increasingly difficult to understand the potential impact of climate variability and climate change on stream flow, as well as to assess in-stream flow requirements and the nature of interaction between surface water and groundwater.

A field experiment was conducted in the Abbotsford aquifer in southwest BC (Figure 2.3) to assess the repeatability (significance of measurement error) of low flow measurements. Discharge was measured through the low flow season (June to September 2005) in two streams to quantify variability across a series of cross-sections, both repeated along a single cross-section (at-a-station measurements), and along multiple consecutive cross-sections from upstream to downstream (downstream measurements). The results have are being published as a portion of a more comprehensive case study on the Abbotsford Aquifer by Berg and Allen (in press).

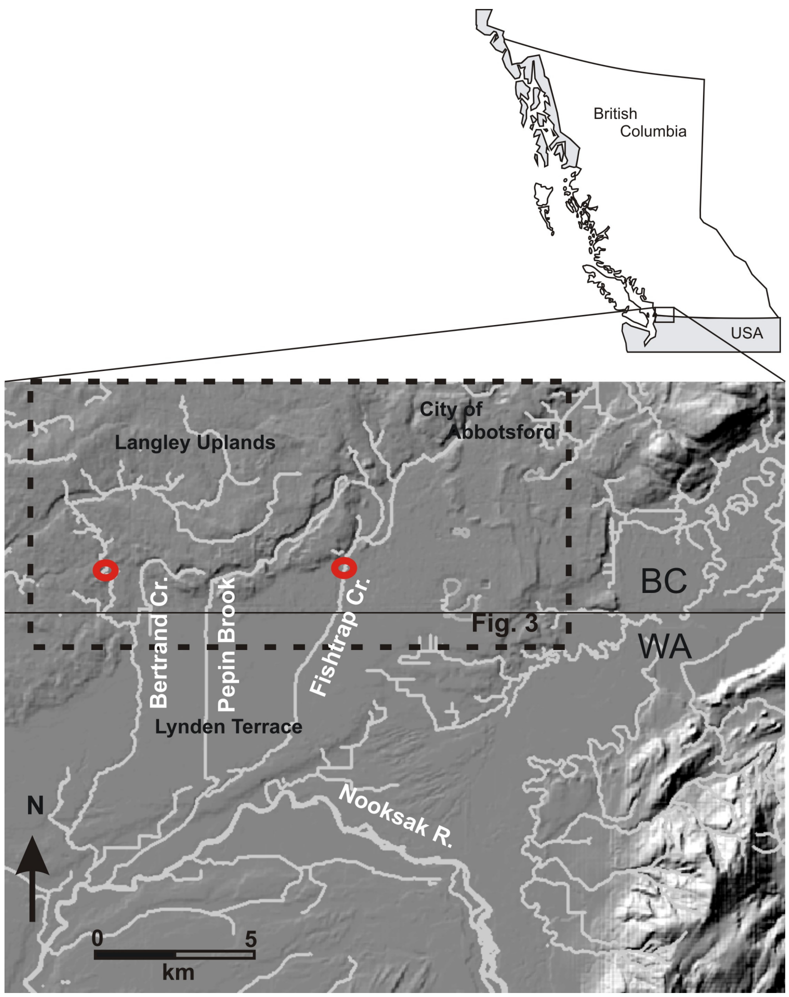


Figure 2.3: Location of study area in southwest British Columbia, Canada. Shown are Bertrand Creek, Pepin Brook and Fishtrap Creek north of the international boundary between BC, Canada

and the WA, USA. Circled are the two study sites for the low flow repeatability experiment. (Berg and Allen, in press)

2.3.2 Field measurements and analysis

Discharge measurements were conducted at approximately one month intervals over the low flow period. The sampling period was from June through September, 2005, and captured a variety of flow conditions. Due to unusually high precipitation during the month of June, typical low flows were not observed until August, and were only observed during that one month.

Environment Canada staff provided training for discharge measurements, and surveying was conducted according to Water Survey of Canada standards. The streams in this study had depths of less than 1.0 m in the sections being sampled. Velocity measurements were taken using a Scientific Instruments 1250 Mini "pygmy" current meter. The pygmy meter is specially designed for use in small, shallow streams; the range of operation is roughly 0.03 to 1.5 m/s. Measured velocities occasionally dropped to below 0.03 m/s, but often these were associated with pooled water, and these measurements were excluded. The average stream velocity ranged from 0.02 m/s to 0.16 m/s, with an average of 0.07 m/s. These velocities generally fall within the optimal range of operating conditions for this particular meter.

The method used to calculate the flow was the sixth-tenths depth method, and the velocity at that point assumed to approximate the average velocity in that section of the channel (Dingman, 2002). The streams were thus sampled using the velocity-area method, in which the width, depth, and velocity are measured and used to calculate the discharge. The width of the channel was measured for each cross-section, and divided into twenty evenly spaced sections. The flow measurement was taken at the mid-point of each of these vertical sections to represent an approximate average of the depth and average velocity in the panel.

The low flow repeatability survey was comprised of two parts; each with a specific purpose. The first part was intended to measure the spatial variability in flow at within a 6m length of stream, such that the variability introduced by channel morphology variations could be assessed. Discharge was measured across five cross-sections, situated 1.5 m apart and oriented perpendicular to the flow. Measurements began at an upstream cross-section and continued downstream (hereafter referred to as downstream measurements). Figure 2.4 illustrates the geometry of the survey. The second part aimed to measure the variability in flow when repeated measurements were made across the middle cross-section (hereafter referred to as were the at-a-station measurement). In this case, the middle cross-section was re-sampled four times, for a total of five measurements (Figure 2.4).

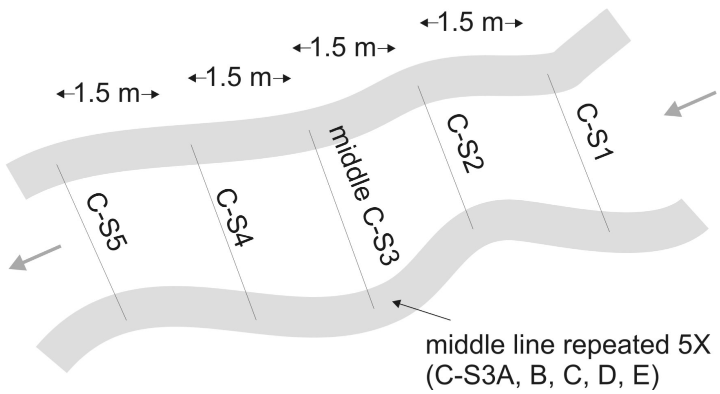


Figure 2.4: Schematic diagram illustrating the locations of cross-sections for the repeatability surveys. Downstream cross-sections C-S1, C-S2, C-S3, C-S4, C-S5), are labeled sequentially downstream, while at-a-station cross-sections across the middle line are labeled (C-S3A, B, C, D, E). (Berg and Allen, in press)

Low flow repeatability values were assessed for the mean flow, the standard error and the coefficient of variation. These calculations were complied separately for the downstream and at-a-station measurements. In both cases, the magnitude of the variation in the calculated discharge was assessed to determine the effect of decreasing flows.

2.3.3 Results

The calculated stream discharge for each cross-section, along with the mean, standard deviation (S.D.), standard error (S.E.), and coefficient of variation (C.V.), are summarized for the downstream measurements and the at-a-station measurements for the Bertrand Creek site (Tables 2.1 and 2.2, respectively), and the Fishtrap Creek site (Tables 2.1 and 2.2, respectively). Missing data for Bertrand Creek reflect values that were highly questionable due to eddies created from vegetation (negative velocities).

Table 2.1: Summary table for the low flow discharge measurements for Bertrand Creek. Cross-section (C-S) locations shown schematically in Figure 2.4. Shading highlights the middle cross-section that was sampled repeatedly for at-a-station measurements (Table 2.2). Missing data due to questionable field results arising from overgrowth.

	Downstream Discharge (m^3/s)			
	15-Jun-05	14-Jul-05	11-Aug-05	10-Sep-05
C-S1	0.077	0.087	0.027	0.023
C-S2	-	-	0.027	0.022
C-S3	0.066	0.078	0.026	0.020
C-S4	0.067	0.082	0.032	0.020
C-S5	0.069	0.055	0.031	0.017
Mean	0.070	0.082	0.028	0.021
S.D.[1]	0.005	0.005	0.003	0.002
S.E.[2]	0.003	0.003	0.001	0.001
C.V.[3] %	7.1	6.1	10.7	9.5

[1] S.D.=standard deviation
[2] S.E. = standard error
[3] C.V.= coefficient of variation

Table 2.2: Summary table for the low flow discharge measurements for Bertrand Creek. Cross-section (C-S) locations shown schematically in Figure 2.4.

	At-a-Station Discharge (m^3/s)			
	15-Jun-05	14-Jul-05	11-Aug-05	10-Sep-05
C-S3A	0.066	0.078	0.026	0.020
C-S3B	0.063	0.087	0.026	0.022
C-S3C	0.062	0.075	0.022	0.017
C-S3D	0.067	0.063	0.029	0.020
C-S3E	0.061	0.074	0.029	0.020
Mean	0.064	0.076	0.027	0.020
S.D.1	0.002	0.009	0.003	0.002
S.E.2	0.001	0.004	0.001	0.001
C.V.3 %	3.1	11.8	11.1	10.0

[1] S.D.=standard deviation
[2] S.E. = standard error
[3] C.V.= coefficient of variation

Table 2.3: Summary table for downstream discharge measurements Fishtrap Creek. Cross-section (C-S) locations shown schematically in Figure 2.4. Shading highlights the middle cross-section that was sampled repeatedly for at-a-station measurements (Table 2.4)

	Downstream Discharge (m^3/s)			
	15-Jun-05	14-Jul-05	11-Aug-05	10-Sep-05
C-S1	0.233	0.159	0.089	0.102
C-S2	0.348	0.184	0.080	0.093
C-S3	0.329	0.206	0.073	0.085
C-S4	0.317	0.314	0.081	0.091
C-S5	0.312	0.311	0.070	0.083
Mean	0.308	0.235	0.078	0.091
S.D.1	0.044	0.073	0.007	0.008
S.E.2	0.020	0.033	0.003	0.004
C.V.3 %	14.3	31.1	9.0	8.8

[1] S.D.=standard deviation
[2] S.E. = standard error
[3] C.V.= coefficient of variation

Table 2.4: Summary table for at-a-station discharge measurements for Fishtrap Creek. Cross-section (C-S) locations shown schematically in Figure 2.4.

	At-a-Station Discharge (m^3/s)			
	15-Jun-05	14-Jul-05	11-Aug-05	10-Sep-05
C-S3A	0.329	0.206	0.073	0.085
C-S3B	0.341	0.207	0.077	0.082
C-S3C	0.315	0.214	0.080	0.091
C-S3D	0.323	0.208	0.085	0.086
C-S3E	0.347	0.206	0.082	0.087
Mean	0.331	0.208	0.079	0.086
S.D.[1]	0.013	0.004	0.005	0.003
S.E.[2]	0.006	0.002	0.002	0.001
C.V.[3] %	3.9	1.9	6.3	3.5

[1] S.D.=standard deviation
[2] S.E. = standard error
[3] C.V.= coefficient of variation

Bertrand Creek

Over the entire measurement period, the maximum discharge at the Bertrand Creek site was $0.087 m^3/s$ (C-S1: July) and the minimum was $0.017 m^3/s$ (C-S3C: September). Overall, discharge decreased from June to September, as evidenced by the decreasing trend mean values; however, the discharge in July was noticeably higher.

Differences between downstream measurements during individual sampling periods ranged from $0.006 m^3/s$ to $0.011 m^3/s$, with an average difference of $0.008 m^3/s$. Downstream measurements exhibit a positive correlation between standard error and mean discharge (slope of 0.04 and an R^2 = 0.86) (excluding July data) (Figure 2.5a). These results suggest that there is sufficient spatial variability within a 6.0 m span between the five transects to influence the measured flow. To better illustrate the relation between discharge rate and degree of repeatability, the coefficient of variation for the discharge was plotted against the mean discharge (Figure 2.5b). For downstream discharge, there was a negative correlation (slope of62.5 and R^2 = 0.93).

The average variation in the at-a-station measurements was 0.010m^3/s. In contrast to the downstream measurements, at-a-station measurements show only a weak positive correlation between the standard deviation and the mean (slope of 0.04 and R^2 = 0.52) (excluding July data) (Figure 2.5a). The coefficients of variation are of similar magnitude to those of the downstream measurements, but with a stronger negative correlation (slope of27, R^2 = 0.05) (Figure 2.5b).

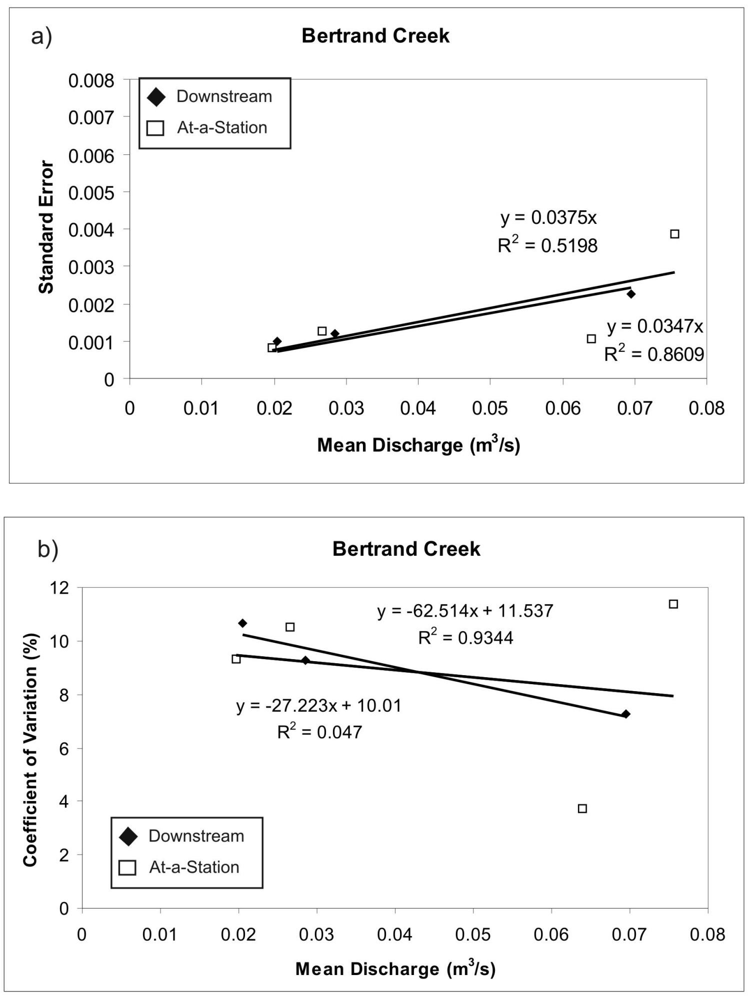


Figure 2.5: (a) Standard error and (b) coefficient of variation (%) as a function of mean discharge for Bertrand Creek measured monthly from June to September, 2005. Downstream measurements and at-a-station measurements shown. (Berg and Allen, in press).

Fishtrap Creek

Over the entire measurement period, the maximum discharge was 0.348m^3/s and the minimum discharge was 0.070m^3/s. There was a consistent decreasing trend in discharge from June to August, as evidenced by the decreasing mean values. Discharge in September increased relative to August.

Differences between downstream measurements during individual sampling periods ranged from 0.016m^3/s to 0.155m^3/s, with an average difference of 0.076m^3/s. The standard error increased with the mean discharge (slope of 0.09 and R^2 = 0.62) (Figure 2.6a), and there is a weak positive correlation between the coefficient of variation and mean discharge (slope of 52 and R^2 = 0.31) (Figure 2.6b).

The average difference between the at-a-station measurements was 0.015m^3/s. The standard error increased with the mean discharge (slope of 0.02 and R^2 = 0.70) (Figure 2.6a), but there is a weak negative correlation between the coefficient of variation and mean discharge (slope of5 and R^2 = 0.15) (Figure 2.6b).

Overall, during the measurement period, the flow levels decreased in both streams by the same relative amount, with both streams experiencing a decrease of 74% in the discharge between July (the month with the highest discharge) and August (the month with the lowest discharge). As well, the discharge in Bertrand Creek was overall considerably lower at all measurement times (by approximately 75% less) than the discharge in Fishtrap Creek.

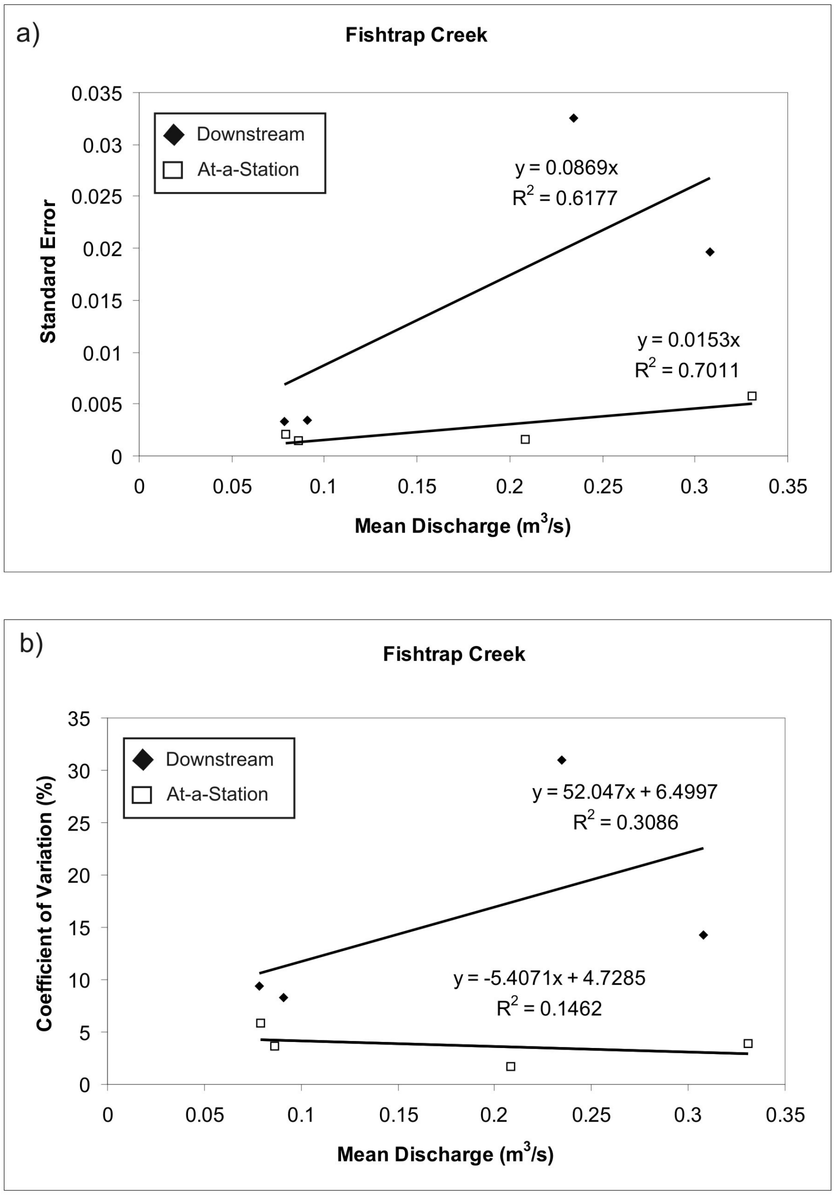


Figure 2.6: (a) Standard error and (b) coefficient of variation (%) as a function of mean discharge for Fishtrap Creek for the monthly visits from June to September 2005. Downstream measurements and at-a-station measurements shown. Berg and Allen (in press).

2.3.4 Discussion

Low flows in Fishtrap Creek, Bertrand Creek, and Pepin Brook occur during the summer months, during the period of least precipitation. During this time, the primary contribution to streamflow is groundwater.

The results of the repeatability experiment indicate that for both Bertrand and Fishtrap Creeks, the standard error decreased as the flow decreased. The degree of change in the standard error was greater for the downstream measurements, as might be expected due to the variability in the channel, including factors such as bed sediment, roughness, in-stream vegetation, groundwater contributions, and changes in channel dimensions. The variability measured at downstream sites was on the order of 7 to 30%, which is over a sampling distance of only 6.0 m.

In contrast, the variability measured at at-a-station sites was on the order of 3 to 10%. Because the measurements are taken at the same location, the results are less influenced by channel variations. The gradual changes in at-a-station measurement made during the day are subtle, and may reflect a combination of simple instrument/measurement error or perhaps diurnal variation. As discussed earlier, the range of operation of the pygmy meter is roughly 0.03 to 1.5 m/s, and measured velocities ranged from 0.02 m/s to 0.16 m/s, with an average of 0.07 m/s. However, velocities did drop below 0.02 m/s, making measurement difficult and uncertain. As well, in some cases the flow was reversed due to small eddies caused by vegetation or debris. In addition, however, diurnal variation can result in small variations in the discharge and, during a daily sampling period, these small variations may account for observed variations. Figure 2.7 shows 15-minute discharge at Bertrand and Fishtrap Creeks for the July 11-18, 2004 period (Environment Canada, 2005). The cause of these diurnal variations is unclear. At Bertrand the magnitude of the diurnal variation is <0.01 m^3/s, which is greater than the standard deviation for both downstream and at-a-station measurements at Bertrand. At Fishtrap, the diurnal variation is as high as 0.03 m^3/s, which is greater than the standard deviation for August and September downstream measurements and all at-a-station measurements. These results suggest that diurnal variations should be considered a factor influencing discharge variability.

The relation between the coefficients of variation and discharge was not consistent between the two streams studied, likely due to the fact that discharge in Fishtrap Creek was substantially higher than that of Bertrand Creek during the low flow season. At Bertrand Creek, there was a strong negative correlation between the coefficient of variation and discharge for the downstream results, and a weak negative correlation for the at-a-station results. At Fishtrap Creek, the correlation was similarly negative for at-a-station results, but positive for downstream results, although the correlations in both cases were weak. Although somewhat inconclusive, the results generally suggest that as the flow in these small streams decreases, the coefficient of variation of the mean discharge tends to increase when the flow is very low (i.e., $<0.1\ m^3/s$), indicating that, as a proportion of the flow, the error is more significant at lower flows. This suggests that repeat measurements should be made during the low flow period to provide a more realistic measure of the error, and to increase the confidence in the measured values. Measurements should also be taken at locations that are consistent from one time to the next, which is simply good practice. In the case of low flow measurements, variations in cross-section location can lead to errors that are measurably higher than those of instrument error alone, even over relatively small distances.

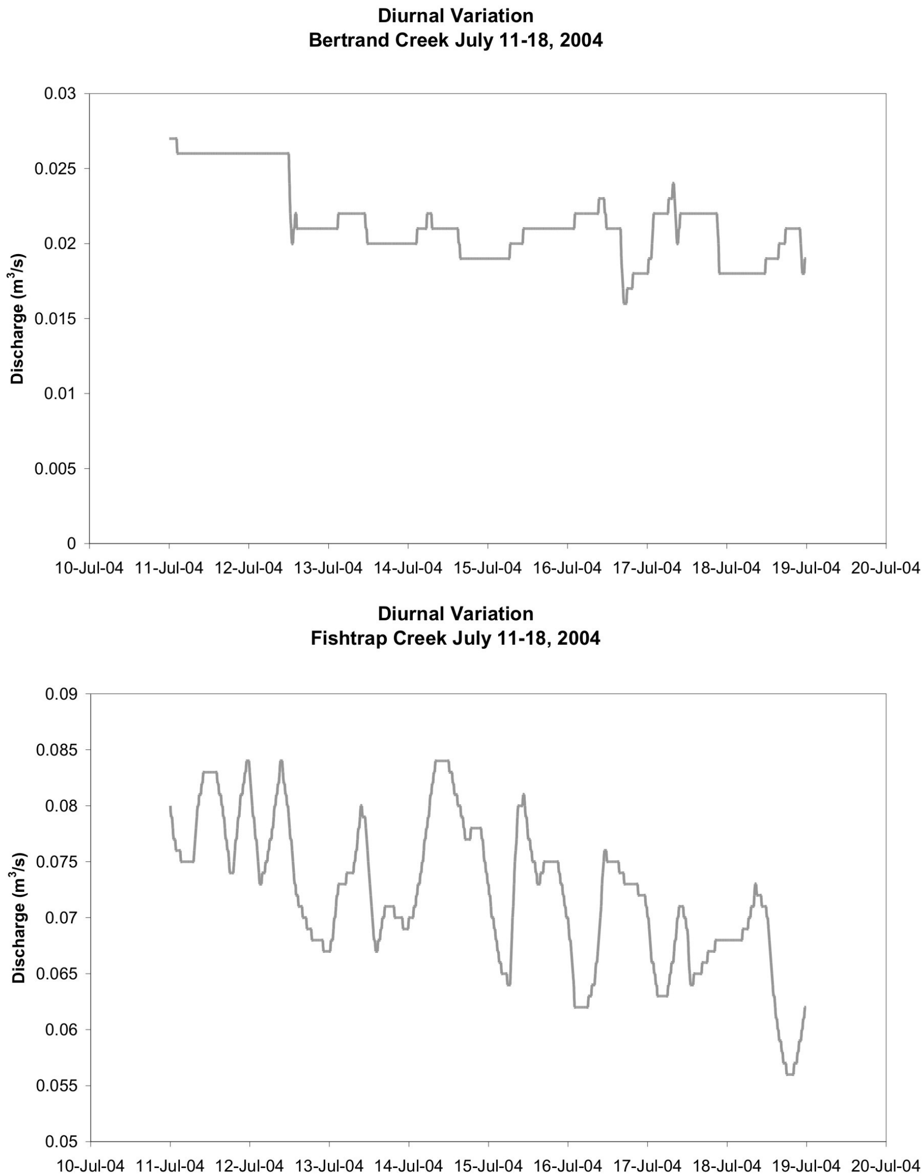


Figure 2.6: Diurnal variations evident from 15-minute discharge measured at Bertrand and Fishtrap Creeks over the period July 11-18, 2004. Berg and Allen (in press).

2.3.6 Conclusions

Accurate assessments of the interactions between groundwater and surface water are essential for management strategies of water resources and for the management of ecologically sensitive habitat. Low flow discharge was found to be associated with higher levels of uncertainty due to a range of factors spanning instrument error, measurement error and possibly diurnal variations. This increase in uncertainty has a potentially negative impact for understanding stream dynamics, and particularly, in quantifying variations in the contributions of groundwater to streamflow over the low flow season. Thus, repeated measurements of stream discharge at the same cross-section location should be done to limit uncertainty in discharge estimates.

2.4 Historical variations in summer climate and streamflow

2.4.1 Introduction

As a first step in studying the effect of climatic variability on late-summer streamflow, we have conducted trend analyses for mean air temperature, monthly total precipitation and mean streamflow for the months of August and September. This section presents these analyses, beginning with a description of the data sources and method of analysis.

2.4.2 Data sources

The Water Survey of Canada's hydrometric network in British Columbia includes 236 gauging stations with 10 or more years of record and with no gaps of more than 6 consecutive years, natural flow conditions, and available land cover data. Of these stations, 113 gauged basins have some glacier coverage (from 0.015% to 61.7%) (Figure 2.7). The record length varies from 10 to 89 years. The land cover information for these basins was derived by Environment Canada from the base thematic map of British Columbia, using Landsat imagery from the 1990s. For analyses that are sensitive to the choice and length of the record, however, a subset of 153 records for the common series from 1976 to 2002 was used.

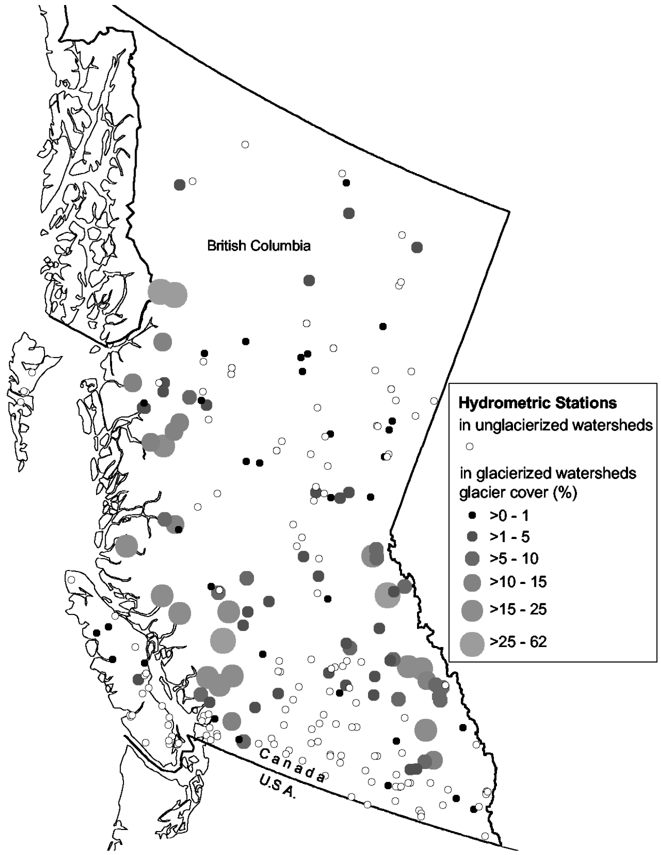


Figure 2.7 Study area and location of hydrometric and climate stations in British Columbia, Canada. Symbols for hydrometric stations also indicate the glacier coverage within the catchment.

A total of 43 climate stations with long records and few missing data from Environment Canada's weather station network were used to estimate summer air temperature and precipitation for the basins. For example, time series of August mean temperatures and August precipitation sums are highly correlated over the region, with average inter-station correlation coefficients of r = 0.89 for temperature and r = 0.76 for precipitation for stations less than 100 km apart. For most hydrometric stations there is a climate station within 100 km that can, therefore, provide a reasonable index of the interannual variations in summer temperature and precipitation.

2.4.3 Data analysis

The Spearman rank correlation coefficient (r_s) was used to detect trends because it does not require normality and can accommodate nonlinear trends. The test has similar power to the Mann-Kendall test (Yue et al., 2002), and has the advantage that the test statistic provides an intuitive indication of the direction and strength of the trend. Significance of a trend at a particular station (local significance) was assessed at $\alpha = 0.05$. The number of observations varies slightly among stations because of missing data, resulting in variations in the critical value of r_s. Data series were not pre-whitened (i.e. serial correlation removed) in light of the consequent reduction in power of trend tests (Fleming and Clarke, 2002) and the low serial correlation of August flows in our dataset.

2.4.4 Results

Trends in August streamflow are mixed, with most basins having weak trends, both positive and negative. However, two regions seem particularly consistent among non-glacierized basins: August flow increased on the Queen Charlotte Islands and decreased in south central BC (Figure 2.8). Glacierized basins exhibited a stronger tendency to decreasing trends, particularly in the Coast Mountains and in the Rockies. Note that basins with less than 1% glacier cover are included in both data sets.

Trends in September streamflow show a stronger spatial coherence, with almost exclusively negative trends in the southern half of the province in non-glacierized basins and in quite a few glacierized basins also, particularly in the Columbia and Rocky Mountains. Trends in September streamflow were dominated by negative trends for all catchments, except for those in a broad region including Haida Gwaii and extending WNW into the northern Rockies west of Williston Lake, where September streamflow increased. September flow increases also occurred in a number of glacier-fed catchments in the Coast Mountains, on both the coastal and inland sides of the heavily glacierized Mount Waddington region (Figure 2.9).

Precipitation during August and September has mostly decreased (Figure 2.10). The decreases are significant in southeastern BC for August and in September for southwestern BC. Temperatures in August have increased most notably in southern BC (Figure 2.11). It should be noted, however, that temperature increases become significant for most stations if years before 1976 are included in the analysis, reflecting a longer term warming trend.

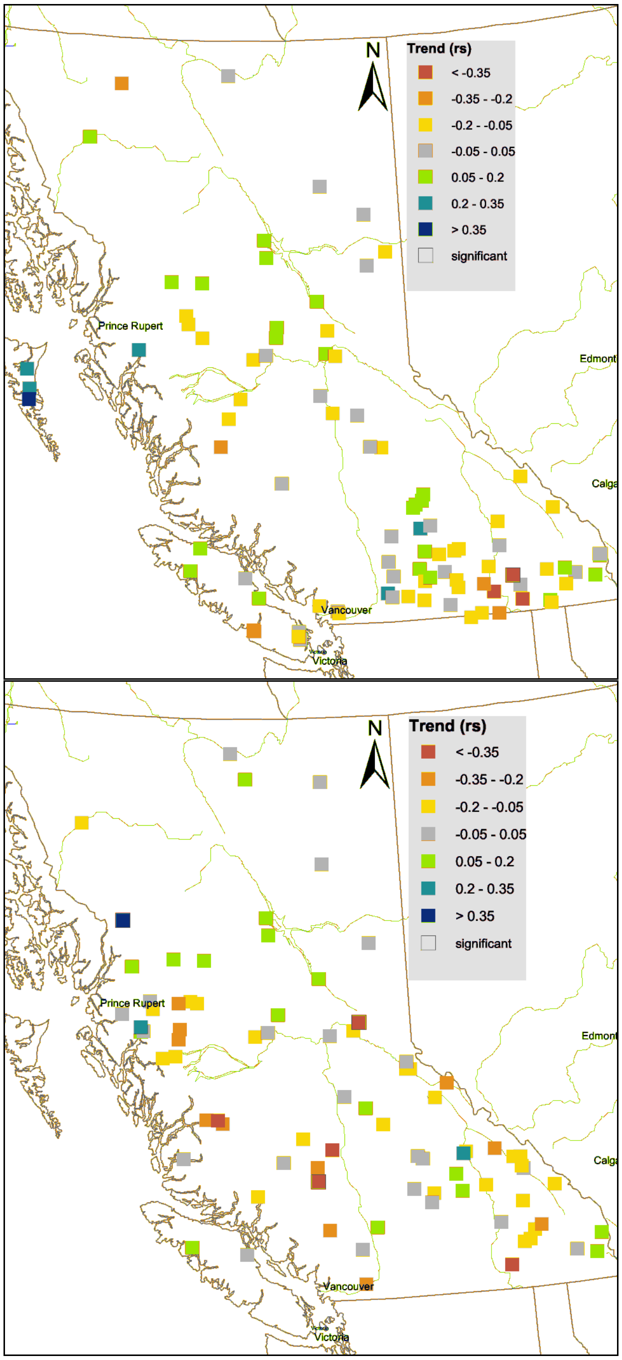


Figure 2.8 Trends in August streamflow for non-glacierized basins (top) and glacierized basins (bottom) for 1976-2002.

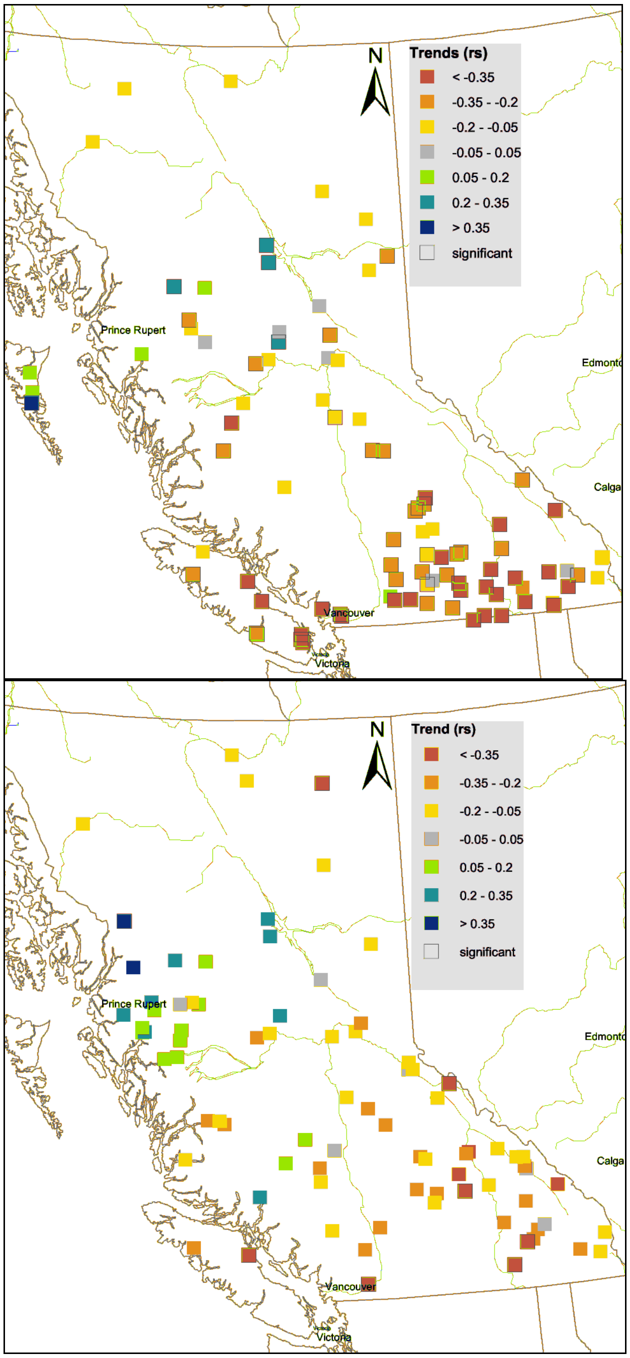


Figure 2.9 Trends in September streamflow for non-glacierized basins (top) and glacierized basins (bottom) for 1976-2002.

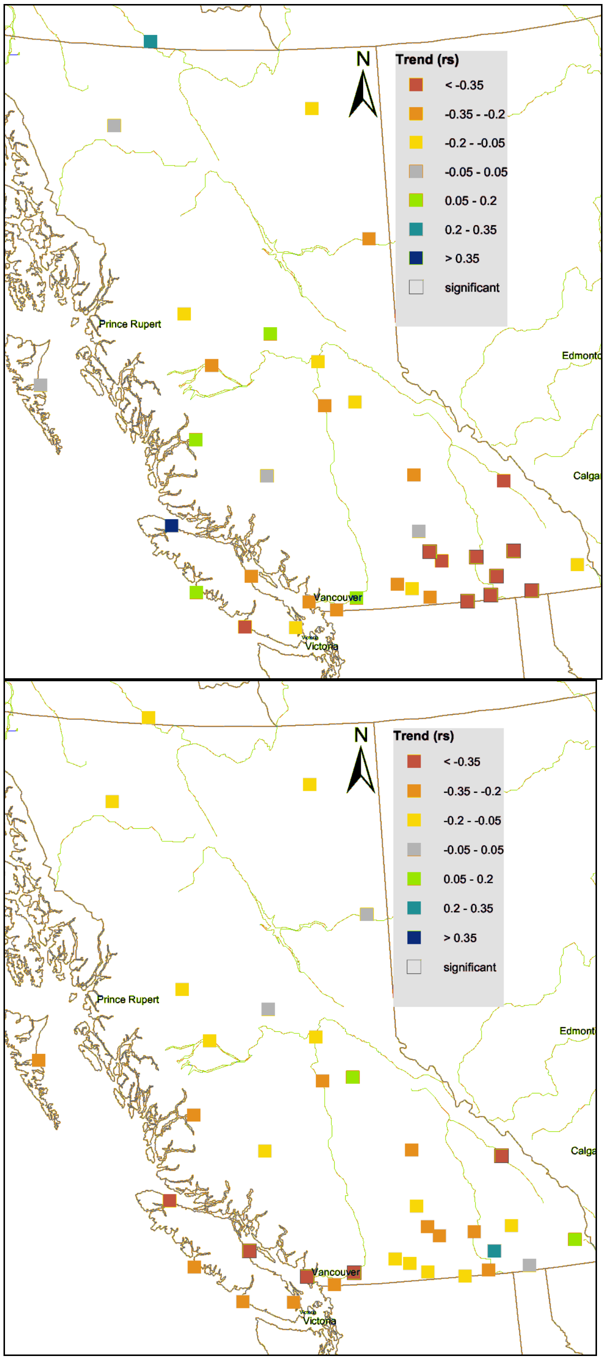


Figure 2.10 Precipitation trends in August (top) and September (bottom) for the period 1976-2002

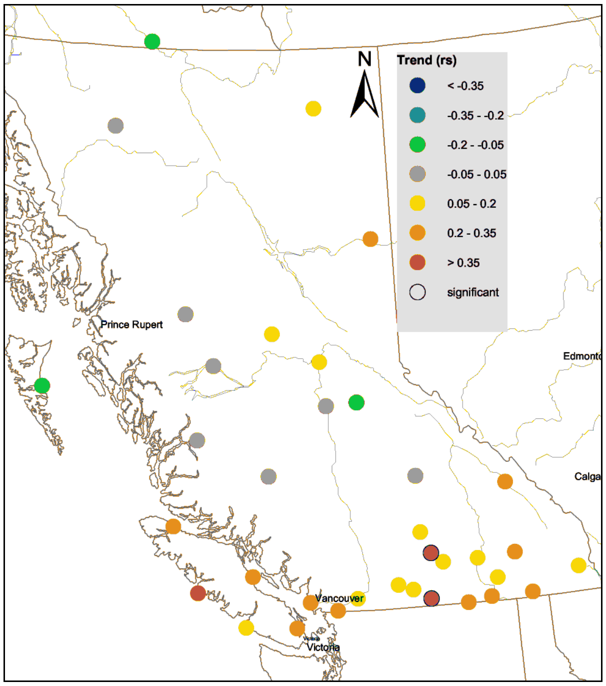


Figure 2.11 Temperature trends in August for the period 1976-2002.

2.4.5 Discussion and conclusions

The climate of British Columbia during August and September tended to become warmer and drier over the period 1976 to 2002. The predominantly negative trends in September streamflow can be attributed in large part to the negative precipitation trends in that month. However, August streamflow for unglacierized basins was not so consistent with this climatic trend: most basins tended to have only weak trends, positive and negative, possibly reflecting the effects of carry-over storage from the spring snowmelt or precipitation in preceding months. However, trends in August flow for glacier-fed catchments tended to be more negative than for the unglacierized catchments, possibly suggesting an influence of the tendency to more strongly negative mass balance and glacier retreat over the period (e.g., Moore and Demuth, 2001).

The next section will explore in more detail the linkages between climatic variability and streamflow through the use of statistical modelling.

2.5 *Sensitivity of late summer streamflow to direct climatic influences and long-term storage changes in groundwater and glaciers*

2.5.1 Introduction

Late-summer streamflow is influenced by factors operating at a range of time scales. These include direct influences of climate in the same year (preceding winter recharge, snow melt, and summer weather) as well as the influence of long-term (multi-year) changes in the storage of glaciers and groundwater. To assess the potential effects of climatic change on late-summer streamflow, it is necessary to combine (1) knowledge of the sensitivity of late-summer flows to hydroclimatic variables such as the timing and magnitude of the snowmelt freshet and post-freshet precipitation, and (2) projections of the changes in the relevant hydroclimatic variables. The sensitivities of late-summer flows to hydroclimatic drivers can be assessed empirically, by fitting statistical models to recorded data, or by application of a deterministic model. The advantage of using the empirical approach is that the streamflow response includes all processes actually acting within the catchment, while deterministic models only include those models understood by the model developer and that were deemed important enough to implement in the model code. The disadvantage is that the historical record may not include the full range of conditions and combinations of factors that may be encountered in the future. On the other hand, deterministic models can, in principle, can be run for arbitrary conditions.

In this section, we apply the empirical approach to assess the hydroclimatic sensitivity of late summer streamflow. In addition, we attempt to infer the influences of longer-term changes in groundwater and glacier storage by examining the temporal variation of the residuals. This method was successfully applied by Moore and Demuth (2001) to attribute decreasing streamflow at Place Creek to glacier retreat. Separate statistical models are fitted for each hydroclimatic regime (pluvial, nivo-pluvial, nival, nivo-glacial) in recognition that the hydroclimatic sensitivities of late summer flows will likely vary among the regime types.

2.5.2 Methods

Different models were applied for the three different hydrologic regimes found in BC. The models describe the interannual variability of August flow, $Q_{Aug}(t)$, as a function of their respective direct influences (x_i). The models can be expressed as

[2.1] $$Q_{Aug}(t) = b_0 + \Sigma\, b_i{\cdot}x_i(t) + d(t)$$

where b_0 is the intercept, b_i are the coefficients for the independent variables x_i representing direct influences, and the $d(t)$ is the residual, representing the sum of storage release from a longer-term reservoir, $g(t)$, and a random component, $e(t)$. That is, the residual can be represented as

[2.2] $$d(t) = g(t) + e(t)$$

Models were fitted to 153 streamflow records in BC that have data from 1976 to 2002. Climate variables were derived from the nearest climate station from a long-term climate dataset assembled by Stahl *et al*. (2006).

Assuming that surface water storages from the previous wet season have been depleted by late summer, flow in unglacierized, rain-dominated basins is considered to be influenced by precipitation in the same and previous months ($x_1 = P_{Aug}$ and $x_2 = P_{Jul}$). In addition, the influence of cumulative precipitation during the previous winter (Oct-Apr) recharge season ($x_3 = P_w$) is considered (Model 1).

In snow-dominated systems, the model included precipitation during the months of July and August (x_1, x_2), and the influence of the magnitude ($x_3 = FM$) and timing ($x_4 = FT$) of the spring snowmelt peak (Model 2). The peak flow and its day of occurrence during a given year were determined from the hydrograph after smoothing by an 11-day moving average. Both models were also fitted to the 10 records with hybrid regimes in the dataset (two peaks from snowmelt and fall/winter rain).

August flow in glacierized rivers is considered to be influenced by the following factors: rainfall-runoff from precipitation during the same month ($x_1 = P_{Aug}$); August temperature ($x_2 = T_{Aug}$), which controls the competing influences of ice melt in the glacierized part of the basin and evapotranspiration losses in the unglacierized part of the catchment; and carry-over surface storage from previous months, which is represented by streamflow in July ($x_3 = Q_{Jul}$) (Model 3).

As all variables are standardized, b_0 is 0 and the coefficients b_i (for i = 1 to 5) indicate the relative sensitivity to these different direct influences. If multi-year changes in groundwater or glacier storage have an important influence on late-summer streamflow, then the residuals should exhibit serial correlation and/or a longer term trend. The residuals were therefore analysed for trends and serial correlation to detect changes and/or multi-year persistence or memory in groundwater or glacier storage. Trends were calculated using the Spearman's rank correlation coefficient r_s (Yue et al., 2002) and multi-year persistence was assessed by a Durbin-Watson test and a runs test for randomness (McCuen, 2003). A significance level of $p < 0.1$ was used to determine local significance.

The following section first presents an overview of the results for all models followed by a detailed analysis and discussion of coefficients and residuals for each regime type and the respective model.

2.5.3 Results

Strength of model fits

The dataset includes only 10 rain-dominated basins, all found on the coast, and 8 hybrid regimes close to the coast. Most regimes are snowmelt dominated (76) or influenced by glacier cover within the watershed (82). There is some overlap between the datasets, as basins with less than 1% glacier cover were also included in the subset of snowmelt dominated basins. Table 2.5 summarizes the model results.

The best model fits were obtained for rain-dominated and glacierized basins, with an average of 68% and 66% of the variability explained, respectively. For 10% of the rain-dominated and glacier-fed basins, the models explain over 84% and 85% of the year-to-year variability of August streamflow, respectively. Streamflow for the snowmelt-dominated regimes is less well predicted, with an average $R^2 = 0.53$. For the hybrid regimes, the model that considers snowmelt peak and timing performed only slightly better than the model with winter precipitation.

Table 2.5 Overview of model results. Values for coefficients are means; the 10th and 90th percentiles are shown in brackets. The table shows the numbers of stations that exhibited statistically significant trends (positive and negative), as well the number for which the runs test found significantly non-random residuals.

Hydrologic regime	**n**	**R^2**	**Coefficients**				**Residuals**	
Rain dominated			*b_1 (P_{Aug})*	*b_2 (P_{Jul})*	*b_3 (P_{winter})*		*Trend (neg/pos)*	*Runs test*
Rain	10	0.68	0.71	0.22	0.14		0/5	0
		(0.42,0.84)	(0.47,0.92)	(0.01,0.51)	(-.06,0.33)			
Rain-snow hybrid	8	0.46	0.51	0.12	0.14		0/3	2
		(0.20,0.63)	(0.20,0.78)	(-.16,0.40)	(-.12,0.36)			
Snowmelt dominated			*b_1 (P_{Aug})*	*b_2 (P_{Jul})*	*b_3 (FM)*	*b_4 (FT)*		
Snow	76	0.53	0.44	0.26	0.27	0.12	3/2	21
		(0.34,0.72)	(0.12,0.71)	(0.05,0.54)	(0.01,0.60)	(-.13,0.37)		
Rain-snow hybrid	8	0.51	0.57	0.16	0.08	0.21	0/1	1
		(0.37,0.62)	(0.25,0.82)	(-.11,0.51)	(-.07,0.24)	(-.07,0.44)		
Glacierized			b_1 (P_{Aug})	b_2 (T_{Aug})	b_3 (Q_{Jul})			
	82	0.66	0.39	0.11	0.62		17/2	12
		(0.47,0.85)	(0.22,0.60)	(-.18,0.47)	(0.38,0.83)			

Model coefficients

Among the hypothesized influences considered in the analysis, August precipitation appears to be the most important in rain-dominated, hybrid and snowmelt-dominated basins, while July streamflow is the dominant influence in glacierized basins. The values for coefficient b_1 (August precipitation), which is the only variable used in all models, are mapped in Figure 2.12. Coefficients are highest in the coastal rain-dominated and hybrid basins. They are also high in the snowmelt-dominated basins of north-eastern BC and in the Columbia Mountains. August precipitation plays a comparatively weaker role in glacierized basins. Among the snowmelt variables, freshet magnitude seems to be somewhat less important in hybrid basins.

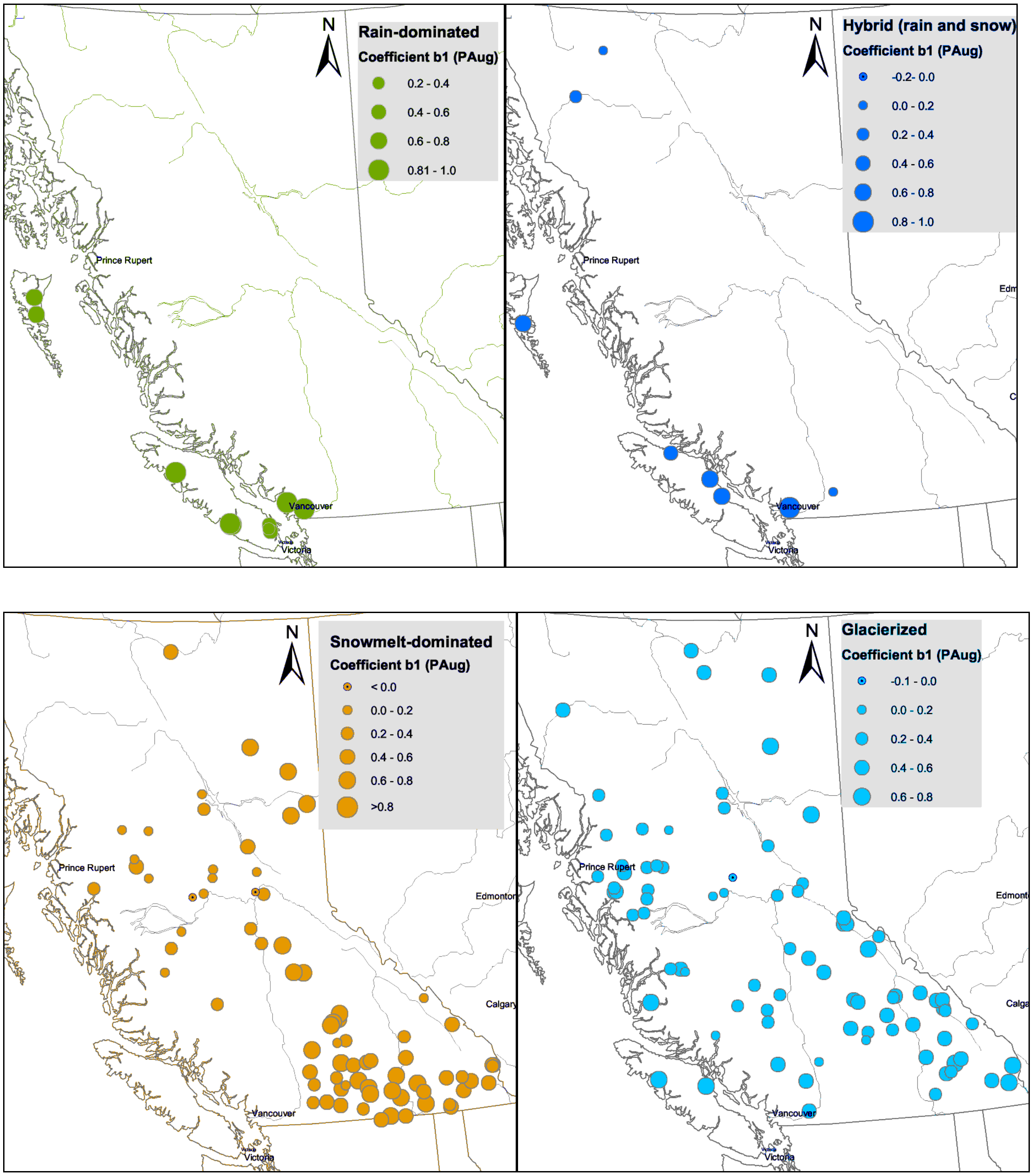


Figure 2.12 Influence of August precipitation, as indicated by the coefficient b_1, on August streamflow in the different models and subsets.

Figure 2.13 illustrates the spatial distributions of coefficient values for snowmelt-dominated catchments. August precipitation has a moderate to strong influence on August streamflow through most of the province, except the west central portion in the lee of the Coast Mountains. July precipitation is particularly important in the southern plateau areas and parts of the central interior. Freshet peak is an important influence in the west central portion of the province, and also has a less strong but spatially consistent influence across southern BC (Figure 2.13).

For the glacierized catchments, the coefficients for August temperature and July were systematically related to catchment glacier cover (Figure 2.14). For coefficient b_2 (August temperature), there was a positive, concave-down relation. There appears to be a threshold glacier coverage of about 3%, below which the coefficients tend to be dominantly negative, indicating that high August temperatures tend to depress August streamflow. Above 3% glacier coverage, the coefficient is dominantly positive, and increases with increasing glacier coverage. Coefficient b_3 (July streamflow) ranges from about 0.3 to over 0.9 (averaging about 0.6) for catchments with less than about 3% glacier coverage. Above that threshold, there is a clear negative relation between b_3 and catchment glacier coverage.

Trends

Statistically significant positive trends were found for the residuals for 5 of the 10 rain-dominated regimes, and 3 of the 8 hybrid regimes for model 1. Two hybrid regimes showed non-randomness as determined by the runs test. For the snowmelt-dominated basins, the number of significant positive and negative trends in the residuals is small and the Durbin-Watson test did not detect any significant positive autocorrelation in the residuals. However, the runs test found significant non-randomness for 21 stations, which suggests the existence of some multi-year persistence in the hydrological systems. In the residuals for the subset of 82 glacierized basins, 17 significant negative trends were found. The residuals of two records from basins with large lakes showed positive serial correlation and those of 12 records show non-randomness.

Figure 2.15 shows box plots of the Spearman's rank correlation coefficients calculated for August flow and the residuals. Trends for August streamflow are predominantly negative in all datasets, except the few hybrid basins. However, the patterns for residuals suggests that after correction for the direct impacts, the trends change to predominantly positive in rain and snowmelt-dominated regimes, while they become even more negative in glacial basins.

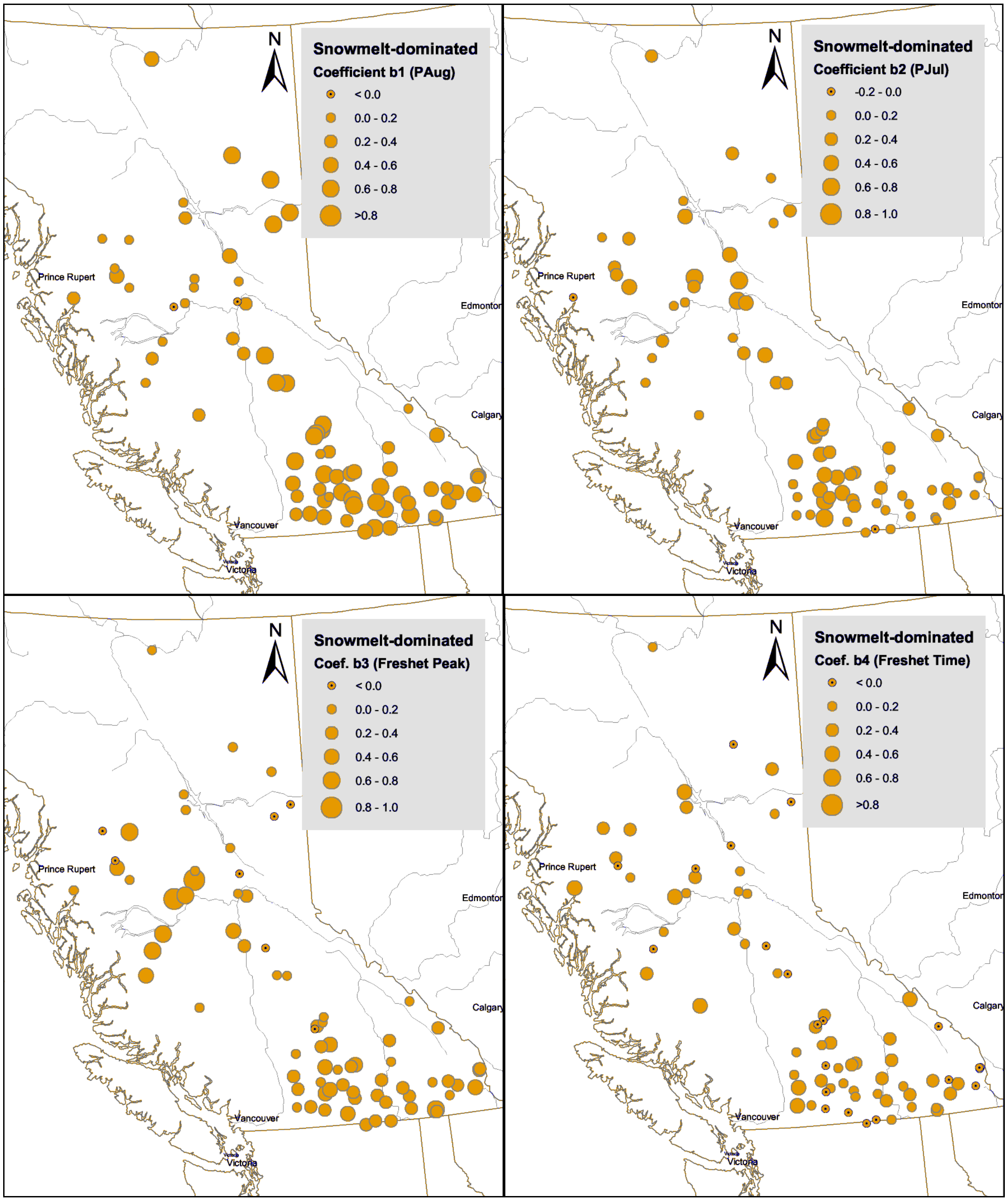


Figure 2.13 Relative influence of the different variables of the regression model for snow-dominated basins.

