The Effect of Annual and Seasonal Variation in Precipitation on Temporal Water Storage Dynamics in Six Headwater Peatland Catchments: Marcell Experimental Forest, Minnesota

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Dedication

For Ian - all the good times we filled with music and conversation; I'll miss them dearly.

Abstract

Using data collected from six headwater peatland catchments at the Marcell Experimental Forest in northern Minnesota, I assessed the relationship between variability in annual precipitation and annual changes in catchment water storage. Three hypotheses are addressed; (1) annual variability in precipitation is a primary driver of catchment storage change, (2) years of below average precipitation drive the relationship between precipitation and catchment water storage change, and (3) winter and fall precipitation variability are significant seasonal drivers of the annual catchment water storage change. The above relationships were analyzed via cross-correlation lag analysis and linear regression analysis of long-term precipitation, peatland water table elevation (WTE), and upland soil moisture (SM) time series, where WTE and SM served to quantify catchment water storage. Results indicate strong correlations between annual water storage change and annual precipitation variability, both in contemporaneous and antecedent years. Concurrent fall precipitation and antecedent winter precipitation were found to have the most influence on a given year's water storage change. Years in which precipitation fell below the catchment average (dry years) exhibited a moderately significant linear relationship with annual catchment water storage change. Results of the above analysis were used to create a series of multivariate linear regression models, both with and without moving-average (MA) errors; these models were able to explain between approximately 50% and 70% of the variance found in the annual water storage change time series. Boreal peatlands play a vital role in the planet's carbon cycle; developing a better understanding of the hydrologic function of these environments will likely prove important to future climate management practices.

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1. INTRODUCTION

The catchment water balance is defined by Thornthwaite and Mather (1957) as the "…balance between the income of water from precipitation and the outflow of water by evapotranspiration." This simple definition can be expanded to account for further hydrologic variables (e.g., streamflow and groundwater interactions) and is often written as:

$$P + GW_{in} - (Q + ET + GW_{out}) = (\pm \Delta S) \qquad \text{Eqn. 1}$$

Where:

P = precipitation (liquid and solid)
$GW_{in/out}$ = groundwater inflow/outflow (liquid)
Q = streamflow (liquid)
ET = evapotranspiration (gaseous)

The water balance is expressed as the total or average over a given time period (i) (Dingman, 2015). Often, the ΔS term is considered stationary and, therefore, overlooked when applying the mean annual and interannual catchment water balance (Wang, 2012). However, changes in catchment water storage can be highly sensitive to variations in hydro-meteorological conditions (Massmann, 2019; Djessou et al., 2022). While the mean and annual change in water storage is likely negligible when considering the annual water balance over large temporal scales, these changes may be a key component of the water balance in years when annual or seasonal precipitation falls below the catchment average (Wang, 2012).

Should global average temperatures continue to increase as projected by current climate models (IPCC, 2014), it is predicted mid and high latitude regions will undergo a shift in precipitation patterns. This shift includes an increase in total annual precipitation (IPCC, 2014) as well as phase transitions from snow dominated to rain dominated regimes (Bintanja et al., 2017). These changes have the potential to alter the hydrologic function of catchments in these regions.

Though only spanning approximately 3% of total land cover, peatlands account for a disproportionate 30% of total soil carbon stores (Krause et al., 2021). Comprising 80% of global peatlands, northern peatlands (north of 45°N latitude) (Alexandrov et al., 2020) have been estimated to contain up to approximately 1.06Gt of carbon alone (Kim et al., 2021); disturbance or destruction of northern peatlands risks the release of these large carbon stores back to the atmosphere. It is estimated that 12% to 40% of the current global carbon budget required to keep warming to manageable levels (1.5°C - 2°C) by the end of the century is stored in peatland systems (Krause et al., 2021; Noon et al., 2022). Due to a long period of recovery, which is on the order of centuries (Krause et al., 2021), peatland emissions represent one of the largest components of what is termed irrecoverable carbon, or a source of permanent debit to the current carbon budget (Noon et al., 2022).

Due to the crucial role northern peatlands play in carbon sequestration, environmental disturbance to these environments due to changes in precipitation patterns and hydrologic function may prove a substantial driver of the ongoing climate crisis in the coming decades. Thus, understanding the response of catchment water storage to changes in precipitation inputs is pertinent to climate change planning, response, and mitigation practices. In this study, I explore the effect of precipitation variability on changes in catchment water storage in six headwater peatland catchments in northern Minnesota. My research assesses the effect of annual and seasonal precipitation inputs on annual water storage change while also investigating the temporal storage dynamics of upland soil moisture and peatland water table elevation. In the following thesis, I address three hypotheses within the framework of small boreal peatland catchments; (1) annual variability in precipitation is a primary driver of catchment storage change, (2) years of below average precipitation drive the relationship between precipitation and catchment water storage change, and (3) winter and fall precipitation variability are significant seasonal drivers of the annual catchment water storage change.

2. STUDY AREA

The Marcell Experimental Forest (MEF; 57°52'N, -93°46'W) encompasses a 1110ha area of Itasca County in north-central Minnesota (Sebestyen et al., 2021). The MEF contains six first-order catchments designated S1 through S6 (Figure 1), each containing either a nutrient-poor raised bog or nutrient-rich fen surrounded by a mineral soil upland (Sebestyen et al., 2021; Table 1). One of the only long-term experimental sites in the U.S. focused on undrained peatland watersheds (Sebestyen et al., 2011a), long-term hydrologic and meteorological monitoring began at the MEF in the early 1960's. Catchments are subdivided into two units, each unit having a separate precipitation record, with catchments S4 and S5 in the North Unit and catchments S1-S3 and S6 in the South Unit. Each unit, North and South, contain a single control catchment, S5 and S2, respectively. All catchments but the control catchments have undergone various land use and forestry treatments since being established (Sebestyen et al., 2011a; Table 2). Topographic relief between the mineral soil uplands and organic soil peatlands within the six research catchments is small, ranging from 412m to 438m between the uplands and the watershed outlets (Dymond et al., 2017). No dominant aspect is present among the MEF catchments (Sebestyen et al., 2011b). Catchments drain to the Gulf of Mexico via the Mississippi River (Sebestyen et al., 2021) except for S4, which also drains to the Hudson Bay due to its position along the continental divide. While most of the catchments drain via ephemeral streams, the S3 fen is unique in containing a perennial stream. Peatlands may also undergo drainage in the form of deep seepage to the regional aquifer (Nichols and Very, 2001), which has been found to account for an average of 40% of the annual water yield for the MEF catchments.

Climate at the MEF is strongly continental, exhibiting moist warm summers with cold, dry, and sunny winters (Sebestyen er al., 2011a). Mean annual air temperature between 1961 and 2019 was 3.5°C with an absolute minimum/maximum temperature of -46°C to +38°C (Sebestyen et al., 2021). Since 1961, air temperatures have increased by 0.4°C per decade (Sebestyen et al., 2011a); most of this warming has occurred in the winter months (Jan-Mar). From 1961 to 2019, average annual precipitation was 787mm (Sebestyen et al., 2021). Annual precipitation is dominated by summer rainfall events (Dymond et al., 2017). Snow cover at the MEF typically lasts from the month of November through March or April (Sebestyen er al., 2021), accounting for approximately one-third of total annual precipitation (Sebestyen et al., 2011a). It is common for the MEF snowpack to undergo a period of mid-winter thaw lasting up to several days.

Annual precipitation has remained unchanged since 1961 (Sebestyen et al., 2011a), though snow water equivalents under upland aspen cover have been declining over the same period.

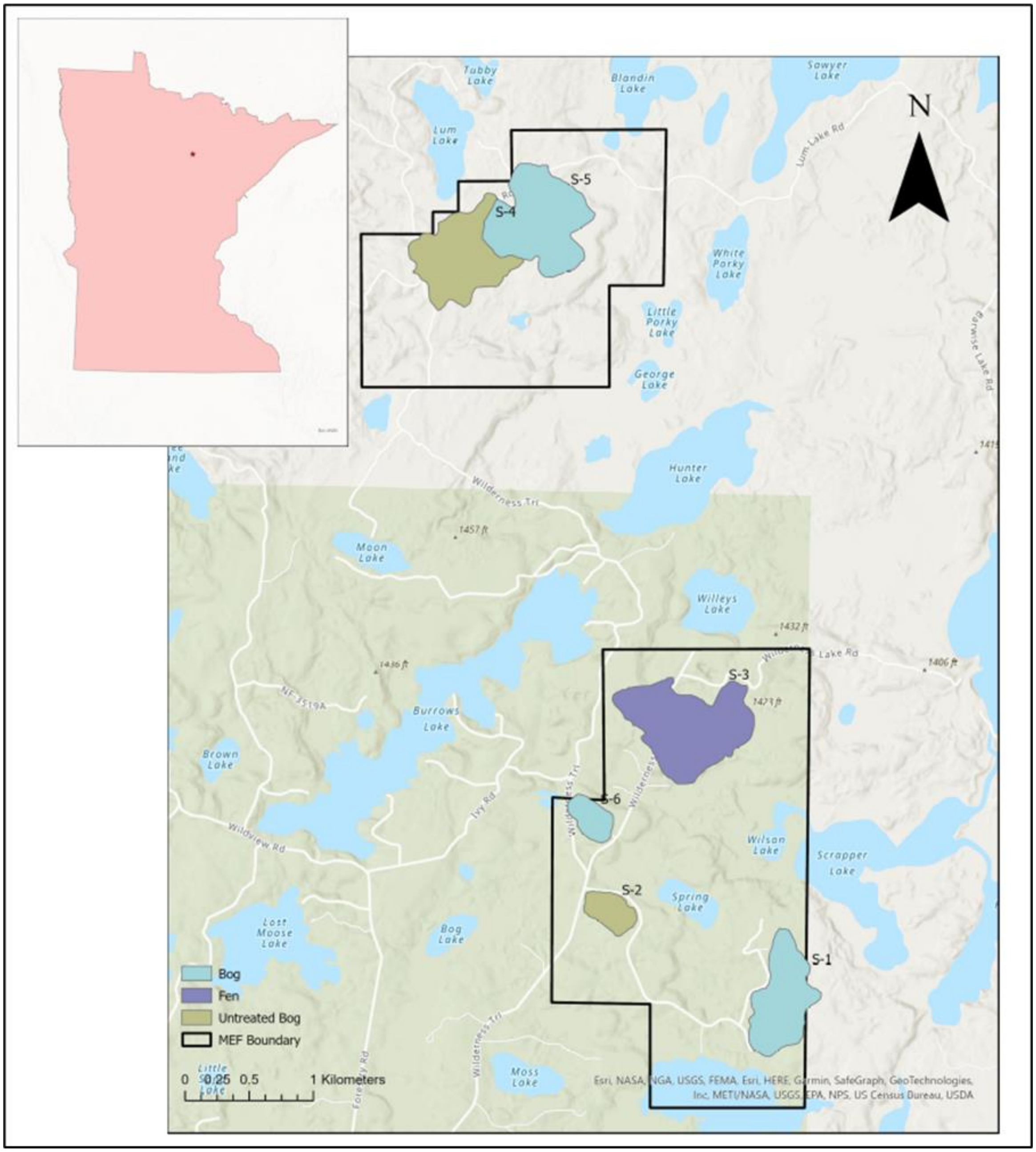

Figure 1: Marcell Experimental Forest study catchments and unit boundaries.

The uplands at the MEF are populated by bigtooth aspen (*Populus grandidentata*) and quaking aspen (*Populus tremuloides*), with intermittent red pine (*Pinus resinosa*) and mixed hardwoods (Dymond et al., 2014). Paper birch (*Betula papyrifera*) may also be found within the upland forest cover (Sebestyen et al., 2021). Overstory species within the peatland bog and fen landscapes are dominated by black spruce (*Picea mariana*) and tamarack (*Larix laricina*), while the understory consists largely of leatherleaf (*Chamaedaphne calyculata*), Labrador tea (*Rhododendron groenlandicum*), cotton grass (*Eriophorum spissum*), and bog rosemary (*Andromeda glaucophylla*) (Sebestyen et al., 2011a; Dymond et al., 2019). The peatlands are dominated by *Sphagnum* mosses and ericaceous shrubs, while fens exhibit a wider variety of moss along with sparse sedge populations.

Table 1: MEF study catchment characteristics.

Catchment	Total Area (km^2)	Type	Elevation Range (m)	Treated
S1	0.33	Bog	412-420	Yes
S2	0.097	Bog	420-430	No
S3	0.72	Fen	412-429	Yes
S4	0.34	Bog	428-438	Yes
S5	0.53	Bog	422-438	No
S6	0.089	Bog	423-435	Yes

Bedrock at the MEF is composed of the intrusive igneous Giants Range Batholith (2.7Ga) portion of the Canadian Shield granite and gneiss, as well as the metamorphic

Ely Greenstone (2.7Ga) (Verry and Jannsens, 2011). These formations are overlain by approximately 40m to 55m of glacial drift deposited during the late Wisconsin glaciation (75Ka-11Ka) (Oakes and Bidwell, 1968; Verry and Jannsens, 2011). This till is covered by up to 35m of sandy outwash composed of soils that include the Menahga sands (mixed frigid, typic Udipsamment), Graycalm loamy sands (isotic, frigid, lamellic Udipsamments), Cutaway loamy sands (fine-loamy, mixed, superactive, frigid oxyaquic Hapludalfs), and Sandwick loamy sands (loamy, mixed, superactive, frigid arenic Glossaqualf) (Sebestyen et al., 2011a); these soils are exposed across approximately one-third of the MEF. The Koochiching clay loam till, which contains fragments of limestone and shale, covers another two-thirds of the MEF (Sebestyen et al., 2011a; Verry and Jannsens, 2011) and ranges in thickness from 3m to 5m. Soils in this till sequence include the Warba sandy clay loam (fine-loamy, mixed, superactive, frigid haplic Glossudalfs), Nashwauk sandy loam (fine-loamy, mixed, superactive, frigid oxyaquic Glossudalfs), and Keewatin fine sandy loam (fine-loamy, mixed, superactive, frigid aquic Glossudalfs) (Sebestyen et al., 2011a). The Koochiching till has an extremely low hydraulic conductivity ($5x10^{-8}cm*s^{-1}$) (Verry and Jannsens, 2011). Both the bog and fen environments are composed of organic soils that range in thickness from 3m in postglacial lake beds to more than 10m in glacial ice block depressions (Verry and Jannsens, 2011). Approximately 75% of the MEF is covered by forested bogs that developed on a dysic typic Borosaprist that grades from a porous *Sphagnum* and ericaceous moss to hemic and sapric members within a depth of 1m (Adams et al., 2004). Forested fen and poor fen peatlands developed on the Mooselake peat, a euic typic Borohemist, and Greenwood peat, a dysic typic Borohemist, respectively. Extensive microtopography can be found in all MEF peatlands in the form of hummocks and hollows (Verry, 1984; Sebestyen et al., 2011a).

