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In the name of Allah

NOTE

EVALUATION OF THE PROFILE METHOD FOR CALCULATION OF EVAPOTRANSPIRATION[1]

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ABSTRACT

To evaluate the profile method in calculation of evapotranspiration, three different methods are used to compute the eddy diffusivity coefficient. Profiles of dry and wet bulb temperatures to a height of 3 m and wind speed to a height of 10 m were measured in a National Weather Service Climatological Station (42°N, alt.1370 m) situated in the center of a broad mountain valley in Utah with a tethered balloon system through a diurnal cycle. Values of stability function, surface friction velocity, eddy exchange coefficient, specific humidity, and air density were deduced from these measurements. Analysis of the hourly data showed that there is a very good agreement in the calculation of evapotranspiration between two of these methods, and

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بنام خدا

ارزیابی روش پروفیل برای محاسبه تبخیر و تعرق

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خلاصه

جهت ارزیابی روش پروفیل برای محاسبه تبخیر و تعرق، از سه روش مختلف جهت محاسبه ضریب پخشیدگی گردبادی استفاده گردید. نیمرخهای دمای خشک و تر تا ارتفاع ۳ متر و سرعت باد تا ارتفاع ۱۰ متر در یک ایستگاه اقلیم شناسی اداره هواشناسی ملی (۴۲ درجه عرض جغرافیائی شمالی و ۱۳۷۰ متر ارتفاع از سطح آزاد دریا) واقع در مرکز یک دره وسیع کوهستانی در ایالت یوتای آمریکا با یک سیستم بالون که با کابل به زمین متصل میگردد در طی یک شبانه روز اندازه گیری شد. مقادیر تابع پایداری، سرعت اصطکاکی سطحی، ضریب تبادل گردبادی، رطوبت ویژه و وزن مخصوص هوا با استفاده از این اندازه گیریها استنتاج گردید. تجزیه و تحلیل داده های ساعتی نشان داد که توافق بسیار خوبی بین دو روش از سه روش محاسبه تبخیر و تعرق وجود داشت و با مقایسه این مقادیر با تبخیر و تعرق پتانسیل حاصله از تشتک تبخیر کلاس A در ۲۰ سپتامبر ۱۹۷۶ ضرایب گیاهی ۰/۶۱ و ۰/۶۸ بدست آمد که ضرایب معقولی برای شرایط تحقیق بشمار میآید.

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crop coefficients of about 0.61 and 0.68 were obtained when comparing the potential evapotranspiration obtained from the class A pan evaporation with the calculated evapotranspiration on September 20, 1976.

INTRODUCTION

A number of methods may be used to determine the movement of water vapor between the surface and the atmosphere. The aerodynamic method makes it possible to calculate the distribution of the water vapor in a vertical profile of the atmosphere as a function of time, source, sink, and atmospheric turbulence characteristic.

Based on the profile method and assuming the eddy diffusivity coefficient to be the same for momentum, sensible heat, and moisture, evapotranspiration (E_t) in cm/sec may be calculated from the relationship (8):

$$E_t = -\rho K_{(z)} (\delta q/\delta z). \qquad [1]$$

When the eddy diffusivity coefficient $K_{(z)}$, in cm^2/sec, the air-density ρ in g/cm^3, and the vertical gradient of water vapor ($\delta q/\delta z$) in cm^{-1} are known, E_t becomes the average vertical transfer of water vapor across a unit area of a horizontal plane. In this investigation, the values of $K_{(z)}$ are calculated using three different formulas developed by Monin and Obukhov (7), Weber *et al.* (10), and Agee *et al.* (1), and the amounts of E_t are compared with potential evapotranspiration E_{tp} obtained from the U.S. Weather Bureau Class A pan evaporation for a complete day.

MATERIALS AND METHODS

A thermocouple psychrometer similar to one developed by Caldwell and Caldwell (2) was constructed for use on a tethered balloon. An insulated water reservoir enables a long period of continuous measurement. The thermocouple was down-linked by insulated leads secured to the tethered line. The psychrometer was carefully calibrated over a wide range of temperatures and humidities and tested for the effects of ventilation and direct solar radiation on the sensors. When in use, it was routinely checked at the surface with respect to an Assmann's psychrometer. An insulated suspension frame was used to mount the apparatus to a tethered helium-filled polyethylene dirigible. Due to the weight of the psychrometer and down-link wires, the additional large outdoor balloons were attached to the dirigible to attain the desired height. The main purpose of this investigation was to model the diurnal variation of the water vapor profile up to 200 m, using a one-dimensional, time-dependent diffusion model.

