AN INVESTIGATION OF GROUNDWATER AND SURFACE WATER INTERACTIONS NEAR A SMALL STREAM IN PRINCE EDWARD ISLAND

by

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ABSTRACT

The province of Prince Edward Island (PEI) currently has a moratorium on the installation of high-capacity irrigation water supply wells, mainly due to concerns over potential interactions with nearby streams. The focus of this research was to conduct an investigation near a small stream in Maple Plains, PEI, and to assess the potential for the site to be used for stream depletion related research. The objectives were to determine whether the stream was gaining or losing, to establish the natural hydraulic gradients, and to characterize the overburden materials.

Stream discharge measurements made in September 2016 indicated that the stream was neither gaining nor losing water. Water elevation data from five drive-point piezometers, two groundwater monitoring wells, an unused residential well, and the stream, revealed that there was an unsaturated zone between the stream and aquifer. The identification of this unsaturated zone only proves that the stream was not directly connected to the underlying aquifer, and does not prove disconnection. The groundwater elevation data also provide evidence that groundwater naturally flows in the same direction as the stream.

Soil samples collected during monitoring well installation showed that there is roughly 3 m of a sand phase till overlying the fractured sandstone aquifer. The hydraulic conductivity of the till appears to decrease with depth. This is also supported by the presence of a perched groundwater system that develops during periods of increased infiltration.
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Finally, I’d like to thank my wife Kate-Lyn Tibbet, for pushing and supporting me at every step of the way.
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LIST OF SYMBOLS, NOMENCLATURE OR ABBREVIATIONS

\( \lambda \) Streambed conductance

\( B \) Aquifer thickness

\( B' \) Aquitard thickness

\( B_s \) Streambed thickness

\( \text{GDP} \) Gross Domestic Product

\( K \) Hydraulic conductivity of the aquifer

\( K_s \) Hydraulic conductivity of the streambed

\( K' \) Hydraulic conductivity of the aquitard

\( L \) Distance between the pumping well and the stream

\( \text{PEI} \) Prince Edward Island

\( Q \) Pumping rate

\( q \) Stream depletion

\( s \) Drawdown

\( S \) Storativity

\( S_s \) Specific storage

\( S_y \) Specific yield

\( T \) Transmissivity

\( t \) Time

\( \text{TOC} \) Top of casing

\( W_s \) Stream width
1 INTRODUCTION

As the global population continues to increase so too does the demand for food and freshwater. Serious issues related to water usage, such as stream flow depletion and groundwater over extraction, have become more commonplace. Consequently, and in part due to the rising costs of food production, the strain on the agricultural sector also continues to climb. In the Canadian province of Prince Edward Island (PEI) the potato industry is an example of this phenomenon. The billion dollar a year potato industry employs 12% of the Island’s workforce and makes up over 10% of the province’s GDP (Cavendish Farms and PEI Potatoes, 2013). Currently, precipitation is the major source of water for agricultural crops in the late summer months, as there is very little to no surface water available in PEI, and the provincial government has issued a moratorium on high capacity irrigation well installation. The province’s potato farmers want the moratorium to be lifted, since producers in other jurisdictions have access to irrigation water allowing for higher crop yield and quality (Cavendish Farms and PEI Potatoes, 2013).

The moratorium was first implemented in 2002 because of the concern that high capacity wells could deplete the fractured sandstone aquifer or affect flow in nearby streams (Department of Environment, Labour and Justice, 2014). Groundwater is the only source of drinking water in the province, and groundwater discharge (baseflow) to streams is important for aquatic ecosystems. There is little detailed hydrogeological data to support either the use of high capacity wells or the moratorium. There are predicative models available that can simulate the effects of pumping wells in unfractured porous aquifers on rivers and streams, but there are very few field studies confirming their results, making the
applicability and accuracy of such models to fractured sandstone aquifers virtually unknown.

1.1 Purpose and Scope

The goal of this research is to advance the understanding and methods for assessing and predicting the impact of groundwater extraction (by pumping wells) on the flow conditions in small streams. Pumping-induced stream flow depletion is defined as the reduction of stream discharge that results from induced infiltration of stream water into the connected aquifer, and/or interception of groundwater that would have otherwise discharged (i.e. baseflow) to the stream. With heightened awareness and pressure in many jurisdictions to better manage the surface water impacts of groundwater extraction, and the increasing evidence of the negative effects of climate change, there is a need for research into the relationship between groundwater extraction and flow conditions in small streams. However, in order to properly assess groundwater and surface water interactions, the degree of connection between the two needs to be well understood.

The focus of this research was to conduct a preliminary investigation near a small stream in central PEI, and to determine the suitability of the area for subsequent, more detailed, stream depletion studies. The objectives of this research were as follows:

- Determine whether the stream was gaining or losing water
- Assess the hydraulic heads in the study area
- Characterize the overburden materials
This study was conducted from September, 2015 to March, 2017, with the field data collection taking place between July and December of 2016.

2 LITERATURE REVIEW

The hydraulic effects of pumping groundwater from an aquifer have been well documented (e.g. Boulton, 1963; Theis, 1940) and can be modelled provided that the aquifer parameters are known. These parameters are most commonly obtained through the analysis of aquifer pumping test data. It is also known that having a pumping well close to a surface water body can induce flow from the water body into the aquifer and intercept groundwater that would otherwise enter the water body (e.g. Theis, 1941; Hantush, 1965; Hunt, 1999). For smaller streams, the pumping of groundwater can significantly reduce the stream flow and thus impact aquatic life. However, the relationship between aquifer pumping and stream flow reduction is far from being completely understood or predictable.

