Determining and Comparing Deep Drainage between Two Years – Dry (2012) and Wet (2013) – using Darcy’s Law

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Abstract

Understanding of the downward flux of water below the plant root zone, known as deep drainage (DD), is significant in agriculture and soil water conservation. It plays a key role to determine the amount of water that travels below the plant root zone and can potentially cause groundwater recharge. The DD in soil varies with location, soil texture, and topography. Thus, the objectives of this study were to determine the unsaturated hydraulic conductivity, soil water storage, and DD for the years 2012 (dry year) and 2013 (wet year) at the University of Guelph’s Arboretum. The depths to the water table data were collected using a Mini Water Level Meter. CS616 sensors were used to determine the soil volumetric water content. The soil temperature was extracted with the use of T107 Temperature Probes. The slug test, based on the Hvorslev method, was performed to determine the field saturated hydraulic conductivity. The soil moisture retention curve was produced based on the data collected in the lab with the use of pressure plate systems, using van Genuchten’s equation. The unsaturated hydraulic conductivity was also determined using van Genuchten’s equation. Darcy’s law was used to determine the specific discharge, which was then converted to the total DD. In general, the soil water storage was 38.5 mm higher in 2013 relative to 2012. The unsaturated hydraulic conductivity was approximately 2 times higher in 2013 than 2012. The average DD was approximately 25 mm higher in 2013. This study provides information needed to better understand the movement and amount of water flux and DD in larger details.

Keywords: soil physical properties; soil moisture; water table depth; hydraulic conductivity; soil water storage; groundwater recharge; deep drainage

Introduction

Groundwater, which is defined as water stored below the earth’s surface, is an important aspect of the hydrological cycle (Hillel, 1998). Rivers, streams and precipitation play an important role in groundwater dynamics. However, to replenish the groundwater supply, deep drainage (DD) is generally required. The term DD is defined as the downward flux of water (volume of water per unit area per unit time) below the plant root zone (Bond, 1998). Estimation of DD is complex and time-consuming. Some of the reasons for the complexity of DD estimation are various soil textures, land relief, and the surrounding environment. There are many physical methods to determine the DD of water, such as lysimeters, zero-flux plane, and soil water balance; however, Darcy’s law was used in this study due to the simplicity of the equation and the data available. Relatively few studies have

been conducted to estimate the DD and potentially the groundwater recharge in Southern Ontario, providing a rationale for this study. Further reasoning for the study was to compare DD between a dry and wet year potentially representing future climatic conditions of southern Ontario.

Literature Review

Understanding of soil water storage and DD is lacking in Ontario, which presents a barrier in guiding effective development of policies requiring groundwater recharge. Numerous studies have been conducted to try to fill the knowledge gaps. For instance, a study by Black et al. (1969) predicted the evaporation, drainage and soil water

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storage for bare Plainfield sand. The study was undertaken under natural rainfall conditions, for a three month period (June-August) using lysimeters. The lysimeters were filled with subsoil and plow layer soil for the upper 25 cm (Black *et al.*, 1969). It was observed that in a uniform soil profile, there was equal drainage of water from all layers. If the drainage estimation is done before the wetting exceeds the soil profile depth (plant root zone) for which the drainage is being calculated, it can lead to overestimation of drainage (Black *et al.*, 1969). Thus, the time lag had to be considered. The process of redistribution of water to drain to the depth below the plant root zone took approximately 2 days (Black *et al.*, 1969). In this study, the estimated and measured soil water storage had a difference of about 0.3 cm. There was 13.2 cm of irrigation and precipitation, whereas the drainage was 10.6 cm and evaporation was 7.5 cm for the period of July 1-September 12. From June 30-September 12, the water storage had decreased from 16 cm to 11.1 cm (Black *et al.*, 1969).

Zebarth *et al.* (1988) studied the soil water flow in a hummocky landscape surrounding two sloughs in central Saskatchewan and estimated the rate of groundwater recharge. Summer fallow (F) and stubble sites were used for the duration of the study from 1985 to 1987. There were piezometer nests installed on the knoll, midslope, and the middle of the slough, which were at depths of 2.5, 5, 10 and 15 m, and were 5.5 m apart. The results were such that the lateral flow was twice the volume and occurred through a smaller area, when compared to the volume of vertical flow. Furthermore, the F site had similar recharge rate all year. This was due to the unsaturated sandy till present at depth, which resulted in consistent vertical hydraulic gradient (Zebarth *et al.*, 1988). Also, 10% of the mean annual precipitation contributed to the regional groundwater recharge rate. The number of sloughs and the strong vertical hydraulic gradient could have possibly resulted in such high recharge rate (Zebarth *et al.*, 1988). The results from this study confirm that hummocky upland areas are great contributors of groundwater recharge.

Parkin *et al.* (1999) estimated the seasonal and annual water surplus (runoff and drainage) in Ontario at four different geographic locations: Harrow, Guelph, Ottawa, and Kapuskasing, using a deterministic model: Simultaneous Heat and Water (SHAW). This model aids to determine the water flow in the soil, while considering the evapotranspiration, plant water uptake, and freeze and thaw conditions (Parkin *et al.*, 1999). All four sites used the SHAW model for daily climate data for a thirty-year period, December 1960-November 1990. Based on various inputs such as crop transpiration, precipitation, water storage and soil hydraulic properties, the output for mean annual evapotranspiration, runoff, and DD (below 1.25 m) were estimated (Parkin *et al.*, 1999).

