Journal of Hydroscience and Hydraulic Engineering
Vol. 7, No. 2, March, 1990, 27-38

A BASIC STUDY ON ESTIMATION OF REGIONAL EVAPOTRANSPIRATION USING REMOTELY SENSED DATA

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SYNOPSIS

In order to develop a new method for estimating regional evapotranspiration rate using remotely sensed data, a preliminary investigation has been done. The method has the following advantages: (1) the estimation procedure is reasonably simple - based on a pixel or an element of divided area, (2) surface soil parameters such as heat conductivity or thermal inertia and aerodynamic parameters such as roughness length or bulk diffusion coefficient, which have spatial variability in general, need not be assumed, and (3) daily averaged values of evapotranspiration can be calculated from a couple of instantaneous, remotely sensed images.

Results of the investigation show the method has good potential. The results also can provide knowledge about more effective ways of data acquisition in present and future operations of satellite observations.

INTRODUCTION

More precise estimation of regional evapotranspiration (actual evapotranspiration) is needed for evaluating a regional or global hydrologic cycle. At the regional scale, potential evapotranspiration can be evaluated from classical meteorological observations. Extension to actual evapotranspiration is more difficult and assumes the existence of a regional equilibrium regime [Bouchet, 1963; Priestry and Taylor, 1972; Brutsaert and Stricker, 1979; Otsuki et al., 1984a,b] or is based on the planetary boundary layer similarity theory [Brutsaert and Mawdsley, 1976; Abdulmumin et al., 1987] or includes soil and vegetation parameters which account for the more or less strong heterogeneities of the ground surface.

With the development of remote sensing techniques well adapted to regional scale observations, many attempts have been made to use remote sensing data in place of meteorological data. The recent representative attempts using remotely sensed data can be divided into two types of approaches. There are those based on a water balance in the ground surface layer and those based on an energy balance at the ground surface.

As regards the water balance method, remotely sensed surface soil moisture is used for solving the Richards's equation as a boundary condition. Bernard et al.[1981] examined this point by using simulated microwave data to measure soil moisture. Camillo et al.[1983, 1984] similarly simulated surface temperature taking into account the interaction between the atmosphere and the ground surface layer, and compared the temperature simulated by a numerical model for estimating evapotranspiration with the temperature measured by the Heat Capacity Mapping Mission (HCMM). They have also developed a method for estimating soil hydraulic parameters [Camillo et al., 1986]. However, these approaches do not seem to be the best way for estimating evapotranspiration on a pixel-by-pixel basis because the hydraulic parameters of surface soil must be determined exactly at each pixel in various surface conditions.

On the other hand, the energy balance method seems to be more simple and effective for the estimation on the pixel-by-pixel basis. Based on the energy balance, Price[1980] has investigated the possibility of using data by the HCMM and Soer[1980] has combined thermal infrared data from an airborne scanner with ground-based measurement. Kotoda et al. [1984] showed an example of an integrated system for estimating regional ($13{,}300km^2$) evapotranspiration. They used Landsat data and the Priestly & Taylor equation. Reginato et al. [1985] estimated instantaneous and daily values of evapotranspiration for 44 cloudless days using Barnes Modular Multiband Radiometer(MMR). Nevertheless, most of the present methods based on energy balance need the aerodynamic parameters such as roughness length for the bulk diffusion coefficient. Moreover, soil parameters such as heat conductivity should be indispensable for estimating instantaneous evapotranspiration.

These parameters have spatial variability in general. The estimation of parameters in a great many sub-area (pixels) which comprise a region is difficult and uncertain. Therefore, a new estimation procedure which avoids the cumbersome steps described above is desired.

In this paper, the daily energy balance at the ground surface is considered first. Next, a new formula for estimating daily averaged net radiation and correction terms in the energy balance equation are examined. Finally, a preliminary study for developing a new method of evapotranspiration using satellite measurement data is presented and the results of the study are discussed.

ENERGY BALANCE AVERAGED OVER 24 HOURS

a. Energy Balance at the Ground Surface

The balance between incoming and outgoing energy fluxes at the ground surface is expressed through the following equation neglecting energy consumption in the surface plant.

$$R_n = H + LE + G \tag{1}$$

where R_n is net radiative flux (net radiation), H is sensible heat flux, LE is latent heat flux and G is soil heat flux.
The following expressions are assumed for the fluxes [Sellers, 1965].

$$R_n = R_S + R_L - R_e \tag{2}$$

where R_S is absorbed solar radiation, R_L is absorbed longwave radiation and R_e is emitted longwave radiation.
R_S, R_L and R_e are given by the following equations.

