

# Atmospheric and soil moisture controls on evapotranspiration from above and within a Western Boreal Plain aspen forest

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## Abstract:

The Western Boreal Plain of North Central Alberta comprises a mosaic of wetlands and aspen (*Populus tremuloides*) dominated uplands where precipitation (P) is normally exceeded by evapotranspiration (ET). As such these systems are highly susceptible to the climatic variability that may upset the balance between P and ET. Above canopy evapotranspiration (ET<sub>C</sub>) and understory evapotranspiration (ET<sub>B</sub>) were examined using the eddy covariance technique situated at 25.5 m (7.5 m above tree crown) and 4.0 m above the ground surface, respectively. During the peak period of the growing seasons (green periods), ET<sub>C</sub> averaged 3.08 mm d<sup>-1</sup> and 3.45 mm d<sup>-1</sup> in 2005 and 2006, respectively, while ET<sub>B</sub> averaged 1.56 mm d<sup>-1</sup> and 1.95 mm d<sup>-1</sup>. Early in the growing season, ET<sub>B</sub> was equal to or greater than ET<sub>C</sub> once understory development had occurred. However, upon tree crown growth, ET<sub>B</sub> was lessened due to a reduction in available energy. ET<sub>B</sub> ranged from 42 to 56% of ET<sub>C</sub> over the remainder of the snow-free seasons. Vapour pressure deficit (VPD) and soil moisture (θ) displayed strong controls on both ET<sub>C</sub> and ET<sub>B</sub>. ET<sub>C</sub> responded to precipitation events as the developed tree crown intercepted and held available water which contributed to peak ET<sub>C</sub> following precipitation events >10 mm. While both ET<sub>C</sub> and ET<sub>B</sub> were shown to respond to VPD, soil moisture in the rooting zone is shown to be the strongest control regardless of atmospheric demand. Further, soil moisture and tension data suggest that rooting zone soil moisture is controlled by the redistribution of soil water by the aspen root system. Copyright © 2013 John Wiley & Sons, Ltd.

KEY WORDS evapotranspiration; boreal forest; *Populus tremuloides*; western boreal plain; soil moisture

Received 10 April 2012; Accepted 29 April 2013

## INTRODUCTION

The clonal nature and spatial distribution of *Populus tremuloides* Michx. (trembling aspen) are drivers that can exert strong controls on evapotranspiration (ET) (Blanken *et al.* 2001) and have the potential to significantly alter the water balance at scales ranging from the stands to landscapes (Wullschlegel and Hanson, 2006; IPCC, 2007). Total ET is defined as the sum of the canopy evapotranspiration (ET<sub>C</sub>) and understory evapotranspiration (ET<sub>B</sub>) (including shrubs, ground cover, and soil surface). ET<sub>C</sub> is often the largest component of both the total evaporative flux to the atmosphere and the overall water budget of forested ecosystems (Schafer *et al.*, 2002).

The sub-humid boreal plain (BP) of Canada's Western Boreal Forest is especially sensitive to changes in water balance. In this region, ET often exceeds precipitation (Devito *et al.*, 2005a), thus changes in the spatial and age distribution of aspen forests and in climate variability could have significant influences on the local to regional water balance influencing forest health and productivity.

ET<sub>C</sub> is significantly influenced by canopy structure, but also varies with other conditions influencing ET, which include radiative transfer of solar energy through the canopy to the understory/ground surface (Chen *et al.* 1997; Blanken *et al.* 2001; Ni-Meister and Gao, 2011), intercepted precipitation (Ahrends and Penne, 2010), understory development (Brown *et al.*, 2010), and within canopy atmospheric turbulence, vapour pressure deficit (VPD), and stomatal conductance, (Hogg and Hurdle, 1997; Dang *et al.* 1997; Blanken and Black, 2004). Chen *et al.* (1993) found large differences in local meteorology such as air temperature, VPD, and soil moisture, among recently clear-cut Douglas fir

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forest, at the clear-cut edge, and within a mature Douglas fir forest sites. Gradients between the factors controlling ET can significantly affect ET fluxes, especially as increasingly larger areas are harvested or altered in some way. For example,  $ET_C$  is linearly controlled by VPD, until a threshold is reached (Hogg and Hurdle, 1997; Hogg *et al.*, 1997; Wullschlegel *et al.*, 1998; Hogg *et al.*, 2000; Kurpius *et al.*, 2003; Bovard *et al.*, 2005), and soil moisture (Wullschlegel *et al.*, 1998; Cinnirella *et al.*, 2002). However, while negative feedbacks occur between VPD and canopy conductance ( $G_C$ ). Stomatal conductance is often unaffected by soil moisture until a deficit occurs, at which point transpiration is limited, as trees close stomata in an effort to conserve water (Wullschlegel *et al.*, 1998; Phillips and Oren, 2001; Cinnirella *et al.*, 2002; Irvine *et al.*, 2002). The relationship between stomatal conductance and soil moisture varies from site to site, and within a stands due to differences in vegetation species type, structure, and soil texture, as different plants respond differently to given soil water states (Roberts, 2000). Barbour *et al.* (2005) also found that canopy wetness was strongly correlated with  $ET_C$ . Within closed canopies, water exchanges between the understory and atmosphere were relatively small due to reduced canopy wetness and incident radiation (Wullschlegel *et al.*, 1998), whereas the opposite was true in open canopies, which were also prone to spatially variable water interception (Unsworth *et al.*, 2004; Barbour *et al.*, 2005). Differences in radiation penetration through canopies due to differences in leaf area also influence the growth and distribution of understory species such as *Rosa acicularis* and *Viburnum edule*, which are prevalent in aspen stands. Proliferation of understory vegetation, including aspen seedlings, which are established through a clonal root system also increase water exchanges between the ecosystem and the atmosphere (Peterson and Peterson, 1992; Blanken *et al.*, 1998).

Trembling aspen may also exacerbate limitations to water balance in the BP because full forests exists over large areas of well-drained soils of a region already with a net moisture deficit ( $ET > P$ ). Therefore, understanding the spatial distribution and heterogeneity of aspen forests, how a forest of aspen is maintained, and impacts to ET and water resources is an important question within this region. Currently, little is known about the sensitivity of ET to variability in the contributions to fluxes within the canopy and the understory in aspen forests of the BP, yet the implications for this research can be significant, given frequent drought periods and infrequent 'wet' periods that occur only about every 20 years (Devito *et al.*, 2005a). Understanding the natural distribution of  $ET_C$  and  $ET_B$  provide important baseline for interpreting the influences of climate change on forest-stand balances, and how natural (i.e. fire) or anthropogenic disturbance of forest canopy influence water resource issues in the region.

The objective of this study is to characterize and quantify the partitioning of ET fluxes within an aspen-dominated upland

within the BP, of Canada's Western Boreal Forest. This study is part of a larger project examining water sink and source dynamics of aspen-peatland-pond ecosystems typical of the BP, a region that persists with a net moisture deficit, climate cycles of infrequent wet periods, deep soils, and large seasonal patterns in precipitation and antecedent moisture conditions. We examine the interaction of vegetation structure, meteorological and soil controls on the ET,  $ET_C$ , and  $ET_B$  within this climate context. It is hypothesized that the evaporation of intercepted water will comprise a significant portion of the daily canopy ET due to the drier atmosphere. Further, the vegetation response to atmospheric demand will ultimately be mediated by hydraulic redistribution of soil moisture from wet areas to the drier aspen stands to meet atmospheric demand. It is hypothesized that driven by diurnal cycles in ET, the clonal roots of the aspen stand will bring moisture to the rooting zone, which will be evident by examining the interactions of soil suction, soil moisture and stand-scale ET.

