RESPONSE OF ECOSYSTEM EVAPOTRANSPIRATION TO WATER-STRESS IN A TEMPERATE DECIDUOUS FOREST IN SOUTHERN ONTARIO

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LAY ABSTRACT

With seasonal droughts occurring more frequently from climate change, water required by tree species may reach limitations to sustain atmospheric demand. In particular, temperate deciduous forests, that are common in the northern hemisphere, may be affected due to predicted future climate scenarios. It is important that we understand how they will respond to future climate conditions. Understanding how forest evapotranspiration is affected by various meteorological and hydrological parameters will help gauge how forests will cope with limiting water resources in the future. An assessment of a temperate deciduous forest in southern Ontario, which is part of the Turkey Point Flux Station, was conducted in this study over 5 years (2012-2016). Over this period the forest experienced varying drought intensities. Results show that air temperature and vapour pressure deficit were the dominant controls on forest ET during times of water limitations. The forest was able to withstand low-moderate droughts by accessing deep soil water stores to keep up with the extreme demand. This study suggests that temperate deciduous forests in the region are well adapted to drought stress and they may be able to cope with similar intensity droughts in the future by maintaining water flow.

ABSTRACT

Seasonal droughts are becoming more prevalent in recent years as a result of climate change and associated fluctuations in precipitation events. Forested ecosystems react very differently to changes in the hydrological cycle in different regions. This study investigates how atmospheric water fluxes react to changes in meteorological variables and subsurface hydrological conditions in a 90-year old temperate deciduous forest in southern Ontario, Canada, using eddy covariance flux data over five years (2012-2016).

Results show that although drought occurrences in 2012, 2015 and 2016 put stress on the forest ecosystem, with significantly varying climate and soil conditions, no substantial impact on forest evapotranspiration appeared to occur. Annual eddy-covariance mean evapotranspiration of the study was 382 ± 46 with a growing season average of 353 ± 36. Highest annual evapotranspiration (422 mm) occurred in 2012 with 389 mm occurring during the growing season while the lowest annual evapotranspiration (341mm) occurred in 2014 with 316mm occurring in the growing season. Air temperature and vapour pressure deficit had the dominant control on evapotranspiration as expected, with the highest sensitivity occurring during drought years. Soil water potential (SWP), as the soil moisture stress proxy, reached maximum values during drought years, but was not found to have a significant control with evapotranspiration. SWP peaked when evapotranspiration values reach maximums combined with warm air temperatures, progressively increasing when precipitation input was minimal. Our results indicate that evapotranspiration rates in this forest ecosystem were predominantly determined by atmospheric controls when accessibility of deep soil water was possible. These findings suggest that temperate deciduous forests were able to sustain ET during low to moderate intensity droughts and observed over the study period due to their ability to access deep soil moisture stores during periods of stress. However, these forests may experience significant declines in ET if severe or multi-year drought conditions arise, similar to those that occurred in the last two years of this study. These conclusions will help to elucidate how ET in deciduous forests will respond to future climate regimes in this region.

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LIST OF ABBREVIATIONS AND SYMBOLS

CPEC = Closed-Path Eddy Covariance

EC = Eddy Covariance

ET = Ecosystem Evapotranspiration

H = Sensible Heat Flux

IPCC = International Panel on Climate Change

LE = Latent Heat Flux

PAR_{dn} = Downwelling Photosynthetically Active Radiation

Ppt = Precipitation

Ppt – ET = Precipitation minus Evapotranspiration Balance

REW = Relative Extractable Water

SWP = Soil Water Potential

T_a = Air Temperature

TPD = Turkey Point Deciduous Forest

TPFS = Turkey Point Flux Station

VPD = Vapour Pressure Deficit

VWC_{0-30} = Shallow-zone (0-30cm) Volumetric Water Content

VWC_{0-100} = Rooting-zone (0-100cm) Volumetric Water Content

1 INTRODUCTION

1.1 Importance of Forest Ecosystems in the Global Water Cycle

Forests cover approximately 30% of Earth's land mass (3.9 billion hectares) and play a significant role in the global water cycle (FAO, 2015). Forest water cycles are reported to be sensitive to changes in climate (Bonan, 2008; Allen et al., 2010) and with IPCC (2013) predictions of increases of global temperatures, combined with less frequent low-intensity precipitation and higher storm frequencies, forests are put at great risk (Pachauri et al., 2014). Warming will not occur uniformly across the globe (Martin-Benito and Pederson, 2015). Temperatures in southeastern parts of Canada are expected to increase by 5 degrees Celsius by the end of this century (Wang et al., 2015). It has also been suggested that the spatially non-uniform changes in climate are causing more frequent intermittent short-term drought events (Bréda et al., 2006; Wu et al., 2013), which may have great impacts on forest water cycles and ultimately ecosystem growth (Choat et al., 2012). As the major contributor of water loss in forest ecosystems, evapotranspiration (ET) plays a critical role in the overall hydrological cycle in forests. Therefore it is important to improve our understanding of the various controls and limitations on ET in forest ecosystems.

Forests provide numerous ecosystem services such as soil resource protection, lumber, food and medicinal products, recreation and most importantly a regulator of the hydrological cycle. Water is the building block of forest ecosystems and the uneven distribution across the Earth is the determining cause of the unique ecozones that we have. Water availability has shaped our planet since the beginning and the dramatic changes in recent years have altered forests as we know them. Without adequate water, effective carbon assimilation in forests would be immensely hindered and thus global warming would be exacerbated. Forests have the ability to dampen or intensify the effects of global warming to a significant degree (Bonan, 2008) acting as climate feedback systems.

As the major contributor of water transfer to the atmosphere in forest ecosystems, evapotranspiration (ET), which is the combination of tree transpiration and evaporation, plays a critical role in the overall hydrological processes in forests. Approximately 39 ± 10% of incident precipitation is returned to the atmosphere as transpiration based on a study spanning 81 sites (Schlesinger and Jasechko, 2014). With up to 50% (in some cases) of the incoming water being released back into the atmosphere through transpiration alone and greater than 100% when the entire process is taken into account (Bracho et al., 2008), minor changes in the controls determining ET in forests can vastly alter the global water cycle. Therefore it is important to improve our understanding of the various controls and limitations on evapotranspiration in forests in a variety of ecozones. Currently the effects of temporally differing droughts on forest water loss and productivity via meteorological and environmental controls are limited (Fisher et al., 2007; Granier et al., 2007). With the increasing likelihood of warming in coming years, these affects need to be further investigated as water loss is directly related to forest productivity.

1.2 Role of Temperate Forests in the Global and Regional Water Balance

Water-stress in forests has been extensively studied on the ecosystem scale: boreal coniferous (Chang et al., 2014; Petrone et al., 2014), mountainous coniferous (Nourtier et al., 2014), temperate coniferous (Mackay et al., 2012; Skubel et al., 2016), temperate mixed (He et al., 2014), subtropical coniferous (Tang et al., 2014), tropical deciduous (Fisher et al., 2007; Mendivelso et al., 2014). However, the resiliency of temperate deciduous forests to maintain ET during water-limited periods is not well documented, although they are considered to be common in the northern hemisphere (Wu et al., 2013; Ma et al., 2014). It has been reported that in temperate forest ecosystems decreasing annual precipitation (PPT) does not typically have a direct profound effect on overall water loss to the atmosphere, but rather only minor changes (Granier et al., 2007; Wu et al., 2013; Skubel et al., 2015). In these ecosystems, ET rates have been found to be largely dependent on vapour pressure deficit (VPD) and air temperature (T_a) (Pejam et al., 2006; McLaren et al., 2008; Brummer et al., 2012). In contrast, soil water content, a factor dependent on precipitation, has been found to have significant effects on overall ET (Brooks et al., 2002; Granier et al., 2007; Domec et al., 2010; Brown et al., 2013; Tang et al., 2014), but not as dominant when compared to atmospheric controls in some cases. With a changing climate and a diverse range of forest ecosystems, controls on forest water loss (ET) can be radically differing, and the effects on the regional water balance is left inconclusive. Overall, it has been reported that different forest ecosystems have varying mechanisms to conserve water and maintain ET in response to drought conditions (Bonan, 2008; Bittner et al., 2010). It is therefore evident that the investigation of a wide array of parameters is of importance when studying forest water loss.

1.3 Controls on Evapotranspiration in Temperate Forests

Evapotranspiration is a complex process that is influenced by several parameters. As documented in the literature, both above and below-ground controls have been found to have dominant controls on ET, thus it is vital to focus on both to fully understand the underlying effects of atmospheric water transfer.

The atmospheric constraints on forest evapotranspiration have been well studied in the literature. Fluctuations in ET have been found to be predominantly controlled by incident radiation and vapour deficits (Barbour et al., 2005; Fisher et al., 2008). Thresholds have been concluded for both variables, whereby significant changes in ET are evident once exceeded (Bovard et al., 2005). Vapour pressure deficit (VPD) being a primary control on ET is consistent throughout the literature, whereas the threshold at which effects on ET are observed varies greatly (Bovard et al., 2005; Mclaren et al., 2009). Air temperature positively correlates with ET (Fisher et al., 2007), being interconnected to radiation and VPD.

The controlling physiological factor of transpiration in forests is the stomata of the leaves, which is affected by the aforementioned variables, and are regulated by both atmospheric and subsurface dynamics. Soil moisture availability, similar to meteorological controls, has been found to have a significant control on ecosystem ET once a lower threshold is reached and thus limiting ET rates (Granier et al., 2007), a direc result of stomatal closure in an attempt to reduce excessive water loss (Wilson and Baldocchi, 2000; Phillips and Oren, 2001; Tang et al., 2014). This occurs when extraction through the roots become too difficult which then restrains the tree to meet atmospheric demand. Thresholds have also been found for relative extractable water (REW), a widely used calculated parameter introduced by Black (1979) derived from water content in the soil, representing extractable water in the rooting zone (Bréda et al., 1995; Granier et al., 2007; Gartner et al., 2009; MacKay et al., 2012, Tang et al., 2014). Another parameter used to identify water uptake stress is soil water potential (SWP). SWP represents the potential for water to do work in the soil matrix, expressing the soil water availability and the ability of tree species to obtain it. Limited literature is available on using SWP as a proxy for controlling ET, but thresholds of -1.6MPa were found to show decreased ET rates (Granier et al., 2007). Subsurface water stress can result in stomatal closure, which causes significant reductions in forest productivity and growth as it results in the reduced carbon-water exchange process through the leaves of the trees.

