CONCEPTUAL WATER MODEL FOR THE HORN RIVER BASIN, NORTHEAST BRITISH COLUMBIA (PARTS OF NTS 094I, J, O, P)

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ABSTRACT

Work was undertaken to develop a conceptual water model for the Horn River Basin (HRB). Water models are needed for resource management because of a rapidly growing water demand associated with shale gas development in northeast British Columbia. Lumped-parameter models are easier to generate, but for the scale of watersheds in the HRB, distributed-parameter models are more appropriate. A representative distributed-parameter model already exists for the Liard Basin. Modeling the spatial distribution and interrelationship between evapotranspiration, permafrost and muskeg is challenging in this relatively flat-lying region of forests, fens, bogs, numerous small and shallow lakes and discontinuous permafrost. Groundwater represents the mechanism by which peatlands retain water, lakes and uplands exchange water, and streamwater quality and quantity is maintained. Information gaps were also identified.

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Key Words: Horn River Basin, surface water, climate, peatland, muskeg, permafrost

INTRODUCTION

Recent shale gas exploration and development in the Horn River Basin (HRB) region of northeastern British Columbia has created a demand for large amounts of water to hydraulically fracture shale and release trapped gas. To sustainably[1] develop surface-water resources to meet this growing demand, sound water models are required. This paper generates a conceptual model that summarizes the current understanding of the key catchment processes, dependencies and impacts on the water resource. It also highlights public information resources for the Horn River Basin.

Developing models in the Horn River Basin region is challenging. The region has low relief with patchy to widespread wetland. The wetland, predominantly sphagnum moss and black spruce forest, is universally referred to as muskeg, but is more properly classified as fens and bogs. Fens have a surface water–groundwater connection and network with lakes and streams to channel water off the landscape, while bogs function as reservoirs in isolation from groundwater and are interconnected only when the water table is sufficiently high. Another challenge to modeling is discontinuous permafrost and heavy seasonal ground frost in the region. Infiltration and overland flow are impacted by the depth to ice and the thickness of ground frost.

The goal of this paper is to outline important information sources for the HRB and identify features that will affect hydrological predictions in the HRB. This paper is divided into three sections: first, a discussion of water model types and regional models in existence for the HRB; second, a review of publicly available data and hydrogeological implications of important complexities particular to the HRB (including climate, permafrost, lakes and streams and muskeg); and third, a discussion of information gaps.

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[1]Sustainability is defined by the Council of Canadian Academies (2009) as the following:

(1) Protection of groundwater supplies from depletion: Sustainability requires that withdrawals can be maintained indefinitely without creating significant long-term declines in regional water levels.
(2) Protection of groundwater quality from contamination: Sustainability requires that groundwater quality is not compromised by significant degradation of its chemical or biological character.
(3) Protection of ecosystem viability: Sustainability requires that withdrawals do not significantly impinge on the contribution of groundwater to surface water supplies and the support of ecosystems. Human users will inevitably have some impact on pristine ecosystems.
(4) Achievement of economic and social well-being: Sustainability requires that allocation of groundwater maximizes its potential contribution to social well-being (interpreted to reflect both economic and noneconomic values).
(5) Application of good governance: Sustainability requires that decisions as to groundwater use are made transparently through informed public participation and with full account taken of ecosystem needs, intergenerational equity and the precautionary principle.

BACKGROUND

Horn River Basin Location and Geography

The Horn River Basin (Figure 1) is located in northeastern British Columbia between Fort Nelson and the Northwest Territories border (mostly in NTS map area 094O eastward into 094P and southward to 094J). Located in the Fort Nelson Lowland of the Alberta Plateau, the area has very low relief (300–730 m above sea level), with the Etsho Plateau forming a minor upland, oriented southeast in the central region (Holland, 1976). The Muskwa Uplands of the Rocky Mountains can just be seen along the southwestern map edge. Two major drainage systems are incised up to 150 m below the general level of the lowland: the Fort Nelson River and the Petitot River, which are tributaries of the Liard River system. In total, the Horn River Basin contains portions of three major watersheds (Figure 2), which drain into the Mackenzie River system. The Fort Nelson and Petitot rivers flow into the Liard River, whereas the Hay River drains into Great Slave Lake. Table 1 gives the overall area of the watersheds and the catchment area within the Horn River Basin (HRB) calculated on the basis of the subwatersheds. Subwatersheds are considered part of the HRB if any portion of the watershed is within the boundary.

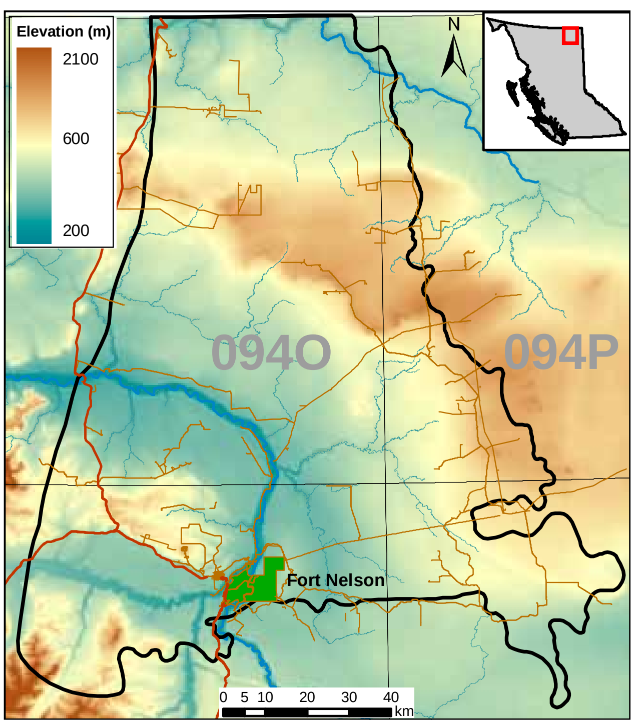

Figure 1. Topography of the Horn River Basin.

The broader HRB area is characterized by muskeg and a black spruce forest. Black spruce bogs are prevalent, especially in the northeast. In drier areas of the HRB, there are stands of white spruce and trembling aspen. Overall, forest productivity is generally low due to long, cold winters and short growing seasons. Winter temperatures average –18°C between November and February. There is discontinuous permafrost throughout the region. The region is relatively dry (annual average precipitation of approximately 450 mm; Environment Canada, 2009). Because of the low relief and muskeg-dominated headwaters, river flows tend to be stable and laminar in highly incised channels and waters have a tannic character. In the summer, flow volumes are low and water temperatures are high (Anderson et al., 2009).

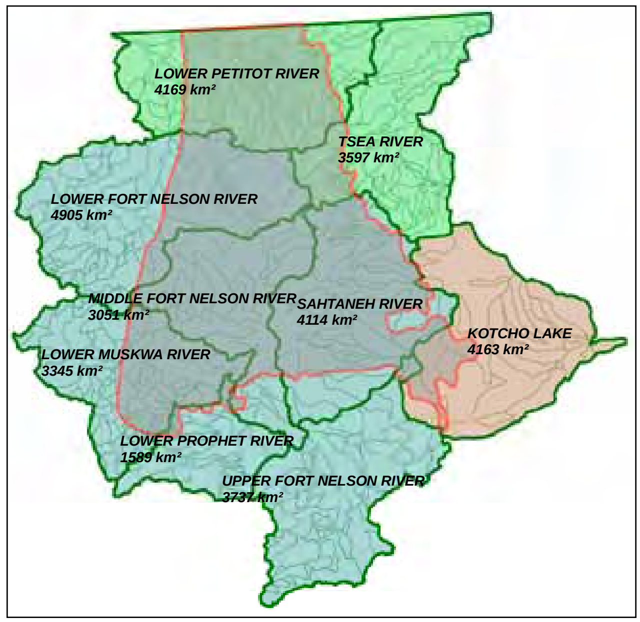

Figure 2. Watersheds of the Horn River Basin. Large colour groupings identify the three major watersheds (i.e., Petitot, Fort Nelson and Hay). Heavy green outlines define the watersheds. Subwatersheds are defined in light grey. The outline of the Horn River Basin is in red.

Demand for water in the HRB

The increased demand for water is primarily owed to hydraulic fracturing, the stimulation technique necessary for the economic development of tight and shale gas. To develop shale gas plays, hydraulic fracturing (herein referred to as 'fracs' or 'fracing') is used to create fractures in the shale that increase borehole access to the gas trapped in the rock. Fracing requires high pressures and injection rates to create fractures around the borehole. A proppant such as sand is added to the water to prevent the fractures from closing again. In the HRB, these fracs are generated with substantial volumes of water with estimates ranging from 1 200 m^3 to over 4 000 m^3 of water per frac depending on

TABLE 1. WATERSHEDS IN THE HORN RIVER BASIN.

Major watershed	Watershed	Total Area (km²)	Area in HRB (km²)
Fort Nelson River	Lower Fort Nelson River	4 905	2 980
	Lower Muskwa River	3 345	2 258
	Lower Prophet River	1 589	612
	Middle Fort Nelson River	3 051	3 001
	Sahtaneh River	4 114	3 780
	Upper Fort Nelson River	3 737	349
	TOTAL	**53 891**	**12 981**
Hay River	Kotcho Lake	4 163	2 812
	TOTAL	**8 098**	**2 813**
Petitot River	Lower Petitot River	4 169	3 447
	Tsea River	3 597	1 095
	TOTAL	**12 083**	**4 542**

geology and engineering (Johnson, 2009; Hayes, 2010). An average well may contain six to eighteen fracs, so the water requirement for one well could be as low as 7 000 m^3 or more than 60 000 m^3. The density of water demand depends partly on the well configuration. The HRB is expected to have three to eight wells per gas spacing unit (approximately 240 ha; R. Stefik, pers comm, 2008).

The amount of water used in multistage fracing of a well varies widely. Unlike vertical wells, whose length is constrained by the thickness of the formation, horizontal well laterals can be extended as far as technology allows. A better metric for water use in horizontal wells is water use 'intensity,' or water volume per unit lateral length. This metric is more informative and is easily scalable to future work (Bene et al., 2007). Early data indicates water use in the Horn River Basin is similar to that in the Barnett shale in Texas, where volumes range from 25 to 40 m^3 per lateral metre drilled.

Permission to use surface water for oil and gas–related activities is either licensed by the British Columbia Ministry of Environment or temporarily granted by the Oil and Gas Commission under Section 8 of the Water Act (Water Act, 2010). There is currently one active water license in the HRB. Temporary water permits are issued for a volume per diem or total volume per permit. As of October 2009, there were 210 active Section 8 permits in northeast BC.

Watershed Modeling

The most reliable decision support tool for water allocation decisions and water resource management is a water balance with supporting water model(s). Watershed models are necessary in the HRB to: 1) characterize and develop an understanding of groundwater and surface-water processes in the watershed and 2) provide spatial and temporal hydrological datasets for use by water-resource managers in decision-making concerning stream habitat, land use and water use.

The province encourages the use of water models in water-management plans. Recently, the Township of Langley, under Part 4 of the Water Act, collaborated with the province to develop British Columbia's first water-management plan. At its core were a conceptual model and a numerical hydrogeological model. The conceptual model is a descriptive model of the system based upon qualitative assumptions about its elements, interrelationships and system boundaries.

Water models are based on a water balance where the amount of water incoming to a region balances with the amount outgoing plus changes in groundwater storage. For example, a lake has a given volume of water that is maintained (balanced) by inputs (precipitation, inflowing streams and groundwater flow into the lake) and outputs (discharge from the lake, evaporation and transpiration and groundwater flow out of the lake). The goal is to generate an accurate and efficient simulation of water-system mechanics within the watershed.

To develop appropriate watershed models, information is required about

- precipitation and drainage area (to determine input volume of water);
- land use, soil types and permafrost (to determine how much water infiltrates to the water table);
- slope (to determine the rate water reaches the drainage outlet);
- land cover of vegetation and lake abundance and size (to determine rates of evapotranspiration);
- climate data (to determine seasonality, snow-water equivalent and runoff); and

- streamflow data and water-level data (to understand discharge and storage diversion data).

