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Daily Evapotranspiration and Soil Moisture Estimates for the Northeastern United States

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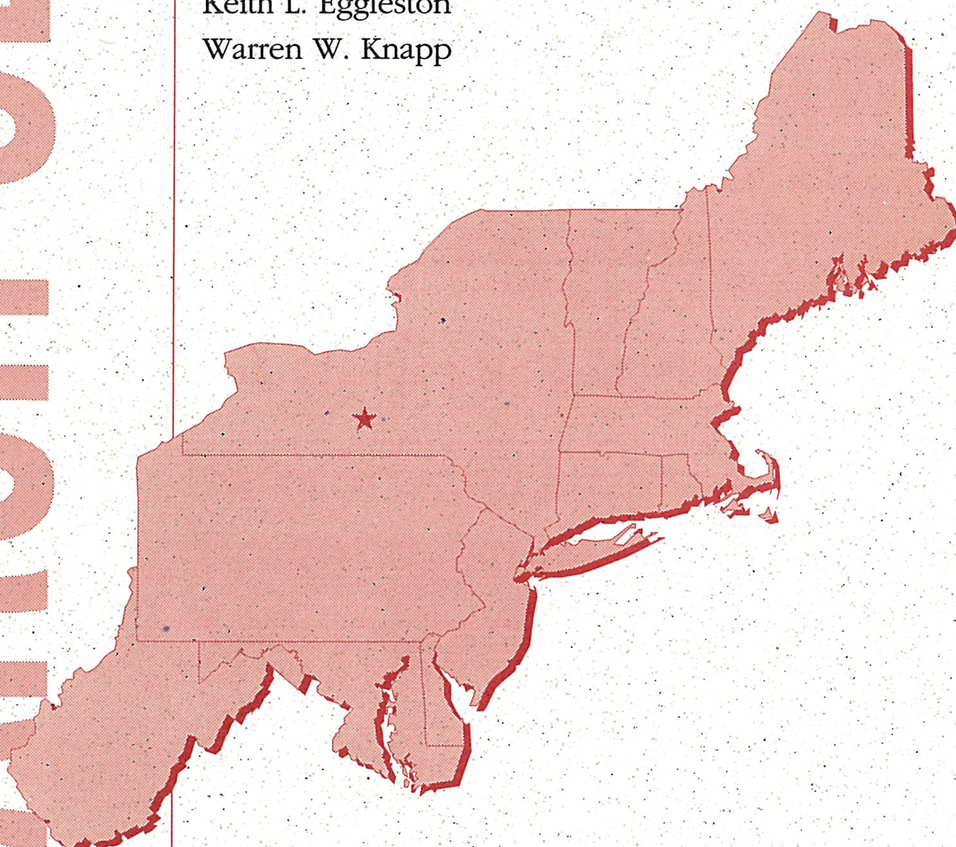

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INTRODUCTION

A recent survey of climate information users in the northeastern United States indicated a strong interest in current evapotranspiration (ET) and soil moisture values. Such data has a wide variety of applications including the planning of agricultural operations, flood potential forecasting, and the scheduling of urban lawn watering. Unfortunately, routine measurements of evaporation and soil moisture are not widely available in the region. Only about 30 cooperative network stations in the Northeast measure pan evaporation, and soil moisture is observed at a limited number of specialized stations. These measurements are not reported, however, with sufficient frequency to be useful in monitoring real-time ET rates or soil moisture status. In addition, the historical records of such observations show many gaps and inconsistencies.

Because the inventory of evaporation and soil moisture measurements is so limited, numerous methods to estimate ET and soil moisture status have been proposed and implemented. Broadly, such methods can be classified as either climatological or physically-based. Climatological methods are designed to estimate potential ET using routinely measured meteorological data such as air temperature, and do not relate well to actual situations where both crop and soil factors affect the rate of ET. Such methods are useful only for estimating maximum ET values over periods of a month or longer. Formally, potential evapotranspiration is defined as the rate of water loss from an extended surface of short green crop assuming that the crop fully shades the ground, exerts negligible resistance to the flow of water and is continually well supplied with water.

