

**GIS ASSISTED DISTRIBUTED HYDROLOGICAL
MODELLING IN THE HIGH BOREAL FOREST
OF NORTHERN MANITOBA**

By

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A Thesis

Submitted to the Faculty of Graduate Studies

In Partial Fulfillment of the Requirements for the Degree of

MASTER OF SCIENCE

Department of Civil Engineering

University of Manitoba

Winnipeg, Manitoba, Canada

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MARC E. J. ST. LAURENT

**A Thesis/Practicum submitted to the Faculty of Graduate Studies of The University
of Manitoba in partial fulfillment of the requirements of the degree
of
Master of Science**

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ABSTRACT

Located at the downstream reaches of the Nelson River Basin, the province of Manitoba is rich in water resources that support the development of hydroelectric projects. Accurate estimates of water yield are necessary to support hydropower planning studies, dam safety studies, and hydroelectric operations. The remoteness and expanse of the region hinders the implementation of additional streamflow gauges, hence deterministic hydrological modelling offers an alternative. The use of a hydrological model is beneficial in ungauged areas where calibration in one area permits translation of the model to other ungauged areas with similar land cover.

The lower Nelson River is located in the High Boreal Forest of northern Manitoba. The region contains complex hydrological processes that shape the streamflow hydrograph. Some of these processes include discontinuous permafrost, unique evapotranspiration (ET) processes, a sub-arctic climate, a landscape that is littered with numerous lakes, marshes, and swamps that may be hydraulically disconnected from the basin due to its hummocky terrain, and land cover consisting of various types of muskeg, forest, and treed rock. Snowmelt processes dominate the hydrograph resulting in peak annual streamflow during the spring freshet and near base flow conditions during the winter.

The main objective of this research is to apply a macro scale deterministic hydrologic model to the High Boreal Forest region of northern Manitoba that is capable of providing accurate water yield estimates. Model parameters for the dominant types of land cover are derived from the literature to initiate model calibration.

The SLURP hydrologic model is a semi-distributed conceptual macro scale model that is proven to work well under a wide variety of basin conditions. The SLURP model was applied to the Taylor River (899 km²) and the Upper Burntwood River (6859 km²). Geographic Information Systems were utilized to develop the physiographic parameters for each of the basin Aggregation Simulation Areas.

The SLURP model was calibrated and verified on the Taylor River for each year between 1985 and 2000. A relatively small number of parameters were adjusted in order to

optimize the modelling performance. Overall, SLURP performed satisfactorily as illustrated by an average annual Nash Sutcliffe Efficiency (NSE) of 61% and Deviation from Runoff Volume (D_v) of $\pm 12\%$. The same model parameters were applied to the Upper Burntwood River with the exception of elevated ET parameters. Using the same test years in the Taylor River application, the resulting NSE = 54% and $D_v = \pm 14\%$.

A number of model deficiencies were determined during the application of the SLURP model to the test basins. A simplified frozen soils model was tested in the SLURP model but it was abandoned since it resulted in a significant amount of overland runoff during the spring which did not sustain the streamflow hydrograph. The date dependent snowmelt rates were replaced with a single constant snowmelt rate. The premature generation of snowmelt in the simulated hydrographs was addressed by incorporating a snowpack temperature deficit model into SLURP in order to simulate the effects of snow ripening. These two modifications provide the modelling flexibility needed to control the timing of initial snowmelt and the rate of snowmelt. The model performance improved during the spring freshet period for nearly all test years in both basins.

This research has demonstrated that the SLURP model is able to provide reasonable estimates of water yield for the Taylor River and Upper Burntwood River. Application of the model to other basins and additional model refinements would continue to improve the model performance and would develop a better understanding of the hydrological processes of the High Boreal Forest.

ACKNOWLEDGEMENTS

I would like to express my sincere gratitude to my supervisor Dr. Caterina Valeo for providing me with the opportunity to undertake this research with her. Her editorial comments, guidance, and patience were much appreciated. I would also like to thank her for always being involved with and contributing to the research project. Special thanks for her continued support and for remaining enthusiastic and committed to this project throughout its duration.

I would also like to thank Efrem Teklemariam, Bill Girling, and Brian Korbaylo for their guidance and support as project line advocates for Manitoba Hydro. Allowing me the flexibility to maintain part time employment to complete this research is also appreciated.

Thanks to Dr. Ken Snelgrove for the fruitful conversations and for his assistance with the frozen ground modelling routines. I would also like to thank Dr. Geoff Kite for his assistance with the SLURP model. Special thanks to David Tang for his GIS assistance and to Craig Fast for his help with SLURPView. A thank you is also extended to Ruth Epp and Nancy Voth for collecting the many journal papers and organizing the Environment Canada data.

Finally, I must thank my wife Jennifer, for her love, support, and patience during the years of my graduate studies. Without her I could not have accomplished this goal.

Funding for this research was provided through a NSERC-PGS A grant, NSERC CRDPJ Grant No. 223339-98, and a Manitoba Hydro Research and Development grant G147. Meteorological data, hydrometric data, and topographic data were provided by Manitoba Hydro. Forest Resource Inventory of Manitoba was provided by Linnet Geomatics.

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LIST OF NOTATION

A	=	coefficient used in the interception equation
b	=	weighting constant for interpolation for precipitation data
B	=	coefficient used in the interception equation
c	=	kinematic wave speed [m/s]
c_i	=	simulated streamflow on day i [cms]
C	=	dimensionless, bulk coefficient that characterizes the effects on infiltration due to differences between model and natural systems
cc	=	proportion of total area that is crop cover relative to soil area [%]
C₀	=	coefficient used in the Muskingum routing method
C₁	=	coefficient used in the Muskingum routing method
C₂	=	coefficient used in the Muskingum routing method
C₃	=	coefficient used in the Muskingum routing method
C₄	=	coefficient used in the Muskingum routing method
CC	=	canopy capacity per unit of LAI [mm]
CC_{max}	=	maximum canopy capacity [mm]
C_r	=	temperature to snow pack cold content conversion factor [mm/day/°C]
d	=	depth of snow pack [mm]
D	=	dimensionless, relative drying power
D_v	=	deviation from runoff volume [%]
DF	=	maximum difference between streamflow in a one year simulation as a result of parameter adjustments [%]
e_a[*]	=	vapor pressure at saturated air temperature [Pa]
e_a	=	vapor pressure at ambient air temperature [Pa]
E_A	=	actual evapotranspiration (ET) demand [mm]
E_P	=	potential ET demand [mm]
E_S	=	energy limited potential ET demand for soil [mm]
E_W	=	wet environment ET demand [mm]
E_{es}	=	energy limited ET rate for soil used in Spittlehouse's ET method [mm]
E_{ew}	=	energy limited ET rate for water used in Spittlehouse's ET method [mm]

E_{swl}	=	soil water limited ET rate used in Spittlehouse's ET method [mm]
EL	=	elevation.[m]
f_A	=	empirical constant used in Morton's modification of Penman's equation [MJ/m ² /d/kPa]
fc	=	field capacity as a decimal
f_u	=	empirical constant used in Granger's modification of Penman's equation [mm/d/kPa]
F_p	=	dimensionless, penetration constant of air temperature into a snow pack
g	=	dimensionless, relative ET in Granger's ET method
G	=	daily soil heat flux density [Mj/m ² /d]
H	=	average change in elevation to-stream and down-stream [m]
i	=	meteorological gauge that contains a measurement
I	=	canopy interception per unit of LAI [mm]
I_t	=	inflow at time step t; used in the Muskingum routing method
Inf	=	infiltration rate [mm/day]
Inf_{max}	=	infiltration capacity [mm/day]
INF	=	cumulative infiltration [mm]
j	=	day
k₁	=	dimensionless, fast storage tank retention constant
k₂	=	dimensionless, slow storage tank retention constant
K	=	time of travel along the channel reach; used in the Muskingum routing Method
L	=	to stream and down stream distance [m]
LAI	=	leaf area index as a decimal
L_{min}	=	minimum to-stream distance [m]
L_{max}	=	maximum to-stream distance [m]
M	=	daily advection term in Morton's ET method [MJ/m ² /d]
n	=	dimensionless, Manning's roughness coefficient
NDVI	=	dimensionless, normalized difference vegetation index
NSE	=	Nash and Sutcliff efficiency [%]
O_t	=	outflow at time step t; used in the Muskingum routing method

P	=	daily precipitation [mm]
P_{LR}	=	precipitation lapse rate [mm/100m]
PF	=	precipitation factor [decimal]
$\bar{q}$	=	average recorded discharge [cms]
q_i	=	discharge recorded on day i [cms]
q_t	=	discharge per unit width at time step t; used in the Muskingum routing method
q₀	=	reference discharge per unit width; used in the Muskingum-Cunge routing method
Q	=	streamflow [cms]
Q_n	=	daily net radiation [w/m ² /day]
Q_{ss}	=	subsurface flow from the fast storage tank [mm]
r	=	precipitation [mm]
R	=	runoff [mm]
R₁	=	degree-day snowmelt rate [mm/day/°C]
R_{1,Jan}	=	degree-day snowmelt rate for January 1 [mm/day/°C]
R_{1,Jul}	=	degree-day snowmelt rate for July 1 [mm/day/°C]
R₂	=	restricted degree-day snowmelt rate [mm/day/°C]
R_{2,Jan}	=	restricted degree-day snowmelt rate for January 1 [mm/day/°C]
R_{2,Jul}	=	restricted degree-day snowmelt rate for July 1 [mm/day/°C]
RG	=	lateral ground water flow from the slow storage tank [mm]
R_H	=	hydraulic radius [m ² /m]
RI	=	interflow [mm]
R_n	=	daily net radiation flux density [MJ/m ² /d]
R_{nl}	=	long wave radiation flux density [MJ/m ² /d]
RO	=	overland runoff [m ³ /d]
RP	=	percolation from fast storage tank into slow storage tank [mm]
R_s	=	daily down welling solar irradiation [MJ/m ² /d]
s	=	slope of the saturation vapor pressure curve [Pa/K]
sc	=	proportion of total area that is soil cover relative to crop area [%]
ΔS	=	change in storage [mm]

S	=	bed slope; used in the Muskingum routing method
S₁	=	water content of fast storage tank [mm]
S_{1,max}	=	storage capacity of fast storage tank [mm]
S₂	=	water content of slow storage tank [mm]
S_I	=	initial soil saturation [decimal]
S_m	=	snowmelt [mm]
S_O	=	surface saturation [decimal]
S_p	=	snowpack [mm]
$\bar{S}_{\text{STREAM}}$	=	average stream slope [%]
$\bar{S}_{\text{LC}}$	=	average slope of land surface occupied by a land cover [%]
SS	=	total soil water content [mm]
SSC	=	capacity of the slow storage tank [mm]
SS_{max}	=	maximum possible total soil water content [mm]
t	=	infiltration opportunity time [hours]
T	=	daily average air temperature [°C]
T_a	=	three day average air temperature [°C]
T_c	=	critical air temperature below which snow melts [°C]
T_I	=	initial soil temperature [°C]
T_{LR}	=	temperature lapse rate [°C/100m]
T_s	=	average temperate of the snow pack [°C]
T'_s	=	average temperature deficit of the snow pack below 0 °C [°C]
u	=	wind speed [m/s]
V	=	average velocity [m/s]
V_M	=	volume of measured streamflow [m ³]
V_C	=	volume of computed streamflow [m ³]
Vol	=	lake volume [m ³]
wp	=	wilting point as a decimal
W_C	=	cold content of snowpack [mm]
x	=	meteorological gauge with missing records
X	=	relative weight of inflow and outflow in computing channel storage; used in the Muskingum routing method

z_0	=	aerodynamic roughness length for momentum [m]
λ	=	latent heat of evaporation [J/kg]
γ	=	psychrometric constant [Pa/K]
α	=	dimensionless, empirical coefficient the Morton application of the Priestly-Taylor equation
α_a	=	dimensionless, conversion factor for potential evapotranspiration to actual evapotranspiration
α_s	=	dimensionless, empirical coefficient for soil in the Spittlehouse-Black application of the Priestly-Taylor equation
α_{sm}	=	snowmelt conversion factor
α_w	=	dimensionless, empirical coefficient for water in Spittlehouse-Black application of the Priestly-Taylor equation
ρ_s	=	average density of the snow pack [gram/cm ³]
β	=	dimensionless, empirical constant used in Spittlehouse's ET method
Δx	=	channel length [m]; used in the Muskingum routing method
Δt	=	time length [s]; used in the Muskingum routing method
θ	=	coefficient that is function of the cross section shape; used in the Muskingum routing method
μ	=	arithmetic mean
σ	=	standard deviation
v_s	=	snow albedo
v_L	=	land albedo

CHAPTER 1

1 INTRODUCTION

1.1 General

The province of Manitoba is rich in water resources. Its location captures the drainage of several large river basins including the Winnipeg River, Red River, Assiniboine River, Saskatchewan River, and Churchill River. With the construction of the Churchill River Diversion (CRD), nearly all of the flow in these rivers combines to create the large Nelson River. This renewable resource has allowed for Manitoba Hydro's substantial development of hydroelectric power on the Nelson River.

As a resource that is considered "renewable", the Nelson River is continually targeted for further hydroelectric development. Large hydroelectric potential still exists on both the Nelson River and CRD and additional development is planned for the near future. Therefore, estimates of water yield are of primary interest to local water resource developers such as Manitoba Hydro. One area of particular interest that requires accurate water yield estimates is the Upper and Lower Nelson River Basins. The extent of this basin is shown in Figure 1.1. Two main inputs to this basin are regulated by Manitoba Hydro control structures. The Notigi control structure regulates the diverted flow from

the Churchill River basin into the Nelson River via the Rat and Burntwood Rivers. This forms the basis of the Churchill River Diversion. The Jenpeg Generating Station regulates outflow from Lake Winnipeg, Manitoba Hydro's largest reservoir. There is little regulated storage downstream of these two control structures and a drainage area of approximately 85,000 km² (Crippen Acres Wardrop, 1990).

The remoteness of the region in conjunction with economic considerations hinders the implementation of additional gauging stations. Estimates of water yield can be used for real time forecasting of flow inputs for system operations, probable maximum flood derivation for dam safety, modelling of ungauged watersheds, and the establishment of long term flow records for hydropower planning studies. Not surprisingly, accurate deterministic hydrological modelling of runoff rates and water yield could be used to support these functions. Deterministic modelling in northern Manitoba is hindered by the remoteness of the region, the variation that exists in both the sub-arctic climate and the low relief terrain as well as the diverse array of land features including small lakes, boreal forest cover, muskeg, and swamps. Accurate runoff predictions can be provided by capturing as much of the variation in factors influencing the formation of runoff, as possible. This approach has been hampered in the past by the large amounts of data required. However, remote sensing provides an alternative to collecting costly ground data for use in hydrological modeling, and is particularly beneficial since it can be collected in remote areas. The use of a hydrological model is beneficial in ungauged areas where calibration in one area can allow translation of the model to other ungauged areas with similar land cover.

The focus of this research is on the application of the SLURP hydrological model to the High Boreal Forest of northern Manitoba situated in the Nelson River sub-basin. The annual stream flow hydrograph of non-regulated rivers in northern Manitoba has a number of unique and consistent features that must be simulated by the hydrological model. These features are controlled largely by the macro-climate which is characterized by long cold winters followed by short mild summers, typical of a continental type climate. During the winter months, November to April, snow accumulates to create a

temporary storage on the ground surface as well as within the vegetation canopy. During the early months of winter, water stored in wetlands, lakes, and peatlands, and upper soil layers depletes. Once that excess surficial storage runs out, the streamflow reaches a minimum and is sustained by baseflow derived by deep groundwater storage for the remainder of the winter. May brings rapidly warming temperatures that rise above the melting point of snow thus releasing large amounts of water over a short period of time. Infiltration of melt water may be impeded by partially frozen ground which creates overland flow. The spring freshet normally creates a very steep rise in the hydrograph that results in the peak annual streamflow. As the ground begins to thaw, melt water enters the soil to replenish the groundwater storage and river flows recede. Throughout the summer the groundwater, wetland, and peatland storage initially recharged by the melt water is continuously depleted causing the river to subside. Groundwater is occasionally replenished by large summer rain events which may result in temporary rises in streamflow. As the fall months approach the groundwater continues to deplete, snowfall begins to accumulate, and water present in the upper soil layer begins to freeze.

The High Boreal Forest contains a number of features that complicate our understanding and modelling of the hydrological processes in that region. During the 1990's a number of researchers studied the complex hydrological processes of the High Boreal Forest in northern latitudes. Two large scale research programs include the NOPEX program in the boreal forest of Sweden (Phersson & Pettersson, 1997) and the Boreal Ecosystem - Atmosphere Study (BOREAS) conducted in the Boreal Shield ecozone of central Canada (Sellers, 1997). BOREAS program researchers conducted a number of field programs where the components of the energy and water balances were monitored. The first of the complex processes found in the region is the formation of discontinuous permafrost which impedes the infiltration of water during the snowmelt and summer season. The vegetation, shallow snow cover, and low temperatures create deep frost penetration causing frozen ground to be spatially variable throughout a river basin (Anderson and Neuman, 1984; Nyberg et al., 2001). Second, many northern river basins contain vast amounts of muskeg/peatland which is a landform with a very unique structure and hydrology. Muskeg has a hydraulic conductivity that varies substantially with depth

which in turn increases the water retention as the water table drops (Radforth & Brawner, 1977). Thirdly, there are the unique evapotranspiration processes in the region controlled largely by the variable water table depth and the discontinuous permafrost. Fourthly, the landscape is littered with small wetlands and ephemeral surface depressions in the open-canopy black spruce forest. Water storage and its subsequent loss through evaporation is a fundamental component of the basin water balance in northern Manitoba basins (Metcalf & Buttle, 1999). Lastly, the microclimate of the Boreal Forest, that climate located within, immediately above, or adjacent to the forest, is controlled by the forest itself by interacting with energy from the sun, wind, and precipitation. Forest influences on the process of energy and precipitation distribution affect rates of transpiration, evaporation, soil freezing, and patterns of snow accumulation and melt (Hetherington, 1987).

Although the NOPEX and BOREAS programs collected vast amounts of hydrological data, literature shows that there are relatively few basin scale model applications in northern boreal forests, specifically in northern Manitoba. Basin scale applications include the application of the SSARR model to the Lower Nelson River and Upper Burntwood River (Crippen Acres Wardrop, 1990); the SLURP model to the Sapochi River Basin (Susilo, 2000); the WATFLOOD model to the Sapochi River Basin (Stadnyk, 2001) and the Churchill River Diversion (Whidden, 1999). In order to obtain accurate estimates of water yield, it is desirable to use a hydrological model that simulates the important hydrological processes relevant to the area of interest. This allows for the application of the model to a wide range of meteorological conditions and terrain and land cover types thus producing results that a water resource engineer can place confidence in.

Due to the remoteness and climate of northern Manitoba, and the large watersheds involved in this study, a hydrological model that utilizes Geographic Information Systems (GIS), remote sensing, incorporates appropriate snowmelt algorithms, and functions with only the most readily available spatial data, is advantageous. One model that may be suitable for deterministic hydrological modelling in Northern Manitoba is the

Semi-distributed Land Use based Runoff Processes model (SLURP) (Kite, 1997). SLURP is a semi-distributed conceptual hydrological model that was initially established for modeling meso-scale Canadian watersheds as an alternative to the use of larger and more complicated hydrological models used in the past (Kite, 1978). The primary advantage of this model is that it is well adapted to using both remotely sensed data and ground truth data. A second advantage of SLURP is its conceptual representation of watersheds as it models each land cover separately within each sub-watershed, thus allowing for the same parameter set to be used in the modelling of neighboring ungauged watersheds. SLURP is also able to simulate the behavior of a watershed continuously and avoids the data and calculation excesses of other distributed models.

1.2 Research Objectives

The primary objective of this thesis is to apply the SLURP macro scale deterministic watershed model to the High Boreal Forest region of northern Manitoba in order to produce water yield estimates. Based on the hydrology of the region of interest it will be shown that the SLURP hydrologic model is an effective model for this region. The following three objectives are formulated in order to support this primary objective:

1. Review the hydrological processes of the High Boreal Forest of northern Manitoba in order to support objectives 2 and 3. There has been a great deal of boreal forest research during several large field studies over the last decade, specifically the BOREAS and NOPEX studies. These research programs covered many aspects of the water, carbon, and energy fluxes. The research that is relevant to this thesis will be reviewed and summarized in order to effectively apply the SLURP model.
2. Calibrate the SLURP hydrological model for the dominant land cover types in northern Manitoba, specifically, the Upper Burntwood River, Rat River, Grass River, and Odei River Regions near Thompson, Manitoba. This will be done in order to develop a database of model parameters that can be used for

hydrological modelling throughout that region. These objectives will be accomplished by drawing knowledge and techniques from a wide variety of disciplines including Civil Engineering, Geography, Meteorology and Physics. Based on the availability of data, basin scale, and location, the SLURP version 11.2 hydrologic model will be applied to the Upper Burntwood River and Taylor River Basins in northern Manitoba. A database for these two basins will be constructed which will include meteorological and physiographic information derived using Geographic Information Systems.

3. Due to the complex hydrological processes found in northern Manitoba, it is foreseeable that no current hydrological model will simulate all of the most important processes influencing water yield. Assessment of the SLURP model performance will facilitate the identification of modelling deficiencies. The model will be enhanced in order to improve the models performance and to increase the physical basis of the model.

The results of this research will increase the accuracy of peak runoff predictions; provide information for flood forecasting purposes; reservoir scheduling and optimization of operations; support dam safety studies; and aid in the construction of long term flow databases which are important for hydropower planning studies.

1.3 Thesis Organization

Chapter 2 presents general background information relevant to this research. The reader is introduced to the hydrologic aspects of the High Boreal Forest. An overview of hydrologic modelling is provided followed by a model comparison. The SLURP hydrologic model premise and underlying equations are presented and model applications are summarized. Finally, the results of a SLURP model sensitivity analysis are provided.

Chapter 3 outlines the boundaries of the study and introduces the Upper Burntwood River and Taylor River. The development of a SLURP model database of the two study basins

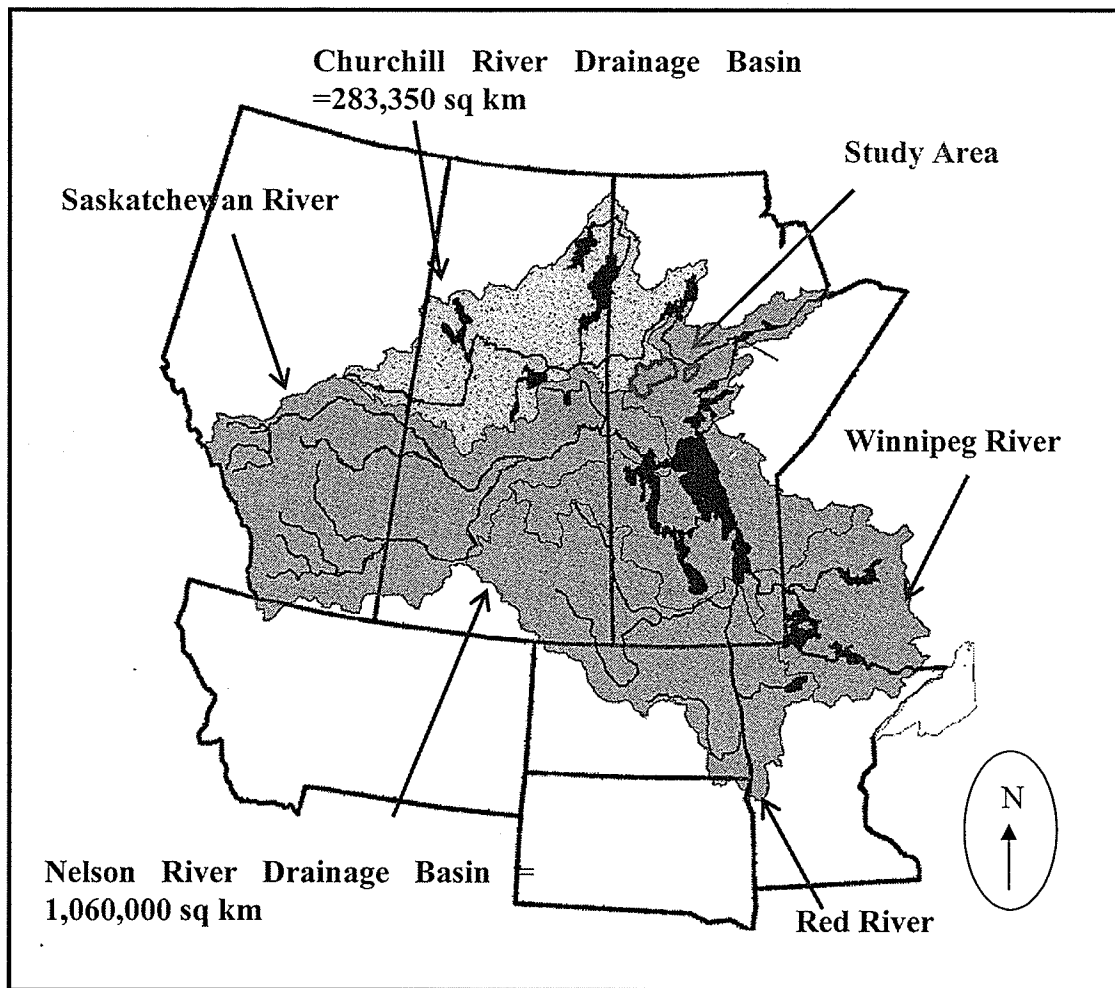
is described. A full description of the data sources and the methods used to prepare physiographic and time series data inputs using tools such as Geographic Information Systems is provided. A database of preliminary model parameter estimates based on values in literature is presented.

Chapter 4 provides details regarding the calibration and validation of the SLURP hydrologic model. The performance of the model is discussed and a final list of model parameters is presented. Model deficiencies are identified as a result of the model application.

Chapter 5 describes modifications made to the SLURP model in order to address some of the model deficiencies. The revised SLURP model is reapplied to the study basins and the results are evaluated and discussed.

Chapter 6 provides conclusions that were drawn from the research and the recommendations for future study.

Figure 1.1 – The Nelson River Basin



CHAPTER 2

2 LITERATURE REVIEW

2.1 Introduction

This chapter provides a summary of the hydrological processes of the High Boreal Forest in northern Manitoba that are presented in the literature in order to support the application of a hydrological model. It provides a general overview of hydrological models and compares the SLURP model to other hydrological models. This chapter also describes the SLURP hydrologic model premise and underlying equations, and summarizes the SLURP applications in the literature. Finally, the results of a model sensitivity analysis are presented.

2.2 Hydrology of the High Boreal Forest of Manitoba

The Nelson River basin and more specifically, the Upper Burntwood River and Taylor River sub-basins (Figure 2.1) are located entirely within the Boreal Forest biome. The Boreal Forest, also known as Taiga is located in the northern latitudes of North America and Eurasia (McKnight, 1993). There are numerous classifications of Canada's landscape and corresponding subcategories of the Boreal Forest. The region is classified as Boreal Forest according to McKnight, (1993); Subhumid High Boreal Forest according

to Welsted et al. (1996); Boreal Predominantly Coniferous Forested according to Metcalf & Buttle (1999), Rennie (1977), Heatherington (1987); Boreal Shield according to the Canadian Council of Forest Ministers (1997); and High Boreal Forest according to Metcalf & Buttle (2001). In the remainder of this thesis, “the study area” will comprise the region containing the Upper Burntwood River and Taylor River sub-basins.

2.2.1 Climate

The province of Manitoba can be divided into several climatic zones. According to the modified Köppen system, the macro-climate of the Nelson River Basin is classified as Humid Continental / Subarctic as the basin lies on the fringe of these two climatic zones (McKnight, 1993). The climate is characterized by brief warm summers followed by cold long winters (Buttle, et. al, 2000). It has moderate precipitation, with the majority falling in the summer. The weather is dominated by westerly winds and storms, especially in the winter (McKnight, 1993). There are no topographic features that have a significant effect on the macro-climate of the region. The major control on northern Manitoba’s climate is its continentality resulting in relatively dry mid-latitude westerly winds that bring the majority of the moisture that arrives in Manitoba. The westerly winds originate in the Pacific and lose most of their moisture in the Rocky Mountains resulting in relatively dry air masses arriving in Manitoba. This results in a relatively dry regime in Manitoba. Polar Easterly winds originating in the Arctic converge with the Westerly winds and often bring with it very cold air masses into Northern Manitoba (Welsted et al.,1996).

Northern Manitoba’s thermal regime exhibits a remarkable amount of day to day variation (Welsted et al.,1996). This is typical of a continental / subarctic climate which exhibits very low temperatures in the winter and, at times, very high temperatures in the summer. Unless otherwise stated, the data reported in this section relate to the period of 1961-1990, the most recent 30-year period for which Climatic Normals have been published. Data is reported at the Thompson Airport meteorological gauge (55°48’ N 97°52’ W) because it is located near the center of the lower Nelson River Basin. The

coolest month of the year, January, experiences an average daily temperature of -25°C , daily maximum of -19.5°C and daily minimum of -30.6°C . The warmest month of the year, July, experiences an average daily temperature of 15.7°C , maximum daily temperature of 22.6°C , and daily minimum temperature of 8.8°C (Environment Canada, 1993).

Precipitation normals show that the wettest month of the year, July, experiences an average total monthly rainfall of 84.3 mm and no snowfall. The driest month of the year, February, experiences 13.9 mm of total precipitation on average, typically all in the form of snow. On an annual basis, average total precipitation is 535.6 mm, total snowfall is 200.9 cm, and total rainfall is 351.6 mm. Of the total annual precipitation 33% is in the form of snow. Roughly 44% of the annual precipitation occurs during June, July, and August. The average snow water equivalent is approximately 200 mm. On average, the month of November receives the most snowfall relative to other months (Environment Canada, 1993).

2.2.2 Geology, Soils and Vegetation

The Nelson River basin has little topographic relief with ground elevations ranging from 239 m – 367 m in the Upper Burntwood River basin and 187 m – 281 m in the Taylor River basin, 50 m at Conawapa, and 0 m at the Hudson's Bay. The terrain is flat and broadly rolling in low regions and hilly in the high areas. The topography reflects the glacially smoothed Pre-Cambrian bedrock surface that has been only minimally modified by glacial drift (NASA, 2002). Terrain can also be described as hummocky with bogs and fens occurring in the low lying areas (Metcalf & Buttle, 1999).

Most of the soils in the Nelson River Basin region are predominantly derived from Glacial Lake Agassiz sediments and consist of clays, organics and some sandy deposits. In low areas, the surficial deposits are mostly varved clays. The deposits are thin or absent in the higher areas. The soil depths vary from bare rock outcrops on the top of hills to deep sedimentation basins in the low areas up to 17 m thick (NASA, 2002).

According to the Manitoba-Canada Soil Survey Map, the Taylor River is contained within a single soil classification polygon comprised of Gray Luvisolic soils on clay with a lacustrine mode of deposition, hummocky terrain, and 10-15° slope. The area surrounding the Burntwood Lake located within the Upper Burntwood River is the same as those soils found in the Taylor River. The head waters of the Upper Burntwood River occur in Dystric Brunisolic sandy soils with a morainal mode of deposition, hummocky terrain, and 16-30° slope. Mid reaches occur in hard rock outcrops, hummocky terrain, and a 16-30° slope. Lower reaches occur in wooded organic mesisols and comprise mainly of flat bogs with little slope (Agriculture Canada, 1986).

Metcalf & Buttle, (1999) characterized the soils in the BOREAS Northern Study Area (NSA), located in close proximity to the Taylor River. They found the wetlands to contain typic and humic mesisols, while the hillslopes are mantled by orthic grey luvisols and orthic eutric brunisols. They found that a typical soil profile consisted of an organic peat layer (mean porosity of 0.813 ± 0.087 standard deviations), overlying a less permeable glacio-lacustrine silty clay (mean porosity of 0.384 ± 0.077 standard deviations). On steeper slopes, these two layers were sometimes separated by a botryoidal layer (mean porosity of 0.536 ± 0.052 s.d.) caused by flocculation during freeze thaw cycles. A simplified profile consisting of a peat layer and a silty clay layer with mean porosity of 0.46 was used to characterize the study area. They found typical organic layer thickness in the order of 0.2 m in Black Spruce Stands, 0.08 m in mixed Black Spruce / Jack Pine stands, 0.12 m in mixed stands with Tamarack, 0.3 m in Treed Muskeg, 0.07 m in treed rock, and 2 m in flooded land. The underlying bedrock consists of Pre-Cambrian gneissic granite that has been glacially polished. There are occasional outcrops of bedrock throughout the region (NASA, 2002).

The Boreal Forest is a simple assemblage of trees that are mainly coniferous and a small proportion of deciduous trees (McKnight, 1993). The predominant tree species throughout the Nelson River Basin is Black Spruce. This tree grows in bogs, muskeg, drier lichen covered sites, and rock outcrops and with stands generally older than 100 years (NASA, 2002). The forest is characterized by normally homogeneous stands of

Black Spruce, Jack Pine, White Birch, and Trembling Aspen (McKnight, 1993). Heights vary from stunted Black Spruce in bog areas to stands as tall as 15 m. Forest cover is broken by varying density, moisture conditions and open treed bogs (NASA, 2002). The undergrowth consists of a layer of deciduous shrubs overtop of a thick layer of sphagnum moss and lichens (McKnight, 1993).

A dominant type of land cover in the northern Manitoba High Boreal forest is Muskeg. It comprises upwards of 50% or more of many northern basins (Radforth, 1977). Muskeg is often referred to as bogland, organic terrain, or peatland. It is always associated with peat – the product of accumulation of organic matter resulting from incomplete decomposition of plants that live and die in the Muskeg ecosystem. In order to form, Muskeg requires a cool and moist climate that allows for peat accumulation where the rate of accumulation is greater than that of decomposition. Radforth and Brawner (1977) present a muskeg classification consisting of nine variations of which four are bog formations and three are fen formations. The surface of bogs are well above the water table and all receiving water is from precipitation. The formation of fens is influenced by ground water as the surface normally coincides with the water table.

2.2.3 Evapotranspiration

Evapotranspiration (ET) includes water that evaporates from the soil surface and water that is transpired from vegetation. ET processes are well understood in southern latitudes but still remains to be poorly understood in northern latitudes. Overall, transpiration rates are controlled largely by: radiant energy, temperature, humidity, soil moisture, land cover physiography, and topography (Buttle, et al., 2000; Hetherington, 1987). The net absorbed energy used for ET goes into transpiration of soil water and evaporation from wet vegetation. The remaining energy goes into evaporation from the ground and water (Hetherington, 1987).

2.2.3.1 Soil Moisture Controls on Evapotranspiration

During the 1995 to 1998 BOREAS study in the NSA, researchers found that there was a relatively small ET in coniferous trees during the growing season. They found that the root zone for conifers is typically less than 0.4 metres in thickness and underlain by a low-permeability substrate. Hence, the upper soil storage fills quickly and additional water inputs run off rather than be used for subsequent abstraction by ET. They also found that the ET rates in the Boreal Forest during the spring melt were overestimated by Global Circulation Models (GCM) because many of these models ignored frozen ground. It was observed that the ground remains frozen during the snowmelt period and prevents the root system from abstracting water resulting in little transpiration (Buttle, et al., 2000).

Hetherington (1987) found that in a muskeg environment, there is an abrupt decrease in transpiration once the capillary fringe loses contact with the rooting zone. This is confirmed by Phersson and Pettersson (1997) who observed that the ET rate in a Swedish Muskeg environment is dependent on the groundwater level. Lafleur et al. (1997) observed an 'on-off' switching system for the ET from a mixed forest in Western Ontario, where a decline in the soil water content to less than 4.5 cm in the upper soil profile resulted in an abrupt decline in ET.

2.2.3.2 Vegetative Physiographic Controls on Evapotranspiration

Sellers et al. (1997) reported that ET rates in the BOREAS NSA were depressed during the middle of the growing season. This is due to the small stomatal conductance's associated minor photosynthetic rates of the local boreal forest tree species.

Pomeroy et al. (1997a) found that Bowen ratios, the ratio of sensible heat flux to latent heat flux, in the BOREAS Southern Study Area (SSA) near Prince Albert, Saskatchewan, remained constant all summer in cut blocks and increased during the summer for treed areas. This indicates that during wet periods, forests use a greater portion of the received energy for evaporation rather than surface heating causing a warming of their localized

environment in order to maintain transpiration rates. This illustrates the forests ability to control evaporative losses.

Metcalf et al. (1999) observed that open canopy Black Spruce tends to have relatively large evaporation fluxes relative to other types of canopies due to their surface roughness, canopy ventilation, and close proximity of water to the ground surface.

Lafleur et al. (1997) divides the ET period into four phases that coincide with the different seasonal growth stages of trees. This classification also helps to illustrate how the ET rates are controlled by physical changes in the vegetation during the growing season. The first phase is "Spring" and includes those periods of snowmelt and ground thaw where ET is effectively zero. The second period is "Preleaf" and extends from the end of ground thaw to the point in time where leaves begin to develop and little ET occurs. The third period is "Green" and is characterized by large radiation inputs and significant photosynthesis resulting in large ET. The last period is "Senescence" and occurs up to leaf fall. Senescence occurs when leaves age, begin to discolor and dry, and photosynthesis decreases. Hence, ET decreases and reaches a minimum when the leaves fall.

The Boreal Forest in the Nelson River basin is characterized by a thick continuous moss-dominated forest floor (Price et al., 1997). Grelle et al. (1997) found that the evaporation from the forest floor made up a very large component of the total evaporation in the BOREAS NSA, roughly 25%.

2.2.3.3 Topographic Controls on Evapotranspiration

Topographical features can exert controls on ET by either limiting or promoting water storage and detention times. Larger transit distances and smaller slopes increase time to concentration of a hydrograph peak resulting in greater ET than that of a basin with smaller transit times and large basin slopes (Metcalf & Buttle, 1999).

Metcalf and Buttle (1999) completed a water balance for a small basin in the BOREAS NSA and found that the greatest overestimation of changes in subsurface storage occurred in areas with a lot of depression storage. They underrepresented the depression storage in their water balance resulting in a significant underestimation of true evaporation. They conclude that small surface storages, and their contribution to total evaporation, is a key component in boreal forest water balance dynamics.

2.2.3.4 Evapotranspiration Rates In Literature

Overall, BOREAS researchers found that ET rates in the NSA were small relative to ET rates in other areas of Canada. From May to September, ET as a percentage of ET from coniferous stands was 0.87 - 0.88 for mixed wood, 0.74 - 0.81 for regenerating stands, and 0.69 - 0.71 for cleared stands (Buttle et al., 2000). Daily mean ET at the Fen and Old Black Spruce sites were 1.68 and 1.27 mm/day respectively (Metcalf & Buttle, 1999).

As a general guideline, Hetherington states that maximum transpiration rates from a forested stand are 0.3 - 0.5 mm/hour and 3 - 5 mm/day for both conifers and hardwoods. However, ET from wet canopies of up to 0.8 - 1.0 mm/hour and over 5 mm/day have been reported. Forest ET is normally 10 - 32% higher than grasslands in Alberta, 6 - 58% higher than tundra, and 36% higher than grasses and shrubs in Saskatchewan. ET during winter is normally less than 3% of annual total ET (Hetherington, 1987).

Price et al. (1997) found that the maximum evaporation rate from the moss layer in the BOREAS NSA Boreal Forest is typically around 0.5-1.0 mm/day. Evaporation was also measured to be negative on days with significant dewfall.

2.2.3.5 Leaf Area Index

Running et al. (1986) state that "Leaf Area Index (LAI), the area of leaf over a given area of ground, is the single variable of greatest importance for quantifying energy and mass exchange by plant canopies over landscapes". Some hydrological models use LAI to estimate evapotranspiration, interception of rain and snow. Today, the most common

method for estimating LAI on a regional basis is through the use of remotely sensed images such as NOAA AVHRR and Landsat TM (Kite, 1998a; Running et al., 1986). Cover types and the corresponding LAI found in literature are summarized in Table 2.1. The summary is restricted to covers that resemble those found in northern Manitoba.

2.2.4 Interception

Intercepted precipitation is the precipitation that does not reach the ground and remains in the foliage of vegetation. It is an important process in water balance dynamics. Rain interception creates a temporary water storage whereby wet surface area is increased and evaporation enhanced. Snow interception is similar in that it increases the air-snow contact area thus increasing sublimation.

Pomeroy et al. (1998) found that snowfall interception varies greatly among different canopy types in the boreal forest of the BOREAS SSA. They found that Spruce stands intercepted 40% more snow than Pine stands, and 300% more snow than mixed tree stands. They surmised that LAI is the vegetation characteristic that has the largest influence on interception efficiency. For the same forest, Pomeroy and Granger (1997b) found that midwinter interception was 40% of cumulative snowfall in a Pine stand, 30% in a regenerating Pine stand, and 6% in mixed wood stand. Intercepted snow sublimation was measured to be 24% in the Pine stand and 6% in the mixed wood stand. Pomeroy et al. (1998) measured the beginning of winter snow interception at the same locations and found it to be 70% for the Spruce stand, 45% for the Pine stand, and 18% for the Mixed stand.

Price et al. (1997) carried out extensive field studies in the BOREAS NSA to characterize interception processes. They found that the moss layer interception was nearly as much as tree interception rates. Extensive measurements showed that of bulk precipitation, 23% was intercepted by the canopy and 18% was intercepted by the moss layer. They also measured the interception capacity of moss at various sites and found them to be

between 15 mm-47 mm, which confirms that moss plays a significant role in the hydrology of the boreal forest of northern Manitoba.

Price et al. (1997) found that stem flow was insignificant in the Spruce forest of northern Manitoba because the tree branches bough away from the tree and funnel water away from the tree. Drip zones created by the tree structure also explains the large variability of throughfall in this forest. Coefficient of Variation in throughfall is 60% for small events, and 24% for the largest events.

Price et al. (1997) found that during small rain events (<10 mm) interception was large and up to 60% of the bulk precipitation. They also found that interception was small, roughly 15-20% for larger rain events. This demonstrates that there is a definitive capacity for the interception storage. It was estimated that the season long 'whole system' interception was 41% of gross precipitation for the summer of 1994. During the same period, 79% of throughfall inputs moved through the moss layer within a few days of input events, and that approximately 21% was retained by the moss to subsequently evaporate during the quiescent periods.

2.2.5 Infiltration

Infiltration is commonly defined as the process of water entry at the land surface into soil from a source such as rainfall or snowmelt (ASCE, 1996). It is an important component of the boreal forest water balance dynamics because it dictates the amount of rainfall excess that is generated verses the water entering the groundwater system. Infiltration is a complex process that is affected by a number of factors such as soil characteristics, surface characteristics, and natural processes such as rainfall intensity, freezing, and soil moisture.

The near surface portion of the forest soils contain a high degree of macropores, pore spaces that are much larger than those of the soil matrix, created by large root densities

and faunal activity. Preferential flow in the macropores can cause infiltrating water to reach depths much quicker than water traveling through macropores (Buttle et al., 2000).

The hydraulic conductivity of the upper soil layer in a muskeg environment reduces exponentially with depth (Radforth and Brawner, 1977; Nyberg et al., 2001; Metcalf and Buttle, 2001). Metcalfe and Buttle (2001) measured the saturated hydraulic conductivity at the BOREAS NSA in northern Manitoba. The largest K_s was observed in the surfacial sphagnum moss layer, a conductivity of 2×10^{-3} m/s. Below the moss, in the organic layer, K_s was observed to be 1.2×10^{-4} m/s to 1.1×10^{-4} m/s. Even lower is the cryoturbated layers, where K_s was measured to be in the range of 1.4×10^{-5} m/s to 9.7×10^{-5} m/s. No infiltration into silty clay layer was detected. Radforth and Brawner, (1977) also quantified K_s in a peatland environment. They list K_s values of 1×10^{-7} m/s for soils below the active layer in raised bog; 0.75×10^{-7} m/s for highly decomposed peat; and 3.81×10^{-7} m/s for undecomposed moss peats close to the surface of active layer.

Once soil temperatures fall below the freezing point of water the soil moisture freezes to create a frost layer. The frost layer may become semi-impermeable and the degree of impermeability is a function of the frozen water content. The frost layer prevents or reduces the infiltration of melt water and rainwater during the snowmelt period. This may significantly increase water yield during the spring snowmelt period. Pomeroy et al. (1997a) measured infiltration rates at the BOREAS SSA and found that all melt water infiltrated into the forest soils if it was not impeded at surface by ice lenses or other obstructions, and if the average initial soil water/ice content of the first meter of soil was less than $0.14 \text{ m}^3/\text{m}^3$. Zhao and Gray (1997) obtained field evidence that showed a rapid decrease in infiltration with time after the start of infiltration in frozen soils, followed by a quasi-steady state regime in which the infiltration rate decreased slowly in time similar to unfrozen soils.

Specific areas including unique types of muskeg forest and bogs are underlain by continuous permafrost, frost that remains at depth throughout the summer months due to the insulating effect of the thick moss layer found in muskeg forest and the thick peat

layer found in bogs. In these areas, water stored within the upper soil zones cannot percolate to the lower soil zones where it would enter the deep ground water storage system (Radforth & Brawner, 1977).

2.2.6 Mechanisms For Streamflow Generation

Stream flow is generated by a number of different mechanisms at the field scale and work to generate overland flow and subsurface flow. Price et al. (1997) studied water fluxes at the BOREAS NSA where a thick moss layer typically covers the mineral soil layer and found the retention capacity to vary with the moss layer depth. Casual observations as well as experimental data suggested that the soil-water interactions at the moss-mineral soil interface are dominated by lateral flow. This indicates that Hortonian type flow is influential in determining the nature of the system hydrologic response. They concluded that the interstorm drainage from the moss carpet is relatively small and the moss has a retention period of 1-3 days prior to moving down to the soil layer. All water that does not move rapidly through the moss within a few days is held by the moss and moves out of the carpet by vapor loss. Overland flow was commonly generated during the snowmelt season because the maximum infiltration rate is greatly reduced during this period as the result of the still frozen ground (Metcalf & Buttle, 1999; Buttle et al., 2000; Nyberg et al., 2001; Anderson & Neuman, 1984; Price & Hendrie, 1983; Stein & Proulx, 1994). Hence, excess melt water may result in overland flow during the snowmelt period.

Metcalf & Buttle, (1999) found that the error term in their water balance at the BOREAS NSA was largely due to the retention of snowmelt and rainwater in perched, clay lined depressions that were not connected to the surface drainage system. This water was subsequently lost via evaporation. However, due to the lack of additional capacity, adding snowmelt or rainwater to the depressions creates runoff for subsequent stream flow generation (Roulet & Woo, 1986; Metcalf & Buttle, 2001).

In a muskeg environment the hydraulic conductivity decreases exponentially with depth where it often decreases thousands of times from the surface of the peat forming layer to the base of the peat forming layer (Metcalf & Buttle, 1999; Whillans, 1987; Phersson & Pettersson, 1997; Radforth and Brawner, 1977). Consequently, there is a very marked decrease in lateral subsurface flow as the water table falls towards the base of the peat forming layer. Although interflow can occur in substantial amounts within the peat forming layer, water is unable to move downwards through the relatively impermeable inert layers. In these landscapes, runoff always ceases once the water table drops to a certain level (Radforth and Brawner, 1977). Hence, the regulation of river flow is likely to be restricted to periods in the summer when there will be an effective storage capacity. The length of time over which this water is released can be very short because of the rapid drawdown of the water table by evaporation. Muskeg is not effective in storing the bulk of annual precipitation since most of it is lost very rapidly in the spring when there is a considerable surplus over the storage capacity (Radforth and Brawner, 1977). While studying a muskeg dominated watershed in Sweden, Phersson and Pettersson (1997) found a large decrease in the peat conductivity when it dried out resulting in a significant decrease in forest floor evaporation when the water table reached low levels.

It is also demonstrated that subsurface stream flow generation in a boreal forest environment is strongly influenced by overburden depth. Devito et al. (1996) found that forested slopes with deep overburden continued to supply groundwater to neighboring wetlands throughout the year. In comparison, they found that slopes with a shallow overburden drained quickly following spring snowmelt causing the baseflow to cease for some streams.

2.2.7 Radiation and Energy Balances

The sun's energy is the primary source of energy for microclimatic processes in the boreal forest environment. Solar energy (short-wave radiation) received by the forest canopy is partly absorbed. In forests with closed canopies, most of the energy absorbed is retained within the canopy and used by the forest to warm their environment for

enhanced ET. Net radiant energy absorbed by the ground beneath the canopy was found to be only 5% of that at the top of the canopy in one study. When moisture supply is adequate, more energy goes into ET. When dry, more energy is used to warm the environment. The amount of energy stored in the vegetation and the soil for photosynthesis usually doesn't exceed 5-10% of net radiant energy on a daily basis (Heatherington, 1987).

In addition to absorption, the forest canopy reflects some short-wave radiation. Heatherington (1987) determined that the amount of reflection during the summer ranges between 7%-16%, with coniferous trees reflecting less than deciduous trees. Granger and Pomeroy (1997) measured summer forest albedo at the BOREAS SSA and ranged from a low of 0.091 for mature jack pine to a high of 0.152 for clear cut land. Phersson and Pettersson (1997) studied energy balances in a muskeg environment in Sweden during the NOPEX studies and found the summer albedo to increase from 0.12 at the beginning of the growing season to 0.22 when the muskeg peats had almost dried out. During the winter months, the snow also reflects a large portion of the sun's solar radiation. The snow albedo diminishes with time as the snow metamorphosis's and accumulates deposits of impurities such as dust and vegetative matter (Leavesley, 1990). Metcalf and Buttle (1999) measured mean areal snow cover albedo at the BOREAS NSA, and found it to equal 0.37.

Of the total incoming solar energy, some of the solar energy passes directly to the ground, the amount ranging from 80% for 10% canopy cover to less than 12% for complete closure of conifers. Forests also receive energy from the atmospheric long-wave radiation, receive energy from sensible heat transfer and condensation of moisture (latent heat), and emits energy (longwave radiation) back to the atmosphere (Heatherington, 1987).

Energy is also used to deplete the snow cover during the winter and spring months. In a southern boreal forest, Pomeroy et al. (1998) observed that sublimation can ablate up to 5 mm of snow water equivalent per day. While studying snowmelt processes at the

BOREAS SSA, Pomeroy and Granger (1997b) concluded that net radiation under the canopy drives most of the melt of forest snow covers and almost all of the melt of the clear-cut snow cover. Correspondingly, Pomeroy and Dion (1996) found melt rates to be slower under forests than in equivalent open areas.

2.2.8 Frost Active Layer

Due to the cold climate of Northern Manitoba, all moisture in the upper soil layers freezes for a portion of each year when soil temperatures fall below 0°C. Permafrost is perennially frost ground at depth and is discontinuous throughout the Nelson River Basin (Peddle & Franklin, 1993). Elsewhere the ground is seasonally frozen, thawing completely at some point after the snowmelt period. Frost development is a complex process and may affect snowmelt runoff characteristics.

Depending on soil moisture conditions, two types of ground frost may form. The first is concrete frost, which forms when very wet soils originating from melt water or an elevated water table, freezes rapidly resulting in impermeable conditions normally at the organic-mineral soil interface (Price & Hendrie, 1983; Stein & Proulx, 1994). The second type of frost, honeycomb or granular frost, forms when relatively dry soils freeze and normally occurs in soils with good aggregation. This frost is loose and friable and maintains or increases infiltration capacities (Price & Hendrie, 1983; Nyberg et al., 2001; Anderson and Neuman, 1984). In a boreal forest basin in Quebec, Stein and Proulx (1994) found that concrete frost was mainly restricted to areas covered with sphagnum and other mosses. According to BOREAS sources, permafrost occurs a few feet below the surface of bogs and occurs at a greater depth under thickly wooded slopes of some clay mantled ridges at the NSA (NASA, 2002).

The frost active layer is that ground that undergoes annual freezing and thawing. In areas with permafrost, the active layer extends from the surface of the ground to the top of the permanently frozen soils. The active layer depth fluctuates in response to changing environmental conditions such as air and surface temperatures, vegetation cover, snow

cover, local moisture regimes, and subsurface properties (Peddle & Franklin, 1993). Metcalf and Buttle (1999) measured active layer depth during the 1995 snowmelt period at the BOREAS NSA. They found mean thaw depth to be greatest in the forested landscape and the smallest in wetlands and open forested areas. They observed that ice lenses in larger wetlands quickly became discontinuous during spring snowmelt and thawed completely by the end of the snow melt period. They also found that the uplands were carpeted by a dry insulating moss layer that maintained permafrost at depth. Metcalf & Buttle (2001) also observed that frost depth was thicker at wetland margins due to the presence of a thinner insulating peat layer.

Perhaps the greatest single environmental condition affecting soil frost depth is snow cover depth (Nyberg et al., 2001). The thermal conductivity of snow is very low thus slowing the exchange of heat from the ground to the air. As the snow depth increases, the insulating effect increases and the resulting frost depth decreases. This is confirmed by Pomeroy et al. (1997a) who observed that the snow depth in cleared forested areas was twice the depth under coniferous trees and resulted in a smaller frost depth and earlier soil thaw in the cleared area due to the snow's insulating effect. During two consecutive years of the NOPEX study, Nyberg et al. (2001) observed frost depths of 0.4-0.8 m under 0.2-0.3 m of snow during the first year and 0.2 m of frost under 0.6-0.8 m of snow at the same location during the second year. The rate of snow accumulation also affects the frost development. In some winters snow accumulates early, insulates the ground, and prevents deep frost penetration. However, in other years where snowfall is late, the ground will freeze to depths of 1-2 metres before significant snow accumulates (Anderson and Neuman, 1984).

According to Brown (1977) muskeg and permafrost are very closely related. This is especially true in the southern fringe of the permafrost region which lies within the Nelson River basin. Freezing and thawing of muskeg (i.e. organic soils) is different than for mineral soils due to vegetation, microrelief, high water table, and thermal properties and dominates the muskeg environment because of the changes in the thermal conductivity of the surface peat throughout the year. During the summer, the surface

layers of peat, except those in proximity to frost, become relatively dry through evaporation. The thermal conductivity of the peat is low hence warming of the underlying soil is impeded. The lower peat layers gradually thaw downward and become wet as the ice layers in the seasonally frozen layer melt. In autumn, surface soil moisture increases due to reduced ET rates. When it freezes, the thermal conductivity of the peat increases considerably. Hence, the peat offers less resistance to the cooling of the underlying soil in winter than to the warming of it in the summer. This also results in lower ground temperatures in peat covered soils relative to adjacent soils that are not covered in peat. Permafrost results when the right conditions under the peat exist such that the temperature remains below freezing throughout the year. According to Brown (1977) the peat thickness varies considerably throughout the boreal forest, however the thickness of the active layer is fairly uniform at 1-2 feet in depth.

Frost effects in the upper soil layers can be related to the blocking effect of ice and redistribution/retention of water. Firstly, there is a dramatic effect if frost substantially reduces the infiltration capacity causing surface runoff above an unsaturated zone. Freezing of infiltrating water can also reduce the soil's hydraulic conductivity. Secondly, cryosuction can redistribute soil water during the winter by drawing it upwards through the soil profile by capillary forces created when water in the largest pores freezes. Thirdly, frost retains water in the soil during the winter and early part of the spring thus preventing it from draining to streams. Once it is released during the spring thaw, frost may become a significant contribution to runoff (Nyberg et al., 2001).

Field studies in the boreal forest suggest that inter-annual variations in snowmelt runoff magnitude are related more to water partitioning at the surface than to differences in pre-melt snow water equivalent (SWE) (Metcalf & Buttle, 2001). This was confirmed by Metcalf & Buttle (2001) at the BOREAS NSA where saturation of muskeg and filling of surface storage prior to freeze up is important. The mean thaw depth resulted in greater infiltration, increased melt water storage on slopes, and reduced downslope transmission of water. Generally, a frost table close to the surface and the presence of ground ice at the peat-silty clay interface ensures rapid down slope movement of snowmelt and rain

water, which will drain to a mosaic of detention storages, ephemeral depression storages, and wetlands of variable size. These surface stores provide the only available storage during the spring melt period when soils are saturated and frozen, hence runoff increases rapidly during spring melt once the surface stores are filled (Roulet & Woo, 1986; Metcalf & Buttle, 2001). For the period immediately following snowmelt and for later spring rainfalls, there is a joint control on stream flow by the active layer development, surface storage, and the magnitude and timing of rainfall inputs (Metcalf & Buttle, 2001).

In a Quebec boreal forest, Stein and Proulx (1994) observed that once all snow melted, the temperature in the organic layer rose above 0°C in 1-9 days after the snow melted compared to 5-25 days in the underlying mineral soils. The delay in the downward progression of the thawing front into the mineral soils resulted because of the ice layer at the organic-mineral soil interface. At the BOREAS NSA, Metcalf and Buttle (2001) found that during the melt period, the absence of percolation below the peat-silty clay interface was attributed to a similar ice layer formation and/or a greater proportion of matrix ice. Infiltration below the frost table occurred once thawing penetrated into the silty-clay layer. Where organic layer thickness was thin, thawing front advanced more rapidly.

2.3 Hydrological Modelling

2.3.1 General Overview of Hydrologic Models

With the advance of computer technology, hydrologic modelling has become an important component in some hydrological studies as demonstrated by Fast (1999). Model selection is an important part of the modelling process due to the sheer number and varying complexities of hydrological models available. The objective is to choose the model and modelling technique that will most closely represent the given system, and provides the required outputs. However, a model should also be chosen that provides the desired level of benefit with the acceptable expenditures associated with model setup and data requirements.

There are five possible sources of variation in hydrologic modelling: time, three space dimensions, and randomness (Maidment, 1993). All abstract models can be classified with respect to these five parameters. With respect to randomness, abstract models are either stochastic or deterministic in nature. Stochastic models account for the uncertainty and random character of hydrologic data by using statistics and frequency analysis to model and quantify random events. Deterministic models do not consider randomness but take into account the physical characteristics of a watershed. A given input will always produce the same output (Chow et al., 1988). This research will use a deterministic model because the output can be reproduced, there are many commercially available deterministic models, and because the complexity of stochastic models is not required.

Hydrological models also vary in the complexity associated with representing the physical processes that occur within a watershed and range from 'Black-Box' to 'Physically-Based' models. Physically based models are developed to realistically represent the hydrological systems and normally use equations based on physics to represent the runoff generating mechanisms that occur with varying magnitude and location within the basin being modelled. Conceptual models use assumptions and mathematical statements to represent generalizations of the perceptual model of the hydrological system (Beven, 1991). Conceptual models are similar to physically based models with the exception that they are more simplified and usually contain fewer parameters. 'Black-Box' models are the simplest form of a hydrological model which are only mathematical propositions that have no physical basis and may only be applicable to the period and conditions of the original data (Kite et al., 1996). Examples are regression type relationships between precipitation or snowpack and seasonal runoff. This study involves simulating a wide range of hydrological conditions, therefore a model that uses both conceptual and physically based equations is required.

The simplification of spatial input data and watershed processes leads to the designation of two main types of deterministic conceptual hydrological models: lumped, and distributed. Models are classified based on how they take into account the spatial

variation of a watershed, as shown in Figure 2.3. Lumped deterministic models simulate streamflow using a single set of parameters that represent basin averaged data, hence ignoring the spatial variation that is inherent in watersheds of increasing size. Distributed models account for at least some degree of spatial variation throughout a watershed. Since physically based models “include the lateral movements of surface and groundwater, they are almost inevitably also distributed models” (Kite et al., 1996). In contrast, Black-Box models are often lumped models due to simplifications in both spatial data and processes.

The final decision depends greatly on the required level of detail, time and money constraints, and, most importantly, data availability. Although fully-distributed models are generally more accurate and realistic than lumped models, if limited data is available for the modelling exercise, no benefits will be gained by using a fully-distributed model, as opposed to a lumped or semi-distributed model. The lumped approach becomes unreasonable for larger watersheds, but the fully distributed models become impractical because of unrealistic data requirements. In these cases it is more practical to use a model that lies between lumped and fully-distributed. These middle-ground models can “incorporate the most important physical laws while retaining simplicity of operation” (Kite et al., 1996).

As stated by Kite et al. (1996), data availability and acquisition are clearly some of the most important challenges in hydrologic modelling:

“The biggest problem facing hydrological modellers is that of obtaining reliable distributed data. The most physically-sophisticated distributed hydrological model will give no better results than the simplest lumped model without accurately distributed input data. Many so-called distributed models are, in fact, simply lumped models run many times over with the same parameters and the same input”.

2.3.2 Model Comparison

Manitoba Hydro selected the SLURP model for this study because of its suitability to the study area and because it provided funding for the development of the model in the past. This section compares the SLURP model to other models to ensure its suitability for this study.

Today there are many different deterministic models available due to the ease with which one can be created with today's computer technology. However, most models were developed for use in a specific geographic location resulting in models that may only account for those hydrological processes that are found in its intended area of application. The model should be able to adequately simulate basins with low relief, muskeg, cold climate, and snowmelt as a major source of runoff. Many models are quickly eliminated as they do not account for some of the dominant processes particular to northern Manitoba. Due to the utility of remotely sensed data in the remote regions of northern Manitoba, there is potential for its use in hydrological studies therefore, the model should have the ability to accommodate this source of data. The model should also have excellent graphical capabilities, and exploit the benefits of a GIS. Some of the continuous distributed models that address some, if not all, of these criteria include: the Canadian models SLURP, WATFLOOD, and UBC; the Swedish model HBV; and the American models HEC-HMS, SRM, SSARR, and NWS.

Most of the models have some characteristic that is desirable for use in northern Manitoba. However the majority of these models have at least one characteristic that makes them inappropriate for the study area. The UBC model (Quick, 1995) is intended for mountainous terrain and is not suitable for analysis of prairie regions or other areas with flat terrain as many of the model routines are a function of altitude. It also models such processes as orographic rainfall and glacier melt, processes that are not present in northern Manitoba. SRM (Rango, 1995) uses remotely sensed data for snow coverage input, but it too is only intended for mountainous regions. The model is entirely Black-Box in nature and subdivides basins based on elevation bands only. HBV (Bergstrom, 1995) was developed for use in Sweden but its suitability has not yet been proven outside

of Scandinavia. It is much like the HBV model in that it also is designed for mountainous terrain and is entirely Black Box in nature. Both SSARR (Speers, 1995) and NWS (Burnash, 1995) are relatively complex conceptual models with numerous parameters. While the models attempt to capture as many factors influencing runoff as possible, the variables used are difficult to measure in the field, and thus, the model's authors recommend using a long historical record to estimate the many parameters involved. This is a daunting task and does not provide user confidence when attempting to translate the model to nearby basins. The SSARR model does not simulate land cover types separately and therefore does not permit studies of land use change. All of the UBC, HBV, SSARR, and NWS models also have poor user interfaces and little or no integration with a GIS. For these reasons, these models would not be suitable for this type of study.

The HEC series of hydrologic modelling programs are probably the most widely used models. Produced by the US Army Corps of Engineers, HEC-1 provides a variety of options for simulating precipitation-runoff processes. HEC-HMS has now superseded HEC-1 by offering a sophisticated graphical user interface. However, this model is Black-Box in nature and does not model snowmelt processes, therefore it was not suitable (US Army Corps of Engineers, 1998).

The WATFLOOD distributed hydrologic model is a physically based model which includes sophisticated snowmelt algorithms (Kouwin, 1998). It has a user-friendly Microsoft Windows format with good graphics capabilities and is designed to use remotely sensed data. However, WATFLOOD consists of a complex modelling system with many parameters that must be determined through calibration.

The SLURP hydrological model is suitable for this study for a number of reasons. The SLURP model has been calibrated on basins world wide and many sizes ranging from 0.47 km^2 - 1.6 million km^2 (Stolte & Van der Kamp, 1999; Kite et al., 1994). These basins contain a wide variety of land cover types and terrain including wetlands, mountainous terrain, prairies, subarctic, and forest. SLURP allows the user to choose

among a variety of evapotranspiration routines. The main advantage of SLURP is that it is proven to work well at many scales and geographic locations and its conceptual representation of the watershed provides sufficient physical basis with relatively few parameters that require calibration. Also, representation of sub-watershed units is flexible and simple to parameterize. SLURP also accommodates remotely sensed data such as snow water equivalent, snow covered area, and cloud cover. The model is also public domain software and includes the source code that can be easily modified for customization.

2.4 SLURP Hydrological Model

The SLURP model was first developed by Dr. Geoff Kite in 1975 while working with the National Hydrology Research Institute, a division of Environment Canada (Kite, 1998a). Today, Dr. Kite continues to improve and support the SLURP model. Recently, Dr. Kite has focused some of his applications of SLURP to semi-arid regions of the world. He has modified SLURP to simulate processes associated with arid regions such as irrigation and associated losses (Kite, 2002). Throughout the course of this study, newer versions of SLURP were released. However, it was deemed that there was no added benefit by using these newer versions, hence only version 11.2 was used for this research.

“SLURP (Semi-distributed Land Use-based Runoff Processes) is a basin model which simulates the hydrological cycle from precipitation to runoff including the effects of reservoirs, dams, regulators, water extractions and irrigation schemes” (Kite, 1998a). It is a semi-distributed conceptual hydrological model (Kite et al., 1996) that was initially established for modelling meso-scale Canadian watersheds as an alternative to the use of larger and more complicated hydrological models used in the past (Kite, 1978). The majority of the models subroutines are conceptual in nature while others are physically based.

2.4.1 Model Premise and Underlying Equations

The SLURP Version 11.2 User Manual written by Geoff Kite (1998a) provides an in depth review of the SLURP model premise and provides details of all underlying equations. The following sections are based on the contents of the SLURP User Manual.

2.4.1.1 Model Structure

The SLURP model divides a basin into hydrologically consistent sub-areas called Aggregation Simulation Areas (ASAs), similar to a sub-watershed. ASAs can be any irregular size or shape since they are based solely on stream network and topography. ASAs are further subdivided into a number of sub-areas, each representing a unique type of land cover. The basic requirements for an ASA are that the land cover distribution is known within the ASA; the elevation for elements within the ASA are known; and that the ASA contributes runoff to a definable stream channel (Kite, 1995).

Each of the land cover sub-areas in each ASA forms an element in a matrix, as illustrated in Figure 2.3. The vertical water balance is then applied to each element of the matrix at a daily time step. The vertical water budget, shown in Figure 2.4, is based upon four reservoirs or tanks. These four tanks are *canopy storage*, *snow storage*, *fast storage*, and *slow storage*, which produce overland runoff, interflow, and groundwater flow. The upper soil layers are modelled using a linear reservoir known as the rapid storage tank. Rainfall and snowmelt enter this tank through infiltration, and exit through percolation to another linear reservoir known as the slow storage tank. Water in the slow storage tank is routed to the nearest stream or leaves the tank through evapotranspiration.

This model structure represents a major advantage for SLURP in that it allows the model to be applied anywhere between Fully-Lumped and Fully-Distributed. In the case of a Fully-Lumped approach, one modelling sub-unit, called an Aggregation Simulation Area (ASA), can be used to represent the watershed. On the other hand, the watershed can be subdivided into many smaller ASAs, the resolution being the choice of the user. As the number of ASAs increases and their corresponding size decreases, a Fully-Distributed

application is approached. Most SLURP applications lie somewhere between Fully-Lumped and Fully-Distributed, as the lack of detailed spatial data normally does not warrant a fully-distributed application.

2.4.1.2 Data Requirements

There are three types of data required by SLURP in order to execute a model simulation. The first group is the physiographic parameters. These parameters are geographic in nature and dictate the spatial structure of the ASAs and the distribution of land covers within the ASAs. The following data are needed for each ASA:

- (a) ASA area;
- (b) fraction of the ASA covered by each land cover;
- (c) mean latitude and altitude of the ASA;
- (d) mean and standard deviation of distance to the nearest stream for each land cover sub-area;
- (e) mean change in elevation to the nearest stream for each land cover sub-area;
- (f) mean and standard deviation of distance within the stream to the ASA outlet for each land cover sub-area;
- (g) mean change in elevation within the stream to the ASA outlet for each land cover sub-area;
- (h) layout of the ASAs within the watershed – i.e. the sequence of the ASAs as they are drained by the stream network.

These parameters are normally derived from a Digital Elevation Model (DEM) and a land cover map. The task of deriving these physiographic parameters is ideally suited for a Geographic Information System (GIS).

The second type of data required by SLURP is time series data. Most of the time series inputs are meteorological in nature. Since the model operates at a daily time step, all of the time series data is required in a daily time step format. The following time series are required for each ASA:

- (a) total daily precipitation [mm];

- (b) daily mean air temperature [$^{\circ}\text{C}$];
- (c) One of:
 - i dewpoint temperature [$^{\circ}\text{C}$];
 - ii relative humidity [decimal]
- (d) One of:
 - i incident global radiation [$\text{w}/\text{m}^2/\text{day}$]
 - ii hours of bright sunshine [hours or percentage of maximum hours]

The next group of time series data is optional for each ASA. If the user does not supply them then these time series are computed internally by SLURP. If the user provides these time series, they replace the internal values:

- (a) cloud cover [percentage of ASA];
 - otherwise SLURP assumes the entire ASA receives precipitation;
- (b) snow covered area [percentage of ASA];
 - otherwise SLURP assumes the entire ASA is covered by snow;
- (c) snow water equivalent [mm]
 - otherwise SLURP distributes a uniform depth of snow throughout the ASA;

The remaining group of time series data is also optional. The user can incorporate them to simulate water additions, abstractions, and lake levels:

- (a) recorded streamflow [m^3/s]
 - (must correspond to the outlet of an ASA; used for model calibration and model assessment)
- (b) abstractions for irrigation [m^3/s];
- (c) water exported from a basin [m^3/s];
- (d) lake evaporation [mm];

The third and last type of data required by SLURP are the model parameters. These parameters are used by the model subroutines in order to simulate the hydrological processes taking place in the watershed. Each of these parameters and how they are

derived is listed in Table 2.2. Each of these parameters will be described in subsequent sections where the modelling equations are presented.

2.4.1.3 Evapotranspiration

At a particular time step, the first operation within the vertical water balance for a particular land cover and ASA is to evaluate the evapotranspiration demand. SLURP provides three alternative methods for computing evapotranspiration (ET).

Each of the three methods uses either hours of bright sunshine or global radiation as inputs for computing ET. If hours of bright sunshine are used, the model computes ASA average daily global radiation. The conversion is based on a complex set of physical equations, their explanation of which would be too lengthy and detailed for the scope of this study. The reader is referred to the SLURP User manual for more information (Kite, 1998a).

The first ET method is Morton's modification of the Complementary Relationship Areal Evaporation Model (CRAE) (Morton, 1983) which computes ET by reducing a modified Penman estimate by an amount proportional to the vapour pressure deficit as stated in the following equation:

$$E_A = 2E_w - E_P \quad (2.1)$$

where E_A is the actual ET [mm], E_P is the potential ET [mm], and E_w is the wet environment ET [mm]. CRAE computes E_P by solving the energy balance and aerodynamic equations at equilibrium temperature using a modification of the Penman equation in which the wind function is replaced with a vapor transfer coefficient as follows:

$$E_p = \frac{sR_n}{\lambda(s + \gamma)} + \frac{\gamma f_A (e_a^* - e_a)}{\lambda(s + \gamma)} \quad (2.2)$$

where s is the slope of the saturation vapor pressure curve [Pa/K], R_n is the daily net radiation flux density [MJ/m²/d], λ is the latent heat of evaporation [J/kg], γ is the psychrometric constant [Pa/K], f_A is an empirical constant [MJ/m²/d/kPa], e_a^* is the vapor pressure at saturated air temperature [Pa], e_a is vapor pressure at ambient air temperature [Pa]. E_w is computed from the slope of the saturation vapor pressure/temperature curve and the net radiation as follows:

$$E_w = \frac{\alpha s (R_n - M)}{\lambda(s + \gamma)} \quad (2.3)$$

where α is an empirical coefficient, and M [MJ/m²/d] is an empirical advection term estimated as:

$$M = (1.37R_{nl} - 0.394R_s) \quad (2.4)$$

where R_s is the daily down welling solar irradiation [MJ/m²/d], R_{nl} is the long wave radiation flux density [MJ/m²/d]. In the SLURP implementation of Morton's method, E_A is set to zero when the surface and soil stores are empty. Otherwise E_A is computed using Equation (2.1). It should also be noted that the Morton estimate of E is not land cover dependent (Barr et al., 1997).

The second ET method is the Granger method (Granger, 1991). It is a variation of the Penman equation:

$$E_p = \frac{sg(R_n - G)}{\lambda(sg + \gamma)} + \frac{\gamma g f_u (e_a^* - e_a)}{sg + \gamma} \quad (2.5)$$

$$G = \frac{1}{0.905 + 0.095e^{6.2D}} + 0.2 * (1 - D) \quad (2.6)$$

$$f_u = 8.19 + 22z_0 + u(1.16 + 8z_0) \quad (2.7)$$

$$D = \frac{f_u(e_a^* - e_a)}{f_u(e_a^* - e_a) + \frac{(R_n - G)}{\lambda}} \quad (2.8)$$

where g is the dimensionless relative ET, f_u is an empirical constant [mm/d/kPa], z_0 is the aerodynamic roughness length for momentum [m], u is the wind speed [m/s], G is the daily soil heat flux density [MJ/m²/d], D is the dimensionless relative drying power.

