Evapotranspiration from Successional Vegetation in a Deforested Area of the Lake Wales Ridge, Florida

By D. M. Sumner

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CONTENTS

FIGURES

TABLES

CONVERSION FACTORS, VERTICAL DATUM, ABBREVIATIONS, AND SYMBOLS

Multiply	By	To obtain
	Length	
millimeter (mm)	0.03937	inch
centimeter (cm)	0.3937	inch
meter (m)	3.281	foot
	Area	
hectare	2.471	acre
	Flux	
millimeter per day (mm/d)	0.03937	inch per day
	Energy	
joule (J)	0.2388	calorie
	Energy flux density	
watt (W/m^2)	0.001433	calorie per square centimeter per minute
	Pressure	
kilopascal (kPa)	0.2953	inches of mercury
	0.1450	pound per square inch
	10.0	millibar
	Photosynthetically-active radiation	
micromoles per square meter per second ($\mu moles/(m^2 \cdot s)$)	6.02×10^{17}	photons per square meter per second

Sea level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Altitude, as used in this report, refers to distance above or below sea level.

Temperature is given in degrees Celsius (^oC), which can be converted to degrees Fahrenheit (^oF) by the following equation:

$$^oF = 1.8(^oC) + 32.$$

Additional abbreviations

CSI = Campbell Scientific, Inc.
ECEBBR = Eddy correlation energy-balance Bowen ratio method
ECEBR = Eddy correlation energy-balance residual method
PAR = Photosynthetically-active radiation
REBS = Radiation and Energy Balance Systems, Inc.
RMY = R.M. Young, Inc.
TDR = Time domain reflectometry
TE = Texas Electronics, Inc.
USGS = United States Geological Survey

List of Symbols

Roman

b	Depth to heat flux plate, in m
C_1-C_8	Site-specific empirical parameters within evapotranspiration models
C_p	Specific heat of air, in J/(g·^{o}C)
$\overline{C}_s$	Mean specific heat capacity of soil above heat flux plate, in J/(m^3·^{o}C)
$\overline{C}_{sd}$	Mean specific heat capacity of dry soil above heat flux plate, in J/(kg·^{o}C)
C_w	Specific heat capacity of water, in J/(kg·^{o}C)
d	Momentum displacement height of vegetation, in m
D	Vapor pressure deficit, in kPa
e	Vapor pressure, in kPa
e_s	Saturation vapor pressure, in kPa
E	Evapotranspiration rate, in g/(m^2·s)
F	Factor used in krypton hygrometer correction that accounts for molecular weights of air and atmospheric abundance of oxygen, equal to 0.229 g·^{o}C/J
G	Soil heat flux at land surface, in W/m^2
h	Average canopy height, in m
H	Sensible heat flux, in W/m^2
k	von Karman’s constant, equal to 0.4, dimensionless
κ_s	Attenuation coefficient of light for the canopy, dimensionless
K_o	Extinction coefficient of hygrometer for oxygen, in m^3/(g·cm)
K_w	Extinction coefficient of hygrometer for water, in m^3/(g·cm)
L	Two-sided leaf area index, in m^2/m^2
L_s	Sunlit leaf area per unit area of land, in m^2/m^2
L_v	Green leaf fraction visible to observer, in m^2/m^2
r_c	Canopy resistance, in s/m
r_h	Aerodynamic resistance, in s/m
R_n	Net radiation, in W/m^2
R_{PAR}	Photosynthetically-active radiation, in µmoles/(m^2·s)
S	Rate of change of stored energy in soil above heat flux plate, in W/m^2
t	Time, in Julian days

T_a	Air temperature, in ^oC
T_i	Mean temperature at time step "i" of soil above heat flux plate, in ^oC
u	Horizontal wind speed, in m/s
u^*	Friction velocity, in m/s
w	Vertical wind speed, in m/s
$\overline{w}_m$	Mean gravimetric water content above heat flux plate, in kg/kg
z	Height of sensors above land surface, in m
z_o	Roughness length of canopy for momentum, in m
z_h	Roughness length of canopy for sensible heat, in m

Greek

α	Priestley-Taylor coefficient, dimensionless
Δ	Slope of the saturation vapor pressure curve, in kPa/^oC
Δt	Time interval between soil-temperature measurements, in s
γ	Psychrometric "constant", in kPa/^oC
λ	Latent heat of vaporization, in J/g
λE	Latent heat flux, in W/m^2
$\overline{\theta}$	Mean volumetric moisture content in rooting zone, in cm^3/cm^3
ρ	Air density, in g/m^3
$\overline{\rho}_b$	Mean soil bulk density above heat flux plate, in kg/m^3
ρ_v	Vapor density, in g/m^3
ω	Angular frequency of Earth's revolution around Sun, in d^{-1}
ψ_h	Stability profile corrector for sensible heat, dimensionless
ψ_m	Stability profile corrector for momentum, dimensionless
ζ	Atmospheric stability, dimensionless

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ABSTRACT

The suitability of three evapotranspiration models (Penman-Monteith, Penman, and a modified Priestley-Taylor) was evaluated at a site of successional vegetation in a deforested area of the Lake Wales Ridge, Florida. Eddy correlation measurements of evapotranspiration made during 22 approximately 1-day periods at a temporal resolution of 20 minutes from September 1993 to August 1994 were used to calibrate the evapotranspiration models. Three variants of the eddy correlation method that ascribe measurement error to three different sources were considered in the analysis. The Penman-Monteith and modified Priestley-Taylor models were successful in approximating measured 20-minute values of evapotranspiration ($r^2 \geq 0.918$). The most successful approaches were the modified Priestley-Taylor model ($r^2 = 0.972$) and a nontraditional and simplified form of the Penman-Monteith model ($r^2 = 0.967$). The Penman approach was unsuccessful as a predictor of evapotranspiration.

The evapotranspiration models were used to estimate evapotranspiration between measurements. When evapotranspiration values measured with a Bowen ratio variant of the eddy correlation method were used for model calibration, estimated daily evapotranspiration rates varied seasonally ranging from 0.2 millimeters per day (0.008 inch per day) in late December 1993 to 5 millimeter per day (0.2 inch per day) in mid-July 1994. Annual evapotranspiration (September 15, 1993, to September 15, 1994) was estimated to be about 680 millimeters (27 inches). Evapotranspiration models calibrated to the standard eddy correlation method and to an energy-balance residual variant provided estimates of annual evapotranspiration that were about 10 percent lower and higher, respectively. These data indicate that of the 1,320 millimeters (52 inches) of precipitation during the 1-year period, about 570 to 700 millimeters (22 to 28 inches) recharged the surficial aquifer. Evapotranspiration at this study site probably defines the lower limit of evapotranspiration from vegetated surfaces in central Florida because of the shallow-rooted plants, rapidly-drained soils, and relatively deep water table.

INTRODUCTION

The importance of evapotranspiration in the hydrologic cycle has long been recognized. In Florida, evapotranspiration is second only to precipitation in magnitude within the hydrologic budget. Of the approximately 1,300 mm of mean annual rainfall in central Florida, 760 to 1,220 mm have been estimated to return to the atmosphere as evapotranspiration (Tibbals, 1990). Despite the importance of evapotranspiration in the hydrologic cycle, the magnitude, seasonal and diurnal distributions, and relation to environmental variables of evapotranspiration remain relatively unknown. This uncertainty in evapotranspiration is particularly apparent in nonagricultural vegetation, such as successional vegetation in deforested areas of the Lake Wales Ridge. This successional vegetation is unique in that vegetative cover can be sparse, the water table is relatively deep, and the sandy soils are rapidly drained; these characteristics minimize evapotranspiration from this vegetation, which in turn, maximizes recharge to the underlying aquifer. Accurate estimates of evapotranspiration are fundamental to improving estimates of aquifer recharge in areas of successional vegetation of the Lake Wales Ridge.

The eddy correlation method has been used successfully to directly measure evapotranspiration in Florida (Bidlake and others (1993) and Knowles (1996)). This micrometeorological method offers several advantages to alternative water-budget approaches (lysimeter or regional water budget) by providing more areal integration and less site disruption than lysimeters, eliminating the need to estimate other terms of a water budget (precipitation, deep percolation, and runoff), and allowing relatively fine temporal resolution (less than 1 hour).

Evapotranspiration can be estimated from evapotranspiration models. These models also provide insight into the relative importance of individual environmental variables in the evapotranspiration process. The Penman-Monteith equation (Monteith, 1965) is a physics-based evapotranspiration model that is based on a coupling of energy-budget and aerodynamic considerations. Although the Penman-Monteith equation is often a theoretically sound model, the equation is computationally cumbersome. Simpler models such as the Penman (Penman, 1948) and Priestley-Taylor (Priestley and Taylor, 1972) equations offer computational ease, but the error introduced by these models must be evaluated to determine model suitability to a particular site.

The U.S. Geological Survey (USGS), in cooperation with the St. Johns River Water Management District and the South Florida Water Management District, began a 4-year study in 1992 to estimate the annual pattern of evapotranspiration at a site containing successional vegetation; to evaluate the implications of these results on estimation of aquifer recharge; and to evaluate the suitability of several evapotranspiration models. This analysis can provide guidance in the estimation of evapotranspiration and aquifer recharge and in the selection of a reliable and efficient evapotranspiration model at other sites with similar environmental characteristics.

Purpose and Scope

This report presents (1) estimates of evapotranspiration during a 1-year period at a site of successional vegetation in a deforested area of the Lake Wales Ridge, Florida; and (2) results of an evaluation of the suitability of three evapotranspiration models—Penman-Monteith, Penman, and a modified Priestley-Taylor—in evapotranspiration simulation. Evapotranspiration measurements were made approximately every 2 weeks from September 1993 to August 1994 at a field site in Orange County, Florida, using eddy correlation instrumentation. Evapotranspiration models were used to estimate evapotranspiration between direct evapotranspiration measurements and to make a comparative analysis of the merits of each model. Estimated annual evapotranspiration and measured precipitation were used to estimate surficial aquifer recharge.

Description of the Study Site

The study site is a field of mostly herbaceous, successional vegetation that typically grows in cleared areas of central Florida. The site location is in west Orange County, Florida (fig. 1). The site (about 450 m by 600 m, or about 27 ha) is bounded by citrus groves, oak trees, cultivated pine trees, and infiltration basins for disposal of reclaimed water. The site contains rolling topography; land surface altitude ranges from about 41 to 46 m above sea level. The site is part of the Lake Wales Ridge physiographic region (White, 1970, plate 1-B), which is characterized by sandy, rapidly drained soils, and hilly, karstic topography.

Vegetation at the site predominantly is natalgrass (*Rhynchelytrm repens*), dog fennel (*Eupatorium* sp.), dwarf horseweed (*Conyza canadensis*), and ragweed (*Ambrosia* sp.). The rooting depth for any of the vegetation rarely exceeds 30 cm. Average canopy height (considering both living plants and dead residual vegetation from the previous growing season) varies seasonally, ranging from about 25 cm in spring to 75 cm in summer (fig. 2). Green foliage density follows a seasonal pattern, reaching maximum values during the summer wet period and minimum values during the winter dry period (fig. 3).

The sandy, rapidly drained soils of this site—Candler series hyperthermic, uncoated Typic Quartzipsamments (Doolittle and Schellentrager, 1989)—allow the shallow-rooted plants to become water stressed during a dry period. Moisture content in the rooting zone of the soil can be as low as 0.02 cm^3/cm^3 and drainage of the soil profile following a rainfall event is relatively rapid (fig. 4). The water table generally is more than a meter below land

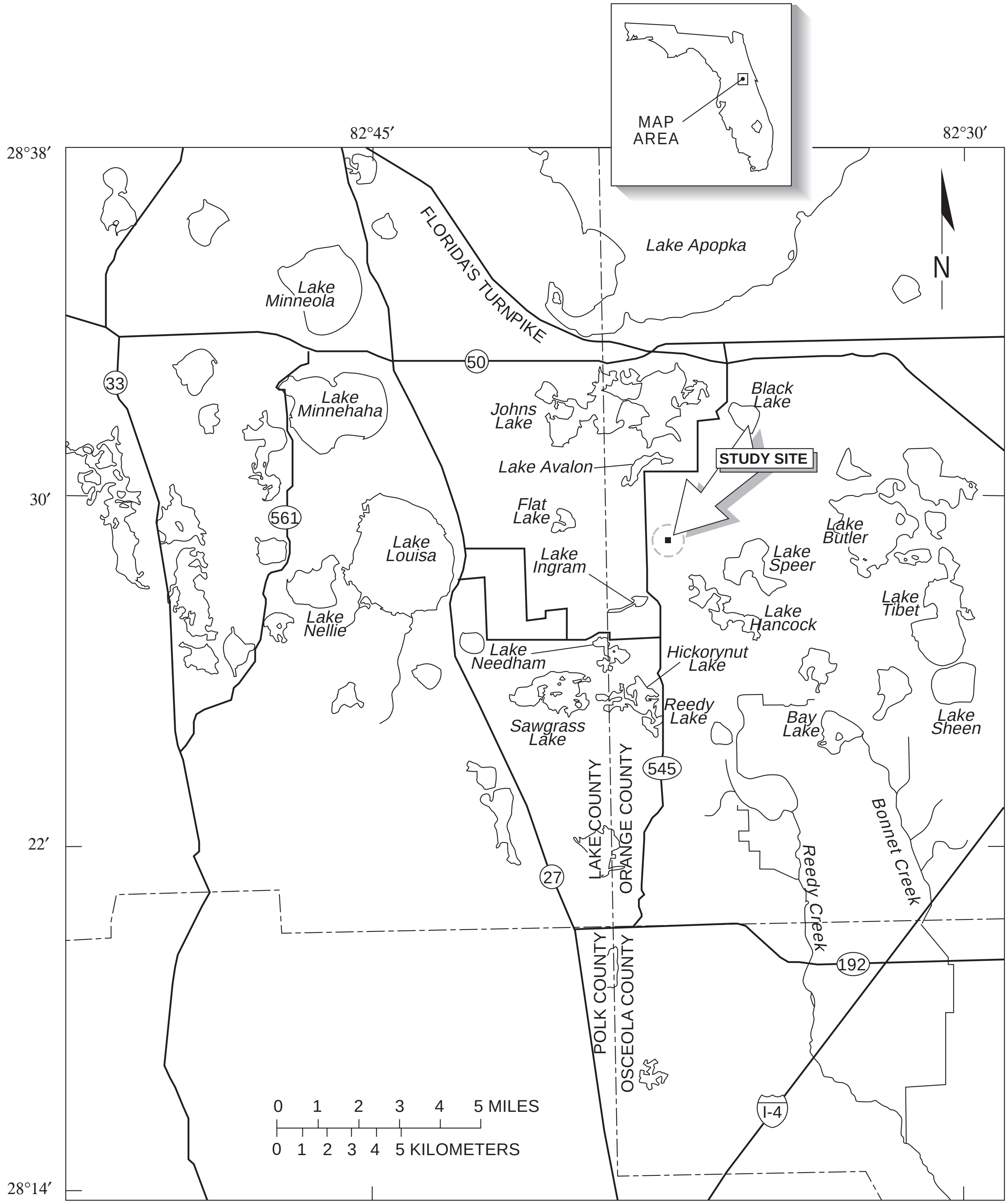


Figure 1. Location of study site in Orange County, Florida.

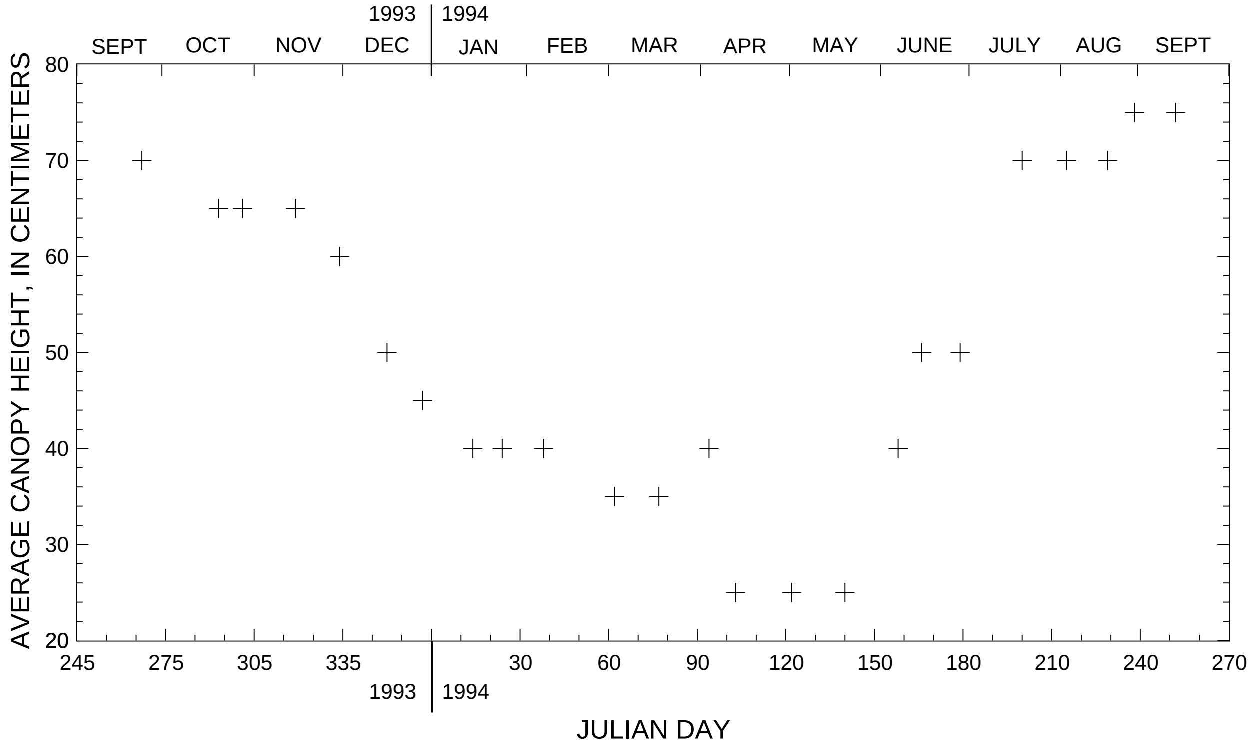


Figure 2. Average canopy height at study from September 15, 1993, to August 28, 1994.

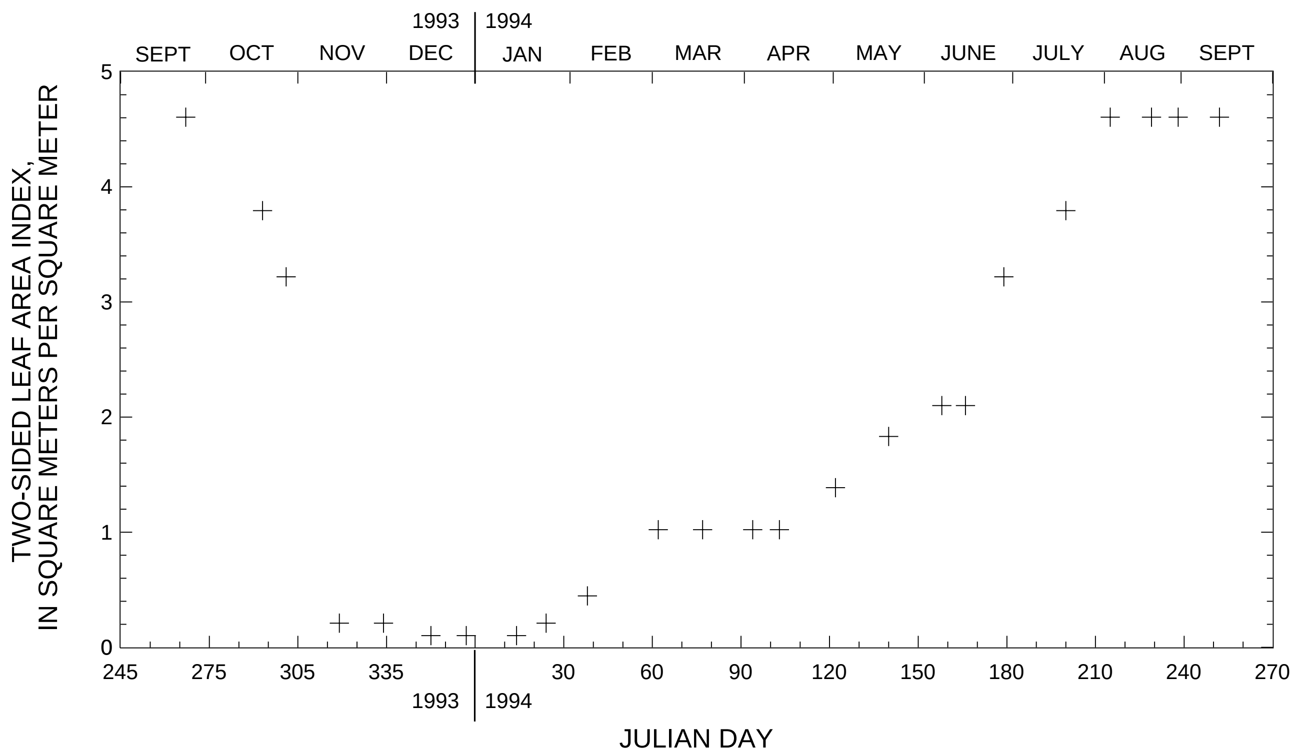


Figure 3. Two-sided leaf area index at study site from September 15, 1993, to August 28, 1994. Note: Leaf area index is defined as the two-sided, green leaf area per unit area of land surface.

