Measuring the Evaporation from a Sand Surface at the HEIFE Desert Station by the Dry Surface Layer (DSL) Method

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Abstract

Most methods of measuring soil-surface evaporation originate from those of measuring water-surface evaporation, and hence they are not necessarily suitable for making measurements of evaporation from dry land surfaces. The major reason is that, as observation shows, when the dry surface layer (DSL) forms on the soil, the constant-flux layer of water vapor is barely developed in the surface air layer. However, during most of the day, the upward water vapor flux in the DSL is nearly constant with depth. The principle of the DSL method is to estimate this upward near-constant flux in the DSL from the depth profiles of temperature and soil water content in the layer.

The DSL method was applied to measurements of evaporation from a sand surface at the HEIFE desert station and they were compared to the decreases in soil moisture in the top 1 m of sand, from which the following results were obtained.

(a) Estimates of daily evaporation about a week after rainfall of 15 mm or so, which were made on the basis of seven measurements taken by the DSL method during the day, were in good agreement with the daily decreases in soil moisture in the top 30 cm of sand.

(b) Daily evaporation from dry soil can be approximately estimated from one measurement of evaporation rate made by the DSL method in the afternoon and the time lapsed since the last rainfall.

1. Introduction

It is rather difficult to measure the evaporation from dry land surfaces because, firstly the evaporation rate is very small and, secondly, constant-flux layers of water vapor are barely developed over dry land surfaces. Although the first must be expected and does not need to be explained further, the second seems to have to be reviewed, because this phenomenon has been recognized only in the past several years.

Recently, many observations of boundary-air-layer processes in arid regions have been made and a “peculiar” phenomenon has been found. Wang and Mitsuta (1990) showed that the flux of latent heat or water vapor over a gobi desert in the HEIFE area was negative (downward) in most of the daytime. Harazono *et al.* (1992) also observed humidity inversions over a sand dune in the Hulunbuir sandy land, China, during the day. However, those observations do not mean that water vapor moves into the ground, because its gradient immediately below

the surface in the dry surface layer (DSL) points downward, that is, soil water moves upward just beneath the surface (Kobayashi *et al.*, 1993). Hu *et al.* (1993) made observations of the profile of water vapor in the 0.25 to 16 m surface air layer over a sand surface at the HEIFE desert station and found there often existed a minimum value of specific humidity near the ground in the daytime, which means that the water vapor flux is not constant with height but even changes its vertical direction in the surface air layer. Nagai *et al.* (1995) made observations of water vapor profile in the air layer from 0.01 to 1 m over the same sand surface at the HEIFE desert station and showed that there was a tendency in the thin air layer in contact with dry soil surfaces for the humidity inversion to occur around nearly the same time of day. They also demonstrated that soil water was moving upward just under the surface when humidity inversions occurred above the surface, which suggests that there existed the counter-gradient flow of water vapor from the dry sand surface into the air. Tsukamoto *et al.* (1995) made observations of the eddy fluxes of sensible heat and water va-

por at two heights (1 m and 9 m) simultaneously in the same sandy place at the HEIFE desert station, and showed that the flux of water vapor changed considerably with height. Teshima and Kobayashi (1995) took observations of boundary-air-layer processes in Kokoshiri ranges of the Qinghai-Xizang plateau, China, and showed that the constant-flux layer of water vapor disappeared when the soil surface dried and unstable conditions prevailed in the air just above the surface. Tamagawa (1995) analyzed the data on wind, temperature and humidity obtained at the HEIFE desert station and raised the question as to whether humidity follows the Monin-Obukhov similarity over dry surfaces.

To discuss the behavior of water vapor in the neighborhood of dry land surfaces is not the purpose of the present paper. However, if the flux of water vapor changes with height in the surface air layer, the profile method or the aerodynamic method, including the bulk method, as well as the eddy correlation method of determining evaporation can hardly be applied to cases under dry conditions. The model or the parameterization of the soil-surface evaporation based on the same concept is also at variance with the reality. On the other hand, the energy balance method is not strictly applicable to dry surface conditions, because the phase transformation of water from liquid to vapor occurs not at the surface but in the soil. Consequently, in order to measure the evaporation from dry soil, it seems most rational to measure the upward flux of soil water in the depths where the constant-flux layer is likely to be formed.

