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Evaluation Of Climatic And Ecohydrological Effects On Longwave Radiation And Evapotranspiration

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EVALUATION OF CLIMATIC AND ECOHYDROLOGICAL EFFECTS ON
LONGWAVE RADIATION AND EVAPOTRANSPERSION

by

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A dissertation submitted in partial fulfillment of the requirements
for the degree of Doctor of Philosophy
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Major Professor: Fidelia N. Nnadi
ABSTRACT

Modern tools, nontraditional datasets and a better understanding of the interaction between climate and ecohydrology are continuously being developed as today’s society is in critical need for improving water management, predicting hydrometeorological hazards and forecasting future climate. In particular, the study of the intra- and inter-annual variations in grass productivity and evapotranspiration caused by variations in precipitation/soil moisture and other biophysical factors is of great significance due to their relation to future climatic changes.

The research presented here falls in three parts. In the first part of the dissertation, a land use adaptable model, based on the superposition of the temperature and water vapor pressure effects, is proposed for the effective clear sky emissivity. Ground radiometer and meteorological data, applicable in the subtropical climate of Saint Johns River Water Management District, Florida, were utilized for the model development over the spring season of 2004. The performance of this model was systematically evaluated by pertinent comparisons with previously established models using data over various land covers.

The second part of the thesis investigates the dynamics of evapotranspiration with respect to its significant environmental and biological controls over an unmanaged bahia grassland. Eddy correlation measurements were carried out at a flux tower in Central Florida over the annual course of 2004. The main focus was on the sensitivity of the water vapor flux to wetness variables, namely the volumetric soil water content and the current precipitation index. It was shown that the time scales involved with the dynamics of evapotranspiration...
were on the order of six days, suggesting that depletion of the soil moisture was mostly responsible for the temporal fluctuations in evapotranspiration. Finally, simple models for the Priestley-Taylor factor were employed in terms of water availability, and the modeled results closely matched the eddy covariance flux values on daily time scale during all moisture conditions.

In the third part of this work, the partitioning between latent and sensible heat fluxes was systematically examined with respect to biophysical factors. It was found that the seasonal variations in leaf area index, soil water content and net radiation were reflected in a strong seasonal pattern of the energy balance. Calculations of the bulk parameters, namely Priestley-Taylor parameter and decoupling coefficient, indicated that evapotranspiration of this grassland was controlled by water supply limitations and surface conductance. At an annual basis, the cumulative evapotranspiration was 59 percent of the precipitation received at the site. The results of this research complemented with other studies will promote better understanding of land-atmosphere interactions, accurate parameterizations of hydroclimatic models, and assessment of climate impact of grassland ecosystems.
To my family:

My beloved parents, Konstantinos and Magdalini,

and my brother Petros
“…all our predecessors called meteorology. It is concerned with events that are natural, though their order is less perfect than that of the first of the elements of bodies. They take place in the region nearest to the motion of the stars. Such are the milky way, and comets, and the movements of meteors. It studies also all the affections we may call common to air and water, and the kinds and parts of the earth and the affections of its parts”.

“Meteorology”

Aristotle, 350 B.C.
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For the first part of my dissertation, I would like to thank mainly George Robinson (SJRWM) for providing the longwave radiation product and access to the ground stations, and Dr. Martha Anderson (University of Wisconsin) for insightful help and for providing GOES radiation data for comparisons. For the later parts of my thesis, I wish to thank USGS (especially the Water Resources Group at Orlando, FL) and SFWMD who funded and assisted in data collection from the WRWX ET and weather stations, respectively. Special recognition to Drs. Martin Wanielista (UCF), Gary Wu (SFWMD) and Louis Murray (USGS) for reviewing the journals related to the evapotranspiration study. In addition, the advice of Dr. Rosvel Bracho and Dr. John Weishampel for software support is acknowledged.
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May 2008

Maria Rizou.

Δόξα εν υψίστος Θεώ
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<tr>
<td>ASOS</td>
<td>Automated Surface Observing System</td>
</tr>
<tr>
<td>CNR</td>
<td>Campbell Scientific Inc. Net Radiometer</td>
</tr>
<tr>
<td>EC</td>
<td>Eddy Covariance</td>
</tr>
<tr>
<td>FPAR</td>
<td>Fraction of Photosynthetically Active Radiation</td>
</tr>
<tr>
<td>GOES</td>
<td>Geostationary Operational Environmental Satellite</td>
</tr>
<tr>
<td>LAI</td>
<td>Leaf Area Index</td>
</tr>
<tr>
<td>MODIS</td>
<td>MODerate resolution Imaging Spectroradiometer</td>
</tr>
<tr>
<td>NOAA</td>
<td>National Oceanic and Atmospheric Administration</td>
</tr>
<tr>
<td>PAR</td>
<td>Photosynthetically Active Radiation</td>
</tr>
<tr>
<td>PM</td>
<td>Penman-Monteith</td>
</tr>
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<td>Priestley-Taylor</td>
</tr>
<tr>
<td>SFWMD</td>
<td>South Florida Water Management District</td>
</tr>
<tr>
<td>SJRWMD</td>
<td>Saint Jones River Water Management District</td>
</tr>
<tr>
<td>USGS</td>
<td>U.S. Geological Survey</td>
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\[
A = \text{Albedo} \\
B = \text{Bowen ratio} \\
C_p = \text{Specific heat of air} \\
\text{CPI} = \text{Current precipitation index} \\
\text{DOY} = \text{Day Of Year} \\
e = \text{Actual water vapor pressure} \\
\dot{e}_o = \text{Screen-level water vapor pressure} \\
\dot{e}_s = \text{Saturation water vapor pressure} \\
\text{ET} = \text{Evapotranspiration} \\
\text{ET}_{\text{pot}} = \text{Potential evaporation} \\
G = \text{Soil heat flux} \\
\dot{g}_a = \frac{1}{r_a} = \text{Aerodynamic conductance}
\]
$g_c = 1/r_c$  
Bulk canopy surface conductance

$H$  
Sensible heat

$k$  
von Karman constant

$K$  
Recession coefficient

$LW_d$  
Downward longwave radiation

$LW_u$  
Upward longwave radiation

MBE  
Mean Bias Error

MSE  
Mean Square Error

NSE  
Nash and Sutcliffe’s Efficiency

$q$  
Specific humidity

$r$  
Pearson correlation coefficient

$R$  
Correlation coefficient

$R^2$  
Coefficient of determination

$R_n$  
Net radiation

RH  
Relative Humidity

RMSE  
Root Mean Square Error

SE  
Standard Error

$SW_{in} = SW_d$  
Incoming or downward shortwave radiation

$SW_u$  
Upward shortwave radiation

$SWC$  
Soil Water Content

$T_a$  
Air temperature

$T_d$  
Dew point temperature

$T_o$  
Surface or screen-level air temperature

$T_s$  
Surface temperature

$u_*$  
Friction velocity

VPD  
Vapor Pressure Deficit

$w$  
Vertical wind speed

WT  
Depth from land surface to water table
\( \alpha \)  
PT coefficient

\( \alpha_{LU} \)  
Offset factor for land use

\( \gamma \)  
Psychrometric constant

\( \Delta \)  
Slope of the saturation vapor-pressure curve

\( \varepsilon_a \)  
Air (atmospheric) emissivity

\( \varepsilon_s \)  
Surface emissivity

\( \lambda \)  
Latent heat of vaporization of water

\( \lambda E \)  
Latent heat flux

\( \lambda E_{eq} \)  
Equilibrium latent heat flux

\( \rho_a \)  
Moist air density

\( \sigma \)  
Stefan–Boltzmann constant

\( \tau \)  
Time constant

\( \Omega \)  
Decoupling coefficient
CHAPTER 1: INTRODUCTION

1.1 Research Needs

Modern tools, nontraditional datasets and a better understanding of the interaction between hydrology and climate are continuously being developed as today’s society is in critical need for improvement of water management, forecasting of future climate and prediction of hydrometeorological hazards. Environmental changes caused by land use changes, greenhouse gas emissions and freshwater resources limitation (mainly due to globalization, agricultural and environmental policies and technological developments) are closely linked to the water and carbon cycles and thus the regional climate. More particularly, climatic variables, such as insolation, temperature, humidity and precipitation affect the physiological functioning and stage of vegetation, the soil and hydrological properties. In return, the type and extent of vegetation and land cover exert feedback on the state of the atmosphere (Wilson and Baldocchi, 2000). Consequently, the transport of atmospheric trace gases is subject to a suite of biophysical controls as they travel between the biosphere and atmosphere. The main greenhouse gas—with about 60% contribution to the greenhouse effect under clear skies—is the water vapor, which is responsible for a dominant feedback in the climate system (Karl and Trenberth, 2003). For instance, as soil water supply or plant stomatal closure limit atmospheric moisture and water vapor flux, the near-surface atmospheric humidity deficit and temperature are increased. In turn, this leads to increased thickness of the atmospheric boundary layer, enhanced entrainment of warm and dry air, and an overall positive feedback for continued surface drying (Entekhabi et al., 1999). There is a
critical research need to better understand the land-atmosphere processes and accurately predict their role to global warming. Following this need, the first part of the dissertation investigates the effect of surface cover on the downward longwave radiation flux (or effective atmospheric emissivity) and subsequently on local warming.

About 40% of the terrestrial natural vegetation includes grassland ecosystems (White et al., 2000), which show significant annual variations in primary production (Knapp and Smith, 2001). Due to a strong link between grass productivity and evapotranspiration, large seasonal and interannual variations in grass evapotranspiration and its biotic and abiotic controls are also observed. In addition, human-induced modifications of the environment, such as land use change, significantly affect these variations. There should be a research focus on understanding how trends and diurnal, seasonal and interannual variations in climatic variables affect the energy and water exchange between terrestrial ecosystems and the atmosphere. The majority of the grassland evapotranspiration studies reported in literature extent mostly to temperate climate zones in North America, such as California (e.g. Baldocchi et al., 2004), Kansas (e.g. Verma et al., 1992), Oklahoma (e.g. Meyers, 2001) and Canada (e.g. Wever et al., 2002). The latter part of the dissertation systematically examines the water vapor dynamics with relation to biophysical factors over an non-irrigated unmanaged grassland at the subtropical region of Central Florida, which is warmer and wetter than most grassland ecosystems in the literature (Table 3.5).
1.2 Background

1.2.1 Background and study area

The studies of this dissertation examine certain energy components, such as downward longwave radiation, and water fluxes, mainly evapotranspiration (ET), in Florida. The climate in Florida is humid subtropical with a rainy, wet season extending from May through October. The long-term annual total precipitation is about 50 (±11.5) in and the annual mean temperature is 22.4 (±0.6) °C based on historical records of a weather station located in Kissimmee, Central Florida (Southeast Regional Climate Center, http://www.dnr.sc.gov/climate/sercc). Sumner and Jacobs (2005) documented that the long-term fraction of annual precipitation returning to the atmosphere as ET in Florida ranges from about 50% in settings of relatively deep water table, shallow rooted vegetation, and sandy soils to almost 110% in lakes. Climatic conditions are influenced by convective systems with dynamic cloud systems during summer-time. The variable cloud cover and precipitation induces diurnal and day-to-day fluctuations in net radiation that partly account for the variability in ET. The temporal variations in ET were found to be less than the variations in precipitation at five sites of Central Florida during a 10-yr study conducted by O’Reilly (2007). A radiation measurement network, which is equipped with Campbell Net Radiometers (CNR1) and operated by Saint Jones River Water Management District (SJRWMD), consists of 11 CNR1 stations within the region of SJRWMD at various land use settings: urban, agricultural, rangeland, forest, open water and wetland (Figure 2.1 of Chapter 2 and Table B.1 of Appendix). In addition, ET is monitored at 21 stations operated mostly by U.S. Geological Survey (USGS)—few stations are operated by University of Florida,
Smithsonian Environmental Research Center and University of Virginia—throughout Florida in various environmental settings (Figure 1.1). There are also several weather stations in Florida operated by various agencies, such as National Oceanic and Atmospheric Administration (NOAA) and South Florida Water Management District (SFWMD) (Figure 2.1 and Figure 3.1).

Evapotranspiration (ET) or latent heat flux ($\lambda E$) is the process by which the earth's surface loses water by evaporation and transpiration by the plants. ET (or $\lambda E$) is an important part of the water and energy balance at the global surface since it accounts for
large portion of the water and energy resources (Figure 1.2). In the hydrologic budget of Florida, ET is the second important component after precipitation (Jones et al., 1984). Accurate knowledge of ET is necessary in evaluating parameterization schemes used in hydrologic and climatic models, quantifying agricultural applications (such as crop yield and water use), assessing the environmental aspects of natural ecosystems and improving water management techniques.

Figure 1.2: a) Mean global water cycle showing storage (regular font) and exchanges (italic font); b) long-term budget of water flows (adopted from Trenberth et al., 2007).
Net radiation ($R_n$) is the main energy flux driver of ET which exhibits significant temporal and spatial variability. Sellers et al. (1990) suggested that estimating the four components of $R_n$ could cause error accumulation, especially when estimating the net longwave flux, because both downwelling and upwelling longwave radiation are large components, so the difference would be small and liable to large uncertainty. Consequently, a high priority should be given to the accurate prediction of the radiation fluxes, especially the downward longwave radiation which has a positive effect on $R_n$ (Section 1.2.5). An improvement of the existing models for estimation of radiation components would increase the accuracy of ET.

1.2.2 Estimation of radiation

As previously mentioned, ET is dominantly controlled by the net radiation reaching the earth surface. When the radiation enters the atmosphere, a fraction is scattered by diffusion, reflected or absorbed by the clouds and other aerosols. A simple schematic of the radiation budget is given in Figure 1.3, where the components of the longwave radiation are given by Stefan’s law.
Figure 1.3: Schematic of the radiation budget.

The radiation balance equation is:

\[ R_n = SW_m (1 - A) + \varepsilon_a \varepsilon_s \sigma T_a^4 - \varepsilon_s \sigma T_s^4 = SW_m (1 - A) + \varepsilon_s LW_d - LW_u \]  \hspace{1cm} (1.1)

where \( SW_m \) (or \( SW_u \)) is the incoming (or downward) shortwave (SW) radiation, \( SW_u \) is the upward shortwave radiation, \( A \) is the surface albedo, \( \varepsilon_a \) is the air (atmospheric) emissivity, \( \varepsilon_s \) is the surface emissivity, \( \sigma \left( = 5.67 \times 10^{-8} \text{ W m}^{-2} \text{K}^{-4} \right) \) is the Stefan–Boltzmann constant, \( T_a(K) \) is the air temperature, \( T_s(K) \) is the surface temperature, \( LW_d \) is the downward longwave (LW) radiation, and \( LW_u \) is the upward longwave radiation. All radiation terms are in \( \text{W m}^{-2} \). The incoming shortwave radiation is contained in the range \( 0.1 - 4 \mu m \), whereas the longwave (terrestrial) radiation is contained in the range \( 4 - 100 \mu m \).
The air temperature \( T_a \) close to the land surface (screen level) is used in lieu of \( T_s \) to estimate \( LW_u \) radiation due to uncertainties involved in the estimation of \( T_s \). The surface albedo, which is the fraction of the shortwave reflected (including the diffuse portion of radiation) to the incoming radiation, depends on the reflectivity and roughness of the surface and the angle of the incoming sun beam. However, when the cloudiness increases, the dependence of \( A \) on the solar angle becomes weaker. Table 1.1 presents mean values of albedo and surface emissivity for different surface types. As shown in Table 1.1, \( s\varepsilon \) is usually close to unity. The atmospheric emissivity under clear skies (\( \varepsilon_a \)) depends on the air temperature and water vapor pressure, which are denoted by \( T_a \) and \( e_v \), respectively, when measured at screen level under the assumption of a homogeneous surface atmospheric slab. Some of the existing clear sky emissivity models are described by the equations shown in Table 1.2.

Table 1.1: Albedo and surface emissivity for various surface types (Brutsaert, 1982).

<table>
<thead>
<tr>
<th>Type of cover</th>
<th>Albedo (( A ))</th>
<th>Surface emissivity (( s\varepsilon ))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>0.04-0.08</td>
<td>-</td>
</tr>
<tr>
<td>Bare soils</td>
<td>0.05 (wet)-0.35 (dry)</td>
<td>0.95-0.98</td>
</tr>
<tr>
<td>White sand</td>
<td>0.30-0.40</td>
<td>-</td>
</tr>
<tr>
<td>Green short vegetation (grass, alfalfa)</td>
<td>0.15-0.25</td>
<td>0.97-0.98</td>
</tr>
<tr>
<td>Dry grass</td>
<td>0.15-0.20</td>
<td>-</td>
</tr>
<tr>
<td>Dry prairie and savannah</td>
<td>0.20-0.30</td>
<td>-</td>
</tr>
<tr>
<td>Forest</td>
<td>0.10-0.30</td>
<td>0.96-0.97</td>
</tr>
<tr>
<td>Snow</td>
<td>0.35-0.90</td>
<td>0.97-0.99</td>
</tr>
</tbody>
</table>
Table 1.2: Clear sky emissivity ($\varepsilon_a$) equations of existing models.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Formulation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Angstrom (1918)</td>
<td>$\varepsilon_a = 0.79 - 0.174 \cdot \exp\left(-0.095 \cdot e_o\right)$</td>
</tr>
<tr>
<td>Swinbank (1963)</td>
<td>$\varepsilon_a = 9.36 \times 10^{-6} \cdot T_o^2$</td>
</tr>
<tr>
<td>Idso-Jackson (1969)</td>
<td>$\varepsilon_a = 1 - 0.261 \cdot \exp\left[-7.77 \times 10^{-4} \left(273 - T_o\right)^2\right]$</td>
</tr>
<tr>
<td>Brutsaert (1975)</td>
<td>$\varepsilon_a = 1.24 \cdot \left(\frac{e_o}{T_o}\right)^{1/7}$</td>
</tr>
<tr>
<td>Satterlund (1979)</td>
<td>$\varepsilon_a = 1.08 \left[1 - \exp\left(-e_o^{T_o/2016}\right)\right]$</td>
</tr>
<tr>
<td>Idso (1981)</td>
<td>$\varepsilon_a = 0.7 + 5.95 \times 10^{-5} \cdot e_o \cdot \exp\left(\frac{1500}{T_o}\right)$</td>
</tr>
<tr>
<td>Bignami (1995)</td>
<td>$\varepsilon_a = 0.653 - 0.00535 \cdot e_o$</td>
</tr>
<tr>
<td>Prata (1996)</td>
<td>$\varepsilon_a = 1 - \left[1 + 46.5 \left(\frac{e_o}{T_o}\right)\right] \exp\left[-\left[1.2 + 139.5 \left(\frac{e_o}{T_o}\right)\right]^{1/2}\right]$</td>
</tr>
<tr>
<td>Zapadka (2001)</td>
<td>$\varepsilon_a = 0.743 \left[1 - \exp\left(0.358 \cdot e_o\right)\right]$</td>
</tr>
</tbody>
</table>

1.2.3 Biophysical controls on ET

The major process that initiates ET is the sufficient moisture availability into the soil-plant system. The driving force for the transport of water vapor flux from the land surface to the atmosphere is the difference between the water vapor pressure of the evaporating surface and that of the bulk air. The vapor pressure and humidity of bulk air depend on the water content and air temperature, while those of the evaporating surface of plants depend on the water potential and leaf temperature (Kramer and Boyer, 1995). The suite of the complex meteorological, hydrological and crop factors affecting ET also includes: incoming solar radiation, wind speed, aerodynamic resistance, outgoing heat conduction into the soil, and
transpiration effects—concentration of CO₂, nutrient supply, Photosynthetically Active Radiation (PAR), Leaf Area Index (LAI), surface conductance, rooting depth, leaf density and other crop characteristics. Table 1.3 outlines some basic biophysical factors and their relation to ET.
Table 1.3: Main climatic, hydrological and plant factors affecting ET (modified from Verstraeten et al., 2008).

<table>
<thead>
<tr>
<th>Climatic and hydrological factors</th>
<th>Relation to evaporation (E) and transpiration (T)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moisture input and Soil Water Content (Precipitation and SWC)</td>
<td>ET increases with moisture availability. Hydraulic conductivity increases with SWC.</td>
</tr>
<tr>
<td>Net Radiation and incoming Shortwave Radiation $(R_n$ and $SW_{in}$)</td>
<td>E increases with $R_n$. T increases with $SW_{in}$. 1-5% of intercepted $SW_{in}$ is used for photosynthesis $^1$.</td>
</tr>
<tr>
<td>Vapor Pressure Deficit and Relative Humidity (VPD and RH)</td>
<td>ET increases with VPD and decreases with RH. But many plants close their stomata and limit T as VPD increases beyond a maximum value.</td>
</tr>
<tr>
<td>Atmospheric temperature $(T_a)$</td>
<td>Water amount in atmosphere increases with $T_a$. ET increases with $T_a$ if initially the surface is warmer than the air (sensible heat &gt;0) and decreases in the opposite case $^2$.</td>
</tr>
<tr>
<td>Wind speed $(w)$</td>
<td>This effect is complex. High $w$ increases E from water surface and decreases surface temperature. T varies with $w$, since high $w$ reduces the thickness and resistance of the boundary layer but also decreases leaf temperature and vapor pressure gradient from leaf to air. T will increase or decrease with $w$ depending on whether $\lambda E/H &gt; \Delta/\gamma$, respectively $^2$, and vary with the stomatal behavior of various plants $^3$.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Ecological factors</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Leaf Area Index; also leaf orientation and rolling, shape, structure, coating and other. (LAI)</td>
<td>T usually increases with LAI. Though soil E may increase with decreasing LAI due to unconcealed soil by the leaves.</td>
</tr>
<tr>
<td>Stomatal conductance $(g_c)$</td>
<td>Stomatal conductance depends mostly on light, as expressed by PAR, and moisture stress. Increased water stress (and VPD) results in stomatal closure, loss of turgor in the guard cells and leaf wilting $^1$, and hence reduces T.</td>
</tr>
<tr>
<td>Rooting density and depth</td>
<td>Plants with denser and deeper roots are more likely to reach water reserves.</td>
</tr>
</tbody>
</table>

$^1$ Verstraeten et al., 2008; $^2$ Monteith and Unsworth, 2008; $^3$ Kramer and Boyer, 1995.
ET is usually classified into two stages by considering the soil water content (SWC) availability (Brutsaert, 1991; Brutsaert and Chen, 1995). First stage evaporation (or potential evaporation, \( ET_{pot} \)) has adequate water such that the combined ET rate from the soil surface and the vegetation is limited by the available energy supply. Second stage evaporation occurs when the soil moisture drops below a critical limit and the evaporation rate is limited by the available soil water coupled with the available energy. Towards the end of the later stage, drying takes place only from the soil surface, since the vegetation is much stressed, and water limitation takes the dominant control on ET. The dependence of ET on SWC via a reduction factor \( \beta \) is shown in Equation (1.2) and illustrated in Figure 1.4.

\[
ET = \beta \cdot ET_{pot}
\]

where \( 0 \leq \beta \leq 1 \) if \( SWC_{wilt} \leq SWC \leq SWC_{cr} \).

Figure 1.4: Schematic of the ET versus SWC relationship for three major types of vegetations. \( SWC_{wilt} \), \( SWC_{cr} \), \( SWC_{fc} \) and \( SWC_{s} \) denote the values of SWC at the wilting point, critical point, field capacity and saturation (modified from Shuttleworth, 1993 and Brandes and Wilcox, 2000).
1.2.4 ET methods

Estimating ET (or $λE$) accurately is a difficult task due to the uncertainty involved in measuring the components associated with ET such as the available energy ($R_n - G$) and sensible heat ($H$). Note that the energy balance on the land surface is written as:

$$R_n - G = H + λE$$  \hspace{1cm} (1.3)

where $G$ is the soil heat flux. For example, accurate estimates of $H$ are very difficult to achieve, mainly when atmospheric effects and surface emissivity are not considered properly (Gowda et al., 2008). Moreover, the vast amount and complexity of data required to estimate these components make ET estimation even more difficult and costly.