Hunt et al. (2001), Kollet and Zlotnik (2003, 2007), Lough and Hunt (2006), and Nyholm et al. (2002), have investigated the relationship between groundwater pumping and stream flow depletion. In all of these studies field data were collected through aquifer pumping tests, collecting drawdown data from monitoring wells or piezometers that surrounded a single pumping well. The stream discharge was also measured in some cases to help confirm the predicted stream flow depletion.

2.1 Well Pumping-Stream Interaction Studies

Hunt et al. (2001) used the Hunt (1999) conceptual model (Figure 1) and associated analytical solution to calculate the aquifer and streambed parameters using field data
collected from an aquifer pumping test near Doyleston Drain (a human-made ditch with a history of depletion due to aquifer pumping), 40 km south of Christchurch, New Zealand. The subsequent analysis lead to the estimation of the aquifer transmissivity and storativity, $T$ and $S$ respectively, and streambed conductance, $\lambda$:

$$\lambda = \frac{W_s K_s}{B_s} \quad [1]$$

where, $W_s$ is the average stream width in metres, $K_s$ is the vertical saturated hydraulic conductivity of the streambed in metres, and $B_s$ is the saturated thickness of the streambed in metres.

Lough and Hunt (2006) repeated the pumping test at the same location as Hunt et al. (2001) and found that the monitoring wells did not sufficiently penetrate the aquifer to yield reliable drawdown data. They also suspected that the aquifer was semiconfined, making the Hunt (1999) model inaccurate for this specific site. Thus the Hunt (2003) model
was used in its place, which has two additional parameters, specific yield of the aquitard, $S_y$, and the ratio of aquitard hydraulic conductivity and thickness, $K'/B'$, to allow for vertical leakage between confining units that contain the water table. The hydraulic conductivity of these units must be significantly smaller than that of the pumped aquifer, so that the assumption of vertical flow is valid, or the estimates of stream depletion could be affected. Finally, Lough and Hunt (2006) demonstrated that stream depletion is very sensitive to $\lambda$, and as a result, it directly controls the stream depletion rate.

Kollet and Zlotnik (2003) used two separate models, which used aquifer drawdown and stream depletion data collected from a pumping test near a natural meandering stream: the Hunt (1999) model, and a homogeneous piecewise model (BZT). They were unsuccessful in characterizing the relationship between stream depletion and aquifer pumping. They name five key factors that could have rendered the models invalid due to violation of the underlying assumptions: streambed conductance, degree of aquifer penetration by the stream, horizontal groundwater flow (Dupuit assumption), uniformity of the aquifer, and degree of aquifer penetration by the pumping well. These factors played a larger part in Kollet and Zlotnik’s experiment because it was conducted on a natural stream and not an engineered ditch, like that of Hunt et al. (2001), and Lough and Hunt (2006). They also express concern that having only one variable to describe the connection between the stream and the aquifer, $\lambda$, is insufficient and that it is more likely an empirical constant that accounts for complex three-dimensional flow, anisotropy, etc., making equation [1] inaccurate.
In 2007, Kollet and Zlotnik conducted a similar experiment to that conducted in 2003, but under different stream flow conditions in an unconfined aquifer. They also examined water temperature data to help quantify stream depletion, which the other studies did not. They found good agreement between the aquifer parameters obtained through analysis with and without the presence of the stream, which validates the use of the stream approximation in the two-dimensional models. They also found that the streambed conductance was not proportional to the stream width, which is important when the stream width changes under different discharge conditions.

Nyholm et al. (2002) created two numerical models for their aquifer-stream interaction study: model A used steady-state conditions without the influence of pumping and model B was calibrated with field data from a pumping test. They found that each model fit the calibration data very well, but could not accurately predict the other model’s data set. When comparing the measured results to the Hunt (1999) model, they found that it overestimated the stream depletion, possibly due to the underlying assumptions not being fulfilled.

The studies mentioned above were all conducted in alluvial sediments, which differ from the fractured sandstone aquifer of this study (i.e. the PEI aquifer). This introduces another factor that could affect the results of a pumping test: secondary porosity. However, if the fracture network is sufficiently developed with good interconnection, it is possible that the aquifer will respond to pumping like a porous medium (Francis, 1989).
2.2 Hydraulic Connection

A very common assumption in regards to modelling of aquifer-stream interactions is that the stream is hydraulically connected to the aquifer via a fully-saturated zone. Each of the studies reviewed above made this assumption but did not discuss its reasonability or the implications of its violation. Brunner et al. (2011) summarized what little knowledge and research has been done on hydraulically disconnected water bodies, and provided a conceptual model of the typical stages of connection between groundwater and surface water (Figure 2).

Figure 2 – Conceptual model of the various types of connection between groundwater and surface water (Brunner et al., 2011).
A water body that is connected to an underlying aquifer can gain or lose water, depending on the direction of the hydraulic gradient. On the other hand, for a disconnected water body, an unsaturated zone exists between it and the aquifer. Brunner et al. (2011) mention that the term “disconnected” does not have a clear definition, and that transitional states of connection also exist between connected and disconnected. They also mention that a disconnected surface water body will not lose water at a constant rate, but will trend towards a maximum flux asymptotically with increasing water table depth. The flux of water that is lost from the stream is dependent on the hydraulic conductivity, thickness, and hydraulic gradient, across the streambed. Additionally, some portions beneath the streambed may be unsaturated, while others remain saturated, and this relationship will change with location. Given the complexity, Brunner et al. (2011) define a surface water body to be disconnected if there is an unsaturated zone beneath the streambed and a measureable change in the infiltration flux does not occur with a change in the depth to water table.

