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Characterizing The Contribution Of Groudwater In The Water Budgets Of Wetlands Using GIS Groundwater Modeling And The Field Measurement Of Water Quality

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CHARACTERIZING THE CONTRIBUTION OF GROUDWATER IN THE WATER BUDGETS OF WETLANDS USING GIS GROUNDWATER MODELING AND THE FIELD MEASUREMENT OF WATER QUALITY

by

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This thesis, submitted by Brianna F. Speldrich in partial fulfillment of the requirements for the Degree of Master of Science from the University of North Dakota, has been read by the Faculty Advisory Committee under whom the work has been done and is hereby approved.

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This thesis is being submitted by the appointed advisory committee as having met all of the requirements of the School of Graduate Studies at the University of North Dakota and is hereby approved.

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Brianna Fuller Speldrich
August 2, 2019

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ABSTRACT

Wetlands provide many benefits to society, including flood attenuation, groundwater recharge, water quality improvement, and habitat for wildlife. Because their structure and function are sensitive to changes in hydrology, characterizing the water budgets of wetlands is crucial to effective management and conservation. The groundwater component of a budget, which often controls resiliency and water quality, is particularly difficult to estimate and can be costly, time-consuming, and invasive. This study used a GIS approach using a digital elevation model (DEM) and the elevations of lakes, wetlands, streams, and hydric soils to produce a water-table surface raster for a portion of the Itasca Moraine. The water-table surface was used to delineate groundwatersheds and groundwater flow paths for lakes and wetlands and map recharge and discharge rates across the landscape. pH and specific conductance, which depend on the hydrological processes that dominate a wetlands water budget, were measured in the field to verify this modeling technique. While the pH of surface waters varied in the study area, specific conductance increased from 16.7 to 357.5 μS/cm down gradient along modeled groundwater flow paths which revealed increased groundwater interaction. These results indicate that GIS tools and public-domain elevation datasets can be used to map and characterize the contribution of groundwater in the water budgets of lakes and wetlands in the Itasca Moraine. Applied elsewhere, this cost-efficient and less invasive modeling technique should be of use to natural resource managers who need to access the

vulnerability of lakes and wetlands to changes in land use, groundwater development, and climate change.

CHAPTER 1: INTRODUCTION

Characterizing the water budgets of lakes and wetlands is crucial for effective management. Wetlands provide many benefits to society, including flood attenuation, groundwater recharge, water quality improvement, and habitat for wildlife including those that are threatened or endangered (Mitsch and Gosselink 2000). Unfortunately, the benefits of wetlands were not adequately understood and the legal protection of wetlands was not a priority until the 1970s. As a result, it is estimated that the United States has lost approximately 50 percent of its wetlands since European settlement (Mitsch and Gosselink 2000). It is especially important to understand the influence of groundwater in these systems as surface water and groundwater are interconnected and wetlands are sensitive to changes in their hydrology (Winter 1999; Winter et al. 1998). Groundwater influence can vary substantially among even adjacent and nearby wetlands, and often groundwatersheds, or groundwater catchment areas (Haitjema 1995), do not coincide with surface watersheds, complicating matters (Winter et al. 2003). Characterizing the groundwater component of the water budgets of wetlands can help natural resource managers better understand how these features will respond to changes in their hydrology and help assure effective wetland conservation.

Traditional techniques of water budget estimation require extensive equipment, long-term field data collection, and copious labor for characterization which can quickly become costly. In addition, some field areas may be hard to access, and field visits and the installation of equipment may be detrimental to the ecosystem. In this report, it is hypothesized that the use of GIS tools and the measurement of pH and specific conductance can be used to characterize the contribution of groundwater in the water budgets of wetlands in a more cost-effective and less invasive way.

For an area within north-central Minnesota's Itasca Moraine, the contribution of groundwater in the water budgets of lakes and wetlands was characterized with a spatial modeling technique using several GIS tools. Specific conductance and pH of surface waters, which depend on the hydrological processes that dominate a wetland's water budget, were measured in the field and used to verify and support this innovative spatial-modeling technique. The technique requires relatively inexpensive field measurement equipment and the use of an industry-standard GIS program. In addition, it requires only readily available, public-domain state and federal data.

There are four stepwise components to this technique. The first component involves the analysis of a digital elevation model (DEM) and lake, stream, wetland, and hydric soil elevation data to generate a local and regional water-table surface. The second component involves using the water-table surface to delineate groundwatersheds for lakes and wetlands. The groundwatersheds can then be compared to topographic (surface) watersheds to compare the relative importance of groundwater and surface water for these features. The third component involves using the water-table surface and analyzing groundwater flow paths as well as where lakes and wetlands lie in relation to these flow paths to characterize the groundwater component of their water budget. The fourth component involves using the water-table surface, hydraulic conductivity data estimated from soils, and the maximum depth of groundwater and surface water interaction estimated from an analytical technique to map recharge and discharge across the landscape. This will identify strong recharge and strong discharge areas as well as their proximity to lakes and wetlands in the study area. Specific conductance and pH measurements collected in the field are used to corroborate with the results of the mapping analysis.

Background and Previous Work

Surface water and groundwater are interconnected in most types of terrain. Because of this relationship, effective water resource management requires an understanding of the interaction that is occurring between the two (Winter 1999). Lakes and wetlands in glacial terrain interact with groundwater in one of three ways. They can recharge groundwater (outflow), receive groundwater discharge (inflow), or both recharge groundwater and receive groundwater discharge (flow-through) (Figure 1). While this interaction commonly corresponds with topography, it depends largely on where the lake or wetland lies within the groundwater-flow system (Winter et al. 1998).

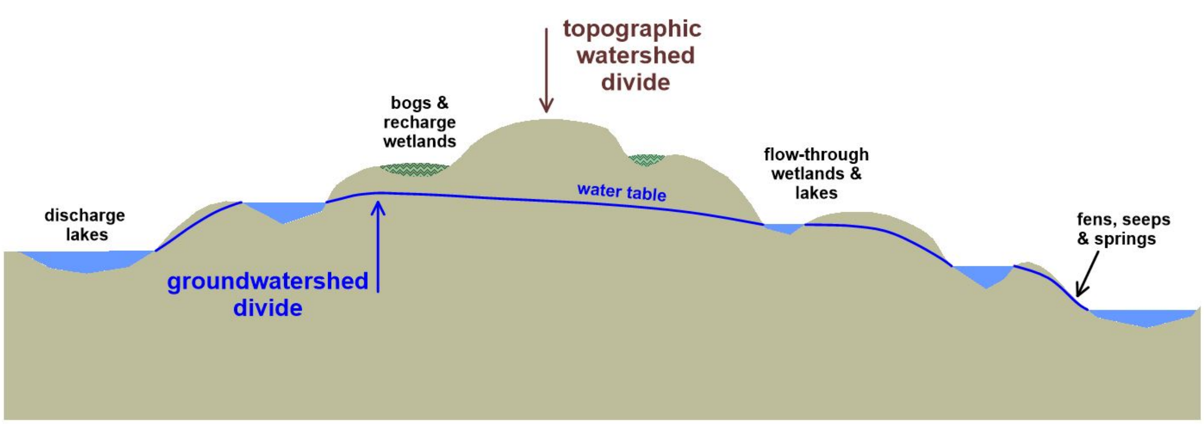


Figure 1. Examples of lake and wetland interactions with groundwater.

For areas with little subsurface information and abundant surface water features, a digital elevation model (DEM) can be useful to characterize the groundwater contribution to wetlands (Gerla 1999). DEMs allow the delineation of topographic (surface) watersheds and provide the elevations of surface water bodies such as lakes, streams, and wetlands, which can serve as a basis for interpolating the surrounding water-table surface. For example, Gerla (1999) used a DEM and a transient numerical groundwater model to map the occurrence of groundwater recharge and discharge in the Shingobee River watershed about 30 kilometers east of Itasca State Park. Interpolation tools available in GIS programs can also be used to generate a water-table surface from surface water feature elevation data derived from a DEM. A similar technique has been used by the Minnesota Department of Natural Resources (2016) to produce the statewide depth-to-groundwater map.

The DEM resolution, or cell size, has been shown to have an effect on hydrological interpretations derived from topographic features (Zhang and Montgomery 1994; Wu et al. 2008; Habtezion et al. 2016). Slope, upslope area, flow lengths, and the areas of watersheds are commonly used in hydrological study and have all found to be sensitive to the resolution of DEMs (Sørenson and Seibert 2007; Wu et al. 2008; Zhang and Montgomery 1994). For example, Wu et al. (2008) found that mean slope decreased with decreasing DEM resolution and upslope area increased with decreasing DEM resolution. Zhang and Montgomery (1994) found that coarser resolution DEMs led to a decrease in depth to the water-table and an increase in peak discharge. These examples emphasize the importance of considering the sensitivity of a model to the DEM resolution and selecting an appropriate resolution for modeling objectives.

Water quality has also been found to be useful in groundwater modeling. Winter and Carr (1980) created a groundwater model for the Cottonwood Lake area of the Missouri Coteau, North Dakota, and measured water quality to evaluate how lakes and wetlands on this moraine interact with groundwater. The Missouri Coteau is similar to the Itasca Moraine in that it is a glacial moraine and that it also serves as a regional recharge area. Lakes with large amounts of total dissolved solids (TDS) received discharge from regional or deep flow systems while lakes with smaller amounts of TDS received discharge from smaller, more localized, flow systems (Winter 1999). Within their study area, Winter and Carr (1980) identified recharge, discharge, and flow-through lakes and wetlands based on data from 1979.

More recently, Jaworksa-Szulc (2016) used a groundwater model and TDS measurements to characterize how lakes interact with groundwater in the Young Glacial Area of northern Poland. Lakes that received groundwater discharge had TDS measurements that ranged from 250-350 mg/L while losing lakes or those that recharge groundwater had measurements less than 100 mg/L. The flow-through lakes identified had TDS measurements that ranged from 170-200 mg/L. In addition, the discharge lakes were commonly found in ancient tunnel valleys or other meltwater derived features while the losing (recharge) lakes were found on ground moraine.

Gerla (2013) has suggested that measuring pH and electrical conductivity (or specific conductance) can be useful in determining what processes are dominant in a wetland's water budget. Unlike TDS, specific conductance can be measured directly in the field. Specific conductance is a measurement of how well a solution conducts electricity and serves as a proxy measure of dissolved minerals. Groundwater interacts with subsurface materials and dissolves minerals as it flows, thereby increasing its specific conductance. Generally, groundwater will have a much higher specific conductance compared to surface water bodies (Black 1996). Because of this, the specific conductance of a surface water body varies and depends on its water budget. If the water budget is dominated by meteoric water (ombrotrophic), then the waters of the wetland will be less mineralized and have a characteristically low specific conductance. If the water budget is dominated by groundwater (minerotrophic), the waters of the wetland will be more mineralized and have a characteristically high specific conductance. (Winter 1999; Seelig and DeKeyser 2006; Schwintzer and Tomberlin 1982; Heinselman 1970).

Meteoric water interacts with carbon dioxide, other gases, and particulates in the atmosphere, which results in a slightly acidic pH of approximately 5.6 (Fetter 2000; Drever 1997). Because of this general trend, wetlands with water budgets dominated by meteoric water, such as ombrotrophic bogs, are generally more acidic (Schwintzer and Tomberlin 1982; Heinselman 1970). Several factors affect surface water and groundwater pH. For example, aquatic plant productivity influences pH as photosynthetic reactions consume carbon dioxide, resulting in a buffered pH of surface water bodies (Glaser 1992; Tucker and D'Abramo 2008). In bogs, vegetation such as sphagnum moss has the opposite effect and maintains acidity (Clymo 1963; Clymo and Hayward 1982; Gorham et al. 1985). This process is enhanced by limited groundwater seepage which otherwise could introduce neutralizing bases (Glaser 1992). Water in wetlands that is contributed from groundwater is also prone to pH variation. The pH of groundwater is influenced by the composition of the soil, sediment, or rock with which the water has interacted. For example, pyrite can react with oxygen and lower the pH of groundwater (Beaty et al. 1996). It also is influenced by subsurface residence time, the contributing basin size, and biological activity (Logan 1995; Winter et al. 1998; Wolock et al. 1997).

The work presented here will add to the existing body of knowledge by providing a new spatial modeling technique for water resource managers. This technique can be of use in areas with abundant surface water features and is more cost-effective and less invasive than traditional techniques of water budget characterization. This research will further explore the interaction occurring between groundwater and lakes and wetlands in the Itasca Moraine and provide insight into whether pH and specific conductance are useful water quality parameters for verifying this spatial modeling technique.

Study Site

Itasca State Park is located in north-central Minnesota's Clearwater County. The park's namesake, Lake Itasca, is the largest and best-known lake in the park, and proclaimed the headwaters of the Mississippi River by Henry Schoolcraft in 1832 (Minnesota Department of Natural Resources 2018a). To the south of Lake Itasca lies the second largest lake in the park, Elk Lake, as well as many small lakes, wetlands, creeks, and groundwater springs. The Elk Lake and Nicolet Creek watershed, which contributes water to the upper end of Lake Itasca, is approximately 47 square kilometers and might be referred to as "the headwaters of the headwaters" (Figure 2).

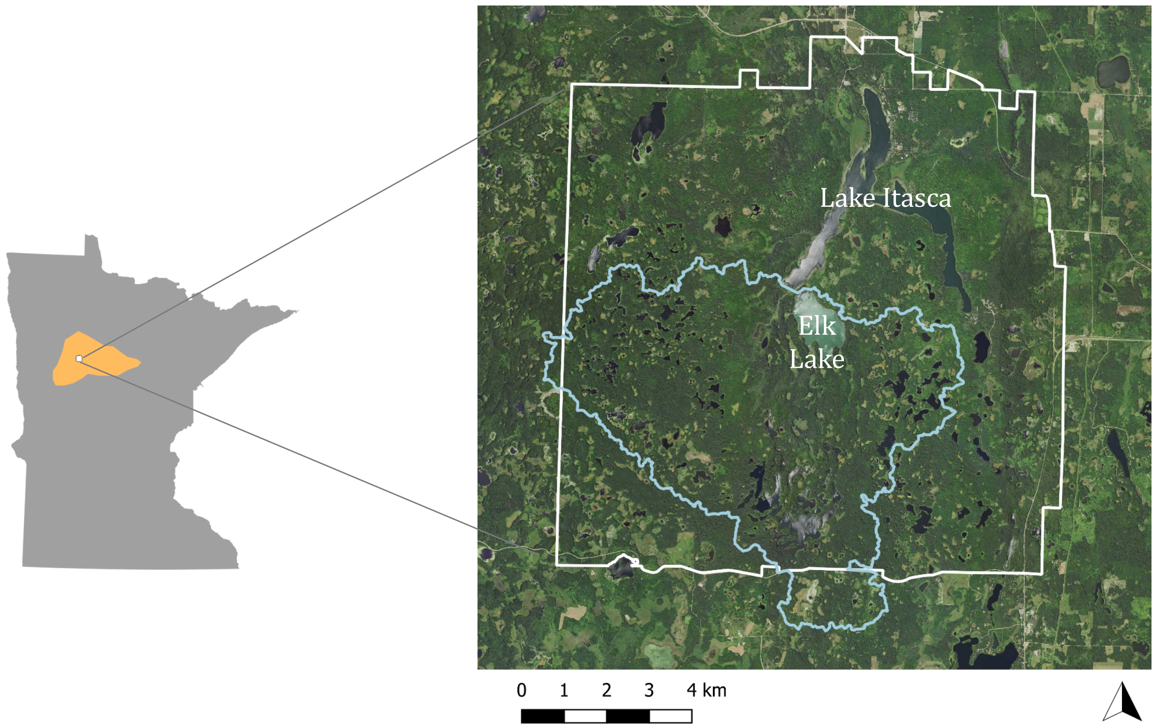


Figure 2. Study area. Itasca Moraine extent is shown in orange. The white outline shows the Itasca State Park boundary and the blue outline shows the Elk Lake and Nicolet Creek (topographic/surface) watershed.

Itasca State Park lies on the northern flank of the Itasca Moraine, which was deposited by the stagnation of the Wadena lobe and was subsequently altered by regional glacial activity up until the end of the Wisconsinan glaciation approximately 11,000 years ago (Wright 1993; Carney and Mooers 1998). The Itasca Moraine exhibits approximately 200 meters of topographic relief and extends for nearly 150 kilometers (Figure 2). It has been suggested that the Itasca Moraine serves as a regional recharge zone and contributes groundwater to the Red Lake Peatlands to the north (Siegel 1981). The Wadena lobe till that composes the moraine is rich in Paleozoic carbonate fragments from southern

Manitoba. It also contains igneous and metamorphic fragments which are believed to have been eroded from the Canadian Shield. It has been suggested that the crystalline fragment contamination of the Wadena lobe resulted when it was diverted by the Rainy lobe (Carney and Mooers 1998; Wright 1993). The water chemistry of Elk Lake is largely influenced by the calcareous glacial drift of the moraine, with Ca^{2+}, Mg^{2+}, and HCO_3^- as dominant ions. The lake precipitates $CaCO_3$ and its varved sediments have been well studied (Megard et al. 1993).

