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ENERGY AND WATER BALANCE IN A DECIDUOUS FOREST IN SOUTHERN ONTARIO

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Abstract

This study discusses energy and water balance in an 80-year-old deciduous Carolinian forest in the Great Lakes region in southern Ontario, Canada. The eddy covariance technique and associated meteorological and soil variables were used to make a year-round measurements of energy and water vapour fluxes from January-December,2012. This site is part of the Turkey Point Flux Station and global Fluxnet. The linear relationship between daily turbulent (sensible heat (H), latent heat (LE)) and radiative fluxes (net radiation (Rn),soil heat (G) and canopy heat storage (S)) has a the slope of 0.75 (intercept of -15.8 Wm^{-2}, and a correlation coefficient, r^2 of 0.93) indicating a 25% deficiency in energy balance closure. The mean value of canopy albedo was 0.16 during the growing season. Maximum daily evapotranspiration (E) rate was 3.8 mm day^{-1} in June, when growing is at its peak in the region. Total annual E was 400 mm, which accounted for 42% of the total annual precipitation of 950 mm. The water storage in upper soil column (1.0 m depth) was approximately 100 mm, indicating that about 450 mm of water was lost from the forest as runoff. Apart from radiation, vapour pressure deficit (D) was the dominant control on E. Maximum value of bulk surface conductance (Gs) was about 18.5 mm s^{-1}. Gs linearly decreased in response to increase in D. The minimum Gs values were recorded when D was maximum, i.e. 3 to 3.5 kPa. Gs also showed high sensitivity to the volumetric soil water content (ϴ), during dry periods, for example the drought event in 2012. In the growing season, the typical value of Priestley-Taylor α ranged between 0.8 to 1.2 with a maximum of 1.8, indicating a wet deciduous forest. However, the LE/Rn relationship showed a linear increase with increasing D with a low (0.26) slope, indicating a conservative response of forest E to atmospheric demand. This study provides insight into energy partitioning, the water balance and their controls in this Carolinian deciduous forest. A better understanding of evapotranspiration

processes and their controls in these forests would help to better quantify water availability at local and regional scales and to evaluate the impacts of future climate change on water resources in the region.

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Acknowledgments

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Chapter 1. Introduction

The deciduous forest region is the most southerly part of Ontario and is situated north of Lake Erie. Common species found in this area include many types of oak and maple, cherry, black walnut, butternut, tulip, magnolia, poplar, black gum, hickory, sassafras and red bud - species commonly found in Ohio, Pennsylvania and the Carolinas (Ontario Ministry of Natural Resources, 2013). Although it covers less than 1% of Canada's land mass, the region is home to more than 25% of Canada's population. This deciduous or Carolinian forest region is important for its ecological, social and economic values. It is rich in terms of species diversity and houses 40% of all of Canada’s vulnerable or endangered species (Ontario Envirothon, 2013). Despite its ecological, social and economic importance, the Carolinian forest ecosystem of Ontario is the most threatened forest type in Canada. It is located in (i) a transition zone between northern cold boreal and southern warm temperate climate regions; and (ii) a densely populated area. Both of these factors increase its vulnerability to climatic and environmental stresses. Therefore, it is important to measure and quantify energy and water balances in this forest ecosystem to explore its sensitivity to seasonal and annual climate variability. This information will help to determine how this forest ecosystem may responsd to future climate change.

Deciduous forests are characterized by phonological stages including leaf emergence and senescence, which can vary in timing and duration between years and alter the exchange properties between the forest and atmosphere (Hutchison et al., 1986). The presence and canopy density of deciduous forest modify a number of factors that influence energy availability and partitioning in the forest. These factors include radiation, temperature, soil water content, and humidity, which together govern evapotranspiration and hence the water balance of the forest. Studying the water balance in a forest ecosystem is biologically more meaningful than traditional climatic metrics. This is done through incorporating the seasonal interactions of energy and water. Measurement of evapotranspiration reflects the availability of biologically usable energy and water, and is therefore an index of forest productivity. When the evaporative demand is not met by available water, the forest ecosystems experiences drought effects. When the water is not used biologically , it becomes surplus and leaves the forest site through runoff or infiltration (Stephenson, 1990). Studying the water balance of forest ecosystems is also important because it not only governs the composition and distribution of vegetation type, but also the climate of a region.

In order to enhance the understanding of energy and water exchange processes in deciduous (Carolinian) forests, long-term year-round measurements of energy, carbon dioxide and water vapour fluxes were initiated in an 80-year-old deciduous forest in southern Ontario, Canada, in January 2012. The site is part of the Turkey Point Flux Station and global Fluxnet. The Eddy Covariance (EC) technique (Baldocchi, 2003; Baldocchi, 2008) was used to measure energy and water fluxes. Evapotranspiration is considered a major component of the water balance (Fisher et al, 2005). Evapotranspiration was determined by energy-partitioning between latent (LE) and sensible heat (H) fluxes . The major source of energy fluxes is the net radiation (Rn). In addition to LE and H (turbulent fluxes), Rn is partitioned into soil heat flux (G) and canopy heat storage (S) (radiative fluxes). Although the data collected through the EC technique is subjected to various corrections (Franssen et al., 2010), it provides a direct measurement of all components of the energy budget(i.e., LE, H, Rn and G+S). The forest energy balance closure

(Rn - G - S = LE + H) was investigated to validate the accuracy of the measured fluxes by the EC system (Foken, 2008). In addition to evapotranspiration, we evaluated other important factors which contribute to the observed water balance of the forest such as soil moisture, bulk surface conductance (Gs) and the role of the soil water content (Barr et al., 2001). The relationship between Gs and Priestley-Taylor α (Priestley and Taylor, 1972), vapour pressure deficit (D), and the ratio of LE to Rn is also discussed.

1.1 Objectives

The objectives of this study are:

1) To measure diurnal and seasonal dynamics of energy and water fluxes and associated meteorological variables in a deciduous (Carolinian) forest in southern Ontario.
2) To estimate the annual evapotranspiration and water budget of this forest.
3) To examine environmental and physiological controls on evapotranspiration in this forest.

Chapter 2. Methodology

2.1. Site Description

The study site (42° 38' 7.124" N, 80° 33' 27.222" W) is located near Long Point Provincial Park in southern Ontario, Canada. The forest is owned and maintained by the Long Point Region Conservation Authority (LPRCA) and known as the Wilson Tract. The study site and the natural heritage woodland properties of LPRCA are shown in Figure 1. Tree species include White Oak (50%), White Pine (20%), soft Maple (10%) and White Ash, Poplar, Black Cherry, Black Oak, Red Oak) and Sassafras make up the remaining 20% of tree species. Trees are approximately

25.7 m high and stand density is 504 ± 18 trees per hectare. Mean stem diameter at breast height (1.3m) is about 22.3cm, while basal area is about 29 m^2 per hectare (Kula, per com.). Leaf area index (LAI) measured by the Plant Canopy Analyzer (model Li-2000, Li-COR Inc.) and Tracing Radiation and Architecture of Canopy, TRAC (developed by Dr. Jing M. Chen's group at the University of Toronto) is 8.0 m^2 m^{-2}.

Soil in the region is classified as a brunisolic grey-brown luvisol (Present and Acton, 1984) with a mid-to-coarse grain size. The thickness of the top organic layer varies from 5 to 10 cm, while the bottom sandy layer may extend to 100 m depth in the area. Soil texture is more than 90% sand with a bulk density of 1.15 g cm^{-3}. Soil is well-drained with low water holding capacity.

Climate of the region is cool continental, with warm summers and cold winters. The mean annual temperature in the area is 7.8˚C with total annual precipitation of about 1010 mm, which is distributed evenly throughout the year (from Environment Canada, 30-year (1971-2000) measurements made at the Delhi weather station). Of the total annual precipitation, about 130 mm falls as snow in the winter months.