The third ET method is the Spittlehouse and Black Method (Spittlehouse, 1989). It is different from Morton and Granger in that it adds a soil store limitation to ET. It determines the soil water limited ET rate [mm], E_{swl} , as well as the energy limited ET rates [mm], for water, E_{ew} and for soil, E_{es} . They are computed as a function of plant transpiration using the Priestly Taylor equation for water and soil separately:

$$E_{ew} = \alpha_w \left(\frac{s}{s + \gamma} \right) (R_n - G) \quad (2.9)$$

$$E_{es} = \alpha_s \left(\frac{s}{s + \gamma} \right) (R_n - G) \quad (2.10)$$

$$E_{swl} = \beta \left(\frac{SS - SS_{max} * wp}{SS_{max} (fc - wp)} \right) \quad (2.11)$$

where α_w is the Priestly-Taylor coefficient from water, α_s is the Priestly-Taylor coefficient from soil, β is an empirical constant [mm/day], SS current soil water contents [mm], SS_{max} is the maximum possible soil water contents [mm], fc is the field capacity

$[\text{m}^3/\text{m}^3]$, and $\mathbf{wp}$ is the wilting point $[\text{m}^3/\text{m}^3]$. The energy limited potential ET demand for soil, $\mathbf{E_s}$, is determined from the relative areas of soil and crop cover:

$$\mathbf{E_s} = \mathbf{sc} * \mathbf{E_{es}} + \mathbf{cc} * \mathbf{E_{ew}} \quad (2.12)$$

where $\mathbf{sc}$ and $\mathbf{cc}$ are the relative areas of soil and crop cover respectively [%]. $\mathbf{E_P}$ is taken as the lesser of $\mathbf{E_s}$ or $\mathbf{E_{swL}}$. Finally, $\mathbf{E_A}$ is determined using the linear function:

$$\mathbf{E_A} = \alpha_a \mathbf{E_P} \quad (2.13)$$

where the α_a is a dimensionless ratio that varies according the contents of the soil storage as illustrated in Figure 2.5. The ratio decreases once the contents of the soil storage increases beyond the Field Capacity because of anoxia, or deficiency in aeration. The plants become less efficient at removing water from the soil (Dijkuis & Berliner, 1988).

2.4.1.4 Interception

The model must first begin where all hydrology begins, precipitation. If the air temperature $\mathbf{T}$, is less than the critical temperature $\mathbf{T_c}$, precipitation is classified as snow. If $\mathbf{T} > \mathbf{T_c}$, precipitation is classified as rain. Both snow and rain are intercepted by vegetation, if present. The magnitude of interception is modelled empirically as follows:

$$\mathbf{I} = \mathbf{LAI} * \mathbf{A} * \mathbf{P^B} \quad 2.14$$

where $\mathbf{I}$ is the interception per unit of LAI [mm], $\mathbf{P}$ is the precipitation [mm], $\mathbf{A}$ and $\mathbf{B}$ are coefficients, and $\mathbf{LAI}$ is the leaf area index as a decimal. Intercepted precipitation is added to the canopy capacity until it reaches its maximum capacity, defined by:

$$\mathbf{CC_{max}} = \mathbf{LAI} * \mathbf{CC} \quad 2.15$$

where CC is canopy capacity per unit of LAI [mm]. Once the canopy capacity is reached, any additional precipitation is directed to the snow storage if it is snow and directed towards the fast storage tank if it is rain.

2.4.1.5 Snow Storage

A snow storage tank with unlimited capacity accumulates snow if there is any precipitation during a day in which the mean temperature, T , during the day is below the critical temperature. The critical temperature, T_c , is the temperature that causes snowmelt, S_m , to form and exit the snow storage tank. The snow-pack, S_p , is depleted using the degree-day approach, which states that:

$$S_m = S_p R_1 (T - T_c) \quad 2.16$$

where R_1 is the snow melt rate [mm/day/°C]. Two snowmelt rates, one for January 1 ($R_{1,Jan}$) and one for July 1 ($R_{1,Jul}$), are specified in the SLURP model. The snowmelt rate for any given day is interpolated between the January 1 and July 1 snowmelt rates.

Alternatively, if net radiation, Q_n [w/m²/day] is provided, a Restricted Degree-Day algorithm is used to calculate snowmelt using the following relationship:

$$S_m = R_2 (T - T_c) + Q_n \alpha_{sm} \quad 2.17$$

where R_2 is the restricted degree-day snowmelt rate, and α_{sm} is a conversion factor. The snow-pack is depleted until the snow storage is empty. In both methods, the snow-pack is depleted until it is empty. This method also utilizes two snowmelt rates, one for January 1 ($R_{2,Jan}$) and one for July 1 ($R_{2,Jul}$), are specified in the SLURP model.

The conceptual representation of snow sublimation is not included in the SLURP manual. Geoff Kite (2002b) provided the explanation through a personal correspondence:

“Snow store evaporation (sublimation) is computed exactly as for ET from a soil store, e.g. a potential is computed using your favorite method and then the model attempts to satisfy this. Since temperatures are low, this is often minimal except for days with high radiation input.”

2.4.1.6 Infiltration

Water reaches the soil interface for infiltration into the subsurface storage tanks if rainfall occurs when the canopy storage tank is full or if snowmelt is generated. Infiltration is modelled using a linear relationship between the infiltration rate, **Inf** [mm/day], and the current contents of the fast storage tank, S_1 [mm], stated as:

$$\mathbf{Inf} = \mathbf{Inf}_{\max} \left(1 - \frac{S_1}{S_{1,\max}}\right) \quad 2.18$$

where $S_{1,\max}$ is the capacity of the fast storage tank [mm], and $\mathbf{Inf}_{\max}$ is the maximum infiltration rate [mm/day].

If the fast storage tank is full (i.e. $S_1 = S_{\max}$) or water supply exceeds **Inf**, then overland runoff, **RO**, is generated for subsequent overland routing.

2.4.1.7 Subsurface Flow

There are two subsurface storage tanks: the fast storage tank, S_1 [mm], and the slow storage tank, S_2 [mm]. Subsurface flow, Q_{ss} [mm], from the fast storage tank consists of lateral interflow, **RI** [mm], and downward percolation, **RP** [mm], into the slow storage tank. These flows are found by solving the following three equations:

$$Q_{ss} = \frac{S_1}{k_1} \quad 2.19$$

$$RP = \frac{Q_{ss}}{\left(1 + \frac{S_2}{SSC}\right)} \quad 2.20$$

$$RP + RI = Q_{ss} \quad 2.21$$

where k_1 is fast storage retention constant, SSC is the slow storage capacity [mm], and S_2 is current contents of the slow storage [mm]. Lateral ground water flow from the slow storage tank, RG [mm], is determined as:

$$RG = \frac{S_2}{k_2} \quad 2.22$$

where k_2 is the retention constant of the slow storage tank.

2.4.1.8 Flow Routing Within ASAs

Water is routed within an ASA in two separate stages. During the first stage water is routed to the nearest stream and during the second stage water is routed down the stream to the ASA outlet.

SLURP computes travel times by assuming the distances to-stream and down-stream fit a Normal distribution. It uses the moments of the two distances to compute the to-stream and down-stream travel times. Figure 2.6 illustrates this routing concept that is used by SLURP. Representative minimum to-stream distance, L_{min} , and maximum to-stream distance, L_{max} , are computed using:

$$L_{min} = \mu - 2\sigma \quad 2.23$$

$$L_{min} = \mu + 2\sigma \quad 2.24$$

The minimum and maximum to-stream travel times are computed using Manning's equation:

$$V = \frac{R_H^{\frac{2}{3}}}{n} \left(\frac{H}{L} \right)^{\frac{1}{2}} \quad 2.25$$

where V is the average velocity [m/s], R_H is the hydraulic radius [m^2/m] which is assumed to be wide and shallow flow, H is the average change in elevation for the to-stream and down-stream distances [m], n is Manning's roughness coefficient, and L is the to-stream and down-stream distances [m].

Travel time down-stream to the ASA outlet is computed using the mean distance down-stream, the change in elevation down-stream, and the stream-velocity computed using Manning's equation. Mean channel depth, channel width, and Manning's roughness coefficient are required for the computation of stream velocity. Total travel time to the outlet is the sum of the to-stream and down-stream travel times.

The total travel times are used in a lognormal smoothing filter to distribute the runoff from each land cover over time. The resulting travel times are weighted for the area of each land cover in the ASA, which then allows for conversion to flow for subsequent addition to the total flow of the ASA.

2.4.1.9 Flow Routing Between ASAs

Once all ASA flows is routed to the ASA outlet, it must then continue to be routed through additional downstream ASAs, if they exist, towards the watershed outlet. Flows are also combined with flows from the other ASAs in order to arrive at the total watershed stream flow. SLURP provides three alternatives for between ASA flow routing. The first option is to accumulate ASA flows with no delay or attenuation. The second option is the Muskingum routing technique. The method utilizes only two constants; the relative weight of inflow and outflow in computing channel storage, X , and

the time of travel along the channel reach, $\mathbf{K}$. These two parameters are specified for each ASA in the system. The outflow is computed using:

$$\mathbf{O}_2 = \mathbf{C}_1 \mathbf{I}_2 + \mathbf{C}_2 \mathbf{I}_1 + \mathbf{C}_3 \mathbf{O}_1 + \mathbf{C}_4 \quad 2.26$$

where $\mathbf{O}_t$ is the outflow at time step t , $\mathbf{I}_t$ is the inflow at time step t , and $\mathbf{C}_0$, $\mathbf{C}_1$, $\mathbf{C}_2$, $\mathbf{C}_3$, $\mathbf{C}_4$, are constants determined using the following set of equations:

$$\mathbf{C}_0 = \mathbf{K} - \mathbf{KX} + \frac{\Delta t}{2} \quad 2.27$$

$$\mathbf{C}_1 = -\frac{\mathbf{KX} - \frac{\Delta t}{2}}{\mathbf{C}_0} \quad 2.28$$

$$\mathbf{C}_2 = \frac{\mathbf{KX} + \frac{\Delta t}{2}}{\mathbf{C}_0} \quad 2.29$$

$$\mathbf{C}_3 = \frac{\mathbf{K} - \mathbf{KX} - \frac{\Delta t}{2}}{\mathbf{C}_0} \quad 2.30$$

$$\mathbf{C}_4 = \frac{0.5(\mathbf{q}_1 + \mathbf{q}_2)\Delta\mathbf{x}\Delta}{\mathbf{C}_0} \quad 2.31$$

where $\mathbf{q}_1$ and $\mathbf{q}_2$ are the discharges per unit width at time 1 and 2 respectively [$\text{m}^3/\text{s}/\text{m}$], $\Delta\mathbf{x}$ is the channel length [m], and $\Delta\mathbf{t}$ is the time length [s].

The third option for between ASA routing is the Muskingum-Cunge method. This method may be used where channel geometry is known or can be approximated. These parameters include channel length, change in elevation along the channel, average

channel width, average channel depth, and average channel roughness. The application is similar to the Muskingum method and only differs in the fact that K and X are calculated using the following equations:

$$K = \frac{\Delta x}{c} \quad 2.32$$

$$c = 1.27\theta \frac{S^{0.3}}{q_0^{0.4}} n^{0.6} \quad 2.33$$

$$X = 0.5 \left(1 - \frac{q_0}{c} S_0 \Delta x \right) \quad 2.34$$

where c is the kinematic wave speed [m/s], θ is a function of the cross section shape, (SLURP assumes a rectangle with θ equal to 5/3), S is bed slope, and q_0 is the reference discharge per unit width.

2.4.2 SLURP Applications in the Literature

Although the SLURP model has been chosen for use in this research, literature shows there are relatively few basin scale model applications in northern boreal forests, specifically in Northern Manitoba. This is unfortunate considering that the NOPEX and BOREAS programs collected vast amounts of hydrological data. The only basin scale applications in Northern Manitoba that the author is aware of include the application of SSARR to the Lower Nelson River (Crippen Acres Wardrop, 1990); SLURP to the Sapochi River Basin (Susilo, 2000); WATFLOOD to the Sapochi River Basin (Stadnyk, 2001); and WATFLOOD to the Churchill River Diversion (Snelgrove, 2001).

Since the first version of SLURP was developed in 1975, the SLURP model has been used to successfully simulate historical streamflow records for basins with varying size, climate, terrain, and geography. The characteristics of the watersheds modelled with

SLURP, which are described in the literature, are listed in Table 2.3. The scale of the SLURP applications range from micro-scale to meso-scale. The smallest watershed modelled has an area of 0.47 km^2 and the largest watershed modelled has an area of 1.6 million km^2 . The types of terrain vary significantly between watersheds and include wetland, prairie, mountainous and subarctic, and agricultural. As listed in Table 2.3, SLURP has been applied several times in North America, South America, Eastern Europe, Africa, and Asia. The remaining portion of this section describes a number of these SLURP applications.

Kite (Kite, 1991; Slough & Kite, 1992) calibrated the SLURP model to the $7,100 \text{ km}^2$ Kootenay watershed located in the Rocky Mountains. The application of the SLURP model was part of a large study that compared the performances of lumped, semi-distributed and fully distributed hydrological models on the same basins over the same periods (Kite, 1991). The secondary purpose of the application was to study the benefits of using remotely sensed data in the SLURP hydrological model. The watershed was divided into three ASAs. The model was calibrated for all of 1988 and verified for January to July of 1989 using stream flow data from one hydrometric gauge. NOAA AVHRR was used to estimate snow cover extent and cloud cover for days where data was available. Compared to model runs using no remotely sensed data, results showed that when cloud cover and snow cover extent was used as inputs, the efficiency increased by 1% and decreased by 4% for the calibration and verification periods, respectively. Results also showed that when only snow cover extent was used, the efficiency increased by 2% for both the calibration and verification periods. Kite concluded that "using satellite snow cover produced the greatest effect while using cloud cover as an indicator of precipitation distribution was not as successful" (Kite, 1991). This study also showed that snow-course data did not improve the model output. In a 1992 study, Kite and Slough (Kite & Slough, 1992) demonstrated that using DMSP SSM/I microwave estimates of snow water equivalent produced better results verses snow course data in the application of SLURP to the Kootenay watershed.

Kite successfully used SLURP to simulate historical streamflow records in the Upper Columbia watershed for three separate studies (Kite, 1995; Kite, 1997; Kite, 1998b). The 1995 study examined how remotely sensed data could be used to enhance the results of the watershed simulation by applying snow cover area and snow water equivalent data. The 1997 study examined the use of meteorological data generated by Numerical Weather Prediction (NWP) models at various scales, for input to the SLURP model in order to overcome “the inadequacy and inaccuracy of the existing climate observation network” (Kite, 1998b). The 1998 study examined the use of a High-Resolution Boundary Layer (HRBL) model in order to extend the hydrological analysis to longer periods. The watershed has a total area of 36,000 km² and the terrain is mainly mountainous. In all studies, the watershed was divided into 48 ASAs with a mean area of 785 km² and ranged in size from 15 km² to 7300 km². Five and six land cover classes were derived using Landsat MSS and NOAA AVHRR scene classifications (Kite, 1997). Snow cover extent was estimated using NOAA AVHRR scenes, and snow water equivalent was estimated using DMSP SSM/I data. The monthly leaf area index was derived using NOAA AVHRR data. The calibration period used was between 1986 and 1990. The Nash and Sutcliffe efficiencies calculated for the calibration period for the twenty ASAs with streamflow gauges averaged 0.36 with a minimum of -1.91 and a maximum of 0.83. From the three studies, Kite concluded that SCA and SWE are extremely important variables in the operation of the SLURP model. The SWE values computed by SLURP and the NWP models were found to be similar. Finally, it was shown that SLURP can reproduce historical stream flows based on inputs from atmospheric models such as NWP and HRBL models, thus making SLURP a useful tool for climate change studies.

The largest scale application of the SLURP model is that of the Mackenzie River Basin (Kite et al., 1994; Haberlandt & Kite, 1998a; Kite et al., 1996). It has a total area of 1,600,000 km² and is located in northern Canada. The SLURP model has been calibrated to this basin a number of times because it is the site of many studies in the Canadian input to the Global Energy and Water Experiment (GEWEX) international project (Kite et al., 1994). The primary reason for the most recent application was to investigate methods of

improving the precipitation database for hydrological simulations (Haberlandt & Kite, 1998a). Methods of improvement included the use of geostatistical techniques to distribute precipitation data, and the use of data generated by atmospheric models. The watershed was divided into 39 ASAs, where five ASA outlets correspond to the locations of five Water Survey of Canada stream flow gauges. A total of ten land cover classes were derived using NOAA-AVHRR scene classifications (Haberlandt & Kite, 1998). The study found that geostatistical techniques can improve hydrological simulations. It also found that using broad-scale atmospheric precipitation data in combination with observed precipitation improved flow simulations. One study investigated the use of SLURP to estimate the lateral transfer of water within the land phase in order to improve output from global circulation models (Kite et al., 1994). This model used five ASAs and five land covers classes. The study demonstrated that streamflow estimates are greatly improved when SLURP is used to estimate the land surface water transfers for the GCM, when compared to stream flows calculated by the GCM routines. Kite et al. (1994) also found that “using the CC GCM II data as distributed input to SLURP produces a hydrograph closer to the recorded one than using observed climatic data”.

The smallest scale application of the SLURP model is that of a small prairie watershed containing a wetland located in Saskatchewan (Stolte & Van der Kamp, 1999). The primary reason for the application was to gain a better understanding of the water balance in wetlands by studying wetland water level fluctuations. Since the few wetland hydrology models available are not suitable for this study, a number of modifications were made to SLURP in order to customize its application. Due to the fluctuating size of the wetland, the program was modified so that the size of the wetland is a function of the water level. Another notable change was a different coefficient of saturated permeability in the fast tank for frozen and unfrozen ground. The watershed was divided into three ASAs: one for wetland, upland, and grassland cover. The “simulation results indicate that the modified SLURP model can adequately represent water level variations in prairie wetlands”. The authors concluded that SLURP “has the potential to be used for exploring the effects of possible climate changes and land use on wetland resources in the future.

The Drainage and Flood Control Committee of the Upper Assiniboine River Basin Study attempted to apply the SLURP model to the Upper Assiniboine River Basin (Weins, 2000). The primary reason for the study was to determine to what extent the changes to the landscape in the Upper Assiniboine River basin landscape has changed through agricultural drainage and bush clearing and evaluating the impact of those changes on downstream flood hydrographs” (Weins, 2000). Fifteen hydrological models were assessed for their appropriateness for study based on model features and capabilities. The SLURP model was chosen because of its use of limited data, snowmelt algorithms, and the local support from the National Hydrology Research Centre. Six land cover classes including large lakes, cropland, grass, shrub, wetland, and woodland were modelled with a total of 25 ASAs. Results of the modelling study showed that the model was not sensitive to land cover changes. It was also apparent that there were a significant number of potholes and sloughs that are disconnected from the drainage system. This feature was not considered in the model application and contributed to modelling error.

There have been a number of SLURP applications beyond North America as indicated in Table 2.3. One of those applications is that of the Lower Satluj catchment located in the western Himalayas of India and Tibet (Sanjay et al., 1998). Within the Indo-Gangetic plains, irrigation, hydropower generation, and water supply are dependent on the availability of water therefore, this study examined the use of SLURP to estimate streamflow for an area of 5513 km². The study area is characterized by tall mountains that surround a valley floor that is developed for agricultural purposes and has a total relief of 3500m. It was divided into ten ASAs and three land cover classes based on Indian Remote Sensing imagery. Thiessen polygons were used to distribute precipitation data from four rainfall gauges. The model was calibrated for January to December of 1992 and verified for January to December of 1993. The standard error for the two periods were 1.77 and 1.48, respectively. Authors concluded “that the SLURP hydrological model can be applied to a basin in India using available meteorological data, topographic data, and satellite imagery” (Sanjay et al., 1998).

The 11th largest river in the world, the Mekong River of southeast Asia, is another large scale application of SLURP (Kite, 2001). The basin has a total area of approximately 800,000 km² and is located in southern China, Burma, Thailand, Laos, Cambodia, and finally draining into the ocean through Vietnam. The application was part of a larger study trying to determine the “implications of developments on interests such as fisheries and the environment”. One area of special interest for the modeling is the water level in the Tonle Sap, an area of Cambodia containing 60-75% of that nations inland fishery. Lake levels and river flows were modeled reasonably well.

The SLURP model was applied to the Gediz and Küçük Menderes basins of western Turkey (Kite, 2000). The Gediz basin covers 17,220 km² and the Menderes covers 3,617 km². Both basins contain large areas of agricultural crops including cotton, cereals, grapes, vegetables and fruits, olives, tobacco, and melons. Hence, large lands are irrigated and place heavy demands on the local reservoirs. The modelling focused on predicting the evaporation and transpiration losses for the various crops. ET was estimated for 37 land covers and the temporal scale extended from 1 day to 11 years. Most of the spatial data used in the study was public domain data that was obtained for free from the Internet. Results showed that ET varied “across the basin on any one day and over the irrigation season” and that the SLURP model “can be used to evaluate the effectiveness of the various irrigation strategies used in the basins” (Kite, 2000).

The SLURP model was also applied to the Olifants Basin of South Africa. The Olifants river basin covers an area of 54,600 km² and is located within three provinces of north-eastern South Africa, reaching the border with Mozambique. The basin is mainly agricultural with a large degree of irrigation. The main issues identified in the Olifants River basin are availability of water in relation to demand, the quality of water, and the impact of land use on the water resources. The application of SLURP to this basin will play a role studying the existing water resources management practices and the investigation of alternative scenarios of management as well as alternative climate and water supply scenarios (Kite, 2002).

The previous sections summarized a number of applications of the SLURP hydrological model. Based on the review of literature, it is apparent that SLURP hydrologic model is a flexible and robust model as it has been applied to a number of basins at various scales ranging from 0.47 km² to 1.6 million km². It is also demonstrated that SLURP can be successfully applied to basins all around the world that comprise a wide range of climatic and geographic conditions including sub-arctic, agricultural, irrigated, mountainous terrain, and wetland. Hence, a general review of these SLURP applications has increased the writer's confidence with respect to the appropriateness of the model for this study.

2.4.3 Sensitivity Analysis

A sensitivity analysis aims to quantify the influence that model parameters have on model output. In the case of hydrological models, the objective is to determine how each model parameter changes the flow, volume, and temporal characteristics of the output hydrograph. A basis of comparison (BOC) hydrograph is compared to all other hydrographs resulting from parameter modifications. The parameters will have varying impacts on the hydrograph either causing flows to increase or decrease as the parameter changes. Hence, the sensitivity analysis identifies the parameters that the model is sensitive and insensitive to and the degree of that sensitivity. Special consideration is then given to the values of the influencing parameters since a small degree of error can have an adverse affect on model results. This facilitates the determination of the parameter set that best simulates the hydrograph for the area of interest. These parameters are used for model calibration since a small change in the parameter value creates a significant change in the output.

Complex sensitivity analyses have been carried out to assess parameter variations on the Nash and Sutcliffe efficiency, Garrick efficiency, and Previous-day criterion (Kite, 1991). However, this task requires a model that is calibrated to an acceptable degree of accuracy. This research used a simple approach to assess the sensitivity of parameters in order to facilitate model calibration. Here, only the impact on flow magnitude relative to the BOC flow is evaluated and is deemed to be sufficient for the purposes of this research.

A preliminary SLURP application to the Taylor River Basin was used as the BOC. The model contains nine ASA's delineated using the TOPAZ physiographic parameterization tool (Garbrecht and Campbell, 1997). The model uses the Spittlehouse Black evapotranspiration method. For the exercise meteorological data from the Thompson Airport, located approximately 50 kilometers northeast of the basin, is used for model input. The simulation period is January 1993 to December 1993. For simplicity the land cover is assumed to be 100% forest cover throughout the basin. Parameter estimates for this model were extracted from the Kootenay River Basin model, contained within the SLURP manual (Kite, 1998a). Some of the model parameter values were calibrated using the SLURP automatic calibration routine. The Nash and Sutcliffe efficiency for the BOC is 34%. Although the efficiency is low, and the resulting hydrograph is significantly different from the observed hydrograph, it is assumed to be at an acceptable level for the purposes of conducting a sensitivity analysis.

The sensitivity analysis examined the 22 model parameters listed in Table 2.4. The parameters were adjusted by six equal increments of 5% from the BOC parameter set resulting in a parameter set that is -15%, -10%, -5%, +5%, +10%, and +15% of the BOC values. This was conducted in order to examine the rate of change in model outputs. Figure 2.7 is a sample graphical representation of the model sensitivity to variations in the Retention Constant of the Slow Storage Tank. In order to rank and classify the sensitivity of each parameter, streamflow was computed using the -15% and +15% BOC parameters, resulting in the $Q(-15\%)$ [m^3/s] and $Q(+15\%)$ [m^3/s] hydrographs respectively. Using these two hydrographs, the difference in the flows for each day, j , in the one year simulation period were computed and compared to the BOC flow for the same day, $Q(\text{BOC})$ [m^3/s]. Finally, the maximum difference within the one year period, DF [percent], was determined:

$$\text{DF} = \text{Max} \left[\text{abs} \left(\frac{\{Q(-15\%)_j - Q(+15\%)_j\} * 100}{Q(\text{BOC})_j} \right) \right] \quad 2.35$$

The sensitivity classification for each parameter was arbitrarily assigned based on the percentage difference values. Four classes were used: Very Low ($DF < 5$), Low ($5 < DF < 10$), High ($10 < DF < 20$), and Very High ($DF > 20$). Table 2.4 lists the results and lists the parameters in descending sensitivity. The model was very sensitive to Slow Storage Tank parameters including the Retention Constant (k_2), Initial Storage, and the Capacity (SSC). Also very sensitive are the Retention Constant (k_1) of the Fast Storage Tank, and the snowmelt rates ($R_{1,Jan}$, $R_{1,Jul}$). The model was insensitive to the Priestly-Taylor Coefficients (α_s , α_w , β) Canopy Storage Capacity (CC), Snow and Land Albedos (v_s , v_L). No modelling sensitivity was observed for the infiltration capacity (Inf_{max}), indicating that the infiltration capacity was never reached during the various simulations.

2.5 Chapter Summary

The High Boreal Forest of Northern Manitoba has unique topographic, meteorological, and land cover characteristics that shape the hydrological processes of the region which include:

1. Terrain that is hummocky, has little relief, and is covered mainly with muskeg, coniferous and deciduous forest. A typical soil profile consists of an organic moss and peat layer overlying a less permeable silty clay layer.
2. Hummocky terrain that contains many small perched clay lined depressions that are not hydraulically connected to the drainage system. Their contribution to total evaporation is a key component of the boreal forest water balance dynamics. Once the depressions are filled with water, runoff is generated with additional precipitation and snowmelt inputs.
3. A moss layer rain interception that is significant and results in considerable evaporation losses. The relatively large tree and moss interception stores the small precipitation events.
4. A hydraulic conductivity of the Muskeg organic soils which increases exponentially with depth resulting in a very marked decrease in interflow as the water table falls towards the relatively impermeable inert layers. Runoff from Muskeg virtually ceases once the water table reaches a limiting depth.

5. The ability of the Boreal forests to enhance ET by releasing energy stored in the canopy when moisture levels are high. During dry conditions, the energy is released to warm the environment. ET from coniferous trees is relatively small during the growing season.
6. The formation and degradation of a frost active layer which is a dynamic process and often forms in Muskeg. It freezes and thaws in response to the air and surface temperatures, vegetation covers, snow cover, local moisture regimes, and subsurface properties. Soil frost affects the water regime by modifying the infiltration characteristics, temporarily storing water as ice, and varies the upper soil storage capacity as the thaw depth changes. A frost layer develops over the course of the winter, the depth varies inversely with snowpack depth. Unique types of muskeg are underlain by frost year round.
7. Overland flow that is commonly generated during the snowmelt period because infiltration capacities are greatly reduced when concrete frost is produced. Honey comb frost forms in dry soils resulting in no change in or increased infiltration capacities.

The SLURP hydrologic model was chosen for application because it is shown to be a flexible model that performs well at a variety of scales, at a wide variety of basins and types of land covers located around the world, of which some basins are similar to the area of interest. The conceptual representation of the watershed is physically based and contains relatively few calibration parameters. A sensitivity analysis revealed that the model is sensitive to changes in only a few of the model parameters.

Figure 2.1 – Location of the Upper Burntwood River and Taylor River.

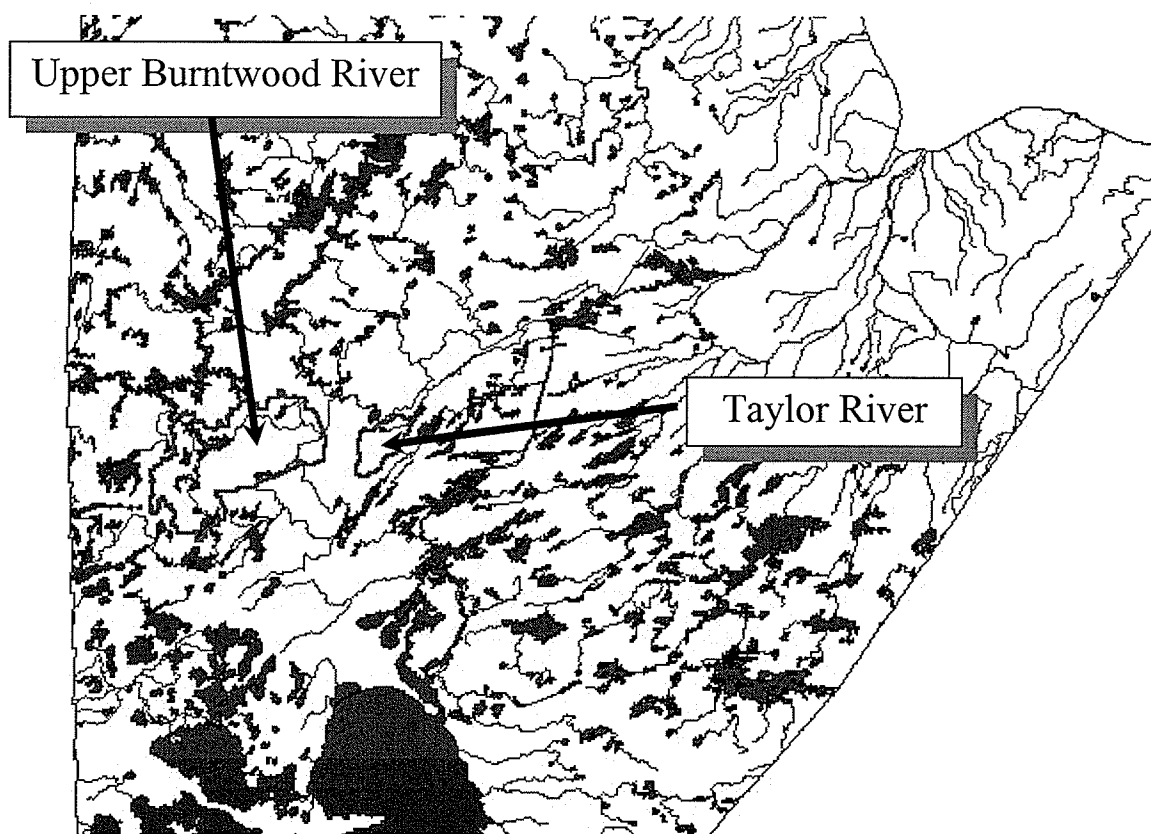


Figure 2.2 – Comparison of lumped and distributed hydrologic models.
(Fast, 1999).

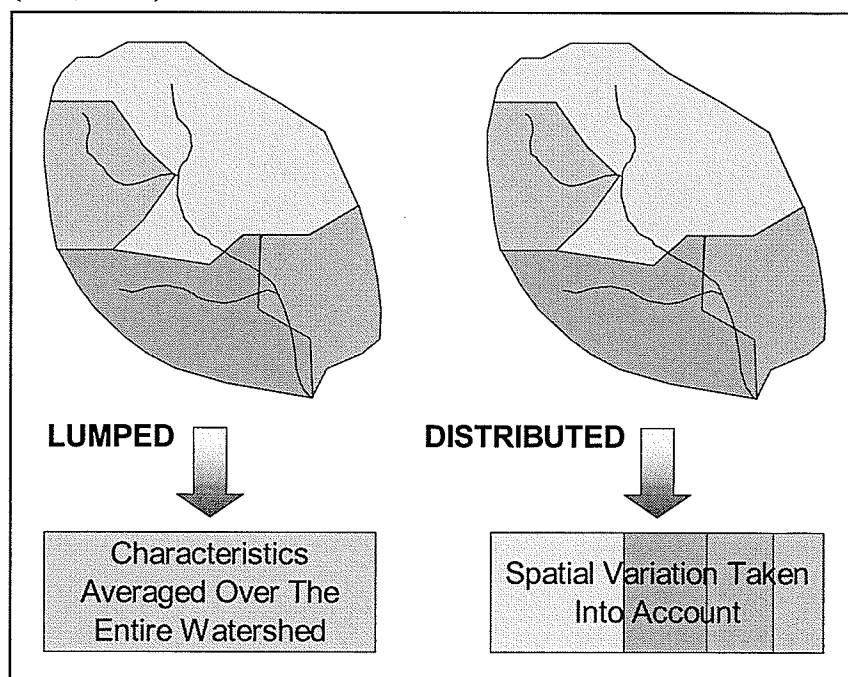


Figure 2.3 – Aggregation of land cover units into ASA's in the SLURP model.

(Modified from Kite, 1998a)

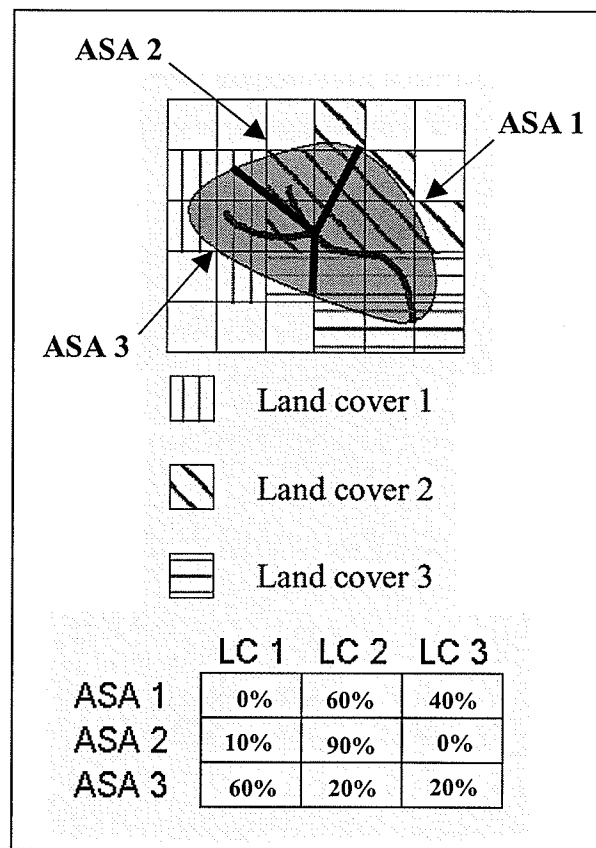


Figure 2.4 – Simplified vertical water balance applied to each land class within an ASA in the SLURP model.

(Modified from Kite, 1998a)

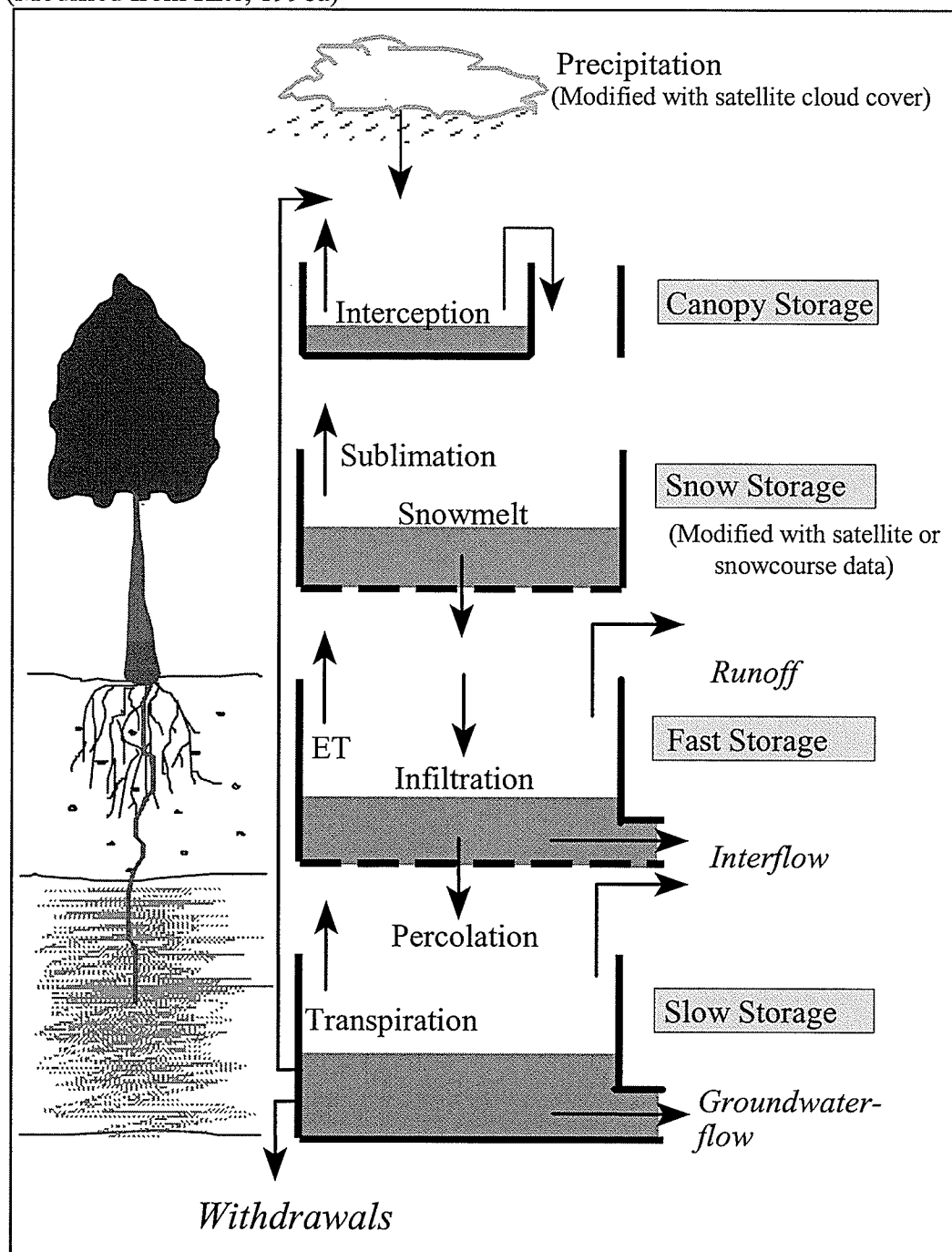


Figure 2.5 – The Spittlehouse-Black conversion of Potential ET to Actual ET.

(Modified from Kite, 2002)

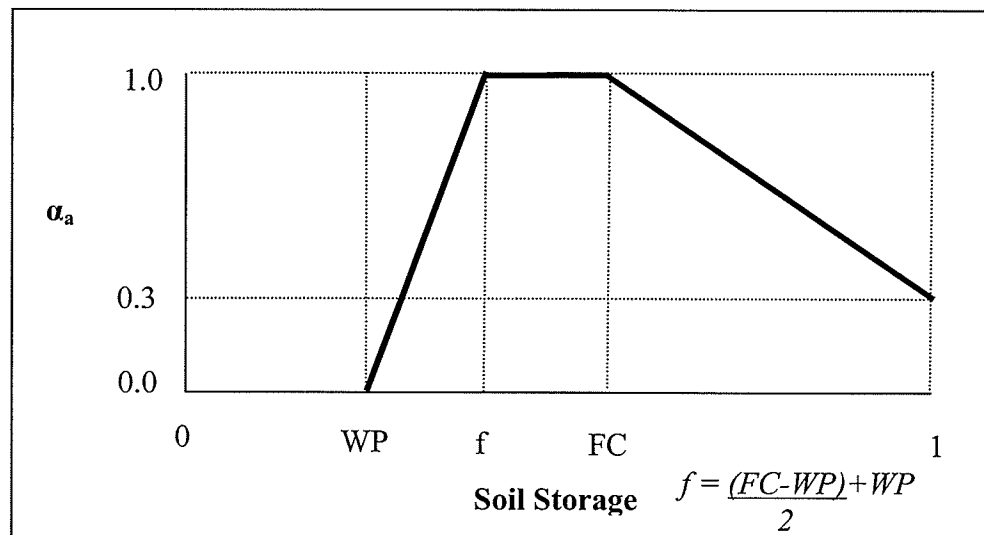


Figure 2.6 – Concept of the within-ASA routing scheme in the SLURP model.

(Modified from Kite, 1995)

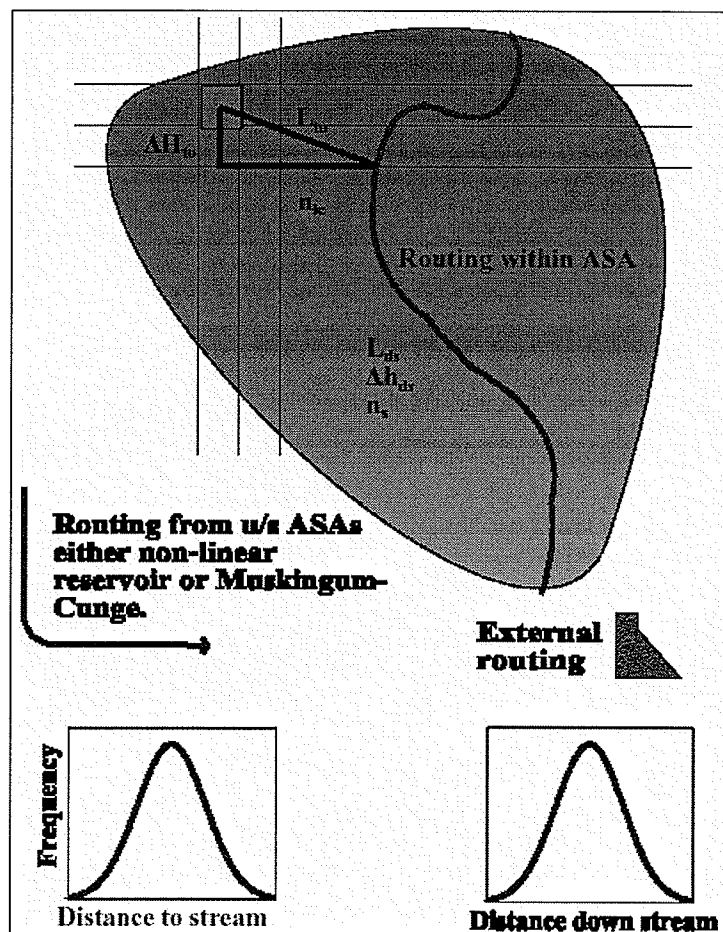


Figure 2.7 – The sensitivity of the simulated flow to variations in the Retention Constant for the Slow Storage tank.

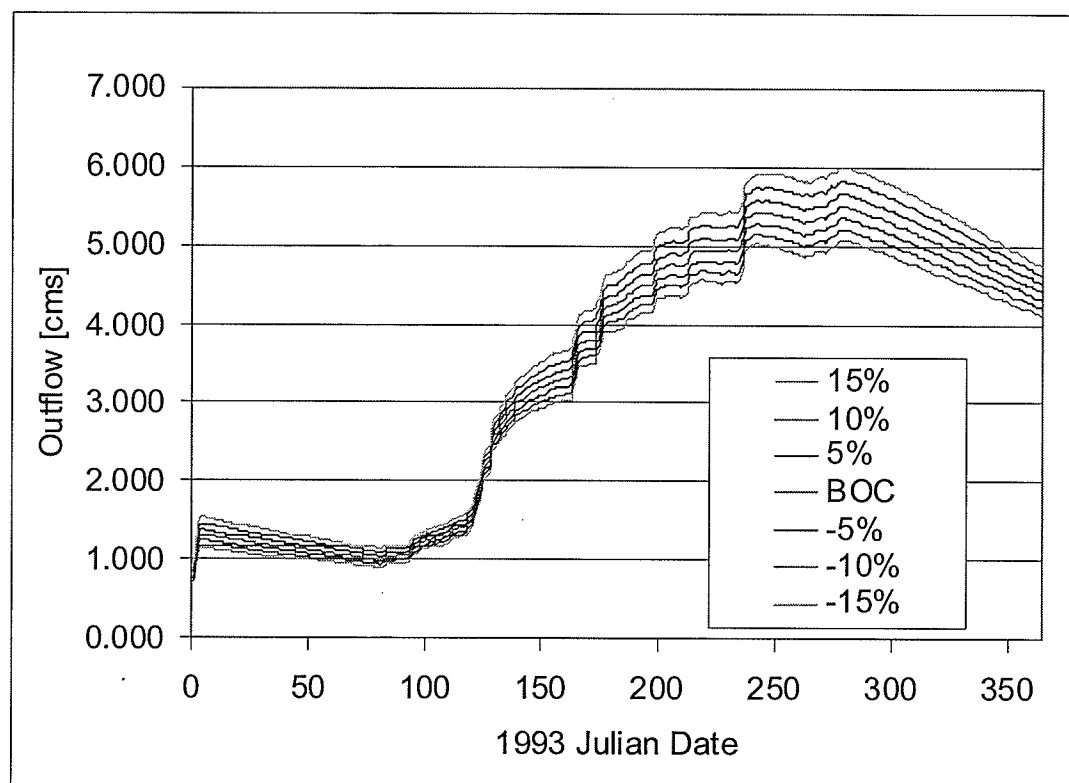


Table 2.1 – Values of LAI in the literature.

Land Cover Type	Location	LAI	Source
Coniferous	Kootenay River, B.C.	2-3	Kite, 1998a
Spruce & Jackpine	BOREAS NSA, Manitoba	1.9-4.3	Newcommer et al., 2000
Spruce	Wolf Creek - Yukon	4	Kite & Spence, 1994
Mixed Deciduous & Coniferous	BOREAS SSA, Saskatchewan	Range:2.0-7.8 Average: 4.9	Newcommer et al., 2000
Deciduous	High Latitude	8	Kite, 2002
Impervious	Kootenay River, B.C.	0.0-0.5	Kite, 1998a
Grassland	Montana	2	Running et al. 1986
Shrubland	Wolf Creek, Yukon	5	Kite & Spence, 1994
Tundra	Wolf Creek, Yukon	2	Kite & Spence, 1994

Table 2.2 – Parameters in the SLURP model. (modified from Kite, 1998a)

Parameter Name	How Derived
Initial contents of snow store*1 [mm]	Measured (Field) / Estimated
Initial contents of slow store*1 [mm]	Estimated
Maximum infiltration rate*1 [Inf_{max} , mm/day]	Estimated
Manning roughness surface*1 [n]	Estimated
Retention constant fast store*1 [k_1]	Estimated
Max. capacity fast store*1 [$S_{1,max}$, mm]	Measured (Field) / Estimated
Retention constant slow store*1 [k_2]	Estimated
Max. capacity slow store*1 [SSC, mm]	Measured (Field) / Estimated
Precipitation factor*1 [PF, decimal]	Estimated
Snow melt temperature [T_c , °C]	Estimated
Temperature lapse rate [T_{LR} , °C/100m]	Measured (Field) / Estimated
Precipitation lapse rate [P_{LR} , mm/100m]	Measured (Field) / Estimated
Initial contents of canopy store [mm]	Initially set to zero
Initial contents of fast store [mm]	Initially set to zero
Max. capacity of canopy store [CC_{max}]	Estimated
Albedos (surface, snow) [v_s , v_L]	Measured (RS) / Estimated
Leaf area index [LAI]	Measured (RS) / Estimated
Interception coefficients [A, B]	Estimated
Soil heat flux [G, MJ/m ² /d]	Estimated
Areas of the land covers [km ²]	Measured (GIS-DEM)
Elevations of the land covers [m]	Measured (GIS-DEM)
Distances to stream [m]	Measured (GIS-DEM)
Distances downstream [m]	Measured (GIS-DEM)
Changes in elevation to stream [m]	Measured (GIS-DEM)
Changes in elevation downstream [m]	Measured (GIS-DEM)
Snow melt rates [R_1 , R_2 , mm/day/°C] and radiation coefficient [α_{sm}]	Estimated
Muskingum routing coeff. x, K*1	Estimated
River geometry	Measured (Field) / Estimated
Priestley-Taylor coefficients*2 [α , β]	Estimated
Aerodynamic roughness or wind function*2	Estimated
Field capacity and wilting point*2 [fc, wp]	Measured (Field) / Estimated
*1 : Parameter can be included in the automatic calibration procedure	
*2 : Parameter is optional	

Figure 2.3 – Applications of the SLURP Hydrological Model.

Name	Area (km ²)	Geographic Type	Location	Source
Wetland	0.47	Wetland	Saskatchewan	Stolte & Van der Kamp, 1999
Wolf Creek	194	Mountainous	Michigan, USA	Kite, 1998
Carp River Watershed	250	unknown	Ontario	Sabourin & Associates, 1996
Küçük Menderes Basin	3,617	Agricultural with highlands	Western Turkey	Kite, 2002
Lower Satluj Catchment	5,513	Mountainous	India, Tibet	Sanjay K. Jain, et al., 1998
Kooteney Basin	7,129	Mountainous	Rocky Mountains of B.C.	Kite, 1998
Gediz Basin	17,220	Agricultural with highlands	Western Turkey	Kite, 2002
Assiniboine River	22,000	Prairie/Agricultural	Manitoba & Saskatchewan	Weins, 2000
Upper Columbia River	36,000	Mountainous	British Columbia, and northern USA	Kite, 1995
Olifants Basin	54,600	Agricultural and undeveloped	Republic of South Africa	Kite, 2002
Souris River	58,700	Prairie	Manitoba & Saskatchewan	Kite, 2002
Mekong River	800,000	Agricultural with highlands	China, Burma, Thailand, Laos, Cambodia, Vietnam	Kite, 2001
Mackenzie Basin	1,600,000	Subarctic	Northern Canada	Kite et al, 1994
St.Maurice	unknown	unknown	Eastern Canada	Kite, 2002
St.Denis	unknown	unknown	Manitoba & Saskatchewan	Kite, 2002
Magdalena	unknown	unknown	South America	Kite, 2002
Malewe	unknown	unknown	South Africa	Kite, 2002
Martin Watershed	unknown	Subarctic	Northwest Territories	Kite, 2002
Uda Walawe	unknown	unknown	Sri Lanka	Kite, 2002

Table 2.4 - Sensitivity of the SLURP simulated streamflow to changes in model parameters.

#	Parameter Name	DF	Classification
1	Retention Constant Slow Storage	30.67	Very High
2	Initial Contents of Slow Storage [mm]	29.99	Very High
3	Maximum Capacity of Slow Storage [mm]	29.95	Very High
4	Retention Constant Fast Storage	19.30	High
5	No Radiation Snowmelt Rate - January [mm/C/day]	17.40	High
6	No Radiation Snowmelt Rate - July [mm/C/day]	16.70	High
7	Initial Contents of Snow Storage [mm]	15.60	High
8	Maximum Capacity of Fast Storage [mm]	11.73	Medium
9	Wilting Point	11.66	Medium
10	Manning Roughness Coefficient	11.34	Medium
11	Snowmelt Temperature [C]	11.25	Medium
12	Minimum Snow Albedo	10.05	Medium
13	Interception Coefficient: B	8.38	Low
14	Priestley-Taylor Coefficient: Alpha Water	7.36	Low
15	Field Capacity	5.63	Low
16	Interception Coefficient: A	5.38	Low
17	Priestley-Taylor Coefficient: Beta	4.56	Very Low
18	Maximum Contents of Canopy Storage [mm]	2.83	Very Low
19	Maximum Snow Albedo	1.24	Very Low
20	Priestley-Taylor Coefficient: Alpha Soil	0.99	Very Low
21	Surface Albedo	0.77	Very Low
22	Maximum Infiltration Rate [mm/day]	0.00	Very Low

CHAPTER 3

3 PREPARATION OF THE SLURP MODEL DATABASE

3.1 Introduction

This chapter presents an overview of the two study basins used in this research, the Upper Burntwood River watershed and the Taylor River watersheds. It describes the development of the SLURP model databases for the two study areas including the computation of physiographic parameters, meteorological inputs, and hydrometric inputs. The GIS procedures used derive the physiographic parameters are described. Finally, the chapter presents a database of parametric values based on literature.

3.2 Upper Burntwood River and Taylor River Watersheds

The Upper Burntwood River and Taylor River watersheds are located in northern Manitoba west of the City of Thompson. They are both tributaries of the Churchill River Diversion (CRD) (Figure 3.1), a project whereby Manitoba Hydro diverts the majority of the natural streamflow from the Churchill River into the Burntwood River via the Rat River system. The Burntwood River ultimately flows into the Nelson River for hydro power production.

Manitoba Hydro has expressed an interest in using a deterministic hydrologic model to simulate streamflow on the Upper Burntwood River and Taylor River watersheds in order to support hydropower planning studies, dam safety studies, and operations. Both basins contain the topography, land cover, and hydrology that are characteristic of the High Boreal Forest region surrounding the CRD and Upper Nelson River regions. Long term flow records, digital topography, and digital land cover exist for both basins facilitating the development of a SLURP model database and the calibration and the verification process. These two basins will form the basis of the study areas of this research.

The Upper Burntwood River is that portion of the Burntwood River upstream of Threepoint Lake as shown in Figure 3.2. The watershed area is 6959 km² and it is the largest tributary of the CRD. The basin has 128 m of relief as the elevation ranges between 239 m – 367 m above sea level. The Upper Burntwood River contains a significant amount of lake storage with several very large lakes in the upper portions of the basin. The largest of these being the Burntwood Lake which is 140 km² in area.

The Manasan River is another major tributary of the CRD. The mouth of this river is located 11 km upstream of Thompson. The Manasan River contains many sub-basins, the largest being the Taylor River sub-basin, shown in Figure 3.2. The Taylor River watershed is 899 km² in area and has approximately 94 m of relief as the elevation ranges between 187 m – 281 m above sea level. The largest lakes in this basin are Bison Lake (9 km²) and Tullibee Lake (7 km²).

3.3 Database Construction

This section describes the data sets and methodologies used to construct the SLURP model databases for the two study areas. These data sets include physiographic data (i.e. elevation, area, and land cover), time series data (i.e. meteorological and hydrometric), and parametric data (i.e. coefficients or constants used within model equations).

3.3.1 Physiographic Data

3.3.1.1 Digital Elevation Model

The SLURP model requires a number of physiographic parameters that are derived from elevation data. An efficient method for determining these parameters is to use a GIS to summarize a digital model of regularly spaced point height values, known as a “Digital Elevation Model” (DEM) or Raster DEM.

A number of agencies that publish DEMs were investigated as a potential source for this study. However, due to the remoteness of the study area, high resolution DEMs (< 1 km) were not available. Alternatively a high resolution (100m) DEM was created using the ARC/INFO and ArcView GIS software packages. The method requires relatively detailed hypsographic and hydrographic data. A total of 24 digital National Topographic System (NTS) maps at a scale of 1:50,000 were utilized (See Figure 3.3 for spatial coverages). The NTS data utilizes the Universal Transverse Mercator (UTM) North American Datum 83 (NAD 83) projection Zone 14. This projection was used for all other data sets utilized in the remainder of this study. All NTS maps were obtained in an AutoCAD format. The hypsography-contour layer, hypsography-points layer, and hydrography layer were imported into the ArcView GIS software. Each layer from the 24 NTS maps was merged into a single layer using the ArcView Geoprocessing Wizard.

DEMs that are used in hydrological process studies should accurately and realistically depict the topography of the area of interest. Sinks or pits are normal by-products of DEM interpolation methods. Steps should be taken to avoid sinks when creating DEMs in order to facilitate the delineation of the watershed. ARC/INFO includes an interpolation method developed by Hutchinson (1989), which is an efficient finite difference method of interpolating Raster DEMs from contour line data, scattered surface-specific spot elevations, and hydrographic information. The method uses a drainage enforcement algorithm that automatically removes spurious sinks or pits and calculates streamlines and ridgelines from points of locally maximum curvature on contour lines. The drainage enforcement algorithm “has been found in practice to be a powerful condition which significantly increases the accuracy, especially in terms of their drainage characteristics”

(Hutchison, 1989). The interpolation is accessed in ARC/INFO using the **TOPOGRID** command. It utilizes four sources of information to develop the raster DEM as shown in Figure 3.4. These are: a points coverage representing spot elevations; a polyline coverage representing contours; a polyline coverage representing the stream network; and a polygon coverage representing lakes. All of the NTS data merged in ArcView were imported into ARC/INFO. The command used to import the ArcView shapefiles is:

Arc: SHAPEARC <input_shape_file> <output_coverage>

This command converts an ArcView Shapefile into an ARC/INFO coverage. The hypsography coverage for both basins is shown in Figure 3.5. The Forest Resources Inventory (FRI) of Manitoba lake polygons were used as the lakes coverage for DEM construction, shown in Figure 3.6. Details regarding the FRI data can be found in Section 3.3.1.2 Land Cover. By including a lake coverage, **TOPOGRID** determines the minimum elevation at the perimeter of the lake and assigns this elevation to all interior points in order to create a flat lake.

The hydrography coverage required a significant amount of preprocessing before the DEM could be constructed. **TOPOGRID** requires a continuous well-defined stream network with directional topology. That is, all stream arcs must be connected and oriented downstream. This was not established in the original NTS maps since the streams were drawn for visual purposes using AutoCAD. Using the ARC/INFO **ARCEDIT** and **FLIP** commands were used to flip those arcs pointing in the wrong direction on an arc by arc basis. Additionally, the stream network must also continue through the lakes. New arcs were drawn through the lakes to create a continuous stream network. Figure 3.7 shows a portion of the stream network that is oriented in the proper direction. Figure 3.8 and Figure 3.9 include the original NTS hydrography and the new topologically corrected hydrography for both basins.

An ARC/INFO Arc Macro Language (AML) script was written to execute the **TOPOGRID** function with the appropriate parameters. The script named "TOPOGRID

AML” is included in Appendix A. Please refer to ARC/INFO help files for an in depth explanation of the code.

A TOPOGRID deficiency was found while creating initial versions of the DEMs. The function ignores lake islands and assigns all interior lake points the same elevation that is assumed to represent the water level of the lake. The study basins contain a few very large lakes with large islands. This was overcome by creating two DEMs, one including the lakes coverage (DEM#1) and one excluding the lakes coverage (DEM#2). The island elevations from DEM#2 were overlaid onto DEM#1 to create the final DEM. The final DEM is shown in Figure 3.10 and Figure 3.11. The resolution of the DEM is 100 m.

The DEM elevations were verified prior to using the DEMs. A series of contours were created using the DEMs within ArcView. These contours were overlaid on top of the original NTS contours as shown in Figure 3.12. Overall, the DEM contours are consistent with the NTS contours.

3.3.1.2 Land Cover

This study uses highly detailed and fine resolution Forest Resource Inventory (FRI) of Manitoba data to represent the land cover in the SLURP model. The data, produced by Manitoba Conservation’s Forestry Branch, is interpreted from 1983-1988 areal photos taken at a 1:15,840 scale. The data is distributed in ArcView Polygon Shapefile format on a township by township basis, of which 193 townships were utilized for this study.

The FRI data is produced in order to identify the quantity and type of productive forest in Manitoba. Within the 193 townships used for this study, the unproductive forest polygons are classified using 32 unique land cover classes shown in Table 3.1. These comprise 35,541 polygons or 43% of all polygons (~50% of the area). The remaining 46,703 polygons, 57% of all polygons (~50% of the area), are classified as productive forest using mixed classes. There are a large number of mixed classes within the FRI polygons. A mixed classification indicates the proportion of up to four different types of plots to the nearest 10% within the polygon. For example, a polygon is classified as

“BS5JP3TA1W1” that indicates the polygon is 50% Black Spruce, 30% Jack Pine, 10% Tamarack, and 10% Willow. The various types of plot species, type, and habitat are listed in Table 3.2. In an effort to reduce the number of classifications, the 46,703 Productive Forest polygons were reclassified into a few unique classes based on the following criteria:

1. If there is one type of plot that comprises 80% or more of the polygon, it is reclassified as that plot type. Result = 18,627 polygons (2,788.2 km²).
2. If the first criterion is not satisfied, and 80% or more of the polygon is comprised of Coniferous Forest, the polygon is reclassified as Coniferous Forest. Result = 19,272 polygons (4,840.6 km²).
3. If the first criterion is not satisfied and 80% of the polygon is comprised of Deciduous Forest, the polygon is reclassified as Deciduous Forest. Result = 135 polygons (19.5 km²).
4. All remaining polygons that do not satisfy the first three criteria are classified as Mixed forest (i.e. both Coniferous and Deciduous Forest). Result = 8,669 polygons (1,462.6 km²).

The reclassification resulted in 43 classes for 82,244 polygons. It was deemed impractical to include all 43 unique land cover types in the modelling process as the parameters for many of these classes could not be distinguished in the modelling process. Therefore, all of the polygons were reclassified into seven land cover classes based on the similarity of vegetation types and hydrologic response. These seven land cover classes include coniferous treed, deciduous treed, mixed treed, muskeg, treed rock/impervious, lakes/rivers, and marsh/beaver flood. The result of the reclassification is shown in

Table 3.3. The final land cover polygon shapefiles were converted into a raster grid with a resolution of 50 m shown in Figure 3.13.

3.3.1.3 ASA Delineation and Parameterization

The delineation of the ASAs was conducted using SLURPView (Fast, 1999). SLURPView is an extension to the ArcView GIS package that links the SLURP model to ArcView. The integration was developed to replace the original SLURP-TOPAZ system and allows the user to conduct their hydrological modelling and physiographic data derivation needs in one platform. Figure 3.14a shows the SLURPView system and the available options within the main ArcView menu. SLURPView was developed using the Avenue programming language and is expected to facilitate the application of the SLURP model to other basins in the region. One of the many parameter entry screens is shown in Figure 3.14b. The primary advantage to SLURPView is that it allows for the delineation of ASAs and the derivation of the ASA physiographic data when supplied with a DEM and a land cover grid theme. Figure 3.14c shows the physiographic data derivation tool that uses ArcView and the Spatial Analyst extension to derive such parameters.

The DEM was used as input for the SLURPView watershed delineation process which is based on a flow accumulation grid. To create an accurate representation of flow direction and accumulated flow, it is best to use a “depressionless” or “sink free” DEM. A sink is a cell or set of spatially connected cells whose flow direction cannot be assigned one of the eight valid values in a flow direction Grid. This can occur when all neighboring cells are higher than the processing cell, or when two cells flow into each other creating a two-cell loop. SLURPView determined that the Taylor DEM contained 207 sinks and the Burntwood DEM contained 1136 sinks. SLURPView was used to fill in all sinks by raising the elevations of those cells.

Once a depressionless DEM was generated, SLURPView was used to create a flow direction and flow accumulation grid for both study basins shown in Figure 3.15 and Figure 3.16 respectively. The flow accumulation grid indicates the number of cells that flow into the grid cell of interest. An outline of all grid cells flowing into the grid cell of

interest forms a subwatershed. A Pour-Points theme was created to specify the location of the outlet for each ASA. SLURPView used the Pour Points theme along with the flow accumulation grid to delineate the ASAs. The location of the pour points was chosen so that each ASA within the basin was of similar size and so that the outlet of certain ASAs coincided with the location of streamflow gauges. A total of 7 ASAs were delineated for the Taylor River watershed and 11 ASAs were delineated for the Upper Burntwood River watershed (see Figure 3.17). Table 3.4 and Table 3.5 provide the areas of each of the ASAs.

3.3.1.3.1 ASA Area

The area of each ASA and the area of each land cover within each ASA was determined using the SLURPView program. In order to compute the ASA areas, SLURPView first creates a series of zones, each representing one of the delineated ASAs in the watershed grid. The “ZonalGeometryTable” command then computes the area of each zone.

SLURPView executed the “TabulateArea” command to determine the land cover fraction of each ASA. The command computes cross-tabulated areas between the land cover grid and the Watershed Grid. Each area is then divided by the total area of the ASA and multiplied by 100 to obtain the percentage of the ASA area occupied by the given land cover class. The area related physiographic parameters for the Taylor and Upper Burntwood River basins are listed in Table 3.4 and Table 3.5 respectively.

3.3.1.3.2 ASA Distance

The SLURP model requires the Average and Standard Deviation To-Stream Flow Distance (DSTS) and Average and Standard Deviation Down-Stream Flow Distance (DSDS) for the area occupied by each of the Land cover class within each of the ASAs as illustrated in Figure 3.18. The parameters were determined using ArcView GIS and ARC/INFO GIS software with the following list of ordered GIS procedures:

1. The SLURPView program was used to create a Flow Direction grid (Figure 3.15) using the “FlowDirection” command. The resulting grid illustrates which of the surrounding eight grid cells the grid cell of interest flows into.
2. A Downstream Total Flow Length grid was created using the ARC/INFO command “FlowLength” which computes flow length of each grid cell to the boundary of the input grid. The function uses the Flow Direction grid and Stream Network Grid (Figure 3.19) as input grids. All cells on the stream network were reclassified to a value of one in order to base the flow length impedance on horizontal distance only. The resulting Downstream Total Flow Length grid is shown in Figure 3.20.
3. A Downstream To Stream Flow Length grid was developed similarly to the previous grid. The difference being that the input Flow Direction grid was modified so that all cells situated on the stream network were reclassified to NO DATA. This results in the “FlowDirection” command terminating its flow length calculations on the NO DATA cells. The resulting Downstream To Stream Flow Length grid is shown in Figure 3.21.
4. The Downstream ASA Outlet Flow Lengths were extracted from the Downstream Total Flow Length grid using the ArcView Identify tool. The values were subtracted from the Downstream Total Flow Length grid in order to determine the downstream flow distance to the ASA outlet.
5. The two flow distance grids were summarized in order to determine the mean and standard deviation of the DSTS and DSDS for the area occupied by each land cover within each ASA. The ARC/INFO commands that were used to determine these parameters are “ZonalMean” and “ZonalStd” which determine the mean and standard deviation of the grid given an input mask. These commands were executed successively using the AML files “DSTS AML” and “DSDS AML” provided in Appendix A. The AML file creates a summary output text file with all of the distance values.

The resulting DSTS and DSDS parameters for the Taylor and Upper Burntwood River watersheds shown are in Table 3.7 and Table 3.9, respectively.

The mean UTM northing and easting of each ASA are required in order to spatially interpolate the time series input data. This was carried out with ArcView using the XYCoord extension downloaded as freeware from the ESRI website (www.esri.com). The extension determines the centroid of all polygons and adds the northing and easting to the attribute table. The resulting coordinates are shown in Table 3.11 and Table 3.12.

3.3.1.3.3 ASA Elevation

The mean elevation of each ASA and the mean elevation of the area occupied by each Land Cover class within each of the ASAs were computed using SLURPView. It uses the “ZonalStatsTable” command for both parameters which computes the mean ASA elevation by summarizing the DEM elevation grid using the Watershed grid as zones. The command computes the mean elevation occupied by each Land Cover by summarizing the DEM elevation grid for each zone that is defined by the Watershed grid and Land Cover grid. The resulting parameters are listed in Table 3.6.

The SLURP model also requires the Average and Standard Deviation To-Stream Change In Elevation (CHELTS) and Average and Standard Deviation Down-Stream Change In Elevation (CHELDS) for the area occupied by each of the land cover classes within each of the ASAs. These elevation parameters are illustrated in Figure 3.18. This was accomplished using ArcView and ARC/INFO using the following list of GIS procedures:

1. The “Derive Slope” command was used in ArcView to create a Slope grid (Figure 3.22). The Slope in degrees from horizontal is computed as the rate of maximum change for each cell to its neighbors, calculated using a 3 by 3 cell window.
2. The Average Stream Slope ($\bar{S}_{\text{STREAM}}$) within each ASA was determined using a mask representing the Stream Network grid. The new grid was created by reclassifying all cells lying on the stream network grid to a value of one and all other cells to the NO DATA value. Finally, the “Summarize Zone” command was used to summarize the Slope grid using the Stream Network mask and the Watershed grid. The resulting $\bar{S}_{\text{STREAM}}$ are shown in Table 3.8 and Table 3.10.

3. The Average Slope of the area occupied by each Land Cover ($\bar{S}_{LC}$) was determined using the ARC/INFO “ZonalMean” and “Zonal Minority” commands. The commands were compiled into an AML script named “Average Slope AML” so that they could be applied for each land cover within each ASA. The AML script also creates a summary output text file with all of the slope values. The AML file is provided in Appendix A. The resulting $\bar{S}_{LC}$ are shown in Table 3.8 and Table 3.10.
4. The elevations at the ASA outlets were extracted from the DEMs using the Identify tool found in ArcView.
5. The total change in elevation to the ASA outlets (CHELTT) was determined by subtracting the ASA outlet elevation from the DEM elevation. The elevations were then summarized using the “Change In Elevation To ASA Outlet AML” file provided in Appendix A. The resulting CHELTT are shown in Table 3.8 and Table 3.10.
6. The $\bar{S}_{STREAM}$ values were multiplied by the DSDS to obtain CHELTS’ which is an initial estimate of CHELTS. The $\bar{S}_{LC}$ was multiplied by the DSTS to obtain the CHELDS’ which is an initial estimate of CHELDS. Finally, both CHELDS and CHELTS were corrected so their sum equaled CHELTT. The CHELDS and CHELTS are shown in Table 3.8 and Table 3.10.

3.3.2 Time Series Data

3.3.2.1 Meteorological Data

Meteorological data collected by Environment Canada is used for this study. The stations where data was collected are listed in Table 3.13 and shown in Figure 3.23. These are all stations that are within 215 km of the study area centroid (UTM coordinates 477,283N 6,131,556E) and contain records between 1984 and 2000. During the initial stages of this research it was expected that remotely sensed data would be used to derive model inputs. Hence, meteorological data prior to 1984 was not utilized due to the lack of remotely sensed data prior to 1984. Additionally, it is shown in Section 3.3.2.3 that the period between 1984 and 2000 is sufficient to represent the wide range of meteorological conditions that occur in the region of interest. Daily average temperature and precipitation time series were utilized. All other parameters were hourly values.

3.3.2.2 Hydrometric Data

Hydrometric data is required in order to calibrate and assess the validity of the SLURP hydrological model output. The Water Survey of Canada (WSC) operates a hydrometric gauge on the Upper Burntwood River named “Burntwood River Above Leaf Rapids”, WSC ID 05TE002. This gauge was installed on August 26, 1985. Manitoba Hydro operates the “Gate Falls” hydrometric gauge located in close proximity to the WSC gauge and was used to fill in data prior to the operation of the WSC gauge for this study in order to create a continuous time series during 1984-2000. It was observed that the Gate Falls discharge measurements are 5-30% larger than the Leaf Rapids discharge measurements for the 1985-1998 period of record. In order to fill in the Leaf Rapids flow record prior to August 26, 1985 the Gate Falls flow is adjusted using a monthly reduction factor shown in Figure 3.24. The monthly reduction factor is determined using the ratio of the Gate Falls discharge to the Leaf Rapids discharge (see Figure 3.25) over the 1985-1998 period.

The WSC also operates a hydrometric gauge on the Taylor River called “Taylor River Near Thompson” WSC ID 05TG002. There are 6 gaps in the Taylor River hydrograph: May 2 - 5, 1988; July 12 - September 27, 1989; August 21 - October 23, 1990; September 31 - December 31, 1991, November 9, 1998 - April 7, 1999, and October 8 - November 7, 1999. Unlike all other test years, the 1999 discharge remains near baseflow well into the summer months. The data is suspected to be erroneous since the Upper Burntwood hydrograph includes substantial snowmelt runoff. Therefore, 1999 will not be used for assessment of the Taylor River modelling results.

The locations of the hydrometric gauges are shown in Figure 3.28. Figure 3.26 and Figure 3.27 provide the hydrographs for the Upper Burntwood River and Taylor River over the 1984-2000 period respectively.

3.3.2.3 Selection of Test Periods

A hydrologic model that is to be used for continuous flow simulation should have the ability to operate under a wide range of meteorological conditions. Hydrologists often choose periods of wet, dry and average levels of precipitation on which to conduct tests. How a year is determined as “wet” must be resolved (Valeo, 1998).

Canadian Climatic Normals published by Environment Canada were referenced for information on levels of precipitation. A Climatic Precipitation Normal is an average precipitation over a 30 year period. Normals were used to classify each year between 1984 and 2000 as being either wet, average, or dry. The precipitation Normals used for this study are for the period of 1961-1990. A total of six precipitation gauges, herein termed as index gauges, contain a nearly complete set of precipitation data between 1984 and 2000 and were used to classify the regional climate and are listed in Table 3.14. No gauge operates within the study boundary. Since the six gauges surround the basins (Figure 3.23), it is assumed that they will provide a reasonable indication of the regional climate in the vicinity of the watersheds. A precipitation Normal for the period of 1961-1990 is not available for the Norway House Airport gauge. Instead, a mean was calculated for the period of 1974-1998, and used as the normal precipitation for this gauge. The climatic precipitation normal for each of the six index stations surrounding the study area are shown in Table 3.14. Based on these six index gauges, the mean annual precipitation is 498.5 mm and the standard deviation is 29.4 mm. This indicates that there is a relatively small variance in precipitation levels in the region hence the six gauges can be used as a reasonable indicator of total precipitation for the Upper Burntwood and Taylor River watersheds.

The majority of the six index gauges are missing precipitation records. In the cases where two gauges are very close to each other, for example the Flin Flon Airport and Flin Flon gauges, the adjacent gauge is used to fill in the missing records. In other cases, the Inverse Distance Method was used to fill the missing data points because “the inverse distance method has been advocated to be the most accurate of all” simple interpolation methods (Singh, 1988). The method is also found to be valid for relatively flat areas

since it does not consider elevation as a factor. The total range of elevation for the six gauges is relatively small at 140 meters therefore, elevation should not influence precipitation measurements. The method is implemented using the following equation:

$$r_x = \frac{\sum_{i=1}^n \frac{r_i}{D_i^b}}{\sum_{i=1}^n \frac{1}{D_i^b}} \quad (3.1)$$

where x denotes the station with the missing precipitation, i is the neighboring stations with known precipitation, r is precipitation, D is the distance between station x and i , b is the weighting constant normally between 1.5 and 2.0. For this exercise a value of 2.0 was used for b and the three closest precipitation recordings were utilized in order to calculate the missing data points.

