

Evaporation from a Drying Soil

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ABSTRACT

A combination formula for evaporation that uses the surface temperature measured by infrared thermometers as a boundary condition is successfully tested against the evaporation obtained from a detailed energy balance and Bowen ratio measurements. The transfer coefficients used in the combination formula are obtained from aerodynamic similarity and include the effect of the diabatic turbulence. Two simple resistive models which attempt to account for the reduction of evaporation due to the water vapor desaturation of the soil surface are analyzed, but fail to describe correctly the transfer processes through the dry upper layer of the soil.

1. Introduction

The evaporation from bare soils has been studied in detail almost exclusively by observing packed soil columns. The main reason for this approach is the difficulty of estimating evaporation *in situ*. Covey and Bloodworth (1966) indicate that the variation of the atmospheric conditions have a considerable effect on evaporation from bare soils in contrast to successful soil physics approaches which predict evaporation on the basis that capillary flow alone limits evaporation. Reliable and fairly easy methods to determine evaporation from undisturbed soil surfaces, over short periods, are thus needed for evaluating in detail the effect of meteorological variables upon the evaporation. Detailed information on vapor transfer processes in the vicinity of the soil-atmosphere interface is needed since this is where evaporation controls can be best exerted. Studies of evaporation from soil are also important to transpiration research by distinguishing soil evaporation from transpiration.

A form of the combination method proposed by Tanner and Fuchs (1967), which involves only measurements in the atmospheric surface layer, including a measurement of soil temperature, can be applied to estimate the evaporation from bare soils. Starting from the energy balance equation

$$R_n + G + E + H = 0, \quad (1)$$

and assuming equality between the transfer coefficient h for heat and water vapor, so that

$$H = \rho c_p h (T_s - T_0), \quad (2a)$$

$$E = (\lambda \rho \epsilon / P) h (e_s - e_0), \quad (2b)$$

we showed that the evaporation is given by

$$E = [(\gamma + s)/s] E_p - (\rho c_p / s) h (e_s - e_0^*), \quad (3)$$

where the potential evaporation E_p is defined by

$$E_p = [\rho c_p h (e_s - e_s^*) - s(R_n + G)] / (\gamma + s). \quad (4)$$

The symbols appearing in the equations are listed at the end of the article.

The derivation of (3) involves no assumption on the saturation deficit at or underneath the surface. This does not imply that the evaporation is independent of the saturation deficit at the surface, but that its effect is accounted for by the surface temperature, which is directly related to e_0^* . As a soil surface dries, T_0 increases and e_0^* , which is the saturation vapor pressure corresponding to T_0 , increases sharply, even though e_0 may decrease.

A suitably accurate estimate of h is required for (3). If similarity for water vapor, heat and momentum transfer in the atmospheric boundary layer holds, h can be found from the wind velocity V_z at height z , and the roughness length z_0 of the surface. Businger (1956), Tanner and Pelton (1960), Monteith (1963) and Van Bavel (1966) have computed h over watered vegetation by assuming near adiabatic conditions. However, the large temperature lapses that develop in the air during sunny days over bare soils create considerable instability which increases the rate of eddy transfer processes.

The diabatic profile has been described in several ways, one of which is the KEYPS function (e.g., Sellers, 1965), which yields the following expression for the transfer coefficient:

$$h = C_z V_z = k^2 V_z / [\Phi + \ln(z + z_0)/z_0]^2, \quad (5)$$

where the diabatic influence (Lettau, 1962; Panofsky, 1963) function Φ at height z is then given by

$$\Phi = \int_0^z [(\varphi - 1)/(z + z_0)] dz, \quad (6)$$

where

$$\varphi = (1 - 18 \text{ Ri})^{-0.25}, \quad (7)$$

and

$$Ri = (g/\bar{T})[(\ln(z+z_0)/z_0)(T_z - T_0)/V_z^2]z. \quad (8)$$

Our purpose is to test experimentally if (5) yields an adequate estimate of h , and consequently, that (3) can be practically used to determine the evaporation. Stearns (1966) found that the KEYPS function satisfactorily accounted for instability in predicting heat transfer from dry sands.