It is clear from this discussion that groundwater and surface water interaction modelling must take the type and degree of connection into account. The simplistic streambed conductance term, which is commonly used in groundwater and surface water interaction modelling, may be insufficient in characterizing stream depletion due to groundwater abstraction.

3 BACKGROUND

The province of PEI has a population of 146,447 (Prince Edward Island Statistics Bureau, 2015), and is 5750 km² in size. About 40% of this area is agricultural land, and
about 20% has potatoes in the production rotation (Jiang and Somers, 2008). PEI receives roughly 1100 mm of precipitation each year, approximately 370 mm/year of which is aquifer recharge, accounting for 34% of the total precipitation. Province wide only 25.9 mm, or 7%, of the yearly recharge is currently being extracted. The current water extraction permitting policy states that groundwater extraction should not reduce the mean summer baseflow in the main branch of streams by more than 35%, and that 70% of the monthly median stream discharge must be maintained (Department of Environment, Labour and Justice, 2014). In other words, 35% of the yearly aquifer recharge is available for extraction, such that 70% of the monthly median stream discharge is maintained.

The bedrock in PEI is primarily a red sandstone formation, with the top fractured portion creating the Island’s primary aquifer. Hydraulic conductivity is typically very low around a depth of 160 m (Jiang et al. 2004) and decreases by an order of magnitude every 60 m (Francis, 1989). This is due to the decrease in porosity and degree of fracturing with depth. The porosity of the aquifer is approximately 0.20, the specific yield is estimated at 0.10, and the hydraulic conductivity ranges between $10^{-3}$ to $10^{-7}$ m/s, from high to low elevation (Francis, 1989; Rivard et al., 2008). Such a porous bedrock matrix could exhibit a delayed yield response due to pumping, but Francis (1989) and Rivard et al. (2008) analyzed drawdown data from pumping tests in the area and found that the response was similar to the Theis ideal response curve. Thus, the fracture network is likely very well developed, with many interconnections. Due to the horizontal bedding of the sandstone, and the horizontal nature of the fractures, the vertical hydraulic conductivity ranges between one and three orders of magnitude less than that of the horizontal component (Francis, 1989).
A thin layer of glacial till covers roughly 75% of the Island, ranging on average between one and five metres in thickness (Jiang et al., 2004). This till is mostly sandy and originates from the sandstone bedrock, with some silt, little clay, and a highly variable portion of gravel (Francis, 1989). The porosity ranges between 0.20 and 0.35, with an average hydraulic conductivity of $10^{-7}$ m/s (Francis, 1989).

The regional water table generally follows topography (Jiang et al., 2004), falling typically within the bedrock aquifer at higher elevations, and within the till near streams at lower elevations (Francis, 1989). There are two major recharge events throughout the year, in April with the spring snowmelt and freshet, and a smaller event around October with the increased precipitation and relatively low evapotranspiration (Jiang et al., 2004). This causes water table fluctuations anywhere from one to four metres throughout the year (Francis, 1989). During the dry summer months, baseflow accounts for approximately 80% of stream flow, with a typical streambed having a vertical hydraulic conductivity of $2.8 \times 10^{-5}$ m/s and an average thickness between 1 to 1.5 metres (parameters determined for Winter River, Prince Edward Island; Francis, 1989).

### 3.1 Study Area Description

A significant portion of this study involved the selection of a new field site. After discussions with technical staff at the PEI Department of Communities, Land and Environment, and Agriculture and Agri-Food Canada, a site was identified in central PEI near Maple Plains, approximately 20 km south-east of the Town of Summerside. The land use in this region is predominantly agricultural, including potato production. A small, accessible, stream reach was selected for further investigation; the stream is 1 to 2 m wide,
with an estimated discharge of 10 to 20 L/s. The stream is a tributary of Southwest Brook, which eventually flows into the Dunk River. The stream lies in a small, shallow sloping valley, in an area with no existing high capacity wells or other factors that may alter the natural hydraulic conditions. Furthermore, the thickness of the overburden was thought to be representative of the average thickness in PEI. The study area location is shown in Figure 3.

![Figure 3 – Maple Plains study area location in central PEI.](image)

Site investigations and preliminary field work began in July 2016, after obtaining permission to access the site from the property owner (Jason Webster) and the East Prince Agri-Environment Association. A wetland and unused residential well (Webster well) exist
roughly 200 m to the south-east of the property, as shown on Figure 4. The stream flows from east to west, passing through a roadway culvert before entering the Webster property, and trends along a relatively straight line for roughly 200 m before beginning to change direction to the south-west. A photo of the stream is shown in Figure 5.

Figure 4 – Maple Plains study site showing the stream location, small wetland, and unused well. Ground elevation contours (in metres) are shown by the brown lines, property boundaries in yellow, and the stream in blue.
Figure 5 – A photo of the stream located on the Webster property, facing upstream.

4 METHODS

The discharge was measured at 5 locations along a 1.6 km section of the stream, over a 2-day period, to determine whether the stream was gaining or losing water. Five drive-point piezometers were installed in the streambed and two groundwater monitoring wells were installed near the stream, which were used to assess the hydraulic gradients at the site. To characterize the overburden, soil samples collected during monitoring well installation were tested for moisture content and grain size distribution.
4.1 Stream Discharge Measurements (Seepage Run)

A seepage run was conducted between September 14 and 15, 2016, to identify whether the stream reach near the study site was gaining water from, or losing water to, the groundwater system. Previous researchers, such as Simonds et al. (2002), have used seepage runs (multiple measurements of stream discharge along a stream reach) to provide an estimate of the volume of water exchanged between surface water and groundwater, given stable hydrologic conditions.