Table 2: Treatment history at the MEF study catchments (Sebestyen et al, 2011a; Perala and Verry, 2011)

Catchment	Treatment
S1	strip clear cutting of black spruce (1969, 1974)
S2	control site
S3	upland logging (1960’s) upland controlled burn and replanting with red pine/white spruce (1963) fen clearcutting; 86% controlled burn (1972-1973) fen seeded with black spruce (1974)
S4	upland harvesting (1970-1972) upland fertilization with ammonium-nitrate (1978)
S5	control site
S6	upland clearcutting (1980) upland cattle grazing (1980-1982) upland replanted with red pine/white spruce (1983) removal of upland willow/paper birch/hazel with Garlon 4 herbicide (1983) sulfate added to downstream half of peatland (2001-2008)

3. BACKGROUND

3.1 Storage

While storage is generally considered to be any non-atmospheric water at the surface or within the subsurface, soil water and groundwater are often the primary storage components of a catchment (Rice and Emanuel, 2019). Long-term storage of water as snowpack may also be considered in colder climates due to its delayed contribution to runoff and storage recharge as snowmelt (Kelleher et al., 2015). The consistent partitioning of incoming precipitation (P) into runoff (Q) and evapotranspiration (ET), and the resulting change in groundwater and soil water storage (ΔS), are the basic parameters that drive hydrologic dynamics within a catchment (Dingman, 2015). Together, these parameters define the basic catchment water balance (Eqn. 1).

Due in large part to the impractical nature of direct observation, hydrologists have typically neglected the storage parameter ($\pm\Delta S$) in catchment studies (Wang, 2012). By choosing long observation periods and manipulating the beginning and end of the water year, researchers can greatly minimize or eliminate the need to consider change in storage over a given study period (Dingman, 2015). With the advent of satellite technology, specifically the GRACE based storage data set (Sharma et al., 2020; Abiy and Melesse, 2017), the problem of direct observation has been somewhat mitigated. However, large uncertainties and relatively coarse spatial resolutions are inherent in satellite-derived measurements, making their use inadvisable for watersheds with a drainage area at or below 200,000 km^2 (Thomas et al., 2016). These observational limitations often result in the continued assumption of a hydrologic steady state in catchment water balance calculations, when catchment water storage dynamics are a primary driver in the relationship between groundwater stores and stream channel baseflow (Thomas et al., 2016; Shaw et al., 2012; Sayama et al., 2011) while also being intrinsically linked to terrestrial hydrologic outflows (e.g., evapotranspiration) (Nippgen et al., 2016; Rice and Emanuel, 2019).

Environmental controls also play an important role in catchment storage dynamics (Sayama et al., 2011; Pan et al., 2020). Small watersheds often have variations in local topography, forest cover, and soils; these factors interact across space and time to influence the movement of water through the system. Catchments exhibiting smaller drainage areas, lower degree slope from upland to outlet, and lower average elevations can be linked to positive changes in the catchment water storage capacity (Pan et al., 2020). Steep local topography will generally lead to deeper water tables and larger hydrologic fluxes, while relatively flat topography exhibits shallower water tables and smaller hydrologic fluxes (Condon and Maxwell, 2015); this relationship is more pronounced in humid climates where there is a higher amount of groundwater recharge and groundwater tends to follow the local rather than regional topography. While increasingly steep topographic gradients may increase local drainage and contribute to water storage deficits, decreases in slope angle, as well as catchment microtopography, can result in the creation of a saturation wedge (Lanni et al., 2011) and result in a backwater effect at the topographic channel; this phenomenon results in a delayed contribution from hillslope soils to catchment streamflow that has been shown to sustain discharge from B/C soil horizons for up to 42 days following the contraction of the saturation wedge (Weyman, 1973).

Rice and Emanuel (2019) posited an interannual tendency toward a steady storage state in forested watersheds independent of hydroclimatic setting, demonstrating a positive correlation between forest cover type (deciduous and coniferous) and evapotranspiration. This relationship is likely driven by feedback regimes wherein downregulation of transpiration and the lower hydraulic conductivity of soils mitigate stress on water storage in drier more arid environments (Nippgen et al., 2015; Nippgen et al., 2016), while higher ET as well as increased connectivity of the watershed reinforced steady state storage dynamics in wetter conditions. Stand composition can also influence storage dynamics, with coniferous watersheds exhibiting lower annual storage values than deciduous watersheds due to higher annual rates of evapotranspiration (Nippgen et al., 2016). It has also been demonstrated that lag times in changes to the catchment water storage capacity during periods of prolonged drought are longer in catchments with a lower concentration of broadleaf evergreen species.

Bedrock permeability can also be a considerable influence on both active and total storage in a catchment (Pfister et al., 2017). Catchments exhibiting highly permeable bedrock have been shown to be positively correlated with storage values while also dampening high flow peaks due to higher bedrock aquifer storage potential and its ability to sustain a larger baseflow (Uchida et al., 2006). Underlying bedrock characteristics may even change the dynamic between slope and catchment storage, having been found by Sayama and others (2011) that total storage change was positively correlated with median slope at two Northern California watersheds; this dynamic was hypothesized to be due to a highly permeable bedrock layer as other studies found negative correlations between storage capacity and topography in watershed study areas with sharp soil-bedrock transitions and less permeable bedrock (Troch et al., 2003; Hopp et al., 2009). On seasonal time scales the catchment soil type, surficial geology (Quaternary, 2.6*BP*) and bedrock geology could have a more prominent effect on catchment storage than land cover (Peskett et al., 2020), with freely draining soils being positively correlated with storage values while exhibiting an inverse relationship where these soils overlay impermeable bedrock as opposed to more permeable glacial tills.

3.2 Hydrologic Lag Effects

Though it has been shown that mean annual storage change may be negligible when considering the mean annual water balance (Wang, 2012), interannual variability of catchment water storage may be a key component of the annual water balance in particularly wet or dry years (Wang, 2012; Nippgen et al., 2016; Rice and Emanuel, 2019). Nippgen et al. (2016) found that catchment storage generally increased in years of high precipitation and low runoff at the Coweta Hydrologic Laboratory, while net losses were seen in years at or below average precipitation with high runoff; this relationship indicates a memory effect in which the antecedent storage state of a watershed has a significant influence on the following annual water balance. Wang and Alimohammadi (2012) found that disregarding the annual water storage change resulted in overestimations of evapotranspiration in the annual water balance and that water storage was the most sensitive component in the water balance to precipitation, particularly under water-limited conditions. Groundwater storage can play a significant role in drought years, acting as a buffer to water stress in the presence of exceedingly high evaporation ratios (Wang, 2012). Furthermore, interannual storage carryover has been found to improve annual water balance predictions (Jothityangkoon and Sivapalan, 2009). In boreal peatlands it has been shown that deficits in soil moisture in previous years can be linked to reduced spring runoff the following season (Holden, 2006); this lag response was found to be less prominent in drained blanket peat environments within the United Kingdom (Woo and Young, 1998; Holden, 2006).

3.3 Peatlands

Peatlands cover approximately 5-million km^2, contain 10% of surficial freshwater stores, and account for a third of total global soil carbon stores (Tarnocai and Stolobovoy, 2006). Peat soils consist of decaying organic matter that has collected under water-saturated conditions, typically developing in areas where precipitation or groundwater input exceed evapotranspiration processes (Holden, 2006). Northern peatlands tend to develop within boreal and upland temperate zones and can also be found in lowland areas subject to high precipitation input, low topographic gradients, and low conductivity substrates. The southern extent of the northern boreal peatland zone is generally considered to be ~40°N latitude (Tarnocai and Stolobovoy, 2006, Alexandrov et al., 2020).

Peatlands are classified according to their hydrologic and chemical characteristics. Ombrotrophic peatlands, or bogs, are maintained by precipitation inputs, requiring an annual precipitation of greater than 600 mm. Water and nutrient cycling in minerotrophic fens is driven largely by groundwater exchange (Holden, 2006). Chemically, bogs are acidic (pH < 4) while fens are acidic to circumneutral, with a pH range between 4.5 (poor fens) and exceeding 6.9 (rich fens).

The basic structure of a peatland can be described using the acrotelm-catotelm model in which the peat is subdivided into two distinct sections (Holden, 2006) (Figure 2). In the highly conductive upper layers (acrotelm) the partially living peat is subject to occasional aeration, facilitating the aerobic microorganisms necessary for peat formation to thrive. The acrotelm can be defined at its base by the lowest average water table depth.

Below the acrotelm lies the catotelm, an anerobic zone void of living plant material with the exception of some aerenchymatous rooting from overlying helophytic angiosperms (Ingram, 1983; Holden, 2006). The water content of the catotelm is generally considered temporally invariable within the acrotelm-catotelm model (Holden, 2006).

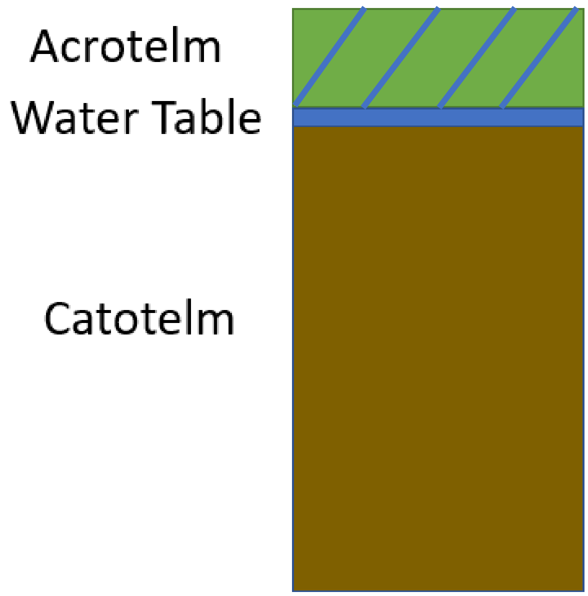

Figure 2: Acrotelm-catotelm model of a peatland. Average water table depth is indicated by the blue horizontal bar. Variability in the water table is indicated by the diagonal blue lines.

The preservation of the catotelm peat requires saturated conditions to inhibit decomposition processes. Saturation is maintained by the water table, which is variable between the top of the catotelm and the top of the acrotelm (Holden, 2006; Verry et al., 2011; Figure 2) in the case of undrained peatlands like those found at the MEF. Though the peat can maintain a 90% − 98% water content by volume, even above the water table, little of this stored water is partitioned to maintain catchment baseflow (Holden, 2006). Therefore, dry weather and low streamflow may not be indicative of water limited conditions within the peat unless there is a connection to a wider hydrologic system. The saturated conditions of peatlands also lead to stormflow regimes dominated by saturation excess overland flow, with much rarer occurrences of infiltration-excess overland flow (Holden and Burt, 2002; 2003). Peatland moisture retention may also play a part in modifying downstream peakflows (Ogawa et al., 1986).

Evapotranspiration and streamflow are considered the primary drivers of peatland water loss (Holden, 2006), with evapotranspiration typically being the largest factor (Baird et al., 2004). Following drops in water table level of approximately 8 cm or more, any additional drops in water level have been shown to be almost entirely driven by evapotranspiration (Evans et al., 1999; Charman 2002), with loss via drainage after this point becoming a minimal factor. Increases in the peatland water table are primarily driven by precipitation (Holden, 2006), with minimal rainfall required to raise water levels. At the MEF, deep seepage to the regional aquifer has also been shown to be an important factor in peatland water table levels (Nichols and Verry, 2001).

3.4 Midlatitude Cold Climates

Midlatitude regions can be considered cold when snow and ice have at least a seasonal presence (Gelfan and Motovilov, 2009). Regional hydrology is appreciably affected by the presence of water in its solid phase; this includes the seasonal accumulation of a winter snowpack and the subsequent spring melt of this snowpack, which is a large contributor to annual stream flow (Aygün et al., 2020, Jones et al., 2023). Spring freshet provides approximately 53% of annual runoff in the Western United Sates and up to 70% in areas located in more mountainous terrain (Li et al., 2017). The seasonality of cold climate regions also affects the nature of catchment runoff, with the degree of subsurface freeze determining the fraction of surface and sub-surface runoff in these regions during the melting period (Lundberg et al., 2016; Aygün et al., 2020).