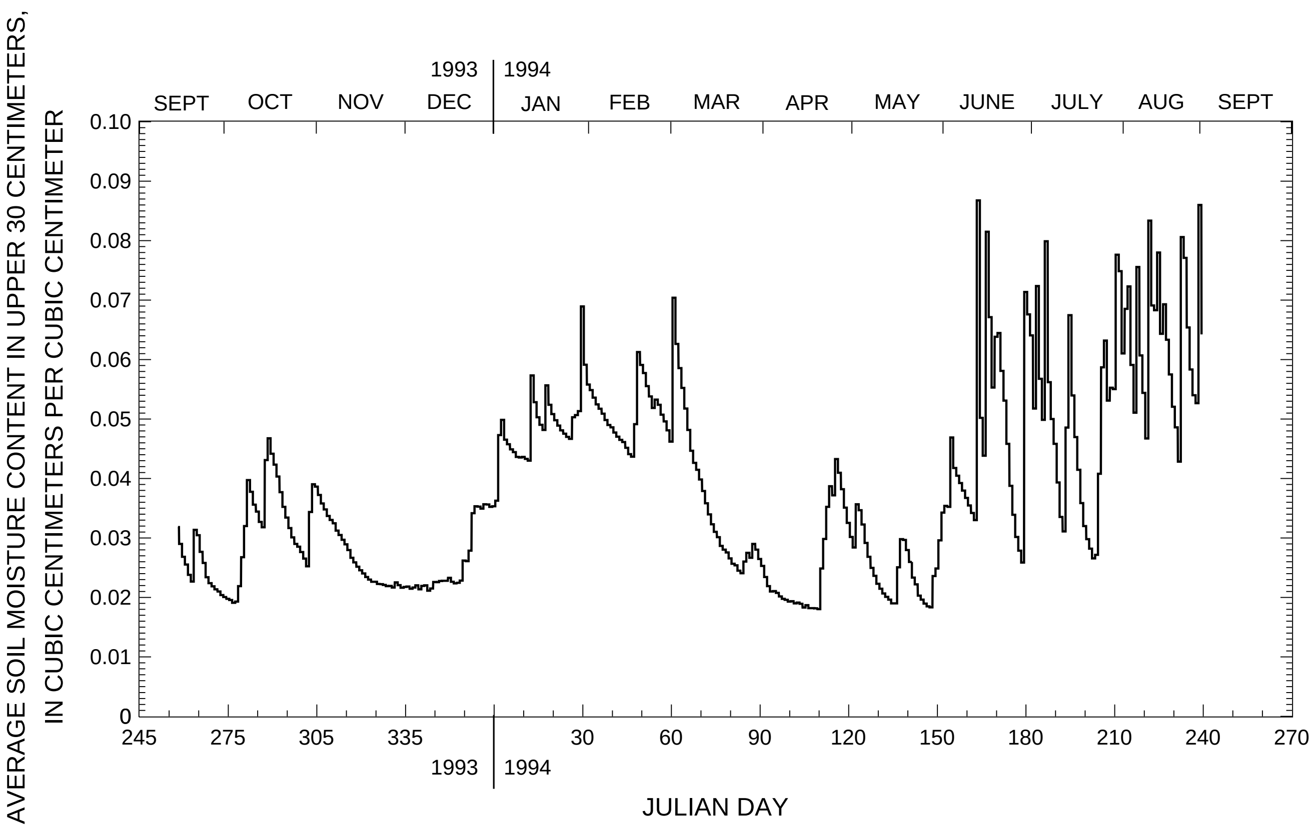


Figure 4. Average soil moisture content in upper 30 centimeters of soil profile at study site from September 15, 1993, to August 28, 1994.

surface throughout the site. In the upland areas of the site, water-table depth can be more than 5 m. Fluctuations in the water table exhibit a strong seasonal pattern in response to the seasonal distribution of rainfall (figs. 5 and 6).

The climate of central Florida is humid subtropical and is characterized by a warm, wet season (June-September) and a mild, relatively dry season (October-May). More than 50 percent of the annual rainfall generally occurs during the wet season (figs. 6 and 7) when diurnal thunderstorm activity due to differential heating of the land and ocean is common. During the dry season, precipitation commonly is associated with frontal systems. Mean air temperature at the site is about 22 ^{o}C, ranging from occasional winter temperatures below 0 ^{o}C to summer temperatures approaching 35 ^{o}C (fig. 8). Diurnal temperature variations at the site average about 12 ^{o}C. Strong seasonal and diurnal patterns also are exhibited by photosynthetically active radiation and the vapor-pressure deficit of the air (figs. 9, 10, and 11). Diurnal solar heating also leads to a diurnal pattern in wind speed and turbulent mixing (high during day and low at night).

METHODS FOR MEASUREMENT AND ESTIMATION OF EVAPOTRANSPIRATION

Evapotranspiration was measured at the study site using the eddy correlation method in a manner similar to Bidlake and others (1993). Eddy correlation instrumentation was deployed for 22 measurement events from September 15, 1993, to August 27, 1994. The length of each measurement event was about 1 day, with 20-min resolution. A station also was installed 15 m from the eddy correlation instrumentation to continuously measure meteorological data needed for evapotranspiration modeling and to provide ancillary data for the eddy correlation analysis. Measured values of evapotranspiration were used to calibrate evapotranspiration models (Penman-Monteith and modified Priestley-Taylor). Evapotranspiration was

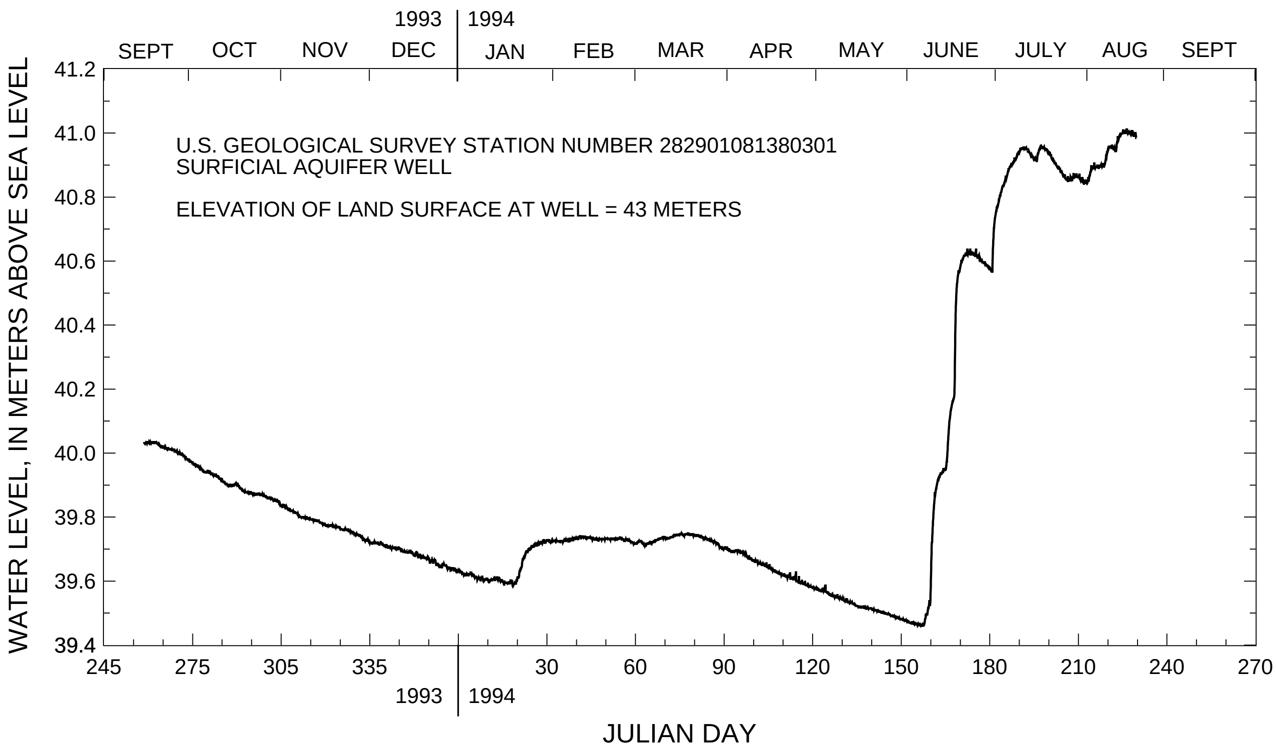


Figure 5. Water level in surficial aquifer at study site from September 15, 1993, to August 28, 1994.

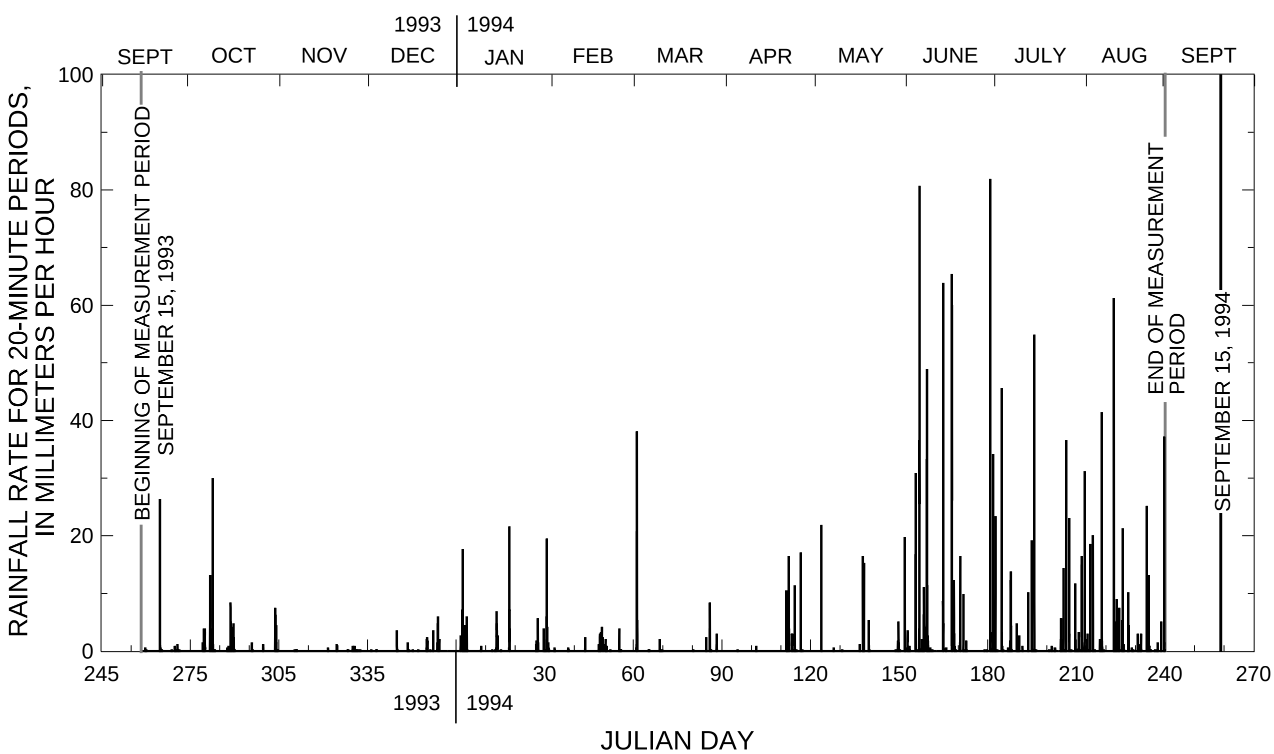


Figure 6. Rainfall rate at 20-minute resolution at study site from September 15, 1993, to August 28, 1994.

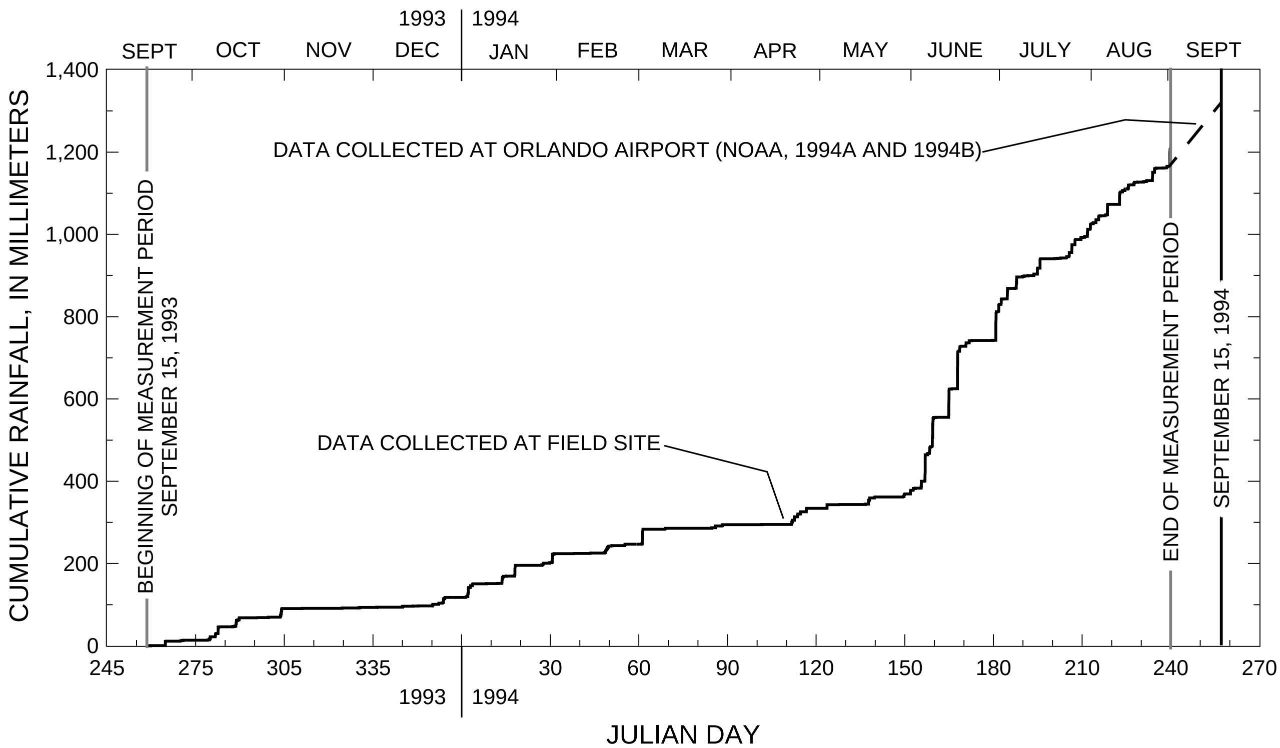


Figure 7. Cumulative precipitation at study site from September 15, 1993, to September 15, 1994.

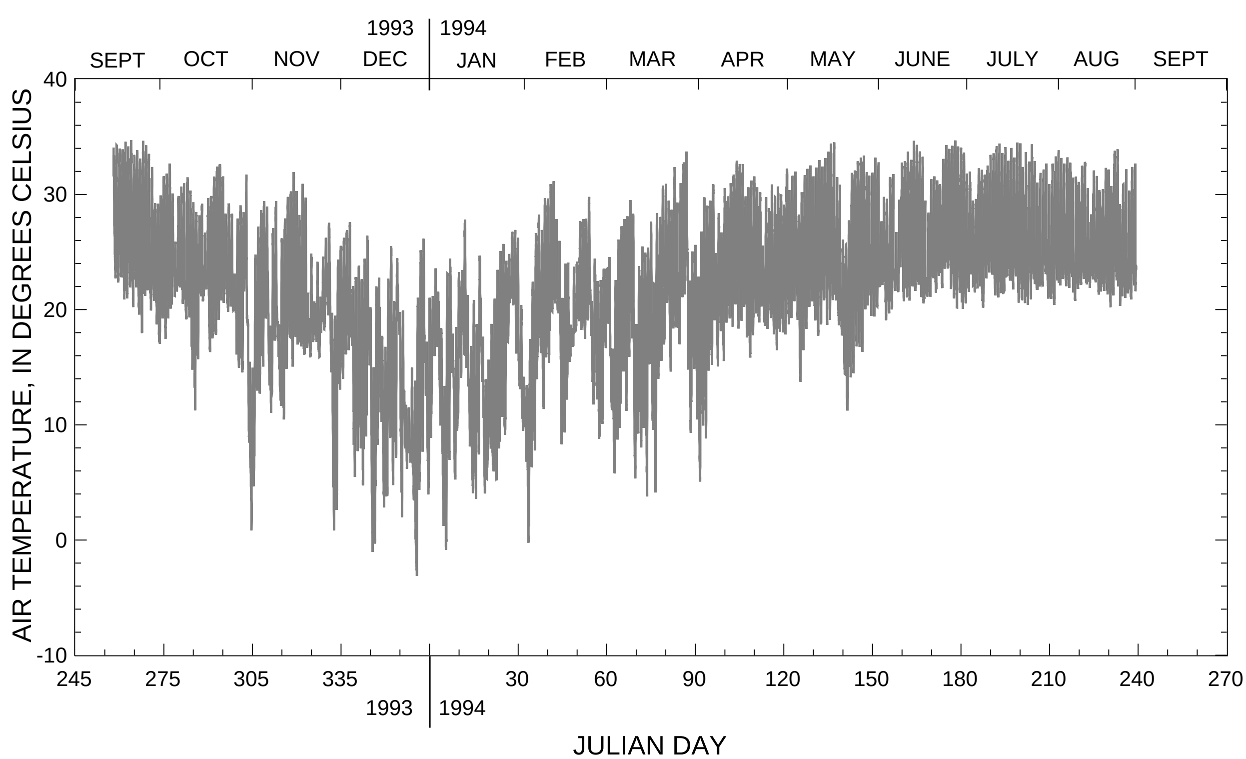


Figure 8. Air temperature at study site from September 15, 1993, to August 28, 1994.

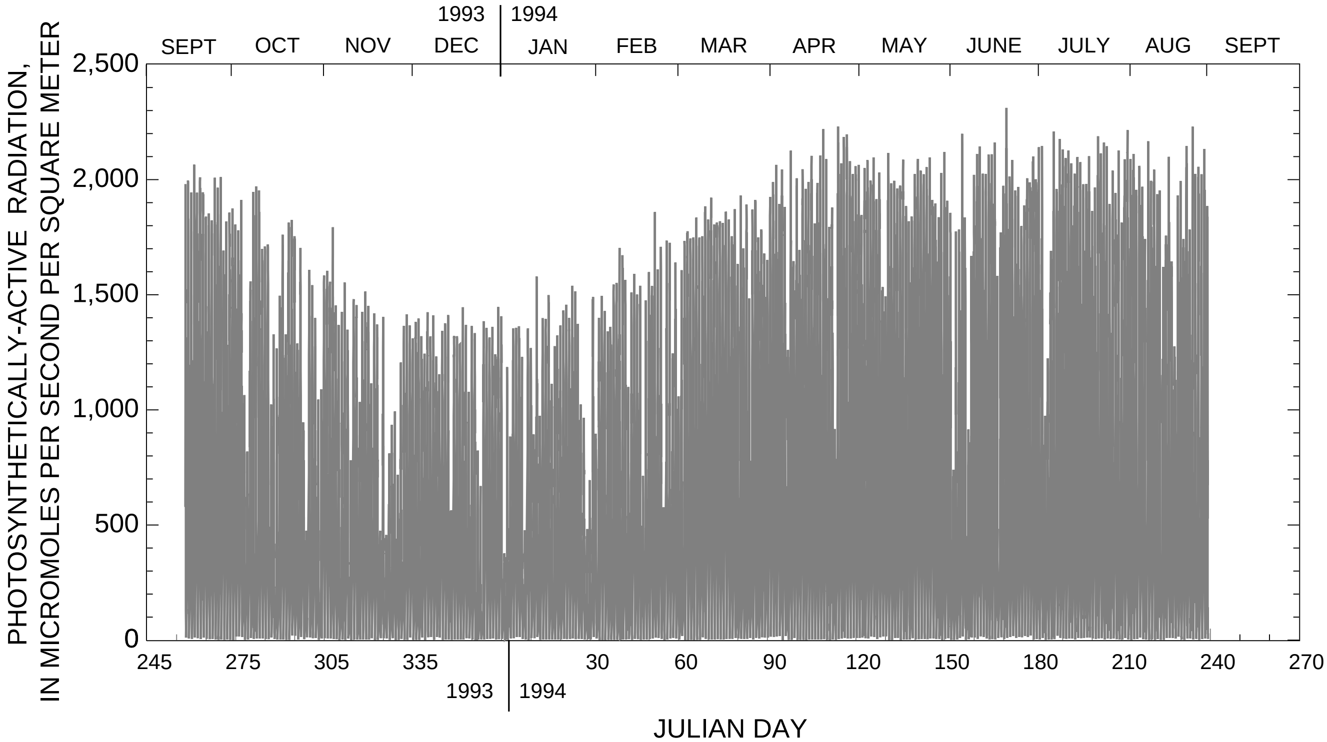


Figure 9. Photosynthetically active radiation at study site from September 15, 1993, to August 28, 1994.

