Estimating surface moisture availability for evaporation on bare soil from routine meteorological data and its parameterization without soil moisture

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Abstract

Surface moisture availability is the most important parameter for estimating evaporation from bare soil using the bulk transfer method. In this study, we developed a simple method for the estimation and parameterization of surface moisture availability over a period ranging from 10 days to several months on bare soil using only routine meteorological data without using sensible and latent heat flux and soil moisture data. If the daily mean soil temperature is observed with other routine meteorological data of daily values (air temperature, solar radiation, precipitation, wind speed, humidity, etc.), surface moisture availability can be estimated by using a simple soil temperature model without using the sensible and latent heat flux observation data. The simple soil temperature model is composed of the heat balance equation at the ground surface, bulk transfer method and modified version of the force-restore model for estimating daily mean soil temperature. The estimation period of surface moisture availability ranges from 10 days to several months according to precipitation conditions that reflect soil wetness. The antecedent precipitation index (*API*) is an effective indicator by which the soil wetness condition can be judged without using soil moisture. This enables classification of proper estimation periods of surface moisture availability. The surface moisture availability is estimated by minimizing the sum of the squares of the residual between the observed value of the daily mean soil temperature and the calculated value under the estimation period as classified by *API*. Furthermore, for the purpose of estimating daily mean soil temperature and cumulative evaporation on bare soil using only routine meteorological data, we propose that the surface moisture availability can be parameterized using the ratio of precipitation to potential evaporation with estimation periods ranging from 10 days to several months.

Key words: Bare soil, Evaporation, Parameterization, Soil temperature, Surface moisture availability.

1. Introduction

Evaporation from the soil surface is an important factor for energy and water balance over land surfaces. Surface moisture availability is the most important parameter for either estimating evaporation from bare soil using the bulk transfer method or for estimating soil temperature using the combined heat balance equation at the ground surface and the bulk transfer method (e.g., Kondo, 1994). A method for estimating

evaporation using the bulk transfer formula is expressed as the following:

$$E = \rho \beta C_H U (q_{sat}(T(0, t)) - q_a) \quad (1)$$

where E is the evaporation (mm s^{-1}), C_H is the bulk transfer coefficient for sensible heat flux (dimensionless number), U is the wind speed (m s^{-1}), $C_H U$ is the exchange speed for sensible heat flux (m s^{-1}), q_a is the specific humidity of the air (kg kg^{-1}), $q_{sat}(T(0, t))$ is the saturation specific humidity at the ground surface temperature (kg kg^{-1}) and time t , ρ is the air density

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(kg m^{-3}), and β is the surface moisture availability, which ranges from 0 to 1. Equation (1) is called the β method.

Modeling evaporation by the bulk transfer method has the advantage of a simple equation form; it is direct and makes use of routine meteorological data easily measured in situ or remotely sensed data to calculate the potential evaporation (when $\beta=1$). According to Kondo *et al.* (1990), Mahfouf and Noihan (1991) and Lee and Pielke (1992), evaporation from bare soil strongly depends on surface soil moisture content. Therefore, β is usually treated as a function of the volumetric water content of the upper few centimeters. Ye and Pielke (1993) proposed the parameterization formula for β using not only surface soil moisture content but also more detailed soil physical conditions.

However, the bulk transfer method has some disadvantages. Experimental estimation of β from Eq. (1) under in-situ field conditions for calibration or verification of β usually requires the observed latent heat flux or evaporation, ground surface temperature and exchange speed $C_H U$ (or bulk transfer coefficient C_H and wind speed U), in addition to routine meteorological data. Therefore, the bulk transfer method requires detailed, site-specific measurements or the inclusion of sensible and latent heat flux for the initial parameterization or verification of β . The bulk transfer method thus requires expensive instrumentation along with that of data logging capability and, as such, is limited to only well-monitored research locations. In addition, even if β can be parameterized using surface soil moisture content, it is seldom monitored on a routine basis as with other atmospheric variables. Before the bulk transfer method can be used as a practical tool in various disciplines such as hydrology, climatology or agronomy, the method for determining surface moisture availability β needs to be greatly simplified.

