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Potential Evapotranspiration as Influenced by Wind[1]

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ABSTRACT

The contribution of wind to calculated potential evapotranspiration was investigated with applications for the climate of the Great Plains. A revised combination model proposed by van Bavel for computing instantaneous potential evapotranspiration was used. The model contains two terms that are expressions for the portions of potential evapotranspiration primarily due to net radiation and wind, respectively. With ambient water vapor pressure 20 mb, temperature 30C, wind 2, 4, and 6 m sec^{-1} at 2 m with a roughness length of 1 cm, contribution of wind dominant term to evaporation from a wet surface is 0.15, 0.30, and 0.45 mm hr^{-1}, respectively. At 10-mb vapor pressure and the same temperature, the corresponding evaporation rates are 0.22, 0.43, and 0.65 mm hr^{-1}. On representative and consecutive "nonwindy" and "windy" days at Manhattan, Kans. (average daily windspeeds at 45 cm were 0.88 and 2.26 m sec^{-1}), the wind dominant term contributed 33 and 113%, respectively, as much as the radiation dominant term to the total calculated potential evapotranspiration. For these 2 days the ratio of potential evapotranspiration to net radiation was 0.98 and 1.60.

Additional index word: **Evaporation.**

POTENTIAL evapotranspiration was defined by Penman (14) as "the amount of water transpired in unit time by a short green crop, completely shading the ground, of uniform height and never short of water." This definition has been extended by many agriculturists to include all crops, including forest (16). The fundamental condition that defines potential evapotranspiration as expressed by van Bavel (18) "is that the surface vapor pressure can be found from the surface temperature." He explains further that "when the surface is wet and imposes no restriction upon the flow of water vapor, the potential value is reached."

Knowledge of potential evapotranspiration has application in several areas of hydrological and agricultural research. It has been used extensively as an estimate of the amount of water required by crops. Tanner and Pelton (16) observed: "Since maximum yields of many of the agricultural crops appear to obtain when water is not limiting, potential evapotranspiration estimates are valuable in scheduling irrigation and in interpreting the results from many agronomic experiments."

When potential evapotranspiration is high, soil water potential must be maintained at a higher level so that the soil can supply water fast enough to meet the demands without placing the plant under stress. A good example of this is given by Denmead and Shaw (3). They found the average soil suction in the corn root zone when actual transpiration rate fell below the potential rate varied from 12 bars when the potential transpiration rate was 1.4 mm/day to 0.3 bar when the potential rate was 6 to 7 mm/day.

Potential evapotranspiration is also used as a basis for determining actual evapotranspiration (5).

In arid climates, heat advected from warm surrounding regions into cropped areas well supplied with irrigation water induces extremely high evapotranspiration rates (9, 19).

In recent studies in Great Plains States – Colorado (12), Nebraska (15), and Oklahoma (Personal communication with R. H. Griffin II) – energy used in evapotranspiration greatly exceeded the energy of net radiation. Even in subhumid Missouri, advection often contributes to evapotranspiration (2). These high evapotranspiration rates were generally associated with hot, dry winds blowing over soil and cropped surfaces.

In an investigation on evaporation of water from soils with wind or radiation, Hanks et al. (11) adjusted wind and radiation intensity so that evaporation rates from soil at the start were equal under both conditions (wind vs. radiation). Aristotle is credited (14) with asking whether sun or wind is the most important factor in evaporation and answering in favor of the wind because it carries the vapor away.

The purpose of this study is to characterize the contribution of wind to potential evapotranspiration for a climate typical of the Great Plains.

MODEL

Various methods have been proposed for estimating potential evapotranspiration from meteorological data. Tanner and Pelton (16) tested the energy balance approximation of Penman (14) for estimating potential evapotranspiration. They found that although the Penman estimates were highly correlated with detailed energy balance measurements, the absolute values of the Penman estimates were much too small. Others (17) have also found this to be so. In order to obtain suitable daily estimates of evapotranspiration with the Penman method, Tanner and Pelton (16) used a wind functoin that accounted for surface roughness and made direct measurements of radiation.

Van Bavel (18) tested a revised Penman version that included surface roughness and a wind function term for water vapor transfer. Tests of the model in Phoenix, Ariz., using open water, wet bare soil, and well-watered alfalfa gave excellent agreement of calculated and measured values on an hourly and daily basis under a variety of conditions including strongly advective.

