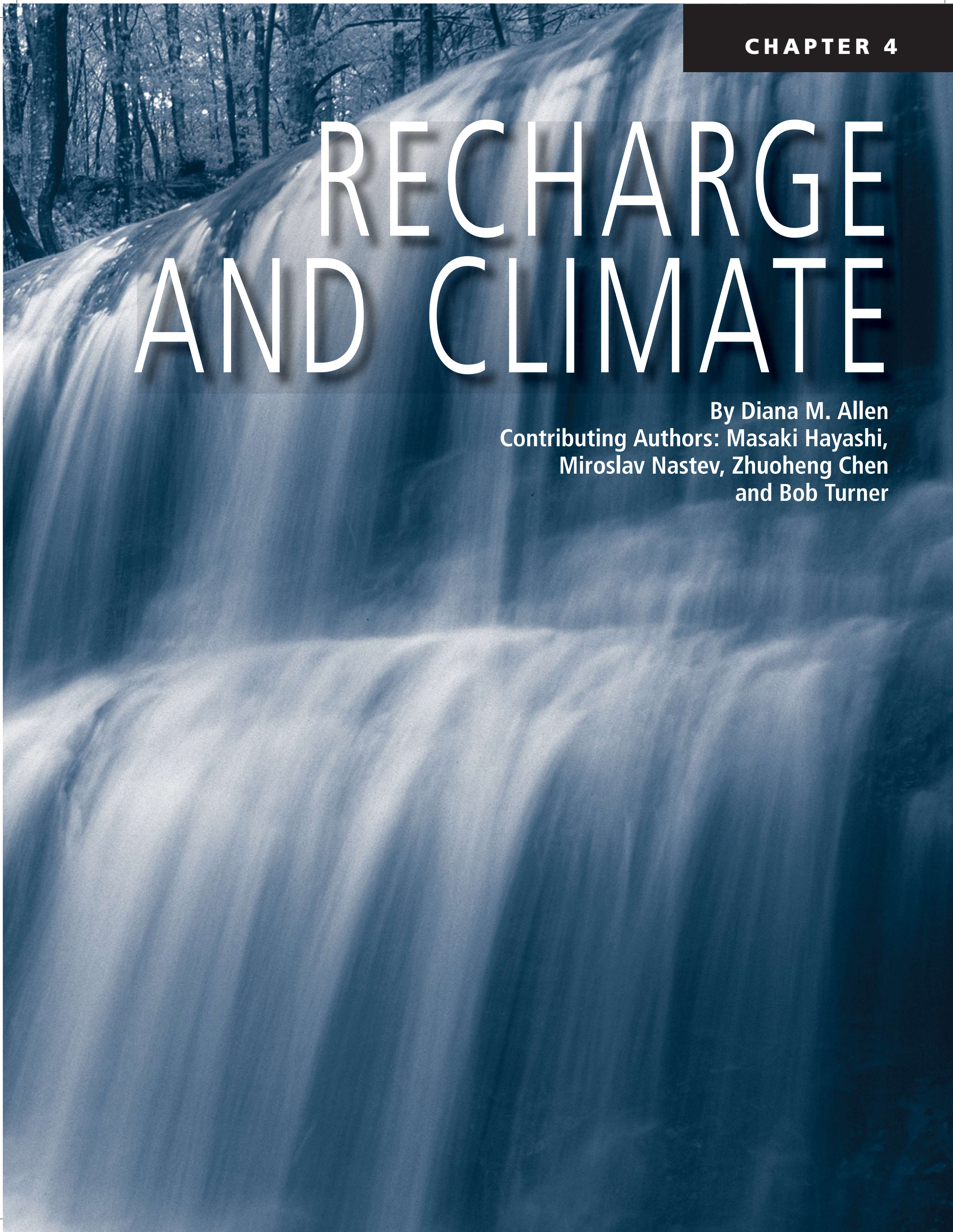

CHAPTER 4

RECHARGE AND CLIMATE

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4.1 INTRODUCTION

Recharge, the water that enters an aquifer, is a critical parameter for understanding, modelling and protecting groundwater systems from over-exploitation and contamination (Lerner, 1997; Lerner et al., 1990; Scanlon et al., 2002). Defining recharge rates and their temporal variability in response to climatic fluctuation and anthropogenic stresses is integral to groundwater management and planning (Allison, 1988; Sami and Hughes, 1996; Simmers, 1987). The spatial patterns of recharge determine where groundwater recharge occurs, which is crucial for protecting groundwater (i.e., source protection) and for remediating contaminated sites (Allison, 1988; Bradbury et al., 1992). Contrary to widespread belief, the amount of recharge does not indicate the sustainable rate of groundwater extraction, because if groundwater is extracted at the rate equal to recharge, there will be no water left to sustain stream baseflow and the riparian and wetland vegetation that are dependent on the shallow water table (Sophocleous, 2000; Bredehoeft, 2002; Devlin and Sophocleous, 2005).

Recharge is controlled by a number of factors, ranging from climate, land cover / land use, topography, and the characteristics of the soil and geologic substrate. Some of these factors vary with time (such as climate and vegetation), while all vary spatially. In addition, recharge to shallow aquifers may be subsequently extracted by evapotranspiration without contributing to long-term aquifer replenishment (de Vries and Simmers, 2002). Therefore, recharge is temporally and spatially variable, making its estimation difficult.

As a whole, the scientific community has developed simple and sophisticated methods for measuring and estimating recharge (e.g., Scanlon et al., 2002), but because natural systems are so dynamic, and the processes so complex, accurate estimates of recharge are often difficult to obtain. Simmers (1987) identified one of the main problems with recharge estimation: "No single comprehensive estimation technique can yet be identified from the spectrum of methods available; all are reported to give suspect results." With this in mind, it is important to recognize the limitations of groundwater recharge estimation techniques and the consequent limitation on our ability to predict how an aquifer system will respond to natural and anthropogenic change.

4.2 WHAT IS GROUNDWATER RECHARGE?

Recharge is the process by which groundwater is replenished. Thus, recharge contributes to the overall volume of fresh water available in the ground. Water can enter the ground either directly from rainfall and snowmelt (direct recharge), indirectly from influent streams and rivers (indirect recharge), or from a concentrated source resulting from the horizontal surface flow of water in the absence of well-defined channels (localized recharge) (Lerner et al., 1990).

Groundwater is found in both unconfined and confined aquifers (Figure 4.1). Unconfined, or water-table, aquifers are shallow and frequently overlie one or more confined aquifers. They are recharged through permeable soils and subsurface materials above the water table. Unconfined aquifers usually contain younger water (light blue arrows in Figure 4.1) with shorter travel paths, and less mineralization. Surface contamination, originating, for example, from agriculture or other land surface activities, can potentially be transported down to the water table of an unconfined aquifer and pose a threat to water quality. Consequently, unconfined aquifers are generally more vulnerable to contamination originating from surface or near surface activities. Confined aquifers usually occur at depth, and may overlie other confined aquifers. Confined aquifers may be recharged at some distance from a point of extraction and, in some cases, very deep aquifers may be recharged in remote mountain ranges. Typically this groundwater is

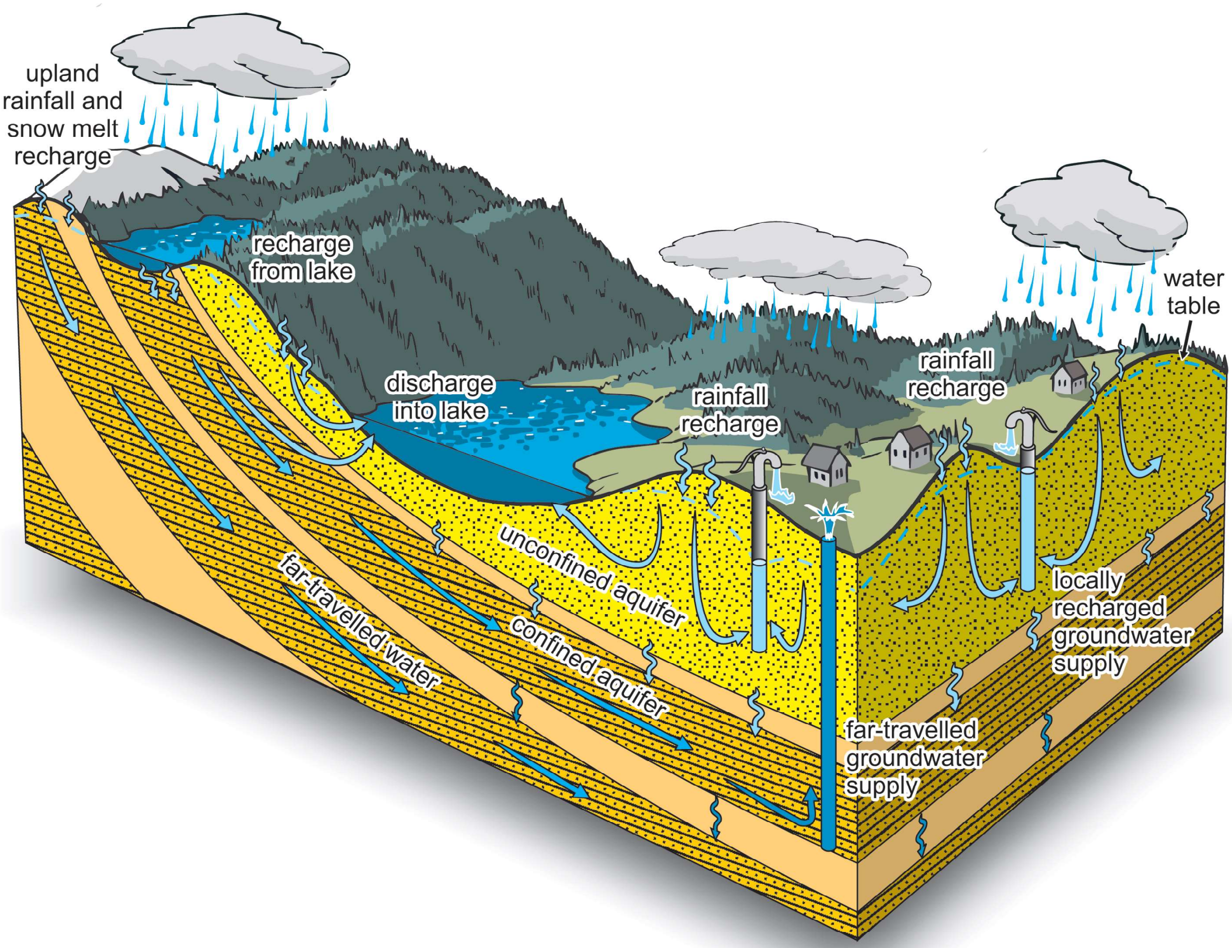


Figure 4.1 Groundwater in unconfined and confined aquifers showing various rrecharge and discharge pathways. Groundwater is recharged both at high elevation and low elevation from snowmelt and rain. Groundwater typically flows from areas of high to low elevation, and often discharges in lakes or streams situated at low elevation.

older (dark blue arrows in Figure 4.1) and often more mineralized due to its longer contact with the rocks and sediments. Water infiltrating bedrock in the mountains may flow downward and then move laterally into confined aquifers. In some regions, confined aquifers extend for many hundreds of kilometres beneath the land surface, and may cross natural surface watershed boundaries or jurisdictional boundaries. As such, groundwater in a confined aquifer may recharge in one watershed and discharge in another, or perhaps recharge in one province or nation, and discharge in another. Confined aquifers can also be recharged through cracks or openings in the less permeable layers above or below them. Even low permeability clay tills have been shown to provide pathways from the surface to confined aquifers (van der Kamp, 2001). Confined aquifers in complex geological formations may be partly exposed at the land surface, or the low permeability confining layer may be breached, allowing direct recharge from infiltrating precipitation.

Direct and indirect recharge processes in the shallow subsurface are illustrated in Figure 4.2, which shows water directly reaching the water table of an unconfined aquifer through the vadose

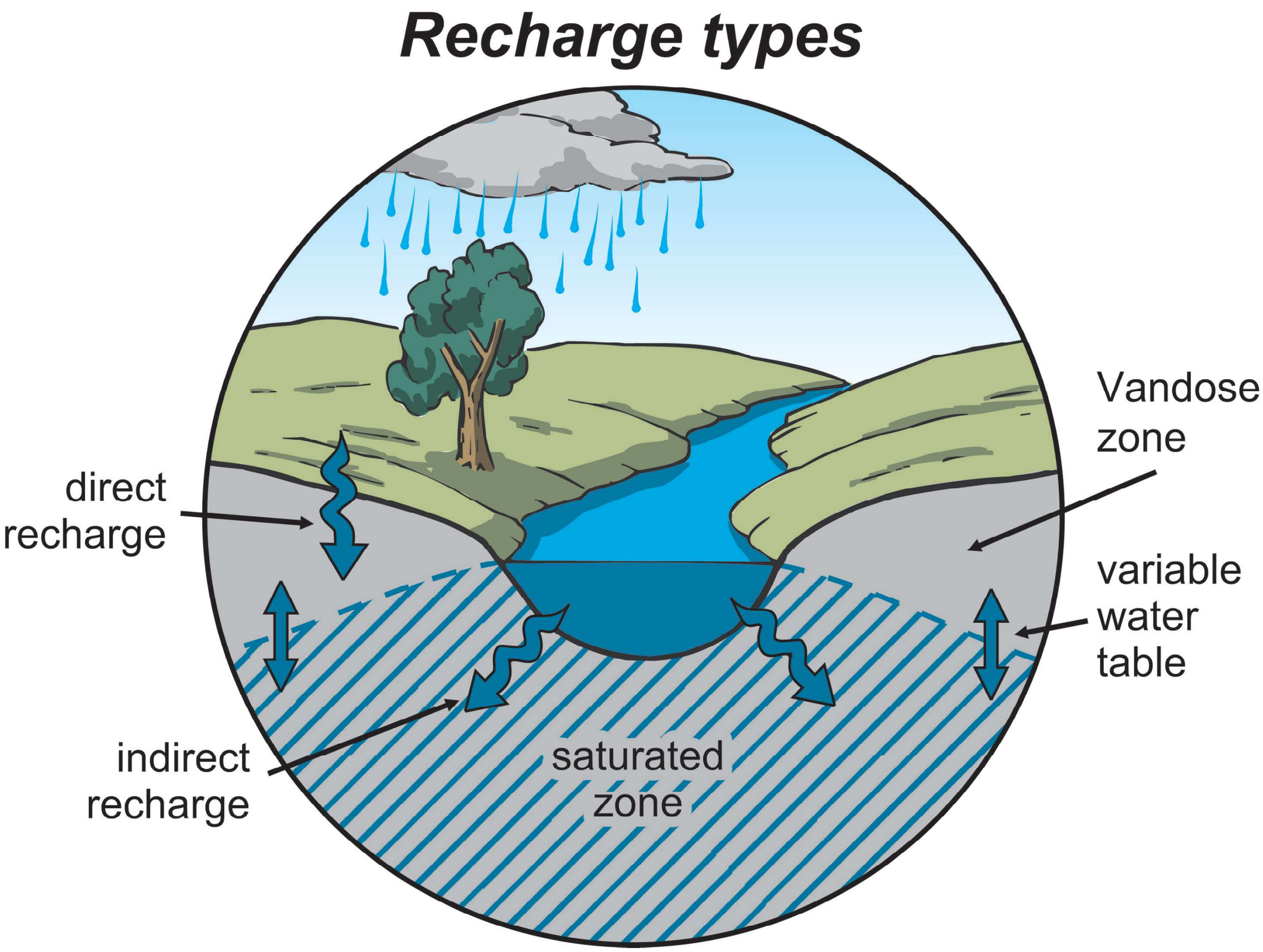


Figure 4.2 Direct recharge occurs directly from precipitation (rainfall, snowmelt) when water enters the unsaturated zone and percolates downward to the saturated zone. Indirect recharge occurs through influent streams, rivers and lakes, contributing water to the saturated zone. In both cases, the water table responds dynamically, and its height varies seasonally.

zone and also entering the groundwater system via an influent stream. In some areas, groundwater systems are recharged by both direct and indirect mechanisms. For example, the unconfined Grand Forks aquifer in south-central British Columbia (Box 4-1) receives recharge both directly from precipitation and from the Kettle River, which meanders through the valley.

Direct recharge is the focus of this chapter and, hereafter, is referred to simply as groundwater recharge. A detailed treatment of indirect mechanisms of groundwater recharge, through the interaction with surface water systems, is discussed in Chapter 5.

The rate of direct recharge in arid and semiarid environments is very small because evapotranspiration demands often exceed available precipitation. Short-term additions of groundwater to shallow aquifers may be subsequently extracted by evapotranspiration without contributing to aquifer replenishment (de Vries and Simmers, 2002). As a result, indirect and localized recharges play important roles in these environments. For example, in southern parts of the Prairie Provinces, the driest region in Canada, lateral flow of snowmelt and storm runoff concentrates water into topographically closed

depressions, causing depression-focused groundwater recharge (Box 4-2).

4.3 RECHARGE, DISCHARGE AND GROUNDWATER FLOW

A **recharge area** is where the net direction of groundwater flow is downward, thereby contributing to groundwater storage in the aquifer. Most areas, unless composed of solid rock or covered by pavement in developed areas, allow a certain amount of infiltrated water to percolate through the unsaturated zone and reach the aquifer. Areas that transmit the most water are often referred to as "**high**" or "**critical**" **recharge areas**. In the case of unconfined aquifers, precipitation moves downward to the water table. In the case of confined aquifers, groundwater from distant areas, or from overlying or underlying aquifers, can contribute to recharge. The geometry of the aquifer (where it outcrops) and the permeability of the overlying and underlying geologic units all determine where recharge to confined aquifers occurs. Recharge in shallow aquifers can also occur in association with streams, rivers and lakes as discussed above.

Discharge areas are the opposite of recharge areas. They are the locations where groundwater leaves the aquifer and perhaps flows to the surface. Groundwater discharge occurs where groundwater flow is directed upward. In shallow aquifers, discharge occurs where the water table intersects the land surface. Springs and seeps may flow into freshwater bodies, such as lakes or streams, or they may flow into saltwater bodies. Groundwater can also move vertically upward from the water table as a consequence of evapotranspiration mechanisms. These processes effectively remove water from the saturated and unsaturated zones. Discharge from deep aquifers can occur through semi-permeable confining beds into shallower aquifer systems, and likewise, leakage from shallow unconfined aquifers can recharge deeper aquifers. Pumping wells are also an anthropogenic cause of groundwater discharge, affecting the rates as well as areas of recharge and discharge.

Groundwater usually flows from areas of high elevation to areas of low elevation (see Figure 4.1). However, mountains are not needed for recharge to occur, or for groundwater to flow. Recharge can occur at low elevations, and even small topographical changes can influence groundwater flow at local scales. Generally, the more permeable the rock or sediments and the steeper the topography, the faster the groundwater flow. The presence of pumping wells can also artificially increase the rate at which groundwater flows, because groundwater is drawn towards the well using a pump.

4.4 HOW OLD IS THE GROUNDWATER?

The time that groundwater spends in the ground is referred to as the **residence time**, and this can vary from days to millions of years, depending on the geology and physiographic setting. During its time in "residence", groundwater is said to exist in **storage**, although, while in storage, groundwater continues to flow from its recharge area to a discharge area or point of extraction.

Groundwater flow in aquifers comprised of low permeability rock, even in steep terrain, will often be sluggish, moving at rates of only a few centimetres per year. As a result, groundwater in deep, confined aquifers can be hundreds, or even millions, of years old. Very old groundwater is found in deep aquifers throughout the Prairies (Bachu and Underschultz, 1995) and other areas of Canada, such as the Canadian Shield (Clark et al., 2000; see also Chapter 11). Should this old groundwater be extracted, it could take thousands or millions of years to replenish the aquifer by natural processes.

Groundwater in unconfined, high permeability aquifers flows much more quickly than in confined aquifers, particularly when the topography is steep. Residence times for shallow groundwater are typically on the order of months or tens of years. These types of aquifers are replenished more quickly than confined aquifers, often on an annual basis, and, because of their shallow nature, are often connected to surface water bodies.

4.5 WHAT FACTORS CONTROL SHALLOW GROUNDWATER RECHARGE?

Groundwater recharge to shallow aquifers occurs when water from precipitation enters the soil's unsaturated zone (infiltration), percolates downward under the force of gravity through the root zone (drainage), and is ultimately added to the saturated zone of the groundwater system (recharge). The complex series of processes that control shallow groundwater recharge are both time dependent and spatially variable (Balek, 1988). This is how recharge contributes to the temporary or permanent increase in groundwater storage.