Under ongoing influence of anthropogenic climate change, it has been predicted there will be changes to the winter snowpack and spring melt dynamic in midlatitude cold regions due to changes in the timing of these events (Aygün et al., 2020), an increase in annual precipitation overall (IPCC, 2014), and a decrease in the fraction of annual precipitation that is snow or ice (Bintanja et al., 2017). Historical trends in the Northern Hemispheres show marked decreases in snow cover duration (Dye, 2002; Choi et al., 2010) and snow water equivalent (Gan et al., 2013); similar trends in snow cover extent have also been reported across the Northern Hemisphere (Déry and Brown, 2007; Brown and Robinson, 2011; Thackeray et al., 2016). Changes to the precipitation regime are projected to increase winter and spring soil moisture levels (Mishra et al., 2010) while decreasing these levels in the summer months (Hayhoe et al., 2007; Kellomäki et al., 2010). Decreases in total annual groundwater recharge have also been projected for multiple Canadian catchments (Sulis et al., 2011; Rivard et al., 2014).

4. METHODS

4.1 Data Sets

4.1.1 Precipitation

Daily precipitation is recorded at two primary locations, a south location within the S2 catchment and a north location within the S5 catchment (Sebestyen et al., 2011a). Data collection has been uninterrupted through 2020 at the North and South units since

1962 and 1961, respectively. Both locations contained a Belfort Universal Recording Precipitation Gauge chart recorder until their replacement with NOAH IV precipitation gauges in 2010 (South) and 2014 (North). Weekly values are verified using a paired standard rain gauge at each site (Sebestyen et al., 2021). The previous instrumentation collected data at daily intervals while the current digital gauges collect data at 15-minute intervals that are summed to daily values. Snowfall is measured as liquid by adding antifreeze and oil to the gauge collection buckets to induce melting while limiting evaporation (Sebestyen et al., 2011a). No distinction is made between precipitation occurring as rain and precipitation occurring as snow. Catchments were assigned separate precipitation records based on their associated meteorological unit.

4.1.2 Soil Moisture

Seasonal upland soil moisture has been measured three times annually as early as 1966, depending on the catchment. Measurements are generally taken in May, September, and November. Collection intervals for the soil moisture readings represent the approximate times preceding deciduous leaf emergence, the end of summer before leaf fall, and the time after leaf fall preceding soil freeze (Sebestyen et al., 2011a). At the time of analysis, measurements were available through calendar year 2019 and available only in part for 2020. Each measurement is taken using a neutron probe technique (Sebestyen et al., 2011a) at intervals of 30.5cm from a depth of 15.2cm to 305.0cm. Soil moisture in the top 15cm of the soil column could not be measured by neutron probe (Dymond et al., 2021) due to the escape of fast neutrons through the soil surface (Chanasyk and Naeth, 1996), soil moisture in the top 15cm of the soil column were measured by sampling mineral soil at three random points surrounding the neutron probe access tubes using a T bar sampler; these samples were weighed before and after being oven dried to obtain a gravimetric moisture content that was subsequently converted to volumetric moisture (Dymond et al., 2021). The maximum soil moisture measurement depth varies by catchment and individual site. Reported values have been converted from percent volumetric soil moisture to available water (cm) by multiplying percent soil moisture by the measurement depth and subtracting a 15-bar wilting point constant for the dominant overstory species from the resulting values (Dymond et al., 2021).

Due to inconsistencies in the maximum depth to which soil moisture measurements were taken, records were only utilized to 250cm across all study sites. Any missing observations above the 250cm cut-off were filled using a seasonal rolling mean across the period of record at each observation depth to account for any potential trends in the data set. Extraneous observations, mostly present toward the earlier portion of the record, were removed such that only three observations were included for each water year.

Because soil moisture measurements were collected based on observed seasonal phenomena (i.e., leaf fall), they were not taken at consistent dates. Therefore, seasonal measurements varied in month over the period of record. Soil moisture was summed across the soil column at each site. Catchments S1 through S2 and S4 through S6 contained two usable sites each; after calculating the sum at each site, the results were then averaged across the available sites in each of the catchments. Soil types were homogeneous between sites for catchments where soil moisture readings were averaged (Figure 3).

4.1.3 Water Table Elevation

Daily water table elevation has been measured in the central peatland of each catchment since the early or mid-1960's. All data are recorded as meters above mean sea level (Sebestyen et al., 2021). Water table elevation was measured near the center of each peatland in sheltered pools via FW-1 strip chart recorders and subsequently verified by weekly manual measurements (Sebestyen et al., 2011a). The elevation of the water table is calculated by taking the difference between the pool shelter elevation relative to mean sea level and depth to the water table.

Measurements for all sites were multiplied by an average drainable porosity value to account for the influence of peatland pore space on the observed water table elevation (Verry et al 2011). While there have been multiple studies and recommendations regarding either derivation or choice of appropriate drainable porosity values for peatland catchments (Boelter, 1965; Paivanen 1973; Gafni, 1986; Nichols and Verry, 2001), a drainable porosity of 0.57 was chosen based on the results of an unpublished experiment conducted at the MEF (see Appendix A).

4.2 Analysis

The relationship between the annual change in catchment water storage and variations in precipitation at the MEF was explored using a combination of cross-correlation (lagged Pearson's correlation), bivariate time series regression, and bivariate time series regression with autoregressive integrated moving average (ARIMA) errors. The influence of climatological conditions (wet/dry) and seasonal precipitation (winter/growing/fall) were also assessed. All analyses were based on a November through October water year, which was chosen to capture the state of water stored in the system after annual inputs from spring snowmelt and summer/fall liquid precipitation had entered and subsequently drained from the system or entered the storage pool. The record length used in the analysis varied by catchment and was dependent on the maximum length of the seasonal soil moisture records (Table 3). All analyses were conducted using the R statistical computing platform version 4.1.2 (”Bird Hippie”).

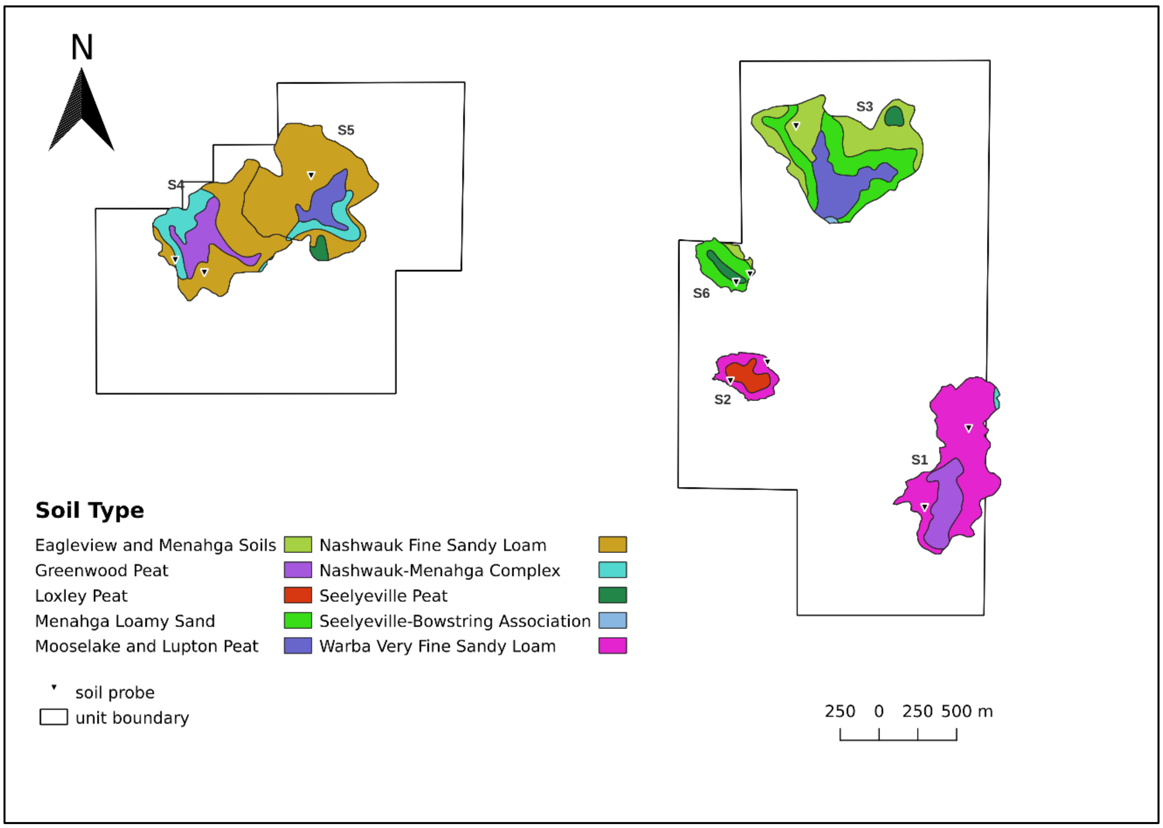

Figure 3: Soil series across the MEF catchments (Soil Survey Staff, 2021). Includes both mineral upland and peatland organic soil series. Soil moisture probe sites used in this study are indicated by black arrows at their approximate location in each catchment.

4.2.1 Hydrologic Variables

Prior to the analysis, upland soil moisture and peatland water table elevation values for each catchment were upscaled by multiplying each observation by the total upland area and total peatland area, respectively. Total annual catchment water storage change was defined at each catchment as the sum of the total annual change in upland soil moisture and the total annual change in peatland water table elevation and calculated as:

$$\pm\Delta S_i = \pm\Delta S_{SM_i} + \pm\Delta S_{WT_i} \qquad \text{Eqn. 2}$$

where:

ΔS = annual water storage change

ΔS_{SM} = annual soil moisture storage change

ΔS_{WT} = annual water table storage change

i = water year

Table 3: Water year ranges used for analysis by catchment. Ranges were based on availability of soil moisture data. The difference in record length for the S6 catchment is due to the use of multiple soil moisture observation sites, one of which contained a large gap of several years prior to 1985.

Catchment	Water Year
S1	1968-2019
S2	1969-2019
S3	1967-2019
S4	1969-2019
S5	1971-2019
S6	1986-2019

The annual change in the upland soil moisture component of the total annual water storage change variable was calculated as:

$$\pm \Delta S_{SM_i} = SM_{FZ_{i-1}} - SM_{FZ_i} \qquad \text{Eqn. 3}$$

where:

ΔS_{SM} = annual soil moisture storage change
SM_{FZ} = soil moisture measurement prior to freeze-up
i = water year

Lastly, the annual change in the peatland water table elevation component of the total annual water storage change variable was calculated using Eqn.4:

$$\pm \Delta S_{WTE_i} = WTE_{i-1} - WT_{Ei} \qquad \text{Eqn. 4}$$

where:

ΔS_{WTE} = annual water table storage change
WTE = water table elevation on November 1st
i = water year

Total annual precipitation for a catchment was defined as the sum of all precipitation events for a given water year (Nov. 1 – Oct. 31) within its respective unit (North or South). The precipitation anomaly was calculated using these totals to establish the magnitude of departure from the mean annual precipitation over the period of record (Eqn.5).

$$\pm P_i^* = P_i - P_u \qquad \text{Eqn. 5}$$

where:

P^* = annual precipitation anomaly
P = total annual precipitation
P_u = mean annual precipitation over the period of record

i = water year

4.2.2 Annual Correlation and Lag Relationships

Cross-correlation analysis was used to determine what, if any, interannual lag relationship exists between annual catchment water storage change and variations in precipitation at the MEF. Cross-correlations were calculated using the precipitation anomaly variable as a predictor of total annual water storage change for each of the six catchments. Cross-correlations were calculated using a 95% confidence interval (α = 0.05) from a lag of zero (comparing storage values to precipitation values in the same year) up to a maximum of five annual time lags (comparing storage values to precipitation values in previous years). This analysis was repeated for both components of the total annual water storage change – annual peatland water table elevation change and annual upland soil moisture change – using the same annual precipitation anomaly data. All correlations were calculated as Pearson's coefficients.

Prior to conducting the analysis, the precipitation anomaly data sets for each catchment were tested for stationarity using the Augmented Dickey-Fuller test from the R "tseries" package (v0.10-51; Trapletti and Hornik, 2022); autocorrelation and partial autocorrelation (ACF/PACF) plots were also examined to determine if autocorrelation existed in these time series. While none of the precipitation data exhibited autocorrelation, the S6 data set was determined to be non-stationary at the 95% confidence interval. Using the "Arima" function from the R Forecast package (v8.21; Hyndman et al., 2023), a first difference was applied to the S6 precipitation anomaly to account for the trend. The S6 precipitation anomaly was subsequently fitted with a single autoregressive (AR) error to account for autocorrelation introduced by the first-difference.

4.2.3 Seasonal Correlation and Lag Relationships

Cross-correlation was also used to assess how variation in precipitation within certain seasons related to annual changes in total annual catchment water storage, total annual soil moisture storage, and total peatland water table elevation. Seasons were defined according to the observation schedule of the soil moisture data, resulting in a three-season water year consisting of a Winter season (November-April), a combined Spring and Summer growing season (May-August), and a brief Fall season (September-October). Using the three seasons, total seasonal precipitation in each water year was calculated then used to calculate three new annual precipitation anomalies for each of the three seasons. The new annual precipitation anomalies (P^*_s) were defined as:

$$\pm P^*_{s\,i} = P_{s\,i} - P_{s\,U} \qquad \text{Eqn. 10}$$

where:

P_s = total annual precipitation in season s
P_{Su} = mean seasonal precipitation
i = water year

The resulting seasonal precipitation anomalies were used as predictor variables in the cross-correlation analysis. Precipitation data was again subject to the Augmented Dickey-Fuller test as well as autocorrelation assessment. It was found that the fall precipitation anomaly data for S6 was not stationary and was corrected using a first-difference.

Seasonal cross-correlation was only assessed up to the highest significant lag found in the annual cross-correlation analysis.