2.2. Flux, meteorological and ancillary data collection

Energy and water fluxes were measured using the eddy covariance technique at half- hour time intervals (Baldocchi, 2008; Wilson and Baldocchi, 2000) from January to December 2012. The site was established in January, 2012, therefore, this study report first year of data for this site. The instruments designed to measure fluxes were placed on a scaffold tower, 36 m above ground surface. Wind velocity and temperature fluctuations were measured with a three-dimensional sonic anemometer (model CSAT3, Campbell Scientific, Canada, Inc. (CSI)). Fluctuations in humidity and CO_2 concentrations were measured with an enclosed high performance gas analyzer (model LI-7200; LI-COR, Lincoln, NE, USA). The LI-7200 is a compact, enclosed CO_2/H_2O analyzer that combines the benefits of open and closed path gas analyzers. It is an integrated system designed to provide measurements in harsh weather conditions and environments, with impressively low power consumption.

Temperature and relative humidity were measured using a temperature/relative humidity probe (Model HMP 155A, CSI). Photosynthetically active radiation (PAR) was measured above and below the canopy with a quantum sensor (model PQS, Kipp & Zonen B.V, Netherland). Net radiation above the canopy was measured using a net radiometer (model CNR4 Net radiometer, CSI). Soil moisture was monitored at two locations by soil water reflectometers (model CS650, CSI.), which were buried at depths of 2, 5, 10, 20, 50 and 100 cm. At the same depths and locations, soil matric potential was also measured (model 253-L; CSI). Soil heat flux was measured using soil heat flux plates (model HFT3, CSI) buried 3 cm below the soil surface. All meteorological and soil data were recorded at 0.5 hour intervals using a data loggers (model CR3000; CSI). Precipitation was measured using an all season tipping bucket raingauge (model CS 700H; Hydrological Services Pty. Ltd) in an open area in the forest. It also contains an internal snow detection sensor that is activated when the air temperature drops below 4°C. Precipitation was also measured using all weather accumulation raingauge (model T200B; Geonor Inc.), as well as tipping bucket rain gauge (model TE525, Texas Inst.) about 20 km away at the Turkey Point Provincial Park. In addition precipitation was measured on top of the Turkey

Point Flux Station TP39 site (~20 km away) using a tipping bucket raingauge (model CS700, CSI).

2.3 Gap filling models and data processing

All flux and meteorological data were quality controlled, once data collection from factory-calibrated monitoring equipment was complete. . Gaps were filled following guidelines provided in the Fluxnet Canada Research Network data analysis protocols with some modification as described below. Gaps in meteorological variables were filled using data from three nearby (within 20 km radius) Turkey Point Flux Station sites. We linearly regressed data variables measured this site with similar variable at the other three site and used data for gap filling from the site which gave best R^2 values

These meteorological variables included air temperature (^oC), relative humidity (%), wind speed (ms^{-1}), surface pressure (kPa), soil temperature (^oC), volumetric water content (m^3m^3). Because of limited availability of precipitation data from raingauge at the deciduous site due to late installation of sensor and initial problem associated with power supply and communication connections, we used precipitation data from our weighted raingauge (~20 km away) in this study. Data gaps in precipitation measured by the weighted raingauge were filled using data from tipping bucket raingauge installed besides it. Currently, precipitation is being fully measured at the deciduous site, using heated raingauge and these data will be used in future studies. Before using data from weighted raingauge, we compared precipitation from all four raingauges of the Turkey Point Flux Station and an Environment Canada raingauge at Delhi, Ontario Data from these gauges compared fairly well. Precipitation data from from weighted raingauge was used because it had minimum gaps.

The LE and H data were filled by neural networks using the matlab neural network toolbox (The MathWorks Inc.). For LE, the neural network used 30 network nodes and soil temperature at 5 cm, wind speed, net radiation, soil moisture for top 30cm layer and vapour pressure deficit. The model output was subsequently spike-filtered to remove noisy data that occured occasionally during the non-growing season. Any remaining gaps in LE data are filled using a windowed linear regression between LE and available energy less sensible heat (Rn - G - S - H). This linear relationship is then used to predict and fill missing LE data.

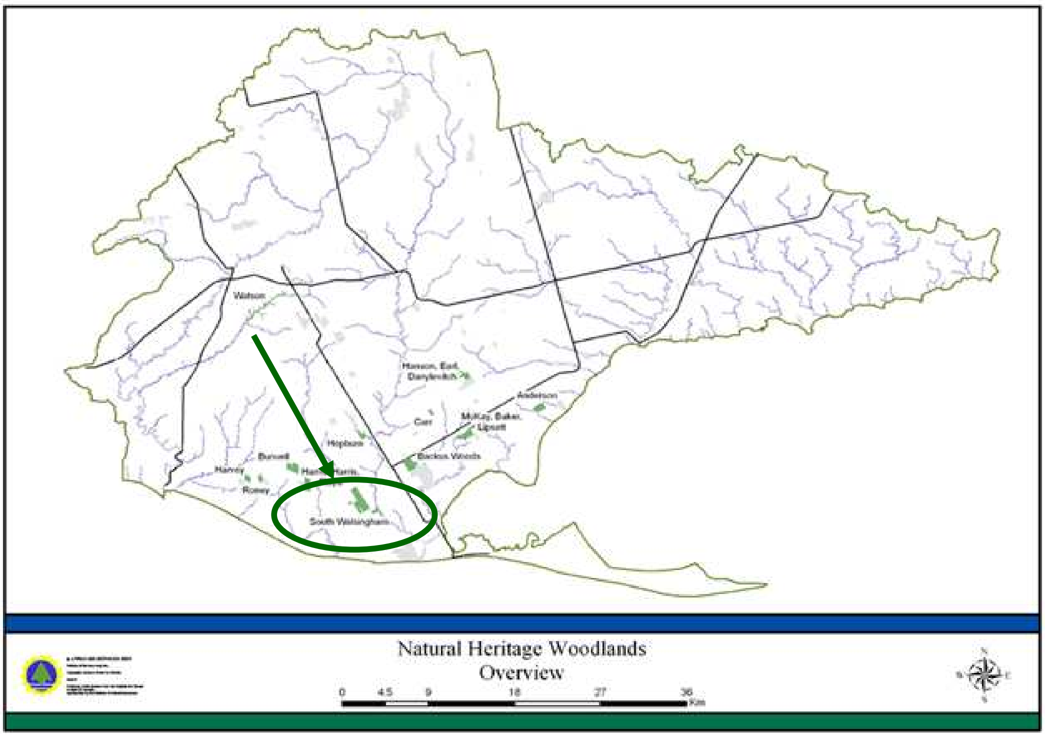
For H, the neural network used 30 nodes, and PAR, net radiation, air temperature and LE data. A similar method was applied to this data as described above for LE for spike filtering. The remaining gaps were filled using a windowed linear regression between available energy (Rn - G - S) and H. This linear relationship is then used to predict and fill missing H data.

2.4. Water balance calculation

The water balance of forest ecosystems is given by

$$P = E + \Delta S_w \qquad \text{Equation (1)}$$

where (P) is precipitation, (E) is evapotranspiration and (ΔS_w) is the change in water stored in the soil column. The term E accounts for transpiration, soil evaporation and water intercepted by the canopy. ΔS_w was calculated by taking the difference between soil water content values at each measurement depth (i.e. 5, 10, 20, 50, 100 cm).



1. Anderson Tract- 35 ac - 14 ha	6. Mackay, Baker, Lipsett, Penner Tract - *105 ac - 43 ha*
2. Backus Woods- 788 ac - 319 ha	7. Hepburn Tract - *78 ac - 32 ha*
3. Harris, Harris, Floyd Tract- 150 ac - 61 ha	8. Harvey Tract - *90 ac - 36 ha*
4. Carr Tract - 50 ac - 20 ha	9. South Walsingham Sand Ridges - *195 ac - 79ha*
5. Earl, Danylevich Tract - 35 ac - 14 ha	10. Watson CA - *307 ac - 124 ac*
	11. Burwell Tract - *100 ac - 41 ha*

Figure 1. The location of study site (Wilson Tract) and other natural heritage woodland properties of the Long Point Region Conservation Authority (LPRCA).