The revised combination model for instantaueous potential evapotranspiration rate is shown here as given by van Bavel

$$LE_o = -\frac{\Delta/\gamma\, H + LB_v d_a}{\Delta/\gamma + 1} \text{ cal cm}^{-2} \text{ min}^{-1} \quad [1]$$

where L is the latent heat of vaporization in cal g^{-1}. Δ is first derivative of saturated water vapor pressure—temperature curve in mb C^{-1}, γ is psychrometric constant in mb C^{-1}; ratio of Δ/γ is a dimensionless number depending on the air temperature at elevation z_a. H is the sum of net radiation (R_n) and soil heat flux (S) in cal cm^{-2} min^{-1}, d_a is the vapor pressure deficit at elevation z_a in mb, and B_v is a turbulent transfer co-

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efficient for water vapor in cal cm^{-2} min^{-1} mb^{-1}, which was defined as

$$B_v = \frac{\rho \epsilon k^2}{p} \frac{u_a}{[\ln(z_a/z_o)]^2} \text{ g cm}^{-2} \text{ min}^{-1} \text{ mb}^{-1} \quad [2]$$

where ρ is the density of air in g cm^{-3}, ε the water-air molecular weight ratio, k the von Karman constant, p the ambient pressure, u_a the windspeed at height z_a in cm min^{-1}, z_a the elevation above the surface, and z_o the roughness parameter in cm.

The two terms in equation [1] give expressions for the portion of potential evapotranspiration primarily due to radiation and wind, respectively:

$$LE_r = \frac{\Delta/\gamma\, H}{\Delta/\gamma + 1} \text{ cal cm}^{-2} \text{ min}^{-1} \quad [3]$$

$$\text{and } LE_w = \frac{LB_v d_a}{\Delta/\gamma + 1} \text{ cal cm}^{-2} \text{ min}^{-1} \quad [4]$$

LE_r does not represent radiation component exactly. From equation [3] one can see that this radiation dominant term, LE_r, equals energy input multiplied by $(\Delta/\gamma)/(\Delta/\gamma + 1)$. This modifying fraction is temperature dependent and has values of 0.74, 0.78, and 0.82 at 25, 30, and 35C, respectively. Soil heat portion of energy input is usually small in comparison to net radiation especially if the soil is covered with vegetation.

The wind dominant term, LE_w, shows the contribution wind makes to total potential evapotranspiration for specified ambient air temperatures and water vapor pressure deficits. It is not identical to sensible heat term, A, of energy balance equation

$$LE = -(H + A) \text{ cal cm}^{-2} \text{ min}^{-1} \quad [5]$$

except when energy input (H) is zero. By equating equations [1] and [5] and solving for sensible heat we obtain:

$$A = \frac{LB_v d_a}{\Delta/\gamma + 1} - H\left(1 - \frac{\Delta/\gamma}{\Delta/\gamma + 1}\right) \text{ cal cm}^{-2} \text{ min}^{-1} \quad [6]$$

This shows sensible heat equivalent to wind dominant term of combination model less some fraction of energy input which is dependent upon temperature. At 25, 30, and 35C the fraction is 0.26, 0.22, and 0.18, respectively. Therefore for sensible heat to contribute positively to energy of evapotranspiration, the wind term should be roughly 1/5 to 1/4 as large as energy input. At night the energy input is often negative.

The van Bavel (18) version of the combination concept, equations [1] and [4], was used in this study for calculating potential evapotranspiration and the contribution of wind to the total potential evapotranspiration for various meteorological conditions.

EXPERIMENTAL

An observation site was established at Manhattan, Kans., on a 100- by 200-m field of clipped sudangrass. Sampling probes containing two copper-constantan thermocouples each were positioned at 5 and 45 cm. Each probe consisted of an outer tube with 3.8-cm (1½-inch) outside diameter and 30.5-cm (12-inch) length and an inner tube with 1.9-cm (¾-inch) outside diameter and 24.9-cm (9-inch) length. The outer tube was painted white to give high emissivity for longwave radiation and low absorption of solar shortwave radiation. The inside of the outer tube and the inside and outside of the inner tube were covered with aluminum foil for low emissivity for longwave radiation.

In each sampling probe one thermocouple, which was covered with a white cotton shoelace and connected to a water reservoir, was used for wet-bulb temperature measurements. Ambient air temperature measurements were obtained from the other thermocouple. To ventilate, air was sucked through the sampling probes over the sensors. Windspeed past the wet thermocouple was greater than 3 m sec^{-1}.

Sensitive cup anemometers were positioned on a mast at the same elevations as the temperature and humidity probes.

For evaluation of roughness length z_o, vertical wind profiles were measured periodically with anemometers spaced at 12, 26, 47, 79, 126, 197, and 303 cm above soil surface.

