Evapotranspiration from Areas of Native Vegetation in West-Central Florida

By W.R. Bidlake, W.M. Woodham, and M.A. Lopez

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CONVERSION FACTORS, VERTICAL DATUM, ACRONYMS, AND ABBREVIATIONS

Multiply	By	To obtain
	Length	
nanometer (nm)	0.00003937	mil
micrometer (µm)	0.03937	mil
millimeter (mm)	0.03937	inch
centimeter (cm)	0.3937	inch
meter (m)	3.281	foot
	Length per unit time	
millimeter per second (mm/s)	0.03937	inch per second
millimeter per day (mm/d)	0.03937	inch per day
millimeter per year (mm/yr)	0.03937	inch per year
meter per second	3.281	inch per second
	Area	
hectare (ha)	2.471	acre
	Energy per unit mass	
joule per gram (J/g)	0.2388	calorie per gram
	Specific heat	
joule per gram per degree Celsius (J/g•°C)	0.1327	calorie per gram per degree Fahrenheit
joule per kilogram per degree Celsius (J/kg•°C)	0.1327	calorie per kilogram per degree Fahrenheit
	Energy per unit area	
megajoule per square meter (MJ/m^2)	23.88	calorie per square centimeter
watt per square meter (W/m^2)	0.001433	calorie per square centimeter
	Pressure	
kilopascal (kPa)	0.2953	inches of mercury
	0.1450	pound per square inch
	10.0	millibar
	Mass per unit area	
kilogram per square meter (kg/m^2)	0.2049	pound per square foot
	Mass per unit volume	
gram per cubic centimeter (g/cm^3)	0.03613	pound per cubic inch
gram per cubic meter (g/m^3)	0.00006245	pound per cubic foot
	Miscellaneous	
kilogram per kilogram (kg/kg)	1.0	pound per pound
cubic meter per gram per centimeter ($m^3/g{\bullet}cm$)	40,680	cubic foot per pound per inch
meter per second per volt (m/s•V)	3.281	foot per second per volt
gram per square meter per second ($g/m^2{\bullet}s$)	0.0002049	pound per square foot per second

For temperature conversions from degrees Celsius (°C) to degrees Fahrenheit (°F), use the following:

1.8 • °C + 32 = °F

1.8 • degrees Celsius per volt (°C/V) + 32 = degrees Fahrenheit per volt

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD 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.

ACRONYMS AND ABBREVIATIONS

DBH	diameter at breast height
ECEBBR	eddy correlation energy-balance Bowen ratio
ECEBR	eddy correlation energy-balance residual
EBBR	energy-balance Bowen ratio
EBWSP	energy-balance wind and scalar profile
Hz	hertz
sd	standard deviation

SYMBOLS

Symbol	Meaning	Dimension
C	Energy-balance closure	dimensionless
C_m	Specific heat of biomass	J/kg•°C
C_p	Specific heat of air at constant pressure	J/g•°C
C_s	Specific heat of soil particles	J/g•°C
C_w	Specific heat of water	J/g•°C
D	Roughness element spacing	m
d	Zero plane displacement height	m
E	Evapotranspiration flux	$g/m^2•s$
E_p	Potential evapotranspiration flux	$g/m^2•s$
e	Vapor pressure	kPa
e_s	Saturation vapor pressure	kPa
G	Subsurface heat flux	W/m^2
G_z	Subsurface heat flux at a fixed depth	W/m^2
H	Sensible heat flux	W/m^2
h	Canopy or instrument height, above ground	m
K_o	Effective attenuation coefficient for oxygen	$m^3/g•cm$
K_w	Effective attenuation coefficient for water vapor	$m^3/g•cm$
k	von Karman constant	dimensionless
L	Latent heat of vaporization for water	J/g
LE	Latent heat flux	W/m^2
M_b	Canopy biomass	kg/m^2
n	Sample size	dimensionless
R_n	Net radiation	W/m^2
r	Correlation coefficient	dimensionless
r_h	Boundary layer resistance	s/m
r^2	Coefficient of determination	dimensionless
S	Rate of canopy heat storage	W/m^2
S_b	Rate of biomass heat storage	W/m^2
S_l	Rate of latent heat storage	W/m^2
S_s	Rate of sensible heat storage	W/m^2
$S_{Y•X}$	Standard error of regression	
s	Slope of the saturation vapor-pressure function	kPa/°C
T	Air temperature	°C
T_b	Biomass temperature	°C
T_k	Absolute temperature	K
T_s	Soil temperature	°C
T_w	Water temperature	°C
t	Time	s
u	Horizontal windspeed	m/s
V	Krypton hygrometer output voltage: proportional to attenuated radiation flux	volts
V_o	Krypton hygrometer output voltage: proportional to unattenuated radiation flux	volts
w	Vertical windspeed	m/s

SYMBOLS—Continued

Symbol	Meaning	Dimension
x	Krypton hygrometer path length	cm
z	Soil or water depth	m
z_h	Roughness length for heat transport	m
z_m	Roughness length for momentum transport	m
z_r	Height at which windspeed is measured	m
β	Bowen ratio	dimensionless
γ	Psychrometer constant	kPa/°C
ρ_a	Air density	g/m^3
ρ_b	Soil bulk density	g/m^3
ρ_v	Vapor density	g/m^3
ρ_w	Water density	g/m^3
σ_{ρ_v}	Standard deviation of vapor density	g/m^3
σ_T	Standard deviation of air temperature	°C
σ_w	Standard deviation of vertical windspeed	m/s
θ	Gravimetric soil water content	dimensionless
$'$	Momentary fluctuation from the mean	
$\overline{}$	Time average	
Δ	Change with time or difference across space	

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By W.R. Bidlake, W.M. Woodham, *and* M.A. Lopez

Abstract

A study was conducted to evaluate the suitability of three micrometeorological methods for estimating evapotranspiration from selected areas of native vegetation in west-central Florida and to estimate annual evapotranspiration from areas having a specific vegetation type. Evapotranspiration was estimated using the methods of energy-balance Bowen ratio (EBBR) and eddy correlation. Potential evapotranspiration was computed using the Penman equation. Field measurements were made intermittently from February 1988 through September 1990.

The EBBR method was used to estimate evapotranspiration from unforested and forested sites. A mean-gradient Bowen ratio system was used to measure and average vertical air temperature and vapor-pressure gradients, and the Bowen ratio was computed using the mean air temperature and vapor-pressure gradients. The Bowen ratio estimated in this manner was then used to compute evapotranspiration by the EBBR method. Computations based on objective review criteria indicated that the Bowen ratio, computed using measurements made using the mean-gradient Bowen ratio system, was not always realistic. During a period of extended operation at a dry prairie site, 9 percent of measured available energy during the daytime occurred when the Bowen ratio obtained using the mean-gradient Bowen ratio system was unrealistic. During 5 out of 14 days of continuous operation at a marsh site, more than 30 percent of measured available energy during the daytime occurred when the Bowen ratio obtained using the mean-gradient Bowen ratio system was unrealistic. One of the primary causes of unrealistic Bowen ratios at the unforested sites was condensation of moisture within the tubing of the mean-gradient Bowen ratio system. Measurements made using the mean-gradient Bowen ratio system at a forested pine flatwood site indicated that vapor-pressure gradients were too weak to be resolved by the system. As a result, the Bowen ratio computed for the forested sites was unreliable when it was obtained using the mean-gradient Bowen ratio system.

Direct estimates of sensible and latent heat flux that were computed from eddy correlation measurements were generally insufficient to account for measured available energy at all sites. Analysis of eddy correlation and energy-balance data indicated that the sum of sensible and latent heat fluxes accounted for 68 percent of available energy at dry prairie and marsh sites, 74 percent of available energy at a pine flatwood site, and 45 percent of available energy at a cypress swamp site. Because specific causes of the energy-balance discrepancies could not be quantified, corrections to the direct eddy correlation flux estimates could not be made, and eddy correlation data were combined with other energy-balance data to yield two alternative evapotranspiration estimates. The first alternative evapotranspiration estimate was computed by combining sensible heat flux obtained from eddy correlation with measurements of available energy to compute latent heat flux as the

residual of the equation for the surface energy balance. The second alternative evapotranspiration estimate was computed by using direct sensible and latent heat flux estimates that were obtained from eddy correlation measurements to compute the Bowen ratio. The Bowen ratio obtained from eddy correlation measurements was then combined with measurements of available energy to compute evapotranspiration by the EBBR method. Of the three alternative evapotranspiration estimates obtained from eddy correlation measurements, the estimate computed using the EBBR method, with the Bowen ratio computed from eddy correlation measurements, agreed most strongly with the corresponding evapotranspiration estimate computed using the EBBR method with the Bowen ratio obtained from the mean-gradient Bowen ratio system. It is probable that actual evapotranspiration was within a range defined by the standard eddy correlation computation, which consistently indicated the smallest evapotranspiration, and the energy-balance residual computation, which consistently indicated the largest evapotranspiration.

Daily potential evapotranspiration, as computed by the Penman method, and daily evapotranspiration, as computed by the EBBR method, did not seem to correlate with each other at a dry prairie site during late spring and summer; however, the two were correlated with each other at a marsh site during late spring and summer. Evapotranspiration was approximately 57 percent of potential evapotranspiration at the marsh site. The correlation between evapotranspiration and potential evapotranspiration at the marsh site, and the fact that evapotranspiration approached potential evapotranspiration, indicated that the Penman method can be useful for estimating evapotranspiration from marshes in west-central Florida.

Annual evapotranspiration estimates were developed for each vegetation type by pooling EBBR and eddy correlation measurements among sites and among the 3 years during which field measurements were made. Three different estimates, which correspond to the three eddy correlation computation methods, were made for each vegetation type. The centric estimates, which were calculated by using the EBBR method with the Bowen ratio obtained from either a mean-gradient system or from eddy correlation measurements, were 1,010 millimeters per year for the dry prairie type, 990 millimeters per year for the marsh vegetation type, 1,060 millimeters per year for the pine flatwood type, and 970 millimeters per year for the cypress swamp type.

INTRODUCTION

Quantification of the major components of the hydrologic budget is essential when planning for the development of water resources of an area, estimating water-supply potential, and understanding the ecological effects of development. Demand for water supply in west-central Florida continues to increase; however, the nature of hydrologic budgets in areas where water supplies are being developed remains uncertain.

Basinwide studies have demonstrated that evapotranspiration is second only to precipitation in importance to terrestrial hydrologic budgets of Florida (Jones and others, 1984). Evapotranspiration can vary considerably among basins that exhibit different types of vegetative cover or different proportions of open-water surfaces. This variation has implications for water-supply potential because surface-water and ground-water supplies are affected by the amount of evapotranspiration. Little work has been done to describe evapotranspiration for different types of native vegetation in Florida.

This study was conducted in cooperation with Sarasota County government, the Southwest Florida Water Management District, and the West Coast Regional Water Supply Authority. These agencies are aware of the need to define hydrologic budgets in conjunction with resource development. This study was conducted to examine the suitability of selected techniques for estimating evapotranspiration from selected types of native vegetation and to estimate evapotranspiration from the vegetation types.

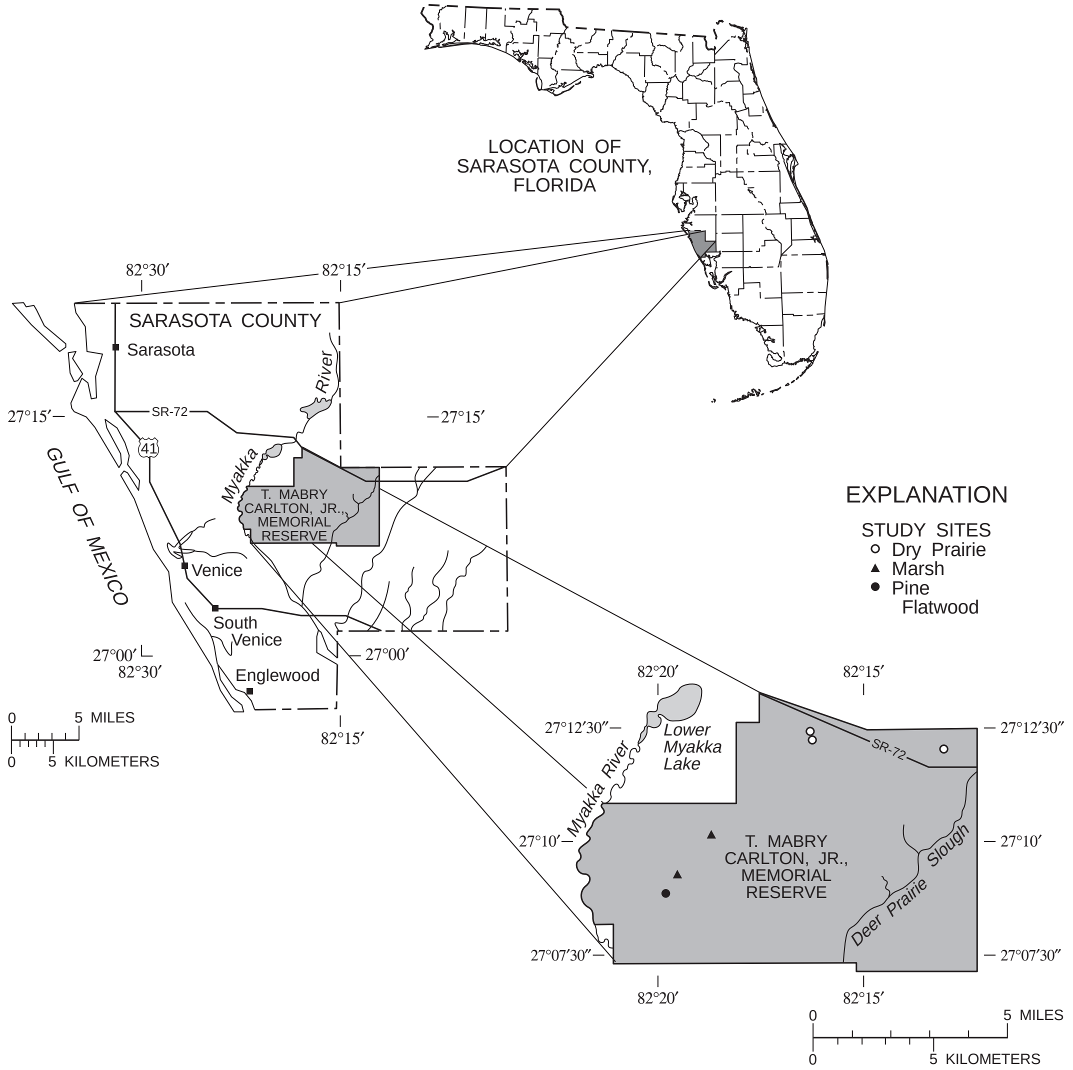

Figure 1. Location of study area and study sites on the T. Mabry Carlton, Jr., Memorial Reserve, Sarasota County.

Purpose and Scope

The purpose of this report, which presents the results of a 5-year study, is to examine the theory, application, and suitability of three micrometeorological methods for determining evapotranspiration and to estimate annual evapotranspiration from selected areas within four types of native vegetation. The report presents and interprets results from evapotranspiration measurements that were based on the micrometeorological methods of energy-balance Bowen ratio (EBBR) and eddy correlation and from measurements of potential evapotranspiration that were made using the Penman method. The native vegetation types studied were dry prairie, marsh, pine flatwood, and cypress swamp.

Description of the Study Area

The study was conducted in two coastal counties in west-central Florida—Sarasota and Pasco Counties, both of which are in the Coastal Plain Province (Brooks, 1981). In general, the near-surface stratigraphy is limestone rock that lies beneath a thin layer of surficial sand. The Sarasota County study area is the T. Mabry Carlton, Jr., Memorial Reserve (Carlton Reserve; fig. 1). The Carlton Reserve lies on a sloping plain that ranges in altitude from 10 to 30 m above sea level. The plain is traversed by the Myakka River and numerous sloughs that drain to the Gulf of Mexico. The Pasco County site has an altitude of less than 30 m above sea level and lies on a sloping plain that drains to the Gulf of Mexico (fig. 2).

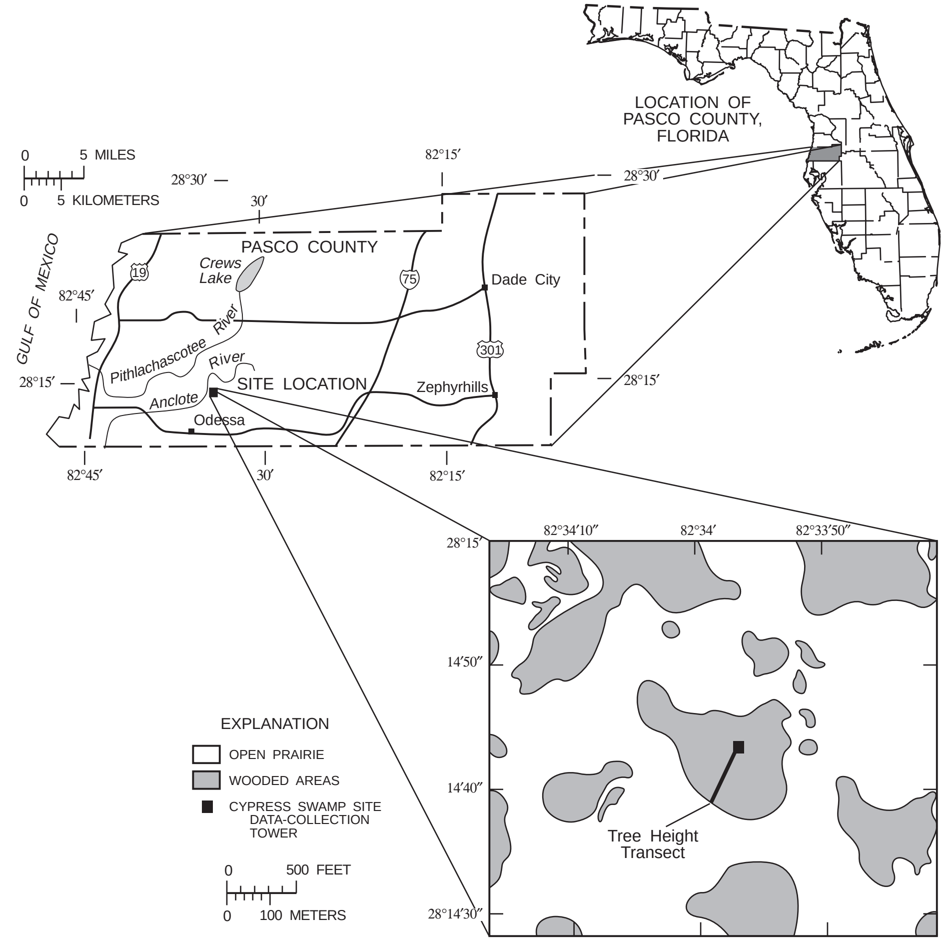

Figure 2. Location of the cypress swamp study site, Pasco County.