Shallow groundwater recharge is controlled by a number of factors. The climate of an area exercises the most important control on recharge because it determines not only the amount and timing of precipitation, but also temperature, relative humidity, and wind and air movement, all of which are important factors influencing evapotranspiration. Local physical and biological conditions at the land surface affect the amount of water infiltration. These conditions include topography (slope), the nature of the land use or land cover (vegetation or

TABLE 4.1 TOTAL PRECIPITATION FOR SELECTED MONTHS WITH ANNUAL TOTALS AND MEAN DAILY TEMPERATURE FOR DIFFERENT CANADIAN CITIES

PLACE	PRECIPITATION (MM)					MEAN DAILY
	JANUARY	APRIL	JULY	OCTOBER	TOTAL ANNUAL	TEMPERATURE (°C)
Vancouver, B.C.	153.6	84.0	39.6	112.6	1199	10.1
Summerland, B.C.	29.7	25.6	30.2	18.0	326.7	9.0
Edmonton, Alta.	22.7	26.3	95.2	19.8	482.7	2.4
Calgary, Alta.	11.6	23.9	67.9	13.9	412.6	4.1
Regina, Sask.	14.9	23.5	64.4	21.8	388.1	2.8
Saskatoon, Sask.	15.2	23.9	60.1	16.7	350.0	2.2
Winnipeg, Man.	19.7	31.9	70.6	36.0	513.7	2.6
Toronto, Ont.	52.2	68.4	74.4	64.1	792.7	7.5
Ottawa, Ont.	70.2	72.4	90.6	79.4	943.5	6.0
Montreal, Que.	78.3	78.0	91.3	77.8	978.9	6.2
Quebec, Que.	89.8	81.2	127.8	101.7	792.7	7.5
Moncton, N.B.	119.2	99.3	103.3	103.8	1223.2	5.1
Halifax, N.S.	149.2	118.3	102.2	128.7	1452.2	6.3
Charlottetown, P.E.I.	106.4	87.8	85.8	108.6	1173.3	5.3
St. John's, Nfld.	150.0	121.8	89.4	161.9	1513.7	4.7
Whitehorse, Yukon	16.7	7.0	41.4	23.8	267.4	-0.7
Yellowknife, N.W.T.	14.1	10.8	35.0	35.0	280.7	-4.6
Iqaluit, N.T.	21.1	28.2	59.4	36.7	412.1	-9.5

All figures based on the 30-year period 1971 to 2000 inclusive. Source: Environment Canada (2002). Stations typically located at airports.

pavement), characteristics of the soil and geologic substrate, depth to the water table, and other factors. Recharge is promoted by cool wet climates, natural vegetation cover, flat topography, permeable soils, a water table that lies at some depth (not at surface), and the absence of low permeability confining beds. Some of these factors vary with time (e.g., land use), while all vary spatially.

4.5.1 Climate across Canada

The climate of a region also plays an important role in groundwater recharge. Although climate is influenced by regional to very local effects, it is most commonly described in terms of precipitation and temperature.

Precipitation (P) is a major component of the hydrologic cycle, and the most important climate variable controlling recharge. Precipitation reaches the Earth's surface in many different forms, including rain, freezing rain, snow, sleet, and hail. Some 71% of total precipitation in southern Canada comes from rainfall events. In northern Canada, more than 50% of total precipitation comes from snowfall events (Zheng et al., 2001). The form of precipitation and its timing are as important as the amount that falls.

Canada's land area does not receive uniform precipitation. The west coast receives 2,500 to

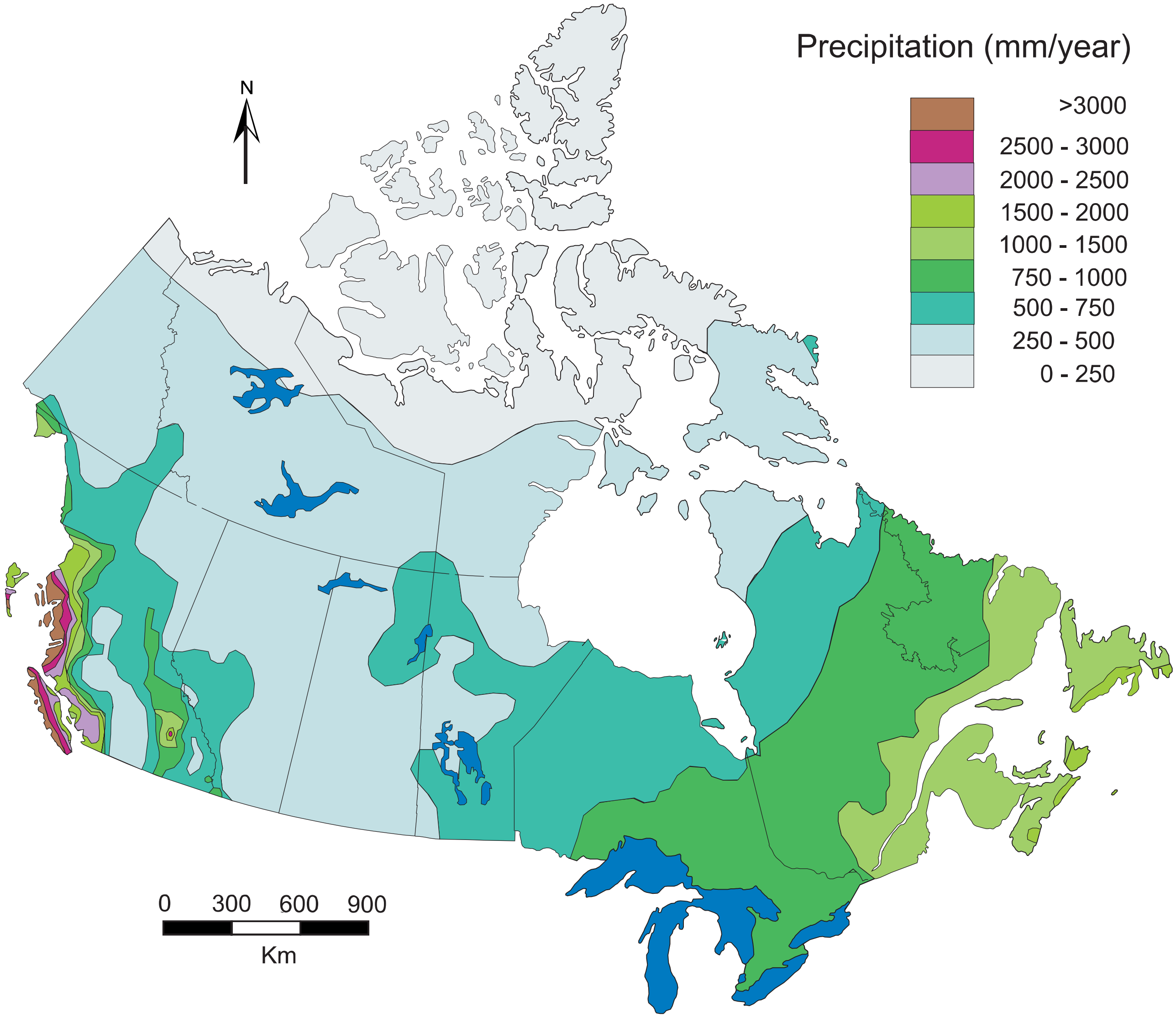


Figure 4.3 Variation in mean annual precipitation (mm/year) across Canada (based on Canada precipitation map, Department of Energy, Mines and Resources, Forestry (now Natural Resources Canada), 1991).

over 4,000 millimetres of precipitation annually, compared with 1,000 to 1,250 millimetres on the east coast, and 250 to 500 millimetres on the Prairies (Figure 4.3). The extreme north receives little precipitation, no more than 120 to 150 millimetres per year, which falls mainly as snow. Table 4.1 provides representative seasonal precipitation values (for the months of January, April, July, and October) as well as total annual precipitation for different cities across Canada.

Temperature (T) also varies regionally, although not as dramatically as precipitation. Maximum summer temperatures range from 15°C to about 20°C, and winter lows range from about 5°C on the west coast to below -20°C in Iqaluit. Mean daily temperatures for selected Canadian cities are shown in Table 4.1.

4.5.2 Evapotranspiration and potential recharge

Evapotranspiration (ET) is the sum of evaporation and plant transpiration, and varies both regionally

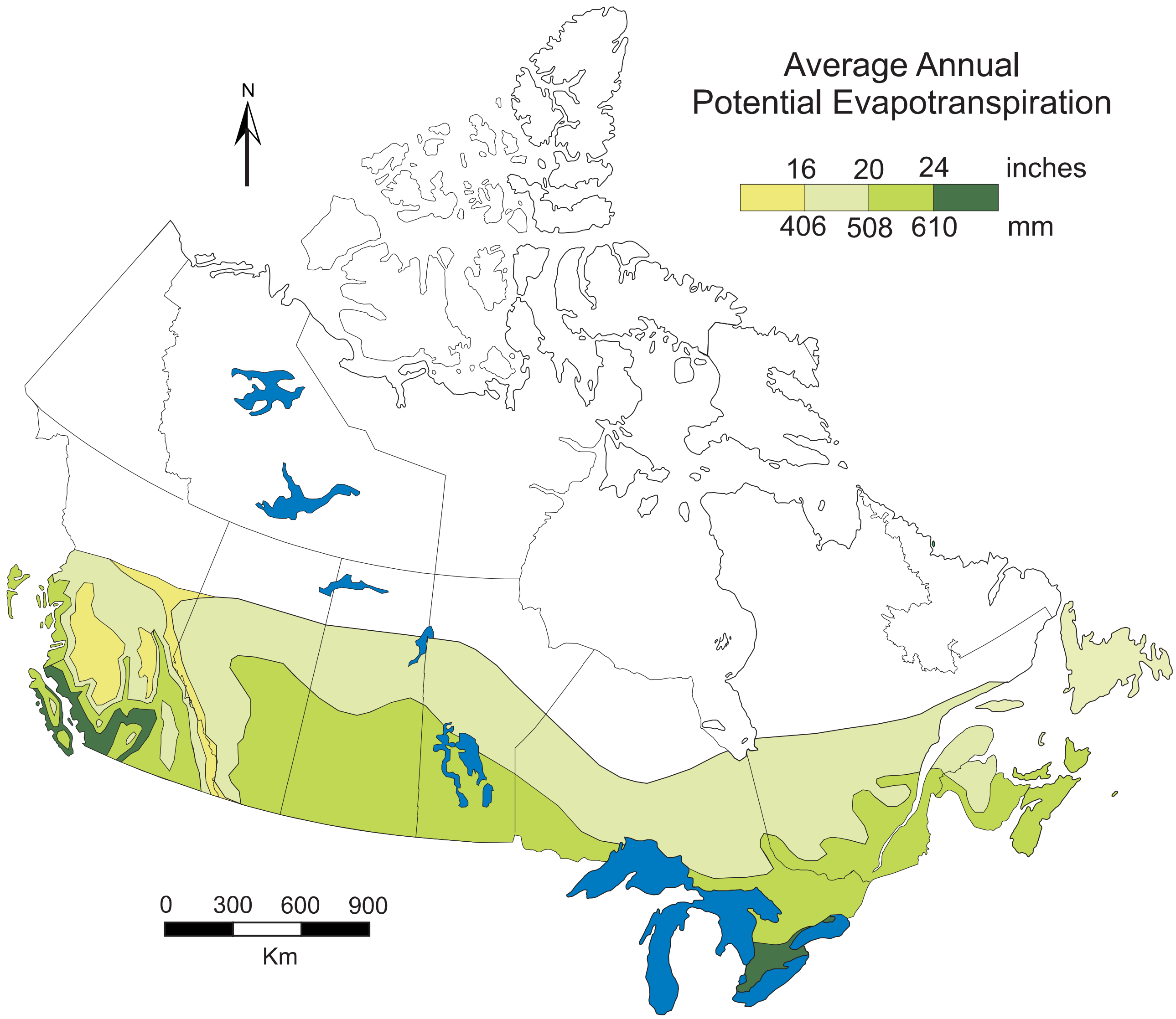


Figure 4.4 Average annual potential evapotranspiration (PET) in millimetres and in inches from ground and plant surfaces for areas where there is a continuous vegetation cover and sufficient soil moisture for plant use (based on National Atlas of Canada, 1974).

and seasonally. Evapotranspiration is closely tied to both precipitation, in terms of moisture availability, and temperature. Evaporation accounts for the movement of water to the air from sources such as the soil, canopy interception, and water bodies. Transpiration accounts for water movement within a plant and subsequent loss of water as vapour through stomata in its leaves. Evapotranspiration is an important part of the water cycle. It is not only closely related to plant growth and carbon uptake but also an important hydrological component affecting runoff, atmospheric circulation, and groundwater recharge and discharge. ET is of key concern in climate change research.

Potential evapotranspiration (PET) is a measure of the atmosphere's ability to remove water from the surface through evaporation and transpiration, assuming an unlimited water supply. Actual evapotranspiration (AET) is the quantity of water actually removed from a surface due to

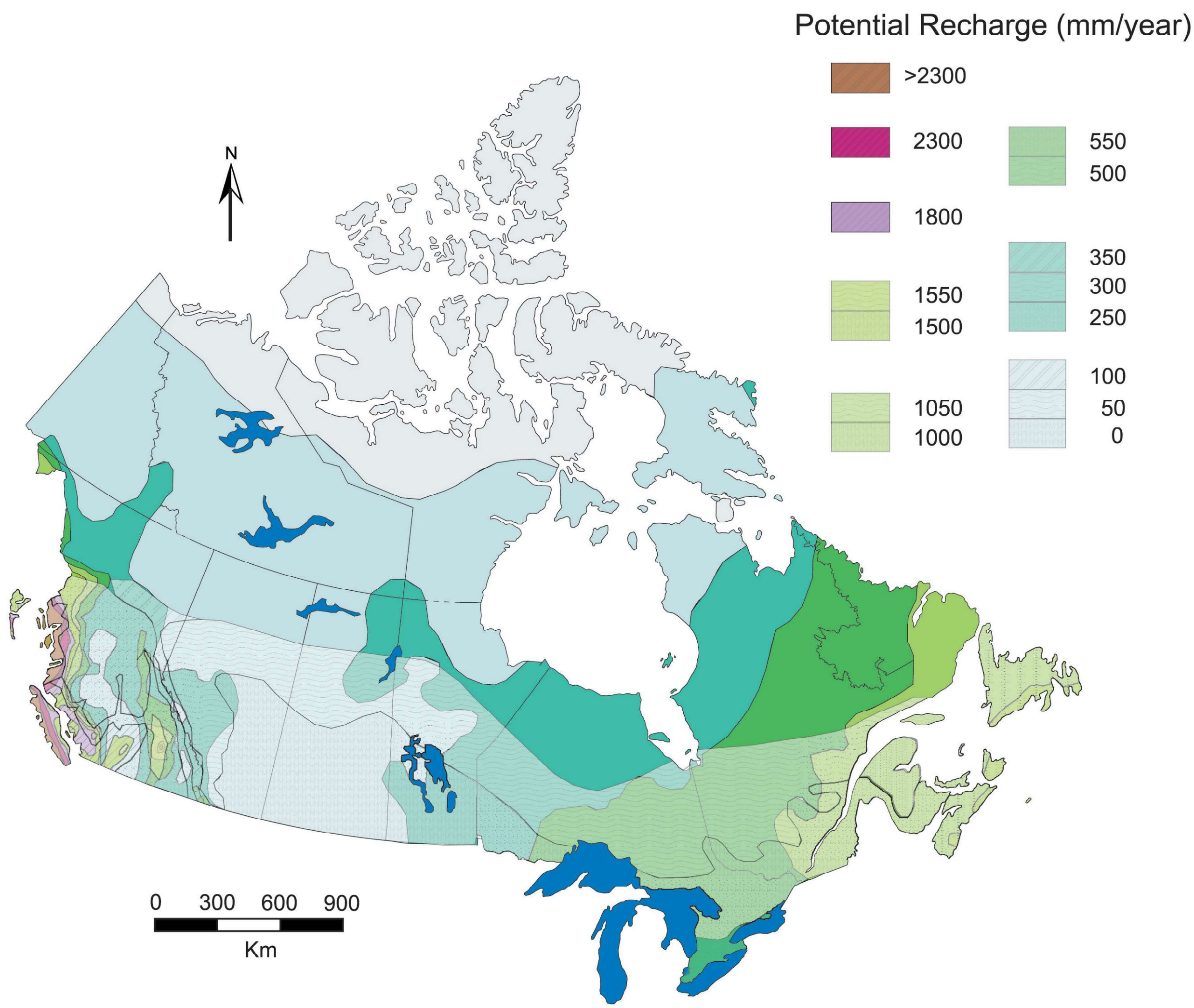
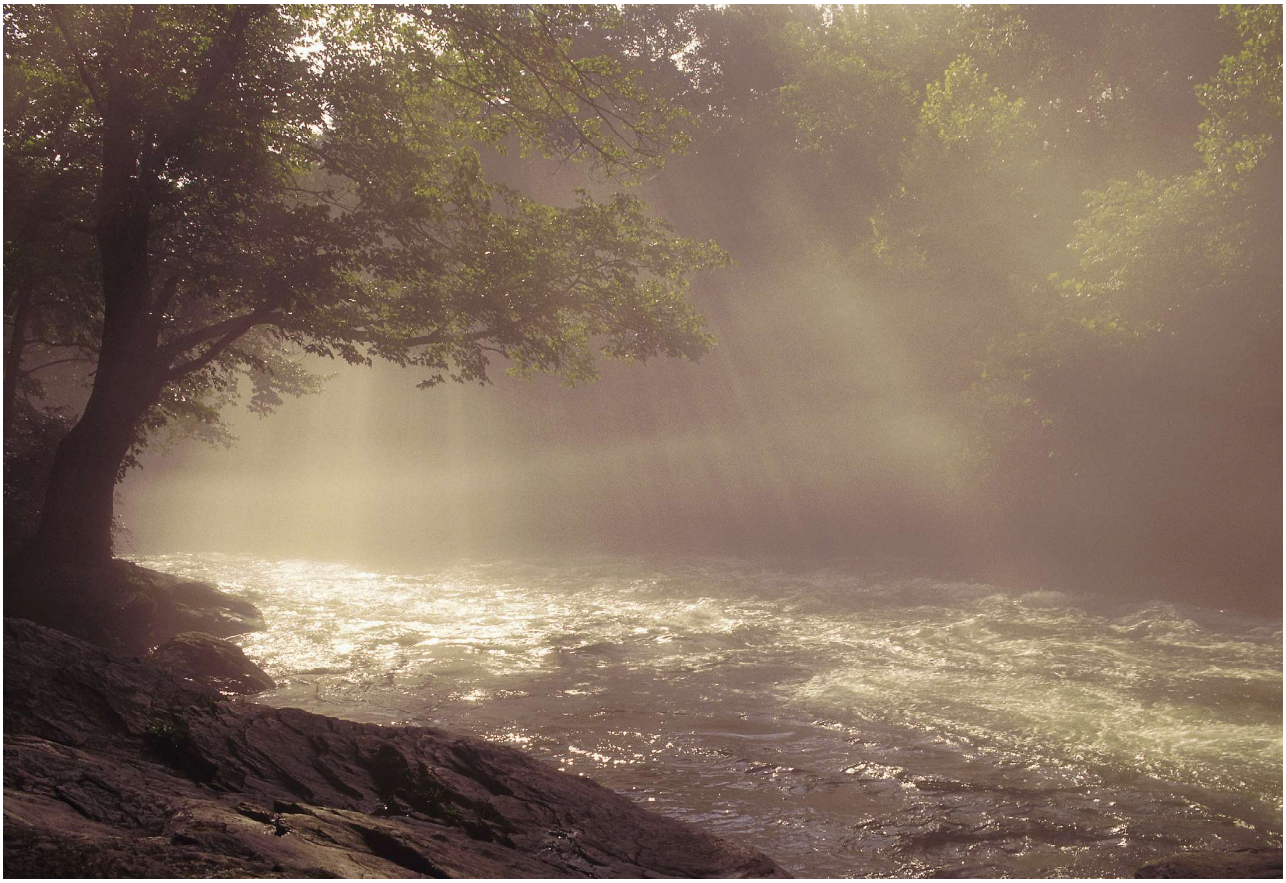


Figure 4.5 Annual potential recharge (precipitation minus potential evapotranspiration) across Canada.

evaporation and transpiration. AET can never be greater than PET, and it can be lower if there is not enough water available to be evaporated, or if plants are unable to readily transpire. Average annual AET cannot exceed average annual precipitation because precipitation sets the limit on how much water is available. Figure 4.4 is an average annual potential evapotranspiration map of Canada.

Potential recharge is the difference between precipitation and potential evapotranspiration, and represents the net amount of water available for groundwater recharge. Not all potential recharge may enter the ground due to runoff. Figure 4.5 is a map of potential recharge constructed by subtracting the mid-range mean annual PET from mean annual precipitation in each Canadian zone. (A more accurate representation of potential recharge would require estimates of AET; however, Fig 4.5 provides a regional picture of how recharge potential might vary across the country.) Potential recharge for Northern Canada is not shown due to lack of potential evapotranspiration data. Recharge potential is expected to be



greater in areas that are moist and that have relatively low evapotranspiration (e.g., West Coast, Kootenays, Rockies, Southern Ontario and St. Lawrence Lowlands, and Maritime provinces). Canada's climate generally favours groundwater recharge (compared to that of other countries around the world), particularly in our coastal regions where moist conditions and relatively low evapotranspiration are common.

Of course, the climate of any region includes variability over time, and these changes invariably affect groundwater recharge (see end of this chapter).

Local surface conditions within a region, including topography, vegetation, soil, and aquifer permeability, also affect recharge. Recharge can vary substantially even when the same climate conditions prevail throughout a region.

4.5.3 Physical and biological controls on infiltration

Infiltration is the process by which water on the ground surface enters the soil. Infiltration is governed by gravity and capillary forces. While smaller pores offer greater resistance to gravity, very small pores will pull water through capillary action in addition to and even against the force of gravity. Infiltration occurs via two different types of mechanisms: piston or translatory flow (wherein precipitation stored in the unsaturated zone is displaced downward by the next infiltration or percolation event without disturbance of the moisture distribution), and preferential flow, in which flow occurs through preferential pathways or macropores (e.g., root channels, animal burrows).

The rate at which water can infiltrate soil depends on a number of factors, including soil texture and structure, presence of preferential pathways, vegetation types and cover, soil temperature, water content of the soil, topography, and rainfall intensity. Coarse-grained sandy soils have large spaces between the grains that allow water to infiltrate quickly. Macropores (large pores) can greatly enhance the permeability of fine-grained soils by forming preferential pathways for water.

