

FACTORS AFFECTING POTENTIAL EVAPOTRANSPIRATION

by

W. O. Pruitt*

As the title indicates, in this paper we will be considering basic factors which affect potential evapotranspiration. The radiation balance will be dealt with very briefly, with considerable attention then being given to the energy balance and the affect of various climatic factors on the partitioning of net radiation into latent and sensible heat fluxes.

THE RADIATION BALANCE

The primary source of energy for almost all of the natural phenomena occurring on the earth is the sun. In particular, the largest direct source for energy required in evaporation of water is radiant energy coming either directly or indirectly from the sun. Of primary interest, then, in considering the process of evaporation (or evapotranspiration), is the radiation balance.

Looking at the relevant radiation sources and sinks with respect to the earth's surface we can write

$$(Q + q)(1 - a) + I^+ - I^- = R_n \quad (1)$$

or $(S + \Delta)(1 - a) + R_a - R_g = R_n$

where $S = Q$ = direct solar radiation received on a horizontal plane at the earth's surface

$\Delta = q$ = diffuse solar radiation incident on the earth's surface

$a = \alpha$ = the surface reflectance or albedo expressed as a fraction

$R_a = I^+$ = downcoming longwave radiation

$R_g = I^-$ = outgoing longwave radiation from the surface

R_n = the total net radiation exchange

The radiation balance is affected most dramatically by the reflectance characteristics of the surfaces, e.g. ocean or lake surfaces reflecting only 5-10% of the incoming solar radiation while fresh snow surfaces may reflect up to 90%. It is thus obvious that with a given amount of solar radiation the net radiation of different surfaces may be drastically different. For surfaces covered with vegetation, however, this variation is much less. Albedo values reported by Gates and Hanks (1) for cropped surfaces range between 14 to 29% with most of the crops averaging around 20%. Variation in evapotranspiration due specifically to albedo differences are discussed in a recent article by Segner (2). Using a computer

*Lecturer and Irrigation Engineer, Department of Water Science and Engineering, University of California, Davis, California.

model to test surfaces with albedos of 0.25 and 0.40, he found for clear sky conditions an approximate 30% reduction in R_n with the higher albedo with a similar percentage reduction in evapotranspiration.

In addition to the loss of solar radiation by reflectance, the outgoing longwave radiation normally exceeds the downcoming atmospheric radiation. This net exchange of longwave radiation is significantly affected by surface temperature, cloudiness, and the amount of water vapor, CO_2 , etc., in the air. Typical clear-weather values of net longwave radiation noted for a grass surface in mid-summer at Davis, California, show a range of from around -0.07 ly/min at night to -0.20 ly/min at midday (3). For surfaces drier and much hotter than grass, the midday net longwave radiation could be significantly more negative due to higher losses to the cold sky.

Although the net radiation of cropped surfaces is very much a function of incident solar radiation, the examples in Fig. 1 for studies over grass or pasture provide ample evidence of a wide range of response under different climatic conditions. The only location where $R_n/(Q + q)$ remains fairly

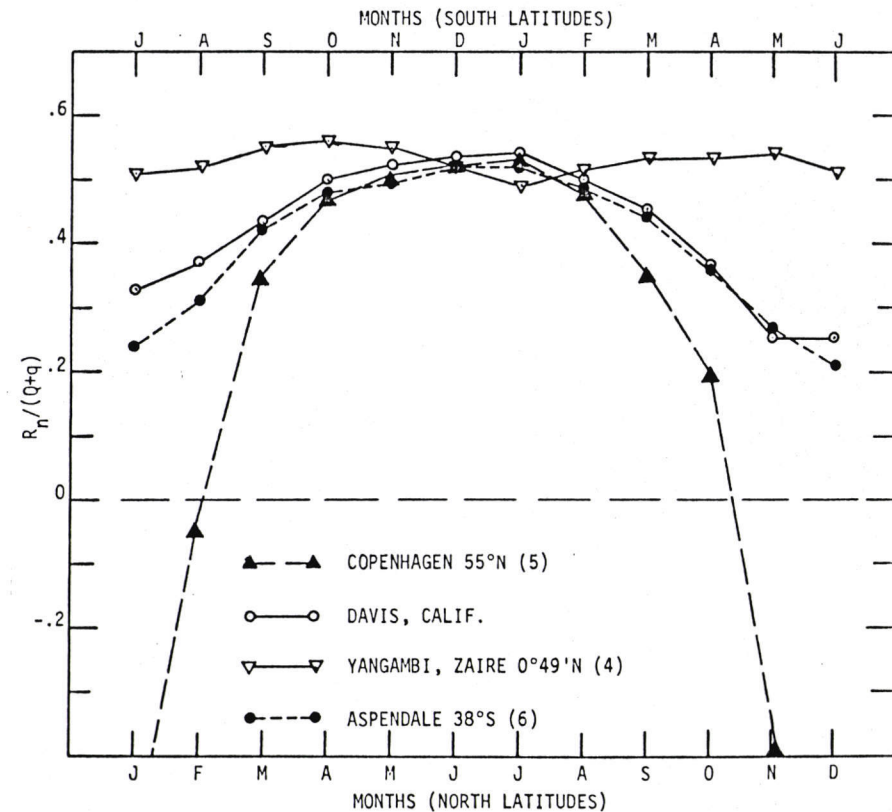


Fig. 1. The Fraction of Incident Solar Radiation, $(Q+q)$ Represented by Total Net Exchange Radiation, R_n for Grass or Pasture Surfaces at Four Locations.

steady throughout the year is at the tropical site (4). At 55°N latitude monthly means of net radiation for a grass surface (snow-covered at times) are actually negative during winter months (5). Patterns are very similar for Australian (6) and Davis, California studies. With the wide variation in $R_n/(Q + q)$ with seasons except in the tropics, it is no wonder that considerable empirical adjustment is needed in evaporation prediction formulas based largely on solar radiation.

THE ENERGY BALANCE

Based on the principles of the conservation of energy, if we consider a soil or water column extending from the surface to some depth where vertical heat exchange is negligible, we can write

$$R_n + H + LE + G + F_i - F_o = 0 \quad (2)$$

where R_n = net exchange radiation (all wave lengths)

H = sensible heat transfer, negative when transfer is a loss from surface to the air, positive if transfer is from air to surface

LE = latent heat flux due to evaporation (-) or condensation (+), where L is the latent heat of vaporization (about 590 cal g⁻¹)

G = net rate at which the heat content of the soil or water column is changing

F_i = rate of horizontal transfer of heat into the column

F_o = rate of horizontal heat transfer out of the column

The last two terms are important only for oceans and other water bodies where currents can transport considerable sensible heat. Over land, ΔF would effectively be zero leaving

$$R_n + G + H + LE = 0 \quad (3)$$

The energy-balance equation as given in (2) or (3) applies over any time period from minutes to years. It is an approximate equation, however, since some components (normally quite insignificant) are neglected. Sellers (7) gives estimates of the magnitude of other items such as snow melt, sources or sinks of heat from precipitation, photosynthesis, respiration, etc. For vegetative surfaces the most significant is likely to be photosynthesis.