Climate in the study area is humid subtropical. Summers are warm and moist; winters are warm and dry. Mean annual and mean monthly air temperatures are similar for both study areas. Mean annual temperature at Tarpon Springs near the Pasco County site was 22°C for 1951–80 (National Oceanic and Atmospheric Administration, 1982). The mean June temperature was 27°C, and the mean December temperature was 16°C. Diurnal temperature variation is modest, typically 5–6°C in summer and winter. Mean annual precipitation at Tarpon Springs was 1,315 mm for 1951–80. Mean annual precipitation at the Myakka River State Park, which borders the Carlton Reserve, was 1,440 mm for 1951–80. About 50 percent of the annual rainfall in each county occurs during the months of June through September.

Vegetation and Soils

Study sites were established within three vegetation types on the Carlton Reserve (fig. 1). The dry prairie, marsh, and pine flatwood vegetation types were identified from a habitat type classification system that has been adopted by the government of Sarasota County (Biological Research Associates, 1986). A cypress swamp in Pasco County was also selected for this study (fig. 2).

The dry prairie vegetation type is common on upland soils where the water table is rarely near or at the land surface. Dry prairie vegetation is dominated by saw palmetto (*Serenoa repens*) and wax myrtle (*Myrica cerifera*), although as many as 45 species can be present (Biological Research Associates, 1986). Many species exhibit an evergreen habit. The treeless state of the prairies probably results from historically short return intervals of wildfire. Saw palmetto has adapted to survive wildfire; the plant has a massive root system and sprouts prodigiously after fire.

Soils of the dry prairie vegetation type are typically EauGallie fine sand (sandy, siliceous, hyperthermic Alfic Haplaquod) or Myakka fine sand (sandy, siliceous, hyperthermic Aeric Haplaquod).

These poorly drained mineral soils are found on nearly level terrain, and the solums of the two soils consist of fine sands. The thickness of the EauGallie solum is 130 cm or more, and the thickness of the Myakka solum is 150 cm or more. Depths to the water table seasonally range from 20 to more than 100 cm below the soil surface.

Three study sites were established in the dry prairie vegetation type (fig. 1). The first site was used from February 1988 until it was burned in a wildfire in April 1989. The second site was used from May 1989 until it was burned in March 1990. The third site was used from March to September 1990. Vegetation at the first two sites was dominated by saw palmetto and widely scattered wax myrtle. The wax myrtle at the first two sites were about 2 m in height; the saw palmetto were about 0.7 m in height. Vegetation at the third site was dominated by saw palmetto that were approximately 0.7 m in height.

Marsh and slough vegetation types occur in areas that are partially flooded for most of the year and are common throughout the Carlton Reserve. Depth of standing water increases systematically from near the edge to the center of these wetlands. The duration of annual floods and the average depth of water are thought to be major controls of floral composition in these wetlands.

Vegetation zones form roughly concentric bands about the center of the wetlands. Major vegetation zones are (starting from the upland border) *Hypericum*, *Panicum-Rhynchospora*, mixed emergent, and *Cladium*. The *Hypericum* zone is dominated by the evergreen shrub St.-Johnswort (*Hypericum fasciculatum*). The *Panicum-Rhynchospora* zone is dominated by the grasses maidencane (*Panicum hemitomon*), redtop panicum (*Panicum* spp.), and Tracy's beakrush (*Rhynchospora tracyi*). The mixed emergent zone is dominated by pickerelweed (*Pontederia cordata*) and arrowhead *(Sagittaria lancifolia*). The *Cladium* zone is located near the center of the larger, deeper wetlands; sawgrass (*Cladium jamaicensis*) is the dominant species in that zone.

Vegetative growth, senescence, and death at marsh sites exhibit seasonal variations that seem to be related to fluctuations of water levels in the marshes. A vegetation study, which included measurement of percent plant cover, on the Carlton Reserve during 1985 by CH2M Hill, Inc. (1988a), indicated that average percent cover in the *Hypericum* zone of 12 study wetland areas increased from 22 percent in May to 65 percent in August, remained nearly constant until September, and then increased to 75 percent in December. Other vegetation zones in the study areas had similar patterns of canopy development. Water levels in the study wetlands were lowest in April, May, or June; increased sharply in late June with the onset of the summer rainy season; and decreased slightly from October through December (CH2M Hill, Inc., 1988b). Similar patterns of vegetative growth and water level in marsh sites were observed during this study. It was noted that the plant canopies at the marsh sites had a high proportion of senescent and dead standing stems in them during periods of low water levels.

Substrate characteristics in the wetlands vary in a concentric pattern similar to that of the vegetation zones. Sandy soils, such as Delray fine sands (loamy, siliceous, hyperthermic Grossarenic Argiaquoll), are common in the *Hypericum* and *Panicum-Rhynchospora* zones. The interior vegetation zones are underlain by a peat layer.

Two marsh sites were used during the study (fig. 1). The first site was located in the *Cladium* vegetation zone near the center of a large (17 ha) marsh. Measurements were made at this site from August 1988 to June 1990. The second site was in a much smaller (4 ha) marsh. Energy-balance Bowen ratio and eddy correlation measurements were made at this site from November 1988 to May 1989.

The pine flatwood vegetation type is found extensively throughout the Carlton Reserve on upland soils where the water table is rarely at or above land surface. Overstory vegetation is dominated by slash pine (*Pinus elliottii*), and understory vegetation is dominated by saw palmetto and wax myrtle. Canopy height and stand density of the slash pine overstory are highly variable. In many areas, the slash pines are scattered as isolated individuals that range in height from about 2 to 12 m.

Soils of the pine flatwood vegetation type are EauGallie and Myakka fine sands and Pineda fine sand (loamy, siliceous, hyperthermic Arenic Glossaqualf). Pineda fine sands are nearly level and poorly drained. The solum is 100–200 cm thick and consists of fine sands. The depth to the water table seasonally ranges from 30 to more than 100 cm below the soil surface.

Vegetation at the pine flatwood site (fig. 1) consisted primarily of an overstory of slash pine and an understory of saw palmetto. The slash pine overstory was sparse. Twelve circular plots (0.04 ha) were used to sample stand characteristics of the slash pine overstory in each cardinal direction for a distance of 250 m from a tower on which study instruments were mounted. Averaged between plots, mean tree height was 7.8 m (standard deviation (sd) 3.1 m), mean maximum tree height was 10.1 m (sd 2.7 m), mean stand density was 96 trees per hectare (range 25–250), and mean tree diameter at breast height (DBH; 1.37 m) was 17.2 cm (sd 6.5 cm). Overstory coverage was estimated at less than 20 percent. Evapotranspiration measurements were made at the site from September 1988 to June 1990.

Cypress swamps typically form in shallow depressions that are partially to fully flooded for half of the year or more. Cypress (*Taxodium distichum*) comprises a tall, dense, monospecific overstory. The trees are tallest near the center of the swamp and become progressively shorter toward the edges. This tree height profile, combined with the low height of vegetation outside the swamp, gives the canopy a dome-shaped appearance when it is viewed from a distance. The tree is deciduous. Leaf expansion at the study site begins in late March and leaf fall begins in late October.

Soils in the cypress swamps are typically Sellars mucky loamy fine sand (sandy, siliceous, hyperthermic Cumulic Humaquept). The solum is characterized by an organic surface horizon approximately 5 cm thick that overlies a horizon of mucky loamy fine sand that is approximately 25 cm thick. The remainder of the solum, which is approximately 50 cm thick, consists of fine sands that are very permeable but poorly drained due to the persistence of a high water table.

In the cypress swamp that was selected for this study, it was estimated that the projected crown area of the cypress overstory was about 80 percent of the total swamp area. An instrument tower was installed near the center of the swamp and near the apex of the canopy. A transect was made from the tower to the edge of the swamp to describe the canopy shape. The transect was made in the general direction of prevailing winds at the site (fig. 2). Canopy height was estimated from the measured heights above ground of dominant and codominant trees (fig. 3). The average height of dominant and codominant trees near the tower ranged from 27.8 to 30.2 m and averaged 28.3 m. Evapotranspiration measurements were made at the site from July 1988 until October 1989.

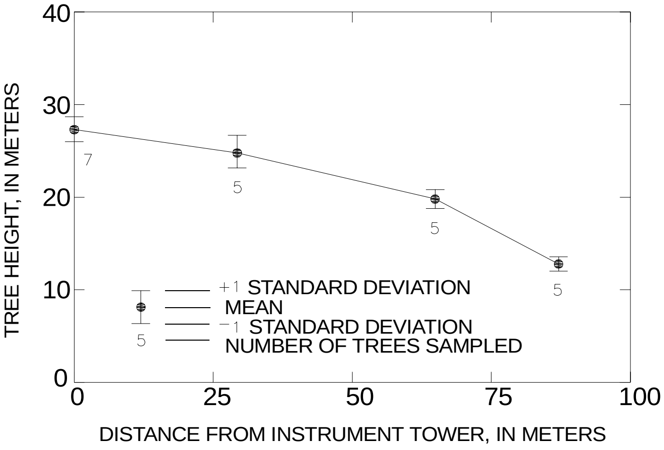

Figure 3. Height of dominant and codominant cypress trees along a transect from the instrument tower to the edge of the swamp at the cypress swamp site.

METHODS, INSTRUMENTATION, AND FIELD MEASUREMENTS

Evapotranspiration was estimated using the micrometeorological methods EBBR and eddy correlation. These two different methods were used so that comparisons between them could be used to examine their suitability for different vegetation types. In the absence of independent knowledge of actual evapotranspiration, agreement between the methods supports at least heuristic arguments that each method furnishes accurate results.

The micrometeorological methods presented in this report rely on measurements that are typically made within a few to a few tens of meters above a surface and within the surface sublayer of the atmospheric boundary layer. The surface can be bare or vegetated soil, or it can be water. Under steady conditions and in the absence of horizontal gradients in vertical fluxes of momentum, heat, and water vapor flux, vertical fluxes of heat and water vapor within the fully turbulent surface sublayer are not appreciably different from fluxes at the surface (Brutsaert, 1982, p. 54). Fluxes of heat and water vapor from a land or water surface can be estimated by determining vertical fluxes of those scalars within the surface sublayer.

The absence of horizontal gradients in vertical fluxes of momentum, heat, and water vapor is an important condition for the one-dimensional EBBR and eddy correlation methods used in this study. When the wind has passed over a uniform land or water surface for a sufficient distance, a layer of air develops near the surface where horizontal gradients in vertical fluxes of momentum, heat, and water vapor are small fractions of the vertical fluxes themselves. This layer of air can be said to be in equilibrium with the surface because windspeed, temperature, and water-vapor profiles reflect surface roughness, temperature, and moisture conditions. EBBR and eddy correlation measurements should be made within the layer of air that is in equilibrium with the land or water surface.

As the wind passes from one type of land or water surface to another, such as from a prairie to a marsh, horizontal gradients can develop across the leading edge of the downwind surface because the airstream begins to exchange momentum, heat, and water vapor with the different downwind surface. Downwind from the surface change boundary, the layer of equilibrated air begins to rebuild from the surface. The thickness of this layer increases with distance downwind.

The instrument height-fetch rule is an operational guide that determines where instruments can be placed to ensure that they will operate within the equilibrated layer of air. Based on this guideline, measurements should be made at a height above a surface not greater than a specified fraction of the distance downwind from a surface change boundary. This fraction has been reported variously as 0.01 (Campbell, 1977, p. 40), 0.01 to 0.003 (Tanner, 1988), and 0.05 (Heilman and Brittin, 1989), depending on such factors as the abruptness of the surface roughness change and atmospheric stability.

Potential evapotranspiration methods can be used to obtain reasonable estimates of evapotranspiration under some field conditions. The data and instrumentation requirements for these methods are typically much less stringent than are requirements for other methods, such as EBBR and eddy correlation. In this study, potential evapotranspiration was computed by the Penman method (Penman, 1956) and compared to evapotranspiration as estimated by the EBBR method to evaluate the suitability of the potential evapotranspiration method for estimating actual evapotranspiration.

Energy-Balance Bowen Ratio

The EBBR method (Bowen, 1926) has been used to estimate evapotranspiration from many different land surfaces, including crops (Tanner, 1960) and forests (McNaughton and Black, 1973) and other types of wildland vegetation (Duell, 1990). It remains as one of the least demanding micrometeorological methods for estimating evapotranspiration from partially wet surfaces.

A description of the EBBR method begins with the equation of the surface energy balance. An equation for the the surface energy balance that omits energy storage due to photosynthesis can be written as

$$R_n - G - S - H - LE = 0, \tag{1}$$

where

R_n is net radiation, in watts per square meter;
G is subsurface heat flux, in watts per square meter;
S is rate of heat storage in the plant canopy, in watts per square meter;
H is sensible heat flux, in watts per square meter; and
LE is latent heat flux, in watts per square meter,
  where L is latent heat of vaporization for water, approximately 2,450 J/g at 20°C and
  E is evapotranspiration flux, in grams per square meter per second.

Energy storage due to photosynthesis is generally thought to be less than a few percent of the surface energy balance. Heat storage in plant canopies can be a substantial portion of the energy balance for periods less than a day, particularly for massive canopies such as forests.

The sign conventions for flux directions are as follows: R_n is positive when the net radiation is directed toward the surface, H and LE are positive when the fluxes are directed away from the surface, G is positive when the subsurface heat flux is directed downward, and S is positive when heat is being stored by the plant canopy.

The Bowen ratio is the ratio of sensible to latent heat flux, H/LE as defined for equation 1. If temperature and vapor pressure are each measured at two heights above a surface, and if vertical profiles of temperature and vapor pressure are similar, the Bowen ratio can be obtained from

$$\beta = H/LE = \gamma \Delta T / \Delta e, \tag{2}$$

where

β is the Bowen ratio, dimensionless;
γ is the psychrometer constant, in kilopascals per degree Celsius;
T is air temperature, in degrees Celsius;
e is water vapor pressure, in kilopascals; and
Δ represents the difference over a vertical interval.

If R_n, G, S, and β are known, equation 1 can be solved for LE by

$$LE = (R_n - G - S)/(1+\beta). \qquad (3)$$

The quantity $R_n - G - S$ can be referred to as "available energy" because radiant energy, less energy that is partitioned to subsurface heat flux and canopy heat storage, is a major impetus for fluxes of sensible and latent heat at the surface. Once LE is known, equation 1 can be solved for H. The latent heat of vaporization varies less than 4 percent over the range of air temperatures encountered in this study. Latent heat flux was converted to evapotranspiration flux by assuming a value for L of 2,450 J/g. By assuming the density of water to be 1 g/cm^3, evapotranspiration flux (gram per square meter per second) can be multiplied by 10^{-3} to express it as a velocity (millimeter per second). Evapotranspiration velocity can be integrated with time to yield daily or annual evapotranspiration.

Standard application of the EBBR method involves the mean-gradient approach because it uses Bowen ratio measurement systems that average temperature and vapor-pressure differences over time in order to calculate the Bowen ratio. The Bowen ratio systems discussed in this section use the mean-gradient approach.

A significant challenge in the application of the EBBR method is to resolve the small temperature and vapor-pressure differences typically present over a short vertical interval above a plant canopy. Daytime air temperature differences are typically a few tenths of a degree, and daytime vapor-pressure differences are typically a few hundredths of a kilopascal. One of the greatest sources of error in these measurements is sensor bias (Fuchs and Tanner, 1970). This error can be reduced by switching sensor positions periodically (Sargeant and Tanner, 1967), or by sampling air from different heights alternately and measuring with the same sensor (Tanner, 1988).

Temperature and vapor-pressure gradients frequently are so small in magnitude at dusk, at night, and at dawn that sensor sensitivity is insufficient to measure them accurately, causing the computed Bowen ratio to be unreasonable at those times. However, little radiative energy is available to drive the exchange of latent heat at those times, and the percentage of the daily flux missed usually is small.