Vegetation influences recharge through interception and transpiration, and other less commonly characterized, yet potentially significant, processes such as stemflow and throughfall (Le Maitre et al., 1999; Taniguchi et al., 1996). The vegetation canopy and the top layer of undecomposed leaf litter create porous soils by protecting the soil from pounding rainfall, which can close natural gaps between soil particles. Plant roots also play an important role in the recharge process by enabling plants to draw water from deep in the vadose zone (and even from the saturated zone) and by creating preferential flow paths and channels that aid infiltration (Le Maitre et al, 1999).

When the soil temperature drops to below freezing, a layer of frozen soil can develop below the ground surface. This frozen soil inhibits infiltration, especially if it is so heavily saturated that the pores are filled with ice. Frozen soils throughout most of Canada inhibit water filtration from melting snow, leading to a spring pulse of runoff and streamflow.

When soil becomes saturated during rainfall or snowmelt, its moisture content increases, resulting in a lowering of the soil's capacity to accept infiltrating water. The infiltration process can continue only if there is room available for additional water at the soil surface. Available volume for additional water in the soil depends on the soil's porosity and on the rate at which previously infiltrated water can move away from the surface. The maximum rate at which water can enter soil in any given condition is described as the **infiltration capacity**. When this rate is less than the infiltration capacity, all the water will infiltrate. When rainfall rate exceeds infiltration capacity, surface ponding occurs. Porosity is followed by **runoff** over the ground surface once depression storage is filled. Runoff includes water travelling over land and through small rivulets to reach a stream, and **interflow**, water that infiltrates the soil surface and travels by means of gravity towards a stream channel (always situated above the main groundwater level), eventually empting into that channel. Technically, interflow is not groundwater because it occurs above the water table.

Land surface topography also plays an important role in determining infiltration. Generally, steep terrain favours runoff, and flat terrain favours infiltration. Microtopography can result in depressions that hold water or snow, effectively allowing more time for infiltration to occur because runoff water is contained (see Box 4-2).

Horton (1933) suggested that infiltration capacity declines rapidly during the early part of a storm and then, after a couple of hours, tends towards an approximately constant value for the remainder of the event. Water that had previously infiltrated fills the available storage spaces and reduces the capillary forces drawing water into the pores. Clay particles in the soil may swell as they become wet and thereby reduce pore size. On ground which is not protected by a layer of forest litter, raindrops can detach soil particles from the surface and wash fine particles into surface pores where they can impede infiltration.

Rainfall intensity is perhaps a less obvious control on infiltration. One might expect that the more

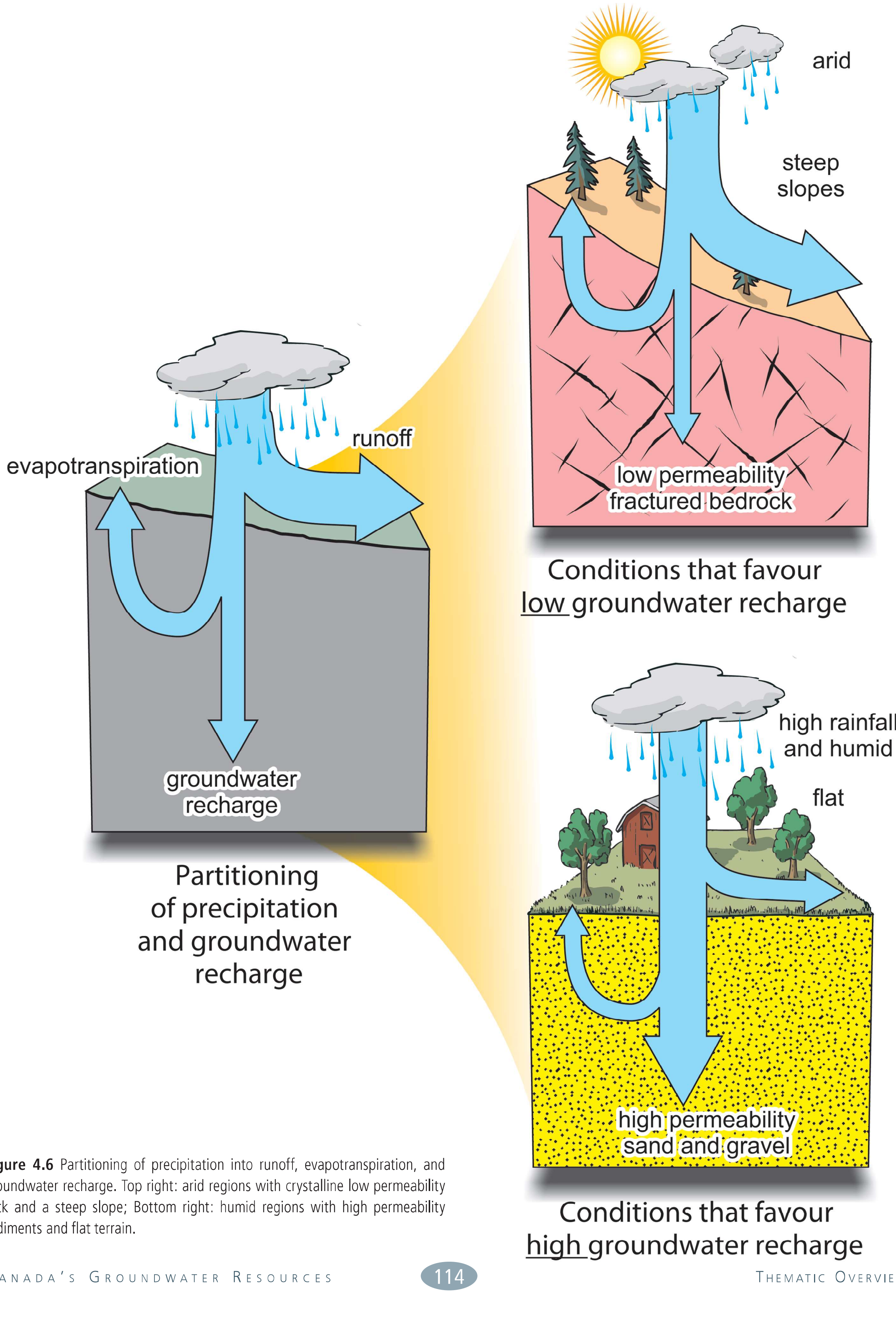


Figure 4.6 Partitioning of precipitation into runoff, evapotranspiration, and groundwater recharge. Top right: arid regions with crystalline low permeability rock and a steep slope; Bottom right: humid regions with high permeability sediments and flat terrain.

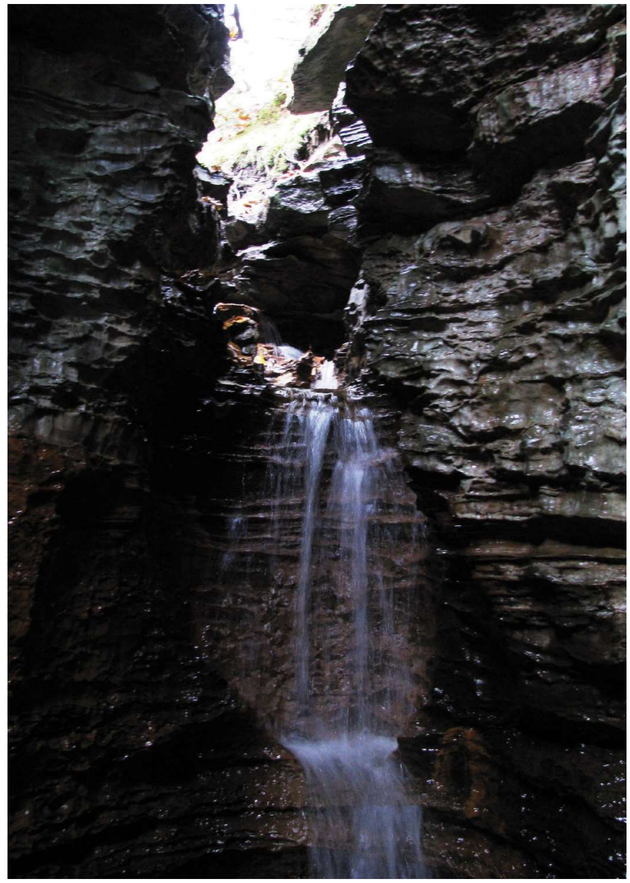
it rains, the more infiltration will occur. Intense rain events, however, often result in more runoff because the precipitation rate exceeds the infiltration rate. Climate change projections suggest that while many areas of Canada will receive more annual precipitation, this precipitation may occur as extreme rainfall events. Thus, projections for increased mean annual or mean monthly rainfall may not result in greater groundwater recharge.

4.5.4 The shallow recharge process

Precipitation landing on the ground can do one of three things: infiltrate, reside as depression storage, or run off. Once water has infiltrated, it may remain as water in soil storage. This water is also available to plants for uptake through their root systems, or it can be lost by evaporation (upward movement of water by capillary action through the soil matrix), or percolate down to the water table, where it becomes groundwater recharge.

The general partitioning of precipitation into a runoff component, evapotranspiration, and groundwater recharge is illustrated in Figure 4.6. The top right figure represents conditions that favour low groundwater recharge, including arid climate, bare soil and/or exposed bedrock, and steep topography. The lower right figure represents conditions that favour high groundwater recharge, including a humid climate with high precipitation, and generally flat topography with sparse vegetation.

The role of vegetation is not explicitly considered in this figure. Forested areas may be areas of high evapotranspiration. Because water transpired through leaves comes from the roots, plants with deep reaching roots can transpire water more constantly, especially during dry periods when the shallow root zone dries out. In dry conditions



herbaceous plants transpire less than woody plants because herbaceous plants lack a deep taproot. Woody plants keep their structure over long winters while herbaceous plants in seasonal climates must grow up from seed contributing little to evapotranspiration during the spring. Factors that affect evapotranspiration include a plant's growth stage or level of maturity, percentage of soil cover, solar radiation, humidity, temperature, and wind. Forests reduce water yield through evapotranspiration.

Landslope and permeability of sediments or rocks lying within the unsaturated zone exercise important controls on the amount and timing of groundwater recharge. Materials with high permeability, such as sand and gravel, favour groundwater recharge: unconfined aquifers comprised of sand and gravel typically have high groundwater

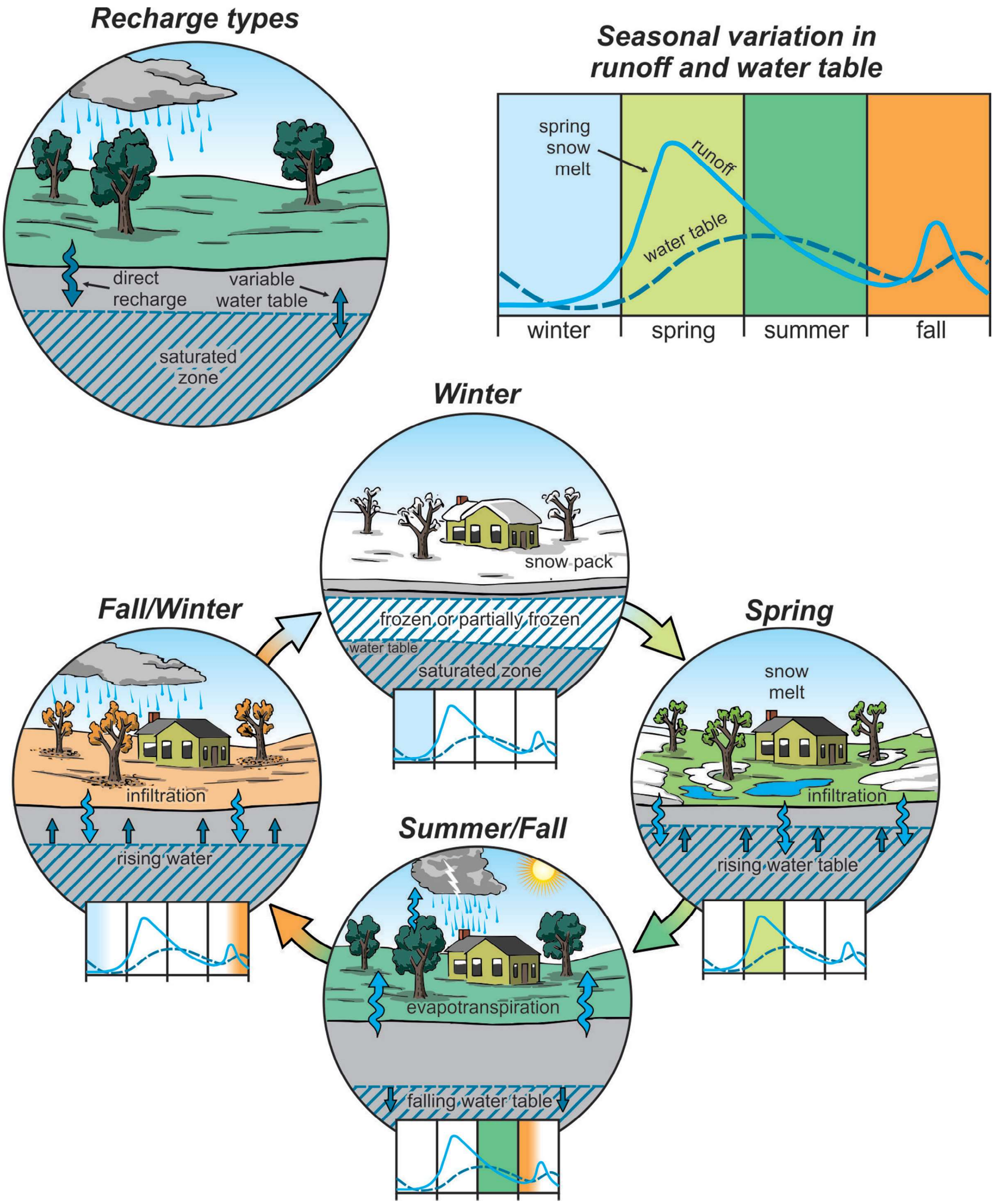


Figure 4.7 Seasonal responses of the water table in response to variable groundwater recharge, or no recharge.

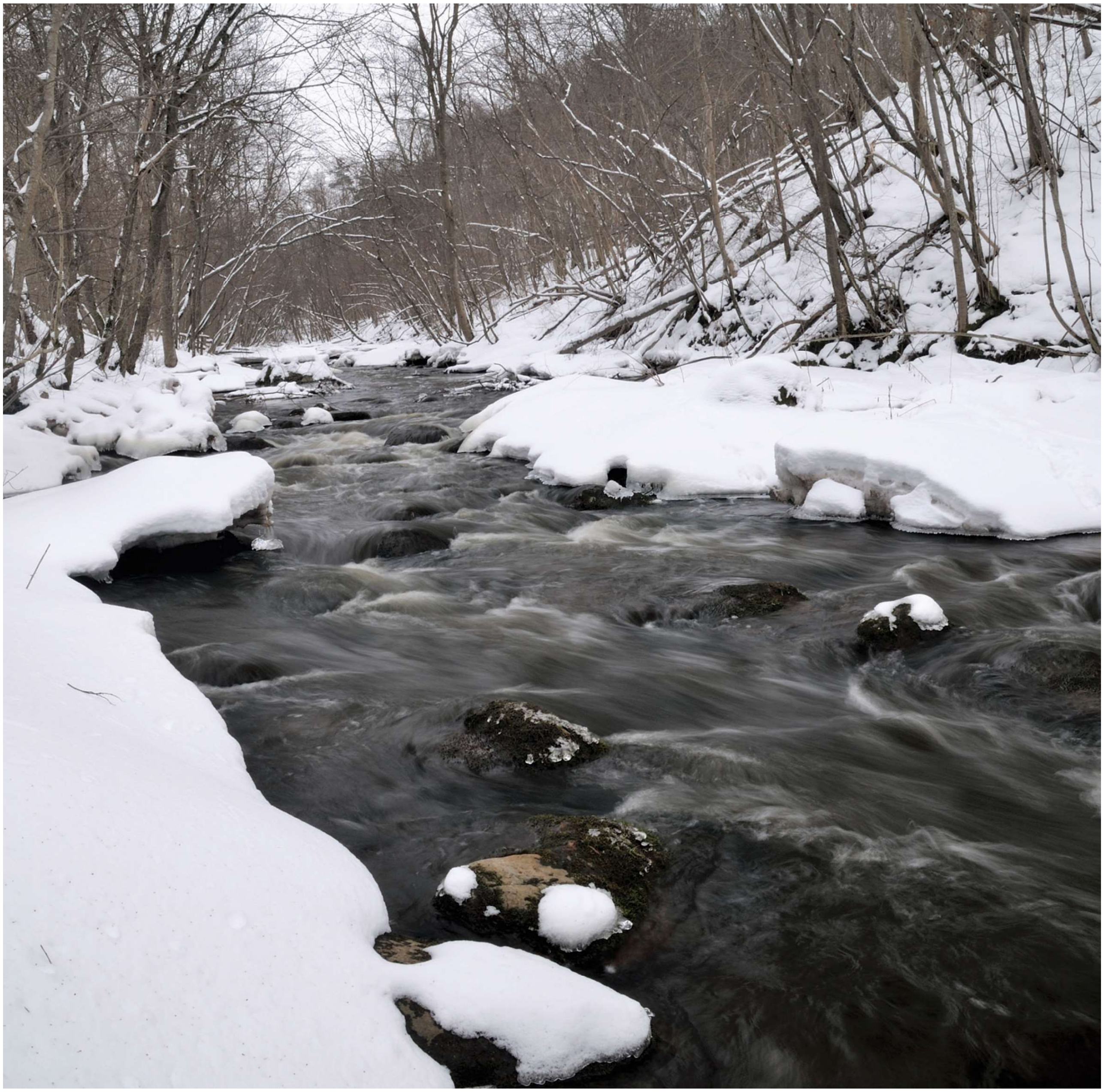
recharge rates. In contrast, materials of low permeability, such as crystalline rock, tend to inhibit groundwater recharge. Consequently, groundwater yields from fractured rock aquifers are generally low and can become unsustainable when pumped for long periods. Fractures and faults, however, have been observed to act as conduits for groundwater recharge and, when tapped by a well, will often yield higher quantities of groundwater when compared to unfractured areas. Should the water table lie within bedrock, infiltrated water will likely move as interflow within the unsaturated zone, ultimately discharging down slope. Bedrock recharge is uncertain in many areas, particularly in mountain regions. Even in these terrains, however, some percipitation can infiltrate the bedrock to recharge groundwater (Smerdon et al., 2009).

Anthropogenic factors, such as roads, buildings, agriculture, or forest harvesting, can impact natural infiltration and groundwater recharge. Urban locations often have reduced infiltration and groundwater recharge: in response, some developers and/or municipalities across the country are experimenting with innovative engineering designs for capturing storm water runoff in collection ponds or through the creation of infiltration galleries. The impacts of agriculture and forestry on groundwater recharge are complex. Removal of natural vegetation for agriculture often leads to greater runoff with soils becoming more vulnerable to erosion and compaction. Agricultural regions often require some form of artificial precipitation, or irrigation, during summer. When potential evapotranspiration is greater than actual precipitation, the soil will dry out, unless irrigation is used. The effect of logging on groundwater is not as wel-understood. Although the removal of trees effectively reduces evapotranspiration losses, thereby promoting groundwater recharge (e.g., Bent, 2001), the loss of the tree canopy and root systems can be expected to bolster increased runoff. Insufficient scientific studies exist on this subject.

4.6 AQUIFER RESPONSES TO INFILTRATION AND RECHARGE

Recharge is usually accompanied by a rise in water level within aquifers. In unconfined aquifers, this rise tends to coincide with higher stream flows or lake levels. Groundwater level peaks, however, are often delayed in comparison to peak levels of surface water bodies. Figure 4.7 provides a conceptual model of an unconfined aquifer illustrating seasonal variations in direct groundwater recharge and the resultant water table response. The model does not represent any specific area of the country, but demonstrates runoff and recharge processes which cause water level variations in aquifers.

Except for those areas with milder climates, most ground in Canada is frozen during the winter, and precipitation falls as snow. Runoff is at a minimum. Water levels in an aquifer, however, may still remain high as a result of the previous year's fall and early winter recharge. This situation continues until spring, when the ground thaws and melt water from snow or spring rain are finally able to infiltrate the soil. Surface runoff typically increases during this time and may even persist into summer depending on the runoff source. Snow pack and/or glaciers at high elevation, for example, melt late in spring and deliver water via stream runoff to lower elevations well into summer. Most of Canada is warm during the summer, and evapotranspiration rates are high. Runoff is generally low, apart from the occasional summer precipitation event. These conditions produce a gradual decline in the water table, which persists into the fall. High spring-summer water



demands in Canada for plant growth, combined with higher evaporative losses from soil and surface water bodies, result in declining water levels. The fall usually brings somewhat wetter conditions to most parts of the country, with consequent small increases in runoff and water table levels. In colder regions, this fall precipitation may occur as snow.

The magnitude and timing of increasing water table levels vary from year to year, depending on the climate and anthropogenic factors. Some years are simply wetter than others, and some are colder. Therefore, groundwater levels, as depicted in a well hydrograph, vary seasonally from year to year when viewed over the long term.