$$R_S = SV(1-\alpha)(\sin\delta\sin\phi + \cos\delta\cos\phi\cos\omega) \tag{3}$$

$$R_L = \varepsilon\sigma T_a{}^4 (a + b\sqrt{e_a}) \quad \text{(Brunt)} \tag{4}$$

$$R_L = \varepsilon\sigma T_a{}^4 \times 0.533 \cdot e_a{}^{1/7} \quad \text{(Brutsaert)} \tag{5}$$

$$R_e = \varepsilon\sigma T_S{}^4 \tag{6}$$

where S is solar constant, V is atmospheric transmittance, α is surface albedo, δ is solar declination, ϕ is latitude of the point observed, ω is solar time angle, ε is surface emissivity, σ is Stefan-Boltzmann constant, a and b are constants, e_a is atmospheric vapor pressure, T_a is air temperature and T_s is surface temperature.

$$H=\gamma C_1 U(T_s-T_a) \tag{7}$$

$$LE=C_1 U(e_s-e_a) \tag{8}$$

where U is wind speed, γ is psychrometric constant, e_s is surface vapor pressure and C_1 is bulk diffusion coefficient for neutrally stable atmospheric conditions. C_1 is expressed as

$$C_1=\frac{\rho C_p \kappa^2}{\gamma \ln^2\{(z-d)/z_0\}(1+r_s/r_a)} \tag{9}$$

where ρ is density of air, C_p is heat capacity of air, κ is von Karman's constant, z is height of measuring, z_0 is roughness length, d is displacement height, and r_a and r_s are aerodynamic and stomatal resistance, respectively. The aerodynamic resistance is given as

$$r_a=\frac{\ln^2\{(z-d)/z_0\}}{\kappa^2 U} \tag{10}$$

If we can assume that r_s/r_a is either small or a certain constant, approximately, in a day, eq. (9) gives

$$C_1=\frac{\rho C_p \kappa^2}{\gamma\ c \ln^2\{(z-d)/z_0\}} \tag{11}$$

where c is a constant ($c \geq 1$).
And,

$$G=\lambda_1 (T_1-T_s)/z_1 \tag{12}$$

where λ_1 is thermal conductivity of the ground surface layer, T_1 is temperature at the center of the surface layer and z_1 is depth to the center of the surface layer.

b. Energy Balance Averaged over 24 hours

Daily averaged evapotranspiration may be one of the most basic factors for evaluating the water budget in a watershed. By time averaging (1), a simple result can be obtained. Assuming that no net heating or cooling of the earth occurs over a 24-hour period, the average of G over 24 hours vanishes. Therefore, one finds

$$\langle R_n\rangle=\langle H\rangle+\langle LE\rangle \tag{13}$$

where $\langle\ \rangle$:24-hour average.

The most important factor in the energy balance at the ground surface is net radiation. And absorbed solar radiation R_s is dominant in equation (2). Equation (3) for R_s is used in the case of clear sky. For more general estimation, it is necessary to take into account the effect of cloud cover on R_s averaged over sunshine hours. If we use a correction factor based on the cloud index and cloud albedo, which are daily averaged values, averaged R_s can be expressed approximately as

$$\begin{aligned}\langle R_s\rangle &= \frac{1}{24} SV(1-\alpha)(1-C_d\alpha_c)\int_{-\omega_0}^{\omega_0}(\sin\delta\sin\phi+\cos\delta\cos\phi\cos\omega)\frac{12}{\pi}\,d\omega \\ &= \frac{1}{\pi} SV(1-\alpha)(1-C_d\alpha_c)(\omega_0\sin\delta\sin\phi+\cos\delta\cos\phi\sin\omega_0)\end{aligned} \tag{14}$$

where $\omega=\left(\frac{t-12}{12}\right)\pi$, t is time (hour),

$\omega_0=\cos^{-1}(-\tan\phi\tan\delta)$, C_d is cloud index (0 ~ 1.0) and α_c is cloud albedo.
and

$$\langle R_L\rangle=\langle\varepsilon\sigma T_a{}^4(a+b\sqrt{e_a})\rangle \tag{15}$$

If we express T_a and e_s as $T_a=\langle T_a\rangle + T_a'$ and $e_a=\langle e_a\rangle + e_a'$ respectively, when $\langle T_a\rangle \gg T_a'$, $\langle e_a\rangle \gg e_a'$, the following expression is obtained.

$$\langle R_L\rangle=\varepsilon\sigma\langle T_a\rangle^4(a+b\sqrt{\langle e_a\rangle}) \tag{16}$$

Similarly,

$$\langle R_e\rangle=\varepsilon\sigma\langle T_s\rangle^4 \tag{17}$$

Figure 1 illustrates the results of the calculation using the formula of Brunt (a=0.605, b=0.048 and ε=0.95 [Sellers, 1965]) or Brutsaert. Focusing on