ET can be measured with various devices and techniques that utilize physical-based principles. The most complex measurement technique is the Eddy Covariance (EC) method which requires extensive instrumentation of high frequency. In addition, several methods based on analytical or empirical approaches have been developed for estimating ET from meteorological and surface data. These approaches can be based on temperature, radiation or on a large set of data (combination methods). The combination methods require more data than the temperature and radiation models and hence they are more accurate in a variety of vegetative and meteorological conditions. The simplest of the combination/radiation models is the Priestley-Taylor (PT) equation, where the aerodynamic and surface considerations are incorporated into an environmental factor $α$. Table 1.4 presents a wide gamut of techniques used to assess ET including computer models which can utilize remote sensing data.
Table 1.4: A suite of ET techniques classified by the type of required data/applied principles. The spatial scale and some equations of methods are also presented; Table 1.4 is modified from Wallace (1995) and Verstraeten et al. (2008).

<table>
<thead>
<tr>
<th>Method Category</th>
<th>Examples</th>
<th>Scale</th>
<th>Description</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Humidity</td>
<td>Dalton</td>
<td>Point/leaf &amp; plant/field</td>
<td>Requires meteorological data ($T_\alpha$, $e_\alpha$, $w$).</td>
<td>Dalton, 1802</td>
</tr>
<tr>
<td>Temperature</td>
<td>Blaney-Criddle</td>
<td>&quot;</td>
<td>Requires meteorological data ($T_\alpha$).</td>
<td>Blaney and Criddle, 1950</td>
</tr>
<tr>
<td>Radiation</td>
<td>Makkink, Priestley-Taylor</td>
<td>One-component</td>
<td>Requires meteorological data ($R_n$, $T_\alpha$).</td>
<td>Makkink, 1957; Priestley and Taylor, 1972</td>
</tr>
<tr>
<td>Energy combined with mass (water) balance</td>
<td>Penman</td>
<td>&quot;</td>
<td>Requires meteorological data ($T_\alpha$, $e_\alpha$, $R_n$, $w$, $r_a$).</td>
<td>Penman, 1948</td>
</tr>
<tr>
<td></td>
<td>Penman-Monteith</td>
<td>&quot;</td>
<td>Requires meteorological and canopy data ($T_\alpha$, $e_\alpha$, $R_n$, $w$, $r_a$, $r_c$).</td>
<td>Monteith, 1965</td>
</tr>
<tr>
<td></td>
<td>FAO-24; FAO-56</td>
<td>&quot;</td>
<td>Based on PM model. Also require crop factors and SWC.</td>
<td>Doorenbos and Pruit, 1984; Allen et al., 1998</td>
</tr>
<tr>
<td></td>
<td>Sap flow</td>
<td>&quot;</td>
<td>Requires heat and temperature.</td>
<td>Kostner et al., 1992</td>
</tr>
<tr>
<td>Energy balance</td>
<td>Energy Balance</td>
<td>&quot;</td>
<td>Requires humidity and temperature at two heights to estimate $B$ and $\lambda E$.</td>
<td>Bowen, 1926</td>
</tr>
<tr>
<td></td>
<td>Bowen Ratio (EBBR)</td>
<td>&quot;</td>
<td></td>
<td>Bowen, 1926</td>
</tr>
<tr>
<td></td>
<td>Scintillometer</td>
<td>&quot;</td>
<td>Based on light propagation and atmospheric turbulence.</td>
<td>de Bruin et al., 1995</td>
</tr>
<tr>
<td>Method Category</td>
<td>Examples</td>
<td>Scale</td>
<td>Description</td>
<td>Reference</td>
</tr>
<tr>
<td>-----------------</td>
<td>---------------------------</td>
<td>----------------</td>
<td>------------------------------------------------------------------------------</td>
<td>----------------------------</td>
</tr>
<tr>
<td>Energy balance</td>
<td>Eddy covariance (EC)</td>
<td>Landscape</td>
<td>Requires covariance between 3D wind speed and humidity to determine $\lambda E$.</td>
<td>Swinbank, 1951</td>
</tr>
<tr>
<td>Mass (water)</td>
<td>Porometer</td>
<td>Point/leaf</td>
<td>Measurement of humidity and temperature to estimate water vapor loss from a leaf in a closed chamber.</td>
<td>Kanemasu et al., 1969</td>
</tr>
<tr>
<td></td>
<td>Water Balance</td>
<td>Plant/field &amp; Landscape</td>
<td>Measurement of water balance components.</td>
<td>Thornthwaite, 1948</td>
</tr>
<tr>
<td>Simulation models (energy-mass balance)</td>
<td>WAVE; SWAP</td>
<td>Plant/field</td>
<td>Simulation of the vertical water flow in the soil based on the Darcy flux law and mass conservation. Upper and lower boundary data are required (such as ET$_{pot}$, rainfall, groundwater level).</td>
<td>Vanclouoster at al., 1996; Van Dam et al., 1997</td>
</tr>
<tr>
<td></td>
<td>SWAP; MIKE-SHE; SWAT; SEBAL</td>
<td>Landscape</td>
<td>1-2-3 D water fluxes in the soil compartment applied on a grid or using hydrological response units. Upper and lower boundary conditions are required.</td>
<td>Van Dam et al., 1997; DHI,1999; Arnold et al., 1998; Bastiaanssen et al., 1998</td>
</tr>
<tr>
<td></td>
<td>SEBAL; PROMET</td>
<td>Regional &amp; Continental</td>
<td>Include remote sensing data from optical, thermal and microwave sensors.</td>
<td>Bastiaanssen et al., 1998; Mauser and Schadlich, 1998</td>
</tr>
</tbody>
</table>

Table 1.4 Equations:

(1) Priestley-Taylor: $\lambda E = \alpha \frac{\Delta (R_n - G)}{\Delta + \gamma}$

(2) Penman: $\lambda E = \frac{\Delta (R_n - G) + \frac{\rho_a C_p (e_s - e)}{r_a}}{\Delta + \gamma}$

(3) Penman-Monteith: $\lambda E = \frac{\Delta (R_n - G) + \frac{\rho_a C_p (e_s - e)}{r_a}}{\Delta + \gamma \cdot \frac{(r_c + r_a)}{r_a}}$

(4) Energy Balance-Bowen Ratio: $\lambda E = \frac{R_n - G}{1 + B}$ where $B = \frac{\delta T_a}{\delta e}$

(5) Eddy Covariance: $\lambda E = \lambda \rho_a \bar{w}'q'$ and $H = \rho_a C_p \bar{w}'T'_a$

Where

$\lambda E$ is the latent heat flux ($W m^{-2}$), $\lambda$ is the latent heat of vaporization of water ($J g^{-1}$), $H$ is the sensible heat flux ($W m^{-2}$), $R_n$ is the net radiation ($W m^{-2}$), $G$ is the soil heat flux ($W m^{-2}$), $\Delta$ is the slope of the saturation vapor-pressure curve dependent on air temperature ($KPa °C^{-1}$), $\gamma$ is the psychrometric constant dependent on atmospheric pressure and temperature ($KPa °C^{-1}$), $\alpha$ is the PT coefficient, $\rho_a$ is the moist air density dependent on air temperature, air pressure and vapor pressure ($g m^{-3}$), $C_p$ is the specific heat of air dependent on specific humidity ($J g^{-1} °C^{-1}$), $e_s$ is the saturation water vapor pressure dependent on air temperature ($KPa$), $e = e_s \cdot RH$ is the water vapor pressure ($KPa$), $RH$ is the relative humidity, $r_a$ is the aerodynamic resistance ($s m^{-1}$), $r_c$ is the canopy resistance ($s m^{-1}$), $B$ is the Bowen ratio, $T_a$ is the air temperature ($°C$), $\delta T_a$ and $\delta e$ are air temperature and vapor pressure differences between two heights above the canopy, $w$ is the vertical wind speed ($m s^{-1}$), and $q$ is the specific humidity ($g g^{-1}$). Bars and primes on the variables of the EC technique denote means over the sampling period and deviations of the mean, respectively.

The PT model is a simplification of the Penman method with the hypothesis of a saturated atmosphere ($e = e_s$), and a further reduction of the Penman-Monteith (PM)
equation with the additional assumption of the canopy resistance being negligible \( r_c = 0 \).
The PT coefficient \( \alpha \) is usually equal to 1.26 at wet surfaces. Several ET studies (Flint and Childs, 1991; Stannard, 1993) relaxed the assumption of a dense, well-watered canopy by allowing \( \alpha \) being less than 1.26. The PT approach has been reported to outperform the PM model in wetland and grass sites as it requires less meteorological data, is computationally more efficient, and simulates ET successfully (Stagnitti et al., 1989; Stannard, 1993; Sumner, 1996; Sumner and Jacobs, 2005). Stannard (1993) suggested that the PT approach was superior to the PM model when he compared both models to evapotranspiration measured over wildland vegetation in a semiarid area. This conclusion was also supported by the studies of Sumner (1996) and Sumner and Jacobs (2005), in which the two aforementioned methods were tested over grasslands located in Lake Wales Ridge and Floral City, respectively, in Central Florida. Such studies justify the suitability of the PT model for the ET analyses of the grass ecosystem examined in Chapter 3.

1.2.5 Interactions between atmosphere, land and vegetation

The main plant characteristics that cause interaction with the atmosphere are: (1) the absorption and reflection of incoming shortwave radiation and emission of longwave radiation by the vegetation; (2) the vegetation’s physical presence, which affects the roughness length; (3) the plant’s transpiration, which generates \( \lambda E \); and (4) the plant’s photosynthesis, which generates CO₂ flux (McPherson, 2007). Anthes (1984) indicated that an increase in green vegetation increases the following surface-atmospheric variables: \( \lambda E \),
atmospheric humidity, surface emissivity, absorption of SW radiation, roughness length, and turbulence. Greener LAI may be also coincident with higher probability of cloud formation and convective rainfall (Freedman et al., 2001). On the other hand, an increase in vegetation cover decreases the sensible heat flux and Bowen ratio, diurnal temperature range, surface albedo, emission of LW radiation, surface winds, and runoff (Anthes, 1984).

The radiative effects of vegetation in terms of SW and LW fluxes are illustrated in Figure 1.5 (a). McPherson (2007) documented that healthy green vegetation with adequate leaf water content absorbs strongly SW radiation and scatters most of the downward LW radiation. However, as the plants mature and produce senescent material (brown LAI) the reflectance of SW flux from grasses increases. In the case of sparse canopies that leave the bare soil exposed, the albedo is mainly influenced by the soil moisture (Table 1.1). Moreover field practices such as grazing and burning, which tend to decrease the leaf density (defoliate) and expose the bare soil, cause an increase in reflectance of SW and a reduction in emissivity of LW radiation. Figure 1.5 (b) shows the energy interactions with the plant-surface system. As previously mentioned greener LAI increases $\lambda E$ and reduces $H$. The Bowen ratio ($B = H/\lambda E$), which is an indicator of both the environmental stress and the vegetation conditions, ranges from infinity under very dry conditions to almost zero in wet regions. The distribution and magnitude of the surface energy fluxes depend on: (a) the magnitude of net radiation and cloud cover; (b) the moisture availability; and (c) the type, development stage and LAI of vegetation.
Figure 1.5: SW and LW radiation (a) along with latent and sensible heat (b) fluxes in the air-land-vegetation system (adopted from McPherson, 2007).
The interactions between the energy exchange components and relevant climatic variables exhibit nonlinear complex behavior (Figure 1.6). This complexity becomes even higher when considering the effects of inhomogeneous landscape and shifts in vegetation. The positive (solid arrows) and negative (dotted arrows) feedbacks are justified on the basis of: (1) corollary of the radiation and energy budget equations, Equation (1.1) and Equation (1.3), respectively; (2) the formulae from which the variables are defined or calculated (e.g. by definition the soil heat flux is proportional to the soil temperature above the heat flux plate); (3) observations (e.g. Jacobs et al. (2002) reported that increased cloudiness attenuates the net radiation as measured by ground pyranometers and estimated by GOES images); or (4) numerical simulations (e.g. Vidale et al., 1997).

Figure 1.6: Interactions between energy fluxes and related climatic components (modified from Eugster et al., 2000).
1.3 Dissertation Objectives and Organization

This dissertation addresses two main objectives: (1) proposes a new downward longwave radiation model that is land use dependent; and (2) examines evapotranspiration and energy partitioning over a Floridian grassland at an annual time-scale with respect to the controlling factors on evapotranspiration dynamics. More specifically, this work, which consists of three concatenated independent studies, sought to:

1. Develop a new air emissivity model (*Chapter 2*). It is noteworthy that the existing longwave radiation models only use atmospheric variables as input.
   - Present a record of downward longwave radiation ground measurements above various land uses in Florida.
   - Investigate the variations of $LW_d$ radiation on different landscapes and introduce a new clear sky effective emissivity model. This model has fixed coefficients but allows one free variable to adjust to the land use type.
   - Verify the new emissivity model by performing comparisons to existing formulations based on data collected in the SJRWMD region, and validate the new model with $LW_d$ fluxes measured over various land uses in different geographic regions.

2. Identify the main biotic and abiotic controls of evapotranspiration, and propose wetness parameterizations for the Priestley-Taylor (PT) parameter over a non-irrigated grassland in Central Florida (*Chapter 3*).
   - Present year-long, daytime mean estimates of measured and potential ET rates using EC and PT methods, respectively.
• Quantify the temporal response of ET to diminishing wetness, namely SWC and Current Precipitation Index (CPI), within the wet season.

• Investigate the significant canopy and environmental effects on the water vapor exchange over the grass site on a yearly basis.

• Present simple parameterizations for the PT coefficient in terms of SWC and CPI, which show good agreement with the actual data over the study site.

3. Characterize the variations of energy partitioning and biophysical controls over the grass ecosystem (Chapter 4).

• Investigate the major biotic and environmental controls on grass evapotranspiration during the energy- and water-limited stages of ET.

• Describe the diurnal patterns of the energy budget components.

• Characterize the seasonal patterns of the energy balance and determine the main biophysical factors that modulate the energy partitioning on seasonal time-scale.

• Describe the diurnal and seasonal cycles of canopy characteristics, mainly canopy conductance and decoupling coefficient, and also relate them to environmental conditions and stomatal control of ET.
References


2.1 Introduction and Objectives

2.1.1 Introduction

A precise and long-term knowledge of the downwelling longwave (LW) radiation is necessary in forecasting temperature variation and cloudiness, evaluating parameterization schemes used in climate models, and estimating climate change and global warming. This knowledge is also valuable when studying the earth energy budget, assessing the status of ecological systems in terms of photosynthesis and crop growth, and improving management of water resources or energy systems.

Many of the existing downward LW radiation formulas are derived from ground spot observations recorded on specific time of day on a sparse geographical scale. On the other hand, the instantaneous satellite radiation estimates are spatial averages of an extensive area, composed of different scatterers. Since 1980’s, there has been a progress in understanding and modeling the influence and importance of vegetation and nature of landscape on surface exchanges of energy, water, and carbon (Pielke and Avissar, 1990; Schneider and Eugster, 2005). The land use feedback on surface climatology was examined in several studies by
using synoptic meteorological observations (Holmer and Eliasson, 1999) or Advanced Very High Resolution Radiometer (AVHRR) images (Carlson and Arthur, 2000). There is an imminent need to conduct further research on the deficiencies of the ground and satellite radiation measurements and algorithms, explain any discrepancies between the two, and attempt to correlate the radiation values with the land surface conditions.

Sellers et al. (1990) suggested that estimating the four components of net radiation from satellite algorithms could cause error accumulation especially in estimating the net LW flux because both downwelling and upwelling components of LW radiation are large, so the difference would be small and liable to wide uncertainty. Diak et al. (2000) noticed that great discrepancies between satellite and pyrgeometer LW radiation values are likely in semi-arid or bare-soil areas under clear conditions where significant temperature gradients above the land surface are present. It was suggested that an effort for increasing the accuracy of the upward LW flux would rely on any information predicting the land surface temperature (versus air temperature), which is dependent on the nature of the landscape. On the other hand, Stanhill and Cohen (1997), who analyzed long-term time series of radiation measurements in the Antarctic, concluded that changes in surface temperature cannot be explained satisfactorily without measurements of LW radiation components.

Routine ground-based measurements of some radiation components—mainly total incoming shortwave (SW) radiation—or their net balance take place above specific land types such as grass and short vegetation in common climates, bare soils in arid areas, snow in cold regions, forests, few crops, and sea locations, while they usually exclude urban and water-covered areas (Kessler, 1985; Kessler and Jaeger 1999; Barr and Sisterson, 2000; GAPP project, accessed December 2005, http://www.ogp.noaa.gov/mpe/gapp/). It is apparent
that more attention needs to be paid to the long-term monitoring and modeling of the radiation components, especially the LW flux over various types of land surfaces simultaneously.

Moreover, the variable land characteristics (mainly due to urbanization and changes in forest cover and cultivated land) affect not only the regional atmosphere but also the observed and simulated global climate depending on the areal extent of the land use change (Kessler and Jaeger, 2003). Pielke and Avisar (1990) reported that, in simulations performed by Shukla and Mintz (1982), the ground surface temperature was 15 to 25°C higher, the precipitation was about four times less, and the surface atmospheric pressure was about 50 to 150 hPa lower over dry than wet soils of most of North America.

In conclusion, the LW fluxes and surface meteorology interaction needs to be further investigated by developing an expanded grid of LW radiation measurements and a continuous mapping of the surface vegetation and soil characteristics.

2.1.2 Existing formulations and objectives

The clear sky longwave emissivity can be concisely determined as follows (Brutsaert, 1982):

\[
\varepsilon_a = \frac{LW_d}{\sigma T_o^4}
\]

(2.1)

where \(\varepsilon_a\) is the effective atmospheric emissivity, \(\sigma = 5.67 \times 10^{-8} \text{ W/m}^2\text{K}^4\) is the Stefan–Boltzmann constant, \(T_o\text{ (K)}\) is the surface or screen-level air temperature, and \(LW_d\text{ (W/m}^2\) is the downward LW radiation. Many investigators presented parameterizations for this ratio of actual to potential atmospheric LW radiation, based on empirical relationships derived from
Angstrom’s (1918) equation, which was the first emissivity parameterization, used an exponential water vapor pressure term. Swinbank (Swinbank, 1963; Deacon, 1970) used a quadratic temperature term to perform emissivity formulations based on monthly mean values of water vapor and temperature. When daily values were used in his model, the correlation between the air temperature and the amount of precipitable water vapor was found to be smaller. Idso and Jackson (1969) developed a model with an exponential quadratic temperature term that allows the emissivity to increase bidirectional about the temperature of 273 K.

Brutsaert (1975) derived a clear sky emissivity formula based on the radiative transfer equation by assuming a standard atmosphere. The main characteristics of his model, which is a power function of the humidity parameter \( \frac{e_o}{T_o} \), where \( e_o \) is the water vapor pressure, are its insensitivity to the temperature effect and the shortcoming of not considering a residual emissivity due to gases such as CO\(_2\) and O\(_3\). Prata (1996), who also adapted the humidity parameter in his quasi-empirical model, overcame the last limitation by setting a non-zero lower limit to account for emissivity due to aerosols, and upper limit of a unity to assure that the downward radiation never exceeds the blackbody radiation at the same temperature. However, the lower limit in the absence of the vapor path leads to temperature-invariant (constant) emissivity in his model.

Other models used individual terms for water vapor pressure and temperature (Satterlund, 1979; Idso, 1981). In Satterlund’s (1979) formula, the emissivity at low temperatures becomes solely function of the vapor pressure. Idso’s (1981) equation violates
the upper limit of black body radiation for below freezing temperatures. A complete
evaluation of the clear sky emissivity models was done by Prata (1996).

The vast majority of the empirical models are based on data of different geographic
regions, mostly land (Aase and Idso, 1978) or sea (Bignami et al., 1995; Zapadka et al.,
2001), but the model performance depends on the atmospheric conditions that prevail locally.
The equations of the discussed models are presented in Table 1.2 of Chapter 1.
The objectives of this study are to:

1. Present some results—available up to date—of LW_d radiation ground measurements
carried out by Saint Jones River Water Management District (SJRWMDS) above
various land uses. These land uses include among others, urban sites in the vicinity of
anthropogenically altered surfaces such as streets and buildings, and wetland surfaces.
The incoming LW flux on these surfaces differs from the one measured on the
standard grass surface sites on diurnal, seasonal and annual time scale.

2. Investigate the variations of LW_d radiation on different landscapes and introduce a
new clear sky daily emissivity model. This model has fixed coefficients and allows
one free variable to adjust to the land use type.

3. Verify the new emissivity model by performing comparisons to the other
formulations based on data in the SJRWMD region, and validate the new model with
LW_d fluxes measured over various land uses in different geographic regions.
2.2 Data and Measurements

2.2.1 Experimental sites and database

The data utilized for the model development covered an almost three-month period in spring 2004 (Julian days 75-144) and came from two sources: weather data, namely air temperature, dew point temperature and cloud cover from the National Climatic Data Center of National Oceanic and Atmospheric Administration (NOAA, accessed March 2005, http://www.ncdc.noaa.gov/o/accdc.html), and LW\textsubscript{d} radiation from ground radiometers installed in the SJRWMD region.

The ground radiation stations, namely CNR1 stations equipped with CNR1 Net Radiometers, are strategically positioned inside the SJRWMD in Florida, and represent different land uses (urban, agricultural, rangeland, forest, open water and wetland) as illustrated in Figure 2.1 and Figure 2.2. Their geographic location is spread over latitudes of 27.58° to 30.32° North and longitudes of 80.60° to 82.07° West.

The Deland radiation sensor, installed in a wastewater treatment plant, represents an urban land use. In the vicinity there is a paved road, grass, few trees and shrubs. The Jarboe Park sensor represents a lower density urban land use and is located on the edge of a city park; there is irrigated grass, a paved road and a small pond in the surrounding area. Orange Creek has a rangeland primary land use type. It is covered by bahia grass including oak and pine trees in the adjacent perimeter of 1 km. Ocklawaha Prairie is a wetland covered by willow, saw grass, cattail, lily pads and wire grass. There is also mixed forest in the 1 km vicinity of the wetland. The sensor is installed on the side of an observatory dock that leads to the marsh. Lindsey Citrus is an agricultural site (citrus grove) with short grass beneath the
tree canopy, which is under regular irrigation schedule. Additional information for the CNR1 and the adjacent NOAA stations is provided in Table 2.1.
Figure 2.1: Location of the CNR1 and weather stations in the SJRWMD region.
Figure 2.2: Radiation sensors installed on two diverse land use environments, wetland and residential: a) Ocklawaha Prairie and b) Jarboe Park. The aerial maps are shown on the left and the CNR1 sensors on the right. The circles with a radius of 20 m show the footprint of the radiation signal. The aerial maps are supplied by Google (Google Local, accessed January 2006, http://www.google.com).
Table 2.1: Site information for the CNR1 and NOAA stations along with CG3 sensor characteristics (CNR1 stations with data of bad quality or insufficient clear day sample are not shown).

<table>
<thead>
<tr>
<th>CNR1 station</th>
<th>Location</th>
<th>Lat/Long (°)</th>
<th>Primary landuse</th>
<th>Landuse within 1 km radius</th>
<th>Elevation of CG3 sensor above land (m)</th>
<th>Temporal resolution (min)</th>
<th>Factory accuracy</th>
<th>NOAA station</th>
<th>Latitude distance to CNR1 (km)</th>
<th>Elevation a.s.l. (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DeLand STP</td>
<td>Wastewater treatment plant-DeLand, Volusia Co.</td>
<td>29.01/-81.30</td>
<td>Residential Density population =1,298 people/m²</td>
<td>95% urban and 5% mixed forest</td>
<td>2.0</td>
<td>30</td>
<td>±20W/m², ±10%</td>
<td>Orlando Sanford Airport</td>
<td>25.2</td>
<td>15.0</td>
</tr>
<tr>
<td>Jarboe Park</td>
<td>Edge of a city park-Neptune Beach, Duval Co.</td>
<td>30.32/-81.40</td>
<td>Residential Density population =1,069 people/m²</td>
<td>95% urban and 5% recreational</td>
<td>2.0</td>
<td>30</td>
<td>±20W/m², ±10%</td>
<td>Jacksonville Craig Municipal Airport</td>
<td>1.9</td>
<td>13.5</td>
</tr>
<tr>
<td>Lindsey Citrus</td>
<td>Citrus grove, Indian River Co.</td>
<td>27.58/-80.60</td>
<td>Citrus</td>
<td>100% citrus</td>
<td>6.0</td>
<td>30</td>
<td>±20W/m², ±10%</td>
<td>Vero Beach Municipal Airport</td>
<td>7.8</td>
<td>6.0</td>
</tr>
<tr>
<td>Orange Creek</td>
<td>District land, Alachua Co.</td>
<td>29.48/-82.07</td>
<td>Rangeland (bahia grass)</td>
<td>50% rangeland and 50% mixed forest</td>
<td>2.0</td>
<td>30</td>
<td>±20W/m², ±10%</td>
<td>Gainesville Regional Airport</td>
<td>27.0</td>
<td>45.5</td>
</tr>
<tr>
<td>Ocklawaha Prairie</td>
<td>District land, Marion Co.</td>
<td>29.10/-81.91</td>
<td>Wetland (cattail, sawgrass, and other aquatic vegetation)</td>
<td>60% wetland, 20% rangeland and 20% mixed forest</td>
<td>2.0</td>
<td>30</td>
<td>±20W/m², ±10%</td>
<td>Leesburg Municipal Airport</td>
<td>31.4</td>
<td>23.5</td>
</tr>
</tbody>
</table>

* according to United States Census Bureau's 2000 Census.