Soil heat flux was determined calorimetrically in the surface 10 cm and with heat flow transducers below 10 cm. To obtain heat capacity for calculating the heat storage term, the soil was sampled frequently and water content gravimetrically determined. The average soil temperature was measured within the top 10 cm at four locations with four vertically spaced temperature probes at each location forming a 16-junction parallel thermopile. The average soil temperature (0 to 10 cm) was referenced against the soil temperature 1 m below surface under the instrument trailer.

Net radiation was measured with Fritschen (6, 8) net radiometers and total or global solar radiation with an Eppley pyranometer.

The output from the various transducers was read and recorded at 15-min intervals with a data acquisition system similar to that described by Fritschen and van Bavel (10).

RESULTS AND DISCUSSION

Data were obtained through much of July and August 1967. Figures 1 and 2 show a comparison of daily variation of net radiation, potential evapotranspiration, portion of potential evapotranspiration due

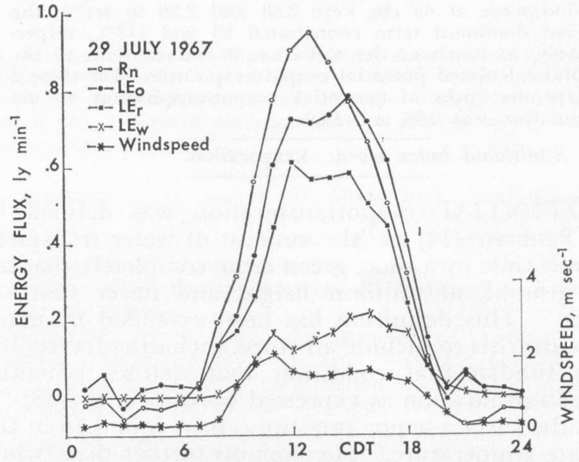

Fig. 1. Daily variation of net radiation, R_n; calculated potential evapotranspiration, LE_o; portions of calculated potential evapotranspiration due to radiation dominant term, LE_r and to wind dominant term, LE_w; and windspeed.

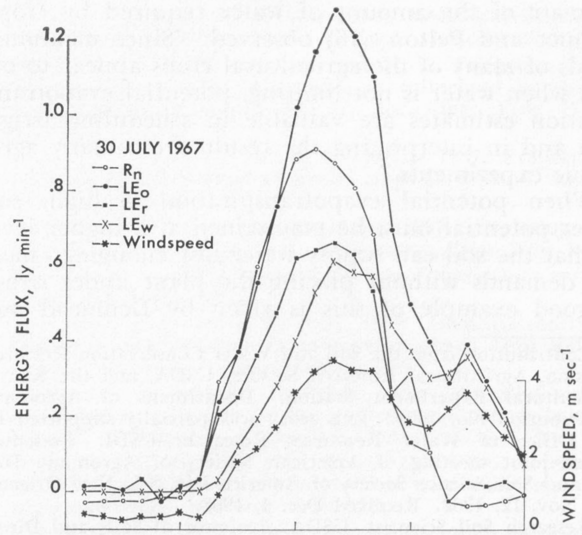

Fig. 2. Daily variation of net radiation, R_n; calculated potential evapotranspiration, LE_o; portions of calculated potential evapotranspiration due to radiation dominant term, LE_r and to wind dominant term, LE_w; and windspeed.

Table 1. Energy flux in cal cm⁻² day⁻¹ (R_n, net radiation; LE_o, calculated potential evapotranspiration; LE_r, portion of calculated potential evapotranspiration due to radiation; LE_w, portion of calculated potential evapotranspiration due to wind) and average daily windspeed $\overline{U}_a$ in m sec⁻¹.

Day	R_n	LE_o	LE_r	LE_w	$\overline{U}_a$
7-29-67	395	388	291	97	.88
7-30-67	401	643	302	341	2.26
8-11-67	321	336	235	101	.83
8-14-67	337	585	254	331	2.66

to radiation, portion of potential evapotranspiration due to wind, and windspeed on consecutive "nonwindy" and "windy" days (July 29 and 30) when the soil moisture tension was low.

Potential evapotranspiration on the 29th (Fig. 1) lagged net radiation and was slightly less. Air temperature lag behind radiation tends to cause a lagging of evapotranspiration when the wind dominant term is contributing significantly to total evapotranspiration.