Water table elevation also varies on short time scales although these variations depend on recharge nature, aquifer permeability, depth to the water table, and amount of available storage in the aquifer. Long and moderate-volume precipitation events tend to result in greater overall recharge as compared to intense, short-duration rain events. A comparison between two aquifers of the same material reveals that the one with the deeper water table will experience a delay in water level response due to deeper water level. High-permeability materials are able to transmit water more readily than low-permeability materials and, consequently, the response is more rapid. Aquifer storage availabilty also acts to mediate the response, as aquifer materials with high storage capacities have a much smaller overall change in water level following a recharge event. Consewuently, some aquifers record precipitation events very effectively, showing high frequency variation. Most aquifers, however, only record long-term water level changes accompanying seasonal precipitation variations. The recharge response of confined aquifers tends to be less

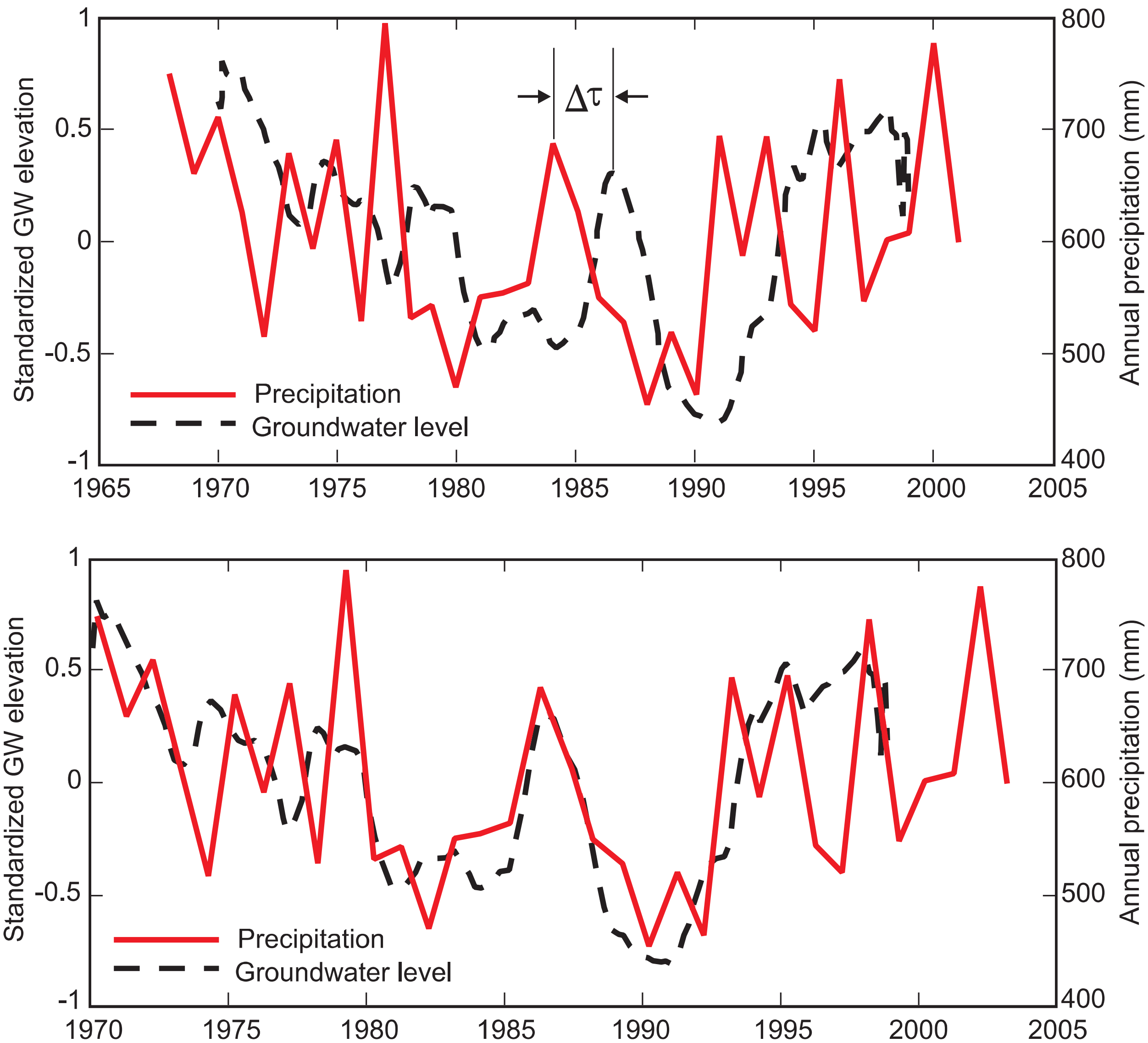


Figure 4.8 Annual precipitation and average standardized groundwater levels in 24 monitoring wells in the Winnipeg area.

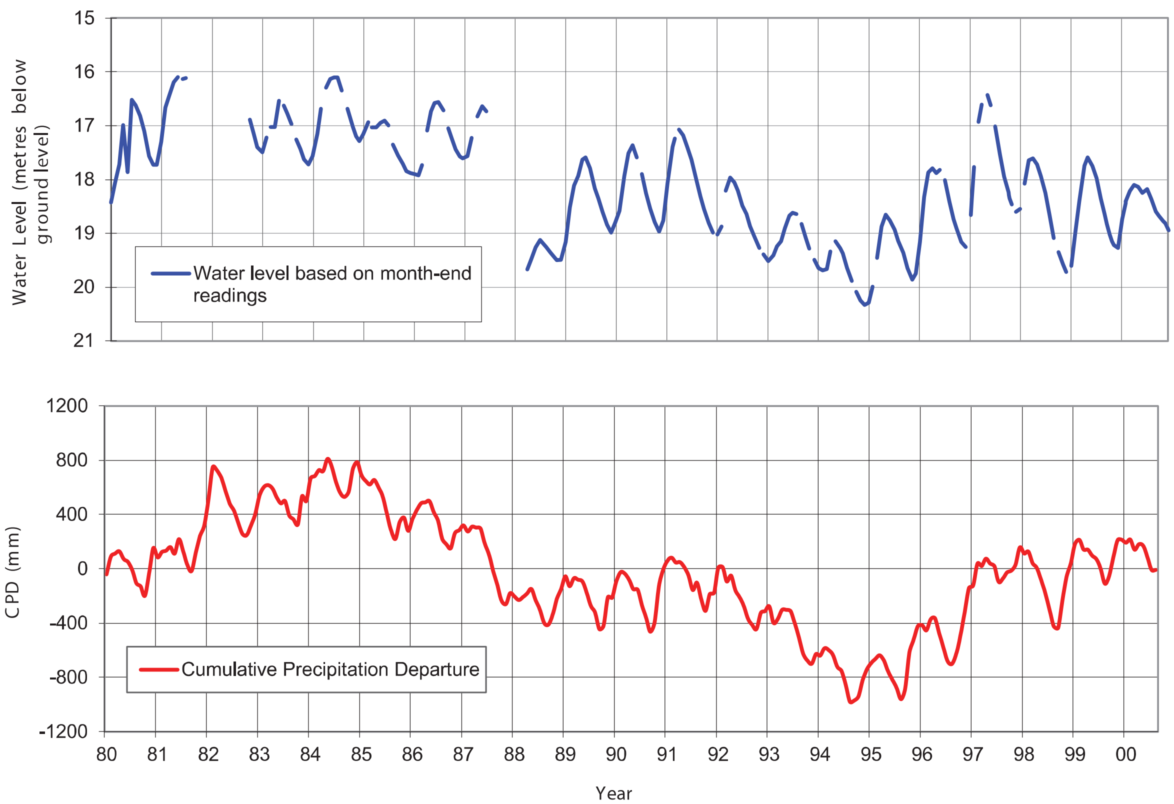


Figure 4.9 Aquifer water level and cumulative precipitation departure (CPD) for Abbotsford, British Columbia. Trends in groundwater level in this highly permeable aquifer are linked to precipitation trends as evidenced by the strong correlation between groundwater level and CPD prior to 1992. These trends are believed to be associated with phases of the El Niño Southern Oscillation (ENSO), namely, El Niño and La Niña (Fleming and Quilty, 2006). Water levels in the aquifer do not appear to correlate as well after 1990, suggesting some other factors, such as anthropogenic influences.

dramatic than that of unconfined aquifers. This is because the former are generally isolated from the near surface climate effect, or subject to recharge in distant areas. Confined aquifers, do, however, exhibit a strong response to barometric pressure variations and tides (Spane, 2002).

Two examples are provided to illustrate aquifer response to precipitation. Figure 4.8 charts annual precipitation (solid red) at the Winnipeg International Airport and standardized average water level (dashed black) calculated from 24 groundwater monitoring wells in the Winnipeg area. Standardizing groundwater levels is accomplished by determining the average water level and plotting the deviation from that water level at each measurement point. The values are referenced to zero deviation. The upper graph in Figure 4.8 illustrates the relationship between the variation of annual precipitation and generalized groundwater response. The groundwater response to recharge has a time delay (Δt) of approximately 2.2 years. When the annual precipitation graph is shifted backward by 2.2 years, the two curves correlate more closely (see lower graph). Groundwater flow systems have different abilities to retain and transport water, and goundwater residence time can vary from days to tens of thousands of years.

Figure 4.9 shows water level variation and cumulative precipitation departure (CPD) for a provincial observation well in Abbotsford, British Columbia. Cumulative precipitation departure can be used to assess water level fluctuations in observation wells completed in shallow unconfined aquifers. The CPD method involves calculating the difference between monthly precipitation and the mean monthly precipitation for a given historic period. A strong correlation between a CPD curve and a hydrograph indicates that precipitation has a major influence on water level at that well. A poor correlation indicates that the water level is controlled by another factor, such as a nearby river or anthropogenic influences (e.g., overwithdrawal). CPD is established for an arbitrary reference date and, therefore, comparison of different aquifers is possible only if consistent dates are used. For example, groundwater levels in Abbotsford are shown to correlate with the CPD in years prior to 1970. After 1970, the CPD curve and groundwater levels begin to diverge, with groundwater levels becoming lower up to 1976, and then rising to the end of the period of record. The reason for these differences in trends is uncertain, but could be related to anthropogenic causes such as pumping up to 1976, followed by intensive irrigation application as this area of the Lower Fraser Valley has intensive agriculture, particularly raspberry production.

Thus, while changes in groundwater level, in terms of year to year and season to season variability, are most commonly associated with relatively short-term variability in precipitation and temperature, groundwater levels also respond to anthropogenic and other influences. Most notable is the response of groundwater level to pumping. During the summer months it is not uncommon for groundwater levels to decline because of groundwater use. If the rate of pumping is sustainable, then groundwater levels will generally recover to levels representative of natural recharge conditions. The topic of groundwater sustainability is discussed in more detail in Chapter 6; however, it is

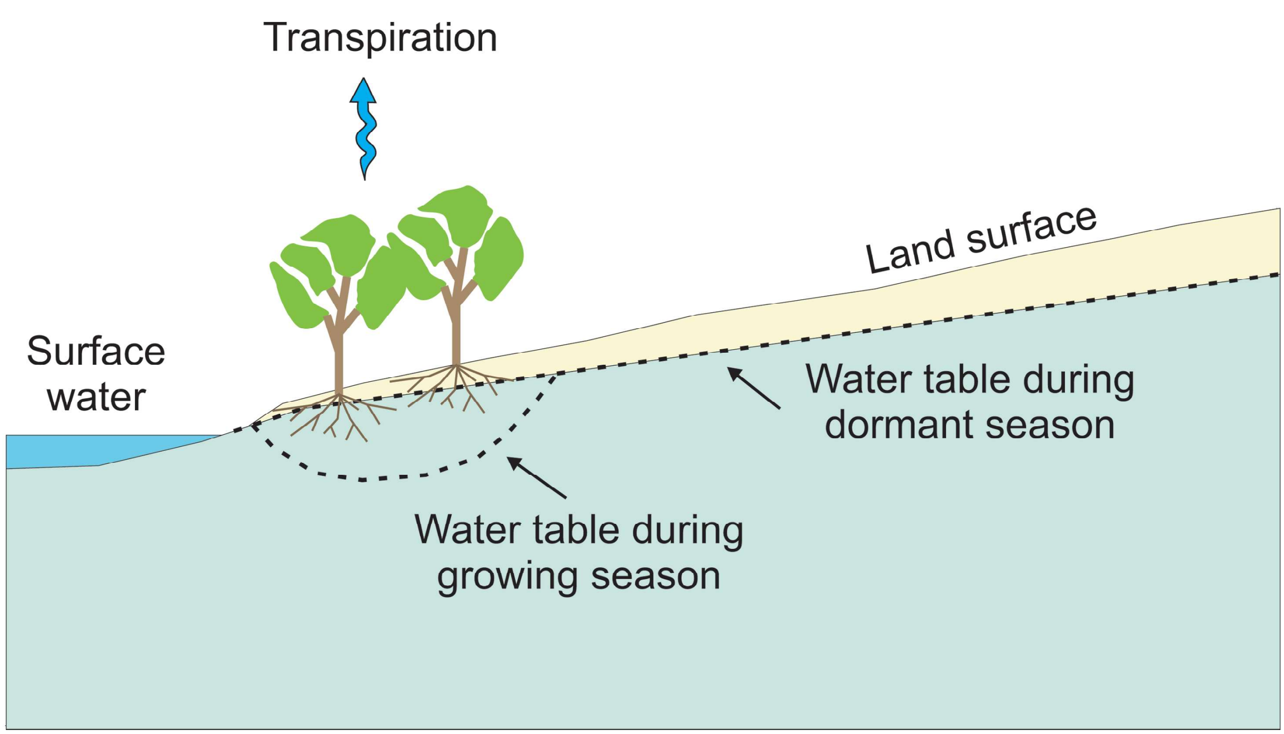


Figure 4.10 The effect of plants on the position of the water table (modified from United States Geological Survey, 2006).

worth mentioning here that some methods used to estimate recharge rely on groundwater fluctuation records which show water level rise and decline. When anthropogenic factors come into play, the natural system becomes difficult to interpret.

One more factor that can lead to natural shifts in groundwater level is vegetation demand. Vigorous vegetation growth during the summer months leads to higher transpiration losses. When the water table is below the depth of the plant's roots, plants will be dependent on water supplied by precipitation or irrigation. Figure 4.10 shows that where the water table is near the land surface (close to lakes and streams) roots can penetrate into the saturated zone below the water table, allowing transpiration directly from the groundwater system. Transpiration of groundwater commonly results in a drawdown of the water table much like the effect of a pumping well (cone of depression—the dotted line surrounding the plant roots in Figure 4.10) Daily fluctuations in groundwater level have also been observed to the extent that water levels vary by over one metre within 24 hours in response to the opening and closing of leaf stomata (Meyboom, 1966).

4.7 METHODS FOR ESTIMATING RECHARGE

There are as many methods available for quantifying recharge as there are different sources and processes of recharge. Each of these methods has its own limitations in terms of applicability and reliability. Recharge mechanisms, rates, and patterns in porous media are reasonably well studied and understood (Allison, 1988; Scanlon et al., 2002), but constraining and quantifying recharge rates and patterns in fractured rock systems are less studied, and generally less understood (Cook and Robinson, 2002; Scanlon et al, 2002).

The objective of any recharge study should be identified prior to the selection of an appropriate study method as this objective may dictate the required space and time scales of the recharge estimates (Scanlon et al., 2002). Groundwater resource evaluations, for example, require more information on recharge at large spatial and temporal scales, whereas assessments of aquifer vulnerability to pollution require more detailed information at local and shorter time scales.

Recharge estimation procedures are usually classified into three main categories: 1) physical—involving the direct determination of soil moisture flux, or catchment water flux, 2) chemical—utilizing natural solutes, tracers, and isotopes, and 3) modelling—often a combination of approaches providing greater confidence in the resultant recharge estimates.

Most recharge estimation methods are based on using measured or estimated values for some components of the water balance equation (Eq. 4.1) in order to predict other missing components. Determination of water balances can be carried out annually at the catchment (or aquifer) scale, or within a soil column located in the unsaturated zone. The various components of water balance equations are measured in millimetres per year. Freeze and Cherry (1979) calculate the water balance equation for a catchment as:

$$P = Q + ET + \Delta S_W + \Delta S_S + \Delta S_G \qquad (4.1)$$

where P is precipitation input, Q is the sum of the surface and groundwater (unsaturated and saturated) runoff, ET is evapotranspiration, ΔS_W is the change in storage of surface water, ΔS_S is the change in the soil moisture or change in storage in the unsaturated zone, and ΔS_G is the change in storage in the saturated groundwater zone. This equation assumes that the catchment is closed, and that there is no net inflow or outflow of groundwater to or from other catchments. Total change in storage is calculated as:

$$\Delta S = \Delta S_W + \Delta S_S + \Delta S_G \qquad (4.2)$$

Assuming no long-term climate change effects or groundwater and/or surface water mining, the changes in storage should approach zero in a natural system ($\Delta S = 0$). Therefore, Equation 4.1 can be simplified to:

$$P = Q + ET \qquad (4.3)$$

Equation 4.1 can be subdivided further to describe differences in the recharge and discharge areas. Thus, in the recharge area:

$$P = Q_W + Q_S + R + ET_R \qquad (4.4)$$

where Q_w is the surface water component of average annual runoff (the portion that does not infiltrate the soil), Q_s is the component of runoff that moves laterally through the unsaturated zone, R is the average annual recharge (water that percolates into the saturated zone), and ET_R is the average annual evapotranspiration in the recharge area.

Rearranging this equation gives an expression for groundwater recharge:

$$R = P - ET_R - Q_W - Q_S \qquad (4.5)$$

The total amount of runoff in the discharge area can be calculated by:

$$Q = Q_W + Q_S + D - ET_D \quad (4.6)$$

where D is the average annual discharge, and ET_D is the evapotranspiration in the discharge area (note that the sign of ET_D is negative). P is negligible in the discharge area compared to the other terms as discharge areas normally constitute a small percentage of the catchment area.

The change in groundwater storage is zero under natural steady-state conditions, thereby balancing recharge R and discharge D:

$$\Delta S_G = 0 = R - D \quad (4.7)$$

Short-term changes in storage ΔS_G can be realized, however, during spring and fall in many regions of Canada, as groundwater recharge occurs in response to spring melt or significant rainfall. Groundwater levels rise and the change in groundwater storage remains positive for some period of time. This rise results in a pulse of groundwater flow that gradually moves through the system, eventually discharging into streams, rivers, lakes, or the ocean. Annually, or perhaps, on average, over several years, the amount of recharge generally equals the amount of discharge, otherwise we would see a gradual rise or fall of water levels in the aquifer. Over time, the amount of groundwater held in storage usually remains constant. Of course, unsustainable pumping can lead to gradual water table declines and a loss of groundwater from storage. Irrigation over long periods may lead to a rise in the water table (water table mounding). This is becoming a significant problem in many arid regions around the world and is linked to groundwater salinization. Climate change may also alter the long-term stability of groundwater recharge and discharge dynamics.

The use of catchment water balance approaches for determining groundwater recharge requires careful measurements of all the various components.

4.7.1 Physical methods

Physical methods for estimating recharge all seek to quantify recharge directly from precipitation. Many rely on measuring components of the water balance equations above to determine recharge. In soil moisture methods, for example, recharge is estimated from the soil profile when losses to evapotranspiration (ET) are subtracted from the precipitation to give an **effective precipitation** value (e.g., Penman, 1948; Grindley, 1967). Soil moisture methods, therefore, rely on accurate measurements of actual evapotranspiration (AET) (Rushton and Ward, 1979), using instruments such as lysimeters, neutron probes, and time-domain reflectometry (TDR), which measure soil moisture content (Howard and Lloyd, 1979). Alternatively, actual evapotranspiration can be calculated as a fraction of potential evapotranspiration (PET), which is calculated from meteorological data using various models (Ragab et al., 1997).

The water balance approach is also used to estimate recharge for a watershed or catchment. Commonly, recharge is determined based on the assumption that the amount of baseflow in a stream is equal to the amount of recharge to the catchment. When the stream is at its lowest level (baseflow only), the surface runoff component is assumed to be zero, so that all the water in the stream is sourced from groundwater. Baseflow represents groundwater being discharged from aquifer storage (although the assumption may not be true for short time periods). Furthermore, the

A **lysimeter** is a measuring device, which can be used to measure the amount of actual evapotranspiration. By recording the amount of precipitation that an area receives and the amount lost through the soil, the amount of water lost to evapotranspiration can be calculated. Lysimeters are of two types: weighing and non-weighing. For a weighing type, the lysimeter consists of a buried container of soil equipped with a weighing device and a drainage system to measure evapotranspiration and percolation. The idea is that precipitation is measured locally, and the drainage through the lysimeter and volume of water stored in the lysimeter are measured by weighing periodically. The amount of water lost by evapotranspiration can be worked out by calculating the difference between the weight before and after the precipitation input.

A **neutron probe** is a device used to measure the quantity of water present in soil. A typical neutron probe contains a pellet of americium-241 and beryllium. Americium-241 is unstable and decays by alpha particle emission, with a by-product of gamma rays. In the neutron probe, the alpha particles collide with the light beryllium nuclei, producing fast neutrons, which then collide with hydrogen nuclei present in the soil as water molecules. During this collision process, they lose much of their energy. The detection of slow neutrons returning to the probe allows an estimate of the amount of hydrogen present. Since water contains two atoms of hydrogen per molecule, this therefore gives a measure of soil moisture.

Time-domain reflectometry (TDR) is an electronic instrument used to estimate soil water content. A very fast step voltage is introduced into a probe that is inserted into the soil or other porous medium. The velocity at which the pulse travels is related to the dielectric constant (characterizing the ability to store rather than to conduct energy) and this is a function of the soil moisture content. The technique can be used in combination with lysimeters and neutron probes to estimate evaporation from soils if measurements are taken at different intervals.

water balance method is based on identifying the groundwater recession portion of a streamflow hydrograph, using hydrograph separation techniques (e.g., Mau and Winter, 1997; Hannula et al., 2003; Halford and Mayer, 2000; Cherkauer and Ansari, 2005; Rutledge, 2007; Lim et al., 2005).

Catchment scale water balances are also typically based on assumptions that surface water and groundwater divides (imaginary boundaries across which water does not flow) coincide, which does not happen in many regions. Divides, in a surface water system, are often identified by the stream network, and the network of streams draining into a single stream defines the catchment area. In some cases, the groundwater catchment boundary coincides with the surface water catchment, although groundwater often moves more regionally, passing from one catchment to the next. In this case, there may be a net gain of groundwater to and a net loss of groundwater from the catchment. Some underlying assumptions for these catchment-scale water balance methods include: 1) the stream fully penetrates the homogeneous and isotropic aquifer; 2) the recharge is uniform over the aquifer; 3) the aquifer is underlain by impermeable rock; 4) there are no groundwater losses from evapotranspiration; and 5) there are no upstream flow diversions or flow controls.