** ±20W/m² for individual measurements, ±10% for daily totals (Kipp & Zonen, 2000).

† the radiation field does not change significantly in a homogeneous region, where the north-south distance to a weather station is less than 50 km (Allen et al., 1998).
2.2.1.a CNR1 radiation

Unlike the conventional Radiation and Energy Balance Systems (REBS) radiometers, the CNR1 radiation sensors—manufactured from Kipp and Zonen (K&Z) — measure all four components of radiation, namely, incoming and reflected solar radiation, and incoming and outgoing LW radiation. The CNR1 sensor is a four-component radiation system housing upward-facing and downward-facing pyranometers (CM3 radiometers) and pyrgeometers (CG3 radiometers with spectral range 5–50 \( \mu \text{m} \)). Some of the CG3 sensor characteristics are shown in Table 2.1.

The manufacturer-provided expected accuracy of the CG3 sensor is ±10% for daily totals and ±20 W/m\(^2\) for individual measurements (Kipp & Zonen, 2000). SJRWMD sets an even more stringent target for the CG3 sensor error, which is ±3% accuracy, during the quality assurance process (described below). Several intercomparison studies have been conducted acknowledging the fact that there are no international standards for pyrgeometers. Brotzge and Duchon (2000) showed that large differences can be observed between K&Z NR-Lite single net radiometers and Eppley four-component radiometers, which are mainly the result of the influence of dew, wind and solar contamination. Nevertheless, van den Broeke et al. (2004a) compared radiation measurements by the K&Z CG3 sensor and the Eppley PIR pyrgeometer (as reference sensor) under controlled conditions at a station in Antarctica. The above comparison yielded a RMS difference of 1.2% (2.7 W/m\(^2\)) for the daily average LW\(_d\) radiation showing that the CG3 sensor performs much better than the manufacturer’s specification (±10% accuracy).
Some CNR1 stations with data coded as low quality or yielding insufficient clear day sample size, during the three-month period, were omitted and not considered in this study. The half-hourly ground-based $LW_d$ data were processed in order to filter radiation errors and then were averaged over a clear day. The range of the CNR1 $LW_d$ radiation data is shown in Table A.1 of the Appendix A.

The quality assurance and control of the CNR1 data is performed through the following steps (Robinson G., 2006, personal communications, SJRWMD, Palatka, Florida):

- Simultaneous measurements are collected from the field and a reference sensor and they are compared with each other twice a month. Measurements that differ by more than $\pm 3\%$ are documented and the sensor is closely monitored to determine if recalibration is needed.

- Collected data are compared with data from other regional sensors for consistency. The incoming LW radiation data are also compared with incoming SW radiation data. Peaks in LW that coincide with nadirs in SW radiation are usually indicative of a shadowing effect on the sensor. These data are removed from the dataset.

- In addition, LW data are also compared with sensor temperature data and low battery voltage reports. During periods when the sensor heater has been deactivated due to low battery reading, the LW data are compared with incoming SW data.

- When precipitation and dew evaporate from the sensor surface, residual debris remains, which can alter the absorptive or reflective properties. In addition, water droplets on the sensor dome may refract or reflect radiation, creating spurious values. If conditions indicate possible water formation on the dome, the data are coded unverifiable.
2.2.1.b Air temperature, water vapor pressure and cloud fraction

Considering limitations such as the poor vertical resolution of the water vapor data (AGU, 1995), the humidity and air temperature only at screen level were used for the emissivity model development having the underlying assumption of a homogeneous surface atmospheric slab of thickness of several hundred meters. The atmospheric structure of higher levels is of less importance for clear sky LW flux calculations at the surface (Prata, 1996; Diak et al., 2000). The NOAA data are part of the Weather Bureau Army and Navy (WBAN) network. At almost all stations, the vicinity to the CNR1 radiometer locations varies from 1 to 17 minutes of latitude (Table 2.1).

The Automated Surface Observing System (ASOS) temperature measurements are made by a HO-83 hygrothermometer, which uses a resistive temperature device (Root Mean Square Error (RMSE): 0.5°C, max error: 1°C) to measure air temperature, and a chilled mirror (RMSE: 0.6-2.6°C, max error: 1.1-4.4°C) to measure dew point temperature. The sampling frequency is 1 min and the averaging interval is 5 min. The cloud amount is determined by a laser beam ceilometer with a vertical reporting range of 3,600 m—at this height the beam’s width is 18 m. The ASOS cloud sensor operates at a wavelength of 0.9 microns, and it has a nominal pulse frequency of 770 Hz, sampling frequency of 30 sec and averaging interval of 30 min. The cloud fraction is recorded in oktas with a maximum error of 5% (ASOS program, 1998).

From each set of data within the three-month period, a sample is selected that consists of “almost clear days”. The sample size varies from ten to twenty-six clear days across the stations. Daily averaging was applied on the air and dew point temperature NOAA data. The
water vapor pressure at the surface can be calculated with an equivalent expression given by Shuttleworth (1993):

\[
e_o = 6.1078 \cdot 10^{\frac{7.5 T_d}{T_o + 237.3}}
\]

(2.2)

where \( e_o \) (hPa) is the actual water vapor pressure at the surface and \( T_d \) (°C) is the dew point temperature. The range of temperature and water vapor pressure data used for the model development is presented in Figure 2.10 of Section 2.5 and Table A.1 of the Appendix A.

2.2.1.1 Collocation of radiation and temperature/humidity values

The following assumptions ensure consistency of the radiation measured at each CNR1 site with the meteorological data at the adjacent NOAA station, which is located closer than 50 km in latitude (Table 2.1). The study region is small and thus the horizontal variability of the air masses controlling cloudiness and vapor pressure is minimal; likewise the physiography of the land surface is almost homogeneous within each study region. Allen et al. (1998) proposed a method for replacing missing data required in the daily ET calculation applied to a lot of studies. Trnka et al. (2005) remarked that the explained variability between the proxy (approximated based on nearby site) and measured daily global solar radiation values in Central Europe decreases by about 1.3% per 10 km distance to the adjacent station.

Though, the transferability of data should be done carefully when convective conditions are dominant in the region (Allen et al., 1998). In fact, spring season in Florida is characterized by some convective activity that influences mostly the cloud cover and LW radiation. The transferable quantities of water vapor pressure and air temperature for clear
skies are more homogeneous than the radiation within the study domain (Sumner D., 2006, personal communication, USGS, Orlando, Florida).

Supporting evidence on the atmospheric homogeneity of the region containing the CNR1 and the corresponding NOAA station, with regard to the radiation and meteorological variables, can be found in Table A.1 of the Appendix A. It is observed that extreme air temperature and/or water vapor pressure at the NOAA station results to a corresponding extreme LWd flux at the CNR1 site during the same clear day at selected locations.

2.2.2 Validation sites and database

The data utilized for the model validation represent three different land uses, agriculture, rangeland and urban, that are common types in the continental U.S. Two datasets are provided by the GAPP program (Gewex America Prediction Project) which is part of the GEWEX (Global Energy and Water cycle EXperiment). Observation variables at the GEWEX air SURFace eXchange (SURFX) sites, which include the fluxes of energy and carbon along with surface meteorology, are supplied from a web source (EOP, accessed January 2006, http://www.joss.ucar.edu). The time coinciding with the initiation of the GAPP project contains hourly meteorological and radiation data for the period of 1 July to 30 September 2001. The cloud cover, which is the criterion for screening the clear days, was supplied by the nearest NOAA station. The sample size varies from eighteen to forty-eight clear days.

One validation SURFX site is located at Bondville, Illinois (40.01 °N, 88.29 °W) and represents an agricultural setting with corn and soybeans. The nearest NOAA station is located at Champaign/Urbana Willard Airport and has an elevation of 230 m a.s.l. An
additional SURFX site is at Fort Peck Reservation, Montana (48.31°N, 105.10°W), and it has rangeland land use. The cloud cover is recorded at an adjacent NOAA station in Wolf Point Clayton Field, MT with an elevation of about 604 m a.s.l. Another LW_d radiation validation dataset is from the CNR1 station at Deland, Florida and covers the period from 1 January to 5 June 2005. Orlando Sanford Airport is the nearest NOAA station that provides air and dew point temperature and cloud cover records (other station characteristics are shown in Table 2.1). The range of the clear sky meteorological and radiation data over the above sites is shown in Table A.1 of the Appendix A.

2.3 Emissivity Model Development

2.3.1 All-sky comparison with GOES radiation

A limited LW_d radiation from Geostationary Operational Environmental Satellite (GOES-East) supplied by Martha Anderson (Anderson M., 2004, personal communication, Department of Soil Science, University of Wisconsin-Madison, Wisconsin), was used for selective comparison with the ground truth CNR1 radiation during all-sky days. The cloud cover was extracted from the nearest NOAA stations as previously mentioned.

The satellite product is computed with the use of a radiative transfer model (Chou and Arking, 1980), which was developed by CIMSS (Cooperative Institute for Meteorological Satellite Studies) Regional Assimilation System based on atmospheric profiles of temperature and relative humidity. The spatial resolution is approximately 20 km and the temporal resolution is 1 hr. More details on the derivation and validation of the synthetic LW radiation database are provided by Diak et al. (2000).
Ground-sensed downwelling LW radiation shows evidence of consistency with the satellite radiation especially during clear days (see also Jacobs et al., 2004). This agreement verifies the quality of both databases and supports the utilization of the CNR1 LW$_d$ radiation as a reliable basis for any improvement in radiation calculations. The presence of clouds results in warmer air temperatures, thus increases the LW$_d$ radiation, and yields more noise in the diurnal pattern of radiation.

The ground and satellite all-sky LW$_d$ radiation measured at a wetland site, Ocklawaha Prairie, for a week in March 2004, is shown in Figure 2.3. Overall during “almost” clear days, the satellite values overestimate the ground-based LW$_d$ estimates. The clear sky data exhibit a peak between 1400 and 1800 hours. During cloudy days, large diurnal and interdiurnal variability is introduced in the LW$_d$ radiation due to reasons described in relevant literature (Gu et al., 1999; Offerle et al., 2003; van den Broeke et al., 2004b). The modeled LW$_d$ radiation, which is calculated for the clear days according to the model of Section 2.4, follows well the diurnal cycle of CNR1 flux.
Figure 2.3: Ground, satellite and modeled downwelling LW radiation at Ocklawaha Prairie. The modeled LW$_d$ radiation is presented for the clear days.

2.3.2 Background and regression modeling

The main goal of this study is to examine the land use effect on LW$_d$ radiation and develop a land use adapted model. The above was implemented based on clear sky radiation meteorology. That was also the case in many field studies (Oke and Fuggle, 1972; Aida and Yaji, 1979; Kobayashi, 1982; Estournel et al., 1983; Lindgren, 1997; Holmer and Eliasson, 1999), which examined the urban-rural differences in LW$_d$ fluxes of clear skies, since the cloudy atmosphere weakens the urban heat island intensity and keeps the net LW radiation close to zero for various land covers. Besides, the dynamic variability of the cloud properties, especially in the tropics, makes the prediction of the incoming LW flux a complicated task,
and leads to radiation estimates that are not well correlated to the ground measured values (Gu et al., 1999; Offerle et al., 2003). This might have got even more complex, if land use differences had to be considered for cloudy days in an effective LWd radiation model.

A starting point for developing a new effective clear sky emissivity model is the Beer’s law applied in a homogeneous slab (thin medium) where the absorption coefficient is wavelength-invariant. At the slab output (land surface) the intensity loss (absorbed wave) is an exponential function of the slab’s optical thickness and the complementary function is the intensity gain (emitted wave). The following equation holds true for the slab emissivity (Elachi, 1987):

\[ \varepsilon_s(D) = 1 - I_o e^{-aD} \]  

where \( I_o \) is the incoming wave intensity, \( a \) is the total extinction coefficient (including absorption and emission) and \( D \) is the slab thickness. The term \( aD \) is usually called the optical thickness or depth.

The positive effect of temperature \( (T_o) \) on the water vapor \( (e_o) \) is widely affirmed (Brutsaert, 1975; Makarieva et al., 2003), especially at many places near sea level (Deacon, 1963). In addition, theory suggests that the optical depth is a positive function of the water vapor pressure and so is the air emissivity according to Equation (2.3). For that reason, Angstrom (1918) used an exponential emissivity model in terms of water vapor pressure. Similarly, Idso and Jackson (1969) found that the air emissivity is an exponential positive function of the air temperature.

However, by omitting one of the highly correlated atmospheric effects (mainly \( e_o \)), or combining those into a confounding parameter, any information about the importance of that
effect in predicting the air emissivity will not be entirely revealed, and the model accuracy over wide geographic locations may be reduced (Hatfield et al., 1983; Iziomon et al., 2003). Even the models that use the humidity parameter, such as Brutsaert’s (1975) and Prata’s (1996) models, may not be able to capture all LW\textsubscript{d} radiation over wide climates due to the compensating effect of $T_o$ and $e_o$ (Prata, 1996). Based on the hypothesis that the two effects (though interrelated) can be superpositioned, their concurrent positive exponential effect on the emissivity is concluded. The following empirical equation for the daily mean emissivity can be generated:

$$
\varepsilon = 1 - \left( C_1 e^{-T_o/C_2} + C_3 e^{-e_o/C_4} \right)
$$

(2.4)

where $C_1$, $C_2$, $C_3$ and $C_4$ are site-specific constants. With the use of multiple nonlinear regression analysis the values of the parameters were obtained for all sites.

The output emissivity for the site of Lindsey Citrus has regression parameters as follows: $C_1=40.1$, $C_2=52$, $C_3=0.2$ and $C_4=15$. Figure 2.4, which illustrates modeled emissivity isolines for Lindsey Citrus, shows that the clear sky emissivity is a positive function of the two effects, $T_o$ and $e_o$. The bivariate function of emissivity in Equation (2.4) satisfies the general criterion for superposition, which is described by \( \frac{\partial^2 \varepsilon}{\partial e_o \partial T_o} = 0 \) (Ott, 1978). The iso-emissivity curves show increasing nonlinearity as the $T_o$ and $e_o$ effects become larger, except for the supersaturation vapor pressure region (shaded area in Figure 2.4).
Figure 2.4: The iso-emissivity lines as a function of $e_0$ and $T_0$ by using the developed model along with overlaid actual emissivity for Lindsey Citrus. The shaded area indicates above saturation vapor pressure.

2.4 Land Use Enhanced Emissivity Model

Intercomparisons of the ground-sensed daily LW$_d$ radiation were attempted for detecting similar parameterization behavior within each land use class, specifically for the two urban, two rangeland, and two wetland land use sites. Although the two sites within each land use group are located from 1.2 ° to 1.4 ° of latitude apart, they exhibit similar measured LW$_d$
radiation trends on the same clear days. To justify this similarity, the Cochran-Cox t-test is applied to samples of LW_d flux measured over the two locations of each land use group during coincident clear days. The conditions of the two radiation samples being independent and following normal probability distribution are met. There is sufficient evidence not to reject the null hypothesis of equality of the means at all cases within 95% confidence interval (\( \alpha=0.05 \)).

Specifically, there is strong evidence that the rangeland measured radiation means are equal, i.e. the two-tailed \( p\)-value is 0.489 which is greater than the significance level \( \alpha \). Similarly, the wetland radiation means have also large \( p\)-value of 0.349 and the urban sites pass the test marginally with \( p\)-value of 0.088.

For the purpose of establishing the aforementioned similar radiation behavior within each land use class, and due to the existence of an agricultural site with “ideal” conditions such as dense grass cover, scheduled irrigation, and radiation sensor installed above the canopy height, another cycle of regression runs were performed. In these simulations, Lindsey Citrus was the “reference” site from where the coefficients of Equation (2.4) are obtained and set fixed. Then, an offset factor \( (\alpha_{LU}) \) is introduced to adapt for different land use as follows:

\[
\varepsilon_a = 1 - \left(1 - \alpha_{LU}\right) \left(40.1 \cdot e^{-T_o/52} + 0.2 \cdot e^{-e_o/15}\right)
\]

(2.5)

The \( \alpha_{LU} \) factor represents an amplification (in the case of urban setting) or reduction (in the case of wetland) applied on the atmospheric effects, \( T_o \) and \( e_o \), and eventually on the effective air emissivity. Equation (2.5) is obtained by adding the land use offset correction term \( \alpha_{LU} \cdot \left(40.1 \cdot e^{-T_o/52} + 0.2 \cdot e^{-e_o/15}\right) \) to Equation (2.4).
The $\alpha_{LU}$ factors along with the RMSE and Mean Bias Errors (MBE) of the modeled versus ground-based LW$_d$ radiation for a varying number of clear days are summarized in Table 2.2. Modeled LW$_d$ radiation requires the emissivity calculation using Equation (2.5) and Equation (2.1) solved for LW$_d$. The RMSE provides information on the short-term model performance without indicating overestimation or underestimation, while the MBE evaluates the long-term model performance. The obtained values of the average overestimation (MBE) for the modeled LW$_d$ flux lie in the range of 0.9 to 6.4 W/m$^2$ (0.3 to 2.1% of the average LW$_d$), and the RMSE errors vary from 5.7 to 9.3 W/m$^2$ (1.8 to 3.0%) across the sites. These RMSE deviations are even smaller than the CG3 sensor error (accuracy of 3% achieved by SJRWMD) indicating the good accuracy of the proposed model.

Table 2.2: Regression and error analysis summary of LW$_d$ radiation.

<table>
<thead>
<tr>
<th></th>
<th>Deland</th>
<th>Jarboe Park</th>
<th>Lindsey Citrus</th>
<th>Orange Creek</th>
<th>Ocklawaha Prairie</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha_{LU}$, %</td>
<td>11.6</td>
<td>4.5</td>
<td>0</td>
<td>-3.7</td>
<td>-13.4</td>
</tr>
<tr>
<td>RMSE, W/m$^2$ (RMSE, %)</td>
<td>8.31 (2.5)</td>
<td>5.72 (1.8)</td>
<td>7.39 (2.4)</td>
<td>7.23 (2.4)</td>
<td>9.34 (3.0)</td>
</tr>
<tr>
<td>MBE, W/m$^2$ (MBE, %)</td>
<td>2.59 (0.8)</td>
<td>1.10 (0.4)</td>
<td>6.43 (2.1)</td>
<td>0.92 (0.3)</td>
<td>3.30 (1.0)</td>
</tr>
<tr>
<td>NSE</td>
<td>0.9</td>
<td>1.0</td>
<td>0.9</td>
<td>0.9</td>
<td>0.9</td>
</tr>
<tr>
<td>n</td>
<td>18</td>
<td>26</td>
<td>14</td>
<td>24</td>
<td>23</td>
</tr>
<tr>
<td>Land Use</td>
<td>Urban (medium density)</td>
<td>Urban (lower density)</td>
<td>Agriculture</td>
<td>Rangeland</td>
<td>Wetland</td>
</tr>
</tbody>
</table>
The Nash and Sutcliffe’s (1970) Efficiency (NSE) is chosen as a reliable estimator to assess the model performance in lieu of the correlation coefficient (R). The NSE is a normalized form of the Mean Square Error (MSE) with respect to the variance of observations ($s^2_{actual}$) and can be determined through the following relationship:

$$\text{NSE} = 1 - \frac{\sum (y_{actual} - y_{model})^2}{\sum (y_{actual} - y_{actual})^2} = 1 - \frac{MSE}{s^2_{actual}}$$

(2.6)

where $y_{actual}$ and $y_{model}$ are observed and predicted emissivities respectively. The range of NSE lies between $-\infty$ and 1 (perfect fit). A negative NSE indicates that the mean observed variable would have been a better predictor than the modeled values. If the model is unbiased, NSE is equivalent to $R^2$. The NSE scores, shown in Table 2.2, are 0.9 to 1.0 throughout the sites, suggesting that the new model estimates are a good fit to the observed climatology.

The dissimilar urban $\alpha_{LU}$ values of Table 2.2 as well as the low p-value of the t-test on the measured urban radiation means can be justified from the diversity of the urban mosaic in the two sites. The sensor in Jarboe Park is installed in a city park, whereas the Deland sensor is located next to a wastewater treatment facility and a collector road, which can generate the urban heat island effect.

According to Equation (2.5), for identical weather effects (i.e. constant $T_o$ and $e_o$) the clear sky effective emissivity increases with the presence of an urban environment and decreases over a wetland landscape leading to less LW$_d$ flux in the later case. Lines representing equal modeled emissivity along with overlaid data of actual emissivity for all land uses are shown in Figure A.1 and Figure A.2 of the Appendix A. The slopes of
emissivity get steeper with the change of the environment from urbanized to natural when one atmospheric effect ($T_o$ or $e_o$) remains constant. Therefore, the emissivity of the urban environment is more sensitive to temperature or water vapor variations than the one of the wetland. The above sensitivity will be further discussed in Section 2.5.3.

2.5 Model Evaluation and Discussion

2.5.1 Model verification

By substituting an appropriate $\alpha_{LU}$ value from Table 2.2, and the air temperature and water vapor pressure for a specific calibration land use site into Equation (2.5), the clear sky emissivity can be obtained. A preliminary graphical inspection (Figure 2.3) indicates that the modeled LW$_d$ radiation follows closely the corresponding CNR1 flux not only on daily but also on shorter time scale (semi-hourly) over all locations.

Overall, there is a good agreement between the actual clear sky daily emissivity values, calculated by using Equation (2.1), and the values obtained by using the new model at all sites, as can be verified from Table 2.3. In other words, the new model exhibits the least errors and the higher prediction scores, thus it performs relatively better than the others with the exception of Deland where Brutsaert’s equation has the best performance. For instance over Jarboe Park, the new model yields the best prediction for emissivity with a normalized MBE error of 0.2%, an RMSE error of 1.8% and a NSE statistic of 0.9. The next best predictions for emissivity in descending order are achieved by the models of Brutsaert (1975), Prata (1996), Satterlund (1979) and Idso (1981). Moreover, intercomparisons of LW$_d$ radiation (similar to Table 2.3) show that the difference between the compared models—in
particular the MBE difference between an existing model and the new formulation—falls outside the observational error of the CG3 sensor (±3% accuracy) with the exceptions of Deland and Jarboe Park. This confirms the superior performance of the proposed model in most of the sites.

Table 2.3: Comparison of model predictions with actual clear sky emissivity (Note: the MBE and RMSE are normalized errors in %).
Subsequently, an individual model performance analysis is performed based on model prediction errors with respect to actual emissivity values. In Figure 2.5, each model’s residuals, i.e. predicted minus actual emissivity, are presented for the data over Jarboe Park in the form of relative frequency histograms. The main observation from Figure 2.5 is that the new model has the least absolute errors which are distributed near the zero-error line. Under the assumption of normal error distribution, 68% of the individual errors should be within the range with bounds $MBE \pm RMSE$. These findings are in agreement with Table 2.3 results. Similar to Jarboe Park residual analyses were performed for the other land use sites, and show the overall better performance of the new model compared to others. It is noteworthy that Brutsaert’s and Prata’s models were found to be the best emissivity predictors for the tropical climatic region. This was based on the results of comparisons between several formulations and actual data obtained from an extensive radiation meteorology database—including polar and desert data (Prata, 1996).
Figure 2.5: Relative frequency of errors of clear sky emissivity as predicted by various models with respect to actual values at Jarboe Park; the zero-error line is also shown.