1.4 Significance of Study

Forest evapotranspiration is a topic of concern due to the uncertainty about forest ecosystems' resiliency to drought in relation to predictions of climate change. With current IPCC warming predictions due to increases in CO_2 concentrations from pre-industrial era values, significant changes can be expected in forest ecosystems across the globe. Changes in temperature and precipitation can significantly influence meteorological and environmental variables that directly influence ecosystem ET. With lower precipitation and higher temperatures, vapour pressure deficit in the atmosphere will increase the demand for more water efflux from the land surface including forests. With insatiable water demands from the atmosphere, the draw of soil water stores from deeper soil layers will have to occur, but without sufficient replenishment, these stores will eventually decline to the point where atmospheric demand will begin to have detrimental effects on the overall physiology of the trees. These detrimental effects to forest ecosystems in this region may become severe as multi-year droughts occur, which we found during this study time frame, an event that is rare in this region. It is thus important to quantify the resiliency of forest ecosystems to a combination of above and below ground stressors as alarming climate predictions have great potential for intensifying feedback cycles in forests (Bonan, 2008).

In this study, we analyzed water fluxes measured by the eddy covariance (EC) technique and associated meteorological and subsurface variables over five years (2012-2016) to assess water-stress sensitivity in a temperate deciduous forest in southeastern Canada. The primary goal of the study is to close the knowledge gap regarding the resiliency of a temperate deciduous forests ability to maintain evapotranspiration during droughts of differing magnitudes to improve our understanding on how future climate change will affect hydrologic cycles in forests. In particular we focus on understanding how the mutual effects of both atmospheric and subsurface factors control the overall ecosystem evapotranspiration during drought and non-drought years.

1.5 Research Objectives

The overall objective of this study is to describe the controls of evapotranspiration in a managed temperate deciduous forest in southern Ontario, Canada. This study is unique in terms of its geographic location in the temperate zone. It is part of the Carolinan ecozone in Canada, overlain on well-drained sandy soils. Over the term of the study period, the forest underwent three annual drought events and two climate-normal years where comparisons could be studied. The automated Eddy Covariance data collection method combined with meteorological and environmental instruments allowed for the delineation of specific ecosystem controls on the hydrological cycle and overall water loss of the forest ecosystem.

The specific objectives of this research are to i) advance our understanding of how hydrologic cycles in temperate deciduous forests respond to both above and below-ground water-stress and ii) what are the main controls on ET during water stress periods.

We hypothesized that both annual and growing season ecosystem evapotranspiration would increase during drought years as strained water availability combined with warmer atmospheric temperatures would put stress on the forest tree species. Air temperature, which drives the metabolism rate, and VPD, which represents the atmospheric water demand, will be the dominant controls on ET during drought years, and not a major factor during non-drought years. Below ground soil parameters such as volumetric water content (VWC) and soil water potential (SWP) will have profound effects on ET during drought years, as water extraction will become progressively more difficult through drying soils, but not a significant regulator of ET during non-drought years. Deep SWP will have stronger control on overall ET compared to shallow SWP during drought (Domec et al., 2010), as it has been documented in the literature that older stands with much more complex root networks will be able to withdraw water from deeper depths and be more resilient to droughts (Coners, 2002).

2 MATERIALS AND METHODS

2.1 Study Site Details

The study site (42^{o}38’7” N, 80^{o}33’27” W; 220 m.a.s.l.) is located 5km northwest of Long Point Provincial Park on the northern shore of Lake Erie in southern Ontario, Canada. The study site is part of the Turkey Point Flux Station, which includes three age-sequence coniferous (78, 43 and 15 year old) and a >90-year old deciduous forest, known as Turkey Point Deciduous (TPD) site. It is characterized as part of the Great Lakes Carolinian ecozone, located in the transition zone between the boreal and broadleaf deciduous forests, making it geographically unique. The site is managed by the Ontario Ministry of Natural Resources and Forestry (OMNRF) and Long Point Region Conservation Authority (LPRCA). Flux measurements at this site commenced on January 5, 2012 and are ongoing. The forest is naturally grown, with some small portions regenerated from abandoned agricultural land and natural sandy terrain. Topography of the region is defined as lightly undulating, not exceeding 2m in depth inclusive of sand dunes and plains. The site predominantly consists of hardwood species including white oak (*Quercus alba*) being the dominant species along with other scattered broadleaf Carolinian species including sugar and red maple (*Acer saccharum, Acer rubrum*), American beech (*Fagus grandifolia*), black and red oak (*Quercus velutina, Quercus rubra*), white ash (*Fraxinus americana*) and yellow birch (*Betula alleghaniensis)* and scattered conifer species, mostly white pine (*Pinus strobus*) (~5% of the trees). The understory is dominated by young deciduous tree species with a variety of forest floor plants. Based on measurements conducted in 2012, average tree height is 25.7 m, average tree diameter at breast height is 22.3 cm with a basal area of 29 m^2 ha^{-1} and stand density of 504 ± 18 trees ha^{-1}. The climate of the region is humid cool-temperate with warm summers and cold winters. A concise list of physical site details can be found in Table 1.

The 30-year (1981-2010) mean annual air temperature is 8.0^{o}C and the total precipitation is 1036 mm, estimated using weather data measured by the Ministry of Environment and Climate Change Canada at Delhi, Ontario (25km northwest of the study site). Approximately 13% of the precipitation falls as snow during the year.

The soils predominantly consist of Brunisolic Gray Brown Luvisol sandy soils with a soil organic layer reaching a depth of 6 cm. The soils have low to moderate water holding capacity, resulting in well-drained conditions. The root system reaches down to at least a 1m depth with the bulk of the root system residing in the top 0-30cm soil layer, with a root biomass of 1.5 g m^{-2} in the 0-15 cm range and 0.73 g m^{-2} in the 15-35 cm range (Khalid, S, unpublished data, 2015). Further specific details on the soils at the Turkey Point Flux Station are explained by Arain and Restrepo-Coupe (2005).

2.2 Water flux, meteorological and hydrological measurements

Water vapour fluxes were continuously measured using the closed-path Eddy Covariance (CPEC) technique from January 2012 to December 2016 (Baldocchi et al., 1988). The EC system comprised of a three-dimensional sonic anemometer (CSAT3, Campbell Scientific Inc. (CSI)) and a closed-path infrared gas analyzer (Li-7200, LI-COR Inc.)

with an analyzer interface unit (LI-7550, LI-COR Inc.) and a flow module pump (7200-101, LI-COR Inc.). The pump generates an airflow of 15 L/min through the optical path of the infrared gas analyzer (IRGA). A 1m-gas intake tube connects the IRGA to reduce interference from the CSAT3. The EC sensor were installed at 35m height on top of a self-supporting scaffold tower made of galvanized steel, with interior stairs and a flat platform to work on at the top. The turbulent fluxes were measured at 20 Hz and averaged into half-hourly values. The raw data is sent to a field trailer located just beside the scaffold tower and stored on a PC. Measurements are taken year-round at this site.

Meteorological variables were also continuously measured on top of the tower. Wind speed and direction were measured using an ultrasonic heated 2 axis anemometer (model 85000, R.M. Young Inc.) without moving parts during the initial 3 years and standard wind sensor (model 05103, R.M Young Inc) thereafter. Air temperature (*Ta)* and relative humidity were measured by a probe, which contains a platinum resistance temperature detector and a humicap capacitive relative humidity sensor (HMP155A, Campbell Scientific Inc.). The probe is cased in a shield that protects the sensor from radiation interference. A constant supply of air is generated to the probe through an energy efficient continuous blower. Downward and upwelling photosynthetically active radiation (PAR) (solar radiation between wavelengths 0.4 μm and 0.7 μm), was measured with two quantum sensors facing opposite directions, upwards and downwards (PQS1, Kipp and Zonen Inc.). The PAR sensor was installed on a metal boom to reduce interference from the scaffold tower. Precipitation (*Ppt*) was measured in an open-area approximately 400m southwest of the site using an all-season heated tipping-bucket rain gauge installed at a height of 2m in cleared forest area (CS700H, Campbell Scientific Inc.). The rain gauge is protected by a wind shield (Alter Wind Screen 24).

Volumetric soil water contents (VWC) were measured at two locations using paralleling 30cm soil moisture probes (model CS650, Campbell Scientific Inc.) and recorded in half hourly intervals. In addition, the soil water potential (SWP) was monitored by a Soil Moisture Block Watermark (model 253-L, Campbell Scientific Inc.) at the same locations. Soil sensors were buried at depths of 2 cm, 5 cm, 10 cm, 20 cm, 50 cm and 100 cm. Using the VWC measurements at various depths, average VWC for the shallow rooting zone (0-30 cm) and entire rooting zone (0-100 cm) were calculated using depth-weighted averages (McLaren et al., 2008); the entire rooting-zone is assumed to be from the 0 cm to 100 cm depth profile for this study. Based on other studies conducted at the Turkey Point Flux Station in the past with similar soil compositions (Peichl et al., 2010), and other standard literature (Oke, 1978), soil moisture levels were considered to be below the wilting point at VWC of 0.01 m^3m^{-3} and at field capacity when VWC reached 0.16 m^3m^{-3} corresponding with SWP values of -1500 kPa and -33 kPa, respectively. These values were used to define water stress and soil water saturation in the subsurface zone.

All meteorological, environmental and subsurface data were sampled at 1-5 second intervals and averaged at half hourly intervals using a data logger (CR3000, Campbell Scientific Inc.).

Relative extractable water (REW) calculations, a re-expressed parameter obtained from volumetric water content, were used as the indicator of drought for the study. The parameter was introduced by Black (1979) and has since been used extensively throughout the literature to assess water deficits in the soil matrix (Bréda et al., 1995; Granier et al., 2007; Gartner et al., 2009; MacKay et al., 2012). For this study, a value of 0.4 for REW was used as a threshold for water stress, coinciding with a study conducted at a site part of the TPFS by MacKay (2012) as well as other studies in the literature (Black, 1979; Granier et al., 2007). For this study, REW was calculated as follows:

$$REW = \frac{\theta - \theta_{wp}}{\theta_{fc} - \theta_{wp}}$$

where θ is the volumetric water content (m^3m^{-3}) of the entire rooting zone (0-100cm), θ_{wp} is the volumetric water content at wilting point (m^3m^{-3}) and θ_{fc} is the volumetric water content at field capacity (m^3m^{-3}). As described above, the wilting point and field capacity values used for the REW calculations were 0.01 m^3m^{-3} and 0.16 m^3m^{-3}, respectively, following other studies conducted at the TPFS (Peichl et al., 2010; MacKay et al., 2012). Although REW dropped below the threshold in 2013 and 2014, climate conditions and minimal consecutive drought days did not warrant drought conditions, therefore are not being considered as drought years for the interest of this study

2.3 Data analysis, processing, quality control and gap-filling

All data were quality controlled, cleaned, and gap-filled. Fluxes, meteorological and subsurface data were run through a cleaning method for invalid data and outliers. Firstly, thresholds were set where any data outside of norm values were excluded for all variables where then the data was visually investigated in detail to exclude any further unrealistic or false values. After data quality control, gap-filling was conducted using protocols outlined by the Fluxnet Canada Research Network and Brodeur (2014). Gaps in latent heat (LE) and hence evapotranspiration fluxes were filled using an artificial neural network in Matlab that utilized net radiation, soil temperature at 5cm, wind speed, volumetric water content in the 30cm soil layer and vapour pressure deficit. Any remaining gaps or missing LE data were filled following Amiro et al. (2006) by windowed linear regression and/or windowed mean diurnal variation approaches. Larger gaps in meteorological data were filled using cross-referencing data from other sites. For example, in 2014, a large gap in wind speed data was filled using data from the nearby coniferous site planted in 1939 due to malfunction of wind sensor from October 31st to December 5th. In addition, for January 2014, missing precipitation data were gap-filled using data from the Environment and Climate Change, Canada weather station at Delhi.

