EVAPORATION FROM NATURAL NONSATURATED SURFACES

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ABSTRACT

Following a development similar to that used by Penman (1948) a general combination equation is derived to describe evaporation from nonsaturated surfaces. To account for departure from saturated conditions, the equation makes use of the concept of relative evaporation, the ratio of the actual to the potential evaporation, defined here as the evaporation rate which would occur under the existing atmospheric conditions if the surface were saturated at the actual surface temperature. A relationship is established between the relative evaporation and a dimensionless parameter called the relative drying power, the ratio of the drying power to the sum of the drying power and the net available energy (net radiation plus soil heat flux). The relationship is non-dimensional and appears to be single-valued. The combination of this relationship with the general evaporation equation derived constitutes a simple model for obtaining estimates of evaporation from nonsaturated surfaces; no prior estimate of the potential evaporation is required, and the surface conditions of temperature and humidity need not be known.

INTRODUCTION

A standard approach for estimating the "actual" evaporation (evapotranspiration) \( (E) \) in operational practice is to moderate the "potential" rate \( (E_p) \) according to soil and vegetative characteristics. Modelling evaporation by this procedure has the advantages that it is simple, direct and makes use of easily measured atmospheric, soil and crop variables. A major problem with the approach however is that it requires a prior calculation of "potential" evaporation and despite its importance, potential evaporation has yet to be clearly defined in a universally-accepted manner. One can distinguish two schools of thought regarding potential evaporation, as evidenced by the two major types of definition found in the literature. On the one hand are those hydrologists who would have potential evaporation defined from atmospheric or energy considerations alone, and on the other are those who would require that it be governed as well by surface temperature and moisture conditions. Potential evaporation used in the models is generally calculated from atmospheric conditions.
variables and the expression most frequently-used for this purpose is the combination equation presented by Penman (1948). It was Penman’s intent that the combination equation would provide a useful and reliable estimate of evaporation from a body of water with which actual evaporation might be correlated. The relationship was thus developed as a “model” of the evaporation from a wet environment with an unlimited water supply. Its application for providing an accurate estimate of $E_p$ for surfaces where soil, surface and atmospheric conditions differ appreciably from these is questionable. The wet surface (open water or saturated soil) case represents the only situation for which an expression for evaporation (such as Penman’s) can be developed in such a way that the surface parameters are excluded. This is possible only because the relationship between surface temperature and surface vapor pressure is unique and known for a saturated surface. For a nonsaturated surface the relationship between surface temperature and vapor pressure is not unique. An approach such as Penman’s can work for nonsaturated surfaces only if some other relationship is introduced which allows for the elimination of the surface parameters. Considering the current state-of-the-art, such a relationship would necessarily be an empirical one between the measured atmospheric parameters and the nonmeasured surface parameters. Bouchet (1963) and Morton (1983) for example suggest that in fact the atmospheric conditions do reflect the surface parameters, and thus establish the basis for such an empirical relationship.

In this paper, much in the same manner as Penman considered a saturated surface and developed a combination equation describing evaporation, the case of a nonsaturated surface is considered and a general equation describing evaporation is developed. This general equation makes use of the concept of relative evaporation (the ratio of actual to potential evaporation). An empirical expression relating the relative evaporation to readily-measured atmospheric parameters is developed from field observations. This expression eliminates the need for observations of surface temperature and vapor pressure for calculating actual evaporation.

**THEORY OF EVAPORATION FROM NON-SATURATED SURFACES**

A nonsaturated surface is characterized by the fact that the vapor pressure at the surface, $e_s$, is less than the saturation vapor pressure at the surface temperature, $e_s^*$. The energy balance in the vertical direction at a horizontal surface, assuming advection is negligible, can be written (with the terms expressed as an equivalent depth of evaporation) as:

$$E + H = Q$$ (1)

where $E$ is the evaporation, $H$ is the sensible heat, and $Q$ is the total energy available from net radiation and soil heat. Applying the Bowen ratio:
\[ \beta = \frac{H}{E} = \frac{\gamma(T_s - T_a)}{(e_s - e_a)} \]  

(2)

and the slope of the saturation vapor pressure curve, \( \Delta = \frac{de^*}{dT} \), equal to:

\[ \Delta = \frac{(e_s^* - e_a^*)}{(T_s - T_a)} \]  

(3)

to eqn. (1) gives:

\[ Q = E \left[ 1 + \frac{\gamma(e_s^* - e_a^*)}{\Delta(e_s - e_a)} \right] \]  

(4)

where: \( \gamma \) = psychrometric constant; \( e_s^* \), \( e_a \) = saturation and actual vapor pressures at the evaporating surface at a temperature, \( T_s \); and \( e_s^* \), \( e_a^* \) = saturation and actual vapor pressures of the air at a temperature, \( T_a \).