Seasonal Patterns of the Energy Balance

The partition of the net radiant energy at the earth's surface is very strongly influenced by the availability of water. Many micro-meteorological studies have been conducted where dry-surface conditions severely restricted the conversion of net radiation into latent heat with most of the energy going into sensible heat transfer into the air and into the soil. Surface temperatures exceed air temperatures during

almost the entire daylight period under such conditions. With air near the surface hotter and thus less dense than above, unstable conditions prevail in the lower few meters of the atmosphere during the entire day. If a source of moisture at the surface is nonexistent and heat flux into the ground (or artificial surface) can be measured, the convective heat flux into the air is easily deduced.

One might think that in an analogous way, if water were completely available (i.e. a very thin film of water on the surface or a complete canopy of vegetation with insignificant stomatal resistance), there would be very little convective heat transfer involved due to a near 100 percent depletion of net radiation by evaporation and soil heat flux. Surface temperature should remain at or near air temperature and near-neutral stability conditions might be expected during much of the daytime period in contrast to the highly unstable conditions noted earlier. This concept concerning the partition of net radiation for vegetative surfaces is quite prevalent in the literature. Perhaps it evolved from the concept of potential evapotranspiration first presented by C. W. Thornthwaite (8). The expression, potential evapotranspiration, was redefined by Penman (9) 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." There is nothing in this definition which says anything about the relationship of latent heat flux to net radiation. It is obvious, however, from later literature that researchers came to think of potential evapotranspiration as a condition under which essentially all of the net radiation (minus soil heat flux) would be used up in the evaporative process, e.g. Lemon et al. (10), Pelton et al. (11). Although the effect of advection in producing ratios of latent heat, LE , to net radiation, R_n , well over 1.0 has long been recognized, it seems that ratios well under 1.0 were only expected for cases where very significant stomatal resistances were likely or where plant cover was only partially shading the ground, or just upwind from cold ocean or lake surfaces.

The concept of potential evapotranspiration has served a useful purpose in the past and will, I believe, continue to do so. However, it may need re-defining as McIlroy suggests (12). Certainly some of the common ideas about it that have crept into the literature need to be corrected. Perhaps the most serious misconception is that whenever potential ET is realized, it should be about equal to net radiation, even on a short-term basis.

For purposes of illustration we turn our attention now to some evapotranspiration studies where for the most part, full cover, vegetative surfaces with low stomatal resistance characteristics were involved. The climates due to latitude and other factors represent a wide range of conditions. Figure 2 presents the mean of 3½ years of data for a well-irrigated, 10-20 cm tall perennial ryegrass cover at Davis, California. Negative LE , G and H are plotted for easier comparison with trends of R_n . During spring months significant energy is going into heating the air above the crop, hence LE is only 70-80% of R_n . For mid-summer and through September the mean daily LE corresponds closely with R_n , with LE exceeding R_n thereafter until January. Soil heat flux G , is rather insignificant although later studies (13) suggest the 1960-63 measurements may have underestimated G by 100-200%. Regardless, the pattern shows heat losses to the soil from mid-March until mid-September, with heat gained from the soil during the other months.

See Monteith (1985) in attached paper

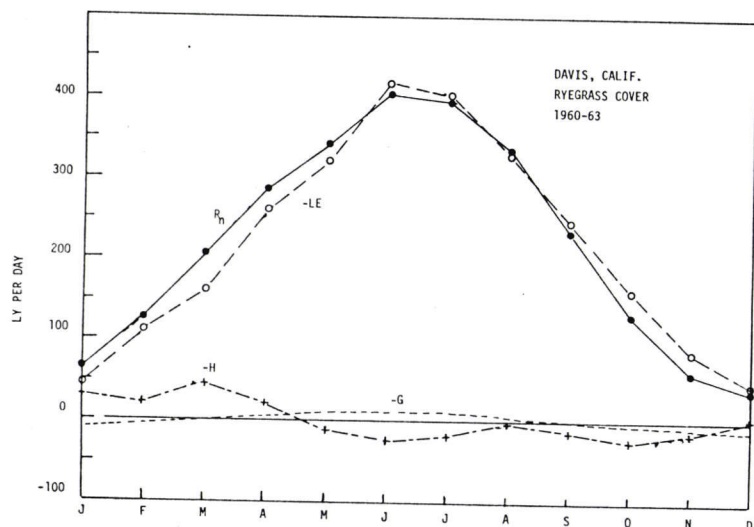


Fig. 2. Energy Balance for a Well-Watered, Perennial Ryegrass Cover at Davis, California. (R_n , LE, and G were measured with H computed from the energy balance.)

Results published by Aslyng and Jensen (14) for a 10-year study in Denmark of a short clover-grass mixture, are reproduced in Fig. 3. Not only is net radiation much lower than at Davis, but the fraction of R_n

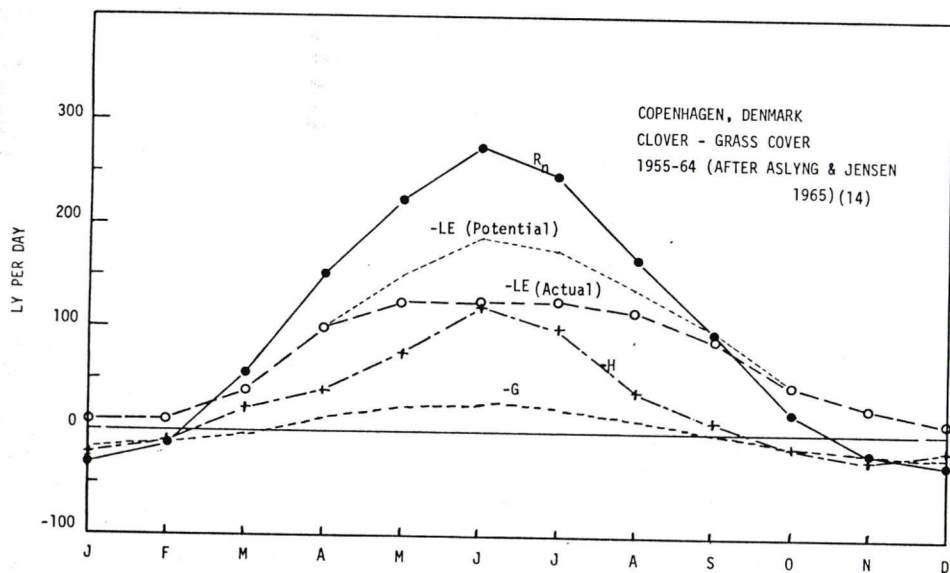


Fig. 3. Energy Balance for 10 Years for a Clover-Grass Mixture at Copenhagen, Denmark.

used as LE remains well under 1.0 during most of the year. At this high latitude, R_n goes negative for the four winter months, the energy for LE coming largely from the air mass moving in from relatively warmer oceans. During the summer months almost one-half of the R_n goes into heating up the cold air mass moving off of the relatively colder ocean surfaces.