We propose also to analyze two simple resistive models that attempt to describe quantitatively the internal mechanisms that restrict evaporation from bare soils. Both models assume the presence of a saturated water vapor source at a plane underneath the surface. The first model considers that a dry porous layer of finite thickness covers the moist soil and that water vapor saturation obtains at the boundary between the dry and the moist layer. The resistance of the dry layer to water vapor transfer is then given by

$$r_i = (\lambda\rho\epsilon/P)(e_0 - e_i^*)/E. \quad (9)$$

The second model, which has been proposed by Monteith (1963) for a vegetal canopy and has been applied by Covey (1965) to bare soil, expresses the surface resistance as

$$r_i = (\lambda\rho\epsilon/P)(e_0 - e_0^*)/E. \quad (10)$$

This model results from (9) by the additional assumption that the thermal conductivity of the dry layer is infinite so that $T_0 = T_i$ and $e_0^* = e_i^*$.

Both models considerably simplify the real situation in a drying bare soil. We propose to test if they yield answers that are physically reasonable.

2. Experimental procedure

Measurements were made on a 5×5 m plot of Plainfield sand near the middle of a disked and smoothed, 100×100 m area. During the measurements, the wind directions were from the SW, W and NW. The minimum fetch was 60 m of bare sand. Grass forages and grain stubble extended for another 100 m beyond the sand. The Plainfield sand has low organic matter content, and is over 90% quartz sand and less than 3% clay.

The evaporative flux to which all our other determinations are compared was obtained from measurements of the energy balance and the Bowen ratio. This method has been demonstrated earlier to be satisfactory at this site by comparison with lysimeter measurements (Suomi and Tanner, 1958).

The net radiation was measured with a ventilated Suomi-type radiometer (Suomi *et al.*, 1954) from a height of 150 cm.

To evaluate the soil heat flux through the 5-cm depth, 5 flux plates with high thermal conductance were connected in series. The flux plates were individually calibrated in the laboratory in a sand tank by using a sheet electrical heater. The average value of the cali-

bration coefficient was within 5% of the value obtained from the direct computation of the heat flux using a steady-state temperature gradient measured across the sand and the heat conductivity of the quartz sand. Additional errors could have resulted during the field use of the flux plates (Philip, 1961); however, the relative error was checked *in situ* by comparing 30-min averages of the flux plate values against the soil heat flux computed from the temperature gradient in a 3.8-cm layer of soil immediately above the plates during a fairly steady cooling period. The flux plate measurement was 20% higher than the computed value, but part of the discrepancy is due to the phase difference resulting from the slightly different depth of the two measurements, to the neglected heat storage, and the error in the estimation of the thermal conductivity of soil. In view of all these considerations the relative error associated with the use of the heat flux plates is estimated to be between 5 and 10%.

The fraction of soil heat flux due to the change of the heat stored in the 5 cm of soil above the flux plates was determined by calorimetry. The temperatures were measured by differential thermopiles of 5 copper-constantan junctions between depths of 0, 1.2, 5.0 and 30.0 cm. The thermocouple wires were 0.025 cm in diameter and were embedded in epoxy in a plane. The junctions in contact with the dry surface were free from epoxy and were dusted with dry sand. The absolute temperature at -30.0 cm was measured with a germanium diode. The heat capacity of the soil layers between 0 and 1.2 cm, and 1.2 to 5 cm, were computed from their water content.

The Bowen ratio measurement was made with the sampling equipment described by Sargeant and Tanner (1967). The values of ΔT and Δe were obtained from differential wet-bulb and dry-bulb temperatures. The relative error in E contributed by temperature measurement error was 0.02β where β is the Bowen ratio. The sampling heights were at 32 and 100 cm so that the uppermost instrument was at a minimum fetch-height ratio of 60:1. The absolute wet-bulb and dry-bulb temperatures were measured at both levels so that both T_z and e_z at 100 cm were available.