On "windy" day, July 30, calculated potential evapotranspiration was of the same magnitude and appeared in phase with net radiation in forenoon. However, calculated potential evapotranspiration was much higher at midday and continued higher throughout the day. The ratio of calculated potential evapotranspiration to R_n was 0.98 and 1.60, respectively, for July 29 and 30. The corresponding average daily windspeeds at 45 cm were 0.88 and 2.26 m sec^{-1} (see Table 1).

On the 29th the wind dominant term contributed only one-third as much as the radiation dominant term to the total calculated potential evapotranspiration, whereas on the following day the wind dominant term contributed 13% more than the radiation dominant term to the total. Using the revised Penman version for the conditions of this study shows that wind contributes a much larger influence on evapotranspiration than is reported for conditions of northwestern Europe (14) or DeVries' and Van Duin's (4) interpretation of Akron, Colo., data.

Daily totals for energy flux and average windspeed for another "nonwindy-windy" pair of days in August are given in Table 1. The same pattern is apparent.

It would be desirable to compare the calculated potential to the actual evapotranspiration measured with accurate weighing lysimeters. Lysimeters were not available and we used the Bowen ratio method for estimating actual evapotranspiration. Fritschen (7) showed that Bowen ratio data agreed well with lysimeter data except as windspeed increased, then computed values tended to underestimate measured values.

Note the agreement between calculated potential evapotranspiration and evapotranspiration computed by Bowen ratio method as shown in Fig. 3 for July 29. The potential exceeds Bowen ratio evapotranspiration slightly. The condition for actual evapotranspiration to proceed at the potential rate was not fully met. The soil was not completely shaded by the plants, and the vapor pressure of some of the evaporating surfaces would have been somewhat less than the vapor pressure of water at the same temperature.

On the following "windy" day (Fig. 4), Bowen ratio evapotranspiration was considerably higher than the previous "nonwindy" day and yet fell far below the calculated potential evapotranspiration. Evapotranspiration (Bowen ratio determination) was able to keep up with potential reasonably well early in the day. It appeared that the soil and crop could not supply water at a rate greater than about 0.75 mm/hour even though the demand went to almost 1.3 mm/hour.

Given the various ambient environmental conditions of temperature, water vapor pressure, windspeed at z_a, and roughness length, one can compute the contribution of wind to potential evapotranspiration. This was done for various windspeeds and temperatures. The results for ambient water vapor pressures of 10 and 20 mb are shown in Fig. 5 and 6, respectively.

Summer temperatures and windspeeds during the growing season are generally within the limits of Fig. 5 and 6, 15 to 40C and 1 to 6 m sec^{-1}, respectively.

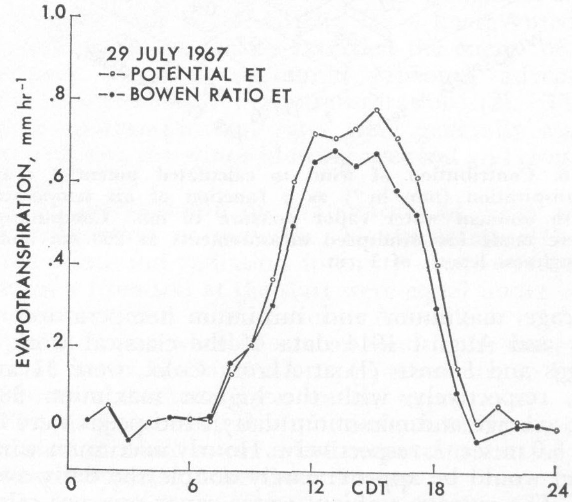

Fig. 3. A comparison of calculated potential evapotranspiration versus evapotranspiration computed by Bowen ratio method.

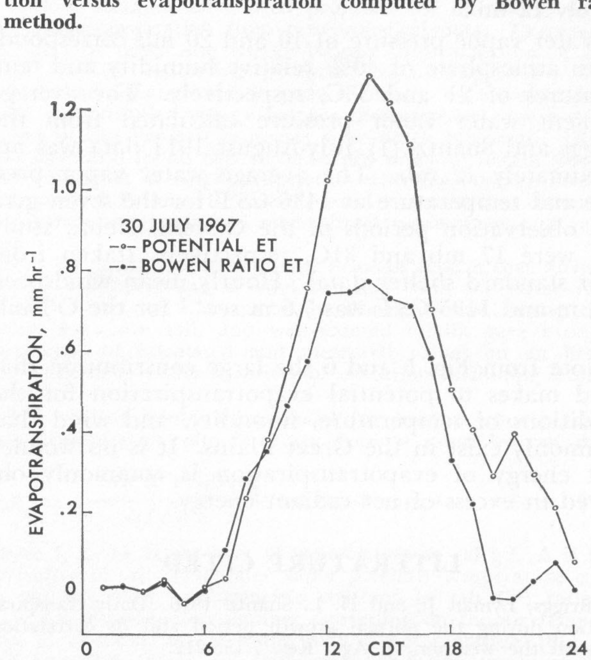

Fig. 4. A comparison of calculated potential evapotranspiration versus evapotranspiration computed by Bowen ratio method.