4.2.4 Climatological Variation

The significance of variation in climatological conditions (i.e., wet/dry) on annual water storage change was also assessed. Wet and dry years were defined based on the annual precipitation anomaly, where:

$$DRY = P_i^* < 0cm \qquad \text{Eqn. 8}$$

and

$$WET = P_i^* > 0cm \qquad \text{Eqn. 9}$$

This categorization did not contain a transitional category or give distinction to years with average precipitation. Total annual catchment water storage change was subset into two new data sets corresponding to the climate category for each year over the period of record for all six sites. A regression analysis was performed between the precipitation anomaly and water storage change data for years determined to be dry and years determined to be wet. Resulting regression models were considered significant at or below the 95% confidence interval (α = 0.05). The analysis was repeated for the annual peatland water table elevation change and annual upland soil moisture change components of the total annual catchment water storage change variable.

The data were further categorized depending on whether a given year was preceded by either a wet or dry year as defined in Eqn. 8 and Eqn. 9, and given a numerical identifier of 1 or 2. The water storage data in each group was then regressed against a lagged version of the precipitation anomaly to determine what, if any, relationships existed between the climatological conditions of a previous year and water storage change in a given year. Because the nature of this analysis involved deconstruction of the original time series, neither stationarity nor auto-correlation assessment was necessary.

4.2.5 Final Models

Results of the previous analysis were used to help construct a series of water storage models using bivariate time series regression. Models were first produced by regressing the water storage values against various lags of the precipitation anomaly as determined by the correlation analysis. The residuals of these models were then examined for MA, AR, and trend structures using ACF/PACF analysis. If any of these structures were found, the regression was remodeled using autoregressive integrated moving average errors as determined by the examination of the ACF/PACF plots. The significance of the model was assessed by regressing the observed water storage variables against the calculated fitted values of the regression with ARIMA error models. The strength of each model was assessed based on the adjusted R^2 and p-value results (α = 0.05).

5. RESULTS

5. 1 Precipitation Anomaly and Storage Time Series

In 1976, a precipitation anomaly of -28.7cm was accompanied by an annual change in total water storage of -446.9cm^3 in catchment S4 and -769.2cm^3 in catchment S5 (Figure 3). An example can also be seen in 2006, where a precipitation anomaly of -17.9cm was accompanied by a total water storage change of -142.5cm^3 in S4 and -

1049.7cm^3 in S5. In most years with less than average precipitation there were similar negative changes in total catchment storage; an exception to this can be seen in 1980, which had a precipitation anomaly of -9.78cm while catchments S4 and S5 saw positive changes in water storage of 144.1cm^3 and 173.8cm^3, respectively.

The South Unit time series (Figure 4) indicates a similar relationship between the annual precipitation anomaly and total annual water storage. A negative precipitation anomaly of -31.8cm can be seen in 1976, with all available catchments showing negative changes in total water storage in that year (S1 = -614.7cm^3, S2 = -135.0cm^3, S3 = -397.2cm^3). After its record begins in water year 1986, total catchment storage in S6 responded in a similar fashion to instances of below average precipitation as the other study catchments.

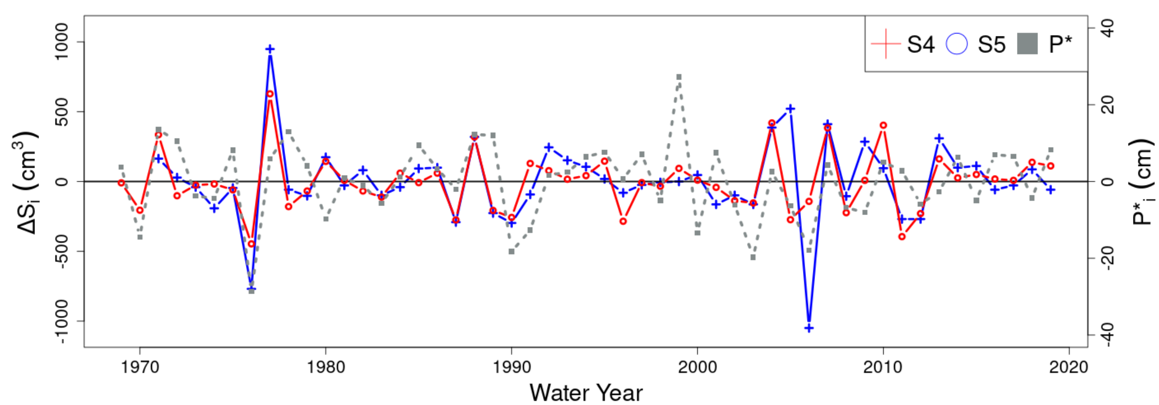

Figure 4: Time series for the annual precipitation anomaly (grey) and the S4 (red) and S5 (blue) total annual water storage change. The primary y-axis represents change in total annual storage while the secondary y-axis represents the precipitation anomaly.

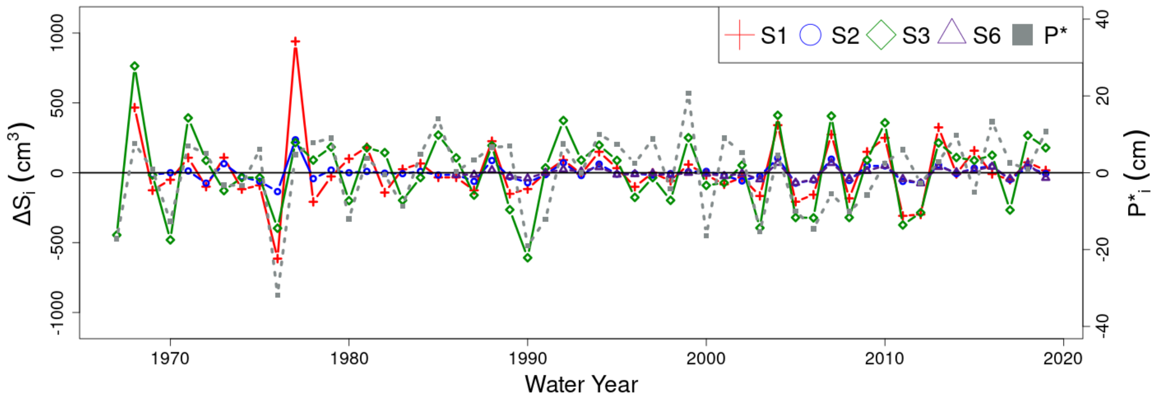

Figure 5: Time series for the annual precipitation anomaly (grey) and the S1 (red), S2 (blue), S3 (green), and S6 (purple) total annual water storage change. The S6 time series does not begin until the late 1980's. The primary y-axis represents change in total annual storage while the secondary y-axis represents the precipitation anomaly.

While there does appear to be a positive correlation between the precipitation anomaly and total water storage change in both units (Figures 4 and 5), it seems highly dependent on the storage state of the antecedent year. For example, there were large positive changes to total water storage in water year 1977 when the precipitation anomaly was only just above average (S_{1977} = 4.72cm, N_{1977} = 5.91cm) following large negative changes to water storage in the previous year. In contrast, high positive precipitation anomalies in water year 1999 (S_{1999} = 20.66cm, N_{1999} = 27.25cm) were accompanied by little to no positive change in total catchment water storage for any of the research catchments. In contrast to water year 1977, there was little to no negative change to water storage in the year(s) preceding water year 1999.

5.2 Correlation and Lag Analysis

5.2.1 Annual Precipitation Anomaly

There was a significant positive correlation ($\alpha = 0.05$) between the annual precipitation anomaly and total annual catchment water storage change for all six catchments (Figure 6). The S3 fen had the highest correlation coefficient ($r = 0.62$), while correlation coefficients ranged between 0.34 and 0.40 for the five raised bog catchments. When lagging the precipitation anomaly by one annual time step, significant negative correlations were found across all six catchments; correlation coefficients ranged between -0.50 and -0.60, with catchment S4 in the North Unit exhibiting the strongest correlation. Catchments S4 and S6 also exhibited a negative correlation at a lag of four water years ($r = -0.40$; $r = -0.44$). Non-significant correlations at the remaining lag steps were primarily positive, except for catchments S3 and S5 at lags two and five, respectively, and catchment S2 at a lag of three.

Results of the analysis on the annual upland soil moisture change component of the storage parameter show a positive correlation with the annual precipitation anomaly at a lag of zero ($r = 0.25$ to 0.54) for all catchments, excluding S1 and S2, which exhibit non-significant positive correlations (Figure 7). All catchments had a significant negative correlation at a lag of one, with correlation values ranging from -0.34 to -0.59. Similar to the total annual water storage lag analysis, S4 and S6 were found to have a significant negative correlation at a lag of four ($r = -0.39$; $r = -0.41$). The remaining catchments have non-significant negative correlations at a lag of four. The S6 catchment was also found to have a positive correlation at the terminal lag of five ($r = 0.48$). The direction of the remaining non-significant correlations at lags two, three and five are identical to the total annual water storage change results except for S3 at lag five, which is positive in this case.

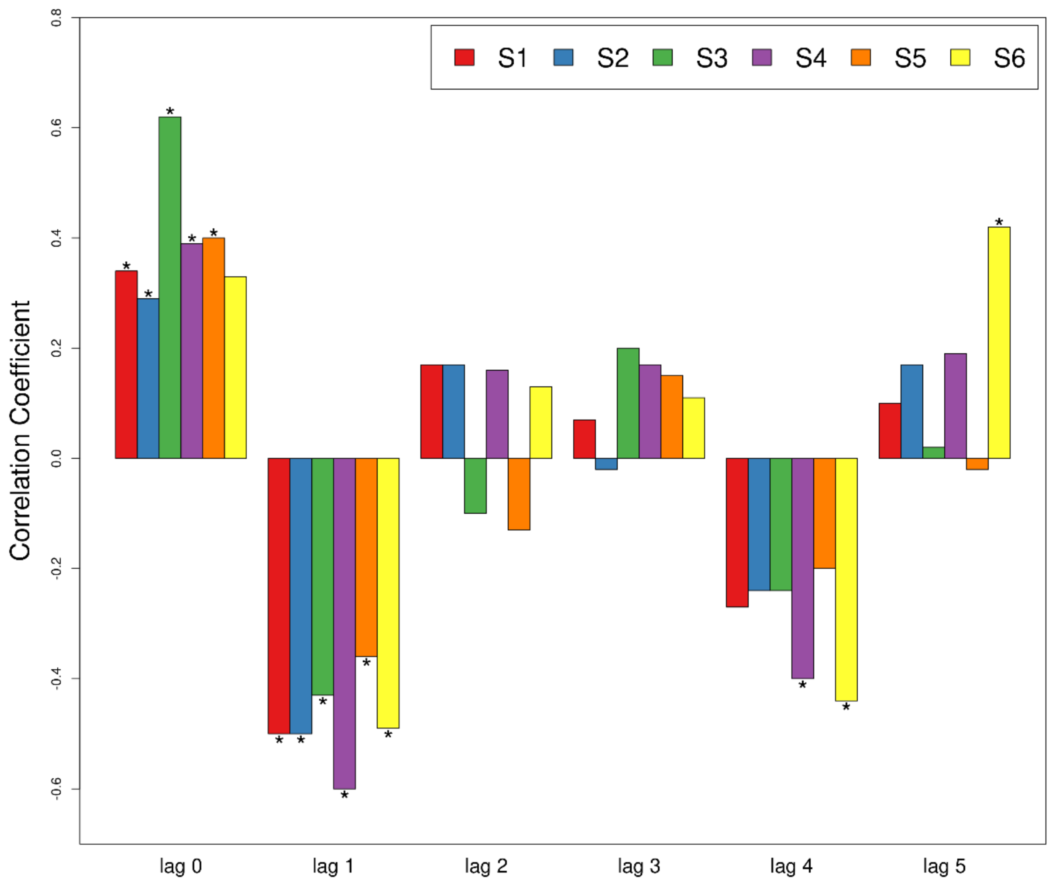

Figure 6: Annual catchment water storage change correlation analysis for all catchments, S1 through S6. Asterisks indicate a significant correlation at α=0.05. Catchments are identified by color, indicated in the upper right legend.

The peatland water table elevation lag analysis showed a positive correlation across all six catchments at a lag of zero (Figure 8); the S3 fen had the largest coefficient (r = 0.68), with the remaining catchments coefficients ranging between 0.42 and 0.45. At a lag of a single year, significant correlation coefficients ranged between -0.41 and -0.45, with non-significant negative correlations found at catchments S3 and S6. Catchment S6 also exhibited a significant positive correlation at a lag of three (r = 0.38) and a significant negative correlation at a lag of four (r = -0.44). At a lag of four, there were also significant negative correlations at catchments S4 (r = -0.38) and S5 (r = -0.30).