Groundwater recharge can also be estimated using records of water table fluctuations (e.g., Healy and Cook, 2002). This procedure is referred to as a water table fluctuation method or well-hydrograph method, and is commonly used (Freeze and Cherry, 1979) with varying degrees of success (Sophocleous, 1985). The onset of a rain event causes a rapid rise in water table elevation to a peak level, followed by a steep recessional curve towards a new equilibrium value (Mew et al., 1997). The rate of elevation change in the water table during a groundwater recharge period is a function of groundwater recharge and the aquifer's specific yield (Sy). Kazman (1988) and Sophocleous (1985) point out that there is no precise correlation between a change in water table elevation and rainfall. Although levels rise during most rainfall events, they will not always produce the same water level change within a particular aquifer. This is because the specific yield may vary depending on moisture content, depth to the water table, and the rate at which these parameters change (Nachabe et al., 2005). A constant value of specific yield is usually employed in this procedure, although this can lead to overestimation of recharge (Sophocleous, 1985).

Johansson (1988) demonstrated that when the water table is deep enough, equilibrium water content develops in the upper portion of the soil profile. As a result the actual specific yield will be close to a constant value and will overestimate less. Regardless of the potential for changes in Sy, this particular parameter is often difficult to estimate with confidence, leading to uncertain recharge estimates.

4.7.2 Chemical methods

Estimation of groundwater recharge is often carried out by using solute or isotopic tracers to estimate vertical water movement in the subsurface. Hydrograph separation methods for determining baseflow and stormflow components can also be used. Uncertainties exist in all of these procedures, particularly related to the groundwater sampling.

Recharge estimations can be based on the concentration, and spatial and temporal distribution of natural tracers, such as tritium (^{3}H), the ratio of helium-3 to tritium ($^{3}He/^{3}H$), chlorine-36 (^{36}Cl), chloride (Cl), and gasses, such as chlorofluorocarbons (CFCs), that are introduced into the groundwater system through precipitation (Cook and Solomon, 1997; Cook and Bohlke, 2000). A multitude of other tracers artificially introduced into the ground may also be used. The rate of tracer movement is directly related to the rate of groundwater movement. Measurements of these tracer concentrations in both unsaturated and saturated zones may allow for direct groundwater dating, and identification of different sources for groundwater recharge (Clark and Fritz, 1997).

Many studies use a method based on chloride mass balance to estimate recharge. Tracer studies in the unsaturated zone utilize the position of the tracer peak, the shape of the tracer profile through the soil, and the total tracer concentration to estimate recharge (e.g., Sharma and Hughes, 1985; Scanlon et al., 2007). Recharge estimates in the saturated zone can be made by measuring concentration of chloride in groundwater if the amount of chloride deposited by the atmosphere is known. This determination assumes that chloride is conservative and is deposited in both wet and dry periods. A portion of the chloride will be recharged to the groundwater by percolation; therefore, the amount of recharge can be calculated by dividing the annual amount of wet and dry chloride deposition by the average chloride concentration of the groundwater. Net infiltration rates are then estimated from measured chloride concentrations using the relationship:

$$I = (P \times C_0) \times C_s \tag{4.8}$$

where I is average net infiltration (mm/year); P is average annual precipitation (mm/year); C_0 is the effective average Cl concentration in precipitation (mg/L), including the contribution from dry fallout; and C_s is the measured Cl concentration in subsurface water (mg/L), which can be pore water, perched water, or groundwater. This calculation provides an actual groundwater recharge mean when several groundwater samples are used (Johansson, 1988). Although the method has been used successfully around the world, particularly in arid areas (Beekman and Xu, 2003), it cannot be used in areas where there are chloride sources other than precipitation (e.g., saltwater intrusion, saline soils). For those locations where chloride data is lacking, specific conductance data collected at stream-gauging sites can be used as a proxy.

Radioactive isotopes such as ^{3}H and ^{36}Cl are also useful for groundwater recharge studies. These isotopes decay, leading to lower and lower concentrations over time. Nuclear testing in the mid-20th century increased atmospheric isotope levels, resulting in precipitation which contained elevated isotope concentrations. For example, the peak tritium concentration in rainfall occurred during 1963. Recognition of this peak in an unsaturated soil column can be used to determine the rate of recharge. Ratios of ^{3}He/^{3}H can also be used to determine recharge rates. ^{3}He is the daughter product of ^{3}H decay, and, in the unsaturated zone, is lost to the atmosphere. When ^{3}H decays in the saturated zone, ^{3}He is isolated from the atmosphere and its concentration increases as the groundwater becomes older (Cook and Solomon, 1997). The ^{3}He/^{3}H ratio method is advantageous when the unsaturated zone is too thin and infiltration rates too high, resulting in the peak not being ovserved. Additionally, this dating method, similar to others (using carbon-14, krypton-81 and chlorine-36), provides groundwater ages considered to be "apparent ages" because parcels of groundwater with different ages are frequently mixed in aquifers, often causing significant uncertainty and non-uniqueness in models of groundwater transport and mixing. The number and accuracy of groundwater age dating methods have dramatically increased within the last two decades, and groundwater age dating is now a fairly mainstream hydrogeological tool.

Chlorofluorocarbons or CFCs have long atmospheric residence time and their atmospheric concentrations are spatially uniform and fairly well known (Plummer and Busenberg, 2000). CFC concentrations peaked in the late 1990s, and have been decreasing since then. Measured CFC concentrations in groundwater can be compared with their atmospheric concentration to obtain an apparent CFC age and, therefore, an apparent recharge rate.

Separation of the baseflow and stormflow components of a stream hydrograph can also be accomplished using chemical methods. The basic principle is that the older water is groundwater stored in the catchment prior to a rainfall event, while the new water is that added during a particular event. The concentration of one or more tracers, such as chloride, oxygen-18 or deuterium, is calculated at many points throughout the hydrograph period, and the release of each component determined from the total discharge coupled with initial and current concentrations. The hydrograph can then be separated into each of its components: baseflow, new stormflow and old stormflow at different times.

4.7.3 Modelling approaches

The third group of recharge estimation methods involves modelling. Modelling can be done in a direct sense, whereby recharge estimates are determined through calculations involving a number of climate and soil input parameters. Or it can be done indirectly, using methods for modelling recharge which require independent means to verify the results (Hill and Tiedeman, 2007). Sometimes this involves using other models, or it may be done by model calibration to field measurements. Inverse modelling methods involve varying the recharge rates to reproduce a set of field observations (e.g., Howard and Lloyd, 1979; Rushton and Ward, 1979). Such procedures often lead to non-unique recharge estimates because different parameter combinations can lead to the same result.

The modelling methods are based on equations generally solved numerically with a computer. Some of these methods are based on a simple water balance equation (Eq. 4.1), while others rely on partial differential equations, such as Richards' Equation (see Gardner, 1972). A growing number of computer codes are available for estimating recharge, each with varying degrees of sophistication. Some simple models are one-dimensional, considering only vertical water movement within a soil column. The Hydrologic Evaluation of Landfill Performance (HELP) model (Schroeder et al., 1994), for example, uses a water balance approach to determine net recharge at the base of a soil column. This code has been used in a number of Canadian studies (e.g., Box 4-1 and Box 4-3), and has been linked with groundwater flow models to predict changes in groundwater levels and baseflow. Research is ongoing to combine climate data, land surface/hydrology models, and groundwater models, at ever increasing scales.

4.8 REGIONAL VARIATIONS IN RECHARGE

Groundwater recharge across Canada is not uniform because of wide varities in land surface conditions and climate. Neither is recharge constant at any particular location. Nevertheless, a description of the general character of recharge for each of Canada's hydrogeological regions is possible (see Chapter 8).

Table 4.2 charts mean annual precipitation, estimated mean annual recharge, and method of estimation for several aquifers across the country. Recharge estimates were generally based on field measurements or modelling. Figure 4.11 depicts these results graphically, categorized according to hydrogeological region.

Below are the summaries of the climate and recharge conditions for each region.

4.8.1 Cordillera

Cordillera climate varies from mild, humid conditions along the southwest coast to subarctic conditions in high mountains and in the north. Climate is dominated by Pacific Ocean air masses and orographic (mountain) effects. Precipitation can exceed 4,000 mm along the Pacific coast, although locally it can be as low as 600 mm (Gulf Islands). Eastward from the coast, annual precipitation decreases from 1,200 to 1,500 mm in

TABLE 4.2 MEAN ANNUAL PRECIPITATION AND ESTIMATED GROUNDWATER RECHARGE FOR DIFFERENT STUDY REGIONS ACROSS CANADA

LOCATION	HYDROGEOLOGICAL REGION	UNCONFINED AQUIFER TYPE (S-SURFICIAL) (B-BEDROCK)		MEAN ANNUAL PRECIP. AT NEAREST STATION	MEAN ANNUAL RE-CHARGE	RECHARGE AS % OF PRECIP.	METHOD USED
		S	B	MM/YR	MM/YR		
Grand Forks, BC[1]	Cordillera	X		510	81	16%	Recharge modelling
Abbotsford, BC[2]	Cordillera	X		1573	1018	65%	Recharge modelling
Gulf Islands, BC[3]	Cordillera		X	883	137	16%	Recharge modelling
Oliver, BC[4]	Cordillera	X		328	60	18%	Recharge modelling
Belcarra, BC[5]	Cordillera		X	2331	757	32%	Hydrograph analysis
Northern Alberta, AB[6]	Plains	x		424	45	11%	Modelling and observations
Prairies, SK[7]	Plains	X		350	2		Field measurement
Dalmeny Aquifer, SK[8]	Plains	X		350	5		Flow rates from springs
Sandilands, MB[9]	Plains	X		610-539	217	36%-40%	CFC age dating, average of three sandy sites
Toronto-Oak Ridges Moraine, ON[10]	Southern Ontario Lowlands	X		886	60-210	7%-24%	Water balance and numerical 3D model
Grand River Basin, ON[11]	Southern Ontario Lowlands	X		950	200	21%	Recharge modelling
Ottawa, ON[12]	St. Lawrence Lowlands		X	914	238	26%	Stream hydrograph analysis
Chateauguay, QC[13]	St. Lawrence Lowlands		X	956	103	11%	Recharge modelling
Mirabel, QC[14]	St. Lawrence Lowlands		X	1065	70	7%	Inverse modelling
Cote Nord, QC[12]	Precambrian Shield		X	1009	428	42%	Stream hydrograph analysis
Portneuf, QC[12, 15]	St. Lawrence Lowlands		X	1056	244-426	20%-40%	Stream hydrograph analysis
Rimouski, QC[12]	Appalachians		X	972	302	31%	Stream hydrograph analysis
Sherbrooke, QC[12]	Appalachians		X	1238	327	26%	Stream hydrograph analysis
Saguenay, QC[12]	Shield		X	1116	119	11%	Stream hydrograph analysis
Trois Rivieres, QC[12]	St. Lawrence Lowlands		X	1041	316	30%	Stream hydrograph analysis
Miramichi, NB[12]	Appalachians		X	1081	369	34%	Stream hydrograph analysis
Sussex, NB[12]	Appalachians		X	1044	626	60%	Stream hydrograph analysis
Grand Falls, NB[12]	Appalachians		X	1029	289	28%	Stream hydrograph analysis
Nepisiguit Falls, NB[12]	Appalachians		X	1041	286	27%	Stream hydrograph analysis
Pennfield, NB[12]	Appalachians		X	1374	479	35%	Stream hydrograph analysis
Moncton, NB[12]	Maritimes		X	1008	317	31%	Stream hydrograph analysis
Fredericton, NB[12]	Maritimes		X	1186	383	32%	Stream hydrograph analysis
Annapolis, NS[16]	Maritimes		X	1209	170	14%	Stream hydrographs, water budget, modelling
Bangor, PEI[12]	Maritimes		X	1289	364	28%	Stream hydrograph analysis
O'Leary, PEI[12]	Maritimes		X	1084	316	29%	Stream hydrograph analysis
Wilmot, PEI[10]	Maritimes	X	X	1078	400	37%	Unknown
Maritimes Basin, NB, NS, PEI[17]	Maritimes		X	1100	148	13%	Stream hydrographs, water budget method, modelling

1 Scibek and Allen (2006a) 2 Scibek and Allen (2006b) 3 Appaih-Adjei and Allen (2009) 4 Toews and Allen (2009) 5 Holt (2004) 6 Smerdon et al. (2008) 7 Hayashi et al. (1998b) 8 van der Kamp and Hayashi (1998) 9 Hinton (2003) 10 Rivera (personal communication, 2007) 11 Jyrkama and Sykes (2007) 12 Michaud et al. (2002) 13 Croteau (2006) 14 Nastev (personal communication) 15 Larose-Charette (2000) 16 Rivard et al. (2007) 17 Rivard et al. (2008)

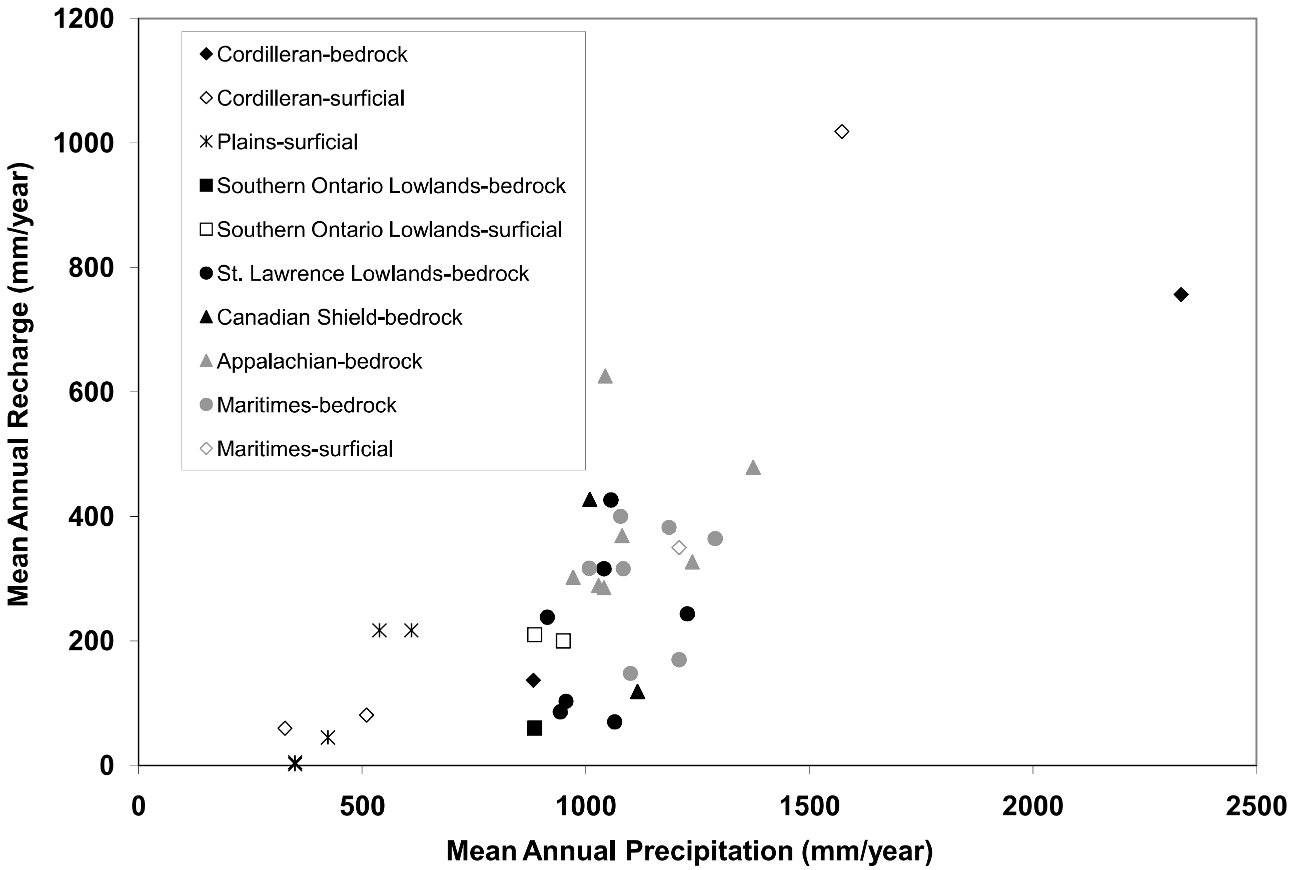


Figure 4.11 Graph showing mean annual recharge and mean annual precipitation obtained at different locations within the country's hydrogeological regions. A variety of methods were used to obtain these recharge estimates.

mountains to less than 300 mm in some intermontane valleys. Coastal regions experience highest precipitation during the winter months. Much of this precipitation falls as rain (temperatures are above freezing), except at high elevation, where it generally falls as snow (temperatures below freezing). In these coastal regions, groundwater recharge tends to occur during winter when evaporation and transpiration rates are at their lowest. Consequently, natural groundwater levels in coastal aquifers show a seasonal high during winter or early spring, and generally decline from spring to late fall (see Figure 4.9). By contrast, interior region records highest precipitation during summer (mostly as rain). Much of this precipitation is not available for recharge, however, due to high evaporation and transpiration. As the mean daily air temperature climbs, no excess water is available to infiltrate past the root zone for aquifer recharge. In the interior region and at higher elevations, snow accumulation during the winter months contributes to recharge during spring and summer as snowmelt (e.g., Okanagan Valley). Natural groundwater levels in aquifers located in the interior generally are at a seasonal high in late spring or early summer, and decline over summer and early fall. The groundwater level usually reaches a seasonal low during winter, when precipitation at the land surface is frozen and not available for recharge. Groundwater discharge, however, continues to maintain streamflows.

4.8.2 Plains

The climate of the Plains is a cool continental regime, dominated by semiarid conditions, but ranging from sub-humid to semiarid. Regional climate is zoned from south to north and from west to east as a result of Pacific or Arctic air masses combined with variations in solar radiation. Summers are short and warm, whereas winters are very cold. The western prairies have the lowest precipitation, and the highest evapotranspiration. Precipitation increases from 250 mm in southwestern Alberta to about 700 mm eastward in Manitoba. The region's basin-edge topography influences regional flow systems by introducing fresh meteoric water as recharge from isolated uplands. Local scale flow systems are driven by minor topographic variations, given the low topography of the region as a whole, its flat-lying stratigraphy, and the high bedrock heterogeneity. Depression-focused recharge tends to occur predominantly during spring snowmelt (Box 4-2). Annual recharge rates range from almost zero to about 50 mm in the prairies and up to about 200 mm in the northern forested portion of the Plains. The geochemical composition of groundwater has been found to reflect the overlying till composition (Grasby et al., 2010). Groundwater throughout the Plains region was recharged during the Pleistocene; this "fossil" groundwater can still be found at depth (Grasby and Chen, 2005).

4.8.3 Precambrian Shield

The Precambrian Shield is characterized by a wide range of precipitation, from some 1,400 mm in the south to about 400 mm at the northern tip of the Ungava Peninsula. The percentage of total precipitation falling as snow varies from about 25% in the south to more than 50% in northern parts of Quebec. Evapotranspiration also decreases with latitude, even more notably than precipitation, and the ratio of runoff to precipitation is often higher in the northern Shield. One characteristic of the Precambrian Shield Region is that roughly half of its northern domain is underlain by permafrost. The presence of a perennial or seasonal frozen layer reduces water infiltration and groundwater recharge by a considerable margin, both in surficial deposits and in bedrock. Canadian Shield bedrock is covered (for the most part) by a thin layer of Quaternary deposits, which can exceed 100 m in thickness in some bedrock valleys. Estimates of groundwater recharge within the permafrost terrain is reflected by a one order of magnitude decrease in stream baseflow with increasing latitude, from about 157 mm/year in the southern part of discontinuous permafrost, to about 15.7 mm/year and approaching zero in continuous permafrost areas (Lapointe, 1977; van Everdingen, 1987) (see Chapters 11 and 15).

4.8.4 Hudson Bay Lowlands

Climate of the Hudson Bay Lowland region is continental and strongly influenced by cold, moisture-laden Hudson Bay and polar air masses. It is characterized by short cool summers and cold winters. Mean annual temperature ranges from -4 to -2°C, although it is closer to -7°C in Manitoba. Annual precipitation averages from 400 to 800 mm, increasing from northwest to southeast. Up to 75% of the area is underlain by wetlands. There is little hydrogeological data for this region because it contains few water wells, and little development has occurred. Therefore, we know very little about groundwater recharge. Nevertheless, recharge is expected to be enhanced in those areas where solution features are well developed within carbonate-evaporite rocks insofar as Holocene karst locally enhances groundwater recharge via sinkholes. Lower recharge can be anticipated in the more massive, fine-crystalline carbonate rocks.

4.8.5 Southern Ontario Lowlands

Southern Ontario has a temperate climate with warm summers and mild winters. Mean temperature ranges from 5°C to 8°C. Mean annual precipitation ranges from 720 to 1,000 mm, plus extremes due to major storms. Precipitation is higher east of major lakes and lowest away from these lakes. Longer, frost-free growing seasons (2,550 degree-days) near Lake Erie lead to higher evapotranspiration (about 600 mm/year) compared to the cooler (1,750 degree-days) northern interior areas (about 500 mm/year). Available moisture surplus for direct runoff to streams or groundwater infiltration varies from about 200 to 400 mm/year with local variation.