Studies have been conducted on estimating surface moisture availability β from routine meteorological data (air temperature, solar radiation, precipitation, wind speed and humidity) and ground surface temperature without using flux observation data. For example, Kimura and Shimizu (1994) developed a linear heat budget model to estimate diurnal variation in latent and sensible heat flux and surface moisture availability on bare soil from the diurnal variation in the ground surface temperature and downward radiation, when the bulk transfer coefficient for sensible heat is given.

Matsushima and Kondo (1995) developed a linear heat budget model to simultaneously estimate daytime means of bulk transfer coefficient for sensible heat and surface moisture availability on bare soil using the daily variation in routine meteorological data and ground surface temperature. Furthermore, there is renewed interest in applying surface moisture availability to data assimilation for soil moisture through the surface heat budget model and ground surface temperature observation including the vegetation layer (Mahfouf, 1991; Lakshmi, 2000). Casteli *et al.* (1999) developed a robust method with respect to noise errors that is based on the methods developed by Mahfouf (1991) and Hu and Islam (1995). The objective of these studies was to estimate daily surface moisture availability and diurnal variation in latent and sensible heat flux. At this time scale, the accuracy of ground surface temperature is crucial; a systematic error of only $\pm 1^\circ\text{C}$ in ground surface temperature can lead to erroneous estimates of these variables (Matsushima and Kondo, 1995). In practice, precise ground surface temperature estimation is difficult to achieve and is generally far from within $\pm 1^\circ\text{C}$ from the satellite data (Lakshmi and Susskind, 2000).

Hirota *et al.* (1995) proposed a method for estimating cumulative evaporation and surface moisture availability β from 10 days to several tens of days using routine meteorological data and daily mean soil temperature. They also suggested that the surface moisture availability β may be parameterized using routine meteorological data such as precipitation without needing soil moisture values for the purpose of estimating cumulative evaporation and daily mean soil temperature. Their method offers the possibility of avoiding the problem of flux measurement, the difficulty of obtaining accurate ground surface temperature estimation and the need to use soil moisture data even though this application is limited to a longer time scale (more than 10 days, seasonal, annual or inter-annual).

We begin with a brief introduction of the method developed by Hirota *et al.* (1995) for estimating cumulative evaporation and averaged surface moisture availability value for a period ranging from 10 days to several tens of days using routine meteorological data and daily soil temperature values. Next, we explain how we improved our method in order to more efficiently estimate the averaged surface moisture availability. Lastly, we introduce the parameterization method for

surface moisture availability β without using soil moisture. The significance of this is that once we establish the parameterization for β , cumulative evaporation on bare soil for a long period (more than 10 days) can be estimated from routine meteorological data (air temperature, solar radiation, precipitation, wind speed and humidity) through the heat balance equation at the ground surface. Daily mean soil temperature can also be estimated, meaning that cumulative evaporation can be estimated without needing soil temperature after establishing parameterization of β . The estimation of daily mean soil temperature is also an especially useful and important factor for agriculture.

2. Overview of previous work in Hirota *et al.* (1995) and Hirota and Fukumoto (1996)

A method for estimating evaporation using a modified force-restore model (mFRM) for estimating daily mean soil temperature was developed by Hirota *et al.* (1995) from the original model by Bhumralkar (1975) (see Appendix for complete details). The modified version is for bare soil and uses routine meteorological data and daily soil temperature values (a few cm to a few tens of cm order). It is not always necessary to use the observed ground surface temperature (surface skin temperature or depth of a few mm order). This feature is an advantage because an accurate ground surface temperature is not easy to observe (Bittelli *et al.* 2008), as a large temperature gradient often exists near the surface in dry soil conditions (Saigusa *et al.* 1993) and it is also difficult to accurately estimate within $\pm 1^\circ\text{C}$ from the satellite data. The study, reported by Hirota *et al.* (1995), showed that if the daily mean soil temperature can be accurately simulated, then the cumulative evaporation can also be simulated, without the measurement of latent and sensible heat flux. Also, the surface moisture availability value β could remain constant for periods of 10 days or longer according to weather conditions (especially precipitation amount or soil wetness condition), meaning that the daily value of β , or that of measured soil moisture, is not needed.