A plot in Fig. 4 of the energy balance data for alfalfa at Tempe, Arizona, given by van Bavel (15), provides evidence of an opposite extreme to that at Copenhagen. The value of R_n is not only much greater than at

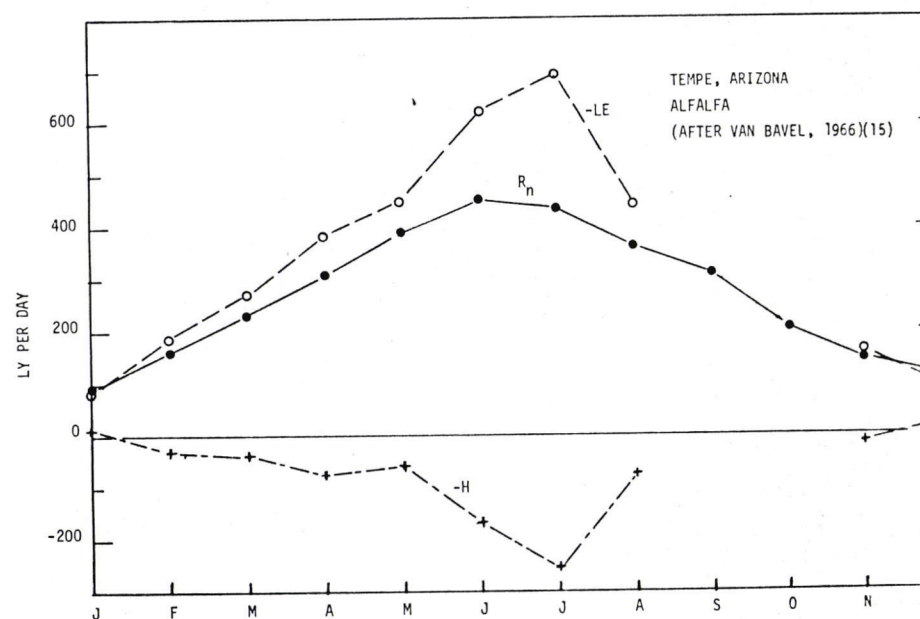


Fig. 4. Energy Balance for Alfalfa Maintained at Near Full-Cover Condition Throughout Most of the Year at Tempe, Arizona. (R_n and LE measured with H computed from the energy balance assuming G to be insignificant.)

Copenhagen, but with the exception of January and December, LE exceeds R_n every month of the year. This is indicative of consistent strong advection of heat from nearby desert areas at Tempe. Only during December and January did the crop surface serve as a net supplier of heat to the air passing over the field on a 24-hour basis.

In Fig. 5, the ratio of LE/R_n for all three locations is shown. The upper curve for Davis is based on data from Fig. 2. The lower Davis curve represents results when the data from approximately 100 strong-advection days are excluded. This results in considerably lower LE/R_n

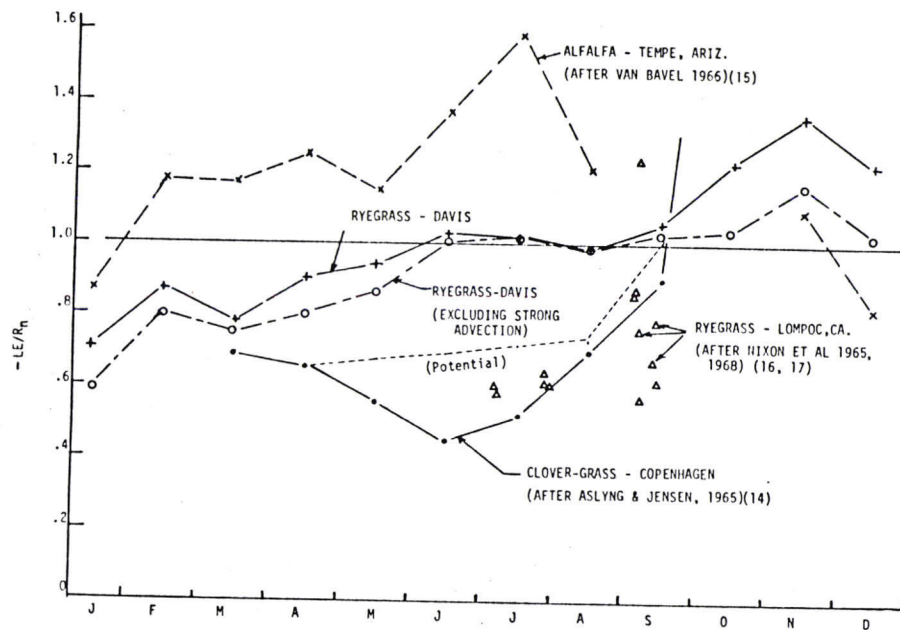


Fig. 5. The Fraction of Net Radiation, R_n Used as Latent Heat, LE for Several Crop Covers and at Several Locations. (Lower curve for Davis excludes all strong advection days.)

ratios for Winter and Spring months, but values very close to 1.0 from June through December (excepting November). Data for a few days from a study by Nixon (16) (17) near Lompoc, California are plotted for comparison. The results are for perennial ryegrass grown about 10 miles inland from the Pacific Ocean. With the exception of one heavy overcast day, the Lompoc data are in close agreement with the data from Denmark with around 40 percent of the net radiation being used in heating up the air mass moving in from the cold ocean. With low radiation on cloudy days, the air off the ocean apparently serves as a source of advected heat resulting in $-LE/R_n$ ratios > 1.0 . Further inland, e.g. at Davis, this does not appear to be the case, and $-LE/R_n$ normally is < 1.0 on cloudy days.

In all of the prior examples we were dealing with mean monthly data which included both day and nighttime hours. Since R_n is normally negative at night, ratios of $-LE/R_n$ for 24-hour periods are higher than for daytime periods only. Thus, if we look at the ratios in Fig. 6 based on daytime data only, we see significantly lower values than those indicated for Davis in Fig. 5. Additionally, the pattern is quite different in shape, with peak ratios occurring in June and July, and quite low ratios in winter months (0.4-0.5). Except for mid-summer, a considerable portion of the daytime net radiation is going into heating the air with, of course, some into heating the soil.