## STUDY SITE

### *Utikuma region study area*

The study site is located within the Utikuma Region Study Area (URSA), 320 km north of Edmonton, Alberta, Canada (56°4' N, 115°28' W; Figure 1). URSA is situated approximately 150 km south of the discontinuous permafrost zone (Woo and Winter, 1993) within the BPs region of the Western Boreal Forest. The climate within URSA is characterized by cold winters (−14.6 °C) and warm summers (15.6 °C) (Environment Canada, 2005). Annual ET averages 517 mm (Bothe and Abraham, 1993), which is slightly higher than the average annual precipitation of 481 mm at URSA (Environment Canada, 2005). Snowfall in this area averages less than 150 mm yr<sup>−1</sup> and represents less than 30% of the total annual precipitation for the region (Marshall *et al.*, 1999; Devito *et al.*, 2005a). Although a water deficit exists during most years, in the sub-humid BP, annual precipitation surpasses ET every 10–15 years, with significant runoff only every 20 years or more (Devito *et al.* 2005a). Approximately 50% of annual precipitation occurs from June through August, coincident with the period of maximum evaporative demand, and is preceded and followed by a relatively dry spring and fall periods, respectively (Devito *et al.*, 2005a; Petrone *et al.*, 2007).

URSA is characterized by the presence of rolling moraines of low relief, low lying clay plains and coarse-textured outwash plains where glacial overburdens range from 20 to 240 m in thickness and overlie the Upper Cretaceous Smoky Shale Group (Vogwill, 1978). This study was conducted on the moraine portion of the region (Figure 1). The moraine landscape was chosen as these systems are dominated by the presence of vertical exchanges of water (Devito *et al.*, 2005a; Redding and

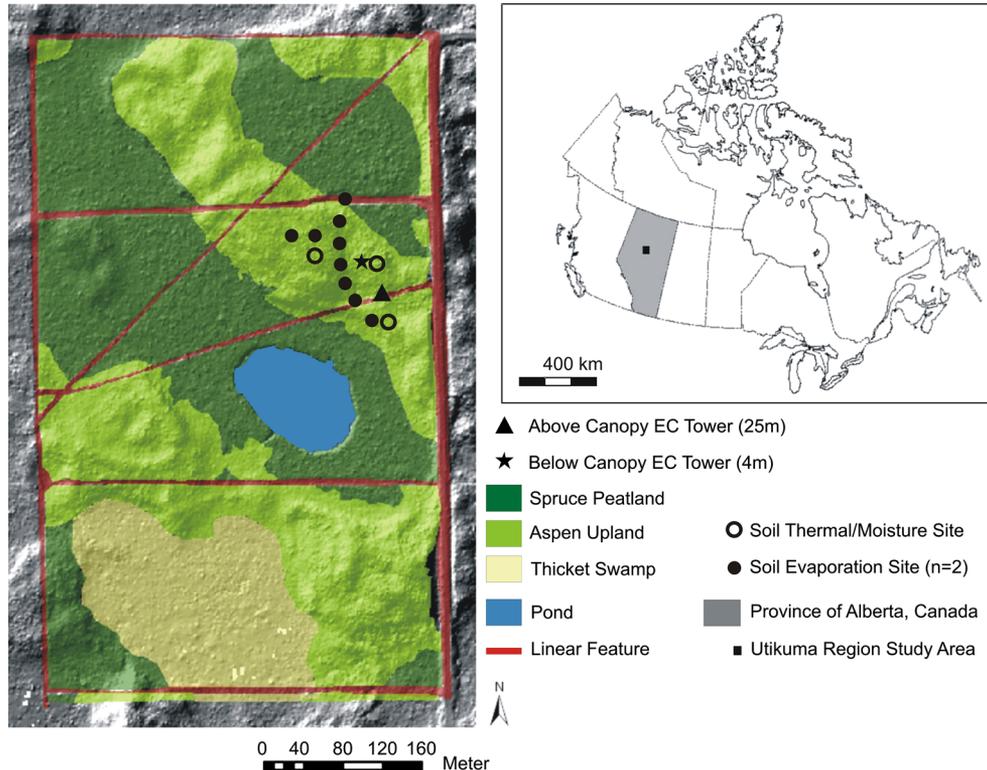


Figure 1. Schematic diagram of the research catchment, location of the instrumentation and dominant vegetation landcover units comprising the catchment, Pond 40, Utikuma Region Study Area (URSA), Alberta, Canada

Devito, 2006; Redding and Devito, 2008) making uplands highly susceptible to fluctuations in climate (Devito *et al.*, 2005b).

#### Study catchment

The dominant overstory species of the study catchment is trembling aspen (*Populus tremuloides*) (covering ~68% of the catchment, determined from remote sensing data), with sporadic clusters of balsam poplar (*Populus balsamifera* L.) in depressions (Figure 1). Fire in 1962 reset much of the ecosystem, resulting in an average stand age in both study areas of approximately 45 years. The understory vegetation at the site is well developed with a shrub layer dominated by prickly rose (*Rosa acicularis*) and low bush-cranberry (*Viburnum edule*) and the herbaceous layer by twinflower (*Linnea borealis*) and sarsaparilla (*Aralia nudicaulis*). The forest floor surface is dominated by aspen litter with negligible amounts of mosses and lichens. Above canopy and within canopy eddy covariance measurements were established on a permanent tower and a mobile tower in an upland area (~7 m total change in elevation across the study catchment; Figure 1) consisting of Gray Luvisolic soils (Soil Classification Working Group, 1998) developed from disintegration moraine deposits, which are typically silt-rich but spatially heterogeneous, with zones of high clay or sand contents (Fenton *et al.*, 2003; Redding and Devito, 2008).

#### METHODS

*Above and within canopy ET measurements.* The eddy covariance technique was used to measure continuous ET fluxes above an aspen-dominated upland stand during the 2005 and 2006 snow-free seasons (DOY 187 to 291 and DOY 123 to 283, respectively). Eddy covariance instrumentation was located 7.0 m above the aspen trees, which ranged in height from approximately 17 m to 21 m, on average. The measurement system consisted of a three-dimensional sonic anemometer (CSAT3; Campbell Scientific, USA), an open-path infrared gas analyzer (LI-7500; Li-Cor, USA) and a 25  $\mu\text{m}$  fine-wire thermocouple, sampled at a frequency of 10 Hz. Mean horizontal ( $v$ ) and vertical ( $w$ ) wind velocities were mathematically rotated to zero following Kaimal and Finnigan (1994), and ET was calculated as described by Webb *et al.* (1980). Mean flux and atmospheric turbulence statistics were computed every 30 min and stored on a CR23X data-logger (Campbell Scientific, USA). Net all-wave radiation ( $Q^*$ ) (NR-lite; Kipp and Zonen, Netherlands), air temperature and relative humidity (HMP45C; Vaisala Oyj, Finland), and photosynthetic active radiation (PAR) (PAR-lite; Kipp and Zonen, Netherlands) were also measured above the canopy on the permanent tower 7.0 m above the tree crown.

Soil moisture and energy flux data were collected at three sites located 20 m, 25 m, and 85 m from the tower

(Figure 1). Soil heat fluxes ( $Q_G$ ) were measured using two heat flux transducers (HFT-03; Campbell Scientific, USA) at each site, buried 0.05 m below the litter fall horizon (LFH)–soil interface. Soil temperature and heat storage in the upper 0.05 m were measured using a thermopile (TCAV-L; Campbell Scientific, USA). Soil moisture ( $\theta$ ) (CS616 TDR; Campbell Scientific, USA; calibrated for study site soils) and soil temperatures (107B Thermistors; Campbell Scientific, USA) were recorded at each site at depths of 0.01, 0.10, 0.30, 0.50, and 1.0 m below the LFH–mineral soil interface. Soil water matric potential ( $\psi_m$ ) was measured in profile at 0.1, 0.3, and 0.5 m using soil tensiometers (SW-033; Soil Measurement Systems, USA). Precipitation was measured 30 m south of the tower over the period using a RM Young Tipping Bucket (52202; R.M. Young, USA) with a series of Tru-Chek metric rain gauge (Edwards Manufacturing, USA) as backup for any missing data points.