4.8.6 St. Lawrence Lowlands

Climate within the central St. Lawrence Lowlands ranges from continental in the west to maritime in the east. Northeastern areas are notably cooler due to effects of the Labrador Current. Mean annual temperatures range from 2.5°C to 5°C, while mean annual precipitation ranges from 800 to 1,100 mm. Spring arrives in April in the western areas of the region, while in the east snow may linger into May. The St. Lawrence Lowlands has generally flat-lying topography, which rarely

rises above 150 m elevation. Recharge occurs predominantly during snowmelt, and results in peak groundwater levels during late spring. Recharge may be enhanced along faults and within the karst openings of carbonate rocks, most notably where sediment cover is thin, or along major rivers or escarpments.

4.8.7 Appalachians

The Appalachian Region is one of the wettest areas in Canada. On average, it receives 1,150 mm/year of precipitation (except in Nova Scotia where the average is around 1,400 mm/year), of which 20% to 25% falls as snow. The climate is humid continental, with long, cold winters, and warm summers. Large seasonal temperature variance (up to 35^{o}C) is common. Due to the region's relatively low relief, the major influence on weather is lands' distance to the sea. Coastal areas are cooled in the summer and warmed in the winter by the ocean, causing sudden temperature changes and frequent freeze-thaw cycles during the winter. Snowfalls are often heavy. The glacial till, the most common surficial deposit in this region, allows a fair amount of the precipitation water to infiltrate. Therefore, it is often assumed, as a first estimate that evapotranspiration, surface runoff, and infiltration each account for one third of total precipitation. However, the net recharge is often less than that, except where permeable deposits overlie permeable rocks. Average recharge rates range from 115 to 250 mm/year over large regions (representing 10% to 22% of precipitation), although these rates can reach 300 to 350 mm/year in some areas like Prince Edward Island. Usually the aquifers in eastern Canada supply streams on a regular basis, even during the summer.

4.8.8 Maritimes Basin

Climate of the Maritimes Basin is humid continental with long winters and warm summers. This region is one of the wettest of Canada, some 25% of precipitation occurs as snowfall. Because of low basin relief, distance to the sea is the major influence on weather. Northumberland's coastal strait areas are cooled in summer and warmed in winter by the ocean. Daily average air temperature varies between 17°C and 24°C in summer and between -12°C and -4°C during winter. Average precipitation is between 900 and 1,500 mm/year; highest values occur along the Bay of Fundy. Mean annual evapotranspiration varies from 345 to 440 mm/year and results in a large water surplus. The relief is commonly less than 150 m asl, although locally it can rise to 300 m asl in New Brunswick and Nova Scotia. Recharge through the sedimentary and volcanic rocks is enhanced locally by fracturing or, where terrain is overlain by sand and gravel deposits, associated with major streams and glaciofluvial corridors. Till is thin but widespread, and adequately transmissive to permit significant recharge to bedrock aquifer systems. Bedrock outcrops are rare. Potential recharge rates are between 100 and 400 mm/year.

4.8.9 Northern Canada

The permafrost region covers Canada's north, and is characterized by rock and/or soil temperatures that remain at or below 0°C through the summer, with the result that pore water is normally frozen. The southern margin of the region is irregular because secondary features, such as type of vegetation cover or snow depth, exercise some control as to where permafrost occurs. Climate is dominated by continental and polar maritime (influenced by the ocean) subtypes, and is additionally affected by the extreme solar radiation conditions of high latitudes. Mean annual temperature ranges from -20°C on Ellesmere Island to -6°C along the southern boundary. Mean annual precipitation varies from 100 mm in the north to 600 mm in the southeast. Precipitation of the central arctic is the lowest in Canada and, consequently, this area is often referred to as a polar desert. The primary hydrogeological function of permafrost is to act as a barrier to groundwater flow. Therefore, recharge is generally limited to areas where the active layer is thawed, or beneath lakes where thick taliks may form, resulting in only local groundwater occurrences.

4.9 CLIMATE VARIABILITY AND CLIMATE CHANGE

Canada's climate has varied historically over many time scales. The last of the great ice sheets began to melt 15 thousand years ago, creating significant runoff and likely significant groundwater recharge. Prior to that time, when the Canadian landmass was largely covered by ice, there was likely little groundwater recharge, although glacial meltwater may have generated recharge in some areas. Thus, climate change and climate variability play important roles in groundwater recharge variability. Whereas climate variability is generally observed on short time scales (few years to decades), climate change is manifested as a longer-term, more persistent change (decades and longer).

Because groundwater moves at very low rates through the ground, old groundwater still remains in some of the deeper aquifers, for example, within the Precambrian Shield. In such environments, the low frequency record of climate change is preserved. Shallow groundwater systems, by contrast, tend to be replenished by groundwater at faster

rates, and thus respond much more quickly to climate variations. Shallow groundwater systems often record higher frequency climate variations.

4.9.1 Climate variability and recharge

Climate variability impacts groundwater recharge over relatively short time scales. In Canada precipitation responses are associated with two extreme phases of the El Niño Southern Oscillation (ENSO), namely, El Niño and La Niña (Shabbar et al., 1997). Using the best available precipitation data from 1911 to 1994, these authors demonstrated that precipitation extending from British Columbia, through the Prairies, and into the Great Lakes region, is significantly influenced by ENSO phenomena. This variability decreases towards the pole, and is greatest in the central prairies. The pattern among the mountains and valleys of British Columbia is irregular.

Zheng et al. (2001) examined the spatial and temporal characteristics of heavy precipitation events over Canada (excluding the high Arctic) for the period 1900-1998, and discovered that about 71% of total precipitation in southern Canada comes from rainfall events. In northern Canada, more than 50% of total precipitation comes from snowfall events. Heavy rainfall and snowfall events were defined for each season and station separately by identifying a threshold value that was exceeded by an average of three events per year. Annual and seasonal time series of heavy event frequency were then obtained by counting the number of annual

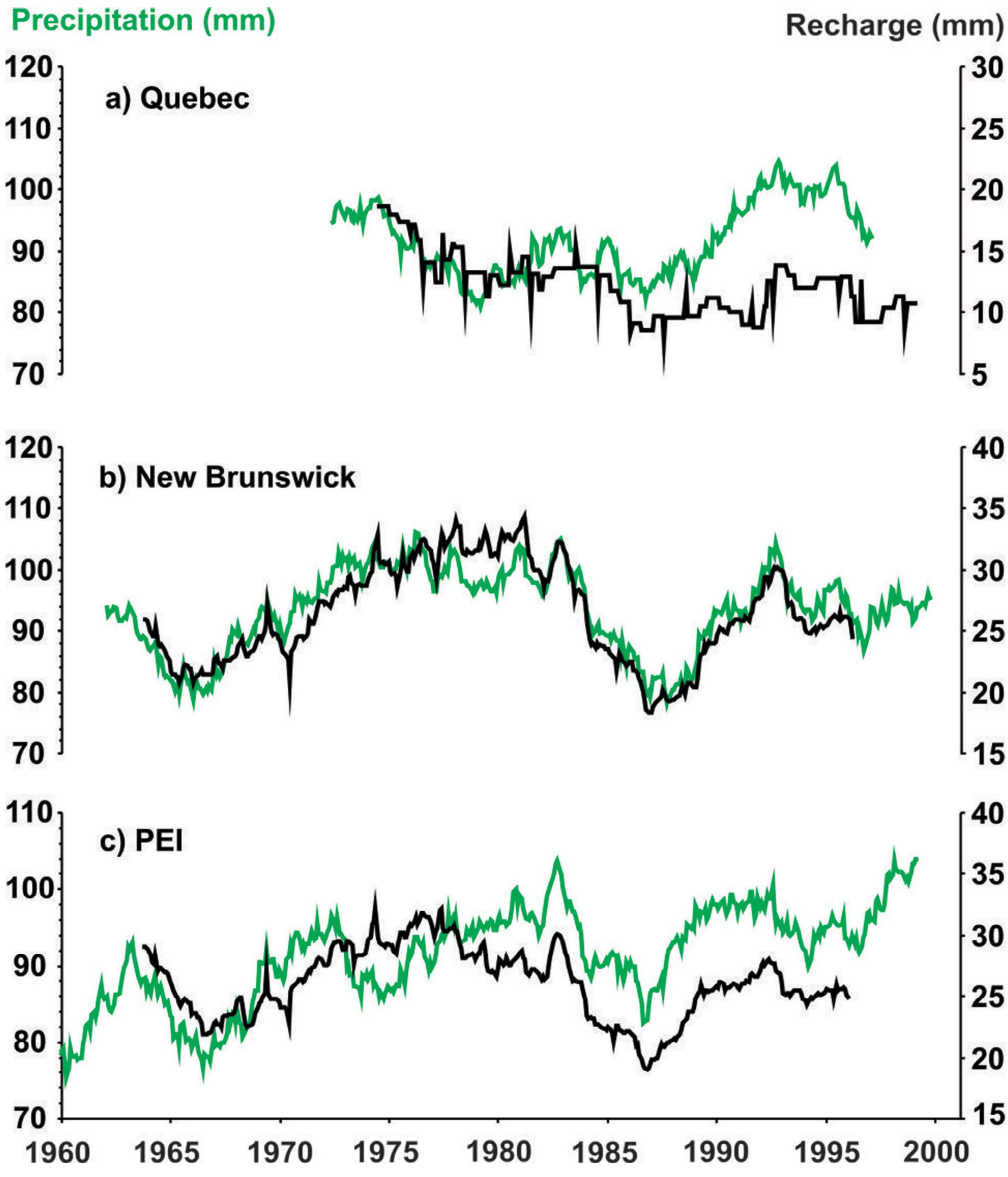


Figure 4.12 Precipitation and recharge variations in three aquifers in eastern Canada.

exceedances. Characteristics of the intensity of heavy precipitation events were investigated by examining the 90th percentiles of daily precipitation, the annual maximum daily value, and the 20-year return values.

Decadal variability is the dominant feature in both frequency and intensity of extreme precipitation events across the country, although for Canada as a whole, there appeared to be no identifiable trends in extreme precipitation (either frequency or intensity) during the last century. The observed increase in precipitation totals during the twentieth century was attributed largely to the increase in the number of small to moderate events. Stations with coherent temporal variability in the frequency of heavy precipitation events were grouped by cluster analysis and examined on a regional basis. Results show stations belonging to the same group are generally located in a continuous region, indicating that the temporal distribution of the number of precipitation events is spatially coherent. It was also found that heavy snowfall events are more spatially coherent than heavy rainfall events. Indices representing temporal variations of regional heavy precipitation display strong interdecadal variability with limited evidence of long-term trends, and vary markedly depending on the precipitation type, season, and region. Heavy spring rainfall events over eastern Canada have shown an increasing trend superimposed on the strong decadal variability. However, heavy rainfall events in other seasons or regions are generally not associated with any such trends. The number of heavy snowfall events in southern Canada reveals an upward trend from the beginning of the 20th century until the late 1950s–1970s, followed by a downward trend continuing to present day. Heavy snowfall events in northern Canada have been increasing with marked decadal variation over the last 50 years. The majority of stations have a significant positive correlation between the total amount of snowfall contributed by heavy events versus that contributed by non-heavy events. This relationship is strongest over western Canada. On the other hand, relatively few (<20%) stations had a significant correlation between total rainfall in heavy events and non-heavy events. These results suggest that the amount of precipitation falling in heavy and non-heavy events increases or decreases coherently for snow, but not for rain.

Recent studies in southern Manitoba reveal that short-term precipitation trends (wet-dry cycles) have good correlations with groundwater level variations (Chen et al., 2002, 2004). The calculated average correlation coefficient between a three-year moving average precipitation and annual groundwater levels is 0.85. 85% of the wells have a coefficient greater than 0.8. Statistically, a 0.85 coefficient means that more than 70% of the variations in the groundwater levels can be explained by the variations in annual precipitation. Annual mean air temperature displays a significant negative correlation with the annual groundwater levels. The mean value of the correlation coefficients between temperature and groundwater level from 72 monitoring wells is 0.72 (Chen et al., 2004).

The potential impact of increasing annual mean temperature on the groundwater levels in Manitoba was examined by studying the correlation coefficients between groundwater levels and temperatures in different time periods, in which the annual mean temperature had about 1.5°C difference. Calculations showed that the correlation coefficients between temperatures and groundwater levels increased about 15% from the cooler period to the hotter period, while the correlation coefficients between precipitation and groundwater levels in these two time periods changed little (<5%) (Chen et al., 2004).

Average growing season precipitation values in the Canadian Prairies range from over 300 mm in west-central Alberta and eastern Manitoba, to less than 200 mm in southern Alberta and into southwestern Saskatchewan (Bonsal et al. 1999). This precipitation is critical to several environmental processes and economic activities and, most notably, to agriculture. Also important is the temporal distribution of precipitation within the growing season. Over the agricultural region of the Plains, maximum rainfall normally occurs from mid-June to early July. Variations in this temporal distribution can also have severe effects.

In eastern Canada, variability in precipitation and recharge occurs on decadal time scales. Precipitation curves for representative climate stations in Quebec, New Brunswick and Prince Edward Island are illustrated in Figure 4.12. Inter-annual and longer-term (several decades) variability is evident. Groundwater recharge is found to vary similarly to precipitation, particularly at the New Brunswick site, although recharge diverges from precipitation at the Quebec and PEI sites.

4.9.2 Climate change impacts on recharge

It is expected that global changes in temperature and precipitation will alter groundwater recharge to aquifers (Zektser and Loaiciga, 1993). As a first response, climate change will cause shifts in water table levels in unconfined aquifers (Changnon et al., 1988).

Studies that examine historical precipitation and groundwater level variability lend insight into the physical processes that control groundwater recharge, and help scientists make predictions for future climate change conditions. Detection of the effects of long-term climate change remains in its early stages, and, as a result, many of the approaches used for predicting the responses of aquifers to climate change are largely modelling-based (e.g., Scibek and Allen, 2006a; Jyearkama and Sykes, 2007; Toews and Allen, 2009). The approach employed in some predictive studies (e.g., Box 4-1) utilizes future climate data generated from global climate models (GCMs) as input to recharge models. Representative climate data for specific time periods, such as the 2050s or 2070s, for example, is generated, rather than using a continuous data series spanning several decades. The resultant modelling results are limited in that they do not illustrate the progressive changes in groundwater recharge, but rather provide snapshot views of recharge for different periods in the future.

Cumulative effects of climate change are not fully accounted for in these models. Long lasting severe dry weather conditions, for example, may change an aquifer's hydraulic properties (Larocque et al., 1998), significantly altering recharge rates for major aquifer systems and affecting the sustainable yield of groundwater in certain regions. Another limitation lies in the recharge models themselves, because they commonly use daily, rather than hourly, climate data as input and, therefore, may miss capturing those extreme precipitation events which occur at time scales of less than one day. Although high intensity rainfall events, contribute, on average, to greater precipitation for any particular day or month, these events can also ultimately lead to less recharge because most of the precipitation generated turns into runoff and does not infiltrate the ground.

Much more research on the effects of extreme events on groundwater recharge is needed, particularly given the premise that future climate change will probably result in more extreme precipitation events.

4.10 CONCLUSIONS

Although we may know much about the science of groundwater recharge in general, only a few groundwater studies have been conducted across the country to provide estimates of recharge. As a result, we lack a comprehensive nation-wide groundwater recharge assessment, although a preliminary compilation of information was undertaken in the writing of this chapter. The authors compiled recharge estimates for different regions across the country as shown in Figure 4.11; these estimates were based on field measurements or modelling. In addition, the authors constructed a map detailing potential annual recharge (Figure 4.5), based on annual precipitation and potential evapotranspiration. Such compilations should be expanded as more data becomes available.

Recharge varies considerably across Canada, ranging from 0 to over 1,000 mm/year. The magnitude and timing of this recharge are strongly dependent on the region's climate, as well as the physical properties of the soil and aquifers, topography, and the nature of the land cover. Both short-term and long-term precipitation and temperature trends, as well as changes in land use and land cover, can be expected to impact on the magnitude and timing of recharge in most regions across the country. Except for a small sampling of aquifers, long term effects of climate change on groundwater recharge across Canada are not well studied.

BOX 4-1 DIRECT AND INDIRECT GROUNDWATER RECHARGE AND IMPACTS OF FUTURE CLIMATE CHANGE

The Grand Forks aquifer in south-central British Columbia (Figure 4.13) is contained within the mountainous valley of the Kettle River along the Washington State border. At Grand Forks, the climate is semiarid and most precipitation occurs in the summer months during convective activity. In the winter, much of the precipitation at high elevation is as snow: the observing sites at valley bottom record less snowfall. The valley fill consists of glaciolacustrine and glaciofluvial sediments. The top-most sediments, forming the unconfined aquifer, are comprised of Holocene age gravels and sands. Groundwater in the valley is used extensively for irrigation and domestic use (Wei et al., 1994).

Within the Grand Forks valley, the Kettle River is a meandering gravel-bed river incised into glacial outwash sediments. Recent research has demonstrated that the aquifer water levels are highly sensitive to water levels in the Kettle River (Allen et al., 2004; Scibek et al., 2007). Figure 4.14 shows a conceptual model of the interaction between the Kettle River and the aquifer. At peak flow, during the spring freshet (regional snowmelt), the rise in river stage causes water to move laterally away from the channel, resulting in groundwater levels rising over a broad area. The rate of inflow to groundwater from the river along the floodplain zone follows very closely the river hydrograph during the rise in river stage. Water is stored in the aquifer during this time. Within approximately 10 days following peak discharge, river levels begin to fall, and the groundwater flow direction is reversed. At this time, the rate of inflow from the aquifer to the river increases, and this

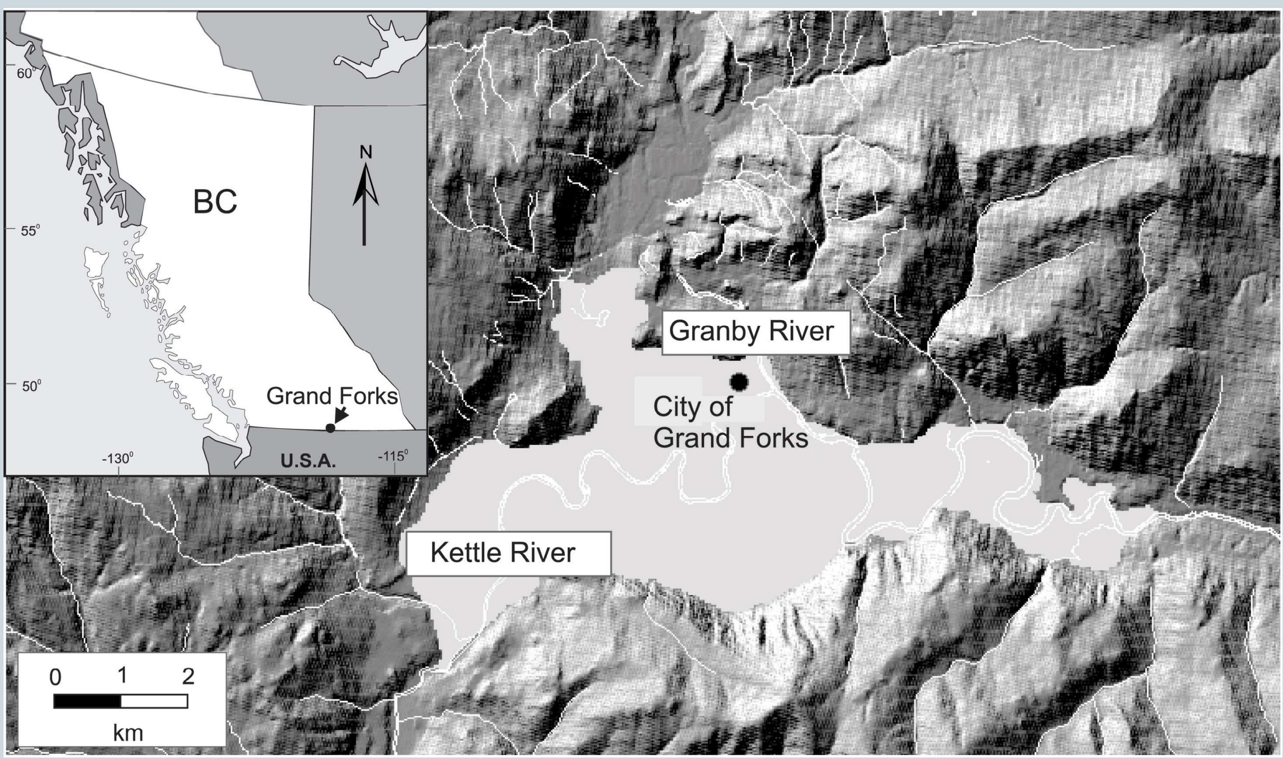


Figure 4.13 Location map of the Grand Forks aquifer in south-central British Columbia. The unconfined aquifer is hydraulically connected to the Kettle and Granby Rivers, receiving both direct recharge from precipitation and indirect recharge from streamflow.