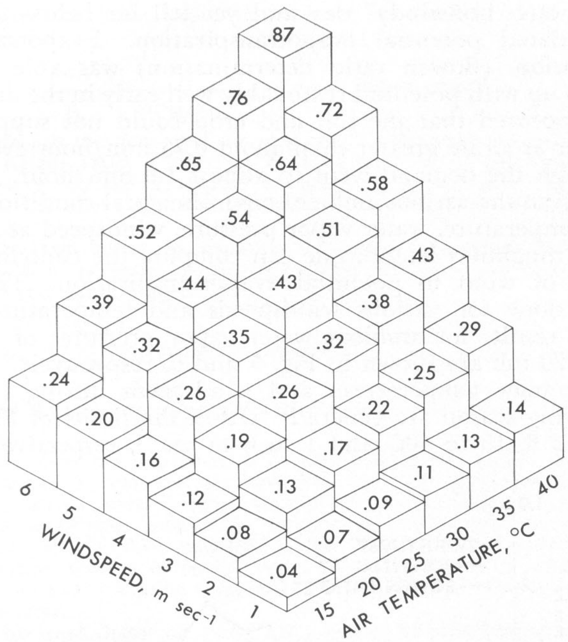

Fig. 5. Contribution of wind to calculated potential evapotranspiration (mm hr^{-1}) as a function of air temperature with ambient water vapor pressure 10 mb. Computations were made for windspeed measurements at 200 cm and a roughness length of 1 cm.

Average maximum and minimum temperatures of July and August 1914 data of the classical work of Briggs and Shantz (1) at Akron, Colo., were 31 and 14C, respectively, with the highest maximum 38C. The average and maximum daily windspeeds were 2.8 and 5.0 m sec^{-1}, respectively. Hourly maximum windspeed would be approximately double the daily average. The average ambient water vapor pressure calculated from the Briggs and Shantz data was approximately 12 mb.

Water vapor pressure of 10 and 20 mb corresponds to an atmosphere of 40% relative humidity and temperatures of 21 and 33C, respectively. The average ambient water vapor pressure calculated from the Briggs and Shantz (1) July-August 1914 data was approximately 12 mb. The average water vapor pressure and temperature at 1430 CST for the seven general observation periods of the O'Neill, Nebr., study (13) were 17 mb and 31C, respectively (taken from their standard shelter data). Hourly mean windspeed at 2 m and 1435 CST was 7.6 m sec^{-1} for the O'Neill site.

Note from Fig. 5 and 6 the large contribution that wind makes to potential evapotranspiration for the conditions of temperature, humidity, and wind that commonly exist in the Great Plains. It is no wonder that energy of evapotranspiration is commonly observed in excess of net radiant energy.

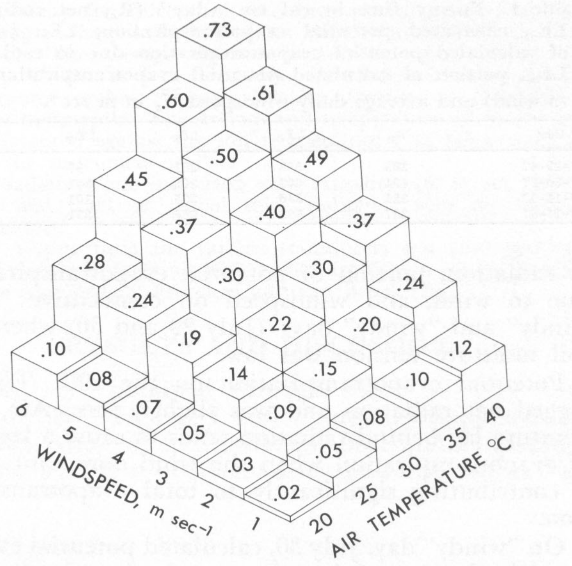

Fig. 6 Contribution of wind to calculated potential evapotranspiration (mm hr^{-1}) as a function of air temperature with ambient water vapor pressure 20 mb. Computations were made for windspeed measurements at 200 cm and a roughness length of 1 cm.

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