In Parkin *et al.* (1999), on average, the results were such that there was some DD that had occurred every year in the summer at all four sites. However, run-off had only occurred in/for a few years, which could possibly be due to the

assumption made of 0.5% slope (Parkin *et al.*, 1999). Harrow had an average annual DD of 163 mm, Guelph had 283 mm, Ottawa had 201 mm, and Kapuskasing had 153 mm (Parkin *et al.*, 1999).

The highest DD, in Guelph, was possibly due to the loam soil, while the others were clay loam and silty clay loam. Thus, soil at Guelph had a greater hydraulic conductivity than the other sites, which led to higher DD through the soil profile and plant root zone (Parkin *et al.*, 1999). The lower hydraulic conductivity, however, led to higher runoff in Kapuskasing (206 mm), due to the lower water holding capacity of the soil (Parkin *et al.*, 1999). There was greater DD than runoff in the summer and fall, which could have been a result of flat slope assumed in the model (Parkin *et al.*, 1999). The amount of precipitation seemed to be the major factor that determined the water surplus, rather than the soil type, possibly because the soil hydraulic properties were similar for each site (Parkin *et al.*, 1999).

A study by Delin *et al.* (2000) examined the effects of surface topography and variations in soil properties on groundwater recharge at two sites in an agricultural field. The upland and lowland sites were located in a sand plain setting in Minnesota in a cropped field. The upland and lowland sites were approximately 78 m apart. Hydrograph analysis, chlorofluorocarbon age dating (CFC-12), and an unsaturated zone water balance (UZWB) were used for groundwater recharge estimation (Delin *et al.*, 2000). The lowland site had coarser textured soil at depths greater than about 1.5 m and there were lamellae present around the upland site (Delin *et al.*, 2000). All the methods used in the study led to recharge of 10-40% of natural precipitation, and larger rates of recharge for the lowland area in comparison to the upland site. The lowland annual recharge exceeded upland recharge by approximately 30% based on hydrograph analysis, 80% based on CFC-12 data, and 60% based on the UZWB method (Delin *et al.*, 2000).

McCoy *et al.* (2006) estimated the soil water budget by using automated soil water content measurements for 2001, 2002 and 2003 at a site near Elora, Ontario. The study developed a method to estimate the monthly drainage and runoff rate by the use of automatically collected precipitation, soil temperature, evapotranspiration and soil water storage measurements. The proposed method also compared the soil water budget differences in No-Tillage (NT) and Conventional Tillage (CT) treatments. Water content reflectometers (WCR) were used for soil water content measurements. These measurements were then converted to soil water storage, which were then used to estimate the monthly DD.

The 3-year average DD was approximately 202 mm (McCoy *et al.*, 2006). The results were such that there was higher run off and interception for NT treatment than CT. There was also higher amount of soil water storage for NT when compared to CT, but similar amounts of monthly drainage for both treatments, with 10% higher drainage for CT

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than NT treatment (McCoy *et al.*, 2006). This can be expected since the undisturbed soil pores (NT) have maintained their natural soil structure, thus resulting in higher soil water storage and lower DD. In the study, the average amount of evapotranspiration was used for both CT and NT treatments, thus it could have possibly overestimated the DD for CT and underestimated for NT treatments in the spring and fall (McCoy *et al.*, 2006).

Brown *et al.* (2013) studied the temporal and spatial variability in estimated water surplus in 12 different regions in Ontario, Canada under different soil profile conditions. A deterministic hydrologic model (DRAINMOD), which estimates not only evapotranspiration (ET), runoff (RO), infiltration, seepage, and subsurface drainage, but also freeze and thaw conditions, was used in the study for estimation of the water surplus (Brown *et al.*, 2013). The positive linear relationship between DD and precipitation seemed to get stronger as the soil texture became coarser (sandy) and weaker as the soil texture became finer (clay). This was most likely due to the higher water retaining capacity of clay soils, in comparison to sandy and loam soils, which led to lower chances of DD for a given precipitation event.

There was stronger potential for DD to occur in coarser-textured soil than finer-textured soil due to higher hydraulic conductivity (Brown *et al.*, 2013). Thus, the mean annual DD was higher than the RO rate at sites with loamy or sandy soil; however, the RO rate was higher than DD at sites with clay or silty clay soils (Brown *et al.*, 2013). The RO for sandy soils rarely exceeded 150 mm. The DD was higher than 300 mm (400 mm mostly) in many years for 7 of the 12 sites, and the precipitation was higher than 1000 mm at 4 of these 7 sites in majority of the years, when the soil was assumed to be sandy (Brown *et al.*, 2013).

Miller *et al.* (2014) measured the vertical unsaturated water flux below the plant root zone at mid and lower slope positions within an inclined landscape of dark brown soil zone in southern Alberta for 1985 and 1986. There were two sites for this study, one of which was at mid slope with annual cropping and located in a local groundwater recharge area, whereas the lower slope had permanent grass cover and in a groundwater discharge area (Miller *et al.*, 2014). Darcy's law was used to calculate the unsaturated water flux (q):

$$q = -Kdh/dL \quad [1]$$

This study follows the method of Miller *et al.* (2014) to estimate DD during successive dry and wet years at a forested site in Guelph, Ontario. Overall, Miller *et al.* (2014) determined the vertical water flux to be mostly downward at the mid slope site for the two years, ranging from <10-11 to 10-10 m s⁻¹, located in a recharge area. However, in 1985 there was higher upward flow, whereas downward flow was more dominant in 1986. Overall, the water flux for both years ranged from <10-11 m s⁻¹ to 10-9 m s⁻¹ for the lower slope site. According to Miller *et al.* (2014), the upward flow was

possibly due to high evaporation rate at the soil surface, along with a shallow water table. The total water flow as a percentage of annual precipitation at the mid slope site was approximately 0.8-1.8%, whereas at the lower slope it was 0.3-0.5% (Miller *et al.*, 2014).

