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Landscape controls on lake-peatland-forestland hydrology within the Boreal Plains

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ABSTRACT

Shallow lakes provide the world's richest bird communities, and provide major critical habitat in the adjacent wetlands, forests and downstream ecosystems that they support. Their resilience is strongly linked to stability of water levels, yet shallow lakes are diminishing in the face of climatic warming and local land use pressures, which will modify hydrological connectivity and negatively impact vital habitats. However, the patterns and controls on shallow lake water levels are poorly understood in relation to their hydrogeoclimatic setting and long-term weather cycles. This PhD thesis investigates how climate, geology, landcover and landscape position interact to control spatiotemporal variability of shallow lake water depth. Findings are used to inform a conceptual framework for defining landscape controls on boreal hydrology based upon mechanistic understanding within a complex heterogenous hydrogeoclimatic Boreal Plains (BP) setting. This is achieved through empirical and numerical evidence of the patterns and controls on 20 years of lake, peatland and forestland water tables in 26 study catchments. The primary research outcomes are: (1) BP lake water levels exhibit low synchronicity as high and spatially variable landscape storage of water produces high interannual variability; (2) surficial geology incorporates important landscape parameters that define the source, magnitude, and continuity of hydrological connectivity and interactions with land cover and lake geometry further modify lake water level patterns; (3) spatiotemporal variation in landscape memory of climate, driven by differences in storage and antecedent conditions, can cause wet or dry climatic conditions to propagate into extreme events; (4) reduced-complexity modelling founded upon HRAs and HUs provide sufficient functionality to accurately simulate multi-decadal, interannual and seasonal lake-peatland-forestland connectivity and water table patterns, and provides a valuable tool to transpose knowledge to ungauged catchments and evaluate spatial variability of hydrological sensitivity to climate. This research provides significant new knowledge of the dominant controls on boreal lake, peatland and forestland hydrological interactions of diverse hydrogeoclimatic settings. It characterises spatial differences in lake water level responses to seasonal and decadal climate cycles

and demonstrates the role of storage-driven landscape memory in driving asynchronous spatiotemporal variability. Empirical and simulated results have strengthened conceptual understanding of the HRA-HU framework for defining hydrologic landscapes and yields crucial insights and model development for evaluating hydrological sensitivity to climate.

This thesis is lovingly dedicated in memory of my father,

Lyndon Probert

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CHAPTER ONE: INTRODUCTION

1. Boreal Plains: A Lake-Wetland-Forestland Mosaic

Glacial lakes constitute almost half of Earth's lake surface area and represent extensive and diverse wetland and peatland environments where shallow lakes form behind glacial moraines, glacial outwash or in kettle holes and infill with terrigenous sediments (Cohen, 2003). Despite their omnipresence across mid-latitudinal belts, and representing the most abundant lake type globally, shallow lake extent has been historically overlooked (Meerhoff and Jeppesen, 2009; p645). Shallow lakes exhibit greater productivity, fish biomass and species richness per unit area than deeper lakes (Bayley and Prather, 2003) and the expansive adjacent wetlands, forests and downstream ecosystems that they support (Ireson *et al.,* 2015; Leibowitz, Mushet, and Newton, 2016) provide extensive crucial habitat for mammals, aquatic organisms and the world's richest bird communities (Macdonald *et al.,* 2006; Morissette *et al.,* 2018).

The distribution of shallow lakes coincides with regions enduring greater non-stationary climatic forcing (Stralberg *et al.,* 2020) and changes to precipitation timing, phase, and magnitude (Tetzlaff *et al.,* 2013), in regions already experiencing high evaporation:inflow ratios (Fergus *et al.,* 2020), in addition to accelerating anthropogenic disturbance (Webster *et al.,* 2015). This is particularly of consequence in Canada's Boreal Plains (BP); climate pressures are being compounded by the large footprint of oil and gas extraction, mining, agriculture and timber harvesting (AEP 1998), and lake water levels are predicted to deplete over the next century and are at risk of permanent terrestrialization (Thompson, Mendoza, and Devito, 2017; Bertassello *et al.,* 2019). Subsequent modifications to hydrologic processes, hydrological connectivity and nutrient and mineral fluxes are set to have critical consequences for downstream water supply (Ireson *et al.,* 2015), water quality, wildlife (Leibowitz, Mushet, and Newton, 2016; Webster *et al.,* 2015) and ecosystem state (Cobbaert, Wong, and Bayley, 2014; Scheffer *et al.,* 2001).

Historically, the BP represents a system of substantial resilience to climate change, with high ecological stability over the past 6000 years (Chapin *et al.,* 2010). As such, shallow lakes are ubiquitous throughout the BP of Canada. However, maximum ecological sensitivity is anticipated in the 21st century (Ireson *et al.,* 2015) as the Canadian Boreal is projected to warm between 2-5°C by 2050 compared to mean temperatures in 2000 (Price *et al.,* 2013). Significant drying has already been observed since the 1950s (Wang *et al.,* 2014) with each of the past three decades being successively warmer than previous records (IPCC, 2013). Furthermore, boreal and arctic soil carbon stocks are orders of magnitude more sensitive to climatic warming than any other biome (Laudon *et al.,* 2017), yet with one third of global terrestrial carbon stored within boreal forests (Bradshaw and Warkentin, 2015), they play a critical role in regulating global water and carbon cycles and in turn, global climate (Kurz *et al.,* 2013; Lemprière *et al.,* 2013).

This unprecedented climatic warming, coupled with catchment disturbances could exceed thresholds of hydrological function in the BP, and is set to modify hydrological connectivity, feedbacks, and interactions, and could result in abrupt and unexpected shift in ecosystem function (Johnstone *et al.,* 2010; Laudon *et al.,* 2017; Scheffer and Carpenter, 2003; Wang *et al.,* 2014), owing to strong coupling with ecological and biogeochemical and processes (Waddington *et al.,* 2009). Frequent water deficits, where annual potential evapotranspiration (PET) typically exceeds annual precipitation (P) (Devito, Creed, and Fraser, 2005) with pronounced seasonal and decadal cycles of water availability (Mwale *et al.,* 2009) makes the BP highly sensitive to modifications to the water budget (Brown *et al.,* 2014; Thompson *et al.,* 2015). However, warming will affect all components of the hydrological cycle (Laudon *et al.,* 2017). Shorter snow-covered seasons and subsequent longer temperature-driven growing seasons will increase evaporative demand, resulting in more substantial water deficits and subsequent falling water tables (Waddington *et al.,* 2009; Brown *et al.,* 2010; Thompson, Mendoza, and Devito, 2017). Increases in evaporative demand under warmer temperatures will significantly outstrip projected increases in precipitation (Wang *et al.,* 2014). Groundwater recharge will diminish as spring

melts reduce in magnitude, and the timing and magnitude of precipitation is set to shift to the growing season during peak ET (Thompson, Mendoza, and Devito, 2017), causing reduced soil water storage and severe soil moisture deficits in these regions.

By nature of their lower volumes, shallow lakes also represent the most sensitive lake ecosystems to water level fluctuations (Kolding and van Zwieten 2012), and are more susceptible to catastrophic shifts in ecosystem states, whereas deep lakes gradually respond to environmental pressures (Scheffer *et al.,* 2001). Stability of water levels is considered a strong driver of lake resilience (Bayley and Prather, 2003; Scheffer *et al.,* 2001; Cobbaert, Wong, and Bayley, 2014), which is supported through landscape storage of water (Carey *et al.,* 2010). Storage provides a crucial self-sustaining mechanism that facilitates both substantial water transmission in wet periods and water retention and landscape scale redistribution during dry periods between water sources and sinks (Devito, Mendoza, and Qualizza, 2012). The water sink-source dependency evident within boreal systems highlights the importance of hydrological connectivity between landscape units (Wells, Ketcheson, and Price, 2017) for maintaining the mosaic of lakes, peatlands and forestlands.

Forecasting the impacts of future climate change on these boreal lakes is imperative to highlight important zones for conservation, or support landscape restoration within long-term management (Thompson, Mendoza and Devito, 2017), yet this challenge represents substantial complexity due to the hydrogeo-climatic setting of the BP. The BP sub-humid climate operates with a long-term water deficit, with wet years occurring on a 10–15 year cycle (Ferone and Devito, 2004) which interacts with heterogeneous surficial geology. The BP also exhibits three broadly defined surficial geologies: coarse-textured glacio-fluvial and glacio-lacustrine deposits; fine-textured clay-rich hummocky moraines; and fine-textured clay-rich glacio-lacustrine plain deposits (Devito *et al.,* 2017). These contrasting glacial deposits determine spatial variations in surface-groundwater interactions and water table configurations, through their significant differences in storage and transmission properties (Plach *et*

al., 2016; Hokanson, Mendoza, and Devito, 2019), that result in landscape-scale wetting and drying cycles in response to seasonal and decadal climate signals (Devito *et al.,* 2012). The contrasting hydrological function of lakes, peatlands and forestlands is then superimposed as a mosaic upon these geological interactions with climate. Each land cover exhibits distinct differences in storage and actual ET (AET; Thompson *et al.,* 2015), in a system where the water table configuration does not mirror local topography, and unsaturated zone storage, vegetation water demand or ET, and vertical flow are the dominant hydrological fluxes, with minimal contribution of lateral flow to hillslope water balances (Devito *et al.,* 2005). However, large surface runoff contributions can occur infrequently (Devito, Creed, and Fraser, 2005), potentially occurring only once every 25 years (Redding and Devito, 2008). This complexity has been encompassed within a conceptual framework that compartmentalizes complex processes and spatial heterogeneity into meaningful and process orientated classifications (Carey *et al.,* 2010), or '*hydrologic landscapes'* (Winter, 2001), derived from conceptual syntheses of BP (Devito *et al.,* 2012). These include *Hydrological Response Areas* (HRAs) to represent spatial differences in surficial geology, and *Hydrolgical Units* (HUs) which represent surface landcovers of lakes, peatlands and forestlands. This conceptual representation of the BP is employed for is for this thesis, which is then tested at landscape scales and built upon to aid in accurate representation of dominant processes in numerical models, and inform the hierarchy of hydrological controls (Devito *et al.,* 2005), within a hydrologic context-composition-connectivity structure. Improved understanding of spatial variability in how interactive landscape controls govern the partitioning of precipitation into evapotranspiration, runoff, recharge or surface water storage will also provide greater capacity to manage resilient landscapes by recognising and conserving the mechanisms that facilitate long-term maintenance of lake levels and facilitate their recovery following dry periods.

2. Research Gaps

Our current understanding of landscape scale hydrology in the BP is a significant limitation for prediction of future system responses (Ireson *et al.,* 2015). Knowledge of ecohydrological and climatic feedbacks that shape the resilience of boreal forests has advanced significantly in recent years (Price *et al.,* 2005; Waddington *et al.,* 2009), however, this knowledge is yet to be fully applied and evaluated at landscape scales (Johnstone, Chapin, *et al.,* 2010; Spence, 2010). Moreover, the prospect of novel landscapes developing under future hydro-geoclimatic settings may demand understanding of HRA and HU configurations not currently observed (Schneider *et al.,* 2016; Thompson, Mendoza, and Devito, 2017). The hydrological complexities, dominance of storage and lack of topographic control on runoff within the BP (Devito *et al.,* 2005) limits the application of topographic index models (e.g., TOPMODEL; Wells, Ketcheson, and Price, 2017). However, understanding the processes governing hydrological interactions between boreal lakes, peatlands and forestlands and with the wider catchment is of increasing importance for current and future catchment management (Tetzlaff *et al.,* 2013; Webster *et al.,* 2015). The following points outline priority research areas identified and addressed within this thesis (detailed literature reviews presented in chapters 2-5):

Long term observations and characterisation of hydrologic variability including spatially heterogenous landscapes: Spatial differences in basin storage, water budget components and lake-landscape connectivity will produce spatially variable responses to decadal weather cycles, climate change and disturbance. While the impact of lake hydrology on ecosystem health is well studied, the patterns of shallow lake water levels themselves are poorly defined in relation to their hydrogeoclimatic setting and long-term weather cycles (Coops, Beklioglu, and Crisman, 2003). Studies incorporating heterogenous glacial deposit landforms have exposed spatially diverse hydrological patterns under the same climate signal (Cobbaert, Wong, and Bayley, 2015; Prepas *et al.,* 2001), yet it remains unclear how these signals will appear in lakes across decadal scales. The poor characterisation

of lake water patterns hinders catchment comparisons and regionalisation approaches, which is necessary to transpose scientific knowledge to ungauged catchments and predict and mitigate ecological and societal impacts of hydrological changes. Thus, frameworks for examining lake hydrology must capture climate, geology and topography (Devito *et al.,* 2005; Winter, 2001; Wolock, Winter, and Mcmahon, 2004) and address varying relative importance of topology, topography and typology controls across spatial and temporal scales (Buttle, 2006). Long term observations are vital for providing the foundation for this knowledge (Tetzlaff *et al.,* 2017), which has previously represented a significant obstacle to advancing scientific enquiry.

Interaction of geology, landcover and landscape position controls on the scales, magnitudes, sources and patterns hydrologic connectivity, and the subsequent spatiotemporal variability in shallow lake water levels: Research by Devito *et al.,* (2017) highlights the need for further testing of current conceptualisations by placing HUs in the context of HRAs as the magnitude and timing of impacts of catchment disturbance and climatic forcing on hydrologic function will vary considerably between different HRA-HU configurations. Transposing knowledge from site-specific studies to landscape scales has long been recognized as a major research challenge (Price *et al.,* 2005), and until recently, there has been little focus on the role of interaction between peatlands and forestlands and with the atmosphere, surface, and subsurface processes at these wider scales (Hwang *et al.,* 2018). A potential reason for this is the need for data collection and empirical studies of sufficient length to observe both seasonal and decadal climate variability, in a setting where storage antecedent moisture conditions is so crucial for hydrological functioning (Wells *et al.,* 2017), which is a limitation that has only recently been overcome. Consequently, much of the previous research exploring hydrology of the BP has been conducted at local scale and largely focused on individual or few hydrological processes (i.e., evapotranspiration, runoff, recharge). Contrastingly, many long-term, regional and global studies include diverse hydrogeologic settings (Ali *et al.,* 2015; Carey *et al.,* 2010; Tetzlaff *et al.,* 2015), however, given the first-order importance of climate on hydrologic setting (Devito *et al.,* 2005), the

challenge remains to unravel landscape controls from climatic controls. The BP of Canada provide the ideal study location to associate lake water level fluctuations to landform and weather patterns due to diverse lake types within the same climate signal.

Dynamic controls associated with storage – the importance of antecedent storage and climate memory: Runoff is non-linear, hysteretic, threshold-mediated, and connectivity-controlled process (Phillips, Spence, and Pomeroy, 2011; Spence, 2010). Water storage causes landscape hydrologic memory of antecedent conditions, which is an important control on threshold driven connectivity and watershed resilience (Sayama *et al.,* 2011). Understanding landscape hydrologic memory is important due to the disproportionate impact of memory on the occurrence and clustering of extreme events (Bunde *et al.,* 2005; Khanal *et al.,* 2019), and dramatic increase in the uncertainty of related hydrological processes (Koutsoyiannis and Montanari, 2007; Koutsoyiannis, 2005). Typically, non-linear response functions such as hysteresis are observed and studied at an event scale, such as rainfall-runoff, storage-discharge or soil water dynamics for one storm or across a single year (Elliott *et al.,* 1998; Cuenca, Stangel, and Kelly, 1997). However, long term interannual lake level fluctuations in some systems (Van Der Kamp, Keir, and Evans, 2008) highlights the need to understand multi-year climate memory. Lakes act as important storage bodies in the landscape and are indicative of the landscape response (hysteresis, non-linear, threshold) to climate. Lakes thus act as important test beds to examine core conceptualisations of hydrological memory that operate at a catchment scale over multi-annual to decadal time periods.

Reduced complexity modelling with lower parameterisation requirements for model adaptability and application to a wide range of landscape configurations: Low runoff responses within the BP may mask the overall impacts of disturbance and external forcings, necessitating exploration of storage dynamics to assess current and projected impacts of aspen harvesting (Carrera-Hernández *et al.,* 2011), wildfire (Johnstone *et al.,* 2010; Kettridge *et al.,* 2015; Lukenbach *et al.,* 2017), road construction

and oil and gas exploration (Strack *et al.,* 2019; Webster *et al.,* 2015), climate change (Dieleman, Branfireun, Mclaughlin, and Lindo, 2015; Ireson *et al.,* 2015; Mirus *et al.,* 2017; Price *et al.,* 2013; Schindler, 1998), and to predict the success of reclaimed oil mines (Elshorbagy *et al.,* 2005; Keshta, Elshorbagy, and Carey, 2009; Alam *et al.,* 2017). Examining system dynamics through numerical modelling enables assessment of how each hydrological process operates and fluctuates (Carrera-Hernández *et al.,* 2011). However, studies that have addressed hydrological processes holistically (e.g. water balances, system dynamics), often explore these dynamics within individual catchments and demand a high level of site specific calibration and parameterization to model numerically (Thompson *et al.,* 2015), thus limiting transferability to other catchments. One-dimensional property-specific models have enabled model transferability to different geologies and soil types (Carrera-Hernández, Smerdon, and Mendoza, 2012), however, this method simulates hydrologic units in isolation, thus discounting dynamic connectivity. Yet, advancements in storage-water balance models coupled with the increasing ability to quantify storage dynamics (Ali *et al.,* 2015; Mcnamara *et al.,* 2011; Sayama *et al.,* 2011; Spence, 2010) could be used to simulate long-term water-level dynamics (Price *et al.,* 2005) to address the challenge of modelling heterogeneous environments with lower parameterisation requirements if dominant controls and processes are represented in a reduced complexity model and conceptual framework.

3. Aim and Objectives

This thesis aims to determine how climate, geology, landcover and landscape position interact to control spatiotemporal variability of shallow lake water depth to inform a conceptual framework for landscape controls on boreal hydrology, based upon mechanistic understanding within a complex heterogenous hydrogeoclimatic BP setting. The following objectives were addressed in each research chapter:

Objective 1. Characterise lake water level fluctuations of the sub-humid boreal plain, addressing synchronicity of lake water levels spatially; the importance of intra-annual versus interannual variation; the impact of long-term water level fluctuations on hydrological thresholds that characterise habitat characteristics; whether surficial geology can be utilised for regionalisation of lakes (Chapter 2).

Objective 2. Test the controls on the spatial variation in lake water levels within decadal climate cycles, to evaluate hydrologic interaction of lakes with their catchment, geology and climate and to inform a framework based upon mechanistic understanding of lake water level (Chapter 3).

Objective 3. Evaluate how multi-year landscape memory of precipitation drives temporal variation in shallow lake water level patterns and how they vary spatially between the typical lake-landscape type variations of a heterogeneous sub-humid Boreal Plains landscape (Chapter 4).

Objective 4. Simulate hydrological function of multiple HRA-HU configurations and associated storage-connectivity dynamics within a versatile, reduced-complexity system dynamics numerical model to further test conceptual understanding of dominant simulated fluxes and landscape interactions and assess spatial differences in landscape runoff sensitivity to the long-term climate (Chapter 5).

4. Thesis Approach and Structure

Twenty-six shallow lake study catchments spanning diverse hydrogeologic settings, which have been the subject of extensive and investigative biogeochemical monitoring were focal research interest for this thesis. The thesis approach utilises long-term (20-year) records of climate, lake stage, peatland

and forestland water tables, hydrochemistry data (stable isotopes and major cations and anions) in addition to field and GIS based catchment pedological data collection. This is combined with a priori conceptual understanding of hydrologic function within boreal and peatland dominated landscapes to produce novel insights into BP hydrology to inform scientific knowledge, catchment management and landscape reclamation. Empirical evidence and knowledge gained from analysis is then applied numerically within a physically based model to further test conceptual understanding and provide insights under a wider range of climatic conditions.

This thesis is presented in research paper format, consisting of four original research chapters (Chapters 2-5) to address each of the thesis objectives. Each research chapter is self-contained, comprising of a literature review, methodology, analysis and discussion, and may be considered independently from other chapters, however, insights gained from preceding chapters are referenced within subsequent chapters (see Figure 1-1). Supplementary material is also provided for Chapters 2 to 5 to support relevant sections of the main text. This is presented in addition to a broad introduction (Chapter 1) and synthesis of research chapters (Chapter 6) which summarises and contextualises main and novel findings, discusses the research implications and outlines future research avenues.

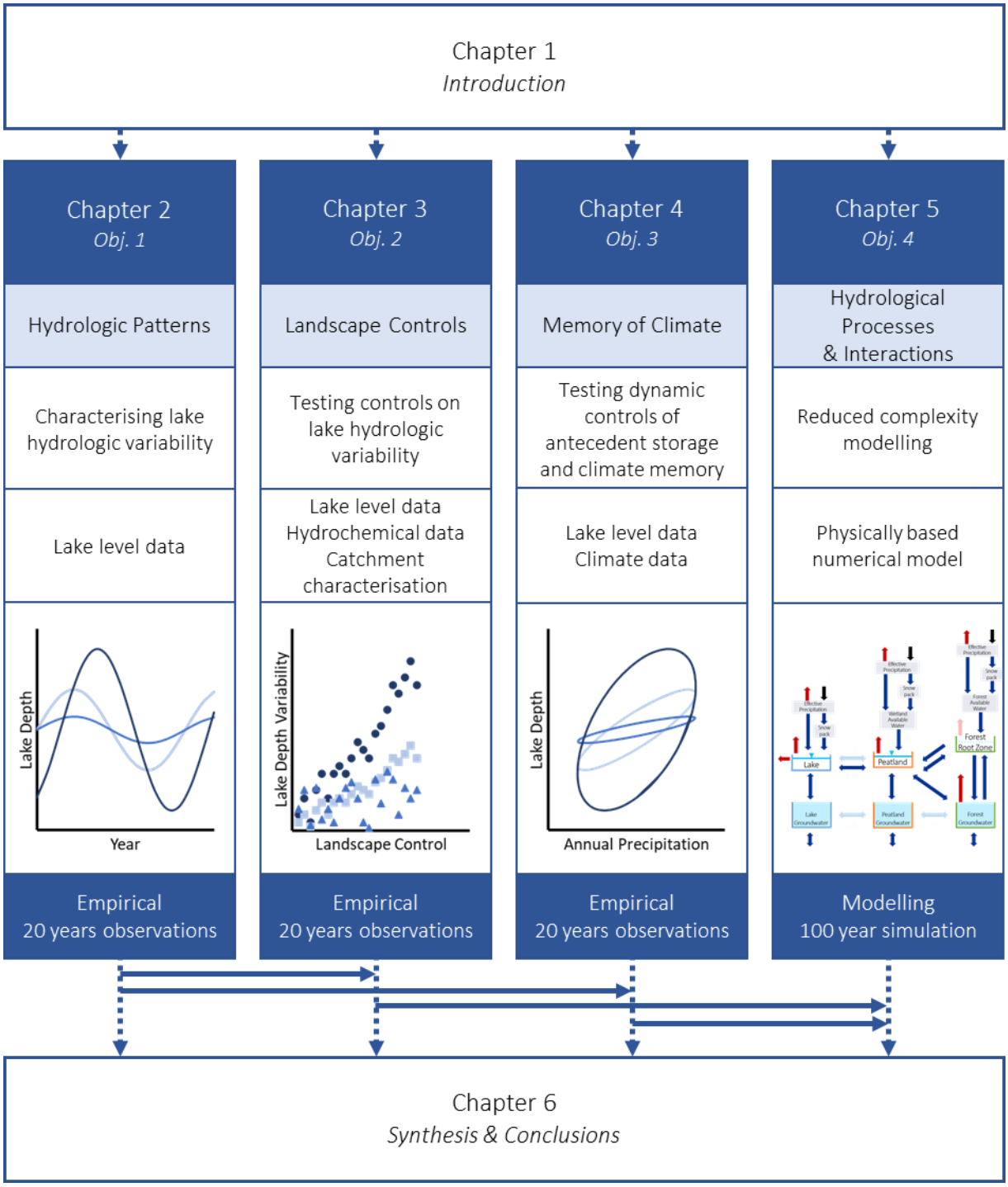


Figure 1-1. Thesis structure including relevant objective, key theme, data requirements and conceptual graphical visualisation of each research chapter, and how each chapter relates to others. Colours on conceptual diagrams represent different types of lakes which may have different dominant controls and lake water level patterns in response to the climate cycle.

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CHAPTER TWO: DIVERSE RESPONSE OF SHALLOW LAKE WATER LEVELS TO DECADAL WEATHER PATTERNS; RECOGNISING THE IMPORTANCE OF HYDROGEOLOGIC SETTING

Abstract

Shallow lakes represent the most sensitive lake ecosystems to water level fluctuations. Spatial differences in lake geometry, water balance components and lake-landscape connectivity will produce spatially variable responses to decadal weather cycles, climate change and disturbance. However, the patterns of shallow lake water levels are poorly defined in relation to their hydrogeoclimatic setting and long-term weather cycles. Here we utilise hydrologic datasets that transcend short-term to intra-annual, inter-annual and interdecadal scales and encompass diverse hydrogeologic settings to demonstrate low synchronicity in shallow lake water levels within the Boreal Plains (BP) of Canada. We found that surficial geology may incorporate important landscape parameters that control lake water level patterns. We also show that interannual variability is significantly greater than intra-annual variability. This demonstrates that long-term weather patterns are important in lake ecosystem functioning, and crucially illustrates that the magnitude of water level variability can be an artefact of study length and initiation, if studies are not sufficiently long to capture full response to decadal climate cycles across all lake types. This clearly highlights the need for studies with sufficient breadth in both temporal and spatial observations to capture climate, geology and topography so that a hydrological framework can be developed. Such a framework is crucial in regionalisation approaches to catchment management in order to mitigate ecological and societal impacts of hydrological changes under climate and anthropogenic impacts. We anticipate our findings and characterisation of lake hydrologic patterns to provide a vital foundation for testing the varying relative importance of topology, topography and typology controls across spatial and temporal scales.

1. Introduction

Glacial lakes comprise almost half of the surface area of Earth's lakes and represent extensive and diverse wetland and peatland environments where shallow lakes form behind glacial moraines, glacial outwash or in kettle holes and infill with terrigenous sediments (Cohen, 2003). They occur abundantly in mid-latitudinal belts as a product of climate and glaciation, also coinciding in distribution with regions experiencing greater non-stationary climatic forcing (Stralberg *et al.,* 2020) and changes to precipitation timing, phase, and magnitude (Tetzlaff *et al.,* 2013) and accelerating anthropogenic disturbance (Webster *et al.,* 2015). Consequently, lakes are projected to decrease in size (Bertassello *et al.,* 2019), and changes to hydrologic processes will have important societal impacts and ecological consequences (Ireson *et al.,* 2015).

Although historically underestimated (Meerhoff and Jeppesen, 2009), shallow lakes exhibit greater productivity, fish biomass and species richness per unit area than deeper lakes (Bayley and Prather, 2003) and provide the world's richest bird communities (Macdonald *et al.,* 2006; Morissette *et al.,* 2018). Conversely, shallow lakes are also the most sensitive lake ecosystems to water level fluctuations (WLF; Kolding and van Zwieten, 2012), and their resilience is strongly linked to stability of water levels, inferred from WLF amplitude (Bayley and Prather, 2003; Scheffer *et al.,* 2001; Cobbaert, Wong, and Bayley, 2014). Modifications to lake depth, area and hydroperiod are critical for migratory bird nesting and breeding grounds (Bellisario, Cerfolli, and Nascetti, 2014), macroinvertebrates (Gleason and Rooney, 2017), fish (Curry and Devito, 1996), and vegetation communities (Bayley and Prather, 2003) and water chemistry (Cobbaert, Wong, and Bayley, 2014). Subsequently, reduced hydrological connectivity and nutrient and mineral fluxes will have damaging consequences for downstream water supply, water quality and wildlife (Leibowitz, Mushet, and Newton, 2016; Webster *et al.,* 2015) and ecosystem state (Cobbaert, Wong, and Bayley, 2014; Scheffer *et al.,* 2001).

While the impact of lake hydrology on ecosystem health is well studied, the patterns of shallow lake water levels are poorly defined in relation to their hydrogeoclimatic setting and long-term weather cycles (Coops, Beklioglu, and Crisman, 2003). Large fluctuations in shallow lakes over decadal time scales is not a novel concept; several meters of change have been observed in homogenous closed prairie lakes (Van Der Kamp, Keir, and Evans, 2008), and shallow lakes have been observed to undergo temporary terrestrial wetland phases in response to decadal climate cycles (Devito, Mendoza, and Qualizza, 2012). However, spatial differences in basin storage, water budget components and lake-landscape connectivity will produce spatially variable responses to decadal weather cycles, climate change and disturbance. Studies incorporating heterogenous glacial deposit landforms have exposed spatially diverse hydrological patterns under the same climate signal (Prepas *et al.,* 2001; Cobbaert, Wong, and Bayley, 2015). Yet, it remains unclear how these signals will appear in lakes across decadal scales. The poor characterisation of lake water patterns hinders catchment comparisons and regionalisation approaches, which is necessary to transpose scientific knowledge to ungauged catchments and predict and mitigate ecological and societal impacts of hydrological changes. Thus, frameworks for examining lakes hydrology must capture climate, geology and topography (Devito *et al.,* 2005; Winter, 2001; Wolock, Winter, and Mcmahon, 2004) and address varying relative importance of topology, topography and typology controls across spatial and temporal scales (Buttle, 2006).

The successful utilisation of such hydrological frameworks in comparative hydrology demands hydrometric datasets that encompass natural WLF from short-term to intra-annual, inter-annual and interdecadal scales (Coops, Beklioglu, and Crisman, 2003), and span diverse hydrogeologic settings to identify spatio-temporal variability in shallow lake WLF patterns. Studies with this breadth in both temporal and spatial observations are generally lacking, instead exhibiting strength in either spatial or temporal insights. Long term studies have revealed the importance of both short and long-term climate cycles for shallow lake WLF (Van Der Kamp, Keir, and Evans, 2008), which cannot be determined in short term studies. Alternatively, high spatio-temporal resolution lake studies often

concentrate on water quality metrics (Gibson *et al.,* 2016), isotopes (Gibson, Yi, and Birks, 2019) or remote sensed data (Sass *et al.,* 2007), instead of hydrometric field observations. Contrastingly, long-term, regional and global studies include diverse hydrogeologic settings (Ali *et al.,* 2015; Carey *et al.,* 2010; Tetzlaff *et al.,* 2017) however, given the first-order importance of climate on hydrologic setting (Devito *et al.,* 2005), the challenge remains to unravel landscape controls from climatic controls.

The Boreal Plains of Canada (BP) provide the ideal study location to associate lake WLF to landform and weather patterns due to diverse lake types within the same climate signal. Deep, low-relief heterogeneous glacial deposits, combined with sub-humid climate, and poor drainage networks produce widespread shallow headwater lakes and lake-wetland complexes of varying characteristics across the BP. The BP encompasses lake types found across Boreal and low-relief dominant interior glaciated systems found globally, comprising of moraine, glaciofluvial outwash, and lacustrine-peat lakes and a combination of closed, flow-through and spilling lakes. Subsequently, BP lakes exhibit spatially diverse eco-hydrological functionalities (Ireson *et al.,* 2015), which engender the high productivity, biodiversity and richness and thus resilience of these ecosystems. However, the threat of climatic warming to ecosystem health (Laudon *et al.,* 2017; Price *et al.,* 2013) is of greater consequence in the BP, where industrial activities and resource extraction are heavily modifying the landscape at unprecedented scales (Webster *et al.,* 2015).

Due to a long-term water balance that oscillates around zero, WLFs represent a larger component of BP lake water balances relative to runoff or hydroperiod hydrometrics, therefore collapsing catchment heterogeneity and complex hydrological processes into simpler and more readily available storage observations (Carey, *et al.,* 2010; Mcnamara *et al.,* 2011; Shaw *et al.,* 2012), relative to deeper, runoff-dominated humid Boreal Shield lakes, or less permanent arid Prairie lakes. Subsequently, we hypothesise inconsistent WLF amplitude between years as a direct function of seasonal climate superimposed onto the long-term climate signal (Soulsby, Birkel, and Tetzlaff, 2016). Despite evidence

for spatially contrasting WLF amplitudes in short-term BP lake studies (Cobbaert, Wong, and Bayley, 2015; Plach *et al.,* 2016; Ferone and Devito, 2004; Smerdon, Devito, and Mendoza, 2005), we question whether the domination of precipitation and evapotranspiration will observe synchronous long-term WLF amplitudes across BP lakes if the study period is sufficiently long to collapse spatial heterogeneities in memory and lag responses to climate. Alternatively, while closed glacial lakes mimic the balance of precipitation and evaporation (Van Der Kamp, Keir, and Evans, 2008), storage-driven thresholds and non-linear processes that govern lake-catchment connectivity (Figure 2-1; Spence, 2010) could produce more complex and diverse non-linear hydrological relationships between heterogenous BP lake types and contrasting climate-lake hydrological relationships (hockey stick, step or sigmoid functions; Ali *et al.,* 2013; Figure 2-1b) based upon up-scaling of process knowledge from fine-scale empirical studies. Previous research indicates the importance of surficial geology (Plach *et al.,* 2016; Ferone and Devito, 2004; Smerdon, Devito, and Mendoza, 2005), and has given rise to the concept of Hydrologic Response Areas (HRAs), which classify surficial geology into landscape units of homogenous hydraulic properties, which could be applied to BP lakes.

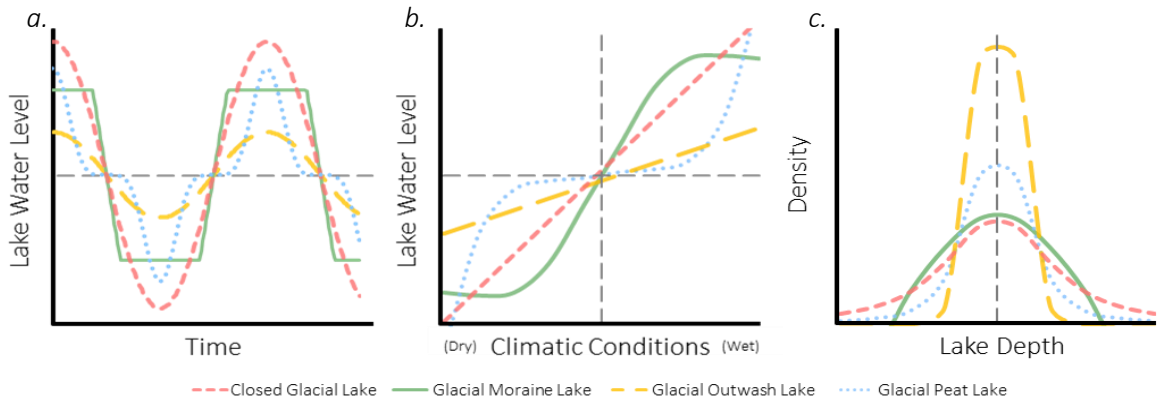


Figure 2-1. Hypotheses of (a) time series of lake water levels across decadal scale weather pattern (b) lake water level response to long-term climatic gradient and (c) subsequent distribution of lake water levels (assuming normally distributed climate) for different glacial lake types.

The role of groundwater supply in glacial outwash has been extensively studied (Winter, 2001; Winter, Rosenberry, and Labaugh 2003; Winter, 1986; Labaugh *et al.,* 1997; Labaugh, 1986; Jaquet, 1976) and shown to potentially moderate climate in outwash systems (Smerdon, Devito, and Mendoza, 2005), and may act to reduce the long-term amplitude of WLF in outwash lakes. Internal negative feedback mechanisms within peatlands (Waddington *et al.,* 2015) have also been shown to moderate climate within peat systems (Dixon *et al.,* 2017; Kettridge *et al.,* 2016) and even maintain saturated conditions during extreme drought (Kettridge and Waddington, 2014), however, the low storage in peat-dominated systems could magnify WLF responses across decadal scales when including extreme wetting and prolonged drought conditions. Moraine lakes also differ in hydrologic function; fine hummocky moraine lakes have been shown to respond strongly to climate (Thompson *et al.,* 2015), however, the occurrence of both lake desiccation and outflow generation in shallow spilling moraine lakes is expected to constrain the amplitude of WLF relative to other BP lakes across long-term observations, but could be critically omitted if study periods do not span full climate cycles.

The objective of this study was to characterise lake water level fluctuations of the sub-humid boreal plain. We also use this information to compare lake water level patterns in the BP with other Boreal and glaciated regions, and to discuss the impact of long-term lake water level variations on hydrological thresholds, stability and resilience. Given the high diversity in lakes and their catchments, and the presence of seasonal and long-term climate cycles, we use 20 years of water levels of lakes from various geology and landscape position to address questions: (1) the synchronicity of lake water levels across the BP; (2) the importance of intra-annual versus interannual variation and how these results are affected by study period timing and length; (3) the impact of long term water level fluctuations on hydrological thresholds that characterise habitat characteristics; (4) whether surficial geology incorporates parameters that distinguish lakes topologically and hydrologically, if these patterns differ from other Boreal and glaciated regions and therefore whether surficial geology can be utilised for regionalisation of lakes. This characterisation of lake WLF may provide critical insights required for the

classification of lakes, and to study the impacts of climate and anthropogenic disturbance on lakes and their ecosystem health and ecosystem services.

2. Study Area

This study was conducted in the Utikuma Region Study Area (URSA; 56°N, 115°W), within the Mid-Boreal Uplands Ecoregion of the Boreal Plain ecozone (Ecological Stratification Working Group, 1996), situated 150km south of the discontinuous permafrost zone (Devito *et al.,* 2012) and 370 km north of Edmonton, Alberta, Canada (Figure 2-2). The climate of the study region is sub humid, characterised by cold winters and warm summers (January and July temperatures of -14.5°C and 15.6°C respectively; Thompson *et al.,* 2015) with long term annual average PET (513 mm) exceeding annual average precipitation (483 mm; Environment Canada, 2003) as peak vegetation growth and ET demand coincides with 50–60% of annual precipitation between June and August (Brown *et al.,* 2010). Superimposed on this seasonal climate cycle is a long-term strong climate cycle, with wet years (PET:P ratio < 0.9) occurring infrequently on a 10–15 year cycle (Ferone and Devito, 2004). This decadal climate cycle dominates the regional hydrology, resulting in orders of magnitude variability in annual runoff over decadal time scales (<5 to 300 mm; Redding and Devito, 2010).

The landscape is characterised by a mosaic of forestlands and peatlands that vary in their dominance and size through the landscape. Within the URSA region 46.3% of landcover is classified as wetland or peatland and a further 12.6% is open water (see appendices). Peatlands are characterised by the accumulation of organic soils, and a forest canopy of black spruce and tamarack. Forestlands are characterised by low productivity pine, predominantly situated upon well-draining sandy regions, and a combination of aspen and spruce forests (white and black spruce), present within finer grained mineral soils. URSA overlies deep (45-240 m), low relief (<80 m) Quaternary glacial deposits (Hokanson, Mendoza, and Devito, 2019), and the Smoky Group shale unit of the Upper Cretaceous period (Vogwill, 1978; Ceroici, 1979; Plach *et al.,* 2016). The surficial geology varies spatially and can be simply

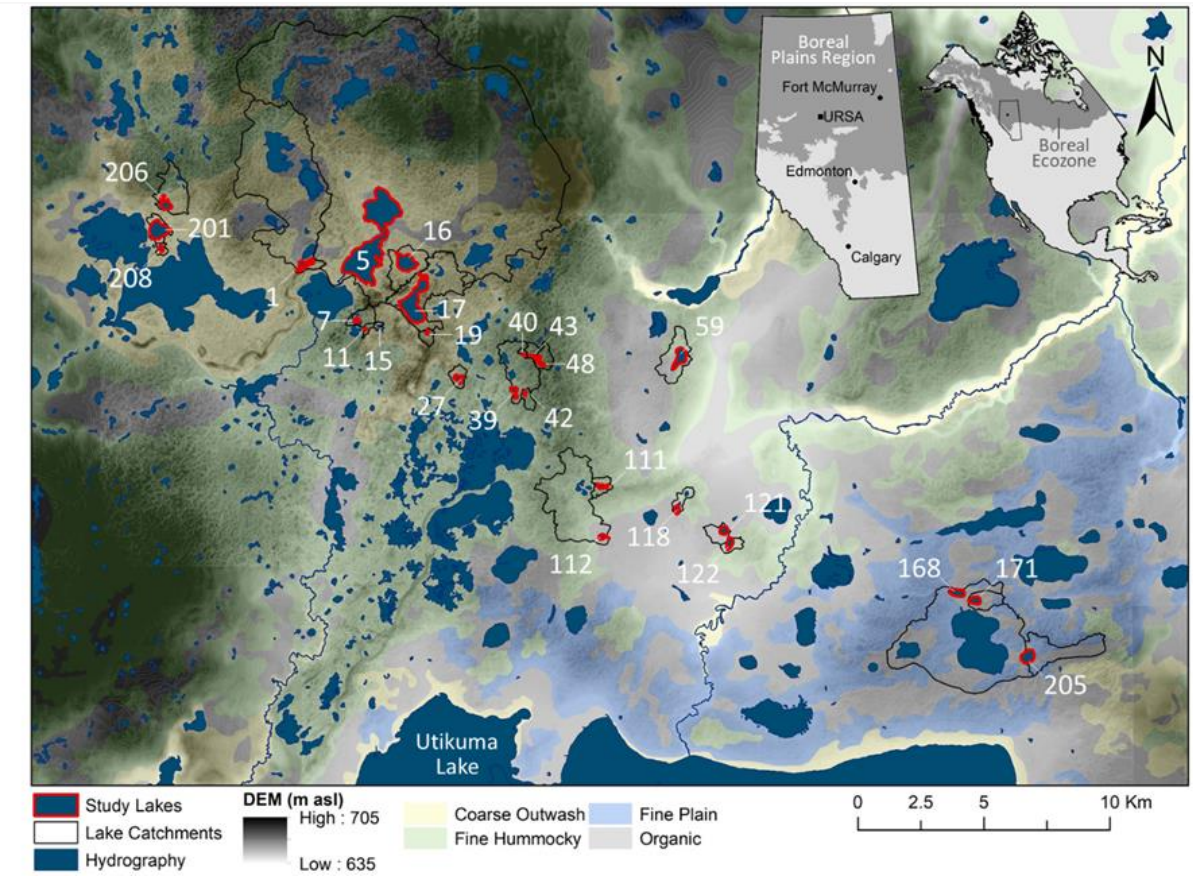


Figure 2-2. Site map of HRAs derived from surficial geology; hydrography, topography and surface catchment boundaries of 26 study lake catchments in the URSA (Utikuma Region Study Area), within the Boreal ecozone of Canada, and plains region of Alberta (Alberta Sustainable Resource Development, Alberta Environment, Alberta Community Development and Agriculture and Agri-Food Canada, June 2005).

categorised into three primary classes (Devito *et al.,* 2017): (1) coarse-textured glacial–fluvial outwash (CO); (2) fine-textured clay rich glacio-lacustrine plains (FP); (3) fine-textured hummocky disintegration moraine deposits (FH) and two transitional classes: (4) the transition from hummocky disintegration moraine to glacial fluvial outwash (CO/FH); (5) the transition from disintegration moraine to glacio-lacustrine plains (FP/FH).

The URSA region has been exposed to a range of natural and anthropogenic disturbance. Approximately 5.3% of the landscape within the URSA region has been harvested since 1950 (ABMI, 2019; see appendices) and oil and gas roads and seismic lines are also prominent features within the landscape (Strack *et al.,* 2019). Wildfire also occurs episodically, with a notably large (~90,000 ha)

Utikuma Complex wildfire within spring, 2011 (Kettridge *et al.,* 2019). The Utikuma Complex wildfire burned a large proportion of URSA region; notably both the fine textured hummocky disintegration moraine and fine textured glacio-lacustrine plains regions to the south of the URSA region. Excluding the Utikuma Complex fire, with the most recent extensive wildfire to burn within the URSA regions was in 1985.

3. Methods

3.1. Sampling strategy

Within the URSA, lake hydrometrics were collected for 26 study lakes between 1999 and 2018. Sites were selected strategically to represent the diversity in surficial geology, land cover, topography, lake geometry and lake types observed within the Boreal Plains (Table 2-1). This sampling strategy was developed to allow for factorial analysis to test for fundamental differences in lake hydrologic behaviour, based on key hypothesised controls on catchment hydrological function within thick low-relief glacial deposits and sub-humid climates (Buttle, 2006; Devito *et al.,* 2005). While varying landscape positions, peatland extent, surface roughness and lake surface area are captured within each HRA, almost all lakes are shallow (approximately 2 metres deep or shallower) and subsequently polymictic, and lakes situated at a high landscape position are absent within coarse HRAs.

3.2. Hydroclimatological measurements

Climate of the URSA and wider study region has been recorded for 23 years (1995-2018). Precipitation and evaporation have been monitored within the URSA with tipping rain bucket gauges, throughfall gauges, snow buckets, hand gauges and evaporation pans. Tipping rain bucket gauges were adjusted for snowfall by using antifreeze and cross checking with snow buckets. This has been supplemented with local government weather stations at Peavine and Red Earth Creek prior to 1999

(https://climate.weather.gc.ca/historical_data/search_historic_ data_e.html), and utilised for missing data, to produce a daily precipitation values and mean, maximum and minimum air temperatures record for the 20-year period. The precipitation cumulative deviation from the mean (CDM) is calculated for a moving 365-day window from the long-term annual mean (1995-2018) to represent a drought index for the study period.

A staff gauge was installed and referenced for elevation above mean sea level within each of the 26 lake catchments in 1998. All replacements of staff gauges within the study period were related to this original staff gauge. Manual lake water level measurements were collected each summer relative to staff gauges across all sites and calculated for level above mean sea level. Some lake water levels were measured multiple times throughout the ice-free period, and also logged by pressure transducers within staff gauges, further to an assortment of piezometers in adjacent or ephemerally connected peatlands and forests to monitor groundwater levels and pressure heads (see Devito *et al.,* 2017) for further details of wider research activities).

Annual lake sampling surveys were conducted in the final week of June or first week of July to capture peak lake water levels. Lake water level amplitude metrics were insensitive to length of sampling window within summer months (see supplementary material), therefore mean summer water levels were calculated from June 1st to August 31st to maximise sample size (number of years per lake). Where lake staff gauge measurements were unavailable in a given year, water levels in proximal shallow (<1m depth) piezometers which exhibit similar dynamics to the lake, were utilised and referenced above mean sea level. Where no water level measurements were available, lake water levels and hydrometrics were omitted for the given year. While gaps occur within this long-term record in some sites, they occur within mesic years (2005-2010) and are expected to have minimal impact on measures of water level variability used within this study.

Four key study lakes were sampled year-round and on a higher frequency. Each lake represents a key surficial geology landform, which also corresponds with Hutchinson's glacial lake types of: (1) a fine textured glacial moraine spilling lake (FH - '*fine hummocky*', lake 43); (2) a fine textured and peat dominated glaciolacustrine clay plain spilling lake (FP – '*fine plain*', lake 171); (3) a coarse textured glaciofluvial outwash flow through lake (CO – '*coarse outwash*', lake 16); (4) a closed lake situated in the transition zone between glaciofluvial and glacial moraine deposits (CO/FH - '*coarse outwash/fine hummocky*', lake 19). High frequency records were used to assess temporal patterns of lake water levels and enable comparisons of interannual and intra-annual dynamics, further to validating assumptions used within spatial analysis of the 26 study lakes (i.e., use of mean summer water levels; addressing data gaps).

3.3. Lake and Catchment Characteristics

Physiographic characteristics representing important topology, typology and topographic variables for lake and catchment hydrologic function were derived. Some metrics are developed relative to the Hydrologic Response Area (HRA) the given lake is situated in or relative to their topographically delineated catchment boundary and are outlined below. Additional maps of each lake catchment showing surficial geology, land cover, topography and catchment delineations are given in the supplementary material.

Lake Types: HRA Delineation of Surficial Geology

Hydrologic Response Areas (HRAs) are landscape classifications of surficial geology with similar hydraulic properties (Devito *et al.,* 2012) determined from Alberta Geological Survey surficial geology mapping (Fenton *et al.,* 2013). We define five HRA classifications: fine textured clay plains (FP); fine textured hummocky moraines (FH); coarse textured outwash (CO) and transitional classes between coarse outwash and fine hummocky moraine (CO/FH) and between fine hummocky moraine and fine

clay plains (FP/FH). The primary texture classification was used to define the texture of surficial geology units (coarse = gravel, sand, sandy-gravel, silty-sand, silty-gravel; fine = silt, clay, sand-silt-clay, sandy-silt, sandy-clay, silty-clay, clayey-silt), or determined from geomorphic class and landform genesis if no primary texture was provided (CO = fluvial, glacial-fluvial, aeolian, and littoral glacio-lacustrine landforms; FH = non-sandy stagnant ice moraines and moraines with hummocky geomorphic class; FP = lacustrine, distal glacio-lacustrine, and thrust moraine landforms) (Fenton *et al.,* 2013; Atkinson, Utting, and Pawley 2014). These HRAs represent significant differences in storage and transmission properties, where hydraulic conductivity averages 4.6×10^{-5} ms^{-1} in coarse and 4.2×10^{-7} ms^{-1} in textured surficial geology (Hokanson, Mendoza, and Devito, 2019). The HRA classification of each lake is given for both the topographically delineated catchment and for the lake itself. Lake HRA classes largely correspond with Hutchinson's lake types (Table 2-1), where all FP/FH and FP lake catchments are 'peat lakes'; FH lake catchments are moraine lakes and CO lake catchments are outwash lakes. CO/FH lake catchments are closed lake systems underlain by fine substrates to enable surface waters to persist at high landscape positions within the CO/FH transition zone; no spilling lakes were recorded within this class during the study period.

Lake and Catchment Characteristics

Catchment boundaries are poorly delineated by traditional flow accumulation tools associated with topography due to low relief and deep substrates within the BP. Therefore, manual delineation of catchment boundaries was undertaken by utilising a high resolution 2 x 2 meter DEM combined with satellite imagery (ESRI, 2020), historical aerial photography, Boreal Plain Wetland Classification (Ducks Unlimited Canada; Smith *et al.,* 2007), and observations or anecdotal evidence of culvert placement and flow direction.

HRA and lake catchment delineations were used to calculate topology, typology and topographic variables. Topographic variables of relief, surface roughness and slope were calculated from the DEM of the given catchment. Due to the greater relative importance of smaller scale relief over regional landscape position within the low-relief BP (Hokanson, Mendoza, and Devito, 2019), HRA subunits were defined to indicate the relative elevation of a given lake within its flow system, interpreted from land surface satellite imagery, field observations of flow and analysis of digital elevation models with GIS tools. Relative elevation is then given as a percentile elevation of the lake water level in 2012 within the HRA sub-unit. Land cover classifications (Brisco *et al.,* 2011; see Devito *et al.,* 2017 for details) were utilised to determine the presence of a 50m or 100m peatland buffer around the lake perimeter. Lake surface areas and perimeters were measured from 2013 land surface satellite imagery. Where high resolution satellite imagery was not available in 2013, 2015 imagery is used. Similar mesic climatic conditions and straight sided bathymetry is expected to have minimal impact on these metrics. Lake sill heights are defined as the elevation at which the lake spills relative the elevation of the base at the centre of the lake. In lakes where outflows are absent, the sill height is calculated from a DEM; in spilling lakes, the sill height is assumed to be the maximum observed lake depth, inferred from lake water level records and spot lake depth measurements in 2002 and 2012.

3.4. Data Analysis

Synchronicity

Mean summer lake water levels were calculated relative to 2012 water levels (WLD2012) for all 26 lakes. This standardised water level deviation metric enables visual assessment of synchronicity temporally (i.e., the timing of peaks and troughs), long term trends (drying or wetting) and the amplitude of long term interannual fluctuations. Water levels were standardised to 2012 due to observations of flow generation at landscape scales in this year (Devito *et al.,* 2017) further to the 1-

Table 2-1. Surficial geology, land cover, topography, lake geometry and lake types observed within the Boreal Plains.

Lake ID	Lake Type	Lake HRA Class	Catchment HRA Class	Lake Area (m^2)	Lake Sill Height (m)	Lake Perimeter (m)	Topographic Catchment Area (m^2)	Relative Elevation (%)	Slope (%)	Peat Lake	Coarse Lake	Outflow
208	Glacial Outwash	CO	CO	37142	1.5	726	178668	26.91	2.98	✗	✓	✓
1	Glacial Outwash	CO	CO	112090	2.5	3114	15912271	22.48	1.37	✗	✓	✓
206	Glacial Outwash	CO	CO	136775	3	2355	1834926	50.92	3.44	✗	✓	✓
201	Glacial Outwash	CO	CO	382239	2.8	2562	720999	26.91	1.11	✗	✓	✓
16	Glacial Outwash	CO	CO	465820	2	3069	1852940	8.41	2.09	✗	✓	✓
17	Glacial Outwash	CO	CO	1059549	5.5	6984	4718789	25.12	2.62	✗	✓	✗
5	Glacial Outwash	CO	CO	3244530	3.3	14158	90559288	4.97	2.08	✗	✓	✓
15	Isolated Glacial Outwash/ Moraine	CO/FH	CO/FH	485	4.5	93	77928	77.21	2.45	✗	✗	✗
11	Isolated Glacial Outwash/ Moraine	CO/FH	CO/FH	10680	2.5	374	129472	73.34	1.01	✗	✗	✗
7	Isolated Glacial Outwash/ Moraine	CO/FH	CO/FH	28671	3.35	762	632350	69.16	4.08	✗	✗	✗
19	Isolated Glacial Outwash/ Moraine	CO/FH	CO/FH	37390	4.65	820	295189	42.74	3.86	✗	✗	✗
40	Glacial Moraine	FH	FH	7725	1	341	82182	94.81	3.42	✗	✗	✓
43	Glacial Moraine	FH	FH	12773	0.8	623	274136	96.42	2.64	✗	✗	✓
42	Glacial Moraine	FH	FH	14037	2.3	485	303315	89.22	2.9	✗	✗	✓
48	Glacial Moraine	FH	FH	49779	1.2	1356	798440	97	3	✗	✗	✗
111	Glacial Moraine	FH	FH	58463	2	1262	463452	95.56	1.41	✗	✗	✓
39	Glacial Moraine	FH	FH	63994	2.5	880	2760538	82.19	2.23	✗	✗	✓
27	Glacial Moraine	FH	FH	67743	1.4	1226	436515	84.98	1.9	✗	✗	✓
112	Peat Lake	FP	FH	48472	1.6	1125	6052994	61.92	1.7	✓	✗	✓
118	Peat Lake	FP	FH	91424	1.6	1271	444833	55.63	1.22	✓	✗	✓
122	Peat Lake	FP	FH	91998	1.6	1232	373987	43.7	1.08	✓	✗	✓
59	Peat Lake	FP	FH	240951	2.2	2273	1674675	72.25	1.17	✓	✗	✓
121	Peat Lake	FP	FP	86767	1.5	1064	563027	38.79	0.88	✓	✗	✓
171	Peat Lake	FP	FP	107893	1	1246	823504	62.33	0.71	✓	✗	✓
168	Peat Lake	FP	FP	135410	1.6	1512	18486283	62.33	0.62	✓	✗	✓
205	Peat Lake	FP	FP	204979	1.3	1668	2672777	81.78	0.83	✓	✗	✓

year precipitation CDM aligning with multi-year precipitation CDM values, potentially accounting for spatially differences in memory.

Classification of Lake Characteristics

Physiographic variables included within the PCA reflect different topological, topographical, and typological parameters hypothesised to be important for lake hydrodynamics. Measures of landscape characteristics were combined with binary lake-type parameters “coarse lake” and “lacustrine” to assess how surficial geology classifications correspond with physiography. Prior to analysis, all

variables were tested for collinearity, scaled, and non-normally distributed variables were identified with Shapiro-Wilk normality tests and log transformed accordingly. Agglomerative and divisive hierarchical clustering methods were utilised to assess ability of HRAs to incorporate physiographic variables and further characterise these landforms.

Patterns and Behaviour in Water Level Fluctuations

The importance of interannual versus intra-annual fluctuations and the length and initiation of observations were determined by calculating the water level (WL) range (maximum WL minus minimum WL) relative to each metric for the four high frequency study lakes. For intra-annual range, the maximum minus minimum WL is calculated for each year for each of the main study lakes. Within each lake, only years with 6 months or more data are retained to prevent low range caused by low sample size. For interannual range, the maximum minus minimum WL from 1999 to 2018 is calculated for each month. Only months with 5 or more years data are retained to prevent low range caused by low sample size. To assess the effect of sampling window length and initiation, the interannual approach is applied using all water level records. The 20-year range is a single value for the study period 1999-2018. The 5 and 10-year ranges are timeseries of 5 and 10-year windows commencing 1999-2014 and 1999-2009 respectively. The range value of a given window is represented on the initial year of that window (e.g., 1999-2003 window plotted at 1999).

Lake water level probability distribution functions (PDF) are calculated relative to different habitat thresholds to provide the following WLF metrics: fluctuations relative to the height at which the lake spills (% sill height); fluctuations relative to the maximum depth of water observed (% maximum depth); fluctuations in lake depth (m); and fluctuations standardised to mean summer 2012 water levels (WLD 2012 m). These are presented as a PDF plot for all recorded water levels for the study period to assess any threshold behaviour and the shape of the distribution of responses. This is then

repeated for the 26 study lakes using summer means for 2000-2018 and presented as boxplots for individual lakes and as a PDF plot for grouped lakes within HRA classes.

4. Results

4.1. Climate of the Boreal Plains

Long term climate records the URSA and wider Boreal Plains (BP) study area reveal cyclical patterns in precipitation that can fluctuate +/- 200 mm around the long-term mean (+444 mm) (Figure 2-3). Snowfall and rainfall totals and timing also indicate strong seasonality of climate within the study area. Wet years (PET:P ratio < 0.9) occur infrequently, and 1-year CDM of precipitation is only observed to

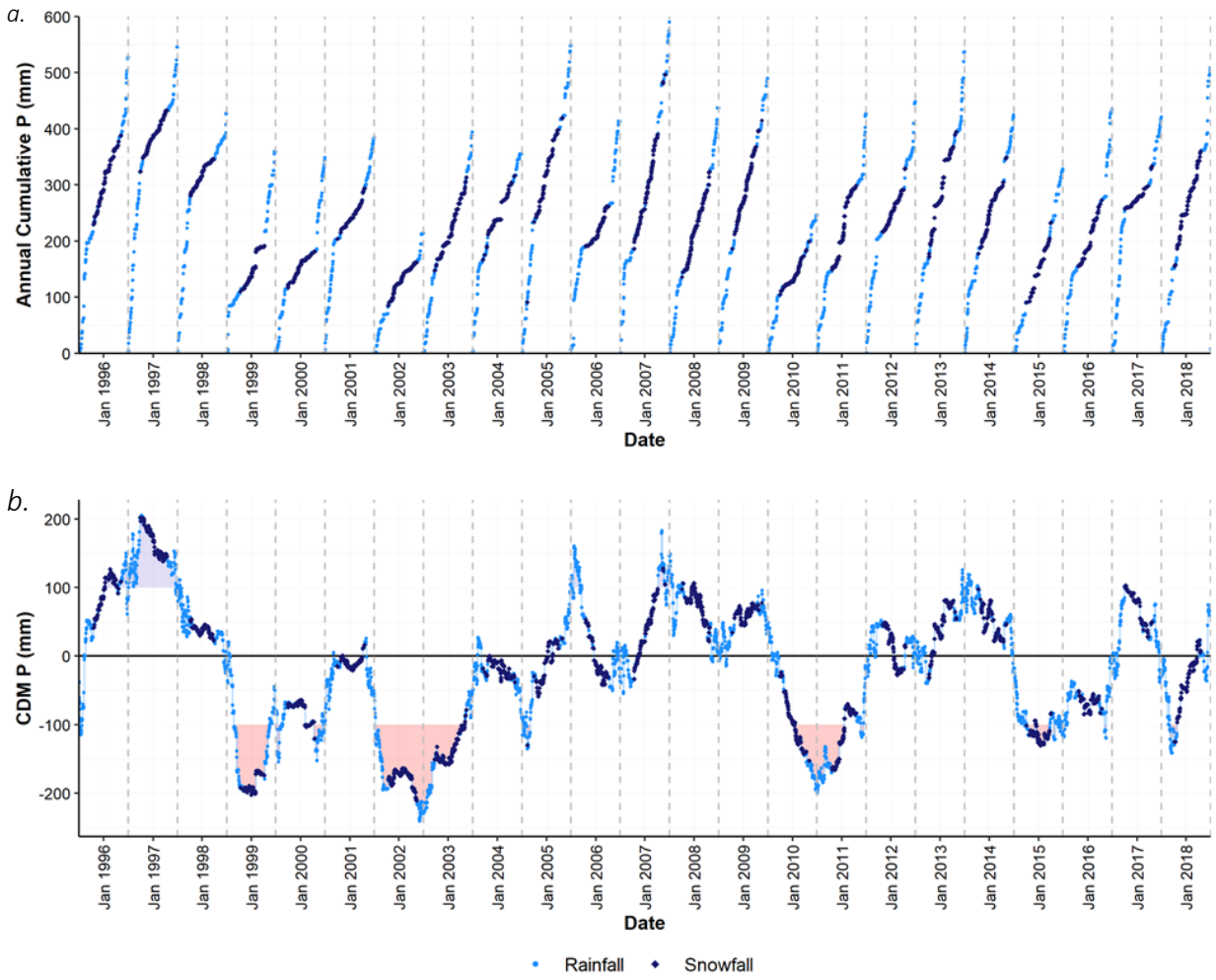


Figure 2-3. Climate of the Boreal Plains region of Alberta, Canada: a) annual cumulative precipitation 1996-2019 (hydrologic year commencing July 1st); b) precipitation 1-year cumulative difference from the long term mean. Red area = CDM P < -100 mm; blue area = CDM P > 100 mm.

exceed 100mm in 1996-97 within the 23-year record. Dry years (PET:P ratio >1.2) and mesic years (PET:P ratio of 0.9-1.2) and cycle every ~4 years, producing four dry periods of 1-year CDM below -100 mm within the 23-year record. The greatest amplitude in climate occurs early within the study period, where multiple dry years 1999-2003 succeed the 1996-97 wet period. Brief dry periods in 2010 and 2015 follow multiple mesic years 2004-2009 and 2012-2014. The 20-year lake water level observations (1999-2018) therefore capture a full decadal weather cycle and multiple shorter interannual water cycles that are commonly observed within the BP.

4.2. Lake WL Synchronicity

Lake water level deviation from 2012 levels (WLD2012) across the 26 study lakes reveal low synchronicity in lake fluctuations spatially and temporally (Figure 2-4). The amplitude and timing of peaks and troughs in WLD2012, and overall trend is inconsistent between individual lakes. However high similarity occurs within outwash lakes throughout the study period, with driest water levels occurring in 2003 or 2004, whereas the driest conditions occur in most peat and moraine lakes in 2002. Differing long-term trends are also observed between groups of lakes: closed lakes are shown to be drying throughout the study period, indicating higher memory or lagged response to the 1996-97 wet period, while peat and moraine lakes show trends of wetting throughout the study period following lows in the 2002-03 drought period. There is no distinct long-term upward or downward trend within the CO lakes and short-term cycles also occur as low magnitude interannual fluctuations. Therefore, the URSA exhibits diverse patterns in lake water levels despite consistent climatic conditions across the study region.

4.3. Characterisation of HRAs and Lake Classification

PCA analysis and hierarchical classification indicate that surficial geology strongly incorporates physiographic parameters defined within this study for determining spatial differences in hydrologic

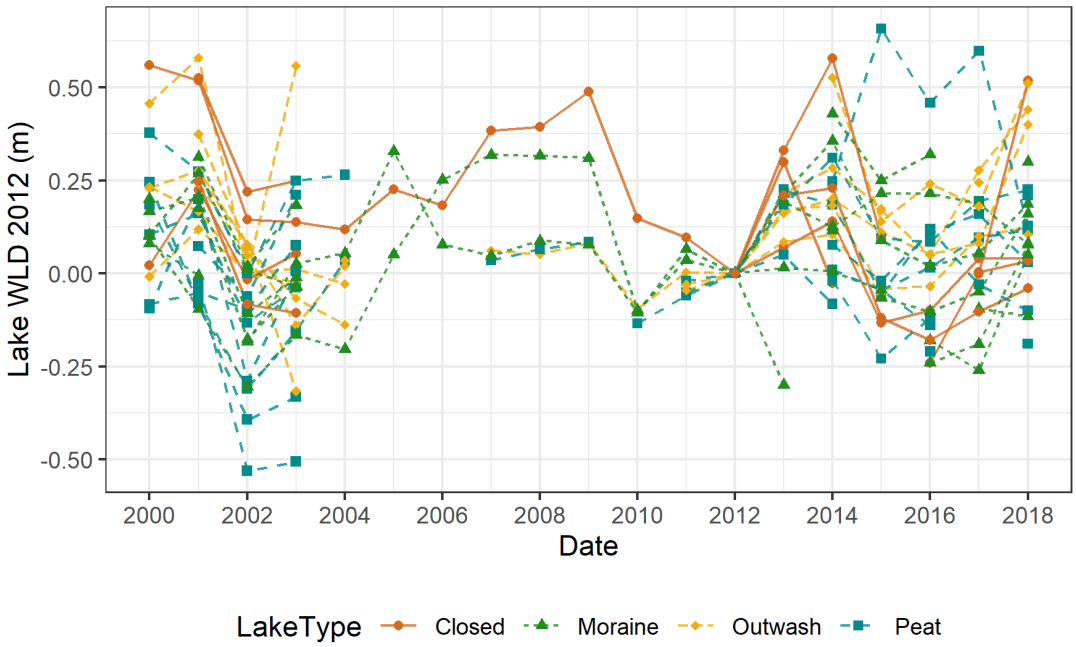


Figure 2-4. Mean summer (1st June – 31st August) lake water level devitaion 1999-2018 from mean 2012 lake water level of 26 study lakes in the URSA. Colour indicates Hutchinson's lake type (Table 2-1). See supplementary material for lake water level relative to sea level.

behaviour (Figure 2-5; see hierarchical classification in supplementary material). Peat lakes are concentrated on the lacustrine plains (FP) and are characterised by flat catchments with low surface roughness, low relief and more than 100 m width of peat around the perimeter of the lake. FP/FH lakes are also peat lakes, but with moraine catchments, yet exhibit greater association with FP than FH due to the presence of peatland buffers around lakes and gentler topography. Moraine lakes (FH) are associated with higher landscape positions, smaller lake areas and lake catchments, low sill heights, and high relief, roughness, and slope. FH lakes are also surrounded by narrower peatland buffers and exhibit a greater proportion of forested upland. CO/FH lakes are closely associated with the characteristics of FH, however greater topographic roughness, slope and sill heights produce small, closed basins. CO lakes are situated at low landscape positions and are characterised by large lake and catchment areas, with higher sill heights and deeper lakes. Lake 208 has an exceptionally small catchment as a low-lying CO lake, where a local hillslope may be dominating over larger scale characteristics in this catchment, causing greater association with FH. Overall, there is an evident

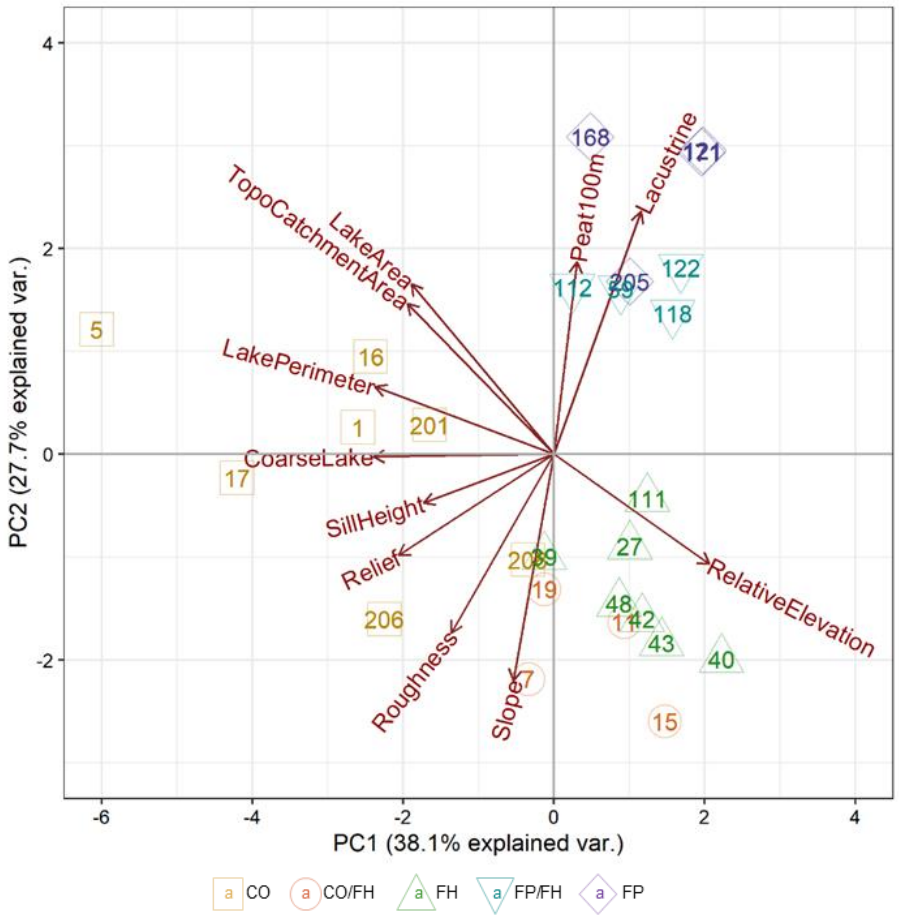


Figure 2-5. PCA analysis of lake and catchment topology (data given in Table 2-1.) variables of 26 study lakes in the URSA. Lake label colours indicate the HRA class of lakes based on surficial geology, which closely correspond with types of lake formation.

distinction in physiographic characteristics between lake types situated on different HRAs, which may be important for spatial patterns in hydrologic behaviour.

4.4. Temporal Water Level Fluctuation Patterns - Key Study Lakes

Interannual vs Intra-annual Variability

High frequency observations of four key study lakes of different lake types/ HRA classes indicate poor synchronicity in timing, magnitude and overall trend of lake water level patterns at seasonal, interannual and decadal time scales (Figure 2-6 and 2-7), in line with lower resolution observations across 26 study lakes (Figure 2-4). Based on the full set of observations for each key study lake, we

show that interannual variability is greater than intra-annual variability in all lakes (Figure 2-7a versus Figure 2-7b), however, this is less significant in CO (lake 16). A comparison of the year with most intra-annual variability with the 20-year interannual range indicates that within year variability is 12% less than between year variability in CO, compared to 29%, 56% and 49% less intra-annual variability in CO/FH, FH and FP respectively.

Interannual variability assessed by month also demonstrates that differences in summer water levels (May to August) drive high ranges in interannual fluctuations, corresponding with spring snowmelt and peak evapotranspiration losses during growing season. This pattern is not evident for CO, although slightly raised variability in May could indicate influence of snowmelt magnitudes. This indicates that the long-term range is a function of seasonal variability superimposed onto the long-term climate.

By varying the length and starting point of the sampling window, we show that study length and initiation are important for true interpretation of magnitude of variability (Figure 2-7c). Our results show that 5-year and 10-year studies are substantially insufficient to capture the range in WLF observed across 19-20 years in all key study lakes and demonstrates how different initiation dates may result in different interpretations on lake stability. A 5-year study may underpredict variability by as much as 69% (0.7m) in FP (lake 171); 56% (0.6m) in CO/FH (lake 19); 54% (0.22m) in CO (lake 16) and 46% (0.43m) in FH (lake 43). On the other hand, short studies conducted following the wet 1996-97 wet period that include the 2002-03 drought could represent the full range of conditions with FP and FH lakes, however, comparisons drawn with CO and CO/FH lakes for the same period may be limited due to spatial differences in memory and lag responses, and greater interannual variability captured at the end of this study period (2014) for CO and CO/FH.