Water models can range from simple to very complex. There are two styles of watershed models: spatially lumped (low-resolution) models using basin-averaged input data and spatially distributed (high-resolution) models. Watershed models are strengthened if they can account for heterogeneity of vegetation, soils and land-use characteristics in the watershed; however, each additional component increases the complexity of the model. For the size of watersheds in the HRB, a spatially distributed model would be most appropriate because distributed models have resolutions on the order of 150 m. The most commonly used distributed hydrological models, in order of decreasing popularity, are HBV, HEC–HMS, UBCWM, TOPMODEL, HSPF, SWAT, SHE, SAC–SMA, VIC, DHSVM and WATFLOOD (Beckers et al., 2009).

A ‘conceptual model’ defines the base assumptions employed to make the model simulate reality. The assumptions used when modeling can dramatically affect the results. Table 2 exemplifies the sensitivity of a single predicted variable, evapotranspiration, to different assumptions in a low-relief northern boreal forest (northern study site for BOREAS; Soulis and Seglenieks, 2005). For three land-cover types, evapotranspiration was calculated three ways: 1) a simple model with no subsurface water transport and no permafrost, 2) a model allowing for subsurface lateral transport of water and no permafrost and 3) a model allowing for subsurface lateral transport of water and ice in the soil. The resulting estimates of evapotranspiration vary substantially across the three models. Differences are not consistent; they change with the season (summer versus fall) and the saturation of the soil (wet versus dry forest). The conceptual model dictates whether permafrost or lateral flow is important.

Modeling Complexities of the HRB

Water modeling in the HRB is complicated by several considerations, including low topography, peatland, discontinuous permafrost and lack of independent data from monitoring stations.

The HRB is an area of low relief. In high-relief areas, drainage pathways are clearly delineated, rivers respond quickly to rainfall events and infiltration periods are limited. In low-relief areas, total discharge is low because snow water is retained on the landscape and stored in soils above frost and in wetlands through spring. Drainage is slow, water infiltration can occur across broad areas and streams can be difficult to identify (Figure 3). The drainage characteristics are typified in the following account:

> *“Few of the reaches sampled were found to be streams (17%). The remainder of the reaches assessed were wetlands, unchannelized drainage areas, or areas where no visible channel was found…Watercourses identified as intermittent on TRIM maps (as indicated by discontinuous lines) were often found to be swamps or to have no visible channel…”*
>
> *(Golder Associates Ltd., 1998)*

TABLE 2. EVAPOTRANSPIRATION FOR THE BOREAS STUDY AREA USING DIFFERENT WATCLASS MODELS (SOULIS AND SEGLENIEKS, 2005).

Modeled Evaporation		**No lateral flow, no permafrost**	**Lateral flow, no permafrost**		**Lateral flow, permafrost**	
		mm	mm	***% change***	mm	***% change***
a) Dry Forest	Spring	77	92	***19***	169	***119***
	Summer	156	134	***−14***	203	***30***
	Fall	10	3	***−75***	24	***137***
	Annual	243	229	***−6***	400	***65***
b) Wet Forest	Spring	115	102	***−11***	149	***30***
	Summer	148	124	***−17***	156	***5***
	Fall	9	−3	***−130***	13	***45***
	Annual	274	224	***−19***	321	***17***
c) Wetland	Spring	118	105	***−11***	159	***35***
	Summer	155	126	***−19***	166	***7***
	Fall	10	−1	***−108***	13	***31***
	Annual	285	232	***−19***	341	***20***

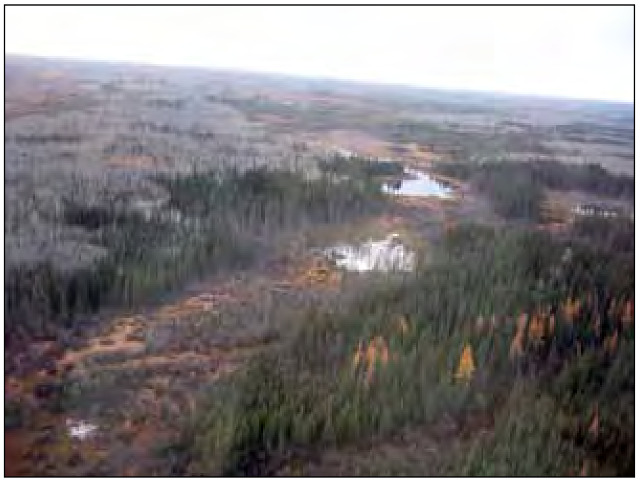

Figure 3. Unchannelized drainage in wetland. Photo by Elizabeth Johnson.

The presence and type of wetland is very important to hydrological modeling. For most land-cover types, infiltration allows the downward flux of precipitation to the water table. In peatland, however, subsurface shallow flow is mainly horizontal and there is virtually no downward conductivity of water. Rainfall runoff is routed through bogs and fens. Fens promote lateral flow, whereas bogs efficiently store water. The quantity of fen coverage on the landscape is directly related to runoff, whereas the prevalence of bogs is inversely related to runoff. Lack of topographic relief, absence of well-defined channels and shallow groundwater tables all combine to make peatlands behave hydrologically like unregulated shallow reservoirs. Beyond wetlands themselves, the depth and texture of surficial deposits influence the extent, ephemeral nature and type of flowpath connecting slopes to streams, wetlands and lakes (Devito et al., 2005).

The HRB lies in an area of discontinuous permafrost. The distribution of permafrost is important to hydrological modeling as the presence of ice impedes the downward infiltration of surface water and limits water storage to a thin surficial layer. The discontinuous nature of permafrost makes it difficult to assess soil as a lumped parameter or to define similar hydrological response units.

Finally, modeling of the hydrological systems in the HRB is limited by the lack of ground-based monitoring for calibration. River gauge stations tend to be developed in densely populated areas and on large water systems. The Horn River Basin is a remote region with mostly smaller, slow-moving water systems (other than the Fort Nelson and Petitot rivers). The most pertinent active station for the HRB is on the Liard River at Fort Liard in the Northwest Territories. There are discontinued stations on the Fort Nelson and Petitot rivers. Weather data is monitored hourly at Fort Nelson, BC and Fort Liard, Northwest Territories dating back to 1953 and 1973, respectively. Groundwater data from shallow-water wells is located near Fort Nelson.

AVAILABLE DATA

Water availability can be estimated from models that incorporate climate (precipitation, temperature, evapotranspiration, wind, radiation, pressure, etc.), land cover, soil type, topography and stream discharge. The remainder of this paper details available data useful for structuring and parameterizing models. Much of this information is available indirectly. One source of information widely available but generally not considered by resource managers are large-scale hydrological models. These models provide coarse-resolution information and constrain unknown variables for the region. There are several models available for cross reference and the input data and results are often available without fee.

Hydrogeological modeling

Acquiring the necessary data can be difficult. There are many global and regional databases that provide data and hydrological models on major drainage systems of the world. The Water Systems Analysis Group at the University of New Hampshire provides a comprehensive listing of global hydrological consortiums that provide data (Water Systems Analysis Group, 2010). In particular, ArcticRIMS (2000; Rapid Integrated Monitoring System) combines several well-established datasets to produce time-varying, region-wide land surface water budgets across the pan-Arctic drainage region (including the Mackenzie Basin). Algorithms include vapour flux convergence, a satellite-derived snow product, a permafrost water-balance model, a water transport model and simulated river networks. Products include components of the water cycle (atmospheric convergence, precipitation, evapotranspiration, change in soil, snowpack, shallow groundwater, runoff and river discharge) and estimates of potential error. Nominal resolution is 25 km with daily time steps.

Researchers within the Mackenzie GEWEX (Global Energy and Water Cycle Experiment) Study (MAGS) developed distributed hydrological models for larger basins with the Mackenzie River drainage (e.g., Liard River, Peace River, Athabasca River) using WATFLOOD and WATCLASS hydrological models (Burn et al., 2004; Soulis and Burn, 2004). The model was constructed at 20 km resolution. Notice the excellent agreement between observed discharge and modeled discharge for the Liard River at Fort Liard in Figure 4. The Nash goodness-of-fit coefficient is 0.77 (where values less than 0 indicate the observed value is a better predictor and 1 represents a perfect fit; Soulis and Seglenieks, 2005). This station is just downstream of the confluence of the Fort Nelson River and the Liard River.

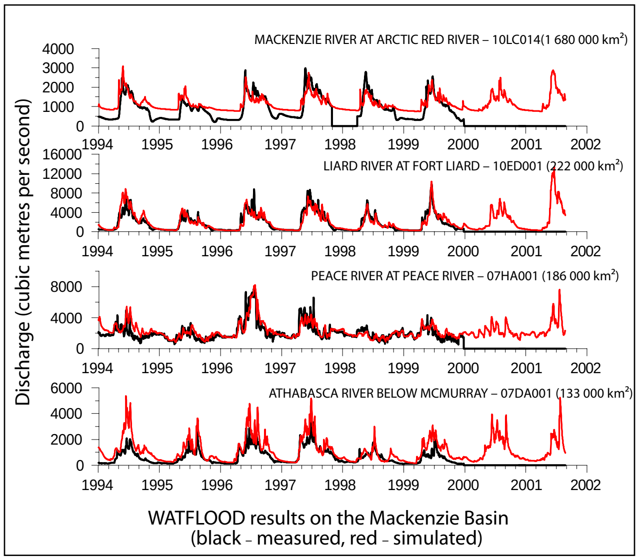

Figure 4. Observed and simulated discharge in the Mackenzie Basin using the WATFLOOD distributed hydrological model (Soulis and Burn, 2004).

A separate independent method for validating the WATFLOOD hydrological model of the Liard Basin involved a comparison of estimated and observed water storage. Remotely sensed data from GRACE (the Gravity Recovery and Climate Experiment) produces integrated geopotential anomalies that relate directly to stored water. In 2002, the GRACE satellite observations showed a gain in storage of 10 mm compared to 7 mm for the WATFLOOD model. In 2003, the GRACE satellite estimated a loss of –1.3 mm, whereas WATFLOOD averages were –3.1 mm (Soulis and Seglenieks, 2005). Recent work by Yirdaw et al. (2009) indicates 70% correlation in the Liard Basin between GRACE total water storage and storage calculated from an atmospherically based water balance model between 2002 and 2005 (Figure 5).

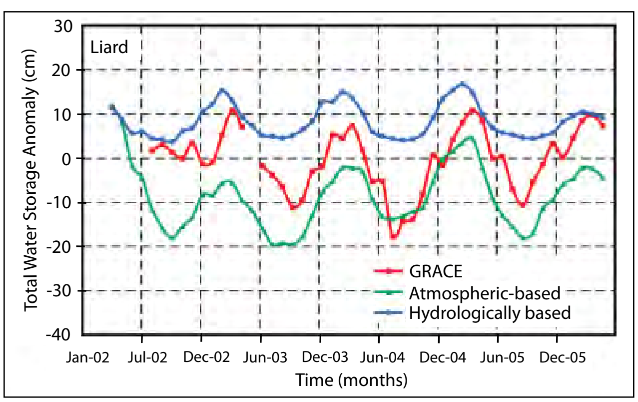

Figure 5. Total water storage anomalies derived from GRACE, atmospheric-based water balance techniques and WATCLASS for the Liard sub-basin (Yirdaw et al., 2009).

The Dartmouth Flood Observatory (DFO) provides another independent means of validating estimated discharge. River discharge is estimated using a surface wetness method that relates changes in river width to changes in river volume. The DFO's River Watch tool uses satellite-based brightness temperatures to estimate river discharge at over 2500 selected river measurement sites globally. The Liard River at Fort Liard (coordinates 61.3703, –121.8324) is one of those sites. They have collected data from 2002 to 2009 (Figure 6).

Models for watersheds in the 10–50 km² range should be consistent with models for the larger Liard River watershed. They require finer resolution on spatial input parameters like vegetation and climate and are much more sensitive to the effects of muskeg, soil, water-table elevation and local topography.