Chapter 3. Results

3.1. Meteorological and Soil Moisture Trends

Daily average trends of photosynthetically active radiation (PAR), air temperature (Ta), and specific humidity (Qs), soil temperature at 2 cm and 5 cm depths (Ts) and precipitation for 2012 are shown in Figure 2(a-e). During the year, the Ta fluctuated from -11 ^{0}C up to 30 ^{0}C. Ta was below 0 ^{0}C at the end of December and it reached as low as -11 ^{0}C in January. In March, substantial fluctuations in Ta occurred when an unusual warm front passed through the area. During this period maximum Ta reached 20.1 ^{0}C. By the beginning of April, the Ta started to increase steadily and reached a maximum of 30 ^{0}C by July. Ts followed a similar trend to Ta, except during the winter months when Ts remained at 0 ^{0}C, due to the isothermal properties of snowpack which insulated the soil and maintained a constant temperature with minimal diurnal variability. Once the snow pack melted, Ts increased and showed diurnal variability similar to Ta. Ts showed maximum variability at lower depths of 2 cm and 5 cm. Qs was influenced by the incoming radiation and Ta. The daily average values of Qs followed the trend of Ta. Maximum daily P values remained below 20 mm from January to May. Most of the large precipitation events were recorded in spring and summer months. From June to December, substantial variation in precipitation was recorded, when few daily P events exceeded 40 mm.. The Maximum daily precipitation event of 50 mm occurred in September. Only 13% of the cumulative P fell in June and July indicating a dry period during peak growing season. Cumulative P from August through the end of the year was 60% of the annual total P while cumulative P from January to May was 30% of the annual P. Annual P values of 950 mm over this ecosystems is slightly less than the 30-year mean annual precipitation of about 1010 in this area. Therefore, the 2012 is considered an average year.

The mean daily volumetric soil water content (ϴ) is shown in Figure (3). ϴ followed the trend of P, especially in the top 2cm of the soil since it is the closest layer to the surface. Shallow (5 cm beneath the surface) ϴ ranged from 0.25 $m^3\ m^{-3}$ in the wet season to about 0.0.3 $m^3\ m^{-3}$ in the drought period (June to the beginning of August). The shallow ϴ varied markedly during the year in response to P events while the deep soil moisture (100 cm) declined gradually during the drought event. In this dry period, the evaporative demand was not met by the available water, hence the forest ecosystems experienced drought condition.

Generally, Θ had approximately the same value from January to May at 50 and 100 cm depths. Θ sharply decreased from May to July because of high rates of E. At the end of July and early August recorded the lowest Θ of the year with a lowest value of 0.023. Then, Θ starts to increase in late August approaching maximum value towards the end of the year (Fig. 3).

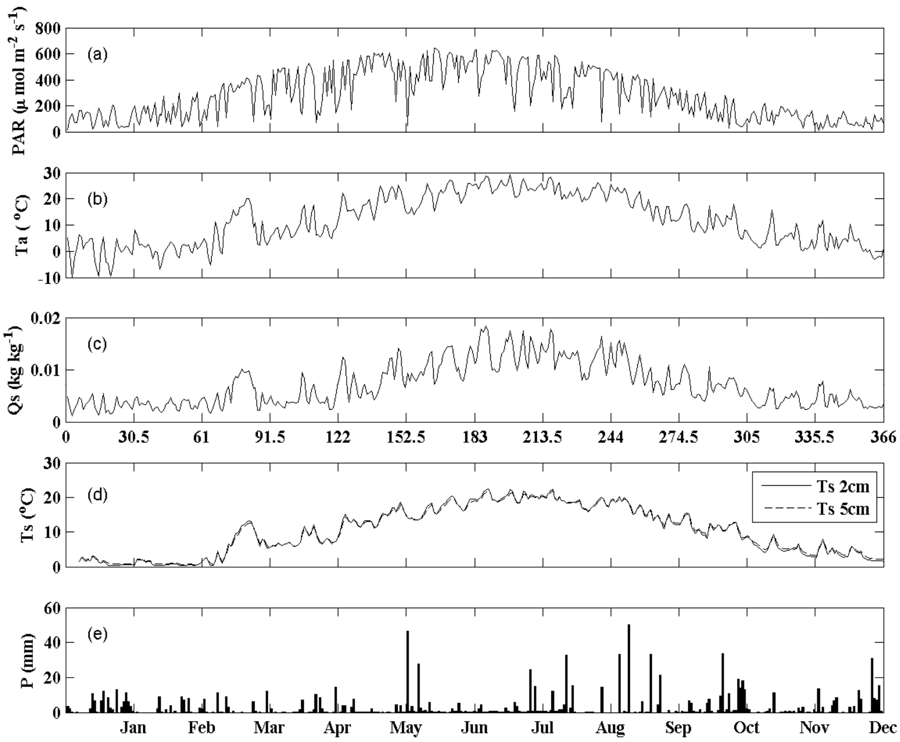


Figure 2. Daily mean values of (a) Photosynthetically Active Radiation, PAR (μmolm^{-2}s^{-2}), (b) air temperature, Ts (^{0}C), (c) Specific humidity, Qs (kgkg^{-1}), (d) Soil temperature, Ts (^{0}C) at the 2 cm depth soil (solid) plot, soil temperature Ts (^{0}C) at 5 cm depth soil (dashed) plot, (e) precipitation, P (mmd^{-1}).

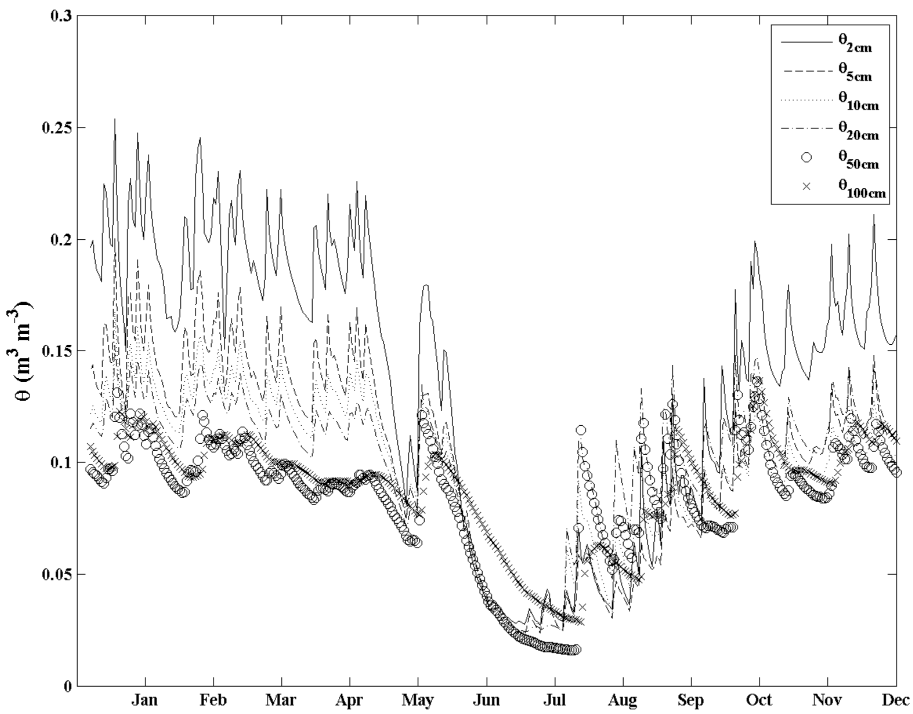


Figure 3. Daily mean values of volumetric water content, Θ (m^3m^{-3}) for soil depths from 2 cm up to 100 cm.

3.2. Energy Balance

Forest energy balance closure was estimated by comparing the sum of turbulent fluxes (H+ LE) and radiative fluxes (Rn - G - S). The relationship between the daily average values of gap-filled (H + LE) and (Rn - G - S) values is shown in Figure 4. The slope of the regression line is 0.75, with an intercept of -15.8 Wm^{-2}, and a correlation coefficient, r^2, of 0.93. Non-closure of the energy balance (deviance from a unity slope) is a common feature of eddy covariance measurements above forest ecosystems (Foken, 2008; Franssen et al., 2010; Restrepo and Arain, 2005). Causes of this non-closure may include the absence of fully developed turbulence during calm conditions (i.e. $u_* < 0.35$), localized large turbulent structures that cannot be spatially averaged by point measurements, measurement errors, and uncertainties associated with Rn and G+S measurements (Foken, 2008; Restrepo and Arain, 2005).