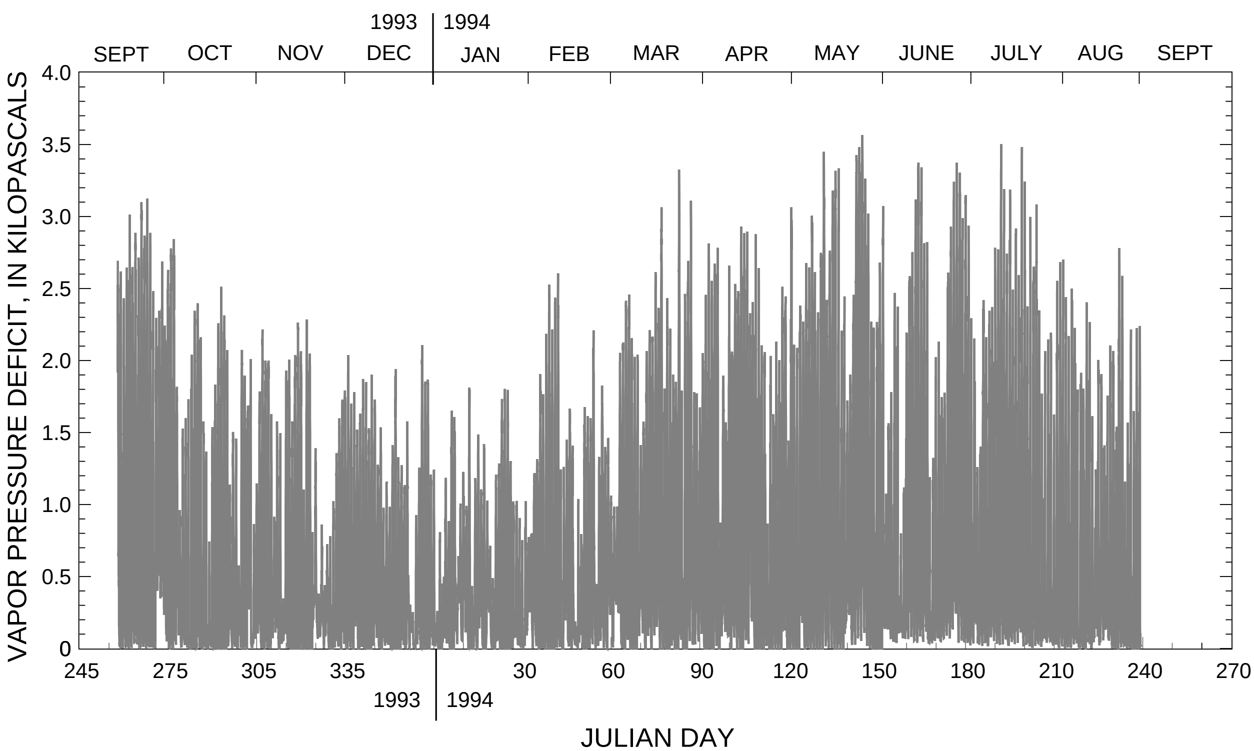


Figure 10. Vapor-pressure deficit at study site from September 15, 1993, to August 28, 1994.

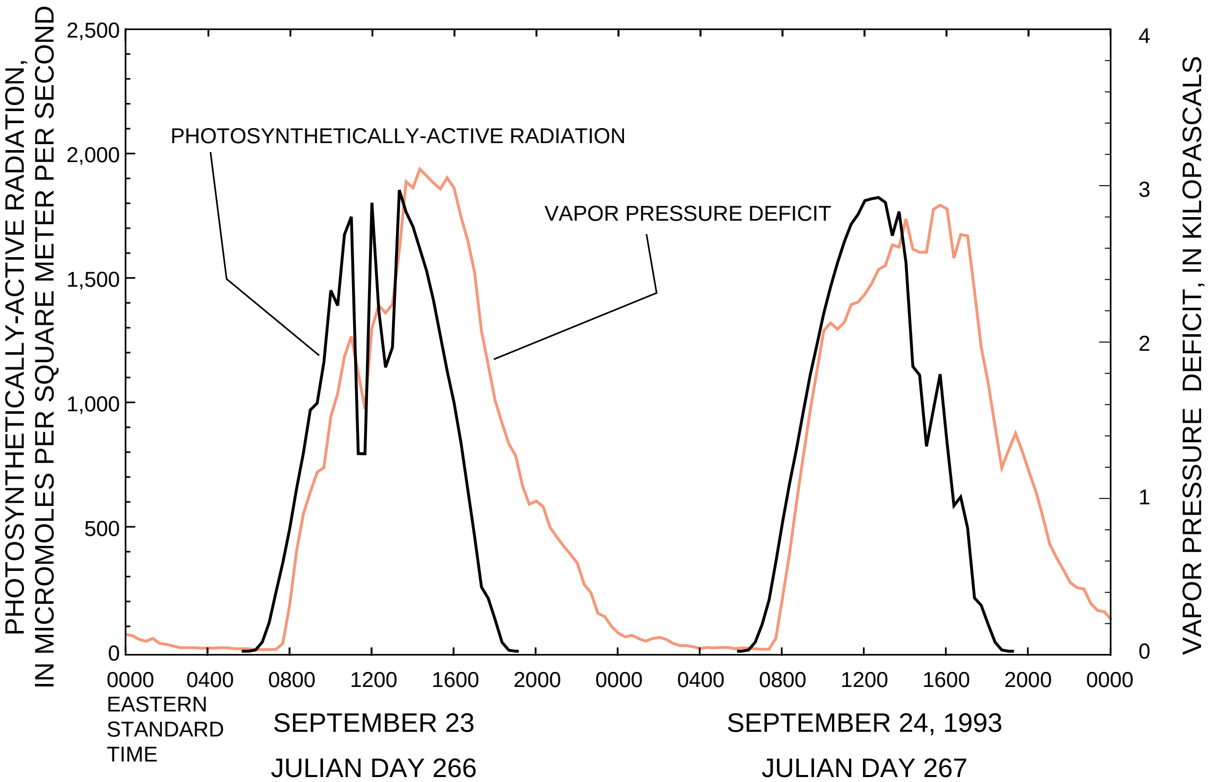


Figure 11. Diurnal variation in photosynthetically active radiation and vapor pressure deficit at study site on September 23-24, 1993.

estimated during the unmeasured periods between events (and at times of identifiable and substantial measurement error during the measured events) using the calibrated evapotranspiration models. Instrumentation used in the study is detailed in table 1.

Meteorological Station

Data collected at the meteorological station included air temperature, relative humidity, net radiation, wind speed, photosynthetically active radiation (PAR), soil heat flux, soil temperature, and soil moisture content. These data, except for soil moisture, were monitored by a datalogger at 15 s intervals and the resulting 20-min means were stored. Soil moisture was measured and stored hourly.

Air temperature and relative humidity were measured using a probe positioned about 1.7 m above land surface. Net radiation (R_n) was monitored with a net radiometer deployed at a height of about 1.6 m above land surface. Wind-speed measurements were made at a height of about 1.8 m with an anemometer and were extrapolated to the height of the net radiometer (assuming logarithmic wind speed profile) and used to correct for the variations in apparent net radiation caused by the convective effects of wind (empirical correction from C. Fritschen, REBS, oral commun., 1995). A quantum sensor was deployed to measure PAR.

Soil heat flux at land surface was estimated based on measured values of subsurface soil heat flux, soil temperature, and soil moisture. A trench was dug about 8 m from the meteorological station and probes for soil heat flux, temperature, and moisture measurement were installed in the undisturbed trench sidewall. The trench was backfilled with native soil 2 months before the initiation of data collection at the site. Instrumentation consisted of a soil heat-flux plate at a depth of 8 cm, averaging soil thermocouple probes at depths of 2 and 6 cm, and Time Domain Reflectometry (TDR) probes at depths of 4, 10, 18, and 30 cm.

Table 1. Study instrumentation

[CSI, Campbell Scientific, Inc.; REBS, Radiation and Energy Balance Systems, Inc.; RMY, R. M. Young, Inc.; TE, Texas Electronics, Inc.]

Type of measurement	Instrument
Air temperature/relative humidity	CSI Model HMP35C temperature and relative humidity probe
Net radiation	REBS Model Q-6 net radiometer
Wind speed/direction	RMY Model 05305-5 Wind Monitor-AQ
Photosynthetically-active radiation	LI-COR, Inc. LI90SB quantum sensor
Soil heat flux	REBS Model HFT-1 heat flux plate
Soil temperature	REBS Model TCAV thermocouple probe
Soil moisture	CSI time domain reflectometry system with Tektronix, Inc. 1502B cable tester
Precipitation	TE Model 525 tipping-bucket rain gage with L Model 260-952 windshield
Water level in well	Druck, Inc. Model PDCR950 pressure transducer
Evapotranspiration	CSI eddy correlation system with Model CA-27 sonic anemometer, Model KH20 krypton hygrometer, and fine-wire chromel-constantan thermocouple
Datalogging	CSI 21X datalogger

Two TDR probes were placed at each depth about 2 m apart. Each TDR probe consisted of two 30-cm stainless steel rods and a twin-lead cable. Twin-lead cable was connected through a balancing transformer to 50-ohm coaxial cable which led to two levels of multiplexing that served a cable tester. The rods were pressed into the soil parallel to each other and to land surface with 3-cm spacing between rods. The measured values are integrated averages of the moisture contents of the soil in the near-vicinity of the rods.The empirical expression of Topp and others (1980) was used to relate the apparent dielectric constant to volumetric moisture content. The mean moisture content of the soil above the heat flux plate was estimated based on the measured values of moisture content at depths of 4 and 10 cm (moisture-content values determined from equal-depth probes were averaged; moisture content from 0 to 4 cm depth was assumed equal to that measured at 4 cm; linear change in moisture content was assumed between 4 and 10 cm). The estimated value of mean moisture content (volumetric basis) was converted to a value of water content (gravimetric basis). Soil heat flux at land surface was estimated as the sum of the heat flux measured at an 8-cm depth and the rate of change in stored energy in the soil above the heat flux plate:

$$S = \frac{(T_i - T_{i-1})\, b\bar{C}_s}{\Delta t}, \qquad (1)$$

where

S is the rate of change of stored energy within the soil above heat-flux plate, in watts per square meter;

T_i is the temperature of the soil above heat flux plate at the end of the current time interval, in degrees Celsius;

$T_{i\text{-}1}$ is the temperature of the soil above heat flux plate at the end of the previous time interval, in degrees Celsius;

b is the depth to heat flux plate, equal to 0.08 meter;

Δt is the time interval between temperature measurements, equal to 1,200 seconds; and

$\bar{C}_s$ is the mean specific heat capacity of soil above heat flux plate, in joules per cubic meter per degree Celsius and is given by:

$$\bar{C}_s = \rho_b\,(\bar{C}_{sd} + w_m C_w), \qquad (2)$$

where

$\bar{\rho}_b$ is mean dry soil bulk density above heat flux plate, estimated as 1.45 kilograms per cubic meter;

$\bar{w}_m$ is mean gravimetric water content of soil above heat flux plate, in kilograms of water per kilogram of soil;

$\overline{C}_{sd}$ is the mean specific heat capacity of the dry soil above heat flux plate, estimated as 840 joules per kilogram per degree Celsius based on the mineral nature of the soil only 1.8 percent organic matter in upper 8 cm of soil (J. A. Tindall, USGS, oral commun., 1995); and

C_w is the specific heat capacity of water, equal to 4,190 joules per kilogram per degree Celsius.

The soil moisture measurements were also used to define the mean soil moisture within the rooting zone (upper 30 cm) in a manner similar to that used for the estimation of the mean moisture content of the soil above the heat flux plate.

Precipitation and water table depth were monitored at the site to further characterize the hydrometeorologic regime during the study period. Rainfall was measured using a tipping-bucket rain gage with an accompanying windshield. The water level in a surficial-aquifer well about 90 m from the micrometeorological instrumentation was monitored using a pressure transducer. Measurements of rainfall and surficial-aquifer water level were recorded with a datalogger at temporal resolutions of 20 and 60 min, respectively.

Eddy Correlation Measurement of Evapotranspiration

The eddy correlation method (Tanner and Greene, 1989) was used to measure two components of the energy budget of a plant canopy: latent and sensible heat fluxes. Latent heat flux (λE) represents the energy removed from the canopy in the liquid-to-vapor phase change of evapotranspired water and is the product of the heat of vaporization of water (λ) and the evapotranspiration rate (E). Sensible heat (H) represents the heat energy removed from the canopy as a result of a temperature gradient between the canopy and the air. Both latent and sensible heat fluxes are transported advectively by turbulent eddies in the air. The energy available to generate turbulent fluxes of vapor and heat is equal to the difference of net radiation (R_n) and heat flux into the soil (G). The net radiation incident on the top of the canopy is the difference of incoming radiation (shortwave solar radiation and longwave atmospheric radiation) and outgoing radiation (reflected shortwave radiation and longwave canopy radiation). Energy is transported to and from the base of the canopy by conduction through the soil. Assuming that net horizontal advection of energy and canopy storage of energy are both negligible quantities, the energy budget equation takes the following form:

$$R_n - G = H + \lambda E, \qquad (3)$$

where

the left-hand side of equation 3 represents the available energy, the right-hand side represents the turbulent flux, and

R_n is net radiation to/from plant canopy, in watts per square meter;

G is soil heat flux at land surface, in watts per square meter;

H is sensible heat flux, in watts per square meter;

λE is latent heat flux, in watts per square meter; and

the sign convention is such that R_n and G are positive downwards and H and λE are positive upwards.

The eddy correlation method is applicable in instances of adequate fetch. Fetch is adequate if there is sufficient upwind homogeneity in the vegetative cover so that the surface layer has equilibrated with the vegetative cover from ground surface to at least the height of the instrumentation. This upwind distance generally is considered to be about 100 times the instrumentation height (Stannard, 1993).

The eddy correlation method is a conceptually simple, one-dimensional approach. The time-averaged product of measured values of vertical wind speed (w) and vapor density (ρ_v) is the estimated vapor flux (evapotranspiration rate) during the averaging period, assuming that the net lateral advection of vapor is negligible (fetch is adequate). Because of the insufficient accuracy of instrumentation available for measurement of actual values of vertical wind speed and vapor density, this procedure generally is performed by monitoring the fluctuations of vertical wind speed and vapor density about their means, rather than monitoring their actual values. This formulation is represented by the following equations:

$$E = \overline{w\rho_v} = \overline{(\bar{w} + w')\,(\overline{\rho}_v + \rho_v')}, \qquad (4)$$

$$= \overline{\overline{w}\overline{\rho_v}} + \overline{\overline{w}\rho_v'} + \overline{w'\overline{\rho_v}} + \overline{w'\rho_v'}, \tag{5}$$

and

$$\cong \overline{w'\rho_v'} = covariance\,(w, \rho_v)\ , \tag{6}$$

where

overbars and primes indicate means over the averaging period and deviations from means, respectively;

E is evapotranspiration rate, in grams per square meter per second;

ρ_v is vapor density, in grams per cubic meter; and

w is vertical wind speed, in meters per second.

The first term of equation 5 is approximated as zero because mass-balance considerations dictate that mean vertical wind speed is zero—assuming constant air density (correction for temperature-induced air-density fluctuations will be noted later in this report). The second and third terms are zero based on the definition that the mean fluctuation of a variable is zero. Therefore, it is apparent from equation 6 that vertical wind speed and vapor density must be correlated for a non-zero vapor flux to exist. The turbulent eddies that transport water vapor (and sensible heat) produce fluctuations in both the direction and magnitude of vertical wind speed. The ascending eddies must generally be "wetter" than the descending eddies for evapotranspiration to occur; that is, upward air movement must be positively correlated with vapor density and downward air movement must be negatively correlated with vapor density.

Instrumentation capable of high-frequency resolution must be used in an application of the eddy correlation method because of the relatively high frequency of the turbulent eddies that transport water vapor (and sensible heat). Instrumentation included a single-axis sonic anemometer; a fine-wire, chromel-constantan thermocouple; and a krypton hygrometer to measure or infer variations in vertical wind speed, air temperature, and vapor density, respectively. The sonic anemometer relies on the sonic transducers, placed 10 cm apart vertically, to detect wind-induced phase shifts in emitted sound waves. The thermocouple was laterally displaced 4 cm from the middle of the sonic transducer pathline. The hygrometer relies on the attenuation of ultraviolet radiation, emitted from a source tube, by water vapor in the air along the 1-cm path to the detector tube. The instrument pathline was laterally displaced 10 cm from the middle of the sonic-transducer pathline and 8 cm from the thermocouple. Hygrometer voltage output is proportional to the attenuated radiation flux and fluctuations in attenuated radiation flux can be related to fluctuations in vapor density by Beer's Law (Weeks and others, 1987). Eddy correlation instrument-sampling frequency was 10 Hz with 20-min averaging periods. Data were processed and stored in a datalogger. The susceptibility of the sonic anemometer transducers and the thin-wire thermocouple to damage by rainfall prevented the continuous deployment of the eddy correlation instrumentation in the field.

Eddy correlation instrumentation was placed about 2.0 m above land surface in a relatively flat area of the site. This measurement height places the sensors at least 1 m above the canopy at all times, avoiding sonic transducer "echo" effects. The northern and eastern boundaries of the site were adequately spaced (about 290 and 430 m, respectively) from the eddy correlation instrumentation, easily satisfying the 1:100 instrument height-to-fetch rule. However, this ratio was as low as 1:85 and 1:90 for the southern (170 m distant) and western site (180 m distant) boundaries, respectively; arcs of 70 and 57 degrees (to the south and west, respectively) did not satisfy the 1:100 fetch rule. Because the 1:100 fetch rule is an empirical approximation, data collected during times when the wind direction was within one of these arcs were not immediately rejected. Rather, energy-budget closure criteria (discussed later in this section) were used to eliminate data that did not satisfy the assumption of one-dimensional vertical transport of turbulent energy.

Latent and sensible heat fluxes were estimated based on a modified form of equation 6 as:

$$\overline{\lambda E} = \bar{\lambda}\left(\overline{w'\rho'_v} + \frac{\overline{\rho_v}\bar{H}}{\bar{\rho}\,\overline{C_p}\,\overline{(T_a + 273.15)}} + \frac{FK_0\bar{H}}{K_w\,\overline{(T_a + 273.15)}} \right) \tag{7}$$

and

$$\bar{H} = \bar{\rho}\,\overline{C_p}\,\overline{w'T_a'}\quad , \tag{8}$$

where

λE is latent heat flux, in watts per square meter;

λ is latent heat of vaporization of water, estimated as a function of temperature (Stull, 1988), in joules per gram;

H is sensible heat flux, in watts per square meter;

ρ is air density, estimated as a function of air temperature, total air pressure, and vapor pressure (Monteith and Unsworth, 1990), in grams per cubic meter;

C_p is specific heat of air, estimated as a function of specific humidity (Stull, 1988), in joules per gram per degree Celsius;

T_a is air temperature, in degrees Celsius;

F is a factor used in krypton hygrometer correction that accounts for molecular weights of air and oxygen, and atmospheric abundance of oxygen, equal to 0.229 gram-degree Celsius per joule;

K_0 is extinction coefficient of hygrometer for oxygen, estimated as 0.0045 cubic meters per gram per centimeter (Tanner and others, 1993);

K_w is extinction coefficient of hygrometer for water, equal to the manufacturer-calibrated value, in cubic meters per gram per centimeter;

and overbars and primes indicate means over the averaging period and deviations from the means, respectively.

The second and third terms of the right-hand side of equation 7 account for temperature-induced fluctuations in air density (Webb and others, 1980) and for the sensitivity of the hygrometer to oxygen (Tanner and Greene, 1989).

Previous investigators (Weeks and others, 1987; Bidlake and others, 1993; and Stannard, 1993) have described a recurring problem with the eddy correlation method—an occasional inconsistency in the measured latent and sensible heat fluxes with the energy-budget equation (eq 3), presumably because of measurement or conceptual errors. The more usual case is that measured turbulent fluxes (H and λE) have not been sufficient to account for the measured available energy. In a manner similar to Stannard (1993), the energy-budget equation was used as a means of filtering data with large energy-budget discrepancies from the data set. If the turbulent flux for a given 20-min time period was more than 20 percent and 20 W/m^2 different than the available energy, the evapotranspiration estimate was considered unacceptable. Evapotranspiration during a period of unacceptable energy-balance closure was estimated using evapotranspiration models described later in this report.