Climate Data

Climate data are available from several sources. Climate data from the Fort Nelson weather station (Fort Nelson A at 58.84°N, 122.6°W) are available from Environment Canada. Daily data exists from 1937 to present. Ground snow data collection began in 1955. The British Columbia Ministry of Environment provides Fort Nelson snow pillow data from 1966 to present. Meteorological researchers have used climate observations from across Canada to create gridded data at roughly 50 km spacing. A widely available 30 year set of temperature and precipitation has been created for the interval of 1961 to 1990 (Hopkinson, 2000; IPCC, 2008; Girardin et al., 2006). Agriculture Canada has recently released 10 km gridded climatic data from across Canada for 2004–2008, which may be useful for finer-resolution watershed models. Additionally, there are many gridded regional and global datasets where climate is estimated from models including the National Centers for Environmental Prediction — Global Reanalysis 2 (NCEP-R2), the European Centre for Medium-Range Weather Forecasts 40 year Global Reanalysis (ERA-40) and the Canadian Meteorological Centre (CMC) Global Environmental Multiscale model (GEM) Regional Analysis, to name a few. These models incorporate multiple parameters of observed and remotely sensed data to estimate climate variables. This modeled data is commonly available at coarse resolution (2.5°) but modeled climate is a rapidly advancing field of study and finer-resolution models (0.5°) are becoming available.

Spatial Analysis

Gridded data are useful for understanding the spatial variability of climate across the Horn River Basin. The Meteorological Service of Canada interpolated data from 1961

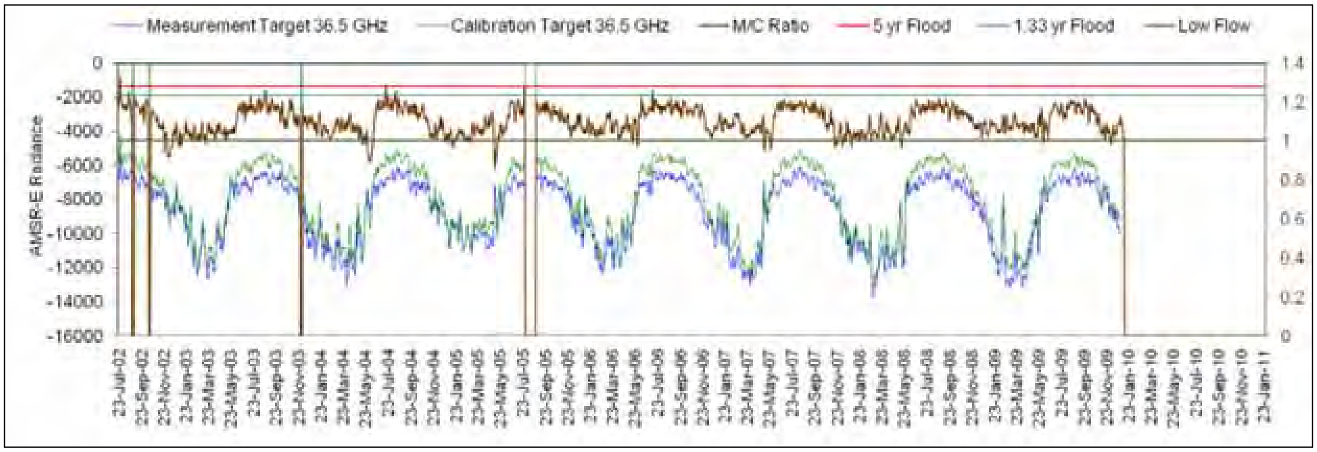

Figure 6. Discharge from the Liard River at Fort Liard as measured by 36 GHz brightness temperature. Discharge is estimated directly via a rating equation from the remote-sensing data (M/C ratio) shown below. The green line records an upwelling microwave emission from a land parcel (~10 km × 10 km) near the river and the blue line from a parcel centred over the river. The brown line is their ratio (M/C; scale on right) and is used to estimate river discharge via a rating equation (Dartmouth Flood Observatory, 2009).

to 1990 from climate observation stations across Canada. Monthly mean temperature and total precipitation data were interpolated to a 50 km grid on polar stereographic secant projection true at 60°N aligned north along 111°W (Hopkinson, 2000; IPCC, 2008). The size of the Horn River Basin (over 150 km long and from 50 to 140 km wide) is large enough to include more than ten cells.

Mean monthly temperature and precipitation data were averaged for three-month seasons (December-January-February, DJF; March-April-May, MAM; June-July-August, JJA; and September-October-November, SON). Figures 7 and 8 show the spatial variation in surface air temperature and precipitation, respectively, across the Horn River Basin for all four seasons.

In this northern region of low topographic relief, temperature varies mostly with latitude as a function of solar radiation: colder in the north, warmer in the south. The temperature is moderated by the Muskwa Ranges of the Rocky Mountains to the west of the HRB (see Figure 1). The air is warmer proximal to the base of the mountains in the winter.

Precipitation is governed by the Rocky Mountains. Precipitation patterns run parallel to the mountains with regions proximal to the mountains having heavier snow accumulation in winter and greater rainfall in summer. Annual precipitation (Figure 9) varies widely between 420 and 510 mm per year with the driest areas along the eastern edge of the HRB. The spatial distribution of wetland and lakes does not correlate with precipitation abundance. For example, land cover over the Petitot Plain is dense wetland, yet that region receives far less precipitation than in the southwest.

Evapotranspiration (ET) is the dominant means of water loss in the area, yet it is very difficult to quantify precisely. Studies in the Liard River Basin from just north of the HRB indicate that evapotranspiration accounts for two-thirds to three-quarters of the annual precipitation input (Quinton and Hayashi, 2005). Quinton and Hayashi (2005) estimated annual ET rates of 241, 245, 271 and 297 mm in the nearby Birch, Blackstone, Jean-Marie and Scotty watersheds, respectively. They confirmed this estimate using a chloride mass-balance approach in Scotty Creek (282 mm/yr). There is more than 5% error between different ET measurement styles for Scotty Creek, but this is an excellent correlation given the uncertainties inherent in ET measurement.

TIME SERIES ANALYSIS

Data were reorganized and analyzed to generate a mean, maximum, minimum and standard deviation for each calendar day (Julian day) for up to 61 years of data. Melting degree days (the temperature difference above 0°C over time) were calculated from temperature data. Snow on the ground measurements were converted to snow water equivalent (SWE) using mean snow density measurements from the British Columbia Ministry of Environment's Fort Nelson snow pillow data (1966–present) (BC Ministry of Environment, 2009). The Canadian Drought Code (CDC) was calculated using temperature and potential evapotranspiration.

Seasonal Breaks

Seasonal breaks were evaluated using melting degree days, snow on the ground and mean temperature and precipitation type. Figure 10 shows the changing volume of snow throughout the year with accumulation beginning in September and loss (melting) beginning in March and continuing until no snow exists in May. Rainfall begins in April and continues through November.

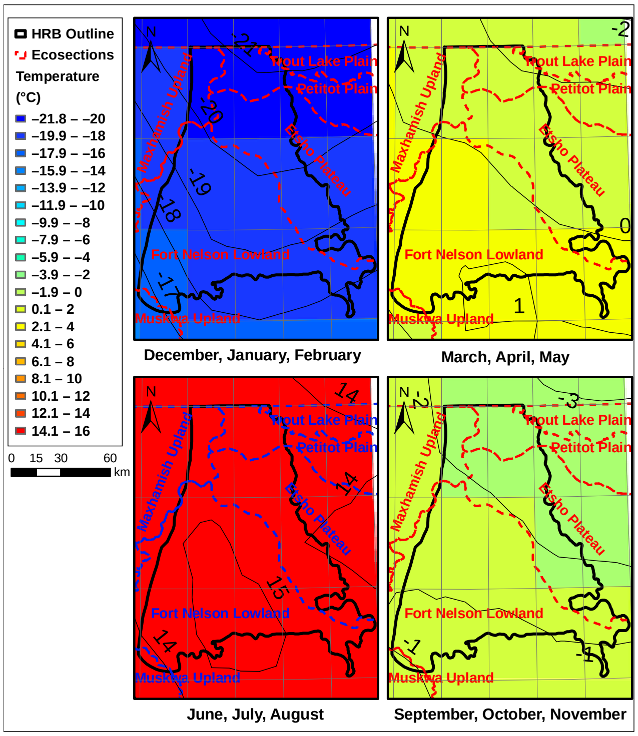

Figure 7. Seasonal surface air temperature in the HRB.

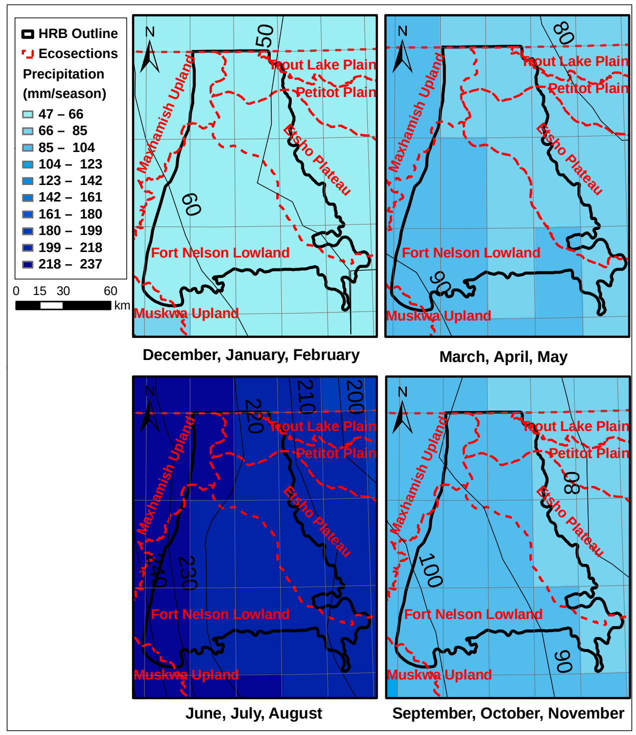

Figure 8. Seasonal precipitation as the sum of three months.

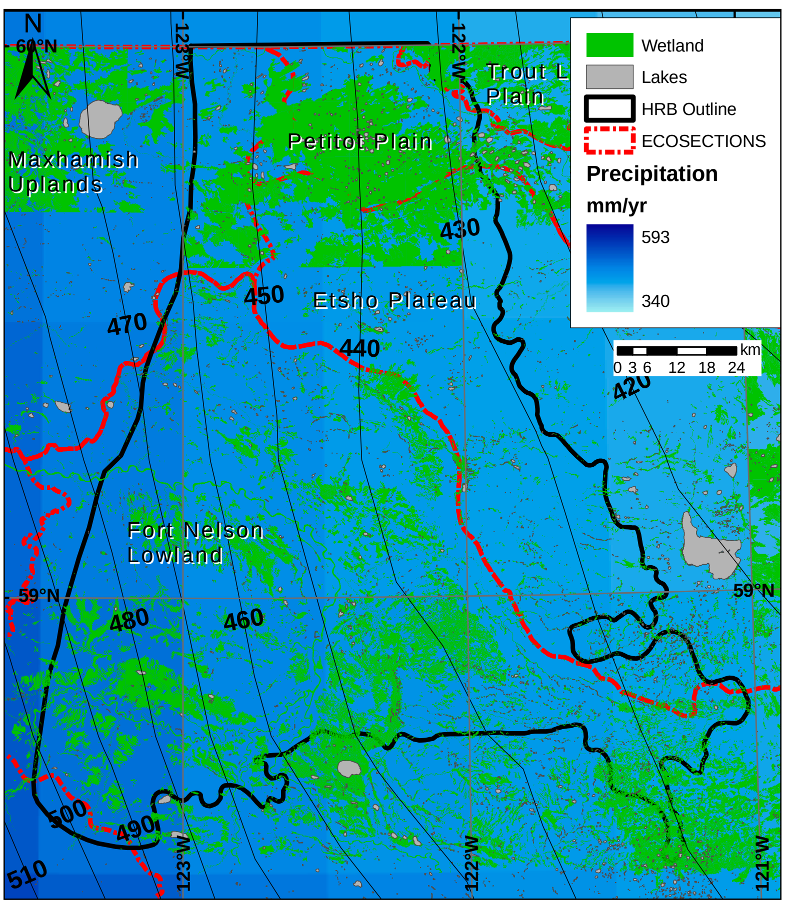

Figure 9. Annual precipitation in the HRB. Precipitation decreases to the east. Wetland and lake distribution is unrelated to regional precipitation patterns.