How a year is classified as wet, average or dry is important. This study maintains that if the precipitation is less than or greater than the climatic normal by a certain number of standard deviations, than it can be considered dry or wet (Valeo, 1998). However, standard deviations are not published with the precipitation normals. Instead, the standard deviations of total annual precipitation (January 1 – December 31) were computed for each year between 1984 and 2000. Total precipitation for each year between 1984 and 2000 were computed and standardized (i.e. $(x - \mu)/\sigma$). The mean standardized precipitation using all six index gauges was computed for each year. This time series is plotted in Figure 3.29. Since most years were between -1 and 1 standard deviations it was decided to further divide this range into slightly dry, average, and slightly wet, in order to improve the classification of each year. From these observations, the author considers 1990, 1992, 1994, and 1998 to be dry, 1987, 1998, 1993, and 1995 are slightly dry, 1984, 1985, 1989, 1991, and 1996 are average, 1986 and 1997 are slightly wet. No year is classified as wet although 1997 has a mean standardized precipitation of 0.90, which is nearly classified as a wet year. The year 1996 will be used to calibrate the models since it is average in comparison to the other years and contains a

complete measured hydrograph that is required for model assessment. The years 1986 and 1997 will be used to validate the models for wet conditions. The years 1992, 1995, and 1998 will be used to validate the models for dry years, and 1991 as an average year verification. The remaining years, 1985, 1987, 1988, 1989, 1990, 1993, 1994, 1999, and 2000 will also be used to assess the model results by computing the performance statistics and inspecting the simulated hydrographs.

The objective of the watershed model calibration and validation process is to accurately reproduce historical stream flow records while maintaining physicality. Precipitation inputs are the most important factor affecting model output. Hence, calibration and verification rely on the assumption that the historical precipitation is correlated with the historical streamflow records. A qualitative assessment was conducted to insure that the historical streamflow and precipitation records are correlated since all meteorological gauges are located outside of the study basins. Figure 3.30 shows a typical chart generated for each year for this assessment. The Taylor River streamflow was scaled up by a factor of 4.5 in order to observe the correlation with the Upper Burntwood streamflow and the precipitation levels. Precipitation recorded at the Thompson Airport and the Flin Flon Airport were included since they are the closest gauges to the basins, they are positioned east and west of the watersheds, and they have complete records. It is expected that the Taylor River flows will show a stronger correlation with precipitation recorded in Thompson since the watershed is significantly closer to this gauge. The same relationship is expected between the Upper Burntwood River flows and the precipitation recorded at the Flin Flon Airport.

All charts indicate that both watersheds show little response to precipitation for the first few months and last few months of the year since the majority of precipitation occurs as snow, thus accumulating into snow storage. In general, the flows for both basins show a strong response to the observed precipitation at the Thompson Airport and Flin Flon Airport precipitation gauges. The Taylor River watershed shows a stronger correlation to the Thompson Airport precipitation gauge, as does the Upper Burntwood River watershed to the Flin Flon Airport precipitation gauge. It should be noted that the

precipitation at the Thompson gauge and the Flin Flon gauge are very similar during the days with large amounts of precipitation and often with a time lag between the two gauges. However, there are a greater number of instances where the Upper Burntwood River watershed shows a response to precipitation measurements at the Thompson gauge than does the Taylor River watershed to the Flin Flon gauge. This is due to the small size of the Taylor River watershed and its close proximity to Thompson. On the other hand, the Upper Burntwood covers a much larger area and its centroid is located near the centroid of the Flin Flon and Thompson gauges. Hence, it is influenced similarly by both gauges. This qualitative assessment has shown that the historical streamflow and precipitation records are correlated and therefore, modelling of the rainfall – runoff process using this data is feasible.

3.3.2.4 Time Series Model Input Files

The SLURP model reads the time series variable data into the model from a series of text files, each text file representing a different time series variable and ASA. Using the three closest gauges spatial interpolation was conducted using the Inverse Distance Method with a weighting factor of 2.0 at the centroid of each ASA and for each day. All other variables were spatially interpolated using the mean of the three closest gauges that have data for the day of interest. Visual basic macros were created within Microsoft Excel in order to spatially interpolate and generate the time series text files.

3.3.3 Parametric Data

3.3.3.1 Routing Parameters

The SLURP hydrological model requires both stream width and depth in order to facilitate the within ASA river routing algorithms. A simple method was used to estimate these parameters since the sensitivity analysis showed that these parameters show very little sensitivity. The Strahler Stream Ordering system was used to estimate the stream width and depth (Singh, 1988). A depth of 5 m and width of 40 m was measured from the Taylor River cross section plot at the Taylor River gauge. That reach of river corresponds to a Strahler Order of 4. The width of the river reaches with a Strahler Order

of 3 or higher was measured directly using the FRI data. Widths of 5 m and 10 m were assumed for Strahler Orders of 1 and 2 respectively. Using the Taylor River gauge section a Depth/Width ratio of 0.125 was computed. This ratio is used to establish the stream width for the lower Strahler Order streams and the depths for all Strahler Stream Orders. Two of the Upper Burntwood River ASAs could not be classified using the Strahler Stream Order because they are comprised mainly of Lake. In these cases, a width of 300m and depth of 10m is assumed. Finally, the average Strahler Order for each ASA was estimated from the 1:50,000 NTS mapping. Table 3.15 and Table 3.16 present the average Strahler Order, width, and depth for each study basin.

The second set of routing parameters required by the SLURP hydrological model are those that support the between ASA routing algorithms. Since the rivers of interest are relatively sinuous, have an abundance of shoreline vegetation, and contain numerous small and large lakes as well as marshes, there is certainly some attenuation of the flood hydrograph. Hence, the “no attenuation” option for river routing will not be used. As pointed out in the previous paragraph, there is a general lack of data that details the river geometry, hence the Muskingum routing method is preferred here rather than the Muskingum Kunge routing method, which requires detailed river geometry. Linsley (1982) indicates that most natural streams and rivers that have mild slopes, well defined channels, and streamflow that is confined within the river banks have a weighting constant, X , between 0.1 and 0.3. Since the Taylor and Upper Burntwood are rather sinuous and have very mild slopes, an X of 0.3 will be used for ASAs without significant lake storage. Those ASAs that have significant lake storage will be assigned an X of 0.2. The time of travel constant K is initially assumed to be 1 day for all ASAs. Both K and X will be adjusted during the calibration process to increase the accuracy of the river routing. Table 3.15 lists the initial estimates of the river routing parameters.

3.3.3.2 LAI Parameters

The SLURP model requires an LAI value for each month of the year for each land cover. A time series of LAI based on field data could not be found within the literature for the types of land covers that resemble those found in the study area. LAI are often derived

from Normalized Difference Vegetation Index (NDVI) maps for many large scale hydrological model applications (Kite, 2002). Yin and Williams (1997) assumed a linear relationship between LAI and NDVI of the form:

$$LAI_i = LAI_{MAX} * \left(\frac{NDVI_i - NDVI_{MIN}}{NDVI_{MAX} - NDVI_{MIN}} \right) \quad (3.2)$$

where **i** indicates the land cover, **LAI_{MAX}** is the maximum LAI for land cover **i** and **max** and **min** indicate the maximum and minimum values of NDVI measured in a series of images for that land cover.

A North America NDVI map for each month of the year was downloaded from the USGS website (USGS, 2002). These maps have a 1 km resolution and were created using a series of AVHRR scenes. Files were downloaded as ERDAS Imagine files in Byte format. The 12 files were converted to ARC/INFO Grids using PCI image processing software, reprojected from Lambert Azimuthal Equal Area to UTM NAD 83 Zone 14 projection, and clipped to include only the portion of the scene in Manitoba. The files were converted to ASCII grid files and imported into ArcView. The NDVI Byte values corresponding to each respective land cover were extracted using ArcView. The average NDVI for each cover and month was computed and summarized in Table 3.17.

The maximum LAI values used in Equation 3.2 are normally obtained by direct measurement or from literature (Kite, 2002). The following section lists the methodology used to establish LAI_{MAX} for the various types of land cover modelled in this study:

- A. Coniferous – BOREAS researchers measured maximum LAI values at the NSA ranging between 1.9-4.3. A value of 4.3 is assumed (Newcommer et al., 2000; Nickeson et al., 2002).
- B. Mixed Deciduous & Coniferous – BOREAS researchers measured maximum LAI values at the SSA ranging between 2.0-7.8 with an average of 7.8. A value of 7.8 is assumed (Newcommer et al., 2000; Nickeson et al., 2002).

- C. Deciduous – An LAI of 8.0 is recommended by Kite (2002) for high latitude deciduous forest.
- D. Muskeg – No LAI for Muskeg was found in literature. This cover contains a large number of Spruce trees but is less dense than a Coniferous forest. A value of 3.5 is assumed since this is similar to the Coniferous LAI.
- E. Impervious – Kite and Spence (1994) measured LAI of 2.0 for these two covers at Wolf Creek, Yukon. The Impervious cover is assumed to have similar vegetation characteristics as Tundra and grassland therefore an LAI of 2.0 is assumed.
- F. Lake/Rivers – It is assumed that water bodies such as lakes and rivers do not contain vegetation, hence an LAI of 0.0 is assumed.
- G. Marsh/Beaver Flood – It is assumed that the LAI for this cover is less than the Muskeg LAI but greater than the Lakes/Rivers LAI. An intermediate value of 2.0 is assumed.

Price et al. (1997) found that the evaporation from the forest floor made up a very large component of the total evaporation in the BOREAS NSA, roughly 25%. They also found that the moss layer also greatly contributed to the interception of precipitation. In order to account for the effects of the moss layer on ET and interception, the LAI_{MAX} was increased by 25% for the Coniferous, Mixed, Deciduous, and Muskeg land covers. Table 3.18 lists the final LAI_{MAX} values. Using Equation 3.2 a time series of LAI for each type of land cover was computed and summarized in Figure 3.31.

3.3.3.3 Canopy Coefficients

SLURP requires three parameters to represent the canopy processes: maximum canopy capacity at maximum LAI, and the interception coefficients A and B. Since water has no canopy, all three parameters are set to zero for this cover. Price et al. (1997) found that moss interception was nearly as much as forest interception and found a definitive moss interception capacity in the range of 15 mm to 47 mm. According to Price et al. some of the intercepted precipitation does move down to the soil layer after a few days. This suggests that the 15 mm - 47 mm range should be reduced if it is to be used in SLURP as it attempts to evaporate all intercepted precipitation. Other studies (Grelle et al., 1997;

Kite, 1998a) have found the coniferous canopy capacities to be less than the 15 mm-47 mm range reported by Price et al (1997). A conservative moss capacity of 15 mm is chosen for the forest covers and 0 mm for the Impervious and Marsh covers due to the general lack of moss in these covers. Since Price et al (1997) demonstrated that the moss does contribute appreciably to ET, the composite canopy capacity will comprise both the canopy and moss capacities for this model study. This results in a composite canopy capacity of 30 mm for the Coniferous Treed cover at maximum LAI. Using the maximum LAI, the canopy capacities for the other land covers are determined using the Coniferous canopy capacity as an index:

$$CC_{MAX,i} = \left(\frac{LAI_{MAX,j}}{LAI_{MAX,Coniferous}} \right) * CC_{MAX,Coniferous} \quad (3.3)$$

where *i* denotes the land cover of interest and $LAI_{MAX,i}$ is the maximum LAI for the canopy and moss combined.

SLURP requires the Canopy Capacity per unit LAI (CC). This was computed using the following relation:

$$CC_i = \frac{CC_{MAX,i}}{LAI_{MAX,i}} \quad (3.4)$$

The resulting estimates of CC are shown in Table 3.19.

Kite (2002) provides canopy parameters on the SLURP 12.2 CD for six different basins located around the world. Due to the relative insensitivity and non-physical basis of the interception coefficients A and B, Kite has used the same parameter values in all modelling studies. These values are A=0.25 for Tundra and Shrubland, and A=0.5 for all other types of cover. A value of B=1 was used for all types of cover. Similarly for this study, B=1 is used for all covers, and A=0.5 for all covers except the Impervious cover where A=0.25 will be used.

3.3.3.4 Snowmelt Rates

Price and Hendrie (1983) measured maximum snowmelt rates between 2.9-5.3 mm/hr in a mixed forest near Ottawa, Ontario. Metcalfe and Buttle (2001) observed maximum snowmelt rates in excess of 10 mm/day at the BOREAS NSA. Also reported are snowmelt rates used in the SLURP applications to the Gediz, Kootenay, Wolf, Olifants, Mekong, and Menderes basins (Kite, 2002). These applications used a snowmelt rate of 1 mm/day/°C for January and 1.7-2.2 mm/day/°C for July. All models except Kootenay used a July snowmelt rate of 2.0 mm/day/°C. Metcalfe and Buttle (2001) provide sufficient information to allow for the calculation of the snowmelt rates [mm/day/°C] during the 1994 snowmelt period at the BOREAS NSA. They provide mean melt rates for open wetland (8 mm/day), wetland with willow (7 mm/day), open forest (5.5 mm/day), and closed forest (3.5 mm/day), melt period (April 11 – May 5), and number of melt days (14 days). The melt rates [mm/day/°C] were back calculated by dividing the average melt rate [mm/day] during the melt period by the average number of degree days during the melt period [°C] using the daily average temperatures recorded by BOREAS at the NSA. The result is average melt rates of 4.60, 2.90, 1.69, and 0.97 mm/day/°C for the open wetland, wetland with willow, open forest, and closed forest covers respectively. These values are assumed to be acceptable in magnitude and will be used for the initial estimates during modelling (see Table 3.20). However, since this parameter is one of the most sensitive parameters in SLURP and largely controls the timing of the spring freshet, it will be necessary to calibrate this parameter in order to best simulate the spring freshet.

Literature provides enough information to characterize how the snowmelt rates should vary among the different types of land cover. Metcalfe and Buttle (2001) observed snowmelt patterns at the BOREAS NSA and found that the snow first melted in wetlands followed by patches of wetland with willow and that these areas were entirely free of snow before the appearance of any bare ground in forested areas. This snowmelt pattern reaffirms the initial estimates for snowmelt rates.

3.3.3.5 Albedo

Values of summer, maximum snow, and minimum snow albedo were chosen using values found in the literature and assigned based on similarity of vegetation structure.

Hetherington (1987) provides a typical albedo of 0.130 for northern Black Spruce forests and 0.160 for deciduous forests. Granger and Pomeroy (1997) measured summer time albedo at the BOREAS SSA and reported values of 0.145 for mixed forests, and 0.152 for clear cut areas. For this study, the latter value is assumed to be representative for the Impervious and Marsh summer time albedos. No reports of Muskeg albedo were found in the literature. It is assumed to be 0.140 which is between the Coniferous and clear cut albedos. No summer albedo is required for the Lakes land class.

Leavesley (1990) states that “the snow albedo diminishes with time as the snow metamorphoses and accumulates deposits of impurities such as dust and vegetative matter”. Therefore, one could then expect a low snow cover albedo to occur near the end of winter and a high snow albedo to occur near the beginning of winter or immediately after fresh snowfall. Metcalfe and Buttle (1999) measured an areal average end of winter albedo equal to 0.370 at the BOREAS NSA. This value is assumed to represent the minimum albedo for all land covers in this study since it is an areal averaged estimate and no other end of winter albedo was found in literature. Hetherington (1987) provides an albedo estimate of 0.780 for snow covered Tundra. This value is assumed to represent the maximum snow albedo for the Impervious, Marsh, and Lakes land covers due to similarities in vegetation composition. It is assumed that the maximum snow albedo for the forest will be less than that of open areas. A value of 0.500 is assumed for the maximum snow albedo of the forest covers. The final summer and snow albedo estimates for each land cover are shown in Table 3.21.

3.3.3.6 Soil Properties

Estimates of Field Capacity and Wilting Point are based on a typical soil profile for the study area. This soil profile, generalized by Metcalfe and Buttle (2001), consists of a peat layer overtop of a silty-clay layer. The ASCE Hydrology Handbook (1996) specifies

an average Field Capacity of 0.36 and Wilting Point of 0.21 for silty-clay soils. The USDA specifies a Field Capacity of 0.42 and Wilting Point of 0.26 for the same soils (Saxton, 2002). Since no field data describing Field Capacity and Wilting Point in the vicinity of the study area could be located for use in this study, an average of these two sources will be used. This study uses a Field Capacity of 0.39 and a Wilting Point of 0.23 for all land covers. Lack of data does not permit the differentiation of Field Capacity and Wilting Point among the various types of land covers.

3.3.3.7 ET Parameters

Barr et al. (1997) evaluated the three ET methods available in the SLURP model by applying the model to the Kootenay Basin over a five year period. Although they found that each method affected the streamflow differently, they recommended the Spittlehouse Black method for ET computations. They found that the Spittlehouse Black method gave the most plausible values and attributed the success of this method to its ability to simulate a gradual decline in ET with diminishing soil water. This feature helped the Spittlehouse method to simulate the seasonal variations in both ET and stream flow correctly. Additionally, this method gave a more realistic estimate of the soil water limitation to ET because it incorporates a sensitive soil water feedback to ET. Finally, the Spittlehouse method provides a better physical representation of ET processes as opposed to the CRAE and the Granger methods because it incorporates a soil water limitation to ET. This is important in a Muskeg dominated environment where ET is sharply reduced once the soil moisture falls below a certain level (Buttle et al., 2000; Goode et al., 1977; Phersson & Pettersson, 1997). Hence, the Spittlehouse method of estimating ET is used for this study.

The Spittlehouse ET method requires Priestly Taylor Coefficients for Water (α_w), Soil (α_s) as well as an empirical constant β . Spittlehouse (1989) gives values for α of 1.26 for crop/grassland and 0.80 for B.C. coniferous forest. Spittlehouse reported values of β that ranged from 10 mm/day for Douglas Fir to 34 mm/day for pasture. If no data exists, Spittlehouse recommends estimating β using the assumption that soil moisture begins to limit ET once the soil store fraction = 0.30 and $ET/\alpha = 4.0$ mm/day. This assumption

may be expressed as $\beta/\alpha = 13$ mm/day. This results in values of β equal to 10 mm/day when $\alpha = 0.80$ and β equal to 16 mm/day when $\alpha = 1.26$. For this study, the first set of values will be used for the forest covers, and the second set of values will be used for the Impervious and Marsh covers. The latter values are adequate since these two covers occupy only a small fraction of the study area.

3.3.3.8 Subsurface Parameters

3.3.3.8.1 Fast & Slow Storage Capacities & Retention Constants

The retention constants and storage capacities of the fast and slow storage tanks are among the most sensitive parameters in the model. Upon reviewing various applications of SLURP, it was observed that all of these parameters vary greatly, often by orders of magnitude among the different geographic locations and types of land cover.

The capacities of the Fast and Slow storage tanks can be estimated based on field data. These values in turn can be used as initial model values and also provides insight to the relative sizes of the storage tanks amongst the different types of cover. Metcalf and Buttle (2001) measured the organic layer thickness, mean porosity of the organic layer (0.813), and mean porosity of the silty-clay layer (0.314) at various locations during the BOREAS studies. Hetherington (1987) states that “most of Canada has been glaciated, which has resulted in extensive wetlands in the boreal forest region and shallow soils, often less than 1 m deep, in most areas presently occupied by forests”. However, it was shown in the Wuskwatim EIS that the depth to bedrock can be as much as 22 m in the vicinity of Wuskwatim Lake (Manitoba Hydro, 2002). To simplify this exercise it is assumed that the upper meter of soil represents the fast storage tank. The initial estimate for the slow storage tank will equal that of that of the fast storage tank. In a simplified manner, the mean porosity for the organic and silty clay layers were applied to the estimated soil thickness for each cover, listed in Table 3.22, to estimate storage capacities.

The following general comments can be made regarding the relative Fast Storage Capacities:

- The Marsh cover, due to the presence of a thick organic layer, has a fast storage capacity that is nearly twice as large as the forest covers and four times as large as the impervious cover.
- The Impervious cover was limited to 0.5 m in thickness. This results in a low storage capacity that is roughly half the size of the forest covers.
- The Storage Capacity of all four forest covers are within 20% of each other. The maximum storage occurs in the Muskeg cover and the minimum storage occurs in the Mixed and Deciduous covers.

The model default values are used as initial estimates of the retention constants for the Fast and Slow Storage tanks. These parameters will not be estimated as they are basin and cover dependent and will certainly change through calibration. However, based on the literature review, one can deduce the relative magnitudes of the retention constants for the different types of cover. These deductions will be used during model calibration to maintain the relative magnitudes of the storage capacities and retention constants.

Due to the inability of many types of Muskeg to maintain an elevated water level, one can presume that the retention constant for the Fast storage tank of Muskeg will be significantly smaller than that of the forest covers. However, once the water table reaches a low level, the water retention increases significantly. This is characteristic of a non-linear storage tank, a feature that SLURP does not accommodate. Since only a single retention constant is used for each tank, using a larger retention constant for the slow storage tank may aid in simulating the reduced interflow that may otherwise take place in the upper soil zone. It can also be deduced that the Marsh cover will have a very low retention constant relative to the other covers due to the very thick (~2 m) highly porous organic layer.

3.3.3.8.2 Initial Slow Storage Contents

The initial content of the slow storage tank is a difficult parameter to initialize because it depends entirely on the antecedent moisture conditions leading up to the period of interest. All model runs will be preceded by an additional year of simulation time in order to initialize or "Spin Up" the slow storage contents.

3.3.3.8.3 Maximum Saturated Infiltration Rate

The SLURP model requires a saturated infiltration rate in a unit of mm/day for each type of land cover. Metcalf and Buttle (2001) found that the peat thickness ranged between 0.07-0.30 m in forested areas and 0.3-2.0 m in the flooded areas of the BOREAS NSA. In the forested soil profiles they measured the saturated hydraulic conductivity (K_{sat}) to range between 2.0×10^{-3} m/s ($\sim 200,000$ mm/day) at the surface to 2.0×10^{-7} m/s (~ 17 mm/day) at the silty clay interface. In the flooded area soil profiles, they found that the K_{sat} was too rapid to measure and near zero at the silty clay interface. This translates to a nearly limitless infiltration rate into the organic layers in both the forested and wetland environments. In general, the Maximum Saturated Infiltration Rate will be maintained at a relatively high value so as not to create overland runoff under normal conditions. A value of 100 mm/day is used for all types of land covers.

3.4 Chapter Summary

The study area consists of the Upper Burntwood River and Taylor River watersheds. Both basins are located in the High Boreal Forest of northern Manitoba. A 100 m resolution digital elevation model was created for the study area using a GIS. High resolution land cover data was obtained from the Forest Resource Inventory of Manitoba and was reclassified into seven land cover classes of which each will be modelled using SLURP. Both ARC/INFO and SLURPView were used to delineate the Aggregated Simulation Areas and to derive all physiographic parameters.

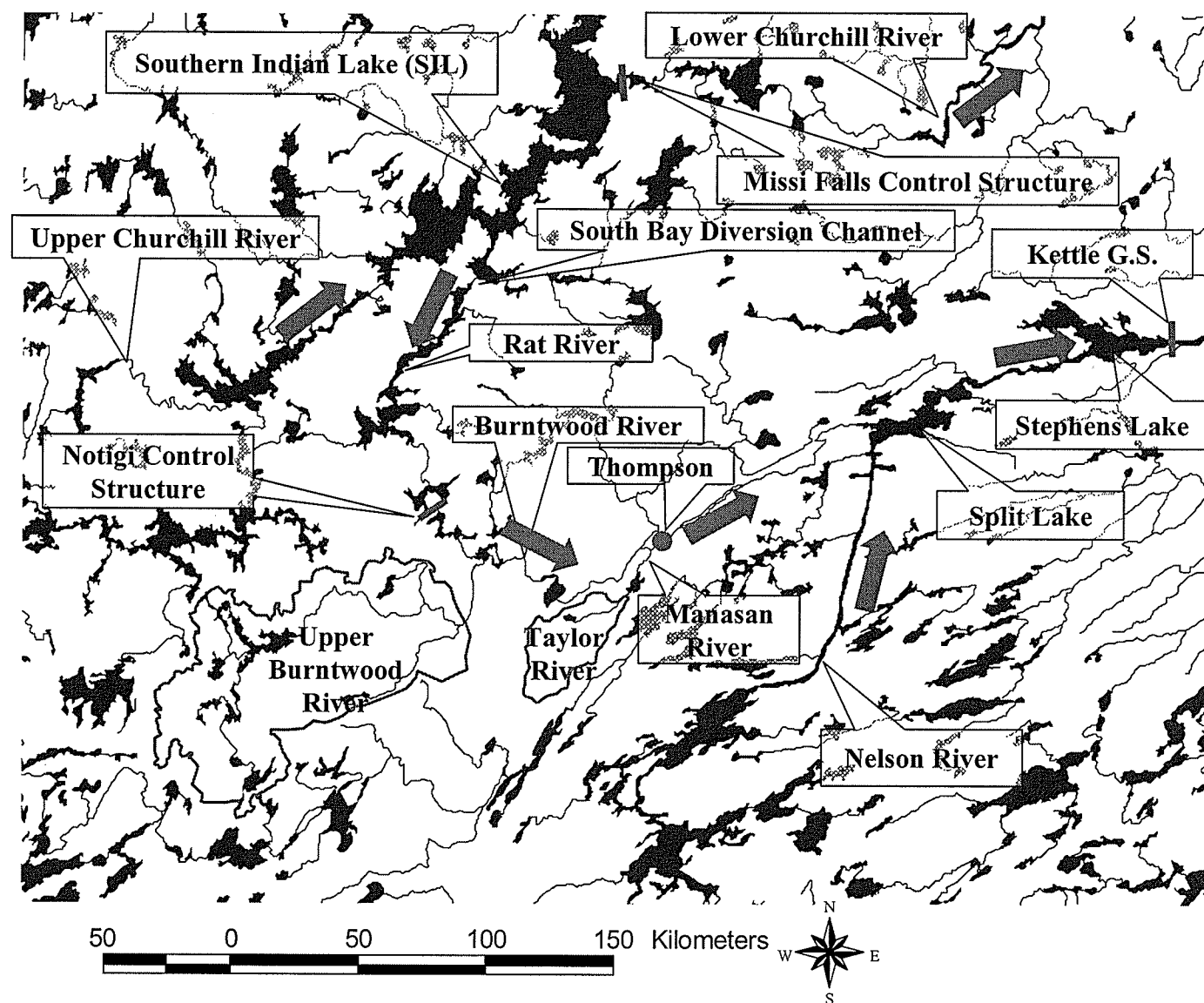
Meteorological inputs for each basin ASA were interpolated from the surrounding Environment Canada meteorological gauges. A total of 15 years from 1985-2000 were

chosen for model calibration and validation. The 15 year period contains a wide range of meteorological conditions however, it is a relatively dry period since only two years are slightly wet and no years are wet. The year 1996 will be used for model calibration and all other years for model validation.

Each study basin contains a Water Survey of Canada streamflow gauge near the outlet of each basin which will be used for model calibration and validation over the 1985-2000 period.

A preliminary database of model parameters was estimated based on values found in the literature.

Figure 3.1 –
Churchill River
Diversion.



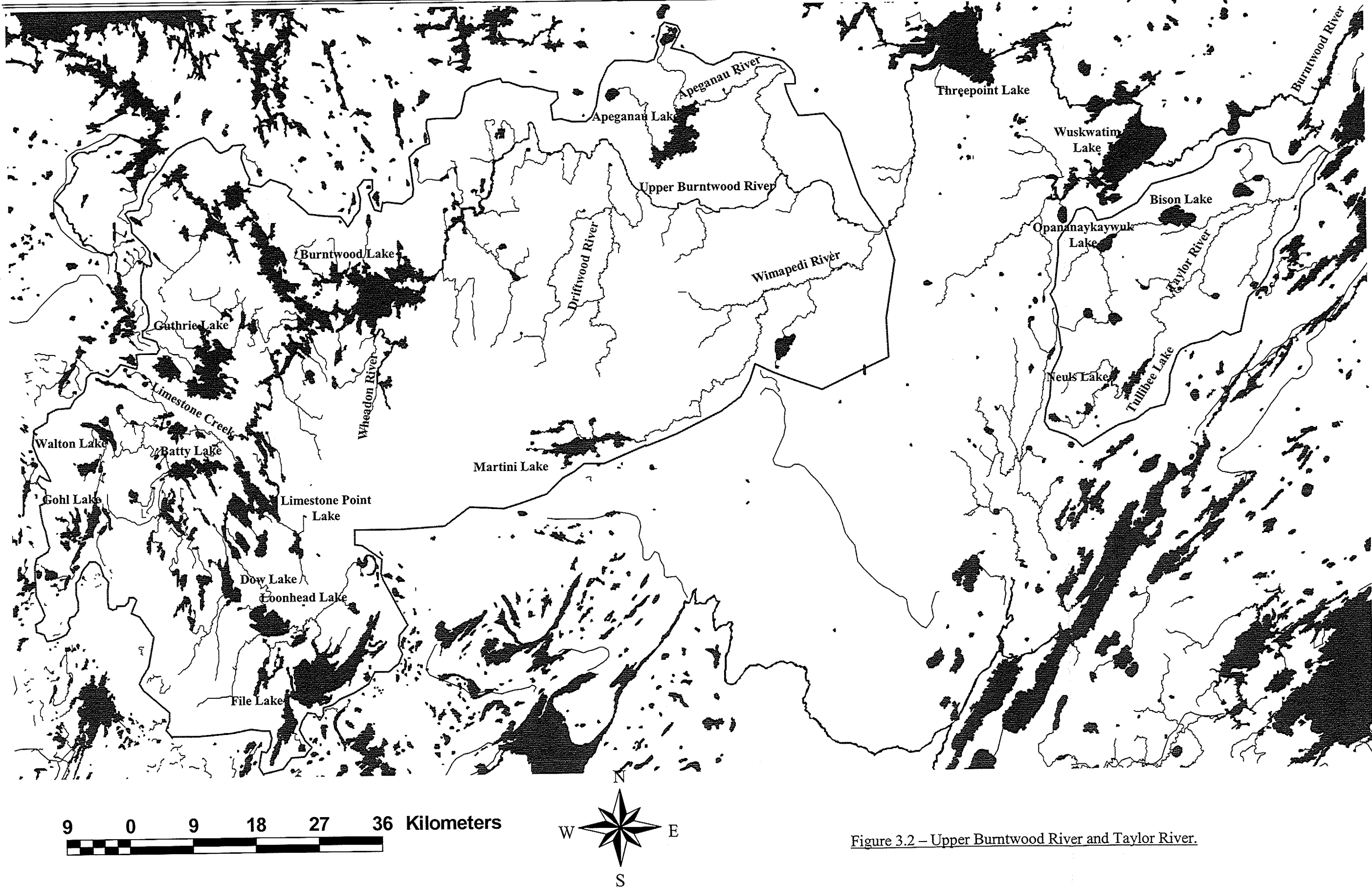


Figure 3.2 – Upper Burntwood River and Taylor River.

Figure 3.3 - Coverage of 1:50,000 NTS Maps used to create DEMs.

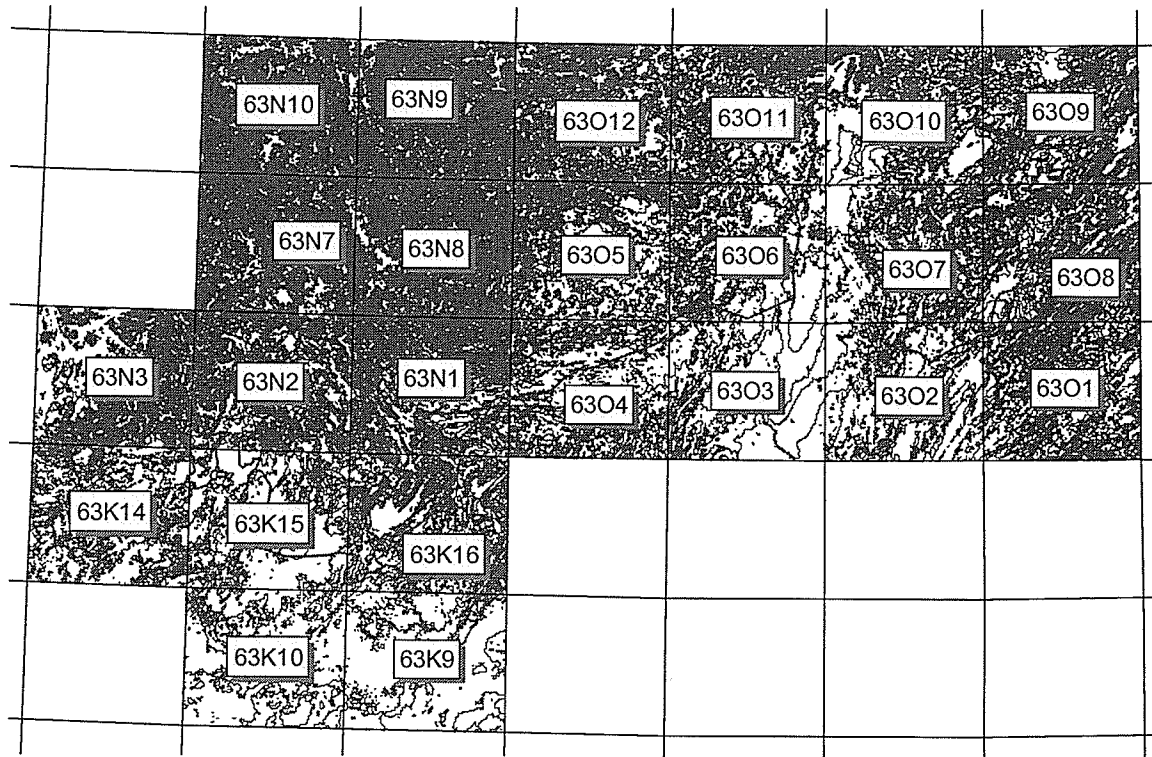


Figure 3.4 - TOPOGRID input data: contours, point elevations, lakes, and rivers.

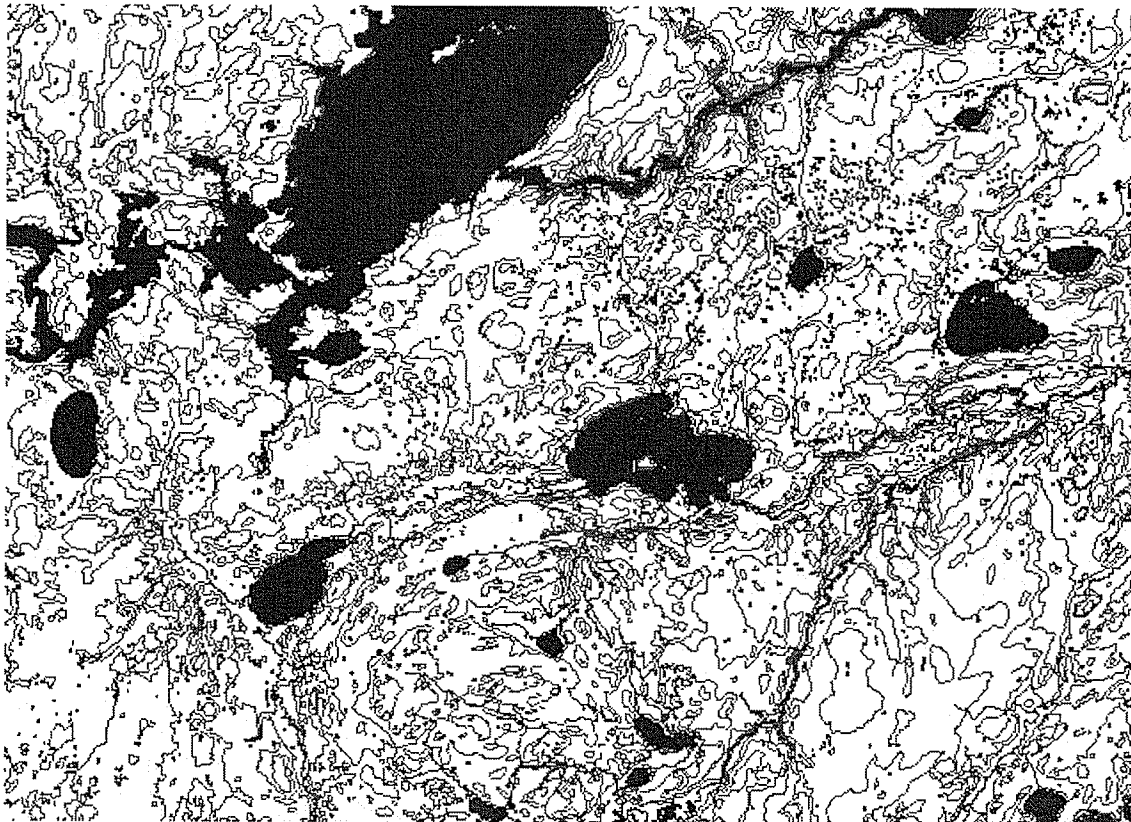


Figure 3.5 – Hypsography coverage for Taylor River and Upper Burntwood River.

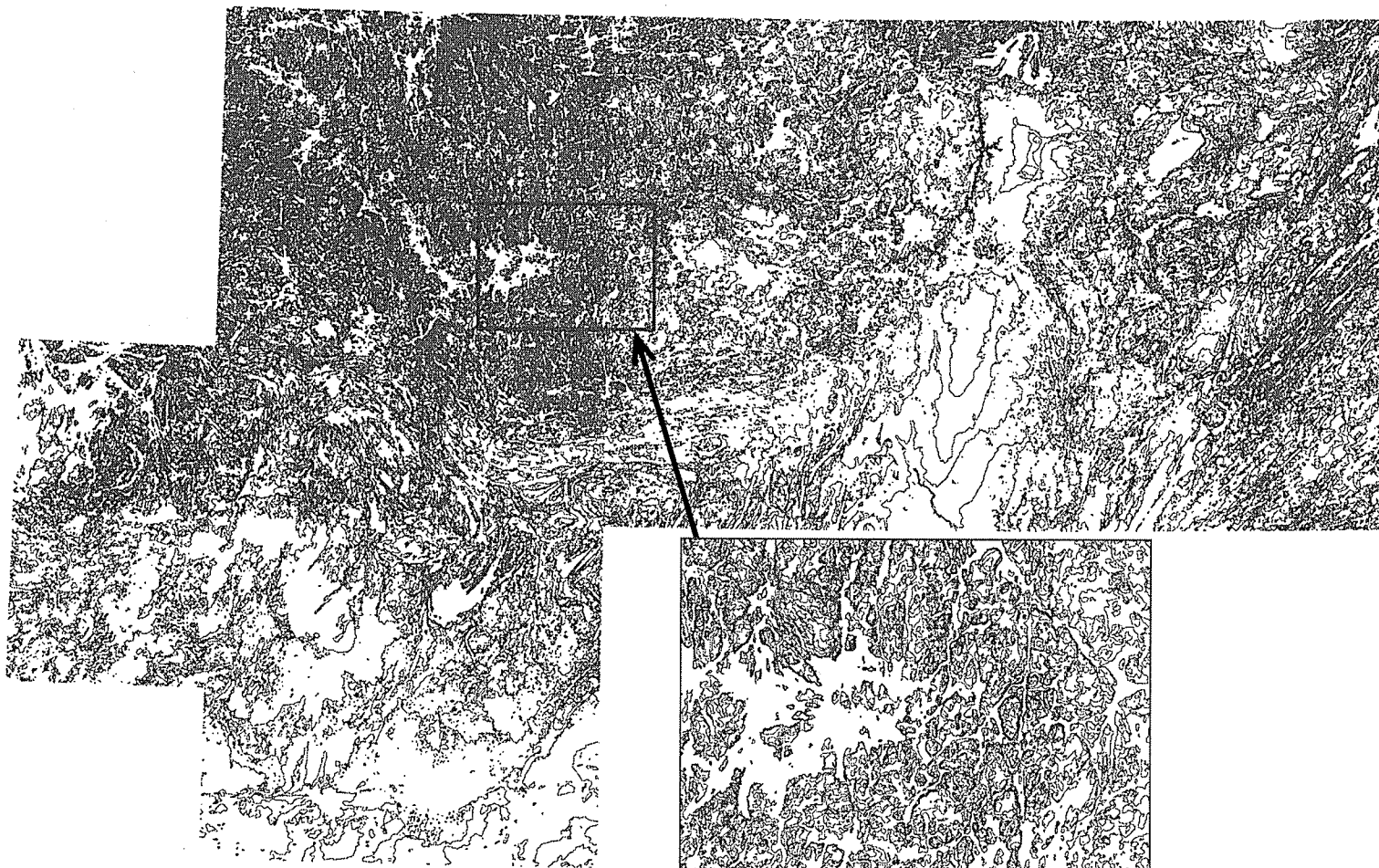


Figure 3.6 – Forest Resources Inventory lakes coverage for Taylor River and Upper Burntwood River.



Figure 3.7 – Sample stream network with corrected topology.

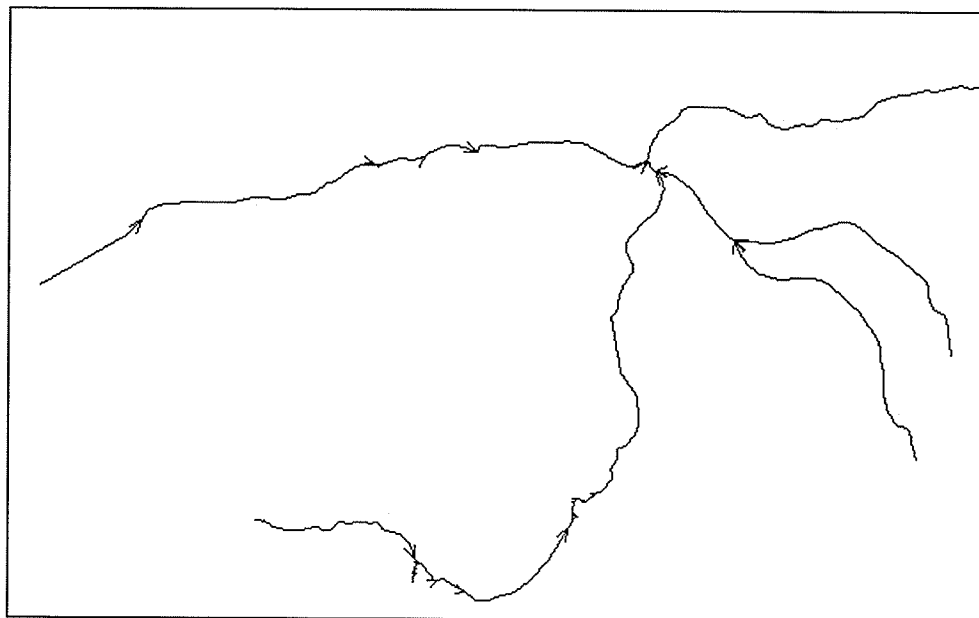
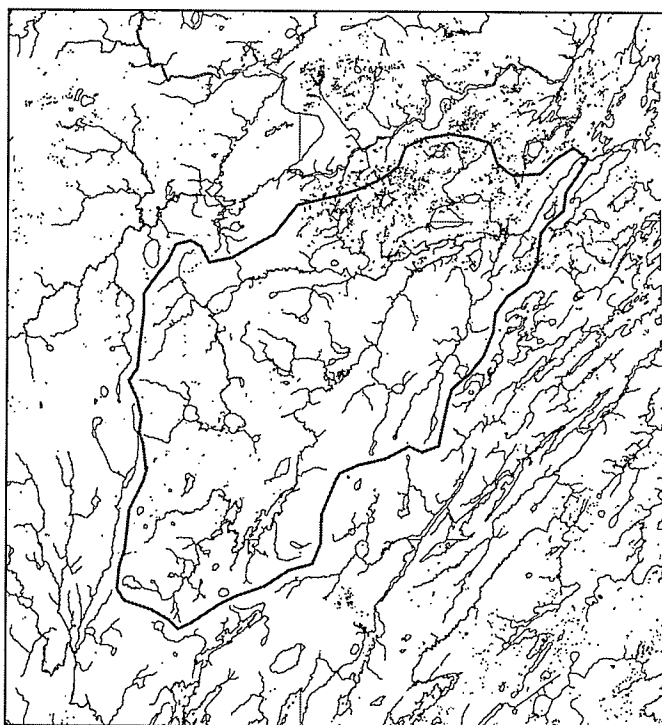
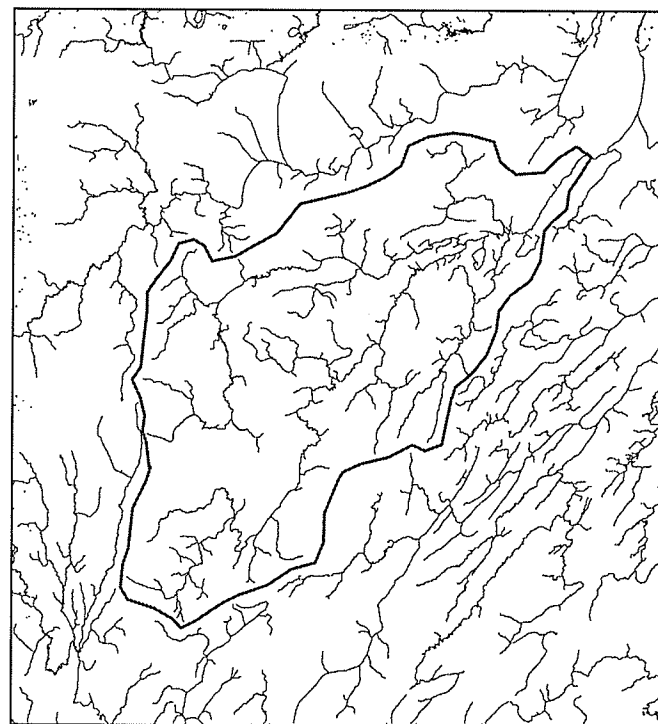


Figure 3.8 - Taylor River: NTS hydrography vs. Topologically Corrected stream network.

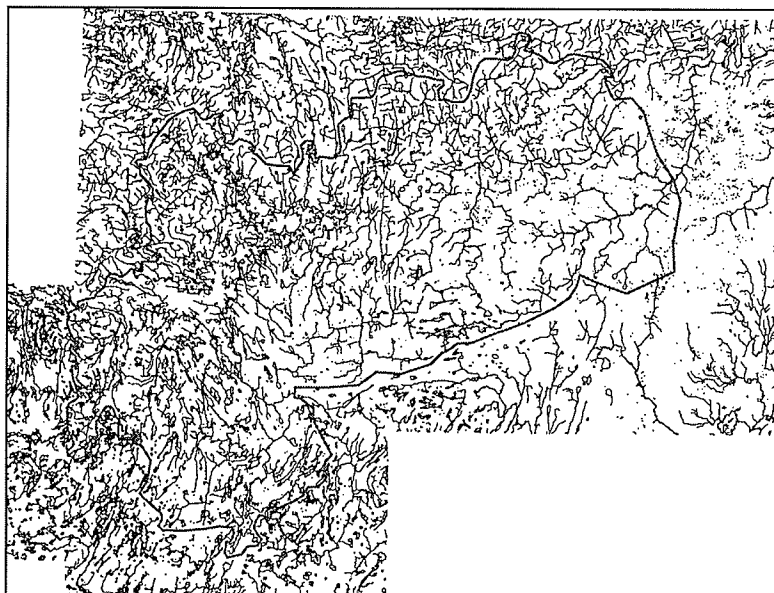


NTS Hydrography

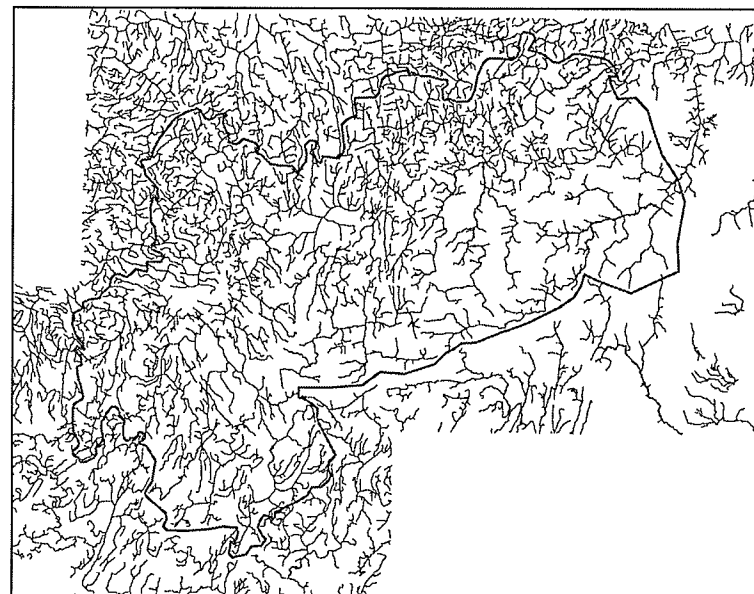


Topologically Corrected Stream Network

Figure 3.9 – Upper Burntwood River: NTS hydrography vs. Topologically Corrected stream network.



NTS Hydrography



Topologically Corrected Stream Network

Figure 3.10 – Taylor River digital elevation model.

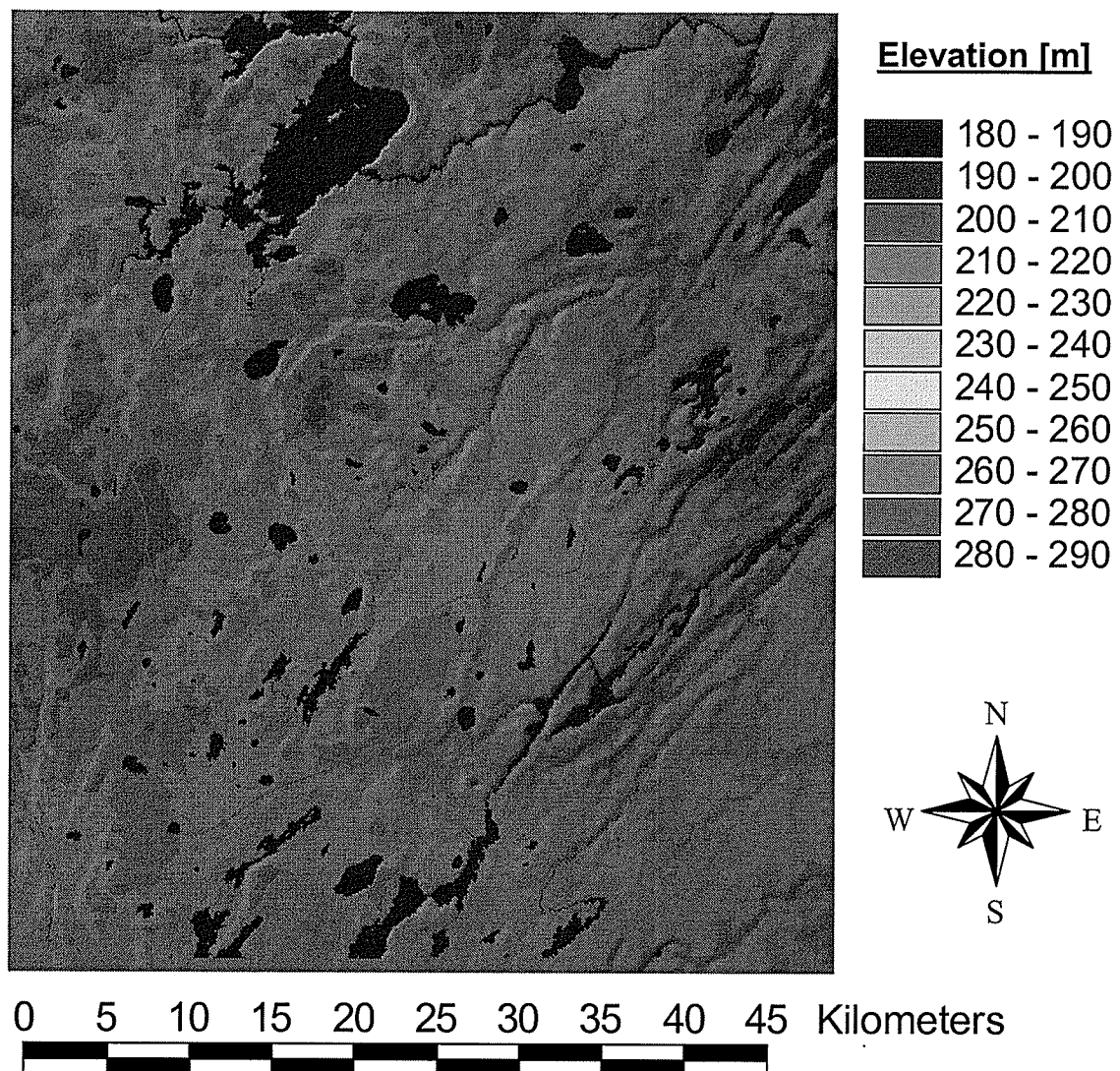


Figure 3.11 – Upper Burntwood River digital elevation model.

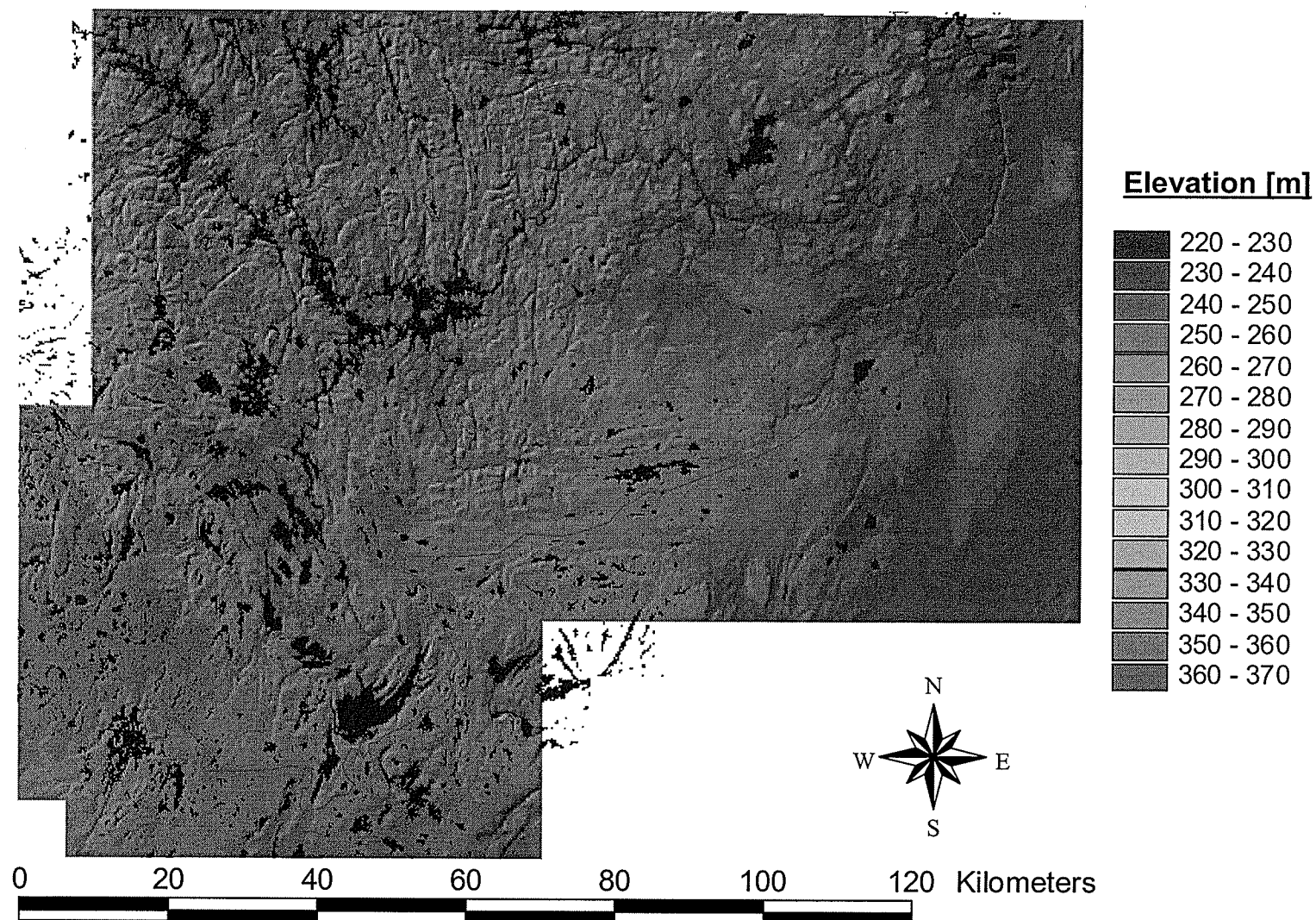


Figure 3.12 – Verification of DEM using 1:50,000 scale NTS contours.

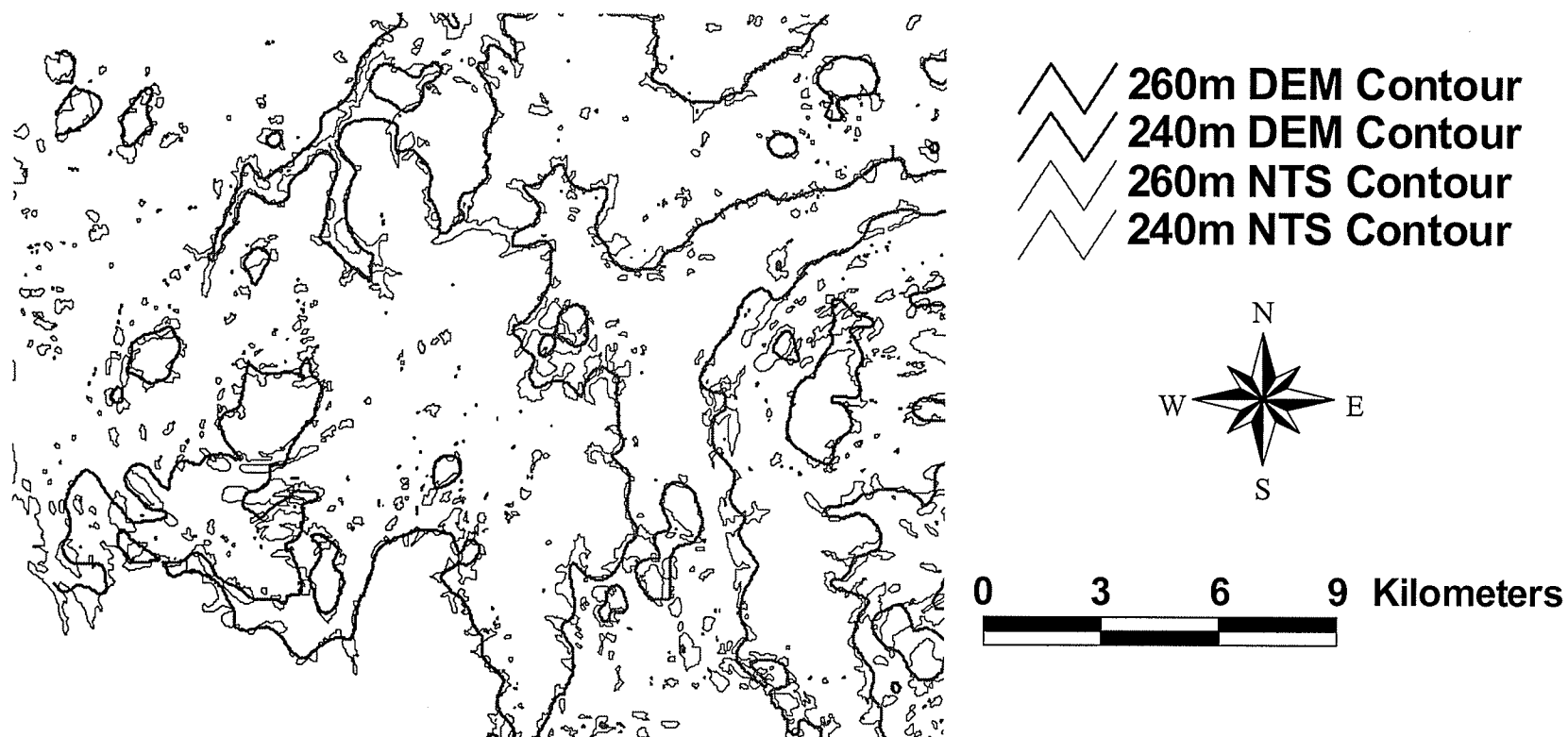


Figure 3.13 – Final reclassification of the Forest Resources Inventory coverage.

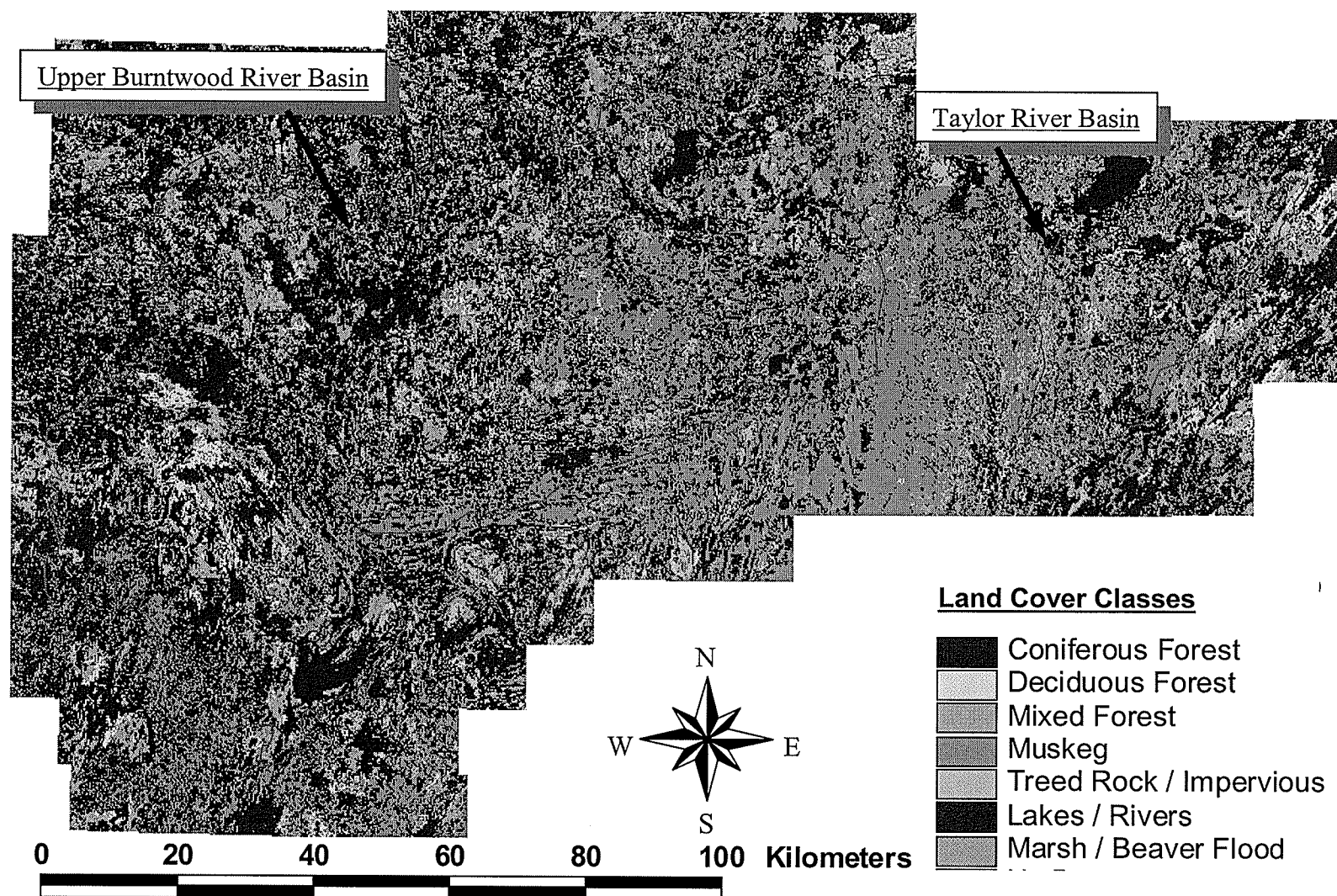
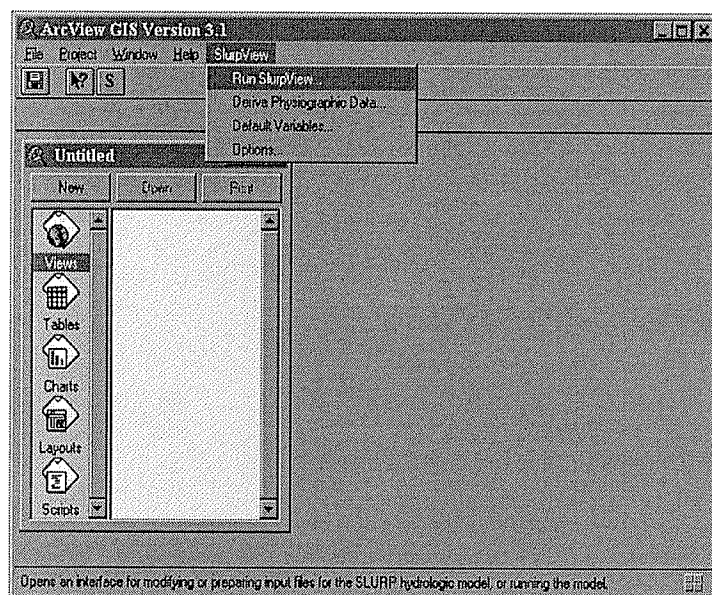
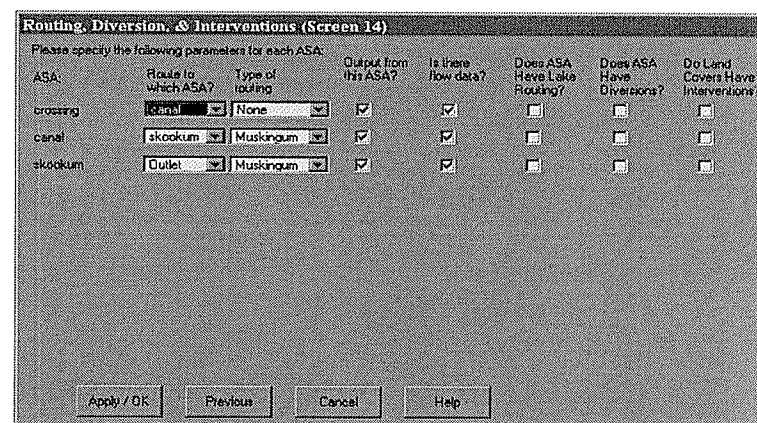


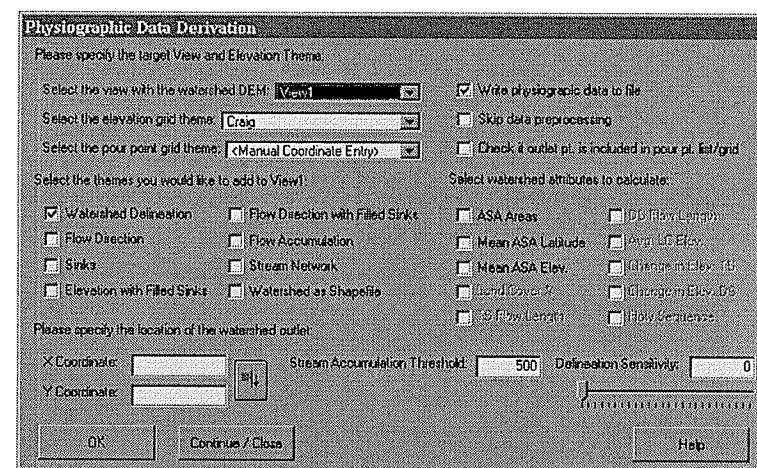
Figure 3.14 – SLURPView Menus.



(a) Running the SLURP model as an extension to ARCVIEW.



(b) A SLURPView parameter entry screen.



(c) SLURPView's physiographic data derivation tool.

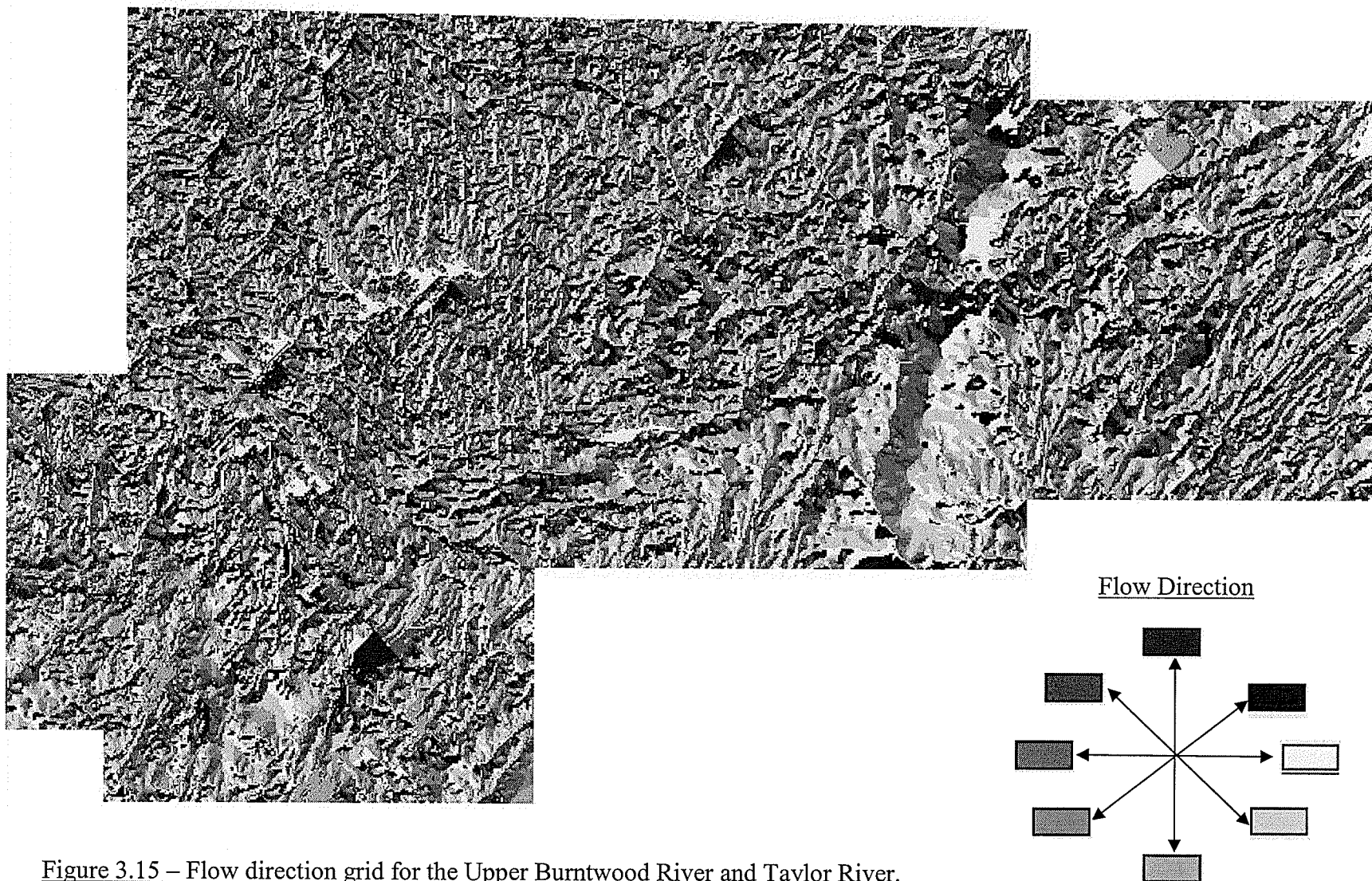


Figure 3.15 – Flow direction grid for the Upper Burntwood River and Taylor River.

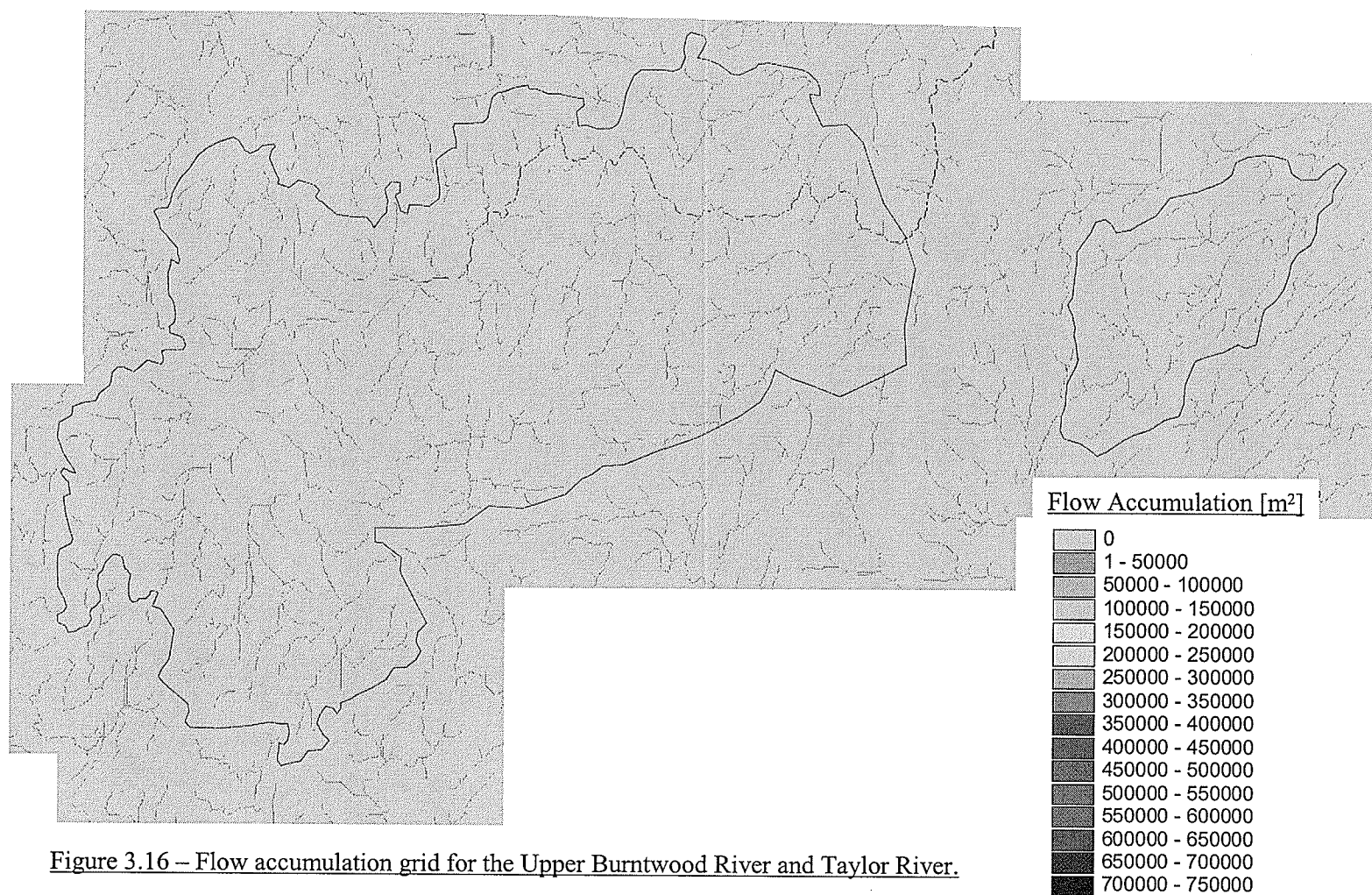


Figure 3.16 – Flow accumulation grid for the Upper Burntwood River and Taylor River.