Another significant challenge in application of the EBBR method is the determination of available energy on a short-term basis (hourly or less). This generally is necessary because both available energy and the Bowen ratio fluctuate throughout the day. Under these conditions, average daily available energy and the average daily Bowen ratio, as computed from equation 2, are not appropriate for computing average daily evapotranspiration. The rate of canopy heat storage and the subsurface heat flux in water are difficult to determine on a short-term basis, yet they can compose a substantial part of the available energy for forested or marsh sites. One solution is to use an average Bowen ratio method to compute the appropriate average Bowen ratio over a period of a day or longer, for which the average rate of canopy heat storage and the average subsurface heat flux for shallow water and soil tend to be nearly zero (Black and McNaughton, 1972). The average Bowen ratio methods still require that the temperature and vapor-pressure differences of equation 2 be measured continuously during the period that canopy heat storage and subsurface heat flux are presumed to average zero. Energy-balance Bowen ratio data generally were not collected for full 24-hour cycles during this study because of instrument difficulties encountered during nighttime operation. For this reason, the short-term rate of canopy heat storage and subsurface heat flux were estimated for some sites. Knowledge of available energy also is useful for implementing the eddy correlation and potential evapotranspiration methods. The methods and instrumentation used for estimating available energy are presented in the section titled "Estimation of Available Energy."

Two different Bowen ratio systems were used during the study. The first was a psychrometric system similar to one described by Duell (1990). As it was configured and operated in this study, the system yielded erratic results and little information about evapotranspiration could be discerned from the data. This report contains no further reference to the psychrometric Bowen ratio system.

The second Bowen ratio system is described by Tanner (1988). Air temperature differences over a fixed vertical interval were measured with a pair of chromel-constantan thermocouples (diameter 76 µm). Vapor-pressure differences were measured with a single dewpoint hygrometer (Model Dew–10, General Eastern, Inc.). A pump alternately drew air from upper and lower sampling ports at 2-minute intervals during a

20-minute averaging period, and the air was introduced into the hygrometer for determination of the dewpoint. The vapor pressure of each air stream was determined from an equation for saturation vapor pressure evaluated at the dewpoint temperature (Lowe, 1977). The Bowen ratio was estimated for each averaging period from air temperature and vapor-pressure differences and by use of equation 2. Measurement and system control functions were performed with a data logger (Model 21X, Campbell Scientific, Inc.).

Unreliable Bowen ratio data, which were detected in this study by using the arguments presented by Ohmura (1982), were excluded from EBBR flux computations (eq. 3). An alternative energy-balance method was used to estimate latent heat flux for periods when the Bowen ratio method was not suitable. By use of the energy-balance wind and scalar profile method (EBWSP) (Stricker and Brutsaert, 1978), the surface energy-balance equation can be solved for the latent heat flux *(LE)* by

$$LE = R_n - G - S - H. \tag{4}$$

Available energy $(R_n - G - S)$ is estimated as it is for the EBBR method. The sensible heat flux is calculated from measurements of air temperature at two heights and measurement of windspeed at one height. In this study, air temperature was measured by using the Bowen ratio system and windspeed was measured with a cup anemometer (Model 12102, RM Young Co., or Model 014A, Met One, Inc.). Details of the method for calculating sensible heat flux are given by Stricker and Brutsaert (1978).

Estimation of Available Energy

Net radiation typically is the largest component in the surface energy balance during daylight hours, and it is the largest component for 24-hour periods. Net radiation was measured at all sites in this study. Data from net radiometers of two different designs were used in this study (Model 3035, Weathertronics, Inc., and Model Q–5, Radiation and Energy Balance Systems, Inc.). The Model 3035 net radiometers were used for EBBR, eddy correlation, and potential evapotranspiration measurements from the beginning of the study through October 1989. The Model Q–5 net radiometers were used during EBBR, eddy correlation, and potential evapotranspiration measurements from November 1989 through September 1990. Performance of the two radiometer models was compared at one of the marsh sites during summer 1990. The radiometers were positioned 1 m above the plant canopy with about 3 m of horizontal separation between the instruments. Measurements were averaged hourly for 48 hours. Net radiation measured with the Model 3035 radiometer was approximately 10 percent less than net radiation measured with the model Q–5 radiometer. Horizontal heterogeneity of canopy structure could have been partly responsible for the measurement differences.

Estimation of subsurface heat flux and the rate of heat storage was difficult because of the complex physiognomies of the vegetation types in this study. For example, the rate of canopy heat storage at forested sites can compose a substantial part of the available energy at those sites, particularly for periods of less than 24 hours. At marsh sites, water is the primary medium for heat storage when the water table is above land surface. When the water table recedes, however, the peat on the floor of the marsh assumes a more important role. Because of the difficulties involved with direct measurement of heat storage rates in plant canopies and in water, these rates were not measured routinely, but were estimated using techniques described in this section.

Estimation of subsurface heat flux and canopy heat storage was more straightforward at the dry prairie sites than at any of the other sites in this study. This simplest case is considered first. The subsurface heat flux *(G)* for a mineral soil can be estimated from

$$G = G_z + z\rho \times_b (C_s + C_w \times \theta)\Delta T_s / \Delta t, \tag{5}$$

where

G_z is measured heat flux at depth z in the soil, in watts per square meter;
z is soil depth, in meters;
ρ_b is soil bulk density, in grams per cubic meter;
C_s is specific heat of soil particles, in joules per gram per degree Celsius;
C_w is specific heat of water, in joules per gram per degree Celsius;
θ is gravimetric soil water content, dimensionless;
ΔT_s is the change in average soil temperature to depth z during an averaging period, in degrees Celsius, and;
Δt is length of an averaging period, in seconds.

The second term on the right side of equation 5 is included to account for the rate of heat storage in the layer of soil above the heat flux measurement.

Subsurface heat flux at the dry prairie sites was estimated from measurements with soil heat flux plates and soil temperature probes. The heat flux plates (Model HFT–1, Radiation and Energy Balance Systems, Inc.) were buried at either 5- or 10-cm depths. The averaging soil temperature probes (Model TCAV, Campbell Scientific, Inc.) were constructed from four chromel-constantan thermocouples that were connected in parallel. The thermocouples were spaced vertically at about equal intervals throughout the soil layer above the heat flux plates. For the sandy mineral soils encountered in this study, C_s was estimated to be 0.9 J/g•°C, and ρ_b was estimated to be 1.5×10^{6} g/m^3. The specific heat of water (C_w) is 4.18 J/g•°C. Soil samples were collected, weighed, dried for several days at approximately 100°C, and weighed again to determine gravimetric soil water content (θ). Subsurface heat flux at the pine flatwood and cypress swamp sites was estimated in the same manner as that described for the dry prairie sites.

Measurements from the heat flux plates alone were used to estimate subsurface heat flux for all EBBR, eddy correlation, and potential evapotranspiration data sets collected during 1988 and until July 1989. Soil temperature probes were installed in July 1989 to enable calculation of the rate of soil heat storage for all EBBR and eddy correlation data sets. The rate of soil heat storage was not considered for potential evapotranspiration data sets that were collected on a long-term, continuous basis because the effect of subsurface heat flux on computed potential evapotranspiration becomes negligible for periods of several days or more (Campbell, 1977, p. 138).

Estimation of available energy on a short-term basis at forested sites is complicated by the fact that changes in canopy heat storage are often substantial (Aston, 1985). The rate of canopy heat storage was estimated for the forested sites in this study. An equation for the rate of canopy heat storage *(S)* can be written (Stewart and Thom, 1973) as

$$S = S_s + S_l + S_b, \tag{6}$$

where

S_s is the rate of sensible heat storage in the canopy air, in watts per square meter;

S_l is the rate of latent heat storage in the canopy air, in watts per square meter; and

S_b is the rate of biomass heat storage, in watts per square meter.

The rate of sensible heat storage in the canopy air (S_s) can be estimated from

$$S_s = \rho_a C_p h \Delta T / \Delta t, \tag{7}$$

where

$\rho{\times}_a$ is air density, in grams per cubic meter;

C_p is specific heat of air at constant pressure, in joules per gram per degree Celsius;

h is sensor height, in meters;

ΔT is change in air temperature averaged from the ground to height h, in degrees Celsius; and

Δt is as defined in equation 5.

The rate of latent heat storage in the canopy air (S_l) can be estimated from

$$S_l = L h \Delta\rho{\times}_v / \Delta t, \tag{8}$$

where

$\Delta\rho{\times}_v$ is the change in vapor density averaged from the ground to height h, in grams per cubic meter; and

other terms are as previously defined.

A rigorous determination of sensible and latent heat storage would require that air temperature and vapor pressure be monitored at several heights within the canopy. However, temperature and vapor-pressure measurements above the canopy probably are suitable for rough estimates of the rates of sensible and latent heat storage. Moore and Fisch (1986) found sensible and latent heat storage rates in air within a forest canopy to be strongly correlated with changes in air temperature and specific humidity measured above the canopy. The rainforest canopy studied by Moore and Fisch was 50 percent taller and had approximately three times as much standing biomass as any stand in this study. Results from that study indicate that direct application of above-canopy measurements can lead to overestimation of heat storage rates. In this study, the rate of canopy heat storage during an averaging period was estimated using measurements of air temperature and vapor density that were made between 1 and 3 m above the top of the canopy.

The rate of biomass heat storage (S_b) can be estimated from the equation

$$S_b = C_m M_b \Delta T_b / \Delta t, \tag{9}$$

where

C_m is the mass-averaged specific heat of canopy biomass, in joules per kilogram per degree Celsius;
M_b is canopy biomass, in kilograms fresh biomass per square meter;
ΔT_b is change in the mass-averaged biomass temperature, in degrees Celsius; and
Δt is as previously defined.

The main difficulty in estimating biomass heat storage is the estimation of the mass-averaged rate of biomass temperature change. Aston (1985) and Moore and Fisch (1986) used tissue temperature measurements to develop site-specific empirical models for biomass heat storage. Plant tissue temperature was not measured in this study, and an approximating method was used to estimate the rate of biomass heat storage. First, it was assumed that the change in biomass temperature during an averaging period was equal to the change in air temperature measured between 1 and 3 m above the canopy. Second, it was recognized that the temperature of larger branches and stems would not respond fully to short-term changes in canopy air temperature, so the rate of biomass heat storage was limited to ±×30 W/m×^{2}(Moore and Fisch, 1986).

The specific heat of canopy biomass for the coniferous slash pine and cypress overstories was estimated for this study. The specific heat of canopy biomass varies principally with moisture content (kilograms of moisture per kilogram of dry mass) and is not particularly sensitive to moisture content over the range of moisture content found in living trees. When using an equation presented by Moore and Fisch (1986), for example, computed biomass specific heat increased only 13 percent when moisture content increased by 60 percent (from 0.5 to 0.8 kg/kg). The specific heat of the slash pine and cypress biomass used in this study was that reported by Stewart and Thom (1973), approximately 2,900 J/kg•°C.

Overstory biomass was estimated based on field measurements and allometric relations determined during previous investigations. At the pine flatwood site, overstory biomass was calculated from measurements made at twelve 0.04-ha plots and equations given by Swindel and others (1979) for estimating slash pine biomass from DBH. The equations given by Swindel and others (1979) yield total tree dry biomass; fresh biomass was computed assuming an average pine water content of 0.55 kg/kg. Based on these calculations, overstory biomass at the pine flatwood site was estimated to be 3 kg/m^2. A single sampling point was selected at a representative location at the cypress swamp site. A variable plot-size sampling method was used to select trees for measurement (Husch and others, 1972), and total tree biomass was calculated from DBH measurements and equations given by Nessel and Bayley (1984) for cypress. From these calculations, cypress biomass was estimated to be 22 kg/m^2.

The potential importance of canopy heat storage for energy-balance calculations of daily evapotranspiration at the dry prairie and marsh sites was examined. Because of the small canopy volume and biomass at these sites, the importance of canopy heat storage was estimated to be minimal, and canopy heat storage was neglected.

Several unsuccessful attempts were made to estimate subsurface heat flux at the marsh sites when there was standing water. First, chromel-constantan thermocouples were suspended in the water column and subsurface heat flux *(G)* was computed from the equation

$$G = \rho_w C_w \, z \Delta T_w / \Delta t, \tag{10}$$

where

ρ_w is density of water, in grams per cubic meter;
ΔT_w is the change in the average temperature of the water column during an averaging period, in degrees Celsius;
z is water depth, in meters; and
other terms are as previously defined.

Unless the water is only a few centimeters deep, however, the temperature measurements must be very precise in order to obtain reasonable estimates of subsurface heat flux by using this method. For example, a 0.1°C error in determining water-temperature change for a 30-cm depth of water during a 20-minute period results in an error in G of over 100 W/m^2. Subsurface heat flux calculated using thermocouple data was erratic and of unreasonable magnitude during daylight hours. The unacceptable performance of the thermocouple sensors likely was due to the inherently small sensor output, temperature gradients along the thermocouple reference junction, or improper sensor shielding. Because of the unreliable results of early measures, subsurface heat flux in water was estimated from empirical relations based on measurements of net radiation and an experiment in which water temperature was measured with greater precision.

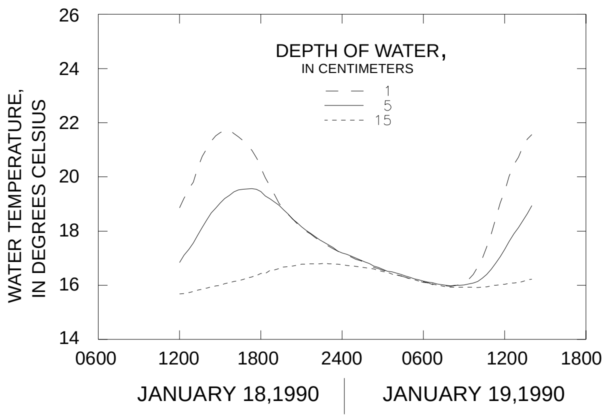

Figure 4. Diurnal variation of water temperature at selected water depths at a marsh site, January 18–19, 1990.

Depth of water near the center of a marsh site in January 1990 was approximately 30 cm. Water temperature was measured at the beginning of each averaging period using shielded temperature sensors constructed from glass bead thermistors (Model AL03006–5818–97–G1, Keystone Carbon Co.) suspended in the water column at depths of 1, 5, and 15 cm. The diurnal trace of water temperature, as measured by the thermistors, was smooth and similar to the typical trace for soil (fig. 4). The amplitude of diurnal temperature fluctuation decreased with depth, and the daily maximum temperature occurred at a later time at greater depth (Campbell, 1977, p. 17). Diurnal temperature variation at the 15-cm depth was nearly damped out and the heat flux below that depth probably was minimal.

Temperature changes at the 1-, 5-, and 15-cm depths were assumed to be equal to the average temperature changes of layers of water that were 3, 7, and 10 cm thick. An average temperature change for the entire 20-cm thickness was estimated from the individual temperature changes weighted by layer thickness. The subsurface heat flux *(G)* for each averaging period was calculated from equation 10 and was regressed against average net radiation (R_n). The regression equation was

$$G = -23.46 + 0.303R_n$$

$$r^2 = 0.94,\ S_{Y\bullet X} = 15.4\ \text{W/m}^2,$$

where

r^2 is coefficient of determination, dimensionless; and

$S_{Y\bullet X}$ is standard error of regression, in watts per square meter.

This regression equation was used to estimate subsurface heat flux *(G)* for all EBBR and eddy correlation data sets collected at the marsh sites when there was standing water.

Another experiment was conducted in June 1990 to determine the relation between net radiation and subsurface heat flux when there was no standing water in the marsh. Peat on the marsh floor was the primary medium for subsurface heat flux when there was no standing water. Heat flux plates and temperature probes were buried in the peat as described previously for the upland sites. The measured gravimetric water content of the peat was 14.1 kg water/kg peat. This value was probably low because some water was squeezed out of the peat during sampling. Because of the large water content, the volumetric specific heat of the moist peat was assumed to be equal to that of water. Subsurface heat flux *(G)*, which was calculated using measured heat flux and equation 5, was regressed against average net radiation (R_n). The regression equation was

$$G = -10.48 + 0.082R_n$$

$$r^2 = 0.65,\ S_{Y\bullet X} = 13.7\ \text{W/m}^2,$$

where terms are as previously defined.

The regression equation was used to estimate subsurface heat flux for all EBBR and eddy correlation data sets collected in the absence of standing water. The relation between net radiation and subsurface heat flux at the marsh sites probably varied depending on the stage of canopy development and the fraction of solar radiation that was intercepted by the plants and the peat surface; however, this was not considered in developing the estimates of subsurface heat flux.

Eddy Correlation

General equations for turbulent transport can be simplified when they are applied to vertical atmospheric transport near an extensive homogeneous surface (Brutsaert, 1982, p. 190). In the absence of horizontal gradients and by ignoring transport by molecular diffusion, equations for sensible *(H)* and latent heat flux *(LE)* can be written as

$$H = \rho_a C_p \overline{w'T'} \tag{11}$$

and

$$LE = L\overline{w'\rho'_v} \tag{12}$$

where

w is vertical windspeed, in meters per second;
ρ_v is vapor density, in grams per cubic meter;
$'$ represents a momentary fluctuation from the mean;
$\overline{}$ signifies the mean of an averaging period; and
other terms are as previously defined.

In the standard application of the eddy correlation method, the variables w, T, and ρ_v are sampled simultaneously, and a series of samples are used to compute the covariances $\overline{w'T'}$ and $\overline{w'\rho_v'}$. Covariance is a measure of the strength of association between two variables.

A significant challenge to the application of the eddy correlation method is the development of adequate sampling schemes. Fluctuations of w, T, and ρ_v each populate frequency distributions that must be adequately sampled if correct fluxes are to be computed. The sampling must be frequent enough to detect high-frequency fluctuations, and the averaging period must be long enough to adequately sample the low-frequency fluctuations (McBean, 1972). In order to adequately sample the high-frequency fluctuations, instruments used to sense the fluctuations must exhibit fast responses, and the data acquisition system must be able to query the sensors and store or process the data at a sufficient rate. Tanner (1988) suggests a 10-Hz sampling frequency and averaging periods from 10 to 20 minutes for work within a few meters of the land surface. Sampling frequencies of either 5 or 10 Hz and averaging periods of either 20 or 30 minutes were used in this study.