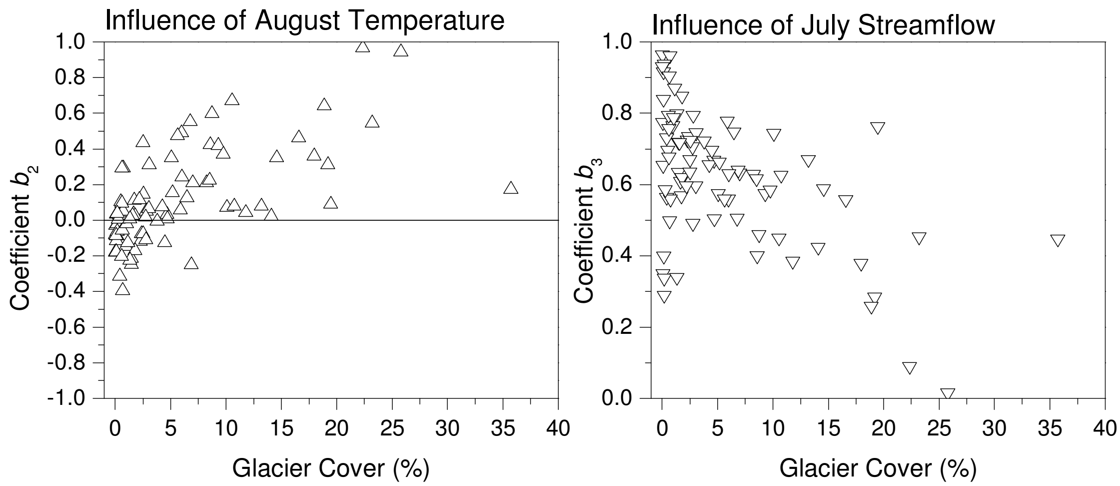


Figure 2.14 Variation of model coefficients with glacier cover.

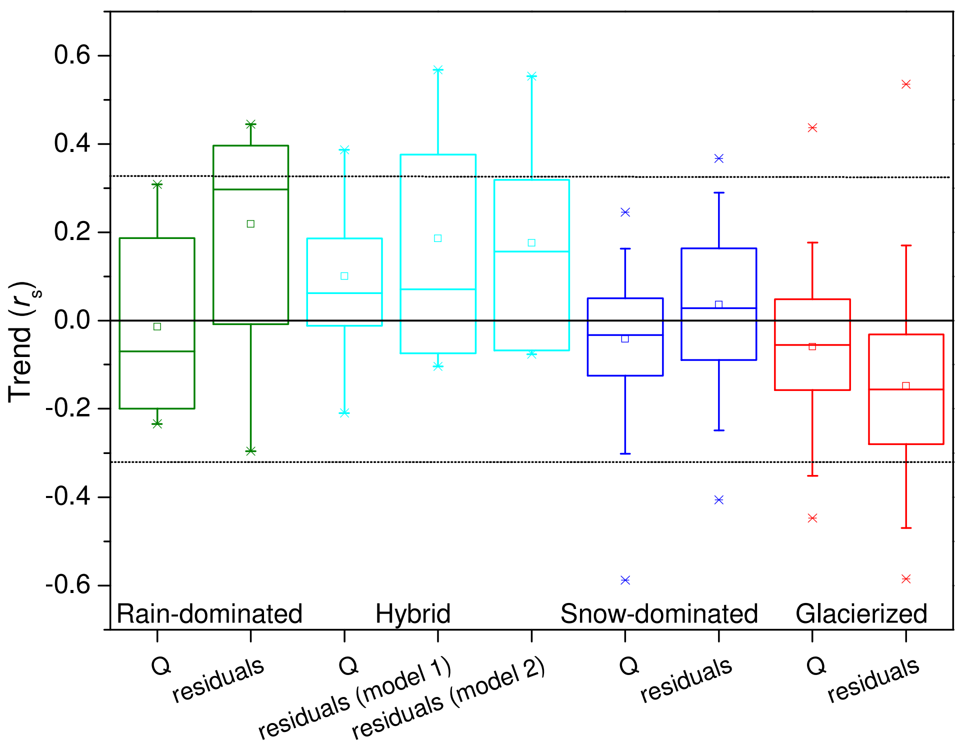


Figure 2.15 Box plots of trends for August streamflow and model residuals for the different regimes.

For comparison, Table 2.6 summarizes trend analyses on the independent variables in the regression models. For the climate variables, the numbers of stations reported for each regime is less than the number of hydrometric stations, because in many cases two or more hydrometric stations shared the closest climate station. Several significant negative trends were found for August precipitation. August temperature trends were positive for all climate records used for the subset of glacierized basins; however, only two were statistically significant. Freshet magnitude seems mainly to have increased, though the trend is only significant in 6 basins, whereas the results for the timing of the highest peak show similar numbers of increases and decreases.

Table 2.6 Trends of variables in models (significant negative/positive tests).

Hydrologic regime	Climate Variables					Streamflow derived variables			
	n	P_{Aug}	P_{Jul}	P_{winter}	T_{Aug}	n	Q_{Jul}	FM	FT
Rain dominated	8	1/0	0/0	0/0	-	10	-	-	-
Hybrid	10	0/1	0/0	0/2	-	10	-	0/0	0/0
Snowmelt	37	10/0	1/0	-	-	77	-	0/6	3/3
Glacier	31	4/0	-	-	0/2	81	0/1	-	-

Although most of the trends for August streamflow are not statistically significant, the majority are negative, especially for catchments with glacier cover greater than 3% (Figure 2.16). This tendency is stronger for the residuals, for which a larger number of stations exhibit statistically significant negative trends.

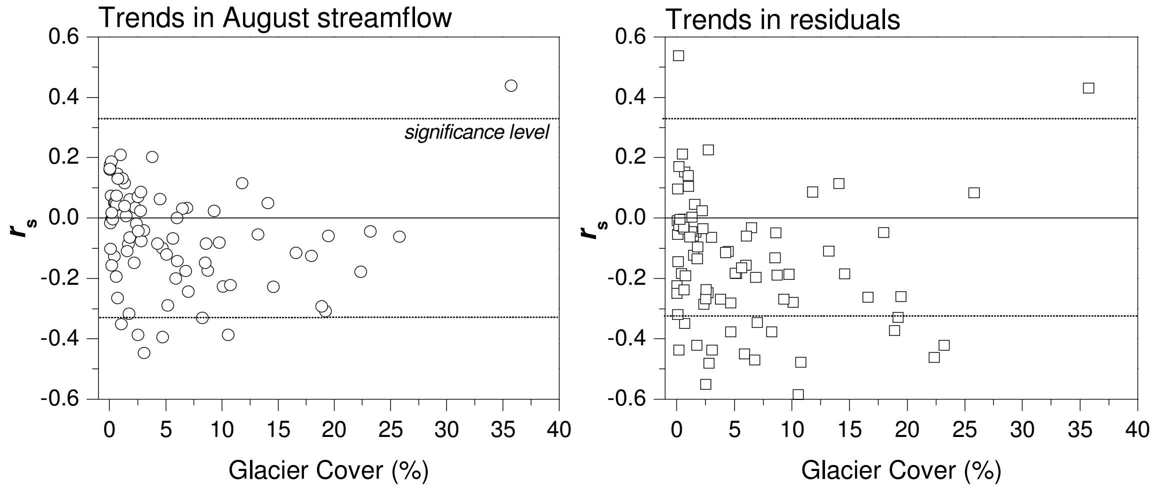


Figure 2.16 Relation between trends in August streamflow (left) and the residuals (right) and glacier cover.

2.5.4 Discussion

Direct influences of August weather

The coefficients in the regression models suggest that August streamflow is relatively sensitive to variations in August precipitation for all regime types, but especially so for rain-dominated catchments. This finding is particularly significant given that the fitted coefficients may not fully represent the strength of the relations, as in many cases the closest climate station may not be fully representative of precipitation over the catchment, especially for larger catchments in areas with sparse climate station coverage. It is commonly acknowledged that General Circulation Models perform less well at simulating precipitation than air temperature. This problem will likely be even more severe for summer precipitation, much of which is generated by convective uplift, because convection operates at spatial scales that are not resolved by GCM's. Therefore, it appears that our ability to predict August streamflow under future climate scenarios will be limited by the uncertainty in the projections of summer precipitation. The ongoing development of higher-resolution Regional Climate Models (RCM's) may help in this regard.

The dependence of coefficient b_2 (August air temperature) on glacier cover reveals the important moderating role of glaciers. For catchments with little glacier cover, the coefficient tends to be negative, reflecting the depressing effect of warmer weather on streamflow, caused by increased evapotranspiration and a tendency to lower precipitation (as air temperature and precipitation tend to be negatively correlated in summer). However, for catchments with more than about 3% glacier cover, the increased glacier melt during warm weather increases August streamflow. The considerable scatter in Figure 2.14 results from sampling variability of the regression coefficients (i.e., each value of bi is only an estimate of the "true" value for the catchment) as well as real differences in the coefficients among basins. For example, the sensitivity of August streamflow to August air temperature depends not just on glacier coverage, but also glacier hypsometry and aspect.

Lagged effects in the preceding year

The positive and generally high coefficients for July discharge in glacerized catchments suggest that seasonal carry-over storage plays an important role in controlling August streamflow. In some cases, this carry-over storage might be in the form of late-lying snow at high elevations, but it is more likely to reflect water originating as meltwater during the freshet, which recharged groundwater, lakes or wetlands, and then is being gradually released through the summer and autumn.

For catchments with more than about 3% glacier cover, the sensitivity of August streamflow to July streamflow generally decreased with increasing glacier cover. This relation probably results from the negative correlation between snow accumulation and glacier runoff (Fountain and Tangborn, 1985). In years with high snow accumulation, unglacierized portions of a catchment will tend to have high flows in July due to direct meltwater runoff from late-lying snow and delayed runoff from melt in previous months. However, on glaciers, a deeper snowpack will persist later in the season. Because snow has a higher albedo than glacier ice, the longer-persisting snowpack will reduce glacier melt by reflecting a greater portion of incident solar radiation, and thus offset to some extent the effect of carry-over storage from unglacierized portions of the catchment.

For snow-dominated basins, lagged contributions from rainfall and snowmelt were expressed through the relations with July precipitation and the magnitude of the snowmelt freshet, which tended to have similar influences on August streamflow. However, these lagged effects were less important than August rainfall. The timing of the snowmelt freshet had a weaker influence than the magnitude, perhaps because it tends to be correlated with freshet magnitude (i.e., later peaks tend to occur in years with higher snow accumulation).

In rain-dominated basins, precipitation during the preceding July and during the preceding had positive but weak influences on August streamflow. Hybrid basins similarly had positive but weak sensitivities to precipitation in July and in the preceding winter, but the relations tended to be weaker than for rain-dominated basins.

Multi-year storage effects

There was mixed evidence for multi-year storage effects in the residuals. The clearest signal was in the negative trends for the residuals for the glacierized catchments, which contrasted with the lack of overall trend for catchments without glacier cover. This contrast indicates that the trends are not simply related to the interannual variations in temperature or precipitation, but are associated with changes in the glaciers. Specifically, the negative trends in the residuals are consistent with declining glacier area, which would reduce the area of ice available for melting. The positive trends in the residuals found for northwest BC, which are consistent with findings of positive trends of annual streamflow in the St. Elias Mountains of southwest Yukon by Fleming and Clarke (2003), possibly suggest a regional contrast between southern and northern BC. A potential explanation is that the typically larger glaciers in northwest BC and the St. Elias Mountains exhibit a lag in their areal response to changes in mass balance.

Previous analyses of long streamflow records in Canada (e.g. Burn and Hag Elnur, 2002) have shown decreasing trends in most spring and summer months and attributed them mainly to earlier and lower spring freshets. The negative streamflow trends in late summer in glacier-fed catchments found in this study suggest that, should the current warming trend continue and glaciers continue to recede, summer streamflow in British Columbia will further decline.

Snowmelt-dominated catchments exhibited mixed signals. While the Spearman's correlation and Durbin-Watson tests found little evidence for multi-year trends or serial correlation, the runs test found significantly non-random behaviour in the residuals for 21 of 77 stations. Further research should examine this finding in more detail. It would be useful to subject these results to a test of field significance (Livezey and Chen, 1983; Burn and Hag Elnur, 2002).

The results for rain-dominated and hybrid catchments are puzzling. While the August flows themselves exhibited no significant trends, the residuals tended to have positive trends. If these trends were ascribed to changes in groundwater storage, it would suggest that there was a long-term increase in groundwater storage, which, in turn, would suggest that precipitation should be increasing and/or evapotranspiration decreasing. The climate trends in Table 2.6 show, at best, only weak support for this suggestion. Alternatively, if the groundwater response were very sluggish, an increasing trend could be a transient response to a step increase in recharge at some time in the past. However, analyses reported in section 3 indicate that many groundwater levels have been declining. It is, therefore, entirely possible, if not probable, that the positive trends in the residuals reflect an incomplete or incorrect model. For example, some of the relations may be nonlinear, a variable may be missing, or there may be interactions among the variables that are not incorporated into the model. We did attempt to include spring precipitation and also winter precipitation from the preceding winter in the model; neither changed the result. Another possible cause of spurious trends could be the choice of period analysed. For example, a few high precipitation amounts occurred at the end of the period of record. After extending the period to 1966-2002 (i.e., by 10 more years), trends were still positive, but not statistically significant. Further research should address whether these positive trends really do reflect changes in groundwater storage, and if so, what is the cause.

2.5.5 Conclusions

August precipitation is the most important control on August streamflow for all streamflow regimes in British Columbia. Because General Circulation Models cannot predict summer rainfall with great accuracy, projections of August streamflow under future climate scenarios will involve substantial uncertainty. August air temperature is important in glacierized catchments, as it is associated with increased glacier melt and higher flows. Seasonal lagged effects related to summer precipitation, snowmelt freshet magnitude and timing, and the previous winter's precipitation were important in all of the unglacierized regimes, but much less so than August precipitation.

The clearest indication of the effects of multi-year storage change were for glacierized catchments, where both the raw streamflow and the residuals from the statistical model exhibited dominantly negative trends. The trends were stronger and more highly significant for the residuals. These negative trends are consistent with the glacier retreat that has occurred over the last few decades, and which has reduced the area of glacier available for meltwater generation.

There was little indication of trend or autocorrelation for either August streamflow or the model residuals for the snowmelt-dominated catchments. However, there was some evidence of non-randomness in the residuals that should be examined further.

Both rain-dominated and hybrid streams exhibited little evidence of trend for August streamflow. However, the residuals tended to have positive trends. While it is possible and even likely that the trends are spurious, and related, for example, to some mis-specification of the regression model, it is possible that the trends are real and related to some ongoing increase in groundwater storage. Further research should examine data for individual stations in more detail to identify whether the trends are spurious and, if not, to identify the physical cause.

2.6 Summary

Streamflow patterns in BC can be broadly classified as rain-dominated, hybrid (rain and snow), snow-dominated, and glacierized, each having distinct low flow seasons. In this study, we chose to focus on the summer/early autumn low flow period, as that has particular economic and ecological significance.

Unglacierized catchments are generally exhibiting declining trends in September streamflow through most of British Columbia, but not in August. Glacierized catchments, on the other hand, show the opposite pattern, with dominantly negative trends in August but not September. The decreases in September flows are broadly consistent with a deline in September precipitation over the period studied.

For all four regimes, the most important control on August streamflow is August precipitation. August streamflow is also positively but more weakly related to lagged variables including July precipitation and the previous winter's precipitation.

There is clear evidence for the effect of multi-year storage change only for the glacierized catchments. Glacier retreat over the last few decades has apparently reduced glacier area sufficiently so as to reduce meltwater generation and thus streamflow. Snowmelt-dominated catchments showed no tendency to trends or serial correlation in the model residuals, though the runs test suggested a substantial number of stations had non-random residuals. Both rain-dominated and hybrid catchments tended to have positive residuals in August, suggesting either an increasing trend in groundwater storage or, more likely, a spurious cause such as model mis-specification.

This section has examined late-summer streamflow trends and attempted to draw inferences about the links to groundwater and glaciers. Sections 3 and 4 will focus directly on the behaviour of groundwater and glacier storage, and their linkages to streamflow.

3. Linkages between groundwater, low flows and climate

3.1. Introduction

As discussed in the introductory section to this report, British Columbia (BC) is one of the most hydro-climatically complex places in North America. It is provided with moisture and warmth by the Pacific Ocean and has steep terrain, which create strong precipitation and temperature gradients. Thus, the influence of climate variability and change is not homogeneous across the province (e.g., Whitfield and Cannon, 2000, Whitfield, 2001, Fleming et al., 2006, Kiffney et al., 2002, Regonda et al., 2005). The three predominant hydro-climatic zones in BC are pluvial, nival, and hybrid (a mixture of pluvial and nival) (Whitfield et al., 2002). Some regions in BC have also been defined as nivally supported-pluvial because water is primarily supplied by rainfall, but is supplemented by snowmelt (Fleming et al., 2006). This melt water is not sufficient to augment low flows in summer as might be the case in a hybrid or nival regime. Other areas are influenced not only by snowmelt but also by glacier melt. This water tends to arrive after snow pack has diminished, creating a second peak in the hydrograph that would otherwise have had only one peak in the spring from snowmelt.

Natural cycles of variability acting on the region are the El Nino Southern Oscillation (ENSO) (Shabbar et al., 1997; Cayan et al., 1999), the Pacific Decadal Oscillation (PDO), and the Pacific North American pattern (PNA) (Stahl et al. 2005). Cycles in PDO are thought to last for 20-30 years (Mantua et al., 1997) and those of ENSO are in the range of 2-5 years. A warm phase of the PDO started in the mid 70s (Mantua et al., 1997). Previous to that was a cool phase that started in the 40s. Consensus on the phase of the PDO presently affecting BC has not been reached.

In BC, variations in temperature and precipitation associated with ENSO (El Niño Southern Oscillation) and PDO (Pacific Decadal Oscillation) influence the amount and form of water that drives streamflow (e.g., Redmond and Koch, 1991). Although, the relationship between climate variability and change, and trends in precipitation, temperature, and streamflow have been tested, the limited record length of groundwater wells has largely prevented relationships between natural climate variability and groundwater levels to be rigorously tested for trends. Fleming and Quilty (2006) used climatological composite analysis to investigate ENSO signals in long-term groundwater level observations from four wells in the lower Fraser Valley of BC. ENSO precipitation impacts were largely limited to winter and spring, with higher lower rainfall occurring, respectively, under cold-phase and warm-phase episodes. Relative to the surface hydrologic systems considered in that study, the aquifers were seen to retain a strong memory of seasonal ENSO-related precipitation anomalies. Changes were potentially extended through to the following summer.

Over the longer term, climate change could have major regional effects on air temperature, precipitation, evapotranspiration, and ultimately runoff (e.g., Whitfield et al., 2003). Whitfield and Taylor (1998) found that streams in coastal areas of BC respond to small variations in climate. In recent decades, decreases in stream discharge during early spring and late summer, and increases in winter runoff have been observed (Whitfield and Taylor, 1998). Lower and extended low flow periods during the late summer and early autumn are of particular concern as they threaten not only water supplies, but also reduce effluent dilution, increase the likelihood of algal blooms, and damage wetlands and aquatic habitats.

With extended low flow periods, streams may be more strongly influenced by interaction with groundwater. Changes to the groundwater regime in respect to the timing and amount of natural recharge, greater groundwater use, and higher summer evapotranspiration could result in a lowering of groundwater levels in some areas. Therefore, low flows in the streams may be exacerbated by the decreasing groundwater levels. Consequently, during low flow periods, streamflow may become inadequate to meet needs for economic uses such as domestic consumption, irrigation and effluent dilution, as well as ecological functions such as in-stream habitat.

3.2. Purpose

Assessments concerning the future predicted impacts of climate change on groundwater have to-date largely been limited to modeling studies. Groundwater in the Grand Forks aquifer was found to be tied closely to streamflow in the Kettle River (Scibek et al., 2006). In that study, changes in streamflow (river stage) and groundwater recharge from precipitation were modeled for future climate periods and used as boundary conditions within a 3-dimensional groundwater flow model. Results suggest that variations in recharge to the aquifer under the different climate change scenarios have a much smaller impact on the groundwater system than changes in the river stage elevation of the Kettle and Granby Rivers (Scibek and Allen, 2006a). By the 2050s, the change in groundwater levels is projected to be less than 0.5 m away from floodplain, but can be greater than 0.5 m near the river. In a similar study, Scibek and Allen (2006b) found that groundwater recharge to the Abbotsford-Sumas Aquifer, in the Central Fraser Lowland of BC, Canada and Washington State, US, would decrease by 12.7 to 14.6 % by the 2050s relative to historic values, and could result in a reduction in groundwater flow to the various streams that drain the aquifer.

In these studies, and other studies that aim to predict the effects of changing groundwater levels (either by climate change or pumping) on streamflow, there lies an inherent limitation in respect of the boundary conditions used in the models. Currently, most models require that boundary conditions be assigned explicitly for surface water features so as to emulate the connection between the groundwater and surface water. While these boundary conditions can be changed temporally within the model (i.e., by invoking a shift in stream stage, or changing the discharge) to accommodate some external forcing, the boundary conditions are nonetheless assigned by the modeler and can give a false prediction of the true response of the surface water body (i.e., discharge or stage) to changes in the groundwater regime.

Thus, despite efforts to model such complex systems and make predictions on the potential consequences of climate change, there still remains uncertainty in the model predictions. This uncertainty can be gauged or evaluated by undertaking historical analysis of aquifer systems in which both surface water and groundwater respond to climate.

The aim of this component of the study is to explore the relations between groundwater fluctuations and past climatic variations using available climate data, hydrometric data, and data from the recording well network in BC. One of the objectives of the study is to gain insight on the mediating ability of groundwater to regulate summer low flows. This type of assessment is complicated due to the varied nature of the climate in different parts of the province, which determines the hydrologic regime (pluvial, nival, glacierized, mixed) and, perhaps more importantly, the complexity of groundwater system, which influences the nature of the connection between groundwater and surface water. Furthermore, within BC, records of groundwater levels extend back only to the 1970s. Although some 80,000+ wells are operated in BC (as reported in the well database, although this is likely an underestimate of the actual number), only a few of these are monitored (BC MoE, 2006). A limited few of these have been operated within aquifers that have maintained natural conditions throughout the length of their record.

3.3. Overview

This section begins by discussing the nature of groundwater-surface water interactions from a hydrogeological perspective. Interactions are classified in a broad sense according to Aquifer-Stream System type, using aquifer characteristics derived from a provincial aquifer classification system.

The responses of aquifer-stream systems to climate are then explored using data from selected high quality provincial observation wells and used to classify the wells according to Groundwater Response type based on nival, pluvial and mixed responses. This type of analysis enables defining those characteristics of the recharge and discharge, which separate the hydrogeologic effects of the site from those that are determined by the hydro-climatology. As such, an appropriate technique for defining the relationships between streamflow, precipitation, and groundwater can be selected and applied to interpret the responses at observation wells that might have incomplete records. The analysis also permits an evaluation of the dominant driver acting on the groundwater (be it precipitation, streamflow or a combination of both) so that it can be further analyzed for trends to trace the corresponding trends in the groundwater. This allows selection of specific features of these records to determine changes in these drivers, rather than features that are physically controlled.

Climate trends in BC and historic trends in groundwater and stream levels are then discussed. These trends are considered from a regional perspective, and interpreted with respect to groundwater response type and aquifer-stream system type. To deal with the site-specific nature of these inter-relations, the links between low flow magnitudes, groundwater and climate fluctuations are explored in detail for selected sites.

3.4. Groundwater-stream interaction

Groundwater-surface water interaction is a general term for a continuum of aquifer-surface water connections observed over a range of scales. These connections range from the kilometre (km) scale of regional groundwater flow, to the metre (m) to millimetre (mm) scale of riffle-pool streambed flow (Winter et al., 1998). Groundwater-surface water interactions play an important role in the function of riparian ecosystems (Harvey and Bencala, 1992; Wroblicky et al., 1998; Winter et al., 1998; Bencala, 2000).

Within alluvial valleys, such as those found throughout BC, groundwater-river interactions are controlled by aquifer and valley geometry, aquifer materials, and river characteristics. Highly permeable floodplain deposits (e.g., cobble gravel) allow for the rapid response of floodplain water levels to river stage changes. This response occurs through lateral and vertical flow of water from the channel, thereby encouraging movement of channel waters into or out of the aquifer. In this way, highly permeable floodplains can develop large hyporheic zones. In contrast, low permeability floodplain deposits (silts and clays) limit river-aquifer flow, and thus, restrict hyporheic zone development. These controls are evident over a range of scales, from reach to regional. At the regional scale, discontinuous hydrostratigraphic units and/or highly variable bedrock topography may result in vertical groundwater flow. Thickening hydrostratigraphic units and subsurface bedrock troughs can cause groundwater downwelling, whereas thinning hydrostratigraphic units and subsurface bedrock ridges may cause groundwater upwelling (Harvey and Fuller, 1998).

The rivers, themselves, also exercise a variable control on their interaction with groundwater depending on their morphology, bedload material, discharge, slope, and floodplain material. Complex river planform morphologies develop complex cross-sectional and downvalley channel water surface slopes, as well as complex floodplain water tables. Flows between these heads are correspondingly complex. Large topographic regional slopes, common within Pacific coastal watersheds, encourage regional groundwater flowpaths. Permeable channel and floodplain sediments can deflect groundwater flowpaths both towards and away from the channel depending on floodplain water position (Woessner, 2000). These sediments can allow for the rapid transference of hydraulic head changes to or from the floodplain.

Streams either gain water from inflow of groundwater (gaining stream; Figure 3.1a) or lose water by outflow to groundwater (losing stream; Figure 3.2b). Many streams do both, gaining in some reaches and losing in other reaches. Losing streams can be connected to the groundwater system by a continuous saturated zone (Figure 3.1b) or can be disconnected from the groundwater system by an unsaturated zone (Figure 3.1c).

As discussed above, change in stream morphology and/or the aquifer properties can affect the nature of the interaction between the stream and the aquifer. Thus, changes in either of these can alter spatially whether a stream gains or loses water along its course. Some reaches may be gaining and others losing. Furthermore, the flow directions between groundwater and surface water can change seasonally as the elevation of water table changes with respect to the stream stage or can change over shorter timeframes when rises in stream surfaces during storms cause recharge to the streambank (Alley et al., 1999). Under natural conditions, groundwater makes some contribution to streamflow in most physiographic and climatic settings. Thus, even in settings where streams are primarily losing water to groundwater, certain reaches may receive groundwater inflow during some seasons. An important feature of streams that are disconnected from groundwater is that pumping of groundwater near the stream does not affect the flow of the stream near the pumped well.

In BC, unglacierized catchments tend to show low flows during late summer. In low elevation coastal regions, summer is the main low flow time of the year, while in snow-dominated mountain or interior regions, late-summer is the secondary low flow period (besides winter low flow due to freezing and storage of precipitation as snow). During summer low flow conditions it can be expected that stream flows are mainly fed by groundwater. Therefore, the observed inter-annual variation of August and September streamflow will depend, to a large extent, on the status of the groundwater system. Groundwater in coastal rain dominated regimes is recharged during the wet winter season, while in interior regions groundwater is recharged by snowmelt and spring/early summer precipitation.

Thus, the type of responses that are measured in streams and aquifers are inter-dependent and depend, to a large degree not only on the character of the aquifer-stream system but also on the climatology of the region, which affects the overall response of the groundwater system. As such, in the following two sections, the interaction between groundwater and surface water is classified according to 1) the aquifer-stream system type, and 2) the groundwater response type.

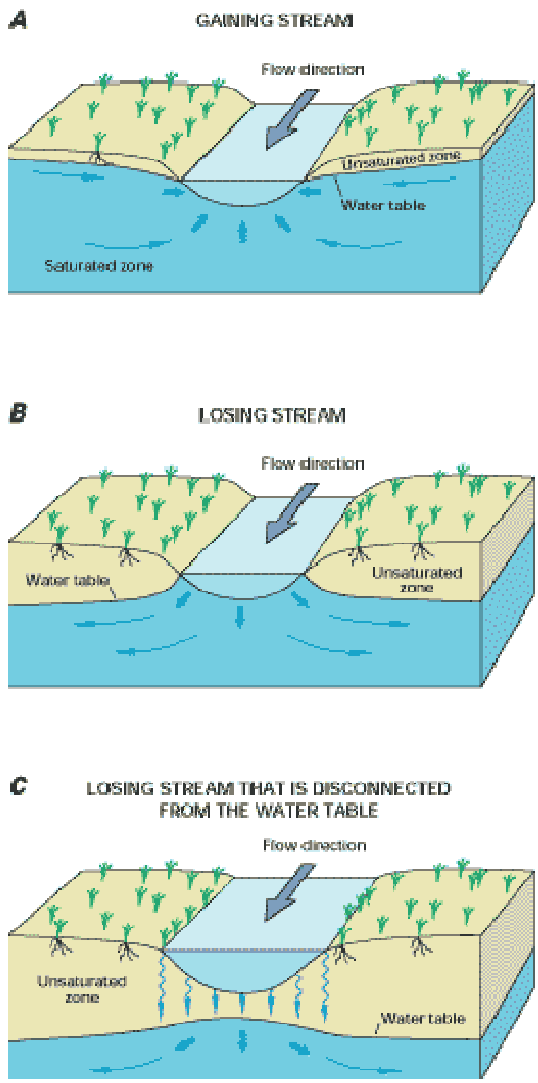


Figure 3.1. Groundwater-surface water interaction showing a) gaining stream, b) losing stream (from Alley et al., 1999 as modified from Winter et al., 1998).

3.5. Classification of aquifer-stream system types

The geologic character of the aquifer plays an important role in determining both the magnitude and timing of the response in the aquifer. The different aquifers types found in BC have recently been categorized (Wei et al., in prep) and are classified here as different Aquifer-Stream System types based on the anticipated responses of these various aquifer types to streamflow.

3.5.1. Main aquifer types in BC

In 1994, the Ministry of Environment developed the British Columbia Aquifer Classification System (BCACS) to identify and classify developed aquifers in BC as a means of providing summary information to assist with management of groundwater (Kreye and Wei, 1994). Developed aquifers are aquifers wherein wells have been completed to utilize groundwater. As of March 31, 2006 some 815 aquifers have been identified and classified in BC (Wei et al., in prep). Information on specific aquifers in BC can be obtained at the BC Water Resources Atlas website http://srmapps.gov.bc.ca/apps/wrbc/.

This inventory and its associated information, together with information contained in previous studies, has enabled the known aquifers in the province to categorized into various types. Each aquifer type is expected to have unique hydrogeologic characteristics, such as the nature of its origin, size and location, typical well depths, well yields, permeability, vulnerability, and <u>potential connection to surface water</u>. Six main aquifer types (four with sub-categories) are identified (Wei et al., in prep):

Unconsolidated aquifers

1. predominantly unconfined fluvial or glaciofluvial aquifers along river or stream valleys;
    - 1a. aquifers along major rivers of higher stream order, where there is potential of being hydraulically influenced by the river;
    - 1b. aquifers along rivers of moderate stream order, where there is potential of being hydraulically influenced by the river; or
    - 1c. aquifers along confined, lower order (< 3-4) streams where aquifer thickness and lateral extent are more limited;
2. Predominantly unconfined deltaic sand and gravel aquifers;
3. Alluvial fan, colluvial sand and gravel aquifers;
4. Sand and gravel aquifers of glacial or pre-glacial origin;
    - 4a. unconfined glaciofluvial outwash or ice contact sand and gravel aquifers; or
    - 4b. confined sand and gravel aquifers of glacial or pre-glacial origin; or
    - 4c. confined sand and gravel aquifers associated with marine environments;

Bedrock aquifers

5. Sedimentary rock aquifers;
    - 5a. fractured sedimentary bedrock aquifers; or
    - 5b. karstic limestone aquifers;
6. Crystalline rock aquifers;
    - 6a. flat-lying or gently-dipping volcanic flow rock aquifers; or

6b. crystalline granitic, metamorphic, meta-sedimentary, meta-volcanic and volcanic rock aquifers.

3.5.2. Aquifer-stream system types

Each aquifer type can be expected to interact to varying degrees with streams that flow through them. Two main aquifer-stream system types are proposed:

1. **Stream-Driven Systems** in which groundwater flow to and from streams is bi-directional, and varies seasonally depending on the magnitude of the streamflow and precipitation. These include valley bottom unconfined fluvial or glaciofluvial aquifers that are found in association with major streams/rivers, and aquifers that are found at a break in slope, such as a valley bottom or a coastal plain, in which streams flow variably across the top or possibly "disappear" into the sediments.

2. **Recharge-Driven Systems** in which the aquifer is raised above the surrounding land surface and which drains to lower elevation. In this type of system, groundwater is recharged solely by precipitation and dominantly discharges to streams during periods of low flow (e.g., Abbotsford-Sumas aquifer).

Stream-driven systems

Due to the mountainous nature of much of the province, many aquifers are situated in confined valley bottoms or at the base of a stream or river that originates at higher elevation. fluvial/glaciofluvial (category 1), deltaic (category 2), and alluvial/colluvial (category 3) aquifers tend to fall in this category.

The fluvial/glaciofluvial aquifers are defined based on their occurrence near streams and rivers in sediment-filled valley bottoms that are typically confined by mountainous bedrock (e.g., Grand Forks aquifer; Scibek et al., 2007). In such environments, a stream entering the valley, encounters permeable aquifer sediments, effectively increasing the area through which water can flow. In essence, the aquifer becomes an extended river channel.

Deltaic aquifers, by the very nature of their depositional environment, will have some interaction with surface water (older deltaic deposits of glacial origin are included in category 4). Deltas are found in low-lying areas at the mouth of rivers or streams. Areas with coarse grained channel deposits will have good connection to the surface waters, whereas, areas with finer-grained deposits may be less likely to interact with surface water.

Alluvial/colluvial fan deposits are often found at the base of slopes and, where exposed at surface, are often found in association with a discharging stream (e.g., Okanagan valley side fan aquifers), but excludes older alluvial or colluvial aquifers at depth, which are considered under category 4.

The key characteristic of these stream-driven systems is the dependence of the groundwater levels on surface water contributions. Rivers recharge the aquifers (this is termed indirect recharge) when their stage is high (losing streams), but then as the hydrologic regime shifts to a low flow regime, the direction of flow is reversed and the groundwater discharges into the stream (gaining stream). Consequently, these systems are bi-directional, and can exhibit a regime shift (losing to gaining) depending on season.

Recharge-driven systems

Other aquifers types do not lend themselves as readily to interaction with streams, either because they tend not to be found in association with a stream or because the morphology or permeability does not lend itself readily to such interactions. These include sand and gravel aquifers of glacial or pre-glacial origin (category 4), sedimentary rock aquifers (category 5) and crystalline rock aquifers (category 6).

Sand and gravel aquifers of glacial or pre-glacial origin include unconfined glaciofluvial outwash or ice contact sand and gravel aquifers, generally formed near or at the end of the last period of glaciation, and confined sand and gravel aquifers underneath till, in between till layers, or underlying glaciolacustrine and glaciomarine deposits (Quadra Sand deposits). In some cases, confined sand and gravel aquifers may be found in association with fluvial/glaciofluvial aquifers.

Sedimentary rock aquifers include both clastic and limestone rocks, in which flow generally occurs through discrete fractures or solution cavities (e.g., Gulf Islands).

Crystalline bedrock aquifers includes both flat-lying to gently-dipping volcanic flow aquifers in which groundwater flow often occurs within weathered zones between flows, and fractured crystalline rocks in which groundwater flow is mostly along joints, fractured and faults.

The key characteristic of these recharge-driven systems is their lack of interaction with surface water and their stronger reliance on recharge from precipitation, but the reason differs for each. In the case of sand and gravel aquifers of glacial or pre-glacial origin, many of these deposits are found in coastal regions (where outwash was generated by melting glaciers). This resulted in their somewhat unique morphology in that they are raised above the surrounding land surface (e.g., Brookswood aquifer, Abbotsford aquifer). As such, they are not fed by streams, but rather any streams originate within them and act to drain the aquifer. Consequently, the primary recharge to the aquifer is precipitation (during the rainy season or during spring snowmelt), and only during the summer months is there strong connection (not interaction *per se*) with the streams in the form of baseflow contribution (gaining streams). Confined sand and gravel aquifers of glacial or pre-glacial origin may occur in association with fluvial/glaciofluvial unconfined aquifers, and as such might be expected to share similar responses to those type 1 aquifers.

In the case of sedimentary rock or crystalline rock aquifers, the lack of interaction with surface water is more a function of the low permeability of these rocks relative to sand and gravel deposits. Certainly, these aquifer types are commonly found nearby streams, but the function of the streams in relation to the aquifer is quite different. Due to their low permeability, any streams flowing within these aquifers will largely flow on the surface. The rock does not offer a viable pathway for water flow, and so water prefers to remain in the stream channel. These types of aquifers consequently promote high runoff and limited infiltration, although some precipitation does recharge the aquifers. As in the case of the sand and gravel deposits of glacial or pre-glacial origin, there is a much greater likelihood that groundwater will discharge to streams during the low flow period.

3.6. Classification of groundwater response types

The magnitude and timing of the response of the groundwater system as reflected in changes in groundwater levels depends on the nature of the driving force, that is, whether the system is stream-driven or recharge-driven, and also on the distribution and permeability of the aquifer materials. However, in order to characterize these responses, it is important to consider a second factor that controls the response, namely the climatology of the region.

In this section, the responses of aquifer systems to climate are explored using data from selected high quality provincial observation wells and then used to classify different groundwater response types. As discussed earlier, the relationships between climate, streamflow and groundwater are complicated. An attempt is made to first classify the groundwater responses according to:

1. the hydrologic regime (pluvial, nival, glacierized, mixed), and

2. whether the response is indicative of hydraulic connection with surface water as predicted by the aquifer-stream system types defined in the previous section.

The source of groundwater level information was obtained from the BC Observation Well Network (http://srmapps.gov.bc.ca/apps/gwl/disclaimerInit.do). The Observation Well Network was established in 1961 and was comprised of a number of unused dug and drilled wells in the Lower Fraser Valley and the Okanagan Valley. As of September 2006, there are 158 active observation wells in the network covering major groundwater areas of the province. Each is identified with a unique observation well number. Hereafter, wells are referred to by their unique BC observation well number.

The primary purpose of the Observation Well Network is to collect, analyze and interpret ground water hydrographs and ground water quality data from various developed aquifers in BC. A select few of these wells monitor aquifers in pristine areas that reflect natural variability; the others have been influenced by human activity making them less representative. Observation wells are equipped with automatic level recorders or data loggers that monitor water level fluctuations on a continuous basis. The periods of record for these observation wells vary. The earliest records date back to 1962, and are still being monitored. As groundwater development increased over time, new observation wells were installed in other regions, hence their periods of record are shorter.

3.6.1. Site descriptions

Of the 158 active observation wells, only nine had records that were considered suitable for detailed analysis in that daily time series data were available for a long period of record. The nine wells are located across southern BC; they span from the western coastal area eastward (Figure 3.2; Table 3.1). Meteorological and hydrological stations were selected based on their close proximity to these wells; see Table 3.1 for these related stations.

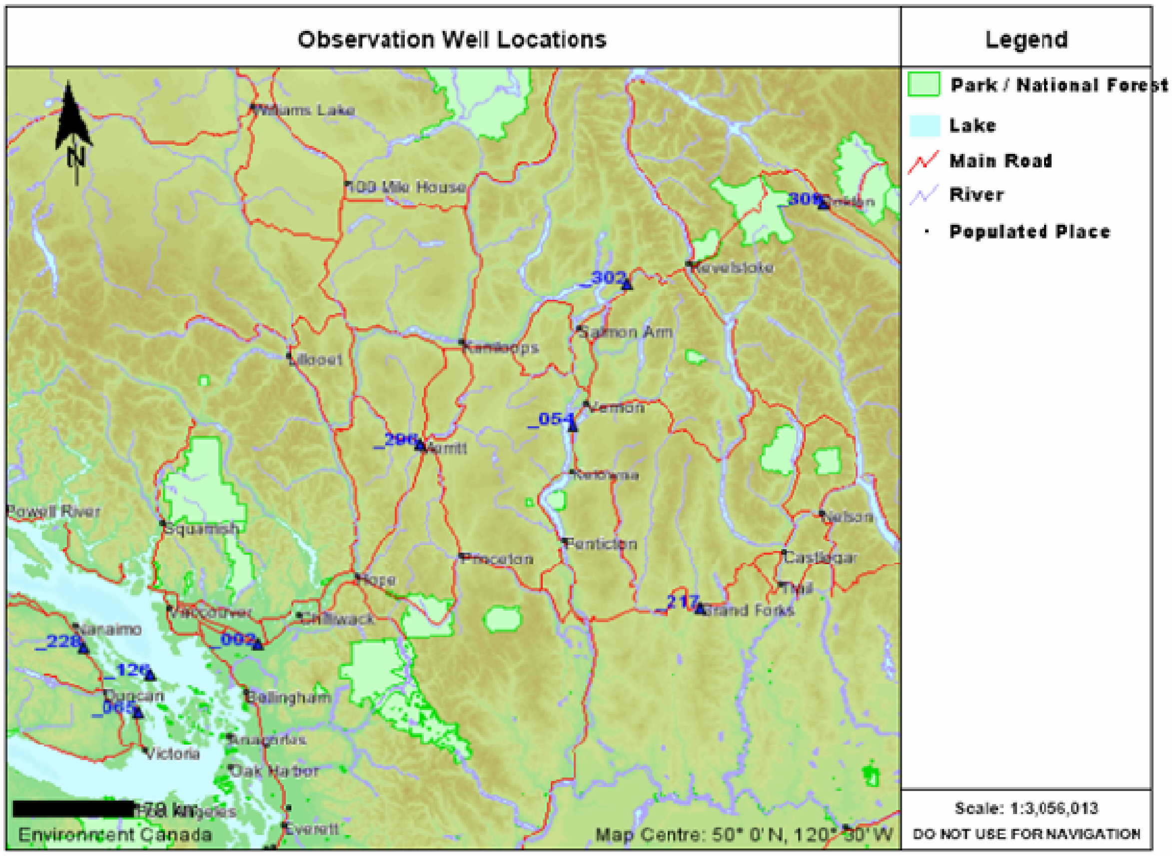


Figure 3.2: The study area for detailed analysis of observation well data.

Table 3.1: The observation wells and their associated hydro-met and climate stations.

Station#	Station name	Latitude	Longitude	Record Length		Hydro	Climate	Station name	Latitude	Longitude	Record Length	Drainage Area
_002	Fraser Valley	49.017	-122.343	1962-1998	1	08MH029		SUMAS RIVER NEAR HUNTINGDON	49.003	-122.231	1935-2005	149
					C		1100030	Abbotsford A	49.025	-122.363	1953-2002	
_054	Carrs Landing	50.145	-119.401	1969-1998	1	08NM174		WHITEMAN CREEK ABOVE BOULEAU CREEK	50.212	-119.539	1971-2005	112
					C		1128958	Winfield	50.037	-119.416	1971-2002	
_065	Saanich	48.649	-123.413	1972-1998	1	08HA011		COWICHAN RIVER NEAR DUNCAN	48.773	-123.712	1960-2005	826
					2	08HA001		CHEMAINUS RIVER NEAR WESTHOLME	48.879	-123.702	1914-2005	355
					C		1018620	Victoria Int'l A	48.647	-123.426	1953-2002	
_126	Mayne Island	48.861	-123.292	1973-1998	1	08HA011		COWICHAN RIVER NEAR DUNCAN	48.773	-123.712	1960-2005	826
					2	08HA001		CHEMAINUS RIVER NEAR WESTHOLME	48.879	-123.702	1914-2005	355
					C		1014931	Mayne Island	48.845	-123.320	1970-2002	
_217	Grand Forks	49.023	-118.434	1977-1998	1	08NN012		KETTLE RIVER NEAR LAURIER	48.984	-118.215	1929-2005	9840
					2	08NN013		KETTLE RIVER NEAR FERRY	48.981	-118.765	1928-2005	5700
					C		1133270	Grand Forks	49.026	-118.466	1941-2002	
_228	Cassidy	49.045	-123.873	1978-1998	1	08HB034		NANAIMO RIVER NEAR CASSIDY	49.069	-123.887	1965-2005	684
					C		1025370	Nanimo A	49.052	-123.870	1954-2002	
_296	Merritt	50.109	-120.788	1987-1998	1	08LG007	1125079	NICOLA RIVER NEAR MERRITT	50.144	-120.885	1911-2005	4350
					2	08LG010		COLDWATER RIVER AT MERRITT	50.110	-120.803	1913-2005	914
					C		1125079	Merrit STP	50.114	-120.802	1968-2002	
_302	Malakwa	50.935	-118.794	1988-1998	1	08ND013		ILLECILLEWAET RIVER AT GREELEY	51.014	-118.083	1963-2005	1170
					2	08LE024		EAGLE RIVER NEAR MALAKWA	50.936	-118.800	1913-2005	904
					C		1166945	Salmon Arm A	50.667	-119.217	1982-2002	
_309	Golden	51.259	-116.919	1989-1998	1	08NA002		COLUMBIA RIVER AT NICHOLSON	51.244	-116.912	1903-2005	6660
					2	08NA006		KICKING HORSE RIVER AT GOLDEN	51.300	-116.968	1911-2005	1850
					C		1173210	Golden A	51.298	-116.982	1991-2002	

Following is a brief description of each observation well. The maps are screen captures from the BC Water Atlas (http://www.env.gov.bc.ca/wsd/data_searches/wrbc/index.html) and are intended for illustration only. Information on well depth, aquifer materials and the aquifer type, as determined by Wei et al. (in prep) is given.

WELL 002

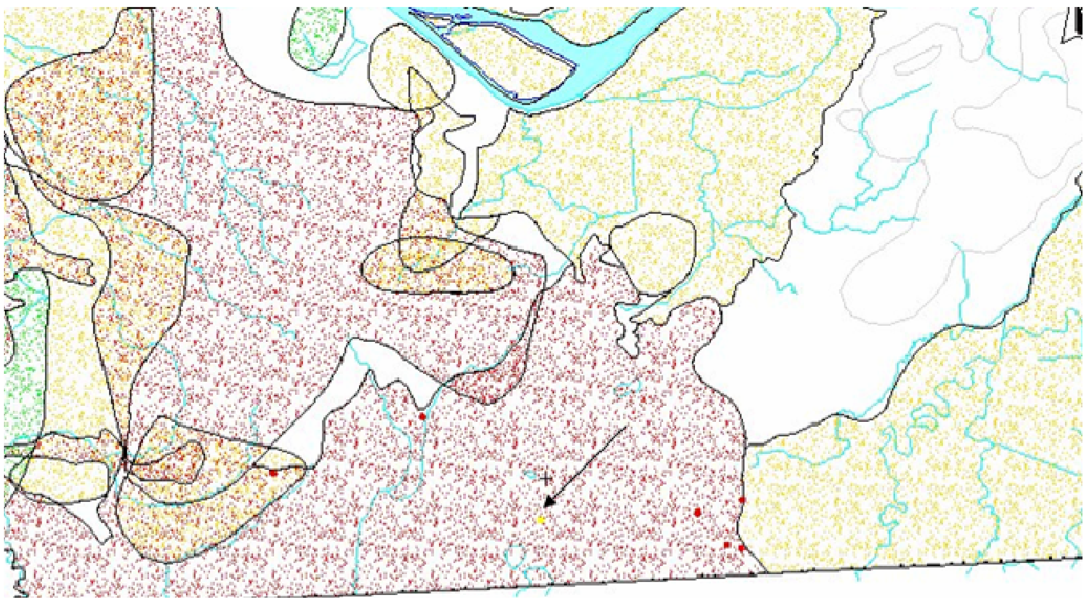


Figure 3.3: Well 002 situated in Abbotsford on Huntingdon Rd. The red stippled polygon at the bottom along the border is the Abbotsford aquifer (Aquifer #015). This aquifer is classified as Type 4a: Unconfined glaciofluvial sand and gravel. The well is 63 ft deep and completed in sandy gravel deposits of the Abbotsford glacial outwash. The Abbotsford aquifer is unconfined, is highly productive and its substrate is predominantly sand and gravel (Scibek and Allen, 2006b). The well is situated to the south east of the Abbotsford International airport, in the Abbotsford Uplands between Walmsley Lake (North) and Laxton and Judson Lakes to the south. The well is situated roughly 3km to the east of Fishtrap Creek (the nearest creek), and is likely not influenced by stream flow at this distance.

WELL 065

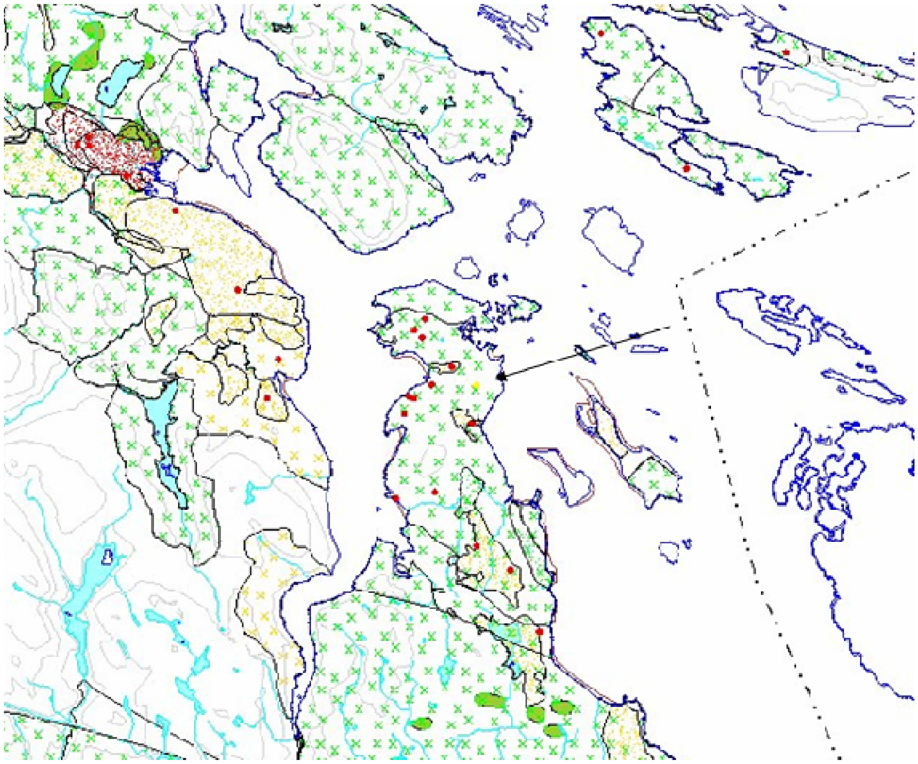


Figure 3.4: Well 065 is situated on the Saanich Peninsula, on Beacon Rd. (Aquifer #608). This aquifer is classified as Type 6b: Fractured crystalline rock (Island Plutonic Suite). The well is 505 ft deep and completed in competent bedrock (recorded as granodiorite). The aquifer has low productivity. The well is situated roughly 1.3 km from the ocean coast, and it is not situated close to any streams.

WELL 126

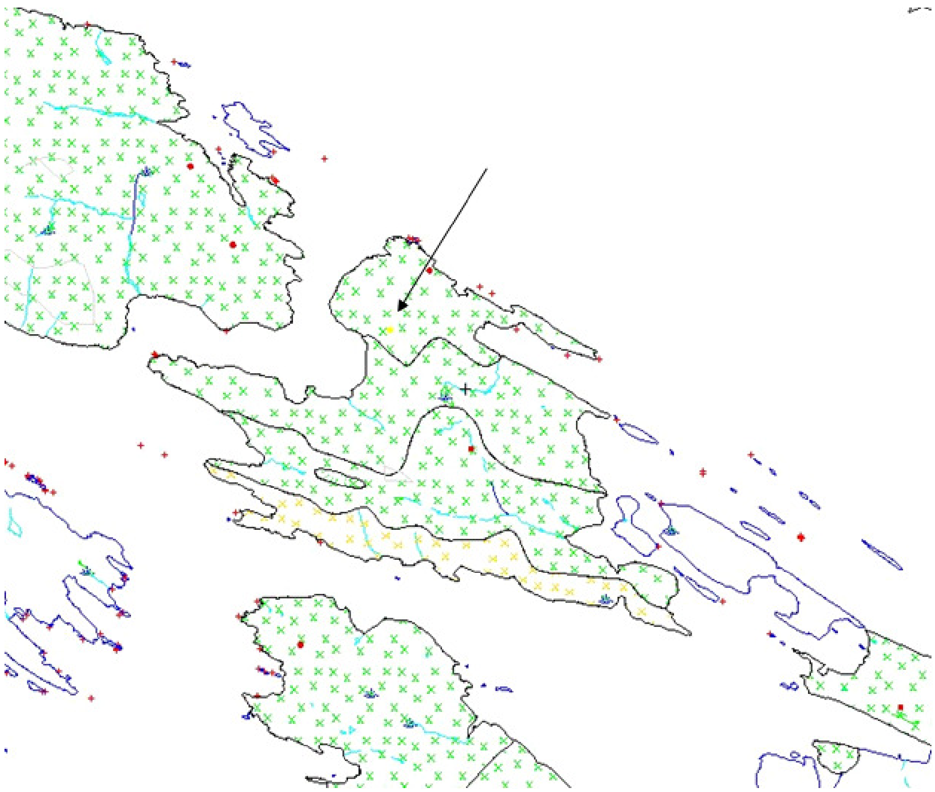


Figure 3.5: Well 126 is situated on Mayne Island, within the Gulf Islands, at relatively high elevation on Georgia Point-Hall Hill (Aquifer #447). This aquifer is classified as Type 5a: Clastic rock aquifers. The well is 233 ft deep and completed in sandstone. The observation well on Mayne Island is completed within Gabriola Formation rocks, which are dominantly sandstone and considered a relatively low productivity formation. The well is not situated near any stream.

WELL 228

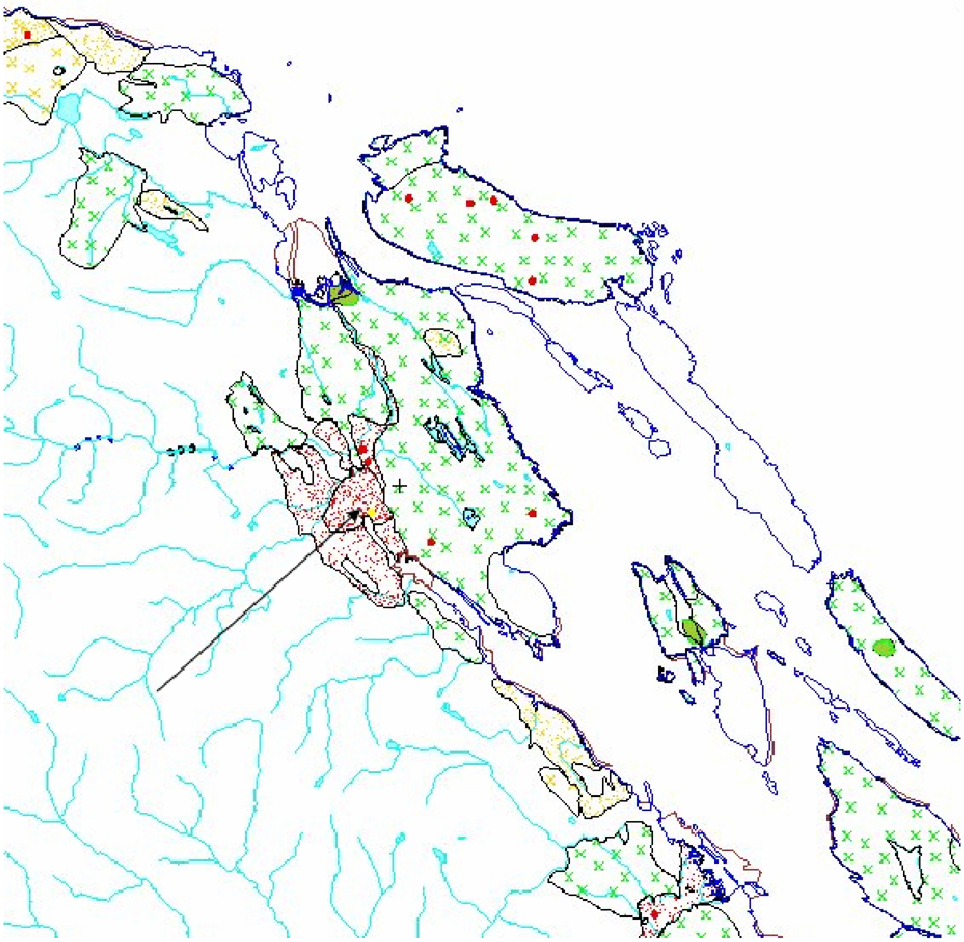


Figure 3.6: Well 228 is situated in the Cassidy aquifer (Aquifer #160) in the center of the inner red speckled area. This aquifer is classified as Type 4b: Confined sand and gravel aquifers of glacial or pre-glacial origin (glaciofluvial or glaciolacustrine). The well is 195 ft deep and is completed in a mixture of sand and gravel with some clay lenses. Distance to the nearest stream is approximately 107 m. The well is also situated approximately 2 km from a potentially tidally influenced bay.

WELL 54

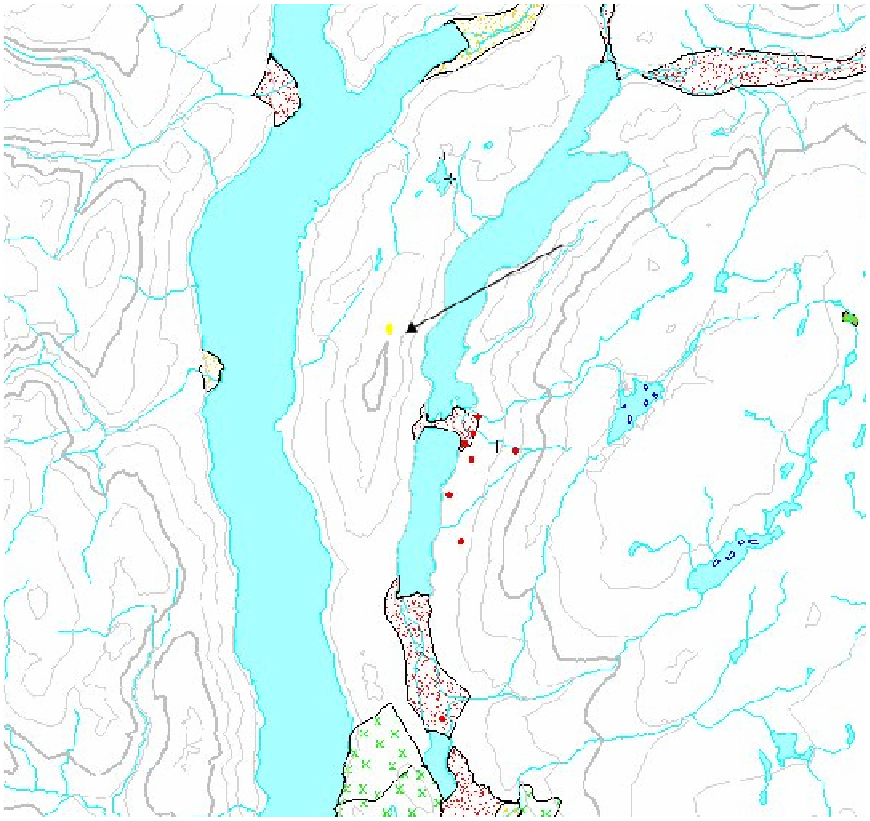


Figure 3.7: Well 054 is situated near Carr's Landing, north of Kelowna in the Okanagan (Aquifer is unmapped). Based on location, this aquifer is likely either Type 5a or 6b: Clastic sandstone or fractured crystalline bedrock. The well is 45 ft deep and the well log indicates undifferentiated bedrock. The well appears to be situated on a bedrock high that separates Okanagan Lake from an adjacent Lake. There are no nearby streams with which this well might interact.