In order to determine dominant controls on ET, simple linear regression analyses were conducted using annual and growing season ET values with PAR_{dn}, T_a, VPD, Ppt, VWC and SWP. Variables that were included in the regression models were considered significant at p-values that were less than 0.1. In addition a Principal Component

Analysis was also conducted to identify the influence of the control variables on ET. Rain-free daily values of the control variables (when PAR_{dn} >200 μmol m^{-2} and Ppt = 0 mm) were transformed into a vector (eigenvector) and an arbitrary value (eigenvalue) between values -1 and 1, which were then plotted together for each year (Figure 4). The influence of control variables on one another is determined by their closeness to their eigenvector. On the contrary, eigenvectors that are orthogonal to each other means there is no significant influence between the two variables. This can be concluded as the purpose of a Principal Component Analysis is to reduce the amount of data being inputted into a regression through the combination of highly correlated variables into principal components based on their closeness. Running means (7-days) for ET were used to better elucidate the lagging effects on ET and reduce stochastic changes in daily differences for some plot analyses (Tang et al., 2014).

The start of the growing season was defined as when there were five consecutive days where daily average air temperatures surpassed $5^{o}C$ and the end of the growing season was defined when daily average air temperatures drop below $5^{o}C$ for 5 consecutive days (Mitchell and Hulme, 2002). A period in the winter of 2016 was disregarded as an outlier, when 5 consecutive days exceeded $5^{o}C$ starting on March 13 but then remained below $5^{o}C$ for >5 days. Using this approach, the growing seasons durations and days for each of the years of the study period were calculated (Table 2). All data was processed and analyzed using software package Matlab 8.5 (Math Works Inc.).

3 RESULTS

3.1 Climatic and Hydrologic Conditions of the Forested Site

The study site's annual climatic conditions are shown in Table 3 and the seasonal and drought dynamics in Figure 1. There was little variation in the dynamics of daily average photosynthetically active radiation (PAR_{dn}) between years (Figure 1(a)). However, there was a noticeable decrease in daily average PAR_{dn} during the growing seasons in 2013 (760µmol m^{-2}) and 2015 (781µmol m^{-2}) compared to the other years of the study that stayed at 800µmol m^{-2} and above. Mean growing season averages of PAR_{dn} exceeded 820µmol m^{-2} in 2012 and 2016. Maximum monthly mean PAR_{dn} occurred in June of 2016 with a value of 1010.7µmol m^{-2}. Maximum monthly PAR_{dn} corresponded with highest monthly ET only in 2012 and 2015.

The mean annual air temperature of the study-period was 9.8 ^{o}C (16.3^{o}C in the growing season), with all years exceeding the 30-year mean of 8.0 ^{o}C in the region (Table 3). 2012 was the warmest year with an annual average air temperature of 3.8 ^{o}C greater than the 30-year mean value. 2016 was the next warmest year with a 2.6^{o}C higher air temperature than the 30-year mean value. Growing season average air temperature was highest in 2016 (17.4^{o}C) followed by 2012 (17.2^{o}C) with all other years averaging at 15.7^{o}C. 2015 had annual average and growing season average air temperatures similar to non drought years (2013 and 2014), closely resembling the 30-year average, compared to the higher extremes shown in the drought years of 2012 and 2016 (Figure 1(b)).

Vapour pressure deficit (VPD) was highest in 2012, coinciding with the drought during this period (Figure 1(c)), reaching a growing season mean of 0.77 kPa, followed by 0.60 kPa in 2016. The lowest growing season mean of the study occurred in 2014, with a mean of 0.47 kPa. Daytime daily mean VPD during rain free periods exceeded 1.0 kPa for 109 days in 2012, 34 days in 2015 and 63 days in 2016 compared with a total of 61 days for the non-drought years combined over the study. Predominantly, the surges in VPD overlapped with drought conditions as shown by REW (Figure 1).

Annual precipitation was lowest in 2015 with an annual value of 750 mm (540 mm falling in the growing season), which was much lower than the 30-year mean of 1036 mm (Table 3). 2012 (801 mm) and 2016 (908 mm) also had lower annual as well as growing season precipitation than the 30-year average as opposed to the non-drought years 2013 (954mm) and 2014 (991mm), which were slightly less (<100mm) than the 30-year average (Table 3). Significant precipitation events with daily maximum values of 69mm, 61mm and 61mm occurred in May 2013, July 2014 and October 2015 (Figure 1(d)). Highest monthly precipitation occurred in July of 2014, totaling 178mm.

High values of volumetric water content (VWC) occurred throughout the winter months during recharge. The onset of the growing season caused clear declines in the VWC (Figure 1(e)) across all years. This decline was much more abrupt and severe in drought years. For example, daily average VWC in the 0-30 cm and 0-100 cm profiles were the lowest in the drought years in 2012, 2015 and 2016 where $VWC_{0\text{-}30}$ reached 0.024 m^3m^{-3}, 0.029 m^3m^{-3} and 0.024 m^3m^{-3} and $VWC_{0\text{-}100}$ reached 0.023 m^3m^{-3}, 0.028 m^3m^{-3} and 0.023 m^3m^{-3} respectively. In contrast, the non-drought years $VWC_{0\text{-}100}$ did not drop below 0.03m^3m^{-3}. Soil moisture levels for the 0-100cm zone did not fall below the designated wilting point of 0.01 m^3 m^{-3} but fell below 0.04m^3 m^{-3} for 35 days in 2012, 28 days in 2015 and 57 days in 2016, where only 9 days in 2014. Overall, $VWC_{0\text{-}30}$ followed highs and lows of the $VWC_{0\text{-}100}$ but fluctuations were more extreme. There were significant declines in both $VWC_{0\text{-}30}$ and $VWC_{0\text{-}100}$ in the winter of 2014-2015 whereby the soil moisture in both profiles dropped below 0.1 m^3m^{-3} from January 20th – March 14th. This decline in soil moisture may have been caused due to frozen soil upper soil layers due to extreme cold temperatures experienced in the beginning of 2015, whereby soil moisture sensors may have underestimated the VWC.

Soil water potential followed the general trends of VWC in terms of decreasing through the onset of drought. Shallow SWP (5cm) surpassed deep SWP (100cm) values in 2012 (30 days), 2013 (1 day), 2015 (1 day) and 2016 (45 days) during initial stretches of rain-free periods (Figure 3). SWP exceeded -1.0MPa for 23 days in 2012, 3 days in 2014, 66 days in 2015 and 47 days in 2016 in the 50cm zone. Daily maximums of SWP_{50cm} during drought years were -1459 kPa in 2012, -1783 kPa in 2015 and -2212 kPa in 2016. SWP_{50cm} during the non-drought years 2013 and 2014 did not exceed -1.2MPa, reaching maximums -956kPa and -1130 kPa respectively. In all years, maximum SWP values occur in deeper zones of the soil profile (>50cm). Minimum SWP_{50cm} occurred in 2012 reaching -124 kPa. The remaining years reached similar SWP_{50cm} values of approximately ~175kPa.

We used a REW threshold of 0.4 to characterize soil water stress. Subsurface water deficits occurred for 67 days in 2012, 110 days in 2015 and 120 days in 2016. Although REW in 2015 fell below the threshold of 0.4 for more days than 2012, the drought occurred later in the season and did not reach as significant deficits as seen in 2012 and 2016. Subsurface water stress was first reached in July for 2015 while initial deficits occurred in June for 2012 and 2016.

The growing season of all years, with the exception of 2012, started approximately at the same time in mid-April, but extended into varying times ranging from November 12 to December 3. The growing season in 2012 started 31 days earlier than the average of the other four years spanning March 11 through November 2 (Table 2). The longest growing season occurred in 2012, having 236 days, followed closely by 2015 with 235 days, where the shortest growing season occurred in 2014 with 216 days.

3.2 Dynamics of Evapotranspiration

Above-canopy evapotranspiration had similar seasonal patterns for all five years of the study, with annual values ranging from 341 mm to 422 mm (Figure 2&3). The highest annual evapotranspiration occurred in 2012, corresponding with the hottest year during the growing season and third driest (behind 2015 and 2016) of the study period, reaching almost 422 mm annually with 389 mm occurring during the growing season (Figure 3). 2016 followed closely behind 2012 with an annual evapotranspiration of 417mm with 380mm occurring during the growing season. Annual ET only surpassed 400mm in 2012 and 2016. The lowest evapotranspiration occurred in 2014 with an annual total of 341 mm corresponding with the coolest and wettest year of the study period. Evapotranspiration rose significantly earlier in 2012 compared to the other years indicating a much earlier start of the growing season. Monthly ET in March rose above 15 mm in 2012 and 10 mm in 2016, whereas in other years it did not exceed 7.5 mm. July was the month of highest ET for all years with the exception of 2013 whereby August ET was higher. The largest monthly drop in total ET throughout the duration of the study occurred in 2012 from September to October, dipping approximately 34mm.

The Precipitation-evapotranspiration (Ppt-ET) balance exhibited significant differences between years (Figure 4). Evapotranspiration accounted for 53% of the precipitation in 2012 and 46% in 2015 and 2016, representing about half of incoming water leaving the ecosystem for the year compared to approximately 35% during the non-drought years (Figure 4). Although annual precipitation in 2016 was more than 100 mm higher than 2012, the difference in annual ET was only 5 mm. Ppt exceeded ET by a minimum of 380 mm (2012) for all years of the study with the largest excess of Ppt occurring in 2014 (650 mm). The study average of Ppt-ET was 500 ± 150 mm. Drought years exhibited lower Ppt-ET compared with non-drought years by as much as 250mm (2012 compared to 2014). Pronounced declines in Ppt-ET occurred in drought years during the growing seasons, decreasing by as much as 118mm from June to August in 2012. 2012 was the only year where mid-growing season Ppt-ET fell below 100mm. All significant declines in Ppt-ET correspond to drought occurrences as indicated by lows in REW. For example, on August 10 in 2012 where Ppt-ET was at its lowest, REW was well below the drought threshold (0.4) at 0.13 with SWP_{50cm} exceeding -1.0MPa. In contrast, minimal declines in the Ppt-ET balance during 2013 and 2014 indicate minimal limitations on water availability.