The above measurements determined E and H through the energy balance. The combination formula further requires the wind velocity at 100 cm, z_0 , T_0 and e_0 . The wind velocity at 100 cm was measured by a 3-cup, sensitive Casella anemometer of the contact closure type. The impulses were accumulated over 15-min periods with Sodeco impulse counters.

The soil surface temperature T_0 was determined with the thermocouples described previously and with two 8-13 μ infrared (IR) thermometers with fields-of-view (f.o.v.) of 7° and 30°, respectively. They were mounted on masts at a height of 180 cm and viewed the soil at an angle of incidence of 45°, with an orientation to prevent shadows in their f.o.v. The effect of the surface emissivity in the sensitivity band of the instrument and of

the surroundings' radiation was found as given in Fuchs and Tanner (1966). The infrared emissivity of the Plainfield sand was found to be 0.900 ± 0.005 . This value has been confirmed by a spectral reflectivity curve made on a sample of the upper horizon of the soil.¹ The spectral curve also shows that the reflection band of quartz around 9μ is sufficiently attenuated by the colloidal coating of the sand grain to cause no detectable errors in the measurements. The two radiometers generally agreed within 0.2C.

The roughness length can be determined from a 3-level wind profile or from empirical relations between z_0 and the silhouette area of the roughness elements of the surface (Kutzbach, 1961). Wind profile measurements were not available in our experiments and we used the roughness length of similar sands ($z_0 = 0.04$ cm) measured by Stearns (1966).

We had no instruments to determine e_0 to test (9) and (10), but used similarity to extrapolate the e_{100} and e_{32} to e_0 in the same relation that T_{100} and T_{32} extrapolated to T_0 . This similarity assumption is necessary to the combination model and the Bowen ratio approach. In view of the very small vapor gradients, e_0 estimates were reliable within ± 1 mb, regardless of errors in the similarity assumption.

3. Energy balance results

The terms of the energy balance equation measured on 26 August 1966 are plotted in Fig. 1. The volumetric water content was 0.5% in the top 1.5 cm of soil, 7% in the layer from 1.5 to 5 cm, and 10% beneath 5 cm. The evaporation curve shows a daily rhythm indicating that the process is dependent upon the atmospheric evaporation conditions.

A large portion of the net radiation goes into the soil heat flux. Errors associated with the soil heat flux mea-

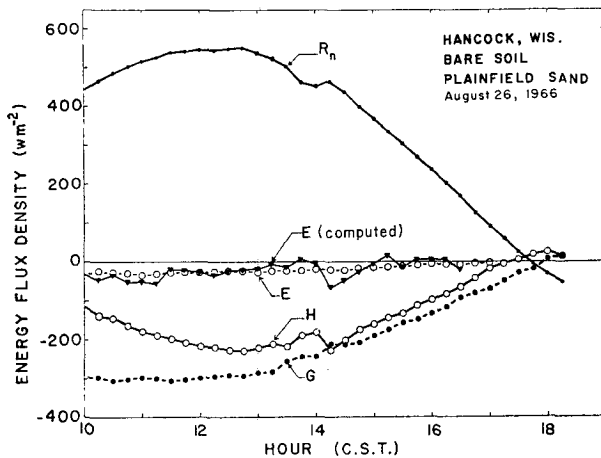


FIG. 1. Daily trend of the energy balance terms. The curve E (computed) is obtained with the combination formula.

¹ This determination was made by W. A. Hovis, NASA, Goddard Space Flight Center, Greenbelt, Md.

surement become critical in the late afternoon when the net radiation and the soil heat flux are approximately equal. The effect of these errors appears clearly after 1730 CST when a sensible heat flux toward the surface is indicated, whereas the inversion of the temperature occurs 30 min later (Fig. 2). This is most likely due to an overestimate of the soil heat flux. The phase lag is small indicating that the midday errors are negligible.