Barnes and Allison (1983) developed a mathematical model describing the shape of H_2^{18}O and HDO depth profiles and asserted that three independent methods for estimating water fluxes in the soil with dry surfaces, or evaporation rates from dry soil, could be developed by using the shape of the ^{18}O or deuterium profile. Allison and Barnes (1985) applied this technique to the estimation of evaporation from the floor of the normally "dry" Lake Frome, a salt lake in northern South Australia, and obtained a mean evaporation rate from the lake of 170 mm yr^{-1} .

Kobayashi *et al.* (1989) proposed a method of measuring the upward water vapor flux in the dry surface layer (DSL) of sand in which water moves exclusively in the vapor phase. Only the measurements of temperature and water content at the surface and the thickness of the DSL are required to calculate the flux by this method. Kobayashi and Miyagawa (1992) generalized this technique to be applicable to soils other than sand and called it the "DSL method". If the DSL is formed on the soil, the upward water vapor flux in the DSL can be determined from the depth profiles of temperature and water content in the layer by the DSL method

(Kobayashi, 1990).

In the present study, the DSL method is applied to the measurement of evaporation from a sand surface at the HEIFE desert station and its accuracy is checked through comparisons to the decrease in soil moisture in the top 1 m of sand. A simple method is also presented for estimating daily evaporation from dry soil on the basis of one estimate of evaporation rate made by the DSL method in the afternoon and the time since the last rainfall.

2. The DSL method

The principle of the DSL method is summarized in this section (Kobayashi, 1990; Kobayashi and Miyagawa, 1992). The depths at which the phase transformation of water from liquid to vapor occurs penetrate into the soil as the soil becomes progressively drier, resulting in the formation of a dry surface layer (DSL) in which water moves only in the vapor phase. We shall suppose that the following conditions are fulfilled to a sufficient degree of approximation.

(a) The upward vapor transfer in the DSL can be considered as a near-uniform and steady-state process; that is, the DSL is a constant-flux layer of water vapor.

(b) The water vapor in a soil pore in the DSL is in equilibrium uniquely with the liquid and/or the adsorbed water in the same pore.

(c) The water vapor is transferred by isothermal molecular diffusion in the DSL.

If these three conditions are satisfied, the upward water vapor flux in the DSL, q_d (cm s^{-1}) is approximately written as

$$\begin{aligned} q_d &= (D_a \nu \alpha a / \rho_w) (\partial \rho_v / \partial z) \\ &= (D_a \nu \alpha a \rho_v g / \rho_w R_w T) (\partial \psi / \partial \theta) (\partial \theta / \partial z) \\ &= D (\partial \theta / \partial z), \end{aligned} \quad (1)$$

in as much as the Condition (b) yields the relation

$$\rho_v = \rho_s(T) \exp(\psi g / R_w T),$$

where z : depth (cm, positive downward),

D_a : molecular diffusivity of water vapor in air ($\text{cm}^2 \text{s}^{-1}$),

ν : mass-flow factor,

α : tortuosity factor ($=2/3$),

a : volumetric air content ($\text{cm}^3 \text{cm}^{-3}$),

ρ_v : density of water vapor (g cm^{-3}),

ρ_w : density of liquid water (g cm^{-3}),

$\rho_s(T)$: density of saturated water vapor at the temperature T (K) (g cm^{-3}),

g : gravitational acceleration (cm s^{-2}),

Table 1. Physical properties of the dune sand at the HEIFE desert station (Shibata, 1992).

Water content		Bulk density	Particle density	Porosity
pF=				
3.5	4.22 (vol. %)	1.59	2.65	0.399
3.8	3.91			
4.0	3.73	(g cm ⁻³)	(g cm ⁻³)	(cm ³ cm ⁻³)

R_w : gas constant of water vapor (erg g⁻¹K⁻¹),

ψ : matric potential of soil water (cm) being considered as a function of the temperature T (K) and the volumetric water content θ (cm³cm⁻³),

D : a kind of diffusion coefficient (cm²s⁻¹), which is expressed in the form of a function of T and θ , and is determined by experiment for each soil (Kobayashi *et al.*, 1986; Kobayashi, 1990).

In the present paper, conventional units are used for getting quantitative understandings of processes with ease.

Since the function $D(T, \theta)$ is approximately separable, that is

$$D(T, \theta) = C(T)G(\theta), \quad (2)$$

we can get Eq. (3) by integrating Eq. (1) with q_d constant over the depth in the layer between $z = d_1$ ($\theta = \theta_1$) and d_2 (θ_2) within the DSL,

$$q_d = \left\{ \int_{\theta_1}^{\theta_2} G(\theta) d\theta \right\} / \left\{ \int_{d_1}^{d_2} C^{-1}(T(z)) dz \right\}. \quad (3)$$

By this is meant that the upward water vapor flux through the DSL can be estimated from the vertical distributions of temperature and moisture in the DSL.