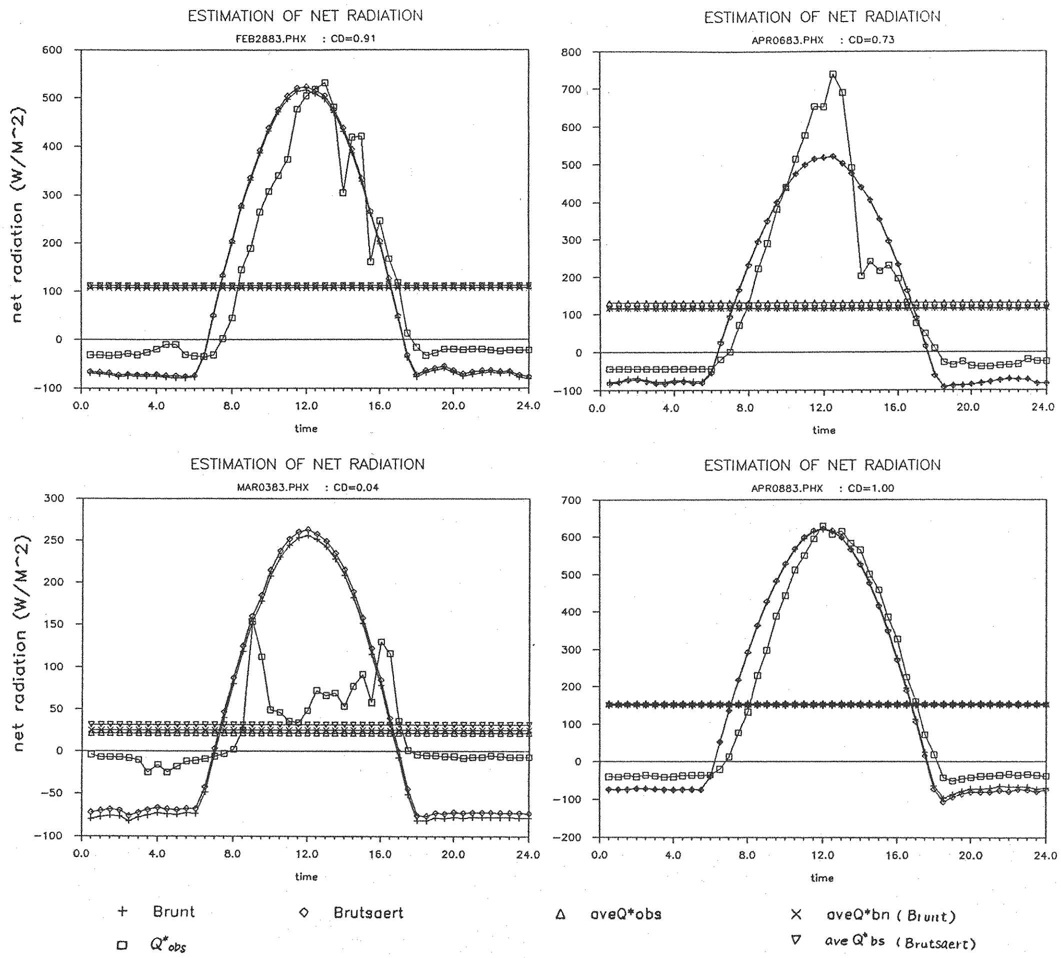

Fig. 1 Daily averaged net radiation (aveQ*) estimated by the developed formula.