Figure 3.17 – Upper Burntwood River and Taylor River ASAs.

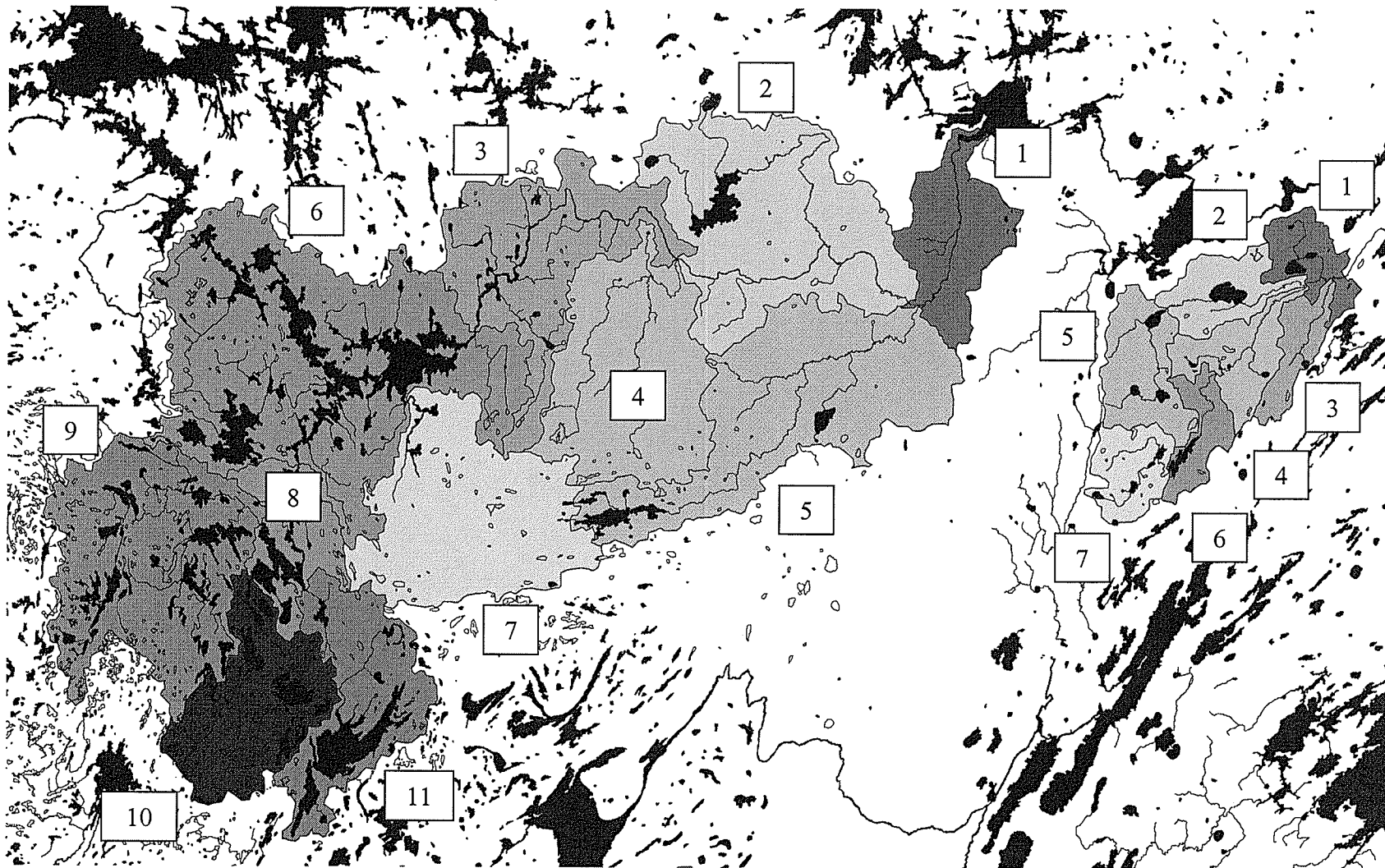


Figure 3.18 - ASA distance and elevation physiographic parameters.

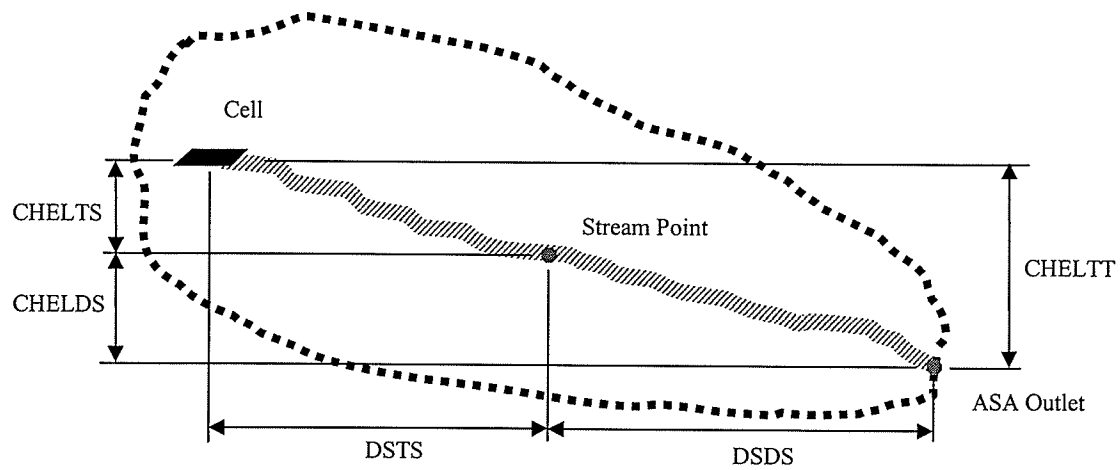
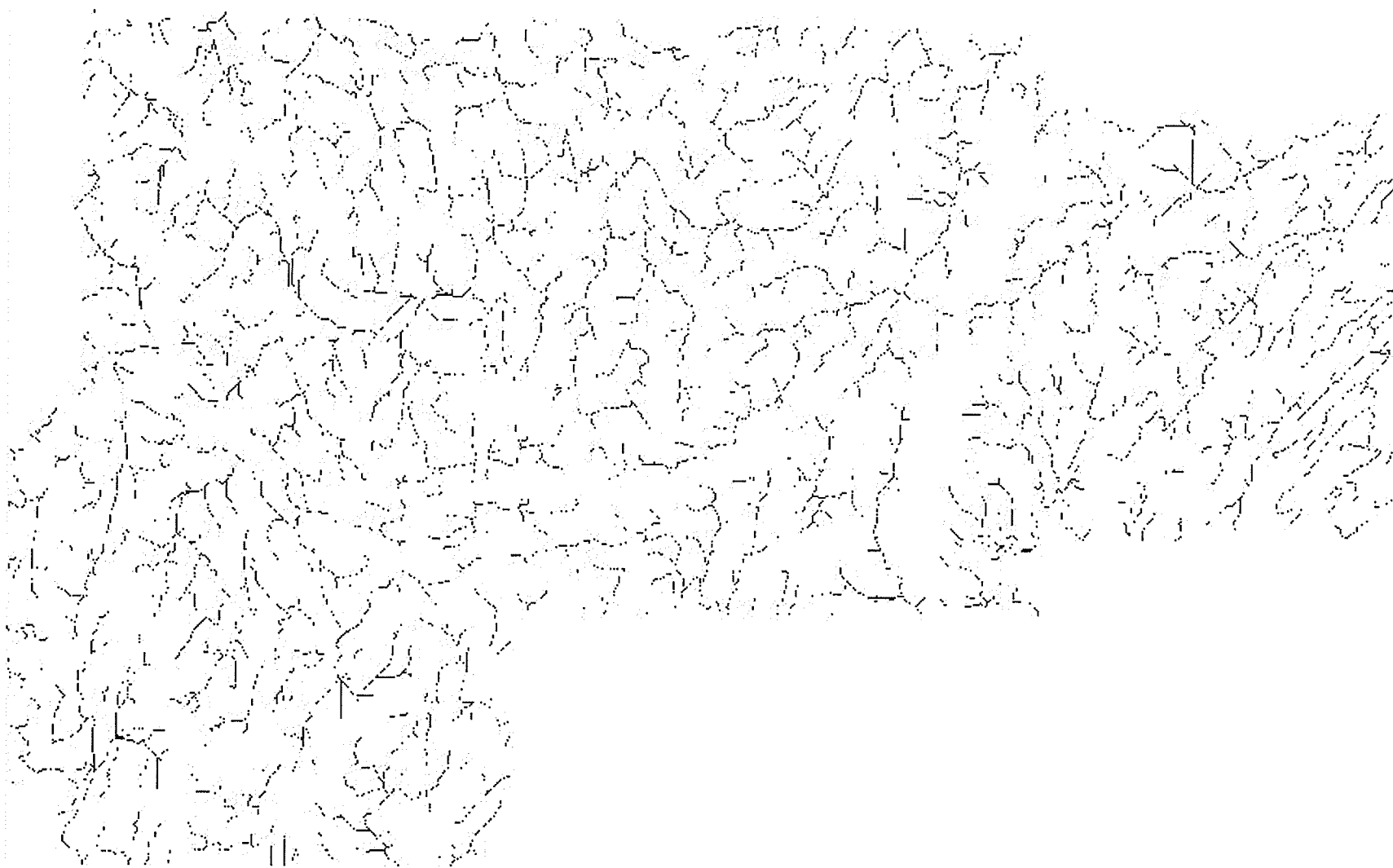


Figure 3.19 – Stream network grid for the Upper Burntwood River and Taylor River.



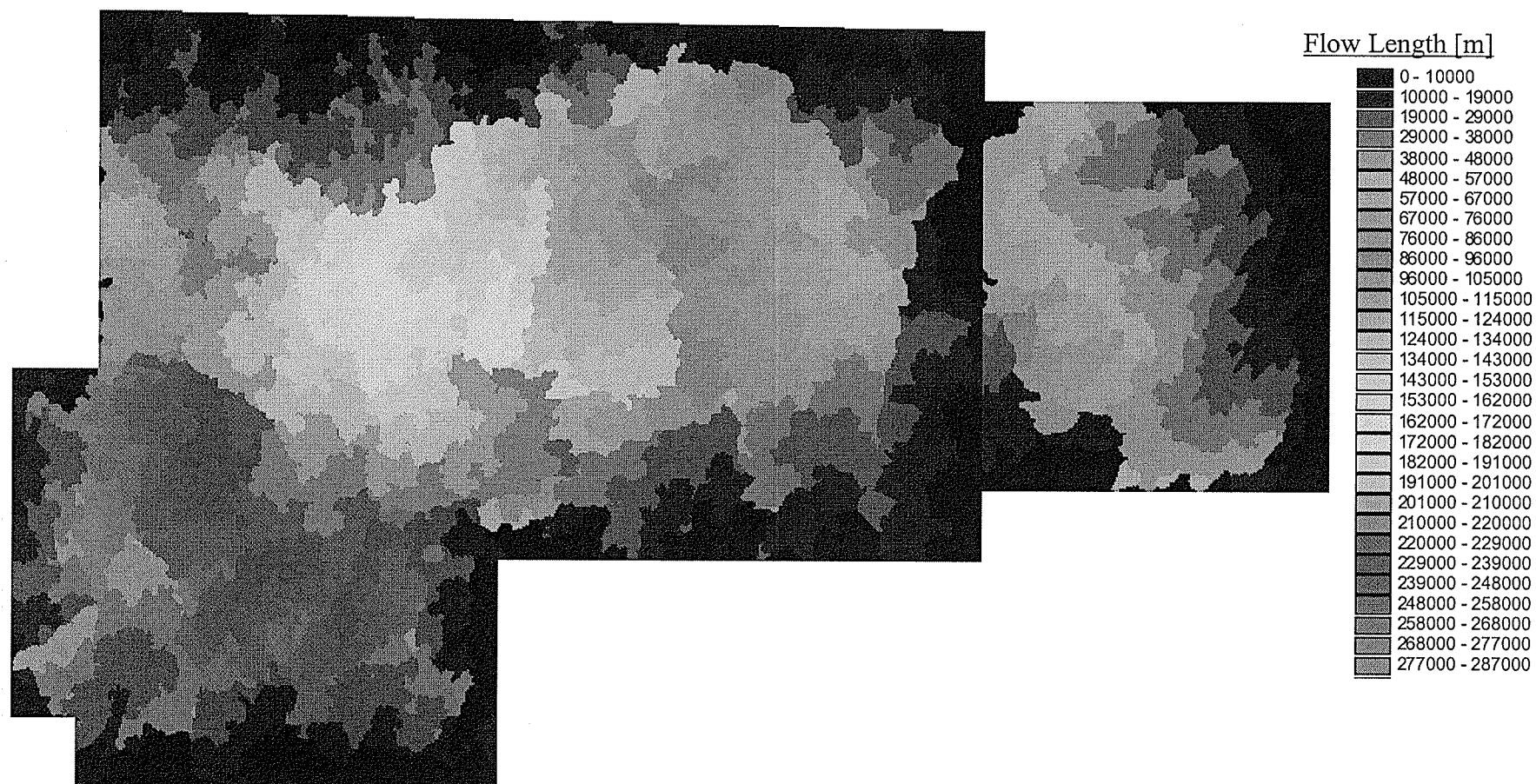


Figure 3.20 – Total flow length to outlet grid for the Upper Burntwood River and Taylor River.

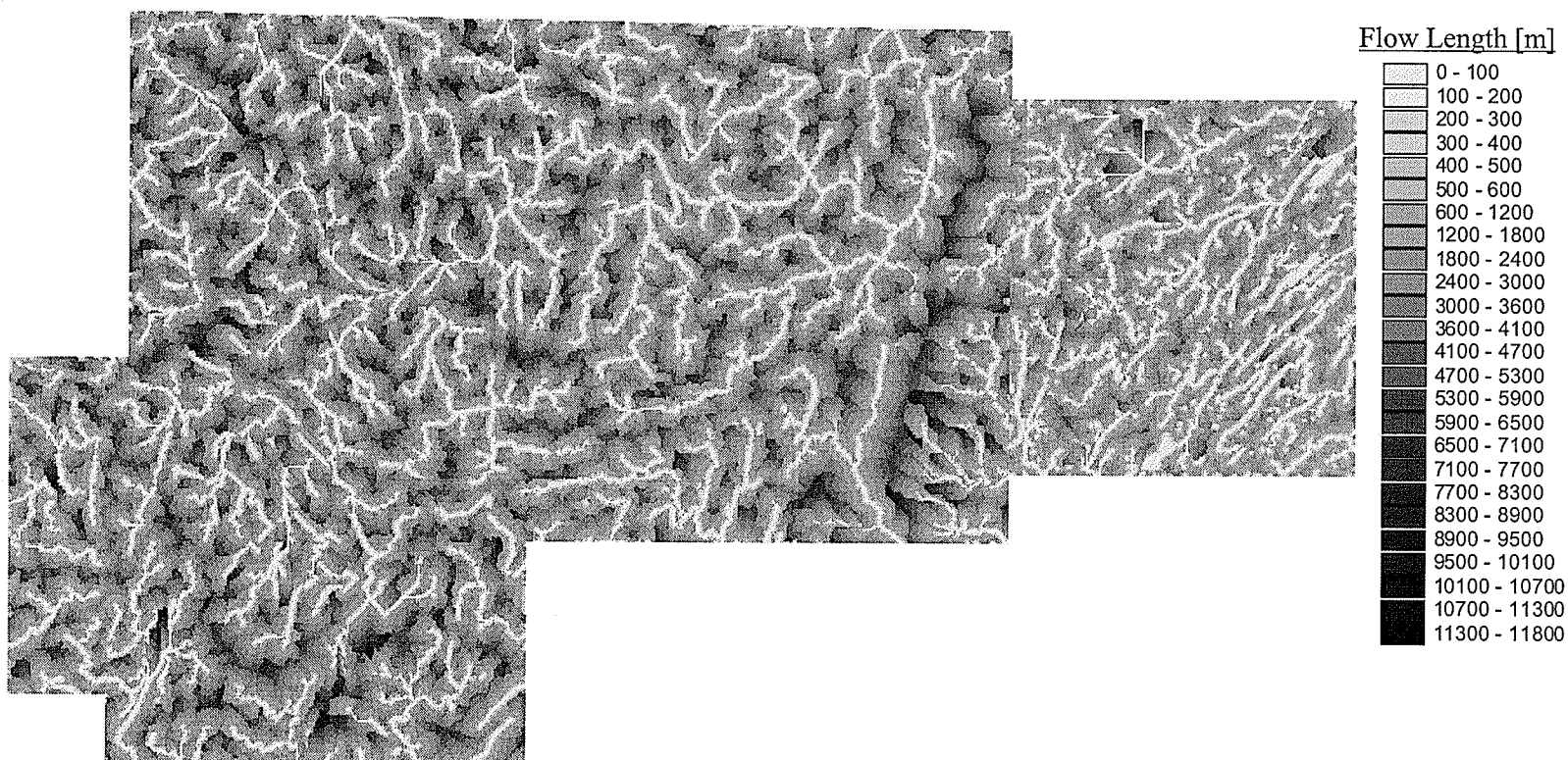


Figure 3.21 – To stream flow length grid for the Upper Burntwood River and Taylor River.

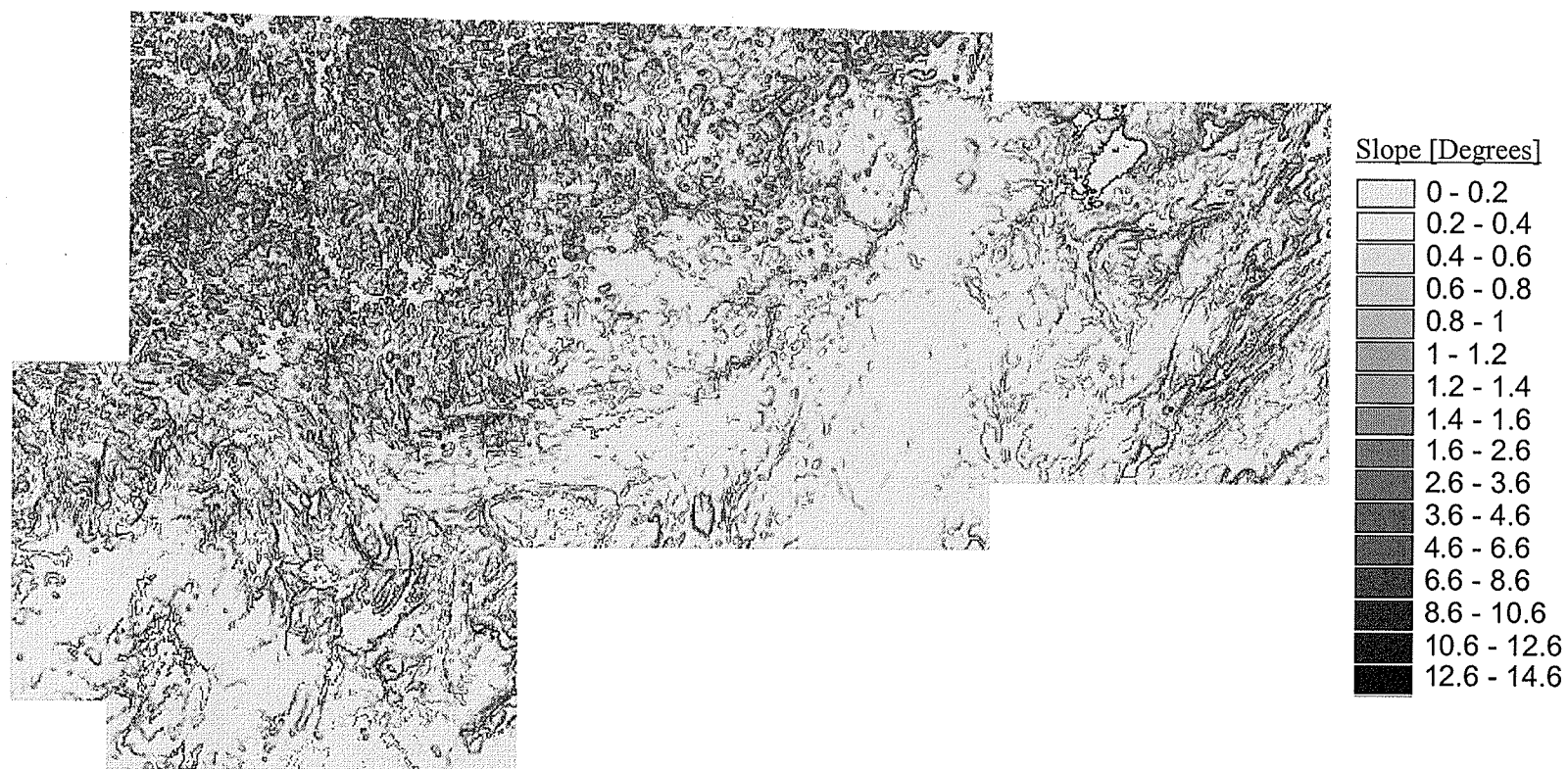


Figure 3.22 – Slope grid for the Upper Burntwood River and Taylor River.

Figure 3.23 - Environment Canada meteorological gauges.

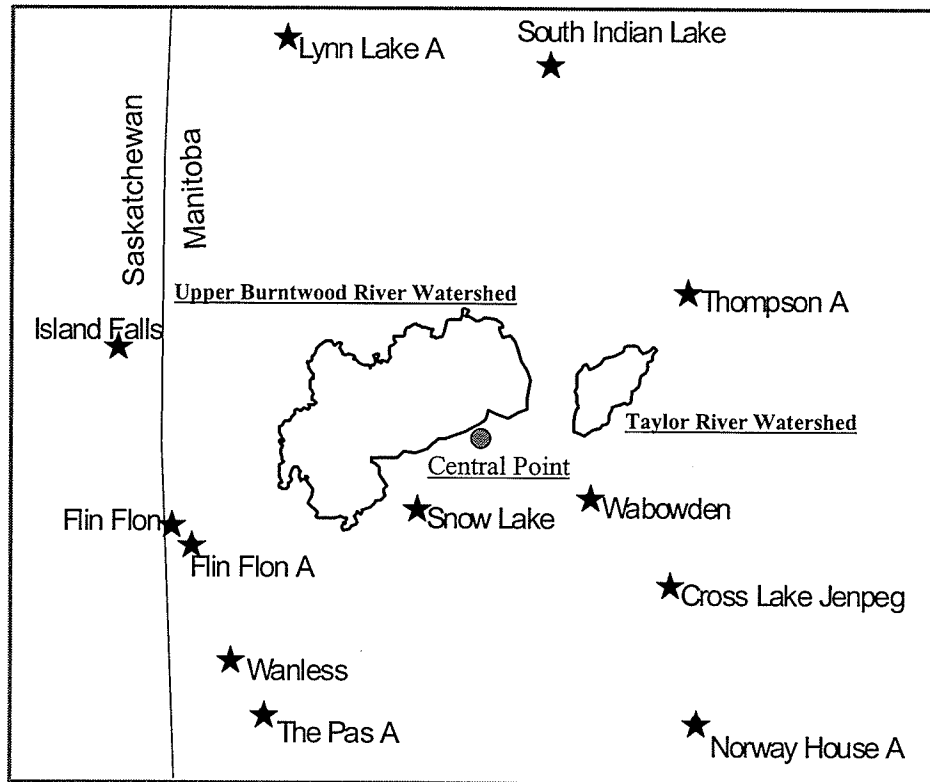


Figure 3.24 - Upper Burntwood River at Gate Falls discharge reduction factor.

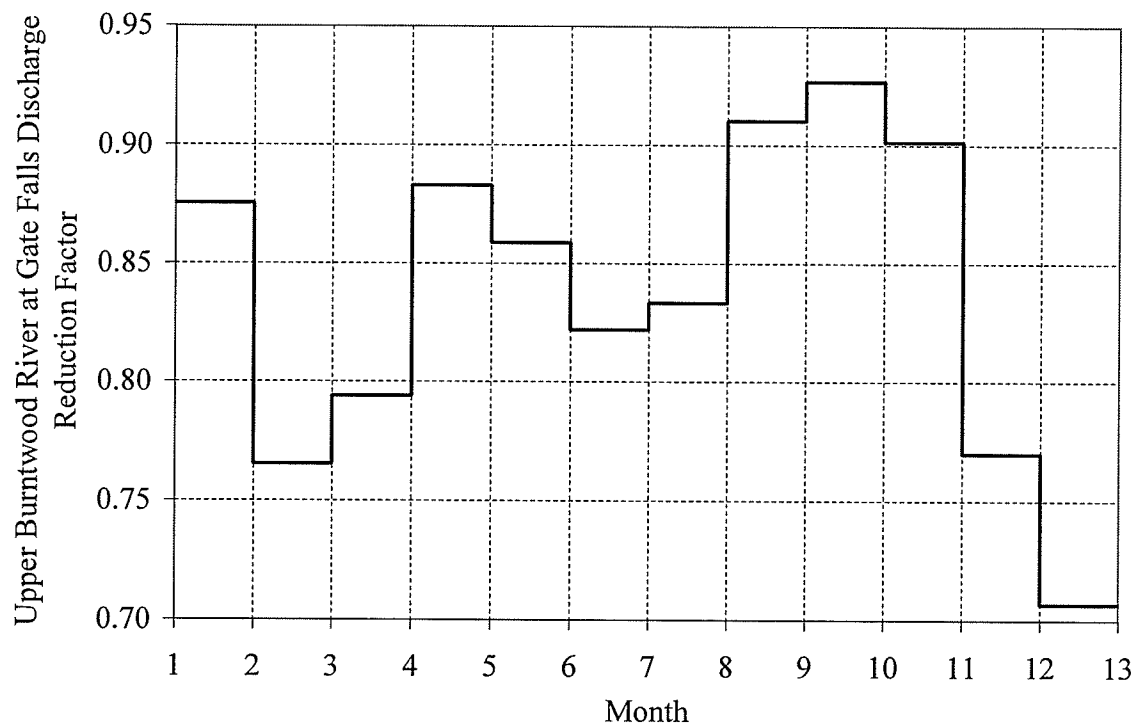


Figure 3.25 – Leaf Rapids to Gate Falls discharge ratio.

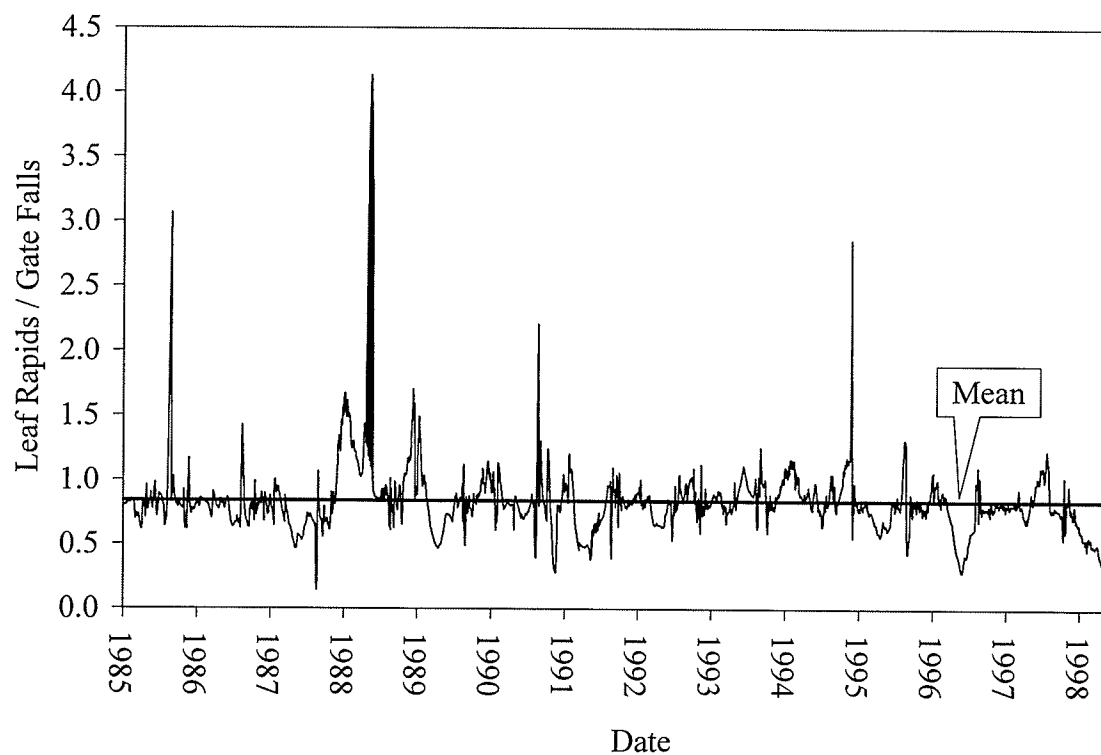


Figure 3.26 - Upper Burntwood River discharge hydrograph at the WSC Leaf Rapids gauge.

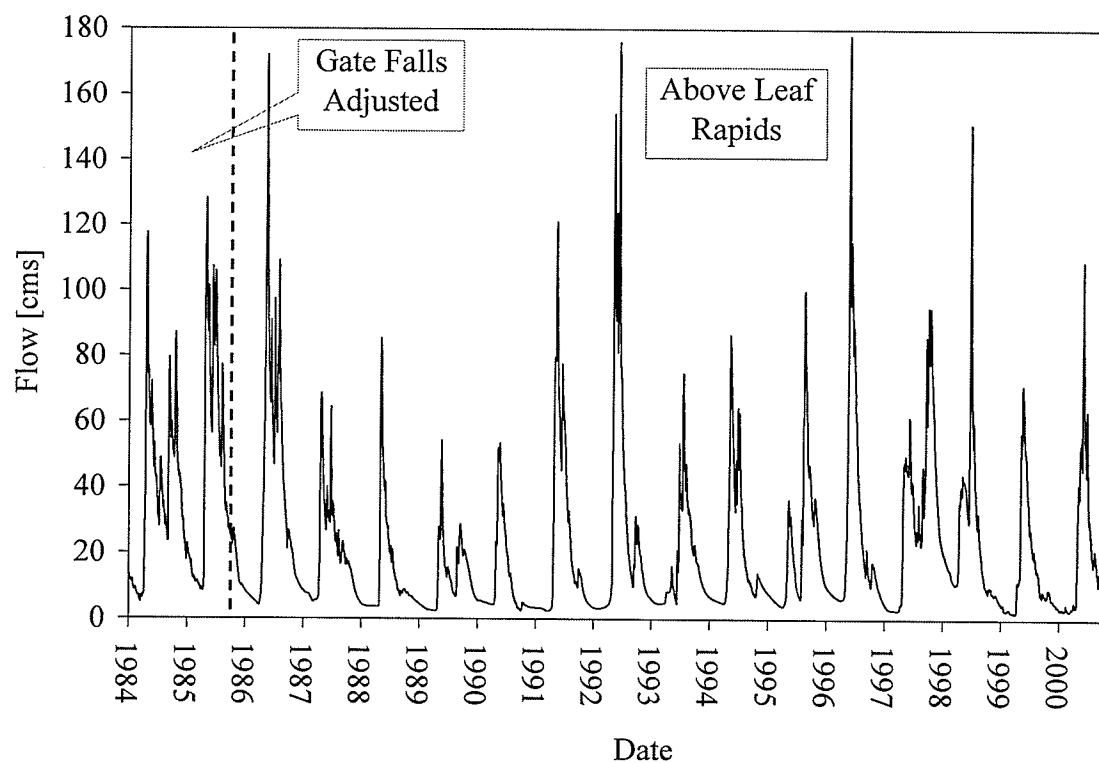


Figure 3.27 – Taylor River discharge hydrograph at the WSC Taylor River gauge.

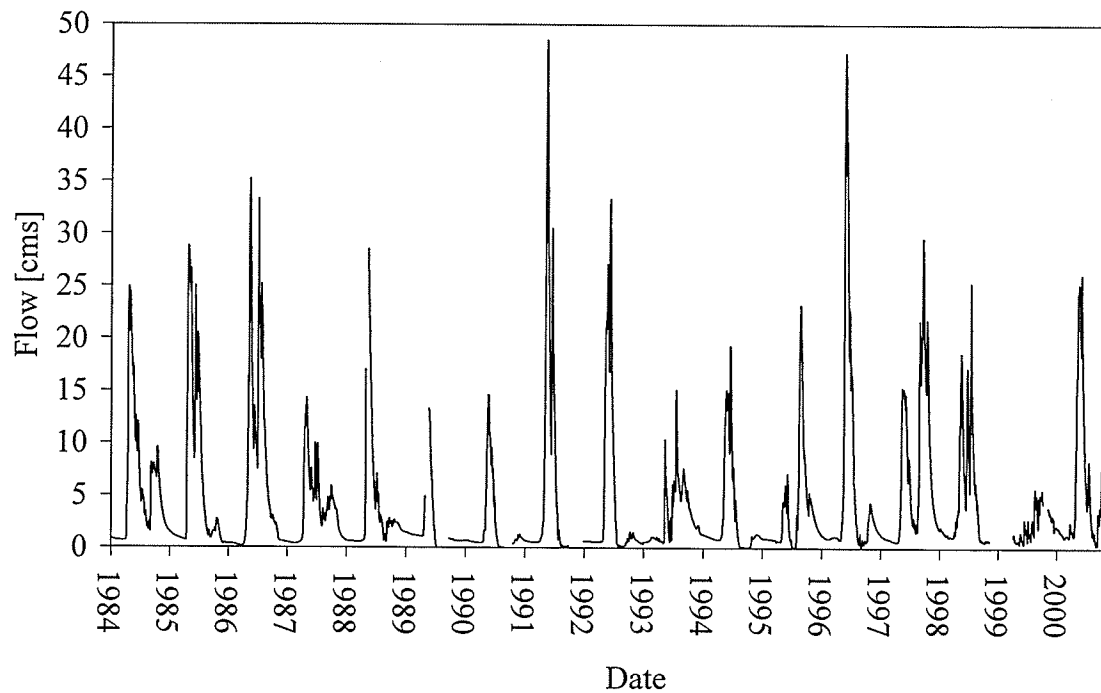


Figure 3.28 – Hydrometric gauge location plan.

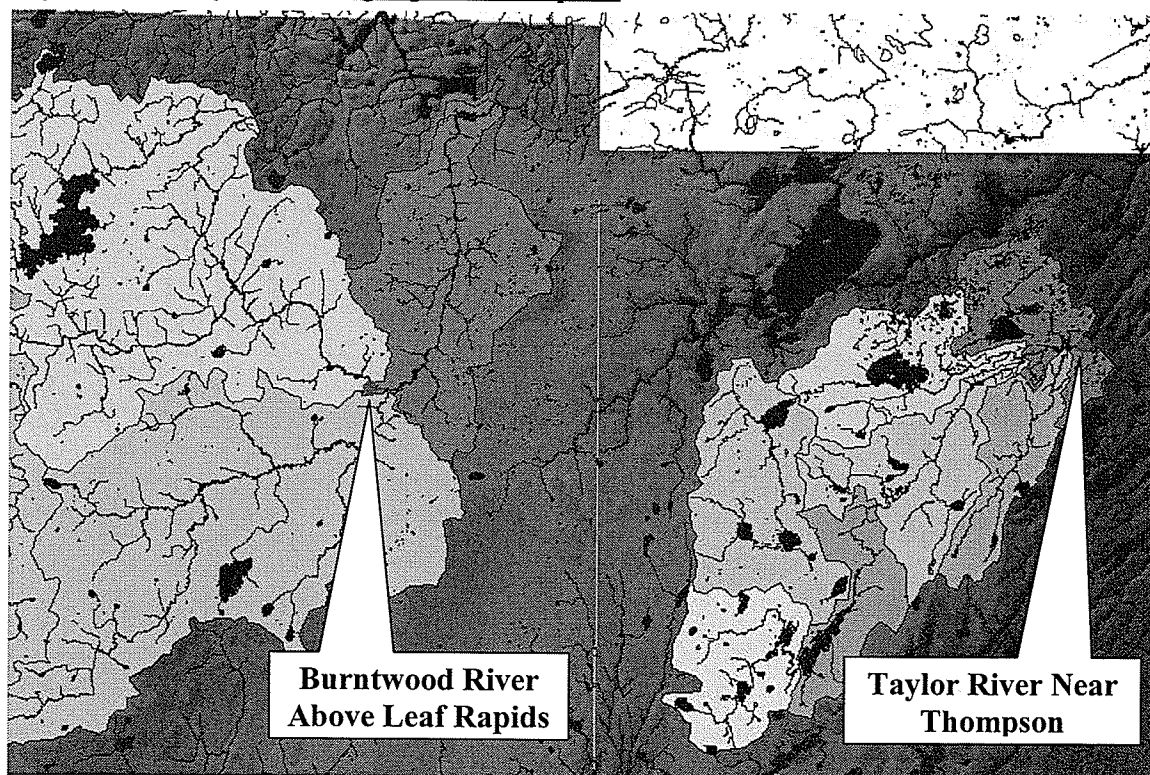


Figure 3.29 - Classification of years 1984-2000 relative to climatic precipitation normal.

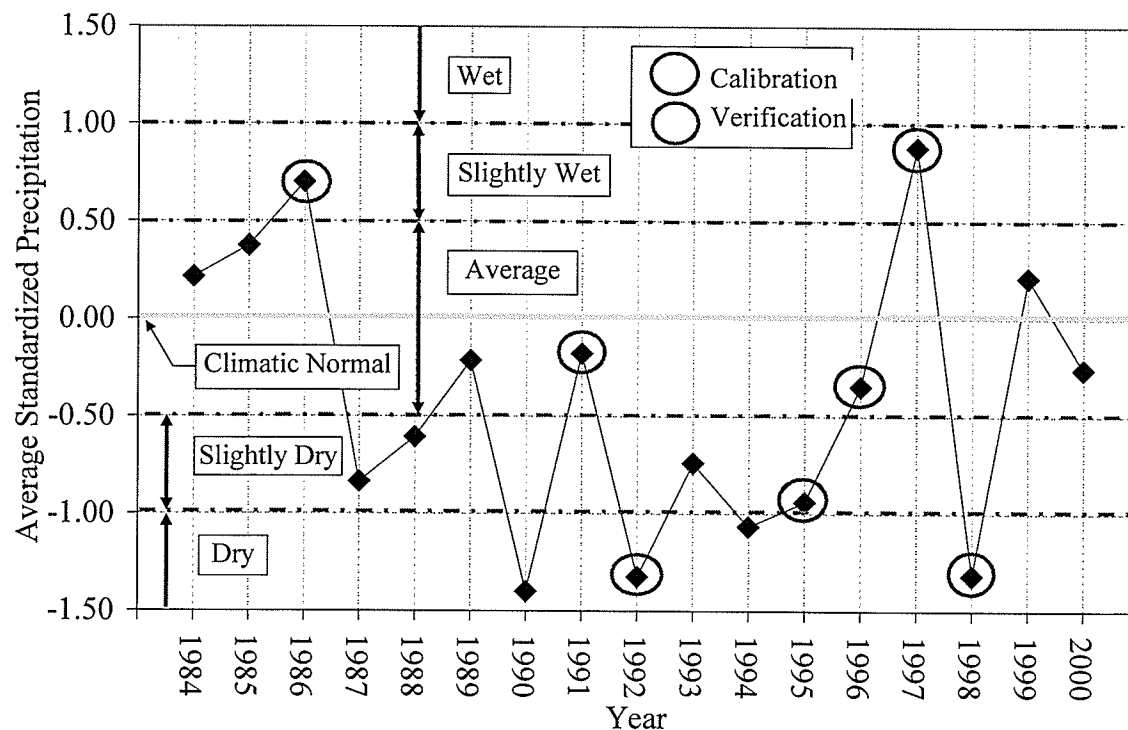


Figure 3.30 - Correlation test between precipitation and streamflow.

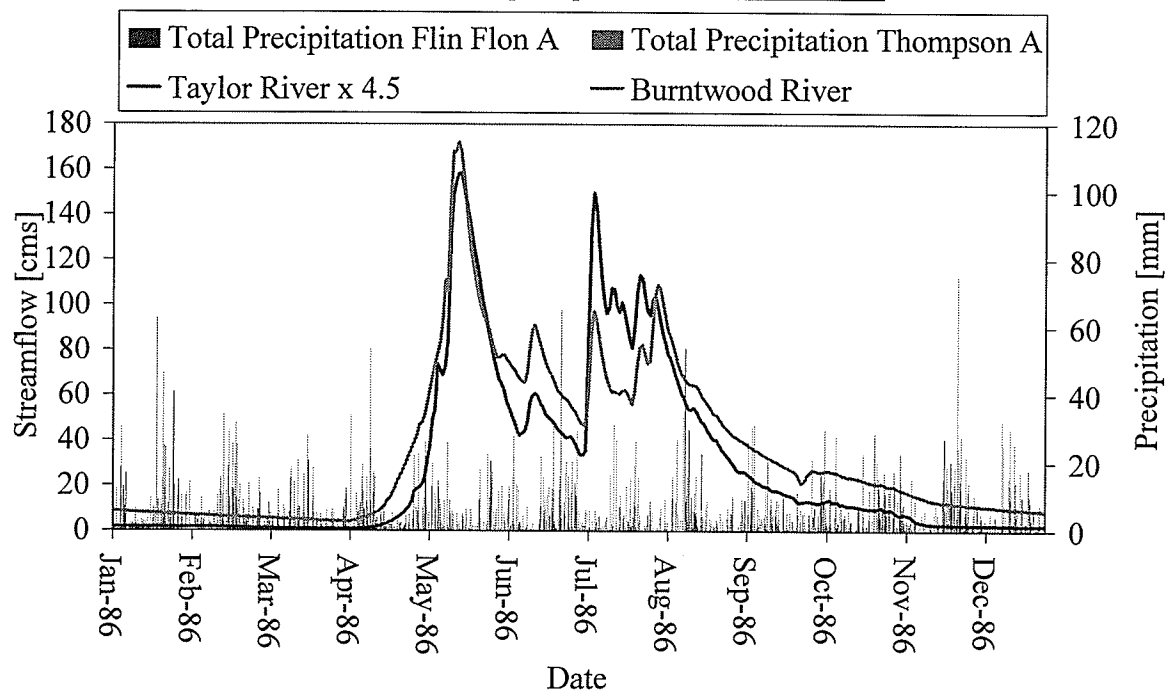


Figure 3.31 - Time series of LAI for the major types of land cover in the study area.

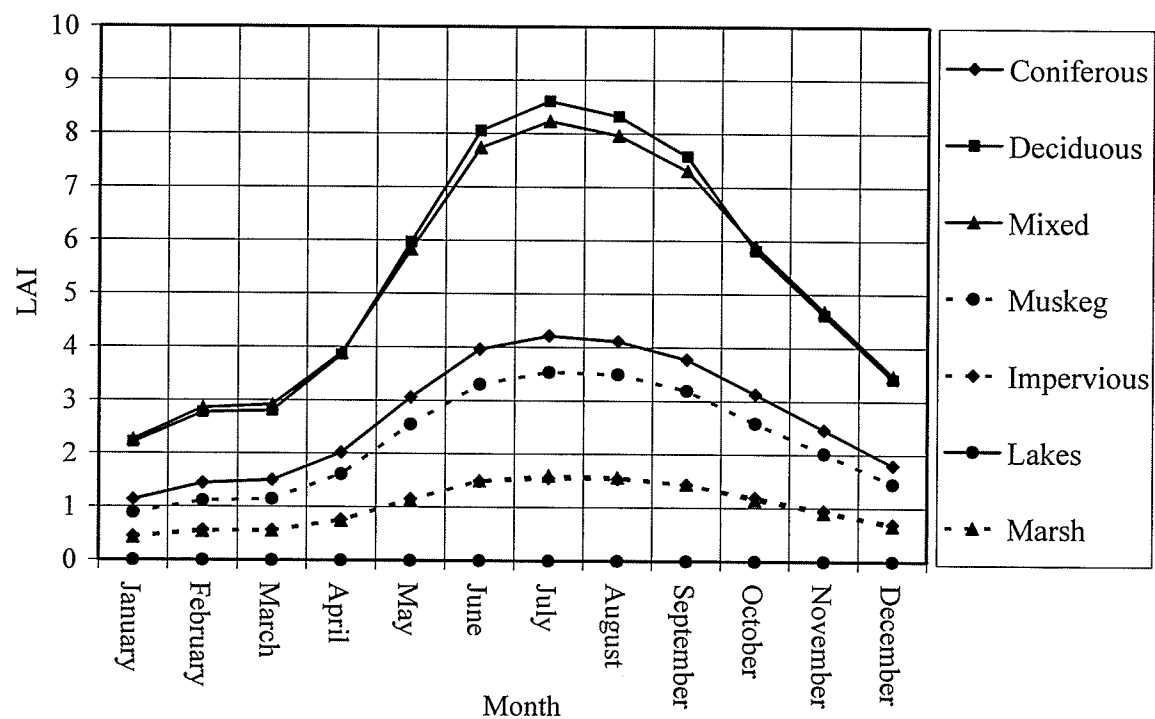


Table 3.1- Unique Forest Resources Inventory Classes.

Unique FRI Classes
Black Spruce Treed Muskeg
Tamarack Larch Treed Muskeg
Jack Pine Treed Rock
Black Spruce Treed Rock
Hardwood Treed Rock
Willow
Alder
Recreational Sites
Small Islands
Precipitous Slopes/Fragile Sites
Barrens - Tundra
Bare Rock - Igneous
Land Clearing In Progress
Abandoned Cultivated Land
Dry Upland Ridge Prairie
Moist Prairie
Wet Meadow
Muskeg
Sting Bogs
Marsh
Mud/Salt Flats
Sand Beaches
Town sites/Residential Sites
Airstrips
Roads/Railroads
Transmission Lines/Pipelines
Gravel Pits/Mine Sites
Drainage Ditches
Beaver Flood
Dugouts/Water Holes
Water
Rivers

Table 3.2- Forest Resources Inventory plot species and attributes.

Plot Species	Type	Habitat (Johnson, 1995)
Black Spruce	Coniferous	variety: generally wet poorly drained sites with tamarack (south); moist soils with many others in north
Jack Pine	Coniferous	coarse sands, shallow soils, rock outcrops, permafrost
Trembling Aspen	Deciduous	variety of soils
White Birch	Deciduous	forest edges, lakeshores, roadsides; variety of soils
White Spruce	Coniferous	variety of soils and climatic conditions with many other trees
Balsam Poplar	Deciduous	river valleys, moist rich low lying ground
Tamarack	Coniferous	cold, wet, poorly drained sites such as sphagnum bogs and muskeg. Grows best in moist well drained light soils
Balsam Fir	Coniferous	variety of soils and climatic conditions with many other trees

Table 3.3 – Reclassification of FRI dataset.

Coniferous	Area %	Deciduous	Area %	Mixed	Area %	Lakes / Rivers	Area %
Black Spruce	23.9427	Trembling Aspen	2.2751	Mixed	6.5291	Water	9.4781
Jack Pine	14.2805	Willow	0.4559	Small Islands	0.0546	Rivers	0.6553
Coniferous	4.7815	White Birch	0.0168				
White Spruce	0.1808	Deciduous	0.0164				
Tamarack	0.0590	Balsam Poplar	0.0094				
Balsam Fir	0.0023	Alder	0.0023				
Total	43.2468		2.7760		6.5837		10.1334

Muskeg	Area %	Treed Rock / Impervious	Area %	Marsh / Beaver Flood	Area %
Black Spruce Treed Muskeg	22.3154	Jack Pine Treed Rock	5.4697	Beaver Flood	4.3317
Tamarack Larch Treed Muskeg	3.1334	Black Spruce Treed Rock	0.1043	Marsh	0.1830
Muskeg	1.3954	Gravel Pits/Mine Sites	0.0936	Dugouts/Water Holes	0.0009
Transmission Lines/Pipelines	0.0521	Roads/Railroads	0.0884		
Sting Bogs	0.0167	Townsites/Residential Sites	0.0303		
Wet Meadow	0.0083	Bare Rock - Igneous	0.0184		
Drainage Ditches	0.0048	Hardwood Treed Rock	0.0041		
Moist Prairie	0.0035	Airstrips	0.0001		
Land Clearing In Progress	0.0024				
Mud/Salt Flats	0.0013				
Sand Beaches	0.0007				
Barrens - Tundra	0.0006				
Dry Upland Ridge Prairie	0.0006				
Abandoned Cultivated Land	0.0003				
Recreational Sites	0.0001				
	26.9356		5.8089		4.5156

Table 3.4- ASA areas and land cover fractions for the Taylor River watershed.

ASA	Area km ²	Conifer %	Decid %	Mixed %	Muskeg %	Imperv %	Water %	Marsh %
1	109.6	43.8	6.0	7.4	24.3	5.4	8.3	4.8
2	140.7	29.0	1.0	4.2	44.6	7.9	5.5	7.7
3	91.4	36.3	4.7	8.5	33.3	1.6	5.4	10.3
4	151.2	43.8	2.0	2.6	37.7	4.3	6.6	2.9
5	187.5	42.1	3.3	10.1	32.7	3.7	2.5	5.6
6	94.8	29.8	4.0	10.0	31.9	7.8	4.8	11.7
7	124.2	10.4	0.1	0.0	57.9	21.5	4.1	6.0

Table 3.5- ASA areas and land cover fractions for the Upper Burntwood River watershed.

ASA	Area km ²	Conifer %	Decid %	Mixed %	Muskeg %	Imperv %	Water %	Marsh %
1	280.1	29.1	1.3	3.8	48.4	11.5	2.2	3.7
2	806.3	46.6	2.2	11.3	24.8	6.1	4.4	4.6
3	570.6	55.5	2.3	6.1	16.1	9.5	3.6	6.9
4	631.7	37.2	2.0	3.1	45.7	7.7	1.5	2.8
5	806.8	39.9	1.0	1.3	45.8	4.0	3.9	4.1
6	1347.1	47.4	2.7	10.1	12.1	8.0	15.7	4.0
7	682.0	48.0	1.5	4.2	28.8	7.0	2.7	7.8
8	222.8	44.8	5.3	9.1	15.8	1.7	19.7	3.6
9	791.2	48.2	7.5	8.5	17.1	1.7	12.5	4.5
10	411.3	48.3	4.2	7.3	22.4	1.1	12.3	4.4
11	409.1	40.4	2.1	8.9	20.4	3.8	19.8	4.6

Table 3.6- ASA average elevation for the Upper Burntwood River and Taylor River.

ASA	Average Elevation [m]	Average Elevation [m]
	Upper Burntwood River	Taylor River
1	264.278	219.363
2	282.361	237.278
3	300.299	230.936
4	295.144	239.291
5	284.344	244.369
6	302.395	244.297
7	309.311	247.547
8	296.789	-
9	313.633	-
10	305.108	-
11	306.104	-

Table 3.7- ASA To-Stream and Down-Stream Distance physiographic parameters for the Taylor River Watershed.

ASA	Land Cover	DSDS Mean [m]	DSDS Standard Deviation [m]	DSTS Mean [m]	DSTS Standard Deviation [m]
1	Coniferous	8116	5121	1312	912
1	Deciduous	5304	3450	849	843
1	Mixed	5026	3663	1357	1098
1	Muskeg	8150	5475	1241	915
1	Impervious	2772	3053	1468	1038
1	Marsh	9163	2750	666	510
1	Lake/River	9427	5126	1230	924
2	Coniferous	14520	5749	1163	1000
2	Deciduous	14837	5554	711	759
2	Mixed	14123	4956	1079	941
2	Muskeg	16233	4220	1329	926
2	Impervious	16690	3831	1515	986
2	Marsh	12946	2296	1036	939
2	Lake/River	14350	5686	741	719
3	Coniferous	11052	5599	1054	894
3	Deciduous	13160	6932	645	605
3	Mixed	10836	6050	1034	954
3	Muskeg	12565	6369	1226	841
3	Impervious	14199	5975	803	602
3	Marsh	14609	4405	1221	909
3	Lake/River	12601	4404	578	764
4	Coniferous	16409	6593	905	750

Table 3.7 continued

ASA	Land Cover	DSDS Mean [m]	DSDS Standard Deviation [m]	DSTS Mean [m]	DSTS Standard Deviation [m]
4	Deciduous	18825	4430	507	664
4	Mixed	15918	8112	747	720
4	Muskeg	19830	4803	1329	969
4	Impervious	20578	6145	2119	1115
4	Marsh	20581	4690	424	517
4	Lake/River	17239	6127	819	738
5	Coniferous	11614	7054	1332	986
5	Deciduous	9846	5875	906	845
5	Mixed	8482	5512	953	823
5	Muskeg	13617	6819	1640	1174
5	Impervious	13861	6906	2590	1152
5	Marsh	11344	8101	347	451
5	Lake/River	12168	6436	988	873
6	Coniferous	10798	6483	1076	849
6	Deciduous	11677	4499	386	566
6	Mixed	10707	7367	912	748
6	Muskeg	9142	5519	1681	1193
6	Impervious	14255	6102	1341	865
6	Marsh	15736	2908	518	588
6	Lake/River	12921	6718	810	667
7	Coniferous	10160	5031	1209	963
7	Deciduous	7170	4127	786	638
7	Mixed	7444	4355	963	923

Table 3.7 continued

ASA	Land Cover	DSDS Mean [m]	DSDS Standard Deviation [m]	DSTS Mean [m]	DSTS Standard Deviation [m]
7	Muskeg	11411	5776	1471	1050
7	Impervious	10412	6932	1809	1082
7	Marsh	11129	4539	940	891
7	Lake/River	10768	4145	806	798

Table 3.8- ASA change in elevation physiographic parameters for the Taylor River Watershed.

ASA	Land Cover	SLC [%]	SST- REAM [%]	CH- ELDS' [%]	CHE- LTS' [%]	CHE- LTT [%]	CHEL DS [%]	CHEL TS [%]
1	Coniferous	0.96	0.42	34.1	12.6	31.0	22.6	8.4
1	Deciduous	1.16	0.42	22.3	9.8	22.0	15.3	6.7
1	Mixed	1.03	0.42	21.1	14.0	27.0	16.2	10.8
1	Muskeg	0.64	0.42	34.2	7.9	29.0	23.6	5.4
1	Impervious	0.54	0.42	11.6	7.9	25.0	14.9	10.1
1	Marsh	0.48	0.42	38.5	3.2	16.0	14.8	1.2
1	Lake/River	0.68	0.42	39.6	8.3	27.0	22.3	4.7
2	Coniferous	1.21	0.47	67.5	14.1	31.0	25.7	5.3
2	Deciduous	1.24	0.47	69.0	8.8	22.0	19.5	2.5
2	Mixed	1.34	0.47	65.7	14.5	28.0	22.9	5.1
2	Muskeg	0.76	0.47	75.5	10.1	33.0	29.1	3.9
2	Impervious	1.08	0.47	77.6	16.4	46.0	38.0	8.0
2	Marsh	0.37	0.47	60.2	3.9	12.0	11.3	0.7
2	Lake/River	0.77	0.47	66.7	5.7	27.0	24.9	2.1
3	Coniferous	0.87	0.41	45.2	9.2	30.0	24.9	5.1

Table 3.8 continued

ASA	Land Cover	SLC [%]	SST-REAM [%]	CH-ELDS' [%]	CHE-LTS' [%]	CHE-LTT [%]	CHEL DS [%]	CHEL TS [%]
3	Deciduous	0.71	0.41	53.8	4.6	29.0	26.7	2.3
3	Mixed	1.02	0.41	44.3	10.6	30.0	24.2	5.8
3	Muskeg	0.59	0.41	51.4	7.3	32.0	28.0	4.0
3	Impervious	0.70	0.41	58.1	5.7	34.0	31.0	3.0
3	Marsh	0.54	0.41	59.8	6.6	28.0	25.2	2.8
3	Lake/River	0.69	0.41	51.6	4.0	25.0	23.2	1.8
4	Coniferous	1.07	0.54	89.3	9.7	30.0	27.1	2.9
4	Deciduous	0.94	0.54	102.4	4.8	22.0	21.0	1.0
4	Mixed	1.42	0.54	86.6	10.6	28.0	24.9	3.1
4	Muskeg	0.57	0.54	107.9	7.6	34.0	31.8	2.2
4	Impervious	1.06	0.54	111.9	22.4	46.0	38.3	7.7
4	Marsh	0.83	0.54	112.0	3.5	18.0	17.5	0.5
4	Lake/River	0.67	0.54	93.8	5.5	29.0	27.4	1.6
5	Coniferous	0.72	0.27	31.0	9.6	19.0	14.5	4.5
5	Deciduous	1.02	0.27	26.3	9.2	16.0	11.8	4.2
5	Mixed	0.97	0.27	22.6	9.2	17.0	12.1	4.9
5	Muskeg	0.53	0.27	36.3	8.6	22.0	17.8	4.2
5	Impervious	0.54	0.27	37.0	13.9	33.0	24.0	9.0
5	Marsh	0.45	0.27	30.3	1.6	6.0	5.7	0.3
5	Lake/River	0.53	0.27	32.4	5.3	14.0	12.0	2.0
6	Coniferous	0.60	0.27	29.0	6.5	23.0	18.8	4.2
6	Deciduous	0.46	0.27	31.3	1.8	17.0	16.1	0.9
6	Mixed	0.88	0.27	28.7	8.0	21.0	16.4	4.6

Table 3.8 continued

ASA	Land Cover	SLC [%]	SST-REAM [%]	CH-ELDS' [%]	CHE-LTS' [%]	CHE-LTT [%]	CHEL-DS [%]	CHEL-TS [%]
6	Muskeg	0.37	0.27	24.5	6.2	24.0	19.2	4.8
6	Impervious	0.66	0.27	38.3	8.9	27.0	21.9	5.1
6	Marsh	0.27	0.27	42.2	1.4	17.0	16.5	0.5
6	Lake/River	0.42	0.27	34.7	3.4	20.0	18.2	1.8
7	Coniferous	0.45	0.17	16.9	5.4	9.0	6.8	2.2
7	Deciduous	0.27	0.17	11.9	2.1	4.0	3.4	0.6
7	Mixed	0.55	0.17	12.3	5.3	6.0	4.2	1.8
7	Muskeg	0.37	0.17	18.9	5.4	11.0	8.5	2.5
7	Impervious	0.51	0.17	17.3	9.1	14.0	9.2	4.8
7	Marsh	0.19	0.17	18.5	1.8	6.0	5.5	0.5
7	Lake/River	0.39	0.17	17.9	3.1	6.0	5.1	0.9

Table 3.9- ASA To-Stream and Down-Stream Distance physiographic parameters for the Upper Burntwood River Watershed.

ASA	Land Cover	DSDS Mean [m]	DSDS Standard Deviation [m]	DSTS Mean [m]	DSTS Standard Deviation [m]
1	Coniferous	19789	6769	2020	1863
1	Deciduous	20250	10559	1199	1221
1	Mixed	21951	8030	982	1219
1	Muskeg	23642	7129	2296	1508
1	Impervious	19596	5903	3408	1382
1	Marsh	15185	10542	661	1449
1	Lake/River	22814	6170	1855	1121
2	Coniferous	40783	20859	1803	1454
2	Deciduous	40969	21937	1054	1412
2	Mixed	32988	19919	1471	1387

Table 3.9 continued

ASA	Land Cover	DSDS Mean [m]	DSDS Standard Deviation [m]	DSTS Mean [m]	DSTS Standard Deviation [m]
2	Muskeg	42937	23018	2226	1456
2	Impervious	38554	21312	2430	1429
2	Marsh	60062	15267	1491	1509
2	Lake/River	36612	21750	1584	1319
3	Coniferous	41177	19669	1972	1567
3	Deciduous	43642	13706	967	1008
3	Mixed	43725	17256	1367	1183
3	Muskeg	39673	23185	2418	1645
3	Impervious	45804	19079	2735	1782
3	Marsh	47408	14482	1280	1594
3	Lake/River	45604	15819	1608	1341
4	Coniferous	30493	11672	1795	1455
4	Deciduous	25252	8607	547	880
4	Mixed	15731	8984	1257	1191
4	Muskeg	31786	11275	1939	1405
4	Impervious	34235	12200	2688	1476
4	Marsh	43647	10280	1651	1424
4	Lake/River	26150	9647	1405	1269
5	Coniferous	33758	18015	1968	1535
5	Deciduous	32357	16742	1014	1421
5	Mixed	21640	16063	1642	1528
5	Muskeg	31817	18254	2170	1552
5	Impervious	40711	19305	3054	1813
5	Marsh	55303	19525	2037	1483
5	Lake/River	26264	12704	1390	1232
6	Coniferous	34978	17621	2315	1701
6	Deciduous	32303	17817	1933	1635
6	Mixed	34867	15972	1990	1491
6	Muskeg	36456	17941	2603	1788
6	Impervious	37945	18718	2592	1799

Table 3.9 continued

ASA	Land Cover	DSDS Mean [m]	DSDS Standard Deviation [m]	DSTS Mean [m]	DSTS Standard Deviation [m]
6	Marsh	31493	16987	1207	1107
6	Lake/River	32905	17883	1728	1539
7	Coniferous	31366	12352	2028	1512
7	Deciduous	15436	9961	1414	1293
7	Mixed	16639	9312	1468	1231
7	Muskeg	37622	10332	2223	1526
7	Impervious	29983	11569	2389	1553
7	Marsh	28263	15060	1472	1424
7	Lake/River	28575	12265	1134	1166
8	Coniferous	14849	8208	2057	1411
8	Deciduous	22028	7286	1550	1151
8	Mixed	17771	8904	1983	1400
8	Muskeg	15975	8550	2181	1428
8	Impervious	20523	11944	2800	1423
8	Marsh	21063	7830	1064	963
8	Lake/River	16803	8628	2050	1363
9	Coniferous	34264	13871	2098	1548
9	Deciduous	19931	10905	1632	1409
9	Mixed	24285	13253	1914	1523
9	Muskeg	34963	14446	2437	1596
9	Impervious	35466	14213	2054	1568
9	Marsh	31411	16299	1475	1386
9	Lake/River	27000	12511	1507	1298
10	Coniferous	24879	10346	1963	1540
10	Deciduous	14681	10883	1241	1231
10	Mixed	17700	10461	1695	1344
10	Muskeg	27559	9508	2357	1539
10	Impervious	19382	8724	2160	1632
10	Marsh	21974	11551	1528	1366
10	Lake/River	19449	10366	1460	1363

Table 3.9 continued

ASA	Land Cover	DSDS Mean [m]	DSDS Standard Deviation [m]	DSTS Mean [m]	DSTS Standard Deviation [m]
11	Coniferous	23839	11163	2616	1924
11	Deciduous	23248	11728	2699	1980
11	Mixed	25851	13652	2077	1743
11	Muskeg	21579	11974	2422	1769
11	Impervious	27172	11114	2677	1617
11	Marsh	27504	6997	1836	1543
11	Lake/River	23870	12971	1687	1604

Table 3.10- ASA change in elevation physiographic parameters for the Upper Burntwood River Watershed.

ASA	Land Cover	SLC [%]	SST- REAM [%]	CH- ELDS' [%]	CHE- LTS' [%]	CHE- LTT [%]	CHEL DS [%]	CHEL TS [%]
1	Coniferous	0.86	0.65	129.0	17.3	23.0	2.7	20.2
1	Deciduous	0.96	0.65	132.0	11.5	16.8	1.3	15.4
1	Mixed	1.13	0.65	143.1	11.1	16.5	1.2	15.3
1	Muskeg	0.50	0.65	154.1	11.4	25.1	1.7	23.3
1	Impervious	0.49	0.65	127.7	16.7	30.3	3.5	26.8
1	Marsh	1.13	0.65	99.0	7.5	4.4	0.3	4.1
1	Lake/River	0.40	0.65	148.7	7.3	23.7	1.1	22.6
2	Coniferous	1.18	0.51	208.9	21.3	35.4	3.3	32.2
2	Deciduous	1.20	0.51	209.8	12.6	27.7	1.6	26.2
2	Mixed	1.28	0.51	168.9	18.8	30.6	3.1	27.6
2	Muskeg	0.70	0.51	219.9	15.5	35.5	2.3	33.2
2	Impervious	1.62	0.51	197.4	39.4	47.9	8.0	40.0
2	Marsh	0.44	0.51	307.6	6.6	27.5	0.6	26.9
2	Lake/River	0.61	0.51	187.5	9.6	28.6	1.4	27.2

Table 3.10 continued

ASA	Land Cover	SLC [%]	SST-REAM [%]	CH-ELDS' [%]	CHE-LTS' [%]	CHE-LTT [%]	CHELDS [%]	CHELTS [%]
3	Coniferous	1.64	0.81	335.3	32.4	41.3	3.6	37.7
3	Deciduous	1.75	0.81	355.3	16.9	33.3	1.5	31.8
3	Mixed	1.67	0.81	356.0	22.9	35.9	2.2	33.7
3	Muskeg	1.09	0.81	323.0	26.4	39.9	3.0	36.9
3	Impervious	1.88	0.81	372.9	51.4	55.0	6.7	48.3
3	Marsh	1.39	0.81	386.0	17.8	23.6	1.0	22.5
3	Lake/River	1.10	0.81	371.3	17.7	34.9	1.6	33.3
4	Coniferous	0.93	0.39	117.6	16.6	36.0	4.4	31.5
4	Deciduous	0.70	0.39	97.4	3.8	22.4	0.8	21.6
4	Mixed	1.37	0.39	60.7	17.3	22.5	5.0	17.5
4	Muskeg	0.56	0.39	122.6	10.9	35.0	2.8	32.2
4	Impervious	1.23	0.39	132.1	33.1	50.5	10.1	40.4
4	Marsh	0.67	0.39	168.4	11.1	42.4	2.6	39.7
4	Lake/River	0.59	0.39	100.9	8.3	25.5	1.9	23.5
5	Coniferous	0.74	0.39	131.5	14.6	42.0	4.2	37.8
5	Deciduous	0.91	0.39	126.0	9.2	33.1	2.3	30.8
5	Mixed	1.14	0.39	84.3	18.7	28.7	5.2	23.5
5	Muskeg	0.48	0.39	123.9	10.4	40.8	3.2	37.6
5	Impervious	0.81	0.39	158.5	24.8	54.4	7.4	47.1
5	Marsh	0.43	0.39	215.4	8.8	50.4	2.0	48.4
5	Lake/River	0.52	0.39	102.3	7.2	31.9	2.1	29.8
6	Coniferous	1.72	1.04	364.0	39.9	32.7	3.2	29.4
6	Deciduous	2.00	1.04	336.2	38.7	24.7	2.5	22.2

Table 3.10 continued

ASA	Land Cover	SLC [%]	SST-REAM [%]	CH-ELDS' [%]	CHE-LTS' [%]	CHE-LTT [%]	CHEL DS [%]	CHEL TS [%]
6	Mixed	2.02	1.04	362.9	40.1	25.9	2.6	23.3
6	Muskeg	1.12	1.04	379.4	29.2	32.5	2.3	30.1
6	Impervious	2.19	1.04	394.9	56.7	46.5	5.8	40.7
6	Marsh	1.04	1.04	327.8	12.6	6.6	0.2	6.4
6	Lake/River	1.12	1.04	342.5	19.3	23.6	1.3	22.3
7	Coniferous	1.17	0.50	157.4	23.7	31.0	4.1	27.0
7	Deciduous	1.63	0.50	77.4	23.0	18.7	4.3	14.4
7	Mixed	1.62	0.50	83.5	23.8	21.5	4.8	16.7
7	Muskeg	0.62	0.50	188.7	13.7	31.9	2.2	29.8
7	Impervious	1.63	0.50	150.4	38.9	38.7	8.0	30.8
7	Marsh	0.75	0.50	141.8	11.1	19.7	1.4	18.2
7	Lake/River	0.78	0.50	143.4	8.9	20.5	1.2	19.3
8	Coniferous	1.24	0.64	94.5	25.6	24.2	5.2	19.1
8	Deciduous	1.85	0.64	140.1	28.6	13.6	2.3	11.3
8	Mixed	1.60	0.64	113.0	31.8	18.9	4.1	14.7
8	Muskeg	0.90	0.64	101.6	19.5	25.9	4.2	21.7
8	Impervious	1.61	0.64	130.5	45.0	28.3	7.3	21.0
8	Marsh	0.52	0.64	134.0	5.5	4.4	0.2	4.2
8	Lake/River	0.90	0.64	106.9	18.4	14.9	2.2	12.7
9	Coniferous	1.16	0.55	188.3	24.4	37.2	4.3	32.9
9	Deciduous	1.33	0.55	109.5	21.6	19.6	3.2	16.3
9	Mixed	1.22	0.55	133.5	23.3	25.1	3.7	21.4
9	Muskeg	0.68	0.55	192.1	16.5	38.7	3.1	35.6

Table 3.10 continued

ASA	Land Cover	SLC [%]	SST-REAM [%]	CH-ELDS' [%]	CHE-LTS' [%]	CHE-LTT [%]	CHELDS [%]	CHELTS [%]
9	Muskeg	0.68	0.55	192.1	16.5	38.7	3.1	35.6
9	Impervious	1.59	0.55	194.9	32.6	47.0	6.7	40.3
9	Marsh	0.67	0.55	172.6	9.8	23.2	1.3	21.9
9	Lake/River	0.75	0.55	148.4	11.3	21.9	1.5	20.4
10	Coniferous	0.87	0.50	124.4	17.0	27.9	3.4	24.5
10	Deciduous	1.28	0.50	73.4	15.9	14.4	2.5	11.8
10	Mixed	1.21	0.50	88.5	20.6	17.8	3.3	14.4
10	Muskeg	0.50	0.50	137.8	11.7	30.5	2.4	28.1
10	Impervious	1.10	0.50	96.9	23.7	27.0	5.3	21.7
10	Marsh	0.48	0.50	109.8	7.3	10.8	0.7	10.1
10	Lake/River	0.79	0.50	97.2	11.6	18.3	2.0	16.3
11	Coniferous	1.17	0.84	200.2	30.6	32.1	4.3	27.8
11	Deciduous	1.70	0.84	195.2	45.9	22.8	4.3	18.5
11	Mixed	1.39	0.84	217.1	28.8	23.8	2.8	21.0
11	Muskeg	0.78	0.84	181.2	18.8	32.1	3.0	29.1
11	Impervious	1.48	0.84	228.2	39.5	36.8	5.4	31.4
11	Marsh	0.52	0.84	231.0	9.5	10.7	0.4	10.2
11	Lake/River	0.91	0.84	200.5	15.3	25.0	1.8	23.2

Table 3.11- UTM coordinates of Taylor River ASA centroids.

ASA	Northing	Easting
1	6151012	546690
2	6147674	535440
3	6139204	544388
4	6137933	536413
5	6137190	524148
6	6127769	531014
7	6122197	521478

Table 3.12- UTM coordinates of Upper Burntwood River ASA centroids.

ASA	Northing	Easting
1	6154660	497313
2	6156612	472022
3	6145404	442815
4	6137937	450435
5	6129380	470923
6	6135560	411124
7	6118202	429960
8	6110886	404154
9	6106871	385838
10	6091578	398986
11	6089283	413557

Table 3.13 – Meteorological gauges surrounding the study areas.

Station ID	Station Name	Elevation [m]	UTM X	UTM Y	Years	Temperature	Precipitation	Hours Of Bright	Solar Radiation	Relative Humidity	Dewpoint Temp	Wind Speed
5050960	Flin Flon A	304	327023	6062860	55-p	X	X			X	X	X
4063560	Island Falls	299	288598	6157332	45-p	X	X					
5061646	Lynn Lake A	357	373001	6304478	59-p	X	X	X				
506b047	Norway House A	223	575471	5978571	73-p	X	X					
5052880	The Pas A	271	362245	5981855	43-p	X	X	X	X	X	X	X
5062922	Thompson A	222	571046	6184402	67-p	X	X	X				
506b0m7	Norway House Forestry	217	578658	5984188	70-p	X	X					
5062706	Snow Lake	296	436890	6080397	83-p	X	X					
5061550	Leaf Rapids	282	437360	6258484	91-96	X	X					
5051986	Nome Lake	290	318525	6258484	86-91	X	X					
5062529	Ruttan Lake	289	459969	6260065	73-91	X	X					
5060623	Cross Lake Jenpeg	219	562551	6043293	72-p	X	X					
5050920	Flin Flon	335	316658	6072553	27-p	X	X					
5062734	South Indian Lake	259	502037	6293269	76-89	X	X					
5062736	South Indian Lake Airport	289	506107	6295128	90-p	X	X					
5063040	Wabowden	233	522435	6085574	82-p	X	X					
5053080	Wanless	271	345562	6006507	35-97	X	X					
5062111	Pikwitonei A	192	615561	6161233	87-p	X	X					

Table 3.14 - Climate Normal Precipitations for the meteorological stations surrounding the Upper Burntwood and Taylor River Basins.

Station	Period	Annual Normal Precipitation [mm]
Thompson A	1967-1990	535.6
Lynn Lake A	1959-1990	492.4
Flin Flon A	1955-1990	484.1
The Pas A	1943-1990	451.9
Island Falls	1929-1990	519.1
Norway House A	1974-1990	508.0
Mean		498.5
Standard Deviation		29.5

Table 3.15 – Estimates of the river routing parameters for each ASA of the Taylor River.