The validity of the assumptions and approximations which the DSL method is based on was verified by comparison of the evaporation rates made by direct weighing and by the present method in indoor experiments (Kobayashi *et al.*, 1989; Li *et al.*, 1989; Kobayashi and Miyagawa, 1992).

3. Rate equation of evaporation

Physical properties of the dune sand at the HEIFE desert station (Wang and Mitsuta, 1993) are shown in Table 1 (Shibata, 1992). The diffusion coefficient $D(T, \theta)$ of the dune sand, which can be obtained by a kind of trial and error using these properties in Table 1, is expressed as

$$D(T, \theta) = \exp(0.0614T - 38.86)\theta^2(\theta - 4.16)^6, \quad (4)$$

where and hereafter θ is expressed in percent (Kobayashi *et al.*, 1986).

Since the DSL of dune sand can be distinguished from the moist underlayer by the sharp contrast in

shade of color, its thickness, δ (cm) can be measured with a rule. Therefore, it is convenient to use $z = 0$ and δ for d_1 and d_2 in Eq. (3), respectively. Further, because the thermally induced vapor transfer can be ignored in the DSL (Kobayashi, 1993), the upward water vapor transfer can be considered as an isothermal process at the mean temperature for the DSL, T_m . Thus, substitution of Eq. (4) into Eq. (3) using Eq. (2) gives

$$q_d = \exp(0.0614T_m - 38.86) \{f(\theta_\delta) - f(\theta_0)\} / \delta, \quad (5)$$

where

$$\begin{aligned} f(\theta) &= \int_0^\theta \theta^2(\theta - 4.16)^6 d\theta \\ &= (1/9)\theta^9 - 3.12\theta^8 + 37.08\theta^7 - 239.97\theta^6 \\ &\quad + 898.45\theta^5 - 1868.78\theta^4 + 1727.58\theta^3. \end{aligned} \quad (6)$$

The value of θ at $z = \delta$, θ_δ corresponds to the hygroscopic coefficient ($pF = 4.5$) (Kobayashi, 1990); however, since the soil moisture changes quickly near the interface between the DSL and the moist underlayer and the depth $z = d_2 = \delta$ must be located completely within the DSL, 2.5 % corresponding to $pF = 5.0$ is used as θ_δ in the present study, though the value of θ_δ does not have such a large influence on q_d as that of θ_0 does. The value of θ_δ can be regarded as one of the soil water constants, while the value of θ_0 exhibits a diurnal variation (Fig. 4), and hence it is desirable that measured values of θ_0 are used in calculating q_d by Eq. (5).

4. Sampling and methods

The HEIFE desert station (about 1300 m ASL) is covered with sand dunes 1m to 4 m high. The present experiment was conducted on a gentle slope of a rather big and flat sand dune on which a 20 m meteorological tower was installed. Two turbulence data sampling systems also were put on the same slope (Tsukamoto *et al.*, 1995).

The observed elements were the DSL thickness; temperatures at a height of 20 cm, at the surface and at the center depth of the DSL; soil water contents at the surface (0 to 0.2 cm) and at depths of 5, 10, 15, 20, 25, 30, 40, 60 and 100 cm; humidities at a height of 2.6 cm and at depths of 0.6 and 2.6 cm. The DSL thickness was measured with a rule as mentioned

Fig. 1. Variation in time and depth of the soil moisture in the top 40 cm of a sand surface at the HEIFE desert station. The numbers labeling the curves indicate the volumetric soil water content in percent.

in the foregoing section. The temperature measurements were taken with thermistor thermometers, except for the air temperature which was measured with an Assmann ventilated psychrometer, and the surface temperature was measured by dusting the sensor with dry sand. The soil water content was determined by the oven dry method. The mean temperature for the DSL, T_m was identified with the arithmetic mean of the temperatures at the surface and at the center depth of the DSL instead of the latter because the temperature decreases not linearly but near-exponentially with depth in the layer during the day. The humidity was measured with HUMICAP[®] H type (VAISALA) sensors at 5-minute intervals during the period August 23 to 31, 1994. The other measurements were made as a rule once a day at 1630 BST (local time is about one hour behind) from August 20 to 28, 1994; in addition, special intensive measurements were made for two days (Aug. 27–28). It began to rain in the early dawn of Aug. 19 and the total rainfall was estimated to be about 15 mm, though the precise value was not taken because the rain gauge at the HEIFE desert station was not working on the day. The weather was mostly fine throughout the period except for a few cloudy days and a windy day (Aug. 28).