A second (mobile) eddy covariance tower was used to quantify  $ET_B$  of the understory and ground surface (Eugster *et al.*, 1997; Brown *et al.*, 2010). Continuous half-hourly ET fluxes were measured at 4 m above the ground, and aspen understory, but below the canopy for both the 2005 and 2006 snow free seasons. The mobile tower was situated 20 m north of the above canopy (permanent) tower for a 25 day period during green up in 2005 (DOY 124 to 149) and 33 days in 2006 (DOY 123 to 156) before being rotated to new locations every two weeks from June 3 to October 5. The purpose was to characterize fluxes from varying sites within the catchment over both seasons at the temporal scale of synoptic weather patterns (Eugster *et al.*, 1997). The mobile tower had identical instrumentation and sampling protocols, with the exception of instrument height, as that of the permanent tower system.

*EC data filtering and gap filling.* A simple flux footprint parameterization (Kljun *et al.*, 2004) was used to estimate temporally varying (30 min) contribution areas and associated vegetation and topographical characteristics to eddy covariance instrumentation. Analysis of the above canopy flux source indicates that the majority of the fluxes originated from the north-westerly upwind area (72% and 73% of all half hour periods in the 2005 and 2006 study periods, respectively), encompassing a predominantly aspen-dominated area (Chasmer *et al.*, 2011). The 80% probability density function of the total half-hourly flux extended from between 160 m and 600 m during highly convective and near neutral atmospheric conditions, respectively, indicating that fluxes represent aspen-dominated upland (Chasmer *et al.*, 2011). The location of the maximum probability of flux, determined from the footprint probability density function ( $X_{max}$ ), averaged 78 m from the tower above the canopy during both study seasons.

The aspen area studied was homogeneous to a minimum distance of 85 m from the tower in the westerly direction, and a maximum distance of approximately 250 m in the northwesterly and southwesterly directions. Results from the footprint model considering wind direction and the 80% footprint probability density function indicate that fluxes originated from the aspen areas approximately 73% of the time and were used to filter-out data that originated outside the area of the aspen stand of interest. To quantify the extent of the area capture by the below canopy system, a footprint analysis was done using an analytical solution to the diffusion equations according to Schuepp *et al.* (1990). This analysis showed that the location of the maximum probability of flux ( $X_{max}$ ) averaged 8 m from the below canopy tower during both study seasons. To ensure that both the above and below canopy sensors captured a representative flux, spectral density functions were computed using high-frequency EC data to ensure sensor location and sampling intervals were sufficient to capture the full size range in eddies (Petrone *et al.*, 2001).

Eddy covariance data were first corrected for density effects (Webb *et al.*, 1980; Leuning and Judd, 1996) and sensor separation (Blanford and Gay, 1992; Leuning and Judd, 1996). Energy balance closure (including canopy energy storage) was calculated based on the slope of the relationship:  $Q^* - Q_G$  and the total turbulent (sensible ( $Q_H$ ) and latent ( $Q_E$ ) energy) flux and canopy energy storage ( $J_c$ ) (Blanken *et al.*, 1997)). For both years, EC underestimates turbulent fluxes by approximately 19%, for all measurement periods, with no significant differences associated with wind direction, or between years, in the closure estimates. Data were then filtered for periods when the atmosphere was highly stable using  $u_*$  (friction velocity)  $< 0.2 \text{ m s}^{-1}$  (above canopy) and  $u_* < 0.29 \text{ m s}^{-1}$  (below canopy), determined using a 99% threshold condition for nighttime data (23h00 – 5h00) (Petrone *et al.*, 2001; Reichstein *et al.*, 2005; Papale *et al.*, 2006). Additional flux measurements were also removed during periods of rainfall, early morning flush of CO<sub>2</sub> from the ecosystem, and when rapid and unexpected changes in state variables occurred over half-hour intervals based on 1.5 standard deviations from the mean for that time period (Restrepo and Arain, 2005). Finally, a de-spiking method was used (Papale *et al.*, 2006), where night time data were selected based on  $20 \text{ W m}^{-2}$  radiation thresholds and a z-value of 5.5. Since this method is mainly used on high-frequency data rather than half-hourly averages, a supplemental de-spiking process was used where data exceeding 2 standard deviations from the mean were removed. For reported cumulative flux data, short half-hour breaks were filled by linear interpolation, while longer breaks (i.e.  $> 12$  half-hour periods) were filled using the mean diurnal variation method with a 14-day moving window (Falge *et al.*, 2001). Missing or rejected data occurred for a total of

19% of all possible time periods during the study, mostly during major precipitation events and nocturnal periods.

**Interception measurements.** Throughfall instrumentation used to estimate canopy wetness contributions to ET differed between 2005 and 2006. In 2005, three 1.5 × 0.1 m V-shaped plastic troughs draining into tipping bucket rain gauges (52202; R.M. Young, USA) were placed in the gap between the adjacent tree canopies, under the full canopy and close to the trunk of a representative aspen. The averages of the throughfall measured in the gap, canopy, and near the trunk of the aspen were similar to longer troughs used the following year which integrated spatial difference with in the aspen stand. In 2006, 10 m × 0.1 m U-shaped troughs were randomly placed within 30 to 200 m from permanent tower. Each trough sampled the integrated precipitation throughfall among several aspen trunks, stems, and differing canopy structural (foliage, leaf area). Three 10 m troughs were located in two separate aspen stands. The throughfall coefficient (p) was determined as the slope of the regression between throughfall and cumulative precipitation (P) over 3 month period, during the two study seasons (Grelle *et al.*, 1999). Water movement down stems (stemflow) was collected in a split plastic hose wrapped in a cork-screw fashion three or four times and affixed to the main tree stem using galvanized staples and silicone from a height of 0.75 to 2.5 m of the trunk. A tube trained into a collection container at the ground surface for measurement of volume collected. Average stem flow from three aspen trees associated with the canopy and with diameter of 15 cm (small), 25 cm (medium) and 35 cm (large) were collected through the growing season. (c.f. Crockford and Richardson, 2000; Toba and Ohta, 2005). Storage capacity (S) was determined as the x-intercept of the throughfall-precipitation regression (Grelle *et al.*, 1999). Evaporation of intercepted P was calculated using a modified Gash Model (Gash, 1979; McLaren *et al.*, 2008).

**Mensuration and leaf area measurements.** Forest mensuration plots were established at the study site in order to assess variations in leaf area index (LAI) and other forest stand characteristics within the catchment during the peak growth season (Aug 17 – 22) (Chasmer *et al.*, 2008). Three 15 m × 15 m plots were sampled for tree height and live crown length using a vertex sonic hypsometer; for crown apices that were not distinct, the average of three measurements was recorded, and diameter at breast height (DBH) (using a standard DBH caliper at 1.3 m). Crown diameter was measured along a north–south and east–west crown axes using a survey tape measure. All trees with a DBH greater than 9 cm were included within the survey as trees with a DBH of less than 9 cm were determined to not constitute a significant element in the overall canopy (Table I).

Table I. Summary table of catchment forest characteristics for the upland aspen-dominated forest situated at 'Pond 40', Utikuma Region, Alberta Canada. *LAI* is leaf area index ( $\text{m}^2 \text{m}^{-2}$ ), *PAI* is plant area index ( $\text{m}^2 \text{m}^{-2}$ ), and breast height is defined as 1.3 m. Measurements consisted of three 15 m × 15 m plots situated on the south facing slope (SFS) of the catchment within the calculated footprint of the EC tower

Max. leaf area index, <i>LAI</i>	1.45
Max. plant area index, <i>PAI</i>	0.92
Diameter at breast height, <i>DBH</i> (cm)	12.97
Base diameter (cm)	15.45
Tree height (m)	16.55

LAI was collected using digital hemispherical photography at five sites (one taken at the centre of the plot, and four were located 11.3 m from the centre along cardinal (N, S, E, and W) directions, determined using a compass bearing and measuring tape following Fluxnet-Canada and the Canadian Carbon Program protocol (Fluxnet-Canada, 2003) within three representative plots throughout the study area (Figure 1). All photographs were taken at a height of 1.3 m above the ground during either diffuse daytime conditions, or 30 min before dawn or dusk, and under-exposed by one f-stop to reduce the influence of sun brightness and apparent leaf reduction within the photograph (Zhang *et al.*, 2005; Chen *et al.*, 2006).