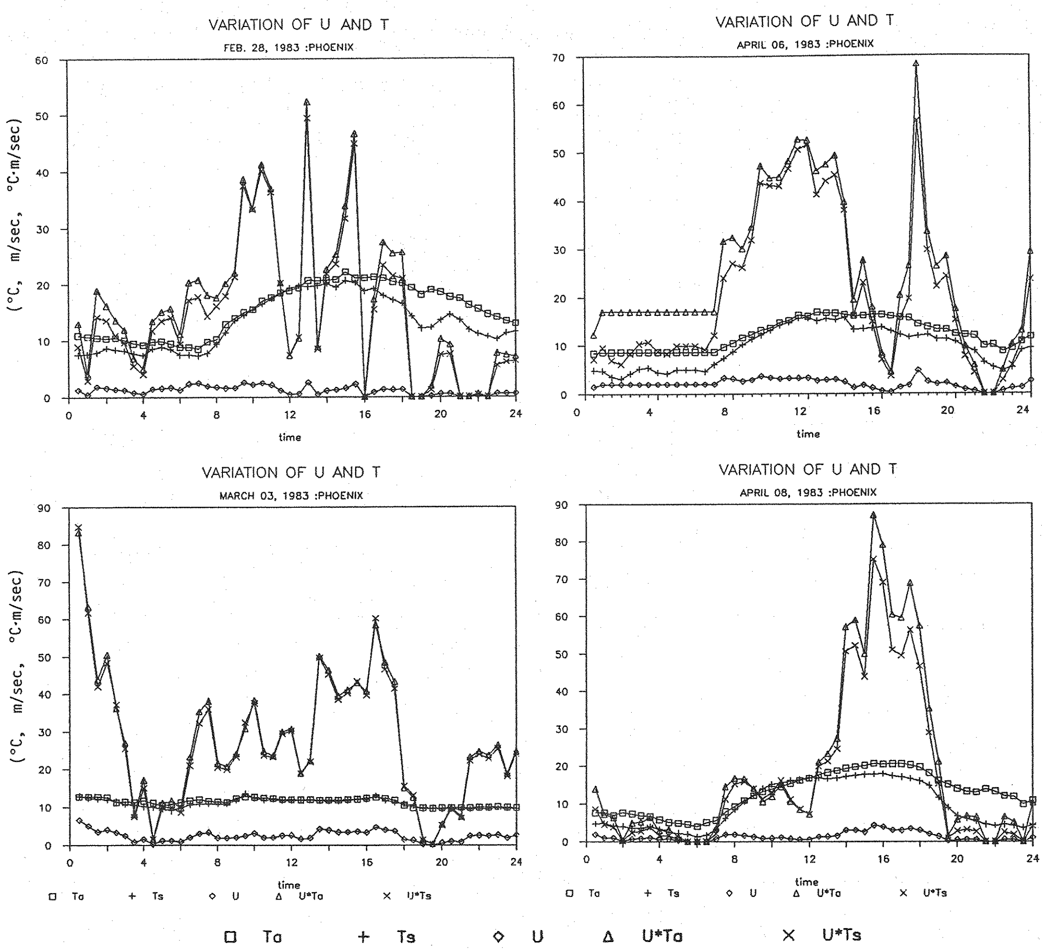

Fig. 2 Diurnal variations of U, T_a, T_s, UT_a and UT_s.

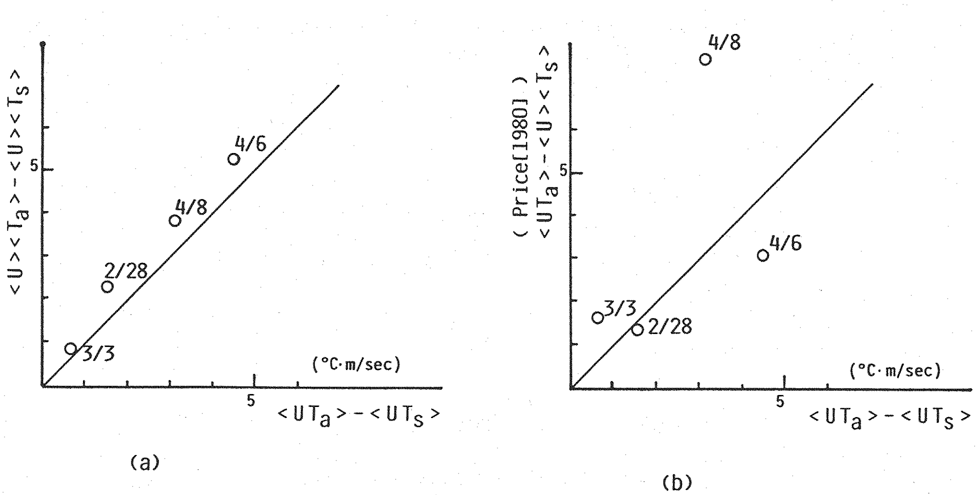

Fig. 3 Simplification of $\langle UT_a \rangle - \langle UT_s \rangle$.

the averaged net radiation (aveQ*), instantaneous values are not of concern to us now, the estimations(symbol: x or ▿) show good agreement with the values(symbol: Δ) observed at the experimental field in Arizona. The conditions of the field and the observation is described later.