5. Results and discussion

5.1 Depth profile of soil moisture

Figure 1 shows the time variation of the soil moisture distribution in the top 40 cm of sand. Soil samples for use to determine the water content were taken from different test holes in a region several meters in size, and hence there appears in Fig. 1 some inconsistent isopleth patterns which reflect the spatial variation in soil moisture. However, we can see the effect of the rainfall event of Aug. 19 on the depth profile of soil moisture, which moved downward to a depth of about 25 cm and became vague

Fig. 2. Temporal variations of the amount of soil water contained in the surface layer. The solid squares show the measurements in the top 30 cm and the open ones those in the 0 to 100 cm surface layer. The curves are fits by least squares.

as time progressed. The depth to which rain water reaches increases with increasing rainfall depth or its intensity, and also depends on the soil moisture condition; the drier the soil, the shallower the depth to which the rain water reaches.

Figure 2 shows the temporal variations of the amount of soil water contained in the top 30 cm of sand and that in the 0 to 100 cm surface layer. The solid curve, which was determined by fitting an exponential regression equation through the data from the top 30 cm (solid squares), seems to be a good fit of the variation in time, which suggests that the spatial variation in soil moisture in the upper 30 cm was small. The curve shows that the decrease in soil moisture in the surface layer was about 10 mm in depth during the period from Aug. 20 to 28. On the other hand, the amount of soil water in the surface layer 100 cm thick (open squares) shows some scatter, which reflects the large spatial variation in soil moisture in the moist underlayer (Berndtsson and Chen, 1994). The measurements seem to be able to be classified into two groups, an upper and a lower, and are bounded above and below by curves (dotted lines) which are least-squares regressions obtained from the respective groups. These curves show that the decrease in soil moisture in the surface layer 100 cm thick during the 8-day period was in the range of about 10 to 20 mm in depth.

As has been discussed above, the effect of rainfall on the depth profile of soil moisture was not recognized in layers below 30 cm. Since the profile of soil water content had a minimum value at a depth of about 30 cm during the observation period (Fig. 1), the soil water flux at 30 cm must have been very

Fig. 3. Diurnal variations of the thickness of the DSL and the evaporation rate measured by the DSL method.

small, though its direction can not be specified. It is probable that the soil water flux at 100 cm directed downward, because summer is the rainy season in this area and soil moisture may be replenished in the moist underlayer mostly in summer. Therefore the total amount of soil-surface evaporation during the 8-day period was estimated to be slightly larger than or nearly equal to the decrease in soil moisture in the top 30 cm of sand.

5.2 DSL evaporation rate

Intensive observations were made on Aug. 27 and 28. Figure 3 shows the diurnal variation of the thickness of the DSL and the upward water vapor flux in it measured by the DSL method. Even a week after the rainfall of about 15 mm, the DSL thickness exhibited a diurnal variation with a range of about 2 cm on Aug. 27. The DSL is moistened mainly by the moisture moving upward from the moist underlayer and is reduced in thickness (Kobayashi *et al.*, 1991). The evaporation rate increased quickly in the morning and reached a maximum around the noon by the local time and gradually decreased in the afternoon so long as the upward water vapor flux in the DSL, q_d can be regarded as the soil-surface evaporation rate. This postulate of equality is justified provided that the temperature at the surface is higher than in the air surface layer (Fig. 4) (Kobayashi, 1993). The reason that the evaporation rate is relatively large in the morning is that the DSL absorbs water vapor moving upward from the moist underlayer during the night and the absorbed moisture is released in the following morning.

Figure 4 shows the diurnal variations of the air temperature at a height of 20 cm, the temperature

Fig. 4. As in Fig. 3, except for the air temperature at 20 cm height, the temperature and volumetric soil water content at the surface.

and soil water content at the surface (T_0 , θ_0) measured simultaneously in the same sandy place. It is clear that T_0 and θ_0 were out of phase each other and θ_0 decreases (increases) as T_0 increases (decreases). Evaporation from bare soil proceeds in three stages as the surface layer dries and not only the rate but also the mechanism of evaporation change from stage to stage (Kobayashi, 1990). The influence of weather conditions on the rate of evaporation from dry soil (during the third stage of evaporation) is exerted through T_0 and θ_0 . According to the rate equation (5), it looks like the evaporation rate increases with decreasing θ_0 , that is, the drier the soil surface the greater the evaporation rate is. This is crucially different from the second stage of evaporation during which the evaporation rate decreases as the soil surface dries and the relative humidity at the surface falls. In reality, since the thickness of the DSL (δ) increases as θ_0 decreases, the evaporation rate is determined by the balance of the two quantities along with the temperature in the DSL.