In agreement with the error analysis of Table 2.2, Figure 2.6 illustrates best-fit comparisons with the measured clear sky LW\textsubscript{d} radiation over the study sites. As previously
mentioned, the new model’s predictions for \( LW_d \) radiation are associated with MBE and RMSE errors of less than 2.1% and 3.0% respectively, and NSE statistics of greater than 0.9.

![Graph](image.png)

Figure 2.6: Comparison of the new model’s \( LW_d \) radiation and the actual \( LW_d \) radiation at the calibration sites.

Figure 2.7, Figure 2.8 (a) and Figure 2.8 (b) show daily \( LW_d \) radiation comparisons applicable to two urban, two rangeland, and two wetland land use sites, respectively. As discussed in Section 2.4, the two sites within each land use group exhibit similar ground-sensed \( LW_d \) radiation trends during the same clear days. The two-tailed t-test is also applied to the samples of modeled \( LW_d \) radiation for the two locations of each land use group. The \( p \)-values are 0.786, 0.544 and 0.057 for the wetland, rangeland and urban sites respectively, indicating that the modeled radiation means are equal within each land use (\( \alpha=0.05 \)).
Figure 2.7: Downward LW radiation at urban sites for Julian days 78-105 (a) and 106-135 (b).
Figure 2.8: Downward LW radiation at rangeland (a) and wetland sites (b).
2.5.2 Model validation

The mean daily relative humidity along with the mean daily air temperature, supplied by GAPP, yields the water vapor pressure according to the formula of Daniel (1962). The daily clear sky LW$_d$ radiation, calculated by the model of Equation (2.5) using $a_{LU}$ values of $+0.116$ (for Deland), 0 (for Bondville) and $-0.037$ (for Ft. Peck), is then compared to the daily actual LW$_d$ flux.

Table 2.4 summarizes the errors and the NSE statistic of the LW$_d$ radiation models tested versus actual data over the aforementioned sites. It is apparent that the proposed land use adapted model exhibits the least errors and the highest NSE scores for LW$_d$ radiation at all cases. Throughout the sites, the new model exhibits the best prediction efficiency when applied to the climatology of Bondville, IL (maximum NSE=0.7). It is also the only formulation among the existing models that predicts the LW$_d$ flux better than the actual climatology of Ft Peck, MT (NSE>0). Note that Satterlund’s (1979) and Idso’s (1981) model predictions are worse than the actual data at all cases.

<table>
<thead>
<tr>
<th>Model</th>
<th>Site, Landuse:</th>
<th>Actual LW $s_{av}$ 308 W/m$^2$</th>
<th>Site, Landuse:</th>
<th>Actual LW $s_{av}$ 313 W/m$^2$</th>
<th>Site, Landuse:</th>
<th>Actual LW $s_{av}$ 302 W/m$^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rangeland</td>
<td></td>
<td>Agriculture</td>
<td></td>
<td>Urban</td>
<td></td>
</tr>
<tr>
<td>Brutsaert (1975)</td>
<td>Ft. Peck, MT</td>
<td>8.5</td>
<td>5.1</td>
<td>6.4</td>
<td>6.4</td>
<td>9.7</td>
</tr>
<tr>
<td>Satterlund (1979)</td>
<td>July-Sept. 2001</td>
<td>12.8</td>
<td>8.6</td>
<td>9.7</td>
<td>9.7</td>
<td>10.6</td>
</tr>
<tr>
<td>Idso (1981)</td>
<td>Deland, FL</td>
<td>13.8</td>
<td>10.5</td>
<td>12.6</td>
<td>12.6</td>
<td>13.7</td>
</tr>
<tr>
<td>Prata (1996)</td>
<td>Bondville, IL</td>
<td>9.2</td>
<td>5.4</td>
<td>6.7</td>
<td>6.7</td>
<td>8.1</td>
</tr>
<tr>
<td>New model</td>
<td>July-Sept. 2001</td>
<td>4.4</td>
<td>1.2</td>
<td>5.2</td>
<td>5.2</td>
<td>5.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>6.7</td>
<td>3.7</td>
<td>6.5</td>
<td></td>
<td>0.6</td>
</tr>
</tbody>
</table>

Table 2.4: New model and existing model predictions of LW$_d$ radiation.
Figure 2.9 shows box plots of the LW_d radiation obtained from the new model and existing models along with pyrgeometer data. It is affirmed, by inspection of the errors, shown in Table 2.4, and the interquartile ranges, shown in Figure 2.9, that the new model overestimates the actual LW_d radiation by 1.2% at the agricultural site, 4.4% at the rangeland location and 5.2% at the urban setting. The RMSE errors are 3.7%, 6.7% and 6.5%, and the NSE scores are 0.7, 0.5 and 0.6 at the corresponding sites. Based on the MBE errors of Table 2.4, it is apparent that the difference between the various models falls outside the experimental noise of the CG3 sensor (±3% accuracy). This shows the best performance of the proposed model with the exception of Deland, where the new model performs relatively better than Brutsaert’s and Prata’s models.
Figure 2.9: $LW_d$ radiation whisker box plots of the proposed model and existing models along with actual radiation over Deland (a), Bondville (b) and Ft. Peck (c). The means are shown with dashed lines and the outliers with circles.
2.5.3 Land use effect on emissivity and local warming

The effective air emissivity, which is the efficiency of the air medium for emitting downwards the infrared solar radiation onto the land surface, follows a “growth” pattern with respect to the effects of air temperature and water vapor. Besides the effects of the air medium, the local scale land environments consisting of materials of different thermal and physical properties affect the surface roughness, exhibit inter-element emissions and reflections of LW radiation, and consequently have an aggregate response on the energy exchange and the air emissivity. The landscape feedback on the surface energy component of LW$_d$ radiation is incorporated in the new model of Equation (2.5) with the use of land cover coefficient $\alpha_{LU}$.

More specifically, the land covers with their surface roughness introduce a degree of mechanical and thermal turbulence (Oke, 1990). On one hand, the urban microclimate is influenced by the presence of buildings, which induce mechanical turbulence due to increased surface roughness, and the street canyons, which induce thermal turbulence due to radiation trapping and shadowing. On the contrary, the wetland environment has the highest inertial resistance to temperature or water vapor fluctuations. In fact, urban environments demonstrate surface temperatures significantly greater than the air temperatures during the daytime and significantly lower at night (Djak et al., 2000; Jonsson et al., 2006). Likewise the diurnal water vapor pattern over the urban atmosphere in several studies exhibits a moisture deficit at daytime and a moisture surplus during the night (Holmer and Eliasson, 1999).
Figure 2.10 shows the measured range and means of air temperature (a), water vapor pressure (b) and air emissivity (c) over the calibration and validation sites. The following discussion refers to the radiation meteorology used for the model development (left portion of Figure 2.10). Over the nearly three-month observed data ensemble, the standard errors ranged from 0.65 to 1.11 hPa for water vapor, 0.66 to 0.70 K for air temperature, and 0.007 to 0.009 for air emissivity. Within the local scale, the urban air contains moderate amounts of water vapor (mean of about 12 hPa). Besides, high temperatures are developed due to the urban heat island effect (mean of about 291 K). The total atmospheric effect is intensified and the effective clear sky emissivity is high. In this case, the atmosphere and the land use have synergetic effect and the land cover factor $a_{LU}$ is positive in the emissivity model. The outcome is higher LW$_d$ radiation and enhanced local warming.
Figure 2.10: Observed atmospheric variables of a) air temperature, b) vapor pressure and c) clear sky emissivity over the calibration (on the left) and validation (on the right) sites. Means and standard error bars as well as ranges are shown. The dashed ellipse depicts the maximum observed variable across the calibration land uses.
The scenario on the wetland landscape is the opposite. The cloudless atmosphere above the wetland exhibits large temperature (mean of 294 K) and water vapor pressure effects (mean of 15 hPa), which could potentially have resulted in high emissivity. However, the “apparent” downward LW flux is effectively reduced by transfer of heat from the warmer surrounding atmosphere to the evaporating surface (high water level) and by transfer from storage. In other words, the clear sky emissivity over the wetland “appears” smaller—as opposed to the large atmospheric effects. The damping of the atmospheric effects is achieved by a negative land use factor $a_{LU}$. This phenomenon promotes local cooling. As a conclusion from the previous discussion, the clear sky emissivity increases with the change of the landscape from wetland to agriculture and to various degrees of urbanized land.

Furthermore, even under the assumption of identical daily weather effects, the cloudless urban sky appears to emit more LW$_d$ radiation than the sky above a rural site. Relevant rationalization may be based on the refractive index (RI) of the land materials in different settings. The refractive indices of typical land materials are in decreasing order: 2.6 for concrete, about 1.5 for soil, 1.4 for leaf and 1.3 for water. An increased RI is associated with higher reflection of the LW$_d$ radiation by the elements of the ground surface. This reflected energy is absorbed by the radiatively active atmospheric constituents and a portion is emitted back to the ground resulting in increased effective air emissivity. Thus, the radiative properties of the agglomerated land surface materials contribute in enhancing (in the case of concrete) or damping (in the case of water) the effective atmospheric emissivity.

These urban-rural differences in radiation meteorology agree with earlier field studies conducted in latitudes ranging from 35 °N to 57 °N (Oke and Fuggle, 1972; Aida and Yaji, 1979; Kobayashi, 1982; Estournel et al., 1983; Lindgren, 1997). In these studies, an excess of
18-40 W/m² has been reported for the clear sky LW$_d$ flux over the usually warmer, more humid and more polluted urban air (Holmer and Eliasson, 1999). A maximum range of [3.1, 35.9 W/m²] represents the LW$_d$ radiation surplus measured over an urban in contrast to a rural site in the present study. More specifically, the significant upper limit (35.9 W/m² or about 11% of the average of the two daily radiation means) refers to the LW$_d$ radiation excess measured over Deland and Orange Creek during the Julian clear day 97. Besides, the lower limit refers to the surplus measured between Jarboe Park and Orange Creek during the clear day 83. The corresponding extremes of the modeled LW$_d$ flux are 35.1 W/m² and 9.7 W/m² for the aforementioned sites and clear days.

2.6 Summary

In this work, a new effective clear sky emissivity model is proposed. The model is land use adapted by incorporating an offset factor ($\alpha_{LU}$) and was developed based on LW$_d$ radiation and meteorological data applicable to the SJRWMD region, Florida for the spring season of 2004. Both air temperature ($T_o$) and water vapor pressure ($e_o$) effects are taken into consideration with specific weights, and the model appears to work well even at low values of $e_o$. It is shown that the developed model, which requires two readily available meteorological variables and one landuse-dependent coefficient as inputs, has good proximity to the actual LW$_d$ radiation.

Moreover, not only does this model predict the clear sky emissivity in SJRWMD region but is it also transferable to other regions of various land uses. The model is tested in three land uses including mid latitude sites, and it is shown that it is the most suitable
predictor of the downward LW radiation with normalized MBE and RMSE errors of less than 5% and 7%, respectively, and NSE statistic greater than 0.5.

The landscapes, which are examined in this study, exhibit different effectiveness with which they absorb LW\textsubscript{d} radiation. It is upon the discretion of the model user to select one of the developed $\alpha_{LU}$ coefficients based on the dominant land cover or to choose a $\alpha_{LU}$ factor between two tabular values, if the site is of composite land use and detailed land cover data are available. Future studies need also to address the model applicability in more diverse climatic regions than the present study. Further future work may include incorporating vegetative and impervious cover in the proposed model and utilizing it for year-around ET estimation over different land covers.
References


CHAPTER 3:
CONTROLS AND PARAMETERIZATIONS OF EVAPOTRANSPIRATION
AT A GRASSLAND IN FLORIDA

This chapter has been submitted for publication with the following citation: Rizou, M., Nnadi, F.N., Sumner, D.M., 2008. Agricultural and Forest Meteorology (under review, May 2008).

3.1 Introduction

In spite of the fact that grasslands account for about 40% of the global terrestrial land cover (White et al., 2000), their contribution to the surface exchanges of energy, water and carbon in local and regional scale is so far uncertain. Grasslands show the largest interannual variation in primary production of the major ecosystem types in the continental US, caused primarily by variation in precipitation patterns, and may be the most responsive to future climatic changes (Knapp and Smith, 2001). Rainfall supply is particularly a water-stress inducing factor for vegetations on soils of low water-holding capacity (San Jose, 2001).

Evapotranspiration (ET) or latent heat flux ($\lambda E$) is a major component of the hydrologic cycle, which is linked to changes in soil moisture storage and precipitation. Over the global land surface, the long-term ET represents about two-thirds of the annual rainfall input (Trenberth et al., 2007). In the hydrologic budget of Florida, ET is the second important component after precipitation (Jones et al., 1984). Information on the estimation of the evapotranspiration losses provides feedback between land surface and climate. For instance, Anderson and Chung (2006) reported that a 3°C increase in air temperature results in 3-6%
enhancement in the reference ET over the entire state of California. In general, accurate knowledge of ET is necessary in evaluating parameterization schemes used in hydrologic and climatic models, quantifying agricultural applications (such as crop yield and water use), assessing the environmental aspects of natural ecosystems and improving water management techniques.

Several approaches are available for determining field-scale ET. Combination methods require more data than the temperature and radiation methods, and hence they are more accurate under a variety of vegetative and meteorological conditions. The original Penman (1948) and Penman-Monteith (Monteith, 1965) methods require extensive input parameters. The Priestley-Taylor (Priestley and Taylor, 1972) approach has been reported to outperform the Penman-Monteith (PM) model in wetland and grass sites as it requires less meteorological data, is computationally more efficient, and simulates ET successfully (Stagnitti et al., 1989; Stannard, 1993; Sumner, 1996; Sumner and Jacobs, 2005). In fact, the Priestley-Taylor (PT) equation with PT coefficient of 1.26 (usually applied to wet surfaces) and PM model yield almost equal ET estimates when applied over surfaces with surface conductance greater than $17 - 20 \, mm \, s^{-1}$ (McNaughton and Spriggs, 1989). Furthermore, deBruin and Stricker (2000) found that the PT formula performs the best among five ET formulations, when compared to ET measurements over a Dutch grass site.

The PT approach depends on accurate prediction of the PT coefficient $\alpha$, which is a function of the Bowen ratio. This constant is used to describe the regional interaction between the surface and the boundary layer (Blanken et al., 1997). In the literature, $\alpha$ is correlated to the soil moisture (Davies and Allen, 1973; Flint and Childs, 1991), the water deficit or supply to vegetation expressed as accumulated actual evaporation minus
precipitation (Priestley and Taylor, 1972) or antecedent precipitation index (Mawdsley and Ali, 1985), and the green leaf coverage under potential ET conditions (Ritchie, 1972). The maximum documented values of $\alpha$ over grass range between 0.7 and 1.5 under all moisture conditions (Mukammal and Neumann, 1977; deBruin and Holtslag, 1982; Chen and Brutsaert, 1995; deBruin and Stricker, 2000; Wever et al., 2002; Baldocchi et al., 2004; and others).

The objectives of this study are four fold: (1) to provide year-long, daytime mean estimates of measured and potential ET rates from a non-irrigated grassland in Central Florida using Eddy Correlation (EC) and PT methods, respectively; (2) to quantify the temporal response of ET to diminishing wetness, namely soil water content (SWC) and current precipitation index (CPI), within the wet season; (3) to investigate the effects of canopy characteristics and environmental factors on the water vapor exchange over the grass site; and (4) to present simple parameterizations for the PT coefficient in terms of SWC and CPI, which show good agreement with the actual data in the study site.

3.2 Experimental Area and Database

3.2.1 Site description and climatology

The study was conducted at a nearly flat (elevation of 18 m above the mean sea level), non-irrigated site within the Disney Wilderness Preserve, Polk County, Florida (28.05 N, 81.40 W). EC instrumentation was operated by U.S. Geological Survey (USGS) within 100 m of a South Florida Water Management District (SFWMD) weather station as shown in Figure 3.1. The study period covered the year 2004.
The dominant plant of about 70% coverage at the study site is bahia grass (Paspalum notatum) that varies from a lush green during the summer to a drab brown during the winter. The bahia grass is ungrazed with grass height of up to 40 cm and an average root depth of 1 m. Other plants at the study site, occurring as distinct patches, include: dog fennel (Eupatorium capillifolium), broomsedge bluestem (Andropogon virginicus), saw palmetto (Serenoa repens), hemlock witchgrass (Dichanthelium portoricense), manyspike flatsedge (Cyperus polystachyos), and flat-topped goldenrod (Euthamia caroliniana). In addition, there are few scattered trees (pine and oak), and extensive forested areas occur within 100 m of the study site. However, the effects of these forests were filtered from this analysis.
The EC flux station was sited to measure ET from the non-forested part of the study area. The source area of an EC measurement is generally considered to extend an upwind distance of about 100 times the sensor height (Campbell and Norman, 1998). With this criterion and a sensor height of 3.4 m, the source area to the South and West of the station (arc of 120° to 360° clockwise) includes both the intended grass cover as well as forested areas (Figure 3.1). Flux data obtained when wind direction was from the arc that contains forested areas were culled from the measured data.

The soil at the study site has fine sands of the Archbold and Immokalee series (USDA, 1990) with soil properties shown in Table 3.1. The water table fluctuated from -0.02 to 1.26 m below the land surface during the study period, but it has generally shallow depth (less than 1 m).

<table>
<thead>
<tr>
<th>Property</th>
<th>Value/characteristic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dominant plant species</td>
<td>bahia grass</td>
</tr>
<tr>
<td>Soil texture</td>
<td>fine sand</td>
</tr>
<tr>
<td>Porosity (cm³ cm⁻³)</td>
<td>0.395*</td>
</tr>
<tr>
<td>Soil heat capacity (J cm⁻³ K⁻¹)</td>
<td>1.134</td>
</tr>
<tr>
<td>Wilting point (cm³ cm⁻³)</td>
<td>0.07*</td>
</tr>
<tr>
<td>Field capacity (cm³ cm⁻³)</td>
<td>0.17*</td>
</tr>
</tbody>
</table>

* Adopted from Dingman (2002)

The climate of the study location is subtropical and humid. During 2004, the regional mean air temperature was 22.2 °C with annual extremes of $T_{\text{max}} = 27.9 °C$ and $T_{\text{min}} = 16.5 °C$, and the mean annual precipitation was about 59 in (1496 mm). The aforementioned values were deduced from historical records of the Southeast Regional
Climate Center (SERCC) program (http://www.dnr.sc.gov/climate/sercc/) based on a weather station located in Kissimmee, Florida. The rain gauge adjacent to the EC flux tower captured about 64 in (1629 mm) of rainfall, of which about 80% fell from May through October (wet season) as shown in Figure 3.2. An unfrequented number of hurricanes affected Florida’s climate during the 2004 wet season. Four hurricanes, ranging from category 2 to category 4 storms, made landfall in the time period from middle of August to end of September.

![Cumulative precipitation and evapotranspiration during the course of 2004](image_url)

Figure 3.2: Cumulative precipitation and evapotranspiration during the course of 2004 (hurricane events are shown with triangles).

3.2.2 Experimental database

The EC flux measurements are shown in Table 3.2. The measured variables include air temperature, relative humidity, wind speed and direction, soil temperature, soil moisture, soil heat flux, net radiation and incoming solar radiation. EC methods (Dyer, 1961; Tanner and Greene, 1989) rely on high-frequency measurements of fluctuations in wind speed, vapor density and air temperature to infer fluxes of vapor and sensible heat. The 3D sonic
anemometer and krypton hygrometer were monitored at 8 Hz, and the latent and sensible heat fluxes were computed at 30-min resolution and then logged on a CR10X datalogger. Water table depth was measured (KPSI Series 500 pressure transducer, Pressure Systems Inc., Hampton, VA) in an adjacent well. Precipitation at the nearby SFWMD site was measured using a tipping-bucket rain gauge (model 6011-A, All Weather Inc., Sacramento, CA) and daily totals were recorded on a datalogger (model CR10X, Campbell Scientific Inc., Logan, UT).

<table>
<thead>
<tr>
<th>Type of measurement</th>
<th>Instrument (model and manufacturer)</th>
<th>Height above land surface (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Evapotranspiration</td>
<td>EC system including Model CSAT3 3-D sonic anemometer and Model KH20 krypton hygrometer (CSI, Logan, UT)</td>
<td>3.4</td>
</tr>
<tr>
<td>Air temperature and relative humidity</td>
<td>Model HMP45C temperature and relative humidity probe (CSI, Logan, UT)</td>
<td>1.2</td>
</tr>
<tr>
<td>Net radiation</td>
<td>Model Q-7.1 net radiometers [2] (REBS, Seattle, WA)</td>
<td>3.4</td>
</tr>
<tr>
<td>Incoming solar radiation</td>
<td>Model LI200X pyranometer (LICOR Inc., Lincoln, NE)</td>
<td>3</td>
</tr>
<tr>
<td>Soil heat flux</td>
<td>Model HFT-3 soil heat flux plates [2] (CSI, Logan, UT)</td>
<td>-0.08</td>
</tr>
<tr>
<td>Soil moisture</td>
<td>Model CS615 water content reflectometer [2] (CSI, Logan, UT)</td>
<td>0 to -0.08 and 0 to -0.30</td>
</tr>
<tr>
<td>Soil temperature</td>
<td>Model TCAV averaging soil thermocouple probes [2] (CSI, Logan, UT)</td>
<td>0 to -0.08</td>
</tr>
<tr>
<td>Wind speed/direction</td>
<td>Model 05305-5 Wind Monitor-AQ (R.M. Young, Traverse City, MI)</td>
<td>3.6</td>
</tr>
<tr>
<td>Data logging</td>
<td>Model CR10X dataloggers [2] with 12 V deep-cycle batteries and 20 W solar panels (CSI, Logan, UT)</td>
<td>0 to 1</td>
</tr>
</tbody>
</table>

Notes: Numbers inside brackets are instrument counts. Negative height is depth below land surface. CSI stands for Campbell Scientific Inc., and REBS stands for Radiation and Energy Balance Systems.
3.2.3 Data corrections and processing

3.2.3.a EC flux data

The net all-wave radiation ($R_n$) is partitioned into latent heat flux ($\lambda E$), sensible heat flux ($H$) and soil heat flux ($G$) while assuming that the energy storage within the canopy ($\Delta S$) is negligible. The available energy input to the canopy is estimated as the difference between the measured net radiation and the ground heat flux into the soil surface, and it is given by the energy budget equation:

$$R_n - G = \lambda E + H + \Delta S$$  \hspace{1cm} (3.1)

Soil heat flux at the surface was based on measurements from soil heat flux plates buried at a depth of 8 cm, coupled with estimates of the transient energy storage change in the upper 8 cm of soil. This latter term was estimated based on measurements from thermocouples buried at depths of 2 and 6 cm, and a soil heat capacity with estimated value of 1.134 (Table 3.1 and Table 3.2). The other two components of the energy budget, $\lambda E$ and $H$, were quantified according to the EC method (Monteith and Unsworth, 2008) as follows:

$$H = \rho_a C_p \overline{wT_a'}$$ \hspace{1cm} and \hspace{1cm} (3.2a)

$$\lambda E = \overline{\lambda \rho_a w' q'}$$ \hspace{1cm} (3.2b)

where $\rho_a$ is the moist air density ($kg m^{-3}$), $C_p$ is the specific heat of air ($J kg^{-1} C^{-1}$), $\lambda$ is the latent heat of vaporization of water ($J kg^{-1}$), $w$ is the vertical wind speed ($m s^{-1}$), $T_a$ is the air temperature ($^\circ C$), and $q$ is the specific humidity ($kg kg^{-1}$). Primes and bars on the variables denote fluctuations from means and means over the sampling 30-min interval, respectively.
Estimates of water vapor flux were corrected for temperature-induced fluctuations in air density (Webb et al., 1980) and for the sensitivity of the hygrometer to oxygen (Tanner and Greene, 1989). Sensible heat flux was corrected for the error introduced by measuring fluctuations in “sonic” rather than actual air temperature (Schotanus et al., 1983). Any misalignment of the sonic anemometer with the air stream was corrected by the coordinate rotation procedure (Baldocchi et al., 1988). Measurements for which the misalignment was greater than 10° were culled from the measured data. Rainfall, dew and scaling caused by exposure to the atmosphere may obscure the hygrometer windows, particularly in humid environments, and result in missing data. The problems associated with dew formation are particularly acute at night in humid environments. Generally, nighttime ET is negligible because of the low energy availability in the absence of solar radiation. Exceptions can occur in open water or wetlands in which nighttime energy for ET can be derived from cooling of standing water. Nighttime ET was not considered for the analyses of this study.

