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ABSTRACT

The antecedent precipitation index (API) has been a useful indicator of soil moisture conditions for watershed runoff calculations and recent attempts to correlate this index with spaceborne microwave observations have been fairly successful. We show that the prognostic equation for soil moisture used in some of the atmospheric general circulation models (GCM) together with Thornthwaite-Mather parameterization of actual evapotranspiration leads to API equations. The recession coefficient for API is found to depend on climatic factors through potential evapotranspiration and on soil texture through the field capacity and the permanent wilting point. Climatological data for Wisconsin together with a recently developed model for global insolation are used to simulate the annual trend of the recession coefficient. Good quantitative agreement is shown with the observed trend at Fennimore and Colby watersheds in Wisconsin. This study suggests that API could be a unifying vocabulary for watershed and atmospheric general circulation modelers.

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1.0 INTRODUCTION

Soil moisture condition just before a rainfall has long been recognized as a major factor in watershed runoff predictions, although actual runoff depends upon many other factors, e.g., physical land features and vegetation. It is readily understood that when the soil is wet, runoff is greater than when it is dry. Soil moisture condition is also a major consideration in general circulation models (GCM) used in global models of the atmosphere because over the land the atmosphere exchanges heat with the surface and derives atmospheric moisture at a rate depending upon soil wetness. Local weather such as sea breeze and heat islands are also known to depend upon the state of the soil moisture conditions.

Since conventional methods for soil moisture estimation are impractical for large areas on a timely basis, hydrologists (see Linsley *et al.*, 1949) 'indexed' the antecedent precipitation as a means of estimating the moisture conditions in a watershed. These index values can as well be applied to the large area represented by grid cells used as input locations for global models of the atmosphere. Cells used in a GCM are generally larger than most watersheds, making the estimation of the average soil moisture even more difficult when using conventional measurements. The antecedent precipitation index (API) suggested by Lindsley has been correlated to microwave measurements made from space platforms (McFarland, 1976 and Blanchard *et al.*, 1981) thus offering the possibility of repeat measurements on a timely basis that could be readily converted to input for global models. Before extensive use of the API index, we need to determine the relation between soil moisture input in GCM and the index.

Clearly, soil left to itself will keep drying, and it is precipitation which can alter the moisture conditions to any significant degree when large areas are considered. Saxton and Lenz (1967) expressed the API in the following recurrence form,

$$API_j = K(API_{j-1} + R_{j-1}) \qquad (1)$$

where K is the recession coefficient and R_{j-1} is the amount of rainfall on (j - 1)th day. Once the recession coefficient and an initial value of API are given, the API equation forms a powerful model for simulating the moisture conditions. The model to date remains conceptual, since no formal basis for this equation has been provided. Starting with the prognostic equation for soil moisture used in some of the atmospheric GCMs we derive the API equation. Using climatological and soil texture data we siimulate the recession coefficient and show agreement with observations.

2. API MODEL

For horizontally-homogeneous vertically-stratified bare soils with known soil physical characteristics one can use the Philip-deVries equations (Philip and deVries, 1957) to simulate the moisture conditions. The numerical methods for the solution of these equations are complex, and becomes computationally quite expensive if 'exact' solutions are sought. Even accepting the inherent assumptions of these equations (e.g., no hysterisis), it becomes rather difficult to provide the required basic data for the solution when areas of a few hectares and larger are considered. We realize point measurement and calculation of soil moisture could be exact, but for large areas which may contain moisture impermeable areas, depressions and vegetation of varied type and density the meaning of 'exact soil moisture' becomes unclear (Engman, 1981). It is probably more appropriate to describe the moisture conditions through some statistical attributes. The API equation implicitly describes the moisture conditions as a stochastic variable following the first-order markov process (see discussion below).

In formulating the atmospheric GCM, Manabe (1969) considered the following prognostic equation for moisture conditions,

$$Z \frac{d\theta}{dt} = P(t) - E_a(t) \qquad (2)$$

where Z is the thickness of a fairly deep soil surface zone, θ is volumetric moisture, P(t) and $E_a(t)$ are, respectively, the rates of precipitation and actual evapotranspiration. Some other GCMs (e.g., Washington and Williamson, 1977) also include this soil moisture equation. It is essentially a one-layer moisture budget equation where the sub-layer remains passive with regard to moisture dynamics.