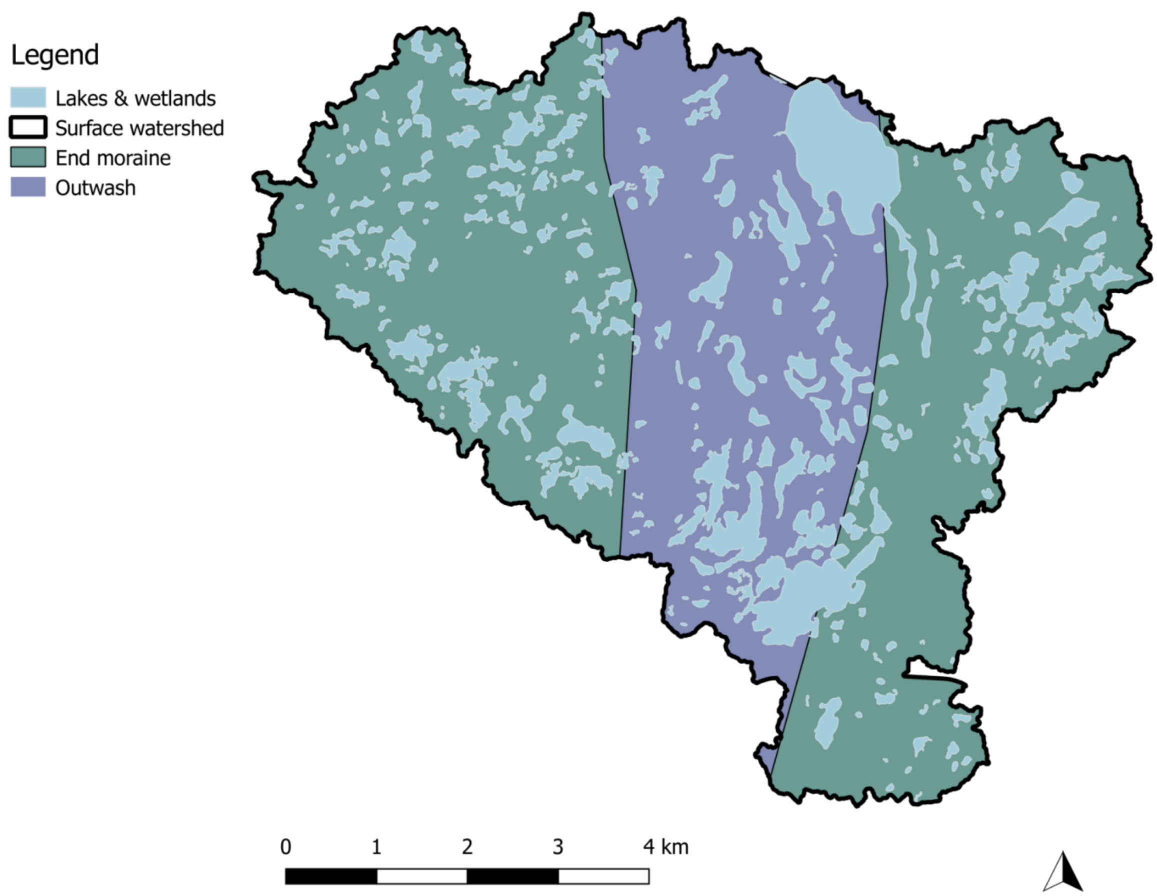


Figure 3. Surficial Quaternary geology map for the study area (Hobbs and Goebel 1982).

On the state surficial Quaternary geology map (Hobbs and Goebel 1982), the study area includes end moraine and outwash deposits (Figure 3). Overall, the study area has a hummocky topography and numerous small lakes and wetlands that were formed when ice blocks broke off from the ice lobe (Wright 1993). Larger lakes, such as Lake Itasca and Elk Lake, are believed to have formed in the coarse remnant tunnel valleys (outwash channels) that once drained the ablating ice lobe (Wright 1993). Landform assemblages for the study area were later described by Carney and Mooers (1998) and consist of outwash channels and a supraglacial complex. The outwash channels are primarily sand and gravel and well sorted while the supraglacial complex is primarily sandy loam with some sand and gravel that are poorly to moderately sorted.

CHAPTER 2: METHODS

Data

Public domain datasets were used for the groundwater model, including a 3-meter DEM derived from LiDAR (Light Ranging and Detection) and obtained from the Minnesota Geospatial Information Office's MnTOPO Viewer (2018) (Figure 4). Lake and wetland data were obtained from the Minnesota National Wetland Inventory Update (Minnesota Department of Natural Resources 2019a), which began in 2008 and was funded by the Environmental and Natural Resources Trust Fund. The wetland update used high-resolution aerial imagery as well as LiDAR (Light Ranging and Detection) to provide more accurate mapping. The Minnesota Department of Natural Resources states the update is, "... the most comprehensive, current, and accurate wetland inventory in the country," (2018b). The northwest region, which includes Itasca State Park, was released to the public in January 2019. Summer and fall 2017 statewide aerial imagery (1-meter resolution) provided by the Minnesota Geospatial Information Office's Web Mapping Service (2017) was used to identify and verify lakes and wetlands. A shapefile including streamlines was obtained from the National Hydrography Dataset (U.S. Geological Survey 2019). Soil polygons (1:20,000 and 1:24,000 scale) for the study area from the Soil Survey Geographic Database (SSURGO) were obtained from the U.S. Department of Agriculture and the Natural Resource Conservation Service's Web Soil Survey (Soil Survey Staff 2019). The SSURGO soil data table was used to estimate hydraulic conductivity and produce the recharge-discharge map (Soil Survey Staff 2019; Wieczorek 2014).

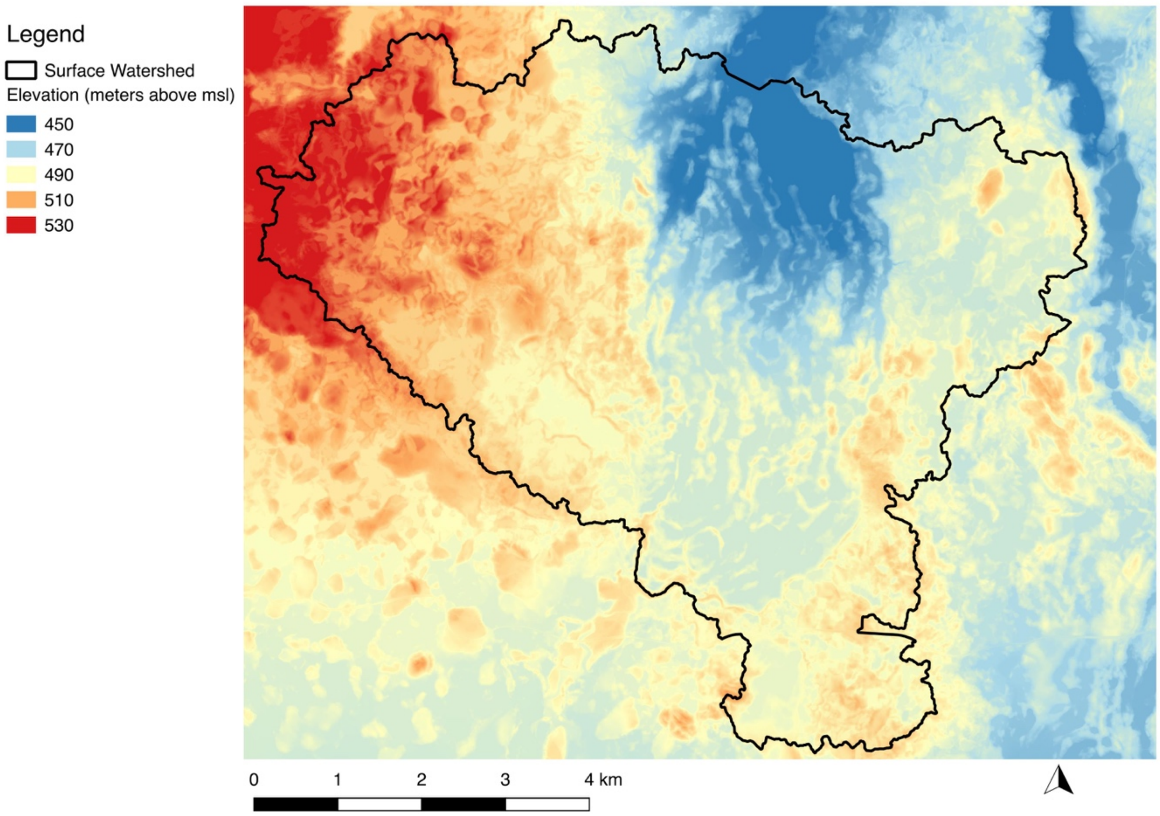


Figure 4: 3-Meter Digital Elevation Model of the study area (Minnesota Geospatial Information Office 2018).

Water-Table Surface

Lakes and wetlands were the primary sources of elevation data for the creation of the water-table surface (Appendix I). Wetlands with specific hydrologic modifier codes were selected to create the water-table surface as they are assumed to be more persistent and thus a more accurate representation of the water-table. Selected hydrological modifier codes for the study area include: F-semi-permanently flooded, G-intermittently exposed, and H-permanently flooded (Cowardin et al. 1979). All other wetlands were removed from the dataset. The edited wetland layer was underlain by aerial imagery (Minnesota Geospatial Information Office 2017) and wetland polygons were observed to verify their extents. An additional dataset (raster) that included the minimum elevation value for each polygon was created. This ensured that the lowest elevation present in each wetland polygon was assigned to the entire polygon in case there was elevation variability within the DEM.

Streams were also used to create the water-table surface. A stream-line dataset was obtained from the National Hydrography Dataset (U.S. Geological Survey 2019) and was used to add elevation control and gridding break-lines. The stream-lines were inspected and edited to ensure that they did not intersect any of the wetlands. Many of the stream-lines were redrawn to include more vertices or redrawn to better correspond with stream channels observed using the aerial imagery.

Iterative finite difference interpolation was used to create the water-table raster from the lake, wetland, and stream data using GIS tools that are based on the ANUDEM algorithm (Hutchinson 2011). This algorithm is discussed in more detail in the next section. For comparison, the water-table surface raster was subtracted from the DEM and the resulting raster was visually inspected for areas of the water-table that lay above the DEM. Areas where this occurred were noted and alternative datasets were collected to provide more elevation control/constraint on the water-table at these sites.

One example of the alternative datasets includes the SSURGO hydric soils (Soil Survey Staff 2019). A hydric soil is a soil that has developed anaerobic, or oxygen deficient, conditions as a result of being saturated during the growing season (Natural Resources Conservation Service 2019a). Soil polygons were selected if they met the following criteria:

1. The polygon was listed as a hydric soil and was described as frequently ponded. According to the National Technical Committee for Hydric Soils, this is a closed depression that contains water and has more than a 50 percent chance of occurring in any given year under normal weather conditions (Natural Resources Conservation Service 2019a).
2. Standing water, drainage features, or wetland-type vegetation were observed on the aerial imagery within the polygon.
3. The soils represented by the polygon covered small parts of the raster where additional water-table elevation control was necessary. This included areas where the water-table surface lay above the DEM.

Features were also created with the DEM and the aerial imagery. Polygons were constructed in areas where standing water was observed in the aerial imagery and streamlines where dendritic drainage features could be discerned. In areas where there was dense vegetation, a color ramp was assigned to the DEM to enhance areas with drainage features or low-lying depressions for easier identification. These features were then either represented by a line or a polygon feature and added to the existing dataset.

Interpolating the water-table surface was an iterative process. In some areas, especially those with steep topography, initial interpolation created areas where the modeled water-table rose to unreasonably high levels above the ground surface (Figure 5). These were corrected by adding additional control points represented by lines and polygons where the water-table elevation could be estimated from occurrences of hydric soils, streams, and wetlands that were not included in the initial lake and wetland datasets. Only 2-5 additional features were added for elevation control at a time before re-running the interpolation tool. The resulting water-table surface was then observed and compared to the DEM to determine what improvements had been made to the resulting surface.

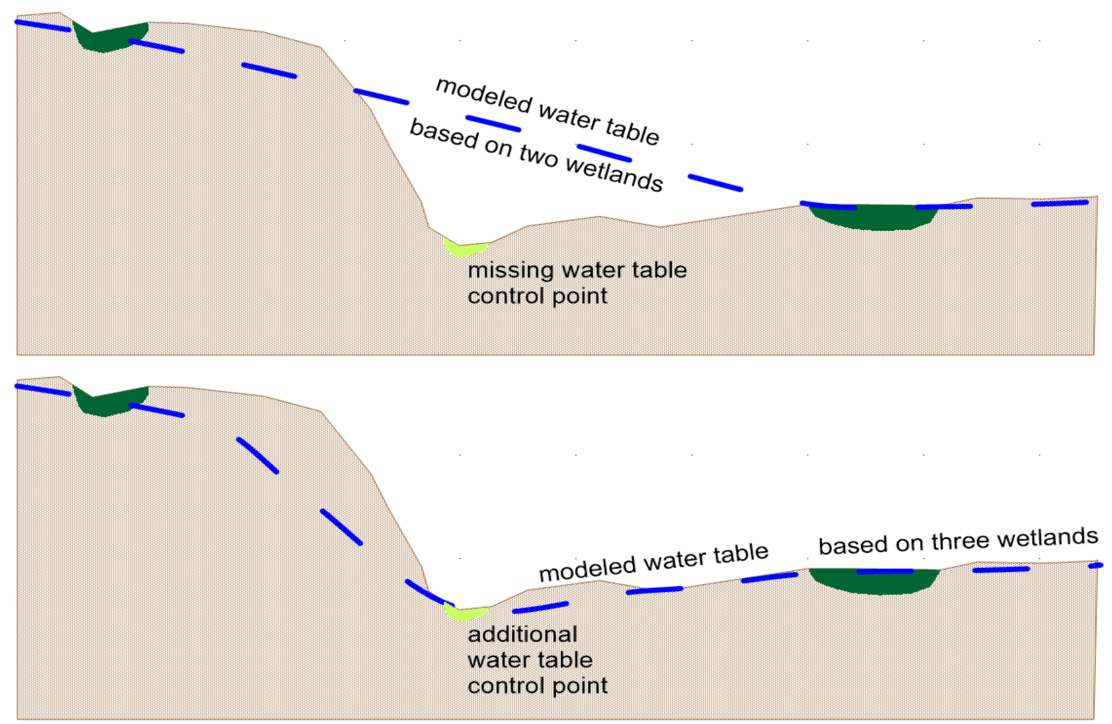


Figure 5. Profile diagram showing the effect of adding additional water-table control in an area of complex topography.

In total, 883 polygons were used to interpolate the water-table surface, including 761 wetland inventory polygons, 102 hydric soil polygons, and 20 polygons created from the aerial imagery and the DEM (Figure 6). It took approximately 30 trials using the interpolation tool to create the final water-table surface.

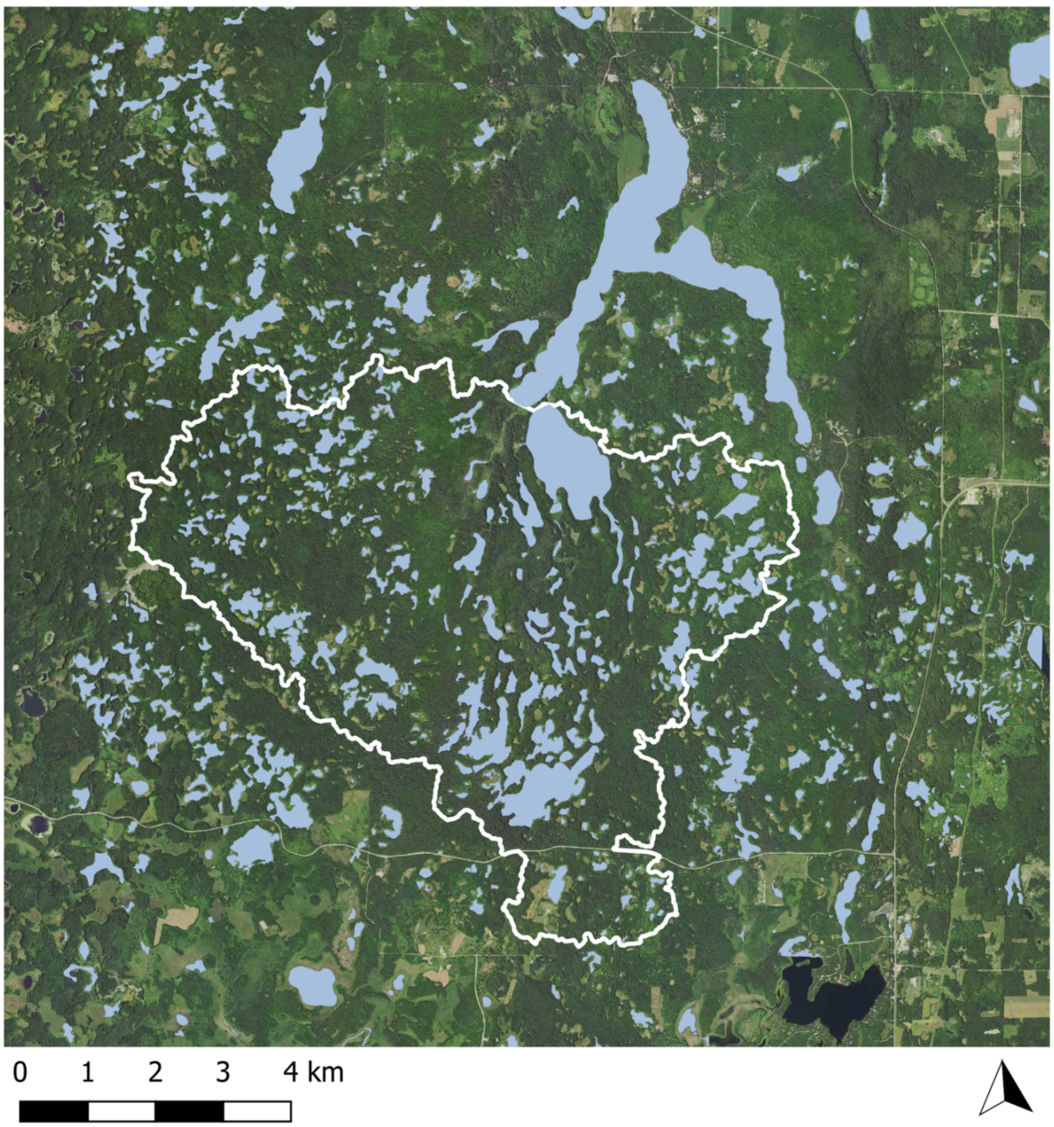


Figure 6: Polygons used to interpolate the water-table surface (in blue) (Elk Lake and Nicolet Creek topographic/surface watershed in white).