The daily trends of the temperature of the air, the surface and the soil are plotted in Fig. 2, showing the large temperature gradients in the air and in the soil during most of the day. The surface temperatures plotted in Fig. 2 are those measured with the IR thermometers. The thermocouple measurements of the soil surface temperature were 2-4C higher than the IR thermometer measurements. The IR measurements were

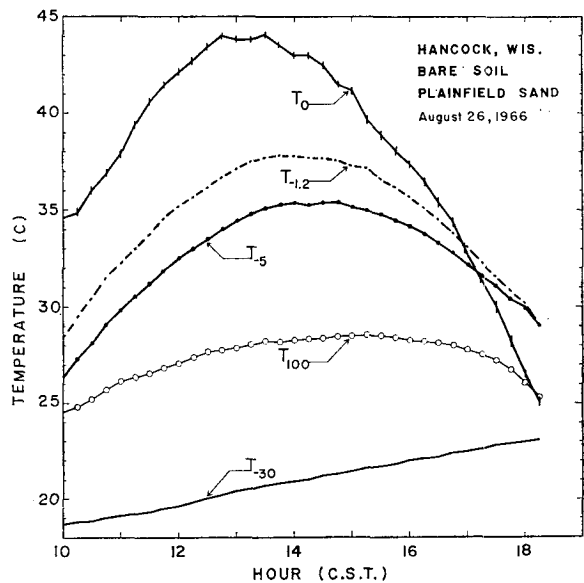


FIG. 2. Temperature of the soil (-1.2, -5, -30 cm), of the surface, and of the air (100 cm).

taken as more reliable because they were within 0.1C of the air temperature when the air was nearly isothermal, whereas the thermocouple readings were still high. We believe the higher thermocouple readings were due to poor thermal contact and to radiation error.

4. Transfer coefficient results

Using the sensible heat flux H from the energy balance and the measurements of the air temperature T_{100} and the surface temperature T_0 , the experimental value of h was obtained from (2a). This is compared with the transfer coefficient computed from (5) with $z_0 = 0.04$ cm. The large negative Richardson numbers that prevailed during the measurements resulted in a large diabatic correction. The transfer coefficients computed from the simple logarithmic profile averaged 25% smaller than those obtained from (5).

Similar computations were made from intermittent measurements taken on 17 August, which was slightly windier. The negative Richardson numbers were smaller than on 26 August but distributed over a larger range, from -0.09 to -0.22 . Corresponding to these values of Ri , the transfer coefficients obtained from the simple logarithmic profile varied from 0.87 to 0.77 of the experimentally measured h . No systematic difference obtained between the values of h from (5) and those measured during the midday period, the mean difference being only 3%.

The daily trend of the computed $h [= C_z V_z]$ and the experimental $h [= H/\rho c_p(T_z - T_0)]$ are plotted in Fig. 3. The differences between the two values are small, and well within the experimental errors. The decrease observed in the morning and late afternoon values of h resulted from the low wind observed during these periods, and the somewhat decreased atmospheric instability as indicated by smaller negative Richardson

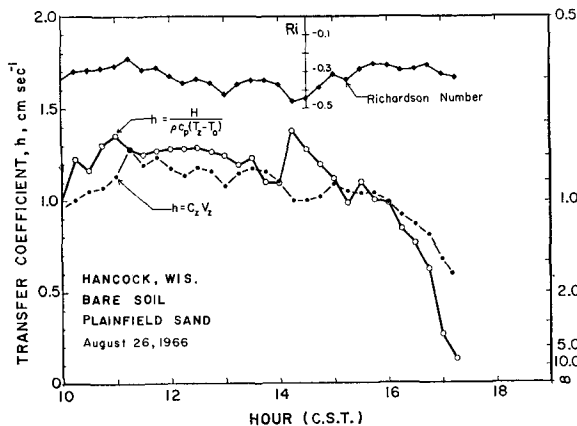


FIG. 3. Comparison between the measured values of the transfer coefficients and those predicted from aerodynamic similarity.

numbers. The decrease of the aerodynamically computed h lags behind the experimental h ; however, this occurs at a time when the energy fluxes become smaller and when the overestimate of G tends to minimize the size of the measured h .