Perhaps the most widely used climatological method for estimating ET is that of Thornthwaite (1948). Based on mean monthly temperature (T_i), in °C, Thornthwaite's method estimates potential ET (ETP) for month i using the formula:

$$ETP_i = 1.6\ (10(T_i)/I)^a \tag{1}$$

where ETP has units of centimeters and the exponent a is given by:

$$a = 6.75\times10^{-7}\ (I^3) - 7.71\times10^{-5}\ (I^2) + 1.79\times10^{-2}\ (I) + 0.49. \tag{2}$$

I is a heat index given as:

$$I = \sum_{i=1}^{12} (T_i/5)^{1.514} \tag{3}$$

Application of the Thornthwaite method to estimate ETP over periods shorter than a month leads to significant errors because short-term temperature means are an unsatisfactory surrogate for net radiation, which physically drives evapotranspiration. Other climatological methods have been proposed by Blaney and Criddle (1950), and Jensen and Haise (1963).

Physically-based methods for estimating evaporation allow for more reliable, short-term estimates at the expensive of requiring considerably more meteorological input data. This is exemplified by the commonly used Penman Method (Penman, 1948) given in simplified form as:

$$E = (\Delta H + \gamma E_a)/ (m + \gamma) \tag{4}$$

where E is daily evaporation from an open water surface, Δ is the slope of the saturation vapor pressure curve at the mean air temperature, H is an estimate of net radiation and γ is the psychromatic constant. E_a, which relies on the saturation and actual vapor pressures, e_s and e_a, respectively, and the mean wind speed, u, is given as:

$$E_a = 0.35(e_s - e_a)(1 + u \times 10^{-2}). \tag{5}$$

Other physically-based methods for estimating evaporation, include those proposed by Thornthwaite and Holzman (1942), Swinbank (1951), and Suomi and Tanner (1958). A physically-based method for estimating evapotranspiration was developed by Monteith (1963).

Combining the methods of Penman and Monteith provides a physically-based means of estimating ET from surfaces other than open water. The Penman-Monteith equation (Monteith, 1965) calculates ET as:

$$\lambda E = \frac{\Delta(R_n - G) + \rho c_p (e_s - e_a)/r_a}{\Delta + \gamma (1 + r_s/r_a)} \tag{6}$$

where E = rate of water loss ($Kg\ m^{-2}s^{-1}$)
Δ = rate of change of e_s with temperature (mb °C^{-1})
R_n = net radiation (Wm^{-2})
G = soil heat flux (Wm^{-2})
ρ = air density ($Kg\ m^{-3}$)
c_p = specific heat of air at constant pressure (1005 JKg^{-1} °C^{-1})
e_s = saturation vapor pressure (mb)
e_a = actual vapor pressure (mb)
λ = latent heat of vaporization ($2.465 \times 10^6\ JKg^{-1}$)
γ = psychromatic constant (0.66 mb °C^{-1})
r_s = surface resistance (sm^{-1})
r_a = aerodynamic resistance (sm^{-1})

Due to its physical basis and its ability to provide reliable daily ET estimates, the Penman-Monteith equation is commonly used to estimate ET for a variety of surface types and locations. The Penman-Monteith equation forms the basis for the British Meteorological Office Rainfall and Evaporation Calculation System (MORECS) (Thompson et al., 1981). MORECS is used operationally in Great Britain to obtain weekly and monthly estimates of average evaporation and soil moisture deficits over

40 km x 40 km grid squares. The system relies on routinely observed daily meteorological data as its input. An important feature of MORECS is a scheme designed to determine potential and actual ET over a variety of different surface types. Using MORECS, such estimates can be obtained for open water, bare soil, grass, cereals, potatoes, deciduous trees, conifers, orchards and pastures.

The Penman-Monteith equation is also the primary means by which ET is calculated in the CERES-Maize corn simulation model (Jones and Kiniry, 1986). This model is used operationally by the Midwest Climate Center to estimate soil moisture status under corn crops in the midwestern U.S. (Kunkel, 1990). Although well suited for use in the Midwest, where corn is a widely grown agricultural crop, such a crop-specific model is not a good choice the Northeast due to the wide variety of land uses.