There is no rigid prescription for choosing the thickness of the surface layer, Z. For bare soils a thickness of 0.2m may be appropriate (Sellers, 1965; Jackson *et al.*, 1973), but for vegetated soils moisture extraction may occur from a deeper soil layer. (Note that actual root depth vary with plant species, climatic factors and soil factors.) Manabe (1969) considers the maximum water holding capacity of the surface layer, $Z\ \theta_{fc}$ = 0.15m (θ_{fc} being the field capacity), which for Colby (θ_{fc} = 0.482) and Fennimore (θ_{fc} = 0.364) watersheds to be studied here correspond to Z values of 0.32 and 0.42m. Noting that the analysis of recession coefficients for these watersheds is based on the soil moisture in the surface 0.31m layer, a compromized value of Z = 0.35m is thought to be reasonable. (The implication of choosing a fixed value of Z rather than $Z\ \theta_{fc}$ is that the rate of drying would vary with soil texture; sandy soils will dry faster than clayey soils. If $Z\ \theta_{fc}$ is fixed then the rate of drying for all soils will be equal, but the thickness of the soil layer which is drying would depend upon soil type. We have opted to fix Z in order to make comparisons of predicted rate of drying with Saxton and Lenz' observed rate of drying).

Actual evapotranspiration depends upon evaporative demand of the atmosphere, transpiration through the plants and soil water availability. Although the Penman equation (Penman, 1948) and some of its derivatives (e.g., Priestley and Taylor, 1972; Jury and Tanner, 1975; Thom and Oliver, 1977) are found to give fairly good estimates of potential evapotranspiration, no such equation has yet been found for actual evapotranspiration. Several linear and nonlinear relations between soil moisture and the ratio of actual and potential evapotranspiration are documented (Budyko, 1956; Thornthwaite and Mather, 1957; Holmes and Robertson, 1963;

Eagleman, 1971; Davies and Allen, 1973; Barton, 1979 and Marsh *et al.*, 1981). Whether any one of these relations is universally applicable is not known, but noting the simplicity and the assessment of Lowry (1959), Holmes and Robertson (1963) and Yaron *et al.* (1973), the Thornthwaite-Mather relation is chosen for this study

$$E_a = \left(\frac{\theta - \theta_w}{\theta_{fc} - \theta_w}\right) E_p \tag{3}$$

where θ_w is moisture at the permanent wilting point and E_p is potential evapotranspiration.

With eqn.(3), the solution of eqn.(2) can be written as

$$W(t + \Delta t) = \phi(t + \Delta t,\, t)\, W(t) + \int_t^{t + \Delta t} \phi(t + \Delta t,\, t')\, P(t')dt' \tag{4}$$

where

$$W(t) = Z[\,\theta(t) - \theta_w\,]$$

$$\phi(t + \Delta t,\, t) = \exp\left\{- \int_t^{t + \Delta t} \frac{E_p(t')dt'}{Z(\theta_{fc} - \theta_w)}\right\}$$

$$\Delta t = \text{time interval}$$

Note that if one postuates that the moisture conditions will vary stochastically according to a first order markov process then eqn.(4) can be written without going through the derivation. Our derivation of this equation considers the underlying physics and assumptions.

If time interval Δt is chosen as one day and j is used as the day index then eqn.(4) can be expressed in several alternate forms,

(i) If during the previous day there was no rain,

$$W_j = K\, W_{j-1} \tag{5a}$$

where

$$K = \exp\left(- \frac{E}{Z\,(\theta_{fc} - \theta_w)}\right) \tag{6}$$

E = daily total potential evapotranspiration for the previous day.

(ii) If rain during previous day occurred after sunset (See Baier and Robertson, 1966)

$$W_j \approx K\,W_{j-1} + R_{j-1} \quad (5b)$$

where R_{j-1} is the amount of rainfall. (Note that evapotranspiration occurs mostly during the sunshine hours, and it is practically zero during the rain).

(iii) If rain intermittently occurred during the previous sunshine period (compare with eqn. (1)).

$$W_j = K(W_{j-1} + R_{j-1}) \quad (5c)$$

Clearly, any one of the above equations would be applicable depending upon the nature of the rainfall occurrance. The choice of a particular form for use throughout the year should be made only after analyzing the rainfall statistics of the location If rainfall statistics show that the occurence of rain is strongly baised toward night time hours then (5b) may be used, and if there is no bais then (5c) would be more appropriate.

3. SIMULATION OF RECESSION COEFFICIENT

The recession coefficient as given by eqn.(6) can be simulated knowing soil texture and climatic data. From Saxton and Lenz (1967) the maximum available water, $Z(\theta_{fc} - \theta_w)$, is 0.0735m and 0.105m respectively in the top 0.35m soil layer for Fennimore and Colby watersheds. Climatological data, actually observed for these watersheds are unavailable, and were therefore synthesized (Table 1) from published sources (Reitan, 1960; Kung *et al.*, 1964; Flowers *et al.*, 1969; Bryson and Hare, 1974, and Climatological data of the U.S. Weather Bureau). Recognizing the inherent data variability and the associated errors in calculating the potential evapotranspiration, a consistency check will be discussed.