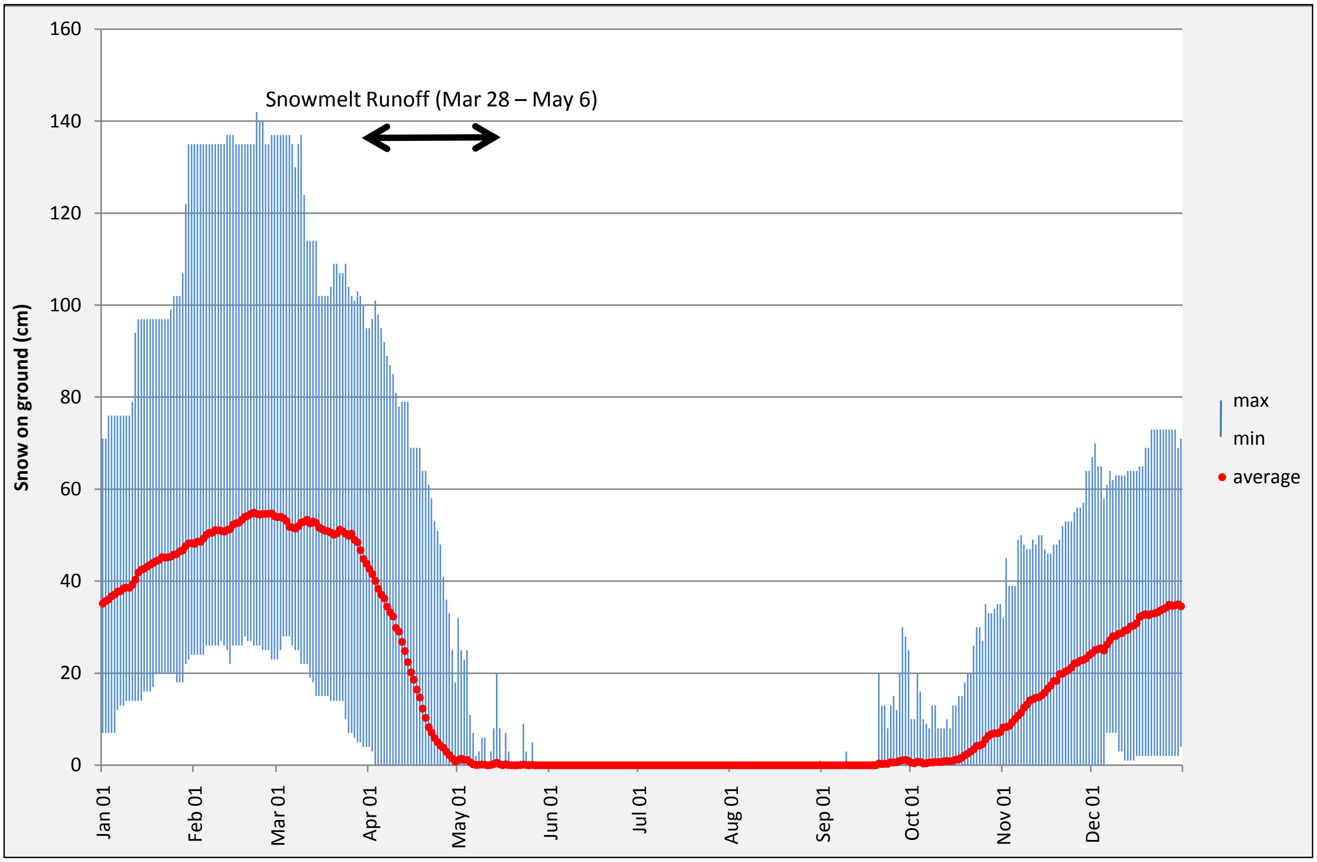

Figure 10. Snowpack volume changes throughout the year based on 54 years of data from the Fort Nelson weather station. Values in snow-water equivalent.

For hydrogeological modeling purposes, the change in season from winter to spring begins is when snowmelt runoff begins. The change in season is marked by a precipitous decrease in the snow on the ground or SWE after ambient daily temperatures rise. Melting degree days is a metric that represents the degrees centigrade above melting over the time causing melting. In Fort Nelson, SWE drops after an average of 2.5 melting degree days (Figure 11).

The snowmelt was charted against degree days to determine the distribution across a calendar year. Figure 12 shows a traditionally shaped curve of snowmelt relative to melting degree days. The historical averages were cross referenced with the melting degree day to determine the calendar date for the change of season. Because there is noise at the beginning and end of the warm season, a normalized curve was used to cross reference 2.5 melting degree days with March 28. Snowmelt runoff declines after eight melting degree days (Figure 11), which approximately corresponds to May 6 (Figure 12).

If climate change was not a factor, then March 28 would be the fixed date for the start of snowmelt. Figure 13 charts the number of melting degree days between February 1 and April 1 over time. In 1948, there was an average of three melting degree days before April 30, but there are currently five melting degree days. Regression analysis shows an increase of one melting degree day for every 27.4 years.

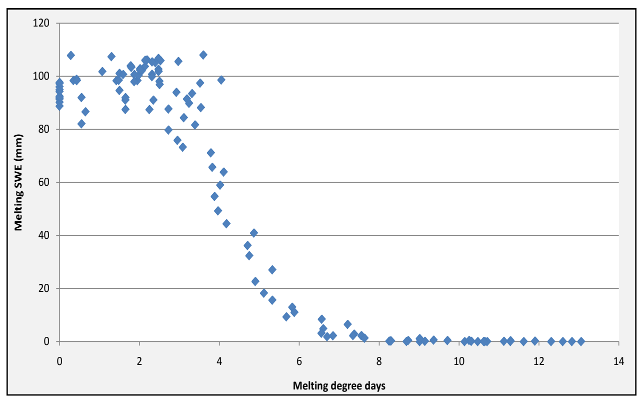

Figure 11. Snowmelt as a function of melting degree days. The graph demonstrates the rate at which the snowpack decreases with increasing numbers of days at temperatures above 0°C.

Fort Nelson is becoming progressively warmer earlier in the season and runoff is occurring earlier.

The final seasonal break occurs when rainfall diminishes and snowfall is retained on the landscape. Figure 14 shows the three runoff phases of the year with predominantly snowmelt-based runoff from March 28 to May 6, mainly rainfall runoff from May 7 to October 15 and winter low flow from October 16 to March 27.

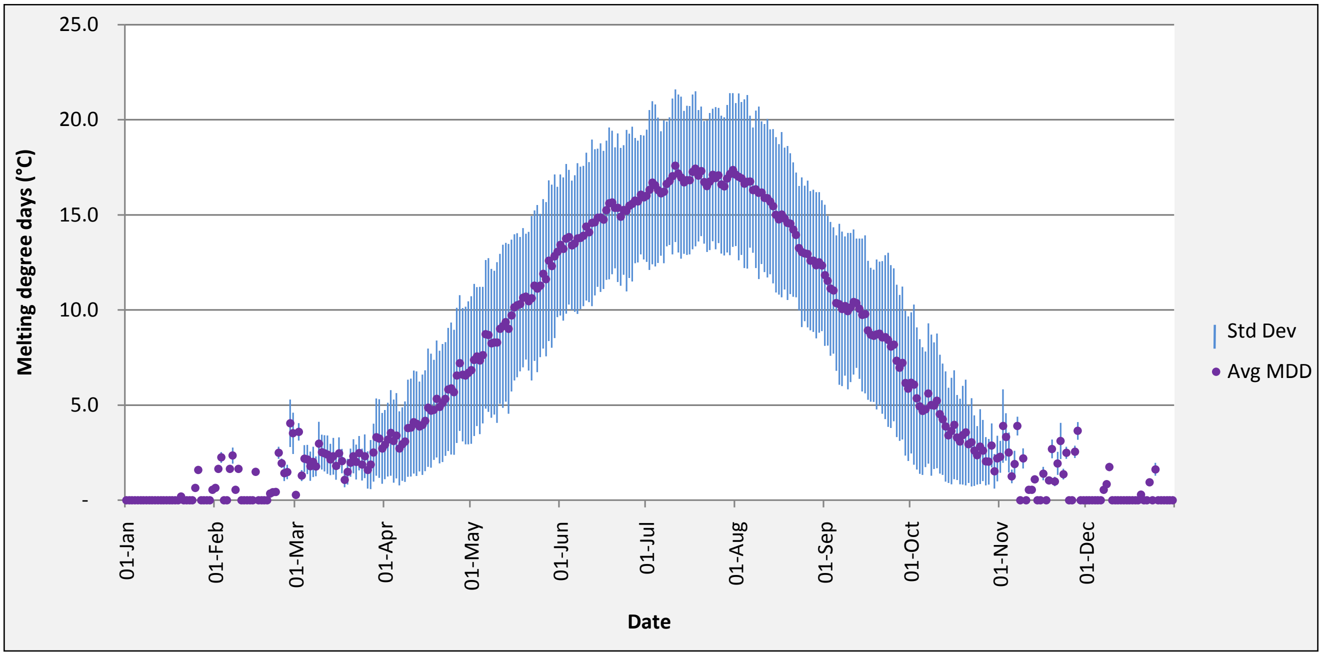

Figure 12. Melting degree days in the Horn River Basin indicate the portion of the year when melting will occur.

Drought

The Canadian Drought Code (CDC) is a meteorological estimate that uses evapotranspiration and precipitation to model water stored in the soil. It estimates soil dryness at an average depth of 20 cm and serves to warn when lower layers of deep partly decomposed organic material may be drier than the upper layers (Girardin et al., 2006). It was designed to indicate slow drying in Canadian boreal forests as a part of the Canadian Forest Weather Index system. The CDC calculation makes no allowance for seasonal changes in vegetation but does account for daylight length. The index accounts for the effect of snowmelt and provides an indicator of water-table depth (Girardin et al., 2004). Figure 15 shows variation in CDC across 61 years. A minimum CDC value of zero represents soil saturation, whereas a rating of 200 indicates high drought and a rating of over 300 indicates severe drought. Each year, CDC values begin low and climb as the cumulative effects of heat and lack of precipitation reduce soil moisture. The month during which maximum drought severity is attained is September (Figure 16).

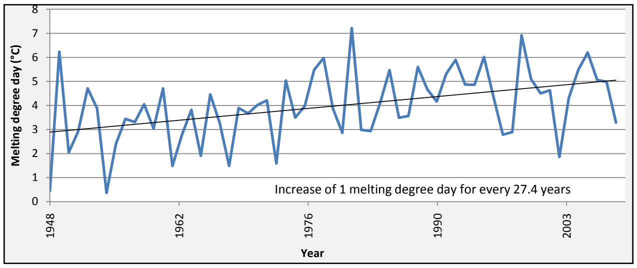

Figure 13. Average annual number of melting degree days each spring between February 1 and April 30. Regression analysis shows an increase of one melting degree day for every 27.4 years.

Using Fort Nelson climate data, there is no significant trend to drought severity over time except for a slight decrease in drought severity. There is a cyclical pattern to dryness severity with a periodicity of approximately 10 years. This cyclicity does not align with El Niño or La Niña weather events. The wettest years are 1948, 1957, 1962, 1977, 1984, 1988, 1997 and 2007. Currently the climate is in a wetter phase. This should be taken into consideration when collecting new baseline data in the HRB. Lake and river levels measured now are significantly higher than during drought years. Current and accepted water withdrawal volumes may not be tenable during a drought period.

