

Spatiotemporal Patterns & Climate Change Impacts to Groundwater
Recharge & Hydrology in a Low-Mountain Area, Nova Scotia, Canada.

by

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Dalhousie University is located in Mi'kma'ki, the
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TABLE OF CONTENTS

List of Tables.....	v
List of Figures	vii
Abstract	xi
List of Abbreviations and Symbols Used.....	xii
Acknowledgments.....	xvi
Chapter 1 - Introduction.....	1
1.1 Importance of Groundwater Resources	1
1.2 Groundwater in Cold Regions	1
1.2.1 Seasonality of Cold Region Groundwater Processes.....	2
1.2.2 Montane, Geologic, & Landcover Influences on Cold Region Groundwater Resources	5
1.3 Climate Change Threats to Cold Region Groundwater Resources.....	10
1.4 Climate Change Impacts to Groundwater Resources & Processes.....	12
1.4.1 Annual Recharge.....	12
1.4.2 Seasonal Recharge	13
1.4.3 Groundwater Exfiltration.....	15
1.5 Climate Change Impacts to Streamflow	17
1.6 Knowledge Gaps.....	19
1.7 Research Objectives.....	21
Chapter 2 - Methods.....	23
2.1 Study Site.....	23
2.1.1 Topography & Landcover	24
2.1.2 Historical Climate	25
2.1.3 Geology & Soil	26
2.1.4 Hydrology	27
2.2 Field Data Collection	28
2.2.1 Hydrological Monitoring Network	28
2.2.2 Meteorological Data and Snow Monitoring	29
2.2.3 Soil Moisture & Temperature	29

2.3	Historical Snow Melt-Out Timing Trend Analysis	31
2.4	Recharge modelling	31
2.4.1	Model Selection	32
2.4.2	Model Overview	33
2.4.3	Key Model Equations and Numerical Methods.....	33
2.4.4	Model Domain and Parameterization	36
2.4.5	Initial & Boundary Conditions.....	41
2.4.6	Model Calibration & Validation.....	42
2.5	Historical & Future Recharge Analysis	46
2.5.1	Historical & Future Recharge Simulations	46
2.5.2	Recharge Quantification	47
2.5.3	Hourly Graphical Recharge Analysis.....	47
2.5.4	Annual & Seasonal Recharge Analysis.....	48
2.6	Sensitivity & Uncertainty Analysis.....	49
2.6.1	Sensitivity Analysis.....	49
2.6.2	Uncertainty Analysis.....	50
Chapter 3 - Results.....		51
3.1	Historical Snow Melt-Out Timing Trend Analysis	51
3.2	Observational Data.....	52
3.3	SHAW Model Calibration & Validation	60
3.2.1	Snow Depth & Density Simulation Performance	61
3.3.1	Soil Temperature Simulation Performance	63
3.3.2	Soil Liquid Water Content Simulation Performance	66
3.4	Hourly Graphical & Seasonal Recharge Processes Analysis.....	69
3.4.1	Historical Snow Cover and Soil Frost	69
3.4.2	Historical Soil Volumetric Moisture Content.....	73
3.4.3	Historical Recharge and Recharge Efficiency	74
3.4.4	Snow Cover and Soil Frost under Climate Change	77
3.4.5	Projected Recharge and Recharge Efficiency under Climate Change	81
3.5	Historical and Future Trends of Annual Recharge and Recharge Efficiency ...	83
3.5.1	Historical Annual Recharge and Recharge Efficiency	83

3.5.2	Projected Annual Recharge and Recharge Efficiency under Climate Change.....	84
3.6	Uncertainty & Sensitivity Analysis Results.....	84
3.6.1	Sensitivity Analysis.....	84
3.6.2	Uncertainty Analysis.....	85
Chapter 4 - Discussion		89
4.1	Historical Snowmelt Runoff Timing Trends.....	89
4.2	Topographic and Landcover Effects on Historical Recharge Rate and Efficiency	90
4.2.1	Snowpack Thickness & Snow Cover Duration	92
4.2.2	Soil Frost Presence.....	92
4.2.3	VMC & Recharge	93
4.3	Climate Change Impacts on Recharge Rate & Efficiency	95
4.3.1	Evapotranspiration	97
4.3.2	Snow Depth & Duration	97
4.3.3	Frost Depth & Duration	99
4.3.4	Recharge	101
4.4	Limitations, Recommendations, & Future Work	103
4.4.1	Model Calibration & Validation Limitations	103
4.4.2	Recharge Analysis Limitations	108
4.4.3	Recommendations & Future Work	109
Chapter 5 - Conclusion		113
Bibliography		115
6. Appendix A – Meteorological Data Cleaning		137
Temperature, Solar Radiation, Humidity		137
Windspeed.....		138
7. Appendix B – Supplementary Figures, Tables & Equations.....		140

LIST OF TABLES

Table 1: Calibrated Parameters & Values (Colour denotes calibration type, blue is snow, yellow is soil, green is both; Soil values presented as a range).	39
Table 2: CT Linear Regression Results for Cape Breton Streamflow Stations (Rows highlighted green are statistically significant)	52
Table 3: Calibration & Validation Performance for Snow Depth, Soil Temperature, & Soil Liquid Water Content.	62
Table 4: Uncertainty Analysis Results for Median Values in the Historical Period.	87
Table 5: Uncertainty Analysis Results for Median Values in the SSP2-4.5 2055-2064 Period.	88
Table 6: Temperature and humidity statistics at the Wetland and the ECCC Eskasoni, Louisbourg and Port Hawkesbury weather stations from 2015 to 2025.	137
Table 7: Solar radiation statistics at the Wetland site and from NASA POWER 2015 to 2025.	138
Table 8: Groundwater Monitoring Well Specifications.	140
Table 9: Non-Calibrated Model Parameter Values	141
Table 10: Low-Elevation Forest Model Depth Specific Soil Layer Parameters	142
Table 11: High-Elevation Forest Model Depth Specific Soil Layer Parameters	143
Table 12: Wetland Model Depth Specific Soil Layer Parameters	144
Table 13: Soil Moisture & Temperature Observations Used in Model Calibration & Validation.	145
Table 14: Groundwater Level Observations Used in Model Calibration & Validation. .	145
Table 15: Snow Depth Observations for Model Calibration & Validation.	146
Table 16: PEST Adjustable Node Depths for Each SHAW Model	146
Table 17: Wetland Model Calibration Objective Function Weights	147
Table 18: High-Elevation Forest Model Calibration Objective Function Weights	148
Table 19: Low-Elevation Forest Model Calibration Objective Function Weights	149

Table 20: Wetland Model Calibrated Parameters.....	156
Table 21: Acceptable Error Metric Ranges for SHAW models.	157
Table 22: High-Elevation Forest Model Calibrated Parameters.	158
Table 23: Low-Elevation Forest Model Calibrated Parameters.....	159
Table 24: Wetland Site Calibration & Validation Soil LWC & Temperature Performance.	160
Table 25: High-Elevation Forest Model Soil LWC & Temperature Calibration & Validation Performance.....	161
Table 26: Low-Elevation Forest Model Soil LWC & Temperature Calibration & Validation Performance.....	162
Table 27: Coefficient of Determination of Simulated Snow Density & the Sampling Event Median Observed Snow Density for each Site.	162
Table 28: Wetland Model Historical Sensitivity Analysis 10 Most Sensitive Parameter Changes.....	163
Table 29: Wetland Model SSP2-4.5 2055 – 2064 Sensitivity Analysis 10 Most Sensitive Parameter Changes.....	164
Table 30: High-Elevation Forest Model Historical Sensitivity Analysis 10 Most Sensitive Parameter Changes.....	165
Table 31: High-Elevation Forest Model SSP2-4.5 2055 – 2064 Sensitivity Analysis 10 Most Sensitive Parameter Changes.....	166
Table 32: Low-Elevation Forest Model Historical Sensitivity Analysis 10 Most Sensitive Parameter Changes.....	167
Table 33: Low-Elevation Forest Model SSP2-4.5 2055 – 2064 Sensitivity Analysis 10 Most Sensitive Parameter Changes.....	168

LIST OF FIGURES

Figure 1: Conceptual Diagram of Cold Region Groundwater Recharge (Cook & Brunner, 2025).	2
Figure 2: Conceptual Diagram of (A) Midwinter Melt Event During Low Moisture Content and Frozen Soil Resulting in Water Infiltrating and Recharging, (B) Subsequent Freezeback Event Resulting in Highly Saturated Frozen Soil, (C) Subsequent Melt Event Resulting in Overland Flow due to Low Permeability Frozen Soil Layer (Hyman-Rabeler & Loheide II, 2023).	4
Figure 3: Conceptual Diagram of (A) Physical Configuration of the Basin Floor, Mountain Front, and Mountain Block, (B) Mountain Front Recharge Dominated System, (C) Mountain Block Recharge Dominated System, and (D) Mountain Front and Mountain Block Recharge System (modified from Bresciani et al., 2018).	7
Figure 4: Conceptual Diagram of Snow Distribution Due to Wind (Arrows Indicate the Prevailing Wind Direction) and Resulting Frost Distribution in a Field Between Forests (Lundberg et al., 2016).	9
Figure 5: Conceptual Diagram of the Hydrological Buffering and Storage Capacity of Wetlands (Turner & Franklin, 2005).	10
Figure 6: Conceptual Diagram of Groundwater-Surface Water Interactions throughout the Year: (A) Early Winter, (B) Spring, (C) Summer, (D) Fall (modified from Huntington & Niswonger, 2012).	16
Figure 7: Conceptual Diagram of Climate Change Impacts to Groundwater Recharge...	20
Figure 8: (A) Cape Breton Island Stream Gauging Stations & Christmas Brook Watershed Location, (B) Christmas Brook Watershed Landcover Classification & Monitoring Locations, (C) Christmas Brook Watershed Hillslope & Sub-site Locations, (D) Christmas Brook Surficial Geology & Fault Aquifer Position.....	24
Figure 9: Conceptual Diagram of the Christmas Brook Watershed Regional Hydrochemical Cross Section (modified from EXP Services Ltd., 2010).	26
Figure 10: SHAW Model Soil Layer Domains	37
Figure 11: Significant CT Date Trends of Cape Breton Island Streams.	51
Figure 12: Wetland Observed Snow Depth at Weather Station and from Snow Sampling Campaigns Compared to SHAW Simulated Snow Depth for Calibration (2023 - 2025) and Validation (2026) Periods (Year Label is the Calendar Year, e.g., 2023 is the 2022/2023 Winter Season).	54

Figure 13: High-Elevation Forest Observed Snow Depth at SDS-Forest and SDS-Wetland and from Snow Sampling Campaigns Compared to SHAW Simulated Snow Depth for Calibration (2023 - 2025) and Validation (2026) Periods (Year Label is the Calendar Year, e.g., 2023 is the 2022/2023 Winter Season).....	55
Figure 14: Low-Elevation Forest Observed Snow Depth at SDS-Forest and SDS-Wetland and from Snow Sampling Campaigns Compared to SHAW Simulated Snow Depth for Calibration (2015 - 2024) and Validation (2025-2026) Periods (Year Label is the Calendar Year, e.g., 2023 is the 2022/2023 Winter Season).	56
Figure 15: Observed Soil Volumetric Liquid Water Content at the (A) Wetland, (B) High-Elevation Forest, (C) and Low-Elevation Forest and Observed Soil Temperature at the (D) Wetland, (E) High-Elevation Forest, (F) Low-Elevation Forest.	59
Figure 16: Measured Snow Density from Snow Sampling Campaigns and Simulated Snow Density (The Low-Elevation Forest Only has Two Sampling Events as Snow Depth was too Low for Two of the Four Other Sampling Campaigns).....	63
Figure 17: Wetland Model Soil Temperature Calibration Performance.	64
Figure 18: High-Elevation Forest Model Soil Temperature Calibration Performance.	65
Figure 19: Low-Elevation Forest Model Soil Temperature Calibration Performance.....	65
Figure 20: Wetland Model Soil LWC Calibration Performance.	67
Figure 21: High-Elevation Forest Model Soil LWC Calibration Performance.	68
Figure 22: Low-Elevation Forest Model Soil LWC Calibration Performance.	68
Figure 23: Wetland Site Hourly (A) Precipitation, (B) Temperature, (C) Snow Depth, (D) Frost Depth, (E) Runoff, (F) VMC ₀ , (G) VMC _{NOI} , and (H) Recharge (line) and Cumulative Recharge (line and area) in Historical and SSP2-4.5 2055 to 2064 Scenarios for 2016 and 2022 Base Years.....	70
Figure 24: High-Elevation Forest Site Hourly (A) Precipitation, (B) Temperature, (C) Snow Depth, (D) Frost Depth, (E) Runoff, (F) VMC ₀ , (G) VMC _{NOI} , and (H) Recharge (line) and Cumulative Recharge (line and area) in Historical and SSP2-4.5 2055 to 2064 Scenarios for 2016 and 2022 Base Years.	71
Figure 25: Low-Elevation Forest Site Hourly (A) Precipitation, (B) Temperature, (C) Snow Depth, (D) Frost Depth, (E) Runoff, (F) VMC ₀ , (G) VMC _{NOI} , and (H) Recharge (line) and Cumulative Recharge (line and area) in Historical and SSP2-4.5 2055 to 2064 Scenarios for 2016 and 2022 Base Years.	72

Figure 26: Wetland, High-Elevation, and Low-Elevation Annual Average Snow Depth (A), Days with Snow Cover (B), Annual Average Frost Depth (C), and Days with Frozen ground (D) for all Climate Scenarios.....	73
Figure 27: Wetland, High-Elevation, and Low-Elevation Seasonal Precipitation (A), Evapotranspiration (B), Runoff (C), Recharge (D), and Recharge Efficiency (E) for all Climate Scenarios.	76
Figure 28: (A) Historical and (B) SSP2-4.5 2055 to 2064 Cumulative Recharge for each year in the Wetland, High-Elevation Forest, and Low-Elevation Forest sites.....	77
Figure 29: Hourly Frost Depth for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites for the (A) Historical Scenario and (B) SSP2-4.5 2055-2064 Scenario, and (C) the Frost Depth Difference between Historical & SSP2-4.5 2055-2064 Scenarios (Positive = Increased from Historical, Negative = Decreased from Historical).	80
Figure 30: Wetland, High-Elevation, and Low-Elevation Annual Precipitation (A), Recharge (B), and Recharge Efficiency (C) for all Climate Scenarios.....	83
Figure 31: Wetland Site Observations of (A) PNC Groundwater Levels and Precipitation, (B) Soil LWC from SMT 1 and SMT 2 (LWC observations above 0.657 were removed as they exceeded the instrument measurement limits).	150
Figure 32: Wetland Site Observations of (A) Air Temperature, (B) SMT 1 Soil Temperature, and (C) SMT 2 Soil Temperature.....	151
Figure 33: High-Elevation Forest Site Observations of (A) MW1A Groundwater Levels and Precipitation, (B) Soil LWC from SMT 3 and SMT 4.	152
Figure 34: High-Elevation Forest Site Observations of (A) Air Temperature, (B) SMT 3 Soil Temperature, and (C) SMT 4 Soil Temperature.	153
Figure 35: Low-Elevation Forest Site Observations of (A) MW4 Groundwater Levels and Precipitation, (B) Soil LWC from SMT 5 and SMT 6 (MW4 has pumping well influences; MW4 GWL observations were recorded for the entire period but unavailable at the time of plotting).....	154
Figure 36: Low-Elevation Forest Site Observations of (A) Air Temperature, (B) SMT 5 Soil Temperature, and (C) SMT 6 Soil Temperature.	155
Figure 37: : Wetland Uncertainty Analysis Parameter Distributions	169
Figure 38: High-Elevation Forest Uncertainty Analysis Parameter Distributions.....	169
Figure 39: Low-Elevation Forest Uncertainty Analysis Parameter Distributions.	170

Figure 40: : Histogram of the Wetland Historical Model 100 Monte Carlo Simulations.	170
Figure 41: Histogram of the Wetland SSP2-4.5 2055-2064 Model 100 Monte Carlo Simulations.	171
Figure 42: Histograms of the High-Elevation Forest Historical Model 100 Monte Carlo Simulations.	171
Figure 43: Histograms of the High-Elevation Forest SSP2-4.5 2055-2064 Model 100 Monte Carlo Simulations.	172
Figure 44: Histograms of the Low-Elevation Forest Historical Model 100 Monte Carlo Simulations.	172
Figure 45: Histograms of the Low-Elevation Forest SSP2-4.5 2055-2064 Model 100 Monte Carlo Simulations.	173

ABSTRACT

In cold regions, snowmelt is a prominent source of groundwater recharge and spring streamflow. As the climate warms, the factors governing snowmelt recharge dynamics are expected to shift. Thus, the amount and timing of groundwater recharge may be altered through changes in temperature and soil ice content, and the predicted shift from less snow towards more winter rain. The study was conducted in the low-mountain area of Christmas Brook watershed located in Eskasoni, Nova Scotia. The objectives were: (1) determine if historical snowmelt runoff timing in streams throughout Cape Breton Island are shifting earlier in the year, (2) quantify differences in the historical recharge amount and efficiency across various topographic and landcover settings, and (3) project future changes in annual and seasonal groundwater recharge and recharge efficiency and compare the response between different topographic and landcover settings. Through these objectives, we evaluated whether increased winter rain can offset snowmelt losses in a warming climate.

We monitored precipitation, snow depth, groundwater level, soil moisture and temperature, and streamflow across three landscape types at differing topographic positions. We applied the Simultaneous Heat and Water (SHAW) model to the three sites over historical, mid-century, and end-of century decadal periods to quantify differences in seasonal and annual recharge and recharge efficiency. We also evaluated whether enhanced rainfall in a warming climate can offset recharge losses due to diminished snowmelt.

We found the High-Elevation Forest site had the highest recharge (972 mm/year) under the current climate, followed by the Low-Elevation Forest site (500 mm/year), and lastly the Wetland site (407 mm/year). We found recharge in the snow-free season under climate change remained stable as the enhanced precipitation compensated for increases in evapotranspiration. However, in the snow season, recharge became more sensitive to the timing and amount of precipitation events due to altered frost presence and snowmelt. Under climate change, snow depth significantly decreased across all sites, but the change in frost depth and duration was site and scenario specific and indicated that there is a warming threshold at which the loss in snow depth insulation is compensated by warmer air temperatures. At two of our three sites, we found that snow season recharge remained stable or slightly increased (-0.2% to 6.8%) under climate change due to increased precipitation, but the recharge efficiency was reduced between -0.02% to -2.8%. Despite the stability in recharge, these hydrological changes can have important implications for water resources management (e.g., shifting seasonal water availability), and contrasts some projections for other cold region climates.

LIST OF ABBREVIATIONS AND SYMBOLS USED

Abbreviation	Description
% UC	Percent uncertainty
AEP	Air entry potential (SHAW parameter)
B	Campbell's pore size distribution parameter (SHAW parameter)
BIOMASS	Dry biomass of the canopy
BLBR	Bras d'Or Lake Biosphere Region
BLBRA	Bras d'Or Lake Biosphere Region Association
c	Solute concentration in soil
c_i	Specific heat of ice
c_l	Specific heat capacity of water
CMIP6	Coupled model intercomparison project phase 6
CT	Center-of-Volume
C_s	Volumetric heat capacity of the soil
DCHAR	Plant characteristic leaf dimension (SHAW parameter)
ECCC	Environment & Climate Change Canada
EFWC	Eskasoni Fish & Wildlife Commission
ENSO	El Nino and the Southern Oscillation
ET	Evapotranspiration
g	Gravitational acceleration
GWL	Groundwater level
GW-SW	Groundwater-surface water
h₁	50 m height
h₂	WS-Wetland measurement height
h_d	Equal to 0.65 times the canopy height
HGS	HydroGeoSphere model
h_{z0}	Roughness length
IPCA	Indigenous protected & conserved area
IPCC	Intergovernmental Panel on Climate Change
k	Von Karman constant
K	Unsaturated hydraulic conductivity
k_s	Soil thermal conductivity
K_s	Saturated hydraulic conductivity
k_{sp}	Thermal conductivity of snow
KZ	Vertical hydraulic conductivity (SHAW parameter)
LAI	Leaf area index (SHAW parameter)
LEAF_OR	Leaf orientation (SHAW parameter)

L_f	Latent heat of fusion
L_{sp}	Latent heat of sublimation
L_v	Latent heat of vaporization
LWC	Volumetric liquid water content
masl	Meters above sea level
mbgs	Meters below ground surface
MBR	Mountain block recharge
MFR	Mountain front recharge
MW	Monitoring well
MW1A	Monitoring well 1a
MX_PND	Maximum ponding depth (SHAW parameter)
NASA	National Aeronautics and Space Administration
NOAA	National Oceanic and Atmospheric Administration
NOI	Recharge evaluation node depth of interest
NSE	Nash-Sutcliffe efficiency
O_{new}	New model output value from sensitivity analysis
O_{original}	Original output value from sensitivity analysis
O_{0.05}	The 5th percentile of the analytical output value
O_{0.95}	The 95th percentile of the analytical output value
P_{ALB}	Canopy albedo (SHAW parameter)
P_{CLUMP}	Plant clumping parameter (SHAW parameter)
P_{HEIGHT}	Canopy height (SHAW Parameter)
P_{INT}	Canopy interception (SHAW parameter)
P_{new}	New model parameter for sensitivity analysis
P_{original}	Original model parameter value for sensitivity index equation
PBIAS	Percent bias
PDO	Pacific Decadal Oscillation
PEST	Model-Independent Parameter Estimation and Uncertainty Analysis software
p_l	Specific heat of liquid water
PNA	Southeast piezometer located in the peatland
PNC	Northwest piezometer located in the peatland
PORO	Saturated water content (SHAW parameter)
POWER	Langley Research Center's Prediction of Worldwide Energy Resources
P_x	Total snow and rain precipitation for period x.
q_i	Streamflow at the time, i
q_l	Liquid water flux

q_v	Downward water vapor flux
R	Universal gas constant
R²	Coefficient of determination
RE_x	Recharge Efficiency for period x
RMSE	Root-mean squared error
R_n	Net radiation
ROOT_D	Rooting depth (SHAW parameter)
R_x	Recharge for period x, defined as vertical water flux below the evapotranspiration extinction depth
S_SR	Soil surface roughness (SHAW parameter)
SDS-ECCC	ECCC snow depth sensor
SDS-Forest	Forest snow depth sensor
SDS-Wetland	Wetland snow depth sensor
SFS	Snow-free season
SGS-Lower	Stream gauging station located in the lowlands.
SGS-Upper	Stream gauging station located in the high-elevation plateau
SHAW	Simultaneous Heat & Water (model)
SI	Sensitivity index
SMT	Soil moisture and temperature
SNW_TEMP	Snow threshold temperature (SHAW parameter)
SS	Snow season
SSP	Shared socio-economic pathway
SUM_LAI	Summer leaf area index (SHAW parameter)
SUTRA-ICE	Saturated and Unsaturated Transport Model
SWE	Snow-water equivalent
SWRC	Soil water retention curve
T	Snowpack temperature
T_c	Soil temperature in Celsius
t_i	Timestep for CT method
T_k	Soil temperature in kelvin
U	Source sink term for water flux (i.e., root extraction)
U(h)	Windspeed at height, h
U.S.	United States of America
UC	Uncertainty
USDA	U.S. Department of Agriculture
VMC	Volumetric moisture content, VMC _{depth} is the VMC at some depth, either NOI or 0 (soil surface)
WIN_LAI	Winter leaf area index (SHAW parameter)
WS-Louis	ECCC station in Louisbourg, station ID = 54578

WS-ECCC	ECCC Eskasoni Meteorological Station (station ID: 49748)
w_{sp}	Volumetric liquid water content of snow
WS-PHawk	ECCC station in Port Hawkesbury, station id = 48668
WS-Wetland	EFWC Meteorological Station located in the highland plateau wetland.
x	Period of interest, can either be annual (A) or seasonal (SS/SFS)
z₀	Surface roughness
z_R	Windspeed reference height
α	Empirical value which is commonly set to 0.2
θ_i	Volumetric ice content of the soil
θ_l	Volumetric liquid water content of soil
θ_s	Saturated water content
κ	Dielectric constant
μ^*	Friction velocity
μ_1	Windspeed at 50 m
μ_2	Windspeed at the WS-Wetland measurement height
ρ_i	Density of ice
ρ_{sp}	Density of the ice fraction in the snowpack
ρ_v	Vapor density within the layer
Ψ	Soil matric potential
Ψ_e	Air entry potential

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CHAPTER 1 - INTRODUCTION

1.1 IMPORTANCE OF GROUNDWATER RESOURCES

Sustainable groundwater development and use is an emerging concern for many communities. Groundwater is a critical source of drinking water; around 30% of Canadians, and up to 50% of residents in the Canadian Maritimes, are dependent on groundwater for drinking water (Statistics Canada, 2016). Additionally, groundwater is an important source of streamflow and an essential resource for many aquatic ecosystems. In groundwater-dependent streams, groundwater discharge, also known as exfiltration, is vital for maintaining streamflow during periods of low flow and stream temperature and nutrient levels (Dierauer et al., 2018; Dugdale et al., 2013). Cold-water fish are especially reliant on cold groundwater inputs for thermal refuge in warm summer months (Dugdale et al., 2013; Smith et al., 2023).

Changes in groundwater withdrawal, recharge, or discharge dynamics can have serious adverse socioeconomic and environmental impacts now and into the future (Konikow & Bredehoeft, 2020). Two major threats to groundwater resources include overextraction and climate change. Overextraction can cause reduced aquifer storage and depleted surface water resources among other environmental consequences (Konikow & Bredehoeft, 2020). Climate change can alter recharge and discharge processes, thereby altering the total groundwater storage and seasonal water availability (Jasechko et al., 2014). Therefore, to develop sustainable groundwater management practices, communities must first understand the current state of their groundwater resources and the potential threats to these resources from climate change.

1.2 GROUNDWATER IN COLD REGIONS

Groundwater resources and processes are primarily shaped by the local climate, topography, landcover, and geology (Cook & Brunner, 2025). However, in cold regions, hydrological processes are also significantly governed by the seasonal or permanent presence of snow, ice, and frozen ground (Springer Nature, n.d.; Figure 1). Snow, ice, and frozen ground presence can also be influenced by these local drivers (i.e., local climate, topography, landcover, and geology), thus, to develop effective groundwater management

programs in cold regions with varied geology and landcover, detailed knowledge of these local effects is required. Mountains are often characterized as cold regions as they typically contain large stores of water as snow and ice. Mountains have unique hydrological processes due to their high relief, complex near surface stratigraphy, and high water table gradients (Somers & McKenzie, 2020). Additionally, groundwater and surface water systems can often be closely tied (Winter et al., 1998). For these reasons, the hydrology of mountain environments and of groundwater-surface water (GW-SW) interactions can provide critical context in understanding cold region groundwater processes.

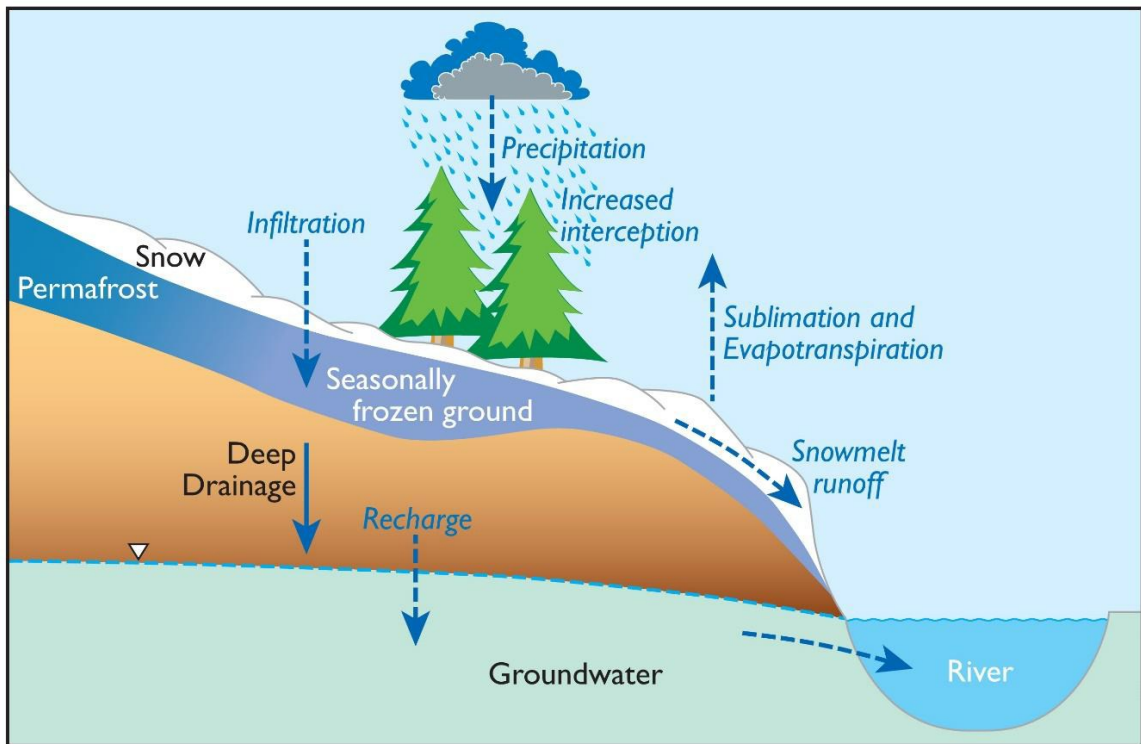


Figure 1: Conceptual diagram of cold region groundwater recharge (Cook & Brunner, 2025).

1.2.1 Seasonality of Cold Region Groundwater Processes

In cold regions, snow generally contributes disproportionately to total annual recharge in comparison to its proportion of total annual precipitation (Jasechko et al., 2014; Earman et al., 2006; Mohammed et al., 2019). The importance of snowmelt in cold region hydrology means that groundwater processes in these watersheds exhibit distinct

seasonality which are strongly influenced by the seasonal accumulation and melt of snow, and soil freezing and thawing dynamics (Mohammed et al., 2019). For this reason, GW-SW interactions in cold regions also follow seasonal patterns.

During winter, frozen ground and snowpack are often present in cold regions, which inhibits infiltration and thus recharge (Jasechko et al., 2014; Boumaiza et al., 2022). As a result, winter typically has the lowest rates of recharge in cold regions, despite it being a period with low evapotranspiration rates (Huntington & Niswonger, 2012; Scibek & Allen, 2006). As temperatures start to warm in the spring, the snow melts and the ground thaws, promoting recharge (Cook & Brunner, 2025). During the snowmelt period, recharge is often limited by the soil infiltration capacity as water availability is high (Cook & Brunner, 2025). Infiltration capacity of a soil is controlled by numerous factors including soil texture, density, moisture content, and ice content (Hyman-Rabeler & Loheide II, 2023).

The interactions between snow depth, snowmelt timing, melt rate, and soil ice content can result in complex recharge processes in cold regions. For instance, snowpack can act as an insulator to the soil surface, thus the snowpack depth impacts the depth of frost and ice content in the soil (Lundberg et al., 2016). Furthermore, the timing of snowmelt versus ground thaw influences recharge. For example, Mohammed et al. (2019) found at grassland sites in the Canadian prairies, snowmelt that occurred when the ground was still frozen exceeded the soil infiltration capacity and encouraged ponding in surface depressions and increased focused recharge. Likewise, midwinter melt events and subsequent freezeback events can result in low permeability ice layers which limit the infiltration and recharge of future melt events (Hyman-Rabeler & Loheide II, 2023; Figure 2). Therefore, controls on recharge in cold regions can be complex as they can include snowmelt processes, soil freeze-thaw dynamics, and the interaction between the two (Hyman-Rabeler & Loheide II, 2023).

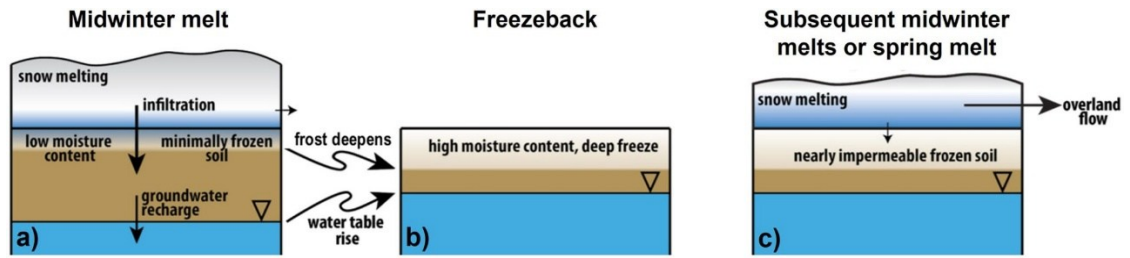


Figure 2: Conceptual diagram of (A) midwinter melt event during low moisture content and frozen soil resulting in water infiltrating and recharging, (B) subsequent freezeback event resulting in highly saturated frozen soil, (C) subsequent melt event resulting in overland flow due to low permeability frozen soil layer (Hyman-Rabaler & Loheide II, 2023).

Higher temperatures and vegetation growth during summer results in higher evapotranspiration (ET) rates that reduce the amount of water available for recharge (Jasechko et al., 2014). The antecedent moisture conditions of the soil can play a significant role in the ability for precipitation to infiltrate and become recharge. For example, dry soil has adequate capacity to infiltrate water but has lower unsaturated hydraulic conductivity and results in water being held in the upper layers of the soil profile due to capillary action rather than continuing downward as recharge (Freeze & Cherry, 1979). In contrast, more saturated soils can limit infiltration due to their lower infiltration capacity and promote runoff but typically these conditions also promote recharge generation as the soils reach field capacity (Cook & Brunner, 2025; Freeze & Cherry, 1979). Thus, historically in the summer, infiltration capacity is driven by antecedent moisture conditions which, along with the precipitation intensity and ET rate, drives the infiltration rate (Ren et al., 2025).

In cold regions, the ratio of recharge to precipitation inputs, also known as recharge efficiency (RE), is thought to be higher for snow precipitation than rain (Earman et al., 2006; Jasechko et al., 2014). For instance, at four sites in the southwestern United States, Earman et al. (2006) found snow, which made up 25-50% of total annual precipitation, accounted for 40-70% of the total annual groundwater recharge. Likewise, in Nova Scotia, Canada, recharge in the winter and early spring accounted for 40% of annual recharge but only 32% of the annual precipitation (Chai et al., 2026). Historically, the highest rates of recharge in cold regions are typically in the early spring, when ET

remains relatively low, the ground is thawing, and snowpack is melting (Jasechko et al., 2014; Sulis et al., 2011). Whereas rainfall in cold regions can occur year round, but often it occurs in the warmer months during times with higher ET rates. Furthermore, winter rainfall may not be able to infiltrate due to snow cover and frozen ground. However, as highlighted in the review by van Tiel et al. (2024), it is unclear whether the historical RE differences between snow and rain are truly due to the phase of the precipitation or due to seasonal conditions, including the lower ET and higher soil moisture.

Groundwater exfiltration in cold regions is also closely linked to snowmelt processes and soil freeze-thaw dynamics as they influence the seasonal GW-SW interactions. Spring historically tends to be a period of peak streamflow in cold regions due to the rapid runoff and interflow from rapid snowmelt. This high flow in turn influences GW-SW interactions (Jasechko et al., 2014; Huntington & Niswonger, 2012). During spring freshet, the stream has a high hydraulic head which suppresses groundwater exfiltration into the stream (Huntington & Niswonger, 2012). Exfiltration occurs once snowmelt recedes and the stream's hydraulic head decreases (Huntington & Niswonger, 2012). This baseflow becomes essential in maintaining streamflow, especially during periods of low flow (Berghuijs et al., 2014; Liu et al., 2018; Dierauer et al., 2018).

Under the current climate in cold regions, snowmelt is a major component of the hydrologic system throughout the entire year and is a critical source of groundwater recharge and discharge. However, the interaction between snow and soil conditions is complex and can significantly alter these recharge processes. Taken together, understanding cold region groundwater resources requires identifying the roles of snow processes and soil-freeze and thaw dynamics.

1.2.2 Montane, Geologic, & Landcover Influences on Cold Region Groundwater Resources

Cold region groundwater processes are influenced by their elevation, topographic relief, geologic formations, soil characteristics, and vegetation cover. These characteristics can be spatially variable thus introduce spatial heterogeneity in groundwater processes. For this reason, understanding how these characteristics drive

differences in groundwater recharge and discharge rates and locations is essential to understanding the groundwater processes in the watershed as a whole.

1.2.2.1 Montane & Geologic Influences

Mountains and highlands often experience cooler temperatures due to their higher elevations and therefore are characterized by cold region hydrology. For these same reasons, snowmelt is a dominant source of mountain recharge (Earman et al., 2006; Somers & McKenzie, 2020). In combination with lower ET rates, mountains generally receive more precipitation due to orographic lift, resulting in enhanced potential for recharge (Dobre et al., 2025; Ward & Robinson, 2000; Somers & McKenzie, 2020). As such, mountain watersheds are frequently referred to as “water towers” due to their enhanced recharge capabilities and key role in providing water to adjacent downstream areas (Viviroli et al., 2007).

Mountain watersheds have unique and complex hydrogeological processes due to their large hydraulic gradients, complex hydrogeologic stratigraphy, and high topographic relief (Somers & McKenzie, 2020). Groundwater recharge in the mountain system can follow a shallow path before discharging to mountain streams and/or flow through deeper bedrock toward the basin aquifer (Somers & McKenzie, 2020). Groundwater in mountains can be a significant source of recharge and baseflow to adjacent lowland aquifers and streams, respectively (Somers & McKenzie, 2020; Hayashi, 2020). Mountain groundwater recharges to adjacent lowlands through mountain block recharge (MBR), recharge occurring in the mountain block which flows through the subsurface to an adjacent lowland aquifer, or mountain front recharge (MFR), recharge provided by infiltrating surface water at the transition between the mountain and valley floor that feeds a basin aquifer (Markovich et al., 2019; Somers & McKenzie, 2020; Figure 3). Differences in recharge sources and flow paths translates to differences in the magnitude and response time of low-elevation aquifers and streams to changes in high-elevation hydrologic processes (Markovich et al., 2019). For this reason, there is increasing work in literature towards quantifying groundwater flow paths and recharge and exfiltration locations (Markovich et al., 2019; Somers & McKenzie, 2020).

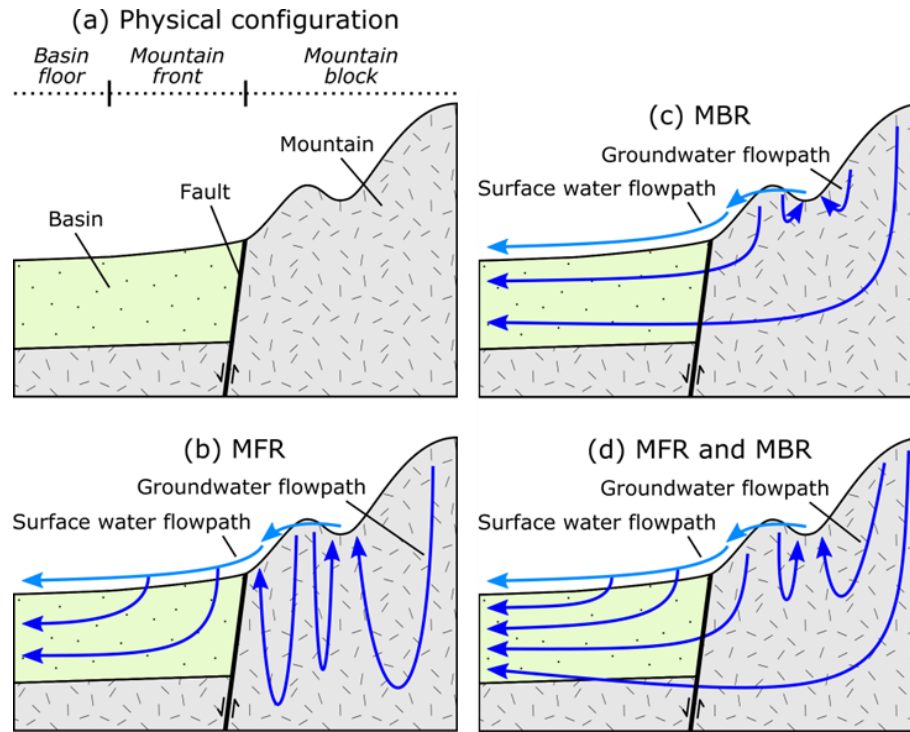


Figure 3: Conceptual diagram of (A) physical configuration of the basin floor, mountain front, and mountain block, (B) mountain front recharge dominated system, (C) mountain block recharge dominated system, and (D) mountain front and mountain block recharge system (modified from Bresciani et al., 2018).

Groundwater flow processes are strongly driven by the hydraulic conductivity of a watershed's soil and underlying geology and the hydraulic head gradients from local and regional topography. Mountains experience a disproportionately large amount of weathering, erosion, and mass wasting which can result in a variety of hydrogeologic features with varying hydraulic conductivities (Somers & McKenzie, 2020). For instance, high permeability coarse deposits, which are common features in mountainous watersheds, often channel flow and can act as groundwater stores due to their high porosity (Somers & McKenzie, 2020; Hayashi, 2020). Whereas deeper groundwater flow through bedrock, which often has lower permeability, typically increases with topographic relief (Somers & McKenzie, 2020; Toth, 1963). Although bedrock permeability, and thus groundwater flow, is also influenced by the lithology, stratigraphy, and structural features (Freeze & Cherry, 1979).

More broadly, soil characteristics can also be variable within watersheds and can drive differences in soil infiltration and freezing processes which in turn impact groundwater recharge. Soil water retention curves (SWRC) represent the relationship between a soil's matric potential and volumetric water content in unsaturated conditions and affect the hydraulic conductivity. The soil texture impacts the nature of this relationship with coarser texture soils typically retaining less water than finer textured soils at the same matric potential (Ochsner, 2019). Soil texture is defined by the proportions of sand, silt, and clay which are governed by the soil parent material, climate, vegetation cover, topography, and weathering processes (Ben Mahmoud & Zurqani, 2021). A higher bulk density also impacts SWRC characteristics, in that compacted soils typically have lower porosities and more negative air entry potentials (Ochsner, 2019). Bulk density is dependent on the soil texture, organic matter content, and the inherent densities of the soil material but also their packing arrangement and the soil depth (United States Department of Agriculture [USDA] Natural Resources Conservation Service, 2008). Typically, bulk density increases with depth due to reduced organic matter and root penetration, and increased compaction due to the weight of the overlying soil (USDA Natural Resources Conservation Service, 2008). Soil texture also impacts the freezing processes of a soil. For instance, Stuurop et al., (2022) found that intermediate soil textures may have higher ice saturation compared to other soil textures due to their high water retention capacity, preventing drainage, but also because their matric potentials are not strong enough to reduce the water freezing point.

1.2.2.2 Landcover Influences

In cold regions, and mountain watersheds, the difference in recharge rates between areas of the watershed are also closely tied to landcover differences impacting snow accumulation and melt. In high relief watersheds, snow processes are mainly governed by the interaction of wind, vegetation, and topography (Pomeroy et al., 2004; Essery & Pomeroy). For example, more snow tends to accumulate on the leeward side of a mountain or, on a smaller scale, a forest (Lundberg et al., 2016; Figure 4). Furthermore, snow accumulation and melt can be impacted by vegetation canopy characteristics like the species, stand density, and leaf area index (LAI); Increased forest density can reduce

the amount of incoming shortwave radiation to the snowpack and wind-induced sublimation, but this may be partially offset by increased longwave radiation emitted from the canopy (Pomeroy et al., 2009). However, the precipitation interception and net incoming radiation typically increase with canopy cover, thereby decreasing snow accumulation (Dobre et al., 2025). For example, Pomeroy et al. (2002) found snow accumulation decreased with increasing LAI. Likewise, Lundquist et al. (2013) found that forest cover reduces snow cover duration in comparison to open areas, as the forest increases longwave radiation in the spring thus accelerates melt. While there are many studies investigating the influence of coniferous canopy cover on snowpack (Lundquist et al., 2013; Pomeroy et al., 2002, 2009; Dobre et al., 2025), few studies focus on the impacts of a mixed canopy on snow (Pastore et al., 2025). In one such study, Pastore et al. (2025) investigated the effects of forest type and canopy characteristics on snow accumulation and melt in Acadian National Park in Maine, USA. They suggest that mature mixed forests with moderate canopy cover promote the deepest snowpacks as they provide the optimal balance between limiting snowfall interception while still providing snowpack sheltering from wind and solar insolation. Ultimately, cold region recharge is dominated by snowmelt processes, thus the impact of terrain and landcover on snow accumulation and melt directly impact the local recharge rates.

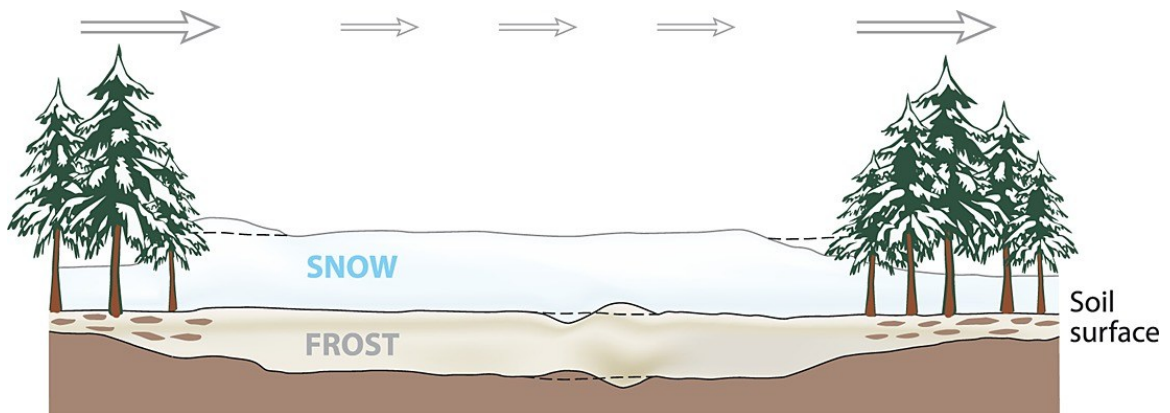


Figure 4: Conceptual diagram of snow distribution due to wind (arrows indicate the prevailing wind direction) and resulting frost distribution in a field between forests (Lundberg et al., 2016).