The study also showed that although the effects of thermometric parameters (soil heat capacity c , damping depth D_a) on the calculation of diurnal variation in soil temperature are sensitive, those on the calculation of daily mean soil temperature or cumulative evaporation are relatively insensitive. Therefore, the determining of soil thermometric parameters does not need detailed parameter characteristics such as dependency on soil

moisture. These soil parameters can be treated as a constant value under each soil type. Also, if mFRM is used, the soil thickness ranges from several cm to several tens of cm, and hence it is not necessary to know the soil thermometric parameters of the deep layer.

The exchange speed $C_H U$ can be estimated from the roughness length for momentum, roughness length for sensible heat, zero plane displacement, wind speed measurement, and universal function of the stability parameter (or Richardson number). However, the roughness length for sensible heat is not always easy to correctly estimate because the value is affected by the scale of fetch even under the same surface conditions (*e.g.*, Kondo, 1994). Hirota and Fukumoto (1996) showed that the exchange speed for sensible heat $C_H U$ on bare soil can be estimated by using routine meteorological data without using observed sensible heat flux. If the soil surface is sufficiently wet after a day of precipitation, assuming that $\beta=1$ and daily mean soil heat flux is negligible, then $C_H U$ can be calculated by the following equation:

$$C_H U = \frac{(1-a)R_s + \varepsilon \{L_d - \sigma(273.15 + T_s)^4\}}{c\rho p \{T(0, t) - T_a\} + l\rho \{q_{sat}(T_s) - q_a\}} \quad (2)$$

where R_s is the solar radiation (W m^{-2}), L_d is the atmospheric long wave radiation (W m^{-2}), a is the albedo of soil surface, ε is the emissivity of the ground, σ is the Stefan-Boltzmann constant ($\text{W m}^{-2} \text{K}^{-4}$), l is the heat of vaporization (J kg^{-1}), q_a is the specific humidity of the air (kg kg^{-1}), $q_{sat}(T_s)$ is the saturation value of the specific humidity (kg kg^{-1}) at the ground surface temperature and T_s is the daily mean ground surface temperature ($^\circ\text{C}$). In this case, T_s can be the soil temperature at the 1-2 cm depth. The reason is that the temperature gradient of the daily mean value is not large in wet condition, then the ground surface temperature is not sensitively dependent on the soil depth for daily mean values under wet conditions (Hirota and Fukumoto, 1996).

Therefore, the method of Hirota *et al.* (1995) is valid for the heat balance at the ground surface because it provides reasonable $C_H U$. The two previous studies (Hirota *et al.* 1995; Hirota and Fukumoto, 1996) suggest that the surface moisture availability β can be easily estimated using only routine meteorological data without needing observed latent and sensible heat flux data. This is possible as other required

parameters (exchange speed for sensible heat flux, soil thermometric parameters) can be easily determined from the previous paragraphs in this section. At least a 10-day period is necessary for estimation of β because without the measurement of soil moisture, precipitation conditions are needed. Here, we concretely show the parameterization of surface moisture availability β over a long period (more than 10 days) without using soil moisture. In addition, before estimating β over a long period (more than 10 days), a proper calculation period should be determined using soil moisture status without using soil moisture. In the method of Hirota *et al.* (1995), the calculation period was determined through trial and error by comparing the observed soil temperature and estimated soil temperature. However, in this case, the calculation procedure is comparatively complex and a more efficient procedure is desirable. If the proper calculation period can be determined without using the trial and error method, the procedure for estimating β would be easier. We also show how, in this study, the proper calculation period is efficiently determined according to soil moisture status but without using soil moisture.

3. Data

Meteorological data

The following daily data from the meteorological station at the National Agricultural Research Center for Hokkaido Region (43.0°N, 141.4°E) (Sameshima *et al.*, 2008) was used for the estimation of IE and β in this study.

- (1) Solar radiation R_s (W m^{-2})
- (2) Sunshine duration N (h)
- (3) Mean air temperature T_a ($^{\circ}\text{C}$)
- (4) Mean wind speed U_{10} (m s^{-1}) at 10 m height
- (5) Mean specific humidity q_a (kg kg^{-1})
- (6) Amount of precipitation Pr (mm)
- (7) Mean soil temperature T_g ($^{\circ}\text{C}$) at a depth of 0.05 m in bare soil (volcanic ash soil)

3.2 Parameters

Parameters required to estimate the surface moisture availability over a long period using the proposed method are shown in Table 1. Parameters c , D_a , $C_H U$, and a were determined based on field and laboratory experiments and were adopted from the averaged value (Fukumoto and Hirota, 1994; Hirota *et al.*, 1995; Hirota and Fukumoto, 1996). The emissivity of ground ε is assumed to be unity in this study.