The various results and methods used for these different studies aid in understanding the complexity when it comes to measuring or estimating soil DD. There are different factors that play key roles in influencing the results; for example, the land relief and soil texture of one location may give different results for another location with slight differences in topography and soil type. Most of the studies that have been undertaken to understand DD in Ontario examined water surplus from various theoretical and qualitative perspectives, which may not always accurately represent the real and complex environment, rather than quantitatively using field data. Hence, the objective of this study was to determine the soil hydraulic conductivity for a field site, along with the soil water storage, hydraulic gradient, and the DD using Darcy's Law for a two-year period of 2012 (dry year) and 2013 (wet year) at the University of Guelph's Arboretum.

Methods

This study is part of another larger study that estimates the groundwater recharge (GWR) with the use of the Water-Table Fluctuation (WTF) method performed by Thair Patros, University of Guelph. This was also a continuation of another study that compared the soil water status between two years, 2012 (dry) and 2013 (wet), conducted by Tamang *et al.*, (2014). Hence, the site description along with the instrument descriptions follow closely with the descriptions given in the previous study by Tamang *et al.* (2014).

The larger study had two field sites, one of which was at the Elora Research Station, located 25 km northwest of Guelph, Ontario at 43°38'27.05"N, 80°24'18.69"W. The other field site was at the University of Guelph's Arboretum, located northeast of the campus, at 43°32'39.06"N, 80°12'57.78"W. The site at the Arboretum was also used for the purposes of this study.

The Arboretum had approximately 60% sand, 10% clay, and 30% silt, which was determined to be Granby sandy loam. This soil type is from the Great Group of Dark Grey Gleysolic family and Orthic Dark Grey Gleysolic sub group, with the parent material being calcareous medium sand. In general, this soil type tends to have a pH of 7.6, A-horizon up to 17.8 cm thick, B-horizon up to 35.6 cm thick, and the rest is C-horizon. The excavation of the pond near the site had altered the soil profile. This was discovered when a pit was dug approximately in the middle of the site, which revealed buried A and B horizons.

Figure 1 shows the study site at the Arboretum, which was located in a forest setting. The site had an abundance of trees, a large amount of shrubs, plants, and

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bushes, and a tree canopy. There were 13 monitoring wells and 3 piezometers installed at the Arboretum site at various locations and elevations. A survey of site and instrument elevations was conducted using a Total Station. The site had three levels of elevation: upper gentle starting at an elevation of 334.954 m, intermediate/middle steeper surface at 334.251 m, and a narrow plain relatively flat low lying area at 332.853 m. The low lying area bordered a pond on the southern edge of the site. The slope of the site was 8.69%, which included the elevation from 332.853 m (Piezometer 3) to 335 m (Base) (Figure 1A). The survey “base” shown in Figure 1A was an arbitrary elevation set at 100 masl, assumed to be the elevation of the site for the survey measuring purposes of this study. The accurate elevations from the survey were used in calculations of hydrologic data by converting the elevation for the wells and piezometers in relative proportion to 335 masl (which was the approximate elevation of the base station), as indicated in Table 1A. Further description of the location and the elevation of the wells and piezometers that were installed can be found in Table 1A, Table 2A, Figure 1A and Table 3A.

Some of the instruments installed at the site were Solinst® Levelloggers, and Campbell Scientific's CS616 and T107. They were used to measure the groundwater level, soil volumetric water content, and temperature, respectively. Wells AW1R, AW2R, AW3, AW4, and AW9 contained the Levelloggers. These Levelloggers sensed and automatically recorded the water level every 5 minutes, which were then downloaded with the use of Levellogger software after 55 and 110 days and reprogrammed on the 110th day. This method helped to determine the water level fluctuation over time in the five wells. However, these data were unavailable for use, thus the data for depth to the water table used in this study were collected by the manual use of Solinst® Mini Water Level Meter (Model 102M).

The water level measurements were taken approximately once to twice a month. The depth to the water table data used for the previous study was only based on Arboretum Well 1 (AW1) due to time constraints for the project, and since AW1 was located closest to the CS616 sensors, it was assumed to be the most representative well for the entire study site. However, that is inaccurate, since each well was located at different elevation and surrounded by different types of soil and plants; thus, assuming homogeneity in the soil type for all the wells would lead to erroneous results. Hence, for this study, the hydraulic conductivity for each monitoring well installed at the site was used.

The CS616 sensor is composed of two 30-cm long stainless steel rods that measure the elapsed travel time of the electromagnetic pulse generated by the CS616, along with the pulse reflection. The measurements collected from it were used to calculate the soil volumetric water content (θ_v). There were ten CS616s installed horizontally in the wall of a soil pit by the AW1 and AW1R location, at 50-140 cm below the ground surface (bgs). Each sensor was 10 cm apart. The

measurements were automatically taken every 5 minutes and stored on a Campbell Scientific model CR-10X datalogger. They were then downloaded once every 24-27 days. The data collected [wave period (P) in microseconds(μ s)] were then converted to θ_v using equation [2]. There were data points (P) missing for 2012, from days 54-75, 151-164, and 330-337 due to technical issues with the instrument. For 2013, the data from days 100-197 for sensor 3 (at the depth of 70 cm) were missing, along with days 9, 107 and 273. The missing data were filled in by the use of simple linear interpolation between the existing data points:

$$\theta_v = -0.0663 - 0.0063 \cdot P + 0.0007 \cdot P^2 \quad [2]$$

where θ_v is assumed to be in soil having bulk electrical conductivity lower than 0.5 dS.m⁻¹, clay content less than 30% and the bulk density lower than 1.55 g.cm⁻³ (Campbell Scientific's CS616 Manual, 2002).