The ANUDEM Algorithm

This algorithm was originally developed by the Australia National University (Hutchinson 2011) for the construction of hydrologically correct DEMs. This algorithm is more efficient to use than other popular global interpolation methods, including thin plate spline and kriging when using large datasets. Even though its efficiency is similar to that of local methods, it still maintains the high-quality surface continuity that the other popular global methods provide (Hutchinson 1988; Hutchinson 2011).

When this technique is applied to create a water-table surface, the algorithm uses the known elevations of the input surface water features to estimate the elevation of the water-table between these features. The algorithm does this by assigning the elevation data supplied by the surface water features to the nearest grid point. Gauss-Seidel iteration with overrelaxation (Young 1971; Golub and Van Loan 1983) is used to calculate nearby grid points that do not have elevation data assigned to them. The process starts with a coarse grid and the grid spacing is then decreased until the final grid resolution (specified by the user) is obtained. For each grid spacing, 25 iterations (unless otherwise specified) occur before the elevation values are linearly interpolated to the following finer resolution grid (Hutchinson 1998; Hutchinson 2011). A drainage enforcement algorithm operates simultaneously and works to remove sinks in the surface through imposing drainage conditions. The ANUDEM algorithm was used by the Minnesota Department of Natural Resources when creating their depth to groundwater map (Minnesota Department of Natural Resources 2016).

To create the water-table surface, the lake and wetland elevation data were input as both contours and lakes. In both cases, the features were assigned the lowest elevation value from the DEM. The input contours are used at first to determine the general morphology of the water-table surface using points of maximum curvature. The contours are then used to estimate the elevations of cells in between input features and the generalized morphology is updated iteratively throughout this process (Hutchinson 1988; Hutchinson 2011). The input lakes, which consist of the lake and wetland polygons, are also treated as a contour at first and the elevation values at the boundary of the lake are used to estimate the elevation of the lake surface. Inputting the lakes and wetland polygons as lakes also ensured that interior cells lie below the lake boundary while outside cells lie above it (Hutchinson 1988; Hutchinson 2011). The input of stream data assists the interpolation process by serving as a gridding break line and ensures the removal of data that conflicts with the stream-lines' descent (Hutchinson 1988; Hutchinson 2011). The elevations of the stream line vertices are also assigned to the nearest grid point to assist with interpolation (Hutchinson 2011).

Groundwater Flow Path Lines

The water-table surface was used to generate groundwater flow path lines using the same techniques that are well established for topographic DEM channels (Tarboton et al. 1991) (Appendix II). In order to produce a connected drainage network, it is necessary to fill sinks present in the water-table surface. Sinks often result from the rounding of elevation values when generating a surface but can also represent actual depressions in the landscape. In any case, these features result in a discontinuous drainage network if they are not filled (Tarboton et al. 1991; Jenson and Domingue 1988).

An eight-direction (D8) flow model was then used to determine the direction of flow out of each cell which is determined by the steepest descent (Jenson and Domingue 1988). This information is then used to determine the flow accumulation, or the number of cells that flow into a particular cell. Cells with high accumulation are areas were flow is concentrated and can be used as flow paths (Tarboton et al. 1991; Jenson and Domingue 1988). Cells that had values greater than 250 were used as flow lines as they are areas where the flow is more concentrated. A threshold of 250 was selected by trial-and-error as it showed a greater level of detail than higher thresholds. This higher level of detail ensured that all sampled wetlands had a flow path in their vicinity. The cumulative flow path length for each sampled water body was then calculated by adding up the lengths of associated flow path lines. If multiple flow paths converged at a lake or wetland, the length for the longest reach was summed. In the case of Elk Lake, which had two large tributaries, a weighted average was calculated using the contributing (groundwater) drainage area for each tributary.

The flow path lengths were then plotted against the average specific conductance and pH measurements for each water body for statistical comparison. Linear regression was used to analyze the strength of the relationship between groundwater flow path lengths and pH and specific conductance measured in the field. The strength of the relationship was evaluated with R^2 and the p-value. R^2 provided the percent of the variance of pH and specific conductance that is explained by the effect of the groundwater flow path length. The p-value provided the likelihood of obtaining water quality measurements when the null hypothesis is true (Helsel and Hirsch 2002). In this case, the null hypothesis was that there is no relationship between the groundwater flow path length and either pH or specific conductance for lakes and wetlands. If the p-value was computed to be less than the selected significance value of 0.05, the results were assumed to be statistically significant and the null-hypothesis was rejected. Lakes or wetlands that were not located on or touching flow paths were not compared because they did not have quantifiable lengths associated with them.

Watersheds

Groundwatersheds were delineated in a manner that was very similar to surface watersheds. The only difference between delineating the surface watershed and the groundwatershed is the source of the elevation data. The surface watershed elevation data was supplied by a 3-meter DEM while the groundwatershed elevation data was provided by the interpolated water-table surface. The same steps that were used to determine the groundwater flow paths including filling the surface, determining flow direction, and calculating flow accumulation, are necessary when delineating watersheds (Appendix III). Once these were completed, the outlet for the watershed was identified. The outlet or outlets selected should be based on the watershed of interest. One common example of an outlet is a cell within a stream flowing out of a lake. In the case of a wetland with no outlet streams, the wetland feature itself can be selected. Either case resulted in a raster that outlined the upslope area or watershed to the designated outlet. The surface watershed to groundwatershed area ratio was then computed and plotted against the average specific conductance and pH measurements for each water body for statistical comparison. Linear regression was used to analyze the strength of the relationship between the ratio computed from the surface watershed and groundwatershed area and pH and specific conductance measured in the field. The strength of the relationship was evaluated with R^2 and the p-value with a significance value of 0.05.

Recharge-Discharge Map

Darcy's Law describes flow through a porous medium and can be combined with the conservation of mass to determine which cells have groundwater inflow that exceeds groundwater outflow (discharge to surface water) and groundwater outflow that exceeds groundwater inflow (recharge of surface water). Darcy's Law calculates the discharge from hydraulic conductivity, area, and the hydraulic gradient (Fetter 2000).

$$Q = -KA\frac{h_2 - h_1}{L}$$

Q= discharge (length^3/time), K= hydraulic conductivity (length/time), A =area (length^2), h= hydraulic head (length), and L= the flow length (length) (Fetter 2000)

The interpolated water-table surface can be used to create a recharge-discharge map with GIS tools (Appendix IV). This procedure gives a groundwater volume balance residual raster in which the residuals are the differences between inflow and outflow of the cells (Figure 7). The flow balance of the cell must be accommodated by either surface water inflow or outflow to comply with the conservation of mass (ArcGIS for Desktop 2019). For example, a negative residual indicates that recharge from surface water exceeds discharge out of the cell while a positive residual indicates that discharge from the cell exceeds recharge from surface water. In two dimensions, the residual for each cell in the raster is computed with the following formula:

$$R_{vol} = Q_1 - Q_2 + Q_3 - Q_4$$

R_{vol}= the residual (length^3/time), Q=the discharge through each respective cell wall (length^3/time) (ArcGIS for Desktop 2019)

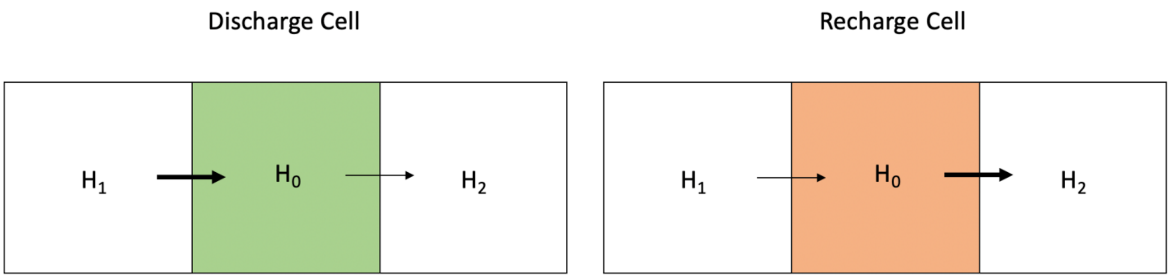


Figure 7: Example of groundwater recharge and discharge cells in the volume balance residual raster. The bold arrow represents a higher magnitude of flow.

Groundwater flow in the Itasca Moraine was assumed to be unconfined. This is because of the thick, relatively coarse surficial glacial deposits (Wright 1993) and the variety of lakes and wetlands which are assumed to be expressions of the water-table, or the unconfined aquifer. Therefore, hydraulic heads were squared to account for unconfined conditions. This was necessary because, under unconfined conditions, saturated thickness and hydraulic head depend on one another. Squaring the hydraulic heads linearizes this dependency by invoking the Dupuit approximation (Dupuit 1857; Fetter 2000). This allows the calculation of specific discharge from hydraulic conductivity and the hydraulic gradient using squared hydraulic head values.

$$q'L = -\frac{K}{2}\left(\frac{h_2^2 - h_1^2}{L}\right)$$

q'L= flow per unit width (length^2/time), K = hydraulic conductivity (length/time), h^2 = squared hydraulic head values (length), and L = flow length (length)

The hydraulic conductivity values were estimated using SSURGO soil polygons (Soil Survey Staff 2019) (Figure 8). These polygons are each assigned a map unit that corresponds to the dominant soil type present in the mapped polygon. This map unit can be found in the National Soil Information System Database which includes a plethora of information about the soil which has been determined from both field and laboratory

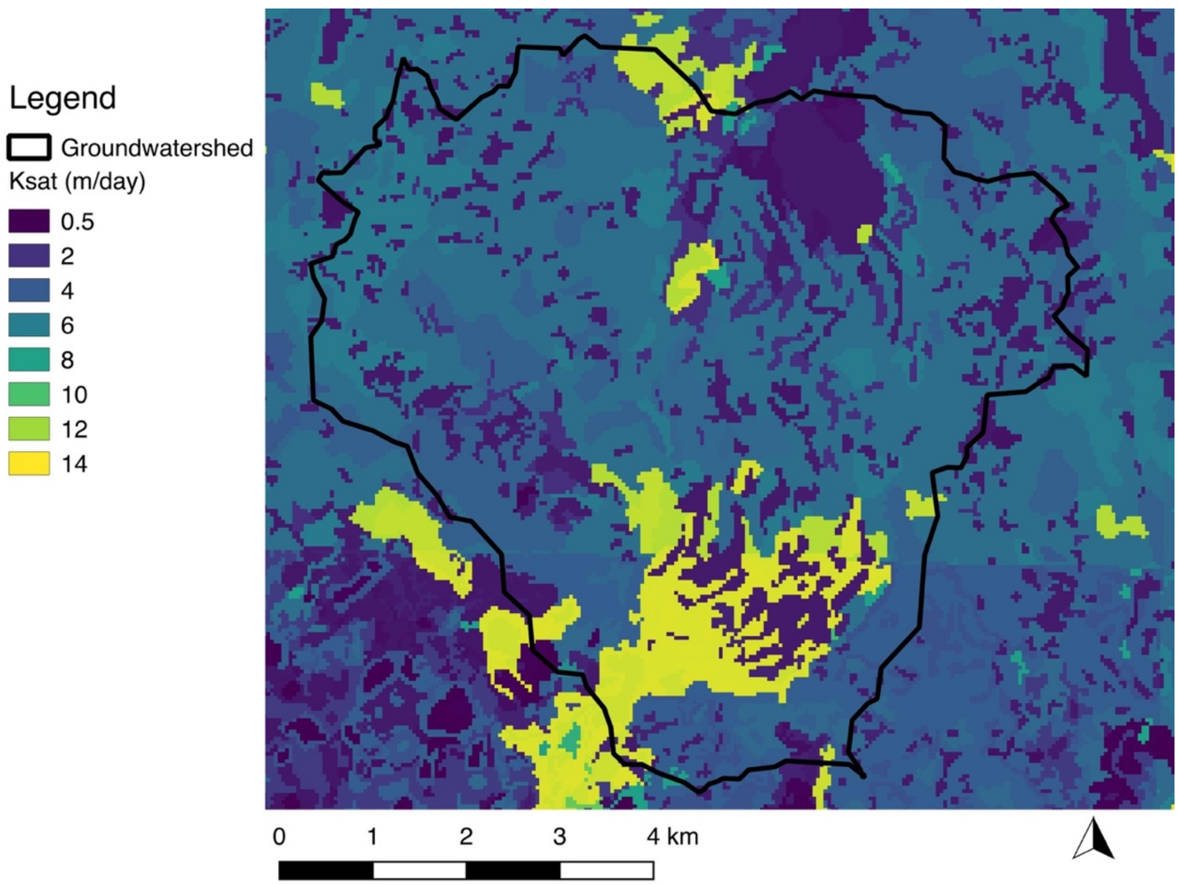


Figure 8. K_{sat} values for the study area from the SSURGO soil polygons (Soil Survey Staff 2019; Wieczorek 2014).

investigation of the soil's characteristics (Natural Resources Conservation Service 2019b). Saturated hydraulic conductivity (K_{sat}) refers to the ease at which water is transmitted through the pores of a saturated soil (Natural Resources Conservation Service 2019a). Wieczorek (2014) with the U.S. Geological Survey used soil information from SSURGO (Soil Survey Staff 2019), including K_{sat}, and produced an area and depth weighted table of soil characteristics that can be appended to soil polygons for any area of interest in the United States. This was done specifically for studies that require spatial estimates of soil hydraulic properties (Wieczorek 2014). No K_{sat} values were available for lake beds within the study area, so they were set to be similar to that of wetlands in the study area at 15 μm/sec. Units for the K_{sat} values were converted from μm/sec to m/day for convenience before they were used to produce the recharge-discharge map.

The interpolated water-table surface and an estimated aquifer base elevation were needed to create the hydraulic head raster. In three transects of the Itasca Moraine within the park, Tschann (2019) applied an analytical model developed by Bresciani et al. (2016) which ultimately is a modification of Hooghoudt's drainage equation (Bresciani et al. 2016; Hooghoudt 1940). Tschann (2019) used saturated hydraulic conductivity estimates provided by the SSURGO soils data (Soil Survey Staff 2019), recharge estimates from Smith and Westenbroek (2015), and water level elevations based on the 3-meter DEM. With these data, the analytical model produced a maximum depth to surface water and groundwater interaction of 441 meters (above mean sea level) for the transect that crossed the study area. This elevation was adopted as the aquifer base elevation for the study area.

The hydraulic head raster was created by subtracting the aquifer base elevation, 441 meters, from the interpolated water-table surface. A few lakes on the north end of the DEM, outside of the study area, had elevations below the aquifer base elevation resulting in negative values. All negative values were set to zero and then the heads were squared to account for unconfined conditions.

Recharge-Discharge Map Sensitivity

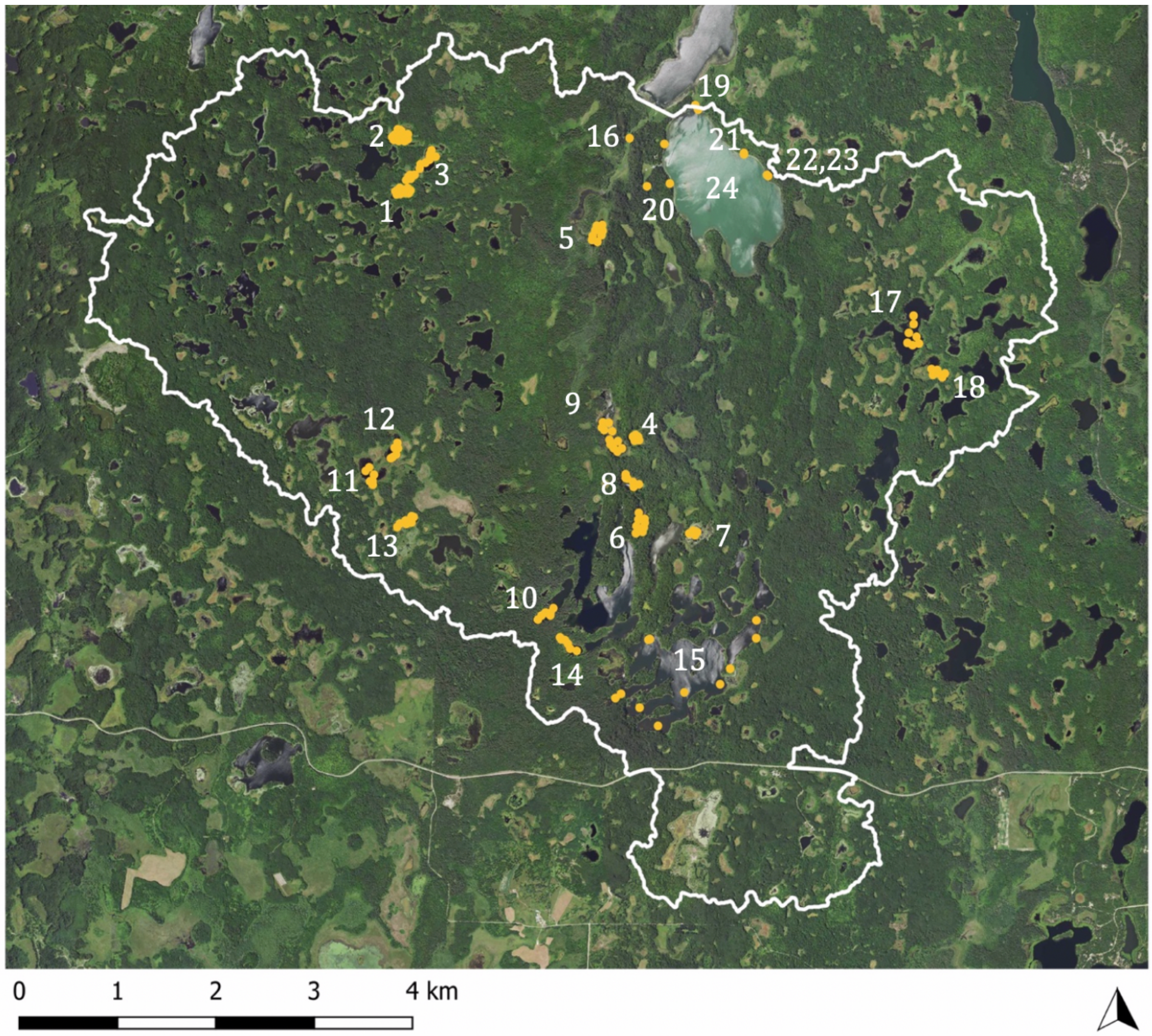
Because hydrological interpretations have been shown to be sensitive to the resolution of elevation rasters (Sørenson and Seibert 2007; Wu et al. 2008; Zhang and Montgomery 1994), the sensitivity of the recharge-discharge map to the water-table surface cell size was determined. This was accomplished by summing the water balance for all cells within the groundwatershed (Appendix V) and then comparing it to streamflow measurements (personal communication, P. Gerla, 2019) at Nicolet Creek and the Elk Lake outlet, Chambers Creek. This approach ensures that the water balance for the groundwater volume balance residual raster is similar to the discharge leaving the two outlets. In an effort to better match the discharge measurements of the two outlets, the cell size for the water-table surface raster was varied. Sizes selected for the sensitivity analysis included: 10, 20, 30, 40, and 50 meters.