Dry spells have a direct effect on runoff. Soil moisture deficits during dry years can have a significant impact on the magnitude of the subsequent spring runoff. The length of the dry periods may control minimum runoff more than the actual values of rainfall or evapotranspiration (Metcalfe and Buttle, 1999).

Permafrost

The entire HRB lies in a region of discontinuous permafrost. Systematic observations of the distribution and thickness of the permafrost were made in the HRB along a traverse extending northeastward from Fort Nelson across the southwest-facing Etsho Escarpment and to the boundary of the Northwest Territories (Figure 17; Crampton, 1977). The thickness and hardness of permafrost increases with increasing latitude, but decreases in areas with increased insulation (e.g., southwest-facing slopes). In general, permafrost throughout the HRB exists at 51 cm (20 in.) below the surface and varies from 38 to 102 cm (15–40 in.).

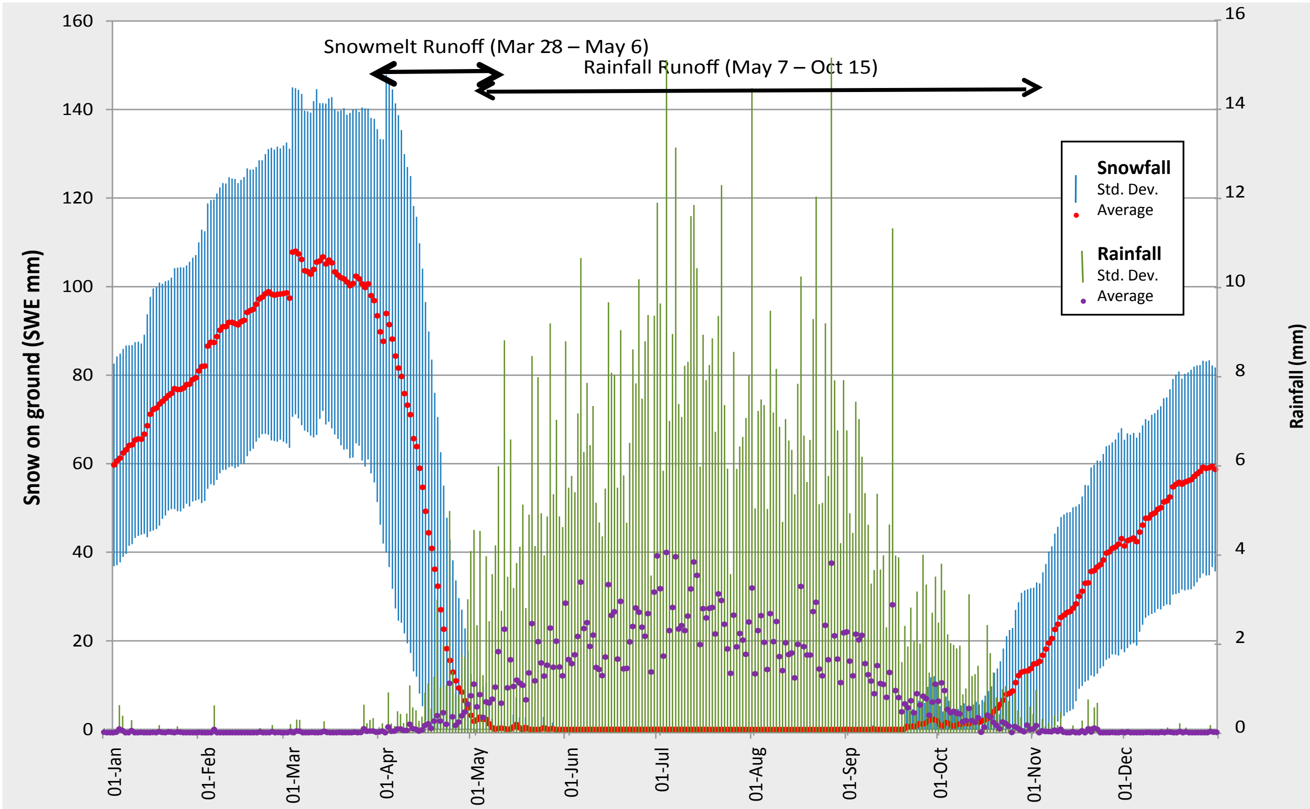

Figure 14. Precipitation in Fort Nelson as three runoff phases of the year with mainly snowmelt runoff from March 28 to May 6, mainly rainfall runoff from May 7 to October 15 and winter low flow from October 16 to March 27.

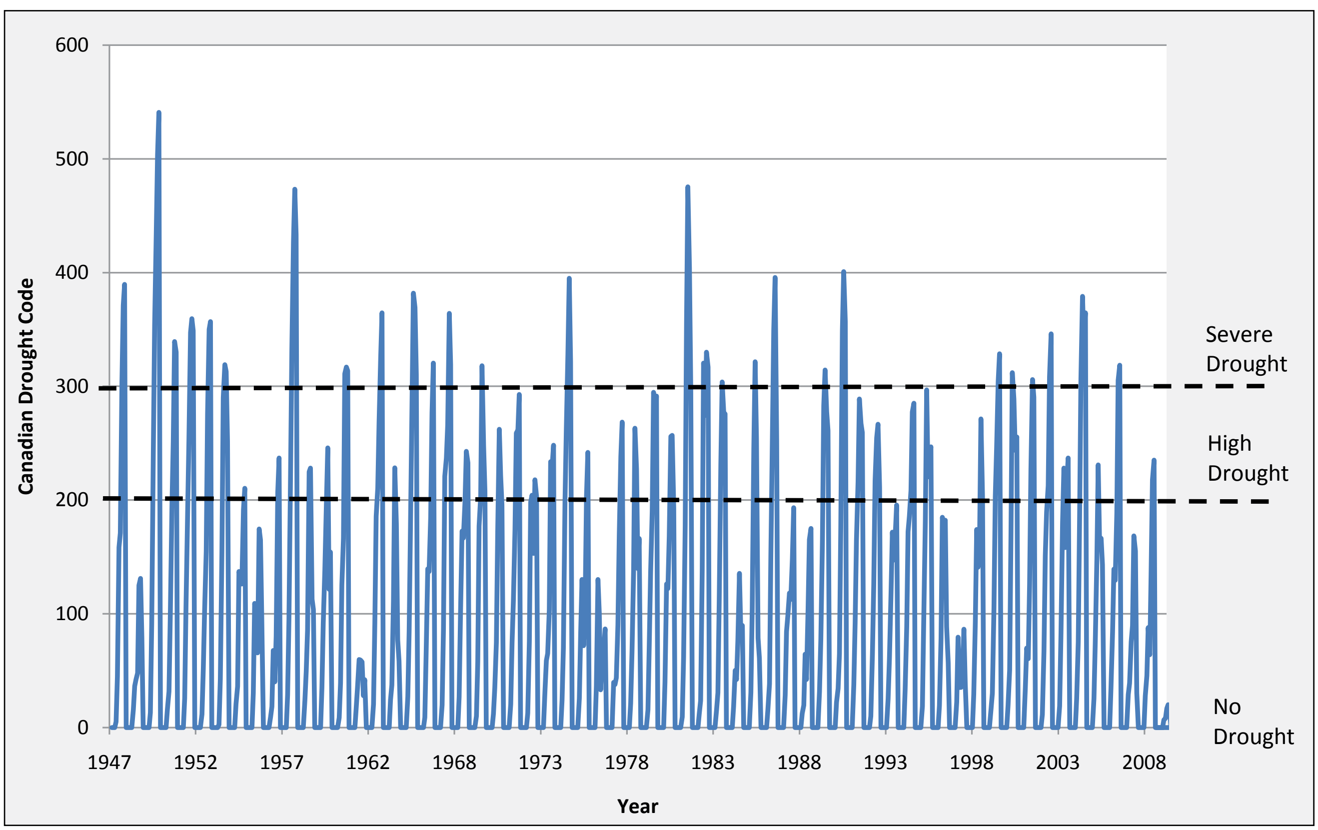

Figure 15. Canadian Drought Code averages for each month from 1948 to 2009.

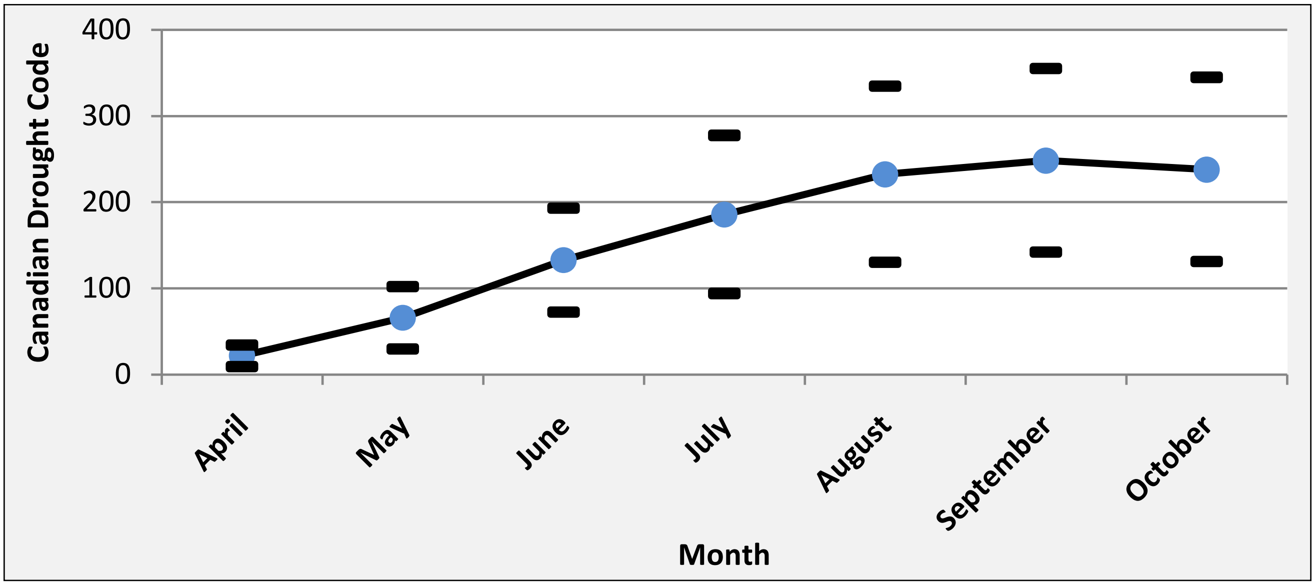

Figure 16. Average (with error bars) of the mean monthly drought index value (April–October) for the period of 1948 to 2009.

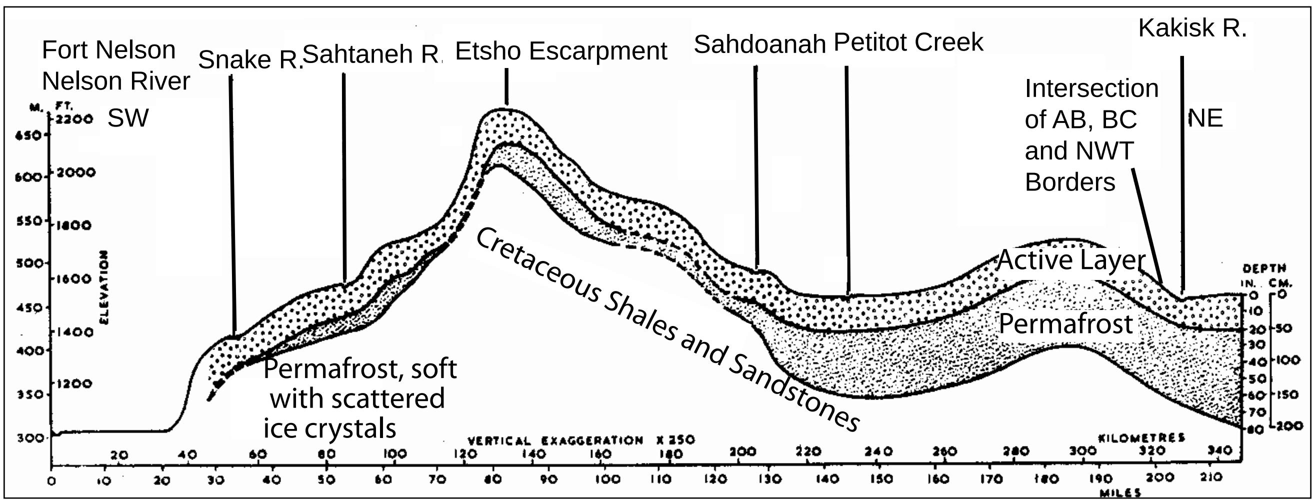

Figure 17. Longitudinal profile of the terrain across the Etsho Escarpment showing the distribution of permafrost. Vertical scale is 1:250 (Crampton, 1977).