Daily evapotranspiration surpassed 4mm for the most days (5) in 2013 compared with 1 day in 2015 and 2016 and 2 days in 2012, predominantly occurring in July. To signify the earlier start of the growing season in 2012, cumulative ET exceeded 70mm on May 16th as compared the beginning of June for the other years of the study. Growing season daily ET was highest in 2016 (1.72 mm day^{-1}) with a study average of 1.56 ± 0.17 mm day^{-1}. Daily maximum growing season ET of the study ranged from 3.65 mm in 2014 to 4.47mm in 2016. Daily maximum ET of the study occurred in 2016 (1.4 mm) when assessing VPD on a daytime-no rain (Ppt = 0 mm & PAR_{dn} >200 μmol m^{-2}) basis.

3.3 Meteorological and Subsurface Controls on Evapotranspiration

Annual and growing season correlations of various meteorological variables on ET are shown in Figures 5 and Table 4. Of the meteorological variables tested only T_a and VPD showed significant correlations (p value <0.1) with ET (Table 4). PAR_{dn} (Figure 5(a)) did not exhibit a significant interannual correlation with ET. Minimal annual differences in average PAR_{dn} (max 40µmol m^{-2}) between years did not have a large effect on overall ET. Annual averages of T_a had a positive effect on overall ET (Figure 5(b)), with 2012 and 2016 having >2^{o}C higher annual mean T_a than all other years coinciding with an increase of more than 40 mm of total ET compared to other years. Annual ET increases with increasing VPD (Figure 5(c)). Similar to T_a, VPD was significantly higher in 2012 and 2016 in comparison to the other years in the study coinciding with the highest annual ET values. With an annual average of 0.57 kPa in 2012 and 0.42 kPa in 2016, ET was significantly higher than the other years that showed combined averages of 0.36 kPa. Delineating the daytime (with PAR_{dn} > 200 µmol m^{-2}), 2012 showed a significantly greater annual average of 0.75 kPa, compared to the combined average of 0.46 kPa for the other years. No significant interannual correlation was found for Ppt and ET (Figure 5(d); Table 4). The largest variation in Ppt occurred in 2014 with 991 mm compared to 2015 with 750 mm. The minimal control of Ppt on ET is evident as annual Ppt in 2012 and 2016 differed by 100 mm whereby no consequential difference in ET occurred (4mm). Interrelated variables $VWC_{0\text{-}100}$ and SWP_{50cm} both exhibited insignificant correlations on an interannual basis with a slightly higher negative correlation observed in $VWC_{0\text{-}100}$ (Figure 5(e) & (f)). No identifiable pattern was observed for higher SWP levels in drought years on annual bases whereby lower $VWC_{0\text{-}100}$ was found in drought years by a minimum of $0.1m^3m^{-3}$ (Figure 5(f)) compared to non-drought years.

Interannual correlations between variables showed significantly different results compared to when winter months were included in the analysis and when separating drought years and non-drought years (Table 4). Drought analysis results show PAR_{dn} having the most significant change, with a growing season r^2 value of 0.59 compared with 0.73 whereby on an interannual basis, Ppt showed the most radical change from a r^2 of 0.45 from 0.024 on a growing season basis. The majority of the parameters became more correlated with ET during a drought-specific analysis. Both shallow and deep profiles for subsurface parameters exhibited higher correlation when non-drought years were excluded from the analysis (p value < 0.01) with the exception of SWP_{100} that showed similar p-values in both analyses and a slight increase in r^2.

Principal Component Analysis results for ET during daytime (10:00 to 16:00) and rain-free periods (Ppt = 0) of the growing season are shown in Figure 6. T_a had the highest influence on ET for all years of the study, with the closest eigenvectors to ET. VPD and PAR_{dn} interchanged for the second greatest relationship with ET for 2012, 2013 and 2015 whereby SWP_{50} was found to have a higher influence on ET compared to VPD and PAR_{dn} in 2014 and similar to PAR_{dn} in 2016. SWP_{50} showed a fairly insignificant control to growing season ET in 2012, 2013 and 2015 with virtually orthogonal relationships. No underlying pattern of correlation on ET was observed for drought years vs. non-drought years through the principal component analysis.

Daily ET was positively correlated with daytime VPD (10:00 – 16:00) during rain-free, photosynthetically active periods (Ppt = 0 mm & PAR_{dn} >200 μmol m^{-2}) throughout the growing season (Figure 7). All relationships showed positive correlations (p-values <0.1) with the strongest correlation occurring in 2012 with an r^2 value of 0.63. The response of ET to VPD for 2013-2016 was relatively similar with r^2 values of 0.43 ± 0.06. The ET-VPD relationship exhibited the lowest slope value in 2012, being the warmest year, encompassing 9 of the maximum VPD values of the study.

Generalized early (April 15 – April 30), during (July 15 – July 30) and late (Oct 15 – Oct 30) -growing season periods were investigated as 2012, 2015 and 2016 had significantly different timing of growing seasons (Figure 8). Correlations between ET and VPD for the drought years concerning different stages of the growing season were positive with mid-summer values exhibiting the highest r^2 values for all years. Correlations between ET and VPD for all stages of the growing season showed similar patterns in 2015 and 2016, having comparable growing season start and end dates, with 2016 exhibiting slightly higher r^2 values. Early and late growing season responses were opposite in 2012 compared to 2015 and 2016; the response of ET to increasing VPD was greater in the early-growing season compared with post-growing season in 2012. This corresponds with the earlier growing season start and end in 2012 compared to 2015 and 2016. A significant amount of the VPD values for mid-growing season in 2012 were greater than 1.5 kPa, with 8 days in 2012, 1 day in 2015 and 2 days in 2016.

Although no significant correlations were found with ET and subsurface parameters, generalized patterns were observed throughout the growing seasons of the study period (Figure 2). Soil water potential became progressively more negative (higher stress on water uptake) as the growing season progressed for all years of the study. 2016 showed the earliest initial increase in SWP. Although max SWP occurred much earlier in 2012 due to the dynamics of the growing season, SWP at varying depths were significantly more negative in 2015 and 2016 by a minimum of -0.5MPa in both years at the deeper levels (50cm & 100cm). During all years of the study, deeper SWP generally exceeded shallow SWP with the exception of periods of rapid increases in ET rates where SWP_{5cm} is found to surpass SWP_{50cm} and SWP_{100cm}. Short-term time frames (~10 days) leading up to SWP_{5cm} surpassing SWP_{50cm} and SWP_{100cm}, 2012 and 2016 exhibited an average of 0.08 mm day^{-1} increase in daily ET. Shallow SWP only exceeded -1.0 MPa in 2016 whereas deep SWP (50cm & 100cm) exceeded -1.0 MPa in all drought years.

Despite the early rise in matric potentials during late June in 2013 and early July in 2014, there were no signals of significant stress from the soils during these years. 2015 showed water-stress latest in the growing season from an evident lack of hydrological input in relation to other years with deep SWP exceeding -1.0 MPa well into the fall. 2016 exhibited the most significant subsurface hydrological stress as well as for the longest time-frame during the growing season.

4 DISCUSSION

4.1 Sensitivity of Evapotranspiration to Climate Dynamics

Currently the available knowledge on how temperate forests respond to changing climate scenarios is limited in the literature (Greco and Baldocchi, 1996; Wu et al., 2013; Ma et al., 2014; Xie et al., 2015). These past studies found that droughts had both positive and negative impacts on forest ET. Total ET in our forest increased during drought years where air temperatures were amplified compared to the cooler non-drought years. This may be partly due to growing seasons commencing earlier in the spring (White et al., 1999; Chmielewski and Rötzer, 2002; Schwartz et al., 2006) with a later autumnal senescence period (Taylor et al., 2007), resulting in an overall longer growing season length. At our site, we found that the growing season in 2012 started approximately 33 ± 3 days earlier than the other years of the study (Table 2) due to higher springtime temperatures. This resulted in an earlier surge of ET, reaching almost 55 mm by the end of April compared to 31±5mm for all other years. Although heightened ET in 2012 may be explained by the earlier commencement of the growing season, 2016 follows general growing season dynamics as the non-drought years (2013 and 2014). In contrast, the dynamics of autumnal leaf senescence are not well understood (Taylor et al., 2007; Xie et al., 2015; White et al., 1999). We observed an earlier onset of autumnal senescence in 2012 compared to all others years (Table 2; Figure 3) as indicated by the largest monthly decline of ET occurring from September to October, dropping 33 mm compared to 24±5 mm for the other years. In contrast, the growing season in 2015 began on average with the other years but lasted on average 18 days longer than the other years of the study. No significant change in ET was observed in 2015 in comparison to non-drought year 2014, which had a much earlier senescence period. A phenomenon exists where longer growing season lengths combined with later autumnal senescence periods can result in excess water depletion, which may reduce early-growing season soil water availability for the following year (Xu et al., 2006). This agrees with our study where the years following the longest growing season (2012) and the latest senescence period (2015) exhibited springtime lows of $VWC_{0\text{-}100}$ in the following years (2013 and 2016 respectively) by at least 0.02 m^3m^{-3} compared with other years. Although 2013 and 2016 exhibited springtime $VWC_{0\text{-}100}$ lows, no significant variance in annual ET was observed between years. Our results show that although growing season length and dynamics can help predict the progression of ET (White et al., 1999), there are more significant factors influencing ET dynamics.

The response of ET to the timing of droughts has been established in the literature. For example, springtime droughts were found to have substantially different effects on ecosystem dynamics compared to droughts that occurred later in the growing season (Kwon et al., 2008; Dong et al., 2011; Jia et al., 2016), which are common in this region (Wullschleger and Hanson, 2006). These findings agree with our study (Figure 1) where the earlier onset of drought in 2012 and 2016 (by 42 days on average) exhibited higher annual ET as compared to a late-season drought that occurred in 2015, whereby ET was significantly less. Although ET in 2015 was much lower than the other drought years, it matched very closely with that of non-drought year 2014 (5 mm difference) which was a year with 240mm more precipitation and a slightly cooler annual mean air temperature of 8.0^{o}C compared with 9.2^{o}C in 2015. The evidence found in this study continues to supports the fact that the early- growing season droughts have significant effects on ecosystem ET. Thus, the timing of droughts is integral to the understanding of how forest ET will react to various climate conditions.