Principal sensing components of the eddy correlation system used in this study were a one-dimensional anemometer, an air temperature sensor, and a hygrometer. The sensors were constructed to provide a fast response to enable sampling of high-frequency fluctuations. The sonic anemometer (Model CA–27, Campbell Scientific, Inc.) was similar to one described by Campbell and Unsworth (1979). The anemometer measured vertical windspeed by detecting frequency shifts in sound waves emitted and received by two sonic transducers that were spaced 10 cm apart. Air temperature fluctuations were measured with a chromel-constantan thermocouple (diameter 13 µm) with the reference junction located in the instrument mount. The thermocouple was an integral part of the anemometer and was laterally displaced 4 cm from the middle of the 10-cm path between the sonic transducers. The calibration constant of the anemometer was set by the manufacturer at 1 m/s•V, where V is volt. The calibration constant of the air temperature sensor was set by the manufacturer at 4°C/V. Additional operational characteristics of the anemometer and air temperature sensor are given by Weeks and others (1987) and Tanner (1988).

The hygrometer (Model KH20, Campbell Scientific, Inc.) measured vapor density over a path length of approximately 1 cm by sensing the attenuation of ultraviolet radiation using the principles given by Buck (1976). The ultraviolet radiation was emitted by energized krypton gas in a source tube and was detected by a tube fixed at the other end of the measurement path. Water vapor strongly absorbs radiation in the 123.6-nm wavelength radiation line emitted by krypton gas, and water-vapor density fluctuations can be determined from fluctuations in attenuation of radiation emitted by the gas. Attenuation of parallel, monochromatic radiation can be described by Beer's law (Campbell, 1977, p. 47), and the law can be written in terms of sensor output to relate attenuation of the ultraviolet radiation to water-vapor density (Weeks and others, 1987) using the equation

$$V = V_o \exp(-xK_w\rho_v), \qquad (13)$$

where

V is hygrometer output voltage, which is proportional to the attenuated radiation flux, in volts;
V_o is hygrometer output voltage that would be obtained if the radiation was not attenuated by water vapor, in volts;
x is the atmospheric path length, in centimeters;
K_w is the effective attenuation coefficient for water vapor, in cubic meters per gram per centimeter; and
ρ_v is as defined in equation 12.

Taking the logarithm of both sides, differentiating and solving for $d\rho_v/dV$ yields

$$\frac{d\rho_v}{dV} = \frac{1}{-xK_wV}.$$

If mean voltage remains nearly constant during an averaging period, $d\rho_v/dV$ can be evaluated at the mean voltage $(\overline{V})$, and this permits writing

$$\rho_v' = \frac{(V - \bar{V})}{-xK_w \bar{V}} = \frac{V'}{-xK_w \bar{V}}.$$

The covariance $(\overline{w'\rho_v'})$ of equation 12 can then be written as

$$\frac{-1}{xK_w \bar{V}} \overline{w'V'}.$$

Operating characteristics of the hygrometer are described by Campbell and Tanner (1985), Tanner and others (1985), and Tanner (1988).

A data logger (Model 21X) was used to monitor sensor outputs and to compute summary statistics. Computations for sensible heat and water vapor fluxes were similar to those used by Weeks and others (1987). The computation formula used in this study for sensible heat is equation 11. Latent heat flux was corrected for the effects of sensible heat flux (Webb and others, 1980) and hygrometer sensitivity to fluctuations in concentration of molecular oxygen (Tanner and Greene, 1989). The computation formula used in this study for latent heat flux is

$$LE = L\left(\frac{\overline{w'V'}}{-xK_w\bar{V}} + \frac{\rho_v H}{\rho_a C_p T_k} + \frac{0.229K_o H}{K_w T_k}\right), \qquad (14)$$

where

T_k is air temperature, in kelvin;

K_o is the attenuation coefficient for molecular oxygen, in cubic meters per gram per centimeter; and

other terms are as previously defined.

The middle term in the parentheses of equation 14 is a correction for the effects of sensible heat flux on the latent heat measurement, and the right-hand term is a correction for hygrometer sensitivity to fluctuations in the density of molecular oxygen. The hygrometer path length and attenuation coefficient for water vapor were obtained from periodic calibration of the hygrometer by the manufacturer. The attenuation coefficient for molecular oxygen was assumed to be 0.0085 m^3/g•cm (Harold Weaver, U.S. Geological Survey, oral commun., 1990). Mean vapor density and air temperature were computed from measurements made with an air temperature and relative humidity sensor (Model 207, Campbell Scientific, Inc.), or from measurements made with a wet- and dry-bulb psychrometer.

Standard application of the eddy correlation method yields sensible and latent heat fluxes that are determined independent of the remaining components in the energy balance; therefore, the energy balance can be arranged to check the consistency of the ensemble of energy-balance measurements. Energy-balance closure *(C)* is evaluated by the equation

$$C = \frac{H + LE}{R_n - G - S}. \qquad (15)$$

Any value of *C* other than 1 indicates that the ensemble of energy-balance measurements is inconsistent and that either additional energy-balance components need to be considered or that there were measurement or sampling errors in determining the existing components, or both. In two other studies that used eddy correlation instrumentation similar to that used in this study (Weeks and others, 1987; Duell, 1990), investigators reported values for *C* that were consistently less than 1. If *C* is computed to be less than 1, one or both of the turbulent fluxes *(H* and *LE)* probably were underestimated, or available energy $(R_n–G–S)$ was overestimated. When presented with substantial energy-balance discrepancies, Weeks and others (1987) and Duell (1990) used an alternative evapotranspiration computation to supplement the direct flux measurements. The form of the computation is identical to equation 4 and it is referred to as the “eddy correlation energy-balance residual” (ECEBR) computation. The rationale for discarding the latent heat flux as measured directly by eddy correlation is that measured latent heat flux is subject to more error than is the measured sensible heat flux (Weeks and others, 1987).

Potential Evapotranspiration by the Penman Method

Potential evapotranspiration can be defined as the evapotranspiration that can be sustained from an extensive area densely occupied by active vegetation that is well supplied with water. Methods for determining potential evapotranspiration are useful for estimating evapotranspiration when it is limited primarily by meteorological factors. Meteorological factors typically control evapotranspiration when water is readily available at the surface. Dense, well-watered stands of physiologically active crop plants and open water are examples of surfaces where water is readily available for evapotranspiration. Under these conditions, namely potential conditions, meteorological factors that control evapotranspiration include net radiation, atmospheric vapor-pressure deficit, windspeed, and air temperature.

Nonpotential conditions exist when water availability at the surface limits evapotranspiration to rates below those imposed by meteorological factors. Available energy is partitioned more to sensible heat flux and less to latent heat flux when nonpotential conditions prevail; this is reflected in the air temperature and vapor-pressure profiles that develop above the surface. The air that is in equilibrium with the surface under nonpotential conditions will probably be warmer and dryer than it would be if potential conditions prevailed. Variables measured to compute potential evapotranspiration typically include air temperature and vapor pressure, and attempts to estimate potential evapotranspiration under nonpotential conditions will result in a computed potential evapotranspiration value greater than potential evapotranspiration values computed for potential conditions.

Many different methods have been used to estimate potential evapotranspiration. The method advanced by Penman (1956) has a sound physical rationale and was used in this study. The Penman equation for potential evapotranspiration (E_p) can be written (Monteith, 1965)

$$E_p = \frac{s\,(R_n - G - S) + \rho_a C_p\,(e_s - e)/r_h}{L\,(s + \gamma)}, \qquad (16)$$

where

s is slope of the saturation vapor-pressure curve, in kilopascals per degree Celsius;
e_s is saturation vapor pressure, in kilopascals;
e is atmospheric vapor pressure, in kilopascals;
r_h is boundary layer resistance, in seconds per meter; and
other terms are as previously defined.

The right-hand side of equation 16 can be broken down into radiation and aerodynamic components. The component $s(R_n - G - S)/(L[s+\gamma])$ is determined largely by net radiation. The component $\rho_a C_p\,(e_s - e)/(r_h L[s+\gamma])$, which is determined by aerodynamic transport, is an indication of atmospheric evaporative demand. The quantity $e_s - e$ is the atmospheric vapor-pressure deficit.

Boundary layer resistance (r_h) can be calculated using the equation given by Campbell (1977, p. 138) as

$$r_h = \frac{\ln\left([z_r - d + z_h]/z_h\right)\,\ln\left([z_r - d + z_m]/z_m\right)}{k^2 u}, \qquad (17)$$

where

z_r is height where windspeed is measured, in meters;
d is zero plane displacement height, in meters;
z_h is roughness length for heat transport, in meters;
z_m is roughness length for momentum transport, in meters;
k is von Karman constant ($k = 0.4$), dimensionless; and
u is windspeed, in meters per second.

In this study, R_n and G were calculated using procedures and instruments described above, S was neglected, and T was measured with a Model 207 sensor. At the marsh sites, vapor pressure was calculated from measurements of air temperature and relative humidity (Model 207 sensor) using an equation for saturation vapor pressure given by Lowe (1977). At the dry prairie sites, atmospheric vapor pressure was estimated as described above for the marsh sites, or it was estimated by assuming that atmospheric vapor density is constant over a day and equal to the saturation vapor density at the daily minimum air temperature (Campbell, 1977, p. 138). The daily minimum air temperature was estimated by assuming it was equal to the daily minimum of the average hourly air temperature. Vapor pressure (e) was calculated from vapor density (ρ_v) and absolute temperature (T_k) using the equation $e = 4.62 \times 10^{-4}\,\rho_v T_k$. The slope (s) of the saturation vapor-pressure curve was evaluated using the equation $s = 5307 e_s / T_k^{\,2}$. Windspeed was measured with a cup anemometer (Model 12102 or Model 014A). Plant canopy height above ground was measured in several places and the average measured height was used to estimate d, z_h, and z_m using empirical relations given by Campbell (1977, p. 38).

Field Measurements

Evapotranspiration measurements were made intermittently with the EBBR and eddy correlation instruments from February 1988 to September 1990. One EBBR and two eddy correlation measurement systems (eddy correlation systems 1 and 2) were used for field measurements. The EBBR system was used at the pine flatwood site during October 1989; no other EBBR measurements were made with that system at the forested sites. Both EBBR and eddy correlation measurements were made simultaneously during a total of nine visits to the dry prairie and marsh sites. On 3 days, both eddy correlation systems were operated with the EBBR system at the same site. The two eddy correlation systems were operated simultaneously at the same dry prairie or marsh site on six occasions beginning in November 1989 so that their performance could be compared. The systems were operated at the same height during the comparisons, but the distance between them varied from 2 to approximately 20 m. EBBR or eddy correlation instruments were set up individually or in tandem for periods of 24 hours or longer during each site visit. Water damaged the replaceable transducers on the sonic anemometer, and eddy correlation measurements could not be made during rain. Intense rainstorms sometimes interrupted eddy correlation measurements, particularly during the summer months, resulting in several abbreviated eddy correlation data sets. To prevent instrument damage, the eddy correlation systems were operated only at night when rain was not anticipated.

The EBBR system was deployed for extended periods of unattended operation during the period April to September 1990 at the third dry prairie site and the first marsh site. The system was serviced or moved between sites at intervals of between 10 and 16 days. Water vapor sometimes condensed in the hygrometer and the system tubing during nocturnal operation; therefore, the vapor-pressure sampling system was not operated at night. Nocturnal evapotranspiration was estimated using the EBWSP method. Data also were collected during these operations to enable calculation of potential evapotranspiration.

Potential evapotranspiration was calculated from data collected at the second dry prairie site from September 1989 to March 1990 and at the third dry prairie site from March to September 1990. No potential evapotranspiration calculations were performed for the two forested sites.

EBBR, eddy correlation, and other micrometeorological instruments were mounted on supporting towers to position them above the plant canopy. Instruments were 1–2.5 m above the plant canopy at the dry prairie sites, and they were 1–2 m above the plant canopy at the marsh sites. Instruments were mounted 2–3 m higher than neighboring trees at the pine flatwood site, except during October 1989 when they were 4–5 m higher than neighboring trees. Instruments at the cypress swamp site were approximately 1 m higher than neighboring trees. Raupach and others (1980) reported that instruments placed close to aerodynamically rough surfaces, such as forests, could be within the roughness sublayer of the atmospheric boundary layer where substantial horizontal gradients can bias turbulent flux measurements.

RESULTS OF EVAPOTRANSPIRATION MEASUREMENTS

Data from field measurements were processed to compute evapotranspiration and potential evapotranspiration for averaging periods and for daily periods. The daily results were then used to evaluate suitability of the various methods for estimating evapotranspiration for each vegetation type, to generate estimates of annual evapotranspiration, and to provide information with which to discuss uncertainties in evapotranspiration estimates.

Energy-Balance Bowen Ratio

Examples of EBBR measurements at dry prairie and marsh sites are presented in figure 5. The four energy fluxes exhibited diurnal fluctuations. At the dry prairie site, for example, net radiation and subsurface heat flux were negative at night. After sunrise, net radiation became positive and increased rapidly. During the daytime, net radiation is strongly related to the incoming flux of short-wave radiation, and this can be seen in the net radiation fluctuations during the day that were caused by clouds that alternately excluded and admitted the solar beam. Net radiation decreased after noon and became negative at about the time of sunset. Subsurface heat flux remained negative in the morning until after net radiation was positive, and then increased to approximately 10 percent of net radiation during the day. Subsurface heat flux became negative

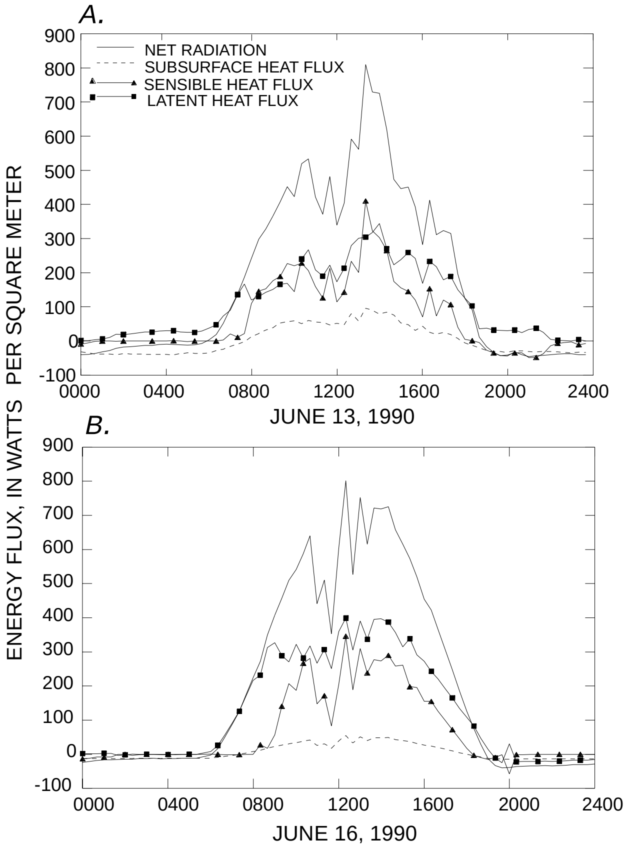

Figure 5. Diurnal variation of surface energy-balance components in conjunction with energy-balance Bowen ratio measurements at *(A)* a dry prairie site on June 13, 1990, and *(B)* a marsh site on June 16, 1990.

in the early evening before net radiation became negative. Sensible heat flux was zero or less at night and changed sign at about the same time as net radiation. The latent heat flux was usually zero or slightly greater at night. The maximum computed nighttime latent heat flux was 37 W/m^2. The latent heat flux increased rapidly with increasing net radiation in the morning and was generally greater than the sensible heat flux during the day.

Diurnal fluctuations of energy-balance components at the marsh site generally were similar to diurnal fluctuations at the dry prairie site. Estimated subsurface heat flux was nearly zero at night and was never greater than 7 percent of net radiation during the day. Subsurface heat flux was estimated from net radiation measurements using the regression equation developed for the waterless condition that existed at the marsh at the time of the EBBR measurements. Latent heat flux was greater than sensible heat flux during the daytime. Latent heat flux and net radiation were about equal during the first 2 hours of daylight, indicating that available energy was partitioned almost exclusively to evaporating water during that time.

The relative importance of energy-balance components for computing daily evapotranspiration can best be examined by summing the energy fluxes for 24-hour or daytime periods (table 1). One obvious effect of summing the dry prairie data for 24 hours was that subsurface heat flux was small in comparison to the remaining fluxes. Measured subsurface heat flux was less than 3 percent of the latent heat flux during 24 hours. At the dry prairie site, daytime subsurface heat flux was approximately 18 percent of the latent heat flux. Daytime subsurface heat flux at the dry prairie site was about 10 percent of net radiation, which is the percentage suggested by Campbell (1977, p. 138) for crops.

Evapotranspiration computed for 24-hour periods at the dry prairie and marsh sites was 3.87 and 4.61 mm, respectively. Computed daytime evapotranspiration at the dry prairie and marsh sites for the examples shown (fig. 5) was 3.56 and 4.69 mm, respectively.

Results from the EBBR method at the pine flatwood site during October 1989 discouraged further use of that method at the forested sites in this study. The dewpoint hygrometer in the Bowen ratio system was

Table 1. Surface energy-balance components estimated in conjunction with energy-balance Bowen ratio measurements and summed over 24-hour and daytime periods for a dry prairie site on June 13, 1990, and for a marsh site on June 16, 1990

[MJ/m^2, megajoules per square meter; R_n, net radiation; *G*, subsurface heat flux; *H*, sensible heat flux; *LE*, latent heat flux]

Measurement period	R_n (MJ/m^2)	*G* (MJ/m^2)	Bowen ratio	*H* (MJ/m^2)	*LE* (MJ/m^2)
Dry prairie site					
24-hour	15.7	0.3	0.63	5.9	9.5
Daytime	16.7	1.6	.74	6.4	8.7
Marsh site					
24-hour	18.1	[1].6	.55	6.18	11.3
Daytime	19.0	[1]1.1	.55	6.34	11.5

[1] Subsurface heat flux at the marsh site was estimated from measurements of net radiation as described in the section titled "Estimation of Available Energy."

capable of resolving a vapor-pressure difference of about 0.005 kPa. The measured vapor-pressure difference over a 2-m interval above the pine canopy was less than 0.005 kPa for 65 percent of the daytime averaging periods, which indicated vapor-pressure gradients above the aerodynamically rough forest were too weak for successful use of the Bowen ratio system.