The daily average albedo throughout the growing season (May-October) is shown in Figure 5. The daily average albedo was calculated as a ratio of upward and downward shortwave radiation. The average daily albedo is 16 %, and it falls within the range of albedo's value for temperate deciduous forests (i.e. 15-17%).

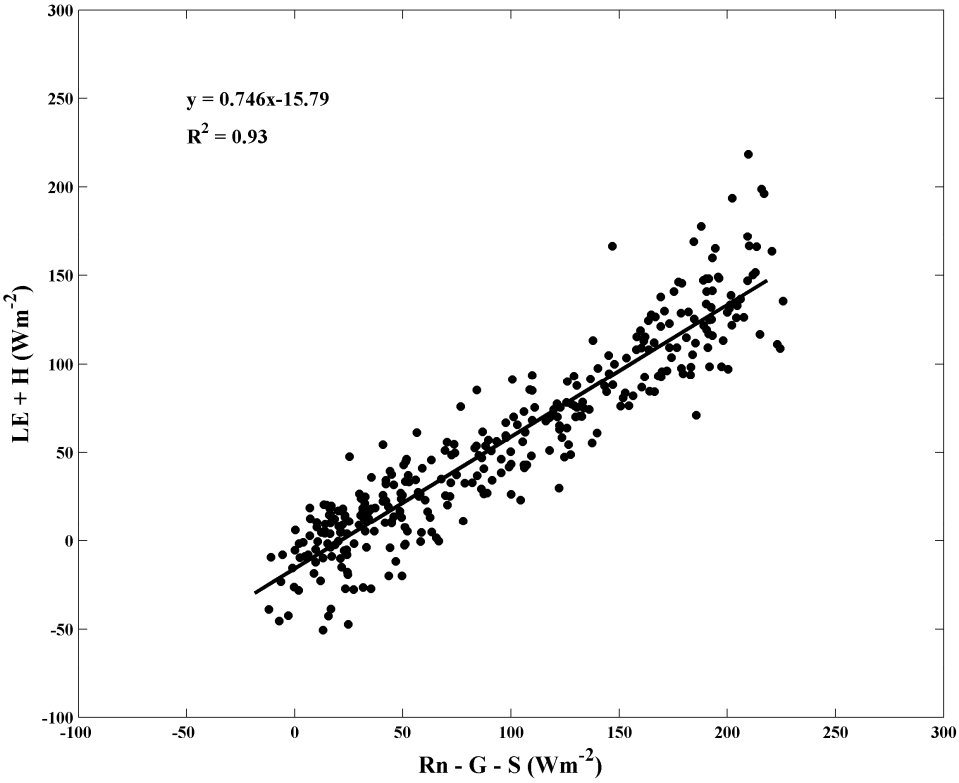


Figure 4. Available energy (net radiation, *R*n (Wm^{-2}) minus the sum of soil heat flux, G (Wm^{-2}) and canopy heat storage, S (Wm^{-2})) plotted against (the sum of latent, LE (Wm^{-2}) and sensible heat (*H*) (Wm^{-2}) fluxes. The linear relationship is fitted by the equation Y =0.746X-15.79, with regression of r^2= 0.93.

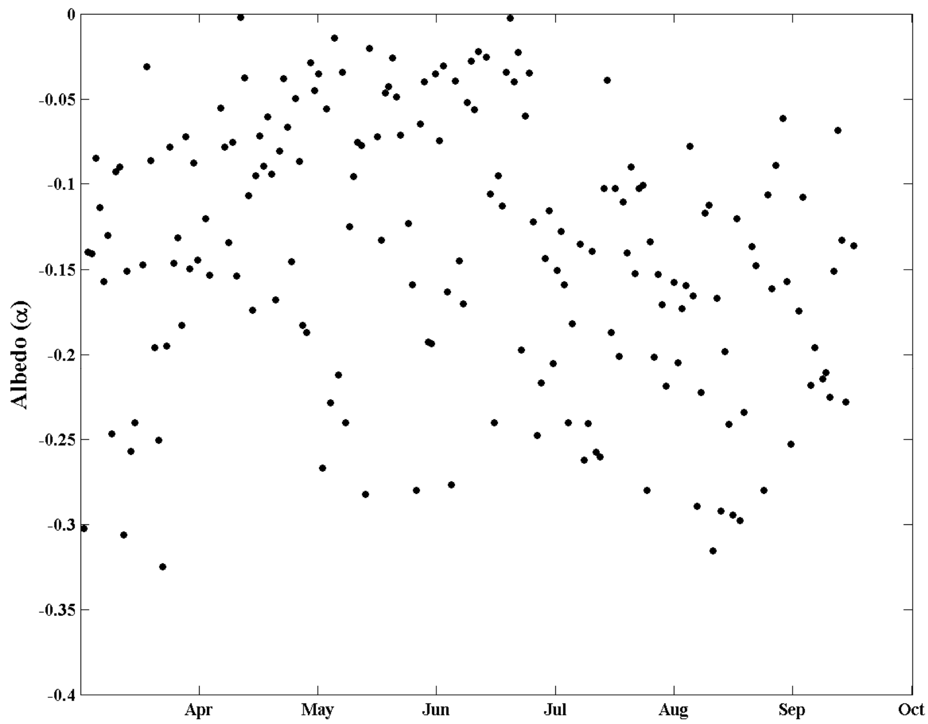


Figure 5. Daily mean values of Albedo (α) throughout the growing season.

The sequence of daily average values of Rn, H, LE, and G + S is shown in Figure 6 (a–d). Solar radiation is the main energy source of the ecosystem and microclimate. During the winter, the minimum Rn value reached -27 Wm^{-2}, while in the summer the maximum value reached 228 Wm^{-2} day^{-1}. On an annual basis the site received about 34,533 Wm^{-2} per year of solar energy. The average of the daily mean values of Rn is 95 Wm^{-2}. From March through the end of August there was a substantial variation in the daily values of the Rn values, in response to cloud cover and precipitation events. More than 90% of Rn was partitioned as H and LE. The remainder of the energy flux was contributed by G+S. This is a trait typical of a mature forest ecosystem (Restrepo and Arain, 2005).

There was a variation in the daily average values of H throughout the year. The average of the daily mean values of H is 27 Wm^{-2}, with a maximum value of 135 Wm^{-2}, and a minimum value of -48 Wm^{-2}. There was a trend in the daily average values of LE, with minimal values in the dormant season and maximum values in the growing season. The LE reached its maximum value in June with a mean daily value of 134 Wm^{-2}. Daily mean values of G+S were the lowest among the energy fluxes. Their values varied throughout the year with a minimum value of -27 Wm^{-2} and a maximum value of about 30 Wm^{-2}. Table 1 summarizes the statistics of different energy is partitioned over the deciduous forest.

Table 1. Summary of different energy partitioned statistics in the study

Fluxes Type	The annual average of the daily fluxes (Wm^{-2})	Maximum value of the fluxes (Wm^{-2})	Minimum value of the fluxes (Wm^{-2})	Annual value of the fluxes (Wm^{-2})
Net Radiation (Rn)	95	228	-27	34,333
Latent Heat Flux(LE)	31	134	-2.0	11,321
Sensible Heat Flux (H)	27	135	-48	9697
Soil and Canopy Heat Flux (G+S)	0.67	30	-27	244.5

The mean monthly diurnal cycles of Rn, H, LE and G + S in the growing season are shown in Figure 7. These data was calculated by daily averaging of half-hourly fluxes over each month. The mean daily positive value of Rn is between 06:00 and 18:30. The lowest daily positive value of Rn was recorded in October (on average between 07:30 and 17:00). In general, there was a variation in the partitioning of different energy fluxes throughout the growing season. Although H and LE followed the pattern of Rn, H was the dominant turbulent flux and followed the trend of the diurnal pattern of Rn in April, May, and October. On the other hand, LE was the dominant turbulent flux in June, July, August, and September.