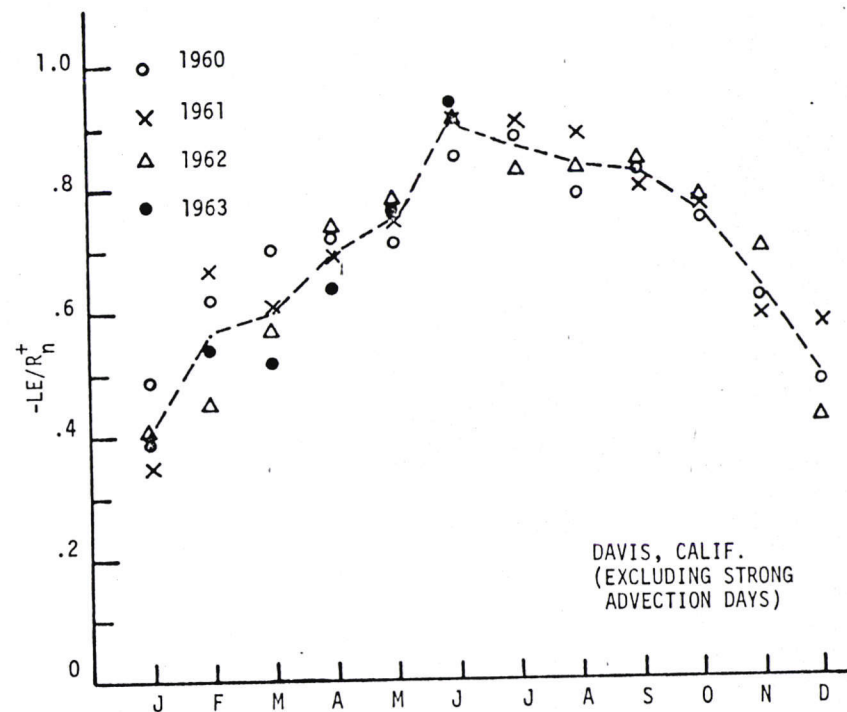


Fig. 6. Ratio of Monthly Mean Latent Heat, LE , to Monthly Mean Daytime Net Radiation, R_n^+ (positive only), at Davis, California.

Diurnal Patterns of the Energy Balance

Even wider variation of $-LE/R_n$ ratios occur diurnally. Some Davis results of Fig. 7 show data for typical clear days during spring and summer with various crops involved. The lower three curves are $-LE/(R_n + G)$ ratios for three calm to light-wind days. The upper curve in Fig. 7 is for the strong, dry-wind day of October 13, 1966 with rather severe advection.

fluid motion

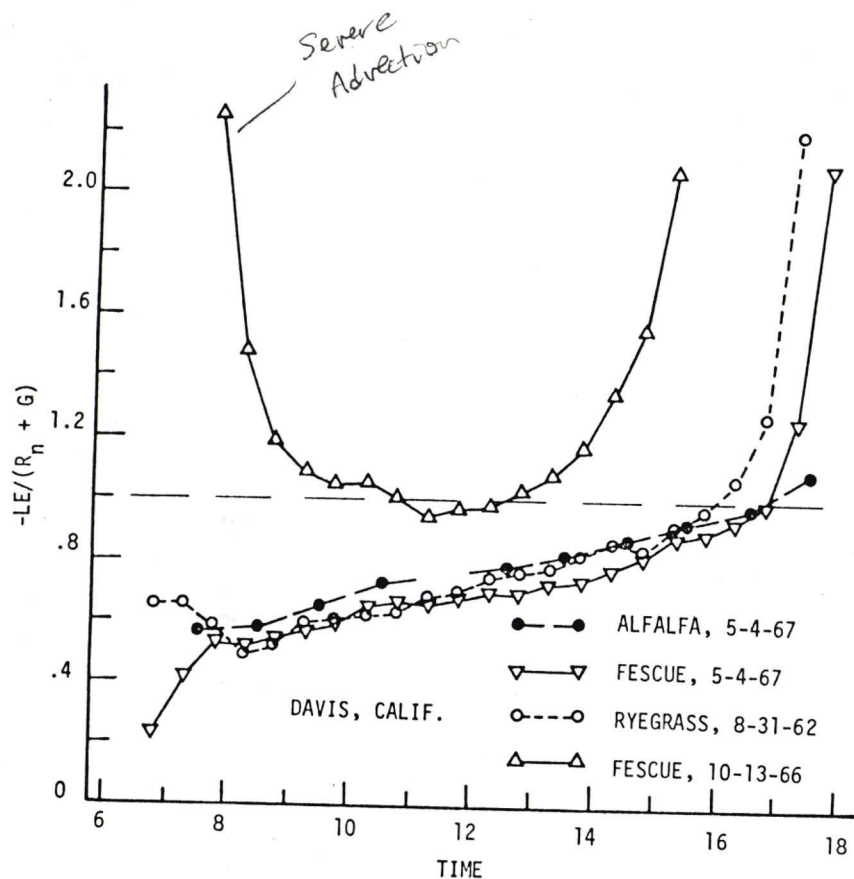


Fig. 7. Variation of the Fraction of $(R_n + G)$ Used as Latent Heat Flux, LE , for Two Crops, on 4 May, 1967. (Also shown are data for perennial ryegrass from an earlier study. Clear skies with light to mild-breeze conditions persisted for all three days. 13 October 1966 was a dry, strong-wind day with clear skies.)

Although actual energy balance data are plotted in Fig. 8, these results reproduced from Sellers (7) reveal a wide variation in the relationship of LE and R_n . The data for Tempe, Arizona, show results similar with regard to $-LE/R_n$ ratios, to that for the day of October 13, 1966 in Fig. 7. By contrast, the data for Wisconsin indicate almost as much heat going into the air as going into latent heat, LE , although this is likely due in part to partial cover or low soil moisture since Tanner and Pelton (18) indicate LE values most of the time close to R_n . The lower one-third of the figure gives a typical pattern of the various energy-balance components for a very dry surface condition on a clear day.