Well 217

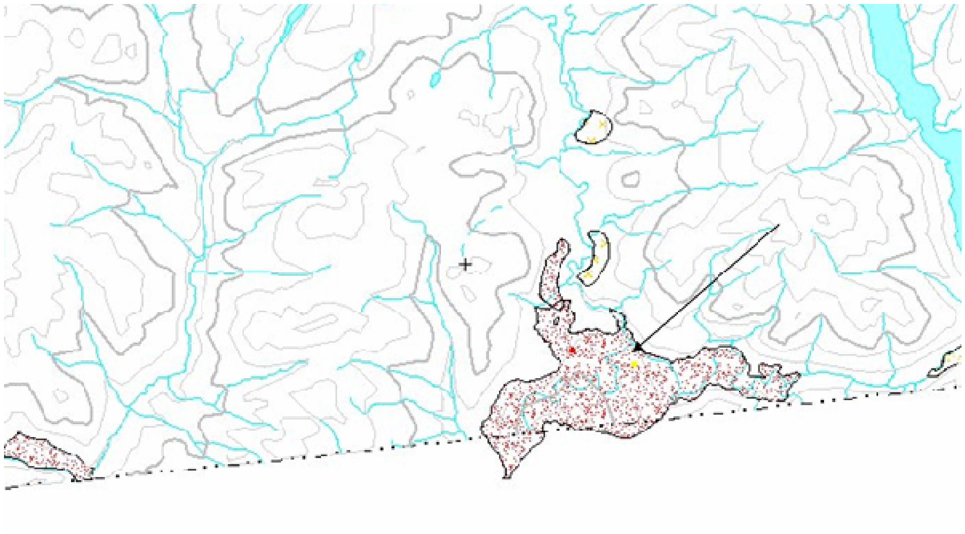


Figure 3.8: Well 217 is situated within the Grand Forks aquifer (Aquifer #158). This aquifer is classified as Type 1a: Along major higher stream order river valleys influenced by surface water (high energy sand and gravel aquifer). The well is 29 ft deep and completed in gravel. The Grand Forks aquifer is a highly productive alluvial (dominantly sand and gravel) aquifer, situated in a bedrock valley. It is relatively continuous across the valley bottom and in close hydraulic connection with the Kettle River, which meanders through the valley (Scibek et al., 2006; Scibek and Allen, 2006a). The well is situated approximately 700 m to the south of the Kettle River, in the curve of a river meander, thus it is influenced on all sides by stream flow interaction.

WELL 309

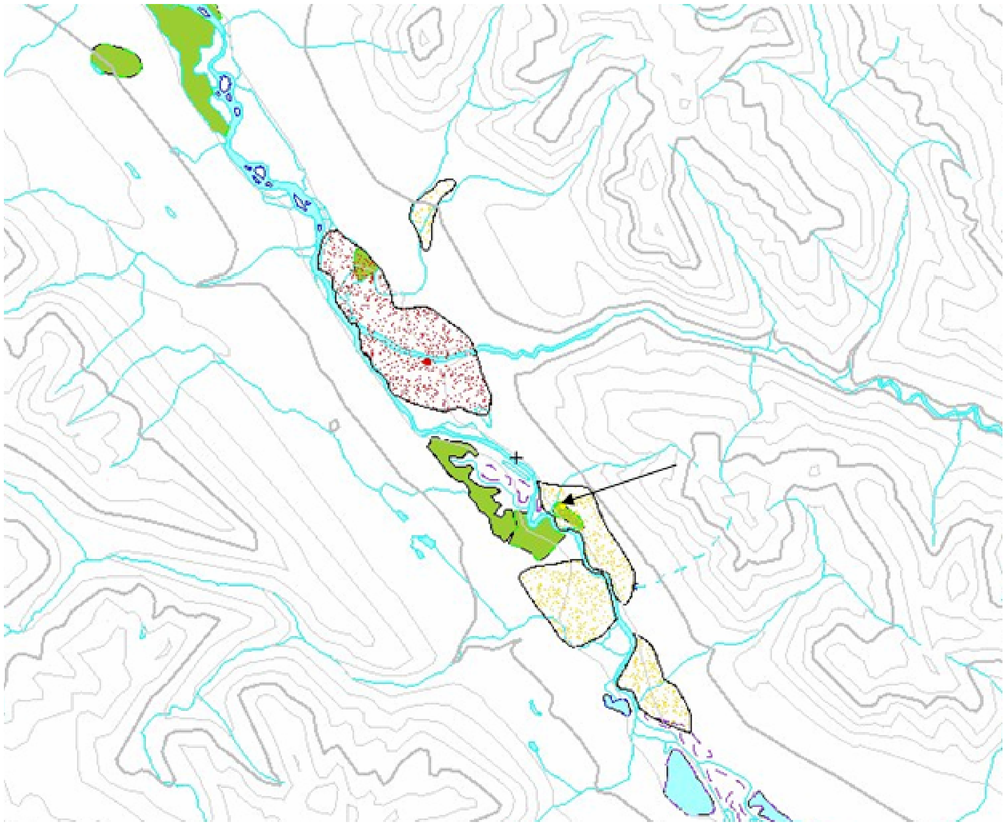


Figure 3.9. Well 309 is situated near Golden on Nicholson Creek (Aquifer #450). The aquifer is classified as Type 4b: Confined sand and gravel aquifers of glacial or pre-glacial origin (glaciofluvial or glaciolacustrine). The well is 148 ft deep and completed in primarily sandy gravel. The well is situated 400 m from a river and right on an ephemeral stream. The aquifer is moderately productive.

Well 296

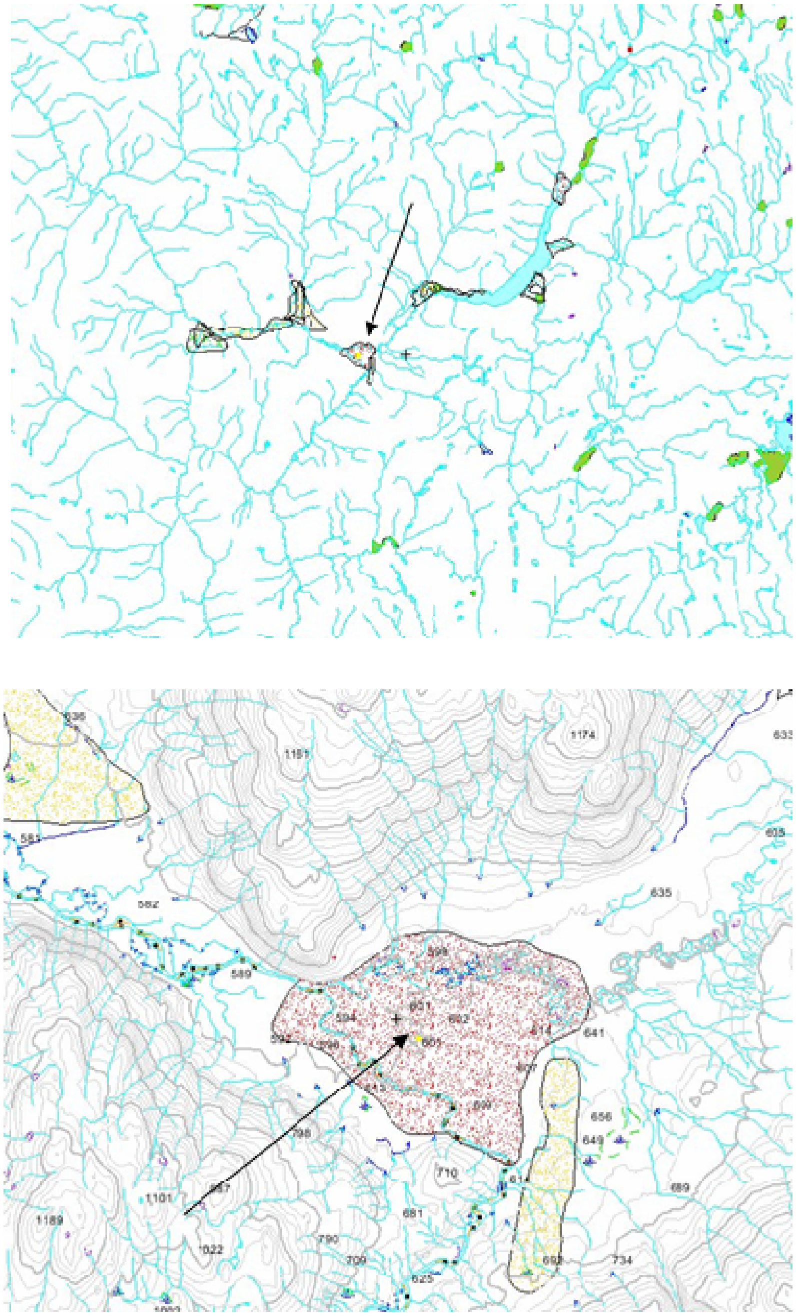


Figure 3.10. Well 296 is situated with the Merritt aquifer (Aquifer #074). This aquifer is classified as Type 1a: Along major higher stream order river valleys influenced by surface water (high E sand and gravel aquifers). The well is 56 ft deep and situated 1.2 km from the confluence of two steams (to the east). One stream flows to the north, the other to the south of the well. The aquifer is a highly productive, fan deposit. The aquifer is comprised primarily of permeable sand and gravel.

Well 302

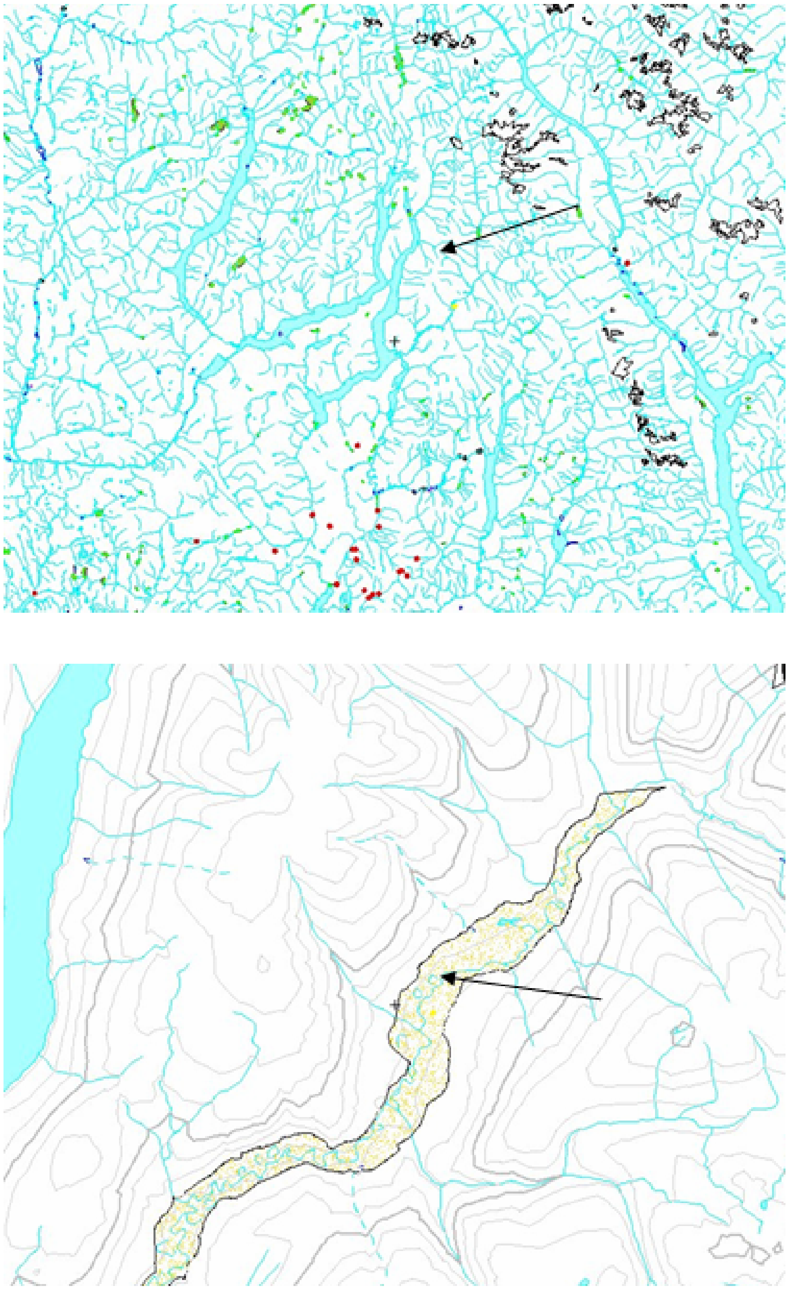


Figure 3.11: Well 302 is located in Salmon Arm (Malakwa) (Aquifer #307). This aquifer is classified as Type 1a: Along major higher stream order river valleys influenced by surface water (high E sand and gravel aquifers). The well is 75 ft deep and is completed in fine to coarse sand and gravel. The well is situated roughly 300m from a stream, thus due to the high permeability of the sediments. The aquifer is highly productive and consists of a sand and gravel river valley bottom aquifer.

The nine observation wells were classified into aquifer type (Table 3.2).

Table 3.2: Classification of observation wells considered in the detailed analysis based on aquifer type.

Observation Well	**BC Aquifer Number**	**Aquifer Type**	**Stream-Driven (SD) or Recharge-Driven (RD)?**
002	015	Type 4a: Unconfined glaciofluvial sand and gravel	RD
065	608	Type 6b: Fractured crystalline rock	RD
126	447	Type 5a: Clastic rock aquifers	RD
228	160	Type 4b: Confined sand and gravel aquifers of glacial or pre-glacial origin (glaciofluvial or glaciolacustrine)	RD or SD
054	N/A	Type 5a or 6b: Clastic sandstone or fractured crystalline bedrock.	RD
217	158	Type 1a: Along major higher stream order river valleys influenced by surface water (high energy sand and gravel aquifer	SD
309	450	Type 4b: Confined sand and gravel aquifers of glacial or pre-glacial origin (glaciofluvial or glaciolacustrine)	RD or SD
296	074	Type 1a: Along major higher stream order river valleys influenced by surface water (high E sand and gravel aquifers).	SD
302	307	Type 1a: Along major higher stream order river valleys influenced by surface water (high E sand and gravel aquifers)	SD

3.6.2. Methodology

Groundwater records were collected by the BC Ministry of the Environment (BC MoE). Well water levels were measured continuously with Stevens Type F chart recorders (accuracy level ± 1 mm) (BC MoE, 2006) and are referenced from ground level (BC MoE, 2006). These records were provided to Environment Canada where they were manually digitized using Water Survey of Canada (WSC) procedure. Daily average water levels were (manually) extracted, and data were carefully verified. Missing values were infilled using linear approximation and mean annual hydrograph fitting.

Meteorological Survey of Canada (MSC) climatological stations were then selected based on their close proximity to the well of interest and their period of record. Data were retrieved from the 2002 CDCD West compact disc (Canadian Daily Climate Data, 2002). Hydrological records were downloaded from the WSC Archived Hydrometric Data site (WSC, 2006) for the station nearest the observation well. Careful consideration of the spatial relationship between well and stream location was made when selecting hydrometric data. This was done to ensure that the stream selected was truly that which had the potential to interact with the groundwater monitored by the well in question.

For each well, the groundwater level data and discharge from the associated stream were plotted in R (R Development Core Team, 2006). The resulting plots present the data over an 18 month period, classifying the median, 5th-10th, 10th-25th, 25th-75th, and the >95th quantiles for the period of record. Traditional hydrograph plots display only 12 months making it difficult to visualize key aspects of the hydrology of a system. This is especially true if the main water events take place during the threshold between one calendar year and the next, which is often the case in North America, where most precipitation arrives in the Nov., Dec., Jan. period. The plots provide an easily understandable representation of the variation in water levels for a given well, from year to year, and display the characteristic time periods when recharge and discharge events occur in the aquifer. By this method, the groundwater plots are easily compared to their associated stream hydrographs. Additionally, the average timing of high and low water periods in one well versus the other is easily recognized. If stream water levels were available they were used in place of stream discharge to make the comparison between water levels in the stream and in the ground more linear.

Boxplots of temperature normals were plotted for each meteorological station using the plot.seas.temp function in the ‘seas’ package for R (Toews et al., 2007). These plots present the median value as the horizontal line through the box, the first and third quartile as the top and bottom of the box, the maximum and minimum as the ends of the whisker, and (if they are present) the circles denote outliers (Chambers et al., 1983). The widths of the boxes are proportional to the square root of the sample sizes in each boxplot.

This plot highlights the variation of air temperature over the year and within a given 11-day group. Traditional boxplots of air temperature might average values over months or seasons. This approach loses some of the information that can be valuable in accessing key features and trends in data at a given station (Whitfield et al, 2003). The 11-day bins maintain a high level of information while not overwhelming the viewer with visual detail.

Precipitation was plotted using the plot.seas.norm function in the ‘seas’ package for R (Toews et al., 2007). Again 11-day bins were used to reduce the data volume while making it easier to compare from one station to another and to visualize seasonal trends. Median was used rather than mean because it is more robust. The precipitation normals were first computed using the precip.norm function in ‘seas’, which calculates normals for precipitation, rain, and snow. The height represents the mean values of total precipitation and the grey colour (where present) represents the snow amount from the rainfall fraction. The thickness of the line represents the median number of days with active precipitation. This gives an indication of the intensity of precipitation over the 11-day period. The precipitation normals plots were extended to span 18 months so that they are comparable to both the hydrographs of groundwater and stream discharge.

Groundwater level data for all 9 wells were compared to find a period when all wells had complete data. This was found to be Oct 30th 1989 to Feb 28th 1992. Using R, a scatter plot of groundwater level versus the log of the discharge volumes from the corresponding stream was created. Each month was given a separate symbol, and each water year (Oct 1st to Sept 30th) was given a different colour. This graph facilitates the evaluation of the response, or lack of response of the groundwater levels to the stream discharge. It presents those trends which are similar from year to year and those that may change from year to year in response to varying water availability.

Breakpoints in the groundwater record were then identified through evaluating the precipitation, groundwater and streamflow hydrograph plots. These breakpoints were used to set the days when rainfall was summed to compare precipitation rates to recharge rates, and then precipitation rates to discharge rates. Both 5 day and 11 day bin averages of recharge rates, rainfall, temperature, and stream discharge were created. Finally cross correlation analysis of precipitation and recharge rates, and stream discharge and recharge rates were performed in R using the ccf() function. This analysis demonstrated correlation between the two variables accounting for the time lag between rainfall and recharge rates, and stream discharge and recharge rates.

3.6.3. Groundwater response types

Similar to hydro-climatic zones, groundwater levels have attributes that are characteristic to the area in which they are situated in the province. Wells 002, 065, 126, and 228 are all found between 48.6 and 49.1 degrees latitude and123.9 and122.3 degrees longitude, in the most southwestern portion of the study area (see Figure 3.2). The wells display their lowest water levels in the Sept. to Nov. period. After this period groundwater levels increase and peak during Feb. or Mar. Groundwater levels decline soon after this peak period. In well 002, there is a much more gradual rise in groundwater level early in the season as compared to 065, 126 and 228, and similarly a more graduate decline in groundwater levels. Throughout the year, the variability in groundwater levels is relatively constant in 002, 065 and 126, as evidenced by the distribution of percentile ranges (Figure 3.12, 3.13 and 3.14). Well 228, however, has a lower range of variability during the recession period (Figure 3.15).

Wells 054, 217, 309, and 296 are situated between 49.0 and 50.9 degrees latitude and120.8 and116.9 degrees longitude (Figure 3.2). All are characteristic of the climate of the Interior region of BC in that the lowest water levels are reached in Jan.-Mar., and begin to climb in late Mar., and increase through to Jun. when peak water levels are reached. The response at 054 (Figure 3.16) is slightly different from that at the other wells (Figure 3.17 318, 3.19) in that the recession appears to be more gradual. Also, there appears to be considerable variability in the groundwater levels throughout the year. Well 296 possibly shows greater variability in Jan.-Feb., which may be mis-interpreted to be a second peak, but this is likely not the case. The quality of the data for this well is questioned.

Well 302 is also located in the interior, but are at lower elevation where rainfall and snowmelt can influence water levels. Well 302, Malakwa, shows a peak in Oct. to Jan. and another increase from Mar.-May (Figure 3.20).

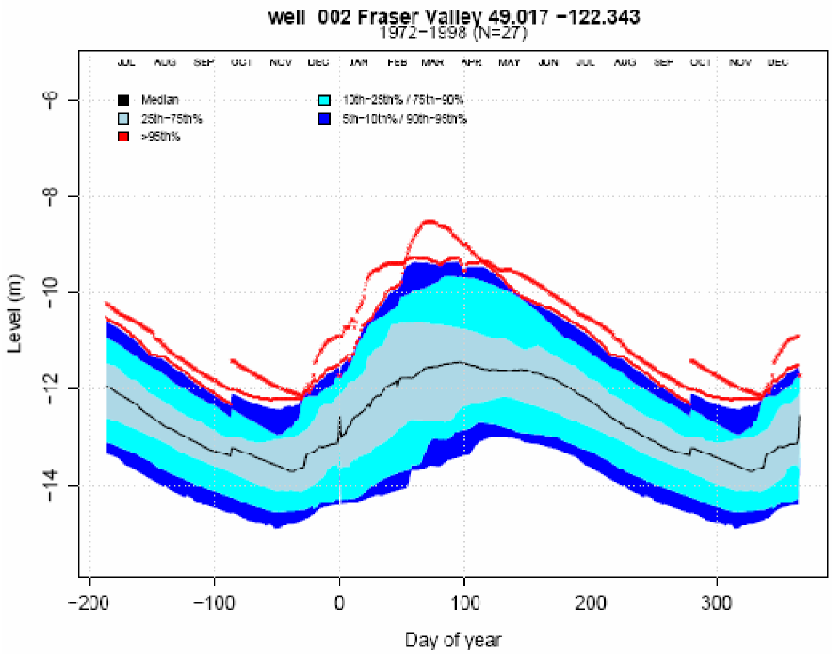


Figure 3.12: Groundwater level hydrograph for well 002.

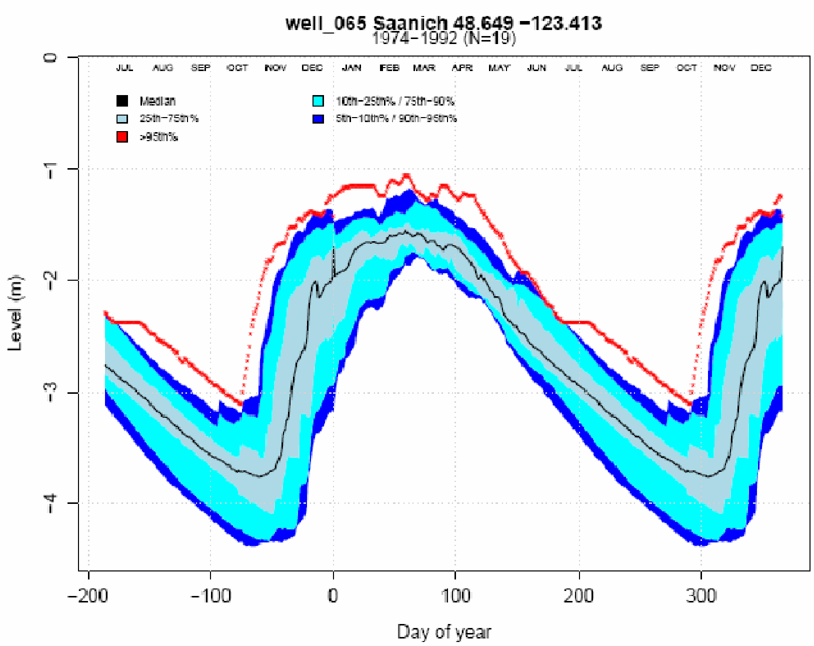


Figure 3.13: Groundwater level hydrograph for well 065.

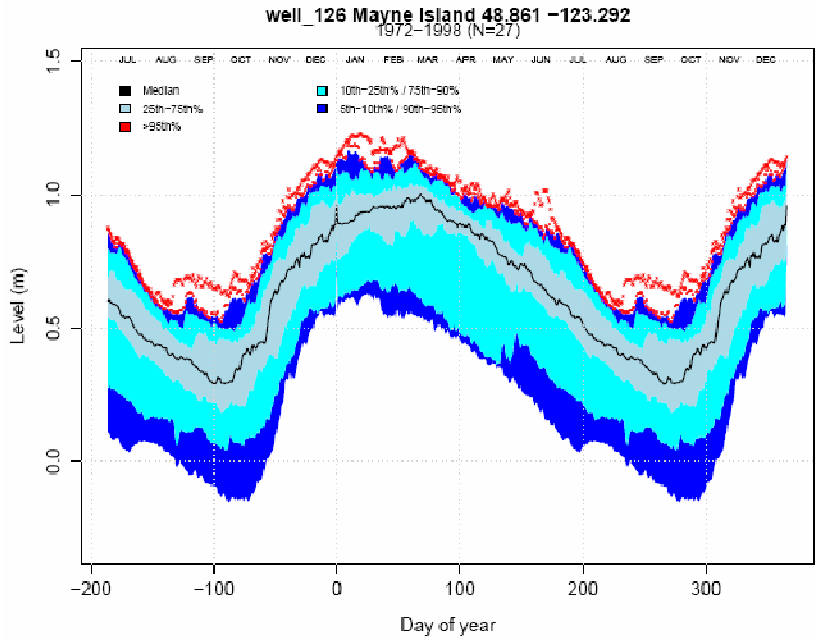


Figure 3.14: Groundwater level hydrograph for well 126.

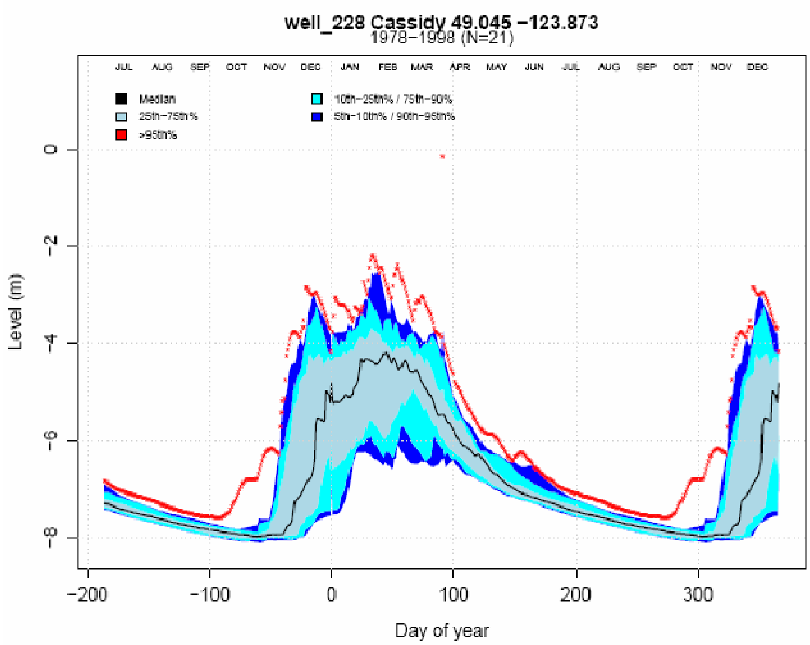


Figure 3.15: Groundwater level hydrograph for well 228.

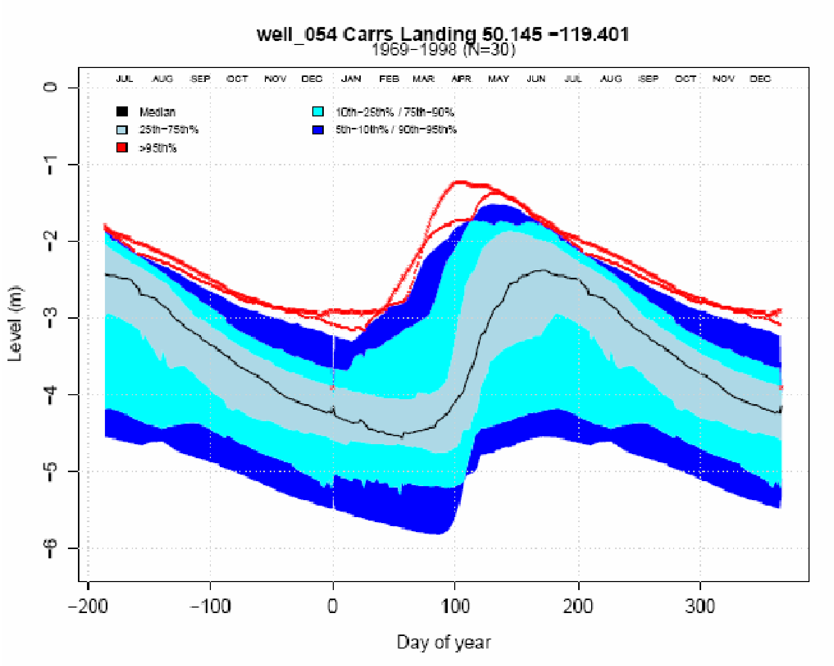


Figure 3.16: Groundwater level hydrograph for well 054.

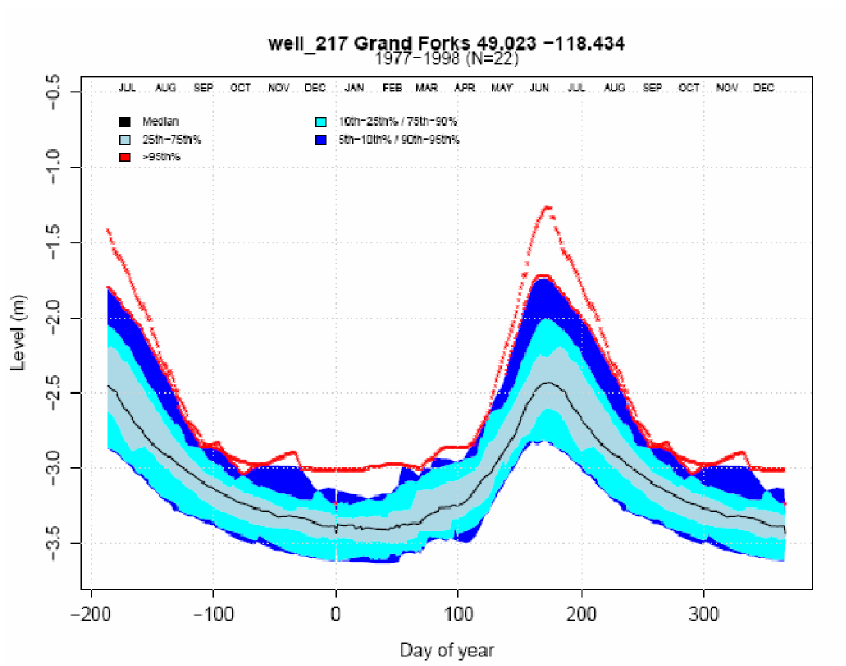


Figure 3.17: Groundwater level hydrograph for well 217.

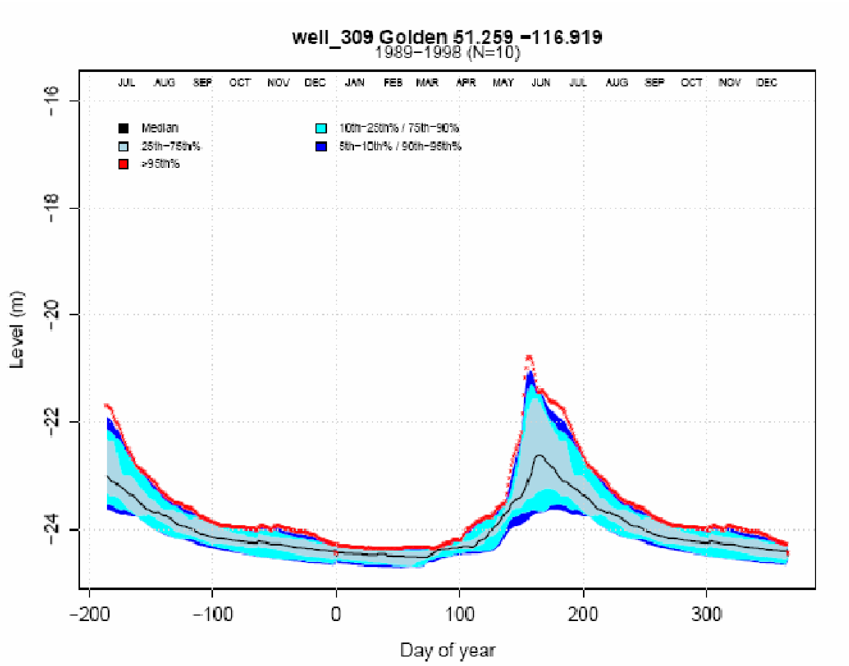


Figure 3.18: Groundwater hydrograph for well 309.

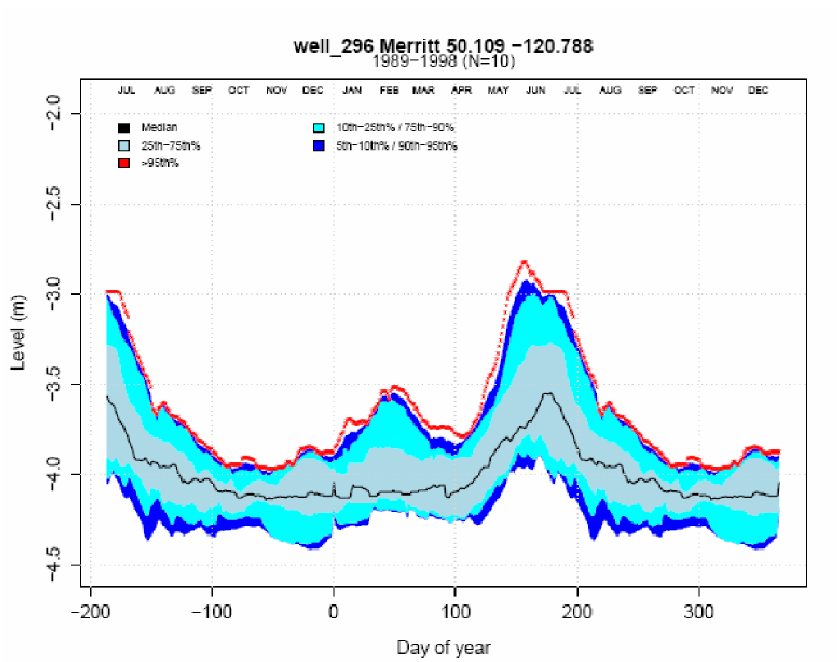


Figure 3.19: Groundwater level hydrograph for well 296.

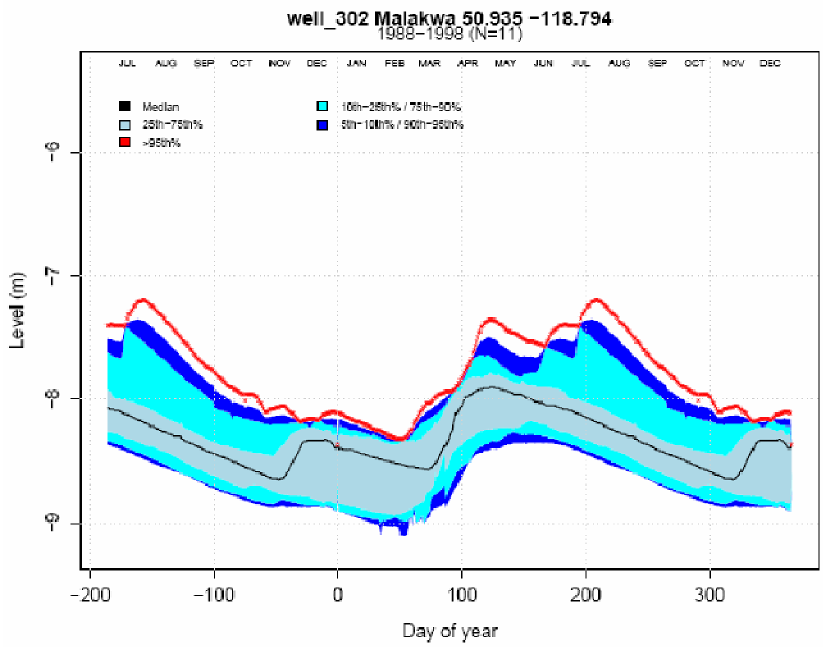


Figure 3.20: Groundwater level hydrograph for well 302.

Overall, three response groups are identified: those that have high water levels in Nov.-Mar. (rainfall regime); those that peak in Mar-Jun (snowmelt regime); and one that possibly shows evidence of two separate peaks in the year (hybrid regime), although the third response group is not as strongly defined as only a single well with this type of response was available. Within each group, there are observable differences in the shape of the hydrograph, which may be attributable to the aquifer hydraulic properties or to the connection of the aquifer to surface water. Following is a discussion of each regime and the wells that characterize each. The results are discussed with consideration given to the aquifer type, such that the proposed classification of stream-aquifer system types might be evaluated.

Rainfall regime

Rainfall regimes in BC are characterized by winter precipitation and dry summers. Within a rainfall regime, snow pack is minimal and streams are not supported by snowmelt in the summer months. Many of these streams have zero or very low discharge during the months of Jul, Aug, and Sep. Where there is streamflow, often it is supplied by groundwater discharge.

Wells 002 (Abbotsford), 065 (Saanich), 126 (Mayne) and 228 (Cassidy) have responses that are typical of a rainfall regime.

Well 002 is drilled into a sand and gravel unconfined aquifer that is classified as a Type 4a. This type of aquifer is not expected to be stream-driven (Table 3.2) for the simple reason of aquifer morphology. Significant recharge to such “raised” aquifers occurs directly from precipitation. Groundwater levels rise fairly rapidly in the aquifer following the onset of the rainy season because the aquifer is so permeable. As well, the storage capacity of these aquifers is high, leading to more gradual changes in water level as compared to aquifers with a lower storage potential for the same amount of water that is added (recharged) or removed (discharged). Consequently, for well 002 we notice a gradual rise in groundwater level and a similarly gradual decline in groundwater level over the course of a year (Figure 3.21). It is important to note that well 002 is located at some distance from any stream, and perhaps if the well had been closer to a stream, even in this type of aquifer, there may have been some component of stream-driven recharge. Nonetheless, based on this well’s response, it appears to be classified as a recharge-driven system.

Wells 065 and 126 are drilled into competent bedrock, with aquifer Types 5/6. These aquifer types are not expected to be stream-driven (Table 3.2) for the main reason that the aquifer permeability and storage coefficients are typically very low in bedrock (as compared to sand and gravel deposits). The low storage of these aquifers means that when it starts to rain in the fall, there is a limited capacity for infiltration due to the relatively low permeability of the rocks. Thus, runoff is high and the streams respond quickly to precipitation events. The groundwater levels in the aquifer rise at a rate dictated by the permeability, and reach their maximum value within a relatively short period of time due to the low storage capacity of the rocks. Similarly, due to the low permeability of the rocks, there is limited capacity for surface water in a stream to enter the aquifer. Thus, streamflow remains relatively confined to the channel. Therefore, the response of a well in a bedrock aquifer is best attributed solely to groundwater recharge from precipitation. This is also supported by comparing the plots of precipitation intensities to the groundwater levels (Figures 3.22 and 3.23, respectively for 065 and 126). In both cases, the timing of the peak intensity matches those of peaks in groundwater level. The pattern of groundwater level decline is also similar in wells 065 and 126. It takes several months for groundwater levels to drop down to their minimum levels.

The plots of groundwater level versus the log of stream discharge for these two wells and the Cowichan River station (08HA001) (which arguably cannot be used for comparison to Mayne Island, but there is no hydrometric station on Mayne Island) show that rising stream flow volumes do not correspond directly to rising water levels in the aquifer (Figure 3.24a and b). Although difficult to interpret, the reader should track the response over the course of a three year period by following the black (1989-1990), red (1990-1991) and green (1991-1992) traces for each month (different symbols are used for each month). In these two wells, the points are scattered, which illustrates the lack of correlation with streamflow. For example, as stream discharge increases, the groundwater levels remain the same for some time (e.g., there is a horizontal line of data points above the x-axis). As stream discharge decreases groundwater levels do not drop, instead they remain constant for some time. Cross-correlation is relatively low ($R^2 < 0.50$) and the groundwater response lags that of the streamflow (Figure 3.25a, b, c).

The last well in this group, 228, is situated in highly productive aquifer, with potentially high permeability and storage coefficients. The aquifer is a Type 4b, which is considered to be possibly stream-driven, particularly if a stream is located nearby and is connected to the confined aquifer (Table 3.2). Comparing the groundwater level to climate data (Figure 3.26) as well as the hydrographs for well 228 and station 08HB034 on the Nanaimo River, respectively (Figure 3.27), it is apparent that water levels in the aquifer start to rise roughly a month later than streamflow levels start to rise. Furthermore, there is a consistent correlation between the shape of the two hydrographs throughout the year. These results suggest that well 228 is stream-driven; Haslam Creek is situated 107 m from the observation well. Figure 3.25c shows the cross-correlation plot for well 228 and hydrometric station 08HB034, which suggests that the correlation is weak ($R < 0.5$), and that there is a lag.

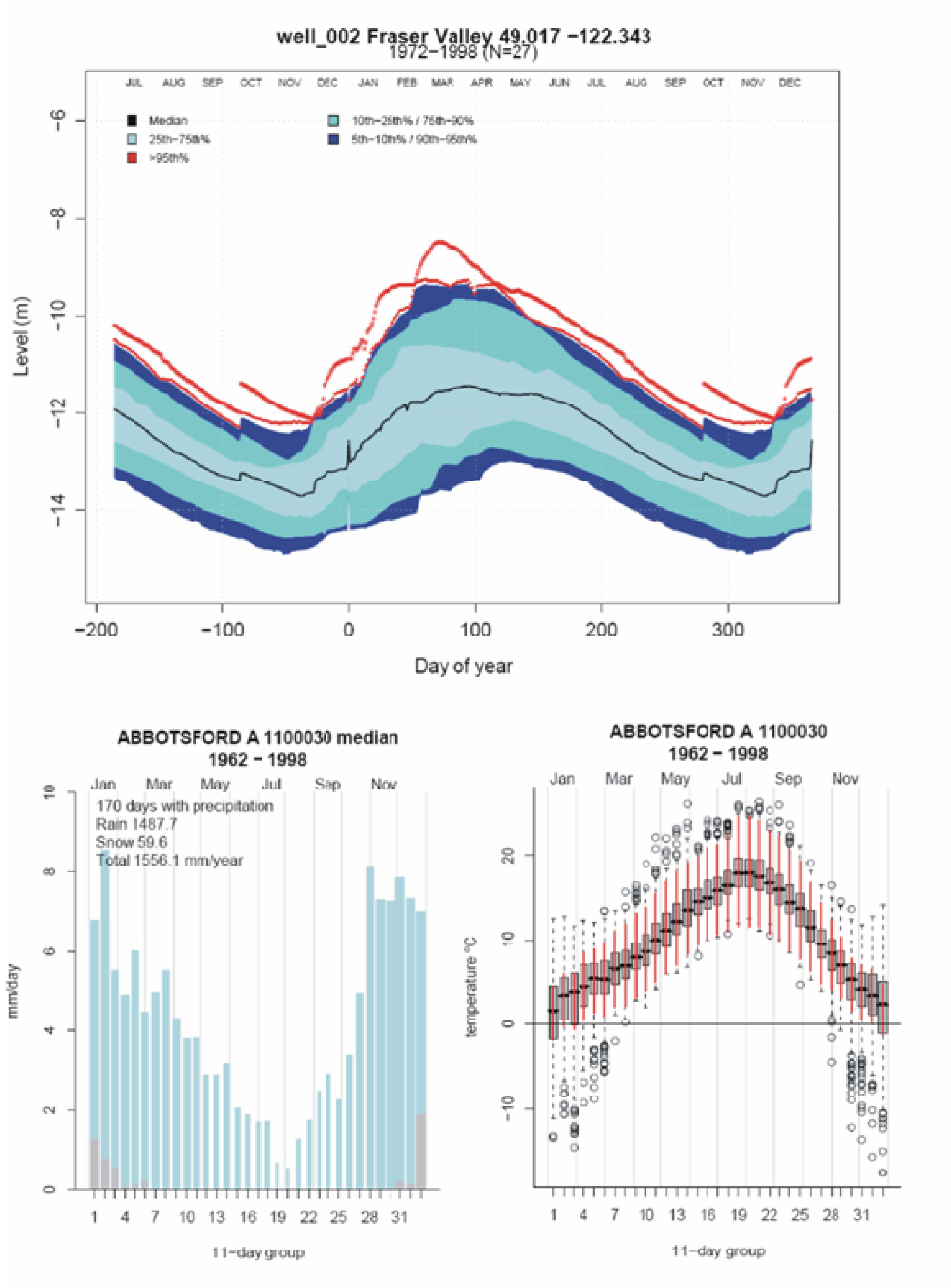


Figure 3.21: (a) Well hydrograph for well 002, and (b) precipitation and (c) temperature normals for Abbotsford International Airport.

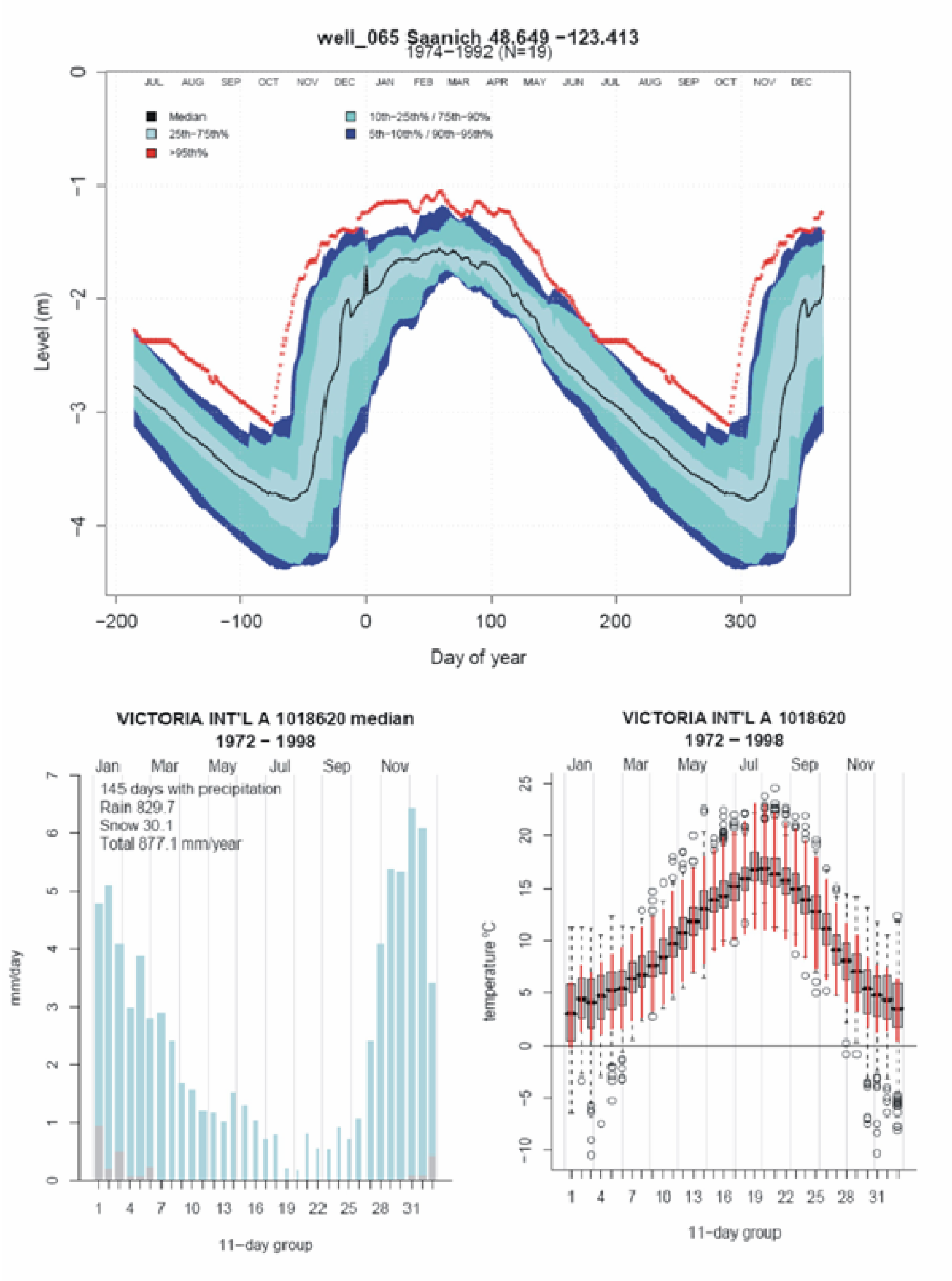


Figure 3.22: (a) Well hydrograph for well 065, and b) precipitation and (c) temperature normals for Victoria International Airport.

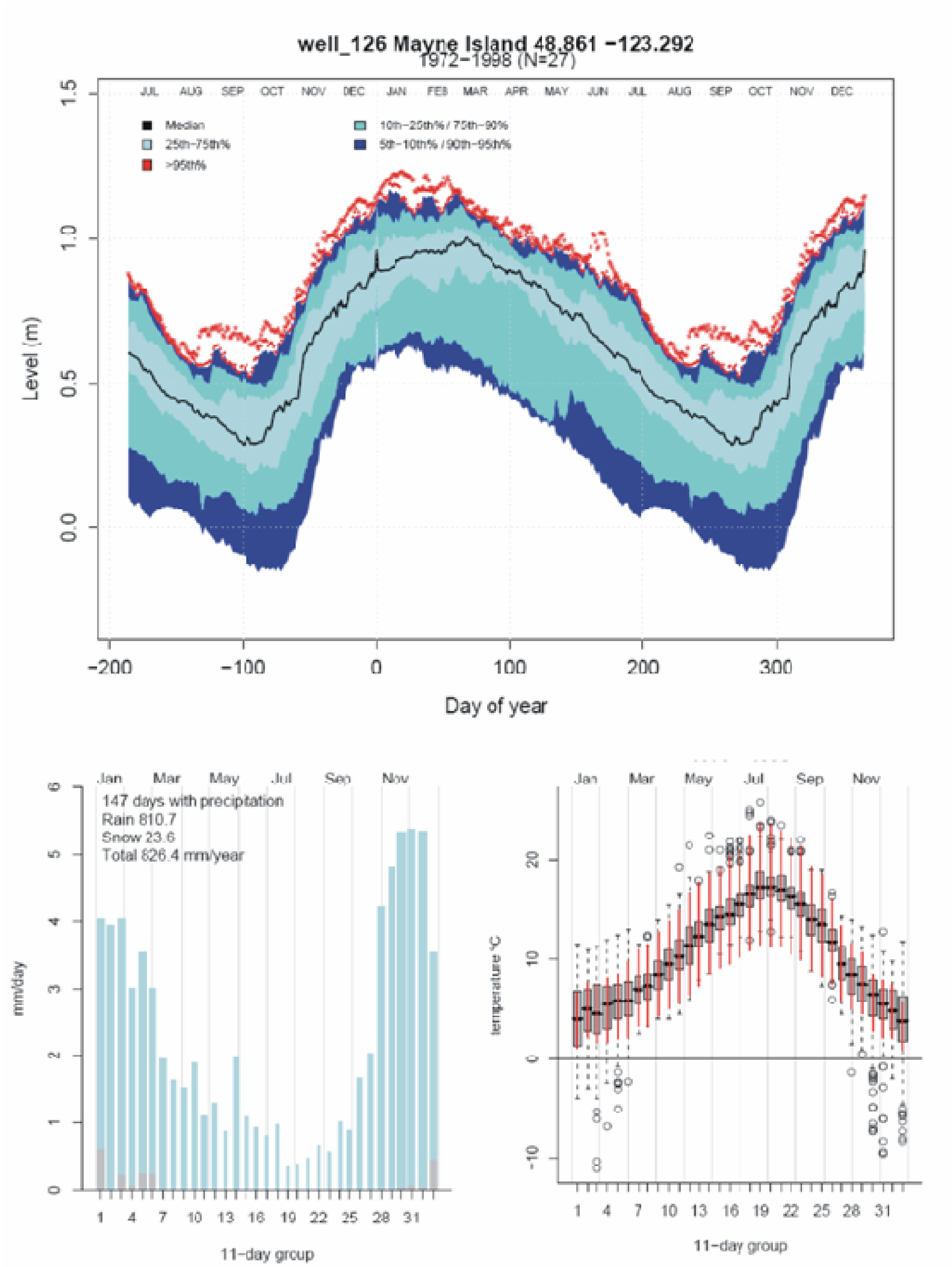


Figure 3.23: (a) Well hydrograph for well 126, and b) precipitation and (c) temperature normals for Mayne Island.

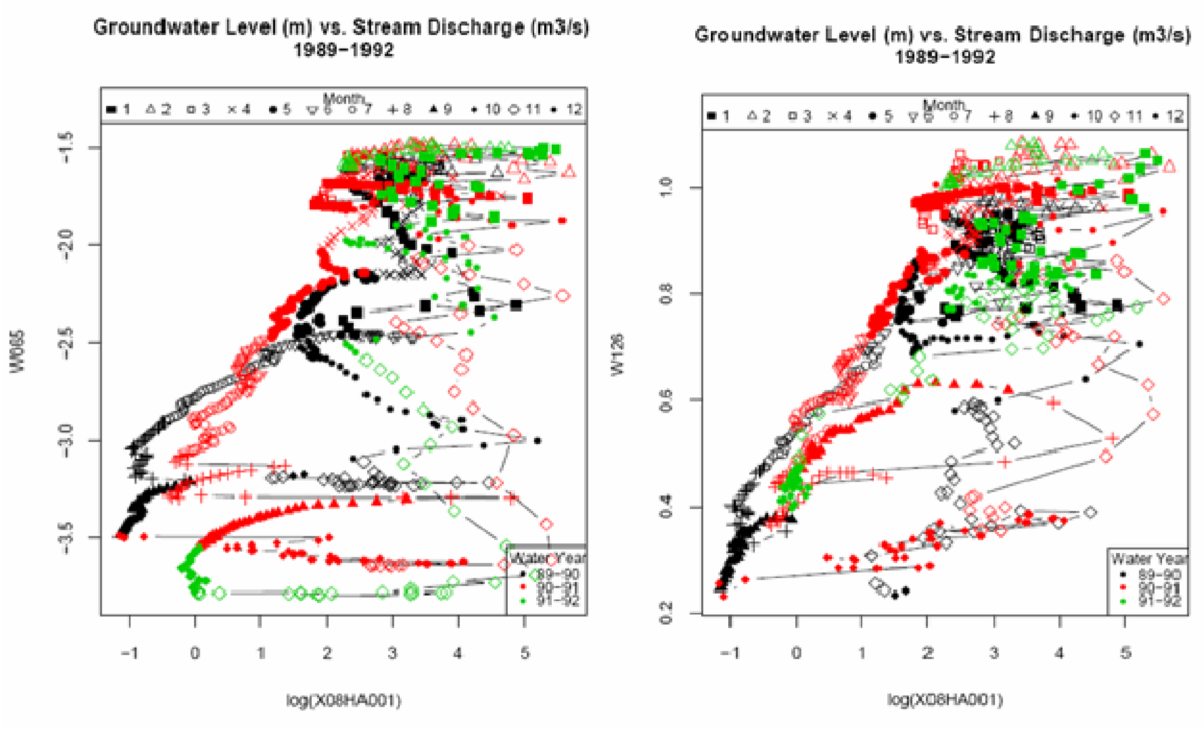


Figure 3.24: Groundwater level versus log stream discharge for (left) 065, and (right) 126.

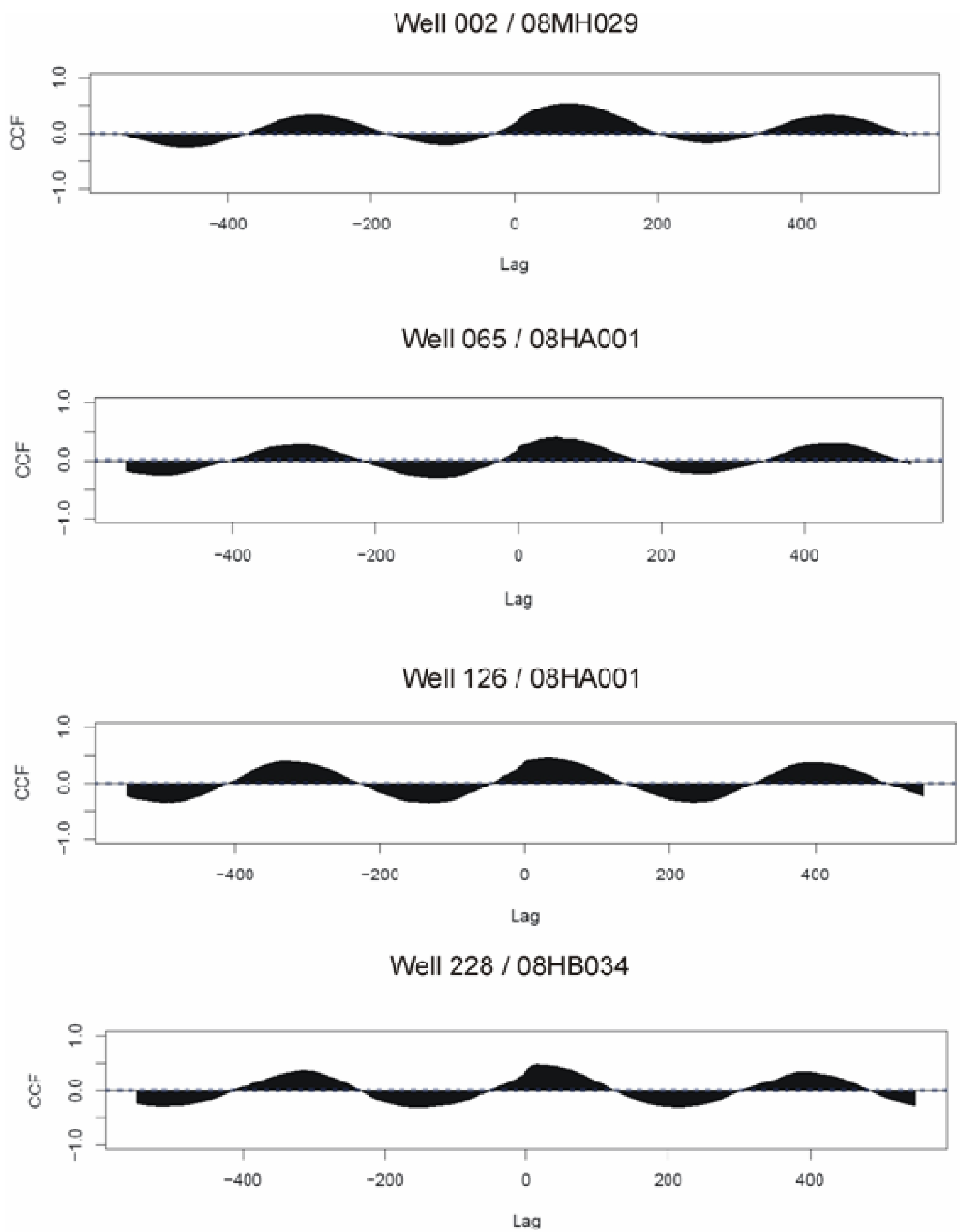


Figure 3.25: Cross-correlation for wells 002, 065, 126 and 228 with their respective hydrometric stations.