During the seepage run two handheld acoustic Doppler velocity instruments (FlowTracker, SonTek) were used to estimate the stream discharge. The instruments were used to determine the velocity and depth at several locations (typically spaced every 5 cm, or 20 locations total) across the stream section, and these data were used to compute the stream discharge. Two discharge measurements were obtained by two operators at each location so that an estimate of reproducibility could be made.

A stationary bottom-mounted depth-velocity instrument (StingRay 2.0, Greyline Instruments) was also deployed during the seepage run. It was placed in the center of the stream cross section just upstream of each discharge measurement location. Unlike the multiple velocity-area measurements obtained with the handheld acoustic Doppler velocity instruments, the stationary instrument reports a single water depth and velocity.

The seepage run consisted of five measurement locations, spaced along an approximate stream length of 1.6 km (Figure 6). There were no surface water tributaries or diversions to the stream in this section. There is a roadside ditch that could allow surface
runoff to enter the stream; however, it was dry during the entirety of the seepage run. Measurements were taken at two locations (SS-DP1 and SS1) on September 14, but additional measurements were canceled due to 9.8 mm of precipitation (recorded at the Summerside airport) that started that afternoon. Measurements were made at all five locations on September 15, under the assumption that the stream would have recovered to steady-state overnight. An additional measurement was taken at SS2 in the evening of September 15 to verify this assumption.

Figure 6 – Stream discharge measurement locations for the September 2016 seepage run.

The stream flow direction is from east to west (SS4 toward SS1).
4.2 Hydraulic Heads

Two “deep” drive-point piezometers were installed to depths of between 1.3 m and 1.5 m in the streambed sediments on July 28, and three “shallow” drive-points were installed to depths of about 1 m on August 22. The drive-point tips were made of stainless steel, measuring 38.1 mm in diameter. A roughly 18 cm section of screen follows the tip, which is in turn attached to the casing sections. The casing sections are made of steel and have lengths varying from 60 to 80 cm. The drive-points were installed via a jackhammer (deep class) or sledge hammer (shallow class). Figure 7 is a photo of a typical drive-point tip.

Figure 7 – Typical drive-point piezometer tip.
The drive-point piezometers were installed in the streambed at three locations along the stream (i.e. DP1, DP3, and DP5). Refer to Figure 8 for their locations. Both of the deep drive-points (i.e. DP1D and DP3D) were found to have very little to no water using a manual water level tape (Model 101 Water Level Meter, Solinst). Stream water was intentionally added to DP1D to evaluate whether the screen may have been clogged with fine material during installation. Water was not added to DP3D to determine if water would naturally enter the piezometer. The three shallower drive-points (i.e. DP1S, DP3S, and DP5S) were installed to evaluate whether the deep drive-points had passed into, or below, some low permeability layer and reached an unsaturated zone. The details of the drive-point piezometers are summarized in Table 1.
Figure 8 – Location of the drive-point (DP) piezometers and groundwater monitoring wells (MW) at the Maple Plains study site. The stream flow direction is from east to west.
Table 1 – Drive-point piezometer screen depths and stick-up (relative to the streambed).

<table>
<thead>
<tr>
<th>Drive Point ID</th>
<th>Mid-Screen Depth (m)</th>
<th>Stick-Up (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DP1S</td>
<td>1.000</td>
<td>1.020</td>
</tr>
<tr>
<td>DP1D</td>
<td>1.315</td>
<td>1.200</td>
</tr>
<tr>
<td>DP3S</td>
<td>1.000</td>
<td>0.593</td>
</tr>
<tr>
<td>DP3D</td>
<td>1.541</td>
<td>0.907</td>
</tr>
<tr>
<td>DP5S</td>
<td>1.000</td>
<td>1.184</td>
</tr>
</tbody>
</table>

Three mini-piezometers were later installed in the streambed to shallower depths (about 0.5 m) than the drive-points. The mini-piezometers consisted of a length of polyethylene tubing (roughly 25 mm in diameter) attached to a similar tip to that of the drive-point piezometers. The bottom portion of the tubing was punctured and wrapped with a nylon fabric to create a screen. The tubing was passed through the same casing used for the drive-points, until the tip was positioned firmly against the casing. A sledge hammer was then used on the casing to drive the tip into the streambed. Once at the desired depth, the casing was removed, exposing the screen.

On September 15, 2016, two standard groundwater monitoring wells (50 mm diameter PVC screens and casing) were installed using a conventional drilling rig at distances of about 15 m from the stream (Figure 8). MW1 was installed midway between DP1 and DP3 to a depth of 7.3 m, and MW2 was installed close to DP5 to a depth of 7.2 m. The wells were installed using a hollow stem auger and sampled by split spoon every 1.5 m, until refusal. The monitoring well PVC pipe was then lowered into the holes, with a 1.5 m screen length at the bottom, and backfilled with a clean sand surrounding the screen.
and a mixture of bentonite chips and drill cuttings to surface. Diagrams of both monitoring wells with approximate dimensions are shown in Figure 9.

Figure 9 – A diagram of the MW1 and MW2 monitoring wells. mbgs indicates metres below ground surface. Water levels correspond to measurements made on September 15, 2016.

Pressure transducers/data loggers (Levelogger, Solinst) were used to collect water elevation data, which included the use of a pressure transducer (Barologger, Solinst) at the surface to compensate for changes in barometric pressure. A manual water level tape (Model 101 Water Level Meter, Solinst) was used to collect manual water elevation readings to calibrate and confirm the pressure transducer data. Pressure transducers were not installed in the three mini-piezometers as the casing diameter was too small; however,
the water elevation in the mini-piezometers was checked periodically with a manual water level tape. Pressure transducers were also installed in the Webster well, the stream near DP1, and the wetland (Figure 8). The Webster well (127 mm diameter) was probed with a manual water level meter until resistance was met, at a depth of roughly 27 m from the top of casing (TOC), though no other details are known (i.e. casing depth, and drilled depth).