5.3 Daily evaporation

The daily amount of evaporation estimated from the decrease in soil moisture in the top 30 cm is shown in Fig. 5 (dotted line). The dashed line shows that estimated from the soil moisture decrease in the layer from 0 to 100 cm (dashed line in Fig. 3). However, as has been discussed above, the true rate of evaporation can be thought to agree well with the decrease in soil moisture in the upper 30 cm in the present case. The difference between the two estimates is due to the local variation in soil moisture in

Fig. 5. Estimations of daily evaporation made by various methods. See the text for explanation.

the layer from 30 cm to 100 cm, and also, probably, is due to the downward soil water flux at 100 cm depth. The explanation of the solid lines in Fig. 5 will be given in Subsection 5.4.

The open circles in Fig. 5 show the estimates of daily evaporation made based on seven measurements per day obtained by the DSL method assuming that soil-surface evaporation occurred during the 10-hour period from 0930 to 1930 BST, where it should be emphasized again that the local time is about one hour behind. Figure 6 shows the variations of humidity measured with HUMICAP[®] H type (VAISALA) sensors in the air and in the soil near the surface of the sandy place. Since the atmosphere (Fig. 6) and the soil surface were very dry and the air temperature was nearly equal to or lower than the surface temperature during the remaining 14-hour period of the day (Fig. 4), it is conjectured that no absorption by the soil of water vapor in the air above the surface occurred and the amount of evaporation was very small for the 14-hour period. As can be seen, the estimates made from the seven measurements for 2 days (Aug. 27–28) are in fairly close agreement with those made from the data on soil water content in the top 30 cm, which means that evaporation rates obtained by the DSL method are reliable. Although the accuracy of the DSL method was demonstrated by indoor experiments (Kobayashi *et al.*, 1989; Li *et al.*, 1989; Kobayashi and Miyagawa, 1992), this is the first time for the method to be checked by an outdoor experiment.

The crosses in Fig. 5 show the daily evaporation estimated from eddy fluxes obtained with the tur-

bulence data sampling system installed at a height of 1 m on the same place (Tsukamoto *et al.*, 1995). The estimate made on a windy day (Aug. 28) agrees well with the daily decrease in soil moisture in the upper 30 cm. However, those made for the other calm days were smaller than the daily decreases. As has been discussed in the Introduction, the major cause of these underestimates seems to be related to the conditions under which constant-flux layers of water vapor are not developed. In general, dry land surfaces are surrounded by regions with moist surfaces without regard to the size of a stretch of land and, hence, water vapor is transported to a dry region from the surroundings. On the other hand, as mentioned above, dry soil surfaces do not absorb water vapor in the air, at least during the daytime, thus air masses conditioned in moist regions do not easily adjust to dry soil surfaces. Consequently, the constant-flux layer can barely be developed over dry land surfaces, and hence the eddy flux of water vapor in the upward direction is not in agreement with the real rate of evaporation. However, at present we are not sure why the eddy flux is smaller, rather than larger than the real rate.

5.4 Daily mean evaporation rate

Figure 7 shows the day-to-day variations of the DSL thickness and the DSL evaporation rate measured at 1630 BST, but at 1500 BST on Aug. 22. As shown in Fig. 3, the thickness exhibits a diurnal variation with a maximum value at about 1600 local time (Kobayashi *et al.*, 1991), around which time the soil-surface evaporation can be regarded as a near steady-state process, thus 1630 BST was accepted as the observation time in the present experiment. The thickness of the DSL increased linearly up to 5 cm as time passed following the rainfall event of Aug. 19. The DSL evaporation rate decreased quickly within a few days after the rainstorm and continued to reduce exponentially.

The relationship between the evaporation rate at 1630 BST and the daily mean rate is thought to change as the soil becomes drier. Since the air was very dry (Fig. 6), it is believed that evaporation occurs even during the night for a while after a rainstorm. However, as time goes on in the dry spell, evaporation occurs only in the daytime and the surface absorbs water vapor moving upward from the moist underlayer at night. Thus it seems reasonable that the ratio of the daily mean evaporation rate to the rate at 1630 BST, denoted by R_{dm} , decreases as the day progresses during a dry period.