Field Methods

The wetland inventory discussed previously was used to select wetlands for sampling. A Hanna pH/SC/TDS/Temperature meter (product HI9813-6) was used to measure pH and specific conductance at approximately 20 groundwatershed wetlands in September 2018. The meter was calibrated prior to the start of fieldwork using a 7.01 pH buffer and a 1413 μS/cm conductivity standard. Ultimately, lakes and wetlands sampled were selected by their Cowardin et al. (1979) hydrologic modifier code (F-semi-permanently flooded, G-intermittently exposed, and H-permanently flooded), location within the watershed, and accessibility. Only lakes and wetlands with codes selected for the water-table raster were sampled and effort was made to select an adequate distribution of lakes and wetlands across the study area. A kayak was used so that measurements could be taken in a variety of locations in each waterbody. Measurements were taken at 5-10 locations in each waterbody depending on its size. This included at least one measurement in the center and in each quadrant (Figure 9).

The following steps were taken when conducting field measurements:

1. The instrument probe was rinsed with deionized water.
2. The instrument probe was placed approximately 10 cm below the surface of the water for each measurement of pH and specific conductance.
3. Algae and floating vegetation were avoided when taking measurements.
4. The instrument probe was swirled in the water and readings were recorded after approximately 60 seconds once the values had stabilized.
5. The coordinates were recorded for each location with a GPS unit (Garmin eTrex 30x).
6. At least one measurement was taken near the center of the waterbody and additional measurements were taken in each quadrant.
7. Care was taken to ensure that bottom sediments were not disturbed when taking measurements.
8. The probe was rinsed with deionized water prior to sampling the next waterbody.



1. Unnamed NW Pond 1
2. Unnamed NW Pond 2
3. Unnamed NW Pond 3
4. Bog near Whipple Lake
5. Nicolet Lake
6. Triplet Lakes
7. Unnamed Pond near Little Elk Lake
8. Unnamed Pond near Whipple Lake
9. Whipple Lake
10. Augusta Lake
11. Unnamed SW Pond 1
12. Unnamed SW Pond 2
13. Unnamed SW Pond 3
14. Horn Lake
15. Hernando DeSoto Lake
16. Nicolet Creek
17. Clark Lake
18. Unnamed Pond near Clark Lake
19. Chambers Creek
20. Elk Lake Spring 1
21. Elk Lake Spring 2
22. Elk Lake Spring 3
23. Elk Lake Spring 4
24. Elk Lake

Figure 9. Field measurement locations.

Measurements of Elk Lake were taken from both shore and from a dock rather than with the kayak due to weather and wave conditions. Creeks and groundwater springs in the study area were also measured for comparison. The sampling of creeks was similar to that of lakes except the measurements were taken while standing on the banks of the creek rather than with the kayak. For groundwater springs, the probe was submerged as much as possible in the flowing spring water without being submerged in the underlying sediments. These procedures were adapted from the Minnesota Pollution Control Agency's Wetland Monitoring Protocol Handbook (Minnesota Pollution Control Agency 2015) and the HANNA instrument operation manual (2018).

Assumptions

For the interpolation of the water-table surface, the lakes, wetlands, streams, and hydric soils selected were assumed to be expressions of the water-table and have elevations that accurately represent the water-table surface. The wetlands with specific codes, F-semi-permanently flooded, G-intermittently exposed, and H-permanently flooded (Cowardin et al. 1979), were selected as they are assumed to be the most persistent. The elevations of hydric soil polygons that were described as frequently ponded and could be verified by aerial imagery were assumed to be an accurate representation of the water-table surface. For the recharge-discharge map, it was assumed that the aquifer is unconfined, laterally isotropic, and has a base elevation of 441 meters. The K_{sat} values obtained from the SSURGO soils (Soil Survey Staff 2019; Wieczorek 2014) were also assumed to be a reasonable estimate for hydraulic conductivity and the lakes in the study area were assumed to have a K_{sat} value similar to that of wetlands.

CHAPTER 3: RESULTS

Water-Table Surface

In the study area, the water-table is the lowest at approximately 448 meters near Elk Lake. The water-table reaches its maximum elevation of approximately 520 meters at the far west end of the study area. The groundwatershed for Nicolet Creek at its outlet on Lake Itasca and Elk Lake was delineated from the interpolated water-table surface (Figure 10).

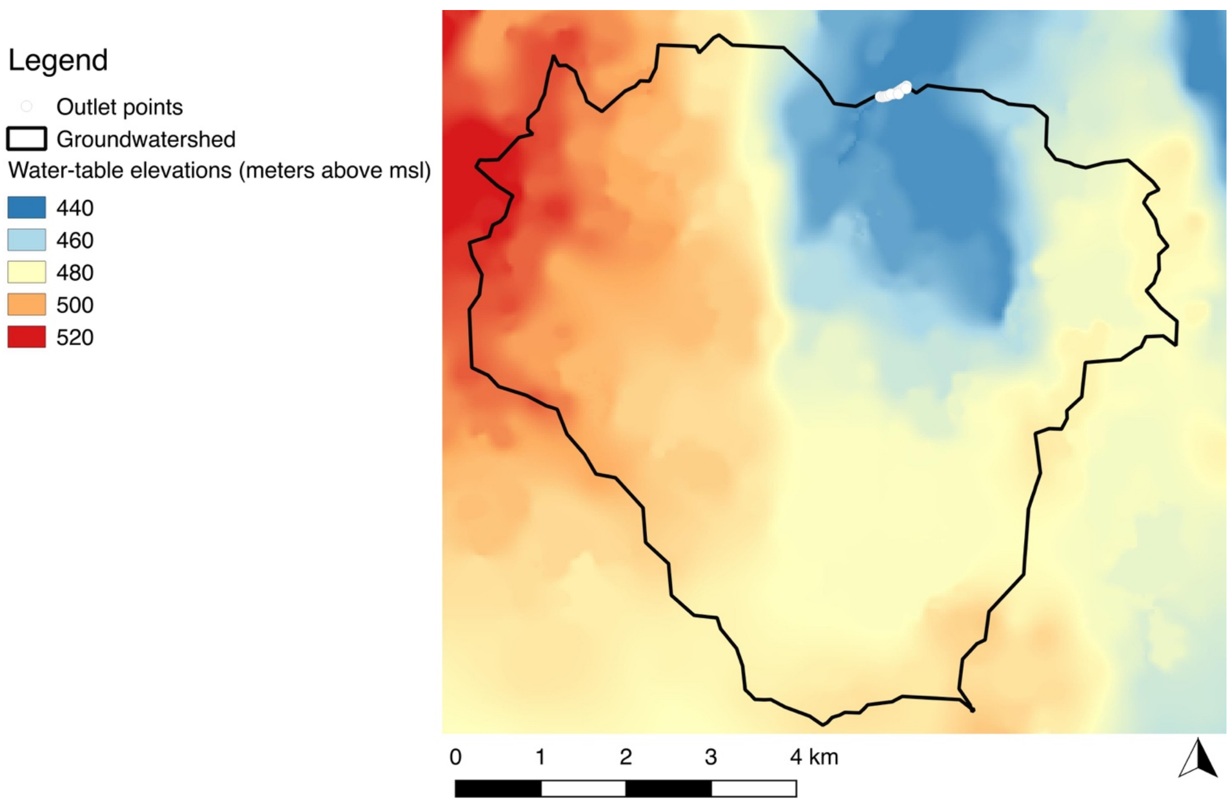


Figure 10. The interpolated water-table surface.

The best-fit interpolated water-table surface was subtracted from the DEM to map the depth to the water-table and identify areas where the water-table surface lies above the surface topography (Figure 11). There were a few scattered areas where the interpolated water-table surface lies above the surface topography, although it occurred in less than 2% of the watershed raster cells. Of these cells, 46% fell within 0.5 meters above the DEM, 51.5% fell between 0.5 and 3 meters above the DEM, and 2.5% fell more than 3 meters above the DEM. No cells exceeded 5 meters above the DEM.

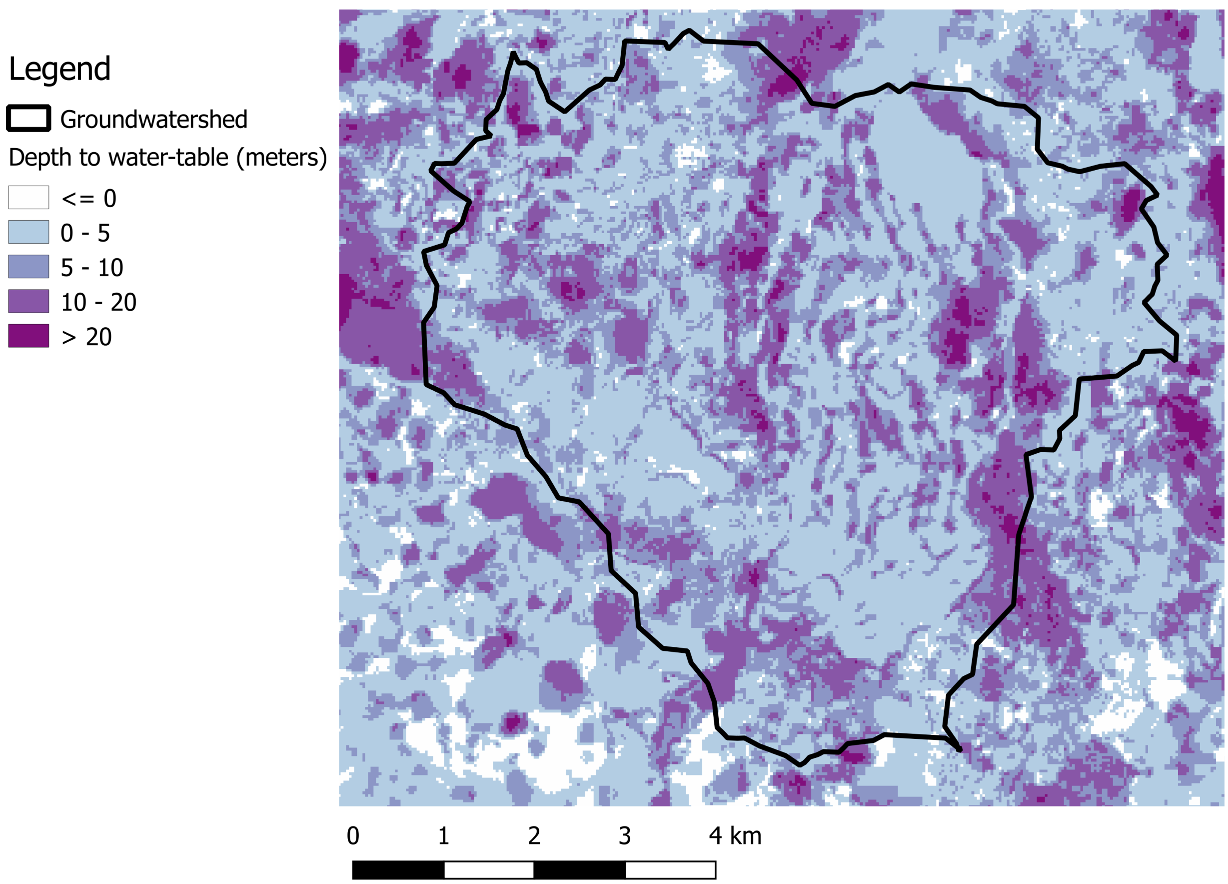


Figure 11. The depth to the interpolated water-table surface for the study area. Shown in white are areas where the interpolated water-table surface lies above the DEM.

Calculation of the water-table slope, or hydraulic gradient, shows several areas with greater than a 3-degree slope on the north end of the study area (Figure 12). These elongate, north-to-south oriented areas range to almost 3 km long, with the largest feature lying to the west of Nicolet Creek. Several smaller features lie to the east and south-east of Elk Lake with maximum slopes that also exceed 3 degrees. These features appear to

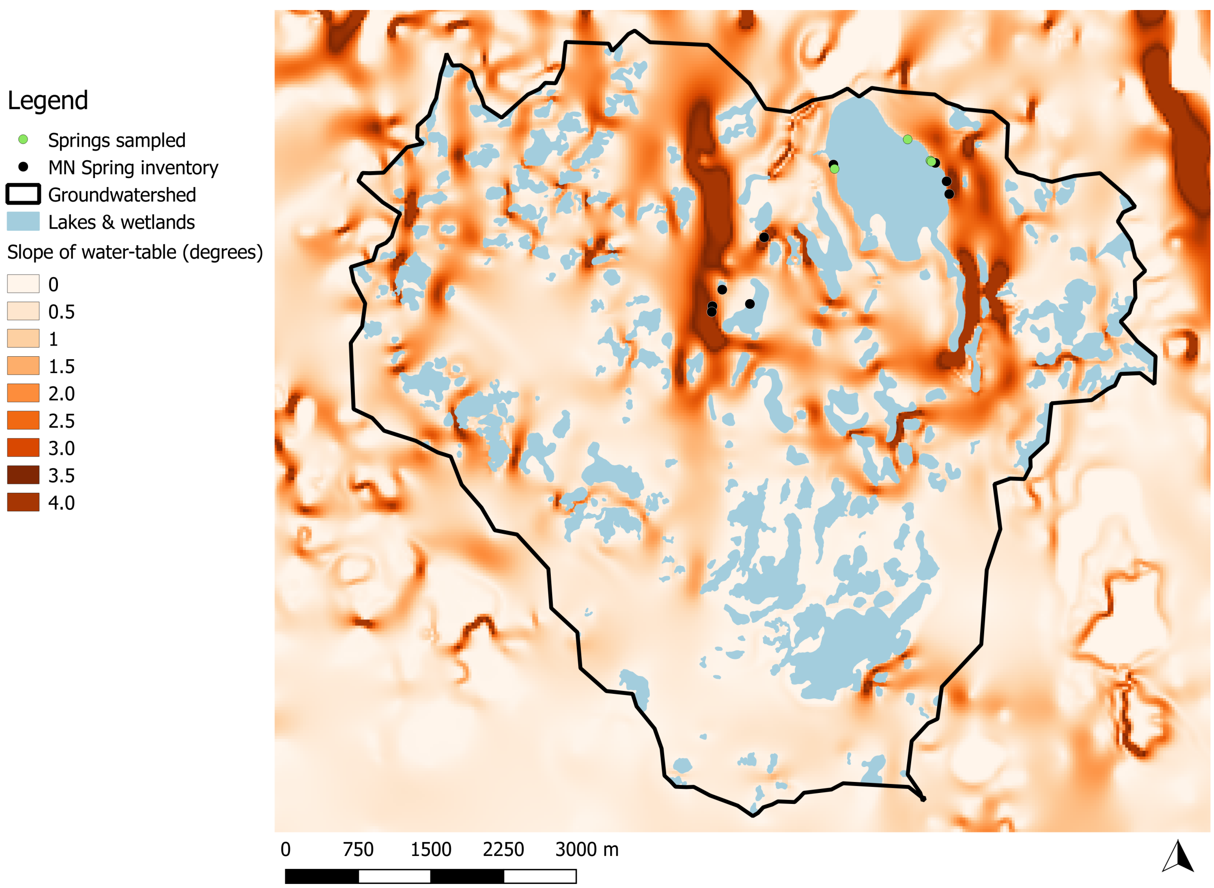


Figure 12. Water-table slope for the study area. Springs sampled for this report are shown in green and those reported in the Minnesota Spring Inventory (Minnesota Department of Natural Resources 2019b) are shown in black.

correspond to mapped contacts between end moraine and glacial outwash sediments that extend north-to-south through the study area (Hobbs and Goebel 1982). Locations of groundwater springs from the Minnesota Spring Inventory (Minnesota Department of Natural Resources 2019b) and springs that were sampled in the study area generally coincide with areas at the base of large water-table slopes (Figure 12). Based on the results, additional springs and seeps are likely to occur in the area southeast of Elk Lake and along the east edge of the feature west of Nicolet Creek.