Relatively small energy-balance discrepancies commonly remain in an eddy correlation-estimated evapotranspiration data set after filtering for removal of large discrepancies. Bidlake and others (1993) present a technique that acknowledges these discrepancies and offer three alternative estimates of evapotranspiration based on differing assumptions regarding the source of these discrepancies. The standard eddy correlation method assumes that error in the measurement of available energy is the source of energy-balance discrepancies and that the measured latent and sensible turbulent fluxes are correct. The eddy correlation energy-balance Bowen ratio method (ECEBBR) assumes that error in the sonic anemometer measurements of vertical wind speed accounts for the discrepancies. If the error in wind speed estimation is assumed to apply proportionally to both latent and sensible heat fluxes, the Bowen ratio ($H/\lambda E$) is preserved and the energy-budget equation allows solution of turbulent fluxes. The eddy correlation energy-balance residual method (ECEBR) assumes that error in the krypton hygrometer measurements of vapor density accounts for the discrepancies. With this assumption, the measured sensible heat flux and the energy-balance equation are used to estimate latent heat flux. The three alternative eddy correlation methods provide a range of evapotranspiration estimates—the standard eddy correlation method generally provides the lower bound of this range, while the ECEBR method generally provides the upper bound. Bidlake and others (1993) noted that evapotranspiration estimates from traditional Bowen ratio instrumentation and the ECEBBR variant of the eddy correlation method were in approximate agreement.

The measurement of nighttime latent heat flux with the eddy correlation method generally is not accurate. Dew formation on the windows of the hygrometer can produce erroneously high measurements of vapor density. Also, the small nighttime energy fluxes can be relatively large compared to instrumentation noise. Fortunately, the fraction of total nighttime evapotranspiration is relatively small.

Air-flow distortions in the vicinity of the sonic anemometer can contribute to error in the eddy correlation method. The krypton hygrometer must be placed near the sonic anemometer in order that nearly the same turbulent eddies are monitored at any one time. A negative side effect from this instrument proximity is the flow distortion produced by the hygrometer when this instrument is upwind of the anemometer.

"Poor" wind directions for a typical sensor arrangement comprise a 75^{o} sector. To minimize flow distortion error, the eddy correlation instrumentation was positioned relative to the prevailing wind at the time of deployment for each of the 22 measurement events. However, varying wind directions during an event could have produced flow-distortion errors. Measurements with this type of error were assumed to have been culled from the data set with the energy-balance criteria.

Penman-Monteith Estimation of Evapotranspiration

Penman (1948) developed an expression for evaporation from open water based on energy budget and aerodynamic principles. That equation has since been applied to predict evapotranspiration from a well-watered, dense agricultural crop ("potential" evapotranspiration). In the Penman expression, the transport of latent and sensible heat fluxes from a "big leaf" to the sensor height is subject to an aerodynamic resistance. The big leaf assumption implies that the plant canopy can be conceptualized as a single source of both latent and sensible heat at a given height and temperature. Inherent in the Penman approach is an assumption of a net one-dimensional transport of vapor and heat between the canopy and the instrumentation, a condition that is met under adequate fetch. Monteith (1965) modified the Penman equation, allowing for the possibility of an additional "canopy" resistance to vapor transport due to plant-canopy resistance. The resulting "Penman-Monteith" equation takes the following form:

$$\lambda E = \frac{\Delta (R_n - G) + \dfrac{\rho C_p (e_s - e)}{r_h}}{\Delta + \dfrac{\gamma (r_c + r_h)}{r_h}}, \qquad (9)$$

where

λE is latent heat flux, in watts per square meter;

Δ is slope of the saturation vapor-pressure curve, estimated as a function of air temperature using the empirical relation of Lowe (1977), in kilopascals per degree Celsius;

e_s is saturation vapor pressure, estimated as a function of air temperature using the empirical relation of Lowe (1977), in kilopascals;

e is vapor pressure, equal to the product of relative humidity and e_s, in kilopascals;

C_p is specific heat capacity of the air, computed as a function of vapor pressure and atmospheric pressure (Stull, 1988), in joules per gram per degree Celsius;

γ is the psychrometric "constant," computed as a function of atmospheric pressure and air temperature (Fritschen and Gay, 1979), in kilopascals per degree Celsius;

r_h is aerodynamic resistance, in seconds per meter; and

r_c is canopy resistance, in seconds per meter.

The first term in the numerator of the Penman-Monteith equation is known as the energy term and the second is known as the aerodynamic term. The aerodynamic resistance to the transport of vapor and heat from the plant canopy to the sensor is assumed to be the same for both constituents—a consequence of the big leaf assumption and similarity theory. Aerodynamic resistance was computed in a manner similar to that described by Stannard (1993) in which an initial assumption of neutral boundary-layer conditions was made and the following equations were iteratively solved in the order indicated until convergence was achieved:

1.

$$u^* = \frac{ku}{\left[\ln\left(\dfrac{z - d + z_0}{z_0}\right) + \Psi_m\right]} . \qquad (10)$$

2.

$$r_h = \frac{\ln\left(\dfrac{z - d + z_h}{z_h}\right) + \Psi_h}{ku^*} . \qquad (11)$$

3. Estimate latent heat flux with Penman-Monteith equation (eq 9).
4. Estimate sensible heat flux with energy-balance equation (eq 8).

5.

$$\zeta = \frac{-kg\,(z - d + z_0)\,(H + 0.07\lambda E)}{\rho C_p T u^{*3}} \quad . \qquad (12)$$

6. Estimate stability profile correctors.
   For unstable ($\zeta < 0$) and neutral ($\zeta = 0$) conditions:

$$X = (1 - 16\zeta)^{1/4} \quad , \qquad (13)$$

$$\psi_m = 2\,\mathrm{atan}X - \frac{\pi}{2} - \ln\left\{\left[\frac{\left(1 + X^2\right)}{2}\right]\left[\frac{(1 + X)}{2}\right]^2\right\} \quad ,(14)$$

and

$$\psi_h = -\ln\left\{\left[\frac{(1 + X^2)}{2}\right]^2\right\} \quad . \qquad (15)$$

For stable conditions ($\zeta > 0$):

$$\psi_m = \psi_h = 5\zeta \quad , \qquad (16)$$

where
- u^* is friction velocity, in meters per second;
- u is horizontal wind speed at sensor height, in meters per second;
- k is von Karman's constant, equal to 0.4, dimensionless;
- z is height of sensors (wind speed, air temperature, and relative humidity) above land surface, in meters; the height of the anemometer (1.8 m), rather than that of the temperature and relative humidity sensors (1.7 m), was used for this variable, because vertical gradients of the latter are relatively small;
- d is momentum displacement height of vegetation, in meters;
- z_0 is roughness length of canopy for momentum, in meters;
- z_h is roughness length of canopy for sensible heat, in meters;
- ψ_m is stability profile corrector for momentum, dimensionless;
- ψ_h is stability profile corrector for sensible heat, dimensionless;
- ζ is atmospheric stability, dimensionless.

The momentum-displacement height and roughness lengths change with canopy height and density. Average canopy height, h, was estimated by visual inspection at the field site about every 2 weeks during the study period. The empirical relations of Campbell (1977) for dense canopies were used to estimate momentum-displacement height and roughness lengths:

$$d = 0.64h \quad , \qquad (17)$$

$$z_0 = 0.13h \quad , \qquad (18)$$

and

$$z_h = 0.20z_0 \quad . \qquad (19)$$

The canopy resistance within the Penman-Monteith equation is the equivalent resistance of all the plant stomatal resistances in parallel and therefore the resistance increases with increasing stomatal constriction and/or decreasing stomatal (foliage) density. By dynamically adjusting the size of the stomatal openings in response to PAR, leaf-air vapor pressure difference, leaf-water potential, ambient carbon dioxide concentration, and leaf air temperature (Jarvis, 1976), plants optimize their growth and survival. During plentiful PAR and high ambient concentrations of carbon dioxide, the stomata of plants (excluding succulents) tend to be relatively open to increase the carbon dioxide flux to the plant while the energy to photosynthetically fix carbon is available. Plant stomata tend to constrict under conditions of high atmospheric demand (large leaf-air vapor pressure difference) or low leaf-water potential (reduced plant turgor pressure, leading to wilting). Stomatal opening is restricted at low and high leaf temperatures and is least constricted at intermediate temperatures. As foliage or leaf density increases through seasonal or growth cycles of the plant, the number of stomata available for plant-to-atmosphere vapor transport changes, therefore changing the canopy resistance to vapor flux.

Canopy resistance generally is estimated by either (1) upscaling porometer measurements of stomatal conductance (reciprocal of resistance) on individual plant leaves, or by (2) use of the Penman-Monteith equation (eq 9) along with an independently estimated value of evapotranspiration (for example, using eddy correlation or Bowen ratio methodologies) and other measured meteorological variables to "back-solve" for canopy resistance. The first method has been used successfully (Abtew, 1995), but is subject to the subjectivity of upscaling approaches and to the impracticality of achieving high temporal resolution over long time periods (field personnel required at site to perform porometer measurements). The second method is not subject to these limitations and has been used (Stannard, 1993) to obtain estimates of canopy resistance that are consistent, within the framework of the Penman-Monteith equation, with measured values of evapotranspiration. This second method was used in the investigation described in this report.

Eddy correlation measurements made periodically from September 1993 to August 1994 were evaluated for acceptable energy-balance closure. Only those measurements that met the closure criteria were used to estimate canopy resistance. Previous investigators have used parameterized forms of canopy-resistance functions (Jarvis, 1976; Avissar and others, 1985; Stewart, 1988; and Stannard, 1993). The form of these functions generally was based on the results of experiments conducted in controlled, laboratory environments with a plant monoculture. A variety of functional forms of canopy resistance were evaluated in this study as a means of expressing the dependence of this variable on environmental variables identified by Jarvis (1976) as being important factors in stomatal constriction. As part of this evaluation, field measurements were made of these environmental variables or of more easily measured variables that are highly correlated with the variables identified by Jarvis. Field-measured environmental variables were PAR, vapor-pressure deficit of air (a surrogate for leaf-air vapor pressure difference), soil moisture content (a surrogate for leaf-water potential), and air temperature (a surrogate for leaf temperature). The effect of carbon dioxide concentrations on canopy resistance was not considered in this study. Parameters within the canopy-resistance functions were estimated through an optimization procedure—the Powell algorithm (Press and others, 1989)—which minimized the sum of squared differences between the Penman-Monteith simulated and measured values of evapotranspiration. This procedure was followed for each of the three alternative estimates of the eddy correlation method. Next, the "calibrated" canopy-resistance functions were used in conjunction with the Penman-Monteith equation and environmental data monitored during the study period to estimate evapotranspiration at times during which the eddy correlation data did not meet energy-balance closure criteria and at times between eddy correlation measurement events.

The Penman-Monteith equation was not applied to nighttime computation of evapotranspiration; an evapotranspiration rate of zero was assumed during this time. Similar to Stannard (1993), this assumption was made because (1) evapotranspiration from vegetation is relatively small at night and (2) the iterative calculation of u^* in the Penman-Monteith model often becomes unstable for nighttime data, preventing the accurate estimation of evapotranspiration and dew formation. PAR was used to provide a quantitative day-night demarcation (nighttime defined as PAR < 1 $\mu\text{moles}/(\text{m}^2 \cdot \text{s})$).

Dew formation is frequent in humid areas such as Florida. If dewfall is substantial but nighttime vapor fluxes are not explicitly measured, evapotranspiration will be overestimated (nighttime dewfall is not measured, but early-morning dew evaporation is measured). Therefore, a methodology to exclude early-morning dew evaporation was needed in this investigation. Abtew and Obeysekera (1995) noted that the mean time to evaporate dew from a cattail marsh in south Florida was 78 minutes after the onset of positive net radiation. In the present study, this criterion (assumed time-invariant and rounded to 80 minutes) was used to prescribe evapotranspiration as zero during times of expected dew evaporation. This approach assumes that dew formed nightly during the study period and that negligible transpiration occurred in the presence of dew.

Leaf-area index was estimated (by visual inspection) for possible use within the canopy-resistance function in a manner similar to that described by Stannard (1993). This method is based on an application of Beer's Law for attenuation of radiation in a medium (Monteith and Unsworth, 1990):

$$L_s = \frac{\left(1 - e^{-\kappa_s \frac{L}{2}}\right)}{\kappa_s} , \qquad (20)$$

where

L_s is sunlit leaf area per unit area of land, in square meters per square meter;

κ_s is attenuation coefficient of light for the canopy, dimensionless;

L is two-sided leaf area per unit area of land (leaf-area index), in square meter per square meter.

Assuming that the leaf-angle distribution is ellipsoidal, the attenuation coefficient is equal to about one when the solar elevation is 30^{o} (Monteith and Unsworth, 1990, p. 76). To an observer viewing the canopy at an angle of 30^{o} below the horizon, the green leaf-area fraction visible, L_v, is equal to the sunlit leaf-area fraction at a solar elevation of 30^{o}. Therefore,

$$L = -2 \ln (1 - L_v) \quad . \qquad (21)$$

To estimate leaf-area index using this method, site vegetation was photographed at an angle of 30^{o} below the horizon about every 2 weeks during the study period. Leaf-area index was estimated at unmonitored times with linear interpolation of measured values.

Simpler Methods for Estimation of Evapotranspiration

Two simpler alternatives to the Penman-Monteith method were considered in this study. These alternatives - the Penman and Priestley-Taylor methods - require less meteorological data and are less computationally demanding than the Penman-Monteith method.

As discussed previously, the Penman method estimates potential evapotranspiration assuming a dense canopy without moisture stress. Computationally, this assumption is equivalent to assuming a negligible canopy resistance within the Penman-Monteith approach. Consequently, the challenge of estimating this empirical term is eliminated. The Penman equation takes the following form:

$$\lambda E = \frac{\Delta (R_n - G) + \dfrac{\rho C_p (e_s - e)}{r_h}}{\Delta + \gamma} \quad , \qquad (22)$$

where

all variables have been previously defined (see eq 9).

Although the assumption of ample soil moisture and dense canopy coverage often is not met in the upland, nonirrigated areas of Florida, the possibility remains that potential evapotranspiration might exceed actual evapotranspiration by some consistent and quantifiable degree and therefore could serve as a reliable estimator of actual evapotranspiration. The usefulness of the Penman equation in this regard was examined within this investigation.

Priestley and Taylor (1972) proposed a simplification of the Penman equation for the case of saturated atmosphere ($e = e_s$), where the aerodynamic term of the Penman equation (eq 22) is zero:

$$\lambda E = \frac{\Delta (R_n - G)}{\Delta + \gamma} \quad . \qquad (23)$$

However, Priestley and Taylor (1972) noted that empirical evidence suggests that evaporation from extensive wet surfaces is greater than this amount, presumably because the atmosphere generally does not attain saturation. Therefore, the Priestley-Taylor coefficient, α, was introduced as an empirical correction to the theoretical expression (eq 23):

$$\lambda E = \alpha \frac{\Delta (R_n - G)}{\Delta + \gamma} \quad . \qquad (24)$$

The value of α was estimated to be 1.26. Eichinger and others (1996) have shown that the empirical value of α has a theoretical basis—a nearly constant value of α is expected under the existing range of Earth-atmospheric conditions.

Previous studies (Flint and Childs, 1991; Stannard, 1993) have examined the use of a modified form of the Priestley-Taylor equation. The approach in these studies relaxes the assumption of a free-water surface or a dense, well-watered canopy by allowing α to be less than 1.26 and to vary as a function of environmental factors. Stannard (1993) noted that the Priestley-Taylor approach to simulation of observed evapotranspiration rates was superior to that of Penman-Monteith for a site of wildland vegetation in semiarid rangeland. In this investigation, the utility of the Priestley-Taylor equation to estimate evapotranspiration was evaluated. The functional form of the Priestley-Taylor α was estimated in a manner analogous to that described for the Penman-Monteith r_c.

RESULTS OF EVAPOTRANSPIRATION MEASUREMENT AND ESTIMATION MEASUREMENT OF EVAPOTRANSPIRATION

The 20-min resolution eddy correlation measurements of evapotranspiration that met the energy-balance criteria provided good temporal coverage—both diurnally and seasonally (fig. 12). There were a total of 538 daytime (following estimated time of complete dew evaporation) measurements that were considered acceptable; about half of the daytime evapotranspiration measurements were discounted using the energy-balance criteria. Results for the first of the 22 measurement events are shown in figures 13-15. All energy terms—net radiation, soil heat flux, sensible heat flux, and latent heat flux—exhibit a strong diurnal pattern (fig. 13). Net radiation generally is the dominant energy term. During daytime, net radiation and soil heat flux are relatively high and positive (net downward flux) and are relatively low and negative (net upward flux) during nighttime as the soil radiates heat to the atmosphere. The effect of intermittent cloud cover is evident as sharp variations in these energy fluxes.

Absolute energy-balance closure (available energy minus turbulent flux) was relatively good at night (low energy flux environment) (fig. 14). However, the relative energy-budget closure (turbulent flux energy divided by available energy) was poor, probably because of dew formation on the hygrometer and the relative magnitudes of energy fluxes and instrumentation noise. Conversely, daytime absolute energy-balance closure was relatively poor, but the relative energy-budget closure was good. The absolute energy-balance closure generally was positive, indicating either an overestimation of available energy or an underestimation of turbulent flux. The three variants of the eddy correlation method, which ascribe energy-balance error to three different sources, are exemplified in figure 15. Because of the generally positive absolute energy-balance closure, the highest, intermediate, and lowest estimates of evapotranspiration are provided by the ECEBR, ECEBBR, and standard eddy correlation methods, respectively.

Model Calibration Using Measured Evapotranspiration

The 538 measurements of daytime evapotranspiration deemed acceptable provided a standard with which to calibrate the evapotranspiration models—Penman-Monteith, Penman, and modified Priestley-Taylor equations. The functional form of Penman-Monteith canopy resistance that was initially considered was based on a composite of empirical models proposed by Jarvis (1976) and Stewart (1988); terms are included in this model to account for the dependence of canopy resistance on leaf-area index, vapor-pressure deficit, PAR, air temperature, and rooting zone soil moisture. Canopy resistance within this traditional approach (“method 1” in this report) takes the following form:

$$r_c = \left[C_1 L\left(\frac{C_2}{C_2 + D}\right)\left(\frac{R_{PAR}}{R_{PAR} + C_3}\right)\left(\left(T_a - C_5\right)\left(C_6 - T_a\right)^{C_4}\right)\left(1 - e^{C_7\left(\bar{\theta} - C_8\right)}\right)\right]^{-1} \qquad (25)$$

where

D is vapor pressure deficit (equal to e_s-e), in kilopascals;

R_{PAR} is photosynthetically active radiation, in micromoles per square meter per second;

T_a is air temperature, in degrees Celsius;

$\bar{\theta}$ is mean moisture content in rooting zone (upper 30 cm of soil), in cubic centimeters per cubic centimeter;

C_1-C_8 are site-specific empirical parameters; and other terms are as previously defined.

The parameter C_8, equal to the moisture content at which canopy resistance is infinite, was set equal to zero; evapotranspiration was measured at the site when $\bar{\theta}$ was as low as 0.02 cm^3/cm^3, implying that canopy resistance was finite above this moisture content.