The investigation took place at the Utah State University Animal Husbandary Farm (42°N, alt. 1370 m), about 8 km Southwest of Logan, Utah. This is also the site of a National Weather Service Climatological Station, near the center of a wide, flat valley bottom, free of nearby topographic obstructions.

Experimental periods in August and September 1976 encountered low wind conditions in the lowest 200 m of the valley atmosphere. Totalizing anemometers yielded windspeeds at heights of 0.6, 3.0, and 10.0 m above the surface.

Referring to Eq. [1], the eddy diffusivity coefficient $K_{(z)}$ could be found from the micrometeorological analysis in different methods. According to Monin and Obukhov (7):

$$K_{(z)}=ku_*z/\phi_m \qquad [2]$$

or

$$K_{(z)}=k^2\ z^2\phi_m^{-2}\ (\delta u/\delta z) \qquad [3]$$

where k=0.41 is the von Karman dimensionless constant, u_* the friction velocity in cm/sec, z the height above ground in cm, ϕ_m stability function, and $\delta u/\delta z$ the vertical wind gradient in sec^{-1}. The values of $\delta u/\delta z$ are obtained by dividing the average amounts of wind speed at a given time and at heights of z_1 and z_2 by (z_2-z_1). The stability function related to the dimentionless Richardson's number R_i. Pruit *et al.* (9) related ϕ_m to R_i as follows:

$$\phi_m=(1+16\ R_i)^{1/3} \text{ for stable conditions} \quad [4]$$

and

$$\phi_m=(1-16\ R_i)^{-1/3} \text{ for unstable conditions} \quad [5]$$

with

$$R_i=(g/\bar{\theta}).(\delta\theta/\delta z)/(\delta u/\delta z)^2 \quad [6]$$

where g=980.35 cm/sec^2 is the acceleration of gravity for the experimental site, $\bar{\theta}$ the average of θ, potential temperature $\theta=T_d/(P_a/1000)^{.286}$ in °K, T_d dry bulb temperature in °K, and P_a the air pressure in mbar.

According to Weber *et al.* (10), the vertical profile of friction velocity u_*, in neutral barotropic conditions can be expressed as:

$$u_*=u_{*(0)}-3.83\text{x}10^{-4}z \quad [7]$$

where $u_{*(0)}$ is the surface friction velocity in cm/sec. Substituting in Eq. [2] yields:

$$K_{(z)}=(u_{*(0)}kz - 3.83 \text{ x } 10^{-4}\ kz^2)/\theta_m. \quad [8]$$

$u_{*(0)}$ could be calculated as proposed by Kao (4):

$$u_{*(0)}=k[(\bar{u}_2-\bar{u}_1)(z_3-z_2)-(\bar{u}_3-\bar{u}_2)(z_2-z_1)]/[(z_3-z_2)\ln(z_2/z_1)-(z_2-z_1)\ln(z_3/z_2)] \quad [9]$$

where $\bar{u}_1$, $\bar{u}_2$, and $\bar{u}_3$ are the mean windspeeds at heights z_1, z_2, and z_3, respectively.

Agee *et al.* (1) showed that $K_{(z)}$ may be expressed by the following model:

$$K_{(z)} = a\ [\exp(-bz/z_T) - \exp(-bcz/z_T)] \qquad [10]$$

where a, b, and c are arbitrarily chosen parameters that primarily affect the value of $K_{(z)}$, and the ratio of the maximum diffusivity to diffusivity at the top of the Ekman layer, z_T. The above parameters were determined by Lewis (5) in different stability conditions for the lower 10 m of the experimental site with z_T=2400 m. In this investigation the values of $K_{(z)1}$, $K_{(z)2}$, and $K_{(z)3}$ were computed using Eqs. [3], [8], and [10], respectively.