Measured fluxes were further corrected for consistency with the canopy energy budget (Twine et al., 2000; Wilson et al., 2002; Baldocchi et al., 2004), while preserving the measured Bowen ratio \( B = H/\lambda E \). The regression on the 30-min fluxes shown by Equation (3.3) produced an intercept of 8.81 \( W m^{-2} \), a slope of 0.73 and a coefficient of determination \( R^2 \) of 0.97 (Figure 3.3). Though imperfect, this energy balance ranks among sites with average degree of energy closure according to an analysis of FLUXNET sites reported by Wilson et al. (2002). The ratio of the turbulent fluxes to the available energy was 0.79 in annual totals.

\[
\lambda E + H + G = aR_{\phi} + b
\]  

(3.3)
Figure 3.3: Test of energy balance closure. 30-min values of net radiation ($R_n$) are plotted against the sum of latent ($\lambda E$), sensible (H) and soil heat (G) fluxes.

Missing or culled ET data (under conditions of inadequate fetch, excessive misalignment of sonic anemometer, obscured hygrometer windows or excessive energy budget correction) were gap-filled with a regression technique. Similarly to previous studies (Sumner, 2001; Sumner and Jacobs, 2005), multiple linear regression was used to determine the dependence of $\alpha$ on environmental variables such as: air temperature, wind speed, vapor-pressure deficit, incoming solar radiation, depth to water table, and soil moisture at top 30 cm of the soil. The most significant explanatory variables for the PT function at this site were incoming solar radiation and depth to water table. A seasonal bias in the residual error
of the simulated ET was reduced through incorporation of a sinusoidal function of the day of year (DOY). The final formulation of the PT function \( \alpha \) used for gap-filling is:

\[
\alpha = (a \cdot SW_{in}^2 + b \cdot SW_{in} + c \cdot WT + d) \cdot g(\text{DOY}),
\]

\[
g(\text{DOY}) = e \cdot \sin\left(\omega(\text{DOY} - f)\right) + 1
\]

where \( SW_{in} \) is incoming solar radiation \( (W m^{-2}) \), \( WT \) is depth from land surface to water table \( (m) \), and \( \omega \) is the annual period \( (\omega = 2\pi/365 \text{ day}^{-1}) \). The best-fit parameters are:

\[
a = 5.52 \cdot 10^{-7}, \quad b = -9.32 \cdot 10^{-4}, \quad c = -0.259, \quad d = 1.23, \quad e = 0.24 \quad \text{and} \quad f = 126.
\]

The performance of the Equation (3.4) utilized in the PT model (Section 3.3.1) was associated with a \( R^2 \) of 0.94 and a Standard Error (SE) of 0.98 mm day\(^{-1}\).

About 67\% of the 30-min daytime ET values were missing and required gap-filling. The data analyses were conducted on daytime measurements, since daytime ET is generally much higher than nighttime ET. Daytime 30-min data correspond to times when the net radiation is greater than 5 \( W m^{-2} \), and these data were averaged over a daily basis. More specifically, the nighttime fluxes were deemed missing and not equal to zero. Data tests involving ET parameterizations, ET classification into water availability stages and other regression analyses used “good” days only. “Good” day is considered any day with more than 65\% of 30-min intervals of measured daytime ET values, with the missing 30-min gaps filled as described above. A record of 101 “good” days of the year 2004 was selected for deriving factor relationships, whereas the yearly time series of gap-filled ET data were used for other analyses such as temporal variations and summations.
3.2.3.b LAI/FPAR data

Leaf Area Index (LAI) measures the one-sided density of the grass leaves on the underlying ground surface. For this study, LAI and Fraction of Photosynthetically Active Radiation (FPAR) data were retrieved from MODerate resolution Imaging Spectroradiometer (MODIS 15) land cover product obtained through the USGS EROS Data Center (EDC) Distributed Active Archive Center (DAAC) (http://lpdaac.usgs.gov, accessed July 2007). The satellite images are provided at 1-km spatial resolution and 8-day temporal resolution. The land products were re-projected and converted to appropriate format with the aid of MODIS Reprojection Tool (MRT) software, and subsequently were processed in Environment for Visualizing Images (ENVI 4.3) software.

The product files contain not only LAI and FPAR fields but also quality control (QC) flags providing information about the overall quality of the product such as algorithm path, cloud state and aerosols. With the application of a QC mask, only the pixels with good QC flags remained in the dataset. Next, the valid LAI field digital numbers, which range from 0 to 100, were converted into physical LAI units by multiplying by the factor of 0.1. The point of interest was a 1x1 km grid cell with homogeneous vegetation cover in the vicinity of the EC flux tower. A complete description of the MODIS algorithms and data derivations can be found in Knyazikhin et al. (1999).
3.3 Theoretical Background

3.3.1 Priestley-Taylor model and coefficient

This work employs the PT method for ET estimation. The PT evapotranspiration is an extension of the equilibrium evaporation of Slatyer and McIlroy (1961) obtained for a homogeneous, well-watered surface in dynamic equilibrium with a saturated atmosphere under minimal regional advection (Priestley and Taylor, 1972; Brutsaert, 1991). It has been shown that the daily $\alpha$ is ideally a constant of the order of 1.26, and in that case, Equation (3.5a) assigns the aerodynamic term of the Penman equation to a constant proportion of the dominant radiation term (Stewart, 1983). In fact, the net solar radiation or available energy has dominant importance in ET calculations (McNaughton and Spriggs, 1986). For instance, the available energy accounted for 86% of the evaporation of unstressed grass in a Dutch site reported by De Bruin and Stricker (2000), and a maximum of 62% of the latent heat flux over serpentine grass in California (Valentini et al., 1995). Moreover, $R_n$ satisfied 60 to 96% of the annual evaporative demand in a total of 30 forest studies as reported by Stagnitti et al. (1989). The PT model and coefficient are given by:

$$\lambda E = \alpha \frac{\Delta (R_n - G)}{\Delta + \gamma} \quad (3.5a)$$

and

$$\alpha = \frac{\lambda E}{\lambda E_{eq}} = \frac{\Delta + \gamma}{\Delta (B + 1)} \quad (3.5b)$$

where $\lambda E$ is the latent heat flux ($W m^{-2}$), $\lambda$ is the latent heat of vaporization of water ($J kg^{-1}$), $R_n$ is the net radiation ($W m^{-2}$), $G$ is the soil heat flux ($W m^{-2}$), $\alpha$ is the
PT coefficient, $\Delta$ is the slope of the saturation vapor pressure-temperature curve of air $(KPa^oC^{-1})$, $\gamma = C_p \cdot p / 0.622 \cdot \lambda \approx 0.067$ is the psychrometric constant $(KPa^oC^{-1})$, $C_p$ is the specific heat of air at constant pressure $(Jkg^{-1}oC^{-1})$, $p$ is the atmospheric pressure $(KPa)$, $\lambda E_{eq} = [\Delta/(\Delta + \gamma)] \cdot (R_n - G)$ is the equilibrium latent heat flux $(Wm^{-2})$, and $B = H/\lambda E$ is the Bowen ratio.

The PT coefficient represents the normalization of the actual to the equilibrium evapotranspiration as shown in Equation (3.5b). The daily average $\alpha$ under the absence of local advection falls in the range $1 < \alpha < \Delta/(\Delta + \gamma)$. As mentioned before, several field studies found that $\alpha$ approaches the value of 1.26 in humid climates under minimum advection and no edge effects (Priestley and Taylor, 1972; Rouse and Stewart, 1972; Dingman, 2002). It is suggested that under the above conditions, ET defined by the PT model represents accurately the evaporation under “equilibrium” wet surface conditions (Davies and Allen, 1973; Eichinger et al., 1996). However in reality, there is almost always advection and deviations from a wet surface, and equilibrium ($\alpha = 1$) rarely occurs. Abundant literature estimated $\alpha$ for various surface conditions. Baldocchi and Meyers (1998) reported that relatively large $\alpha$ values ($\alpha > 0.9$) are associated with healthy crops or temperate and tropical canopies, whereas low $\alpha$ ($\alpha < 1$) with canopies of low LAI or low photosynthetic capacity. Alternatively, $\alpha > 1$ represents advective enhancement of ET, whereas $\alpha < 1$ represents advective suppression or strong surface control through the stomatal resistance of the leaves (McNaughton et al., 1979). Details about the $\alpha$ values of the current study and comparisons with literature findings will be discussed in later sections.
3.3.2 Canopy resistance

The bulk canopy surface conductance \((g_c)\) was calculated in order to assess the physiological control on ET losses by the procedure outlined below. First, the friction velocity \((u_*)\), which represents a measure of the turbulent velocity fluctuations in the air, is related to the mean wind speed \((u_z)\) measured at a height \(z\) through the following logarithmic law (Dingman, 2002):

\[
u_* = \frac{k u_z}{\ln \left( \frac{z - d_{\text{max}}}{z_o} \right)} \quad (3.6)
\]

Where \(z = 3.6\ m\), \(z_{\text{veg}} = 0.4\ m\) is the vegetation height, \(d_{\text{max}} = 0.67 \cdot z_{\text{veg}} = 0.27\ m\) is the zero-plane displacement height, \(z_o = 0.1 \cdot z_{\text{veg}} = 0.04\ m\) is the roughness height for momentum, and \(k = 0.40\) is the von Karman constant.

Subsequently, the total aerodynamic resistance \(r_a\) \((s\ m^{-1})\) and the bulk canopy surface resistance \(r_c\) \((s\ m^{-1})\) are calculated using Equations (3.7) and (3.8), respectively (Monteith and Unsworth, 2008). Equation (3.8) is an inverted form of the PM equation.

\[
r_a = \frac{1}{g_a} = \frac{u_z}{u_*^2} + 6.2 u_*^{-0.67} \quad (3.7)
\]

\[
r_c = \frac{1}{g_c} = r_a \left[ \frac{\rho_a C_p (e_s - e)}{\gamma^\lambda E} + \frac{\Delta}{\gamma} - 1 \right] \quad (3.8)
\]
where $g_a$ is the aerodynamic conductance ($m{s^{-1}}$) between the effective canopy surface and the reference height at which micrometeorological measurements are made, $g_c$ is the bulk canopy surface conductance ($m{s^{-1}}$), $\rho_a$ is the moist air density ($kg\ m^{-3}$), $C_p$ is the specific heat of air under constant pressure ($J\ kg^{-1}\ C^{-1}$), $e_s$ is the saturation vapor pressure ($KPa$), $e = e_s \cdot RH$ is the actual vapor pressure ($KPa$), $RH$ is the relative humidity, $\Delta$ is the slope of the saturation vapor pressure-temperature curve of air ($KPa\ oC^{-1}$), $\gamma$ is the psychrometric constant ($KPa\ oC^{-1}$), $B$ is the Bowen ratio, and $6.2\ u_o^{-0.67}$ is the excess aerodynamic (leaf boundary layer) resistance.

For the calculations of $g_a$ and $g_c$, the 30-min daytime data when $g_c < 15\ mm\ s^{-1}$ were used in this study, in order to exclude abnormal atmospheric conditions such as low solar radiation. Similar screening criteria were also used in the literature (Goodrich et al., 2000; Sumner and Jacobs, 2005; Li et al., 2007). About 9% of the 30-min conductance data during “good” days were filtered out prior to modeling.

3.3.3 Antecedent precipitation

The CPI concept was introduced by Smakhtin and Masse (2000) following the original idea of the antecedent precipitation index (API) by Kohler and Linsley (1951). The CPI is a continuous function of precipitation that abruptly increases during rainy days and exponentially decays during dry periods. The CPI for any day $t$ is defined as follows:

$$CPI_t = K CPI_{t-1} + P_t$$ (3.9)
where $CPI_t$ is the current precipitation index for the day $t$ (mm), $P_t$ is the precipitation for the day $t$ (mm), and $K$ is the recession coefficient. This constant usually varies from 0.85 to 0.90 over most of the Central and Eastern US (Kohler and Linsley, 1951).

The generation of the annual time series of daily CPI values requires assumptions for the initial CPI value and the recession coefficient. The recession coefficient was assumed to be 0.9. The initial value of CPI, however, does not influence the resultant CPI time series, because several time series calculated with different initial values will converge within several weeks (Smakhtin and Masse, 2000). The CPI methodology is promising in hydrologic simulations, such as storm runoff, that require knowledge of both the amount of precipitation and its distribution over time. Since the CPI accounts for the current precipitation history of a catchment, it will also be utilized for modifying the PT parameter of this grassland ecosystem.

### 3.4 Results and Discussion

#### 3.4.1 Rainfall, soil water and ET

The temporal variations of daily soil water, i.e. SWC, WT and daily change of WT, are presented in Figure 3.4, and the daily rainfall pulses and CPI along with the daytime average gap-filled ET are illustrated in Figure 3.5. During the wet season, the volumetric SWC at the top 30 cm of the soil exceeded its field capacity (which is 0.17) attaining a maximum value of 0.52 on DOY 250 (6 September). The minimum SWC value of 0.07 (wilting point) was reached on days 153-155 (1-3 June) at the end of the dry period 1 shown in Figure 3.5. The
variation of the WT had similar pattern with the SWC. The WT varied from a minimum level of 1.26 m below the ground on DOY 156 (4 June) to 0.02 m above the ground (minor flooding) on DOY 249 (5 September). The WT dropped below the root zone of the bahia grass (which is 1 m) for about 51 days mainly over the dry May. The falling and rising of WT is associated with negative and positive daily change in depth to WT, respectively. An interesting remark refers to the high values of ET and daily change of WT, shown by the upward arrows in Figure 3.5 and Figure 3.4, respectively. The average daytime ET peaked either the day of or the day following a peak in daily change of WT. In addition, peaks in CPI caused by rainfall events resulted in peaks of soil water and ET, indicating the possible coupling between the above water budget components. Other field studies also observed the coupling between precipitation (and soil water) and evapotranspiration (Scott et al., 2006; Hao et al., 2007).
Figure 3.4: Daily averages of volumetric SWC for the top 30 cm of soil and WT level, along with the daily change of WT during the course of 2004. The lines indicating surface, root depth, and 17% SWC are shown. The wet season is delineated with the dotted vertical lines.
Figure 3.5: Daily total precipitation and CPI along with daytime average ET. The CPI value of 51 mm and three numbered interstorm events are depicted. The wet season is delineated with the dotted vertical lines. The nighttime ET was deemed missing and not equal to zero.
It is apparent from Figure 3.5 that ET exhibited large average daytime values (greater than 6 $\text{mm day}^{-1}$) during the period of DOY 158-283 within the wet season. The total rainfall of the wet season was 1273 mm. Three major interstorm periods were delineated to examine the effect of drying on ET (Figure 3.5). These events were defined as periods after large rain (greater than 8 mm) with small rainfall accumulation (less than 2 mm). During the drydown, CPI, SWC and ET decreased exponentially. More details on the water vapor dynamics during dry periods will be discussed in the following section.

The critical wetness values, illustrated in Figure 3.4 and Figure 3.5, were determined to classify ET into two water availability stages (Section 3.4.4). Water limited conditions, when SWC was less than 0.17 or CPI was less than 51 mm, were observed for more than 65% of the days in 2004. This provides a preliminary indication that water supply was a limiting factor on ET of this grass ecosystem during most of the year.

3.4.2 Dynamics of ET and wetness variables during dry intervals

Based on evidence that there is no significant surface or subsurface runoff in the study site, ET is a component of the water balance coupled with changes in soil moisture storage and rainfall. Previous literature studies estimated the dry-out dynamics of ET computed by meteorological methods (McAneney and Judd, 1983; Hunt et al., 2002; Kurc and Small, 2004) or simulations models (Scott et al., 1997; Lohmann and Wood, 2003). Similarly in this study, a first order system equation was used in order to determine the temporal responses of CPI and SWC in relation to the ET response during dry-down periods. This equation is given by the following expression:
where \( y(t) \) is the variable under consideration, i.e. CPI, SWC or ET, at \( t \) days after the rainfall event, \( y_i \) and \( y_f \) are the initial and final values of the variable observed at the first and end day of the drydown period, respectively, and \( \tau \) is the time constant to be determined from best-fit regression. The coefficient \( \tau \) indicates the time required for \( y(t) \) to decline to 1/e or 37% of its range \((y_i - y_f)\).

The minimum antecedent rain amount of 8 mm, sufficient to stimulate a response by the moisture variables and ET, was applied in selecting dry periods with insignificant rainfall accumulation (less than 2 mm). The cutoff value of 8 mm was chosen as the least rain to cause a substantial ET decay with time \((R^2 > 0.55)\). The longest period without significant rain lasted 53 days (2 October-23 November) following a storm of 12.7 mm rain. The duration of the above event—interstorm event 3 in Figure 3.5—was truncated to 30 days to coincide with the end of the wet season and be no longer than dry intervals observed in 2004. In addition, two major dry events occurred after storms of 24.1 mm of rain on 2-3 May (event 1) and 35.6 mm of rain on 8-9 September (event 2); two shorter dry events were also included in the time series.

The time constants for the function of daytime ET during “good” days, \( y = ET (mm/day) \), versus the corresponding amounts of the antecedent rain for the five aforementioned dry events are illustrated in Figure 3.6. Regression analysis indicates that \( \tau \) is a negative linear function of the amount of rain \((R^2 = 0.97)\). In other words, ET declination rates are more rapid following larger rainfall in the wet season. This is explained
by the fact that more rainfall provides additional soil moisture and depression storage (mainly captured on the grass leaves), and subsequently causes enhanced photosynthetic and water uptake rates by the plants, as well as, greater bare soil evaporation. Therefore, it is suggested that moisture is a main limiting factor for ET dynamics following rain pulse events.

Figure 3.6: Correlation between the time constant (\(\tau\)) of the ET drydown model and the amount of antecedent rain. The regression line is also shown.

Figures 3.7(a), 3.7(b) and 3.7(c) show the exponential decrease in CPI, SWC and daytime ET, respectively, during a dry period of 1–30 days. These data are composites of the five dry events. It is apparent from these graphs, that ET following a rain of 8 mm reduces to a minimum average value of about 5.5 mm day\(^{-1}\) in 11 days when the SWC is 0.11. The time constants are 9.5 days for CPI and 6.1 days for both SWC and ET during the wet season. The exponential curves, given by Equation (3.10), fit the moisture data \(R^2 = 0.98 - 0.99\) better than they fit the actual ET data \(R^2 = 0.71\).
Some interesting remarks relevant to the water cycle dynamics of the bahia grassland during the wet season can be mentioned here. The time constants for the reduction in both SWC and ET following rain events are about 6 days, shorter than the timescale over which CPI declines. The equality of the ET and SWC timescales suggests that depletion of the soil moisture is mostly responsible for the temporal fluctuations in ET, whereas the dissimilarity to the CPI timescale may suggest that CPI has moderate control on ET dynamics. In addition, the duration of the composite dry event is longer than the timescales on which the CPI and SWC drop to relatively low values and ET decreases. Possible results of prolonged soil dryness are the diminishing of latent heat flux and water vapor in the atmosphere and the reduction of downwelling longwave radiation (Rizou and Nnadi, 2007).

When comparing these findings to previous studies conducted in New Zealand, the time constant of the ET model in this study is half of that ($\tau \approx 12$ days) obtained in an irrigated rye grass/white clover during summer (McAneney and Judd, 1983), and is equal to
that ($\tau = 6.1$ days) obtained in a summer tussock grass (Hunt et al., 2002). It is also three times larger than the time constant ($\tau = 2.1$ days) of a summer semiarid grassland, dominated by Black Grama, in New Mexico (Kurc and Small, 2004). The daytime ET may thus be more dynamic at the present study site than the ET over the managed pasture in New Zealand, and less dynamic than that of the semiarid grass in New Mexico.

3.4.3 Canopy effects on ET

LAI development of the bahia grass exhibited three phases that were in synchrony with the precipitation pattern (Figure 3.8(a)) and the daytime ET fluctuations (Figure 3.8(b)). LAI maintained an average value of about 1.4 $m^2 \text{m}^{-2}$ up to early July (approximately DOY 189), and increased to an average of about 2.1 $m^2 \text{m}^{-2}$ until the second half of September (DOY 265). Next, LAI increased precipitously to values greater than 3.0 $m^2 \text{m}^{-2}$ (maximum of 3.6 $m^2 \text{m}^{-2}$ on DOY 266) and then dropped towards the end of the year. This variation in LAI was associated with differences in moisture availability and therefore changes in ET. The extra precipitation received in early June led to high CPI and soil moisture that extended for the following three months during which time LAI increased. Later, LAI reached its peak due to the rainfall of a hurricane event on the already saturated soil. Though, the drop in LAI was out of phase with the decrease in CPI and ET after the end of the rainy season. This might be due to root water extraction from deep water reserves, which preserved the greenness of the grass while the evaporative water losses were reduced.
Figure 3.8: LAI variation with respect to CPI (a) and daytime average ET (b) during 2004.

Figure 3.9 and Figure 3.10 demonstrate the dependence of ET, expressed as PT coefficient $\alpha$, on canopy physiology, as represented by LAI and $g_c$, respectively. The 8-day
composite LAI and the daily means of the daytime $g_c$ values during “good” days of 2004 were used. It was found that $\alpha$ increased as a linear function of LAI (Figure 3.9) and a logarithmic function of $g_c$ (Figure 3.10). The PT coefficient was about 1 for green LAI greater than 1.8. The maximum cut-off LAI value of 2.7 was used here, since surface conductance and ET are often more strongly controlled by environmental conditions (such as water supply and vapor pressure deficit) after a maximum LAI is reached. This confounding effect of LAI has also been reported in the literature (Ritchie, 1972; Verma et al., 1992; Wever et al., 2002; Suyker and Verma, 2008). This might also explain the decrease of the slope of $\alpha$ versus $g_c$ curve in Figure 3.10. For high canopy surface conductance, when $g_c$ was averaged approximately 8 $mm \ s^{-1}$, measured ET was close to equilibrium rates. On the other hand, as $g_c$ decreased $\lambda E/\lambda E_{eq}$ declined rapidly and approached 0.5, when $g_c$ was about 2 $mm \ s^{-1}$. 
Figure 3.9: Daytime means of Priestley-Taylor coefficient $\alpha$ as a function of LAI during “good” days. The best-fit line is also shown.

$$\alpha = 0.77 \cdot \text{LAI} - 0.50$$

$$R^2 = 0.78$$

Figure 3.10: The relationship between daytime averages of Priestley-Taylor coefficient $\alpha$ and daytime means of canopy surface conductance during “good” days. The best-fit line is also shown.

$$\alpha = 0.35 \cdot \ln(g_c) + 0.24$$

$$R^2 = 0.82$$
The high positive correlation between $\alpha$ and canopy development is justified as follows. The greenness of the leaves increases when the soil moisture is ample, causing an enhancement of $g_c$ and ET, and thus resulting in higher $\alpha$ values. This finding is consistent with other grassland studies, such as the simulation study of McNaughton and Spriggs (1989), and the observations of Kim and Verma (1990), Verma et al. (1992), Kelliher et al. (1995), Valentini et al. (1995), Burba et al. (1999), and Wever et al. (2002). In addition, $g_c$ of the present grassland was in the range 2-11 $\text{mm s}^{-1}$, which is in agreement with values observed in other grasslands. More specifically, Kelliher et al. (1995) reported maximum $g_c$ values in the range of 4-12 $\text{mm s}^{-1}$ for various temperate grasslands and Kim and Verma (1990) and Zenker (2003) (cited in Schwarzel et al., 2006) estimated maximum $g_c$ to be in the range 1.3-15 $\text{mm s}^{-1}$ and 11-13 $\text{mm s}^{-1}$, respectively. In addition, Verma et al. (1992) and Burba et al. (1999) found that wetland grass evapotranspired at potential rates when $g_c$ was greater than 10 $\text{mm s}^{-1}$.