The results of estimation of the parameters in equation 25, using evapotranspiration estimates provided by the ECEBBR method, are shown in table 2. Alternative, simplified versions of equation 25, in which individual terms were eliminated, were used to examine the extent to which these terms were superfluous. Results indicated that all terms (particularly the term involving PAR) make a significant contribution to the description of canopy resistance. Values of parameters C_5 and C_6 selected by the parameter optimization procedure produced the best match between simulated and measured values of evapotranspiration;

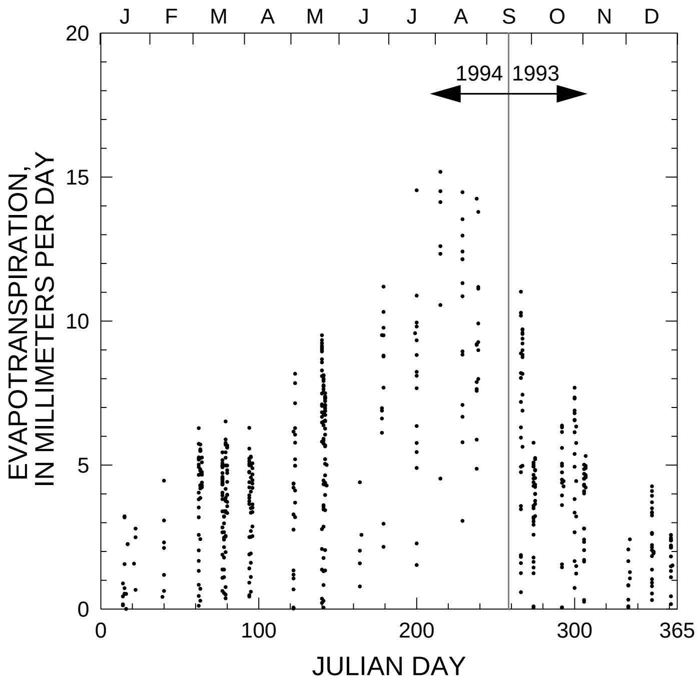
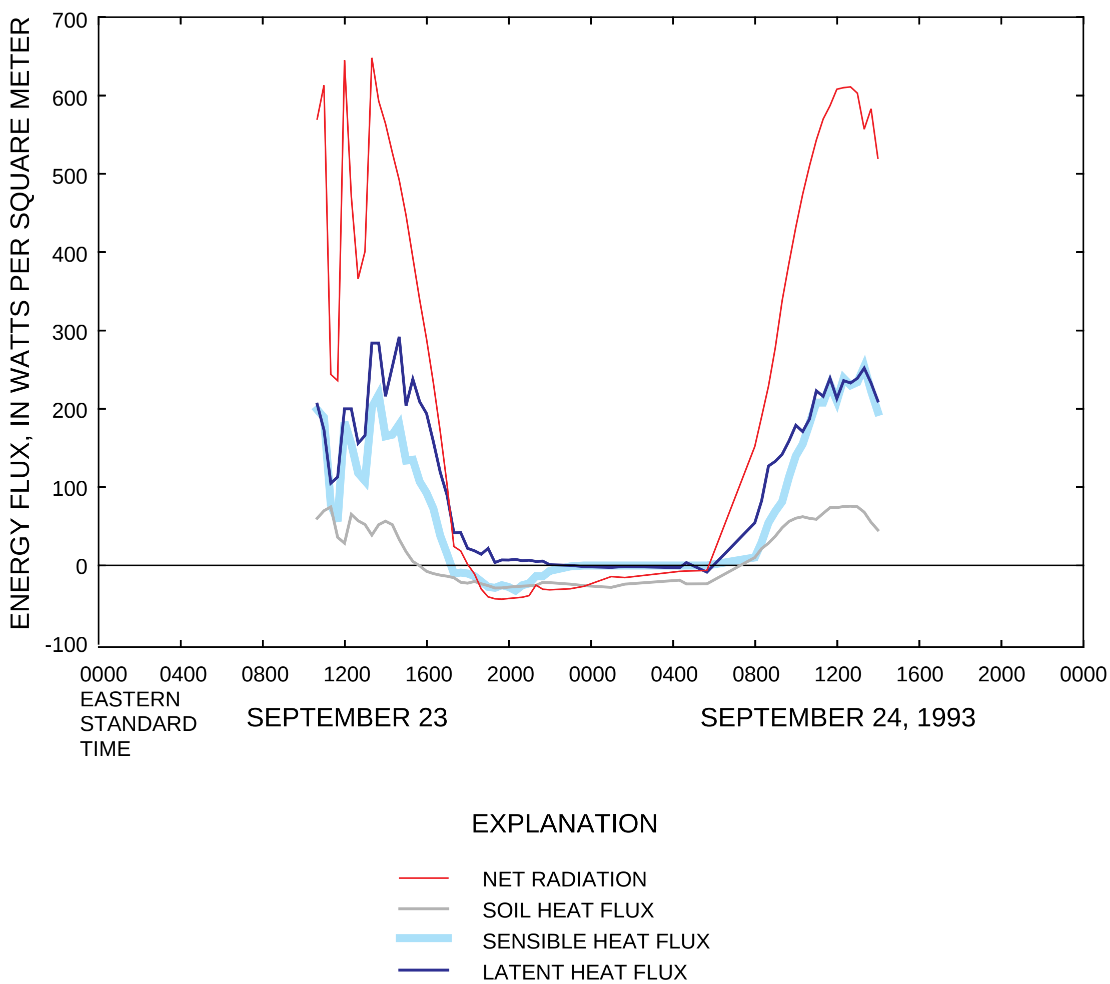


Figure 12. Evapotranspiration measured by eddy correlation energy-balance Bowen ratio method at study site from September 1993 to August 1994. Note: Measurements shown include only daytime (following estimated time of complete dew evaporation) values that met the energy-balance criteria.

Figure 13 (Below). Surface energy-balance components on September 23-24, 1993, at study site. Note: Sensible and latent heat flux were estimated with the standard eddy correlation method.



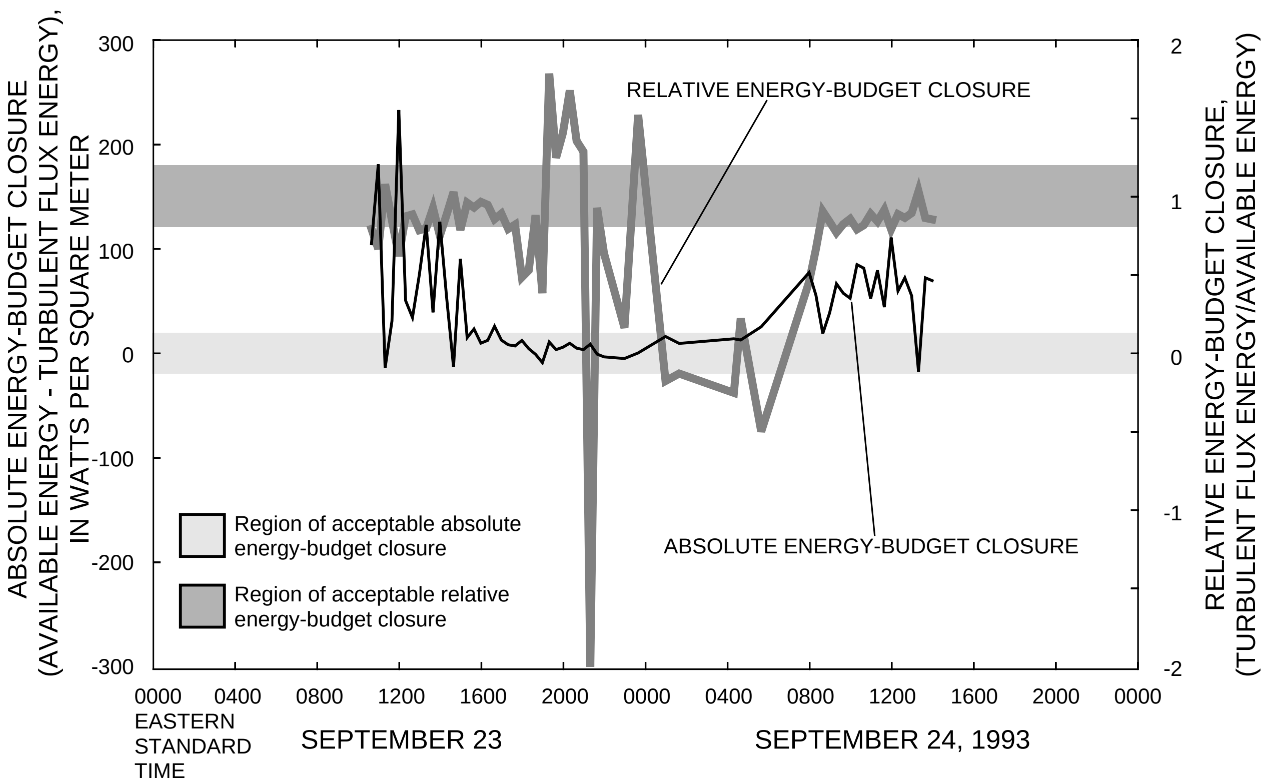


Figure 14. Absolute and relative measures of energy-budget closure for eddy correlation data collected on September 23-24, 1993, at study site. Note: Evapotranspiration measurements were considered “unacceptable” if the measurement did not meet either the absolute energy-budget closure criterion (turbulent flux within 20 W/m^2 of available energy) or the relative energy-budget closure criterion (turbulent flux energy within 20 % of available energy.

however, these values probably are not physically realistic. The coefficients C_5 and C_6 are defined as the lower and upper temperatures, respectively, at which canopy resistance is infinite. The coefficient C_5 is equal to -91 ˚C, implying that full stomatal closure does not occur until that temperature is reached; stomatal closure probably would occur just below freezing, at the coldest. Also, the selected values of C_4, C_5, and C_6 imply that the optimal (lowest) value of canopy resistance occurs at a temperature of -39 ˚C; optimal values of canopy resistance are expected to occur at temperatures above freezing. This anomalous relation of canopy resistance to air temperature probably is the result of inadequacies in the model (eq 25) or because of correlation between the independent variables of the model that obscure a unique definition of the model parameters.

Comparisons between modeled and measured values of evapotranspiration and canopy resistance are shown in figures 16 and 17, respectively. The model generally performed better in estimating evapotranspiration than in estimating canopy resistance. Because simulation of evapotranspiration—rather than canopy resistance—was the objective of the study, the mismatch in canopy-resistance values points out the importance of calibrating the canopy-resistance model on measured values of evapotranspiration, rather than “measured” values of canopy resistance. As also noted by Stannard (1993), the mismatch in canopy-resistance values generally occurred during times of low turbulent flux (shortly after sunrise or before sunset) when instrumentation error is relatively high. Because the latent heat flux was low at these times, error in simulating canopy resistance did not seriously degrade estimating daily evapotranspiration. The model exhibited substantial seasonal bias (fig. 18), generally over-predicting evapotranspiration during the months of July to October and under-predicting evapotranspiration during other periods. An example of the ability of the model to simulate diurnal patterns of evapotranspiration is shown in figure 19.

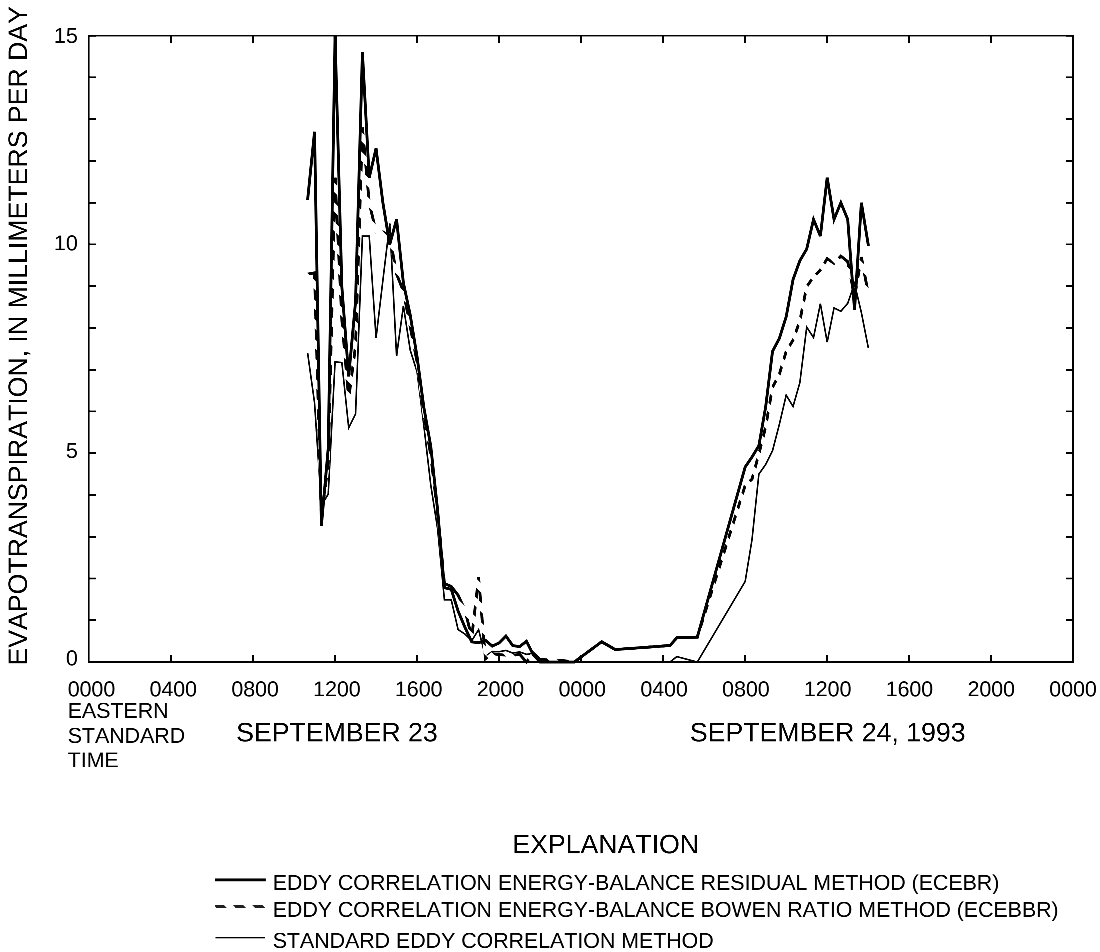


Figure 15. Eddy-correlation-measured evapotranspiration on September 23-24, 1993, at study site.

Table 2. Summary of parameters and error statistics for evapotranspiration models

[Method 1, see equation 25; method 2, see equation 26; Priestley-Taylor, see equation 27; SEE, Standard error of estimate (in millimeters per day); --, not applicable]

	Penman-Monteith (method 1)	Penman-Monteith (method 2)	Priestley-Taylor
	Parameters		
C_1	0.000000388	-0.144	-0.0000978
C_2	1.30	157	.0118
C_3	1,800	-1,920	4.64
C_4	1.78	313.	.508
C_5	-91.0	.504	-.349
C_6	53.5	305	294
C_7	-32.0	--	--
C_8	.0	--	--
	Error statistics		
r^2	.918	.967	.972
SEE	1.33	.84	.77

A nontraditional model for canopy resistance also was considered, referred to as “method 2” in this report. This model assumed a linear relation between all environmental variables and canopy resistance. Also, a sinusoidal function of Julian day with a wavelength of 1 year was used as a surrogate for leaf-area index:

$$r_c = \left(C_1 R_{PAR} + C_2 D + C_3 \bar{\theta} + C_4\right)\left(C_5 \sin\left(\omega\left(t - C_6\right)\right) + 1\right) \qquad (26)$$

where

ω is $2\pi/365$ per day;

t is Julian day; and

C_1-C_6 are site-specific empirical parameters.

The results of parameter estimation are shown in table 2. An additional linear term in equation 26

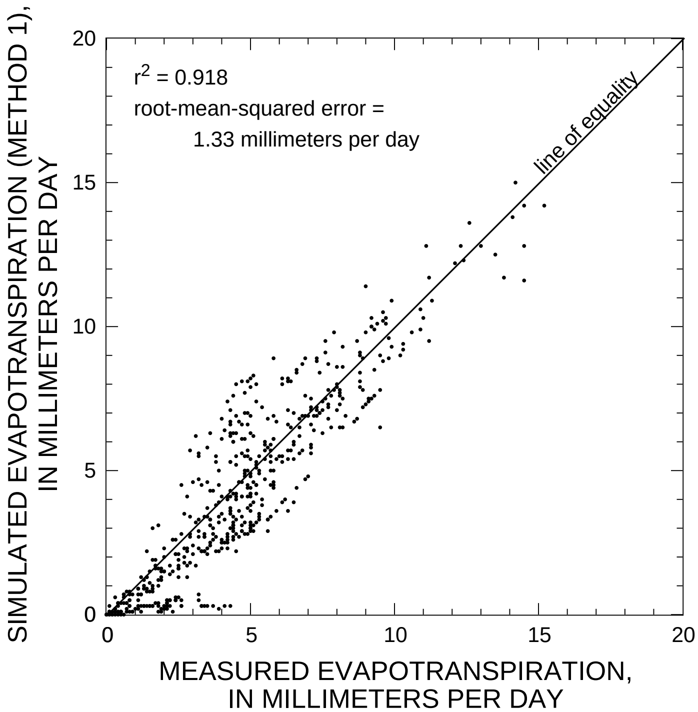


Figure 16. Comparison of Penman-Monteith-simulated evapotranspiration (method 1) and measured (eddy correlation energy-balance Bowen ratio method) evapotranspiration at study site from September 1993 to August 1994.

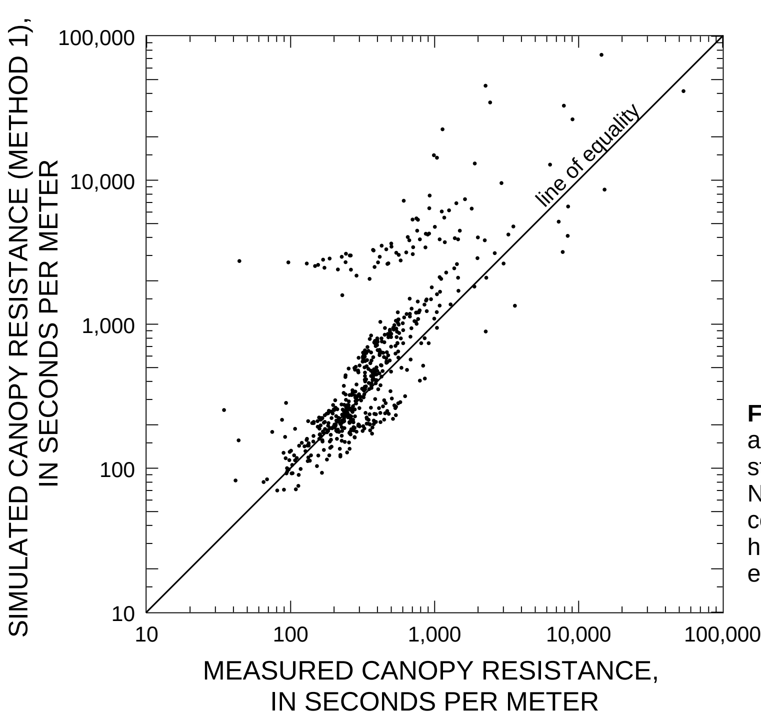


Figure 17. Comparison of simulated (method 1) and measured values of canopy resistance at study site from September 1993 to August 1994. Note: Measured values of canopy resistance computed using values of latent and sensible heat flux measured with the eddy correlation energy-balance Bowen ratio method.

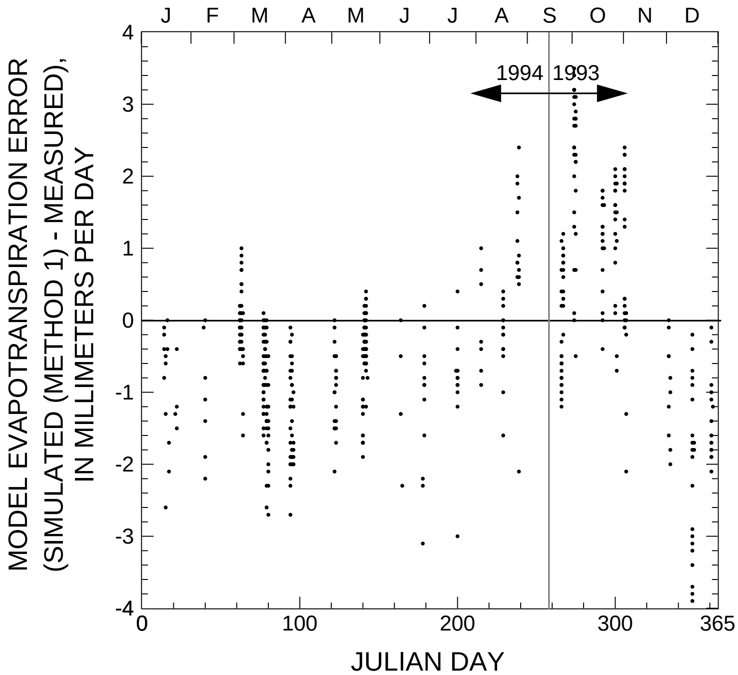
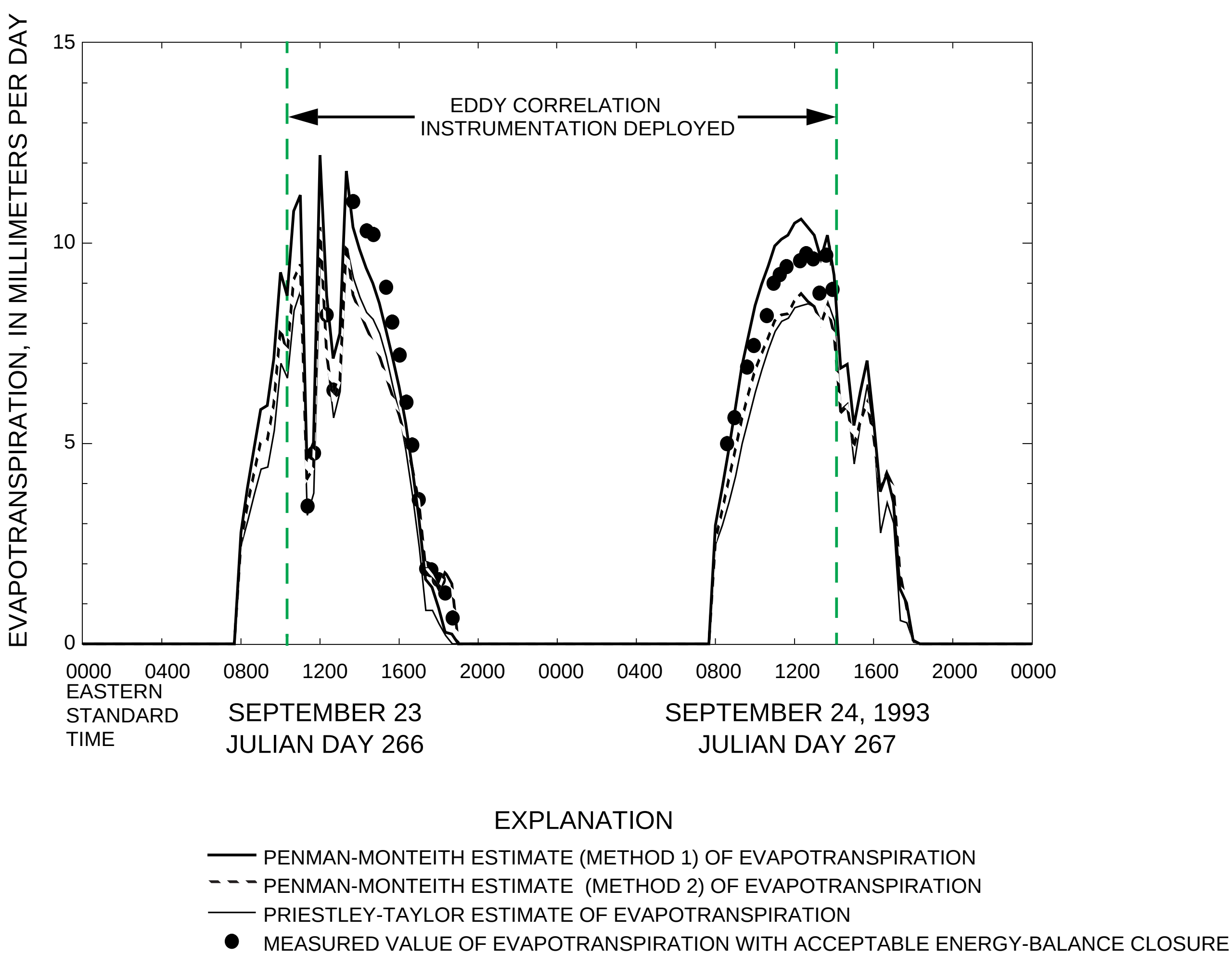


Figure 18. Seasonal distribution of discrepancy between Penman-Monteith-simulated evapotranspiration (method 1) and measured (eddy correlation energy-balance Bowen ratio method) evapotranspiration at study site from September 1993 to August 1994.