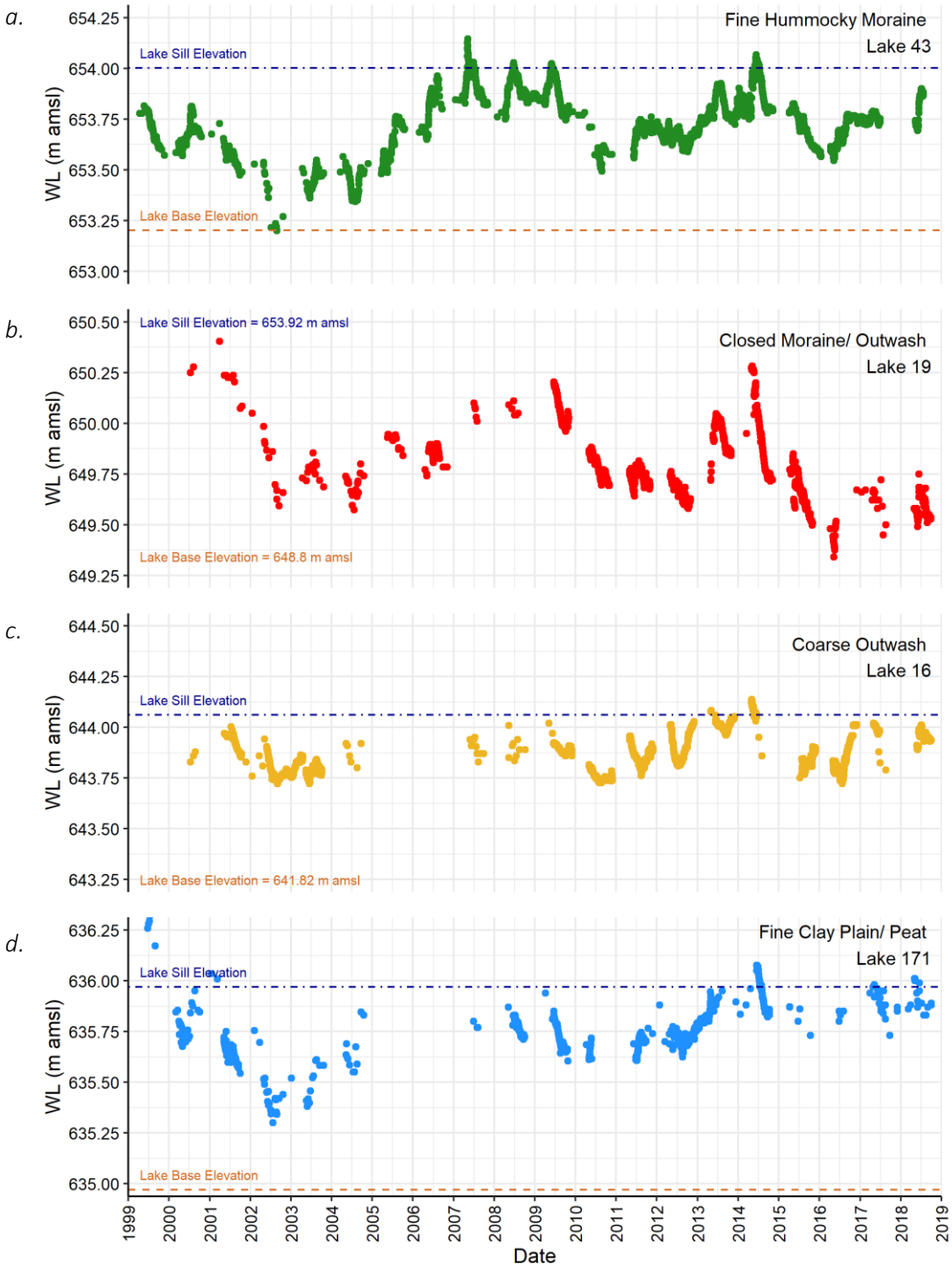


Figure 2-6. Lake water level observations 1999-2019 above mean sea level of 4 key study lakes in the URSA: a) Lake 43 – fine hummocky moraine lake; b) Lake 19 – moraine/outwash closed lake; c) Lake 16 – coarse outwash lake; d) Lake 171 – fine clay plain peat lake. Lake sill and lake bottom elevations are given for each lake.

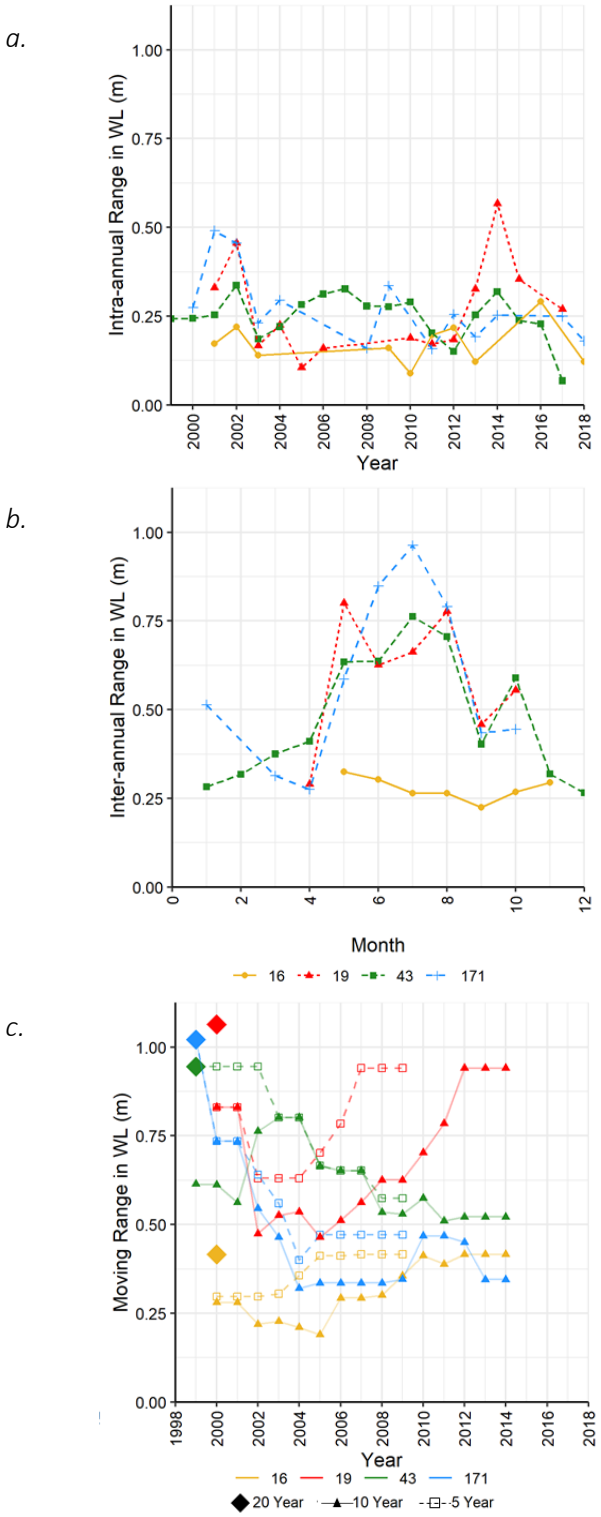


Figure 2-7. Metrics of lake WLF amplitude across different sampling periods: a) Intra-annual range; maximum minus minimum lake WL within each year 1999-2018. Within each lake, only years with 6 months or more data retained to prevent low range caused by low sample size. b) Inter-annual range: Maximum minus minimum lake WL within each month from 1999 to 2018. Within each lake, only months with 5 or more years data retained to prevent low range caused by low sample size. c) Inter-annual range across varied sampling windows: Maximum minus minimum lake WL within sampling windows. The 20 year range is a single value for the study period 1999-2018. The 5 and 10 year ranges are timeseries of 5 and 10 year windows commencing 1999-2014 and 1999-2009 respectively. The range value of a given window is represented on the initial year of that window (e.g. 1999-2003 window plotted at 1999).

Characteristics of Water Level Fluctuations

The shape of water level probability distribution functions (PDF) for key study lakes (Figure 2-8) strongly correspond with hypothesised distributions of lakes of different landforms (Figure 2-1). These distributions when plotted relative to key habitat thresholds or standardised to a single point in time provide insight into hydrological mechanisms controlling their patterns and indicate key spatial differences in lake stability, including: frequency and magnitude or distance of interaction with the sill (Figure 2-8a); stability inferred from percentage range in depth (Figure 2-8b); proximity or frequency of desiccation (Figure 2-8c); and differences in memory or lag responses (Figure 2-8d).

CO water levels exhibit the narrowest range and distribution indicating low variability in fluctuations; water levels are maintained around the sill elevation and consistently produce outflow annually. As the deepest key study lake with the lowest range in depth (0.42 m), CO fluctuates through just 18% of its maximum observed storage.

FP lakes exhibit infrequent but potentially high magnitude exceedance of their sills, and similar drying behaviour of low frequency, high magnitude response, producing tapered tails within its water level distribution. While water level fluctuations are moderate for most climatic conditions, extreme fluctuations occur during drought conditions in 2003, and following a wet period 1996-97. This results in a high range in depth (1.02 m), representing fluctuation through 75% of the maximum observed storage. FH also infrequently exceeds the sill, however the PDF of FH is restricted by both outflow generation and desiccation due to shallower depths, thus fluctuating through 100% (0.94 m) of its maximum observed storage and a more truncated distribution relative to FP. CO/FH presents the widest distribution in depth as the only key study lake not producing outflow due to high sill height, but also greater distance from desiccation than FH due to deeper water depths. Subsequently, CO/FH fluctuates through 66% (1.06 m) of its maximum observed storage.

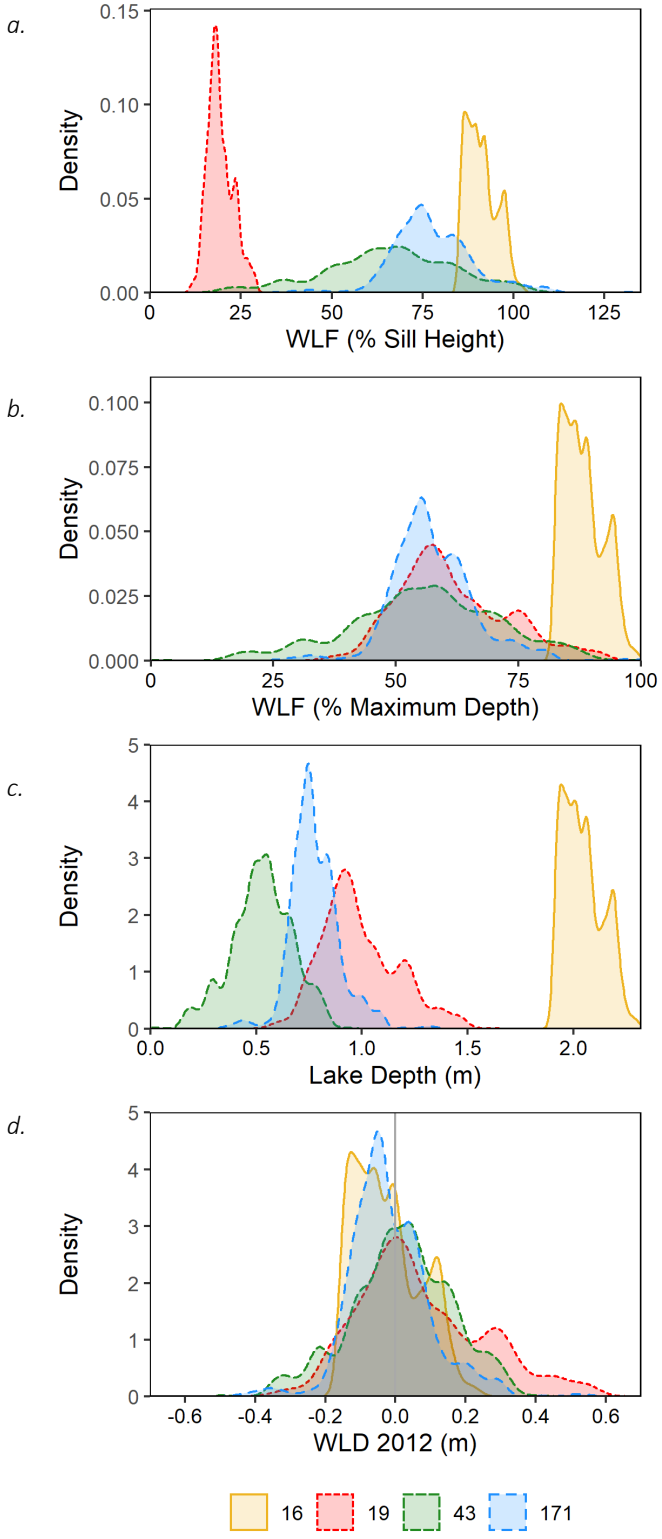


Figure 2-8. Probability distribution functions (PDF) of lake WL observations in the context of: a) water depth as a percentage of potential storage (sill height) to indicate frequency or distance to outflow; b) water depth as a percentage of maximum observed storage to indicate stability of lake depth; c) water depth (m) to indicate frequency or distance to desiccation; d) lake water level relative to summer mean levels in 2012 to indicate synchronicity of lakes (positive values indicate higher water levels relative to 2012).

Differences in WLD2012 (Figure 2-8d) between lakes reinforces findings of differences in memory or lag response to climate between lakes resulting in low synchrony in the timing of peaks and troughs. These echo findings of moving window ranges, whereby CO/FH WLD2012 is skewed towards positive values due to memory of the 1996-97 wet period, dryer conditions in 2012 and rapid wetting in 2014. In contrast, CO, FP and FH exhibit skews towards negative WLD2012 values of varying degrees and distribution shapes, indicating different timing and magnitude responses to recurring drought conditions.

4.5. Spatial Water Level Fluctuation Patterns - 26 Study Lakes

The distributions of 20-year mean summer lake water levels across the 26 study lakes relative to important habitat thresholds and standardised to 2012 mirror the findings of the main study lakes within their surficial geology classification (Figure 2-9). CO lakes display a narrow distribution and low range in fluctuations in both absolute depth and as a percentage of maximum storage. All CO lakes fluctuated around their sill elevation for the study duration (Figure 2-9a), indicating frequent runoff annually. FP, FP/FH and FH lakes infrequently exceed their sill while closed CO/FH lakes never exceed their sills within the study period. CO/FH and FH exhibit the greatest percentage fluctuations relative to their maximum depth (Figure 2-9b), however this appears to be a function of lake depth, where shallow lakes exhibit greater percentage fluctuations. In turn, these shallower lakes display the closest proximity to or propensity for desiccation (Figure 2-9c). While the characteristics of WLF relative to outflow and desiccation are similar amongst HRA classes, the absolute variability and WLD2012 is less distinct, indicating additional controls on magnitude responses and lag responses that vary within HRA class (Figure 2-9d).

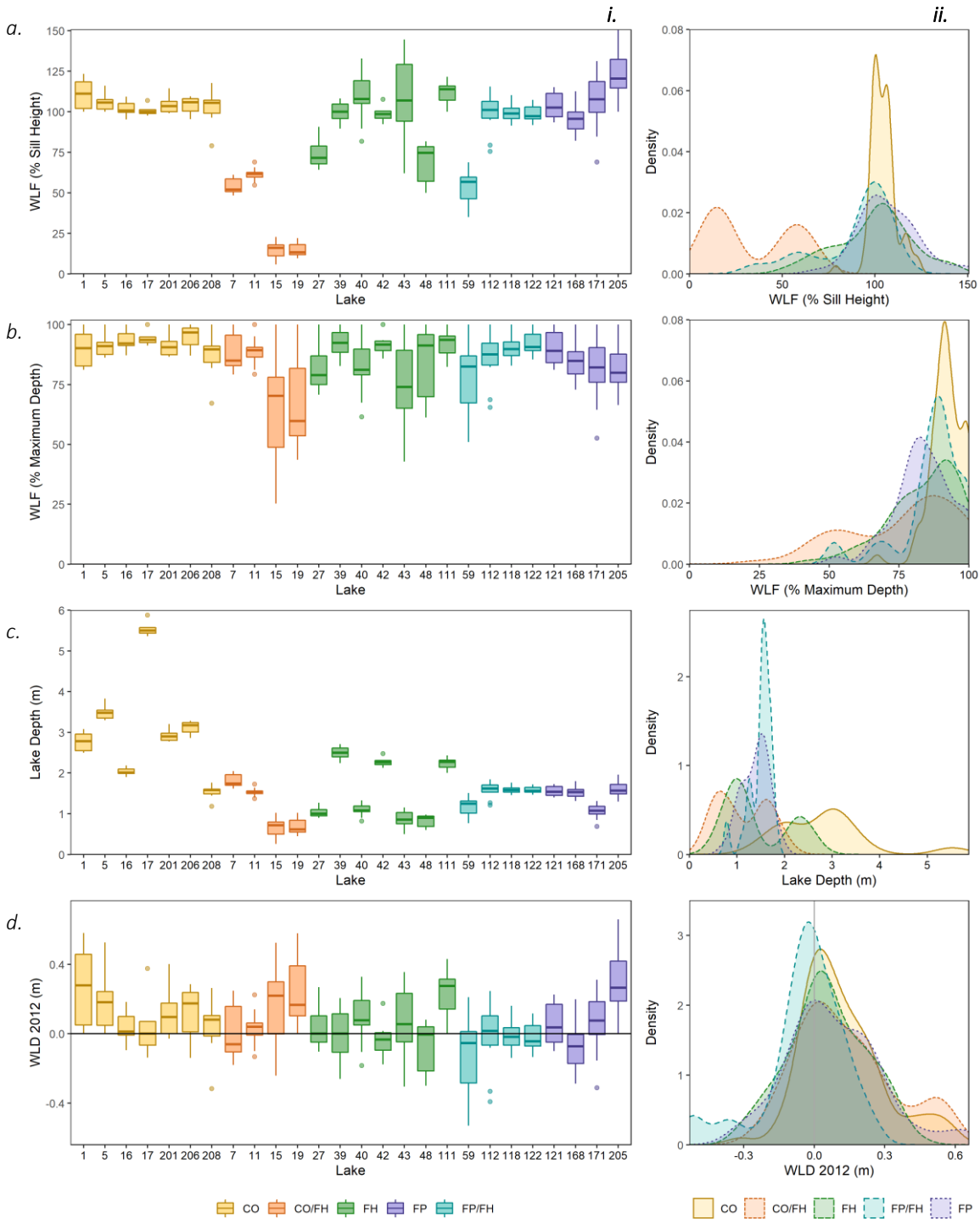


Figure 2-9. Maximum summer (1st June- 31st August) lake water level observations 1999-2018 above mean sea level of 26 study lakes in the URSA as: i) boxplots for individual lakes and ii) probability distribution function (PDF) plots for HRAs. Lake WL observations are given in the context of: a) water depth as a percentage of potential storage (sill height) to indicate frequency or distance to outflow; b) water depth as a percentage of maximum observed storage to indicate stability of lake depth; c) water depth (m) to indicate frequency or distance to desiccation; d) lake water level relative to levels in 2012 to indicate synchronicity of lakes responses to climate (positive values indicate higher water levels relative to 2012). Colour indicates surficial geology classification of lake (HRA).

5. Discussion

5.1. Synchronicity of BP lakes water levels across space and time

Interannual versus intra-annual lake water level patterns

This study clearly demonstrates greater interannual than intra-annual variability in Boreal Plain (BP) lake water levels across a range of lake types, signifying the importance of long-term weather patterns in lake ecosystem functioning. Lake water level fluctuations (WLF) are governed by seasonal climate patterns in regions with high relief and high precipitation, such as the humid Boreal Shield (Oswald, Richardson, and Branfireun, 2011); snow-dominated Montane (Lee *et al.,* 2015) and Alaskan (Arp *et al.,* 2012) regions and temperate European Boreal (Karlsen *et al.,* 2019). In these hydro-geoclimatic settings, inputs frequently exceed storage thresholds, producing runoff annually and maintaining stable interannual fluctuations. However, poor drainage and greater storage of water in the low relief, sub-humid BP causes memory of antecedent conditions within shallow glacial lakes. Therefore, peak water levels within a given year reflect the seasonal climate superimposed onto the long-term climate cycle, which fluctuates between positive and negative water balances on four-year cycles, with wet years returning on a 20-year cycle in the BP (Devito *et al.,* 2017). Moreover, we show that large differences in interannual variability are driven by summer water levels, signifying the importance of precipitation timing relative to evaporative demand within the growing season, in contrast to systems where spring snowmelt drives WLF.

Storage-driven memory is a widely understood phenomenon in arid closed Prairie depression basins (Van Der Kamp, Keir, and Evans, 2008), where waterbodies vary immensely from transient puddles to semi-permanent lakes (Shook and Pomeroy, 2011). In turn, storage concepts derived within the Prairies may lend to BP lakes better than complex non-linear threshold-mediated runoff dynamics developed in Boreal Shield studies (Oswald, Richardson, and Branfireun, 2011; Spence and Woo, 2006) and global northern catchment studies (Ali *et al.,* 2015) where intra-annual fluctuations dominate. In

closed Prairie basins, WLF reflect cumulative departures from the long-term average of precipitation (CDM), minus evaporation (Van Der Kamp, Keir, and Evans, 2008), similar to the three-year CDM signal observed within BP groundwater levels (Hokanson, Mendoza, and Devito, 2019).

While high storage gives rise to long term trends in hydrological responses across the whole region, BP lakes are substantially more complex as connectivity to wetlands is ephemeral across varying timescales and heterogeneous surficial geology results in varied scales and magnitudes of groundwater connectivity (Plach *et al.,* 2016). Our research demonstrates the need for hydro-geoclimatic frameworks that examine the interaction of climate with geology and storage to characterise interannual responses to long-term weather patterns and climate cycles in order to assess how lake water levels may differ between varying hydrogeologic settings under future climates.

Spatially diverse hydrological threshold behaviours

Empirical evidence within this study demonstrates low synchronicity in shallow lake water levels within the BP, with spatially diverse amplitudes of WLF and varied temporal responses to climate. Consistent threshold-related lake water level behaviour is observed within Hydrological Response Areas (HRAs) derived from surficial geology classifications. This indicates that surficial geology may incorporate important landscape parameters that control WLF patterns and characteristics, in line with individual field studies for runoff (Devito *et al.,* 2017); groundwater levels (Hokanson, Mendoza, and Devito, 2019); lake-groundwater dynamics (Plach *et al.,* 2016; Smerdon, Mendoza, and Devito, 2007) and modelling research (Carrera-Hernández, Smerdon, and Mendoza, 2012; Thompson, Mendoza, and Devito, 2017). Yet, this is the first study to examine hydrometrics of diverse lake and catchment types over a full long-term climate cycle and provide evidence for spatially diverse water level patterns under one climate signal.

In studying diverse lake types within one climatic region, the effects of climatic variability are removed, and instead hydrologic patterns associated with physiographic characteristics can be examined. Threshold metrics are increasingly recognised as useful tools for defining hydrological signatures (Spence, 2007) for catchment classification (Wagener *et al.,* 2007) and strengthening knowledge of relationships between climate and physiography (Ali *et al.,* 2013). Within this study, coarse textured glaciofluvial outwash lakes (CO) produced runoff annually and exhibited the narrowest range and distribution of water levels, corresponding with the behaviour of groundwater patterns in BP coarse outwash systems (Hokanson, Mendoza, and Devito, 2019) and larger scale groundwater flow in BP outwash modelling work within systems (Smerdon, Devito, and Mendoza, 2005). Fine textured glaciolacustrine clay plains lakes (FP) also exhibit moderated water levels for a wide range of climatic conditions, corresponding with observations of regulated hydrologic conditions in peat dominated systems (Kettridge and Waddington, 2014). However, FP lakes also displayed infrequent but high magnitude exceedance of their sills, and similar low frequency, high magnitude drying behaviour under extreme climatic conditions, producing large differences in behaviour during different phases within a climate cycle. Fine textured hummocky moraine (FH) systems are recognised for mimicking climate (Thompson *et al.,* 2015) due to lower peatland cover and receiving lower groundwater contributions cover than FP, and are regarded as the least stable systems under future climate projections (Thompson, Mendoza, and Devito, 2017). However, we show that the magnitude of WLF are constrained by desiccation and outflow; the resulting lower capacity to accumulate climate causes rapid recovery of lake water levels following drought upon rewetting. This is contrary to closed fine textured lakes which characterise CO/FH lakes, where high sill heights enable greater water level fluctuation and in turn reflect the cumulative climate. This demonstrates the importance of storage controls on connectivity and hydrological thresholds for driving patterns in lake water levels, similar to varying types of prairie lakes, where closed basins reflect the long-term climate; small hummocky plateaus frequently dry out; end receiving lakes spill infrequently upon increased connectivity across

the catchment, while large, deep lakes spill annually (Shaw *et al.,* 2012). These contrasting hydrological patterns between surficial geology textures and dominant land covers within the BP highlight the need for geology and landcover frameworks in order to extrapolate empirical findings and modelling research to ungauged catchments and to future climates in order to develop management tools in these landscapes.

5.2. Importance of long-term observations concurrent with different lake types to accurately assess stability

Measures of interannual variability across different timeframes clearly show that the magnitude of water level variability is an artefact of study length and initiation in BP lakes. Our results indicate that short-term studies (<10 years) are insufficient to capture the range in WLs observed across 20 years. Moreover, the non-linear relationship between climate and hydrometrics in some lake types means that shorter studies initiated on different dates of the same study site result in different interpretations of hydrologic behaviour as a product of their position within the long-term climate cycle. Further, spatial differences in lag responses to climate suggests that short term studies of diverse lake types will capture lakes in different phases and thus not necessarily responding to climatic conditions recorded within the study or the same scale of climate cycle (annual, multi-annual, decadal). On the other hand, our 20-year data set collapses spatial differences in long-term absolute variability; high spatial variability in lake water level patterns only become apparent when examining temporal dynamics, or the distributions of water levels and assessing them relative to important hydrological thresholds across long-term data sets. While averages or ranges of long-term data may expose spatial differences in landscape controls in systems dominated by intra-annual variability (e.g. the shield, Cobbaert *et al.,* 2014; Mwale *et al.,* 2009; Sass *et al.,* 2007), this cannot be assumed for systems exhibiting interannual variability. Instead, summary statistics may obscure contrasting hydrological

behaviour within large spatio-temporal studies and lead to incorrect conclusions of high synchronicity in diverse landscapes exposed to the same climate.

Artefacts of study timing and length have long been speculated as the cause for conflicting findings in lake stability. While there has long been a demand for more long-term hydrology observations to fully understand northern catchments (Tetzlaff *et al.,* 2017; Laudon *et al.,* 2017), this is the first study to examine hydrometrics of diverse lake and catchment types over a full long-term climate cycle and provide evidence for changes in magnitude response at different points within the climate cycle. Subsequently, we suggest that climate-landscape controls can only be explained upon approaching these observations with a priori knowledge of heterogeneities in hydrogeoclimatic setting in large spatial datasets (Hokanson, Mendoza, and Devito, 2019). In doing so, study lakes can be placed in the context of their phase response to the long-term climate based upon their surficial geology and dominant land cover. We exemplify this with BP case study examples, where inconsistent interpretations of lake water level patterns can be made as a result of differing study length and initiation.

We provided Figure 2-1 as a priori visualisation of hydrologic relationships, which highlights the importance of threshold behaviour that may not be captured under a study observing a narrower range of climatic conditions (Figure 2-1b). While spatial differences in storage-driven connectivity and thus memory will result in far more complex temporal patterns than posited (Figure 2-1a), our results support hypothesised climate-lake level hydrologic relationships across long-term climate cycles. Most strikingly, moving 5-year water level ranges erroneously indicate that FP is the least variable lake based on observations commencing in 2014, yet the most variable within 5-years commencing in 1999. The former conforms with common understanding of stability in peat systems; however, this study presents compelling evidence of highly variable FP systems depending on timing within study. Here, FP shows the greatest response following the wettest period in 1996/97 through to the driest period

2002/03. Conventionally, FH lakes are considered to respond greatest to climatic conditions, however, minimal memory of the 1996/97 wet period, likely due to outflow losses, and desiccation in 2002 results in lower variability in FH than FP. Similarly, the selected length of study results in conflicting conclusions of WLF magnitudes in the closed CO/FH lake basin due to high memory of the wet 1996/97 period, which declines over a very long period by evaporation (Van Der Kamp *et al.,* 2008), followed by rapid wetting in 2014. Subsequently, 5 and 10-year ranges show increasing interannual variability in CO/FH throughout the study period, whereas FH and FP decrease in variability throughout the study period. Timing and length of study is of less significance in CO, which corresponds with observations in low interannual variability. Lower 5 and 10-year ranges at the beginning of the study could signify diminishing buffering capacity of groundwater as an extended lag response to the 1996-97 wet period, which could be restored upon the next wet period. These findings clearly exemplify the need for long term studies in order to produce and apply frameworks.

5.3. Implications for determining stability and interpreting resilience

Defined as low deviation from its average state, stability in lake systems is commonly inferred from the magnitude of water level fluctuations (Cobbaert, Wong, and Bayley, 2015; Bayley and Prather, 2003; Cobbaert, Wong, and Bayley, 2014). We expose two problems with this definition for stability, the first being a methodological issue, where studies of insufficient length may produce values unrepresentative of true long-term variability as outlined above. The second issue lies with the logic of using variability to define stability, or by extension, to interpret resilience. Despite identifying diverse hydrologic behaviour between HRAs, absolute magnitude of variability is comparable across the study region, therefore the periodicity and timing of fluctuations may be of greater importance for inferring stability. Stability is often associated with hydrological thresholds which shape habitat characteristics such as desiccation or outflow generation (Sass *et al.,* 2007; Cobbaert, Wong, and Bayley, 2015). These represent important markers both ecologically and chemically, by governing

habitat availability, the movement and concentration of nutrients, and the downstream supply of water. As a result, it is a common misconception that if a lake dries up is hydrologically unstable or exhibits low resilience; however, drying out may be part of a lakes natural cycle, acting as a negative feedback mechanism to enable rapid recovery following droughts and thus shape the resilience of the given lake. Furthermore, the high community richness, diversity and productivity of BP ecosystems is a result of heterogeneity in lake habitats, which derive from spatially diverse lake responses to climate producing lakes of varying depths. Given the dominance of interannual WLF, and that WLF may be operating on different timescales due to spatial differences in storage-driven memory, the long-term interactions of hydrological thresholds and how they shape landscape resilience could be overlooked in short studies. If we can improve our understanding of hydrometrics, then findings from resilience and stability studies could be extrapolated to the full climate cycle. However, if behaviour in lake water levels is not truly represented due to artefacts of study period length and initiation, then lake water level patterns cannot be fully understood, or predicted under future climates.

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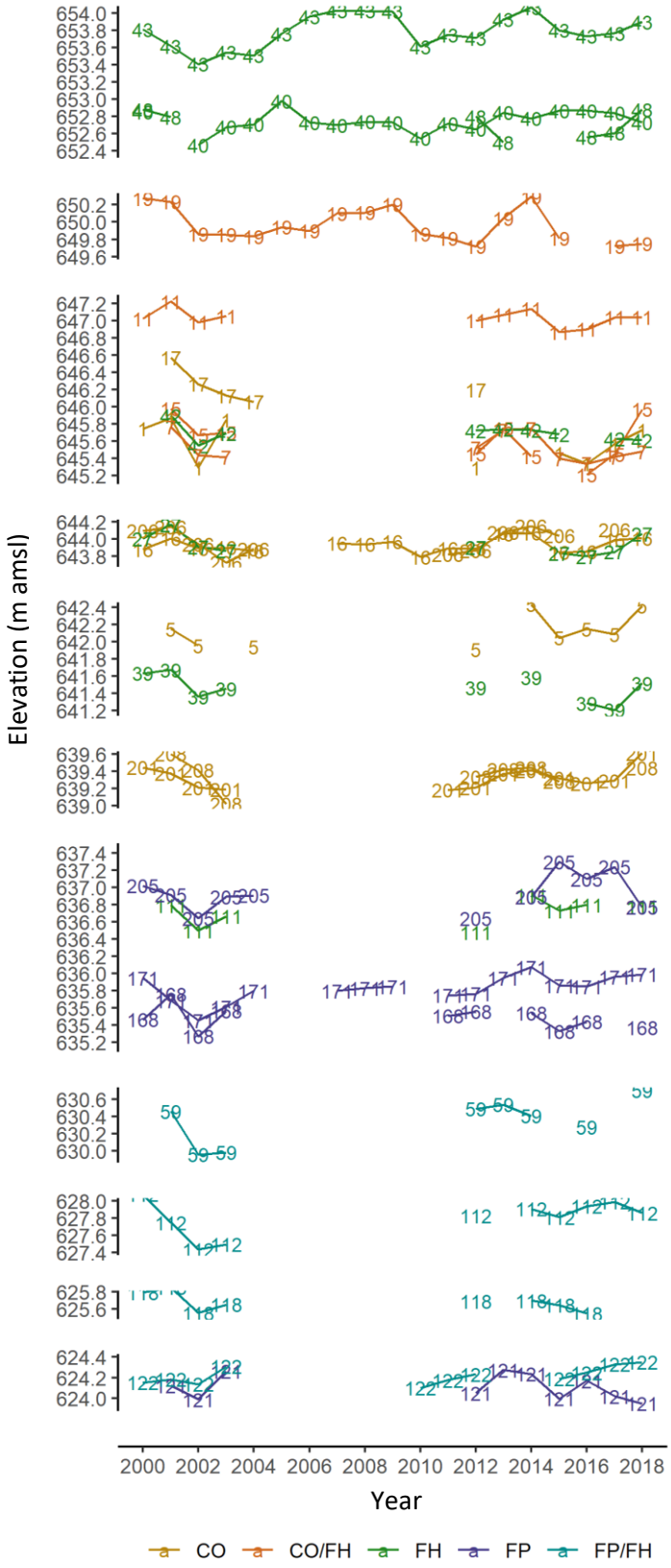
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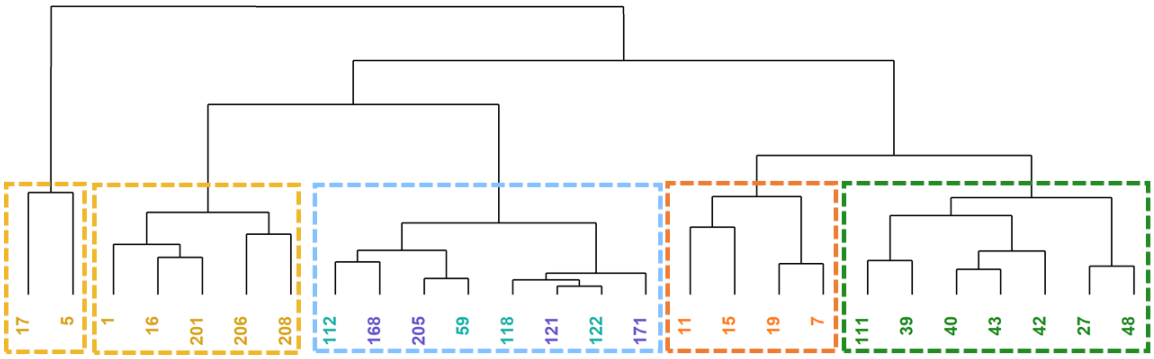
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7. Supplementary Material



Supplementary Figure 2-1. Mean summer (1st June – 31st August) lake water level observations 1999-2018 above mean sea level of 26 study lakes in the URSA. Colour indicates surficial geology classification of lake (HRA).



Supplementary Figure 2-2. Agglomerative hierarchical clustering of lake and catchment topology (data given in Table 2-1.) variables of 26 study lakes in the URSA. Lake label colours indicate the HRA class of lakes based on surficial geology, which closely correspond with types of lake formation. Dashed boxes indicate groups identified by hierarchical clustering.

CHAPTER THREE: CONTROLS ON SHALLOW LAKE WATER LEVEL RESPONSES TO DECADAL CLIMATE CYCLES IN CANADA'S BOREAL PLAINS

Abstract

Shallow lakes are significant ecosystems across glaciated, low relief interior regions and are highly susceptible to large water level fluctuations caused by climate and anthropogenic pressures. Seasonal and decadal climate cycles within the sub-humid Boreal Plains (BP) result in decadal lake water level patterns and temporary shifts to terrestrial wetland phases. Before we can understand the impacts of disturbance, we must first define natural variability and improve understanding landscape controls. Interactions of geology, land cover and landscape position in heterogeneous glaciated landscape may result in contrasting hydrological thresholds and connectivity throughout the landscape. There is a need to develop lake management frameworks built on process understanding of interacting controls on lake hydrology. Whilst individual components of the water budget and bathymetry directly determine lake level deviations, the interaction of lakes with their catchment, geology and climate provide the underlying control on these water level deviations and form the foundation of frameworks. This study tests and refines concepts and hypotheses of BP hydrological functioning, using data sets which transcend long term climate cycles and are inclusive of all landscape configurations. Analysis of landscape and lake characteristics show that hydrogeological setting drives scales of lake-groundwater connectivity and interactions with land cover and lake geometry further modifies hydrological connectivity, lags and memory of BP shallow lakes. This insight into the dominant controls and sources of flow in low relief glaciated regions can be incorporated within physically based models by deriving empirical functions that link hydrologic function with catchment structure, geology, landcover and landscape position for each classification.

1. Introduction

Shallow lakes are the most abundant lake type globally, but have been historically underestimated (Meerhoff and Jeppesen, 2009, p645), despite greater productivity, fish biomass and species richness per unit area than deeper lakes (Bayley and Prather, 2003). They cover expansive areas and support adjacent wetlands, forests and downstream ecosystems (Ireson *et al.,* 2015; Thompson, Mendoza, and Devito, 2017; Leibowitz, Mushet, and Newton, 2016), providing major critical habitat for mammals, aquatic organisms and the world's richest bird communities (Macdonald *et al.,* 2006; Morissette *et al.,* 2018). Shallow lakes naturally promote spatial diversity in lake water level that is critically important for bird habitat, as modifications to lake area, depth and hydroperiod impact on availability and quality of suitable nesting and breeding grounds (Bellisario, Cerfolli, and Nascetti, 2014). But by nature of their lower volumes, shallow lakes are more vulnerable to drying out, and are more susceptible to catastrophic shifts in ecosystem states, whereas deep lakes gradually respond to environmental pressures (Scheffer *et al.,* 2001). This potential for catastrophic shifts is magnified in low relief dominant interior ecosystems, where the distributions of shallow lakes, high evaporation:inflow ratios (Fergus *et al.,* 2020), and highly susceptibility to high non-stationary climate-change velocities (Stralberg, Carroll, and Nielsen, 2020) are overlapping. This is particularly of consequence in Canada's Boreal Plains (BP); lake water levels are predicted to deplete over the next century, and are at risk of permanent terrestrialization (Thompson, Mendoza, and Devito, 2017; Bertassello *et al.,* 2019), where climate pressures are being compounded by local land use pressures. The footprint of natural resource sectors including oil and gas, mining, agriculture and timber harvesting is set to increase by 50-60% by 2030 in Canada's Boreal (Webster *et al.,* 2015). In addition to their significant demand on water, natural resource extraction and associated road networks significantly compromise the essential role of hydrological connectivity in ecosystem services (Leibowitz, Mushet, and Newton, 2016; Cobbaert, Wong, and Bayley, 2014), including climate and water regulation, food supply, nutrient cycling and habitat provision (Webster *et al.,* 2015).

Despite a sub-humid climate, shallow headwater lakes and lake-wetland complexes are ubiquitous across the Boreal Plain (BP) natural subregion of Alberta, Canada, due to low relief glacial deposition landscapes and poor drainage networks. At less than 2 meters deep, shallow lakes are commonly labelled wetlands (Canadian Wetland Classification; National Wetlands Working Group, 1997) as shallow depths enable domination of submerged (Scheffer and Van Nes, 2007) and hydrophytic vegetation (Rydin and Jeglum, 2013, p3). This ambiguity in definitions leads surface cover estimates to range from ~10% for ponded water up to 50% for wetlands (Kuhry *et al.,* 1993; Vitt, Li, and Belland 1995). BP shallow lake water levels are misconstrued as persistent due to more humid climates relative to the semi-arid Northern Prairies Region (Hayashi, Kamp, and Rosenberry, 2016). However, within the BP, fifth order streams dry up during dry periods in multi-decadal weather cycles, and runoff return periods are greater than 20 years (Devito, Mendoza, and Qualizza, 2012). As part of this cycle, BP shallow lakes demonstrate longer fluctuations in water levels (Chapter 2; Van Der Kamp, Keir, and Evans, 2008) and undergo terrestrial wetland phases (Devito *et al.,* 2012). Therefore, before we can assess ecosystem sensitivity to hydrologic disturbances (Mirus *et al.,* 2017), to infer the stability or persistence of these pond-wetland systems (hereafter termed "shallow lakes"), we first require understanding and definition of "normal" long term natural ranges (Chapter 1) and the controls on the translation of existing weather patterns in shallow lakes water levels.

There has been extensive work undertaken across the globe examining fluctuations in lake water levels (e.g. Fergus *et al.,* 2020; Jones *et al.,* 2016; Nõges *et al.,* 2008; Owor *et al.,* 2011; Pasquini, Lecomte, and Depetris, 2008; Torabi Haghighi and Kløve, 2015) and Holocene lake-level changes (e.g. Harrison, Prentice, and Guiot, 1993; Holzhauser, Magny, and Zumbühl, 2005). Whilst individual components of the water budget and bathymetry directly determine lake level deviations (Hayashi and Van Der Kamp, 2000; Minke, Westbrook, and Van Der Kamp, 2010), the interaction of lakes with their catchment, geology and climate provide the underlying control on these water level deviations and form the foundation of frameworks. Climate modifies stream and groundwater inputs and evaporative losses

from lakes (Hayashi and Van Der Kamp, 2007) and non-stationary climatic forcing has caused lake levels to decline globally (Schindler, 2009), whereas cyclical weather patterns may produce long-term lake water level fluctuations or temporally variable hydroperiods in response to decadal meteorological cycles (Hayashi and Van Der Kamp, 2007; Molinos *et al.,* 2015; Thompson *et al.,* 2015). Within the BP, climate interactions with lake geometry governs evaporative losses and lateral losses to riparian vegetation (Hayashi *et al.,* 2016; Smerdon, Devito, and Mendoza, 2005). Inflow is controlled by scales of groundwater interaction between fine and coarse-textured surficial geologies (Plach *et al.,* 2016), and varying surface connectivity with adjacent wetland and forest landscape units (Devito *et al.,* 2017). Lower lake capacity-inflow ratios typically exhibit greater lake level stability (Torabi Haghighi *et al.,* 2016). Landscape position is important to determine position within flow systems and magnitude of groundwater inputs (Winter, 2001; Kratz *et al.,* 1997) and may indicate whether lakes function as discharge or recharge lakes (Van Der Kamp and Hayashi, 2009).

Numerous climate-geology frameworks have been devised to capture these processes (e.g. Dahl, Nilsson, Langhoff, and Refsgaard, 2007; Devito *et al.,* 2005; Euliss *et al.,* 2004; Wagener *et al.,* 2007; Winter, 2001), however, the application of these frameworks will fail to represent BP lakes due to the need for combined interactions between catchment, lake geometry, geology and climate controls, which have been largely overlooked to date. Studies incorporating heterogeneous hydrogeological settings often span different regions with different weather conditions or climatic zones (e.g. Ali *et al.,* 2015; Carey *et al.,* 2010; Gibson *et al.,* 2005; Gibson, Prepas, and Mceachern, 2002; Gibson, Yi, and Birks, 2020; Mwale *et al.,* 2011; Sprenger *et al.,* 2018; Tetzlaff *et al.,* 2015; Tetzlaff *et al.,* 2017; Welch, Smith, and Stadnyk, 2018; Winter, 2001); subsequently, long term studies on shallower systems that allow for teasing out climate-geology interaction are generally lacking.

BP lakes can borrow concepts from Boreal Shield and prairie hydrology of dynamic storage and fill-spill connectivity (Spence, 2007; Spence, 2010) as a simplified framework. Yet, the more hydrologically

permanent, primarily surface driven Boreal Shield lakes lack both long term memory observed in closed basins and larger scale groundwater flows seen in coarse glacial deposits due to steeper topography and absence of deep, heterogenous glacial deposits (Oswald, Richardson, and Branfireun, 2011; Peters *et al.,* 1995; Spence and Woo, 2006). There is well-established management and understanding of hydrologic memory and lags associated with closed basins in the semi-arid Prairie Potholes Region (PPR) (Conly and Van Der Kamp, 2001; Hayashi *et al.,* 2016; Labaugh *et al.,* 2016, 2018; Labaugh, Winter, and Labaugh, 1998; Leibowitz *et al.,* 2016; Shaw *et al.,* 2012; Shook and Pomeroy, 2011; Van Der Kamp, Keir, and Evans, 2008). However, PPR models of geographically isolated lakes with poor surface water connection (Gleason and Rooney, 2017; Bertassello *et al.,* 2019) cannot fully explain modifications to hydrologic memory and lags caused by semi-continuous peatland water supply within the cooler BP climate. Process understanding and controls on groundwater-surface water interactions has been studied extensively in coarse textured outwash sediments in Minnesota (Labaugh *et al.,* 1997), Wisconsin (Jaquet, 1976; Kratz *et al.,* 1997), and Nebraska (Labaugh, 1986; Winter, 1986) and can be largely transposed to coarse outwash lakes within the BP. However, the BP features more diverse scales of groundwater connectivity due to heterogeneous glacial deposits (Ferone and Devito, 2004; Plach *et al.,* 2016; Smerdon *et al.,* 2005). Therefore, there is a need to develop a lake management framework for the BP that builds on this understanding of interacting controls on lake hydrology gained from lake studies in glaciated landscapes.

The goals of this study were to test the controls on the temporal variation in lake water levels within decadal climate cycles, to evaluate hydrologic interaction of lakes with their catchment, geology and climate and to inform a framework based upon mechanistic understanding of lake water level. High PET within the subhumid BP climate represents the dominant water loss mechanism (Brown *et al.,* 2014; Devito *et al.,* 2005; Hokanson *et al.,* 2019; Redding and Devito, 2010; Redding and Devito, 2008) and lake evaporation likely provides a key control on lake water level dynamics, relative to any water inputs from the wider catchment through groundwater or near surface water. Therefore, we

hypothesise lake geometry and relative lake area to be important controls on evaporative losses (Hayashi *et al.,* 2016), in addition to climate conditions (Petrone, Silins, and Devito, 2007). Groundwater inputs are theorized to exhibit spatial differences in dominant controls, corresponding with different sources of groundwater flow between textures of surficial geology (Plach *et al.,* 2016). Greater groundwater dominance within coarse-textured is expected to buffer the climate signal and minimise lake level variability (Smerdon, Mendoza, and Devito, 2007); while greater peatland cover is hypothesised to maintain lake levels during droughts (Devito *et al.,* 2017; Ferone and Devito, 2004), however due to low storage capacities in peatlands, this flow may reverse and amplify lake decline in prolonged water deficit conditions. In contrast, hydrologically closed basins are expected to mimic climate (Hayashi and Van Der Kamp, 2007) due to limited flow contributions from the catchment. The Boreal Plains of Canada presents a significant opportunity to unravel these interactions of climate with contrasting geology, land cover, landscape position and lake geometry due to diverse configurations of these attributes within one climate signal.

2. Study Area

The study was conducted in the Utikuma Region Study Area (URSA; 56°N, 115°W), situated 370 km north of Edmonton, Alberta, Canada within the BP ecozone (Figure 3-1). The study region is characterised by a sub humid climate, with the long term annual average PET (513 mm) exceeding annual average P (483 mm) (Plach, Ferone, Gibbons, Smerdon, and Mertens, 2016; Environment Canada, 2003). Superimposed upon this average is a strong climate cycle, of wet, mesic and dry years, with wet years occurring infrequently on a 10–15 year cycle (Ferone and Devito, 2004); although 20 year frequencies have also been reported (Mwale *et al.,* 2011; Carrera-Hernández *et al.,* 2011; Plach *et al.,* 2016). This decadal climate cycle dominates the regional hydrology, with annual runoff over a

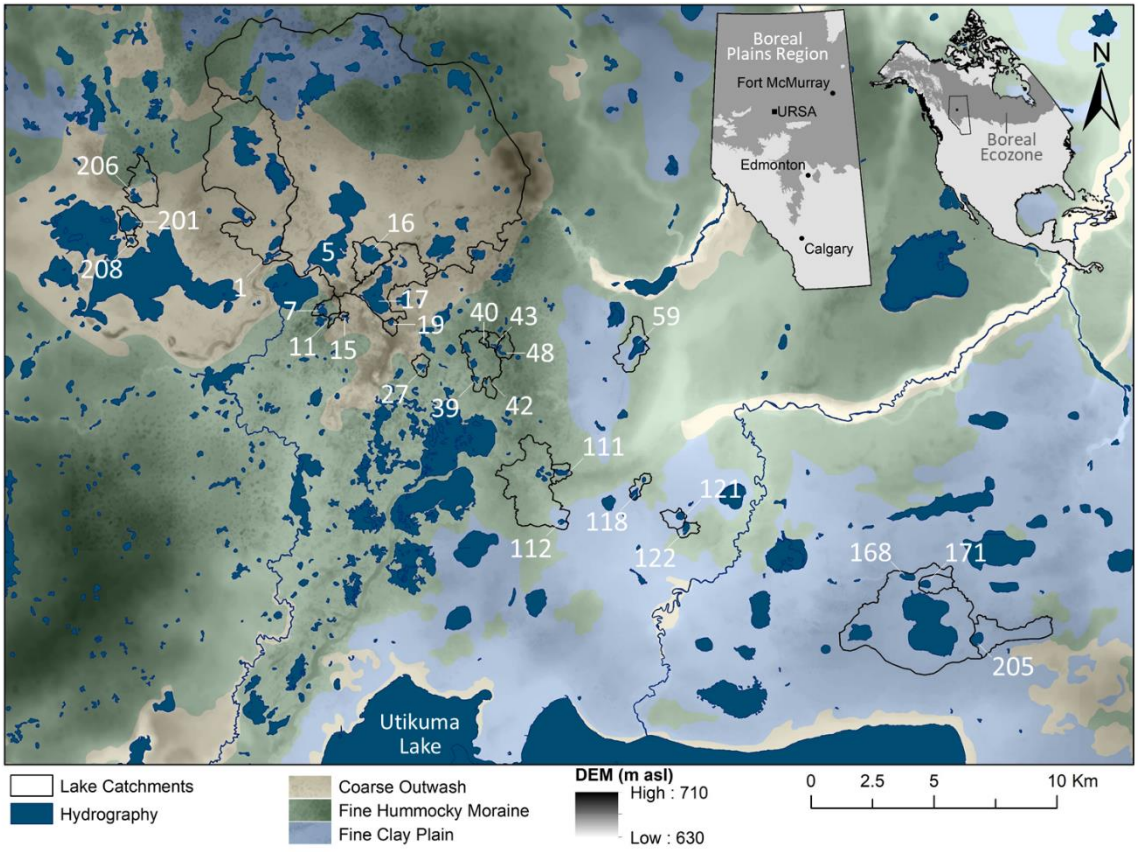


Figure 3-1. Site map of HRAs derived from surficial geology; hydrography, topography and surface catchment boundaries of 26 study lake catchments in the URSA (Utikuma Region Study Area), within the Boreal ecozone of Canada, and plains region of Alberta (Alberta Sustainable Resource Development, Alberta Environment, Alberta Community Development and Agriculture and Agri-Food Canada, June 2005).

decadal time scale varying in response by orders of magnitude (<5 to 300 mm; Redding and Devito, 2010).

URSA overlies deep (45-240 m), low relief Quaternary glacial deposits (Hokanson, Mendoza, and Devito, 2019), and the Smoky Group shale unit of the Upper Cretaceous period (Vogwill, 1978; Ceroici, 1979; Plach *et al.,* 2016). The surficial geology varies spatially and can be simply categorised into three primary classes (Devito *et al.,* 2017): (1) coarse-textured glacial–fluvial outwash (CO); (2) fine-textured clay rich glacio-lacustrine plains (FP); (3) fine-textured hummocky disintegration moraine deposits (FH) and two transitional classes: (4) the transition from hummocky disintegration moraine to glacial fluvial outwash (CO/FH); (5) the transition from disintegration moraine to glacio-lacustrine plains (FP/FH). URSA exhibits very low topographic relief (<80 m), characteristic of the BP. On the surface, the

landscape is characterised by a patchwork of forestlands and peatlands that vary in their dominance and size through the landscape. Forestlands are characterised by low productivity pine, predominantly situated upon well-draining sandy regions, and a combination of aspen and spruce forests (white and black spruce), present within finer grained mineral soils. In both fine and course texture regions, peatlands are characterised by the accumulation or organic soils, and a forest canopy of black spruce and tamarack.

The URSA region has been exposed to a range of natural and anthropogenic disturbance. Since 1950 harvesting represents 5.3% of the landscape within the URSA region (ABMI, 2019; see appendix) and roads and seismic lines for oil and gas exploration are also prominent (Strack *et al.,* 2019). Wildfire is also an episodically occurring feature of the landscape. The most notably fire within the regions was the ~90,000 ha Utikuma Complex wildfire (Kettridge *et al.,* 2019; see appendix) that burned in the spring of 2011. The Utikuma Complex wildfire burned a large proportion of URSA region; notably both the fine textured hummocky disintegration moraine and fine textured glacio-lacustrine plains regions to the south of the URSA region. Excluding the Utikuma Complex fire, with the most recent extensive wildfire to burn within the URSA regions was in 1985.

3. Methodology

3.1. Sampling strategy

Along a 50 km transect that spans the URSA, 26 study lakes were measured and sampled between 1999 and 2018 (Figure 3-1). Lakes were selected based on a systematic sampling strategy that identified ponds that were located within different HRAs, with lake catchments that have a range of different peatland land cover proportions (low and high) and located within catchment at different landscape positions (low and high elevation). This sampling strategy was defined to represent hypothesised key controls on catchment hydrological function identified by Devito *et al.,* (2017) and

hypothesized controls on lake hydrological function from local scale investigations (Plach *et al.,* 2016; Smerdon *et al.,* 2005; Brown *et al.,* 2014; Devito, Creed, and Fraser, 2005; Devito *et al.,* 2005; Hokanson *et al.,* 2019; Redding and Devito, 2010; Redding and Devito, 2008). For further information with regards to lake and associated catchment attributes, see supplementary material. Whilst this three-dimensional sampling strategy represents the idealised approach, not all components of the factorial analysis were present within the URSA landscapes. Most notably, lake catchments at a high landscape position located within coarse HRAs.

3.2. Hydroclimatological measurements

Climate of the URSA and wider study region has been monitored for 23 years (1995-2018) with the use of tipping rain bucket gauges, throughfall gauges, snow buckets and evaporation pans, together with local government weather stations at Peavine and Red Earth Creek, to produce a dataset for daily precipitation values and mean, maximum and minimum air temperatures for the 20-year period. Local government stations are used prior to 1999 and URSA data from 1999 onwards. Precipitation was collected using 2-3 tipping bucket rain gauges and adjusted for snowfall by using antifreeze in the reservoirs and validating against snow buckets. Peavine and Red Earth Creek local government weather stations were utilised for missing data based on which station tracked URSA local weather best in the given period. Potential evapotranspiration (PET) was then estimated using Hamon's method (Hamon, 1963) from day lengths and daily mean temperatures within the 23 year climate records.

A staff gauge was installed within each of the 26 lake catchments in 1998. Staff gauges were complemented with a diverse range of piezometers and wells installed in these lakes and their associated peatlands and forests to monitor groundwater levels and pressure heads (see Devito *et al.,* (2016) for further details of wider research activities). All replacements of instrumentation during the study period were related to this original staff gauge. Lake level measurements relative to the staff

gauge were manually collected each summer across all sites and, from measured stake elevations, used to determine the 'water level', lake water level elevation above sea level. In addition, some water levels were measured multiple times through the summer, and others logged by pressure transducers installed in staff gauges (in addition to wider wells and piezometer installed at the study sites).

Annual sampling trips were undertaken at the last week of June to first week of July to target peak lake water levels. The summer lake water level was assumed equal to the summer maximum (June 1st to August 31st). This period was used to capture peak water levels and maximise the number of years and number of lakes within the dataset, however long-term lake water level metrics are insensitive to the data window selected (see supplementary material). Where lake measurements related to the staff gauge were not available in a given year, water levels in wells or shallow piezometers (< 1 m depth) that are both in close proximity to the lake and follow similar dynamics to the lake water levels were used. Where such measurements were not available, water levels of the given lake were excluded from the analysis. As with all long-term data records, gaps are evident within this long-term record for some sites. These gaps are expected to have a minimal impact on the subsequent analysis, occurring in years of limited extremes (2005-2010), but should be borne in mind in subsequent analysis.

Surface water chemistry was sampled within each lake every July (2000-2018), with some lakes sampled more frequently throughout the ice-free period. At each lake, three surface water samples were collected approximately 30 cm below the water surface and 3 m from the lakeshore edge using a sampling pole. Samples were analysed for stable isotopes $\delta^{18}O$ and $\delta^{2}H$, and major cations and anions including Ca^{2+}, Mg^{+}, Na^{+}, K^{+}, CO_3^{2-}, HCO_3^{-}, SO_4^{2-}, Cl^{-}. Hydrochemical and isotopic analysis were undertaken at Biogeochemical Analytical Service Laboratory, University of Alberta, Edmonton, AB, Canada (http://www.biology.ualberta.ca/basl/). In situ measurements of EC, temperature and pH were taken in the field with a calibrated temperature/level/conductivity meter (Solinst, Canada).

Analysis of the summer lake water level focuses on the deviation in the lake water level from its level in 2012, here defined as the water level deviation (WLD). WLD was defined relative to the year 2012, when runoff peaked regionally, and study lakes were observed as generating outflow (see definitions in Table 3-1). This year therefore provides a benchmark for maximum water level thresholds regionally. Metrics representing variability in this WLD for each lake were then used to calculate metrics for long term lake water level variability (WLD_V) including range (WLD_R), standard deviation (WLD_{SD}), interquartile range (WLD_{IQR}), mean absolute deviation (WLD_{MAD}) and interdecile range (WLD_{IDR}). Pearson's correlation was then calculated between WL_V metrics to assess the suitability of each metric for further analysis.

3.3 Lake and Catchment Characteristics

A diverse range of metrics representing characterises of each lake, catchment, and the wider landscape and position within have been derived. Exact details of each metric and the rational for its measurement (its hypothesised influence on lake water level deviations) is outlined in Table 3-1. Below we provide further detail of the classification and measurement approaches applied to derive each metric.

HRA Delineation

Hydrologic Response Areas (HRAs) are landscape classifications of surficial geology with similar transmission and storage properties (Devito *et al.,* 2012). We define five HRA classifications; two fine-textured classes of clay plain (FP) and hummocky moraine (FM), one coarse-textured class of coarse outwash (CO), with two further transitional classes: (4) the transition from fine hummocky moraine to coarse outwash (CO/FH); (5) the transition from fine hummocky moraine to fine clay plains (FP/FH). These HRAs were derived primarily from the texture of mapped surficial geology (Alberta Geological

Survey; Fenton *et al.,* 2013), or informed by geomorphic class and landform genesis where texture was not provided (Atkinson, Utting, and Pawley, 2014; Table 3-1).

This results in substantial differences in the ground water hydrological properties between regions with a coarse and fine surficial geology; hydraulic conductivity in the coarse and fine textured surficial geologies averaging 4.6×10^{-5} and 4.2×10^{-7} ms^{-1}, respectively (Hokanson *et al.,* 2019). Within these broad HRAs, we defined sub-units from an assessment of regional flow systems through the interpretation of a number of data layers including digital elevation models, satellite imagery, and field-based interpretation.

Table 3-1. HRA classification derived from texture and geomorphic class and landform genesis, from Alberta Geological Survey surficial geology mapping.

HRA	Texture	Geomorphic class/ landform genesis
Coarse Outwash (CO)	Coarse: • Gravel • Sand • sandy-gravel • silty-sand • silty-gravel	• fluvial • glacial-fluvial • aeolian • littoral glacio-lacustrine
Fine Hummocky Moraine (FH)	Fine: • silt • clay • sand-silt-clay • sandy-silt • sandy-clay • silty-clay • clayey-silt	• non-sandy stagnant ice moraines • moraines with hummocky geomorphic class
Fine Clay Plain (FP)		• lacustrine • distal glacio-lacustrine • thrust moraine

Lake Catchment Boundaries

Catchment boundary identification within the BP is a difficult task. The low relief and deep substrates means that traditional flow accumulations models, whilst able to show broad patterns at the landscape scale, fail to accurately replicated catchment systems. Therefore, delineation of catchment watersheds was undertaken manually from the detailed interpretation of a number of data layers: 2 x 2 m

resolution DEM and associated flow models (ArcMap), satellite imagery (ESRI, 2020), historical aerial photography, Boreal Plain Wetland Classification (Ducks Unlimited Canada; Smith *et al.,* 2007), local knowledge (such as placement of culverts), and observations of flow direction. Spatial analyst in ArcMap was utilised to compute 1 m contours and generate estimates of flow direction, flow accumulation and flow basins from the DEM.

Lake and Catchment Characteristics

Lake catchment delineations were used to calculate topographical metrics, lake catchment proportions and land cover ratios and disturbance impacts. Land cover areas and ratios reflect HU classifications of lake, wetland or upland forest, which represent three main ecosystem types of similar hydrologic function derived from 24 vegetation classes (DUC, 2011; see Devito *et al.,* 2017 for details), and reflect proportions observed in 2013 satellite imagery (or otherwise stated for lake area calculation above). Relative elevation (%) is calculated as the percentile elevation of the lake water level in 2012 within the HRA sub-unit DEM, whereas boundary distance (km) is the straight-line distance from the centre of the lake to the highest point in the HRA sub-unit in the flow system above lake. Topographic variables slope (%) and flat (%; area of catchment with slope <1%) were calculated by applying the slope function to the DEM within ArcMap, whereas relief is calculated by the highest elevation subtracted by the lowest elevation in a catchment to show elevation change within a catchment. Percentage of burnt catchment area and burnt wetland area within the 2011 Utikuma region fire were calculated to examine any potential impact of fire. The percentage of harvested catchment was calculated for the period of 2000-2017 (ABMI, 2019) to examine potential disturbance impacts on lake water levels.

Table 3-2. Variable definitions, their purpose for BP hydrological conceptualisation and indication of use in further analysis following testing for redundancy in conceptual model (Figure 3-4). Table of extended list of all variables explored given in supplementary material.

Variable of Interest		Description	Purpose	Used in PCA
Definitions	WL	Water Level = Water elevation above mean sea level	Time series of lake water levels above mean sea level	✖ - timeseries
	$Depth_W$	$Depth_W$= WL – Lake bottom elevation above sea level (m)	Time series of the lake depth of water	✖ - timeseries
	WLD	Water Level Deviation WLD = WL(t) – WL(2012) (m)	Time series of lake water level deviation from the water level in 2012 (when the region is generating runoff).	✖ - timeseries
	D_{2012}	Observed depth of water at the centre of the lake in July 2012	Indicate potential storage, and capacity for flow generation and retention within the catchment during a wet period within the study	✖ - timeseries
Lake Water Level Variability	WLD_R	Range of mean annual lake WLDs (water level deviation) WLD_R = MAX(WLD) – MIN(WLD)	Indicator for long term hydrologic magnitude of variability of lakes	✓
	WLD_{SD}	Standard deviation of mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✖ - significantly correlated with WLD_R
	WLD_{MAD}	Mean absolute deviation in mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✖ - significantly correlated with WLD_R
	WLD_{IDR}	Interdecile range of mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✓
	WLD_{IQR}	Interquartile range of mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✖ - significantly correlated with WLD_R
Lake Characteristics	$LakeClass_C$	The dominant (50% surficial geology of the lake perimeter; within a 100m buffer). Binary score of Fine = -1, coarse = 1	Indicator for scales of groundwater interactions	✓
	A_L	Area of lake (m^2)	Indicator of lake storage Proxy for lake volume (assuming straight sided bathymetry) (in 2012?)	✖ - incorporated within ratio variables
	P_L	Perimeter of lake (m)	Indicator for potential for lateral water loss or gain	✖ - significantly correlated with all lake characteristics
	H_{Sill}	Lake Sill height = Sill Elevation-Lake Bottom Elevation For Lake with an outflow, SillElevation = WL_{Max}. Else, for lakes with no outflow, SillElevation calculated from DEM.	Indicator of maximum lake level (control on outflow to define fill and spill dynamics)	✓
	V_L	Estimated maximum potential lake volume (area * sill height, assuming straight sided bathymetry)	Indicate lake potential storage	✖ - significantly correlated with all lake characteristics
Landscape Position	Rel Elev	Percentile relative elevation within local HRA sub-unit (%)	Potential for flow accumulation	✓
	A_{CGW}	Area of hyporheic catchment of lakes within coarse textured surficial geology	Potential for groundwater interactions – larger area also indicates lower landscape position in coarse HRA	✖ - significantly correlated with relative elevation
Surficial Geology Ratios	A_{CGW}: A_L	Ratio of area of groundwater capture (hyporheic catchment) of coarse textured lakes, relative to lake area	Potential groundwater interactions	✖ - significantly correlated with groundwater indicator $LakeClass_C$
	A_F: A_L	Ratio of area of fine textured surficial geology within topographic catchment to lake area	Indicate surficial geology texture	✖ - does not distinguish between storage (FH/aspen) and flow accumulation (FP/wetlands)
	A_{FH}: A_L	Ratio of area of fine hummocky surficial geology within topographic catchment to lake area	Indicator for storage relative to lake (higher FH associated with higher slope)	✓
Land Cover Ratios	A_{TC}: A_L	Ratio of lake area to topographically delineated catchment area (excluding study lake)	Lake storage potential relative to the potential for flow accumulation - counter hypothesis to connected wetland ratios driving lake WL response	✓
	A_W: A_L	Ratio of lake to wetland area	Lake storage potential relative to the potential for flow accumulation	✓
	A_{CW}: A_L	Ratio of lake area to "connected" wetland area	Storage of lake relative to the potential for flow accumulation from connected, flow generating HU	✖ - significantly correlated to A_W: A_L and A_{TC}: A_L
	A_{OW}: A_L	Ratio of lake area to area of open water above the study lake the catchment	Proxy for water loss within the catchment	✖

Lake surface areas and perimeters were defined for 16 of the lakes from 2013 satellite imagery (ESRI, 2020); the remainder obtained from 2015 imagery due to availability of high-resolution satellite imagery. Year 2013 represents mesic climate conditions. It also represents a period where annual P-PET approximately equal the 3-year CDM P-PET (P-PET = 43 mm, 3yr CDM P-PET = 29 mm). Therefore, the mesic climate conditions should be reflected in the landscape state too, although this makes assumptions about climate memory within the landscape. Year 2015 represents a similar mesic state, and straight sided lake bathymetry results in negligible change in lake surface area between these years, as is evident from remote sensing (Montgomery *et al.,* 2019). Sill height for lake outflow was assumed to be the maximum observed lake depth (1999-2018) in lakes with outflows. Sill heights for lakes without outflows were calculated using a DEM within ArcMap. Lake volume is estimated as a function of sill height, lake area and a parabolic depression profile to the represent the potential lake storage (Minke *et al.,* 2010; Hayashi and Van Der Kamp, 2000).

3.4. Data Analysis: PCA Analysis

Parameters chosen to be included within the PCA reflect different aspects of land cover, landscape position and topography in addition to catchment scales and WL_V. Prior to analysis, all variables were scaled, and non-normally distributed variables were identified with Shapiro-Wilk normality tests and log transformed accordingly. Independent variables were tested for collinearity, and variables with significant correlations were omitted.

3.5. Data Analysis: Hydrochemistry

Local meteoric water line (LMWL) and local evaporation line (LEL) equations were derived from dual isotope plots (cf. Hokanson *et al.,* 2019). Line conditioned excess (lc-excess) was calculated from the LMWL:

Equation 1. $lc-excess = \delta^2H - a \cdot \delta^{18}O - b;$

where *a* is the slope and *b* is the intercept of the dual isotope plot, equal to 7.25 and -11.37 for URSA, respectively (Hokanson *et al.,* 2019). Lc-excess is used as an indicator for isotopic enrichment, from which dominance of groundwater and evapotranspiration can be inferred. Mean lc-excess values close to 0 indicate isotopic depletion, more negative values indicate greater enrichment, and low standard deviation in lake lc-excess indicates stable water supply through groundwater dominance. LELs were also fitted to each lake using a linear model, where the slope of the fitted model represents an additional indicator of isotopic enrichment to lc-excess.

Major ion concentration plots were employed to give insight in to the differences in water composition and balance of absolute concentration in alkaline earths (Ca^{2+}+Mg^{+}) and alkali metals (Na^{+}+K^{+}) relative to differences in weak acidic anions (CO_3^{2-}+HCO_3^{-}) and strong acidic anions (Cl^{-}+SO_4^{2-}) (Chadha, 1999; see supplementary material). These approaches are intended to classify lake catchments into hydrochemical facies, and in turn associate the spatial variability in these chemically homogenous groups with catchment characteristics and connectivity (Güler *et al.,* 2002). Absolute concentrations of Ca^{2+}+Mg^{+} were plotted against mass of Ca^{2+}+Mg^{+} to infer processes causing changes in ion concentrations. Mass of Ca^{2+}+Mg^{+} is calculated as the product of the lake parabolic volume (Minke *et al.,* 2010) and the concentration of (Ca^{2+}+Mg^{+}). Linear models were then fitted for each lake, where a high slope coefficient of the linear model indicates increased Ca^{2+}+Mg^{+} mass through increased supply due to groundwater dominance. A low slope indicates little or no increase in mass with increase in concentration, indicative of evapoconcentration controls increasing ion concentrations.

4. Results

4.1. Climate

Long term climate records of Fort McMurray reveal decadal cyclical weather patterns (Figure 3-2a). Wet years (PET:P ratio < 0.9) occur infrequently, returning every ~20years, whereas mesic years (PET:P ratio of 0.9-1.2) and dry years (PET:P ratio >1.2) cycle every ~4 years (defined in Hokanson et al 2019, Devito *et al.,* 2017, 2012; Buttle *et al.,* 2005). This produces a long-term moisture deficit with periods of excess moisture (P-ET), which is typical of the climate of the Boreal Plains subregion. The study area climate is comparable with the 100-year record of Fort McMurray (Figure 3-2b); the study period captures mesic and dry conditions, with a wet period (1996-1997) a couple years prior to the study. Therefore, the 20-year study of lake water levels captures full weather cycle conditions observed within the BP.

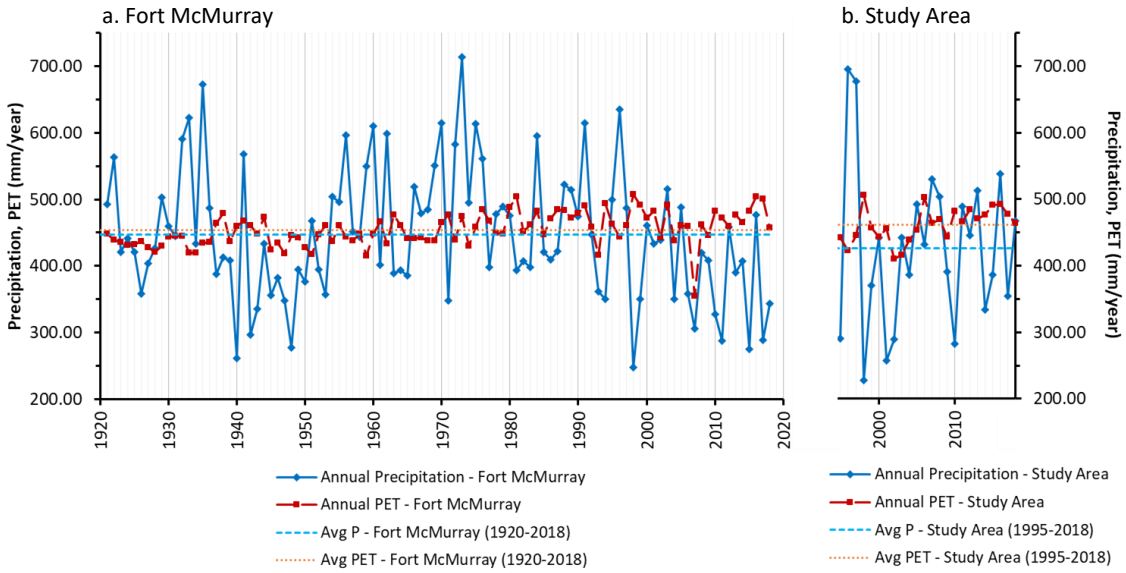


Figure 3-2. Long term climate of the Boreal Plains region of Alberta, Canada: a) Fort McMurray, 1920-2018; b) Study Area, 1995-2018. Annual totals represent the hydrologic year (1st November-31st October). Potential evapotranspiration (PET) is calculated using Hamon's equation.