Thermal conductivity associated with increased wetness decreases permafrost locally. However, in areas of discontinuous or isolated permafrost, ice can be found under bogs with raised sphagnum-covered mounds (M. Geertsema, pers comm, 2010). A rise in the water table (flooding) can cause permafrost degradation and the development of thermokarst terrain in areas of black spruce bog (Rennie, 1978). Areas where permafrost has melted may be marked by the development of thermokarst lakes and a transition to fen conditions (Figure 18). Permafrost will melt in areas of road development because compaction of surface soils reduces the porosity of the overlying soil and its attendant insulating capacity.


Figure 18. Thermokarst lakes in the Horn River Basin. Photo Courtesy of Adrian Hickin.

The presence of permafrost affects watershed discharge. As the temperatures climb in the spring and summer, the depth to the top of the permafrost falls so the thickness of the ice-free soil horizon is increased. The regulation of river flow by the soil horizon is restricted to short periods in the summer while there is an effective storage capacity in the soil horizon, but before the water table is lowered by evaporation.

Modeling subsurface water flow below the top of permafrost is challenging, because ice in the soil horizon affects horizontal and vertical hydraulic conductivity and porosity. The degree of soil saturation needs to be adjusted for ice content. When ice is present, it is considered to be part of the soil matrix that reduces the pore space, thereby increasing effective saturation and reducing pore size and connectivity, which decreases the saturated conductivity (Soulis and Seglenieks, 2005). One hydrogeological modeling strategy for permafrost is to preferentially adjust horizontal and vertical hydraulic conductivity by a modified form of an impedance factor. Horizontal conductivity can be restricted by reducing the thickness of the transmitting layer, which is directly related to the ice fraction. Vertically, hydraulic conductivity is restricted by reducing the width along connected pathways.

Streams and Lakes

STREAM DISCHARGE

The HRB lies within the Liard Basin of the Mackenzie River drainage basin. The primary gauging station for the Liard Basin (222 000 km^2) is located at Fort Liard, Northwest Territories on the Liard River (station number 10ED001). Data can be acquired through the Water Survey of Canada (2010), Environment Canada. Data are also available through GEWEX's Global Runoff Data Centre (2010). Peak flows occur in June, and low flow is from November to April. The greatest percentage of annual stream flow occurs in the spring months due to snow melt. Streamflow is reduced in the summer due to lower precipitation and higher evapotranspiration.

Figure 19 displays gauging stations in the HRB and the surrounding area. The stations are listed in Table 3. There are no hydrostations capturing the outflow of the Fort Nelson River from the HRB prior to its confluence with the Liard River. There was, however, a hydrostation (station number 10DA001) that captured discharge on the Petitot River briefly from 1992 to 1996 in coordination with the MAGS project.

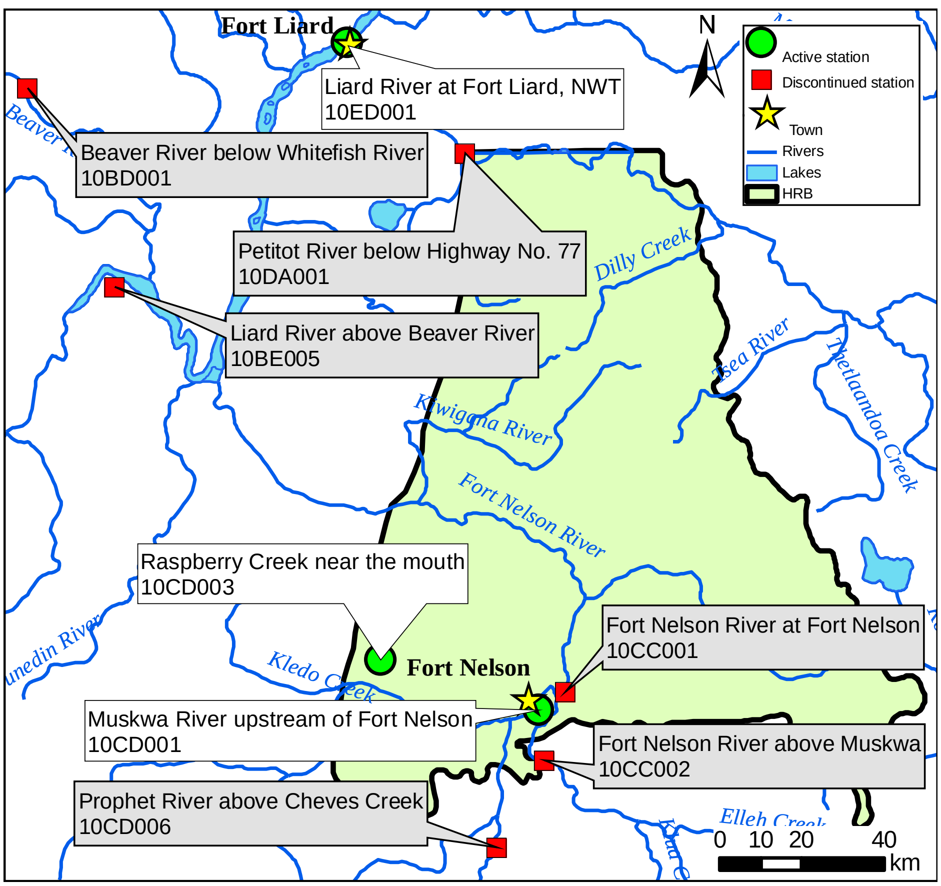

Figure 19. Locations of the streamflow gauging stations near the HRB.

TABLE 3. HYDROSTATIONS NEAR THE HORN RIVER BASIN (ENVIRONMENT CANADA, 2009).

Location	Coordinates	Station	Data available	Volume (km^2)
Liard River above Beaver River	59.70°N, −124.48°E	10BE005	1958–1995	119 000
Fontas River near the mouth	52.27°N, −121.46°E	10CA001	1988–present	7400
Sikanni Chief River near Fort Nelson	57.23°N, −122.69°E	10CB001	1944–2007	2160
Fort Nelson River at Fort Nelson	58.82°N, −122.54°E	10CC001	1960–1978	43 500
Fort Nelson River above Muskwa River	58.67°N, −122.63°E	10CC002	1978–2004	22 800
Muskwa River upstream of Fort Nelson	58.78 °N, −122.65°E	10CD001	1944–present	20 300
Parker Creek near the mouth	58.24°N, −122.80 °E	10CD002	1979–1982	61
Raspberry Creek near the mouth	58.89°N, −123.32°E	10CD003	1979–present	273
Bougie Creek at Km 368 Alaska Highway	58.03°N, −122.72°E	10CD004	1981–2007	332
Prophet River above Cheves Creek	58.48°N, −122.83°E	10CD006	1988–1995	7320
Adsett Creek at Km 386 Alaska Highway	58.11°N, −122.72°E	10CD005	1983–present	109
Petitot River below Hwy 77	60.00°N, −122.96°E	10DA001	1992–1996	22 400
Liard River at Fort Liard	60.24°N, −123.48°E	10ED001	1942–present	222 000
Liard River near the mouth	61.74°N, −121.22°E	10ED002	1974 –present	275 000
Rabbit Creek below Hwy 7	60.46°N, −123.41°E	10ED004	1978–1984	105
Rabbit Creek below Hwy 7	60.46°N, −123.36°E	10ED006	1984–1990	92.7
Liard River at Lindberg Landing	61.74°N, −121.22°E	10ED008	1991–1996	–

LAKES

Most lakes in the HRB are small (<1 km^2) and very shallow, rarely measuring more than 2 m deep. The two largest and deepest lakes in the larger HRB region are Maxhamish Lake (12 m deep) and Kotcho Lake (2 m deep). There are publicly available bathymetric surveys on less than five 'large' lakes in the HRB and its environs. Of those surveyed, three lakes are steep sided and deep, and two are very shallow.

The British Columbia Ministry of Environment's GIS data layer of lakes (EAUBC_LAKES_SP) was compared with the Base Mapping and Geomatic Services Branch of the Integrated Land Management Bureau's layer of lakes (TRIM_EBM_WATERBODIES). The TRIM layer places a minimum size threshold for a lake at 19 m^2 as opposed to EAUBC's threshold of 109 m^2 (Table 4), so TRIM has approximately twice as many lakes (7376 TRIM compared to 3732 EAUBC). The statistically determined average diameter of HRB lakes in EAUBC is 150 m (62.5 m TRIM) and the average area is 24 000 m^2 (2 285 m^2 TRIM). The

TRIM layer's inclusion of the smaller lakes lowers the estimated mean lake size. The volume of standing surface water was estimated assuming an average flat-bottomed lake depth of 1 m and vertical sides. The estimated volume is 90 000 000 m^3 (106 000 000 m^3 TRIM).

Figure 20 shows the size distribution for lakes in the HRB. The red line represents the cumulative total. Note that 65% of all lakes are smaller than 0.5 ha, 80% of lakes are smaller than 1 ha and 95% of lakes are smaller than 4 ha. Figure 19 shows that more than a third of the larger lakes are located in the northernmost part of the basin (north of 59.7°N).

In northeast British Columbia, the maximum lake ice thickness is approximately 1 m. Ice duration averages 200 days (October–April; Rouse et al., 2008). Maximum ice-cover thickness shows differences of only about 10–20 cm for lakes of different depths.

Muskeg

When constructing watershed models for the HRB, it is important to understand that the quantity, location and character of muskeg controls discharge. 'Muskeg' is a traditional Algonquin term for peatland that generally refers to a bog or marsh with thick layers of decaying material. The Canadian Wetland Classification System and special studies from British Columbia (MacKenzie and Moran, 2004) use the ecological wetland classes of bog, fen, swamp and marsh. The differentiation between bogs and fens is important. To quote Quinton and Hayashi (2005), the

> *...contrast between the channel fens and flat bogs suggests that the relative proportion of two these two peatland types should have implications for basin runoff. For example, a basin with a relatively high proportion of flat bogs should generate less runoff than a basin with a lower coverage of flat bogs. [Figure 4] indicates that annual runoff was positively correlated with the percentage cover of channel fens, and negatively correlated with the percentage cover of flat bogs.*

In the Taiga Plains ecoregion of the province, bogs predominate, though fens and swamps occur along the sluggish streams that drain the region. Wetlands have developed in depressions left in thick till by receding glaciers (Vitt et al., 2000). Figure 21 shows wetland distribution throughout the HRB. In detail, ponds tend to be ringed by peat deposits known as 'pond-peatland complexes' (Devito et al., 2005). According to MacKenzie and Moran (2004), a bog is a nutrient-poor, sphagnum moss–dominated peatland ecosystem in which the rooting zone is isolated from mineral-enriched groundwater, soils are acidic and few minerotrophic

TABLE 4. LAKE SIZES FOR THE HORN RIVER BASIN.