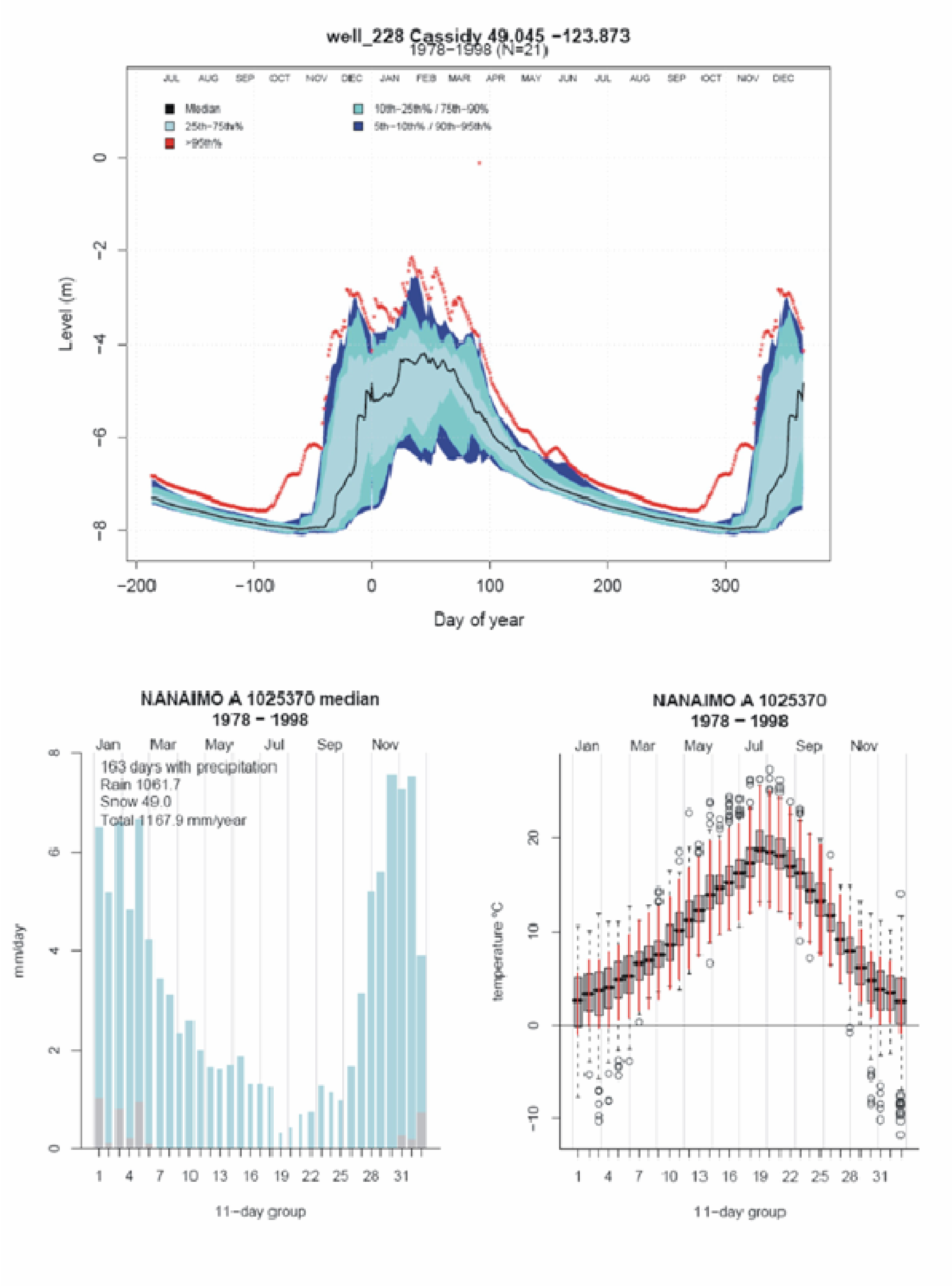
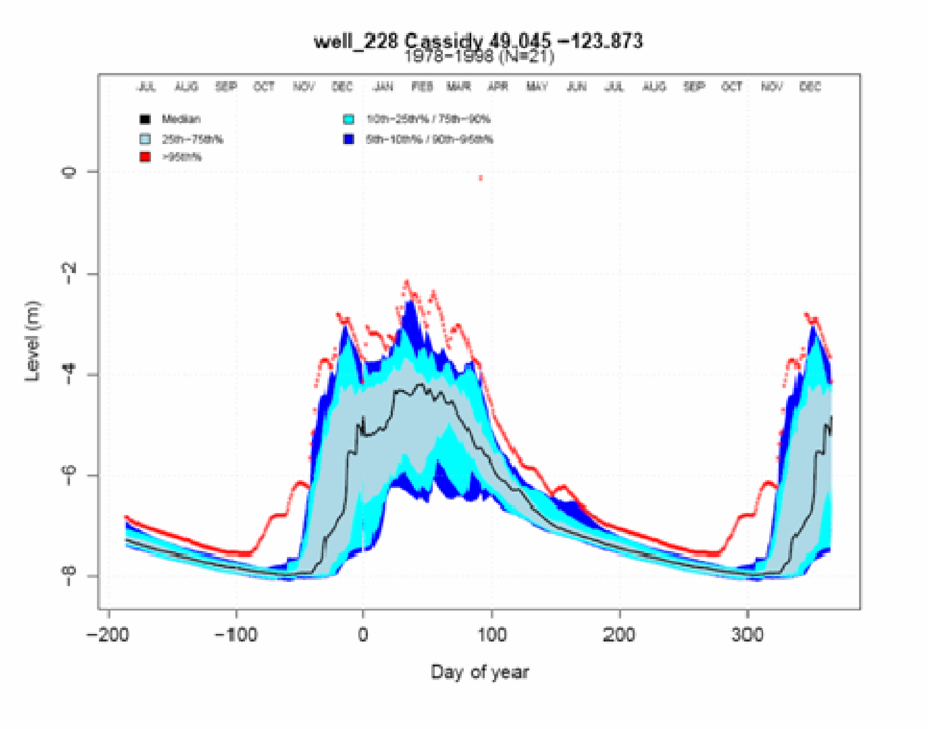


Figure 3.26: (a) Well hydrograph for well 228, and b) precipitation and (c) temperature normals for Nanaimo.



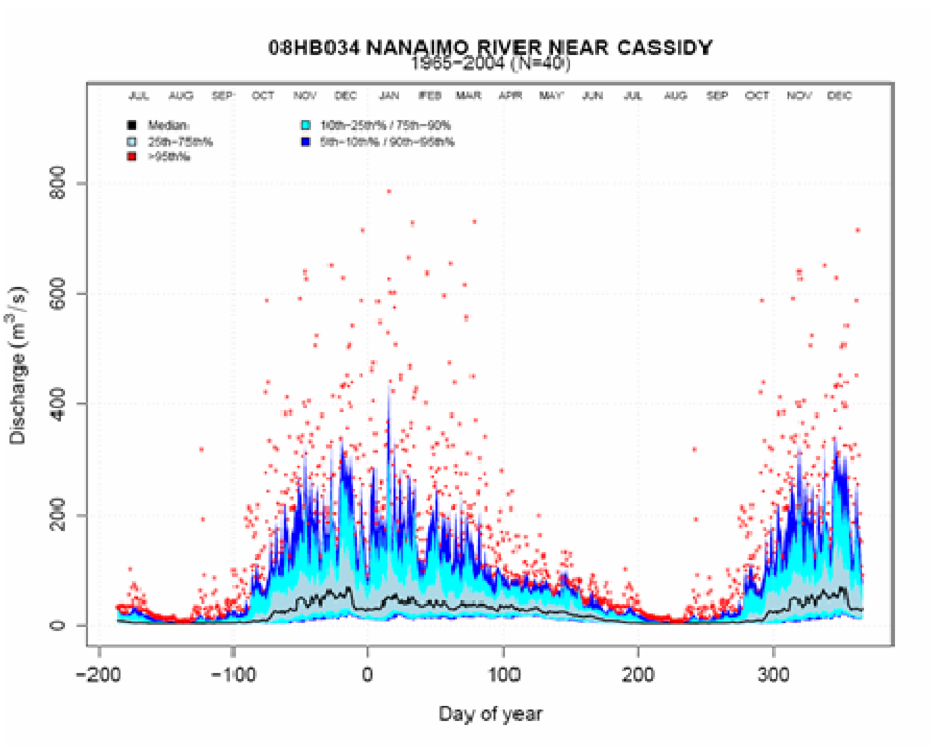


Figure 3.27: (top) Well 228 hydrograph and (bottom) stream hydrograph for 08HB034 Nanaimo River near Cassidy.

Snowmelt regime

Snowmelt regimes in BC are characterized by accumulation of winter precipitation in the form of snow, followed by spring runoff and a dry summer period. Significant snow depths fall from Oct.-Mar. Air temperatures start to increase in mid-Mar., on average, bringing the mean air temperature above the $0^{o}C$ melting point of the available snow pack. Rainfall rates are greatest from May-Jul. in this region. These rainfall events are thought to be brought on by convective rather than frontal patterns, which recycle water from the nearby area limiting the input of new water to the system. Often this rain is evaporated before it has time to infiltrate the land surface. In fact, many regions of the interior (e.g., Okanagan) experience recharge deficits during the summer due to high evaporation.

Within a snowmelt regime, streamflows are supported by snowmelt into late spring and early summer months. Streams peak between April and mid-June in this region. Baseflow in streams is typically supported by discharging groundwater through the months of July, August, and September or by a combination of glacier runoff and groundwater discharge.

Wells 217 (Grand Forks), 309 (Golden), 296 (Merritt) and 054 (Carr's Landing) are situated in this area. Two wells (217 and 296) are situated in highly productive Type 1a aquifers, which are characterized as stream-driven. The response at well 217 is well understood due to previous numerical modeling of the interaction between stream flow and groundwater in the Grand Forks aquifer (Allen et al., 2004b). The modeling demonstrated that groundwater levels in the floodplain area of the aquifer (well 217 is in this area) are strongly correlated with stream levels. At peak flow, river water recharges the aquifer and moves laterally away from the channel, causing groundwater levels to rise over a broad area. Within 30-60 days following the peak discharge, when river levels begin to fall, the groundwater direction is reversed and contributes to baseflow. Water balance results from the model indicate that the river-aquifer interaction has a maximum flow rate between 11 and 20% of river flow during spring freshet – on average, the river contributes about 15% of its spring freshet flow into aquifer storage. Direct recharge to the aquifer from precipitation also results in a rise in water levels; however, the overall effects of direct recharge are very small in comparison to the indirect recharge from the rivers, except in areas removed from the influence of the rivers (i.e. the benches along the valley sides). Figure 3.28 shows the well hydrograph for well 217 along with the precipitation and temperature normals for Grand Forks. Figure 3.29 shows the well hydrograph along with the stream hydrograph for the Kettle River. Groundwater levels lag slightly behind stream levels.

Well 296 (Merritt) is situated in a sand and gravel deposit in between two major streams. The aquifer is highly productive. Figure 3.30 shows the well hydrograph and climate normals. Groundwater levels do not appear to climb until discharge volumes are at their highest in the Nicola River (near Merritt) and then they climb very steeply (Figure 3.31). The response indicates that this aquifer is largely stream-driven. Similar to Grand Forks (well 217), groundwater levels lag slightly behind stream levels.

Well 309 (Golden) is situated within a Type 4b confined aquifer, which may be stream-driven if the well is situated close to a stream. Similar to the Cassidy well (228), this type of aquifer may or may not be stream-driven, depending on the nature of the connection between the confined aquifer and the stream and the location of the well. Figure 3.32 shows the well hydrograph for well 309 and the climate normals, while Figure 3.33 shows the groundwater hydrograph and the hydrograph for the Columbia River at Nicholson. There is a strong correspondence between the two hydrographs, suggesting that the aquifer at this location is stream-driven.

Plots of groundwater level versus the log of discharge for these nival systems (217, 296 and 309) shows their close relationship (Figure 3.34). Water levels increase as soon as stream discharge volumes increase, and decrease as soon as stream discharge volumes decrease. A hysteresis effect is evident in all three aquifers.

The cross-correlation of groundwater levels to stream discharge demonstrates that groundwater lags stream discharge slightly, but that this response is highly correlated for part of the year ($R2 > 0.8$) for wells these wells (Figure 3.35).

Well 054 (Carrs Landing) is completed in a Type 5a or 6b (unknown bedrock) and is situated at high elevation some distance from any stream. Its response would be most likely dominated by snowmelt, the timing of which is likely similar to that reflected in a hydrograph of any nearby streams; however, the response would be delayed due to the low permeability of the rock. In this particular case, the well at 054 is relatively shallow, and so it likely responds a bit more quickly than a deeper well might in the same area. Figure 3.36 shows the well hydrograph for 054 along with the climate normals. A weak cross-correlation is observed (Figure 3.37). The groundwater recession appears to be delayed, similar to what was observed in the other bedrock wells (065 and 126) in the rainfall regime.

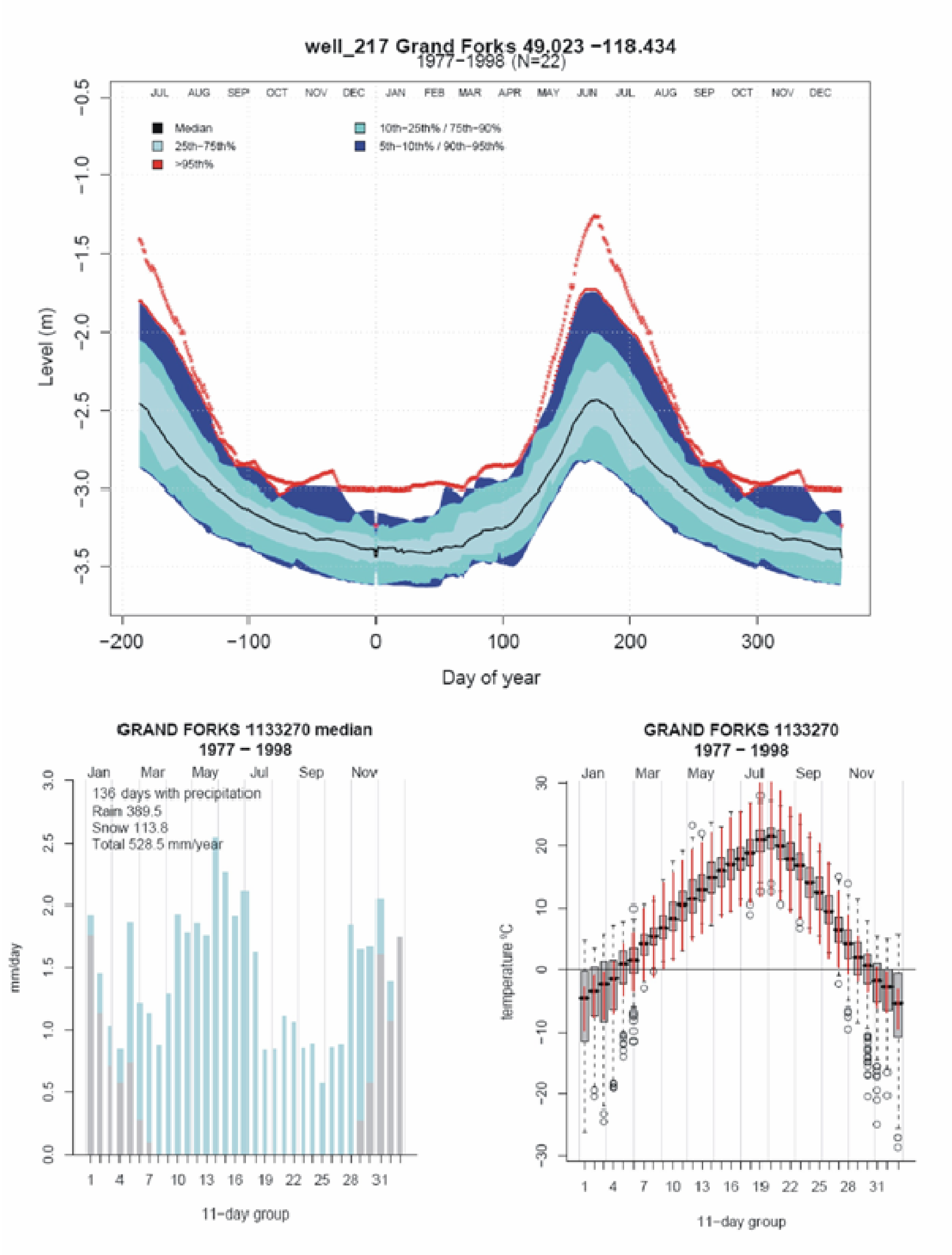
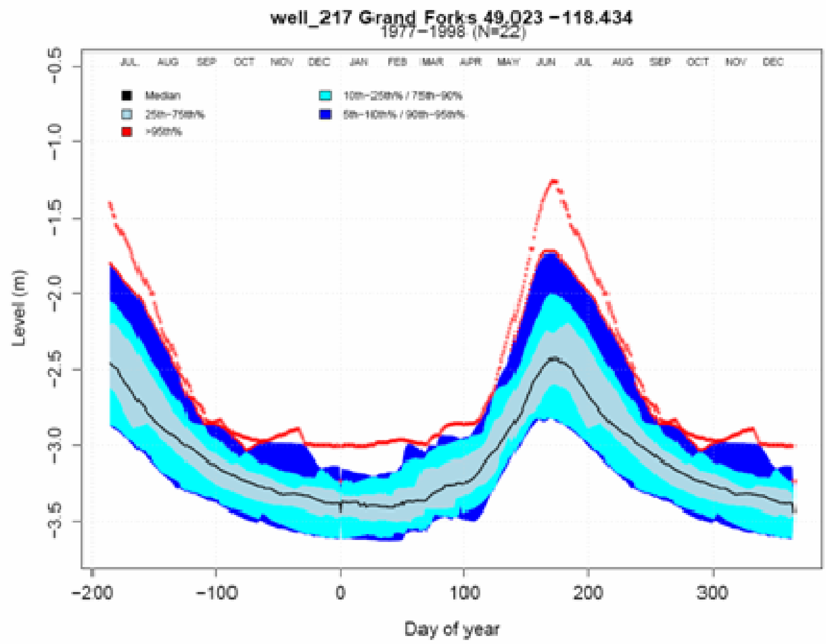


Figure 3.28: (a) Well hydrograph for well 217, and b) precipitation and (c) temperature normals for Grand Forks.



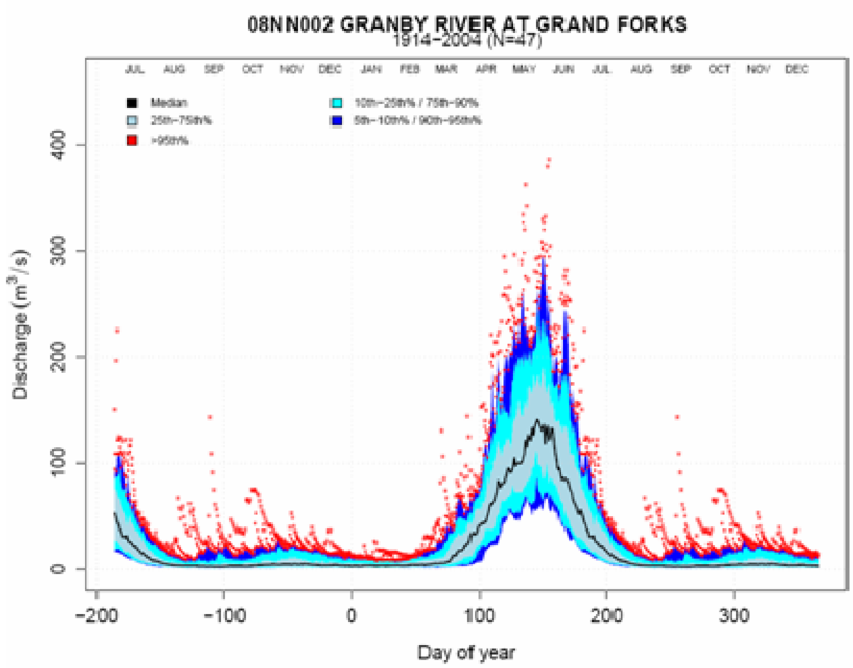


Figure 3.29: (top) Well 217 hydrograph and (bottom) stream hydrograph for 08NN002 Granby River near Grand Forks.

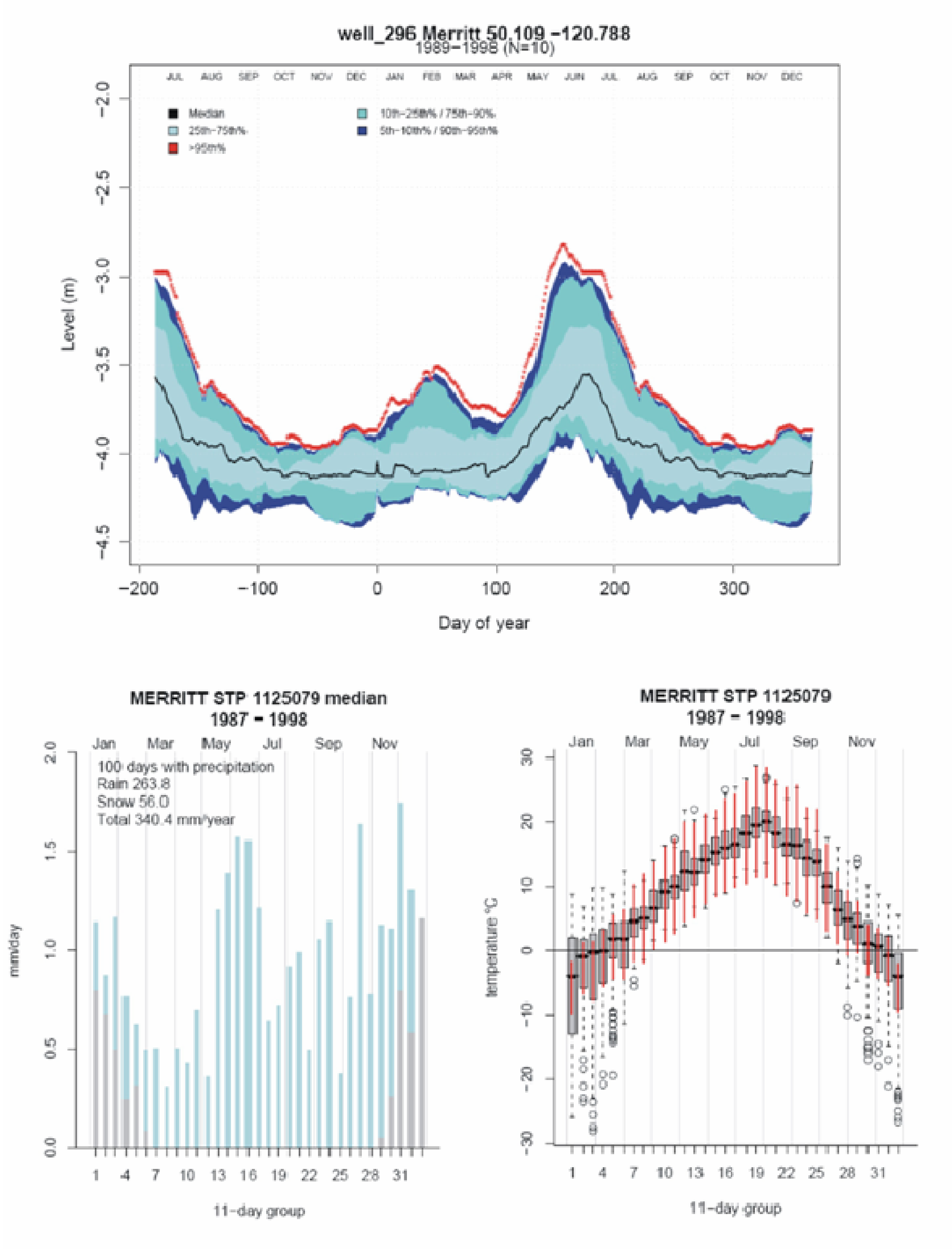
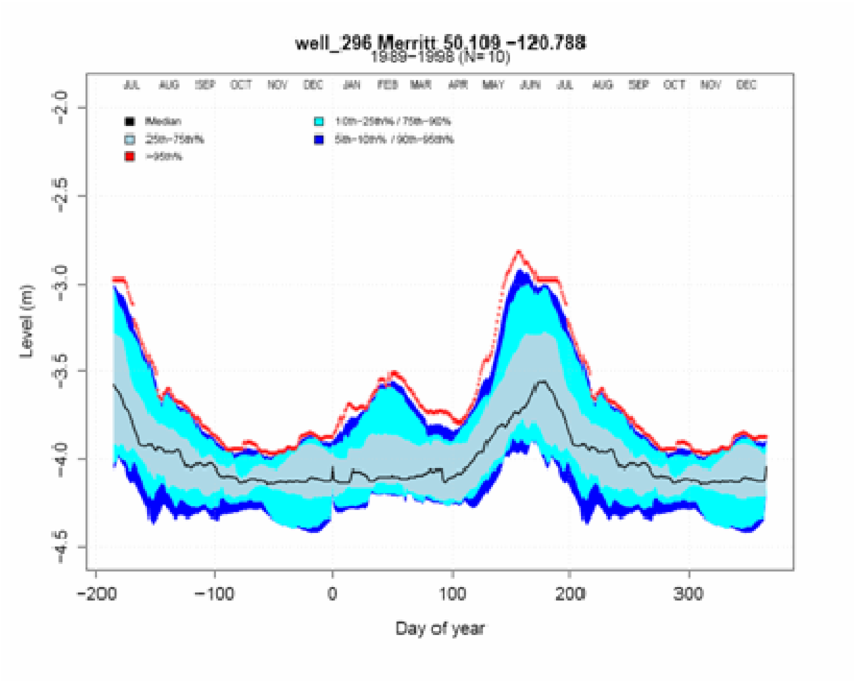


Figure 3.30: (a) Well hydrograph for well 296, and b) precipitation and (c) temperature normals for Merritt.



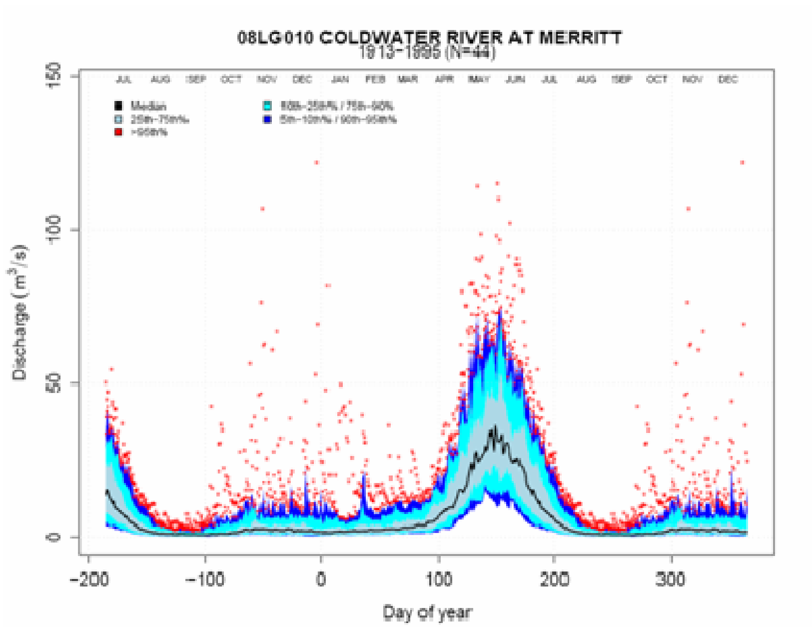


Figure 3.31: (top) Well 296 hydrograph and (bottom) stream hydrograph for 08LG010 Coldwater River at Merritt.

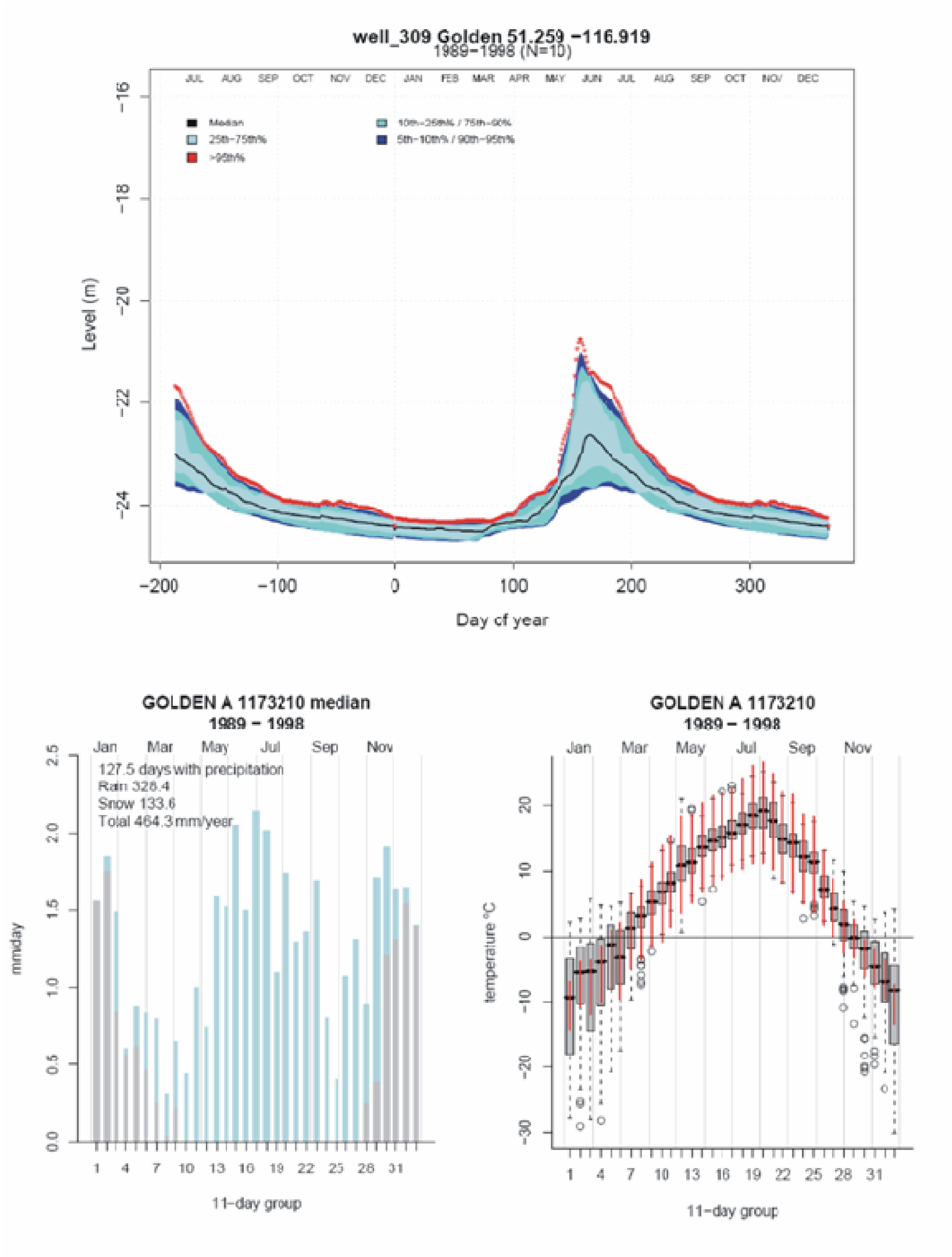


Figure 3.32: (a) Well hydrograph for well 309, and b) precipitation and (c) temperature normals for Golden.

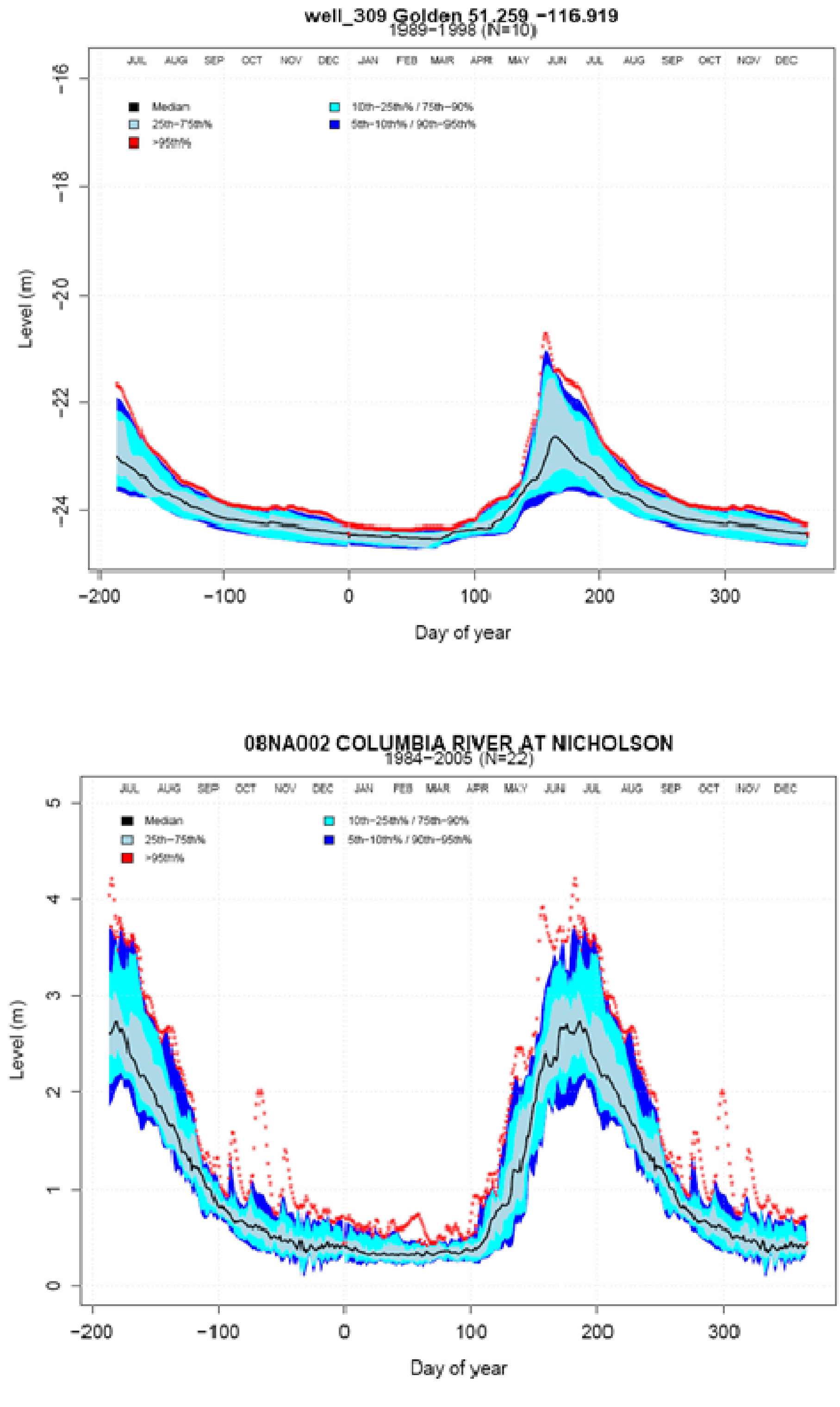


Figure 3.33: (top) Well 309 hydrograph and (bottom) stream hydrograph for 08NA002 Columbia River at Nicholson.

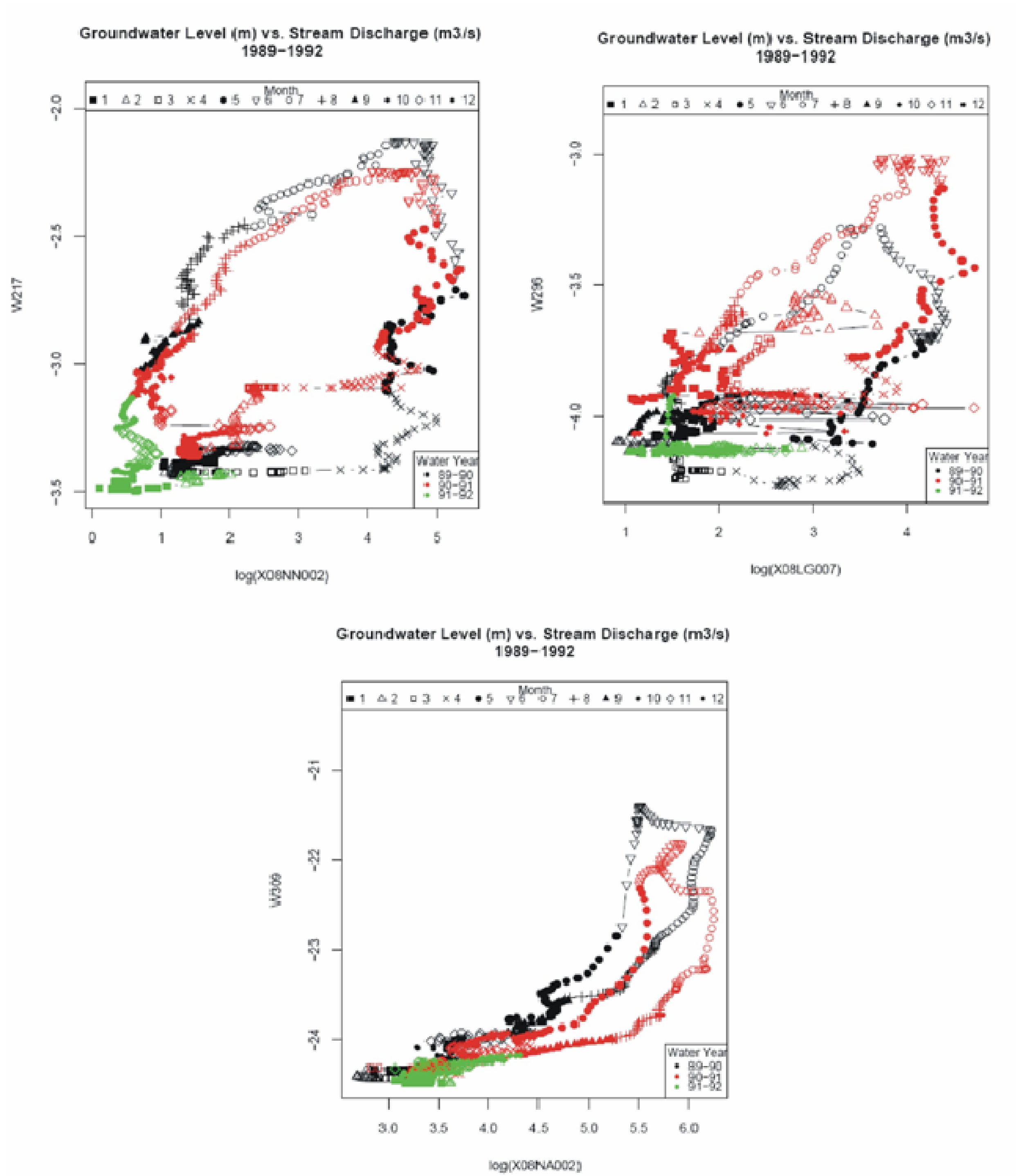


Figure 3.34: Groundwater level versus log stream discharge for 217, 296 and 309.

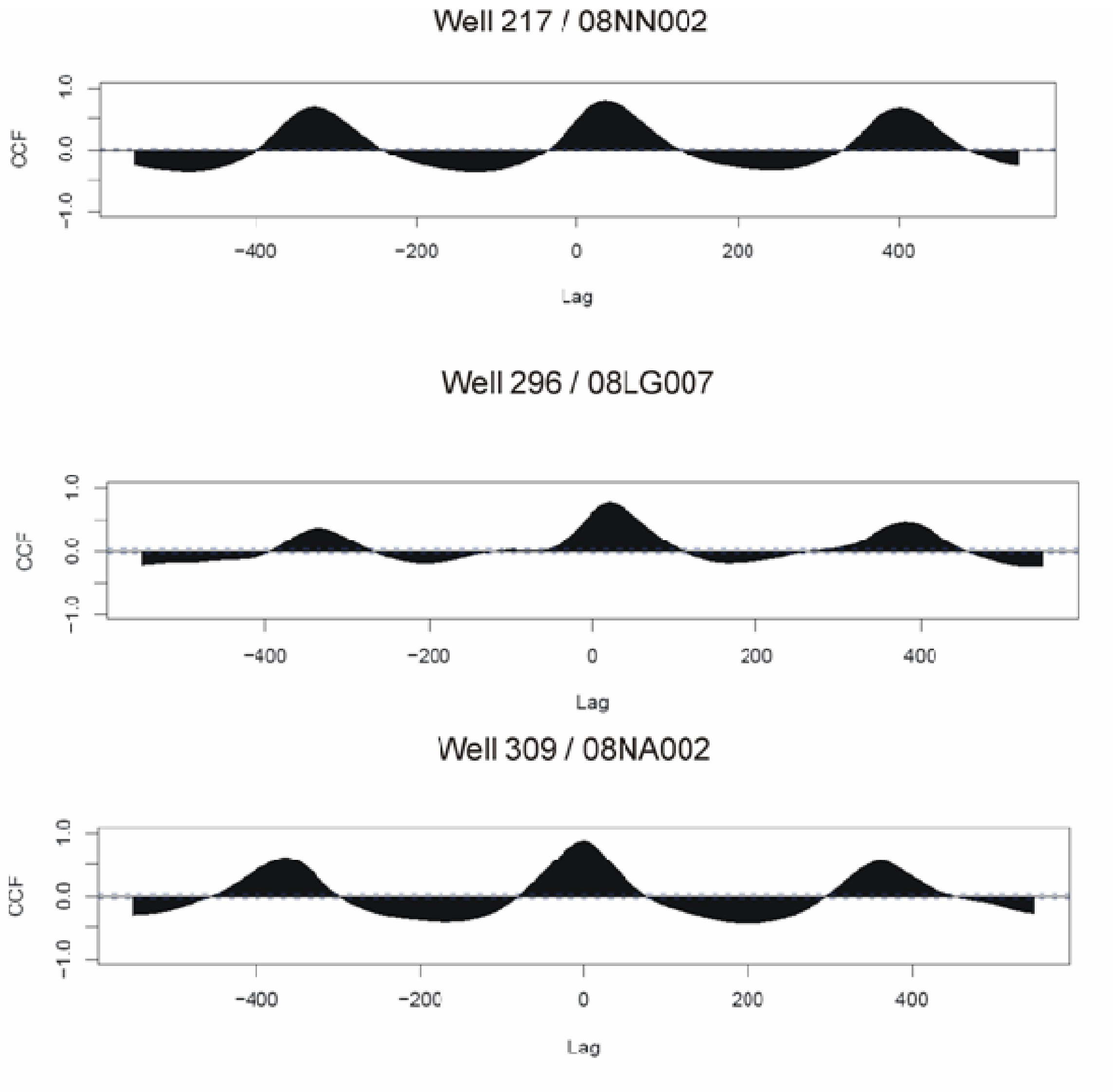


Figure 3.35: Cross-correlation for wells 217, 296 and 309 with their respective hydrometric stations.

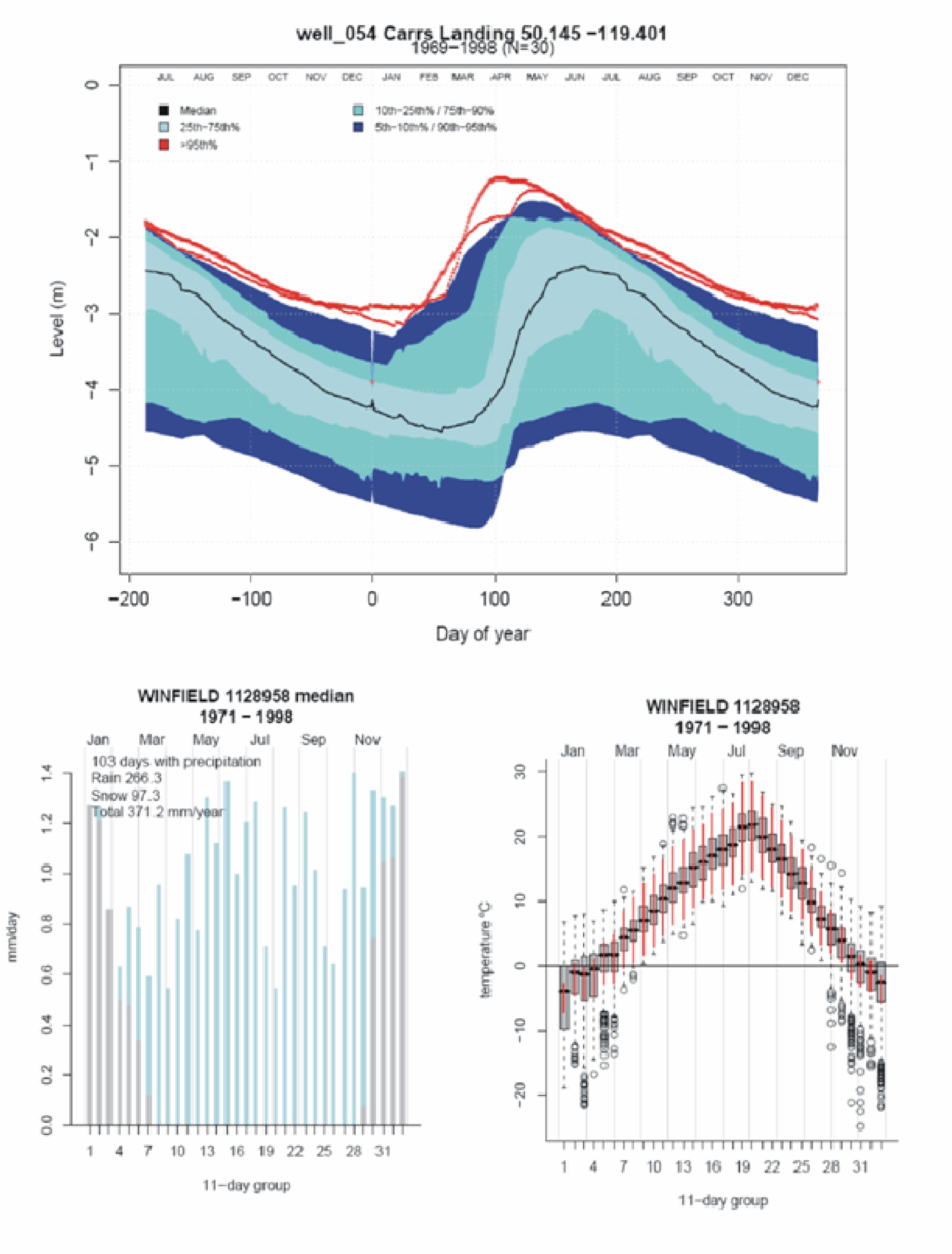


Figure 3.36: (a) Well hydrograph for well 054, and b) precipitation and (c) temperature normals for Winfield.

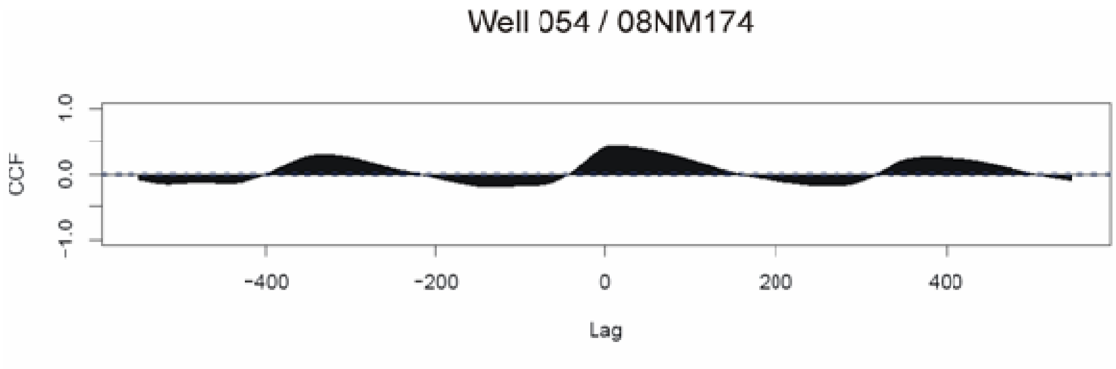


Figure 3.37: Cross-correlation for wells 054 with its respective hydrometric station.

Hybrid regime

Well 302 (Malakwa) is possibly situated in a hybrid regime. Well 302 is situated some 300m from a stream and in a valley bottom aquifer, corresponding to a Type 1a. Thus, the aquifer is expected to be stream-driven. The groundwater levels appear to rise to a certain level by winter rains and are then topped up with stream discharge during May, June and July (Figure 3.38). During some years water levels peak in Jul. and Aug. suggesting the glacier melt is augmenting the water levels in this aquifer. The groundwater level to stream discharge plot for 302 shows that the groundwater response is somewhat tied to the stream discharge of the Eagle River, but the discharge in the river has to increase to a certain level before the groundwater will start to rise (Figure 3.39). The cross-correlation shows that the relationship between the groundwater and stream discharge is strong (R2 > 0.50) and the timing of groundwater level and stream discharge change is closely tied (Figure 3.40). The rise in the groundwater level between Nov.-Mar. in this aquifer can be attributed to the intense precipitation events that occur in early Nov. in this area. This aquifer demonstrates a mixed response to rainfall, snowmelt, and glacier runoff.

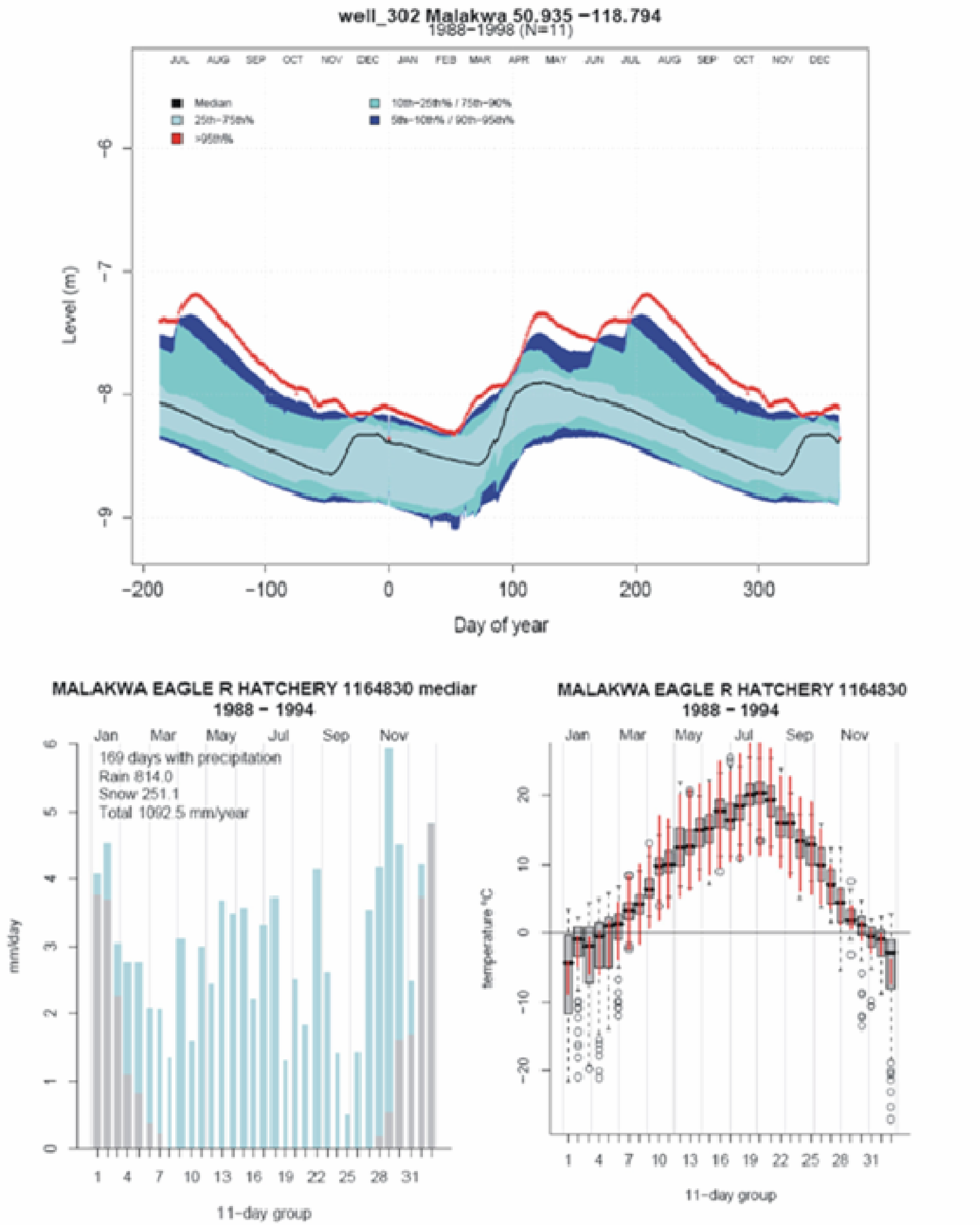


Figure 3.38: (a) Well hydrograph for well 302, and b) precipitation and (c) temperature normals for Malakwa Eagle River Hatchery.

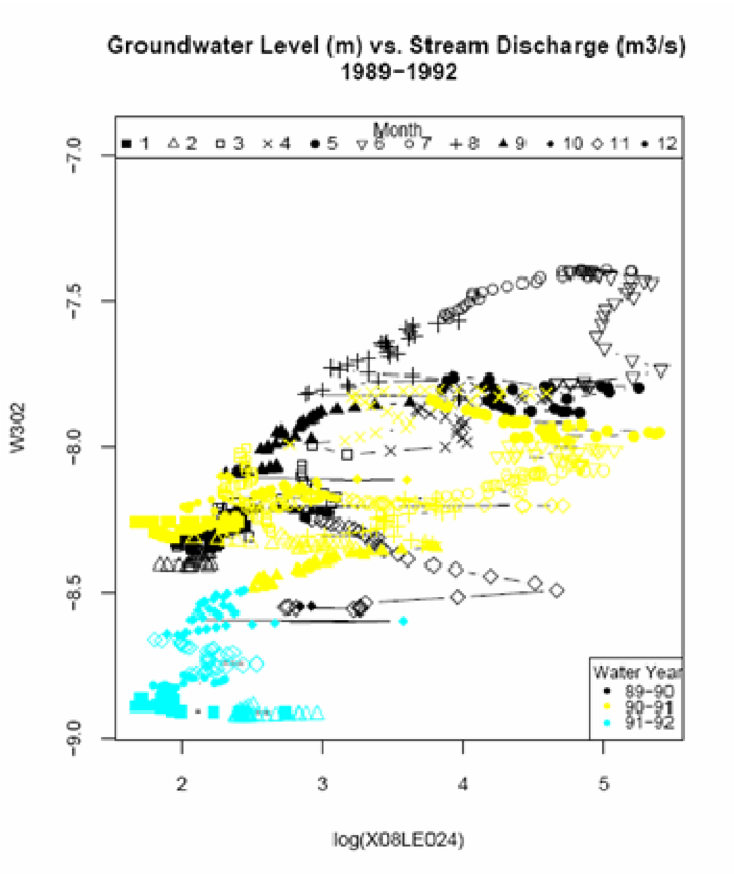


Figure 3.39: Groundwater level versus log stream discharge for 302.

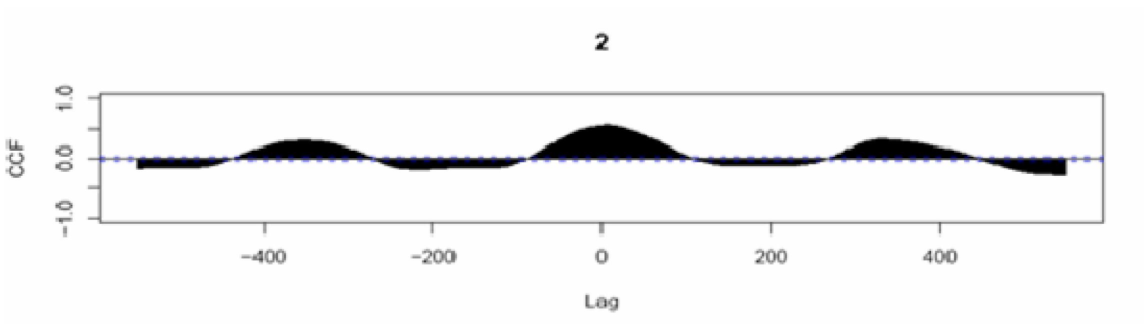


Figure 3.40: Cross-correlation for well 302 and 08LE024.

3.6.7 Summary

The responses of aquifers, as exhibited by groundwater level variations in wells, are a function of hydrogeology of the aquifer, including the aquifer properties and proximity or interconnection with streams, and partly by climate. Two dominant stream-aquifer system types have been identified based on classifying different aquifer types. The recharge-driven systems include primarily bedrock aquifer types (Type 5 and 6) and possibly some occurrence of sand and gravel aquifers of glacial or pre-glacial origin where there is little opportunity for interaction with streams (Type 4). The stream-driven systems include fluvial/glaciofluvial aquifers (in stream valleys), deltaic aquifers (Type 2) or alluvial/colluvial fan aquifers (Type 3). Each one of these aquifer types may be found within any climate regime: rainfall regime, snowmelt regime, hybrid regime, thus, there are many possible combinations of potential responses of wells.

Of the nine observation wells considered in this preliminary analysis, not all combinations were available; thus, the classification scheme, as it is presented cannot be rigorously tested. Nonetheless, the responses of those wells considered appear to correspond to what was anticipated. The classification scheme does provide a framework for evaluating the responses of wells in other regions of the province. Specifically, this framework will be used to classify a larger subset of wells throughout the province for trend analysis as discussed in the following section.

3.7. Groundwater trend analysis

3.7.1 Data selection and preparation

Most long observation well records in British Columbia consist of end-of-month manual readings. Some provincial observation wells are equipped with Stevens recorders (end-of-month readings only), but several were recently equipped with data loggers that record hourly or daily groundwater levels. For the trend analysis presented in this section all available records that cover the time period from 1980 or earlier until at least the year 2002 were extracted from the BC Ministry of Environment website, including end-of-month only series, and series that have mixed end-of-month and (later) daily records. For mixed series, the daily part of the record was aggregated to end-of-month values to obtain longer time series. The procedure is explained in detail in Appendix 1, which also shows the results of a test for representativeness and a discussion of some of the problems involved.

Records with more than 20% missing data were excluded. The time series were plotted and visually inspected for inhomogeneities such as jumps, changes in annual amplitude, etc. Finally, 37 observation well records remained (Table 3.3).

Aquifer information and distance to the closest stream were determined for all wells from the information available in the Water Resources Atlas (http://www.env.gov.bc.ca/wsd/data_searches/wrbc/index.html) and the WELLS database (http://aardvark.gov.bc.ca/apps/wells/). The well records were classified by the criteria discussed in the previous section, i.e. by their seasonality regime (rain, snowmelt) and their recharge mechanism (driven by recharge from precipitation only or additionally through a stream).

For the 37 monthly records, a common period from 1976 to 2002 was chosen for the analyses. Where records started later, the earliest year after 1976 was selected as the starting date. This period was chosen because of the best overlap with the streamflow and climate records used in the project. The starting year coincides with the shift to the positive PDO phase, a situation that is predicted to become more dominant in the future climate.