5. Combination formula results

The evaporation was computed from (3) using the aerodynamic h . The results are plotted in Fig. 1. The values are close to those obtained from the energy balance and Bowen ratio method, although the relative errors are large. It is noted that $(R_n + G)$ is the only term common to both determinations of E . These two determinations of E are correlated in Fig. 4, which includes the data taken on 17 August 1966 when the soil was more moist and had a higher evaporative rate. The largest evaporation rates shown are still an order of magnitude smaller than the rates normally obtained

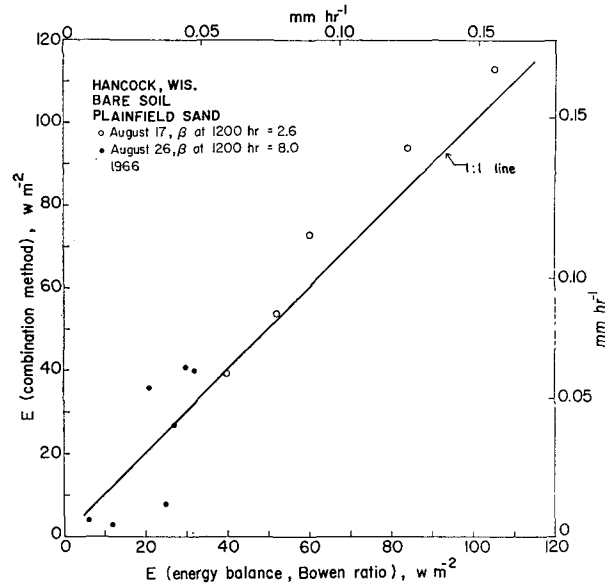


FIG. 4. Comparison of hourly evaporation from drying soils obtained with the new combination formula and with the standard energy balance-Bowen ratio method.

from wet soil. The agreement between the two determinations of E on an hourly basis is good.

6 Internal resistance models

The resistances of the upper dry layer of the soil to water vapor diffusion as defined by (9) and (10) are plotted in Fig. 5. The molecular diffusive resistance which is computed from the molecular diffusivity in air of water vapor ($0.24 \text{ cm}^2 \text{ sec}^{-1}$), the porosity of the dry layer (0.38), and a tortuosity factor of 0.66 is also indi-

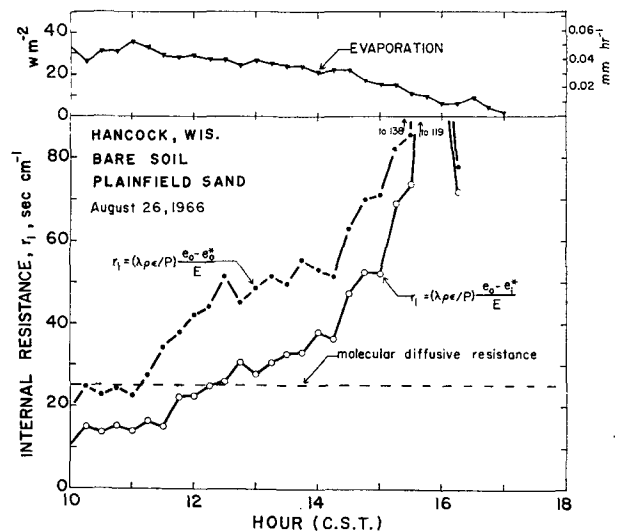


FIG. 5. Daily trend of the internal resistance to water vapor transport from a sandy soil with a dry surface layer of 1.5 cm as predicted by two models.

cated in Fig. 5 and constitutes an upper bound for the resistance of the dry layer.

From 1200 on, both models indicate a steadily increasing resistance that is far greater than the molecular diffusive resistance. Although it is quite possible that the molecular diffusive resistance increased slightly during the day, it could not increase by a factor of 5 as did the resistances predicted by the models. This would have required a dry layer 7.5 cm thick, whereas the observed change in the thickness of the dry layer was too small to be measurable. Furthermore, an evaporation rate 10–15 times larger than that measured would be required to develop a 7-cm dry layer.