There were two T107 Temperature Probes installed to measure the soil temperature. The T107 probes were installed horizontally at depths of 52 and 93 cm bgs, close to the CS616 sensors. The CS616 readings and soil temperatures were recorded automatically every 5 minutes by the Campbell Scientific CR-10X datalogger, and were then downloaded every 24-27 days. The data for 2012 and 2013 precipitation (rainfall and melted snow) and air temperature were recorded at the Environment Canada's Guelph Turfgrass Institute (GTI) weather station located approximately 1.3 km from the study site.

The slug test (called falling-head test) was performed to measure the field saturated hydraulic conductivity (Kfs) based on the Hvorslev method. For this method, the measurements of the initial depth to the water table were taken. In order to determine the rate of water level recovery, water was added to each well until it started overflowing from the top, and the recovery rate was recorded every 30 s until the water level had nearly recovered back down to its initial depth. Although the method suggests measuring the recovery rate until the water recovers back down to its initial depth, due to differences in the soil texture, some wells took longer than others to fully recover. Since the difference between the last measurement point taken and the initial depth to water for the wells was not large (overall approximately 50 cm) and there were enough measurement points taken to estimate the recovery rate, it did not affect the final results. Due to the unavailability of some slug test data, only the measurements for AW1, 2, 5, 6, 7, 8, and 10 were used in the previous study by Tamang *et al.* (2014). However, for this study, the unavailable recovery rate and the data points for AW1R, AW2R, AW3, AW4, AW9 and AW11 were estimated based on the soil texture and the surrounding wells.

These collected and estimated slug test data were then used to determine the field saturated hydraulic conductivity using Hvorslev (1951) equation [3]:

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$$Kfs = (r^2 \ln(L_e/R)) / 2L \epsilon t_{37} \quad [3]$$

where r equals the radius of the well casing (cm), L_e is the effective length of the well screen (cm), R is radius of the well screen (cm), and t_{37} represents the time when the water level falls to 37% of the initial hydraulic head (s).

Soil samples were collected from two pits at the study site, one of which was where the CS616 and T107 sensors were installed (Pit 1), while the other was approximately 5 m north of Pit 1 (Pit 2). The soil sample cores used for the soil moisture retention curve (SMRC) were 2.5 cm in length and 4.7 cm in diameter. On the other hand, the cores used for the measurement of the bulk density (ρ_b), particle density (ρ_s), saturated hydraulic conductivity (K_{sat}) and particle size distribution (PSD) were 5 cm in length and 4.7 cm in diameter. The soil samples from Pit 1 were collected systematically at an interval of 30 cm from 60-150 cm bgs. The soil samples from Pit 2 were collected systematically at an interval of 20 cm from 0-200 cm bgs.

The rest of the data for the ρ_s , ρ_b , and PSD were measured in the laboratory using the collected soil samples (Pit 1 and Pit 2). The procedures for these measurements were based on methods described in Carter and Gregorich (2008). Equation [4] was used to determine the ρ_s using a pycnometer, and equations [5], [5.1] and [5.2] were used to determine the soil texture percentage (% silt, % clay and % sand) using the hydrometer method.

$$\rho_s = [\rho_w (W_s - W_a)] / [(W_s - W_a) - (W_{sw} - W_w)] \quad [4]$$

where ρ_s is the particle density (g/cm³), ρ_w is the density of water (g/cm³), W_s is the weight of the pycnometer plus soil (g), W_a is the weight of pycnometer filled with air (g), W_{sw} is the weight of the pycnometer filled with soil and water (g), and W_w is the weight of pycnometer filled with water (g).

$$\text{Sand \%} = 100 - (R_{40} - RL) \times 100 / \text{oven-dried soil weight in g} \quad [5]$$

$$\text{Clay \%} = (R_7 - RL) \times 100 / \text{oven-dried soil weight in g} \quad [5.1]$$

$$\text{Silt \%} = 100 - (\text{sand \%} + \text{clay \%}) \quad [5.2]$$

where R_{40} is the average of the three 40s readings (g/L), RL is the hydrometer scale reading (g/L) in the calibration solution, and R_7 is the 7-hour reading (g/L).

The soil moisture retention curve (SMRC), which indicates the amount of water retained in a soil under equilibrium at a given matric potential, was based on the data collected through the pressure plate systems. The procedure for the pressure plate systems was also based on the methods described in Carter and Gregorich (2008). The water content of the soil samples was measured gravimetrically (θ_g) using

equation [6], once the objective pressure of the method was reached. The θ_g was later converted to θ_v using equation [7]. The SMRC was produced based on the weighted average of the volumetric content at each applied positive (assumed negative equivalent) pressure head (hp) for both Pit 1 and Pit 2 (Figure 5a, 5b and 5c) at 0, 10.2, 20.4, 30.6, 60, 71.4, 102, 306, 346.7, 509.9, 713.8, 1019.7, 3059.1, 5098.6, 7138, 10197.2, and 15295.7 cm. The experiment at 1019.7 cm was repeated three times due to erroneous data present.