4.2. Surficial Geology Control

HRA classifications are largely overlapping in their distributions of individual WLD metrics (Figure 3-3). However, differences between metrics indicate surficial geology control on response is relative to the range of weather conditions. FP lakes display higher WLD_R than FH lakes but comparable WLD_{IDR} and WLD_{IQR} values. FP lakes also show high within class similarity of WLD_{IQR}. This indicates greater or cumulative responses to extreme weather conditions in FP, but otherwise stable lake water levels, whereas extreme responses are not as evident in FH lakes. CO lakes display mid variability for all WLD metrics, contrary to hypothesis of regional groundwater supply causing lowest WLD. Transitional classes CO/FH and FP/FH exhibit the most within class variability, indicating more extreme WLD than their 100% HRA counterparts. CO/FH exhibit the highest WLD for all metrics as hypothesised, due to

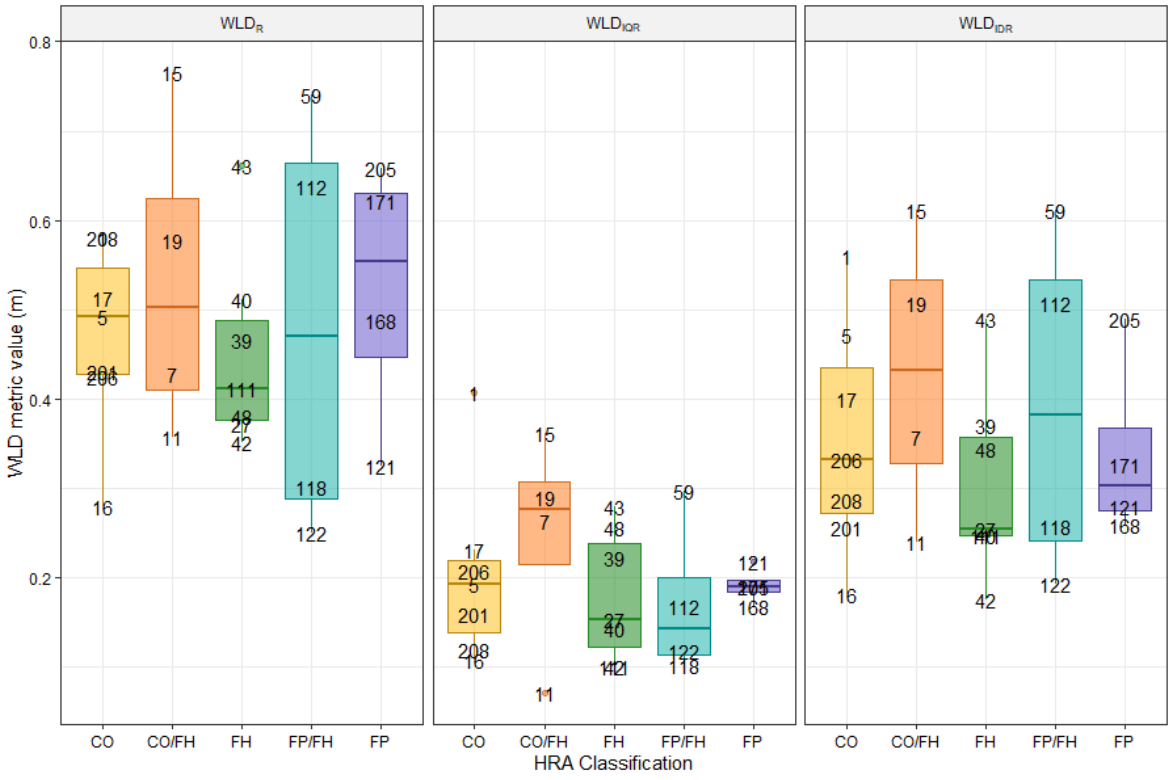


Figure 3-3. Lake hydrology metrics compared between HRA classifications. Each metric is calculated from the maximum lake WL observed within summer months (June 1st-August 31st) each year from 2000-2018 where: i) WLD_R is the maximum minus the minimum deviation in lake WL from the 2012 lake WL; ii) WLD_{IQR} is the interquartile range of deviation in lake WL from the 2012 lake WL; iii) WLD_{IDR} is the interdecile range of deviation in lake WL from the 2012 lake WL.

their high topographic locations and absence of lake outflows creating hydrologically disconnected systems that mimic climate.

4.3. Univariate Analysis

No strong relationships were identified in univariate analysis between WLD metrics and independent variables (Figure 3-4). Spearman's rank tests show that land cover ratios produce moderate

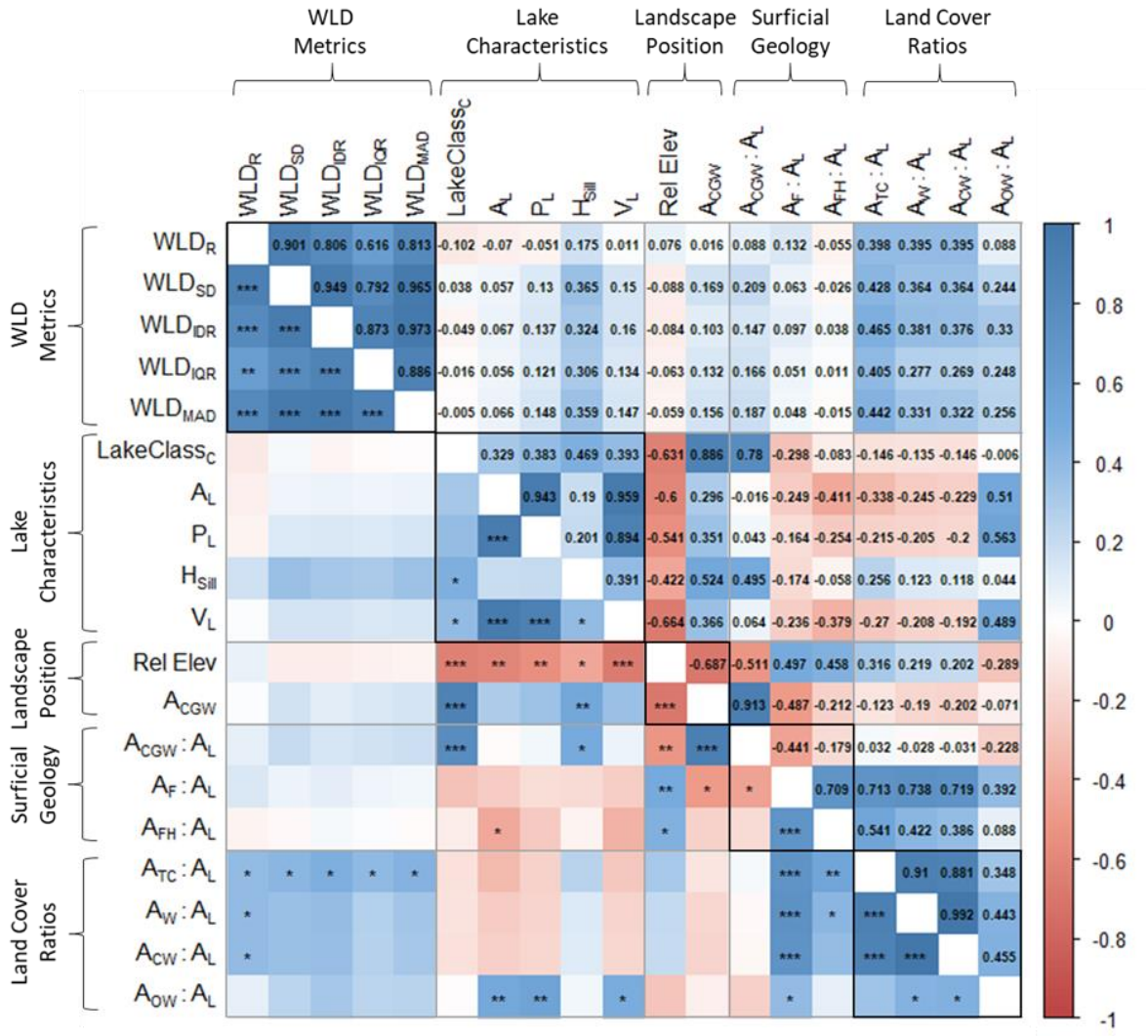


*Figure 3-4. Correlogram of spearman rank tests between all dependent variables and all independent variables hypothesised as important within conceptual framework (full set of variables provided in supplementary material). Collinearity between independent variables highlights any redundancy between variables within conceptual model. Spearman's rho statistic given in upper triangle and by shading, significance is given in the lower triangle ('***' denotes p ≤ 0.001, '**' denotes p ≤ 0.01, '*' denotes p ≤ 0.05). Correlogram of extended list of all variables explored provided in supplementary material.*

relationships with WLD, whereby higher WLD is associated with higher A_W:A_L, A_{CW}:A_L and A_{TC}:A_L ratios. All lake, landscape position, and surficial geology characteristics exhibit weak relationships with WLD metrics across all study lakes due to interactions with surficial geology texture or HRA classification (Figure 3-5). Relative elevation displays a general within HRA pattern of higher WLD_R and WLD_{IDR} at

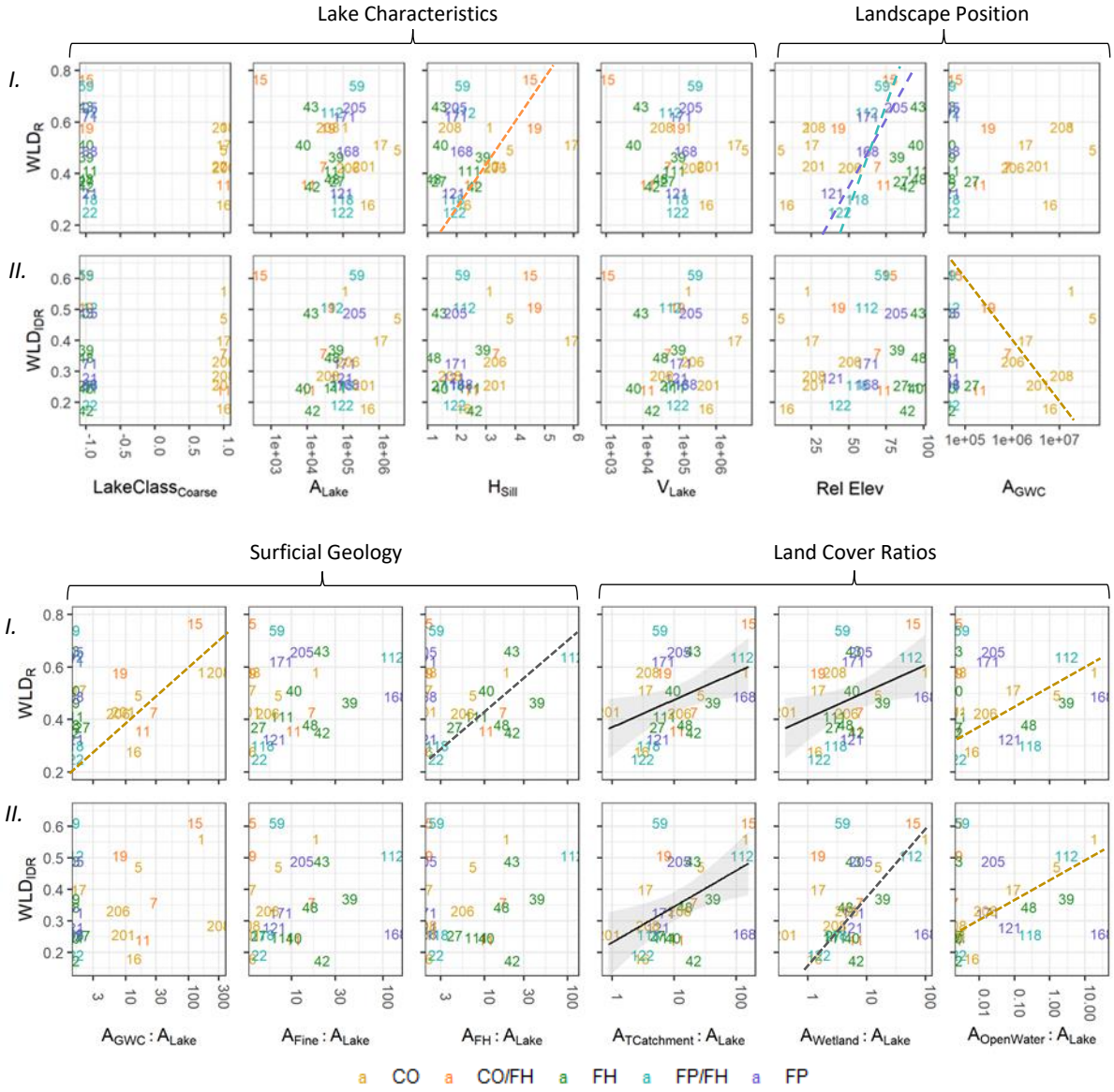


Figure 3-5. Scatter plots between dependent variables: I) WLDR ; II) WLDIDR and independent variables including: lake characteristics, landscape position characteristics, catchment surficial geology characteristics and catchment land cover characteristics. Label colour indicates HRA classification and label text indicates lake ID. Solid black line fitted using generalised linear model (grey band denotes standard error) for significant correlations with spearman's rho >0.3 (Figure 3-4). Dashed lines are manually drawn to qualitatively describe clear relationships that are repressed by outliers or interactions with HRA classification; coloured dashed lines correspond with HRA class, grey dashed lines indicate a relationship across multiple HRA classes.

higher positions. This is most apparent in FP and FP/FH HRAs, indicative of greater importance of surface connectivity in the flatter clay plains catchments. Higher A_{GWC}:A_L is associated with higher WLD_R in CO and CO/FH classes (r = 0.43 when 0 A_{GWC}:A_L values removed), demonstrating the importance of groundwater contributions relative to lake size in coarse textured surficial geology.

CO also demonstrates higher WLD with higher A_{OW}:A_L, due to greater evaporation losses associated with open water. Higher A_{FH}:A_L ratios are associated with higher WLD_R (r = 0.36 when 0 A_{FH}:A_L values removed), implying greater WLD in smaller FH lakes. While A_W:A_L and A_{TC}:A_L ratios demonstrate a significant positive relationship with WLD across all HRAs. These results indicate that there is no singular overarching control on lake WLD.

4.4. Interactions Analysis

Univariate analysis indicated no dominant control of any independent variable on WLD, indicating the potential importance of interactions. In preparation for multivariate testing, a correlogram (Figure 3-4) tests for redundancy of important independent variables within the conceptual framework (see Table 3-2). All lake characteristics and surficial geology variables are moderately associated with relative elevation, whereby lake size decreases with higher landscape positions, and CO exhibit larger lakes. Land cover variables are weakly associated with landscape position. Lake proportions (A_L, P_L, V_L) display high collinearity. Due to this redundancy, A_L is considered relative to surficial geology and land cover variables as ratios in multivariate analysis. A_{FH}:A_L is favoured over A_F:A_L to distinguish between FH and FP; interactions with H_{sill} and $LakeClass_C$ will also distinguish CO, providing a more stable dataset to analyse. A_W:A_L and A_{CW}:A_L represent important hypothesis variables regarding cumulative flow. To eliminate redundancy, A_W:A_L is selected due to the resolution of study, in addition to evidence that even isolated wetlands can influence catchment hydrology (Cohen *et al.,* 2016). A_{TC}:A_L provides the alternative hypothesis variable of topographic control, although A_{TC}:A_L is strongly collinear with A_W:A_L.

The PCA (Figure 3-6) within this study provides a descriptive technique for examining the interactions between lake and catchment characteristics and with lake WLD (90% explained variance across 4 PCs). The dominant loadings of each PC (Table 3-3) strongly represent different key conceptualisations of the landscape hydrologic function: PC1 is strongly associated with cumulative flow (A_W:A_L; 38.5% explained variance); PC2 signifies groundwater dominance with surficial geology texture indicators (H_{Sill} and $LakeClass_C$; 27.3% explained variance); PC3 represents catchment storage and aspen cover (A_{FH}:A_L; 15.4% explained variance); PC4 enables investigation of aspen-wetland interactions within HRA classes (A_{FH}:A_L and A_W:A_L; 7.9% explained variance). The dependent variables WLD_R and WLD_{IDR} also display high loadings for each PC presented, demonstrating the requirement of multiple dimensions to understand the interactive control of variables on lake water level variability.

Table 3-3. PCA loadings associated with Figure 3-6. The dominant loadings of each PC strongly represent different key conceptualisations of the landscape hydrologic function: PC1 = cumulative flow (wetlands); PC2 = groundwater (surficial geology texture); PC3 = aspen/storage (fine hummocky); PC4 = Aspen-wetland interaction. Table fill colours reflect direction of association with each PC (red = negative; green = positive) and strength of association with each PC (weaker association = lighter shades closer to 0).

	PC1	PC2	PC3	PC4
WLD_R	-0.400	-0.225	0.482	-0.179
WLD_{IDR}	-0.413	-0.313	0.310	-0.309
A_{TC}: A_{Lake}	-0.512	-0.043	-0.305	0.299
Relative Elevation	-0.238	0.517	0.154	-0.284
A_{FH}:A_{Lake}	-0.262	0.259	-0.541	-0.622
$A_{Wetland}$: A_{Lake}	-0.489	-0.048	-0.288	0.475
$LakeClass_{Coarse}$	0.205	-0.457	-0.414	-0.246
H_{Sill}	-0.036	-0.551	-0.090	-0.171
Standard deviation	1.755	1.479	1.110	0.797
Proportion of Variance	0.385	0.273	0.154	0.079
Cumulative Proportion	0.385	0.658	0.812	0.891

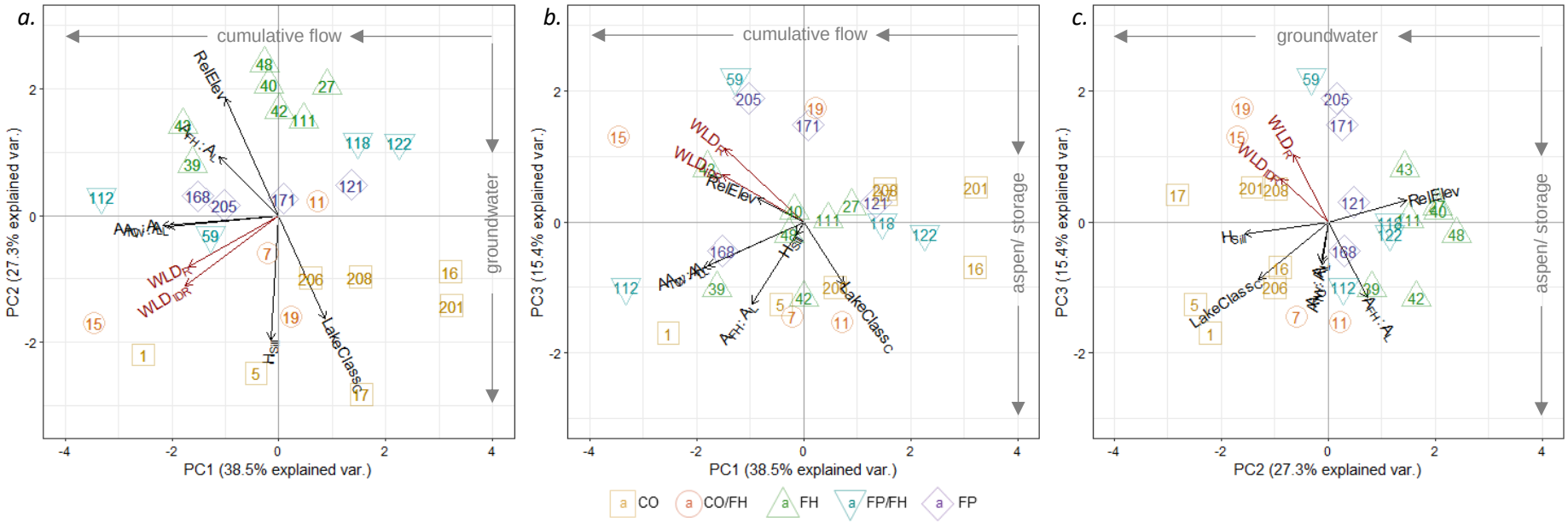


Figure 3-6. Principal component analysis (PCA) of lake and catchment characteristics (black loadings arrows) and lake WLD metrics (red loadings arrows). Data point shape and colour correspond with HRA classification and are labelled by lake. Multiple PCs are presented due to demonstrate the multidimensional interactions between lake and catchment characteristics and lake hydrologic variability: a) PC1 & PC2; b) PC1 & PC3; c) PC2 & PC3. The dominant loadings of each PC strongly represent different key conceptualisations of the landscape hydrologic function: PC1 = cumulative flow (wetlands); PC2 = groundwater (surficial geology texture); PC3 = aspen/storage (fine hummocky). See Table 3-2 for PC loadings and PCA summary; see Table 3-1 for variable abbreviations key.

HRA classifications are clearly distinguished across PC1 and PC2 (Figure 3-6a). CO lakes are predominantly situated in the lower right quadrant, and are associated with low landscape position, high sill height and coarse textured lake surficial geology. Fine classes are dispersed across upper quadrants, whereby lakes lower on the axis of PC1 represent greater cumulative flow associated with higher wetland cover. Relative elevation indicates another environmental gradient between fine textured lakes along PC2; FH lakes located highest along the axis of PC2, indicative of higher aspen cover, higher storage, and higher landscape positions. PC3 further demonstrates the separation between FH and FP, with higher aspen cover and subsequently higher hillslope storage relative to lake area in FH classes.

Examination of WLD loadings with catchment and lake variables indicates higher WLD with higher cumulative flow within fine textured classes (Figure 3-6a). Greater aspen cover and hillslope storage is also associated with lower WLD within fine textured classes (Figure 3-6c). WLD is loaded perpendicular to groundwater indicators (Figure 3-6a and 3-6c), however Figure 3-5 indicates lower WLD at lower relative elevations, which is also associated with higher sills and larger groundwater capture areas and ratios in CO. PC4 further unpicks aspen-wetland interactions within and between fine textured classes. Lakes 118, 121 and 122 exhibit low WLD_R values and correspondingly low A_W:A_L ratios (2.8, 5.5 and 1.5 respectively), whereas lake 112 exhibits high WLD_R and high A_W:A_L (55.3), indicative of higher variability with greater cumulative flow from wetlands. High WLD_R in lake 39 is potentially associated with beaver activity in addition to high A_W:A_L (18.4). High WLD_R is attributed to lack of lake outflow in lake 59, in addition to high sill heights for lakes 19 and 15.

4.5. Chemistry Analysis: groundwater dominance

Enrichment of lake water isotopes is used to test groundwater connectivity relative to HRA classifications (Figure 3-7). CO displays moderate mean lc-excess and very low lc-excess standard

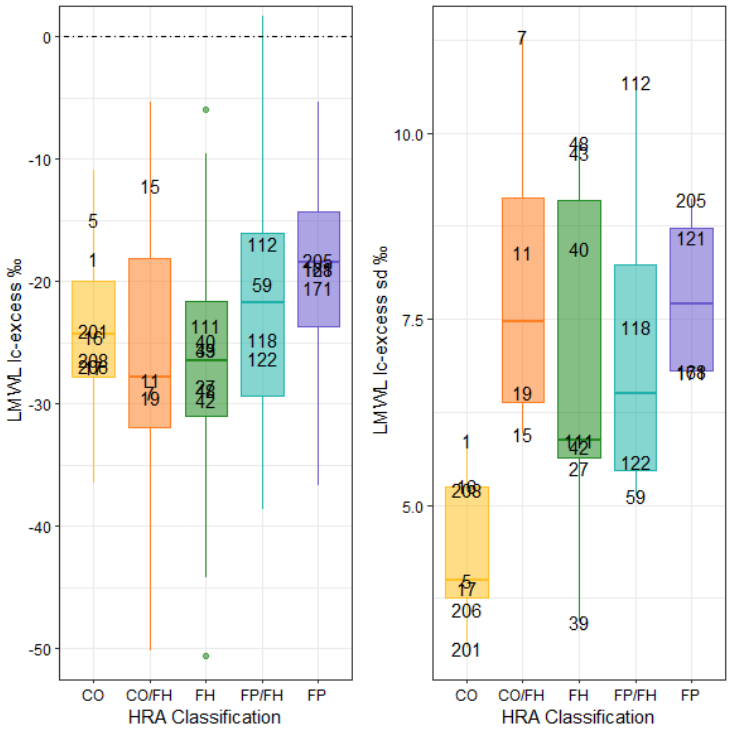


Figure 3-7. Comparison of line-conditioned excess (lc-excess) statistics (a. means; b. standard deviation) relative to the local meteoric waterline (LMWL) of stable isotopes δ^2H and $\delta^{18}O$ (‰). Lc-excess calculated as $\delta^2H - a \cdot \delta^{18}O - b$; where a is the slope and b is the intercept of the dual isotope plot (Hokanson et al., 2019). a) Lake annual means calculated from observations within summer months (June 1st-August 31st) from 1998-2018; text labels indicate the mean of annual means for each lake. Greater negative lc-excess values denote isotopic enrichment of heavier species relative to the LMWL. b) standard deviation in lake annual means; text labels indicate the standard deviation of each individual lake. This study produces LEL slopes between 4.07 and 4.91, in line with the slope of LEL in Alberta of approximately 4.1 (SFM_NCE/TROLS Network , Gibson et al., 2005) and the average value of 4.12 for the URSA in comparison to a LMWL slope of 7.25 (Hokanson et al., 2019).

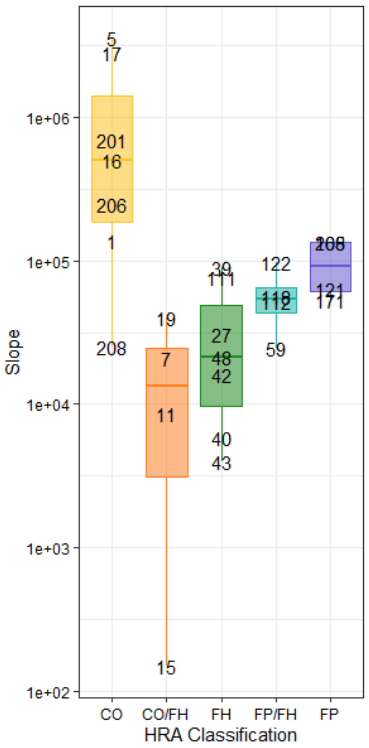


Figure 3-8. Slope coefficient of the linear model derived between lake Ca^{2+} + Mg^{+} concentration (mg L^{-1}) and lake Ca^{2+} + Mg^{+} mass (mg) for each lake. Higher slopes signify greater increase in mass with concentration, indicative of greater inputs of ions with greater concentration. Low slopes signify concentration increases with negligible change in mass due to evapoconcentration, indicative of greater evaporative fluxes relative to groundwater inputs.

deviation, indicating a stable supply of groundwater from an enriched source, reflecting regional groundwater dominance. CO/FH and FH represent the greatest mean enrichment and high lc-excess standard deviation, denoting lower groundwater inputs relative to evaporative losses, and greater sensitivity to atmospheric exchange. This variability reflects resetting of isotopic composition following the lake drying out and then receiving precipitation. FP lakes are the least enriched HRA classification, but also exhibit high lc-excess standard deviation, indicating high but inconsistent groundwater inputs,

reflecting supply of isotopically depleted peatland water inputs that mirrors weather cycles. FP/FH lakes are distributed between FH and FP in both mean and standard deviation of lc-excess, suggesting dominance of peatland water inputs greater than FH and lower than FP.

The slope coefficient of linear relationships between ion concentration and ion mass indicates whether increase in concentration is due to increased supply from groundwater during wet conditions (high slope), or evapoconcentration during dry conditions (low slope). CO lakes incur the highest Ca^{2+}+Mg^{+} concentration-mass slopes, signifying the greatest groundwater inputs, sourced from regional groundwater as inferred from low lc-excess standard deviation. Lake 208 exhibits the lowest slope within the CO class; however, this lake receives localized flow from an adjacent hillslope. An environmental gradient of slope coefficients is present in fine textured HRA classes, indicating decreasing groundwater inputs from peatland sources from FP through to CO/FH. Lake 39 represents an outlier in the FH class with low lc-excess standard deviation and a higher Ca^{2+}+Mg^{+} concentration-mass slopes indicative of consistent water supply, however this can be attributed to beaver activity within the catchment.

The combined interpretation of groundwater source, and the magnitude and stability of groundwater supply inferred from isotopic and ion mass-concentration relationships reinforces conceptual understanding of PCA axes in Figure 3-6, whereby CO lakes are affiliated with regional groundwater indicators, and CO/FH, FH, FP/FH and FP classes represent a variability associated with cumulative flow and storage indicators (see additional supporting graphs in the supplementary material).

5. Discussion

5.1. Conceptualising water movement: On the need for a Climate-Lake-Landscape Framework

This research clearly demonstrates that no single dominant control on lake hydrology exists within the BP. BP shallow lakes exhibit high magnitude lake level fluctuations (up to 0.8m; Chapter 2) comparable with shallow lake fluctuations in PPR (Van Der Kamp, Keir, and Evans, 2008), Shield (Oswald, Richardson, and Branfireun, 2011), and Boreal systems (Karlsen *et al.,* 2019). Spatial heterogeneities in glacial deposits within the BP are understood to produce distinct hydrologic responses to climate in seasonal and interannual studies. While varying temporal patterns are apparent, we show no substantial differences in magnitude response of lake water levels between lake types, as this 20-year study is sufficiently long to capture spatial differences in memory and lag responses to reflect the climate signal in all lakes. Subsequently, coarse-textured outwash lakes, which receive consistent and larger-scale groundwater flow, are not significantly less variable over decadal timescales as hypothesised. We identify spatially variable hydrologic pathways that enable lakes to persist within the sub-humid BP, whereby these processes cause all lake types to fluctuate, but on contrasting timescales. Such mechanisms control the source, magnitude, and continuity of groundwater supply, and modify surface water connectivity; they range in scales from local to regional and internal to external controls, and may only be apparent under certain climatic conditions or across longer study periods and across a range of lake types. This knowledge is key to predict impacts on lake water level variability in the BP landscape under climatic extremes, future climates or landscape modifications and identify landscape configurations of higher sensitivity or at higher risk of shift into new ecosystem states. More crucially, this research suggests management approaches should shift from ascribing buffers based upon lake permanence, to targeted protection of the mechanisms identified that enable different lake types to persist.

Hydrologic Response Areas (HRAs; Devito *et al.,* 2017) delineated by surficial geology provide an exceptional classification tool for the physical characteristics of boreal catchments (Chapter 2). Subsequently, they condense complex processes and spatial heterogeneity into meaningful and process orientated classifications (Carey *et al.,* 2010) and produce distinct hydrological patterns and homogenous isotopic and chemical compositions (Hokanson *et al.,* 2019). Our hydrochemistry results maintain this conceptual understanding of water movement within the BP. The dominant sources of flow vary spatially with surficial geology texture (Plach *et al.,* 2016): coarse outwash lakes receive larger-scale groundwater flow due to high hydraulic conductivity of quartz sand mineral deposits (Smerdon *et al.,* 2005); contrastingly, groundwater inputs are constrained to near-surface flow from organic soils within the fine-textured deposits of clay plains and hummocky moraines, where soil hydraulic conductivity is orders of magnitude lower (Hokanson *et al.,* 2019). HRAs therefore provide the basis for linking lake level variability and catchment characteristics within this study, and offer insight into the dominant controls and sources of flow, akin to the concept of hydrologic landscape frameworks for hypothesising water movement within catchments (Winter, 2001; Devito *et al.,* 2017).

Despite clear existing understanding of mechanisms controlling the scales and sources of flow, lake water level response to the long term climate signal is complex due to interactions with spatially diverse landcover (Devito *et al.,* 2012), lake geometry, landscape position, and geology, which further modifies hydrological connectivity, lags and memory of BP shallow lakes. Landscape position is an important control on magnitude of groundwater supply to coarse-textured lakes and therefore the ability to moderate climate. Open water proportion is also an important control on evaporative loss which may diminish the climatic buffering capacity of larger scale groundwater inputs. Sill height and wetland proportion controls on surface connectivity act to dampen or amplify the climate signal respectively in fine textured lakes. Such characteristics and their associated storage produce high spatial variability in fill-spill thresholds for both groundwater and surface water. We also show high unpredictability of lake fluctuations at surficial geology transition zones. These zones exhibit the

greatest spatial variability within the study and cannot be readily explained by HRA understanding. Therefore, consideration of further climate-landscape interactions in this study extends these existing frameworks by providing understanding of mechanisms controlling lake variability.

Lake hydrochemistry is controlled by interactions with geology, soil type, land cover, vegetation type and topography (Stendera and Johnson, 2006; Güler *et al.,* 2002), while lake-groundwater connectivity operates at a range of scales depending on surficial geology and landscape position (Plach *et al.,* 2016), and catchment scales (Andersson and Nyberg, 2009; Wolock, Fan, and Lawrence, 1997). Therefore, dominant sources of flow and connectivity can be inferred by combining major ion chemistry and stable isotope enrichment with hydrologic metrics (Floriancic *et al.,* 2019). We show high spatiotemporal variability in lake hydrochemistry, indicating that the best geodata-derived predictors of hydrology and hydrochemistry vary markedly between catchments, comparable with European boreal headwaters studies with deep quaternary deposits (Floriancic *et al.,* 2019; Karlsen *et al.,* 2016; Kuentz *et al.,* 2017; Temnerud and Bishop, 2005). Texture of surficial geology provides the overriding control on the source and scale of groundwater flow. Within coarse-textured systems (CO), the high slope coefficient of the linear relation between major ion mass and concentration signifies increased supply from larger-scale mineral ground water sources during wetter periods. In contrast, fine-textured systems display low major ion mass-concentration slope coefficients, signifying low variability in lake major ion total mass. Instead, concentrations increase during dryer conditions, indicative of evapoconcentration processes. However, an environmental gradient in major ion mass-concentration slope coefficients exposes spatial differences in surface connectivity. There is negligible change in mass of major ions where nutrients are trapped and recycled, due to the absence of surface runoff and lake outflows to flush lakes (Van Der Kamp and Hayashi, 2009; LaBaugh *et al.,* 1987; Hayashi, Van Der Kamp and Rosenberry, 2016). Therefore, evaporative concentration is an important process in isolated systems (FH and CO/FH; Cobbaert, Wong, and Bayley, 2014), and concentrations reflect the evaporative demand of the climate signal. Lakes that intermittently generate outflow present variable

flushing capabilities throughout the landscape as a function of fill-spill thresholds associated with lake sill height and bathymetry and the runoff generation capabilities of the catchment. Peatland-dominated systems (FP and FP/FH) generate greater shallow groundwater supply from organic soils, which modifies mass of major ions with climatic conditions, by flushing lakes, altering nutrient supply (Leibowitz, Mushet, and Newton, 2016) and temporarily reducing the effects of evapoconcentration. However, this groundwater supply is semi-continuous, therefore evapoconcentration represents the dominant mechanism during drier periods of reduced surface connectivity. Surface connectivity controls are of less importance for hydrochemistry within the groundwater dominated CO. Instead, landscape position is a good predictor of magnitude of groundwater inputs within CO for inferring position within flow systems, where higher solute concentrations in downgradient lakes arise from greater contributions from deeper groundwater, as opposed to shallow, freshly recharged, less mineralised groundwater, or an accumulative effect of flow from lake to lake (Winter, 2001; Kratz *et al.,* 1997; Webster *et al.,* 1996; Smerdon *et al.,* 2005).

Analysis of isotopic enrichment reiterates hydrochemistry insights into controls on the relative importance of groundwater and evaporation. The stable isotopes δ^2H and $\delta^{18}O$ are sensitive indicators of lake water balances, whereby evaporation enriches surface waters with heavy isotope species, enabling distinction of evaporative versus outflow losses due to kinetic fractionation (Gibson, Prepas, and Mceachern, 2002; Gibson *et al.,* 2005). The hydrologically isolated fine textured basins of FH and CO/FH produce the most enriched δ^2H-$\delta^{18}O$ local evaporation lines (LEL) and most negative line-conditioned excess (lc-excess), associated with greater relative evaporation losses. Despite receiving the greatest groundwater inputs, coarse-textured lakes are not the least enriched class as hypothesised; however, they do exhibit the least variation in isotopic composition, indicating a stable and consistent source of groundwater that has been recharged by enriched sources. High lc-excess variability within fine-textured lakes signifies higher sensitivity to atmospheric exchange. Lower volume lakes will inherently exhibit greater sensitivity, however, this is amplified in lakes with lower

permanence. Lakes that completely dry out reset isotopic compositions from highly enriched prior drying out, to highly depleted upon precipitation inputs. Isotopic analysis therefore supports interpretation of the effects of lake permanence on the magnitude of lake level fluctuations.

5.2. Climate-landscape interactions: processes governing lake water level fluctuations

The Boreal Plains ecoregion is exposed to multi-decadal weather cycles that cause a long-term moisture deficit with infrequent periods of excess moisture (P-ET). Decadal weather cycles and delicate balances between deficit and surplus conditions are common throughout Boreal ecosystems globally and glaciated, low relief dominant interior systems. The URSA experiences 4-year cycles between mesic (PET:P ratio of 0.9-1.2) and dry (PET:P ratio >1.2) conditions, with wet years (PET:P ratio < 0.9) returning approximately every 20 years (defined in Buttle, 2006; Devito, Mendoza, and Qualizza, 2012; Hokanson *et al.,* 2019). The hydrogeoclimatic setting of the Boreal Plains of Canada presents a significant opportunity to unravel the interaction of decadal climate cycles with their landscapes and gain mechanistic understanding of hydrological processes due to diverse geology and catchment configurations within one climate signal.

Groundwater dominance, combined with stable supply of ions from mineral soil weathering processes (Whitfield *et al.,* 2010) results in high within group similarity hydrochemically, indicating insensitivity to local characteristics as coarse-textured systems (CO) converge on a regional signal (Gibson, Prepas, and Mceachern, 2002). Unsurprisingly, lake water level dynamics within CO therefore mimic the controls and patterns of groundwater levels (Hokanson, Mendoza, and Devito, 2019). However, we show that the widely recognised buffering capacity of groundwater dominant CO systems to climate is less evident in long-term observations. Instead, we theorise that the length of this study is sufficient to capture lagged responses to the climate signal caused by groundwater moderating short term lake water level variability in coarse-textured systems. Although groundwater dominance results in low

spatial variability within CO, isotopic enrichment clearly increases with decreasing groundwater connectivity at higher landscape positions. In turn, we observe dampened responses to the climate signal in some lakes at lower elevations, attributed to the high hydrological conductivity of the quartz sand mineral deposits producing well connected, larger scale flowpaths (Smerdon, Devito, and Mendoza, 2005; Plach *et al.,* 2016). However, larger relative evaporative losses may diminish the buffering capacity of groundwater during dry periods where open water represents larger proportions of the catchment. This is in keeping with the concept of BP lakes as 'evaporation windows' on the landscape, where lake evaporation represents the largest hydrologic flux and influences the magnitude of lake-groundwater exchange (Smerdon, Mendoza, and Devito, 2007), and where isotopic signatures depend on landscape position and lake size (Smerdon, Devito, and Mendoza, 2005).

While coarse-textured systems are characterised by high levels of mineral soil groundwater inputs, fine-textured groundwater flow originates from organic soils and represents a semi-continuous and smaller proportion of lake water balances (Olefeldt, Devito, and Turetsky 2013; Plach *et al.,* 2016). Instead, lake fluxes are dominated by precipitation and evapotranspiration (Ferone and Devito, 2004) and groundwater flows are understood to be characterised by intermediate and localised flows for FP and FH respectively (Devito *et al.,* 2012; Hokanson *et al.,* 2019; Plach *et al.,* 2016). An environmental gradient between fine-textured lakes is apparent from hydrochemistry observations, which reflects the transition from isolated to intermittently highly connected systems (CO/FH; FH; FP/FH; FP respectively). This spectrum of catchment types represents variable importance of local morphometry (Gibson, Prepas, and Mceachern, 2002), and switch of dominant controls on lake hydrologic function. These controls subsequently interact with the differences in scales of organic groundwater connectivity that have been widely observed between hummocky moraines and clay plains (Plach *et al.,* 2016).

The presence of an active outflow, or the height of lake outflow sills, provide straightforward mechanisms for explaining lake level variability, whereby closed basins or higher sill heights experience greater long term lake variability due to greater memory of climatic conditions (Van Der Kamp, Keir, and Evans, 2008), whereas active outflows during wet periods will suppress variability. This describes storage driven fill-spill hydrology observed within the Prairies (Leibowitz *et al.,* 2016; Shaw *et al.,* 2012; Spence, 2007; Spence, 2010) and Boreal Shield (Oswald, Richardson, and Branfireun, 2011). The presence of an active lake outflow is also an important control on lake water fluctuations within coarse-textured systems in Minnesota, where flow-through lakes have been shown to buffer climate and minimise lake variability (Jones *et al.,* 2016). Interconnection between lakes also produces low variability "gate keeper" lakes (Gharari and Razavi, 2018; Phillips, Spence, and Pomeroy, 2011; Shook and Pomeroy, 2011; Spence, 2007), whereby flow is attenuated to satisfy storage in higher positioned lakes (Spence, 2010). This disrupts surface connectivity to lower lakes during dry periods, but also extends the effective catchment of the lower positioned lake during wet periods, resulting in higher lake level variability. Consequently, lake storage controls can govern basin scale connectivity (Phillips, Spence, and Pomeroy, 2011).

Lake size and bathymetry are important controls on evaporative lake level decline; smaller lakes are more susceptible to evapotranspiration losses through the moist margin, whereby water flows laterally from the lake to replenish groundwater uptake by vegetation (Hayashi *et al.,* 2016). While this process acts to enhance lake level decline in smaller lakes, this mechanism may explain lower lake variability in characteristically small FH lakes, which were observed to completely dry out within this study, providing a low threshold restriction on lake variability; whereas deeper CO and FP lake levels have greater capacity to potentially decline further.

Peatlands add a component of substantial complexity to this surface connectivity, whereby cumulative flow significantly modifies lags and memory within these systems. We identify latent landscape

controls in wet periods and patent landscape controls in dry periods results in temporally variable controls (Lyon *et al.,* 2012). This may explain why studies have observed opposing effects of peatland:lake ratios on lake water level variability, depending on the distribution of climatic conditions observed. During wet periods, wetland-dominated catchments produce higher mean and maximum runoff ratios than upland dominated catchments (Gibson, Prepas, and Mceachern, 2002), causing greater lake levels increases. Under drought conditions, forests dry parts of the landscape while wetlands maintain wetness (Devito *et al.,* 2012; Lyon *et al.,* 2012; Thompson *et al.,* 2017), reducing lake decline in wetland dominated catchments (Devito *et al.,* 2005; Petrone *et al.,* 2007). We show higher magnitude long-term variability in lake catchments with higher peatland proportions, whereby the ability of peatlands to maintain lake levels diminishes with prolonged water deficit conditions as a function of their low storage capacity and variable source area effects. Low storage capacity enables peatlands to typically function as water sources annually and enables quicker water table recovery following dry periods due to low memory, but also restricts the longevity of water redistribution to lakes. Lukenbach *et al.,* (2017) reports high stability in peatland margin water tables during a wet period but reports rapid drying once the water table fell below the peat surface, in keeping with findings in this study. Changes to surface connectivity throughout climate cycles create a variable source area (Lyon *et al.,* 2012) which amplifies the storage effect of peatlands on the long-term hydrological response of lake catchments. A further possible explanation for greater variation in peatland dominated catchments is that the hydraulic properties of peatland soil profiles occur in bimodal stable ecosystem states that either favour vegetation productivity, or the conservation of water (Dixon *et al.,* 2017). Catchments with hydrologically isolated peatlands may have evolved into conservative systems with negative feedback mechanisms that produce lower evaporation rates due to lower productivity (Waddington *et al.,* 2015) as a product of consistently lower groundwater contributions, and consequently are better adapted under drying cycles. Meanwhile, catchments with highly connected, expansive peatlands are likely to develop ecologically into productive ecosystems,

which observe greater hydrological responses under drying conditions due to historical dependence on shallow groundwater inputs to maintain water tables (Van Der Kamp and Hayashi, 2009). Inclusion of these variables enables study of environmental gradients not previously explored, and demonstrates that changes in hydrochemistry with flow derive from geologic legacy (Bertassello *et al.,* 2019) and a structured mosaic of landscape characteristics, rather than a randomly arranged landscape (Temnerud *et al.,* 2007).

5.3. Insights gained from combined long-term hydrochemistry and implications for management

Increasing susceptibility of BP landscapes to natural resource extraction and non-stationary climatic forcing has placed emphasis on identifying resilient landscapes within research and management strategies. There is common misconception that hydrological variability, or low lake permanence equates with vulnerability, or that stable water levels indicate high resilience. This is further perpetuated by short term studies which capture high variability in frequently fluctuating lakes, while comparable magnitude fluctuations over longer timescales in other systems are overlooked. Here we demonstrate with multi-decadal observations that the magnitude of lake water level fluctuations are less diverse spatially than previously thought, and that all lake types may have the potential to dry out or generate flow. This indicates that management strategies based on permanence or variability may be flawed and instead approaches should focus on conserving landscape mechanisms that enable lakes to persist.

Shallow lakes have evolved to persist for ~6000 years within the Boreal Plains (Schindler, 2009; Zoltai and Vitt, 1990) despite a sub-humid climate. Therefore, mechanisms for resilience must exist within all lake types in order to manage the delicate balance between precipitation and evaporation. We identify spatially variable mechanisms for the persistence of shallow lakes within the BP. Coarse-

textured systems place high dependence upon groundwater contributions to maintain water levels (groundwater dependent ecosystems; Smerdon *et al.,* 2012), with decreasing control at higher landscape positions. Fill-spill mechanisms governed by lake sill height and lake bathymetry control lake levels in isolated and poorly connected fine-textured systems; while greater but intermittent connectivity associated with shallow storage and low memory in peatlands drives lake fluctuations in peatland-dominated fine-textured systems.

These dominant controls on lake-landscape connectivity vary from regional scale to local scale characteristics and processes, and associated differences in memory and lag convey the climate signal across different temporal scales. Consequently, larger scale studies which rely on low resolution, regional scale geodata may result in weak relationships between hydrochemistry and catchment properties where local scale controls override regional controls (Floriancic *et al.,* 2019). Moreover, interactions between catchment descriptors and hydrologic response varies temporally with wetness (Lyon *et al.,* 2012), therefore snapshot hydrochemistry studies lack insight into temporal changes in flowpaths. We also highlight the potential for erroneous conclusions when inferring hydrologic process from major ion concentrations without combining lake water levels, particularly in landscapes where groundwater and evaporation both represent important components of lake water balances. Diverse BP lake types observe comparable major ion concentrations despite orders of magnitude difference in groundwater supply due to evapoconcentration of trapped nutrients in less connected systems. Therefore, analysis of hydrochemistry relies on interpretation of processes at a wide range of scales and wetness conditions in combination with lake water level responses.

Resilience is a measure of the ability of a catchment to return to normal functioning following a disturbance, whereby enhanced threshold behaviour indicates low resilience (Carey *et al.,* 2010). Based upon this definition and insights gained into mechanisms driving lake level fluctuations, it could be argued that isolated systems exhibit autogenic conservative mechanisms that enable them to

recover from perturbations, making them potentially highly resilient and requiring simple management if lake geometry and substrate remain unmodified. Internal negative feedback mechanisms also enable peatlands to maintain lake levels (Waddington *et al.,* 2015) and enhance landscape resilience, whereby low storage and memory limits the longevity of this resistance to drought but also enables rapid recovery upon rewetting. However this assumes that external influences such as wildfire, pests, and anthropogenic disturbance do not outweigh internal regulatory mechanisms (Thompson, Mendoza, and Devito, 2017). Deeper and more permanently ponded water (Gleason and Rooney, 2017) may result in higher resistance of coarse-textured lakes to disturbance, however, these systems would be highly sensitive to modifications within the wider landscape that impede larger scale groundwater flow. Moreover, groundwater moderation of precipitation deficits may conceal responses to disturbance (Smerdon, Devito, and Mendoza, 2005), and ecological implications may only become apparent during prolonged water deficit conditions.

This research clearly demonstrates that a crucial shift in landscape management is needed away from buffering strategies founded on lake permanence and fluctuation magnitudes, and towards focus on recognising and conserving the mechanisms that maintain lake levels and facilitate their recovery following dry periods. Given that these mechanisms operate on a range of spatial and temporal scales, long-term studies are necessary to further understand spatial differences in climate memory and lagged responses in shallow lakes.

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7. Supplementary Material

Supplementary Table 3-1. *Variable definitions, purpose for BP hydrological conceptualisation and indication of use in further analysis following testing for redundancy in conceptual model (Figure 3-4).*

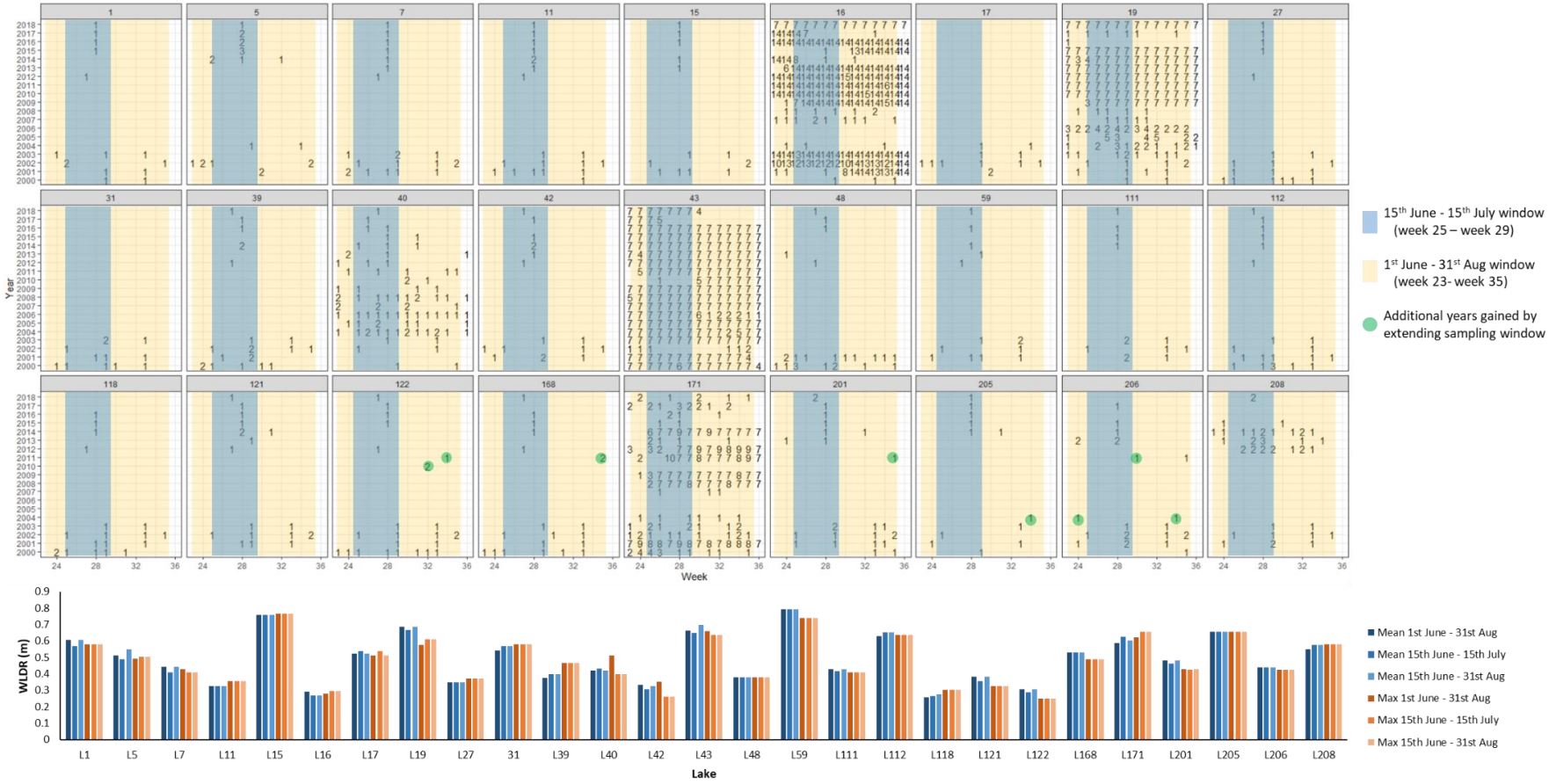
Variable of interest		Description	Purpose	Used in PCA (clear reason for all crosses)
Definitions	WL	WL = Water elevation above mean sea level	Time series of lake water levels above mean sea level	✖ - timeseries
	$Depth_W$	$Depth_W$= WL – Lake bottom elevation above sea level (m)	Time series of the lake depth of water	✖ - timeseries
	Water Level Deviation (WLD)	WLD = WL(t) – WL(2012) (m)	Time series of lake water level deviation from the water level in 2012 (when the region is generating runoff).	✖ - timeseries
	D_{2012}	Observed depth of water at the centre of the lake in July 2012	Indicate potential storage, and capacity for flow generation and retention within the catchment during a wet period within the study	✖ - timeseries
Lake Water Level Variability (dependent variables)	WLD_R	Range of mean annual lake WLDs (water level deviation) WLD_R = MAX(WLD) – MIN(WLD)	Indicator for long term hydrologic magnitude of variability of lakes	✓
	WLD_{SD}	Standard deviation of mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✖ - significantly correlated with WLD_R
	WLD_{MAD}	Mean absolute deviation in mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✖ - significantly correlated with WLD_R
	WLD_{IDR}	Interdecile range of mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✓
	WLD_{IQR}	Interquartile range of mean annual lake WLDs	Indicator for long term hydrologic magnitude of variability of lakes	✖ - significantly correlated with WLD_R
Lake Characteristics (independent variables)	D_μ	Mean depth of observed water at the centre of the lake across study period (mean of summer means)	Indicate the average active storage, and capacity for flow generation and retention within the catchment	✖ - significantly correlated with all lake characteristics
	A_L	Area of lake (m^2)	Indicator of lake storage Proxy for lake volume (assuming straight sided bathymetry) (in 2012?)	✖ - incorporated within ratio variables
	P_L	Perimeter of lake (m)	Indicator for potential for lateral water loss or gain	✖ - significantly correlated with all lake characteristics
	H_{Sill}	Lake Sill height = SillElevation-LakeBottomElevation For Lake with an outflow, SillElevation = $WL_{Max.}$ Else, for lakes with no outflow, SillElevation calculated from DEM.	Indicator of maximum lake level (control on outflow to define fill and spill dynamics)	✓
	V_L	Estimated maximum potential lake volume (area * sill height, assuming straight sided bathymetry)	Indicate lake potential storage	✖ - significantly correlated with all lake characteristics
	$LakeClass_C$	The dominant (50% surficial geology of the lake perimeter; within a 100m buffer) Binary score of Fine = -1, coarse = 1	Indicator for scales of groundwater interactions	✓
Catchment Characteristics (independent variables) — Topography	Relief	Range in catchment elevation (maximum minus minimum elevation)	Indicator of storage (higher relief has higher upland storage)	✖ - significantly correlated to slope (r=0.45, p=0.02) and relative elevation (r=-0.4, p=0.04)
	Slope	Average slope of catchment (%)	Indicator of storage (higher slope has higher upland soil storage)	✖ - significantly correlated to flat (r=-0.89, p<0.001) and relief (r=0.45, p=0.02)
	Flat	Percentage of catchment with slope<1%	Indicator of landscape connectivity and storage (higher flat area is more connected with less storage)	✖ - A_{FH} indicates catchment slope/flat
Catchment Characteristics (independent variables) — Landscape Position	Rel Elev	Percentile relative elevation within local HRA sub-unit (%)	Potential for flow accumulation	✓
	Boundary Distance	Distance from lake to closest upper boundary of HRA (km)	Potential for upstream flow accumulation and distinguishes lakes that accumulate less flow than expected for a given landscape position due to proximity to HRA boundary	✖ - significantly correlated to relative elevation
	A_{CGW}	Area of hyporheic catchment of lakes within coarse textured surficial geology	Potential for groundwater interactions – larger area also indicates lower landscape position in coarse HRA	✖ - significantly correlated with relative elevation

Supplementary Table 3-1 continued. *Variable definitions, purpose for BP hydrological conceptualisation and indication of use in further analysis following testing for redundancy in conceptual model (Figure 3-4).*

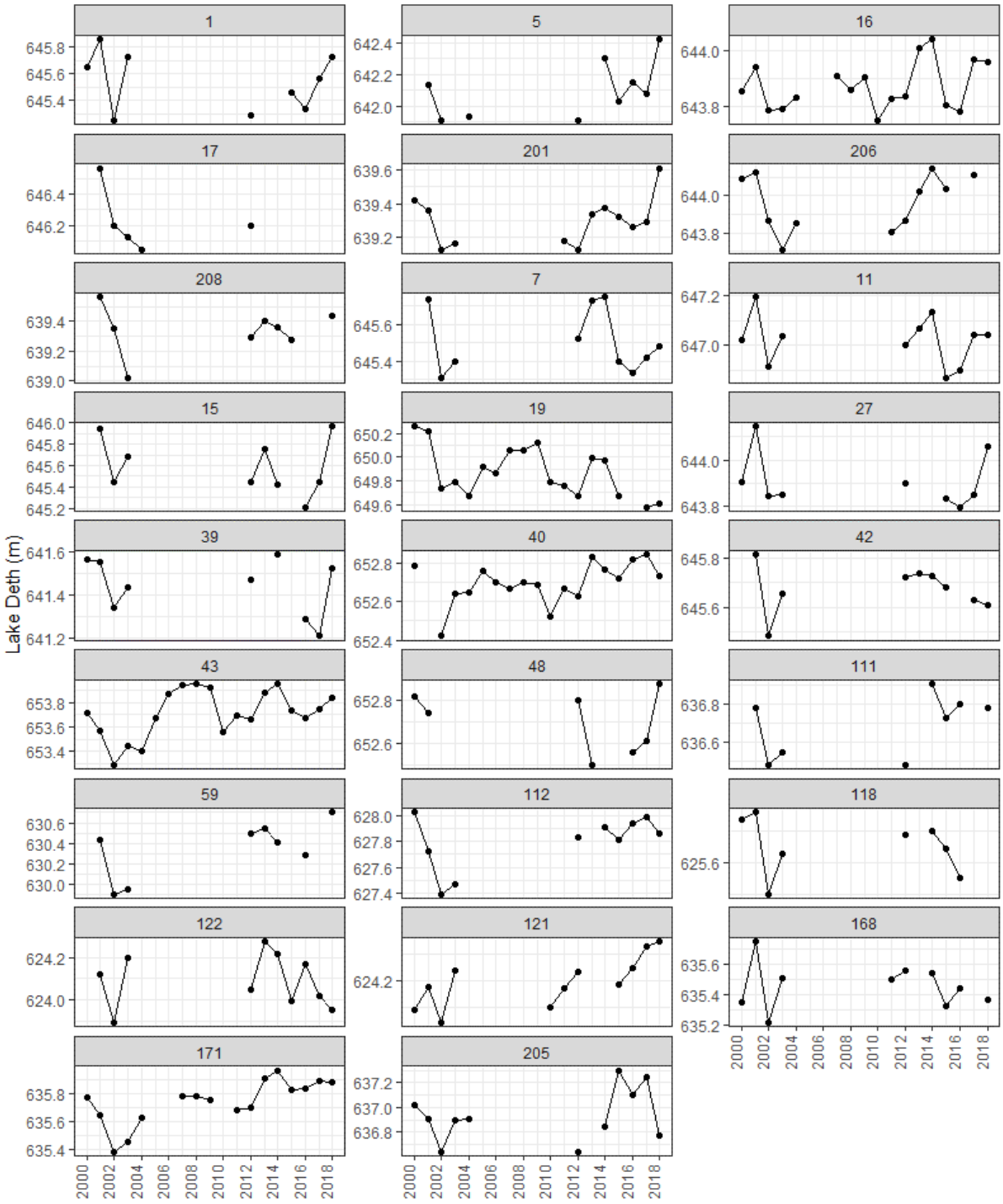
Variable of interest			Description	Purpose	Used in PCA (clear reason for all crosses)
Catchment Characteristics (independent variables)	Surficial Geology Areas and Ratios	A_{CGW}: A_L	Ratio of area of groundwater capture (hyporheic catchment) of coarse textured lakes, relative to lake area	Potential groundwater interactions	✖ - significantly correlated with groundwater indicator $LakeClass_C$
		A_F: A_L	Ratio of area of fine textured surficial geology within topographic catchment to lake area	Indicate surficial geology texture	✖ - does not distinguish between storage (FH/aspen) and flow accumulation (FP/wetlands)
		Fine%	Percentage of fine textured surficial geology within topographic catchment	Indicate surficial geology texture	✖ - ratio to lake better indicator of lake WLD
		A_C: A_L	Ratio of area of coarse textured surficial geology within topographic catchment to lake area	Potential groundwater interactions	✖ - $LakeClass_C$ selected
		Coarse%	Percentage of coarse textured surficial geology within topographic catchment	Potential groundwater interactions	✖ - ratio to lake better indicator of lake WLD
		A_{FH}: A_L	Ratio of area of fine hummocky surficial geology within topographic catchment to lake area	Indicator for storage relative to lake (higher FH associated with higher slope)	✓
		FH%	Percentage of fine hummocky surficial geology within topographic catchment	Indicator for storage relative to lake (higher FH associated with higher slope)	✖ - ratio to lake better indicator of lake WLD
		A_{FP}: A_L	Ratio of area of fine plain surficial geology within topographic catchment to lake area	Indicator for storage relative to lake (higher FP associated with lower slope and higher wetland cover relative to FH)	✖ - ratio to lake better indicator of lake WLD
		FP%	Percentage of fine plain surficial geology within topographic catchment	Indicator for storage relative to lake (higher FP associated with lower slope and higher wetland cover relative to FH)	✖ - A_{FH}: A_L selected
	Land Cover Areas and Ratios	A_{TC}	Area of catchment (m^2)	Area of landscape that the lake potentially accumulates flow from	✖ - significantly correlated to wetland area and connected wetland area (r=0.00, p<0.001)
		A_{TC}: A_L	Ratio of lake area to topographically delineated catchment area (excluding study lake)	Lake storage potential relative to the potential for flow accumulation	✓ - testing counter hypothesis to connected wetland ratios driving lake WL response
		A_W	Area of wetland within the catchment (m^2)	Area of HU with potential for flow generation (lower storage, lower AET)	✖ - significantly correlated to connected wetland area (r=0.99, p<0.001), catchment area (r=0.99, p<0.001)
		A_W: A_L	Ratio of lake to wetland area	Lake storage potential relative to the potential for flow accumulation	✖ - significantly correlated to lake:connectedwetland (r=0.99, p<0.001) and lake:catchment (r=0.97, p<0.001)
		Wetland%	Percentage of wetland within the catchment (m^2)	Percentage of catchment with potential for flow generation (lower storage, lower AET)	✖ - ratio to lake better indicator of lake WLD
		A_{CW}	Area of wetland connected to the lake (m^2)	Area of HU with potential for flow generation that is connected to the lake (lower storage, lower AET)	✖ - ratio to lake better indicator of lake WLD
		A_{CW}: A_L	Ratio of lake area to "connected" wetland area	Storage of lake relative to the potential for flow accumulation from connected, flow generating HU	✓
		CWetland%	Percentage of connected wetland within the catchment (m^2)	Percentage of catchment with potential for flow generation (lower storage, lower AET)	✖ - ratio to lake better indicator of lake WLD
		A_{OW}	Area of open water within the catchment above the study lake	Proxy for water loss within the catchment – lakes as water sinks in the catchment	✖ - ratio to lake better indicator of lake WLD
		A_{OW}: A_L	Ratio of lake area to area of open water above the study lake the catchment	Proxy for water loss within the catchment – lakes as water sinks in the catchment	✓
		OpenWater%	Percentage of open water within the catchment above the study lake	Proxy for water loss within the catchment – lakes as water sinks in the catchment	✖ - ratio to lake better indicator of lake WLD

Supplementary Table 3-1 continued. *Variable definitions, purpose for BP hydrological conceptualisation and indication of use in further analysis following testing for redundancy in conceptual model (Figure 3-4).*

Variable of interest			Description	Purpose	Used in PCA (clear reason for all crosses)
Catchment Characteristics (independent variables)	**Disturbance**	Catchment Fire Area	Area of topographic catchment burnt in 2011 fire	Examine impact of fire within landscape	✖ - t-test of burn vs unburn, and regression of burnt connected wetland area show no significant association with any lake WL metrics.
		Catchment Fire %	Percentage of topographic catchment area burnt in 2011 fire	Examine impact of fire within landscape	✖ - t-test of burn vs unburn, and regression of burnt connected wetland area show no significant association with any lake WL metrics.
		Wetland Fire Area	Area of connected wetland burnt in 2011 fire	Examine impact of fire within connected landscape	✖ - t-test of burn vs unburn, and regression of burnt connected wetland area show no significant associated with any lake WL metrics.
		Wetland Fire %	Percentage of connected wetland area burnt in 2011 fire	Examine impact of fire within connected landscape	✖ - t-test of burn vs unburn, and regression of burnt connected wetland area show no significant associated with any lake WL metrics.
		Harvested Area	Area of topographic catchment area cleared for harvesting 2000-2017	Examine impact of harvesting	✖ - t-test of harvested vs unharvested, and regression of harvested area show no significant association with any lake WL metrics.
		Harvest %	Total percentage of topographic catchment area cleared for harvesting 2000-2017	Examine impact of harvesting	✖ - t-test of harvested vs unharvested, and regression of harvested area show no significant association with any lake WL metrics.