## RESULTS

### *Environmental conditions*

The upland aspen site was cooler and wetter on average in 2005 than 2006. Total precipitation was 491 mm and 432 mm for the two years, respectively, (Figure 2b) based on the hydrologic year, Nov 1 – Oct 31 (Figure 2a). Average daily air temperatures were 2.2 °C (ranging between –34.8 °C and 19.1 °C) in 2005 and 3.8 °C (ranging between –30.5 °C and 23.6 °C) in 2006 (Figure 2a). However the higher precipitation in 2005 was largely due to snowfall. During the June to Augusts period, when historically 50% of the annual precipitation occurs (Environment Canada 2005), both 2005 (231 mm) and 2006 (217 mm) were slightly below normal (248 mm). In both years, July rainfall was slightly above normal, with dry Augusts. August rainfall in 2006 (28 mm) was only half that in 2005 (48 mm), and well below normal August rainfall of 65 mm.

$\theta$  in the rooting zone (0–50 cm) did not vary greatly between years, ranging between 0.2 to 0.3  $\text{m}^3 \text{m}^{-3}$  in response to precipitation events coupled with evaporative demand (Figure 2b). However, with drier end of summer conditions in 2006, soil moisture at the end of September in the top 0.5 m dropped to about 0.11  $\text{m}^3 \text{m}^{-3}$  in 2006, compared to on 0.2  $\text{m}^3 \text{m}^{-3}$  in 2005.

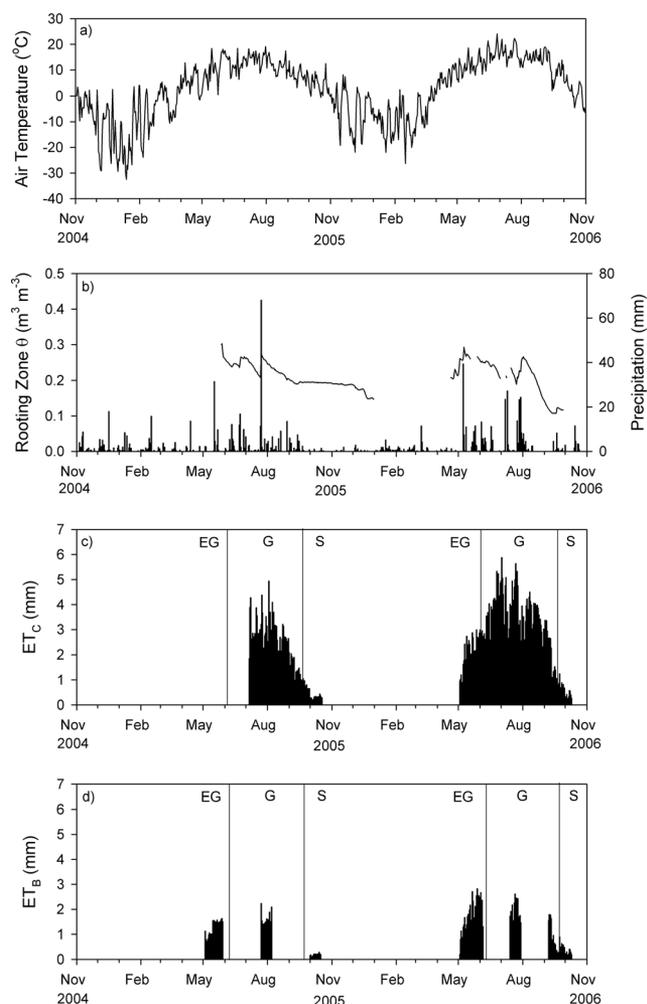


Figure 2. Daily averaged a) air temperature ( $^{\circ}\text{C}$ ), b) rooting zone Soil Moisture ( $\text{m}^3 \text{m}^{-3}$ ) (0–50 cm) and daily precipitation (mm), c) above canopy ET ( $\text{ET}_C$ ) (mm) and d) within canopy ET ( $\text{ET}_B$ ) (mm) for the 2005 and 2006 snow-free seasons, Pond 40, Utikuma Region Study Area (URSA), Alberta, Canada. Transitions between defined growth periods are denoted by vertical lines. EG represents early green period, G represents green period, and S represents senescence

*Seasonal partitioning of total stand ET.*  $\text{ET}_C$  differs significantly between the 2 years (Mann–Whitney:  $P < 0.001$ ,  $U = 5695$ ), and reaches maximum and minimum rates during the green and senescence periods, respectively (Figure 2b). In both the 2005 and 2006 snow-free seasons peaks in daily  $\text{ET}_C$  are seen following precipitation events (Figure 2b) and the magnitude of the precipitation events also affects the maximum rate of daily  $\text{ET}_C$  across both seasons where  $\text{ET}_C$  is larger following larger events. Following times of no precipitation,  $\text{ET}_C$  is reduced and is presumed to follow the controls of transpiration across the upper canopy. As precipitation is intercepted by the canopy during green periods,  $\text{ET}_C$  increases, suggesting the dominance of diurnal evaporation of intercepted water from leaf catch at the canopy. This is supported by the significant

canopy interception rates measured at this site. Throughfall measurements here show that the aspen intercept as much as 25% of incoming precipitation, for events greater than 5 mm (Petroni and Devito, unpublished data) and as much as 15% on an annual basis (Table II). Precipitation intensity – duration and frequency analysis at this site suggests that this aspen stand intercepted 49 mm and 40 mm during the 2005 and 2006 study periods, respectively (Table II). Results from the modified Gash model estimate that evaporation of this intercepted precipitation amounted to 43 mm and 35 mm in 2005 and 2006, respectively. As such, peaks in  $\text{ET}_C$  across both seasons are observed following days where precipitation events occurred that were greater than 10 mm.

$\text{ET}_B$  averages  $1.1 \text{ mm d}^{-1}$  during the early green period and increases at the point of understory vegetation emergence (Figure 2c) (e.g.: DOY 139 in 2005; DOY 141 in 2006).  $\text{ET}_B$  also differed significantly between 2005 and 2006 (Mann–Whitney:  $P < 0.001$ ,  $U = 1864$ ). Green period  $\text{ET}_B$  is approximately 45% of  $\text{ET}_C$  over both study seasons. Senescence period  $\text{ET}_B$  in 2005 is approximately 60% of  $\text{ET}_C$  rates with an average of  $0.4 \text{ mm d}^{-1}$  and is stable across the late season measurement period, in which no precipitation events occur (Figure 2, 3). In 2006,  $\text{ET}_B$  during late green averages  $0.9 \text{ mm d}^{-1}$  peaking during the early portion of late green when canopy development has just begun to cease.