In the previous study by Tamang *et al.* (2014), the SMRC for Pit 1 only was produced due to time constraints and thus was assumed to be representative of the entire site. However, as mentioned earlier, this is not accurate, as the soil texture varies at the site from location to location and with soil depth. Thus, producing SMRC for Pit 1 (lower half of the site) and Pit 2 (intermediate to upper half of the site) was more representative of the entire study site (Figure 5a, 5b and 5c).

$$\theta_g = M_w / M_s \quad [6]$$

where M_w is the mass of water (g) and M_s is the oven dry soil mass (g).

$$\theta_v = \theta_g (\rho_b / \rho_w) \quad [7]$$

where ρ_b is the bulk density (g/cm³) and ρ_w is the density of water at 25°C (g/cm³).

There was variability present between Pit 1 and Pit 2, such as: different soil textures at various depths between the two pits; the intervals at which the soil samples were collected were different for the two pits; the soil cores for Pit 2 (depth of 120-140 cm) were destroyed; and Pit 2 (up until 100 cm below ground surface) was not the actual soil horizon—it was a mixture of organic matter with soil, which may have resulted in erroneous data for Pit 2 above 100 cm. Since the overall soil texture for both pits was Sandy Loam, which was the same as the overall soil texture of the study site, it was considered appropriate for the weighted average of the two pits to be taken for calculating the SMRC, regardless of the variability present.

For the overall average SMRC of the two pits, the average water content at each hp was multiplied by the total depth for each pit. The sum of the two multiplied averages for the two pits was then divided by the total depth of both pits. For example: the average water content at hp = 0 for Pit 1 was 0.4178 and for Pit 2 was 0.4895. These were then multiplied by the depth of each pit: 0.4178 X 90 and 0.4895 X 200. The resulting values were added and then divided by 290 (90 cm + 200 cm). This procedure was then repeated for all the soil cores at each hp, resulting in average θ_v at each applied hp, which was then used to produce the SMRC for the average of Pit 1 and Pit 2 (Figure 5c).

In order to have a continuous SMRC, a line of best fit through the SMRC data points using Russo's (1988) equation was used in the previous study. However, for this

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study, equation [8] (van Genuchten *et al.*, 1980) was used because van Genuchten's equation led to a more accurate and closer best fitting line in comparison to the Russo's equation due to the two unknown parameters (α and n) involved in van Genuchten's equation. The ' α ' used in the equation is related to the pore size distribution of the soil and ' n ' is a parameter that indicates the steepness of the SMRC curve. RETention Curve (RETC), a computer program which determines the line of best-fit of equation [8] to SMRC data and unsaturated hydraulic conductivity, was used to automatically determine the ' α ' and the ' n ' fitting values (van Genuchten *et al.*, 1991).

$$\theta(h_p) = \frac{\theta_s - \theta_r}{[1 + (\alpha|h_p|)^n]^{1-1/n}} + \theta_r \quad [8]$$

where $\theta(h_p)$ is the soil water content at each applied h_p , θ_s is the saturated soil volumetric water content (equivalent to $h_p = 0$), θ_r is the residual soil volumetric water content (approximately equivalent to $h_p = -15300$ cm), and α (1/L) and n (dimensionless) are best-fit parameters using RETC (Parkin, 2014).

Unlike the previous study, which used Gardner's equation, this study used the van Genuchten (1980) equation [9], to determine the unsaturated hydraulic conductivity (K).

$$K(\theta^*) = K_{fs}\theta^{*1/2} [1 - (1 - \theta^{*1/m})^m]^2 \quad [9]$$

where K_{fs} is the average field saturated hydraulic conductivity (cm d^{-1}) based on equation [3]. $\theta^* = (\theta_v - \theta_r) / (\theta_s - \theta_r)$ where θ_v was determined using CS616 sensors, equation [1]; θ_s and θ_r were determined using RETC, m is a fitting parameter of SMRC, $m = 1 - 1/n$.

The drainage calculation based on Darcy's Law (equation [1]) was applied above the water table. Once the average depth to the water table was determined using the wells (excluding the stick up height), the greatest depth of the CS616 sensors was determined. The stick up height was based on the initial measured stick up height, unless it had been updated. If it was updated, the new updated value was used for that particular well, after that specific date (Table 4A).

To determine the vertical hydraulic gradient for the drainage calculation, two CS616 sensors at two different depths were selected: the shallowest depth was always the elevation for the shallowest CS616 sensor (50 cm below soil surface) and the second depth was the depth to the deepest CS616 sensor, based on the sensor that was still above the water table. For example: in Figure 2, the average depth to the water table on that particular day was 105 cm. The elevation for CS616 sensor 1 (50 cm) was used, along with the deepest sensor, CS616 sensor 6 (100 cm). Once the deepest sensor to be used was determined (e.g., sensor 6), the average measured θ_v for that sensor on that specific day (e.g., January 6, 2012) was used to determine the h_p using the average SMRC. The first sensor was always assumed to be CS616 sensor 1. Once the h_p for both sensors (sensor 1 and the deepest sensor) was

determined, the gravitational head (z) was determined for each sensor, assuming the reference level, $z = 0$, was always at the bottom of the deepest CS616 sensor. This led to the determination of the two hydraulic heads (h_1 and h_2) since $h = h_p + z$, for that particular day when the measurements were taken. The hydraulic gradient (dh/dL or i) was then determined using equation [10]:

$$dh/dL = (h_2 - h_1)/(L_2 - L_1) \quad [10]$$

where h_1 and h_2 are the hydraulic heads at sensor 1 (point 1) and the deepest sensor (point 2), respectively. L_1 and L_2 are lengths of the column below ground surface at sensor 1 (point 1) and the deepest sensor (point 2), respectively.

These data were then inserted into Darcy's law, equation [1], along with $K(\theta^*)$ from equation [9], which determined the specific discharge (q , cmd^{-1}) for 2012 and 2013.