Evaporation can also be expressed by the Dalton-type bulk transfer equation:

\[ E = f(u)(e_s - e_a) \]  

(5)

or:

\[ e_s - e_a = \frac{E}{f(u)} \]  

(6)

where \( f(u) \) is a wind speed function. Substituting eqn. (6) into eqn. (4) and rearranging terms yields:

\[ E = Q + \frac{\gamma}{\Delta} f(u)(e_s^* - e_a^*) \]  

(7)

If the vapor pressure of the air, \( e_a \), is added and subtracted to the difference in vapor pressure eqn. (7) becomes:

\[ E = Q + \frac{\gamma}{\Delta} [f(u)(e_s^* - e_a) - f(u)(e_s^* - e_a)] \]  

(8)

The first term in the brackets in eqn. (8), \( f(u)(e_s^* - e_a) \), the product of the wind function and the vapor pressure deficit of the air, is generally referred to as the "drying power" of the air \( (E_s) \). The second term in the brackets, \( f(u)(e_s^* - e_a) \), is the evaporation rate which would occur under the same atmospheric conditions for wind and humidity, if the surface was saturated at the temperature of the surface. This term therefore represents the potential evaporation, \( E_p \) (Van Bavel, 1966). Equation (8) is a general equation relating evaporation to the net energy available by radiation and conduction, the drying power of the air and the potential evaporation. It can be used to calculate \( E \) when an adequate expression for \( E_p \) is available. The formulation for \( E_p \) although quite simple, has the major drawback of requiring a measurement of the surface temperature, a parameter which is rarely observed.
Consider the general case of evaporation from a nonsaturated surface at some rate less than the potential, i.e. \( 0 < E < E_p \). The relative evaporation, the ratio of actual to potential evaporation, \( G = E/E_p \), should be a unique parameter for each set of atmospheric and surface conditions. Using eqn. (5) and the equivalent expression for \( E_p \), \( G \) can be expressed as:

\[
G = \frac{E}{E_p} = \frac{f(u)(e_a - e_s)}{f(u)(e^*_a - e_s)}
\]  

Equation (9) shows that the relative evaporation is governed by the vapor pressure deficit at the evaporating surface, \( e^*_a - e_s \), or by the availability of water at the surface. In other words surface conditions play a predominant role in the partitioning of the available energy to evaporation, and thus also control the potential evaporation. For a wet surface, where \( e_s = e^*_a \), \( G \) will be equal to unity; for a very dry surface, \( e_s \) approaches \( e_a \) and \( G \) will approach zero, thus, \( 0 \leq G \leq 1 \). Substituting \( E = GE_p = Gf(u)(e^*_a - e_s) \) into eqn. (8) and simplifying yields the general expression for evaporation from a nonsaturated surface as:

\[
E = \frac{\Delta GQ}{(\Delta G + \gamma)} + \frac{\gamma GE_a}{(\Delta G + \gamma)}
\]  

Equation (10) is similar in form to the Penman equation, but differs through the inclusion of the relative evaporation, \( G \), which accounts for departures from saturated conditions. It is interesting to note that Barton (1979), in a parallel development, introduced the relative humidity at the soil surface (rather than the relative evaporation) and produced an equation resembling eqn. (10) to account for nonsaturated conditions. He used his expression to attempt to extend the Priestley–Taylor (1972) hypothesis to include nonsaturated surfaces.