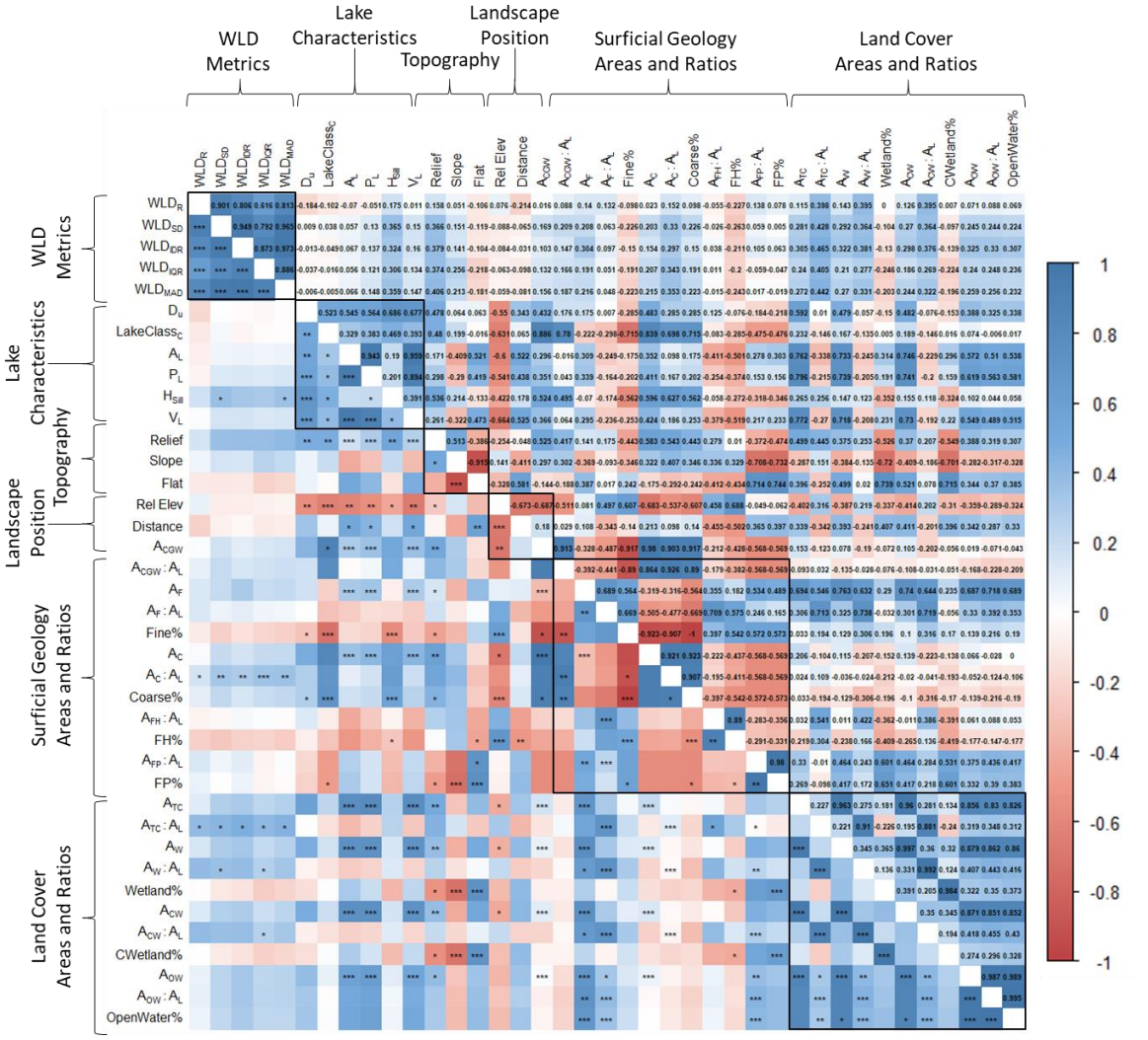
Supplementary Figure 3-1. *(a) Sampling frequency and N values per week per year (weeks 23-36) per lake; (b) sensitivity to different sampling windows for mean WLD_R and maximum WLD_R. Lake 31 is omitted from further analyses. Using a narrower window (blue window) results in lower N (number of years) in 5 lakes (green circles). A wider window enables inclusion of these values, but also causes frequently sampled lakes (e.g., 16, 40, 43, 171) to capture seasonality. A sensitivity analysis of this shows no significant effect of widening the summer sampling window on water level deviation metrics used in this study.*



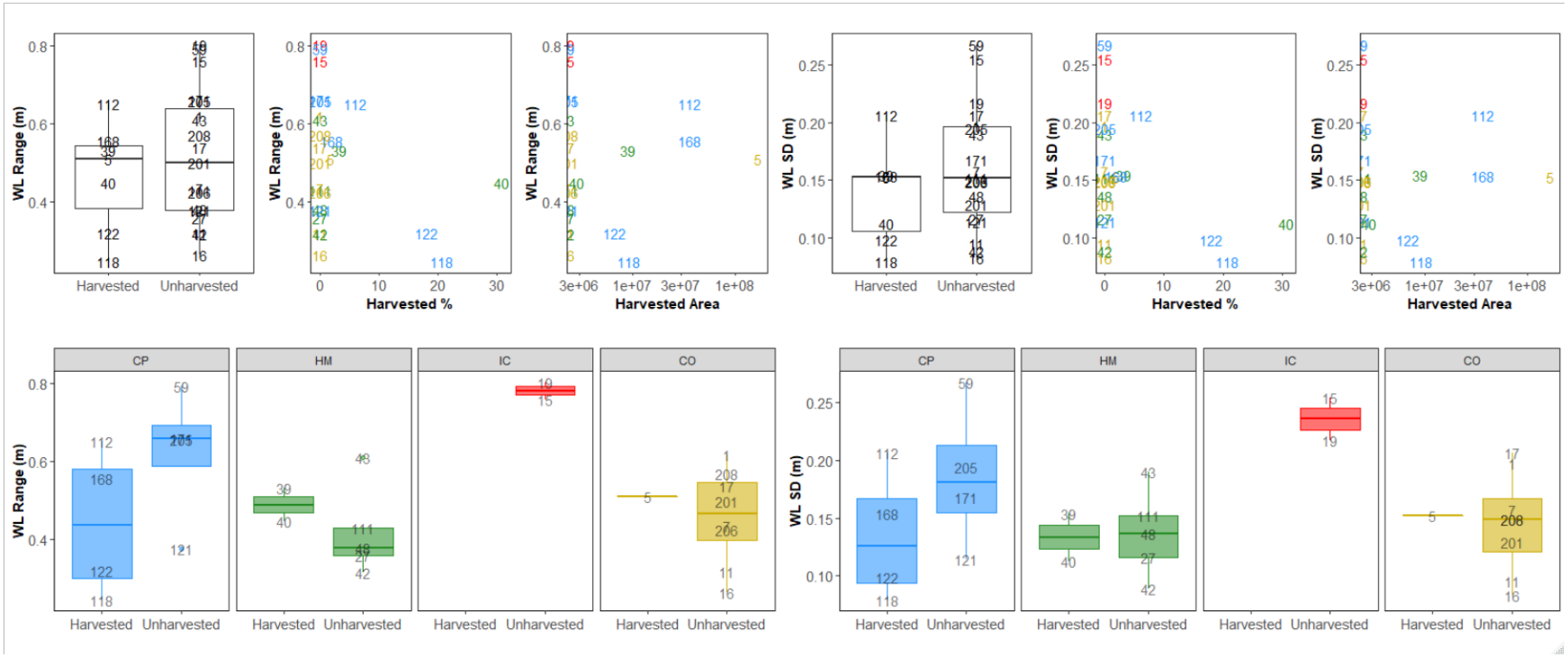
Supplementary Figure 3-2. *Annual WL timeseries of each study lake.*

Supplementary Table 3-2. *Lake catchment characteristics data.*

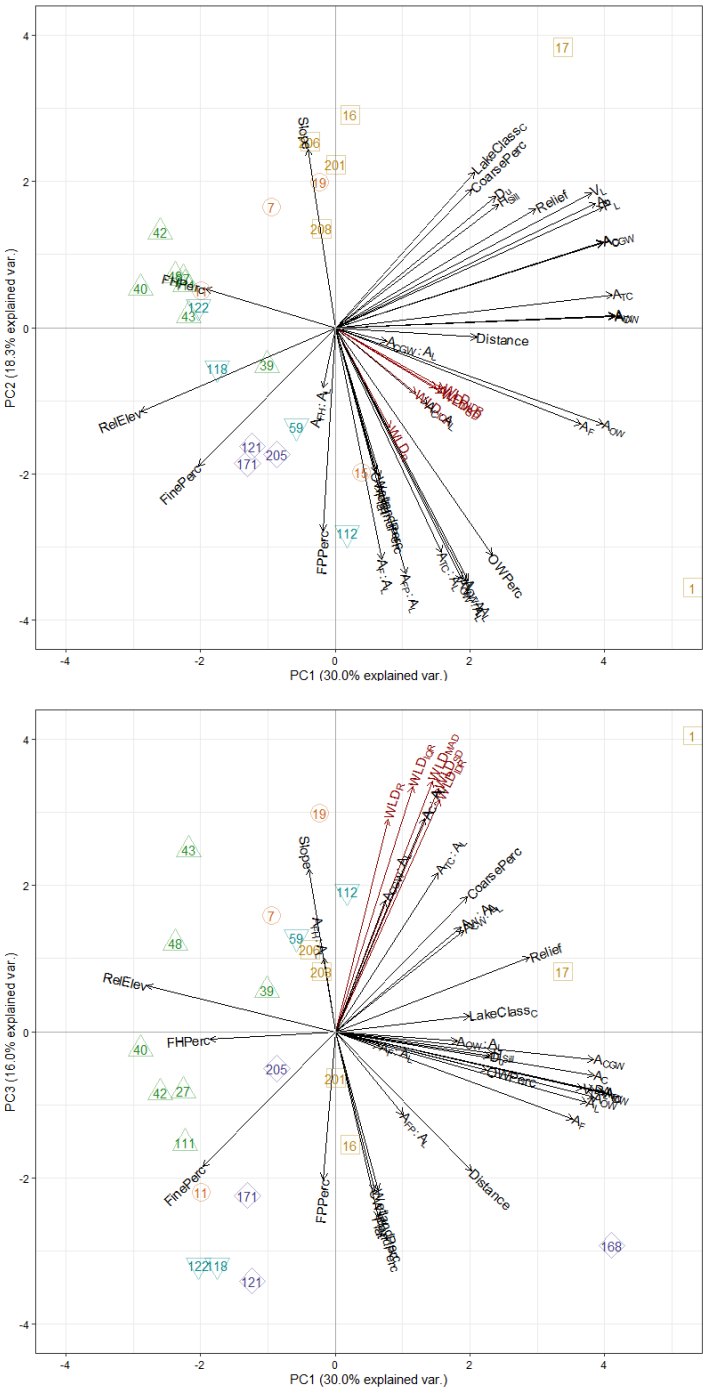
Lake	HRAclass	LakeClassc	WLD_R	WLD_{SD}	WLD_{IDR}	WLD_{IDR}	WLD_{MAD}	D_{μ}	A_{CGW}	P_L	H_{SIII}	V_L	Relief	Slope	Flat	Rel Elev	Distance	A_{CGW}	A_{CGW}/A_{CGW}	A_F	$A_F:A_L$	Fine%	A_C	$A_C:A_L$	Coarse%	A_{FH}	$A_{FH}:A_L$	FH%	A_{FP}	$A_{FP}:A_L$	FP%	A_{TC}	$A_{TC}:A_L$	A_W	$A_W:A_L$	Wetland%	A_{CW}	$A_{CW}:A_L$	CWetland%	A_{OW}	$A_{OW}:A_L$	OpenWater
1	CO	1	0.580	0.222	0.560	0.407	0.202	2.80	112090	3114	3.1	174776	24.56	1.37	52.97	22.48	6.1	17840000	159.16	2068595.2	18.46	13	13843676	123.51	87	2068595	18.455	13	0	0.00	0	15912271	140.960	11000012	98.14	69.6	10810140	96.44	68.4	2015581	17.98	0.13
5	CO	1	0.492	0.176	0.471	0.192	0.140	3.51	3244530	14158	3.8	618245	53.00	2.08	40.12	4.97	9.0	51400000	15.84	23545415	7.26	26	67013873	20.65	74	23545415	7.257	26	0	0.00	0	90559288	26.911	50548602	15.58	57.9	48921022	15.08	56.0	5124060	1.58	0.06
7	CO/FH	1	0.427	0.153	0.358	0.263	0.134	1.99	28671	762	3.4	32035	42.04	4.08	28.90	69.16	3.1	800000	27.90	480586	16.76	76	151764	5.29	24	480586	16.762	76	0	0.00	0	632350	21.055	220645	7.70	36.6	203699	7.11	33.7	0	0.00	0.00
11	CO/FH	1	0.357	0.095	0.240	0.070	0.069	1.63	10680	374	2.5	13350	13.34	1.01	67.14	73.34	2.7	200000	18.73	112640.6	10.55	87	16831.36	1.58	13	112641	10.547	87	0	0.00	0	129472	11.123	70253	6.58	59.1	70253	6.58	59.1	0	0.00	0.00
15	CO/FH	-1	0.766	0.251	0.611	0.361	0.212	0.66	485	93	4.5	247	14.41	2.45	16.28	77.21	2.1	61243	126.27	0	0.00	0	77928	160.68	100	0	0.000	0	0	0.00	0	77928	159.676	29246	60.30	37.8	29246	60.30	37.8	0	0.00	0.00
16	CO	1	0.279	0.077	0.180	0.107	0.066	2.10	465820	3069	2.2	522347	33.43	2.09	42.10	8.41	7.5	6200000	13.31	0	0.00	0	1852940	3.98	100	0	0.000	0	0	0.00	0	1852940	2.978	607116	1.30	43.8	542883	1.17	39.1	2805	0.01	0.00
17	CO	1	0.512	0.196	0.399	0.230	0.162	5.57	1059549	6984	5.9	3135046	41.25	2.62	39.62	25.12	8.1	1900000	1.79	0	0.00	0	4718789	4.45	100	0	0.000	0	0	0.00	0	4718789	3.454	1586712	1.50	43.4	1516224	1.43	41.4	95058	0.09	0.02
19	CO/FH	-1	0.578	0.190	0.507	0.289	0.170	1.95	37390	820	4.7	43720	24.22	3.86	19.15	42.74	8.5	300000	8.02	0	0.00	0	295189	7.90	100	0	0.000	0	0	0.00	0	295189	6.895	54582	1.46	21.2	47406	1.27	18.4	0	0.00	0.00
27	FH	-1	0.371	0.114	0.254	0.153	0.097	1.07	67743	1226	1.4	45164	14.07	1.90	34.02	84.98	3.8	138408	2.04	296830.2	4.38	68	139684.8	2.06	32	296830	4.382	68	0	0.00	0	436515	5.444	181361	2.68	49.2	181361	2.68	49.2	0	0.00	0.00
39	FH	-1	0.465	0.148	0.370	0.222	0.122	2.62	63994	880	2.9	92317	23.05	2.23	30.68	82.19	2.1	0	0.00	2760538	43.14	100	0	0.00	0	2760538	43.137	100	0	0.00	0	2760538	42.137	1179506	18.43	43.7	1161998	18.16	43.1	188039	2.94	0.07
40	FH	-1	0.511	0.118	0.247	0.142	0.090	1.27	7725	341	1.4	5591	9.71	3.42	8.79	94.81	1.1	0	0.00	82182	10.64	100	0	0.00	0	82182	10.638	100	0	0.00	0	82182	9.638	43169	5.59	58.0	43169	5.59	58.0	0	0.00	0.00
42	FH	-1	0.351	0.094	0.174	0.100	0.070	2.49	14037	485	2.6	18364	14.96	2.90	15.92	89.22	2.1	0	0.00	303315	21.61	100	0	0.00	0	303315	21.608	100	0	0.00	0	303315	20.608	93358	6.65	32.3	91317	6.51	31.6	0	0.00	0.00
43	FH	-1	0.660	0.187	0.489	0.280	0.154	1.16	12773	623	1.4	9008	18.03	2.64	17.16	96.42	0.7	0	0.00	274136	21.46	100	0	0.00	0	274136	21.462	100	0	0.00	0	274136	20.462	76108	5.96	29.1	64153	5.02	24.5	0	0.00	0.00
48	FH	-1	0.380	0.146	0.344	0.255	0.138	1.02	49779	1356	1.2	30181	23.77	3.00	17.76	97.00	0.6	0	0.00	798440	16.04	100	0	0.00	0	798440	16.040	100	0	0.00	0	798440	15.040	218351	4.39	29.2	177869	3.57	23.8	12773	0.26	0.02
59	FP/FH	-1	0.741	0.246	0.611	0.296	0.209	1.70	240951	2273	2.2	265046	10.27	1.17	59.97	72.25	3.8	0	0.00	1674675	6.95	100	0	0.00	0	636377	2.641	38	1038298.5	4.31	62	1674675	5.950	934714	3.88	65.2	934714	3.88	65.2	770	0.00	0.00
111	FH	-1	0.411	0.120	0.247	0.100	0.093	2.23	58463	1262	2.4	71023	9.44	1.41	50.19	95.56	6.4	0	0.00	463452	7.93	100	0	0.00	0	463452	7.927	100	0	0.00	0	463452	6.927	166327	2.85	41.1	156541	2.68	38.7	0	0.00	0.00
112	FP/FH	-1	0.638	0.193	0.507	0.168	0.149	2.00	48472	1125	2.3	54567	32.34	1.70	41.80	61.92	4.9	0	0.00	6052994	124.88	100	0	0.00	0	5508225	113.637	91	544769.5	11.24	9	6052994	123.876	2680744	55.31	44.6	2026174	41.80	33.7	140911	2.91	0.02
118	FP/FH	-1	0.301	0.098	0.257	0.101	0.079	1.74	91424	1271	1.9	85152	6.89	1.22	54.99	55.63	7.7	0	0.00	444833	4.87	100	0	0.00	0	244658	2.676	55	200174.9	2.19	45	444833	3.866	268357	2.94	75.9	262421	2.87	74.3	24330	0.27	0.06
121	FP	-1	0.325	0.116	0.279	0.218	0.106	1.69	86767	1064	1.9	81448	3.54	0.88	68.11	38.79	8.7	0	0.00	563027	6.49	100	0	0.00	0	0	0.000	0	563027	6.49	100	563027	5.489	478553	5.52	100.5	478553	5.52	100.5	6174	0.07	0.01
122	FP/FH	-1	0.250	0.079	0.192	0.117	0.069	1.75	91998	1232	1.9	88308	6.72	1.08	61.11	43.70	8.6	0	0.00	373987	4.07	100	0	0.00	0	216913	2.358	58	157074.5	1.71	42	373987	3.065	140745	1.53	49.9	140745	1.53	49.9	0	0.00	0.00
168	FP	-1	0.487	0.132	0.259	0.167	0.105	1.85	135410	1512	2.2	146026	18.76	0.62	77.88	62.33	6.9	0	0.00	18486283	136.52	100	0	0.00	0	0	0.000	0	18486283	136.52	100	18486283	135.521	13328400	98.43	72.6	13186862	97.39	71.9	4060537	29.99	0.22
171	FP	-1	0.621	0.149	0.326	0.190	0.113	1.43	107893	1246	2.0	105896	5.35	0.71	74.09	62.33	6.1	0	0.00	823504	7.63	100	0	0.00	0	0	0.000	0	823504	7.63	100	823504	6.633	632979	5.87	88.5	632979	5.87	88.5	1869	0.02	0.00
201	CO	1	0.430	0.123	0.256	0.158	0.100	3.00	382239	2562	3.3	630617	10.69	1.11	47.40	26.91	2.9	3500000	9.16	0	0.00	0	720999	1.89	100	0	0.000	0	0	0.00	0	720999	0.886	158116	0.41	46.7	158116	0.41	46.7	0	0.00	0.00
205	FP	-1	0.658	0.192	0.489	0.190	0.158	1.62	204979	1668	2.0	200674	14.34	0.83	69.59	81.78	3.5	0	0.00	2672777	13.04	100	0	0.00	0	0	0.000	0	2672777	13.04	100	2672777	12.039	1615773	7.88	65.5	1615773	7.88	65.5	5042	0.03	0.00
206	CO	1	0.424	0.140	0.333	0.207	0.125	3.15	136775	2355	3.3	226068	26.19	3.44	18.20	50.92	1.9	1030000	7.53	752319.7	5.50	41	1082606.3	7.92	59	752320	5.500	41	0	0.00	0	1834926	12.416	624327	4.57	36.8	484819	3.55	28.5	2076	0.02	0.00
208	CO	1	0.580	0.158	0.286	0.119	0.118	1.48	37142	726	1.8	32731	13.03	2.98	32.34	26.91	3.6	11200000	301.55	0	0.00	0	178668	4.81	100	0	0.000	0	0	0.00	0	178668	3.810	106591	2.87	75.3	106591	2.87	75.3	0	0.00	0.00



Supplementary Figure 3-3. *Correlogram of spearman rank tests between all dependent variables and all independent variables devised for study catchments. Collinearity between independent variables highlights any redundancy between variables within conceptual model. Spearman's rho statistic given in upper triangle and by shading, significance is given in the lower triangle ('***' denotes p ≤ 0.001, '**' denotes p ≤ 0.01, '*' denotes p ≤ 0.05).*



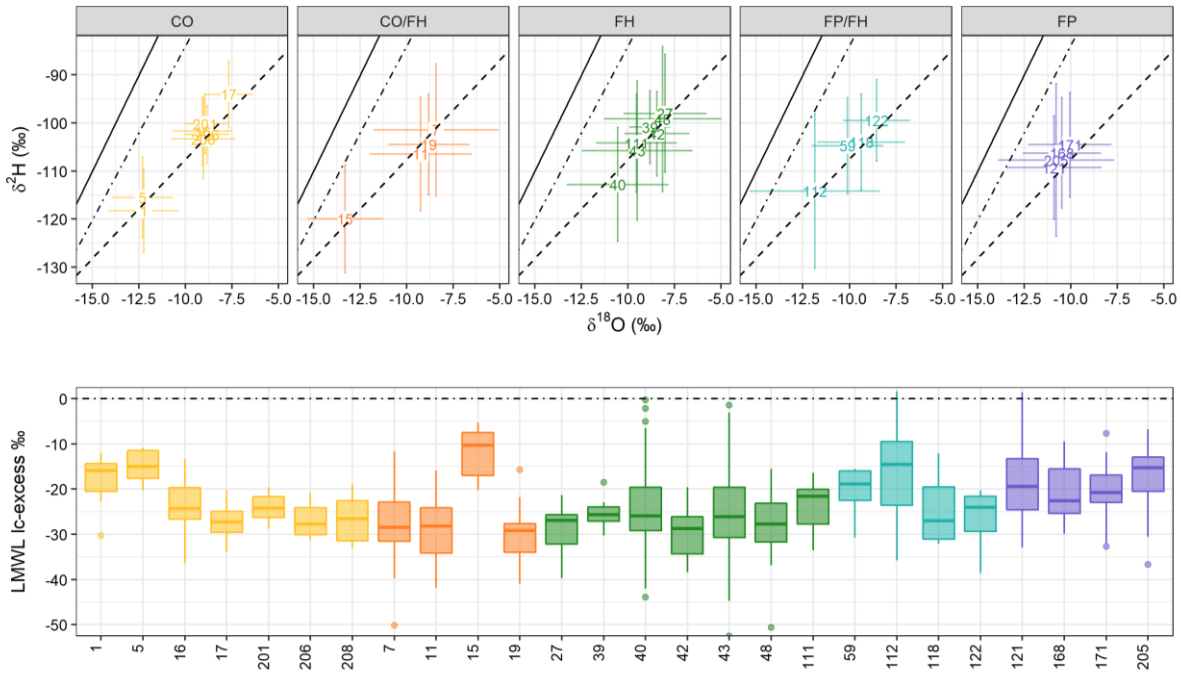
Supplementary Figure 3-5. *Boxplots of harvested versus unharvested catchments, and scatterplots of harvest area. T-tests find no significant differences between harvested area and water level variability. Only CP has significant correlation between harvest % and WL range (r=-0.7657, p=0.02677) and WL sd (r=-0.6675, p=0.0704) but this is likely due to catchment characteristics (122 and 118 similar to 121 and all small, sloping). Harvesting disturbance variables are subsequently omitted from further analysis.*



Supplementary Figure 3-6. *Principal component analysis (PCA) of lake and catchment characteristics (black loadings arrows) and lake WLD metrics (red loadings arrows). Data point shape and colour correspond with HRA classification and are labelled by lake. Multiple PCs are presented due to demonstrate the multidimensional interactions between lake and catchment characteristics and lake hydrologic variability: a) PC1 & PC2; b) PC1 & PC3. Loadings provided in supplementary table 3.*

Supplementary Table 3-3. *Loadings of principal component analysis (PCA) in supplementary figure 6.*

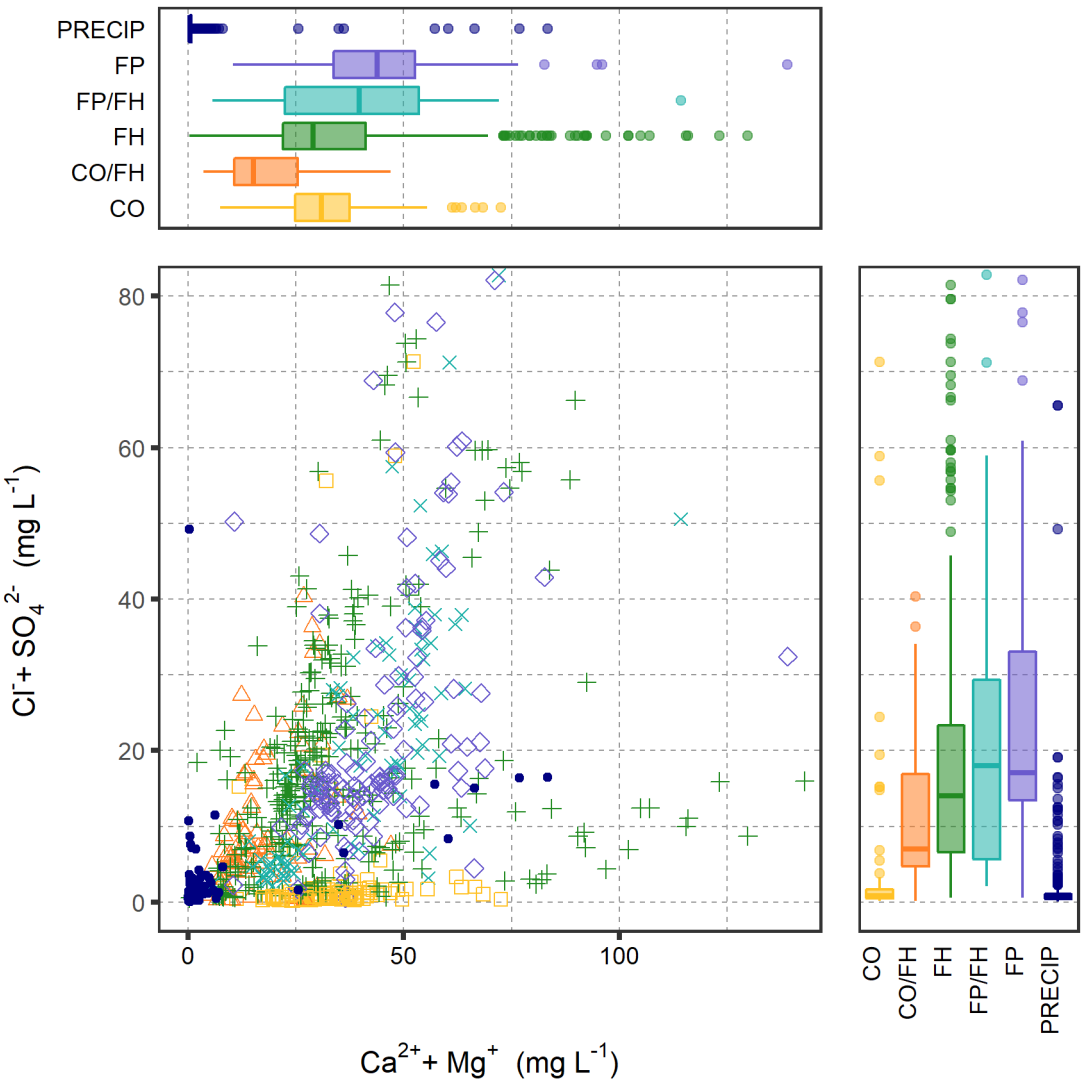
	PC1	PC2	PC3	PC4
WLD_R	0.052	-0.113	0.266	-0.05
WLD_{SD}	0.102	-0.074	0.304	-0.057
WLD_{IDR}	0.104	-0.068	0.29	-0.017
WLD_{IQR}	0.077	-0.074	0.306	-0.068
WLD_{MAD}	0.097	-0.068	0.314	-0.068
D_u	0.153	0.148	-0.031	-0.017
$LakeClass_C$	0.133	0.175	0.019	-0.139
A_L	0.249	0.14	-0.088	0.058
P_L	0.257	0.136	-0.073	0.029
H_{SIII}	0.156	0.139	-0.028	-0.035
V_L	0.245	0.152	-0.071	0.051
Relief	0.192	0.134	0.093	0.178
Slope	-0.026	0.201	0.204	0.152
Flat	0.043	-0.186	-0.232	-0.201
Rel Elev	-0.187	-0.095	0.058	0.255
Distance	0.135	-0.011	-0.174	-0.172
A_{CGW}	0.256	0.096	-0.035	0.02
$A_{CGW}:A_L$	0.049	-0.016	0.164	-0.233
A_F	0.235	-0.109	-0.109	0.174
$A_F:A_L$	0.044	-0.261	-0.018	0.247
Fine%	-0.131	-0.156	-0.169	0.24
A_C	0.256	0.097	-0.055	0.084
$A_C:A_L$	0.088	-0.086	0.266	-0.141
Coarse%	0.131	0.156	0.169	-0.24
$A_{FH}:A_L$	-0.011	-0.068	0.092	0.296
FH%	-0.124	0.043	-0.009	0.392
$A_{FP}:A_L$	0.067	-0.277	-0.104	0.06
FP%	-0.012	-0.229	-0.185	-0.165
A_{TC}	0.265	0.037	-0.075	0.117
$A_{TC}:A_L$	0.102	-0.253	0.199	0.062
A_W	0.267	0.014	-0.076	0.111
$A_W:A_L$	0.125	-0.285	0.131	0.03
Wetland%	0.042	-0.168	-0.198	-0.266
A_{CW}	0.268	0.012	-0.077	0.109
$A_{CW}:A_L$	0.127	-0.285	0.127	0.011
CWetland%	0.039	-0.163	-0.198	-0.28
A_{OW}	0.256	-0.108	-0.082	0.103
$A_{OW}:A_L$	0.121	-0.283	-0.012	0.028
OW%	0.15	-0.256	-0.048	0.059
SD	3.419	2.675	2.495	2.014
Proportion of Variance	0.3	0.184	0.16	0.104
Cumulative Proportion	0.3	0.483	0.643	0.747



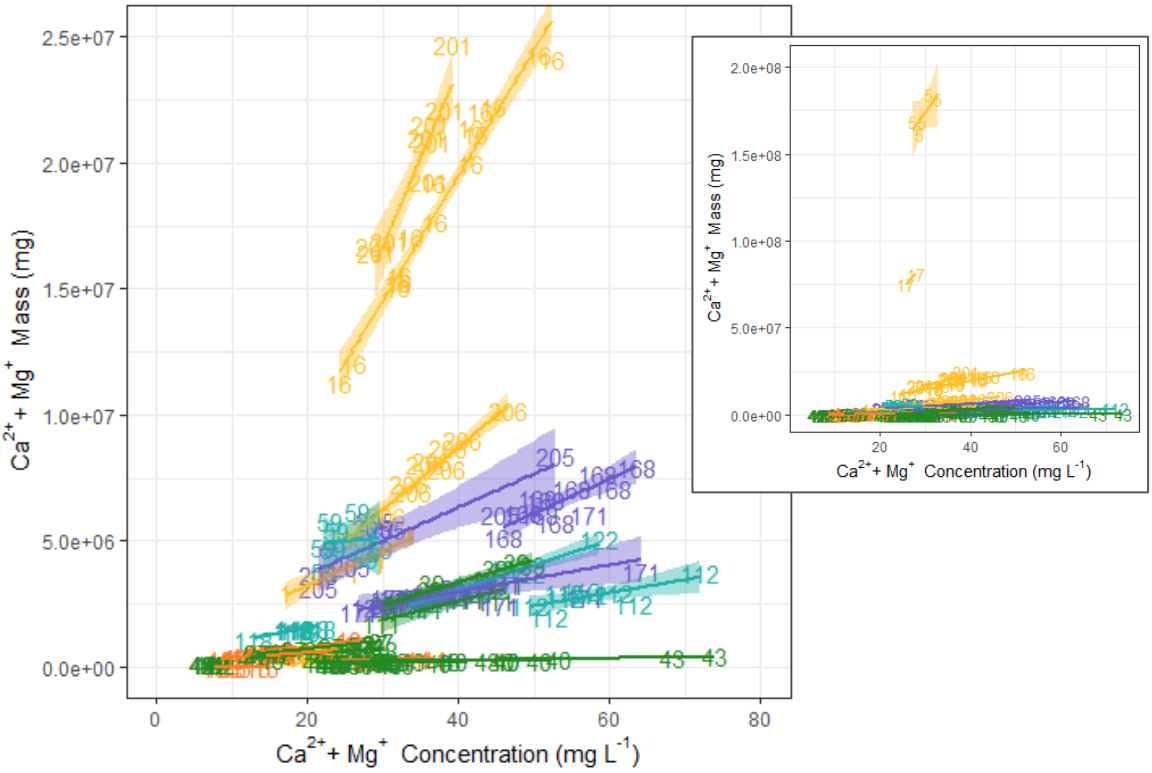
Supplementary Figure 3-7. *Comparison of stable isotopes δ^2H and $\delta^{18}O$ (‰) composition between HRA classifications within a) dual isotope plot; b) line conditioned excess (lc-excess) relative to the local meteoric waterline. Values are of means annually within summer months (June 1st-August 31st) from 1998-2018. Cross hairs denote the standard deviation around the mean of annual means. Lc-excess calculated as $\delta^2H - a \cdot \delta^{18}O - b$; where a is the slope and b is the intercept of the dual isotope plot (Hokanson et al., 2019). Greater negative lc-excess values denote isotopic enrichment of heavier species relative to the LMWL.*

Supplementary Table 3-4. *Coefficients of linear models fitted to lake Ca^{2+} + Mg^{+} concentration (mg L^{-1}) and lake Ca^{2+} + Mg^{+} mass (mg) relationship supplementary figure 8). N = numbers of years both concentration and mass (lake depth x concentration) values available.*

HRA Classification	Lake	N	Slope	Intercept	R^2	p
CO	1	11	137088	481068	0.876	0.000
	16	16	496281	-290334	0.965	0.000
	17	3	2769035	3722323	1.000	NA
	201	11	695985	-4117960	0.900	0.000
	206	11	248509	-1195453	0.958	0.000
	208	9	24914	42002	0.481	0.038
	5	5	3568703	66717342	0.609	0.119
CO/FH	11	12	8546	2034	0.880	0.000
	15	9	151	-25	0.269	0.232
	19	17	40258	-60025	0.933	0.000
	7	11	20887	130215	0.742	0.003
FH	111	8	76388	-444256	0.848	0.009
	27	11	30789	89380	0.972	0.000
	39	11	89395	-224667	0.917	0.000
	40	15	5837	-29381	0.965	0.000
	42	10	15974	9822	0.922	0.000
	43	19	3997	96685	0.826	0.000
	48	9	21487	73978	0.779	0.009
FP/FH	112	9	51648	-168955	0.553	0.022
	118	10	55979	413949	0.870	0.001
	122	10	97383	-828162	0.929	0.000
	59	10	24721	4354144	0.008	0.835
FP	121	9	63116	497542	0.729	0.003
	168	12	135160	-654183	0.755	0.001
	171	17	53188	845830	0.420	0.005
	205	10	133031	1021000	0.819	0.002



Supplementary Figure 3-8. *Chadha's diagram for Lake Ca^{2+} + Mg^+ concentration (mg L^{-1}) versus lake Cl^- + SO_4^{2-} mass concentration (mg L^{-1}) for each lake 1999-2018.*



Supplementary Figure 3-9. *Lake Ca^{2+} + Mg^{+} concentration (mg L^{-1}) and lake Ca^{2+} + Mg^{+} mass (mg) for each lake with fitted linear models (shaded area denoted standard error of fitted line). Higher slopes signify greater increase in mass with concentration, indicative of greater inputs of ions with greater concentration. Low slopes signify concentration increases without mass increase, indicative of evapoconcentration processes. CO lakes 5 and 17 have the greatest slopes with masses orders of magnitude higher, and higher slopes, shown on plot b. Supplementary table 4 provides fitted line coefficients.*

CHAPTER FOUR: MULTI-YEAR CLIMATE MEMORY IN SHALLOW LAKE WATER LEVELS

Abstract

Landscape hydrologic memory of climate can have a disproportionate impact on hydrological responses by propagating clusters of wet or dry conditions into extreme events. The non-linear and hysteretic response functions driven by memory are difficult to define and incorporate within frameworks and hydrologic models, and are often only studied at short temporal scales (event, season) despite larger interannual hydrologic responses evident in some systems. Lakes and lake water level records provide important test beds for conceptual understanding of hydrologic memory; they reflect the landscape response to climate while collapsing complex non-linear and threshold related processes into more easily interpretable metrics. Our results demonstrate the critical importance of interannual landscape memory. We reveal how spatial differences in memory, driven by hydrogeologic setting, affects how drought conditions, or wet years, amplify the magnitude of landscape drying or wetting, thus providing vital information for understanding system sensitivity, duration of recovery and in turn infer resilience. Understanding of the processes governing antecedent moisture conditions is crucial to accurately model lake water levels, groundwater levels or landscape runoff over prolonged periods. These insights will provide invaluable understanding under drought and deluge – periods where uncertainty is currently at its greatest.

1. Introduction

Hysteresis occurs when the equilibrium state is dependent on system history (Hendriks, 2010, p243). Thus water storage causes landscape hydrologic memory of antecedent conditions, otherwise known as long-range dependence, long-term persistence, or the 'Hurst' phenomenon (Hurst, 1951). Understanding landscape hydrologic memory, hereafter termed memory, is important due to the disproportionate impact of memory on the occurrence and clustering of extreme events (Khanal *et al.,* 2019; Bunde *et al.,* 2005), and dramatic increase in the uncertainty of related hydrological processes (Koutsoyiannis, 2005). There has been considerable emphasis in hydrology on characterising non-linear and threshold-mediated hydrological relationships. Typically, non-linear response functions such as hysteresis are observed and studied at an event scale, such as rainfall-runoff or storage-discharge for one storm or across a single year (Ali *et al.,* 2015). However, greater inter-annual variability of water levels in some systems indicates the possibility for multi-year climate memory and hysteresis loops that form over multiple years (Chapter 2 and Chapter 3).

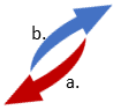
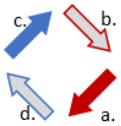
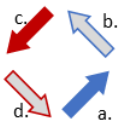
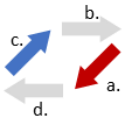
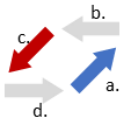
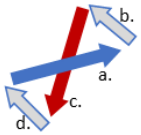
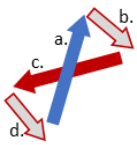
Lakes act as important storage bodies in the landscape and are indicative of the landscape response (hysteresis, non-linear, threshold) to climate. Lakes thus act as important test beds to examine core conceptualisations of hydrological memory that operate at a catchment scale over multi-annual to decadal time periods. Antecedent storage in lakes also represents an important regulator of catchment connectivity and thus also drives further non-linearity and hysteresis of the hydrological system.

Geology controls the processes of landscape memory by determining water storage and transmission capacities. The filling of depression storage (fill and spill) is a widely applied concept in both the humid Boreal Shield and the semi-arid Prairie Pothole Regions and represents a form of cumulative memory; the cumulative deviation from the long term mean in precipitation (CDM P). Within the Prairie Pothole Region this depression storage produces multi-year memory reflecting the balance of P to PET within isolated basins (Shook and Pomeroy, 2011), whereas storage-driven seasonal responses in the

Precambrian Shield bedrock lakes (Spence, Ali, Oswald, and Wellen, 2019) are associated with variable connectivity processes such as gatekeeper lakes (Phillips, Spence, and Pomeroy, 2011). In contrast, water tables in coarse outwash lakes are driven by large scale groundwater flows (Smerdon, Devito, and Mendoza, 2005; Hokanson, Mendoza, and Devito, 2019), whereby memory is associated with lag, an autoregressive delay in response water level to climate, due to transit time through a catchment. While numerous studies have examined hysteresis responses in lakes, ponds and wetland water levels to climate, few studies have attempted to characterise hysteresis characteristics (i.e., loop direction, shape, size) and their associated hydrological processes to assess the role of multi-annual memory across diverse hydrogeologic settings.

Shallow lakes are ubiquitous in the Boreal Plains sub region of Alberta, Canada (BP). They are critical aquatic habitats (Morissette *et al.,* 2018) that require effective management in response to extreme landscape and climate mitigation pressures (Price *et al.,* 2013; Ireson *et al.,* 2015). But Boreal Plain lakes likely exhibit the diversity of memory related behaviour evident across multiple Boreal and glacial lake systems because of their diverse hydrogeologic settings. The low relief sub-humid BP is dominated by a strong contrast in storage due to thick glacial substrates that differ markedly in their textural characteristics, set within a sub humid climate with a strong and decadal climate signature (Devito, Mendoza, and Qualizza, 2012). Previous research predominantly through short term studies has formulated detailed conceptual understanding of the spatially variable and interacting landscape controls on hydrology within the boreal plain, including geology, land cover, lake geometry (Chapter 2 and Chapter 3; Devito *et al.,* 2017; Plach *et al.,* 2016; Smerdon *et al.,* 2005) that transform the climate signal in lake responses. These controls suggest the prevalence of multi-annual hysteresis (BP peatland driven lags to drought, not just storage-deficit driven memory) formulated through memory-related hydrological responses of water levels of varying periodicities across multiple types of Boreal and glacial lake systems due to heterogeneous surficial geology (Chapter 2). Yet this conceptual

Table 4-1. Hypotheses and references to plausible systems for hysteresis climate-lake level responses .X axis on graphs refers to climatic conditions (left = dry, right =wet) and y axis refers to lake water level response (up = wetting, down = drying).

	Climate-WL Hysteresis Loop Shape and Direction		Definition	Theorised Cause/ Process	Hypothesised Lake/ Landscape Type
Linear/ Narrow Ellipse	Clockwise		Linear or near linear response function with minimal lag in water level response to climate	▪ Lower capacity for cumulative memory or lag ▪ Low storage ▪ Surface runoff dominated ▪ Localised flowpaths	Boreal shield lakes: impermeable bedrock, high rainfall results in low storage deficits and runoff dominated systems (based on multi-annual response – memory is expected intra-annually)
	Anti-Clockwise				
Diamond	Clockwise		Exhibits periods of a negative relationship between cumulative precipitation and lake water levels (arrows b and d)	▪ Cumulative memory of previous climatic conditions or lag responses ▪ Response to external disturbance or an extended lag e.g., beaver dam build and break ▪ High storage ▪ Long transit times ▪ Groundwater dominated ▪ Snowmelt	Coase outwash lakes: groundwater dominated systems results in long transit times (CO – Figure 4-1b)
	Anti-Clockwise				
Parallelogram	Clockwise		Exhibits periods of no change in water level despite significant changes to cumulative precipitation (arrows b and d)	▪ Threshold exceedance, e.g., lake desiccation, lake outflow ▪ Variable catchment connectivity, e.g., upstream storage deficits, upstream gatekeeper lakes	Shallow hummocky moraine low storage capacity lakes: lake spilling and desiccation (FH - Figure 4-1b) Prairie Potholes Region: coalescing and desiccating lakes
	Anti-clockwise				End lakes/ receiving lakes: Wetting lag due to catchment storage deficits, ephemeral peatland connectivity, or gate keeper lakes upstream; drying lag due to groundwater supply
Figure-eight	Slow wetting, rapid drying		Exhibits both clockwise and anticlockwise loops caused by differing rates of wetting and drying response to the same climate gradient (arrows a and c). May contain diamond or parallelogram loops.	▪ Rapid drying and slow wetting – productive, low drought resilient landscape, e.g., forest dominated	Forest dominant systems: rapid drying due to high evaporative demand, large storage (forest upland hummocks) results in slow rewetting and high memory
	Rapid wetting, slow drying			▪ Rapid wetting and slow drying – efficient, high drought resilient landscape, e.g., peatland dominated	Peatland dominant systems: rapid wetting due to low storage, slow drying due to negative feedback mechanisms, results in quick recover from drought (FP – Figure 4-1b) Isolated systems: can only dry via ET

Key:



Climate and lake levels wetting
Lake levels wetting under drying climate
No lake level change under wetting climate



Climate and lake levels drying
Lake levels drying under wetting climate
No lake level change under drying climate

understanding has not been tested effectively at the landscape scale and over multi annual to decadal time periods to assess the integrative response of these memory processes.

The aim of this paper is to evaluate how multi-year landscape memory of precipitation (P) varies spatially within shallow lake water levels (WL) of the typical lake-landscape type variations of a heterogeneous sub-humid Boreal Plains landscape, the most common of which are presented in Figure 4-1, in order to address the following research questions:

1. What are the general characteristics of precipitation-water level (P-WL) hysteresis and how do they reflect hypothesised hysteresis loops based on different conceptualised storage-threshold relationships between different lake-landscape characteristics (Figure 4-1)?
2. How does landscape memory of climate emerge at different temporal scales ranging from seasonal to interannual and therefore what additional insights are gained by testing landscape memory at multiple temporal scales?
3. Does landscape memory cause temporally different short term lake level response to climate due to position within the long-term climate?

We devised hypothesised hydrographs and interannual hysteresis loops by combining different lengths of memory variables, CDM and lag, to map out the entire parameter space associated with lags and memory. This is then mapped onto a priori conceptual understanding of spatial differences in response functions of four lake types that are common within the URSA (Chapter 2) and other glaciated or peat dominated regions (Table 4-1; Figure 4-1). Subsequently, we predict varying types of P-WL hysteresis loops with different lake types due to differing dominant sources of flow, transit times and storage-threshold mechanisms. Low storage capacity within peatland dominated fine plain systems (FP) is hypothesised to produce short landscape memory of climate, with rapid recovery following drought conditions (Lukenbach *et al.,* 2017). However, peatland feedback mechanisms are predicted to produce complex lags and rates of drying as a function of drought length and severity (Kettridge and

Waddington, 2014) producing produce irregular figure-eight shaped hysteresis loops due to changes in the response gradient between wetting and drying periods (Figure 4-1b). Shallow lakes within fine hummocky surficial geology (FH) are hypothesised to exhibit frequent landscape memory resetting as the lakes spill and dry out, but high storage capacity within forested hummocked is predicted to produce multi-year cumulative responses within the range of conditions between runoff generation and lake desiccation, consequently producing parallelogram loops due to periods of no lake level change due to threshold exceedance.

In addition to climate-WL response functions modifying hysteresis loop shape, loop size is also determined by the magnitude of lake level variability. High storage within closed, forest dominated fine textured basins is expected to produce the longest landscape memory and thus large elliptical hysteresis loops as lake water levels only deplete via evapotranspiration (Shook and Pomeroy, 2011). Within coarse outwash systems (CO), the important control of larger scale groundwater flows (Chapter 3; Hokanson, Mendoza, and Devito, 2018) could produce consistent low magnitude WL responses and landscape memory spatially, and subsequently smaller hysteresis loops. However, receiving CO lakes in low landscape positions may display more variable memory effects temporally due to gatekeeper processes (Winter, 1999; Gibson, Prepas, and Mceachern, 2002) modifying landscape connectivity at different points along the climate cycle. Further, we predict that the differences in P-WL responses between lake types will be amplified under dry conditions due to contrasting storage capacities and ratios of runoff generating landscape to lake proportions, but synchronise during extended wet cycles as landscape storage deficits are filled and spatial differences in antecedent conditions become less important. Characterising landscape memory within the context of the long-term climate cycle, and across multiple common lake types will provide greater understanding of the impact of memory on extreme landscape responses (drought and deluge) and significantly improve integration of antecedent conditions and associated hydrological processes within hydrological modelling at multiple temporal scales.

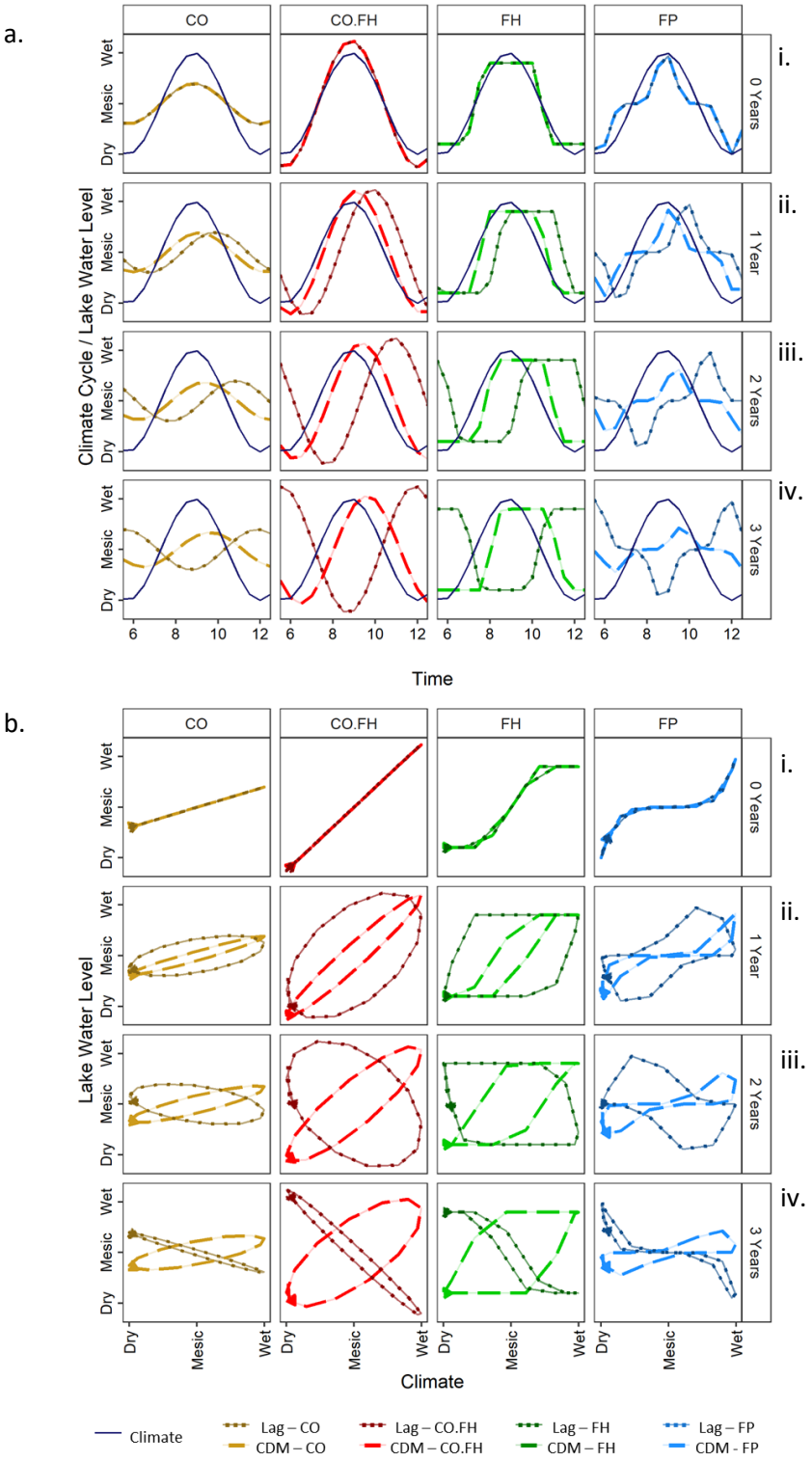


Figure 4-1. (a) Hypothesised time series plots of theorized sinusoidal climate cycle (solid navy line) and hypothetical lake water level response and (b) climate-lake water level hysteresis relationships for (i) 0 years lag; (ii) 1 year; (iii) 2 years and (iv) 3 years lag or cumulative precipitation climate variables. Relationships are derived from conceptualised response functions of four lake types (Chapter 2; 0 years lag response). Arrow on hysteresis plots indicates direction with time, situated at time 6. Time is a dimensionless unit, however a long-term climate cycle (one wavelength) occurs over approximately 6 timesteps in this conceptualisation, with each climate cycle simulated with the same sine wave amplitude and wavelength.

2. Study Area

The study was conducted in the Utikuma Region Study Area (URSA; 56°N, 115°W), situated within the subhumid Boreal Plains (BP) ecozone of Alberta, Canada, where long term annual average PET (513 mm) exceeds annual average P (483 mm) (Plach, Ferone, Gibbons, Smerdon, and Mertens, 2016; Environment Canada, 2003). The region is exposed to strong decadal climate cycles, with wet years occurring infrequently on a 10–15 year cycle (Ferone and Devito, 2004), in addition to a pronounced seasonal weather cycle with cold winters (January temperatures of -14.5°C) and warm summers (July temperatures of 15.6°C; Thompson *et al.,* 2015). Peak vegetation growth and evaporative demand overlaps with the majority of annual precipitation, with 50-60% falling between June and August (Brown *et al.,* 2010), driving the long-term water deficit. Although contributing a smaller proportion of total annual precipitation, snowfall and snowmelt and the occurrence of heavy spring rains provide a crucial component of the water balance by replenishing soils and recharging groundwater (Devito *et al.,* 2012).

The URSA is characterised by heterogenous deep (45-240 m), low relief Quaternary glacial deposits (Hokanson, Mendoza, and Devito, 2019), that have been classified into (1) coarse-textured glacial–fluvial outwash (CO); (2) fine-textured clay rich glacio-lacustrine plains (FP); (3) fine-textured hummocky disintegration moraine deposits (FH); (4) the transition from fine-textured hummocky disintegration moraine to coarse-textured glacial fluvial outwash (CO/FH); (5) the transition from fine-textured disintegration moraine to fine-textured glacio-lacustrine plains (FP/FH) (see Chapter 3 for more information). The overlying vegetation is characterised by a high density of wetlands, ponds, and shallow lakes, with wetlands dominated by black spruce (*Picea mariana*) and tamarack (*Larix laricina*), and forestlands are characterized by trembling aspen (*Populus tremuloides*) or jack pine (*Pinus banksiana*) communities (Devito *et al.,* 2012). This study comprises of 26 study lakes, with lake extents ranging 0.0005 to 3.2 km^2 and topographic catchments ranging in area from 0.08 to 90 km^2; varying

from isolated headwater systems to large regionally connected systems and ranging in proportion of peatland and forestland cover (Chapter 3).

3. Methods

3.1. Climate variables and climate-driven temporal study design for memory testing

Daily precipitation values and mean, maximum and minimum air temperatures have been monitored for 23 years (1995-2018) within the URSA and wider study region (see Chapter 3 for more information). Climate memory variables for testing landscape memory were generated from daily moving windows of one-year cumulative precipitation records, relative to the long-term mean (444 mm year^{-1}). Lag variables are devised to identify autoregressive responses associated with varying transit times, whereas cumulative variables are formulated to associate antecedent conditions and storage capacities with response. Precipitation lag value relative to the long-term mean (Lag P), and precipitation cumulative deviation from the long-term mean (CDM P) variables for memory periods of 0 – 5 years and 1 - 6 years respectively, were generated using the following equations:

Equation 1. $$\text{Lag } X\text{yr } d = \left(\sum_{d-X-1year}^{d-X} P \right) - \mu;$$

Equation 2. $$\text{CDM } X\text{yr } d = \left(\sum_{d-X}^{d} P \right) - (\mu \times X);$$

Where:

X is the length of time for which the climate variable is calculated (in years);

d is the date of the calculated climate variable

P is daily precipitation total (mm);

μ is the long-term mean of annual precipitation (November hydrologic year), given as 444 mm for the study region.

Table 4-2. Summary of climate in key study years used to classify short term climate for use in analysis, corresponding with Figure 4-2. Annual and multiannual CDM variables are calculated from 1st July (relative to 1920-2018 1yr average 444 mm); CDM growing season is given for May 1st – September 30th (relative to 1920-2018 average 303 mm). Conditional formatting provides a wetness indicator for each variable for each year; red = dry, blue = wet, white = mid.

CDM 1yr Trend	Year	1yr P	CDM 1yr P	CDM Growing Season P	CDM 3yr P	CDM 5yr P	Summary
Dry	2002	214	-230	-91	-389	-487	Dry year, dry summer, dry long-term climate
	2010	244	-198	-108	-159	-43	Dry year, dry summer, mesic long-term climate
Mesic to Dry	2004	356	-88	-14	-369	-528	Mesic to dry year, mesic summer, dry long-term climate
	2001	385	-59	-147	-240	-155	Mesic to dry year, dry summer, dry-mesic long-term climate
Dry to Mesic	2003	393	-50	-17	-340	-521	Dry to Mesic year, mesic summer, dry long-term climate
	2011	427	-17	32	-169	-30	Dry to Mesic year, wet summer, mesic long-term climate
Wet to Mesic	2014	424	-20	-132	77	-138	Mesic year, dry summer, mesic-wet long-term climate
	2008	437	-7	-11	103	126	Mesic year, mesic summer, wet long-term climate
	2012	449	5	-28	-213	-172	Mesic year, mesic summer, dry long-term climate
	2009	492	48	-104	188	263	Mesic year, dry summer, wet long-term climate
Mesic to Wet	2018	516	72	-6	40	-95	Wet year, mesic summer, mesic-wet long-term climate

CDM P (mm) for 1 – 6 years is presented in Figure 4-2 for 1997-2019. Analyses are conducted over three temporal scales for this study. For the first and largest temporal scale, the full study period is used to assess spatial variability in responses at the decadal scale. Based upon long-term cumulative precipitation relative to 1-year cumulative, and relative to the long-term mean (wet, >100 mm P CDM; mesic, -100 mm < P CDM < 100 mm; dry, <-100 mm P CDM), three multi-annual periods were identified for comparative analysis. These three multi-annual periods represent enable spatial and temporal comparison of multi-year responses in the context of the decadal climate cycle. For intra-annual analyses, 11 growing seasons (May 1st – September 30th) were identified and classified based on the directionality (wetting or drying) and value of the CDM P 1yr (mm) timeseries for that period (Table 4-2), then subsequently used for testing for differences in landscape memory within the context of their long-term climatic conditions.

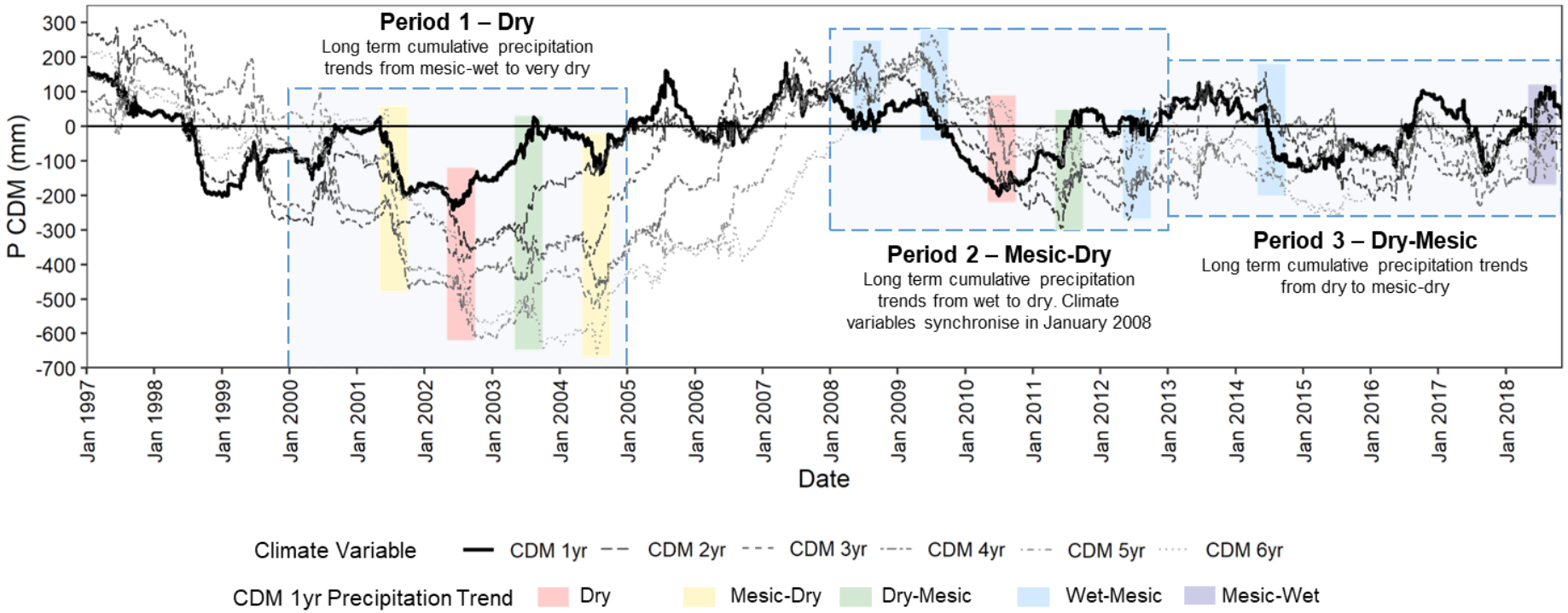


Figure 4-2. Climate of the URSA during the study period given as the difference in cumulative precipitation from the long term mean (444 mm year^{-1}). Cumulative precipitation is calculated as a daily moving window for 1 to 6 years (see equation 1). Figure annotations delineate the paper's temporal methodology by dividing the 19-year climate record, which span one long-term climate cycle, into three periods of multi-annual wetting-drying cycles. Intra-annual patterns are then examined by placing short term climate (CDM 1yr P and within year growing season P) within the context of long-term climate (CDM 2-6yr P).

3.2. Satellite Imagery

World Satellite Imagery was obtained from Esri and combined with historical aerial photography of the URSA. Lake surface area was then delineated for each satellite image of Lake 19 and regression analysis was conducted between lake area and CDM P 1-6yr (mm) to determine the effects of memory at interdecadal scales.

3.3. Lake Classifications and Conceptualisations

Lake classifications employed for this analysis were taken from Chapter 3. Classification was executed using a principal component analysis and clustering techniques that incorporated different aspects of land cover, lake geometry, landscape position, topography, and catchment scales in addition to magnitude of lake water level variability for the period 2000-2018. Whilst classifications included a range of topological, typological and topographical variables; the final classes largely reflect their surficial geology of coarse-textured glacial–fluvial outwash (CO); fine-textured clay rich glacio-lacustrine plains (FP); fine-textured hummocky disintegration moraine deposits (FH), the transition from fine-textured hummocky disintegration moraine to coarse-textured glacial fluvial outwash; the transition from fine-textured disintegration moraine to fine-textured glacio-lacustrine plains (FP/FH). Each classification can be generalised as the following: CO lakes exhibit large lake and catchment areas, CO/FH lakes are small, isolated basins, FH lakes are typically small, spilling and situated at high landscape positions and hummocky terrain, FP lakes are situated on flat plains and their catchments are dominated by wetlands, and FP/FH lakes represent and interface between FP and FH lakes, exhibiting a dominance of wetlands characteristic of FP, but sloping terrain associated with FH.

3.4. Data Analysis

Prior to cross correlation analysis, daily and annual climate records of URSA were tested for to establish whether the cyclical climate patterns produce autocorrelation (see supplementary material). Autocorrelation and cross correlation were executed using the CCF package within RStudio.

Cross correlation and hysteresis between precipitation within the URSA and lake water levels were conducted for 26 study lakes. Analysis was undertaken at three scales, representing an trade-off between spatial and temporal scales: (1) interannual summer water levels of 26 study lakes across 19 years; (2) interannual summer water levels of 4 key study lakes across multiannual wetting drying cycles representing dry (2000-2004), dry-mesic (2008-2012) and mesic (2013-2018) conditions within the long-term climate cycle; (3) intra-annual water levels of 4 key study lakes in years representing contrasting short term precipitation values (CDM 1yr P) placed in the context of long term cumulative precipitation relative to the long term mean (CDM 2-6yr P). This is visualised in Figure 4-2, which also exemplified the methodological approach to the temporal scales implemented within this study.

For interannual summer water levels analyses, the average summer water level is correlated with precipitation and cumulative precipitation corresponding with the July hydrologic year, and a lag of 1 corresponds to 1 year. Where higher frequency water levels are tested, the cumulative precipitation (CDM P) relative to the long-term mean is correlated against water levels of the same day. The graphical results of cross correlation and hysteresis analysis are interpreted within the context of their classification and conceptual understanding of their hydrologic functioning ascertained in previous studies (Chapter 2 and Chapter 3; Devito *et al.,* 2017; 2012; Hokanson *et al.,* 2019).

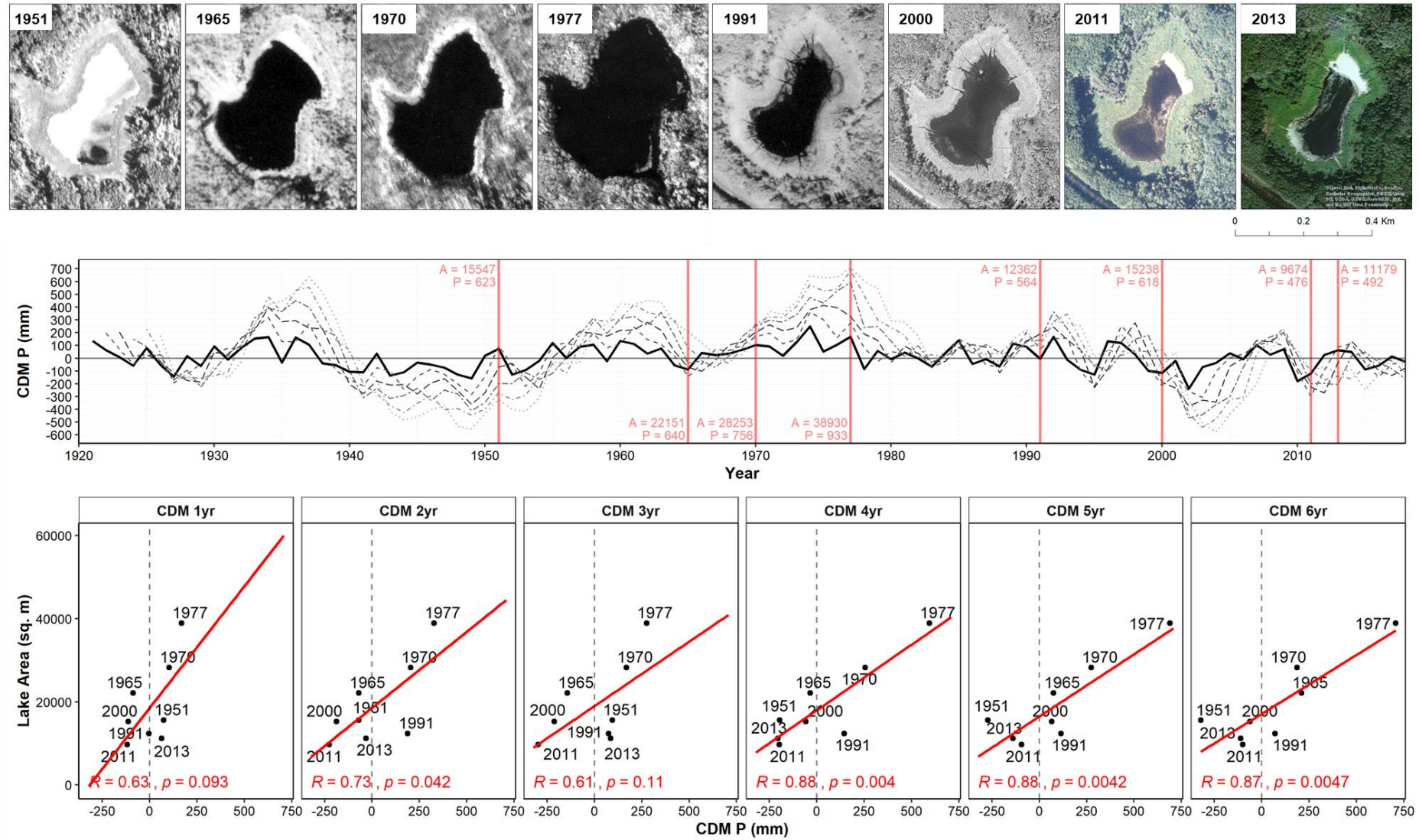


Figure 4-3. Satellite imagery of lake surface water extent (summer months, 1951-2013) within a forest dominated isolated basin with fine textured surficial geology overlying coarse deposits (Lake 19) within the URSA, Alberta, Canada. Surface water area (m2) is compared to the cumulative precipitation record of Fort McMurray (1920-2018) and URSA precipitation record (1996-2018) for 1-6 years (June 1st).

4. Results

4.1. Interdecadal memory

Historic satellite imagery of the CO/FH Lake 19 has revealed more than 4-fold change in lake surface area extent since 1951 due to interdecadal water level fluctuations (Figure 4-3), ranging from 9674 m^2 in 2011 to 38930 m^2 in 1977, with perimeter varying two-fold between these records (476 m – 933 m). Climate-lake water level correlation strength and significance were considerably greater with longer cumulative climate variables of CDM P 4yr (r=0.88, p=0.004), CDM P 5yr (r=0.88, p=0.0042) and CDM P 6yr (r=0.87, p=0.0047) than one-year cumulative deviation from the long term mean in precipitation (CDM P 1yr; 444 mm; r = 0.63, p=0.093). Relatively low lake surface area occurred in 1951 despite above average rainfall in 1951 (74mm CDM P 1yr) as landscape antecedent conditions reflect a long-term precipitation deficit following 8 consecutive years of below average rainfall (CDM 6yr P = -320mm in July 1951). Similarly, 1965 surface water area is greater than expected for a CDM 1yr P value of -88 mm. Instead, 1965 lake water levels reflect memory of ~10 years of above average rainfall and thus a CDM 6yr P value of +210mm. This clearly demonstrates the importance of multi-annual memory of climate within this landscape, whereby clustering of wet or dry years propagate into extreme wetting and drying of the landscape.

4.2. Spatial variability in memory between different lake-landscape characteristics across 19 years

Interannual summer lake water level response to annual precipitation totals (July hydrologic year) produced hysteresis loops in all 26 study lakes (Figure 4-4). Graphical comparisons reveal high variability in the shape, size and direction of hysteresis loops spatially between lakes (see summary in Table 4-3). The varied characteristics of hysteresis loops indicate that complex and spatially variable memory is present in the landscape in time and in space.

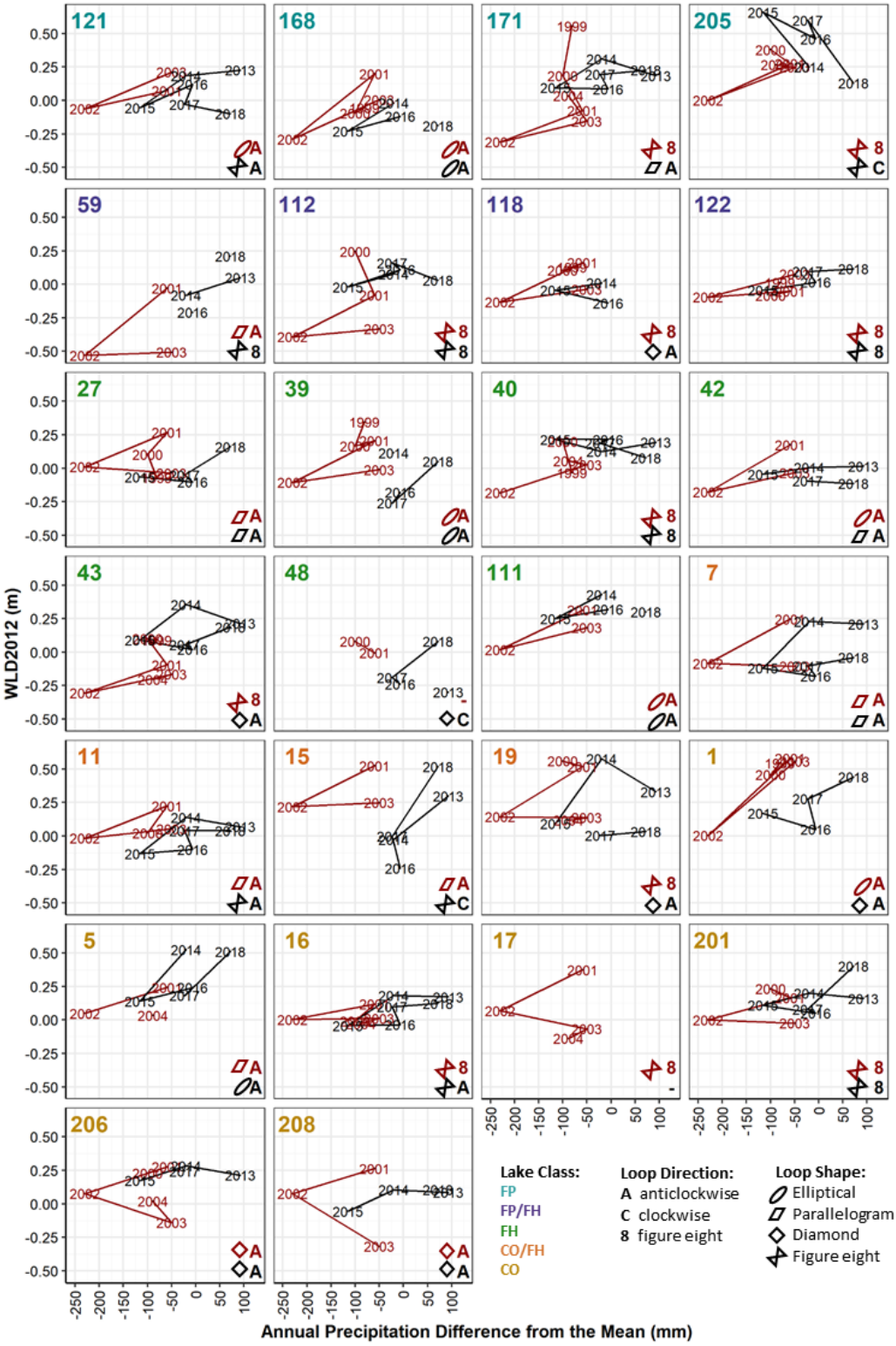


Figure 4-4. Hysteresis loops of annual precipitation relative to mean summer water levels relative to 2012 (WLD2012) in 26 study lakes (red: 1999-2004 – long term climate transitioning from wet to very dry; black: = 2013-2018 – long term climate mesic). Lakes are presented in order of their classification (Chapter 3). Each loop is classified in shape and direction.

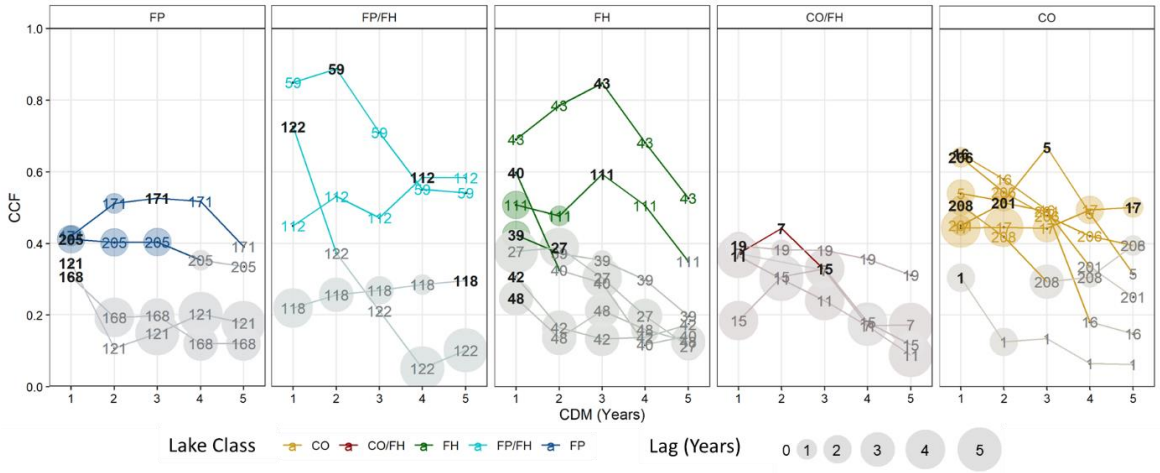


Figure 4-5 . Maximum cross correlation function (CCF) value of water levels of each lake with CDM precipitation variable (years) combined with lag (years). CCF was calculated for each combination of CDM (1-5 years) and Lag (1-5 years) against mean summer (June-August) water levels relative to 2012 levels (WLD2012) at 26 study lakes; the optimal lag per CDM of each lake is presented. Labels denote the lake ID; black labels signify the best CCF per lake; coloured labels/points signify CCF ≥ 0.4; grey labels/points signify CCF < 0.4. Study lakes are presented in 5 groups representing their classification.