The DD (mm) was then determined using equation [11], where Δt is the change in time over which q was estimated in equation [1].

$$DD = q \times \Delta t \quad [11]$$

Results and Discussion

The pattern between the air temperature and the amount of precipitation for the years 2012 and 2013 was similar. The air temperature fluctuated based on the seasonal changes, given in Figure 3. The air temperature for 2012 and 2013 started to increase around May, reaching its peak around July and then decreasing again. Although the precipitation patterns were also similar for 2012 and 2013, both reaching their peaks around July to October, the amount of precipitation in 2013 was higher by 175.6 mm compared to 2012. The daily maximum precipitation for 2012 occurred on Day 248 (September 4th) and was 39.6 mm. On the other hand, the daily maximum precipitation for 2013 occurred on Day 264 (September 21st) and was 40.4 mm.

The soil water storage for the years 2012 and 2013 were based on the average of all the CS616 sensors for each time (day) the measurements were taken. The daily soil water storage in 2013 was higher than in 2012 by an average of 38 mm due to higher precipitation and infiltration, resulting in shallower depth to the water table for 2013 (Figure 4). Also, higher air temperature in 2012 may have resulted in greater evapotranspiration, further reducing the soil water storage in this dry year. Depth to the water table followed the same pattern as soil water storage between the two years (Figure 4).

As shown in Figure 4, the soil water storage was higher at the beginning of 2012, which could be due to higher snowmelt in 2012 than in 2013. The gradual decrease in storage in the summer of 2012 was possibly due to evapotranspiration exceeding the amount of precipitation in that year compared to 2013. The average shallowest depth to

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the water table for all the wells in 2012 was 1.3 m (May) and the deepest was 2.1 m (October), whereas in 2013, the average shallowest depth to the water table was 0.710 m (May) and the deepest was 1.6 m (September). The sharp increase in the soil water storage around day 300 in 2012 and 2013 indicates the time when groundwater recharge possibly started. The soil water storage and the depth to the water table fluctuate proportionally. The increase in depth to the water table results in higher soil water storage and decrease in the water table results in lower soil water storage. This is an indication of the soil moisture content affected by precipitation and its influence on the soil water storage.

The difference between using only one well (AW1) (Appendix B; Figure 3B) and all the wells (Figure 4) in assessing the depth to the water table can be noticed. In Appendix B; Figure 3B, the depths to the water tables in 2012 and 2013 are higher than they should be for the entire site. This could be due to AW1 being close to the pond at the low elevation, which may have caused the water table to be higher than the rest of the wells. Thus, using only one well is not representative of the entire site.

The histogram showing the soil water storage for 2012 and 2013 is shown in Appendix B; Figure 2B. Overall, in 2013, the soil water storage ranged from 385 to 480 mm, the most frequent amount being around 450 mm, whereas in 2012 the soil water storage ranged from 340-450 mm and the most frequent amount was approximately 340-345 mm. The longer time required for the recharge to occur in spring 2013 could have been a result of low soil water storage in 2012 (McCoy *et al.*, 2006).

The SMRC, in Figures 5a, 5b and 5c, were plotted based on the average θ_v at various applied pressure heads, from 0 – 15300 cm, that were calculated based on the data collected from the pressure plate system. The values used in the van Genuchten's equation are shown in Table 2A. The solid line represents the line of best fit determined by using equation [8] in RETC. The unsaturated hydraulic conductivity (K), on average, was roughly 2 times higher in 2013 than in 2012 based on equation [9]. The differences in the unsaturated K values for the two years could possibly be a result of higher water content in 2013.

The slug test data that were not available had to be estimated by using the data of the wells that were available. The estimation was based on various components: the wells surrounding the wells with the missing data, comparison of their soil texture to determine if it was the same soil type, and the equivalent elevation at the bottom of the well screen. The data of the closest well with minimal differences were used. However, some of these available data have a possibility of errors being involved, since some of the graphs made for the wells based on the slug test data did not have an accurate trend line fitting through the data points. These water level versus time data points may have been affected by unsaturated soil conditions. The Hvorslev method states that the well screen has to be in the saturated zone; however, a portion of

some of the well screens at the site were above the water table, which may have resulted in erroneous data in the field. The first datum after removing these data points was used as the reading at time 0. This resulted in a better fitting trend line. These sources of errors may have resulted in an inconclusive average Kfs of 6.104 cm/day.

In 2013, most of the θ_v at hp1, extracted from CS616, were higher than 0.3902 ($h_p = 0$ in Pit 1), but not higher than the average value using SMRC from both pits. This issue was detected later in the calculation process. The θ_v for 2013 could be inaccurate due to the use of equation [2]. For example, using equation [2] assumes the bulk density to be lower than 1.55 g.cm⁻³, which does not match with the bulk density of the overall soil layers in Pit 1, which was 1.639 g.cm⁻³, or Pit 2, which was 1.638 g.cm⁻³. Although there was heterogeneity in the soil texture and variability in Pit 1 and 2, the various reasons for choosing to average the two pits were mentioned earlier. Since the h_p from the average of the two pits and the Kfs, with possibilities of errors, was used in equation [9] to determine the unsaturated K, this may have resulted in inaccurate values for the unsaturated K.