A very interesting occurrence of our study was the multi-year drought from 2015 to 2016, which has not been commonly observed in this region in recent years. This occurrence supports the predictions of changing climate scenarios for this region (Wang et al., 2015). Although the prolonged growing season and fall drought of 2015 may have caused a decline in the available soil moisture for the spring of 2016, the precipitation events did not appear to alleviate the high SWP of deeper layers (Figure 2). The higher SWP values seen in 2016 compared with 2012 could be a result of the multi-year drought as no significant droughts were observed in this region prior to 2012 (Skubel et al., 2016), correlating with less negative SWP values in the deeper layers during the dry spell of that year. The exhaustion of deep soil water through multi year drought may have significant impacts on the physiology (i.e. cavitation and reduced xylem conductivity) for deciduous species (Hogg et al., 2013), risking the possibility of long-term ET-reduction and thus productivity. Future studies of forest ET in regions where multi-year droughts are occurring should be investigating to provide further data on how they affect overall productivity and hydrological processes in forest ecosystems.

4.2 Meteorological and Subsurface Controls on Evapotranspiration

It has been well documented in the literature that higher atmospheric demand during dry periods (due to high VPD) has been found to substantially contribute to higher rates of ET across many forest types, including temperate forests (Greco & Baldocchi, 1996; Granier et al., 2007; McLaren et al., 2008; Brown et al., 2013; Meinzer et al., 2013; Wu et al., 2013). This agrees with the findings of our study as it is well depicted in 2012 and 2016 where daily, growing season and annual VPD levels were highest, coinciding with the highest rates of ET for the study period (Figure 1c & 5). In 2012, a growing season average of 0.8 kPa, followed by 0.6 kPa in 2016 resulted in growing season ET values on average 50 mm higher than other years. Increased VPD were able to sustain high ET rates despite a lack of summertime precipitation in these drought years (Greco and Baldocchi, 1996). In contrast, during non-drought years, where precipitation input was more abundant thus reducing atmospheric demand during the growing season (0.47 kPa in both non-drought years) a much reduced ET rate resulted. This has also been shown in other studies focusing on temperate forests such as one by Nobuhiro et al. (2009) where daily ET values were 1.5 mm higher during dry periods compared to wet periods in a lowland coniferous forest. Integrated over the growing seasons, the droughts increased overall ET in 2012 and 2016. What was interesting in our study was that ET in 2015 was found to be low relative to both drought years and non-drought year 2014. This may have been a direct result of a lack of precipitation combined with low air temperatures similar with non-drought years (Table 3), which may have influenced the lower growing season VPD (0.48 kPa). To further explain this, decreasing Ppt correlated with increases in VPD

(Figure 1) in the spring for 2012 and 2016 whereas 2015 did not follow this trend. Growing season values in 2015 exhibited days with a mean VPD greater than 1 kPa (32±5), that was very similar to 2013 and 2014, resulting from sufficient input of precipitation, thus reducing atmospheric demand. A large difference in PAR_{dn} occurred in 2015 compared with 2012 and 2016, being almost 40μmol m^{-2} lower throughout the growing season. In addition, the monthly average of PAR_{dn} during June-August was the lowest of the study period for 2015. These differences between PAR_{dn} and ET may explain the variance between the drought years (Oren and Patiki, 2001; Brown et al., 2013), despite having a low interannual correlation ($r^2 = <0.1$; Figure 5(b)).

When assessing VPD on a daily timescale, we observed a considerable spike in ET in 2012 early in the year compared to other drought years where the correlation between ET and daytime VPD was higher in the early growing season compared to late growing season (Figure 8). For all drought years, the highest value of regression coefficients were found for T_a and VPD during the peak growing season (July 15 – July 30), as expected. Of the drought years analyzed, 2012 exhibited the highest regression coefficient, possibly attributed to a threshold that has been found between ET and increasing VPD. Several studies have shown that once VPD surpasses a certain level (varies between climate ecozones), ET rates begin to stabilize (O'Brien et al., 2004; Bovard et al., 2005; McLaren et al., 2008) due to physiological adaptations to minimize water-stress thus reducing water loss (Bréda et al., 2006). During the peak-growing season in 2012 (July 15 - July 30), daytime daily mean VPD values surpassed 1.0 kPa all but two days, with some days even surpassing 2.0 kPa. Although the slope of the regression was lower than the other drought years, no evidence of ET rate stabilization was found for the daily time scale during this high VPD period. This finding suggests that at this temperate deciduous forest site, the tree species are able to withstand atmospheric water deficits surpassing 2.0 kPa without any obvious signs of significant stomatal closure; no thresholds for limitations between ET and VPD were concluded for this study.

Precipitation was unable to explain the large overall differences in ET, with an insignificant interannual correlation (Table 4), although the slightly decreasing trend (Figure 5(d)) justifies the aforementioned analyses of wetter years having lower ET. For example, annual ET varied little between 2014 and 2015 (341.4 mm and 346.0 mm) despite the large variation in interannual precipitation (ranging from 991 to 750 mm). This was also found across many forest ecosystems including both conifer (Körner, 2012; Swidrak et al., 2013) and deciduous forests (Wu et al., 2013; Meinzer et al., 2013) where in both cases precipitation was not found to solely influence ET. Although this was found, meteorological variables directly influenced by Ppt such as VPD and soil water content have been found to be dominant controls on ET (Granier et al., 2007). Incoming Ppt decreases the level of vapour deficit in the atmosphere through saturation of the air and reduces air temperatures through latent heat flux and increase moisture content in the soil. For example, a substantial surge of precipitation occurred during the early growing season (May-June) of 2015 (219 mm) compared with 2012 and 2016 (65±2 mm). This significantly reduced the atmospheric demand of water in 2015, resulting in, on average, 30 mm less ET for each of those months (Figure 3). This provides evidence that ET can be highly sensitive to the combined influence of certain meteorological controls in temperate deciduous forests, also found in other ecosystem-based studies such as Restrepo and Arain (2005), and Arain (2003).

In addition to meteorological and environmental controls on ET, the complexity of the subsurface water availability on forest functionality is vital to understanding how an ecosystem will respond to drought. In many forest studies, soil moisture has been found to become the major controlling factor on ecosystem ET during times of drought (Kelliher et al., 1993; Oren and Pataki, 2001; Fisher et al., 2008; Gartner et al., 2009; Wallace and McJannet, 2010; MacKay et al., 2012). In our study, although soil moisture was found to have significant correlations during drought years (Table 4), it was not the most significant. This may be attributed to the lack of stress the trees are facing due to their ability to access deeper water stores (Leuschner et al., 2010). The available water in deep soil layers allows them to remain sensitive to atmospheric demand through prolonged dry spells, which ultimately allows them to maintain high rates of ET (Greco and Baldocchi, 1996; Bonan, 2008; Tang et al., 2014; Skubel et al., 2016). With a changing climate, this physiological adaptation is expected to be more frequent and tree species will become more dependent on tapping these water sources during periods of water stress (Meinzer et al., 2013; Wu et al., 2013). This was found at our study site during drought years where times that daily values of VWC continually decreased, increasing SWP did not hinder ET rates but rather increased them (Figure 2). For example, in 2016 VWC_{0-100} and SWP_{100cm} reached values as low as 0.22 m^3m^{-3} and -1.9MPa respectively, which translated to ET rates of more than 2 mm day^{-1} during the same time frame. This leads to the conclusion that the soil water storage within the 0-100 cm soil matrix combined with the tree's rooting systems ability to obtain this water is able to mitigate the effects of the drought severity seen during this time frame (2012 – 2016).

The aforementioned analysis becomes further apparent when assessing the wilting point thresholds for this forest. Daily ET rates remained normal during times when VWC surpassed the assigned threshold (0.01 m^3m^{-3}), and even occasionally increased. In 2016, from August 29 to August 30, the VWC of the entire rooting zone dropped from 0.026 m^3m^{-3} to 0.025 m^3m^{-3} but daily total ET increased from 2.7mm day^{-1} to 3.2mm day^{-1}. During this same time frame, the SWP_{100} increased by more than -0.05MPa, illustrating that the root systems were pulling water to meet atmospheric demand. Other studies concluded similar results such as in a blue oak (*Quercus douglasii*) dominated forest where effective transpiring activity was reported when SWP dropped well below their assigned SWP threshold of -3MPa (Ishikawa and Bledsoe, 2000). This may allude to the temperate deciduous forest in this study to withstand higher intensity droughts observed in this study timeframe as SWP levels did not drop below -1.6MPa (2016).

When comparing the shallow zone to the deep rooting zone, it is visually apparent that the water at deeper depths has a substantial role in this forest ecosystem, especially during drought years (Figure 2). Although shallow SWP exhibits higher sensitivity to immediate environmental stimuli, the overall matric potential of deep SWP has shown the profound role it has in maintaining ecosystem ET. This defining dynamic has even been evident across many forest ecosystems where the lack of deep rooting networks results in resorting to other hydrological phenomena such as hydraulic redistribution to obtain water (Oliveira et al., 2005; Warren et al., 2007; Domec et al., 2010; David et al., 2013). Overall, levels of soil water availability in drought years did not appear to impose a strong control on ecosystem ET as 2012 and 2016 had highest annual rates of ET but also lowest VWC during the study term.

4.3 Implications of Accessing Deep Water Stores and the Water Budget

The physiological ability of obtaining deep soil water is advantageous for mature broadleaf forests, which makes them resilient to drought comparing to coniferous forests. (Bonan, 2002; Coners, 2002; Giambelluca et al., 2003; Bréda et al., 2006; Burk, 2006;). Although coniferous forests are known to have roots extending deeper into the subsurface, reaching to depths greater than 7m, compared to less than 5m for deciduous forests (Canadell et al., 1996), they are not as effective at obtaining deep soil water (Oren and Patiki, 2001). A coniferous forest under drought showed SWP levels at 1m depth of -0.15 MPa, meaning that water at that depth was not being accessed for water use, or else it would be significantly higher (Brooks et al., 2002). In another study by Coners (2002) a coniferous forest was found to have root zone SWP of only -0.6MPa, where in a deciduous forest, SWP values exceeded -1.2 MPa during drought periods. In our forest,

SWP_{50cm} reached < -1.4 MPa in 2012, < -1.7 MPa in 2015 and < -2.2 MPa in 2016, complementing the aforementioned studies on the effectiveness of deciduous forests to access deep soil water during drought periods. This is also evident in our study where clear gaps between the shallow SWP_{5cm} and deep SWP_{50cm} SWP_{100cm} exist during drought years, while being not so apparent in non-drought years (Figure 2). The largest gap occurred in September of 2016 where SWP_{50cm} exceeded SWP_{5cm} by more than -1.0 MPa, thus exhibiting the pull of water at great depths from the rooting systems well into a dry spell. It is apparent that with a projected changing climate, forests with shallow rooting systems may not be able to cope with water-stress and are susceptible to die-off from lack of water for normal function. Our results demonstrate the importance of deep soil water stores in maintaining ET through a lack of precipitation input, but with persistent drought periods, these stores may not remain at levels to satisfy atmospheric needs.