Table 3.3: Well records

Well	Name	Depth (m)	Geology	Type	Productivity	Vulnerability	Distance to river (m)	Stream-flow data available
Rain-dominated								
002	Fraser Valley	19	Sand and Gravel	4a	high	high	3000	
015	Fraser Valley	96	Sand and Gravel	4a	high	high	350	x
060	Saanich	16	till	4b	na	na	coast	
065	Saanich	152	bedrock, granodiorite	6b	low	moderate	coast	
126	Mayne Island	70	bedrock	5a	low	high	coast	
128	Mayne Island	68	bedrock	5a	low	high	coast	
211	Duncan	31	sand and gravel	4b	high	low	160	
228	Cassidy	65	sand and gravel	4b	high	low	107	x
258	Galiano Island	90	bedrock	5a	low	moderate	coast	
268	Denman Island	312	bedrock	5a	low	high	na	
Snowmelt-dominated								
35	Stump Lake	na	likely sand and gravel	1a	na	na		
45	Westwold	na	sand and gravel	3a	na	na		

Well	Name	Depth (m)	Geology	Type	Productivity	Vulnerability	Distance to river (m)	Stream-flow data available
Snowmelt-dominated								
47	Silverstar U	90	clay/sand/grav over metamorphic	6b	na	na	40	
53	Carrs Landing	15	Likely bedrock	5a or 6b	na	na	250	
54	Carrs Landing	14	sand/gravel bedrock depth 15, up on ridge at 1000 m.a.s.l.	5a or 6b	na	na	230	
81	83 Mile	150	bedrock (Basalt)	6a	moderate	moderate		
96	Osoyoos	11	sand and gravel	4a or 4b	moderate	high	next to little pond	
100	Osoyoos	19	sand and gravel	4a or 4b	moderate	high	between lake and canal/ditch	
105	Osoyoos	13	sand and gravel	4a or 4b	moderate	high	between lake and canal/ditch	
107	Osoyoos	10	sand and gravel	4a or 4b	moderate	high	between lake and canal/ditch	
115	Mission Creek	22	till and boulders, fractured bedrock, 1800m.a.s.l.	6b	na	na	340 m to alpine lake,	
117	Armstrong	188	sand silt	4a or 4b	na	na	na	
118	Armstrong	471	sand/silt	4a or 4b	moderate	low	150	
119	Armstrong	177	sand and gravel	4a or 4b	moderate	low	100	

Well	Name	Depth (m)	Geology	Type	Productivity	Vulnerability	Distance to river (m)	Stream-flow data available
Snowmelt-dominated								
153	Summerland	7	sand and gravel	2	low	moderate	90m to lake on alluv fan	
154	Summerland	15	sand and gravel	2	low	moderate	on alluvial fan into Ok. Lake	
162	Kalawoods	4	Na/ 9m to bedrock	4a or 4b	na	na	70	
172	Kalawoods	20	gravel	4a or 4b	high	high	35	
173	Kalawoods	19	gravel	4a or 4b	high	high	70	
174	Kalawoods	41	gravel	4a or 4b	high	high	120m to lake	
175	Kalawoods	23	na	4a or 4b	na	na	275	
176	Kalawoods	51	na	4a or 4b	na	na	280 to lake	
185	Salmon River	10	sand and gravel	1a	mod-high	low-high	345	x
217	Grand Forks	9	sand and gravel	1a	high	high	600	x
264	Mt. Kobau	na	bedrock	6b	na	na	na	
296	Merritt	17	sand and gravel	1a	high	high	240	x
309	Golden	44	sand and gravel	4b	moderate	moderate	1	x

3.7.2. Methodology: seasonality and trends

To obtain a seasonal cycle comparable among the different wells, monthly groundwater levels were standardized and averaged over the period. The mean monthly levels were then used to manually classify the wells into rainand snowmelt-dominated seasonal cycles.

Then, temporal trends of groundwater levels and a simple recharge measure were calculated using the non-parametric Spearman's rank correlation coefficient, rs, and a p-value of <0.05 to determine whether the trend was significant. Trends for the observation well levels were calculated for annual series of each individual calendar month. To determine trends in the annual recharge, first series of differences between the annual minimum and maximum levels were derived. For rain-dominated basins, the minimum level during summer was deducted from the maximum level in the following winter. For snow-dominated systems, the minimum level during winter was deducted from the maximum of the following spring/summer season. Hence larger values should indicate more recharge.

In addition to the trend analysis on the entire dataset, for selected wells, where a hydraulic connection between groundwater and river flow is presumed and where data are available, trend analyses were also performed on the precipitation and streamflow records for the longest overlapping time period. This facilitates the attribution of the trends and allows the results of the trend analysis to be discussed within the context of the stream-aquifer system type.

The following results (Section 3.7.3) are an overview of the seasonality and trends found for the entire provincial dataset. Sections 3.7.4 and 3.7.5 then take a closer look at regional cases in rain and snowmelt dominated system where streamflow and groundwater records were available. Section 3.7.6 discusses trends in wells not influenced by streamflow.

3.7.3 Results: overview of seasonality and trends

Seasonality

Figure 3.41 shows the final dataset of observation wells and the primary classification into rainor snowmelt-dominated seasonality of the recharge. Rain-dominated groundwater systems are found on Vancouver Island or along the coast, while all wells in the interior reflect the winter snow storage and recharge by spring snowmelt. Unfortunately, there are no long records available in the central and northern parts of the province.

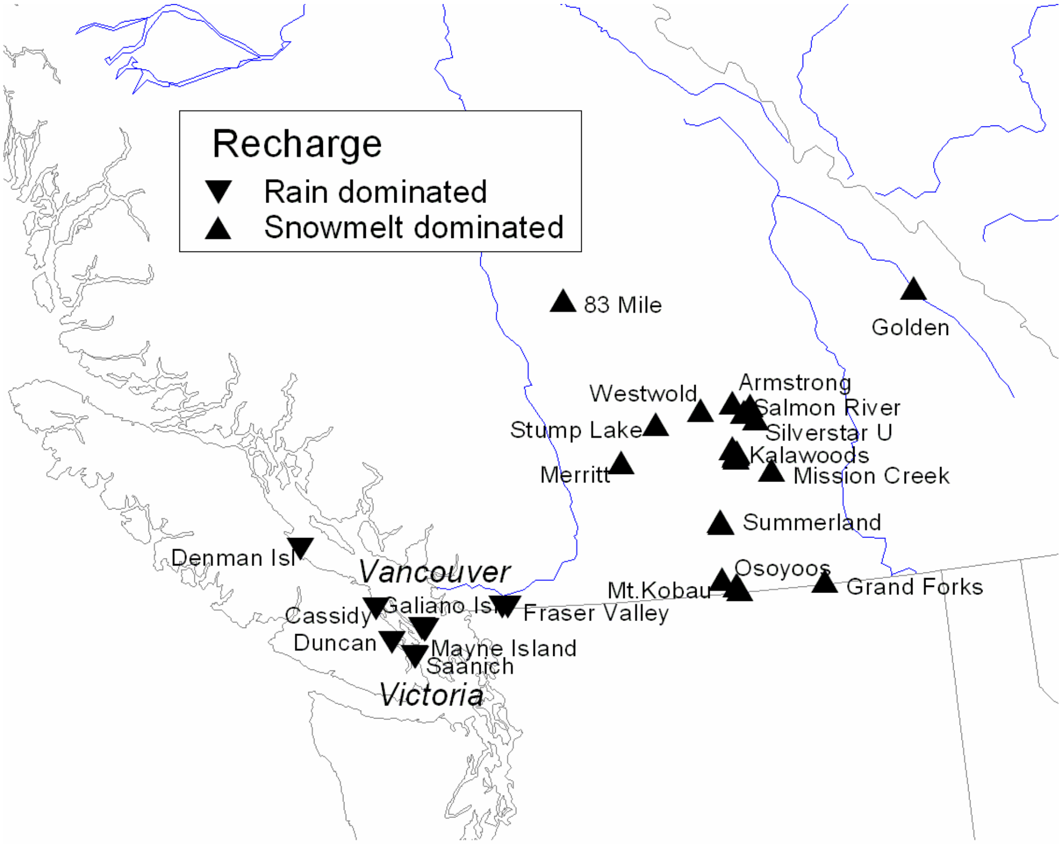


Figure 3.41: Map of final dataset of observation wells with location names and dominant recharge mechanism

Figure 3.42 shows the seasonality of the groundwater level fluctuations for several of the raindominated groundwater wells. Only two of the wells (#15 in Fraser Valley, and #228 in Cassidy) in rain dominated regions are relatively close to a stream, in an alluvial aquifer, and hence potentially connected to rivers. They are distinguished with thicker lines in blue colors in the figure. The remaining wells are completed in bedrock. As already discussed previously there is variability in the seasonality because the response of the aquifers depends on the local aquifer characteristics and their potential connection to surface water. However, the wells on the Gulf Islands and southwest Vancouver Island follow very similar annual cycles with maximum levels in January and minimum levels in September. The fluctuations in Abbotsford (Fraser Valley) lag behind with a maximum level in March and minimum level in October.

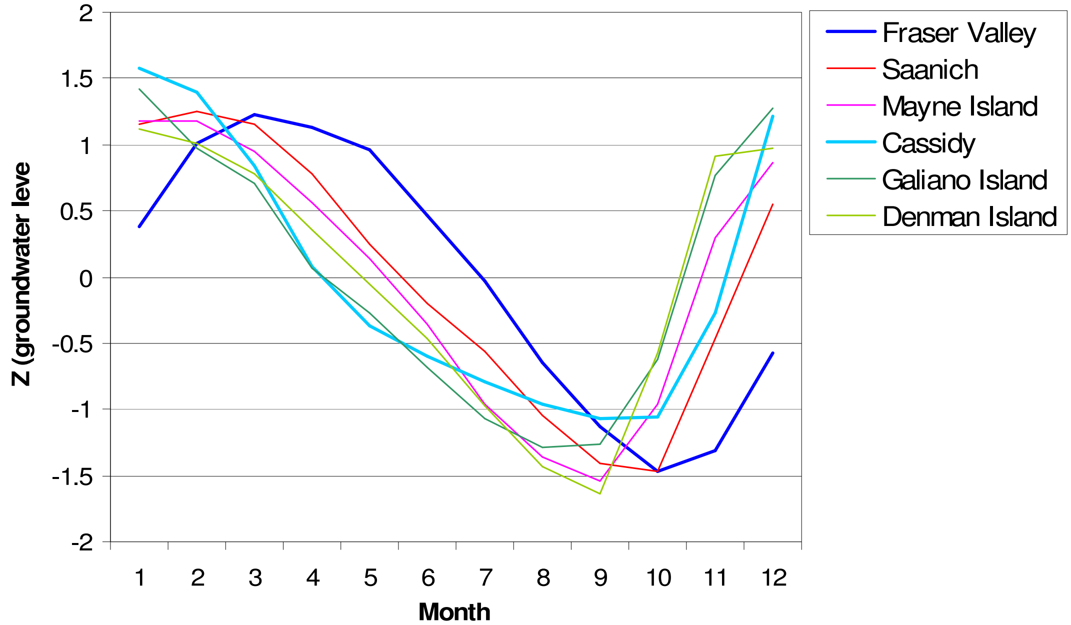


Figure 3.42: Seasonality of selected groundwater levels with rain dominated annual cycle (wells in alluvial aquifers close to streams are marked by bold lines)

Among the snow-dominated systems there is also considerable variability. Recharge depends on the timing of the snowmelt, which depends on elevation, topography, etc. and on aquifer characteristics. The wells in Grand Forks (#217), Salmon River (#185), Merritt (#296) and Golden (#309) are in alluvial gravel aquifers where hydrometric records of nearby rivers exist. With their fast rise, the groundwater level series closely resembles the streamflow hydrographs.

The groundwater level in Merritt drops first after snowmelt among those four sites, possibly reflecting the earlier and faster spring snowmelt in this dry area with comparably low elevation differences, and/or a very direct hydraulic connection to the stream. Salmon River (#185), Carr's Landing (#53/54), and Summerland (#153/154) have very wide summer peaks. The well in Summerland shows the latest rise and fall. Potentially, the three aquifers could hence feed streamflow until well into the fall season. However, they are in alluvial material, close to lakes and, hence, also close the rivers' mouths. It is therefore not clear whether these hydrographs give a general indication of groundwater levels in the area and whether they could provide information on how groundwater feeds late summer streamflow, or whether they may perhaps be influenced by lake levels. Therefore, they may be of limited value to examine groundwater fed baseflow in rivers that flow from the plateau areas into the Okanagan Valley. Trends and seasonality are discussed in more detail in Section 3.7.5.

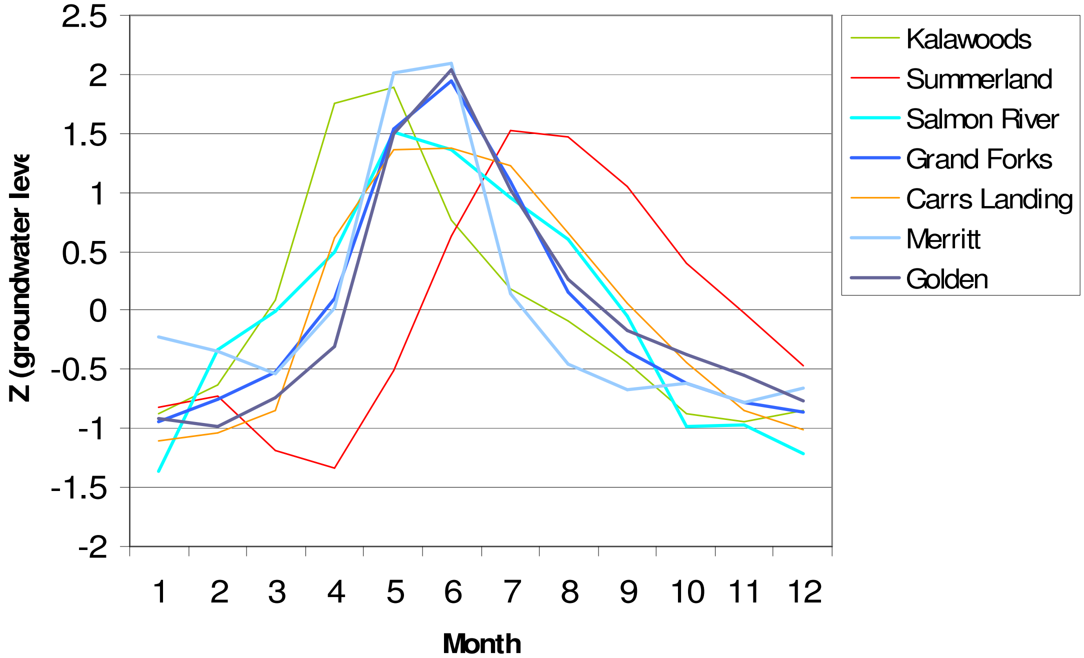


Figure 3.43: Seasonality of selected groundwater levels with snow-melt dominated annual cycle (wells close to gauged streams are marked by bold lines)

Trends

Trends in groundwater well levels were found to be highly variable. However, negative trends, i.e. decreasing groundwater levels, dominate the records (Appendix 1.3). Among the rain dominated systems, one of the few observation wells that shows mainly increasing groundwater levels, particularly in the summer, is Cassidy (#228), close to Nanaimo on Vancouver Island. The seasonality of this well and its trends will be discussed in more detail in the next section.

Negative trends also dominate the snowmelt driven systems. One exception within this subset is the groundwater level at Grand Forks, which has increased, particularly in winter. This trend is also discussed in more detail later. There is no common seasonal pattern in the trends.

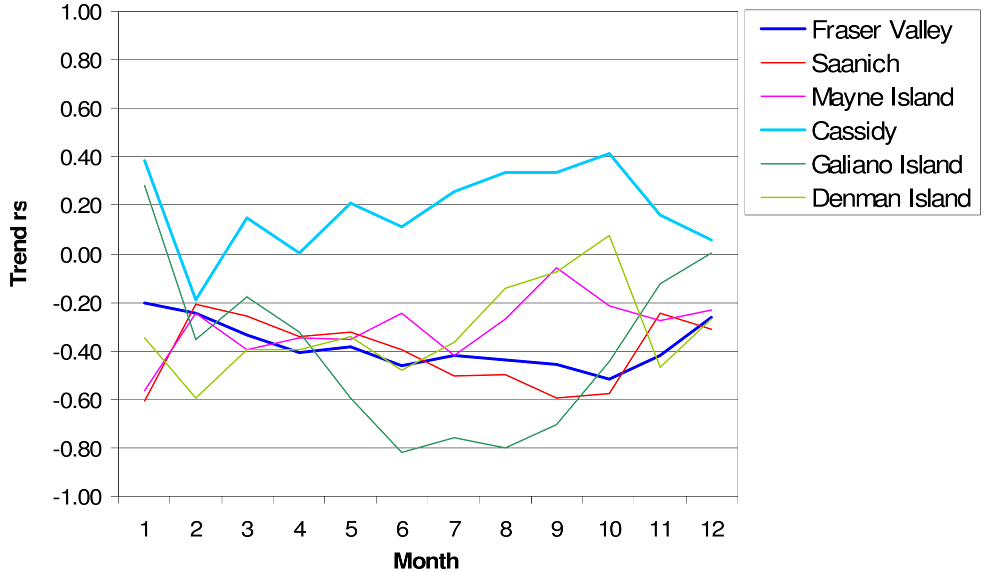


Figure 3.44: Monthly trends in selected groundwater levels with rain dominated seasonality (wells in alluvial aquifers close to streams are marked by bold lines).

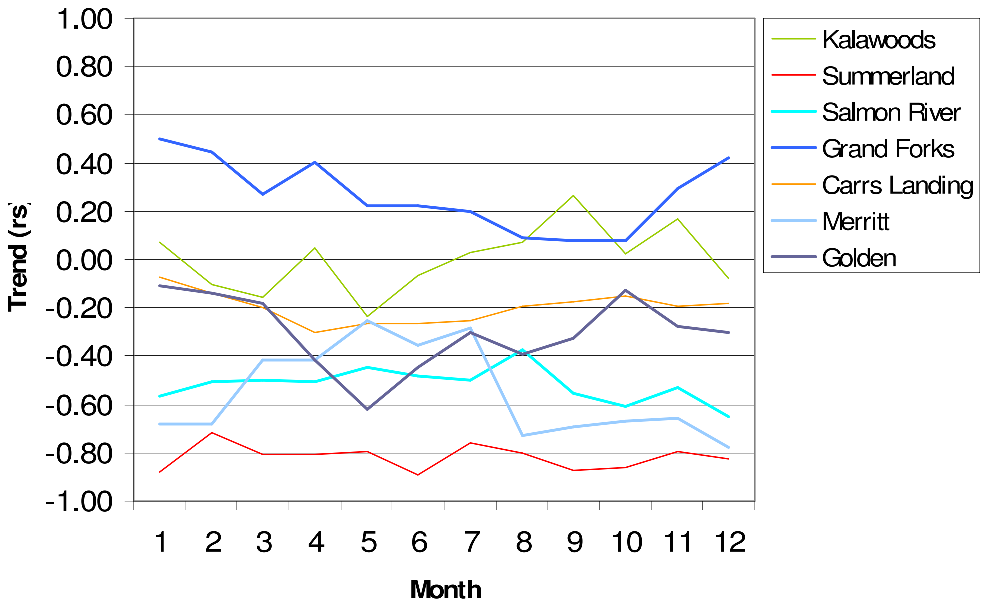


Figure 3.45: Monthly trends in selected groundwater levels with snowmelt dominated seasonality (wells close to gauged streams are marked by bold lines)

The spatial distribution of trends found for the summer months of August and September, which is the time when groundwater most likely feeds streamflow in BC, is mapped in Figure 3.46. The map also shows a predominance of negative trends, but there are exceptions that do not appear to have any spatial pattern. In general, the dominance of negative trends could suggest that the streamflow trends seen during these months (see streamflow trends in Section 2.0) may be linked to groundwater decrease.

Trends in the estimated recharge are more variable (Figure 3.47). Positive trends indicate an increase in the difference between the lowest and the following highest groundwater level, negative trends indicate a decrease. There are more positive trends which, however, are also not spatially consistent.

Hence a lack of recharge does not seem to be the predominant reason at the provincial scale for the late summer groundwater level decrease. Figure 3.48 shows the results of a similar trend analysis applied to recharge-season precipitation as well to April 1st snow water equivalent (SWE) records at provincial snow course sites (http://www.env.gov.bc.ca/rfc/river_forecast/manusurv.htm). Precipitation amounts from October to March have increased. April 1st SWE has increased at some sites and decreased at other sites with only a few significant increasing trends. This hypothesis is also supported by positive trends in recharge-season precipitation.

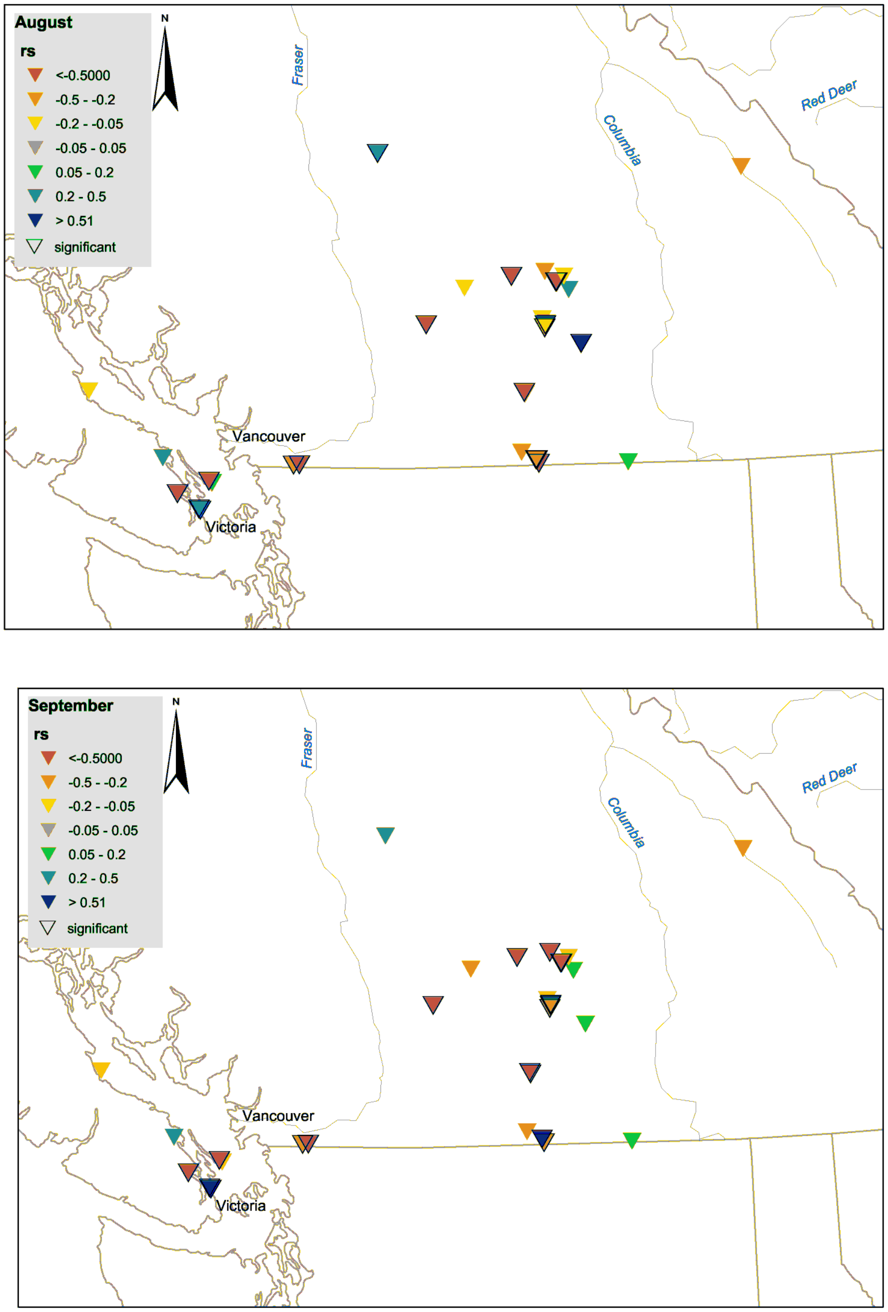


Figure 3.46: Trends in groundwater well levels in August (upper panel) and September (lower panel)

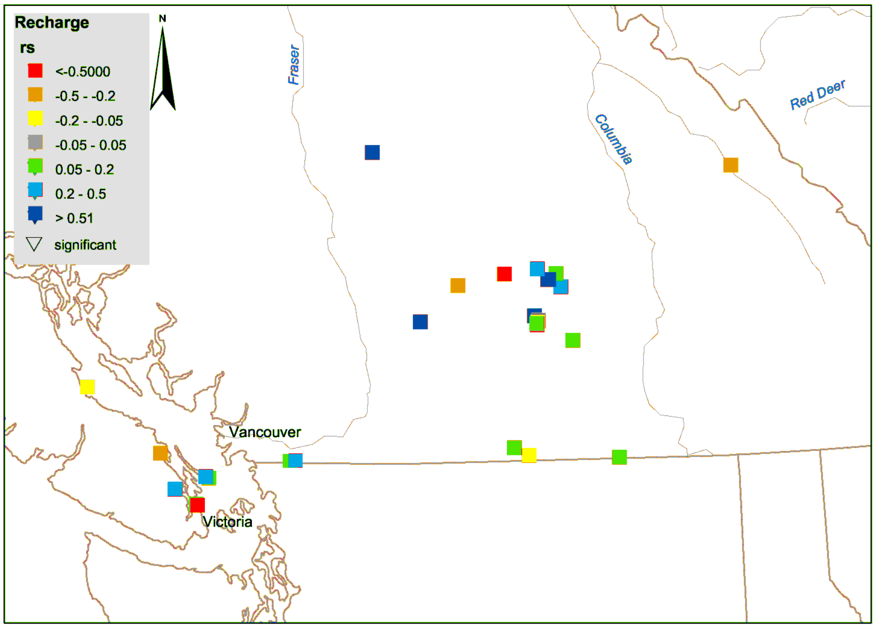


Figure 3.47: Trends in groundwater recharge estimated from annual minimum and maximum groundwater levels.

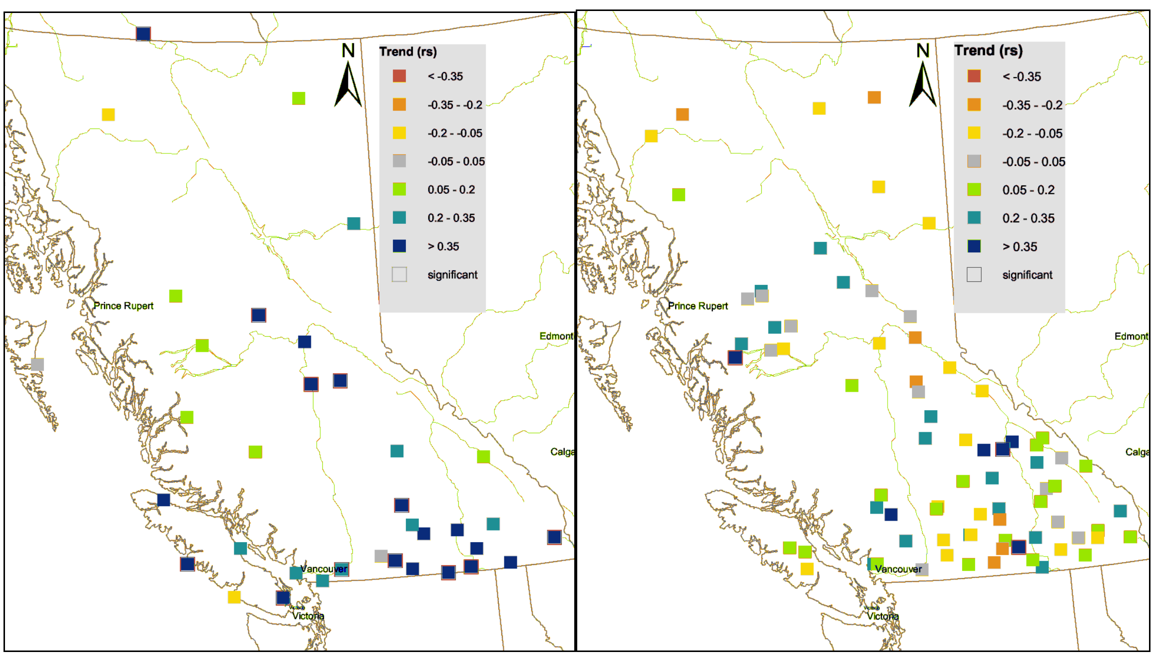


Figure 3.48: Trends in October to April precipitation at selected climate stations (left panel) and April 1st snow water equivalent (SWE) at provincial snow course sites (right panel), both for 1976-2002.

3.7.4 Regional seasonality and trends (rain dominated)

Lower mainland/Fraser valley

There are three provincial observation wells in the Lower Fraser Valley with long records: #002, #014, #015 (Figure 3.49). The wells are completed in an unconfined sand and gravel aquifer that is classified by the province as classified as a Type 4a aquifer. The deeper wells #014 and #015 (96 m and 97 m depth, respectively) are situated at the edge of the aquifer (discharge area). This aquifer is also classified by the province as being 'highly productive' and 'highly vulnerable'.

The most visible change in water level in these wells occurred in 1976 when groundwater levels dropped to the current level of between 10 and 15 m below surface (see Figure 3.49). The drop is also visible in well #002; however, it was not nearly as pronounced, and the water level recovered by 1982. The time of the drop coincides with the shift of the PDO phase and several strong El Nino years. Thus, the trend analysis may be influenced by the PDO shift. Well #015 was not included in the trend analysis as it is no longer active.

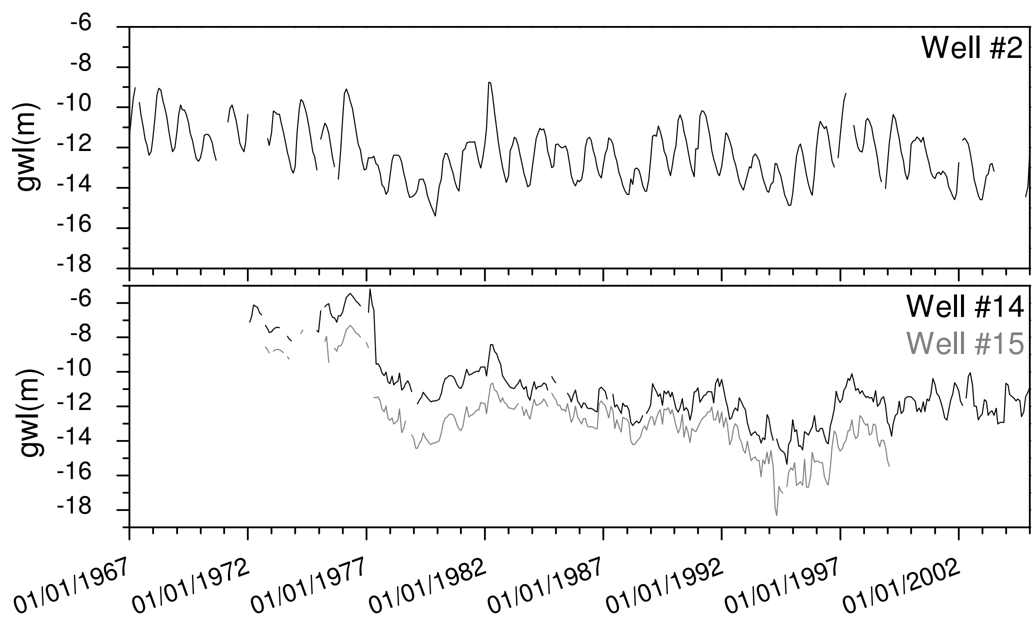


Figure 3.49: Groundwater levels in Abbotsford (Fraser Valley)

Figure 3.50 shows the standardized seasonality and the results for the trend analysis on this relatively short common time series from 1984 to 2003. The closest climate station is the Abbotsford International

Airport, and the closest river though with a short streamflow record (only since 1984) is Fishtrap Creek. The creek's main low flow period is during the dry summer season (July to September), slightly lagging behind the annual cycle of precipitation.

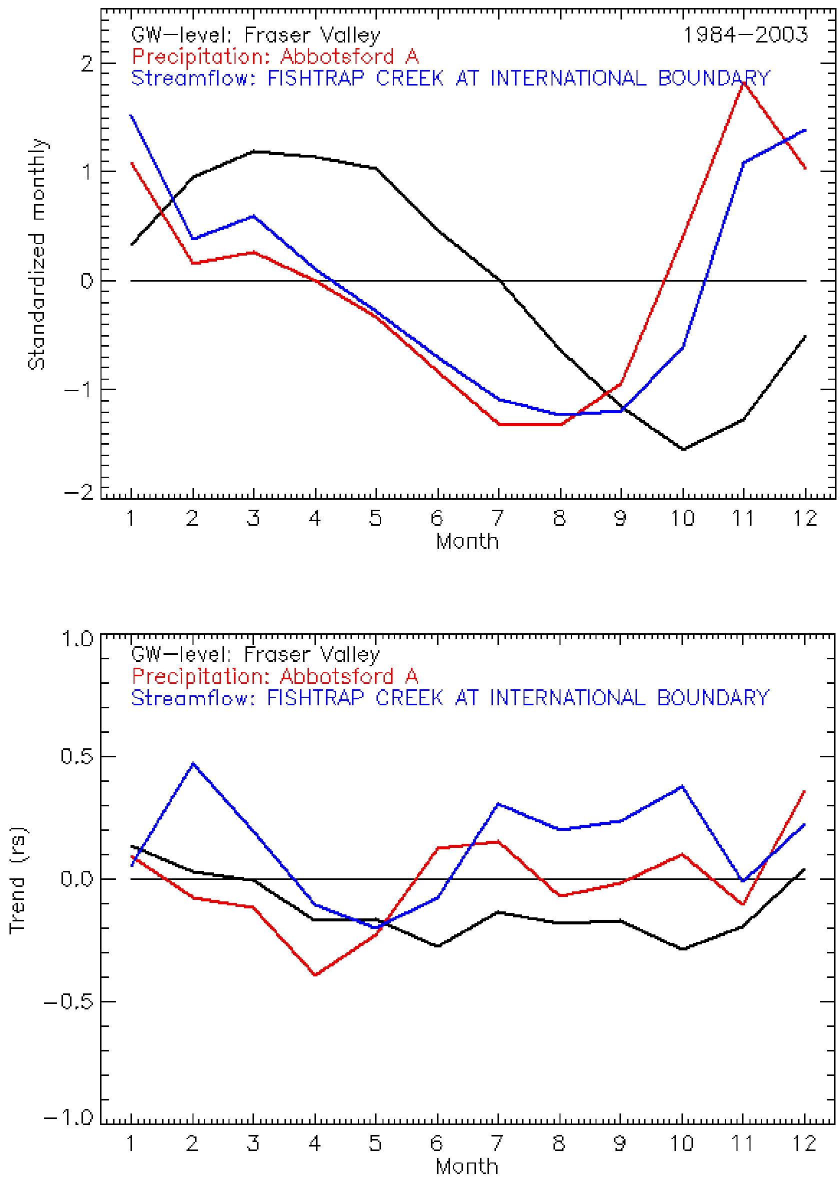


Figure 3.50: Standardized monthly groundwater level (Obs Well #002), precipitation and streamflow (top panel) and monthly trend statistic (Spearman's rho, bottom panel).

Recharge to the groundwater coincides with variation of precipitation from January to August (see also Scibek and Allen. 2005). Then recharge begins to diverge, gradually lagging up to 2 months behind precipitation. Greater evapotranspiration and a deeper water table may explain this shift. The groundwater level in well #002 hence follows the precipitation cycle with a lag of ca. 3 months.

Appendix 2 gives an overview of a field study that was conducted in the Abbotsford Aquifer to investigate the interaction of streams and the aquifer. Stream discharge and water chemistry were measured throughout the summer season to identify summer trends in the contribution of groundwater to streamflow. The research constitutes a portion of a paper by Berg and Allen (in press). Fishtrap Creek, Bertrand Creek and Pepin Brook are found variably fed by groundwater during their summer low flow season largely on account of the variable distribution of surficial sediments comprising the aquifer. Of the three streams, Fishtrap was most strongly connected to the aquifer.

With the exception of increasing February streamflow, no significant trends were found in any of the variables. However, significances require large persistent trends when testing such a short record. Therefore, it is still worth comparing the relative trends found in the three variables.

The streamflow of Fishtrap Creek shows a slightly negative trend in spring following that of precipitation. Streamflow then shows a larger positive trend during the low flow season from July to October. There is only a slight increase in June and July precipitation during the same period, which may not be enough to explain this increase in low flow by direct rainfall runoff processes. The negative trends in the groundwater levels throughout the entire summer from April to November rule out increased groundwater discharge. Hence, the explanation of this increase in summer low flow is not straightforward, and may be influenced by other factors.

To assess the situation, we should consider results from a longer time period and results from surrounding wells and gauges. For the common period 1976-2002 most other rivers in the area show no trend or a non-significant small decreasing streamflow trend in August, and a decreasing trend in September (see Section 2.0). However, there is no record in the dataset from a comparable low elevation river. August precipitation has generally decreased in the region. Groundwater levels during the summer months for the longer period were also found to be mostly negative. Only annual recharge trends and winter precipitation trends are mostly positive.

One possible explanation is that due to the decrease in summer precipitation and increase in summer temperatures, water demand for irrigation and consequently pumping rates on the entire aquifer have increased and have lowered the summer water tables. Increased irrigation may have resulted in increased return flows into Fishtrap Creek augmenting natural summer low flow. However, without any additional information on groundwater abstraction and return flow from irrigation, it is difficult to determine the true reason for the small trends seen.

Southeast Vancouver Island

Two wells (#211 and #228) with fairly long records are located north of Victoria in unconsolidated sand and gravel aquifers (Type 4b), which are classified as 'high productivity' and 'low vulnerability'. The wells are situated close to rivers and, therefore, are possibly influenced by groundwater-surface water interaction. Figure 3.51 shows the time series of the groundwater levels in both wells. The strong trend in Duncan and the various gaps in the record may suggest human influences, therefore, this well is not discussed further.

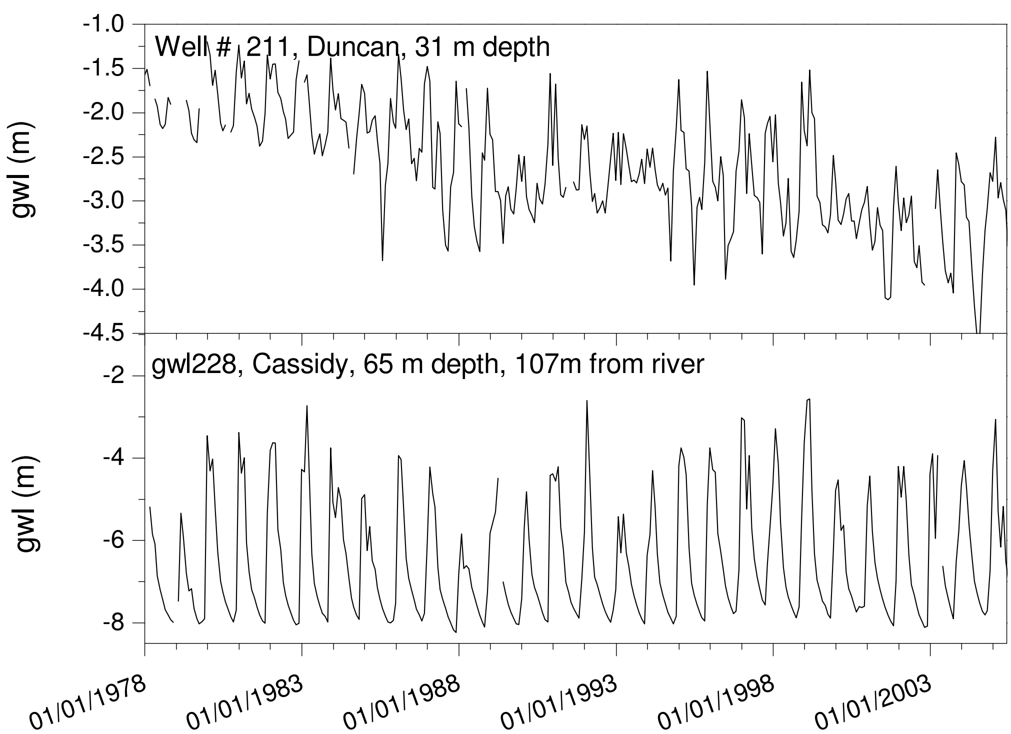


Figure 3.51: Groundwater levels in southeast Vancouver Island

The groundwater well ‘Cassidy’ is close to Haslam Creek (distance of 107 m). Unfortunately, the creek was gauged for only for a few years during the summer and hence cannot be used for comparison. Nanaimo River, which is also plotted in Figure 3.52 is in the neighbouring basin. It is larger, but generally covers similar terrain as Haslam basin. However, it can only be used as a climate-surrogate, not to describe any functional relation between river and aquifer.

Nanaimo Airport climate station is nearby and shows that streamflow and the groundwater level lag the precipitation cycle by about one month. Overall, precipitation, streamflow and groundwater follow the typical rain-dominated seasonal distribution with highest levels between December and April and lowest levels in late summer.

For the longest common period from 1978 to 2002 significant negative precipitation trends in Nanaimo were found only in September (Figure 3.52). There is also a considerable, but not statistically significant, positive precipitation trend in December, January, and July. Groundwater level trends are positive in summer/fall.

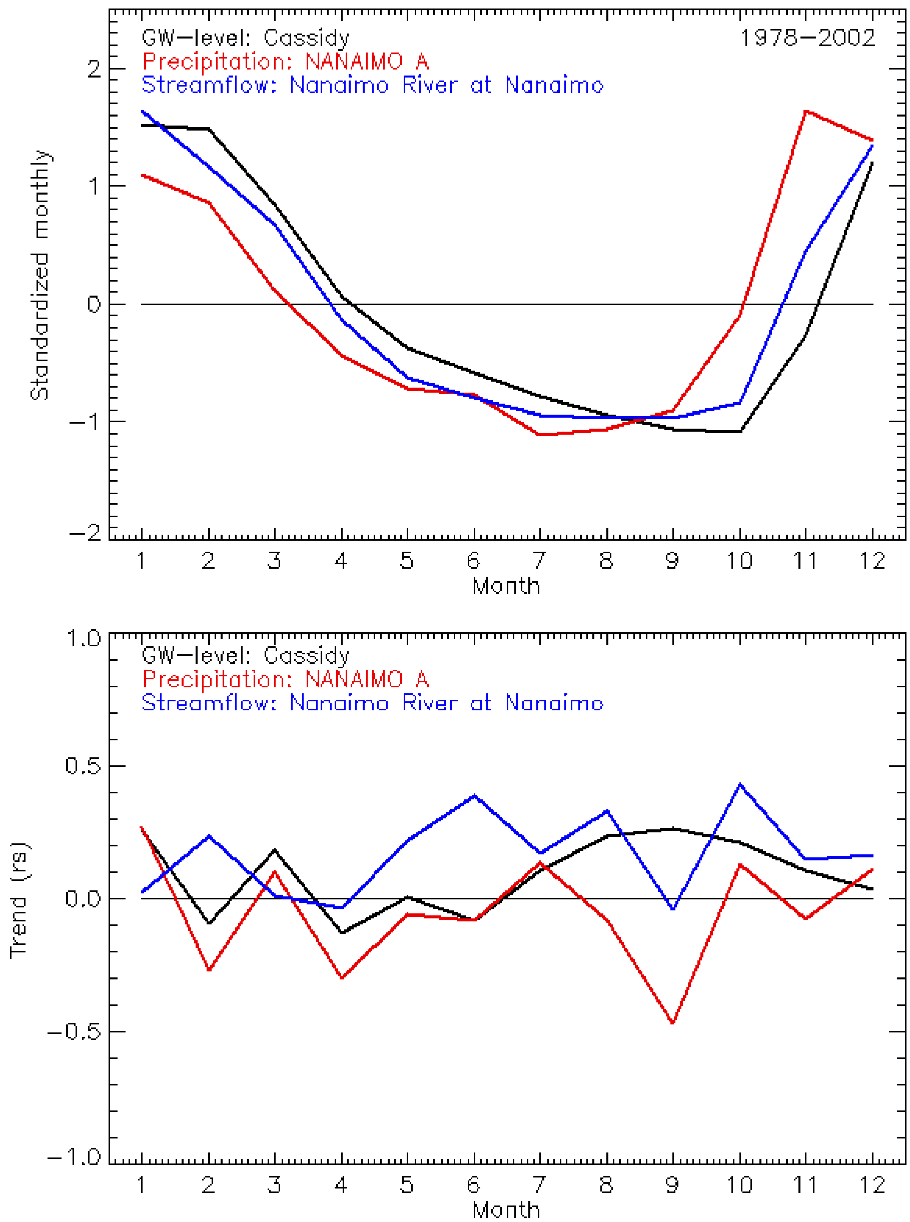


Figure 3.52: Seasonality and trends at groundwater well Cassidy, precipitation in Nanaimo and streamflow of Nanaimo River.

Where rivers are fed by groundwater during the summer they may therefore also show positive trends. Nanaimo River and other rivers to the north do show positive trends in July and August, but negative trends in September. Attribution of the trends is again difficult. Increased recharge during the winter as well as increased early summer precipitation may have played a role. However, the former is not visible in the groundwater trends, and the latter is not a clear precipitation trend at the nearest climate station. Most rivers to the south show opposite (negative) trends in August streamflow. In September streamflow trends throughout the region are negative (despite positive groundwater trends at the well in Cassidy). Thus, the hydraulic connection to many streams may be limited by September.

Another complicating factor is that although the groundwater seasonality (and streamflow at Nanaimo River) is distinctly pluvial, the headwaters of Haslam Creek and Nanaimo River likely get some snow accumulation during the winter.

3.7.5. Regional seasonality and trends (snowmelt dominated)

Merritt Salmon Arm – North Okanagan

There are several observation wells between Merritt and the Northern Okanagan. Well #296 in Merritt is in a sand and gravel (likely the alluvium of the Nicola River) aquifer classified as Type 1a, and is 'highly productive' and 'highly vulnerable'. Well #185 is close to Salmon River (near Salmon Arm) in a sand and gravel aquifer also classified as a Type 1a, but is considered 'moderate-to-high productivity and vulnerability' based on development. The strong seasonal fluctuations indicate that both wells may be influenced by the river's snowmelt peaks (Figure 3.53).

Both wells show a marked downward trend in water levels after ca. 1997. In the broader area there are also several wells in Armstrong which show similar long term fluctuations with a recent water level decrease. One of them, the much deeper well #118, also plotted in the figure, is in a sand and silt aquifer (Type 4a or 4b).

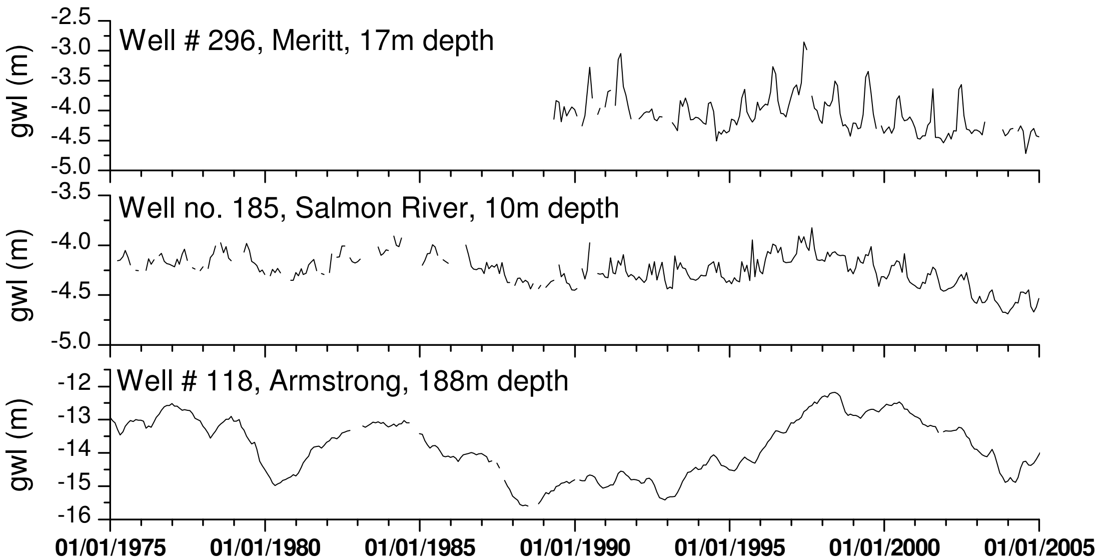


Figure 3.53: Well records from Merritt to the Northern Okanagan

There are streamflow records for the Nicola River upstream of Well #296 near Merritt and for the Salmon River in Salmon Arm. Unfortunately, the record of the observation well in Merritt is very short, resulting in a common period with climate and streamflow only from 1989 to 2002, and the well record at Salmon River has many missing values in the earlier years. A hydraulic connection between groundwater and streamflow is likely at both sites given the proximity of the wells to the river, and that the seasonality of the groundwater level seems to follow the snowmelt peak closer than the precipitation seasonality (Figure 3.54 and 3.55).

In Merritt, precipitation and groundwater show negative trends in August. Groundwater trends remain negative throughout the fall and until January, and streamflow seems to follow the same pattern from September to December. These streamflow trends (not significant) are therefore likely due to the negative precipitation trends in the same season propagating through the hydrological cycle. Possibly lower snowpack and, hence, lower recharge adds a cumulative deficit seen in the groundwater trends in this very dry climate.

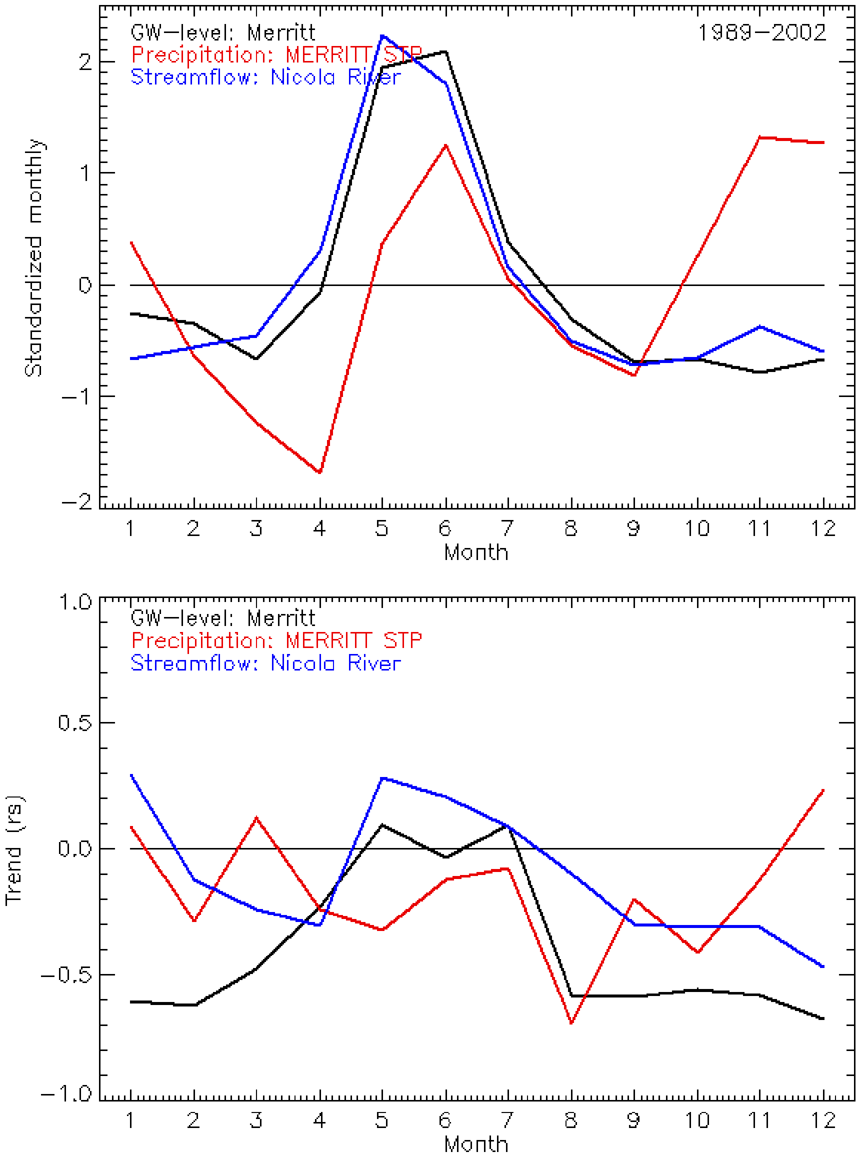


Figure 3.54: Seasonality and trends in Merritt.

The groundwater level seasonality in the well at Salmon River is slightly different. It follows streamflow of Salmon River with a lag, suggesting a connection, and recharge through snowmelt during the spring. However, while streamflow decreases rapidly in the early summer, groundwater levels remain high throughout the summer. It should be noted that the streamflow record is marked as ‘regulated’. The degree of the regulation, however, may be small.

The hydrometric record from Salmon River is relatively short (since 1980). During those 20 years, however, groundwater levels have dropped year round except in August. Streamflow has declined during the late summer low flow season, particularly in September and October). The slight decrease in summer precipitation is not significant and may not be enough to explain the decreases in groundwater level and, subsequently, in streamflow.

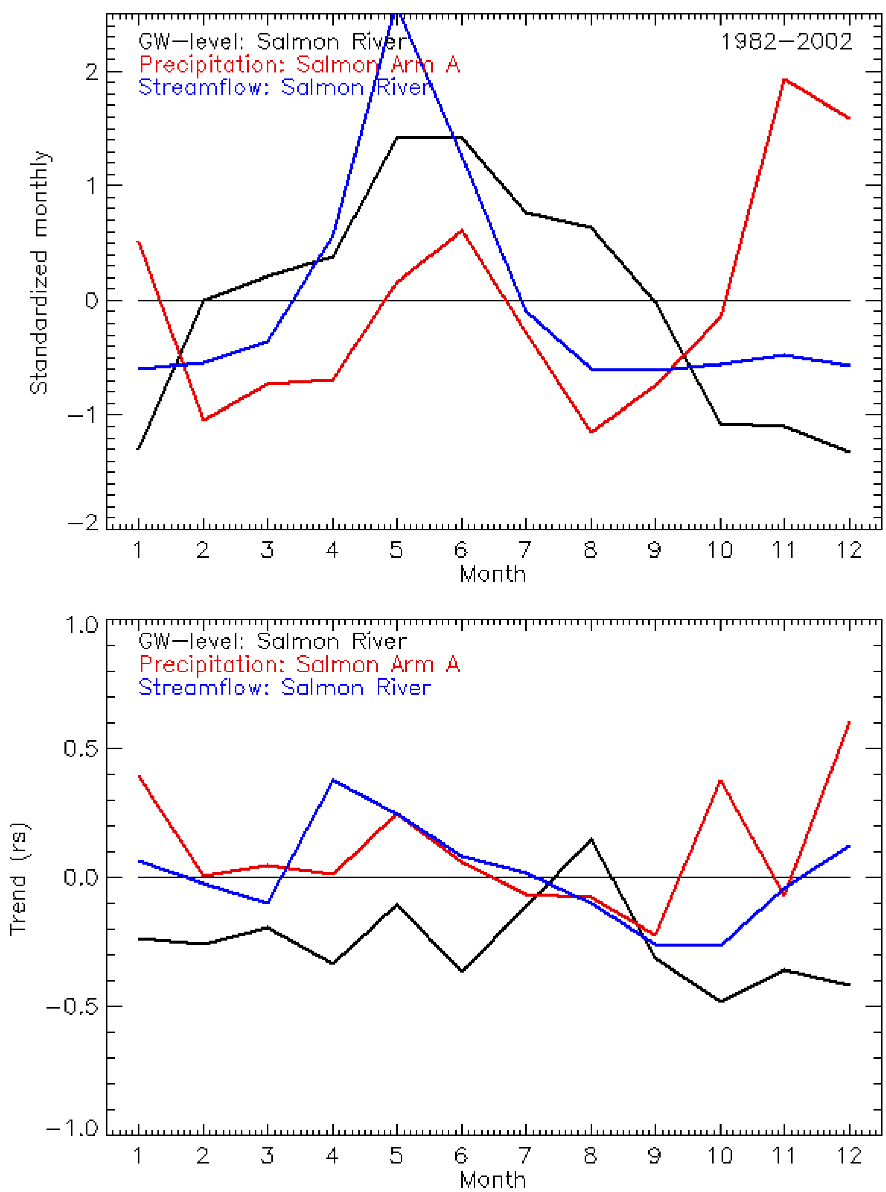


Figure 3.55: Seasonality and trends at Salmon River near Salmon Arm

Southern and southeastern BC

There are two observation wells in southern BC that are in sand and gravel aquifers and likely connected to gauged rivers: Grand Forks (#217) and Golden (#309). The well at Grand Forks is situated in a Type 1a aquifer, which is 'highly productive' and 'highly vulnerable'. The well at Golden is situated in a Type 4b aquifer. Both show strong within year fluctuations and the same recent (since 1997) negative trends in overall groundwater levels as the wells discussed in the previous section.

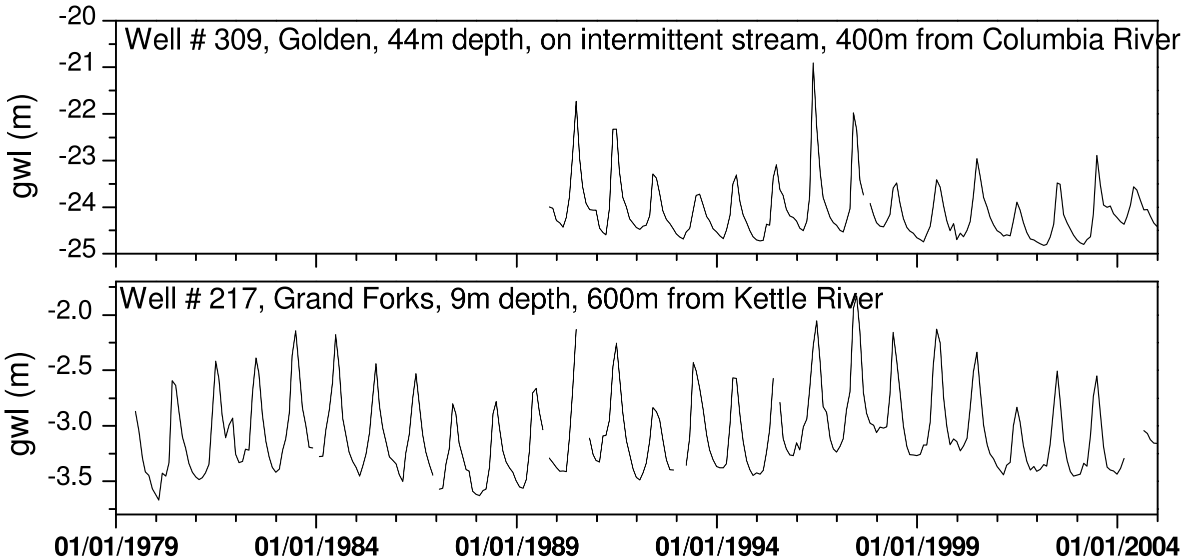


Figure 3.56: Well records for Golden and Grand Forks.

This particular aquifer at Grand Forks was the subject of a comprehensive research project funded by the Climate Change Action Fund by Allen et al. (2004). In that study, future climate change predictions were used to quantify the shifts in hydrologic regime of the Kettle River and of the aquifer. A strong connection between the aquifer and the river was demonstrated through numerical modeling. In the early part of the year, following spring freshet, the river provides most of the recharge to the aquifer. Towards the summer months, the direction of flow reverses, and the groundwater discharges into the stream (low flow period). In the fall, local precipitation recharges the aquifer and this additional water continues to contribute to baseflow, although at a much reduced amount. It is within this context that the results of the trend analysis are discussed.

Figure 3.57 illustrates the seasonality of precipitation, streamflow and groundwater in Grand Forks. Peak flows in the Kettle River occur in May, followed by a groundwater peak in June due to the recharge mechanism described by Allen et al. (2004). There is also a precipitation maximum in May/June. Streamflow then drops fast and the low flow season, during which groundwater flow is reversed to the stream, starts in August and continues to March.

Contrary to the aforementioned groundwater level decrease recently (after 1997), the groundwater level in Grand Forks shows mostly positive trends in the longer time period (since 1977) used in the trend analysis. However the monthly trends are not significant except for January. The groundwater level record still stands out among those discussed because the trends are positive year round.