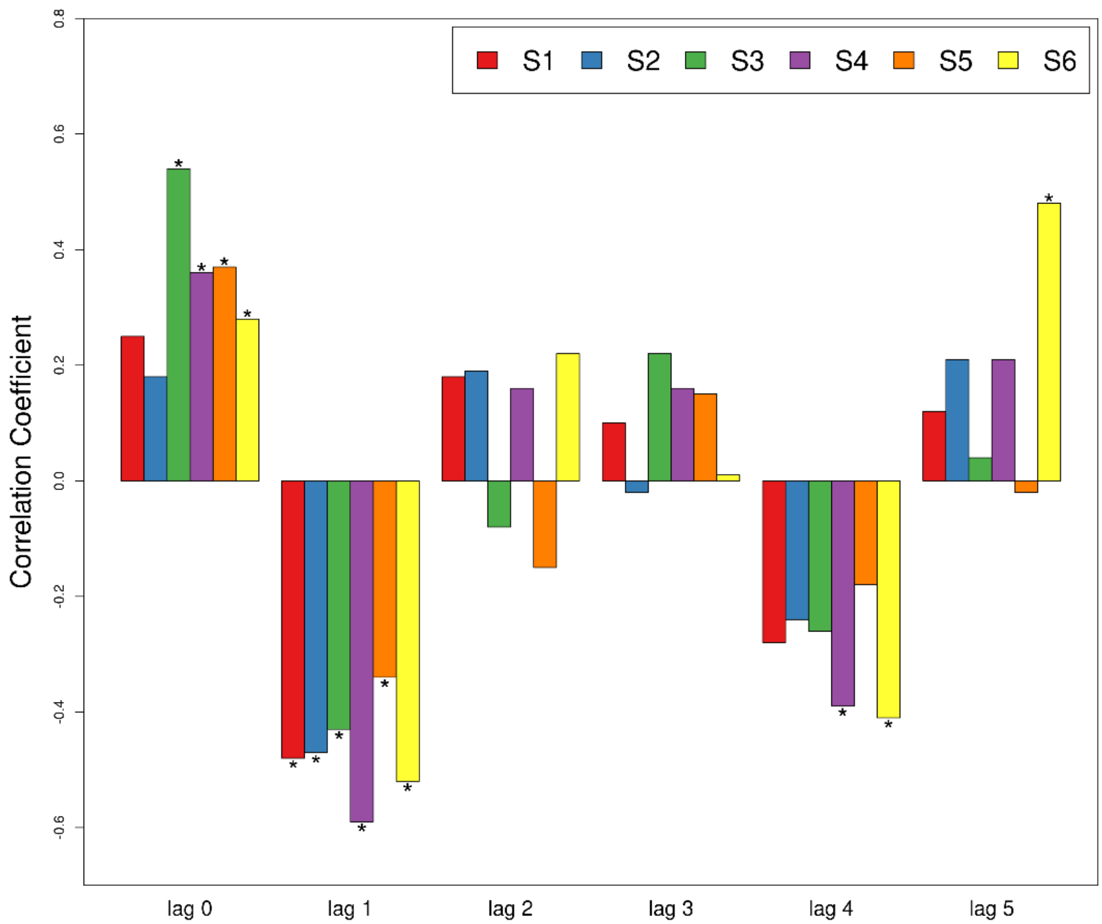

Figure 7: Annual catchment soil moisture change correlation analysis for all catchments, S1 through S6. Asterisks indicate a significant correlation at α=0.05. Catchments are identified by color, indicated in the upper right legend.

At a lag of two, all correlations were non-significant; these correlations were positive at catchment S1, S2, and S4, the remaining catchments exhibiting negative correlations. At lag five, S1, S2 and S6 have non-significant positive correlations while S3, S4, and S5 have non-significant negative correlations.

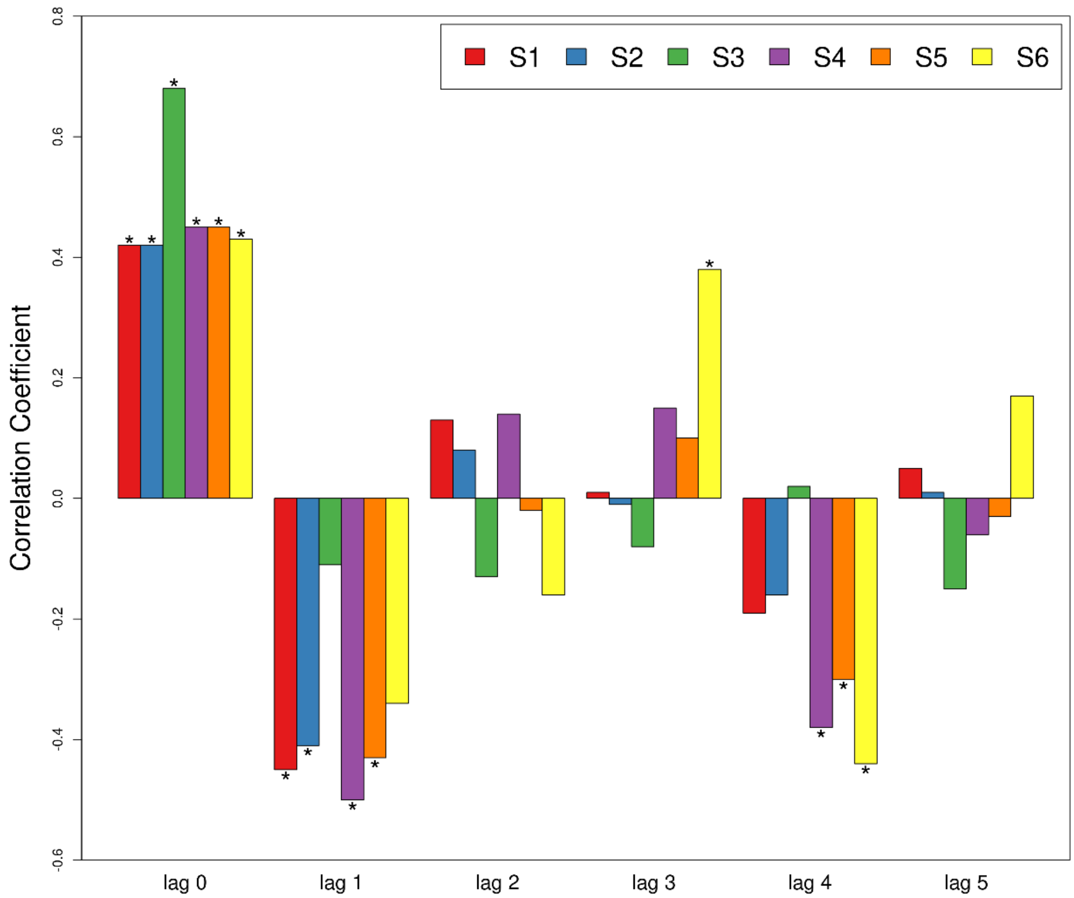

Figure 8: Annual catchment water table elevation change correlation analysis for all catchments, S1 through S6. Asterisks indicate a significant correlation at α=0.05. Catchments are identified by color, indicated in the upper right legend.

5.2.2 Seasonal Precipitation Anomaly

Seasonal analysis was conducted on lag zero and lag one only. At a lag of one, the winter precipitation anomaly for a given year was positively correlated with all catchments (Figure 9), but only significantly at catchment S3 (r = 0.27); total annual water storage change had a significant negative correlation at catchments S1 through S4, with non-significant negative correlations at S5 and S6. All correlations were negative and non-significant in the growing season with the exception of S3 at lag zero, which was

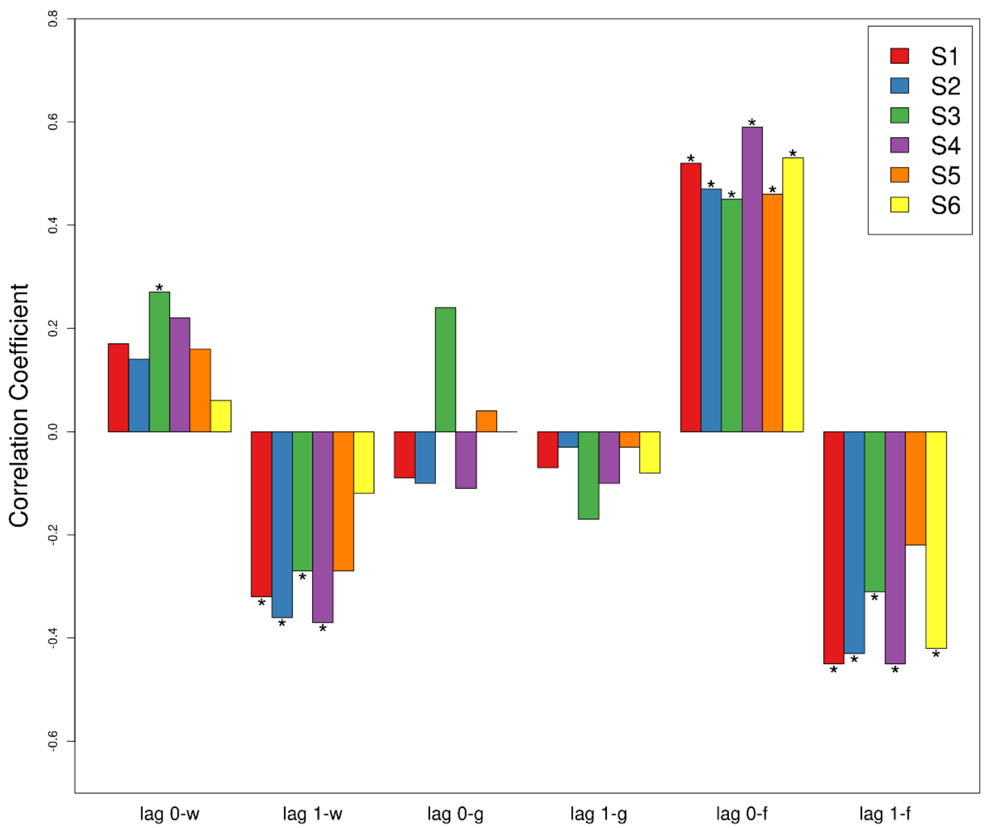

Figure 9: Total annual water storage change and seasonal precipitation anomaly correlation analysis for all catchments, S1 through S6. Seasons are broken up into winter (w), fall(f), and growing(g). Asterisks indicate a significant correlation at α=0.05. Catchments are identified by color, indicated in the upper right legend. Lag relationships are only assessed for lag zero and lag one.

positively correlated and non-significant. The fall precipitation anomaly was positively correlated with all catchments at a lag of zero, with correlation coefficients ranging between 0.45 and 0.59. At a lag of one, the fall precipitation anomaly had a significant negative correlation with all catchments (r=0.31 to r=0.35), except for S5.

Seasonal soil moisture results closely resemble the total annual storage change results (Figure 10). Positive but non-significant correlations were found across all catchments at a lag of zero between the annual soil moisture change and annual winter

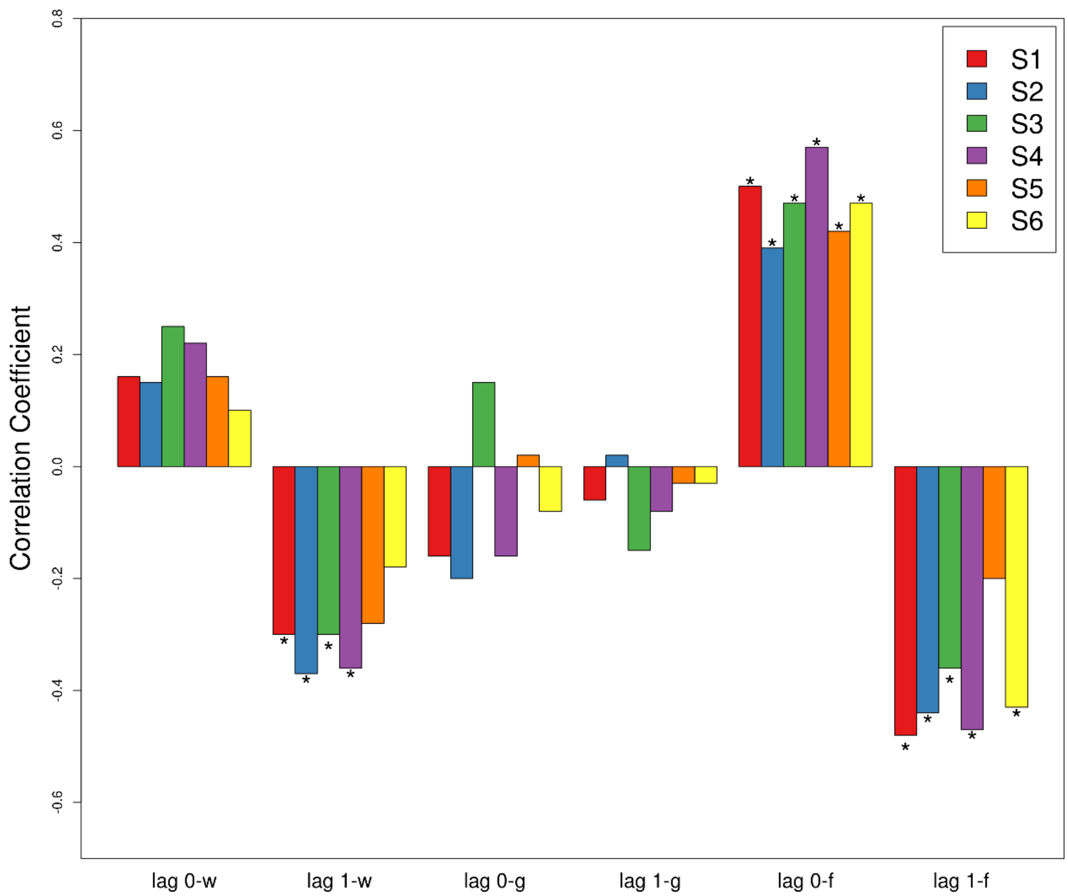

Figure 10: Total annual soil moisture change and seasonal precipitation anomaly correlation analysis for all catchments, S1 through S6. Seasons are broken up into winter (w), fall(f), and growing(g). Asterisks indicate a significant correlation at α=0.05. Catchments are identified by color, indicated in the upper right legend. Lag relationships are only assessed for lag zero and lag one.

precipitation anomaly; at a lag of one, there were significant negative correlations with the winter precipitation anomaly at catchment S1 through S4 (r = 0.30 to r = 0.37) and non-significant negative correlations at catchments S5 and S6. No significant correlations were found with the growing season precipitation anomaly at either lag one or lag zero. Lag zero growing season correlations were all negative except for S3 and S5. Lag one growing season correlations were all negative except for catchment S2. For the fall season precipitation anomaly, all catchments exhibited positive correlations at a lag of zero between 0.39 and 0.57; all lag one correlations were negative, with all but catchment S5 being significant between -0.36 and -0.48.