For the first term of the right hand side of eq. (13),

$$\langle H\rangle=\langle\gamma C_1 U(T_s-T_a)\rangle=\gamma C_1\{\langle UT_s\rangle-\langle UT_a\rangle\} \tag{18}$$

Diurnal variation of UT_s and UT_a of the observed data are exemplified in Figure 2. In order to carry out a simplification of $\langle UT_s\rangle-\langle UT_a\rangle$, $\langle UT_a\rangle-\langle UT_s\rangle$ is examined in Figure 3. Price [1980] has taken $\langle UT_a\rangle-\langle U\rangle\langle T_s\rangle$ as $\langle UT_a\rangle-\langle UT_s\rangle$ (Figure 3(b)). From Figure 3(a), one can adopt the following relation as a first approximation.

$$\langle UT_a\rangle-\langle UT_s\rangle=\langle U\rangle\langle T_a\rangle-\langle U\rangle\langle T_s\rangle \tag{19}$$

Therefore,

$$\langle H\rangle=\gamma C_1\{\langle U\rangle\langle T_s\rangle-\langle U\rangle\langle T_a\rangle\} \tag{20}$$

In order to use this equation under various conditions in general, the simplification described above may be examined further.

ESTIMATION OF EVAPOTRANSPIRATION USING SATELLITE MEASUREMENT DATA

a. Estimation Procedure

Assuming that a couple measurements of surface temperature (Ts) in a day can be given by an observing satellite [Price, 1980; Lo, 1986; Barrett, 1988], HCMM for example, and that diurnal variation of T_s is approximated by a sine-wave, the following simple procedure for estimating evapotranspiration may be developed.

Soil heat flux for the boundary condition of sine-wave temperature at the ground surface is expressed [Sellers, 1965] as

$$G=\Delta T_0(\omega_1 C\lambda)^{1/2}\sin(\omega_1 t+\frac{\pi}{4}) \tag{21}$$

where ΔT_0 is amplitude of the surface temperature wave, C is heat capacity of soil, λ is thermal conductivity of soil, ω_1 is angular frequency of oscillation ($\pi/12$) and $(\omega_1 C\lambda)^{1/2}$ is thermal inertia.

For a pair of data which are provided by local meteorological observations, U, T_a, e_a, and by remote sensing, T_s, in a day,

$$\begin{aligned} R_{n1}&=H_1+LE_1+G_1\\ R_{n2}&=H_2+LE_2+G_2 \end{aligned} \tag{22}$$

where suffix 1 and 2 mean times of observing t_1 and t_2 respectively.
If we have a set of data of 12-hour time interval, $G_2=-G_1$.
Assuming neutrally stable atmospheric conditions, eq. (22) gives

$$\begin{aligned} R_{n1}+R_{n2}=C_1[&\gamma\{U_1(T_{s1}-T_{a1})+U_2(T_{s2}-T_{a2})\}\\ &+\{U_1(e_{s1}-e_{a1})+U_2(e_{s2}-e_{a2})\}] \end{aligned} \tag{23}$$

If we know e_{s1} and e_{s2}, we can determine C_1. We can also obtain the thermal inertia using eqs. (21) and (22) after determination of C_1 (ΔT_0 is assumed to be given).

The value of e_s can be determined by microwave measurement [Jackson et al., 1983,1986 and 1988] or surface temperature difference and soil characteristics [Schmugge et al., 1980 for example]

In the case of atmospheric unstable conditions, we can use the following forms of equations:

$$R_{n1} = K_e (e_{s1} - e_{a1}) + K_h (T_{s1} - T_{a1}) + G_1$$
$$R_{n2} = K_e (e_{s2} - e_{a2}) + K_h (T_{s2} - T_{a2}) + G_2 \qquad (24)$$

where K_e is apparent diffusion coefficient of vapor and K_h is apparent diffusion coefficient of heat.

After determining G_1 (G_2) using eq. (22), K_e and K_h can be calculated.

b. Results of the Estimation and Discussion

The method was applied to the winter wheat field of U S water Conservation Laboratory, Phoenix, Arizona (33deg26'N, 112deg01'W). The elevation of the field is 345m. The equipment was located near the center of the field 62m(N-S) by 78m(E-W). Local meteorological conditions (R_n(Q* in the Figures), T_a, U and e_a), and ground surface temperature (T_s) by infrared thermometry were observed, and sensible and latent heat fluxes were measured by the Bowen ratio method. T_a and e_a were measured at 0.85m height (about 0.5m above the canopy) on days 59 (day of year) and 62, and at 1.45m (about 0.6m above the canopy) on days 96 and 98. Two days of data on days 59 (Feb. 28) and 98 (April 8), 1983 have been used. Figure 4 illustrates the bulk diffusion coefficient (C_1) calculated from eq. (23).