Wetlands are another landcover type which can have an important impact on groundwater processes in cold regions. Peatlands, which are the dominant type of

wetland in northern latitudes, have high porosity and storage capacity that can regulate groundwater levels (GWs) and recharge and exfiltration processes (Karimi et al., 2024; Bring et al., 2022; Locky & Bayley, 2006; Somers & McKenzie, 2020; Figure 5). Mountains often promote the formation of wetlands due to the common presence of poorly draining substrates and higher precipitation rates (Somers & McKenzie, 2020). High-elevation wetlands can be especially valuable sources of water storage as they commonly help to buffer streamflow by attenuating high flows and discharging to streams during periods of low flow (Buytaert & Beven, 2011; Squeo et al., 2006). Moreover, high-elevation wetlands can be recharge zones as they are often constantly saturated (Winter, 1999).

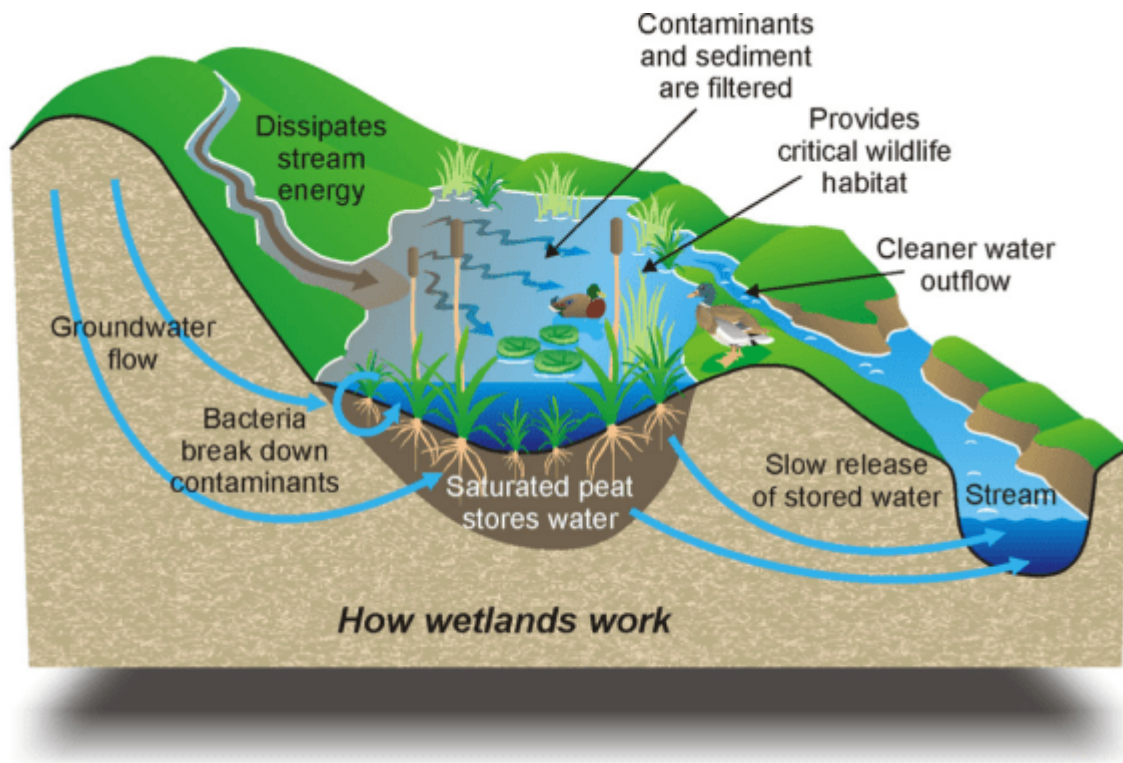


Figure 5: Conceptual diagram of the hydrological buffering and storage capacity of wetlands (Turner & Franklin, 2005).

1.3 CLIMATE CHANGE THREATS TO COLD REGION GROUNDWATER RESOURCES

Climate change has the potential to alter the amount and timing of groundwater recharge through changes in temperature and changes in precipitation amount, timing, intensity, and phase. As the climate warms, it is predicted there will be less snow falling

and accumulating and more rain (Intergovernmental Panel on Climate Change [IPCC], 2019). Changes to snow depth and soil freeze-thaw processes will directly impact groundwater resources in cold regions as these factors historically exert control on groundwater recharge and discharge (Jasechko et al., 2014). Furthermore, it is expected that high-elevation areas, which are typically snow dominated and important water sources, may experience more enhanced climate change effects than low-elevation areas (Vuille et al., 2018; Pepin et al., 2015; IPCC, 2019). As such, there is increasing concern about the implications of climate change on water security and ecosystem health.

Climate change impacts to groundwater may have serious ecological impacts due to changes in streamflow quantity and quality. Summertime streamflow is predicted to decrease and stream temperatures are predicted to increase due to climate change, which, in turn, applies stress to aquatic ecosystems and cold-water fish populations (Dugdale et al., 2013; Smith et al., 2023). However, snowpack provides insulation to the subsurface and groundwater. Thus, thinning snowpacks under climate change may partially mitigate these effects and play an important role in maintaining healthy exfiltration temperatures for aquatic ecosystems (Smith et al., 2023).

Climate change may put stress on current infrastructure and water management systems as it threatens to alter the quantity and seasonal timing of water availability. Moreover, the dependence on current water resources may become even greater in the future with population growth, land use changes, warming temperatures, altered precipitation patterns, and changes to surface water availability, many of which are amplified by climate change (Jyrkama & Sykes, 2007; Sulis et al., 2011; van Engelenburg et al., 2018). Therefore, a complete understanding of groundwater recharge and the mechanisms affecting groundwater recharge is essential to protect and preserve drinking water resources and groundwater-dependent ecosystems (Dierauer et al., 2018; Ndehedehe et al., 2023). The projected declines in summertime low flows could result in an increase in drought frequency and severity, putting stress on agriculture, industry, and drinking water resources (Barnett et al., 2006; Cochand et al., 2019). For example, Barnett et al. (2006) highlighted that the Columbia River system in the Western United States releases water from the McNary Dam to support hydroelectric power generation

and runs for federally protected salmon but with the projected increased severity of summer low flows, the reservoir will be unable to support both functions. Huntington & Niswonger (2012) predicted increased overland flow due to increased rain precipitation, more rain-on-snow events, and increased snowmelt rates (Huntington & Niswonger, 2012). Due to the low hydraulic conductivity of frozen soil in the winter, these changes may also promote flooding (Cochand et al., 2019). Although, other findings suggest snowmelt rates may decline in a warming climate due to less snow persisting until the high energy period when rapid melt historically occurs (Musselman et al., 2017; Barnhart et al. 2016).

1.4 CLIMATE CHANGE IMPACTS TO COLD REGION GROUNDWATER RESOURCES & PROCESSES

The response of groundwater recharge to climate change has been investigated in several studies, but few have directly considered the impact of a shift in winter precipitation phase. Generally, previous studies have found that finding changes to groundwater recharge are due to changes in precipitation amount, the infiltration capacity of the soil, and ET (Jasechko et al., 2014; Jyrkama & Sykes, 2007; Sulis et al., 2011; Meixner et al., 2016; Boumaiza et al., 2022). Below we discuss the possible changes to annual recharge, seasonal recharge and groundwater exfiltration in cold regions and the mechanisms which drive changes therein.

1.4.1 Annual Recharge

Climate change impacts on annual groundwater recharge in cold regions were mainly found to depend on the balance between projected changes in precipitation, infiltration, and increases in ET, but the impact on annual recharge efficiency under a precipitation phase shift is poorly understood (Meixner et al., 2016; Boumaiza et al., 2022; Jyrkama & Sykes, 2007; Samways et al., 2024). Meixner et al. (2016) found diffusive recharge in the Western United States was predicted to decrease for several of their southern study sites due to increased temperature and ET, and decreased precipitation. However, changes to groundwater recharge were more uncertain in northern areas due to the anticipated increase in precipitation (Meixner et al., 2016). Likewise, Samways et al. (2024) tested for temporal trends in GWLs for 171 wells in

mountains across North America and found that 69% of the trends were negative, of which, the largest historical declines in GWLs were correlated with higher average annual temperatures, lower average annual precipitation, and lower elevation regions. In contrast, in Eastern Canada, studies found increasing trends in groundwater recharge but at vastly different rates (between 0.7 mm/year to 100 mm/year; Boumaiza et al., 2022; Jyrkama & Sykes, 2007). In the Boumaiza et al. (2022) study, recharge was found to increase due to an increase in precipitation rather than improved efficiency, as surface runoff and ET also increased under climate change.

The disagreement between annual recharge trends between studies may be in part due to the fact that recharge processes are dictated by several spatially variable factors. Spatially variable factors governing recharge in a warmer climate as cited across literature include ground surface, soil, and bedrock characteristics, groundwater levels, precipitation amount, vegetation type, potential ET, saturated hydraulic conductivity, and the seasonality of rainfall (i.e., the timing and magnitude of rainfall; Jyrkama & Sykes, 2007; Boumaiza et al., 2022; Kim & Jackson, 2012; Samways et al., 2024). For instance, Samways et al. (2024) found sedimentary aquifers showed more negative historical trends in GWLs than crystalline aquifers due to their higher diffusivity and, thus, faster responses to changes in recharge. Altogether, studies indicate there are numerous interrelated factors which govern annual recharge and aquifer responses to climate change. Additionally, the differences in climate change impacts to recharge between locations indicate certain watershed characteristics may increase its vulnerability to climate change.

1.4.2 Seasonal Recharge

The response of groundwater recharge to climate change was also found to vary seasonally as the dominant controls on recharge differ across the seasons and due to the inherent uncertainty in utilizing climate change projections to project groundwater recharge. Under a warming climate, studies generally found wintertime recharge increased and summertime recharge decreased (Huntington & Niswonger, 2012; Cochand et al., 2019; Eckhardt & Ulbrich, 2003; Sulis et al., 2011). The dominant controls

impacting seasonal recharge in a warming climate were primarily ET, snowmelt rate, frozen ground, precipitation amount, and precipitation phase.

With climate change expected to induce even warmer temperatures, ET is anticipated to increase, especially for the spring and summer months and thereby decrease recharge during these seasons (Sulis et al., 2011; Goulden & Bales, 2014). Goulden & Bales (2014) predicted average ET in their mountainous Californian basin could increase by 28% by 2100. Thus, in the summer season, without supplemental precipitation inputs, this increase in ET has been predicted to contribute to a decrease in recharge (Sulis et al., 2011). Eckhardt & Ulbrich (2003) also found spring recharge was anticipated to decrease partially due to an earlier start to the growing season, which would increase ET.

Spring recharge is expected to decrease in a warming climate (Sulis et al., 2011; Eckhardt & Ulbrich, 2003). Sulis et al. (2011) found earlier, and lower intensity snowmelt was responsible for the decrease. Likewise, Barnhart et al. (2016) and Musselman et al. (2017) predict snowmelt rate will decrease in a warming climate, which may reduce the volume of water able to infiltrate deep into the subsurface and increase the volume of meltwater partitioned to ET. In contrast to other studies, Scibek & Allen (2006) found an increase in spring recharge but did not account for an earlier snowmelt or changes to snowpack amount in their hydrological model.

Some studies predict warming temperatures may reduce ground frost and may promote higher rates of winter recharge (Rivard et al., 2014; Jyrkama & Sykes, 2007). Although warmer winter temperatures are anticipated to decrease snowpack depths, which traditionally provides insulation to the ground; Therefore, warmer winters may instead increase ground frost (Hyman-Rabeler & Loheide II, 2023). Hyman-Rabeler & Loheide II (2022) found that melt events occurring following a previous snowmelt and freezeback were correlated with lower recharge. This suggests that, in a warmer climate, the frequency and period of temperature changes above and below zero, or conversely the presence of ground frost, may be a dominant control on wintertime recharge.

Precipitation amount and intensity was also found to be an important control on recharge especially for the summer months (Scibek & Allen, 2006; Rivard et al., 2014).

Scibek and Allen (2006) that their site in south central British Columbia historically had the highest recharge in summer months due to summer rain inputs. They modelled future changes to recharge under warming scenarios and found recharge was predicted to increase in the summer (Scibek & Allen, 2006). In contrast, a study by Rivard et al. (2014) in the Annapolis Valley of Nova Scotia (NS), predicted a decrease in summer recharge primarily due to a decrease in precipitation.

Several studies have highlighted the importance of precipitation phase for groundwater recharge, with conflicting results (Cochand et al., 2019; Eckhardt and Ulbrich, 2003; Godsey et al., 2014; Markovich et al. 2016; Sulis et al., 2011). Several studies have highlighted increased winter rainfall as a driver of increased recharge (Sulis et al., 2011; Eckhardt and Ulbrich, 2003; Cochand et al., 2019; Godsey et al., 2014). For instance, Godsey et al. (2014) studied groundwater in the Sierra Nevada mountains in the U.S. and found increased groundwater recharge in the winter, which was believed to be due to higher frequency and lower intensity winter rain. In contrast, Markovich et al. (2016) suggests a precipitation phase shift, with lower intensity infiltration events, cannot overwhelm soil water retention capacity and allow deep groundwater recharge to occur. Furthermore, Earman et al. (2006) an often-cited isotope hydrology study, proposed an increase in temperature and shift in precipitation phase to more rain and less snow, may significantly decrease groundwater recharge as they demonstrated that snow makes up a disproportionately large component of groundwater recharge compared to rain historically.

1.4.3 Groundwater Exfiltration

Several studies have highlighted climate change impacts to groundwater-surface water interactions which significantly influence summertime low flows. Numerous studies support that climate change will lengthen the period in which groundwater discharges to streams, due to an earlier peak snowmelt, and result in decreased low flows (Huntington & Niswonger, 2012; Godsey et al., 2014; Evans et al., 2018). Huntington and Niswonger (2012) applied twelve general circulation model climate projections to an integrated surface-subsurface model from 2010 to 2100 for three snow-dominated catchments in the Western United States; They found that during periods of rapid runoff

and interflow from rapid snowmelt, streamflow and hydraulic head in a stream are high, causing exfiltration to be suppressed until the end of freshet (Figure 6). Thus, earlier snowmelt recession and peak streamflow will advance the timing of groundwater exfiltration and lengthen the groundwater recession period, which, in combination with increased ET, results in lower summertime flows (Huntington & Niswonger, 2012). Carroll et al. (2024) used an integrated hydrological model of the Colorado River watershed and found climate change induces higher water use in forests which reduces groundwater recharge, groundwater storage, and annual streamflow. In their model of streamflow under a warming climate, warming only accounts for a modest decline in annual streamflow but depleted groundwater storage nearly doubled the annual streamflow loss, primarily due to increased seepage recharge in the winter.

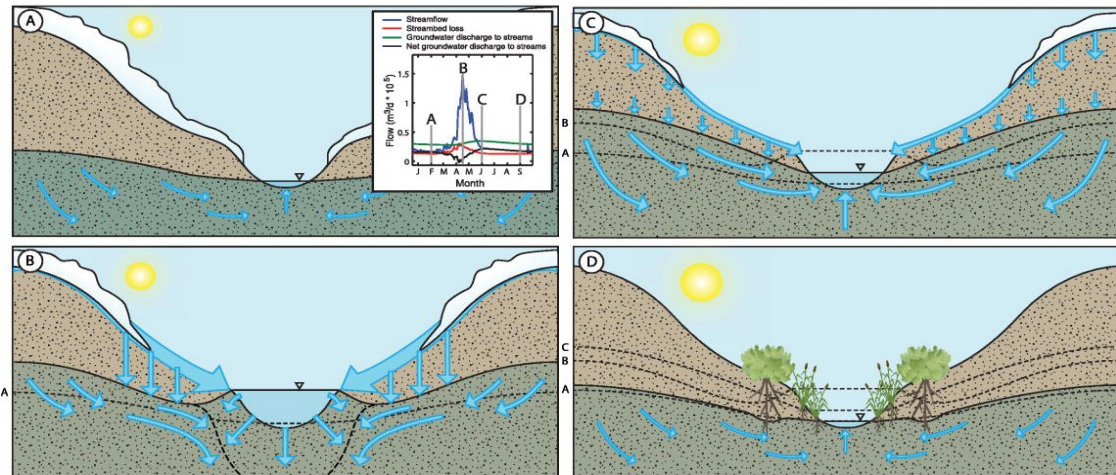


Figure 6: Conceptual diagram of groundwater-surface water interactions throughout the year: (A) early winter, (B) spring, (C) summer, (D) fall (modified from Huntington & Niswonger, 2012).

Studies which investigated the impacts of a precipitation phase shift on exfiltration found it may increase, but the ability of groundwater to offset loss in snowpack water storage and, as a result, the declines in summertime flows remains uncertain (Godsey et al., 2014; Jyrkama & Sykes, 2007; Sulis et al., 2011; Eckhardt & Ulbrich, 2003). For example, Godsey et al. (2014) examined the impacts of a shift in precipitation phase towards more winter rain and found that fall and winter net change in groundwater storage increased (i.e., a larger increase in groundwater storage over these seasons compared to historical conditions) but decreased for the spring months (i.e., more

groundwater storage loss than the historical period). The spring decrease was believed to be due to the loss of snowpack, as snowmelt historically helped to sustain streamflow and ET but in the warmer climate, these were instead compensated by a loss in groundwater storage (Godsey et al., 2014). As such, higher-storage groundwater systems are noted to have a higher capacity for buffering streamflow volume and timing in a warming climate, but at a cost to groundwater storage volumes (Godsey et al., 2014; Markovich et al., 2016). Annual net groundwater storage change in Godsey et al. (2014) increased by as much as 26% in the modelled catchments, but low flows still decreased by around 10% due to the precipitation phase shift. Thus, the increase in net groundwater storage change (via its contribution to baseflow) may be able to help partially replace the lost meltwater contribution to streamflow from diminished snowpacks, but this buffering capacity seems to be limited (Godsey et al., 2014).

Not only will the length and timing of exfiltration change in a warming climate, but the locations of exfiltration may also change (Evans et al., 2018). The study by Evans et al. (2018) was conducted in the Rocky Mountains of Colorado, U.S. using a groundwater flow and heat transport model and 50-year climate warming simulations. A warmer climate was predicted to reduce subsurface ice content by as much as 33% and result in rerouted groundwater flow paths. Groundwater discharge to streams was predicted to increase by 26% and discharge to hillslopes decreased by 4% (Evans et al., 2018).

1.5 CLIMATE CHANGE IMPACTS TO STREAMFLOW

Several studies have found similar seasonal trends between streamflow and groundwater recharge, with an increase in the winter and decrease in the summer (Cochand et al., 2019; Eckhardt & Ulbrich, 2003; Huntington & Niswonger, 2012). The synchronized trends between groundwater recharge and streamflow highlight the close relationship between the two systems and the importance of understanding these interactions. Studies primarily found that changes to streamflow were driven by changes to snowmelt timing and amount, ET, ground frost content, and precipitation phase. However, as noted in the previous section, changes to streamflow were also tied to changes in the groundwater system.

Winter is usually a period of low flow in cold regions, and winter streamflow was found to increase in response to climate change. Studies typically attributed this change to a combination of warmer temperatures promoting more snowmelt, an increase in precipitation, and more precipitation falling as rain (Eckhardt & Ulbrich, 2003; Cochand et al., 2019; Dierauer et al., 2018; Rood et al., 2008). Cochand et al. (2019) used an integrated surface-subsurface hydrological model for the St. Charles River Catchment in Quebec, Canada to project hydrological conditions from 2070 to 2100. They predicted a winter streamflow increase which was attributed to more winter rain combined with warmer temperatures resulting in higher snowmelt rates and, thus, an increase in meltwater contributions to streamflow during the winter. Likewise, Dierauer et al. (2018) conducted a correlation analysis of historical climate controls on streamflow in 63 mountainous catchments in North America from 1983 to 2012 and found less severe winter low flows were historically correlated with warmer and wetter winters.

Studies have found that annual peak streamflow occurred earlier in the year and that the magnitude decreased in response to climate change (Arnell, 1999; Stewart et al., 2004; Eckhardt & Ulbrich, 2003; Cochand et al., 2019; Rood et al., 2008). For instance, Eckhardt and Ulbrich (2003) used a conceptual ecological-hydrological model for a low mountain range catchment located in central Europe. From 2070 to 2099, spring streamflow decreased due to earlier and less snowmelt, warmer temperatures, and an earlier start to the growing season.

Several studies also stress that slower snowmelt rates due to a warming climate may decrease streamflow during the low flow period (Musselman et al., 2017; Barnhart et al. 2016). Barnhart et al. (2016) found higher snowmelt rates were correlated with higher daily streamflow generation overall. They propose the snowmelt rates exceed the field capacity of the soil, promoting percolation and subsurface flow, and in turn greater streamflow. In contrast, slower melt rates result in a higher volume of meltwater being lost to ET (Barnhart et al., 2016). Musselman et al. (2017) found these slower snowmelt rates were caused by less snow persisting until the high energy period which historically drives the faster melt rates. However, Barnhart et al. (2020) proposed runoff losses from slower snowmelt may be counteracted by the earlier timing of snowmelt, when energy

availability is lower, which reduces ET. Cochand et al. (2019) also attributed a decrease in summer low flows to increased ET but linked this to warmer temperatures and unchanged summertime precipitation rather than changes to snowmelt. Likewise, Eckhardt & Ulbrich (2003) also found a decrease in summertime flow due to increased ET, but also partly attributed to a decrease in summer precipitation for their warmest climate scenario.

1.6 KNOWLEDGE GAPS

The sustainable management of our groundwater and surface water resources rely on our understanding of their underlying hydrological processes and the threats they face in a warming climate. Studies evaluated in the literature review have typically found similar seasonal recharge responses to climate change (e.g., increasing wintertime recharge and decreasing summertime recharge) but the mechanisms driving these changes are less consistent and may be linked to differences in site characteristics. Many studies analyzing groundwater recharge impacts under climate change in North America are focused on watersheds in the northwest, whereas few studies evaluate watersheds in the northeast. The northeast regions of the continent can have substantially different landcover, topography, and climate, all of which have implications on groundwater recharge and exfiltration processes. Furthermore, many studies have evaluated snow hydrology in watersheds with forest canopies primarily dominated by coniferous forests but few with mixed forest canopies.

On an annual scale, the impact of climate change to groundwater recharge was less consistent across the different studies evaluated in the literature review. Many found the potential for winter rain to adequately recharge the groundwater system and counteract climate impacts to summer flows, but its ability is believed to be dependent on numerous other factors including ET rates, rainfall amount, intensity, phase, and timing, infiltration capacity, soil water retention, and groundwater storage capacity (Barnett et al., 2005; Jasechko et al., 2014; Jyrkama & Sykes, 2007; Barnhart et al., 2016, Markovich et al., 2016).

The impact of changing precipitation phase on annual and seasonal groundwater recharge in a warming climate is particularly understudied in literature (Figure 7).

Earman et al. (2006) is an often-cited isotope hydrology study which demonstrated that snow makes up a disproportionately large component of groundwater recharge compared to rain. However, the comparison in precipitation recharge ratios made in the Earman et al. (2006) study may not hold true in the future as it does not consider how the future climate may alter rainfall recharge efficiency in the winter. Rainfall in some cold-region climates (e.g. humid continental climates) can occur year-round but often rainfall occurs during periods with higher ET rates. In addition, under normal climate conditions, rainfall in the winter may not be able to properly infiltrate due to snow cover and frozen ground. As highlighted in the review by van Tiel et al. (2024), it is unclear whether the historical recharge efficiency differences between snow and rain are due to a difference in precipitation phase or season conditions. With warming temperatures, rainfall in the winter may more readily infiltrate and recharge into the groundwater system as suggested by several studies (Jyrkama & Sykes, 2007; Sulis et al., 2011; Eckhardt & Ulbrich, 2003).

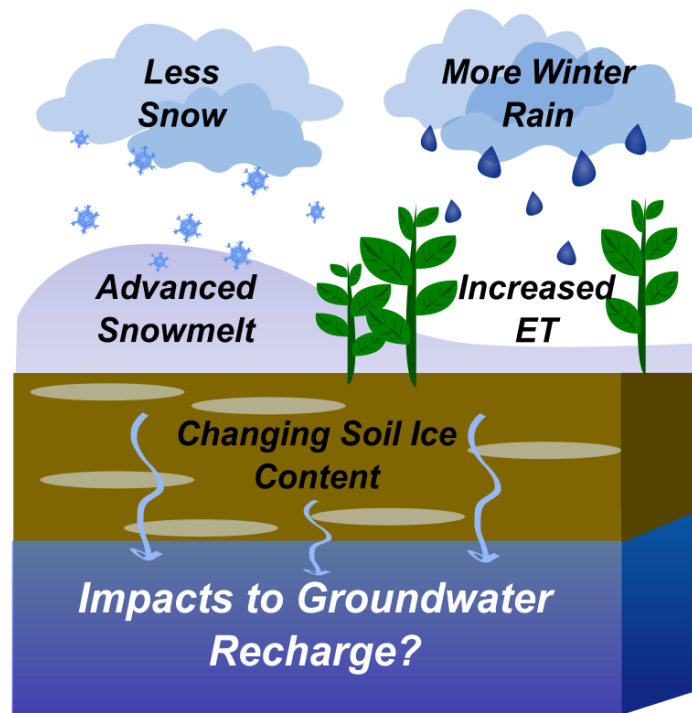


Figure 7: Conceptual diagram of climate change impacts to groundwater recharge.

1.7 RESEARCH OBJECTIVES

The goal of this study is to assess the implications of precipitation phase change on groundwater recharge for Christmas Brook Watershed in Eskasoni, NS. The objectives are to (1) determine if historical snowmelt runoff timing in streams throughout Cape Breton Island are shifting earlier in the year, (2) quantify differences in the historical recharge amount and efficiency across various topographic and landcover settings, and (3) project future changes in annual and seasonal groundwater recharge and recharge efficiency and compare the response between different topographic and landcover settings. Through these objectives, we will understand if increased winter rain can offset snowmelt losses in a warming climate.

We hypothesize that snowmelt timing in streams across Cape Breton Island will shift to earlier in the season (Objective 1) due to a trend in warming temperatures causing decreased snowpack and earlier and more frequent snowmelt events (Dudley et al., 2017; Stewart et al., 2004). To accomplish Objective 1, the center-of-mass method was applied to historical streamflow records to quantify snow melt-out timing dates, and a simple linear regression was calculated to detect trends.

For Objective 2, we hypothesize that there will be a difference in precipitation recharge efficiency between different topographic and landcover settings as recharge is controlled by soil characteristics, vegetation type, and precipitation (Cook & Brunner, 2025; Hyman-Rabeller & Loheide II, 2023; Kim & Jackson, 2012). We hypothesize that for Objective 3, there will be a significantly different response in recharge amount and efficiency to climate change between different landcover types as this response is believed to be dependent on numerous factors like soil properties, groundwater storage capacity, ET rates, and precipitation characteristics (Barnett et al., 2005; Jasechko et al., 2014; Jyrkama & Sykes, 2007; Barnhart et al., 2016, Markovich et al., 2016). Furthermore, we hypothesize that winter rain may offset recharge losses due to diminished snowpack in a warming climate if ground frost is reduced such that the infiltration capacity of the soil remains high. However, the recharge efficiency may be reduced due to the slower rates of snowmelt and higher ET rates in the spring (Hyman-

Rabeller & Loheide II, 2023; Eckhardt & Ulbrich, 2003; Barnhart et al., 2016; Musselman et al., 2017).

Objectives 2 and 3 were accomplished by analysis of historical observations of groundwater levels and one-dimensional modelling results of soil water movement using SHAW. SHAW was applied across three landcover types to quantify historical differences in recharge and recharge efficiency between landcover types under current climate conditions as well as in the future by integrating climate change projections into the models.

CHAPTER 2 - METHODS

2.1 STUDY SITE

The study site for this project is within the Eskasoni First Nation community in Cape Breton, NS, the largest Mi'kmaq community with a population of over 4500 people (Eskasoni Band Council, n.d.). The study site is also within the UNESCO Bras d'Or Lake Biosphere Region (BLBR). The study focuses on the Christmas Brook watershed and three sub-sites of interest: the Wetland, High-Elevation Forest, and Low-Elevation Forest sites (Figure 8). Field monitoring and numerical modelling is centered around these three landscape types.

An inactive geologic fault and the associated damage zone cross the Christmas Brook watershed, forming a fractured bedrock aquifer, which serves as a water source for the community of Eskasoni. Two public pumping wells intersect the aquifer, and pumping has occurred from the aquifer for over 20 years (Baechler, 2015). Our study is in collaboration with the Bras d'Or Lake Biosphere Region Association (BLBRA) and the Eskasoni Fish and Wildlife Commission (EFWC). The study area overlaps with a proposed Indigenous Protected and Conserved Area (IPCA) which is being developed by the EFWC. Study results will aid their understanding of hydrological processes in the IPCA and identify areas of hydrological importance to support source water protection planning by the Eskasoni Band and the Atlantic First Nation Water Authority. More broadly, this research will also improve the understanding of potential climate change impacts on water supplies in Atlantic Canada, low mountain areas and cold regions generally.

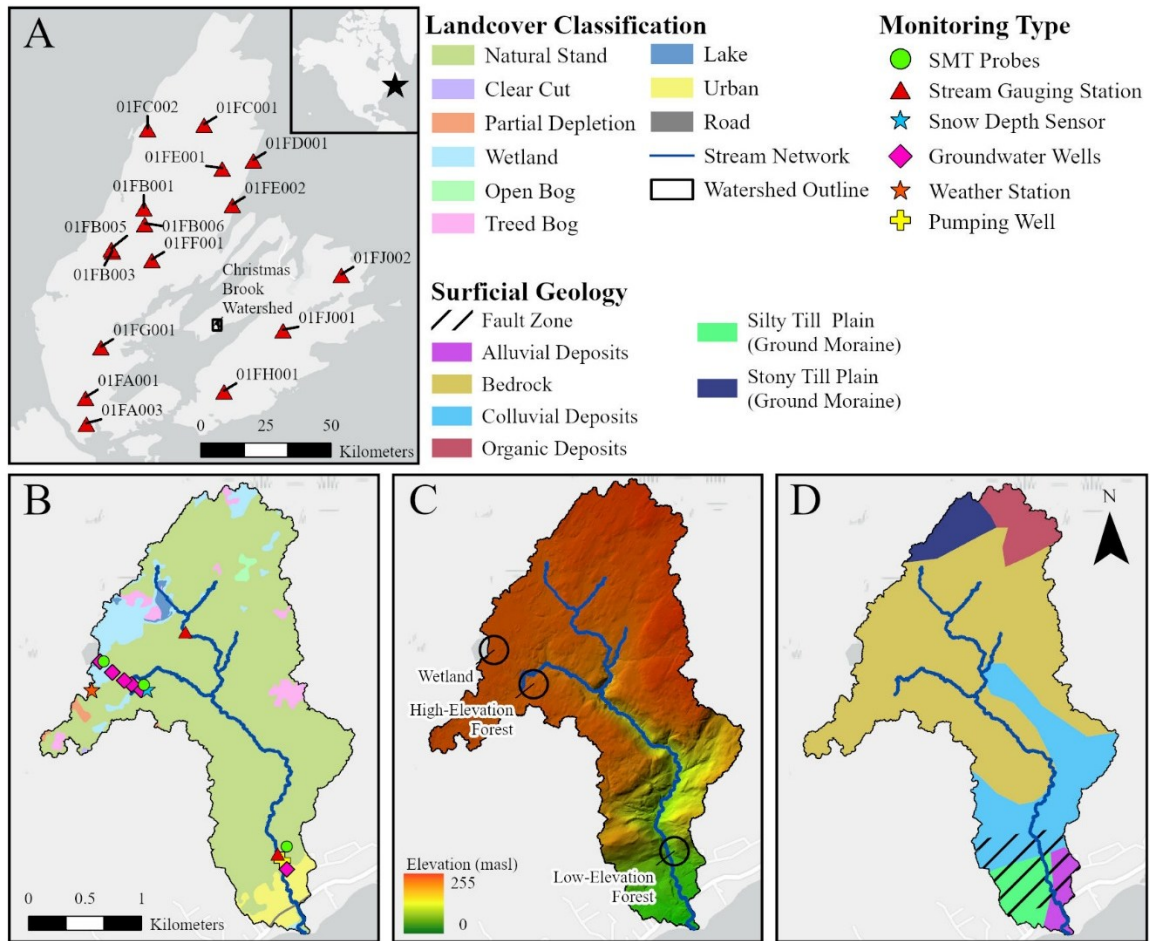


Figure 8: (A) Cape Breton Island stream gauging stations from the ECCC Water Survey of Canada & Christmas Brook watershed location, (B) Christmas Brook watershed landcover classification & monitoring locations, (C) Christmas Brook watershed hillslope & sub-site locations, and the (D) Christmas Brook watershed surficial geology & fault aquifer position.

2.1.1 Topography & Landcover

Christmas Brook watershed is 4.76 km² with an elevation range from sea level to 224 meters above sea level (masl). The watershed includes a higher-elevation (relatively speaking) plateau and steep incised valleys which drop down to a lowland area along the coast of Bras d'Or Lake. Much of the watershed is forested and mainly falls within the Cape Breton Hills ecodistrict. This district is covered predominately by hardwood trees, shrubs, ferns, and mosses. Tree and shrub species include sugar maple, red maple, beech, yellow birch, ironwood, and white ash (Neily et al., 2017). Hardwood Acadian forests are especially prominent in the low elevation and river valley of the watershed (Government

of Canada, 2018a). The average tree species height is around 13.5 m (Nova Scotia Department of Natural Resources and Renewables, Lands and Forestry Division, 2021).

In the high-elevation plateau, the landcover is characterized mainly by softwood boreal forest or mixed wood stands and sphagnum moss bogs. Boreal forests are dominated by balsam fir and white birch trees, and occasionally white and black spruce (Government of Canada, 2018b). Dominant hardwood species in this area are the same species in the lowland areas. The average height of tree species in the area is around 12.5 m. The main bog area is flat with hummocks and hollows, with an average peat depth of 1.08 m up to a maximum of 2.75 m (Nova Scotia Department of Natural Resources and Renewables, Lands and Forestry Division, 2021; EXP Services Ltd., 2011b).

2.1.2 Historical Climate

Cape Breton Island has a maritime boreal climate characterized by hot summers and cold winters (Government of Canada, 2018c). This region receives a mixture of rain and snow in winter, making it an ideal location to investigate the role of precipitation phase on groundwater recharge. Additionally, the range in topography in the site, allows us to evaluate the role that topography plays on historical recharge rates.

The low elevation area has a median annual air temperature of 6.8°C, median total annual precipitation of 1353 mm/yr, and average snow depth of 13 cm between 2015 and 2025 during the historical range of snow cover presence (November to end of April; ECCC, 2026b). The climate of the low-elevation and high-elevation areas in the study site are similar, but the high-elevation tends to be cooler (0.5°C cooler overall and 2°C cooler in winter), more humid, have a deeper snowpack, and around 100 mm more precipitation annually over the 2023 to 2025 period. The Christmas Brook headwaters, at an elevation of 218 masl, have a recorded a median air temperature of 7.1°C, median total annual precipitation of 1218 mm/yr, and average snow depth of 17 cm from 2023 to 2025 during the historical range of snow cover presence (November to end of April; Note air temperature and precipitation medians are over different year ranges thus do not reflect the lapse rate).

2.1.3 Geology & Soil

The bedrock geology of Christmas Brook watershed in the highlands and valleys is characterized by granodiorite originating from the Shunacadie Pluton and biotite-hornblende granodiorite of the Boisdale Hills pluton, as well as rock of the Bourinot Group. In the lowlands of the watershed, the bedrock is primarily sedimentary rock of the Grantmire Formation (Barr & White, 2017). For the three sub-sites of interest, borehole logs in the area classify the bedrock as igneous granodiorite with various degrees of weathering and fractures (EXP Services Ltd., 2011).

The watershed has a faulted contact which is a fracture zone at the interface of two rock units due to their relative movement, one of which is the Neoproterozoic (an era 1000 to 542 million years ago) granodiorite bedrock of plutonic origin and the other the Windsor Group consisting of interbedded evaporites and argillaceous sediment (EXP Services Ltd., 2020). The exact boundaries of the fault aquifer in Christmas Brook watershed have not been precisely delineated although hydrogeological investigations suggest the aquifer is along the slope of the Boisdale hills (Figure 9).

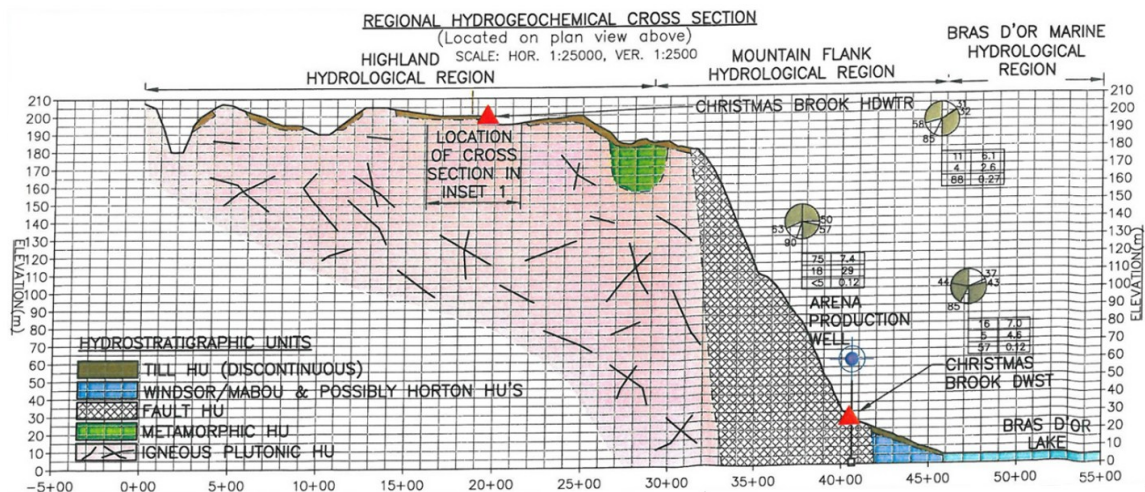


Figure 9: Conceptual Diagram of the Christmas Brook watershed regional hydrochemical cross section (modified from EXP Services Ltd., 2010).

The soils over the majority of the watershed belong to the Thom Soil series, which is a well-drained mineral soil with high porosity. The Thom Soil series is characterized as sandy loam to loam and stoniness typically increases with depth (Cann et

al., 1963). The depth of the soil layer varies across the site but is between 0.6 to 2 m in the highland plateau and becomes as deep as 11.5 m in the lowland area with high levels of stoniness at these lower depths (EXP Services Ltd., 2011b; Nolan, Davis & Associates Limited, 1984). Additionally, wetlands are extensive in the highland plateau area of the watershed. Bogs in this region generally have 25 to 40 cm of sphagnum moss above poorly decomposed peat (Cann et al., 1963). Peat depth probing in the Koqtuk bog found the peat depth ranged from 11 cm deep up to 375 cm, with an average of 108 cm (EXP Services Ltd., 2011b). There are also treed swamp areas within the high-elevation plateau area occurring near the bogs.

2.1.4 Hydrology

The main surface water feature in the watershed is Christmas Brook, a flashy stream with typically higher specific discharge rates in the headwaters but flow is generally higher at the low-elevation station. The watershed is gauged at two locations in the stream: one in the headwaters (SGS-Upper) and the other in the lowland area (SGS-Lower; Figure 8). The daily average flow rate of 0.2 m³/s at SGS-Lower and 0.158 m³/s at SGS-Upper. The GWLs in monitoring wells (MW) in the high-elevation area of the watershed tended to be closer to the surface than the low-elevation MW. The GWL in the wetland remains close to the surface year-round at an average depth of roughly 0.3 mbgs. In the forested region of the plateau, the GWL is typically deeper than the wetland, but remains close to surface at an average depth of 0.9 mbgs across the six MWs in the area. In the forested lowland region, the GWL is deeper in the ground with an average depth of 6.9 mbgs (Dauphinee, 2025).

The surface water and groundwater sources and flow paths in Christmas Brook watershed were characterized in depth by Dauphinee (2025) to highlight important water sources for Christmas Brook and the fault aquifer. Dauphinee (2025) found through hydrochemical analysis that streamflow is mainly composed of groundwater and wetland water, with increasing groundwater contribution with distance downstream. Furthermore, Dauphinee (2025) applied a hydrogeological model of the watershed and found that wetland water had shallow and slow flow paths whereas in the steep river gorge the groundwater flow was shallow and rapid. In the lowlands, Dauphinee's simulations

showed losing stream conditions in which surface water infiltrates into the fault aquifer at the upper fault boundary. However, there was little subsurface flow from the highlands to the lowlands.

2.2 FIELD DATA COLLECTION

2.2.1 Hydrological Monitoring Network

Working with EFWC and the BLBRA, an existing hydrological monitoring network in the Christmas Brook watershed in Cape Breton was utilized and expanded upon. The monitoring network consists of nine groundwater MWs and two stream gauging stations (Figure 8). The MWs are distributed throughout the watershed including at least one in each sub-site of interest. SGS-Upper is located near the plateau wetlands and SGS-Lower is in the lowlands near the Low-Elevation Forest site, both of which record water level every fifteen minutes. Observations of stage from the stream gauging stations were not used in this study's analysis.

Each MW has an associated borehole log from when they were originally drilled which characterize the substrate and well specifications (Table 8 in Appendix B). The bedrock MWs in the high-elevation area of the watershed consisted of three sets of two wells, one shallow (MW1A, MW2A, MW3A; between 4 to 6 mbgs) and one deep (MW1B, MW2B, MW3B; between 15 to 16 mbgs). The other two high-elevation MWs were located in the wetland (PNA and PNC) and monitored porewater in the peat (screened between 0 to 2.75 and 0 to 2.50 mbgs, respectively). Pressure transducers (Solinst LevelloggerTM) were installed in all of the shallow bedrock wells (MW1A, MW2A, MW3A), MW1B, both wetland MWs (PNA, PNC), and the two stream gauging stations to measure water level beginning in the summer of 2022. MW4, located in the lowlands, was also later equipped with a pressure transducer in August of 2023. Two barometric pressure loggers (Solinst BarologgerTM) were launched to record at a 15-minute interval in the summer of 2022, one in the high-elevation plateau and one in the lowlands, to compensate water level observations for air pressure in the MWs.

2.2.2 Meteorological Data and Snow Monitoring

There are two meteorological stations in Eskasoni recording at an hourly interval. Environment and Climate Change Canada's (ECCC) Eskasoni weather station level (WS-ECCC; station ID: 49748) is located near sea-level (elevation = 28 masl, distance to watershed = 3.3 km). The EFWC operated weather station (WS-Wetland) is located in the plateau wetlands of the Christmas Brook headwaters (elevation of 218 masl). WS-ECCC has continuous records since December 1st, 2011, and WS-Wetland was more recently installed in February of 2023. Both weather stations recorded meteorological observations which include air temperature, relative humidity, precipitation amount, wind speed, and snow depth. The Wetlands Station additionally recorded solar radiation. Additional solar radiation data required to supplement the meteorological observations for the project was retrieved from National Aeronautics and Space Administration (NASA) Langley Research Center's Prediction of Worldwide Energy Resources (POWER) for the area (NASA, 2026).

Given the importance of snow depth for model calibration, an additional snow depth sensor was installed at the High Elevation Forest plot in November of 2024. Manual snow surveys were completed in February 2024, February 2025, February 2026 and March 2026 to measure snow depth, density, and snow water equivalent (SWE). The snow surveys were completed using a government snow sampler at our three sub-sites of interest within the watershed. Sampling duplicates ($n = 12$) were performed for each sampling campaign, when possible, over a 10 m wide area in a cross shape (N to S and E to W) to account for spatial variability in snow characteristics. A total of 108 snow samples were conducted over the four campaigns. These survey observations aimed to validate the snow depth observations recorded by the snow depth sensors and were used for validating model simulations of snow density.

2.2.3 Soil Moisture & Temperature

Soil moisture and temperature (SMT) probes were installed in November of 2024 at each sub-site (Figure 8). The SMT probes were 90 cm in length (GroPointTM Profile Multi-depth Soil Volumetric Moisture Content and Temperature Profile Sensor) and recorded soil moisture across the entire rod length and temperature at nine depths with an

accuracy of $\pm 2\%$ and $\pm 0.5^\circ\text{C}$, respectively (RIOT Technology Corporation, 2024). The SMT probes use an antenna and time domain transmissometer to measure the time an electromagnetic wave takes to propagate through the soil. Soil moisture varies the propagation time. Thus, soil moisture is calculated from the wave propagation time measurement (RIOT Technology Corporation, 2024).

Three sets of two SMT probes were installed to record at a 15-minute interval at each sub-site of interest. The SMT probes were installed such that the lower probe number in the set was installed in a lower elevation location, and the higher probe number in a higher elevation. In the wetland, the probes were installed by pushing the probe into the peat directly, whereas in the mineral soil sites the probes were installed with a pilot rod and backfilled with a soil and water slurry. The wetland probes were installed to their full length, but the two mineral soils could only be installed part way (un-installment depth ranged between to 0.3 to 0.4 m). The uninstalled depth was recorded, and all measurement depths were corrected to be relative to ground surface. Measurements including the exposed portion of the rod were not used.