In general, the Bowen ratio system performed reliably during extended periods of unattended operation at the dry prairie sites. EBBR averaging periods rejected on the basis of review criteria of Ohmura (1982) accounted for an average of only 9 percent of the daytime available energy. Lack of steady-state conditions in the surface sublayer could have been responsible for some of the data rejections; however, errors in determining available energy, limited instrument sensitivity, and errant measurements of vapor-pressure difference caused by condensation within the system tubing were likely responsible for most of the rejected data. Considerably more difficulty was encountered during extended operation of the Bowen ratio system at a marsh site. During 5 of 14 days of unattended operation at a marsh site, more than 30 percent of daytime available energy could not be used to calculate evapotranspiration by the EBBR method because of rejected averaging periods. Data from those 5 days were not used in evapotranspiration computations.

Eddy Correlation

Examples of diurnal fluctuations of energy-balance components obtained during eddy correlation measurements are presented in figure 6. Examples of eddy correlation measurements for the dry prairie and marsh sites are for the same day as the examples that are presented for EBBR measurements (fig. 5). Some of the energy-balance components, such as net radiation, and, at the unforested sites, subsurface heat flux, exhibited common diurnal variations that have been described previously. At the forested sites, the sum of subsurface heat flux and estimated canopy heat storage $(G+S)$ generally was less than the sensible and latent heat fluxes.

The relative importance of energy-balance components for computing daily evapotranspiration from the forested sites was examined by summing the energy fluxes for 24-hour and daytime periods (table 2). Subsurface heat flux at the forested sites, like subsurface heat flux at the unforested sites, was small compared to net radiation and sensible and latent heat fluxes for 24-hour periods. Daytime subsurface heat flux was approximately 8 percent of net radiation at the pine flatwood site and was less than 2 percent of net radiation at the cypress swamp site. Estimated canopy heat storage at the pine flatwood was negligible when summed for 24 hours, which indicated that little canopy heat storage was carried from one diurnal cycle to the next. For daytime periods, estimated canopy heat storage was about 2 percent of net radiation at the pine flatwood site and about 3 percent of net radiation at the cypress swamp site. The canopy heat storage estimates are only approximate. The estimation procedures used probably overestimated the rate of canopy heat storage, and the results indicate that canopy heat storage is only a small fraction of the surface energy balance when the balance is summed for daytime periods, or especially for 24-hour periods.

Available energy and the turbulent fluxes were statistically related for each site, and the strength of the relations differed between the unforested and forested sites. Sensible and latent heat fluxes in all of the examples presented generally followed daily patterns that were similar to patterns of available energy. Measured sensible and latent heat fluxes were largest during the daytime when available energy was largest. Measured sensible heat flux was strongly correlated with available energy during the daytime at the dry prairie ($r=0.95$), marsh ($r=0.98$), and pine flatwood ($r=0.88$) sites. Daytime available energy was not as strongly correlated with sensible heat flux at the cypress swamp site ($r=0.79$). The strength of the relation between available energy and latent heat flux was distinctly greater for the unforested sites. The correlation between available energy and latent heat flux was strong for the dry prairie ($r=0.95$) and marsh ($r=0.96$) sites, but was weak for the pine flatwood ($r=0.40$) and cypress swamp ($r=0.74$) sites. The relative decoupling between available energy and latent heat flux at the pine flatwood site also was indicated by a substantial latent heat flux in the early evening (fig. 6). The lack of correlation between calculated available energy and latent heat flux at the forested sites was probably partly due to an inability to estimate available energy accurately for the 20- or 30-minute averaging periods.

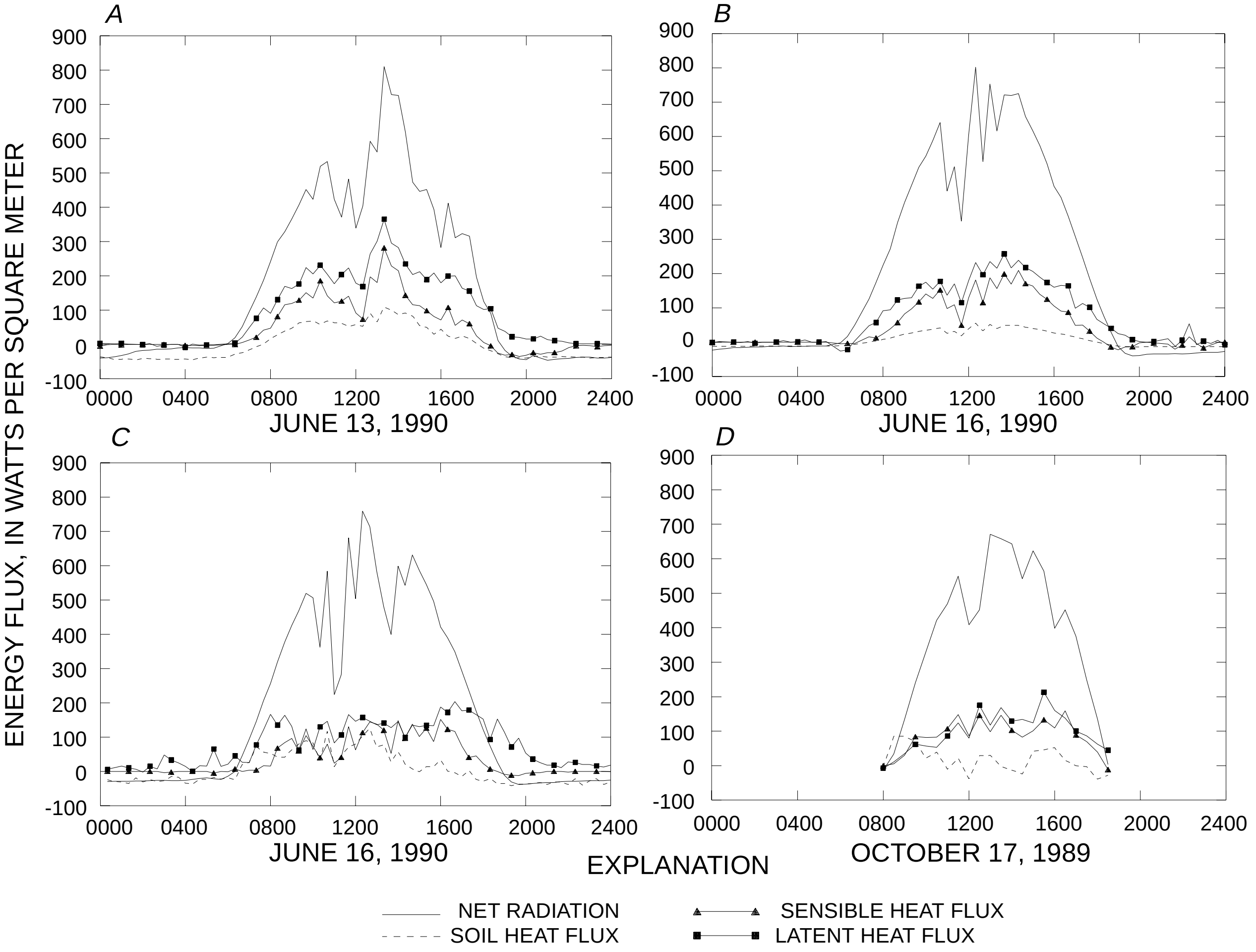

Figure 6. Diurnal variation of surface energy-balance components in conunction with eddy correlation measurements at (*A*) a dry prairie site on June 13, 1990, (*B*) a marsh site on June 16, 1990, (*C*) a pine flatwood site on June 16, 1990, and (*D*) a cypress swamp site on October 17, 1989.

Table 2. Surface energy-balance components estimated in conjunction with eddy correlation measurements and summed for 24-hour and daytime periods for a pine flatwood site on June 16, 1990, and for a daytime period at a cypress swamp site on October 17, 1989

[MJ/m^2, megajoules per square meter; R_n, net radiation; *G*, subsurface heat flux; *S*, rate of canopy heat storage; *H*, sensible heat flux; *LE*, latent heat flux; <, less than]

Measurement period	R_n (MJ/m^2)	*G* (MJ/m^2)	S^1 (MJ/m^2)	*H* (MJ/m^2)	*LE* (MJ/m^2)
		Pine flatwood site			
24-hour	15.8	0.4	<0.1	3.2	7.0
Daytime ...	17.3	1.4	.3	3.3	5.8
		Cypress swamp site			
Daytime ...	15.0	.2	.5	3.4	3.8

[1] The rate of canopy heat storage was estimated using techniques described in the section titled “Estimation of Available Energy.”

Eddy correlation measurements indicated that nighttime evapotranspiration usually was quite small except at the pine flatwood site. Measured nocturnal evapotranspiration, averaged for all nighttime eddy correlation measurements, was 0.03 mm/d at the dry prairie (n=18) and marsh sites (n=9), 0.28 mm/d at the pine flatwood site (n=8), and 0.07 mm/d at the cypress swamp site (n=7).

Calculations of energy-balance closure indicated that turbulent fluxes calculated by the standard eddy correlation computation method were typically insufficient to account for available energy (eq. 15). Mean daytime energy-balance closure *(C)* was 0.68 for the dry prairie sites (n=22), 0.68 for the marsh sites (n=20), 0.74 for the pine flatwood site (n=10), and 0.45 for the cypress swamp site (n=10). Data for some eddy correlation data sets were collected continuously for a day or longer, which permitted the energy balance to be averaged for 24-hour periods. Effects of errors in determining subsurface heat flux and the rate of canopy heat storage on energy-balance closure can be reduced when the energy balance is averaged for 24-hour periods. The reason for this is that much of the energy stored in the soil, water, and plant canopy during the day is discharged into the atmosphere at night. Changes in energy-balance closure for 24-hour periods generally were small. Closure calculations for 24-hour periods yielded means of 0.69 for the dry prairie sites ($n=18$), 0.66 for the marsh sites ($n=9$), 0.80 for the pine flatwood site ($n=8$), and 0.49 for the cypress swamp site ($n=7$).

Energy-balance inconsistencies are difficult to reconcile because there are many potential sources of error in the estimation of energy-balance components. Net radiation typically is the largest component in the energy balance during the daytime, and net radiation measurements have some error associated with them. The Model Q–5 radiometers used in this study were similar to the radiometers used by Weeks and others (1987) except the radiometers in this study had been modified by the manufacturer to improve the response to long-wave radiation. On the basis of independent field measurements, Weeks and others (1987) thought that their radiometers overestimated net radiation by about 10 percent. The Model Q–5 net radiometers used in this study could have overestimated net radiation slightly, but the error was probably less than 10 percent. Less independent information is available about the performance of the Model 3035 net radiometers. The only comparison that was made between the Model Q–5 and Model 3035 radiometers indicated that the latter radiometers measured about 10 percent less net radiation than did the Model Q–5. Net radiation measurement errors of 10 percent are not sufficient to explain the energy-balance closure errors.

Net radiation sampling errors could have contributed to the energy-balance closure errors at the forested sites. The radiometers were positioned 1–3 m above the forest canopy, which likely was not high enough to obtain the canopy-averaged net radiation because the radiometers received radiation from only a small number of trees or parts of trees.

Energy-balance closure errors can be caused by errors in determining either subsurface heat flux or the rate of canopy heat storage. Among results from the four vegetation types in this study, those from the dry prairie sites are the most suitable for examining the potential effect of errors in subsurface heat flux and the rate of canopy heat storage on energy-balance closure. Subsurface heat flux was measured at the dry prairie sites and the role of canopy heat storage in the surface energy balance was minimal at the dry prairie sites. Determination of subsurface heat flux at the dry prairie sites is subject to sampling error because subsurface heat flux varies spatially and measurements of subsurface heat flux were made at only three to six points. The number of measurements that would be needed to adequately characterize subsurface heat flux would probably depend on the magnitude of subsurface heat flux with respect to other components in the energy balance and on the spatial heterogeneity of the plant canopy. The fact that energy-balance closure for 24-hour averaging periods was not improved, as compared to daytime periods, indicated that errors in determining subsurface heat flux and errors due to neglecting canopy heat storage were either offsetting or not important in the energy-balance calculations. Diurnal changes in the rate of heat storage in the soil and the plant canopy typically are much more rapid than changes induced by varying weather conditions. For this reason, it is unlikely that energy-balance closure would be improved significantly if averaging periods were extended to 48 or 72 hours or longer. The significance of energy-balance closure errors for the marsh, pine flatwood, and cypress swamp sites are more difficult to assess because subsurface heat flux and the rate of canopy heat storage were estimated and the estimates could not be independently tested.

The turbulent fluxes can be underestimated for many reasons, including improper sensor calibration, distortion of the wind field caused by the sensors or supporting structures, improper cosine response of the anemometer or tilt of the anemometer with respect to the mean wind vector, insufficient sensor frequency response or averaging period duration, and insufficient fetch over a homogeneous plant canopy. In an effort to solve the energy-balance closure problem, a hygrometer and a sonic anemometer were returned to the manufacturer for calibration. The hygrometer was returned in December 1989, and the anemometer was returned in April 1990. The effective attenuation coefficient for water vapor (K_w) had changed by approximately 2 percent during the 6 months since the previous factory calibration. No malfunction or calibration shift was reported by the manufacturer for the anemometer. No substantial improvement in energy-balance closure was detected when the instruments were returned. Sensor tilt with respect to the mean wind vector and wind-field distortion most likely would result in a nonzero wind component along the measurement path of the sonic anemometer. Unfortunately, the anemometer signal exhibits a temperature-dependent drift that precludes use of the instrument for determining mean wind components. In an attempt to minimize instrument tilt with respect to the mean wind vector, eddy correlation sensors were always leveled in the field. Attempts were made to minimize wind-field distortion by pointing the eddy correlation sensors into the predominant wind direction. The instruments were not attended continuously, and shifts in wind direction at times resulted in the wind blowing past the support structures before it reached the eddy correlation instruments.

There was a systematic difference between turbulent fluxes computed from the two eddy correlation systems when they were operated simultaneously at dry prairie or marsh sites. Daytime evapotranspiration computed using eddy correlation system 1 was, with one exception, larger than evapotranspiration computed using eddy correlation system 2 (fig. 7). The systematic difference between results obtained with the two systems supports a conclusion that errors in estimating the turbulent fluxes were partly responsible for the energy-balance closure errors.

The systematic difference between the two eddy correlation systems was apparent in the energy-balance closure calculated from all of the dry prairie measurements. Eddy correlation systems 1 and 2 were each used the same number of times at the dry prairie sites. The mean 24-hour closure was 0.74 (n=9) when eddy correlation system 1 was used and 0.63 (n=9) when eddy correlation system 2 was used. This closure difference supports a conclusion that errors in estimating the turbulent fluxes were partly responsible for the energy-balance closure errors.

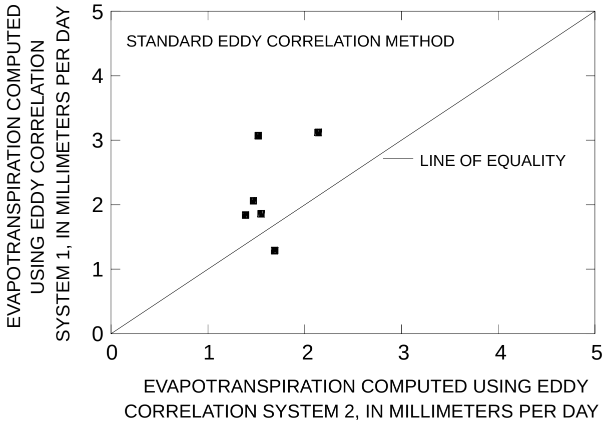

Figure 7. Relation between daytime evapotranspiration computed for two eddy correlation systems using the standard eddy correlation computation method.

As discussed in the presentation of the eddy correlation method, other investigators have had similar difficulties in achieving energy-balance closure, and the eddy correlation energy-balance residual (ECEBR) computation method has been used as an alternative calculation for evapotranspiration. A third evapotranspiration calculation can be based on eddy correlation and other energy-balance measurements. Tanner and others (1985) compared turbulence measurements from six eddy correlation systems that were similar in design to the two systems used in this study. Some of the six systems consistently indicated lower sensible and latent heat fluxes than others, and statistical analyses indicated that much of the variation among systems could be attributed to variation in the measurement of vertical windspeed *(w)*. Based on arguments given by Tanner and others (1985), the equations for sensible heat flux *(H)* and latent heat flux *(LE)* can be written as

$$H = \rho_a C_p \overline{w'T'} = \rho_a C_p r_{w,T} \sigma_w \sigma_T \qquad (18)$$

$$LE = L\overline{w'\rho'_v} = L r_{w,\rho_v} \sigma_w \sigma_{\rho_v} \qquad (19)$$

where

r is the correlation coefficient for w and T and for w and ρ_v, dimensionless;

σ is the standard deviation for w, T, and ρ_v, with dimensions of meters per second, degrees Celsius, and grams per cubic meter, respectively; and

other terms are as previously defined.