**Relative Evaporation**

To determine \( E \) by eqn. (10) requires an estimate of the relative evaporation, \( G \). Rearranging the equation and simplifying gives an expression for \( G \) equal to:

\[
G = \frac{\gamma E}{\Delta Q + \gamma E_a - \Delta E}
\]

Several investigators, for example Priestley and Taylor (1972), Black (1979) and Federer (1979), have related parameters similar to \( G \) to soil water deficits or to available soil water. The major disadvantages of such relationships are they are empirical and neither single-valued nor unique. Thus, the practical application of the model would be greatly enhanced if \( G \) could be related in a universal manner to readily-measured or calculated parameters. These would exclude the surface parameters, such as temperature and vapor pressure, which are rarely measured but which govern both actual and potential evaporation.
Morton (1983) points out that the effects of changes in the availability of soil water on potential evaporation can be assessed by their effects on the temperature and humidity gradients. Since an increase in actual evaporation causes the vapor pressure of the overlying air to increase, the drying power, $E_a$, reflects to some extent the "dryness" of the surface. Further, the sum of $E_a$ plus the available energy, $Q$, should index the "potential" for evaporation. Following this logic it is assumed that $G$ can be related to the ratio $D = E_a / E_a + Q$, the relative drying power. The relationship, which is also dimensionless, is an inverse one; for a dry surface ($G = 0$), $E_a$ is large and $D$ approaches unity. For a wet surface ($G = 1$), $E_a$ and $D$ approach zero.

An investigation of the $G$-$D$ relationship was undertaken using field data monitored at two stations located in the semi-arid climatic zone of western Canada. Both sites, Bad Lake and Saskatoon, include silty clay and heavy clay soils which form part of the Brown soil zone in the Province of Saskatchewan. Relative evaporation was evaluated by eqn. (11) from estimates of $E$, $E_a$, $Q$, $\gamma$ and $\Delta$ calculated from observations made at the stations. A brief review of the procedures used for the calculations is given below.

The data were separated and grouped into two periods: postmelt (the period extending from the end of snowcover ablation to the day of seeding of an annual crop); and the crop growing season. As well the data were stratified according to landuse and vegetative cover, that is as fallow, stubble, growing crop (wheat) and grass. Stratification was necessary for several reasons: (a) the ground heat flux, surface albedo, and soil moisture and temperature regimes and other factors affecting the net energy available for evaporation ($Q$) differ widely between postmelt and growing-season periods; and (b) the calculation of the drying power ($E_a$) includes a wind function $[f(u)]$ that is affected by surface roughness, which in turn depends on the type, density and height of vegetation.

Evaporation was calculated as the residual term in a soil water balance applied at field sites where precipitation and soil moisture were measured. Profiles of changes in soil moisture were obtained with a two-probe gamma, density meter. Complete details on the determination of soil moisture soil density by this method are well-documented in the literature (e.g. see Troxler, undated; Smith et al., 1967; Reginato and Jackson, 1971; Jame and Norum, 1980). In principle the method measures soil wet density and a change in density in a measurement interval is assumed to be due to a soil moisture change. Profiles of soil moisture changes were obtained at 20 mm increments of depth to 1 m, and 40 mm increments between 1 and 1.6 m. Repeatability tests conducted with the equipment in the field gave a standard error of estimate in soil moisture content of 2.5 mm in a 1-m profile. Precipitation estimates used in the budget calculations were daily values collected by a standard Canadian rain gauge.

Although radiation measurements were available at both sites, these were plagued by frequent intervals of missing data. Hence, they were not used directly for calculating the energy flux $Q$, except as a check of values of net radiation estimated by the model developed by Gray and Landine (1988). Soil
heat flux was assumed to be negligible and omitted from the calculation of $Q$ except for those periods during postmelt when a soil was thawing. It was observed that during the "thaw" period a significant fraction of the net radiation is apportioned to soil heat. Soil temperature measurements taken daily at the sites themselves allowed for calculation of this term assuming a simple, steady-state conduction model.