EAUBC Lakes

	Average	Minimum	Maximum
Diameter (m)	149	15	2 435
Area (m^2)	24 000	109	1 740 000
Volume (m^3)	24 000	109	1 740 000

TRIM Lakes

	Average	Minimum	Maximum
Diameter (m)	393	30	9 293
Area (m^2)	14 400	19	1 750 000
Volume (m^3)	14 400	19	1 750 000

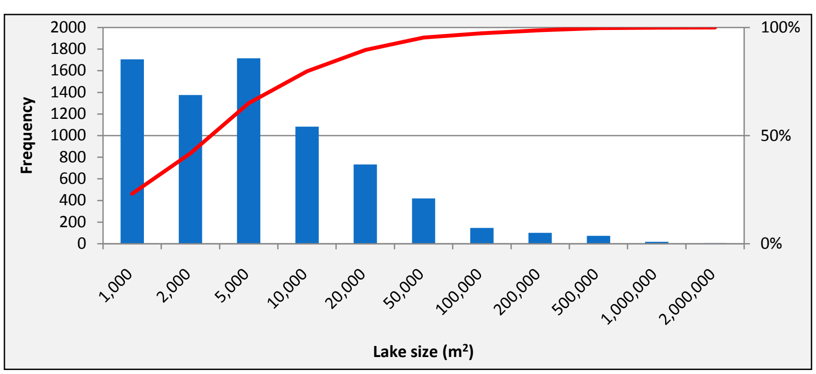

Figure 20. Histogram of lake sizes in the HRB from TRIM data.

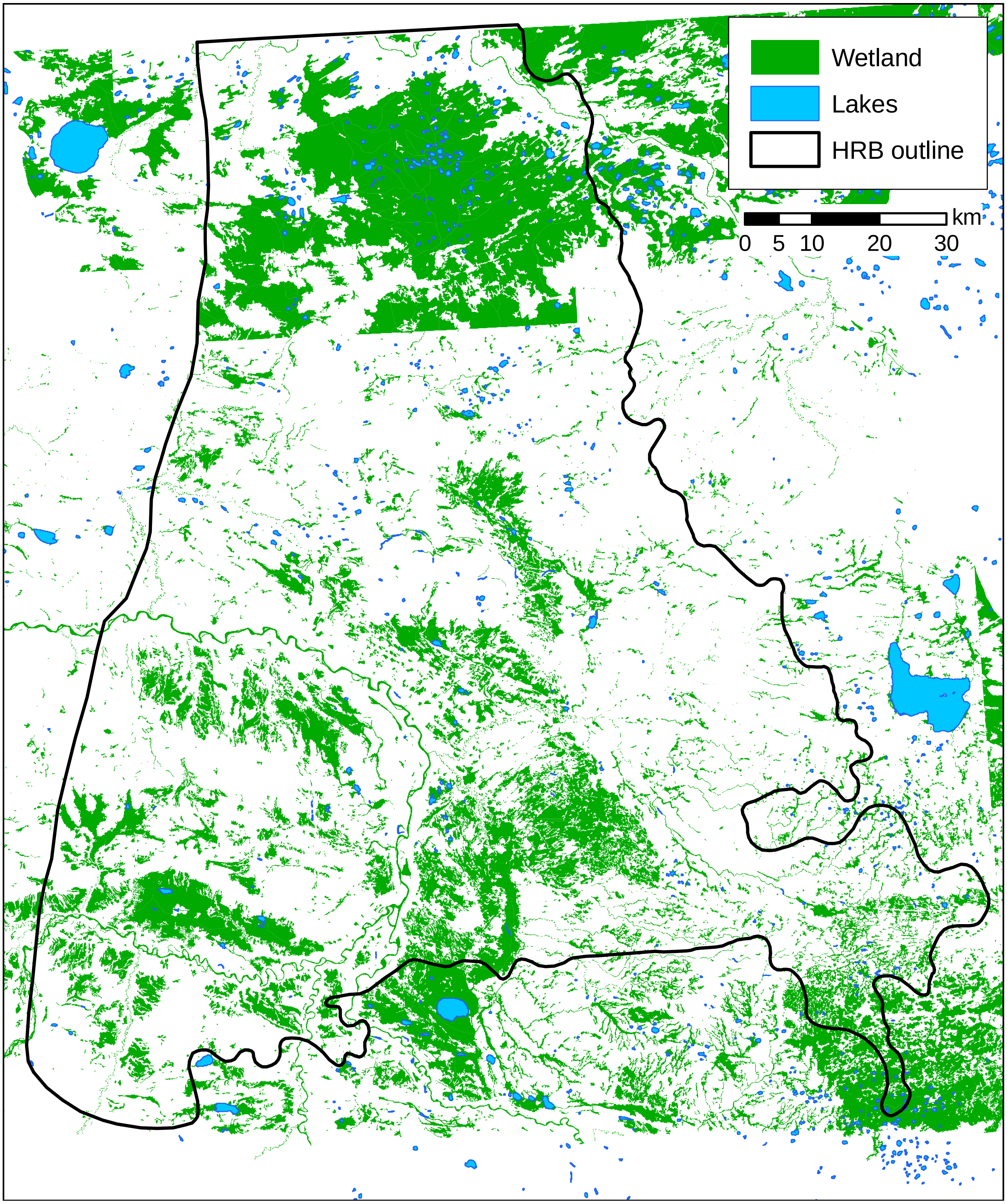

Figure 21. Wetlands and lakes (larger than 4 ha) in the HRB.

plant species occur. A fen is a nutrient-medium peatland ecosystem dominated by sedges and brown mosses, where mineral-bearing groundwater is within the rooting zone and minerotrophic plant species are common (Figure 22).

Figure 22. Bog and fen in the HRB. Photo by Elizabeth Johnson

The region southeast of Fort Simpson, Northwest Territories, is characterized by a mosaic of sphagnum moss and black spruce bogs underlain by permafrost and wet fens without permafrost (Pomeroy, 1985). There is considerable potential for development of thermokarst in the bog terrain (Pomeroy, 1985). The seven categories of muskeg in the lower Liard River valley (Pomeroy, 1985) are shown on the triangle diagram of Figure 23. Vegetation types in the HRB are shown in Figure 24.

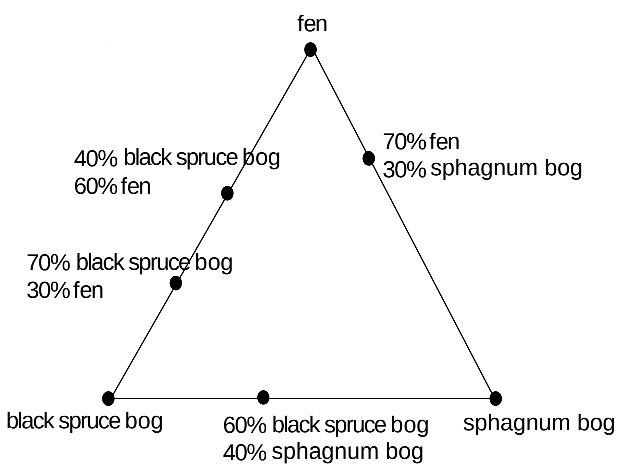

Figure 23. Triangle diagram showing seven categories of muskeg in the lower Liard River valley.

EVAPOTRANSPIRATION

Evaporation is one of the most important factors in developing a water budget. Potential evapotranspiration is often greater than precipitation in summer months (Petrone et al., 2008). Open-water evaporation accounts for 5–60% of total evapotranspiration, depending on latitude and geography (Gibson and Edwards, 2002). Calculated catchment-weighted evaporation losses typically range from approximately 10–15% in tundra areas draining into the Arctic Ocean to as high as 60% in forested subarctic areas draining into the Mackenzie River (Gibson and Edwards, 2002).

Lakes have the highest evaporation rates of any land-cover type. Evaporation for medium and large lakes is significantly greater than for wetlands and small lakes (Table 5). Small lakes have a longer ice-covered period (six to seven months) than large lakes (four to seven months; Rouse et al., 2008). Shallow lakes warm quickly in spring and have very high evaporation rates. Evaporation during the open-water period is an important water-loss component for a small lake of 4 ha and ranges from 70 to 100% of annual precipitation (Gibson et al., 1996). In peatland, water losses through evapotranspiration are far more directly related to evaporation than transpiration. Fen areas can be expected to have evapotranspiration rates that are 10–20% lower than adjacent upland areas during a growing season. This difference may only be about 0.2 mm/day, a total of 30 mm in 150 days. For large drainage basins on the order of 10 000 km^2, this difference represents 3×108 m^3 water annually that is lost to the atmosphere (Barker et al., 2009).

Vegetation type is a large controlling feature in evapotranspiration. Sphagnum moss is prevalent in bogs and widespread in fens. It can hold large amounts of water in its cells and the surrounding area. Evaporation from sphagnum mosses is well below potential evaporation (Campbell and Williamson, 1997) compared to relatively efficient latent heat transfer by (vascular) sedges (Lafleur et al., 1997). Sphagnum can reduce evaporation by changing shape to increase water retention and changing colour to increase the albedo effect. Evaporation from fens is 20–25% greater than sphagnum bogs with dwarf shrubs and 3–10% greater than whole raised bogs.

For modern water-modeling applications, it is important to be aware that evaporative water loss and water-table drawdown can cause groundwater reversals in peatland (Devito et al., 1997). The use of lumped-parameter models for evapotranspiration can result in substantial errors when calculating long-term values for evaporation, particularly for strongly seasonal climates where errors may be as high as 50% for low-throughflow, high-evaporation lakes.

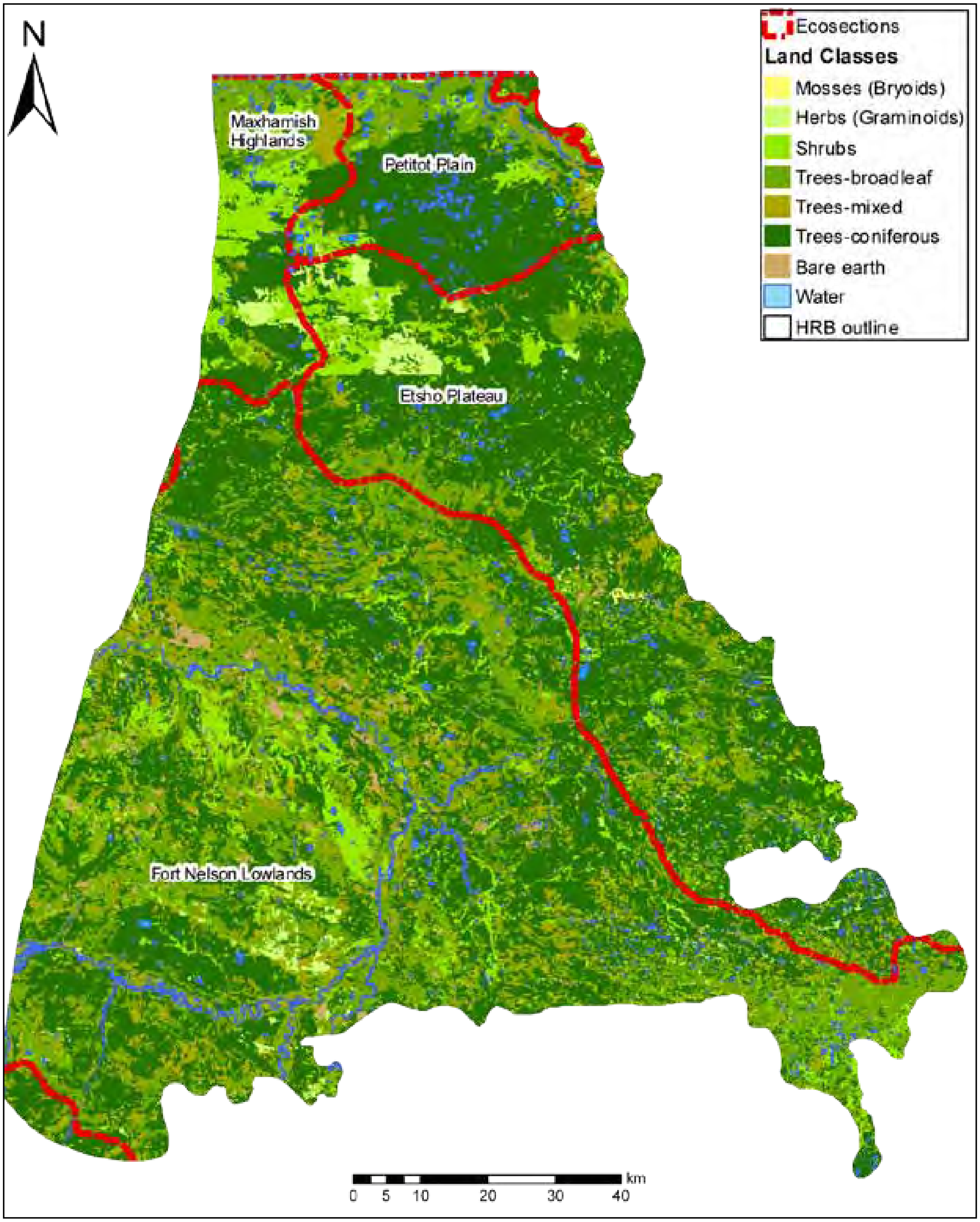

Figure 24. Vegetation map for the HRB.