Correlation coefficients between annual water table elevation change and the winter precipitation anomaly (Figure 11) were positive at a lag of zero and negative at a lag of one; only catchment S3 was significant at a lag of zero (r = 0.32) while only S4 was significant at a lag of one (r = -0.28). At a lag of zero, all catchments were found to have a positive relationship using the growing season anomalies precipitation, with the S3 catchment having the only significant relationship (r = 0.42); at a lag of one, all catchments exhibit a non-significant negative correlation. Using the fall precipitation anomaly, all but the S3 catchment were found to have a significant positive relationship, with coefficients ranging from 0.44 to 0.62; at a lag of one, all catchments have a significant negative relationship (r = -0.27 to r = -0.36) except S3, which was found to have a non-significant positive correlation.

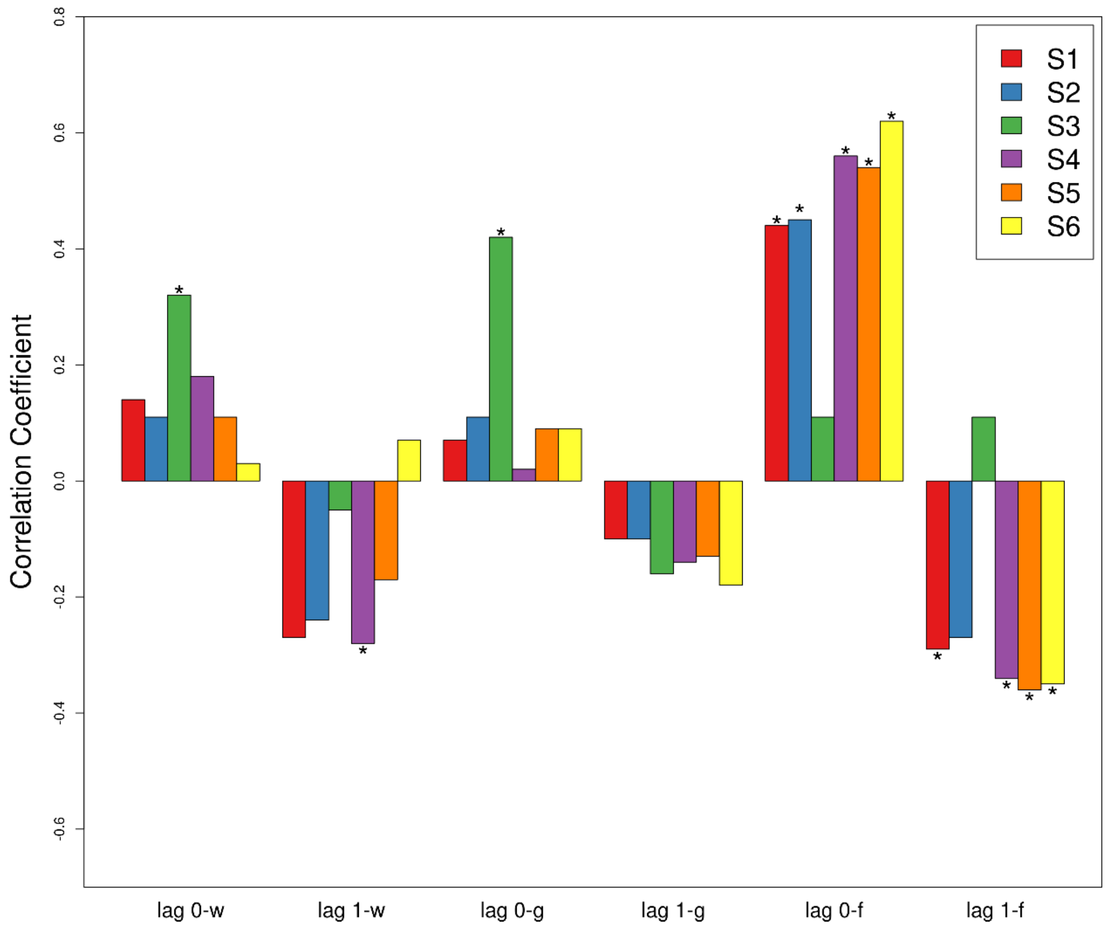

Figure 11: Annual peatland water table elevation change seasonal correlation analysis for all catchments, S1 through S6. Seasons are broken up into winter (w), fall(f), and growing(g). Asterisks indicate a significant correlation at α=0.05. Catchments are identified by color, indicated in the upper right legend. Lag relationships are only assessed for lag zero and lag one.

5.3 Wet/Dry Analysis

Regression analysis revealed a positive linear relationship between total annual water storage change and the annual precipitation anomaly in dry years for all catchments except S6 (Figure 12a). It was also found that a negative linear relationship exists between the precipitation anomaly in antecedent dry years and total annual water storage change in a given year at catchments S1, S2, S4, and S5 (Figure 12b). The coefficient of determination fell between 0.16 and 0.33 at a lag of zero (Table 4). The coefficient of determination fell between 0.24 and 0.42 at a lag of one. There was no significant relationship between the total annual water storage change and the precipitation anomaly in wet years or years preceded by a wet year.

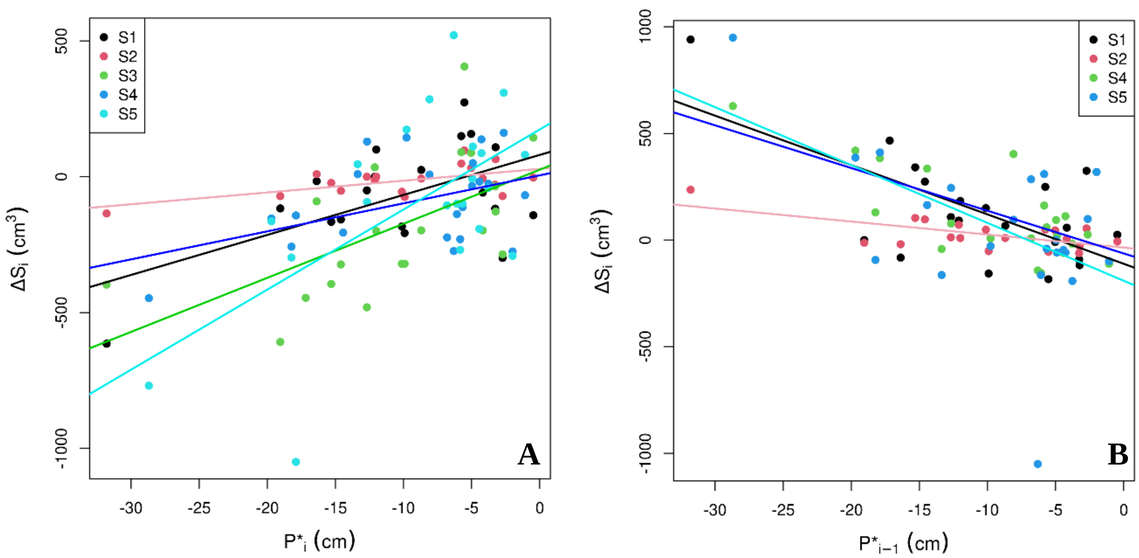

Figure 12: Results of the regression analysis between the total annual water storage change and the dry year precipitation anomaly (A) and the precipitation anomaly in antecedent dry years (B). All plotted regressions are significant.

When looking at the annual upland soil moisture change, there were significant positive relationships found only at catchments S3 (R^2 = 0.21) and S5 (R^2 = 0.28) in dry years (Figure 13a). When regressing annual soil moisture change against the precipitation anomaly in preceding dry years (Figure 13b), catchments S1, S2, S4, and S5 all exhibited a significant negative linear relationship; coefficient of determination for these catchments ranged between approximately 0.2 and 0.38 (Table 5). The precipitation anomaly in wet years and antecedent wet years did not produce any significant relationships.

Table 4: Coefficient of determination and p-value for catchments S1 – S5. Lag 0 indicates the relationship between the precipitation anomaly and water storage change in a dry year. Lag 1 indicates the relationship between water storage change and the precipitation anomaly in a preceding dry year.

Site	Lag 0 R^2	Lag 0 p-value	Lag 1 R^2	Lag 1 p-value
S1	0.27	0.01	0.40	0.002
S2	0.31	0.008	0.38	0.003
S3	0.31	0.0006	NA	NA
S4	0.16	0.04	0.42	0.0006
S5	0.33	0.004	0.24	0.01

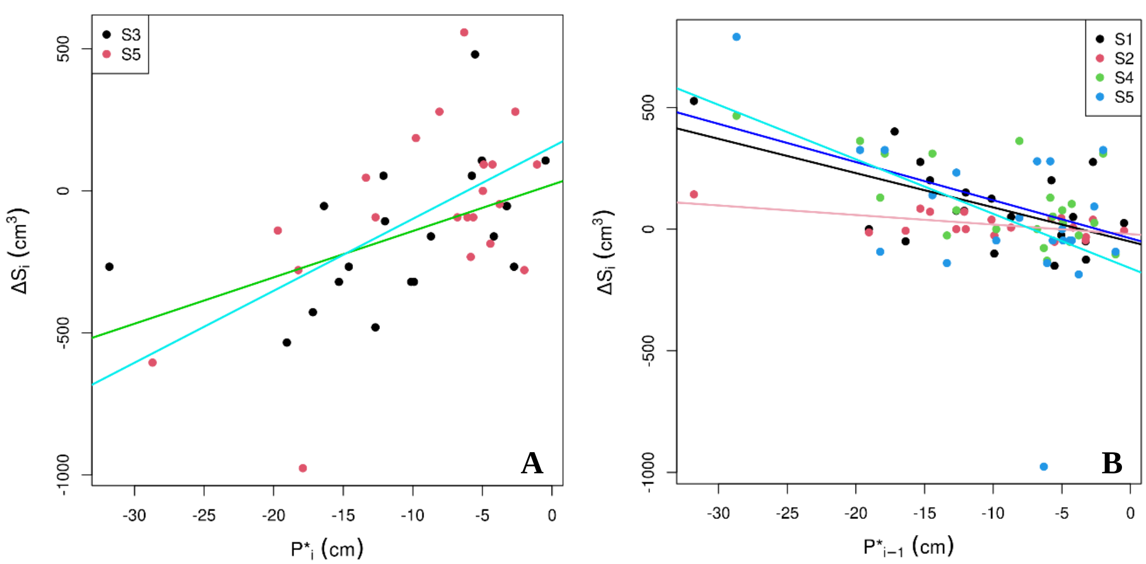

Figure 13: Results of the bivariate time series regression analysis for the annual upland soil moisture change when considering dry years (A) and years that followed dry years (B). All regressions pictured are significant.

Table 5: Coefficient of determination and p-value for catchments S1 – S5. Lag 0 indicates the relationship between the precipitation anomaly and annual upland soil moisture change in a dry year. Lag 1 indicates the relationship between soil moisture change and the precipitation anomaly in a preceding dry year.

Site	Lag 0 R^2	Lag 0 p-value	Lag 1 R^2	Lag 1 p-value
S1	NA	NA	0.31	0.006
S2	NA	NA	0.31	0.008
S3	0.21	0.02	NA	NA
S4	NA	NA	0.38	0.001
S5	0.28	0.008	0.20	0.02

Annual peatland water table elevation showed a significant linear relationship with the dry year precipitation anomaly at all but the S6 catchment (Figure 14a), with coefficients of determination between approximately 0.16 and 0.33 (Table 6). Results of the regression with the antecedent dry year precipitation anomaly resulted in significant negative linear relationships for all catchments except for S3 and S6 (Figure 14b); the coefficient of determination for these relationships was between 0.24 and 0.42 (Table 6). Again, no relationship between the storage component and the wet year or antecedent wet year precipitation anomaly was found.

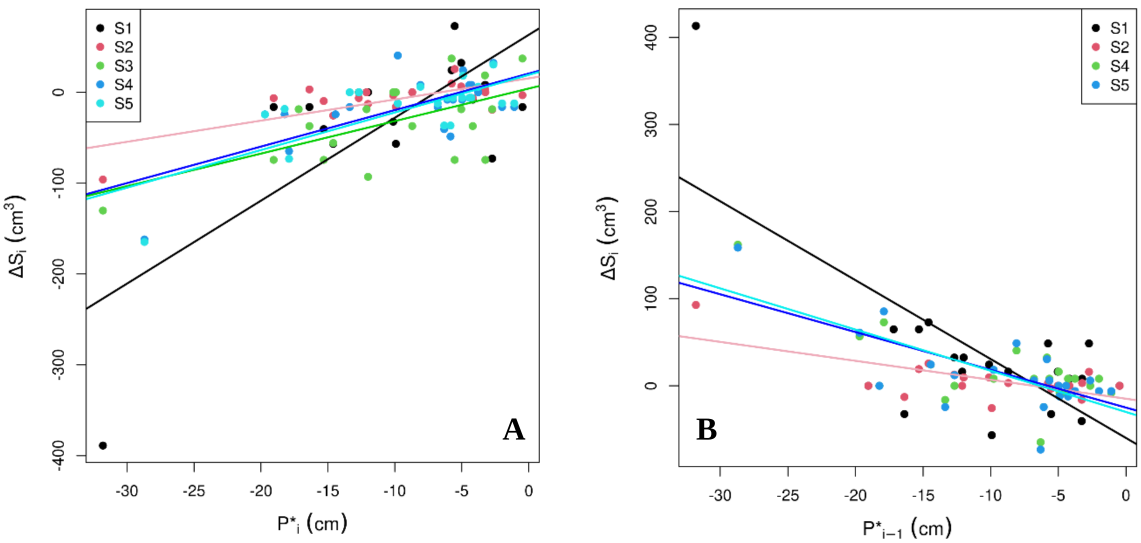

Figure 14: Results of the bivariate time series regression analysis for the annual peatland water table elevation change when considering dry years (A) and years that followed dry years (B). Catchment S6 was not found to be related to the precipitation anomaly in either of these scenarios.