3.3 Surface soil moisture data

Volumetric soil moisture content ($\text{m}^3 \text{m}^{-3}$) in the 0–0.02 m layer of bare soil was measured by the gravimetric method every day at 09:00 to analyze the soil wetness condition. The observation period was from May 15, 1992 to August 14, 1992 (Fukumoto and Hirota, 1994).

4. Method for estimating β

4.1 Estimation of soil wetness condition

The antecedent precipitation index (API) is frequently used for qualitative judgment of the soil wetness condition (Saxton and Lenz, 1967; Iwai and Ishiguro, 1977; Wang, 1985; Choudhury and Blanchard, 1983; Kondo *et al.* 1998; Shinoda *et al.* 2000). Although there are many types of API expressions, the one used in our paper is given by the following equation (Choudhury and Blanchard, 1983):

$$API_j = K_j (API_{j-1} + Pr_j) \quad (3)$$

where API_j is the antecedent precipitation index on Day j , Pr_j is the amount of precipitation (mm) on Day j , and K_j is the recession coefficient defined by

$$K_j = \exp(-Ep_j/W_m) \quad (4)$$

where Ep_j is the daily potential evaporation (mm) on Day j and W_m is the maximum depth of soil available for evaporation (mm). The potential evaporation Ep_j can be calculated from Eqs. (10) to (20) (Appendix) in the case of $\beta=1$ and ground soil heat flux $G=0$.

Table 1. Values of parameters used in this study.

Item	Symbol	Value
Volumetric heat capacity	c ($\text{J m}^{-3} \text{K}^{-1}$)	2.0×10^6
Damping depth	D_a (m)	1.7
Exchange speed	$C_H U$	$0.00229 \times 0.7 U_{10} + 0.00456$
Annual mean soil temperature	T_{ym} ($^{\circ}\text{C}$)	9.0
Surface emissivity	ε	1.00
Soil albedo	a	0.1 (rainy day) 0.25 (other days)

Fig. 1. Temporal variations in soil moisture at 0-0.02 m layer and antecedent precipitation index (*API*).

Fig. 1 shows the relationship between *API*, in the case of $W_m=10$, and soil moisture in the 0-0.02 m soil layer. The variation in *API* is similar to that in soil moisture. As such, the *API* is useful for classifying the soil wetness condition in estimating the average value of β over a long period. In this calculation, an *API* value ≥ 20 corresponds to soil moisture content equivalent to or greater than field capacity. When the *API* value remains continuously <1 , this is regarded as the dry condition, and may correspond to the third stage of evaporation processes over bare soil (He and Kobayashi, 1998; Kobayashi *et al.* 1999). From this analysis of the relationship between soil moisture variation and *API*, we classified soil wetness into the following three conditions so as to estimate β :

- 1: Wet condition (greater than or equivalent to field capacity when $API \geq 20$);
- 2: Semi-dry condition ($20 > API \geq 1$) for more than 10 consecutive days; and
- 3: Dry condition ($API < 1$) for more than 10 consecutive days.

If $API \geq 20$ does not occur for more than 10 consecutive days, the soil condition is classified as semi-dry. Also, if the dry condition ($API < 1$) does not occur for more than 10 consecutive days, the soil condition is classified as semi-dry.

It is considered that W_m is a complex function of soil physical and plant properties (Choudhury *et al.* 1987) and therefore, it is not easy to determine this value in practice. If W_m is changed, the classification value of *API* is also changed. In this case, we assumed that $W_m=10$. If $W_m=20$, the wet condition is defined as $API \geq 30$, the semi-dry condition is defined as $30 > API \geq 5$, and the dry condition is defined as $API < 5$. If W_m is three times the value ($W_m=30$), the wet condition is defined as $API \geq 35$, the semi-dry condition is defined

as $35 > API \geq 10$ and the dry condition is defined as $API < 10$. However, the pattern of variation is the same, as long as W_m is taken as a constant throughout the study period, and it is still easy to roughly classify the soil wetness condition using *API*. In this study, *API* is used to classify the state of soil wetness on bare soil as a qualitative and auxiliary method for determining the calculation period for β over periods ranging from 10 days to several months; it is not used to correctly estimate daily soil moisture values. For the purpose of qualitative judgment of soil wetness condition, we assumed that it is not important to correctly obtain W_m .