Table 4-3. Summary of precipitation-lake water level hysteresis characteristics and best cross correlations between cumulative precipitation variables and lake water levels in 26 lakes across 5 lake classes (see Figure 4-4 and 4-5).

	Class Attributes	***Dominant Loop Type***	***Lag/CDM cross correlation***	***Relating Process to Class Response***
FP	*Large flat plains, peatland dominated catchments*	*Figure-eight dominated*	▪ *Largest lake/catchment and receiving catchment = low CCF (168)* ▪ *Headwater plateau lakes = 3yrs CDM, varying lag (171, 205)*	*Peatland storage, feedback mechanisms and ephemeral connectivity = complex response patterns to varying connectivity*
FP/FH	*Peatland dominated catchments*	*Figure-eight dominated*	▪ *Catchments with larger peatland proportion = high CCF, 0 lag, varying CDM (59, 112)* ▪ *Catchments with larger lake proportion = low CCF (118, 122, also FP lake 121)*	*Peatland storage, feedback mechanisms and ephemeral connectivity = complex response patterns to varying connectivity*
FH	*Shallow lakes, smaller surface areas, with small, hummocky forest dominated catchments*	*Mixed loop types*	▪ *High CCF, 3yr CDM, 0yr lag in Small, spilling, high positioned FH (40, 43, 111)* ▪ *Very low CCF in larger, low positioned FH catchments (39, 27, 42 and 121, 118)*	*CDM/lag varies with point in long term climate due to antecedent storage deficits (therefore differs between lakes due to varying capacity spatially)*
CO/FH	*Closed systems*	*Mixed loop types*	▪ *Poor CCF* ▪ *High lags*	*High storage = poor association with climate; memory/lags vary with point in long term climate*
CO	*Groundwater dominated*	*Diamond and figure- eight loops*	▪ *High CCF (except Lake 1).* ▪ *CDM1-CDM3, 0yr lag* ▪ *Larger lakes (5 and 17) associated with higher cumulative landscape memory.*	*Long transit times = lag and cumulative climate*

A consistent response is observed across the landscape in the dry year of 2002, where large decreases in water levels occur in all lakes, however, subsequent responses to wet conditions in 2003 vary substantially. This is exemplified by shallow fine hummocky (FH) lakes which dried out in 2002, thus could not dry further and produced parallelogram shaped loops. Peatland dominated fine plain lakes (FP) display the greatest lake water level increases in 2003, producing elliptical or rapid-wetting figure-eight loops. In contrast, coarse outwash (CO) lakes are the only class to display continued drying in 2003. Furthermore, comparisons of loops for the two 6-year cycles (period 1: 1999-2004 and period 3: 2013-2018) demonstrate consistent hysteresis responses between periods in some lakes and highly contrasting responses in others. This occurs due to the similar CDM P 1yr pattern for each period, enabling landscapes with low memory to produce consistent hysteresis loops (e.g., FH or FP/FH lakes 122, 121, 118), whereas lakes associated with longer CDM P variables display contrasting hysteresis loops between periods (e.g., CO/FH lakes 7, 15, 19). Therefore, systems with greater memory produced larger hysteresis loops and drier water levels in the long-term dry period of 2000-2005 period.

For each CDM P variable (1-5 years), the cross correlation function (CCF) value is displayed for the optimal lag (highest CCF) in Figure 4-5. Cross correlation plots of CO lakes demonstrate high similarity between lakes, with the exception of lake 1. Generally, CO lakes produced highest CCF values for CDM P 1-3yrs with 0 lag, whereas the largest lakes (5 and 17) exhibit higher CCF for longer CDM variables. Closed lakes that typify CO/FH (11, 19, 15 and FH lake 48) exhibit the poorest CCF values and longest lags, signifying poor association with climate at decadal scales due to interactions with antecedent moisture. FH and FP/FH lakes positioned lower within the landscape and with larger lake surface area for their surficial geology class (39, 27, 42 and 121, 118; Chapter 3) also exhibited low CCF and low synchronicity between lake types. In contrast, small, spilling, high landscape positioned FH lakes (40, 43, 111, and FP lake 171) produced high CCF values at 3year CDM**.** Of the peatland dominant catchments, FP/FH lakes (59, 112 and 122) produced high CCF values at varying CDM variables,

whereas large FP lake catchments (168 and 205) produces low CCF values, which could signal complex response patterns as a function of variable connectivity.

4.3. Spatio-temporal variability: placing short-term responses in context of the long-term climate

Interannual Hysteresis

Interannual and intra-annual responses to climate at different short term climatic conditions (CDM 1yr P) within the long-term climate cycle (CDM 2-6yr P) were compared for four key study lakes of contrasting lake-catchment characteristics. Interannual climate timeseries, water level timeseries and hysteresis loop plots (Figure 4-6) demonstrate contrasting responses between each lake type and between each 5/6 year time period. In each period, CO/FH lake 19 produces the largest hysteresis loops due to the greatest water level declines with no recovery of water levels in 2000-2004 and continued drying in 2008-2012. Lake 43 produces similar limited recovery of water levels in 2000-2004 and 2008-2012, however the magnitude of responses is varied between years. Lower declines in 2002 and greater recovery was possible in 2003 due to the lake drying out in summer 2002, whereas higher antecedent water levels within the dry/mesic period enables greater drying 2009-2010. Similarly, lake 171 produces highly variable responses between periods, producing a large figure-eight hysteresis loop 2000-2004, and a linear, low magnitude response in 2008-2012. In contrast, lake 16 water levels do not decline to the same magnitude during dry periods, and also show an increase during dry years which could reflect lagged response to climate.

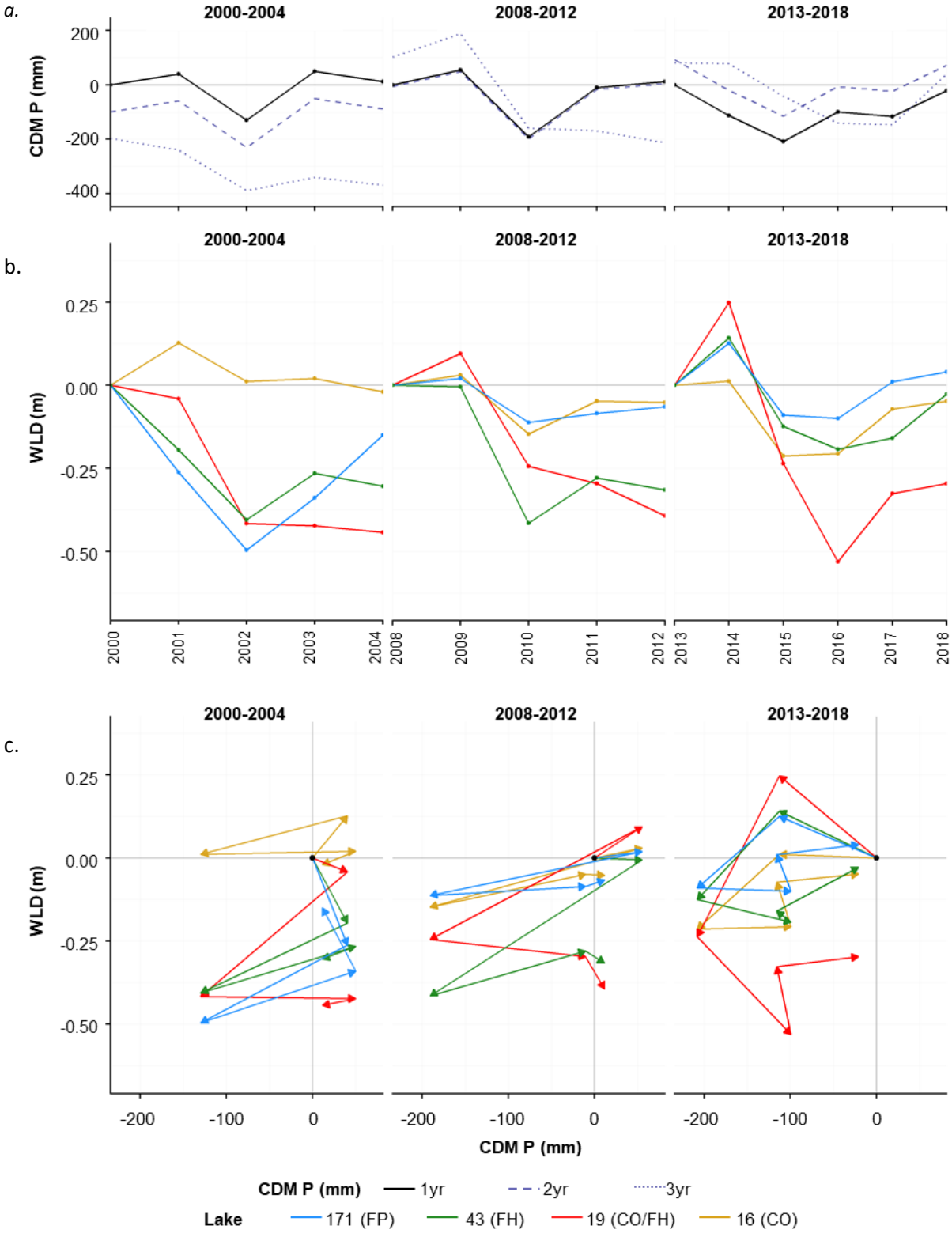


Figure 4-6. (a) Multiyear change in 1-year cumulative difference from the mean in precipitation (CDM P mm); b) change in lakes water levels and (c) corresponding hysteresis loops for three study periods at four key study lakes. CDM (mm) and WL (m) are given as 0 for the first time point in each multiyear period/loop So this is in effect a repeat of the figure 4 but focusing on the four key lakes of interest.

High frequency climate-WL correlation

Regression analysis of high frequency lake water level data against CDM P for 1 to 3 years (Figure 4-7) indicates higher association with CDM 2yr P for lake 19 and lake 43, and CDM 3yr P for lake 16 and lake 171. Whereas other lakes display improved linear relationships, lake 19 produces a hockey stick shape, with low gradient response under precipitation deficit conditions but high response under above average cumulative precipitation. Figure 4-7 also highlights individual years that are better fitted

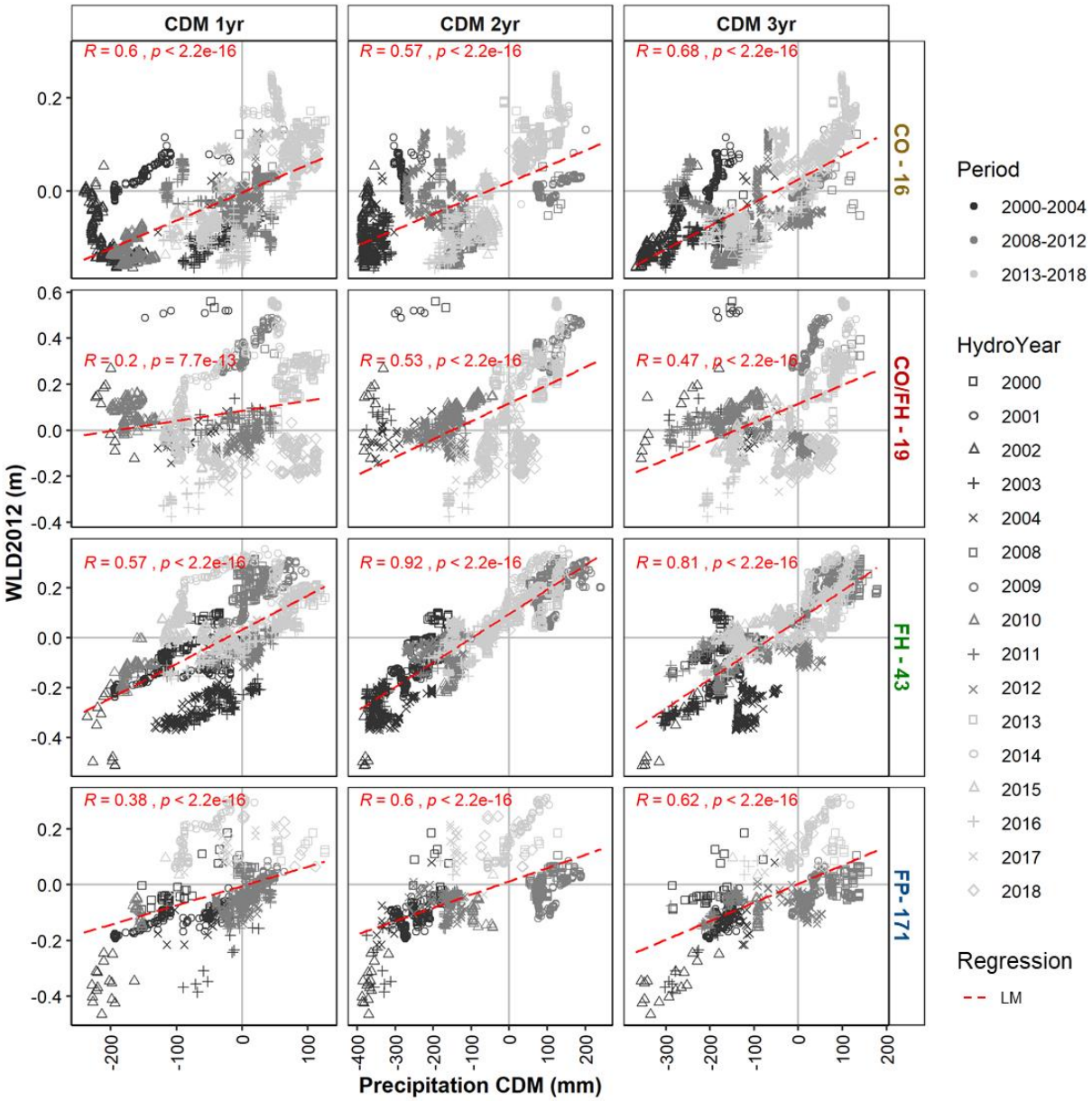


Figure 4-7. Scatter plots of Precipitation CDM variables (mm; 1-3 years) and lake water levels relative to mean summer 2012 levels (WLD2012; m) in four key study lakes across three time periods. Dashed red line represents a fitted linear model (LM) with associated the R and p coefficients given between climate variables and lake water levels.

within the regression analysis under longer cumulative climate variables, which supports interpretation from interannual analysis. Notably, CDM 3yr P significantly improves the residuals of high lake water levels for lake 16 during the long-term dry climatic conditions in 2001. Meanwhile, large negative residuals of low water levels in 2003 for lake 171 and 2003-2004 for lake 43 signify memory of dry conditions which are better predicted with CDM 2/3yr P. High water levels during the dry summer of 2014 are also better represented by CDM 2/3yr P across all four lakes due to memory of wet conditions within 2013, as is also true for 2009 in lake 19 and lake 43 following prolonged above average precipitation 2005-2009.

Intra-annual Climate-WL Responses

Comparison of intra-annual responses by placing short term climate (CDM 1yr P and within year growing season P) within the context of long-term climate (CDM 2-6yr P) highlights contrasting responses between years that are driven by memory (Figure 4-8; Table 2).

Similar short-term net precipitation totals in dry years 2002 and 2010 produced comparable rates in lake levels decline, however, lake water levels display greater magnitude rewetting responses to rainfall events in 2010 when in a wetter long-term climate. Contrasting responses between lakes to rainfall events in 2002 reiterates the rapid post-drought recovery rate of peatland systems (lake 171/ FP) shown in 2003 and 2004 of Figure 4-4 and Figure 4-6. However, intra-annual data provides evidence for greater post-drought rewetting in lake 171 in late summer, when growing season evaporative demand is high but precipitation is typically low. Lake 43 (FH) displays the greatest magnitude water level declines within 2010, which implies lower retention of wet conditions 2005-2009 due to lake outflow generation during this period. Years 2003 and 2011 also have a comparable short term mesic climate and wet summer. Lake 171 continues to wet up the greatest in 2003 following the prolonged precipitation deficit, whereas under mesic long-term conditions, lake 43 and lake 19 display the

greatest increases in lake levels, suggesting changes in the precipitation-lake water level gradients with CDM P value, as hypothesised in response functions within Figure 4-1.

Four short-term wet to mesic years were identified within the study period, each with contrasting growing season P and long-term climate. Both 2009 and 2014 represent dry summers following prolonged wet periods and lake outflow generation in spilling lakes, and produce near linear drying trends throughout the growing season in all lakes. Drier long-term conditions in 2014 than 2009 (111 mm drier CDM 3yr P; 401 mm drier CDM 5yr P) result in steeper gradients and greater magnitude decreases in lake water levels. Under these dry conditions we also observe a change in order of the gradient and magnitude response of each lake in comparison to mesic and wet years. Lake 16 exhibits the lowest drying rate in 2009, but highest in 2014, which could reflect greater association with the long-term climate. Rates and patterns are similar between lakes 43, 19 and 171, however lake 171 water levels dry at a slow rate, diverging from lake 19 and lake 43 in mid-August, whereas lake 19 displays greater rates of water level decline in 2014 than lake 171 and lake 43 throughout the whole growing season. In contrast to 2009 and 2014, mesic summers in the wet years of 2008 and 2012, frequent rainfall events produce several anticlockwise hysteresis loops in all lakes. In both 2008 and 2012, lake 16 shows early season declines in lake water level while other lakes are stable or increasing, but also exhibits late summer increases when other lakes are drying. In both 2008 and 2012, lake 43 displays the lowest lake level decline across the growing season and greatest wetting in response to rainfall events, whereas lake 171 produces larger anticlockwise loops towards a drier state.

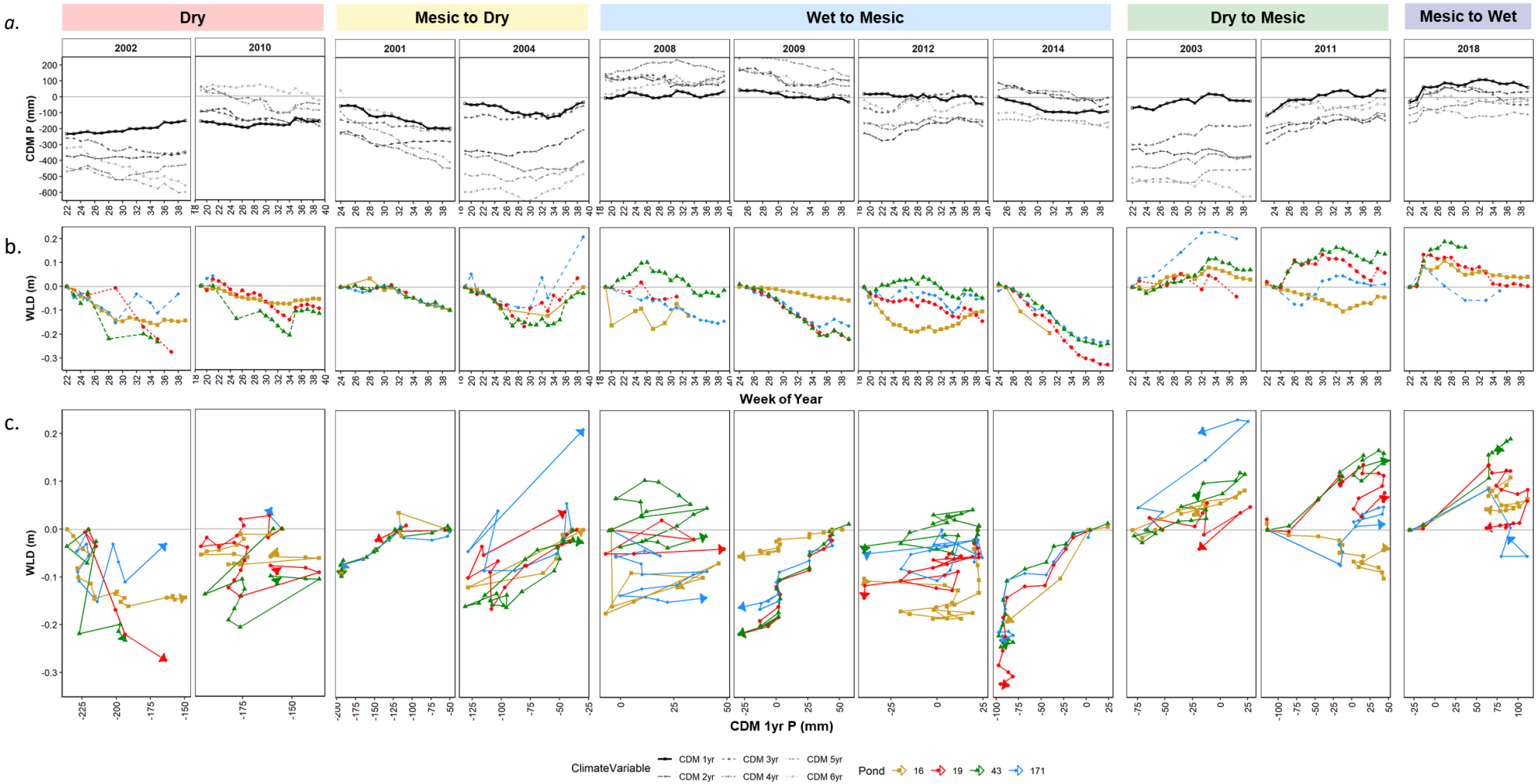


Figure 4-8. (a) Cumulative difference from the mean in precipitation (CDM P mm), cumulative for 1-6 years; b) lakes water level deviation from week X of year; (c) corresponding CDM P-WL hysteresis loops for figure a and b. Individual years representing drying, wetting and variable summers relative to the long term annual precipitation mean, within contrasting long term climatic conditions (see Figure 4-2 for further detail on climatic conditions) at four key study lakes.

5. Discussion

Within this study, multi-annual hysteresis of precipitation-lake water levels clearly demonstrate the importance of understanding interannual landscape memory of climate within the Canadian boreal landscape. The varied characteristics of hysteresis loops indicate that complex and spatially variable memory is present. These spatial variations in landscape memory are driven by differences in storage capacities, which is in turn governed by landscape properties, such as surficial geology, landcover and landscape position (Chapter 3; Devito *et al.,* 2017). Thus, the propensity for drought years or wet years to persist or accumulate into extreme landscape drying or wetting varies significantly between different lake-landscape characteristics.

Fine plains (FP) lakes demonstrated the rapid recovery of peatland dominated systems following drought driven by their low storage capacities and subsequently short memory of dry conditions (Lukenbach *et al.,* 2017). In contrast, desiccation in many fine hummocky (FH) lakes restricted the magnitude of lake water level declines, but low water levels persisted one to two years following droughts. This implies longer landscape memory of dry conditions within FH systems, which fits with current conceptual understanding of deep hillslope soils, high infiltration rates, high evaporative demand and subsequent long runoff return periods (Redding and Devito, 2008). On the other hand, under mesic/wet climatic conditions FH and CO/FH lakes produced greater increases in lake water levels than FP, indicating non-linear precipitation-lake water level response functions with CDM 1yr P value. Small, high landscape positioned FH lakes produced a strong cross correlation with CDM 3yr P as spilling and desiccation frequently 'reset' memory, whereas closed basins (CO/FH) produced poor cross correlations, and a hockeystick shaped response function with a near zero gradient response to CDM 1yr P < 0 mm, but high responses to above average long-term precipitation. This echoes findings within the PPR, whereby depressional shallow lakes exhibit longer memory as they can only dry via evapotranspiration due to the absence of lake outflows (Hayashi and Van Der Kamp 2007). Within

coarse outwash (CO) lakes, high cross correlation with lagged and longer cumulative precipitation variables reflects the dominance of larger scale groundwater as a control on lake water levels within the BP (Smerdon, Mendoza, and Devito, 2007). As such, intra-annual patterns of CO lake water levels often contrast with patterns in fine textured lakes due lags in seasonal dynamics (snowmelt recharge) superimposed on to a lagged response to long term dry conditions (Carrera-Hernández, Smerdon, and Mendoza, 2012; Carrera-Hernández *et al.,* 2011).

While some consistencies in precipitation-lake water level hysteresis characteristics and cross correlation were identified within lake classifications, generally poor cross correlation in a high number of lakes across the 18-year record could signify the complex dynamicity of memory associated with antecedent storage. In depth investigative analysis comparing intra-annual responses of similar short term climatic demonstrate that seasonal climate is an important driver for intra-annual water level patterns, but landscape memory of long-term climate governs the magnitude and gradient of these patterns. This echoes findings of short-term studies that specify that precipitation timing and magnitude relative to PET, rather than annual precipitation totals defined runoff (Devito, Creed, and Fraser, 2005; Redding and Devito, 2008). Yet slight differences in antecedent water table position can propel a hydrologically disconnected catchment into one that yields high proportions of annual runoff in one storm event (Wells, Ketcheson, and Price, 2017), thus understanding of antecedent conditions is crucial as proximity to storage thresholds results in disproportionate effects of storage increase on discharge (Spence, 2010), further to dynamic connectivity between wetlands modifying contributing area (Shaw *et al.,* 2012). Consequently, it is difficult to directly define or infer hydrological processes from analysis that spans multiple climate cycles in landscapes with capacity for high memory. Simultaneously, it is impossible to gain a full understanding of a system's hydrologic function without observing response under a range of climatic conditions where non-linearities exist. Instead, shorter term interannual or intra-annual analyses should be placed within the context of the long-term

climate, where comparisons of responses at different antecedent conditions will provide stronger evidence of landscape memory and associated hydrological processes.

Implications for land management

A direct consequence of multiannual memory is that clustering of wet or dry climatic years can propagate into extreme wetting and drying of the landscape (Khanal *et al.,* 2019). Multi-annual landscape memory of climate within the Boreal Plains demands greater understanding of the processes governing antecedent moisture conditions to accurately model lake water levels, groundwater levels or landscape runoff over prolonged periods (multiannual to multidecadal), especially under drought or deluge. Thus, this knowledge is essential for land management practices at hydroclimatic periods when both uncertainty in hydrological processes is at its greatest (Koutsoyiannis, 2005) and water resource management is at its most critical. Yet one of the main limitations of previous BP research is the poor understanding of dynamic controls such as landscape memory. While research focusing solely on 'static' landscape characteristics controls such as landcover proportions, lake geometry and surficial geology have provided crucial knowledge for conceptualising hydrologic landscapes, static controls cannot fully explain temporal variations in both interannual and intra-annual water level patterns. The important control of storage in catchment hydrograph non-linearity has previously been recognised within the BP (Wells, Ketcheson, and Price, 2017), and a plethora of concepts have been defined to describe these non-linear processes (e.g. Ali *et al.,* 2013; Shaw *et al.,* 2012; Spence, 2007) however this is the first study to characterise landscape memory through both interannual and intra-annual water level analysis across heterogeneous hydrogeoclimatic settings.

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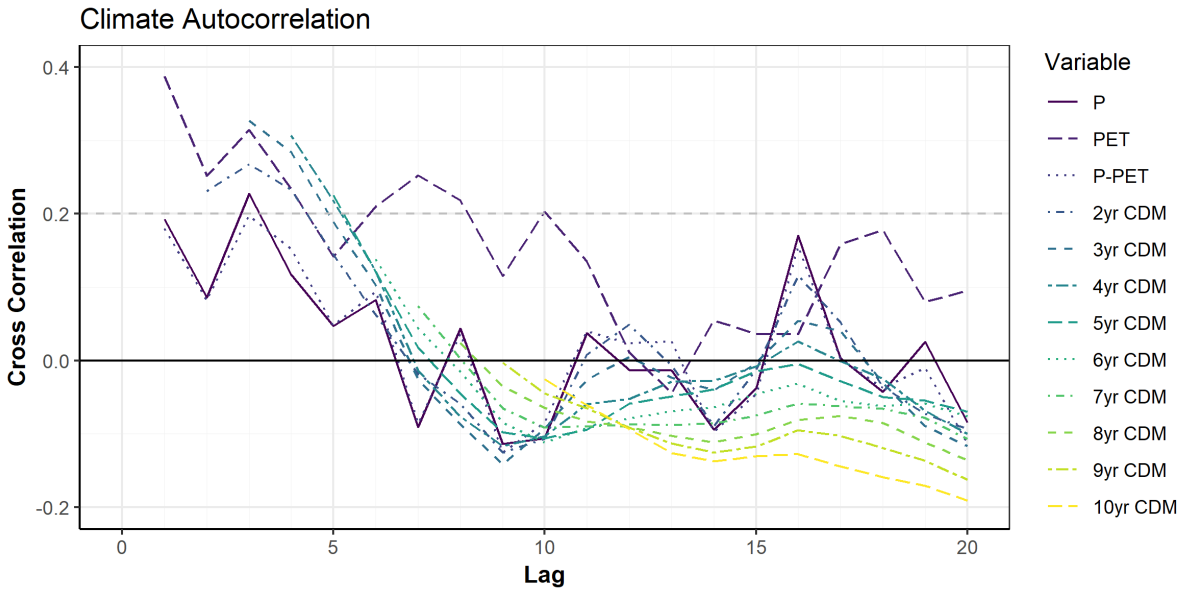
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7. Supplementary Material



Supplementary Figure 1. Autocorrelation of annual precipitation and cumulative differences to the long-term mean (2year to 10year) 1920-2018 (Fort McMurray 1920-1995; URSA 1996-2018). Expected high autocorrelation due to cyclical climate but autocorrelation values are low.

CHAPTER FIVE: REDUCED COMPLEXITY MODELLING OF CATCHMENT SCALE HYDROLOGIC FUNCTION ACROSS HETEROGENOUS BOREAL LANDSCAPES

Abstract

Fundamental controls on catchment scale hydrologic response to climate can be conceptualised using reduced-complexity numerical modelling and low parameterisation by linking dominant hydrological processes and storage-runoff dynamics to landscape units. Hydrologic response areas (HRAs) and hydrologic units (HUs) were employed in this study to evaluate how surficial geology and land cover interact with climate to govern the patterns of lake and groundwater levels and dominant flows across 26 diverse lake-peatland-forestland catchments within the Boreal Plains (BP) of Canada.

Here we show that our reduced-complexity model provides sufficient functionality to accurately represent interannual (20-year) and seasonal patterns of lake, peatland and forestland water tables, lateral flows and atmospheric fluxes. Our results demonstrate that groundwater connectivity significantly controls soil water dynamics as a function of surficial geology transmission properties. Surficial geology interacts with spatial variability in actual evapotranspiration (AET), which is driven by hydroecological feedback mechanisms and landcover proportions. AET is a crucial driver of hydrological connectivity and the water source-sink dependencies within the landscape; peatland regulation of AET during drought conditions generated higher landscape runoff than lakes and forestlands, and enabled runoff generation every year, compared to multidecadal runoff return periods in forests. Our results indicate that the model utilises sufficient complexity to accurately simulate system dynamics across diverse hydrogeologic settings, providing versatility to transpose models to diverse catchment types. We anticipate that our model will provide a valuable tool for researchers and land managers through increased predictive power to evaluate spatial variability of hydrological sensitivity to climate and project exceedance of critical thresholds for ecosystem functioning under future climates and reclaimed landscapes.

1. Introduction

The Boreal Plains (BP) of Canada represent a diverse mosaic of shallow lakes, peatlands, and forested upland habitats. These landscapes are of ecological significance, hosting the world's richest bird communities (Morissette *et al.,* 2018; Macdonald *et al.,* 2006) in addition to having a critical role in regulating global water and carbon cycles (Price *et al.,* 2013); and boreal forests accounting for 21-30% of global terrestrial carbon (Ireson *et al.,* 2015). Despite minimal ecological change over the past 6000 years (Chapin *et al.,* 2010), unprecedented climatic warming, coupled with catchment disturbances of harvesting, mining and wildfire, are modifying all components of the hydrological cycle (Laudon *et al.,* 2017) and subsequently risks catastrophic shift in ecosystem function (Wang *et al.,* 2014). Maximum ecological sensitivity is anticipated in the 21st century (Ireson *et al.,* 2015), owing to strong coupling of climate with ecological and biogeochemical and hydrological processes (Waddington *et al.,* 2009). Consequently, it is imperative to fully characterise the dominant processes controlling water movement for the ecohydrological management and restoration of these ecosystems (Thompson *et al.,* 2015).

The BP's climate operates with a long-term water deficit, with frequent periods of water deficits (Devito, Mendoza, and Qualizza, 2012) interspersed with wet years occurring on a 10–15year cycle (Ferone and Devito, 2004). Subsequently, the BP has evolved within a delicate hydrologic balance between precipitation and evapotranspiration (Brown *et al.,* 2014), where water storage provides a crucial self-sustaining mechanism that facilitates both substantial water transmission in wet periods and water retention and landscape scale redistribution. Thus, storage capacities underline hydrological connectivity, which is crucially important for ecosystem health (Wells, Ketcheson, and Price, 2017). However the dominant sources of flow and their controls is spatially complex due to the hydrogeo-climatic setting of the BP. Contrasting evaporative demand and hydrologic function of lakes, peatlands and forestlands (Devito, Creed, and Fraser, 2005) interacts with deep heterogeneous surficial deposits

of contrasting storage and transmission properties (Devito *et al.,* 2012). Subsequently, landscape-scale wetting and drying cycles occur at contrasting rates, magnitudes and amplitudes in response to seasonal and decadal climate signals (see Chapter 2 and Chapter 4).

This spatiotemporal complexity has been encompassed within a conceptual framework that compartmentalizes the surficial geologies into *Hydrological Response Areas* (HRAs), and surface landcovers of lakes, peatlands and forestlands into *Hydrological Units* (HUs) derived from conceptual syntheses of BP (Devito *et al.,* 2012). Empirical research testing the interactions of HRAs, HUs and landscape position have demonstrated spatially variable dominant sources of flow and controls on lake water levels (Chapter 3) and water tables (Hokanson, Mendoza, and Devito, 2019). Larger scale mineral groundwater dominates coarse outwash lake water tables and dampens the magnitude of the response to climate cycles (Smerdon *et al.,* 2007). Whereas localised and intermittent organic groundwater is the dominant source of flow in peatland dominated fine plains system (Chapter 3; Plach *et al.,* 2016), while deeper soil storage and higher forest productivity in fine hummocky catchments produces lake water level patterns that strongly mimic atmospheric exchange due to high sensitivity to evaporative demand (Devito *et al.,* 2017). Further interactions with landscape position and lake geometry and their influence on surface connectivity further modify hydrologic patterns in these systems. Moreover, low runoff responses within the BP may obscure interpretation of the overall impacts of disturbance and external forcings, necessitating exploration of storage dynamics to assess or project impacts of aspen harvesting (Carrera-Hernández *et al.,* 2011), wildfire (Kettridge *et al.,* 2015), road construction and oil and gas exploration (Strack *et al.,* 2019), climate change (Price *et al.,* 2013), and to predict the success of reclaimed oil mines (Elshorbagy *et al.,* 2005; Elshorbagy, Jutla, and Kells 2007; Keshta, Elshorbagy, and Carey, 2009).

Translating site-specific studies to landscape scales has long been recognized as a major research challenge (Price *et al.,* 2005), and until recently, there has been little focus on the role of interaction

between peatlands and forestlands and with the atmosphere, surface, and subsurface processes at these wider scales (Hwang *et al.,* 2018; Devito *et al.,* 2017). Data collection and empirical studies of sufficient length to observe both seasonal and decadal climate variability, in a setting where storage antecedent moisture conditions is so crucial for hydrological functioning (Wells, Ketcheson, and Price, 2017) has presented a significant barrier to this understanding (Chapter 3). Consequently, much of the previous BP hydrological research has been conducted at fine scale or focused on individual hydrological processes. Studies that have addressed hydrological processes holistically (e.g. water balances, system dynamics), often explore these dynamics within individual catchments and demand a high level of calibration and parameterization to model numerically (Thompson *et al.,* 2015), thus limiting transferability to other catchments. Reduced complexity models are an essential tool to facilitate model transferability. One-dimensional property-specific models as opposed to site specific models have enabled models to be transposed to different geologies and soil types (Carrera-Hernández, Smerdon, and Mendoza, 2012). However, this method approaches hydrologic units in isolation and therefore lacks consideration of dynamic connectivity, the importance of which has been so crucially attested (Spence, 2010). However, the increasing ability to quantify storage dynamics, the recognition of threshold driven connectivity (Chapter 4; Ali *et al.,* 2015), and that storage drives watershed resilience (Sayama *et al.,* 2011) has seen the recent rise in storage theories, which lend well to low parameterisation requirements and reduced complexity modelling. This insight and data availability coupled with storage-water balance models can be used to simulate long-term water-level dynamics (Price *et al.,* 2005). Subsequently, this framework aids in accurate processes representation in quantitative models (Devito *et al.,* 2017), and informs the hierarchy of hydrological controls (Devito *et al.,* 2005), within a hydrologic context-composition-connectivity structure.

This work aims to simulate hydrological function of multiple HRA-HU configurations and associated storage-connectivity dynamics within a versatile, reduced-complexity system dynamics numerical model. This study addresses the challenge of modelling heterogeneous environments with lower

parameterisation requirements and thus tests the plausibility of using reduced complexity modelling within such landscapes with the following objectives:

1. Construct a reduced complexity model that replicates the dominant fluxes and landscape interactions as outlines within conceptual understanding and empirical evidence of dominant hydrological processes within different HRA-HU configurations.
2. Evaluate model performance in simulating water table patterns of diverse HRA-HU landscape configurations and identify lake catchment types poorly represented by reduced complexity modelling.
3. Assess spatial differences in landscape runoff sensitivity to the long-term climate.

The HRA and HU conceptual framework is used as the framework for the model design within this research, in line with findings that land cover, landscape position and surficial geology govern the patterns of lake water levels and dominant flows within the BP (Chapter 2, 3 and 4). The simplicity of the model design within this framework enables model adaptability of application to a wide range of landscape configurations, by exploring the fluxes within and between ecological HUs (i.e., lakes, peatlands, forests) of varying composition, within the context of different geological HRAs (i.e., coarse sand, fine hummocky moraine, fine clay plains). This paper presents the evaluation of a lake-peatland-forestland model for 25 lake catchments with one calibrated catchment, and sensitivity analyses of varying HRA, soil depth and type, and boundary conditions with long term climate data (1920-2018).

2. Study Area

The Utikuma Region Study Area (URSA; 56°N, 115°W), situated 370 km north of Edmonton, Alberta, Canada within the BP ecozone, is characterised by a sub humid climate with a pronounced climate cycle, with wet years occurring infrequently on a 10–15 year cycle (Ferone and Devito, 2004).

URSA overlies deep (45-240 m), low relief Quaternary glacial deposits (Hokanson, Mendoza, and Devito, 2019), and the Smoky Group shale unit of the Upper Cretaceous period (Vogwill, 1978; Ceroici, 1979; Plach *et al.,* 2016). The surficial geology varies spatially and can be simply categorised into three primary classes (Devito *et al.,* 2017): (1) coarse-textured glacial–fluvial outwash (CO); (2) fine-textured clay rich glacio-lacustrine plains (FP); (3) fine-textured hummocky disintegration moraine deposits (FH) and two transitional classes: (4) the transition from hummocky disintegration moraine to glacial fluvial outwash (CO/FH); (5) the transition from disintegration moraine to glacio-lacustrine plains (FP/FH). URSA exhibits very low topographic relief (<80 m), characteristic of the BP. On the surface, the landscape is characterised by a patchwork of forestlands and peatlands that vary in their dominance and size through the landscape. Forestlands are characterised by low productivity pine, predominantly situated upon poorly well-draining sandy regions, and a combination of aspen and spruce forests (white and black spruce), present within finer grained mineral soils. In both fine and course texture regions, peatlands are characterised by the accumulation or organic soils, and a forest canopy of black spruce and tamarack. Shallow lakes are ubiquitous across URSA and the BP, with open water accounting for 12.6% of the URSA and wetlands covering an additional 46.3%.

This study models 26 study lakes along a 50km transect that spans the URSA, strategically encompassing lakes that were located within different HRAs, with lake catchments that have a range of different peatland land cover proportions (low and high), and located within catchment at different landscape positions (low and high elevation). See Chapter 3 for further site descriptions and the appendix for maps of each catchment.

3. Methods

3.1. Field Data

Daily precipitation (mm) and mean daily air temperature (°C) was collected from 1998 to 2017 in the URSA, Alberta. Daily precipitation (mm) and mean daily temperature (°C) were obtained for Fort

McMurray, Alberta (https://climate.weather.gc.ca/historical_data/search_historic_ data_e.html; Station Fort McMurray CS, Climate ID 3062696), between 1920 and 1998 (as in Carrera-Hernández *et al.,* 2011). Potential evapotranspiration (PET) was then estimated using Hamon's method (Hamon, 1963) from day lengths and daily mean temperatures.

Lake water levels were monitored across all 26 study lakes using staff gauges installed in 1998. Lake water levels were manually collected each summer across all sites, with higher frequency manual sampling and/or autologged levels by pressure transducers installed in staff gauges in key study lakes. Peatland and forestland groundwater levels and pressure heads were also monitored with a diverse range of manual piezometer, dipwells and autologgers at varying instrument depths (see appendix). A time series of peatland and forestland water table depths were calculated for the calibration catchment (lake 43) and a further four lakes for evaluation across different HRAs (lake 16, lake 208, lake 40, lake 171).

In addition to water tables, daily lake outflow data for lake 43 were recorded for the period 2000-2010 (Thompson *et al.,* 2015). We also draw on reported values of runoff (Redding and Devito, 2010) and AET from conceptual understanding derived from empirical evidence and water balances developed across the study area (Devito *et al.,* 2012). Lake outflow is reported relative to catchment area; peatland and forestland runoff is reported relative to their respective HU land surface area.

3.2. Numerical Model Framework

A two-dimensional (2D) model was developed to simulate hydrological processes within and Darcian driven water flow and connectivity between HUs. For this study, a generic lake-peatland-forest complex has been developed, so that it can be easily modified to represent different geologies and catchment compositions. With this versatility in mind, the purpose of the 2D model is to simulate the hydrologic interactions occurring between the different landscape units, to explore threshold

mediated dynamic connectivity and in turn to assess landscape scale hydrological functioning of different configurations of HRAs and HUs based on their sensitivity to climatic forcing and modification of specified parameters and boundary conditions. Therefore, it is the system behaviour and interactions of these fundamental hydrologic building blocks, and not the exact or point scale spatio-temporal variation of hydraulic heads, that this research sets out to achieve. In order to accomplish this, HUs are simplified 'buckets' to define fill-spill mechanisms as a function of the interaction of climate, parameterised storage and physical process driven flows, but not represent higher resolution heterogeneities in water table elevations. This work is undertaken on the impetus that simplified models, with 'just enough' functionality, allow for better model interrogation through uncertainty and sensitivity analysis, subsequently enabling better understanding of system behaviour and sensitivities (Brunner and Simmons, 2012; Doble and Crosbie, 2017).

Numerical simulations were completed within the system dynamics simulation environment STELLA Architect (High Performance Systems Inc., 2001), consisting of stocks and flows to represent the accumulation and flux of water driven by physical processes and empirical relations. Figure 5-1 graphically represents one example of the configuration of the model lake-peatland-forestland hydrologic units, demonstrating how the combined elevations, areas, connectivity lengths and node distances to define the 3D structure of buckets produces a simple but realistic representation of hydraulic gradients and hydrologic connectivity between HUs in addition to landcover interaction with the atmosphere. Figure 5-2 displays the structure of the model with physical water stores conceptualised as buckets for each HU. The model can function as two states, a saturated system whereby the groundwater buckets are "full", and a perched system with more than one water table. All flows, storage components and water table calculations are fully defined within the supplementary material, where equations include physically based relationships, conditional statements and threshold mechanisms for alternative flows to switch conditionally based on the state the system.

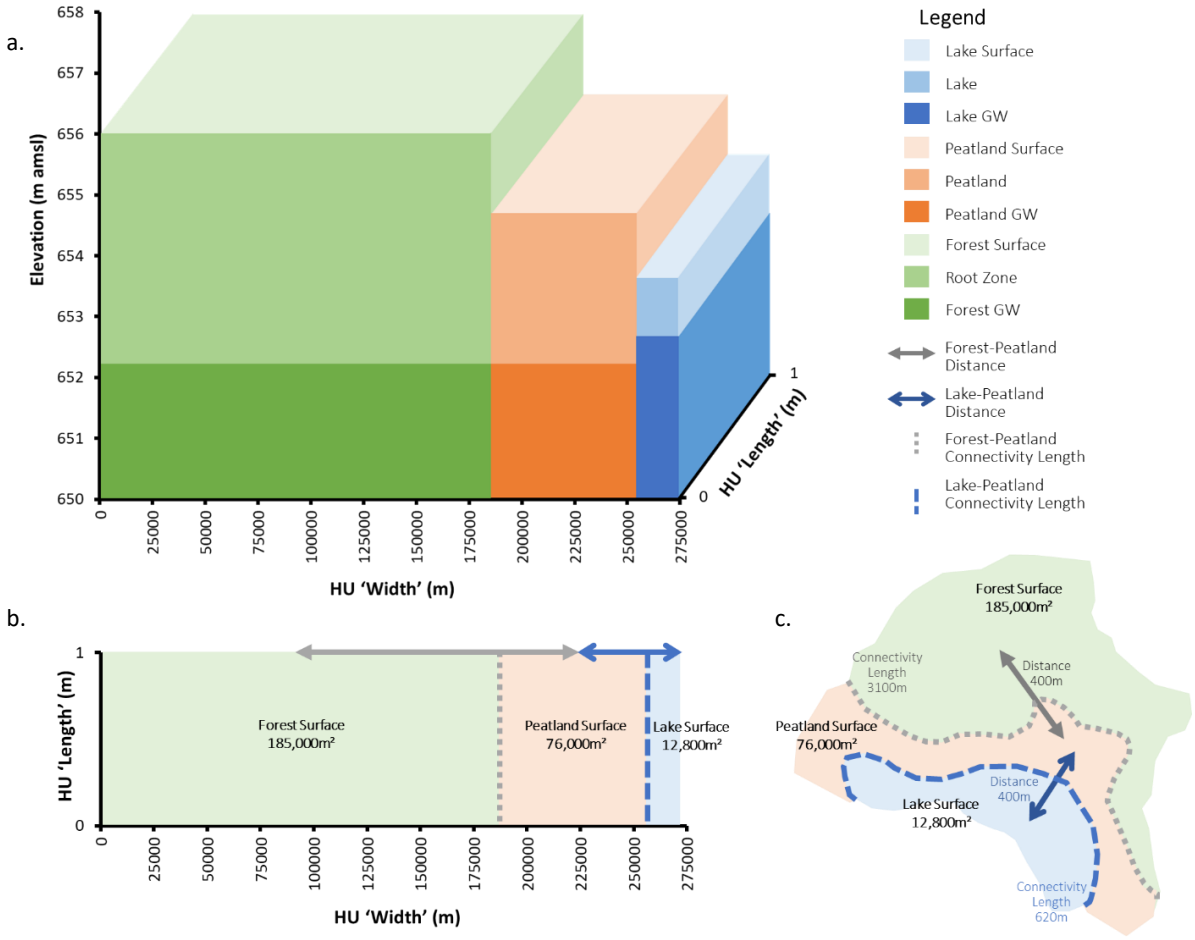


Figure 5-1. Model configuration of hydrologic units (HUs) as developed within STELLA Architect (approximate Lake 43 example values), represented as: a) simplified 3D structure of bucket elevations; b) simplified plan view/ surface dimensions, connectivity lengths and node distances; c) plan view of surface areas, connectivity lengths and node distances.

3.3. Lateral Flows

Lake outflow occurs via an outflow channel as a function of water height above the outflow channel. Throughflow between lake and peatland HUs will occur under saturated conditions and is therefore governed by Darcy's law (1856), allowing for flow reversals depending on the location of the higher hydraulic head. Throughflow between the peatland and forest root zone is governed by Richard's equation as the root zone is unlikely to reach saturated conditions, as the forestland exhibits similar vertical and horizontal K values and therefore favours vertical over lateral flow due to large soil water storage. Consequently, lateral flow via this function is unlikely, however threshold driven rapid lateral

flow from the forest floor active layer may occur when rainfall events or snowmelt events greater than 25mm occur in one day due to infiltration excess (Redding and Devito, 2008). Groundwater flow between HUs operates as a biflow governed by Darcy's law, allowing for flow reversals.

3.4. Vertical Fluxes

The lake and peatland HUs are represented by two buckets; a surface/shallow subsurface store and a deep groundwater store. The forest HU is also represented by 2 buckets, a shallow subsurface root zone store, and a deep groundwater store; the root zone store includes a forest floor 'active layer' with different soil hydraulic properties to determine infiltration excess runoff. Infiltration rates typically exceed rainfall intensities within the western boreal plains, therefore the entire available water store infiltrates into the peatland store/ forest root zone store, with the exception of very high intensity rainfall events (>25 mm day^{-1}) under wet antecedent conditions within the forest.

Within saturated systems, groundwater recharge is governed by Darcy's law. In perched systems, groundwater recharge is defined by Richard's equation, using the Brooks and Corey (1964) relation to relate hydraulic pressure head to θ and K_{sat} (Cuenca, Stangel, and Kelly, 1997; Brunner and Simmons, 2012; Doble and Crosbie, 2017), in line with existing numerical models, such as HydroGeoSphere (Therrien *et al.,* 2010), HYDRUS (Šimůnek *et al.,* 2003) and HYDRUS-based MODFLOW (Twarakavi, Šimůnek, and Seo, 2008). Percolation from the forest root zone is typically macropore flow and occurs when soil moisture thresholds are exceeded (Lukenbach *et al.,* 2019). Uptake from the groundwater is calculated with the Richards (1931) equation, and the Brooks and Corey (1964) relation.

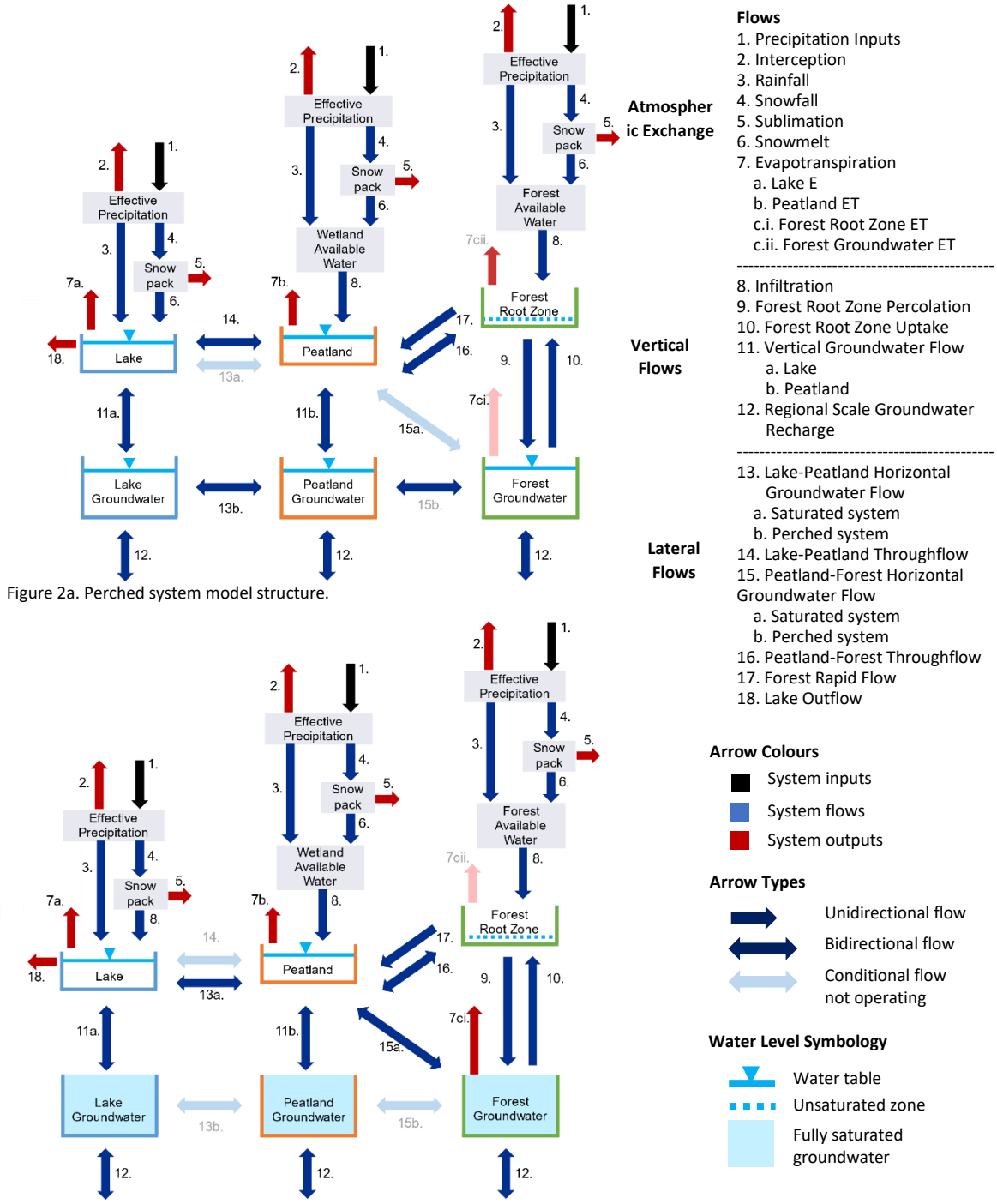


Figure 2a. Perched system model structure.

Figure 2b. Saturated system model structure.

Figure 5-2. Model structure as seen within STELLA Architect. The model utilises flows conditionally to alternate between a) perched systems with multiple water tables and b) saturated groundwater systems (figure 1b). and Arrows indicate flows, buckets indicate water stores, and grey boxes are hypothetical stores for water inputs. Diagram flow numbers correspond to equation numbers.

3.5. Atmospheric Fluxes

Precipitation inputs are translated into rainfall and snowfall as a function of temperature. Temperature controls when precipitation becomes available water due to storage as snow and ice, which is crucial to simulate significant spring melt inputs. Snowmelt is then determined using a degree day model. Interception varies seasonally with canopy growth, however, Thompson *et al.,* (2015) found measured interception to be similar in aspen and black spruce areas, therefore the same seasonal effect on interception was assumed for both the black spruce dominated wetlands, and aspen dominated forestlands. It is assumed that all intercepted water evaporates and is lost.

Actual Evaporation/ Evapotranspiration (AET) is calculated from Potential Evapotranspiration (PET), which in turn is estimated from Hamon's (1963) method as in Thompson *et al.,* (2015), who show Hamon's calculated values to be similar to measured lake ET. Hamon's equation (Hamon, 1963) is a simple relation between mean air temperature and day length to give PET. Within this study, seasonal ET in the lake and peatland HUs is restricted to ice free seasons, which is currently from 1st April to 31st October as estimated by air temperatures. Seasonal ET in forest HUs is constrained to the growing season, which is determined as 15th May to 15th September at present, however, these periods may extend with climate change. For the remainder of the year, evaporation equals 0. Each HU exhibits distinct soil and vegetation characteristics with different levels of evaporation rates, therefore PET is multiplied by α, which is a constant value of 0.8 for black spruce peatlands (Petrone, Silins, and Devito, 2007; Brown *et al.,* 2010), 1.2 for lakes and 1.3 for aspen forestlands (Brown *et al.,* 2014) within the URSA. PET then translates into AET as a function of water table depth (WTD) or plant available water. The WTD function enables the unrestricted peatland evaporation when the water table is at the surface and prevents ET beyond a depth of 1m for peatlands, and at 3.5m from the forest groundwater. Within the forest root zone, transpiration is also constrained under dry conditions, in line with previous research by Thompson *et al.,* (2015) whereby AET occurred at maximum levels under saturated

conditions and decreased towards zero ET at wilting point. Small quantities of forest ET can also occur via the aspen-root-pipeline, whereby roots and higher soil water tensions within the forest root zone draw water from the peatland.

3.6. Material Properties

Saturation curves for hydraulic conductivity (K) and soil water tension (ψ) of different soil types of sand, sandy loam, sandy clay loam and clay, are derived from tabular values from the RETC code for quantifying the hydraulic functions of unsaturated soils (van Genuchten, Leij, and Yates, 1991). Within RETC, soil hydraulic data are defined by the equations of Brooks and Corey (1964) and van Genuchten (1980), with Burdine (1953) and Mualem (1976) defining soil pore-size distributions. Values of horizontal and vertical K for peat, and gyttja were taken from Thompson *et al.,* (2015) and from Hokanson *et al.,* (2019) for HRA saturated K (see supplementary material).

3.7. Model Development and Evaluation

The model was developed for one lake catchment (Lake 43) and calibrated using 20 years of lake level observations (1998-2018), 18 years for peatland and forestland water table observations (2000-2018) and 10 years of lake outflow (2000-2010). The key output variables of interest for calibration targeted physically based empirical relationships of conceptual understanding derived within the BP to accurately represent the balance between evaporation loss and runoff generation. Peatland water table-evapotranspiration feedbacks were developed based on empirical evidence of reduced AET with greater water table depths (Waddington *et al.,* 2015), but higher AET when water tables exceeded the ground surface to simulate open water ET. This required calibration of peat properties and depths within the peatland to accurately represent the water table-ET feedback mechanism to produce realistic peatland AET and peatland runoff, where values of annual runoff from peatland and forestland

HUs and annual AET were also compared against values reported within the study area (Devito *et al.,* 2012). Similarly, soil properties and rooting zone depths were refined within the forestland (Lukenbach *et al.,* 2019) to determine the balance between evaporation loss and runoff generation or groundwater recharge.

The model was then evaluated by applying the model to a further 25 lakes. Each catchment is assumed to receive the same precipitation and temperatures and are therefore simulated with the same climate data sets. It is acknowledged that convection storms common within the region could cause spatial variations in annual precipitation, further to microclimatic variability in temperature, which may lead to more minor discrepancies between observed and simulated water tables. Only parameters reflecting landscape proportions, lake geometry and topography were modified for each lake (see supplementary material). Surficial geology hydraulic conductivity and specific yield were also modified for coarse textured catchments to a fixed value.

Model performance was evaluated for lake water depths in all 26 simulated lakes against observed lake depths 1999-2018 using time series and scatter plots, and performance statistics including Pearson's correlation, root mean square error (RMSE), normalised root mean square error (NRMSE) and mean absolute error (MAE). The same approach was applied to peatland and forestland water tables in four lakes of varying surficial geology and land cover configurations (FP lake 171; FH lake 40; CO lakes 16 and 208). Assessment of water tables in peatland and forestland HUs evaluates the internal functioning of the model and ascertains whether lake water levels are accurately represented by simulating interaction with adjacent HUs. To further explore the internal functioning of the model, mean annual fluxes were plotted for each lake (catchment ET, precipitation, lake outflow, peatland runoff and forest runoff) and assessed against conceptual understanding of dominant hydrological processes derived from empirical evidence (Chapter 3; Devito *et al.,* 2017; Plach *et al.,* 2016). Model performance was also evaluated for historical lake water levels dating back to 1951 by utilising

historical aerial imagery of lake surface water extent. Simulated lake water levels were projected onto historic satellite imagery using contours of a digital elevation model and assessed visually.

3.8. Spatial Variation in Hydrologic Response to Climate

Following model evaluation, simulations using climate data from 1920-2000 were performed to undertake an assessment of spatial variability in hydrological responses to the long-term climate. The long-term climate data set is collected from Fort McMurray, outside of the URSA, however, it represents a similar climate, and also reflects the climate of the oil sands region which requires reclamation. The stationarity of lake water level time series is determined to identify any long-term wetting or drying gradients and compared within lake catchment classifications (Chapter 3). Interannual lake water level variability is assessed relative to intra-annual variability and the climatic conditions that facilitate higher intra-annual variability and/or lake catchment types which propagate this through cumulative multi-annual memory are addressed.

HU runoff generation as cumulative distribution functions (CDF) were used to determine runoff return periods and magnitudes and gain understanding of landscape sensitivity to climatic conditions and how this differs spatially between HRAs and HUs.

4. Results

4.1. Climate

Long term climate records of Fort McMurray and the URSA (Figure 5-3) indicate decadal cyclical weather patterns. Wet years (PET:P ratio < 0.9) occur infrequently, returning every ~20years, whereas mesic years (PET:P ratio of 0.9-1.2) and dry years (PET:P ratio >1.2) cycle every ~4 years (defined in Devito *et al.,* 2017; Hokanson *et al.,* 2019). This produces a long-term moisture deficit with periods of

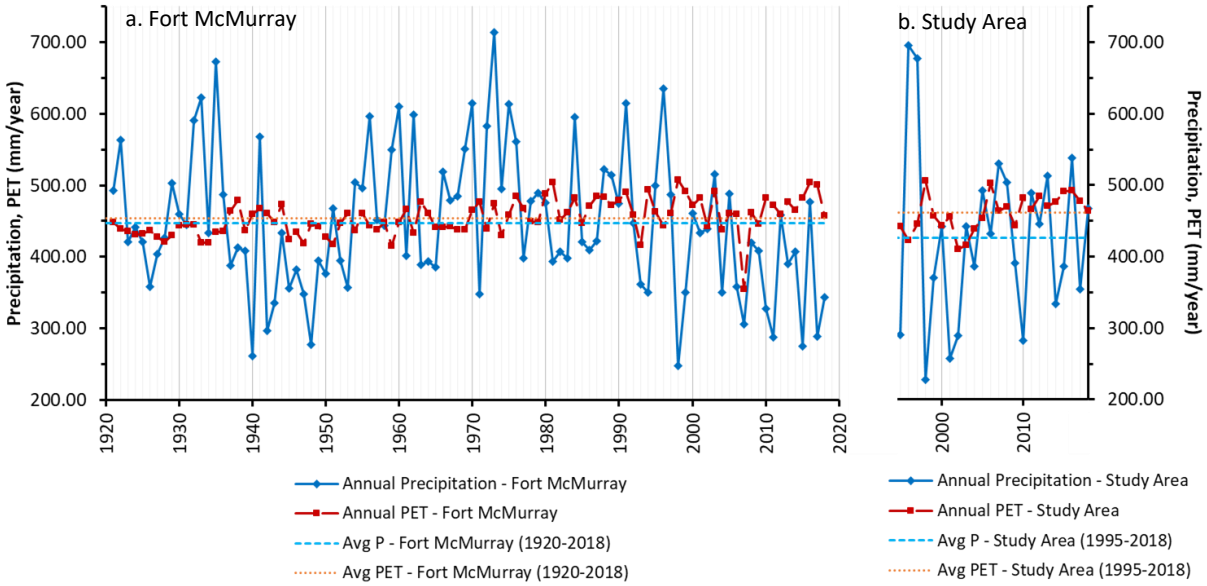


Figure 5-3. Long term climate of the Boreal Plains region of Alberta, Canada: a) Fort McMurray, 1920-2018; b) Study Area, 1995-2018. Annual totals represent the hydrologic year (1st November-31st October). Potential evapotranspiration (PET) is calculated using Hamon's equation.

excess moisture (P-ET). The URSA climate (Figure 5-3a) is comparable with the 100-year record of Fort McMurray (Figure 5-3b); the model calibration-evaluation period captures mesic and dry conditions, with a wet period (1996-1997) a couple years prior to the URSA climate records. Dry years and wet years occur in clusters throughout the climate record, notably, prolonged, consecutive and large deficit years occur throughout the 1940's and to a lesser extent late 1990's and 2010's, whereas prolonged consecutive surplus conditions occur throughout years 1966-1976.

4.2. Model Calibration

The model for lake 43 achieved a reasonable level of accuracy across a number of hydrological outputs, including water tables in all HUs, lake outflow and at both daily and annual timescales (Figure 5-4), signifying the capability for the model to simulate seasonal and long-term interannual trends. Annual atmospheric fluxes are also comparable with values reported within the BP (Thompson *et al.,* 2015). A number of model parameters were predetermined by field characterisation, such as landscape proportions, lake geometry and topography (Chapter 2), or field values reported within the URSA, such

as soil hydraulic properties (Hokanson, Mendoza, and Devito, 2019; Thompson *et al.,* 2015). Subsequently, water tables were reasonably represented prior to refinement of additional parameters. Annual peatland runoff totals (Figure 5-4e) are consistent with valued reported within literature (Devito *et al.,* 2012), with high variation between wet years (e.g., 1997 – 187 mm) and dry years (e.g., 2002 – 0 mm). Annual peatland AET is also representative of values reported within the study area (Brown *et al.,* 2010), with lowest values occurring in dry years (e.g., 2002 – 286 mm) and highest values occurring in wet years (e.g., 2007 – 410 mm), reflecting restricted available water due to deep water tables, and pronounced AET due to standing water.

Deep soils with high hydraulic conductivity in forest uplands contribute to long runoff return periods. Subsequent forestland groundwater dynamics were critical for determining the availability of water for evapotranspiration. Rooting zone depth and soil properties within the forestland HU were defined to ensure high percolation rates and capacities under dry and mesic conditions, but to also provide infiltration excess runoff under high rainfall events (>25mm day^{-1}) in addition to saturation excess under long-term cumulative wet climatic conditions. These processes crucially replicate the long-term runoff return periods reported in Lake 43's catchment (Redding and Devito, 2010), with negligible forest runoff occurring throughout the calibration period (Figure 5-4f). In addition, the rooting zone depth provided a function of creating a hydrologic disconnect to groundwater as a supply for forestland evapotranspiration under low water tables, resulting in water availability limited AET under dry conditions (e.g. 2002 – 291 mm), but enabled aspen trees to evaporate at PET when the groundwater reached the bottom of the rooting zone (e.g. 2013 – 498 mm), which accurately reflects values reported for the catchment forestland and study region forest HUs (Figure 5-5; Devito *et al.,* 2012).

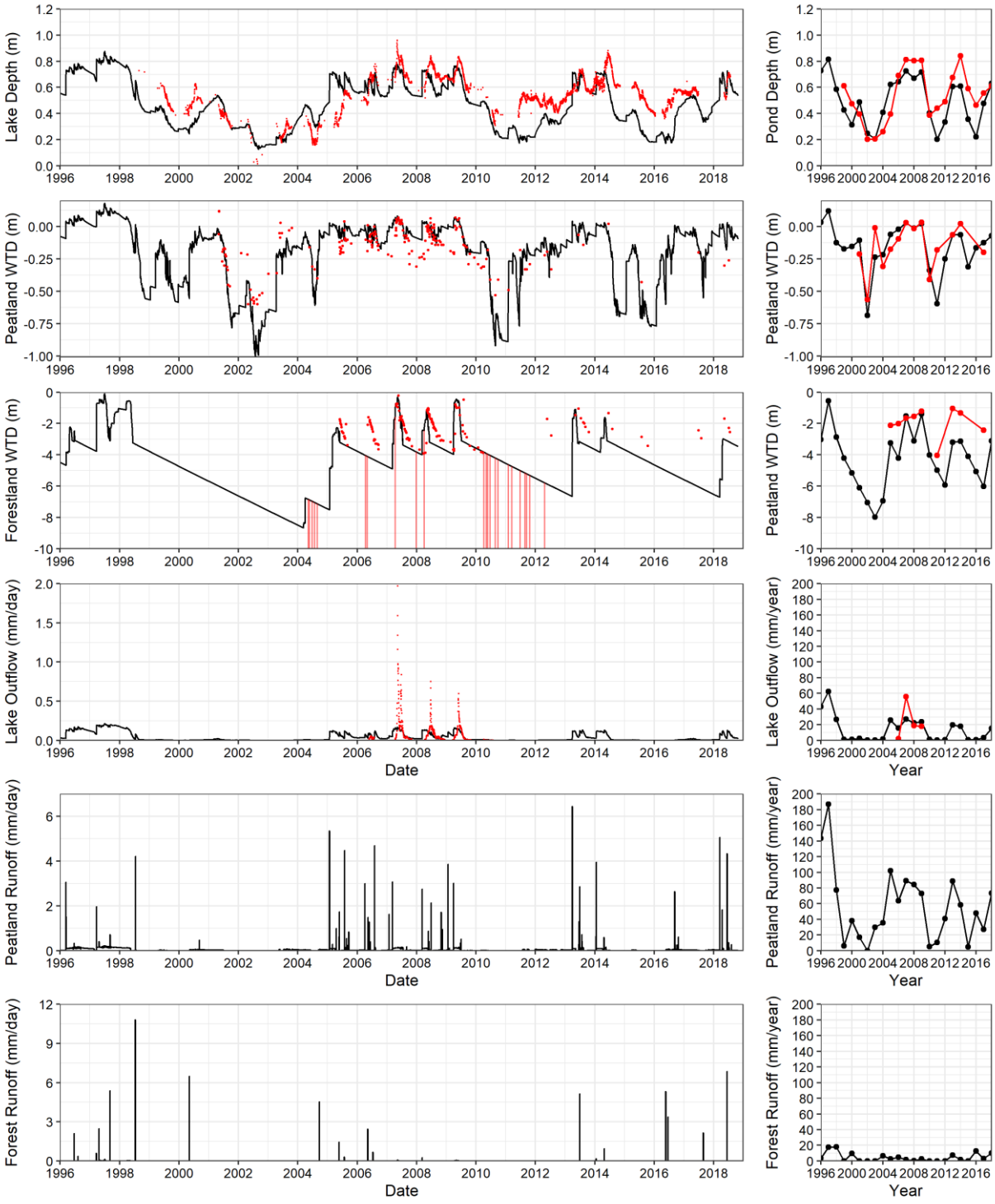


Figure 5-4. Comparison of simulated (black) and observed (red) Lake 43's lake, peatland and forestland daily water tables and mean June water levels, and daily and annual total HU runoff (mm/day). Lake outflow is relative to catchment area, peatland and forest runoff are relative to their HU area. Vertical red lines indicate dry well observations.

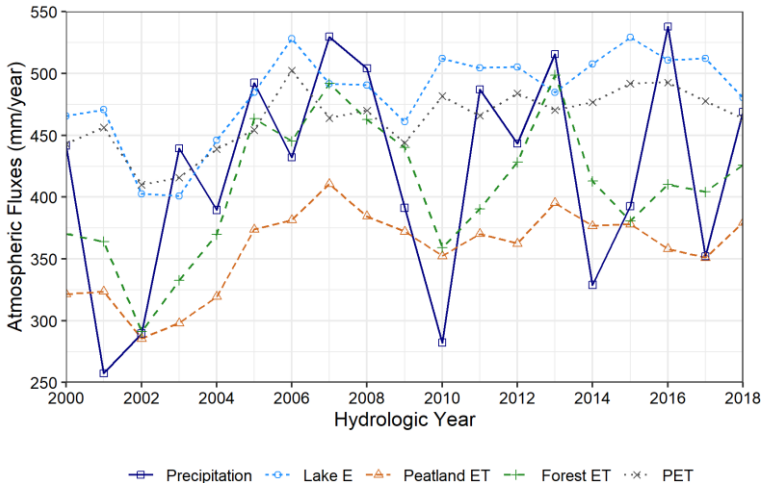


Figure 5-5. Lake 43 annual precipitation, PET and simulated AET of each HU (hydrologic year November 1st – October 31st).

4.3. Model Evaluation

Lake Water Levels

Daily values of simulated lake water depth relative to the lake bottom at the centre of the lake were plotted against observed values for all 26 study lakes (Figure 5-6) and performance statistics reported (Table 1). Simulated lake water depths significantly correlated with observed patterns (1999-2018) in 22 of 26 lakes ($p < 0.001$; p-value < 0.05 in 25 of 26 lakes). Pearson's r coefficient is high across 4 of 5 surficial geology classifications (CO average r = 0.763; CO/FH average r = 0.716; FH average r = 0.731; FP/FH average r = 0.757), however performance is substantially lower across peatland dominated FP class (FP average r = 0.547). Although the performance in the latter is greatly attributed to the model performance of the largest lake and catchment within the study, lake 168 (r = 0.304), which is 36 times the area of the average study lake. NRMSE is also greatest within the FP class, but comparable with other classes with lake 168 removed (FP - 12.8%; CO – 11.4%; CO/FH – 13%; FH – 10%; FP/FH – 12.9%). A comparison of simulated lake water levels with historical lake water levels using surface water extent within aerial imagery combined with digital elevation model contours indicates high model performance for lake 19 (Figure 5-7). The lowest simulated levels in 1951 and highest in 1977 correspond accurately with surface water extent in these years, signifying the capability of the model to simulate a wide range of climatological conditions.