Daily mean E values are shown in Figure 8. E reached a maximum of 3.8 mm day^{-1} in June, and reaches minimal or zero values in the dormant season. The annual cumulative value of E is 400 mm. The maximum values of E coincided with high values of temperature, net radiation, and vapour pressure deficit. In July, E decreased in response to the sharp decrease in soil water deficit (Figure 8).

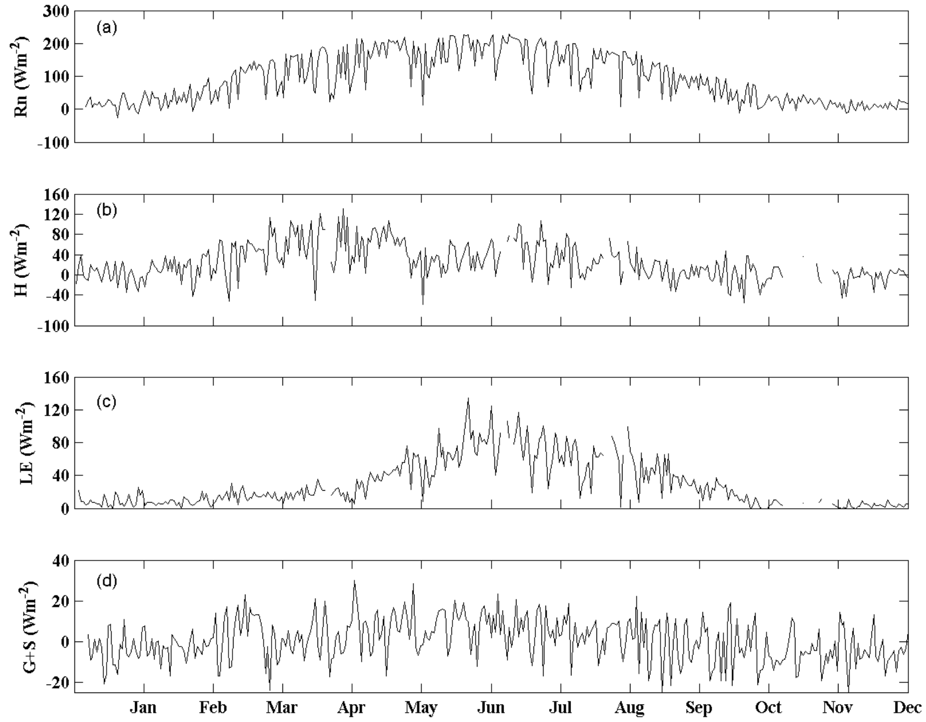


Figure 6. Daily mean values of (a) Net radiation, Rn (Wm^{-2}), (b) Sensible heat flux, H (Wm^{-2}), (c) Latent heat flux, LE (Wm^{-2}), (d) Soil heat flux, G (Wm^{-2}) and canopy heat storage, S (Wm^{-2}).

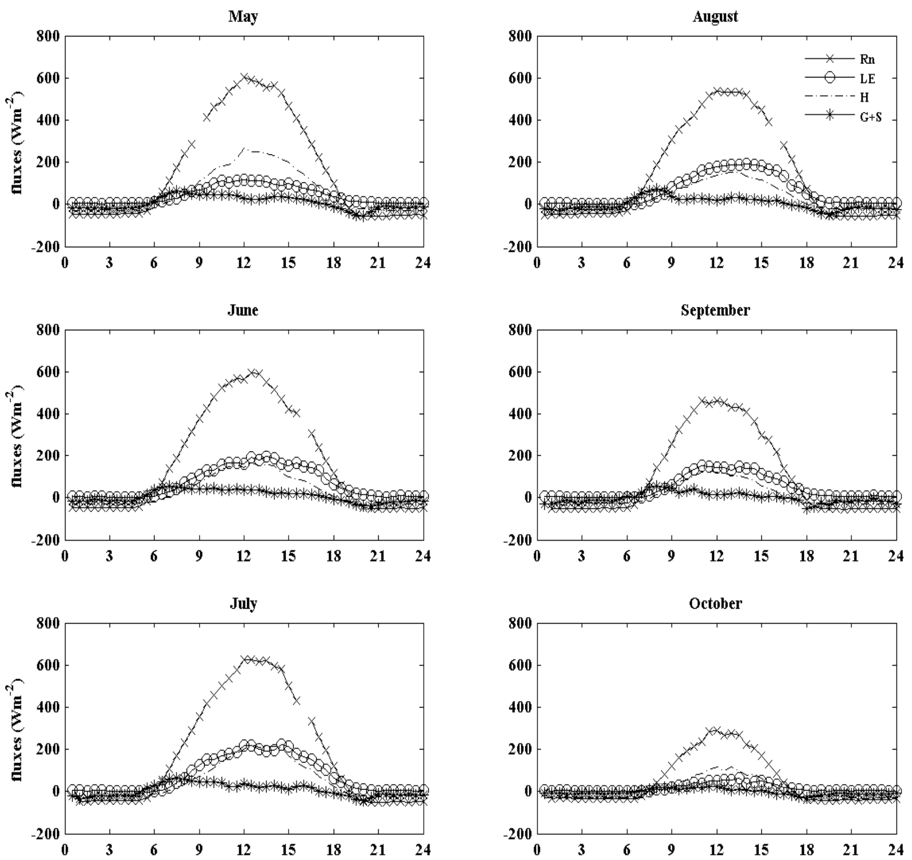


Figure 7. Diurnal values of net radiation, Rn (Wm^{-2}) (cross) plot, latent heat flux, H (Wm^{-2}) (circle) plot, sensible heat flux, H (Wm^{-2}) (dashed) plot, and soil heat flux, G (Wm^{-2}) and canopy heat storage, S (Wm^{-2}) (star) plot.

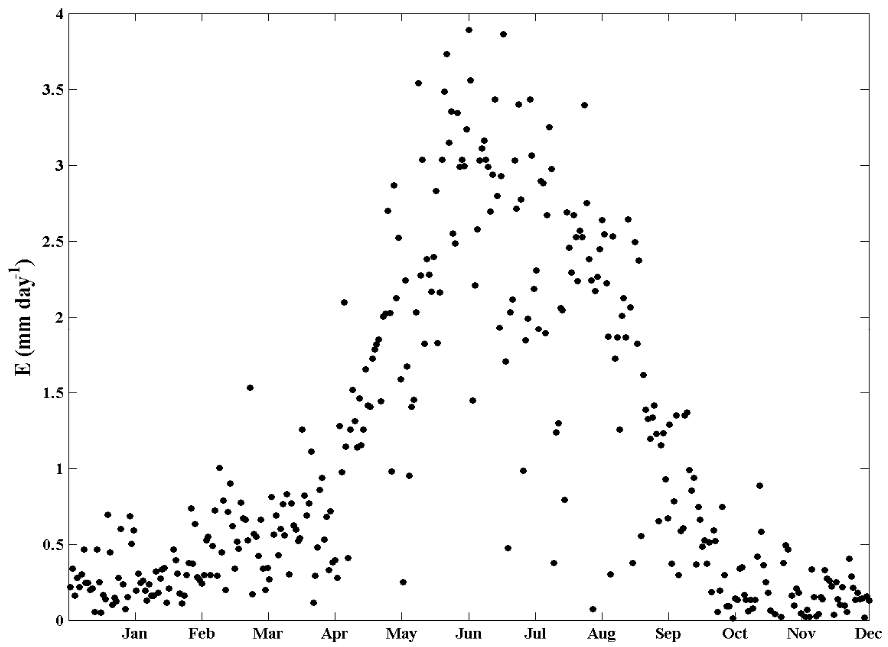


Figure 8. Daily average values of Evapotranspiration, E (mm day^{-1}).