4.2 Estimation and parameterization of β

The mFRM (see Appendix) for estimating daily mean soil temperature can be solved to estimate surface moisture availability β , latent heat flux IE , and sensible heat flux H for given meteorological, soil, and surface roughness conditions (Hirota *et al.* 1995; Hirota, 1999) via the procedures summarized in Fig. 2 and detailed in the following paragraphs.

First, the soil wetness condition is classified using *API*. Second, the exchange speed $C_H U$ is estimated using Eq. (2) for wet conditions ($API \geq 20$) and $C_H U$ is parameterized as a linear function of daily mean wind speed (Hirota and Fukumoto, 1996). Third, the surface moisture availability β is estimated for the daily mean meteorological data and the given soil parameter using the mFRM (Appendix) under given proper estimation period based on the soil wetness condition. The procedure for estimating the surface moisture availability β is as follows. Choose β to minimize the sum of the squares of the residual between the observed value of the daily mean soil temperature and the calculated value, χ^2 , written as

$$\chi^2 = \sum \{T_\delta(\text{calculated}) - T_\delta(\text{observed})\}^2 \quad (5)$$

Here, the summation should be taken over the proper estimation period. The proper estimation period was determined by using *API* information. It should be noted that the recommended calculation period for β is at least 10 days. The proper estimation period β is basically taken from the first occurrence of $API \geq 20$ to the next period just before $API \geq 20$. Furthermore, for the case of estimating β more accurately under dry condition, the period β is taken from the first timing of dry conditions $API < 1$ to the next timing just before $API \geq 20$. If the period between two

Fig. 2. Procedure for parameterization of the surface moisture availability β without flux measurement and soil moisture.

Fig. 3. The soil moisture availability β and the API . (1) wet condition period, (2) semi-dry condition period, (3) dry condition period.

consecutive occurrences of $API \geq 20$ is not more than 10 consecutive days, the soil wetness condition is classified as a semi-dry condition. Also, if $API < 1$ is not more than 10 consecutive days, the soil wetness condition is classified as a semi-dry condition. Thus, API is used to determine the calculation (averaging) period for estimating β (see the example in Fig. 3). This procedure using API is helpful for achieving correct and efficient estimation of β . From these procedures, we can estimate surface moisture availability on bare soil without using observed sensible and latent flux.

In the fourth and final step, we introduce the parameterization method for surface moisture availability β without using soil moisture. β is parameterized using the ratio of the precipitation Pr to potential evaporation Ep , where Ep is calculated from Eqs. (10) to (20) (Appendix) in the case of $\beta=1$ and soil heat flux at the surface $G=0$. This method of parameterization is

Fig. 4. Relationship between Pr/Ep and β (a); between Pr/Ep and E/Ep (b).

often used for the annual mean period (e.g., Budyko, 1974; Kondo and Xu, 1997). In our study, we applied this parameterization to shorter periods ranging from 10 days to several months, which was based on soil wetness conditions determined by API . Fig. 4 shows the relationship between β , E/Ep and the mean value of Pr/Ep using data from 1992 to 1994 from spring to fall during conditions of no snow cover. Fitting equations

describing the relationship between β and Pr/Ep and between E/Ep and Pr/Ep are as follows:

$$\bar{\beta} = 0.84 \tanh\left(\frac{(\sum Pr / \sum Ep)}{2.47}\right), \text{ for } \sum Pr / \sum Ep > 0, \text{ where for } \sum Pr / \sum Ep = 0, \beta = 0.03 \quad (6)$$

$$\overline{(E/Ep)} = 0.88 \tanh\left(\frac{(\sum Pr / \sum Ep)}{1.64}\right), \text{ for } \sum Pr / \sum Ep > 0, \text{ where for } \sum Pr / \sum Ep = 0, \overline{(E/Ep)} = 0.07 \quad (7)$$

where, $\bar{\beta}$ and $\overline{(E/Ep)}$ are each average value during the calculation period based on soil wetness conditions determined by API . $\sum Pr$ and $\sum Ep$ are each summation value during the calculation period based on soil wetness conditions determined by API .