ASA	Strahler Order	Stream Width [m]	Stream Depth [m]	X	K [day]
1	4	40	5.00	0.3	1
2	3	30	3.75	0.2	1
3	2	10	1.25	0.3	1
4	3	30	3.75	0.3	1
5	2	10	1.25	0.3	1
6	3	30	3.75	0.3	1
7	2	10	1.25	0.2	1

Table 3.16 – Estimates of the river routing parameters for each ASA of the Upper Burntwood River.

ASA	Straher Order	Stream Width [m]	Stream Depth [m]	X	K [day]
1	4	90	11.25	0.3	1
2	3	60	7.50	0.3	1
3	4	60	7.50	0.3	1
4	2	10	1.25	0.3	1
5	2.5	20	2.50	0.3	1
6	Mainly Lake	300	10.00	0.2	1
7	2.5	20	2.50	0.3	1
8	Mainly Lake	300	10.00	0.2	1
9	2	10	1.25	0.3	1
10	2	10	1.25	0.3	1
11	3	30	3.75	0.3	1

Table 3.17 – Average NDVI for the land covers in the Upper Burntwood River and Taylor River.

Month	Conifer	Decid	Mixed	Muskeg	Imperv	Water	Marsh
Jan	0.14	0.139	0.148	0.134	0.148	0.064	0.142
Feb	0.178	0.175	0.188	0.168	0.186	0.09	0.179
Mar	0.185	0.177	0.192	0.173	0.187	0.092	0.182
Apr	-	-	-	-	-	-	-
May	0.376	0.376	0.383	0.386	0.381	0.231	0.375
Jun	0.487	0.507	0.508	0.5	0.491	0.304	0.494
Jul	0.518	0.542	0.541	0.534	0.511	0.292	0.527
Aug	0.505	0.524	0.523	0.528	0.508	0.309	0.517
Sep	0.464	0.477	0.48	0.482	0.473	0.276	0.476
Oct	0.385	0.366	0.387	0.39	0.396	0.204	0.38
Nov	-	-	-	-	-	-	-
Dec	-	-	-	-	-	-	-

Table 3.18 – LAI_{MAX} for the major types of land covers in the study area.

Conifer	Decid	Mixed	Muskeg	Imperv	Water	Marsh
5.38	10.00	9.75	4.38	2.00	0.00	2.00

Table 3.19 – Estimates of canopy capacity per unit of LAI [mm].

Conifer	Decid	Mixed	Muskeg	Imperv	Water	Marsh
5.58	4.29	4.33	6.22	2.79	0.00	2.79

Table 3.20 – Estimates of snowmelt rates.

Conifer	Decid	Mixed	Muskeg	Imperv	Water	Marsh
0.97	0.97	0.97	1.69	2.90	4.69	4.69

Table 3.21 – Estimates of albedo.

	Conifer	Decid	Mixed	Muskeg	Imperv	Water	Marsh
Surface	0.130	0.160	0.145	0.140	0.152	-	0.152
Max. Snow	0.500	0.500	0.500	0.500	0.780	0.780	0.780
Min. Snow	0.370	0.370	0.370	0.370	0.370	0.037	0.370

Table 3.22 – Estimates of Fast Storage Capacity.

Cover Type	Organic Layer Thickness [m]	Silty-Clay Layer Thickness [m]	Organic Layer Capacity [mm]	Silty-Clay Layer Capacity [mm]	Fast Storage Capacity [mm]
Coniferous	0.20	0.80	163	307	470
Mixed	0.12	0.88	98	338	435
Deciduous	0.12	0.88	98	338	435
Muskeg	0.30	0.70	244	269	513
Impervious	0.07	0.43	57	165	222
Marsh	1.00	0.00	813	0	813

CHAPTER 4

4 SLURP MODEL APPLICATION

4.1 Introduction

This Chapter describes the application of the SLURP model to the Taylor River and Upper Burntwood River using the parameter set established in Chapter 3. The model is calibrated and verified and the performance of the model is discussed.

4.2 Calibration Procedure

In an ideal hydrological model, all parameters would be physically based and easily measured in the field. This model in turn would not require any calibration. However, hydrological conditions across a watershed are extremely heterogeneous and an ideal state is unlikely ever to be reached. A semi-distributed approach, such as the application of SLURP in this study, lumps all of the hydrological processes of a land cover within each ASA and assumes that the hydrological processes of that area are homogeneous. This method will inherently create some error in the estimation of the runoff yield for the area of interest. Hence, all hydrological models require some degree of parameter calibration.

There exists an infinite number of parameter sets that can be used in this SLURP application, each providing a different level of modeling accuracy. However, many of these parameter sets contain parameters that are not physically realistic. The strategy used during the model calibration is to use parameter values that are within a range that is deemed to be physically realistic and reasonable. This ensures that the model maintains its physical basis.

During the calibration process presented in this study, the goodness of fit of the SLURP model is judged both graphically and statistically. If historic hydrometric data is provided to SLURP, the model computes four measures of efficiency that may be used to assess the accuracy of model output. These are:

1. Nash and Sutcliff Efficiency (NSE) (Nash & Sutcliffe, 1970)
2. Garrick Efficiency
3. Previous-Day Criterion
4. Deviation of Runoff Volumes (D_V)

The NSE and D_V will be used as statistical measures of the goodness of fit of the SLURP model results for this study. This study maintains that both the Garrick Efficiency and Previous-Day Criterion are rarely used in literature because they are too stringent and will therefore not be used in this research. Simply calibrating the SLURP model to obtain desirable NSE and D_V is a difficult and time consuming task. The following equation describes NSE:

$$NSE = 100 * \left(\frac{\left[\frac{1}{n} \sum_{i=1}^n (q_i - \bar{q})^2 \right] - \left[\frac{1}{n} \sum_{i=1}^n (q_i - c_i)^2 \right]}{\left[\frac{1}{n} \sum_{i=1}^n (q_i - \bar{q})^2 \right]} \right) \quad (4.1)$$

where q_i is the observed flow on day i , c_i is the simulated flow on day i , and $\bar{q}$ is the average measured flow over the period in which the NSE is computed. The NSE varies

between 100% and a large negative number. A value of 100% indicates a perfect fit. A value less than zero indicates that the model performs no better than using the average measured streamflow. The following equation describes D_V :

$$D_V = \frac{100 * (V_M - V_C)}{V_M} \quad (4.2)$$

where V_M is the measured volume of runoff and V_C is the computed volume of runoff over the period of interest. This criterion is simply a transformation of the mean computed and observed flows for the simulation period. It indicates the accuracy of the model in terms of a mass balance. A positive value of D_V indicates that the mean computed flow is too high indicating that there is a sink (i.e. evapotranspiration) that is not being accounted for properly. A negative value of D_V indicates that the mean computed flow is too low indicating that there is a source (i.e. precipitation) that is not being accounted for properly.

In addition to considering the NSE and D_V , the peak flow magnitude and time of peak flow will also be examined for goodness of fit. The characteristics of the peak flows will be monitored since it is possible to obtain a good NSE and D_V without a good simulation of peak flows.

Two methods of calibration are used in this research. The first method is to manually adjust the model parameters, and then execute the model, followed by a visual inspection of the resulting hydrographs for goodness of fit. The NSE, D_V , and peak flows will also be monitored to assess model results. The second method of calibration is to use the built in SCE-UA (Kite, 1998a) automatic calibration routine. According to Kite, the SCE-UA method is well suited to hydrological models as it avoids the many local minima that may exist in the solution space.

4.3 Taylor River Application

The parameter values recommended in Chapter 3 were utilized during the initial application of SLURP to the Taylor River watershed. The resulting fit for 1996 was very poor as indicated by the $NSE = -7.9\%$. The $D_V = 60.5\%$ indicates that the volume is greatly overestimated. Upon inspection of the hydrograph (see Figure 4.1), it is apparent that the base flow dominates the hydrograph and it is significantly overestimated by the model.

4.3.1 Model Calibration

During the calibration process for 1996, a number of parameters were adjusted to improve the model performance. The parameters were adjusted both manually and automatically. This parameter set is labeled the "Taylor-1996-Calibration". Figure 4.2 shows a comparison of the hydrograph produced by the Taylor-1996-Calibration parameter set and the observed hydrograph.

SLURP creates a STORAGE file that contains the values of all state variables in the model for each time step and for each land cover component of each ASA. Visual Basic code was created to summarize this large amount of data into an average value for the entire basin of interest. Thus, the model results are also interpreted by observing each state variable with time. Figure 4.3 to Figure 4.5 show the STORAGE file for Taylor-1996-Calibration.

The following subsections describe the adjustments made to the parameters in order to produce the Taylor-1996-Calibration. Table 4.1 lists the modified model parameter values the Taylor-1996-Calibration. Figure 4.6 illustrates how the parameter changes affect the simulated hydrograph.

(a) Retention Constant of Slow Storage Tank (k_2)

The optimum value of k_2 for the Forest covers is 517.3, the highest of all covers. The value of k_2 for the other covers is 393.2 for Muskeg, 262.5 for Impervious, 3 for Lakes, and 50 for Marsh. This parameter change modified the hydrograph by increasing interflow, increasing the slope of the recession limb of the hydrograph,

and decreasing baseflow. Figure 4.3 shows that the recession in Slow Store Contents is sufficient to produce adequate baseflow during winter months. As shown in Figure 4.6, a larger k_2 overestimates baseflow during winter months because too much water is kept in deep storage during the summer months for eventual release during the winter months. This in turn causes a slower recession in baseflow through the fall and into the winter. Figure 4.6 also shows that keeping too much water in deep storage during the spring snowmelt period results in a stunted peak flow and a recession limb that is too gradual. As a result, the discharge during the summer slightly exceeds that of the 1996-Taylor-Calibration simulation. The change in this parameter creates a 9.6% improvement to the NSE and no improvement to the D_V .

(b) Maximum Capacity of Slow Storage Tank (SSC)

The SCC was reduced by approximately an order of magnitude for most covers. The reduction in storage capacity greatly improves the snowmelt runoff portion of the hydrograph by limiting deep ground water storage and generating significantly more snowmelt runoff through interflow. The drawback is an over prediction of discharge during the fall rain events indicating a lack of subsurface storage in the model during this period (see Figure 4.6). The reduced capacity also improves the baseflow during the winter months due to a reduction in available water for this period. The change in this parameter creates a 37.7% improvement to the NSE and a 9.1% improvement to the D_V .

(c) Maximum Capacity of Fast Storage Tank ($S_{1,max}$)

This parameter required relatively little change and was reduced by approximately 13%. It was found that decreasing this parameter improved the snowmelt runoff portion of the hydrograph by creating less available groundwater storage. However, the decreased available storage degraded the simulation of the summer and fall events where increased upper storage was necessary in order to absorb all but the largest rain events. Hence, increasing/decreasing this parameter improves/degrades the simulation of discharge during the snowmelt period and has

the reverse effect during the fall period. The simulation error that occurs during the snowmelt period may be due to the effects of frozen soil rather than the effect of storage, hence the capacity of the Fast Storage Tank is not reduced to improve the snowmelt runoff modelling. The change in this parameter creates no improvement to the NSE and no improvement to the D_V .

(d) Retention Constant of Fast Storage Tank (k_1)

The value of k_1 was reduced by approximately an order of magnitude for most covers. As shown in Figure 4.6, the reduction in k_1 causes the hydrograph peaks to be more pronounced and allows the model to replicate the flashyness of the watershed by significantly increasing the interflow. The drawback is an over prediction of discharge during the fall rain events indicating that too much interflow is being generated during this period rather than keeping it in storage. The change in this parameter creates a 54.4% improvement to the NSE and a 6.9% improvement to the D_V .

(e) January and July Snowmelt Rates ($R_{1,Jan}$ & $R_{1,Jul}$)

The observed hydrograph does not respond to the snowmelt until May 2 although, there are four three-day periods of above zero temperatures prior to May 2. The later response is partly due to the transient time of the snowmelt runoff being routed to the outlet. The main reason is likely that the snow pack must first ripen to a point where additional energy generates snowmelt that exits the snowpack. This process is not represented in SLURP and results in a significant amount of snowmelt generation prior to May 2 as illustrated in Figure 4.6.

The premature generation of snowmelt was somewhat alleviated by adjusting the snowmelt parameters to artificially delay snowmelt. SLURP uses a parabolic interpolation between the January 1 and July 1 snowmelt rates to establish daily snowmelt rates. This interpolation allowed the onset of snowmelt to be delayed by setting the January rate to 0 and the July rate to 0.5. As shown in Figure 4.2, these values eliminated the majority of the runoff due to snowmelt prior to May 2 and

generated the majority of snowmelt after May 2. However Figure 4.4 shows that the snowmelt period is still drawn out over a lengthy period because the lower snowmelt rates deplete the snow pack relatively slowly. The last day of simulated snowmelt is June 6, 1996, 12 days after the observed peak discharge. Additionally, the snowmelt rates will likely present a problem during the model validation phase as the date of initial snowmelt runoff varies from year to year, thus requiring different snowmelt rates to control the timing of initial snowmelt. The change in this parameter creates a 15.5% improvement to the NSE and a 10.0% improvement to the D_V .

(f) Interception Coefficient A

Although the sensitivity analysis determined that the Interception Coefficient A has a low sensitivity, it was determined that interception has a profound affect on the hydrograph. The lower Interception Coefficient A suggested by literature effectively reduces interception and increases inputs to the Fast and Slow storage tanks. The result is a hydrograph that is far too peaky (see Figure 4.6). Additionally, the total ET is reduced as there is far less moisture available in the canopy for evaporation. This in turn results in a significant overestimation of basin runoff. The Interception coefficient A parameter was doubled for all land covers. The change in this parameter creates a 15.2% improvement to the NSE and a 40.5% improvement to the D_V .

(g) Initial Contents of Slow Storage Tank

The Initial contents of the slow storage tank were manually adjusted until the initial simulated flow matched the initial observed flow. Initial contents of Fast Storage cannot be specified within SLURP.

(h) Initial Contents of Snow Storage

The start date for the model simulation was chosen to be September 1 so that the initial contents of the snow storage tank could be set to zero.

(i) Routing Coefficients

Adjustments to the Muskingum routing coefficients improved the simulation performance. The new routing coefficients are listed in Table 4.2.

The different retention constants for the Forest and Muskeg covers is appropriate considering the inability of Muskeg to maintain an elevated water level due to the large variation in hydraulic conductivity with depth as described in Section 3.3.3.8. The low retention constants for the Impervious cover are suitable considering the rapid movement of water through the thin organic and silty clay layers and laterally along the bedrock interface. The retention constants and zero fast storage capacity for the Water are model defaults. The retention constants for the Marsh cover were arbitrarily chosen to be larger than water but somewhat smaller than all other covers due to the very thick and highly porous organic layer characteristic of this cover.

Overall, the Taylor-1996-Calibration produced an $NSE=74.5\%$ and $D_V = 21.8\%$. The NSE is considered to be satisfactory considering the deficiencies observed in the simulated hydrograph. Of the 25.5% deficiency in NSE , 14.7% is due to the snowmelt runoff period (May 17, 1996 – June 30, 1996) and another 8.4% is due to modelling error during the fall period, specifically September 9, 1996 – October 31, 1996. The annual D_V indicates that the annual volume is over estimated by SLURP. However if the D_V is computed over shorter periods of time the resulting D_V is even worse. The D_V is -17.5% (-22.2 mm) for the snowmelt runoff period and D_V is 485.3% (+37.2mm) for the fall period. The two deviations in volume nearly cancel one another out, however the error in volume during the fall period is the most prevalent source of error.

The simulated spring peak discharge of 36.8 cms is 10.5 cms or 22.1% lower than the observed spring peak discharge of 47.3 cms. The simulated time of peak for the spring freshet is 9 days later than the observed. The peak flows during the fall months are greatly over estimated by SLURP. The simulated fall peak discharge of 15.1 cms is 10.7

cms or 1084.6% greater than the observed fall peak discharge of 4.34 cms. The time of peak is 14 days sooner than the observed time of peak during the fall months.

4.3.2 Model Validation

The SLURP model of the Taylor River basin was validated on all years between 1985 and 2000 where historical discharge measurements permit. In order to condense the discussion, a set of five years of varied climatic conditions is closely examined. These are 1986 and 1997 which are considered to be wet; 1991 which is considered to be average; and 1992 and 1995 which are considered to be dry. Annual statistics were determined for all years except 1988-1991 because they are missing lengthy periods of discharge data. Annual statistics are not computed for 1999 since the recorded hydrograph appears to be erroneous.

Upon examining the values of D_v for the 1985-2000 period it was found that the 1996 calibration parameters overestimated the volume of runoff for the majority of the verification years during both the spring and summer periods. This suggests that the model parameters are more suitable for wet years than average and dry years. Although 1996 is an average precipitation year, the snowmelt generated peak flow is the largest of all years. The SLURP model is not removing the required volume of water from the system through the process of ET.

The ET parameters were increased manually until the D_v for all test years was in best agreement with the observed D_v . The new ET parameters are listed in Table 4.3. The increase in ET parameters adversely affected the 1996 simulation hydrograph by increasing the runoff volume during the spring freshet. The change in simulated streamflow due to the adjusted ET parameters was not significant enough to warrant the adjustment of the other model parameters. Hence, this parameter set is considered to be the optimum parameter set for the Taylor River for the meteorological conditions present in the 1985-2000 period. This parameter set will be named the "Taylor-Calibration".

Figure 4.21 show the simulated and recorded hydrographs for the test years. Table 4.4 lists the performance statistics of the test years. Overall the model performed satisfactorily given the wide range of climatic conditions in the test years, given that most parameters were calibrated for a single year, and given that the model was run as a single sequence from September 1, 1984 to December 31, 2000. The average annual NSE for 11 test years is 61% and six of these years have an NSE of 70% or higher, and eight of these years are 50% or higher. The average annual D_v for the 11 test years is $\pm 12\%$ and five of these years have a D_v that is $\pm 15\%$ or less.

The volume of runoff generated in the summer and fall months and the corresponding peaks were well simulated for most of the wet years (1986, 1997) and significantly over estimated for the majority of the dry years (1989-1992, 1994, 1996, 2000). The lack of surface and subsurface storage following the spring freshet is a potential source of error that contributes to this error. The fact that this error is more pronounced during the dryer years illustrates the ability of Muskeg to greatly limit interflow once the water level reaches a minimal elevation where the hydraulic conductivity is exponentially lower than at higher elevations.

In 1992, the majority of the error occurs during the August-November period. The observed flow never exceeds 2 cms and shows little or no response to rain events. The over prediction in late summer and fall time flows is similar to the 1996 calibration period. It is mainly due to the lack of available surface and subsurface storage that is required to absorb the rain events. The observed hydrograph shows very little response to any of the 10 mm and smaller rain events. One difference between the 1992 and 1996 rain events is that most of the 1992 rain events are spread out in time where as the 1996 events are often grouped together. The grouped rain events fill the canopy and a portion of the upper soil zone. The lack of capacity results in the generation of runoff once subsequent rain events occur.

The 1995 simulation is among the worst simulations within the 16 test years. The annual NSE of 58% is quite high considering the annual D_v of 31% is the worst of all 15 test

years. The April-May NSE and D_v are -280% and 126% respectively. This indicates that a major hydrologic process is not being represented properly within the SLURP model. Unique to this year is the unusually low spring discharge volume and peak discharge of 5.5 cms. The corresponding simulated discharge of 12.5 cms is significantly larger. The snow pack during 1995 is comparable to the 1986 snow pack however 1986 has a spring runoff volume several orders of magnitude higher than 1995. The lack of runoff during the spring of 1995 is possibly due to the occurrence of very dry conditions in the fall of 1994 resulting in low soil moisture. The result could be no change or possibly increased infiltration capacities that are characteristic of friable or honey comb permafrost coupled with significant available soil moisture storage capacity. The over estimation of flow during the spring freshet and during the summer and fall periods combine to create the low NSE and high D_v during April and May.

The observed hydrographs for all verification years show a sudden response to snowmelt resulting in a steep rising limb. All of the simulated hydrographs show some response to snowmelt well in advance of the observed hydrograph response to snowmelt. As described earlier, the delay in the observed streamflow response to the snowmelt is characteristic of a ripening snow pack. The Degree-Day snowmelt algorithm included in SLURP generates snowmelt and routes it to the fast storage tank on any day where the temperature exceeds zero degrees.

The observed 1996 hydrograph has a very steep rising limb and a significantly large discharge peak relative to the other test years. While snow ripening likely contributes to the intense snowmelt, it is likely that the rain on snow events that occur during May of 1996 also contribute to the intense snowmelt.

During all years except 1995, the timing of the simulated spring discharge peak is later than the observed spring discharge peak. In nearly all test years the simulated spring freshet discharge peak coincides with the final snowpack depletion. It is likely that the simulated snow pack is not depleting as quickly as it is in nature. This problem is inherent to the date dependent snowmelt rates that SLURP uses. The snowmelt rates

were lowered to alleviate the onset of snowmelt runoff however it is apparent that this has the negative side effect of prolonging the snowmelt runoff period.

The recession rate during the spring freshet and during large summer and fall rain events matches very well in nearly all validation years. This indicates that interflow and percolation are being modelled with reasonable accuracy. The August to December period of 1997 was a very wet period with some very large rain events. The resulting discharges reach 30cms which is higher than any other year. The simulation fits extremely well over this period as indicated by the NSE of 90% and D_v of -6%. The time to peak and the peak flow magnitudes match well. The steep discharge peaks and extended recession limbs are simulated well particularly during the October 14 to December 31 period where the fast storage tank depletes and the discharge approaches base flow conditions.

4.4 Upper Burntwood River Application

The Taylor-Calibration parameters were applied to the Upper Burntwood River for the same calibration and validation periods used for the Taylor River simulations. Table 4.5 lists the performance statistics for this simulation. The NSE and D_v are not as good as the Taylor River simulations for the common test years. Of the test years, 13 of 16 over estimate the volume of runoff indicating insufficient ET. This indicates that there is a significant difference in the hydrological processes present in the Upper Burntwood River and Taylor River basins.

The average annual yield for the Taylor River and the Upper Burntwood River were compared assuming that differences in meteorological conditions are negligible. The average annual yield for each year between 1984 and 2000 is 142.7mm for the Taylor River and 120.3mm for the Upper Burntwood River. This indicates that the ET losses are roughly 20% greater in the Upper Burntwood river basin relative to the Taylor River basin. This may be due to the occurrence of significant lake evaporation in the Burntwood Lake, located in ASA6, and the many other lakes located in the headwaters of this basin, specifically ASA8-11. Many of these lakes receive a continuous supply of

water from the upstream headwaters of the basin. Hence, evaporation from these lakes is generally not limited by water availability and is likely to be near the potential evaporation. The same can be said about the marsh land cover that makes up much of the riparian zone along rivers and lakes. The marshes also receive inflow from the upstream headwaters and exchange water with the adjacent lakes and rivers. Accordingly, SLURP determines too little evaporation because it computes all processes at a single point whereby the only water that the model provides for ET processes is that precipitation that falls directly on the area of interest. It excludes the contribution of water from the upstream headwaters. All water that is in transit and is being routed between ASAs is not available for ET within the SLURP model.

It can be shown that SLURP is not representing the ET processes of the Water cover accurately if it is assumed that rivers and lakes are never limited by the availability of water. Figure 4.22 shows the average ASA total evaporation losses for the water and marsh covers of the Taylor River 1996 Calibration file. The evaporation curve for the Water cover contains numerous periods of zero evaporation. This occurs because the Slow Storage tank, the only storage tank that is represented in the Water cover, depletes entirely between the rainfall events that are also shown in Figure 4.22. The Marsh cover exhibits lower daily ET rates relative to the Water cover because the ET rate of this cover is controlled by the wilting point and field capacity of the soil. Water in the marsh is never fully depleted between rainfall events resulting in continuous ET during the summer months.

Another mechanism that may contribute to higher ET losses from the Upper Burntwood river relative to the Taylor river is the presence of significantly more depression storage. This depression storage that is characteristic of the hummocky terrain may not be hydraulically connected to the rest of the basin at all times resulting in greater evaporative losses. The SLURP model does not simulate ET losses from depression storage.

4.4.1 Model Calibration

In order to account for the processes that contribute to the elevated ET rates within the Burntwood River basin, the ET parameters: Alphas and Betas, were increased for all covers. As a result of the model calibration, the optimum ET parameter values are listed in Table 4.6. These are approximately 29% higher than the ET parameters in the Taylor River calibration.

The headwaters of the Burntwood River contain the very large Burntwood Lake as well as many other large lakes which attenuate the hydrographs. A single set of stage-discharge and stage-volume curves were developed to dampen the peak flows and high recession rates that are present in the simulations. The volume curve, discharge curve, and Muskingum routing coefficients were manually adjusted until the simulated hydrographs displayed a reasonable fit to the observed hydrographs. The resulting curves are shown here:

$$Q = 200*(EL - 271)^{0.49} \quad (4.3)$$

$$Vol = 400*(EL - 271)^{0.99} \quad (4.4)$$

where **Q** is reservoir outflow [cms], **EL** is the elevation of the lake [m], and **Vol** is the volume of the lake [Million m³]. The calibrated Muskingum routing coefficients are listed Table 4.7.

Other model parameters including the Fast and Slow Storage parameters were varied to test for modeling improvements. The simulation performance did not improve significantly and did not warrant the establishment of a unique parameter set for the Upper Burntwood River basin.

4.4.2 Model Validation

Overall, the model performed satisfactorily given the wide range of climatic conditions in the test years and the fact that the majority of model parameters are the same as those

used in the Taylor River model. The only parameter differences are the higher ET parameters and the lake routing scheme. Table 4.8 lists the performance statistics for all test periods. The average annual NSE of the 16 test years is 37% and five of these years have an NSE of 70% or higher, seven of these years are 60% or higher, and eight of these years are 50% or higher. The average annual D_V of the 16 test years is $\pm 21\%$ and nine of these test years are $\pm 15\%$ or less. The results improve significantly if the test years are limited to the same 11 test years used for the Taylor River. The NSE improves to 54% and D_V improves to $\pm 14\%$ which is comparable to the Taylor River performance. The two years with the worst performance were 1988 and 1989, which were not included in the Taylor River verification. The runoff is significantly overestimated for these two years. The average performance statistics for the Taylor River simulation would likely deteriorate if the performance statistics could be determined for all 16 test years.

Figure 4.23 to Figure 4.38 show the simulated and recorded hydrographs for the test periods. The modelling errors observed in the Upper Burntwood River simulated hydrographs are similar to those observed in the Taylor River simulated hydrographs. These include:

- volume of runoff and peak flows are underestimated during the spring freshet for the majority of the test years
- over estimation of flows in late summer for the majority of the dry years (1988-1991, 1994, 1996, 2000)
- premature generation of snowmelt runoff in the majority of the test years
- timing of the spring freshet peak is delayed in the majority of the test years

Potential causes for these deficiencies are discussed in the Taylor River Model Validation Section 4.3.2.

The rate of recession is modeled reasonably well in the Upper Burntwood simulations. In particular, the 1992 hydrograph displays good recession rates during the spring freshet as well as the fall receding limbs. Additionally, the receding limb of the 1995 spring freshet

correlates well with the observed hydrograph. The receding limb of the 1997 fall hydrograph correlates well with the observed hydrograph.

The two test years with the best modeling efficiencies were 1991 which is an average year and 1995 which is a dry year. This is opposite to the Taylor River simulations which found 1995 to be the worst simulation. The 1995 simulation shows good agreement with the observed hydrograph during late summer months and a satisfactory fit during the spring freshet.

The 1989 simulation is among the worst simulations as indicated by the annual NSE of -193% and D_v of -61 which are the worst statistics of all 16 test years. The main source of error is the significant overestimation of flows in the late summer. During this period the peak observed flow of 27 cms is grossly overestimated by SLURP as 82 cms. This type of error is also observed in the 1996 and 1994 test years but the statistics for those years are significantly better as the average annual flow of 1996 and 1994 are significantly greater than the average annual flow of 1989.

4.5 Water Balance Calculations

A water balance calculation is a necessary component of hydrological modelling applications. The water balance calculations consider all water flow across the system boundaries including all flow into and out of the system as well as the storage within the system. The calculation indicates whether or not there is error within the modelling code which causes water to be systematically destroyed or created. Figure 4.39 and Figure 4.40 illustrate that over the long term $P - E_A$ (Precipitation – Evapotranspiration) is equal to $R + \Delta S$ (Runoff – Change in System Storage) and that the model appears to be accurately balancing water for the both the Taylor River and Upper Burntwood River. There is less than 2% error in the water balance for both applications. Table 4.9 provides the average annual water balance components of the SLURP application to the Taylor River and Upper Burntwood River.

4.6 Chapter Summary

This chapter presented the application of the SLURP model to the Taylor and Upper Burntwood River for a continuous streamflow simulation over the 1985-2000 period. The performance of the model was evaluated against the observed hydrographs.

The initial model application to the Taylor River included those parameters based on literature and resulted in a poor model performance. Relatively few model parameters were calibrated to achieve satisfactory modelling results. These parameters included the Fast and Slow Storage Tank Capacities and Retention Constants, the January and July snowmelt rates, and the Interception Coefficient A, and the ET Apha and Beta Coefficients.

Overall the model performed satisfactorily for the Taylor River and Upper Burntwood River Basins considering the general quality and quantity of data used in the simulations. The 16 test years contain a diverse set of climatic conditions. Nearly all model parameters are common for the Taylor River and Upper Burntwood River models with the exception of the higher ET coefficients and the Lake routing used in the Upper Burntwood. The simulation statistics for the Upper Burntwood River are not as good as those for the Taylor River but are respectable.

The simulation hydrographs displayed a number of modeling deficiencies. These include:

- the volume of runoff and peak flows are underestimated during the spring freshet for the majority of the test years
- the flows in late summer are over estimation for the majority of the dry years
- there is premature generation of snowmelt runoff in the majority of the test years
- the timing of the spring freshet peak is delayed in the majority of the test years

Figure 4.1 - 1996 recorded Taylor River hydrograph vs. SLURP simulation using initial parameter set estimated from literature.

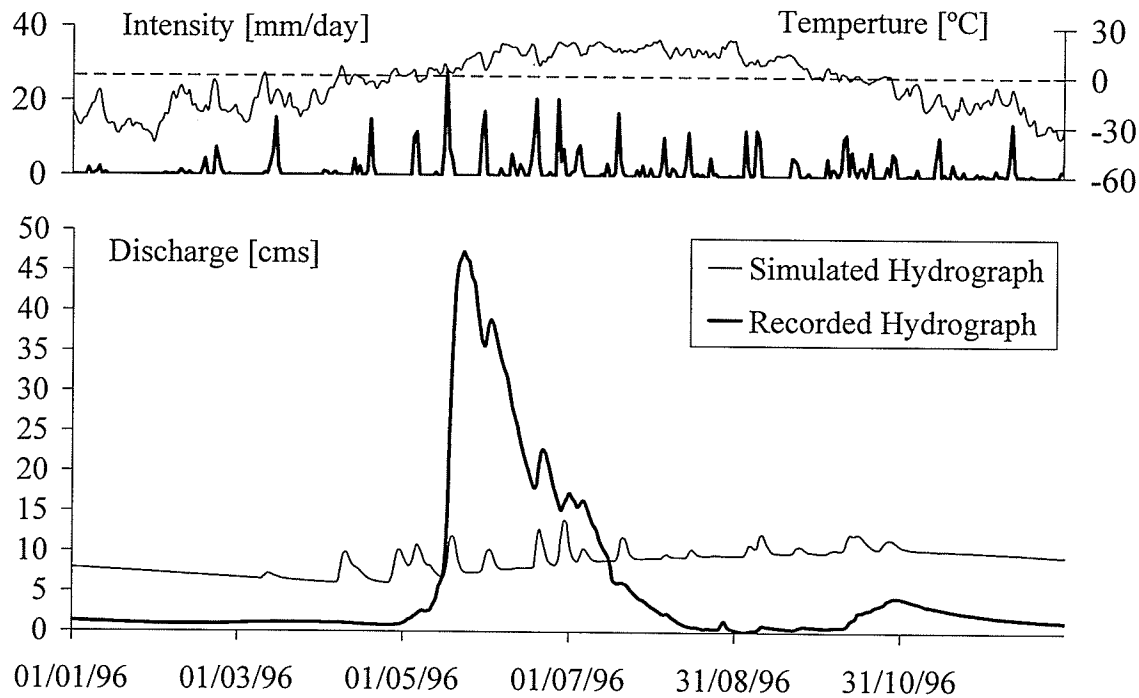


Figure 4.2 - 1996 recorded Taylor River hydrograph vs. SLURP simulation using "Taylor-1996-Calibration".

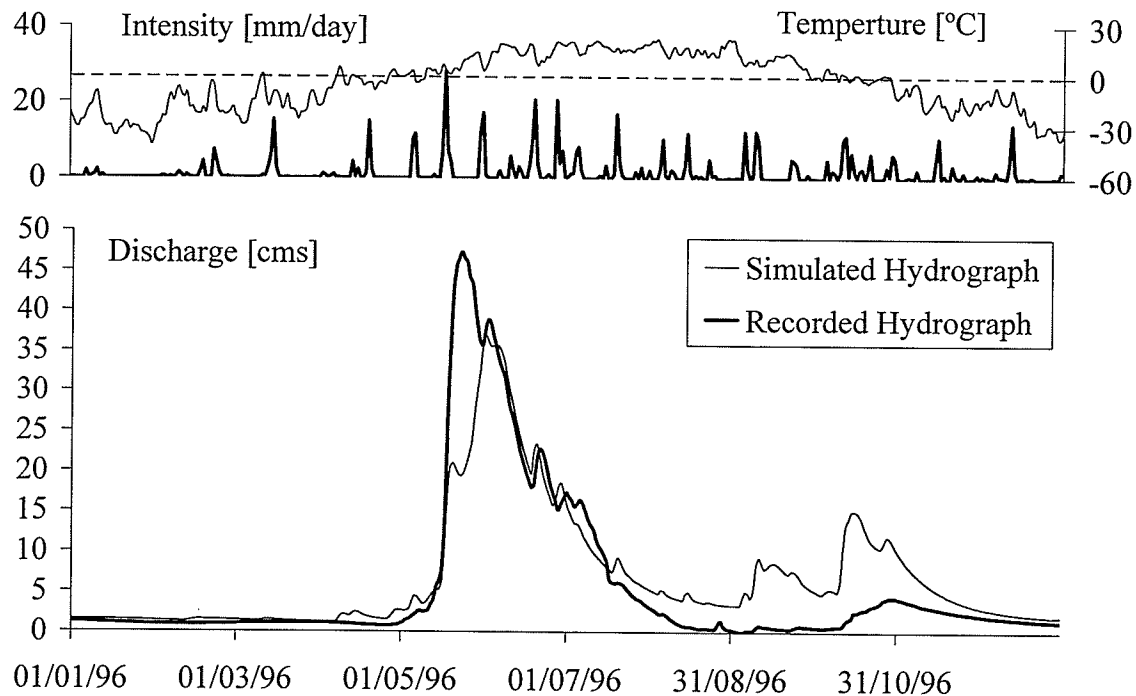


Figure 4.3 - 1996 Taylor River Calibration STORAGE file with "Taylor-1996-Calibration" parameters (1 of 3).

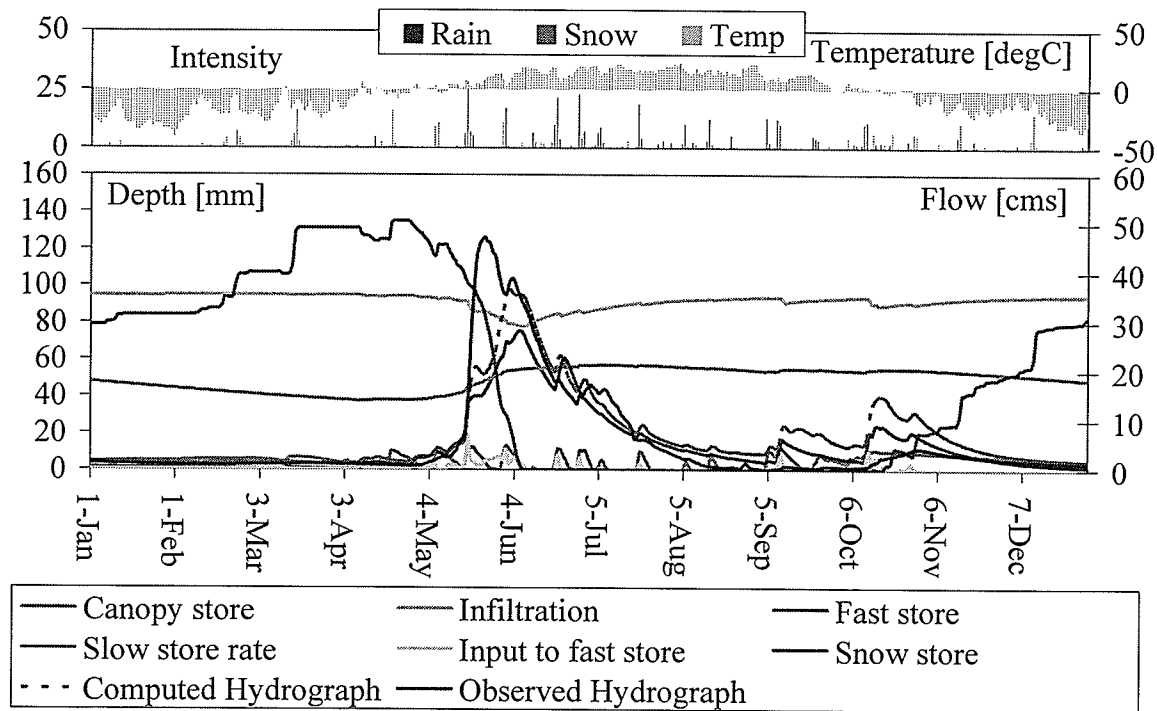


Figure 4.4 - 1996 Taylor River Calibration STORAGE file with "Taylor-1996-Calibration" parameters (2 of 3).

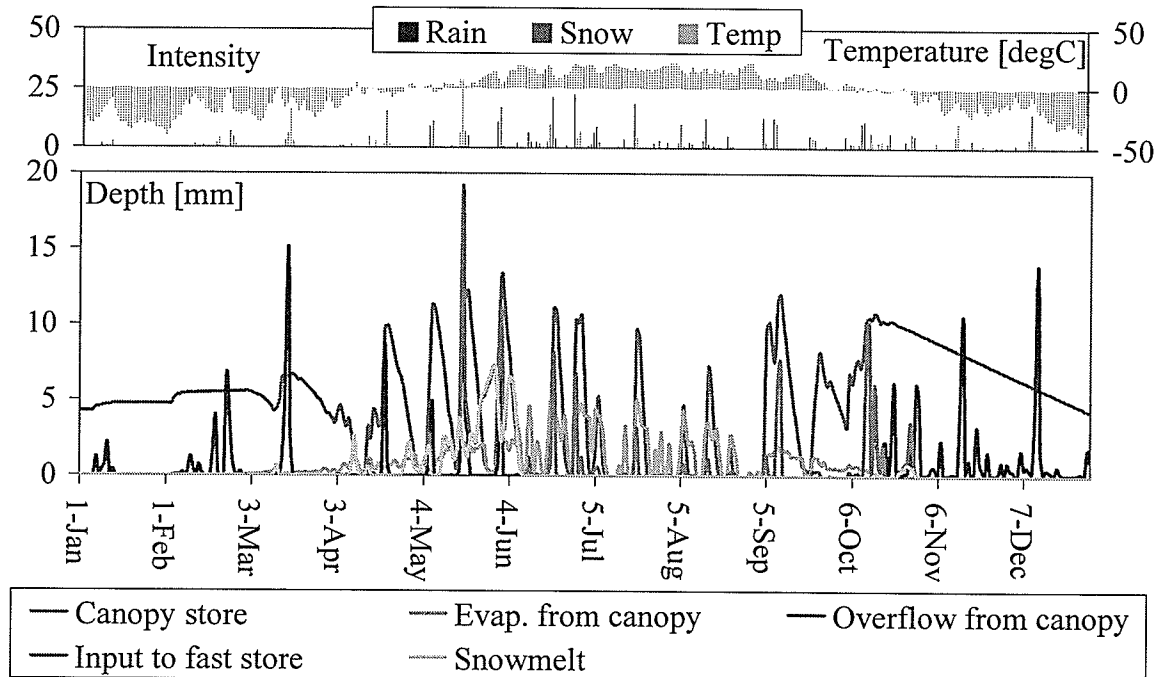


Figure 4.5 - 1996 Taylor River Calibration STORAGE file with "Taylor-1996-Calibration" parameters (3 of 3).

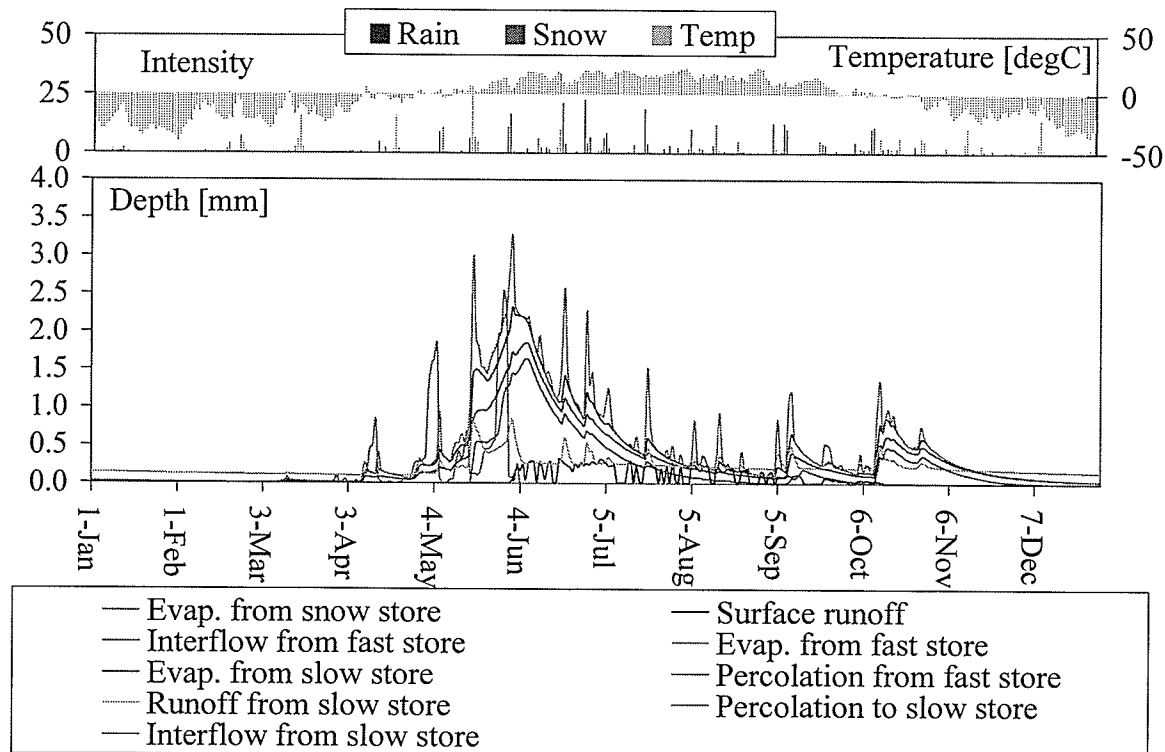


Figure 4.6 - 1996 Observed Taylor River Hydrograph vs. Parameter adjustments to parameter set estimated from literature.

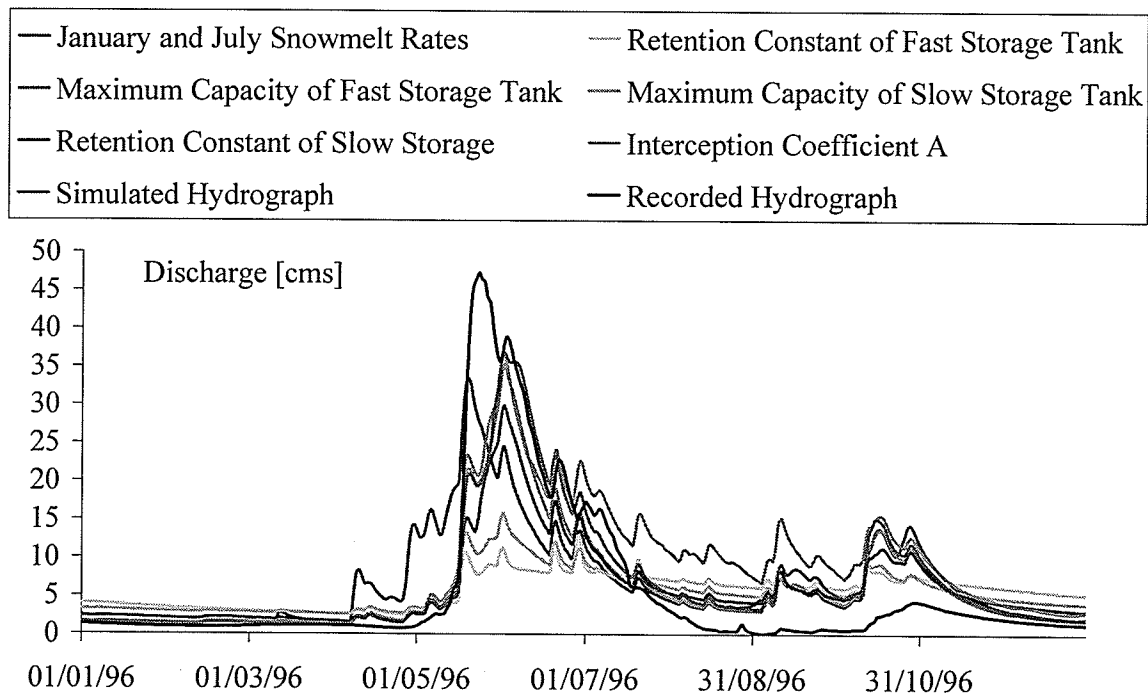


Figure 4.7 - 1985 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

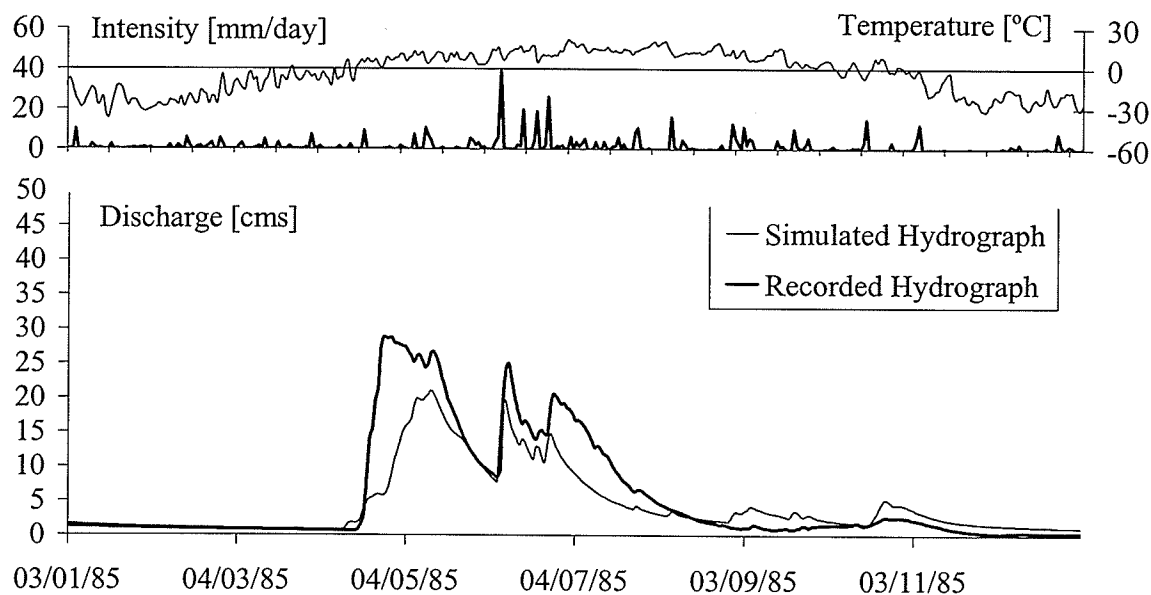


Figure 4.8 - 1986 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

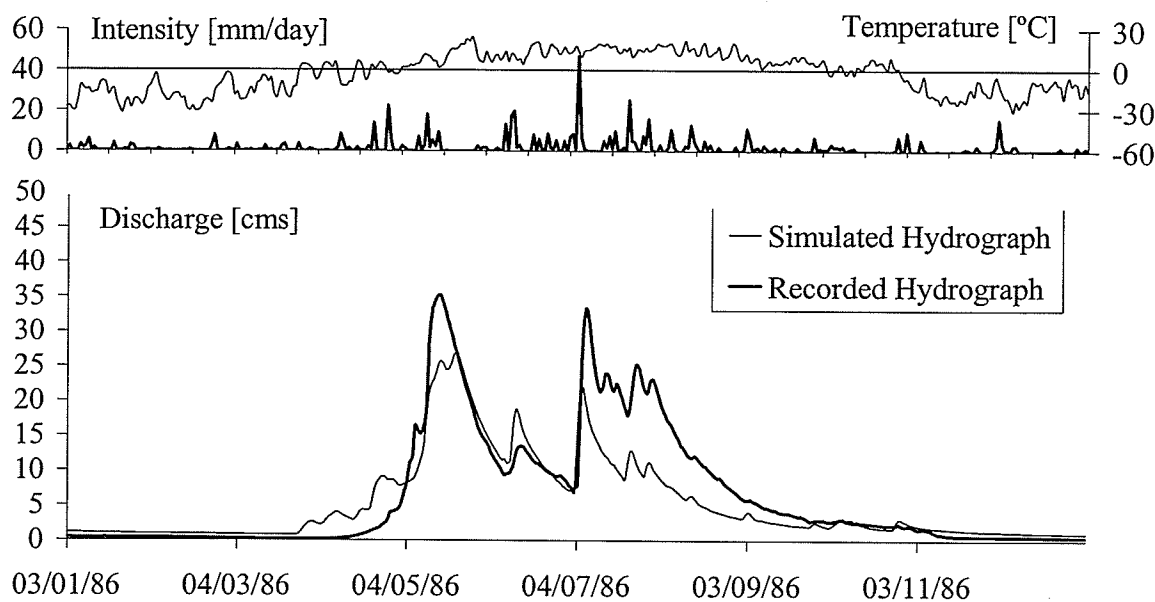


Figure 4.9 - 1987 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

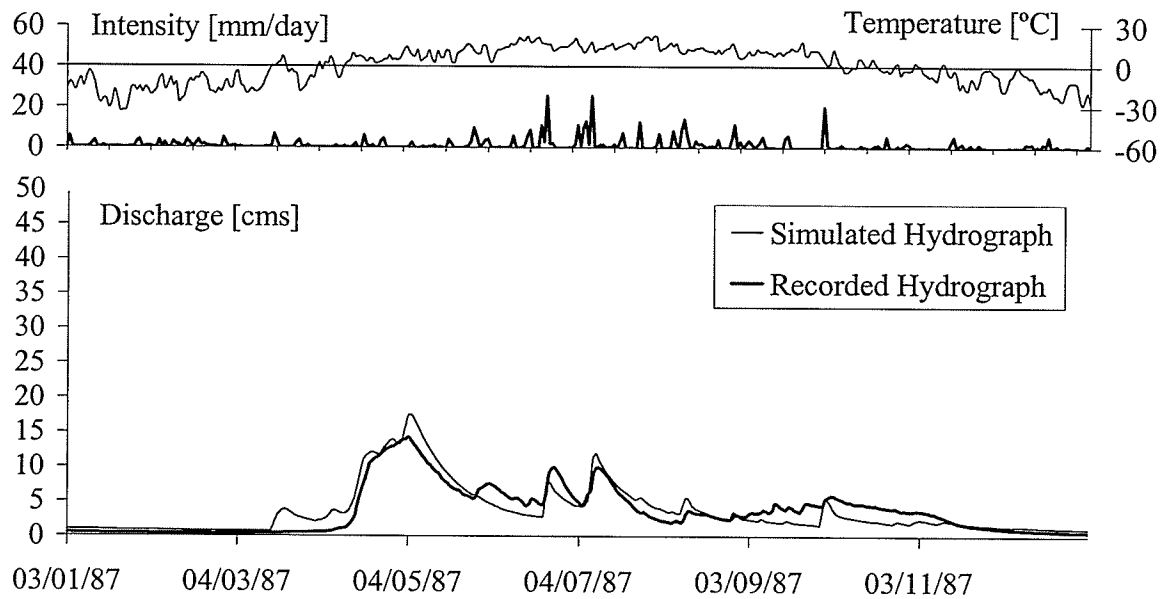


Figure 4.10 - 1988 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

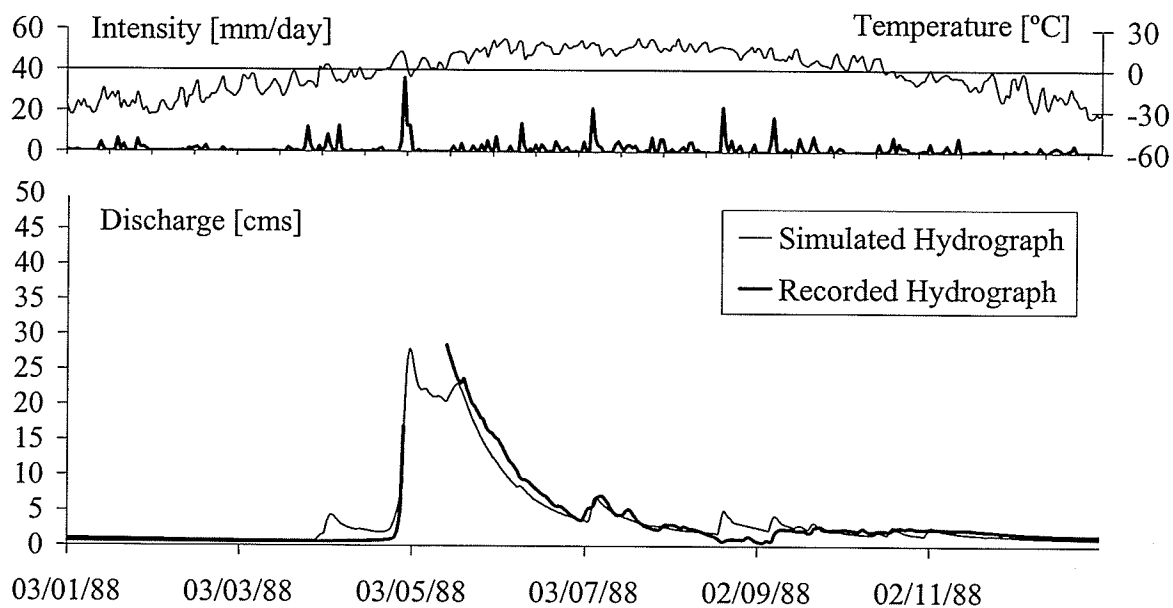


Figure 4.11 - 1989 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

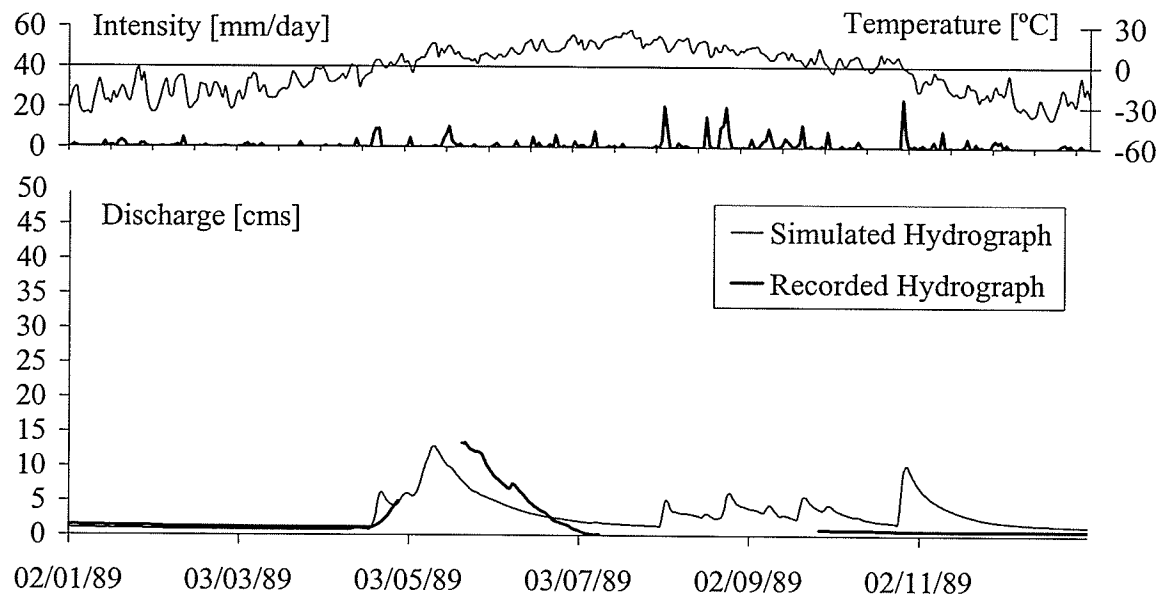


Figure 4.12 - 1990 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

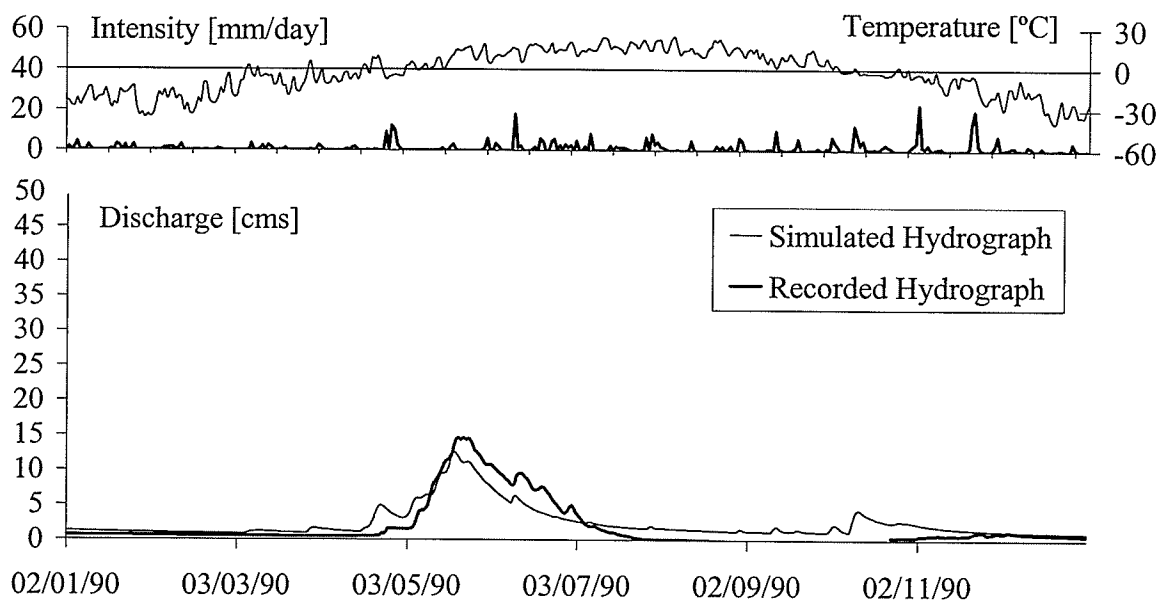


Figure 4.13 - 1991 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

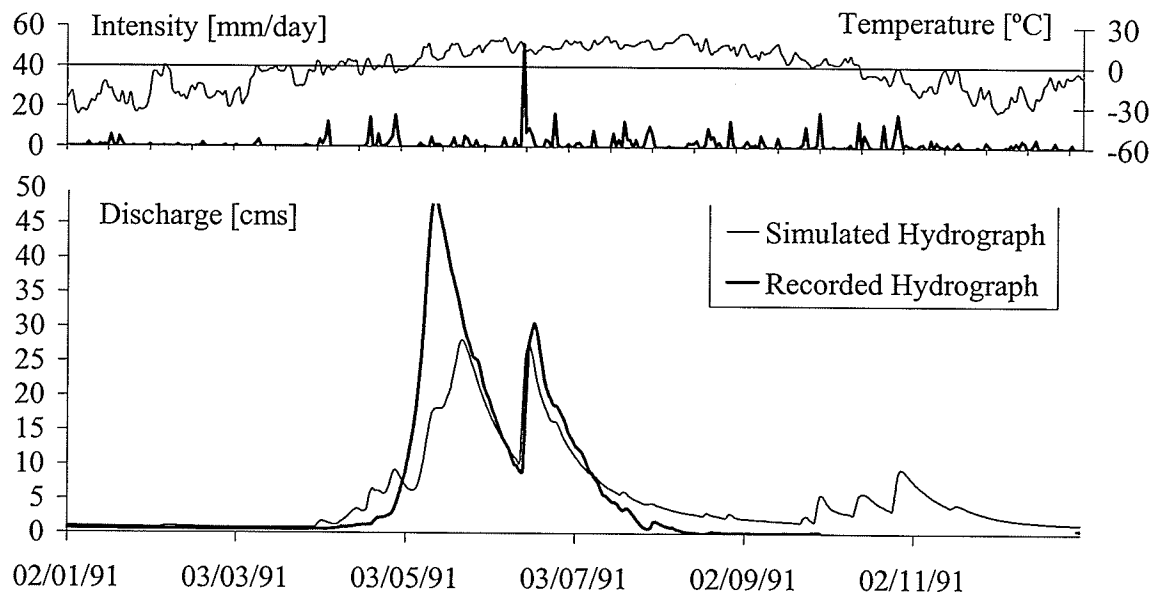


Figure 4.14 - 1992 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

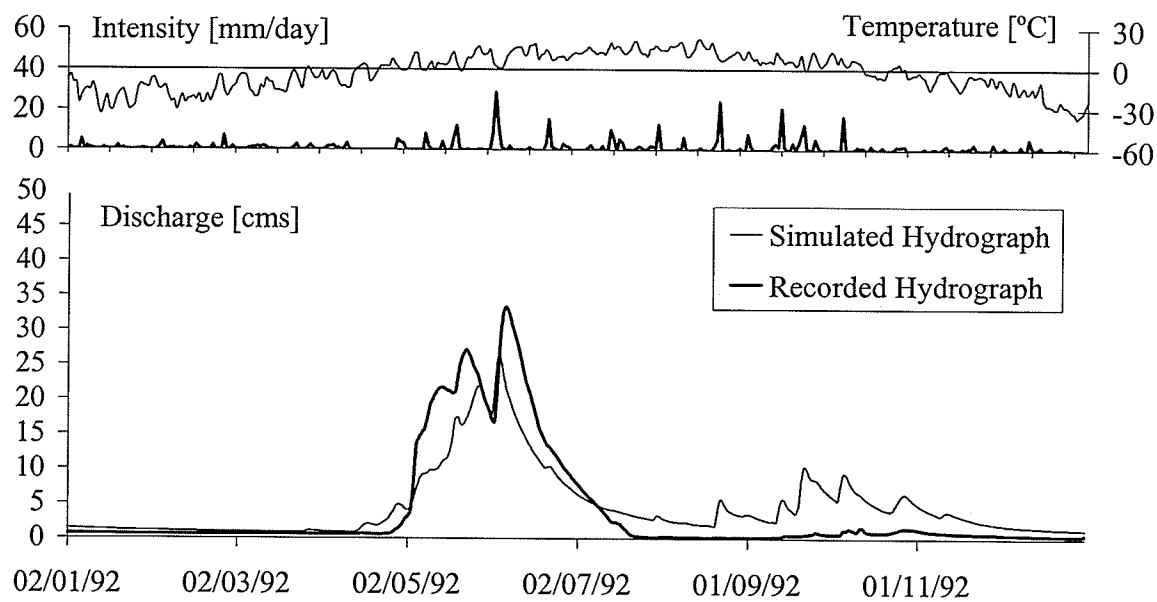


Figure 4.15 - 1993 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

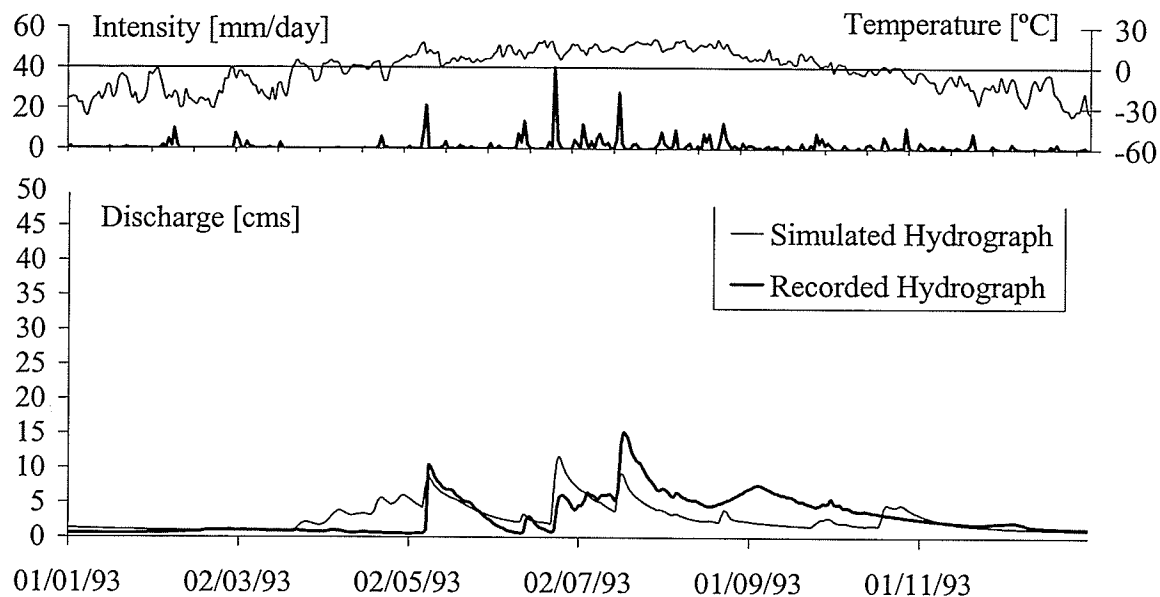


Figure 4.16 - 1994 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

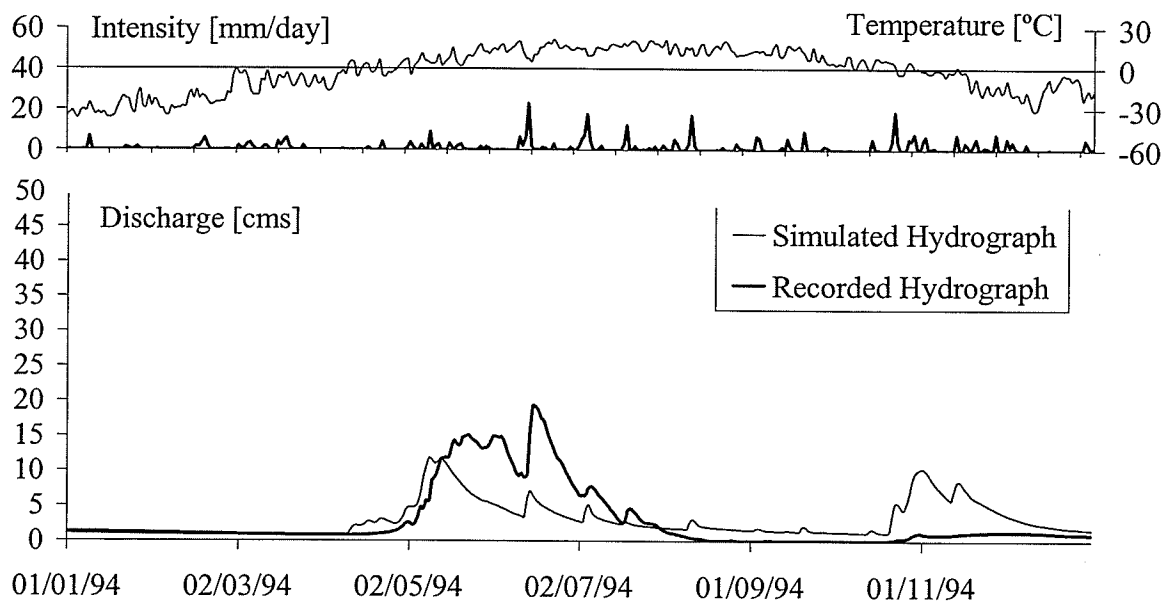


Figure 4.17 - 1995 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

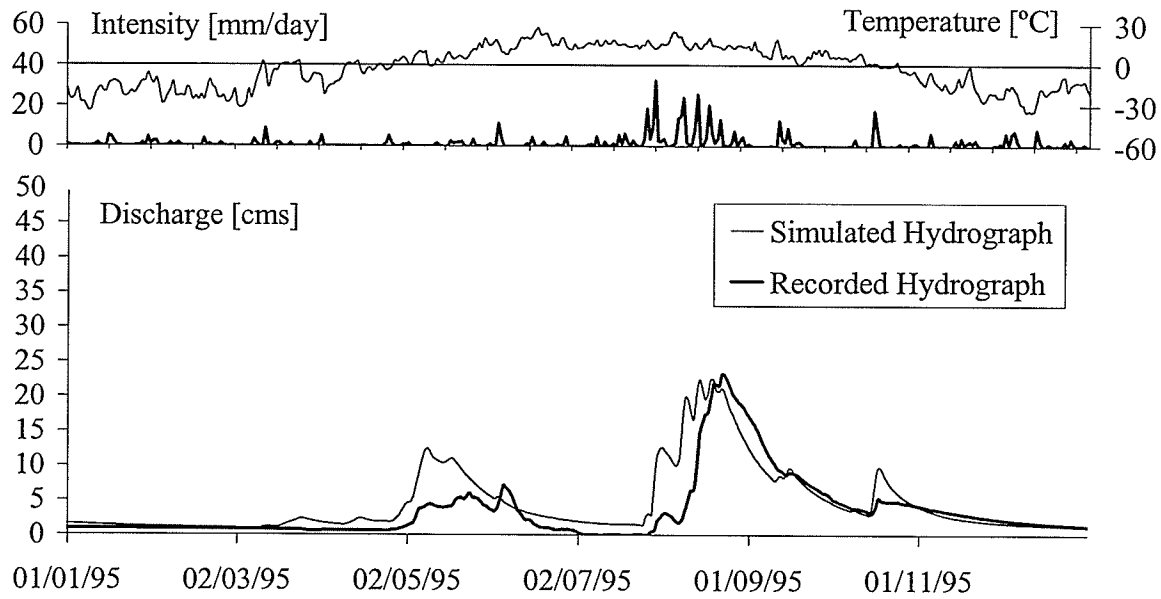


Figure 4.18 - 1996 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

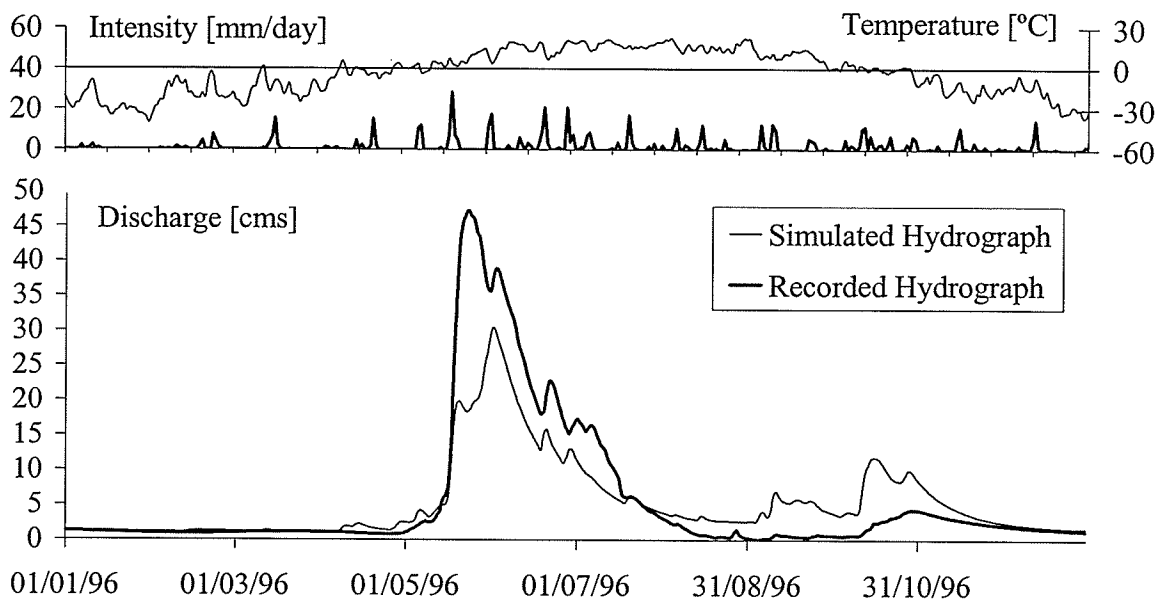


Figure 4.19 - 1997 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

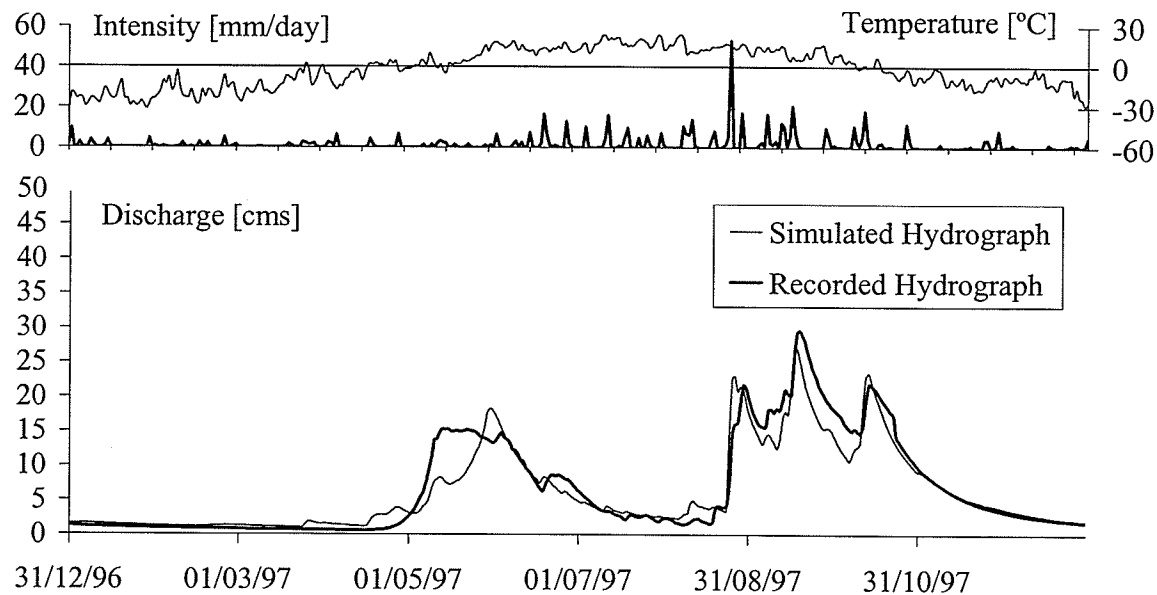


Figure 4.20 - 1998 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

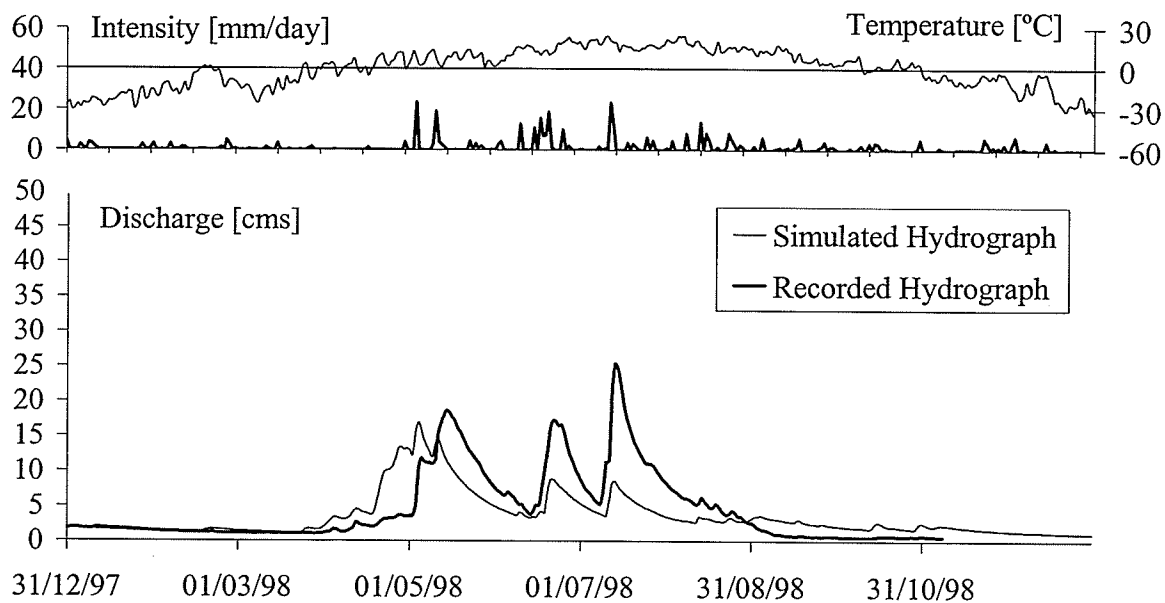


Figure 4.21 - 2000 Taylor River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

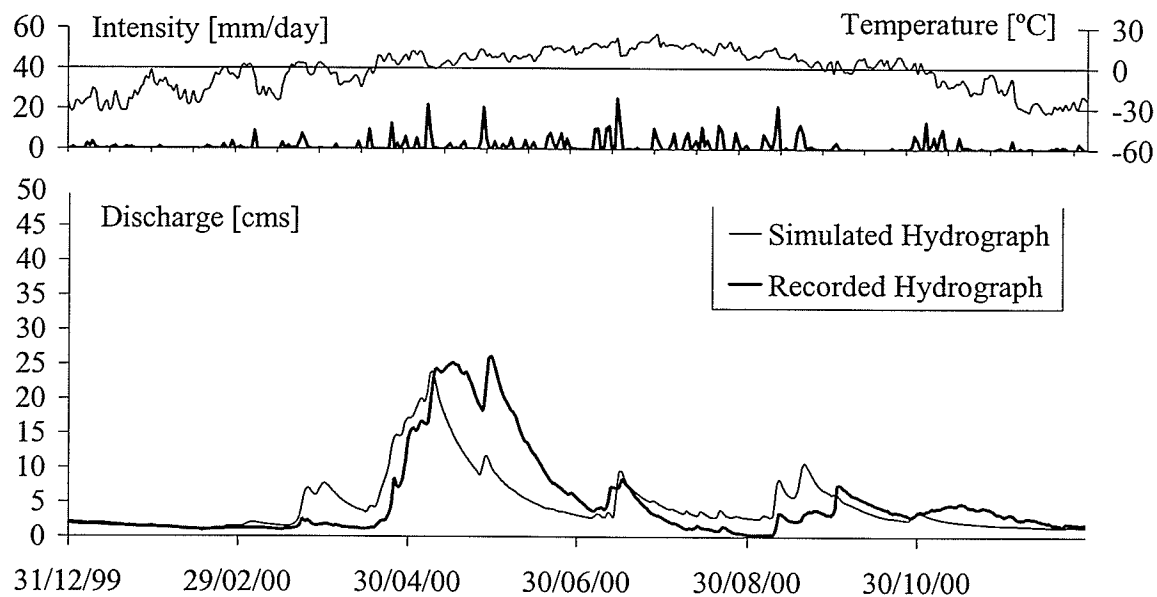


Figure 4.22 - Simulated evaporation for Water & Marsh land covers for the Taylor River Basin during 1996.