Figure 6 shows the relationship between daily DD and unsaturated K. For 2012, as the K increased, so did the DD, and as K decreased, the DD also decreased. A similar pattern can be noticed for 2013 as well. As K value fluctuated throughout the year, so did DD, which is proportional to K. This could be because the higher the K value is, the faster the water moves, resulting in greater DD; a lower K value results in slower water movement, thus lower DD. The DD for both 2012 and 2013 started to decrease around Day 200. This was when the SWS for 2013 was still slightly increasing. However, in 2012, the SWS started to decrease around Day 176. This was the start of the period when the soil was so dry that the SWS decreased and the value for DD became negative, as shown in Figure 6. The negative DD values indicate the flow of water to be upwards (Day 205-291), which is plausible due to evapotranspiration exceeding precipitation.

Overall, as shown in Table 1, the average daily DD in 2013 was approximately 294 mm higher than 2012. This could be due to higher SWS in the year 2013 than in 2012.

The statistical summary of soil water storage, K, and DD for years 2012 and 2013 can be found in Table 1. As seen in Table 1, the sum of the DD for 2013 was higher than 2012. The sum was based on the data available for the days when the depth to the water table measurements were taken. Although the value for the sum may not be precise for the two years, higher DD is expected in 2013 due to higher precipitation. The overall mean of the daily DD for 2013 was higher than in 2012 by 25 mm. The daily median was also higher for 2013 than 2012 by 27.5 mm. The variance and the standard deviation for 2013 was higher than in 2012, indicating fewer differences in the daily amount of water stored for 2012. The data for 2012 and 2013 are both skewed to the right, indicating more frequent months of low DD, however 2013 had higher DD than 2012 (Figure 7a and 7b). The negative kurtosis for 2012

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and 2013 indicate that the frequency of occurrence of DD was less around the mean. The maximum for 2012 (May 6th) is lower than 2013 (May 31st). The minimum is lower in 2012 (October 17th) than 2013 (September 30th). The standard deviation is 81% of the mean for 2012 and 46% of the mean for 2013. These high values for the coefficient of variation could possibly be a result of inaccurate means for 2012 and 2013. However, these results indicate that the variability of DD in 2012 was higher than in 2013, which could be due to inconsistent and lower precipitation in 2012 than in 2013. Miller *et al.* (2014) indicated the dominance of upward flow in 1985 and downward flow in 1986, which is similar to slight upward flow in 2012 and mostly downward in 2013 in the present study, due to higher evaporation at the soil surface and shallow water table (Miller *et al.*, 2014).

Conclusion

The results presented in this paper help to understand how the unsaturated hydraulic conductivity affects the amount of DD in soil at the local scale, where the knowledge of groundwater recharge is lacking or even unavailable for many parts of Ontario. Generally, higher hydraulic conductivity, such as in 2013, leads to higher DD. On the other hand, lower hydraulic conductivity, such as in 2012, leads to lower DD. The DD is extremely important since it helps to estimate the groundwater recharge. The DD can be affected due to temporal and spatial variables, such as soil texture and topography, as found at the study site and discussed earlier.

Overall, there was higher amount of precipitation in 2013 than in 2012, the unsaturated hydraulic conductivity was 2 times higher in 2013, there was approximately 38.5 mm of higher soil water storage on average in 2013, and the average DD was approximately 25 mm higher in 2013. These results indicate that higher precipitation leads to higher DD and, possibly, faster groundwater recharge.

Although this was a small scale study, it can still be applied in a larger context. Since the DD varies with location, the results from this study can be applied to other environments that have similar soil texture, topography, and location. These results can also help to make educated decisions in the future in various fields, such as soil water conservation, to minimize any consequences of future climate change.

Recommendations

This study can also be further extended to understand the movement and amount of water flux and DD in greater details. Some of the things that could potentially be improved in the future are making sure that the assumptions made by Campbell CS616 sensors, such as bulk density lower than 1.55 g.cm⁻³, are met. Another recommendation would be to perform the slug test on all the wells installed at the site, in order to minimize any errors that may have occurred due to

estimating the results for the missing slug test data based on the other wells. In addition, it is recommended to ensure that the water recovers back to its initial depth so an accurate recovery rate can be determined. Another option would be to perform the bail-down test instead, by pumping the water out of the well and instantaneously lowering the water table. The rate of the water rising in the well can then be monitored. This ensures that the bottom of the well screen is under the water table, which meets the assumption of the Hvorslev method. Lastly, it is recommended to determine the unsaturated hydraulic conductivity for each layer at the site (wells), rather than assuming the same unsaturated hydraulic conductivity for each well. This will help to obtain more accurate results since the heterogeneity of soil texture for each well will be considered in the calculations.

Acknowledgements

The first author wishes to thank Dr. Gary Parkin and Thair Patros for their guidance with this project and for their tremendous amount of help to prepare the final paper.

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Tables and Figures

Figure 1: Study site at the Arboretum showing the pond, piezometers and wells (after Patros, T., 2010).

Figure 2: Example for determination of the hydraulic gradient.

Figure 3: Comparison of daily precipitation (mm) and air temperature (°C) for 2012 and 2013.

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Figure 4: Graph showing the daily soil water storage (mm) and depth to the water table (m) for 2012 and 2013.

Figure 5a: Average Soil Moisture Retention Curve (60-150 cm depth) for Pit 1 with the line of best fit using the van Genuchten’s equation [8].

Figure 5b: Average Soil Moisture Retention Curve (0-200 cm depth) for Pit 2 with the line of best fit using the van Genuchten’s equation [8].

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Figure 5c: Soil Moisture Retention Curve for the average of Pit 1 and 2 with the line of best fit using the van Genuchten's equation [8].

Figure 6: Graph showing the daily deep drainage (mm) and unsaturated hydraulic conductivity (cm/day) for 2012 and 2013.

Figure 7a: Histogram showing deep drainage (mm) for 2012.