Because of this need for a more general ET model, the British MORECS has been modified and validated for use in the northeastern United States. Presently, historical and real-time estimates of potential ET from grass, evaporation from bare soil and standard evaporation pans, as well as actual ET from grass- and deciduous tree-covered surfaces are available for the region. In addition, soil moisture deficits can be calculated under grass, bare soil and deciduous trees. ET and soil moisture estimates can also be obtained for a variety of other crops, however the unavailability of reliable verification data for other surface covers has precluded validation of the model for other surface cover types.

MODEL DESCRIPTION

a. Evaporation

Calculation of ET values using Equation 6 requires several supplementary physical and empirical relationships with which to compute values for terms that are not routinely measured. Because solar radiation measurements are not widely available in the Northeast, daily estimates of downward shortwave solar radiation are calculated based on hourly cloudiness, dew point and station pressure observations using the methods described in DeGaetano et al., 1993. Net radiation is obtained by decreasing the short wave radiation estimate according to the surface albedo, and summing this estimate and the net long wave radiation given by Linacre (1968) as:

$$R_{LN} = \varepsilon\sigma T^4 \,[1.35(e_s/T)^{0.143} - 1]\,(0.6) \qquad (7)$$

where R_{LN} = net long wave radiation (Wm^{-2})
ε = emissivity (0.95)
σ = Stefan's constant (5.67×10^{-8} Wm^{-2} $°K^{-4}$)
e_s = saturation vapor pressure (mb)
T = shelter temperature (°K).

The constant 0.6 accounts for cloudiness assuming a constant cloud cover of five tenths.

During daylight hours, G is defined as the flux density of heat into the soil and is calculated as:

$$G_d = (0.3 - 0.03L)R_{Nd} \qquad (8)$$

where L is the leaf area index and R_{Nd} is daytime net radiation. For grass, L varies from 2.0 during winter (December-February) to 5.0 in summer (July-September). However, L is assumed to equal 3.33 when calculating G_d. The leaf area index used to calculate G_d for deciduous trees varies linearly from 0.1 during dormancy to 6.0 at full leaf. A similar linear decrease in leaf area index is assumed during senescence. For bare soil, L = 0.0. At night, an estimate of G is given by:

$$G_n = (D(G_d) - P)/(24 - D) \tag{9}$$

where D is the number of daylight hours and P is the average daily heat storage in soil (Whr m^{-2}). Monthly values of P were empirically determined by Wales-Smith and Arnott (1980) and are given in Thompson et al. (1981). It is assumed that the British heat storage values are suitable for use in the northeastern U.S. When estimating pan evaporation, G is set equal to 0.0.

Using the logarithmic wind profile and assuming neutral stability, r_a is given as:

$$r_a = (6.25/u) \ln(10.0/z_o) \ln(6.0/z_o) \tag{10}$$

where u is the wind speed (ms^{-1}) at a height of 10 m above the ground and z_o is the roughness length (m). Fixed roughness lengths of 1.5×10^{-2}, 5.0×10^{-3} and 5.0×10^{-4} m are assigned to grass, bare soil and water, respectively. For deciduous trees, roughness length varies linearly between 0.2 m at leaf emergence to 1.0 m for full leaf. During autumn, roughness length is linearly decreased from the full leaf value to a defoliated value of 1.5×10^{-2} m. Similarly, roughness length is linearly increased during the period of bud break in spring.

In MORECS, water may be extracted from both the soil and the crop. Thus, the surface resistance term incorporates resistances due to both the crop and soil. Daytime values of crop resistance are prescribed for each surface type. These values reflect a crop that is freely supplied with water and thus represent a minimum resistance associated with each crop type. For deciduous trees, the minimum resistance is set equal to 80 sm^{-1}, while for grass this value varies from 50 during winter to 40 sm^{-1} in the summer months. Despite the lack of crop cover, a relatively high crop resistance value of 600 sm^{-1} is used for evaporation from bare soil to ensure that transpiration is negligible.