Global insolation is a major factor determining potential evapotranspiration, and the quality of Weather Service data for insolation is questionable (Hoyt, 1978). We used a model developed by Choudhury (1981) to calculate the daily total insolation (see Appendix). A comparative illustration of the model for daily total insolation at Rockville, MD, is shown in Fig. 1. The pertinent climatological data (atmospheric precipitable water, turbidity, cloud fraction and optical thickness and surface albedo) used to simulate insolations at Wisconsin are given in Table 1 and the calculated insolations are shown in Fig. 2.

The daily total potential evapotranspiration is calculated using Penman equation (Thom and Oliver, 1977)

$$E = \frac{R_n}{L\,(1 + r)\rho} + \frac{2.6 \times 10^{-4}\, r\,(e_s - e)\,(1 + 0.54U)}{(1 + r)}\ \text{(m)} \qquad (7)$$

where

$$r = 1.192 \times 10^{-7} \left(\frac{P}{e_s}\right) T^2$$

$$R_n = (1 - \alpha)\, S - 4.98 \times 10^{-3}\, \epsilon_s\, T^4\, (1 - \epsilon_a)\left(1 - \frac{1.24\,\tau c^2}{6.62 + \tau}\right)$$

$$e_s = 6.11 \exp \frac{17.27\,(T - 273.2)}{T - 35.86}$$

$$\epsilon_a = 0.7 + 5.95 \times 10^{-5}\, e \exp\left({}^{1500}/T\right)$$

L is the latent heat of evaporation (2.47×10^6 J kg^{-1}), ρ is the density of water (1000 kg m^{-3}), e is vapor pressure (mbar), T is air temperature (K), U is wind speed (m/sec), α is surface albedo, P is surface air pressure (985 mbar), R_n is net radiation (J m^{-2}), ϵ_s surface emissivity (0.97), S is global insolation (Appendix), ϵ_a atmospheric emissivity (Idso, 1981), c is cloud fraction and τ is cloud optical thickness.

As a check of the climatological data and model equations, the *simulated* potential evapotranspiration and net radiation are shown in Fig. 3 together with the regression equation based on *observed* evapotranspiration and net radiation during July through September at Hancock, Wisconsin (Tanner and Pelton, 1960). Although simulated values are more-or-less within the standard error of estimate for the regression equation, we see that the calculated evapotranspiration are gen-

generally lower than what would be expected from the regression equation. Tanner and Pelton (1960) also found that the Penman equation somewhat underestimates the evapotranspiration.

The simulated annual trend of the recession coefficient for Fennimore watershed together with observations (Saxton and Lenz, 1967) is shown in Fig. 4 using both Penman equation and the regression equation from Tanner and Pelton (1960). Except for April-May period, the simulated trend is within the observed values. If instead of using the climatological cloud fraction data, the range of recession coefficient due to global insolation from totally clear to totally cloudy skies is calculated then as shown in Fig. 5 much of the scatter in the observation appear reasonable. One should, however, note that our modification of cloud fraction data without adjusting the other climatologic parameters may not depict a realistic situation since it is known (Sellers, 1965) that air temperature, for example, is correlated with global insolation and hence cloudiness. The higher recession coefficient for April-May period indicates that the rate of soil drying was slower than the calculated rate. During March soil remains largely snowcovered, and dormant and emerging vegetations blanket fairly wet soil during April-May. Evaporation is expected to be slower and probably does not occur from entire 0.35 m surface layer (assumed in this paper). Some discrepancy with observations is understandable. In the versatile water budget model (Baier and Robertson, 1966) the thickness of the soil layer involved in moisture dynamics is adjusted seasonally to take into account the changes in plant root activity and growth.

Saxton and Lenz (1967) observed that recession coefficients for the Colby watershed were generally higher than those for Fennimore. Since these two watersheds are only 240 km apart it is reasonable to assume that the climatological parameters would be about the same. Assuming this climatological equivalence, the difference in the recession coefficients should then arise from the difference in the soil texture which determines the available water, Z $(\theta_{fc} - \theta_w)$. From eqn. (6) one would obtain

$$K_{Colby} = K_{Fennimore}^{0.7} \tag{8}$$

The observed mean recession coefficient is 0.92 during July at Fennimore, which would imply the corresponding value for Colby should be 0.943. This calculated value for Colby agrees well with the observed value of 0.95.

4.0 CONCLUSION

We discussed a formal basis for the API model and a rational method for calculating the recession coefficient. To the extent that one can use the climatological (or pan evaporation) data to obtain potential evapotranspiration and soil texture information, it will be possible to calculate a good first approximation for the recession coefficient, and simulate moisture conditions in large areas. Microwave remote sensing of API would be useful in updating the atmospheric general circulation models and watershed runoff forecasting.