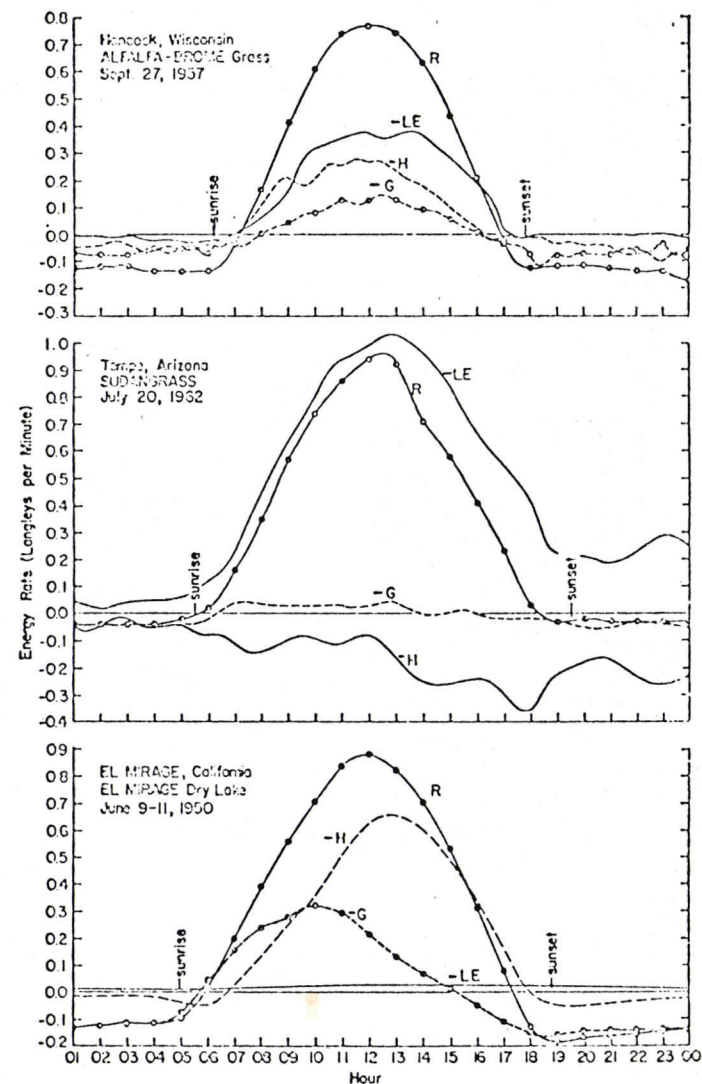


Fig. 8. Diurnal Patterns of the Components of the Surface Energy Balance at Two Cropped Sites and One Desert Site. (After Sellers (7).)

Effect of Air Mass and Surface Conditions

From the preceding figures it is obvious that the partitioning of net radiation into latent and sensible heat fluxes is highly dependent upon prevailing air mass conditions. The magnitude of LE can be much less than that of R_n or it can be considerably larger. In general, it is easy to comprehend why considerably greater ratios of $-LE/R_n$ are experienced at Davis with a given amount of radiation as compared to a coastal location such as Lompoc. Looking at the effect of air temperature alone, it is obvious that the vegetative temperatures (with a given radiation) will be able to reach a much warmer temperature at Davis with 95° air passing over it as compared to coastal vegetation with 65° air coming off the cold ocean. Since the vapor pressure of a surface is a function of its temperature, the vapor gradient from surface to air is going to be considerably greater at Davis than at Lompoc.

To come to a basic understanding of the magnitude of the effect of various climatic factors on evapotranspiration, is considerably more difficult. We would like simple answers to questions such as "what is the effect of wind on evapotranspiration, or of radiation, or of humidity?" Answers in the literature seem to conflict. For example, various studies have suggested that an increase in wind speed above 2 to 3 mph causes no further increase in ET. Other reports, however, suggest considerable influence of wind on ET well beyond this wind speed. The literature of a decade or more back contains many references indicating that for well-watered crops, a near equality of LE and R_n should be expected. Figures 2-8, however, point to a wide range of the ratio of $-LE/R_n$ for such surfaces. Some researchers suggest that most crops with full cover and adequate soil moisture, use about the same amount of water. Others have reported fairly wide differences between crops, supposedly due to differences in the aerodynamic roughness characteristics of crops.

The following discussion should help clarify the reason for some of the above seeming conflicts of opinion yet, at the same time, help the reader to realize why there are no easy answers to the earlier posed questions.

There are two requirements for evaporation which can be stated as being clearly necessary: (1) There must be a source of energy or heat and (2) Diffusion or flow of water vapor away from the surface must be maintained by some mechanism. We can use the energy balance alone in predicting ET by measuring all the other significant terms, or we can predict ET from the use of strictly diffusion or aerodynamic equations. We must realize, however, that (1) and (2) above must be met and that they are dependent upon each other.

In an effort to arrive at a better understanding of how the various climatic parameters affect the partition of energy at a vegetative surface, an evaporation prediction equation, proposed by McIlroy (19) is quite helpful. This method essentially is based on an approach for partitioning R_n , proposed by Bowen (20), where the ratio of convective to latent heat flux H/LE is determined from gradients of air temperature and humidity. In combination with the energy-balance equation of $R_n + G + LE + H = 0$, both LE and H can thus be determined if R_n and G are measured.

The ratio of H/LE is commonly referred to as the Bowen ratio. In recent years, standard procedures for use of the Bowen ratio have involved the partition of $(R_n + G)$ into LE and convective heat flux, H, using the following:

$$H = \rho c_p K_H \frac{\Delta T}{\Delta z} \quad (4)$$

and

$$LE = L \rho K_W \frac{\Delta q}{\Delta z} \quad (5)$$

where ρ is air density, c_p is the specific heat of dry air at constant pressure, K_H and K_W are the eddy transfer coefficients for heat and water vapor, respectively, ΔT is the difference in air temperature between two heights above ground with a difference in elevation of Δz , and Δq is the difference in specific humidity as measured at the same two elevations as for temperature. Dividing (4) by (5) and assuming equality of K_H and K_W , a ratio of convective to latent heat flux, β , is derived giving:

$$\beta = \frac{H}{LE} = \frac{c_p}{L} \frac{\Delta T}{\Delta q} \quad (6)$$

From (6) and the energy balance equation, $R_n + G + LE + H = 0$, the following expression for LE results:

$$LE = - (R_n + G) \frac{1}{1 + \beta} \quad (7)$$

or

$$LE = \frac{-(R_n + G)}{1 + \frac{c_p}{L} \frac{\Delta T}{\Delta q}} \quad (8)$$

McIlroy (19), (12) derived an expression for the Bowen ratio approach in terms of wet-bulb and dry-bulb measurements which with rearrangement to fit sign conventions used herein, gives:

$$LE = - (R_n + G) \left(1 - \frac{\gamma}{s + \gamma} \frac{\Delta T}{\Delta T_w} \right) \quad (9)$$

where ΔT and ΔT_w are measured differences of dry-bulb and wet-bulb temperature, respectively, over the same interval of height, $\frac{\gamma}{s + \gamma}$ is a temperature dependent weighting function where, in consistent units, γ is the ratio of specific heat of air to the latent heat of water and s is the slope of the saturation specific humidity versus temperature curve, at mean wet-bulb temperature.

With respect to equation (9), McIlroy made some very interesting observations and introduced a new term called equilibrium evaporation. Taking the case where over a very extensive uniform moist surface the air close to the surface is saturated (or nearly so), it is obvious that the

Monteith picks up on the significance and value of the McIlroy equation and provides much additional insight (See pp 249).

wet-bulb temperature profile would coincide (or nearly so) with the dry-bulb profile. Consequently, $\frac{\Delta T}{\Delta T_w} \rightarrow 1.0$ and for such conditions LE can be approximated by

$$LE = - (R_n + G) \left(1 - \frac{y}{s+y}\right) = - (R_n + G) \frac{s}{s+y} \quad (10)$$

It is this condition McIlroy describes as an equilibrium evaporation case since the air and the ground would be mutually adjusted to each other in respect to their wetness, at least from the standpoint of relative humidity. Even if the air just above a surface were essentially saturated, a strong gradient upwards of specific humidity would still exist due to radiation heating of the surface. Hence, a significant vapor flux would still result.