The amounts of specific humidity q in Eq. [1] were measured at heights of 0.6 and 3.0 m through:

$$e = e_s - \gamma(T_d - T_w) \qquad [11]$$

where e_s is the saturation vapor pressure in mbar, γ the psychrometric constant in mbar /°C, and T_d and T_w dry and wet-bulb temperatures in °C, respectively. T_d and T_w are measured by the described thermocouple psychrometer.

The psychrometric constant could be computed as:

$$\gamma = (0.24p_a)/(0.622\lambda) \qquad [12]$$

where p_a=1013 - 0.1055z in mbar and λ=595 - 0.51T_w (°C) is the latent heat of vaporization in cal/g. Considering altitude of the experimental site, amounts of actual vapor pressure in Eq. [11] could be computed for different heights, and then:

$$q = 0.622\ e/Pa \qquad [13]$$

where q is specific humidity in grams of water vapor per grams of moist air.

The air density ρ in g/cm^3 could be found by the following equation (6):

$$\rho = 3.4838 \times 10^{-4}(P_a/T_v) \qquad [14]$$

where $T_v = T_d$ (1+0.61 q) is the virtual temperature in °K.

In this investigation the hourly data on September 20, 1976 were used for evaluation of the profile method in calculation of evapotranspiration.

RESULTS AND DISCUSSION

Table 1 shows the micrometeorological data on September 20. The long time intervals indicate the time between launching the balloons up to 200 m and bringing them down to the ground, and the short ones show the time for the next launching. The data show that the night time inversion vanishes a few hours after sunrise and appears around sunset, and that the wind speed is maximum around midafternoon and diminishes near sunset in the lower 10 m of the air layer in the valley atmosphere.

Having the data in Table 1, and using Eqs. [2] through [14], the values of $u_{*(0)}$, ϕ_m, $K_{(z)}$, q, and ρ were computed (see Table 2). $K_{(z)}$ and ρ are computed at screen height at 2 m. The results in Table 2 show the stability ($\phi_m > 1$) of the air layer above the experimental site changes to unstability ($\phi_m < 1$) a few hours after sunrise and that the stability begins around sunset due to the lack of solar radiation. The time dependent amounts of $K_{(z)}$ and specific humidity reach maximum around midafternoon.

The results shown in Table 3 indicate that the downward transfer of water vapor (negative value of E_t) continues about an hour after sunrise; this downward transfer leads to condensation and consequently to formation of dew. This table also indicates that, E_{t2} (upward transfer of water vapor) is very low compared with the potential evapotranspiration (E_{tp}) obtained form the class A pan evaporation on September 20 which was equal to 0.85 x 3.8=3.2 mm. The value 0.85 is called pan coefficient which is obtained from the proper table for the ground cover, level of mean relative humidity, and mean daily wind speed of the

Table 1. Micrometeorological data on September 20.

Time interval MST	T_d (°C) 0.6 m	T_d (°C) 3.0 m	T_w (°C) 0.6 m	T_w (°C) 3.0 m	$\bar{u}$ (cm/sec) 0.6 m	$\bar{u}$ (cm/sec) 3.0 m	$\bar{u}$ (cm/sec) 10.0 m
05:26-06:47†	1.9	2.0	1.7	1.8	10	63	79
06:47-06:57	2.3	3.1	2.4	2.7	28	63	109
06:57-08:40	7.1	7.3	5.8	5.8	31	80	97
08:40-09:07	11.9	11.6	9.1	8.8	12	57	64
09:07-10:34	14.4	13.9	10.6	10.1	44	87	97
10:34-11:07	15.9	15.3	11.3	10.9	78	123	140
11:07-12:38	17.4	16.8	12.2	11.9	105	154	179
12:38-13:04	18.9	18.6	13.4	13.1	128	180	206
13:04-14:32	21.1	19.9	14.5	13.9	108	160	182
14:32-14:45	23.2	21.3	15.0	13.9	102	161	177
14:45-15:59	23.1	22.2	14.0	13.4	78	141	164
15:59-16:28	22.2	21.5	13.2	12.6	75	134	164
16:28-17:36	21.4	20.1	12.5	11.5	96	151	183
17:36-18:01	19.2	18.8	11.1	10.9	52	106	139
18:01-19:17§	16.6	17.2	10.0	10.4	21	78	105

†**Sunrise at 05:43.**

§**Sunset at 18:01.**

Table 2. Amounts of ϕ_m, $u_{*(0)}$, $K_{(z)}$, q, and ρ based on Eqs. [2] to [13].