The SMT probes were developed for mineral soils and therefore required some corrections for organic soils with high saturated water contents. The SMT probes maximum measurable VMC was 55%, thus observations above this threshold were removed from observations for all probes. Additionally, the probe observations in the wetland had to be post-calibrated to reflect organic soil electromagnetic wave transmission behavior rather than the original calibration to a typical mineral soil properties. GroPoint™ Sensors come pre-calibrated with a dry calibration point such that at a dielectric constant (κ) of 5.598, the percent VMC was 7.9% and a wet solution such that a κ of 29.355 had a percent VMC of 41.2%. The sensor then used a linear relationship to yield VMC data (Equation 1).

$$\kappa = 0.7134 * VMC - 0.038. \quad [1]$$

The observations of VMC for the wetland SMT probes were used in this equation to calculate the measured κ , which were then entered into the Pepin et al. (1992) equation (Equation 2) relating κ to VMC in peat using:

$$VMC = 0.85e^{-1} + 1.92e^{-2} * \kappa - 0.95e^{-4}\kappa^2. \quad [2]$$

2.3 HISTORICAL SNOW MELT-OUT TIMING TREND ANALYSIS

Historical daily streamflow observations from streams across Cape Breton Island were obtained from the HYDAT database from ECCC (2026a) using the R package “tidyhydat” (Albers, 2017). All active or inactive stream gauging stations on Cape Breton Island were analyzed if they had at least ten years of streamflow observations and were missing less than five days of data between January and May (Figure 8).

The winter to spring Center-of-mass timing (CT) method was applied to streamflow records to quantify the timing of snowmelt each year. CT is the date on which the center-of-mass of streamflow from January 1st to May 31st passes and is commonly used in literature to detect trends in the timing of melt-out (Dudley et al., 2017; Shen et al., 2018; Stewart et al., 2004, 2005). The CT is calculated as:

$$CT_i = \sum(t_i q_i) / \sum q_i, \quad [3]$$

where CT_i is the CT date in the i^{th} year, t_i is the time-step, which was daily, and q_i is the streamflow for the timestep. The CT dates were then regressed against the year using a simple linear regression from the R Core Team (2023) package “stats”, and a slope value was computed to give the rate of change of the snowmelt runoff timing. A significance level of 0.95, or p-value of 0.05, was used to determine if slopes were statistically significant.

2.4 RECHARGE MODELLING

Due to the complexity of cold region groundwater recharge processes, predicting future groundwater recharge rates requires numerical modelling. The SHAW model was used to quantify seasonal and annual recharge and recharge efficiency across the three sub-sites of interest based on collected field data. Following the characterization of each site under historical climate conditions, the models were run using climate projections to determine the sensitivity of groundwater recharge to decreased snow and increased air temperature, evapotranspiration, and rain due to climate change.

2.4.1 Model Selection

Several popular numerical models exist which simulate soil freezing and thawing processes including: the Simultaneous Heat and Water (SHAW; Flerchinger & Saxton, 1989), Saturated-Unsaturated Transport (SUTRA-ICE; Voss & Provost, 2002), FEFLOW (Diersch, 2013), HydroGeoSphere (HGS, Brunner & Simmons, 2012), and HYDRUS-1D (Šimůnek et al., 2006) models.

SHAW is regarded as one of the more advanced models for simulating snow processes and the impact of snow cover on soil freezing and thawing (Liu et al., 2023; Nassar et al., 2000). SHAW is a physically-based one-dimensional numerical model developed by Flerchinger and Saxton (1989) which is used to simulate water and energy transfer for vertical snow and soil profiles. The accuracy of SHAW simulations of snow depth and soil freeze-thaw processes has been verified throughout the world, with several studies located in the United States, China, and Canada (Flerchinger et al., 1994; McDonald et al., 1996; Campbell et al., 2010; Huang & Gallichand, 2006; Chen et al., 2019; Hayhoe, 1994; Parkin et al., 1999; Nassar et al., 2000). Hayhoe (1994) used SHAW to test the simulation of soil freezing and thawing in Ottawa, Ontario and found the model successfully estimated the snow depth, soil temperature extremes, and changes in total and liquid water content of the soil in comparison to measured values. Likewise, Flerchinger et al., (1994) assessed the model simulations of spatially and temporally variable snowmelt and snow density in Idaho, United States and found it agreed well with measured values and simulated each component of the snow energy balance accurately.

Other available models have some limitations which make them less suitable for understanding and predicting changes to wintertime recharge processes. For instance, SUTRA and FEFLOW are purely subsurface models and do not simulate snow accumulation and melt processes, therefore would require coupling with an additional model. Furthermore, SUTRA and FEFLOW are not used to model the critical zone. The HGS model makes simplified assumptions, such as uniform soil thermal properties and negligible convective heat transfer into the subsurface, that may impede simulation accuracy (Wright & Novakowski, 2022; Schilling et al., 2019). HYDRUS-1D also simplifies melt processes by assuming the snow layer melts proportionally to the air

temperature. Additionally, the temperatures which partition precipitation into snow versus rain is fixed in HYDRUS-1D (Šimůnek et al., 2006) but not in SHAW. For these reasons, SHAW is believed to be the most suitable model for modelling cold region groundwater recharge processes.

2.4.2 Model Overview

SHAW is a one-dimensional hydrological model which simulates vertical water movement through layers of plant canopy, snow, residue, and soil, in which each layer is represented by a node (Flerchinger, 2000). The model inputs include meteorological data for each simulation time step and a number of parameters describing characteristics of each layer in the model. The SHAW model computes soil temperature, total soil moisture content, soil liquid water content, and soil ice content profiles for each soil node depth in the model. Additionally, the model outputs include the vertical water fluxes between soil nodes, and the lateral outflow and liquid water extracted by roots at each node depth. The model also provides a summary of the water and energy balances across the system, and simulations of snow depth and soil frost depth.

2.4.3 Key Model Equations and Numerical Methods

Within SHAW, the energy and moisture fluxes are computed between the nodes for each time step and the water and energy balances across the system are solved. SHAW uses implicit-finite difference form for each layer's water and energy balance equation which are solved simultaneously using the iterative Newton-Raphson method. The iterative approach balances the solving the two equations by alternating between solving the heat flux equations and the water flux equations, with the unknown values being updated after each solution. The iterative process stops when the approximations of the unknown values are within a user-defined prescribed tolerance.

2.4.3.1 Heat Fluxes

Heat fluxes are calculated within the snow and soil layers. The snow routine implemented in SHAW was developed by Anderson (1976). For a small layer within the snowpack, the energy balance equation is given by:

$$\frac{\partial}{\partial z} \left(k_{sp} \frac{\partial T}{\partial z} \right) + \frac{\partial R_n}{\partial z} = \rho_{sp} c_i \frac{\partial T}{\partial t} + \rho_l L_f \frac{\partial w_{sp}}{\partial t} + L_s \left(\frac{\partial p_v}{\partial t} + \frac{\partial q_v}{\partial z} \right), \quad [4]$$

where k_{sp} is the thermal conductivity of snow, T is the snowpack temperature, R_n is the net radiation, ρ_{sp} is the density of the ice fraction in the snowpack, c_i is the specific heat of ice, ρ_l is the density of liquid water, L_f is the latent heat of fusion, w_{sp} is the volumetric liquid water content of the snow, L_s is the latent heat of sublimation, ρ_v is the vapor density within the layer, and q_v is the downward water vapor flux. The equation terms relate to the net thermal conduction into the snow layer, the net radiation absorbed by the snowpack, the specific heat stored due to an increase in temperature, the latent heat for snowmelt, and the net latent heat of sublimation, respectively.

The heat transport processes in a soil layer used in SHAW is given by the energy balance equation:

$$\frac{\partial}{\partial z} \left(k_s \frac{\partial T}{\partial z} \right) + \rho_i L_f \frac{\partial \theta_i}{\partial t} + S = C_s \frac{\partial T}{\partial t} + \rho_l c_l \frac{\partial q_l T}{\partial z} + L_v \left(\frac{\partial p_v}{\partial t} + \frac{\partial q_v}{\partial z} \right), \quad [5]$$

where k_s is the soil thermal conductivity, ρ_i is the density of ice, θ_i is the volumetric ice content of the soil, C_s is the volumetric heat capacity of the soil, c_l is the specific heat capacity of water, q_l is the liquid water flux, and L_v is the latent heat of vaporization. The terms of the equation represent the net thermal conduction into the soil layer, latent heat due to water freezing, specific heat term for a change in energy stored due to a temperature increase, net thermal advection into a layer due to water flux, and net latent heat due to evaporation, respectively.

2.4.3.2 Water Fluxes

Water flux is calculated in the snow and soil layers as follows. Water flux within the snowpack is calculated from the energy balance as the change in liquid water content due to melting. This change is then accounted for at the end of each timestep in the density and thickness of the snow, and the excess water is converted into an outflow, as described in Anderson (1976). Infiltration at the soil surface is calculated using the layered Green-Ampt infiltration method, which assumes infiltration occurs as a distinct wetting front (Green & Ampt, 1911). Excess water not infiltrated in the model is assumed

to pond at the surface to a user specified maximum ponding depth, which, if exceeded, is then lost as runoff.

Liquid water or vapor transport within the soil profile in unsaturated conditions is simulated using a modified form of Richards (1931) equation. In saturated conditions, SHAW uses Darcy's law to drive liquid water flux. The equation for water flux through a soil layer in the model is given by the equation:

$$\frac{\partial}{\partial z} \left[K \left(\frac{\partial \psi}{\partial z} \right) + 1 \right] + \frac{1}{\rho_l} \frac{\partial q_v}{\partial z} + U = \frac{\partial \theta_l}{\partial t} + \frac{\rho_i}{\rho_l} \frac{\partial \theta_i}{\partial t}, \quad [6]$$

where K is the unsaturated hydraulic conductivity of the soil, Ψ is the soil matric potential, θ_l is the volumetric liquid water content of the soil, and U is a source/sink term for water flux. The equation terms represent the net liquid flux into the soil layer, net vapor flux into the layer, a source/sink term, and the rate of change in volumetric liquid water content and ice content, respectively.

The soil matric potential is a function of saturation and computed using a SWRC which can be defined by one of three models in SHAW. To minimize the number of parameters, we used the Campbell (1974) SWRC definition:

$$\psi = \psi_e \left(\frac{\theta_l}{\theta_s} \right)^{-b}, \quad [7]$$

where the Ψ_e is the air entry potential, θ_s is the saturated water content, and b is the pore size distribution parameter. The unsaturated conductivity is then computed from the equation:

$$K = K_s \left(\frac{\psi_e}{\psi} \right)^{\left(2 + \frac{3}{b}\right)}, \quad [8]$$

where K_s is the saturated hydraulic conductivity.

Ice Content

The Clausius-Clapeyron equation is used to define the formation and melt of ice content in the soils of the model. As a result, liquid water content can be defined as a function of temperature by the equation:

$$\theta_l = \theta_s \Psi_e \left(\frac{\frac{L_f T_c}{T_c + 273.16} + c R T_k}{g \Psi_e} \right)^{-\frac{1}{b}}, \quad [9]$$

where g is gravitational acceleration, R is the universal gas constant, T_k is the soil temperature in Kelvin, T_c is the soil temperature in Celsius, and c is the solute concentration in the soil solution. Any excess total water content not accounted for by the calculated liquid water content is assumed to be ice.

2.4.4 Model Domain and Parameterization

Each SHAW model domain extends from the top of the vegetation canopy to 10 m below ground surface (Figure 10). This depth was chosen to minimize the impact of the lower boundary conditions on the simulated recharge. The characteristics of the vegetation layer were primarily defined from biomass surveys at each site, and soil layers and characteristics were defined based on drilling logs and observations from the National Soil Database (Nolan, Davis & Associates Limited, 1984; EXP Services Ltd., 2011a; Agriculture and Agri-Food Canada, 2019a, 2019b). Each of the two forest sites had a top organic horizon layer, followed by a layer of soil, then for the low-elevation site only, a layer of bouldery till, with the bottom layer as weathered granitic bedrock (Figure 10). The Wetland site had a top layer of peat which represented the relatively uncompacted hummock, followed by a denser layer of peat, then weathered granitic bedrock (Figure 10).

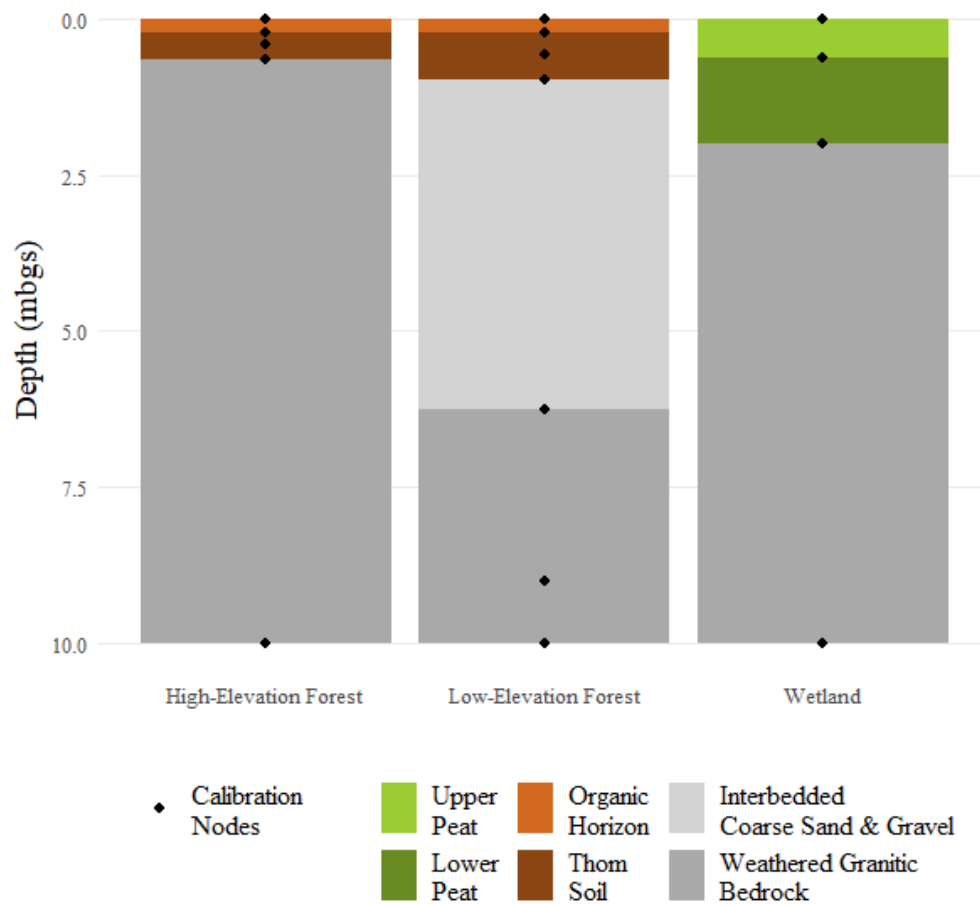


Figure 10: SHAW model soil layer domains.

The model required input parameters defining vegetation, snow, residue, and soil characteristics as well as general site characteristics such as elevation, slope, aspect, and wind surface roughness (Flerchinger et al., 2012). Since SHAW is a physically-based model, many of the parameters could be found either from site specific measured values or literature estimates of similar vegetation or soil characteristics. The uncalibrated parameter values and sources are found in the Appendix B (Table 9).

2.4.4.1 Vegetation Parameters

The main vegetation parameters in the model are listed in Table 3. Several other canopy parameters were required in the model but were parameterized with recommended values by the SHAW documentation (Appendix B, Table 9). A vegetation biomass survey was conducted at each sub-site in July of 2025 to help define several parameters in the canopy layer of the models for the two forest sites, namely DCHAR

and BIOMASS. During the surveys, the number of each dominant tree species in a 100 m² area were counted and the leaf dimensions and trunk diameter were recorded.

P_HEIGHT was obtained for each forest site from the Nova Scotia Forest Inventory shapefile for Victoria County from the Nova Scotia Department of Natural Resources and Renewables (2021). The trunk diameter, species type, and the plant heights of each species were used in the Canadian national tree aboveground biomass equations from Lambert et al., (2005) to calculate each species' and the total site's biomass in the area. The BIOMASS for the wetland site was obtained from the National Soil Database from Agriculture and Agri-Food Canada, (2019a) for the Castley soil (NSCY7~N) for a similar sphagnum peat bog in NS. The P_HEIGHT was set to 10 cm to allow for canopy ET process, but the remainder of the peat was modelled as soil. All other canopy parameters were found from literature estimates. For the forested sites with multiple species, the canopy parameter values (DCHAR, LAI, P_INT, P_ALB, ROOT_D) were calculated as the weighted average of literature estimates or measured values for each species, with the weights corresponding to the relative biomass of the species in the site.

Table 1: Calibrated Parameters & Values (Colour denotes calibration type, blue is snow, yellow is soil, green is both; Soil values presented as a range).

Model	Parameter Abbreviation	Parameter Name	Calibrated Value Range (units)
Wetland	SNW_TEMP	Snow threshold temperature	-0.1 (°C)
	LEAF_OR	Leaf angle orientation	4
	P_CLUMP	Leaf clumping parameter	0.5
	S_SR	Soil surface roughness	0.273
	KZ	Vertical saturated hydraulic conductivity	5E-3 to 45 (cm/hr)
	AEP	Air entry potential	-0.13 to -0.001 (m)
	B	Pore size distribution parameter	4.02 to 8
	LAI	Leaf area index	8.00
	DCHAR	Characteristic plan dimension	2.5 (cm)
	ROOT_D	Rooting depth	0.32 (mbgs)
	MX_PND	Maximum ponding depth	0.169 (cm)
	P_ALB	Canopy Albedo	0.25 (mm)
	P_INT	Canopy precipitation interception	2 (cm)
High-Elevation Forest	SNW_TEMP	Snow threshold temperature	0.0 (°C)
	LEAF_OR	Leaf angle orientation	1.0
	P_CLUMP	Leaf clumping parameter	0.6
	S_SR	Soil surface roughness	2.0 (cm)
	WIN_LAI	Winter leaf area index	3.0
	KZ	Vertical saturated hydraulic conductivity	0.15 to 23.03 (cm/hr)
	AEP	Air entry potential	-0.627 to -0.001 (m)
	B	Pore size distribution parameter	1.52 to 18
	SUM_LAI	Summer leaf area index	6.00
	P_ALB	Canopy Albedo	0.36
	P_INT	Canopy precipitation interception	0.25 (mm)
Low-Elevation Forest	SNW_TEMP	Snow threshold temperature	0.35 (°C)
	LEAF_OR	Leaf angle orientation	5.00
	P_CLUMP	Leaf clumping parameter	0.285
	S_SR	Soil surface roughness	2.00 (cm)
	WIN_LAI	Winter leaf area index	5.06
	KZ	Vertical saturated hydraulic conductivity	1.54E-2 to 21.00 (cm/hr)
	AEP	Air entry potential	-0.35 to -0.022 (m)
	B	Pore size distribution parameter	1 to 6
	PORO	Porosity	0.357 to 0.651 (m ³ /m ³)
	SUM_LAI	Summer leaf area index	5.06
	P_ALB	Canopy Albedo	0.163
	P_INT	Canopy precipitation interception	0.500 (mm)

The LAI for the two forest sites were modified using the optional plant growth input file in the SHAW model to account for the seasonal change in LAI due to leaf emergence and abscission. SHAW linearly interpolates the LAI between the start and end of user specified dates. The timing of leaf emergence was selected to start on May 1st and reach the full LAI by June 1st, and the leaf abscission was selected to start on October 1st and be complete by October 15th (Niinemets & Tamm, 2005; Rozenstein & Adamowski, 2017).

2.4.4.2 Soil Characteristics

The soil water retention curves were developed using observations of matric potential and water content, and saturated hydraulic conductivity in the National Soil Database for the Castley (NSCY7~N) and Thom (NSTHM3~N) soil series for the wetland and forest sites, respectively (Agriculture and Agri-Food Canada, 2019a, 2019b). The SWRC parameters were calibrated (see section 2.4.5) within limits which were selected from various literature sources as detailed in the Appendix. For both mineral and organic soil types, the SWRC was fit to the National Soil Database observation using the Excel Solver tool constrained by the parameter limits. Since each model extended 10 mbgs, the model domain also included a weathered granitic bedrock layer. The SWRC for bedrock for each site was again fit using Excel's Solver fit to soil water retention observations from Katsura et al., (2006) for weathered granitic bedrock, but used the saturated hydraulic conductivity results from in-situ slug tests.

Despite being a one-dimensional model, SHAW optionally allows for the inclusion of lateral hydraulic conductivities and thus lateral outflow and lateral inflow. To minimize complexity in selecting a realistic outflow and inflow rate, no lateral flow was assumed for almost all depths in the model. An exception was made for the wetland model in which it was assumed for the upper peat depths, which represented a hummock, these depths would likely drain into nearby hollows and not receive lateral inflow (Shi et al., 2015). As such, the lateral hydraulic conductivity was assumed to be 2.5 times in the vertical hydraulic conductivity (Guo et al., 2015).

2.4.5 Initial & Boundary Conditions

Initial conditions required for the model included snow depth and density, soil temperature, and water content (or matric potential) at each node at the beginning of the simulation for the soil nodes. A spin-up period of three years was included to minimize the effect of the ICs on the model simulations during the period of interest. Observations of soil LWC from the SMT probes were used to calculate the percent saturation at each node depth, which is equal to the LWC divided by the porosity of the soil node, thus assuming no ice content present. Using the percent saturation at the observation depths, and assuming full saturation at the GWL (as observed in the MWs), a logarithmic regression of saturation against depth was computed for each model. The relationship was then used to compute the percent saturation at each node depth above the water table and in turn the matric potential using Equation 7. In the saturated zone, the initial matric potential was simply equal to the depth below the water table. The temperature ICs for each soil node depth were calculated from a logarithmic regression of observed temperatures with depth. The regression used observations of temperature from each depth in the SMT probes and the assumption that below 6 mbgs the temperature is constant at 8.5°C (Gold, 1964).

The upper boundary conditions of the model are defined by the hourly inputs of meteorological data throughout the entire simulation period. The required meteorological data included air temperature, wind speed, humidity, precipitation, and solar radiation. Each model used a different version of meteorological data to best represent the different site conditions. Meteorological data consisted of observations recorded by the WS-ECCC and WS-Wetland with bias corrections of air temperature, precipitation, and humidity using seasonal lapse rates and windspeed adjustments using a correction factor for differences in canopy height and roughness to adjust for site differences (detailed in Appendix A).

The lower boundary conditions for the soil water content and temperature can either be specified by the user estimated by the model. We used the model-estimated option across all models wherein the water flux is assumed to be due to gravity drainage alone and temperature is estimated from the bottom two soil nodes.

2.4.6 Model Calibration & Validation

Snow depth was calibrated prior to the calibration of soil LWC and temperature given the importance of accurately simulating snow depth for the accurate simulation of soil freezing dynamics (Li et al., 2024). The parameters used in calibration were selected by conducting a preliminary sensitivity analysis on all model parameters and their impact on soil LWC and temperature, and snow depth (Table 1). Parameters were selected if they were impactful on the simulation outputs or had highly uncertain initial estimates. Each parameter used in calibration had specified limits as defined by literature estimates or measured values, detailed fully in Appendix B (Table 20 to Table 24). Due to the Wetland and High-Elevation Forest models being calibrated in part with GWL observations, the porosity for these models had to be pre-defined (i.e., not calibrated) using literature estimates or observed saturated water contents from the SMT probes. For the Low-Elevation Forest site, the model was not calibrated with GWL, thus the soil porosities were a calibration parameter. The resulting parameters used in each model's calibration are summarized in Table 1. Each parameter was constrained by upper and lower limits found either via literature estimates or ranges in measured values detailed in the Appendix (Table 20 to Table 24).

Each model was calibrated first manually for snow depth, then subsequently with PEST (Doherty et al., 2010) for soil LWC and temperature. Due to the limited canopy cover in the wetland site, the model's simulations were not sensitive to any parameters that simultaneously affected both snow depth and soil characteristics. For this reason, the wetland site snow depth calibration was unaffected by the soil characteristic calibration in PEST and could be completed independently. In contrast, the snow depth and soil characteristics of the two forest sites were simultaneously sensitive to canopy albedo and interception, thus the two calibration processes impacted one another. For this reason, the calibration process was iterative. The snow depth was calibrated first, and the overlapping parameters were allowed to change in the PEST calibration but with tighter limits. This process continued until all simulations obtained satisfactory error metrics.

A local search method was used to calibrate the snow depth, in which the initial model inputs were used as a starting point, and the snow depth calibration parameters

were modified one at a time, until converging on a local minimum. Snow depth was calibrated for each of the models using observations from the closest available snow depth sensor (Appendix B: Table 15) for one (High-Elevation Forest) or more (Wetland and Low-Elevation Forest) winters of snow depth observations. An hourly snow depth sampling frequency was available for the SDS-Wetland and SDS-Forest, but they were down-sampled to be at a daily frequency to make calibration targets equivalent to the SDS-ECCC observations. The Low-Elevation Forest model was calibrated prior to the availability of the 2025 snow season; Thus, this year was instead used in the validation process. The High-Elevation Forest model was the only site with only a year of data available for calibration. To avoid over calibrating the model to one snow season, which could potentially be unrepresentative of the average snow season, the simulated snow depth was also compared to the SDS-Wetland observations.

Calibration of the soil LWC and temperature was automated using PEST coupled with SHAW, completed by Strong et al. (2025). PEST iteratively adjusts model parameters within user defined limits to achieve a minimum objective function with user defined weights. Our objective function weights aimed to balance the performance of temperature and moisture simulations across all layers (Appendix B, Table 17). The algorithm used for PEST was the differential evolution algorithm which generates a population of stochastic models and uses vector differences to guide its search (Li & Yin, 2012; Castillo et al., 2020).

Each model was calibrated with one year of soil moisture and temperature data supplemented with GWL observations (Appendix B, Table 13). The soil volumetric moisture content recorded by the probes was compared to the SHAW model's output of volumetric liquid water content (LWC) as it was assumed that the probe could not accurately measure total volumetric water content when ice was present as in other studies (Flerchinger et al., 2006; Strong et al., 2025). The 15-minute SMT and GWL observations were down-sampled to match the hourly timestep of the SHAW models. Differences in the observation depths and number of observations between models is due to the differences in the version of the SMT probes and the installation depths of the probes.

GWLs were integrated into the soil LWC observations for the wetland and high-elevation forest models (Appendix B, Table 14) by assuming soil node depths just below the water table were at their saturated water content value. The high-elevation forested site GWL observations were well below the SMT probes, leaving several depths with scant calibration data. For this reason, the percent saturation was linearly interpolated between the bottom SMT probe depth and the uppermost GWL depth to avoid the model calibrating to only saturated conditions at these depths. The interpolated soil LWC conditions were only used to aid calibration convergence onto a reasonable solution but not used to evaluate calibration performance.

Several of the adjustable parameters used in PEST were soil characteristic parameters, meaning each depth in the model had a distinct set of soil characteristics. Since the number of soil nodes were on the order of tens for any given model, parameterizing the soil characteristics at every node was inefficient. Additionally, sharp changes in soil characteristics between nodes often introduced convergence issues. To resolve both issues, a subset of the soil nodes was selected to be adjustable during calibration, and all other nodes depth values in between the selected nodes were interpolated (linearly for all but hydraulic conductivity which was log-linear). Selected node depths were chosen to align with the top and bottom of the soil/rock layers, or occasionally in the middle for additional model flexibility (Figure 10). For most parameters, a linear interpolation was used, except the hydraulic conductivity in which a log-transformed linear interpolation was applied.

Calibration objective functions were defined by minimizing the Root-Mean Squared Error (RMSE) of soil LWC and temperature for several depths in the soil. The Nash-Sutcliffe Efficiency (NSE) and Percent Bias (PBIAS) metrics were used for snow depth. The RMSE is defined as

$$RMSE = \sqrt{\sum_{i=1}^n \frac{(Y_i^{sim} - Y_i^{obs})^2}{n}}, \quad [10]$$

where Y^{obs} is the array of the observed values, Y^{sim} is the array of simulated values, both of which contain constituents (i) from one to the total number of observations (n). The NSE is defined as

$$NSE = 1 - \left[\frac{\sum_{i=1}^n (Y_i^{obs} - Y_i^{sim})^2}{\sum_{i=1}^n (Y_i^{obs} - \bar{Y})^2} \right], \quad [11]$$

where $\bar{Y}$ is the mean of the observed values. Lastly, the PBIAS is defined as

$$PBIAS = \left[\frac{\sum_{i=1}^n (Y_i^{obs} - Y_i^{sim}) * 100}{\sum_{i=1}^n (Y_i^{obs})} \right]. \quad [12]$$

Model calibrations were considered acceptable if they achieved satisfactory error metrics as defined by limits reported in literature for SHAW or hydrological models more generally (Appendix B, Table 21). RMSE for soil temperature below 2.2°C and LWC below 0.9 m³/m³, and a NSE above 0.5 for snow depth were considered acceptable calibration criteria for this study (Chen et al., 2019; Flerchinger et al., 2003; Flerchinger et al., 2006; Li et al., 2012; Strong et al., 2025; Li et al., 2024; Moriasi et al., 2007). While the PBIAS metric is suggested by Moriasi et al. (2007) for calibrating streamflow, it aided the manual snow depth calibration process by indicating the simulated snow depth tendency to be higher or lower than observations. Snow depth error metrics were calculated by excluding timesteps where both simulated and observed snow depth were zero.

Snow depth and density measurements from the manual snow surveys were used to validate model results. In the two high-elevation sites, the 2025/2026 winter season were used, whereas the Low-Elevation site included both the 2024/2025 and 2025/2026 winter seasons (The Low-Elevation Forest model's snow depth was calibrated prior to the end of the 2024/2025 winter season). The simulated snow density in the models was compared to snow density measurements observed over four sampling campaigns which occurred between February 2024 to March 2026. For each site and sampling date, the median snow density was calculated and compared to the simulated snow density for the same date and respective model by calculating the coefficient of determination (R^2), to be consistent with other studies (Flerchinger et al., 1994), using

$$R^2 = 1 - \frac{\sum (y_{sim} - y_{obs})^2}{\sum (y_{obs} - \bar{y}_{obs})^2}, \quad [13]$$

where y_{sim} , is the simulated snow density, y_{obs} , is the observed snow density, and $\bar{y}_{obs}$, is the average observed snow density.

2.5 HISTORICAL & FUTURE RECHARGE ANALYSIS

2.5.1 Historical & Future Recharge Simulations

The historical simulations used in the recharge analysis ran from 2015 to 2025 with three additional years of spin-up which repeat 2015 meteorological data. The historical simulations highlight the current controls on summer and winter recharge, as well as the differences in recharge and recharge efficiency between landscape types.

The future climate scenarios were generated using the delta-change approach to modify the precipitation and air temperature of the historical climate data and determine the impacts on recharge. The delta-change approach is commonly used in literature to generate climate change data for hydrological modelling, especially for studies which require hourly meteorological data, whereas downscaled climate projections typically provide only daily data (Andaryani et al., 2019; Kuma et al., 2024; Li et al., 2025; Motovilov & Fashchevskaya, 2021; Rössler et al., 2012). For air temperature, the method takes the difference in historical and future air temperatures and adds this change to each day of the historical meteorological data. For precipitation, the relative change is computed and used to multiply the precipitation rate of the historical data. One major limitation of the delta-change approach is that it does not account for weather extremes. However, our analysis is not targeting extreme climatic conditions under climate change.

The climate change projections were retrieved from ClimateData.ca (2025) for the area and we used the 50th percentile 30-year seasonal climate normals from the Coupled Model Intercomparison Project Phase 6 (CMIP6) and selected the Shared Socioeconomic Pathways (SSP) 1-2.6, 2-4.5, and 5-8.5 to represent low, moderate, and high intensity climate change scenarios, respectively. Seasonal climate normals were used for the climate change projections for enhanced precision. The climate change projections were applied to the historical base years to create two additional climate periods representing mid-century (2055 to 2064) and end-of-century (2075 to 2084). Taken together, this resulted in seven climate settings per site (including historical simulations) which represented a range of climate change intensification conditions.

2.5.2 Recharge Quantification

Patterns in recharge (R) were examined at three temporal scales: hourly, seasonal and annual. In all cases, the total vertical water flux at a given soil node is defined as negative for upward flux and positive for downward flux; Thus, the sum of each vertical flux over a period of interest is equal to the net R. The node depths of interest (NOI) to analyze R in our study were selected to be below the lowest rooting depth across all models (0.32 m), such that any upward flux was due to capillary action only. The resulting NOI for the R analysis were 0.366 m for the Wetland and Low-Elevation Forest sites, and 0.361 m for the High-Elevation Forest site. These depths differed only due to differences in node spacing between the models.

2.5.3 Hourly Graphical Recharge Analysis

Hourly-scale graphical analyses of hydrological processes for the snow season (SS) and snow-free season (SFS) were conducted for the historical period and the SSP2-4.5 2055 to 2064 climate scenario. The historical simulations highlighted the drivers of interannual variability in historical R and the differences in R between snow and rain events. The projected simulations were examined to understand the hydrological processes that drove the projected climate change impacts on R. The SSP2-4.5 2055-2064 scenario was selected because it is a moderate change scenario which is perhaps more likely than the best- or worst-case scenarios. The graphical analysis included comparing hourly energy and water balance components such as air temperature, snow depth, frost depth, precipitation, ET, the VMC at the soil surface and NOI, and the vertical water flux across the soil surface and R. The full ten years were used in the graphical analyses for both seasons, but for illustration purposes, only 2016 and 2022 during the SS are presented in the results. 2016 was selected to represent a year with limited frost presence, whereas 2022 had deeper and longer frost presence. Cumulative R was also plotted for each year over the entire water year to show interannual variability in SS and SFS recharge.

2.5.4 Annual & Seasonal Recharge Analysis

Annual and seasonal recharge (R_A and R_S , respectively) and recharge efficiencies (REs) were computed for each year, site, and climate change scenario to quantify the spatial, seasonal, interannual and projected variability in R and RE. The R_A was calculated based on a water year starting on October 1st. Lower precipitation at this time of year aided in minimizing the misallocation of R events between years. Each year was then divided into the SS and SFS to quantify the seasonal recharge ($R_{SS/SFS}$). The beginning of the SS was selected to be the earliest historical date of snowfall across all years and sites. The SS end date was the latest historical date of simulated snowmelt across all years and sites, plus a spring lag time which accounts for the time between snowmelt and the associated recharge. This lag time was calculated by cross-correlating the simulated historical snowmelt in R using the function “ccf” in the R Core Team (2023) package “stats”. A lag of 1 day (rounded up to the nearest integer) produced the highest correlation. This lag was doubled (2 days) to be conservative and was used as the spring lag time. The resulting SS was from Julian day 309 (November 4th/5th) to 131 (May 10th/11th), and the SFS from Julian day 132 (May 11th/12th) to 308 (November 3rd/4th).

RE over some period x (RE_x) was computed as the sum of the total vertical water flux at the NOI (R), divided by the total precipitation (P_x) over the period of interest (x), where x is either annual (A) or seasonal (SS/SFS).

$$RE_x = \frac{\sum R_x}{\sum P_x} \cdot 100\%, \quad [14]$$

For the purpose of this study, we defined the RE as the amount of R occurring in the defined period over the total P occurring in the period, even if this R is later lost as ET outside the window of interest.

The change in snow and frost presence was also computed on a seasonal and annual basis and compared between scenarios and landscapes to highlight the shift in R conditions with climate change intensification and across different landscapes. The median snow depth and median frost depth were computed for each water year by taking

the depth for all time-steps with snow or frost depth above 0, respectively. The total number of days with snow cover and the total number of days with frozen ground were also computed for each SS.

Total R was calculated for each water year over the period of interest (season or water year). As a result, each simulation output of interest consisted of ten R values for each water year that were plotted as boxplots. A final value of ten R values was reported for each simulation output of interest. The percent change was then computed from historical to future conditions for recharge, snow, and frost. Percent change reflects the median difference between specific year pairs with the same base year (e.g., difference between total R of 2016 and total R of 2056), as opposed to the difference in medians calculated across all scenario data (e.g., difference between median of 2016 to 2025 and median of 2056 to 2065).

2.6 SENSITIVITY & UNCERTAINTY ANALYSIS

2.6.1 Sensitivity Analysis

For computational simplicity, a local one-factor-at-a-time sensitivity analysis was used to identify the sensitivity of the results to model parameters. The sensitivity analysis focused on the calibration parameters. A triangular distribution was generated for each parameter with the upper and lower bounds set as the upper and lower limits used in calibration which were defined from literature ranges of most likely values (Appendix B: Figure 38 to Figure 40). Triangular distributions are commonly used in uncertainty estimates and appropriate when there are known upper and lower limits and a preferred value (in this case the calibrated value) but no other information about the distribution (McMurray et al., 2017). The 40th and 50th percentiles of the distribution were used to modify each parameter from its original calibrated value. Each modified model was then analyzed for the total annual recharge, average recharge efficiency, average snow depth, and average frost depth. Each output was then compared to the original model's output, and a first-order sensitivity index (SI) was calculated with

$$SI = \frac{(O_{new} - O_{original})}{O_{original}} \bigg/ \frac{(P_{new} - P_{original})}{P_{original}}, \quad [15]$$

where O_{new} is the new output value, O_{original} is the original output value, P_{new} is the new parameter value, P_{original} is the original parameter value (Razavi & Gupta, 2015). The SI represents the relative change in the output to the relative change in the input parameter.

2.6.2 Uncertainty Analysis

Monte Carlo simulations were used to generate uncertainty bounds on annual recharge analysis results for the historical period and for the SSP2-4.5 2055-2064 scenario for each sub-site. The Monte Carlo method is a type of global uncertainty analysis and was chosen as it is more conservative than a local sensitivity analysis as it does not assume linearity and

simulation was run with 100 iterations and modified the same parameter set as in calibration and using the same triangle distribution as the sensitivity analysis. For each iteration, the parameters were randomly sampled from their triangular distribution to generate new model inputs. The new model was then run, and the total annual recharge, average recharge efficiency, average snow depth, and average frost depth was computed. The 5th and 95th percentiles, or 90% confidence interval, of the resulting distributions were presented as the uncertainty bounds. The percent uncertainty (% UC) was then calculated as

$$\% UC = \frac{(O_{0.05} - O_{0.95})}{2 \cdot O_{\text{original}}} \cdot 100, \quad [16]$$

where $O_{0.05}$ is the 5th percentile output value, and $O_{0.95}$ is the 95th percentile output value (McMurray et al., 2017).

CHAPTER 3 - RESULTS

3.1 HISTORICAL SNOW MELT-OUT TIMING TREND ANALYSIS

The CT linear regression analysis found that all but two streamflow stations in Cape Breton exhibited negative temporal trends (Table 2). However, only six of the sixteen sites had statistically significant slopes, all of which were negative as shown in Figure 11. Statistically significant slopes ranged from -0.091 to -0.307 days/year with R^2 ranging from 0.08 to 0.18. The average CT date for all stations for all years was March 29th. The average significant CT slope was -0.21 days/year, which is equivalent to the CT date moving two days earlier per decade.

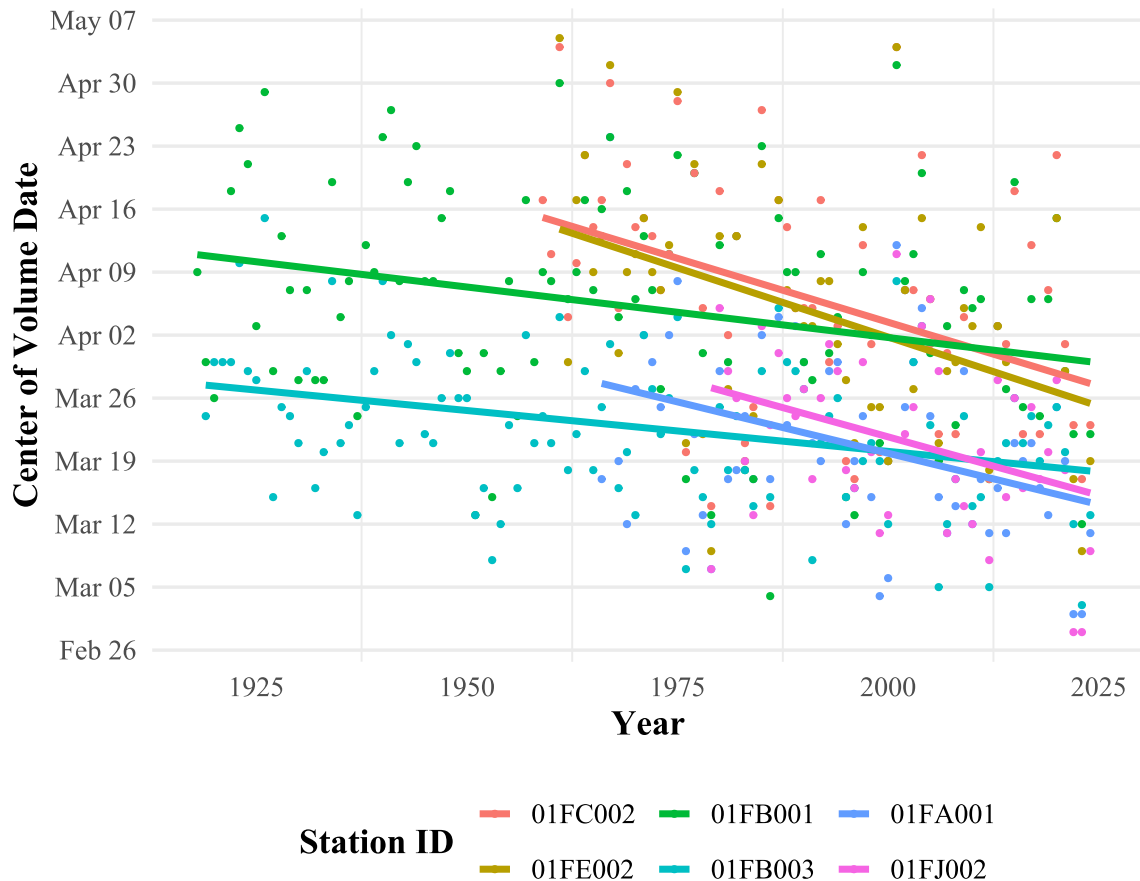


Figure 11: Significant CT date trends of Cape Breton Island streams.

Table 2: CT Linear Regression Results for Cape Breton Streamflow Stations (Rows highlighted green are statistically significant)

Station ID	Station Name	Slope (days/year)	R²	p	Record Length (Years)
01FA001	River Inhabitants At Glenora	-0.2281	0.1425	0.003205	59
01FA003	North Little River Below Beaver Dam Lake	-0.2774	0.0326	0.5032	16
01FB001	Northeast Margaree River at Margaree Valley	-0.1122	0.0785	0.003629	106
01FB003	Southwest Margaree River Near Upper Margaree	-0.0908	0.1111	0.000479	106
01FB005	April Brook At Gillisdale	-1.3881	0.1843	0.1637	12
01FB006	Lake O'Law Brook at Egypt Road	0.8603	0.134	0.1631	16
01FC001	Cheticamp River Below Cheticamp Lake	-0.4211	0.0583	0.3192	19
01FC002	Cheticamp River Above Robert Brook	-0.2839	0.1708	0.000688	64
01FD001	Wreck Cove Brook Near Wreck Cove	-0.1221	0.0048	0.7644	21
01FE001	Indian Brook Near Matheson's Lake	-0.6422	0.0796	0.2726	17
01FE002	Indian Brook at Indian Brook	-0.3074	0.1807	0.000881	58
01FF001	Middle River at MacLennans Cross	-0.5364	0.09	0.1865	21
01FG001	River Denys At Big Marsh	-0.3281	0.0557	0.3306	19
01FH001	Grand River at Loch Lomond	-0.0563	0.017	0.2675	74
01FJ001	Salmon River At Salmon River Bridge	0.0808	0.0051	0.7114	29
01FJ002	Macaskills Brook Near Birch Grove	-0.2598	0.1355	0.01184	46

3.2 OBSERVATIONAL DATA

Snow depth observed from SDS at each site show the two high elevation sites tended to have much thicker snowpacks than the Low-Elevation Forest site (Figure 12 to Figure 14). The two high elevation sites only had two to four seasons of snow depth

observations for comparison, but they showed mid season melt events tended to infrequent, whereas the Low-Elevation Forest site in some years had discontinuous snowpack cover for the winter due to more thawing degree days. Snow depth measured from snow sampling campaigns show high spatial variability in snow depth at each site. The median snow depth of most sampling campaigns tended to be close to the observations of the SDS (e.g., Wetland samples from the 2024/2024 sampling campaign; Figure 12), but occasionally could be largely different (e.g., High-Elevation Forest samples from the 2024/2025 sampling campaign; Figure 13).

Observations of soil LWC and temperature were plotted for the SMT probes along with GWLs used in the SHAW models (Figure 15). Complete observations of soil LWC and temperature for all SMT probes, GWLs, precipitation, and air temperature are available in Appendix B (Figure 32 to Figure 37). Between different probes at the same site, there were slight differences observed in soil temperatures and LWCs. This local variation was likely due to differences in probe installments depths as well as slight spatial variations in environmental conditions (e.g., degree of shading, wind exposure, snow accumulation, etc.). However, at the Wetland site, differences in soil LWC tended to be more significant between probes, likely owing to the fact SMT 1 was installed in a hollow whereas, SMT 2 was installed in a hummock. These local topographic differences resulted in higher LWC in the hollow than the hummock due to its lower elevation (Appendix B, Figure 32). At the Wetland site, the high-porosity of peat and highly saturated conditions also meant the soil LWC observations often exceeded the measurement limit of the SMT probes, especially in the hollow.