At this point, it is interesting to bring out the variation of the weighting function $\frac{s}{s+y}$ as a function of the mean of the surface temperature (assumes effectively moist surface) and of the wet-bulb temperature of the air at some height above the surface. The value of $\frac{s}{s+y}$ ranges, for example, from 0.4 for a mean temperature of 32°F to a value of around 0.8 at 86°F. This indicates that during very humid periods in hot tropical areas, the ratio of $LE/(R_n + G)$ might range from 0.8 to 0.9, whereas for a similarly humid condition in a very cool climate (such as Lompoc, California, or Denmark) the ratio of $LE/(R_n + G)$ should be expected to be as low as 0.4 to 0.6. This, of course, also explains the low daytime ratios of 0.4 to 0.5 during January and February days with high humidity at Davis.

Where air above the ground is not already saturated, it will be taking up moisture in addition to that indicated by (10) and at the same time, of course, be taking up less heat by convection, or under more severe conditions, the air may give up heat to the surface to help meet the latent heat requirement. For a generalized expression, McIlroy (19), (12) derived the following simple combination equation giving:

$$-LE = \frac{s}{s+y} (R_n + G) + h(D - D_o) \quad (11)$$

where D is the wet-bulb depression at some height above the surface in °C or in °F if desired, D_o is the same for the surface, and h is a wind speed dependent heat-transfer coefficient which of course, would also vary with crop roughness and stability conditions of the air mass above. For effectively moist surfaces D_o would drop out and the equation would be a predictor of potential ET.

Although assumptions leading to (11) need further verification, we will now use this equation to help illustrate possible relative effects of temperature, wind, and humidity on the partition of $(R_n + G)$ under varying levels of radiation. The radiation, temperature, and humidity data given in Fig. 9 for the three times of the day were picked from a typical clear, fairly calm day in mid-July in Davis. Predicted LE values, using (11) for calm conditions (near left side of figure), are quite typical of measured values for such days. The indicated split between $h(D)$ and $\frac{s}{s+y} (R_n + G)$ is dependent, of course, upon assumptions

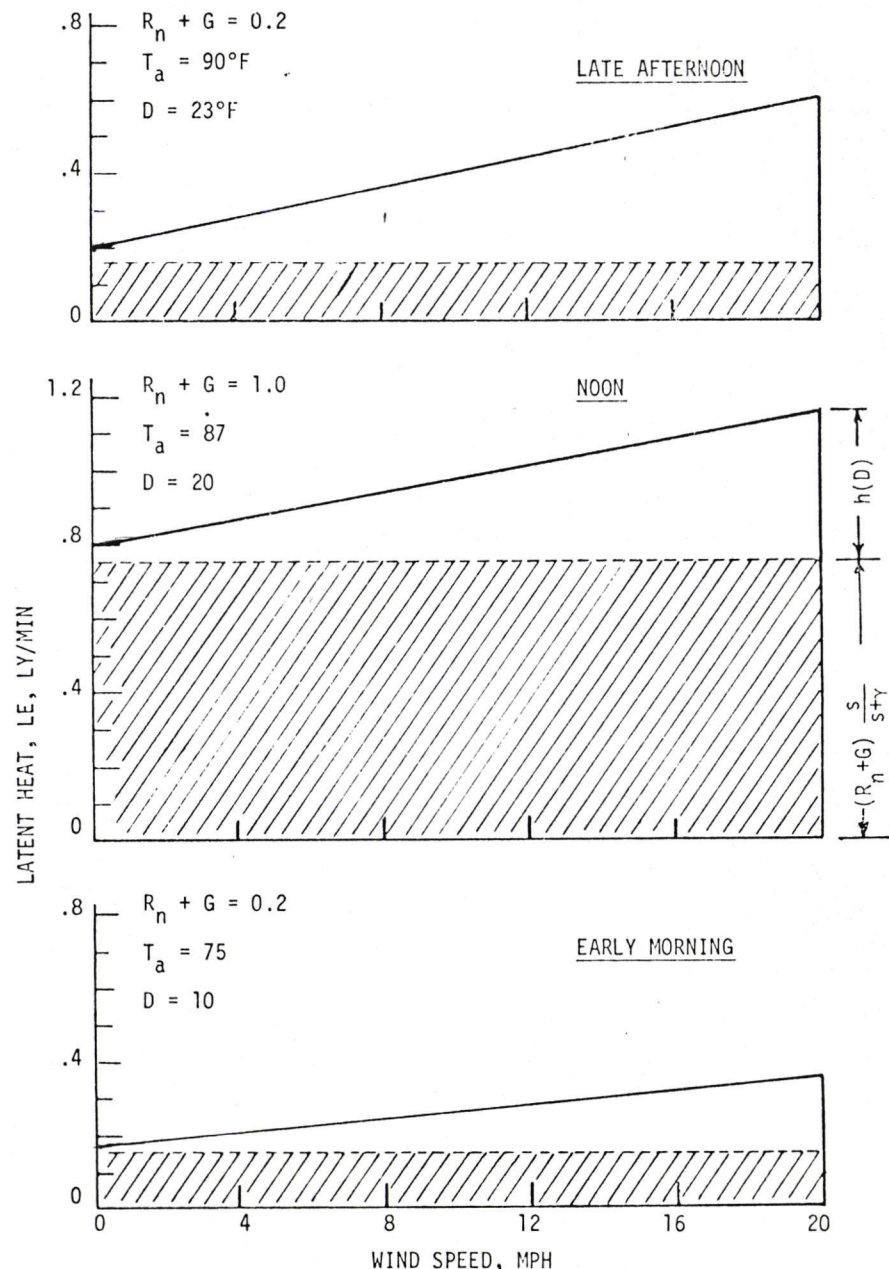


Fig. 9. Estimates of the Effect of Wind on LE at Three Times of the Day, Using Two Portions of the McIlroy Equation and Holding Air Temperature, T_a and Wet-Bulb Depression, D Constant.

FAO PENMAN: $ET_o = W \cdot R_n + (1-W) \cdot f(u) \cdot (e_a - e_d)$

where ET_o is in mm/day,

R_n is in equivalent mm/day (Ly/Day ÷ 59),

W is equivalent to $\frac{s}{s+y}$ above

e_d = actual vapor pressure of air, mb
 e_a = saturated vapor pressure, mb at average

inherent in (11). Three times of the day are examined, a midday period with $(R_n + G) = 1.0$ ly per min., and an early morning and late afternoon period when $(R_n + G) = 0.2$ ly/min.