MORECS assumes two soil moisture reservoirs. Water in the top reservoir (x) is freely available for ET, while water in the second reservoir (y) becomes increasingly more difficult to extract as soil moisture decreases. The contents of each reservoir can be subdivided into water available for evaporation (x_{SOIL} or y_{SOIL}) and water available for transpiration (x_{CROP} or y_{CROP}). In the case of bare soil, water can only be evaporated from x_{SOIL} or y_{SOIL}. Provided water exists in x, the crop resistance remains at the minimum value. Soil resistance, is set to 100 sm^{-1} until x_{SOIL} has been depleted. After this point, soil resistance increases according to the formula:

$$r_{SOIL} = 100Cx_{max}/(x_{SOIL} + x_{CROP} + 0.01Cx_{max}) \tag{11}$$

where r_{SOIL} is the soil resistance, x_{SOIL} and x_{CROP} are the amount of water contained in each reservoir, and Cx_{max} is the maximum amount of water that can be held in x_{CROP}. For potential evapotranspiration, r_{SOIL} remains at 100 sm^{-1} and crop resistance is set at the minimum value for grass.

Once the water in the x reservoir has been exhausted, r_{SOIL} is set to 10^4 sm^{-1} and the crop resistance (r_{CROP}) is increased proportionally to the water deficit of the y reservoir using the formula:

$$r_{CROP} = (r_{CROP})_{min}((2.5y_{max}/(y_{SOIL} + y_{CROP}))-1.5) \quad (12)$$

where y_{SOIL} and y_{CROP} are the amount of water contained in each reservoir, y_{max} is the maximum amount of water that can be held in the y reservoir and $(r_{CROP})_{min}$ is the minimum crop resistance value.

Daytime surface resistance, r_s, is related to r_{CROP} and r_{SOIL} by the expression:

$$r_s = r_{CROP}\, r_{SOIL}/((r_{SOIL}(1\text{-}A))+(r_{CROP}\, A)) \quad (13)$$

where $A = 0.7^L$. At night, when stomata are closed, r_s is given by:

$$r_s = 2500(r_{SOIL})/(r_{SOIL}(L)+2500). \quad (14)$$

However, when the surface is bare soil and all water in x_{SOIL} has been depleted, regardless of the time of day, r_s is specified as:

$$r_s = 100(3.5(1\text{-}(y_{SOIL}/Sy_{max})) + \exp(0.2(Sy_{max}/(y_{SOIL} - 1)))) \quad (15)$$

where Sy_{max} is the maximum amount of water that can be held in y_{SOIL}. For an open water surface, r_s equals 0.0.

b. Precipitation

Each of the soil reservoirs can be replenished by rainfall and, theoretically, by dew deposition. In cases where the soil surface is covered by vegetation, a certain amount of rainfall is intercepted by the plant canopy and is thus unavailable to the soil. The proportion of rainfall that can be intercepted by grass, P, is:

$$P = (1.0 - 0.5^L). \quad (16)$$

The amount of interception, I, is simply the product of P and the daily rainfall. However, I can not exceed 20% of the leaf area index, L (i.e., $I \leq 0.2L$). Particularly during summer, several individual showers may contribute to the daily rainfall total. In such cases, the interception associated with the first shower may evaporate prior to any subsequent rainfall. Thus, Thompson et al. (1981) suggest the calculated value of I be multiplied by an adjustment factor during the months from March through November. These adjustment factors are given in Table 1. During all months, however, I is limited to the daily rainfall total.

Table 1. Adjustment factors to allow for evaporation of interception resulting from multiple daily rainfall events.

Month	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov
Factor	1.2	1.4	1.6	2.0	2.0	2.0	1.8	1.4	1.2

Interception by deciduous trees is treated differently. Helvey and Patric (1965) present a regression-based approach for estimating interception of rainfall in eastern hardwood forests. During dormancy (trees are in a defoliated state), interception is given as:

$$I = 0.086R + 0.015 \tag{17}$$

Interception by trees in full leaf is calculated using:

$$I = 0.099R + 0.031 \tag{18}$$

where R is the daily rainfall and the date of full leaf is obtained using phenological data. During leaf emergence, I is linearly increased from its dormancy value. Conversely, during senescence, I is linearly decreased from its full leaf value.