If the magnitude of fluctuation in vertical windspeed were underestimated, the result would be an underestimation of σ_w. It can be seen from equations 18 and 19 that, if σ_w were underestimated by an eddy correlation system, the turbulent fluxes would be underestimated, but the Bowen ratio would be preserved. If the correct Bowen ratio were determined by eddy correlation and if available energy could be determined, evapotranspiration could be calculated from equation 3 by the eddy correlation energy-balance Bowen ratio (ECEBBR) computation method. When using the ECEBBR method, evapotranspiration calculated from eddy correlation systems 1 and 2 was very similar for the dry prairie and marsh sites (fig. 8). Some of the agreement was due to the contribution of a common available energy component. The agreement between results from the two eddy correlation systems indicates that the Bowen ratios obtained by the two eddy correlation systems were similar. To compare the Bowen ratios obtained from each system, a flux-averaged Bowen ratio ($\bar{\beta}$) was computed as

$$\bar{\beta} = \frac{\bar{H}_{ECEBBR}}{\overline{LE}_{ECEBBR}}$$

where the overbar signifies daytime average. The flux-averaged Bowen ratio was used to compare daily results because it weights the averaging periods when available energy is highest and Bowen ratio measurements are most critical for determining daily energy fluxes. Analysis of variance indicated that, at the 5-percent probability level, there was no statistically significant difference between flux-averaged Bowen ratios obtained from the two eddy correlation systems; the maximum sampled difference was 0.16. These comparisons do not demonstrate with certainty that the Bowen ratios determined by eddy correlation measurements were accurate because there was no independent confirmation of the actual Bowen ratio. A more rigorous test would be to compare results obtained by the ECEBBR computation method with results obtained by the EBBR method implemented using a mean-gradient Bowen ratio system.

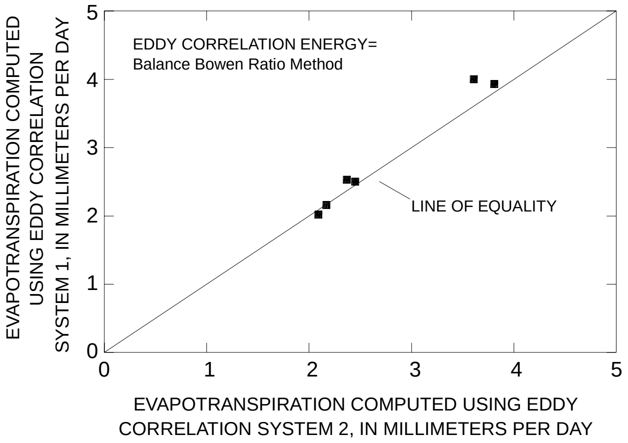

Figure 8. Relation between daytime evapotranspiration computed for two eddy correlation systems using the eddy correlation energy-balance Bowen ratio computation method.

Comparison Between Energy-Balance Bowen Ratio and Eddy Correlation Results

Comparisons between EBBR and eddy correlation results were based on the nine daytime periods when the EBBR system and one or both eddy correlation systems were operated at the same dry prairie or marsh site (fig. 9*A*). Daytime evapotranspiration computed by the EBBR method was always greater than evapotranspiration computed by the standard eddy correlation method, and results from the two methods were not well correlated. The observation that evapotranspiration computed by the EBBR method was systematically greater than evapotranspiration calculated by the standard eddy correlation method was consistent with available energy being overestimated or with latent heat flux from eddy correlation being underestimated, or both.

Comparison between EBBR and ECEBBR results revealed much closer agreement (fig. 9*B*). Daytime evapotranspiration, as calculated by the two methods, seemed to be strongly correlated and there was no indication that one method predicted systematically greater evapotranspiration. Some of the agreement was due to the contribution of a common available energy component in the two evapotranspiration calculations. The favorable comparison between the results from the EBBR and ECEBBR methods does not attest to the accuracy of the methods for estimating evapotranspiration because errors in determining available energy could not be isolated. However, the favorable comparison between results based on turbulence measurements and on mean-gradient measurements supports a conclusion that the two different methods yielded approximately the correct Bowen ratio, at least for the dry prairie and marsh sites.

Comparison between EBBR and ECEBR results revealed that the two seemed to correlate, but the ECEBR method systematically yielded greater estimates of evapotranspiration than did the EBBR method (fig. 9*C*). Some of the correlation was due to the contribution of a common available energy component in the two evapotranspiration calculations. The observation that evapotranspiration calculated by the ECEBR method was greater than evapotranspiration calculated by the EBBR method pointed to the possible overestimation of available energy or the underestimation of sensible heat flux by eddy correlation (eq. 15). The level of agreement between the two methods could not be used to determine the accuracy of either because the accuracy of the calculated available energy is unknown.

The substantial energy-balance closure errors encountered in this study and the poor agreement between results from the EBBR method and the standard eddy correlation computation method cause concern about the equipment and techniques used in this study. Additional research to identify the primary sources of the closure errors in this and similar studies is needed. Because of the similarity of results obtained with the EBBR and ECEBBR methods, results from the these two methods will be emphasized for estimating evapotranspiration from the vegetation types examined in this study.

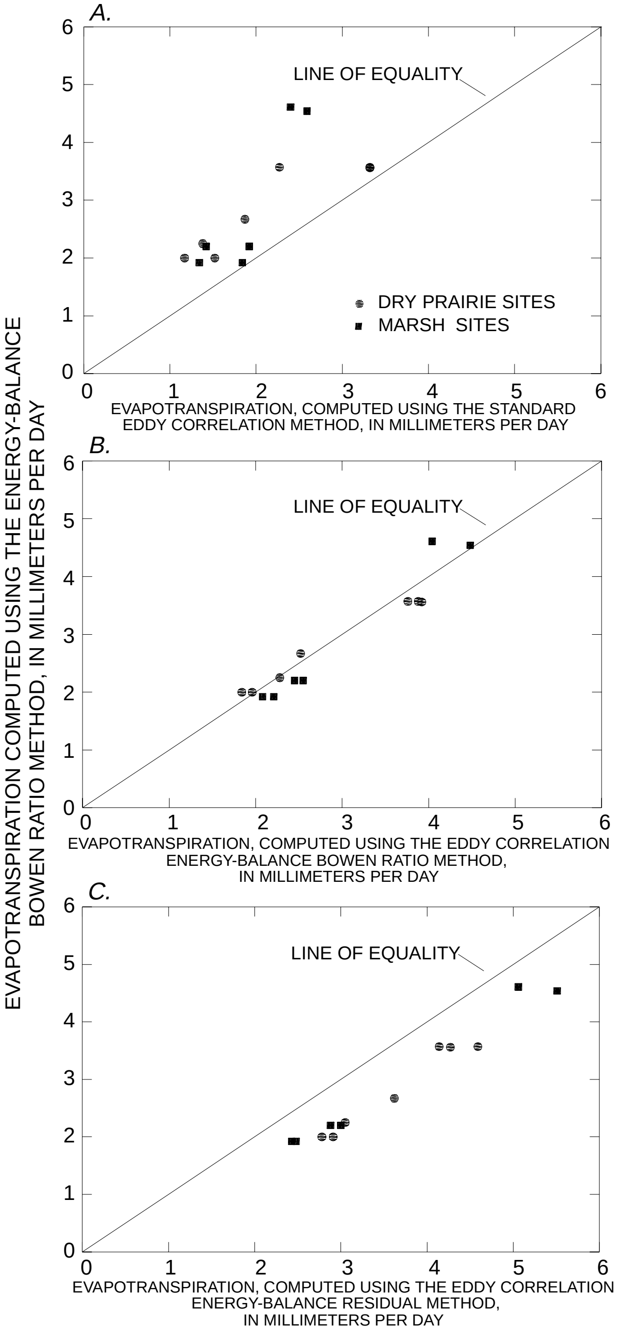

Figure 9. Relation between daily evapotranspiration from unforested sites computed using the energy-balance Bowen ratio method and daily evapotranspiration computed using *(A)* the standard eddy correlation computation method, *(B)* the eddy correlation energy-balance Bowen ratio computation method, and *(C)* the eddy correlation energy-balance residual computation method.

Potential Evapotranspiration

Calculated potential evapotranspiration varied seasonally at the dry prairie sites (fig. 10). The 3 months of highest average potential evapotranspiration were March, April, and June. The 3 months of lowest average potential evapotranspiration were November, December, and January. Annual potential evapotranspiration, calculated from monthly means, was 1,820 mm. The radiation and aerodynamic components of the Penman equation are shown in figure 10. The radiation component was larger than the aerodynamic component for 8 months of the year, and 1,120 mm of the annual potential evapotranspiration

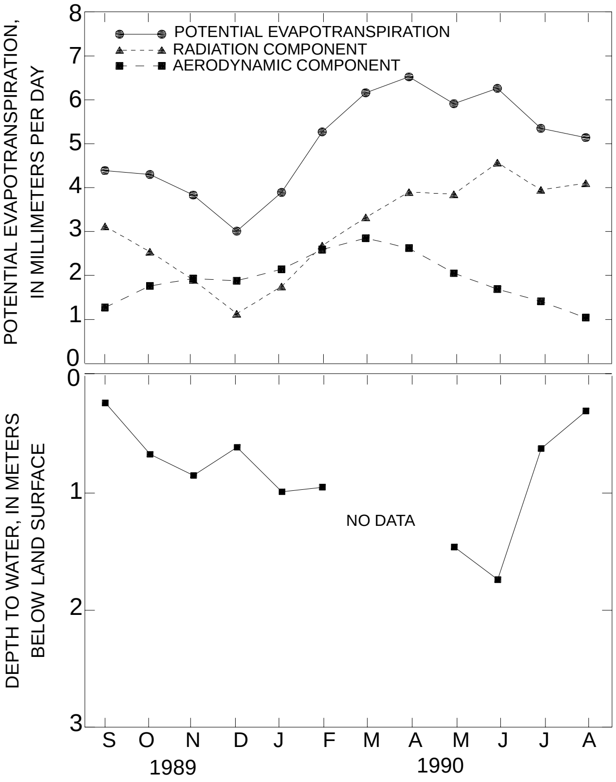

Figure 10. Annual variations in daily potential evapotranspiration and depth to the water table at the T. Mabry Carlton, Jr., Memorial Reserve. (Depth to the water table is the maximum depth to the water table below land surface recorded on the last day of each month. From U.S. Geological Survey, 1990; 1991.)

could be attributed to the radiation component. The aerodynamic component accounted for 700 mm of the annual potential evapotranspiration. This component decreased during late spring and summer, whereas the radiation component increased, primarily as a result of seasonal increases in net radiation. The decrease in the aerodynamic component was due primarily to a decrease in the atmospheric vapor-pressure deficit during the humid months of June to October.

Daytime evapotranspiration at the dry prairie site during the months of April to September 1990, as estimated by the EBBR method, was not correlated with potential evapotranspiration (fig. 11*A*). Potential evapotranspiration roughly defined the upper limit of evapotranspiration. Estimated evapotranspiration ranged from 1.67 to 5.20 mm/d during the period, whereas potential evapotranspiration ranged from 3.10 to 7.84 mm/d.

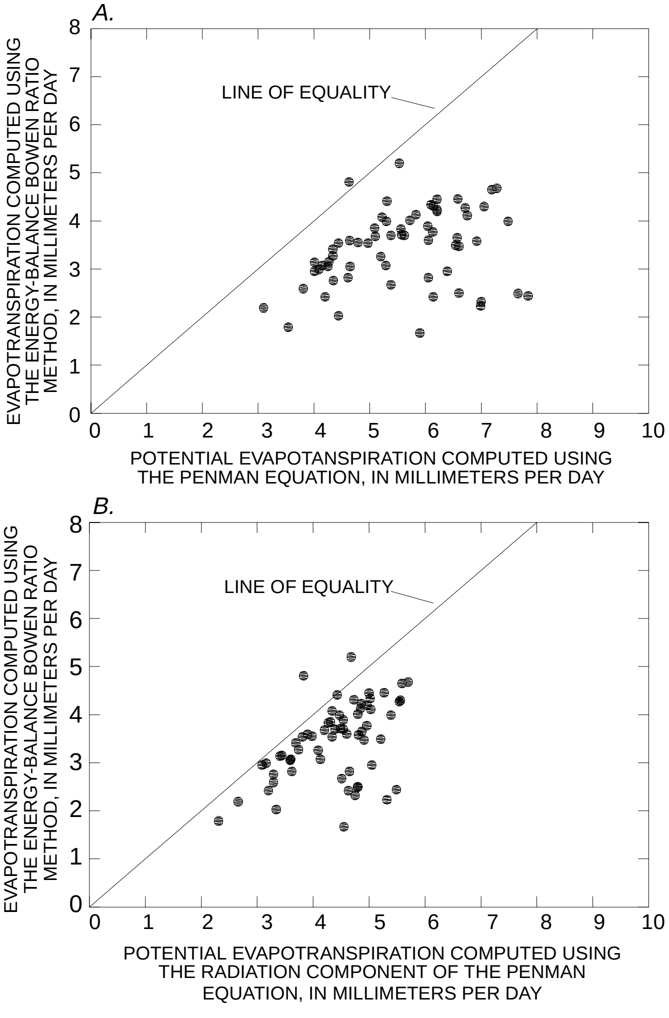

Figure 11. Relation between daily evapotranspiration from a dry prairie site computed using the energy-balance Bowen ratio method and *(A)* daily potential evapotranspiration computed using the Penman equation, and *(B)* daily potential evapotranspiration represented by the radiation component of the Penman equation.

Daytime evapotranspiration at the dry prairie site during the months of April to September 1990 was not correlated with the radiation component of the Penman equation (fig. 11*B*). Differences between evapotranspiration and the radiation component were, on average, not as great as differences between evapotranspiration and potential evapotranspiration.

Daytime evapotranspiration at the marsh site for 9 days during May and June 1990 was correlated with potential evapotranspiration (fig. 12*A*). The slope of the relation was 0.57 when the intercept was constrained to the origin.

Daytime evapotranspiration at the marsh site during May and June 1990 was correlated with the radiation component of the Penman equation (fig. 12*B*). The slope of the relation was 0.83 when the intercept was constrained to the origin.

Analysis of Method Suitability

Evaluation of suitability of the methods used in this study is complicated by uncertainties about the validity of numerous assumptions that were used in the application of the methods and by uncertainties about the performance of sensors that were used. The assumptions, such as those pertaining to the nature of transport in the surface sublayer of the atmospheric boundary layer, were not tested. Because sensor performance and measurement errors are difficult to assess under field conditions, no rigorous testing of sensor performance was done. Resources and arguments that are available for the purpose of evaluating method suitability in the absence of field tests are (1) literature that describes field conditions where the methods would be suitable, (2) the presence or absence of results that are consistent with current knowledge of near-surface atmospheric transport and energy balances, and (3) corroboration of results among different methods or other comparable studies.

Energy-Balance Bowen Ratio.—Results from this study indicate that the EBBR method, applied using a mean-gradient Bowen ratio system, is suitable for estimating evapotranspiration from dry prairie and marsh vegetation sites. First, daytime Bowen ratios computed for these vegetation types were generally

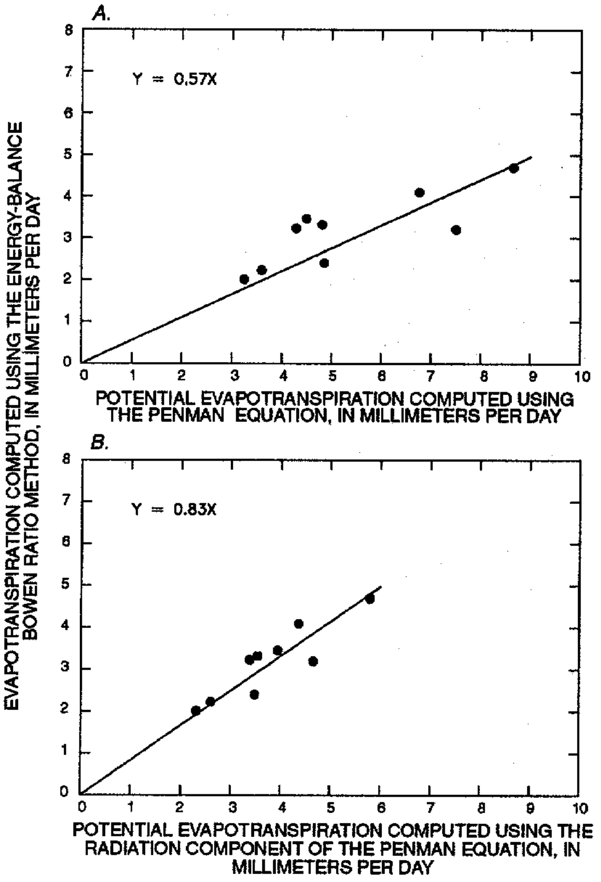

Figure 12. Relation between daily evapotranspiration from a marsh site computed using the energy-balance Bowen ratio method and *(A)* daily potential evapotranspiration computed using the Penman equation, and *(B)* daily potential evapotranspiration represented by the radiation component of the Penman equation.

less than 1. The method works well with mean-gradient systems when the Bowen ratio is less than 1 because this indicates that vapor-pressure differences are relatively large (Angus and Watts, 1984) and are resolved more easily. Second, the Bowen ratio system based on the dewpoint hygrometer consistently gave physically meaningful results during diurnal operation at dry prairie sites. Turbulent fluxes computed by this method usually were rejected only for the early morning hours when the magnitude of vapor-pressure and temperature differences and available energy were relatively small. The percentage of the daily evapotranspiration that was missed because of the erroneous fluxes was usually small. Less reliable evapotranspiration estimates were obtained by the EBBR method when the Bowen ratio system was left at a marsh site for extended periods of unattended operation. However, difficulties in obtaining reliable evapotranspiration estimates at the marsh sites probably were caused by the design and operation of the particular Bowen ratio system and not by a failure of the EBBR method. Finally, the consistent agreement between evapotranspiration estimated by the EBBR and ECEBBR methods at dry prairie and marsh sites indicates that the EBBR method is suitable for estimating evapotranspiration from those vegetation types if available energy can be determined accurately and sites with sufficient fetch are used.