3.4.4 Environmental effects on ET

The water and energy exchange between the grassland and the atmosphere is controlled by various meteorological and soil factors. Table 3.3 shows the Pearson correlation coefficients between the daytime latent heat flux and the daytime environmental conditions observed in the study site, namely $w$, $T_a$, vapor pressure deficit ($VPD = e_s - e$), $RH$, $R_n$, $SW_m$, WT, SWC, and CPI (CPI has by definition daily time scale). The Pearson coefficients show that
factors with positive effect on $\lambda E$ in decreasing order of significance are: $T_a$, CPI, SWC (or negative WT), $R_n$, VPD and $SW_{sw}$. Among the above effects, $T_a$, CPI and SWC are strongly correlated to $\lambda E$. This suggests that elevated latent heat flux is the result of the influence of high soil moisture on surface (or overlying air) temperature. The impact of SWC on ET should be intensified under convective rainfall conditions (enhanced CPI). As reported by Findell and Eltahir (1997), extreme soil moisture availability (or lack) acts as, either a feedback mechanism maintaining the wet (or dry) conditions or as a flag indicative of some process affecting the soil moisture and precipitation regime. This significant impact of SWC ($r=0.66$) and CPI ($r=0.69$) on daytime $\lambda E$, during “good” days, justifies the use of the above moisture variables for ET stage classification (following in the text), and ET parameterizations (next section). A supplementary remark refers to the strong positive correlation observed between several variables, namely $R_n$ and VPD, $SW_{sw}$ and VPD, $SW_{sw}$ and $R_n$, CPI and SWC, CPI and WT (negative correlation).
Table 3.3: Pearson correlation coefficients (r) between daytime $\lambda E$ and its daytime environmental controls during “good” days at level of significance 0.05. Strong correlations are shown with underlined numbers.

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<th>$T_a$ ($^\circ C$)</th>
<th>VPD (KPa)</th>
<th>RH (%)</th>
<th>$R_n$ ($Wm^{-2}$)</th>
<th>$SW_{in}$ ($Wm^{-2}$)</th>
<th>WT ($m$)</th>
<th>SWC ($cm^3 cm^{-3}$)</th>
<th>CPI</th>
<th>$\lambda E$ ($Wm^{-2}$)</th>
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<td>-</td>
<td>-</td>
<td>-0.85</td>
<td>0.81</td>
<td>1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\lambda E$</td>
<td>-</td>
<td>0.82</td>
<td>0.52</td>
<td>0.63</td>
<td>0.50</td>
<td>-0.64</td>
<td>0.66</td>
<td>0.69</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

In Figure 3.11(a) and Figure 3.11(b) the relationship between daytime latent heat flux (normalized by its equilibrium evaporation rate) and moisture variables, volumetric SWC and CPI, respectively, is presented for the grassland site. The normalized ET ratios were grouped into bins (1% 30-min SWC bins and 5 mm daily CPI bins) and then averaged. As a result of the bin averaging, the PT coefficient $\alpha$ of the bahia grass was almost constant and attained a value below 1.26 when soil moisture was ample ($SWC > 0.17 \ cm^3 cm^{-3}$). The SWC limit (0.17), which is also the field capacity of the soil, is about 33% of the SWC at saturation (0.52). When SWC dropped below that critical threshold, $\lambda E/\lambda E_{eq}$ dropped abruptly until
the point when SWC reached 0.07 cm$^3$ cm$^{-3}$. The slope of this decrease (approximately 3.8) was close to the slope (about 3.2) found by Jacobs et al. (2002) for evapotranspiration of a wetland in Florida. In addition, Hunt et al. (2002) documented that evaporative fraction declined by a factor of 4, as soil moisture dropped from 0.12 to 0.04 cm$^3$ cm$^{-3}$, in a tussock grass drought study. Figure 3.11(b) demonstrates similar variation of $\alpha$ for the various CPI values observed at the study site. The data show less scatter than that of Figure 3.11(a). The PT function $\alpha$ reached a value close to unity, when the rainfall input yielded CPI values larger than the critical value of 51 mm.
Figure 3.11: Relationship of daytime PT coefficient with volumetric SWC (a) and CPI (b) during “good” days. The critical moisture limits are also shown.
The daily means of daytime $\alpha$ (without bin averaging) showed a wide range 0.38-1.12 during the “good” days of 2004, but most of the time they were less than unity, indicating that the hydrologic conditions (evapotranspiration) of this grass ecosystem were characterized by the limitations in water supply on annual time basis. The fluctuations of daytime $\alpha$, from the low limit observed on DOY 37 to the upper limit occurred on DOY 350, followed closely the drying and wetting of the soil. The monthly average $\alpha$ was lower than unity (Table 3.4), with the exception of July ($\alpha=1.02$), August ($\alpha=1.13$) and September ($\alpha=1.08$). The variations of the energy flux and climatic components on diurnal and seasonal time scales are discussed in details by Rizou et al. (2008).

Table 3.4: Monthly means of daytime energy fluxes, SWC and PT coefficient $\alpha$. The monthly totals of precipitation are also included.

<table>
<thead>
<tr>
<th>Month</th>
<th>Prec. (mm)</th>
<th>$R_n$ (Wm$^{-2}$)</th>
<th>$\lambda E^*$ (Wm$^{-2}$)</th>
<th>H (Wm$^{-2}$)</th>
<th>G (Wm$^{-2}$)</th>
<th>SWC (cm$^3$cm$^{-3}$)</th>
<th>$\alpha^*$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>88.1</td>
<td>236.1</td>
<td>80.9</td>
<td>148.0</td>
<td>7.2</td>
<td>0.112</td>
<td>0.59</td>
</tr>
<tr>
<td>Feb</td>
<td>69.1</td>
<td>235.4</td>
<td>84.0</td>
<td>138.7</td>
<td>12.7</td>
<td>0.129</td>
<td>0.64</td>
</tr>
<tr>
<td>Mar</td>
<td>41.9</td>
<td>321.5</td>
<td>127.1</td>
<td>179.1</td>
<td>15.3</td>
<td>0.114</td>
<td>0.65</td>
</tr>
<tr>
<td>Apr</td>
<td>57.4</td>
<td>359.4</td>
<td>145.5</td>
<td>201.4</td>
<td>12.5</td>
<td>0.084</td>
<td>0.64</td>
</tr>
<tr>
<td>May</td>
<td>26.9</td>
<td>363.4</td>
<td>158.9</td>
<td>190.2</td>
<td>14.3</td>
<td>0.075</td>
<td>0.67</td>
</tr>
<tr>
<td>Jun</td>
<td>276.6</td>
<td>372.3</td>
<td>216.6</td>
<td>136.5</td>
<td>19.3</td>
<td>0.108</td>
<td>0.85</td>
</tr>
<tr>
<td>Jul</td>
<td>218.9</td>
<td>340.9</td>
<td>237.4</td>
<td>87.2</td>
<td>16.2</td>
<td>0.198</td>
<td>1.02</td>
</tr>
<tr>
<td>Aug</td>
<td>371.6</td>
<td>322.8</td>
<td>249.2</td>
<td>55.3</td>
<td>18.3</td>
<td>0.427</td>
<td>1.13</td>
</tr>
<tr>
<td>Sep</td>
<td>365.5</td>
<td>279.2</td>
<td>201.3</td>
<td>63.1</td>
<td>14.9</td>
<td>0.370</td>
<td>1.08</td>
</tr>
<tr>
<td>Oct</td>
<td>13.0</td>
<td>285.3</td>
<td>178.4</td>
<td>97.9</td>
<td>9.0</td>
<td>0.159</td>
<td>0.91</td>
</tr>
<tr>
<td>Nov</td>
<td>31.0</td>
<td>252.8</td>
<td>124.6</td>
<td>122.9</td>
<td>5.3</td>
<td>0.094</td>
<td>0.75</td>
</tr>
<tr>
<td>Dec</td>
<td>68.8</td>
<td>192.8</td>
<td>75.4</td>
<td>115.7</td>
<td>1.7</td>
<td>0.106</td>
<td>0.68</td>
</tr>
<tr>
<td>Annual sum:</td>
<td>1629</td>
<td>means:</td>
<td>296.8</td>
<td>156.6</td>
<td>128.0</td>
<td>12.2</td>
<td>0.165</td>
</tr>
</tbody>
</table>

* Based on gap-filled 30-min $\lambda E$ data during all days of 2004.
The values of $\frac{\lambda E}{\lambda E_{eq}}$ when moisture at the grassland was not limited, agree with measured or modeled evaporation rates from the majority of literature studies over grass under unstressed plant and soil conditions. On the contrary, the maximum spatial mean of ET flux ratios over five grass sites in Kansas (Chen and Brutsaert, 1995) reached almost 1.5, whereas the upper limit of $\alpha$ reported in some drought studies (Meyers, 2001; Li et al., 2007) was lower attaining a maximum of 0.7. Several values for the PT coefficient and critical SWC limits over short, not inundated vegetation are documented in Table 3.5.
Table 3.5: Values of the PT coefficient over grass.

<table>
<thead>
<tr>
<th>PT coefficient</th>
<th>SWC&lt;sub&gt;cr&lt;/sub&gt;</th>
<th>Surface conditions</th>
<th>Location</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;0.5</td>
<td>1.12</td>
<td>0.17</td>
<td>Non irrigated grass</td>
<td>Poinciana, Central FL, US</td>
</tr>
<tr>
<td>0.04</td>
<td>0.7</td>
<td>-</td>
<td>Typical dry steppe grassland</td>
<td>KBU, Hentiy province, Mongolia</td>
</tr>
<tr>
<td>0.87</td>
<td>1.25</td>
<td>-</td>
<td>Temperate riparian grass during summer</td>
<td>Maryhill, South Ontario, Canada</td>
</tr>
<tr>
<td>-</td>
<td>1.20</td>
<td>0.13</td>
<td>Temperate annual grass</td>
<td>Ione, CA, US</td>
</tr>
<tr>
<td>&lt;0.5</td>
<td>0.8-1</td>
<td>0.40</td>
<td>Northern temperate grass</td>
<td>Lethbridge, Canada</td>
</tr>
<tr>
<td>&lt;0.4</td>
<td>1</td>
<td>0.15</td>
<td>Tussock grassland under summer drought</td>
<td>South Island, New Zealand</td>
</tr>
<tr>
<td>0.2</td>
<td>0.7</td>
<td>-</td>
<td>Range under summer all-moisture conditions</td>
<td>South-west Oklahoma, US</td>
</tr>
<tr>
<td>-</td>
<td>1.28</td>
<td>-</td>
<td>Grass under no moisture stress</td>
<td>Netherlands</td>
</tr>
<tr>
<td>0.4</td>
<td>1.5</td>
<td>0.27</td>
<td>FIFE grass in summer (spatial averaging)</td>
<td>Manhattan, KS, US</td>
</tr>
<tr>
<td>~0.2</td>
<td>0.86</td>
<td>-</td>
<td>Serpentine grass</td>
<td>Stanford, CA, US</td>
</tr>
<tr>
<td>0.29</td>
<td>0.9</td>
<td>0.50</td>
<td>Grass under all-moisture conditions</td>
<td>Manhattan, KS, US</td>
</tr>
<tr>
<td>0.9</td>
<td>1.26</td>
<td>-</td>
<td>Grass</td>
<td>Cabauw, Netherlands</td>
</tr>
<tr>
<td>0.8</td>
<td>1.12</td>
<td>-</td>
<td>Short grass</td>
<td>Netherlands</td>
</tr>
<tr>
<td>0.5</td>
<td>1.29</td>
<td>-</td>
<td>Grass at field capacity</td>
<td>Toronto, Canada</td>
</tr>
<tr>
<td>-</td>
<td>1.27±0.02</td>
<td>0.10</td>
<td>Irrigated ryegrass</td>
<td>Ontario, Canada</td>
</tr>
</tbody>
</table>

* SWC<sub>cr</sub> is critical threshold of SWC.
3.4.5 Wetness parameterizations of PT coefficient

The PT method with $\alpha$ value of 1.26 overestimated the actual evapotranspiration during all “good” days of 2004 (Figure 3.12), since this model was originally designed for well-watered conditions and water limitation had the dominant control on ET over the study site. The remedy to overestimation is to reduce and redefine $\alpha$ as a function of the wetness conditions. Starting from the visual inspection of the plots of daytime average values of $\alpha$ versus wetness, an exponential function of SWC and a linear function of CPI were the most suitable equations for fitting the actual rates of $\lambda E/\lambda E_{eq}$ (graphs not shown). Table 3.6 shows the proposed PT functions along with the error statistics of the actual versus modeled ET on daily time scale for the study site. The SWC and CPI parameterizations had good accuracy (defined by high $R^2$, slope close to unity, and intercept approaching zero), and yielded small Mean Bias Errors (MBE) and Root Mean Square Errors (RMSE) with the later being in the order of 20% of the average daytime latent heat fluxes. As a result, both parameterizations of $\alpha$ provided PT evapotranspiration estimates that were not significantly different from the measured ET rates (Figure 3.12).
Table 3.6: Functions of the PT coefficient along with comparison of daytime actual ET versus modeled ET results. The data were fitted to the regression equation

\[ \lambda E_{\text{model}} = A \lambda E_{\text{actual}} + B; \text{ number of days is } N=101. \]

<table>
<thead>
<tr>
<th>Wetness function*</th>
<th>A (Wm⁻²)</th>
<th>B (Wm⁻²)</th>
<th>R²</th>
<th>RMSE (Wm⁻²)</th>
<th>MBE (Wm⁻²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \alpha = 1.26 \cdot \left( 1 - e^{-1.32 \cdot \frac{SWC}{SWC_{fc}}} \right) )</td>
<td>1.03</td>
<td>11.0</td>
<td>0.84</td>
<td>33.1</td>
<td>16.4</td>
</tr>
<tr>
<td>( \alpha = 0.09 \cdot CPI ,(in) + 0.63 )</td>
<td>0.85</td>
<td>35.9</td>
<td>0.79</td>
<td>31.6</td>
<td>12.4</td>
</tr>
</tbody>
</table>

* \( SWC_{fc} \) is the soil water content at field capacity (0.17) and CPI is in inches.
Figure 3.12: Daytime means of measured latent heat flux plotted versus modeled $\lambda E$ based on the SWC (a) and CPI (b) parameterizations under all moisture conditions for the “good” days of 2004. The 1:1 lines are also shown.
3.5 Summary and Concluding Remarks

In this chapter, the annual course of the water vapor over a bahia grass ecosystem in Central Florida was investigated. The major conclusions of this study are outlined as follows.

ET had decreasing trend during interstorm periods in the wet season. This reduction was higher when the antecedent precipitation was more significant. Dry-down analysis indicated that the decrease in CPI and SWC led to depletion of soil moisture storage and exponential drop in ET at this grassland environment (\( \tau = 6.1 \) days of ET model for a composite dry period). Normalized evaporation rates from the grassland were almost constant when moisture was ample (SWC \( > 0.17 \) cm\(^3\) cm\(^{-3}\) or CPI \( > 51 \) mm) and attained a value below that of the PT constant (1.26). During most of 2004, values of the PT parameter \( \alpha \) were less than unity indicating the dominant control of water availability on ET. These findings agree with most of the literature values over grass, which reached maxima of 0.7 to 1.5 under all moisture conditions.

Regression of the PT coefficient in terms of canopy surface conductance and LAI indicated that ET was coupled with plant physiology in this unmanaged grassland. In addition, Pearson correlation matrix showed that air temperature and moisture supply (SWC and CPI) were the most significant environmental controls on ET. When site-calibrated PT functions were introduced in terms of SWC and CPI, good agreement was achieved between the actual and modeled ET data during the “good” days of 2004. The R\(^2\) coefficients for the SWC and CPI functions were 0.84 and 0.79 (p-value<0.001), respectively.
Point measurements of soil moisture exist but scaling and interpolating these measurements to larger domains is problematic (Kurc and Small, 2004). On the contrary, CPI, which is an indicator of the precipitation history (both rain amount and frequency), can readily be available from extensive rainfall networks. Usually an increase in rainfall frequency, which is associated with repeated peaks of CPI, could hinder water stress and enhance ET rates, even during seasons with the same amount of precipitation. Thus, the introduction of the CPI function in the PT coefficient under daily parameterizations appears quite promising, beside the widely documented use of SWC.

About 59% of precipitation returned to the atmosphere as 24-hr ET over a yearly basis. In concluding these data analyses, ET of this bahia grassland was dependent on its biophysical controls and it was linked to soil moisture and precipitation history.
References


CHAPTER 4: TEMPORAL VARIATIONS AND BIOPHYSICAL CONTROLS OF ENERGY PARTITIONING OVER A GRASS SITE IN FLORIDA

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4.1 Introduction

The partitioning between latent ($\lambda E$) and sensible (H) heat fluxes is critical in determining the hydrological cycle, boundary layer development, weather and climate (Wilson et al., 2002). In particular, evapotranspiration from a vegetated surface is controlled by available energy, atmospheric humidity deficit, atmospheric turbulence and stomatal control of the ability of the surface to transmit water to the atmosphere. Among these factors, stomatal regulation governs, to a large extent, the response of atmosphere to the energy partitioning over terrestrial vegetations (Kelliher et al., 1995). In addition, the response of stomata to environmental factors (such as vapor pressure deficit, radiation, temperature and carbon dioxide concentration) is closely related to the process of photosynthesis, and this justifies the coupling of the water and carbon cycles. Subsequently, comprehensive knowledge on the dynamics of energy partitioning in relation to both canopy and environmental controls over various ecosystems is crucial in forecasting the effects of changes in biogeochemical cycling, local weather and global climate.
Approximately 40% of the terrestrial natural vegetation is comprised of grassland ecosystems (White et al., 2000), which show significant annual variations in primary production (Knapp and Smith, 2001). In addition, grassland productivity exhibits asymmetric response to temporal variations in environmental factors, such as precipitation (Wever et al., 2002). In a recent comprehensive study (Law et al., 2002), it has been demonstrated that grasslands have slightly larger index of water use efficiency, in comparison with other terrestrial vegetation types. This indicator is defined as the annually integrated amount of carbon used for photosynthesis relative to the water lost by evapotranspiration, and it was found to be 3.4 g CO$_2$/kg H$_2$O for grasslands. Due to the aforementioned strong link between grass productivity and evapotranspiration, large seasonal and interannual variations in grass evapotranspiration and its biotic and abiotic controls are also observed. It is noteworthy here, that the systematic continuous investigation of the above variations should be a research priority for understanding the biosphere-atmosphere interactions. The majority of grassland studies reported in literature extent mostly to temperate climate zones in North America, such as California (e.g. Baldocchi et al., 2004), Kansas (e.g. Verma et al., 1992), Oklahoma (e.g. Meyers, 2001) and Canada (e.g. Wever et al., 2002).

The non-irrigated grassland of the present study is located at the subtropical region of Central Florida (southeastern US), and is warmer and wetter than most grassland ecosystems. This work, which is based on year-long eddy correlation flux measurements, has the following objectives: (1) to investigate the major biotic and environmental controls on grass evapotranspiration; (2) to describe the diurnal patterns of the energy budget components; (3) to characterize the seasonal patterns of the energy balance and determine the main biophysical factors that modulate the energy partitioning on seasonal time-scale; and (4) to
describe the diurnal and seasonal cycles of canopy characteristics, mainly canopy conductance and decoupling coefficient, and also relate them to environmental conditions and evapotranspiration.

4.2 Methods

4.2.1 Site description

The study site is located within the Disney Wilderness Preserve, Polk County, Florida (28.05 N, 81.40 W). The energy flux measurements were carried out by U.S. Geological Survey (USGS) at an Eddy Correlation (EC) tower located within 100 m of a South Florida Water Management District (SFWMD) weather station, as shown in Figure 4.1. The terrain is relatively flat with elevation of 18 m above the mean sea level. The soil is composed of fine sands with an overall soil porosity of 40%. The soil water content (SWC) at wilting point is 7% and at field capacity is 17% (Dingman, 2002). About 70% of the vegetation is dominated by bahia grass (Paspalum notatum), while the rest includes a few scattered trees as well as other plants (e.g. Eupatorium capillifolium, Andropogon virginicus, Serenoa repens, Dichanthelium portoricense). The bahia grass is non-irrigated and ungrazed with grass height of up to 40 cm and root system reaching on average 1 m. The greenness of the grass varies from a drab brown during the winter to a lush green during the summer (Figure 4.2). The climate is subtropical and humid. During 2004, average annual air temperature was 21.4 °C and average yearly precipitation was about 1629 mm, of which about 80% fell from May through October (wet season). This site was described in details in a previous study (Rizou et al., 2008).
Figure 4.1: Aerial map showing the location of the EC and weather stations (Google Maps, accessed November 2007, http://www.google.com).

Figure 4.2: Variation of the greenness of the bahia grass from drab brown in February (a) to lush green in July (b).
4.2.2 Measurements

The study period covered the annual course of 2004. The EC system included a CSAT3 3-D sonic anemometer and a KH20 krypton hygrometer, installed at 3.4 m above ground. The tower was also equipped to measure air temperature and humidity\(^1\) at 1.2 m, net radiation\(^2\) at 3.4 m, incoming solar radiation\(^3\) at 3 m, and wind speed and direction\(^4\) at 3.6 m. The EC sensors (3-D sonic anemometer and krypton hygrometer) were monitored at 8 Hz. Latent heat and sensible heat fluxes were computed at 30-minute resolution and then logged on a CR10X datalogger\(^5\).

The soil parameters were measured as follows: soil heat flux at 8 cm depth by soil heat plates\(^6\), soil temperature by averaging soil thermocouple probes\(^7\) placed from 0 to 8 cm depth, soil moisture at 0 to 8 cm and 0 to 30 cm depth by water content reflectometers\(^8\), and water table depth by pressure transducer\(^9\). Precipitation at the adjacent SFWMD site was measured using a tipping-bucket rain gauge\(^10\) and daily totals were recorded on a CR10X datalogger.

Leaf Area Index (LAI) and Fraction of Photosynthetically Active Radiation (FPAR) data were retrieved from MODerate resolution Imaging Spectroradiometer (MODIS 15) land cover product via the USGS EROS Data Center (EDC) Distributed Active Archive Center.

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\(^1\) model HMP45C, Campbell Scientific Inc. (CSI), Logan, UT  
\(^2\) model Q-7.1 net radiometers, Radiation Energy Balance Systems (REBS), Seattle, WA  
\(^3\) model LI200X pyranometer, LICOR Inc., Lincoln, NE  
\(^4\) model 05305-5 wind monitor-AQ, R.M. Young, Traverse City, MI  
\(^5\) model CR10X, CSI  
\(^6\) model HFT-3, CSI  
\(^7\) model TCAV, CSI  
\(^8\) model CS615, CSI  
\(^9\) KPSI Series 500, Pressure Systems Inc., Hampton, VA  
\(^10\) model 6011-A, All Weather Inc., Sacramento, CA
(DAAC) (http://lpdaac.usgs.gov, accessed July 2007). The satellite data are provided at 1-km spatial resolution and 8-day temporal resolution. Description of the MODIS algorithms and data derivations can be found in Knyazikhin et al. (1999). A more detailed description of measurements is included in the complementary study by Rizou et al. (2008).

4.2.3 Data processing and calculations

4.2.3a EC flux data analysis and gap-filling

Evapotranspiration (ET) was measured using an energy budget variant of the EC method (Tanner and Greene, 1989; Twine et al., 2000). Estimates of latent heat flux were corrected for temperature-induced fluctuations in air density (Webb et al., 1980) and for the sensitivity of the hygrometer to oxygen (Tanner and Greene, 1989). Sensible heat fluxes were corrected for differences between the “sonic” temperature and the actual air temperature (Schotanus et al., 1983). Both sensible and latent heat fluxes were corrected for misalignment of the sonic anemometer with the airstream by the coordinate rotation procedure (Baldocchi et al., 1988). The Bowen ratio \( B = H/\lambda E \) method was applied in order to force the measured fluxes to satisfy the canopy energy budget (Twine et al., 2000).

Missing or rejected ET data were gap-filled with a regression equation included in Section 4.3.1. About 67% of the daytime 30-min values were missing and therefore required gap-filling. The data analyses were conducted on daytime measurements, since daytime ET is generally much higher than nighttime ET. The 30-min daytime data correspond to times
when the net radiation is greater than $5 \ W \ m^{-2}$, and they are averaged over a daily basis. More specifically, the nighttime fluxes were deemed missing and not equal to zero. Data tests involving regression analyses used “good” days only. “Good” day is considered any day with more than 65% of 30-min intervals of measured daytime ET values, with the missing 30-min gaps filled. The complete annual time series of the turbulent fluxes were used for other analyses, such as temporal variations.