*Ecosystem responses to atmospheric demand.* Figure 3 shows average diurnal ET, VPD, and PAR for three periods of the snow-free season of 2006 and the differences between above and below canopy fluxes. Both  $\text{ET}_C$  and  $\text{ET}_B$  peak in late afternoon (14:00 MST) during the early green period and at midday (12:00 MST) during green and senescence periods. Diurnal variations of PAR and the phenology of canopy *versus* understory biomass are strongly coupled with ET fluxes ( $r^2 = 0.95 - 0.99$ ;  $p < 0.01$ ), whilst the greatest PAR received by the understory occurs during the green up period before canopy leaves have reached their maximum leaf area. Figure 3 also indicates that the observed midday peaks in ET and PAR occur approximately 1 h before peak VPD during the Green period, as demonstrated in other studies (Blanken *et al.*, 1997; Hogg *et al.*, 1997; Grelle *et al.*, 1999). This synchronicity between ET and PAR, and offset with VPD, persists during the Early Green and Senescence

Table II. Total precipitation, above canopy ET ( $\text{ET}_C$ ) and interception ( $\pm$  standard deviations) for the study period in 2005 and 2006 (JD 187–291 and JD 123–283 in 2005 and 2006, respectively). Utikuma Region Study Area, Alberta, Canada

	Precipitation (mm)	$\text{ET}_C$ (mm)	Interception (mm)
2005	$181 \pm 13$	$205 \pm 2$	$27 \pm 3$
2006	$331 \pm 13$	$442 \pm 3$	$40 \pm 2$

CONTROLS ON ASPEN EVAPOTRANSPIRATION IN THE WESTERN BOREAL PLAIN

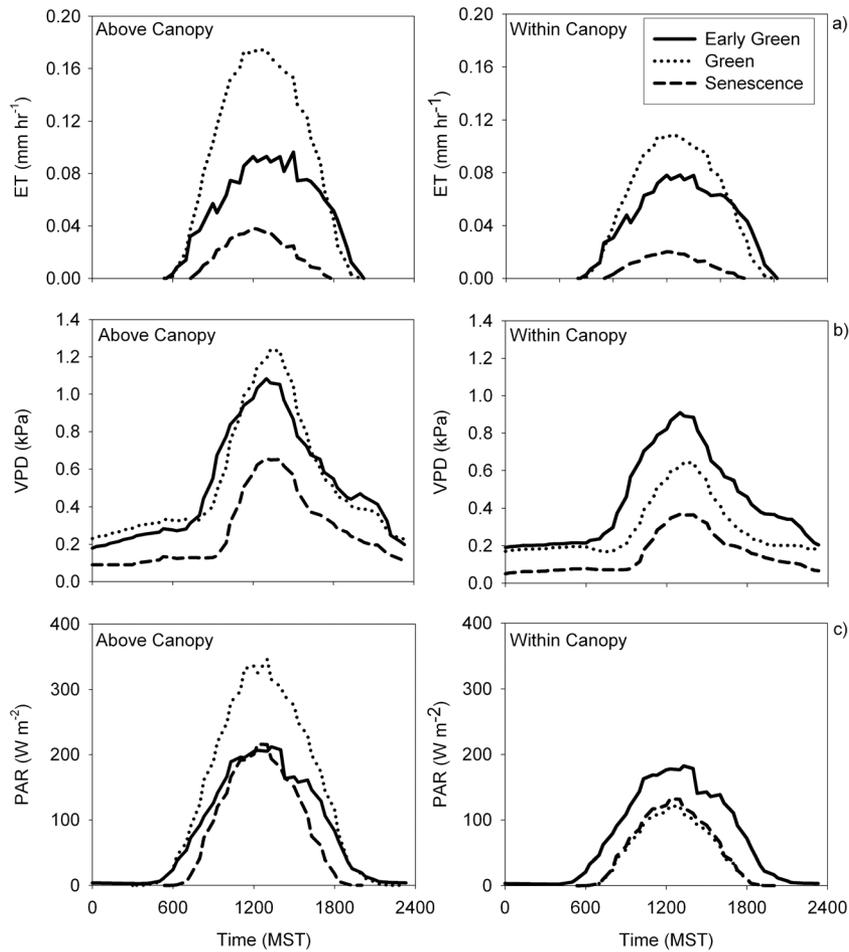


Figure 3. Above and below canopy period averaged diurnal comparisons of a) evapotranspiration (ET) (mm h<sup>-1</sup>), b) vapour pressure deficits (VPD) (kPa) and c) photosynthetic active radiation (PAR) (W m<sup>-2</sup>) Pond 40, Utikuma Region Study Area (URSA), Alberta, Canada

periods; however, the duration of the lag to the peak in VPD is much shorter (Figure 3).

The initial VPD for any given day defines the stomatal response to atmospheric demand, such that large atmospheric demand early in the day, during the onset of transpiration, will cause quicker stomatal restriction to vapour loss under prolonged atmospheric demand and moisture stress during the course of the day (McLaren *et al.*, 2008). The relationship between initial VPD (VPD<sub>0</sub>), defined as the average VPD between 06:00 and 08:00 MST, and ET<sub>C</sub> for both study periods shows a clustering of high, medium, and low ET<sub>C</sub> as a result of variability in VPD (Figure 4). To illustrate these clustered relationships, data are grouped into three categories based on ranges of VPD, whereby (1) VPD<sub>0</sub> LOW is defined as VPD<sub>0</sub> < 0.05 kPa; (2) VPD<sub>0</sub> MID > 0.05 kPa and < 0.15 kPa; and (3) VPD<sub>0</sub> HIGH > 0.15 kPa. Although there is some overlap in ET–VPD relationship among the three groups, days that begin with a high initial VPD clearly result in a greater probability of larger rates for the same range in VPD (Figure 4). Coefficients of determination (*r*<sup>2</sup>) between low,

medium, and high VPD and ET<sub>C</sub> are (adjusted) *r*<sup>2</sup> = 0.45 (*p* < 0.01, RMSE = 0.018 mm h<sup>-1</sup>); 0.44 (*p* < 0.01, RMSE = 0.051 mm h<sup>-1</sup>); and 0.71 (RMSE = 0.093 mm h<sup>-1</sup>), indicating that VPD<sub>0</sub> influences on ET<sub>C</sub> are most significant when VPD is high and other controlling mechanisms to ET<sub>C</sub> are not as important. Maximum rates of ET<sub>C</sub> occur during periods when VPD<sub>0</sub> remained high throughout the day (peak daily maximums > 0.3 mm h<sup>-1</sup>), while days with low VPD<sub>0</sub> had much lower ET<sub>C</sub> (< 0.15 mm h<sup>-1</sup>). These distinct clustering patterns observed between the three categories show that for aspen uplands, the use of VPD<sub>0</sub> as a proxy for ET<sub>C</sub> is an appropriate method for characterizing the daily catchment ET rates. Thus, to investigate whether changes in θ can produce diurnal shifts in maximum ET relative to maximum VPD data were first grouped by daily initial VPD (06:00 and 08:00 MST) (VPD<sub>0</sub>).

Given that understory vegetation receives a fraction of the solar energy received by the canopy (Petroni *et al.*, 2010), differences in VPD<sub>0</sub> due to incident radiation and canopy shadowing can have significant influences on ET, especially where understory and canopy LAI can vary by

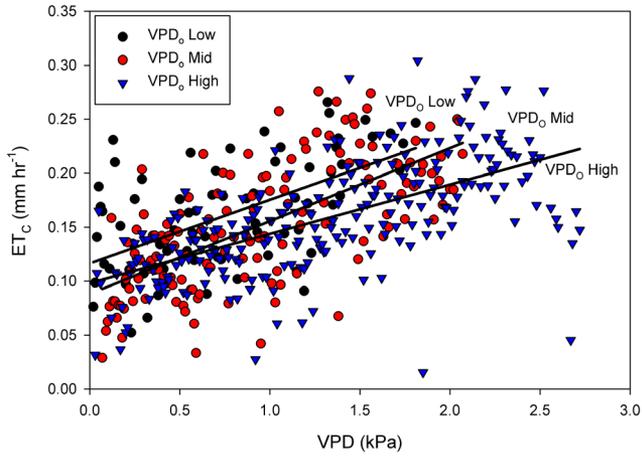


Figure 4. Relationship between above canopy ET ( $ET_C$ ) ( $\text{mm h}^{-1}$ ) and VPD (kPa) based on  $VPD_O$  thresholds (binned in 0.01 kPa intervals) of a)  $VPD_O$  LOW  $<0.05$  kPa, b)  $VPD_O$  MID  $>0.05$  kPa and  $<0.15$  kPa, and c)  $VPD_O$  HIGH  $>0.15$  kPa, Pond 40, Utikuma Region Study Area (URSA), Alberta, Canada. Solid lines represent best fit linear regressions through the respective VPD groups