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APPENDIX: A PARAMETRIC MODEL FOR DAILY TOTAL INSOLATION

The insolation is clearly a major factor in calculating evapotranspiration. In the absence of direct observations, either climatological data or a parameteric model may be used. With a parametric model one can simulate insolation as it is affected by changes in cloud type and cloud cover as well as the atmospheric parameters and surface albedo. A parametric model (Choudhury, 1981) for calculating daily total global insolation is recently developed and tested against observations and a set of exact radiative transfer calculations. Pertinent equations of this model are given below.

The daily total global insolation (KJ m^{-2}) is

$$S = \frac{86.4\, S_0 r}{\pi} \left[h \sin\phi \sin\delta + \cos\phi \cos\delta \sin h \right].$$

$$T\,(m, c, \tau, \alpha, w, u, \beta, \rho) \tag{A-1}$$

S_o = 1353 (W m^{-2}); solar constant

$$r = 1 - 0.0335 \sin\left[\frac{2\pi\,(N - 94)}{365}\right]; \text{ mean sun-earth distance} \tag{A-2}$$

N: Julian date

$$h = \pi - \cos^{-1}(\tan\phi \tan\delta) \tag{A-3}$$

ϕ: latitude

δ: solar declination

$$m = 0.105 + \frac{1.275}{\cos(\phi - \delta)}; \text{ mean air mass} \tag{A-4}$$

c: fractional cloud cover

τ: cloud optical thickness

α_s: surface albedo

w: atmospheric precipitable water (cm)

u: ozone path length (cm)

β: turbidity

ρ: surface air pressure (mbar)

$$T(m, c, \tau, \alpha_s, w, u, \beta, \rho) = \frac{T(m, w, u, \beta, \rho)}{1 - \alpha_s \left\{ \frac{0.0658 + \beta}{0.9606 + \beta} \right\}} \left\{ \frac{1 - c + c\,T_{cl}}{1 - c\,\alpha_s\,\alpha_{cl}} \right\}$$

; atmospheric transmission function (A-5)

$$T_{cl} = \frac{0.97\,(2 + 3/_m)}{4 + 0.6\,\tau}$$; cloud transmission function (A-6)

$$\alpha_{cl} = \frac{0.6\,\tau}{4 + 0.6\,\tau}$$; diffuse albedo of cloud (A-7)

$$T(m, w, u, \beta, \rho) = [1 - \sum_{j=1}^{5} \epsilon_j]\,[(1 - S_a)(1 - S_d) + 0.5\,S_a + 0.75\,S_d]$$

; clear sky atmospheric transmission function for non-reflecting surfaces (A-8)

ϵ_j: absorption coefficients for water vapor (j=1), carbon dioxide (j=2), ozone (j=3), oxygen (j=4) and aerosols (j=5) given in Hoyt (1978)

S_a, S_d: scattering coefficients, respectively, for Rayleigh and aerosols given in Hoyt (1978)

Climatological data sources for the atmospheric parameters are cited in the text. Monthly average values of the cloud optical thickness are obtained by matching calculated and observed insolations.

FIGURE CAPTIONS

Figure 1. Comparison of calculated and observed (.) clear sky insolations at Rockville, MD. The unit for insolation is in accordance with observations; to convert into SI units, 1 ly = 41.87 KJ m^{-2}.

Figure 2. Calculated global insolation and net radiation for south-central Wisconsin.