Groundwater Flow Path Lines

Calculated path lines of groundwater flow show northward flow with convergence towards Elk Lake and Lake Itasca (Figure 13). Lengths of groundwater flow path lines were compared with pH and specific conductance field measurements. pH varied throughout the study area and ranged from 6.7 to 8.9. Linear regression was used to compare average pH with flow path lengths, with the results showing no statistical significance (R^2=0.18) (Figure 14; Table 1).

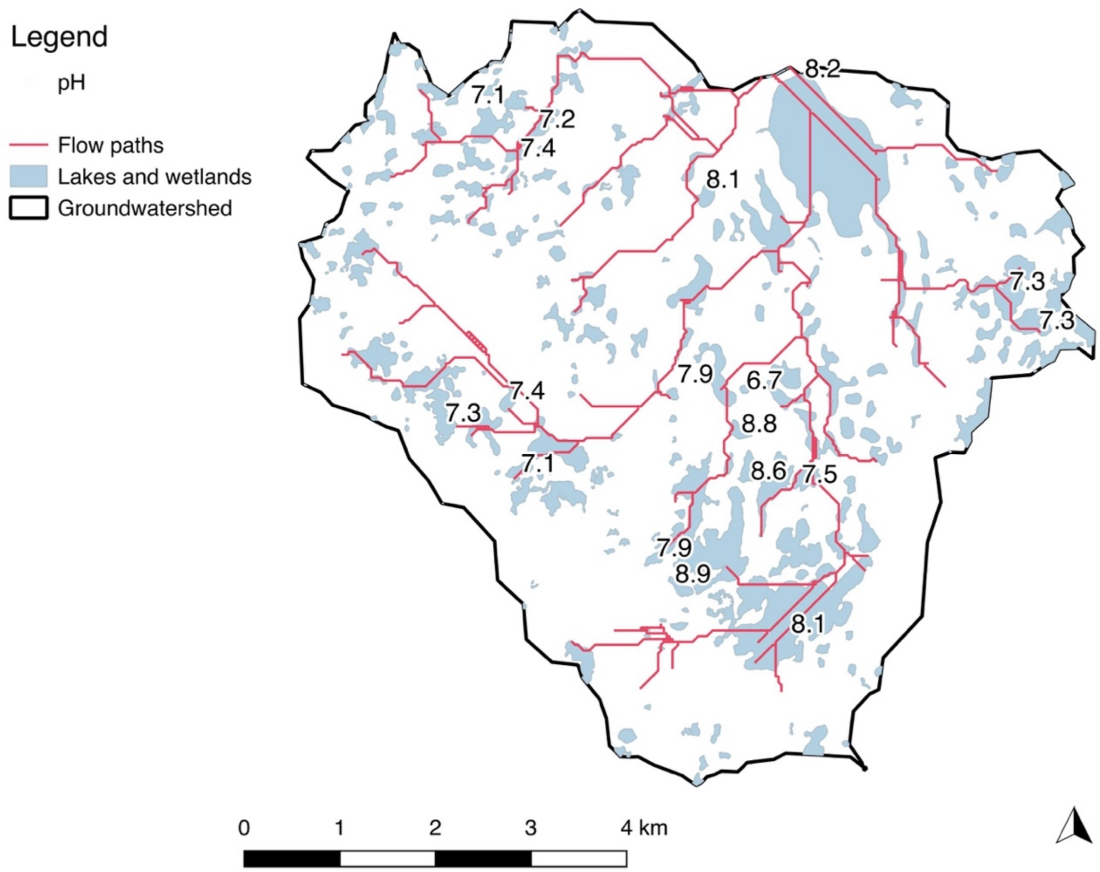


Figure 13. Groundwater flow path lines (in red) for the study area with pH measurements for sampled lakes and wetlands.

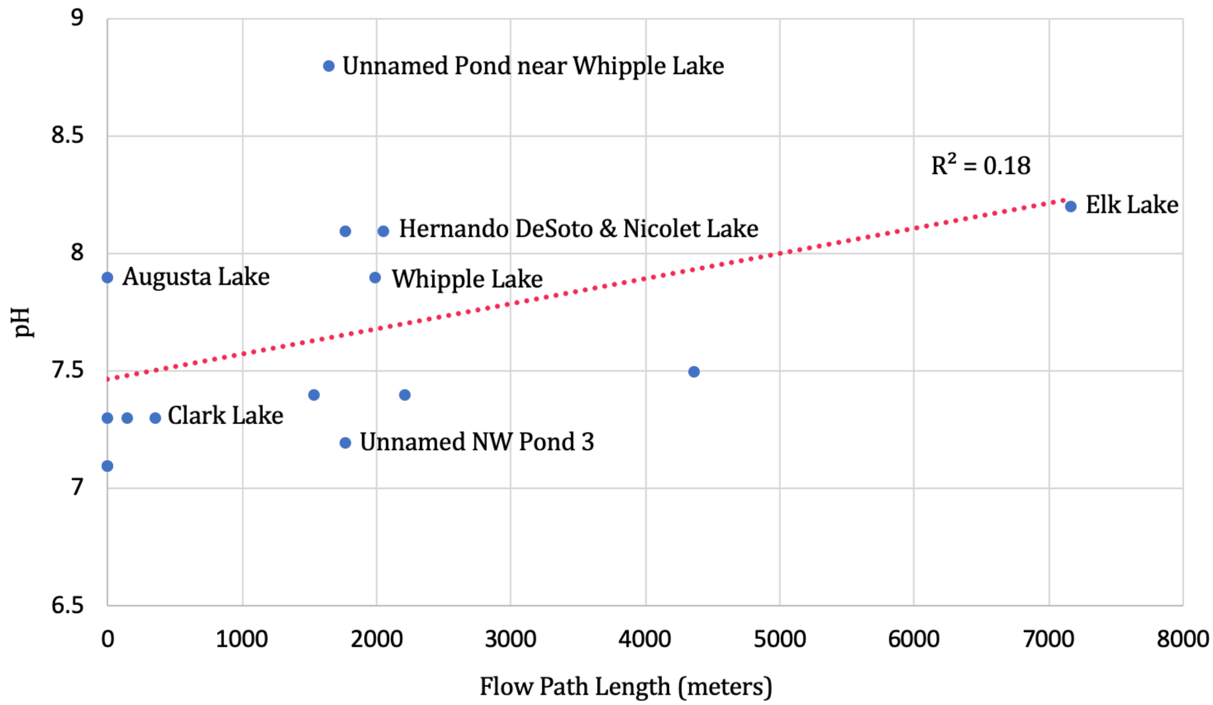


Figure 14. Graph of flow path lengths and pH measurements for sampled lakes and wetlands.

Parameter	pH
R Square	0.18
p-value	0.118

Table 1. Regression statistics computed from flow path lengths and pH measurements.

Lengths of groundwater flow path lines were also compared with average specific conductance measurements. Specific conductance appears to be larger in lakes and wetlands that have longer flow path lengths associated with them, although there are a few lakes (for example, Augusta Lake and Horn Lake) that lie near the beginning of flow path lines and have a relatively large specific conductance. In addition, unnamed ponds in the northwest portion of the study area lie at the middle of flow path lines but have the lowest specific conductance measurements sampled in the study area (Figure 15), although ponds farther upgradient were not sampled and may have even lower specific conductance. Linear regression was also used to compare specific conductance with groundwater flow path lengths. In this case, results were found to be statistically significant (R^2=0.39) (Figure 16; Table 2).

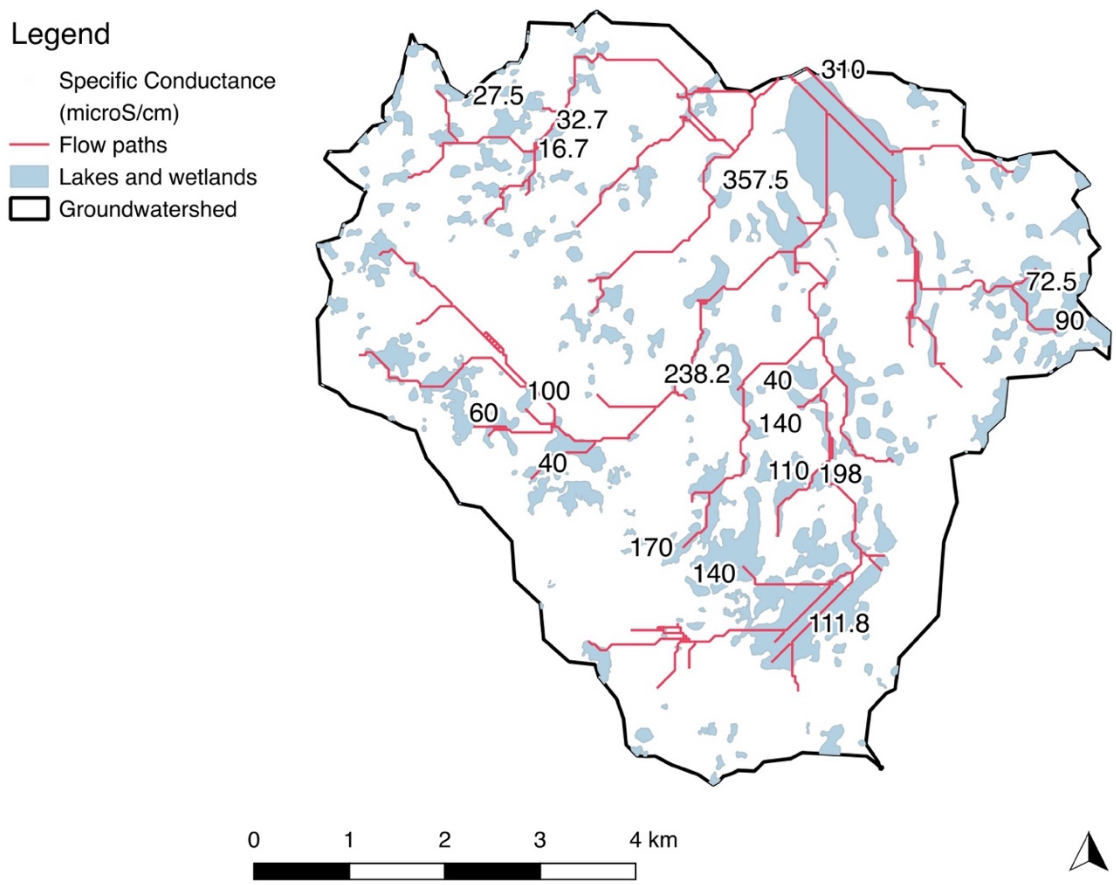


Figure 15. Groundwater flow path lines (in red) for the study area with specific conductance measurements for sampled lakes and wetlands.

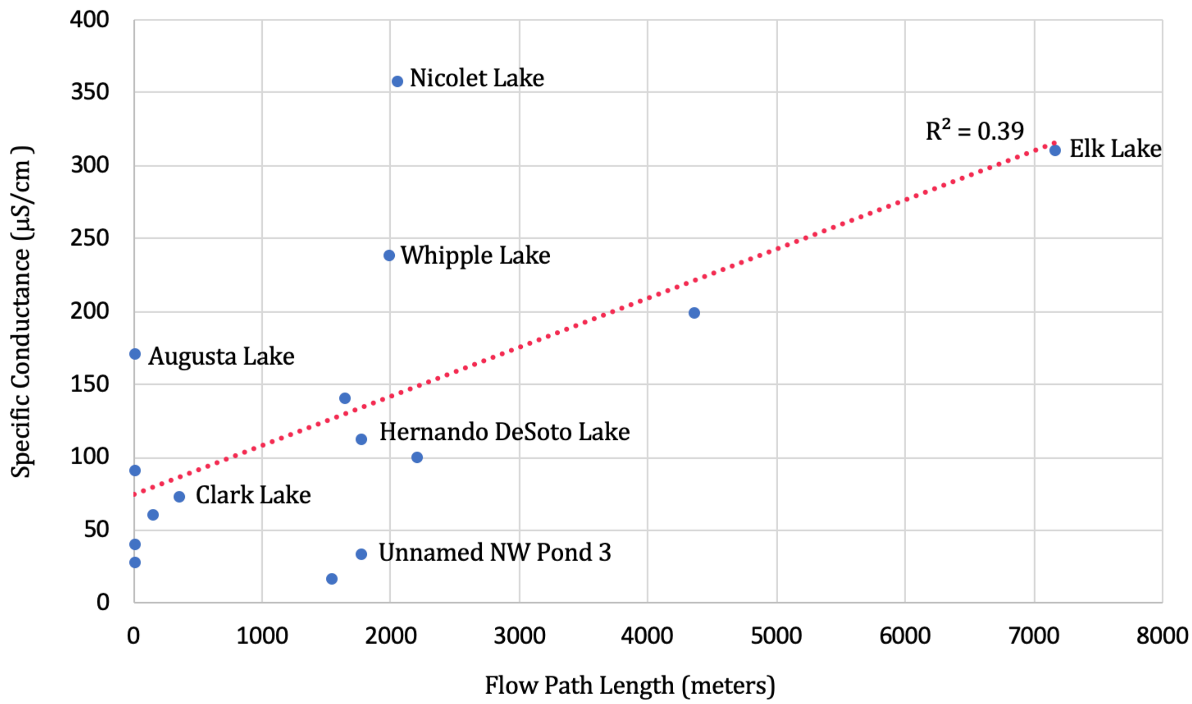


Figure 16. Graph of flow path lengths and specific conductance measurements for sampled lakes and wetlands.

Parameter	Specific Conductance
R Square	0.39
p-value	0.013

Table 2. Regression statistics computed from flow path lengths and specific conductance measurements.

Watersheds

The surface watershed-groundwatershed area ratio was computed for the lakes and wetlands that were sampled for pH and specific conductance. In this case, a large ratio represents a larger surface watershed while a smaller ratio represents a larger groundwatershed for a particular lake or wetland. Ratios close to 1 indicate that the surface watershed and groundwatershed are of similar size.

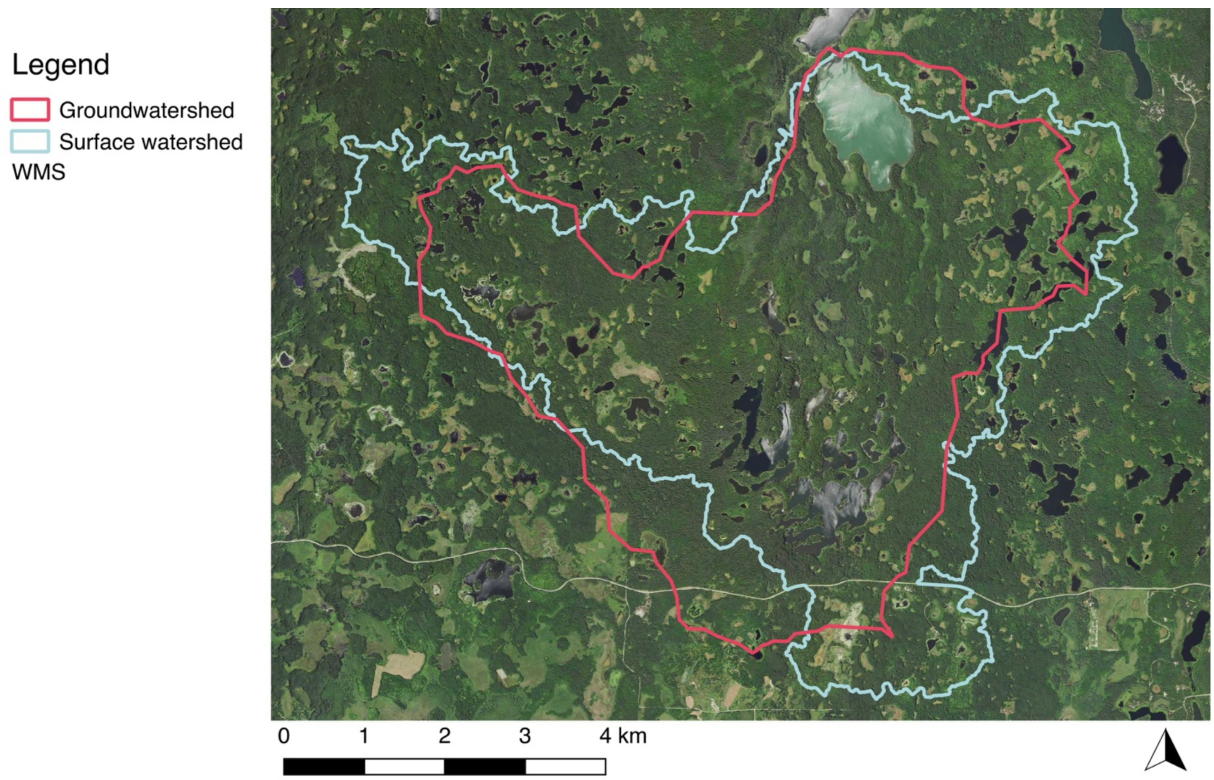


Figure 17. Surface watershed and groundwatershed delineated for Elk Lake.

For example, Elk Lake has a similarly sized surface watershed and groundwatershed therefore a ratio of 1.12 was calculated (Figure 17). In contrast, some of the ponds have a larger surface watershed compared to its groundwatershed (Figure 18.a.) and others have a groundwatershed that is estimated to be 25 times larger than its surface watershed (Figure 18.b.). For the lakes and wetlands that were sampled, ratios ranged from 0.04 to 6.11 (Figure 19).