Figure 19: (Below) Comparison between simulated and measured evapotranspiration on September 23-24, 1993, at study site.



involving air temperature was shown to contribute a negligible amount of unique information to simulation of canopy resistance and was eliminated.

Canopy resistance, modeled with method 2, provided a correspondence between measured and simulated evapotranspiration that was superior to that of method 1 (table 2; figs. 16 and 20), in spite of the relative simplicity of method 2 (6 parameters, rather than 7). The standard error of estimate was reduced from 1.33 mm/d (method 1) to 0.84 mm/d (method 2). Additionally, the strong seasonal error bias noted with method 1 (fig. 18) is not apparent with method 2 (fig. 21). Canopy resistance also is simulated better with method 2 than with method 1 (figs. 17 and 22), although substantial error remains for high resistance/low turbulent flux conditions.

Calibration of the Penman-Monteith (method 2) model with estimates of evapotranspiration provided by eddy correlation variants other than the ECEBBR method (standard eddy correlation and ECEBR methods) produced canopy-resistance functions that were structurally similar to that obtained using the ECEBBR method. Also, the standard errors of evapotranspiration estimate produced using these variants of the eddy correlation method were approximately equal to that produced using the ECEBBR method. However, the canopy-resistance values obtained in this manner were slightly higher with the standard eddy correlation estimates and slightly lower with the ECEBR estimates than with the ECEBBR estimates of evapotranspiration.

Evapotranspiration estimated by the Penman equation was compared to that estimated by the more rigorous Penman-Monteith equation (method 2). The relative discrepancy between the two estimates was somewhat erratic, although a seasonal trend was apparent (fig. 23). This discrepancy was not unexpected given the variable soil-moisture content and canopy coverage noted at the site. The lack of a consistent relation between the evapotranspiration estimates of the Penman and Penman-Monteith equations precludes the use of the Penman equation as a reliable, and simple, estimator of evapotranspiration from non-irrigated vegetation in central Florida. However, Bidlake and others(1993) note that the Penman equation may be useful in estimating evapotranspiration from a marsh vegetation type.

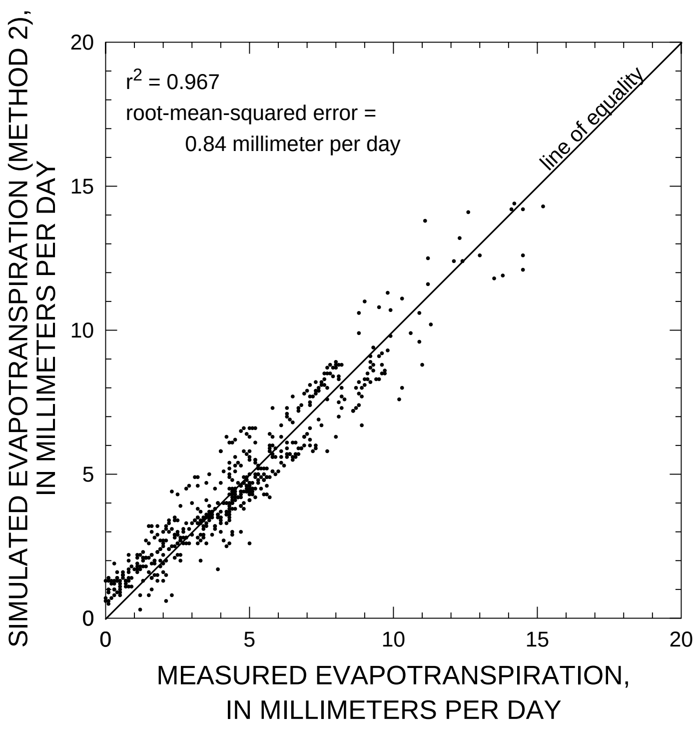


Figure 20. Comparison of Penman-Monteith-simulated evapotranspiration (method 2) and measured (eddy correlation energy-balance Bowen ratio method) evapotranspiration at study site from September 1993 to August 1994.

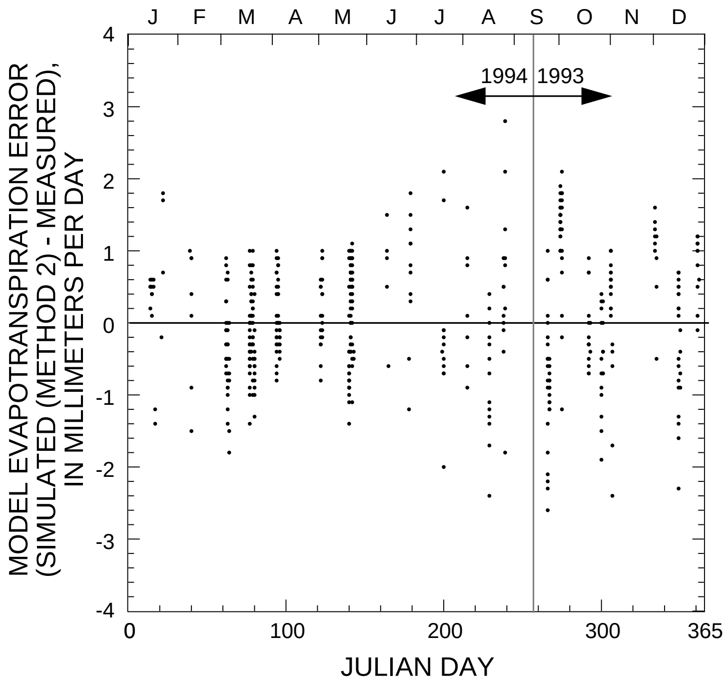


Figure 21. Seasonal distribution of discrepancy between Penman-Monteith-simulated evapotranspiration (method 2) and measured (eddy correlation energy-balance Bowen ratio method) evapotranspiration at study site from September 1993 to August 1994.

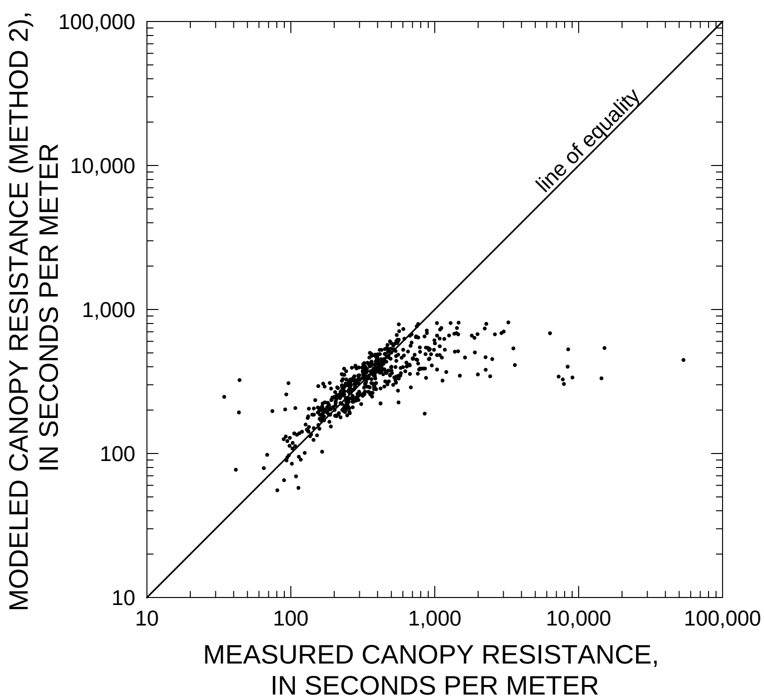


Figure 22. Comparison of modeled (method 2) and measured values of canopy resistance at study site from September 1993 to August 1994. Note: Measured values of canopy resistance computed using values of latent and sensible heat flux measured with the eddy correlation energy-balance Bowen ratio method.

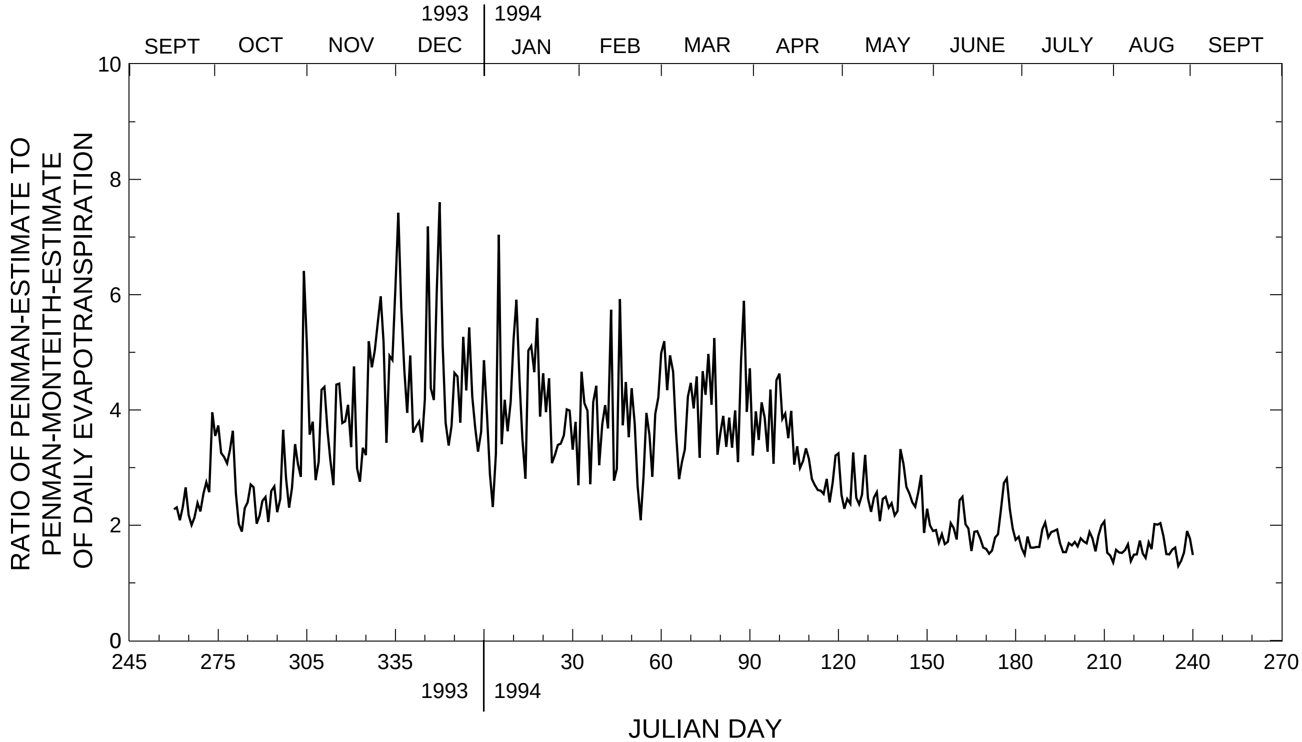


Figure 23. Ratio of Penman-estimate of daily evapotranspiration to Penman-Monteith estimate of daily evapotranspiration at study site. Note: Estimated Penman-Monteith evapotranspiration is based on calibration (method 2) of the Penman-Monteith equation to evapotranspiration values measured by the eddy correlation energy-balance Bowen-ratio method.

Evapotranspiration was modeled relatively well with the modified Priestley-Taylor equation (figs. 24 and 25). The results of this model were slightly superior to those of the Penman-Monteith equation; standard error of estimate was 0.77 mm/d (modified Priestley-Taylor) as compared to 0.84 mm/d (Penman-Monteith - method 2). The Priestley-Taylor α was formulated as a function of several environmental variables (PAR, vapor-pressure deficit, soil-moisture content, air temperature, and Julian day) in a fashion similar to that for Penman-Monteith canopy resistance (method 2):

$$\alpha = \left(C_1 R_{PAR} + C_2 \Delta e + C_3 \bar{\theta} + C_4\right)\left(C_5 \sin\left(\omega\left(t - C_6\right)\right) + 1\right) \qquad (27)$$

Initially, wind speed and air temperature were included as independent variables in this expression. However, these variables contributed negligible information to simulation of measured evapotranspiration values and were eliminated from the function for Priestley-Taylor α. The results of estimation of the parameters in equation 27 are shown in table 2. A simplified 4-parameter version of equation 27, which included only soil-moisture content and Julian day as independent variables, produced results comparable to those with the 6-parameter Penman-Monteith canopy-resistance function (method 2)—standard error of estimate was 0.84 mm/d in both cases. These results were unexpected given the more rigorous nature of the Penman-Monteith model as compared to the Priestley-Taylor model, but are not unprecedented (Stannard (1993)). A comparison of simulated and measured Priestley-Taylor α (fig. 26) reveals that simulation of this parameter generally is good. However, similar to simulation of Penman-Monteith canopy-resistance values, substantial mismatches sometimes occurred during times of low turbulent flux.

The poor performance of the Penman-Monteith model with a traditional formulation of canopy resistance (method 1) relative to that of a nontraditional Penman-Monteith model (method 2) or the modified Priestley-Taylor model was unexpected. Possible reasons for the relative differences in model performance might include:

(1) The models neglect possible synergistic interactions of environmental variables on model parameters (Penman-Monteith canopy resistance or Priestley-Taylor α). This neglect is possibly most

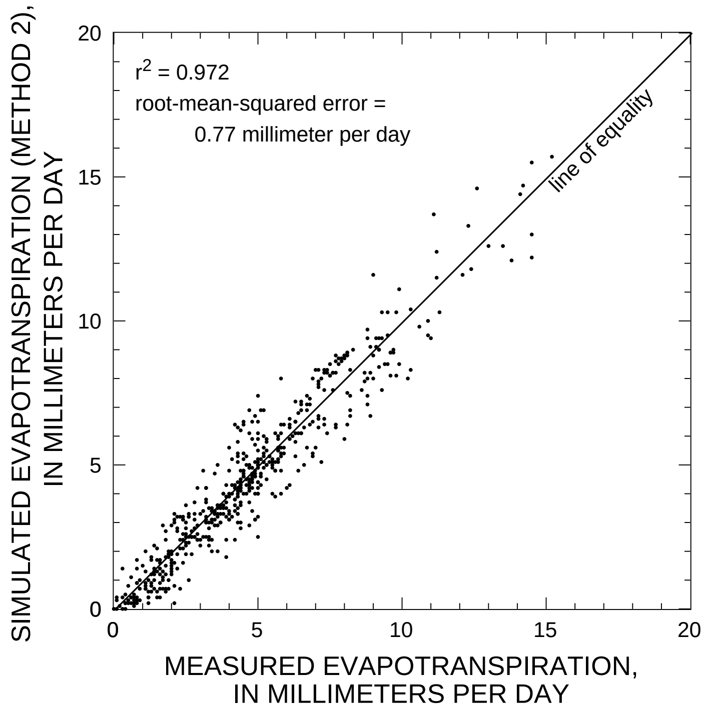


Figure 24. Comparison of Priestley-Taylor-simulated evapotranspiration and measured (eddy correlation energy-balance Bowen ratio method) evapotranspiration at study site from September 1993 to August 1994.

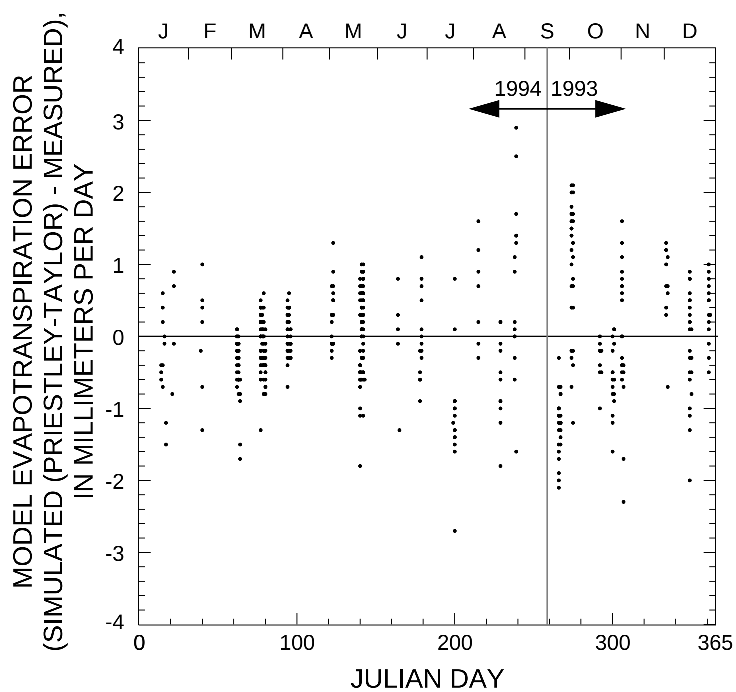


Figure 25. Seasonal distribution of discrepancy between Priestley-Taylor-simulated evapotranspiration and measured (eddy correlation energy-balance Bowen ratio method) evapotranspiration at study site from September 1993 to August 1994.

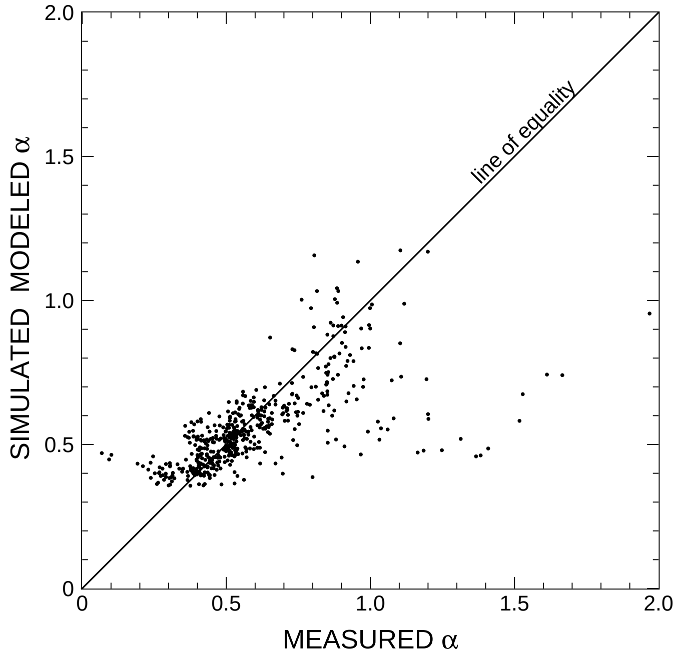


Figure 26. Comparison of simulated and measured values of Priestley-Taylor α at study site from September 1993 to August 1994. Note: Measured values of Priestley-Taylor α computed using values of latent heat flux measured with the eddy correlation energy-balance Bowen ratio method.

apparent in the Penman-Monteith (method 1) model, in which a given change in an environmental variable will produce a constant proportional change in the canopy resistance regardless of the state of the other variables. The structural form of the other models allows for a nonproportional change in the model parameter with a change in an environmental variable, which perhaps more satisfactorily accommodates synergistic interactions.