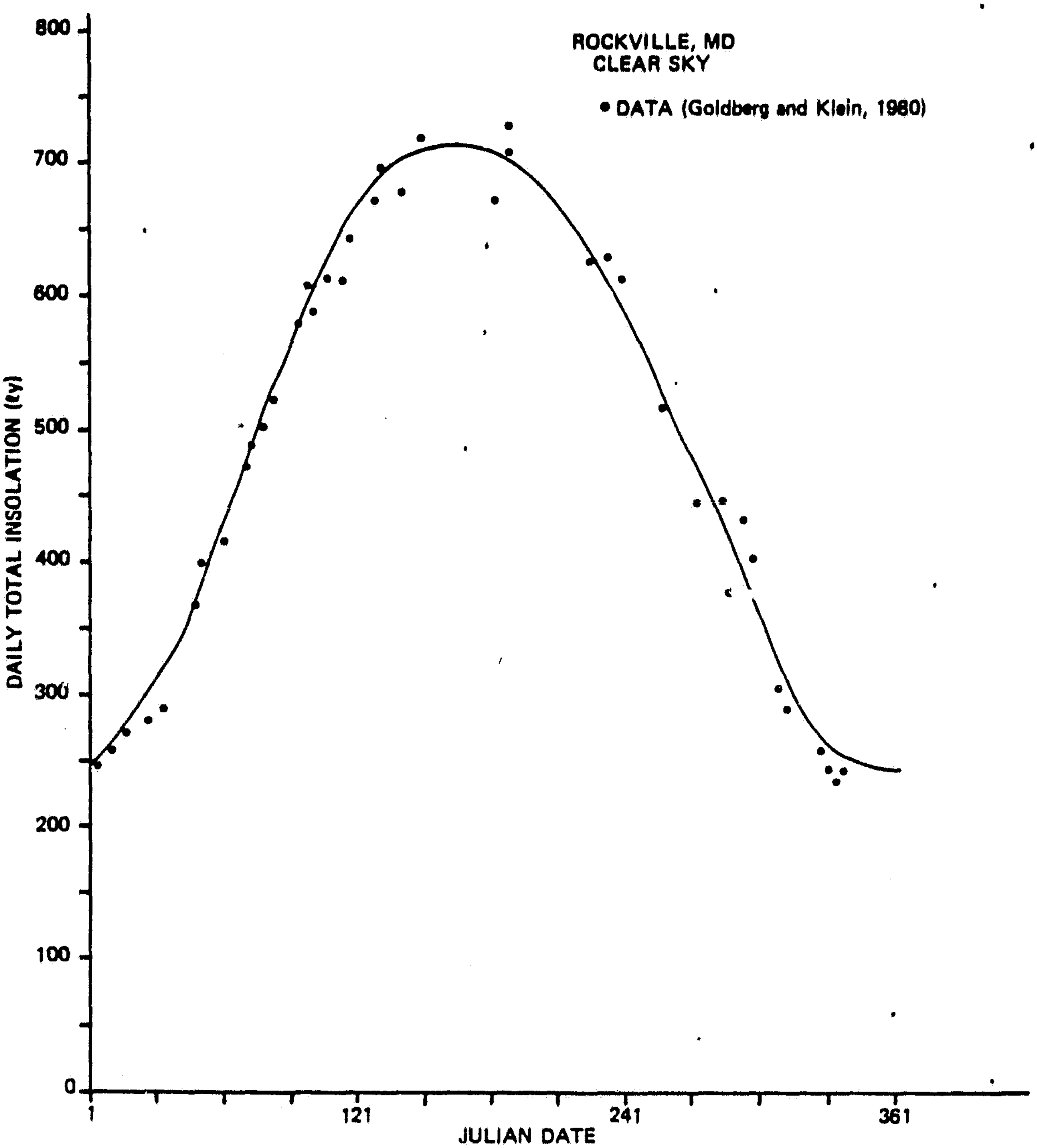
Figure 3. Comparison of simulated (.) Penman evapotranspiration and daily net radiation with the regression equation from Tanner and Pelton (1960).

Figure 4. Comparison of observed (.) and simulated annual trend of the recession coefficient (solid line based on Penman equation, dashed line based on Tanner-Pelton regression equation). The observed data points are from Saxton and Lenz (1967).

Figure 5. The range (vertical bars) of simulated recession coefficients due purely to cloudiness conditions. (Penman equation). The observed data points (.) are from Saxton and Lenz (1967).

TABLE 1: Climatological data for south-central Wisconsin (latitude 44°N) used in the simulation of the recession coefficient.

	March	April	May	June	July	August	Sept	Oct	Nov
Precipitable water (10^2 m)	0.75	1.1	1.7	2.5	2.9	2.9	2.1	1.55	0.95
Turbidity	0.06	0.07	0.08	0.09	0.09	0.08	0.07	0.06	0.05
Surface Albedo	0.4	0.3	0.22	0.22	0.22	0.22	0.22	0.22	0.22
Cloud Fraction	0.69	0.67	0.65	0.61	0.57	0.55	0.57	0.58	0.72
Cloud Optical Thickness	15.0	15.0	12.0	11.0	8.0	12.0	12.0	12.0	13.0
Air Temperature (°C)	−1.6	7.0	13.3	18.8	21.2	20.4	15.1	9.0	1.5
Vapor Pressure (mbar)	5.0	7.0	10.0	15.0	18.0	18.0	12.0	8.0	6.0
Wind Speed (m/sec)	5.0	5.2	4.2	4.0	3.3	3.3	3.4	3.6	4.0