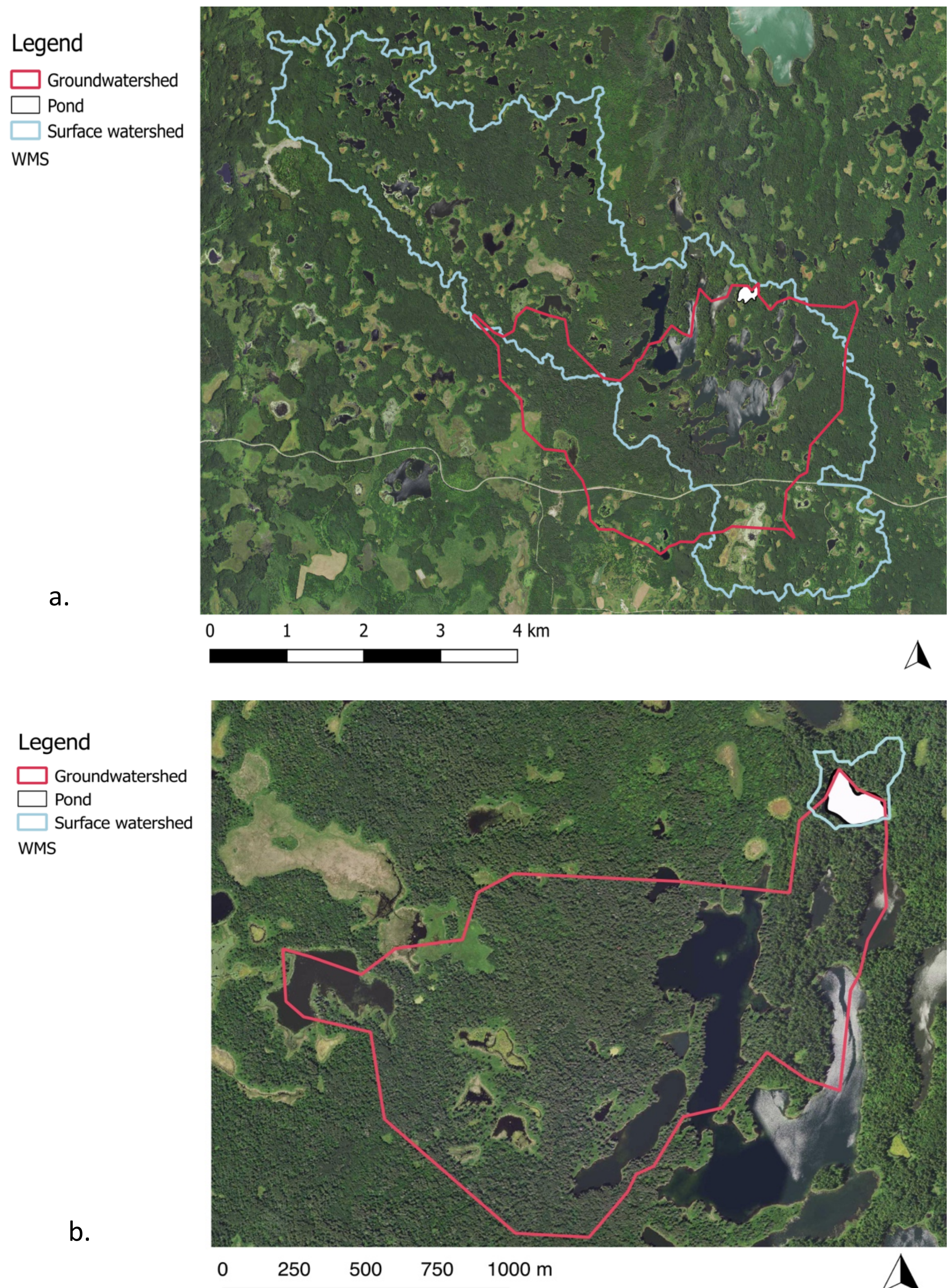


Figure 18. a. Surface watershed and groundwatershed delineated for a sampled pond. This pond had a watershed area ratio of 2.17. b. Surface watershed and groundwatershed delineated for a sampled pond. This pond had a watershed area ratio of 0.04.

Linear regression was used to compare the ratio computed from the surface watershed and groundwatershed areas to the average pH and specific conductance field measurements. In both cases, the results were not found to be statistically significant (R^2=0.02 and R^2=0.07) (Figures 19 & 20, Tables 3 & 4). Although, when the ratio computed from the watershed areas and specific conductance were compared, specific conductance appears to begin to converge and become less variant when the ratio of the surface watershed area to groundwatershed area increases. There also are a few small lakes whose high pH measurements skew the results. For example, if the Triplet Lakes pH data were removed the correlation would improve (R^2=0.23, p-value=0.054).

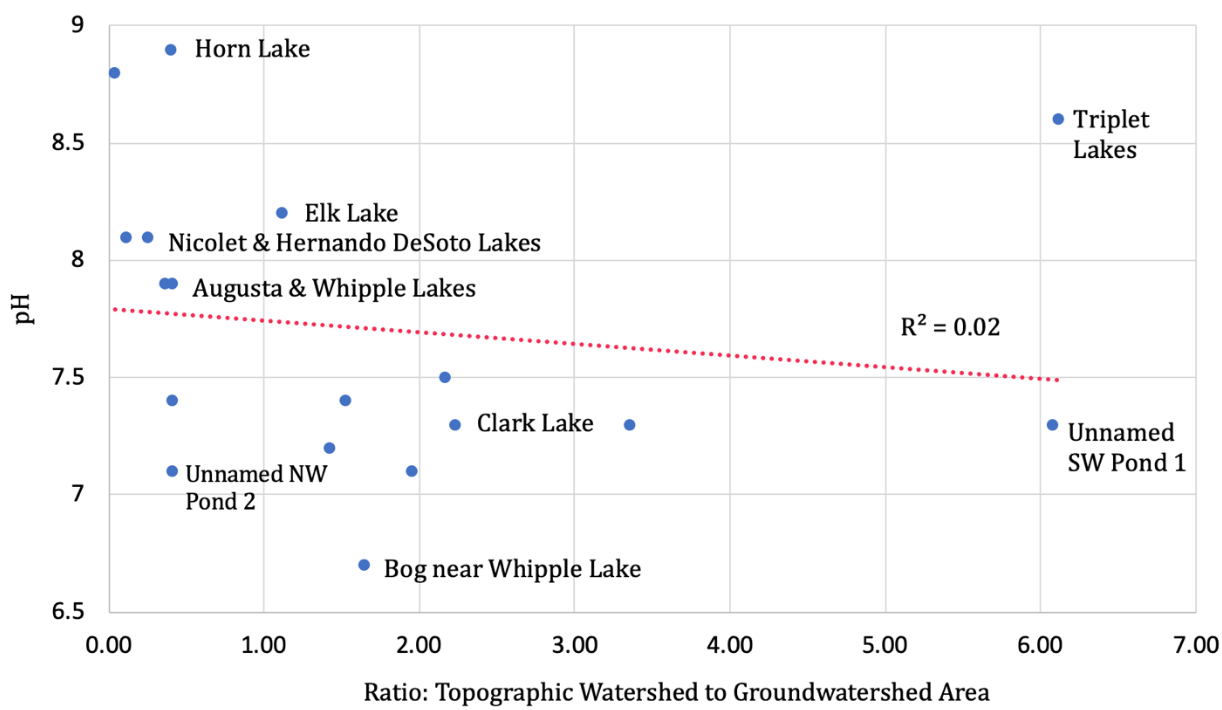


Figure 19. Graph of the watershed area ratio and pH measurements for sampled lakes and wetlands.

Parameter	pH
R Square	0.02
p-value	0.564

Table 3. Regression statistics computed from the watershed area ratio and pH measurements.

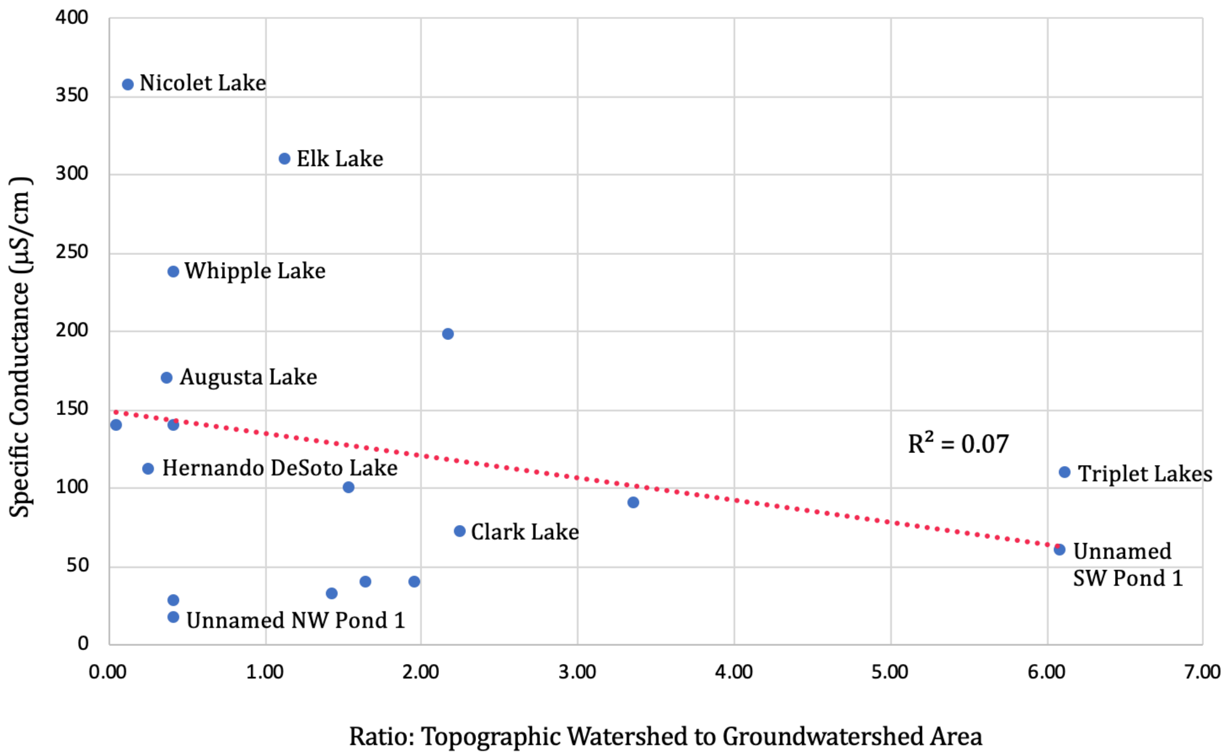


Figure 20. Graph of the watershed area ratio and specific conductance measurements for sampled lakes and wetlands.

Parameter	Specific Conductance
R Square	0.07
p-value	0.279

Table 4. Regression statistics computed from the watershed area ratio and specific conductance measurements.

Field Measurement Location	Name	Date Sampled	Average Specific Conductance (µS/cm)	Flow Path Length (m)	Average pH	Surface Watershed Area (m^2)	Groundwatershed Area (m^2)
1	Unnamed NW pond 1	9/1/18	16.7	1535	7.4	145364	355434
2	Unnamed NW Pond 2	9/1/18	27.5	0	7.1	145364	355434
3	Unnamed NW pond 3	9/1/18	32.7	1764	7.2	4374080	3075606
4	Bog near Whipple Lake	9/8/18	40	-	6.7	13320	8100
5	Nicolet Lake	9/8/18	357.5	2047	8.1	213328	1832891
6	Triplet Lakes	9/8/18	110	-	8.6	201202	32908
7	Unnamed Pond near Little Elk Lake	9/8/18	198	4359	7.5	20758800	9556850
8	Unnamed Pond near Whipple Lake	9/8/18	140	1638	8.8	69337	1725230
9	Whipple Lake	9/8/18	238.2	1982	7.9	874391	2104139
10	Augusta Lake	9/15/18	170	0	7.9	202395	549434
11	Unnamed SW pond 1	9/15/18	60	146	7.3	5199954	855804
12	Unnamed SW pond 2	9/15/18	100	2205	7.4	6065608	3967526
13	Unnamed SW pond 3	9/15/18	40	0	7.1	747904	383190
14	Horn Lake	9/15/18	140	-	8.9	91957	226313
15	Hernando DeSoto Lake	9/18/18	111.8	1771	8.1	2011595	7998188
16	Nicolet Creek	9/18/18	430	-	7.5	-	-
17	Clark Lake	9/22/18	72.5	350	7.3	3873821	1731283
18	Unnamed Pond near Clark Lake	9/22/18	90	0	7.3	1986991	592394
19	Chambers Creek	9/29/18	380	-	8.2	-	-
20	Elk Lake Spring 1	9/29/18	590	-	7.4	-	-
21	Elk Lake Spring 2	9/29/18	570	-	7.2	-	-
22	Elk Lake Spring 3	9/29/18	550	-	7.4	-	-
23	Elk Lake Spring 4	9/29/18	610	-	7.5	-	-
24	Elk Lake	9/29/19	310	7164	8.2	37147655	33211802

Table 5. Field measurement data, calculated flow path lengths, and calculated watershed areas for sampled lakes and wetlands.

Recharge-Discharge Map

The recharge-discharge map shows the residual flow after the subtraction of groundwater input and output (Figure 21). A negative residual (orange) represents areas where recharge must be added to balance groundwater flow, while a positive residual (purple) represents areas where discharge must occur to the surface. Near-zero residuals (white) indicate areas with balanced groundwater flow, similar recharge and discharge rates, or a combination of both. Strong discharge occurs on the east side of Elk Lake and a

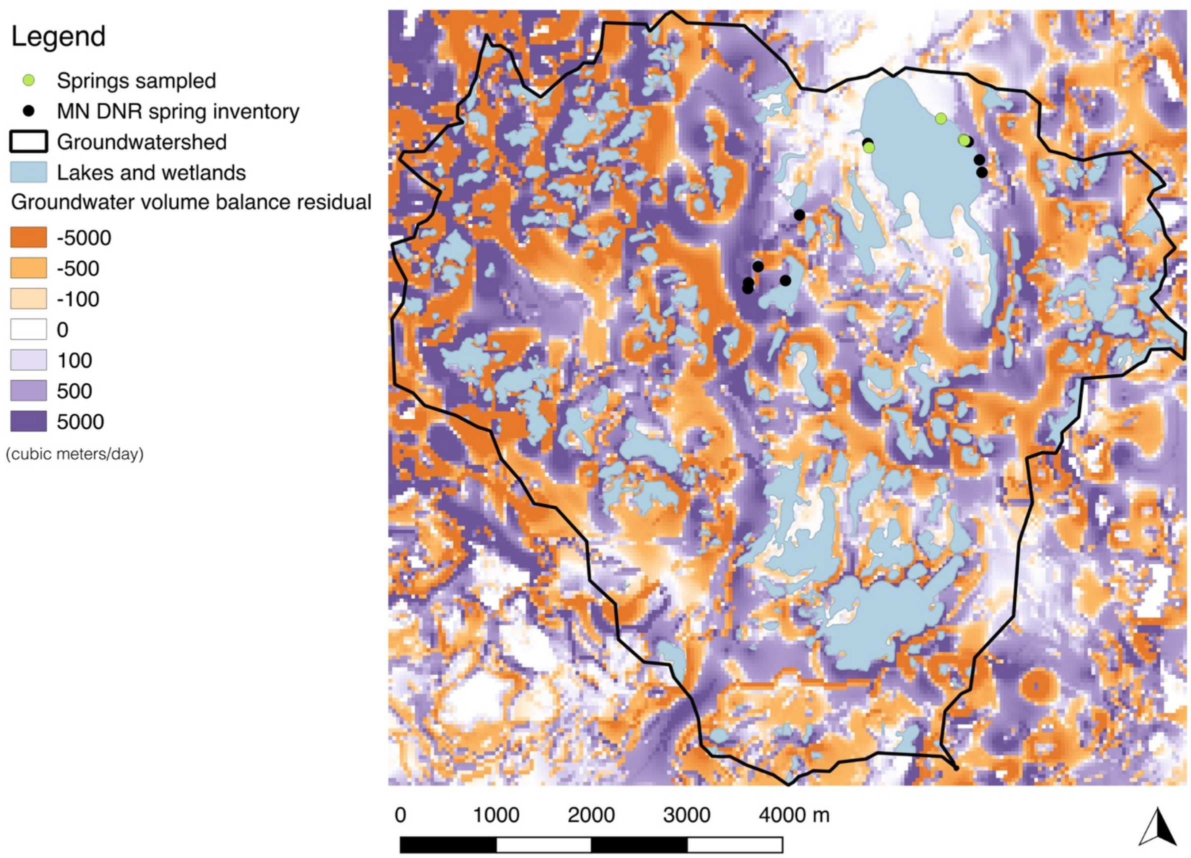


Figure 21. Recharge-discharge map for the study area computed from the unfilled 40-meter interpolated water-table surface.

prominent discharge area to the west of the lake and Nicolet Creek. These features appear to correspond with springs from the Minnesota Spring Inventory (Minnesota Department of Natural Resources 2019b) and those sampled, as well as areas that were mapped as having a water-table slope that exceeds 3 degrees. The springs measured in the study area, as expected from deep groundwater flow, had the highest specific conductance measurements with an average of 580 μS/cm. In addition, one wetland sampled east of Whipple Lake mapped as a strong recharge area surrounded by discharge. Due to low pH measured in the field and the type of vegetation observed, it is believed this wetland may be a raised bog.