Cumulative values of P, E, and the progressions of P -E and ΔS_w for the year are shown in Figure 9. The cumulative E pattern illustrates a steady increase of E from the beginning of March due to the slight increase in temperature. E continues increasing by the start of the growing season in April with temperature increases. Then, a sharp increase in E occurred from May to July with temperature increases and canopy growth. The cumulative E for the year was 400 mm, which accounted for 42% of the cumulative precipitation of 950 mm. Cumulative E shows a stabilization of values by the end of the growing season.

Cumulative ΔS_w was 100 mm at the beginning of the year; then it decreased from June through the beginning of August. During this period, the temperature reached its maximum value of 30 ^{0}C by = mid July. This happened in response to a drought event that occurred in June and July. Despite the soil water deficit, ΔS_w values balanced towards the end of the year, while P approached near normal value (905 mm vs 30-year mean value of 1010 mm). By the end of the year, cumulative ΔS_w values reached a value of about 100 mm.

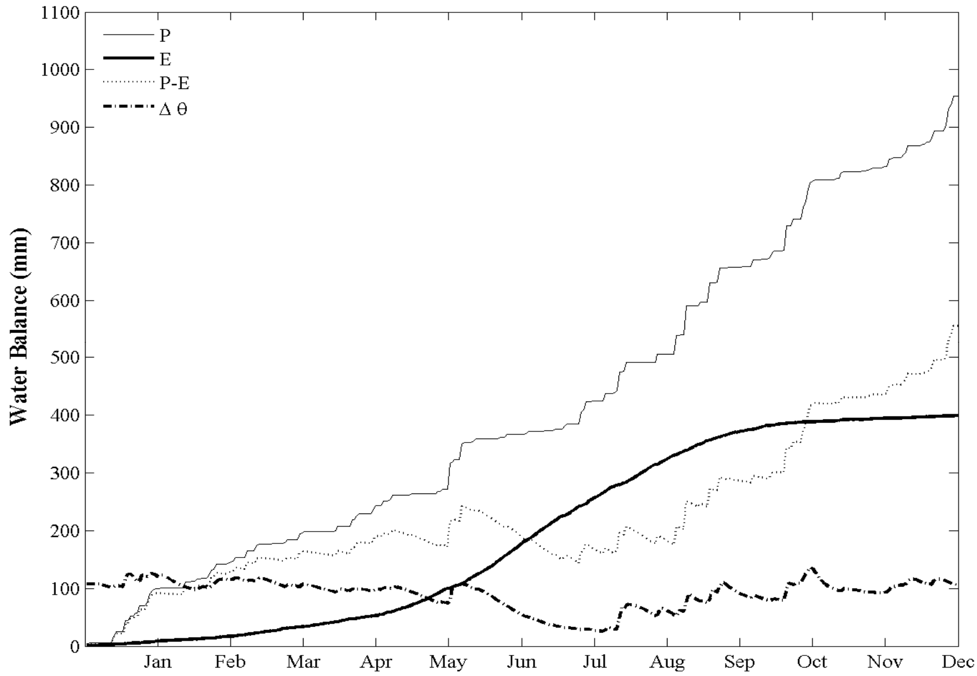


Figure 9. Daily values of cumulative precipitation, P (mm), evapotranspiration E (mm), and precipitation minus evapotranspiration, P – E (mm) and soil water storage change, ΔΘ (mm) throughout the year.

3.3. Environmental Controls on Evapotranspiration

The relationship between the bulk surface resistance (Gs) and diurnal vapour pressure deficit (D) in the growing season is shown in Figure 10. D reached a maximum value of 3.5 kPa and Gs reached a maximum value about 18.5 mms^{-1}. However, Gs values were higher when D values were low. The sensitivity of the Gs response decreased with the increase in D values. The minimum Gs values were recorded for D value between 3 to 3.5 kPa.

The relationship between the daily volumetric water content in the root zone ϴ (m^3m^{-3}) and Gs in the growing season is shown in Figure 11. Maximum ϴ in the root zone reached about 0.14 m^3m^{-3} while Gs reached a maximum value about 18.5 mm s^{-1}. Gs showed variation in response to the increase of ϴ in the root zone. However, there was a decreasing trend in the response of Gs to when ϴ in the root zone values were about 0.03 and 0.06 m^3m^{-3}. This is because of low soil moisture in the June and July. The Gs showed a slight increase when ϴ in the root zone reached about 0.08 m^3m^{-3}. Despite the increase in the ϴ of the root zone, the Gs showed a decreasing trend because the growing season is towards its end.

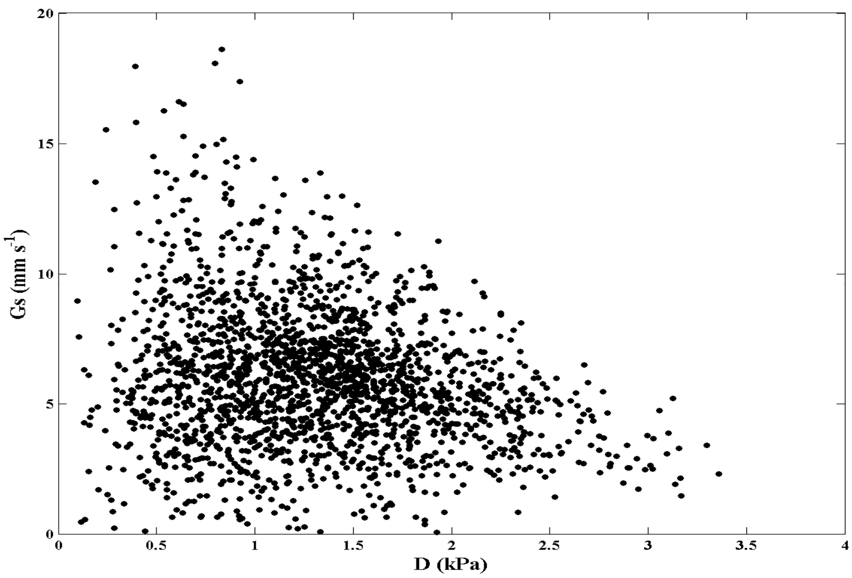


Figure 10. Relationship between daytime mean of vapour pressure deficit, D (kPa), and bulk surface conductance, Gs (mms^{-1}) throughout the growing season.

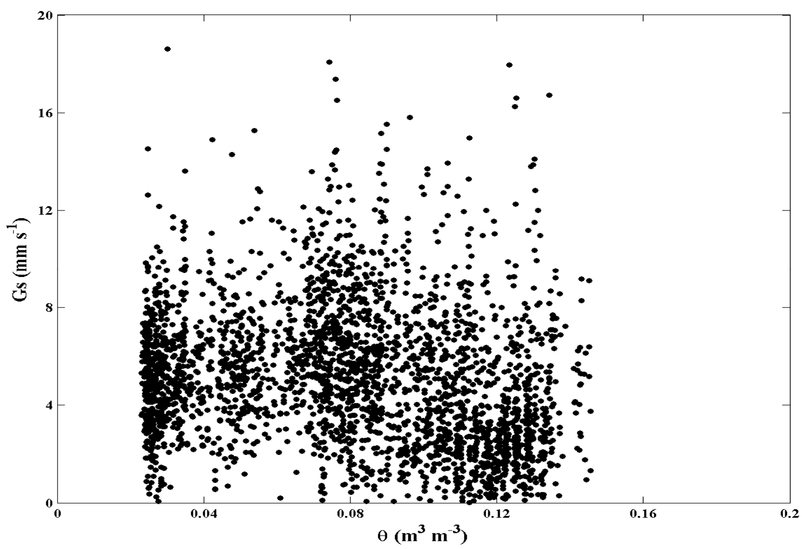


Figure 11. Relationship between volumetric water content, Θ (m^{3}m^{-3}), and bulk surface conductance, Gs (mms^{-1}) throughout the growing season.

The relationship between Pristley-Taylor α and Gs in the growing season is shown in Figure 12. Given that α represents the fraction of surface moisture available for evapotranspiration, the increase in Gs is nonlinearly proportional to the increase in α. In the growing season, the Gs reached a maximum value about 18.5 mm s^{-1} while α reached a maximum value about 1.6. When this forest is assumed to be in hypothetical equilibrium, Gs increases with the increase of α. Under hypothetical equilibrium system, Gs is insensitive to α above a value of 15 mm s^{-1}. Overall, Gs falls well within a range of α between 0.8 to 1.2.