When interception is present, evaporation of the intercepted moisture occurs prior to any evapotranspiration from the soil. After setting r_s to 0.0, the open water value, evaporation is calculated hourly until the foliage is completely dry (no interception). Subsequent hourly ET estimates are calculated using r_s given by Equations 13 and/or 14. If intercepted water still exists after 24 hours, the unevaporated interception is assumed to fall to the soil.

c. Dew Deposition

The formation of dew is assumed when nighttime evaporation is negative. In these instances, dew is treated as open water and nighttime evaporation is recalculated after setting r_s to zero. If this calculation again yields negative evaporation, the deposition of dew is assumed with the amount of dew equal to the absolute value of evaporation. Ensuing calculations treat dew deposition in the same manner as rainfall. If recalculation yields positive evaporation, nighttime evaporation is set to zero. In such cases it is assumed that only dew has evaporated.

d. Runoff

For surface types other than deciduous tress, runoff is assumed equal to zero unless both the x and y reservoirs are at capacity. In the case of trees, runoff is also assumed to occur if the daily rainfall exceeds 1.00 inch or regardless of the daily rainfall total, when x_{SOIL} is greater than zero. These criteria are based on the curve number method (USDA, 1972). Using a simplification of this method, the runoff from a tree covered surface is:

$$RO_{tree} = (R - 0.2D)^2/(R + 0.8D) \tag{19}$$

R is the daily rainfall (cm), and D is given by:

$$D = (x_{max} + y_{max}) - (x_{soil} + x_{crop} + y_{soil} + y_{crop}) \tag{20}$$

where x_{max} and y_{max} are the capacities of the x and y soil water reservoirs.

e. Water Budget Calculations

The maximum total amount of water available for ET from a specific crop (AW) is assumed to fill two soil moisture reservoirs. Water in the x reservoir, 40% of AW, is freely available for ET, while the remaining 60% of AW, which fills the y reservoir, becomes increasing difficult to transpire or evaporate as the contents of y decrease. The amount of water within each reservoir is further subdivided into water available for evaporation from bare soil (x_{soil} or y_{soil}) and water available for ET from a crop covered surface (x_{crop} or y_{crop}). For soil with typical water holding capacity, AW is assigned a value of 20 mm for bare soil; 125 mm for grass and 175 mm for trees. Thus, regardless of crop type, x_{soil} and y_{soil} can not exceed 8 and 12 mm, respectively for a soil with average water holding capacity.

Through the process of ET, water is withdrawn from x_{soil} until this sub-reservoir is empty. Subsequent ET draws water from x_{crop} until the entire x reservoir is exhausted. At this point ET draws water from the y reservoir, depleting y_{soil} before tapping the reserve stored in y_{crop}. Soil moisture is replenished in a similar manner. Rainfall must fill the x_{soil} sub-reservoir to capacity before replenishing any moisture deficit in x_{crop}. Once the x reservoir is at capacity, additional rainfall fills y_{soil} and finally y_{crop}. This sequence of ET and recharge qualitatively represents the decreasing availability of soil moisture for evaporation and/or transpiration. Such an assumption simplifies the process of specifying crop and soil resistances as soil moisture becomes increasingly depleted or recharged.

f. Modifications for winter conditions

Because MORECS was developed for a climate in which snowfall is uncommon, several modifications were required to adapt the model for use in the northeastern U.S. where snowfall is possible throughout the cold half of the year. These modifications are designed to assure that soil moisture conditions are correctly initialized at the start of the growing season. Precipitation is assumed to fall in liquid form throughout the year. Therefore, the liquid water contained in snowfall is assumed to immediately replenish the soil moisture reservoirs. Although these assumptions allow soil moisture conditions to be tracked through the winter, individual daily soil moisture values are overestimated when snow cover is present or the soil is frozen.