In this application, data of e_s were estimated by an indirect method using the maximum surface temperature difference.
First, the maximum temperature difference (ΔT_s) is given by using sin-wave approximation of the surface temperature. The volumetric soil water content of the ground surface (θ) may be estimated by the following empirical relation between θ and ΔT_s [Schumugge et al., 1980].

$$\Delta T_s = 42.7 - 83.7\theta \qquad (25)$$

The matric potential ψ can be calculated from the following expression.

$$\psi = \psi_s (\theta / \theta_s)^{-m} \qquad (26)$$

where θ_s and ψ_s are volumetric soil water content and matric potential at saturation, respectively, and m is constant determined by the soil texture (m=4 was used in this study). The standard relative humidity model as a function of both moisture and temperature is as follows [Camillo et al., 1983]:

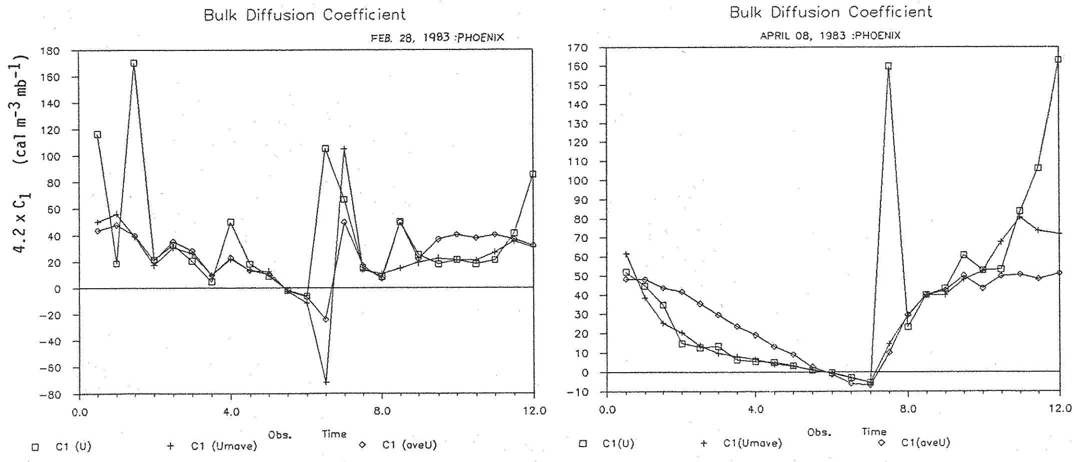

Fig. 4 Calculated bulk diffusion coefficient.

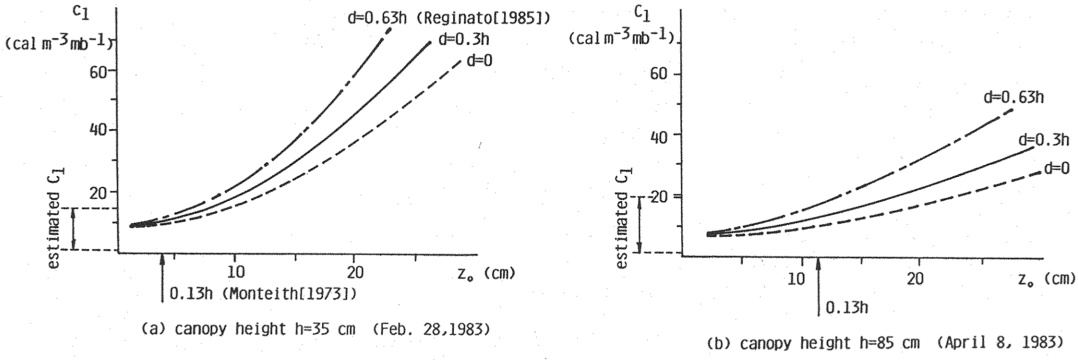

Fig. 5 Comparison of the bulk diffusion coefficient between the calculated values and the physically expected ones.

$$h=\exp[(\psi g)/(RT_s)] \tag{27}$$

where h is the relative humidity, g is gravitational acceleration and R is the gas constant. Then, the actual surface vapor pressure is computed from

$$e_s=h\cdot e_{sat} \tag{28}$$

where e_{sat} is the saturation vapor pressure at the surface temperature T_s.

In Figure 4, "obs. time" means the paired observation with a 12-hour time interval. For example, observation time 4 indicates that C_1 is found from the data at times of 04:00 and 16:00. Variation of instantaneous wind speed (U) is very intense, therefore, two kinds of averaged speeds are also examined. One is the daily averaged wind speed (aveU) as a reference, the other is the moving averaged wind speed over each two hours (Umave). Adoption of the averaged speeds (Umave) seems to be more reasonable than the case of the instantaneous wind speed and to keep the values of C_1 more stable except from 06:00 to 07:00 when the surface temperature cross its mean value.