4.3 Characterization of the Overburden Materials

A total of eight sediment samples were collected during the drilling of the borehole for the installation of MW1. Only two samples were collected from MW2, as the material at both locations was visually similar. Moisture content and grain size analyses were conducted on the eight samples from MW1, according to ASTM D422. Any material retained on the #200 (0.075 mm) sieve was classified according to the Unified Soil Classification System (USCS), and any material that passed the #200 was classified according to the Canadian Foundation Engineering Manual (CFEM).

After installation of the three shallow drive-point piezometers, all five drive-point piezometers were filled with stream water so that a falling-head test could be performed, and trends in the hydraulic conductivity of the overburden could be qualitatively assessed.

5 RESULTS

5.1 Stream Discharge Measurements (Seepage Run)

The seepage run discharge results and associated uncertainties, as provided in the instrument data report, are shown in Table 2. The discharge results obtained with the acoustic Doppler velocity instruments were used to calculate a percent difference relative
to the mean discharge, and to evaluate consistency. The two measurements at each location were considered consistent if the absolute value of their difference was less than the sum of their absolute uncertainties. Key data from each stream discharge measurement are provided in Appendix A.

The results obtained with the stationary depth-velocity acoustic instrument (Stingray 2.0, Greyline Instruments) differed significantly from the results obtained with handheld instruments (Table 2). This is attributed to the fact that the StingRay 2.0 takes the measurement of a single water velocity that is subsequently used to compute discharge. For this reason, the results obtained with the StingRay 2.0 were not used to assess whether the stream was gaining or losing.

Table 2 – Results of seepage run conducted on the study-area stream, presented in chronological order. The stream discharge, Q, measurement locations are shown in Figure 6.

<table>
<thead>
<tr>
<th>Date</th>
<th>Location</th>
<th>2D FlowTracker Q (L/s)</th>
<th>3D FlowTracker Q (L/s)</th>
<th>Consistency</th>
<th>% difference</th>
<th>Stingray Q (L/s)</th>
<th>% difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>September 14th</td>
<td>SS-DP1</td>
<td>3.9</td>
<td>1.4</td>
<td>not</td>
<td>94.3%</td>
<td>5.5</td>
<td>99.8%</td>
</tr>
<tr>
<td></td>
<td>SS1</td>
<td>3.2</td>
<td>2.4</td>
<td>not</td>
<td>28.6%</td>
<td>3.1</td>
<td>99.9%</td>
</tr>
<tr>
<td>September 15th</td>
<td>SS2</td>
<td>6.5</td>
<td>7.8</td>
<td>not</td>
<td>18.2%</td>
<td>4.4</td>
<td>99.9%</td>
</tr>
<tr>
<td></td>
<td>SS-DP1</td>
<td>4.6</td>
<td>3.7</td>
<td>consistent</td>
<td>21.7%</td>
<td>13.4</td>
<td>99.7%</td>
</tr>
<tr>
<td></td>
<td>SS1</td>
<td>4.8</td>
<td>5.3</td>
<td>consistent</td>
<td>9.9%</td>
<td>2.5</td>
<td>100.0%</td>
</tr>
<tr>
<td></td>
<td>SS3</td>
<td>5.3</td>
<td>5.9</td>
<td>consistent</td>
<td>10.7%</td>
<td>1.9</td>
<td>100.0%</td>
</tr>
<tr>
<td></td>
<td>SS4</td>
<td>5</td>
<td>3.8</td>
<td>not</td>
<td>27.3%</td>
<td>9.6</td>
<td>99.8%</td>
</tr>
<tr>
<td>September 15th (evening)</td>
<td>SS2</td>
<td>5.3</td>
<td>5.2</td>
<td>consistent</td>
<td>1.9%</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

During the seepage run the water depths in the stream were approximately 6 cm to 13 cm, which is close to the suggested minimum (> 8 cm, FlowTracker, SonTek), so there were relatively high uncertainties (> 5%), high percent differences, and pairs of inconsistent measurements. The average discharge on September 14 was 2.7 L/s (n = 4), compared to 5.3 L/s (n = 10) on September 15 which suggests that the precipitation that
occurred late on September 14 affected the stream discharge measurements made the following day. The discharge results are shown graphically in Figure 10.

Figure 10 – Stream discharge measured along the study reach. Stream flow is from left to right, with 0 m being the most upstream measurement location. 2D and 3D indicate the two different models of acoustic Doppler velocity instruments, while 14th, 15th, and 15th_evening are the dates in September, 2016. The error bars represent the uncertainty associated with each measurement.

5.2 Hydraulic Heads

All the hydraulic head (water elevation) data collected during this study are presented in Figure 11. All water elevation data were referenced to an assumed top of
casing elevation of 41.0 m for the Webster well (see Appendix C for the elevation survey results). Appendix B contains the manual water level measurements.

The highest hydraulic head was consistently measured in the small wetland, which contained ponded surface water, located approximately 100 m to the south of the stream (Figure 8). The stream water elevation measured at DP1 was about 2.5 metres lower than the water elevation in the wetland. All groundwater hydraulic heads, either from wells or drive-point piezometers, were lower than surface water levels for the duration of the monitoring period.