The lower portion of each section represents the contribution of the first major term of the McIlroy equation to evapotranspiration. The upper portion represents the contribution from the second term of the equation. The expression for h used was $h = 0.002 (1 + u)$ with the wind speed, u , in meters per second. This expression was derived by Pruitt and Lourence (21) using lysimeter and meteorological data from a number of days in 1962 and 1963. Equation (11) with this expression for h , and with D in $^{\circ}\text{F}$ and D_0 assumed negligible, has been tested many times at Davis during 1965 through 1969. Predicted values were seldom greater than $\pm 10\%$ of half-hour or hourly lysimeter measurements.

A word of caution on interpretation of Fig. 9 is needed here. We cannot presume, for example, that on a given day if the wind should quickly change from 2 mph to 20 mph that the change of evapotranspiration with wind speed would follow the patterns indicated. It would not, for the values of T_a and D would also be affected. The patterns in Fig. 9, however, do indicate differences in LE that might be expected to occur on two different days when T_a and D were the same, but with different wind speeds. Thus, the pattern gives an estimate of the effect of wind with the other factors held constant.

As can be seen, for an early morning period LE would be about doubled by a 20 mph wind as compared to a calm condition. With the same radiation but with T_a and D much higher in late afternoon, the effect of the same wind difference would tend to produce three times as much LE. Under the higher noontime radiation condition, however, LE at 20 mph would be only about 45% greater than under the calm condition.

Now if the patterns shown in Fig. 9 are reasonably close to reality, we see that there is no simple answer to the question of what the effect of wind is on evapotranspiration. So much depends upon other climatic factors. The effect of humidity, for example, is dramatically depicted in Fig. 10, where the solid line is a repeat of results for Davis shown in the previous figure. For comparison, predicted LE as a function of wind speed assuming a higher air temperature and larger wet-bulb depression (lower RH) are shown above the data for Davis. For lower wind conditions the predictions would still be representative of most clear, hot days in Davis. A 20 mph wind with 99° maximum air temperature would be a very rare event, however (e.g. only 1 day in last 12 years). In the curves below the Davis data, results using weather conditions which might occur on clear days at Lompoc, California, are shown.

Now, still thinking in terms of the effect of wind on LE we see that under the more humid condition, and with high radiation, 20 mph winds might produce only 20-25% more LE than would be realized under a calm condition; or looking at the upper 1/3 of the figure with lower radiation, a 20 mph wind at a D of 11° would produce twice the LE expected under calm conditions. Under the hotter, drier conditions, however, ($D=32^{\circ}$) the effect of a 20 mph wind would be to produce $3\frac{1}{2}$ times the LE that would be realized under calm conditions. Again we see that the effect

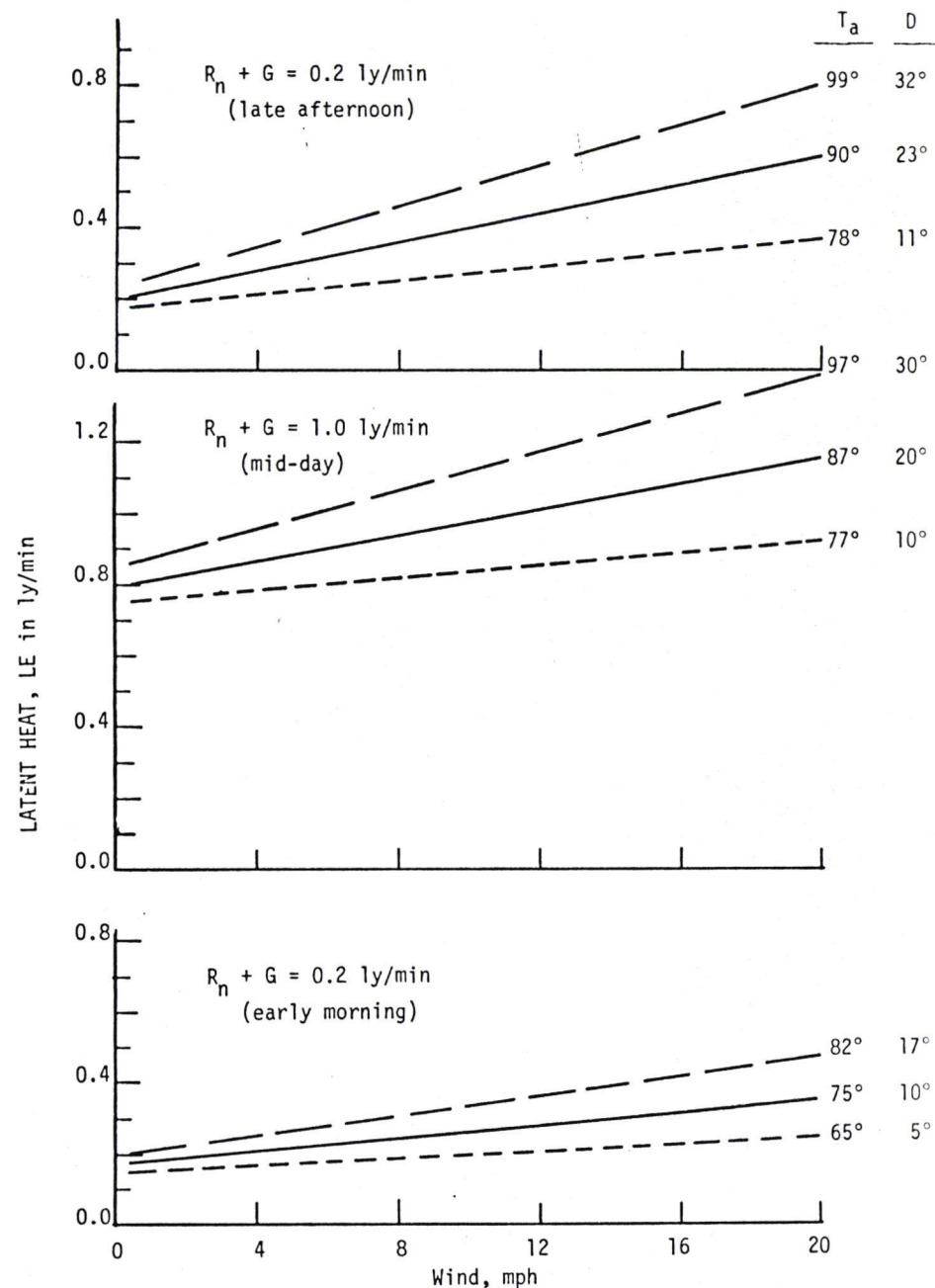


Fig. 10. Predictions by the McIlroy Equation of Latent Heat Flux, LE at Three Times of the Day as a Function of Wind Speed with T_a and D Held Constant. (— Davis data given in previous figure; — hotter, drier conditions; and - - - cool, humid conditions, perhaps as at Lompoc, Calif.)

of wind is very much dependent upon the humidity of the air. Considering the influence of humidity per se, it is obvious that percentagewise the effects of different levels of humidity are much less under high radiation than low radiation and under low wind as compared to high winds.