Table 6: Coefficient of determination and p-value for catchments S1 – S5. Lag 0 indicates the relationship between the precipitation anomaly and annual water table elevation change in a dry year. Lag 1 indicates the relationship between water table elevation change and the precipitation anomaly in a preceding dry year.

Site	Lag 0 R^2	Lag 0 p-value	Lag 1 R^2	Lag 1 p-value	
S1	0.27		0.01	0.40	0.002
S2	0.31		0.008	0.38	0.003
S3	0.31		0.006	NA	NA
S4	0.16		0.04	0.42	0.0006
S5	0.33		0.004	0.24	0.01

5.4 Final Models

Using the results of the preceding analysis as guidelines for model parameters, multivariate time series regression with ARIMA error models were produced for the total annual water storage change. All initial regression models exhibited a first-order moving average structure, requiring the addition of a moving average (MA) error to the model, with the exception of the S3 fen. The regression with first-order MA errors can be written in the form:

$$y_i = c + \beta_1 x_1 + \beta_2 x_2 + n_i \qquad \text{Eqn. 21}$$

where:

y_i = observed total water storage change for water year i
c = regression model intercept
β = standardized regression coefficient
n_i = first order moving average

The simple linear model for catchment S3 lacked any autocorrelative structures, and therefore was implemented as a multivariate linear regression model. When regressing the observed values against the fitted values for catchments S1, S2, S4, and S5 against the fitted values from the final models (Figure 15), coefficients of determination between 0.49 and 0.70 were achieved (Table 7). The coefficient of determination between the total annual water storage change and the annual precipitation anomaly at catchment S3 was 0.48 (Figure 15).

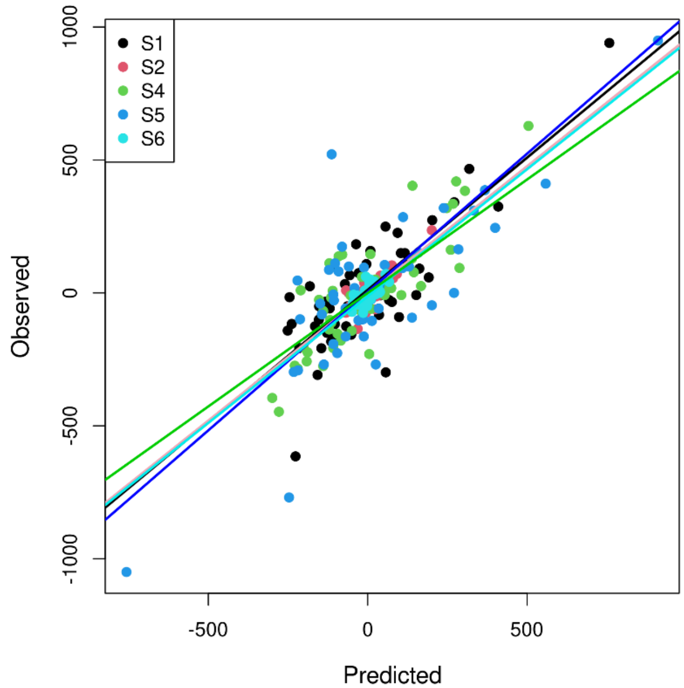

Figure 15: Observed values for the total annual water storage change regressed against the predicted values produced by the regression with first order MA error models. All catchments represented in the plot are those that contain a central raised bog.

Table 7: Coefficient of determination and related p-value for the bog watershed observed/predicted regressions.

Site	R^2	**p-value**
S1	0.64	$6.42e^{-13}$
S2	0.59	$3.30e^{-11}$
S4	0.70	$1.25e^{-15}$
S5	0.62	$1.53e^{-11}$
S6	0.49	2.83^{-6}

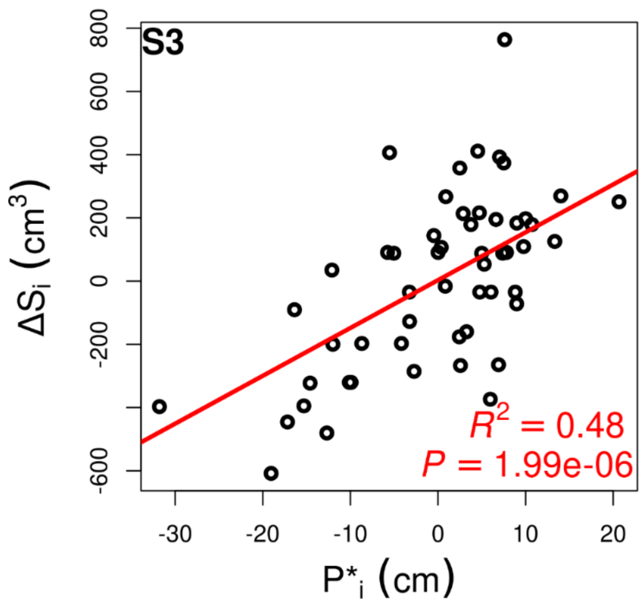

Figure 16: Results of the multivariate linear regression between the S3 total annual water storage change and the South Unit precipitation anomaly. Coefficient of determination and p-value are included in the bottom right corner of the plot. The S3 catchment model was the only model that did not require any accounting for autocorrelation.

Table 8: Final models for each catchment as well as the coefficient of determination and p-value.

Catchment	**Model**	**R^2**	**p-value**
S1	$y_t = 1.22 + 10.12x_1 - 10.43x_2 + n_t$	0.64	$6.42e^{-13}$
S2	$y_t = -1.18 + 2.71x_1 - 2.83x_2 + n_t$	0.59	$3.30e^{-11}$
S3	$y_t = 3.01 + 15.13x_1 - 9.02x_2 + \varepsilon_t$	0.48	$1.99e^{-6}$
S4	$y_t = -9.18 + 8.36x_1 - 10.41x_2 + n_t$	0.70	$1.25e^{-15}$
S5	$y_t = 0.47 + 11.50x_1 - 12.39x_2 + n_t$	0.62	$1.53e^{-11}$
S6	$y_t = -4.13 + 2.40x_1 - 1.76x_2 + n_t$	0.49	2.83^{-6}

6. DISCUSSION

Research at six small headwater peatland catchments at the Marcell Experimental Forest in northern Minnesota found that variability in total annual precipitation was a significant predictor of total annual catchment water storage change at all six of the study catchments. The storage/precipitation relationship can be modeled as a bivariate linear regression fit with a first order MA error for catchments S1, S2, S4, S5, and S6, while the S3 fen can be modeled as a simple bivariate linear regression. All six study catchments had the precipitation anomaly in a given year as well as the precipitation anomaly at a lag of one included as regressors in their respective models.

The cross-correlation analysis used to determine useful regressors for the catchment storage models revealed a moderately strong interannual dependence of the catchment water storage change on precipitation variability. Total catchment storage was found to change in response not only to precipitation inputs in a given year, but also to precipitation inputs from the previous year across all six study catchments, regardless of peatland type. These results point to a degree of watershed memory (the influence of hydrologic conditions in previous years on the hydrologic condition of the current year) influencing overall hydrologic function at the MEF. Regression analysis revealed that these relationships are primarily driven by catchment sensitivity to below average precipitation inputs across all study catchments at both the annual and interannual scale.

Total catchment storage tends to drop in years of below average precipitation while these losses are typically recovered within one or two annual time steps at the majority of the study catchments, suggesting certain homogeneity in hydrologic response.

However, analysis of the individual upland soil moisture and peatland water table elevation as they relate to the annual precipitation anomaly showed variation between catchments in regard to the individual storage pools. Differences between WTE and SM response were more pronounced when only considering total seasonal precipitation anomalies.

6.1 Catchment Response to Antecedent Precipitation

Interannual relationships between hydrologic function and past precipitation have also been shown to be present at the Coweta Hydrologic Laboratory in North Carolina (Nippgen et al., 2016), where researchers found that past precipitation was equally important to a given years runoff ratio as the contemporaneous precipitation at an annual time scale. The same study also found the storage state in antecedent years, as set by precipitation, to be important to annual runoff. Studies in Nebraska (Istanbulluoglu et al., 2012) and the Amazon (Tomasella et al., 2008) produced similar results, in which it was found that variability in precipitation was influential on the subsequent years water balance due to carryover from groundwater storage.

The significant correlations at a lag of one year in my research are all negative, indicating recovery from dry years historically occurs within one year. Regarding the upland soil moisture storage, this is likely due to precipitation in the previous year providing a buffer against evapotranspiration during the growing season. As shown by Hewlett and Hibbert (1963), ~11m^3 saturated soils were able to sustain outflow through a ~14m concrete trough at a 40% slope for up to 145 days, with 5% of the water still draining in the last 95 days. It was concluded that if the results of the experiment were to be scaled up to the catchment level, this rate of drainage could sustain baseflow through a growing season. Upland side slopes at the MEF can be as low as 5% (Verry and Jannsens, 2011) while the Koochiching till and overlying glacial flower are both of low hydraulic conductivity; these features work in conjunction to slow both deep infiltration and subsurface lateral flow of soil water, supporting this buffer capacity within the MEF catchments.

Peatland water table storage also appears to be influenced by this watershed memory effect, as all but the S3 fen and S6 bog exhibit negative correlations with the precipitation anomaly at a lag of one year. Peatland water tables generally remain close to the surface, showing little variation in depth from year to year (~0.5m) (Verry, 1984). High water table levels would leave little room for positive changes in the storage pool, even under the influence of above average precipitation. Furthermore, while some recharge may come from upland soils, these stores are fed primarily by precipitation inputs (Bay, 1968). Water elevation tends to peak following winter snowmelt and ground thaw, while holding steady during the fall season, unless influenced by excess or lack of precipitation inputs (Bay, 1968) that may dampen or exacerbate recession via streamflow and a slowly falling water table. Fall season dynamics imply that should fall precipitation fall above or below the mean seasonal precipitation, the start of the next water year (November) will influence the magnitude and direction of the annual storage change, as it is representative of the next water years (Nov-Oct) starting storage state.

A lack of correlation in the S3 fen water table at the lag one time step may be due to its connection with the regional groundwater aquifer, with water table elevations in

MEF fens known to be influenced by the surrounding groundwater system (Bay, 1966, Bay 1967, Sebestyen, 2011a). While other work has shown the importance of both regional and local recharge (Sampath et al., 2015) to fen water storage, direct recharge from antecedent local precipitation may be overshadowed by regional climatic conditions affecting the regional aquifer. The S3 fen has also demonstrated a lack of overwinter drawdown when compared to the S2 bog (Bay, 1968), further highlighting the regional groundwater influence. The mechanism for the lack of lag one correlation at S6 is unclear, however, it contains the smallest of the six central peatlands, and is composed of the Seelyville organic soil series (Figure 2), which is unique to this catchment. Peat type can be an influence on water table levels (Bay, 1968) due to drivers of storage change (e.g., infiltration) being dependent on peat soil characteristics (Rezanezhad et al., 2016; Stockstad et al., 2001). Differences in the analysis results for the S6 catchment may be due to the effects of upland harvesting that took place from 1980-1981; due to its truncated period of record, the S6 catchment also required time series pre-whitening to account for non-stationarity found in the precipitation data, which may also be of some influence on the final results of the correlation analysis.

Multiple confounding lag responses were found in all three of the annual correlation analysis, particularly at lags for and five. Sometimes cross-correlation can be affected by random functions within a data set, and this is how they were interpreted in this study for the purpose of defining useful regressors in the final storage models. However, there is some evidence for long term storage patterns attached to wet/dry climate cycles in North American wetlands (van der Valk, 2005). Short term, the catchment time series in this study clearly illustrate some degree of oscillatory behavior

(Figure 3; Figure 4), while larger lag correlations may be representative of some larger climate cycle.

6.2 Sensitivity to Below Average Precipitation Inputs

While no significant relationships were found between years of above average precipitation and catchment storage change, hydrologic function was found to be sensitive to below average precipitation, both within a given year and in years that followed a drier year. Periods of below average precipitation can negatively impact the hydrologic function of peatlands (Dise, 2009, Beyer et al., 2021), though peatlands can adjust to water stress, at least in the short term, due to high storage capacity and capillary-wicking (Ingram, 1987; Lapen et al., 2000). This resilience can be seen in the raw time series (Figure 3; Figure 4), with losses during dry years being largely replaced within a year or two of storage drawdown events. Negative lags in the correlation analysis (Figure 5 - Figure 10) also illustrate this resilience, as storage pools tend to move in the opposite direction of the precipitation anomaly in an antecedent year. Though no extended drought periods are present in the data used for this study, other studies have concluded that long term dry periods lead to drawdown, which then leads to increased carbon emissions from peat mass carbon stores (Leifeld et al., 2019). Emissions can be appreciable, particularly when considering the relative land occupancy of global peatlands (Tiemeyer et al., 2016; Tubiello et al., 2016). Though the MEF storage pools appear to recover quickly according to my results, multiple years of well below average precipitation would likely lead to extended periods of low water table elevation.