Two additional modifications are also incorporated when snow cover exists and/or the air temperature falls below 30° F. The surface resistance is set to 0.0 on days with snow cover, because the presence of snow implies that any evaporation will occur from an open, although frozen, water surface. At temperatures below 30° F, the value of λ, used in Eq. 6, is assigned the latent heat of sublimation (2.799×10^6 Jkg^{-1}).

VALIDATION

To assess the accuracy of the ET and soil moisture values estimated by the model, output values were compared with observations at several sites in the northeastern United States. Unfortunately, sets of high-quality soil moisture observations are extremely scarce, so much of the validation analysis is based on pan evaporation measurements. Daily pan evaporation observations were obtained from 4 sites in the region and compared with the corresponding model estimates of open water evaporation. Mean errors (ME), (model - observed), mean absolute errors (MAE) and root mean square errors (RMSE) were computed at each site. These results appear in Table 2.

The results in Table 2 indicate close agreement between the observed and modeled evaporation values. The model appears to be unbiased as indicated by the overall

Table 2. Mean error (ME), mean absolute error (MAE) and root mean square error (RMSE) associated with modeled pan evaporation at the indicated sites. Errors have units of inches. The daily average observed pan evaporation (PAN), period of record (Years) and number of daily observations (OBS) are also given.

Station	ME	MAE	RMSE	PAN	Years	OBS
Beltsville, MD	0.00	0.04	0.06	0.19	1985 - 1990	1043
Emmaus, PA	0.01	0.05	0.07	0.17	1985 - 1991	1278
Ithaca, NY	0.00	0.03	0.05	0.15	1984 - 1990	1277
New Brunswick, NJ	0.01	0.04	0.05	0.17	1985 - 1990	1003
All Stations	0.00	0.04	0.06	0.17		4601

mean error value of zero. Little bias is apparent at the individual stations as well. On average, individual evaporation estimates deviate from the observed value by approximately ±0.04 inches. This value is remarkably consistent among the four stations, as are the RMSE values. Errors were also calculated for monthly periods and for days with and without precipitation reported. Generally, these results were similar to those given in Table 2.

Model-derived soil moisture estimates under grass and bare soil were verified using weekly data collected at Rock Springs, PA (McKee, 1983). Figure 1 compares actual and model-derived soil moisture under grass during 1977. This growing season was generally characterized by a dry spring and moist early summer. Dry weather during the late summer and early autumn was followed by wet late-autumn conditions. Despite these frequent and rather abrupt changes in soil moisture conditions, the modeled values follow the observed values quite closely. The largest deviations between the two curves occur in mid-August and mid-September. During these periods, observed soil moisture exceeds the modeled value. Because coincident meteorological and soil moisture observations were unavailable, these deviations most likely result from differences in the amount of precipitation received at Rock Springs and the rain gauge site at State College, PA which is located approximately 10 miles to the northeast. Prior to 8 August, errors (model - observed) averaged -0.06

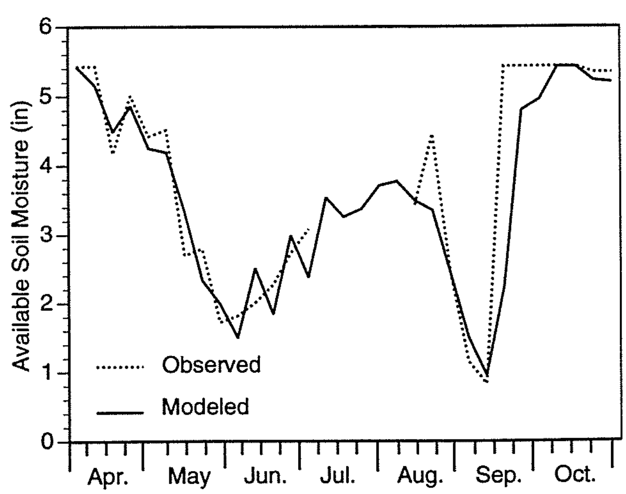

Figure 1. Comparison of modeled and observed soil moisture under grass during 1977 at Rock Springs, PA.

inches, while the mean absolute error and root mean square error were 0.34 and 0.40 inches, respectively. Over this period actual soil moisture averaged 3.40 inches. Similar agreement between observed and model-derived soil moisture values was achieved during other growing seasons and for bare soil.