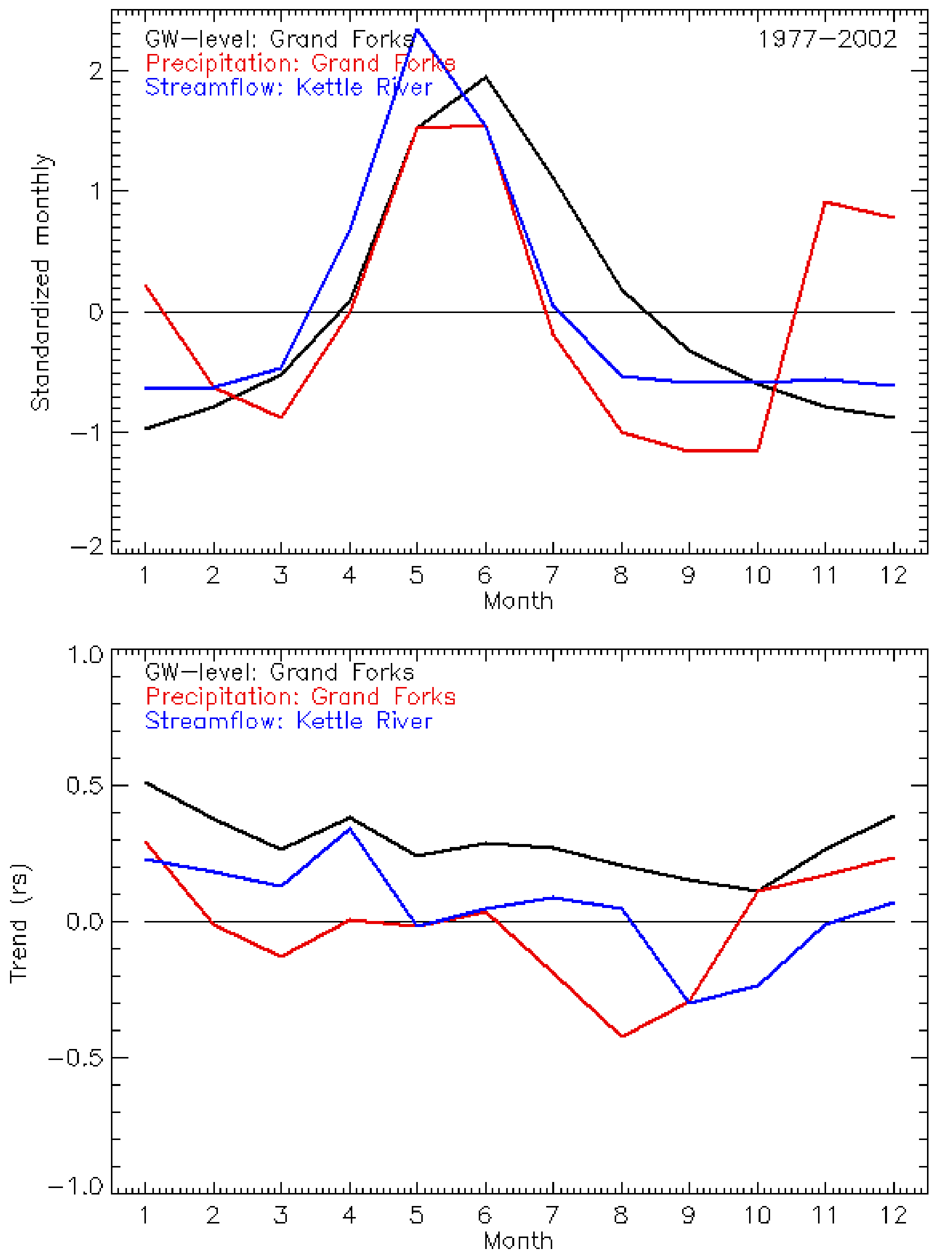


Figure 3.57: Seasonality and trends in Grand Forks

Winter streamflow shows a positive trend due to less freezing/snow storage at higher elevations in the watershed. A shift in peak flow to earlier in the year has been shown for many streams in southeast BC (Whitfield and Cannon, 2000). Thus, due to the earlier freshet, the levels appear to be higher in any given month compared to their historic levels in that month. Due to the close connection of this river with the aquifer, the higher streamflows give rise to a corresponding increase groundwater levels. This result is consistent with the predicted shifts in the groundwater levels in any particular month due to an earlier freshet (Scibek et al., 2006).

September and October streamflows show negative trends, which again are consistent with previous findings (Whitfield and Cannon, 2000). Yet, groundwater levels have slightly positive trends. During the early fall, groundwater is recharged primarily by precipitation, thus, an increase in precipitation during the fall will lead to a corrsponding rise in groundwater level, and perhaps not as significant a change in streamflow. The results are generally consistent with future predicted impacts on groundwater recharge (Scibek and Allen, 2006) under a wetter and warmer future climate scenario.

The seasonalities in the Rocky Mountain Trench are discussed using the Columbia River, the groundwater well at Golden and the climate station at Golden. All three have similar seasonality. The Columbia River has low flows from September to April.

Trends in groundwater levels and streamflow, however, are quite different. At the well in Golden, long term trends are negative, however they are not significant. One reason could be the much shorter time series analyzed, as the groundwater record at Golden is only exists since 1989. For this short common time period, streamflow has increased slightly during the summer. Since groundwater has decreased, this may be due to increased glacier melt in the headwaters of the Columbia River. Groundwater in the area is likely recharged by snowmelt (lower elevation snowmelt causing a groundwater level rise ahead of the river's freshet) and groundwater levels year round have dropped. Possibly snowpack was low in recent years (which would explain negative trends in spring streamflow in the river). The area around Golden also has seen substantial development, so anthropogenic influences cannot be ruled out.

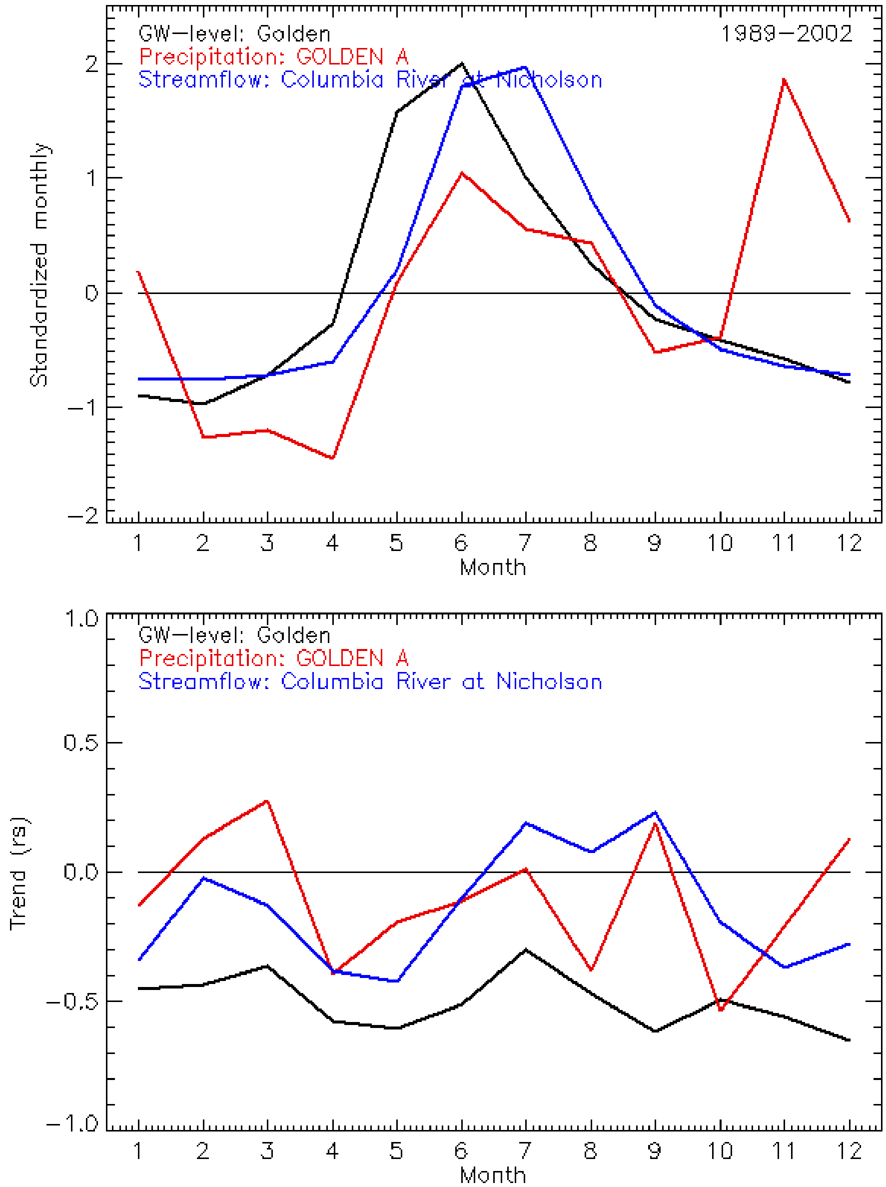


Figure 3.58: Seasonality and trends at Golden.

3.7.6 Regional groundwater trends for wells in bedrock or where no streamflow data are available

Gulf Islands

There are no gauged rivers on the Gulf Islands. Most observation wells are completed in bedrock and, hence, are only influenced by precipitation recharge. The results are therefore mainly of interest as a baseline case of groundwater behavior itself. In general, water tables on the Gulf Islands are shallow with fluctuations following precipitation with a lag of up to a month. All show strong annual cycles with higher levels during the wet winter season and lower levels during the dry summer season.

Wells # 126, #127 and #128 on Mayne Island are close to the coast and drilled into bedrock aquifers classified as Type 5a, and having 'low productivity' and 'high vulnerability'. Well 126 shows positive piezometric levels, suggesting that the piezometric level is above ground surface, and that this well is completed into a highly pressurized confined aquifer. Often, wells situated near the coast on the Gulf Islands are found to be flowing artesian. Groundwater levels near the coast, in general, are shallower than levels inland.

To the north, on Denman Island, well # 168 (312m) also shows positive levels, however, only in winter. In summer the water table drops 1 to 2 m below the surface. Well # 258 (90 m) on Galiano Island has lower water levels and larger absolute fluctuations, but similar timing as the others.

Both observation wells on Mayne Island show negative trends year-round. They are stronger in the first half of the year, but not significant. The same applies to the well on Denman Island. The well on Galiano Island has significant negative trends from June to Sept.

Long-term analyses report negative precipitation trends for the winter recharge season in the area (BC Ministry of Environment, 2006), however, shorter series as the period used in this study indicate more variable monthly precipitation trends with mainly increasing precipitation for the winter half year and a more common signal of negative precipitation trends in the summer (see Section 2.0). A combination could be the cause for the negative trends in water level in addition to human influences to the aquifer through nearby drinking water wells.

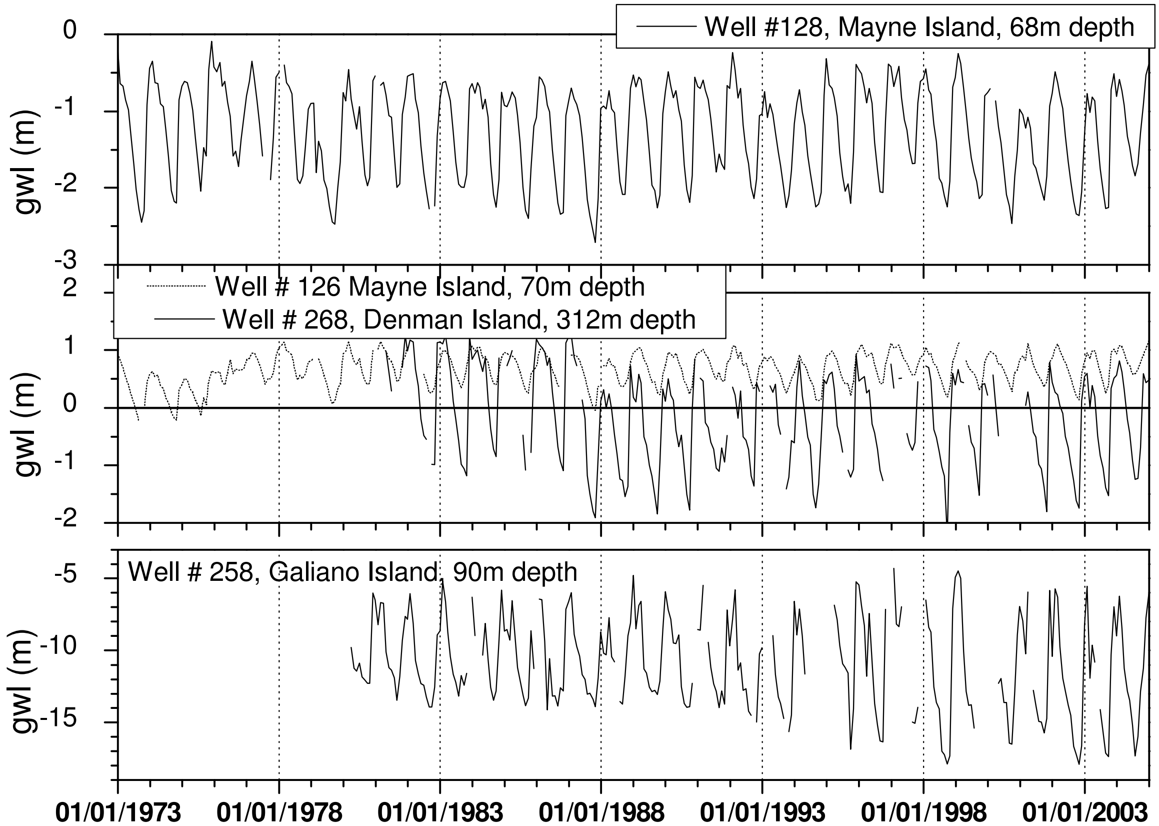


Figure 3.59: Groundwater levels in the Gulf Islands.

Vancouver Island

On Vancouver Island's Saanich Peninsula two wells have long records: #65 and #60. The former is drilled deep into a fractured aquifer (granodiorite); the latter well is shallower in till bedrock where groundwater levels are lower. Groundwater levels in both wells follow the annual precipitation cycle, with stronger fluctuations in well #65 but a stronger multiyear memory in #60 (Figure 3.60). The long-term variability with a drop after 1976, the recovery by 1982 and another low in 1995 is very similar to the well in Abbotsford in the Fraser Valley.

Trends in the two wells records contradict each other. Over the period of record, well #60 shows negative trends year-round, and they are significant in September/October. Well #65 shows the opposite: positive trends year-round, which are significant in June to October. It should be kept in mind that this might be an effect of the different length of their time series.

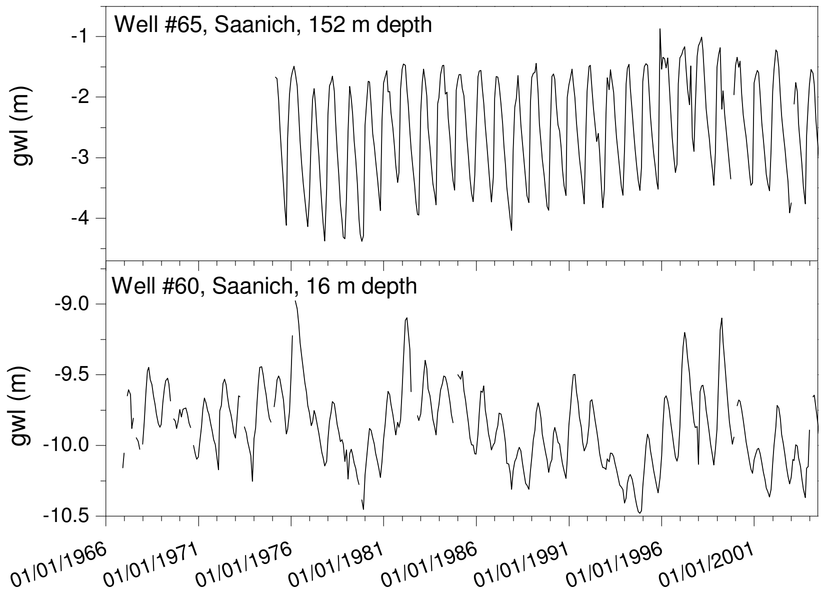


Figure 3.60: Groundwater levels on the Saanich peninsula

Various high-elevation groundwater observation wells

There are a few observation wells on ridges and at high elevations (above 1000 m). They all show relatively similar seasonal cycles, but no obvious common long-term trends (Figure 3.61). They are all recharged by snowmelt with rising water tables from March to May. The water level in observation wells #81 at '83 Mile' (the deepest well, drilled into basalt) and #54 'Carr's Landing' respond in a delayed fashion and drop only slowly over the summer and fall, whereas Silver Star (#47) and Mission Creek (#115) have larger absolute water level differences seasonally and drop quickly in summer, reacting almost like a river hydrograph. They are both in the vicinity of headwater streams, but the wells are in (fractured) bedrock. Mission Creek is gauged (far) downstream and hence a comparative seasonality and trend graph is shown in Figure x.

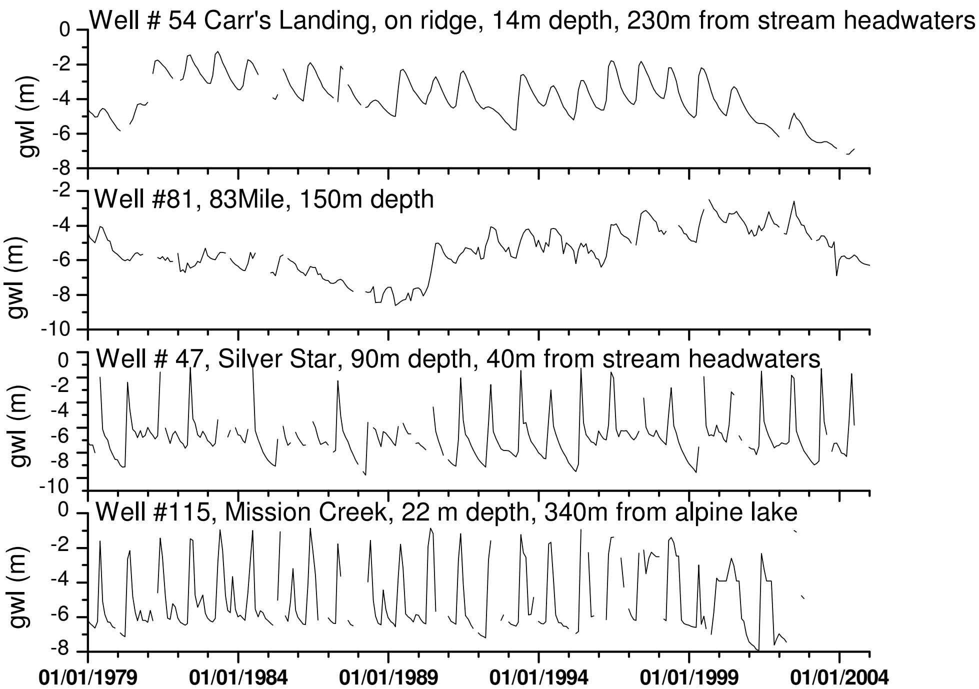


Figure 3.61: Groundwater levels at high elevations in interior BC

Trends for Carr's Landing are negative, likely due to a water table drop after ca. 1999 (however there are some missing values in that stretch as well). 83 Mile has positive trends, significant only in February. The well shows a shift in 1989 that could be anthropogenic.

Figure 3.62 shows that the well at Mission Creek has the same seasonality as the creek's streamflow further downstream near Kelowna. Streamflow and well records show positive trends (marginally significant) in April. As that is the beginning of the snowmelt season, these trends may be a result of streamflow as well as groundwater response to earlier timing of snowmelt. The water table also shows a significant positive trend in August, which may be a direct result of increased early summer precipitation visible in the climate data from Joe Rich Creek and Kelowna. Note that Mission Creek (the Creek) is classified as 'regulated.'

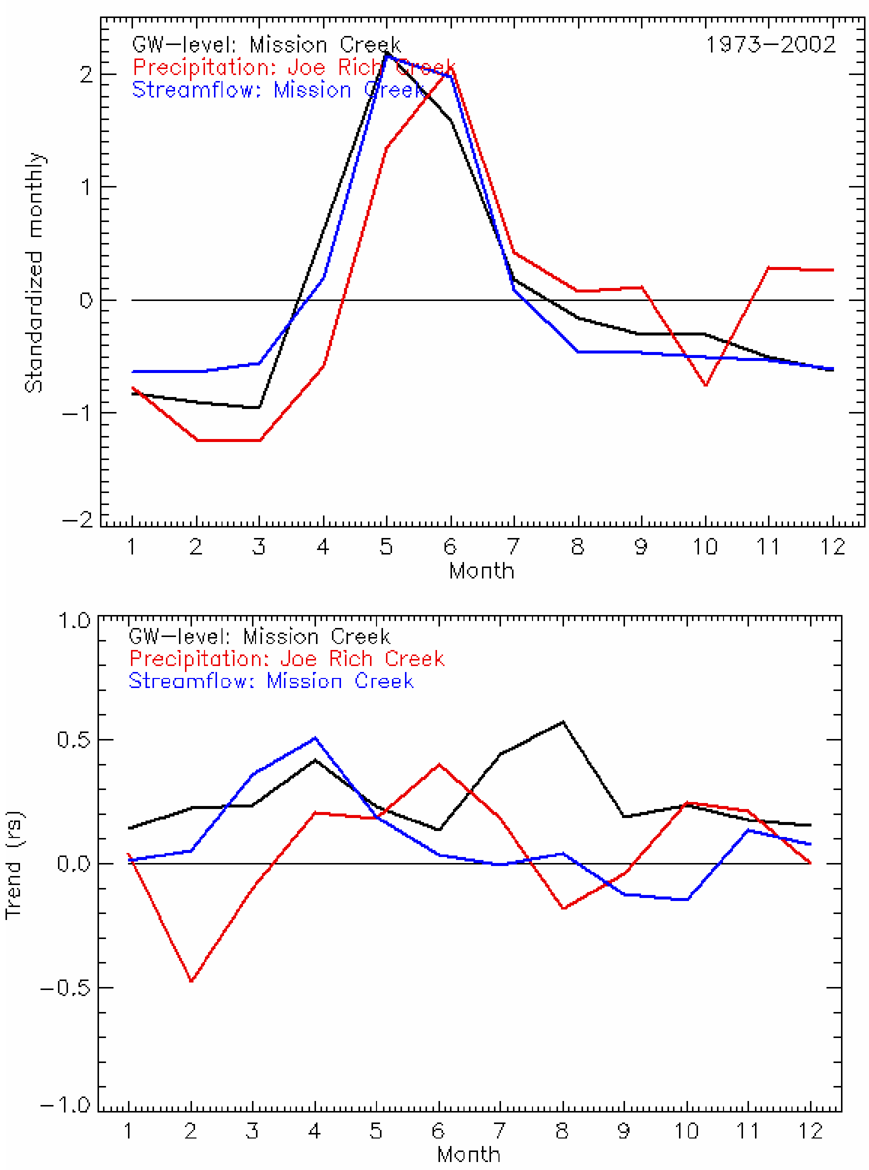


Figure 3.62: Seasonality and trends at Mission Creek well and stream gauge, and precipitation at neighbouring Joe Rich Creek.

South Okanagan

There are several observation wells in Osoyoos and two in Summerland. Most records have many gaps. The best are shown in Figures 3.63 and 3.64. The wells record the lowest groundwater levels from

January to April and the highest from July to September. Due to their proximity to Osoyoos Lake and Okanagan Lake, respectively, it is possible that they are influenced by variations in lake level, which are regulated.

The two wells in Osoyoos are drilled into a sand and gravel aquifer classified as a Type 4a or 4b, which is of ‘moderate productivity’ and ‘high vulnerability’. They are located close to Osoyoos Lake. Well #100 has negative trends in all months but only significant in March and August. Well #96 has negative trends from June to December (significant in June and August), but positive trends in the first five months of the year (not significant).

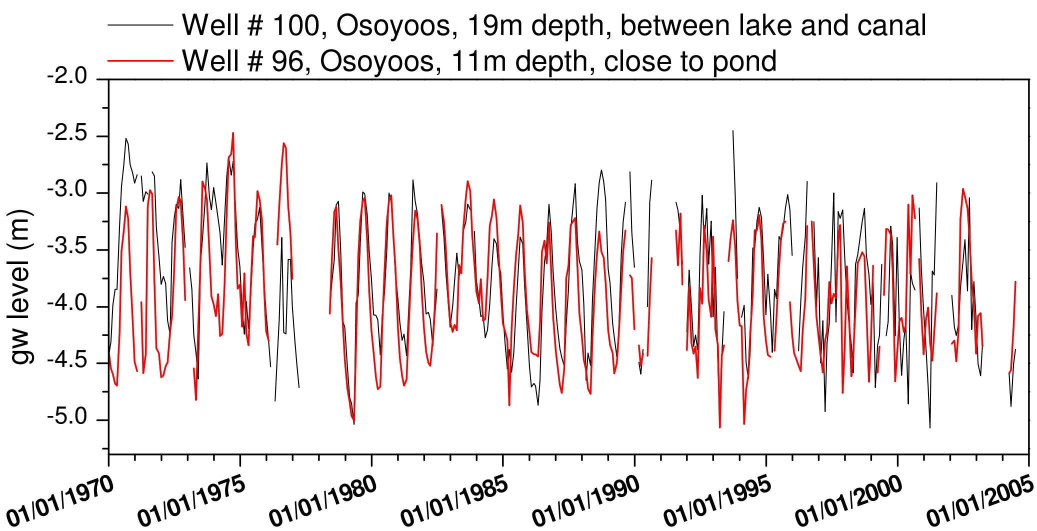


Figure 3.63: Well records in Osoyoos.

The well in Summerland is located on an alluvial fan that borders Okanagan Lake. The sand and gravel aquifer is classified as a Type 2 and is of ‘low productivity’ and ‘moderate vulnerability’. The well record reveals a decreasing groundwater table since the mid-1970s (significant in all months) (Figure 3.66), which is likely due to human influences or possibly lake regulation.

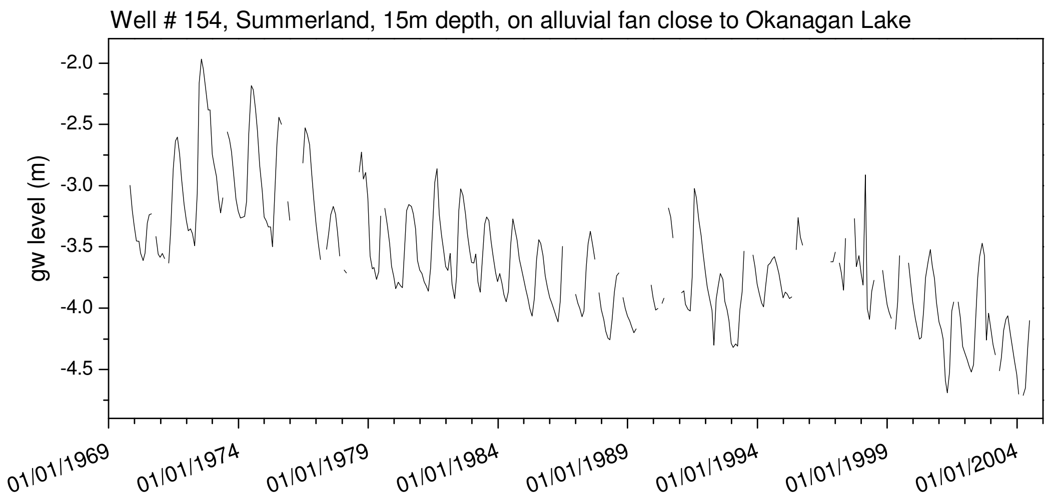


Figure 3.64: Well records in the south Okanagan

3.7.7 Conclusion

This section explored trends in groundwater levels and whether they can be used to attribute trends in summer low flows.

Overall, summer groundwater levels seem to have lowered across the province, despite an increase in winter precipitation and recharge during the same time period. Due to the limited availability of long well records near gauged streams, the attribution of whether and how these changes have affected low flows proves difficult. The available groundwater observation wells are quite different in terms of aquifer properties and the hydraulic connection to rivers and for many, we suspect the natural records may be altered by changes in the abstraction patterns. The few examples where streamflow and groundwater observations are available in the same basin and nearby, provide some of the exceptions to the trends found. In Abbotsford, even though a field study and a modeling study showed that summer low flow is fed by groundwater, low flows increased while over the same time period groundwater levels decreased. The same was found for Grand Forks, where previous studies have shown that streamflow is fed by groundwater in the summer. However, the recharge mechanisms are different in both examples.

4. Climate change, glaciers and streamflow

4.1 Introduction

This component of the study involved the development and application of models for simulating the potential effects of future climate scenarios on streamflow in glacier-fed catchments. This task is challenging because glacier response, especially for smaller glaciers, may occur at a similar rate to the rates at which the climate changes. Many previous studies have applied precipitation-runoff models to estimate the effect of climate change on streamflow in glacier-fed basins, but did not adjust the glacier cover to account for glacier response to the imposed climatic changes (e.g., Moore, 1992; Singh and Bengtsson, 2005). Hagg et al. (2006) examined the effect of an assumed reduction in glacier area in central Asia, but did not model the transient streamflow response associated with changing glacier area. Horton et al. (2006) updated the glacier area for simulation of future conditions assuming a constant accumulation-area ratio, but did not model the transient streamflow response during the period of glacier shrinkage. Only the study by Rees and Collins (2006) appears to have considered the transient response associated with glacier retreat. They assumed a simplified glacier geometry and removed elevation bands as ice thickness depleted.

Glaciologists have used two main approaches to modelling the transient response of glaciers to climate changes. The most physically rigorous is through dynamical modelling (e.g., Oerlemans et al., 1998). However, this approach is computationally intensive and not well suited for extensive regions with sparse climate data and little to no information about past glacier dynamics. The alternative approach is to use a specified relation between glacier volume and glacier area, which was originally derived empirically (Chen and Ohmura, 1990), but which has since been derived through considerations of the physics of glacier flow (Bahr et al. 1997). This approach involves using the volume-area scaling relation to estimate future changes in glacier area that would be associated with changes in glacier volume based on simulated future mass balance. While this approach has been used in many glaciological studies (e.g., van de Wal and Wild, 2001; Radic and Hock, 2006), to our knowledge it has not yet been incorporated into a hydrological simulation.

The objective of this study component was to develop and apply a methodology for estimating changes in streamflow associated with the coupled effects of climatic change and associated glacier response, with a specific focus on transient responses. The approach combines conceptual models of catchment hydrology and glacier mass balance with a model of glacier area evolution based on volume-area scaling. After calibration and validation, the combined model is driven with three different sets of climate scenarios: a scenario created with a weather generator that assumes a continuation of the climate of the last decades;

and two scenarios downscaled from a GCM based on climate response to increased greenhouse-gas concentrations. The approach will allow comparison with the empirical trends presented in Section 2.0 and provide a basis for assessing the hydrologic effects of future glacier changes.

4.2 Study area and data

This study focused on the Bridge River catchment, located in the Southern Chilcotin Mountains, a transition zone from wet coastal mountains to dry interior climate in southern British Columbia, Canada (Figure 4.1). The Bridge River Complex is the third largest hydropower development operated by BC Hydro. Hence, any future changes in streamflow are of considerable socio-economic concern.

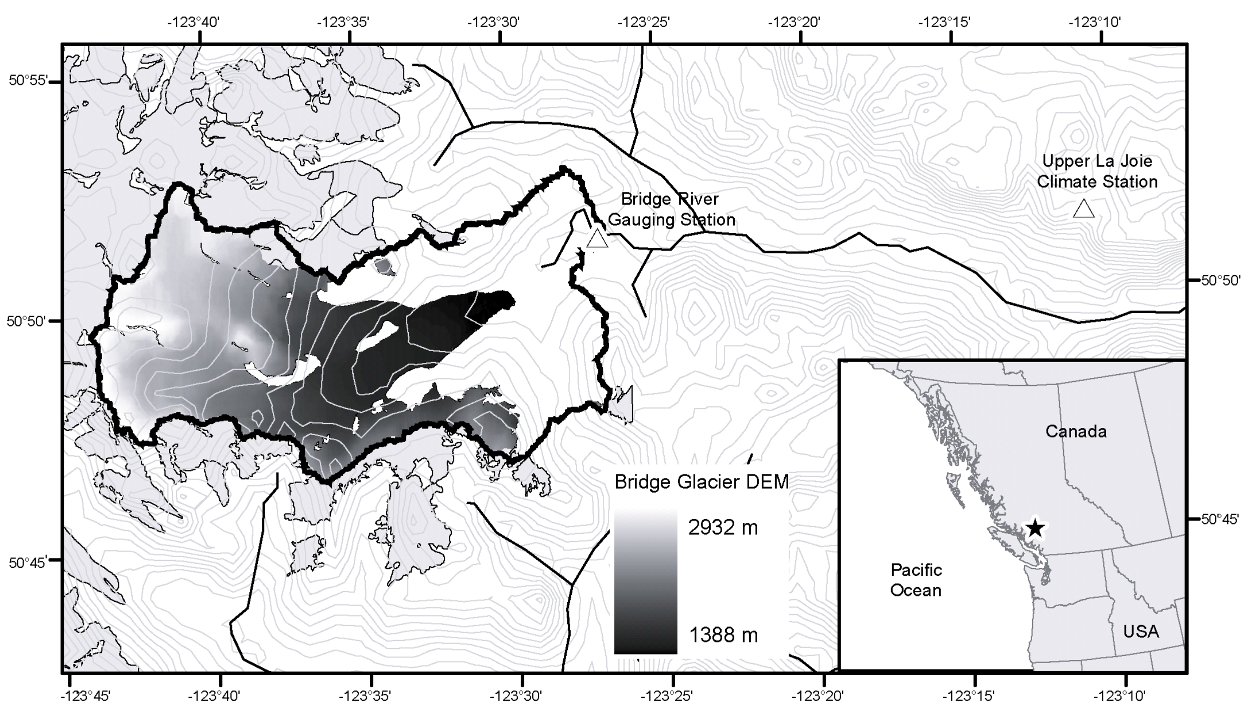


Figure 4.1: Map of Bridge River basin with DEM of Bridge Glacier.

Basin boundary and GIS coverages of land cover information were derived by Environment Canada from the base thematic map (BTM) of British Columbia, which is based on Landsat Imagery from the early 1990s. The basin drains an area of 152.4 km^2, of which the BTM glacier cover is 61.8 % (92 km^2), and spans an elevation range from 1400 m.a.s.l. to 2900 m.a.s.l. Hydrometric data are available since the early 1980s from HYDAT, the database of the Water Survey of Canada. Average annual runoff during the period of record was ca. 2600 mm/year. Climate data are available from a high elevation (1829 m.a.s.l.) weather station ("Upper La Joie"), which has been operated by BC Hydro since 1986, and provides the climate input routinely used in their inflow forecasting models. Daily temperature, precipitation and discharge data of sufficient quality and completeness for application of a hydrological model are available for the period 1985 to 2004.

Mass balance surveys on Bridge Glacier were carried out from 1977 to 1985 (Mokievsky-Zubok et al., 1985; Dyurgerov, 2002). Unfortunately, this period is prior to the period chosen for the hydrological modelling and hence cannot be used directly for validation of the glacier model. However, the total net mass balance record at Bridge Glacier correlates well (r^2= 0.83) with that of Place Glacier (Dyurgerov, 2002), located 80 km to the south-east. For the purpose of this study, Bridge Glacier mass balance values were reconstructed at 100 m intervals (1450 2850 m) using simple linear regressions with the mass balance record at Place Glacier (1850 – 2550 m) at a similar or closest available elevation. Winter and summer mass balance reconstructions were performed separately, and net annual mass balances for each elevation band at Bridge Glacier were calculated as the sum of reconstructed winter and summer balances. The strength of the individual regressions varied from R^2 = 0.41-0.86 (summer balance) to R^2= 0.24 – 0.87 (winter balance), reflecting the broad regional coherence of glacier mass balance.

4.2 Methods

4.2.1 The HBV-EC model

Streamflow was modelled using the HBV-EC semi-distributed conceptual hydrological model (Hamilton et al., 2000), which is based on the Swedish HBV model (Lindström et al., 1997). An earlier version was applied with reasonable success to the 17% glacier-covered Lillooet River catchment, immediately south of Bridge River (Moore, 1993). The current HBV-EC model was integrated into the EnSim™ modelling environment (Canadian Hydraulics Centre, 2006) to leverage data preand post-processing and model visualization capabilities. The model allows for discretization of the watershed into climate zones to account for horizontal gradients in basin climatology. Within each climate zone, the HBV-EC model uses the Grouped Response Unit (GRU) concept to group DEM/GIS grid cells into bins having similar land cover, elevation, slope, and aspect, in order to maintain computational efficiency (Figure 4.2). In this application, Bridge River Basin, which has four types of land cover (glacier, non-forest land, forest, and lake), was discretized into 14 elevation bands of 100 vertical metres, two aspects (north/south) and two slope classes.

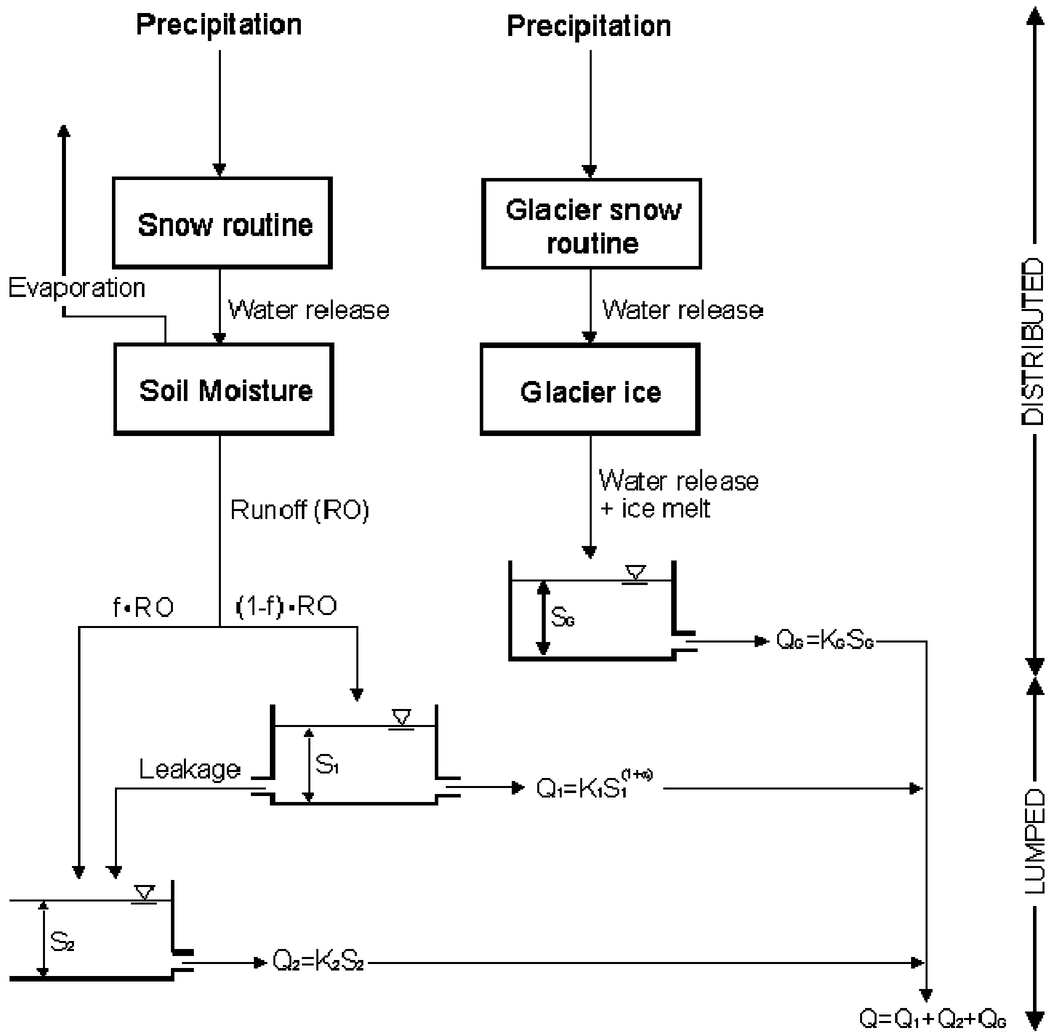


Figure 4.2: Schematic of the HBV-EC model with glacier module.

The model was run using a daily time step, consistent with the resolution of climate data. Correction factors for snowfall (*SFCF*) and rainfall (*RFCF*) adjust recorded precipitation to correct for gauge undercatch as well as bias associated with differences in precipitation between the basin and the climate station. Input climate data were adjusted for elevation using a lapse rate for temperature (*TLAPSE*) and separate gradients for precipitation below (*PGRADL*) and above (*PGRADH*) a threshold elevation (*EMID*). Both temperature lapse rates and precipitation gradients do not vary seasonally. The dominant phase of precipitation (rain vs. snow) occurring within an elevation band is determined by the threshold temperature (*TT*) and mixed-phase precipitation can occur within a temperature interval (*TTI*) around the threshold temperature. In forested areas, interception loss is treated as a constant fraction of precipitation, with separate fractions applied to rain and snowfall.

Snow melt is computed using a temperature index approach based mainly on HBV algorithms. The parameters include a threshold temperature for melt (*TM*) and a base melt factor (C_0) that varies sinusoidally from a minimum value (*Cmin*) on the winter solstice to a maximum value on the summer solstice (*Cmin + DC*). The melt factor for open, flat areas is adjusted for slope (s) and aspect (a) following a simple trigonometric function:

$$[4.1] \qquad C'(t) = C_o(t) \cdot \left(1 - AM \cdot \sin(s) \cdot \cos(a)\right)$$

where *AM* is a calibration parameter. In forested areas, the melt factor is further multiplied by *MRF* (ranging between 0 and 1) to account for the shading and sheltering effects of forest cover on melt rates. The melt factor for glacier GRUs is enhanced by the coefficient *MRG* once seasonal snowpack has ablated to reflect a reduction in surface albedo associated with a bare ice cover.

Refreezing of liquid water can occur when air temperature is below the melt threshold, at a rate governed by the parameter CRFR. Soil evaporation, soil moisture storage and drainage are modelled according to established HBV algorithms (Hamilton et al., 2000).

Runoff routing is computed separately for glacierized and non-glacierized GRUs. For non-glacierized GRU's, water draining from the soil reservoirs enters a common set of lumped reservoirs, representing "fast" and "slow" drainage, as is common in HBV implementations (Hamilton et al., 2000). In contrast, each glacier GRU has a separate reservoir for runoff routing, outflow from which is calculated as:

$$[4.2] \qquad Q_g(t, g) = KG(t, g) \cdot S(t, g)$$

where $S(t, g)$ is the liquid water stored in the reservoir at time t for glacier GRU g (mm) and $KG(t, g)$ is a time-varying outflow coefficient, parameterized as a function of snowpack water equivalent:

$$[4.3] \qquad KG(t, g) = KG_{\min} + dKG \cdot \exp[-AG \cdot SWE(t, g)]$$

where $KG(t, g)$ is the outflow coefficient for time t and glacier GRU g ($time^{-1}$); KG_{min} is a minimum value, representing conditions with deep snow and poorly developed sub-, enand supra-glacial drainage systems ($time^{-1}$); $KG_{min} + dKG$ is the maximum outflow coefficient, representing late-summer conditions with bare ice and a well developed glacial drainage system; *AG* is a calibration parameter (mm^{-1}); and *SWE(t, g)* is the snowpack water equivalent for time *t* and glacier GRU *g* (mm). The streamflow at the basin outlet is the sum of the outflow from the fast reservoir (Q_1), the slow reservoir (Q_2) and the glacier reservoirs ($Q_G(g)$).

4.2.2 Glacier mass balance calculation

The glacier module allows the calculation of annual glacier mass balances by GRU from the modelled time series of snow water equivalent (*SWE*) and ice melt (M_{ice}). By analogy with the stratigraphic method for mass balance computation (Østrem and Brugman, 1991), the winter balance for a given year and GRU, $b_w(t,g)$ is the maximum daily value of $SWE(t,g)$ for that year. The summer balance (b_s) depends on whether or not all of the snow melts off the glacier, with

[4.4] $$b_s(g) = \min[SWE(t,g)] - \max[SWE(t,g)] \qquad \text{if } \min[SWE(t,g)] > 0$$

and

[4.5] $$b_s(g) = -\left[\max[SWE(t,g)] + \sum M_{ice}(g)\right] \qquad \text{if } \min[SWE(t,g)] = 0$$

where $\sum M_{ice}(g)$ is the cumulative ice melt from the date of snow disappearance to the beginning of the next accumulation season.

The net balance ($b_n(g)$) for each GRU is then

[4.6] $$b_n(g) = b_w(g) + b_s(g)$$

and the total mass balance *MB* (as a volume) for the glaciers in the basins in a given year can then be computed as

[4.7] $$MB = \sum_g \left(A_g \cdot b_n(g)\right)$$

where *A* is area and the subscript *g* refers to each glacier GRU.

4.2.3 Model calibration and validation

Calibration was conducted by trial and error for the ten year period from 1986 to 1994 and validated for the period 1995-2004, primarily on the basis of visual inspection of hydrographs. The parameters were varied within physically plausible ranges and chosen to fit not only the streamflow hydrograph but also the reconstructed summer and winter glacier mass balances at different elevations for Bridge Glacier.

4.2.4 Simulating glacier retreat and advance

Glacier advance or retreat was simulated by combining the modelled mass balance with the volume-area scaling relation introduced by Chen and Ohmura (1990), which was based on measured geometries of alpine glaciers around the world:

[4.8] $V = \mathrm{b} \cdot A^{1.36}$

[4.9] $A = (V/\mathrm{b})^{0.735}$

where V is the glacier volume (m·km^2), b = 28.5, and A is the glacier area (km^2). The physical basis of the relation was confirmed by Bahr et al. (1997), who also updated the empirical relation with additional data. Bridge Glacier lies well within the range of glacier sizes that was used to derive the empirical volume area scaling relation. As no observations of the glacier depth exist for Bridge Glacier, the initial volume was derived from the current glacier area using eq. (8). Each decade, the volume was updated by adding the computed change in volume, determined from the accumulated mass balance for the period (MB), and the new area (A_{new}) was computed using the scaling relation (eq. 9). The area that had to be removed or added was computed as

[4.10] $\Delta A = A_{\mathrm{new}} - A_{\mathrm{old}}$

The removal (or addition) of glacier cover from the land use grid was simulated by an algorithm coded in IDL (Interactive Data Language). The glacier area change was first converted into the number of glacier pixels to be removed or added to the glacier. Glacier pixels were then removed or added along the edge of the glacier using a morphologic erosion or dilation operator, a common method in digital image processing to fill holes or remove islands and to expand/grow or reduce/shrink a binary image. The erosion or dilation is performed iteratively, removing one row of pixels in each of “n” passes, stopping when the total number of desired pixels has been removed or added. To reduce glacier area, glacier pixels along the glacier edge are first eroded from the lowest elevation band, then from the two lowest elevation bands, and so on up to the equilibrium line elevation (determined from the glacier mass balance model). This iterative removal mimics the stronger retreat of the glacier tongue and the effect can be increased by setting a higher number for the parameter "n" to erode proportionally more from lower elevations. Newly glacier-free areas were considered non-forest land for subsequent hydrologic simulations. To grow the glacier, the reverse procedure was performed. The glacier was allowed to grow into the area of its historical extent.

4.2.5 Climate scenarios

Three sets of future climate scenarios were established to model future glacier change and its impact on hydrology: S0, a continuation of the observed 1986-2004 climate over the next 200 years; and S1 and S2, representing climate downscaled from the SRES-B1 and A2 scenarios as calculated by the third version of the Canadian Centre for Climate Modelling and Analysis Coupled Global Climate Model (CGCM3) for the IPCC's Fourth Assessment Report (AR4).

The climate input variables for S0 were derived using the stochastic weather generator LARS-WG [*http://www.rothamsted.bbsrc.ac.uk/mas-models/larswg.php*]. LARS-WG generates weather data based on statistics of wet and dry spells, the distribution of rainfall amounts and temperature variability, which is conditioned on the wet and dry series (Semenov, 1998).

Scenarios S1 and S2 were derived using the TreeGen statistical downscaling model. TreeGen relates observed synoptic-scale atmospheric predictor fields to observed surface weather elements, and then, based on these relationships, generates realistic series of weather elements from GCM fields (details in Appendix 3). Predictands in the TreeGen downscaling model were daily minimum temperature, maximum temperature, and precipitation amounts observed at the Upper La Joie weather station from 1985-2005. Gridded synoptic-scale mean sea-level pressure, surface air temperature, and surface precipitation data from the US National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis (Kalnay et al., 1996) were used to define the historical synoptic map-types controlling daily weather conditions at Upper La Joie. Daily mean maps were obtained for the period 1961-2005 for a region covering western North America and the eastern Pacific Ocean (30°N-70°N; 160°W-110°W) on a grid subsampled to a spacing of 5° by 5°. Synoptic-scale fields matching those from the NCEP/NCAR Reanalysis were obtained from transient greenhouse gas plus aerosol runs of CGCM3 for simulated years 1961-2100.

Forcing data for the scenarios S1 and S2 were from the IPCC SRES B1 and A2 scenarios, respectively. The SRES B1 scenario assumes low population growth, rapid changes towards a service and information economy, and the introduction of clean and resource-efficient technologies, whereas the A2 scenario assumes high population growth, an emphasis on regional economic development, and slower technological change. Equivalent CO_2 concentrations for the B1 scenario are projected to increase to 1.5 times year 2000 levels by 2100, while those for the A2 scenario are projected to increase to 2.5 times year 2000 levels by 2100.

For each scenario, ten realisations were computed creating an ensemble of ten members (M1, M2, ..., M10). Each member consists of a time series of daily temperature and precipitation. In the case of S0 these are 200 years long and are based on the statistics of the observed data from 1986 to 2005; in the case of S1 and S2 the time series represent the transient climate change as modelled by the GHG forced GCM scenarios. All scenarios were then used to drive the HBV-EC model with the glacier being re-scaled at the end of every decade based on the decade's cumulative mass balance. The effect of glacier re-scaling on streamflow development was tested by applying a scenario run S1a, which had the same climate input as in S1 but without application of glacier scaling throughout the scenario period. That is, S1a maintained the glacier at its current extent.

4.4 Results

4.4.1 Calibration and validation of HBV-EC

The calibration procedure resulted in a streamflow model efficiency (Nash and Sutcliffe, 1970) of E = 0.88 for the calibration period and E = 0.92 for the validation period. The calibrated model well simulates the interannual variation in snowmelt and the glacial hydrograph in general, but it systematically underestimates the (low) winter streamflow (Figure 4.3). Values for most model parameters (Table 4.1) are comparable to those found in other applications of HBV-based models in glacierized catchments (e.g. Moore, 1993). The range of mass balances and their variation with elevation during the calibration period is reproduced correctly by the model (Figure 4.4), except for the modelled summer balance in the lowest elevation bands, which is lower than the summer balances reconstructed from Place Glacier. However, Place Glacier's lowest elevation is only at around 2000 m.a.s.l. Therefore, the mass balance reconstruction for Bridge Glacier's lower elevation bands had to be carried out based on correlations with much higher elevation bands on Place Glacier and reconstructed values may, therefore, not be as reliable as for higher elevations. Mass balances in the higher elevation bands were represented well and the interannual variability of the mass balance for the entire glacier is particularly well reproduced from the beginning of the 1990s.

Table 4.1 Important HBV parameters

Model Component	**Parameter**	**Description**	**Value**
Climate	RFCF	Precipitation correction factor	1.6
	PGRADL	Fractional precipitation increase with elevation (m^{-1})	0.0013
	PGRADH	Fractional precipitation increase above EMID (m^{-1})	0.0005
	EMID	Mid point elevation separating precipitation gradients (m)	2200
	TLAPSE	Temperature lapse rate (°C m^{-1})	0.0065
	TT	Threshold air temperature to distinguish rain/snow (°C)	0
	TTI	Temperature Interval with mixed rain/snow (°C)	2
Snow	AM	Influence of aspect/slope on melt factor	0.25
	TM	Threshold temperature for snowmelt (°C)	0
	CMIN	Melt factor for winter solstice in open areas (mm·°C^{-1}·day^{-1})	1.6
	DC	Increase of melt factor between winter and summer solstice (mm·°C^{-1}·day^{-1})	1.8
	MRF	Ratio between melt factor in forest and melt factor for open areas	0.7
	CRFR	Rate of liquid water refreezing in snowpack (mm·°C^{-1}·day^{-1})	2
Glacier	MRG	Ratio of melt of glacier ice to melt of seasonal snow	1.5
	AG	Factor controlling relation between SWE on glacier and runoff coefficient (mm^{-1})	0.1
	DKG	Difference between maximum and minimum outflow coefficients for glacier water storage (day^{-1})	0.3
	KGmin	Minimum outflow coefficient (day^{-1})	0.05

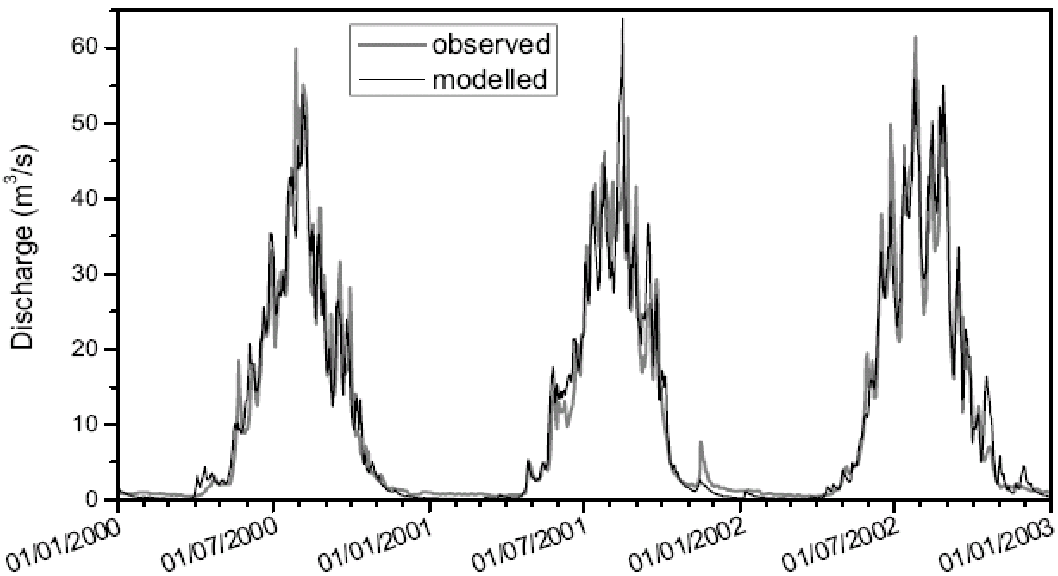


Figure 4.3: Observed and modelled hydrographs for three years of the validation period.

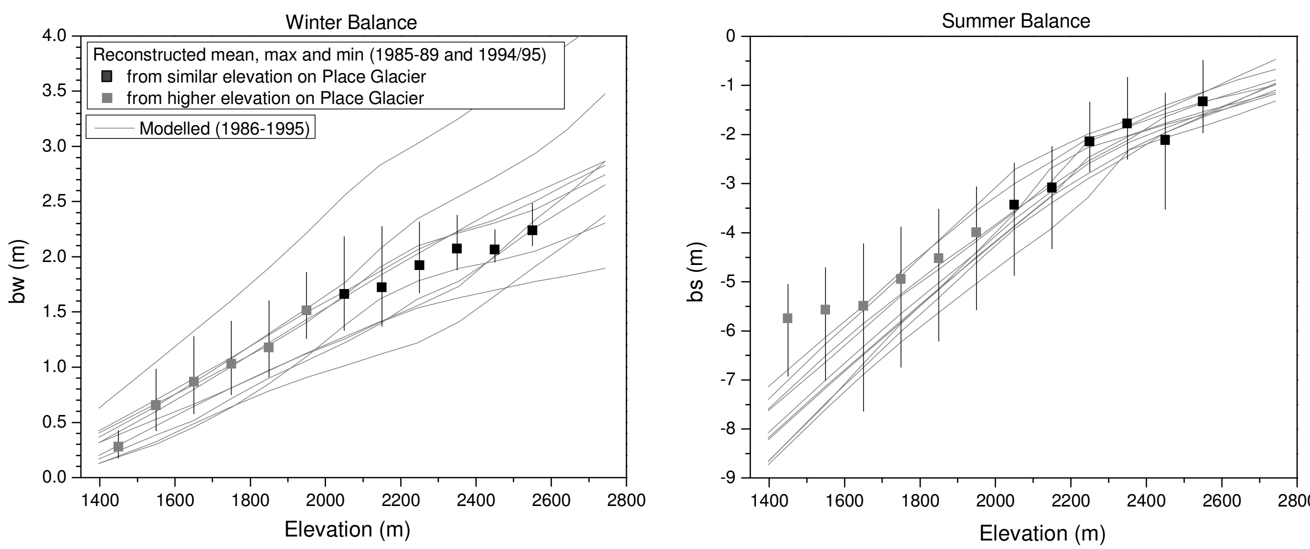
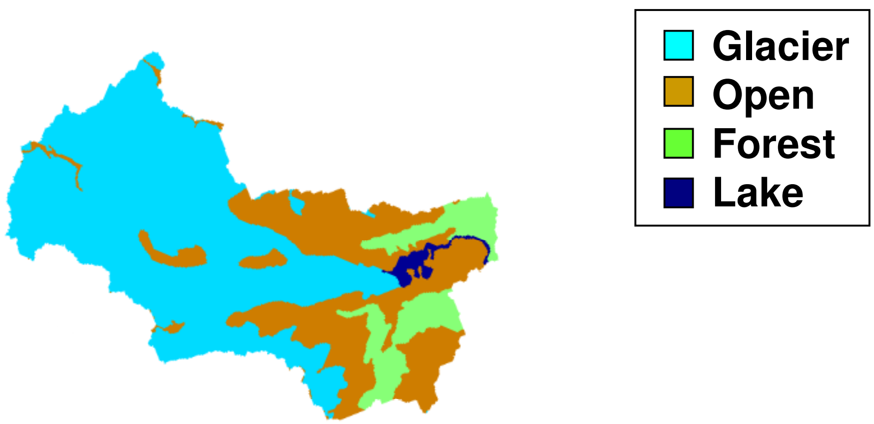


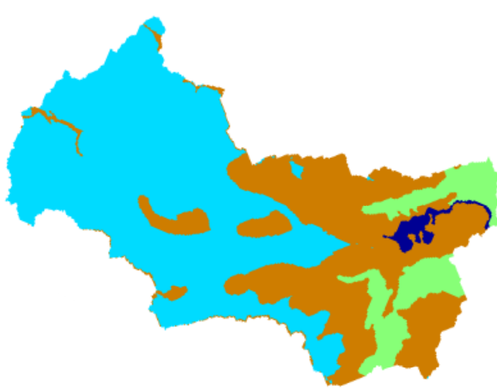
Figure 4.4: Reconstructed and modelled winter (left) and summer (right) mass.

The modelled cumulative net mass balance of the glaciers in Bridge River Basin for the calibration period was10.8 m and15 m for the validation period. The initial volume-area scaling for the validation period resulted in a loss of glacier area of 8 km^2 from the initial 92 km^2 in the 1990s, for an updated area of 83.7 km^2 (Figure 4.5). Modelled glacier change between 1995 and 2004 compares well with recent pictures from Bridge Glacier, though the north branch lost connection with the main glacier tongue during that period.

a) Initial landuse 1990s



b) Landuse 2005 (after validation period)



c) Bridge Glacier in 2006

Figure 4.5: Land cover in Bridge basin: a) initial glacier extent, b) re-scaled glacier after validation period in 2005, c) photo of Bridge Glacier in 2006.