(2) The models do not explicitly incorporate the seasonally changing relative distribution of species at the site within the model parameter function. Also, although vegetation at the site progresses yearly from juvenile to mature status, the possible effect of these phenological changes on the parameter function of the model are neglected. The “sine” term within the parameter functions for the Penman-Monteith (method 2) and Priestley-Taylor models may somewhat incorporate the effects of these seasonal changes, as well as the changing canopy coverage. Foliage seasonality is incorporated in the Penman-Monteith (method 1) model only through a specified leaf-area index, which considers only changes in canopy coverage.

(3) The model inputs are subject to field measurement error. In particular, estimation of leaf-area index, as part of the Penman-Monteith (method 1) model, is prone to subjectivity.

Values of Penman-Monteith r_c or Priestley-Taylor α demonstrated substantial sensitivity to environmental variables (fig. 27). The Priestley-Taylor model incorporates the effects of plant stomatal sensitivity to environmental variables in a different framework than do the Penman-Monteith models, resulting in a different expression of this sensitivity. Correlation between environmental variables further confounds a unique definition of parameter sensitivity. For these reasons, the functions for r_c and α described in this report should be viewed as a general guide, rather than as a definitive description of stomatal response of site vegetation to environmental variables. However, the fact that the models successfully reproduced 538 measurements of site evapotranspiration over a wide range of seasonal and diurnal values lends credence to the ability of the models to estimate evapotranspiration at the site. This observation is particularly appropriate in the case of interpolation of evapotranspiration between measured values. More caution would need to be observed in the case of extrapolation of the models beyond the measured values of evapotranspiration, or beyond the range of meteorologic conditions used to calibrate the models.

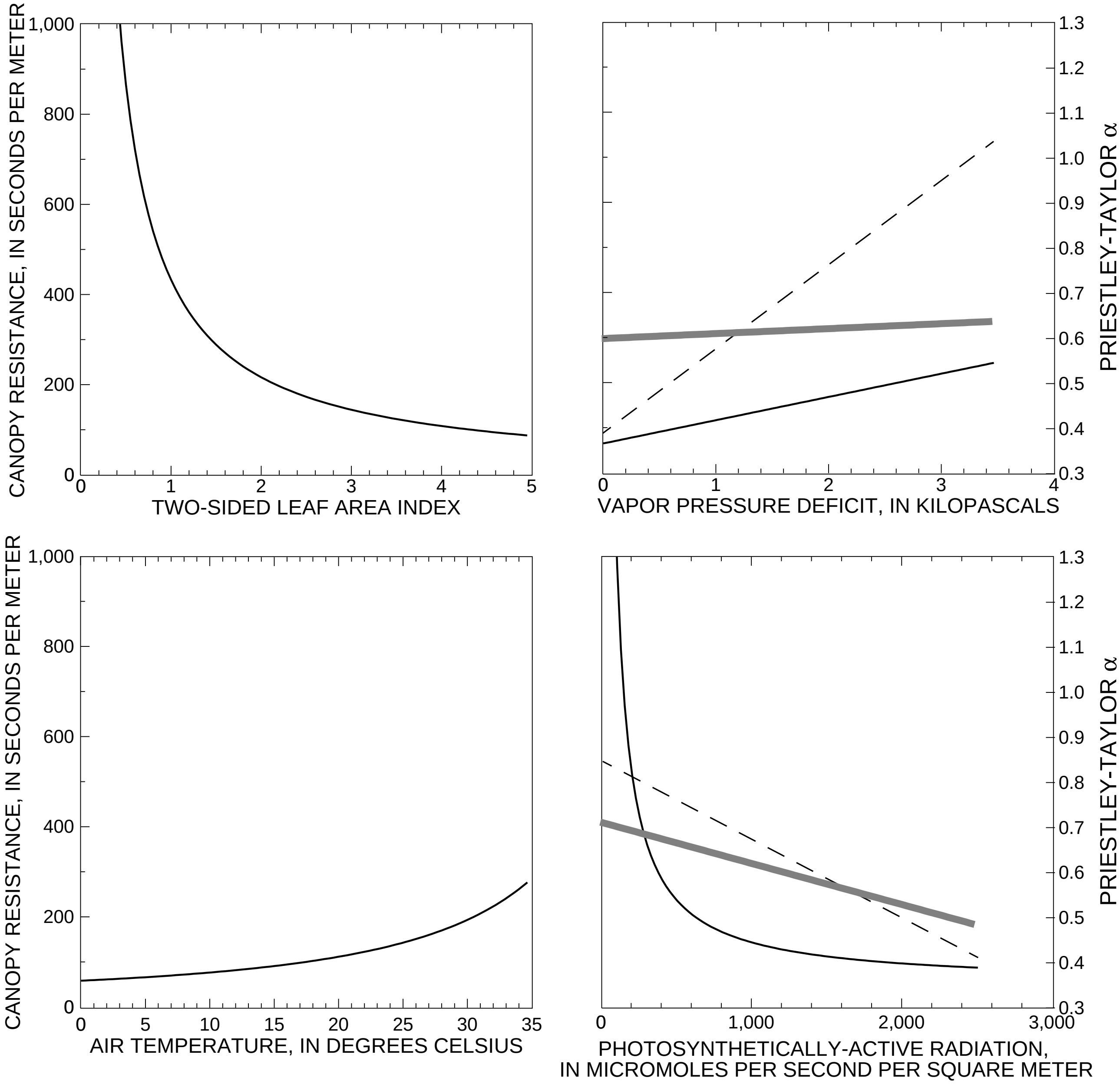


Figure 27. Simulated relations of canopy resistance to environmental variables. Note: Base values are: two-sided leaf area index =3; air temperature = 25 ^{o}C; vapor pressure deficit = 1.5 kPa; photosynthetically-active radiation = 1,000 µmoles/$m^{2}\cdot s$) average moisture content in upper 30 cm of soil = 0.05; Julian day = 100. (Figure 27 continued on next page.)

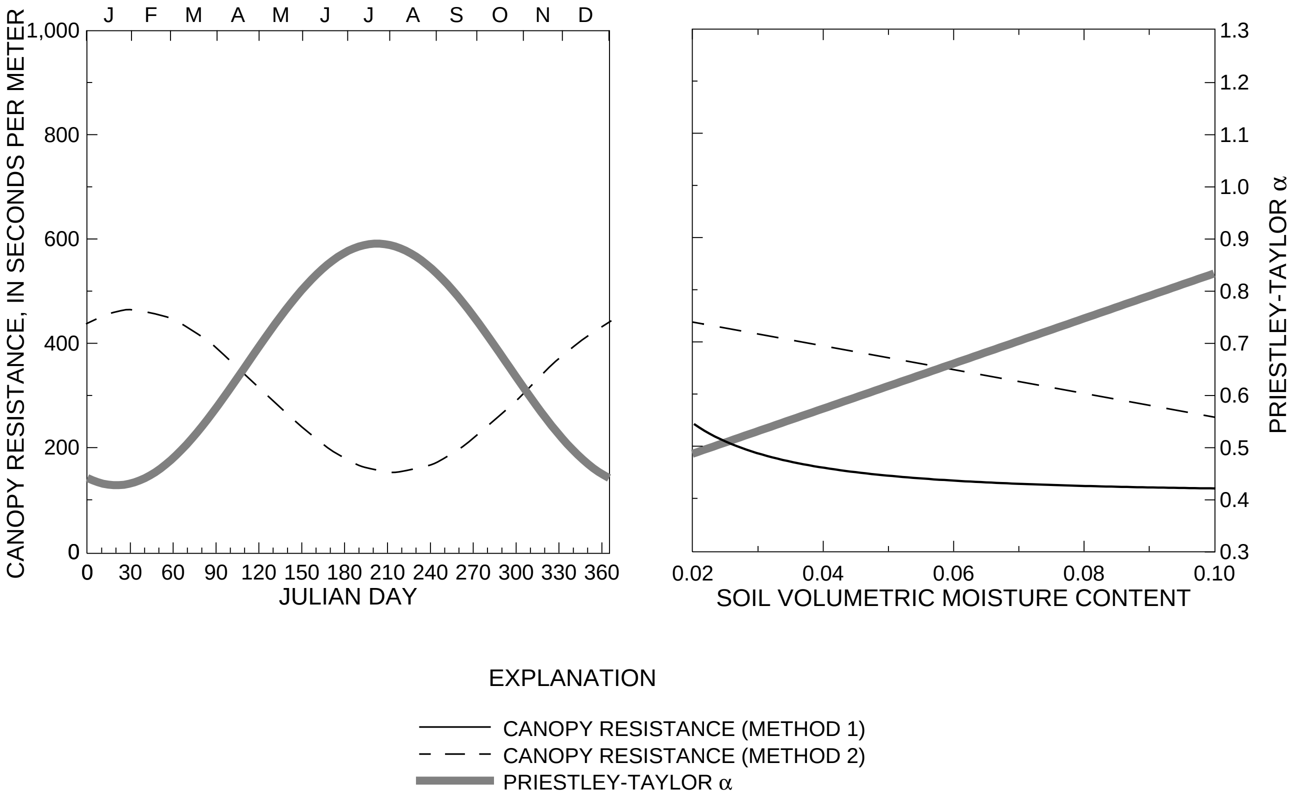


Figure 27. Simulated relations of canopy resistance to environmental variables. Note: Base values are: two-sided leaf area index =3; air temperature = 25 ^{o}C; vapor pressure deficit = 1.5 kPa; photosynthetically-active radiation = 1,000 µmoles/m$^{2\cdot}$s) average moisture content in upper 30 cm of soil = 0.05; Julian day = 100 —Continued.

Application of Evapotranspiration Models

The calibrated evapotranspiration models described in the previous section were applied to the estimation of evapotranspiration, aquifer recharge, partitioning of available energy between latent and sensible heat, Penman-Monteith r_c, and Priestley-Taylor α.

Estimated daily and cumulative evapotranspiration for September 15, 1993, to August 27, 1994, are shown in figures 28 and 29, respectively. The strong seasonal pattern of evapotranspiration is evident in these graphs. Daily evapotranspiration varied over an order of magnitude during the study period (from less than 0.2 mm/d in late December to almost 5 mm/d in mid-July). The cumulative evapotranspiration estimates from this period were linearly extrapolated to estimate annual evapotranspiration. Annual evapotranspiration (September 15, 1993, to September 15, 1994) was estimated to be about 680 mm with either of the more accurate evapotranspiration models (Penman-Monteith (method 2) and modified Priestley-Taylor). Using the less accurate Penman-Monteith (method 1) model, only 560 mm of annual evapotranspiration was estimated. Annual evapotranspiration values estimated using the Penman-Monteith (method 2) model calibrated using the alternatives to the ECEBBR eddy correlation variant were 620 mm (standard eddy correlation method) and 750 mm (ECEBR method). The modified Priestley-Taylor model produced similar alternative estimates of annual evapotranspiration.

The 620 to 750 mm estimated range for evapotranspiration produced by the most successful models (Penman-Monteith method 2 and the modified Priestley-Taylor method) can be used to make a water budget-based estimate of aquifer recharge. The sandy soils of the site readily accept infiltration, producing little overland flow. Therefore, assuming a negligible change in storage in the unsaturated zone during the study period, these data imply that of the 1,320 mm of

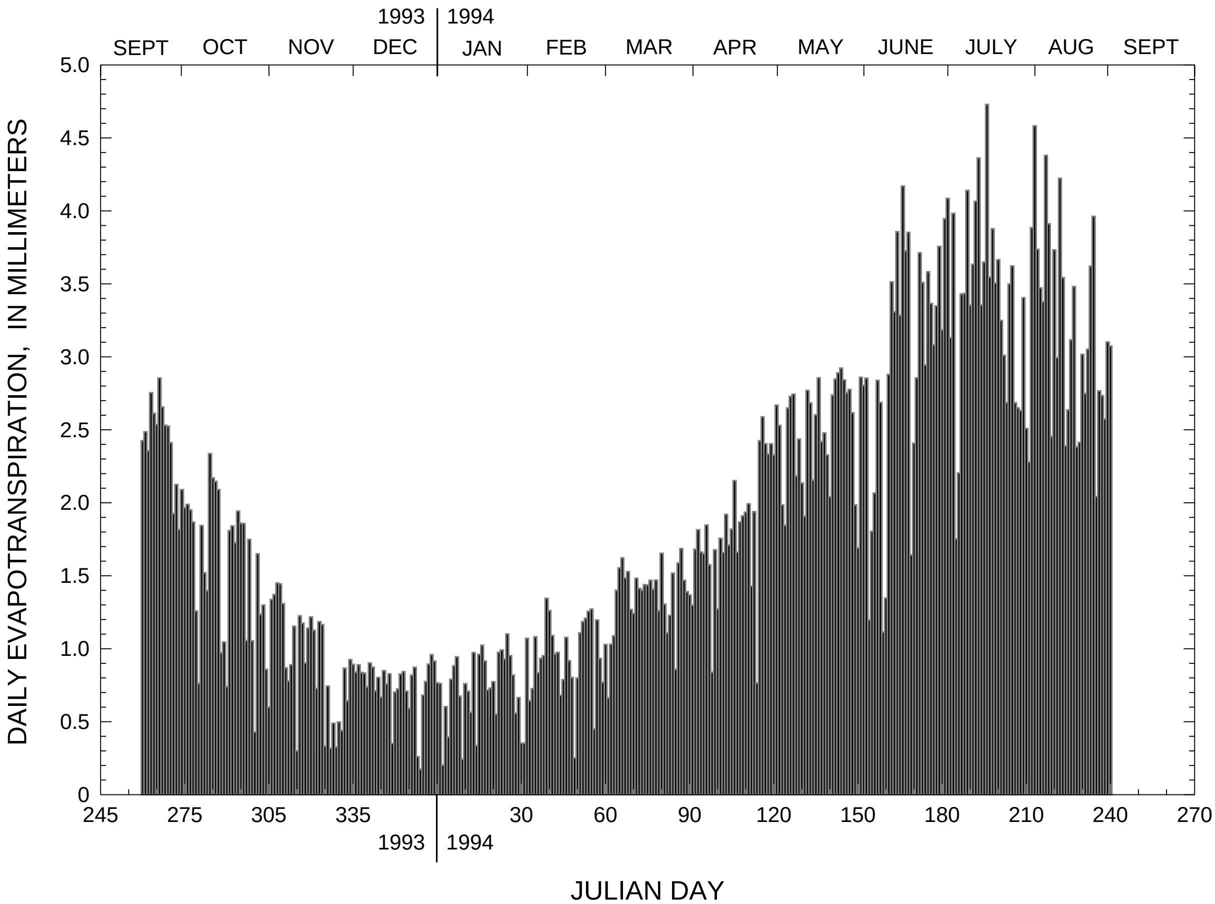


Figure 28. Simulated daily evapotranspiration rate at study site from September 16, 1993, to August 28, 1994. Note: Estimated values based on calibration of Penman-Monteith model (method 2) to evapotranspiration rates measured by the eddy correlation energy-balance Bowen ratio method.

rainfall during the 1-year estimation period about 570 to 700 mm contributed to aquifer recharge. These relative values of evapotranspiration and recharge probably approach the low and high bounds, respectively, for these terms of the water budget in central Florida. The rapidly drained soils, shallow-rooted vegetation, and relatively deep water table of this site are very conducive to minimizing evapotranspiration and maximizing recharge. Results of this investigation indicate that the minimum possible annual evapotranspiration probably is lower for central Florida than that suggested by Tibbals (1990).

Annual recharge estimates may be in error by as much as 20 percent based on the 130-mm error band of annual evapotranspiration determined from the three variants of the eddy correlation method. However, for sites in which evapotranspiration is a higher proportion of precipitation, the method described in this report can produce substantially greater relative error in estimating aquifer recharge. For example, annual evapotranspiration at a marsh site probably approaches the potential value (about 1,200 mm). An error band of 130 mm in estimated evapotranspiration in this case can result in an error of over 100 percent in estimating recharge.

Most of the energy available for turbulent energy fluxes was partitioned as sensible heat from October to May (fig. 30) when there was relatively little canopy coverage (fig. 3) and soil conditions were dry (fig. 4). Most of the available energy was partitioned as latent heat flux during conditions of relatively dense canopy coverage and moist soil from June to September. Therefore, canopy coverage and soil-moisture content act as important constraints on evapotranspiration at the study site.

Site precipitation (figs. 6 and 7) for the study period was about average for the area in both temporal distribution and total magnitude. The distribution and

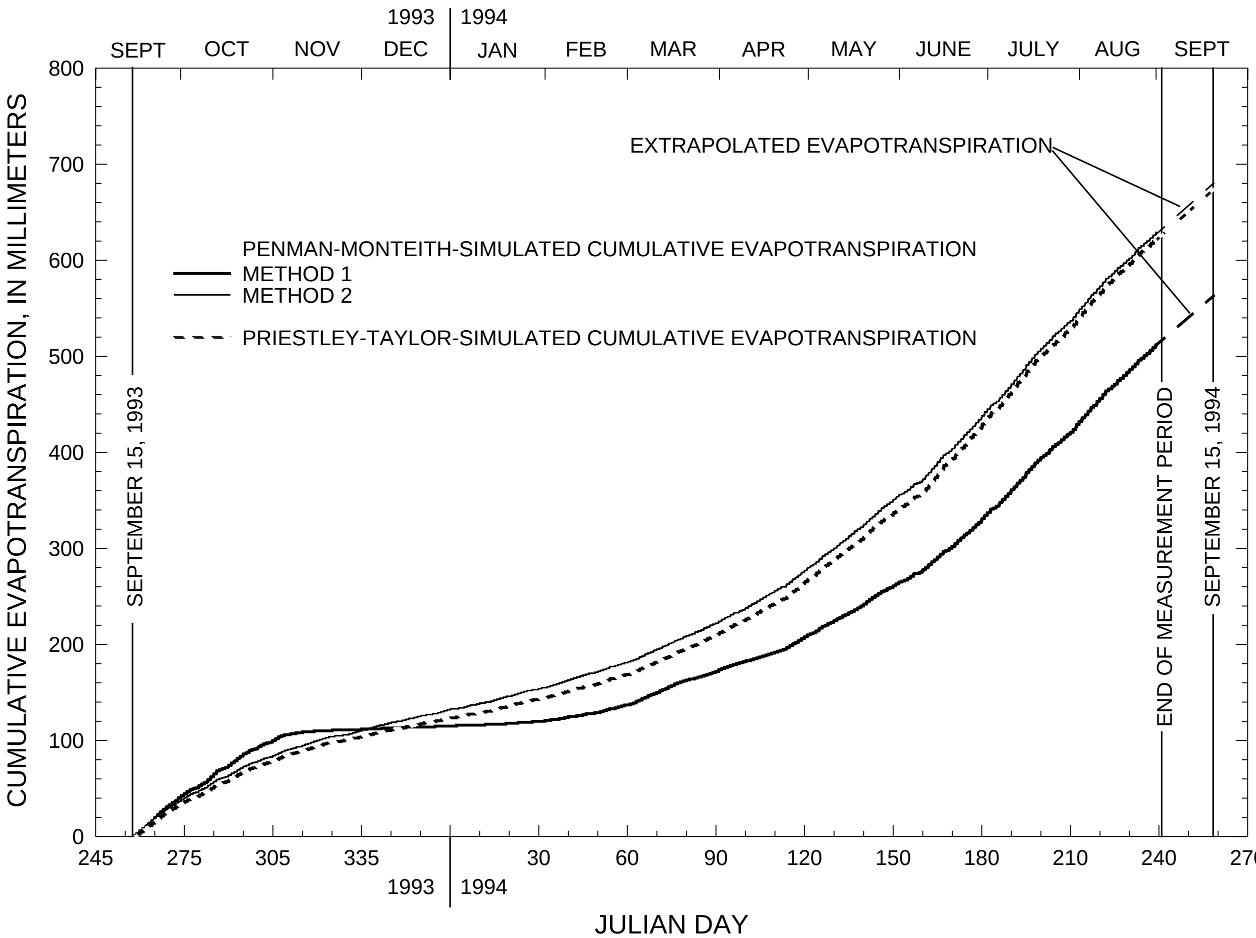


Figure 29. Simulated cumulative evapotranspiration at study site from September 15, 1993, to September 15, 1994. Note: Estimated values are based on calibration of Penman-Monteith and Priestley-Taylor models to evapotranspiration values measured by the eddy correlation energy-balance Bowen ratio method.

magnitude of precipitation can strongly affect soil moisture. Because of the strong dependence of evapotranspiration-model parameters and canopy coverage on soil moisture, the annual value of site evapotranspiration could be higher or lower during wet and dry years, respectively, than that reported here. Also, during a year in which total precipitation is similar to that of the study period, evapotranspiration might be considerably different than that in the study period because of differences in the temporal pattern of rainfall between the two years. A relatively large fraction of the rainfall from a pattern of low-frequency, high-intensity storm events is expected to escape evapotranspiration because much of the water percolating through the sandy soil moves quickly below the shallow rooting zone. Site vegetation is susceptible to stress because of relatively low soil moisture during extended dry periods between storm events, and evapotranspiration is reduced as stomata constrict and canopy coverage decreases. Conversely, a pattern of high-frequency storm events, even those of low intensity, can lead to high evapotranspiration because site vegetation is less susceptible to moisture stress.