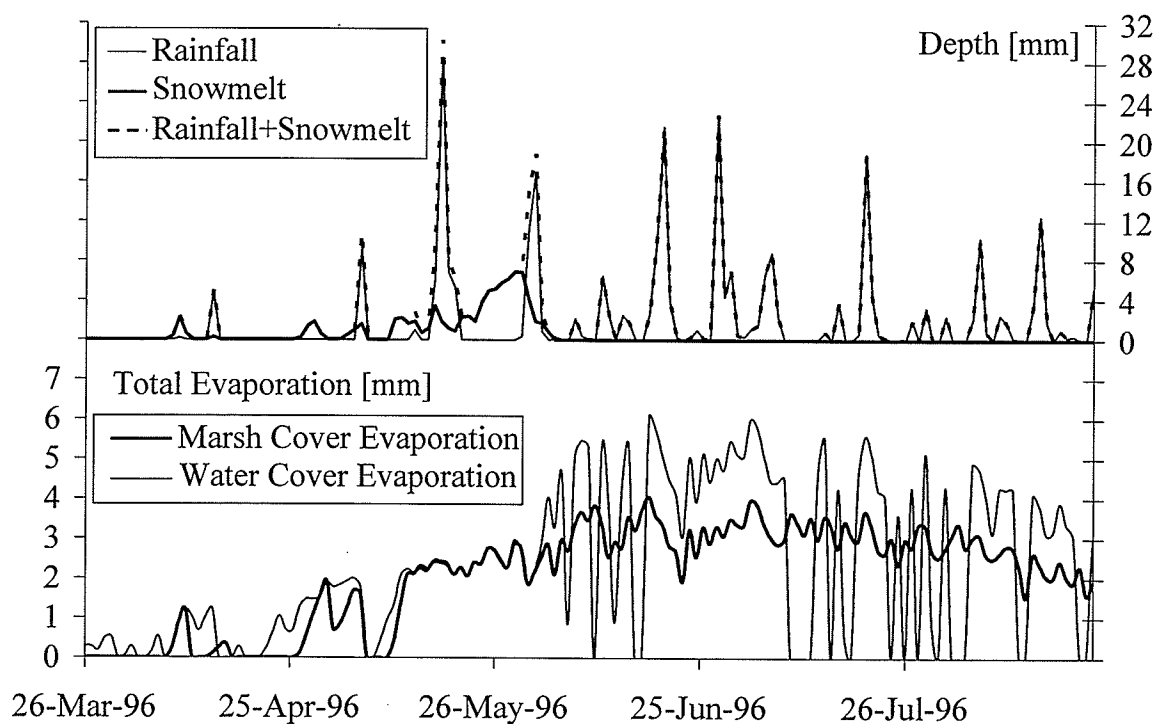


Figure 4.23 - 1985 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

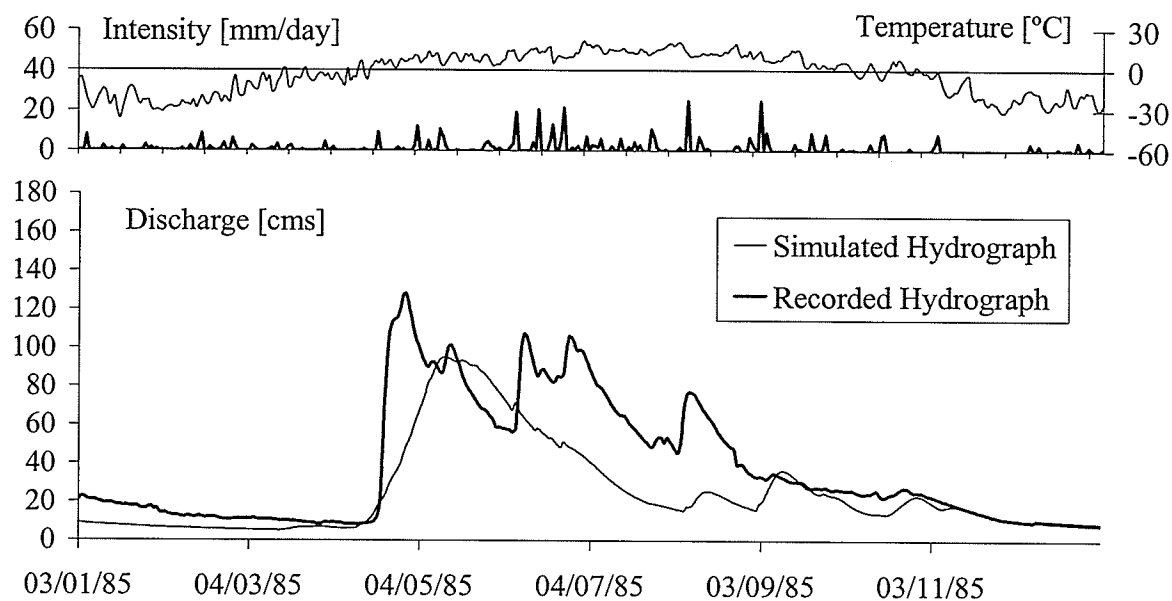


Figure 4.24 - 1986 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

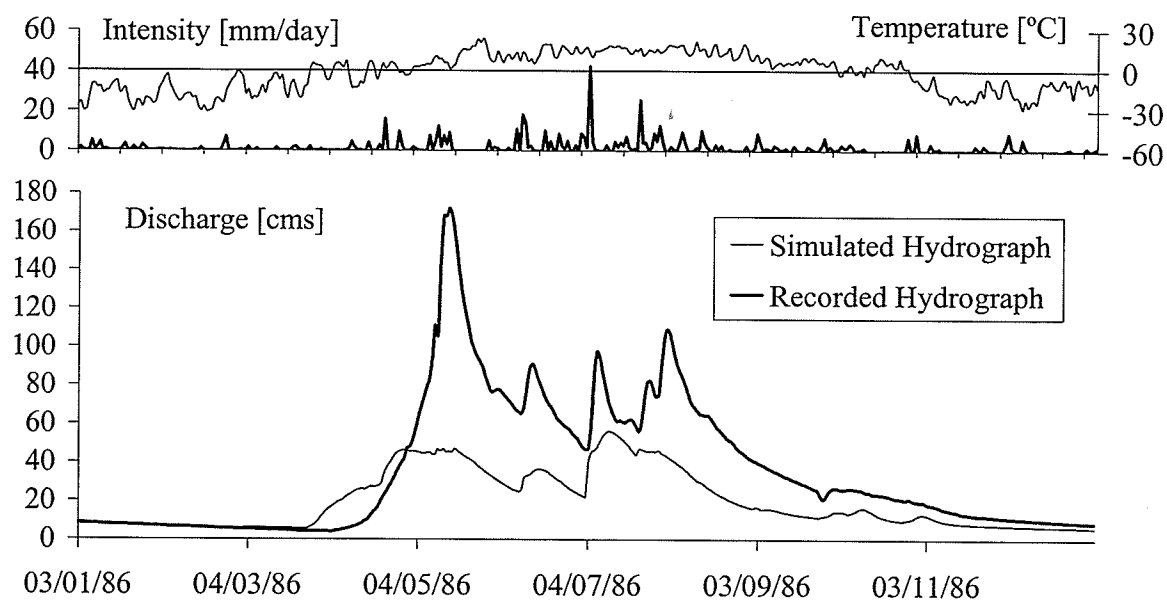


Figure 4.25 - 1987 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

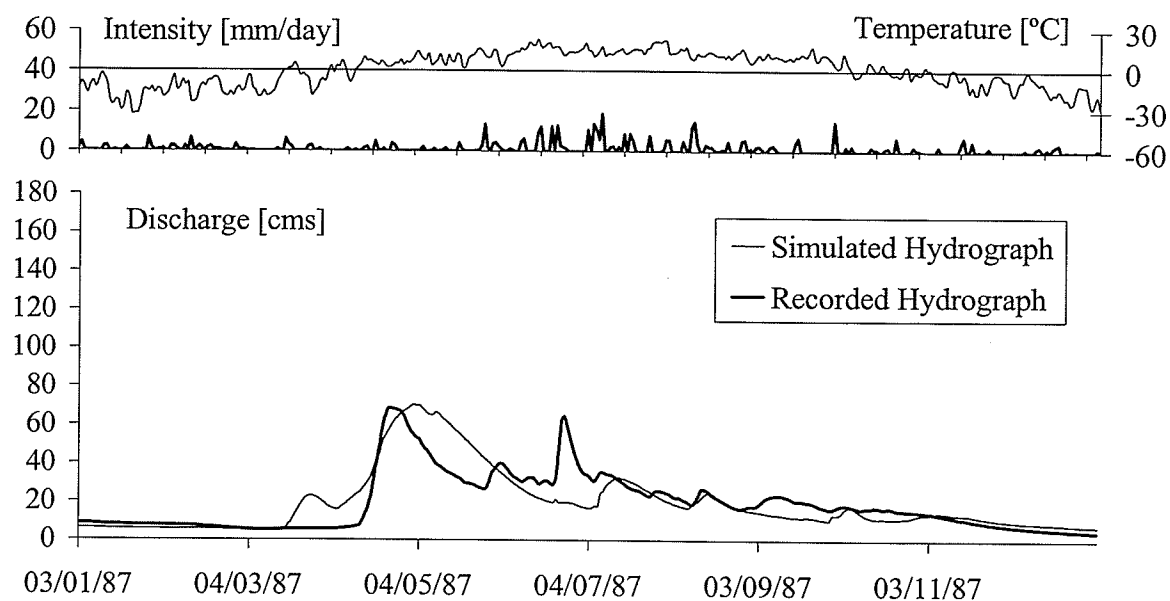


Figure 4.26 - 1988 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

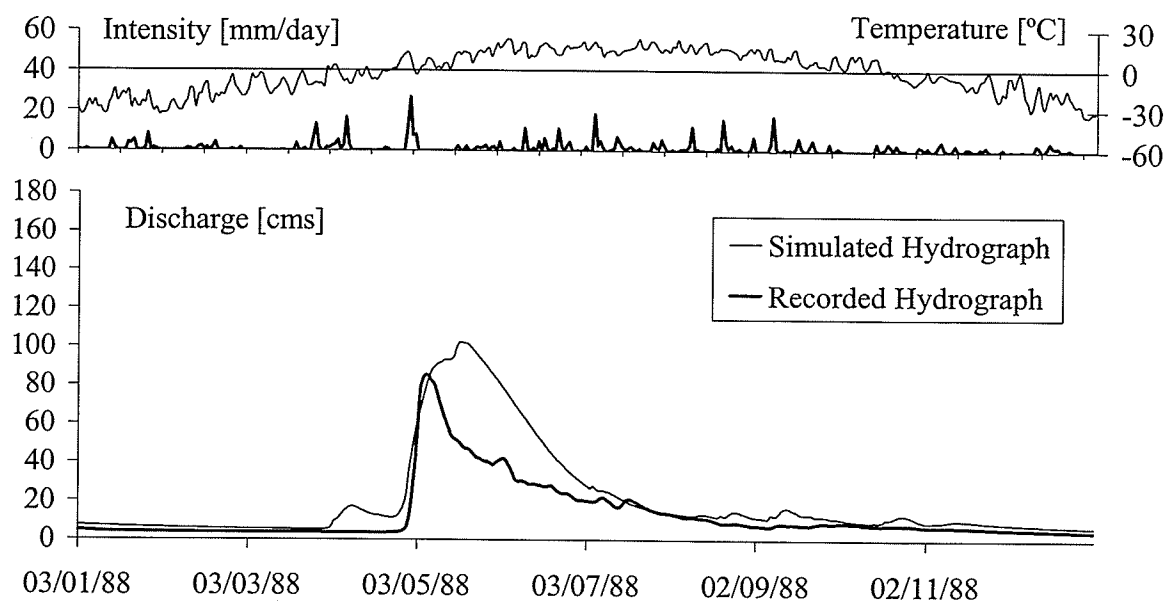


Figure 4.27 - 1989 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

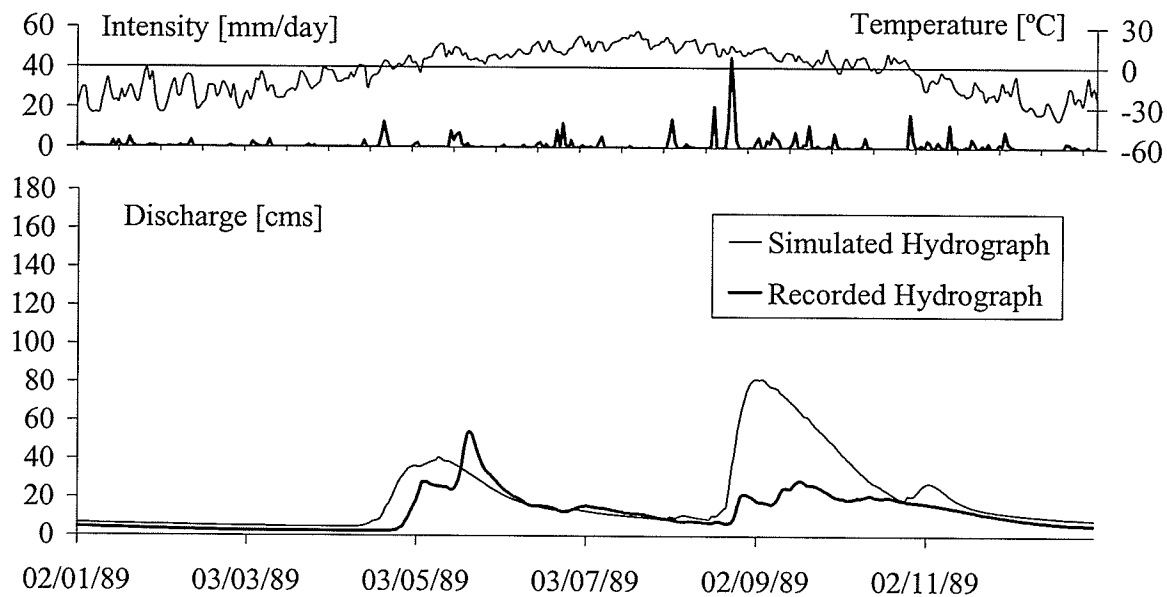


Figure 4.28 - 1990 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

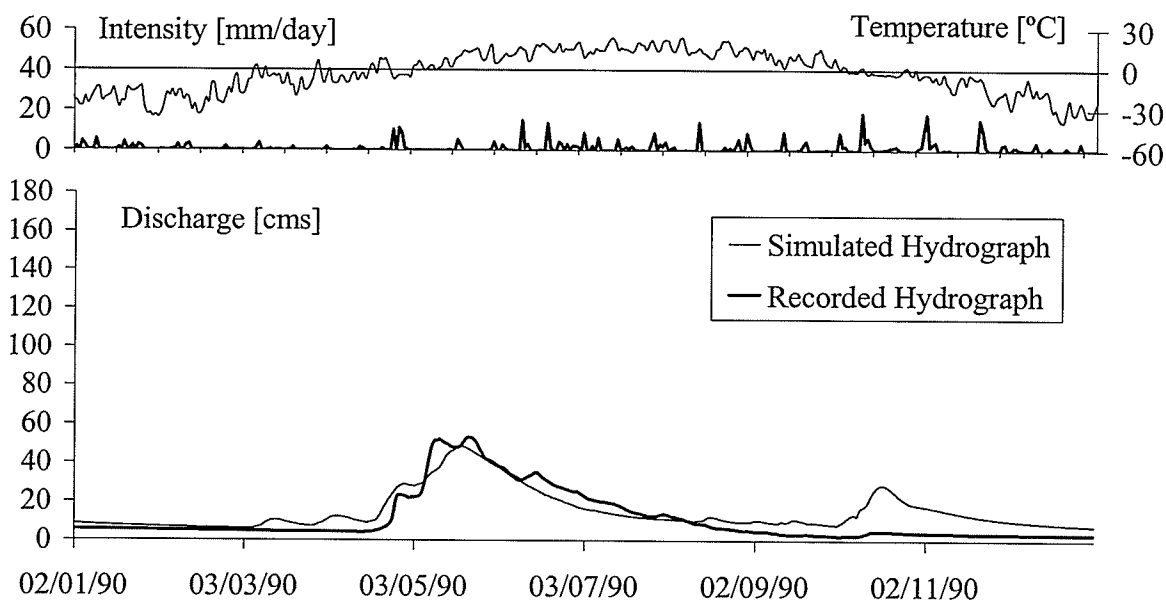


Figure 4.29 - 1991 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

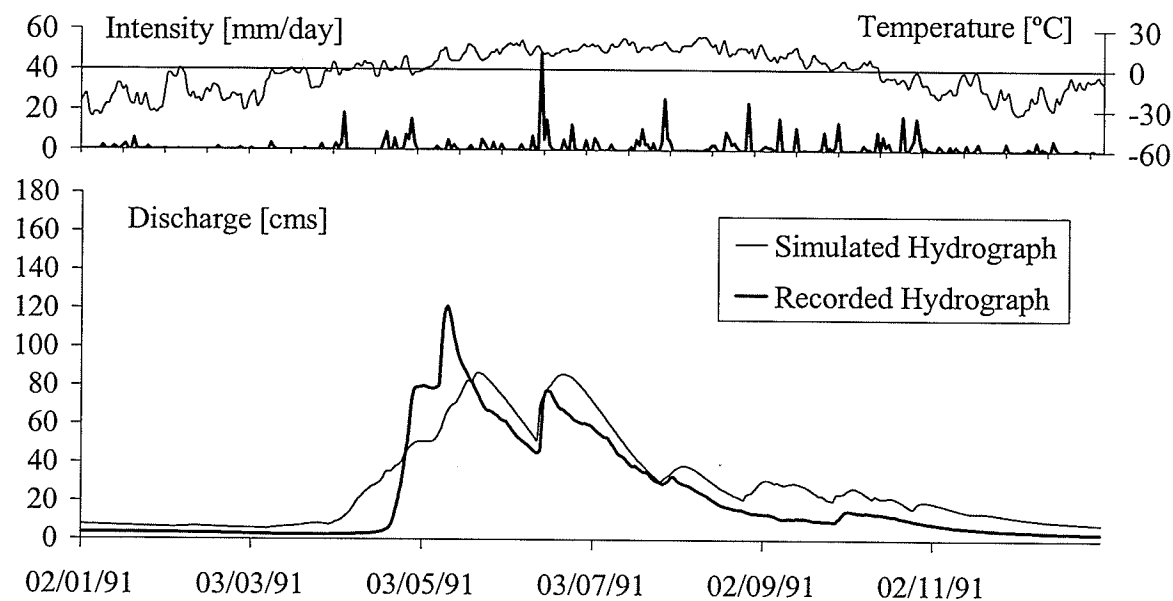


Figure 4.30 - 1992 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

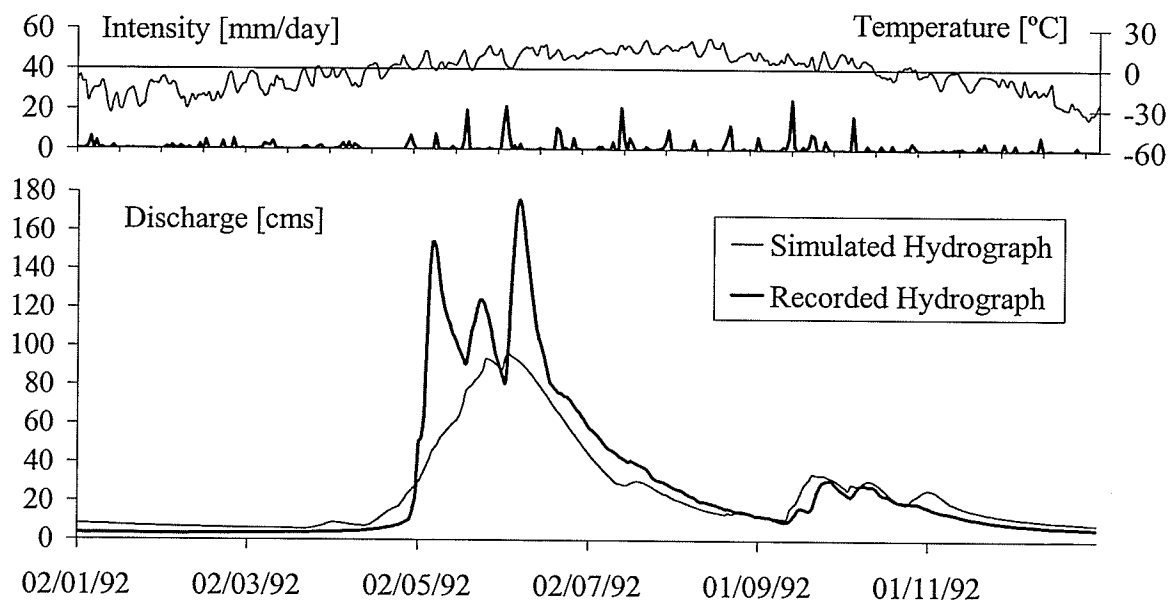


Figure 4.31 - 1993 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

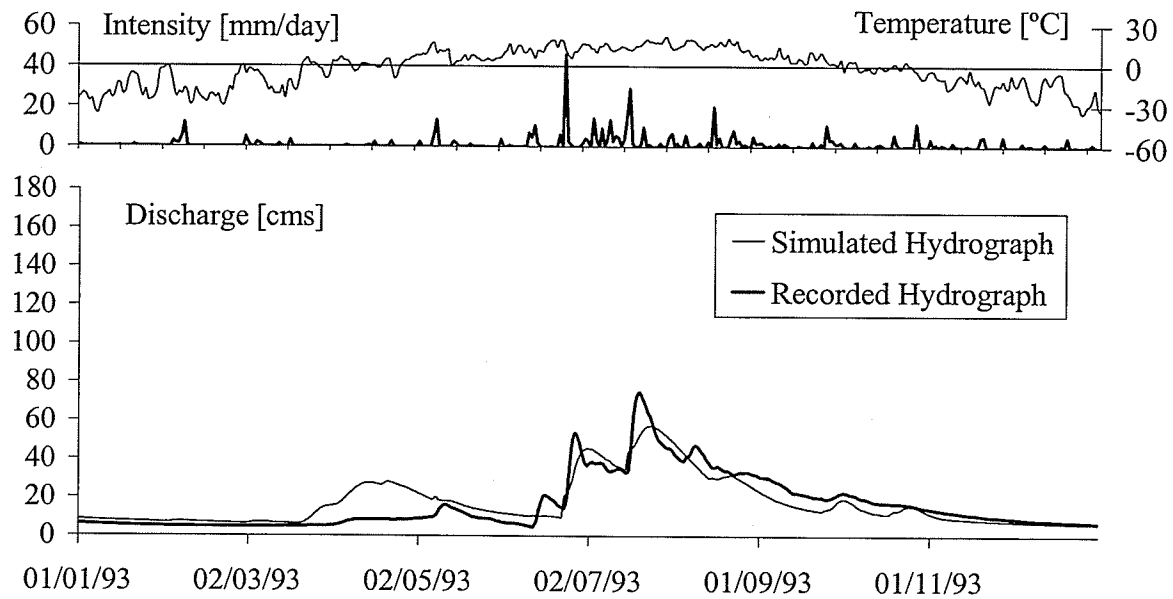


Figure 4.32 - 1994 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

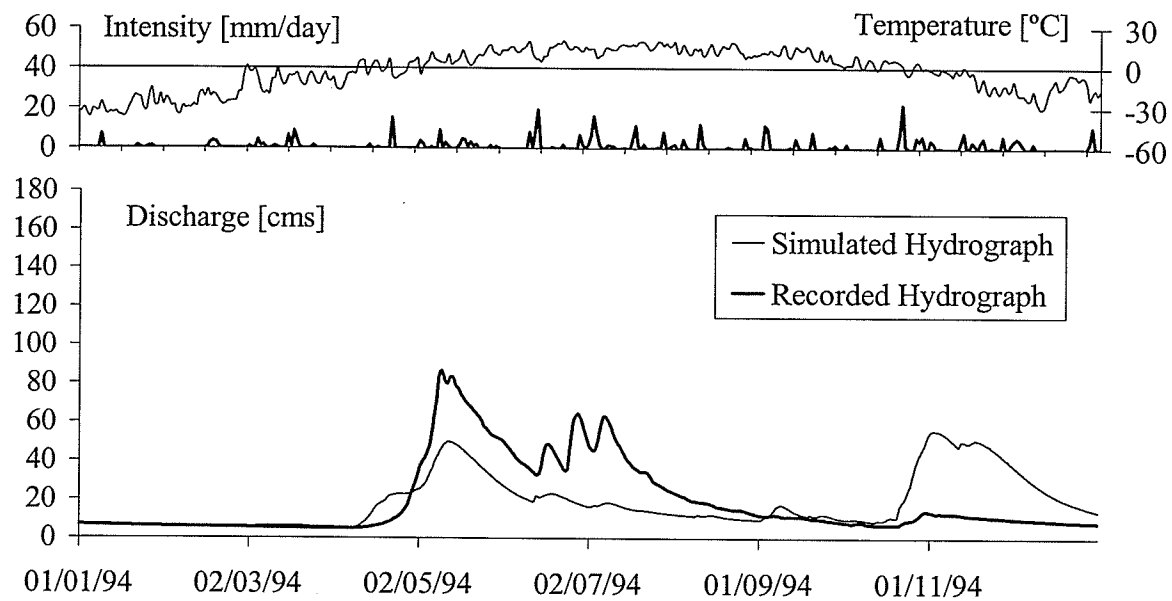


Figure 4.33 - 1995 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

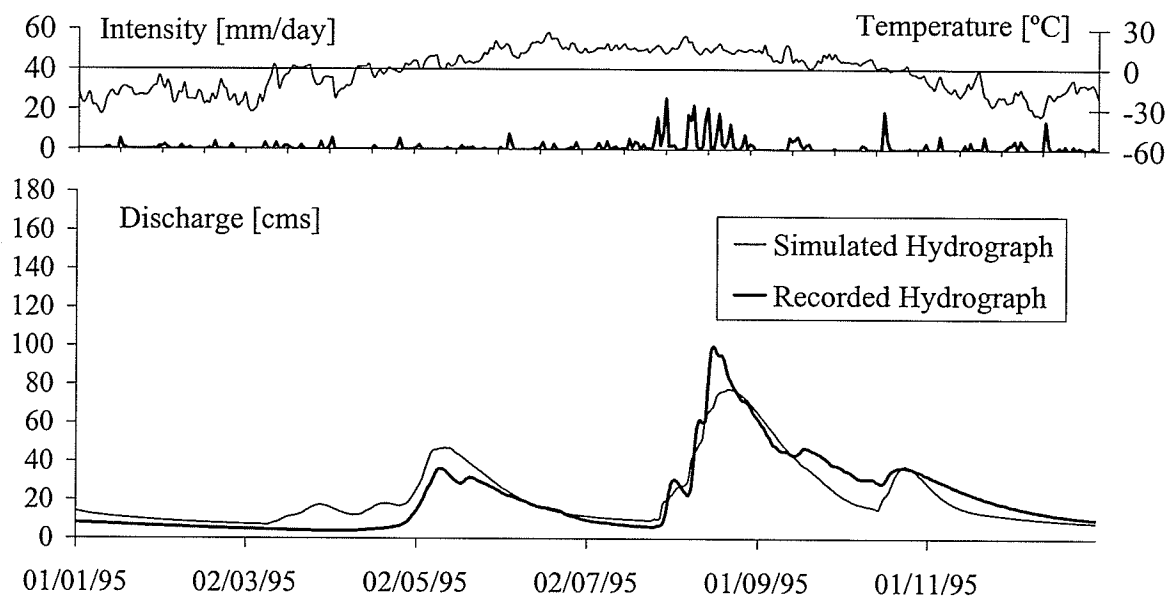


Figure 4.34 - 1996 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

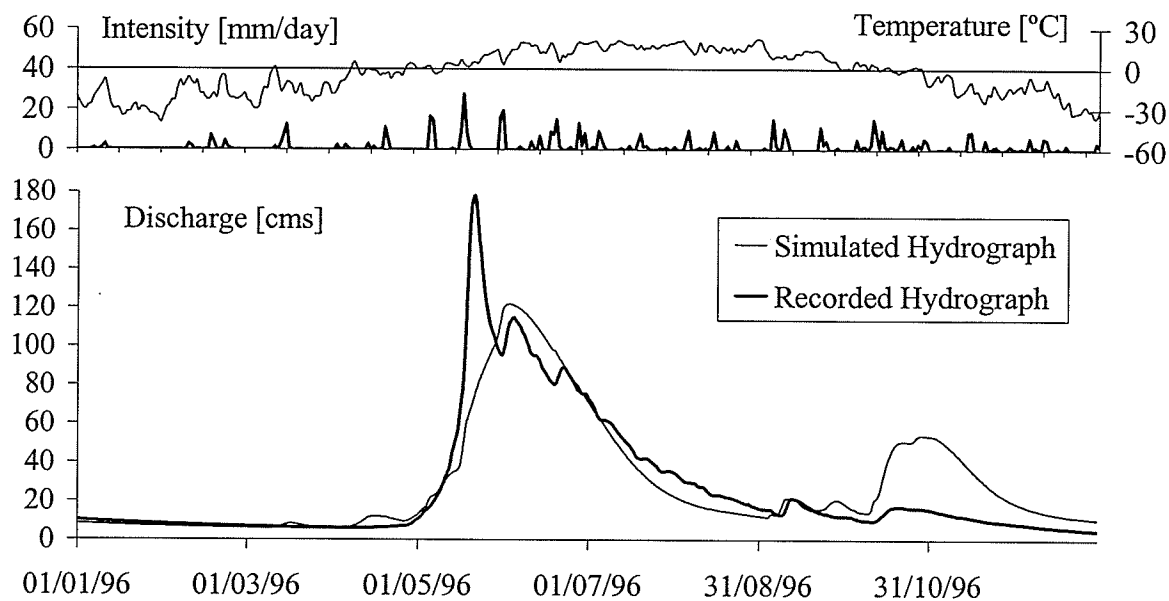


Figure 4.35 - 1997 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

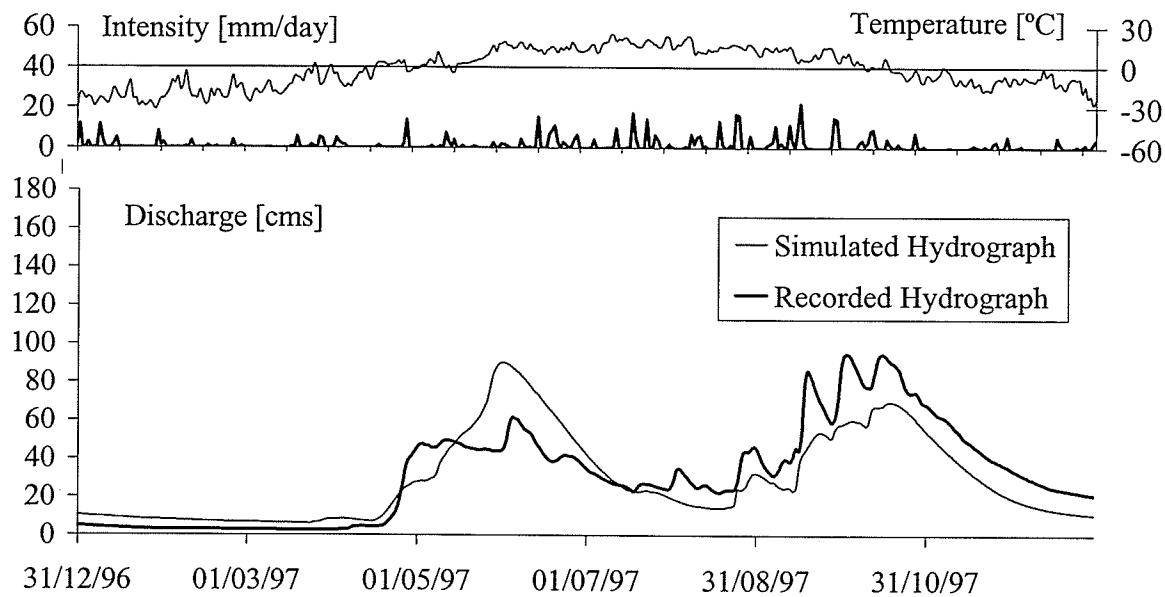


Figure 4.36 - 1998 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

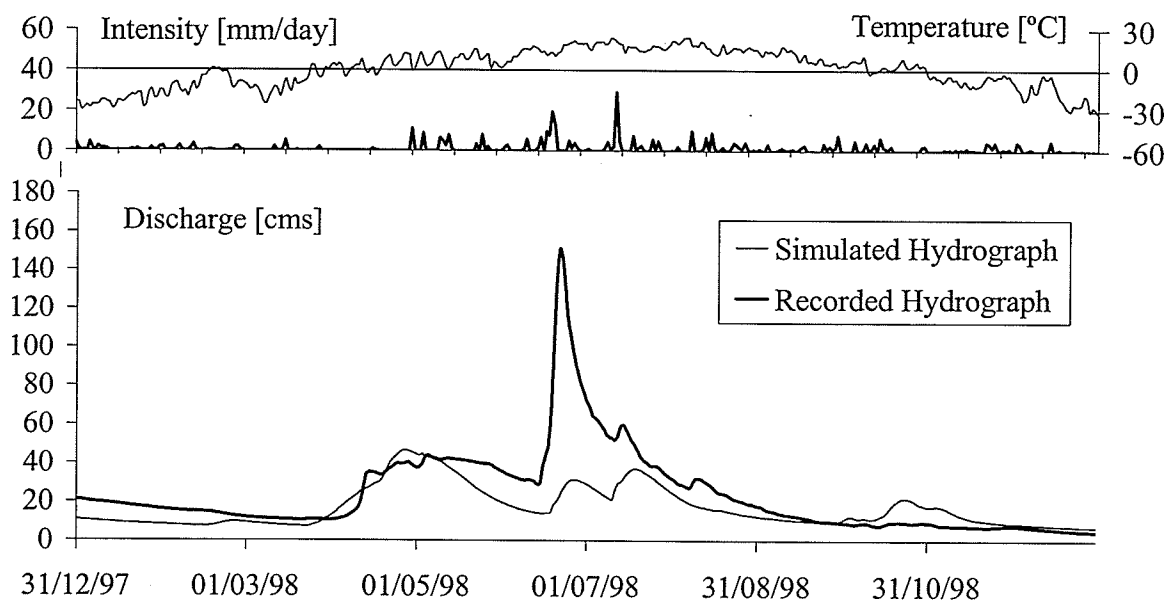


Figure 4.37 - 1999 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

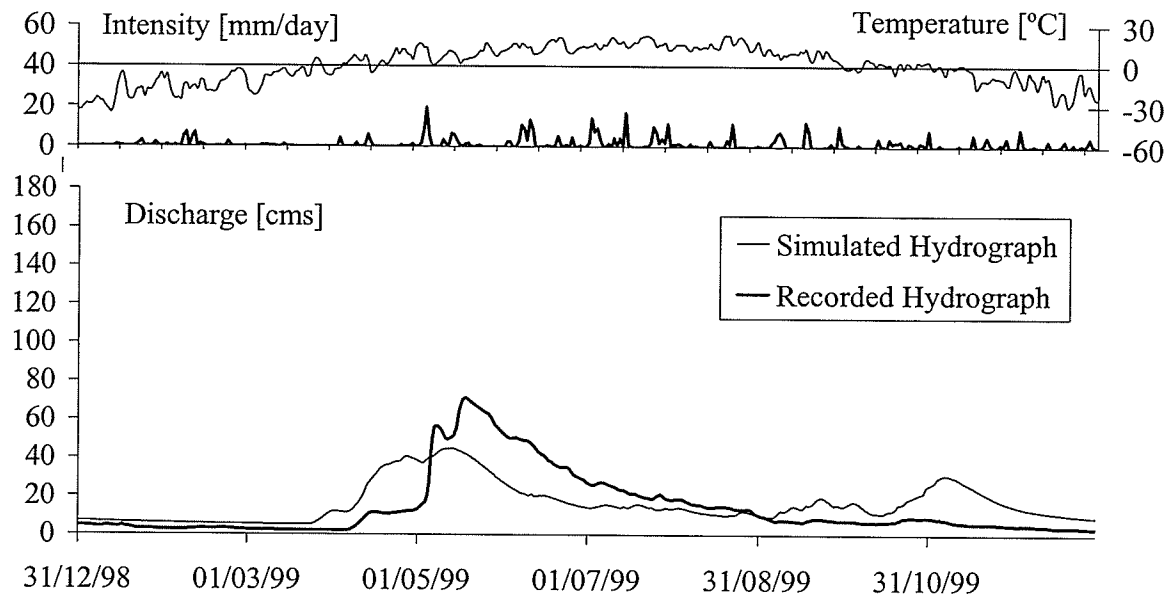


Figure 4.38 - 2000 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP Hydrograph.

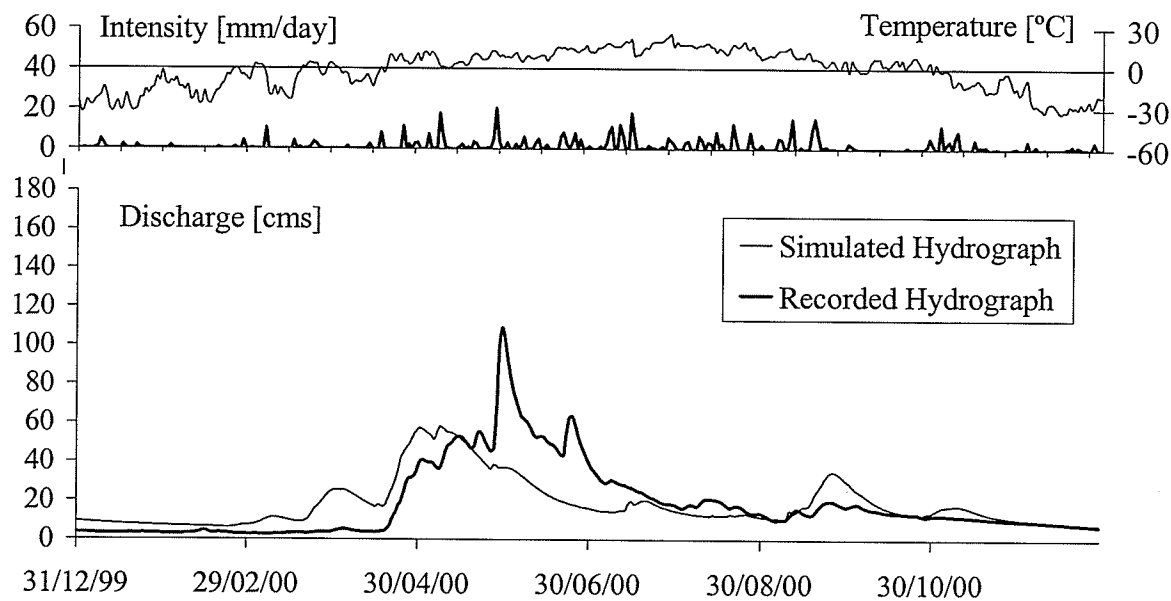


Figure 4.39 – Water balance calculation for the SLURP model application to the Taylor River.

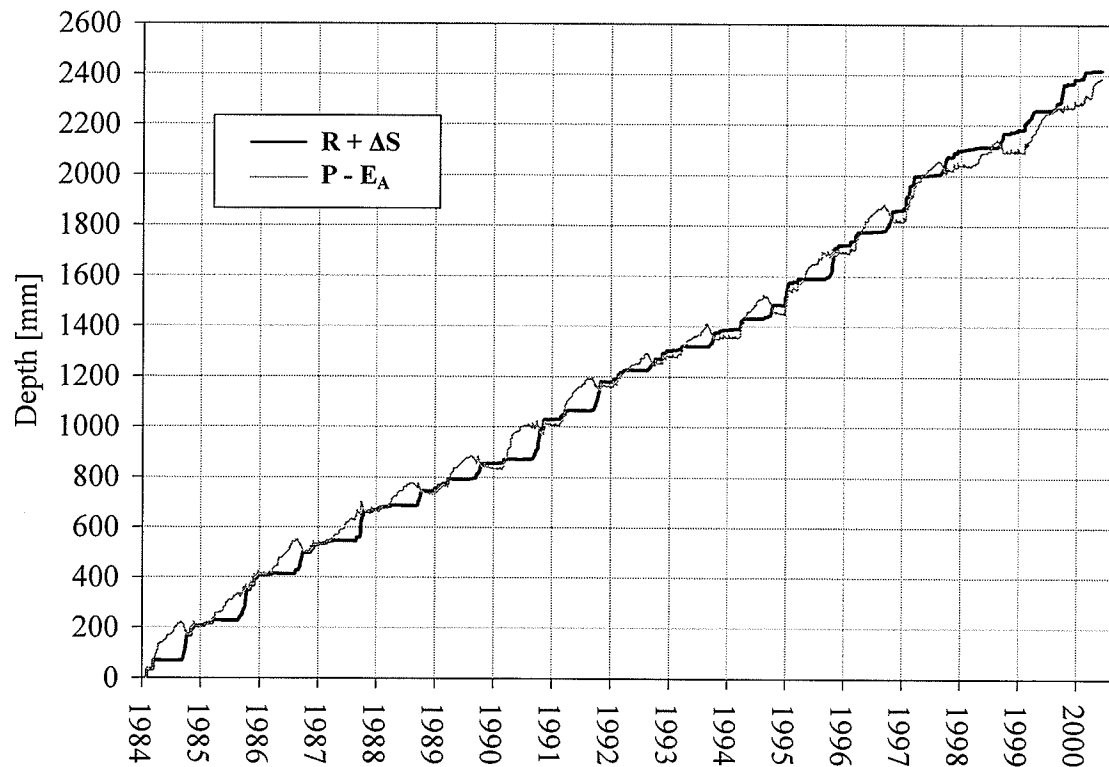


Figure 4.40 – Water balance calculation for the SLURP model application to the Upper Burntwood River.

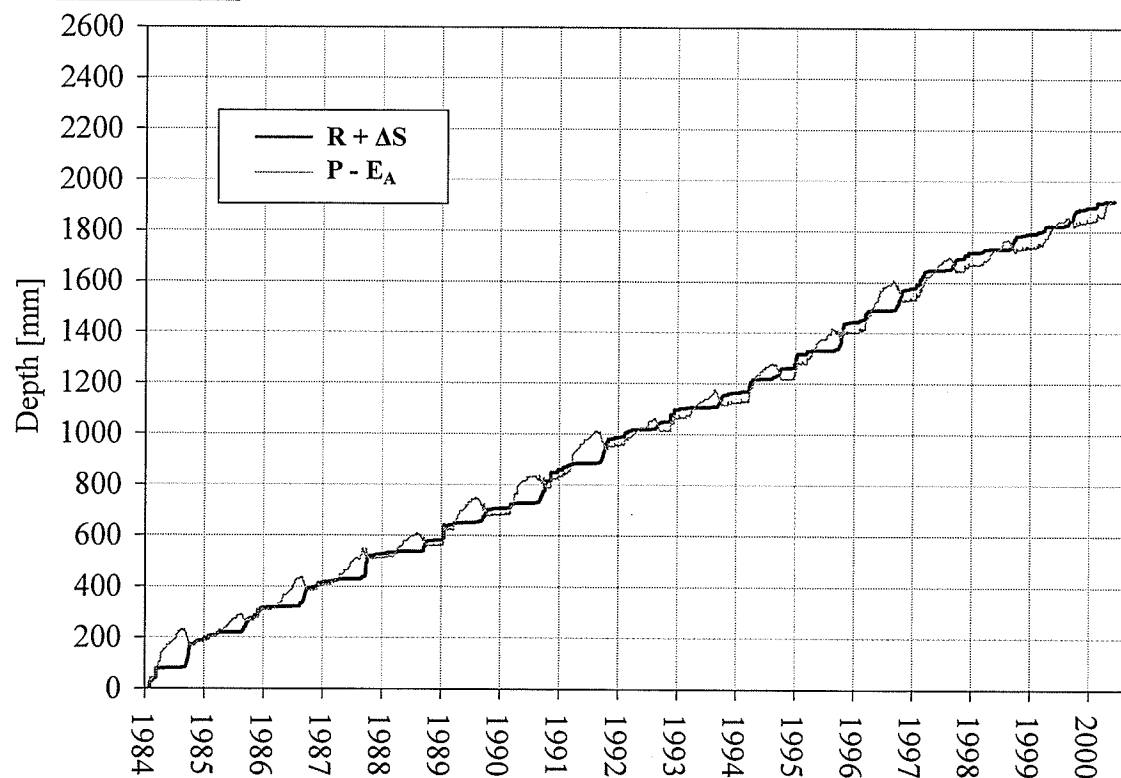


Table 4.1 – Final calibrated parameters for the SLURP application to the Taylor River during for 1996.

Parameter	Conif	Decid	Mixed	Muskeg	Imperv	Water	Marsh
k_2	517.3	517.3	517.3	393.2	262.5	3	50
SSC [mm]	58.4	58.4	58.4	27.6	375	1000	500
$S_{1,max}$ [mm]	407.8	407.8	407.8	492.9	365	0	100
k_1	32.7	32.7	32.7	13.2	18.9	1	5
$R_{1,Jan}$ [mm/day/°C]	0	0	0	0	0	0	0
$R_{1,Jul}$ [mm/day/°C]	0.5	0.5	0.5	0.5	0.5	0.5	0.5
A	1	1	1	1	0.5	0	1

Table 4.2 – Final Muskingum routing coefficients for the Taylor River.

ASA	X	K [day]
1	0.0012	0.707
2	0.0000	0.7856
4	0.5000	0.1663
6	0.2418	0.707

Table 4.3 – Evapotranspiration parameters for final calibration of SLURP to the Taylor River.

Parameter	Conif	Decid	Mixed	Muskeg	Imperv	Water	Marsh
α_w	1.36	1.36	1.36	1.36	2.14	2.14	2.14
α_s	2.14	2.14	2.14	2.14	2.14	2.14	2.14
β	17.0	17.0	17.0	17.0	27.2	27.2	27.2

Table 4.4 – Performance statistics for the SLURP calibration to the Taylor River.

Year	NSE [%]	Dv [%]	Volume Recorded [mm]	Dv [mm]
1985	71	-22	2079	-459
1986	76	-16	2335	-375
1987	79	3	1323	33
1992	75	18	1428	256
1993	27	-10	1134	-109
1994	28	1	1065	12
1995	58	31	1288	397
1996	72	-6	2063	-134
1997	89	-6	2445	-136
1998	42	-16	1446	-225
2000	55	-8	1874	-157
Average	61	+/- 12	1680	+/- 208
Maximum	89	31	2445	397
Minimum	27	-22	1065	-459

Table 4.5 – Performance statistics for the SLURP application to the Upper Burntwood River Basin using the Taylor-Calibration parameter set.

Year	NSE [%]	Dv [%]	Volume Recorded [mm]	Dv [mm]
1985	44	-19	14046	-2684
1986	66	-28	13025	-3645
1987	11	22	7067	1565
1988	-103	88	4823	4253
1989	-503	102	4500	4577
1990	30	64	4242	2706
1991	48	57	8126	4612
1992	81	9	10578	1000
1993	68	22	6053	1317
1994	28	17	7120	1229
1995	63	31	7833	2430
1996	52	33	9775	3263
1997	38	12	11744	1432
1998	40	-16	8837	-1379
1999	30	38	5359	2036
2000	35	20	6808	1380
Average	2	+/- 36	8121	1506
Maximum	81	102	14046	4612
Minimum	-503	9	4242	-3645

Table 4.6 – Evapotranspiration parameters for the SLURP calibration on the Upper Burntwood River.

Parameter	Conif	Decid	Mixed	Muskeg	Imperv	Water	Marsh
α_w	1.76	1.76	1.76	1.76	2.77	2.77	2.77
α_s	2.77	2.77	2.77	2.77	2.77	2.77	2.77
β	22.0	22.0	22.0	22.0	35.2	35.2	35.2

Table 4.7 – Final Muskingum routing coefficients for the Upper Burntwood River.

ASA	X	K [day]
1	0.12	1.50
2	0.16	1.95
3	0.11	2.12
6	0.14	3.77
8	0.13	0.57

Table 4.8 – Performance statistics for the SLURP calibration to the Upper Burntwood River Basin.

Year	NSE [%]	Dv [%]	Volume Recorded [mm]	Dv [mm]
1985	44	-35	14046	-4863
1986	33	-44	13025	-5736
1987	56	1	7067	44
1988	16	57	4823	2772
1989	-193	61	4500	2727
1990	69	29	4242	1250
1991	77	29	8126	2369
1992	75	-15	10578	-1536
1993	76	5	6053	276
1994	11	-5	7120	-389
1995	85	3	7833	254
1996	74	9	9775	836
1997	68	-8	11744	-972
1998	26	-29	8837	-2596
1999	40	11	5359	601
2000	44	-1	6808	-94
Average	37	21	8121	-316
Maximum	85	61	14046	2772
Minimum	-193	1	4242	-5736

Table 4.9 – Average annual simulated water balance components for the SLURP application to the Taylor River and Upper Burntwood River.

Water Balance Component	Taylor River	Upper Burntwood River
Evapotranspiration [mm]	362	372
Precipitation [mm]	501	482
Runoff [mm]	147	115
Observed Runoff [mm]	149	119
Snowfall [mm]	123	115
Rainfall [mm]	378	367
Sublimation [mm]	47	58
Canopy Evaporation [mm]	276	278
Soil Evaporation [mm]	39	36
Snowmelt [mm]	77	56

CHAPTER 5

5 SLURP MODEL MODIFICATION

5.1 Introduction

As the understanding of hydrologic processes increases and hydrologic modelling experience is gained for a particular region revisions are made to the model to improve its performance. A number of SLURP Modelling deficiencies were noted while simulating the hydrology of the High Boreal Forest of northern Manitoba were determined through the literature review of hydrological processes in Chapter 2 and the model application presented in Chapter 4. This chapter describes the attempts made to address a few modelling deficiencies through model revisions. The modified SLURP model is applied to the Taylor and Upper Burntwood Rivers to test for model performance improvement.

5.2 SLURP Model Deficiencies

The following lists the SLURP modeling deficiencies that if addressed in the model would likely improve the modeling performance in the are of interest to this study:

1. Effects of frozen ground on infiltration capacities during the snowmelt period.

2. Date independent snowmelt rates.
3. Delay of initial snowmelt generated runoff due to snow ripening.
4. Rain on snow events.
5. Depression Storage, its periodic disconnection with the basin, and its corresponding evaporation.
6. Non-linear retention of water and soil moisture capacity in the upper soil zone of Muskeg due to an exponential decrease in hydraulic conductivity and porosity.
7. Time varying Fast Storage Tank capacity due to formation of an active frost layer or permafrost layer.
8. Moss interception and evaporation of precipitation.
9. Recharge of water to marshes, lakes, and rivers from headwaters, and the corresponding evaporation.
10. Exchange of water between Marshes/Beaver Flooded Land with Lakes and Rivers.
11. Ability to specify the initial contents of the Fast Storage Tank.

This research attempts to address a few of these modeling deficiencies, and specifically targets the snowmelt period.

5.3 Infiltration into Frozen Soils

5.3.1 Literature Review

Many field studies have shown that frozen soils decrease or increase infiltration rates depending on antecedent moisture conditions (Buttle et al., 2000; Zhao & Gray, 1997; Zhao & Gray, 1999). Chapter 4 showed that during the spring snowmelt period a small storage capacity and/or diminished infiltration capacity is required in order to limit infiltration and generate increased runoff. During the summer months, a much larger soil storage capacity is required in order to absorb all but the largest rain events. Hence, this modeling deficiency impacts both the snowmelt period of the hydrograph and summer period of the hydrograph.

Zhao and Gray present a simple correlation function for estimating snowmelt infiltration into frozen soils developed by using the results from a numerical model named HAWTS (Heat And Water Transport in frozen Soils). The HAWTS model was run for various soil conditions in both a prairie and boreal forest setting. Zhao and Gray show that infiltration into frozen ground has two regimes: transient followed by steady state. The transient flow regime is normally short compared to the steady state regime and is therefore not accounted for in their frozen ground infiltration model. The resulting correlation function is designed so that infiltration into frozen ground could be modeled in operational hydrology (Zhao & Gray, 1999):

$$INF = CS_o^{2.92} (1 - S_i)^{1.64} \left(\frac{273.15 - T_i}{273.15} \right)^{-0.45} t^{0.44} \quad (5.1)$$

where **INF** is cumulative infiltration [mm], **C** is the bulk coefficient that characterizes the effects on infiltration due to differences between model and natural systems, **S_i** is the initial soil saturation [decimal], **S_o** is the surface saturation [decimal], **T_i** is initial soil temperature [°C], and **t** is the infiltration opportunity time [hours].

Through field testing Zhao and Gray (1999) determined that $C = 1.3$ for Boreal forests.

The model assumes the following:

- (a) a homogeneous, isotropic soil
- (b) uniform distributions of initial soil temperature and moisture/ice
- (c) a constant head of melt water at the soil surface.

5.3.2 Conceptual Representation in SLURP

A simplified version of SLURP was constructed in a Microsoft Excel spreadsheet in order to examine the effects of modeling frozen ground. A simplified version of Equation 5.1 was implemented:

$$INF = 1.3 * S_o^{2.92} (1 - S_i)^{1.64} 6.05 * t^{0.44} \quad (5.2)$$

Zhao and Gray state that the effects of the soil temperature on INF is minor compared to the other state variables. Consequently the soil temperature is assumed to be a constant -5°C resulting in a constant equal to 6.05. Zhao and Gray assumed the soil to be fully saturated for their application to a prairie environment. For simplicity, a fully saturated soil surface will be assumed for this research. Zhao and Gray assumed that the infiltration opportunity time was taken as the total time during the day where air temperature was above 0°C , snow temperature was close to 0°C , and solar radiation was greater than 100 W/m^2 . For simplicity this research assumes t to be an average of 10 hours on days where snowmelt exists. This assumptions works out to be similar to the hourly temperature curves presented by Zhao and Gray.

The frozen ground routine was tested for the 1991 and 1996 periods. Figure 5.1 shows the 1996 hydrographs for the Taylor River when frozen ground is modeled. The frozen ground creates a significant amount of overland runoff resulting in a steep hydrograph rising limb, a discharge peak with a short duration, and an equally steep receding limb. The fraction of the basin that contains frozen ground was lowered from 100% to 75%, 50%, 25%, and 10%. Correspondingly, the magnitude of runoff generation decreases as the fraction decreases.

The first peak which on May 19 occurs because there is approximately 46 mm of rainfall and 8 mm of snowmelt over a four day period. The maximum infiltration rate varies between 6.6 and 11.4 mm/day over this period and results in approximately 10.0 mm of overland runoff when the entire basin is frozen soil. If the fraction of the basin with frozen ground is reduced to 25%, the overland flow is decreases to 2.5 mm. The overland flow exits the basin quickly resulting in a peak with a very short duration and a steep rising and receding limb. The observed hydrograph is characterized by a steep rising limb and a slower receding limb. The frozen soil model is not accounting for the process responsible for this hydrograph characteristic. The 1991 test period exhibits the same type of results as the 1996 test period.

The bulk coefficient of 1.3 results in low infiltration capacities on the order of 3.6-17.8 mm/day. Figure 5.2 illustrates that the infiltration capacity increases as the bulk coefficient increases. The infiltration capacity ranges between 7.2-82.3 mm/day if the bulk coefficient is raised to 6.0. Additionally, the infiltration capacity is reduced quickly during the snowmelt period as the contents of the Fast Storage tank approaches the Fast Storage tank capacity.

The second peak which occurs on June 1 is a result of 11 mm of rain on May 31 – June 1 in combination with 74 mm of snowmelt from May 24 – June 6. The result is 20 mm of overland runoff between May 29 and June 6. The total overland runoff is reduced to 7 mm if the fraction of the basin that is frozen is reduced to 25%. The receding limb is much steeper than the observed receding limb. Once the overland runoff exits the basin and the snowpack depletes entirely on June 6, the total streamflow reduces to below the observed discharge indicating that too little snowmelt and runoff is allocated to the soil storage zones.

The simulation of the May 19 and June 1 discharge peaks are complicated by the significant rain on snow events which likely modifies the thermal characteristics of the snowmelt. The analysis carried out for 1991 does not have any significant rain on snow events during the majority of the snowmelt period. However the results for 1991 are similar to the 1996 results.

The inclusion of frozen ground results in significant overland runoff which results from a combination of snowmelt and rainfall for the 1991 and 1996 test periods. The runoff is of short duration and recedes too quickly. The affects are mitigated but still present as the fraction of the basin containing frozen ground is reduced. The frozen ground routine developed by Zhao and Gray does not improve the SLURP hydrograph simulation during the snowmelt period because frozen ground does not likely generate significant amounts of excess runoff during the 1991 and 1996 periods. This modification will not be

included in the modified version of SLURP. The results do seem reasonable given if frozen soil is present hence this modification does warrant additional research.

5.4 Snow Ripening

5.4.1 Literature Review

The majority of the observed hydrographs for both the Taylor River and Upper Burntwood River share three common characteristics:

- (1) the portion of the hydrograph prior to the spring freshet is very flat and shows little or no response to those days where temperatures exceed the melting point of snow;
- (2) the rising limb during the spring freshet is very steep;
- (3) the peak flow during the spring freshet often but does not always occur shortly after the onset of snowmelt runoff.

The SLURP model does not normally replicate these hydrograph characteristics. The simulated hydrographs normally show some hydrograph response to those days where the temperature exceeds the melting point of snow. The simulated rising limb is not as steep as the observed rising limb. The time of rise in the simulated hydrographs are normally longer than the time of rise in observed hydrographs. These deficiencies are likely caused by a few snowmelt processes that are not simulated by the SLURP model.

From the time that fresh snowfall is added to a snowpack until it is entirely converted to snowmelt the snowpack undergoes a constant metamorphosis. Freshly fallen snow is crystalline in structure with sharply defined edges and has a density of roughly 10% of liquid water. Over a period of hours, days, or weeks, the individual snow crystals become rounded and form large ice crystals as the wind and gravity pack them into a smaller space resulting in densities that can reach upwards of 50% of liquid water. In turn the albedo of the snowpack decreases (ASCE, 1996).

The snowpack is depleted by snowmelt and evaporation which are both thermodynamic processes that can be described using an energy balance approach. The energy of snowmelt is derived from:

- (1) net solar radiation
- (2) net longwave radiation exchange
- (3) conduction and convection transfer of sensible heat to or from the overlying air
- (4) condensation of water vapour from the overlying air
- (5) conduction from the underlying soil
- (6) heat supply by indecent rainfall.

Generally, 5 and 6 are minor sources of energy. Exchanges of energy at the surface of the snowpack dominate the snowmelt process (Bedient & Huber, 1992).

The temperature profile of a snowpack is dictated by the atmospheric temperature at the surface, near 0°C at the base, and below 0°C at shallow depths within the snowpack (ASCE, 1996). The ambient temperature eventually raises the temperature of the top layer of the snowpack to 0°C causing some of the snow to be converted to water. Where the snowpack is 0°C there exists three forms of liquid water: Hygroscopic - a thin film of water on the snow crystals; Capillary - water held by the surface tension of water; and Gravitational - water that moves downwards due to the force of gravity (Eagleson, 1970). The gravitational water that moves downwards through the snowpack loses its heat energy to the lower parts of the snowpack in turn causing the melt water to refreeze and the temperature of the lower portion of the snowpack to rise. In time, the pack becomes isothermal at 0°C and reaches a point in time where it contains significant Hygroscopic and Capillary water and no Gravitational water. At this point in time the pack is said to be fully ripe. Any energy added to the fully ripe snowpack results in the production of Gravitational water that exists the snowpack. This entire process is known as “snow ripening”, which causes a delay in the production of melt water that exits the snowpack.

5.4.2 Conceptual Representation in SLURP

The degree day approach for snowmelt modelling included in the SLURP model does not account for snow ripening. The method automatically generates snowmelt that exits the

snowpack on any day where the average temperature exceeds the critical temperature. Figure 4.2 illustrates this deficiency. The observed 1996 hydrograph for the Taylor River basin shows very little response to the snowmelt until May 13. Immediately after May 13 temperatures permit the fully ripe snowpack to melt quickly over a short period of time resulting in a steep rising limb. The simulated hydrograph responds to the snowmelt that is produced on April 12, April 14, April 28. The result is a prolonged snowmelt period resulting in a delayed and stunted hydrograph peak that under estimates the observed peak by more than 25%. Figure 4.3 and Figure 4.4 shows the snowpack depletion curves and the snowmelt inputs to the Fast storage tank as simulated by the SLURP model. If the snow ripening process is included in the SLURP simulation the initiation of the snowmelt runoff would be delayed resulting in a greater release of snowmelt over a shorter period of time. The result would be an improved timing of snowmelt generation and a rising limb and peak that would resemble the observed hydrograph. Hence, there may be benefit to including the snow ripening process in the SLURP model.

The use of a full energy budget is not a practical method for the SLURP model as many of the required inputs are not readily available. Methods for simulating snowpack ripening which use limited inputs have been available for many years and have not undergone many changes (ASCE, 1992; Eagleson, 1970; US Army Corps of Engineers, 1999). One of the most common inputs is the average daily air temperature. The air temperature can be assumed to equal the temperature near the surface of the snowpack. When the air temperature is averaged over a few days in order to allow for the penetration time constant of the thermal transients, this air temperature provides a satisfactory index of the portion of the heat deficit which is due to the capacity of the pack for sensible heat. This capacity is often called the cold content W_c of the snowpack and is defined "as the heat required per unit area to raise the temperature of the snowpack to 0 °C" (Eagleson, 1970). The cold content is normally expressed "in terms of the amount of water needed to be produced at the surface in order to release energy by freezing" (ASCE, 1996). This can be calculated by:

$$W_c = \frac{0.5 \rho_s d T'_s}{80} \quad (5.3)$$

where W_c is the cold content of the snowpack [mm], ρ_s is the average density of the snowpack [gram/cm³], d is the depth of the snow pack [mm], 0.5 is specific heat of ice [cal/(gram °C)], T'_s is the average temperature deficit of the snowpack below 0 °C [°C], and 80 is the latent heat of fusion of water [cal/cm³]

The computation of the cold content relies entirely on the temperature of the snowpack. In the case of a thin snowpack the temperature of the snowpack can be assumed to equal the air temperature for the 1-3 days prior to the forecast time [ASCE]. This study will use a 3 day period prior to forecast time. In the case of deep snow packs it is normally assumed that only the top 60 cm of the snowpack is influenced by the air temperature and the deeper portion of the pack is near -1°C.

The approach taken to model the condition of the snow is to maintain an accounting of the relative temperature of the snowpack below freezing as a function of time. The snowpack is treated as an energy reservoir where snowmelt runoff is generated once the reservoir is full. This state coincides with the snowpack isothermal temperature of 0°C. The cumulative temperature of the snowpack is determined with an index relation (ASCE, 1996):

$$T_s(2) = T_s(1) + F_p [T_a(2) - T_s(1)] \quad (5.4)$$

where T_s is the index of the snowpack surface temperature at times (1) and (2) (°C), T_a is the average air temperature of the three day period prior to the forecast time (°C), F_p is a factor that varies between 0 and 1 which represents the relative penetration of the air temperature into the snowpack. A value of 0 indicates little penetration of the air temperature into the snowpack causing a slow warming or cooling of the snowpack relative to the surrounding air temperature.

The cold content of the snowpack is calculated using the following equation:

$$W_c(2) = W_c(1) + C_r [T_a(2) - T_s(1)] \quad (5.5)$$

where W_c is the cold content at times (1) and (2) [mm] and C_r is the conversion factor [mm/day/°C]. C_r functions identically to the snow melt rate R_1 used in the snow melt degree day method and will be set to equal R_1 for this study. The snow pack becomes fully ripe once W_c reaches zero thus permitting snowmelt to exit the snowpack.

The snow ripening algorithm described in this research was implemented into the SLURP model. Appendix B contains the FORTRAN code that was added to or modified in SLURP. This new version of the SLURP model will be called the Modified SLURP.

5.5 Snow melt Rates

As described in Chapter 3 literature recommends snowmelt rates between 1 and 5 mm/day/°C for various types of land cover. However, it was found that a January melt rate of 0 mm/day/°C and a July snowmelt rate of 0.5 mm/day/°C provided optimum results. These two snowmelt rates minimize the magnitude of premature snowmelt runoff and best simulate the timing of the rising limb of the simulated hydrograph. The drawback of using the low snowmelt rates is a stunted hydrograph resulting from a prolonged snowmelt period.

The implementation of a snow ripening routine as described in the previous section controls the timing of initial snowmelt runoff generation. This eliminates a major modeling constraint when specifying the snowmelt rates thus allowing for much larger and realistic snowmelt rates to be chosen.

The January and July snowmelt rates that are used by the original SLURP were replaced by a single snowmelt rate, R , that is used for the entire year. This removes the date dependency of the original snowmelt rates and eliminates one parameter from the model.

Appendix B contains the FORTRAN code that was added or modified in the SLURP model for this revision.

5.6 Application of Modified SLURP

5.6.1 Taylor River Application

The Modified SLURP model was applied to the Taylor River basin for a continuous simulation period of 1985-2000. The Modified SLURP introduces two new parameters which are a single snowmelt rate R and the snow ripening constant F_p .

The values of R and F_p were varied and the results visually inspected in order to determine their optimal values. It was found that the two variables provided the flexibility required to effectively simulate the spring freshet. R controls the rate at which the snowpack depletes and F_p controls the timing of the initial snowmelt. Due to the low snowmelt rates the original SLURP simulation displayed a lengthy snowmelt period. This resulted in the peak discharge during the spring freshet occurring later than was found in the observed hydrographs.

Figure 5.3 illustrates the sensitivity of R within the Modified SLURP model. It shows that as the value of R is decreased the resulting hydrograph peak decreases and occurs at a later date. The smaller melt rate results in an extended snowmelt period that causes the recession limb to translate further in time and further from the observed recession limb.

The Modified SLURP accommodates the larger snowmelt rates that are recommended in literature because the snow ripening routine prevents premature snowmelt runoff. Through model calibration the optimal snowmelt rates were found to be approximately 30% higher than the snowmelt rates determined by Metcalfe and Buttle (2001). These new rates are 1.3 mm/day/°C for the Coniferous and Muskeg covers, 2.2 mm/day/°C for the Deciduous and Mixed covers, 3.8 mm/day/°C for the Impervious cover, and 6.0 mm/day/°C for the Water and Marsh covers.

Figure 5.4 illustrates the sensitivity of F_p within the Modified SLURP model. It shows that a larger F_p allows the air temperatures to penetrate the snowpack to a greater degree resulting in a faster warming of the snow pack and consequently a sooner snowmelt. A smaller F_p reduces the penetration of air temperatures into the snowpack resulting in a later snowmelt. The optimal value of F_p varied from year to year but was found to be 0.10 for entire 16 year simulation period.

Overall, the Modified SLURP improved the modeling results during the spring freshet period. Figure 5.5 to Figure 5.19 show the hydrographs produced with the original SLURP model and the modified SLURP model for the 1985-2000 period. Every year shows some improvement in the simulated hydrographs.

The timing of initial snowmelt, that point where the cold content of the snow pack is satisfied and snow pack becomes ripe, is improved in nearly all years. This results in a rising limb that occurs over a similar period which is similar to the observed hydrographs. This characteristic combined with the shorter and more intense snow melt period results in a spring hydrograph with improved an improved time to peak for most test years. The increased snowmelt intensity raises the spring hydrograph peaks in nearly all test years. Many of the test years benefit from the higher peak discharge however a few years such as 1987 and 1995 do not benefit.

Figure 5.20 to Figure 5.23 illustrate the snowpack depletion curves developed by the SLURP and Modified SLURP models for the 1991 and 1996 test periods. The curves demonstrate that the snow pack depletes over a shorter period of time in the Modified SLURP model. This causes the snowmelt to translate into runoff over a shorter period of time and results in a steeper rising limb. The slope of the rising limb in the simulation hydrographs is improved and is similar to the rising limbs of the observed hydrographs.

The snow pack depletion curves also demonstrate that the snow pack is larger in the Modified SLURP model relative to the original SLURP model. This occurs because less snowmelt is generated during the fall months as temperatures sometimes rise above 0 °C

and cause the freshly fallen snow to partially ripen and re-freeze. The original SLURP model does not account for this process and generates snowmelt in the fall months on any day where temperatures exceed the critical temperature.