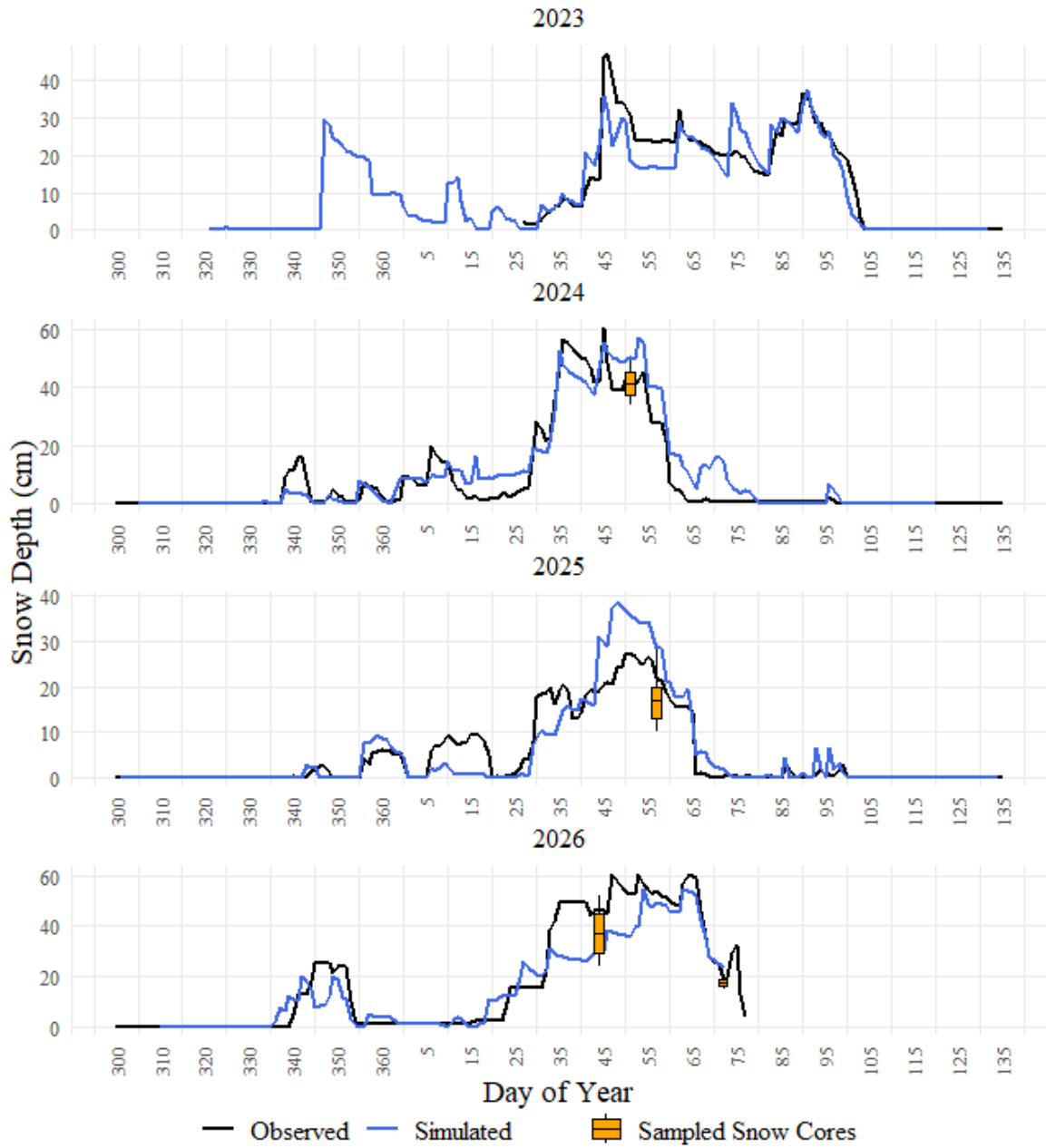


Figure 12: Wetland observed snow depth at weather station and from snow sampling campaigns compared to SHAW simulated snow depth for calibration (2023 - 2025) and validation (2026) periods (year label is the calendar year, e.g., 2023 is the 2022/2023 winter season).

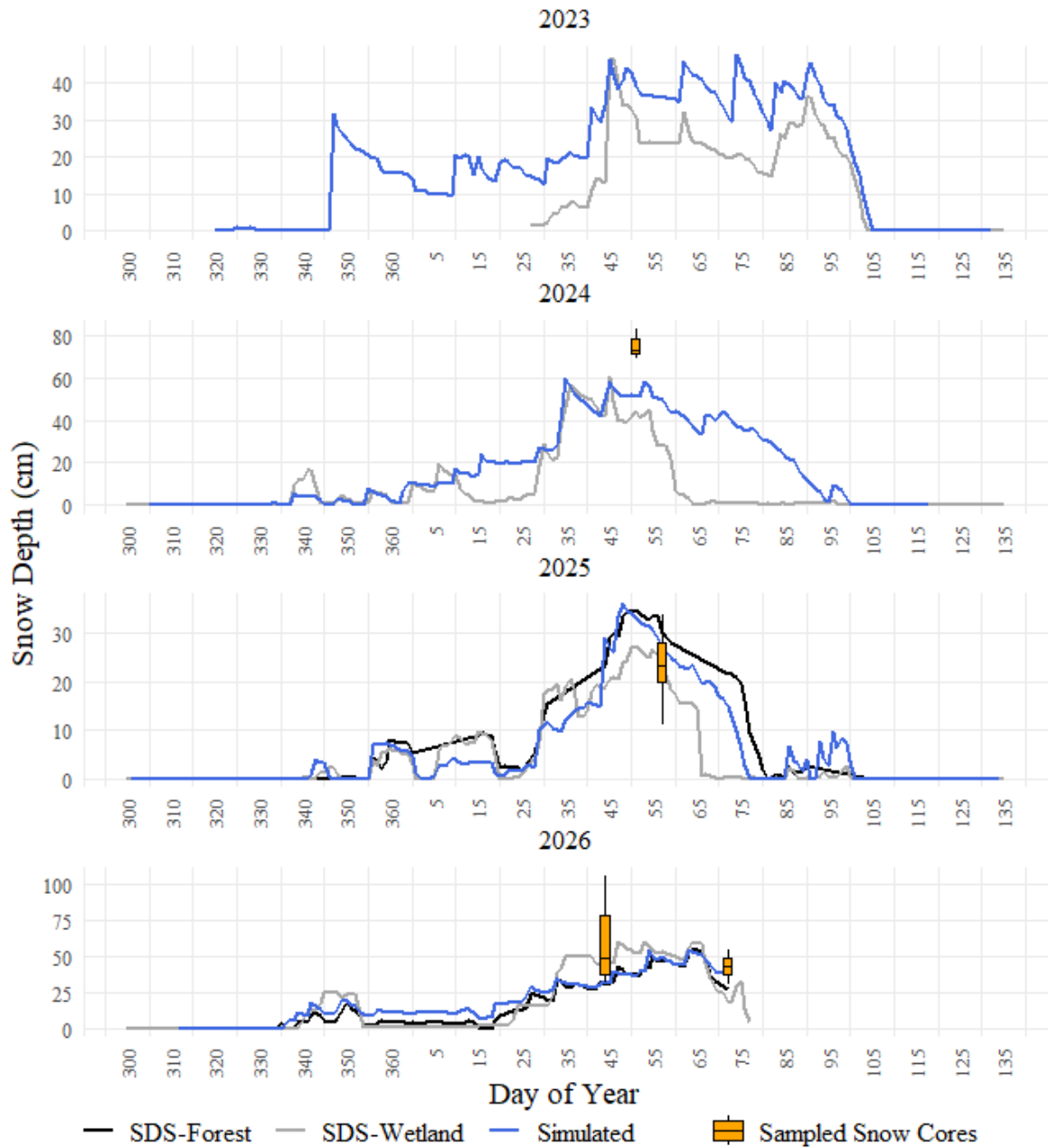


Figure 13: High-Elevation Forest observed snow depth at SDS-forest and SDS-wetland and from snow sampling campaigns compared to SHAW simulated snow depth for calibration (2023 - 2025) and validation (2026) periods (year label is the calendar year, e.g., 2023 is the 2022/2023 winter season).

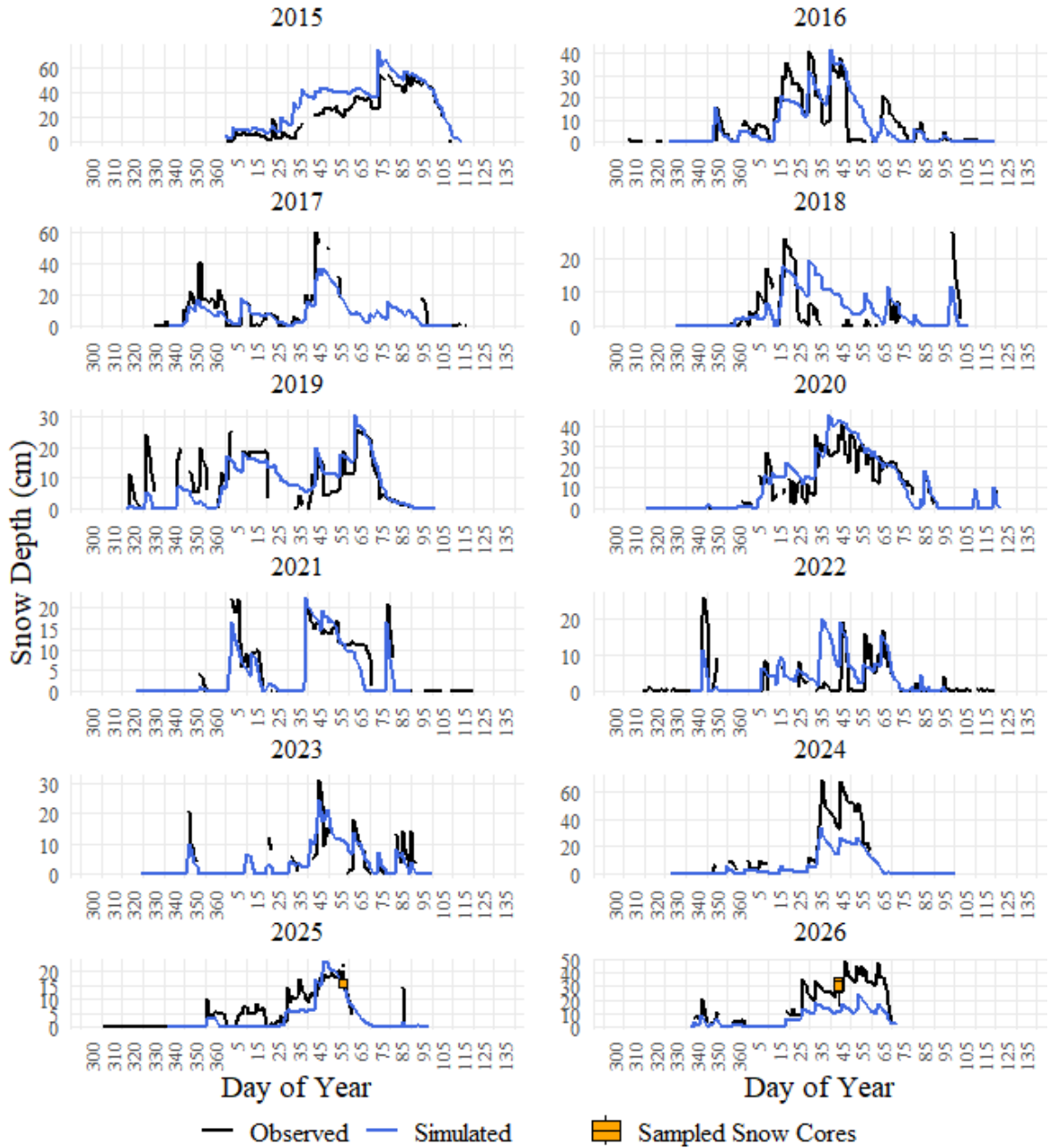


Figure 14: Low-Elevation Forest observed snow depth at SDS-forest and SDS-wetland and from snow sampling campaigns compared to SHAW simulated snow depth for calibration (2015 - 2024) and validation (2025-2026) periods (year label is the calendar year, e.g., 2023 is the 2022/2023 winter season).

Across the three sites GWL and soil LWC observations were substantially different, however the LWC were all closely tied to the timing of precipitation. The Wetland site had the shallowest groundwater levels overall (between 0.1 to 0.5 mbgs) and the highest soil LWC observations in comparison to the other two sites (Appendix B, Figure 32). The

High-Elevation Forest site also had a relatively shallow water table (between 0.5 to 3.5 mbgs) and for deeper observations depths of the SMT probe, the soil reached saturation for extended periods (e.g., SMT 4 0.336 to 0.486 mbgs in the spring of 2025; Appendix B, Figure 34). However, the soil LWC in the High-Elevation Forest site was generally lower than the Wetland due to the presence of lower porosity mineral soil. In the Low-Elevation Forest site, the soil LWC was never saturated for extended periods and had the driest soil conditions in comparison to the other two sites (Appendix B, Figure 36). MW4, which was the closest well (~220 m away) to the Low-Elevation Forest site, had a deep water table (>5 mbgs). MW4 has a significantly lower elevation than the SMT probes installed in the Low-Elevation Forest site (about 20 m lower) and MW4 is influenced by pumping, thus was not representative of actual GWLs for the Low-Elevation Forest site. However, based on the relative elevation of the MW4 GWLs it was anticipated that the GWL at this Low-Elevation Forest site was much lower than the two high elevation sites, which agrees with the low LWC observed at this site. For these same reasons, signals in the soil LWC observations at the Low-Elevation Forest site do not correlate well with signals observed in the MW4 GWL observations. In contrast, the LWC observations in the two high-elevation forest sites appeared well correlated with the respective GWL observations.

The three sites also exhibited differences between observed soil temperatures but were more consistent in comparison to soil LWC observations (Figure 16). The Wetland site tended to have the hottest surface soil temperatures in the summer likely due to its limited shade (Appendix B, Figure 33). In the winter the Low-Elevation Forest site most frequently experienced sub-zero soil temperatures at the soil surface (Appendix B, Figure 37). The Wetland site also experienced sub-zero or near zero soil temperatures in the winter at the surface of the hummock but less frequently. Whereas the High-Elevation site and the hollow in the Wetland site had air temperatures which were typically at or above zero for the majority of the winter season (Appendix B, Figure 35 and Figure 33, respectively). These air temperature differences could be partially attributed to differences in snowpack cover insulation effects or in the case of the Wetland site, the differences in specific conductivity of saturated and unsaturated peat.

Across all sites the variability of soil temperature tended to be higher at the soil-atmosphere interface, which was expected as soil closer to this interface would be more influenced by air temperatures. As such, the shallowest soil depths had the highest and lowest temperatures, in comparison to other measurement depths. Although, all depths show influence from the seasonal air temperature signal. Across all sites, the influence of snowpack thermal insulation on soil temperature can be directly observed. During the winter (i.e., during snow cover presence) the variability of soil temperature decreased, despite air temperature demonstrating similar variability to the rest of the record. Thus, suggesting snowpack has an insulating effect on the soil and weakened the influence of air temperature on soil temperatures. Despite this insulating effect, observations suggest the potential presence of frozen soil at the Wetland and Low-Elevation Forest sites. In the wintertime, LWC at the shallowest depth was observed to decrease or flatline at a low LWC in these sites (Appendix B, Figure 32 and Figure 36). This is likely due to the formation of ice which is poorly measured by the SMT probe. Observations of sub-zero soil temperatures occurred in the upper depths at the same time which further supports this theory (Appendix B, Figure 33 and Figure 37).

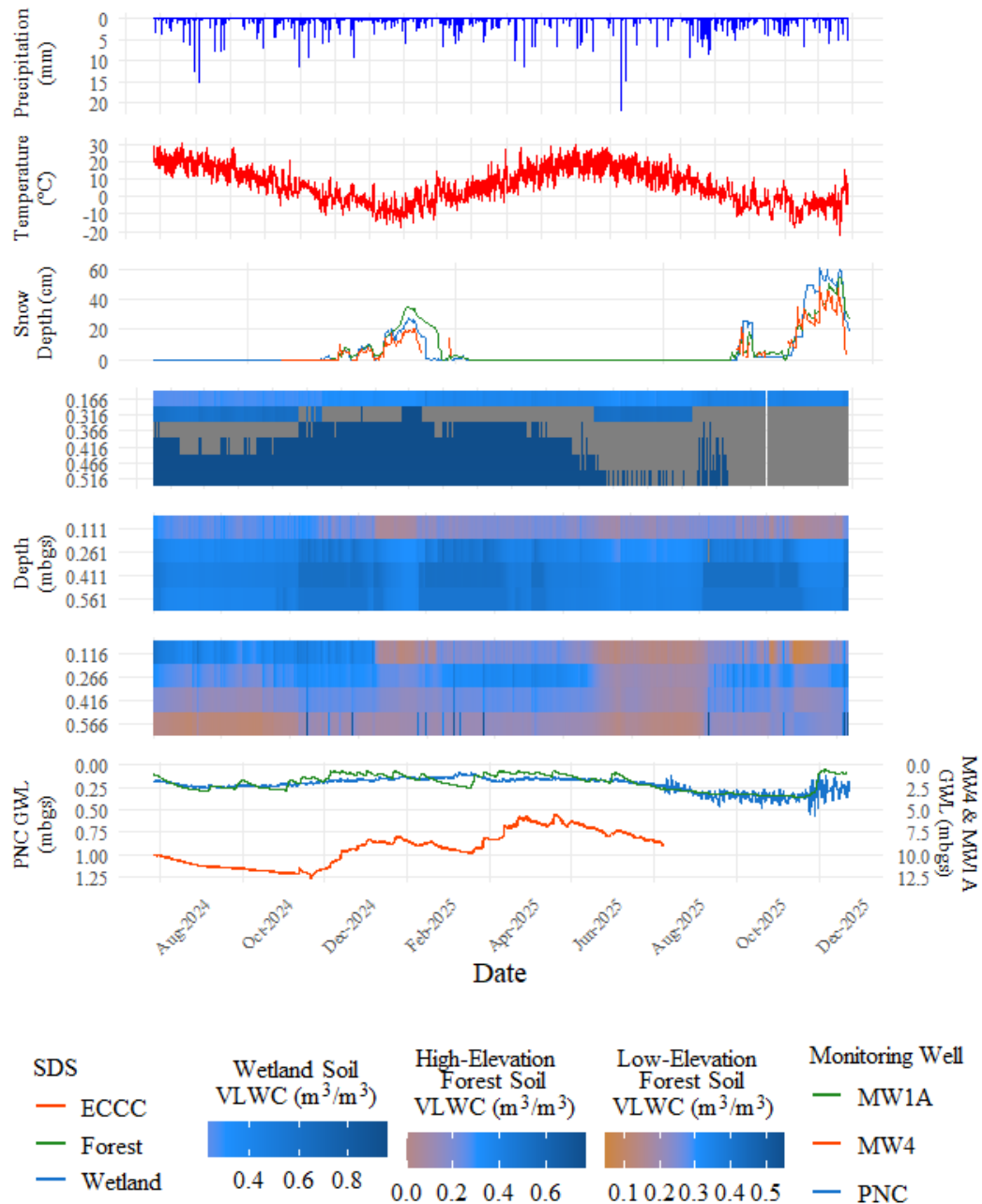


Figure 15: Observations of (A) precipitation and (B) air temperature recorded at the WS-wetland, (C) snow depths at the SDS-wetland, SDS-forest, and SDS-ECCC stations, and soil volumetric liquid water content at the (D) Wetland (SMT2), (E) High-Elevation Forest (SMT6), (F) and Low-Elevation Forest (SMT4) and groundwater levels recorded in MW1A, MW4, and PNC (note: the SLWC observations include corrected measurements from the SMT2 and groundwater level observations from PNC).

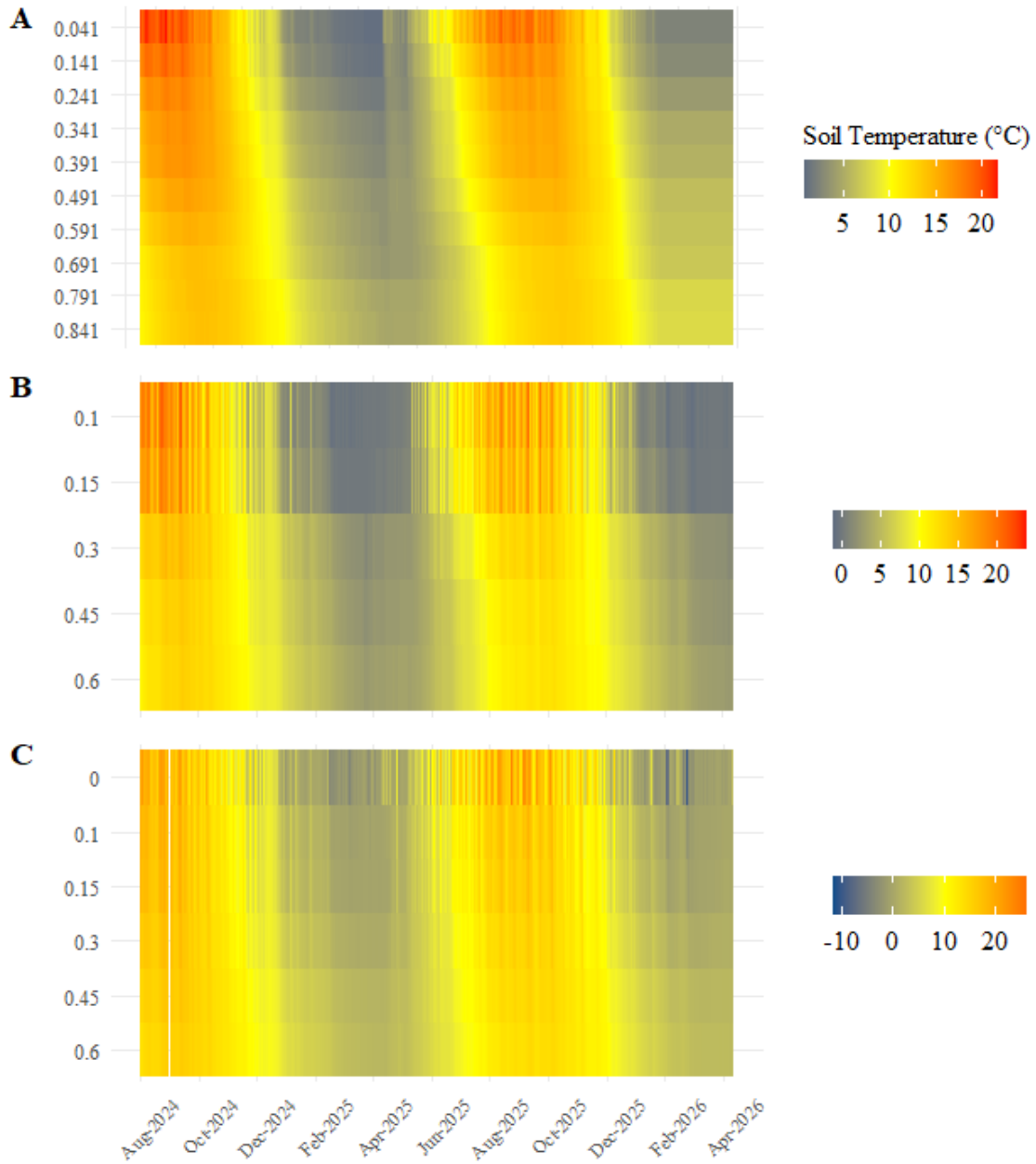


Figure 16: Observations of soil temperature at the (A) Wetland site (SMT2), (B) High-Elevation Forest site (SMT4), and Low-Elevation Forest site (SMT6).

3.3 SHAW MODEL CALIBRATION & VALIDATION

Across all models the snow depth simulations agreed well with snow depth observations. Snow density simulation performance was poor for all models but there were limited observations to compare to through time and of these observations they tended to exhibit high spatial variability in density. The soil temperature performed well across all

models generally, but the uppermost soil temperature tended to perform the worst. Likewise, the uppermost soil LWC simulations also tended to perform the worst in the models. The soil LWC was simulated within acceptable calibration criteria in two of the three sites. The Wetland model was unable to be sufficiently calibrated for the soil LWC however the performance was deemed adequate for interpretation.

3.2.1 Snow Depth & Density Simulation Performance

All models performed within acceptable calibration criteria for snow depth (Table 3). In the Wetland and High Elevation Forest sites the snow depth during the calibration periods had NSE of 0.79 and 0.91, respectively, which corresponds to “very good” fit according to Moriasi et al. (2007). The Low-Elevation Forest had a “satisfactory” fit with an NSE of 0.56. Additionally, all these sites had low RMSE and PBIAS.

As expected, the snow depth typically performed worse in the validation period. The validation period of the High-Elevation and Low-Elevation Forest sites both performed worse than the calibration period whereas the Wetland site performed slightly better in terms of NSE. However, both the High-Elevation Forest and Wetland remained within acceptable calibration criteria in the validation period.

While the PBIAS was low for most sites, some models displayed a over/under-estimation bias across the different years. At the High-Elevation Forest the comparison of simulated to observed snow depths recorded at the WS-Forest and WS-Wetland show the model tended to simulate more snow depth than observed in the WS-Wetland across all years and slightly underestimated the snow depth of the SDS-Forest (Figure 13 and Figure 12). While the WS-Wetland was not used in the calculation of calibration metrics, the comparison to the simulated depths gives context to the performance of the model outside the single year of WS-Forest snow depths. At the Wetland site, the simulated snow depth performed well both in terms of timing and magnitude for all three calibration years in comparison to observations for the WS-Wetland (Figure 12). Although, the 2025 snow season simulation appears to perform the worst with the peak snow depth being over estimated by around 15 cm. Lastly, the Low-Elevation Forest simulated snow depths appear to perform similarly for all years both terms of accuracy as

well as the direction and magnitude of bias (Figure 14). However, during the 2023/24 winter season the Low-Elevation Forest model largely underestimated snow depth

The simulations of snow density had poor performance across all models and indicated the model performed worse than the average of observations (Figure 17). The R^2 of the measured and simulated snow density was -0.6, -0.5, and -1.1 for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites (Appendix B, Table 27).

Table 3: Calibration & Validation Performance for Snow Depth, Soil Temperature, & Soil Liquid Water Content.

Simulation Type	Simulation Period	Model	NSE	RMSE (cm)	PBIAS (%)
Snow Depth	Calibration	Wetland	0.79	6.25	-5.2
		High-Elevation Forest	0.91	3.62	16.88
		Low-Elevation Forest	0.89	5.67	-22.63
	Validation	Wetland	0.82	9.1	17.2
		High-Elevation Forest	0.56	8.63	0.56
		Low-Elevation Forest	0.33	10.36	50.47
Soil Temperature	Calibration	Wetland	N/A	1.54	N/A
		High-Elevation Forest		1.17	
		Low-Elevation Forest		1.95	
	Validation	Wetland		2.08	
		High-Elevation Forest		1.05	
		Low-Elevation Forest		2.32	
Soil Liquid Water Content	Calibration	Wetland	N/A	0.124	N/A
		High-Elevation Forest		0.04	
		Low-Elevation Forest		0.037	
	Validation	Wetland		0.114	
		High-Elevation Forest		0.03	
		Low-Elevation Forest		0.053	

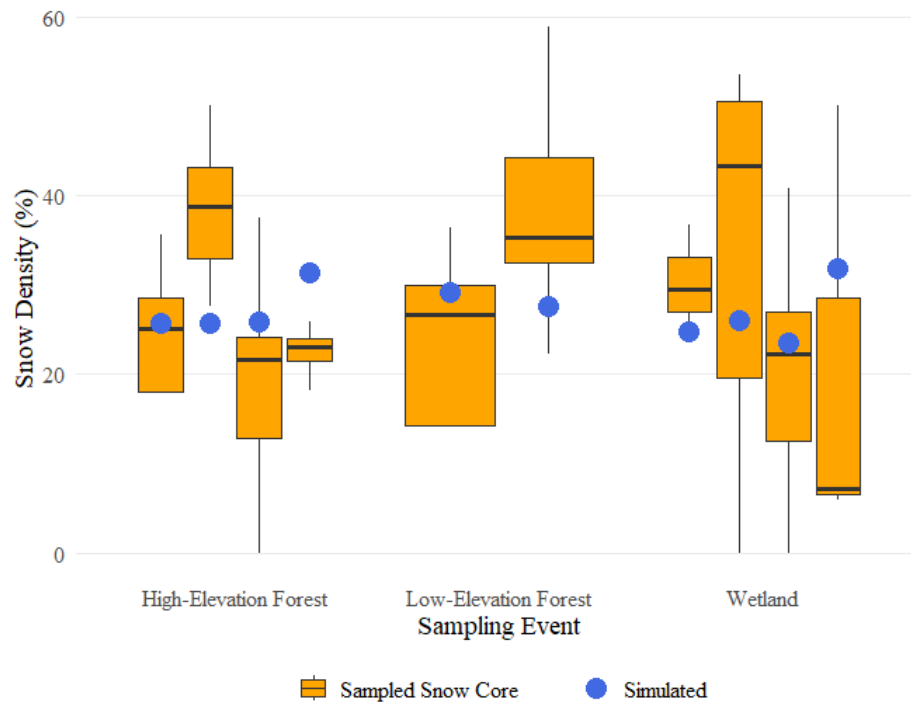


Figure 17: Measured snow density from snow sampling campaigns and simulated snow density (the low-elevation forest only has two sampling events as snow depth was too low for two of the four other sampling campaigns).

3.3.1 Soil Temperature Simulation Performance

The soil temperature during the calibration period was typically within the acceptable performance criteria across all sites and depths except in the near-surface node depths which typically had the highest RMSE values (Appendix B, Table 24 to Table 26). The Wetland model was within the acceptable RMSE for soil temperature for the average across all depths (Table 3) and all depths individually except the soil node at 0.141 mbgs (RMSE between 0.92°C to 2.3°C; Appendix B, Table 24). The Wetland soil temperature tended to closely follow observed temperatures year-round (Figure 19). The average across all depths and the individual soil nodes all had acceptable RMSE for the High-Elevation Forest site (RMSE between 0.89°C to 1.3°C; Appendix B, Table 25). The soil temperature for the upper two depths in the High-Elevation Forest performed better than the lower depths used in calibration (Figure 20). The Low-Elevation site had an acceptable average RMSE, but the topmost soil node was above the acceptable RMSE

(RMSE between 1.55°C to 2.49°C; Appendix B, Table 26). The soil node temperature performance in the Low-Elevation model improved with depth (Figure 20) .

The validation period soil temperature RMSE generally performed worse overall and more of the depths exceeded the acceptable calibration criteria. The Wetland model average RMSE increased in the validation period to 2.08°C which was still within acceptable limits (Table 3). The Wetland model performance tended to decline more for the near surface nodes (RMSE between 1.25°C to 2.86°C; Appendix B, Table 24). The High-Elevation Forest model had slightly improved performance during the validation period on average with RMSE decreasing to 1.06°C (Table 3); The upper nodes had slightly increased RMSE, whereas the lower nodes were largely decreased (RMSE between 0.875°C to 1.32°C; Appendix B, Table 25). The Low-Elevation Forest model had reduced calibration performance on average, with RMSE increasing to 2.49°C (Table 3). All depths had increased RMSE during the validation period but most significantly increased closer to the ground surface (RMSE between 1.90°C to 3.2°C; Appendix B, Table 26).

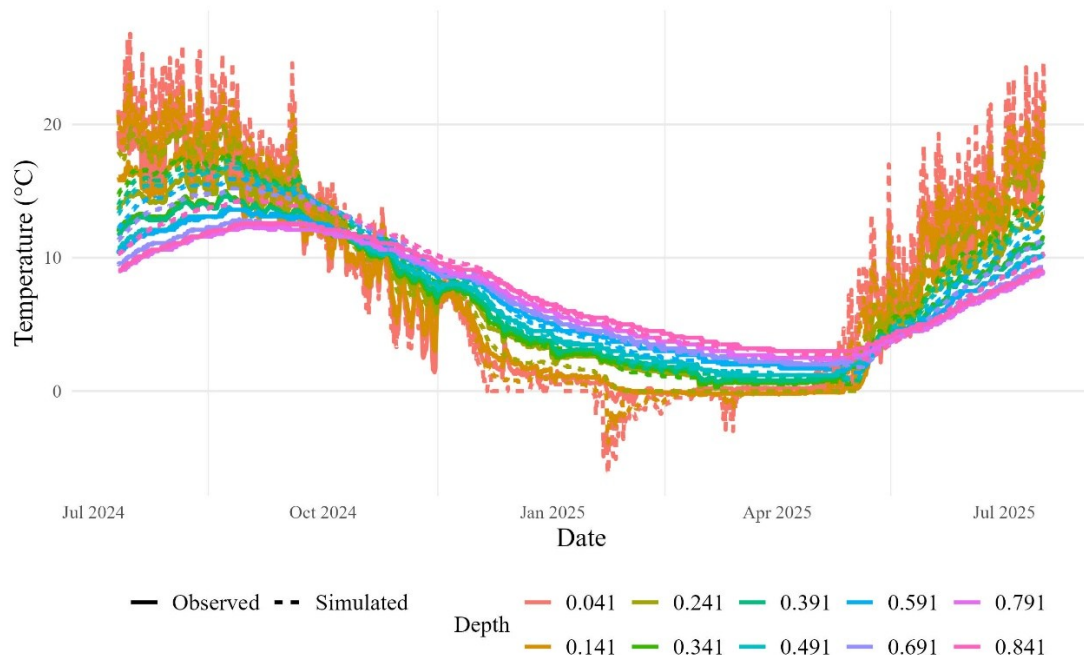


Figure 18: Wetland model soil temperature calibration performance.

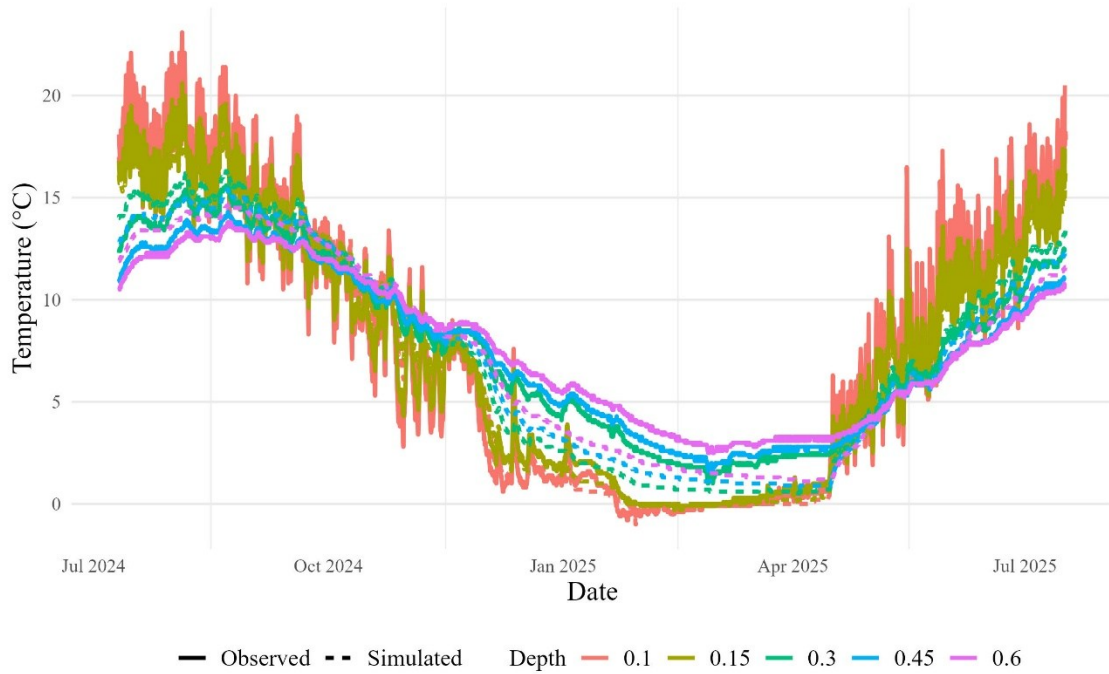


Figure 19: High-Elevation Forest model soil temperature calibration performance.

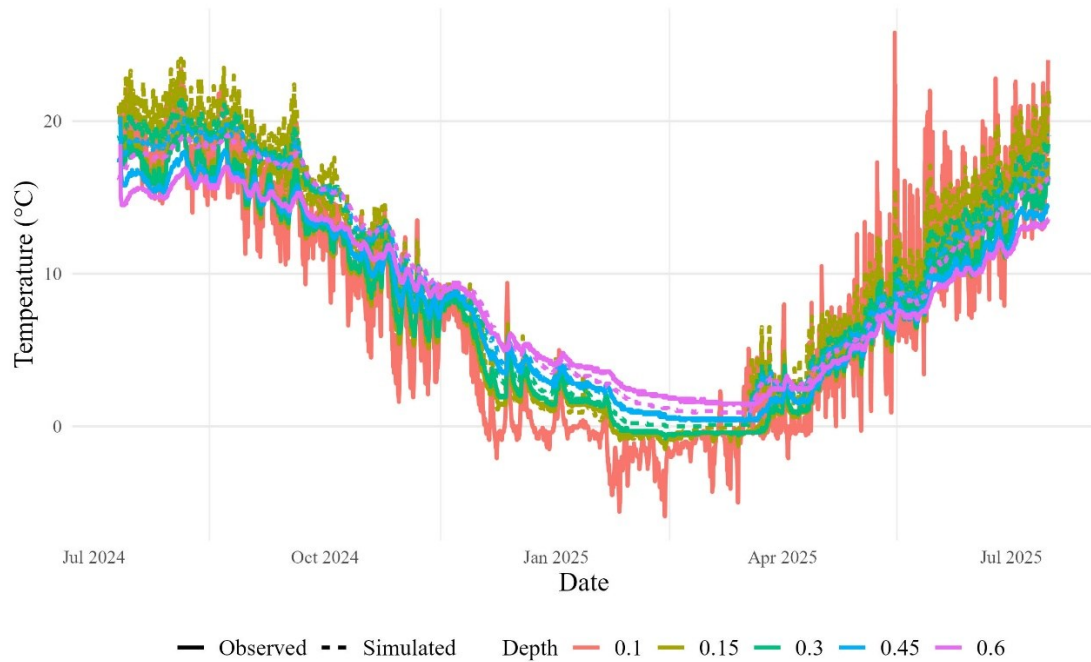


Figure 20: Low-Elevation Forest model soil temperature calibration performance.

3.3.2 Soil Liquid Water Content Simulation Performance

In the calibration period the soil LWC performed within acceptable criteria for two out of the three sites (Table 3). The Wetland site soil LWC failed to meet acceptable performance for most observation depths (RMSE between 0.038 to 0.165 m³/m³; Appendix B, Table 24) and on average (RMSE of 0.124 m³/m³; Table 3). Observed soil LWC in the upper depths of the Wetland site were typically quite dry whereas the lower depths were almost always at saturation, creating a challenging calibration target (Figure 21). The High-Elevation Forest model had an average RMSE of 0.037 m³/m³ which was within acceptable limits (Table 3). The individual node depths in the High-Elevation model were also all acceptable (RMSE between 0.017 to 0.057 m³/m³; Appendix B, Table 25). In the High-Elevation Forest model the soil LWC for the upper two SMT probe depths tended to be underestimated, whereas LWC was overestimated primarily for the lower two depths (Figure 22). The Low-Elevation Forest model had an average RMSE of 0.037 m³/m³ and individually all soil node RMSE were within acceptable calibration limits (RMSE between 0.020 to 0.059 m³/m³; Appendix B, Table 26). The Low-Elevation Forest model's uppermost soil node LWC performed the worst, with it tending to underestimate LWC in the warmer seasons, and overestimate during the winter (Figure 23). Observed infiltration events for the lowest soil node are reflected in the simulations but are highly dampened in comparison to the peaks in the LWC observations.

Overall, the validation performance was generally improved for two out of the three sites in comparison to the calibration period. The validation performance improved the RMSE for some nodes depths in the models but decreased for others. The Wetland site soil LWC RMSE average was 0.114 m³/m³ in the validation period which was a slight improvement from the calibration period (Table 3). The performance was better in the near-surface nodes but worse in the lower depths (RMSE between 0.046 to 0.165 m³/m³; Appendix B, Table 24). The High-Elevation Forest Model had primarily reduced RMSE over the entire soil column in the validation period from the calibration period, except for the uppermost node which increased (RMSE between 0.006 to 0.085 m³/m³; Appendix B, Table 25). On average, the RMSE was 0.033 m³/m³ in the validation period for the High-Elevation Forest model which was lower than the calibration period (Table

3). The Low-Elevation Forest model had increased RMSE over the validation period on average ($0.053 \text{ m}^3/\text{m}^3$; Table 3) and across all soil nodes, but all soil nodes remained within the acceptable calibration criteria except the topmost node (RMSE between 0.030 to $0.105 \text{ m}^3/\text{m}^3$; Appendix B, Table 26)

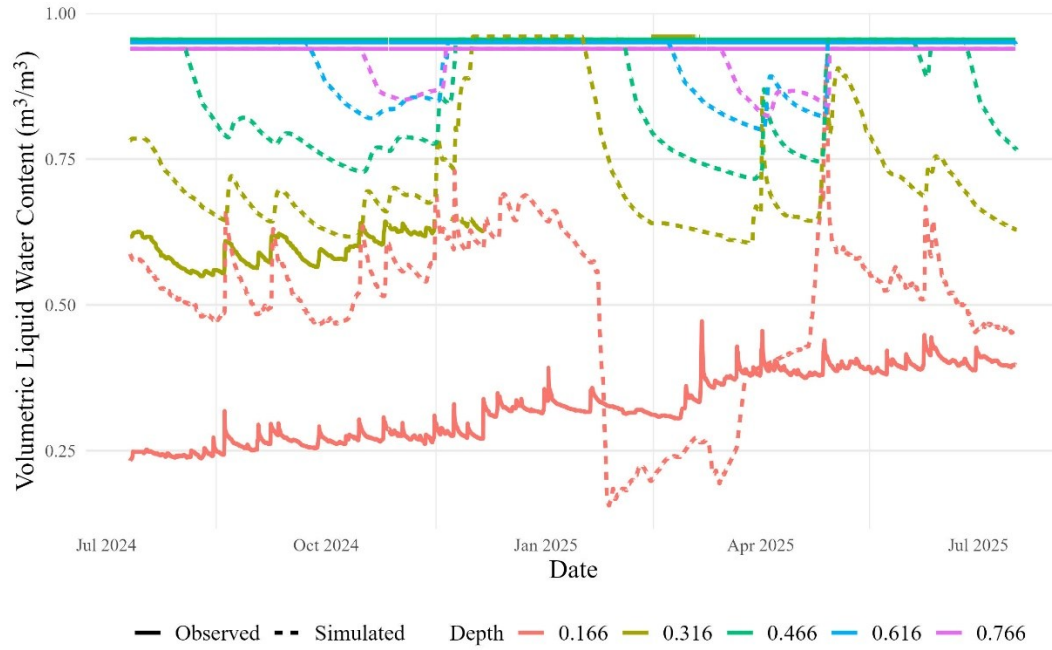


Figure 21: Wetland model soil LWC calibration performance.

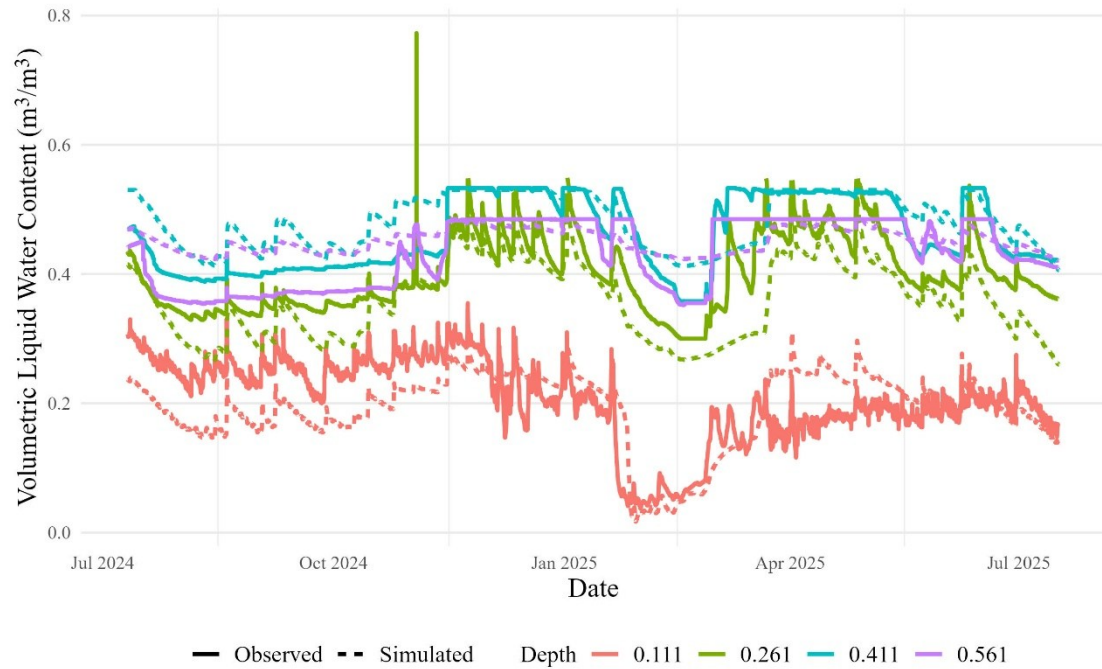


Figure 22: High-Elevation Forest model soil LWC calibration performance.