up to  $6.7 \text{ m}^2 \text{ m}^{-2}$  depending on canopy closure (Chasmer *et al.* 2011). At the upland aspen site,  $ET_C$  is higher than  $ET_B$  for a given VPD (Figure 5). Maximum  $ET_B$  rates reach  $0.16 \text{ mm h}^{-1}$  at  $VPD_O$  levels less than 0.1 kPa. An increase in  $VPD_O$  from below the canopy does not result in an increased maximum daily ET above approximately 0.1 kPa. The same is observed for the relationship between daily maximum  $ET_C$  and  $VPD_O$  with a leveling of  $ET_C$  across a range of  $VPD_O$  values above  $\sim 0.10$  kPa. Clear clustering of daily maximum  $ET_C$  is observed at  $VPD_O$  levels less than 0.10 kPa. However, the range in  $ET_C$  at  $VPD_O$  values between 0.1 and 0.4 kPa is greater than observations in  $ET_B$  (Figure 5). The lower limit of  $ET_C$  is approximately  $0.15 \text{ mm h}^{-1}$  with the exception of

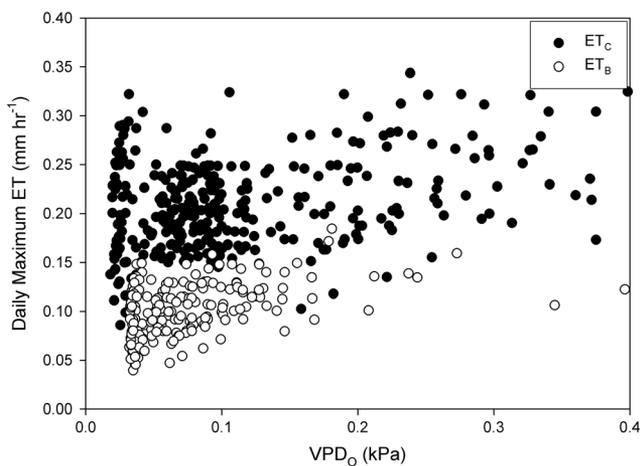


Figure 5. Relationship between daily maximum ET ( $\text{mm h}^{-1}$ ) and  $VPD_O$  (kPa) for above and within canopy layers of the forest canopy during defined green periods of the 2005 and 2006 snow-free seasons, Pond 40, Utikuma Region Study Area (URSA), Alberta, Canada

minimal outliers, which corresponds to the upper threshold of the daily maximum  $ET_B$ . Maximum  $G_C$  rates of approximately  $0.33 \text{ mm h}^{-1}$  are observed when  $\theta > 0.20 \text{ m}^3 \text{ m}^{-3}$  ( $r^2=0.61$ ,  $p < 0.01$ ,  $\text{RMSE}=0.05 \text{ mm h}^{-1}$ ) (Figure 6). Whereas when  $\theta$  is limiting ( $\theta < 0.20 \text{ m}^3 \text{ m}^{-3}$ ), ET only reaches a maximum of approximately  $0.19 \text{ mm h}^{-1}$ , with the majority of the ET clustering below  $0.9 \text{ mm h}^{-1}$  ( $r^2=0.53$ ,  $p < 0.01$ ,  $\text{RMSE}=0.03 \text{ mm h}^{-1}$ ).

**Rooting zone soil moisture controls on  $ET_C$ .** The relationships between rooting zone soil moisture ( $\theta$ ) (average  $\theta$  over the top 50 cm) and soil water matric potential ( $\psi_m$ ) with  $ET_C$  over four selected periods of growth through the 2006 season shows the interconnected dependence of the variables (Figure 7). During the 2006 study season early green period (DOY 135–138)  $\theta$  depletion from 0.255 to  $0.235 \text{ m}^3 \text{ m}^{-3}$  occurs when  $\psi_m$  and  $ET_C$  reach daily maximums of  $-82$  kPa and  $0.18 \text{ mm h}^{-1}$ , respectively (Figure 7a). Similar trends are observed for the green period with  $\theta$  depletion of 0.208 to  $0.187 \text{ m}^3 \text{ m}^{-3}$  coupled with a peak  $\psi_m$  of  $-120$  kPa and maximum daily  $ET_C$  rate of  $0.31 \text{ mm h}^{-1}$  (Figure 7b). Late green period shows a  $\theta$  depletion of 0.203 to  $0.182 \text{ m}^3 \text{ m}^{-3}$  and a maximum daily  $\psi_m$  of  $-96$  kPa and  $ET_C$  of  $0.29 \text{ mm h}^{-1}$  (Figure 7c). Senescence in 2006 again shows limited  $\theta$  depletion with reduced daily maximum  $\psi_m$  ( $\sim -37$  kPa) and limited  $ET_C$  ( $0.12 \text{ mm h}^{-1}$ ) (Figure 7d). This results in limited  $\theta$  depletion within the rooting zone.

## DISCUSSION

**Seasonal partitioning of ET and implications for climate change.** Aspen in the BP are already forced to make the most use of the limited precipitation the region receives

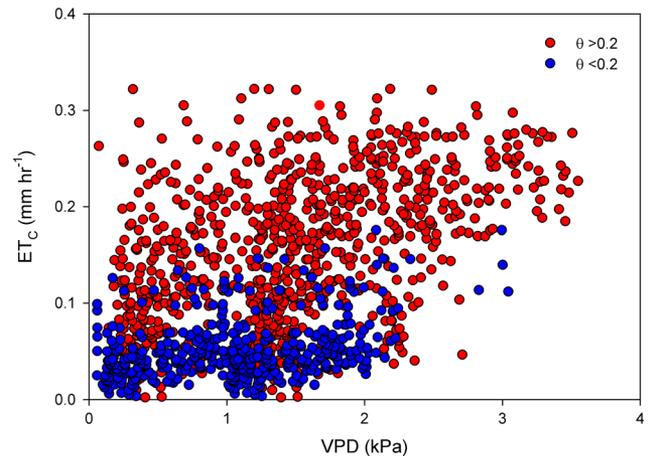


Figure 6. Relationship between above canopy evapotranspiration ( $ET_C$ ) ( $\text{mm h}^{-1}$ ) and VPD (kPa) for periods during the 2005 and 2006 peak growth periods when Soil Moisture ( $\theta$ ) becomes limited and approaches the wilting point ( $0.2 \text{ m}^3 \text{ m}^{-3}$ ) (Saxton *et al.*, 1986), Pond 40, Utikuma Region Study Area (URSA), Alberta, Canada