Figure 5.24 illustrates the SLURP and Modified SLURP snowmelt accumulation curves for the 1985-2000 period for the Coniferous Forest and Water land covers. The curves show that the Modified SLURP generates more snowmelt than the original SLURP model. This occurs because the snowpack is fully depleted at an earlier date for most test years due to the higher snowmelt rates. This results in less opportunity time for evaporation from the snowpack and less evaporative losses from the snowpack. Consequently, a larger volume of runoff is generated during the spring freshet.

There are marked differences in the snowmelt accumulation curves among the different land covers. There is a difference of 100 to 450mm of snowmelt generation over the 1985-2000 simulation period among the land covers. The largest difference occurs for the Water land cover and the smallest for the Coniferous and Muskeg covers. The largest change in snowmelt rate from the original SLURP model to the modified SLURP model occurs for the Water cover. Consequently, the date where the snowpack is fully depleted is the earliest for the Water cover and results in the least amount of snow evaporation. The snowmelt rates for the Muskeg and Coniferous covers show the smallest change of all covers resulting in the smallest change in snow depletion periods.

Table 5.1 lists the performance statistics for the original SLURP application and the Modified SLURP application on the Taylor River. The statistics were computed over January – December, April-May, April, and May periods. The SLURP modification mainly impacted the spring freshet period therefore, an evaluation of the statistics over the shorter periods amplifies any changes in the model results.

Overall, the annual NSE increased an average of 2.5% for each of the 15 test years. Only three years did not improve (1987, 1995, 1998) with an average change in annual NSE of -19.6% for those years. The other 12 years showed an average annual NSE improvement

of 8% which is considered to be a significant improvement. The years of 1987, 1995, and 1998 also showed an average change in April-May NSE of -256%. All other years showed a significant improvement in the April-May NSE with an average of 21.4%. The statistics for May exhibit the same trend with an average change of -978% for the same three years and an average increase in NSE of 39% for the other 12 years. The April simulations showed the most improvement in NSE but most values still remain below zero. The streamflow is normally quite small during April hence small errors in discharge relative to the mean April discharge results in a significant change in the NSE during April.

The average annual D_v increased by 0.7% for the 15 year period. This occurs as the volume of discharge increases during the spring freshet and does not cancel out as much of the excess discharge during the summer and fall months. The average 1985-2000 D_v for April-May, April, May improved by 3.8%, 123%, and 2.4% respectively.

In summary, April showed the most improvement in simulation results due to the delay of snowmelt generation during snow ripening. May also showed improvements in NSE and D_v for most years due to the increased snowmelt intensity and greater snow pack available for snowmelt resulting in a steeper rising limb and a higher spring discharge peak. Also, the snowmelt rates and melting characteristics are similar to those determined through field studies. For instance, open areas such as impervious and marsh areas are free of snow long before forested areas (Buttle et al., 2000).

5.6.2 Upper Burntwood River Application

This section describes the application of the Modified SLURP model towards simulating the hydrologic behavior of the Upper Burntwood River. The value of F_p was raised from 0.100 to 0.125 which causes the snowpack to ripen faster and melt sooner in the Upper Burntwood River. The snowmelt rates were not modified.

Overall, the Modified SLURP improved the modeling results during the spring freshet period in a fashion similar to the Taylor River basin. Figure 5.25 to Figure 5.40 show the

original SLURP and modified SLURP hydrographs for the 1985-2000 period. Nearly every year shows at least some improvement in the simulated hydrographs. The snow ripening routines delayed snowmelt and resulted in a delayed hydrograph response in all 16 test years. Most years benefit from the delayed generation of snowmelt.

There are a few differences in the results of the Taylor and Upper Burntwood River simulated hydrographs. Six of the Upper Burntwood River hydrographs (1987, 1988, 1995, 1997, 1998, 2000) have a snowmelt runoff generated peak that is exaggerated by the intensified snowmelt created by the Modified SLURP. This is unlike the two years (1987 and 1995) on the Taylor River that exhibit exaggerated snowmelt runoff generated peaks.

Table 5.2 lists the evaluation statistics for the original SLURP application and the Modified SLURP application. Overall, the annual NSE decreased by 11.3% for each of the 16 test years. Nine of 16 years did not improve with an average change in annual NSE of -26.6%. The remaining seven years improved with an average change in annual NSE of 8.3%. The average April-May NSE improved by 68.6% however the average still remains below zero at -49.6%. The April NSE showed a significant improvement of 3494.1% but still remains below zero at -1914.2%. The average May NSE did not improve however nearly half of the test years showed some improvement. A few years such as 1987, 1988, 1997, and 1998 show a significant decrease in May NSE.

The average annual D_v decreased by -3.8% for the 16 year test period. The average April-May and May D_v also decreased by -0.8% and -7.5% respectively. The average April D_v showed a significant improvement of 53.5% from $\pm 123.7\%$ to $\pm 70.2\%$. However, it should be noted that the April D_v of $\pm 70.2\%$ is still far worse than the Annual, April-May, and May D_v of $\pm 25.2\%$, $\pm 39.3\%$, and $\pm 40.3\%$ respectively.

In summary, April showed the most improvement in simulation results due to the delay of snowmelt generation during snow ripening. May also showed improvements in many years due to the increased snowmelt intensity and greater snow pack available for

snowmelt resulting in a steeper rising limb and a higher spring discharge peak. The results for the Upper Burntwood River are considered satisfactory considering the general quality and scarcity of meteorological input data, the diversity of and quantity of the test years, and the use of nearly all the same parameters used for the Taylor River simulation.

5.6.3 Water Balance Calculations

Figure 5.41 and Figure 5.42 illustrate that over the long term $P-E_A$ (Precipitation – Evapotranspiration) is equal to $R+\Delta S$ (Runoff – Change in System Storage) and that the modified SLURP model appears to be accurately balancing water for the both the Taylor River and Upper Burntwood River. There is less than 2% error in the water balance for both applications. Table 5.3 provides the average annual water balance components of the Modified SLURP application to the Taylor River and Upper Burntwood River.

5.7 Chapter Summary

This chapter presented three SLURP model modifications in order to improve the modelling performance during the snowmelt period. These include the incorporating the effects of frozen ground on infiltration, snow ripening, and modified snowmelt rates.

Infiltration into frozen soils was addressed by incorporating a modified infiltration model developed by Zhao and Gray (1999). The model was tested on the Taylor River for the 1991 and 1996 test periods. The model generates a significant amount of overland runoff that is not sustained and recedes quickly resulting in a significant underestimation in streamflow once the snowpack is depleted. The affects are mitigated but still persist as the fraction of the basin containing frozen ground is reduced. The frozen ground routine does not improve the SLURP hydrograph simulation during the snowmelt period in the manner required and is therefore not included in the modified version of SLURP.

The premature generation of snowmelt in the simulated hydrographs was addressed by incorporating a snowpack temperature deficit model into SLURP. The January and July snowmelt rates in SLURP were changed to a single time independent snowmelt rate.

The Modified SLURP model was applied to the Taylor River and Upper Burntwood River for the 1985-2000 period. The performance of the model during the snowmelt period improved for both basins in most test years. The most improvement in model performance was realized in the Taylor River while the performance of the Upper Burntwood River improved moderately.

The snow ripening model delays the snowmelt in nearly all test years and prevents the premature generation of snowmelt that was created by the original SLURP model. Additionally, the snow ripening that occurs during the fall months is also accounted for resulting in less snowmelt in the fall and a larger snow pack in the following spring. The larger snow melt rates recommended in literature are accommodated in the Modified SLURP and were raised by 30% to provide optimum snowmelt characteristics. The larger snowmelt rates cause a faster snow melt, resulting in a shorter snowmelt period, less sublimation, and more total snowmelt runoff volume. The slope of the hydrograph rising limb, time to peak, and magnitude of the peak are improved in most test years.

Finally, the model performance for both the Taylor River and Upper Burntwood River improved relative to the original SLURP mode performance. The results are considered to be satisfactory considering the general quality and scarcity of meteorological input data, the quantity and diversity of meteorological conditions in the test years, and the fact that most modeling parameters are common between both basins.

Figure 5.1 – Simulated hydrographs resulting from SLURP modified to include frozen soil, Bulk Coefficient=1.3 – Taylor River, 1996.

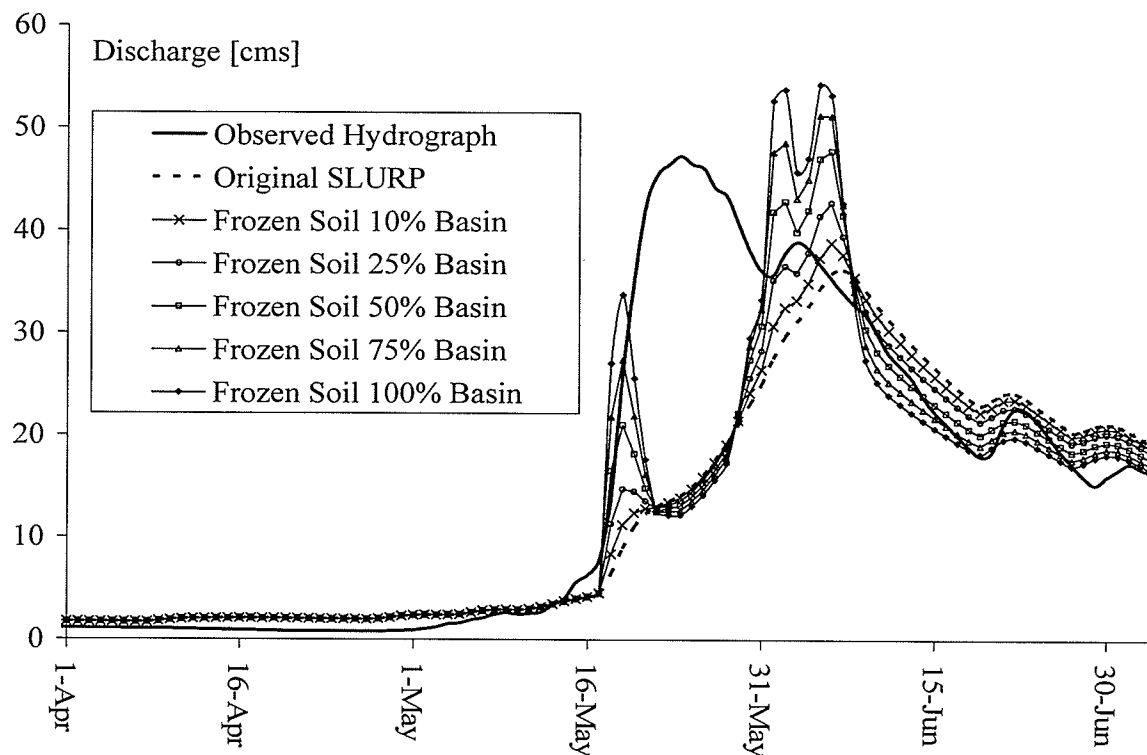


Figure 5.2 – Simulated infiltration capacities resulting from SLURP modified to include frozen soil for a range of Bulk Coefficient, Taylor River, 1996.

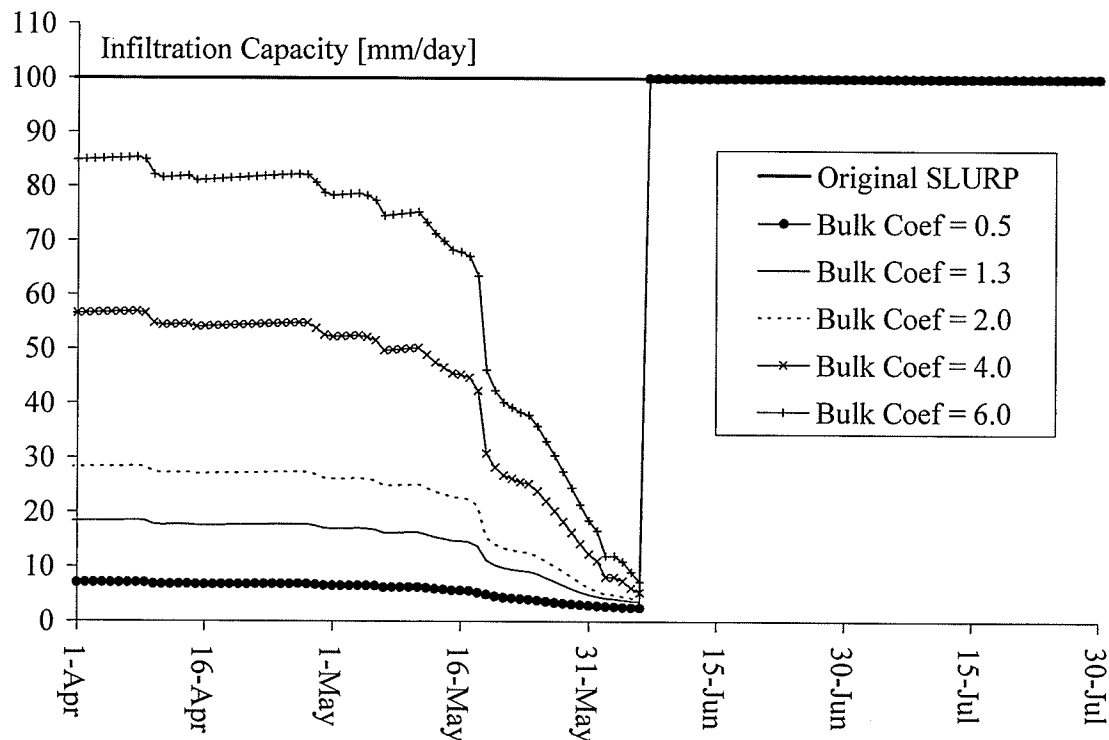


Figure 5.3 – Sensitivity of Snowmelt Rate in Modified SLURP – Taylor River 1996.

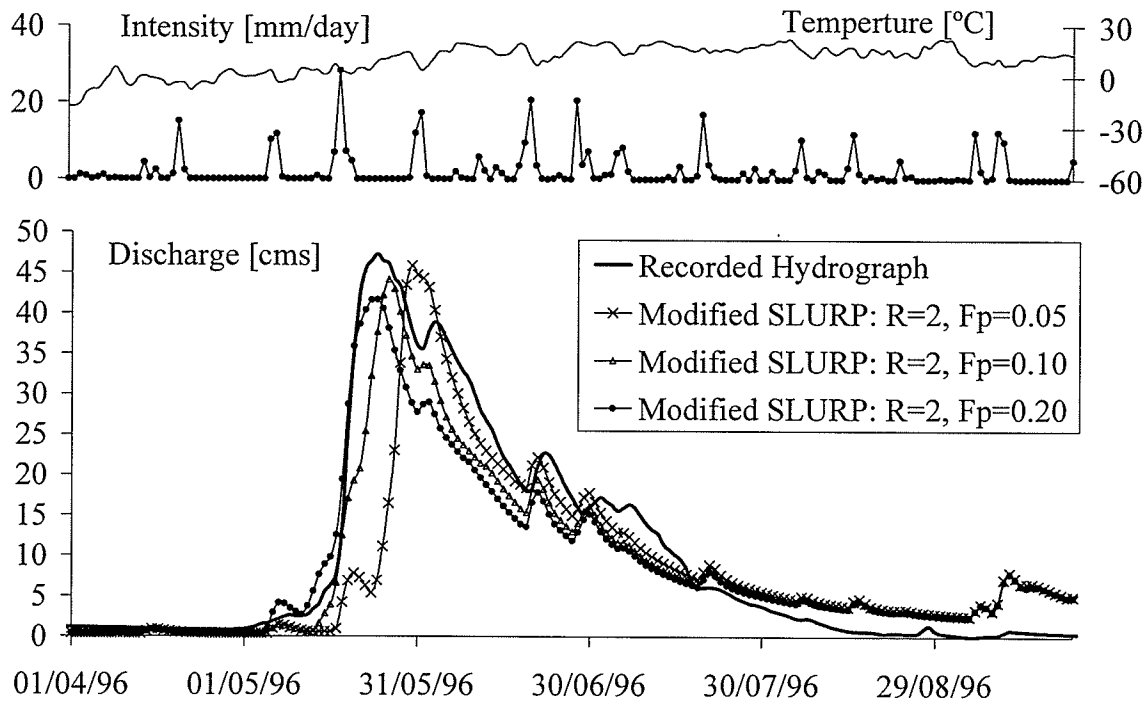


Figure 5.4 – Sensitivity of Fp in Modified SLURP – Taylor River 1996.

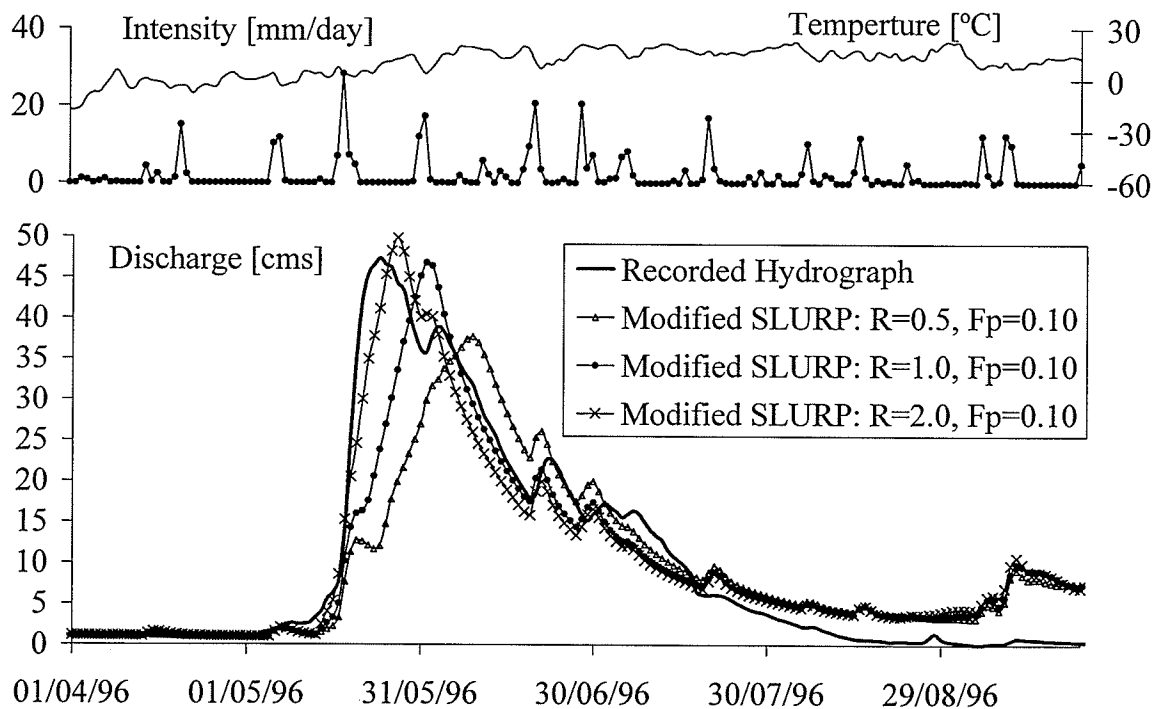


Figure 5.5 - 1985 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

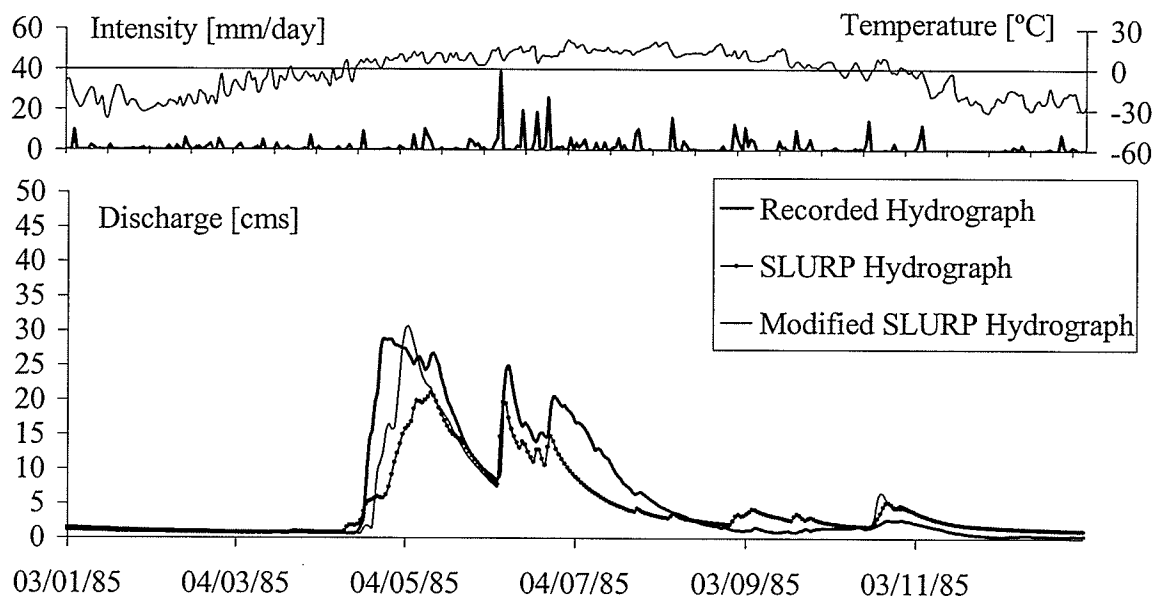


Figure 5.6 - 1986 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

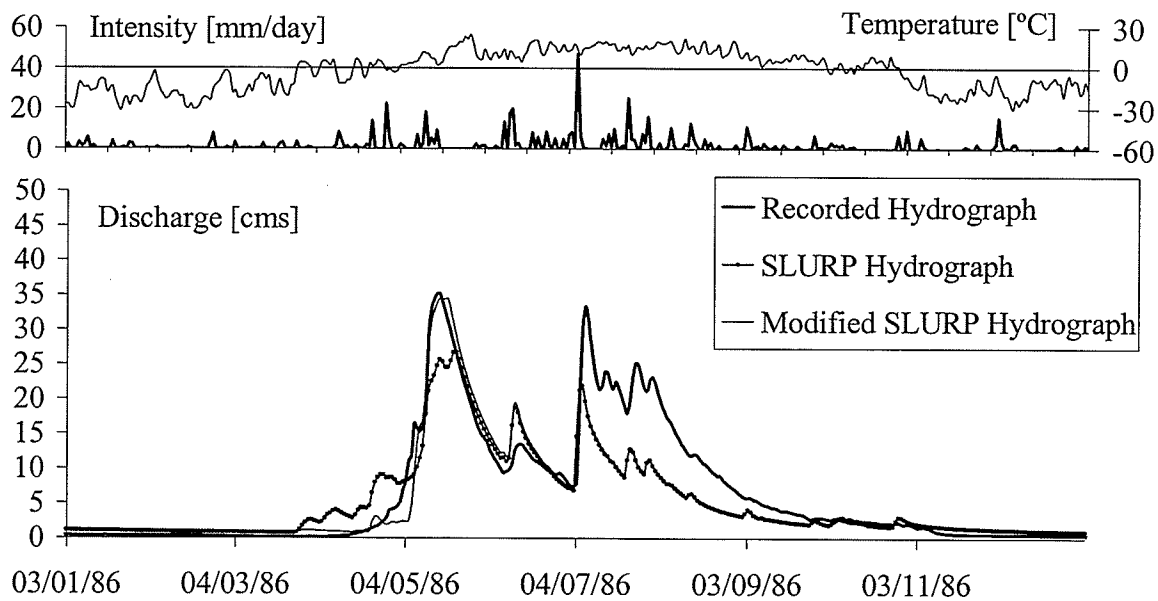


Figure 5.7 - 1987 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

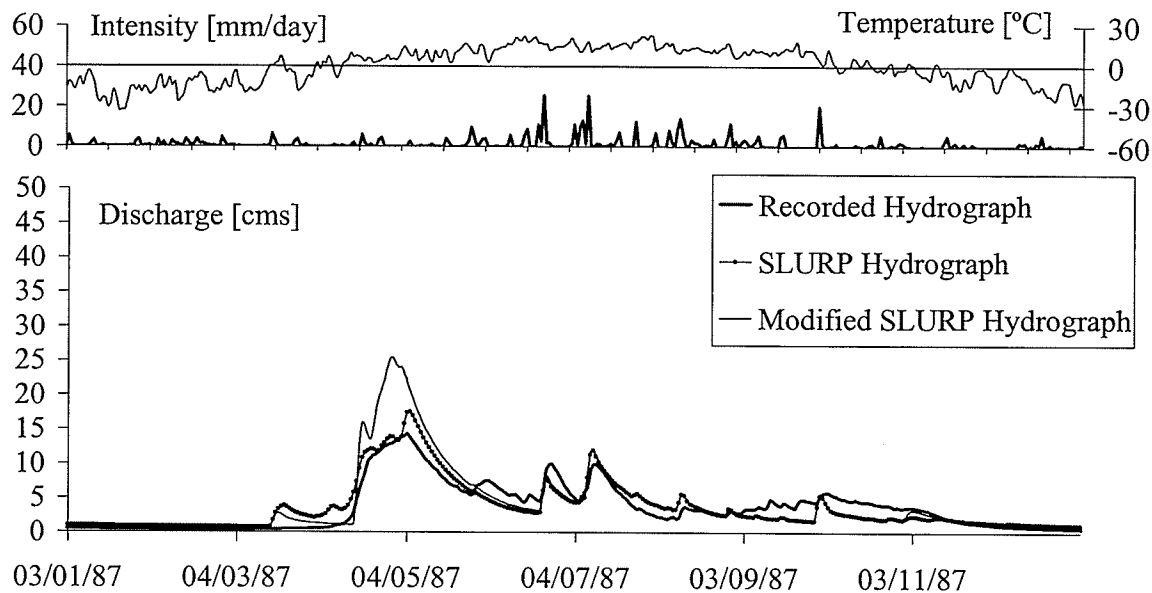


Figure 5.8 - 1988 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

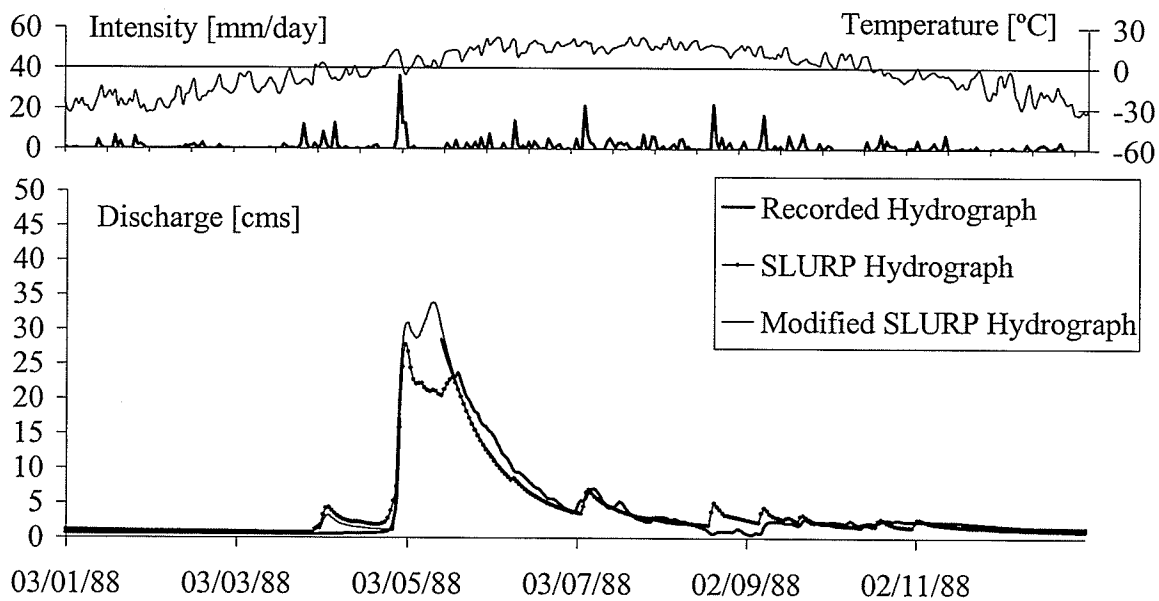


Figure 5.9 - 1989 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

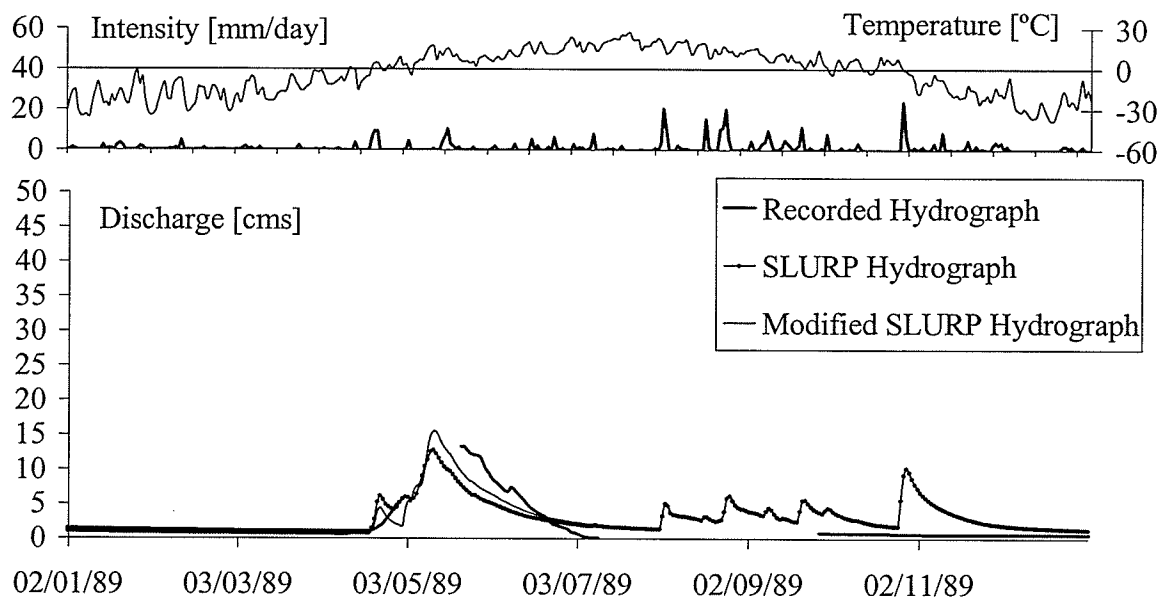


Figure 5.10 - 1990 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

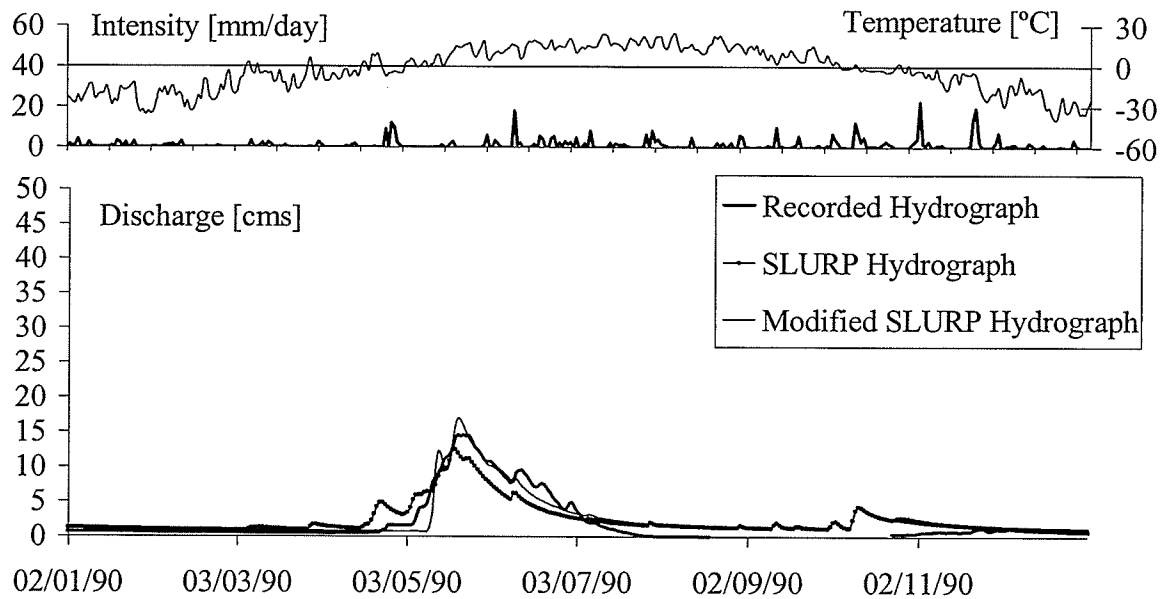


Figure 5.11 - 1991 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

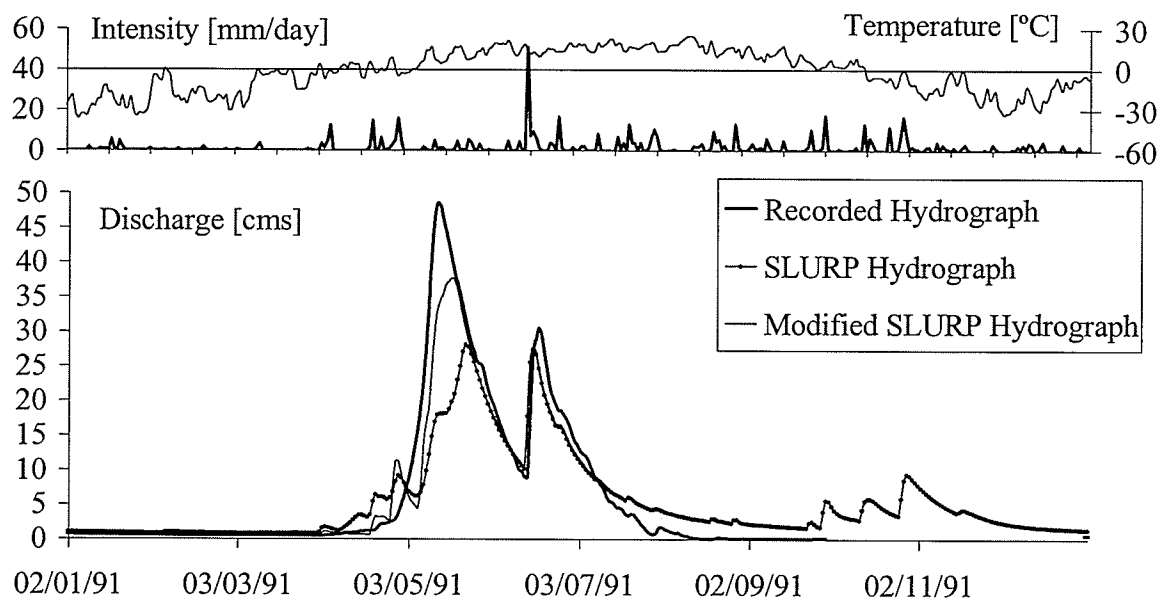


Figure 5.12 - 1992 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

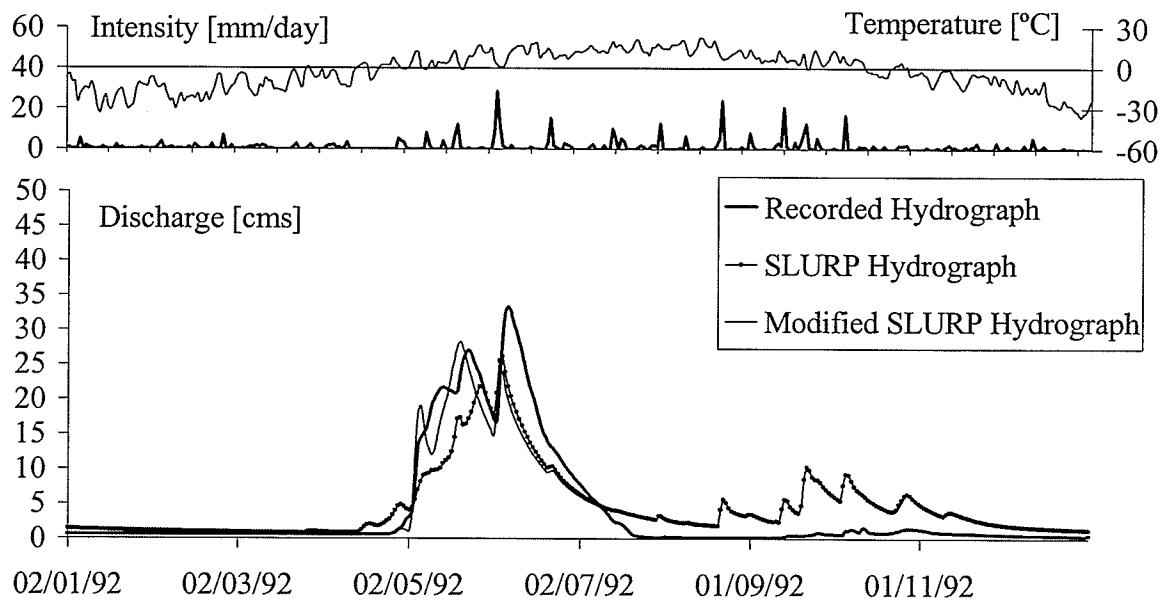


Figure 5.13 - 1993 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

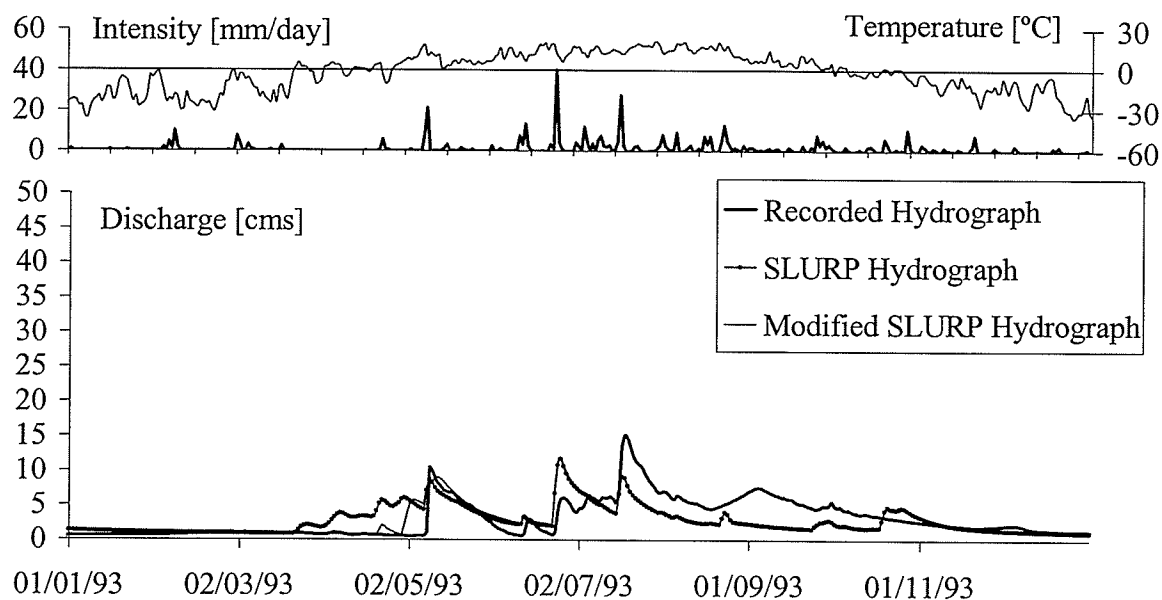


Figure 5.14 - 1994 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

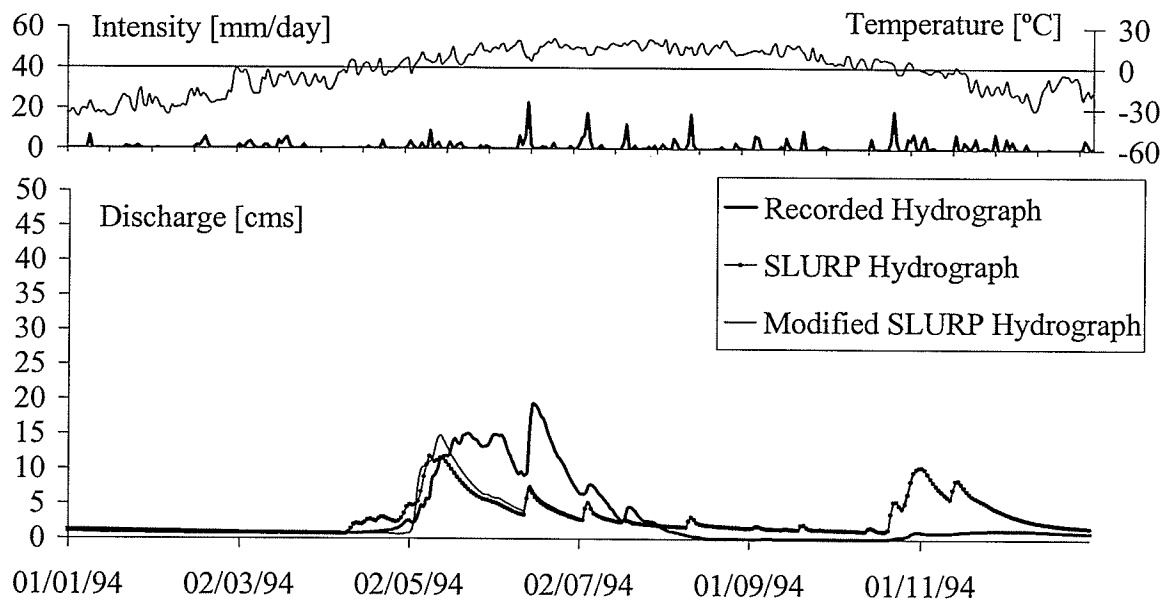


Figure 5.15 - 1995 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

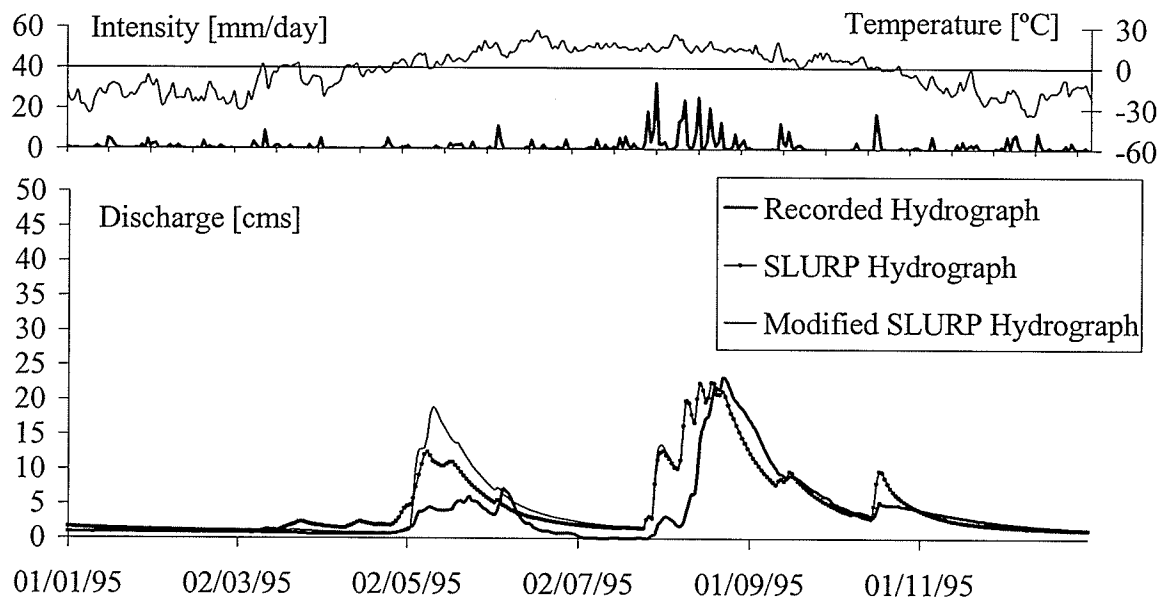


Figure 5.16 - 1996 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

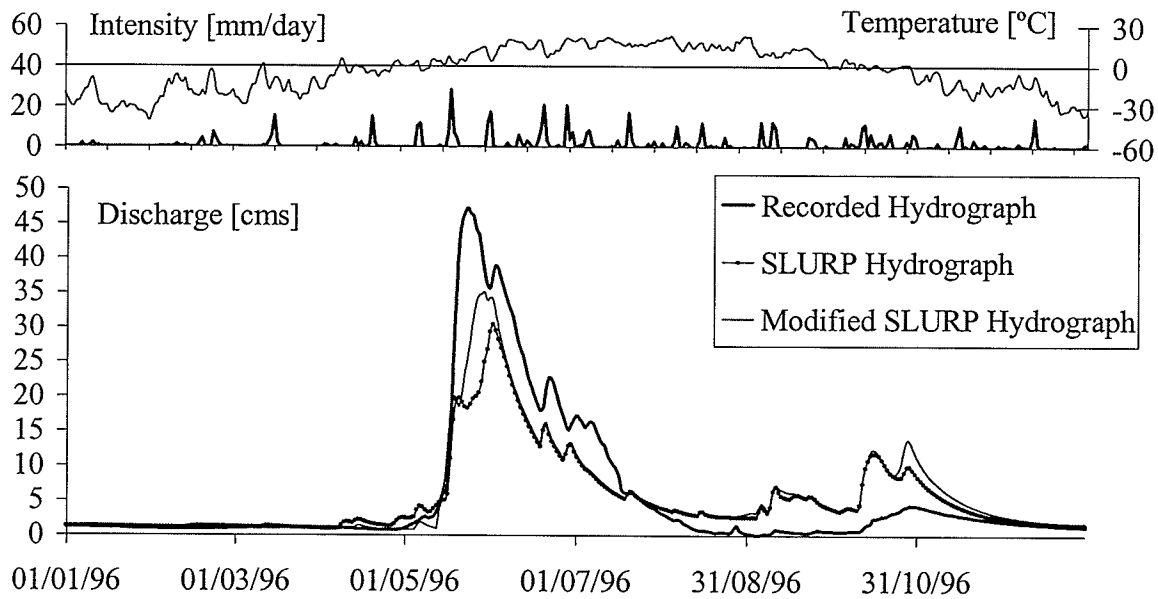


Figure 5.17 - 1997 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

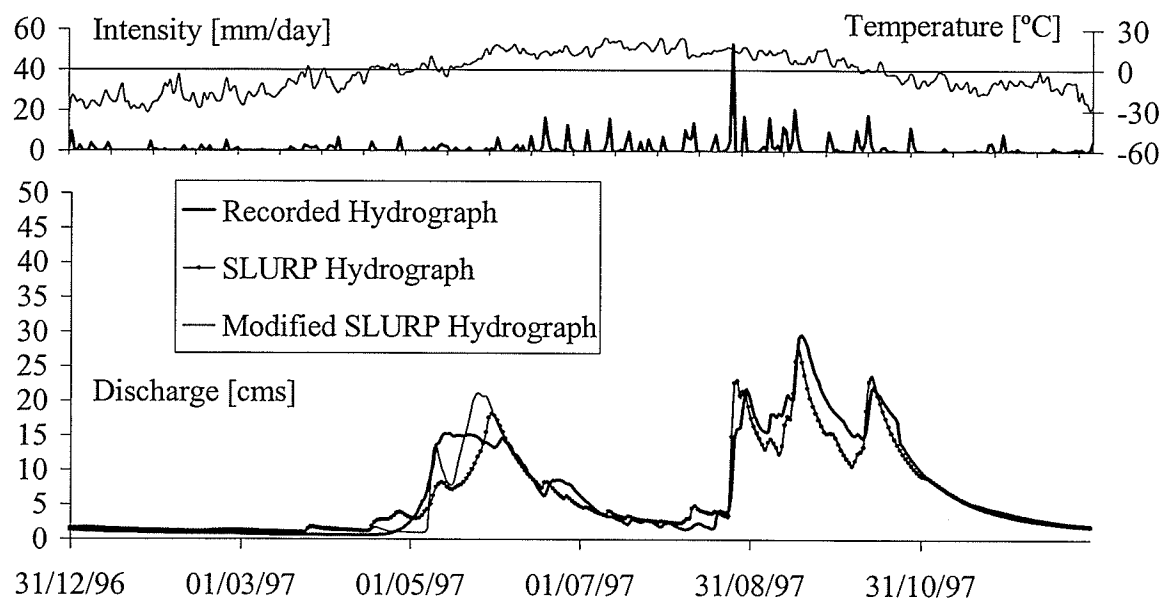


Figure 5.18 - 1998 Taylor River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

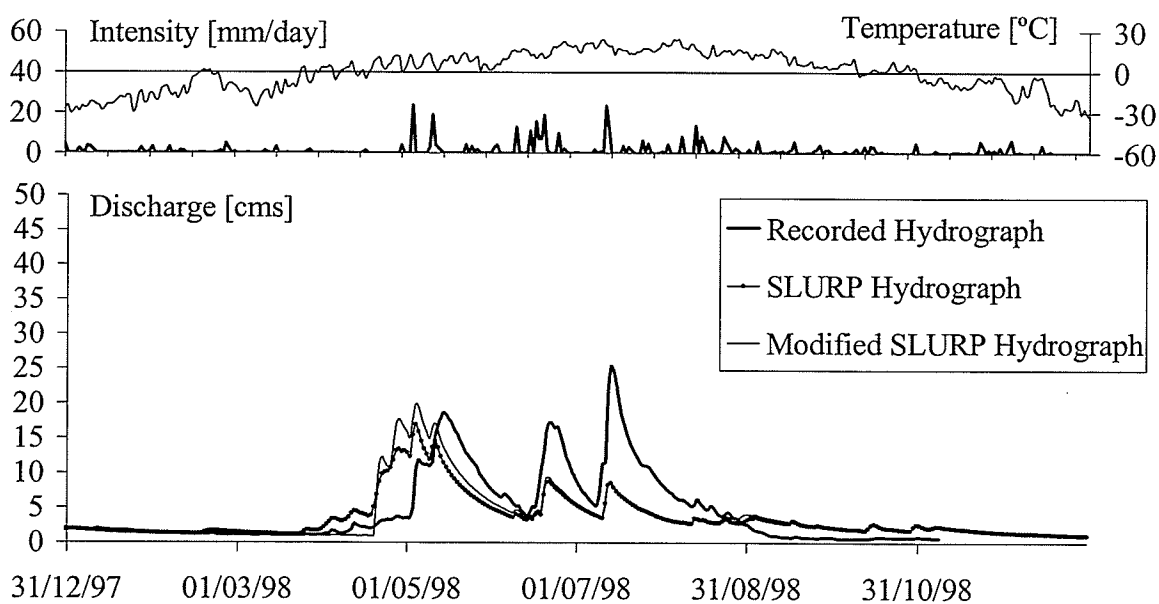


Figure 5.19 - 2000 Taylor River: Recorded Hydrograph vs. Calibrated SLURP And Modified SLURP Hydrographs.

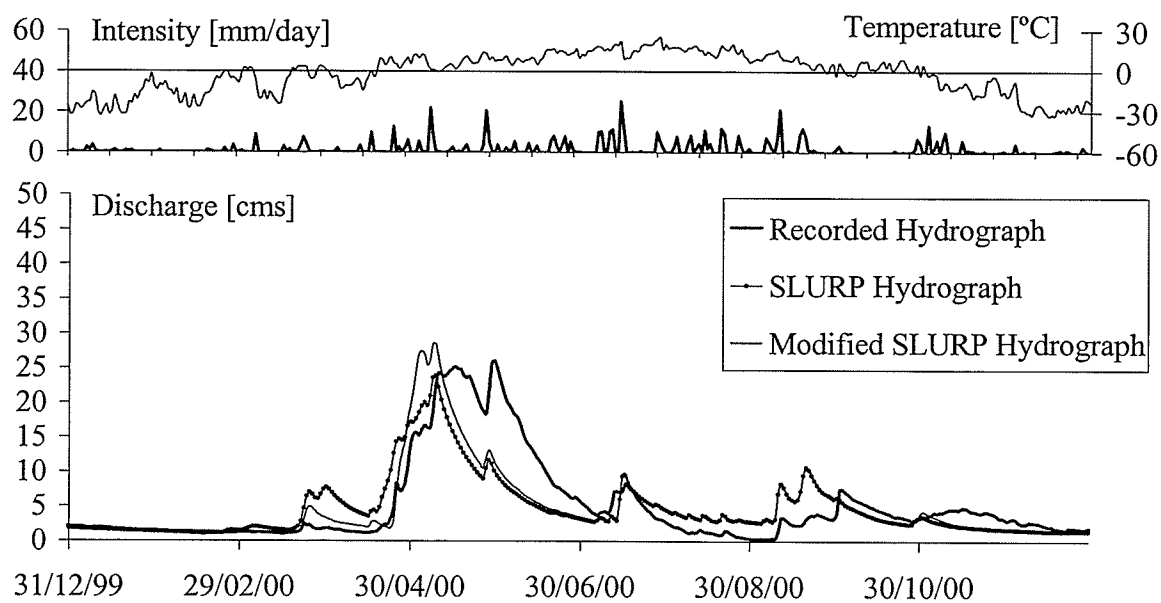


Figure 5.20 – Snowpack depletion curves Original SLURP Model – Taylor River, 1991.

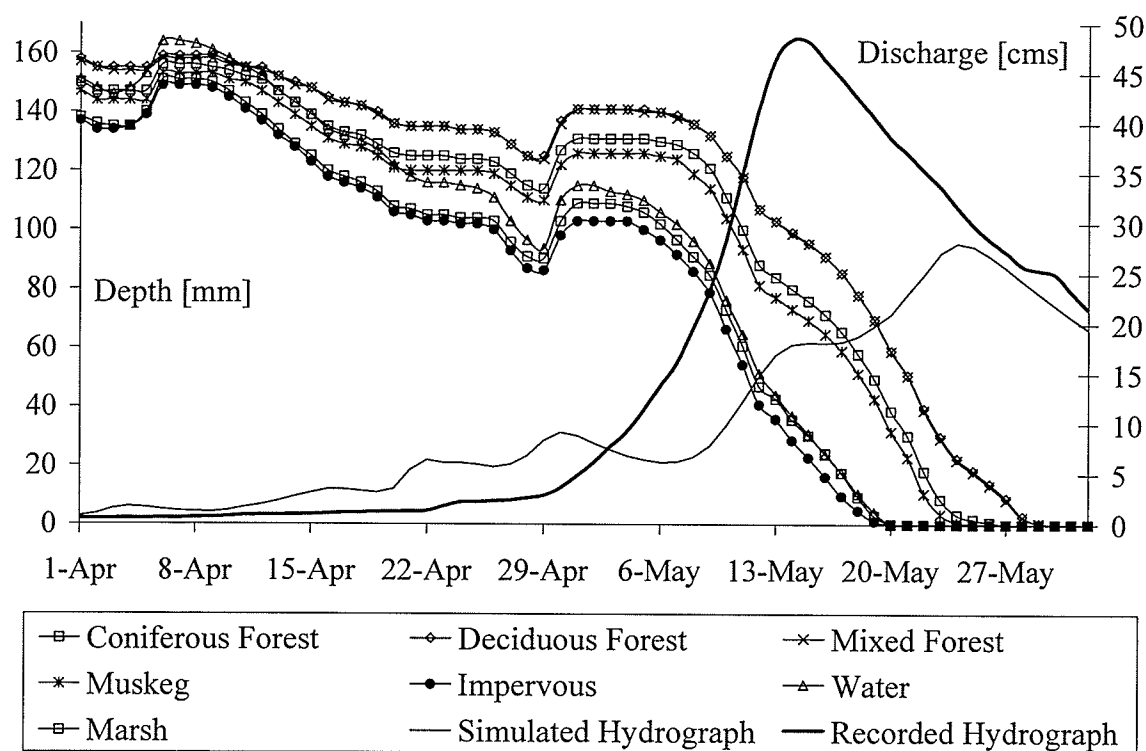


Figure 5.21 – Snowpack depletion curves Modified SLURP Model – Taylor River, 1991.

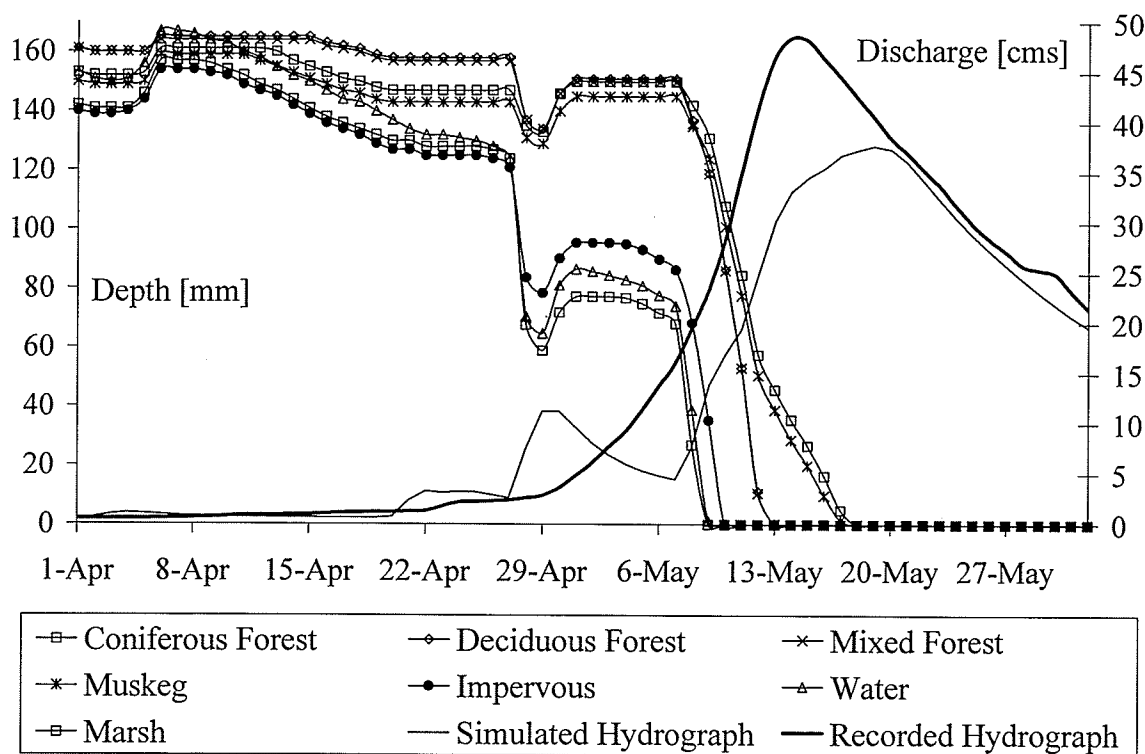


Figure 5.22 – Snowpack depletion curves Original SLURP Model – Taylor River, 1996.

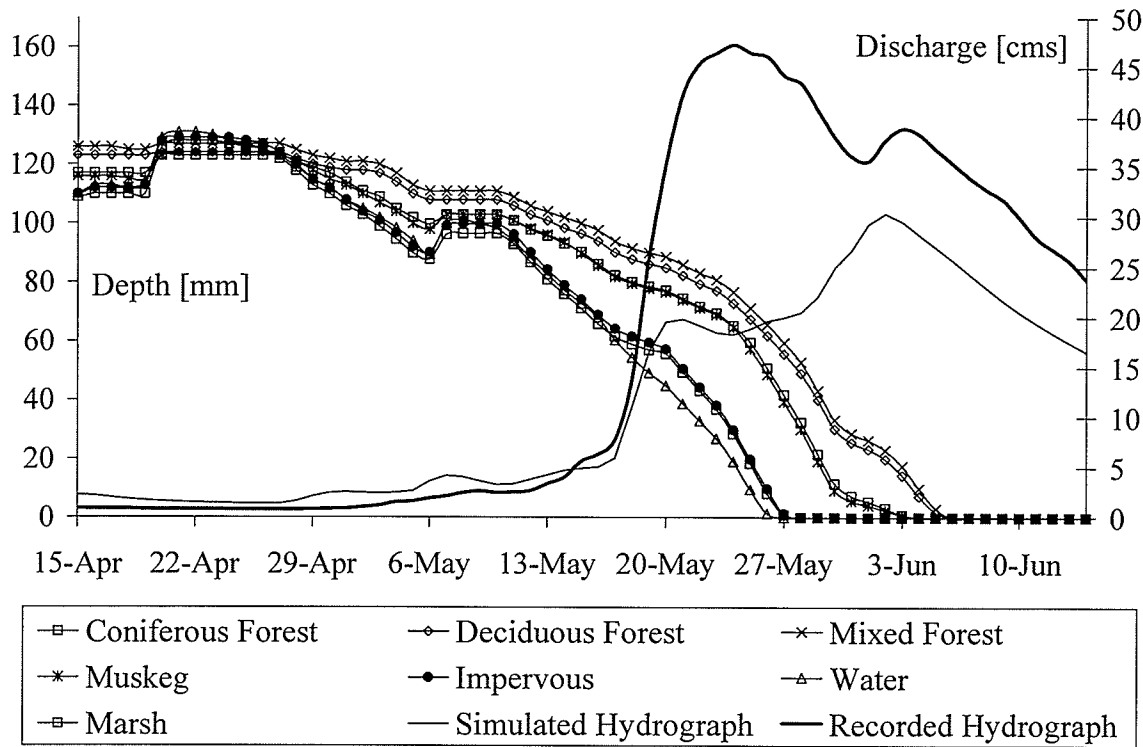


Figure 5.23 – Snowpack depletion curves Modified SLURP Model – Taylor River, 1996.

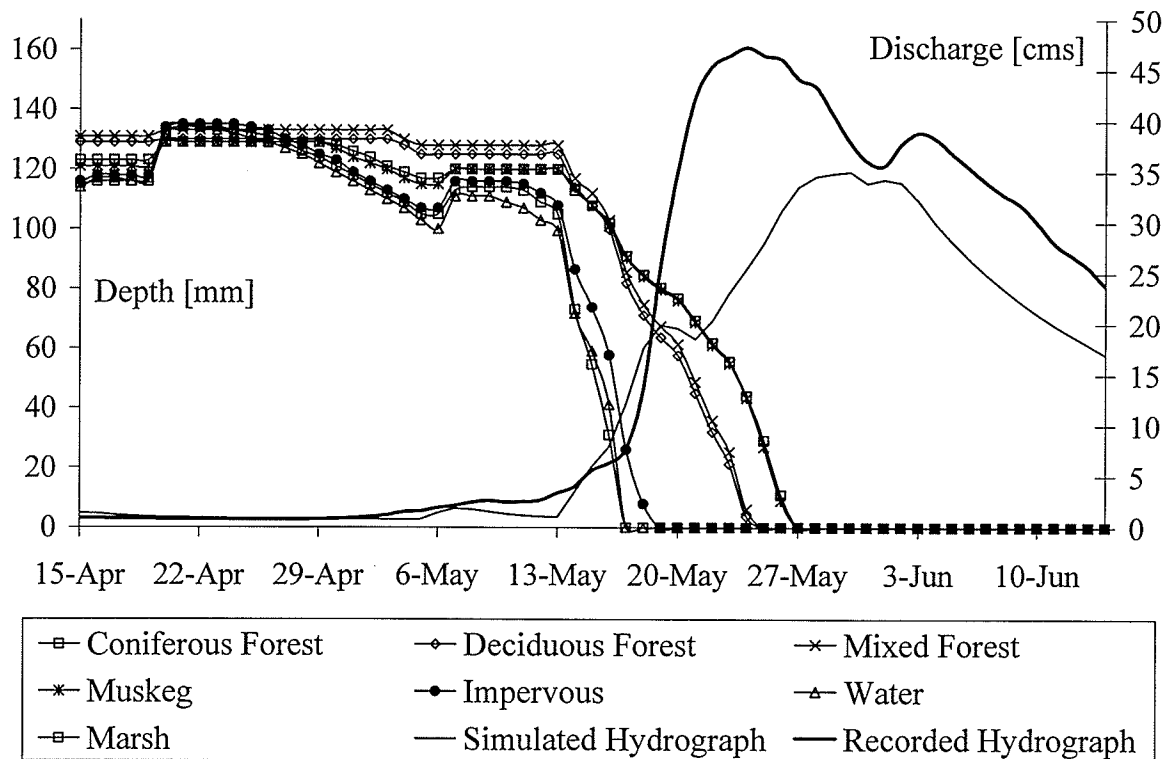
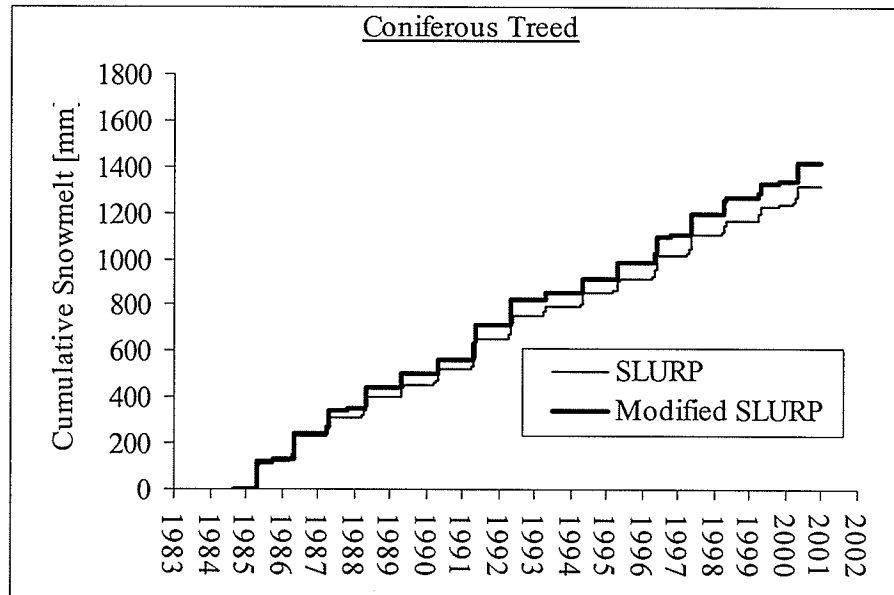


Figure 5.24 – Snowmelt accumulation curves for the SLURP and Modified SLURP Model – Taylor River, 1996; (a) Coniferous Treed; (b) Water.