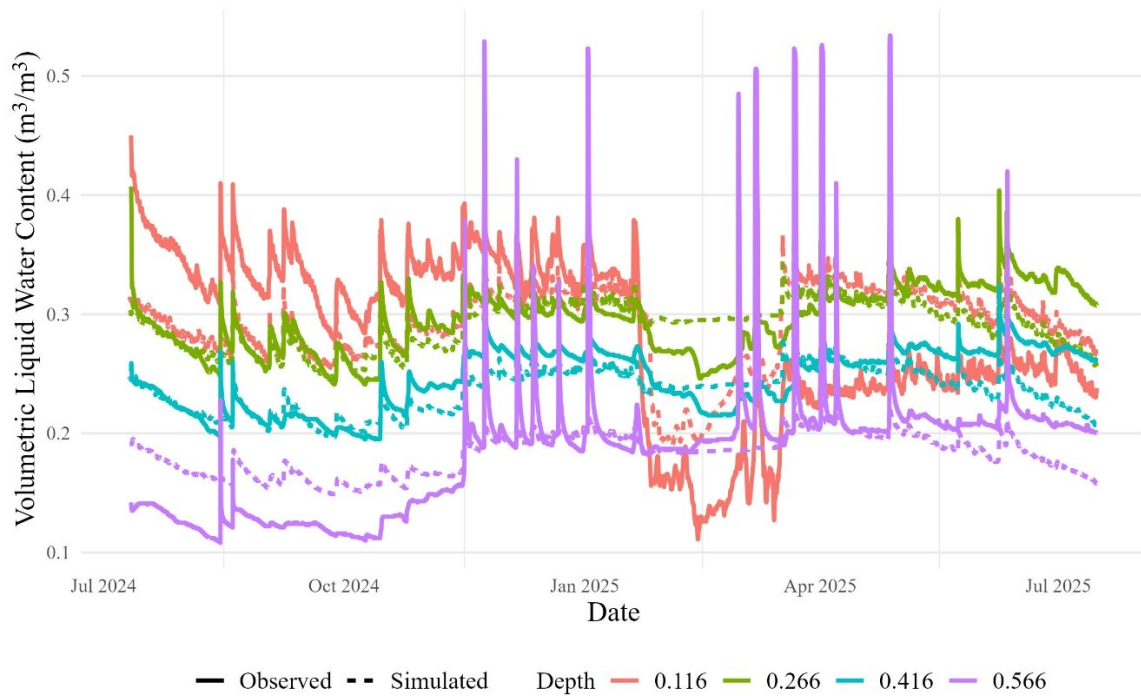


Figure 23: Low-Elevation Forest model soil LWC calibration performance.

3.4 HOURLY GRAPHICAL & SEASONAL RECHARGE PROCESSES ANALYSIS

The hourly graphical recharge analysis results clarify how hydrological processes differ between sites and are changing under climate change while the seasonal plots summarize the overall impacts on seasonal recharge. We therefore present these results concurrently given their inter-related nature.

3.4.1 Historical Snow Cover and Soil Frost

Over the historical period, all sub-sites exhibited a relationship between snow depth and frost depth intra-annually (Figure 24, Figure 25, & Figure 26). In the years with shallower snowpack or colder air temperatures, such as the Low-Elevation Forest Site in 2022, frost depth tended to be deeper for all sites (Figure 26). Likewise, sites with the shallowest snowpack tended to have deepest frost depth and longest frost duration overall.

At the higher elevation sites, more precipitation fell and accumulated as snow than the Low-Elevation Forest site, and snow was able to persist longer in the season (Figure 27). Across the 10-year historical simulation, the two high elevation sites had the deepest snowpacks, but the High-Elevation Forest was substantially higher than the Wetland, (median = 42 cm and 24 cm, respectively, during the SS; Figure 27). The Low-Elevation Forest site had the lowest snowpack thickness (median = 7 cm) and had much less interannual variability than the other two sites (Figure 27). The number of days with snow cover was also highest in the two high-elevation sites (121, 130, and 86 days for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites, respectively) but they had less interannual variability than the Low-Elevation Forest site.

The historical median frost depth and number of days with frozen ground differed substantially between the three sites (Figure 27). The Wetland and Low-Elevation Forest sites had the deepest and most interannually variable average frost depths historically, with median values of 4.4 cm and 4.6 cm, respectively, whereas the High-Elevation Forest site was shallower at 1.7 cm. The median number of days with frozen ground was much higher in the Wetland (146 days), in comparison to the High- (73 days) and Low-Elevation Forest (85 days) sites but exhibited similar interannual spread.

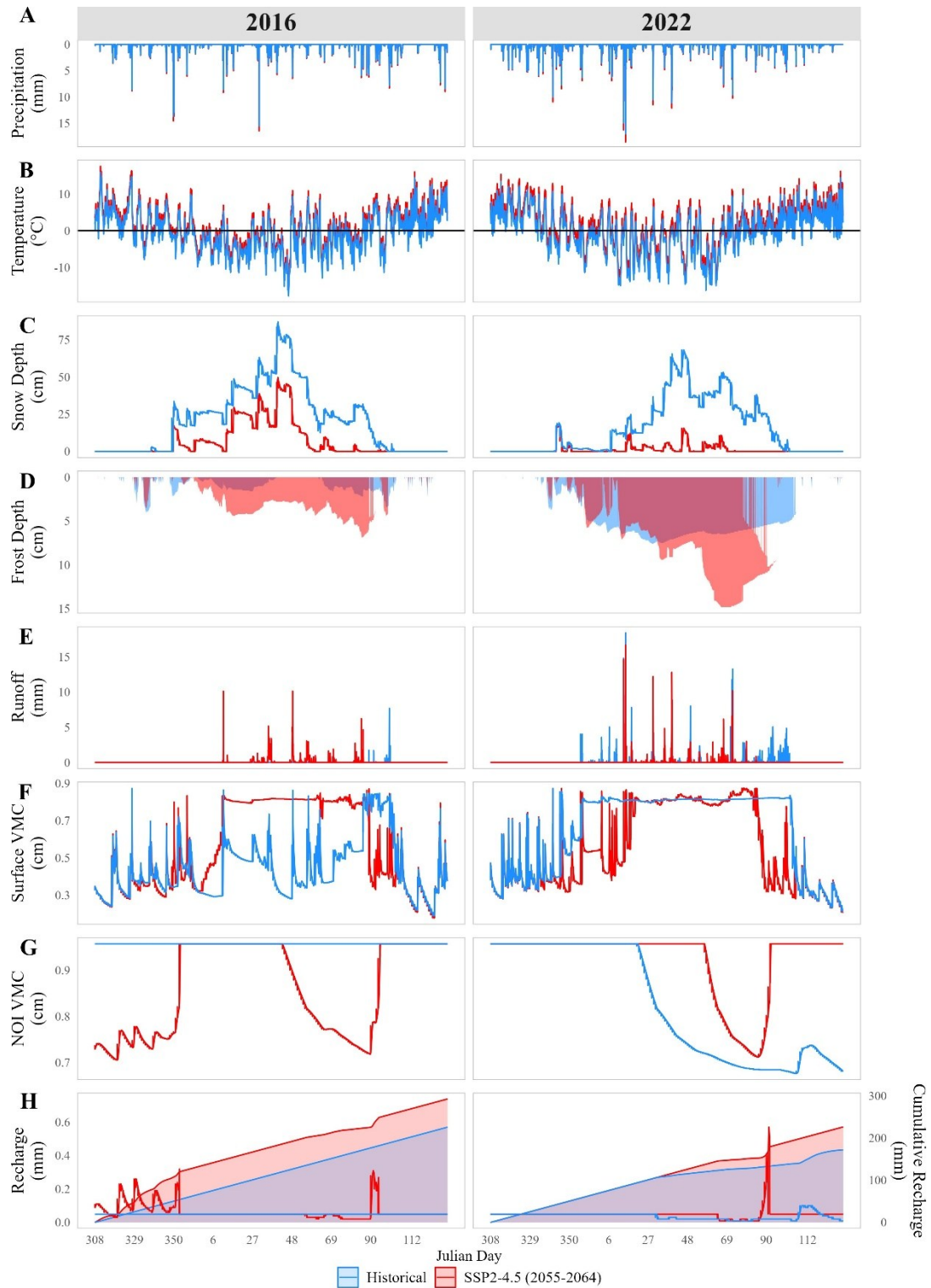


Figure 24: Wetland site hourly (A) precipitation, (B) air temperature, (C) snow depth, (D) frost depth, (E) runoff, (F) VMC₀, (G) VMC_{NOI}, and (H) recharge (line) and cumulative recharge (line and area) in Historical and SSP2-4.5 2055 to 2064 scenarios for 2016 and 2022 base years.

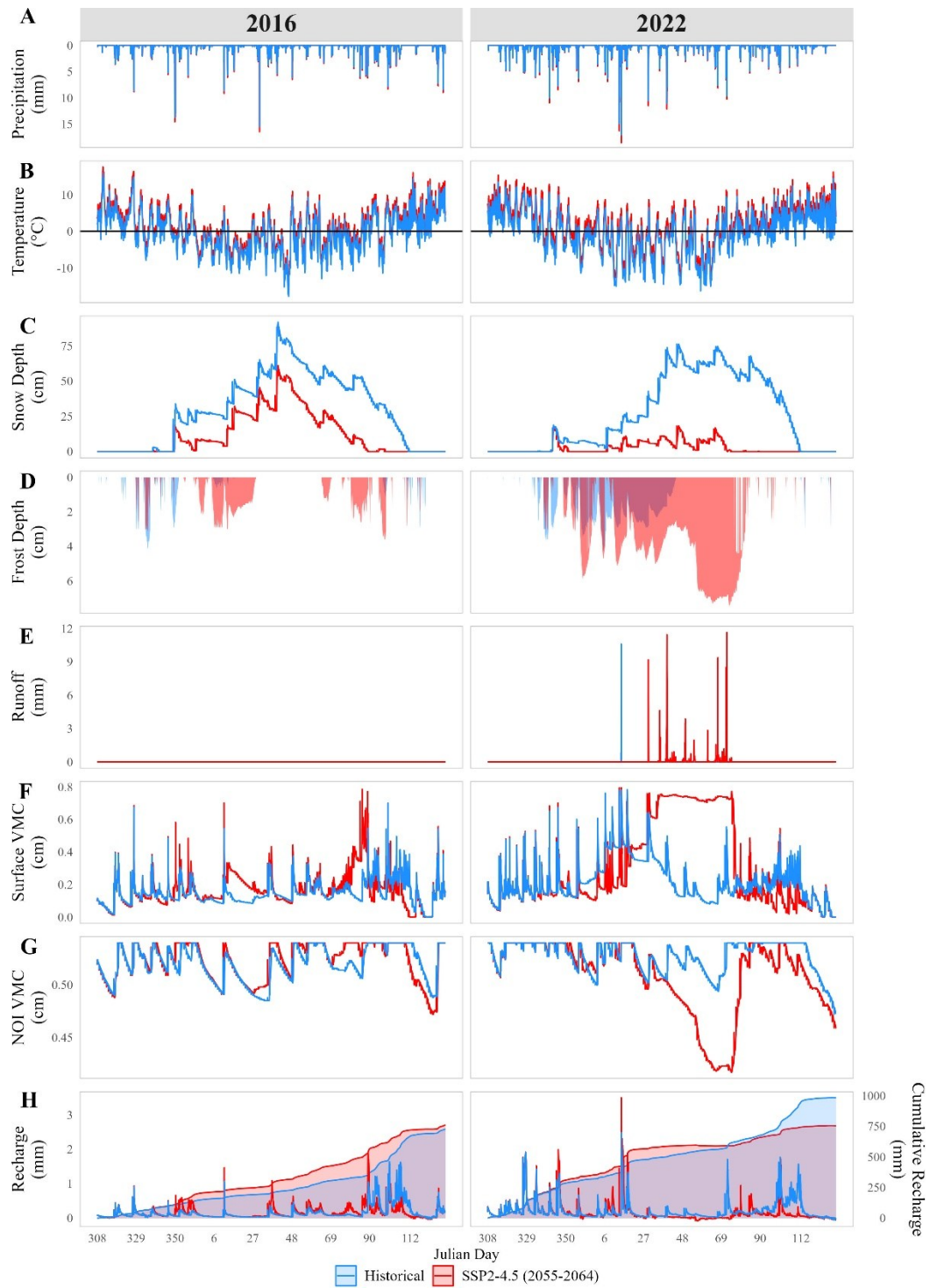


Figure 25: High-Elevation Forest site hourly (A) precipitation, (B) air temperature, (C) snow Depth, (D) frost depth, (E) runoff, (F) VMC₀, (G) VMC_{Noi}, and (H) recharge (line) and cumulative recharge (line and area) in Historical and SSP2-4.5 2055 to 2064 scenarios for 2016 and 2022 base years.

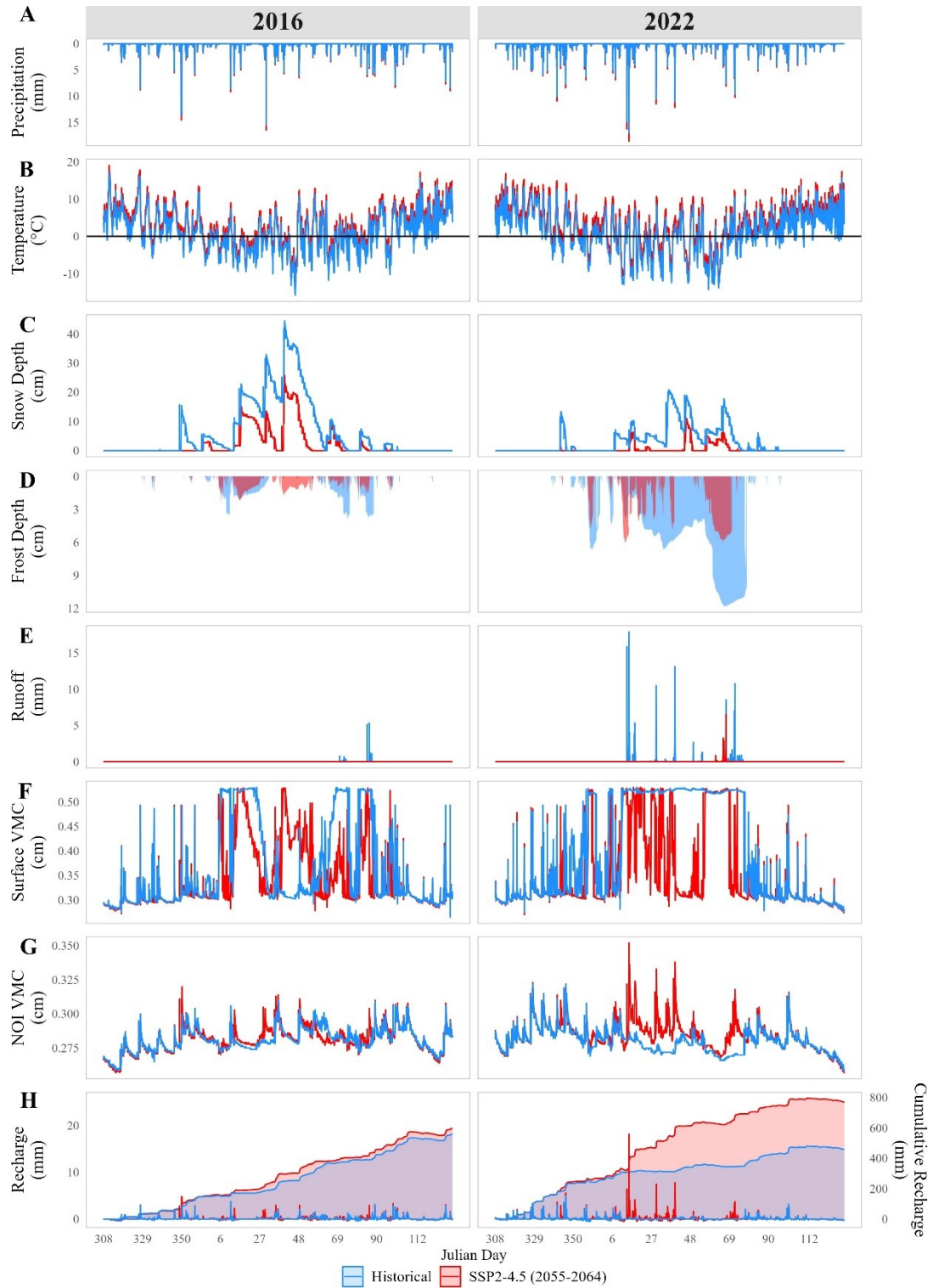


Figure 26: Low-Elevation Forest site hourly (A) precipitation, (B) air temperature, (C) snow depth, (D) frost depth, (E) runoff, (F) VMC₀, (G) VMC_{NOI}, and (H) recharge (line) and cumulative recharge (line and area) in Historical and SSP2-4.5 2055 to 2064 scenarios for 2016 and 2022 base years.

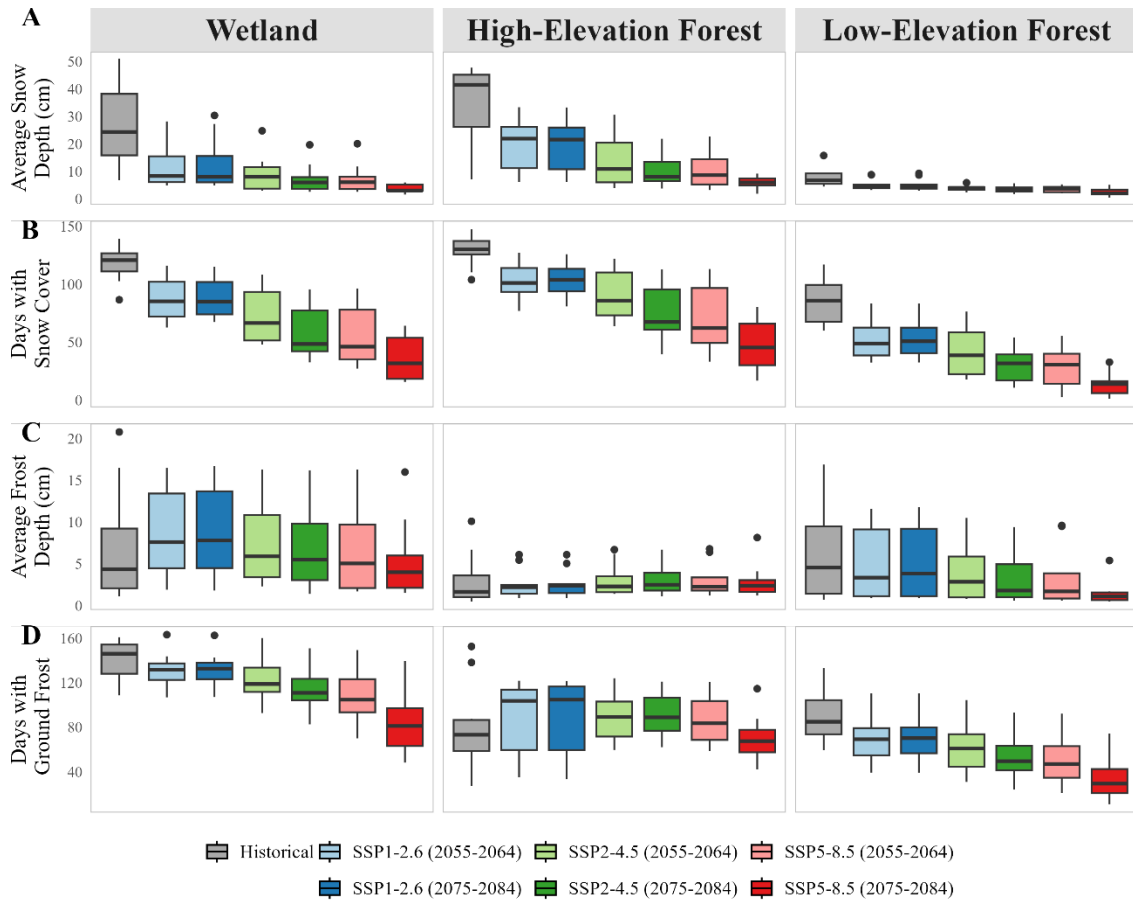


Figure 27: Wetland, High-Elevation, and Low-Elevation annual average snow depth (A), days with snow cover (B), annual average frost depth (C), and days with frozen ground (D) for all climate scenarios.

3.4.2 Historical Soil Volumetric Moisture Content

Under historical conditions, the VMC in the soil was largely different between sites and often exhibited both high inter- and intra-annual variability. The VMC at the recharge NOI (VMC_{NOI} , ~ 0.36 mbgs) for the Wetland site was saturated or near saturated for the majority of the SS but was typically unsaturated at the start of the SS or during deep frost events. The Low-Elevation Forest site VMC_{NOI} was almost always unsaturated, with saturation only occurring with the passing wetting fronts. In contrast, the High-Elevation site VMC_{NOI} was a mixture of saturated and unsaturated conditions, with the SFS typically remaining unsaturated, and the SS fluctuating between the two states. All three sites had peaks in VMC_{NOI} which corresponded with peaks in rainfall or the timing of snowmelt, with some lag (between 2 to 24 hours). Additionally, all sites had

drier soil conditions in the SFS than in the SS. Both of these patterns were also seen in the soil LWC observations.

For all sites during the SS, when frost depths were deep, the VMC_{NOI} decreased to a lower saturation state until the full or partial thawing of soil, likely due water migration due to cryosuction (e.g. Figure 25 in 2022). This VMC_{NOI} drawdown was most pronounced and most frequently simulated in the Wetland site, followed by the High-Elevation Forest site, whereas it infrequently occurred in the Low-Elevation Forest site. In contrast, the VMC at the soil surface (VMC_0) increased during frost formation and the water flux into the soil surface was low until the timing of soil frost thaw. A similar phenomenon was observed in the soil LWC observations from the SMT probes; During the simulated period of deep frost presence, the LWC observations tend to decline across all depths. However, we did not observe the same increase in LWC near the soil surface as seen in simulations.

3.4.3 Historical Recharge and Recharge Efficiency

The distinct seasonality of R was observed across all sites. Historically, the simulated R_{SS} and RE_{SS} were higher than that of the SFS, except for the Wetland site, in which the RE was similar (Figure 28). Higher R and RE in the SS than the SFS were driven by higher winter precipitation and lower ET in the SS. For both seasons, the Low-Elevation Forest had the highest ET, followed by the High-Elevation Forest site, with the lowest in the Wetland site.

Intra-annually the two forested sites followed a similar temporal pattern in R whereas the Wetland site differed considerably (Figure 29). R in the High-Elevation Forest site was generally high before the start of snow accumulation and frost formation, then decreased during the period with snow-cover and frost presence. Following snow melt, the R in the High-Elevation Forest increased. The resulting cumulative R graph for the SS followed an S-Shaped curve. This R pattern was also evident in the Low-Elevation Forest site but was a flatter cumulative recharge curve (Figure 29) as the site experienced more mid-winter R events aligning with the timing of snowmelt events. The Wetland site did not follow this cumulative R graph shape as the NOI was often saturated/below the

groundwater table which meant the recharge was limited by the model's lower boundary drainage rate. Additionally, the Wetland site often had surface frost for much of the SS which restricted the infiltration of mid-winter snowmelt or rain events. As a result, the Wetland site experienced more runoff, and the cumulative R graph followed a more linear path during the SS with step-wise jumps from large R events occurring when the soil node was unsaturated and the ground was thawed. Between the three sites the total SS runoff in the Wetland was the highest, followed by the Low-Elevation Forest site, and the High-Elevation Forest site, which aligns with the pattern of frost duration across the three sites. On a seasonal scale, R_{SS} and RE_{SS} were highest in the High-Elevation Forest site (median = 720 mm and 86%), followed by the Low-Elevation Forest site (median = 436 mm and 53%), and the Wetland (median = 224 mm and 27%), respectively.

In the SFS, the cumulative recharge typically increased throughout the season for the two high-elevation sites but decreased for the Low-Elevation site (Figure 29). On a seasonal scale, the R_{SFS} and RE_{SFS} were highest in the High-Elevation Forest site (median = 186 mm and 31%, respectively), followed closely by the Wetland (median = 179 mm and 27%), and were much lower in the Low-Elevation Forest site (median = -3 mm and -0.7%). The Low-Elevation Forest site tended to have a decreasing cumulative R response in the SFS (and negative R_{SFS} and RE_{SFS}) as the cumulative water lost via capillary action (i.e. ET) outweighed the cumulative precipitation (in the SFS, the runoff was negligible at all sites). The main drivers of these differences were the lower seasonal precipitation and the higher ET demand due to the warmer air temperatures, lower shortwave radiation canopy interception, and lower relative humidity at the lower elevations of the watershed. In some years with higher frequency rainfall events in the late SFS, the cumulative SFS recharge increased for the Low-Elevation Forest site. In contrast, both the high-elevation sites had cumulative P that outweighed the cumulative ET, which resulted in positive cumulative R_{SFS} . In the Wetland site hourly R in the SFS was typically higher than R during the SS as the water table was lower, thus not limited by the lower boundary drainage rate.

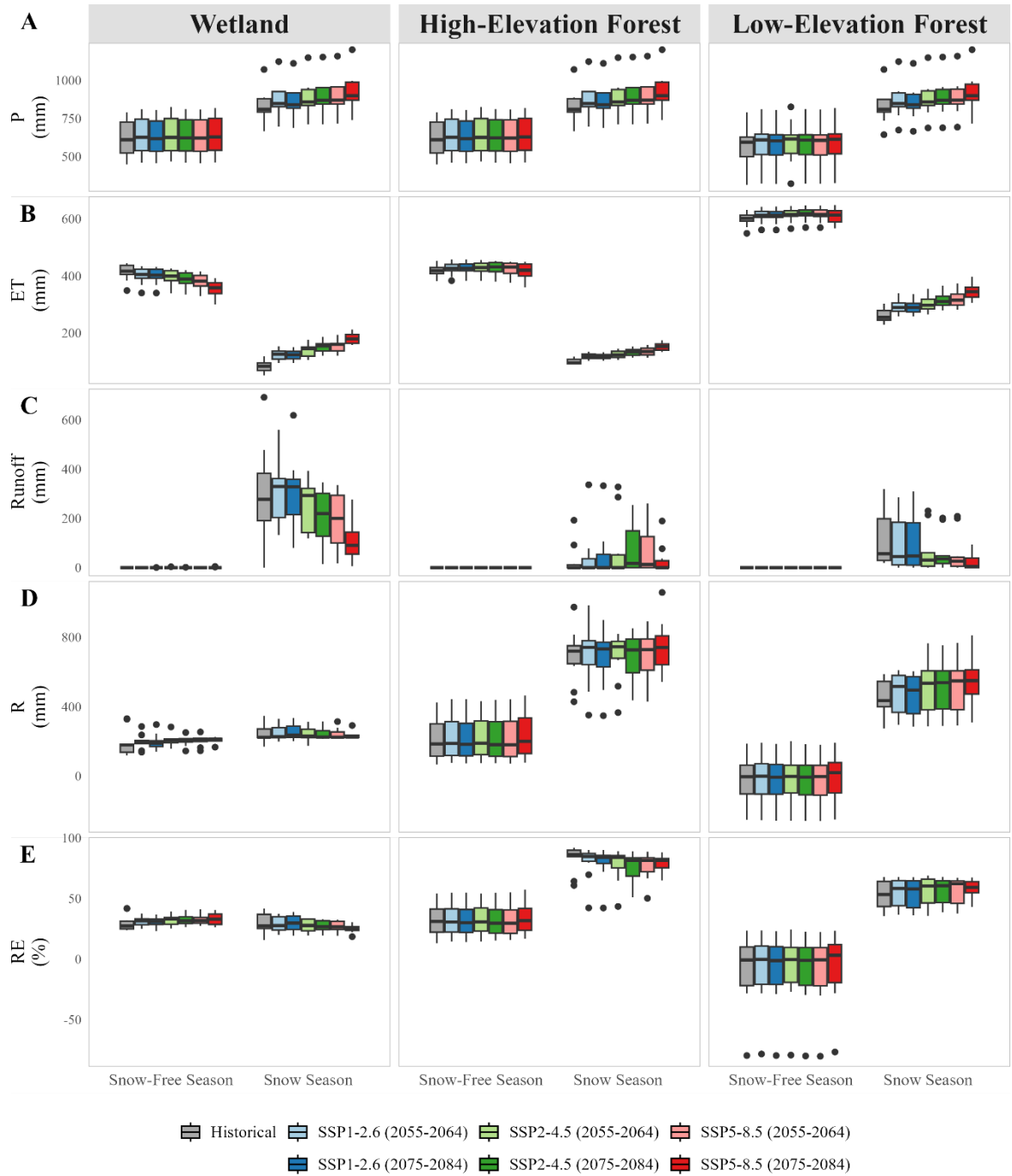


Figure 28: Wetland, High-Elevation, and Low-Elevation seasonal precipitation (A), evapotranspiration (B), runoff (C), recharge (D), and recharge efficiency (E) for all climate scenarios.

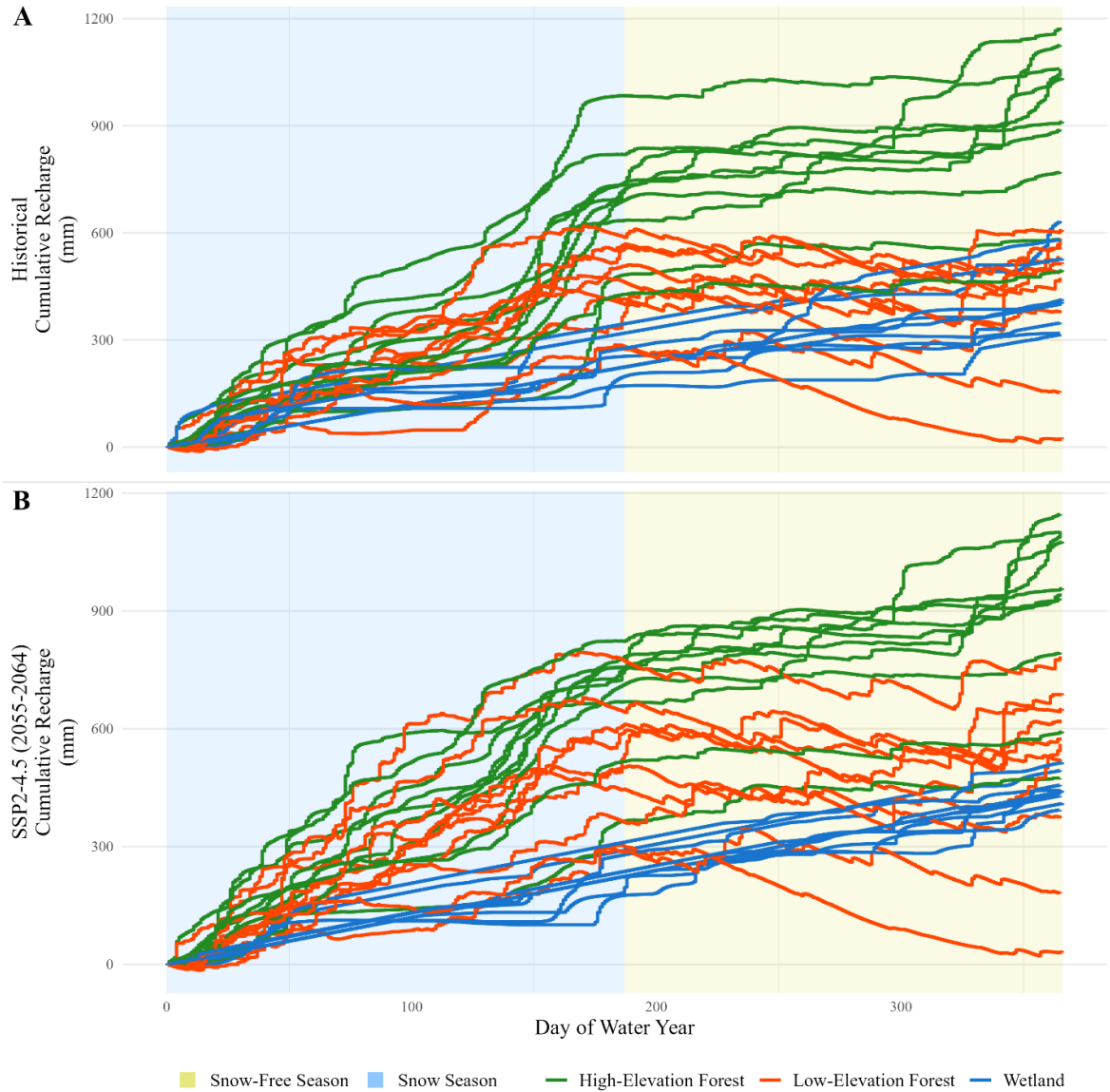


Figure 29: (A) Historical and (B) SSP2-4.5 2055 to 2064 cumulative recharge for each year in the Wetland, High-Elevation Forest, and Low-Elevation Forest sites.

3.4.4 Snow Cover and Soil Frost under Climate Change

The projected reduction in snow presence was consistent across sites. Snow depth and duration were much lower across all sites and years, and there were more mid-winter melt events (Figure 27). The average snow depth and the number of days with snow cover decreased markedly with increasing climate change intensification across all sites (Figure 27). The high-elevation sites had the steepest declines in average snow depth under a warming climate. The median annual snow depth change across all scenarios was

-72%, -63%, and -47%, for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites, respectively. The median annual snow cover duration change across all scenarios was -41%, -30%, and -54% for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites, respectively.

Unlike changes to the snowpack, changes in frost presence under climate change were different interannually, intra-annually, across the three sites, and between different climate change scenarios. The different responses were deemed due to a temperature inflection point at which increased air temperatures go from promoting more soil frost to less.

Interannually, the change in frost depth was more variable than snow depth but was related to the change in thermal insulation provided by the snowpack. In the historical simulations, the years with less snow typically had deeper frost depths during the coldest part of the SS and vice versa. For these same low-snow base years, the snowpack was even shallower in a warmer climate, and frost depth decreased due to the warmer air temperature. In contrast, in base years with historically deep snowpacks, the snowpack was again substantially reduced but the frost depth increased during the coolest part of the SS despite warmer air temperatures. Although, for these deep snowpack years, frost depth was still shallower in comparison to years with historically deep frost depths. This relationship was due to the tighter coupling of soil to air temperatures from decreased thermal insulation. Thus, the counterbalancing effects of air temperature warming and snow depth reduction under climate change suggest an inflection point in which increased air temperatures go from promoting more soil frost to less.

The air temperature inflection point also resulted in intra-annual differences in frost presence changes under climate change. While for some sites during the coldest part of the SS, the frost depth or presence increased, at the start and end of the frost season there was typically less frost presence across all three sites (Figure 30). The shallower snowpack and warmer air temperatures in the shoulders of the frost season, which were near zero historically, reduced frost formation leading into the winter and promoted more rapid frost thaw in the spring. In comparing the historical median start date of the frost season to that of the SSP2-4.5 2055-2064 scenario, the start date was between 10 to 15

days later across the sites. Similarly, at the end of the frost season, the complete thaw of the soil occurred between 6 to 13 days earlier in the spring.

Across the three sites, the impact of climate change on median frost depth and number of days with frozen ground was distinctly different during the coolest period (sub-zero air temperatures) of the SS. The Low-Elevation Forest site typically had reduced frost depth and duration in the winter months of the season. This change was due to the warming air temperatures being above zero more frequently combined with less snow insulation. In contrast, the two-high elevation sites typically had deeper frost depths in the cooler part of the SS as, even with warming air temperatures, air temperature was sub-zero for much of the season and there was less insulation. However, the overall change to frost duration in the SS between the two high elevation sites were different under climate change. Years with increased frost duration in the winter months of the SS were most abundant in the High-Elevation Forest site as it historically was frost-free for much of the SS. Thus, an increase in frost depth resulted in significant increase in the number of days with frozen ground. Whereas the increased frost depth in the Wetland site during the winter months had minimal increase in days with frozen ground as it was typically frozen most of the season under historical conditions.

Lastly, the change in frost depth and duration was also different under the different climate change scenarios. Frost depth decreased in all future climate scenarios for the Low-Elevation Forest site (median relative change of -31%) commensurate with climate change intensity. However, the Wetland and High-Elevation Forest sites experienced non-monotonic responses in average frost depth relative to climate change intensification. The average frost depth for the Wetland site increased for the moderate scenarios, but the most severe scenario (SSP5-8.5 2074 to 2085) returned to near historical levels, resulting in a median change of -0.6%. The High-Elevation Forest site showed an increase in average frost depth with intensification (+34%) but a decrease for the most severe climate scenario. The number of days with frozen ground at the Wetland and Low-elevation Forest sites decreased under all future climate scenarios commensurate with climate change intensity, with median relative changes of -16% and -34%, respectively. Whereas, for the High-elevation forest site, most of the climate

scenarios caused an increase in frozen ground duration (highest in the SSP1-2.6 scenarios), except the most severe case (SSP5-8.5 2074-2085), resulting in a median change of +14% across all scenarios. This non-monotonic response is believed to be due to this air temperature threshold being exceeded in the high intensity climate change scenarios in which warmer air temperature offset snowpack insulation losses.

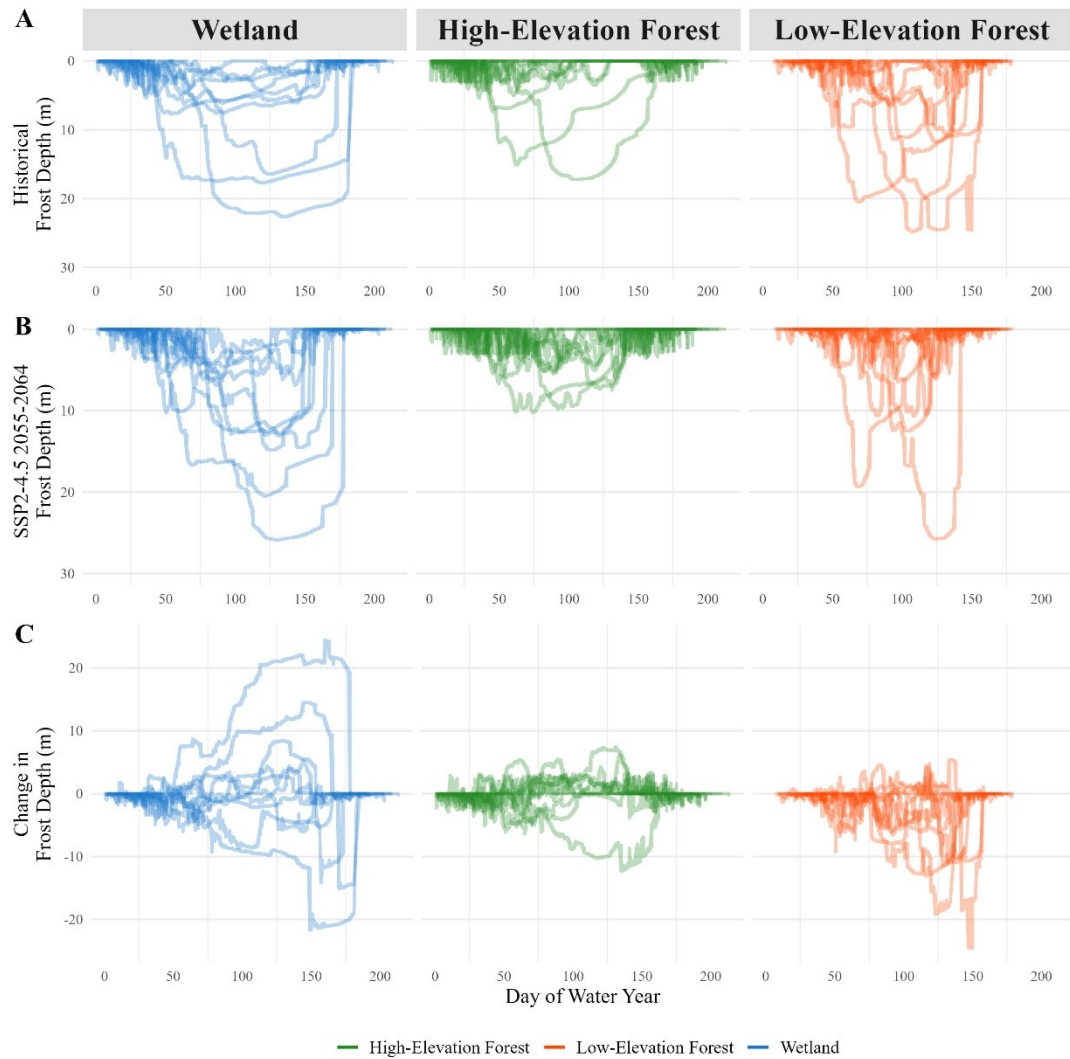


Figure 30: Hourly frost depth for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites for the (A) Historical scenario and (B) SSP2-4.5 2055-2064 scenario, and (C) the frost depth difference between Historical & SSP2-4.5 2055-2064 scenarios (positive = increased from Historical, negative = decreased from Historical).

3.4.5 Projected Recharge and Recharge Efficiency under Climate Change

Under the climate change scenarios, the difference in R from historical conditions was most pronounced in the SS rather than the SFS (Figure 29). In Figure 29, the timing of R during the SS generally shifted earlier in the water year, especially for the Low-Elevation Forest site. Whereas the distribution of R remains similar in the SFS in a warming climate.

In a warming climate, changes to R_{SS} were tied to changes in snow depth and frost depth and the ability for increased precipitation to offset increase ET. ET steadily increased with increasing climate change intensification for all sites during the SS (Figure 28). The Wetland site experienced the highest median seasonal increase in ET of 71%, whereas the Low-Elevation Forest and High-Elevation Forest sites were much lower at 17% and 27%, respectively. Cumulative R shifted earlier in the season for most sites due to the timing of water availability (i.e. snowmelt and rainfall) being more distributed over the season than it was historically. In the Low-Elevation site, due to less frost presence, infiltration capacity improved and there were more mid-season recharge events, as evidenced by the reduced runoff, however this was balanced by increased ET. As a result, the RE_{SS} stayed almost the same as under historical conditions (-0.02%) but the R_{SS} in the Low-Elevation Forest site increased (6.8%) due to higher rainfall amounts. In the two high-elevation sites, there was typically less R_{SS} during periods of increased frost presence, but this was compensated by increased recharge during frost-free periods. Thus, in the High-Elevation Forest, the increased R_{SS} (median change = 3.3%) was due to increased precipitation, but the RE_{SS} decreased (-2.8%) due to the increased frost presence (i.e. increased runoff) and ET. Meanwhile, for the Wetland site, cumulative SS recharge tended to follow a similar pattern as its historical behaviour as the soil NOI was often below the groundwater table. The Wetland site's R_{SS} was limited by the lower boundary drainage rate as the VMC_{NOI} remained saturated for longer of the SS. As a result, this water instead accumulated above the NOI to later be accounted for as R_{SFS} . The RE_{SS} in the Wetland was thereby reduced (-1.9%) but less so than the High-Elevation Forest, likely owing to the Wetlands decreased frost presence duration and, in turn, lower seasonal runoff generation.

Under climate change, the SFS cumulative R tended to follow a similar annual pattern to the historical scenario for all of the sites, but the Low-Elevation Forest site exhibited more interannual variability (Figure 29). The increase in ET due to warming was offset by the increased P under climate change, thus the shape of the cumulative SFS recharge was generally retained. However, due to changes in the R_{SS} , the cumulative R_{SFS} was typically shifted upwards overall for this site. Infrequently, some years exhibited slightly lower R rates than historically, these were typically in years with lower-than-average precipitation and higher air temperatures. During these years, the soil was especially dry due to limited and infrequent precipitation, thus more water was partitioned to ET as it was held in the upper soil layers through capillary forces. This most frequently occurred in the Low-Elevation site and resulted in much more interannual variability in cumulative R_{SFS} in a warming climate.

R_{SFS} and RE_{SFS} due to climate change intensification generally increased for the Wetland site but stayed relatively constant for the forest sites. Across all sites, changes in R_{SFS} and RE_{SFS} were governed by the balance between increased P and changes to ET (runoff was negligible). The R_{SFS} in the Wetland site generally increased more for the more intense climate change scenarios (median change = 12.3%) but stayed relatively constant across scenarios for the High-Elevation Forest and Low-Elevation Forest sites (median change = 2.3% and 1.5%, respectively). In the SFS, there was a median increase in ET of around 3% for the two forest sites, whereas the Wetland site had a median decrease of 5%. In the two forest sites, the increase in ET was compensated by the increased P and resulted in slightly higher R_{SFS} but almost no change in RE_{SFS} . The RE_{SFS} under climate change, the Low-Elevation Forest and High-Elevation Forest sites had similar RE_{SFS} (+0.02% and -0.6%, respectively), whereas it increased for the Wetland site (+3.3%). The Wetland site exhibited the largest increase in R and RE during the SFS as the precipitation increased and the ET decreased from historical rates. The reduced ET was due to the Wetland site having a relatively high permeability; Thus, the higher rainfall amounts for rainfall events (i.e. higher intensity) were able to infiltrate deeper into the soil and resulted in less available water to be lost to ET.

3.5 HISTORICAL AND FUTURE TRENDS OF ANNUAL RECHARGE AND RECHARGE EFFICIENCY

3.5.1 Historical Annual Recharge and Recharge Efficiency

Historically, the Wetland site had the lowest R_A and RE_A , followed by the Low-Elevation Forest site, with the High-Elevation Forest site having the highest (Figure 31). The median historical R_A for the Wetland, High-Elevation, and Low-Elevation site were 407, 972, and 500 mm/year, respectively, and the median historical RE_A were 28, 64, and 36%. Differences in R_A between sites were primarily driven by differences in R_{SS} , except for the Low-Elevation Forest site which had a much lower R_{SFS} than the other two sites. R_{SS} made up 55%, 79%, and 100% of the R_A for the Wetland, High-Elevation Forest, and Low-Elevation Forest site, respectively (based off of proportions of median seasonal recharge).