CONTROLS ON ASPEN EVAPOTRANSPIRATION IN THE WESTERN BOREAL PLAIN

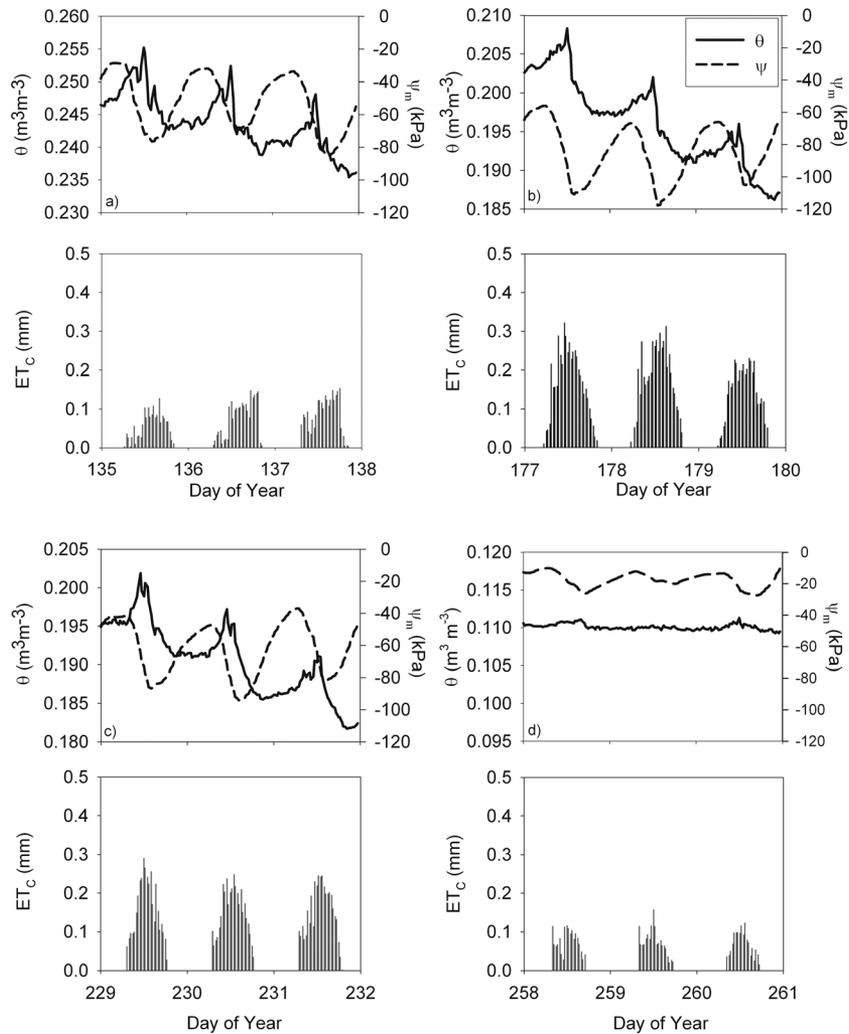


Figure 7. Precipitation free 4-day periods for the a) early green (EG), b) green (G), c) late green (LG), and d) senescence (S) periods of rooting zone soil moisture ( $\theta$  ( $\text{m}^3 \text{m}^{-3}$ ), rooting zone soil water matric potential ( $\Psi_m$ ) (kPa), and above canopy ET ( $\text{ET}_c$ ) (mm) for the 2006 snow-free season, Pond 40, Utikuma Region Study Area (URSA), Alberta, Canada

(Devito *et al.*, 2005b). The clonal nature of aspen (Desrochers and Lieffers, 2001) and sparse canopy (Sucoff, 1982; Davison *et al.*, 1988) are understood to have the most significant influence on the ability of the species to cope to with climatic stress. LAI values observed in this study are well within the range of similarly aged sites studied elsewhere (c.f. Shepperd, 1993; Constabel and Lieffers, 1996; Delong *et al.*, 1997). Canopy structure in general, and LAI specifically, affects environmental factors such as radiation penetration, wind, atmospheric humidity (or VPD), and soil moisture. In both summers of 2005 and 2006 when aspen LAI was at its maximum, a significant decrease within canopy VPD and PAR was observed, which suggests less radiation penetration but also less mixing. Both the suppressed VPD and diminished PAR within the canopy translate to much a much lower  $\text{ET}_B$ , relative to  $\text{ET}_C$  during the

Green period, whereas a minimal difference is observed during Early Green and Senescence periods.

Interception is an important component in water balance of forests as the amount of evaporation from interception is a significant proportion of precipitation and can be large in regions with net moisture deficit. (Toba and Ohta, 2005; Asdak *et al.*, 1998; Scatena, 1990; Crockford and Richardson, 2000). The importance of interception and evaporation in BP aspen forests is indicated in both the 2005 and 2006 snow-free seasons, where peaks in daily  $\text{ET}_C$  occurred following precipitation events. In both years, the magnitude of the precipitation events also affected the maximum rate of daily  $\text{ET}_C$ , as  $\text{ET}_C$  increased following larger events (Crockford and Richardson, 2000). Following times of no precipitation,  $\text{ET}_C$  is reduced and is presumed to follow the controls of transpiration across the upper canopy

(Iacobelli and McCaughey, 1993). However, as precipitation is intercepted by the canopy during green periods,  $ET_C$  increases, suggesting the dominance of diurnal evaporation of intercepted water from leaf catch atop the canopy in maintaining high  $ET_C$  rates. This is supported by the significant canopy interception rates measured at this site. For example, the difference between canopy and canopy + understory LAI varies by as much as 32% along the soil evaporation transect and between above and below canopy eddy covariance sites (Figure 1) (Chasmer *et al.* 2011). Thus, as the BP region is expected to warm 2–4 °C, and experience a 0–20% change in precipitation over the next century (IPCC, 2007), larger VPDs and greater interception loss have the potential to increase aspen stand ET and may significantly alter the watershed water balance.

*Atmospheric controls and ET.* VPD is expected to be the most important environmental factor governing the transpiration of boreal deciduous forests when  $\theta$  and radiation are not limiting (Blanken *et al.*, 1997; Hogg *et al.*, 1997). Both  $ET_C$  and  $ET_B$  peak in late afternoon during the green up period, but peak at midday (12:00 MST) during both the green and senescence periods. Maximum and minimum atmospheric demand as indicated by the maximum of VPD from above and below the canopy can provide insight into the nature of diurnal ET patterns at both scales. Peak VPD corresponds with peak  $ET_C$  and  $ET_B$  during the early green and senescence periods but is delayed compared to peak ET for the green period. A delay in peak ET between the canopy layers is because eventually, the stomata of aspen react to VPD and soil moisture to keep water potential above a critical level in water scarce environments (Lieffers *et al.*, 2001). In this study, the observed midday peaks in ET occurred just before peak VPD and at approximately the same time as PAR, as demonstrated in other studies (Blanken *et al.*, 1997; Hogg *et al.*, 1997; Grelle *et al.*, 1999). Since, aspen has the ability to survive in water stress periods, there must be a mechanism to respond conservatively to the low soil moisture as well as high VPD, which is common in drought areas like the BP.

*Interaction between atmospheric demand and  $\theta$ .* At high transpiration rates, trees will experience water stress due to decreasing water potentials from the roots to the leaves as a result of large VPD and/or high productivity (Sperry *et al.*, 1998). The stomatal dynamics responsible for this transpiration response to VPD variations are a function of the hydraulic response from the soil to leaf where stomatal conductance will decrease as the soil dries (Meinzer and Grantz, 1991; Iacobelli and McCaughey, 1993; Hogg and Hurdle, 1997). Regulation of leaf water potential and transpiration by stomatal conductance in response to VPD maintains a homeostasis of water in the leaves for optimal balance between productivity and  $\theta$  supply (Katul *et al.*, 2003; Ewers *et al.*, 2005). However, while a relationship should be observable between VPD and canopy ET under differing  $\theta$  conditions, VPD and  $\theta$  may not always be independent of each other (Kurpius *et al.*, 2003; McLaren *et al.*, 2008). If VPD is high at the beginning of the diurnal transition cycle (early in the day), stomatal closure will occur earlier in the day regardless of the magnitude of VPD later in the cycle (Kurpius *et al.*, 2003).

The distinct clustering patterns observed between the three categories (HIGH, MID, and LOW) of  $VPD_O$  and maximum  $ET_C$  show that for URSA the use of  $VPD_O$  as a proxy for  $ET_C$  is an appropriate method for characterizing the daily catchment ET rates. Thus, to investigate whether changes in  $\theta$  can produce diurnal shifts in maximum ET relative to maximum VPD data were first grouped by daily initial VPD (06:00 and 08:00 MST) ( $VPD_O$ ).