The daily trend of the two r_i estimates run parallel, but this is merely a consequence of the relation between the surface temperature and the temperature in the soil. The exceedingly large values obtained after 1500 are questionable because of the uncertainty associated with the energy flux measurements in the late afternoon. The model proposed by Monteith and Covey is further off than the other model because of the additional assumption of isothermy in the dry layer involved in the formulation of the model.

Measurements made on 17 August 1966, when the thickness of the dry top layer of the soil varied between 0.2 to 0.4 cm, resulted in similar conclusions on the validity of the resistance models (Fig. 6). The corresponding molecular diffusive resistance was between 3 to 8 sec cm^{-1} , whereas the two models yield resistance values that are about half of those obtained on 28 August.

Physically, the two models are unacceptable. Including a temperature gradient across the dry layer does not improve significantly the model proposed by Monteith and Covey. Obviously, saturation does not obtain at an apparently sharp boundary between the dry and the

moist layers of sand. The water vapor sources in the sand are distributed over a layer rather than being concentrated at a well-defined plane, as the models assume.

The concept of a simple resistive network representation of the diffusion of water vapor from the soil to the atmosphere must be rejected when the vapor sources are distributed. Moreover, the usefulness of the concept of an internal resistance to assess the evaporation is questionable, because the generalized formulation of the combination method as given by (3) described the evaporation without any reference to the internal diffusive resistance of the soil.

List of symbols

- C_z drag coefficient
- c_p specific heat of air
- E latent heat flux density (evaporation)
- E_p potential evaporation
- e_z water vapor pressure in the air at height z
- e_0 water vapor pressure at the surface
- e_i^* saturated water vapor pressure at the water vapor source corresponding to T_i
- e_z^* saturated water vapor pressure corresponding to T_z
- e_0^* saturated water vapor pressure corresponding to T_0
- G soil heat flux density
- g gravity
- H sensible heat flux density
- h transfer coefficient for heat and water vapor from the surface to the height z in the air
- k von Kármán constant (=0.428)
- P barometric pressure
- Ri Richardson number
- R_n net radiation flux density
- r_i internal resistance of the soil surface layer to the transport of water vapor
- s slope of the saturation water vapor curve versus temperature
- T_i soil temperature at the boundary between the dry and moist layer
- T_z air temperature at height z
- T_0 surface temperature
- \bar{T} $(T_0 + T_z)/2$
- V_z wind velocity at height z
- z height in the air from the surface
- z_0 roughness length
- β Bowen ratio
- γ psychrometric constant
- ϵ 0.622
- λ latent heat of vaporization of water
- ρ density of dry air
- Φ profile diabatic influence function defined by (6)
- φ KEYPS function defined by (7)
- Δe water vapor pressure difference between two levels in the air
- ΔT temperature difference between two levels in the air

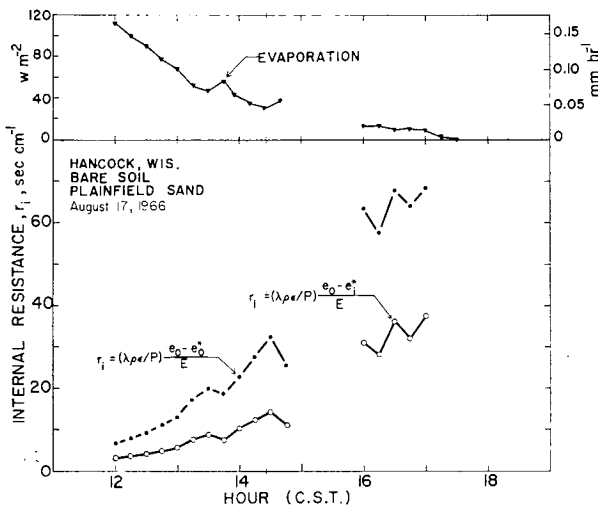


FIG. 6. Daily trend of the internal resistance to water transport from a sandy soil with a thin (0.2–0.4 cm) dry surface as predicted by two models.

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