These results show that $\bar{\beta}$ or $\overline{(E/Ep)}$ for periods of more than 10 days, can be parameterized without using soil moisture. We already showed that this method can be used to correctly estimate cumulative evaporation on bare soil for a long period (more than 10 days) with evidence that supports the heat balance since a valid exchange speed $C_H U$ is given (Hirota *et al.* 1995).

5. Verification results

Once we establish the parameterization for $C_H U$ and β , the daily mean soil temperature and cumulative evaporation can be estimated using routine meteorological data (*e.g.*, air temperature, humidity, solar radiation, sunshine duration, wind speed, and precipitation). In the previous section, we established the relationship between β and Pr/Ep using 1992-1994 data. In this section, in order to verify the parameterization result for β in Fig. 4 (Eq. 6), the rest of the dataset, 1995-1997, was used.

Fig. 5 shows the schematic diagram for estimating daily mean soil temperature and evaporation by using the mFRM (see Appendix). In this procedure, the API also has an important role as it is used to determine the calculation period (averaging period) for estimating β for each soil moisture status. After determining the calculation period, Pr/Ep is calculated for each period, and then β is estimated using Eq. 6 (Fig. 4). An example of the variation in API ($W_m=10$) and the results for estimating β using Eq. 6 are shown in Fig. 3.

Thus, since β can be obtained, the daily mean soil temperature and other heat balance components can be

Fig. 5. Procedure for estimating surface moisture availability, daily mean soil temperature and cumulative evaporation.

Fig. 6. Comparison of daily mean soil temperature at a depth of 0.05 m and estimated surface moisture availability β (1995 year).

Fig. 7. Comparison of daily mean soil temperature at a depth of 0.05 m (1996 and 1997 year).

estimated using the procedure in Fig. 5. Fig. 6 and 7 show the comparison of observed soil temperature and

that estimated by the procedure shown in Fig. 5 for the years 1995 through to 1997. The temporal variation in estimated values of daily mean soil temperature corresponds well with the observed values during 1995, 1996 and 1997. The root mean square error (RMSE) of the annual period varied between 0.82°C and 0.93°C during 1995 to 1997. On the other hand, the case when β is treated as a constant value over the calculation period, the RMSE in 1995 ($\beta=0.35$) was 0.91°C, that in 1996 ($\beta=0.30$) was 1.10°C and that in 1997 ($\beta=0.18$) was 1.06°C. The accuracy was improved by 0.1 to 0.2°C by using the calculation period determined by *API*. These results show that the daily mean soil temperature can be reasonably estimated using our proposed method for β .

6. Discussion and Conclusions

We developed a simple method for estimation and parameterization of surface moisture availability over periods ranging from 10 days to several months on bare soil using only routine meteorological data without needing sensible and latent heat flux and soil moisture data. The data required for our method is daily mean soil temperature, daily mean air temperature, daily solar radiation, sunshine duration, daily precipitation, daily mean wind speed, and daily mean humidity. The surface moisture availability can be estimated by using a simple soil temperature model composed of the heat balance equation at the ground surface, bulk transfer method and modified force-restore model. The estimation period of surface moisture availability can range from 10 days to several months in duration. We showed that the antecedent precipitation index (*API*) was an effective indicator for judging the soil wetness condition and thus for classifying the proper estimation period of surface moisture availability. Furthermore, for the purpose of estimating daily mean soil temperature and cumulative evaporation on bare soil using only routine meteorological data, we show that the surface moisture availability can be parameterized using the ratio of precipitation to potential evaporation based on soil wetness conditions classified using the *API* method. This parameterization does not require soil moisture data.

In this study, we used an empirical method to judge the soil wetness condition. As there are many other *API* methods, further research may be needed to find a more suitable one. Our proposed method cannot be used for shorter time scales such as a diurnal cycle. This is a

disadvantage compared with previous parameterization methods for surface moisture availability based on surface soil moisture content (e.g., Kondo *et al.* 1990; Lee and Pielke, 1992; Mahfouf and Noilhan 1991; Fukumoto and Hirota, 1994). However, our method has the advantage of needing only routine meteorological data without needing special measurements. Also, by using daily mean values, the estimated value cancels out the error of each hourly estimation (Hirota *et al.* 1995; Hirota *et al.* 2001) and the effect of soil thermal properties on the calculation of annual variations in the mean daily soil surface temperature is small, compared to the effect on diurnal variations (Hirota *et al.* 1995; Hirota *et al.* 2002). Therefore, this method can reduce the effect of observation error on the estimated value.