The bulk diffusion coefficient can be determined also by the conditions of the ground surface. Comparison of the bulk diffusion coefficient between the

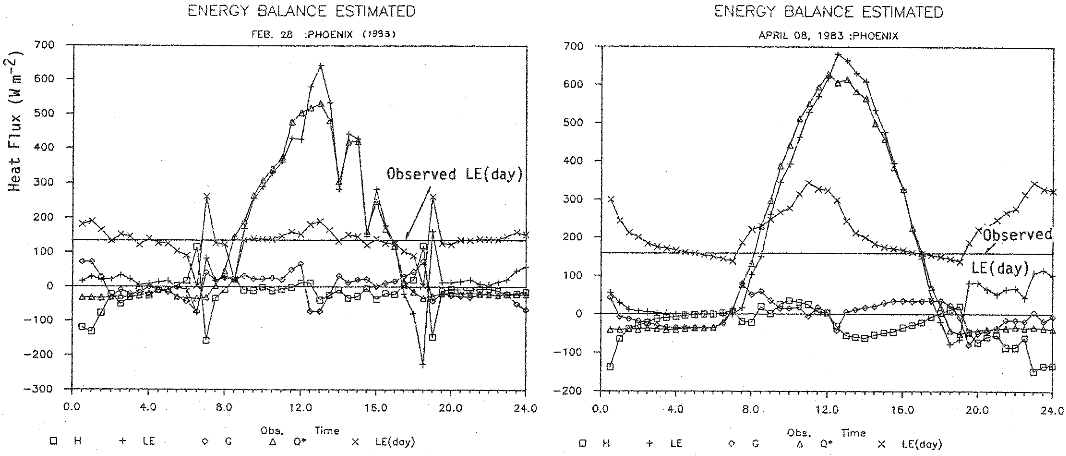

Fig. 6 The energy balance estimated by using C_1.

calculated value (range) and the physically expected value using the parameters of displacement and roughness length proposed by Reginato et al.[1985] are illustrated in Figure 5. The range of the C_l values so calculated is in agreement with the physically estimated region. The result of the calculation by eq. (23) seems to be reasonable.

Figure 6 represents the estimated energy balance. In the Figure, LE(+) shows the estimated instantaneous latent heat (evapotranspiration) at each observation time. And, LE'(day) (×) expresses the daily average evapotranspiration estimated from the daily averaged values of air temperature, vapor pressure and wind speed and the C_l by using the data given at each observation time. Comparing the results of LE'(day) (×) with the observed average latent heat, LE(day): solid straight line, the estimation shows good agreement with the observed data in certain time zones. From the graph, the most effective observation times for estimating evapotranspiration may be from 2:00(&14:00) to 5:00(&19:00).

Figure 7 illustrates the diffusion coefficients, K_e and K_h, calculated from eq. (24). The results give suggestion that data observed from 5:00(&17:00) to

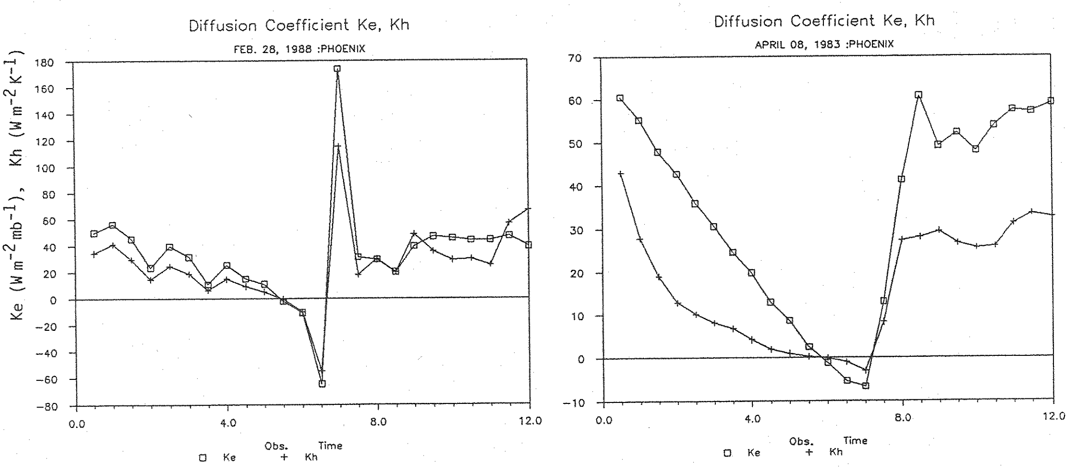

Fig. 7 Calculated diffusion coefficient K_e and K_h.