As the downscaling was performed from the transient CGCM3 scenario for the period 1961 to 2100, climate variables for 1986-2004 could be compared with the observed climate and were used to run the HBV-EC model with the calibrated parameters. This allows the validation of the downscaled climate in terms of the correct reproduction of mass balance and runoff variables (Table 4.2). The average of the ten realisations shows slightly higher mean annual temperatures (+0.15°C) and slightly lower amounts of annual precipitation (-50 to120 mm/year). However, the observed values lie well within the span of the ten realisations. The same applies to modelled mass balances and mean annual runoff.

Table 4.2: Mean annual variables modeled for 1986-2004 (range for 10 different realizations)

Climate Input	**At-station Air Temperature (°C)**	**Basin Precipitation (mm)**	**Cumulative Mass Balance (m)**	**Mean annual Runoff (mm)**
Observed	1.02	1998	-25.8	2640
S1 (M1,...M10)	1.17 (±0.29)	1936 (±182)	-23.2 (±9.7)	2532 (±217)
S2 (M1,...M10)	1.18 (±0.25)	1870 (±137)	-25.1 (±5.5)	2517 (±141)

4.4.2 Future developments under scenarios S0-S2

While Scenario S0 keeps the climate input constant, downscaled scenarios S1 and S2 mainly exhibit changes in temperature (Table 4.3). Annual temperatures for S2 increase particularly strongly after the middle of this century. A projected increase in winter precipitation is offset by a decrease in summer precipitation. Annual precipitation totals therefore remain relatively constant, with the same bias to S0-precipitation as the downscaled values for 1986-2004 showed to observed precipitation (Tables 4.2 and 4.3).

The application of scenario S0 results in several more decades of negative mass balances, which causes a rapid further retreat of the glacier over the next few decades. Mass balances then gradually become smaller and the glacier reaches its equilibrium (where net mass balances fluctuate around zero) after ca. 90 years of simulation time (Figure 4.6). The glacier area decreases from the 83.7 km^2 in 2004 to 67.4 km^2, representing a loss of 20% of the present glacier area. Assuming no further climate change, the glacier coverage of the Bridge River basin could hence be expected to shrink from the original 62 % in the 1990s BTM map to ca. 43% at the end of th2 21st century. This represents the equilibrium state in a climate similar to that of 1986-2004. In the new equilibrium, most of the two present glacier tongues disappear (Figure 4.7). The glaicer shrinks even more rapidly in the downscaled climate change scenarios (Figures 4.6 and 4.7). Mass balances remain negative for S1 and strongly negative for S2 at the end of the century, with glacier area shrinking to 56 km^2 (33% change) and 51 km^2 (39% change), respectively.

Table 4.3 Mean annual climate input and runoff output modeled for future time slices (range for 10 different realizations.

Time Slice	Climate Scenario	At-station Temperature (°C)	Basin Precipitation (mm)	Mean annual Runoff (mm)
2045-2055	S0	1.02 (±0.2)	1976 (±94)	1980 (±55)
	S1	1.92 (±0.23)	1952 (±144)	2069 (±125)
	S2	2.12 (±0.21)	1850 (±146)	2153 (±107)
2085-2095	S0	0.95 (±0.24)	1984 (±130)	1779 (±96)
	S1	2.44 (±0.36)	1922 (±164)	1879 (±125)
	S2	3.53 (±0.20)	1867 (±130)	1977 (±132)

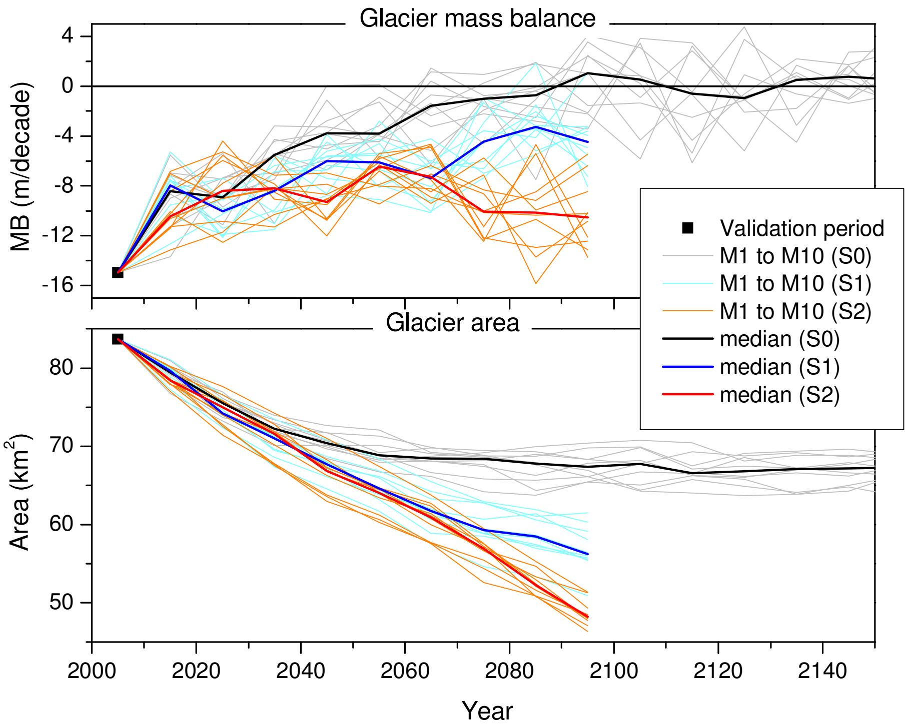


Figure 4.6: Development of simulated decadal glacier mass balance (upper panel) and glacier area (lower panel)

The glacier recession in all scenarios produces a decline in annual streamflow (Table 4.3). The streamflow decline is strongest in the summer months in the next few decades. For scenario S0 the initial drop of flow in August is followed by an asymptotic approach to an equilibrium flow of ca. 26 m^3/s, which is a reduction of 37% from the current 41 m^3/s (Figure 4.8). For S1 and S2 mean August streamflow will be even further reduced to 23 to 24 m^3/s by 2095. The results for scenario S2 reflect the effect of a change in temperature increase by the middle of the century, when mass balances are even more negative and a relative increase in melt offsets the decrease in glacier area reduction.

To illustrate the effect of glacier re-scaling on streamflow development Figure 4.9 shows mean monthly streamflow of three time-slices for S1a and S1. With the same climate input, modelled streamflow increased considerably when the glacier are was kept constant, while it decreased considerably when the glacier area was reduced.

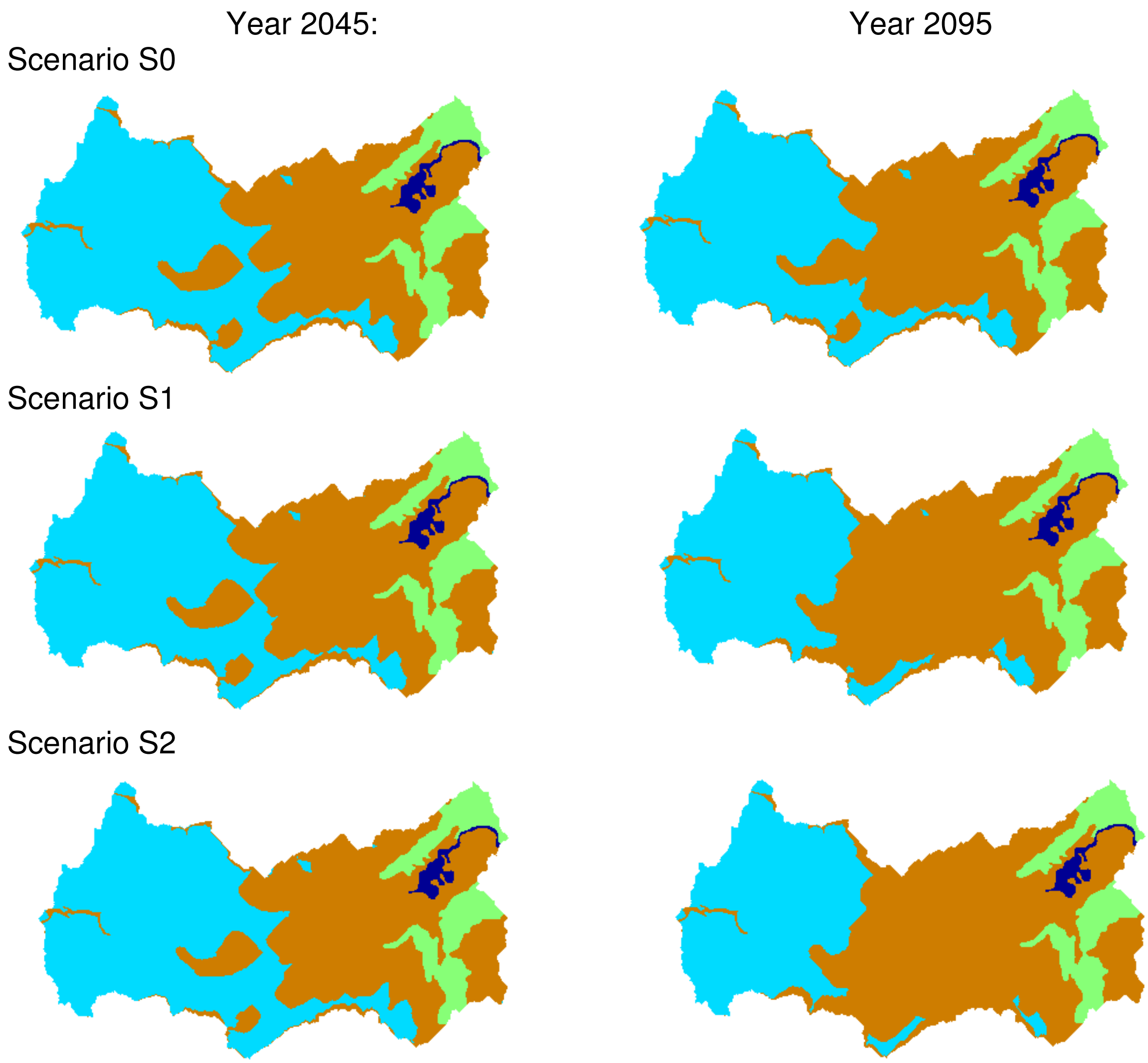


Figure 4.7: Glacier coverage in the Bridge River basin in 2045 and 2095 under different climate scenarios (legend as in Fig. 4.5: light grey: glacier, orange: non-forest, green: forest, blue: current proglacial lake and river).

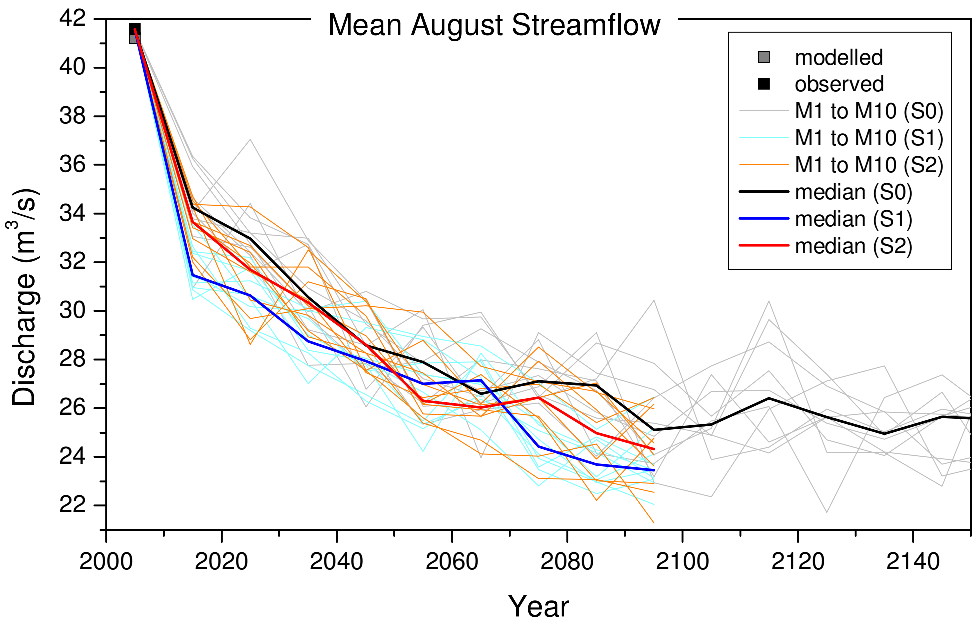


Figure 4.8: Change in decadal mean August streamflow. Individual realizations are shown as faint lines; medians for each scenario are shown as prominent lines.

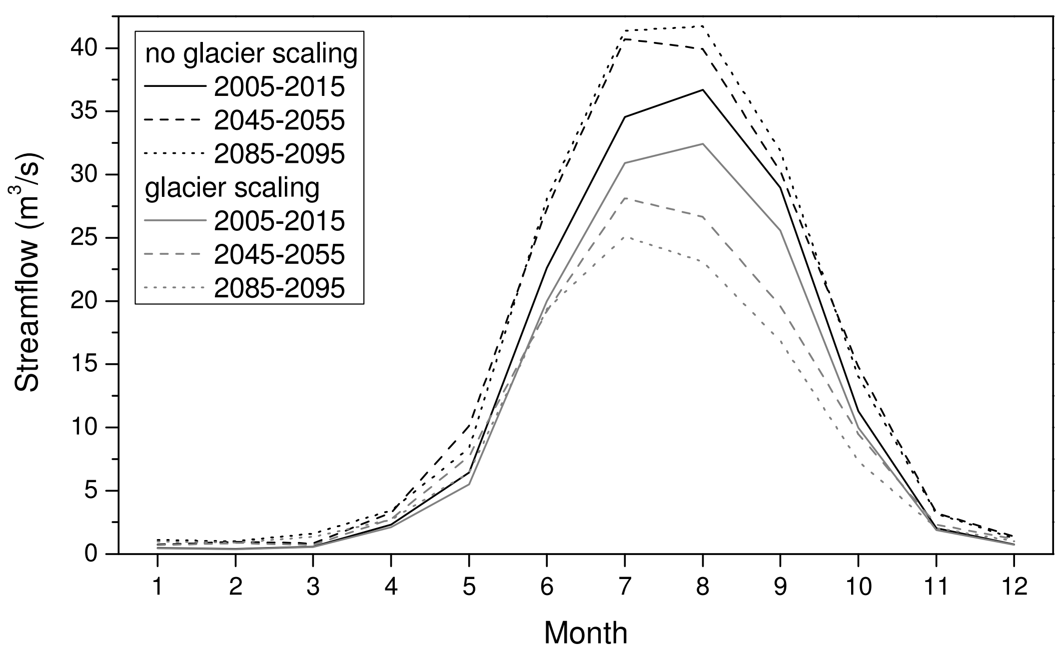


Figure 4.9: Streamflow response to the first member of S1 (with glacier scaling) and S1a (without glacier scaling).

4.5 Discussion

In general the model fit to both streamflow and glacier mass balance is good and within the performance range of similar modelling applications in glacierized watersheds. We are therefore confident that the model is appropriate for estimating glacier changes and glacial meltwater contributions to streamflow. However, the model tended to produce some systematic underand overestimation of discharge, particularly in winter. These errors can most likely be improved in future applications through fine-tuning of temperature lapse rates or alternative meteorological input. The climate station used in this study is located ca. 20 km from the basin outlet at a relatively high elevation of 1829 m. Though this station is fairly close to the basin, considering the remoteness of the area, the strong climatic gradients in the area required that some adjustments be made to the daily weather data. For example, the value of $RFCF = 1.6$ reflects the drier climate at the station location compared to the glacier.

Additional sources of error for the calibration period include the land cover from the mid-1990s (i.e. the end of the calibration or beginning of validation period). Glacier extent during the calibration period (1985-1994) may have been somewhat greater, given that the glacier retreat likely occurred through that period (Menounos and Wheate, unpublished data). Furthermore, mass balance survey data for Bridge Glacier used a slightly different glacier hypsometry and the entire Bridge River basin (and hence the hydrological model) includes additional small glaciers that may cause some difference in the overall modelled mass balance.

The use of volume-area scaling for modelling glacier changes assumes that there are no lags in response. Glacier dynamics can stall retreat or even maintain advance for some time following an initial shift to negative mass balance. Even discounting dynamic effects, mass loss may initially be dominated by thinning rather than terminal retreat. In this application, Bridge Glacier was already in a state of retreat, suggesting an initial stage of thinning had already occurred. A related issue is that the coefficients in the volume-area scaling relation are not specific to the region, but stem from a dataset of glaciers worldwide (Chen and Ohmura, 1990). Though it has been applied in several studies for that purpose, the extent to which the V-A scaling relation can be used to model temporal changes, as opposed to spatial variations, has not been adequately tested on surveyed data. Subsequent model applications therefore need to verify earlier glacier extents using maps and airphotos and test the sensitivity to the assumed volume-area scaling relation.

Additional processes not captured in the current model could influence streamflow in a changing climate. For example, Braun and Escher-Vetter (1996) showed that the retreat of firn cover can have a strong influence on glacier runoff in the period following a shift from positive to negative net balance. However, this process may not be so important in the southern Coast Mountains, where signficant firn depletion already appears to have occurred (Moore and Demuth, 2001). In addition, the current model does not represent the effects of vegetation establishment following glacier retreat or other vegetation changes driven by climate change, such as tree establishment at higher elevations. However, these processes are likely to have a second-order effect on streamflow at Bridge River, relative to the effects of glacier retreat.

Despite these issues, the study clearly shows the importance of re-scaling glaciers when assessing the influence of a warming climate on hydrology. The assumption of a continuation of the climate of the past 20 years in S0 has been found to be useful to determine the theoretical equilibrium geometry of Bridge Glacier (and hence streamflow from Bridge Basin). As the glacier is in fact far from its equilibrium, it will retreat rapidly within the next few decades and if temperatures further rise the retreat will continue thereafter. As the glaciated area in the basin shrinks rapidly, streamflow reductions for the summer months follow the same pattern. The decrease may only temporarily be slowed during times when temperature increases more strongly, as was shown in the second half of the century for S2. Predicted precipitation increases in BC during the winter months are not large enough to offset this development. Some tests with precipitation increases of 5% to 15% added to S0 (not shown), as well as the results from the downscaled scenarios, illustrate this point. This result confirms that glacier retreat appears to be a major cause for the negative streamflow trends in August found across BC (Stahl and Moore, 2006) and that the period of initial meltwater increase has passed.

4.6 Conclusions

In this study we presented a methodology to simulate the transient effects of glacier retreat on streamflow patterns by coupling a semi-distributed hydrological model with a glacier mass balance and glacier scaling model. The volume-area scaling approach has been applied in glaciological studies and is implemented into global climate models, but to date does not appear to have been used in hydrological modelling. While the validity of the volume-area scaling for temporal changes in glacier has not been confirmed empirically, it provides a viable approach to generating first-order estimates of coupled glacier and streamflow response to climate change. In particular, extrapolation of current climatic conditions into the future highlights the extent to which current glacier extent and streamflow regime are out of equilibrium with the current climate. However, further research should focus on testing the validity of the use of volume-area scaling for temporal changes in glacier extent. In addition, algorithms should be developed for simulating the effects of firn dynamics, which appear to have an important effect on glacier runoff in the initial shift from a positive to negative mass balance.

The results of this study suggest that climate warming and associated glacier retreat will have significant implications for water resources and aquatic ecology. Glacier-fed rivers are likely to experience a shift from a glacial regime with high flows in mid and late summer, with an associated moderating effect on stream temperature, to a regime that responds to the summer dry period with streamflow recession, low flows and increased temperatures.

5. Conclusions

5.1 Summary of key findings

Streamflow patterns in BC can be broadly classified as rain-dominated, hybrid (rain and snow), snow-dominated, and glacierized, each having distinct low flow seasons. In this study, we chose to focus on the summer/early autumn low flow period, as that has particular economic and ecological significance.

A field study examined variability in stream discharge was measured over a low flow season. The coefficient of variation of the mean discharge tended to increase when the flow is very low. Variations in measurements are of sufficient magnitude that subtle changes in streamflow due to, for example, climate change may be difficult to detect.

Unglacierized catchments generally exhibited declining trends in September streamflow through most of British Columbia, but not in August. Glacierized catchments, on the other hand, show the opposite pattern, with dominantly negative trends in August but not September. The decreases in September flows are broadly consistent with a decline in September precipitation over the period studied. For all four streamflow regimes, the most important control on August streamflow is August precipitation. August streamflow is also positively but more weakly related to lagged variables including July precipitation and the previous winter's precipitation. It is commonly acknowledged that General Circulation Models perform less well at simulating precipitation than air temperature. This problem will likely be even more severe for summer precipitation, much of which is generated by convective uplift, because convection operates at spatial scales that are not resolved by GCM's. Therefore, it appears that our ability to predict August streamflow under future climate scenarios will be limited by the uncertainty in the projections of summer precipitation. The ongoing development of higher-resolution Regional Climate Models (RCM's) may help in this regard.

There is clear evidence for the effect of multi-year storage change only for the glacierized catchments. Glacier retreat over the last few decades has apparently reduced glacier area sufficiently so as to reduce meltwater generation and thus streamflow. Snowmelt-dominated catchments showed no tendency to trends or serial correlation in the model residuals, though the runs test suggested a substantial number of stations had non-random residuals. Both rain-dominated and hybrid catchments tended to have positive residuals in August, suggesting either an increasing trend in groundwater storage or a spurious cause such as model mis-specification.

The majority of groundwater well observation records in BC start in the late 1970s or 1980s, providing a total of 20 to 30 years of record. A select few of these wells monitor aquifers in pristine areas that reflect natural variability; the others have been influenced by human activity making them less representative. By comparing the relationships between groundwater, climate, and surface water within and between groups of well records from the two major hydro-climatic zones in BC (rain-dominated and snowmelt-dominated), a system for detecting the influence of climate change and variability on groundwater in the absence of long term records is defined, and suitable correlation coefficients are applied to evaluate the strength of these interactions. Different aquifer types are assessed with respect to vulnerability to climate change influences. Overall, summer groundwater levels appear to have lowered across the province, despite an increase in winter precipitation and recharge during the same time period. Due to the limited availability of long well records near gauged streams, the attribution of whether and how these changes have affected low flows proved difficult. The available groundwater observation wells vary substantially in terms of aquifer properties and the hydraulic connection to rivers and for many, we suspect the natural records may be altered by changes in the abstraction patterns. The few examples where streamflow and groundwater observations are available in the same basin and are nearby furthermore provide some of the exceptions to the trends found. In Abbotsford, even though a field study and a modelling study showed that summer low flow is fed by groundwater, low flows increased while over the same time period groundwater levels decreased. The same was found for Grand Forks, where previous studies have shown that streamflow is fed by groundwater in the summer. However, the recharge mechanisms are different in both examples.

A methodology was developed to simulate the transient effects of glacier retreat on streamflow patterns under future climate scenarios by coupling a semi-distributed hydrological model (HBV-EC) with a glacier mass balance and glacier scaling model. The model application shows strong reductions in glacier area and summer streamflow even under the assumption of a continuation of the present climate (i.e., no further warming). Climate warming and associated glacier retreat will have significant implications for water resources and aquatic ecology. Glacier-fed rivers are likely to experience a shift from a glacial regime with high flows in mid and late summer, with an associated moderating effect on stream temperature, to a regime that responds to the summer dry period with streamflow recession, low flows and increased temperatures.

5.2 Recommendations for further work

The availability, quantity and record length of existing data limited our ability to analyse and model the roles of changes in glaciers and groundwater storage in summer low flows, especially in relation to future climate scenarios. The problem was most severe for examining climate-groundwater-streamflow linkages. The majority of well water level records are simply of insufficient length, continuity and quality to provide a clear picture of interannual variability, particularly in relation to climate variations and change. Furthermore, the majority of existing records are impaired by the effects of extraction at nearby wells, and therefore do not represent a clean response to climatic forcing. Linking groundwater and streamflow variations is particularly hampered by the paucity of data, given that there are almost no locations where unimpaired well levels are available within a catchment where streamflow is gauged.

The provincial and federal governments should cooperate to rectify this situation. Monitoring wells should be drilled in aquifers within gauged catchments and instrumented with state-of-the-art equipment and processed using appropriate Quality Assurance/Quality Control protocols. The aquifer classification scheme outlined in Section 3 of this report will provide a robust basis for designing an appropriate scheme for sampling the range of climate-groundwater-streamflow linkages.

While statistical modelling supports the hypothesis that glacier retreat has resulted in negative trends in August streamflow, it will be important to back up these results with deterministic modelling, drawing upon records of changes in glacier area and volume. Such records are currently being produced by the Western Canadian Cryospheric Network, which is funded by the Canadian Foundation for Climate and Atmospheric Science, and should be available for analysis within the next two to three years.

There is also a need for more data to support modelling studies of coupled glacier and streamflow response to future climate scenarios. British Columbia has a woefully sparse network of climate stations, particularly in the glacierized mountain regions. Another problem facing the application of deterministic models is the compensating influences of precipitation and glacier melt. If a model is incorrectly calibrated such that precipitation is overestimated, there will tend to be less ice melt, due to the lower albedo of snow relative to ice. On the other hand, if precipitation is underestimated, increased ice melt can compensate. In the current study, mass balance data for Bridge Glacier proved invaluable as an aid for calibrating the model. However, it may be possible to use other types of glacier data, such as end-of-summer snowlines, as used by Moore (1993) and which can be extracted from remote sensing imagery, or volume and area changes as determined from existing maps and remote sensing products. Further research should examine the value in these alternative types of glacier data as support for model calibration and testing.

6 References

Allen D.M., Scibek, J., Whitfield, P. and Wei, M. 2004b. Climate Change and Groundwater: Summary Report. Final report prepared for Natural Resources Canada, Climate Change Action Fund, March 2004, 404pp. http://www.adaptation.rncan.gc.ca/projdb/index_e.php?class=118Grand Forks

Allen, D.M., Mackie, D.C. and Wei, M. 2004a. Groundwater and climate change: a sensitivity analysis for the Grand Forks aquifer, southern British Columbia, Canada. Hydrogeology Journal 12: 270-290.

Alley, W.M., Reilly, T.E., and Franke, O.L. 1999. Sustainability of Ground-Water Resources, USGS Circular 1186, Denver, CO.

Bahr, D.B., Meier, M.F. and Peckham, S.D. 1997. The physical basis of glacier volume-area scaling. J. Geophys. Res. 102(B9): 20355-20362, doi:10.1029/97JB01696.

Barnett, T.P., Adam, J.C. and Lettenmaier, D.P. 2005. Potential impacts of a warming climate on water availability in snow-dominated regions. Nature 438: 303-309. doi:10.1038/nature04141.

Bencala, K.E., 2000. Hyporheic zone hydrological processes. Hydrological Processes 14: 2797-2798.

Berg, M.A. and Allen, D.M. 2007. Low flow variability in groundwater-fed streams. Canadian Water Resources Journal 32: 1-20.

Berka, C., Schreier, H., and Hall, K. 2001. Linking water quality with agricultural intensification in a rural watershed. Water, Air, and Soil Pollution 127: 389-401.

Beschta, R.L., Bilby, R.E. Brown, G.W. Holtby, L.B. and Hofstra, T.D. 1987. Stream temperature and aquatic habitat: fisheries and forestry interactions, in Streamside Management: Forestry and Fishery Interactions, edited by E.O Salo and T. W. Cundy, University of Washington, Institute of Forest Resources, Contribution No. 57, Seattle, Washington, pp. 191-232.

Birsan, M., P. Molnar, P. Burlando, and M. Pfaundler, M. 2005. Streamflow trends in Switzerland. Journal of Hydrology 314: 312-329. doi:10.1016/j.jhydrol.2005.06.008

Braun, L.N. and Escher-Vetter, H. 1996. Glacial discharge as affected by climate change, in Interpraevent 1996: Protection of Habitat against Floods, Debris Flows and Avalanches, 24-28 June 1996, Garmisch-Partenkirchen, Germany, pp. 65-74, Conference Proceedings, Internationale Forschungsgemeinschaft, Klagenfurt.

British Columbia Ministry of Environment (2006) Observation well network http://srmapps.gov.bc.ca/apps/gwl/disclaimerInit.do

British Columbia Ministry of Environment (2006) WELLS database http://aardvark.gov.bc.ca/apps/wells/

British Columbia Ministry of Environment (2006) Water Resources Atlas http://www.env.gov.bc.ca/wsd/data_searches/wrbc/index.html

BC Ministry of Environment. 2006. British Columbia's Coastal Environment: 2006 http://www.env.gov.bc.ca/soe/bcce/

Brown, L.E., Hannah, D.M. and Milner, A.M. 2005. Spatial and temporal water column and streambed temperature dynamics within an alpine catchment: implications for benthic communities. Hydrological Processes 19: 1585-1610.

Buishand, T.A. and Brandsma, T. 2001. Multisite simulation of daily precipitation and temperature in the Rhine basin by nearest-neighbor resampling, Water Resour. Res. 37: 2761–2776.

Burn, D.H. and Hag Elnur, M.A. 2002. Detection of hydrologic trends and variability. Journal of Hydrology 255: 107-122.

Canadian Hydraulics Centre 2006) EnSim Hydrologic Reference Manual. National Research Council, Ottawa, Ontario, 251 pages.

Cannon, A.J. and Whitfield, P.H. 2002. Downscaling recent streamflow conditions in British Columbia, Canada using ensemble neural network models. J. Hydrol. 259:136-151.

Cannon, A.J., Whitfield, P.H. and Lord, E.R. 2002a.Synoptic map-pattern classification using recursive partitioning and principal component analysis. Monthly Weather Review 130: 1187-1206.

Cannon, A.J., Whitfield, P.H. and Lord, E.R. 2002b. Automated, supervised synoptic map-pattern classification using recursive partitioning trees. In: Preprints, The 16th Conference on Probability and Statistics in the Atmospheric Sciences, American Meteorological Society, p. J210-216.

Cayan, D.R., Dettinger, M.D., Kammerdiener, S.A., Caprio, J.M. and Peterson, D.H. 2001. Changes in the onset of spring in the western United States. Bulletin Amer. Met. Soc. 82:399-416.

Chambers, J., Cleveland, W., Kleiner, B., Tukey, P. 1983. Graphical methods for data analysis. Wadsworth & Brooks/Cole.

Chen, J. and Ohmura, A. 1990a. On the influence of Alpine glaciers on runoff. Hydrology in mountainous regions I. Hydrologic measurements, the water cycle, Proceedings of the two Lausanne Symposia, IAHS Publ. no. 193, 117-125.

Chen, J. and Ohmura, A. 1990b.Estimation of alpine glacier water resources and their change since the 1870s. Hydrology in mountainous regions I. Hydrologic measurements, the water cycle, Proceedings of the two Lausanne Symposia, IAHS Publ. no. 193, 127-135.

Chen, Z., Grasby, S.E. and Osadetz, K.G. 2002. Predicting average annual groundwater levels from climatic variables: an empirical model. Journal of Hydrology 260: 102-117

Collins, D.N. 1987. Climatic fluctuation and runoff from glacierised alpine basins, in The Influence of Climate Change and Climatic Variability on the Hydrologic Regime and Water Resources. IAHS Publ. 168, edited by S.I. Solomon, M. Beran, and W. Hogg, pp. 77-89, International Association of Hydrological Sciences, Wallingford, UK.

Collins, D.N. 2006. Climatic variation and runoff in mountain basins with differing proportions of glacier cover. Nordic Hydrology 37: 315–326.

Demuth, M.N. and Pietroniro, A. 2003. The impact of climate change on the glaciers of the Canadian Rocky Mountain eastern slopes and implications for water resource adaptation in the Canadian prairies Phase I. Headwaters of the North Saskatchewan River Basin. Geological Survey of Canada Open File 4322, 111 pp.

Dingman, S.L. 2002. *Physical Hydrology*. 2nd Ed. Prentice Hall, New Jersey.

Dyurgerov, M. 2002.Glacier mass balance and regime: data of measurements and analysis. Institute of Arctic and Alpine Research, University of Colorado Occasional Paper No. 55.

Eaton, J.G. and Scheller, R.M. 1996. Effects of climate warming on fish thermal habitat in streams of the United States. Limnology and Oceanography 41: 1109-1115.

Environment Canada. 2002. Canadian Climate Normals Online, CDCD West CD. Accessed November 2005. http://www.climate.weatheroffice.ec.gc.ca/prods_servs/cdcd_iso_e.html

Environment Canada. 2004. Water Survey of Canada Archived Hydrometric Data Online. Accessed October 2005. http://www.wsc.ec.gc.ca/hydat/H2O/

Environment Canada. 2005. Unpublished data. Water Survey of Canada Observation Station Data, 15 Minute Intervals.

Environment Canada. 2005. Unpublished data. Water Survey of Canada Observation Station Data, 15 Minute Intervals.

Fleming, S.W. and Clarke, G.K.C. 2002. Autoregressive noise, deserialization, and trend detection and quantification in annual river discharge time series. Canadian Water Resources Journal 27: 335-354.

Fleming, S.W. and Clarke, G.K.C. 2003. Glacial control of water resource and related environmental responses to climatic warming: empirical analysis using historical streamflow data from northwestern Canada. Canadian Water Resources Journal 28: 69-86.

Fleming, S.W. and Clarke, G.K.C. 2005. Attenuation of high-frequency interannual streamflow variability by watershed glacial cover. ASCE Journal of Hydraulic Engineering 131: 615-618.

Fleming, S.W. and Quilty, E.J. 2006. Aquifer responses to el niño-southern oscillation, southwest British Columbia. Groundwater 44: 494-499.

Fleming, S.W. 2005. Comparative analysis of glacial and nival streamflow regimes with implications for lotic habitat quantity and fish species richness. River Research and Applications 21: 363-379.

Fleming, S.W., Whitfield, P.H. Moore, R.D. and Quilty, E.J. 2007. Sensitivity of distinct streamflow regimes across the Georgia-Puget transboundary ecoregion to Pacific ocean-atmosphere circulation. Hydrological Processes, Published Online: 8 May 2007, DOI: 10.1002/hyp.6544

Fountain, A.G. and Tangborn, W.V. 1985. The effect of glaciers on streamflow variations. Water Resources Research 21: 579-586.

Hagg, W., Braun, L.N., Weber, M. and Becht, M. 2006. Runoff modelling in glacierized Central Asian catchments for present-day and future climate. Nordic Hydrology 37: 93-105.

Halford, K.J., and Mayer, G.C. 2000. Problems associated with estimating ground water discharge and recharge from stream-discharge records. Ground Water 38: 331-342.

Hamilton, A.S., Hutchinson, D.G. and Moore, R.D. 2000. Estimating winter streamflow using a conceptual streamflow model. J. Cold Regions Eng. 14:158-175.

Hamilton, A.S., Hutchinson, D.G. and Moore, R.D. 2000. Estimation of winter streamflow using a conceptual streamflow model. Journal of Cold Regions Engineering 14: 158-175.

Harvey, J.W. and Bencala K.E., 1992. Stream-subsurface transport interactions; groundwater model interpretation of lumped parameters for solute storage. Eos, Transactions, American Geophysical Union. 73, 43, Suppl.

Hockey, J.B., Owens, I.F. and Tapper, N J. 1982. Empirical and theoretical models to isolate the effect of discharge on summer water temperatures in the Hurunui River. Journal of Hydrology (New Zealand) 21: 1-12.

Horton, P., Schaefli, B. Mezghani, A. Hingray, B. and Musy, A. 2006. Assessment of climate-change impacts on alpine discharge regimes with climate model uncertainty. Hydrol. Process. 20: 2091-2109. DOI: 10.1002/hyp.6197

Imbert, A. and Benestad, R.E. 2005. An improvement of analog model strategy for more reliable local climate change scenarios. Theoretical and Applied Climatology 82: 245-255.

IPCC (Intergovernmental Panel on Climate Change) 2001a. Climate Change 2001: The Scientific Basis, Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change, edited by J.T. Houghton, Y. Ding, D.J. Griggs, M. Noguer, P.J. van der Linden, X. Dai, K. Maskell, & C.A. Johnson, Cambridge University Press, Cambridge.

IPCC (Intergovernmental Panel on Climate Change) 2001b. Climate Change 2001: Impacts, Adaption and Vulnerability, Contribution of Working Group II to the Third Assessment Report of the Intergovernmental Panel on Climate Change, edited by J.J. Mc Carthy, O.F. Canziani, N.A. Leary, D.J. Dokken & K.S. White, Cambridge University Press, Cambridge.

Johanson, D. 1988. Fishtrap/Pepin/Bertrand Creeks Watershed Management Basin Plan: Groundwater Component. Province of British Columbia Ministry of Environment and Parks, Water Management Branch.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, A. Leetmaa, B. Reynolds, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K.C.

Mo, C. Ropelewski, J. Wang, R. Jenne, and D. Joseph 1996.The NCEP/NCAR 40-year Reanalysis Project, Bulletin of the American Meteorological Society 77: 437-471.

Kiffney, P.M., Bull, J.P. and Feller, M.C. 2002. Climatic and hydrologic variability in a coastal watershed of southwestern British Columbia. Journal of the American Water Resource Association 38: 1437-1451.

Krajina, V.J. 1969. Ecology of Forest Trees in British Columbia. Ecology of Western North America, 2(1): 1-146.

Kreye, R., and Wei, M. 1994. A Proposed Aquifer Classification System for Groundwater Management in British Columbia; BC Ministry of Environment, Lands and Parks, Water Management Division, Hydrology Branch, Groundwater Section, Victoria. http://www.env.gov.bc.ca/wsd/plan_protect_sustain/groundwater/aquifers/Aq_Classification/.

Lindstrom, G., Jansson, B., Persson, M, Gardelin, M. and Bergstrom, S. 1997. Development and test of the distributed HBV-96 hydrological model. J. Hydrol. 301:272-288.

Livezey, R.E. and Chen, W.Y. 1983. Statistical field significance and its determination by Monte Carlo techniques. Monthly Weather Review 111: 46-59.

Mantua, N.J., Hare, S.R., Zhang, Y., Wallace, J. M., & Francis, R.C. 1997. A Pacific interdecadal climate oscillation with impacts on salmon production. Bulletin of the American Meteorological Society 78: 1069–1079.

Mau, D.P. and Winter, T.C. 1997. Estimating ground-water recharge from streamflow hydrographs for a small mountain watershed in a temperate humid climate, New Hampshire, USA. Ground Water 35: 291-304.

McCuen, R.H. 2002. Modeling Hydrologic Change: Statistical Methods. Lewis Publishers, Boca Raton.

Ministry of the Environment (MOE). 2006. Observation Well Network. Water Stewardship Division, Government of British Columbia, Victoria.

Mitchell, R.J., Babcock, R.S., Gelinas, S., Nanus, L., and Stasner. D.E. 2003. Ground water quality: nitrate distribution and source identification in the Abbotsford-Sumas Aquifer, northwestern Washington State. Journal of Environmental Quality 32: 789-800.

Mokievsky-Zubok, O., Ommanney, C.S.L. and Power, J. 1985.NHRI glacier mass balance 1964-1984 (Cordillera and Arctic). Glacier Section, Surface Water Division, National Hydrology Research Institute Internal Report.

Moore, R.D. 1992. Hydrological responses to climatic variations in a glacierized watershed: inferences from a conceptual streamflow model, in G. Kite and D. Harvey (eds..Using Hydrometric Data to Detect and Monitor Climatic Change. National Hydrology Research Institute, Saskatoon, p. 79-88.

Moore, R.D. 1993. Application of a conceptual streamflow model in a glacierized drainage basin. J. Hydrol. 150:151-168.

Moore, R.D. 2006. Stream temperature patterns in British Columbia, Canada, based on routine spot measurements. Canadian Water Resources Journal 31: 41-56.

Moore, R.D. and Demuth, M.N. 2001. Mass balance and streamflow variability at Place Glacier, Canada, in relation to recent climate fluctuations. Hydrological Processes 15: 3473-3486.

Moore, R.D. and Wondzell, S.M. 2005. Physical hydrology and the effects of forest harvesting in the Pacific Northwest: a review. Journal of the American Water Resources Association 41: 753-784.

Nash, J.E. and Sutcliffe, J.V. 1970. River flow forecasting through conceptional models, Part I: a discussion of principles. J. Hyrol. 10: 282-290.

Oerlemans, J. 2005. Extracting a climate signal from 169 glacier records. Science 308 5722: 675 – 677 DOI: 10.1126/science.1107046

Oerlemans, J., Anderson, B., Hubbard, A., Huybrechts, P., Johannesson, T., Knap, W.H., Schmeits, M., Stroeven, A.P., van de Wal, S.W., Wallinga, J. and Zuo, Z. 1998. Modelling the response of glaciers to climate warming. Climate Dynamics 14: 267–274.

Osborn, G. and Luckman, B.H. 1988. Holocene glacier fluctuations in the Canadian Cordillera (Alberta and British Columbia. Quaternary Science Reviews 7: 115-128.

Østrem, G. and Brugman, M.M. 1991 Glacier Mass-Balance Measurements: A manual for field and office work. National Hydrology Research Institute Science Report No. 4, Minister of Suppy and Services, Ottawa.

Pearson, M. P. 2004. The Ecology, Status and Recovery Prospects of Nooksack Dace (Rhinichthys cataractae spp.) and Salish Sucker (Catostomus sp.) in Canada. PhD Thesis. Resource Management and Environmental Studies, University of British Columbia.

Piggott, A., D. Brown, B. Mills and S. Moin, 2001 Exploring the dynamics of groundwater and climate interaction, in Proceedings of the 54th Canadian Geotechnical and 2nd Joint IAH-CNC and CGS Groundwater Specialty Conferences, pp. 401-408, Canadian Geotechnical Society and the Canadian National Chapter of the International Association of Hydrogeologists.

R Development Core Team, 2006. R: A language and environment for statistical computing. R Foundation for Statistical Computing, Vienna, Austria. ISBN 3-900051-07-0, http://www.R-project.org.

Radic, V. and Hock, R. 2006. Modeling future glacier mass balance and volume changes using ERA-40 reanalysis and climate models; A sensitivity study at Storglaciären, Sweden. Journal of Geophysical Research, 111 DOI:10.1029/2005JF000440

Redmond, K.T. and Koch, R.W. 1991. Surface climate and streamflow variability in the western United States and their relationship to large-scale circulation indices. Water Resources Research 27: 2381-2399.

Rees, H.G. and Collins, D.N. 2006. Regional differences in response of flow in glacier-fed Himalayan rivers to climatic warming. Hydrol. Process. 20: 2157–2169. DOI: 0.1002/hyp.6209

Regonda, S.K., Rajagopalan, B., Clark, M. and Pitlick, J. 2005. Seasonal cycle shifts in hydroclimatology over the western United States. Journal of Climate 18: 372-384.

Rutledge, A.T. and Daniel, C.C. III, 1994. Testing an automated method to estimate ground-water recharge from streamflow records. Ground Water 32: 180-189.

Schwarz, G. 1978. Estimating the dimension of a model. Annals of Statistics 6: 461-464.

Scibek, J. and Allen, D. 2005. Numerical Groundwater Flow Model of the Abbotsford-Sumas Aquifer, Central Fraser Lowland of BC, Canada, and Washington State, US. Final Report to Environment Canada.

Scibek, J. and Allen, D.M. 2006a. Modeled impacts of predicted climate change on recharge and groundwater levels. Water Resources Research, 42, doi:10.1029/2005WR004742

Scibek, J. and Allen, D.M. 2006b. Comparing the responses of two high permeability, unconfined aquifers to predicted climate change. Global and Planetary Change 50: 50-62.

Scibek, J., Allen, D.M., Cannon A.and Whitfield, P.H. 2006. Groundwater-surface water interaction under scenarios of climate change using a high-resolution transient groundwater model. Journal of Hydrology, doi:10.1016/j.jhydrol.2006.08.005

Semenov, M.A., Brooks, R.J., Barrow, E.M. and Richardson, C.W. 1998. Comparison of the WGEN and LARS-WG stochastic weather generators in diverse climates. Climate Research 10: 95-107.

Shabbar, A. Bonsal, B. and Khandekar, M. 1997. Canadian precipitation patterns associated with the Southern Oscillation. Journal of Climate 10: 3016-3027.

Singh, P. and Bengtsson, L. 2005. Impact of warmer climate on melt and evaporation for the rainfed, snowfed and glacierfed basins in the Himalayan region. J. Hydrol. 300: 140-154.

Singh, P. and Kumar, N. 1997. Impact assessment of climate change on the hydrological response of a snow and glacier melt runoff-dominated Himalayan river. Journal of Hydrology 193: 316–350.

Smakhtin, V.U. 2001. Low flow hydrology: a review. Journal of Hydrology 240: 147-186.

Stahl, K. and Moore, R.D. 2006.Influence of watershed glacier coverage on summer streamflow in British Columbia, Canada. Water Resources Research, 42, W06201, doi:10.1029/2006WR005022.

Stahl, K., Moore, R.D. and McKendry, I.G. 2005. The role of synoptic-scale circulation in the linkage between large-scale ocean-atmosphere indices and winter surface climate in British Columbia, Canada. International Journal of Climatology DOI: 10.1002/joc.1268

Toews, M.W., Whitfield, P.H. and Allen, D.M. 2007. Seasonal statistics: the seas package for R. Computers & Geosciences. doi:10.1016/j.cageo.2006.11.011

Van de Wal, R.S.W. and Wild, M. 2001. Modelling the response of glaciers to climate change by applying volume-area scaling in combination with a high resolution GCM. Clim. Dyn. 18: 359–366. DOI: 10.1007/s003820100184

Wang, J.Y., Whitfield, P.H. and Cannon, A.J. 2006. Influence of Pacific climate patterns on low-flows in British Columbia and Yukon, Canada. Canadian Water Resources Journal 31: 1-16.

Water Survey of Canada, 2006. Archived Hydrometric Data

Webb, B. W., Clack, P.D. and Walling, D.E. 2003. Water-air temperature relationships in a Devon river system and the role of flow. Hydrological Processes 17: 3069-3084.

Wei, M., Kohut, A.P., Ronneseth, K., Allen, D.M., Turner, B. and Grasby, S. (in prep). Types and general characteristics of aquifers in the Canadian Cordillera Hydrogeologic Region. For submission to Joint CGS-IAH conference, Ottawa, October, 2007.

Wernick, B. G., Cook, K. E., and Schreier, H. 1998. Land use and streamwater Nitrate-N dynamics in an rrban-rural fringe watershed. Journal of the American Water Resources Association 34: 639-650.

Whitfield, P.H. 2001. Linked hydrologic and climate variations in British Columbia and Yukon. Environmental Monitoring and Assessment 67: 217-238.

Whitfield, P.H. and A.J. Cannon, 2000. Recent variations in climate and hydrology in Canada. Canadian Water Resource Journal 25: 19-65.

Whitfield, P.H. 2001. Linked hydrologic and climate variations in British Columbia and the Yukon. Environmental Monitoring and Assessment 67: 217-238.

Whitfield, P.H., and Taylor, E. 1998. Apparent recent changes in hydrology and climate of coastal British Columbia. In: Alila, Y. ed., Mountains to Sea: Human interactions with the Hydrologic Cycle. Proceedings of the 51st Canadian Water Resources Association Conference. 22-29.

Whitfield, P.H., Pilon, P.J., Burn, D.H., Arora, V., Lins, H.F., Ouarda, T., Sellers, C.D., and Spence, C. 2003. Chapter 11: Water availability and climate variability and change –rivers and streams. In: Threats to Water Availability in Canada. Victoria, National Water Research Institute and Meteorological Service of Canada: 85-90.

Whitfield, P.H., Wang, J.Y. and Cannon, A.J. 2003. Modelling Future Streamflows in Present and Future Climates – Examples From Rainfall/Snowmelt Streams in Coastal British Columbia, Canada. Hydrological Science and Technology 2002 AIH Annual Meeting, Hydrological Extremes: Challenges for Science and Management, October 13-17, 2002, Portland Oregon American Institute of Hydrology. Vol. 19, No. 1-4.

Whitfield, P.H., Wang, J.Y. and Cannon, A.J. 2003. Modelling future streamflow extremes — floods and low flows in Georgia Basin, British Columbia. Canadian Water Resources Journal 28: 633-656.

Winter, T.C., Harvey J.W., Franke O.L., Alley W.M., 1998. Groundwater and surface water: a single resource. US Geol Surv Circ 1139.

WMO: World Meteorological Organization, 1974. “International Glossary of Hydrology”. WMO, Geneva.

Woessner, W.W. 2000. Stream and fluvial plain ground water interactions; rescaling hydrogeologic thought. Ground Water 38: 423-429.

Wroblicky, G.J, Campana M.E., Valett H.M. and Dahm C.N. 1998. Seasonal variation in surface-subsurface water exchange and lateral hyporheic area of two stream-aquifer systems. Water Resources Research 34: 317-328.

WWF 2005. An overview of glaciers, glacier retreat, and subsequent impacts in Nepal, India and China, World Wildlife Fund, Nepal Program.

Yue, S., Pilon, P. and Cavadias, G. 2002. Power of the Mann-Kendall and Spearman’s rho test for detecting monotonic trends in hydrological series. Journal of Hydrology 259: 254-271.

Zebarth, B.J., Hii, B., Liebcher, H., Chipperfield, K., Paul, J.W., Grove, G., and Szeto, S.Y. 1998. Agricultural land use practices and Nitrate contamination in the Abbotsford Aquifer, British Columbia, Canada. Agriculture, Ecosystems and Environment 69: 99-112.

Appendix 1 Estimation of end of month groundwater levels from daily data

A1.1 Algorithm

The following steps were followed to estimate end-of-month water levels from daily records:

- find first year and last year
- loop over each year and month
- find how many gw level values exist between 15th of month and 14th of next month (exclude missing)
- if 1 or 2 are found, use the one closest to months' end (some records have readings twice per month, and date of reading varies a lot, therefore the wide search)
- if more than 2 are found (daily data) then calculate the median of all values between the 28th of that month5 days and the 28th +7 days (this seems the most frequent range of end of month readings and the median protects the estimate from outliers)
- write all "end of month" data into a file named according to the well number

Figures A1.1 to A1.9 provide comparisons between end-of-month data as recorded in official Ministry of Environment data files and end-of-month water levels estimated from the 9 digitized daily series. They show that the main type of error that sometimes occurred was a slight smoothing of the highest values due to the relatively long period used to calculate the median (e.g., Well #002). Various period lengths were tested, and the chosen window appears to be the best compromise reflecting the variation in the real end of month readings better than a shorter period. Other errors with this approach mostly occur where there are true errors in the daily data, i.e. values that are completely out of range. To avoid the inclusion of such values, while looping through the list of observation wells the script plots each series,and allows to interactively set an upper and lower threshold above and below which values are then considered missing values for the end of month estimate.

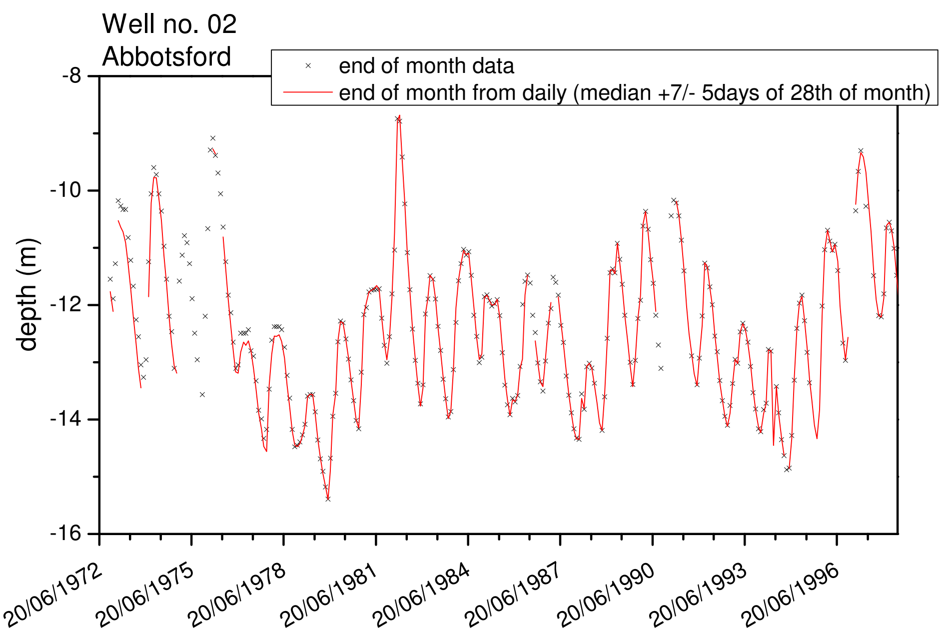


Figure A1.1: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 002, Abbotsford.

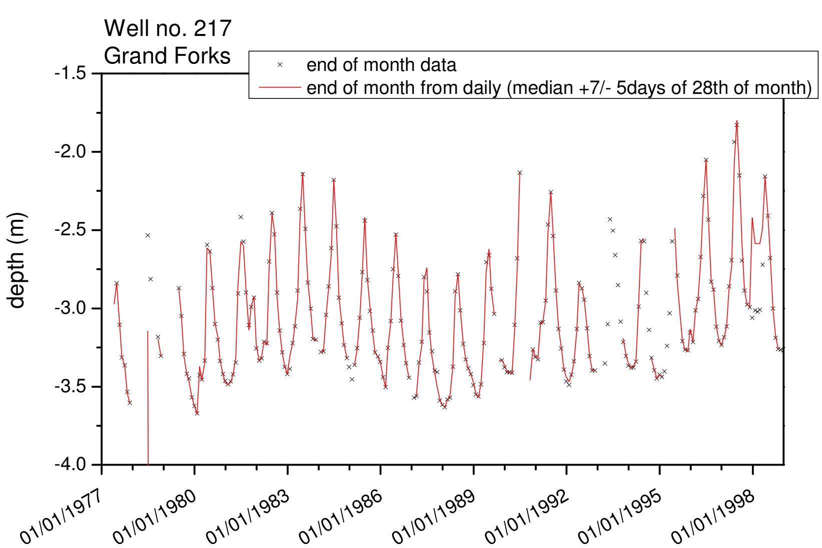


Figure A1.2: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 217, Grand Forks.

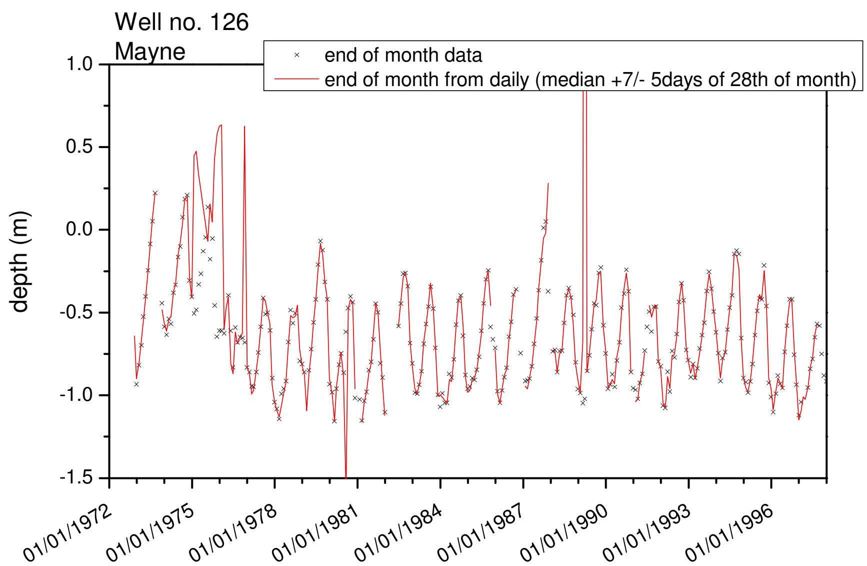


Figure A1.3: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 126, Mayne Island.

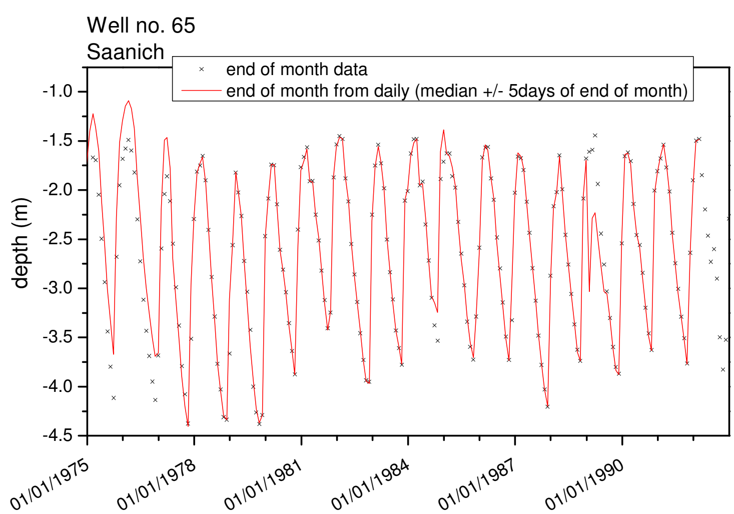


Figure A1.4: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 065, Saanich.

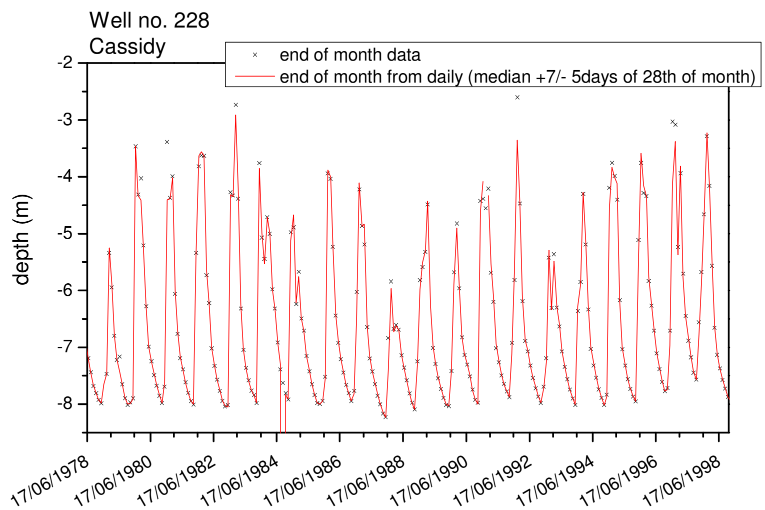


Figure A1.5: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 228, Cassidy.

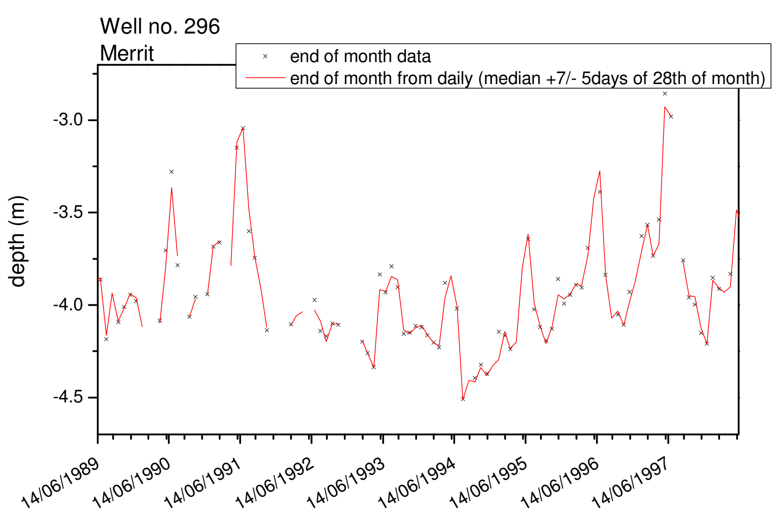


Figure A1.6: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 296, Merritt.