The effect of temperature is another important consideration. Several of the well-known and often relied-upon empirical equations are mainly dependent upon a high correlation of ET with temperature. There may not be a direct relationship between the two, however. A number of writers, e.g. van Wijk and de Vries (22) and Pelton, King and Tanner (11), have been quite critical of such methods, insisting that the seasonal out-of-phase relationship between air temperature and net radiation would produce an out-of-phase relationship between ET and temperature. Such comments came largely before the availability of $-LE/R_n$ data such as those presented herein which show a rather appreciable out-of-phase relationship normally exists between LE and R_n , as well. The McIlroy equation can help our understanding at this point as well. The term $\frac{S}{S+T}$ is temperature sensitive as discussed earlier. Hence, even in very humid areas, at least away from the tropics, the first half of the equation alone would predict that the fraction of R_n used as LE would be greater in the fall months as compared to spring months. Increasing temperature, especially in arid areas, would also indirectly affect the second half of the prediction equation since the higher the air temperature with a given absolute humidity in the air, the greater would be the wet-bulb depression, D . General considerations then lead to the conclusion that for clear weather under very humid conditions we might expect less of a correlation between ET and temperature than in dry windy climates.

A recent analysis similar to the preceding exercise, but using the van Bavel (23) equation, was presented by Skidmore, et al. (24). They chose to pick two levels of vapor pressure and vary temperature and wind, and the results illustrate even more dramatically, the same points discussed above.

Another factor which is interesting to consider is that of the effect of aerodynamic roughness of the crop itself. Although no indication of the magnitude of the effect of this factor is suggested here, it is obvious that under lower wind speeds and higher humidities (low values of D in Equation (11)), the effect of roughness should be much less than with higher wind speeds and lower humidities. This would be true since only the value of h in the McIlroy equation would vary with crop roughness. This helps explain why Penman (9) in England with high humidity, cooler conditions suggests little effect of crop roughness on ET while work in hotter and drier climates seems to point to a significant effect of crop roughness on the evapotranspiration process.

The Penman (25) equation as modified by van Bavel (23) included the use of surface roughness, z_o and zero-plane displacement, d in the log-law wind profile formulation. Although widely acclaimed for elimination of the empiricism within the aerodynamic term of Penman's equation, it appears that this method does not provide a realistic way of investigating the effect of roughness per se on evapotranspiration. Rosenberg (26) reports considerable overestimation of ET for alfalfa using measured z_o values in the van Bavel equation and suggests the equation to be too

sensitive under strong winds. This is also very obvious from an example given by Kovda et al. (27) on page 305 where potential ET values exceeding 2.0 and approaching 3.0 mm/hour are predicted by the van Bavel equation for windy desert conditions, and this, using a z_o of only 1.0 cm, and with an air temperature of only 104°F (conditions only slightly more severe than the extreme case in Fig. 10). Considering the 1 to 2 ly/min sensible heat transfer downward which would be necessary to provide the total energy required for this to take place, the prediction seems entirely out of line*. From a practical standpoint the $f(z_o, d)$ functional relationships developed experimentally by several researchers may provide a more reliable evaluation of the effect of crop height and surface roughness on evapotranspiration (see Tanner and Pelton (18), Rijtema (28), Stanhill and Fuchs (29), Slabbers (30), and Feddes (31). Between the above a wide range of crops and climates is covered.

Summary on the Energy Balance

Although the net radiation of a surface is no doubt the most important single factor affecting evapotranspiration, Davis studies in the early 1960's clearly showed a strong out-of-phase relationship between the two, both diurnally and seasonally (33), (34). It is clear from the results considered herein that the partition of the net radiation into various components of the energy balance is very much a function of other climatic factors such as wind, temperature, and humidity as well as plant factors. On an annual basis, we have noted that LE and R_n can be very close, for example, as at Davis; or LE can exceed R_n considerably even in large irrigated valleys surrounded by desert areas, such as at Tempe, Arizona. Conversely, in areas near cold oceans LE can be considerably lower than R_n such as at Lompoc, California.

Seasonally the ratio of $-LE/R_n$ can vary considerably at a given location, tending to be lower in spring months as compared with fall months. Diurnally this ratio varies even more widely. Under calmer conditions, for example at Davis in mid-summer, $-LE/R_n$ typically climbs from a low of around 0.6 during the first two or three hours after R_n becomes positive, to values well over 1.0 by late afternoon.

Let us now try to relate our discussion to a definition of potential evapotranspiration which evolved from the same meeting where Penman (9) presented the one given earlier in this paper. This revised statement (32) defines it as "the rate of evapotranspiration, primarily determined by the weather, from an extended surface of a short green crop, actively growing, completely shading the ground, of uniform height and not short of water." Just what is meant by "an extensive surface (or fetch)?" Unless one is thinking in terms of perhaps hundreds of miles of a fetch

*Refer to Fig. 8 where H peaked at 0.65 ly/min for the dry desert in June where surface temperature probably exceeds air temperature by several score degrees. Clearly, a heat flux greater than this in the opposite direction could not take place without the surface reaching an unrealistic temperature on the cold side, which in turn would limit the vapor gradient needed to produce the high latent heat flux.

See Monteith (1985)

of similar upwind surface conditions, potential evapotranspiration does not suggest, as some writers have proposed, that LE/R ratios should necessarily be close to 1.0. In fact, with the high relative humidity normally experienced with such an extensive fetch, ratios close to 1.0 should be common only during hot, mid-summer periods. Equations (10) and (11) suggest ratios of LE/R closer to 0.5 would be common for cold winter or spring periods.

If by "extensive fetch" a mile or so of similar upwind vegetation is considered adequate, it is obvious from the data for Lompoc, Davis and Tempe that potential evapotranspiration as defined can vary widely in spite of similar net radiation conditions. One might go further and exclude any fetch requirement in the definition of potential evapotranspiration. Although such a definition has become increasingly in vogue, it seems to have little left to offer as an aid in classification of climate, or in relating to crop water requirements within typical irrigation projects.

The use of the McIlroy energy balance-aerodynamic equation illustrates quite well the reasons there can be no simple answer to a question like "What is the effect on ET of doubling the wind speed?" The evaporation process is a highly complicated process with many interrelated factors. For example, wind may quite drastically affect ET percentage-wise under dry, low-radiation conditions, while under high-radiation, high-humidity conditions the effect of wind may be considerably less important. Under calm conditions with a given amount of radiation, the effect of humidity levels may be relatively unimportant, especially with high radiation levels. Likewise, the importance of temperature differences of even 10-15°F should be relatively unimportant under calm humid conditions but would be very significant under dry windy conditions.

Actual evapotranspiration by crops can, of course, be very significantly affected by stomatal resistance. The degree of plant cover, stage of maturity and moisture status of the surface layer of soil are also important factors.

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See Hantel for consideration of stomatal resistance

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