The solid lines in Fig. 5 show the daily evaporation estimated on the basis of the evaporation rate at 1630 BST shown in Fig. 7. The numbers on the curves refer to the values of R_{dm} . Although the ratio was 0.8 to 0.9 shortly after rain, it became about 0.5 within a week. We can conceive of the ratio as

Fig. 6. Temporal variations of humidity in the air (solid line) and in the soil (dotted and dashed lines).

Fig. 7. Day-to-day variations of the thickness of the DSL and the evaporation rate measured by the DSL method at 1630 BST, but at 1500 BST on Aug. 22.

decreasing progressively toward about 0.3.

Kobayashi and Miyagawa (1992) proposed a method for estimating the daily amount of evaporation by combining a soil-surface evaporation model and the DSL method, and applied it to the evaporation from arcose sand. Since arcose sands are finer than dune sands, it is rather difficult for the DSL to form on the soil, and hence the experiments were conducted when the DSL was developed to be about

1 cm thick under natural conditions. According to their results, obtained in Fukuoka, Japan, the ratio of the daily mean evaporation rate to the mean rate for the afternoon (1300–1700 JST for October and 1300–1800 JST for April and May) was 0.36 four days after a rainstorm (Oct. 2, 1989) and 0.28 and 0.27 a week after rainfall events (Apr. 30 and May 26, 1990), respectively. Taking account of that the mean evaporation rate for the afternoon is bigger than the rate at about 1600 local time, and that air humidity in Japan is rather high and the DSL may absorb water vapor in the air during the night, these results do not contradict the present results.

As a result, the daily evaporation from bare dry soil can be estimated from a measurement of evaporation rate made by the DSL method at about 1600 local time and using the time lapsed since the last rainfall during a spell of fine weather.

6. Concluding remarks

It is essential to clarify whether or not constant-flux layers of water vapor are formed over dry land surfaces as are made over water surfaces before the soil-surface evaporation in arid regions is discussed. A lot of observations made recently, however, have achieved negative results, which suggests that many models developed for the soil-surface evaporation and many techniques used for measuring it can not be applied to the evaporation from dry soil. These results, as a necessity, make us focus attention on the behavior of water under the soil surface rather than

above it. It was demonstrated in the present study that the DSL method, which measures the water vapor flux in the DSL, can be promising technique for use in measuring the evaporation from bare soil with dry surfaces.

The DSL method is applicable literally to cases when a dry layer has formed over the soil surface, and hence the evaporation rate just after a rainfall event must be measured by the other method. As O'kane (1991) also pointed out, the soil-surface evaporation mechanism changes suddenly from atmosphere-controlled to soil-controlled as the soil surface dries. Thus two methods must be used jointly to measure continually the evaporation from bare soil. Kobayashi and He (1995) proposed a rational parameterization of the soil-surface evaporation, in which the bulk method and the DSL method are used in cooperation for expressing the evaporation rates under atmosphere-controlled and soil-controlled conditions, respectively.

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乾燥表土層 (DSL) 法による HEIFE 沙漠観測点の砂地面からの蒸発量評価

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土壌面蒸発測定法の多くは水面蒸発測定法に由来する。したがって、それらは必ずしも乾燥土壌面からの蒸発の測定には適していない。その主な理由は、乾燥表土層 (DSL) が形成されると、接地気層内に水蒸気のコンスタント・フラックス層が形成され難いからである。しかし DSL 内では、日中の大部分の時間帯において、水蒸気の上向きフラックスは深さによらずほぼ一定である。DSL 法の原理はこの準コンスタント・フラックスを同層内の温度と含水量のプロフィールから推定しようとするものである。

DSL 法を用いて HEIFE 沙漠観測点の砂地面からの蒸発量が測定され、表層 1 メートルの土壌水分の減少量と比較された。その結果つぎのような結論が得られた。

(a) DSL 法によって日中に 7 回に測定した蒸発速度から推定した降雨 (約 15 mm) 後約 1 週間の日蒸発量は、表層 30 cm の土壌水分の日減少量とほぼ一致した。

(b) 乾燥土壌面からの日蒸発量の近似値は、DSL 法によって午後に 1 回測定した蒸発速度と降雨後の経過日数から推定できる。