(a)



(b)

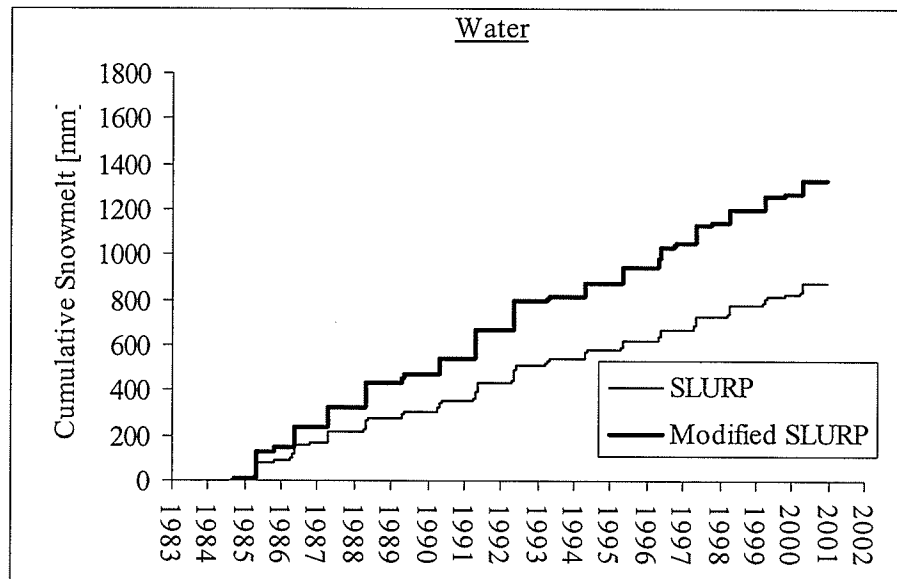


Figure 5.25 - 1985 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

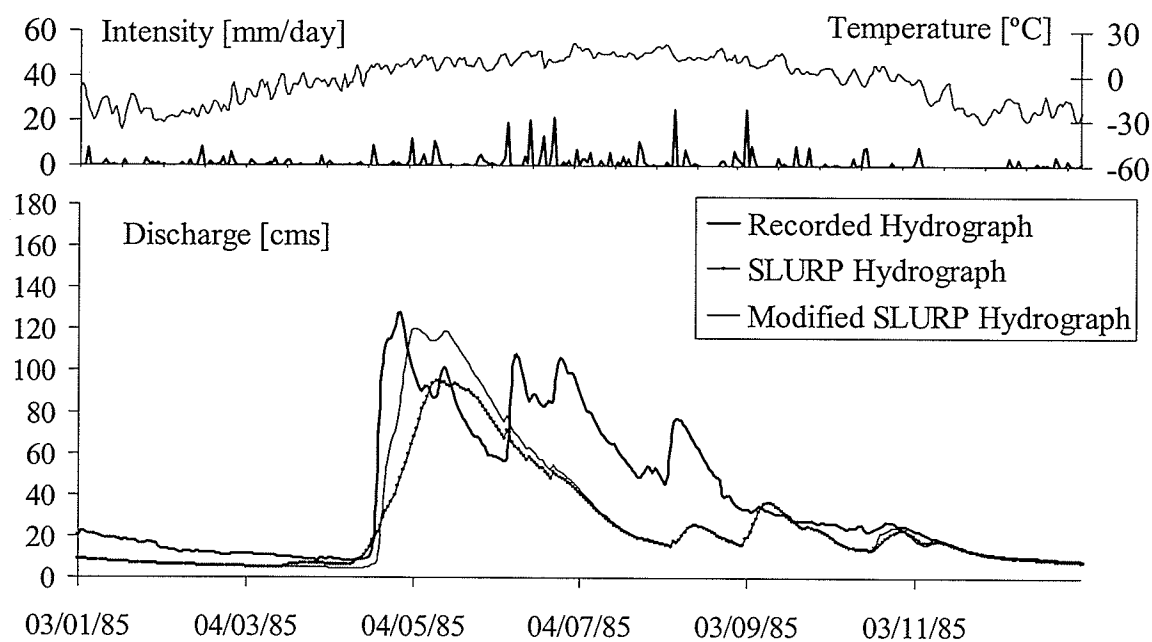


Figure 5.26 - 1986 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

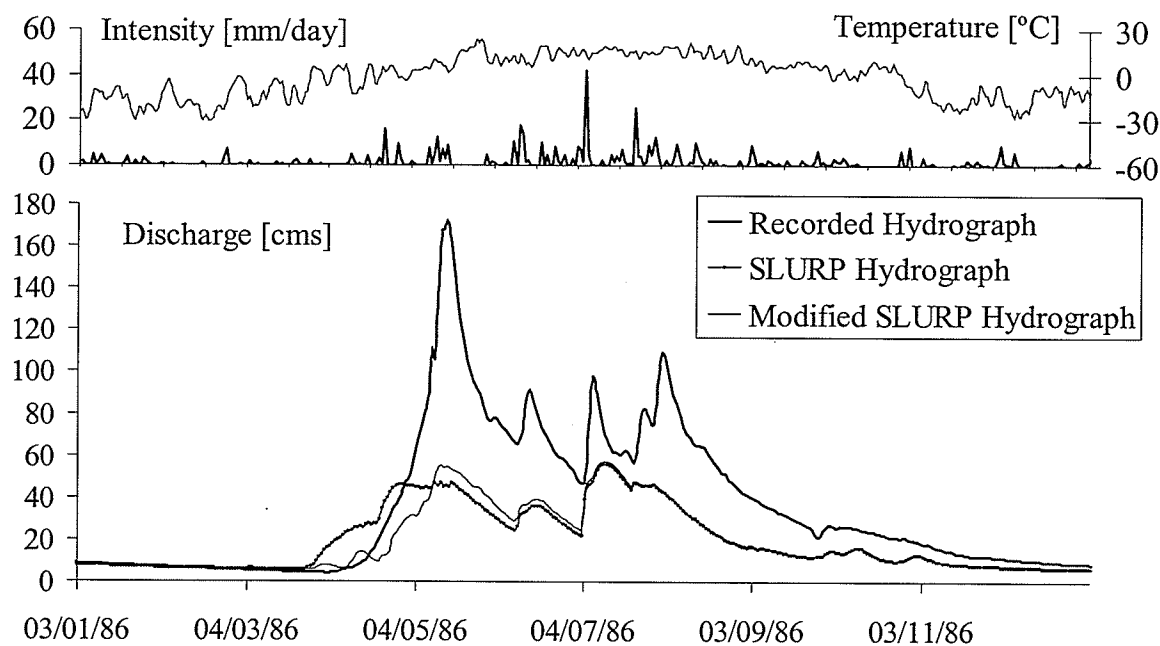


Figure 5.27 - 1987 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

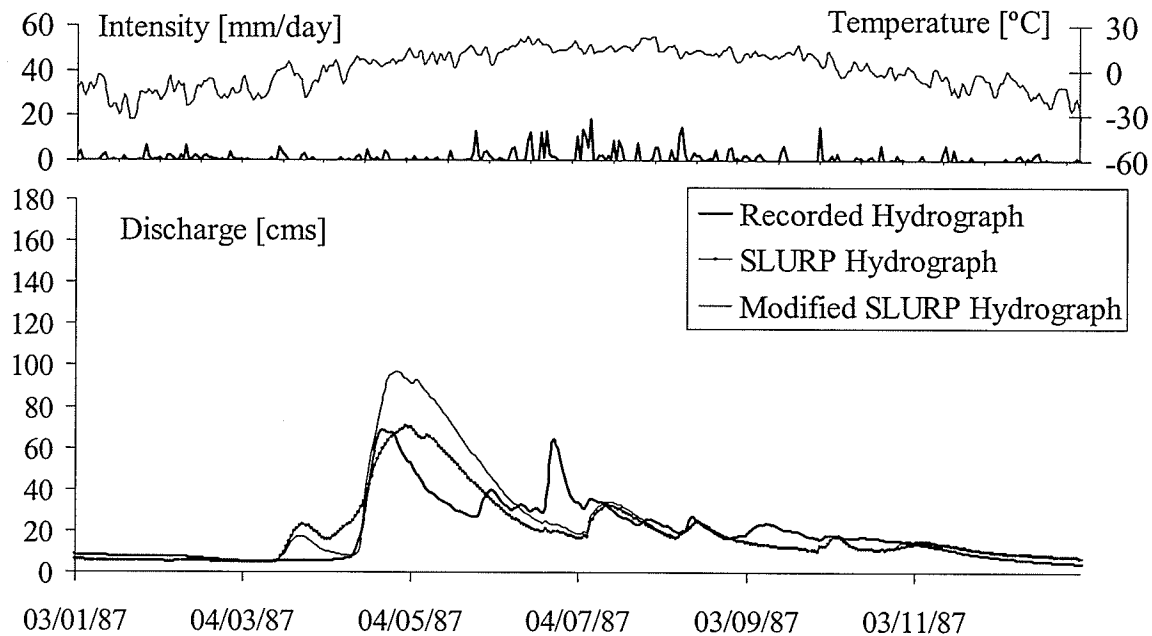


Figure 5.28 - 1988 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

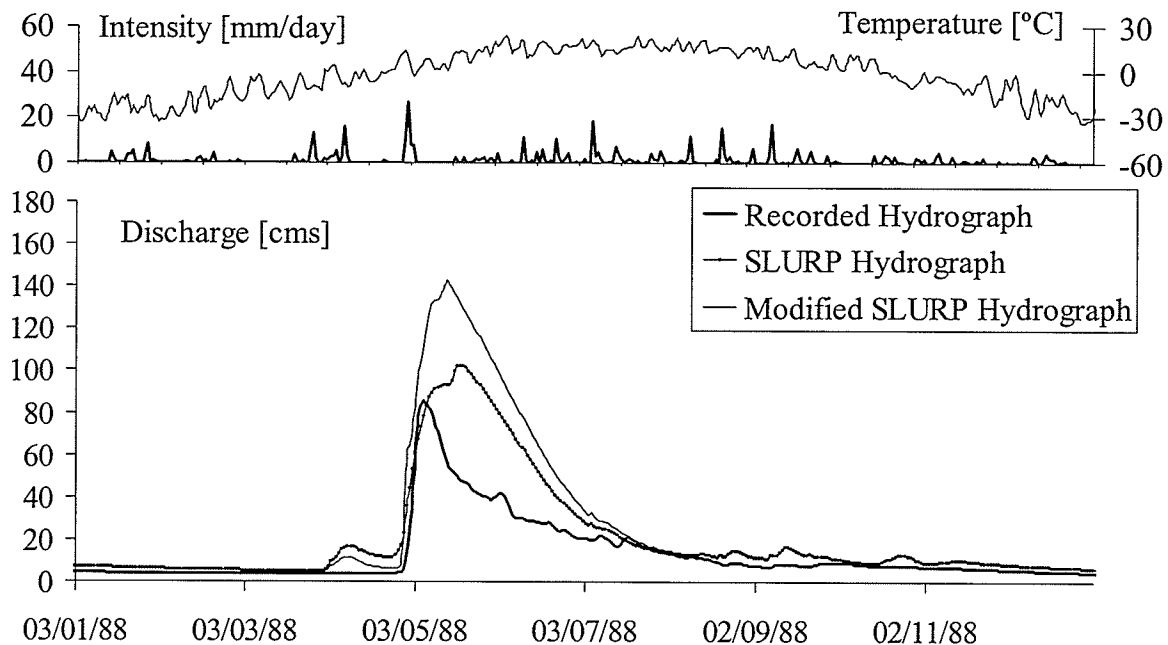


Figure 5.29 - 1989 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

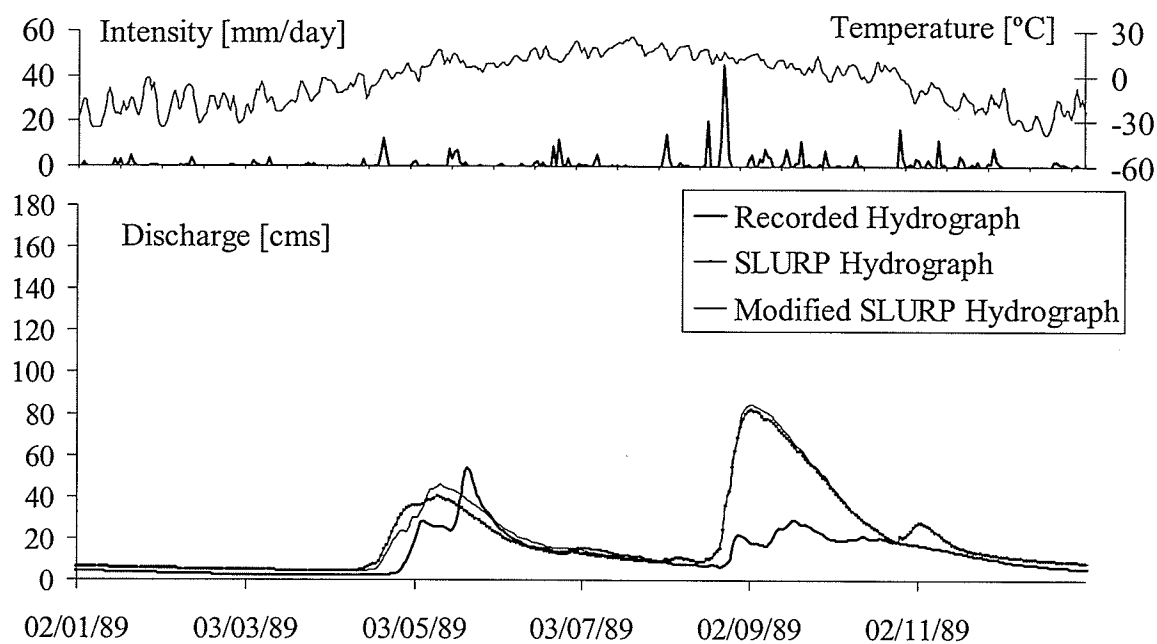


Figure 5.30 - 1990 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

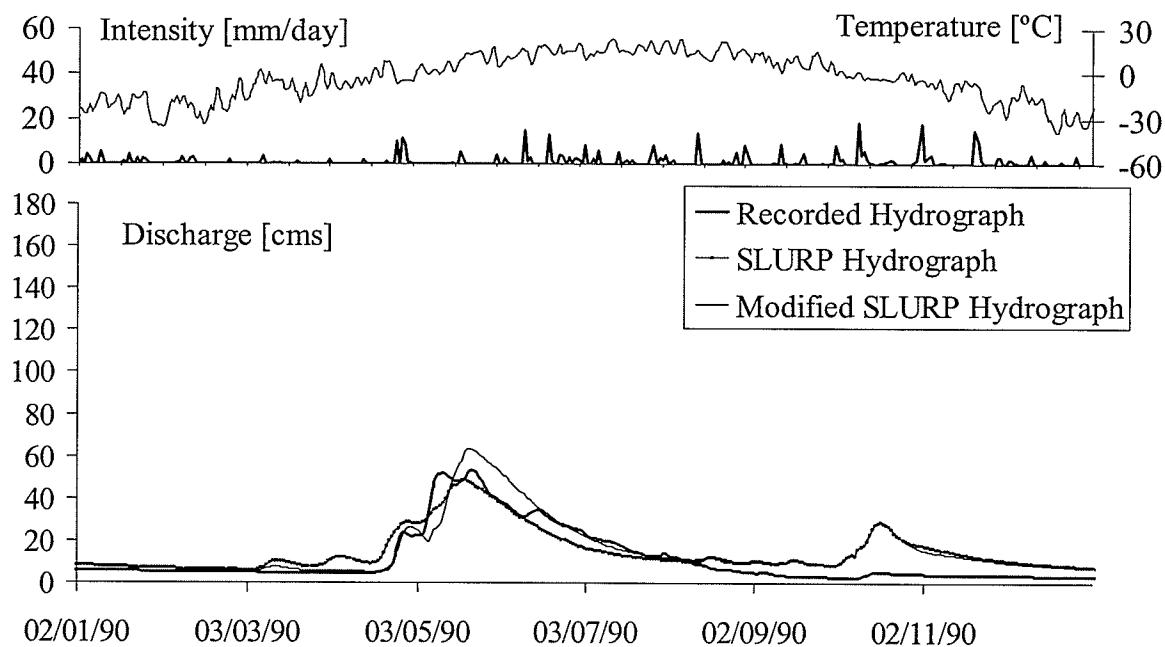


Figure 5.31 - 1991 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

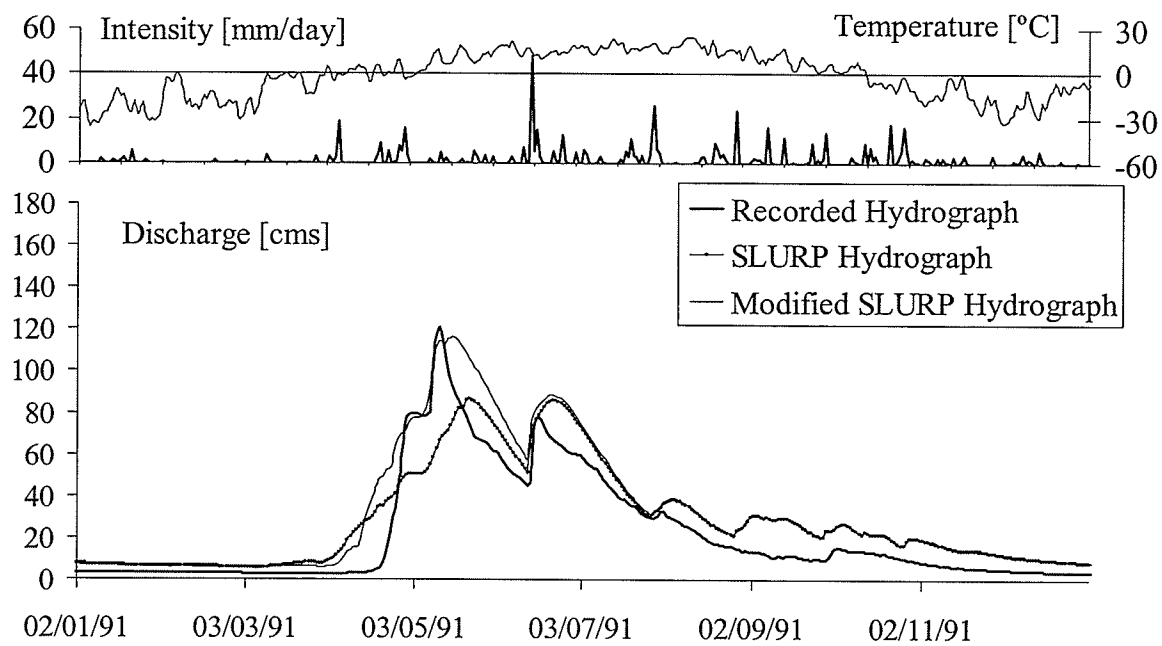


Figure 5.32 - 1992 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

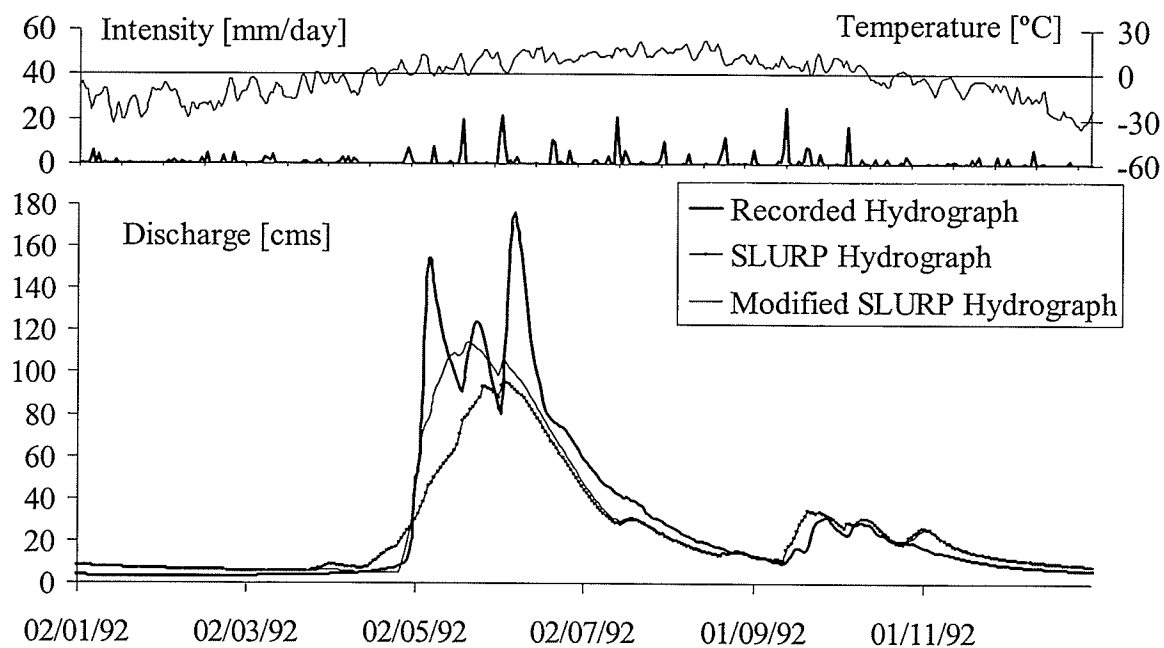


Figure 5.33 - 1993 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

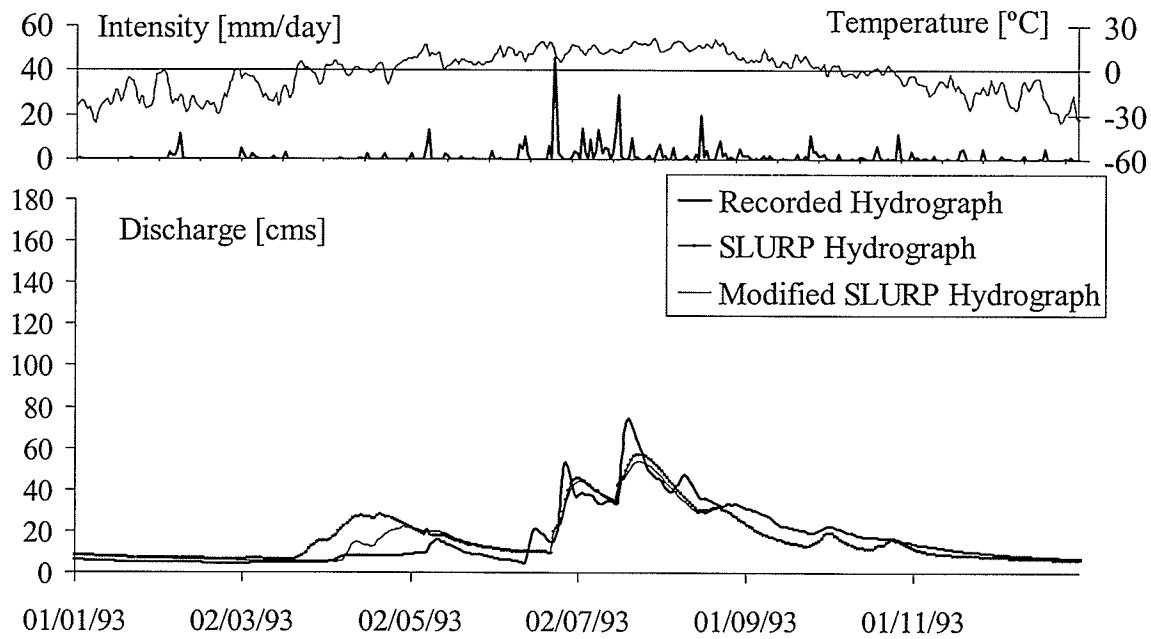


Figure 5.34 - 1994 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

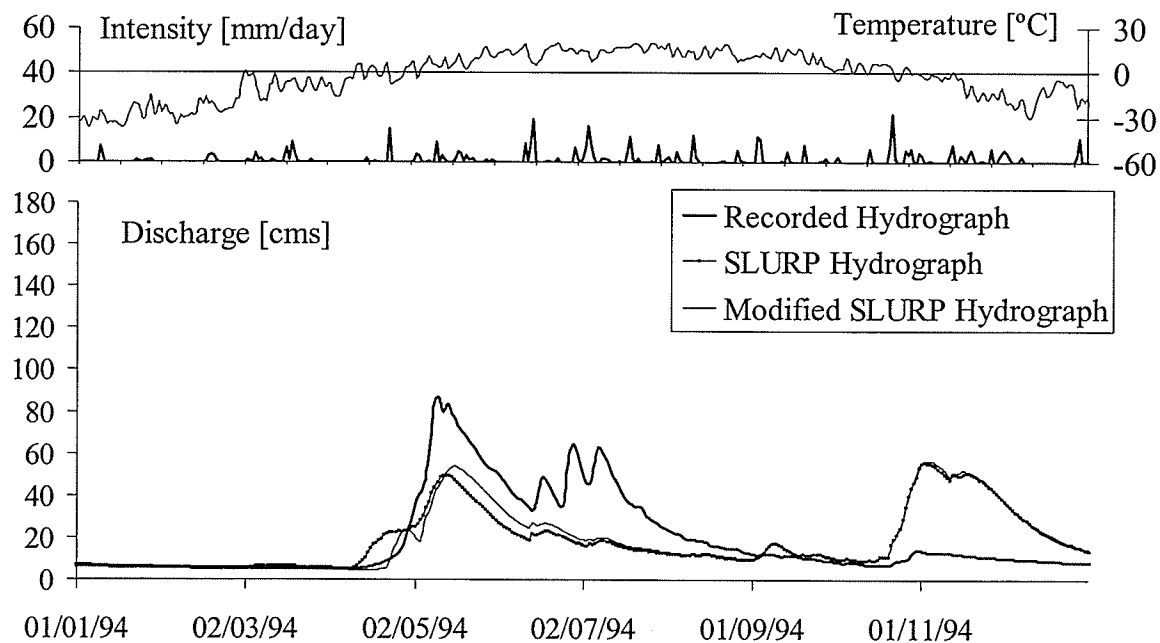


Figure 5.35 - 1995 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

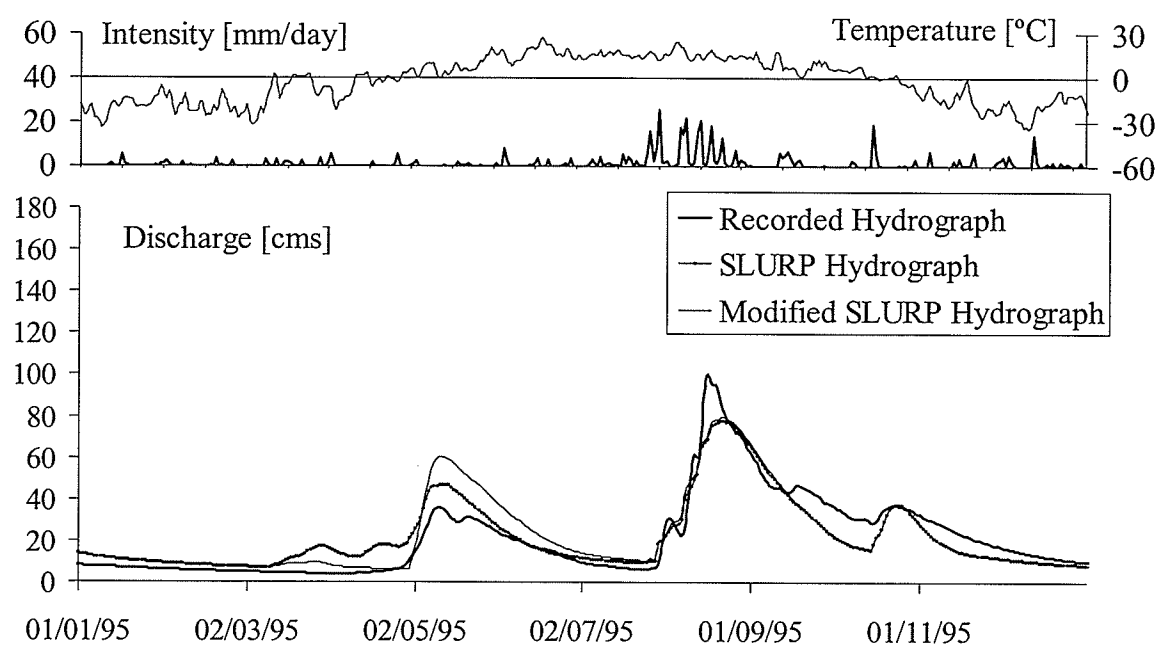


Figure 5.36 - 1996 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

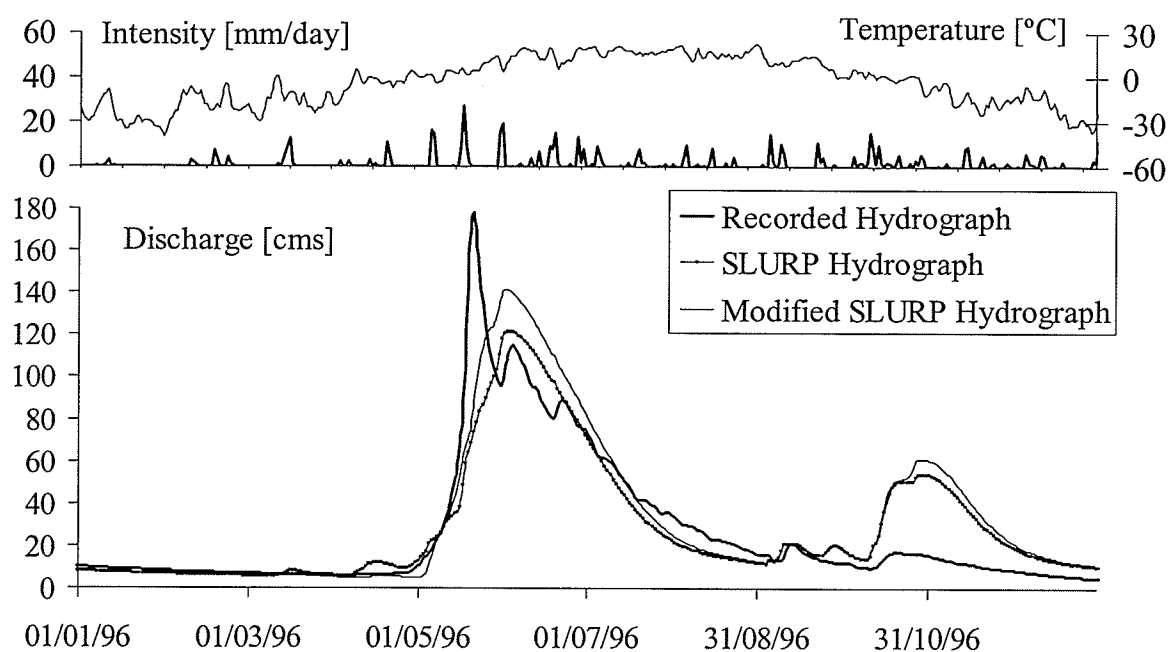


Figure 5.37 - 1997 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

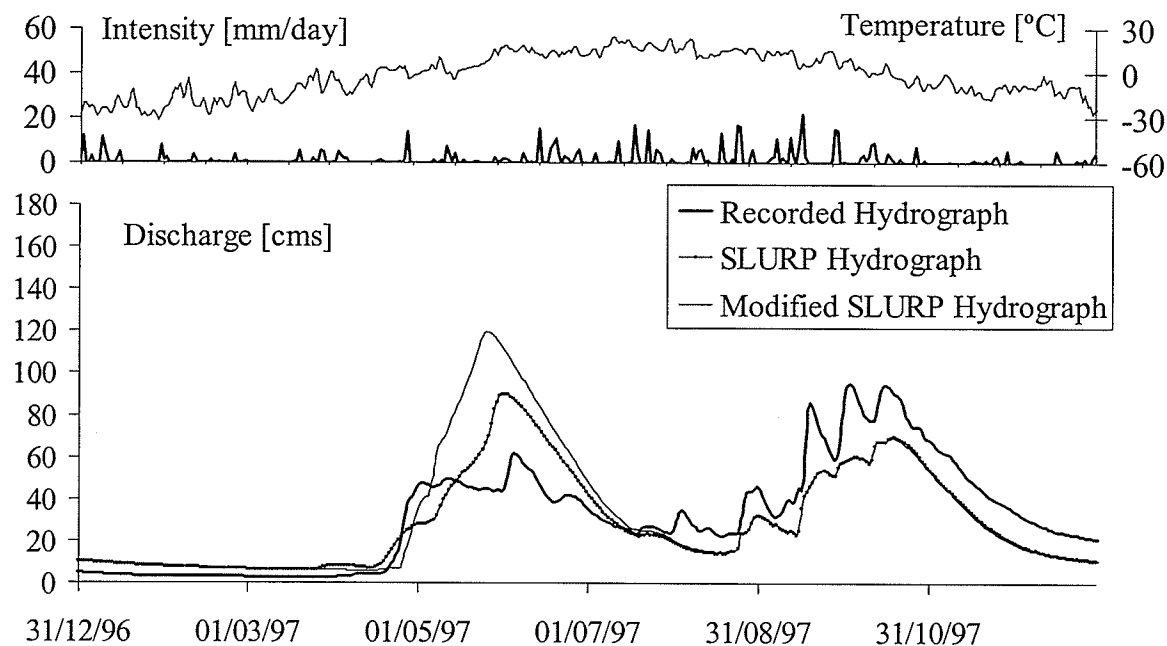


Figure 5.38 - 1998 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

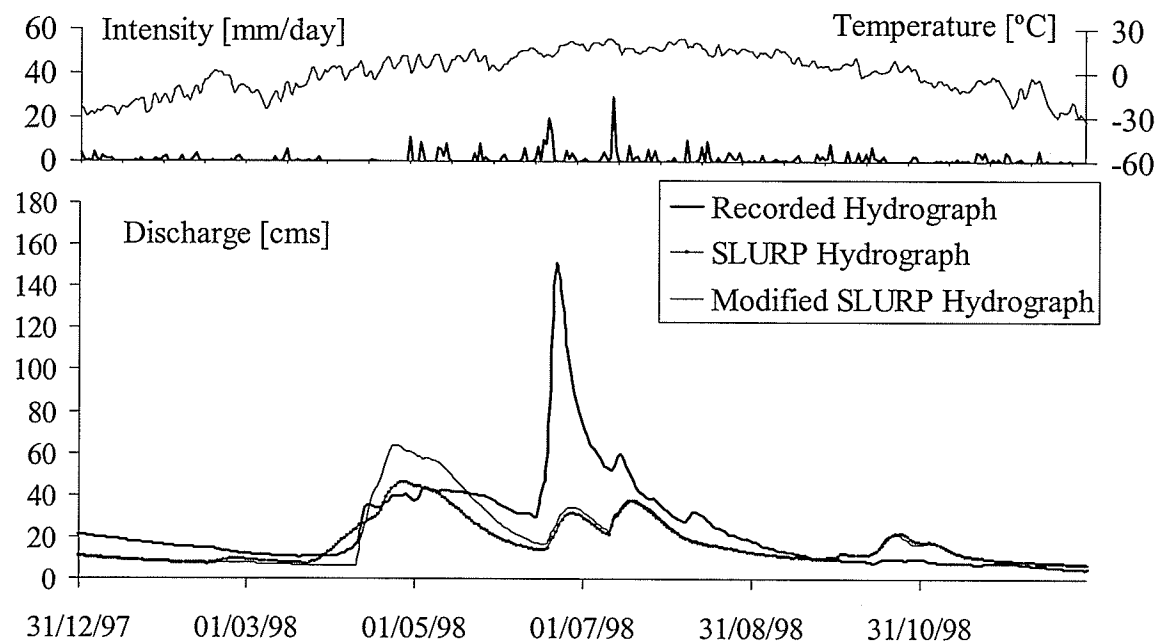


Figure 5.39 - 1999 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP and Modified SLURP Hydrographs.

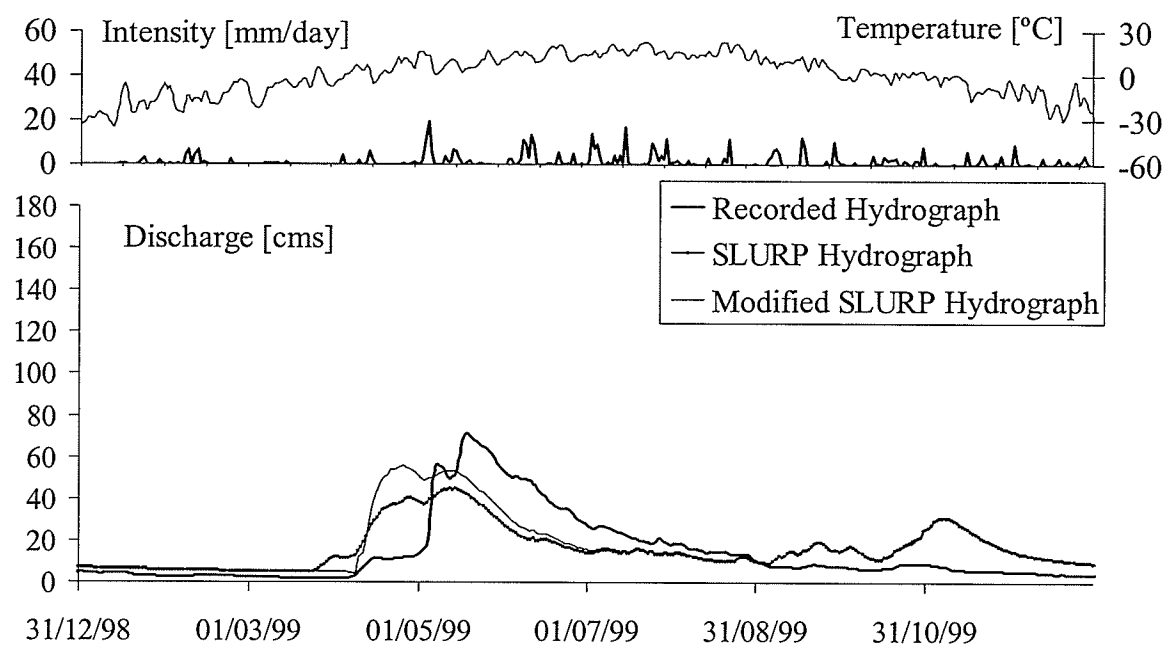


Figure 5.40 - 2000 Upper Burntwood River: Recorded Hydrograph vs. Calibrated SLURP And Modified SLURP Hydrographs.

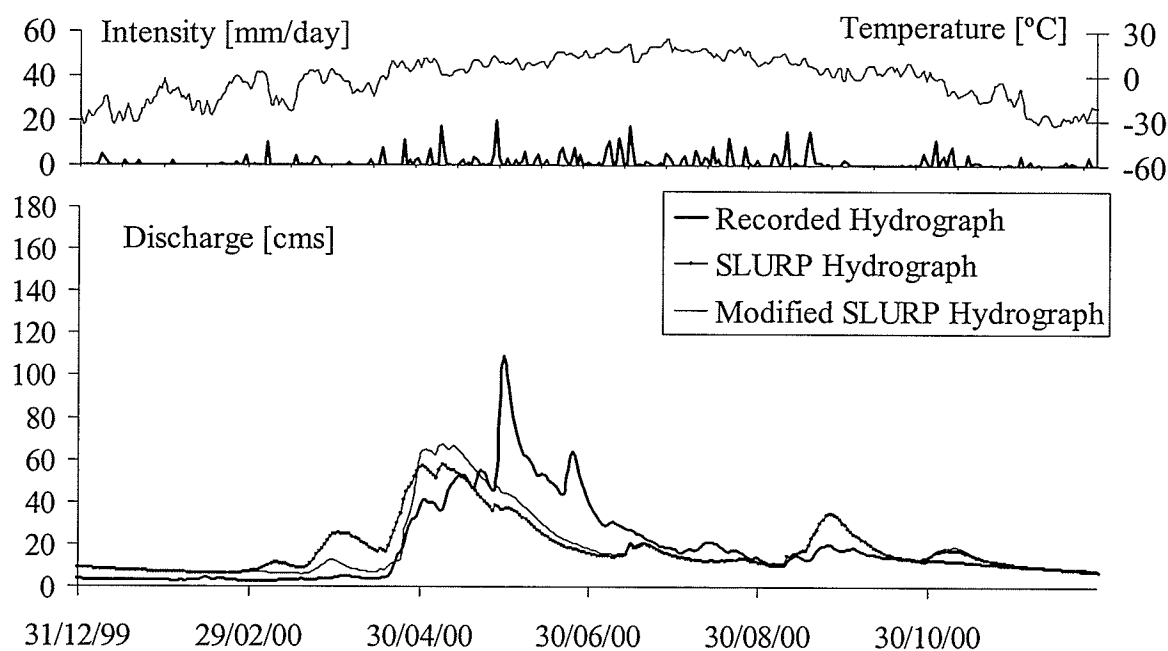


Figure 5.41 – Water balance calculation for the Modified SLURP model application to the Taylor River.

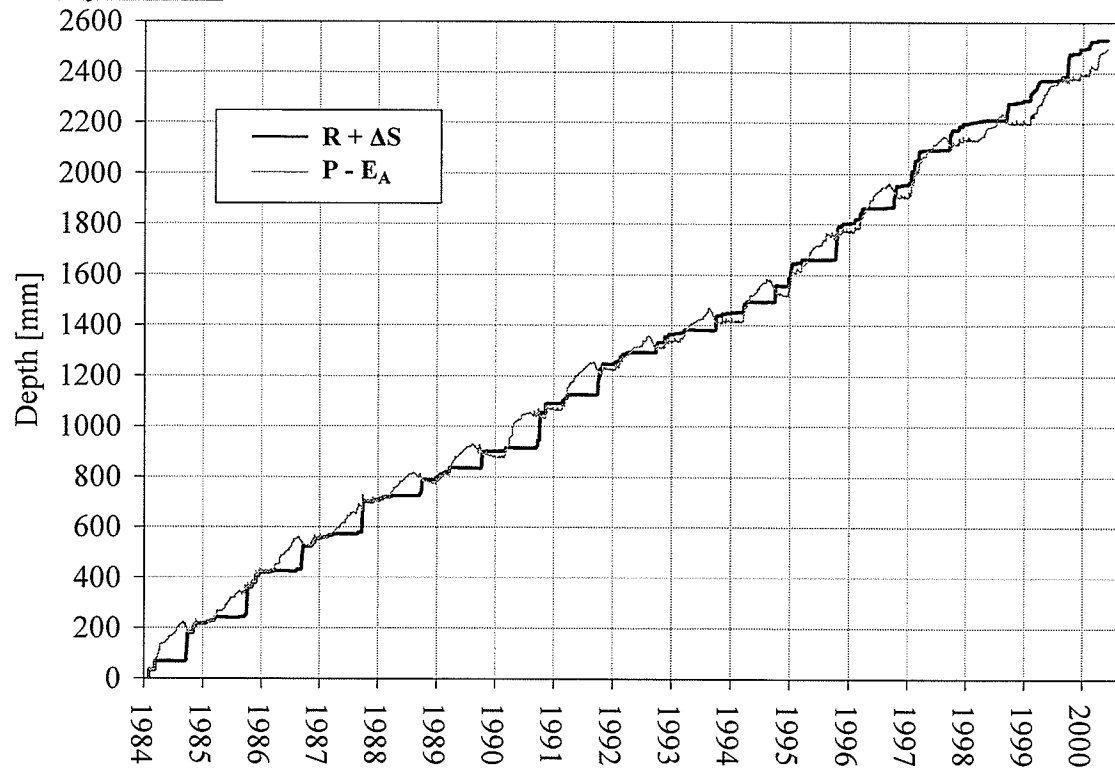


Figure 5.42 – Water balance calculation for the Modified SLURP model application to the Upper Burntwood River.

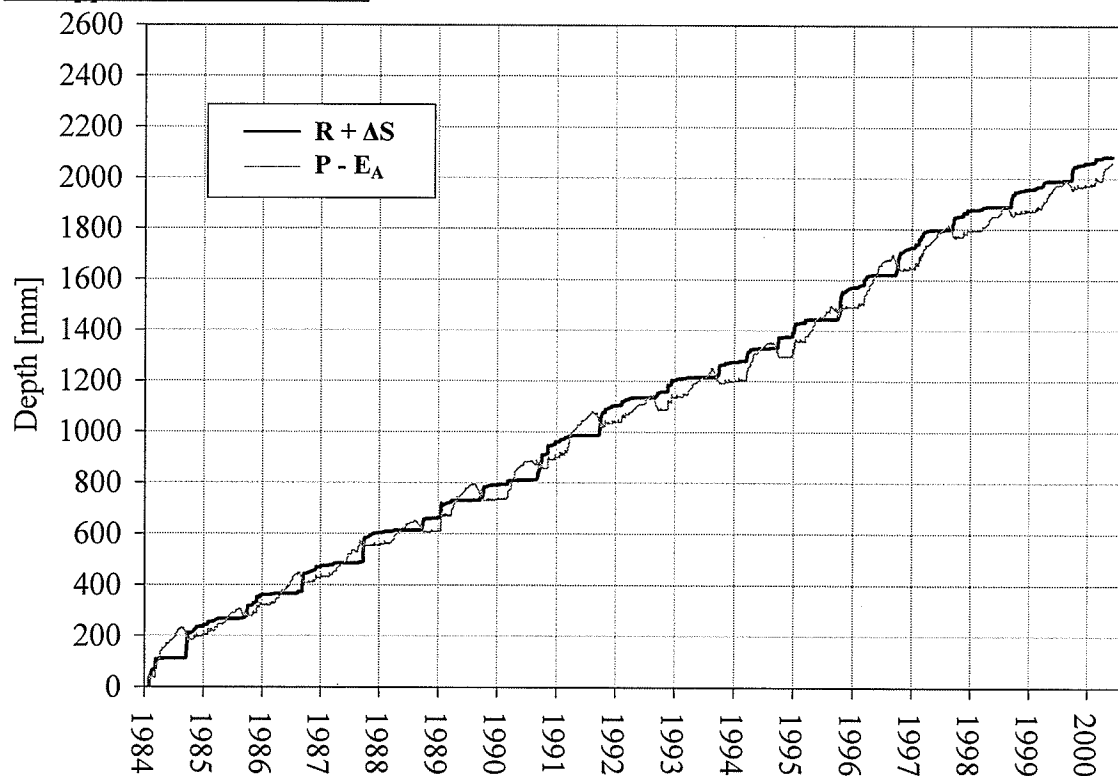


Table 5.1 – Performance Statistics for SLURP and Modified SLURP simulation of the Taylor River streamflow for 1986-2000.

Nash-Sutcliffe Efficiency

Period	1985	1986	1987	1988	1989	1990	1991	1992	1993	1994	1995	1996	1997	1998	2000	Average	Average Increase	Average Decrease
SLURP (Jan-Dec)	71.2	75.6	79.3					74.7	26.9	28.0	57.8	71.6	89.5	41.6	55.2	61.0		
Modified SLURP (Jan-Dec)	81.0	79.5	46.1					78.4	38.7	34.6	37.1	79.2	90.6	36.7	62.7	60.4		
Change in statistic	9.8	3.8	-33.2					3.7	11.8	6.7	-20.7	7.5	1.1	-4.9	7.5	-0.6	6.5	-19.6
SLURP (Apr-May)	37.8	84.9	80.9			84.5	50.4	74.2	-9.5	55.4	-279.8	61.6	65.9	14.4	47.7	28.3		
Modified SLURP (Apr-May)	69.2	95.6	-35.8			91.8	86.4	92.7	56.7	64.4	-906.9	79.8	72.8	-12.2	57.6	-14.4		
Change in statistic	31.4	10.8	-116.7			7.3	36.0	18.5	66.1	9.1	-627.1	18.2	6.9	-26.6	9.9	-42.8	21.4	-256.8
SLURP (Apr)	15.9	-1287.4	83.2	-386.9	-78.6	-2268.0	-1130.3	-5033.5	-71436.9	-2262.2	-44607.3	-4696.4	-2280.1	-3242.6	-242.1	-9256.9		
Modified SLURP (Apr)	49.5	39.3	-24.2	-135.1	-10.2	-13.6	-752.2	29.8	-1266.8	-124.4	-1400.9	-166.3	-261.0	-5150.1	25.6	-610.7		
Change in statistic	33.6	1326.7	-107.3	251.9	68.4	2254.5	378.1	5063.3	70170.1	2137.7	43206.4	4530.1	2019.1	-1907.5	267.7	8646.2	10131.4	-1007.4
SLURP (May)	-60.3	70.0	58.6			78.6	-56.0	2.0	19.9	-19.0	-1371.5	44.3	-18.9	-90.9	-429.8	-136.4		
Modified SLURP (May)	51.2	87.3	-182.4			81.8	59.1	71.3	26.1	0.8	-4035.2	70.6	-1.5	-119.1	-426.0	-332.0		
Change in statistic	111.5	17.2	-241.0			3.1	115.1	69.2	6.2	19.8	-2663.7	26.3	17.3	-28.2	3.8	-195.6	39.0	-977.6

Deviation From Volume

Period	1985	1986	1987	1988	1989	1990	1991	1992	1993	1994	1995	1996	1997	1998	2000	Average	Average Increase	Average Decrease
SLURP (Jan-Dec)	-22.1	-16.1	2.5	-0.6	21.9	13.1	-8.3	17.9	-9.6	1.1	30.8	-6.5	-5.6	-15.6	-8.4	+/- 12.0		
Modified SLURP (Jan-Dec)	-15.9	-16.5	13.8	1.1	27.0	13.5	1.0	22.8	-15.3	2.0	38.1	1.6	-2.9	-11.3	-7.2	+/- 12.7		
	6.2	-0.4	-11.3	0.5	-5.1	-0.4	7.2	-4.9	-5.7	-0.9	-7.3	4.8	2.7	4.3	1.2	-0.6	3.8	-4.5
SLURP (Apr-May)	-35.7	4.1	22.1			10.9	-32.3	-24.5	74.6	-10.5	126.0	-38.7	-18.2	14.9	-4.9	+/- 32.1		
Modified SLURP (Apr-May)	-22.3	-0.6	51.1			-6.3	-16.3	-5.1	32.5	-13.4	163.3	-28.6	-2.1	22.9	-3.7	+/- 28.3		
	13.4	3.6	-29.0			4.6	16.0	19.4	42.1	-2.9	-37.2	10.1	16.1	-8.0	1.2	3.8	14.1	-19.3
SLURP (Apr)	-63.3	385.6	34.1	212.1	20.2	230.9	174.5	162.2	425.5	91.2	187.3	66.1	167.5	163.2	160.0	+/- 169.6		
Modified SLURP (Apr)	-56.2	37.5	64.4	116.8	-14.4	15.2	68.3	19.0	27.0	-27.1	32.4	3.1	54.0	119.5	44.0	+/- 46.6		
	7.1	348.2	-30.4	95.3	5.8	215.7	106.2	143.3	398.4	64.1	154.8	63.1	113.6	43.7	116.0	123.0	133.9	-30.4
SLURP (May)	-25.9	-13.1	15.4			-5.6	-41.5	-31.0	20.4	-20.1	116.2	-43.6	-30.2	-12.3	-26.4	+/- 30.9		
Modified SLURP (May)	-10.2	-2.3	43.6			-7.9	-20.1	-5.9	33.4	-12.1	184.2	-30.0	-5.7	5.2	-9.9	+/- 28.5		
	15.7	10.8	-28.2			-2.3	21.4	25.1	-13.0	8.0	-68.0	13.6	24.5	7.0	16.5	2.4	15.8	-27.9

Table 5.2 – Performance Statistics for SLURP and Modified SLURP simulation of the Upper Burntwood River streamflow for 1986-2000.

Nash-Sutcliffe Efficiency

Period	1985	1986	1987	1988	1989	1990	1991	1992	1993	1994	1995	1996	1997	1998	1999	2000	Average	Average Increase	Average Decrease
SLURP (Jan-Dec)	43.9	32.6	55.5	15.6	-193.3	69.0	77.4	75.4	75.7	11.4	84.6	73.8	68.1	25.6	39.9	44.4	37.5		
Modified SLURP (Jan-Dec)	51.0	39.9	14.7	-117.1	-206.9	67.1	76.6	88.2	83.3	18.9	80.3	70.3	34.1	28.2	31.9	57.7	26.1		
Change in statistic	7.0	7.3	-40.8	-132.7	-13.6	-2.0	-0.9	12.8	7.7	7.5	-4.2	-3.5	-34.0	2.6	-7.9	13.3	-11.3	8.3	-26.6
SLURP (Apr-May)	36.2	4.2	33.8	0.8	35.0	86.5	65.5	54.8	-2520.3	60.0	25.0	68.1	47.0	44.3	32.6	34.7	-118.2		
Modified SLURP (Apr-May)	53.9	16.5	-109.1	-180.2	52.8	79.2	69.1	87.2	-807.7	68.4	-32.3	77.1	-182.8	-40.1	1.6	52.0	-49.6		
Change in statistic	17.8	12.3	-142.9	-181.0	17.7	-7.3	3.5	32.4	1712.6	8.4	-57.3	9.0	-229.7	-84.4	-31.0	17.3	68.6	203.5	-104.8
SLURP (Apr)	21.6	-31.4	82.5	-37064.8	-12585.6	-50.8	-26.6	-1356.2	-23160.6	-201.8	-6815.3	-3182.2	70.3	78.1	-2083.3	-225.5	-5408.2		
Modified SLURP (Apr)	54.0	52.3	76.5	-14789.8	-6377.9	95.3	-101.8	-61.7	-4299.1	59.3	-318.7	-305.0	56.7	-40.8	-4787.0	61.0	-1914.2		
Change in statistic	32.4	83.7	-6.0	22275.0	6207.7	146.1	-75.2	1294.6	18861.5	261.2	6496.7	2877.2	-13.6	-118.9	-2703.7	286.4	3494.1	5347.5	-583.5
SLURP (May)	-166.4	-256.5	-336.0	-346.4	-67.4	61.3	-207.6	-145.8	-744.4	-134.8	-170.6	45.1	-11899.3	-2976.6	-46.5	-81.3	-1092.1		
Modified SLURP (May)	-195.8	-217.1	-1501.7	-1229.8	-44.2	-23.6	-75.5	29.7	-881.5	-98.1	-814.5	60.4	-68460.0	-3987.5	-30.0	-109.4	-4848.7		
Change in statistic	-29.5	39.5	-1165.7	-883.4	23.3	-84.9	132.2	175.5	-137.1	36.7	-644.0	15.3	-56560.7	-1010.9	16.5	-28.1	-3756.6	62.7	-6727.1

Deviation From Volume

Period	1985	1986	1987	1988	1989	1990	1991	1992	1993	1994	1995	1996	1997	1998	1999	2000	Average	Average Increase	Average Decrease
SLURP (Jan-Dec)	-34.6	-44.0	0.6	57.5	60.6	29.5	29.2	-14.5	4.6	-5.5	3.2	8.6	-8.3	-29.4	11.2	-1.4	+/- 21.4		
Modified SLURP (Jan-Dec)	-27.0	-46.4	13.0	85.3	66.0	35.5	44.1	-5.4	-3.3	-3.6	6.4	19.2	2.6	-22.7	21.7	-1.6	+/- 25.2		
Change in statistic	7.7	-2.3	-12.3	-27.8	-5.4	-6.0	-15.0	9.1	1.3	1.8	-3.2	-10.6	5.6	6.7	-10.5	-0.2	-3.8	5.4	-9.3
SLURP (Apr-May)	-20.7	-40.6	38.6	69.9	34.7	7.8	-0.7	-31.4	117.1	-26.6	62.0	-24.8	10.5	-6.1	9.2	24.9	+/- 32.8		
Modified SLURP (Apr-May)	1.8	-53.2	69.6	116.5	36.7	2.2	34.4	-8.8	64.7	-30.3	61.3	-19.3	59.7	17.0	32.2	20.2	+/- 39.3		
Change in statistic	18.9	-12.6	-31.0	-46.6	-2.0	5.6	33.7	22.6	52.4	-3.7	0.7	5.6	-49.3	10.9	-23.0	4.7	-0.8	17.2	-24.0
SLURP (Apr)	-58.0	88.6	21.2	242.2	278.6	94.5	133.3	102.4	206.7	69.2	210.5	41.8	24.2	9.1	228.5	170.4	+/- 123.7		
Modified SLURP (Apr)	-51.8	-20.3	27.6	126.3	201.9	7.1	171.3	11.1	67.0	2.5	39.6	-15.2	-19.3	20.3	296.0	45.3	+/- 70.2		
Change in statistic	6.3	68.4	-6.5	115.9	76.7	87.4	-38.0	91.3	139.7	66.7	170.9	26.7	4.9	-11.2	-67.5	125.1	53.5	81.7	-30.8
SLURP (May)	-4.4	-58.5	51.6	58.1	13.0	-7.0	-18.0	-38.7	55.3	-38.3	36.2	-30.5	7.6	-15.9	-23.4	-3.5	+/- 28.8		
Modified SLURP (May)	25.1	-57.7	100.9	115.8	22.1	1.4	16.7	-9.9	63.1	-34.3	65.1	-19.6	76.1	14.8	-7.0	15.3	+/- 40.3		
Change in statistic	20.6	0.7	-49.3	-57.7	-9.0	5.6	1.3	28.8	-7.9	4.0	-28.9	10.9	-68.5	1.0	16.4	11.8	-7.5	10.1	-36.9

Table 5.3 – Average annual simulated water balance components for the Modified SLURP application to the Taylor River and Upper Burntwood River.

Water Balance Component	Taylor River	Upper Burntwood River
Evapotranspiration [mm]	355	364
Precipitation [mm]	501	482
Runoff [mm]	154	123
Observed Runoff [mm]	150	119
Snowfall [mm]	124	115
Rainfall [mm]	378	367
Sublimation [mm]	37	47
Canopy Evaporation [mm]	276	278
Soil Evaporation [mm]	41	38
Snowmelt [mm]	87	68

CHAPTER 6

6 CONCLUSIONS AND RECOMMENDATIONS

6.1 General

Estimates of water yield in the High Boreal Forest of Northern Manitoba are of primary interest to local water resource users and developers such as Manitoba Hydro. Many of the watersheds in northern Manitoba are remote and ungauged. Meteorological stations that provide the necessary data to support estimates of water yield are sparse. The region contains a variety of land covers, topography, soil conditions, and unique hydrologic behavior. An accurate deterministic hydrological model that is capable of reproducing the regional hydrologic behavior is required in order to support activities such as hydropower operations, hydropower planning studies, and dam safety.

The first goal of this research has been to review the hydrological processes of the High Boreal Forest in northern Manitoba in order to support the application of a hydrological model. Two recent field studies, NOPEX and BOREAS, resulted in a large volume of new information that characterizes the hydrology of the Boreal Forest. Significant contributions from these programs as well as other programs include:

1. The terrain is hummocky, has little relief, and is covered mainly with muskeg, coniferous and deciduous forest. A typical soil profile consists of an organic moss and peat layer overlying a less permeable silty clay layer.
2. The hummocky terrain contains many small perched clay lined depressions that may not be hydraulically connected to the drainage system. Their contribution to the total evaporation is a key component of the boreal forest water balance dynamics. Once the depressions are filled with water, runoff is generated with additional precipitation and snowmelt inputs.
3. Moss layer rain interception is significant and results in considerable evaporation losses. Trees and moss are able to intercept and store small precipitation events.
4. The hydraulic conductivity of the Muskeg organic soils increases exponentially with depth resulting in a very marked decrease in interflow as the water table falls towards the relatively impermeable inert layers. Runoff from Muskeg virtually ceases once the water table reaches a limiting depth.
5. Boreal forests have the ability to enhance ET by releasing energy stored in the canopy when moisture levels are high. During dry conditions, the energy is released to warm the environment. ET from coniferous trees is relatively small during the growing season.
6. The formation and degradation of a frost active layer is a dynamic process and often forms in Muskeg. It freezes and thaws in response to the air and surface temperatures, vegetation cover, snow cover, local moisture regimes, and subsurface soil properties. Soil frost may affect the water regime by modifying the infiltration characteristics, temporarily storing water as ice, and varies the upper soil storage capacity as the thaw depth changes. A frost layer develops over the course of the winter and the depth varies inversely with snowpack depth. Unique types of muskeg are underlain by frost year round.
7. Overland flow is commonly generated during the snowmelt period because infiltration capacities are greatly reduced when concrete frost is produced. Honey comb frost forms in dry soils resulting in no change, or an increase in infiltration capacities.

The second goal of this research has been to apply a deterministic hydrological model to reproduce the hydrologic behavior of the High Boreal Forest of northern Manitoba. The SLURP model is proven to be a flexible model that performs well at a variety of scales and on a wide variety of basins and types of land covers located around the world. Some of those basins are similar to the area of interest. The conceptual representation of the watershed is physically based and contains relatively few calibration parameters.

6.2 Conclusions

The SLURP model was used to simulate the hydrologic response from two basins located in the High Boreal Forest of northern Manitoba. Based on the findings of this research the following conclusions can be drawn:

- 1) The streamflow response from the two basins is highly variable on a seasonal basis. Streamflow approaches base flow conditions during the winter months as the precipitation is stored in the snow pack. The spring freshet hydrograph is characterized by a steep rising limb resulting from a rapid snowmelt period and is followed by a longer recession limb. The spring freshet normally produces the peak annual discharge and the majority of the annual streamflow. During the summer and fall months most small and moderate rain events do not generate any runoff as a result of canopy and moss interception, depression storage and subsequent evaporation. Only the largest rain events or a series of moderate rain events generate runoff. During significant rain events, the summer hydrographs are characterized by a steep rising limb and a longer receding limb but short compared to the spring freshet. This indicates that the region has a flashy runoff response and a relatively small soil storage capacity.
- 2) A conceptual deterministic hydrologic model such as SLURP can be used to simulate the runoff response in the High Boreal Forest of northern Manitoba. A set of model parameters was established for the seven dominant land covers of the Taylor River and Upper Burntwood River. The majority of the parameters

were derived from the literature and the remaining were calibrated. The parameters are considered to be reasonable and physically realistic.

- 3) The SLURP model performs satisfactorily for the 1985-2000 test period considering:
 - the magnitude of the natural streamflow varies considerably from season to season due to the northern climate;
 - that a single parameter set was used for a large number of test years containing diverse meteorological conditions that range from dry to moderately wet;
 - the scarcity of precipitation gauges resulted in using gauges within a 430 km diameter region. Significant variation in precipitation exists between these gauges;
 - a single streamflow gauge was used for each basin during the model calibration and validation process;
 - that the majority of model parameters are common for both test basins;
 - the complex hydrologic processes that exist in the region of interest, some of which are not accounted for in the model;
 - the model was executed as one simulation from September 1984 to December 2000 and no backward correction of model state variables was applied during the simulation.
- 4) The differences in subsurface storage and interflow characteristics among the land covers are reflected in the subsurface model parameters. The forested covers have smaller upper soil storage capacity but higher retention constants relative to the Muskeg cover which is unable to hold water within the highly porous peat at high water levels.
- 5) The physical basis of the model and the modelling performance was improved with a few model modifications. A number of modelling deficiencies that affect the hydrograph throughout the year were identified. Three model

modifications were made of which two were found to improve the model performance during the snowmelt period:

- a) A frozen soil model was tested and discarded as it was found to generate too much overland runoff even when applied to a small portion of the basin;
- b) A snow ripening model was found to minimize snowmelt runoff during the fall months and delay snowmelt during the spring freshet as found in nature;
- c) The January and July snowmelt rates were replaced with a single date independent snowmelt rate.

The original SLURP model required small snowmelt rates in order to delay snowmelt. The snow ripening routine delays premature snowmelt and permits the use of the larger and more appropriate snowmelt rates that are recommended in the literature. The model modification resulted in a delayed streamflow response to snowmelt, a faster rise in the hydrograph, and a peak discharge that occurs earlier in the spring freshet. The modified SLURP model significantly improves the modeling performance during most test years during the month of April and to a lesser extent during the April to May period.

- 6) The parameters are considered to be appropriate for the surrounding region since the majority of the model parameters are common for both test basins. The parameters may be used to estimate water yield in neighboring basins that have similar characteristics. The parameters that do differ are the 23% higher ET parameters of the Upper Burntwood River likely due to greater evaporation from lakes and depressions. The Snowpack Temperature Penetration Constant differs by 25% between the two basins however the impact of this parameter difference is relatively minor.

A methodology is established for developing high resolution DEMs and land cover datasets for hydrological modelling in Manitoba. The method relies on mapping data that is readily available throughout the province of Manitoba and utilizes sophisticated GIS technology.

Manitoba Hydro maintains a long term flow database in order to support hydropower planning studies. The database contains monthly average flows for dozens of basins throughout the Nelson River from 1912 to present, many of which are ungauged. Preliminary work suggests that the SLURP model provides very accurate estimates of monthly average flows for the Taylor River and Upper Burntwood River. Based on this work the SLURP model could be a valuable tool to support the long term flow database.

6.3 Recommendations for Further Study

Many hydrologic models such as the SLURP model are very flexible which permits their application to a wide variety of geographic locations and hydrologic and climatic conditions. However, the gain in modeling flexibility results in a loss of modelling accuracy due the lack of some site specific hydrologic processes. As the understanding of hydrologic processes at a site increases and hydrologic modelling experience of a regions is gained, revisions are made to the model to improve its performance. This research establishes the framework so that further revisions can be made to the model if an improvement in model performance is desired.

A number of modeling deficiencies were identified while learning about the study area and through model application. This research revised the model to account for a few hydrologic processes that resulted in improved modelling performance during the snowmelt period. Firstly, the model should be revised to include depression storage, its periodic disconnection with the drainage basin, and its associated evaporation losses. Secondly, the model would benefit by accounting for the recharge of water to lakes, rivers and marshes by headwaters and the associated evaporation. The majority of marshes and beaver flooded land is adjacent to lakes and rivers. A marsh to river linkage and marsh to lake linkage that simulates the exchange of water between those bodies and

the associated evaporation would improve modelling performance and the physical basis of the model. Thirdly, since Muskeg is a dominant land cover in the High Boreal Forest, attempts should be made to improve the understanding of Muskeg hydrology and to better model the unique hydrology of this cover. For example, the exponentially decreasing hydraulic conductivity of Muskeg results in a nonlinear retention of soil water and unique interflow characteristics. Lastly, ET models are very important in water yield estimates since they determine total basin losses. The various ET methods in SLURP should be tested and compared to each other and compared to field measurements such as those collected during the BOREAS field studies to test for accuracy. This exercise may determine that the SLURP model requires alternative ET models for the region of interest.

Model calibration is a time consuming and exhaustive process. The current version of SLURP contains an automatic calibration routine that lacks flexibility, minimizes only one statistic, and requires a lot of computational time. For these reasons, the calibration routine was not used extensively in this research. New and improved automatic calibration routines such as genetic algorithms could be explored to improve the efficiency of the calibration process.

Collection of field data would facilitate the calibration of the current modified version of the SLURP model. Intensive monitoring programs should be established for those basins that will likely require modeling in the future. The monitoring programs should include the following measurements collected over an entire year:

- Depending on the size of the basin, one or more precipitation gauges located within the basin of interest;
- A number of streamflow gauges strategically located throughout the basin including the headwaters;
- Remote sensing or intensive snow depth and density measurements throughout the basin at all dominant types of land covers and an even greater intensity of

measurements during the snowmelt period to characterize snowmelt patterns and rates;

- Ground water depth measurements in Muskeg and Coniferous forests would assist in characterizing soil moisture and aid in developing a better model for the Muskeg cover;
- Water level measurements in marshes and lakes would assist in characterizing changes in storage and evaporation losses;
- Measurements of organic layer thickness, silty clay layer thickness, and depth to bedrock as well as soil porosities and hydraulic conductivities would assist in defining model soil characteristics;

The preceding field program would also greatly assist in developing a better understanding of the basin hydrologic characteristics and in the revision of the SLURP model. Application of the SLURP model to other basins of varying scale under a variety of climatic conditions would aid in developing modelling experience, develop model confidence, and help in revising the model.

Remote sensing technology has undergone tremendous advances in recent years. Work should be undertaken to examine ways of utilizing this data in hydrological modelling in order to refine estimates of variables such as SWE, snow covered area, LAI, surface temperature, and precipitation. This would be especially useful in data sparse remote regions.

The 1984 - 2000 test period used in this research is a relatively dry period. Efforts should be directed to testing the model on other years that are significantly wetter.

Lastly, the impetus for this research came from Manitoba Hydro who expressed an interest in using hydrological models in various hydropower planning and operations. This research demonstrates that a model can be used to simulate the hydrologic behavior in a region of interest to Manitoba Hydro. However, there is a general lack of research that aims to assess and quantify the economic benefit of hydrological modelling in the

day to day, month to month, and year to year operations of hydro power utilities such as Manitoba Hydro. That research in combination with the type of applications based research presented by the author is necessary in order for this technology to gain acceptance by decision makers.

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APPENDIX A**Arc Macro Language (AML) Scripts**

TOPOGRID AML

```
topogrid tdem 100
reset
enforce on
datatype contour
iterations 40
/*chose tolerance based on Arc Help      - 2.5 def, 1=no
      smoothing, 2.5 for 1:50000maps
tolerances 2.5 1 2.5
xyzlimits 500000 6094790 564000 6178690
boundary extent1
point tr-spots elev_m
contour t_contours2 elev_m
stream tay_stre
outputs tsinks tdrainage tdiag
end
```

DSTS AML

```
&setvar tslength = length2
&setvar asas = asas
&setvar lcgrid = lc-100m
&setvar asa
&setvar lc
&setvar meanb
&setvar stdb

&do asa = 1 &to 11 &by 1
  &do lc = 1 &to 7 &by 1
    zonea = con(%asas% == %asa%, con(%lcgrid% == %lc%,
1, 0), 0)
    zoneb = zonea * 0

    zonalmean = zonalmean(zonea, %tslength%)
    zonalstd = zonalstd(zonea, %tslength%)
    zonalmean2 = int(zonalmean + 0.5)
    zonalstd 2 = int(zonalstd + 0.5)

    mean = zonalminority(zoneb, zonalmean2, DATA)
    std = zonalminority(zoneb, zonalstd2, DATA)

    docell
      a }= mean
    end
    &set meanb [show a]
```

```

&type ASA %asa% LC %lc% ts-dist-mean: %meanb%
&type

docell
  b }= std
end
&set stdb [show b]

&type ASA %asa% LC %lc% ts-dist-stdev: %stdb%
&type

kill (!zonea, zoneb, mean, zonalstd, zonalstd2,
zonalmean, zonalmean2, std!)

&sv record = 'ASA '%asa%' LC '%lc%' mean: '%meanb%'
stdev: '%stdb%'
&sv unit2 = [open DSTS.out openstat -append]
&type %openstat%
&type %unit2% and %record%
&sv writestat = [write %unit2% %record%]
&if [close %unit2%] = 0 &then
&type file closed
&end
&end
end

```

DSDS AML

```

&setvar tslength = length2 /* grid containing to-stream
length
&setvar asas = asas /* grid of ASAs, integer value for
each asa from 1 to many
&setvar lcgrid = lc-100m
&setvar asa = [response 'Enter ASA number for this
computation']
&setvar lc
&setvar outlen = [response 'Enter stream length at ASA
outlet']
&setvar meanc
&setvar stdc

/* downstream length for each cell is total streamflow
length
/* subtract streamflow length at the outlet, subtract the
streamflow
/* length to the stream

```

```
dslength = length - %outlen% - length2
  &do lc = 1 &to 7 &by 1
    zone = con(%asa% == %asa%, con(%lcgrid% == %lc%,
1, 0), 0)
    zone2 = zone * 0

    zonalmean = zonalmean(zone, dslength)
    zonalstd = zonalstd(zone, dslength)
    zonalmean2 = int(zonalmean + 0.5)
    zonalstd2 = int(zonalstd + 0.5)

    meanx = zonalminority(zone2, zonalmean2, DATA)
    stdx = zonalminority(zone2, zonalstd2, DATA)

    docell
      a }= meanx
    end
    &set meany [show a]

    &type ASA %asa% LC %lc% ds-dist-mean: %meany%
    &type

    docell
      b }= stdx
    end
    &set stdy [show b]

    &type ASA %asa% LC %lc% ds-dist-stdev: %stdy%
    &type

    kill (!zone, meanx, zonalmean, zonalmean2,
zonalstd, zonalstd2, zone2, stdx, dslength!)

    &sv record = 'ASA '%asa%' LC '%lc%' mean: '%meany%'
stdev: '%stdy%'
    &sv unit2 = [open DSDS.out openstat -append]
    &type %openstat%
    &type %unit2% and %record%
    &sv writestat = [write %unit2% %record%]
    &if [close %unit2%] = 0 &then
    &type file closed

  &end
end
```

Average Slope AML

```
&setvar slope = slope
&setvar asas = asas
&setvar lcgrid = lc-100m
&setvar asa
&setvar slp
&setvar lc
&setvar meank

&do asa = 6 &to 11 &by 1
  &do lc = 1 &to 7 &by 1

    zonei = con(%asas% == %asa%, con(%lcgrid% == %lc%,
1, 0), 0)
    zonej = zonei * 0

    zonalmean = zonalmean(zonei, %slope%) * 10000
    zonalmean2 = int(zonalmean + 0.5)

    meani = zonalminority(zonej, zonalmean2, DATA)

    docell
      a }= meani
    end
    &set slp [show a]

    &type ASA %asa% LC %lc% ts-dist-mean: %slp%
    &type

    kill (!zonei, zonej, meani, zonalmean, zonalmean
2!)

    &sv record = 'ASA '%asa%' LC '%lc%' mean: '%slp%'
    &sv unit2 = [open slope.out openstat -append]
    &type %openstat%
    &type %unit2% and %record%
    &sv writestat = [write %unit2% %record%]
    &if [close %unit2%] = 0 &then
    &type file closed

  &end
&end

end
```

Change In Elevation To ASA Outlet AML File

```
&setvar elev = burnt-dem /* grid containing elevation
&setvar asas = asas /* grid of ASAs, integer value for
each asa from 1 to many
&setvar lcgrid = lc-100m
&setvar asa = [response 'Enter ASA number for this
computation']
&setvar lc
&setvar outelev = [response 'Enter elevation at ASA
outlet']
&setvar meanb
&setvar stdb

chelev = %elev% - %outelev%

  &do lc = 1 &to 7 &by 1
    zone = con(%asas% == %asa%, con(%lcgrid% == %lc%,
1, 0), 0)
    zone2 = zone * 0

    zonalmean = zonalmean(zone, chelev) * 100
    zonalmean2 = int(zonalmean + 0.5)

    mean = zonalminority(zone2, zonalmean2, DATA)

    docell
      a }= mean
    end
    &set meanb [show a]

    &type ASA %asa% LC %lc% mean change in elev:
%meanb%
    &type

    kill (!zone, mean, zonalmean, zonalmean2, zone2!)

    &sv record = 'ASA '%asa%' LC '%lc%' mean change in
elev: '%meanb%'
    &sv unit2 = [open chelev.out openstat -append]
    &type %openstat%
    &type %unit2% and %record%
    &sv writestat = [write %unit2% %record%]
    &if [close %unit2%] = 0 &then
    &type file closed
```

```
    &end  
kill chelev  
  
end
```

APPENDIX B**Modified SLURP FORTRAN Code**

SLURP contains numerous FORTRAN files. The "Slrp_mod.f" contains the majority of the model routines and consists of roughly 2300 lines of code. Therefore, those sections of code that are modified or added are listed. **Bold text** indicates modified or new FORTRAN code.

```

. . . .
. . . .
      real t_1,t_2,t_3,Ts_cum,Wc_cum,fp,cr,t_3d
      integer t_num
. . . .
. . . .
      t_1=0
      t_2=0
      t_3=0
      t_num=0
      ts_cum=0
      wc_cum=0
. . . .
. . . .
*           set cr to the jan snowmelt rate
*           set fp to the jul snowmelt rate variable
*           use jul snowmelt rate so don't have to add
model variable
*           to be used in the computation of the snowpack
cold content
*
      cr=jan_snowmelt(il)
      fp=jul_snowmelt(il)
. . . .
. . . .
*           Compute 3 day average temperature
*
      if(t_num.eq.1) then
          t_3d=t_1
          t_num=t_num+1
      else if (t_num.eq.2) then
          t_3d=(t_1+t_2)/2
          t_num=t_num+1
      else
          t_3d=(t_1+t_2+t_3)/3
      endif
. . . .
. . . .
      if(rad_in_snowmelt.eq.'Y')then
          snowmelt_rate=res_snowmelt(il)
          factor=net_factor(il)/0.0864
      else

```

```

*               call parabola(jan_snowmelt(il),
jul_snowmelt(il),
*       1               day_of_year,snowmelt_rate)
                        factor=0.0
                        snowmelt_rate= jan_snowmelt(il)
endif

. . . .
. . . .
*
      subroutine snow_store(xin,snow_store_content,t,p10,
1snowmelt_rate,factor,netrad,evap_demand,act_evap,
2runoff,t_3d,fp,cr,ts_cum,wc_cum)
*
*   snowpack store subro utine for all land covers except
"water"
*
*   inputs
*       xin           precipitation input to snowpack
*       snow_store_content  current content of snowpack
*       t             mean daily temperature
*       p10           parameter 10 (rain/snow division
temperature)
*       snowmelt_rate  daily snowmelt rate,
mm/degree/day
*       factor         factor to convert net radiation to
snowmelt
*       netrad         net radiation, MJ/m**2/day
*       evap_demand    maximum possible evaporation
(sublimation)
*       t_3d           3 day average temperature (Celcius)
*
*   outputs
*       act_evap       actual evaporation (sublimation) from
snowpack
*       runoff         actual snowmelt
*
      implicit none
      real xin, snow_store_content, t, temval, p10,
snowmelt_rate
      real evap_demand, act_evap, runoff, factor, netrad
      real t_3d,fp,cr,ts_cum,wc_cum
*
*   increment storage
*
      snow_store_content=snow_store_content+xin
*       xin=0.0
*

```

```
*      if no water, exit
*
*      if (snow_store_content.le.0.0) go to 10
*
*      Determine Cumulative temperature of snowpack.
*
*      if (snow_store_content.eq.0.0) then
*         ts_cum=0
*      else
*         ts_cum=ts_cum+fp*(t_3d*(-1)-ts_cum)
*      end if
*
*
*      Determine the cold content of the snowpack.
*
*      if (snow_store_content.eq.0.0) then
*         wc_cum=0
*      else
*         wc_cum=wc_cum+cr*(t_3d*(-1)-ts_cum)
*      end if
*
*
*      take out runoff (snowmelt) only if cold content,
wc_cum, is less than zero.
*
*      runoff=snowmelt_rate*(t-p10)
*      if(netrad.gt.0.0.and.factor.gt.0.0) runoff=runoff+
factor*netrad
*      if(runoff.lt.0.0) runoff=0.0
*      if (wc_cum.gt.0) runoff=0.0
*      temval=snow_store_content-runoff
*      if (temval.le.0.0) then
*         runoff=snow_store_content
*         go to 20
*      endif
*      snow_store_content=snow_store_content-runoff
*
*      set evaporation equal to maximum possible evaporation
*      and remove as much as possible
*
*      act_evap=evap_demand
*      temval=snow_store_content-act_evap
*      if (temval.le.0.0) then
*         act_evap=snow_store_content
*         go to 30
*      endif
*      snow_store_content=snow_store_content-act_evap
```

```

        go to 40
10      runoff=0.0
20      act_evap=0.0
30      snow_store_content=0.0
40      return
        end

        subroutine lake_store(xin,lake_snow_content,t,p10,
1snowmelt_rate,factor,netrad,evap_demand,snowmelt,
act_evap,
2runoff,t_3d,fp,cr,ts_cum,wc_cum)
*
*      This store is used only for a water "land cover"
*      The store simulates snowpack on a lake as well
*      as the water column of the lake.
*      This store sets act_evap to evap_demand and allows -
ve runoff
*
*      inputs
*      xin          precipitation input to lake/snowpack
*      lake_snow_content  current content of lake
snowpack
*      t            mean daily temperature
*      p10          parameter 10 (rain/snow division
temperature)
*      snowmelt_rate  daily snowmelt rate,
mm/degree/day
*      factor        factor to convert net radiation to
snowmelt
*      netrad        net radiation, MJ/m**2/day
*      evap_demand   maximum possible evaporation
(sublimation)
*      t_3d          3 day average temperature (Celcius)
*      outputs
*      act_evap      actual evaporation (sublimation) from
lake/snowpack
*      snowmelt      actual snowmelt
*      runoff        actual runoff
*
        implicit none
        real xin,lake_snow_content,t,temval,p10,snowmelt_rate
        real
evap_demand,act_evap,runoff,factor,netrad,snowmelt
        real t_3d,fp,cr,ts_cum,wc_cum
*

```

```
*      increment storage
*
      lake_snow_content=lake_snow_content+xin
      runoff=0.0
      act_evap=0.0
      snowmelt=0.0
*
*      if no snow, satisfy all evaporation from lake water,
*      set runoff to -ve and exit
*
      if (lake_snow_content.le.0.0) then
        act_evap=evap_demand
        runoff=lake_snow_content-act_evap
        lake_snow_content=0.0
      else
*
*      Determine Cumulative temperature of snowpack.
*
        if (Sno_depth.eq.0) then
          ts_cum=0
        else
          ts_cum=ts_cum+fp*(t_3d*(-1)-ts_cum)
        end if
*
*      Determine the cold content of the snowpack.
*
        if (Sno_depth.eq.0) then
          wc_cum=0
        else
          wc_cum=wc_cum+cr*(t_3d*(-1)-ts_cum)
        end if
*
*
*      if we have snow, melt it and set evaporation equal
to melt
*      and runoff as any remainder
*
      snowmelt=snowmelt_rate*(t-p10)
      if(netrad.gt.0.0.and.factor.gt.0.0)snowmelt=
snowmelt+
      1      factor*netrad
      if(snowmelt.lt.0.0) snowmelt=0.0
      if (wc_cum.gt.0) snowmelt=0.0
      temval=lake_snow_content-snowmelt
      if (temval.le.0.0) then
        snowmelt=lake_snow_content
```

```
        lake_snow_content=0.0
    else
        lake_snow_content=temval
    endif
    temval=snowmelt-evap_demand
    if (temval.le.0.0) then
        act_evap=snowmelt
        runoff=0.0
    else
        act_evap=evap_demand
        runoff=temval
    endif
endif
return
end
```

*