Time interval MST	ϕ_m	$u_{*(0)}$ (cm/sec)	$K_{(z)1} x10^{-3}$† (cm^2/sec)	$K_{(z)2} x10^{-3}$§ (cm^2/sec)	$K_{(z)3} x10^{-3}$¶ (cm^2/sec)	$qx10^3$ (g/g) 0.6m	$qx10^3$ (g/g) 3.0m	$\rho x 10^3$ (g/cm^3)
05:26-06:47	1.14	16.3	1.14	1.17	0.15	4.88	4.91	1.097
06:47-06:57	1.40	15.8	0.77	0.92	0.15	5.13	5.20	1.093
06:57-08:40	1.28	14.8	0.84	0.94	0.15	6.22	6.16	1.075
08:40-09:07	0.76	15.0	2.18	1.61	4.54	7.23	7.07	1.058
09:07-10:34	0.65	13.6	2.85	1.71	4.54	7.62	7.38	1.049
10:34-11:07	0.60	13.4	3.05	1.82	4.54	7.70	7.53	1.043
11:07-12:38	0.69	14.2	2.88	1.68	4.54	8.13	7.92	1.037
12:38-13:04	0.75	14.8	2.59	1.61	4.54	8.84	8.55	1.031
13:04-14:32	0.53	15.3	5.17	2.36	4.54	9.28	8.96	1.026
14:32-14:45	0.51	18.4	6.36	2.95	4.54	8.95	8.45	1.022
14:45-15:59	0.68	19.3	3.82	2.32	4.54	7.84	7.55	1.020
15:59-16:28	0.75	16.7	2.94	1.82	4.54	7.25	6.66	1.022
16.28-17:36	0.54	15.1	5.28	2.28	4.54	6.73	6.24	1.027
17:36-18:01	0.72	14.6	2.92	1.65	4.54	6.22	6.14	1.033
18:01-19:17	1.05	16.3	1.45	1.27	0.15	6.19	6.28	1.038

† $K_{(z)_1}$ is obtained using Eq. [3].

§ $K_{(z)2}$ is obtained using Eq. [8].

¶ $K_{(z)3}$ can be calculated using Eq. [10] for two **different** conditions. For stable conditions: a=-1.04x10^6 cm^2/sec, b=2.085, c=0.92, and for **unstable** conditions: a=4x10^7 cm^2/sec, b=1.933, and c=1.07.

Table 3. Calculated values of E_{t1}, E_{t2}, and E_{t3} based on Eq. [1].

Time interval MST	E_{t1}x10 (mm)	E_{t2}x10 (mm)	E_{t3}x10 (mm)
05:26-06:47	-0.08	-0.07	-0.01
06:47-06:57	-0.02	-0.02	-0.01
06:57-08:40	0.14	0.16	0.03
08:40-09:07	0.25	0.18	0.52
09:07-10:34	1.56	0.94	2.49
10:34-11:07	1.67	0.27	0.66
11:07-12:38	1.43	0.83	2.25
12:38-13:04	1.28	0.31	0.89
13:04-14:32	2.56	1.70	3.28
14:32-14:45	1.06	0.49	0.75
14:45-15:59	2.09	1.27	2.48
15:59-16:28	1.29	0.80	1.98
16:28-17:36	4.53	0.20	3.89
17:36-18:01	1.51	0.85	2.34
18:01-19:17	-0.26	-0.23	-0.03
Total	19.4	8.0	21.6

experimental site, and 3.8 is the class A pan evaporation in mm on the mentioned day. Comparing E_{t1} and E_{t3} with E_{tp} yields crop coefficients of 194/3.2=0.61 and 2.16/3.2=0.68, repectively, which seem reasonable crop coefficients for this period of year in this region. This investigation confirms the validity of the profile method in calculation of evapotranspiration.

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