4.2.3b Canopy resistance and decoupling coefficient

The half–hourly daytime values of aerodynamic conductance ($g_a$) and bulk canopy surface conductance ($g_c$) were calculated based on an inverted form of the Penman-Monteith (Monteith, 1965) model. The stomatal decoupling coefficient ($\Omega$) was also determined in order to evaluate the effect of a fractional change in surface conductance on transpiration rate (Jarvis and McNaughton, 1986). This coefficient, which lies in the range of $[0,1]$, is given by the following equation:

$$\Omega = \left(1 + \frac{\gamma}{\gamma + \Delta} \cdot \frac{r_c}{r_a}\right)^{-1}$$  \hspace{2cm} (4.1)

where $r_a = 1/\gamma$ is the aerodynamic resistance ($s \ m^{-1}$) between the effective canopy surface and the reference height at which micrometeorological measurements are made, $r_c = 1/g_c$ is the bulk canopy surface resistance ($s \ m^{-1}$), $\Delta$ is the slope of the saturation vapor pressure-temperature curve of air ($KPa ^{\circ} C^{-1}$), and $\gamma \approx 0.067$ is the psychrometric constant ($KPa ^{\circ} C^{-1}$).
The Penman Monteith model with the introduction of $\Omega$ becomes:

$$\lambda E = \Omega \left[ \frac{\Delta}{\Delta + \gamma} (R_n - G) \right] + (1 - \Omega) \left[ \frac{\rho_a C_p (e_s - e)}{\gamma r_c} \right]$$

(4.2)

where $R_n$ is the net radiation ($Wm^{-2}$), $G$ is the soil heat flux ($Wm^{-2}$), $\rho_a$ is the moist air density ($kgm^{-3}$), $C_p$ is the specific heat of air under constant pressure ($Jkg^{-1}^C^{-1}$), $e_s$ is the saturation vapor pressure (KPa), $e = e_s \cdot RH$ is the actual vapor pressure (KPa), $RH$ is the relative humidity, and $(e_s - e) = VPD$ is the vapor pressure deficit (KPa).

Rough surfaces (forests), which are well coupled to the mixed layer, have small $\Omega$ values of less than 0.5, while smooth surfaces (short grass at moderate wind speed) have $\Omega$ values greater than 0.5 (McNaughton and Jarvis, 1983). Similarly to other studies (Goodrich et al., 2000; Sumner and Jacobs, 2005), the 30-min daytime data when $g_c < 15 mm s^{-1}$ were used for the calculations of $g_a$, $g_c$ and $\Omega$ in order to model normal atmospheric conditions. About 14% of the 30-min data were eliminated from the yearly dataset. For a detailed account of data processing see Rizou et al. (2008).

4.3 Results

4.3.1 PT coefficient

In order to characterize the grass evapotranspiration, the Priestley–Taylor (PT) coefficient $\alpha$ (Priestley and Taylor, 1972), defined as the ratio of measured to equilibrium $\lambda E$, was calculated based on measured fluxes and meteorological data. The equilibrium $\lambda E$ is given by the relation: $\lambda E_{eq} = \Delta (R_n - G)/\left(\Delta + \gamma\right)$. A multiple linear regression analysis was applied to determine the dependence of $\alpha$ on environmental variables, as described in previous
studies (Sumner, 2001; Sumner and Jacobs, 2005; Rizou et al., 2008). The following site-calibrated equation for $\alpha$ was used for gap-filling the 30-min latent heat fluxes:

$$\alpha = (a \cdot SW_{in}^2 + b \cdot SW_{in} + c \cdot WT + d) \cdot g(DOY),$$

$$g(DOY) = e \cdot \sin\left[ \omega(DOY - f) \right] + 1$$

(4.3)

where $SW_{in}$ is the incoming solar radiation ($W m^{-2}$), $WT$ is the depth from land surface to water table ($m$), DOY is the Julian day of year 2004, and $\omega$ is the annual period ($\omega = 2\pi/365$ day$^{-1}$). The best-fit parameters are: $a = 5.52 \cdot 10^{-7}$, $b = -9.32 \cdot 10^{-4}$, $c = -0.259$, $d = 1.23$, $e = 0.24$ and $f = 126$. The performance of the regression Equation (4.3) after substitution in the PT equation was associated with a $R^2$ of 0.94 and a Standard Error (SE) of 0.98 mm day$^{-1}$. Equation (4.3) shows that $\alpha$ has a negative linear relation with $WT$, for the bounds of $SW_{in}$ observed in this study, and decreases by about a factor of 1.5 for a given value of solar radiation, over the range of $WT$ values present during the experiment (Figure 4.3).
Figure 4.3: Site-calibrated values of PT coefficient in terms of WT (a) and incoming solar radiation (b). The negative WT indicates flooding.
4.3.2 Biophysical controls on ET by drying stage

Evapotranspiration is controlled by the complex, interacting effect of several biotic and abiotic variables that limit the demand and supply of water vapor to the atmosphere. The major abiotic factors include the atmospheric evaporative demand (net radiation, vapor pressure deficit), moisture supply (soil moisture, water table), air temperature and wind speed. The biotic supply depends on canopy surface conductance ($g_c$), LAI and other ecophysiological features of plant functional type and phenological stage. The daytime average $\lambda E$ at the study site was classified into two stages, energy-limited (wet stage) and water-limited (dry stage), based on a critical SWC value of 0.17 cm$^3$ cm$^{-3}$ (Rizou et al., 2008). Since it was found that water availability has a dominant control on $\lambda E$ of this grassland, the biophysical and $\lambda E$ data were grouped into two classes (stages) according to the volumetric SWC (at top 30 cm of soil). Then, the relation of $\lambda E$ with each of its controls was analyzed using linear regression for 101 “good” days of the year (Figure 4.4 and Table 4.1). The significant effects on $\lambda E$ were: radiation ($R_n$ and $SW_{in}$), air temperature ($T_a$), VPD, LAI and $g_c$. LAI gaps were interpolated to daily intervals by using a third order polynomial.
Figure 4.4: The relationship between daytime evapotranspiration and its controls, namely $R_n$ (a), $SW_{in}$ (b), $T_a$ (c), VPD (d), LAI (e) and $g_c$ (f), during “good” days of 2004. The data were sorted into two classes based on the critical SWC. Each class was fitted to the linear model $Y = a \cdot X + b$, where $Y$ is $\lambda E$ or $\alpha$ and $X$ is the biophysical control.
Table 4.1: Parameters for the linear regression $Y = a \cdot X + b$ at level of significance 0.05. Same as Figure 4.4.

<table>
<thead>
<tr>
<th></th>
<th>$\lambda E = Y$</th>
<th>SWC</th>
<th>a</th>
<th>b</th>
<th>$R^2_{adj.}$</th>
<th>F</th>
<th>N</th>
<th>p</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R_n$</td>
<td>$&lt;17%$</td>
<td>0.42</td>
<td>6.67</td>
<td>0.43</td>
<td>60.5</td>
<td>80</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$&gt;17%$</td>
<td>0.74</td>
<td>-12.35</td>
<td>0.96</td>
<td>503.1</td>
<td>21</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td>SW$_{in}$</td>
<td>$&lt;17%$</td>
<td>0.27</td>
<td>15.86</td>
<td>0.32</td>
<td>38.2</td>
<td>80</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$&gt;17%$</td>
<td>0.54</td>
<td>-14.06</td>
<td>0.89</td>
<td>166.9</td>
<td>21</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td>$T_a$</td>
<td>$&lt;17%$</td>
<td>8.29</td>
<td>-50.79</td>
<td>0.58</td>
<td>110.1</td>
<td>80</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$&gt;17%$</td>
<td>31.51</td>
<td>-665.32</td>
<td>0.77</td>
<td>69.54</td>
<td>21</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td>VPD</td>
<td>$&lt;17%$</td>
<td>66.13</td>
<td>34.14</td>
<td>0.42</td>
<td>57.8</td>
<td>80</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$&gt;17%$</td>
<td>120.07</td>
<td>92.19</td>
<td>0.58</td>
<td>28.7</td>
<td>21</td>
<td>&lt;0.0001</td>
<td></td>
</tr>
<tr>
<td>LAI</td>
<td>$&lt;17%$</td>
<td>41.16</td>
<td>67.82</td>
<td>0.12</td>
<td>11.5</td>
<td>80</td>
<td>0.001</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$&gt;17%$</td>
<td>-235.17</td>
<td>813.92</td>
<td>0.15</td>
<td>4.7</td>
<td>21</td>
<td>0.044</td>
<td></td>
</tr>
</tbody>
</table>

There was strong linear correlation between $\lambda E$ and the abiotic factors when SWC was ample (SWC>17%), as this is expressed by the adjusted coefficient of determination ($R^2_{adj}$). The above correlation along with the slope a (which is an increase of $\lambda E$ per fractional change of the effect) were reduced when water was limited (SWC<17%). More specifically, between 58% and 96% of the variance in $\lambda E$ at the study site was associated with the environmental effects in the wet stage (N=21 days), with net radiation being the most dominant factor on $\lambda E$. On the contrary, between 32% and 58% of the $\lambda E$ variability was explained by the abiotic effects, with air temperature being the most significant effect.
under water limitations (N=80 days). It is also noteworthy that the slopes of the linear regression model with respect to VPD were large (Table 4.1), indicating high sensitivity of $\lambda E$ to VPD (this will be further discussed in Section 4.3.4).

The response of $\lambda E$, as represented by the PT coefficient $\alpha$, to $g_c$ was reverse to the abiotic controls (Figure 4.4 (f)). There was a 70% variance of $\alpha$ accounted for by the variance in $g_c$ under water-stressed conditions, and no correlation during the energy-limited stage. In addition, $\lambda E$ had weak correlation with the daily LAI, and the wet stage data violated the linear regression assumptions for random errors and equal variances (Figure 4.4 (e)). The inadequacy of the regression model in terms of the biotic factors when SWC>17% is explained by the high dependence of $\lambda E$ or $g_c$ on environmental conditions (such as water supply and vapor pressure deficit), when water availability is high. Such observation has been consistently documented in the literature (Ritchie, 1972; Verma et al., 1992; Wever et al., 2002). Though inadequate, the inverse relationship between $\lambda E$ and LAI under ample water supply might be due to soil evaporation, which tends to be inversely proportional to LAI. The above relationship has also been reported by Law et al. (2002).

4.3.3 Atmospheric and surface conditions

Understanding the patterns of the prevailing environmental conditions is a necessary step prior to the analysis of the temporal variations of the water and energy exchange between the atmosphere and the grassland. Figure 4.5 illustrates the daily averages of wind speed (w), air temperature ($T_a$), air humidity (VPD, RH), volumetric soil water content (SWC) and
precipitation during 2004. The daily air temperature showed a clear seasonal pattern with
greater temperatures observed during the wet summer season (Figure 4.5 (b)). The maximum
air temperature was 29.5 °C recorded on 23 June and the minimum was 6.2 °C on 20
December. Seasonal trends were also observed for VPD. Higher humidity deficits occurred
between April and July with maximum values recorded at the end of the dry May (Figure 4.5
(c)).
Figure 4.5: Seasonal variation of daily means of environmental variables; a) wind speed (w), b) air temperature ($T_a$), c) vapor pressure deficit (VDP) and relative humidity (RH), d) volumetric soil water content (SWC) and precipitation.
The variations in SWC were associated with the amount and timing of precipitation (Figure 4.5 (d)). A total of 1629 mm of precipitation was recorded over the year, with the largest single day total of 186 mm occurring on 26 September due to a hurricane event. During the first half of the year, SWC at the top 30 cm of the soil fluctuated around an almost constant value, and it reached the wilting point (0.07) on 1-3 June. Soil moisture level exceeded the soil field capacity during the wet season, attaining a maximum value of 0.52 on 6 September in response to a significant rain at the end of a 2-month rainy season. This was followed by a decrease in SWC during a 10-day scarcity of the precipitation, and an increase to high moisture in response to the largest rain event of the year.

Figure 4.6 illustrates the seasonal variation of the active vegetation status via LAI and FPAR. LAI is defined as the one-side leaf area per unit ground area, whereas the fraction of incoming solar radiation at the photosynthetically active wavelengths (0.4-0.7 mm) absorbed by the plant canopy is defined as FPAR. The variation in LAI and FPAR were associated with differences in precipitation and soil moisture. The summer rainstorms occurring from early June to late September provided water input to the soil and increased water availability to the plants (Figure 4.5 (d)). This resulted in large LAI values (greater than $2 \text{m}^2 \text{m}^{-2}$ on average) during a five month period (approximately DOY 190-347). LAI and FPAR exhibited maxima of $3.6 \text{m}^2 \text{m}^{-2}$ and 89%, respectively, on DOY 266 (22 September).
4.3.4 Diurnal energy partitioning and stomatal control

The ensemble average of the diurnal cycle of $R_n$, $\lambda E$, $H$ and $G$ is depicted in Figure 4.7 for the months of January and August. The highest $\lambda E$ yielding the lowest monthly Bowen ratio (B) was observed during the wet August, whereas the low soil moisture in January resulted in the increase of $H$ and the maximum monthly B (Table 4.2). Latent heat flux was the dominant component of the energy balance at midday of the wet season in response to high water availability and LAI (Table 4.2, Figure 4.5(d) and Figure 4.6). More specifically, latent heat flux was higher than sensible heat flux from DOY 152 to 321, with average midday values of $\lambda E$ approximately $416 \text{ Wm}^{-2}$ during August. In contrast, sensible heat flux dominated the energy budget at midday for the rest of the year with maximum value of $258 \text{ Wm}^{-2}$ in January. Generally over grass, the maxima of $\lambda E$ are coincident with the maxima of $R_n$ at about midday hours (Kelliher et al., 1993), and this was the case for the current
grassland. The peak midday 30-min values of $\lambda E$ and $H$ were 613 $Wm^{-2}$ and 421 $Wm^{-2}$ on DOY 231 and DOY 7, respectively. The course of the soil heat flux component did not show large differences between the two months with midday monthly values reaching approximately 18–38 $Wm^{-2}$.

Figure 4.7: Mean diurnal course of daytime energy budget components for the months of January and August 2004.
Table 4.2: Monthly means of daytime energy flux ratios, moisture (SWC, WT), atmospheric conditions (Rn, SWin, $T_a$, RH, VPD, w), and PT coefficient $\alpha$; the total monthly precipitation is also included.

<table>
<thead>
<tr>
<th>Month</th>
<th>DOY</th>
<th>Prec. (mm)</th>
<th>$R_n$ (Wm$^{-2}$)</th>
<th>$\lambda E/R_n$</th>
<th>$H/R_n$</th>
<th>$G/R_n$</th>
<th>$B=H/\lambda E$ (Wm$^{-2}$)</th>
<th>SWin (cm$^3$cm$^{-3}$)</th>
<th>SWC (cm$^3$cm$^{-3}$)</th>
<th>WT (m)</th>
<th>$T_a$ (°C)</th>
<th>RH (%)</th>
<th>VPD (KPa)</th>
<th>w (ms$^{-1}$)</th>
<th>$\alpha$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>31</td>
<td>88.1</td>
<td>236.1</td>
<td>0.34</td>
<td>0.63</td>
<td>0.03</td>
<td>1.83</td>
<td>366.6</td>
<td>0.112</td>
<td>0.866</td>
<td>18.4</td>
<td>62.7</td>
<td>0.91</td>
<td>2.1</td>
<td>0.59</td>
</tr>
<tr>
<td>Feb</td>
<td>60</td>
<td>69.1</td>
<td>235.4</td>
<td>0.36</td>
<td>0.59</td>
<td>0.05</td>
<td>1.65</td>
<td>348.7</td>
<td>0.129</td>
<td>0.703</td>
<td>20.0</td>
<td>67.1</td>
<td>0.88</td>
<td>2.3</td>
<td>0.64</td>
</tr>
<tr>
<td>Mar</td>
<td>91</td>
<td>41.9</td>
<td>321.5</td>
<td>0.40</td>
<td>0.56</td>
<td>0.05</td>
<td>1.41</td>
<td>477.8</td>
<td>0.114</td>
<td>0.746</td>
<td>22.7</td>
<td>57.5</td>
<td>1.29</td>
<td>2.6</td>
<td>0.65</td>
</tr>
<tr>
<td>Apr</td>
<td>121</td>
<td>57.4</td>
<td>359.4</td>
<td>0.40</td>
<td>0.56</td>
<td>0.03</td>
<td>1.38</td>
<td>537.1</td>
<td>0.084</td>
<td>0.922</td>
<td>23.8</td>
<td>54.9</td>
<td>1.45</td>
<td>2.5</td>
<td>0.64</td>
</tr>
<tr>
<td>May</td>
<td>152</td>
<td>26.9</td>
<td>363.4</td>
<td>0.44</td>
<td>0.52</td>
<td>0.04</td>
<td>1.20</td>
<td>519.4</td>
<td>0.075</td>
<td>1.108</td>
<td>27.8</td>
<td>58.0</td>
<td>1.77</td>
<td>2.4</td>
<td>0.67</td>
</tr>
<tr>
<td>Jun</td>
<td>182</td>
<td>276.6</td>
<td>372.3</td>
<td>0.58</td>
<td>0.37</td>
<td>0.05</td>
<td>0.63</td>
<td>504.7</td>
<td>0.108</td>
<td>0.808</td>
<td>30.1</td>
<td>66.7</td>
<td>1.61</td>
<td>1.9</td>
<td>0.85</td>
</tr>
<tr>
<td>Jul</td>
<td>213</td>
<td>218.9</td>
<td>340.9</td>
<td>0.70</td>
<td>0.26</td>
<td>0.05</td>
<td>0.37</td>
<td>462.9</td>
<td>0.198</td>
<td>0.488</td>
<td>29.8</td>
<td>69.2</td>
<td>1.44</td>
<td>1.8</td>
<td>1.02</td>
</tr>
<tr>
<td>Aug</td>
<td>244</td>
<td>371.6</td>
<td>322.8</td>
<td>0.77</td>
<td>0.17</td>
<td>0.06</td>
<td>0.22</td>
<td>435.1</td>
<td>0.427</td>
<td>0.143</td>
<td>29.4</td>
<td>76.0</td>
<td>1.08</td>
<td>1.6</td>
<td>1.13</td>
</tr>
<tr>
<td>Sep</td>
<td>274</td>
<td>365.5</td>
<td>279.2</td>
<td>0.72</td>
<td>0.23</td>
<td>0.05</td>
<td>0.31</td>
<td>384.0</td>
<td>0.370</td>
<td>0.256</td>
<td>28.5</td>
<td>75.3</td>
<td>1.03</td>
<td>3.8</td>
<td>1.08</td>
</tr>
<tr>
<td>Oct</td>
<td>305</td>
<td>13.0</td>
<td>285.3</td>
<td>0.63</td>
<td>0.34</td>
<td>0.03</td>
<td>0.55</td>
<td>420.1</td>
<td>0.159</td>
<td>0.644</td>
<td>26.5</td>
<td>68.6</td>
<td>1.16</td>
<td>2.2</td>
<td>0.91</td>
</tr>
<tr>
<td>Nov</td>
<td>335</td>
<td>31.0</td>
<td>252.8</td>
<td>0.49</td>
<td>0.49</td>
<td>0.02</td>
<td>0.99</td>
<td>384.1</td>
<td>0.094</td>
<td>0.894</td>
<td>23.7</td>
<td>66.3</td>
<td>1.07</td>
<td>2.4</td>
<td>0.75</td>
</tr>
<tr>
<td>Dec</td>
<td>366</td>
<td>68.8</td>
<td>192.8</td>
<td>0.39</td>
<td>0.60</td>
<td>0.01</td>
<td>1.53</td>
<td>303.4</td>
<td>0.106</td>
<td>0.944</td>
<td>18.6</td>
<td>66.1</td>
<td>0.78</td>
<td>2.6</td>
<td>0.68</td>
</tr>
</tbody>
</table>

Annual sum: 1629
Mean: 296.8 0.52 0.44 0.04 1.01 428.7 0.165 0.710 24.9 65.7 1.21 2.3 0.80
The partitioning of available radiation into sensible and latent heat flux is influenced by changes in vegetation characteristics and moisture availability. It has been reported in a number of studies that a switch in energy partitioning (from $H$ to $\lambda E$ dominated) is associated with larger LAI and wetter soils and canopies (Wilson and Baldocchi, 2000; Wever et al. 2002). This will be further discussed in the next section.

Figure 4.8 presents the diurnal cycle of stomatal response, expressed as $g_c$ and $\Omega$, to Photosynthetically Active Radiation (PAR) and VPD for two months, which were characterized by dissimilar canopy controls on $\lambda E$. Throughout the year, dry May exhibited the highest humidity deficit (average VPD of 1.77 KPa), and wet August was associated with peak $\lambda E$ rates governed by high $g_c$ and $\Omega$. In Figure 4.8 (a), the 30-min resolution PAR was found by a direct relationship to $SW_{in}$, based on field data collected for bahia grass in Central Florida (Sumner, 2001). The approximation used is: $PAR = 2.04 \ SW_{in}$, where PAR is in $\mu$moles m$^{-2}$s$^{-1}$ and $SW_{in}$ is in W m$^{-2}$. 
Figure 4.8: Mean diurnal course of daytime: a) photosynthetically active radiation (PAR) and vapor pressure deficit (VPD), b) canopy surface conductance ($g_c$) and decoupling coefficient ($\Omega$) for the months of May and August 2004.
High values of $\Omega$ indicate that $\lambda E$ is more sensitive to net radiation, whereas small values of $\Omega$ show high sensitivity of $\lambda E$ to ambient humidity deficit and surface conductance, thus implying enhanced coupling between the canopy and the mixed layer in the later case (see Equation (4.2)). On daily basis, $g_c$ and $\Omega$ were highest in the morning but then declined through the rest of the day, suggesting stronger stomatal control of evapotranspiration losses by plants as the day progressed. This pattern, typical of the full-leaf period, is explained by the low-PAR saturation level that allows maximum $g_c$ with maximum photosynthesis and minimum water loss to occur at the early morning hours. Later in the day, $g_c$ and $\Omega$ decreased due to high VPD, indicating a high response of the above parameters to VPD. An increasing degree of the coupling to the atmosphere with time from morning to afternoon has been also reported in various studies (e.g. Korner, 1994; Wright et al., 1995; Blanken et al.; 1997; Wever et al., 2002; Li et al., 2006). Over the mean diurnal course of August, $g_c$ and $\Omega$ of the present grassland declined from 14.4 to 0.9 mm s$^{-1}$ and from 0.82 to 0.36, respectively. These stomatal parameters were lower during May with corresponding ranges 11.4-0.5 mm s$^{-1}$ and 0.93-0.09. Figure 4.8 implies that these low values of $g_c$ and $\Omega$ were stimulated by the larger VPD and PAR observed in May compared to August. Thus, $\lambda E$ was more sensitive to $g_c$ and VPD during May rather than in August.

To expand on the physiological feedback mechanism on evapotranspiration, daily means of daytime $g_c$ and $\lambda E$ are plotted against VPD, and the data are grouped by VPD bins of width 0.2 KPa (Figure 4.9). Despite the noise induced by other environmental effects (especially at low VPD), there is some apparent relationship between $g_c$ and VPD during the
months of June through October. Canopy surface conductance declined with the increase of VPD due to rainless periods and high atmospheric demand. This suggests stomatal closure that restricts transpiration losses. It is apparent from Figure 4.9 that $g_c$ of the bahia grass reached consistently low values and resulted in restricting and leveling out the evapotranspiration at VPD larger than approximately 2 KPa.
Figure 4.9: Daily means of daytime $g_c$ (a) and $\lambda E$ (b) plotted versus VPD during the period from June to October. The closed diamonds denote bin-averaged data (0.2 KPa bins). The bars indicate ± 1 SE.
According to a comprehensive study for various types of vegetation by Law et al. (2002), stomatal closure causes not only a decrease in transpiration but also in photosynthesis, especially when water availability or hydraulic capacity of the whole plant system (leaves, stem, roots) is limited. It was also found that the reduction in water use efficiency of grasslands (i.e. the ratio of gross ecosystem production to ET) during the summer months occurred at VPD larger than about 1.5 KPa, which is close to the value of the present study (2 KPa). In addition, the VPD limit to stimulate stomatal closure in this grassland agrees with the values (about 1 KPa) reported by Baldocchi et al. (1997), Blanken et al. (1997), Anthoni et al. (1999) and Pejam et al. (2006), and the value (about 3 KPa) reported by Scott et al. (2004) for various forest ecosystems. The negative, nonlinear correlation between $g_c$ and VPD, which is inherent in the Penman-Monteith model, was also observed over several grasslands (Stewart and Gay, 1989; Wever et al., 2002; Li et al., 2006). This physiological feedback is more evident on sunny days, and this justifies the stratification of the above relation into PAR or $SW_{in}$ levels in some literature.