Predictions of increasing T_a and a larger variability in Ppt in North America have been made by the IPCC (Pachauri et al., 2014). In addition to this, regional climate models for southern Ontario show reduced summer precipitation with increased spring and winter precipitation (McDermid et al., 2015), similar to what is seen during 2015 in this study. Combined effects of larger gaps between growing season precipitation and increasing winter temperatures can have substantial effects to the regional water balance Schlesinger and Jasechko, 2014). Increased springtime temperatures will reduce effective infiltration of snowmelt due to higher ET rates thus reducing the replenishment of soil water stores for the proceeding growing season (Llorens et al., 2010). In our study, this is seen through an analysis of the Ppt-ET balance. During the warmest years of the study, Ppt-ET saw drastic declines in the growing season from the increasing demand of water from the atmosphere. Ppt-ET dropped lower than 100mm for more than 60 days during the growing season of 2012 implying a several month deficit of the water balance. Ppt-ET noticeably declined throughout the onset of the droughts, influenced from higher atmospheric demands and reduced hydrological input, in all drought years (Figure 4). The significant excess of Ppt in comparison to ET for the duration of this study indicates that a large amount of the excess water is contributing to groundwater recharge and streamflow (Brümmer et al., 2010). Although the current time frame shows a significant excess of Ppt, continued increases in T_a with decreased frequent precipitation for consecutive years, ET may inherently be reduced in the long-run through a reduction of deep soil water recharge from reduced infiltration and runoff (Huntington, 2003).

5 CONCLUSION

This study assesses how a 90-year old temperate deciduous forest in southern Ontario responds to varying levels of water availability over a 5-year period to explore the overall effects of drought on ecosystem evapotranspiration. In summary, drought years of 2012 and 2016 had the highest annual ET rates, while 2015 and 2014 showed annual lows with 2013 falling in between. During drought events, T_a and VPD were intensified thus increasing ET rates. Drought years 2015 and 2016 had two of the three lowest annual precipitation levels of the study, demonstrating that temperate deciduous forests are able to exhibit varying ET rates with low hydrological input from precipitation. Meteorological factors such as VPD and T_a appear to play the dominant role in controlling ET. However, assessment of SWP illustrated that access to deep soil water stores maintained forest ET through times when low precipitation was unable to replenish the shallow root zone water stores to meet atmospheric needs. These findings suggest that mature deciduous forests are able to withstand low to moderate-intensity drought events, with heightened ET rates, but may be at-risk if severe or multi-year droughts occurred simultaneously, continuously depleting water stores, thus altering the regional water budget.

6 FUTURE WORK

Our study will help for better forest management in the future with the knowledge of how the climate is expected to impact such forest ecosystems in the region. Future work should aim to broaden the understanding of how forests respond to drought events through assessing both above-ground and subsurface parameters. Specifically, analyzing subsurface dynamics in different age forests will provide understanding on the proficiency of rooting systems through the progression of the lifespan of the forest. This will allow for better model predictions on the resiliency of young forests ET to consecutive drought events. In addition, longer-term studies should be conducted to assess how these forests respond to more severe droughts, lagging effects and cross-comparisons to other forests in different ecozones. Specifically, soil water potential needs to be implemented into additional forest studies, as available data is very limited. As a good indicator of water stress in the soil profile, SWP provides a good proxy, as it includes a biological aspect, for how effective species are at obtaining moisture in drying soils at varying depths (Granier et al., 2007). Furthering the understanding of these areas will increase the effectiveness and feasibility of best management practices for anticipated future climate scenarios, specifically in terms of water loss, in forest ecosystems.

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TABLES

Table 1. Site Specific Physical Characteristics

Stand Parameter	Condition
Location	42° 38' 7.124" N
	80° 33' 27.222" W
Elevation (m)	210.6
Tower Height (m)	35
Dominant Overstory Species	white oak (*Quercus alba)*
Secondary Overstory Species	red oak (*Quercus rubra)*
	sugar maple (*Acer saccharum)*
	white pine (*Pinus strobus)*
	red maple (*Acer rubrum)*
	American beech (*Fagus grandifolia)*
	yellow birch (*betula alleghaniensis)*
Understory Species	putty root (*aplectrum hyemale)*
	yellow mandarin (*Disporum lanuginosum)*
	Canada mayflower (*Maianthemum canadense)*
	red trillium (*Trillium erectum)*
	black cherry (*Prunus serotina)*
	horsetail (*Equsitum)*
	prostrate tick-trefoil (*Desmodium rotundifolium)*
	wood violet (*Viola palmata)*
Maximum Leaf Area Index (m^2 m^{-2})[a]	8.0
Average Diameter at Breast Height (cm)[a]	22.29 ± 14.02
Average Tree Height (m)[a]	25.7 ± 4.77
Stem Density (ha^{-1})[a]	504 ± 18
Mean Tree Basal Area (m^2)[a]	0.0578 ± 0.0133

[a] (Kula, 2013)

Table 2. Growing season length for each year of the study from 2012-2016.

Year	Interval	Length
2012	11 March - 2 November	236
2013	15 April - 23 November	222
2014	10 April - 12 November	216
2015	12 April - 3 December	235
2016	16 April - 21 November	219

Table 3. Mean values of air temperature (T_a) and total precipitation (Ppt) over the growing season with annual values given in parenthesis. The 30-year (1981 – 2010) mean annual values from the Delhi Weather Station are also displayed.

Year	Ta (^oC)	Ppt (mm)
2012	17.2 (11.8)	544 (801)
2013	15.7 (9.2)	653 (954)
2014	15.7 (8.0)	732 (991)
2015	15.7 (9.2)	540 (750)
2016	17.4 (10.6)	491 (908)
30-year	8.0*	1036*

Table 4. Simple linear regression values for total and drought years annual evapotranspiration (ET) against annual average downwelling photosynthetically active radiation (PAR_{dn}), annual average air temperature above canopy 36m (T_a), annual average vapour pressure deficit (VPD), annual precipitation (Ppt), 50cm and 100cm soil water potential (SWP) and growing season average shallow zone (0-30cm) and growing season rooting zone (0-100cm) volumetric water content (VWC_{0-30} and VWC_{0-100}). Annual values are given in parentheses. *Indicates where only growing season averages were available.

	r^2	p-value	Drought r^2	Drought p-value
PAR_{dn} (μmol m^{-2})	0.35 (0.028)	0.30 (0.79)	0.59 (0.73)	<0.01 (<0.01)
T_a (^oC)	0.83 (0.85)	0.031 (0.026)	0.89 (0.83)	<0.01 (<0.01)
VPD (kPa)	0.75 (0.69)	0.060 (0.082)	0.54 (0.78)	<0.01 (<0.01)
Ppt (mm)	0.45 (0.024)	0.21 (0.80)	0.077 (0.062)	0.82 (0.84)
SWP_{50cm} (kPa)	0.039*	0.75*	0.11*	0.79*
SWP_{100cm} (kPa)	0.039*	0.75*	0.011*	0.93*
VWC_{0-30} (m^3m^{-3})	0.63*	0.11*	0.77*	<0.01*
VWC_{0-100} (m^3m^{-3})	0.37*	0.63*	0.48*	<0.01*

FIGURES

Figure 1. (a) Daily mean downwelling photosynthetically active radiation (PAR_{dn}); (b) daily mean air temperature above canopy (36m); (c) daily mean vapour pressure deficit (VPD); (d) daily total precipitation (Ppt); (e) 0-30cm & 0-100cm soil volumetric water content (VWC); and (f) relative extractable water content (REW) at the Turkey Point temperate deciduous forest in southeastern Canada. Shaded areas signify the drought periods as determined by REW values (<0.4).

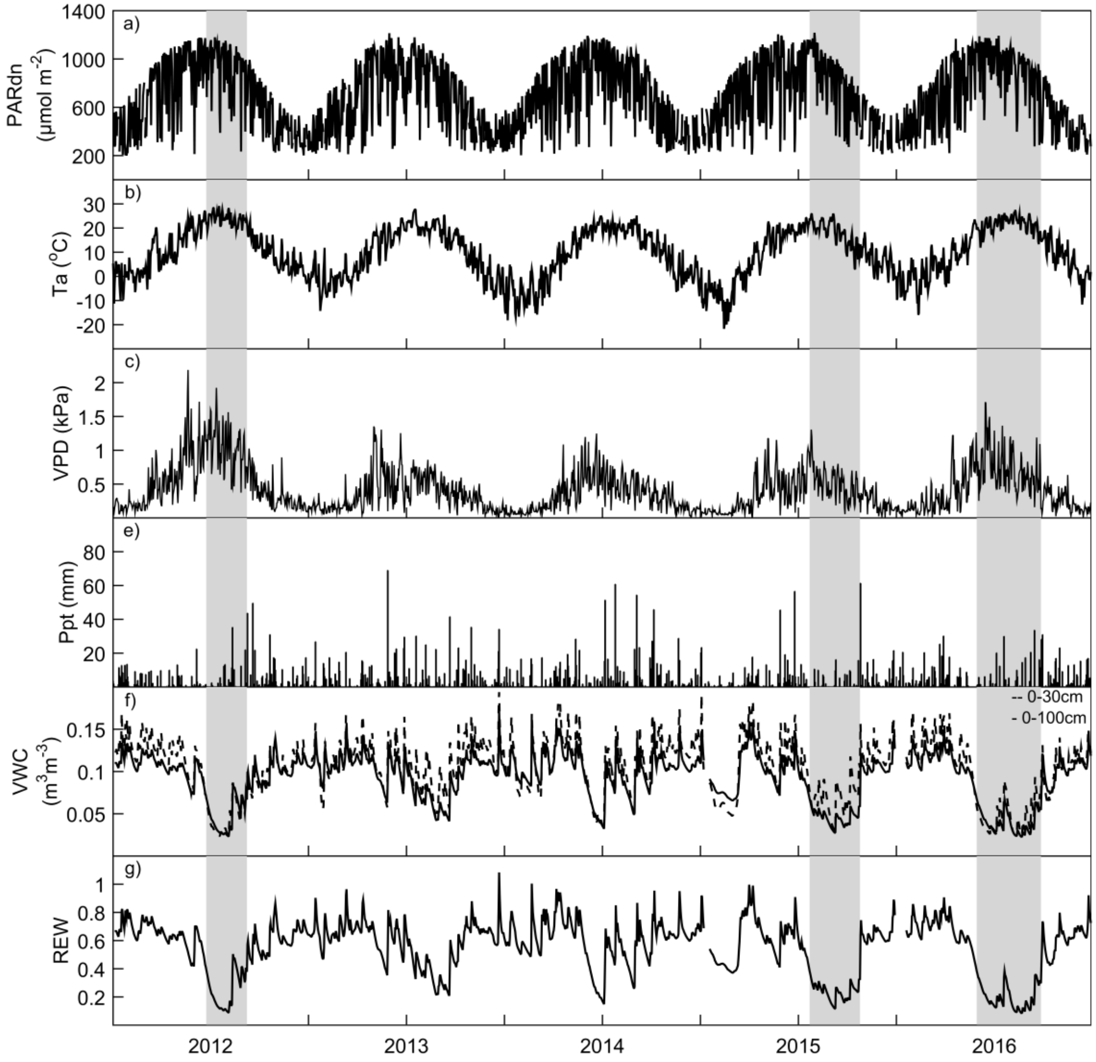

Figure 2. Soil water potential of 5cm (open circles), 50cm (closed circles) and 100cm (compasses) depths and 7-day running average daily evapotranspiration (lines) from April to October in 2012-2016.