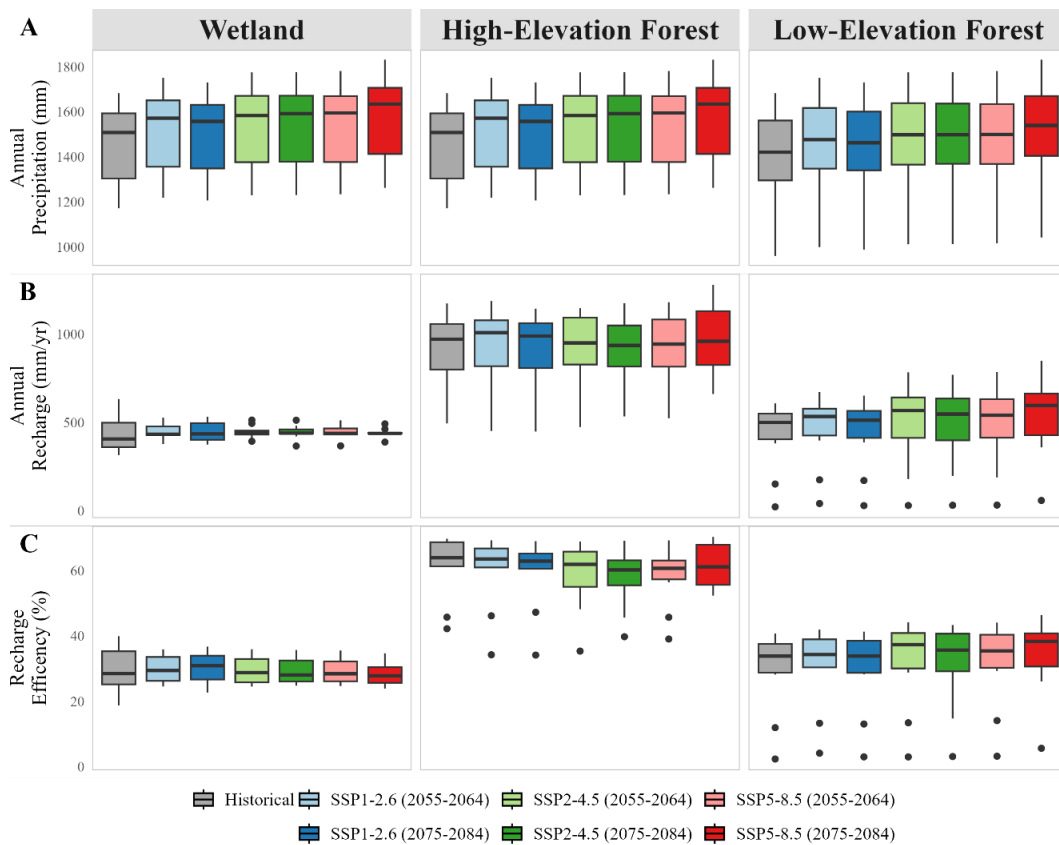


Figure 31: Wetland, High-Elevation, and Low-Elevation annual precipitation (A), recharge (B), and recharge efficiency (C) for all climate scenarios.

3.5.2 Projected Annual Recharge and Recharge Efficiency under Climate Change

The differences in R_A and RE_A between landscape types were much greater than the differences between the historical and future climate scenarios. The R_A and RE_A remained similar to historical levels across all climate change scenarios despite large hydrological changes simulated across sites (Figure 31). The R_A and RE_A remained minimally changed in a warmer climate because the SFS remained stable as increased P offset increases to ET (between 6.7% to 7.6% median percent change) and because changes to R_{SS} were minimal as enhanced P offset both increased ET and any reduction to infiltration capacity. For the Wetland and High-Elevation Forest sites, the median change in P_A was 5.2% from the historical period, and 5.3% for the Low-Elevation Forest site. From the historical period to the future climate scenarios, the median changes in R_A were 6.6, 3.0, and 8.4% and RE_A s were 0.4, -0.9, and 0.7% for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites, respectively. For the Low-Elevation Forest and Wetland site, this increase in R_A and RE_A was enhanced as the intensity of climate change increased. The maximum R_A increase for the Low-Elevation Forest and Wetland across all scenarios was 375 mm (79% increase) and 127 mm (40% increase), and the RE_A increase was 18.2% and 8.6%, respectively. Additionally, the interannual variability of R_A and RE_A for both these sites increased with climate change intensification. In the High-Elevation Forest site, the changes to R_A and RE_A had a less prominent relationship with climate change intensification.

3.6 UNCERTAINTY & SENSITIVITY ANALYSIS RESULTS

3.6.1 Sensitivity Analysis

The results of the local sensitivity analysis showed that the models were commonly sensitive to the SWRC parameters and canopy LAI across all outputs and in both the historical and future climate scenarios (Appendix B, Table 28 to Table 33). In the Wetland site, historically the parameters with the highest SI were decreased hydraulic conductivity in the bottom of the model, increased or decreased LAI, and increased SWRC parameter b (pore size distribution parameter) value at 2 m depth. In the climate change scenario, the parameters with the highest SIs remained similar to the historical scenario. In the High-Elevation Forest, the outputs were historically most commonly sensitive to a decreased

LAI, a decrease in the air entry potential at 0.41 mbgs, a decrease in the plant interception, and an increase in the bottom of the model's hydraulic conductivity. The average snow and forest depth were highly sensitive to a decreased air entry potential at 0.41 mbgs. Under climate change, the sensitivity of the model overall decreased, the most sensitive parameters remained similar except the model was more sensitive to a decreased b value and less to decreased hydraulic conductivity in the bottom of the model. For the Low-Elevation Forest, historically all simulation outputs were most sensitive to an increase in the hydraulic conductivity at the surface or at 0.97 mbgs, and an increase in the porosity at the 6.25 m depth. Under climate change, this sensitivity remained similar for most outputs.

3.6.2 Uncertainty Analysis

The model's uncertainty was relatively high for a number of the analytical outputs investigated in our study which suggests the findings should be interpreted with caution (Table 4 & Table 5). The uncertainty analysis of the historical climate scenarios found the R_A and RE_A had a similar percent uncertainty within a given model (Table 4). The % UC for the R_A and the RE_A for the Wetland site was the highest (42%), followed by the Low-Elevation Forest (29% and 28%, respectively), and lastly the High-Elevation Forest (15%). The % UC was also distinct between the three sites for the median snow depth. Historically, the Low-Elevation Forest had the tightest UC bounds (range of 14 cm) but the highest % UC (100%). The Wetland site also had a relatively high % UC for median annual snow depth (43%). Under climate change, both sites' snow depth had decreased % UC. In contrast, the High-Elevation Forest site had the lowest % UC historically (22%) but under climate change, the High-Elevation Forest site's snow depth becomes significantly more uncertain (124%). The days with snow cover historically had low uncertainty for the two high elevation sites (<8%) and moderate uncertainty in the Low-Elevation Forest site (25%). Under climate change, this shifted to higher % UC across all sites (21% to 38%). Frost depth was highly uncertain historically across all sites (41% to 83%). In the Wetland site, the % UC of frost depth decreased in the climate change scenario (53% or -30% change) whereas is increased in the High-Elevation Forest site (72% or +31; Low-Elevation Site had 1% increase). Lastly, the number of days with ground frost had high uncertainty in the High-Elevation Forest site historically (42%),

whereas the other two sites were low-moderate (6% to 25%). In a warmer climate, the % UC of days with frozen ground remained similar to the historical for the Wetland and Low-Elevation Forest site (10% to 25%) but decreased largely for the High-Elevation Forest (26%).

The Wetland site's base model outputs for both climate scenarios tended to be on the low-end of the UC bounds and was often below the 5th percentile UC bound. The distribution of the uncertainty analysis model outputs for the Wetland site tended to be multi-modal, potentially suggesting a rugged model solution space and a high-degree of uncertainty given the input parameter constraints (Appendix B - Figure 41 & Figure 42).

Table 4: Uncertainty Analysis Results for Median Values in the Historical Period.

Results	Wetland				High-Elevation Forest				Low-Elevation Forest			
	Base	5th	95th	% UC	Base	5th	95th	% UC	Base	5th	95th	% UC
Median												
R_A	407	414	753	42	972	803	1097	15	500	478	769	29
(mm/yr)												
Median												
RE_A (%)	28	27	50	42	64	52	71	15	36	34	53	28
Median												
Snow												
Depth	24	23	44	43	42	35	54	22	7	5	19	100
(cm)												
Median												
Days												
with	121	123	133	4	130	124	144	8	86	72	115	25
Snow												
cover												
Median												
Frost												
Depth	4.4	3.4	10.7	83	1.7	0.5	1.9	41	4.6	1.4	8.2	74
(cm)												
Median												
Days												
with	146	141	159	6	73	44	105	42	85	74	116	25
Ground												
Frost												

Table 5: Uncertainty Analysis Results for Median Values in the SSP2-4.5 2055-2064 Period.

Results	Wetland				High-Elevation Forest				Low-Elevation Forest			
	Base	5 th	95 th	% UC	Base	5 th	95 th	% UC	Base	5 th	95 th	% UC
Median												
R_A	441	494	730	27	951	768	1090	17	568	566	826	23
(mm/yr)												
Median												
RE_A (%)	29	32	47	26	62	49	68	16	37	38	54	22
Median												
Snow												
Depth	8	5	11	38	11	8	36	124	4	3	6	36
(cm)												
Median												
Days with												
Snow	67	71	99	21	86	79	121	24	39	32	62	38
cover												
Median												
Frost												
Depth	5.9	6.0	12.3	53	2.3	1.1	4.5	72	2.9	1.2	5.5	73
(cm)												
Median												
Days with												
Ground	119	114	138	10	89	67	113	26	61	46	77	25
Frost												

CHAPTER 4 - DISCUSSION

4.1 HISTORICAL SNOWMELT RUNOFF TIMING TRENDS

The CT method found statistically significant slopes for six of the sixteen stations, which indicates earlier snowmelt timing between 0.9 to 3.1 days earlier per decade. Several other studies in North America have found a trend towards earlier snowmelt timing, although they generally find a faster rate of change than in Cape Breton (Dudley et al., 2017; Stewart et al., 2004); Dudley et al. (2017) found earlier snowmelt runoff across the contiguous United States, with a median slope of around 8.2 to 8.6 days per decade in the eastern U.S. from observation periods between 1940 to 2014 and 1960 to 2014, respectively. In the western U.S., they found fewer significant streamflow trends, but these trends also tended to indicate earlier snowmelt timing. Basins with elevations less than 1600 m had median trends of 6.8 and 3.4 days per decade, whereas basins above this elevation had median trends of 4.0 to 5.2 days per decade for high-elevation basins for the 1940 to 2015 and 1960 to 2014 periods, respectively (Dudley et al., 2017). Likewise, Stewart et al. (2004) analyzed snowmelt runoff timing for 279 snowmelt-dominated streamflow stations from 1948 to 2000 in western North America and found snowmelt typically occurred around 1.9 to 5.8 days earlier per decade.

There is minimal research investigating the trends in peak streamflow timing in Atlantic Canada, but of the few existing studies, they too support the trend of earlier melt-out. Zhang et al. (2001) used the same CT method across Canadian streams from 1947 to 1996 and found Atlantic Canada had negative trends in CT timing. Goguen & Caissie (2022) studied the peak discharge day for the Northeast Margaree stream from 1929 to 2018 and found a slope of -2.2 days per decade for the high flow date which is roughly double our change in CT date for the station of -1.1 days/decade.

As expected, the R^2 was small for all the linear regressions of CT date against year, indicating variance in the CT date is weakly explained by the year. Snowmelt timing is primarily governed by air temperature and winter precipitation which have high interannual variability that would not be explained solely by climate change trends. For instance, the El Nino and the Southern Oscillation (ENSO) and Pacific Decadal

Oscillation (PDO) have been found to have a statistically significant correlation with streamflow timing in the U.S. (Stewart et al., 2005).

Snowmelt runoff timing is a function of latitude as well as elevation (Stewart et al., 2004), which may explain the difference in the trend magnitudes exhibited across these studies and within our own study. Additionally, trends may be different due to the observation periods used in each analysis. For instance, within our study, statistically significant trends tended to be observed in the longer time series records, and the longest records typically had shallower slopes whereas the more recent shorter records tended to have steeper slopes. This pattern could suggest the impacts of climate change on streamflow are accelerating.

Earlier snowmelt has several hydrological implications, like the lengthening of the baseflow recession period and the resulting intensification of summer hydrologic droughts (Huntington & Niswonger, 2012). Climate change is also expected to result in less snow accumulating and slower melt rates which may further exacerbate drought effects (Barnhart et al., 2020). Barnhart et al. (2020) found the earlier timing of snowmelt may offset these losses partially due to the lower vegetation water use at this time, but this effect may be subject to the site conditions like plant available water storage and energy availability.

4.2 TOPOGRAPHIC AND LANDCOVER EFFECTS ON HISTORICAL RECHARGE RATE AND EFFICIENCY

Our study found substantial differences in R_A between landscape types which were driven mostly by factors impacting SS processes. Our historical simulations confirm literature findings that in cold regions, R_{SS} makes up a significant proportion of R_A ; Many studies have found that in cold regions, snow contributes more to R_A in comparison to its fraction of total annual precipitation (Jasechko et al., 2014; Earman et al., 2006; Mohammed et al., 2019). Our study does not differentiate the RE between precipitation phases, but we had similar findings to past studies in that, under historical climate conditions, RE in the SS, which is dominated by snow, has a much higher RE than the SFS for two of our three sites (median = 54% to 55% higher).

We found high rates of R_A and RE_A in our study that should be verified through observational estimates of recharge, especially for the High-Elevation Forest which had R_A which largely exceeded the rates found in literature. Our historical annual recharge estimates for the area were much higher (medians between 407 to 972 mm/year) than the 260 to 300 mm/yr recharge estimate for the area from Kennedy et al. (2010). Likewise, studies in Ontario and Quebec typically found the areas with the highest simulated R_A were, at most, between 515 to 567 mm/year (Jyrkama & Sykes, 2007; Boumaiza et al., 2022). RE in our study was also generally higher than RE found globally, which was around 0.1% to 25% (Wada et al., 2010; Jasechko et al., 2014) and in Quebec and Ontario, around 20% to 30% (Jyrkama & Sykes, 2007; Boumaiza et al., 2022). However, our simulated REs may be realistic as Jyrkama and Sykes (2007), who simulated recharge throughout their watershed in Ontario, found average RE_A could be as high as 59% in some areas (567 mm/year of recharge and 950 mm/year of average annual precipitation). Likewise, Bouzmaiza et al. (2022) simulated potential groundwater recharge in a watershed in Quebec and found RE_A could be as high as 37% to 55% (350 to 515 mm/year of potential groundwater recharge and 930 mm/year of precipitation), but this occurred in only 6% of the watershed area.

In evaluating our different site observational and simulated data over the same time period at a high temporal resolution, our study was able to effectively distinguish drivers of R_A in different landscape types. We found that the snowpack thickness, soil characteristics, and water table depth are the key drivers of site differences in frost formation and moisture conditions and, as a result, R. Each of these drivers are discussed in detail in the subsequent sections. These factors are commonly cited controls on recharge in cold regions; However, few studies have evaluated catchment-scale spatial variability of recharge in cold regions at a high temporal resolution (Hyman-Rabeler & Loheide II, 2023; Dripps & Bradbury, 2010; Lundberg et al., 2016; Mahmoud et al., 2025).

R_{SFS} , while less than that of the SS, still made up a significant proportion of the R_A , which aligns with findings for other cold region watersheds (Earman et al., 2006; Mahmoud et al., 2025; Jasechko et al., 2014). Our study found that site differences in

R_{SFS} was driven primarily by warmer air temperatures and lower precipitation between the lower elevation and higher elevation sites. This topographically-driven recharge difference adheres to other literature findings that higher elevation locations tend to have cooler air temperatures and more precipitation on the windward side due to orographic lift (Dobre et al., 2025; Ward & Robinson, 2000).

4.2.1 Snowpack Thickness & Snow Cover Duration

As expected, our study found that higher elevation sites had deeper snowpacks and longer durations of snow cover (Dobre et al., 2025; Ward & Robinson, 2000). We also found that increased canopy cover also resulted in deeper snowpacks and longer snow cover duration when comparing the two high-elevation sites. Precipitation interception and net incoming radiation are known to increase with increased canopy cover; Thus, many studies found that increased canopy cover leads to less snow accumulation and accelerates seasonal melt out timing in comparison to an open area (Dobre et al., 2025; Pomeroy et al., 2002 & 2009; Lundquist et al. 2013). However, these studies refer to softwood dominated forests, which have different characteristics than the mixed forest stands in our watershed. One study by Pastore et al. (2025) suggested that mixed forests may promote deep snowpacks due to their optimal balance between limiting snowfall interception while still providing snowpack sheltering from wind and solar insolation. Our results support this statement, but there is limited research on the impacts of a mixed forest on snow accumulation and melt to confirm these findings.

4.2.2 Soil Frost Presence

Snow cover and air temperature exert control over the formation of soil frost as snow acts as an insulator for the ground (Lundberg et al., 2016). This relationship was reflected in our study where the High-Elevation Forest site had the deepest median snowpack and the shallowest frost depths. Meanwhile, the opposite was true of the Low-Elevation Forest site. However, the relationship between frost depth and snow depth was not directly proportional or linear. For example, the Wetland site had more than triple the median snow depth of the Low-Elevation Forest site but only 4% lower median frost depth. The non-linear nature of the snow insulation effect is well documented in literature (Zhang, 2005; Goodrich, 1982). Zhang (2005) performed a literature review of the impact

of snow depth on soil temperature and found that as snow cover increases, the insulating effect increases until a maximum depth around 40 cm, after which the insulating effect begins to plateau. Likewise, Goodrich (1982) found through numerical simulations that a doubling of snow cover from 25 to 50 cm resulted in a 36% increase in ground temperature. Therefore, the frost depth will be more sensitive to air temperature with decreasing snow depth. This explains why the Wetland site, which had cooler winter air temperatures than the Low-Elevation site, had a similar median frost depth despite its deeper snowpack.

The Wetland site had a significantly longer median duration of frozen ground presence than the other sites, which is likely due to the thermal characteristics of peat (Atchley et al., 2016; Beringer et al., 2001; Hinzman et al., 1991; Brown & Williams, 1972). We found upper layers of the peat froze early and rapidly which we attribute to the thermal properties of hummocks that tend to have drier soil conditions and thus a relatively low latent heat of fusion (Brown & Williams, 1972). We also found that once frozen, the frost persisted later in the season than other sites which agrees with literature (Atchley et al., 2016; Beringer et al., 2001; Hinzman et al., 1991; Brown & Williams, 1972). Peat is known to have a lower thermal conductivity for all saturation states compared to that of a mineral soil (Atchley et al., 2016; Brown & Williams, 1972). Furthermore, due to its high porosity, peat can have a high specific heat capacity when at or near saturation (Atchley et al., 2016; Brown & Williams, 1972). Taken together, when peat is saturated, it has a high latent heat of fusion and low thermal diffusivity, which means frost can persist late into the season. For instance, Atchley et al. (2016) modelled the impact of peat depth and saturation on active layer thickness in a permafrost environment and found a thicker peat layer reduced the depth of the active layer thickness regardless of its VMC.

4.2.3 VMC & Recharge

We found that the differences in the R_{SS} and RE_{SS} and SS runoff between sites was directly related to the site differences in depth and duration of frozen ground. Frost formation can significantly decrease infiltration capacity of the soil; Thus, soil frost can promote water partitioning into runoff rather than recharge (Hyman-Rabeler & Loheide

II, 2023). However, macropores often remain open and allow some infiltration unless the ground is extremely wet when freezing occurs (Hyman-Rabaler & Loheide II, 2023).

At all three sites in our study, cryosuction frequently caused the topmost soil layer to reach high saturation conditions when frozen, which largely inhibited wintertime recharge over the period of frost presence and promoted runoff generation. Cryosuction is the process of warmer water being drawn towards a freezing front due to a thermal gradient inducing negative pore pressures (Miller & Black, 2003). Alekseev (2022) found that during the freezing of peat; frost heaving became present at peat saturation greater than 40% and the moisture content of the peat samples increased due to cryosuction. Cryosuction impacts to soil VLWC was not seen in observations; However, the dielectric properties of ice are different than that of liquid water, thus, can lead to underestimated values of ice content when measuring with time-domain reflectometry (Wu et al., 2021). Several studies have found that the formation of a surface ice layer due to saturation from snowmelt or rain-on-snow events followed by a subsequent freezing air temperature can efficiently block infiltration (Kane & Stein, 1983; Stein et al., 1994; Iwata et al., 2011; Hyman-Rabaler & Loheide II, 2023). Furthermore, wetter frozen soils have been found to restrict infiltration more than drier frozen soils (Granger et al., 1984; He et al., 2015). Additionally, it is well known that cryosuction can lead to layers of high ice content (Fu et al., 2022). However, the link between cryosuction and recharge-runoff partitioning is not well studied in seasonally frozen ground.

We found that site differences in water table depth exhibited influence on cryosuction processes whereby a shallower water table was associated with more pronounced cryosuction. In addition to increasing VMC at the freezing front, cryosuction draws water from deeper soil layers which leads to water table recession as a part of this process (Jiang et al., 2023). We found that cryosuction-induced decreases in VMC_{NOI} (or increases in VMC_0) was reduced in sites with deeper water table depths, consistent with other literature findings (Alkhaier et al., 2012; Chen et al., 2019; Hansson & Lundin, 2006; Jiang et al., 2023). Lower GWLs cause a larger zone of low unsaturated hydraulic conductivity, which restricts water migration towards the freezing front (Jiang et al., 2023). While a shallower water table may increase the ice saturation of the frozen soil, it

leads to shallower frost depths as water (liquid or ice) has a higher heat capacity and thermal conductivity than soil particles (Jiang et al., 2023). Thus, water level differences drive differences in ice saturation and overall soil moisture content but the deeper frost depths in the Wetland site compared to the High-Elevation Forest site are not attributable to its shallower water table depth but rather to the lesser snow depth and thermal properties of peat as previously discussed.

Overall, our study found that R is closely tied to frozen soil presence, which is variable between the sites due to topographic, soils, and landcover differences. The interaction between recharge and frozen ground is sensitive to the temporal sequence of precipitation, air temperature, and antecedent soil moisture conditions (Hyman-Rabeller & Loheide II, 2023). As such our study demonstrates the importance of utilizing numerical modelling with a high temporal resolution to quantify wintertime recharge differences between sites. Furthermore, our study highlights that even within a small catchment, recharge and the role of frozen ground can have high spatial variability. Thus, applying a single recharge rate to a study area may not adequately represent recharge processes, or in the case of 3D groundwater models, groundwater flow paths and recharge sources.

4.3 CLIMATE CHANGE IMPACTS ON RECHARGE RATE & EFFICIENCY

The shift from less snow to more rain is a commonly raised concern in literature as it is expected to reduce groundwater recharge and its subsequent discharge to streams given that cold-region wintertime recharge is historically dominated by snowmelt (Earman et al., 2006; Hyman-Rabeller & Loheide II, 2023; Huntington & Niswonger, 2012; Godsey et al., 2014; Jyrkama & Sykes, 2007; Sulis et al., 2011; Eckhardt & Ulbrich, 2003). Despite these concerns, few studies have investigated the specific impact of a precipitation phase shift on groundwater recharge in a warming climate, in part because it is difficult to separate snow and rain recharge in practice. Earman et al. (2006) studied isotope hydrology in the Southwestern United States and found that the historical component of groundwater recharge from snow was between 40% to 70% with snow only accounting for 25% to 50% of the total annual precipitation. For this reason, Earman et al. (2006) proposed that in a future climate, a shift in precipitation phase toward more rain and less snow may reduce groundwater recharge. In our study, we could not isolate

the R and RE for individual snowmelt/rain events because the time lags associated with percolation blended the two sources. However, our findings show R_A remains stable in a shifted precipitation regime, but RE during the SS slightly decreased for the High-Elevation Forest site (median decrease -2.8%). The reduction in RE is due to a combination of higher ET rates as well as reduced infiltration capacity due to increased frost presence. Therefore, our study adds to our understanding of future recharge changes in cold regions, establishing that in a humid, low-mountain setting (Cape Breton, NS) under a warming climate, increased winter rainfall compensates for snowmelt losses but is generally less efficient. Despite the limited impact of climate change on R_A , we found large shifts in the seasonal behaviour of recharge; The timing of recharge shifts to be earlier in the water year as consistent with literature predictions of more winter recharge and less spring recharge (Sulis et al., 2011; Eckhardt & Ulbrich, 2003; Rivard et al., 2014; Jyrkama & Sykes, 2007). Furthermore, the mechanisms governing seasonal recharge experienced significant change in a warmer climate. Taken together, our study supports the idea that exploring climate change impacts on R_A requires considering seasonal or finer temporal resolution due to the distinct seasonality of cold region R (Jasechko et al., 2014; Huntington & Niswonger, 2012; Cochand et al., 2019; Eckhardt & Ulbrich, 2003; Sulis et al., 2011).

Our study also demonstrates the value of utilizing numerical modelling to extend the utility of field data for capturing important spatial and temporal dynamics in hydrological processes. Many studies have evaluated climate change impacts on recharge using numerical models, but they typically use a lower temporal resolution (e.g., monthly or daily) and a larger spatial area (e.g., watersheds between 700 to 13,000 km²) than in our study (Boumaiza et al., 2022; Jyrkama & Sykes, 2007; Jasechko et al., 2014; Sulis et al., 2011). In contrast, our study methodology enables high-resolution temporal comparisons of hydrological processes between sites. As a result, we were able to pinpoint the mechanisms driving spatial differences in recharge changes under a warming climate. Below, we discuss the mechanisms which contribute to changing recharge processes.

4.3.1 Evapotranspiration

Our study found ET on an annual scale increased under a warming climate at a similar rate between the three sites due to warmer air temperatures as congruent with other literature findings (Boumaiza et al., 2022; Jyrkama & Sykes, 2007). The projected ET changes (6.7% to 7.6% median percent change) in a warming climate found in our study were within climate change projections for ET found in other studies in Eastern, Canada (Boumaiza et al., 2022; Jyrkama & Sykes, 2007). For instance, Boumaiza et al. (2022) predicted actual ET trends between 0.130 mm/year to 0.468 mm/year (our study ET trends were between 0.44 to 1.6 mm/year). Likewise, Jyrkama & Sykes (2007) simulated between a 3% to 12% increase in ET.

4.3.2 Snow Depth & Duration

There is abundant evidence to support our finding that climate change has the potential to significantly decrease snow depth, SWE and snow cover duration. Climate change is generally anticipated to cause warmer winter air temperatures, which will result in more precipitation falling as rain instead of snow, more mid-season melt events, decreased snowpack thickness and snow cover duration, and earlier melt-out timing (Jyrkama & Sykes, 2007; Jasechko et al., 2014; Musselman et al., 2017; Mote et al., 2005). The impact of increased precipitation on snowpack is less studied, as climate change impacts to precipitation are less certain and more variable globally (Barnett et al., 2005). For locations where climate change is anticipated to increase precipitation (e.g., eastern North America), this is associated with an increase in winter rain and it is expected that this would further contribute to snowpack melt as rain-on-snow events can significantly increase turbulent, latent, and sensible heat fluxes in the snowpack (Jyrkama & Sykes, 2007; Hotovy et al., 2025).

Our study supports previous research findings in lower-elevation mountain areas that the impact of climate change on snow processes can be especially significant. For example, Grundstein & Mote (2013) found in the western United States that the fraction of stations with significant negative trends in snow depth was greatest in lower elevation areas (<1000 m) between 1949 to 2005. Our results also align with existing studies showing that lower-elevation mountain areas often exist in a rain-snow regime transition

zone and tend to exhibit the most change to snowpack thickness in a warmer climate (Dobre et al., 2025; Klos et al., 2014; IPCC, 2022; Rasouli et al., 2022). Our high-elevation sites (relative to our watershed relief) were located within this precipitation regime transition zone; Thus, we saw larger changes in snow pack depth in a warmer climate. Whereas the Low-Elevation Forest site was historically more rain-dominated and exhibited less snow depth sensitivity to climate change. However, the snowpack presence duration was most reduced in the Low-Elevation site due to its historically low snow depth. However, to our knowledge, our study is one of few which have aimed to quantify the future climate change impacts to snow processes in a low-elevation mountain area and the first study to utilize modelling to predict snowpack changes in Atlantic Canada.

Many studies support that significant decreases in snowpack depths and SWE will occur in a warmer climate, but there are limited studies which predict the magnitude of the trend, especially for low elevation areas (Jyrkama & Sykes, 2007; Jasechko et al., 2014; Musselman et al., 2017; Mote et al., 2005; Brown & Braaten, 1998; Dyer & Mote, 2006). Furthermore, most studies tend to report snow depth changes on a large spatial scale, but our findings support that climate change impacts on snow depth and duration can have substantial spatial variability even in a small catchment. Our results generally exhibited higher rates of snow depth and SWE declines than other historical observational studies and studies evaluating simulated climate change projections (IPCC, 2022; Government of Canada, 2019; Suriano et al., 2019). We predicted a change of median snow depth of 67%, 74%, and 43% (17%, 18%, and 11% per decade) from the historical to the SSP2-4.5 2055-2064 scenario, for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites, respectively. Whereas other studies found between 2% to 9% decrease in snow depth per decade (IPCC, 2022; Suriano et al., 2019). The median SWE in our study decreased by 75%, 77%, and 49% (19%, 19%, and 12% per decade) from the historical to the SSP2-4.5 2055-2064 scenario, for the Wetland, High-Elevation Forest, and Low-Elevation Forest sites, respectively. In contrast, the Government of Canada (2019) observed an average Canada-wide SWE decrease of only 5% to 10% from 1981 to 2015 (1% to 3% per decade). However, they predicted that SWE would decrease at a higher rate of 5% to 10% per decade by 2050 for southern Canada, with the Maritimes being on the higher end of the range (Government of Canada, 2019).

Our results also exhibited higher rates of snow cover duration declines (between -9% to -17% per decade or between -9 to -15 days/decade) than historical observational trends, which were between -5 to -10% per decade across Canada from 1998 to 2015 (Government of Canada, 2019) or -4 days/decade in western North America from 1972 to 2013 (DeBeer et al., 2016); Potentially indicating that climate change impacts are accelerating. However, this difference could also be due to study site differences. Studies which predicted future changes in snow cover duration through modelling aligned closer with our results than observational trends but still tended to be lower (between -7% to -40% per decade; Klos et al., 2014; Mitterwallner et al., 2024 and between -3 to -5 days/decade; Bresson et al., 2026). This discrepancy may coincide with literature findings that studies in the rain-snow transition zone are particularly sensitive to climate change (Klos et al., 2014) and that most studies used for comparison are in higher elevation bands, thus likely more snow dominated. Ultimately our study supports smaller scale site characteristics, like landcover and elevation, can results in unique changes in snow processes from climate change. Therefore, understanding this spatial variability is essential to be able to extrapolate study findings to other watersheds.

4.3.3 Frost Depth & Duration

Our projected change in frost depth and duration in a warming climate was closely related to the change in snow depth and duration, in agreement with other studies (Campbell et al., 2010; Lundberg et al., 2016; Hyman-Rabeler & Loheide II, 2023). Going beyond other studies, we highlight the spatial variability of frost depth and duration changes within a site due to elevation. During the middle of the winter season, we found that air temperature, and the associated temperature lapse rate, became a more important control on high-elevation frost depth and duration due to the decrease in thermal insulation from shallower snowpacks. Although we expect that the negatively correlated trends between frost presence and snow depth will diminish with increasing climate change intensification as air temperature increases will eventually offset the thermal insulation losses from shallower snowpacks. We expect lower elevations (rain-dominated), which historically already experience limited insulation from their shallower

snowpacks, will have shallower frost depths and reduced frost presence over the entire season in a warmer climate.

Our study findings are congruent with others which have investigated freeze-thaw cycle changes, in that we found climate change will result in an overall decrease in the frost season length due to later onset of frozen ground and earlier thaw (Chen et al., 2022; Leduc & Logan, 2025; Bresson et al., 2026). We predicted frost thaw was reduced in the shoulders of the winter months due to thinner snowpacks and more above zero air temperatures, as consistent with literature trends (Chen et al., 2022; Leduc & Logan, 2025; Bresson et al., 2026).

In comparison to other literature estimates in nearby areas, we found the frost season length change (Historical to SSP2-4.5 2055-2064) was similar to that found in our study (between -2.7% to -3.6%/decade). ClimateData.ca (2026) projected a decrease in the median number of days with ground frost of 3% per decade for the CMIP6 SSP2-4.5 scenario in Eskasoni, NS when comparing the 30-year averages from 2001 to 2030 and 2041 to 2070. Likewise, across all of Quebec, Bresson et al. (2026) found the duration of the frost season decreased by around 2% per decade for the SSP2-4.5 scenario between 1991 to 2020 and 2041 to 2070.

Examining the hydrological simulations at hourly, seasonal and annual time scales provided complementary information, particularly for the soil frost processes. For example, despite the similar behaviour observed from the hourly-scale graphical analysis between our two high-elevation sites, we found they had different overall frost depth and duration trends on a seasonal scale. The conflicting trends were deemed due to differences in soil properties between the mineral and peat soils and, in turn, the historical frost presence, rather than a difference in climate change mechanisms; Peat characteristics promoted season-long ground frost in the Wetland site and often deep frost depths during the middle of winter, whereas the High-Elevation forest was often frost free. As such, despite both sites exhibiting increased frost depth during the middle of winter due to snow pack thinning, more days in the High-Elevation site shift from unfrozen to frozen thus the change is more significant. However, this discrepancy between interpreting the hourly versus seasonal results highlights the importance of

analyzing the winter season on a finer temporal resolution to accurately capture the mechanisms driving site differences.

4.3.4 Recharge

We found that the site differences between changes in R_{SS} and RE_{SS} induced by climate change are primarily driven by site differences in changes to frost processes. The Low-Elevation Forest site had favourable recharge conditions under a warming climate, due to decreased frost presence, which promotes R in a more rain dominated SS. Several studies support our results that warmer air temperatures induced by climate change can reduce frozen ground and promote higher rates of winter recharge (Rivard et al., 2014; Jyrkama & Sykes, 2007). In contrast, the two high-elevation sites are within a temperature zone in which, despite an air temperature increase, reduced snowfall and snowpack thickness can promote mid-winter frost formation and create unfavourable infiltration conditions. As a result, the R_{SS} in the high-elevation sites is more dependent on highly temporally variable site conditions (e.g., the interplay between snow depth, frost depth and the mechanisms that govern these processes), thus the timing and amount of P has heightened importance. Few studies have highlighted this complex recharge response. In one such study, Hyman-Rabeller & Loheide II (2023) conducted a statistical historical recharge analysis in the Great Lakes region of the United States and found similar results in which recharge is sensitive to the temporal sequence of air temperature, frost presence, and precipitation.

Changes in SFS P will also become increasingly important in a warmer climate due to the threat of reduced R_{SFS} from increased ET. It is well established in literature that increased air temperatures due to climate change will increase ET (Sulis et al., 2011, Goulden & Bales, 2014). In the Christmas Brook watershed, the P_{SFS} is predicted to increase enough to offset heightened ET losses across all sites. However, the change in summertime P_{SFS} under a warming climate is highly uncertain and, as demonstrated in literature, spatially variable; Rivard et al. (2014) and Sulis et al. (2011) found summer recharge decreased due to increased ET and decreased P , in NS, and Quebec, respectively. Meanwhile, Scibek and Allen (2006) found the dry months in their south-central British Columbia study watershed became wetter and summertime recharge increased in

response to climate change. Additional uncertainty is introduced by the interannual variability of rainfall; Despite the limited overall change in R_{SFS} in our study, in years with prolonged low periods of rainfall, the severity of droughts may be enhanced.

Literature estimates of annual recharge trends change are highly variable and few studies evaluating R_A in a comparable or proximal watershed are available with which to compare our results. In Quebec, Canada, Boumaiza et al. (2022) found an increasing trend of 0.7 mm/year in R_A from 1910 to 2009. Whereas, in Ontario, Canada, Jyrkama & Sykes (2007) predicted an increase in R_A of 100 mm/yr under a range of climate change scenarios. In our study we found the median R_A rate of change (-0.36 mm/year to 0.95 mm/year over all scenarios) agreed more closely with the rate reported by Boumaiza et al. (2022).

Due to the increased sensitivity of R_A to P timing and intensity in both seasons, we expect there will be enhanced interannual variability of R_A under climate change and the spatial variability in R_A rate changes are tied to spatial variability in P. Literature findings support the result that the change in R_A is strongly tied to the precipitation changes and its ability to offset increases in ET and changes to soil infiltration capacity (Meixner et al., 2016; Boumaiza et al., 2022; Jyrkama & Sykes, 2007). Meixner et al. (2016) found increased air temperature and ET, and decreased precipitation resulted in decreased R_A for several of their southern study sites in the western United States. Whereas in their northern study sites, the changes to R_A were more uncertain due to a projected increase in precipitation. Boumaiza et al. (2022) found their increasing R_A trend was due to increased precipitation and snowmelt offsetting increases in annual ET and surface runoff. Likewise, Jyrkama & Sykes (2007) predicted an increasing trend that was attributed to increased precipitation, but also due to enhanced soil infiltration capacity from reduced frozen ground.

Our study found that, despite clear seasonal changes to R processes, the R_A remained similar or slightly higher than the historical climate scenario across all sites. Although R_A is predicted to be similar to historical rates, changes to the underlying mechanisms governing R_A have significant implications on water resources management. For instance, our work highlights that the timing of water availability may shift in a

warming climate which can help aid water withdrawal rates and timing. Additionally, our study suggests RE will be reduced in the winter under climate change, thus water resources managers can proactively expect water availability will be more sensitive to changes in precipitation amount.

4.4 LIMITATIONS, RECOMMENDATIONS, & FUTURE WORK

4.4.1 Model Calibration & Validation Limitations

The main limitation in our study was the relatively short record available for model calibration and validation. Limited observational data can lead to overfitting the models to the calibration period and the model poorly representing the physical processes. However, the advantage of using a physically based model like SHAW is that the model is less reliant on extensive observational records than that of an empirical model (Guinot & Gourbesville, 2003). Several studies have successfully utilized SHAW to assess soil moisture and temperature dynamics with relatively short observational datasets for calibration or validation between less than a year to two years (Flerchinger & Hardegee, 2004; Flerchinger & Hanson, 1989; Li et al., 2012; Zhang et al., 2013). For most of our models we had several years of snow depth observations. Thus, we expect our model simulations of snow depth to be robust to different climatic conditions. Model simulation certainty could primarily be improved through increased record length of soil liquid water content and soil temperature observations. Although, we maximized our observational data by integrating groundwater level observations in model calibration and validation of soil moisture. Furthermore, we used available literature or measured site values to tightly constrain each parameter value utilized in SHAW to minimize the solution space.

4.4.1.1 Soil Temperature & Liquid Water Content

For all landscapes, model errors were typically highest near the soil surface for both temperature and LWC. The variability of soil temperature was higher closer to the surface at the soil-atmosphere interface, typically resulting in the improvement of temperature simulation performance with depth. Additionally, soil temperature was impacted by the soil LWC, thus node depths with higher soil LWC simulation performance also generally had better soil temperature performance. Similar to the

temperature performance, the soil LWC simulation performance typically was the worst at the soil-atmosphere interface. Strong et al. (2025) found SHAW simulations of soil temperature and moisture improved with depth and pointed to model difficulties in simulating soil physics at the soil-atmosphere interface. In the validation period of our study, typically the models remained within acceptable calibration criteria except at the surface soil node. While Strong et al. (2025) simulations were produced at a daily time-step our results confirm similar issues may exist at a finer temporal resolution.

The two forest models met all necessary performance criteria during calibration, but the Wetland model was unable to achieve acceptable performance criteria for the soil LWC, likely due to the numerical limitation of modelling the hydrologic nature of the peat substrate. Observations of LWC and GWL in the wetland varied largely over a small change in depth; Upper depths were quite dry, but the groundwater table fluctuated close to surface, resulting in the close proximity of highly unsaturated and saturated peat. The ability to rapidly shift between dry and saturated conditions is typical of peat but not of mineral soils (Dimitrov & Lafleur, 2021). SHAW is typically used in mineral soils with limited organic content. Thus, the wetland model likely struggled to accurately simulate soil LWC as the porosity and hydraulic conductivity were much higher than typical mineral soils. Peat soil is also known to sharply transition from hydrophilic to hydrophobic with decreasing water content. However, this transition is not solely governed by water content but also from chemical changes in the peat itself (Wu et al., 2020). Therefore, this phenomenon, which is not modelled in SHAW, may also explain the difficulty the model experienced in simulating the extreme change in moisture content over a small change in depth as observed in the wetland site. Peat is also known to have dual-porosity, which consists of a connected macropores and immobile porosity (McCarter et al., 2023). As a result, the soil water retention curves commonly used to describe mineral soils may not adequately represent processes in peat (Rezanezhad et al., 2016). Furthermore, due to the prevalence of macropores in peat, the simulation of soil LWC in cold regions can be challenging as these macropores can allow rapid infiltration despite the presence of soil frost (Rezanezhad et al., 2016; Mohammed et al., 2018). Despite the Wetland model's limitations, the model can provide valuable insights into

soil-freeze thaw processes under highly saturated conditions with highly porous soil properties thus model results were still interpreted, albeit with caution.

At the low-elevation forest site (12 – 15 events), and in one instance at the high-elevation forest site, the observed soil LWC exhibited highly flashy behavior at some node depths that were not captured in the simulation. It is unclear whether the observational response is realistic or a potential measurement error. The timing of these events typically corresponded with the timing of high intensity rainfall events and were observed in the second SMT probe for the similar depth in the soil but typically were smaller in magnitude. During probe installation, the forested sites' soil became had increasingly more cobbles and boulders with depth which could have introduced cavities when moved by the probe during installation. These cavities could fill with water during an infiltration event and produce artificially high soil liquid water content measurements.

Aside from the high-elevation forest site, the simulated soil LWC often performed the worst in the winter, and tended to overestimate the actual soil LWC, likely due to inaccuracies in the simulated soil temperatures or numerical limitations in modelling soil ice content. The soil temperature and ice content was observed to be sensitive to snow depth, thus inaccuracies in snow depth simulations could also drive these errors.

The performance of the models during the validation period were typically worse than over the calibration period, as expected. The validation period (July 2025 to March 2026) included an unusually dry period over the summer and fall. During this period, GWLs across the watershed were much lower than average over the entire record period and remained low for a long duration. The validation period performing worse than the calibration period is likely due to the model not including the response of soil temperature and LWC during drought conditions. As such, this suggests the model is likely more representative of a normal hydrologic year but may not represent extreme hydrologic events as well. However, the performance of the two forest sites during validation met the necessary performance criteria for most soil node depths, suggesting they are relatively robust.

4.4.1.2 *Snow Depth & Density*

All models achieved acceptable NSE metrics for the simulated snow depths but had substantially different levels of performance. The snow depth in the high-elevation forest site performed the best out of the three models, followed by the wetland site, and the low-elevation forest site. The difference in model performance in simulating snow depth could be driven by a variety of factors, such as the observation length and quality, the calibration methods, and the complexity of the model. The high-elevation forest model likely performed the best as snow depth was calibrated primarily to one year. While the WS-Wetland snow depth was used to aid the calibration of the High-Elevation Forest model to avoid overfitting the model to this year, this bias may still exist. Moreover, the variability in snow depth calibration performance could be due to the quality and quantity (10 years) of observational data. The nature of the calibration process of the two forest sites could also hinder the ability to optimize the snow depth simulation. In the two forest models the snow depth and soil conditions calibration processes were interlinked. As a result, the snow simulation performance tended to slightly decrease following the calibration of the soil processes.

SHAW has been proven to be robust at simulating snow depth in several studies and uses the detailed snow routine from Anderson (1976) which accounts for snowpack changes due to compaction, settling, and vapor transfer (Campbell et al., 2010; Flerchinger et al., 1994). Our study further highlights the SHAW model's ability to effectively capture the major mechanisms controlling snow depth and thus accurately simulate snow depth. The quality of performance was best represented in the Low-Elevation Forest model as it had the longest record of snow depth measurements. Although the Low-Elevation Forest model performed the worst comparatively, the model demonstrated that SHAW accurately simulated snow depth across all ten snow seasons. However, as expected, the validation performance of the models was typically worse than over the calibration period. The two high-elevation models still performed well, but the Low-Elevation Forest site was substantially worse. Over the validation period, the Low-Elevation Forest site simulated the 2024/2025 snow season closely but largely underestimated snow depth during the 2025/2026 snow season. The 2025/2026 snow

season had an above average snow depth. The model was calibrated to similar snow depth seasons like 2014/2015, 2016/2017, and the 2023/2024 and also tended to underestimate snow depth for the latter two seasons. Thus, the model may be calibrated such that it poorly simulates high snow depth accumulation years, or this simulation error could be due to other sources of error.

Across the three models, there are few instances of significant over or underestimation of snow depth, but their occasional occurrence suggests there may be some controls on snow depth not well accounted for in the model either numerically or in its parameterization. Due to the one-dimensional nature of the SHAW model, it is unable to account for spatiotemporal interactions of the terrain, wind, and vegetation, which can influence the snow accumulation, redistribution, and melt (Lundberg et al., 2016). Lundberg et al. (2016) summarises many of these effects such as more snow accumulation on the leeward side of a forest or in surface depressions. These errors may also stem from the multi-objective approach used to calibrate the models, which often required compromised between optimal snow depth performance and soil moisture and LWC performance. The calibration RMSE error range of 3.6 to 8.6 cm for the three models is relatively close to the observed sensor RMSE of between 5 cm up to 12 cm when compared to manual snow depth surveys. Thus, some simulation error in the models could also be partially accounted for in snow depth sensor measurement error. Furthermore, rain gauges often record underestimations of precipitation, especially for snowfall due to the impact of wind which could result in lower snow depth simulations, or over-calibrated model parameters to compensate for this issue (Grossi et al., 2017).