Lake-Peatland-Forestland Water Tables

General temporal patterns of daily water tables are accurately simulated in peatlands and forested uplands of CO lakes 16 ($r = 0.84$, $p < 0.001$; $r = 0.67$, $p < 0.001$ respectively) and lake 208 ($r = 0.62$, $p < 0.001$; $r = 0.62$, $p < 0.001$ respectively). Peatland water tables are also well represented in FP lake 171 ($r = 0.52$, $p<0.001$) and FH lake 40 ($r = 0.58$, $p<0.001$), but correlation between observed and simulated water tables for forested uplands is insignificant. Despite the low correlation between observed and simulated forestland water tables, dry conditions where the water table drops significantly below the surface are captured, indicating that the model is reproducing hillslope hydrologic heads and thus hydraulic gradients accurately over longer periods (100 years; Figure 5-8) at a full range of climatic conditions including a decadal climate signals (wet and dry cycles).

Table 5-1. Performance of simulated versus observed lake water depths (see Figure 5-6).

Lake Class	Lake ID	r	P value	R^2	RMSE (m)	MAE (m)	NRMSE (%)
CO	**1**	0.6653	0.0001	0.4426	0.3437	0.2772	23.3
CO	**5**	0.6120	0.0012	0.3745	0.1433	0.1054	8.3
CO	**16**	0.6812	0.0000	0.4640	0.1470	0.1252	10.9
CO	**17**	0.9112	0.0000	0.8304	0.1014	0.0758	4.2
CO	**201**	0.8406	0.0000	0.7066	0.0623	0.0486	4.9
CO	**206**	0.7214	0.0000	0.5205	0.1801	0.1647	18.6
CO	**208**	0.9071	0.0000	0.8228	0.0924	0.0767	9.4
CO/FH	**7**	0.7318	0.0000	0.5356	0.1444	0.1126	16.4
CO/FH	**11**	0.4841	0.0090	0.2344	0.1672	0.1302	12.1
CO/FH	**15**	0.8108	0.0000	0.6575	0.1859	0.1505	16.5
CO/FH	**19**	0.8382	0.0000	0.7025	0.1056	0.0812	7.0
FH	**27**	0.6536	0.0001	0.4272	0.1034	0.0851	5.1
FH	**39**	0.7977	0.0000	0.6364	0.1006	0.0801	6.1
FH	**40**	0.5377	0.0000	0.2891	0.1220	0.0930	14.3
FH	**42**	0.8165	0.0000	0.6667	0.1596	0.1454	16.5
FH	**43**	0.6616	0.0000	0.4377	0.1563	0.1298	17.6
FH	**48**	0.8440	0.0000	0.7123	0.0566	0.0421	3.9
FH	**111**	0.8084	0.0000	0.6535	0.0820	0.0652	6.6
FP/FH	**59**	0.8343	0.0000	0.6961	0.1606	0.1389	12.1
FP/FH	**112**	0.8828	0.0000	0.7793	0.1885	0.1580	20.7
FP/FH	**118**	0.7545	0.0000	0.5692	0.1020	0.0773	10.8
FP/FH	**122**	0.5565	0.0004	0.3097	0.1029	0.0820	8.0
FP	**121**	0.7206	0.0000	0.5193	0.1154	0.0810	11.5
FP	**168**	0.3039	0.0806	0.0923	0.4918	0.4496	25.7
FP	**171**	0.5552	0.0000	0.3082	0.1285	0.1082	16.5
FP	**205**	0.6096	0.0270	0.3716	0.1139	0.0899	10.5

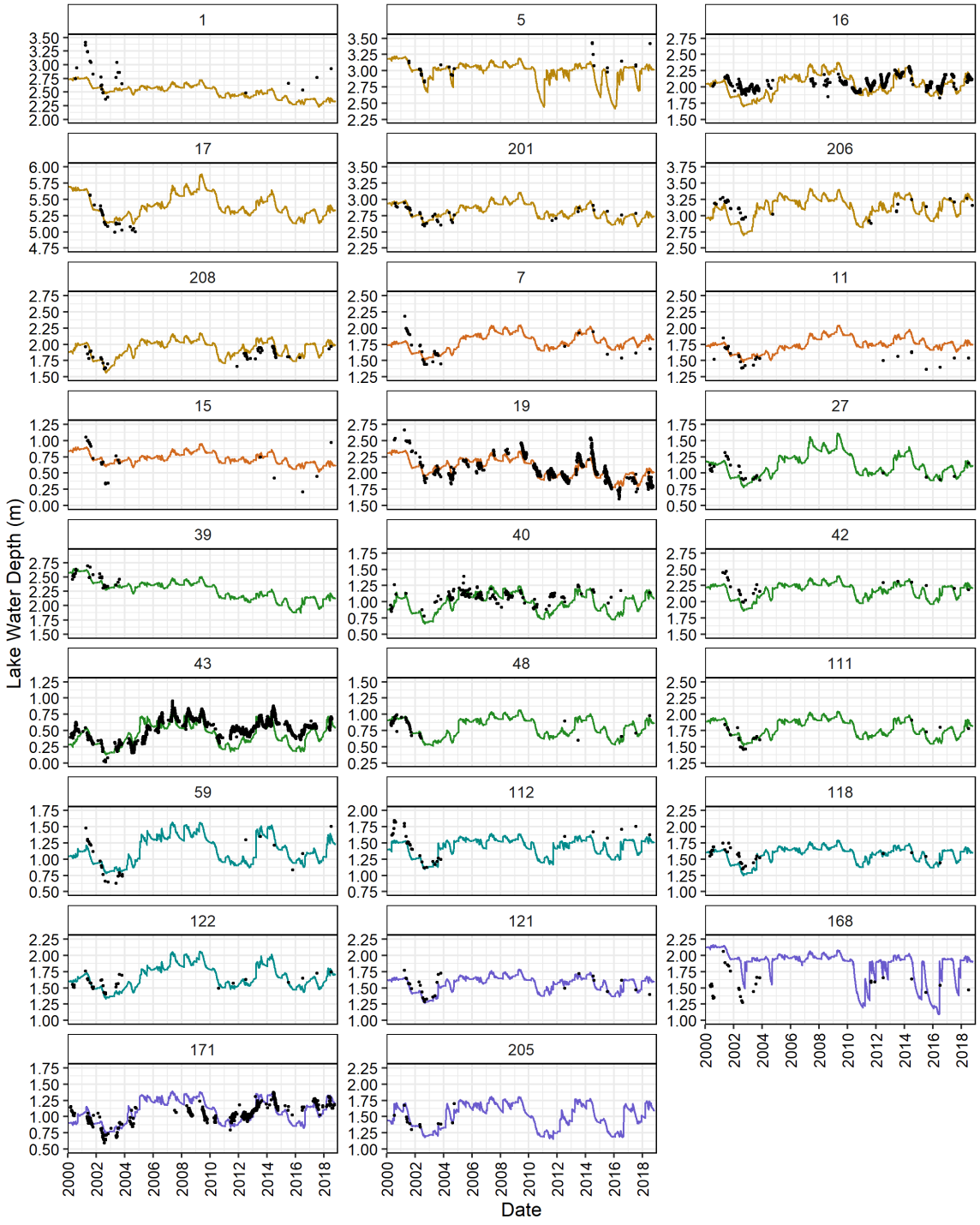


Figure 5-6. Comparison of simulated (coloured line) and observed (black points) daily lake water levels 2000-2018. Model performance is summarised in Table 5-1.

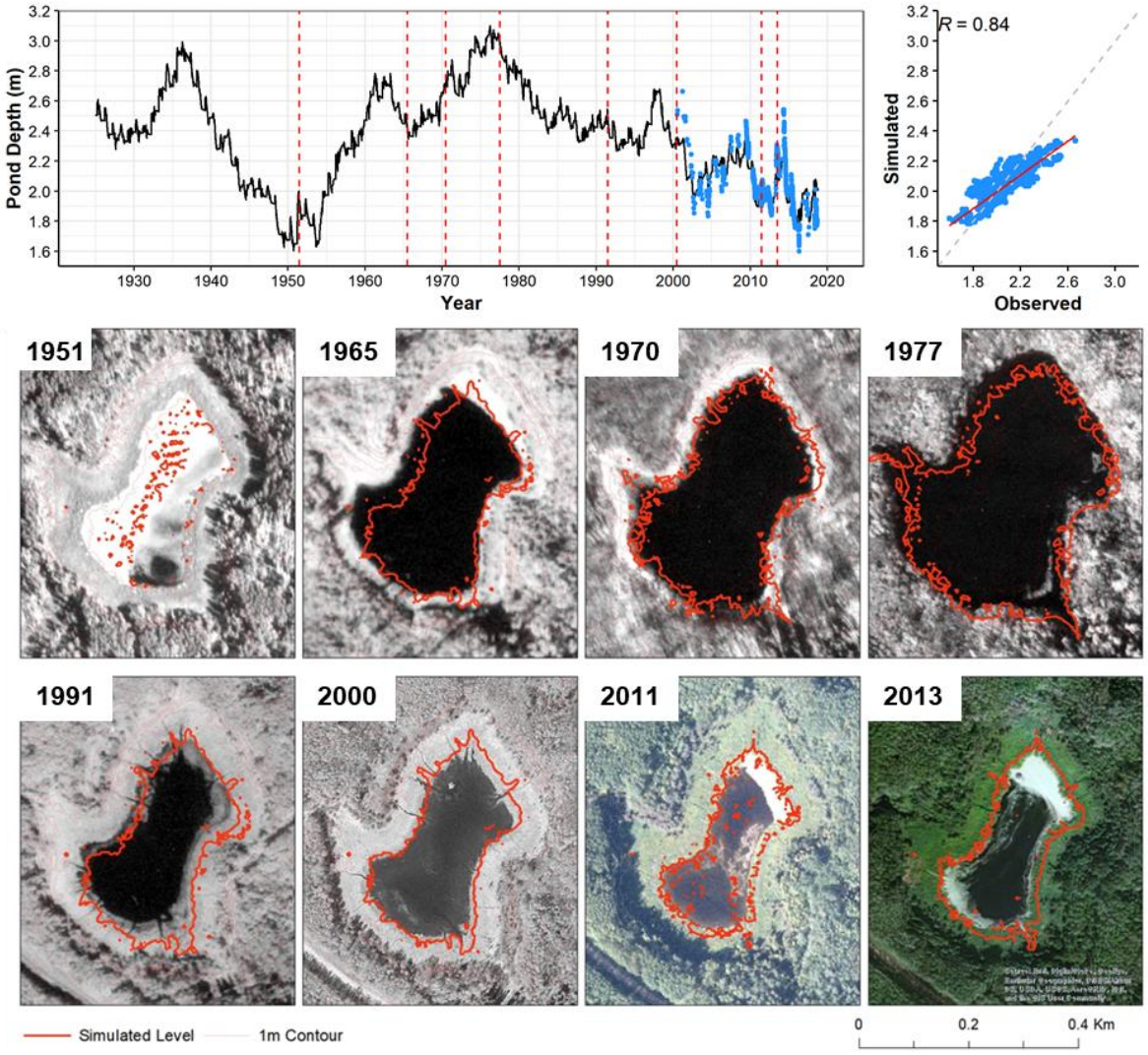


Figure 5-7. Evaluation of model performance for historical water levels at Lake 19 (CO/FH isolated basin). Simulated lake water levels are projected onto aerial imagery using a digital elevation model, visualised with the red contour line.

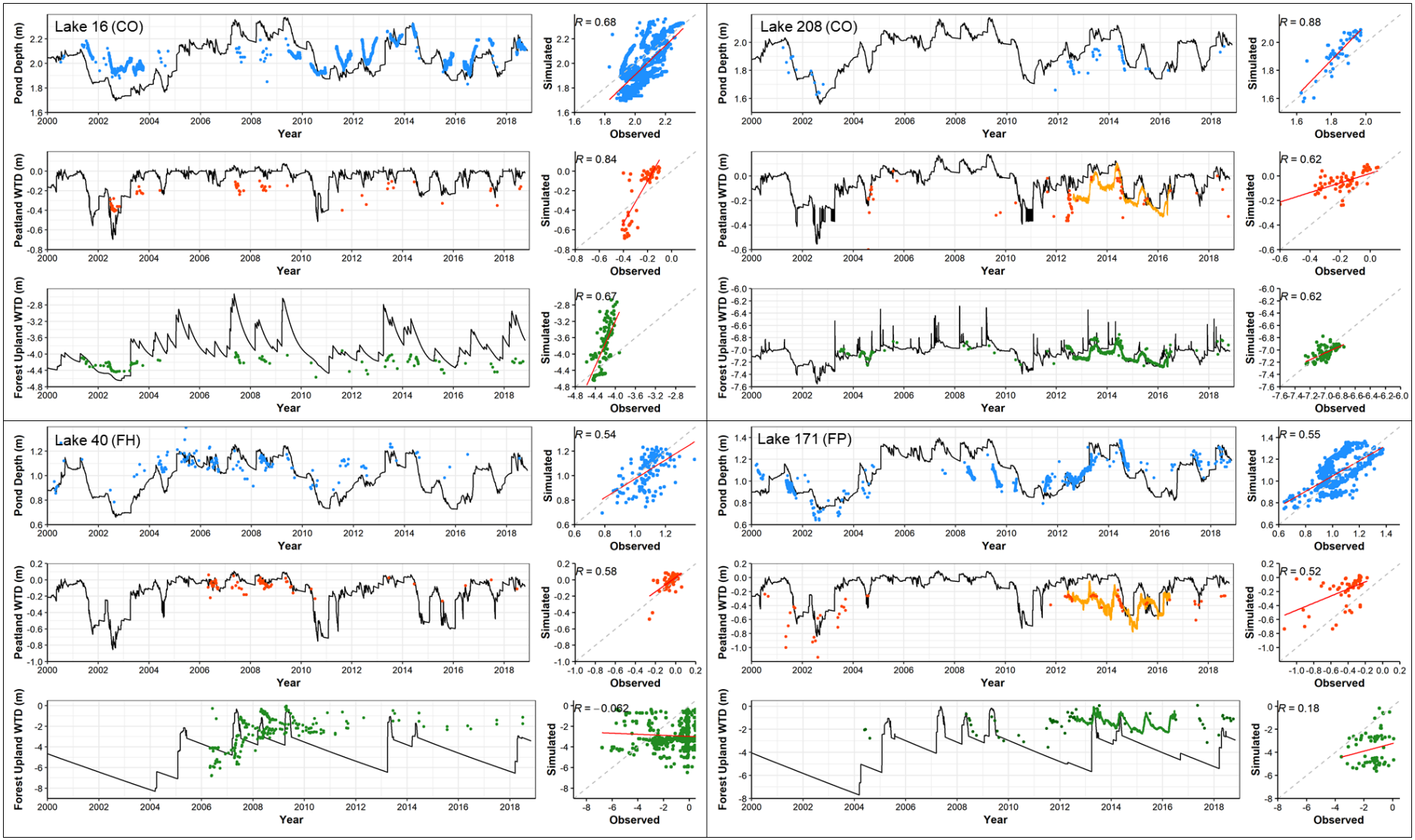


Figure 5-8 – Evaluation of model performance in lake, peatland and forest upland water tables of four lake catchments: lake 171 (FP), lake 40 (FH), lake 16 (CO), lake 208 (CO) presented as a timeseries and a scatter graph of observed versus simulated. Note axis values and ranges differ between all plots.

4.4. Simulated Long term water levels

Daily lake water levels are presented for all 26 lakes (Figure 5-9) and peatland and forestland water tables for four lakes (Figure 5-10) to assess long-term (1925-2018) temporal trends in water levels. All lakes demonstrate long-term stationarity in lake depth (1925-2018: -3.8 to +1.3 mm $year^{-1}$), though a long-term drying trend is apparent 1980-2018 (linear gradient -3.2 to -23 mm $year^{-1}$; see Table 2 for individual lake metrics). There is no clear class trend associated with long term gradients, however lakes with the largest 1980-2018 drying trends are the 5 largest lakes and lakes catchments (lake 168 (FP), lake 1, 5 and 17 (CO) and lake 39 (FH), although FP/FH lake 112 differs from this trend). While larger variability is expected in larger catchments, this could also reflect either poorer model performance of lake 168 and lake 5 or temporal variability in contributing area or threshold change in hydrological response, which is evident from water level patterns during dry periods (1940/50's and 2010's; Figure 5-9).

Interannual lake water level ranges (based on July 1st; Table 5-2) is greater than intra-annual variability in all lakes, signifying the role of interannual memory. There is high spatial variability in interannual range between lakes (coefficient of variation (CV) = 34.6%) in comparison to lower spatial variability in mean intraannual range (CV = 20.3%). Similar intra-annual ranges indicates that lakes with the lowest interannual variability span greater proportions of their range within a year, whereas larger interannual ranges must be driven by cumulative memory. Moreover, years with the highest intra-annual ranges correspond with wetting years, therefore, low interannual variability is associated with rapid rewetting and drought recovery, whereas higher interannual variability is driven by longer memory of dry conditions associated with their lake geometry relative to runoff generating landscape. This is evidenced by the high spatial variability in sensitivity to the 1940/50's dry period; all lakes demonstrate and long-term multiannual drying-wetting trend, however this response is greater in larger or isolated basins with higher magnitude interannual variability.

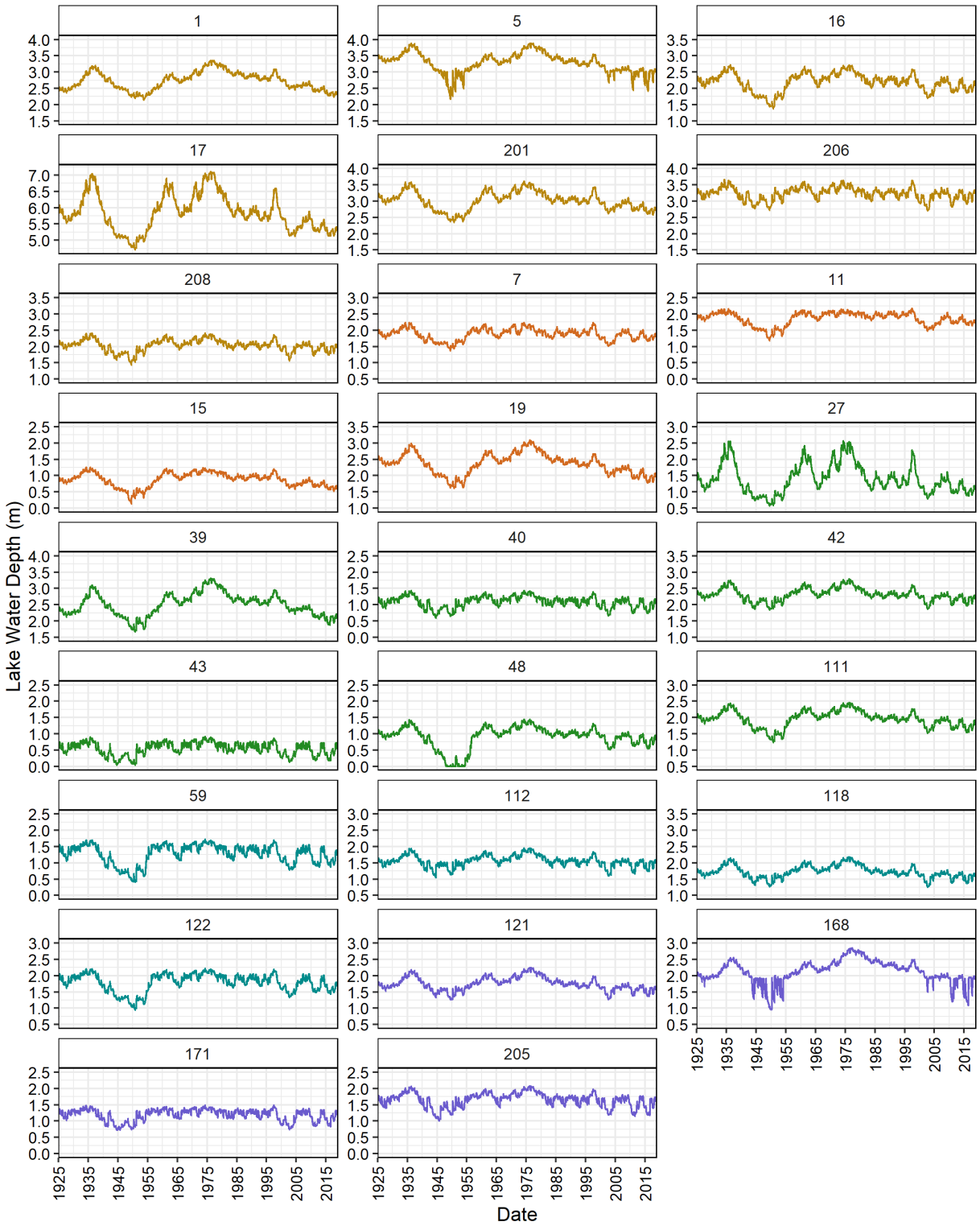


Figure 5-9. Simulated daily lake water depths 1925-2018 across 26 study lakes.

Table 5-2. Lake water level variability and trends.

Class	Pond	Interannual Variability		Stationarity (long term gradient)		Intra-annual Range		
		Interannual Range (m)	Interannual SD (m)	1980-2018 gradient (mm/year)	1925-2018 gradient (mm/year)	Mean (m)	Min (m)	Max (m)
CO	1	1.05	0.27	-19	-0.58	0.14	0.06	0.24
CO	5	1.27	0.28	-17	-3.1	0.20	0.07	0.87
CO	16	1.14	0.26	-6.8	-0.59	0.20	0.09	0.46
CO	17	2.14	0.54	-21	-3.8	0.28	0.08	0.64
CO	201	1.06	0.27	-11	-1.3	0.17	0.07	0.34
CO	206	0.74	0.17	-3.2	-0.87	0.23	0.09	0.46
CO	208	0.73	0.16	-4	-0.57	0.18	0.08	0.50
CO/FH	7	0.67	0.16	-3.5	-0.14	0.16	0.06	0.30
CO/FH	11	0.80	0.19	-6.9	0.29	0.15	0.06	0.37
CO/FH	15	0.93	0.21	-9.8	-0.5	0.14	0.06	0.38
CO/FH	19	1.27	0.31	-17	-2.2	0.17	0.07	0.40
FH	27	1.74	0.43	-8.8	-2.6	0.29	0.08	0.71
FH	39	1.41	0.33	-19	-0.13	0.17	0.07	0.46
FH	40	0.68	0.16	-4.2	-0.66	0.20	0.09	0.48
FH	42	0.77	0.18	-5.5	-1.1	0.17	0.07	0.38
FH	43	0.72	0.18	-4.9	-0.48	0.23	0.09	0.66
FH	48	1.36	0.33	-8.2	1.3	0.18	0.04	0.53
FH	111	1.00	0.24	-9.2	-1.1	0.17	0.08	0.40
FP/FH	59	1.12	0.28	-7.1	0.29	0.24	0.08	0.59
FP/FH	112	0.73	0.16	-4	-1.1	0.18	0.08	0.47
FP/FH	118	0.79	0.18	-6.5	-1.6	0.16	0.07	0.34
FP/FH	122	1.05	0.26	-6.9	0.29	0.21	0.08	0.45
FP	121	0.83	0.19	-8.9	-1.7	0.16	0.07	0.35
FP	168	1.50	0.31	-23	-0.46	0.25	0.08	1.00
FP	171	0.64	0.16	-4.9	-0.46	0.21	0.09	0.51
FP	205	0.88	0.20	-6.6	-1.2	0.23	0.06	0.52

Comparison of peatland and forestland long term water tables demonstrates similar temporal peatland water table patterns across different lake types but contrasting forestland patterns (Figure 5-10). FP lake 171 and FH lake 40 forest water tables exhibit higher magnitude intra-annual variability than CO lakes 16 and 206, and display long-term multiannual groundwater level draw down during dry periods in the 1940's and 2000's.

4.5. Dominant simulated hydrologic fluxes

Average annual simulated hydrologic fluxes of precipitation, catchment ET, lake outflow, peatland runoff and forestland runoff demonstrate spatial variability in dominant flows and their magnitudes (Figure 5-11). The same precipitation and PET model inputs were applied to each lake catchment, however spatial differences in land cover and feedback mechanisms with actual evapotranspiration

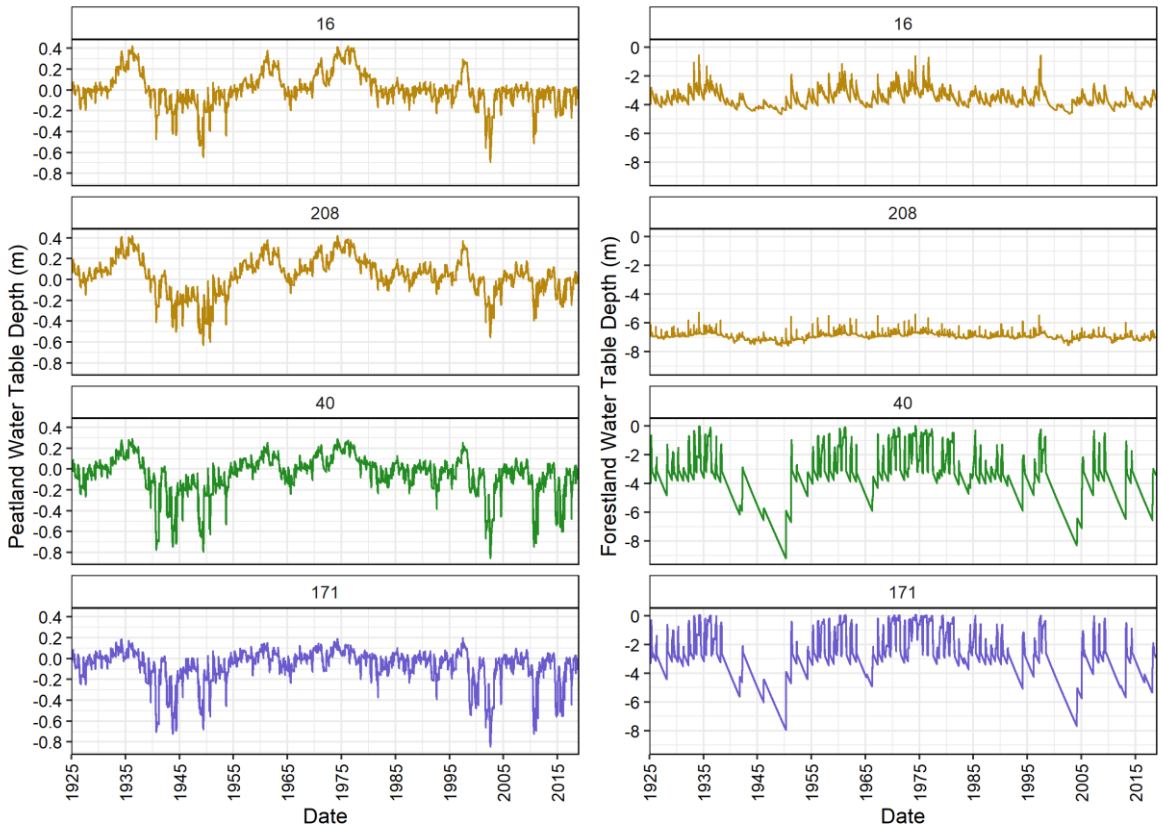


Figure 5-10. Simulated daily peatland and forestland water table depths 1925-2018 across 4 study lakes.

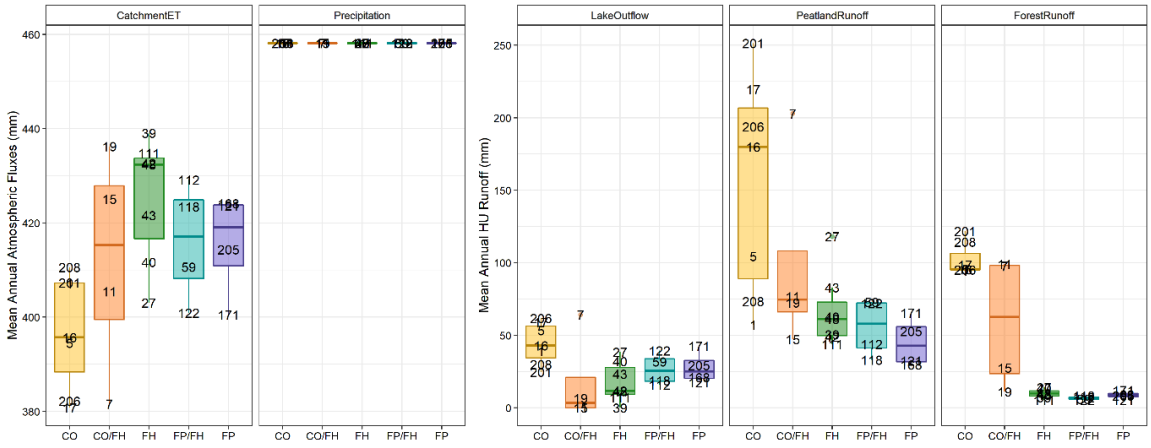


*Figure 5-11. Mean annual simulated fluxes of each lake catchment (mm year^{-1}). Each boxplot represents the lakes within their surficial geology classification and labels signify the lake catchment ID. Lake outflow is expressed relative to the catchment area (i.e. pond outflow volume/pond area * catchment area), peatland runoff is expressed relative to peatland HU area and forest runoff is expressed relative to forest HU area.*

(AET) result in significant differences in AET between surficial geology classifications. CO lakes exhibit on average the lowest AET, whereas forest dominated FH lake catchments display the highest AET. Of the fine textured classes, AET is lower in peatland dominated FP catchments.

CO lakes also exhibit the highest mean annual runoff totals across all HUs. CO/FH display the lowest lake runoff due to isolated basins for the majority of the 94-year simulation. Within fine textured classes, a gradient of low to moderate lake outflow is present from forest dominated FH through to peatland dominated FP. In all fine textured lake catchments, forest runoff is significantly lower than peatland and lake runoff, implying greater peatland cover facilitates higher landscape runoff. In contrast, no significant difference between forest, peatland and lake runoff occurs in coarse textured lake catchments, signifying high forestland contribution in these systems as a product of high mineral groundwater contributions and lower evaporative demand.

4.6. Runoff return periods

Cumulative density functions (CDFs) were derived for each lake and each surficial geology classification for annual HU runoff contributions (1925-2018) to ascertain spatial differences in runoff return periods between HRAs and HUs (Figure 5-12). Similar CDFs were produced within HRAs and are therefore summarised by HRA in Table 5-3. Peatlands generated runoff every year across all HRAs and produced higher average annual runoff than forests and lakes. With the exception of CO, runoff return periods were longer in forest HUs than lake and peatland HUs, occurring every 21 years in FH (flow >0mm $year^{-1}$), and average flows exhibiting return periods greater than 68 years. Subsequently, forest runoff is characterised by infrequent events. Lake runoff is most frequently generated in FP lake catchments. However, CO lakes exhibit the highest magnitude average lake outflows, which reflects the high forest runoff generation in CO. In contrast, CO/FH lakes, are characterised as isolated basins with the absence

of lake outflows within the observed record (1998-2018), however some runoff has been generated under extremely wet conditions (lake 7 and lake 19).

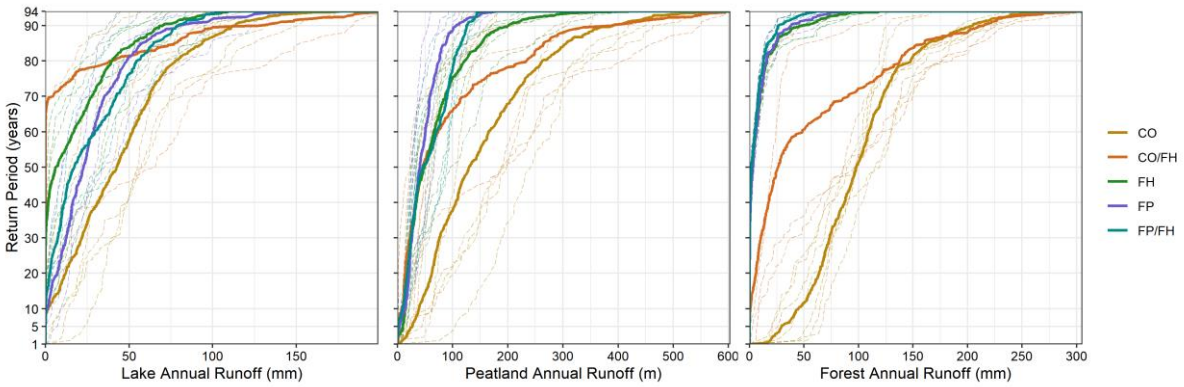


Figure 5-12. Annual simulated runoff totals (mm year^{-1}) of each HU in each lake catchment and across each HRA for 1925-2018, expressed as a cumulative distribution function to determine return periods of flow. Return period summaries of each HRA Class are given in Table 5-3.

Table 5-3. HU annual runoff return period summary based on the cumulative distribution frequency of each HRA (see Figure 5-6). Return periods are expressed in years out of 94 years.

HRA and HU		Return period of flow >0mm year^{-1}	Average flow (mm year^{-1})	Return period of average flow	Maximum flow (mm year^{-1})	Summary
CO	Lake	7	43	51	176	Peatland and forest generating flow every year; high average landscape runoff. Peatland runoff is generated from forest. Flow through system of water generated from forest.
	Peatland	1	151	56	603	
	Forest	1	100	52	281	
CO/FH	Lake	64	17	76	199	Isolated lake basins, no lake outflow
	Peatland	1	95	66	595	
	Forest	4	56	63	305	
FH	Lake	64	17	61	110	Long forest runoff return periods; low average landscape runoff
	Peatland	1	65	59	419	
	Forest	21	9	68	119	
FP/FH	Lake	6	26	58	106	Long forest runoff return periods and short lake runoff return periods
	Peatland	1	55	52	150	
	Forest	19	6	64	63	
FP	Lake	3	27	59	150	Short lake runoff return periods
	Peatland	1	45	55	179	
	Forest	13	8	66	87	

5. Discussion

This study aimed to accurately model long-term lake, peatland and forestland hydrologic interactions and their water tables across heterogeneous environments using a versatile, system dynamics numerical model with lower parameterisation requirement. The model was developed with the scope to provide sufficient functionality by using an appropriate level of complexity, which in turn allows for higher interrogation of model and system behaviour (Doble and Crosbie, 2017). The fundamental importance of landscape connectivity in governing catchment hydrologic response to climate can be conceptualised using low parameterisation by linking dominant hydrological progresses and storage-runoff dynamics to landscape units (Soulsby, Birkel, and Tetzlaff, 2016). The HRA and HU conceptual framework compartmentalises storage within the landscape into units that represent a priori understanding of how surficial geology and land cover govern the patterns of lake and groundwater levels and dominant flows within the BP (Chapter 2 and Chapter 3; Devito *et al.,* 2012). Long-term monitoring coupled with simulated water levels and fluxes across 26 lake catchments of diverse geological settings, topography, landscape cover proportions and lake geometry provides a pioneering test bed for testing and refining the HRA-HU conceptualisation.

5.1. Performance of reduced complexity model across heterogeneous lake catchments

Lake water table simulation of the model was representative of interannual and seasonal patterns across 25 of 26 study lakes for the observation period (2000-2018). Moreover, the model accurately simulated decadal fluctuations in lake water levels when tested against historical aerial imagery, crucially demonstrating the capability of the model to represent multi-annual memory of cumulative wetting and drying cycles and subsequent propagation into extreme system states. Evaluation of simulated peatland and forestland water table dynamics further exemplified the model's capability to simulate interannual and seasonal temporal patterns at catchment scales by representing dominant

processes, thus indicating that the model outputs were accurate for the right reasons (Kirchner, 2006). This feat of reduced complexity modelling was achieved through zero calibration to 25 of 26 lakes catchments, with only minimal parameter variation between catchment models to reflect differences in catchment land cover proportions, lake geometry and topography. Simple modifications to surficial geology hydraulic properties and forest PET scaling constant were sufficient to simulate the contrasting hydrological processes, magnitude of fluxes and dominant sources of flow of coarse and fine textured surficial geology, indicating robust incorporation of peatland and forestland atmospheric and groundwater processes, while utilising laterally homogenous parameterisation within each HU.

Poorer model performance in larger catchments highlights variable source area as a limiting factor in this study, as the model did not incorporate multiple ephemerally connecting peatland and forestland HUs in these landscapes, nor vary any HU area temporally. Integrating variable source area would require additional parameterisation and calibration to characterise temporally dynamic flowpath networks associated with geographically isolated wetlands (Cohen *et al.,* 2016), which is beyond the scope of this work and deters from the model objective for versatility though low parameterisation. However, the simple conceptual model configuration of lake-peatland-forestland sequence largely applies to most catchments within the Boreal Plains (BP), whereby peatlands are typically adjacent and/or encompass lakes as riparian zones, with forestlands forming on uplands (Devito *et al.,* 2012). Thus, hydrologic interactions between HUs and runoff generated from each of the HU buckets corresponds with conceptual understanding (Devito *et al.,* 2012) and empirical evidence of long runoff return periods in forestlands (Redding and Devito, 2010; Redding and Devito, 2008), but frequent runoff generation from peatlands (Devito *et al.,* 2017) and frequent flow reversals between lakes and peatlands (Ferone and Devito, 2004). This matches with sink-source dynamics observed within the BP, which is largely associated with spatial differences in AET between the HUs (Devito *et al.,* 2017; Thompson, Mendoza, and Devito, 2017), driven by ecohydrological feedbacks.

5.2. Modelling-based insights into hydrologic function in Boreal Plains

Examining system dynamics through numerical modelling enables assessment of how each hydrological process operates and fluctuates (Carrera-Hernández *et al.,* 2011) and extends insights to a wider range of climatic conditions. Due to the long-term P-PET water deficit within the BP, spatial variability in AET driven by hydroecological feedback mechanisms is a crucial driver in hydrological connectivity and the source-sink dependencies within the landscape. Thus, accurate representation of AET was an integral component within the model and required strong conceptual understanding of how best to represent controls with low parameterisation and maintain transferability between catchments.

Diminishing AET with increasing water table depth (WTD) provides an autogenic negative feedback mechanism to conserve water within peatland HUs during dry conditions. While we represented this physical relationship with a simple empirical relationship, the feedback mechanism also integrates many additional processes that drive peatland resistance to drought including: increased moss surface resistance to evaporation; decreasing transmissivity and hydraulic conductivity with depth to reduce lateral flow; reduced ability to conduct water upwards via capillarity beyond a threshold depth; and decreasing specific yield and porosity with depth, which governs the magnitude of water level response to changes in storage (Waddington *et al.,* 2015). As a result of this mechanism, we show that per unit of area, peatlands generated higher landscape runoff than lakes and forestlands, and generated runoff every year. This demonstrates how ET has the largest influence on the water balance (Sterte, Johansson, Sjöberg, Karlsen, and Laudon, 2018), and how peatland feedback mechanisms resist climate-induced shifts in hydrological conditions, however, this conclusion is founded on the assumption that these internal feedbacks will not be overcome by ongoing external stresses such as fire, or anthropogenic disturbance that modify flows from local to landscape scales (Thompson, Mendoza, and Devito, 2017). While peatlands are regarded as important for maintaining surface

waters within the BP during dry periods (Hwang *et al.,* 2018), flow reversals were also present within wet periods. Previous groundwater studies have demonstrated lateral subsurface water redistribution mechanisms (Waddington *et al.,* 2009), such as lake-peatland flow reversals between wet and dry cycles (Ferone and Devito, 2004), which dominate controls on lateral lake–groundwater interaction (Smerdon, Mendoza, and Devito, 2007). In addition, tension driven soil moisture redistribution at night from peatlands to forests alleviates daily soil moisture declines to meet aspen transpiration demands (Brown *et al.,* 2014). We observe this mechanism within our model, which results in shorter forest runoff return periods in peatland dominated FP catchments, although peatlands within forest dominated FH catchments generate higher peak flows as a product of greater cumulative capacity of wet years.

In BP forestlands, it is well understood that soil depth governs storage as a function of capacity, and recharge-upflux ratios, whereby shallower soils permit more frequent and higher quantities of upflux, whereas deeper soils promote groundwater recharge (Carrera-Hernández *et al.,* 2011). Varying soil depths also control recharge lag times in wet periods, ranging from months to multiple years (Carrera-Hernández, Smerdon, and Mendoza, 2012). However, we show that texture of surficial geology was the dominant control on forest runoff. Groundwater connectivity significantly controls BP soil water dynamics, and is a function of surficial geology transmission properties (Plach *et al.,* 2016), and landscape position relative to groundwater flow system (Gibson *et al.,* 2016), which determine whether systems have a recharge, flow-through or discharge function (Price *et al.,* 2005). Larger scale groundwater contributions associated with higher hydraulic conductivity of quartz dominate water balances in lake catchments situated within coarse textured surficial geology. Furthermore, long-term climate interactions with surficial geology texture determine vegetation type (Van Der Velde, Lyon, and Destouni, 2013) and acts to reduce AET in coarse textured systems as high recharge rates limit vegetation access to water (Smerdon, Mendoza, and Devito, 2007). In contrast, aspen dominated forested uplands formed on steeper slopes within fine-textured systems drive higher AET and higher

soil-water storage and subsequently lower average runoff magnitudes and longer runoff return periods (Devito *et al.,* 2017; Redding and Devito, 2008). However, it is notable that temporal variability driven by the long-term climate was as important as surficial geology texture, with prolonged multi-annual periods of very deep forestland water tables driven by cumulative memory of consecutive dry years.

5.3. Applications and implications for industry

Petroleum exploration and extraction is having profound impact on the Boreal ecozone regions of Western Canada (Strack *et al.,* 2019; ABMI, 2019; Webster *et al.,* 2015), where the Oil Sands Region (OSR) of Alberta spans 142,200 km^2 (Chasmer *et al.,* 2021), and is projected to impact 200,000km^2 of boreal landscape once fully operational (Wells and Price, 2015). Oil sands companies have a regulatory obligation to return the post-mined landscape to land uses with equivalent capabilities to those that existed prior to industrial development (Government of Alberta, 1993). Numerous models have been developed to estimate the performance of reclaimed catchments, many of which focus on soil water dynamics in 1-D soil profiles (Keshta, Elshorbagy, and Carey, 2009) or individual slopes (Elshorbagy *et al.,* 2005; Ketcheson *et al.,* 2016; Alam, Barbour, and Huang, 2020) and consider the controls of texture and depth of soil cover (Lukenbach *et al.,* 2019; Carrera-Hernández *et al.,* 2012). Although many studies are monitoring hydrology processes of peatland reconstruction (e.g. Biagi and Carey, 2020; Biagi, Clark, and Carey, 2021; Volik *et al.,* 2021), very few studies have tackled the challenge of modelling reclaimed peatlands (Nwaishi *et al.,* 2015; Price, McLaren, and Rudolph, 2010) and even fewer peatland restoration studies are founded upon use of natural analogues for baseline hydrologic understanding (Wells and Price, 2015).

This study presents the first example of catchment scale hydrological modelling specifically based upon natural analogues that has the capability to be transposed to reclaimed landscapes, which can be

achieved by using HRAs and HUs as fundamental landscape building blocks (Devito *et al.,* 2012). Use of the HRA-HU conceptual framework as the building blocks for a reduced complexity model enables model adaptability of application to a wide range of landscape configurations, by exploring the fluxes within and between ecological HUs (i.e., lakes, peatlands, forests) of varying composition, within the context of different geological HRAs (i.e., coarse sand, fine hummocky moraine, fine clay plains). Not only does the reduced complexity model serve to increase predictive power through rapidly and efficiently computing multiple model configurations, but it assists in understanding of physical processes and increases stakeholder accessibility through a highly visual and user-friendly interface (Malard *et al.,* 2017). This is especially important for emphasising the role of peatlands as the main sources of flow in this sub-humid landscape, contrary to perpetual misunderstanding that forested uplands generate runoff to sustain downgradient wetlands (Price *et al.,* 2010; Sutton and Price, 2020). This concept is vital for the success of reclaimed landscapes, as we demonstrate the role of forested uplands as water sinks in the boreal landscape, in addition to anticipation that landscape configurations of greater water sink to source ratios will be more susceptible to shift in hydrological function under future climates (Schneider *et al.,* 2016; Waddington *et al.,* 2015). Future work using the model presented within this study has the potential to test how different landscape configurations of spatial extent, arrangement and ratios of lake, peatland and forest hydrologic units' impact on hydrological sensitivity to climate and subsequent exceedance of critical thresholds for ecosystem functioning.

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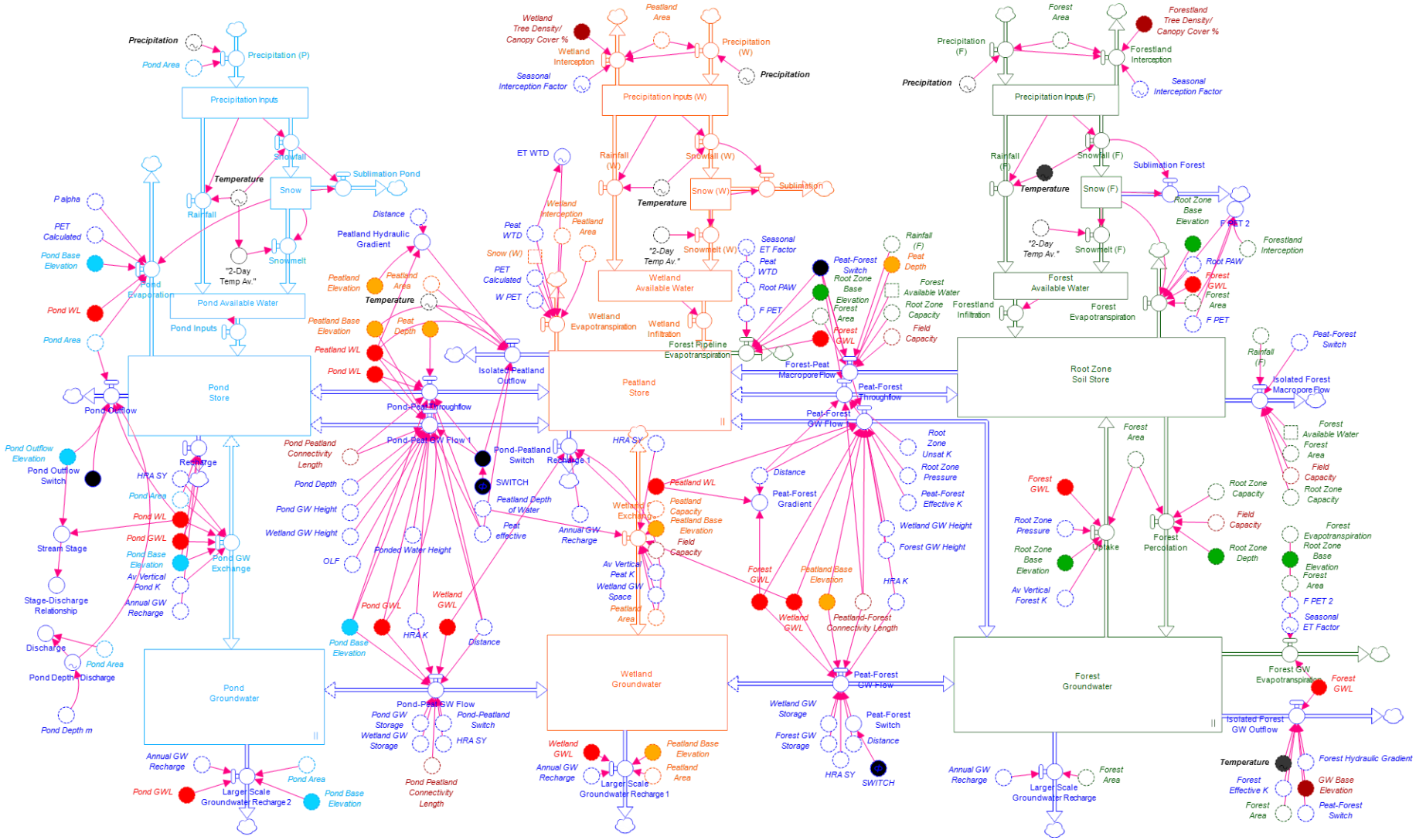
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7. Supplementary Material

Abbreviations, parameter values, and all model equations are provided within the supplementary material below.



Supplementary Figure 5-1. Model structure as seen within STELLA Architect. The model utilises flows conditionally to alternate between perched systems with multiple water tables and saturated groundwater systems. Arrows indicate flows, buckets indicate water stores, and pink links show interactions with converters, denoted by circles, which can represent data inputs, fixed parameters, empirical relationships, initial conditions or thresholds.

Supplementary Table 5-1. Abbreviations and interpretations of variables. Constant parameters are given in Supplementary Table 5-2 and Supplementary Table 5-3.

	Variables	Interpretation
Climatological	P	Observed precipitation (mm day^{-1})
	P_E	Effective precipitation (mm^3 day^{-1})
	P_I	Precipitation Inputs (mm^3 day^{-1})
	I	Interception (mm^3 day^{-1})
	T	Daily mean temperature (°C)
	T_{Av2}	Two-day average temperature (°C)
	AET	Actual Evapotranspiration (mm^3 day^{-1})
	PET	Potential Evapotranspiration (mm day^{-1})
	e_s	Saturated vapor pressure at the given T (mbar)
	ρ_{sat}	Saturated vapor density at the daily mean air temperature T (g m^{-3})
	L_{day}	daytime length in multiples of 12 hours
Water Storage	S_{Lake}	Lake Store (mm^3)
	S_{LakeGW}	Lake Groundwater Store (mm^3)
	S_{Peat}	Peatland Store (mm^3)
	$S_{WetlandGW}$	Peatland Groundwater Store (mm^3)
	$S_{RootZone}$	Root Zone Store (mm^3)
	$S_{TransientZone}$	Transient Zone Store (mm^3)
	$S_{ForestGW}$	Forest Groundwater Store (mm^3)
	D_{Lake}	Depth of the Lake (mm)
	GWH_{Lake}	Depth of Lake groundwater storage (mm) (accounting for specific yield)
	$GWH_{Wetland}$	Depth of peatland groundwater storage (mm) (accounting for specific yield)
	GWH_{Forest}	Depth of forest groundwater storage (mm) (accounting for specific yield)
	$WH_{Acrotelm}$	Depth of peatland acrotelm storage (mm) (accounting for specific yield)
	$WH_{UpperCatotelm}$	Depth of peatland acrotelm storage (mm) (accounting for specific yield)
	$WH_{LowerCatotelm}$	Depth of peatland acrotelm storage (mm) (accounting for specific yield)
Water Level Elevations	WL_{Peat}	Peatland water level elevation asl (mm)
	WL_{Lake}	Lake water level elevation asl (mm)
	$WL_{RootZone}$	Root zone water level elevation asl (mm)
	GWL_{Forest}	Forest Groundwater Elevation asl (mm)
	GWL_{Lake}	Lake Groundwater Elevation asl (mm)
	$GWL_{Wetland}$	Peatland Groundwater Elevation asl (mm)
	WTD_{Peat}	Peatland water table depth (mm)
	WTD_{Forest}	Forest water table depth (mm)
Soil Parameters	K	Horizontal unsaturated hydraulic conductivity (mm day^{-1})
	$K_{RootZone}$	Unsaturated hydraulic conductivity of the root zone (mm day^{-1})
	PAW	Plant Available Water (mm)
	θ	Volumetric Soil Moisture Content (%)
	θ_{Peat}	Peatland Soil Moisture Content (%)
	$\theta_{RootZone}$	Root Zone Soil Moisture Content (%)
	$h_{Peatland}$	Peatland Pressure (mm)
	$h_{RootZone}$	Root Zone Pressure (mm)
	$h_{TransientZone}$	Transient Zone Pressure (mm)
	$h_{WetlandGW}$	Wetland Groundwater Pressure (mm)

*Supplementary Table 5-2. Constant parameter values for model of each lake catchment dimensions, landscape proportions and elevations. *CO/FH classes were partially calibrated due to uncertainty of HRA texture proportions; lakes 19 and 15 were assigned fine textured parameters with a perched over coarse set up; lakes 7 and 11 were assigned parameters given to coarse textured lake catchments. Additional constant parameter values for catchment structure are given in table x for each lake catchment.*

Constant Parameter		Estimated Value			Interpretation	Source Reference
		Coarse	Perched*	Fine		
Climatological	I_{Season}	0mm (14/Sep-28/Apr) 0.5mm (29/Apr-29/May) 2mm (30/May-13/Sep)	0mm (14/Sep-28/Apr) 0.5mm (29/Apr-29/May) 2mm (30/May-13/Sep)	0mm (14/Sep-28/Apr) 0.5mm (29/Apr-29/May) 2mm (30/May-13/Sep)	Seasonal Interception Factor (mm)	Thompson *et al.,* 2015
	α	Lake = 1.2 Peatland = 0.8 Forest = 1	Lake = 1.2 Peatland = 0.8 Forest = 1.5	Lake = 1.2 Peatland = 0.8 Forest = 1.5	Scales PET by Hydrologic Unit (Dimensionless)	Devito *et al.,* 2012
	β	0.5	0.5	0.5	Scales PET under dry conditions (Dimensionless)	
Soil Parameters	$D_{ActiveLayer}$	0.625	0.75	0.75	Depth of the forest floor active later (m)	Smerdon *et al.,* x. Carrera-Hernandez et at., 2011
	$D_{RootZone}$	2.5	3	3	Depth of the forest root zone (m)	Smerdon *et al.,* x. Carrera-Hernandez et at., 2011
	D_{Peat}	2	1	2	Depth of peat (m)	
	$D_{Acrotelm}$	0.2	0.2	0.2	Depth of peat acrotelm layer (m)	
	$D_{UpperCatotelm}$	0.8	0.35	0.8	Depth of peat upper catotelm layer (m)	
	$D_{LowerCatotelm}$	1	0.45	1	Depth of peat lower catotelm layer (m)	
	K_{Sat}	7×10^{-5}	1×10^{-8}	1×10^{-8}	Horizontal saturated hydraulic conductivity of the HRA soil (mm day^{-1})	Hokanson *et al.,* 2019
	$K_{ActiveLayer}$	1×10^{-4}	1×10^{-4}	1×10^{-4}	Hydraulic conductivity of the forest floor "active layer" (mm day^{-1})	Thompson *et al.,* 2015., Forest floor
	$K_{RootZone}$	1×10^{-7}	1×10^{-7}	1×10^{-7}	Hydraulic conductivity of the forest root zone (mm day^{-1})	Thompson *et al.,* 2015, Glacial Till vertical upper layer
	$K_{Acrotelm}$	3×10^{-3}	3×10^{-3}	3×10^{-3}	Horizontal saturated hydraulic conductivity of the peat within the acrotelm (mm day^{-1})	Thompson *et al.,* 2015, Peat 0-0.1m depth
	$K_{UpperCatotelm}$	2×10^{-5}	2×10^{-5}	2×10^{-5}	Horizontal saturated hydraulic conductivity of the peat within the upper catotelm (mm day^{-1})	Thompson *et al.,* 2015, average of peat 0.3-2m depth
	$K_{LowerCatotelm}$	3×10^{-8}	3×10^{-8}	3×10^{-8}	Horizontal saturated hydraulic conductivity of the peat within the lower catotelm (mm day^{-1})	Thompson *et al.,* 2015, peat at 2m depth
	$K_{VerticalLake}$	1×10^{-8}	1×10^{-8}	1×10^{-8}	Vertical hydraulic conductivity across the HRA and the gyttja Lake base (mm day^{-1})	Approximation from multiple horions in Thompson *et al.,* 2015; gyttja 0-2m
	$K_{VerticalPeat}$	1×10^{-8}	1×10^{-8}	1×10^{-8}	Vertical hydraulic conductivity across the peat and peatland base (mm day^{-1})	Approximation from multiple horions in Thompson *et al.,* 2015; vertical gyttja 0-2m/ vertical peat >1.5m
	SY_{HRA}	0.07	0.01	0.01	Specific yield of the superficial geology	Determined from RETC
	$SY_{ActiveLayer}$	0.1	0.1	0.1	Specific yield of the forest floor active later	Determined from RETC
	SY_{Soil}	0.03	0.03	0.03	Specific yield of the forest root zone soil	Determined from RETC
	$SY_{Acrotelm}$	0.35	0.35	0.35	Specific yield of the peatland acrotelm	Determined from RETC
	$SY_{Catotelm}$	0.15	0.15	0.15	Specific yield of the peatland catotelm	Determined from RETC
	$SY_{LowerCatotelm}$	0.025	0.025	0.025	Specific yield of the peatland lower catotelm	Determined from RETC

*Supplementary Table 5-3. Constant parameter values for model of each lake catchment dimensions, landscape proportions and elevations. *CO/FH classes were partially calibrated due to uncertainty of HRA texture proportions; lakes 19 and 15 were assigned fine textured parameters with a perched over coarse set up; lakes 7 and 11 were assigned parameters given to coarse textured lake catchments. Additional constant parameter values for catchment structure are given in table x for each lake catchment.*

Lake	1	5	16	17	201	206	208	7	11	15	19	27	39	40	42	43	48	111	59	112	118	122	121	168	171	205
Texture	Coarse							Perched (Coarse)		Perched (Fine)		Fine														
Lake Class	CO							CO/FH				FH							FP/FH				FP			
Lake Area (m^2)	112090	3244530	465820	1059549	382239	136775	37142	28671	10680	485	37390	67743	63994	7725	14037	12773	49779	58463	240951	48472	50625	91998	86767	135410	107893	204979
Peatland Area (m^2)	10810140	47217293	607116	1586712	163064	624327	107983	203699	76999	29246	61140	181361	312947	54697	91317	86695	187365	166327	934714	2026174	354600	281732	478566	13693334	638471	1702353
Forest Area (m^2)	4800169	38513825	780000	2072528	175696	1073824	35055	383034	41210	31512	196659	187411	1745368	29516	195920	174668	530310	238662	499010	3295046	39608	0	0	5177374	77140	765445
Lake Base Elevation (m amsl)	643.0	638.0	641.9	640.0	637.0	641.0	638.0	643.0	645.5	645.0	648.8	642.9	638.9	651.5	643.5	652.5	651.6	636.6	629.0	627.0	624.6	624.1	623.9	635.3	634.5	635.8
Lake Sill Height (m)	1.0	0.3	2.0	5.0	2.8	2.7	2.0	1.8	2.0	0.5	2.5	0.9	2.5	0.8	2.3	0.9	1.0	2.0	1.0	1.4	1.6	1.6	1.6	0.8	0.9	1.3
Peatland Elevation (m amsl)	645.0	641.0	644.2	647.0	641.5	644.2	640.0	644.5	647.0	645.5	651.0	653.9	641.0	652.7	645.7	654.2	652.5	638.5	630.5	628.5	626.2	626.1	625.5	637.2	635.8	637.5
Peatland Base Elevation (m amsl)	643.0	639.0	642.2	645.0	639.5	642.2	638.0	642.5	645.0	644.5	650.0	651.9	639.0	650.7	643.7	652.2	650.5	636.5	628.5	626.5	624.2	624.1	623.5	635.2	633.8	635.5
Forest Elevation (m amsl)	668.0	663.0	648.4	652.0	644.0	652.0	647.0	653.0	655.5	659.0	655.0	657.0	645.5	655.0	648.0	656.0	655.0	639.0	635.0	633.0	627.0	625.0	625.0	650.0	637.0	644.0
Root Zone Base Elevation (m amsl)	665.5	660.5	645.9	649.5	641.5	649.5	644.5	650.5	653.0	656.0	652.0	654.0	642.5	652.0	645.0	653.0	652.0	636.0	632.0	630.0	624.0	622.0	622.0	647.0	634.5	641.0
GW Base Elevation (m amsl)	620.0	520.0	620.0	620.0	620.0	620.0	620.0	620.0	620.0	620.0	620.0	630.0	620.0	640.0	620.0	640.0	640.0	620.0	620.0	620.0	620.0	620.0	620.0	620.0	620.0	620.0
Lake-Peatland Distance (m)	200000	500000	800	2000	500	700	200	600	200	100	150	250	1000	200	300	400	300	250	8000	1500	500	400	800	3000	400	2000
Peatland-Forest Distance (m)	200000	500000	800	2000	500	700	200	600	200	150	250	250	1000	200	300	400	300	250	8000	1500	500	400	800	3000	400	2000
Lake Peatland Connectivity Length (m)	3114	14158	3069	6984	2562	2355	726	762	374	93	820	1226	880	341	485	623	1356	1262	2273	1125	1271	1232	1064	1512	1246	1688
Peatland Forest Connectivity Length (m)	66800	668000	13613	27000	2000	15000	1500	6000	1000	1100	1800	5061	9898	750	2880	3170	6302	5156	6000	22000	3000	0	0	69380	2600	9000

Flows and Fluxes

This section defines the equations for all flows and fluxes, corresponding with numbered flows in Figure 5-2.

Equation	*Explanation*	
Precipitation Inputs ($mm^3\ day^{-1}$) (1) $P_I = P \times A_{HU}$	Precipitation x Area	***ATMOSPHERIC EXCHANGE***
Interception ($mm^3\ day^{-1}$) (2) IF($P_I > 0$) THEN($I_{Season} \times T_D \times A_{HU}$) ELSE(0)	On days with precipitation, interception loss varies seasonally x Tree density x Area Interception loss = 0.5mm/day on day 120-150; 2mm/day on day 151-257; and 0mm/day on day 258-119	
Rainfall ($mm^3\ day^{-1}$) (3) IF($T > 0°C$) THEN(P_E) ELSE(0) Where $P_E = P_I - I$	If temperature>0°C, Then, rainfall = effective precipitation, Else, 0mm rainfall Where effective precipitation = precipitation inputs - interception	
Snowfall ($mm^3\ day^{-1}$) (4) IF($T < 0$) THEN(P_E) ELSE(0) Where $P_E = P_I - I$	If temperature<0, Then, snowfall = effective precipitation, Else, 0mm snowfall Where effective precipitation = precipitation inputs - interception	
Sublimation ($mm^3\ day^{-1}$) (5) Snowfall x 0.25	25% of snowfall sublimates	
Snowmelt ($mm^3\ day^{-1}$) (6) IF($T_{Av2} > 0°C$) THEN(IF(PREVIOUS(SELF=0) THEN(Snow/2) ELSE(Snow)) ELSE(0)	If the two day average temperature is greater than 0 degrees C, Then, If there was no snowmelt yesterday, Then, half the snowpack melts, Else, melt the remaining snowpack. Else, the two day average temperature is less than 0, and no snowmelt occurs. This enables snow to fully melt across two consecutive days greater than where two-day average temperature exceeds 0°C, but prolongs snowmelt duration when daily temperatures fluctuate around 0°C.	

Equation	*Explanation*	
Lake Actual Evaporation (AET; $mm^3\ day^{-1}$) (7a) IF($S_{Lake} > 0$) THEN(PET x α x A_{HU}) ELSE(0) Where $PET = 0.1651 \cdot L_{day} \cdot \rho_{sat}$ $\rho_{sat} = \frac{216.7 \cdot e_s}{T + 273.3}$ $e_s = 6.108 \cdot e^{\frac{17.26 \cdot T}{T+273.3}}$	If there is water in the Lake, Then, Hamon's PET x PET multiplier x Area, Else, the Lake is dry, and no evaporation occurs.	ATMOSPHERIC EXCHANGE
Wetland Actual Evapotranspiration (AET; $mm^3\ day^{-1}$) (7b) IF($WTD_{Peat} < 1000$) THEN(IF($WTD_{Peat} > 0$) THEN((PET x α x β x A_{HU}) – I) ELSE((PET x A_{HU}) – I) ELSE(0) Where $PET = 0.1651 \cdot L_{day} \cdot \rho_{sat}$ $\rho_{sat} = \frac{216.7 \cdot e_s}{T + 273.3}$ $e_s = 6.108 \cdot e^{\frac{17.26 \cdot T}{T+273.3}}$ And β declines linearly with WTD_{Peat}	If the WTD is less than 1m, Then, If the WTD is more than 0 (i.e. below the surface and not ponded) Then, (Hamon's PET x PET multiplier x WTD restriction x Area) minus interception loss. Else, WTD is less than 0 (ponded surface), Hamon's PET x Area, minus interception loss. Else, WTD is greater than 1m, ET is limited by WTD and equals 0. PET multiplier (α) is 0.8 for WTD below the surface. Once ponded, PET multiplier equals 1, therefore not included. WTD restriction (β) declines linearly from 1 at 0.2m, to 0 at 1m depth.	

Equation	*Explanation*	

Forest Actual Evapotranspiration (AET; mm^3 day^{-1})
(7ci) Forest Root Zone Evapotranspiration (perched system)
 	
IF($Snow_D$<1)
THEN(
IF(GWL_{Forest} > $Elevation_{RZB}$)
THEN(
IF((PET x α – (I/A_{HU}))*0.25 < PAW)
THEN(((PET x α x A_{HU})-I)*0.25)
ELSE(PAW x A_{HU}))
ELSE(
IF((PET x α – (I/A_{HU})) < PAW)
THEN((PET x α x A_{HU}) – I)
ELSE((PAW x A_{HU})+(0.5 x A_{HU} x (PET x α – (I/A_{HU})))))
ELSE(0)
 	
Where $PET = 0.1651 \cdot L_{day} \cdot \rho_{sat}$
 	
$\rho_{sat} = \frac{216.7 \cdot e_s}{T + 273.3}$
 	
$e_s = 6.108 \cdot e^{\frac{17.26 \cdot T}{T+273.3}}$
	Forest Root Zone Evapotranspiration (perched system)
	
	If there is no snow on the ground,
	Then,
	
	If the groundwater level is within the root zone,
	Then,
	If plant available water is greater than 0.25 of PET x Hamon's multiplier minus interception loss,
	Then sufficient water, ET equals 0.25 of PET x Hamon's multiplier minus interception loss,
	Else, insufficient water, ET equals plant available water,
	
	Else, the groundwater is below the root zone,
	If plant available water is greater than PET x Hamon's multiplier minus interception loss,
	Then, sufficient water, ET equals PET x Hamon's multiplier minus interception loss,
	Else, insufficient water, ET equals plant available water plus 0.5 of PET x Hamon's multiplier minus plant available water minus interception loss.
	
	Else, there is snow, ET equals 0.
	
	This enables the root zone to evaporate based on available moisture, and also determines that additional AET will occur from both the root zone and groundwater (see 7cii) when the groundwater level is within the root zone.
		ATMOSPHERIC EXCHANGE

(7cii) Forest Groundwater Evapotranspiration (saturated system) IF(GWL_{Forest} < $Elevation_{RZB}$) THEN(0) ELSE((PET x α x A_{HU} x ET_{Season}) – I – ET_{RZ})	Forest Groundwater Evapotranspiration (saturated system) If the groundwater level is below the root zone, Then, perched system, ET equals 0, Else, saturated system, ET equals PET x Hamon's multiplier minus interception loss and root zone ET and is limited seasonally. ET_{Season} increases linearly from 0 on day 121 to 1 on day 138, then decreases linearly from 1 on 240 to 0 on day 257 to create shoulder seasons.	

Equation	*Explanation*
Infiltration ($mm^3\ day^{-1}$) (8) Infiltration = Available Water Available Water = Rainfall + Snowmelt	Infiltration rates typically exceed rainfall intensities within the western boreal plains, therefore the entire available water store infiltrates into the peatland/ forest.
Forest Root Zone Percolation ($mm^3\ day^{-1}$) (9) IF($\theta_{RootZone} > \theta_{FC}$) THEN($S_{RootZone}$ x (($\theta_{RootZone} - \theta_{FC}$) /100)) ELSE(0)	The movement of water from the root zone occurs when soil moisture thresholds are exceeded. Any water content above field capacity percolates.
Uptake ($mm^3\ day^{-1}$) (10) ($K_{VerticalForest}$ x A_{HU} x (((($h_{RootZone}$ - 0)/($Elevation_{RZB}$ - GWL_{Forest})) +1) x -1))	The movement of water from the forest groundwater to the root zone driven by pressure, related to soil moisture within the root zone.
Vertical Groundwater Flow ($mm^3\ day^{-1}$) (11a) IF(WL_{Lake} > $Elevation_{LakeBase}$) THEN(IF($Elevation_{LakeBase}$ > GWL_{Lake}) THEN($K_{VerticalLake}$ x A_{HU}) ELSE(($Elevation_{LakeBase}$ - GWL_{Lake}) x A_{HU} x SY_{HRA})) ELSE(0)	If there is water in the Lake, Then, If the groundwater level is below the Lake base, Then, vertical Darcy's flow, Else, the groundwater is above the Lake base, move to the Lake bucket. Else, the Lake is dry, no flow.
(11b) IF(GWL_{Peat} < $Elevation_{PeatBase}$) THEN(IF(($K_{VerticalPeat}$ x A_{HU} x (((S_{Peat}/A_{HU})/ ($Elevation_{PeatBase}$ - GWL_{Peat})) +1)) < ((($Elevation_{PeatBase}$ - GWL_{Peat}) x SY_{HRA}) x A_{HU})) THEN($K_{VerticalPeat}$ x A_{HU} x (((S_{Peat}/A_{HU})/ ($Elevation_{PeatBase}$ - GWL_{Peat})) + 1)) ELSE((($Elevation_{PeatBase}$ - GWL_{Peat}) x SY_{HRA}) x A_{HU})) ELSE((($Elevation_{PeatBase}$ - GWL_{Peat}) x SY_{HRA}) x A_{HU}))	If the groundwater level is below the peatland base, Then, If calculated vertical Darcy's flow is less than available space in the GW bucket Then, vertical Darcy's flow Else, only flow the amount until the GW is full Else, the groundwater is above the peatland base, move this excess from the groundwater bucket to the peatland bucket.
Larger Scale Regional Groundwater Recharge ($mm^3\ day^{-1}$) (12) ($Recharge_{Annual}$/365 x A_{HU})	10mm of recharge per year recharges the regional groundwater.

VERTICAL FLOWS

Equation	*Explanation*	

Lake-Peat GW Flow ($mm^3\ day^{-1}$)
(13a) Saturated System
IF(GWL_{Lake} > $Elevation_{LakeBase}$)
THEN(0)
ELSE(
IF(($WL_{Peatland}$ < $Elevation_{Peatland}$)
THEN((K_{Sat} x ((WL_{Lake} - $WL_{Wetland}$)/ $Dist_{Lake\text{-}Peat}$) x $L_{Lake\text{-}Peat}$ x (($GWH_{Wetland}$ + GWH_{Lake})/ 2)) + ($K_{PeatEffective}$ x ((WL_{Lake} - $WL_{Wetland}$)/ $Dist_{Lake\text{-}Peat}$) x $L_{Lake\text{-}Peat}$ x (((WL_{Lake} - $Elevation_{LakeBase}$) + D_{Lake})/ 2)))
ELSE(OLF)))
 	
Where $Dist_{Lake\text{-}Peat}$ = (W_{Lake} + W_{Peat})/2
$L_{Lake\text{-}Peat}$ = (L_{Lake} + L_{Peat})/2
OLF = IF(WL_{Lake} > $Elevation_{Peatland}$)
THEN((WL_{Lake} - $WL_{Wetland}$) x A_{Lake} x ($A_{Peatland}$ / ($A_{Peatland}$ + A_{Lake})))
ELSE(MAX((WL_{Lake} - $WL_{Wetland}$) x A_{Lake} x ($A_{Peatland}$ / ($A_{Peatland}$ + A_{Lake})), -1 x ($WL_{Peatland}$ – $Elevation_{Peatland}$) x $A_{Peatland}$))
	Groundwater flow between the Lake and peatland HUs operates as a bi-flow governed by Darcy's law, allowing for flow reversals. In saturated systems where the groundwater buckets are "full", hydraulic heads of the surface buckets are used, and groundwater flow operates between the surface buckets (eq. 13a). In perched systems with more than one water table, the hydraulic heads of the groundwater are used for the calculations, and flow operates between the lower buckets (eq. 13b) in conjunction with Lake-peatland throughflow between the surface buckets (eq. 14).
	