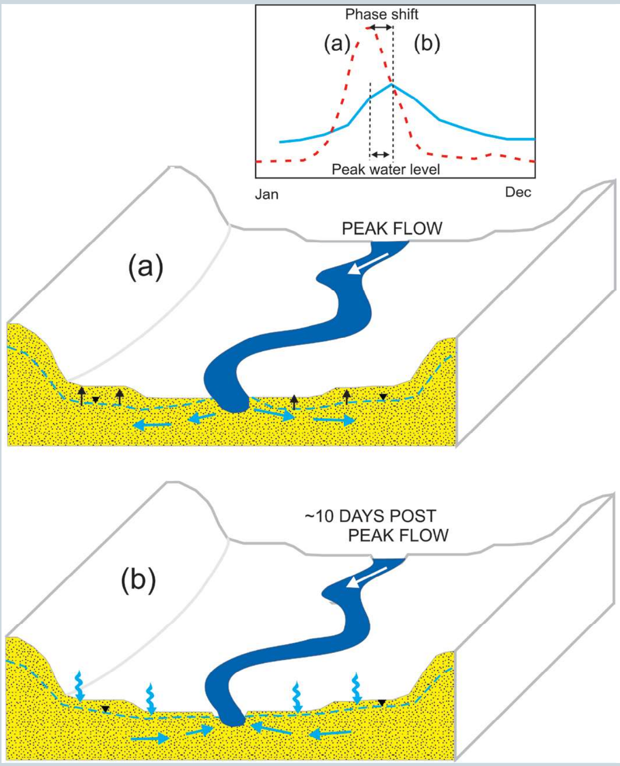


Figure 4.14 This schematic drawing illustrates the interaction between the Kettle River and the Grand Forks aquifer. (a) At peak flow, river water recharges the aquifer and moves laterally away from the channel, causing groundwater levels to rise over a broad area. (b) Within a relatively short period of time (about 10 days) following the peak streamflow, when river levels begin to fall, the groundwater flow direction is reversed and groundwater contributes to baseflow (adapted from Scibek et al., 2007).

condition prevails for the rest of the year, as water previously stored in aquifer drains back to the river as baseflow seepage. Modelling results indicate that river puts about 15% of its spring freshet flow into storage in Grand Forks valley aquifer, and most of that water is released back to the river as baseflow, within 30 to 60 days. Under natural conditions, the net exchange is roughly zero over the year. When pumping wells are included in the model, however, there is a small reduction in the baseflow component to the Kettle River.

Estimates of direct recharge to the shallow aquifer were obtained by modelling. Mean annual recharge, determined spatially using the HELP

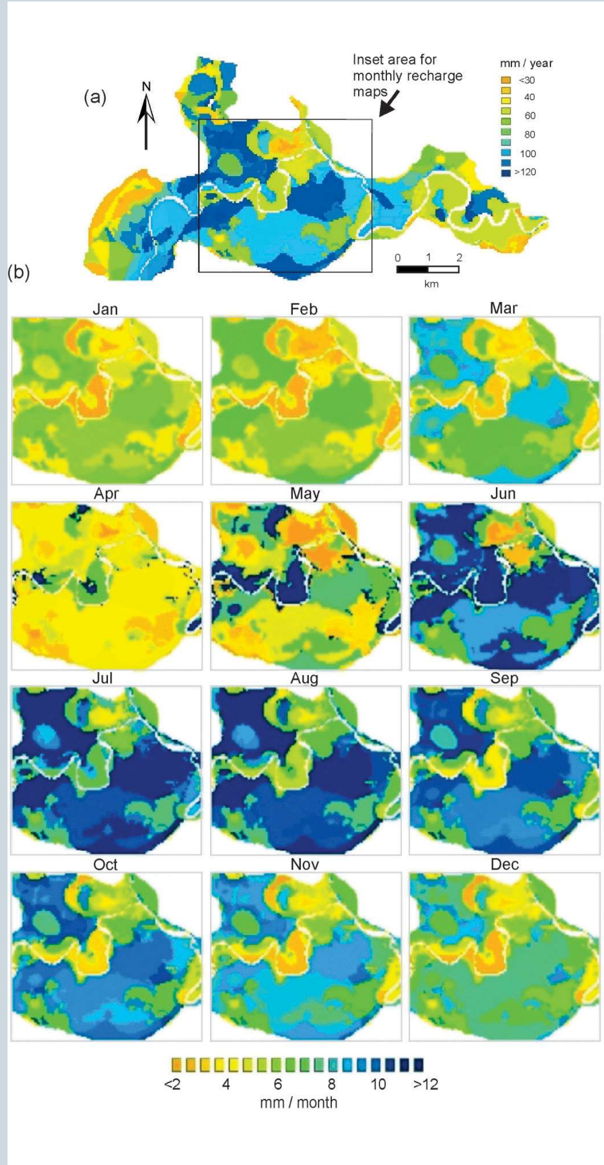


Figure 4.15 (a) Mean annual recharge to the Grand Forks aquifer. **(b)** Mean monthly recharge maps for inset area (central portion of valley). From Scibek and Allen, 2006a.

infiltration model (Schroeder et al., 1994) varies considerably across the aquifer (Figure 4.15), ranging from slightly less than 30 mm/year to over 120 mm/year (corresponding to 6% and 24% of mean annual precipitation, respectively). Mean monthly recharge (to the inset area shown) follows the annual distribution of precipitation. The ground is frozen in winter, and snowmelt infiltration does

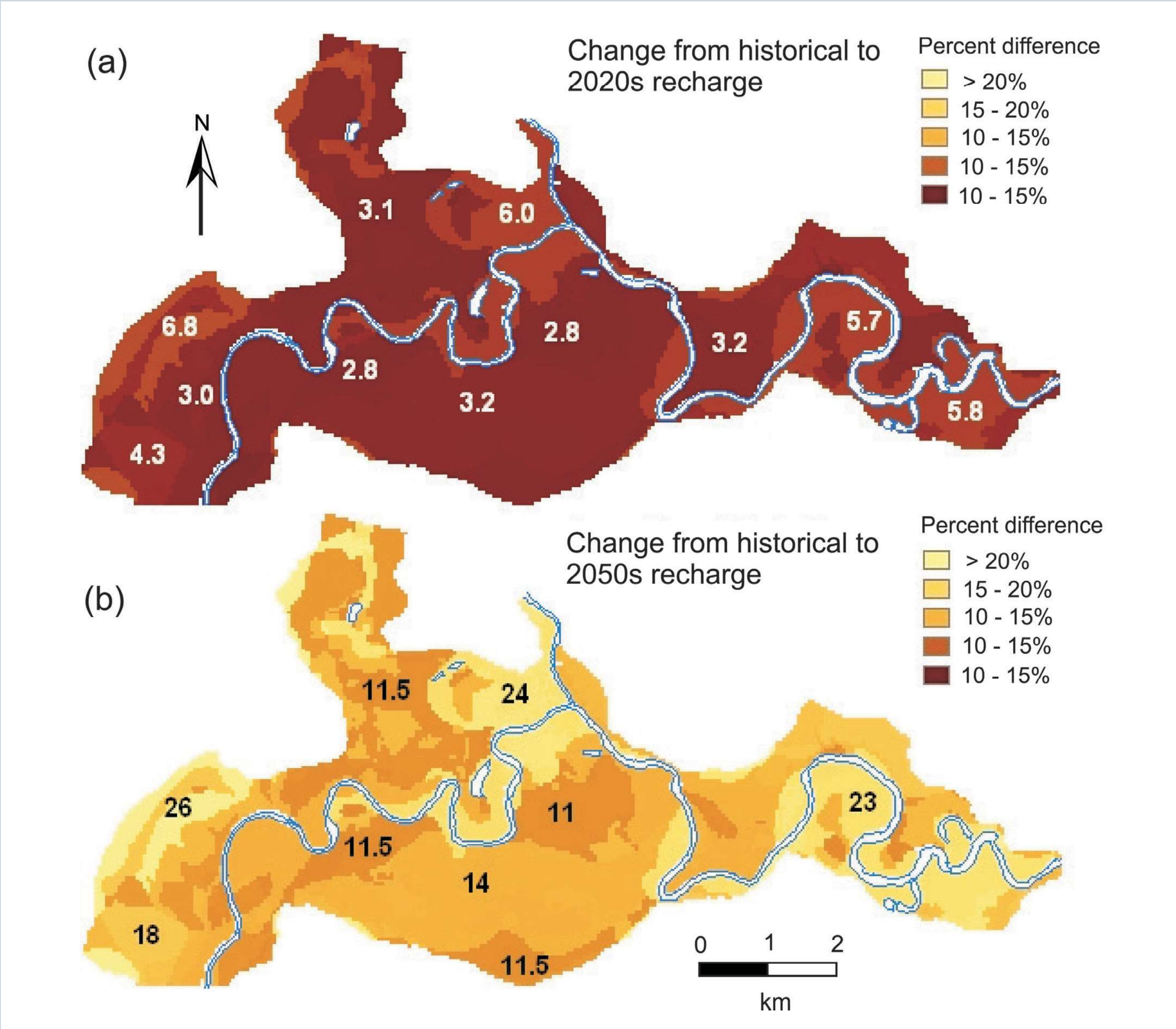


Figure 4.16 Percent change in mean annual recharge for (a) the 2020s, and (b) the 2050s, relative to historical recharge. From Scibek and Allen, 2006a.

not occur. Most of the recharge is received in late spring and summer through snowmelt and summer rainstorm events. Autumn has moderate recharge, less than in early summer. Monthly recharge varies from less than 2 mm/month to over 12 mm/month. The range in percentages (i.e., 6% to 24%) of mean annual precipitation is smaller than the range in percentages (10% to 80%) of monthly precipitation, due to seasonal variation in precipitation and averaging on annual time scales. Should a high intensity event, such as a thunderstorm, occur during the late summer most of the water will infiltrate the aquifer. When it rains more slowly, and over a longer time, much more water is lost through evaporation. This relationship may be very different in other climate regions, and in other aquifers, as high intensity rainfall events may lead to increased runoff and less infiltration. The overall effects of direct recharge in the Grand Forks aquifer are very small in comparison to indirect recharge from the Kettle River, except in those areas, such as the terraces along the valley sides, removed from the river influence.

The impact of predicted future climate change on groundwater in this aquifer has also been estimated. Three year-long climate scenarios were run, each representing one typical year in the present, and in future (2020s and 2050s), through the perturbation of historical weather according to downscaled CGCM1 global climate model results. Results projected for the 2050s showed that the largest increase in recharge relative to present would occur in late spring, by a factor of three or more. Recharge throughout most areas of the aquifer would evidence a 50% increase in summer months; a 10% to 25% increase in autumn, and a reduction during winter (Figure 4.16). CGCM1 downscaling was also used to predict basin-scale runoff for the river. Future climate scenarios suggest a shift in the hydrograph peak to an earlier date. Although the peak flow would remain the same, the baseflow level would be lower and of longer duration. Groundwater levels near the river floodplain are predicted to be higher earlier in the year due to an earlier onset of peak flow, but considerably lower during the summer months (Figure 4.17.). Away from rivers, groundwater levels increase slightly due to the predicted increase in recharge.

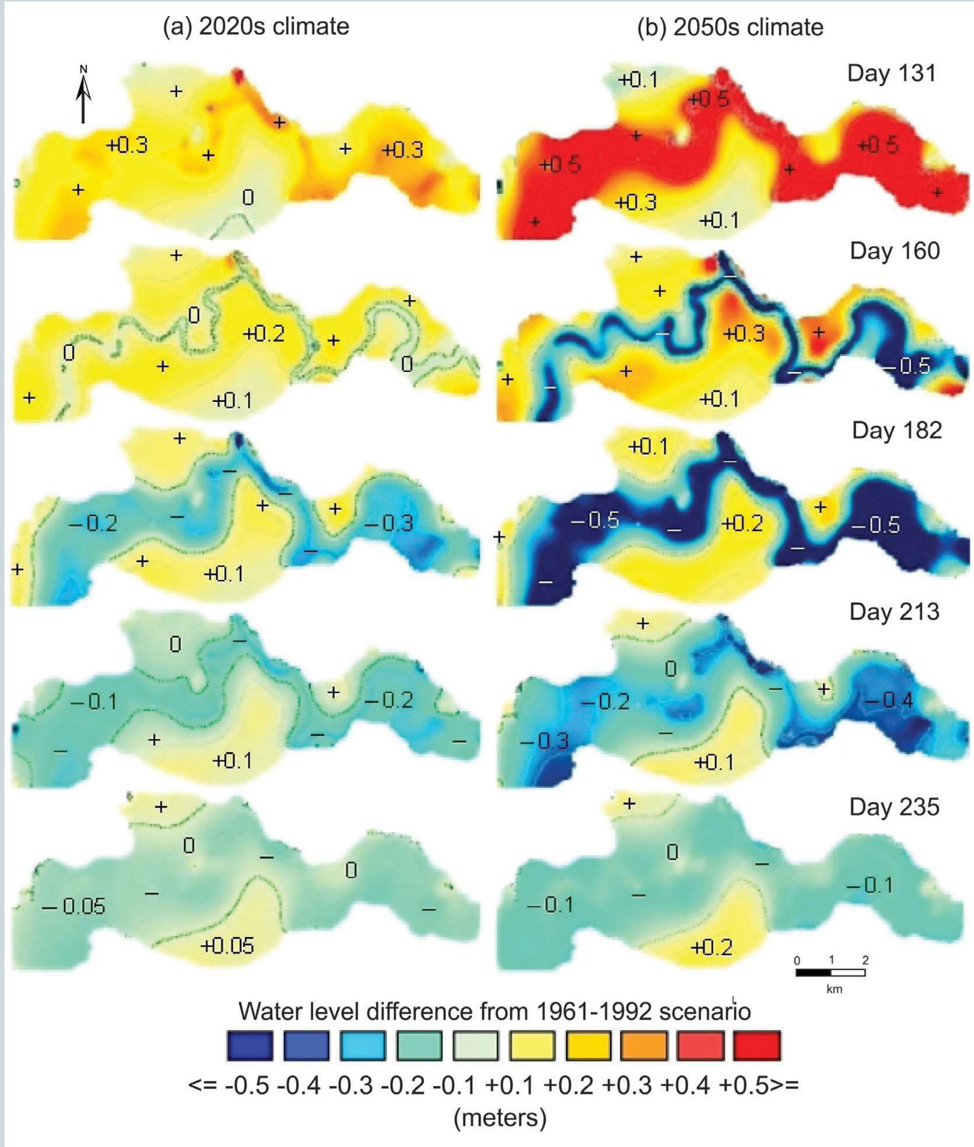


Figure 4.17 Water level differences between (a) future (2020s) and historical climate, and (b) future (2050s) and historical climate under pumping conditions. Maps represent different Julian Days. Darkest blue colours indicate values less than -0.5 m (along rivers only). From Scibek and Allen, 2006a.

BOX 4-2 DEPRESSION-FOCUSED GROUNDWATER RECHARGE

Depression-focused recharge is commonly observed in arid and semiarid regions around the world including the Southern High Plains of Texas (Scanlon and Goldsmith, 1997), the La Plata region of Argentina (Logan and Rudolph, 1997), and the Sahel region of Niger (Desconnets et al., 1997).

Several decades of studies in the Canadian prairies (e.g., Meyboom, 1966; Zebarth et al., 1989; Hayashi et al., 2003) have established a reasonably good understanding of the recharge processes at least at the scale of individual depressions, as presented below in a summary of detailed field studies conducted at the St. Denis National Wildlife Area near Saskatoon.

The climate of the Canadian prairies is characterized by high moisture deficit and cold winters with temperatures frequently below -30ºC. Annual precipitation in Saskatoon is only 360 mm, which is greatly exceeded by potential evaporation of 700 mm (Hayashi et al., 1998a). As a result, annual runoff in the prairies is typically less than 25 mm (CNC/IHD, 1978), of which a large portion is generated during snowmelt, when the rapid depletion of snowpack over frozen, relatively impermeable soil generates a relatively large amount of runoff over a period of a few weeks (van der Kamp et al., 2003). Storm runoff rarely occurs in summer and fall, and the base flow in first-order streams originating within the prairies is primarily sustained by groundwater discharge.

The continental ice sheet covered much of the prairies during the last glaciation, depositing a thick (up to 200 m) blanket of poorly sorted material referred to as glacial till. Glacial till has relatively high clay content (20%–40% by weight): the hydraulic conductivity of unweathered till is in the order of 10^{-9}–10^{-11} m/s (Keller et al., 1989). Conductivity of the weathered till in the shallow zone is usually much higher owing to the presence of fracture network and macropores (Hayashi et al., 1998a; Parsons et al., 2004). Glaciated terrains generally have undulating or hummocky topography with numerous depressions. Ephemeral ponds and wetlands form in these depressions (Figures 4.18a and 4.18b), resulting in depression-focused infiltration and groundwater recharge.

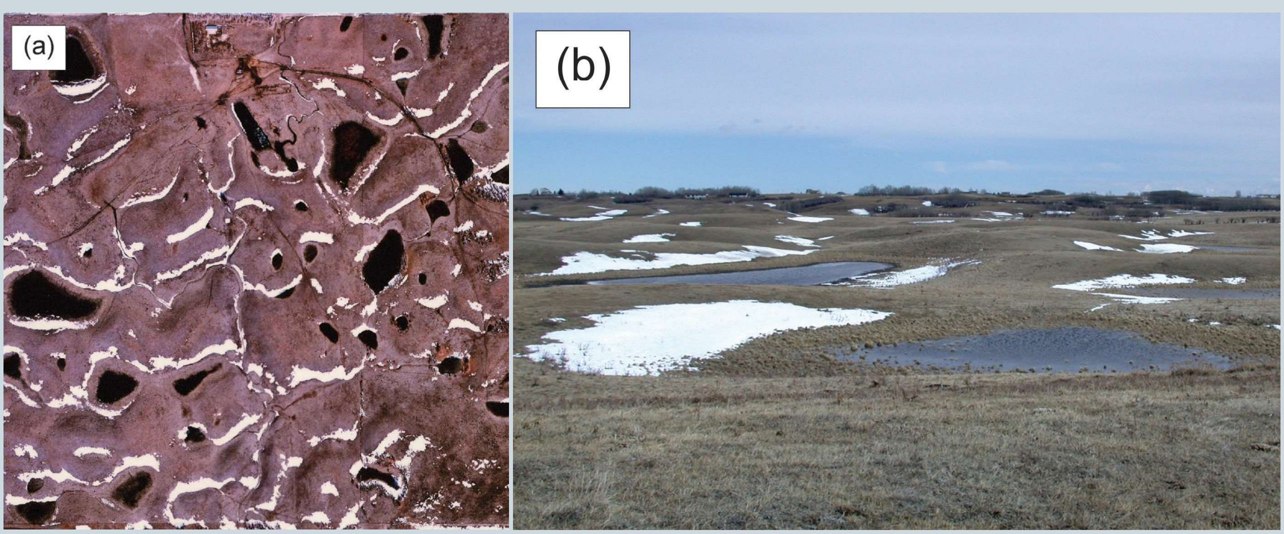


Figure 4.18 (a) An infrared (IR) aerial photograph. **(b)** A ground-based colour photograph of a pasture in the West Nose Creek watershed, located 20-km north of Calgary, Alberta. The photograph, taken on March 25, 2003, shows numerous depressions filled by snowmelt water. The water remains in depressions for a few weeks while the underlying soil thaws.

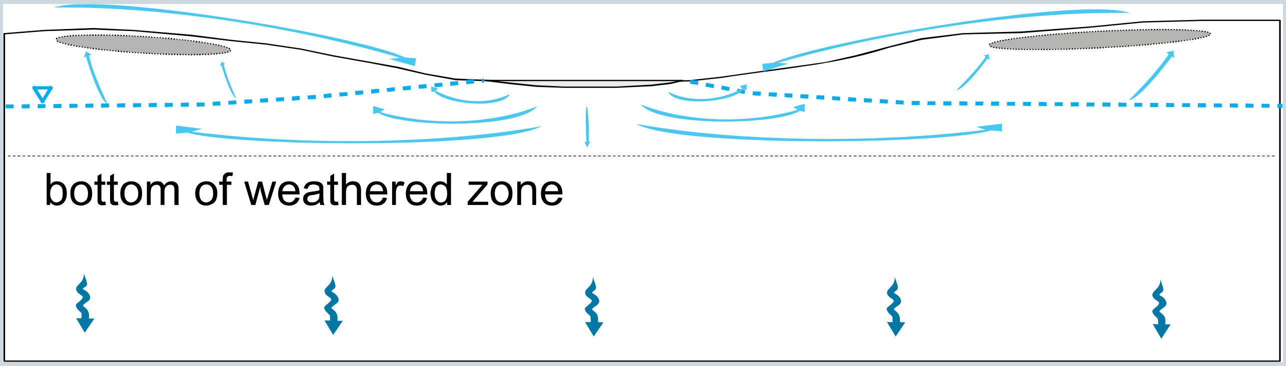


Figure 4.19 Conceptual model of snowmelt runoff, groundwater flow, and solute transport under a seasonal wetland. Arrows indicate the surface and subsurface pathways of water and dissolved salts, and the shaded areas on uplands indicate the occurrence of high soil salinity (modified from Hayashi et al., 1998b).

By definition, infiltration under depressions is regarded as shallow groundwater recharge although the recharge rate of deeper aquifers is much lower than the infiltration rate. Hayashi et al. (1998b) estimated a recharge rate of 1–3 mm/year for a sand aquifer located 25 m below the surface. This is because plants growing around depressions take up shallow groundwater to sustain relatively high transpiration rates during summer months, thereby inducing lateral groundwater flow towards the uplands (Figure 4.19). The flow usually occurs in the weathered zone, which has relatively high hydraulic conductivity. Evapotranspiration on the uplands concentrates dissolved salts in soil water, resulting in the formation of “saline ring” around depressions (Mills and Zwarich, 1986; Berthold et al., 2004). Grasses and agricultural crops growing on uplands can consume most infiltrated water during the growing season, allowing very little downward flux below the root zone. The presence of saline rings (Figure 4.19) implies that the downward flux under uplands is essentially zero averaged over a long period.

These studies suggest that the dominant mode of recharge is depression-focused. However, it is also clear that much of the shallow groundwater recharge under the depression flows laterally, to be consumed by evapotranspiration without recharging deeper aquifers. Water-holding depressions typically occupy 10%–20 % of the land within their respective watersheds (Hayashi et al., 1998b). When expressed as the recharge rate over an entire watershed, depression-focused recharge represents 1–45 mm/year of water input to deep aquifers in several zones under study in Canada and the United States (Hayashi et al., 1998b). The range of recharge rates is consistent with published values of groundwater recharge for prairie aquifers (van der Kamp and Hayashi, 1998), suggesting that depression-focused recharge may account for the majority of aquifer recharge. However, the amount of snowmelt infiltration and runoff on uplands is strongly dependent on land use (van der Kamp et al., 2003), and the effects of land use change on groundwater recharge are still not well understood. Future research is required before we can explain how land use change and climate change may affect the hydrologic cycle in the prairies, and depression-focused groundwater recharge.