The suitability of the EBBR method for estimating evapotranspiration from the forested vegetation types in this study is less certain because application of the method is more difficult for tall canopies. First, temperature and vapor-pressure gradients are usually small above aerodynamically rough surfaces such as forests; therefore, successful use of the EBBR method is contingent upon the ability of sensors to resolve very small differences. The sensors used in this study apparently lacked sufficient resolution for use over forests. Second, the EBBR method is predicated on the assumption that temperature and vapor-pressure profiles above the canopy are similar; however, dissimilar profiles can develop over tall forest canopies because of vertical separation of sources of sensible heat and water vapor within the canopy. Brutsaert (1982, p. 211) suggests that the dissimilarity can persist for heights of three to five times the canopy height. Measurement heights of these magnitudes are not practical for the forest stands in this study because the instrument height-fetch requirement could not be met.

One method for identifying dissimilarity between temperature and vapor-pressure profiles is to measure temperature and vapor-pressure differences over more than one interval above the canopy (Brutsaert, 1982, p. 211). If it can be demonstrated that the profile dissimilarities do not persist to heights that would exceed the instrument height-fetch requirement, then the EBBR method might be suitable. Unfortunately, the Bowen ratio system used in this study was designed to measure temperature and vapor-pressure differences over one interval only.

Another difficulty encountered when fluxes from an aerodynamically rough surface are estimated from measurements made within the surface sublayer is the presence of local horizontal gradients in the vertical fluxes that are induced by individual roughness elements. The presence of significant horizontal gradients invalidates the assumption of one-dimensional turbulent transport of heat and water vapor necessary for the EBBR method. Raupach and others (1980) suggest that instruments be placed at a height (z) of $z = h + 1.5D$, where h is height of the roughness elements, in meters, and D is average spacing between the roughness elements, in meters, in order to avoid effects of horizontal inhomogeneities. The dominant roughness elements at the forested sites in this study were the trees. Tree spacing was not measured in this study; however, average spacing between dominant and codominant trees at the pine flatwood site was probably on the order of 10 m. Instrument heights indicated by the equation of Raupach and others (1980) would be too great to meet the instrument height-fetch requirement for the pine flatwood site in this study. If an instrument height-fetch ratio of 0.01 is assumed and the criterion of Raupach and others (1980) is adopted, a suitable pine flatwood stand would have a fetch of about 1,500 m. The pine flatwood forests on the Carlton Reserve are patchy, cutover, and uneven-aged, and they are intermixed with marshes, sloughs, and hardwood forests. A homogeneous pine flatwood stand with a width of 1,500 m probably does not exist on the Carlton Reserve. A casual inspection of the dome-shaped cypress canopy indicated that the fetch at the cypress swamp site was not suitable for the use of the EBBR method (fig. 3).

Eddy Correlation.— Fetch requirements for the eddy correlation method are similar to those of the EBBR method; therefore, eddy correlation is probably a suitable method for estimating evapotranspiration from the dry prairie and marsh vegetation types, but probably is not suited for most sites within the pine flatwood and cypress swamp vegetation types. If suitable sites could be found, the eddy correlation method might be useful; however, there are operational difficulties associated with the use of the instruments on tall towers. First, it is difficult to ensure that the sonic anemometer and net radiometer are level when they are installed at the top of a tall tower. Sensor misalignment by only a few degrees can cause substantial errors in the computed turbulent fluxes (Brutsaert, 1982, p. 192). Net radiation measurements can be substantially in error if the radiometer is not level. Second, the eddy correlation instruments should be pointed in the predominant wind direction to avoid distortion in the wind field caused by the tower. If a retractable tower is used, it must be brought down to turn the instruments, which can lead to many periods of missing eddy correlation data if the wind shifts frequently. A system that could sense wind direction and turn the instruments automatically would be useful.

Potential Evapotranspiration.—Potential evapotranspiration methods can be used to obtain realistic estimates of evapotranspiration only if a relation exists between the two variates. Results from this study did not indicate that a relation existed between calculated potential evapotranspiration and evapotranspiration for the dry prairie sites. First, calculated annual potential evapotranspiration for the dry prairie vegetation type was 80 percent greater than annual evapotranspiration estimated using EBBR and ECEBBR measurements. The large difference between calculated potential evapotranspiration and evapotranspiration indicates that nonpotential conditions prevailed at the dry prairie sites. Relations between calculated potential evapotranspiration and evapotranspiration can be expected to be weak or nonexistent when nonpotential conditions prevail because water availability, not meteorological factors, limits evapotranspiration (Campbell, 1977, p. 141).

Second, patterns of seasonal variation in calculated potential evapotranspiration did not match patterns of seasonal variation in evapotranspiration (fig. 10). Potential evapotranspiration was lowest during the winter months and reached a maximum in late spring before the onset of the summer rainy season in west-central Florida. This seasonal pattern has been reported in other Florida studies (Smajstrla and others, 1984). Evapotranspiration was lowest during the winter months; however, it did not reach annual maximums until after the start of the summer rainy season in June. Similar seasonal patterns have been reported in basinwide water-balance studies for south Florida (Knisel and others, 1985). The reason for these differing patterns is probably that water availability at the surface, not meteorological factors, limits daily evapotranspiration during a substantial part of the year. The dry season extends into early summer as reflected by the depth to the water table (fig. 10). If water-table depth can be used qualitatively as an indicator of water availability at the surface, it is apparent that the maximum potential evapotranspiration occurs when water availability is near the annual minimum.

Finally, comparisons between calculated potential evapotranspiration and evapotranspiration do not indicate a relation between the two variates for the dry prairie sites (fig. 11*A*). Conceptually, potential evapotranspiration establishes the upper limit for evapotranspiration; results from the dry prairie sites are consistent with this concept. The lack of correlation between computed potential evapotranspiration and evapotranspiration in this study indicates that daily evapotranspiration usually was limited by moisture availability rather than by meteorological factors.

Potential evapotranspiration methods probably are not suitable for estimating evapotranspiration from pine flatwood or cypress swamp sites. Although potential evapotranspiration was not calculated for the forested sites, theory indicates that evapotranspiration can be appreciably uncoupled from available energy in conifer canopies. Conifer leaves are small and typically exhibit high vapor diffusion resistances, and transpiration from them responds much more strongly to changes in atmospheric vapor-pressure deficit than to changes in leaf radiation balance (Campbell, 1977, p. 143). Uncoupling between evapotranspiration and available energy was indicated by the examples of eddy correlation measurements discussed previously, which indicated weaker correlations between latent heat flux and available energy for the forested sites than for the unforested sites. Available energy typically dominates potential evapotranspiration calculations (fig. 10); therefore, potential evapotranspiration methods should be even less suited for estimating evapotranspiration from the forested vegetation types in this study than they are for the dry prairie vegetation type.

Potential evapotranspiration methods might be useful for estimating evapotranspiration for the marsh vegetation type. The correlation between calculated potential evapotranspiration and measured evapotranspiration during late spring and early summer at a marsh site in this study indicates that potential conditions were approached at that site and that meteorological factors were at least partly limiting to daily evapotranspiration (fig. 12*A*). The peat on the floor of the marsh was very wet when the measurements were made, but there was no standing water. It was noted during the measurements that the plant canopy was composed largely of senescent and standing dead vegetation. Above ground physiologic activity and associated transpiration from canopy plants probably were near annual minimums. As a result, the plant materials most likely had a mulching effect by shading the moist peat surface and by obstructing eddy transport between the surface and the atmosphere. Because of the vegetative mulch, it would be anticipated that evapotranspiration from the marsh would not be as great as the evapotranspiration from a dense stand of physiologically active plants with similar aerodynamic properties; however, the correlation between potential evapotranspiration and measured evapotranspiration did indicate that daily evapotranspiration was controlled by meteorological factors to an extent that made calculated potential evapotranspiration a relevant predictor for evapotranspiration. Because of seasonal changes in moisture availability caused by such factors as changes in marsh water levels, plant canopy composition, and the growth stage of canopy plants, the relation between potential evapotranspiration and evapotranspiration could change during the year. Attempts to estimate seasonal and annual evapotranspiration at marsh sites from calculated potential evapotranspiration should be accompanied by further investigation to determine the degree to which the relation between those two variates changes during the year.

It might be possible to obtain sufficiently representative estimates of evapotranspiration for the marsh sites through the use of methods that are less demanding than the method described Penman (1956). The radiation component of the Penman equation was correlated with measured evapotranspiration (fig. 12*B*). If the radiation component is used to predict evapotranspiration, measurements of vapor pressure and windspeed and estimates of canopy aerodynamic properties are unnecessary.

The slope of the relation between the radiation component of the Penman equation and evapotranspiration has been an object of several evapotranspiration studies. Brutsaert (1982, p. 218) has summarized some of these studies and cites slopes that range from approximately 1.29 for open water to 1.05 for moist soil. The slope of the relation obtained for the marsh site in this study during late spring and early summer (0.83) was lower than the slopes given by Brutsaert (1982). When the slope is determined from data collected under potential conditions, it typically is greater than 1 because of the implicit contribution of factors in the aerodynamic component of the Penman equation. If moisture availability at the surface were partially limiting to evapotranspiration, the experimental slope should reflect this, and the slope would be less than that determined under potential conditions.

The slope of the relation between the radiation component of the Penman equation and evapotranspiration could vary during the year as moisture availability varies. Seasonal variation in the slope could also be caused by seasonal changes in atmospheric vapor deficit, air temperature, and windspeed, as well as the composition of the plant canopy and the growth stage of canopy plants. The slope could be greater than 0.83 during the summer rainy season when plants in the marsh are physiologically active. Attempts to predict seasonal and annual evapotranspiration at marsh sites from the radiation component of the Penman equation should be accompanied by an investigation to determine the degree to which the slope of the relation changes during the year.

Estimates of Annual Evapotranspiration

Seasonal variation of evapotranspiration was described for each vegetation type by pooling EBBR and eddy correlation data among sites and years. Eddy correlation data were used to compute evapotranspiration by the standard eddy correlation, ECEBBR, and ECEBR methods. Measured evapotranspiration was averaged by month, and monthly values were connected with lines to approximate the annual curve for evapotranspiration. Annual evapotranspiration was calculated from the area under the connected line segments. Except for the pine flatwood site, nighttime evapotranspiration was assumed to be negligible and was not included in the annual estimates. The annual estimate for the pine flatwood site was first calculated based on daytime measurements and was then increased by 11 percent, which was the average nighttime evapotranspiration estimated by using the standard eddy correlation method.

Daily evapotranspiration for all four vegetation types varied seasonally (fig. 13). The monthly average of measured evapotranspiration remained near the annual minimums from January to April of the composite year. Measured evapotranspiration

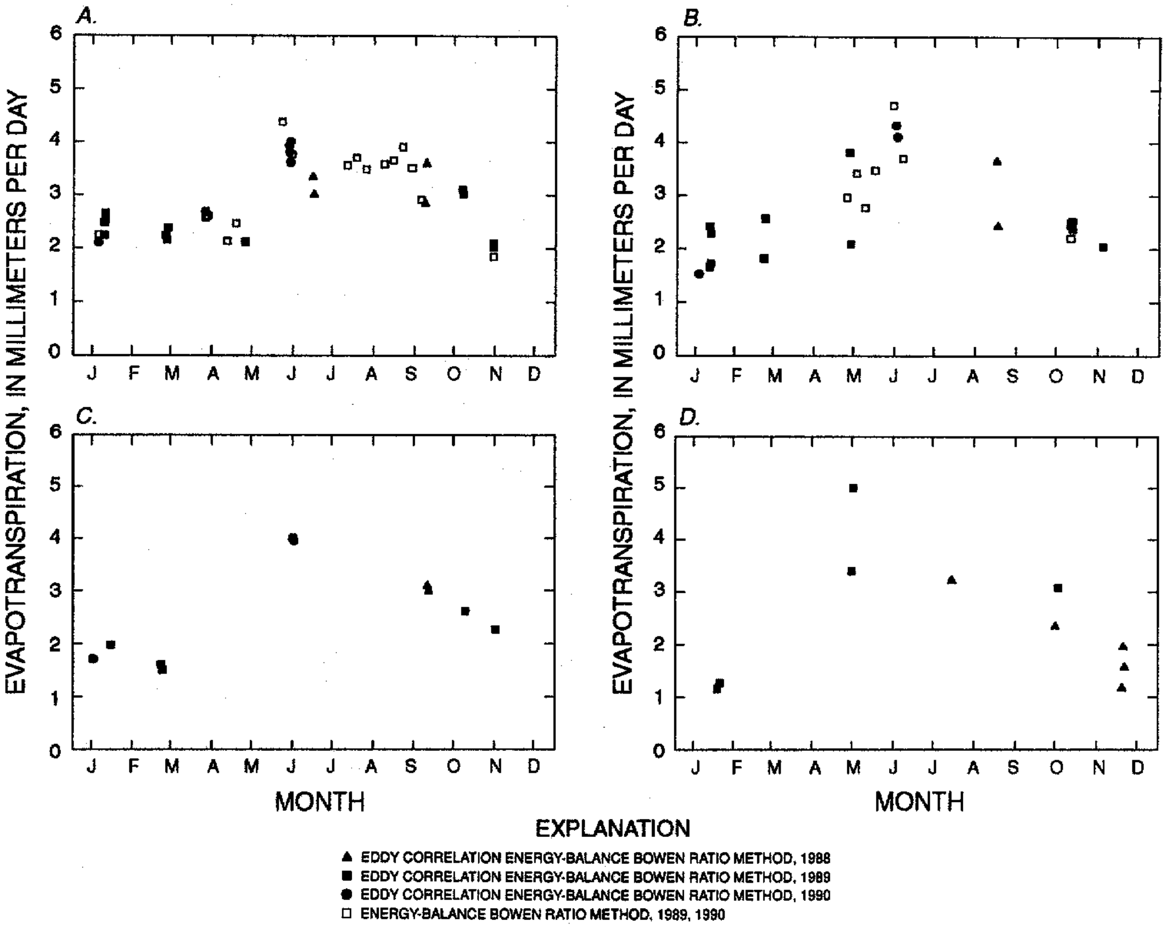

Figure 13. Daytime evapotranspiration estimated using the energy-balance Bowen ratio method and the eddy correlation energy-balance Bowen ratio method for sites within *(A)* dry prairie, *(B)* marsh, *(C)* pine

increased to near the annual maximum during May or June for each vegetation type and returned to near the annual minimum during September, October, or November.

Data sets with which annual evapotranspiration could be calculated were not available in equal number for each vegetation type. The dry prairie vegetation type produced the largest number of data sets, and evapotranspiration measurements were available for all months except February and December. Work at the marsh sites yielded evapotranspiration measurements during 8 months of the year. Data sets from the cypress swamp vegetation type yielded only 10 daily evapotranspiration measurements made during 5 months. Work at the pine flatwood site produced 10 daily evapotranspiration measurements made during 6 months. Results of daytime evapotranspiration measurements are given in tables 3 and 4.

Estimated annual evapotranspiration varied depending on the computation method used for eddy correlation data. Annual estimates were lowest for all sites when evapotranspiration was computed by the standard eddy correlation method; estimates were highest when evapotranspiration was computed using the ECEBR method (table 5). The effect of the eddy correlation computation method on annual estimates varied by vegetation type. The difference ranged from 260 mm/yr for the dry prairie vegetation type to 1,060 mm/yr for the cypress swamp vegetation type.

Uncertainties in Annual Evapotranspiration Estimates

The annual evapotranspiration estimates developed in this study were subject to errors in estimating daily evapotranspiration and to sampling errors. Errors in estimating evapotranspiration by the EBBR and eddy correlation methods can occur if the nature of turbulent transport in the surface sublayer where the measurements are made departs from the ideal conditions on which the methods are based. For example, assumptions on which the two methods are based are not valid if there are substantial horizontal gradients in vertical fluxes of momentum, heat, or water vapor. These assumption-based errors could not be quantified in this study. Errors in estimating evapotranspiration also can arise due to errors in measuring or estimating the variables that are necessary for the application of the EBBR or eddy correlation methods.

To provide some insight about effects of typical measurement errors, a sensitivity analysis was performed using the three most important input variables for the EBBR method: net radiation and the vertical air temperature and vapor-pressure differences. The analysis was conducted for EBBR measurements made at a dry prairie site from July 31 to August 9, 1990. One of the challenges to performing a meaningful sensitivity analysis is the determination of typical measurement errors. Measurement errors are difficult to determine under field conditions and they often must be estimated. The potential measurement errors used in the sensitivity analysis were estimated based on how large they reasonably can be expected to be under typical field conditions.

Based on results from a study by Weeks and others (1987), who used a net radiometer similar in design to the Model Q–5, a 10–percent error for net radiation measurements was used in the sensitivity analysis. Errors in estimating evapotranspiration were roughly proportional to net radiation measurement errors, with the effect of overestimating net radiation being slightly greater (table 6). Resolution of the vertical air temperature difference using the mean-gradient system was 0.006°C under ideal conditions. In the field, differential heating of the thermocouple connectors and contamination of the thermocouple junctions could have caused greater measurement errors. A 0.1°C error is approximately 20 percent of the typical vertical air temperature difference measured at midday and was assumed to be a maximum error for the sensitivity analysis. The effect of a positive 0.1°C measurement error was to decrease computed evapotranspiration, and the effect of a negative measurement error was to increase computed evapotranspiration. The vertical vapor-pressure difference was determined from the difference in dewpoint temperature at two heights. Resolution of the dewpoint temperature difference was limited by the electronic stability of the dewpoint hygrometer (±0.05°C). The effect of a positive 0.05°C error in determining the dewpoint difference was to increase the computed evapotranspiration, and the effect of a negative 0.05°C error was to decrease computed evapotranspiration. The error in computing the dewpoint could have been much greater when condensation formed in the system tubing; however, the available energy usually was small when that happened and the effect on computed evapotranspiration was minimal.