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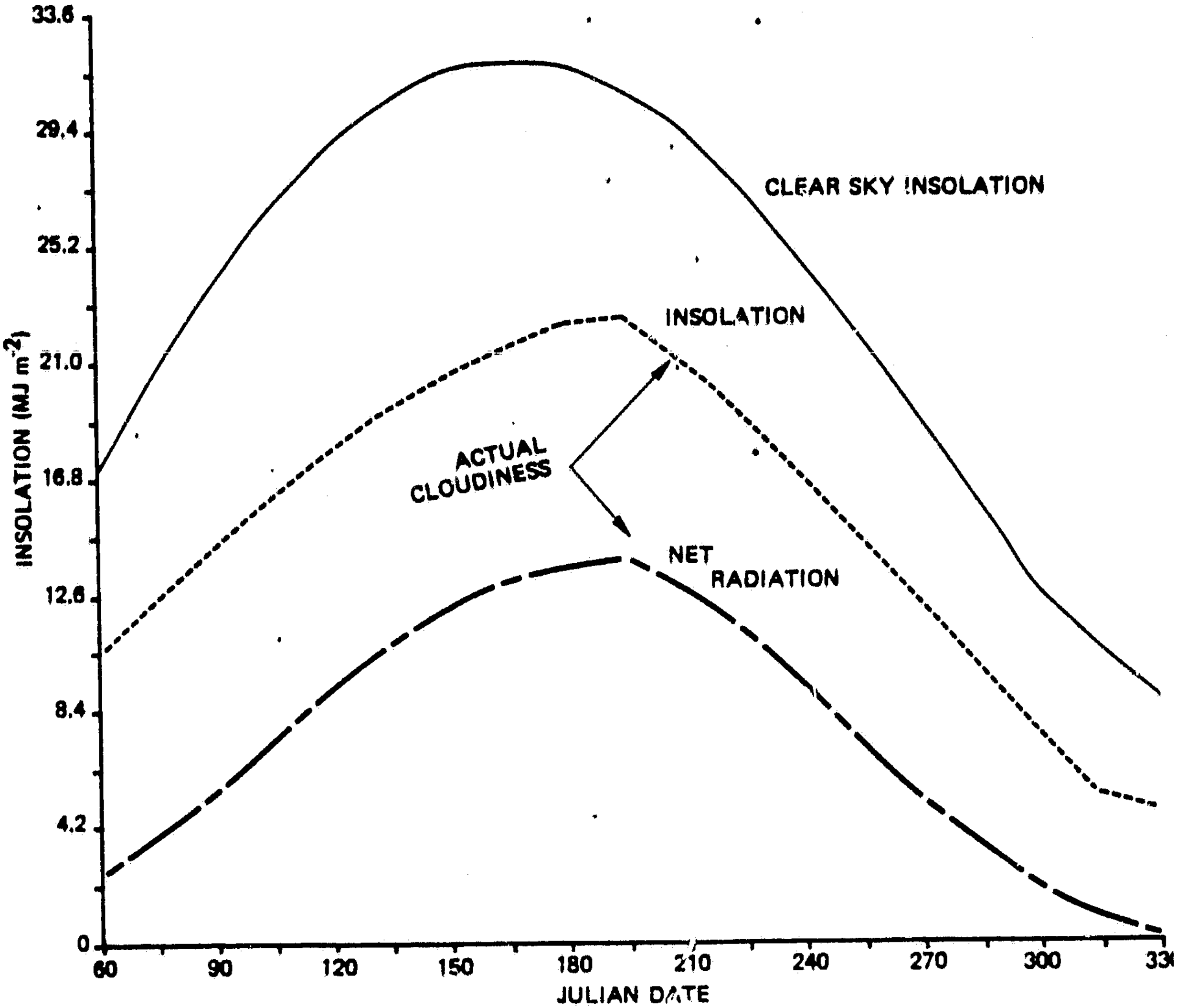
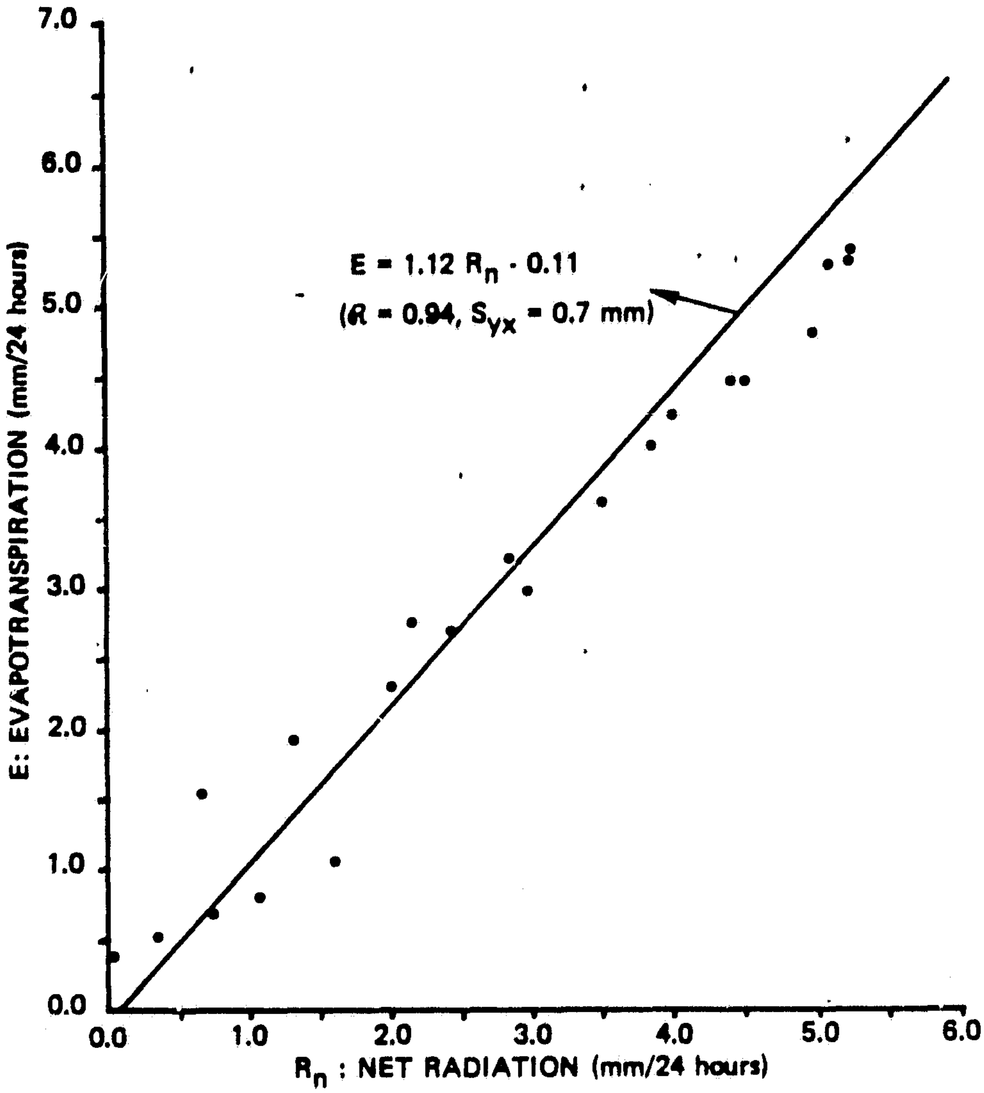