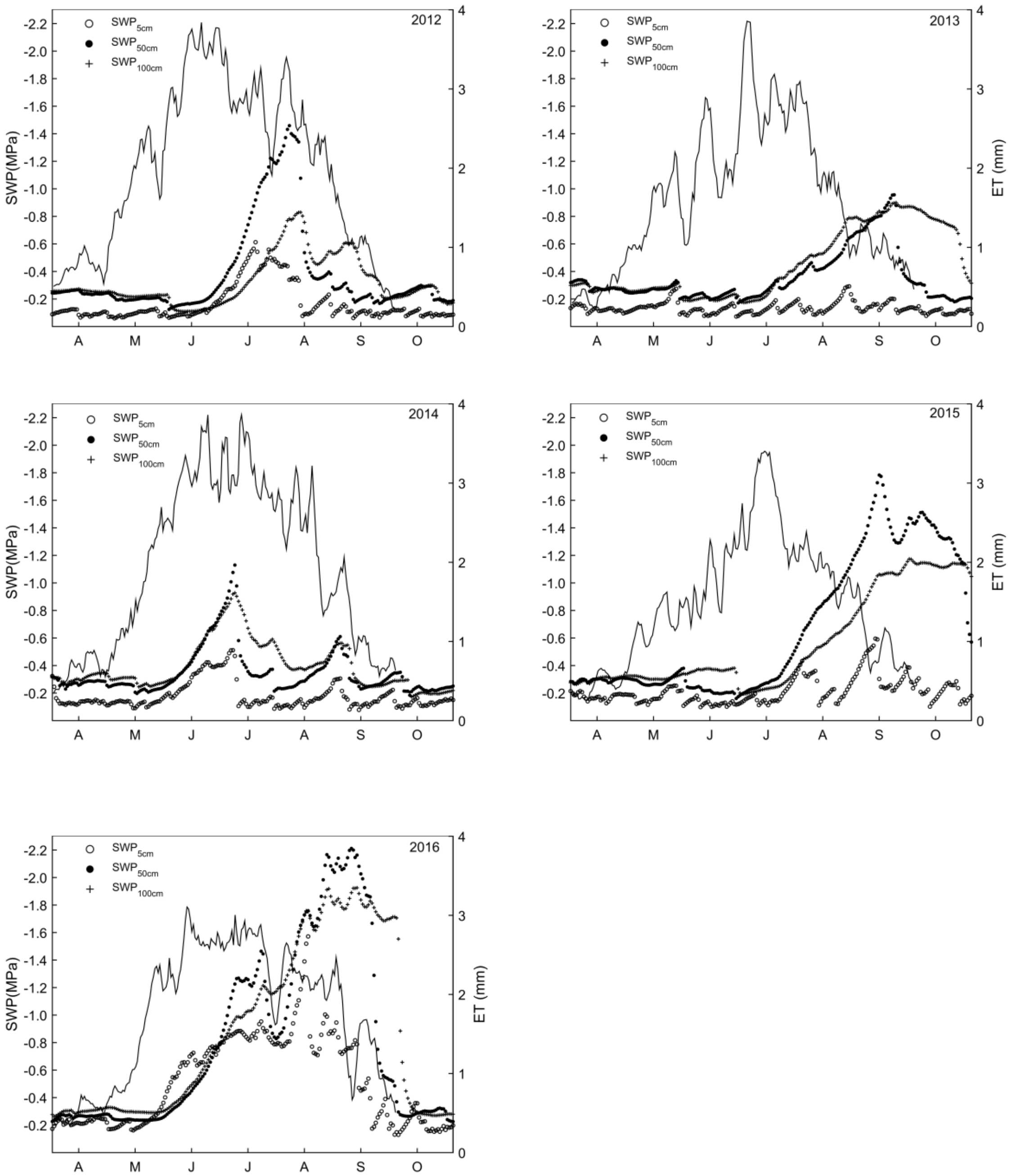

Figure 3. Mean diurnal and seasonal dynamics of evapotranspiration (ET) from 2012 to 2016 (a-e). Cumulative and monthly mean values of ET are also shown (f).

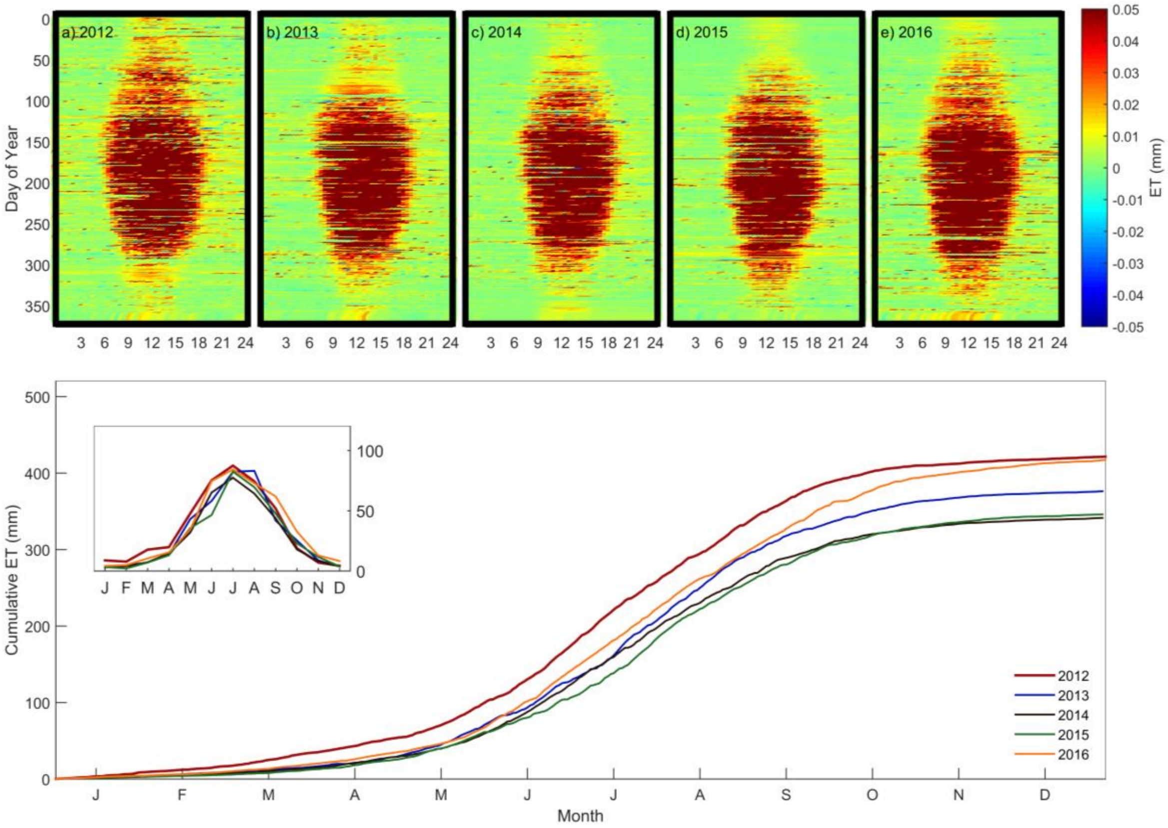

Figure 4. Cumulative values of daily precipitation (Ppt), evapotranspiration (ET) and precipitation – evapotranspiration (Ppt-ET) from 2012 – 2016 in progression through the hydrological year (October – September).

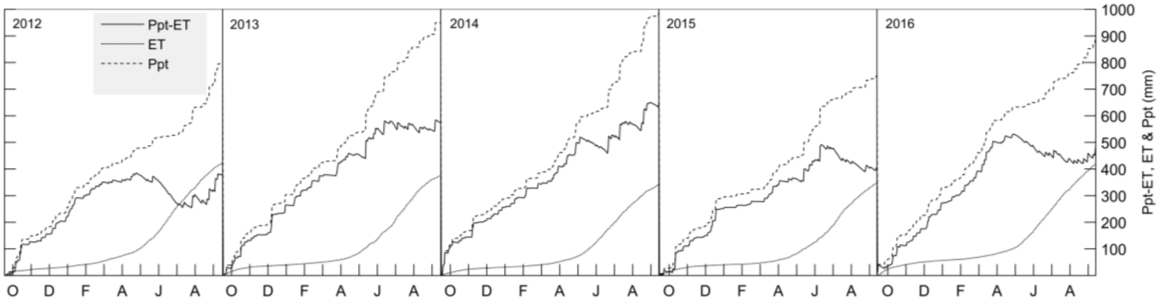

Figure 5. Relationships of annual total evapotranspiration with (a) average downwelling photosynthetically active radiation (b) average air temperature, (c) average vapour pressure deficit (d) total precipitation (e) growing season average 50cm soil water potential and (f) growing season average 0-100cm volumetric water content. Dashed lines represent no significant correlation (p value>0.1). Solid lines represent significant correlations (p value<0.1).