![Water Elevation at Maple Plains Site](image)

Figure 11 – Observed groundwater and surface water elevations. Manual water level readings are shown by symbols. Precipitation data are the daily totals reported for the Summerside airport.
After approximately two weeks, the water that had been added to DP1D had fallen 102 cm and was below the transducer, signifying that the screen was not clogged. Also during this same time period, no water had risen above the transducer in DP3D and manual readings showed that if water was present, that it was near the base of the piezometer screen. The data from DP1D and DP3D were interpreted to mean that the surrounding material was unsaturated. The three shallow (1.0 m deep) drive-point piezometers exhibited very similar behavior in that no standing water (above the pressure transducers) was detected within 14 weeks of installation.

The data recorded by the pressure transducers between August 22 and December 23 showed that the groundwater elevation increased in late November by roughly 50 cm in the Webster well, and by roughly 30 cm in the two monitoring wells (Figure 11). In late November, water also rose above the pressure transducers in two (DP3S and DP5S) of the shallow drive-point piezometers for the first time during the monitoring period.

All mini-piezometers consistently produced water levels that were very near the elevation of the stream water, suggesting that the piezometer screens may not have been hydraulically isolated from the stream. Water may have seeped into the hole created by installing the mini-piezometer and for this reason the mini-piezometer data are not reported.
5.3 Characterization of the Overburden Materials

Grain size analyses were conducted on the eight samples collected during the installation of MW1. The results of this testing are summarized in Table 3 and the grain size distribution curves are shown in Figure 12.

Table 3 – Summary of soil test results for samples taken from borehole MW1.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Depth (m)</th>
<th>Gravel (%)</th>
<th>Sand (%)</th>
<th>Silt (%)</th>
<th>Clay (%)</th>
<th>Moisture Content (%)</th>
<th>Material</th>
</tr>
</thead>
<tbody>
<tr>
<td>SPT1</td>
<td>0.6 to 1.2</td>
<td>9.6</td>
<td>41.4</td>
<td>36.5</td>
<td>12.5</td>
<td>11.9</td>
<td>Till</td>
</tr>
<tr>
<td>SPT2</td>
<td>1.2 to 1.8</td>
<td>16.8</td>
<td>39.5</td>
<td>31.8</td>
<td>11.9</td>
<td>13.3</td>
<td>Till</td>
</tr>
<tr>
<td>SPT3</td>
<td>1.8 to 2.4</td>
<td>5.5</td>
<td>48.9</td>
<td>32.3</td>
<td>13.3</td>
<td>14</td>
<td>Till</td>
</tr>
<tr>
<td>SPT4</td>
<td>2.4 to 3.0</td>
<td>33.3</td>
<td>51.2</td>
<td>13.2</td>
<td>2</td>
<td>9.2</td>
<td>Sandstone</td>
</tr>
<tr>
<td>SPT5</td>
<td>3.0 to 3.7</td>
<td>30.4</td>
<td>49.4</td>
<td>16.8</td>
<td>3.4</td>
<td>10.6</td>
<td>Sandstone</td>
</tr>
<tr>
<td>SPT6</td>
<td>3.7 to 4.3</td>
<td>32.7</td>
<td>50</td>
<td>14.6</td>
<td>2.7</td>
<td>11.3</td>
<td>Sandstone</td>
</tr>
<tr>
<td>SPT7</td>
<td>4.3 to 4.9</td>
<td>21.3</td>
<td>53.2</td>
<td>21.2</td>
<td>4.2</td>
<td>12.4</td>
<td>Sandstone</td>
</tr>
<tr>
<td>SPT8</td>
<td>4.9 to 5.5</td>
<td>30.6</td>
<td>41.4</td>
<td>23.8</td>
<td>4.3</td>
<td>13.1</td>
<td>Sandstone</td>
</tr>
</tbody>
</table>

Figure 12 – Grain size distribution curves of samples collected during the installation of MW1.
The results of the soil testing indicate an increase in gravel content and decrease in both silt and clay content at a depth of 2.4 m below ground surface. The three samples above 2.4 m have very similar grain size distributions, and the same can be said of the five samples below 2.4 m. The moisture content also changed abruptly at 2.4 m, dropping by roughly 5%.

To qualitatively evaluate trends in the hydraulic conductivity of the overburden, each drive-point piezometer was filled with water, and the water level decline was measured with time. Although a conventional falling-head slug test analysis (Fetter, 2001) cannot be conducted using these data because the soil surrounding the piezometer screens was unsaturated, the data were normalized by the height of water ($H_o$) initially added (Figure 13).
Figure 13 – Normalized head above mid-screen versus time for all drive-point piezometers.

The head in drive-point DP3D dropped to about $H/H_0 = 0.5$ within the first day, before the rate of decline changed to something more similar to that of the other locations. This behavior could have been caused by a lose connection between lengths of casing allowing water to drain from the piezometer more quickly, so this portion of the data was removed and the initial change in head was adjusted. As a result, the adjusted slope of DP3D is in the same range as DP1S and DP3S, which seem to be screened in very similar material (refer to Appendix C for the corrected plot). It is also noted that the rate of water level decline (i.e. slope of $\log(H/H_0)$ versus time, Figure 13) increases near the end of each
curve. This is thought to occur when the water surface reaches the pressure transducer, causing the head to fall more quickly in this interval of reduced cross sectional area. This hypothesis is supported by the data, in that the transition occurs in a roughly 15 cm interval (the length of a Levelogger), followed by an indication that the transducer is above the water surface.

6 DISCUSSION

The data collected during the seepage run suggest that the stream discharge increased by about 2.6 L/s between September 14 and 15, after a brief precipitation event, and remained stable on September 15. The average discharge for all measurement locations was used for this comparison because of the high percent difference and inconsistencies measured in the stream discharge data. Any error associated with the discharges may be larger than the actual difference in discharge between locations.