The models were further validated using measured snow density estimates. However, all models had below zero R^2 , thus performed worse than each site's averages. Flerchinger et al. (1994) validated their model with measurements of snow density and found significantly higher R^2 of 0.69. However, Flerchinger et al. (1994) had significantly more snow density observations over time, whereas we were limited to four snow surveys. We conducted numerous samples for each snow survey to account for spatial variability in density which tended to be large. However, due to the poor performance of snow density simulations, it is recommended for future work that a higher

temporal frequency of snow surveys be prioritized over a high spatial resolution. Furthermore, these observations should be used to calibrate the model because calibrating to snow depth alone does not achieve acceptable snow density performance. Snow density is a primary driver of the thermal insulation of the snow and thus poor representation of snow density in the model could result in poor simulations of soil temperature and frost formation, thus impacting our estimates of recharge amount and efficiency.

4.4.2 Recharge Analysis Limitations

In our literature review and even between the different sites studied in Christmas Brook watershed, we find climate change impacts to hydrological processes were spatially variable. While our findings can be extrapolated to similar watersheds in Eastern Canada, their broader applicability remains unclear. Two out of three models were able to meet all acceptable criteria for calibration, but we found a high degree of uncertainty in our analytical outputs, which reduces the reliability of the climate change projections.

For all models, except the topmost layers of the Wetland site, we made the simplifying assumption that there was no lateral water movement. While SHAW allows for the simulation of lateral water flow out of the model as well as the input of lateral flow entering the system, balancing the two requires iteratively solving the model to balance upslope and downslope water fluxes. Alternatively, representing these 2D processes would require adequate observational data of these processes to properly calibrate the model. We found we had insufficient data to include lateral flow in the models and iteratively balancing later fluxes was deemed too complex for the scope of this study. The two high-elevation sites were in low-slope areas, therefore lateral water movement was expected to be negligible. We allowed the Wetland's topmost layers to have lateral drainage as we anticipated the hummock could drain laterally into the hollows but not the reverse. In the Low-Elevation Forest site, within the immediate area of the SMT probes the site had a moderate slope but on a larger scale was located within the much steeper slopes of the river gorge area. Thus, there could be significant lateral flow at this site both from the upslope areas and to the downslope areas not accounted for

in our model. However, this model was fully above the water table and therefore lateral flows would be mostly immediately following precipitation events (i.e., interflow).

In our recharge analysis, the separation of years for the annual analysis and seasons for the seasonal analysis could introduce error into the total recharge and RE estimates due to the delayed response of R to precipitation inputs. However, this error is expected to be minimal relative to the overall R and RE. Furthermore, precautions were taken to avoid the misallocation of events between years and seasons by separating years/seasons during periods with minimal precipitation and recharge and with no snow or frost cover.

Our study constrained the calibration process using measured site values or literature estimates of input parameters. However, as shown in the uncertainty analysis, there was a high degree of uncertainty in the analytical results. Based on the sensitivity analysis, the canopy LAI and the soil water retention curve characteristics tended to have the most impact on the analytical results. Therefore, certainty in our results could be improved through the tighter constraint of these parameters estimates by directly measuring site values. Additionally, during model development, it was expected that increasing the depth of the lower boundary in the model would minimize the impact of the lower boundary conditions on the model. However, as shown in the sensitivity analysis, especially for the two high-elevation sites, the drainage rate out of the model bottom significantly impacted water movement through the system. Therefore, decreasing the depth of the model may improve model performance and reduce the uncertainty by simplify calibration processes through reducing the number of soil layers required for calibration and the model solution space.

4.4.3 Recommendations & Future Work

Understanding how changes in R across different landscape types will impact community water supply requires integrating our results with a broader understanding of the watershed and aquifers of interest in Christmas Brook watershed and beyond. Dauphinee (2025) conducted a previous study of Christmas Brook watershed in which there were three main water sources to the pumping well supply in the low-elevation

area: local precipitation recharge within the fault area, MFR due to stream water entering the subsurface, and MBR. Historically, both the wetland and forested gorge areas were found to be important sources of streamflow (Dauphinee, 2025). Furthermore, Dauphinee (2025) determined wetland water accounted for 73% of streamflow, despite the area only accounting for 11% of the total watershed area according to end-member mixing analysis results. Taken together this previous work highlighted the need for our study to evaluate the three different locations, which were sources of recharge, in the watershed.

The current study complements previous work with important implications for water resource management. We found that under climate change, the amount of recharge at our three sites remained similar to historical levels on an annual and seasonal basis across all sites. However, the timing of recharge and changes to the amount of streamflow generated from site runoff may reduce or shift the water availability at the pumping well in the low-elevation area. Under climate change, we found the SS recharge in the wetland remains similar to historical levels, but the winter runoff may be significantly reduced. Thus, the amount of streamflow generated in a warming climate may shift in terms of timing and magnitude. Moreover, we found the timing of diffusive recharge in the Low-Elevation Forest site, an important well recharge area according to Dauphinee (2025), shifted earlier in the season. Taken together, the timing of water availability at the pumping well may be shifted in a warming climate and possibly reduced depending on changes to streamflow. Pumping rates from the fault aquifer may need to account for these water availability shifts. Further research could be completed which integrates the SHAW model recharge results from our study into a transient version of the Dauphinee (2025) MODFLOW model of Christmas Brook watershed to understand the implications of climate change impacts to groundwater recharge timing throughout the year on streamflow and groundwater availability in the fault aquifer.

The implication of a precipitation phase shift on groundwater discharge to streams was outside of the scope of the study but could also have important hydrological and ecological impacts. Groundwater is important for sustaining streamflow year-round but especially during low-flow periods (Dierauer et al., 2018); Likewise, in Christmas Brook watershed, Dauphinee (2025) found baseflow was an important component of streamflow

in the summer. Earlier snowmelt recession is anticipated to advance the timing of groundwater discharge to streams resulting in lower summertime flows due to a longer groundwater recession period (Huntington & Niswonger, 2012). Therefore, despite similar recharge rates at the high-elevation sites, the earlier freshet timing may still lead to reduced streamflow and in turn reduced mountain-front recharge in the low-elevation watershed area. Moreover, the ecological implications of a precipitation phase shift are not well understood; In Qamsipuk (Indian) river located in a nearby watershed in Eskasoni, the EFWC is actively working to restore Atlantic salmon habitat, a species listed as Endangered under the Committee on the Status of Endangered Wildlife in Canada. In groundwater-dependent streams, groundwater discharge provides stream temperature and nutrient levels regulation which benefits Atlantic salmon and other cold water fish (Dugdale et al., 2013; Smith et al., 2023). Thus, further research or monitoring of groundwater discharge locations to streams may aid their aquatic habitat restoration efforts.

While our study investigated the change in RE between historical and future climate conditions, our study also could not compute the change in RE for individual precipitation phases (i.e., the RE of snow versus rain) as the distinct precipitation components of the R signal were not separable. The R response was typically so delayed from the precipitation/snowmelt events that these events consistently overlapped. Furthermore, in cases where the events were adequately separated in time, the R response was not expected to be the same as the case when there are mixed events occurring. The impact of RE for the different precipitation phases under warmer climate conditions could be investigated using modelling. However, it would require extensive modelling of a variety of antecedent conditions and precipitation/snowmelt inputs to get a comprehensive idea of possible recharge conditions and their probability of occurring in the historical versus the future climate.

Overall, we recommend that the monitoring program in Christmas Brook watershed be continued. Long-term hydrological observational records are extremely valuable for future analytical work and for validating the accuracy of completed analytical work. Additionally, continuous monitoring of the watershed can be used to

actively inform groundwater resources management decisions and encourage emergency preparedness to extreme hydrological events like floods or droughts.

Our study had significant uncertainty associated with the results, and the RE_A and R_A were generally higher than what was found in literature. Thus, it is also recommended that future work be completed to estimate of R and RE using observational data, such as the water table fluctuation method. Not only would this help to verify the historical model estimates of R and RE but would also help validate the results of the climate change projections and conclusions drawn from these results.

Our results demonstrated that recharge processes in a warmer climate increasingly rely on rainfall precipitation patterns, thus monitoring precipitation may become increasingly important. Climate models have lower certainty in predicted impacts to P than air temperature, due to its high spatial and temporal variability (Bush & Lemmen, 2019). Additionally, historical P in our watershed was highly variable interannually. The temporal complexity of wintertime recharge processes due to freeze-thaw dynamics combined with the inherent uncertainty in P patterns means the monitoring and management of water resources will likely become more difficult in a warming climate. For this reason, real-time SMT measurement networks may be a useful tool for water managers in warming cold regions. Additionally, the increased importance of summertime precipitation at offsetting increased ET , means drought severity and duration may increase in a warming climate. Droughts are becoming an increasing concern in the Canadian Maritimes. Motamedi et al. (2026) found in their trend analysis of precipitation, GWLs, and baseflow in NS, extreme dry days are increasing, and more severe droughts are becoming more frequent and widespread throughout the province.

CHAPTER 5 - CONCLUSION

Our study demonstrated the benefit of utilizing modelling to reveal the complex mechanisms that control groundwater recharge and how less snow and more rain will impact groundwater resources. We used observations of precipitation, snow depth, groundwater level, soil moisture and temperature to develop three SHAW models of different landscape types and elevations in the Christmas Brook watershed, located in Eskasoni, NS. We compared historical recharge patterns between these three sites to quantify site variability and understand historical controls on recharge. We then applied the delta change method to generate future climate data under a range of climate change scenarios for use in the SHAW models. We compared the changes in seasonal and annual recharge due to climate change to the historical scenario as well as analyzed the dominant mechanisms driving these changes.

We found that there was a high degree of spatial variability in recharge and recharge efficiency even within our relatively small study catchment. Under climate change, we found recharge in the snow-free season remained similar to historical rates as increased precipitation balanced the increased evapotranspiration. In the snow season, we found site-wide decreased snow depth and duration due to a warming climate. However, the change in frost depth and duration was different between sites and scenarios; Two of our three sites had increased frost depth during the coolest months. The differences in frost depth changes suggest there is a temperature threshold at which frost depth will decline where the loss of snow insulation is compensated by warmer air temperatures. This change to seasonal infiltration capacity associated with frost depth meant the timing of rainfall and snowmelt during the winter became more important. As such, their recharge efficiency decreased for the snow season and annually. However, enhanced wintertime precipitation resulted in an overall increase in snow-season and annual recharge amount for all sites in our study.

We found enhanced wintertime rain can compensate for snowmelt losses in a warming climate at our site but with reduced efficiency and altered timing. These changes may have important watershed-scale hydrological effects such as shifts in seasonal water availability timing or increased susceptibility to droughts due to enhanced

dependence on rainfall timing and amount. Overall, our results suggest a shifting hydrological regime under climate change whose future behavior will require sustained observation and careful interpretation.

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6. APPENDIX A – METEOROLOGICAL DATA CLEANING

Meteorological data was mainly used from the two nearby weather stations and solar radiation from NASA POWER (NASA, 2026). The Wetland and High-Elevation Forest sites used weather data from the WS-Wetland which had data available starting in February of 2023. For 2015 to 2023, and for any other data gaps the climate data was filled with bias corrected temperature, humidity, solar radiation, and windspeed from the WS-ECCC, from December 1st, 2011, to March 13th, 2026, and solar radiation data from NASA POWER (ECCC, 2026b; NASA, 2026). The bias correction processes for each meteorological component are described in the following subsections. The Low-Elevation Forest site used weather data from the WS-ECCC, and data gaps were filled with bias corrected WS-Wetland meteorological data. Remaining gaps less than three hours long were filled with linear interpolation. Data gaps larger than three hours were filled with bias corrected metrological data from the ECCC stations in Port Hawkesbury (WS-PHawk; station id = 48668) and Louisbourg (WS-Louis, station ID = 54578; ECCC, 2026c, 2026d).

Temperature, Solar Radiation, Humidity

Seasonal statistics of historical weather observations of temperature, solar radiation, and humidity were calculated for each weather station to bias correct data prior to gap filling. The temperature and humidity statistics are reported in Table 6. The seasonal divisions were categorized as winter from Julian days 1 to 60 and 331 to 365, spring from day 61 to 150, summer from 151 to 240, and fall from 241 to 330.

Table 6: Temperature and humidity statistics at the Wetland and the ECCC Eskasoni, Louisbourg and Port Hawkesbury weather stations from 2015 to 2025.

Statistic	Period	WS-Wetland ^a	WS-ECCC	WS-Louis	WS-PHawk
Median Temperature (°C)	Fall	11.2	11.4	10.9	10.5
	Spring	2.4	3.5	2.6	2.6
	Summer	17.3	17.6	15.9	17.4
	Winter	-3.3	-1.4	-1.1	-2.5
	Annual	7.1	6.8	5.9	6.1
Maximum Temperature (°C)	Annual	30.4	32.5	30.9	33.5

Minimum Temperature (°C)	Annual	-25.7	-23.6	-23.9	-27.4
Median Humidity (%)	Fall	92	84	89	86
	Spring	89	80	86	83
	Summer	91	86	94	86
	Winter	91	80	84	84
	Annual	91	83	88	85

^a2023 to 2025 years available only.

Solar radiation data was available from NASA POWER and directly measured at the WS-Wetland. A simple linear regression between the two sets of solar radiation data was computed with an R^2 of 0.7. The linear regression was used to substitute one source for the other. The two high-elevation sites preferentially used the WS-Wetland solar radiation whereas, the low-elevation site preferentially used the NASA POWER station.

Table 7: Solar radiation statistics at the Wetland site and from NASA POWER 2015 to 2025.

Statistic	Period	Wetland^a	NASA POWER
Mean Solar Radiation (W/m²)	Fall	116	112
	Spring	165	178
	Summer	207	219
	Winter	58	60
	Annual	142	142
Maximum Solar Radiation (W/m²)	Annual	1290	991

^b2023 to 2025 years available only

Windspeed

For the different weather stations linear models of windspeed and wind direction from one site was used to calculate windspeed at the other station and this was used to correct and fill windspeed data gaps.

Weather data in the High-Elevation Forest site was assumed to be the same except for the windspeed. Since the WS-Wetland was in an open area it was assumed that the windspeed in the High-Elevation Forest site would be more sheltered due to the presence

of forest canopy cover. Thus, the forest canopy was converted by adjusting it to account for the frictional differences between sites. The following equation, commonly used in wind turbine windspeed height adjustment methods, was used to convert the wetland windspeed to 50 m above ground (Jung & Schindler, 2021).

$$\frac{\mu_1}{\mu_2} = \left(\frac{h_1}{h_2}\right)^\alpha \quad [17]$$

Where μ_1 is the windspeed at one height, h_1 (50 m), and μ_2 is the windspeed at the WS-Wetland measurement height, h_2 , and α is an empirical value which is commonly set to 0.2 for studies in Canada (Yang et al., 2024). The WS-Wetland windspeed was first converted to 50 m elevation as it was assumed the frictional effect differences on windspeed would be reduced and could be assumed negligible between the Wetland site and the High-Elevation Forest sites. Following this assumption the windspeed at 50 m above was converted to windspeed just above the canopy using the following two equations for the friction velocity and the windspeed at 16 m elevation in the High-Elevation Forest site (Stull, 2017; Sypka & Starzak, 2013).

$$\mu_* = \frac{k * h_1}{\ln(z_R/z_0)} \quad [18]$$

Where μ_* is the friction velocity, k is the von Karman constant equal to 0.4, z_R is the reference height, z_0 is the surface roughness and is equivalent to 0.1 times the canopy height. μ_* was then used in the equation

$$U(h) = \frac{\mu_*}{k} \cdot \ln\left(\frac{h - h_d}{h z_0}\right) \quad [19]$$

Where $U(h)$ is the windspeed at the height, h (16 m), and h_d is equal to 0.65 times the canopy height and $h z_0$ is the roughness length and is equal to z_0 . Under these assumptions the forest windspeed at 16 m elevation was equal to 0.62 times the WS-Wetland station windspeed.

7. APPENDIX B – SUPPLEMENTARY FIGURES, TABLES & EQUATIONS

Table 8: Groundwater Monitoring Well Specifications.

Well ID	Ground Surface Elevation (masl)	Well Bottom Elevation (masl)	Screen Length (m)	Screened Interval (mbgs)	Screened Material
MW1A	194.16	187.46	1.52	5.18 – 6.70	Fractured bedrock
MW1B	194.12	177.02	1.52	15.58 – 17.1	Fractured bedrock
MW2A	192.32	185.92	1.52	4.88 – 6.4	Fractured bedrock
MW2B	192.33	175.42	1.52	15.39 – 16.91	Fractured bedrock
MW3A	196.86	192.29	1.52	3.05 – 4.57	Fractured bedrock
MW3B	196.95	181.41	1.52	14.02 – 15.54	Fractured bedrock
MW4	19.34	0.93	Unknown	Unknown	Fault damage zone bedrock
PNA	200.04	197.29	2.75	0 – 2.75	Peat
PNC	202.00	198.84	2.50	0 – 2.50	Peat

Table 9: Non-Calibrated Model Parameter Values

Parameter	Low-Elevation Forest	High-Elevation Forest	Wetland	Source
Latitude (degrees, minutes)	45, 57	45, 58	45, 57	a
Slope (%)	47	10	6	
Aspect (° clockwise from North)	181	180	180	
Solar Noon		12.5		b
Elevation (masl)	40	193	200	a
Rooting Depth (m)	0.25	0.25	Calibrated	
Windspeed Measurement Height (m)	15	16	6	
Dead plant water potential coefficient & exponent (m)		-53.72, 1.32		c
Saturated water content of dead plant material* (kg/kg)		0		
Critical Temperature for Plant Evapotranspiration		5		d, e
Stomatal Resistance (s/m)		100		c
Exponent for Leaf Potential		5		
Critical Leaf Water Potential		-300		
Resistance of Leaves		100000		
Resistance of Roots		200000		
Solar Radiation Parameter** (W/m2)		143.4		f, g
Lower Temperature Limit** (°C)		4		
Upper Temperature Limit** (°C)		30		
Optimum Temperature** (°C)		20		
Stomatal Reduction Limit**		0.15		
Stomatal Reduction Coefficient** (kPa)		0.3		
Height of Plant Species (m)	13.5	12.5	0.1	a
Characteristic Plant Dimension	4.05	2.2	Calibrated	a
Canopy Dry Biomass (kg/m ³)	13.92	22.76	2	a
Thickness of Snow* (m)		0.05		h
Temperature of Snow* (°C)		0		
Depth of liquid water of snow* (m)		0		
Bulk density of Ice Fraction* (kg/m ³)		200		
Soil Bulk Density (kg/m ³)		Equation [20		c
Albedo of Dry Soil	0.402	0.402	0.2	c,i,j,k
Albedo Exponent	1.75	1.75	1.5	c,i
Percent Sand, Silt, Clay (%)**		33, 33, 34		
Percent Gravel (%)	See Table		0	l,m,n
Percent Organic Matter (%)	10	See Table 11	See Table 12	l,m,n, o,p,q
Porosity	Calibrated			

*At the beginning of simulation.

**Parameters in Stewart-Jarvis function for stomatal conductance controls.

***The percent composition had negligible impact on model, thus was split evenly across soil textures.

^aMeasured in-situ, ^b(National Oceanic and Atmospheric Administration [NOAA], Global Monitoring Laboratory, n.d.), ^cSHAW recommended value, ^d(Sottocornola & Kiely, 2010), ^e(Liu et al., 2016), ^f(Link et al., 2004), ^g(Whitley et al., 2008), ^hAssumed initial conditions, ⁱ(Ohkubo et al., 2021), ^j(Berglund & Mace Jr., 1976), ^k(Kellner, 2001), ^l(Agriculture and Agri-Food Canada, 2019a, 2019b), ^m(Nolan, Davis & Associates Limited, 1984), ⁿ(EXP Services Ltd., 2011a), ^oSMT observations, ^p(Rouxel et al., 2012), ^q(Katsura et al., 2006)

Soil Bulk Density

$$= \frac{1 - \text{Porosity}}{\frac{1 - \text{Organic Matter Fraction}}{2650} + \frac{\text{Organic Matter Fraction}}{1300}} \quad [20]$$

Table 10: Low-Elevation Forest Model Depth Specific Soil Layer Parameters

Depth	Percent Gravel (%)	Percent Organic Matter (%)	Depth	Percent Gravel (%)	Percent Organic Matter (%)
0		28	4.25	11.429	
0.066	0	18.667	4.5	10	
0.116		9.333	4.75	8.571	
0.216			5	7.143	
0.266			5.25	5.714	
0.366	25		5.5	4.286	
0.416			5.75	2.857	
0.516			6	1.429	
0.566			6.25		
0.666	26.25		6.5		
0.766	27.5		6.75		
0.866	28.75		7		
0.966	30		7.25		0
1.016	30		7.5		
1.266	28.571	0	7.75		
1.5	27.143		8		
1.75	25.714		8.25	0	
2	24.286		8.5		
2.25	22.857		8.75		
2.5	21.429		9		
2.75	20		9.25		
3	18.571		9.5		
3.25	17.143		9.75		
3.5	15.714		9.95		
3.75	14.286		10		
4	12.857				

Table 11: High-Elevation Forest Model Depth Specific Soil Layer Parameters

Depth	Percent Gravel (%)	Percent Organic Matter (%)	Porosity	Depth	Percent Gravel (%)	Percent Organic Matter (%)	Porosity
0	0	28	0.9	3.261	0	0	0.38
0.061			0.74	3.511			0.38
0.111			0.6	3.761			0.38
0.161			0.59	4.011			0.38
0.211			0.58	4.261			0.38
0.261			0.57	4.511			0.38
0.311			0.56	4.761			0.37
0.361			0.54	5.011			0.37
0.411			0.53	5.261			0.37
0.461			0.52	5.511			0.37
0.511	30	0	0.5	5.761	0	0	0.37
0.561			0.49	6.011			0.37
0.611			0.47	6.261			0.36
0.661			0.45	6.511			0.36
0.711			0.43	6.761			0.36
0.761			0.42	7.011			0.36
0.811			0.4	7.261			0.36
0.861			0.4	7.511			0.36
0.911			0.4	7.761			0.35
0.961			0.4	8.011			0.35
1.011	0	0	0.4	8.261	0	0	0.35
1.261			0.4	8.511			0.35
1.511			0.4	8.761			0.35
1.761			0.39	9.011			0.35
2.011			0.39	9.261			0.34
2.261			0.39	9.511			0.34
2.511			0.39	9.761			0.34
2.761			0.39	9.95			0.34
3.011			0.39	10			0.34

Table 12: Wetland Model Depth Specific Soil Layer Parameters

Depth	Percent Organic Matter (%)	Porosity	Depth	Percent Organic Matter (%)	Porosity
0	100	0.97	1.466	100	0.889
0.066		0.968	1.516		0.886
0.116		0.967	1.566		0.882
0.166		0.965	1.616		0.879
0.216		0.963	1.666		0.875
0.266		0.962	1.716		0.871
0.316		0.96	1.766		0.868
0.366		0.958	1.816		0.864
0.416		0.957	1.866		0.861
0.466		0.955	1.916		0.857
0.516		0.953	1.966		0.854
0.566		0.952	2		0.85
0.616		0.95	2.5	75	0.82
0.666		0.946	3	50	0.79
0.716		0.943	3.5	25	0.76
0.766		0.939	4	0	0.73
0.816		0.936	4.5		0.7
0.866		0.932	5		0.67
0.916		0.929	5.5		0.64
0.966		0.925	6		0.61
1.016		0.921	6.5		0.58
1.066		0.918	7		0.55
1.116		0.914	7.5		0.52
1.166		0.911	8		0.49
1.216		0.907	8.5		0.46
1.266		0.904	9		0.43
1.316		0.9	9.5		0.4
1.366		0.896	9.95		0.37
1.416		0.893	10		0.34

Table 13: Soil Moisture & Temperature Observations Used in Model Calibration & Validation.

Model	Wetland	High-elevation Forest		Low-elevation Forest		
Source	SMT1	SMT4		SMT6		
Calibration Date Range	July 2024 – July 2025					
Validation Date Range	July 2025 – March 2026					
Sampling Frequency	15 minutes (Hourly)					
LWC Depths (mbgs)	0.166	0.111		0.116		
	0.316	0.261		0.266		
	0.466	0.411		0.416		
	0.616	0.56		0.566		
	0.766					
Temperature Depths (mbgs)	0.041	0.491	0.00	0.30	0.00	0.30
	0.141	0.591	0.10	0.45	0.10	0.45
	0.241	0.691	0.15	0.60	0.15	0.60
	0.341	0.791				
	0.391	0.841				

Table 14: Groundwater Level Observations Used in Model Calibration & Validation.

Model	Wetland	High-elevation Forest
Source	PNC	MW1A
Calibration Date Range	July 2024 – July 2025	
Validation Date Range	July 2025 – November 2025	
Sampling Frequency	15 minutes (Hourly)	
Depth Range (mbgs)	0.661 - 3.261	0.266 – 0.516

Table 15: Snow Depth Observations for Model Calibration & Validation.

Model	Wetland	High-elevation Forest	Low-elevation Forest
Source	SDS-Wetland	SDS-Forest	SDS-ECCC
Calibration Date Range	2023-2025 winter seasons	2025 winter season	2015-2024 winter seasons
Validation Date Range	2026 winter season	2026 winter season	2025 & 2026 winter seasons
Sampling Frequency	Hourly (Daily)	Hourly (Daily)	Daily

Table 16: PEST Adjustable Node Depths for Each SHAW Model.

PEST Adjustable Node Depths (mbgs)		
Wetland	High-Elevation Forest	Low-Elevation Forest
0	0	0
0.616	0.211	0.566
2	0.411	0.966
10	0.661	6.25
	10	9.00
		10.0

Table 17: Wetland Model Calibration Objective Function Weights

Objective Type	Depth (mbgs)	Weight
Soil LWC	0.166	0.1
	0.316	0.5
	0.366	1.5
	0.416	1.5
	0.466	1.5
	0.516	1.25
	0.566	1.25
	0.616	1.25
	0.666	1.25
	0.716	1.25
	0.766	1.25
Soil Temperature	0.041	2.00
	0.141	1.00
	0.241	1.00
	0.341	1.00
	0.391	1.00
	0.491	1.00
	0.591	1.00
	0.691	1.00
	0.791	1.00
	0.841	1.00

Table 18: High-Elevation Forest Model Calibration Objective Function Weights

Objective Type	Depth (mbgs)	Weight
Soil LWC	0.111	4.00
	0.261	4.00
	0.411	4.00
	0.561	4.00
	0.661	1.00
	0.711	1.00
	0.761	1.00
	0.811	1.00
	0.861	1.00
	0.911	1.00
	0.961	1.00
	1.011	1.00
	1.261	1.00
	1.511	1.00
	1.761	1.00
	2.011	1.00
	2.261	1.00
	2.511	1.00
	2.761	1.00
	3.011	1.00
	3.261	1.00
Soil Temperature	0.00	1.10
	0.10	0.50
	0.15	0.50
	0.30	0.50
	0.45	0.50
	0.60	0.50

Table 19: Low-Elevation Forest Model Calibration Objective Function Weights

Objective Type	Depth (mbgs)	Weight
Soil LWC	0.116	1.50
	0.266	1.50
	0.416	1.50
	0.566	1.50
Soil Temperature	0.00	1.00
	0.10	1.00
	0.15	1.00
	0.30	1.00
	0.45	1.00
	0.60	1.00

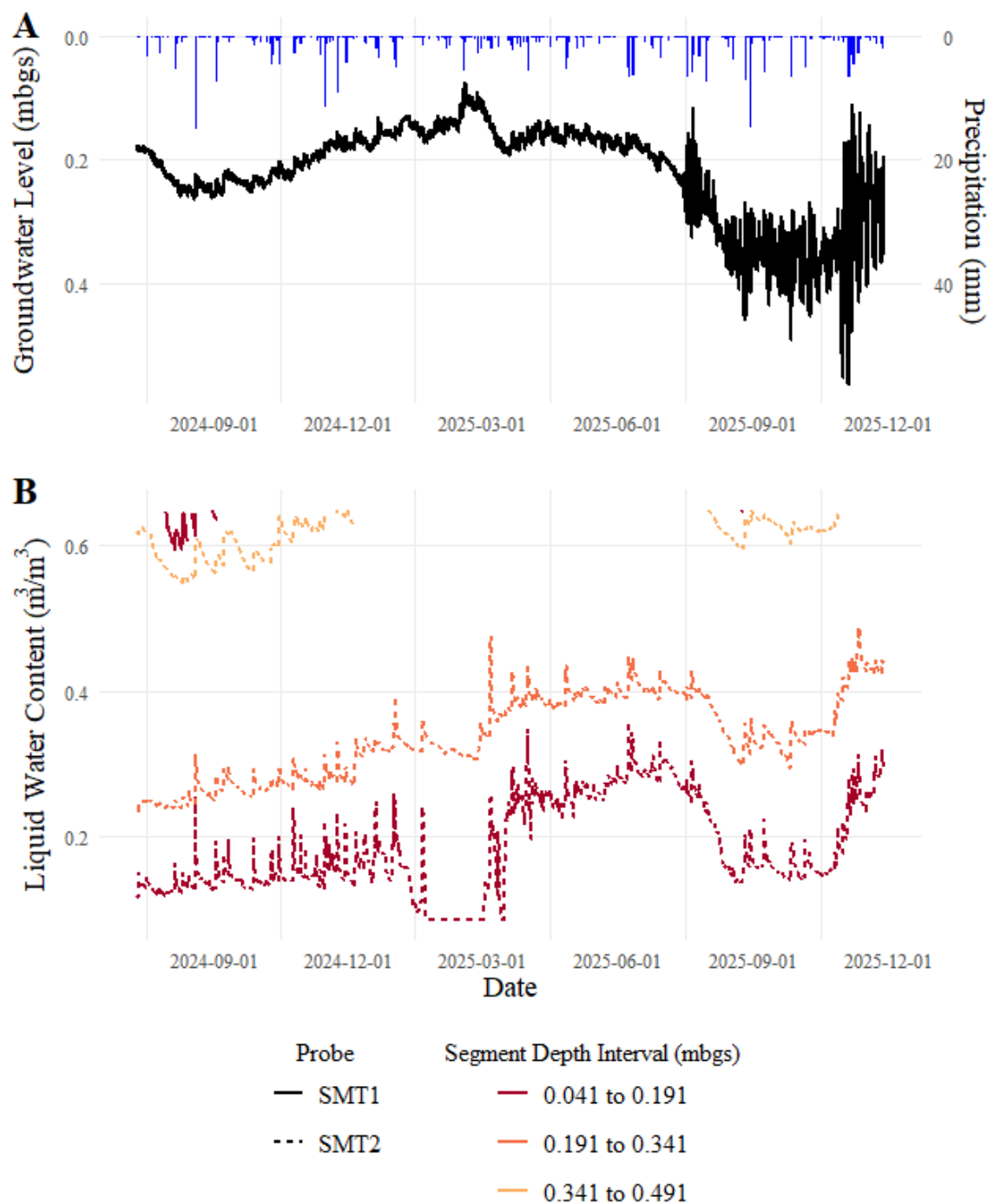


Figure 32: Wetland Site observations of (A) PNC groundwater levels and precipitation, (B) soil LWC from SMT1 and SMT2 (LWC observations above 0.657 were removed as they exceeded the instrument measurement limits).

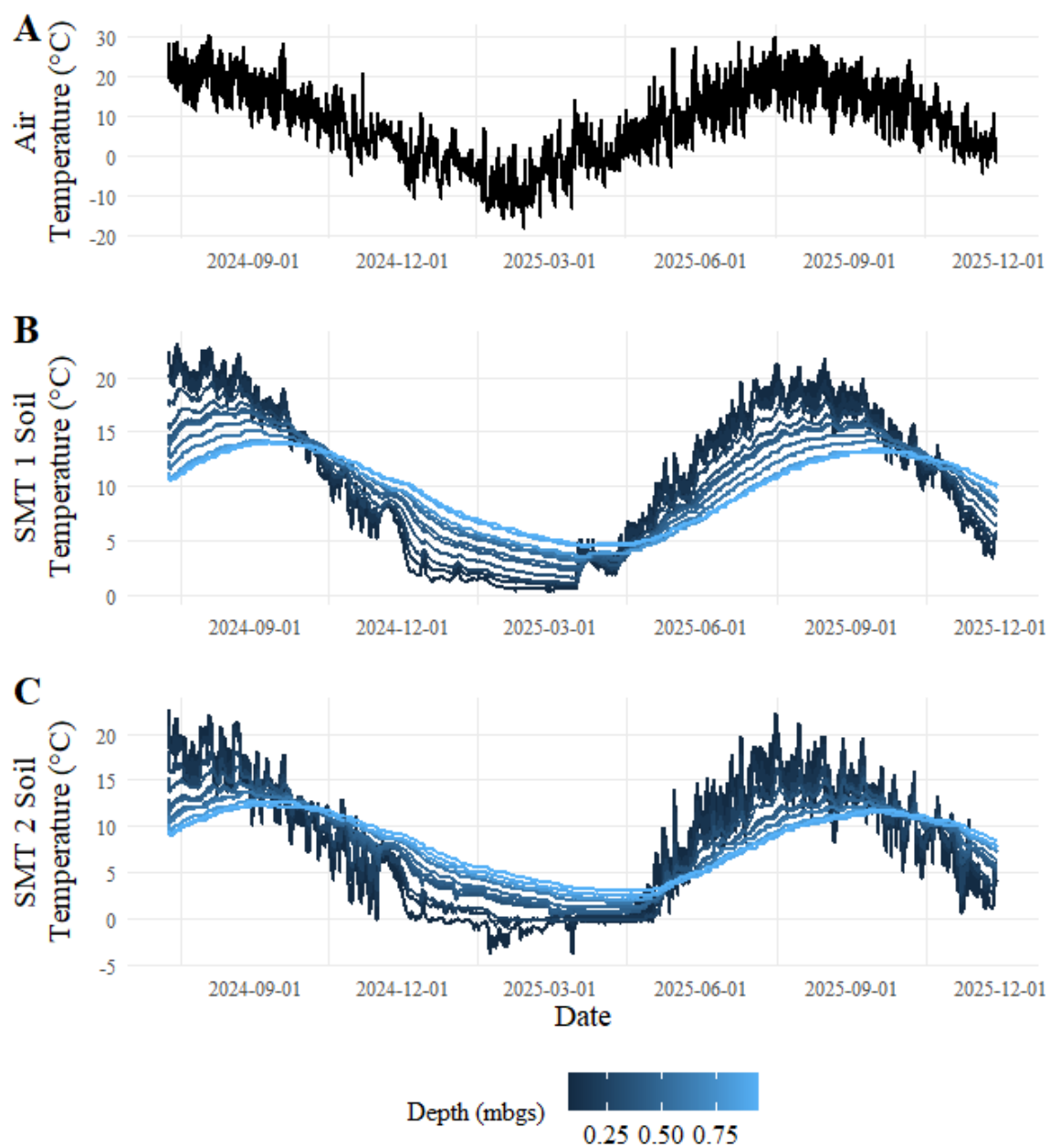


Figure 33: Wetland site observations of (A) air temperature, (B) SMT1 soil temperature, and (C) SMT 2 soil temperature.

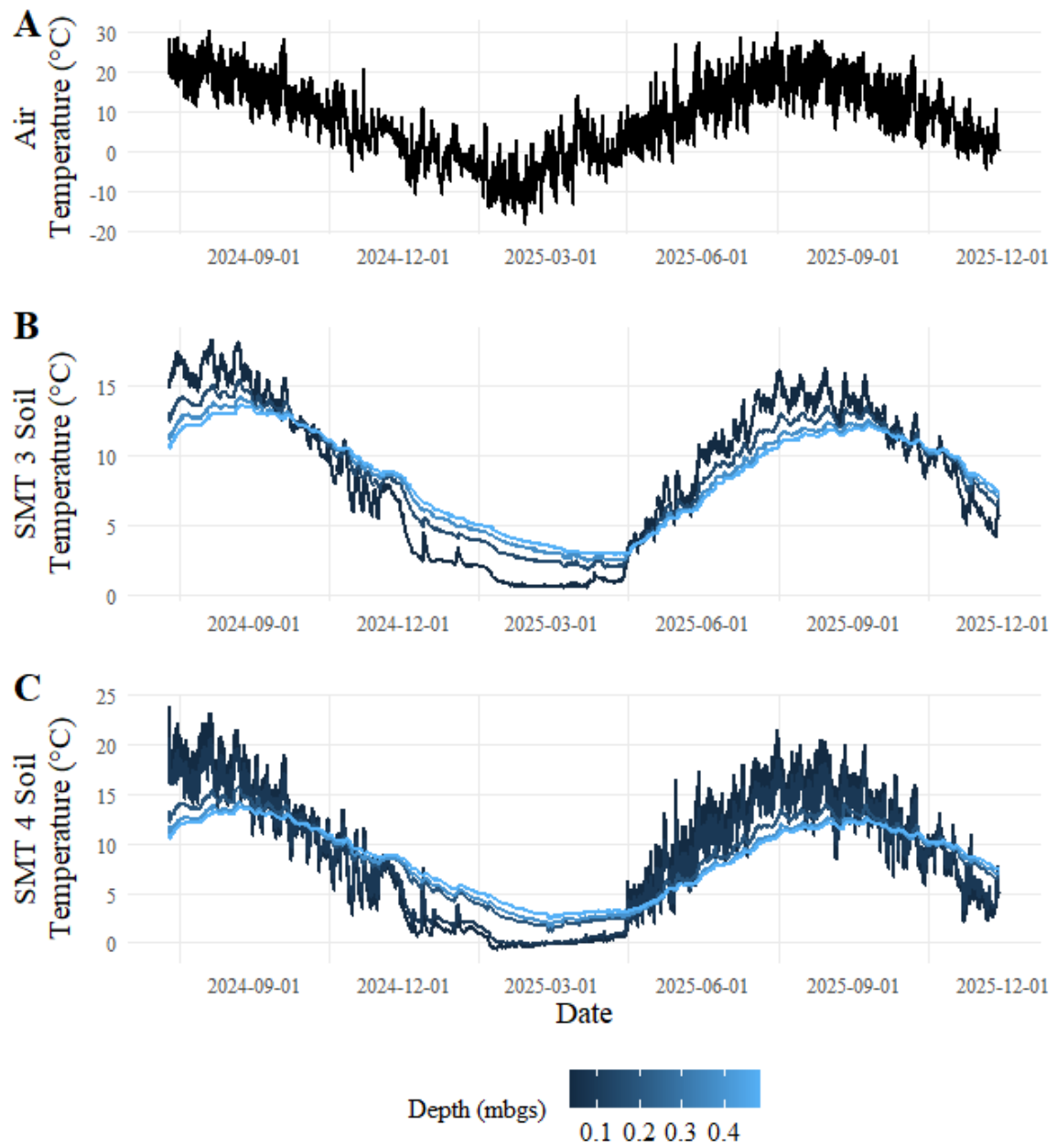


Figure 35: High-Elevation Forest site observations of (A) air temperature, (B) SMT3 soil temperature, and (C) SMT4 soil temperature.

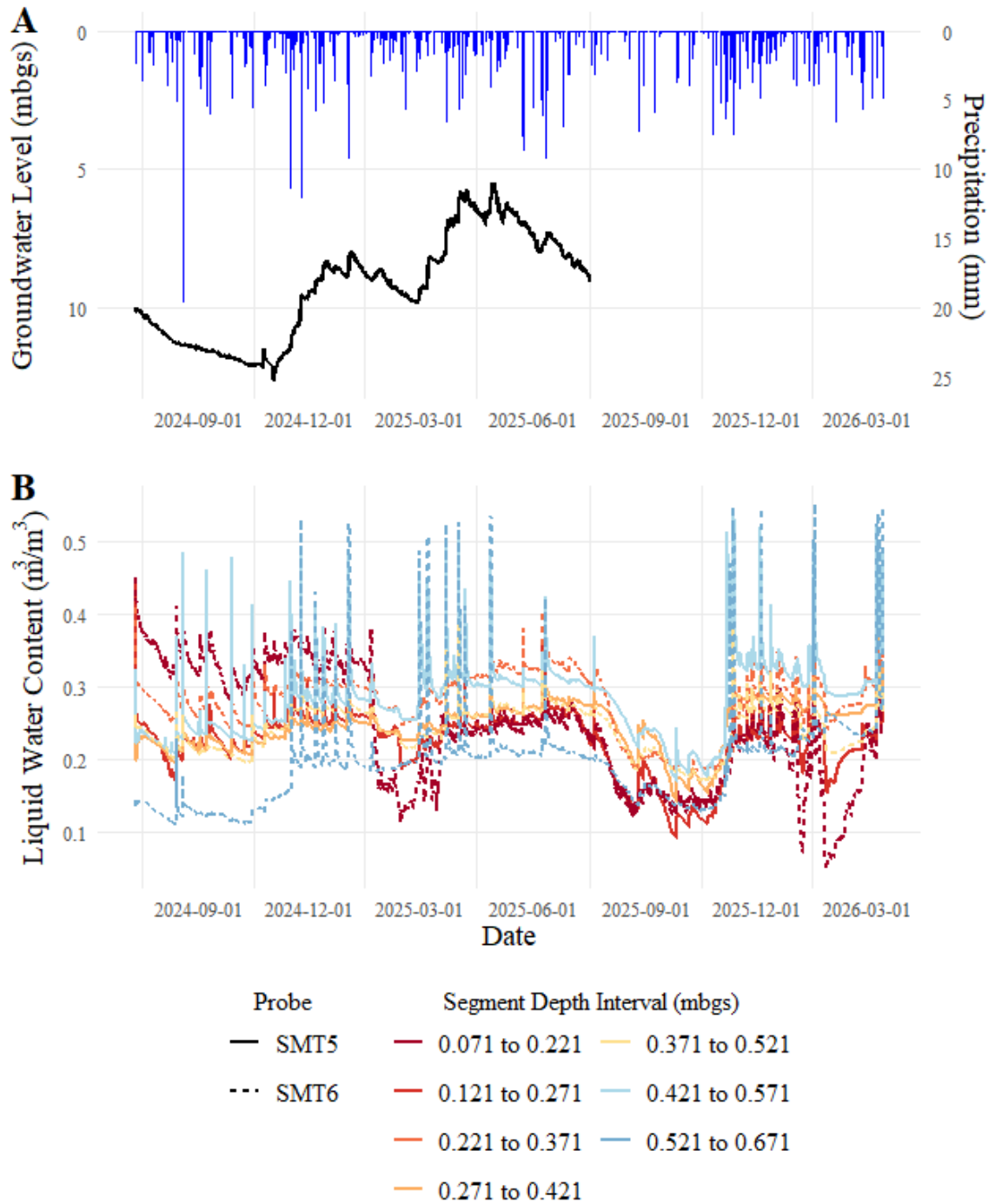


Figure 36: Low-Elevation Forest site observations of (A) MW4 groundwater levels and precipitation, (B) Soil LWC from SMT5 and SMT6 (MW4 has pumping well influences; MW4 GWL observations were recorded for the entire period but unavailable at the time of plotting).

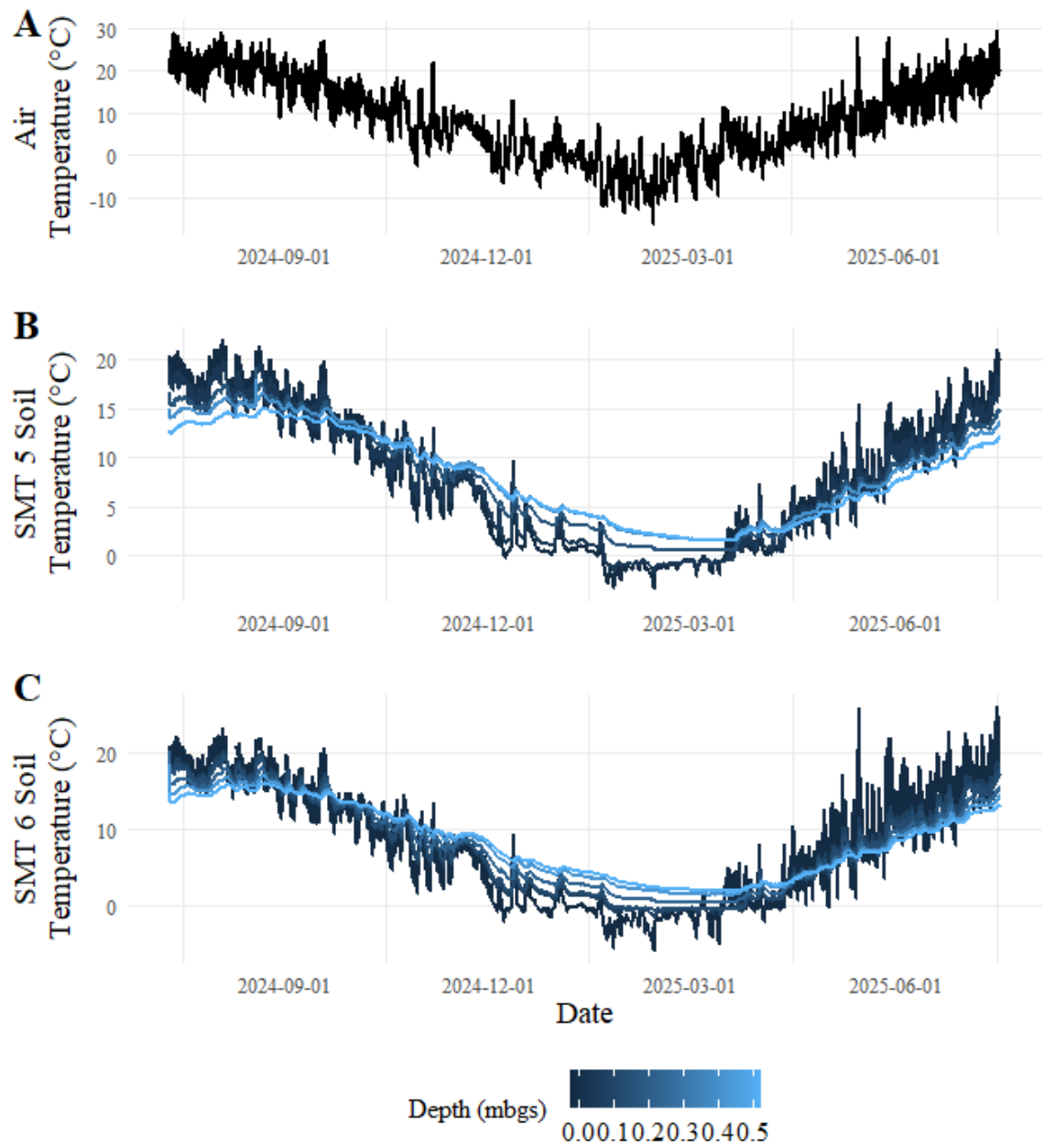


Figure 37: Low-Elevation Forest site observations of (A) air temperature, (B) SMT5 soil temperature, and (C) SMT6 soil temperature.