Table 3. Daytime evapotranspiration calculated using eddy correlation and energy-balance measurements at dry prairie, marsh, pine flatwood, and cypress swamp sites

[Values are in millimeters. ECEBBR, eddy correlation energy-balance Bowen ratio; ECEBR, eddy correlation energy-balance residual]

Date[1, 2]	Standard eddy correlation	ECEBBR	ECEBR
First dry prairie site			
July 1, 1988 (1)	2.15	3.35	4.42
July 2, 1988 (1)	1.84	3.02	4.14
September 23, 1988 (1)	1.98	2.86	3.47
September 24, 1988 (1)	2.39	3.61	4.26
January 24, 1989 (1)	1.67	2.48	2.94
January 24, 1989 (2)	1.47	2.24	2.96
January 25, 1989 (1)	2.19	2.53	2.79
January 25, 1989 (2)	1.75	2.65	3.36
March 11, 1989 (2)	1.72	2.24	2.98
March 12, 1989 (2)	1.36	2.15	3.42
March 13, 1989 (2)	1.16	2.38	3.61
Second dry prairie site			
May 10, 1989 (2)	1.65	2.11	3.67
October 21, 1989 (2)	1.90	3.02	3.51
October 22, 1989 (2)	1.98	3.10	3.82
November 14, 1989 (1)	1.29	2.02	2.33
November 14, 1989 (2)	1.69	2.09	2.20
January 20, 1990 (2)	1.37	2.10	2.73
Third dry prairie site			
April 10, 1990 (1)	2.44	2.68	3.59
June 12, 1990 (1)	3.12	3.93	4.34
June 12, 1990 (2)	2.14	3.81	4.75
June 13, 1990 (1)	3.07	4.00	4.45
June 13, 1990 (2)	1.52	3.61	4.62
First marsh site			
August 31, 1988 (1)	2.15	3.67	4.44
September 1, 1988 (1)	1.75	2.44	3.30
January 26, 1989 (2)	1.02	1.66	2.25
January 27, 1989 (2)	1.05	1.73	2.31
March 18, 1989 (1)	2.16	2.00	1.90
March 19, 1989 (1)	2.66	2.76	2.82
May 12, 1989 (2)	2.57	3.81	4.55
October 26, 1989 (1)	1.84	2.50	2.88
October 26, 1989 (2)	1.39	2.45	3.03
First marsh site—Continued			
October 27, 1989 (1)	2.06	2.53	2.84
October 27, 1989 (2)	1.47	2.37	2.88
November 19, 1989 (2)	1.23	2.05	2.60
January 18, 1990 (2)	.89	1.53	2.23
June 16, 1990 (2)	2.56	4.33	5.54
June 17, 1990 (2)	2.43	4.11	5.14
Second marsh site			
January 26, 1989 (1)	1.21	2.42	2.59
January 27, 1989 (1)	1.58	2.29	2.71
March 8, 1989 (1)	1.61	1.82	2.00
March 9, 1989 (1)	1.92	2.57	3.07
May 12, 1989 (1)	1.77	2.09	3.50
Pine flatwood site			
September 26, 1988 (1)	3.11	3.11	3.25
September 27, 1988 (1)	2.61	3.01	3.23
January 29, 1989 (2)	1.07	1.98	2.69
March 8, 1989 (2)	1.22	1.60	1.95
March 9, 1989 (2)	1.31	1.51	2.09
October 24, 1989 (2)	1.49	2.62	3.17
November 16, 1989 (2)	1.36	2.27	3.25
January 16, 1990 (1)	1.69	1.71	1.68
June 15, 1990 (1)	2.48	4.00	5.14
June 16, 1990 (1)	2.65	3.95	4.88
Cypress swamp site			
July 29, 1988 (1)	1.96	3.23	4.66
October 15, 1988 (1)	1.30	2.36	3.37
December 4, 1988 (1)	.52	1.19	3.01
December 5, 1988 (1)	.54	1.97	3.37
December 6, 1988 (1)	.43	1.58	3.58
February 3, 1989 (1)	.64	1.16	2.86
February 4, 1989 (1)	.40	1.27	2.85
May 15, 1989 (1)	1.53	3.39	5.54
May 16, 1989 (1)	2.22	5.00	5.91
October 17, 1989 (2)	1.58	3.07	4.46

[1] When measurements were not made for a complete daytime period, a composite daytime period was generated from measurements made during consecutive days. The day listed was the day when most of the measurements were made.

[2] The number in parentheses indicates eddy correlation system 1 or 2.

Table 4. Daytime evapotranspiration, potential evapotranspiration, and potential evapotranspiration represented by the radiation component of the Penman equation at dry prairie and marsh sites

[Values are in millimeters. --, no data]

Date	Energy-balance Bowen ratio	Potential evapotrans-piration	Radiation component evapotrans-piration
Second dry prairie site			
November 14, 1989	1.84	--	--
January 20, 1990	2.25	--	--
Third dry prairie site			
April 9, 1990	2.03	4.44	3.34
April 10, 1990	2.67	5.38	4.51
April 11, 1990	2.59	3.81	3.29
April 12, 1990	2.95	6.39	5.05
April 13, 1990	2.82	6.05	4.65
April 27, 1990	2.49	7.66	4.79
April 28, 1990	1.79	3.54	2.31
April 29, 1990	2.23	6.99	5.32
April 30, 1990	1.67	5.90	4.55
May 1, 1990	2.42	6.14	4.63
May 2, 1990	2.44	7.84	5.49
May 3, 1990	2.32	7.00	4.75
May 4, 1990	2.50	6.60	4.80
May 5, 1990	3.47	6.60	4.91
June 5, 1990	4.68	7.28	5.70
June 6, 1990	3.77	6.13	4.96
June 7, 1990	3.49	6.54	5.21
June 8, 1990	4.81	4.63	3.83
June 9, 1990	5.02	5.53	4.68
June 10, 1990	4.41	5.31	4.43
June 11, 1990	3.60	6.05	4.60
June 12, 1990	3.65	6.57	4.87
June 13, 1990	3.56	6.92	4.81
July 28, 1990	3.99	7.48	5.39
July 29, 1990	2.19	3.10	2.66
July 30, 1990	3.27	4.34	3.74
July 31, 1990	4.65	7.19	5.59
August 1, 1990	4.20	6.21	4.96
August 2, 1990	3.05	4.65	3.59
August 3, 1990	4.30	7.05	5.56
August 4, 1990	4.27	6.71	5.54
August 5, 1990	3.54	4.97	4.34
August 6, 1990	3.05	4.25	3.59
August 7, 1990	4.45	6.21	5.00
August 8, 1990	3.70	5.62	4.54
Third dry prairie site—Continued			
August 9, 1990	2.42	4.20	3.20
August 10, 1990	4.08	5.22	4.34
August 11, 1990	3.15	4.27	3.44
August 22, 1990	4.46	6.58	5.27
August 23, 1990	2.99	4.09	3.16
August 24, 1990	3.71	5.57	4.49
August 25, 1990	3.14	4.01	3.41
August 26, 1990	3.54	4.44	3.81
August 27, 1990	4.13	5.83	4.85
August 28, 1990	3.41	4.34	3.69
August 29, 1990	3.99	5.30	4.47
August 30, 1990	3.59	4.64	3.90
August 31, 1990	3.07	4.15	3.60
September 1, 1990	3.85	5.09	4.31
September 2, 1990	3.70	5.38	4.39
September 3, 1990	4.31	6.14	4.73
September 4, 1990	4.11	6.75	5.03
September 5, 1990	3.26	5.20	4.09
September 6, 1990	4.34	6.10	5.02
September 7, 1990	4.01	5.72	4.80
September 8, 1990	3.55	4.79	3.98
September 9, 1990	4.23	6.21	4.86
September 10, 1990	2.95	4.01	3.09
September 11, 1990	3.68	5.10	4.20
September 12, 1990	3.89	6.04	4.54
September 13, 1990	2.82	4.61	3.62
September 18, 1990	3.83	5.56	4.27
September 19, 1990	2.76	4.35	3.29
September 20, 1990	3.07	5.29	4.13
First marsh site			
October 26, 1989	2.20	--	--
May 22, 1990	2.01	3.25	2.31
May 23, 1990	3.23	4.29	3.38
May 24, 1990	2.22	3.60	2.60
May 25, 1990	3.20	7.50	4.65
May 26, 1990	2.40	4.86	3.48
May 27, 1990	3.46	4.50	3.94
June 16, 1990	4.69	8.65	5.79
June 17, 1990	4.10	6.76	4.36
June 18, 1990	3.32	4.81	3.54

Table 5. Estimates of annual evapotranspiration for sites within dry prairie, marsh, pine flatwood, and cypress swamp vegetation types developed by linear interpolation of daily evapotranspiration estimates for the years 1988–90

[Eddy correlation data were analyzed by three different methods: standard eddy correlation method; ECEBBR, eddy correlation energy-balance Bowen ratio; ECEBR, eddy correlation energy-balance residual. For the dry prairie and marsh vegetation types, eddy correlation results were pooled with results obtained from a mean-gradient energy-balance Bowen ratio system. Values are in millimeters]

Vegetation	Standard eddy correlation	ECEBBR	ECEBR
Dry prairie	820	1,010	1,180
Marsh	720	990	1,180
Pine flatwood	780	1,060	1,300
Cypress swamp...........	480	970	1,540

Eddy correlation measurements also were potential sources of errors, and the persistent energy-balance closure problems left much uncertainty concerning the magnitude of those errors. Because the data indicate that the energy-balance closure errors occurred as a result of failure to accurately determine available energy, or because turbulent fluxes were underestimated, the range of uncertainty in evapotranspiration could be approximated by assuming that actual evapotranspiration was within a range bounded by results from the standard eddy correlation computation and the ECEBR result. Those ranges were quite broad for all of the vegetation types (table 3).

Sampling errors also were potential sources of errors in the annual evapotranspiration estimates. The evapotranspiration measurements were necessarily made during relatively few days and at relatively few sites within each vegetation type. The low temporal and spatial sampling intensity, which is typical of the studies employing the labor-intensive EBBR and eddy correlation methods, limits the confidence that can be ascribed to the annual evapotranspiration estimates. The authors are not aware of any objective method with which to quantify errors associated with the annual evapotranspiration estimates that were due to incomplete sampling.

Annual evapotranspiration could have varied during the 3 years of data collection, and the lack of complete yearly data sets did not allow for comparisons between years. Annual differences were probably not large. Eddy correlation measurements made at about the same time during different years generally were similar (fig. 13).

Because of the uncertainty concerning errors associated with the annual evapotranspiration estimates and because estimates developed for each vegetation type using the ECEBBR method differed by 70 mm/yr or less (table 5), it probably is not reasonable to conclude, on the basis of this study, that annual evapotranspiration varied significantly among the dry prairie, marsh, and pine flatwood vegetation types. Although the contrasting physiognomies indicate that differences in evapotranspiration should have existed among the vegetation types, the available evidence is insufficient to conclude that differences did exist.

Despite the degree of uncertainty that exists in the annual estimates developed in this study and the uncertainty about how evapotranspiration varies from year to year, it is useful to note that the values obtained using the EBBR and ECEBBR methods agreed well with estimates obtained by water-budget studies for similar areas in west-central Florida. Long-term studies for the Upper Kissimmee basin in central Florida and the Green Swamp area in west-central Florida indicated average evapotranspiration was 1,080 and 1,020 mm/yr, respectively (Jones and others, 1984). Both areas are similar to the Carlton Reserve in that they contain numerous wetlands.

Table 6. Sensitivity of daily evapotranspiration computed by the energy-balance Bowen ratio method to changes in selected input variables

[Evapotranspiration response is expressed as a percent change of average daytime values computed for a dry prairie site for the period August 31 to September 9, 1990. The average computed evapotranspiration for the period was 3.80 millimeters per day. R_n, net radiation; ΔT_{dew}, dewpoint temperature difference; ΔT, air temperature difference; °C, degree Celsius]

Change in input variable	Percent change in average computed evapotranspiration
Decrease R_n by 10 percent	–10
Increase R_n by 10 percent	+13
Decrease ΔT_{dew} by 0.5°C	–6
Increase ΔT_{dew} by 0.5°C	+7
Decrease ΔT by 0.1°C.....................	+13
Increase ΔT by 0.1°C......................	–9

SUMMARY AND CONCLUSIONS

A study was conducted to evaluate the suitability of three micrometeorological methods for estimating evapotranspiration from selected areas of native vegetation types and to estimate annual evapotranspiration from each vegetation type. Evapotranspiration was estimated using the energy-balance Bowen ratio (EBBR) and eddy correlation methods. Potential evapotranspiration was computed from field measurements and the Penman equation. Field measurements were conducted intermittently from February 1988 to September 1990.

Results from this study support the conclusion that the EBBR method is suitable for estimating evapotranspiration from the unforested dry prairie and marsh vegetation types. The Bowen ratio obtained with the mean-gradient Bowen ratio system was very similar to the Bowen ratio obtained by turbulence measurements with the eddy correlation systems. The primary impediment to the use of this method for either vegetation type is the accurate determination of available energy. Computations for evapotranspiration are sensitive to the net radiation values that are used, and some uncertainties remain concerning the degree of measurement accuracy that is achievable under field conditions. The determination of subsurface heat flux at marsh sites presents particular challenges because the medium in which that flux travels changes seasonally from water to moist peat. Determination of heat storage in a water column requires the use of precise water-temperature measurements.

Eddy correlation might be a suitable method for estimating evapotranspiration from some of the vegetation types that were studied. The method can be used to compute evapotranspiration without relying directly on the surface energy balance; evaluation of energy-balance closure is useful for checking the consistency of measured energy-balance components, which include net radiation and subsurface heat flux, as well as sensible and latent heat fluxes as estimated by eddy correlation. Evaluation of energy-balance closure in this study consistently indicated that available energy was overestimated or that the turbulent fluxes were underestimated. Comparisons between two eddy correlation systems of the same design indicated that errors in measuring the turbulent fluxes were partly responsible for the closure errors.

Eddy correlation data were combined with measurements of net radiation and estimates of subsurface heat flux and canopy heat storage to compute evapotranspiration by two alternative methods. The rationale for use of the eddy correlation energy-balance Bowen ratio (ECEBBR) computation method was that systematic underestimation of the vertical wind component by eddy correlation could lead to fluxes being proportionately underestimated. The Bowen ratio would be preserved under such circumstances and could be used with the EBBR method to compute evapotranspiration. Comparison between evapotranspiration computed using a mean-gradient EBBR system with evapotranspiration computed using the ECEBBR method supported the conclusion that the eddy correlation measurements could be used to estimate the Bowen ratio. The rationale for the eddy correlation energy-balance residual method (ECEBR) was that the closure errors were due to an inability to measure fluctuations accurately in vapor density so that the latent heat flux was underestimated by eddy correlation. The alternative methods were useful for estimating possible ranges in evapotranspiration for the dry prairie and marsh sites, but the breadth of those ranges was large with respect to the estimates themselves, and this pointed out the need for future work to determine how instruments and procedures can be improved so that energy-balance closure can be improved.

The suitability of EBBR and eddy correlation methods for the forested vegetation types examined in this study is questionable. The users of these methods are confronted with several problems when an attempt is made to measure evapotranspiration at forested sites. These problems include (1) determination of the rate of canopy heat storage; (2) procurement of mean-gradient EBBR systems that can resolve the weak temperature and vapor-pressure gradients that are typically present above forests; and (3) the positioning of sensors close enough to the canopy to avoid horizontal gradients in the vertical fluxes of momentum, heat, and water vapor that are caused by advection while still placing them high enough to avoid local horizontal gradients that can be induced by individual roughness elements. This last problem can only be addressed by selecting sites that have sufficient fetch. Such sites might not exist among the heterogeneous forest communities that are typical of west-central Florida.

This study did not yield evidence to indicate that potential evapotranspiration estimates could be used to obtain realistic estimates of evapotranspiration for the dry prairie vegetation type. Potential evapotranspiration and evapotranspiration estimated by the EBBR method were not correlated on a daily basis; also, seasonal patterns of the two variates were dissimilar, and annual potential evapotranspiration was 80 percent greater than annual evapotranspiration estimates based on EBBR and eddy correlation measurements. Daily potential evapotranspiration was correlated with evapotranspiration estimated by the EBBR method at a marsh site during late spring and summer, indicating that potential evapotranspiration could be useful for estimating evapotranspiration for the marsh vegetation type during at least part of the year. Future comparisons can reveal how the relation between potential evapotranspiration and actual evapotranspiration varies by season at marsh sites.

Daily evapotranspiration, as estimated by the EBBR and ECEBBR methods, varied seasonally for each vegetation type. Maximum evapotranspiration occurred during May, June, or July for each vegetation type, and minimum evapotranspiration occurred during the months of November through March.

Annual evapotranspiration was estimated for each vegetation type by pooling daily EBBR and eddy correlation estimates among sites and among the 3 years during which field measurements were made. Three different annual estimates, which corresponded to the three different computation methods for eddy correlation data, were developed for each vegetation type. When the ECEBBR computation method was used, the annual evapotranspiration estimates ranged from 970 mm/yr for a cypress swamp site to 1,060 mm/yr for a pine flatwood site.

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