Figure 2. Calculated global insolation and net radiation for south-central Wisconsin.



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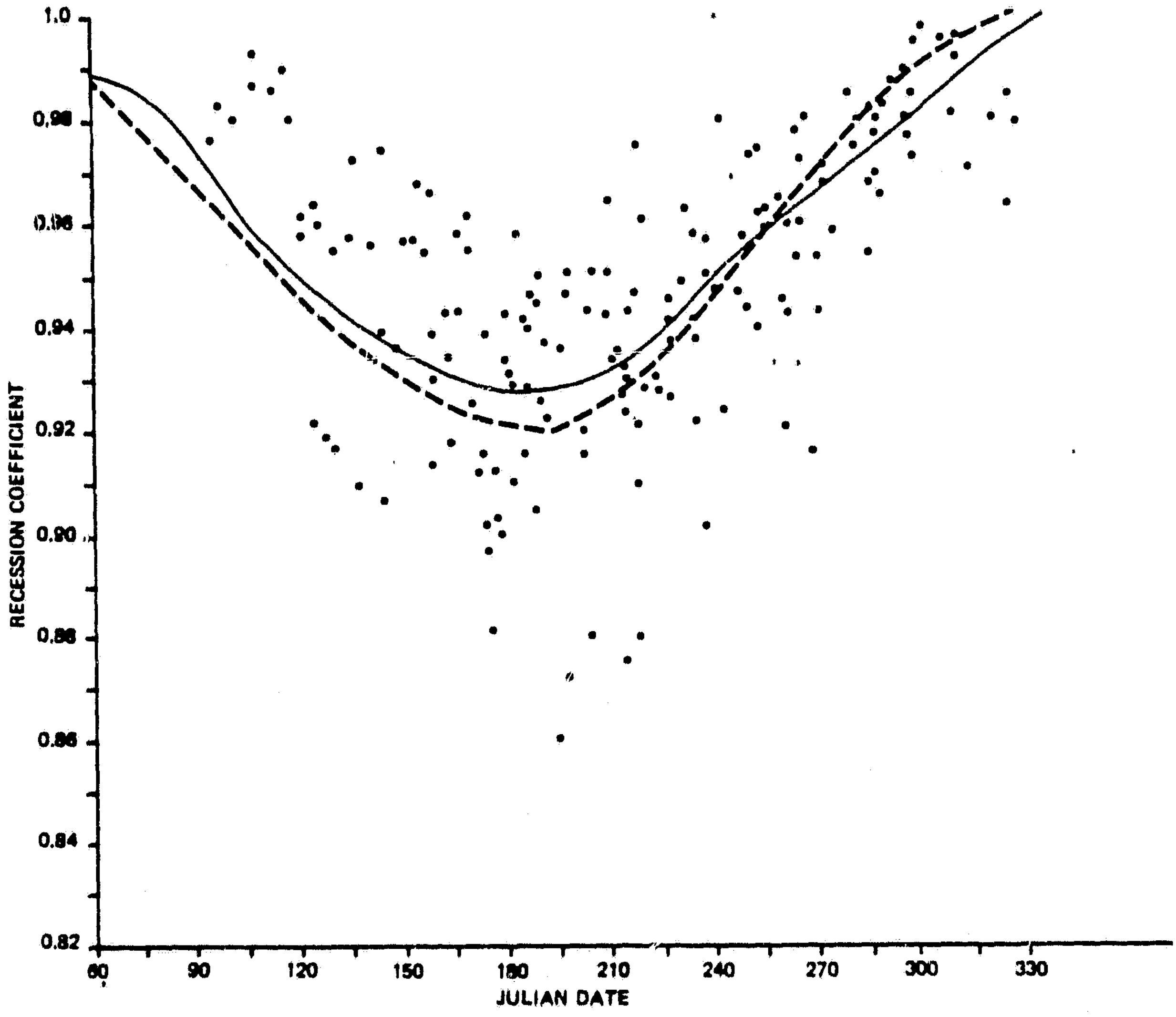