	If the Lake groundwater level is above the Lake base (saturated system)
	Then, 0
	Else,
	If the peatland water level is below the peatland surface,
	Then, Darcy's flow
	Else, the peatland water level is above the surface, overland flow.
		LATERAL FLOWS

(13b) Perched System ` IF(GWL_{Lake} < $Elevation_{LakeBase}$) THEN(K_{Sat} x (((GWL_{Lake}) - ($GWL_{Wetland}$))/ $Dist_{Lake\text{-}Peat}$) x $L_{Lake\text{-}Peat}$ x (($GWH_{Wetland}$ + GWH_{Lake})/2)) ELSE(0) Where $Dist_{Lake\text{-}Peat}$ = (W_{Lake} + W_{Peat})/2 $L_{Lake\text{-}Peat}$ = (L_{Lake} + L_{Peat})/2	If the Lake groundwater level is below the Lake base (perched system) Then, Darcy's flow Else, no flow	

Equation	*Explanation*	
Lake-Peat Throughflow ($mm^3\ day^{-1}$) (14) IF($GWL_{Wetland} < Elevation_{PeatBase}$) THEN ($K_{PeatEffective}$ x (($WL_{Lake} - WL_{Peat}$)/ $Dist_{Lake\text{-}Peat}$) x $L_{Lake\text{-}Peat}$ x D_{Peat}) ELSE(0) Where $Dist_{Lake\text{-}Peat} = (W_{Lake} + W_{Peat})/2$ $L_{Lake\text{-}Peat} = (L_{Lake} + L_{Peat})/2$	Throughflow between the Lake and peatland will occur under saturated conditions and is governed by Darcy's law, allowing for flow reversals.	***LATERAL FLOWS***
Peat-Forest GW Flow ($mm^3\ day^{-1}$) (15a) Saturated System IF($GWL_{Wetland} > Elevation_{PeatBase}$) THEN((($K_{HRA}$ x (((WL_{Peat})-(GWL_{Forest}))/ $Dist_{Peat\text{-}Forest}$) x $L_{Peat\text{-}Forest}$ x (($GWH_{Wetland}$ + GWH_{Forest})/2)) + (($K_{P\text{-}F\text{-}effective}$ x (((WL_{Peat})-(GWL_{Forest}))/ $Dist_{Peat\text{-}Forest}$) x $L_{Peat\text{-}Forest}$ x (((WL_{Peat} - $Elevation_{PeatBase}$)+ GWH_{Forest})/2)) ELSE(0)) Where $Dist_{Peat\text{-}Forest} = (W_{Forest} + W_{Peat})/2$ $L_{Peat\text{-}Forest} = (L_{Forest} + L_{Peat})/2$	Groundwater flow between the forest and peatland HUs operates as a bi-flow governed by Darcy's law, allowing for flow reversals. In saturated systems where the groundwater buckets are "full", hydraulic heads of the surface buckets are used, and groundwater flow operates between the peatland surface bucket and the forest groundwater bucket (eq. 15a). In perched systems with more than one water table, the hydraulic heads of the groundwater are used for the calculations, and flow operates between the lower buckets (eq. 15b). If wetland groundwater level is above the peatland base (saturated system) Then, Darcy's flow Else, no flow	
(15b) Perched system IF($GWL_{Wetland} < Elevation_{PeatBase}$) THEN($K_{Sat}$ x (($GWL_{Wetland}$ - GWL_{Forest})/ $Dist_{Peat\text{-}Forest}$) x $L_{Peat\text{-}Forest}$ x (($GWH_{Wetland}$ + GWH_{Forest})/2)) ELSE(0) Where $Dist_{Peat\text{-}Forest} = (W_{Forest} + W_{Peat})/2$ $L_{Peat\text{-}Forest} = (L_{Forest} + L_{Peat})/2$	If wetland groundwater level is below the peatland base (saturated system) Then, Darcy's flow Else, no flow	

Equation	*Explanation*	
Peat-Forest Throughflow ($mm^3\ day^{-1}$) (16) $((K_{RootZone} \times ((0 - h_{RootZone})/ Dist_{Peat\text{-}Forest})) \times L_{Peat\text{-}Forest} \times D_{Peat}) \times -1$ Where $Dist_{Peat\text{-}Forest} = (W_{Forest} + W_{Peat})/2$ $L_{Peat\text{-}Forest} = (L_{Forest} + L_{Peat})/2$	Unsaturated flow between forest and peatland driven by hydraulic heads and pressure and unsaturated K (determined by soil moisture of the forest root zone)	***LATERAL FLOWS***
Forest Rapid Flow ($mm^3\ day^{-1}$) (17) IF(Rainfall > 25 x A_{HU}) THEN(IF(Rainfall > (θ_{FC} - $\theta_{RootZone}$) x A_{HU}) THEN(0.5 x Rainfall x (θ_{FC} - $\theta_{RootZone}$)) ELSE(0)) ELSE(0)	Infiltration excess runoff based on empirical evidence of forest floor "active layer" (Smerdon *et al.,* x) If rainfall event greater than 25mm/day, Then, If rainfall event greater than the inputs required to exceed field capacity, Then, half of all inputs exceeding field capacity runoff, Else, insufficient inputs, no runoff, Else, insufficient inputs, no runoff.	
Lake Outflow ($mm^3\ day^{-1}$) (18) IF(WL_{Lake} > $Elevation_{LakeOut}$) THEN((WL_{Lake} – $Elevation_{LakeOut}$) x A_{HU} x C) ELSE(0)	If the Lake water level is above the outflow elevation, then water level minus outflow elevation x area x constant to define flow rate.	

Storage and Water Levels

This section defines the equations for all storage components, including volumes, water levels and moisture contents for stores shown in Figure 5-2.

Lake Store (mm^3)

(19) ∇S_{Lake} = Rainfall + Snowmelt – Evaporation – Lake-Peatland Throughflow – Lake Outflow - Groundwater $Flow_V$

Lake Water Level (mm asl)

(20) $WL_{Lake} = Elevation_{LakeBase} + (S_{Lake}/ A_{HU})$

Lake Groundwater Store (mm^3)

(21) ∇S_{LakeGW} = Groundwater $Flow_V$ – Lake-Peatland GW Flow – GW Outflow

Lake Groundwater Level (mm asl)

(22) $GWL_{Lake} = Elevation_{GWBase} + (S_{LakeGW}/ A_{HU})$

Peatland Store (mm^3)

(23) ∇S_{Peat} = Available Water – Evapotranspiration – Aspen Pipeline – Groundwater $Flow_V$ + Lake-Peatland Throughflow + Forest Macropore Flow – Peatland-Forestland GW Flow

Peatland Water Level (mm asl)

(24) WL_{Peat} =

IF($S_{Peat}/ A_{HU} < WH_{LowerCatotelm}$)
THEN($Elevation_{PeatBase} + (S_{Peat}/ A_{HU}) / SY_{LowerCatotelm}$))
ELSE(
IF($S_{Peat} < WH_{UpperCatotelm} + WH_{LowerCatotelm}$)
THEN($Elevation_{PeatBase} + D_{LowerCatotelm} + ((S_{Peat}/ A_{HU}) - WH_{LowerCatotelm})/ SY_{UpperCatotelm}$))
ELSE(
IF($S_{Peat} < WH_{Peatland}$)
THEN($Elevation_{PeatBase} + D_{UpperCatotelm} + D_{LowerCatotelm} + ((S_{Peat}/ A_{HU}) - (WH_{UpperCatotelm} + WH_{LowerCatotelm}))/ SY_{CatotelmAcrotelm}$))
ELSE($Elevation_{PeatBase} + (S_{Peat}/ A_{HU}) - (WH_{Peatland})$))

If volume of peatland water is less than the storage capacity of the lower catotelm
Then calculate water table using lower catotelm hydraulic properties only
Else
If volume of peatland water is less than the storage capacity of the lower plus upper catotelm
Then calculate water table using lower and upper catotelm hydraulic properties
Else
If the volume of peatland water is less than the storage capacity of the peatland (acrotelm plus upper and lower catotelm)
Then calculate the water table using acrotelm and lower and upper catotelm hydraulic properties
Else the peatland is fully saturated, calculate height of standing water plus peatland surface elevation

Peatland Water Content (%)

(25) $\theta_{Peat} = ((S_{Lake}/ A_{HU})/D_{Peat})*100$

Peatland Groundwater Store (mm^3)

(26) $\nabla S_{WetlandGW} = \text{Groundwater Flow}_V + \text{Lake-Peatland GW Flow}_H - \text{Peatland-Forest GW Flow}_H$

Peatland Groundwater Level (mm asl)

(27) $GWL_{Wetland} = \text{Elevation}_{GWBase} + (S_{WetlandGW}/ A_{HU})$

Forest Root Zone Store (mm^3)

(28) $\nabla S_{RootZone}$ = Available Water – Evapotranspiration – Forest Macropore Flow + Lake-Peatland Throughflow – Percolation + Uptake

Forest Root Zone Water Level (mm asl)

(29) $WL_{RootZone} = \text{Elevation}_{RZB} + ((S_{RootZone}/ A_{HU})/SY_{Soil})$

Forest Root Zone Water Content (%)

(30) $\theta_{RootZone} = ((S_{RootZone}/ A_{HU})/D_{RootZone})*100$

Forest Groundwater Store (mm^3)

(31) $\nabla S_{ForestGW} = \text{Percolation}_{Lower} - \text{Uptake}_{Lower} - \text{Peatland-Forest GW Flow}_H$

Forest Groundwater Water Level (mm asl)

(32) GWL_{Forest} =

IF(($Elevation_{GWBase}$ + ($S_{ForestGW}$ / SY_{Soil}))< $Elevation_{RZB}$)	If forest groundwater level below root zone base
THEN($Elevation_{GWBase}$ +(($S_{ForestGW}$ / A_{HU}) / SY_{HRA}))	Then (calculate level based on HRA soil hydraulic properties only): elevation of groundwater base (model base) + volume of groundwater divided by forest area, divided by HRA SY
ELSE(	Else
IF($Elevation_{RZB}$ +((($S_{ForestGW}$ / A_{HU}) - GWH_{Forest})/ SY_{Soil})< $Elevation_{ActiveLayerBase}$)	If groundwater level above root zone base, but below active layer base
THEN($Elevation_{RZB}$ +((($S_{ForestGW}$ / A_{HU}) - GWH_{Forest})/ SY_{Soil}))	Then (calculate level based on HRA and root zone soil hydraulic properties): elevation of root zone base plus (volume of groundwater divided by forest area, minus groundwater stored in groundwater bucket, divided by soil SY)
ELSE	Else
IF($S_{ForestGW}$ < GWH_{Forest} + $GWH_{ActiveLayer}$ + $GWH_{RootZone}$)	If volume of groundwater is less than total storage capacity of groundwater bucket, plus root zone plus active layer
THEN(($Elevation_{ActiveLayerBase}$ +($S_{ForestGW}$ - GWH_{Forest} - $GWH_{RootZone}$)/ $SY_{ActiveLayer}$))	(calculate level based on HRA and root zone and active layer soil hydraulic properties) Then elevation of active layer base plus (volume of groundwater divided by forest area minus groundwater stored in groundwater bucket and in root zone, divided by active layer specific yield
ELSE($Elevation_{Forest}$ +($S_{ForestGW}$ -(GWH_{Forest} + $GWH_{ActiveLayer}$ + $GWH_{RootZone}$))))	Else (calculate level as ponded water based on exceeding capacity of HRA and root zone and active layer soil hydraulic properties) Forest ground surface elevation plus (volume of groundwater divided by forest area, minus groundwater stored in groundwater bucket and in root zone and in active layer)

CHAPTER SIX: KEY RESEARCH FINDINGS, SYNTHESIS AND FUTURE DIRECTIONS

1. Introduction

Our knowledge of the ecohydrological and climatic feedbacks that shape the resilience of boreal ecosystems has advanced significantly in recent years (Price *et al.,* 2005; Waddington *et al.,* 2009). Yet application of this knowledge at landscape scales remains a significant limitation for prediction of future system responses (Ireson *et al.,* 2015). The research presented within this thesis was motivated by the need to test and advance hydrological frameworks at landscape scales by unravelling landscape controls from climatic controls, made possible by the availability of extraordinary data records of sufficient length to observe both seasonal and decadal climate variability across a range of diverse catchments under one climate signal.

The aim of this research was to determine how climate, geology, landcover and landscape position interact to control spatiotemporal variability of shallow lake water depth to inform a framework based upon mechanistic understanding within a complex heterogenous hydrogeoclimatic BP setting.

This thesis addressed four interlinked research objectives. Chapter 2 (objective 1) characterised spatiotemporal patterns in lake water level fluctuations intra-annually and interannually, and catchment classification techniques were employed to provide meaningful and process orientated classes with distinct hydrological patterns and threshold behaviour. Chapter 3 (objective 2) tested landscape controls on lake water level variability at decadal timescales to inform a framework based upon mechanistic understanding of hydrologic interaction of lakes with their catchment, geology and climate. Chapter 4 (objective 3) evaluated the role of landscape memory of climate in controlling the magnitude of hydrological response to periods of drought or deluge and how this varied between different lake-landscape types. Chapter 5 (objective 4) produced a versatile, reduced-complexity system dynamics numerical model to further test conceptual understanding of dominant simulated fluxes and landscape interactions and assess spatial differences in landscape runoff sensitivity to the

long-term climate. This research generated new understanding of the dominant controls on boreal lake, peatland and forestland hydrological interactions; strengthened conceptual understanding from previous site-specific studies transposed to catchment scales; and underpins crucial insights and provides tools for evaluating hydrological sensitivity to climate. In this chapter the key research findings are identified, and synthesised, and future research avenues are discussed.

2. Key Research Findings

The research presented within this thesis was unique in: (1) first study to examine hydrometrics of diverse lake and catchment types over a full long-term climate cycle and provide evidence for spatially diverse water level patterns under one climate signal; (2) demonstrating how spatial variability in lake water level patterns, connectivity and threshold dynamics are explained by interactions of climate with surficial geology, land cover, landscape position and lake geometry, and evidencing dominant processes with both hydrologic and hydrochemical monitoring over decadal time scales; (3) providing the first study analysing the impacts of dynamical controls (climate memory) in addition to static controls (landscape characteristics) across diverse hydrogeologic catchments under one climate signal; (4) producing a versatile numerical model capable of accurately representing catchment scale hydrological fluxes and water tables and successfully transpose the model to multiple diverse landscape configurations. Major research outcomes are as follows:

1. Shallow BP lakes set within the diverse hydrogeologic landscapes demonstrate low synchronicity of lake water level patterns. High storage of water gives rise to high spatial variation in lake water levels, which emerge over long-term (decadal) times scales as interannual variability in lake water levels is significantly greater than intra-annual variability, in a system that experiences both strong decadal and seasonal weather cycles. BP lake water level patterns are exceptionally complex as connectivity to wetlands is ephemeral across

varying timescales and heterogeneous surficial geology results in varied scales and magnitudes of groundwater connectivity.

2. Surficial geology incorporates important landscape parameters that control lake water level patterns and characteristics. Thus, no single dominant control on lake hydrology exists within the BP. Hydrogeological setting drives scales of lake-groundwater connectivity and interactions with land cover and lake geometry further modifies hydrological connectivity. We identify spatially variable hydrologic pathways that enable lakes to persist within the sub-humid BP, whereby these processes cause all lake types to fluctuate over contrasting timescales. Such mechanisms control the source, magnitude, and continuity of groundwater supply, and modify surface water connectivity; they range in scales from local to regional and internal to external controls, and may only be apparent under certain climatic conditions or across longer study periods and across a range of lake types.
3. Multi-annual landscape hydrologic memory of climate can cause a disproportionate impact on hydrological responses by propagating clusters of wet or dry conditions into extreme events. Spatial differences in memory, driven by hydrogeologic setting, affects how drought conditions, or wet years, amplify the magnitude of landscape drying or wetting, thus providing vital information for understanding system sensitivity (magnitude of response), duration of recovery and in turn infer resilience. Consequently, it is difficult to directly define or infer hydrological processes from analysis that spans multiple climate cycles in landscapes with capacity for high memory, while simultaneously, it is impossible to gain a full understanding of a system's hydrologic function without observing response under a range of climatic conditions where non-linearities exist. Instead, shorter term interannual or intra-annual analyses should be placed within the context of the long-term climate, where comparisons of responses at different antecedent conditions will provide stronger evidence of landscape memory and associated hydrological processes.

4. Reduced-complexity numerical modelling founded upon the HRA-HU conceptual framework provides sufficient functionality to accurately represents seasonal and interannual lake-peatland-forestland water tables, lateral fluxes and atmospheric exchange over decadal time scales by simulating dominant controls and processes, while also providing high versatility in order to represent multiple diverse landscape configurations due to low parametrisation. Within fine textured catchments, peatlands provides a crucial driver in both spatial and temporal variation in landscape runoff generation and lake water level variability through AET regulation, with shorter runoff return periods in peatland dominated systems, whereas greater groundwater contributions associated with higher recharge and higher hydraulic conductivity of surficial geology maintains lake levels and water tables in coarse textured systems.

This research also presents important findings which have major consequences for the way studies approach hydrological and hydrochemical analyses in the BP:

1. Interannual variability is an artefact of study length in systems where interannual variability exceeds intra-annual variability. Due to the presence of interacting seasonal and decadal climate cycles, and spatial differences in landscape memory of climate, lake water level responses to multi-annual climate cycles transpire over different periods. Landscapes dominated by ET, low lateral connectivity and low storage mimick the immediate climate patterns, whereas systems with complex groundwater connectivity with regional flows and/or local organic groundwater sources can result in cumulative responses that only transpire over a full decadal climate cycle. Consequently, short term studies may be presenting spatial differences in hydrologic patterns as a product of differences in lag response to climate, rather than different sensitivity (short term magnitude response) to climate, and cannot use these observations to directly infer resilience.

2. Metrics commonly applied to analyse predominantly intra-annually variable systems such as long-term averages or ranges will compound interpretation in systems of greater interannually variability. In systems dominated by seasonal climate cycles, single value metrics that summarise within season averages will expose spatial differences in hydrologic function. However, applying such metrics where strong decadal climate cycles cooccur with multi-annual memory will mask longer term variability and will either produce misleading results as an artefact of study length, or if the study captures responses to a full climate cycle, could collapse spatial differences in long-term absolute variability, thus high spatial variability in lake water level patterns may only become apparent when examining temporal dynamics.
3. In landscapes where evapotranspiration and groundwater both present large but spatially varying components of the water balance, water chemistry such as stable isotopes and concentrations of major cations and anions could lead to incorrect conclusions if used as a proxy for groundwater connectivity. In lakes that receive low groundwater inputs and thus low nutrient inputs, the evaporative signature is sufficiently strong and flushing sufficiently irregular to produce comparable concentrations through evapoconcentration to those systems where groundwater contributions dominate the lake water balance. Similarly, systems dominated by regional groundwater connectivity can present comparable isotopic compositions to hydrologically isolated surface water basins due to high in-lake enrichment processed in large coarse out-wash lake ‘evaporation windows’, while frequent resetting of isotopic composition in small shallow isolated lakes, either through desiccation or their volumes are sufficiently low for rainfall to have a significant impact. Thus, using isotopic depletion as an indicator for high groundwater connectivity could be erroneous. We demonstrate how pairing hydrochemical analyses with hydrological monitoring provides a much clearer understanding of sources and magnitudes of groundwater contributions in highly enriched landscapes.

3. Synthesis

Lake water levels fluctuate directly as a function of individual water budget components and lake bathymetry (Hayashi and Van Der Kamp, 2000; Minke, Westbrook, and Van Der Kamp, 2010). Runoff contributions from the catchment are governed by the interactions of geology, landcover, and landscape position with climate (Chapter 3), thus metrics characterising lake water level fluctuations collapse spatial catchment heterogeneity, non-linearity of runoff generation and complex hydrological processes into simpler and more readily available storage observations (Chapter 2; Carey *et al.,* 2010; Mcnamara *et al.,* 2011; Shaw *et al.,* 2012). The research presented within this thesis demonstrates that cumulative memory of climate within the landscape is a significant driver of lake water level patterns (Shook and Pomeroy, 2011; Wells *et al.,* 2017) and indicates the possibility for multi-year climate memory and hysteresis loops that form over multiple years (Chapter 4). Accurate representation of antecedent conditions within numerical models is crucial as proximity to saturation thresholds results in disproportionate effects of storage increase on discharge (Spence, 2010), further to dynamic connectivity between wetlands modifying contributing area (Shaw *et al.,* 2012). This impact of memory emphasises the need to understand static controls (landscape characteristics) and dynamic controls (antecedent moisture) (Phillips, Spence, and Pomeroy, 2011), so that increased uncertainty under extreme conditions does not generate misleading statistical interpretations of water level patterns (Koutsoyiannis and Montanari, 2007) and erroneous inference of resilience (Bayley and Prather, 2003; Scheffer *et al.,* 2001; Cobbaert, Wong, and Bayley, 2014).

Storage metrics have been employed for catchment classification (Wagener *et al.,* 2007; Sayama *et al.,* 2011) to divide catchments into units of varying physiography representing differing storage capacity, to better facilitate fill–spill runoff analyses when coupled with precipitation amounts and antecedent conditions (Shaw *et al.,* 2012). Hydrologic Response Areas (HRAs; Devito *et al.,* 2017) delineated by surficial geology provide an exceptional classification tool for the physical characteristics of boreal

catchments (Chapter 2), and also produce distinct hydrological patterns and homogenous isotopic and chemical compositions (Hokanson *et al.,* 2019; Figure 6-2). Subsequently, we demonstrate that no single dominant control on lake hydrology exists within the BP (Chapter 3; Gibson *et al.,* 2016; Gibson, Prowse, and Peters, 2006) and that landscape-scale wetting and drying cycles occur at contrasting rates, magnitudes and periodicities in response to seasonal and decadal climate signals (see Chapters 2 and 4). This results in significant spatial variability in the propensity for drought years or wet years to persist or accumulate into extreme landscape drying or wetting states (Devito *et al.,* 2012). Consequently, research into landscape controls must be placed into context of landscape antecedent conditions and position within climate cycle, in addition to conducting analyses within a framework of BP hydrologic landscapes (Winter, 2001; Wolock *et al.,* 2004) headed by a priori knowledge of the controls of surficial geology and their interactions with landcover and landscape position (Devito, Creed, and Fraser, 2005).

Storage dynamics are largely determined by surficial geology texture (Plach *et al.,* 2016) and the proportion of peatland to forested upland cover (Chapter 3), as contrasting evaporative demand and hydrologic function of lakes, peatlands and forestlands (Van Der Velde, Lyon, and Destouni, 2013) interacts with deep heterogeneous surficial deposits of contrasting storage and transmission properties (Plach *et al.,* 2016; Hokanson, Mendoza, and Devito, 2019). Coarse outwash lakes receive larger-scale groundwater flow due to high hydraulic conductivity of quartz sand mineral deposits (Smerdon *et al.,* 2005); whereas groundwater inputs are constrained to near-surface flow from organic within the fine-textured deposits of clay plains and hummocky moraines, where soils hydraulic conductivity is orders of magnitude lower (Hokanson *et al.,* 2019). These insights coupled with storage-water balance models can be used to simulate long-term water-level dynamics (Price *et al.,* 2005) (Chapter 5).

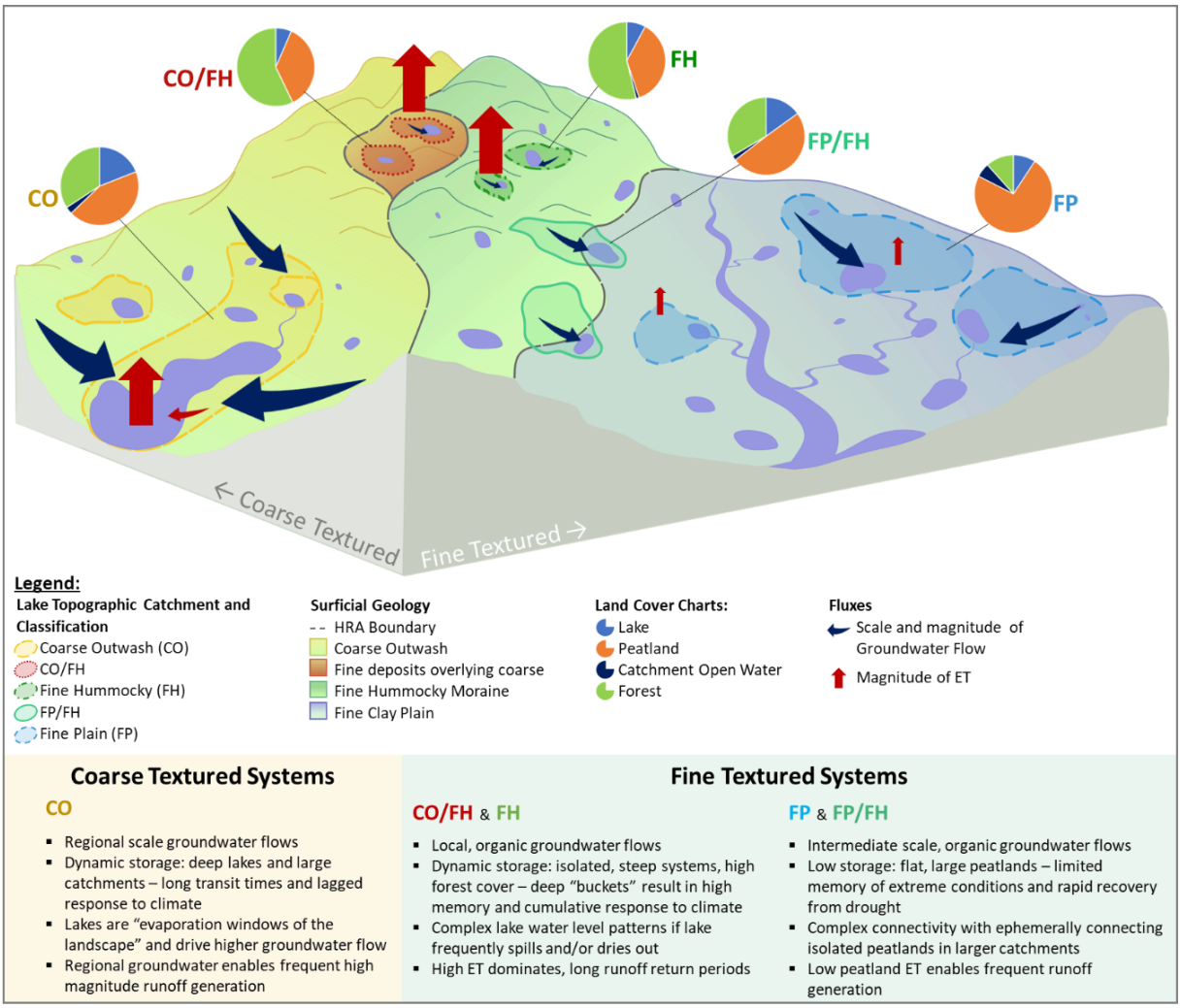


Figure 6-1. Graphical representation of hydrologic response areas (HRAs) within the boreal plains, which incorporates important landscape parameters (surficial geology, typical land cover proportions, topography) for defining hydrologic processes and dominant fluxes (scales and sources of groundwater connectivity, evapotranspiration component of the water balance, storage capacity).

Despite clear empirical evidence of mechanisms controlling the scales and sources of flow (Chapter 3), lake water level response to the long-term climate signal is complex due to interactions with spatially diverse landcover, lake geometry, landscape position, and geology, which further modify hydrological connectivity, landscape memory of climate. Numerical simulations enable greater understanding of system behaviour and sensitivities through model interrogation and sensitivity analyses (Chapter 5; Brunner and Simmons, 2012; Doble and Crosbie, 2017), and corroborates that increased peatland landcover generates greater and more frequent runoff than forestlands (Tetzlaff, Seibert, and Soulsby, 2009; Karlsen *et al.,* 2016b; Van Der Velde, Lyon, and Destouni 2013). Subsequently, peatland

resistance to drought through autogenic feedback mechanisms (Dixon *et al.,* 2017; Kettridge and Waddington, 2014; Waddington *et al.,* 2015) provides vital sources of water that enable shallow lakes to persist within sub-humid boreal landscapes (Chapter 5).

4. Future Research Avenues

4.1. Conceptual framework for hydrologic landscapes

A key finding of this research stipulates that landscape and lake management approaches should shift from ascribing buffers based upon lake permanence, to targeted protection of the mechanisms that enable different lake types to persist. This is because we show that all lakes exhibited high magnitudes of variability, but lagged or cumulative responses driven by landscape memory of climate meant that these patterns emerged over contrasting timescales and at different time points. Memory driven temporally variable responses to climate proves to explain contradictory results within the literature. Thus, inference of resilience founded solely upon variability metrics to indicate sensitivity to climate is fundamentally flawed. Consequently, future research should generate process-based knowledge that can be placed in the context of longer-term climate cycles and utilise hydrologic frameworks, so that that it can be transposed spatially and temporally. We suggest the following as priority research avenues that researchers may wish to address:

- Shorter term interannual or intra-annual analyses should be placed within the context of the long-term climate, where comparisons of responses at different antecedent conditions will provide stronger evidence of landscape memory and associated hydrological processes. Emphasis should be placed on long-term observations to understand responses in systems where non-linearities transpire over longer time scales, or otherwise research should make a concerted effort to define antecedent conditions and dominant controls on storage dynamics to facilitate study comparisons.

- Combining lake water level monitoring with hydrochemistry to examine mass versus concentration of nutrients provided new insights into the relative controls of varied groundwater pathways and evaporation on lake water levels (Chapter 3). Future research should explore the temporal dynamics of these observations in order to identify thresholds for shifts in connectivity, improve understanding of memory effects and in turn identify the dominant processes that enable lakes to persist within the sub-humid BP.

4.2. Applications of reduced-complexity numerical model

The Oil Sands Region (OSR) is projected to impact 200,000 km^2 of the Boreal ecozone regions of Western Canada (Wells and Price, 2015). Oil sands companies have a regulatory obligation to return the post-mined landscape to landuses with equivalent capabilities to those that existed prior to industrial development (Government of Alberta, 1993). Boreal plains hydrology presents a challenge to conventional topographically driven hydrological models due to the dominance of vertical fluxes over lateral flows, and water tables that do not mirror surface topography due to interactions of the sub-humid climate with deep soils, low topographic relief and high peatland cover. Consequently, hydrological model development for reclamation of boreal ecosystems within the oil sands region is often formulated in regions with very different controls on hydrology, and thus leads to complex translation to oil sands regions. The reduced-complexity numerical model presented within this research provides novel and substantial opportunities to increase exploratory and investigative power through rapidly and efficiently computing multiple model configurations, thus enabling development of more resilient landscapes in the face of climate change and anthropogenic impacts on hydrologic connectivity.

While the benefits of the model for land managers and stakeholders are clearly evident, the model also presents many avenues of opportunity for research. Within the bounds of the current model presented within this work, we suggest the following as future research efforts to address:

- sensitivity analyses and interpretation of all hydrological fluxes to pinpoint both spatial thresholds and temporal climatic tipping points within hydrologic response, which will further improve understanding of system response in ungauged basins, and recognise potential sensitivities under future climates.
- evaluate ecosystem vulnerabilities through the periodicity and magnitude of exceedance of important ecohydrological or hydrochemical thresholds, such as lake desiccation implications for wildlife, water table depths for tree mortality, or risk of wildfire.

The inclusion of just enough functionality to represent fundamental controls has produced a versatile, physically based numerical model that provides the foundations for efficient and wide-reaching predictive capacity across a multitude of hydrology-related applications within boreal, peatland dominated and/or low-relief dominant interior landscapes. Therefore, the model presents potential to be extended to simply incorporate and address the following:

- ecohydrological feedback mechanisms, such as plant dieback in response to plant available water, or landscape evolution and tree growth following reclamation, and subsequent impacts on rainfall partitioning into AET and runoff.
- simulate peatland carbon accumulation or loss as a function of water table depth and assess future risk to peatland and boreal forest carbon stores.

5. Final Remarks

The research presented within this thesis has advanced understanding of the dominant controls on boreal hydrology and how they vary in time and space within the context of an oscillating climate and

a heterogeneous hydrogeologic setting. This new knowledge has been implemented within a versatile numerical model, which contributes significantly towards informing how different landscape configurations respond to climatic forcing by enabling higher interrogation of system behaviour and subsequent sensitivities. As such, this research contributes substantially to informing land managers on the role of sinks and sources of water within the BP landscape and provides both knowledge and tools to manage for more resilient shallow lakes and boreal landscapes.

6. References

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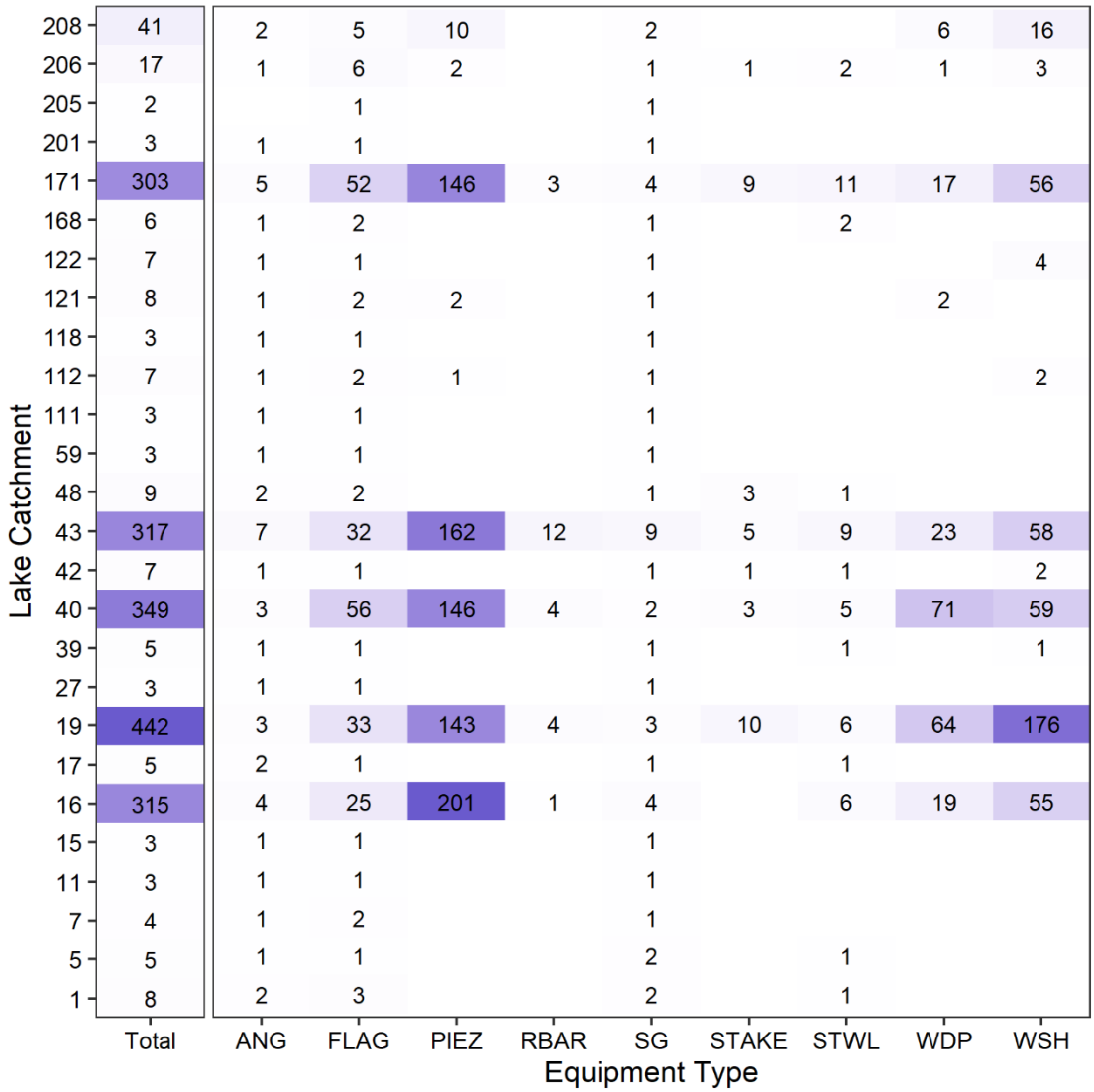
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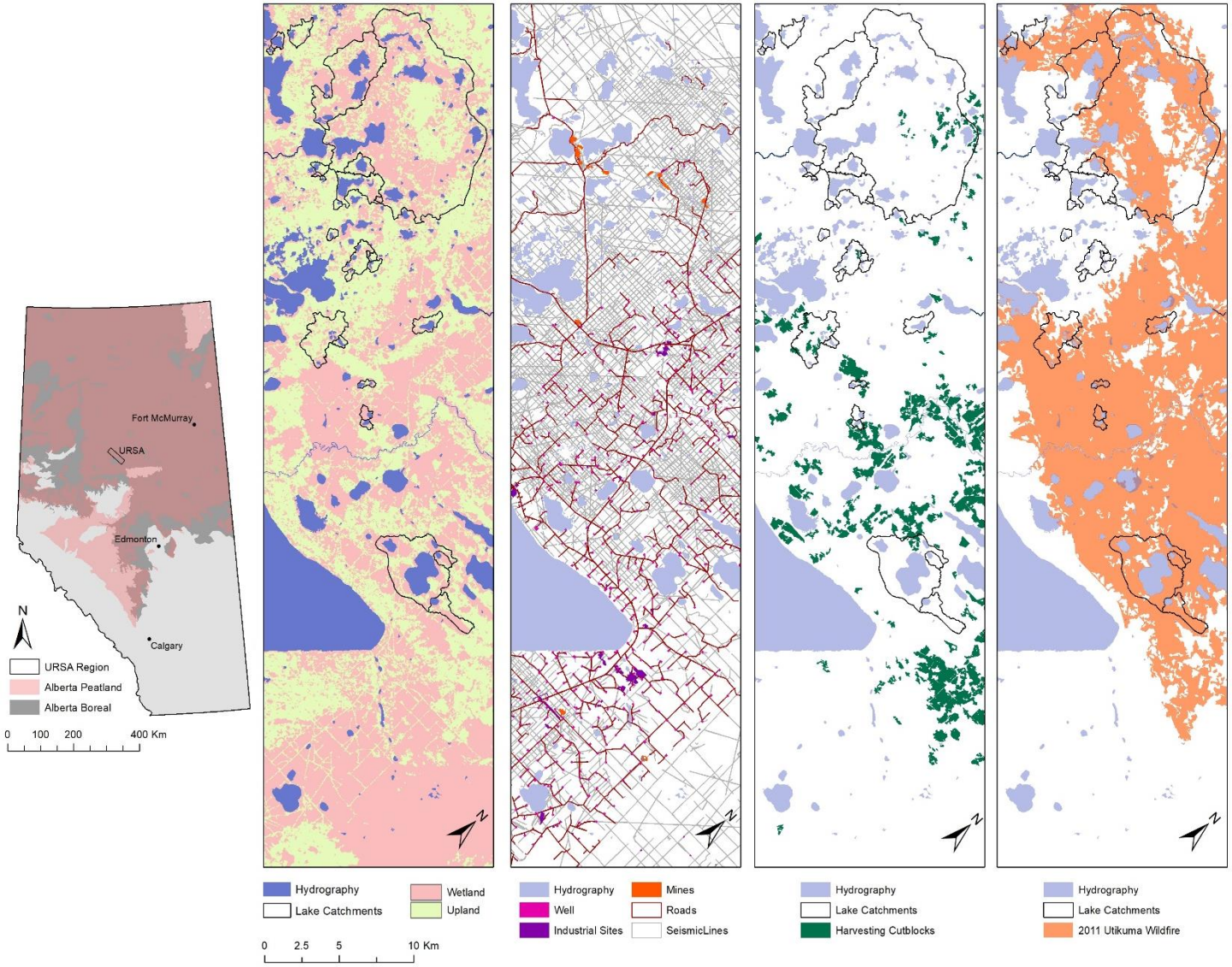
APPENDIX

URSA Catchment Mapping and Hydrological Monitoring Network

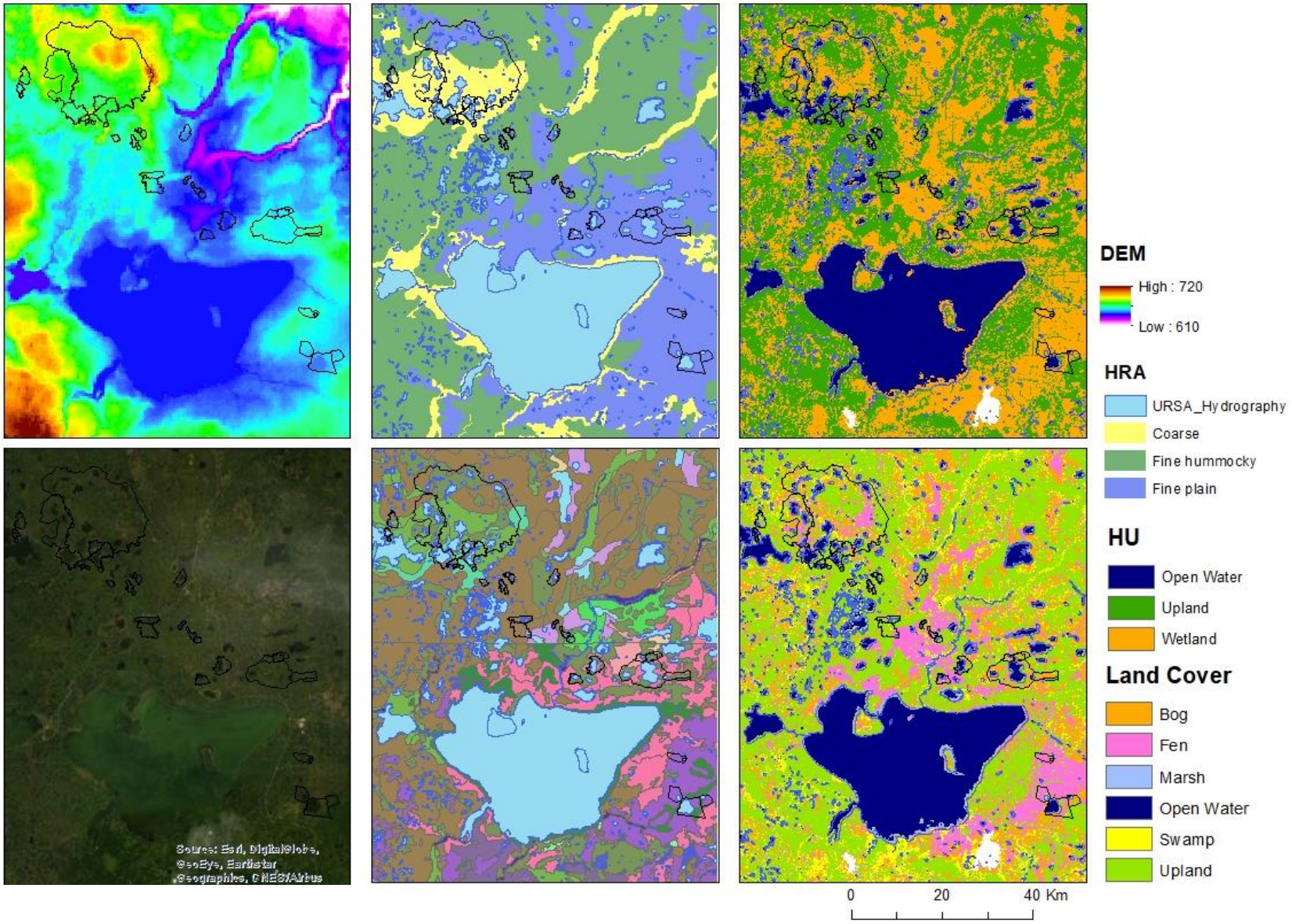
This appendix presents additional information on the URSA study lakes and their catchments including the extent of the hydrological monitoring network, and the spatial extent of disturbance, including human footprint variables (mining, harvesting, roads, wells and seismic lines; ABMI, 2020) and the extent of the 2011 Utikuma wildfire. A collection of 6 additional maps is presented for each of the 26 lake catchments and URSA to present the distributions of surficial geology, HRAs, land cover, HUs, topography and satellite imagery (ESRI, 2020).



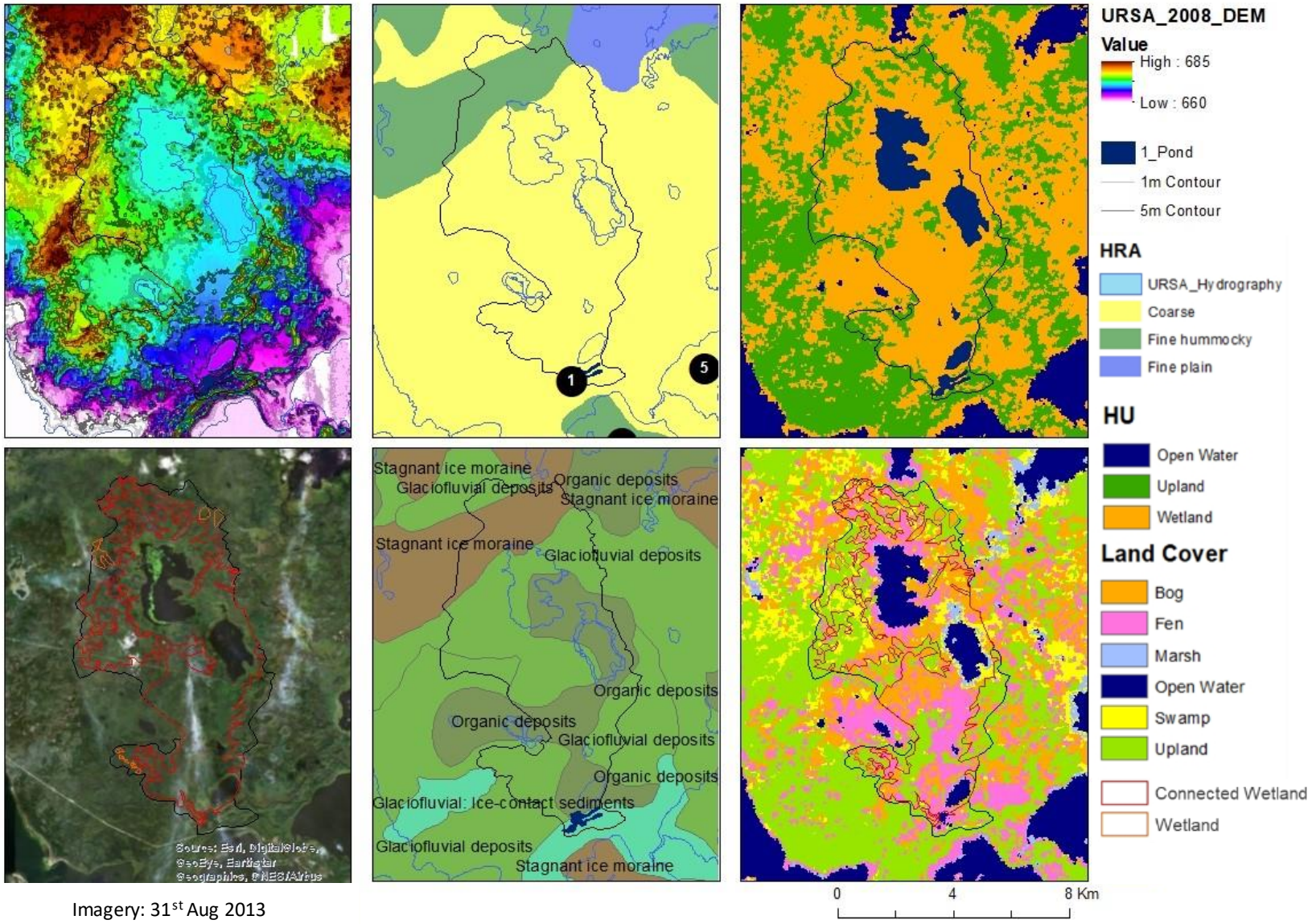
Appendix Figure 1. Number of hydrological monitoring sites and their equipment type per lake catchment. Lake level monitoring includes staff gauges (SG), angle irons (ANG), rebar (RBAR), staked sites (STAKE), temporary flagged sites (due to extreme water levels; FLAG) and some stilling wells (STWL). Groundwater monitoring includes stilling wells (STWL), deep wells (WDP) shallow wells (WSH) and piezometers (PIEZ) all of varying equipment depth.



Appendix Figure 2. URSA distribution of peatlands, forestlands and open water and the extent of disturbance, including anthropogenic landuse change due to industrial activities, and the extent of the Utikuma 2011 wildfire.

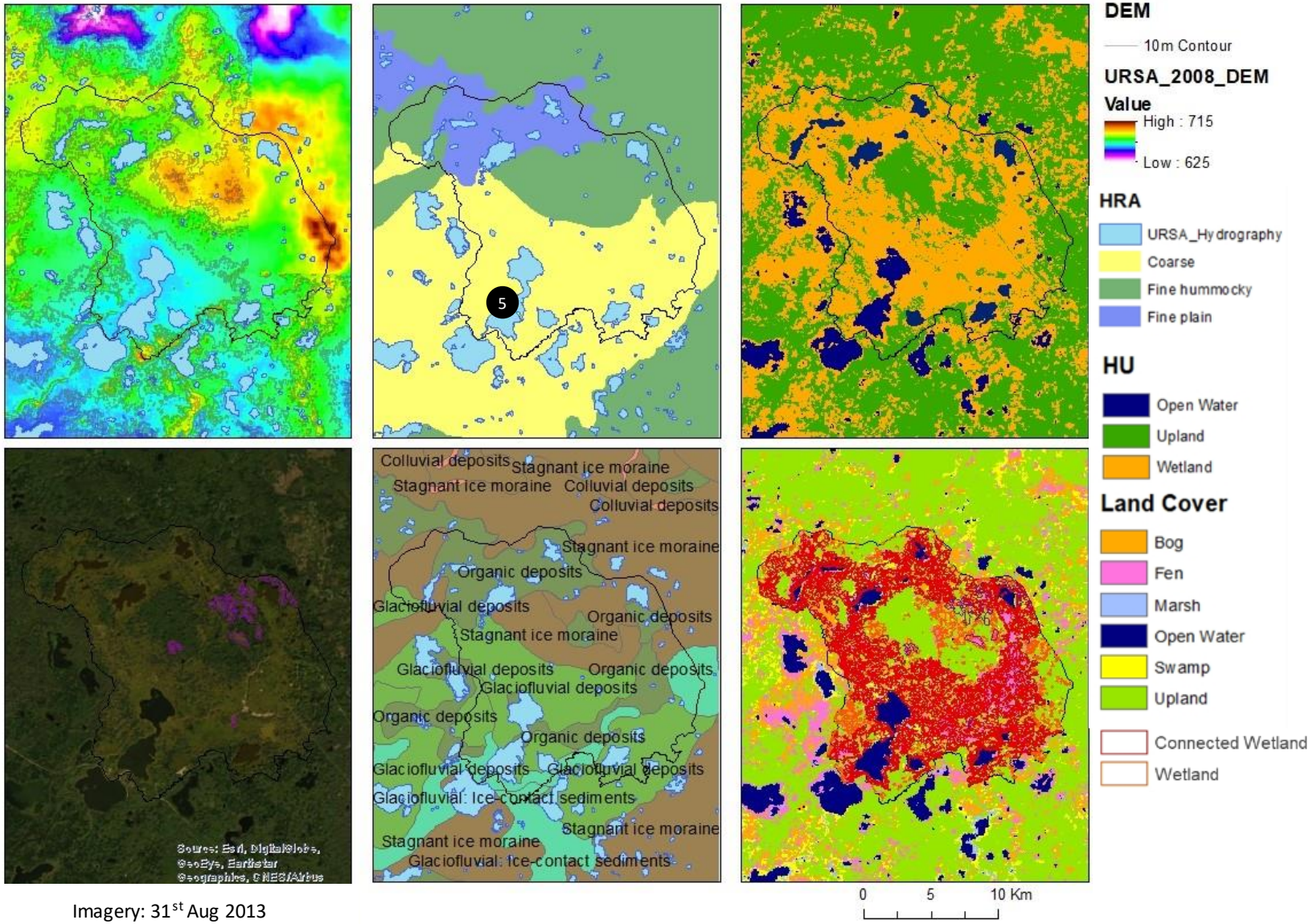


Appendix Figure 3. URSA topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



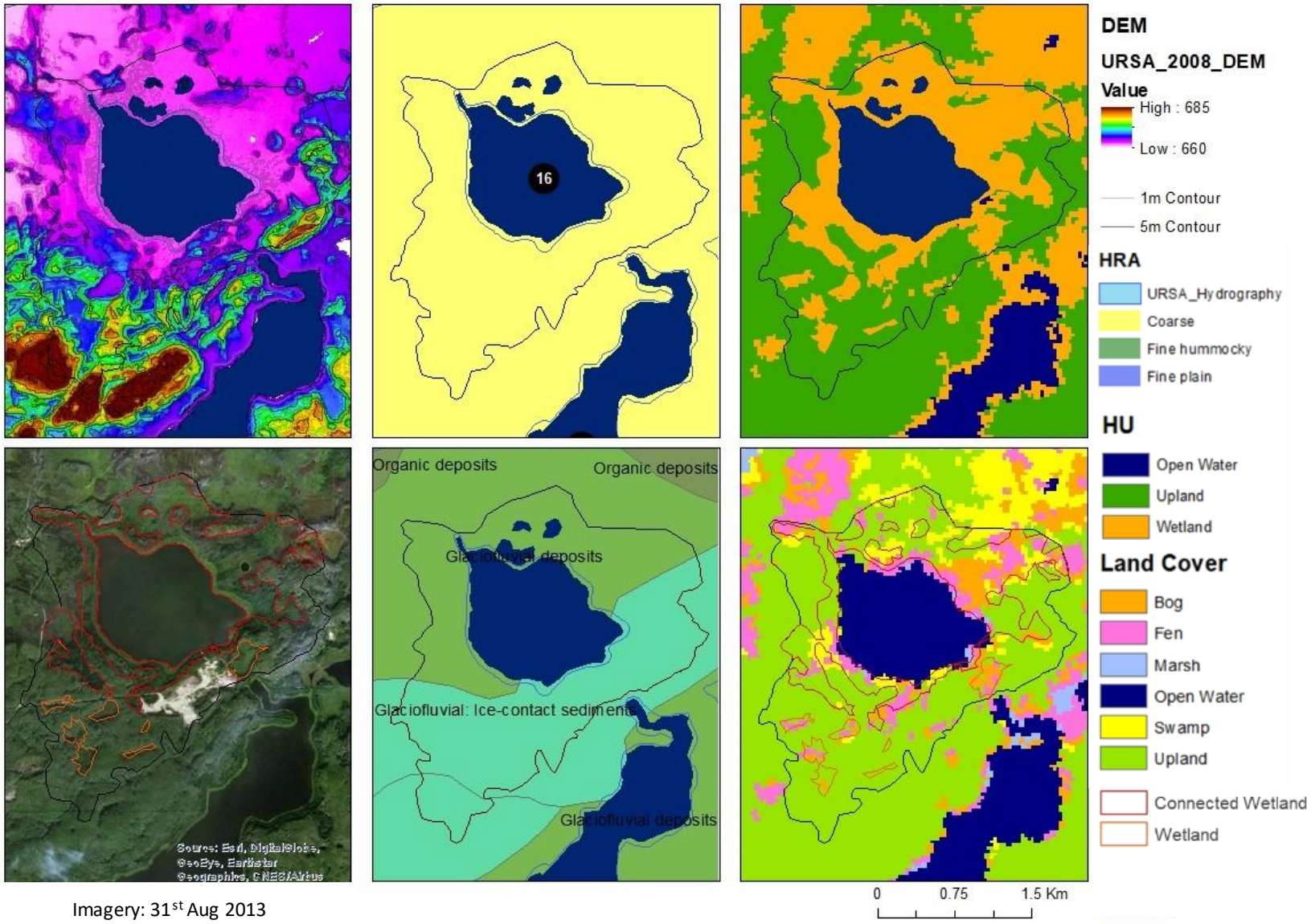
Imagery: 31st Aug 2013

Appendix Figure 4. Lake 1 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



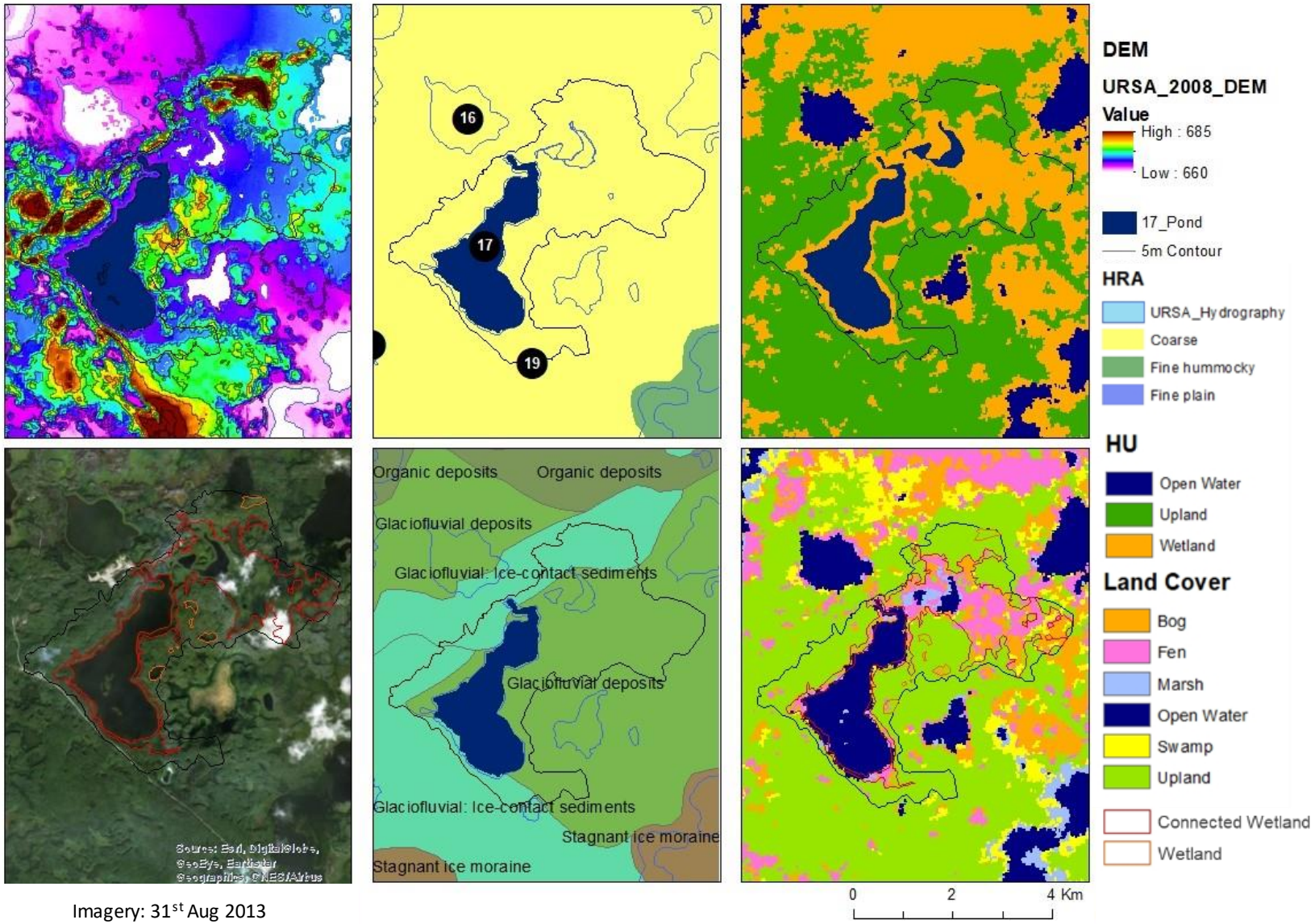
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Appendix Figure 5. Lake 5 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



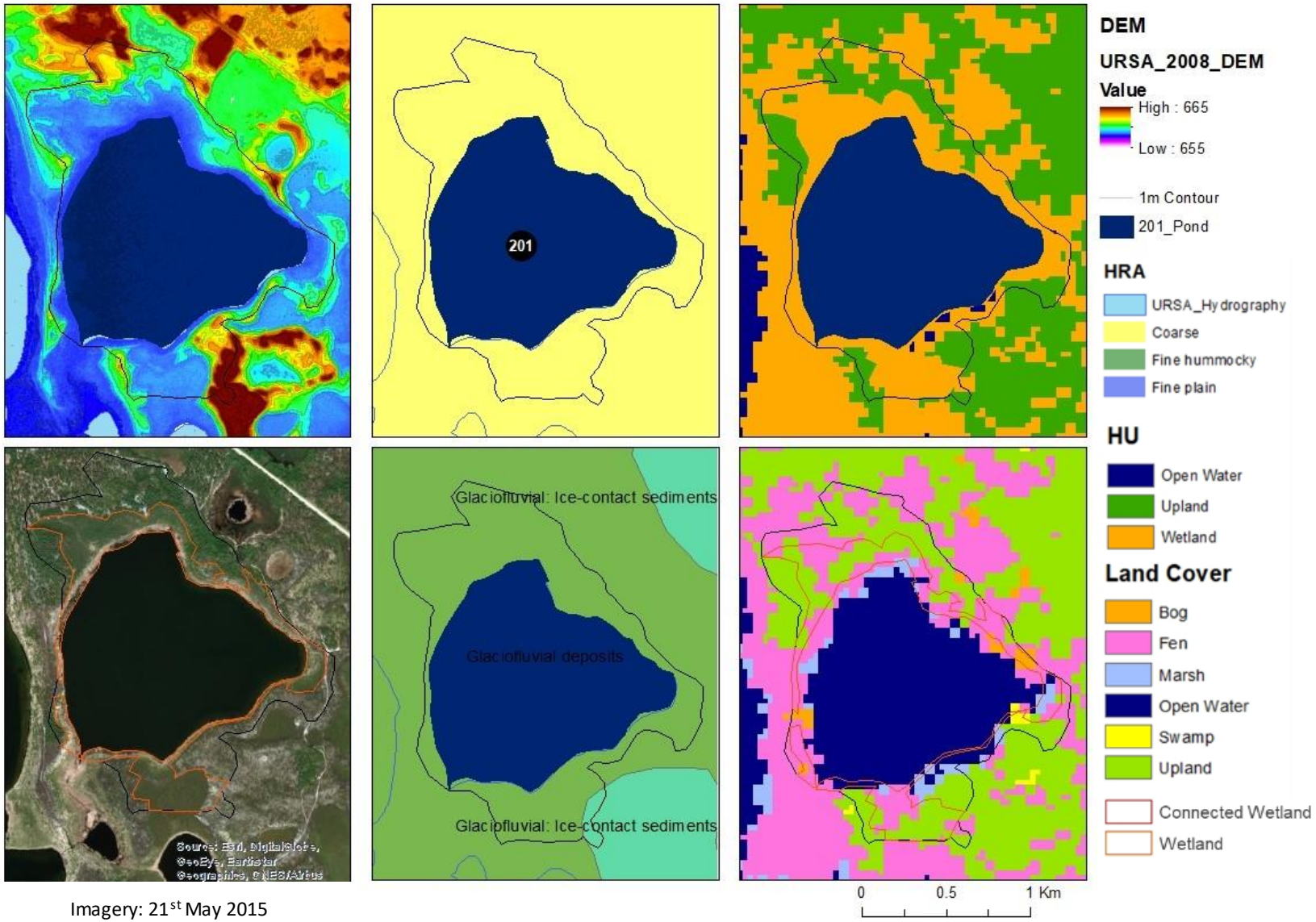
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Appendix Figure 6. Lake 16 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



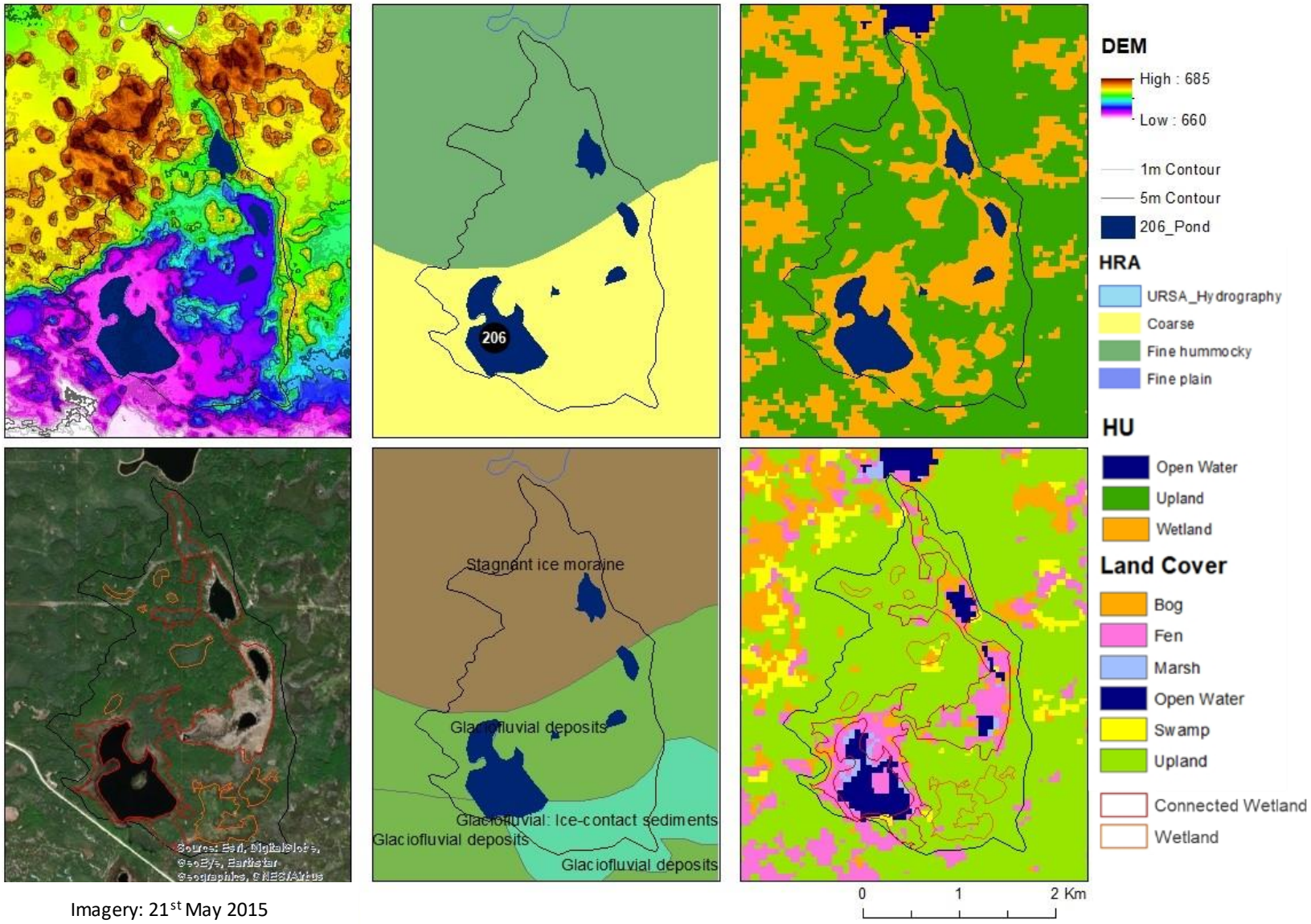
Imagery: 31st Aug 2013

Appendix Figure 7. Lake 17 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



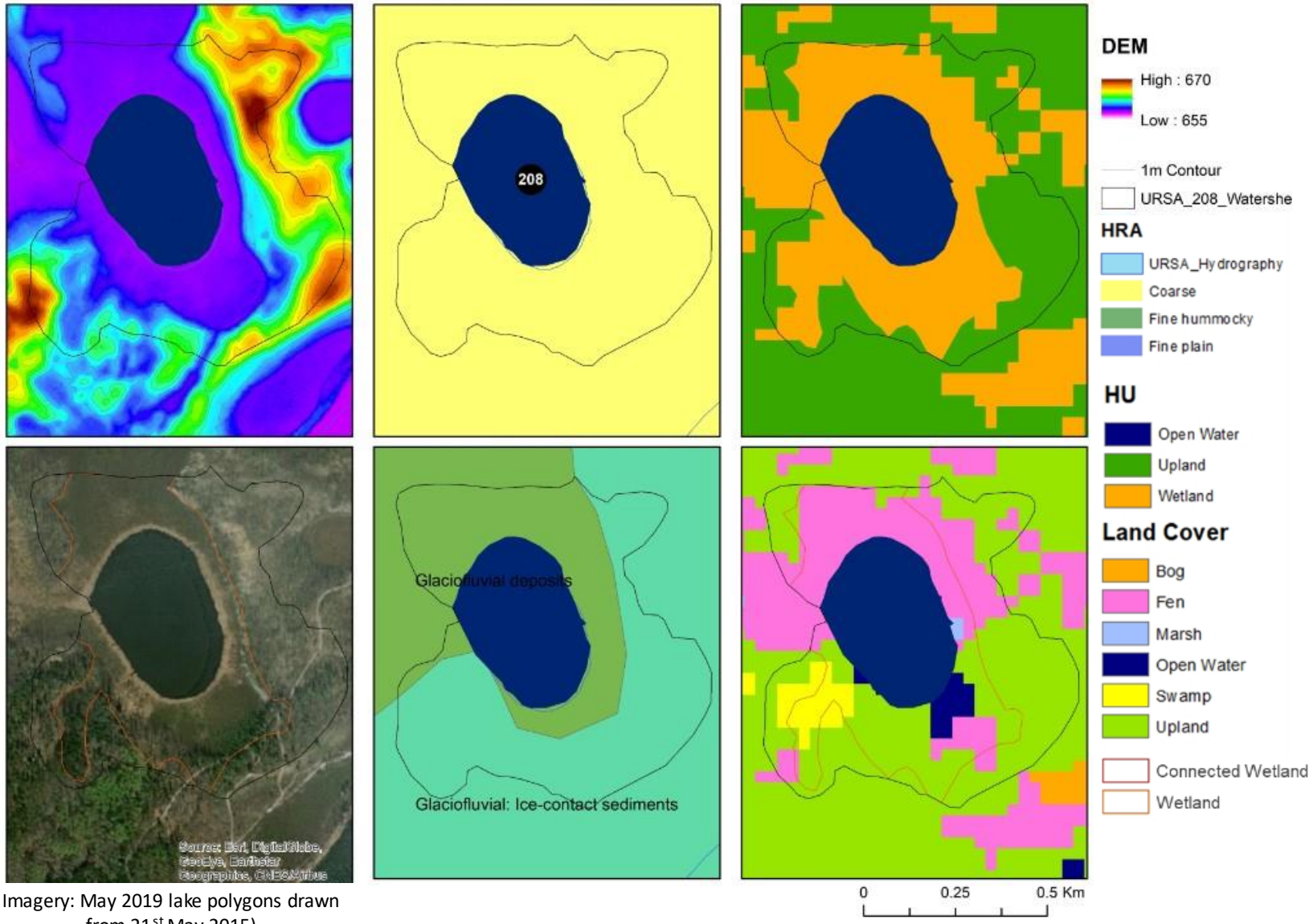
Imagery: 21st May 2015

Appendix Figure 8. Lake 201 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



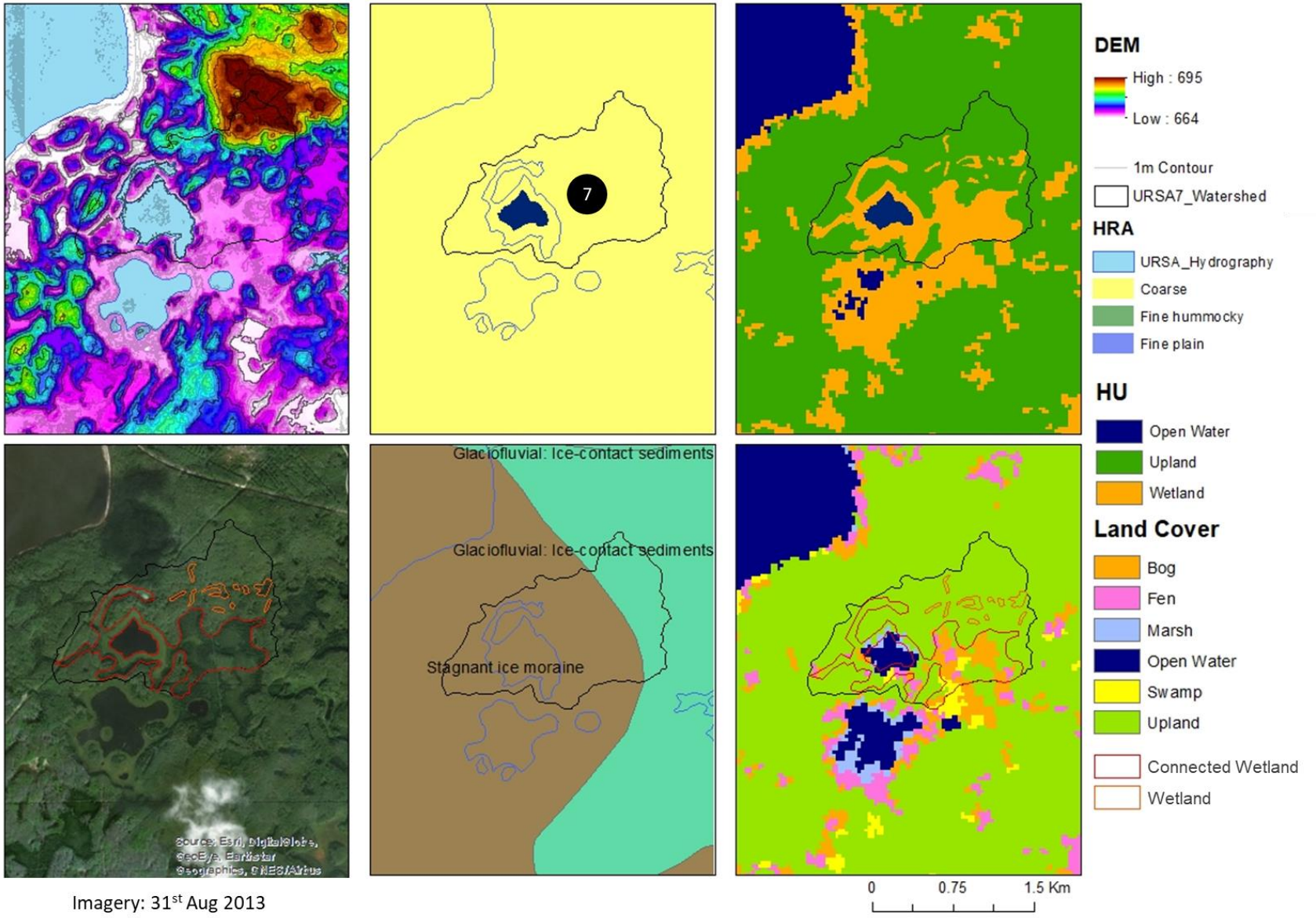
Imagery: 21st May 2015

Appendix Figure 9. Lake 206 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.