Application of the evapotranspiration models for simulation of Penman-Monteith canopy resistance and Priestley-Taylor α is useful for understanding the models. Estimates of canopy resistance (method 2) showed strong seasonal and diurnal fluctuations (figs. 31 and 32). Canopy resistance is relatively low during the wet summer season (June-September) when foliage is dense, soil is moist, and PAR is high. Diurnal fluctuations in canopy resistance generally follow the diurnal pattern of vapor-pressure deficit (fig. 11), increasing from low morning values to a peak value a few hours after midday and decreasing to low values at sunset. The effect of diurnal variations in PAR on estimated canopy resistance is second to that of vapor-pressure deficit and opposite in direction. Canopy-

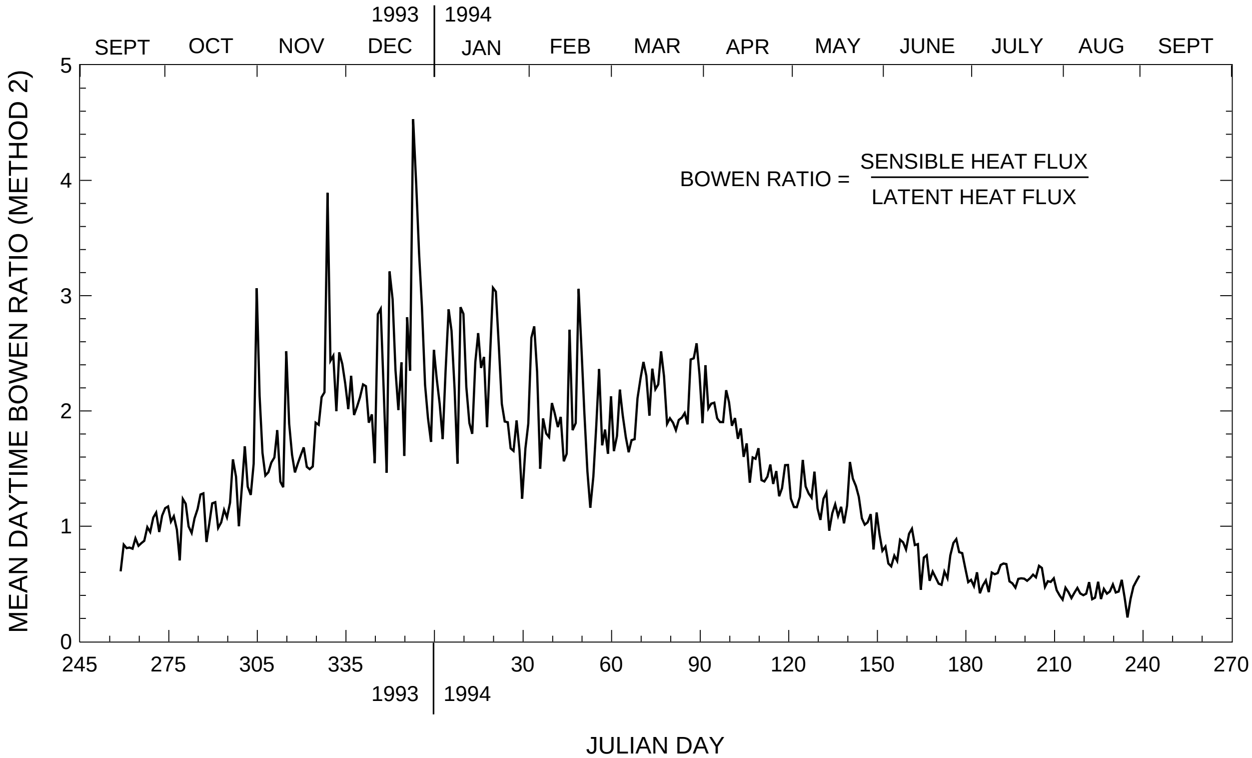


Figure 30. Simulated mean daytime Bowen ratio at study site from September 16, 1993, to August 28, 1994. Note: Estimated values are based on calibration (method 2) of the Penman-Monteith equation to evapotranspiration values measured by the eddy correlation energy-balance Bowen ratio method.

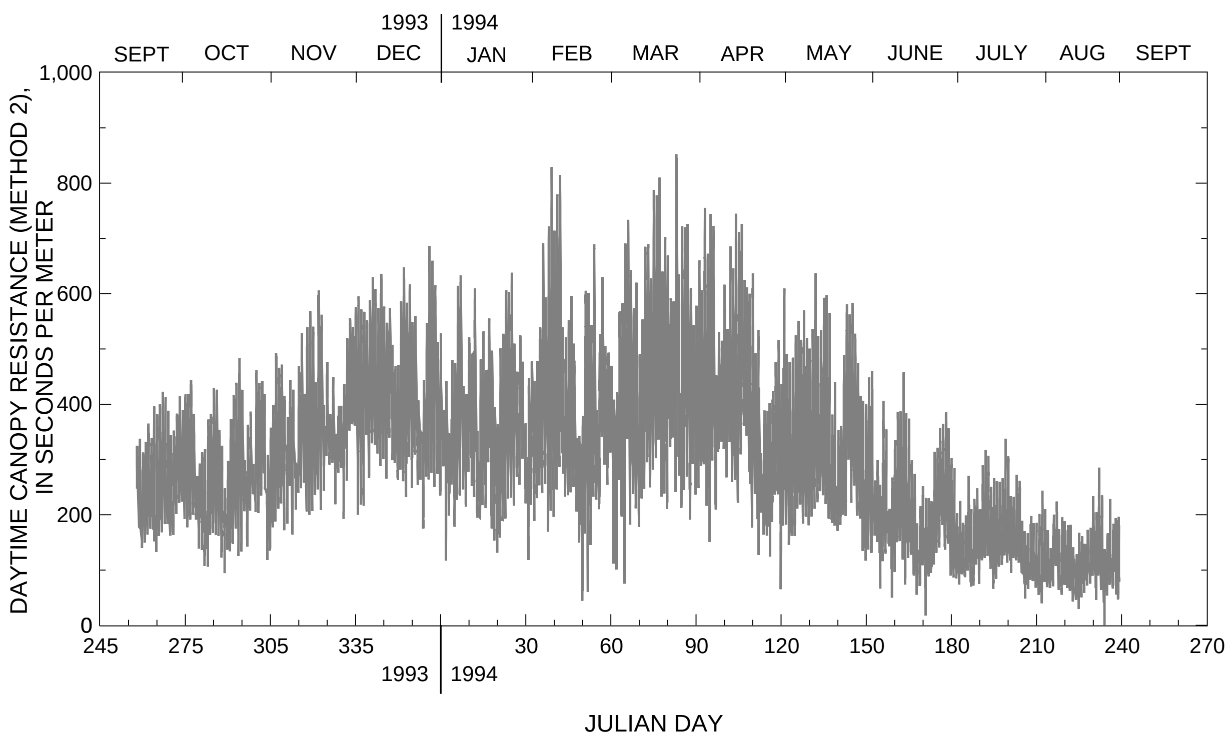


Figure 31. Simulated daytime canopy resistance at study site from September 15, 1993, to August 28, 1994. Note: Estimated values are based on calibration of the Penman-Monteith equation (method 2) to evapotranspiration values measured by the eddy correlation energy-balance Bowen ratio method.

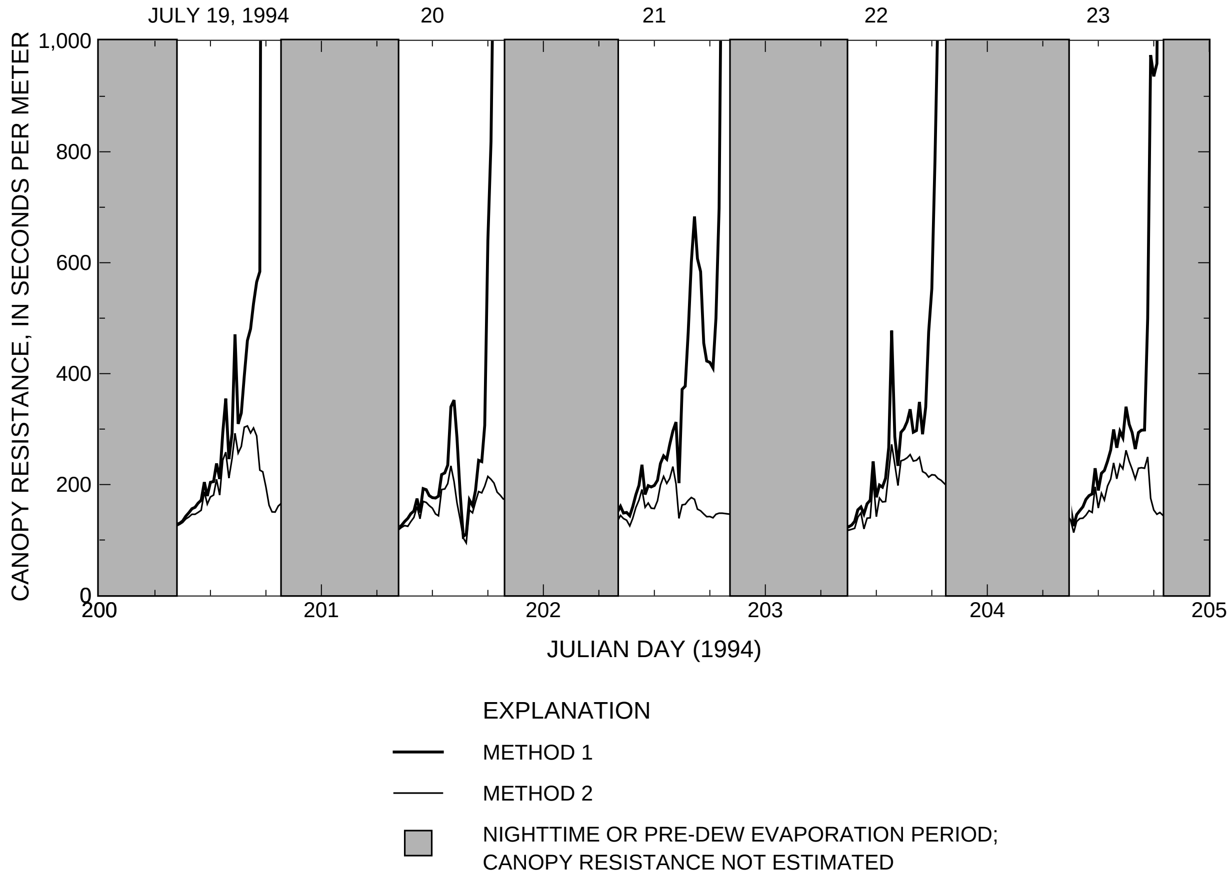


Figure 32. Simulated canopy resistance at study site from July 19-23, 1994.

resistance values simulated by method 1 are similar to those simulated by method 2 during most of the day. However, canopy-resistance values simulated using method 1 rose rapidly as sunset approached, but simulated values using method 2 decreased slightly at this time (fig. 32). A similar effect would be noted in the morning were it not for the exclusion of the early-morning period of dew evaporation from model simulation.

The Priestley-Taylor α generally was well below the empirical value of 1.26 for a well-watered canopy, although for short periods in the wet season (June through August) this value was approached (fig. 33). This parameter was relatively low during the dry season. Diurnal fluctuations in α are exemplified in figure 34. The diurnal pattern of α is “concave up” and responds largely to variations in PAR. Vapor-pressure deficit has only a small effect on estimated Priestley-Taylor α. in contrast to the strong sensitivity evident in the estimated Penman-Monteith canopy resistance.

SUMMARY AND CONCLUSIONS

A study was conducted to: (1) estimate evapotranspiration and aquifer recharge over a 1-year period at a site of successional vegetation in a deforested area of the Lake Wales Ridge, Florida; and (2) to evaluate the suitability of several evapotranspiration models (Penman-Monteith, Penman, and modified Priestley-Taylor) for evapotranspiration simulation. Vegetation at the site consisted of natalgrass, dog fennel, dwarf horseweed, ragweed, and other nonagricultural plants. Site evapotranspiration probably defines the lower limit of evapotranspiration from vegetated surfaces in central Florida because of the shallow-rooted plants, rapidly drained soils, and relatively deep water table. Eddy correlation measurements of evapotranspiration were made during 22 events from September 1993 to August 1994. Each measurement event was about 1 day in length with a temporal measurement resolution of 20 minutes. Eddy correlation measurements were used to “calibrate” evapotranspiration models to estimate evapotranspiration at unmeasured times.

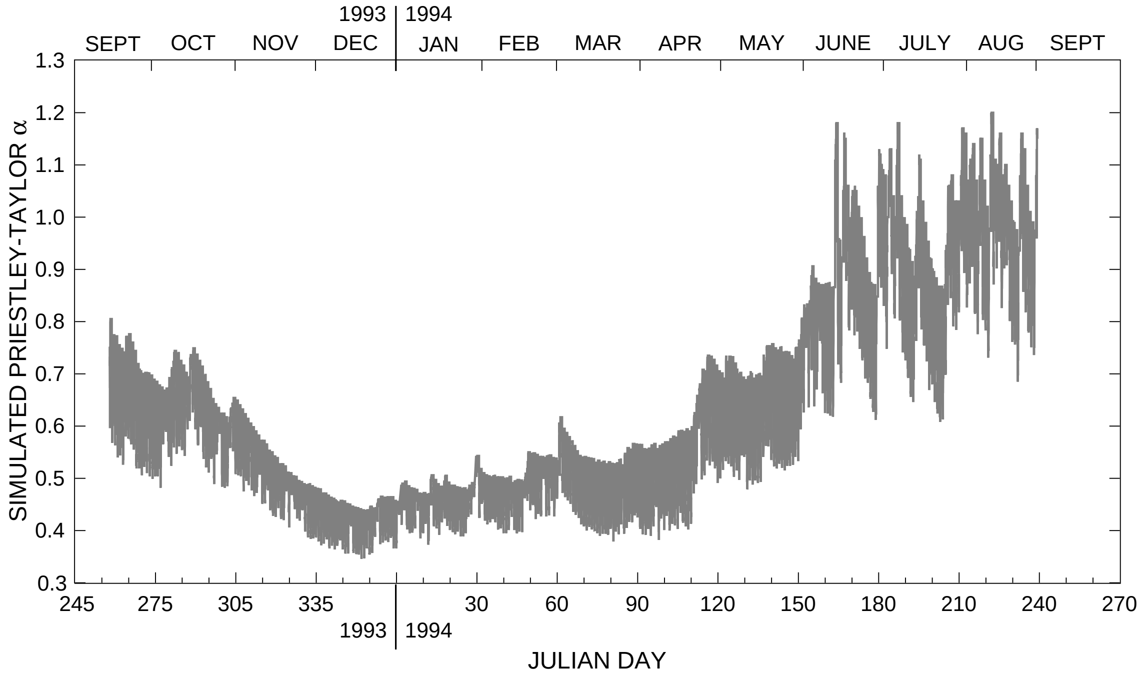


Figure 33. Simulated daytime Priestley-Taylor α at study site from September 15,1993, to August 28, 1994. Note: Estimated values are based on calibration of the Priestley-Taylor equation to evapotranspiration values measured by the eddy correlation energy-balance Bowen ratio method.

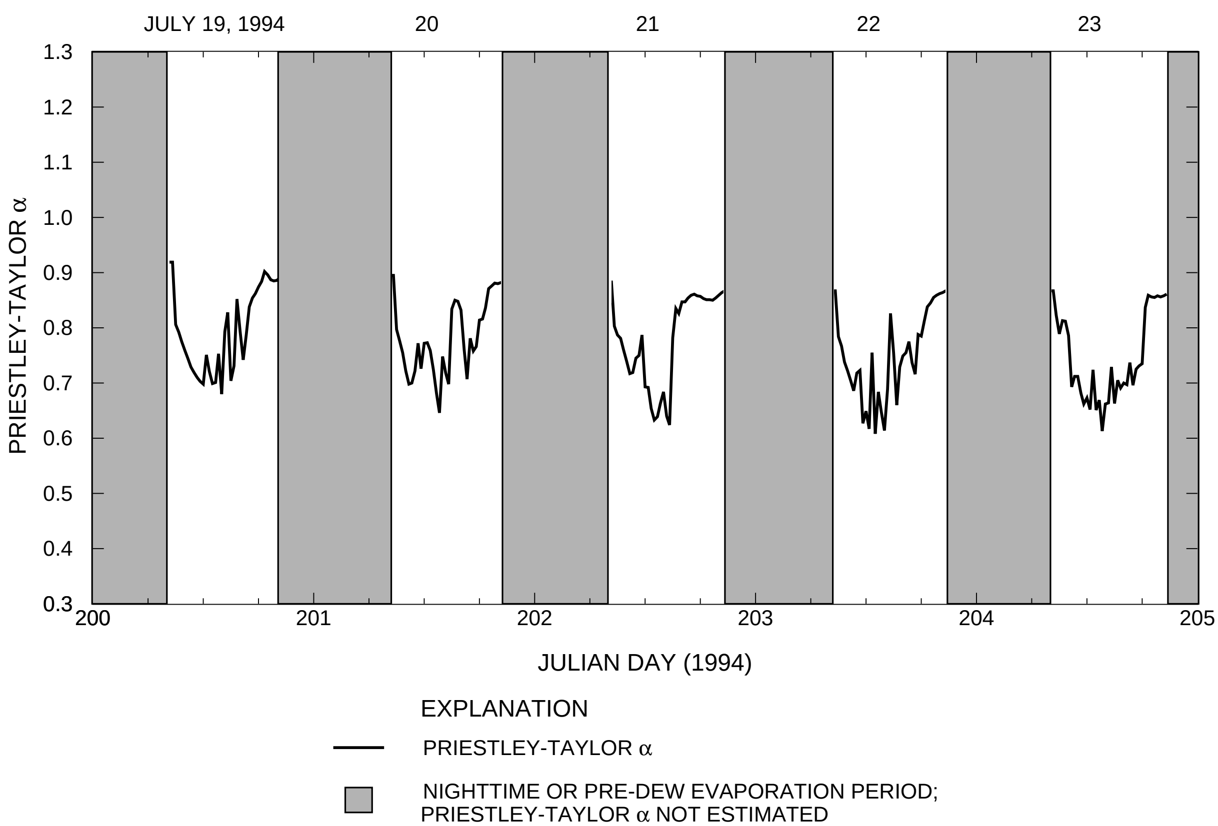


Figure 34. Simulated Priestley-Taylor α at study site from July 19-23, 1994.

The Penman-Monteith and Priestley-Taylor models reproduced the 538 measured values of evapotranspiration very well. A nontraditional functional form of the Penman-Monteith canopy resistance proved superior (r^2 = 0.967; standard error of estimate = 0.84 millimeter per day (mm/d)) to forms used by previous investigators (r^2 = 0.918; standard error of estimate = 1.33 mm/d) in evapotranspiration simulation at the site. This nontraditional approach conceptualized canopy resistance as a simple linear function of photosynthetically active radiation, vapor-pressure deficit, mean moisture content in the rooting zone, and a seasonal term intended to incorporate changes in canopy coverage, relative species density, and plant phenological stages. A simple modified Priestley-Taylor model simulated evapotranspiration slightly better (r^2 = 0.972; standard error of estimate = 0.77 mm/d) than did the best Penman-Monteith model. Thus, the simple Priestley-Taylor model would be preferable to the Penman-Monteith model for general usage at sites similar to the study site because the Priestley-Taylor model simulates evapotranspiration well, is computationally much simpler than the Penman-Monteith model, and does not require collection of wind speed and relative humidity data.

The Penman model of evapotranspiration was a poor predictor of measured evapotranspiration at the study site. No relation was evident between Penman-simulated evapotranspiration and measured evapotranspiration. As expected, the discrepancy between this model and measured values was most extreme when canopy coverage and soil moisture were relatively low.

Daily evapotranspiration rates (simulated by nontraditional Penman-Monteith and Priestley-Taylor models calibrated to a Bowen ratio variant of the eddy correlation method) varied seasonally, ranging from about 0.2 mm/d in late December 1993 to about 5 mm/d in mid-July 1994. Most of the available energy for turbulent energy fluxes is distributed to sensible heat during times of relatively sparse canopy coverage and dry soil (generally October-May). Most of the available energy is dissipated as latent heat flux during times of relatively dense canopy coverage and moist soil (generally June-September). Annual evapotranspiration (September 15, 1993, to September 15, 1994) was estimated to be about 680 millimeters (mm). Evapotranspiration models, calibrated to the standard eddy correlation method and to an energy-balance residual variant, provided estimates of evapotranspiration that were about 10 percent lower and higher, respectively, than the above estimates provided by the Bowen ratio variant.

Measured precipitation over the study period (1,320 mm) was about equal to mean annual precipitation for the area (1,300 mm). Neglecting changes in storage of water above the water table and assuming negligible surface runoff, the measured data imply that about 570 to 700 mm of recharge to the surficial aquifer occurred over the 1-year study period. However, the dependence of evapotranspiration-model parameters on environmental variables (particularly soil moisture content) suggests the possibility that evapotranspiration during years of nontypical rainfall might vary from that measured in this study.

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