Further, given that the vegetation comprising the understory will differ in the amount of solar radiation received and root structure, it can be expected that this relationship will differ for  $ET_C$  and  $ET_B$ . The lower limit of  $ET_C$  is approximately  $0.15 \text{ mm h}^{-1}$  with the exception of minimal outliers, which corresponds to the upper threshold of the daily maximum  $ET_B$ . This data suggests that VPD tends to constrain ET within a much narrower range in the understory (Schwartz *et al.*, 2006). Despite the controls of atmospheric demand on stomata opening, the root systems of the aspen are still able to maintain critical hydraulic gradients from the soil to the leaf (Schwartz *et al.*, 2006). Thus, it is expected that  $\theta$  will

Table III. Soil texture properties for the rooting zone (0–50 cm) of this silty clay loam soil below the litter fall horizon based on soil cores and taken during installation (May 2005), soil texture classifications, and soil moisture properties based on Saxton *et al.*, 1986

Soil texture properties	Sand (%)	7.0
	Clay (%)	28.0
	Silt (%)	65.0
Soil moisture properties	Wilting point ( $\text{m}^3 \text{ m}^{-3}$ )	0.15
	Field capacity ( $\text{m}^3 \text{ m}^{-3}$ )	0.34
	Bulk density ( $\text{g cm}^{-3}$ )	1.29
	Saturation ( $\text{m}^3 \text{ m}^{-3}$ )	0.51
	Soil hydraulic conductivity ( $\text{cm h}^{-1}$ )	0.72
	Available water ( $\text{m}^3 \text{ m}^{-3}$ )	0.18

influence this relationship between VPD and ET and will likely be the limiting factor on ET regardless of atmospheric demand. When data was binned into two  $\theta$  categories,  $\theta > 0.20 \text{ m}^3 \text{ m}^{-3}$  and  $\theta < 0.20 \text{ m}^3 \text{ m}^{-3}$ , roughly corresponding to the wilting point for aspen in this type of soil (Table III; Saxton *et al.*, 1986), it can be observed that ET rates are limited when  $\theta$  approaches the wilting point of the soil ( $0.15 \text{ m}^3 \text{ m}^{-3}$ ) regardless of VPD (Saxton *et al.*, 1986; Oren *et al.*, 1998). During periods when  $\theta$  is not limited in the rooting zone, ET rates can be maintained at higher levels, even at higher VPDs (Meinzer and Grantz, 1991). Further, the more gradual overall decrease in  $G_c$  with increasing VPD (Figure 6), even under the drier soil conditions, is indicative of the deeper rooting clonal nature of the aspen stand, as they are collectively capable of progressively drawing water from deeper layers (Bréda *et al.*, 1995; Oren *et al.*, 1998).

*Interaction of  $\theta$  and  $\psi$  within the rooting zone and the influence on  $ET_C$ .* The existence of many plant species is driven by water availability, as this is the key factor in limiting carbon dioxide fixation and subsequent growth (Horton and Hart, 1998). This is especially important in sub-humid and arid regions, where there are many periods of water shortage. Therefore, in a sub-humid region like the BP, it is important to understand where the aspen stands/clone is obtaining water to drive ET. Under most conditions within their range, aspen stands are characterized by an extensive clonal root system (Barnes, 1966; DesRochers and Lieffers, 2001) that could increase the ability of aspen to cope with climatic stress (DesRochers and Lieffers, 2001). Thus, despite having a relatively high transpiration rate (averaging as high as 3.5 mm/d in this study), aspen might be able to succeed on dry sites as a result of its deep and extensive root system (Kemperman and Barnes, 1976; Sucoff, 1982; Strong and La Roi, 1983).

One of the important factors that could affect the water balance associated with an extensive clonal root system is hydraulic redistribution (Burgess *et al.*, 1998). During hydraulic redistribution, roots take up water from deep soil layers, or laterally from adjacent wet areas during low ET (at night) and distribute it through the root system and release (bleed) the water in areas where soil water potentials are much lower than root water potentials (Horton and Hart, 1998). Hydraulic lift and hydraulically redistributed water have positive implications not only for those species that are capable of doing so, but also for associated (understory) vegetation (Horton and Hart, 1998). Thus, forest stands comprised of trees capable of hydraulic redistribution have the potential to maintain a higher stand transpiration rate and have greater success during periods of water shortage (Horton and Hart, 1998). Studies (c.f. McLaren *et al.*, 2008) have suggested that the diurnal increases in  $\theta$  are evidence of hydraulic lift. After

calibration and temperature corrections, the error in TDR data reported here are approximately  $\pm 0.005$ . While the downward trend in  $\theta$  over the 3 day period does provide evidence of general decline in rooting zone  $\theta$ , the diurnal patterns in  $\psi_m$  provide a clearer indication of the occurrence of hydraulic lift occurring in response to ET demand. This is especially evident when comparing green and senescence period data where ET and  $\theta$  diurnal variation and maximums are decreased but the amplitude of the daily changes in  $\theta$  remains basically the same.

A delay exists in the response of  $\theta$  to morning decreases in  $\psi_m$  throughout the snow-free season of 2006. As evaporative demand begins during early morning hours a decrease in  $\psi_m$  is observed along with ET. However, over all periods the  $\theta$  response is delayed until  $ET_C$  reaches peak daily values at which time  $\psi_m$  increases until daily  $ET_C$  is limited and a 'bleeding effect' of soil water demand permits the capillary rise and moisture release from the roots to occur, causing  $\theta$  to increase gradually (McLaren *et al.*, 2008). In our study, these subtle changes in  $\theta$  occur throughout the snow-free seasons of 2006, but are most pronounced during peak growth when water demand from the aspen roots is at its greatest (Caldwell *et al.*, 1998). While during senescence periods when root water uptake is all but removed a limited response is observed in  $\theta$  within the rooting zone.

## CONCLUSIONS

Understanding the controls on the partitioning of ET within aspen forests of the BP are essential to interpreting the influences of climate change on stand water balances, and how these stands may influence the hydrology of adjacent wetland areas, in this region. Strong seasonality is observed in  $ET_C$  across both snow-free seasons, with maximum daily averages in  $ET_C$  of 3.08 and 3.45  $\text{mm d}^{-1}$  occurring during the green periods of 2005 and 2006, respectively.  $ET_B$  can be an important contribution of overall canopy ET although  $ET_B$  varies seasonally, largely a function of the ability of the aspen canopy to permit radiation to reach the understory and the aspen root systems influence on the availability of root zone  $\theta$ . For the 2005 and 2006 snow-free seasons, understory ET was the dominant flux during the early green period, which was a direct result of the timing of understory canopy shrub layer development. For the remainder of the growth periods,  $ET_B$  averaged 32% of  $ET_C$ . Further, the evaporation of intercepted water comprises a significant portion of the daily canopy ET as a result of the large atmospheric demand for water vapour in this climate.

Similar responses to VPD are observed in both the above and within canopy ET fluxes. However, the relative increase in ET for a given increase in VPD is greater from the understory, especially during the green period, which

is a function of access to available rooting zone  $\theta$  and muted fluctuations in VPD within the canopy microclimate. Further, in all levels of the canopy,  $\theta$  exerts the ultimate control on ET. That is, regardless of the degree of atmospheric demand, ET can be maintained at higher levels if there is sufficient  $\theta$  in the rooting zone. Finally, soil tension and moisture data within the rooting zone suggest that some form of hydraulic redistribution is occurring in addition to capillary recharge from below at night, transporting soil moisture from wetter areas to meet the demands of the transpiring aspen canopy. When aspen transpiration decreases this redistributed moisture becomes available within the rooting zone for the within canopy vegetation and mitigating the daily decline in soil moisture.

## ACKNOWLEDGEMENTS

The authors would like to thank the HEAD2-NSERC Collaborative Research and Development grant, Canadian Oilsands Network for Research and Development, Forest Producers Association of Canada, Ducks Unlimited, Alberta Pacific Forest Products, and Tolko for funding for this research. We would also like to thank the anonymous reviewers for their insightful comments.

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