Because the raster scale affects the calculated magnitude of local groundwater flow paths, the cell size of the water-table surface raster influences the overall water balance for the groundwatershed. When the sum of the raster cell residuals within the groundwatershed was compared to the surface water discharge measurements made at the outlets in the field, a water-table surface raster with a cell-size of 40 meters provided the best results. This consisted of 0.54 m^3/sec discharging the model, compared to an observed average of 0.38 m^3/sec discharging from the two outlets (Table 6; Table 7) (Figure 22).

Cell Size	Sum (raw)	Sum (filled)
10	-3.84	-4.35
20	-6.08	-6.13
30	-1.16	-2.46
40	-0.54	-0.56
50	-15.89	-16.14

Table 6. Results of the sensitivity analysis (m^3/sec).

Discharge	Min	Max	Mean
Nicolet Creek	0.03	0.18	0.09
Chambers Creek	0.03	1.03	0.29
Total	0.07	1.21	0.38

Table 7. Minimum, maximum, and mean discharge measurements made at the outlets (m^3/sec).

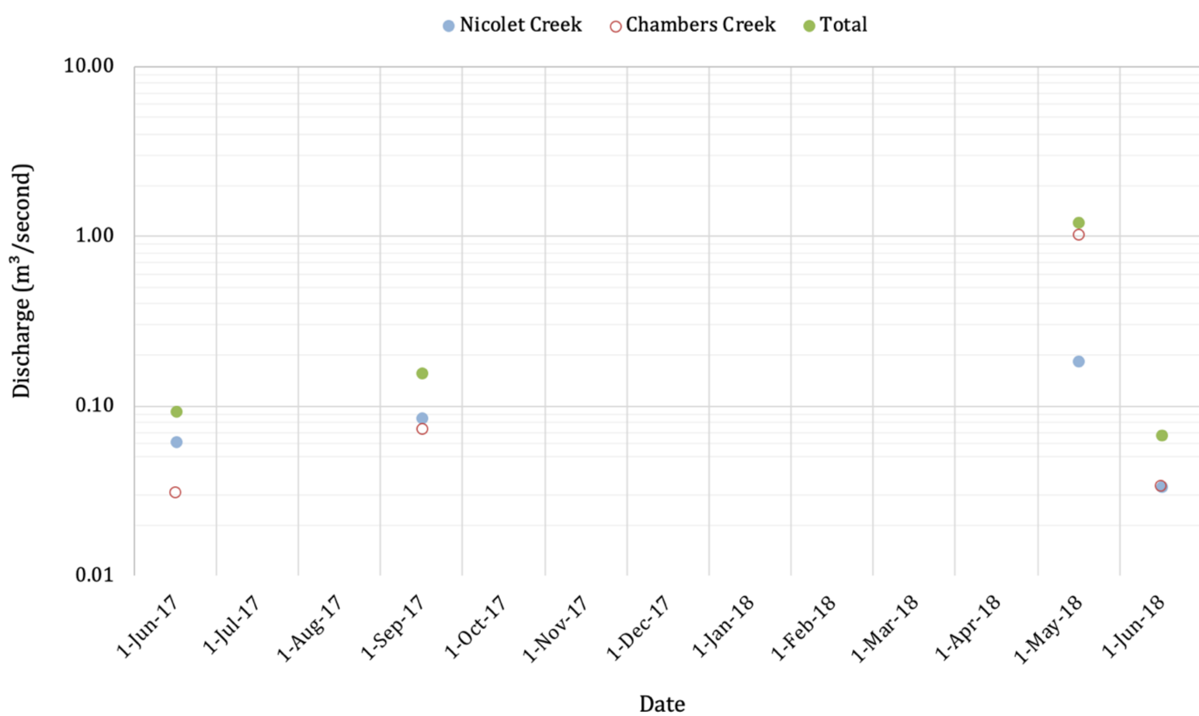


Figure 22. Discharge measurements made at the outlets.

CHAPTER 4: DISCUSSION

Water-Table Surface

Results suggest that the interpolated water-table surface performed well to delineate groundwatersheds, groundwater flow paths, and the mapping of recharge and discharge in the study area. The wealth of lakes, wetlands, and streams and the availability of a 3-meter DEM for the study area provided adequate data for this modeling technique. One challenge with groundwater modeling is attempting to characterize a complex natural system, oftentimes with limited data availability. There are no deep borings or groundwater wells in the study area with detailed soil and sediment descriptions, so the elevations of lakes, wetlands, streams, and hydric soils were the sole source of data used for interpolating the water-table surface and for estimating the depth to the base of the unconfined aquifer. While wetlands with specific codes were selected in an effort to identify the most persistent water-table elevations, it is possible that a portion of the wetlands included in the dataset could be perched or disconnected from the water-table and do not have water elevations that truly reflect the permanent water-table surface. This is also a possibility for the hydric soil polygons that were used for mapping the water-table.

Several areas where the slope of the water-table exceeds 3 degrees were identified. These areas appear to correspond with springs that were sampled as well as those included in the Minnesota Spring Inventory (Minnesota Department of Natural Resources 2019b). The springs in the study area are likely caused by either change in terrain where the water-table intercepts the surface or focused flow related to the occurrence of less permeable deposits (Fetter 2000). The springs could also be caused by preferential flow through high permeability material (Winter et al. 1998).

Groundwater Flow Paths and Field Measurements

pH field measurements were variable in the study area and did not have statistical significance when compared with groundwater flow path lengths. pH is influenced by a variety of factors some of which include aquatic plant productivity, biological activity, and subsurface material (Glaser 1992; Logan 1995; Tucker and D’Abramo 2008). Because aquatic plant productivity and surface water pH in ponds is coupled with diurnal processes (Tucker and D’Abramo 2008), it is possible that correlation may have been observed if pH measurements had been taken at the same time of day at each site, preferably in the early morning hours. This would have significantly increased field sampling days and expense for this study, however.

Specific conductance correlates with groundwater flow path lengths at the 95% confidence limit and was a useful approach in this case. These results suggest it is likely that lakes and wetlands lying on distal portions of groundwater flow paths have budgets that are strongly influenced by groundwater and are more mineralized; those lying on proximal portions of a groundwater flow path are not strongly influenced by groundwater and are less mineralized. A few exceptions in the model need to be highlighted. Lakes that lie at the beginning of flow path lines, including Augusta Lake and Horn Lake, had large specific conductance relative to their path line position. In addition, Nicolet Lake, which had the highest specific conductance measurement in the study area, lies toward the middle of a modeled groundwater flow path. These specific conductance measurements suggest that these lakes are at the end or middle of local flow paths, which is contradictory to the modeled flow paths. It is possible that Nicolet Lake is being fed by significant seeps or springs. The area of its groundwatershed is approximately nine times larger than the area of its surface watershed. The lake also lies at the base of the steeply sloping zone of the water-table and a groundwater spring has been mapped just south of the lake. Unnamed ponds in the northwest portion of the study area had the smallest specific conductance measurements but lie toward the middle of modeled flow paths. The specific conductance measurements suggest these ponds are at the beginning of local flow paths but again contradict the modeled flow paths. It would be useful to sample ponds farther upgradient to determine if they have even lower specific conductance.

The distance groundwater has traveled in the subsurface is only one of many factors that have the ability to influence water quality, including specific conductance. For example, the residence time is also important to consider because the longer water remains in the subsurface the more likely its dissolved solids concentration will increase (Freeze and Cherry 1979). The composition of the subsurface materials and the sequence at which the groundwater encounters these materials are also important to consider as certain soil, rock, or sediment types may have minerals that are more prone to dissolution than others (Freeze and Cherry 1979). For example, some shale fragments that match the composition of drift from the St. Louis sublobe have been found in the park and could influence mineralization differently than the otherwise calcareous drift. They are believed to have been deposited when outwash from the St. Louis sublobe utilized drainage features, such as tunnel valleys, created by the wastage of the Wadena lobe (Wright 1993). While the mineralization of water is a natural occurrence in the subsurface, it can be enhanced by human activities. For example, the use of salt on roads (Novotny et al. 2007) or other human activities such as the extraction of natural resources (Cormier et al. 2013) can increase the specific conductance of surface water bodies. Although specific conductance and pH are influenced by several factors, one advantage they have over other types of water quality sampling is that they are relatively easy and inexpensive to measure.

Watersheds and Field Measurements

Neither pH nor specific conductance were found to correlate with the ratio calculated from the surface watershed and groundwatershed areas. While more than half of the lakes or wetlands sampled in the study area had a significant size difference between their surface watershed and groundwatershed, the pattern of pH and specific conductance field measurement results could not be predicted from the surface watershed to groundwatershed ratio. While basin size has been positively correlated with residence time (Wolock et al. 1997), which is an important factor in mineralization (Winter et al. 1998) and pH change (Wolock et al. 1997), other properties of the watershed can influence the contribution to lakes and wetlands. For example, the surface watershed contribution is subject to soil properties, precipitation amounts, and losses from evapotranspiration. The groundwatershed contribution is subject to subsurface properties and recharge and discharge rates (Minnesota Department of Natural Resources 2019c; Winter et al. 1998). It is still useful to delineate both surface watersheds and groundwatersheds, however, especially for management purposes. For example, there can be large differences between the two (e.g., Figure 18) and knowing the catchment area for both is useful when considering land use changes, groundwater development, and contaminant transport. It is likely that the areal extent of groundwatersheds differs between dry and wet conditions due to fluctuations of the water-table surface that result from changes in recharge and discharge (Winter et al. 1998). This is also an important consideration for both modeling and management. If the elevations of surface water features were available for wet or dry conditions, the water-table surface could be reconstructed using this technique to determine the changes that result in the aerial extents of groundwatersheds.

Recharge-Discharge Map

Strong discharge areas in the study area appear to correspond to mapped groundwater springs and areas of the water-table with steep gradients. It would be useful to explore other strong discharge areas in an attempt to identify additional springs. In addition, one wetland sampled east of Whipple Lake, believed be a raised bog due to low pH and vegetation, mapped as a strong recharge area surrounded by discharge. This type of recharge-discharge pattern could also be further explored in an attempt to identify additional bogs within the Itasca Moraine and elsewhere.

For the recharge-discharge map, it was assumed that the K_{sat} value obtained from the SSURGO soils (Soil Survey Staff 2019) was a reasonable estimate for hydraulic conductivity of the shallow glacial sediments and that the unconfined aquifer is laterally isotropic. Because there are no observation wells or data, no slug or pump tests, nor information about heterogeneities in the subsurface, the K_{sat} value obtained from the SSURGO soils, mapped at a scale of 1:20,000 and 1:24,000 (Soil Survey Staff 2019) is the most spatially complete and detailed source for the hydraulic conductivity of surficial sediments available in the study area. In addition, no K_{sat} values were available for lake sediments within the study area so they were assigned a K_{sat} value similar to that of wetlands in the study area. This is assumed to be reasonable as many lakes in the study area have accumulated large amounts of organic sediments. According to Wright (1993), the organic sediments of lakes in the area can range from 8 to 12 meters thick. The hydraulic conductivity assigned to the lakes is in range of what Międsiak-Wójcik et al. (2018) found in lake bottoms composed of fine organic sediment with low percentages of fine sand. The aquifer base elevation, which was calculated by Tschann (2019) using an analytical model developed by Bresciani et al. (2016), was also calculated using the K_{sat} values (Soil Survey Staff 2019) and recharge estimates (Smith and Westenbroek 2015). Therefore, the aquifer base elevation is not based on observational data of a confining layer because detailed subsurface information for the study area is not available.

The computed recharge and discharge rates shown in the recharge-discharge map (Figure 21) are greater than regional estimates (Smith and Westenbroek 2015). The large fluxes likely resulted from excessive hydraulic conductivity estimates based on soils (Figure 8) rather than glacial sediments, and/or aquifer saturated thickness values derived from a constant aquifer base elevation. Because of this, the recharge-discharge map (Figure 21) should only be used to qualitatively map recharge and discharge zones and not to quantify recharge and discharge rates in the study area.

Future work with this method could use a constant thickness hydrologically active zone for the aquifer rather than relying on a fixed aquifer base elevation across the entire study area. This approach would reduce the large residuals that occur in more elevated areas where the aquifer thickness is greatest. It also would be useful to explore how the flux values respond to changes in the assigned hydraulic conductivity. It is possible that the values estimated from the soils are too high and do not adequately represent the more compacted underlying sediments (Cravens and Ruedisili 1987; van der Kamp and Hayashi 2009). Deep borings and data supplied by slug or pump tests would be useful to refine the hydraulic conductivity and the aquifer thickness estimates but would significantly increase the expense of this study.

The results of the sensitivity test suggested that the 40-meter cell size and the raw (unfilled) water-table surface provided the best match to the discharge measurements from the two outlets for the study area. Based on the comparison to surface water discharge from the watershed, it is likely that the analysis using 40-meter cells best characterizes the overall water budget of the local flow systems that occur in the study area. While this is much coarser than the DEM used to determine the surface watersheds, it has been suggested that rasters with a coarser resolution are more appropriate for the water-table surface. This is based on the assumption that the water-table surface is smoother and more subdued than the surface topography (Wolock and Price 1994; Sørensen and Seibert 2007). While filling of the water-table surface was appropriate for delineating groundwatersheds, it did cause increased net residual discharge for the study area in the groundwater volume balance residual raster. This is because the filling of the surface results in small increases in the heads of various cells that were once sinks (Tarboton et al. 1991). The difference between the discharge of the filled and raw (unfilled) water-table surface was as little as 0.02 m^3/sec for the 40-meter cell size trial and as much as 1.3 m^3/sec for the 30-meter cell size trial. Because DEM surfaces are filled to ensure a connected drainage structure and therefore enable delineation of watersheds and groundwater flow path lines, using a filled raster for the production of the recharge-discharge map is probably unnecessary. In addition, these areas could represent actual groundwater sinks and may warrant further investigation. One example of a groundwater sink is a wetland that loses a greater amount of water to evapotranspiration than it gains from groundwater seepage, resulting in a cone of depression (Winter et al. 1998). Detailed physical monitoring of the wetland's budget would be needed to confirm whether or not it is a true sink.

Specific Conductance, Flow Paths, and Mapped Glacial Sediments

When comparing the flow paths and average specific conductance measurements to the mapped glacial deposits, measurements of 100 μS/cm or less were recorded in lakes and wetlands located on the more elevated areas along the moraine axis. The bog near Whipple Lake is an exception and lies within an area of generally greater surface water conductance. Specific conductance measurements greater than 100 μS/cm were measured in the lakes and wetlands situated on the glacial outwash or remnant tunnel valley features that pass through the center of the study area. Similarly, in the Young Glacial Area in northern Poland, Jaworksa-Szulc (2016) found that recharge or losing lakes with low total dissolved solids (TDS) were commonly identified on till while discharge or gaining lakes with high TDS were commonly identified within glacial tunnel valleys. These results show the importance of hydraulic conductivity and position within the groundwater flow system for characterizing surface water and groundwater interaction (Winter et al. 1998).

Discussion of Specific Conductance Variability Between Sites

As discussed previously, the lakes and wetlands of the Cottonwood Lake area of the Missouri Coteau, North Dakota, and the Young Glacial Area of northern Poland have been studied in an effort to determine their interaction with groundwater. While these areas and the ablation moraine in Itasca State Park consist of similar glacial terrain, they have characteristically different magnitudes of surface water mineralization.

The ponds, often referred to as prairie potholes, in the Cottonwood Lake area have been extensively studied with specific conductance measurements during 33 years from 1979 to 2012 (LaBaugh et al. 2016). For comparison to the Itasca site, six Cottonwood ponds were selected (P1, P8, T3, T4, T5, T8) (Figure 23). Ponds P1 and P8 have been described as discharge ponds, T3 and T4 as flow-through ponds, and T5 and T8 as recharge ponds (Winter and Rosenberry 1998). In this region, precipitation is exceeded by evapotranspiration (Winter et al. 2001) and the wetlands, on average, are much more mineralized than those found in Itasca State Park. The primary inputs of water to the Cottonwood potholes consist of precipitation and snowmelt. The higher evapotranspiration rate in this area is partly responsible for the increased specific conductance as it causes salts to accumulate in the ponds. The ponds in this area are believed to lose little water to groundwater recharge and the groundwater input, or seepage, to the ponds is believed to be small (Shjeflo 1968; Sloan 1972). In cases where the ponds do interact with groundwater, dissolved minerals can be removed from the ponds through groundwater recharge while large amounts of minerals can be supplied to the ponds through groundwater seepage (Eisenlohr 1972). The two discharge ponds selected, P1 and P8, had an average specific conductance of 1900 μS/cm. The two flow through ponds, T3 and T4, had an average of 1400 μS/cm while the two recharge ponds, T8 and T5, had an average of 190 μS/cm.

The glacial lakes of the Young Glacial Area of northern Poland have total dissolved solids (TDS) measurement data from 2010 to 2012 (Jaworska-Szulc 2016). Of these lakes, 15 were described as being discharge lakes, three as flow-through lakes, and seven as recharge lakes. Lakes that were described as losing and gaining and/or losing with the pond situated slightly above the aquifer were not included, as well as lakes with waters that were described as impaired. The TDS data for the 25 lakes included for comparison were empirically converted to an approximate specific conductance measurement by dividing the TDS values by 0.6 (Hem 1985) (Figure 23). This site has been described as having a precipitation rate nearly equal to that of evapotranspiration (Jaworska-Szulc 2016) and the lakes are, on average, less mineralized than those found in the Cottonwood Lake area. The 15 discharge lakes (#1-15) had an average calculated specific conductance of 460 μS/cm. The three flow-through lakes (#26, 34, 35) had an average of 180 μS/cm and the seven recharge lakes (#16-19, 21-23) had an average of 110 μS/cm.