4.3.5 Seasonal energy partitioning and stomatal control

Energy partitioning into latent and sensible heat fluxes showed a distinct seasonal pattern. Figure 4.10 (a) presents the annual course of the 7-day average daytime values of the energy budget components. The monthly means of the energy and climatic components are given in Table 4.2. Throughout the year, daytime $R_n$ varied from 25 $Wm^{-2}$ (25 December) to 451 $Wm^{-2}$ (26 June) with an average of 297 $Wm^{-2}$. It is apparent from Figure 4.10, that the
period from early June to middle of November (DOY 152-321) was characterized by a dominant latent heat flux and a suppressed sensible heat flux component. During this $\lambda E$-dominated period, variation in $\lambda E$ traced the variation in $R_n$ very closely, and a maximum daytime $\lambda E$ value of 337 $Wm^{-2}$ (11.9 $mm day^{-1}$) was observed on DOY 223 (10 August) corresponding to a high value of $R_n$. Similarly during the $H$ dominated period, the pattern of $H$ followed that of the $R_n$, and a maximum daytime $H$ value of 268 $Wm^{-2}$ was observed on DOY 106 (15 April). The fluctuations of $R_n$ are mainly caused by the increased cloud cover that is typical of the Florida summer-time convective systems. During cloudy periods, such as the intervals DOY 198-203, 249-250 and 284-287, reductions in $R_n$ suppressed the latent and sensible heat fluxes. Several researchers have suggested that this effect of the cloud cover might also be responsible for energy flux inconsistencies (Burba et al., 1999; Jacobs et al., 2002; Wu et al., 2007).
To account for variations in available energy, the fluxes were normalized with respect to the net radiation (Figure 4.10 (b)). During the winter month of January, a large portion of the average daytime net radiation, i.e. 63%, was partitioned to sensible heat flux, 34% to evapotranspiration, and the remainder to ground heat flux (Table 4.2). On the contrary, most of the $R_n$ was used in $\lambda E$ throughout the wet season. Specifically during August, 77% of the
net radiation was partitioned to $\lambda E$, 17% to H, and the remainder to G. On monthly time scale, the average daytime evapotranspiration ranged from 2.7 to 8.8 mm day$^{-1}$ for December and August, respectively. Over the annual course, more than 90% of $R_n$ was converted to turbulent fluxes and the rest was a small contribution by G (monthly average of ground heat flux was 1 to 6%).

Bowen ratio becomes lower when moisture supply is high. On a seasonal basis, the mean daytime B was decreased by a factor of 3 from the dry to the wet season (to a value of 0.51), and its variation formed a “U” shape with a broad plateau occurring in the wet season. On a monthly basis, B changed from a value of 1.83 in the dry January to 0.22 in the wet August (Table 4.2). The Bowen ratio exhibited an annual average of 1 and values larger than 2 during few dry days mainly at the beginning of the year.

Besides $R_n$, soil moisture and vegetation phenology are key variables that modulate $\lambda E$ transformation. The shift from an H- to a $\lambda E$-driven system on early summer (about DOY 152) can be attributed to the increased soil moisture (during DOY 156-282) and enhanced LAI (during DOY 190-347). As shown in Figure 4.10, the changes in $\lambda E$ and B were in synchrony with the variations in SWC and LAI rather than with $R_n$ within the wet season.

Similar findings about the impact of seasonal variations of $R_n$, soil moisture, and vegetation activity on the magnitude and distribution of $\lambda E$ and H fluxes were reported for several ecosystems, such as forests (Blanken et al., 1997; Wilson and Baldocchi, 2000; Wu et al., 2007) and grasslands (Kim and Verma, 1990; Verma et al., 1992; Valentini et al, 1995; Ham and Knapp, 1998; Saigusa et al., 1998; Burba and Verma, 2001; Wever et al., 2002; Frank, 2003; Li et al., 2006; Hao et al, 2007). Among the above studies, there are few
(Blanken et al., 1997; Ham and Knapp, 1998; Wilson and Baldocchi, 2000) that examined the sole contribution of canopy characteristics on the energy partitioning during non-limiting water conditions. Furthermore, the variations in the energy partitioning due to soil moisture availability and mainly canopy activity exert feedback on the properties of the atmospheric boundary layer. Greener LAI reduces sensible heat flux, and often synchronizes with increased humidity and smaller diurnal temperature range, as well as higher probability of cloud formation (Schwartz, 1996; Wilson and Baldocchi, 2000; Freedman et al., 2001).

The seasonal dynamics of the grass daytime $g_c$ with relation to VPD variation is shown in Figure 4.11 (a). The mean daytime canopy surface conductance increased during the wet summer season, and it ranged from 1.6 to 12 mm s$^{-1}$ throughout the year. More specifically, during the period from DOY 156 to 347, when a pattern of increased LAI and SWC was emerged, the enhanced $g_c$ averaged 6.5 mm s$^{-1}$. Following its peak, $g_c$ began to decrease rapidly at approximately DOY 254 suggesting stomatal closure caused by the low soil moisture during a dry event (DOY 254-263). The maximum value of $g_c$ as well as its seasonal trends observed in this study are similar to findings of other studies conducted in grasslands (Verma et al., 1992, Ham and Knapp, 1998; Li et al., 2006) and forests (Blanken et al., 1997; Wilson and Baldocchi, 2000).
Figure 4.11: Seasonal variation of daytime average $g_c$ and VPD (a) along with decoupling coefficient and PT coefficient $\alpha$ (b) during 2004. DOY on the x-axis is the center point of a 7-day average. The arrows represent drops in $g_c$ and $\Omega$.

The seasonal course of $\Omega$ and $\alpha$ is presented in Figure 4.11 (b). The values of $\Omega$ range from 0 to 1, with the stomatal control of transpiration becoming weaker as $\Omega$
approaches 1. In this study, the annual mean of daytime $\Omega$ value was 0.5 indicating moderate degree of coupling between the vegetation and the atmosphere. The lowest daytime $\Omega$ value was approximately 0.2 (enhanced coupling) on DOY 7, and the maximum of 0.9 (weak coupling) recorded on DOY 222. In similar way, the Priestley-Taylor $\alpha$ is used to describe the regional interaction between the surface and the boundary layer (Blanken et al., 1997), with the control of water availability on evaporation growing stronger as $\alpha$ becomes much below unity. Over the year, the mean daytime $\alpha$ was 0.8 (with bounds of 0.4 and 1.5) indicating some water supply limitation on evaporation and stomatal control on transpiration. Both $\Omega$ and $\alpha$ values increased at the beginning of the wet season in association with an increase in SWC and reduction in VPD, and then declined for the remaining of the season due to reduction in water supply (Figure 4.5 and Figure 4.11).

The values of $\Omega$ obtained in the present study agree with literature values, which were documented to be in the range 0.1-0.8, for serpentine grassland (Valentini et al., 1995) and maize crop (Steduto and Hsiao, 1998). However, this grassland shows lower sensitivity of evapotranspiration to surface conductance and VPD than some other grasslands (Wever et al., 2002; Hao et al., 2007) and forests (Blanken et al., 1997; Wilson and Baldocchi, 2000). The later studies reported $\Omega$ values in the range 0.1-0.5. In addition, the $\alpha$ values of this study compare well with literature findings outlined in Rizou et al. (2008).

4.4 Summary

This study has documented the annual dynamics of energy fluxes and biophysical controls on evapotranspiration, and investigated their relationship over an unmanaged grassland in Central Florida. With regard to the stated objectives, the following conclusions are made.
Evapotranspiration is controlled by a combination of abiotic and biotic variables. In this work, the response of $\lambda E$ to environmental and stomatal controls sorted by drying stage was evaluated. Net radiation was the most dominant abiotic control on $\lambda E$ in the energy-limited stage ($R_{adj}^2 = 0.96$), whereas the air temperature was the most significant effect during the water-limited stage ($R_{adj}^2 = 0.58$). Regarding canopy effects, canopy conductance ($g_c$) was strongly positively correlated to $\lambda E$ ($R_{adj}^2 = 0.70$), and LAI had a weak positive correlation under water-stressed conditions.

Latent heat flux was the dominant sink of the available energy during the wet season (specifically from early June to mid November), and H occurred dominantly during the rest of the year. The maximum daytime $\lambda E$ value was $337 \text{ Wm}^{-2}$ ($11.9 \text{ mmday}^{-1}$) observed in August and the peak daytime H was $268 \text{ Wm}^{-2}$ in April. The seasonal variation in Bowen ratio was clearly U-shaped and the annual B averaged 1. The variations in B, which were in close relationship with variations in soil moisture, LAI and net radiation with the later affected by cloud cover, have significant impacts on properties of the atmospheric boundary layer. Furthermore, the seasonal dynamics of bulk parameters, such as $g_c$, decoupling coefficient $\Omega$ and Priestley-Taylor parameter $\alpha$, was affected by variations in soil moisture conditions and saturation deficits. The annual means of $\Omega$ and $\alpha$, which were 0.5 and 0.8, respectively, indicated some stomatal control on ET relatively weaker than other studies. The diurnal pattern of $g_c$ and $\Omega$ was typical of the full-leaf period with decreasing values from the morning to the evening, indicating an increasing degree of coupling to the atmosphere as the day progressed. The canopy conductance responded to high vapor pressure deficit and restricted transpiration losses (stomatal closure) after an approximate value of 2 KPa.
Cumulative ET during the study year was 59% of the annual precipitation (1629 mm) received at the site. The results of this investigation for the subtropical grassland complemented with other studies, will lead to better understanding of biosphere-atmosphere energy interactions and accurate parameterizations of hydrologic models.
References


CHAPTER 5:
SUMMARY OF RESEARCH AND FUTURE IMPLICATIONS

5.1 Summary of Work

The existing models for downwelling longwave (LW) radiation employ different formulations for clear sky emissivity calculation. Even though both the temperature and water vapor affect the emissivity, the majority of the models use either a lump humidity parameter or only one effect due to existing correlation between the two parameters. These parameters are further affected by heterogeneous land use patterns and temporal changes in atmospheric circulation patterns. In Chapter 2, a model considering the nonlinear temperature and water vapor pressure effects superpositioned in one equation to account for the net impact on clear sky emissivity was investigated. Furthermore, the developed model is enhanced to become adaptable to different land use. Ground radiometer and meteorological data applicable in the subtropical climate of Saint Johns River Water Management District (SJRWMD), Florida during the spring season are utilized for the model development. The new model is tested against pyrgeometer data gathered above a crop in Bondville, IL, a rangeland in Ft. Peck, MT, and an urban setting in Deland, FL. When the new parameterization is validated using the 24hr average downwelling LW flux, it yields a Nash-Sutcliffe Efficiency (NSE) greater than 0.5, and normalized Mean Bias Errors (MBE) and Root Mean Square Errors (RMSE) of less than 5% and 7%, respectively with the latest being the smallest deviations with respect to existing formulations.
In *Chapter 3*, the modulation of evapotranspiration (ET) by several abiotic and biotic factors over a non-irrigated, ungrazed grass site in Central Florida was systematically investigated. The main focus was on the sensitivity of the water vapor flux to wetness variables, namely the volumetric soil water content (SWC) and the current precipitation index (CPI). Eddy correlation measurements were carried out at a flux tower adjacent to a weather station, during the course of 2004. The soil is composed of fine sands and it is mainly covered by bahia grass. Leaf Area Index (LAI) and daytime canopy surface conductance \( (g_c) \) were positively correlated to ET and reached maximum values of 3.6 \( m^2 m^{-2} \) and 11 \( mm s^{-1} \), respectively. Throughout the year the average ET was 5.5 \( mm day^{-1} \) (daytime hours), while about 59% of the total precipitation returned to the atmosphere as 24-hr ET. Decreases in SWC and daytime ET after rainfall were quite dynamic depending on the amount of the antecedent rain (exponential time coefficient of 6.1 days for a composite dry event). The daytime average ET was classified into two stages, first stage (energy-limited) and second stage (water-limited). The daytime average Priestley-Taylor (PT) coefficient varied from a low bound of 0.4 to a peak of 1.12, but most of the time was less than unity indicating that ET of this grass ecosystem was characterized by the limitations in water supply. Simple models for the PT factor were employed in terms of water availability, and the modeled results closely matched the eddy-covariance flux values on daily time scale during all moisture conditions.

*Chapter 4* complemented the work of Chapter 3 with the study of dynamics of daytime turbulent energy fluxes over the bahia grass site. The significant biophysical constraints on latent heat flux \( (\lambda E) \) were examined for the two water availability stages. The
seasonal variations in LAI, SWC and net radiation ($R_n$) were reflected in a strong seasonal pattern of the energy balance. $\lambda E$ was the main sink of the energy during DOY 152-321, accounting for approximately 50-80% of the monthly mean $R_n$. The annual maximum $\lambda E$ was 249 $W m^{-2}$ over August (monthly mean, daytime basis) with a peak midday value of 416 $W m^{-2}$ (monthly ensemble, 30-min basis). Over the year, more than 90% of $R_n$ was converted to turbulent fluxes, and the Bowen ratio had an average value of one. The daytime canopy surface conductance, decoupling coefficient ($\Omega$) and PT parameter $\alpha$ increased at the beginning of the wet season in association with an increase in SWC and a reduction in vapor pressure deficit, and then they declined towards the end of the rainy season. Over the diurnal course, $g_c$ and $\Omega$ were highest in the morning and lowest in the evening, suggesting stronger stomatal control of $\lambda E$ losses as the day progressed. The annual means for $\Omega$ and $\alpha$ were 0.5 and 0.8, respectively, with the first value indicating some stomatal control on transpiration, and the $\alpha$ value signifying water supply limitations on $\lambda E$. Indeed, the stomatal control was demonstrated with a reduction in $g_c$ and $\lambda E$ losses as the atmospheric humidity deficit increases (beyond 2 KPa).

5.2 Future Implications

Over the last decade, the study of the climate has been the topic of intense investigation. Several papers showed the limitations of the global climate models, including some theoretical treatments suggesting flaws in the models. For instance, there is an argument suggesting that higher greenhouse gas levels, which are on the focus of international energy-
regulatory campaigns, are the result of higher temperatures, rather than vice versa. Indeed temperature and greenhouse gases can be mutually reinforcing and interact in a dynamic atmosphere (Mulkey, 2007). In addition, the limitations of the global climate models of not predicting the fine regional scale and the local climate variability have been highlighted. The land use change, although not a primary climate-forcing factor is relatively so far quite unexplored as a cause for local climate change and variability of water resource systems (e.g. Schneider and Eugster, 2005). Human-induced land use change can have big impact on both the downwelling and upwelling components of LW and can dramatically influence heat convection, wind flow, precipitation, soil wetness and ET on a local and regional basis.

Future studies need to address the applicability of the land-use adaptable clear sky emissivity model, described in Chapter 2, in more diverse climatic regions than the present work. Further future efforts may incorporate the vegetative and impervious cover into this formulation. A possible extension of the atmospheric emissivity model for annual ET estimation over various land covers should consider to:

1. Collect data of radiation meteorology from existing or proposed meteorological and radiation sensors in a study region, namely surface temperature ($T_s$), air temperature ($T_o$), water vapor pressure ($e_o$), upwelling LW radiation ($LW_u$), $LW_d$ and albedo. Complement with ET database from eddy correlation stations. Correlate the collected data with land use and soil properties;
2. Establish trends of land use versus $LW_u$ and $LW_d$ radiation;
3. Develop empirical land use-enhanced LW radiation models similar to the model proposed by Rizou and Nnadi (2007);
4. Characterize the local and regional warming effect based on the agglomerated land surface material as well as local heterogeneities;
Test the sensitivity of local warming at varying levels of measured greenhouse gases. Then compare this effect with the land use effect described previously in step (4); (6) Adapt the simple PT ET model to vegetation type. There are limited ET parameterizations based on the PT formulation that consider land cover in literature (e.g. de Bruin and Stricker, 2000); and (7) Implement land use management alternatives to minimize the effects of local warming and the loss of evaporative water.

Regarding the material presented in Chapters 3 and 4 the following suggestions can be made. The parameterizations of the PT parameter considered in this work include the individual functions of SWC and CPI. In fact, CPI is an indicator of the precipitation history (both rain amount and frequency) of a catchment and can readily be available from extensive rainfall networks. On the other hand, the widely used SWC is a point, “current” estimate of water availability, and its scaling to larger grids is rather problematic (Kurc and Small, 2004). In addition, the amount of SWC extracted by the plants depends on the root depth of vegetation and soil properties. Thus the soil moisture dynamics should be considered when studying the surface flux components. A possibility for future research aiming to improve ET estimation would be the development of a more complex model incorporating both precipitation history and soil moisture dynamics (refer to the stochastic model by Laio et al., 2001).

Additional knowledge on the carbon exchange, productivity and physiological functioning of this grassland would have provided insight to the carbon cycle, which is coupled to the water cycle, and led to comprehensive understanding of the atmosphere-biosphere interactions. Finally, a promising idea would be to compare the surface energy components across different vegetation types in regions with similar climatic and soil
variables. This idea must be taken into serious consideration, since dynamic (in space and time) shifts in vegetation and land use occur rather frequently. For instance, Beringer et al. (2005) suggested that transitions in vegetation caused by climate warming alter the energy balance and will result in a positive feedback to further warming in the Arctic.
References


APPENDIX A:
BACKGROUND INFORMATION OF NEW LW_d MODEL
The range of the weather conditions, and measured and modeled LW_d flux over the calibration and validation sites is given in Table A.1.

Table A.1: Range of weather conditions and radiation at the study sites during clear days.

<table>
<thead>
<tr>
<th>Location, year</th>
<th>e_o (hPa)</th>
<th>T_o (K)</th>
<th>LW_r-measured (W/m²)</th>
<th>LW_r-modeled (W/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deland, 2004</td>
<td>Min 8.7</td>
<td>288.3</td>
<td>297.5</td>
<td>298.8</td>
</tr>
<tr>
<td></td>
<td>Max 18.1</td>
<td>297.5</td>
<td>373.4</td>
<td>369.1</td>
</tr>
<tr>
<td>Jarboe, 2004</td>
<td>Min 6.6</td>
<td>284.5</td>
<td>271.4</td>
<td>266.1</td>
</tr>
<tr>
<td></td>
<td>Max 21.4</td>
<td>297.1</td>
<td>371.9</td>
<td>360.3</td>
</tr>
<tr>
<td>Lindsey, 2004</td>
<td>Min 9.1</td>
<td>288.3</td>
<td>283.7</td>
<td>289.1</td>
</tr>
<tr>
<td></td>
<td>Max 20.1</td>
<td>296.4</td>
<td>356.6</td>
<td>354.8</td>
</tr>
<tr>
<td>Orange, 2004</td>
<td>Min 6.1</td>
<td>285.3</td>
<td>262.4</td>
<td>262.0</td>
</tr>
<tr>
<td></td>
<td>Max 17.5</td>
<td>297.6</td>
<td>350.8</td>
<td>355.8</td>
</tr>
<tr>
<td>Ocklawaha, 2004</td>
<td>Min 8.7</td>
<td>287.3</td>
<td>266.5</td>
<td>267.4</td>
</tr>
<tr>
<td></td>
<td>Max 22.9</td>
<td>299.4</td>
<td>376.3</td>
<td>364.2</td>
</tr>
<tr>
<td>Deland, 2005</td>
<td>Min 6.9</td>
<td>279.6</td>
<td>240.3</td>
<td>257.7</td>
</tr>
<tr>
<td></td>
<td>Max 23.4</td>
<td>299.5</td>
<td>374.3</td>
<td>383.7</td>
</tr>
<tr>
<td>Bondville-IL, 2001</td>
<td>Min 8.1</td>
<td>285.3</td>
<td>276.9</td>
<td>269.5</td>
</tr>
<tr>
<td></td>
<td>Max 24.2</td>
<td>298.0</td>
<td>354.5</td>
<td>371.0</td>
</tr>
<tr>
<td>Ft. Peck-MT, 2001</td>
<td>Min 4.7</td>
<td>279.4</td>
<td>245.8</td>
<td>229.3</td>
</tr>
<tr>
<td></td>
<td>Max 26.4</td>
<td>301.7</td>
<td>379.1</td>
<td>382.8</td>
</tr>
</tbody>
</table>

The iso-emissivity lines by using Equation (2.5) and land use factors α_{LU} for the calibration sites are presented below. For Lindsey Citrus refer to Figure 2.4.
Figure A.1: The iso-emissivity lines for a) Deland and b) Jarboe Park by using the new model along with overlaid actual emissivity data. The shaded area indicates above saturation water vapor pressure.
Figure A.2: The iso-emissivity lines for c) Orange Creek and d) Ocklawaha Prairie. Same as Figure A.1.
APPENDIX B:
LAND USE OF ALL CNR1 STATIONS
Table B.1: Land use and location of CNR1 stations.

<table>
<thead>
<tr>
<th>Site Name</th>
<th>Location</th>
<th>Longitude/Latitude (°)</th>
<th>Picture</th>
<th>Primary land use</th>
<th>Secondary land use</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jarboe Park</td>
<td>Edge of a city park</td>
<td>30.32/-81.40</td>
<td></td>
<td>Residential</td>
<td>Irrigated grass, paved road, sidewalk, pond</td>
</tr>
<tr>
<td>Hastings IFAS</td>
<td>Row crop potato setting in UF research facility</td>
<td>29.69/-81.45</td>
<td></td>
<td>Agriculture</td>
<td>Pine flatwoods</td>
</tr>
<tr>
<td>Orange Creek Restoration</td>
<td>District land</td>
<td>29.48/-82.07</td>
<td></td>
<td>Pasture Land (baha grass)</td>
<td>Rangeland/pastureland, mixed hardwood and pine</td>
</tr>
<tr>
<td>Denver Road</td>
<td>District land on Denver Rd</td>
<td>29.38/-81.55</td>
<td></td>
<td>Silviculture</td>
<td>Pine forest, pond, and agriculture</td>
</tr>
<tr>
<td>Ocklawaha Frame</td>
<td>District land</td>
<td>29.10/-81.91</td>
<td></td>
<td>Agriculture (cattail, sawgrass, and other aquatic vegetation)</td>
<td>Wetland, rangeland/pastureland and mixed forest</td>
</tr>
<tr>
<td>Site Name</td>
<td>Location</td>
<td>Longitude/Latitude (°)</td>
<td>Picture</td>
<td>Primary land use</td>
<td>Secondary land use</td>
</tr>
<tr>
<td>--------------------</td>
<td>---------------------------</td>
<td>------------------------</td>
<td>---------</td>
<td>--------------------------------------------</td>
<td>----------------------------------------</td>
</tr>
<tr>
<td>DeLand STP</td>
<td>Wastewater treatment plant</td>
<td>29.01/-81.30</td>
<td></td>
<td>Agriculture (grass, a few trees and shrubs)</td>
<td>Residential and mixed forest</td>
</tr>
<tr>
<td>Lake Apopka Center</td>
<td>Near the center of Lake Apopka</td>
<td>28.63/-81.63</td>
<td></td>
<td>Open water</td>
<td>Open water, wetland, and urban</td>
</tr>
<tr>
<td>Lake Washington</td>
<td>Eastern shore of Lake Washington</td>
<td>28.15/-80.73</td>
<td></td>
<td>Open water and marsh</td>
<td>Open water, wetland, and urban</td>
</tr>
<tr>
<td>Bull Creek</td>
<td>Distact land</td>
<td>28.09/-80.96</td>
<td></td>
<td>Agriculture (balsa grass and palmetto)</td>
<td>Rangeland and pine flatwoods</td>
</tr>
<tr>
<td>Mulberry Marsh</td>
<td>Distact land</td>
<td>27.91/-80.78</td>
<td></td>
<td>Agriculture (cattail and some willow)</td>
<td>Wetland</td>
</tr>
<tr>
<td>Lindsey Citrus</td>
<td>Citrus grove</td>
<td>27.59/-80.60</td>
<td></td>
<td>Citrus</td>
<td>Citrus</td>
</tr>
</tbody>
</table>