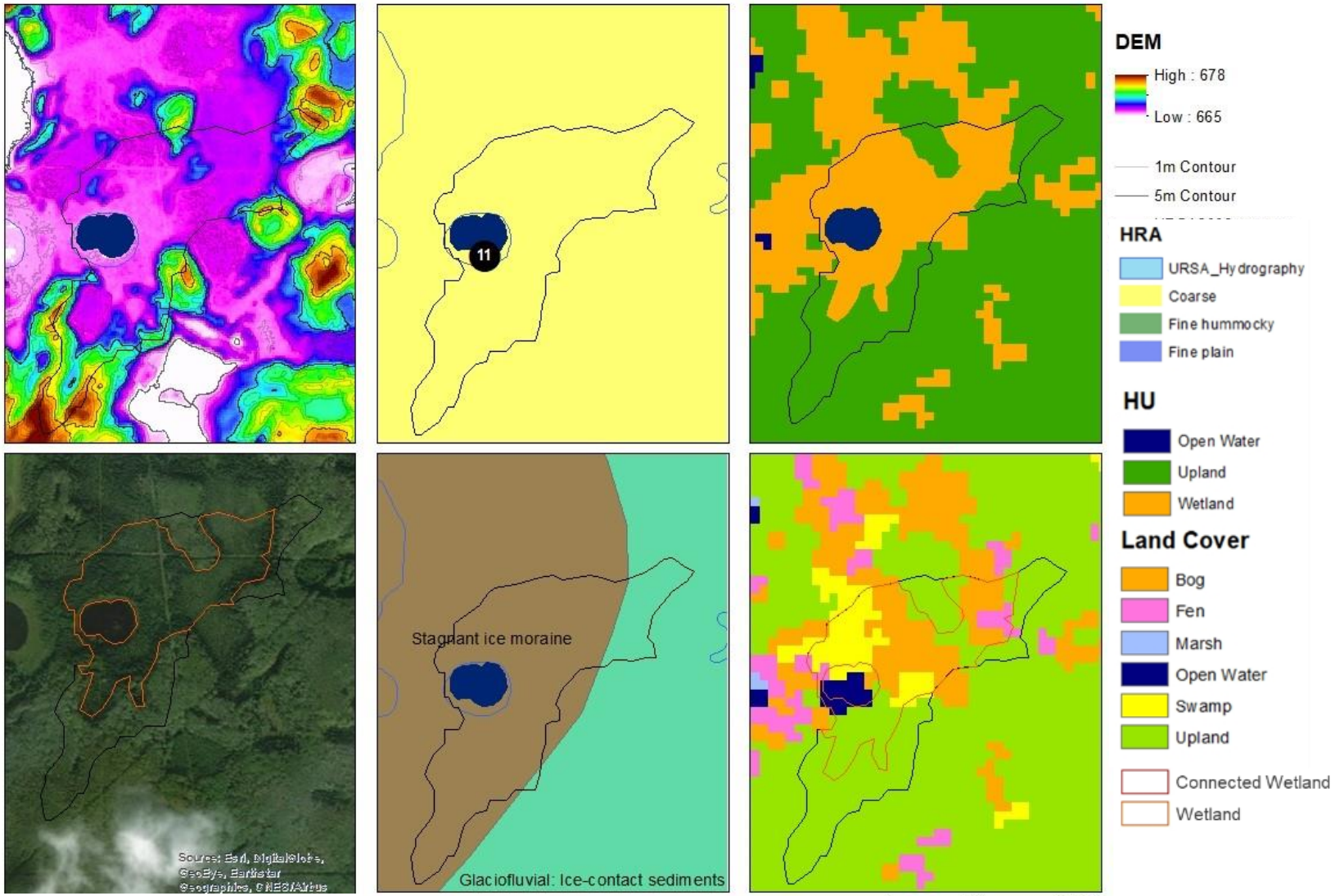


Imagery: May 2019 lake polygons drawn from 21st May 2015)

Appendix Figure 10. Lake 208 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.

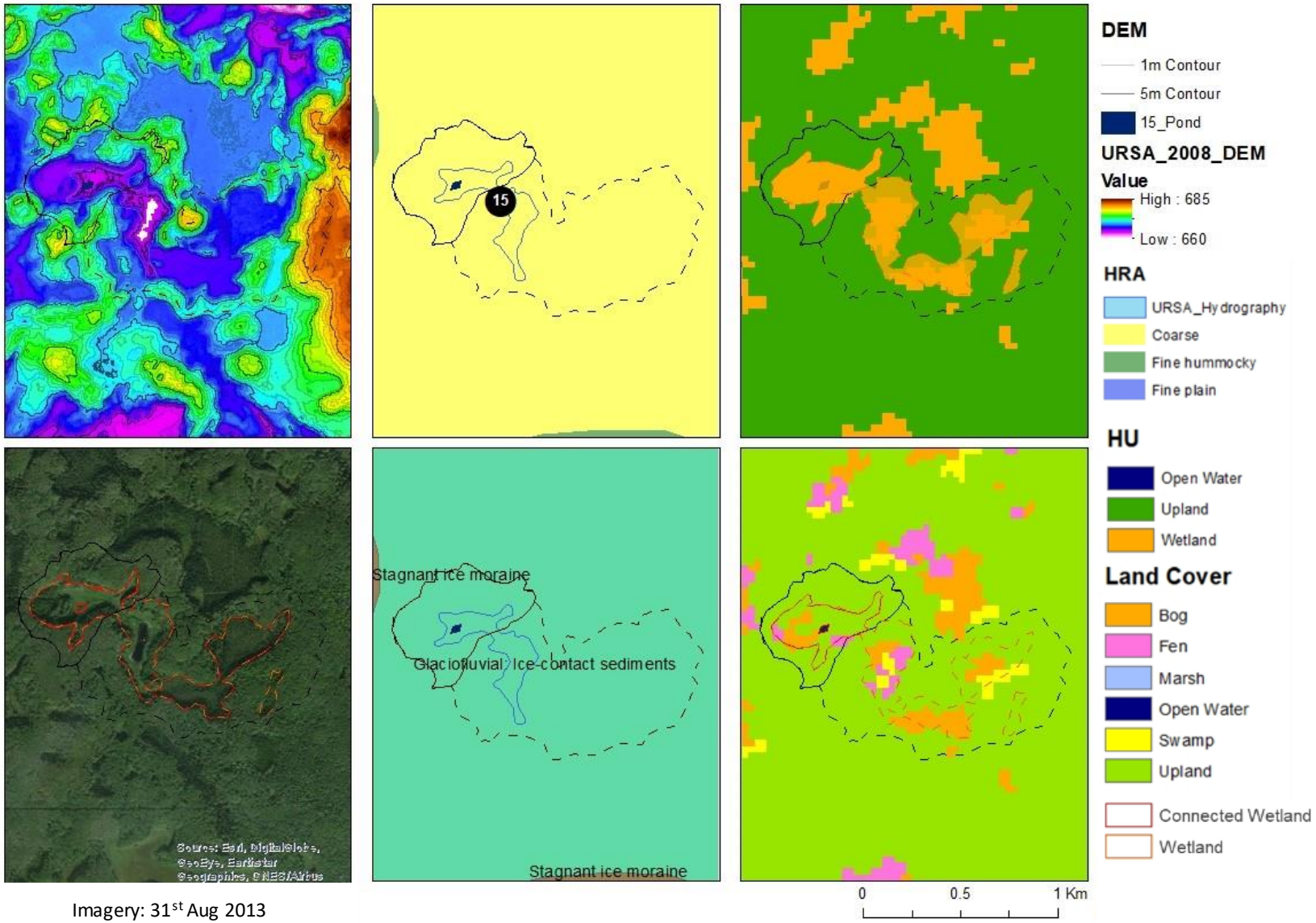


Appendix Figure 11. Lake 7 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



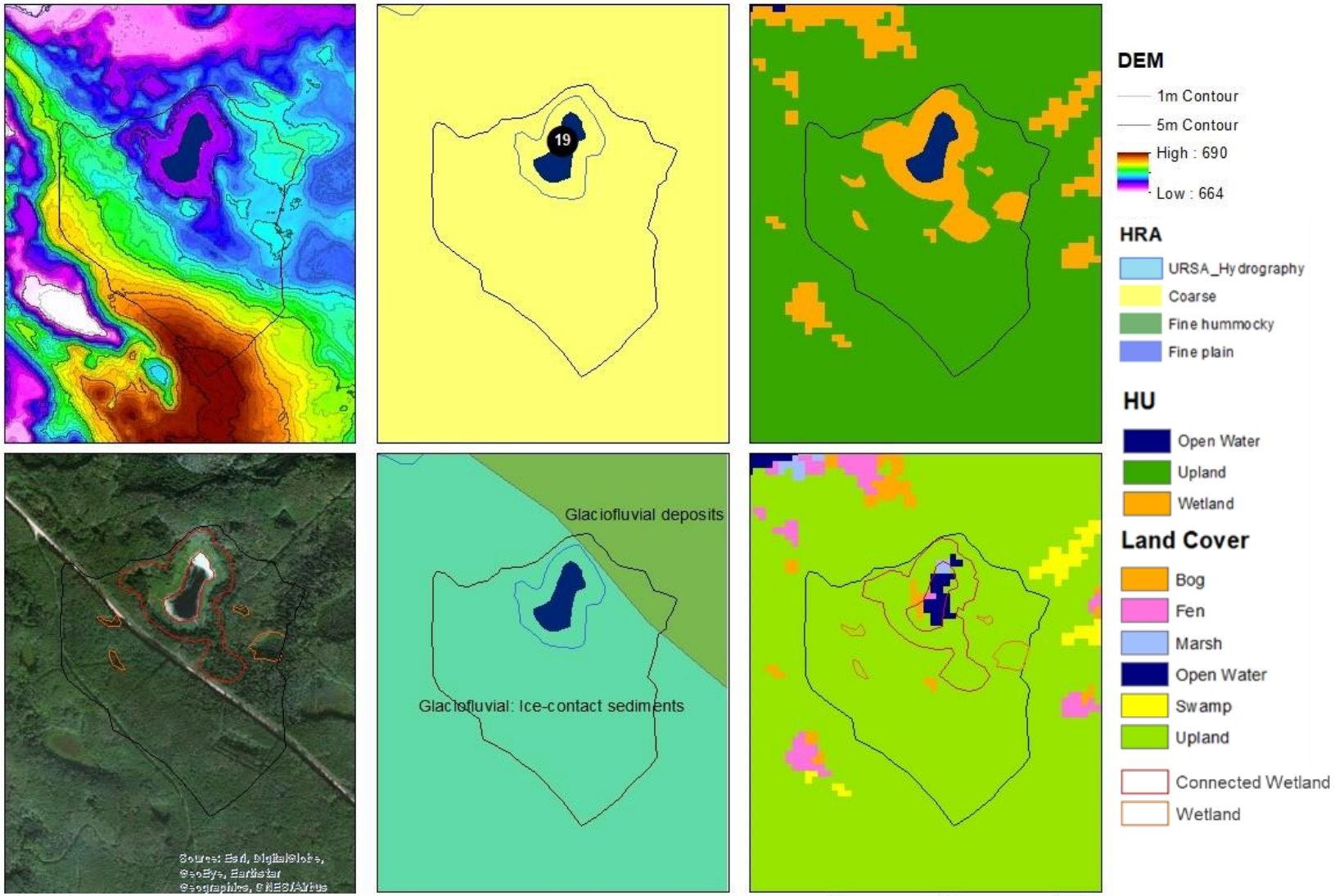
Imagery: 31st Aug 2013

Appendix Figure 12. Lake 11 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



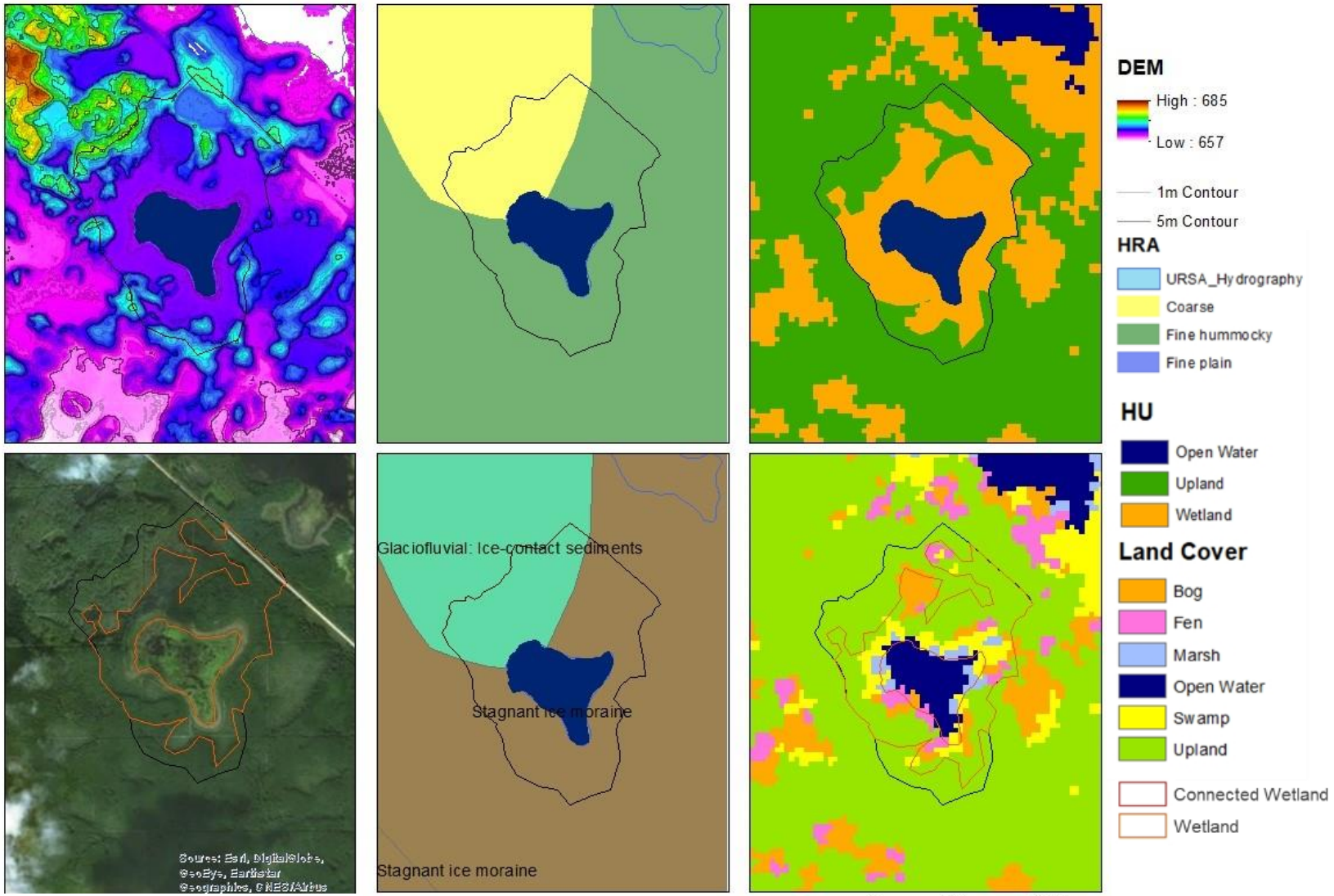
Imagery: 31st Aug 2013

Appendix Figure 13. Lake 15 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



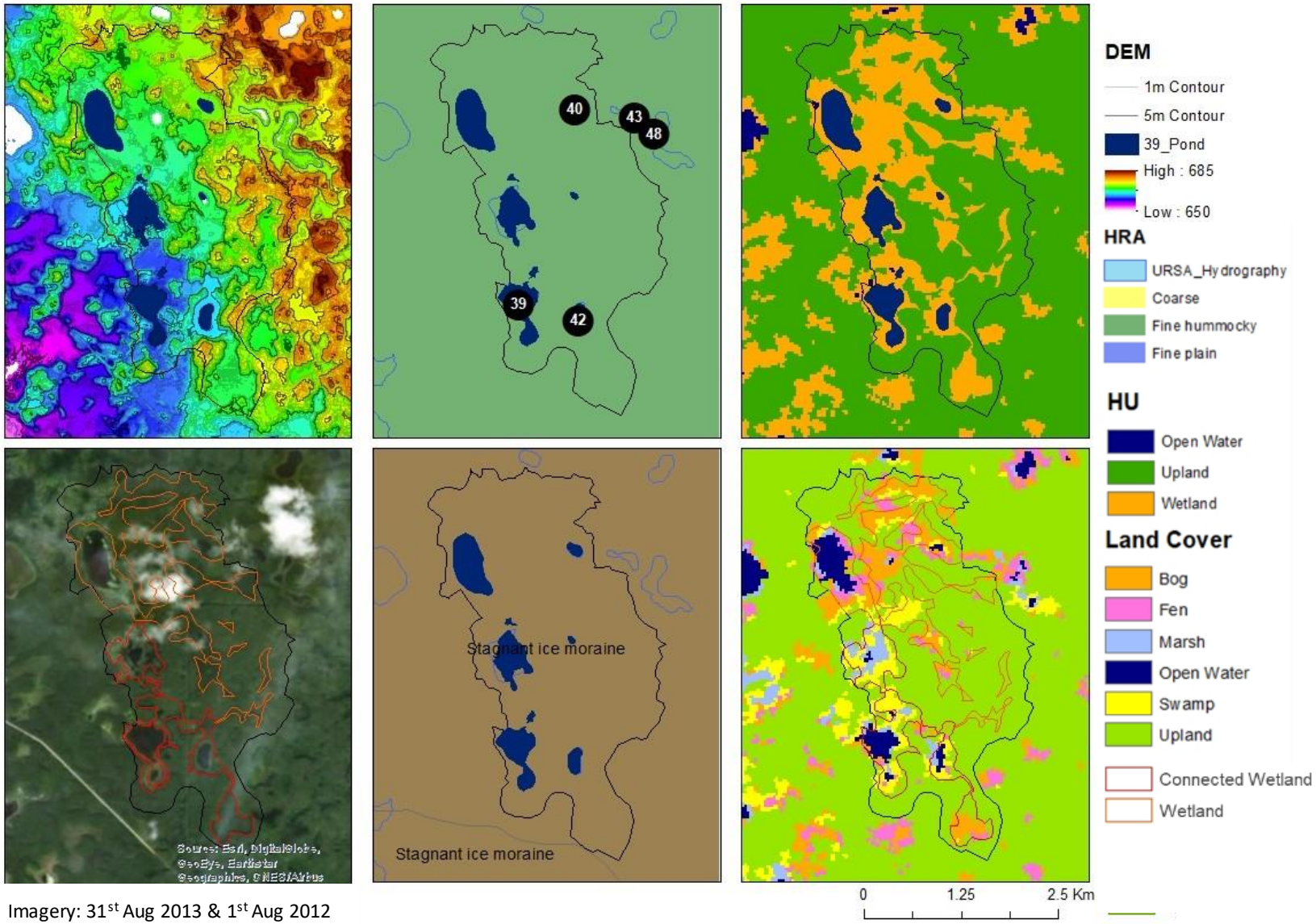
Imagery: 31st Aug 2013

Appendix Figure 14. Lake 19 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



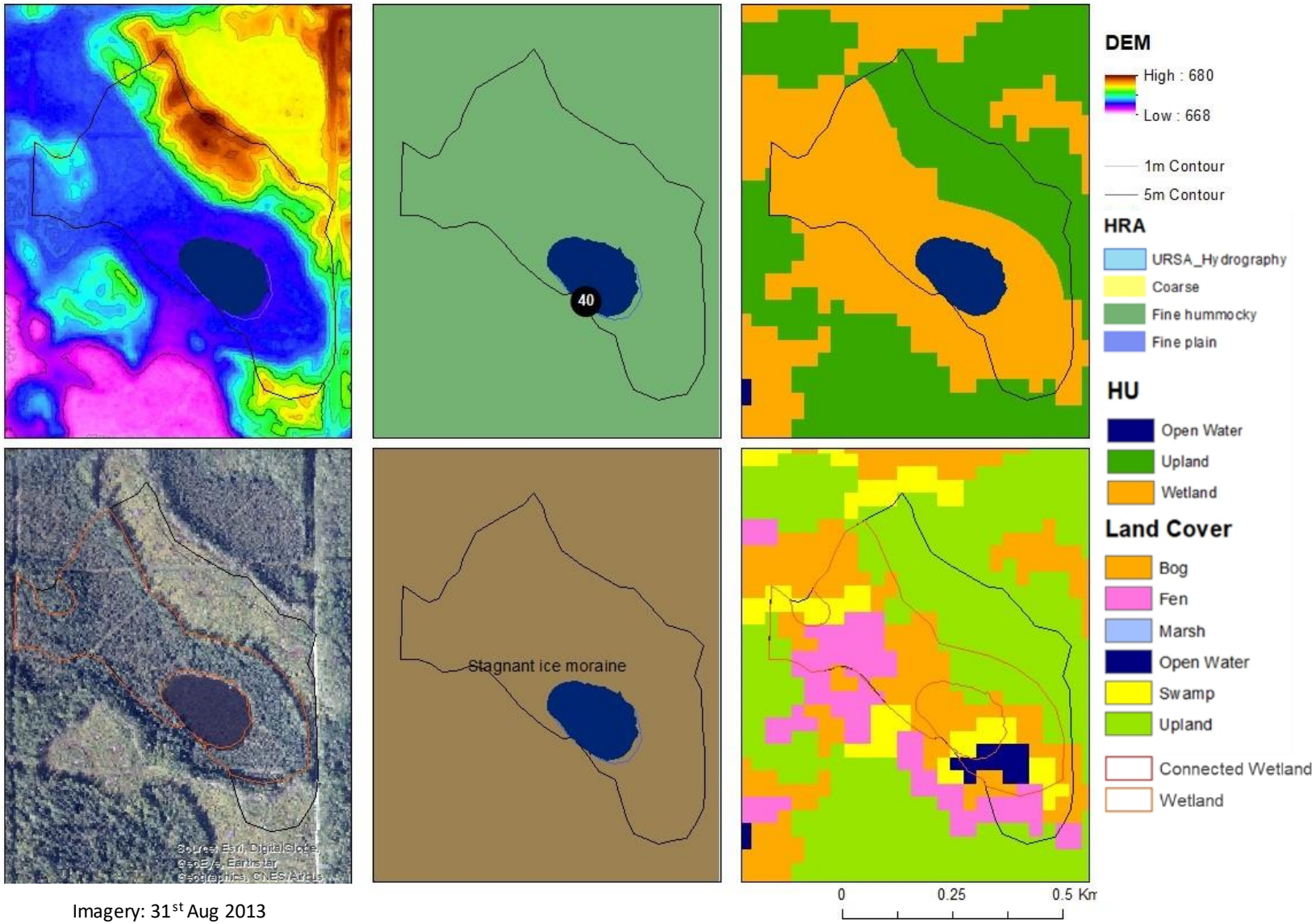
Imagery: 31st Aug 2013 & 1st Aug 2012

Appendix Figure 15. Lake 27 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



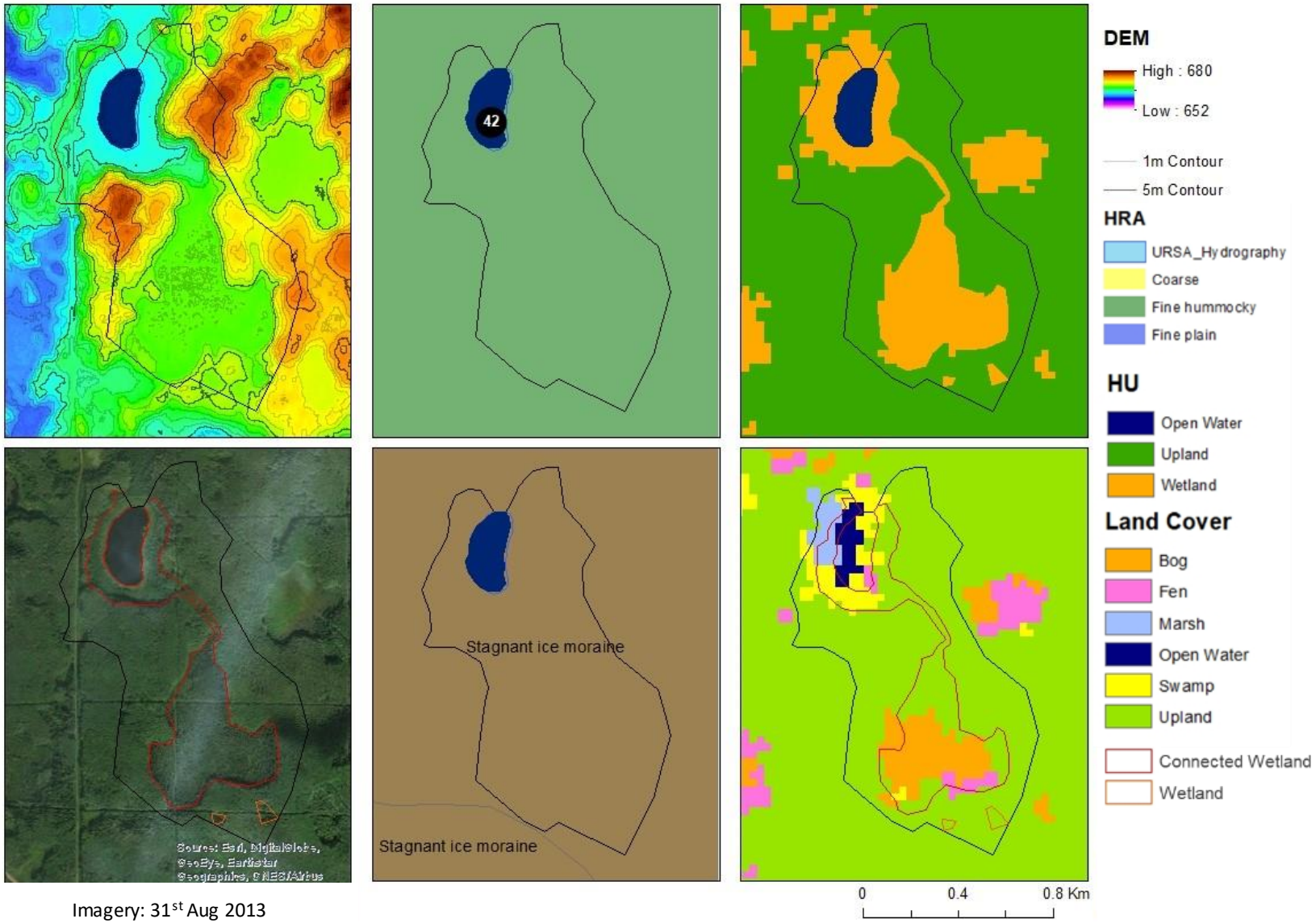
Imagery: 31st Aug 2013 & 1st Aug 2012

Appendix Figure 16. Lake 39 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



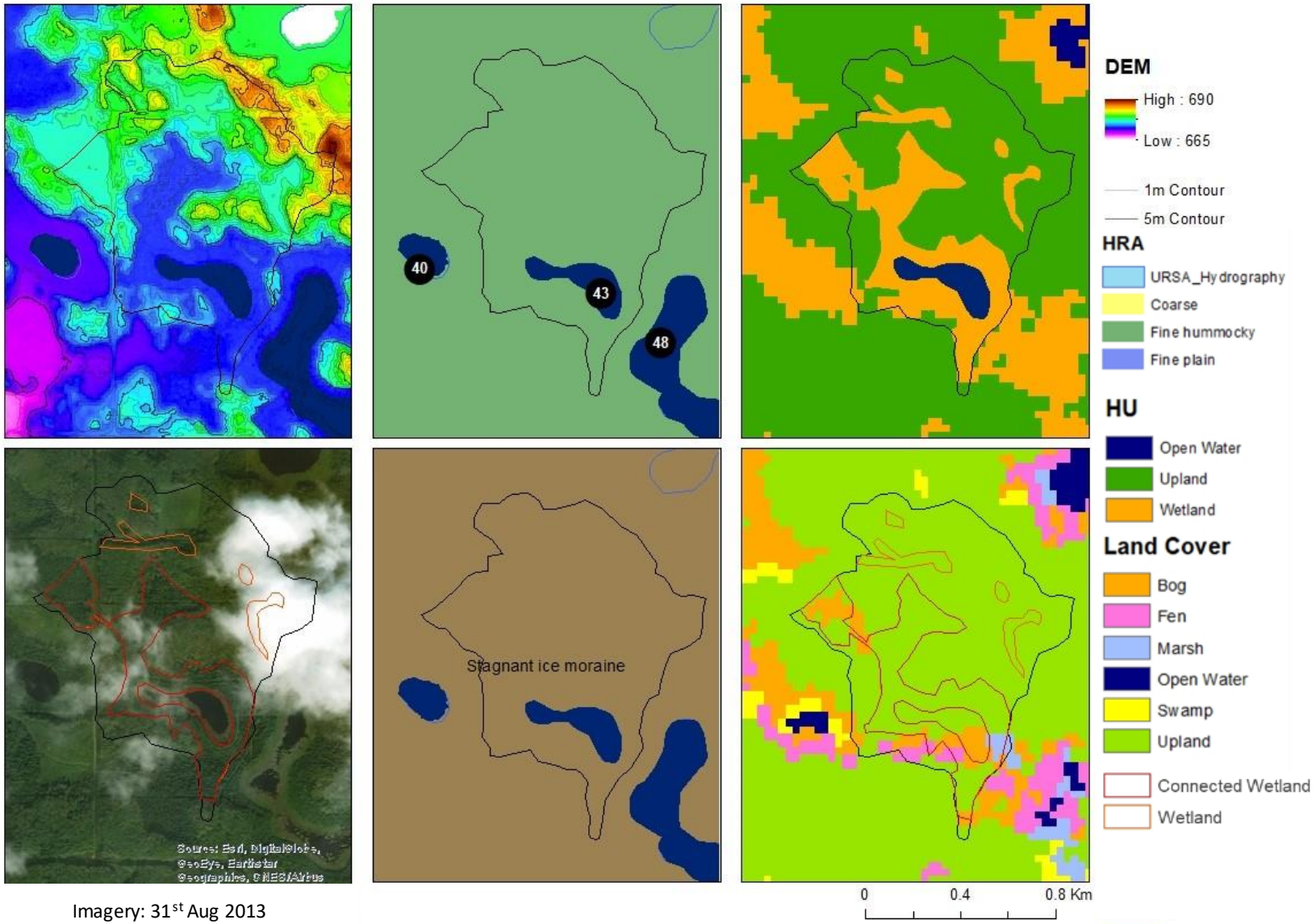
Imagery: 31st Aug 2013

Appendix Figure 17. Lake 40 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



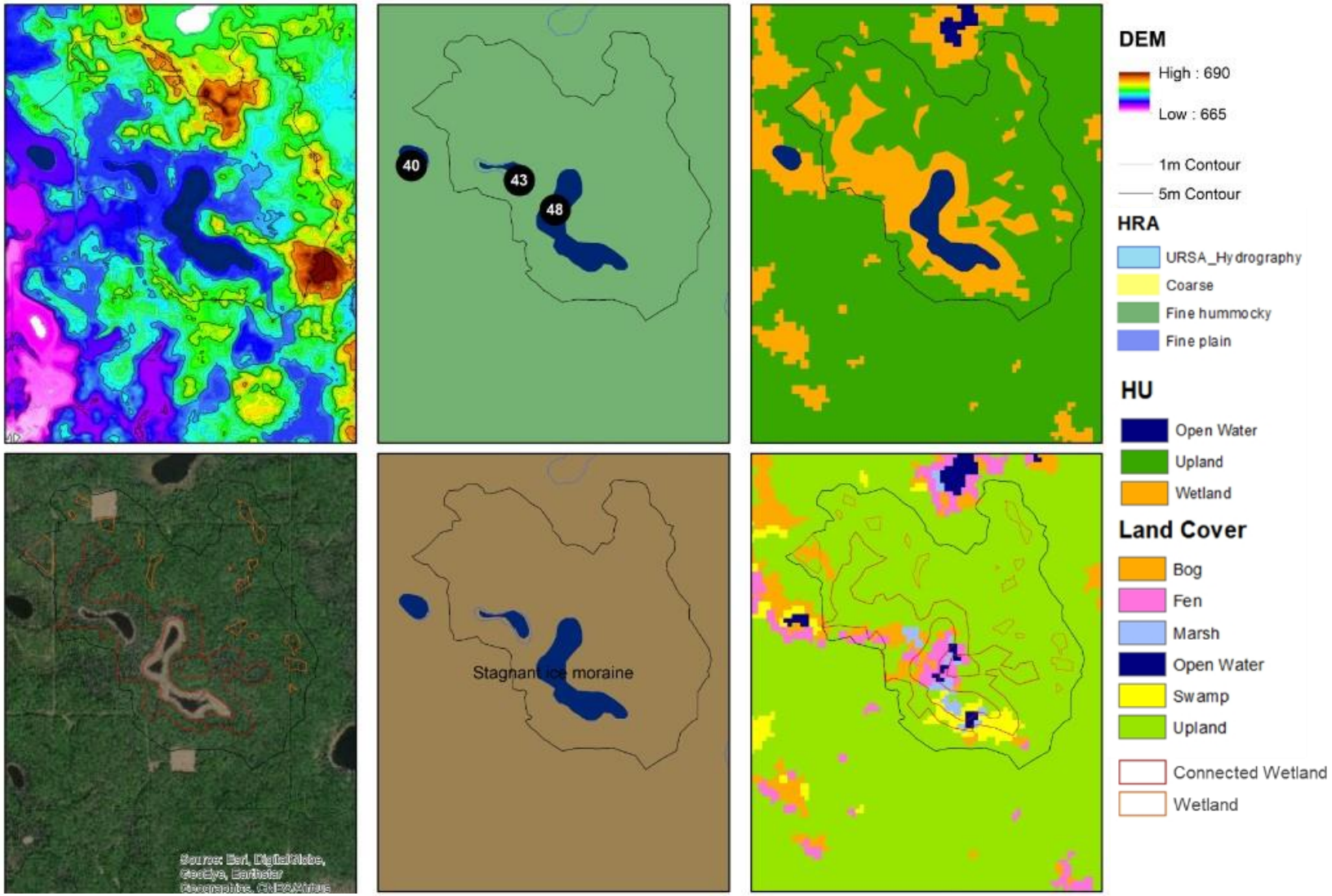
Imagery: 31st Aug 2013

Appendix Figure 18. Lake 42 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



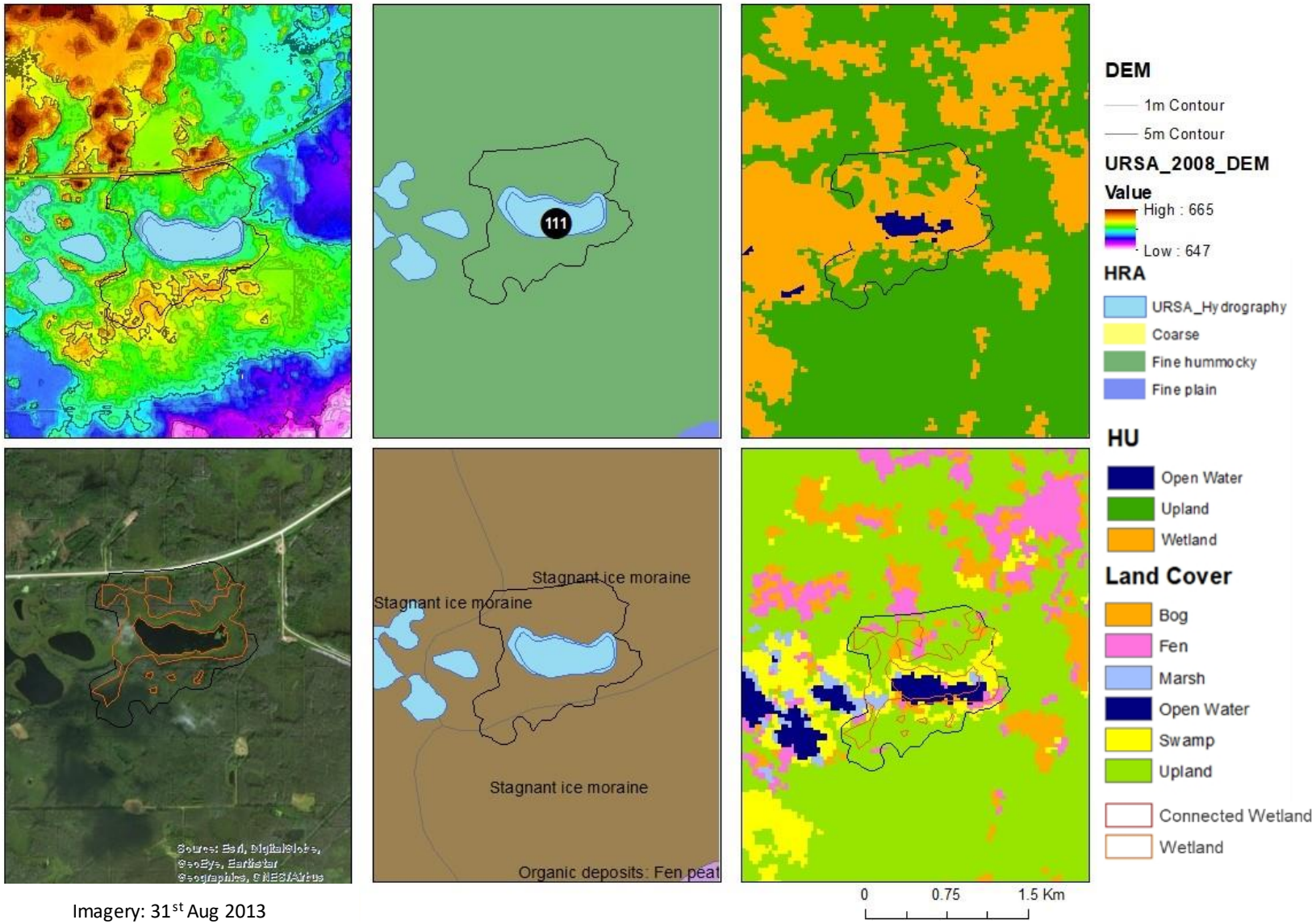
Imagery: 31st Aug 2013

Appendix Figure 19. Lake 43 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



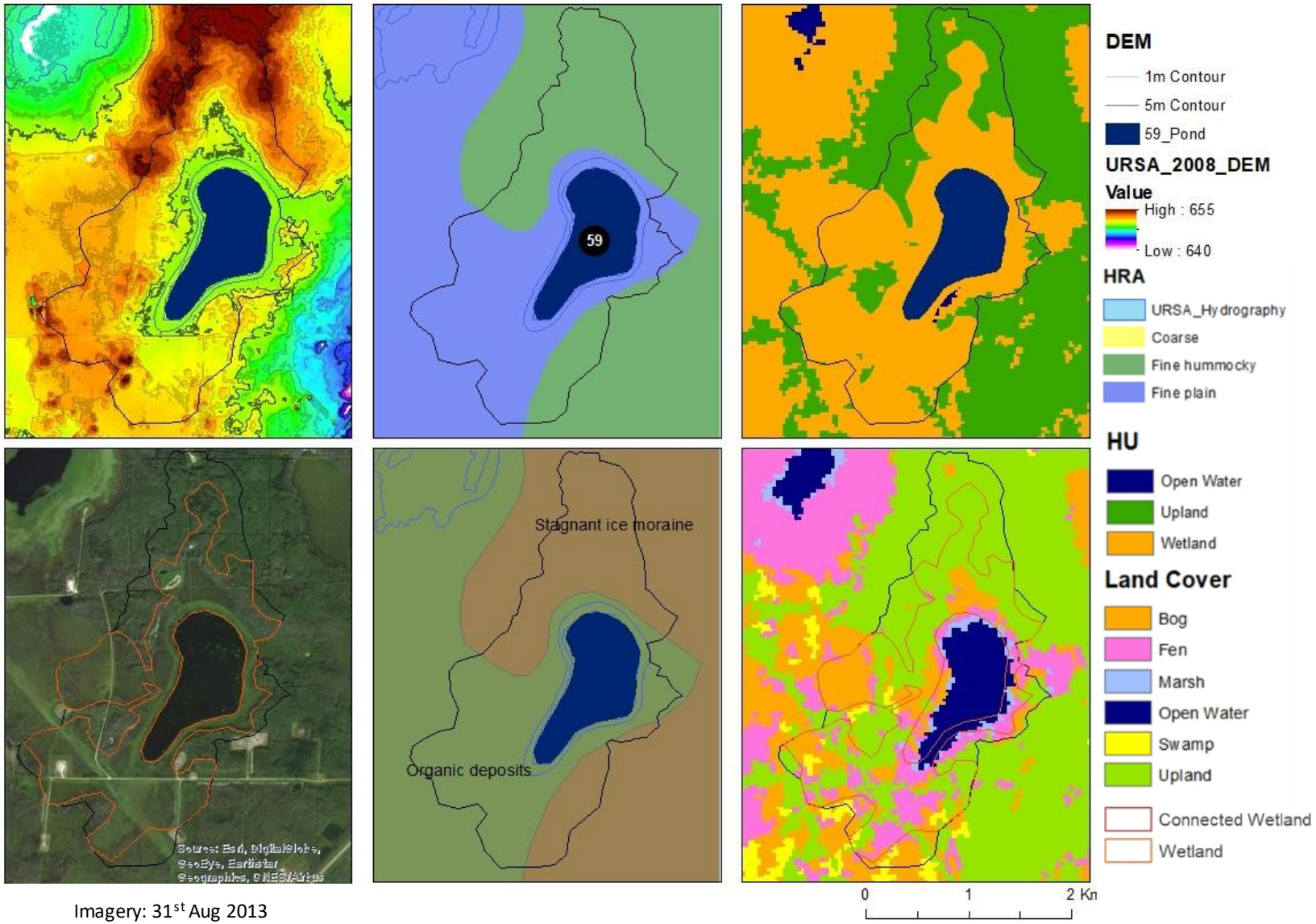
Imagery: May 2019 (lake polygons drawn from August 2013)

Appendix Figure 20. Lake 48 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



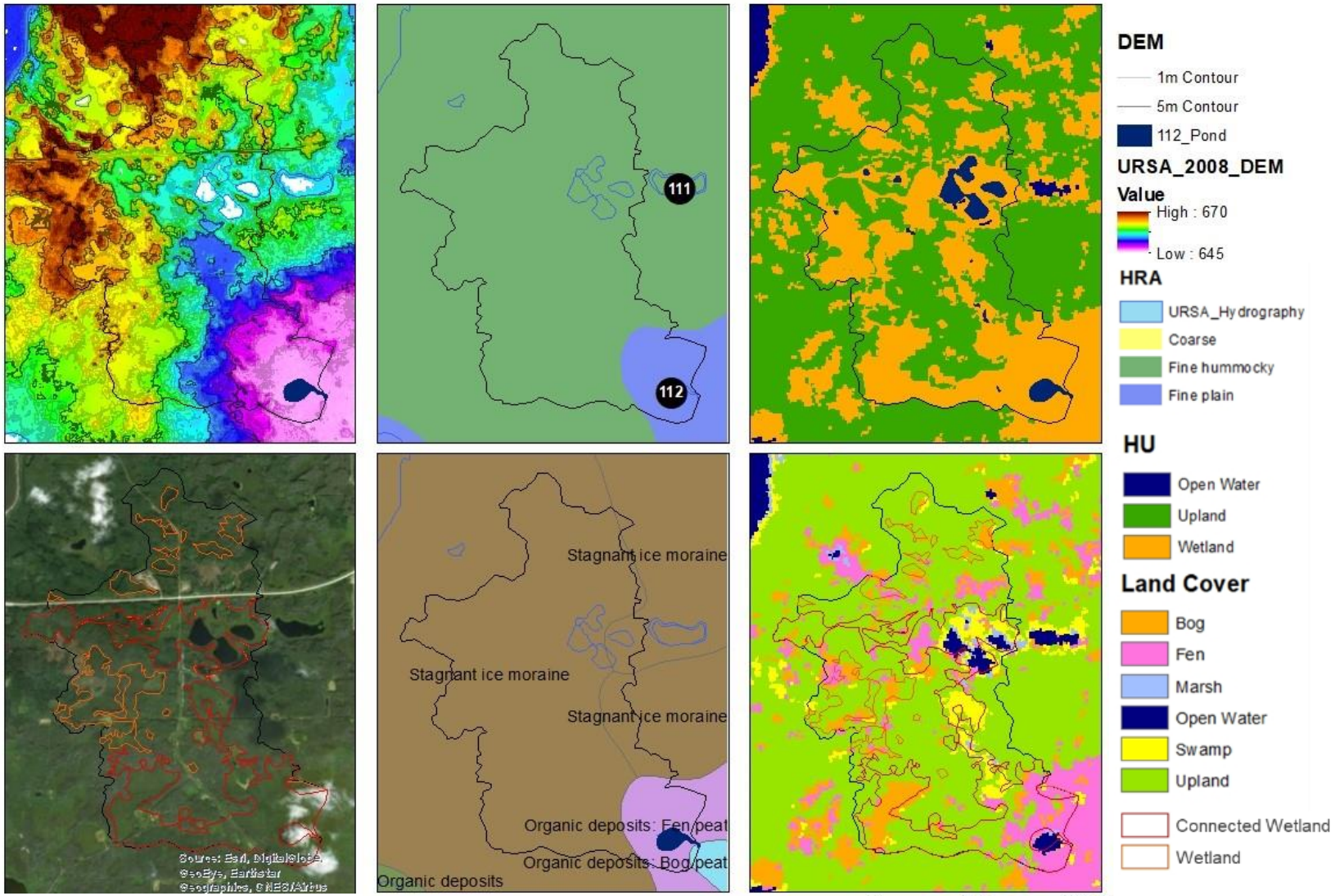
Imagery: 31st Aug 2013

Appendix Figure 21. Lake 111 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



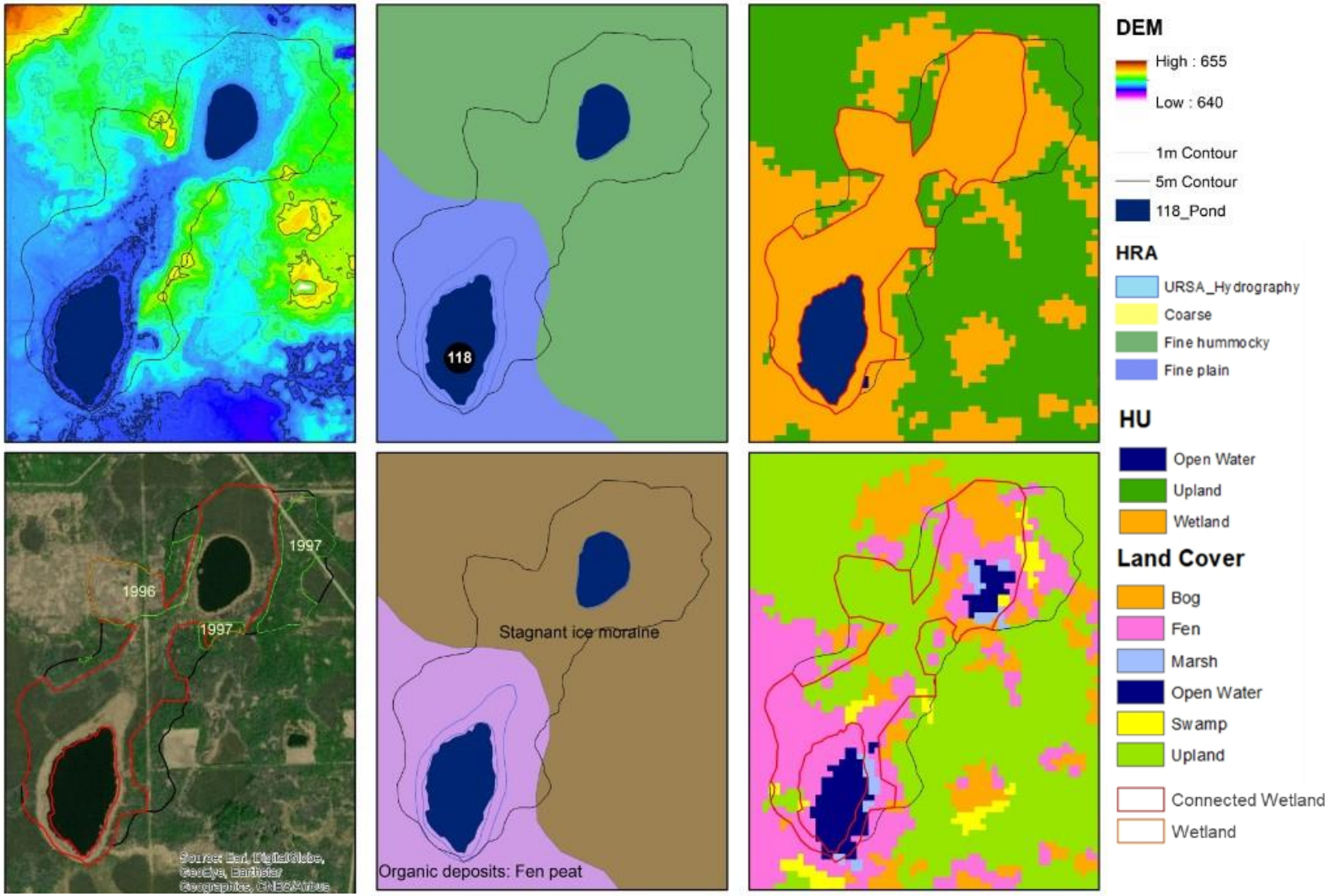
Imagery: 31st Aug 2013

Appendix Figure 22. Lake 59 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



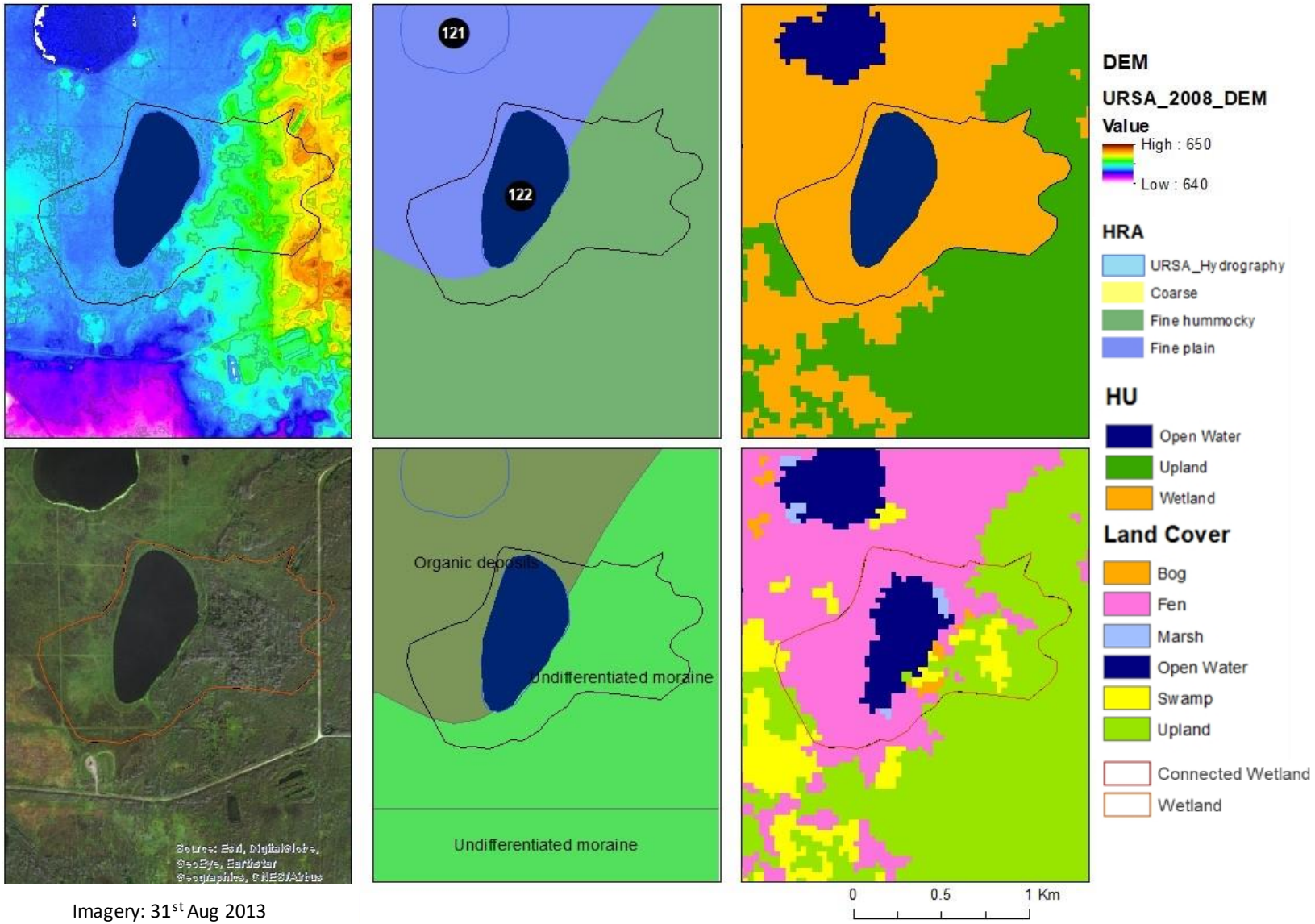
Imagery: 31st Aug 2013

Appendix Figure 23. Lake 112 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



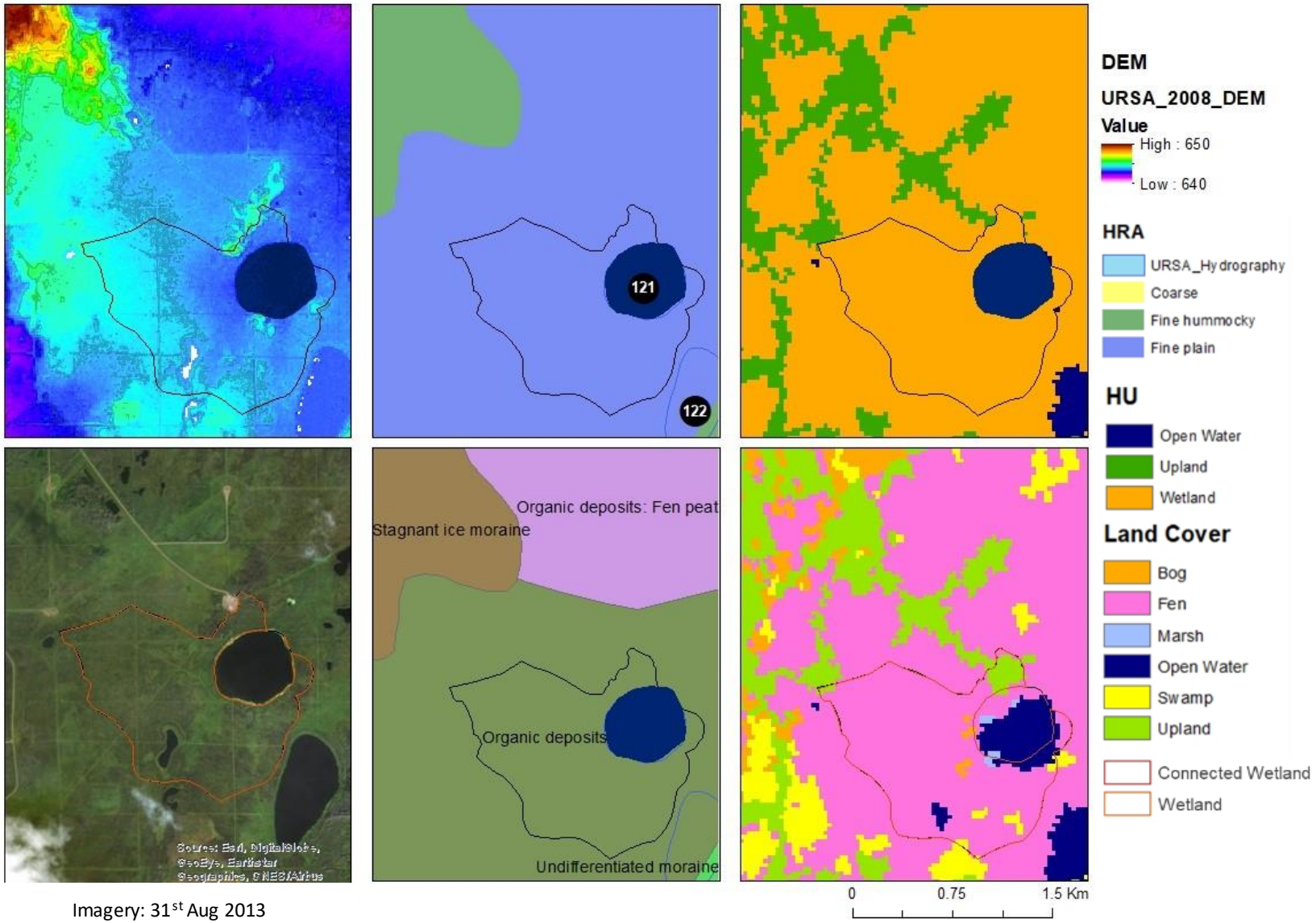
Imagery: May 2019 (Lake polygons drawn from 31st Aug 2013)

Appendix Figure 24. Lake 118 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



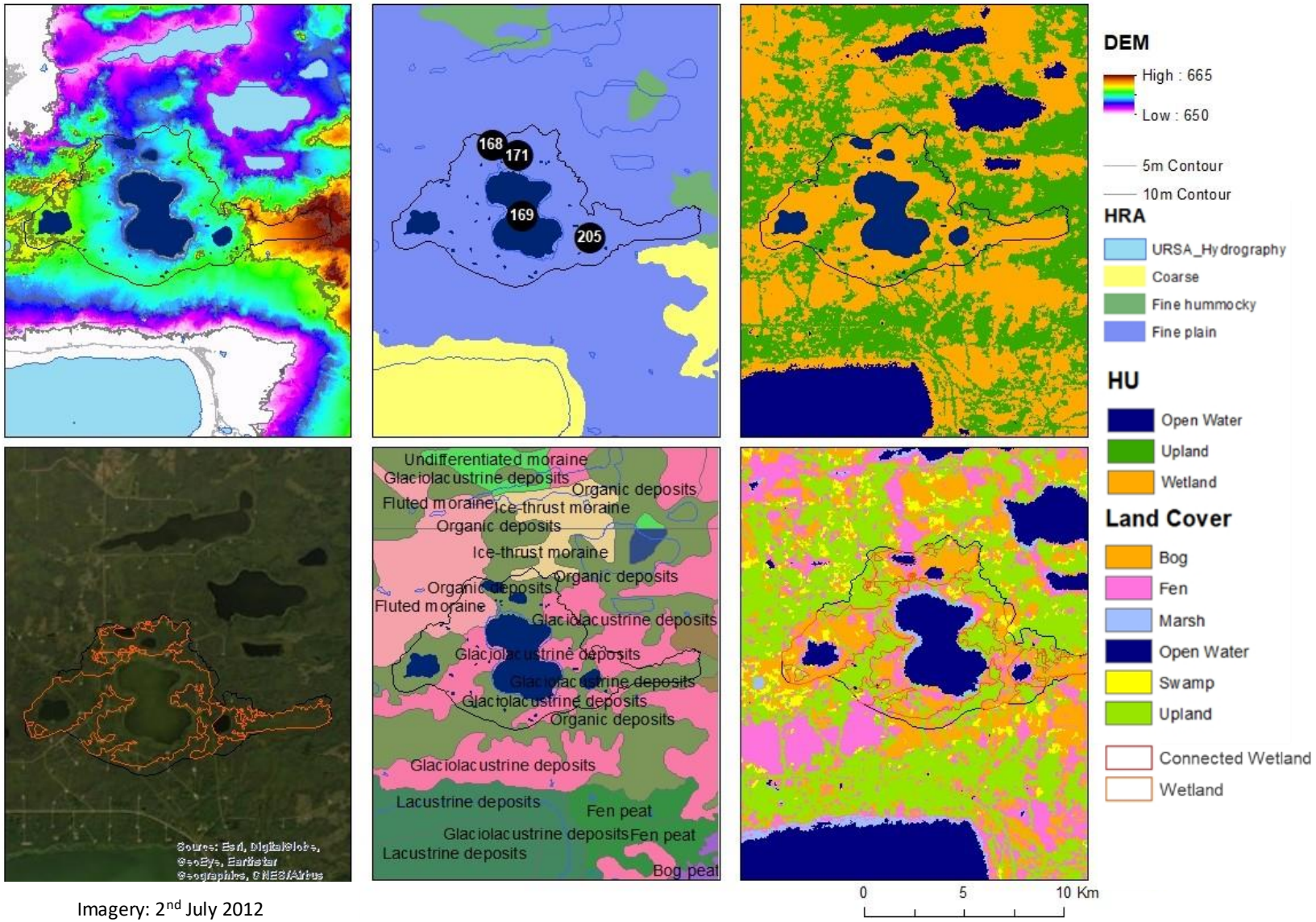
Imagery: 31st Aug 2013

Appendix Figure 25. Lake 122 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



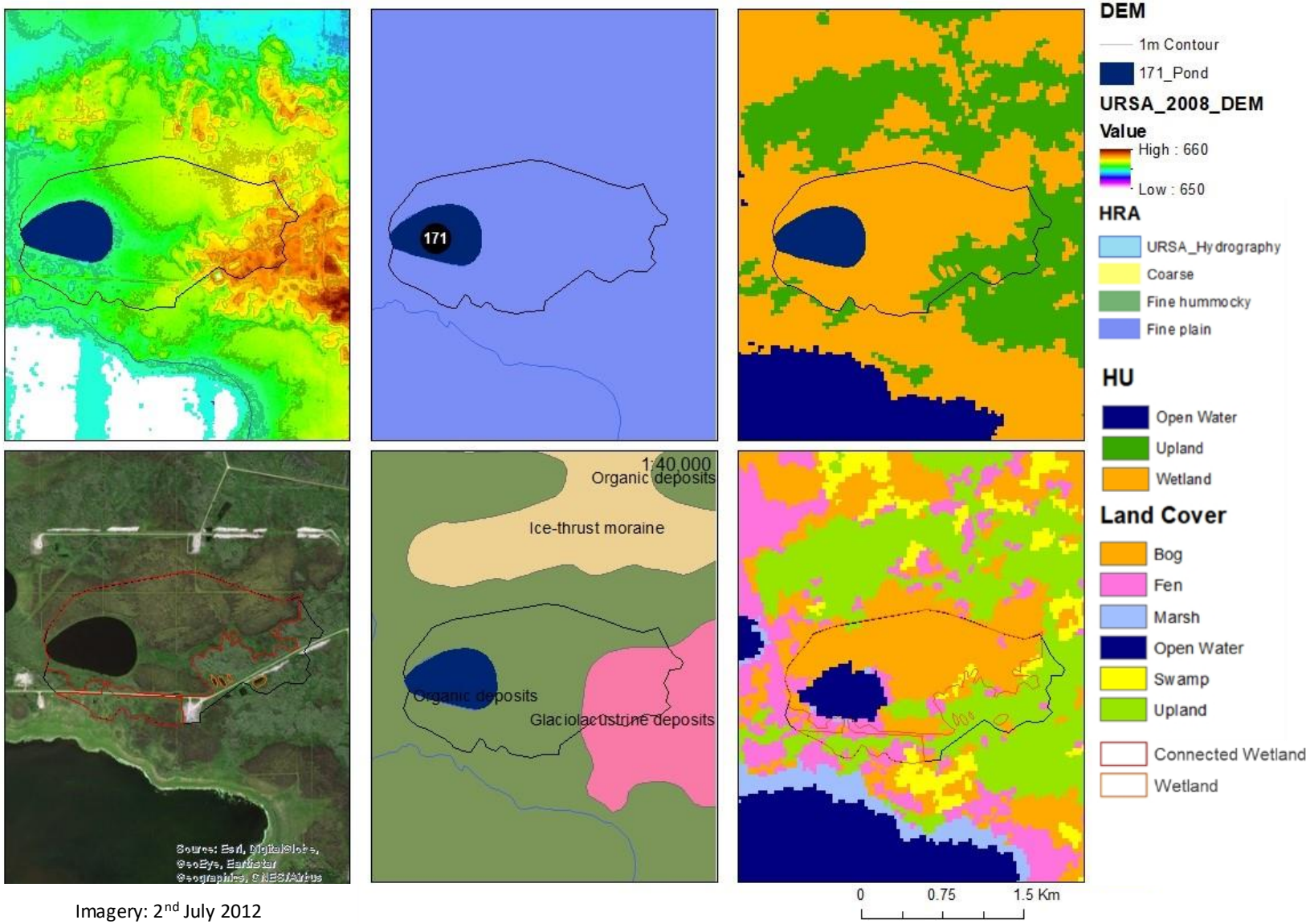
Imagery: 31st Aug 2013

Appendix Figure 26. Lake 121 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



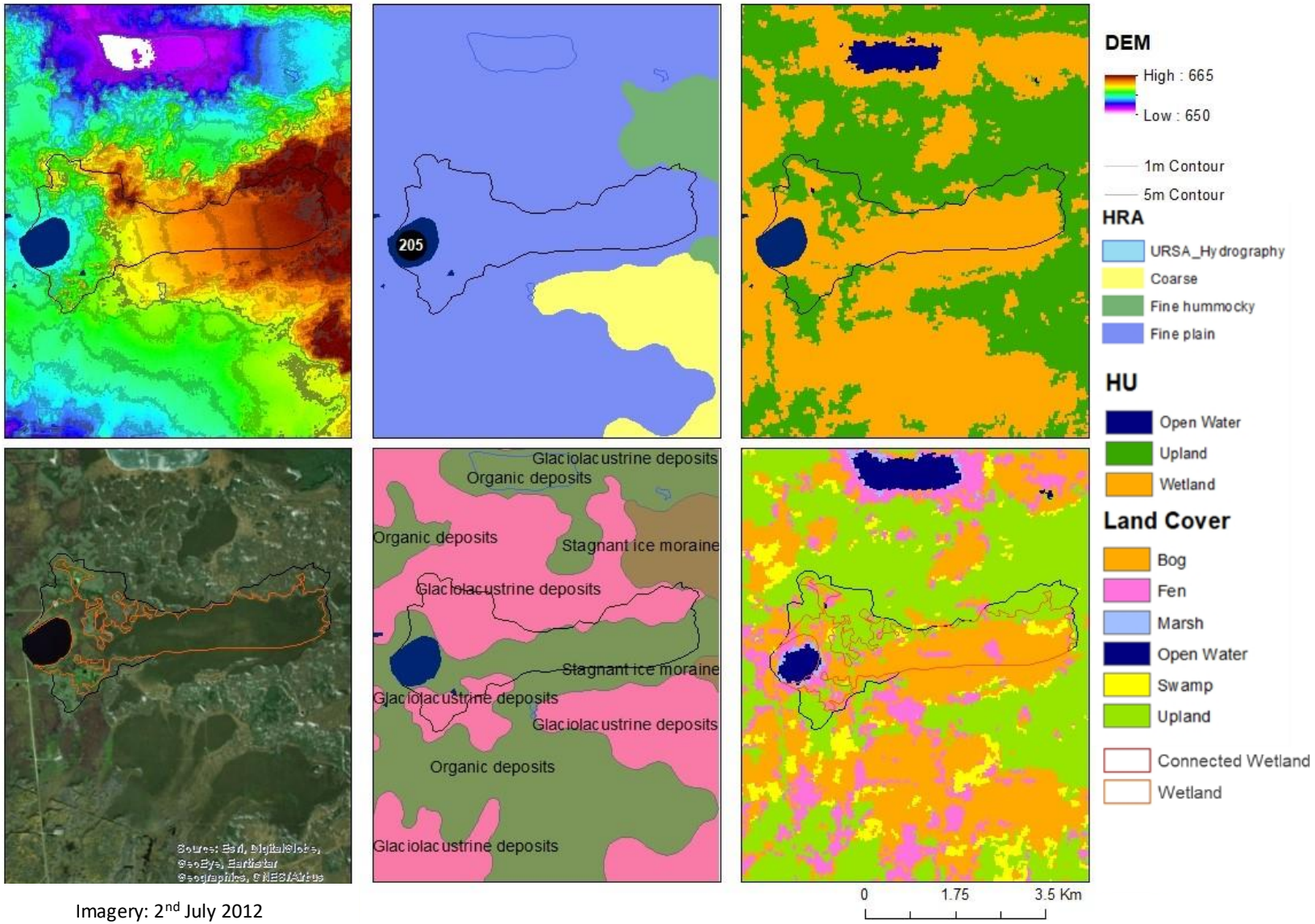
Imagery: 2nd July 2012

Appendix Figure 27. Lake 168 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



Imagery: 2nd July 2012

Appendix Figure 28. Lake 171 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.



Imagery: 2nd July 2012

Appendix Figure 29. Lake 205 topography (top left), HRAs (top middle), HUs (top right), satellite imagery (bottom left), surficial geology (bottom middle, legend omitted), and land cover classification.