As discussed previously, the gaining lakes at this site are the most mineralized, the flow-through lakes have intermediate mineralization, and the losing lakes have the least mineralization (Jaworska-Szulc 2016). This also is likely the general trend in the Cottonwood Lake area and Itasca State Park, although the magnitude of mineralization is strikingly different among the three sites (Figure 23).

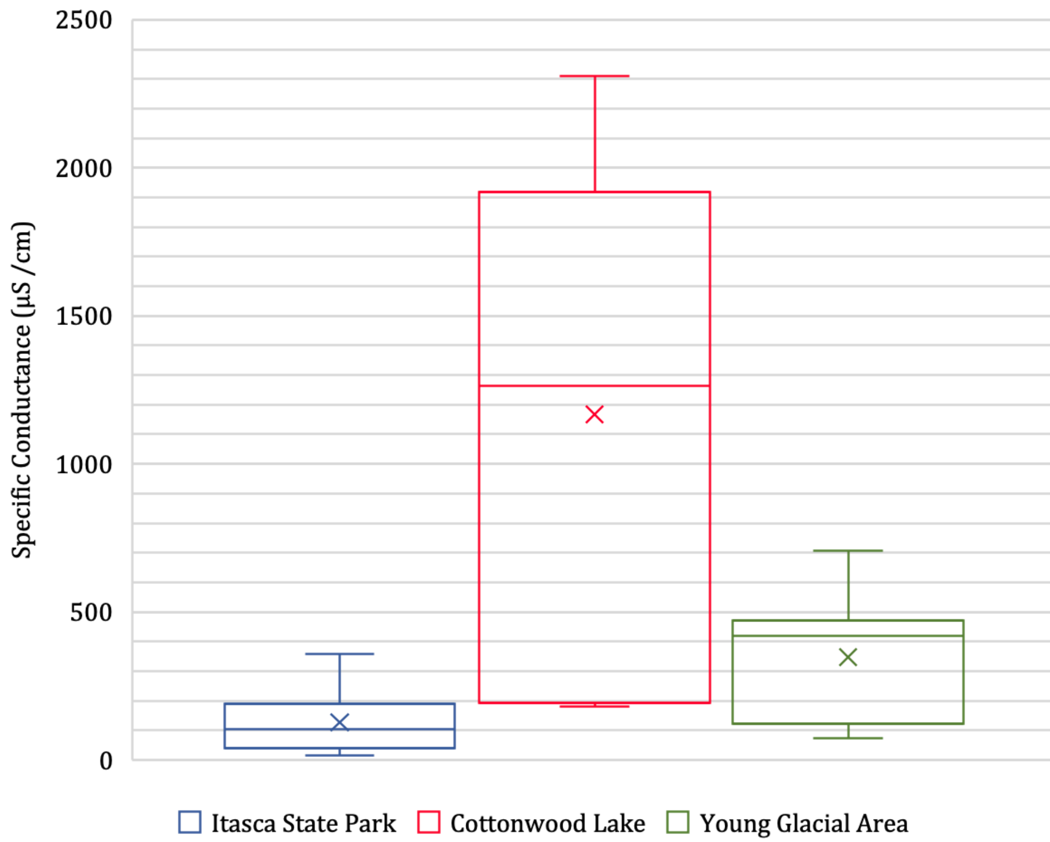


Figure 23. Specific conductance measurements for the Cottonwood Lake area (North Dakota), Young Glacial Area (northern Poland), and Itasca State Park (Minnesota) using averaged values for the types of ponds and lakes sampled. In the plots, the x represents the mean, the middle bar the median, the upper box the third quartile, the lower box the second quartile, the upper whisker the fourth quartile and the maximum, and the lower whisker the first quartile and the minimum.

In this comparison, it is important to emphasize that the ponds in the Cottonwood Lake area have evapotranspiration rates that exceed precipitation by approximately 30 cm/year (Winter et al. 2001), whereas these rates are approximately equal for the Young Glacial Area (Jaworska-Szulc 2016) and precipitation exceeds evapotranspiration by up to

5 cm/year in the Itasca State Park region (Vaughan 2017). It is possible the that lower evapotranspiration rates and the increased interaction with groundwater occurring at Itasca State Park and the Young Glacial Area are resulting in a flushing effect which has kept dissolved minerals from becoming concentrated in the lakes and wetlands. The Cottonwood Lake area ponds have been described as having less interaction with groundwater in terms of their water budget which is one mode at which minerals are imported and exported. They also have increased evapotranspiration which can lead to mineralization of water in the ponds through concentration (Sloan 1972; Shjeflo 1968). When water levels are high, the potholes can fill with water and spill into adjacent basins which also can export minerals (Leibowitz and Vining 2003). Heagle et al. (2013) also found that the sediments in prairie potholes can have a large influence on pond salinity in Saskatchewan. This particular wetland is moderately saline but is believed to have little surface water inflow and interaction with groundwater. They suggested that past dry periods have caused a large mass of sulfate to accumulate in the wetland's sediments and that the sediments release ions to the surface water during wet periods which maintains salinity. These are all factors that could contribute to the higher mineralization found in the prairie potholes.

It is also important to consider land use when comparing sites. The Itasca State Park location is relatively undisturbed forest while the Cottonwood Lake area is primarily native prairie (LaBaugh 2016). The Young Glacial Area is a mix of forested and agricultural land (Jaworska-Szulc 2016). Depending on the type of land use or cover, the runoff or overland flow from the land surface can increase the specific conductance in lakes and wetlands in some cases, even with minimal groundwater interaction. This scenario was shown not to be the case in the native prairie of the Cottonwood Lake area but has been observed at the wetland site in Saskatchewan (LaBaugh 2016; Nachshon et al. 2014) where Heagle et al. (2013) found large concentrations of sulfate in the wetland sediments. This site was found to experience increased specific conductance during intense precipitation events. This has been suggested to be a result of its catchment which is efficient at shedding water due to past land use (Nachshon et al. 2014; LaBaugh et al. 2016). In addition, three lakes that were described as losing or recharge lakes in the Young Glacial Area also had increased specific conductance and were described as being contaminated or impaired (Jaworska-Szulc 2016).

It is interesting to consider climate change when comparing the mineralization between sites. Elk Lake in Itasca State Park has been the focus of many paleoclimate studies as it lies approximately 80 kilometers east of the forest-prairie boundary. Elk Lake is unusually deep (>30 meters) and has well-developed annual varves that have provided key information on past climate variability (Anderson 1993; Whitlock et al. 1993; Bartlein and Whitlock 1993; Clark 1993). The results of studies on these varves suggest that the Itasca State Park area is sensitive to climate change and was once a prairie in the warm and dry mid-Holocene (Anderson 1993). Although the lakes and wetlands in Itasca have the lowest mineralization among the three sites, it is possible that the lakes and wetlands present in this area in the mid-Holocene were similar hydrogeochemically to the conditions that occur currently in the Cottonwood Lake area.

CHAPTER 5: CONCLUSION

Application of this GIS-based groundwater modeling technique in the Itasca Moraine provides valuable water budget information and would likely produce good results in other areas with abundant streams, lakes, wetlands, and hydric soils. Unfortunately, subsurface geological information was not available for the Itasca Moraine site, but having these data would strengthen model outcomes. Flow path length and specific conductance showed a significant correlation, suggesting that this relatively easy and inexpensive technique that couples GIS modeling and field measurements can help characterize the contribution of groundwater in wetlands.

Although the ratio of watershed areas was not shown to be significantly correlated to either pH or specific conductance, knowing the catchment for both surface water and groundwater is still imperative to make informed management decisions. This is because the two catchments do not always correspond (Winter et al. 2003) and changes to either of the catchments, in the form of land use or groundwater development for example, can directly affect the quantity and quality of water within a lake or wetland. Mapping recharge and discharge also can be of use to management for identifying groundwater springs and sensitive wetlands such as bogs. In addition, knowing the distribution of strong recharge and discharge sites can be of use when making decisions that affect land use within a watershed.

The Cottonwood Lake prairie site and the Young Glacial Area in northern Poland also show a trend of increased specific conductance in lakes and wetlands with increased groundwater interaction. Results for the Itasca Moraine suggest that this GIS modeling technique would function similarly in those areas, even with contrasting precipitation-evapotranspiration ratios and the different magnitudes of surface water mineralization. In addition, the prevalent land use or land cover of the watershed and the nature of sediments in the wetlands should be considered in the GIS analysis as they been shown to influence specific conductance measurements of surface water.

APPENDICES

Appendix I:

Water-Table Surface GIS Methods

All of the following described datasets were analyzed using the NAD 1983 UTM Zone 15N projection. The following general steps were taken in ArcGIS to interpolate the water-table surface:

1. The Minnesota National Wetland Inventory Update (Minnesota Department of Natural Resources 2018b) was clipped to the extent of the DEM. To accomplish this, a polygon was constructed that matched the extent of the DEM and used to clip the wetland inventory shapefile using the *Clip* tool. The DEM extent was input as the *Clip Feature* and the wetland inventory was input as the *Input Feature*.
2. *Snap to Raster* and *Extent* were set to the DEM in the *Environment Settings.*
3. Wetlands that did not have the specific hydrologic modifier codes (Cowardin et al. 1979) were selected in the attribute table and removed from the dataset by deleting them from the attribute table while in edit mode.
4. The edited wetland layer was underlain by aerial imagery (Minnesota Geospatial Information Office 2017) and wetland polygons were observed to verify their extents.
5. If more than one polygon existed for each wetland, smaller or encompassed polygons were merged into the feature with larger extent. If polygons had an island or internal feature that was excluded from the polygon, this feature was removed. If polygons were created as a multi-part feature, they were separated and assigned their own unique objectID.

6. Once the edits to the updated wetland inventory were complete, wetland polygons were converted to a line file using the *Feature to Line* tool*.*

7. The wetland polygons were then converted to a point file using the *Feature to Point* tool. Before doing so, a box was checked to ensure that the point was placed in the inside of the polygon.

8. It was then necessary to check and ensure that there were an identical number of features in the polygon, line, and point files and that the objectIDs for the corresponding features matched. If there were excess features in the line or point files, the wetland polygons were checked to ensure that no nested polygons or island-type features remained. If they were found, they were removed.

9. The wetland polygons were then converted to a raster using the *Polygon to Raster* tool. The *Input Feature* was the wetland polygon shapefile while the *Value Field* was set to the objectID and the *Cellsize* to the DEM.

10. Next, the *Zonal Statistics* tool was used to create an additional raster that included the minimum elevation value for each polygon. For the *Zonal Statistics* tool, the raster created in the previous step was used as the *Input Raster*, the *Value* was set as the zone field, the DEM was set as the *Input Value Raster*, and Minimum was selected for the *Statistics Type*.

11. The *Extract to Points* tool was then used to extract the elevation from the Zonal Raster to the corresponding point within each wetland polygon.

12. The extracted elevations were joined with the line file created previously and the polygon file using the corresponding objectID. The line and polygon files were then checked to ensure that the features had reasonable elevation values assigned to them.

13. A stream-line shapefile from the National Hydrography Dataset (U.S. Geological Survey 2019) was added. The stream-lines were inspected and edited to ensure that they did not intersect wetland polygons. In many cases, it was easier to re-draw the stream line than edit the vertices. Stream-lines with few vertices were redrawn to include more vertices or redrawn to better correspond with stream channels observed using the aerial imagery.
14. The *Feature Vertices to Points* tool was then used to convert the stream-line vertices to a point file.
15. The *Extract Values to Points* tool was used to extract the elevation from the DEM to each of the corresponding stream vertices (points).
16. The *Topo2Raster* tool which is included in the *Spatial Analyst* tools and based on the ANUDEM algorithm (Hutchinson 1988; Hutchinson 2011) was used to interpolate the water-table surface. The wetland polygon line file was input as *Contours* and the DEM elevation field was selected. The wetland polygons were input a second time but as *Lakes*. The stream-line file was input as *Streams* and the stream-line vertices (point) file was input as *Point Elevation* with the DEM elevation field selected. The *Output Cell Size* was set to 30 meters to assure convergence.
17. Once the water-table surface was generated, it was compared with the DEM using the *Raster Calculator* tool. For comparison, the water-table surface raster was subtracted from the DEM and the resulting raster was visually inspected for areas of the water-table that lay above the DEM.

18. SSURGO soil polygons (Soil Survey Staff 2019) were added to provide more elevation control/constraint on the water-table at sites that lay above the DEM.

19. In addition, polygons were constructed in areas where standing water was observed in the aerial imagery and stream lines where dendritic drainage features could be discerned. In areas where there was dense vegetation, a color ramp was assigned to the DEM to enhance areas with drainage features or low-lying depressions for easier identification. These features were then either represented by a line or a polygon feature and added to the existing dataset.

20. Interpolating the water-table surface was an iterative process. Only 2-5 additional features were added for elevation control at a time before re-running the *Topo2Raster* tool. The resulting water-table surface was then observed and subtracted from the DEM using the *Raster Calculator* tool. It took approximately 30 trials using the Topo2Raster tool to create the final water-table surface.

Appendix II:

Groundwater Flow Path GIS Methods

All of the following described datasets were analyzed using the NAD 1983 UTM Zone 15N projection. The following general steps were taken in ArcGIS to model the groundwater flow path lines:

1. The *Fill* tool was used to fill depressions in the water-table surface raster.
2. *Snap to Raster* and *Extent* were set to the water-table surface in *Environment Settings.*
3. The direction of flow from each raster cell was determined using the *Flow Direction* tool.
4. The *Flow Accumulation* tool was then ran using the results of the *Flow Direction* tool. Cells that had values greater than 250 were used as flow lines and were inspected visually using a classified color ramp.
5. The flow lines were then converted to vector line features using the *Raster Calculator* and the *Raster to Polyline* tools. The flow line length was calculated for each line feature and added to the attribute table using the *Calculate Geometry* option.

Appendix III:

Watersheds GIS Methods

All of the following described datasets were analyzed using the NAD 1983 UTM Zone 15N projection. Surface watersheds and groundwatersheds were delineated with the following general steps in ArcGIS:

1. The *Fill* tool was used to fill depressions in the DEM and water-table surface.
2. The *Flow Direction* and *Flow Accumulation* tools were ran with the same procedure listed previously in the Groundwater Flow Path GIS methods.
3. The Flow Accumulation raster was observed along with the aerial imagery and an outlet or outlets were selected based on the results of the flow accumulation raster. Point features were placed at these cells. If delineating the watershed for a wetland, the wetland polygon was converted to a raster using the *Polygon to Raster* tool.
4. Either the outlet point(s) or wetland raster described above were used as the outlet with the Flow Accumulation raster in the *Watershed* tool. This produced a raster that outlined the watershed.
5. The watershed raster was then converted to a vector using the *Raster to Polygon* tool and the watershed area was calculated and added to the attribute table using the *Calculate Geometry* option.

Appendix IV:

Recharge-Discharge Map GIS Methods

All of the following described datasets were analyzed using the NAD 1983 UTM Zone 15N projection. The recharge-discharge map and its necessary input rasters were created with the following general steps in ArcGIS:

1. The saturated thickness raster is not necessary for unconfined conditions but is needed as an input to the Darcy Flow tool. This raster was created by subtracting the aquifer base elevation, from the interpolated water-table surface raster using the *Raster Calculator* tool. Ultimately, the saturated thickness is the same as the water-table elevation above the aquifer base elevation (datum).
2. The hydraulic head input raster was created by squaring the saturated thickness raster using the *Raster Calculator* tool.
3. The transmissivity raster is not necessary for unconfined conditions but is needed as an input to the Darcy Flow tool. It was created using a hydraulic conductivity raster and the saturated thickness raster. Transmissivity is a product of saturated thickness and hydraulic conductivity, so these two raster datasets were multiplied together using the *Raster Calculator* tool. Prior to using the *Darcy Flow* tool, units for K_{sat} provided by the SSURGO soils polygons (Soil Survey Staff 2019) were converted from μm/sec to m/day by dividing the values by 11.574. The hydric soils shapefile was then converted to a raster using the *Polygon to Raster* tool. The output cell size was set to match the water-table raster.
4. The effective porosity is not needed for this application as we are interested in specific discharge rather than the groundwater velocity. A raster with cell values of

1 was used for the effective porosity raster to achieve this. This raster was created with the *Raster Calculator* tool by dividing the water-table surface by itself.

Appendix V:

Recharge-Discharge Map Sensitivity GIS Methods

All of the following described datasets were analyzed using the NAD 1983 UTM Zone 15N projection. The following general steps were taken in ArcGIS for the sensitivity analysis:

1. The *Topo2Raster* tool was rerun and the output cell size was changed to the desired size (10, 20, 40, or 50 meters).
2. If the sensitivity analysis was focused on the results of a filled water-table raster (not raw), the *Fill* tool was used.
3. The input rasters needed for the *Darcy Flow* tool were recreated using the water-table surface raster with the updated cell size.
4. The hydraulic conductivity (K_{sat}) raster was recreated with an updated corresponding output cell size. This was achieved using the *Polygon to Raster* tool.
5. The *Darcy Flow* tool was rerun using the updated rasters.
6. The resulting groundwater volume balance residual raster was clipped to the study area using the *Clip* tool and the Elk Lake and Nicolet Creek groundwatershed polygon.
7. The *Raster Calculator* tool was used to create a raster that served as a zone to calculate statistics using the *Raster Calculator* tool. This resulted in a raster with values of zero that was the same size and shape as the groundwater volume balance residual raster.
8. The *Zonal Statistics as Table* tool was used to calculate the sum of the recharge and discharge. The raster created in the previous step was used as the *Input Zone* and the groundwater volume balance residual raster was used as the *Input Value Raster.*

9. The sum of the raster cells was recorded and used for comparison.

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