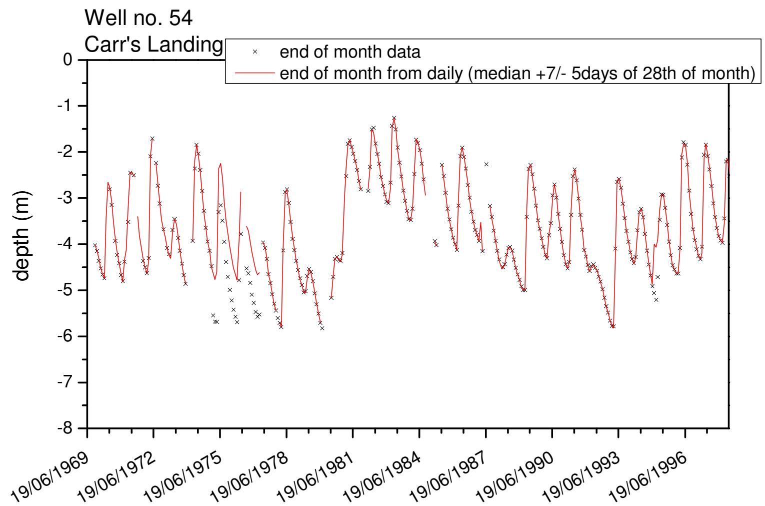


Figure A1.7: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 054, Carr's Landing.

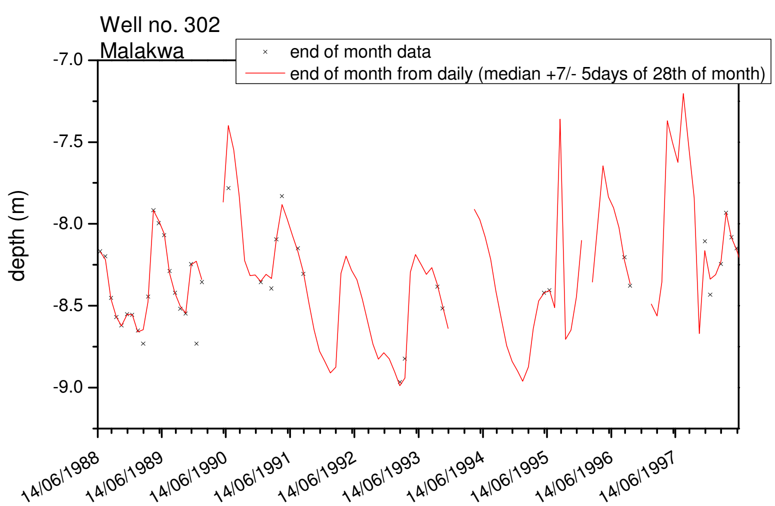


Figure A1.8: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 126, Mayne Island.

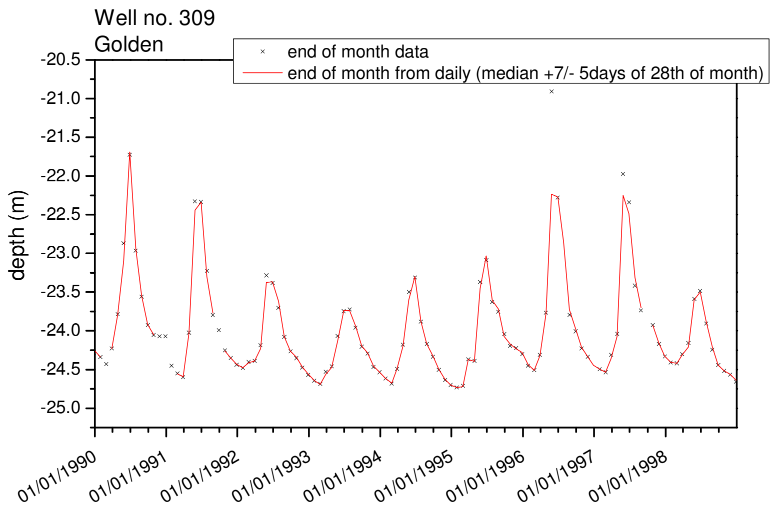


Figure A1.9: Comparison of end-of-month data to calculated end-of-month data from daily data for Observation well 309, Golden.

A1.2 Final dataset

Observation wells metadata (note that where no information is given in the last three columns, the entire series is end-of-month data; i.e. they don't change to daily as do the others)

Well	Location	Well Tag Number	Start	End	Missing (%)	Mean (end of month)	Mean (daily)	Signifi-cance (t-test)
Recharge-dominated								
2	Fraser Valley	26787	1965	2006	11.6	-12.25	-12.22	
14	Fraser Valley	24413	1972	1999	14.2	-12.76		
15	Fraser Valley	24415	1972	2006	10.8	-11.01	-11.21	
60	Saanich	20143	1966	2006	7.5	-9.88	-9.84	
64	Saanich	25936	1974	1998	9.4	-2.46		
65	Saanich	25891	1975	2006	3.2	-2.51	-2.50	
69	Saanich	32624	1975	1998	6.2	-6.33		
126	Mayne Island	26713	1972	2006	7.6	-0.66	-0.87	x
127	Mayne Island	29606	1972	1996	8.7	-0.99		
128	Mayne Island	24845	1972	2006	6.1	-1.27	-1.07	
228	Cassidy	3155	1978	2005	1.5	-6.51	-6.50	
258	Galiano Island	44593	1980	2006	15.1	-10.86	-11.95	x
211	Duncan	33651	1976	2006	10.6	-2.54	-3.23	x
268	Denman Island	62734	1981	2006	15.3	-0.15	-0.15	
Snowmelt-dominated								
35	Stump Lake	20598	1968	2006	16.0	-3.50	-3.75	
45	Westwold	19522	1965	2005	6.3	-8.97		
47	Silverstar U	49951	1966	2004	14.0	-6.75		
53	Carrs Landing	19772	1966	2004	13.2	-3.06		
54	Carrs Landing	22692	1969	2004	9.8	-3.95		
81	83 Mile	20697	1967	2006	15.4	-5.42	-5.25	
96	Osoyoos	22769	1969	2004	17.1	-3.89		
100	Osoyoos	22733	1969	2004	16.4	-3.77		
105	Osoyoos	22702	1969	2004	16.2	-2.98		

Well	Location	Well Tag Number	Start	End	Missing (%)	Mean (end of month)	Mean (daily)	Significance
107	Osoyoos	22706	1969	2004	15.7	-3.03		
115	Mission Creek	23980	1973	2004	19.1	-5.17		
117	Armstrong	20648	1971	2006	7.4	-16.28	-16.66	x
118	Armstrong	24080	1971	2006	8.3	-13.74	-14.04	
119	Armstrong	24104	1971	2003	13.3	-31.15		
153	Summerland	22796	1969	2004	18.1	-1.33		
154	Summerland	22792	1969	2004	18.6	-3.60		
162	Kalawoods	26248	1972	2006	5.4	-1.58	-2.16	x
172	Kalawoods	8596	1977	2006	6.0	-15.48	-15.67	
173	Kalawoods	8374	1972	2004	6.3	-13.90		
174	Kalawoods	8608	1972	2004	17.7	-36.18		
175	Kalawoods	8623	1972	2004	13.8	-20.74		
176	Kalawoods	8533	1972	2004	12.0	-39.08		
185	Salmon River	30438	1974	2006	16.1	-4.23	-4.54	x
217	Grand Forks	14947	1977	2006	13.2	-3.08	-3.05	
264	Mt.Kobau	62733	1981	2000	10.1	-30.31		
296	Merritt	56918	1989	2005	12.0	-4.03	-4.38	x
309	Golden	59115	1989	2005	9.4	-24.11	-24.06	

A1.3 Groundwater Trends

Monthly groundwater trends (r_s values). Red and blue trends are significant for 1976-2002.

Well name	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Recharge
rain-dominated													
2 Fraser Valley	-0.20	-0.24	-0.33	-0.41	-0.38	-0.46	-0.42	-0.44	-0.46	-0.52	-0.42	-0.26	-0.09
15 Fraser Valley	-0.54	-0.62	-0.65	-0.61	-0.60	-0.54	-0.57	-0.62	-0.63	-0.52	-0.43	-0.41	0.29
60 Saanich	-0.04	-0.05	-0.16	-0.30	-0.31	-0.38	-0.45	-0.50	-0.59	-0.51	-0.34	-0.31	0.03
65 Saanich	-0.61	-0.21	-0.26	-0.34	-0.32	-0.39	-0.50	-0.50	-0.59	-0.58	-0.25	-0.31	-0.54
126 Mayne Island	-0.57	-0.25	-0.40	-0.35	-0.35	-0.24	-0.42	-0.27	-0.06	-0.21	-0.27	-0.23	-0.60
128 Mayne Island	0.13	-0.39	-0.22	-0.23	-0.38	-0.27	-0.01	0.05	-0.07	0.09	0.05	-0.17	0.00
211 Duncan	-0.57	-0.70	-0.61	-0.81	-0.78	-0.84	-0.84	-0.85	-0.88	-0.77	-0.61	-0.64	0.01
258 Galiano Island	0.28	-0.35	-0.18	-0.33	-0.60	-0.82	-0.76	-0.80	-0.70	-0.44	-0.12	0.00	0.77
268 Denman Isl.	-0.35	-0.60	-0.40	-0.39	-0.34	-0.48	-0.36	-0.14	-0.08	0.07	-0.47	-0.27	-0.02
snowmelt-dominated													
228 Cassidy	0.38	-0.19	0.15	0.00	0.21	0.11	0.26	0.34	0.34	0.41	0.16	0.06	-0.30
35 Stump Lake	-0.26	-0.21	-0.59	-0.41	-0.32	-0.16	-0.22	-0.12	-0.23	-0.26	-0.09	-0.24	-0.60
45 Westwold	-0.08	0.07	0.08	0.39	-0.01	-0.62	-0.78	-0.76	-0.72	-0.50	-0.49	-0.23	-0.75
47 Silverstar U	0.13	-0.01	-0.01	-0.09	0.11	-0.07	0.04	0.22	0.14	-0.10	0.20	0.03	-0.01
53 Carrs Landing	-0.20	-0.33	-0.19	-0.46	-0.45	-0.29	-0.15	-0.15	-0.19	-0.14	-0.10	-0.19	-0.04
54 Carrs Landing	-0.07	-0.14	-0.20	-0.30	-0.26	-0.26	-0.25	-0.19	-0.18	-0.15	-0.19	-0.18	0.00
81 83 Mile	0.40	0.28	0.51	0.39	0.39	0.27	0.39	0.35	0.27	0.35	0.30	0.37	0.60
96 Osoyoos	0.46	0.24	0.01	0.22	0.08	-0.55	-0.43	-0.73	-0.46	-0.20	-0.25	-0.24	-0.36
100 Osoyoos	-0.15	-0.37	-0.52	-0.09	-0.21	-0.13	-0.20	-0.58	-0.18	0.01	-0.14	0.03	na
105 Osoyoos	0.02	-0.30	-0.23	-0.05	-0.28	-0.23	-0.50	-0.40	-0.10	-0.21	-0.39	-0.54	0.17
107 Osoyoos	0.26	-0.08	0.47	0.49	-0.17	0.01	-0.07	-0.38	0.62	0.68	0.69	0.41	-0.47
115 Mission Creek	0.14	0.25	0.23	0.46	0.23	0.18	0.44	0.57	0.19	0.24	0.18	0.15	0.05
117 Armstrong	-0.61	-0.61	-0.62	-0.62	-0.63	-0.73	-0.78	-0.82	-0.77	-0.71	-0.73	-0.69	0.64
118 Armstrong	-0.14	-0.12	-0.09	-0.07	-0.06	-0.07	-0.11	-0.13	-0.18	-0.12	-0.18	-0.14	0.20
119 Armstrong	-0.83	-0.84	-0.83	-0.85	-0.89	-0.86	-0.87	-0.90	-0.89	-0.89	-0.89	-0.85	0.71
153 Summerland	-0.70	-0.69	-0.57	-0.70	-0.47	-0.57	-0.18	-0.28	-0.44	-0.46	-0.55	-0.60	0.44
154 Summerland	-0.88	-0.72	-0.81	-0.81	-0.80	-0.89	-0.76	-0.80	-0.87	-0.86	-0.80	-0.82	na
162 Kalawoods	0.06	-0.36	-0.15	-0.13	-0.42	-0.50	-0.46	-0.71	-0.42	-0.47	-0.32	-0.39	-0.62
172 Kalawoods	0.07	-0.11	-0.16	0.05	-0.23	-0.07	0.03	0.07	0.26	0.03	0.17	-0.08	-0.10
173 Kalawoods	-0.74	-0.62	-0.80	-0.33	-0.15	-0.25	-0.13	-0.10	-0.49	-0.55	-0.43	-0.68	0.37
174 Kalawoods	-0.75	-0.68	-0.67	-0.36	-0.27	-0.20	-0.11	0.00	-0.33	-0.43	-0.49	-0.69	0.25
175 Kalawoods	0.41	0.61	0.43	0.68	0.48	0.57	0.44	0.72	0.70	0.60	0.49	0.53	-0.23
176 Kalawoods	-0.61	-0.54	-0.59	-0.31	-0.20	-0.18	-0.20	-0.20	-0.36	-0.52	-0.64	-0.58	-0.10
185 Salmon River	-0.57	-0.51	-0.50	-0.50	-0.45	-0.48	-0.50	-0.37	-0.56	-0.61	-0.53	-0.65	0.35
217 Grand Forks	0.50	0.44	0.27	0.41	0.22	0.22	0.20	0.09	0.08	0.08	0.29	0.42	0.15
264 Mt.Kobau	-0.28	-0.28	-0.12	0.03	-0.09	-0.10	-0.20	-0.22	-0.23	-0.26	-0.33	-0.28	0.06

296 Merritt	-0.68	-0.68	-0.42	-0.41	-0.26	-0.35	-0.29	-0.73	-0.69	-0.67	-0.65	-0.78	0.88
309 Golden	-0.11	-0.14	-0.18	-0.42	-0.62	-0.44	-0.30	-0.39	-0.33	-0.13	-0.28	-0.30	-0.36

Appendix 2 Case Study: Abbotsford-Sumas Aquifer

A2.1 Introduction

This case study summarizes the results of a field investigation in which the magnitude of the variation in discharge over the low flow season (June to September 2005) was measured in three streams (Bertrand Creek, Pepin Brook and Fishtrap Creek) to quantify variability along the length of the stream and relate those variations to differing groundwater contributions over the low flow season. Water quality data were also collected to comment on the temporal variation in contribution from groundwater. The case study comprises a portion of a paper in Canadian Water Resources Journal (Berg and Allen, in press), which also includes the low flow measurement field study discussed in Section 2.0 of this report.

The streams drain the Abbotsford aquifer in southwest BC. This important trans-national aquifer lies within the Lower Fraser Valley in southwest BC, Canada and northwest Washington State, USA. The aquifer is approximately 160 km2 and is roughly bisected by the Canada-USA border. The three streams (Figure A2.1) flow south to the Nooksack River in Washington State. During the low flow period (late summer and early fall), the streams are largely maintained by contributions from groundwater.

A2.2 Geology of the study area

The surficial geology of the Abbotsford aquifer consists of Quaternary age glacial sediments, underlain by Tertiary sedimentary rock. The low permeability Tertiary sequences are effectively aquitards with limited groundwater flow (Scibek and Allen, 2005). The most extensive geologic units identified in the study area are the Fort Langley Formation and the Sumas Drift, with lesser amounts of Salish Sediments. The Fort Langley Formation consists of glaciomarine unsorted pebbly silt in a clay layer, which has been interpreted as a being continuous at depth, and forming a confining layer (Mitchell et al., 2003). Sumas Drift includes glaciofluvial sand and gravel deposits, with intermittent lenses of till. This unit becomes coarser in the Abbotsford area, and encompasses what is called the Abbotsford outwash. The sand and gravel deposits of the Sumas Drift, including the Abbotsford outwash, are the units that contain the Abbotsford-Sumas aquifer and, in the area of Abbotsford, overlie the Fort Langley Formation. The Salish Sediments also occur intermittently in the area of these streams, and comprise fluvial, lacustrine, and slope sediments (Scibek and Allen, 1995; Johanson, 1988).

The Abbotsford aquifer is predominantly unconfined. The thickness of the aquifer is generally 15-25 m thick, but ranges from 5 to 70 m in thickness. Soils over the aquifer are thin (generally less than 70 cm thick), with permeability that is greater than average precipitation rates (Mitchell et al., 2003). The hydraulic conductivity of the aquifer has been estimated to have a mean value of 80 m/day, but ranges from 2 to 2380 m/day. Thus, the aquifer is highly productive. The mean lateral velocity of the groundwater has been estimated at 0.75m/day (Mitchell et al., 2003). Groundwater flow is generally to the southwest, but is strongly controlled by local topographic variations, as well as by stream interactions.

Groundwater extraction within the aquifer is generally limited to a few large capacity production wells, with a cumulative pumping rate of roughly 77,000 m3/day. None of these large capacity production wells are in close proximity to Fishtrap Creek, Pepin Brook or Bertrand Creek and, thus, are not likely to influence the flow in these streams due to the high productivity of the aquifer. Domestic wells are few in number and generally operate intermittently and at low capacity.

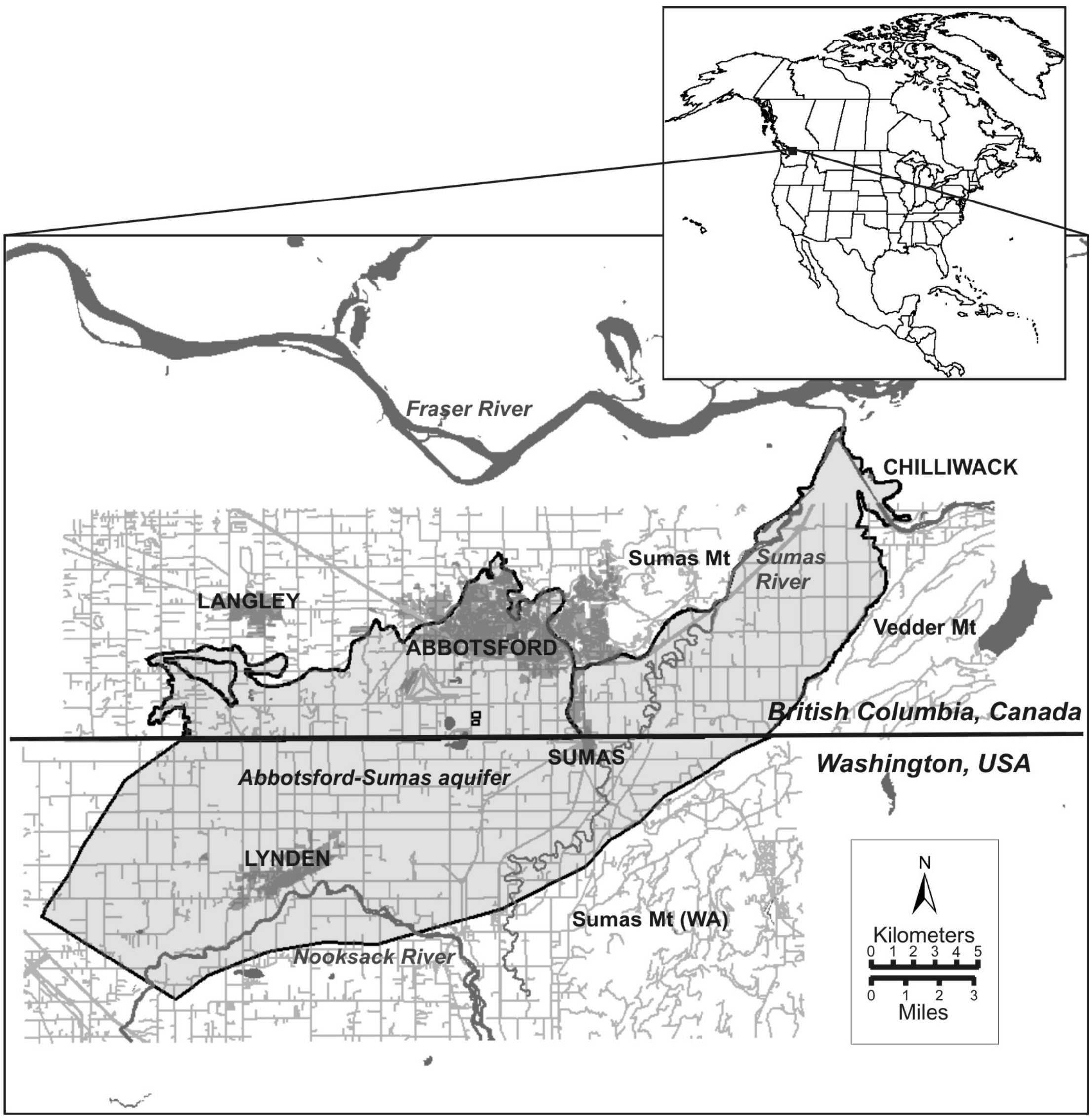


Figure A2.1: Location of the Abbotsford-Sumas aquifer in the Lower Fraser Valley. The gray zone indicates the extent of the aquifer.(From Berg and Allen, in press)

A2.3 Climate

The climate of the Abbotsford area is dominated by moderate temperatures and high annual precipitation. The ocean moderates the temperatures throughout the year, and during the November to May period, the prevailing westerly winds bring precipitation-laden clouds inland. The annual average precipitation is 1500 mm/yr, with less than 100 mm as snow. Approximately 70% of precipitation falls between October and May (Figure A2.2) (Berka et al., 2001; Zebarth et al., 1998; Environment Canada, 2002). Only approximately 6% of the precipitation falls in July and August (Wernick et al., 1998).

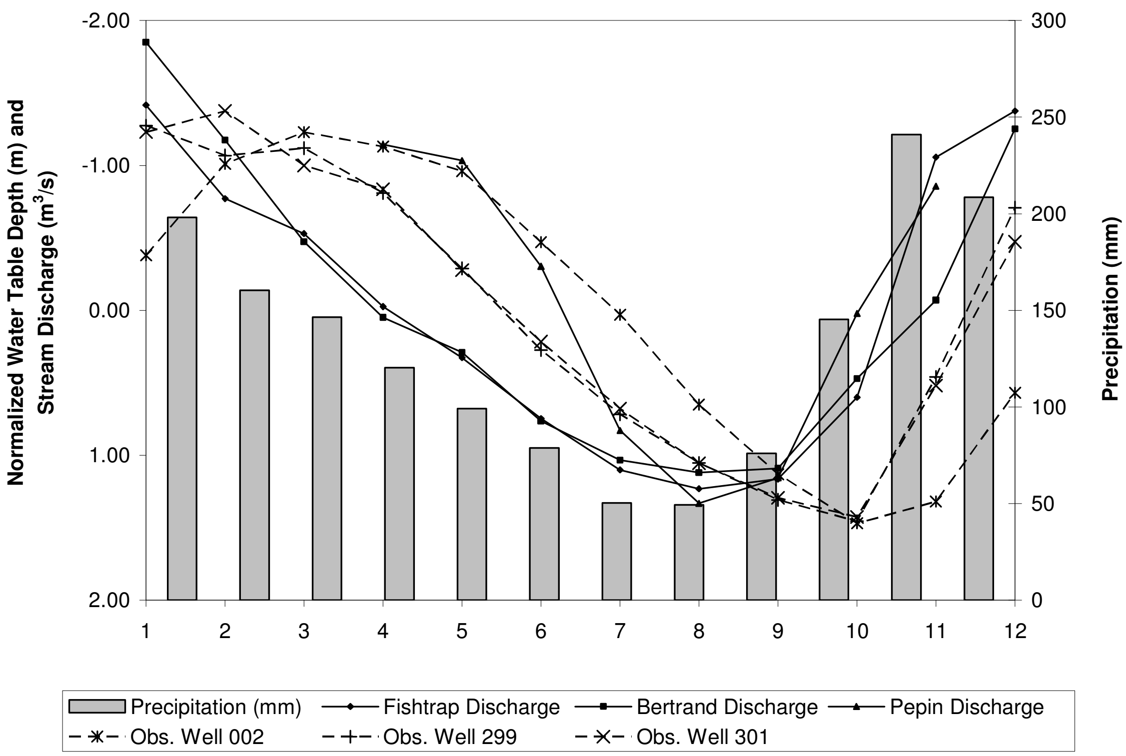


Figure A2.2: Mean monthly discharge for Fishtrap, Pepin and Bertrand Creeks, and mean monthly groundwater levels for the Province of BC observation wells 002 , 299 and 301 (see Figure 3 for location). Mean monthly precipitation at Abbotsford International Airport is also shown. The water levels and the discharge follow the pattern of the precipitation, but exhibit a lag period due to the time required for the precipitation to percolate down into the aquifer. Mean values were calculated over the period of record (1988-2002). (From Berg and Allen, in press)

A2.4 Hydrology

Fishtrap Creek, Bertrand Creek, and Pepin Brook originate at low elevation (slightly above mean sea level), and the flow regime is driven by precipitation and the interaction with the groundwater. All respond similarly to precipitation as the primary driving force (Figure A2.2). The flow regime follows the timing of precipitation, with a time lag of approximately one month. Peak flow occurs between October and May, corresponding to the period of highest precipitation (Environment Canada, 2002; 2004). Minimum precipitation occurs in July and August, and the lowest stream flows, or even dry conditions, occur during August, shortly after the minimum precipitation. At this time, the streams become susceptible to drought conditions.

Recharge to the aquifer is predominantly by precipitation (Scibek and Allen, 2005). Water levels in the aquifer fluctuate seasonally (see Figure A2.2 for three provincial observation wells), as well as annually, due to variations in precipitation and recharge. The average annual variation in the aquifer levels is 2 m, with a maximum variation of approximately 3m (Scibek and Allen, 2005). There is a time lag of roughly three months between the minimum precipitation and the minimum water table level.

The streams have varied interactions with groundwater as evidenced by complex flow paths simulated with a regional groundwater flow model (Scibek and Allen, 2005). Variations in aquifer and streambed sediment type, local groundwater gradients, and stream geometry are likely the main factors controlling these interactions (see Woessner, 2000). While it has been suggested that groundwater discharges into the streams and maintains the baseflow (Johanson, 1988; Pearson, 2004), a detailed account of the nature of these interactions is currently lacking. Average monthly groundwater levels lag stream discharge by approximately 1.5 months (Figure A2.2). During this period, the discharge is presumed to be sustained by baseflow as there is no other major source of water to the streams. Stream discharge reaches a minimum in August, and the minimum groundwater level is evident two months later.

A2.5 Regional site descriptions

The measurement sites along Pepin Brook, Bertrand Creek, and Fishtrap Creek were selected primarily on the basis of accessibility and spatial coverage. Limitations to site selection included access restrictions (private properties) or hazards (steep embankments). Regional measurement sites span the Langley Uplands, south to the Canadian border, where the topography generally levels off. The sites were spaced along each stream so as to facilitate a general characterization of varying groundwater contributions to discharge along each stream. Water quality measurements were made at a total of thirteen sites (see Figure A2.3 for site locations). Due to difficulties in accessing some sites, discharge measurements were made at only 8 of these sites: F2, F3, P1, P2, P4, B1, B2, and B3. The low flow repeatability sites, F2 and B2, are included among the regional flow sites.

A2.5.1 Fishtrap Creek

Discharge was measured at two locations along Fishtrap Creek. At the lower reach site (F2), the creek flows through an agricultural area dominated by berry production. This lower section of Fishtrap Creek had a deep (<70 cm water depth in June) and moderately wide channel (5m) of fairly uniform shape, which may be due, in part, to dredging in 1990 (Pearson, 2004). Low flow repeatability measurements were also taken at site F2. The upper reach site (F3) is situated in the Langley Uplands, downstream of a residential area. In June, the water depth was <40cm and the channel width was 4.6m. The tributary stream, Waechter Creek (W1), enters Fishtrap Creek between these two sites (W1 was a water quality sampling site only).

A2.5.2 Bertrand Creek

Bertrand Creek has the longest length of channel of the three streams; measurements were taken at three sites. The first site (B1) consists of a straight section of channel, with dominantly sand bed material, and some angular boulders along the right bank. In June, the water depth was <60cm, and the channel width was 5m. The second site (B2) has lateral sand and gravel bars, and well-developed riffle-pool morphology (water depth in June <50 cm; width 5m). The channel here is shallow and wide, and dominated by fine gravel. The third site (B3) is a deep pool section of the stream (water depth in June <80cm, width 6m). The sediments are dominantly fines, with some sand, and the banks were observed to be composed of clay in the exposed portions.

A2.5.3 Pepin Brook

Pepin Brook had three sites; two in the lower reaches in the outwash plain, and one in higher reaches in the Langley Uplands. The site in the lowest reach (P1) was at the Water Survey of Canada gauging station where the stream flows through a section of pasture. In June, the stream was narrow (3m) and deep (<60cm), with dominantly sand and fine gravel sediments. There is a large amount of in-stream vegetation that completely covers the surface of the stream beyond the area cleared for flow measurements. The second site (P2) is within the Aldergrove Lake Regional Park, through which a large section of the Canadian portion of Pepin Brook flows. The stream in this section has a riffle-pool morphology, and a shallow (<30cm in June), wide channel (6.3m). The sediments are dominantly gravel with sand, and there was no in-stream vegetation at this location. The third discharge site (P4) was a narrow channel (3.5m) with <30cm water depth. This site was strongly controlled by beaver activity. The sediments consist mainly of sands and fines. The stream at this location flows through the Langley Uplands, and is smaller and more confined than in the lower reaches.

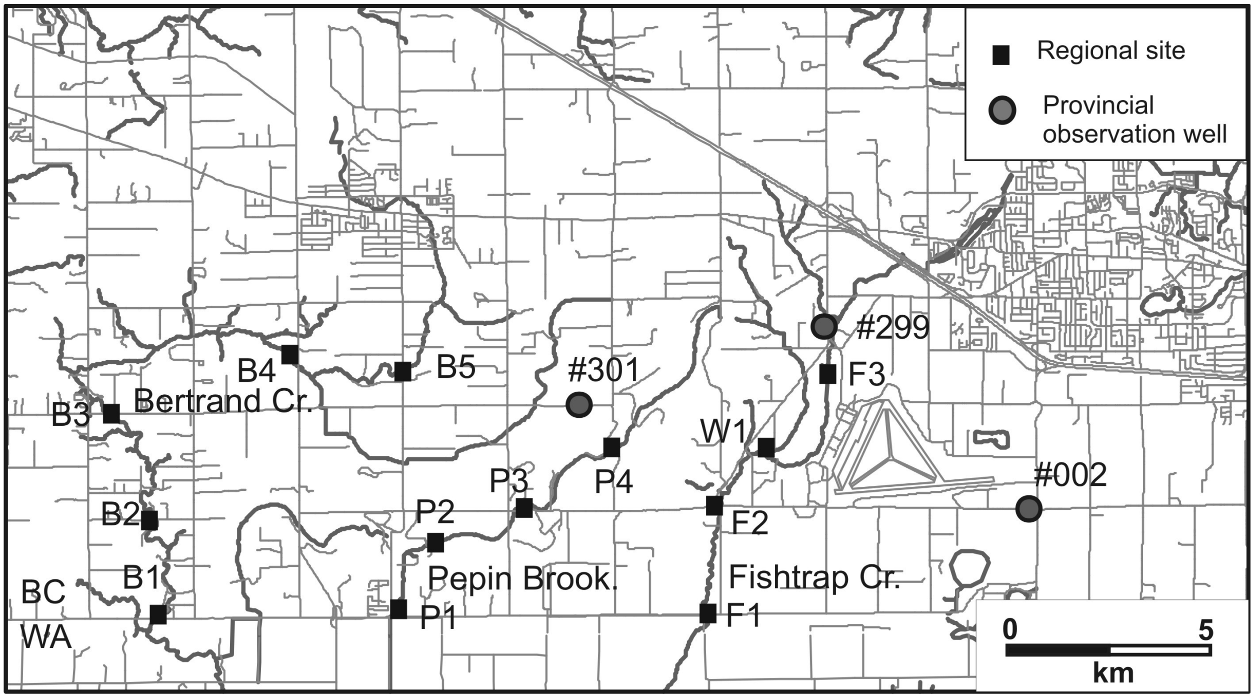


Figure A2.3: Location of the sampling sites for the regional flow and chemistry sampling. The low flow repeatability sites are F2 and B2. Provincial observation wells situated in the Abbotsford study area include 002 (18 m deep), 299 (16 m deep) and 301 (41 m deep). All are completed in sand and gravel. (From Berg and Allen, in press)

A2.6. Field measurements and analysis

Discharge measurements were conducted at approximately one month intervals over the low flow period. The sampling period was from June through September, 2005, and captured a variety of flow conditions. Due to unusually high precipitation during the month of June, typical low flows were not observed until August, and were only observed during that one month.

Environment Canada staff provided training for discharge measurements, and surveying was conducted according to Water Survey of Canada standards. The streams in this study had depths of less than 1.0 m in the sections being sampled. Velocity measurements were taken using a Scientific Instruments 1250 Mini "pygmy" current meter. The pygmy meter is specially designed for use in small, shallow streams; the range of operation is roughly 0.03 to 1.5 m/s. Measured velocities occasionally dropped to below 0.03 m/s, but often these were associated with pooled water, and these measurements were excluded. The average stream velocity ranged from 0.02 m/s to 0.16 m/s, with an average of 0.07 m/s. These velocities generally fall within the optimal range of operating conditions for this particular meter.

The method used to calculate the flow was the sixth-tenths depth method, and the velocity at that point assumed to approximate the average velocity in that section of the channel (Dingman, 2002). The streams were thus sampled using the velocity-area method, in which the width, depth, and velocity are measured and used to calculate the discharge. The width of the channel was measured for each cross-section, and divided into twenty evenly spaced sections. The flow measurement was taken at the mid-point of each of these vertical sections to represent an approximate average of the depth and average velocity in the panel.

The regional flow sites consisted of one cross-section, at which the flow was measured once during each site visit. Regional flow measurements were used to determine the average change in discharge over the summer season, as well as the average change in the discharge per unit length of stream channel. Water chemistry sampling was undertaken to obtain estimates of water quality parameters over the summer season, and to evaluate the interaction between the surface water flows and the groundwater along the length of the stream channel. At each of the thirteen sites, the water was sampled for temperature, pH, dissolved oxygen, alkalinity, and nitrate. The temperature, dissolved oxygen, and pH were sampled directly in the flowing stream water using portable meters. Water samples were tested for nitrate-nitrogen using a portable colourimetric meter. Total alkalinity (HCO_3^-), was determined by titration using hydrochloric acid and bromcresol green as an indicator according to standard practice.

A2.7 Results: discharge

Regional flow measurements were used to assess the contributions to baseflow from discharging groundwater over the low flow season. At any particular measurement time, discharge generally increased in the downstream direction, with the exception of Bertrand Creek between sites B3 and B2, which decreased. For example, in June, Bertrand Creek had a discharge of 0.157 m^3/s at the upstream site (B3), 0.066 m^3/s at middle site (B2), and 0.102 m3/s at the downstream site (B1) (Table A2.1). Fishtrap Creek had a discharge of 0.070 m^3/s at the upstream site (F3), while the downstream site (F2) had a significantly higher discharge of 0.300 m^3/s. Pepin Brook had a discharge of 0.064 m^3/s at the upstream site (P4), 0.182 m^3/s at the middle site (P2), and 0.239 m^3/s at the downstream site (P1).

Stream discharge generally decreased throughout the sampling season at each site, although site B2 showed a slight increase in July relative to June, and both Fishtrap (F3) and Pepin (P4) showed increases in discharge in September relative to August (Table A2.1). Table A2.2 also shows the average change per month for each site. All values are negative. Generally, water depth decreased for all sites over the sampling season (except P4 in July and P2 in August), and water depths increased for all but one site in September relative to August levels (depth data not shown). Over the low flow season, discharge decreased in all streams, on average, from between 0.009 and 0.073 m^3/s (Table A2.2).

Table A2.1: Measured flows and average rate of change in discharge at each site relative to the previous month.

Site	**June (m^3/s)**	**July (m^3/s)**		**Aug (m^3/s)**		**Sept (m^3/s)**		**Average change per month (m^3/s)**
		Flow	**Change[2]**	**Flow**	**Change[2]**	**Flow**	**Change[2]**	
B3[1]	0.157	0.001	-0.156	-	-		-	-0.056
B2	0.066	0.078	+0.013	0.014	-0.064	0.022	-0.004	-0.018
B1	0.102	0.088	-0.014	0.026	-0.062	-	-	-0.038
F3	0.070	0.041	-0.029	0.009	-0.032	0.043	+0.034	-0.009
F2	0.300	0.206	-0.094	0.085	-0.121	0.081	-0.004	-0.073
P4	0.064	0.040	-0.024	0.047	-0.007	0.062	+0.015	-0.005
P2	0.182	0.145	-0.037	0.116	-0.029	0.114	-0.002	-0.023
P1	0.239	0.156	-0.083	0.120	-0.036	0.101	-0.019	-0.046

[1] only two measurements available for B3 and three for B1 due to non-ideal conditions (overgrowth, pooling) for flow measurements.

[2]relative to the previous month

The average rate of increase in discharge per unit length of stream over the low flow season was calculated (Table A2.2). Along Fishtrap Creek, the average rate of increase in discharge between F3 and F2 was $3.17 \cdot 10^{-5}$ m^3/s per m, although some portion of this may be attributed to the inputs from Waechter Creek (located between F2 and F3). Along Pepin Brook, the average rate of increase in discharge between P2 and P1 was $2.00 \cdot 10^{-5}$ m^3/s per m, and between P4 and P2 it was $2.23 \cdot 10^{-5}$ m^3/s per m. Along Bertrand

Creek, the discharge increased slightly between B2 and B1 at an average rate of $8.71 \cdot 10^{-6}$ m^3/s per m, but decreased between B3 and B2 at an average rate of$3.51 \cdot 10^{-5}$ m^3/s. Thus, these three, apparently similar creeks appear to have varying contributions from groundwater along their lengths. As discussed later, the degree to which groundwater contributes to streamflow is dependent largely on the nature of the surficial geology. An interesting result is that the difference in discharge between two consecutive sites tends to decrease through the summer season (Table A2.2).

Table A2.2: Difference in stream discharge between consecutive sites for each month, and the average rate of change of discharge (Q) per unit length of stream over the summer season.

Station (from – to)	Separation (m)	ΔQ_{June} (m^3/s)	ΔQ_{July} (m^3/s)	ΔQ_{Aug} (m^3/s)	ΔQ_{Sept} (m^3/s)	Average $\Delta Q/m$[1] (m^3/s per m)
F3-F2	4014	+0.230	+0.165	+0.076	+0.038	3.17E-05
P2-P1	1238	+0.057	+0.011	+0.044	-0.013	2.00E-05
P4-P2	3854	+0.118	+0.105	+0.069	+0.052	2.23E-05
B2-B1	2221	+0.036	+0.010	+0.012	-	8.71E-06
B3-B2	2596	-0.091	-	-	-	-3.51E-05

[1]Along the measured length of the stream.

A2.8. Results: water chemistry

A summary of the water chemistry results from June to September are shown in Table A2.3. During the low flow period, it is anticipated that there will be greater inputs to the discharge from groundwater and corresponding trends in water chemistry.

Table A2.3: Summary of water chemistry sampling for June to August.

	June					July				
Site	Temp. (°C)	DO[1] (mg/L)	NO_3-N[2] (mg/L)	pH	Total Alk.[3] (mg/L)	Temp. (°C)	DO[1] (mg/L)	NO_3-N (mg/L)	pH	Total Alk. (mg/L)
B1	14.4	8.96	2.9	6.75	63	16.7	9.99	2.4	6.86	80
B2	19.7	10.98	4.8	6.85	93	19.1	10.69	2.5	6.93	75
B3	13.9	10.08	2.7	7.56	85	16.7	9.68	2.9	6.71	78
B4	14.8	8.36	2.1	6.79	74	16.4	5.31	2.3	6.64	88
B5	15.2	7.87	2.0	6.83	85	16.2	5.27	1.8	6.64	94
P1	13.3	8.21	1.2	7.38	124	14.2	7.31	1.2	7.64	132
P2	13.5	7.54	1.4	7.08	123	14.9	7.08	0.6	6.54	132
P3	14.3	6.21	1.5	7.14	128	15.5	4.87	1.1	7.33	116
P4	15.4	3.31	2.7	7.09	121	16.1	0.76	0.7	6.64	130
F1	12	6.46	4.6	6.52	84	13.4	5.74	4.9	6.61	79
F2	12.7	6.45	2.5	6.34	83	14.8	6.55	2.9	6.88	86
F3	14.6	6.85	2.6	6.14	73	16	6.37	1.9	6.29	82
W1	12	9.38	3.2	6.79	120	12.3	8.66	1.9	7.42	106
	August					**September**				
Site	**Temp. (°C)**	**DO[1] (mg/L)**	**NO_3-N[2] (mg/L)**	**pH**	**Total Alk.[3] (mg/L)**	**Temp. (°C)**	**DO[1] (mg/L)**	**NO_3-N (mg/L)**	**pH**	**Total Alk. (mg/L)**
B1	16.8	7.37	2.2	6.64	70	16	8.47	2.6	6.84	73
B2	18	9.91	1.8	6.8	67	14.1	6.99	1.8	6.98	98
B3	16.6	7.84	1.3	6.68	91	16.4	8.94	1.9	8.32	117
B4	16.4	3.67	1.3	6.78	121	15.3	5.39	2.8	7.32	142
B5	16.4	4.21	1.3	6.08	119	14.5	5.72	4.4	7.22	129
P1	14.8	6.9	1.5	6.68	138	14.1	6.99	1.7	6.98	142
P2	15	7.41	1.3	6.96	147	13.9	7.36	1.5	7.17	142
P3	15.2	3.86	1.5	6.94	150	14.1	5.02	1.1	7.16	153
P4	15.9	0.85	2.0	6.94	139	14.8	1.75	1.9	7.02	135
F1	14.1	5.53	1.6	7.15	92	12.5	4.77	2.4	6.53	93
F2	14.3	5.58	1.9	6.95	88	13.1	5.63	1.9	6.32	92
F3	16.4	6.93	1.8	7.27	98	14.4	6.37	1.0	6.71	82
W1	14	8.33	1.9	7.38	102	12.1	8.63	2.6	6.33	110

[1]DO=Dissolved oxygen
[2]NO_3-N=nitrate-nitrogen
[3] Total Alk=Total alkalinity as HCO_3^-

Stream temperature ranged from 12.0 to 19.7 °C over the summer. Temperature increased consistently from June to July for all sites except B2. The average increase in temperature was 1.3 °C. From July to August, roughly 50% of the streams increased in temperature, while the balance decreased; there were no consistent trends. From August to September, all streams decreased in temperature, on average, by 1.4 °C. Bertrand Creek had the highest water temperatures of the three streams, and was the most variable throughout the season. Fishtrap Creek had lower temperatures in the lower reaches, and the temperature at sites F1 and F2 remained relatively stable throughout the summer, with a variation of only 2°C. Pepin Brook had moderately variable water temperature throughout the summer, but the water was considerably cooler than the temperatures measured in Bertrand Creek.

Dissolved oxygen (DO) ranged from 0.76 to 10.98 mg/L over the summer. Sites B1, B2 and B3 maintained relatively high dissolved oxygen concentrations throughout the season, suggesting limited groundwater input. DO decreased at sites B4 and B5 through August, but then increased slightly in September. At most sites along Fishtrap Creek, dissolved oxygen generally decreased throughout the summer, but interestingly there was no correlation with temperature. DO in Pepin Creek tended to decrease from June to July, but no discernable trends are evident from July to September.

Nitrate-nitrogen (N-NO3) ranged from 0.6 to 4.8 mg/L. Concentrations were highest in Bertrand Creek. Otherwise, there are no strong trends in concentration either along streams or over the summer season.

The pH ranged from 6.08 (B5: August) to 8.32 (B3: September). Variations in pH both between sites along the same stream and over the low flow season were subtle. Pepin Creek generally had the highest pH. Otherwise there were no discernable trends. Increases in stream pH are often associated with greater contributions from groundwater; but this parameter is not a strong indicator of such interactions in this study.

Total alkalinity, which tends to be related to pH, ranged from 63 to 153 mg/L. Total alkalinity was lowest is Bertrand Creek, although it was only slightly lower than the average values along Fishtrap Creek. Pepin Brook had the highest total alkalinity, and concentrations tended to increase throughout the summer.

Waechter Creek, a tributary to Fishtrap Creek, had slightly different chemistry, with higher dissolved oxygen, higher pH, and higher total alkalinity. This variable chemistry may have influenced the downstream trends in the main channel of Fishtrap Creek.

A2.9 Discussion of field study

Low flows in Fishtrap Creek, Bertrand Creek, and Pepin Brook occur during the summer months, during the period of least precipitation. During this time, the primary contribution to streamflow is groundwater. Groundwater tends to flow in a southerly direction, from a groundwater high in the Langley Uplands to the Lynden Terrace, although, as mentioned earlier, the flow directions at a local scale can be more complex due to the nature of the channel geometry and the heterogeneity of the aquifer sediments. The nature of the interaction between groundwater and streamflow in Fishtrap, Pepin and Bertrand Creeks is considered to be a largely a function of the surficial geology.

Generally, streamflow was observed to have increasing contributions from groundwater further downstream, as evidenced by increasing discharge between regional sites (except between B2 and B3 where discharge decreased). The most significant increase in discharge was in Pepin Brook (between sites P1 and P2). In this section, the stream flows out of the Langley Uplands and onto Lynden Terrace. Not only is there a topographic change, but there is also a transition from lacustrine and fluvial Salish Sediments to glaciofluvial sand and gravel of the Sumas Drift. Pepin Brook sites P1, P2, and P4 are in a shallow water table zone (<3m), thus, there is a much greater likelihood of strong interaction between stream and aquifer.

Fishtrap Creek generally flows over uniform sand and gravel from the Abbotsford Outwash deposit. These sediments are highly permeable and would, therefore, promote strong (un-hindered) interaction between the aquifer and streams. As suggested by Johanson (1988), Fishtrap Creek appears to be dominant groundwater discharge zone as it maintains the highest flows of the three streams studied.

Bertrand Creek at site B2 had a decrease in the discharge that was more significant than gains between site B1 and B3. Bertrand Creek flows over complex deposits. The stream channel is underlain by Fort Langley Formation, which is a stony till deposit with generally low permeability, but with interlayered high permeability sand and gravel lenses. Thus, depending on the deposits within the immediate vicinity of the stream, the nature of the interaction between groundwater and surface water can be quite variable. The water table is also deeper along this portion of the channel, with the average depth to the water table being between 3-8m immediately adjacent to the stream. This suggests limited connection between the stream and the aquifer. It is worth noting, however, that the effect of pumping near the channel cannot be ruled out as a possible cause for the declines in streamflow at B2, despite the fact that groundwater pumping is not expected to be significant in this area.

Discharge also tended to decrease over the summer season, except in Bertrand Creek where discharge increased in July, and at Pepin and Fishtrap where discharge increased in September. The decreasing contributions of groundwater to streamflow over the low flow season are not unexpected as groundwater levels similarly decrease over the summer, thereby lowering the hydraulic gradient to the streams and reducing the flux. The opposite trends in Bertrand in July may point to a delayed response owing to the less permeable aquifer materials, although this is uncertain. Increasing streamflow in September relative to August at Fishtrap and Pepin point to a wet period between August 11 and September 10, 2005. During this one month period, a total of 35.1 mm of rain fell. However, from June 15 to August 11 (2 months) a total of 73.4 mm fell (36.7 mm per month), so this explanation is not likely. An alternative explanation is irrigation return flow, which would add to runoff and groundwater recharge. As this is raspberry production area, irrigation is often used during the summer to supplement minimal rainfall. These results imply that the contributions from groundwater are both spatially varying and temporally varying within the study area.

Interestingly, the difference in discharge between two consecutive sites tended to decrease through the summer season. This result suggests that there is less and less of a difference between the streamflow between consecutive sites. This is interpreted to be the result of a declining water table over the summer season. In the early summer, the water table in the Langley Uplands is at higher elevation than later in the season. In this area, the water table is generally deep, and there are significant fluctuations in groundwater level (greater than 2 m) as evidenced by the provincial observation well data, which are all at high elevation. Whereas, as lower elevation areas, such as the Lynden Terrace, groundwater levels are much shallower. For example, at Pepin Brook, the water table is <3 m deep. Thus, as the water table drops over the summer, it does so more significantly at higher elevation, leading to a gradual decline in the hydraulic gradient, and a consequent gradual decline in the flux of groundwater entering the streams at progressively lower elevations.

The water chemistry results further support the variable contributions of groundwater to surface water. Under natural conditions, groundwater often has low dissolved oxygen, higher total alkalinity, and a more stable and colder temperature than surface water in summer. The high water temperatures and high dissolved oxygen in Bertrand Creek indicate that this stream does not receive a significant groundwater input. This indicates that Bertrand Creek has limited groundwater interaction, as corroborated by the lower discharge measurements at B2 relative to the other sites. In Fishtrap Creek and Pepin Brook, the discharge increased downstream and the temperatures were more stable, indicating groundwater contributions to streamflow. The dissolved oxygen was also lower in both of these streams. The dissolved oxygen decreased during the summer, and the alkalinity increased, both supporting increased groundwater inputs relative to streamflow. Therefore, despite the fact that less and less groundwater is contributed over the summer season between consecutive sites, the relative amount of groundwater to surface water increases.

Appendix 3 Description of TreeGen downscaling algorithm

The TreeGen downscaling algorithm consists of four steps: (1) common principal component analysis (PCA) of observed/GCM predictor fields (Imbert & Benestad, 2005); (2) synoptic map-type classification via a multivariate regression tree (Cannon et al. 2002a,b); (3) a nonparametric weather generator based on conditional resampling of weather elements from each map-type (Buishand & Brandsma, 1999); and (4) within-type extrapolation based on stepwise linear regression and linear trend analysis [*Imbert & Benestad*, 2005]. Each step is described in turn.

To mitigate potential biases between the NCEP/NCAR Reanalysis and GCM simulated predictors, common PCA was applied to the two datasets (Imbert & Benestad, 2005). First, predictor fields from the reanalysis and GCM were standardized so that the time series for each grid point had zero mean and unit variance during a common 1961-2000 baseline period. Second, standardized data from the reanalysis and the GCM were concatenated to form a single data matrix. Third, PCA was applied to the correlation matrix of the concatenated predictors. Finally, common PC scores from the GCM were rescaled so that their means and variances in the simulated baseline period matched observed values from the same period. To ensure consistency of the simulated seasonal cycle, rescaling was performed on each month separately.

Following the common PCA, synoptic map-types were defined using a multivariate regression tree (MRT) model that recursively split the observed data into groups based on thresholds in the common PC scores (Cannon et al. 2002a,b). Values of the thresholds were optimized so that the associated daily surface weather elements at the Upper La Joie station were placed into groups (or map-types) that were as homogeneous as possible. Thirteen map-types were selected via cross-validation using the '1-SE'criterion described by Cannon et al. (2002a).

Once the synoptic map-types were defined from the historical record, common PC scores from the GCM were then entered into the MRT and each day was classified into one of the map-types. Next, surface weather conditions on a given day were predicted using a nonparametric weather generator based on conditional resampling from cases assigned to that day's map-type (Buishand & Brandsma, 2001). The probability $p(i)$ of randomly selecting weather elements observed on day i as the predicted values on day t was taken to be inversely proportional to the Euclidean distance $d(t-1, i-1)$ between the predicted values on the previous day $t-1$ and historical values of the weather elements on day $i-1$,

$$p(i) = \frac{1/(d(t-1, i-1)+b)^a}{\sum_{h \in I} 1/(d(t-1, h-1)+b)^a},$$

where I is the set of historical days assigned to the predicted map-type occurring on day t and a and b are parameters that modify the shape of the nonparametric probability density function. In this study, values of $a = 7$ and $b = 0.1$ were selected to minimize the absolute bias between predicted and observed lag-1 and lag-2 autocorrelations.

As the nonparametric weather generator sampled cases from the historical dataset, future trends in surface climate conditions were due exclusively to changes in the frequency and timing of the synoptic map-types simulated by the GCM. Additional processing was thus needed to generate values above/below the highest/lowest records in the historical dataset and to accurately reflect trends occurring within map-types. To capture both betweenand within-type trends, a modified version of the extrapolation algorithm described by Imbert & Benestad (2005) was adopted. For each map-type, multiple linear regression equations linking the common PC scores and the surface weather elements were first created via stepwise regression based on the Bayesian Information Criterion (Schwarz, 1978). For precipitation, separate models were built for occurrence of precipitation (via regression estimation of event probabilities) and log-transformed precipitation amounts on wet days. Linear trends in predicted temperatures and precipitation amounts relative to baseline from the regression equations were then estimated for each map-type. Finally, these trends were superimposed onto the time series derived from the conditional resampling.

Appendix 4 Outputs from the project

A4.1. Overview

Outreach activities associated with this project have involved a range of outputs, aimed at different audiences. For example, we solicited feedback from participants at the Climate Change and Fisheries conference in Victoria at the outset of the project in regard to important questions of relevance to fisheries biologist. We communicated with the scientific community by presenting results at meetings of scientific organizations, such as the Canadian Geophysical Union; by organizing a special session at the 2006 Annual Meeting of the American Geophysical Union; and by preparing manuscripts for publication in peer-reviewed scientific journals. Exposing this research to the international scientific community helps to validate the work and establish its credibility.

We have also aimed our research at users of the results. This audience includes government scientists involved in research on and management of aquatic resources, including fisheries and water; hydroelectric power providers such as BC Hydro and operators of independent power projects, many of which are located on glacier-fed streams; and consultants who provide technical and scientific support to the managers. We have reached out to this audience by presenting our work at meetings such as the Canadian Water Resources Association – BC Branch annual meeting, by organizing a one-day workshop focused on the results of the project, and by participating in a workshop on low flow risk assessment, run by Dr. Martin Carver of the BC Ministry of Environment.

Details of these activities are provided below.

A4.2. Detailed summary

A4.2.1 Workshop organization and participation

Canadian Climate Impacts and Adaptation Research Network (C-CIARN) Fisheries Sector, and American Fisheries Society – Pacific Northwest Branch. Climate and Fisheries: Impacts, Uncertainty and Responses of Ecosystems and Communities. Victoria Conference Centre, Victoria, BC October 26-28, 2005. We organized a special session and workshop on "Climate, Low Flows and Fish."

American Geophysical Union, 2006 Fall Meeting, San Francisco, CA December 2006. We organized a special session on "Low flows and climate change." The session was co-sponsored by the Global Change and Public Policy sections of AGU.

Climate Change and Low Flows. Vancouver, March 12, 2007. This was a one-day workshop at which the research team presented the project results to an invited audience comprising scientists, managers and consultants involved in managing aquatic resources, followed by a workshop to discuss management adaptations and ongoing information needs. The audience represented agencies including Fisheries and Oceans Canada, BC Ministry of Environment, BC Hydro, Knight Piesold Consulting and David Suzuki Foundation. See Appendix 4 for more details.

BC Ministry of Environment, Workshop on Risk Assessment Modelling – Low Flows. Vancouver, July 12, 2007. Several members of the research team (Allen, Hutchinson, Moore, Stahl, Whitfield) were invited participants in this workshop to help BC MoE scientists develop a risk assessment model. Allen and Stahl made presentations based on the project results to initiate discussion.

A4.2.2 Conference and workshop presentations

Allen, D.M. 2007. Groundwater – Stream Flow Interactions: A British Columbia Context. Invited presentation. Workshop on Low-Flow Hazard Modelling. BC Ministry of Environment, Vancouver. July 12, 2007.

Allen, D.M. 2007. Groundwater-Surface Water Interaction: Groundwater Contribution to Low Flows. Invited speaker and panelist. Speaking for the Salmon Series "Salmon and Groundwater". Simon Fraser University, Burnaby, BC, March 6, 2007.

Allen, D.M. 2007. Groundwater – Stream Flow Interactions: Climate Change Linkages. Workshop on Climate Change and Low Flows: Influences of Glaciers and Groundwater, Vancouver, March 12, 2007.

Moore, R.D. 2007. Hydrologic consequences of glacier retreat in British Columbia. Invited presentation to the Department of Earth and Space Science, University of Washington, Seattle, May 10, 2007.

Moore, R.D. 2005. Will glacier recession cause stream warming? Climate and Fisheries. Canadian Climate Impacts and Adaptation Research Network (C-CIARN) Fisheries Sector, and American Fisheries Society – Pacific Northwest Branch, Victoria, October 2005.

Stahl, K., Moore, R.D., Shea, J., Hutchinson, D. and Cannon, A. 2007. Coupled modelling of glacier and streamflow response to future climate scenarios. Presented at the Annual Meeting of the Canadian Geophysical Union. St. Johns, Newfoundland, June 2007.

Stahl, K. 2007. Low flow processes, characteristics, and issues in mountainous regions of Western Canada. PUB (Prediction in Ungauged Basins) Low Flow Workshop. The Canadian Society for Hydrological Sciences. Quebec, QC, April 2007.

Stahl K., Moore, R.D. and Allen, D. 2007. Empirical analysis of climate influences on late-summer streamflow. Workshop on Climate Change and Low Flows, Vancouver, March 2007.

Stahl, K., Moore, R.D., Shea, J., Hutchinson, D. and Cannon, A. 2007. Coupled modelling of glacier and streamflow response to future climate scenarios. Workshop on Climate Change and Low Flows: Influences of Glaciers and Groundwater, Vancouver, March 2007.

Stahl, K. 2007. Streamflow changes in BC: impact of climate change and glacier retreat. A workshop to determine research priorities for declining Eulachon stocks. Richmond, BC, February 2007.

Stahl, K., Moore, R.D, Shea., J.M., Hutchinson, 2006. D. Effects of glacier retreat on summer low flow. Presented at the Fall Meeting of the American Geophysical Union, San Francisco, December 2006.

Stahl, K., Moore, R.D, Shea., J.M., Hutchinson, D. 2006. Effects of glacier retreat on summer streamflow in selected basins in British Columbia, Canada. Presented at the Annual Meeting of the Canadian Water Resources Association–B.C. Branch Conference, Vancouver, October 2006.

Stahl K. and Moore, R.D. 2006. Influence of watershed glacier coverage on summer streamflow in British Columbia, Canada. Presented at the European Geophysical Union General Assembly, Vienna, April 2006.

A4.2.3 Journal articles (published, submitted and in preparation)

Allen, D.M., Werner, A. and Whitfield, P. (in prep).Groundwater responses in mountainous terrain: regime classification, and linkages to climate and streamflow.

Berg, M.A. and Allen, D.M. (in press). Low flow variability in groundwater-fed streams. Canadian Water Resources Journal, 32(3): 1-20.

Stahl, K. and Allen, D.M. (in prep). Historic trends in groundwater levels in British Columbia.

Stahl K. and Moore, R.D. 2006. Influence of watershed glacier coverage on summer streamflow in British Columbia, Canada. Water Resources Research 42, W06201, doi:10.1029/2006WR005022.

Stahl, K., Moore, R.D., Shea, J., Hutchinson, D., Cannon, A. Coupled modelling of glacier and streamflow response to future climate scenarios. Water Resources Research (in revision).

Stahl, K. Moore, R.D., Allen, D.A. (in prep.) Sensitivity of summer streamflow to short-term climate and long-term storage changes.

A4.2.4 Further contributions related to project

Burn D.H., Buttle J.M., Caissie D., MacCulloch G., Spence C. and Stahl K. The processes, patterns and impacts of low flows across Canada. Canadian Water Resources Journal (submitted)

Stahl, K., van Lanen, H.A.J, Uhlenbrook, S. Chapter 4: Processes and Regimes. In: Manual on the estimation and prediction of low flows. (ed. by Gustard, A. and Demuth, S.) World Meteorological Organization, HWRD. (in review)