Based on the data collected on September 15 (Figure 10), a period of the year when stream discharge would normally be expected to be dominated by groundwater baseflow, there did not appear to be any significant changes in discharge along the 1.6 km reach of stream investigated. This indicates that the stream was not gaining significant amounts water from the bedrock aquifer through a saturated zone beneath the stream, or that the hydraulic conductivity of the stream bed sediments is low enough to create a relatively impermeable layer between the stream and aquifer.

Based on the hydraulic heads observed in the two monitoring wells screened in the aquifer, and the unsaturated conditions observed in the streambed drive-point piezometers,
the stream and wetland both appear to be perched, with an unsaturated zone separating the surface and groundwater systems. A shallow perched groundwater system may also develop within the overburden till during periods of increased infiltration, as evidenced by the water levels in DP3S and DP5S in late November 2016. During this period, however, the water elevations in the shallow drive-point piezometers remained below the stream water elevation and about 2.5 metres above the hydraulic heads in the aquifer (Figure 11). During this period the downward vertical hydraulic gradient between the stream and DP5S was about 0.6, and between DP5S and MW2 it was also about 0.6.

Figure 14 shows a plan view of the site with hydraulic head contours drawn for the aquifer and the perched system, and stream elevations in December 2016. The groundwater elevation data (Figure 11) show that there was a horizontal gradient in the aquifer from the Webster well north-west toward MW1 and MW2 at all times during the monitoring period. There was also a horizontal gradient observed that would suggest flow from DP3S towards DP5S, when a perched system was present. Both of these gradients are consistent with the direction of flow in the stream. The data produce a horizontal hydraulic gradient of 0.007 in the aquifer, 0.01 in the perched groundwater system, and a slope of 0.005 for the stream.
Figure 14 – Plan view of the Maple Plains study site with elevation contours for the aquifer (green) and perched (red) groundwater systems. The stream elevation is indicated by the results in blue font.

The identification of an unsaturated zone between the stream and aquifer suggests that the stream is disconnected from the aquifer. However, as discussed by Brunner et al. (2011), identifying that a stream is disconnected requires that there be no change in flux for a change in the water table depth, which can be difficult to prove. An aquifer pumping test with a pumping well close to the stream, that produces a measurable lowering of the water table beneath the stream, would be required to prove disconnection. It would also be necessary to accurately measure the stream discharge, and to conduct this pumping test during a period of relatively stable discharge and groundwater elevation. An extra challenge here is that the stream discharge may vary on a daily basis, and there are
generally large errors associated with stream discharge measurements, which may be larger than any change due to stream depletion.

While it is possible that this stream is disconnected from the aquifer, this does not mean that the two systems are separate. Brunner et al. (2011) note that groundwater abstraction can increase the length of the disconnected portion of the stream, which affects stream discharge. Conversely, infiltration flux will change with a change in the depth and width of the stream, which will affect the groundwater system. Therefore, groundwater and surface water should be jointly managed, since the presence of an unsaturated zone between a stream and water table does not indicate that the systems behave independently.

The results of soil testing on eight samples collected during the installation of MW1 reveal a sudden drop in moisture content, and silt and clay percentage, below 2.4 m. This is interpreted to represent the change from the till overburden to the sandstone bedrock. The grain size analysis results for the overburden agree with the sand phase till description of Prest (1973). The high percentage of gravel in the sandstone samples was caused by the presence of sandstone clasts that had not been pulverized during driving of the split spoon sampler. The decrease in moisture content at 2.4 m could be caused by an increase in porosity that sandstone would provide. If this change in moisture content was removed, the moisture with depth would increase at a relatively constant rate. No samples were obtained below the water table, as the sandstone became too competent to sample by split spoon. The fact that the upper sandstone could be sampled by split spoon at all is indicative of significant weathering near the top of the formation.
The data presented in Figure 13 suggest that the hydraulic conductivity of the streambed is heterogeneous, due to the varying slopes at each location. The shallower slopes found in the deep drive-point data show evidence that K decreases with depth. It is unlikely that any of the drive-points were screened in the sandstone, as it would be difficult to penetrate the sandstone, and it is expected that the drainage times would be smaller by an order of magnitude or more. Therefore, it is thought that the till is generally more weathered near the surface.

As a result of sediment sampling during the monitoring well installations, it was found that a fractured bedrock aquifer exists below a till overburden. An unsaturated zone exists within the overburden and shallow bedrock. The absence of groundwater during the summer months in the drive-point piezometers that were installed beneath the stream bed indicates that the overburden will permit seepage into the bedrock, but does not hold water during dry periods of the year. During periods of significant infiltration, a perched groundwater system develops. This is thought to occur as a result of lower hydraulic conductivity near the bottom of the till. The presence of water in some drive-points and its absence in others, suggests that the hydraulic conductivity and thickness of the till is heterogeneous. The presence of water in DP3S while DP3D was dry, is thought to be due to DP3D being screened within the low hydraulic conductivity layer. Figure 15 presents an initial conceptual model of the site.
The stream water travels along the top of the overburden, so the streambed must also act as a relatively low hydraulic conductivity layer, otherwise water from the stream would be constantly feeding a perched system below the stream, which was not observed. A streambed with a low hydraulic conductivity could have developed as a result of natural depositional processes.

7 CONCLUSION

The goal of this study was to assess the degree of connection between groundwater and surface water in a small stream typical of central PEI. The main objectives of this research were as follows:

- Determine whether the stream was gaining or losing water
- Assess the natural hydraulic gradients
- Characterize the overburden materials
Investigations at the study site showed about 2 to 3 m of sandy glacial till overburden overlying the fractured sandstone bedrock. The hydraulic conductivity of the overburden was not measured directly, but based on water level declines in drive-point piezometers is likely heterogeneous and decreasing with depth. The bottom portion of the till overburden may have sufficiently low hydraulic conductivity to allow a shallow perched groundwater system to form during periods of increased infiltration. However, during low recharge (summer) periods no perched groundwater system was observed.