BOX 4-3 MODELLING GROUNDWATER RECHARGE AND WATER BUDGETS

The spatial and temporal distribution of recharge rate has been estimated for the Chateauguay River watershed (Croteau, 2006). This regional aquifer system consists of fractured sedimentary rocks covered by heterogeneous Quaternary sediments of various thicknesses (For details on the watershed see Box 6-1, Chapter 6). Recharge was defined as the amount of precipitation water percolating through unconsolidated sediments to reach regional aquifer units. Water budget components (see Equation 4.1 in main text) were estimated with the physically based HELP infiltration model (Schroeder et al., 1994). Collected input data (climate, vertical Quaternary stratigraphy with corresponding physical properties, drainage properties, land use and vegetation cover) were integrated over a 250 × 250 m grid (47,616 cells). The model was calibrated against the runoff, and baseflow components of the hydrographs recorded over the considered period 1963–2001. The Champan (1999) baseflow separation technique was used.

Results indicate an average regional recharge rate of 86 mm/year or 9% of the mean total precipitation of 943 mm. Evapotranspiration was the most important water budget component accounting for 52% of precipitation, whereas combined surface and subsurface runoff accounted for 39% of precipitation. The average recharge varied from zero, along streams and other surface waters considered as gaining streams, up to 404 mm/year at rock outcrops and rock covered with a shallow sandy layer (Figure 4.20).

The daily evolution of water budget components averaged over the basin is shown in Figure 4.21. The year 1985 was selected as representing typical climate conditions for the region. Temperature has

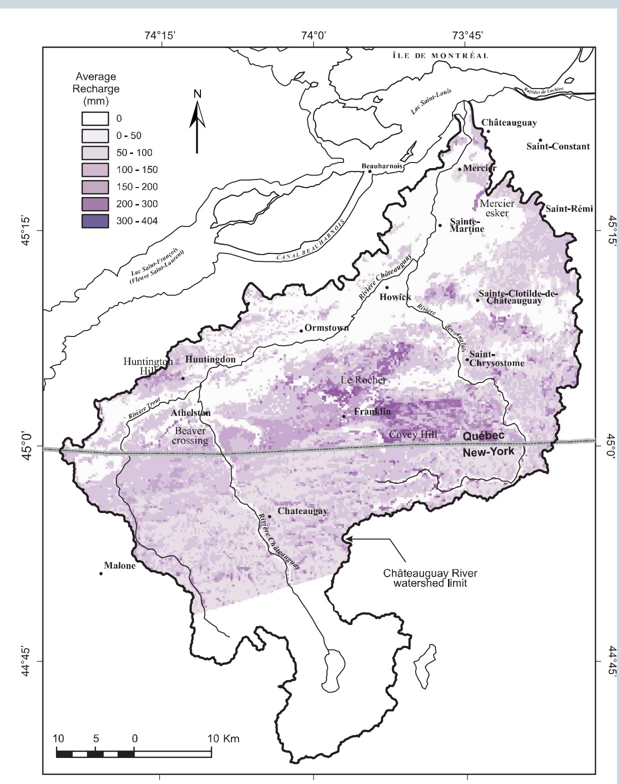


Figure 4.20 Spatial distribution of the average annual recharge rate.

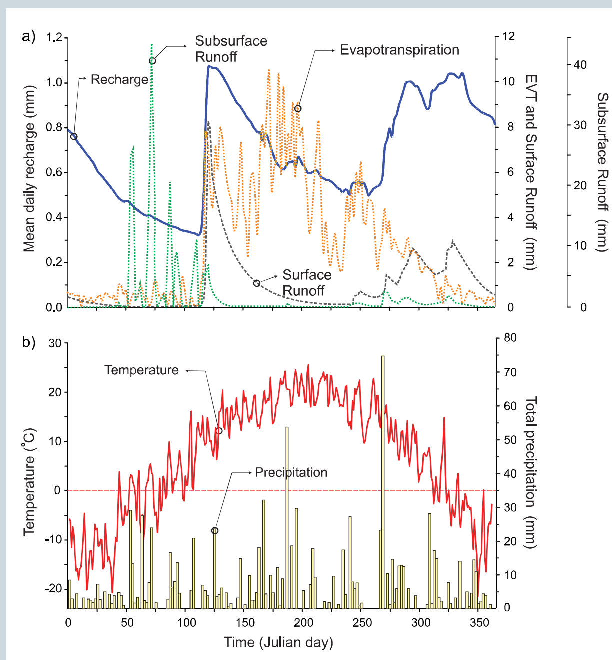


Figure 4.21 **(a)** Daily evolution of the recharge, surface runoff, subsurface runoff, and evapotranspiration rates compared with **(b)** daily variations in climate conditions for the year 1985.

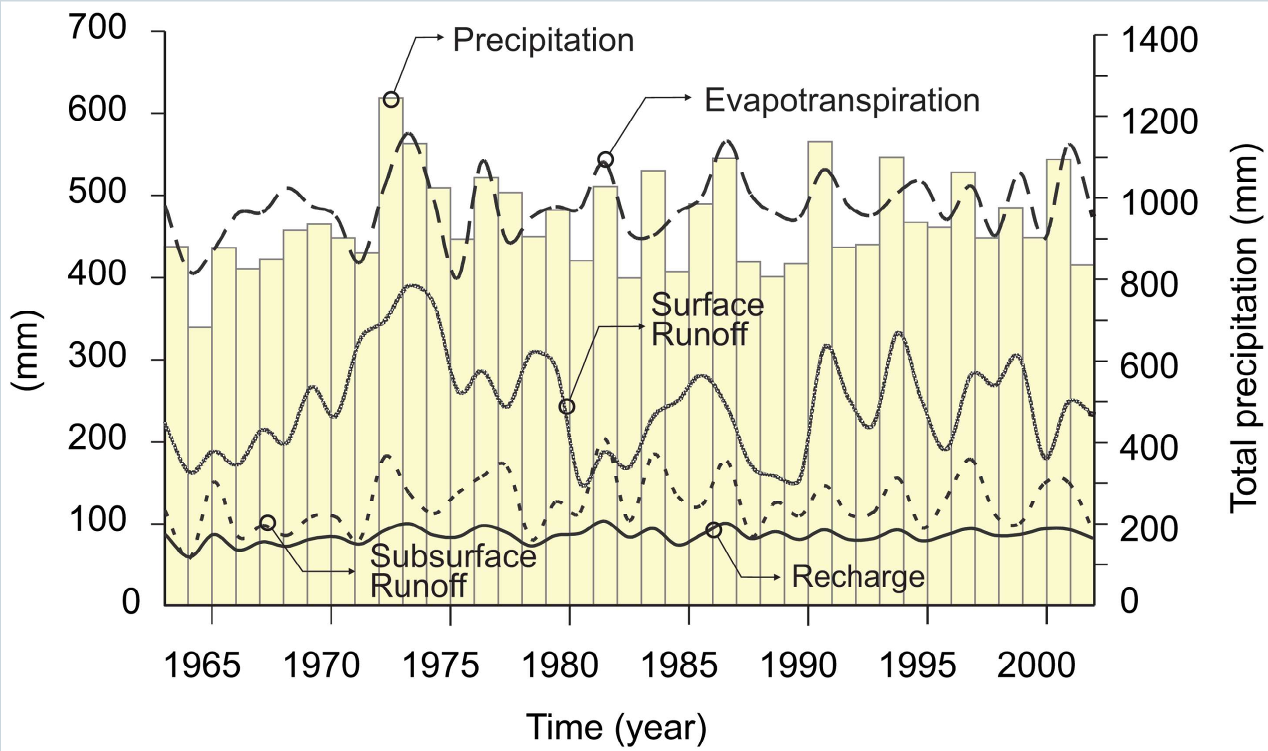


Figure 4.22 Annual variations in the water budget components.

a stronger influence on the daily values of the water budget components than precipitation. Two major recharge periods can be observed. In spring, when the temperatures rise above zero, the snowmelt contributes to a rapid increase of the recharge, subsurface runoff, and the evapotranspiration rates. In late winter, prior to this increase, surface runoff rate peaks as melted water runs off the still frozen soil. The second recharge peak occurs during the fall, as increased air humidity and decreased temperatures contribute to a reduced evapotranspiration rate which, in turn, increases the recharge rate. During this period, the vegetation cover decreases water uptake by vegetative roots and evaporation are minimal and more water becomes available for runoff and percolation.

Annual variations of the average water budget components for the considered period are illustrated in Figure 4.22. Evapotranspiration and surface runoff variations show direct correlation to precipitation fluctuations. The groundwater recharge rates are relatively constant and vary in a narrow range, with a standard deviation s = ± 9 mm/year. Recharge indicates a subdued response to precipitation fluctuations. The storage capacity of soil media overlying the regional aquifer units acts as a reservoir that releases water during the dry years and stores water surplus during the wetter years. The mean annual temperature remains fairly constant over the computed period, i.e., 6.3°C, in opposition to the daily variations of the water budget components where it assumes a major role, whereas in Figure 4.23, it is considered to have negligible impact on the total annual values.

The calibrated recharge model has been used to simulate water budget components under those extreme climate conditions considered

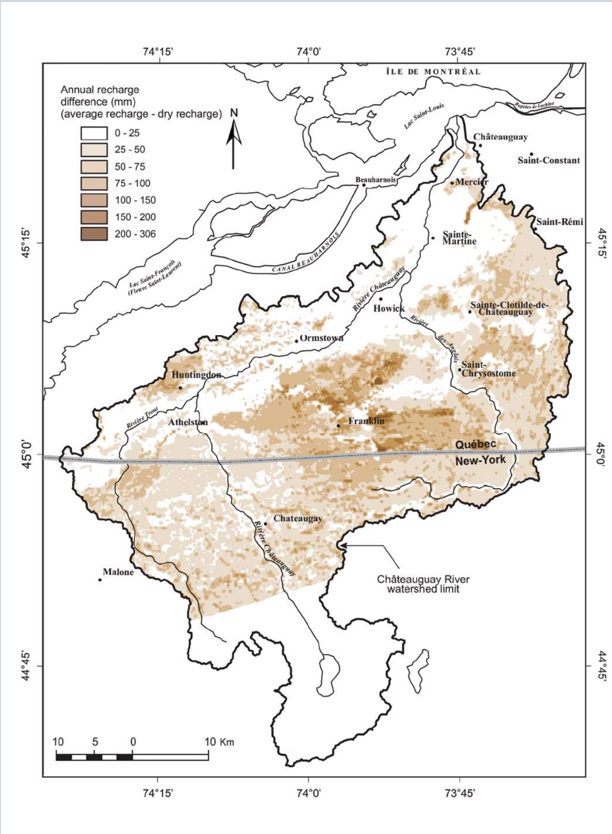


Figure 4.23 Spatial variation between average and dry recharge rates. Those areas with the greatest difference are thought to be sensitive to future dry climate conditions.

as representative for possible climate changes. Two scenarios, drought and prolonged humid conditions, were considered. Real climate data was used for these scenarios because recorded climate conditions varied considerably over the considered period, 1963–2001. Extreme conditions were defined by imposing climate data recorded for the driest (1964, precipitation=683 mm) and the wettest (1972, precipitation=1245 mm) year to the model (see Figure 4.22). The model was then run with the same input data repetitively over a period of six years, which allowed for stabilization of the soil saturation and output parameters. The results for the sixth year were those used for the analysis. These indicated a mean annual recharge rate of 51 mm/year for the drought and 99 mm/year for the humid scenario. In both cases, the spatial distribution of the recharge rate corresponded to that of the mean recharge rate shown in Figure 4.20. In other words, those areas that yielded the highest recharge rates under average climate conditions also did so under other climate conditions. A comparison to the average recharge rate for the basin (86 mm/year, mean annual precipitation=943 mm) revealed the fact that that the regional aquifers have an upper limit of how much groundwater can be replenished on an annual basis. A 32% precipitation increase yields a 15% increase in recharge rate. Further increase in precipitation would result in only limited recharge increase, and the current humid climate conditions yielded recharge rates close to this threshold. Conversely, recharge rate decreases more rapidly with precipitation decrease as a 37% decrease of precipitation yields a 67% decrease in the recharge rate.

Figure 4.23 shows the spatial difference between the average and the dry recharge rates computed for each grid cell. This map actually indicates those areas where, in absolute terms, the maximum decrease in recharge rate will occur in the case where dry climate conditions prevail. These areas coincide generally, but not necessarily, with those where highest recharge rate occurs under average climate conditions. The recharge map and the recharge sensitivity map define and confirm spatial distribution of those regions where most attention must be paid to the environment protection and its groundwater resource.

CANADA'S GROUNDWATER RESOURCES

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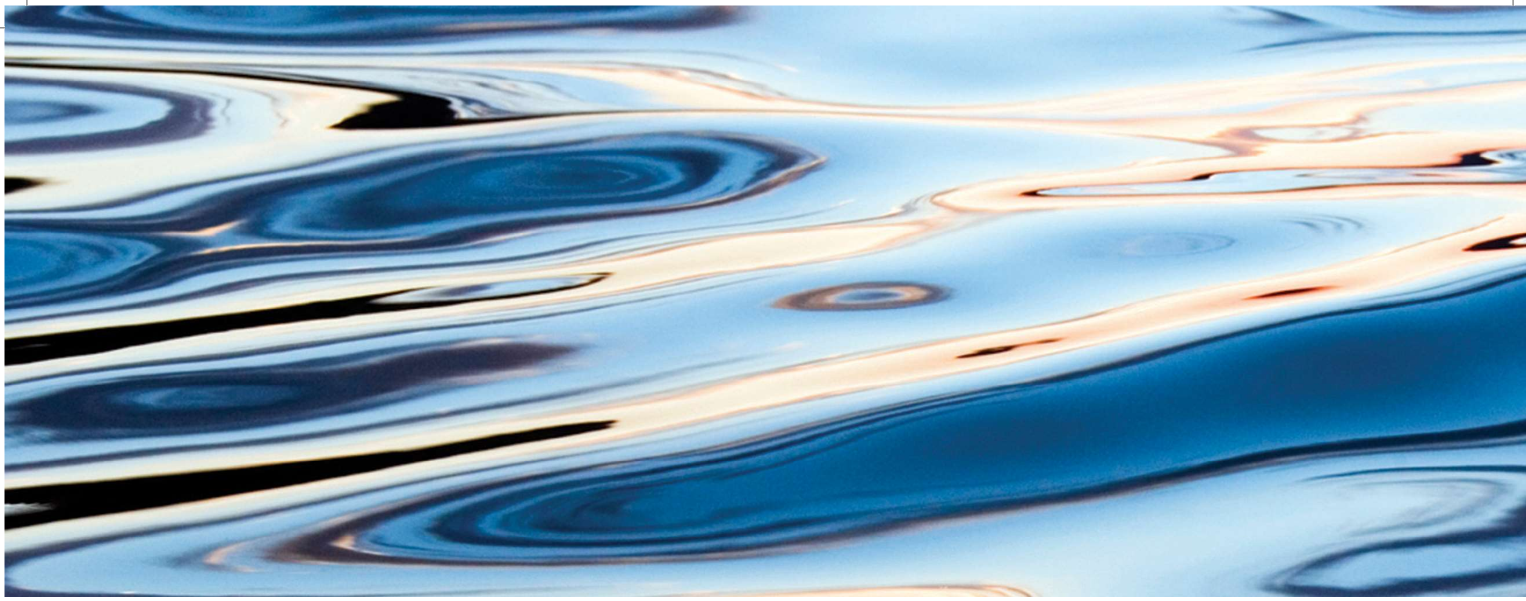
November, 2013



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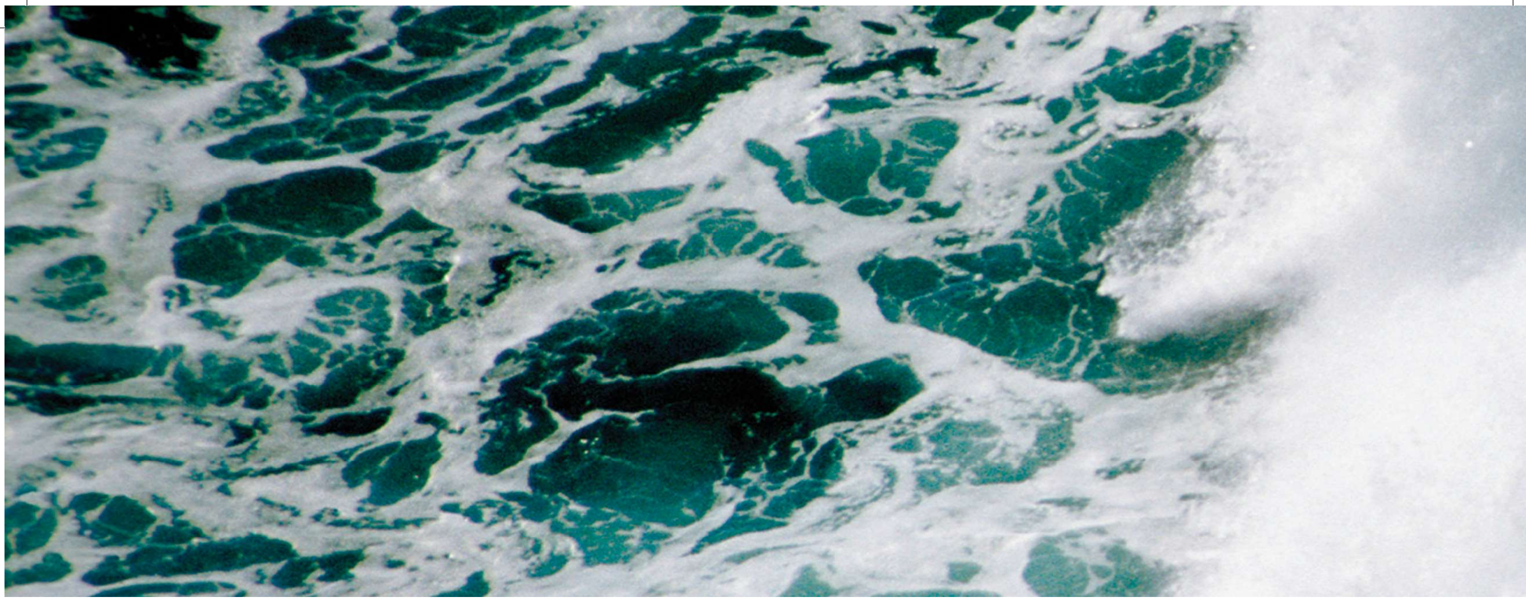

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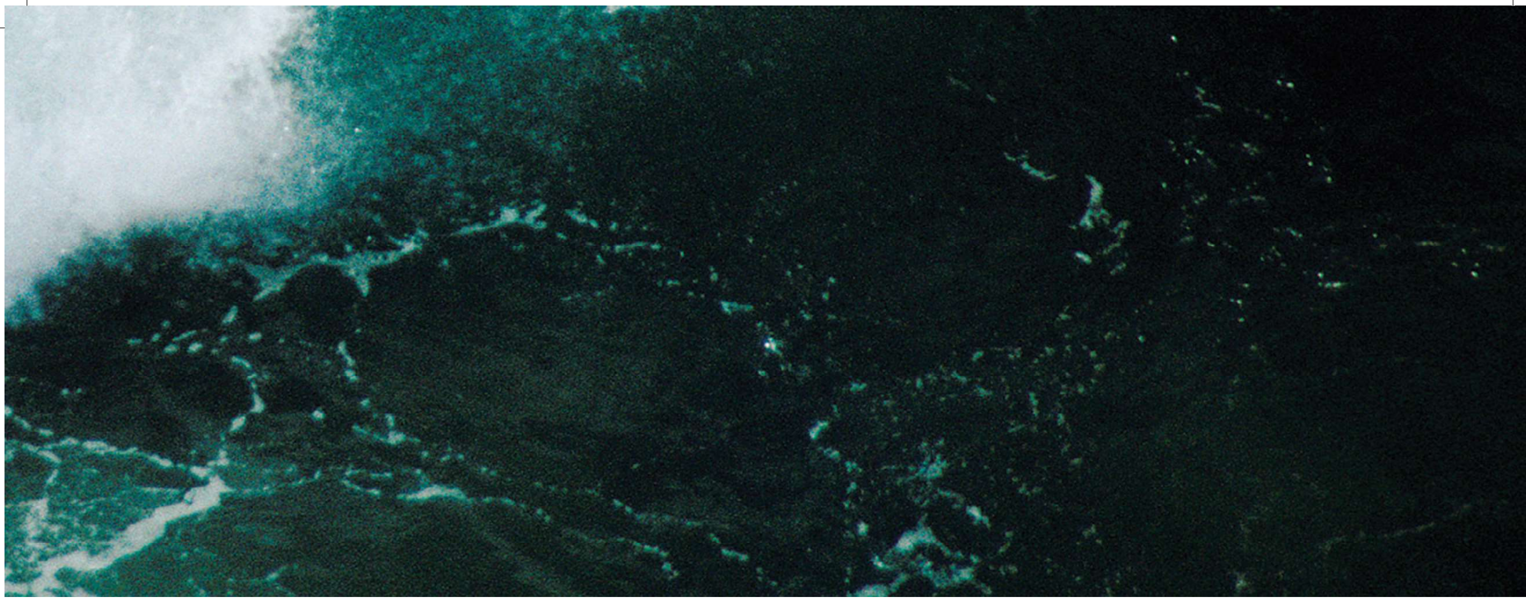

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