The relationship between the ratio of latent heat flux to net radiation (LE/Rn) and D in the growing season is shown in Figure.13. Understanding the relationship between the LE/Rn and D is important to examine the response of forest evaporation to canopy resistance and factors influencing it such as soil water content. The LE/Rn relationship showed a linear increase with increasing D. The slope of the regression line is 0.26, with an intercept of 0.09 and a correlation coefficient, r^2 of 0.66. Low value of the slope of regression line indicates a slow response of evapotranspiration rate to Gs over the growing season.

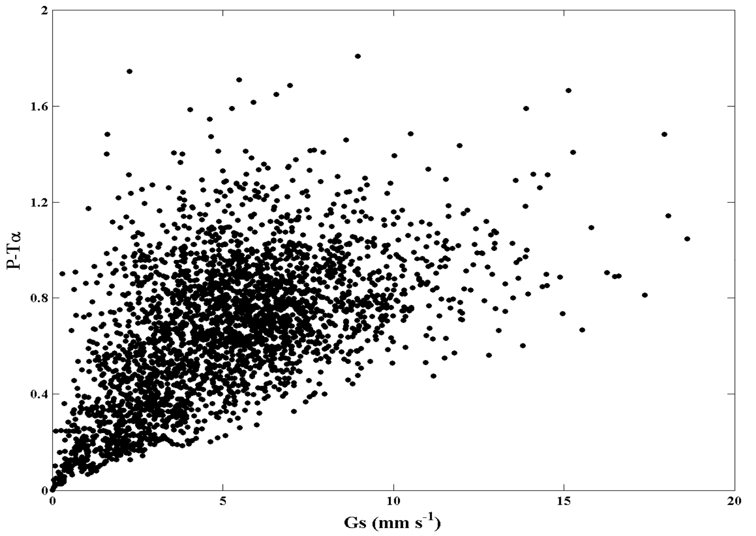


Figure 1. Relationship between bulk surface conductance, Gs (mms^{-1}) and Priestley-Taylor alpha, α throughout the growing season.

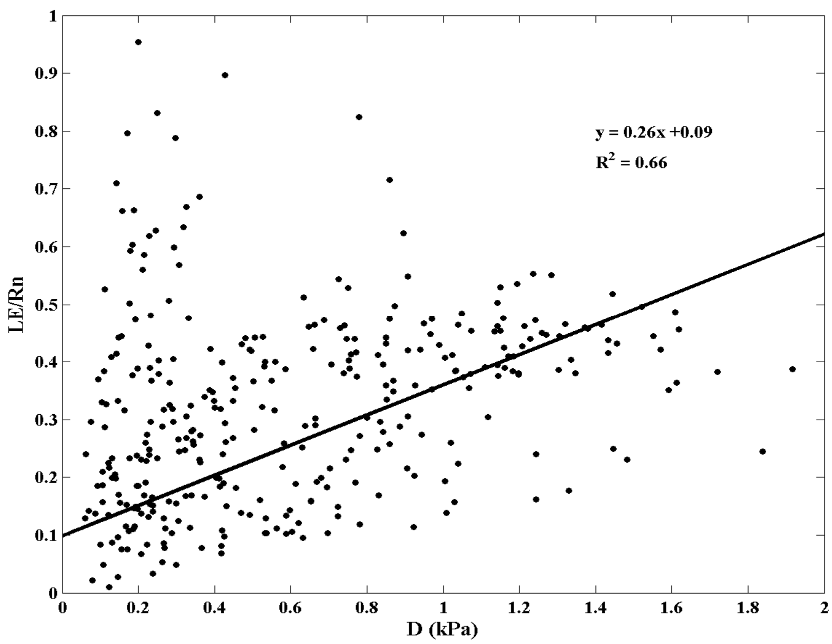


Figure 13. Relationship between daytime mean vapour pressure deficit, D (kPa), and the ratio of latent heat flux to net radiation, LE/Rn throughout the growing season.

Chapter 4. Discussion

4.1. Energy Balance Closure

The energy balance closure of measured fluxes is a common problem related to large-scale convection in forest ecosystems, where measuring sensors are installed at high elevations (Foken, 2008; Franssen et al., 2010). As mentioned earlier, the most common cause of this energy balance closure is weak turbulence or low friction velocities (u*) that result in low sensible and latent heat fluxes (Rocha and Goulden, 2004). Studies have shown that energy balance closure improves during time periods when turbulence is strong or when u* is high (Franssen et al., 2010). This is because convection is not suppressed under unstable conditions. The lack of energy balance closure may not necessarily indicate poor CO^2-flux measurements (Baldocchi, 2008). It may also be caused by (1) differential attenuation of water vapour when air passes through the tube, (2) differences in flux footprints for the energy sensors and the eddy flux instruments and (3) spatial and statistical sampling of net radiation, soil heat and storage fluxes (Baldocchi, 2008). Often radiation is measured at a single point in the forest near the tower while turbulent fluxes are measured for a much larger foot print. The analysis, in this study, shows an energy balance closure value (slope of the regression line) of 0.75. Several past energy balance closure assessment studies across many sites have found that energy balance closure might be underestimated by 10 to 30% (Twine et al. 2000; Wilson et al. 2002; Li et al. 2005; Barr et al., 2006; Oncley et al. 2007; Baldocchi, 2008). Therefore the energy balance closure underestimation of 25%, in this study, is within the range reported in the literature. The fluxes have not corrected to close the energy balance as has been done in past studies, such as Barr et al. (2006).

4.2. Seasonal Patterns of Water and Heat Fluxes

Rn was the main controller of day-to-day variation in LE and H, with high values of turbulent fluxes occurring in the growing season. H reached a maximum before or around noon time, and was typically negative at night. LE reached a maximum shortly after noon time and approached zero at night. Half-hourly LE peaked at 600 Wm^{-2} in May and June while half-hourly H peaked at 300 Wm^{-2} in April. After the start of growing season, H started to decrease with the increase of LE fluxes.

Mean daily values of Rn ranged from -27 to 228 Wm^{-2} with annual cumulative value of 34,333 Wm^{-2}. The inter- and intra-seasonal dynamics of LE and H followed Rn. The seasonal pattern of H was broadly similar to the seasonal pattern of Rn. Annual total H was 28% (9697 Wm^{-2}) of the annual total Rn. H recorded minimum daily mean values of -48 Wm^{-2} in the middle of April and the beginning of May. The maximum mean daily H value of 130 Wm^{-2} was observed in April. The variance in H was slightly larger in the dormant season than the growing season, a pattern that attributed to the increase partitioning of available energy as LE due to the increased water loss in the growing season. Cumulative value of Rn over the dormant season was 8343 Wm^{-2}. LE reached a maximum value of 1109 Wm^{-2} in June, and the annual total LE was 33% (11,321 Wm^{-2}) of annual Rn. Diurnal and seasonal patterns of fluxes observed in this forest are similar to that reported by Wilson et al. (2008) in a deciduous forest. In forest ecosystems, evaporation represents the sum of evaporation from soil surfaces, vegetation leaves through their stomatas and wet plant surfaces after precipitation (Rocha & Goulden, 2004). In this study mean daily values of E ranged from -0.08 to 3.8 mm day^{-1} with an annual cumulative value of 400 mm $year^{-1}$. Our maximum daily E value was similar to many other mature temperate forests growing in wet environments such as 4.4 mm day^{-1} reported by Barbour et al. (2005), 4.0 mm day^{-1} reported by Grelle et al. (1997), 3.7 mm day^{-1} reported by Humphreys et al. (2003) and 3.6 mm day^{-1} reported by Unsworth et al. (2004). The maximum daily E value was also similar to the maximum E value of 4 mm day^{-1} (excluding two extreme days) observed by McLaren et al. (2008) in a nearby white pine forest.