The results of a seepage run conducted along a 1.6 km section of the stream in September 2016 suggested that the stream was not gaining nor losing water. Using water elevation data collected from the stream, a nearby existing residential well, drive-point piezometers installed beneath the streambed, and monitoring wells screened in the aquifer, it was determined that an unsaturated zone of approximately 3 m exists between the stream bed and the water table. The identification of an unsaturated zone beneath the stream indicates a downward vertical hydraulic gradient, and thus losing conditions. Identification of an unsaturated zone does not automatically mean that the stream is disconnected from the aquifer, only that it is not connected. Disconnection requires proof that a change in the height of the water table does not induce a change to the flux of water from the stream to the aquifer. This proof would require additional work, such as conducting an aquifer pumping test during stable hydraulic conditions that would produce a measurable drawdown of the water table beneath the streambed, and monitoring of the stream discharge over this period.
Although a low degree of connection between the stream and aquifer may be interpreted to mean that the two systems do not interact, a small downward flux of water still exists from the stream. Pumping the aquifer and subsequently lowering its water table may not affect this flux in already disconnected sections of the stream, but can extend the length of these disconnected sections, increase the flux in sections that are not disconnected, or cause connected sections to change from gaining to losing.
REFERENCES


Appendix A – Key FlowTracker Data

![Discharge Measurement Summary](image)

Figure 16 - FlowTracker measurements from the 2D model, on September 14, at location SS1

(46°18'17.88" N, 63°35'10.62" W).
Figure 17 – FlowTracker measurements from the 3D model, on September 14, at location SS1

(46°18'17.88" N, 63°35'10.62" W).
Figure 18 – FlowTracker measurements from the 2D model, on September 14, at location SSDP1 (46°18′19.92″ N, 63°34′34.50″ W).
Figure 19 – FlowTracker measurements from the 3D model, on September 14, at location SSDP1
(46°18’19.92” N, 63°34’34.50” W).
Figure 20 – FlowTracker measurements from the 2D model, on September 15, at location SS1

(46°18'17.88" N, 63°35'10.62" W).
Figure 21 – FlowTracker measurements from the 3D model, on September 15, at location SS1

(46°18'17.88" N, 63°35'10.62" W).
Figure 22 – FlowTracker measurements from the 2D model, on September 15, at location SS2

(46°18'22.20" N, 63°34'9.18" W).
Figure 23 – FlowTracker measurements from the 3D model, on September 15, at location SS2

(46°18'22.20" N, 63°34'9.18" W).
Figure 24 – FlowTracker measurements from the 2D model, on September 15, at location SS3

(46°18'20.95" N, 63°34'47.03" W).
Figure 25 – FlowTracker measurements from the 3D model, on September 15, at location SS3

(46°18'20.95" N, 63°34'47.03" W).
Figure 26 – FlowTracker measurements from the 2D model, on September 15, at location SS4

(46°18'25.57" N, 63°33'57.00" W).
Figure 27 – FlowTracker measurements from the 3D model, on September 15, at location SS4

(46°18'25.57” N, 63°33'57.00” W).
Figure 28 – FlowTracker measurements from the 2D model, on September 15, at location SSDP1

(46°18'19.92" N, 63°34'34.50" W).
Figure 29 – FlowTracker measurements from the 3D model, on September 15, at location SSDP1

(46°18'19.92" N, 63°34'34.50" W).
Figure 30 – FlowTracker measurements from the 2D model, on September 15 (evening), at location SS2 (46°18'22.20" N, 63°34'9.18" W).
Figure 31 – FlowTracker measurements from the 3D model, on September 15 (evening), at location SS2 (46°18'22.20" N, 63°34'9.18" W).
Appendix B – Manual Water Level Data, and GPS Coordinates

Table 4 – Manual water level data, in metres below reference point. Reference points and coordinates shown in Table 5.

<table>
<thead>
<tr>
<th>Date</th>
<th>Location ID</th>
<th>Stream (at DP1)</th>
<th>Wetland</th>
<th>Webster Well</th>
<th>MW1</th>
<th>MW2</th>
<th>DP1S</th>
<th>DP1D</th>
<th>DP3S</th>
<th>DP3D</th>
<th>DP5S</th>
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<td>7/28/2016 18:18</td>
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<td>1.284</td>
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Notes:
1 - Unreliable readings that were responding to wet silts.
2 - The casing was filled with water at this time.
3 - A reading could not be successfully detected.
4 - A temporary stake was used as a reference point.

Table 5 – GPS coordinates and reference locations.

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<th>Reference Location</th>
<th>Coordinates</th>
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<td>46° 18' 19.92 N 63° 34' 34.50 W</td>
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<td>DP3S</td>
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<td>46° 18' 20.82 N 63° 34' 39.12 W</td>
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<tr>
<td>DP5S</td>
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<td>MW2</td>
<td>top of PVC casing</td>
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## Appendix C – Elevation Survey

Table 6 – Results of elevation surveys conducted on the Maple Plains study site.

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Appendix D – Corrected Normalized Head Plot

Figure 32 – Normalized head above mid-screen versus time for all drive-point piezometers, with DP3D corrected to neglect data above a suspected leak in the casing.
CURRICULUM VITAE

Wesley Tibbet

University of New Brunswick, 2011-2015, B.Sc.E. - Civil Engineering