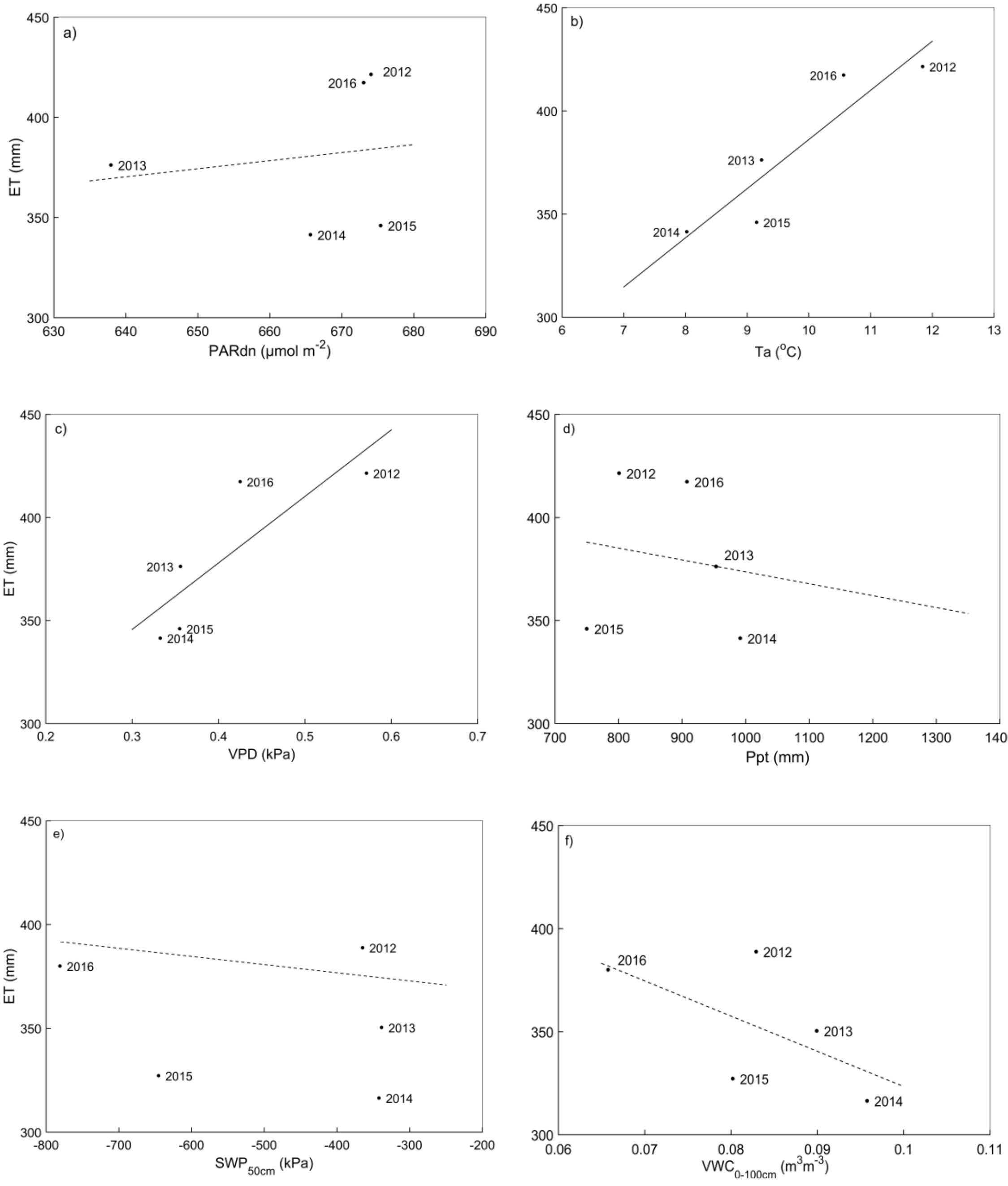

Figure 6. Principal Component Analysis for daily total evapotranspiration (ET) from air temperature (T_a), downwards photosynthetically active radiation (PAR_{dn}), vapour pressure deficit (VPD) and 50cm soil water potential (SWP_{50}).

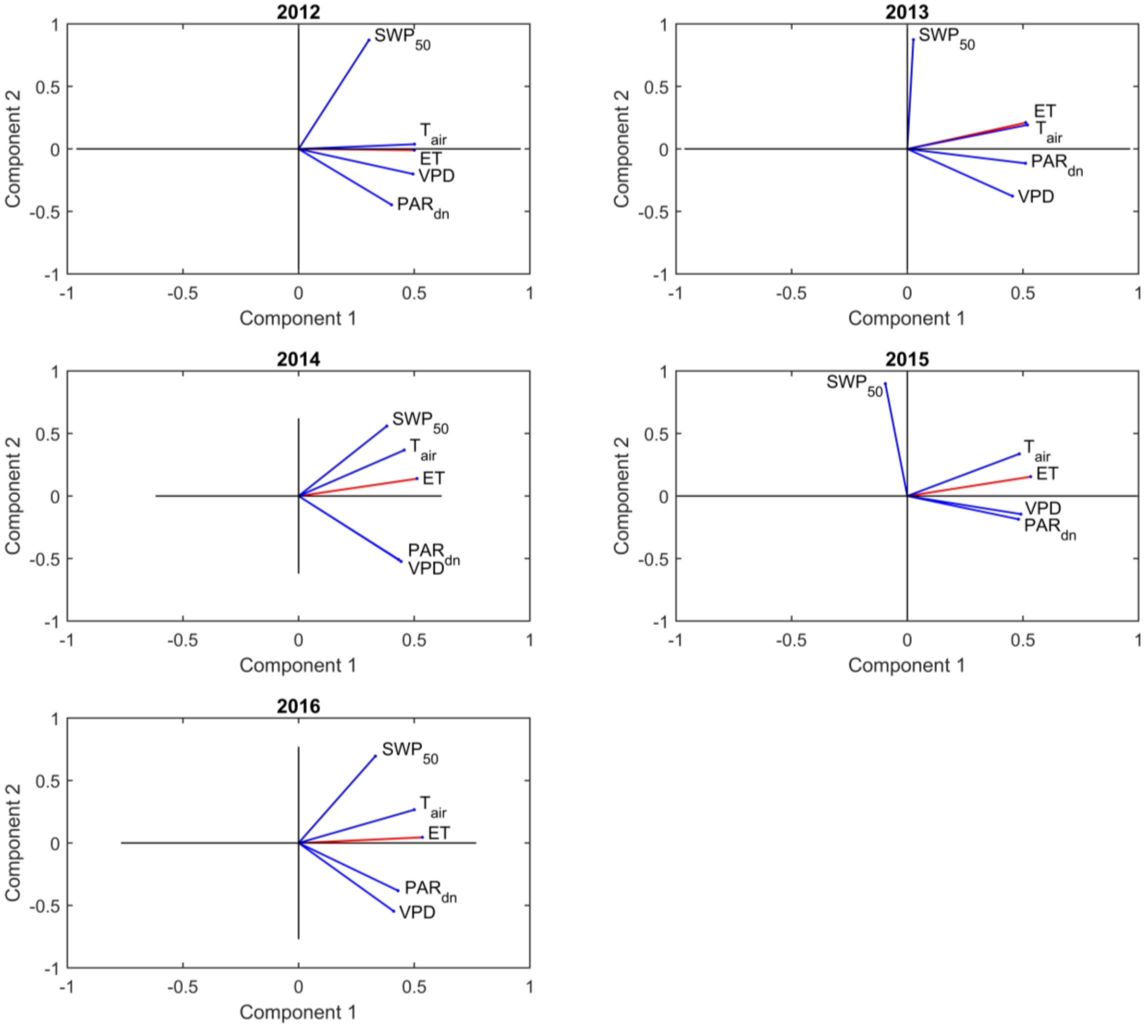

Figure 7. Effect of daily daytime (10:00 – 16:00) average vapour pressure deficit on daily daytime total evapotranspiration for the photosynthetically active part of the day (>200 µmol m^{-2}) with no influence from precipitation (Ppt = 0) during the growing season for years 2012-2016. Closed circles show data in 2012, open circles show data in 2013, closed triangles show data in 2014, open triangles show data in 2015 and compass show data in 2016.

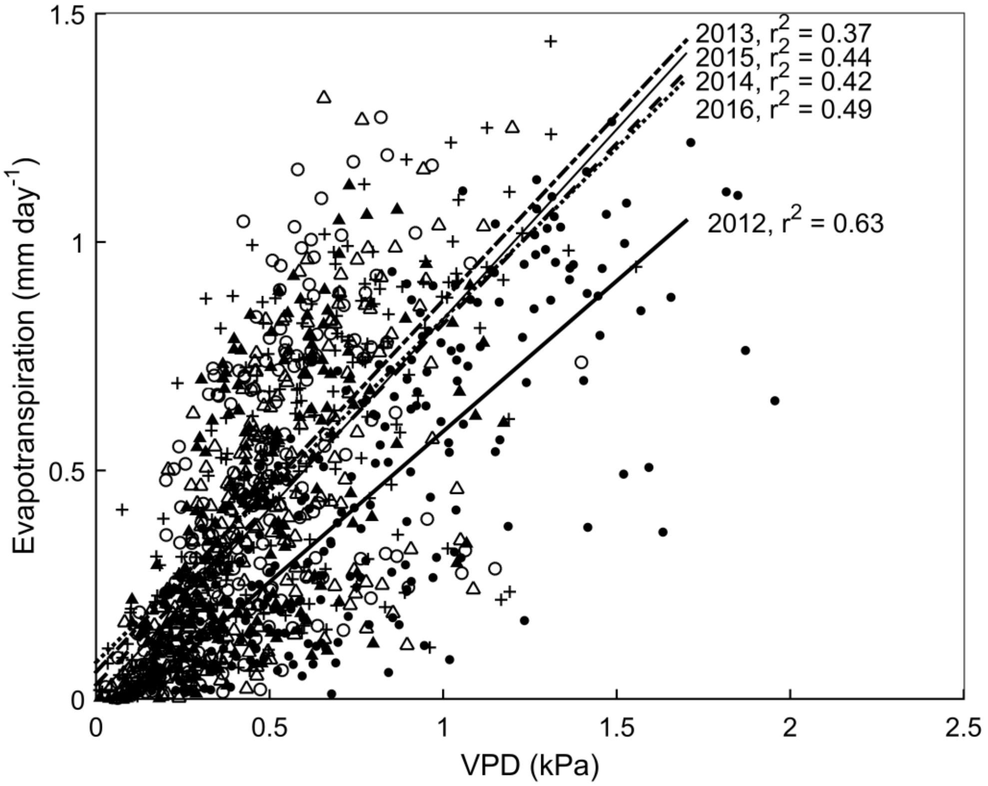

Figure 8. Relationship between daily average vapour pressure deficit and daily total evapotranspiration for the photosynthetically active part of the day (>200 µmol m^{-2}) in different stages of the growing season in drought years 2012, 2015 and 2016. Open circles show early-growing season data from April 15 to April 30. Closed circles show data during growing season July 15 – July 30. Open triangles show late-growing season data from October 15 to October 30.