Although verification in this study is limited to only bare soil and relatively humid conditions, we consider that application of our parameterization method can be expanded to drier and vegetation fields. The function form of *Pr/Ep* is similar to the radiative index of dryness proposed by Budyko (1974), which can be applied to not only bare soil but any vegetation and climatic region for estimating regional and annual evaporation (Gutman *et al.* 1984). In this study, we show that the function of *Pr/Ep* could be applied to a shorter time scale ranging from 10 days to several months on bare soil by using *API* for classifying soil moisture status. Future work using the method in this study can be expanded to vegetation fields by combining the functions of *Pr/Ep* and *API*. In order to expand our method for vegetation areas, using remote sensing data such as surface temperature and normalized differential vegetation index (*NDVI*) may be more effective than the use of soil temperature for bare soil.

APPENDIX

Modified version of force-restore model for estimating daily mean soil temperature (Hirota *et al.* 1995; Hirota, 1999)

The modified version of the force-restore model for estimating daily mean soil temperature using daily mean or daily meteorological data is expressed by

$$C1(\delta) \frac{\partial T_g(\delta, t)}{\partial t} = \frac{2}{cD_a} G(0, t) - \frac{2\pi}{\tau_y} (T_g(\delta, t) - T_{ym}) \quad (8)$$

where $T_g(\delta, t)$ is the average temperature ($^{\circ}\text{C}$) in a soil layer with a thickness of δ (m), defined as,

$$T_g(\delta, t) = \frac{1}{\delta} \int_0^{\delta} T(z, t) dz \quad (9)$$

$T(z, t)$ is the soil temperature over some vertical depth, z , and time t , c is the volumetric heat capacity of soil ($\text{J m}^{-3} \text{K}^{-1}$), D_a is the damping depth for annual value (m), $G(0, t)$ is the soil heat flux at the surface (W m^{-2}), T_{ym} is the annual mean soil temperature ($^{\circ}\text{C}$), τ_y is the annual period (365 days), and $C1(\delta) = (1 + 2\delta/D_a)$. T_{ym} is relatively invariant with depth until a depth of several meters (Hirota *et al.* 1995; Hirota *et al.* 2002), therefore T_{ym} can be adopted for any depth up to several meters.

The force-restore model (FRM) was originally developed for the simple prediction of diurnal variation in ground surface temperature (Bhumralkar, 1975). Regarding the annual variation in the daily mean value, we have shown that the force-restore model not only accurately estimates ground surface temperature but also soil temperature in a layer up to approximately 0.5 m thick for agricultural land (Hirota *et al.* 1995; Hirota, 1999).

Equation (8) is used in combination with the following Eqs. (10) to (20).

In the case of $\delta \rightarrow 0$

$$\frac{\partial T_g(0, t)}{\partial t} = \frac{2}{cD_a} G(0, t) - \frac{2\pi}{\tau_y} (T_g(0, t) - T_{ym}) \quad (10)$$

where $T_g(0, t)$ is the ground surface temperature ($^{\circ}\text{C}$). The heat balance at the ground surface is given by:

$$G(0, t) = R_n - IE - H \quad (11)$$

$$R_n = (1 - a) R_s + \varepsilon (L_d - \sigma (273.15 + T(0, t))^4) \quad (12)$$

where R_n is the net radiation (W m^{-2}), R_s is the solar radiation (W m^{-2}), L_d is the atmospheric long-wave radiation (W m^{-2}), IE is the latent heat flux (W m^{-2}), H is the sensible heat flux (W m^{-2}), a is the soil albedo, ε is the emissivity of the ground, and σ is the

Stefan-Boltzmann constant ($\text{W m}^{-2} \text{K}^{-4}$). The latent heat flux IE and the sensible heat flux H are expressed by the following bulk formula (Kondo, 1994):

$$H = c_p \rho C_H U (T(0, t) - T_a) \quad (13)$$

$$IE = l \rho \beta C_H U (q_{sat}(T(0, t)) - q_a) \quad (14)$$

where l is the latent heat of vaporization (J kg^{-1}), β is the surface moisture availability, C_H is the bulk transfer coefficient for sensible heat flux, U is the wind speed (m s^{-1}), $C_H U$ is the exchange speed for sensible heat flux (m s^{-1}), q_a is the specific humidity of the air (kg kg^{-1}), $q_{sat}(T_g(0, t))$ is the saturation value at the ground surface temperature (kg kg^{-1}), c_p is the specific heat of the air ($\text{J kg}^{-1} \text{K}^{-1}$), ρ is the air density (kg m^{-3}), and T_a is the air temperature ($^{\circ}\text{C}$).