Soil moisture observations taken within a deciduous forest were available from the Hubbard Brook Experimental Forest in New Hampshire. Figure 2a compares actual and model-derived soil moisture at this site during 1971. During this growing season, dry weather during the spring and early summer was followed by generally wet conditions during late-summer and autumn. As was the case for grass, the modeled values follow the observed values quite closely. However, particularly from July onward, a tendency for the model to overestimate soil moisture is apparent. This bias most likely results from runoff characteristics which are specific to the Hubbard Brook site. For instance, sharp increases in estimated soil moisture follow rainfalls of over

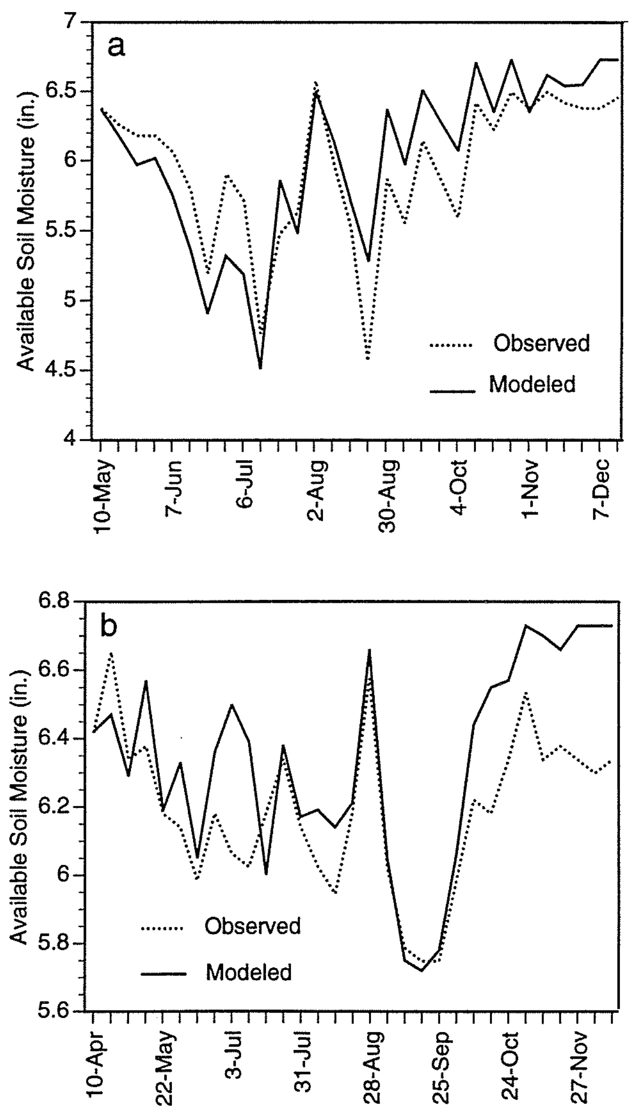

Figure 2. Comparison of modeled and observed soil moisture under a deciduous forest during (a) 1971 and (b) 1972. Growing season observations are at approximately seven day intervals.

1.8 inches on 27-28 August; 1.90 inches over the period from 12 to 14 September and 2.13 inches on October 9. Sharp increases in soil moisture are also associated with frequent rainfall events during 1972, particularly in autumn and the late spring and early summer (Fig. 2b). During 1971, mean errors averaged -0.07 inches, while the mean absolute and root mean square errors were 0.28 and 0.33 inches, respectively.

SUMMARY

Evapotranspiration and soil moisture measurements in the northeastern United States are relatively few in number and have limited periods of record. Since this semi-physical model requires only standard hourly surface observations and daily precipitation as inputs, current ET and soil moisture estimates can be generated at several dozen hourly observing sites within the region. Historical ET and soil moisture estimates dating back to 1948 can also be computed at approximately 30 sites in the Northeast. This data base of estimated ET and soil moisture values will provide essential data for applications ranging from drought and flood monitoring to the scheduling of urban lawn watering.

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