The upland soils show a less cohesive response to dry years across the six sites, with only catchments S3 and S5 exhibiting linear relationships with the precipitation anomaly in dry years. However, this is not true when considering the precipitation in an antecedent dry year, as all but the S3 fen and S6 bog show a significant correlation with the precipitation anomaly. Results of the correlation analysis indicate stronger seasonal components to the soil moisture/precipitation variability rather than annual, which are expressed in the data primarily as negative correlations with seasonal precipitation anomalies. These correlations are discussed in the following section.

6.3 Seasonal Inputs and Catchment Storage

6.3.1 Winter Precipitation

Results of this study highlight the importance of the winter snowpack and subsequent spring melt to maintaining storage levels at the MEF. Significant negative correlations were found between the annual winter precipitation and the total annual water storage change across all catchments at a lag of one, with the exception of the S5 and S6 bogs. This relationship is primarily present as changes in upland SM, with little correlation at either lag one or zero between the seasonal precipitation anomaly and peatland WTE change. Water introduced in the winter months at the MEF is primarily snow, which acts as a temporary water storage pool (Musselman et al, 2021); this results in delayed introduction to the wider storage pool until the spring, when snowmelt is reintroduced as liquid water and can be partitioned into the systems hydrologic outputs. The presence of soil frost can inhibit infiltration of snow melt in the central organic soils during the Spring melt, decreasing recharge to the water table storage pool and largely being partitioned as runoff, and subsequently streamflow (Jones et al., 2023).

All catchment WTE lack correlation with the antecedent winter precipitation anomaly, with the exception of the S4 bog. The influence of winter precipitation in the bog catchments is possibly inhibited by the perched nature of these water tables, resulting in higher frost content of the acrotelm, causing a large fraction of snowmelt to be partitioned as streamflow rather than storage recharge (Jones et al., 2023). The S3 fen also lacks this relationship, but as a fen is by definition connected to the aquifer (Klove et al., 2011, Sebestyen, 2011a), with groundwater from this aquifer a significant contributor to the water budget in this catchment (Bay, 1967; Sander, 1978; Boeye and Verheyen, 1992). Catchment S3 is also the only catchment to have a significant correlation with winter precipitation at a lag of zero, indicating a higher level of infiltration in the spring.

Snowmelt is a highly important hydrologic control on soil moisture in cold climate regions with seasonal snowfall, and has been shown to be important to the wetting and restoration of catchment hydrologic connectivity following the winter season (McNamara et al., 2005; Blankinship et al., 2014). However, the influence on hydrologic function is highly dependent on timing and amount of the seasonal snowpack (Yeh et al., 1983; Shinoda, 2001), as large snow amounts lead to later snow disappearance and higher soil moisture levels (Douville and Royer, 1996, Shinoda, 2001). The amount of infiltration into the mineral soil is also reliant on the properties of the soil series (Ambadan et al., 2017), as well as the degree of seasonal soil freeze. It has been shown that changes in the soil moisture content of a catchment can be linked to precipitation variability, while the influence of infiltration via snow water equivalent from the winter snowpack decays relatively quickly at monthly and seasonal scales (Ambadan et al., 2017). Results from my study at the MEF show a larger interannual soil memory effect, with winter precipitation variability having a larger influence on the following years storage change than in the contemporaneous year. On the annual scale, we are effectively looking at storage states at two time points, the start and end of a water year. By approaching storage in this fashion, we see the effect winter precipitation has on the starting storage state of next year's storage pool. Keeping in mind that dry years are more influential to the precipitation/storage relationship, its likely these correlations are illustrating how a below average snowpack in a previous year will result in larger potential gains in the storage pool the following year, as little to no buffer is available during growing season increase in evapotranspiration outflows.

6.3.2 Fall Precipitation

Seasonal correlations show a significant relationship between the total annual water storage change and the fall precipitation anomaly, both at a lag of zero and a lag of one. Water years in this study begin and end in November, immediately following the defined fall season. This results in fall precipitation, the shortest of the three defined seasons, contributing to both the starting and ending storage state in a given water year to a greater degree than the other seasons. Bog water tables have been shown to remain relatively stable during the fall season, baring out of the ordinary precipitation patterns (Bay, 1968). Again, recalling the sensitivity of the storage pool to below average precipitation, and keeping in mind the position of the fall season relative to the defined water year, it follows that losses to water storage in the fall would affect both a given year and the next years storage state. Winter months coincide with low points in annual storage (Bay, 1968). Though water yield to streamflow ceases at all but the S3 catchment in winter (Sebestyen, 2011a), there is still an element of deep seepage present at the bog catchments (Nichols and Verry, 2001); fall precipitation is likely working to replenish or maintain storage levels at the end of the year, while also buffering against potential water yield over the winter season. The S3 fen WTE lacks any correlation with either lag of the fall precipitation anomaly, highlighting the differences in hydrologic function between the two catchment types.

7. CONCLUSION

My research indicates significant influence of both antecedent precipitation and seasonal precipitation on hydrologic function of both the upland mineral soils and central peatland organic soils at the MEF. These relationships are driven by seasonal winter and fall precipitation inputs, as well as dry climatic conditions. If predicted changes to precipitation regimes come to pass in mid to high latitude cold climate regions, these relationships stand to be altered. Alterations to timing and magnitude of precipitation events are likely to disrupt typical hydrologic functioning for catchments in these regions. As disproportionately large carbon sinks, boreal peatlands are of particular concern when considering these possibilities. While there is some evidence to suggest a certain resilience in peatland catchments to losses in water storage that can lead to peat degradation and ultimately carbon store emissions, further research is needed to assess this dynamic on long term scales. In particular, research into the effects of multiyear drought and response to timing of snowmelt could help to further describe future scenarios under the influence of climate change.

The models produced in this study to explore the precipitation/storage relationship at the MEF were able to describe a significant amount of the overall water storage change variability within the study catchments; future modeling efforts should consider the inclusion of antecedent precipitation values. Models could also be refined by modeling the WTE and SM pools separately, using their unique seasonal correlations as regressors rather than approaching them as a combined variable.

Peatland environments have unique hydrological properties that can influence the catchment water balance. In general, peatlands have a high-water storage capacity due to their high organic matter content, which allows them to retain water for longer periods of time. This means that during dry years, peatlands have a potential buffer to storage drawdown. However, the relationship between peatlands and water availability is complex and can vary depending on several factors, including climate, topography, vegetation, and land use disturbances. Overall, the effect of precipitation variability on storage and water availability in peatland catchments depends on the specific characteristics of the peatland and the surrounding ecosystem, as well as the local and regional climatic conditions. Further research is needed to better understand the complex interactions between peatlands, upland mineral soils, and meteorologic inputs under different climate scenarios.

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APPENDIX A

The value used in translating the observed peatland water table elevation to the water filled porosity of the acrotelm (WFP) was derived from an unpublished experiment that took place in the S1 catchment at the MEF beginning in September 2011. A 133m^2 corral and reservoir (Figure A-1) was built on a section of the central peatland and flooded with 6,000gal of water to examine water table elevation response. Water levels

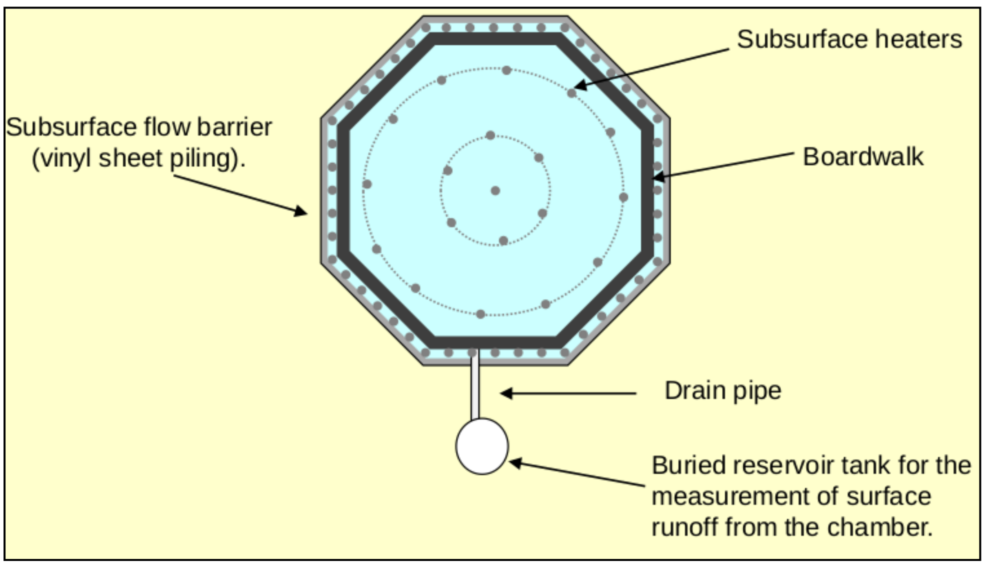

Figure A-1:Corral and reservoir schematic. The octagonal corral is 133m^2 in area. Water table elevation was measured to a depth of 3m. The total volume considered was 39.9m^3 .

were logged hourly using nested piezometers (Figure A-2) to a 3m depth and verified by weekly manual measurements. The WFP accounted for approximately 57% of the acrotelm volume (Eqn. 22; Eqn. 23) with the remaining 43% occupied by organic matter.

$$133m^2 * 0.30m = 39.9m^3 = 10,540gal \qquad \text{Eqn. 22}$$

$$6{,}000gal \div 10{,}540gal = 0.57 \qquad \text{Eqn. 23}$$

There was found to be two separate hydrologic regimes at the site (Figure A-3); this was suggested to be the result of the low hydrologic conductivity of the deep catotelm peats. There was little to no response at the 3.0m and higher water table elevation in the surface piezometers compared to the deeper observation depths.

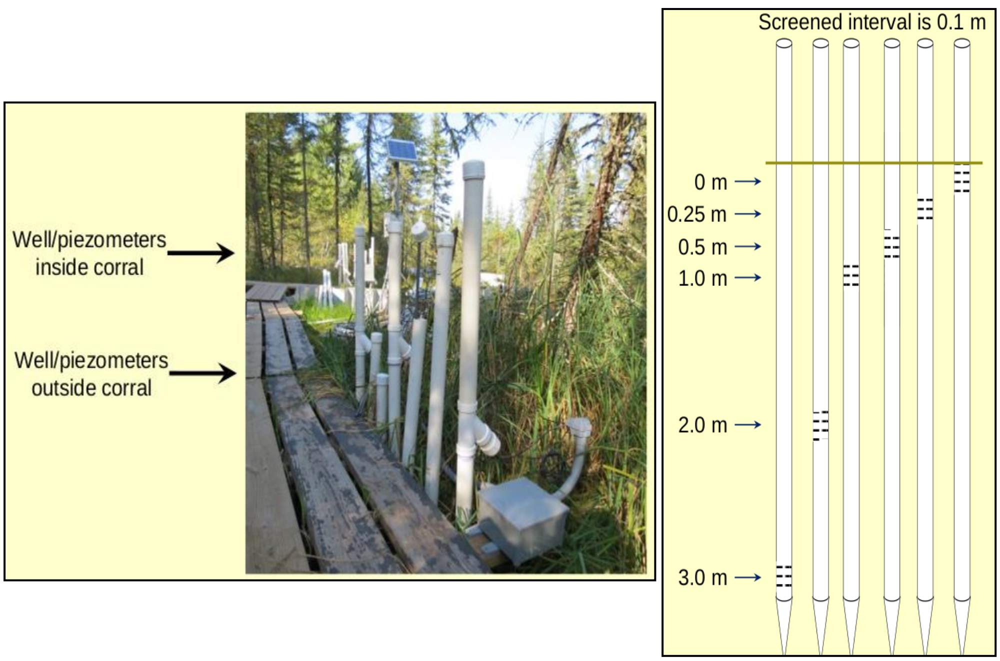

Figure A-2: Nested piezometers in the field (left). Measurements were taken hourly to a depth of 3m (right).

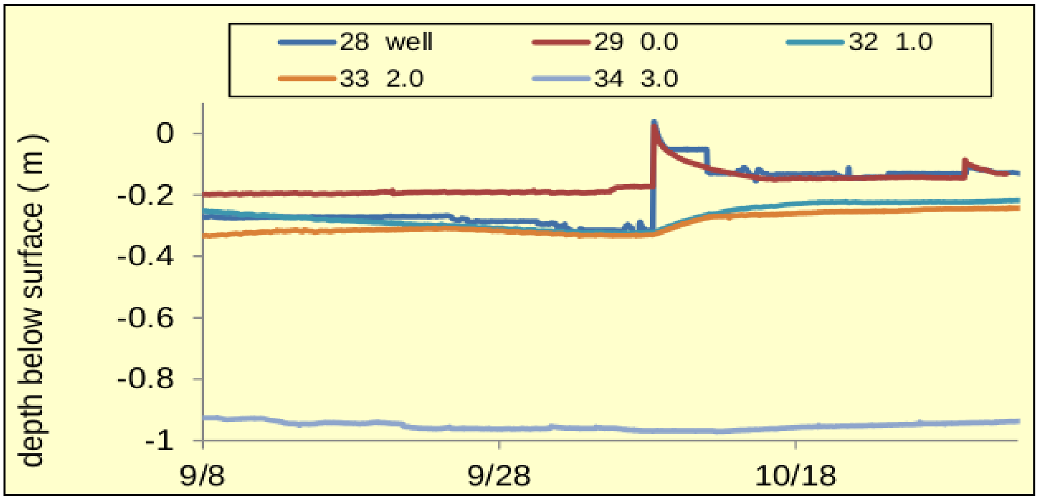

Figure A-3: Water table depth below surface (m). In the top legend, well ID is indicated by the left number and depth of measurement is indicated by the right number.