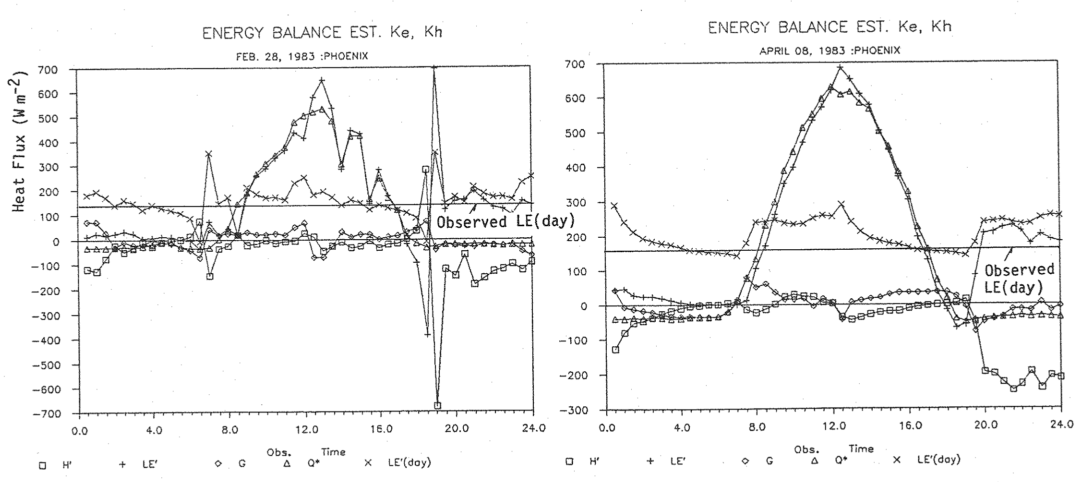

Fig. 8 The energy balance estimated by using K_e and K_h.

7:00(&19:00) is inappropriate for the estimation. Figure 8 shows the energy balance estimated by using K_e and K_h. The estimation of LE seems to be slightly more stable than the previous case by C_1; however, the trend of the graph is similar to the previous ones.

CONCLUDING REMARKS

Results of this preliminary investigation show that the method discussed above offers a promising new way of estimating regional evapotranspiration. The method has the following advantages: (1) the estimation procedure is reasonably simple - based on a pixel or an element of divided area, (2) surface soil parameters such as heat conductivity or thermal inertia and aerodynamic parameters such as roughness length need not be assumed at each pixel, and (3) daily averaged values of evapotranspiration can be calculated from a couple of instantaneous remotely sensed images. At present stage, the method is to be used under the condition that the plants water stress is low. The results of the applications, however, may indicate some information for more effective data acquisition of earth surface measurement. To establish the method, further study should be continued to verify it with some other conditions of the ground surface.

ACKNOWLEDGMENTS. The authors would like to ackowledge the administrative support from Prof. N. Buras, the preceding Head of the Department of Hydrology and Water Resources, University of Arizona. They also wish to ackowledge to Mr. J. Washburne, Graduate Student of the Department, for his excellent computational assistance.

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APPENDIX - NOTATION

The following symbols are used in this paper:

C =heat capacity of soil;
C_1 =bulk diffusion coefficient for neutrally stable atmospheric conditions;
C_d =cloud index ($0 \sim 1.0$);
C_p =heat capacity of air;
d =displacement height;
e_a =atmospheric vapor pressure;
e_s =surface vapor pressure;
e_{sat} =saturation surface vapor pressure;

G	=soil heat flux;
g	=gravitational acceleration;
H	=sensible heat flux;
h	=relative humidity;
K_e	=apparent diffusion coefficient of vapor;
K_h	=apparent diffusion coefficient of heat;
LE	=latent heat flux;
R	=gas constant;
R_e	=emitted longwave radiation;
R_L	=absorbed longwave radiation;
R_n	=net radiative flux (net radiation);
r_a	=aerodynamic resistance;
r_s	=stomatal resistance;
S	=solar constant;
t	=time (hour);
T_a	=air temperature;
T_s	=surface temperature;
T_1	=temperature at the center of the surface layer;
ΔT_0	=amplitude of the surface temperature wave;
ΔT_s	=maximum surface temperature difference;
U	=wind speed;
V	=atmospheric transmittance;
z	=height of measuring;
z_0	=roughness length;
z_1	=depth to the center of the surface layer;
α	=surface albedo;
α_c	=cloud albedo;
γ	=psychrometric constant;
δ	=solar declination;
ε	=surface emissivity;
θ	=volumetric soil water content;
θ_s	=volumetric soil water conttent at saturation;
κ	=von Karman's constant;
λ	=thermal conductivity of soil;
λ_1	=thermal conductivity of the surface layer;
ρ	=density of air;
σ	=Stefan-Boltzmann constant;
ϕ	=latitude of the point observed;
ψ	=matric potential of soil water;
ψ_s	=matric potential of soil water at saturation;
ω	=solar time angle;
ω_0	=solar angle at sunrise and sunset;
ω_1	=angular frequency of oscillation($\pi/12$); and
$\langle\ \rangle$	=operator of 24-hour averaging.