Table 20: Wetland Model Calibrated Parameters.

Parameter Name	Soil Node Depth (m)	Calibration Type	Lower Limit	Upper Limit	Calibrated Value	Units
SNW_TEMP^a	-	Snow	-0.4	2.4	-0.1	°C
LEAF_OR^b	-	Snow	0	5	4	-
P_CLUMP^b	-	Snow	0	1	0.5	-
P_ALB^{c,d,e}	-	Snow	0.092	0.35	0.273	-
P_INT^{b,f}	-	Snow	0	2.7	0.25	mm
S_SR^g	-	Snow	2	3	2	cm
KZ^{h,i,j,k}	0.000	Soil	0.0127	45	45	cm/hr
	0.616		0.0127	20	0.05	
	2.000		0.0127	20	0.1	
	10.00		0.0001	10	5E-3	
AEP^{i,j,l,m}	0.000	Soil	-0.1	-0.001	-0.001	m
	0.616		-0.1	-0.001	-0.05	
	2.000		-0.1	-0.001	-0.004	
	10.00		-0.13	-0.006	-0.13	
B^{n,o,p}	0.000	Soil	1	8	4.02	-
	0.616		1	8	8	
	2.000		1	8	7.62	
	10.00		1	18	8	
DCHAR^q	-	Soil	0.8	2.5	2.50	cm
ROOT_D^{q*}	-	Soil	0.1	0.5	0.32	mbgs
LAI^{r,s}	-	Soil	3.5	10	8.00	-
MX_PND^{t,u,v}	-	Soil	0	0.5	.179	cm

*Rooting depth based on the minimum and maximum acrotelm depth since sphagnum moss has no roots. Upper and lower limit sources: ^a(Jennings et al., 2018), ^b(Flerchinger, 2000), ^c(Ohkubo et al., 2021), ^d(Thompson et al., 2015), ^e(Berglund & Mace Jr., 1976), ^f(Porada et al., 2018), ^g(Kellner, 2001), ^h(Menberu et al., 2021), ⁱ(Agriculture and Agri-Food Canada, 2019a), ^j(Katsura et al., 2006), ^k(EXP Services Ltd., 2011b), ^l(Rouxel et al., 2012), ^m(McCarter et al., 2023), ⁿ(Rawls et al., 1992), ^o(Pittaki-Chrysodonta et al., 2018), ^p(Timlin et al., 2004), ^q(Hayward & Clymo, 1982), ^r(Garisoain et al., 2023), ^s(Thielen et al., 2021), ^t(Xu et al., 2019), ^u(Tsihrintzis & Hamid, 1998), ^v(Krebs et al., 2013)

Table 21: Acceptable Error Metric Ranges for SHAW models.

Calibration Type	Applicable Error Metric	Calibrated Value Ranges	Source
Soil liquid water content	RMSE (m ³ /m ³)	0.003 to 0.081	(Chen et al., 2019)
		0.040 to 0.090	(Flerchinger et al., 2003)
		0.010 to 0.084	(Flerchinger et al., 2006)
		0.005 to 0.096	(Li et al., 2012)
		0.020 to 0.050	(Strong et al., 2025)
Soil temperature	RMSE (°C)	0.101 to 2.139	(Chen et al., 2019)
		1.10 to 3.96	(Flerchinger et al., 2003)
		0.5 to 2.2	(Li et al., 2012)
		0.61 to 1.57	(Strong et al., 2025)
Snow Depth	NSE	≥ 0.5	(Li et al., 2024; Moriasi et al., 2007)
	PBIAS	≤ 25	(Moriasi et al., 2007)

Table 22: High-Elevation Forest Model Calibrated Parameters.

Parameter Name	Soil Node Depth (m)	Calibration Type	Lower Limit	Upper Limit	Calibrated Value	Units
SNW_TEMP^a	-	Snow	-0.4	2.4	0.0	°C
LEAF_OR^b	-	Snow	0	5	1.0	-
P_CLUMP^b	-	Snow	0	1	0.6	-
S_SR^b	-	Snow	0.1	10	2.0	cm
WIN_LAI	-	Snow	0	SUM_LAI	3.0	-
KZ^{c,d,e,f}	0	Soil	0.25	50	23.03	cm/hr
	0.211		0.25	21	21.00	
	0.411		0.25	21	21.00	
	0.661		0.0001	21	0.15	
	10		0.0001	5	4.35	
AEP^{g,h}	0	Soil	-0.35	-0.001	-0.001	m
	0.211		-0.35	-0.001	-0.295	
	0.411		-0.64	-0.001	-0.627	
	0.661		-0.35	-0.001	-0.009	
	10		-0.130	-0.060	-0.007	
B^{g,i,j}	0	Soil	1	6	2.30	-
	0.211		1	6	1.52	
	0.411		1	6	4.25	
	0.661		1	18	18.00	
	10		1	18	6.35	
SUM_LAI^{k,l}	-	Soil	0.1	6.00	6.00	-
P_ALB^{m,n,o,p}	-	Both	0.1	0.9	0.36	-
P_INT^{b,q,r,s}	-	Both	0	1.5	0.25	mm

Upper and lower limit sources: ^a(Jennings et al., 2018), ^b(Flerchinger, 2000), ^c(Freeze & Cherry, 1979), ^d(Agriculture and Agri-Food Canada, 2019b), ^e(Katsura et al., 2006), ^f(EXP Services Ltd., 2011b), ^g(Rawls et al., 1982), ^h(Rouxel et al., 2012), ⁱ(Rawls et al., 1992), ^j(Pittaki-Chrysodonta et al., 2018), ^k(Brown et al., 2021), ^l(Chen et al., 2006), ^m(Sakai et al., 1997), ⁿ(Rautiainen et al., 2012 & 2009), ^o(Hari et al., 2013), ^p(Lukeš et al., 2013), ^q(Hadiwijaya et al., 2021), ^r(Alivio et al., 2023), ^s(Anys & Weiler, 2024)

Table 23: Low-Elevation Forest Model Calibrated Parameters.

Parameter Name	Soil Node Depth (m)	Calibration Type	Lower Limit	Upper Limit	Calibrated Value	Units
SNW_TEMP^a	-	Snow	-0.4	2.4	0.35	°C
LEAF_OR^b	-	Snow	0	5	5.00	-
P_CLUMP^b	-	Snow	0	1	0.285	-
S_SR^b	-	Snow	0.1	10	2.00	cm
WIN_LAI	-	Snow	0	SUM_LAI	5.06	-
KZ^{c,d,e,f}	0	Soil	0.25	50	46.64	cm/hr
	0.216		0.25	21	21.00	
	0.566		0.25	21	17.41	
	0.966		0.25	21	20.44	
	9.000		0.0001	21	1.54E-2	
	10.00		0.0001	10	0.864	
AEP^{g,h}	0	Soil	-0.35	-0.001	-0.112	m
	0.216		-0.35	-0.001	-0.350	
	0.566		-0.35	-0.001	-0.350	
	0.966		-0.35	-0.001	-0.350	
	9.000		-0.35	-0.001	-0.022	
	10.00		-0.130	-0.060	-0.082	
Bg^{i,j}	0	Soil	1	6	6.00	-
	0.216		1	6	6.00	
	0.566		1	6	4.12	
	0.966		1	6	4.78	
	9.000		1	18	1.00	
	10.00		1	18	3.29	
PORO^{c,d,g,k,l,m}	0	Soil	0.3	0.9	0.549	m ³ /m ³
	0.216		0.3	0.5	0.500	
	0.566		0.3	0.5	0.359	
	0.966		0.2	0.5	0.357	
	6.250		0.3	0.7	0.651	
	9.000		0.2	0.5	0.441	
	10.00		0.34	0.57	0.522	
SUM LAI^{n,o}	-	Soil	0.1	6.00	5.06	-
P_ALB^{p,q,r,s}	-	Both	0.1	0.9	0.163	-
P_INT^{b,t,u,v}	-	Both	0	1.5	0.500	mm

Upper and lower limit sources: ^a(Jennings et al., 2018), ^b(Flerchinger, 2000), ^c(Freeze & Cherry, 1979), ^d(Agriculture and Agri-Food Canada, 2019b), ^e(Katsura et al., 2006), ^f(EXP Services Ltd., 2011b), ^g(Rawls et al., 1982), ^h(Rouxel et al., 2012), ⁱ(Rawls et al., 1992), ^j(Pittaki-Chrysodonta et al., 2018), ^k(Nolan, Davis & Associates Limited, 1984), ^l(Woessner & Poeter, 2020), ^m(Zhang et al., 2011), ⁿ(Brown et al., 2021), ^o(Chen et al., 2006), ^p(Sakai et al., 1997), ^q(Rautiainen et al., 2012 & 2009), ^r(Hari et al., 2013), ^s(Lukeš et al., 2013), ^t(Hadiwijaya et al., 2021), ^u(Alivio et al., 2023), ^v(Anys & Weiler, 2024)

Table 24: Wetland Site Calibration & Validation Soil LWC & Temperature Performance.

Observation Type	Depth (mbgs)	Calibration RMSE	Validation RMSE	Units
Soil temperature	0.041	2.009	2.376	°C
	0.141	2.267	2.862	
	0.241	1.964	2.463	
	0.341	1.608	2.299	
	0.391	1.45	2.105	
	0.491	1.615	2.323	
	0.591	1.235	1.792	
	0.691	1.286	1.844	
	0.791	1.091	1.503	
	0.841	0.917	1.247	
Soil liquid water content	0.166	0.228	0.158	m ³ /m ³
	0.316	0.158	0.046	
	0.466	0.135	0.165	
	0.616	0.061	0.114	
	0.766	0.038	0.086	

Table 25: High-Elevation Forest Model Soil LWC & Temperature Calibration & Validation Performance.

Observation Type	Depth (mbgs)	Calibration RMSE	Validation RMSE	Units
Soil temperature	0.111	1.155	1.319	°C
	0.161	0.887	1.063	
	0.311	1.218	0.875	
	0.461	1.284	1.008	
	0.611	1.289	0.989	
Soil liquid water content	0.111	0.057	0.085	m ³ /m ³
	0.261	0.051	0.052	
	0.411	0.044	0.058	
	0.561	0.045	0.035	
	0.661	0.017	0.006	
	0.711	0.024	0.013	
	0.761	0.028	0.014	
	0.811	0.029	0.015	
	0.861	0.031	0.016	
	0.911	0.029	0.017	
	0.961	0.028	0.017	
	1.011	0.027	0.017	
	1.261	0.031	0.022	
	1.511	0.034	0.025	
	1.761	0.039	0.033	
	2.011	0.04	0.035	
	2.261	0.041	0.038	
	2.511	0.042	0.040	
	2.761	0.045	0.044	
	3.011	0.046	0.047	
	3.261	0.052	0.056	

Table 26: Low-Elevation Forest Model Soil LWC & Temperature Calibration & Validation Performance.

Observation Type	Depth (mbgs)	Calibration RMSE	Validation RMSE	Units
Soil temperature	0.10	2.485	3.213	°C
	0.15	2.141	2.285	
	0.30	1.861	2.157	
	0.45	1.722	2.044	
	0.60	1.552	1.895	
Soil liquid water content	0.116	0.059	0.105	m ³ /m ³
	0.266	0.023	0.030	
	0.416	0.020	0.030	
	0.566	0.046	0.047	

Table 27: Coefficient of Determination of Simulated Snow Density & the Sampling Event Median Observed Snow Density for each Site.

Model	R ²	Median Snow Density (%)
Wetland	-0.6	25.4
High-Elevation Forest	-0.5	25.8
Low-Elevation Forest	-1.1	27.6

Table 28: Wetland Model Historical Sensitivity Analysis 10 Most Sensitive Parameter Changes

Output	Parameter	Direction of Change	SI	Output	Parameter	Direction of Change	SI
Median Annual Recharge	ksat_58	Decrease	0.71	Snow Duration	lai	Increase	0.11
	lai	Increase	0.22		lai	Decrease	0.05
	leaf_or	Increase	0.2		alb	Increase	0.04
	p_clump	Decrease	0.19		p_clump	Increase	0.03
	b41	Increase	0.15		b1	Increase	0.03
	alb	Decrease	0.08		b41	Increase	0.03
	b1	Increase	0.08		p_clump	Decrease	0.02
	ksat_1	Decrease	0.07		alb	Decrease	0.02
	p_clump	Increase	0.07		dchar	Decrease	0.02
Median Annual Recharge Efficiency	leaf_or	Decrease	0.06	Average Frost Depth	b1	Decrease	0.02
	b41	Increase	2.01		lai	Decrease	0.48
	ksat_58	Decrease	0.64		b1	Decrease	0.42
	lai	Increase	0.51		p_clump	Decrease	0.34
	alb	Increase	0.47		p_clump	Increase	0.27
	leaf_or	Increase	0.42		airent13	Increase	0.19
	lai	Decrease	0.23		leaf_or	Increase	0.18
	rootdp	Decrease	0.21		lai	Increase	0.18
	b13	Decrease	0.19		alb	Increase	0.16
Median Snow Depth	b1	Decrease	0.18	Frost Duration	int	Decrease	0.14
	p_clump	Increase	0.18		ksat_58	Decrease	0.14
	b41	Increase	0.84		b41	Increase	0.14
	lai	Increase	0.2		lai	Decrease	0.09
	leaf_or	Increase	0.17		lai	Increase	0.07
	alb	Increase	0.16		ksat_58	Decrease	0.06
	rootdp	Decrease	0.07		alb	Increase	0.05
	rootdp	Increase	0.07		b1	Increase	0.05
	alb	Decrease	0.06		p_clump	Decrease	0.04
	dchar	Decrease	0.06		p_clump	Increase	0.04
	ksat_13	Decrease	0.06		b1	Decrease	0.03
	airent41	Increase	0.06		leaf_or	Increase	0.03

Table 29: Wetland Model SSP2-4.5 2055 – 2064 Sensitivity Analysis 10 Most Sensitive Parameter Changes

Output	Parameter	Direction of Change	SI	Output	Parameter	Direction of Change	SI
Median Annual Recharge	ksat_58	Decrease	0.85	Snow Duration	b41	Increase	0.42
	p_clump	Decrease	0.05		lai	Decrease	0.12
	lai	Increase	0.05		b1	Decrease	0.11
	b41	Increase	0.04		p_clump	Decrease	0.09
	lai	Decrease	0.03		leaf_or	Increase	0.09
	b1	Decrease	0.03		p_clump	Increase	0.09
	p_clump	Increase	0.03		alb	Increase	0.07
	alb	Increase	0.03		alb	Decrease	0.06
	int	Decrease	0.02		ksat_58	Decrease	0.05
	dchar	Decrease	0.02		lai	Increase	0.05
Median Annual Recharge Efficiency	ksat_58	Decrease	0.85	Median Frost Depth	p_clump	Increase	0.61
	p_clump	Decrease	0.05		snw_temp	Decrease	0.42
	lai	Increase	0.05		lai	Increase	0.27
	b41	Increase	0.04		alb	Increase	0.24
	lai	Decrease	0.03		p_clump	Decrease	0.2
	b1	Decrease	0.03		lai	Decrease	0.06
	p_clump	Increase	0.03		rootdp	Increase	0.06
	alb	Increase	0.03		alb	Decrease	0.05
	int	Decrease	0.02		b1	Decrease	0.05
	dchar	Decrease	0.02		leaf_or	Decrease	0.03
Median Snow Depth	b41	Increase	1	Frost Duration	lai	Increase	0.17
	lai	Decrease	0.18		b41	Increase	0.14
	alb	Increase	0.18		p_clump	Decrease	0.12
	leaf_or	Increase	0.15		lai	Decrease	0.1
	p_clump	Decrease	0.12		p_clump	Increase	0.1
	p_clump	Increase	0.11		b1	Increase	0.05
	rootdp	Increase	0.09		alb	Decrease	0.03
	b1	Increase	0.08		b1	Decrease	0.03
	rootdp	Decrease	0.07		ksat_58	Decrease	0.02
	b1	Decrease	0.07		snw_temp	Decrease	0.01

Table 30: High-Elevation Forest Model Historical Sensitivity Analysis 10 Most Sensitive Parameter Changes

Output	Parameter	Direction of Change	SI	Output	Parameter	Direction of Change	SI
Median Annual Recharge	sum_lai	Decrease	0.1	Snow Duration	airent9	Decrease	0.55
	int	Decrease	0.07		win_lai	Decrease	0.08
	alb	Increase	0.07		p_clump	Decrease	0.07
	win_lai	Decrease	0.06		ksat_58	Increase	0.07
	p_clump	Increase	0.06		airent58	Increase	0.07
	p_clump	Decrease	0.05		leaf_or	Decrease	0.06
	win_lai	Increase	0.05		airent5	Decrease	0.06
	int	Increase	0.05		p_clump	Increase	0.06
	b9	Increase	0.05		alb	Decrease	0.05
	alb	Decrease	0.04		alb	Increase	0.05
Median Annual Recharge Efficiency	sum_lai	Decrease	0.09	Median Frost Depth	airent9	Decrease	17.71
	win_lai	Decrease	0.06		ksat_58	Increase	2.35
	int	Decrease	0.06		airent58	Increase	2.11
	p_clump	Increase	0.05		airent5	Decrease	1.88
	win_lai	Increase	0.05		b5	Decrease	1.03
	alb	Increase	0.05		b9	Increase	0.86
	int	Increase	0.05		b14	Decrease	0.5
	b9	Increase	0.05		b9	Decrease	0.46
	airent9	Increase	0.04		airent58	Decrease	0.44
	p_clump	Decrease	0.03		b1	Decrease	0.42
Median Snow Depth	airent9	Decrease	2.87	Frost Duration	airent9	Decrease	2.46
	ksat_58	Increase	0.38		ksat_58	Increase	0.41
	airent5	Decrease	0.36		b1	Decrease	0.38
	airent58	Increase	0.34		b5	Decrease	0.32
	win_lai	Increase	0.16		airent58	Increase	0.3
	p_clump	Increase	0.15		airent5	Increase	0.22
	alb	Increase	0.15		airent9	Increase	0.21
	b9	Increase	0.14		win_lai	Decrease	0.19
	b5	Decrease	0.13		p_clump	Increase	0.15
	win_lai	Decrease	0.09		alb	Decrease	0.12

**Table 31: High-Elevation Forest Model SSP2-4.5 2055 – 2064 Sensitivity Analysis
10 Most Sensitive Parameter Changes**

Output	Parameter	Direction of Change	SI	Output	Parameter	Direction of Change	SI
Median Annual Recharge	b58	Decrease	0.41	Snow Duration	airent9	Decrease	0.34
	sum_lai	Decrease	0.11		b58	Decrease	0.19
	airent5	Decrease	0.09		airent5	Decrease	0.17
	p_clump	Decrease	0.08		alb	Increase	0.17
	alb	Increase	0.07		alb	Decrease	0.08
	win_lai	Decrease	0.06		win_lai	Increase	0.08
	int	Decrease	0.06		b1	Decrease	0.07
	airent9	Increase	0.06		b9	Increase	0.07
	alb	Decrease	0.05		win_lai	Decrease	0.06
	b5	Decrease	0.05		sum_lai	Decrease	0.06
Median Annual Recharge Efficiency	b58	Decrease	0.4	Median Frost Depth	airent58	Increase	0.52
	airent9	Decrease	0.11		airent5	Decrease	0.46
	sum_lai	Decrease	0.1		b1	Increase	0.32
	int	Increase	0.07		int	Decrease	0.26
	p_clump	Decrease	0.06		p_clump	Increase	0.26
	win_lai	Decrease	0.06		b5	Decrease	0.25
	int	Decrease	0.06		alb	Increase	0.23
	p_clump	Increase	0.06		b9	Increase	0.21
	win_lai	Increase	0.06		win_lai	Decrease	0.18
	alb	Increase	0.06		sum_lai	Decrease	0.18
Median Snow Depth	airent9	Decrease	0.91	Frost Duration	airent9	Decrease	0.89
	b58	Decrease	0.28		win_lai	Decrease	0.26
	int	Decrease	0.2		airent9	Increase	0.23
	win_lai	Increase	0.2		alb	Decrease	0.18
	airent5	Decrease	0.19		p_clump	Decrease	0.16
	b5	Decrease	0.16		b1	Decrease	0.16
	p_clump	Increase	0.16		airent5	Increase	0.15
	p_clump	Decrease	0.13		ksat_58	Increase	0.12
	ksat_58	Increase	0.12		b5	Decrease	0.11
	airent58	Increase	0.11		airent58	Increase	0.11

Table 32: Low-Elevation Forest Model Historical Sensitivity Analysis 10 Most Sensitive Parameter Changes

Output	Parameter	Direction of Change	SI	Output	Parameter	Direction of Change	SI
Median Annual Recharge	ksat_1	Increase	0.55	Snow Duration	ksat_13	Increase	0.57
	win_lai	Decrease	0.48		p35	Increase	0.52
	sum_lai	Increase	0.37		ksat_1	Increase	0.27
	int	Decrease	0.32		sum_lai	Increase	0.21
	p_clump	Increase	0.24		p51	Increase	0.2
	ksat_9	Increase	0.19		snw_temp	Decrease	0.19
	b1	Decrease	0.18		p46	Increase	0.14
	win_lai	Increase	0.18		alb	Decrease	0.11
	int	Increase	0.17		p9	Decrease	0.11
	alb	Decrease	0.16		p9	Increase	0.11
Median Annual Recharge Efficiency	ksat_13	Increase	8.72	Median Frost Depth	ksat_1	Increase	1.21
	ksat_1	Increase	3.78		win_lai	Decrease	1.02
	p35	Increase	3.23		p51	Increase	0.94
	p51	Increase	2.61		p1	Increase	0.75
	p46	Increase	1.79		p_clump	Decrease	0.65
	sum_lai	Increase	1.62		p46	Increase	0.65
	p9	Decrease	1.43		p1	Decrease	0.48
	ksat_9	Increase	1.33		win_lai	Increase	0.47
	b13	Increase	1.04		p13	Decrease	0.4
	airent51	Increase	0.89		airent51	Increase	0.32
Median Snow Depth	ksat_13	Increase	4.69	Frost Duration	ksat_13	Increase	0.46
	p35	Increase	1.72		p_clump	Increase	0.17
	p51	Increase	1.4		b1	Decrease	0.15
	ksat_1	Increase	1.19		ksat_1	Increase	0.15
	p46	Increase	0.95		p46	Increase	0.14
	p9	Decrease	0.87		p35	Increase	0.13
	sum_lai	Increase	0.61		p51	Increase	0.13
	ksat_9	Increase	0.55		p_clump	Decrease	0.11
	win_lai	Increase	0.54		p1	Decrease	0.07
	airent51	Increase	0.48		airent1	Increase	0.06

**Table 33: Low-Elevation Forest Model SSP2-4.5 2055 – 2064 Sensitivity Analysis 10
Most Sensitive Parameter Changes**

Output	Parameter	Direction of Change	SI	Output	Parameter	Direction of Change	SI
Median Annual Recharge	win_lai	Decrease	0.22	Snow Duration	ksat_13	Increase	9.97
	int	Decrease	0.19		p51	Increase	4.52
	win_lai	Increase	0.19		ksat_1	Increase	3.79
	int	Increase	0.18		p35	Increase	3.78
	p_clump	Increase	0.17		p46	Increase	3.08
	sum_lai	Increase	0.15		sum_lai	Increase	2.23
	sum_lai	Decrease	0.13		win_lai	Increase	2.05
	p_clump	Decrease	0.09		p9	Decrease	1.66
	p1	Increase	0.09		airent51	Increase	1.53
	ksat_13	Increase	0.08		ksat_9	Increase	1.35
Median Annual Recharge Efficiency	ksat_13	Increase	1.39	Median Frost Depth	ksat_1	Increase	9.56
	ksat_1	Increase	0.54		p35	Increase	9.25
	p35	Increase	0.48		ksat_13	Increase	5.03
	p51	Increase	0.44		win_lai	Increase	4.07
	win_lai	Decrease	0.33		p9	Decrease	3.77
	p46	Increase	0.29		sum_lai	Increase	3.33
	int	Decrease	0.25		p1	Increase	2.37
	p9	Decrease	0.24		p46	Increase	2.05
	ksat_9	Increase	0.2		p35	Decrease	1.98
	sum_lai	Decrease	0.18		p51	Decrease	1.98
Median Snow Depth	ksat_13	Increase	1.82	Frost Duration	ksat_13	Increase	3.04
	p35	Increase	1.34		ksat_1	Increase	1.31
	ksat_1	Increase	0.69		p51	Increase	1.05
	sum_lai	Increase	0.54		p35	Increase	0.93
	p51	Increase	0.54		p46	Increase	0.75
	p46	Increase	0.37		sum_lai	Increase	0.56
	b1	Decrease	0.33		p9	Decrease	0.51
	p9	Decrease	0.3		b13	Increase	0.45
	int	Increase	0.26		ksat_9	Increase	0.41
	p9	Increase	0.26		p4	Decrease	0.38

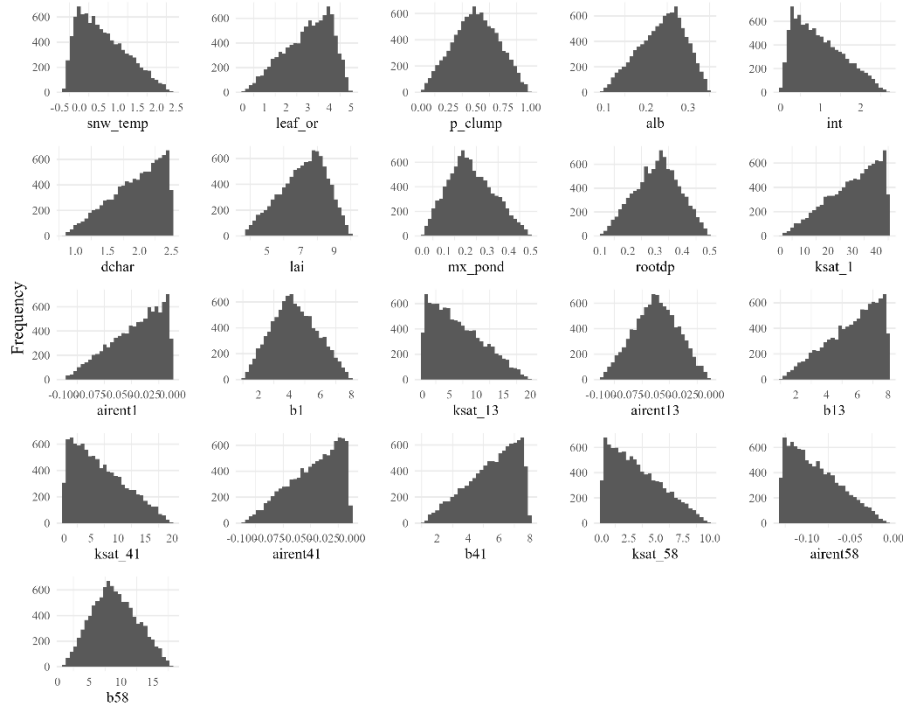


Figure 38: : Wetland uncertainty analysis parameter distributions.

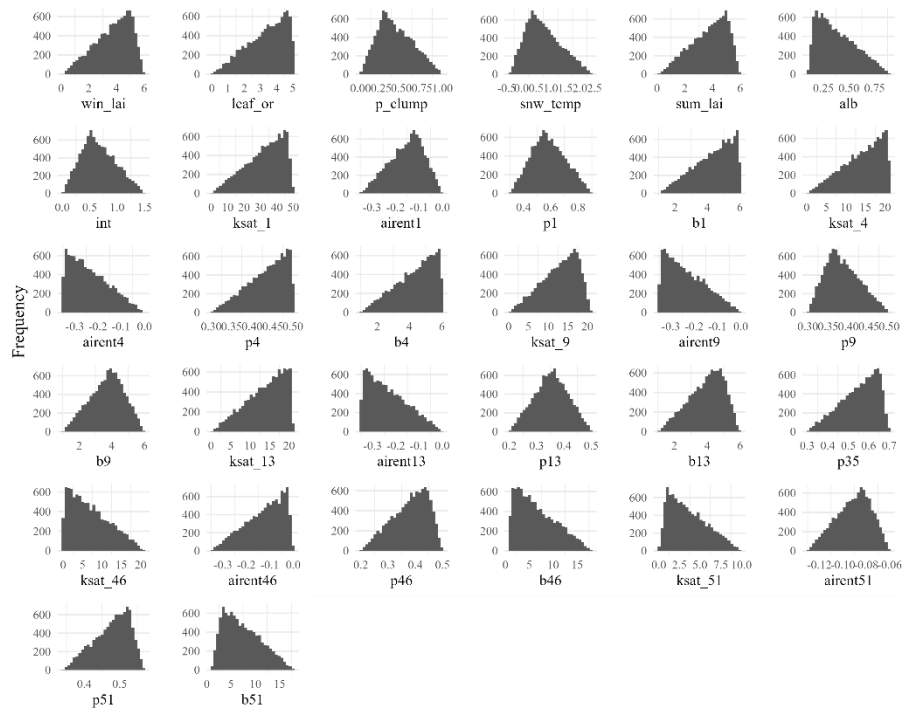


Figure 39: High-Elevation Forest uncertainty analysis parameter distributions.

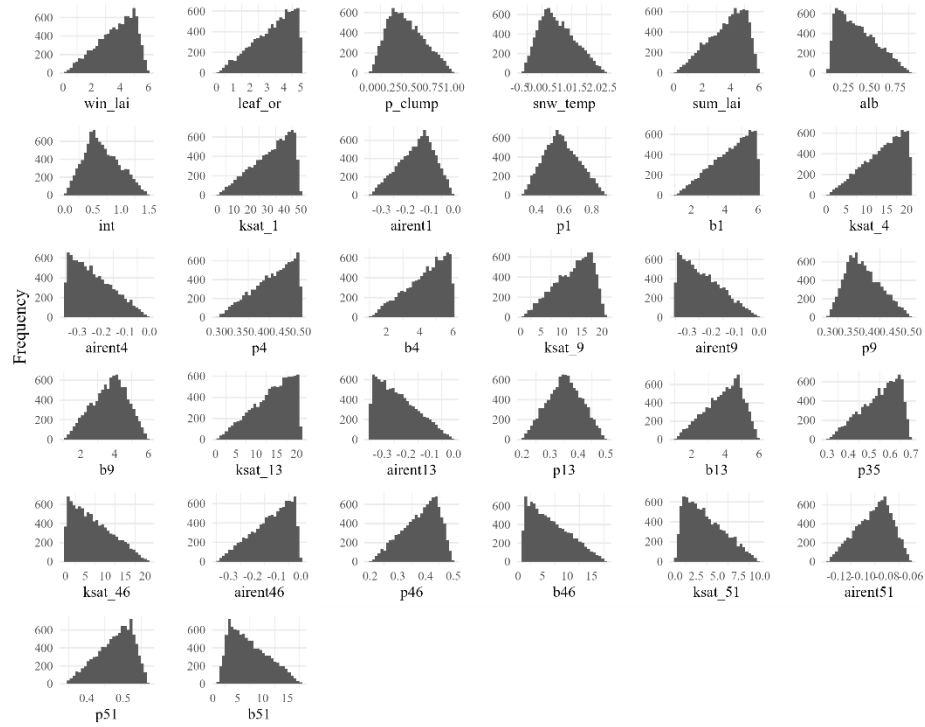


Figure 40: Low-Elevation Forest uncertainty analysis parameter distributions.

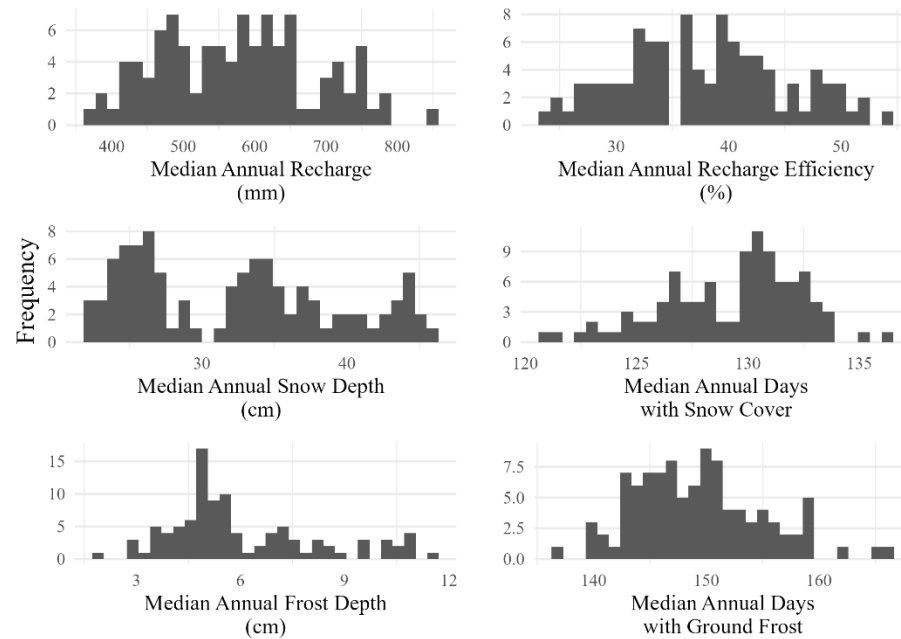


Figure 41: : Histogram of the Wetland historical model 100 Monte Carlo simulations.

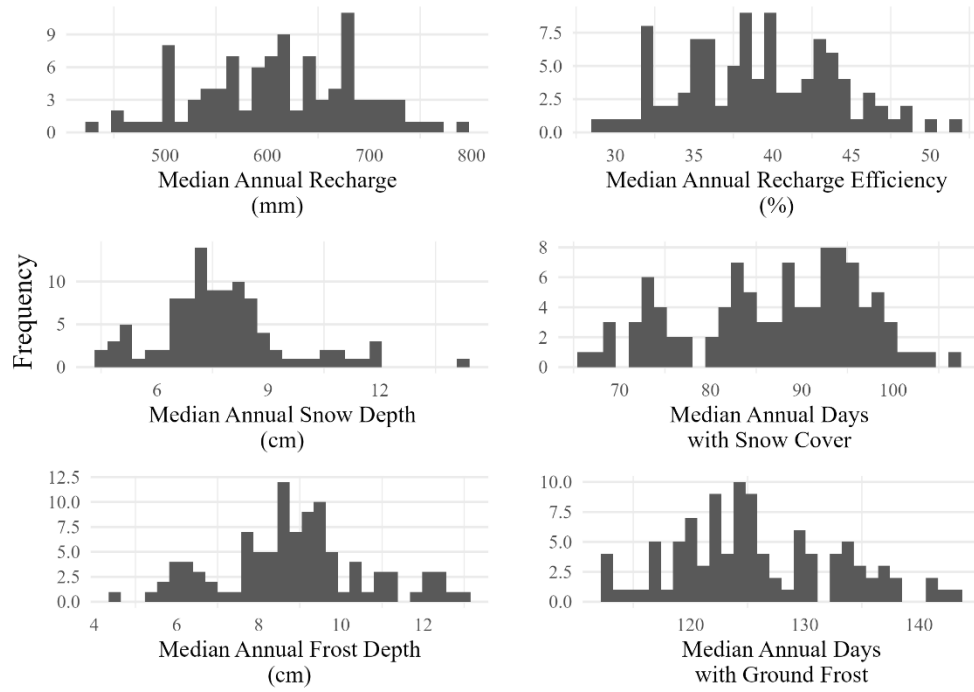


Figure 42: Histogram of the Wetland SSP2-4.5 2055-2064 model 100 Monte Carlo simulations.

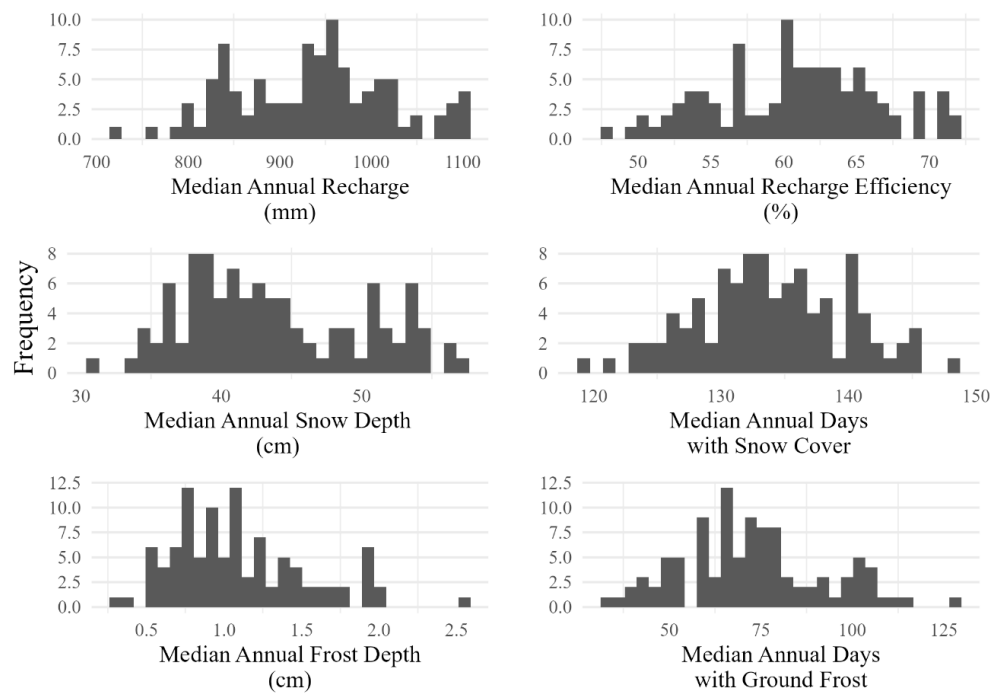


Figure 43: Histograms of the High-Elevation Forest Historical model 100 Monte Carlo simulations.

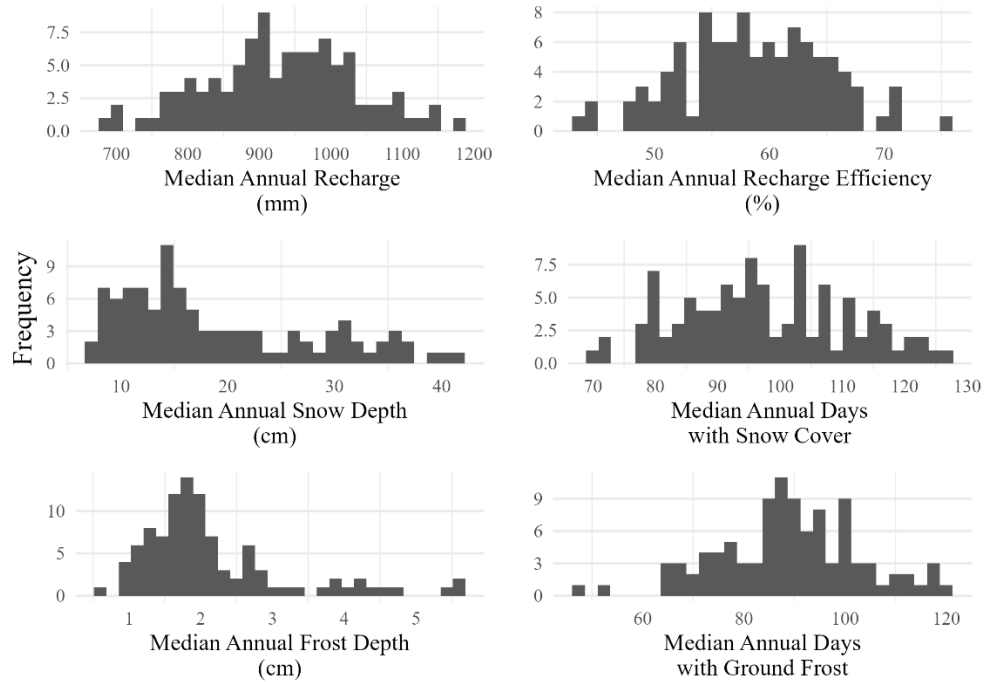


Figure 44: Histograms of the High-Elevation Forest SSP2-4.5 2055-2064 model 100 Monte Carlo simulations.

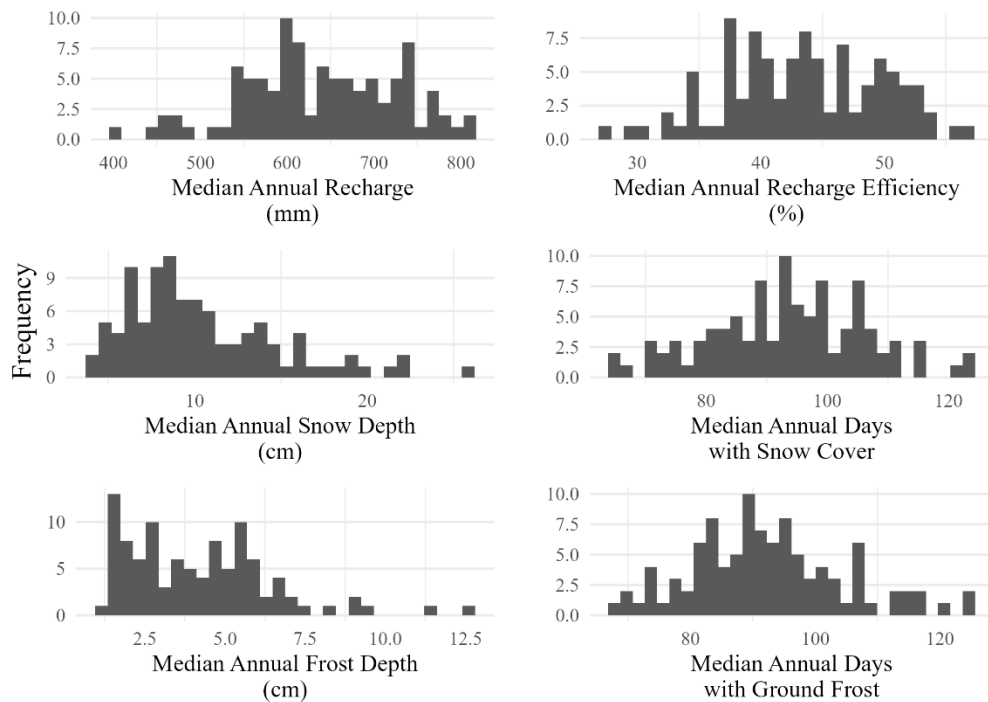


Figure 45: Histograms of the Low-Elevation Forest Historical model 100 Monte Carlo simulations.

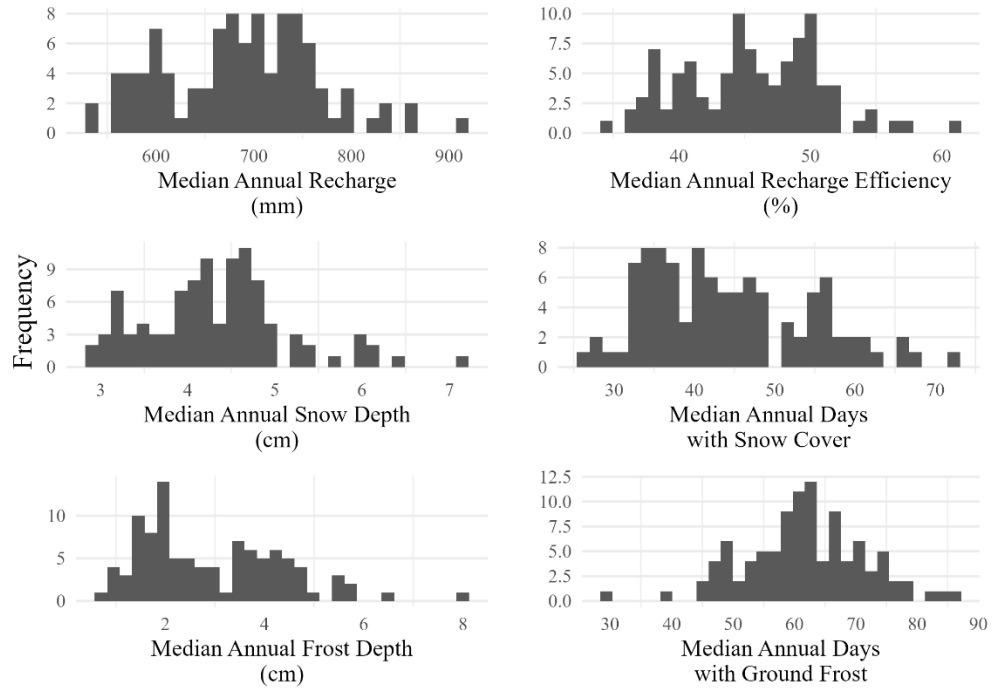


Figure 46: Histograms of the Low-Elevation Forest SSP2-4.5 2055-2064 model 100 Monte Carlo simulations.