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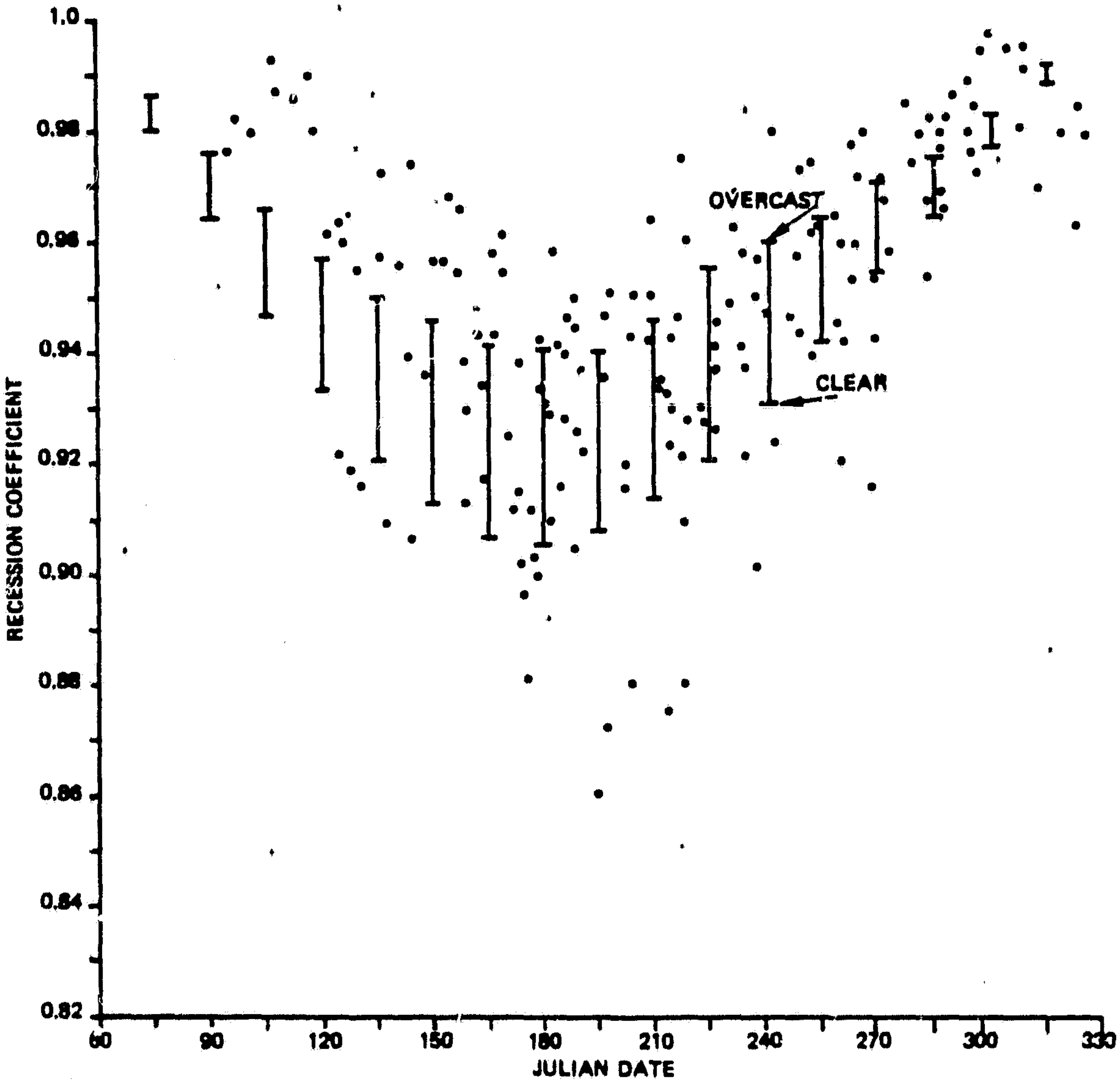

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