The cumulative E was 42% of the annual P of 950 mm. Annual E in our deciduous forest was similar to other mid-latitude temperate forests in North America. Humphreys et al. (2002) reported annual *E* values of 434 mm and 435 mm for 1998 and 1999, respectively in a temperate Douglas-fir forest in British Columbia, Canada. Anthoni et al. (1999) reported an annual *E* values of 430 mm in 1996 and 400 mm in 1997 over a temperate Ponderosa pine forest in Oregon, USA. Their annual precipitation values in respective years were 595 mm and 188 mm. Restrepo and Arain et al. (2005) reported annual *E* value of 465 mm, from February 2002 to March 2003, in a white pine forest in the area (at Turkey Point), which accounted for 47% of the cumulative precipitation of 996 mm. The study indicated mean water storage of about 100 mm in the soil zone (0-100 cm). It indicates that about 450 mm of water was lost from the forest in terms of runoff, ground waterflow.

4.4. Factors Influencing Evapotranspiration

One of the most important environmental variables to which stomata respond is the vapour pressure deficit (D) between a leaf and air (Addington et al., 2004). The bulk surface conductance reported in our study is derived from the inverted Penman-Monteith equation (Monteith and Unsworth, 1990). It is directly related to the Gs of the individual leaves (Stewart, 1988). Stomata close as D increases and the response is often depicted as a nonlinear decline in

Gs (Addington et al., 2004). Although the mechanism of this decline is not known, some studies suggest that stomatal closure with increasing *D* occurs as a feedback response to some aspect of transpiration (bulk leaf or epidermal) and water loss from the leaf, rather than as a direct response to humidity (Addington et al., 2004). Also, this response supports the feed-forward hypothesis, which states that stomatal conductance (or Gs) decreases directly as D increases as a result of stomata being able to sense an increasing D (Nereu, 2003).

At thesite, Gs was high when D values were between 0.5 and 2.0 kPa. The stomatal responses to atmospheric conditions such as temperature, humidity and incident radiation occur at sub-diurnal and diurnal scales. On the other hand, over longer terms such as days and seasons these responses are influenced or controlled by variation in soil water content and rooting characteristics (Pataki et al., 1998). Gs values showed a high sensitivity toward the variability in soil water content over the growing season. In particular water stress or a drought event in June and July at the site reduced Gs values and exerted a strong control E.

The radiation term of the Penman-Monteith equation expresses the ability of the surface in capturing the incoming radiation, and it sets the lower limit of E if soil water supply is not limiting and if it is not influenced by upwind or overhead conditions (Pereira, 2004). Under such conditions, E is defined as the equilibrium evapotranspiration (Eeq). Based on the Penman–Monteith equation, Allen et al. (1998) defined reference evapotranspiration (E_0) as the amount of water used from a well-watered reference surface. This surface is a hypothetical grass reference crop with specific characteristics. Under such conditions, advection has the least effect on the evapotranspiration. On average, Eeq represents about 80% of E_0, (McNaughton and Jarvis, 1983). Priestley and Taylor (1972) neglected the aerodynamic term and corrected Eeq by a dimensionless coefficient, α (the Priestley–Taylor parameter) as $E_0 = \alpha$ Eeq. Experimental results from several sites around the world, including vegetated surfaces and large water bodies (lake and oceans), gave P-T α values between 1.08 ± 0.01 and 1.34 ± 0.05, with an average of 1.26. Variability or deviations from the idealized 'wet surfaces' are created largely by changes in surface conductance, which is related to changes in leaf area and stomatal conductance. These variations alter P-T α values. Wet ecosystems, with an unlimited supply of water, may have values of P-T $\alpha >= 1.26$, while dry ecosystems, with lower than potential evaporation rates, have values of P-T $\alpha < 1$ (Restrepo and Arain, 2005).

In our forest, P-T α values were sensitive to the surface conductance throughout the growing season. At our site, typical growing season value of P-T α ranged between 0.8 to 1.2 with maximum value reaching 1.8. These values are typical for well watered deciduous forest (Wilson and Baldocchi, 2000; Rao et al., 2011).

At the canopy surface, net radiation is partitioned into LE, H, G and S (Campbell & Norman, 1998). The advection of LE and H can also influence the partitioning of radiation if the canopy is not horizontally homogeneous, the terrain is irregular or if ambient conditions are not at steady-state (Baldocchi, 2008). Because, LE is a dominant turbulent flux in forest ecosystems, the relationship between the ratio of LE and Rn (i.e. LE/Rn) and vapour pressure deficit (D) helps to examine the interaction of LE with the canopy resistance. The LE/Rn ratio increases as D increases, and this reflects the facts that evaporation of a well-watered deciduous forest can increase with increasing D over a wide range of conditions (Baldocchi, 1989). Usually, the slope of this linear relationship indicates a slow or fast response of the evapotranspiration rate to atmospheric demand. In this study,the slow response of the evaporation to D in vegetation ecosystem can be explained by lower coupling between leaves and canopies and their environment (Baldocchi, 1989). It also indicates the sensitivity of vegetation ecosystems to the soil water stress over the growing season.

In our forest the relationship between LE/Rn and D showed an increase in evapotranspiration with increasing D. Because the slope of the regression line is low (i.e. 0.26), it indicated a conservative rate of evapotranspiration over the growing season. This response may be due to drought-induced reduction in soil water, which leads to stomatal closure and reduces canopy evaporation (Baldocchi, 1989). It may be caused by the drought event that occurred at the site in June and July when the forest growth is maximum in the region.

4.5 Significance of the Study

About 15 to 20% of landscape in southern Ontario is covered by the Carolinian deciduous forest and planted conifer forests. Much of these forested areas are comprised of unconnected patchy woodlots in primarily agricultural landscape and their areas range from few ha to 10s of ha (Restrepo and Arain 2005). Energy partitioning and evapotranspiration from these patchy deciduous and conifer forests have important implication for regional water and carbon budgets, through alterations in the boundary layer and atmospheric feedback processes (Barr et al, 2001; McNaughton and Spriggs, 1989). The study provides insight to energy partitioning, water balance and their controls in this Carolinian deciduous forest ecosystem. A better understanding of evapotranspiration processes and their controls in these forests would help to better quantify water availability at local and regional scales and to evaluate the impacts of future climate change on water resources in the region, which is heavily populated and industrialized with rapidly growing water demands. Data and knowledge acquired from this study will help in the assessment of hydrologic models. It can also be used for validation of catchment-scale and regional ecosystem models. Our study as well as flux and meteorological data will help to evaluate the impacts of climate change and extreme weather events on the deciduous forest in the region. It will also help to determine the vulnerability and sustainability Carolinian deciduous forest ecosystem in the Great Lakes region area in the face of future climate change. Indirectly, it will also help in the conversational efforts for the natural (ecosystem) heritage of this area.

Chapter 5. Conclusions

In this study, characteristics of energy partitioning and water balance and their controls were examined in temperate deciduous forest. Key conclusions of this study are given below:

1. Evapotranspiration (E) increased in the growing season and decreased in the dormant season. Maximum daily E rate of 3.8 mm day^{-1} was observed in June when growing is its peak in region.

2. Annual mean value of evapotranspiration was 400 mm, which accounted for 42% of the annual total precipitation of 950 mm. Mean soil water storage was about 100 mm in the soil zone (0-100 cm). It indicates that about 450 mm of water was lost from the forest in terms of surface runoff, subsurface lateral flow or infiltration.

3. Maximum value of bulk surface conductance (Gs) was about 18.5 mms^{-1}. Gs linearly decreased in response to increase in vapour pressure deficit (D). The minimum Gs was recorded when D was maximum at about 3 to 3.5 kPa.

4. Gs and hence E showed high sensitivity to the volumetric water content (ϴ), during dry periods of the growing season. There was a decreasing trend in Gs values when ϴ in the root zone reached low values of about 0.03 m^3m^{-3} during June and July when forest experienced a drought event.

5. In the growing season, typical value of Priestley-Taylor α ranged between 0.8 to 1.2 with a maximum of 1.8, indicating a wet deciduous forest. However, low response of LE/Rn relationship to increasing D values indicated a conservative response of forest water loss to atmospheric demand.

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