Since a variety of surface conditions were encountered, use of a single wind function such as proposed by Penman (1948) was inappropriate. Therefore, the function was calculated from boundary layer and surface roughness considerations based on the well-known aerodynamic formula:

$$ E = \frac{0.622k^2 \rho u_1(e_s - e_2)}{P \ln \left[ \frac{z_1 - d_0}{z_0} \right] \ln \left[ \frac{z_2 - d_0}{z_0} \right]} \quad (12) $$

in which $E$ is the evaporation rate ($\text{cm s}^{-1}$), $k$ is the von Karman constant (0.4), $\rho$ is the air density ($\text{g cm}^{-3}$), $P$ is the atmospheric pressure (mb), $u$ is windspeed ($\text{cm s}^{-1}$) and $e$ is vapor pressure (mb). The subscripts 1 and 2 refer to the measurement heights $z_1$ and $z_2$. $d_0$ is the zero-plane displacement of the vegetative surface (cm) and $z_0$ is the roughness height (cm). According to eqn. (12) the wind function of the Dalton-type equation can be taken equal to:

$$ f(u) = \frac{0.622k^2 \rho u_1}{P \ln \left[ \frac{z_1 - d_0}{z_0} \right] \ln \left[ \frac{z_2 - d_0}{z_0} \right]} \quad (13) $$

In addition to atmospheric measurements, eqn. (13) requires information on the roughness elements of the ground surface. The displacement and roughness heights were estimated from measurements of crop and stubble height in the following expressions (Brutsaert, 1982):

$$ d_0 = 0.67 \bar{h} \quad (14) $$

and:

$$ z_0 = \frac{\bar{h}}{7.35} \quad (15) $$

in which $\bar{h}$ is the mean height of the vegetation. Values of $d_0$ and $z_0$ assigned to fallow were 0 and 0.006 m, respectively, and to grass surfaces 0 and 0.16 m.

Values for $E$, $Q$ and $E_s$ estimated by these procedures were used in eqn. (11) to calculate $G$ for 158 periods varying from 2 to 30 days in which soil moisture changes had been monitored. These calculations used the mean values of atmospheric and soil parameters observed during the measurement interval. Relative evaporation is plotted against the relative drying power for each individual period in Fig. 1. The data for the different landuses have been pooled because the differences between the means of $G$ for the different landuse groups are small and the scatter in $G$-values within groups is of similar magnitude. In
Fig. 1. Plot of relative evaporation against relative drying power showing all data points.

fact the relationship between $G$ and $D$ should be independent of landuse provided the various terms used in their calculation properly account for the effects of surface condition on the energy and mass exchange processes. Figure 1 shows a trend for $G$ to decrease nonlinearly with increasing $D$, however the shape of the decrease is masked by the scatter of the data. A clearer picture of this trend is shown in Fig. 2 in which the mean values of $G$ for different class intervals of $D$ are plotted. It appears that the relationship is S-shaped, a shape similar to the curve of the ratio of actual to potential evaporation versus soil moisture content. Unfortunately, the lack of measurements from wet environments ($G > 0.7$) do not allow the development of a functional relation between $G$ and $D$ that can be treated with confidence over the entire range in $G$. An exponential function fitted to the data gave the expression:

$$G = \frac{1}{1 + 0.028e^{8.045D}}$$

with a standard error of the estimate of $G$ of 0.051. Equation (16) can be used with eqn. (11) to calculate evaporation from nonsaturated surfaces. The equations are independent of surface parameters (temperature and vapor pressure) and they do not require an estimate of potential evaporation.
SUMMARY

Penman (1948) developed a combination equation which allows the calculation of evaporation from a wet surface without the need for the surface parameters (temperature or vapor pressure). This study considers the general case of evaporation from nonsaturated surfaces and, using a procedure similar to that followed by Penman, develops a general "combination" expression for evaporation based on energy budget and aerodynamic principles. This expression requires knowledge of the ratio of actual to potential evaporation, termed the relative evaporation $(G)$.

An expression describing $G$ as a function of the relative drying power $(D)$, the ratio of the drying power, $E_a$ to the sum of $E_a$ plus the net energy supplied to the ground surface by net radiation and soil heat conduction $(Q)$ is presented. $G$, $E_a$ and $Q$ are calculated from observations of atmospheric variables, soil moisture change, soil temperature and height of vegetation monitored over fallow, stubble, grass and wheat during spring and summer at two locations in western Canada. The relationship between $G$ and $D$ is curvilinear with a strong trend to be single-valued and independent of landuse. This relationship can be used in the general "combination" equation to obtain an estimate of evaporation.
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