TABLE 5. EVAPORATION FOR DIFFERENT-SIZED LAKES IN THE MACKENZIE BASIN COMPARED TO UPLANDS AND WETLANDS (AFTER ROUSE ET AL., 2008).

	Evaporation (mm)	Days of open water
Upland	227	
Wetland	314	
Small lake	346	154
Medium lake	406	170
Large lake	422	228

TOPOGRAPHY

Topographic relief is important to modeling drainage in discontinuous permafrost areas. Peatlands occur extensively in the headwaters of many streams and rivers, but those headwaters can be in mountainous regions or low-lying plains. High-relief catchments efficiently drain water while lowland catchments store water (McEachern et al., 2002). Most of the HRB is low relief with less than 5% gradient. The edge of the Etsho Plateau has higher relief, up to 10%. Slope analysis shows that only a few streams appear to exceed that gradient in 094O/08 (A, F, K and L) and 094P/05E.

In high-relief catchments, snowmelt comprises half of the annual discharge. Initially, discharge is predominantly groundwater for the first week melting, but by the second week, discharge comprises mainly precipitation. Overall, discharge is low after snowmelt until mid-summer because lowering of the permafrost layer allows the creation of substantial soil volume for snowmelt infiltration (McEachern et al., 2002).

In lowlands, the depth to the water table governs streamflow response to moisture input and evapotranspiration losses. Peatland operates as a single source area with rapid response for spring runoff when the water table exceeds the depression storage capacity of wetland pools. Bogs are not generally capable of storing all of the annual precipitation because most precipitation occurs in the spring when there is a considerable surplus over storage capacity (Goode, 1977). Following snowmelt, streams in the lowlands follow predictions for saturated catchments with rapid response in streamflow for even small precipitation events. As water levels decline in lowland catchments, stream discharge becomes increasingly dominated by organic sources. Contributions to stream discharge from groundwater tends to remain relatively constant but the majority of runoff is generated from surficial water stored in peatland (McEachern et al., 2002). Relatively slow stream responses occur when pools become disconnected into separate microcatchments during drier periods (Goode, 1977; Quinton and Roulet, 1998).

HYDRAULIC CONDUCTIVITY

Peat has a strong ability to retain water by shrinking (compressing) to reduce pore size and increase water retention by matrix forces. The associated changes in internal pore structure can alter hydraulic properties such as bulk density and hydraulic conductivity (Price and Waddington, 2000).

Porosity is so great in peatland that precipitation will cause an immediate change in the water table. Larger-diameter soil pores of the living vegetation and lightly decomposed peat near the surface of the peat offer much less resistance to water motion than the finer-grained peat deeper in the profile (Quinton et al., 2000). Sphagnum-covered peat has macroporosity down to 20 cm depth. The macropores have been found to transport greater than 50% of the flux in a fen (Martini, 2006). In an overall context, however, evapotranspiration and groundwater–surface water interactions are more important to the water balance than changes in peat volume (Petrone et al., 2008).

Subsurface shallow flow is mainly horizontal, while flow in the lower anaerobic layer provides a negligible contribution to streamflow. The hydraulic conductivity close to the surface is often thousands of times greater than at the base of the peat layer. Hydraulic conductivity in the lowermost inert horizons of bogs is lower than that of glacial tills. There is a very marked decrease in lateral subsurface flow as the water table falls toward the base of the active layer. Although interflow can occur in substantial amounts within the active layer, water is unable to move downwards through the relatively large impermeable layers (Goode, 1977).

Subsurface drainage is strongly affected by the position and thickness of the saturated zone within the peat matrix. A first approximation for a model of the flow regime may consider a peat profile with depth-varying resistance properties with respect to subsurface flow (Quinton et al., 2000). Goode (1977) shows that vertical fluctuation is greater within the active layer of ridges than in adjacent pools over the same period of time. It is important to note that knowledge of the ratio of areas occupied by positive and negative relief elements (ridges and pools) together with their storage capacity is essential to the calculation of total runoff.

The water table often does not mirror topography in the boreal plain. Water-table gradients (counterintuitively) slope against topography. Surficial landforms influence the scale of groundwater interactions, water-table configuration and the distribution of discharge and recharge locations (Devito et al., 2005). The blanket bogs are often recharge zones, while forested wetlands are often discharge zones. The amounts of wetland recharge and discharge can be very small. A study in Alaska found that recharge from wetlands to viable aquifers was less than 1% of the total annual recharge to the aquifer system. The amount of groundwater discharge to streams from wetlands was too small to measure (Siegel, 1989). Although groundwater plays a relatively minor part in the water balance, the mechanisms by which the peatlands retain water and exchange groundwater with adjacent ponds or uplands are important to their maintenance and the quantity of water in the ponds and hill slope soils (Petrone et al., 2008).

Groundwater participates in many peatland water budgets, but its role is difficult to quantify. The quantity of groundwater in a wetland can affect vegetation, water chemistry and biogenic gas production. The presence of permafrost (or thick and persistent ground frost) adds to this complexity. Frost hinders peat shrinkage above the water table until late summer. The ground frost also regulates horizontal and vertical moisture exchanges within the peatland and between the peatland and nearby ponds (Petrone et al., 2008).

ANTHROPOGENIC EFFECTS ON PEATLAND

Peatlands are developed and maintained when a positive water balance exists and there is a surplus of peat production over decomposition. In northeast British Columbia, precipitation exceeds evapotranspiration because evapotranspiration is hindered by weak sunlight and shortened growing seasons. There is very little downward loss of water because peat tends to have low vertical permeability and because there are areas of discontinuous impermeable permafrost throughout the HRB. Muskegs in the region will often not include a major outflow of water as they are found on almost completely flat land. The arctic nature of their climate severely limits peat production and the decomposition rate of peat. Little live peat grows each year, so there is that much less that can die off and decompose annually.

Anthropogenic affects on peatland are mostly associated with artificially raising or lowering the water table. When the water table is raised, forests die as they are flooded out, fens develop, CH_4 emissions increase and more CO^2 is sequestered. When the water table drops, CO^2 is released to the atmosphere and stream chemistry is significantly more affected by the acidic, mineral-rich waters of the peatland.

Peatlands are one of the largest terrestrial carbon reservoirs in the world (Whitfield et al., 2009). When the addition of plant material exceeds decomposition, peatlands represent a long-term net transfer site for the removal of carbon from the atmosphere. However, peatlands have the potential to become immense sources of greenhouse gases (Waddington et al., 2009). Plant material does not decompose quickly in waterlogged, airless, acidic conditions in peatlands but when the water table is lowered, significant amounts of plant material are exposed to the air and the rate of decomposition increases dramatically. During hot dry summers when there is a drop in moisture availability, peatlands can become a net source of atmospheric CO^2 as photosynthesis is decreased and respiration loss enhanced (Price and Waddington, 2000). Drainage of peat also cause increases in summer baseflow, suspended sediments, maximum stream temperature, specific conductivity, pH, and NH^{4+}, NO^{3-},Ca^{2+},Mg^{2+} and Na+ stream concentrations (Prevost et al., 1999).

In muskegs containing discontinuous permafrost, such as those in the HRB, road development can cause flooding. Flooding will cause permafrost degradation, development of thermokarst terrain in areas of black spruce bog and a rapid transition to fen conditions (Rennie, 1978). For example, along portions of the Liard Highway, the roadbed has acted as a dike. Surficial flows have ponded against the road increasing the area of fen and a decreasing the area of black spruce (Pomeroy, 1985).

INFORMATION GAPS

This paper has identified the important features for inclusion in any hydrogeological model(s) in the HRB. However, the transition from this conceptual model to a representative numerical model requires more information in the following areas:

1. Muskeg: The identification of wetland and delineation of fens and bogs. Muskeg in the HRB allows very little downward infiltration of water. Fens channel water laterally towards the basin outlet, whereas bogs retain water within the basin. Some basins in the HRB are identified as having more than 25% wetland. The appropriate characterization of wetland will dramatically affect the accuracy of the model.
2. Permafrost: Location and distribution of discontinuous permafrost. In permafrost areas, water flows in a shallow horizon with little storage capacity, while in nonpermafrost areas water may infiltrate to depth and directly affect the water table. Permafrost dictates regions of surface water–groundwater interaction. Better delineation of fens and bogs will aid in identifying potential permafrost locations.
3. Climate: The spatial distribution of evapotranspiration. Evapotranspiration is one of the largest parameters affecting water balance and is poorly understood across most of the HRB. It varies spatially with the distribution of vegetation type (e.g., upland forest, fen, bog), lakes and water-table depth.
4. Stream discharge: Increased monitoring of stream levels and watershed discharge. Greater ground-based knowledge will help calibrate any numerical model(s).

5. Water table: Increased monitoring of water-table levels basin-wide. Greater ground-based knowledge will help calibrate any numerical model(s). It will also identify spatial sensitivity to seasonal climatic variation. Water-table height dictates evapotranspiration and lateral flow. Lowering of the water-table fragments subsurface channels in fens and causes shallow lakes to disappear.
6. Topography: Delineation of lake depth. Understanding lake bathymetry will clarify the volume of water stored in lakes.
7. Lake-wetland interconnection: Identification of lakes that are connected to wetland via subsurface lateral flow and those that are isolated. Lakes are often identified as ready, replenishable water sources. Identifying isolated lakes will focus water withdrawal toward lakes that can better support the demand.
8. Identification of thermokarst lakes: Thermokarst lakes form where permafrost melts and tend to be associated with a transition to fen.

CONCLUSIONS AND APPROACH FORWARD

The purpose of this report was to generate a conceptual model for the HRB. Generating a sound numerical water model is challenging for northeast British Columbia because of the complications introduced by discontinuous permafrost, widespread patchy muskeg, low relief and a lack of ground-based observations. Overcoming these challenges requires applying all resources available.

Information sources available relate to international, government and academic studies on global and Canadian climate and surface hydrological process modeling, the Mackenzie Basin, Canadian boreal peatland function and carbon storage modeling for peatland. An extensive international study on the Mackenzie Basin has generated a good regional water model for the Liard Basin and provided much information on surface and subsurface forcing factors that affect hydrological models. Time series and gridded climate data are available from Canadian research institutes. Gridded data are becoming available at increasingly finer resolutions. Water movement in peatland-pond complexes has been the subject of much recent research because it affects the carbon-storing capability Canadian boreal peatlands.

Climate analysis indicates that precipitation is not uniform across the HRB. Evapotranspiration is a very important component in a water balance, at times exceeding precipitation, but it varies with water table, lake distribution and vegetation type. The depth, thickness and duration of ground frost and permafrost control subsurface flow, infiltration and recharge rates. The distinction between fen and bog is imperative to understanding and predicting stream discharge. Fens indicate interaction with laterally flowing groundwater, whereas bogs represent the storage of water and regions of potential permafrost. Water models should account for groundwater. While the groundwater component in the water balance may be minor, it represents the mechanism by which peatlands retain water, lakes and uplands exchange water, and streamwater quality and quantity is maintained.

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