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Figure 7b: Histogram showing deep drainage (mm) for 2013.

Table 1: Statistical summary of daily soil water storage, Unsaturated K and Deep Drainage for years 2012 and 2013.

	SWS (mm)		K (cm/day)		DD (mm)	
	2012	2013	2012	2013	2012	2013
Sum (mm)	145743	159415	12	22	108	402
Daily Mean (mm)	398	437	1	2	6	31
Daily Median (mm)	404	440	1	2	8	36
Variance (mm ²)	1208	610	0.1	0.3	26	199
Standard Deviation (mm)	35	25	0.3	1	5	14
Skewness	-0.3	-1	-0.4	-1	-0.5	-1
Kurtosis	-1	-0.3	-2	-1	-2	-1
Max (mm)	446	478	1	2	13	49
Min (mm)	341	380	0.4	1	-1	7
Coefficient of Variation (%)	9	6	35	30	81	46

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Table 1A: Elevation for the wells and piezometers installed at the study site in the Arboretum

Well/Piezometer	Elevation (masl)
AW1	333.773
AW1R	333.781
AW2	334.606
AW2R	334.666
AW3	332.794
AW4	334.954
AW5	334.845
AW6	334.165
AW7	334.699
AW8	334.251
AW9	334.277
AW10	334.365
AW11	334.328
P1	332.946
P3	332.853
P6	332.763

Table 2A: Brief summary of the information of the site and the values used in calculations

Sand	60%	PD Pit 1 (g/cm³)	2.533
Clay	10%	PD Pit 2 (g/cm³)	2.574
Silt	30%	BD Pit 1 (g/cm³)	1.639
Soil Texture	Granby Sandy Loam	BD Pit 2 (g/cm³)	1.638
pH	7.6	Porosity Pit 1 [-]	0.353
A – Horizon	< 17.8 cm thick	Porosity Pit 2 [-]	0.364
B – Horizon	17.8 cm - 35.6 cm thick	K-Sat Pit 1 (cm/s)	0.009
C – Horizon	> 35.6 cm thick	K-Sat Pit 2 (cm/s)	0.006
Elevation of site	335 masl	K_s (cm/day)	6.104
Upper slope elevation	334.954 masl	θ_s	0.449
Intermediate slope elevation	334.251 masl	θ_r	0.054
Low slope elevation	332.853 masl	α	0.034
Slope	8.69%	n	1.286

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Table 3A: Description of the wells and piezometers installed at the Arboretum study site (after Patros, T., 2010).

Group	Well/Peizo	Description					DBMP (m)	Installation Method
		SL (m)	Le (m)	B (m)	S (m)	DBSS (m)		
I	1	0.70	0.95	0.30	0.30	1.55	2.3	Polaris/Hand-Augered
	1R	*1.50	1.89	0.15	—	2.04	2.75	Polaris
II	2	*1.50	1.57	0.15	0.10	1.82	2.57	Hand-Augered
	2R	*1.50	2.32	0.15	—	2.47	3.01	Polaris
III	3	0.70	0.83	0.20	—	1.03	2.22	Polaris
IV	4	*1.50	1.565	0.85	—	2.42	3.265	Polaris
V	5	*1.50	1.59	0.81	—	2.40	3	Polaris
VI	6	*1.50	1.60	0.47	—	2.07	3.47	Polaris
VII	7	*1.50	1.90	0.50	—	2.40	3.52	Polaris
VIII	8	*1.50	1.51	0.55	—	2.06	3.52	Polaris
IX	9	*1.50	1.63	0.60	—	2.23	3.63	Polaris
X	10	*1.50	1.60	0.80	—	2.40	3.52	Polaris
XI	11	*1.50	1.48	0.60	—	2.08	3.52	Polaris

Abbreviation	Definition
AW	Arboretum Well
Peizo	Peizometer
R	Replicate
SL	Screen Length
Le	Effective Screen Length
B	Bentonite
S	Soil
DBSS	Depth Below Soil Surface
m	Meter
Elev.	Elevation
masl	Meter above sea level
SW	Surface Water
DBMP	Depth Below Mounted Point

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Table 4A: Stick up height (cm) for each well which varied depending on the date (after Patros, T., 2010).

Date : July-27-10 - June-20-14		Date: Sept-24 - Sept-27-10		Date: July-20-12 - June-20-14		Date: Dec-13-12 - Apr-26-13		Date: May-03-13 - June-20-14	
AW	Stick up Value (cm)	New Stick up Value (cm)	Date	New Stick up Value (cm)	Date	New Stick up Value (cm)	Date	New Stick up Value (cm)	Date
1	75								
1R	71			124	July-20-12				
2	75								
2R	54					203	Dec-13-12	112.5	May-03-13
3	119			164	July-20-12				
4	85					208	Dec-13-12	112.5	May-03-13
5	60								
6	140	118	Sept-24-10						
7	112								
8	146	119	Sept-27-10						
9	140	102	Sept-27-10			226.5	Dec-13-12	133	May-03-13
10	112	85	Sept-27-10						
11	144	117	Sept-27-10						

Appendices

Appendix A

Figure 1A: Contour map of the study site at the Arboretum indicating the location of the wells (AW) and piezometers (P) installed at various elevations (masl) (after von Bertoldi, P., 2014).

Appendix B

Figure 1B: Comparison of daily soil temperature (°C) with soil volumetric water content for 2012 and 2013.

Figure 2B: Histogram showing soil water storage for 2012 and 2013.

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Figure 3B: Graph showing the daily soil water storage (mm) and depth to the water table (m) using AW1 for 2012 and 2013.