The daily mean atmospheric long-wave radiation L_d is estimated by the following equations (Kondo, 1994):

$$L_d = C_c L_{df} + (1 - C_c) \sigma (273.15 + T_a)^4 \quad (15)$$

$$C_c = A_1 (n/N)^3 - A_2 (n/N)^2 + A_3 (n/N) + A_4 \quad (16)$$

where C_c is the cloudiness factor, N is the possible sunshine duration and n is the sunshine duration. Here, $A_1=2.1$, $A_2=-3.241$, $A_3=2.291$, and $A_4=0.291$, as coefficients obtained at the Hokkaido National Agricultural Research Center for Hokkaido Region from observed data in 1992 (Hirota *et al.* 1995).

L_{df} is the daily mean atmospheric long-wave radiation under clear sky conditions (W m^{-2}) and is estimated from air temperature and effective water vapor amount using the following equations (Kondo, 1994):

$$L_{df} = (0.74 + 0.19x + 0.07x^2) \sigma (273.15 + T_a)^4 \quad (17)$$

$$x = \log_{10} w_{TOP}^* \quad (18)$$

$$\log_{10} w_{TOP}^* = 0.0315 T_{DEW} - 0.1836 \quad (19)$$

$$T_{DEW} = \frac{237.3 \log_{10} \left(\frac{e_a}{6.11} \right)}{7.5 - \log_{10} \left(\frac{e_a}{6.11} \right)} \quad (20)$$

where w_{TOP}^* is the effective water vapor amount (cm), T_{DEW} is the dew point temperature ($^{\circ}\text{C}$) and e_a is the

daily mean vapor pressure (hPa).

The model was typically used to estimate daily mean ground surface temperature or daily mean soil temperature in a layer with a thickness of δ . From our soil temperature observation value, we found that $T_g(\delta, t) \approx T(\delta/2, t)$ within 0.4 m thickness (Hirota *et al.* 1995). Therefore, in this model verification study, model output value $T_g(0.1, t)$ was compared with observed $T(0.05, t)$.

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ルーチン気象観測データからの裸地面の蒸発効率の推定と 土壌水分データを用いないパラメータ化

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要 約

バルク法による裸地面からの蒸発量を推定する場合、蒸発効率は最も重要なパラメータとなる。本研究では、10 日以上から数ヶ月程度の蒸発効率の期間平均値を、顕熱と潜熱フラックスおよび土壌水分のデータを用いずにルーチン気象観測データから簡易に推定しパラメータ化する方法を開発した。地温データがルーチン気象観測データ(気温、日射量、降水量、風速、湿度等)と共に観測されて、これらの日平均値の値が提供されるならば、蒸発効率は簡易な地温推定モデルにより潜熱、顕熱フラックスデータを用いなくても推定することができる。地温推定モデルは地表面の熱収支式とバルク法、および日平均地温を推定するための force-restore 法で構成されている。蒸発効率の推定期間は、土壌の湿潤状態を反映

する降水量の条件によって 10 日程度から数ヶ月程度となる。その際、先行降雨指数は土壌水分データを用いないで、土壌の湿潤状態を判定できる効果的な指標となる。この先行降雨指数により蒸発効率を土壌の湿潤状態に応じた適切な推定期間に分類することができる。蒸発効率は先行降雨指数によって分類された期間の下での日平均地温の推定値を観測値の差の二乗値の合計値が最小になるように推定する。さらに、裸地面の日平均地温と積算蒸発量を標準的な気象観測データから推定することを目的として、10 日から数ヶ月平均の蒸発効率を降水量と可能蒸発量の比からパラメータ化する手法を提案する。
キーワード：蒸発効率、蒸発量、地温、パラメータ化、裸地