The Impact of Soil Moisture on Dispersion–Related Characteristics

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ABSTRACT

This study investigates the impact of soil moisture availability on dispersion–related characteristics: surface friction velocity (u_*), characteristic scales of temperature and humidity (T_* and q_*), the planetary boundary layer height (h) and atmospheric stability classified by Monin–Obukhov length (L), Kazanski–Monin parameter (μ) and convective velocity scale (w_*) during daytime convective condition using a one–dimensional primitive equation with a refined soil model.

I. INTRODUCTION

Pollutants emitted from stacks will be dispersed in the planetary boundary layer (PBL) rapidly or slowly, depending on the pollution–dispersion–related characteristics in PBL during daytime hours. These dispersion–related characteristics involve the height of the PBL,h, the friction velocity (u_*). and characteristic scales of temperature and atmospheric humidity (T_* and q_*).The pollutant dispersion is catalogued according to atmospheric stability characteristics measured by one of the following six major methods, commonly used in atmospheric diffusion discipline: (a) the Pasquill stability category (from A to G), (b) the temperature gradient or the Richardson number Ri, (c) the variance of horizontal wind direction (σ_θ), (d) the Monin–Obukhov length (L), (e) the Kazanski–Monin parameter (μ), and (f) the convective velocity scale (w_*). All of them can be directly or indirectly expressed by u_*, T_*, q_* and have:

$$\frac{\partial\theta}{\partial z}=\frac{T_*}{kz}\varphi_h(\xi) \tag{1}$$

$$\sigma_\theta=\frac{kz}{\varphi(\xi)}(12+\frac{h}{2|L|})^{1/3} \tag{2}$$

$$L^{-1}=L_\theta^{-1}+L_q^{-1} \tag{3}$$

$$\mu=\frac{gkT_*}{T_v f u_*} \tag{4}$$

$$w_*=(\frac{g}{T_v}u_*T_*h)^{1/3} \tag{5}$$

where g is gravitational acceleration, $\xi=Z/L$, is non– dimensional height, φ_h is the non–dimensional Obukhov similarity functions for heat, f is Coriolis parameter, k is the Karman constant ,and L_θ and L_q are defined as

$$L_\theta^{-1}=\frac{gkT_*}{Tu_*^2}$$

$$L_q^{-1}=\frac{0.61gkq_*}{u_*^2} \tag{6}$$

where T and T_v are temperature in dry air and virtual temperature in humidity air, respectively.

$$T_v=T+0.61Tq \tag{7}$$

Equations (1)–(5) directly depict the relationship between atmospheric stability and elements (u_* ,T_* ,q_* and h), whereby these elements are called as dispersion–related elements.

Generally speaking, the Pasquill stability category is simple, qualitatively estimated from common meteorological elements: wind speed ,three levels of amount of solar radiation and cloud coverage, they are easy to obtain from routine meteorological station. Therefore it has been widely used for classifying atmospheric stability for past several decades. The σ_θ method is not only directly correlating to lateral dispersion σ_x ($\sigma_x=\sigma_\theta x$) near the source, but also correlates well with vertical dispersion (σ_z). The temperature gradient method was shown to be essentially uncorrelated with daytime diffusion rates (Briggs,1988). The scaling of dispersion in terms of Pasquill category and temperature gradient is out of date with many shortcomings, while the last three methods, however, have the great advantages in characterizing daytime diffusion developed since 1970 (Briggs 1988; Sutherland et al.,1987; Weil and Brower,1984; Golder 1972 ;and others).It is not a purpose of this paper,however,to assess in detail that which one of the six methods mentioned above, used to characterize atmospheric stability is relatively better.The purpose of the present study lies in exploring the impact of soil moisture availability on atmospheric dispersion ability measured by the six methods in terms of determining the influence of soil moisture availability on the turbulent characteristics:u_* ,T_* ,q_* and h. The soil moisture availability (m) is defined as follow:

$$m=\frac{\eta_w}{\eta_{ws}}$$

where η_w is the volumetric fraction of water in soil, and η_{ws} is the value of η_w at saturated moisture condition.

The impact of soil moisture availability on u_* ,T_* ,q_* and h is conducted by following processes: (a) The latent heat flux (LE), partitioned from surface net radiation, increases and the sensible heat flux(H_s), partitioned from surface net radiation, decreases when the value of m increases (see Ye and Jia, 1989 in detail), where H_s and LE are defined as:

$$H_s=\rho c_p u_* T_* \tag{8}$$

$$LE=\rho L u_* q_* \tag{9}$$

the PBL's height (h)is determined by sensible heat flux and described as (see Ye et al., 1987):

$$h(t)=\left(\frac{2Q_s}{\beta_0}\right)^{1/2} \quad (10)$$

$$Q_s=\int_0^t -\frac{H_s}{\rho c_p}dt \quad (11)$$

where $\beta_0=\frac{\partial\theta_0}{\partial z}$ is the background thermal stability. (b)The surface albedo (A) over soil is influenced by soil moisture availability based on the formula given by Idso et al.(1975):

$$A=0.31-0.17\text{m} \quad (12)$$

Therefore, the soil moisture availability influences the amount of solar energy (Rs) absorbed by earth's surface as

$$R_s=S(1-A)$$

where S is the incoming solar energy. (c) The soil consists of minerals, organic material, water and air. The variation of soil moisture availability would result in the corresponding change of overall soil heat capacity (C_s) and heat conductivity (K_s) (details see Ye and Jia, 1989), whereby the value of m exerts an impact on soil heat flux (G), based on equation:

$$G=-C_sK_s\frac{\partial T_s}{\partial z}=-C_sK_s\frac{\partial T_s}{\partial z}$$

In this paper, several aspects of the influence of soil moisture availability on dispersion–related elements (u_*, T_*, q_* and h) and the relationships between elemets (u_*, T_*, q_* and h) the and dispersion–related atmospheric stability have briefly been discussed. In Section 2, the one–dimensional primitive equation model used in current study is presented. A summary of the results is given in Section 3 and the main conclusion is presented in Section 4. Finally, the study focuses on daytime convective situations involved with a horizontally and vertically homogeneous distribution of soil moisture availability.

II. THE MODEL

A set of numerical model simulations,designed to investigate the impact of soil moisture availability on the dispersion–related characteristics,such as u_*, q_*, T_* and h, as well as their combinations (L, μ and w_*), which characterize the atmospheric stability features, was carried out under the conditions of synoptic flow $u_g=3$ m / s, surface roughness, $z_0=5$ cm and background thermal stability, $\beta_0=3.5$ K / km, with soil moisture availability, $m=0.01$, 0.05, 0.1, 0.5 and 1.0, respectively. A set of hydrostatic primitive equation model was used, whose formulation is given in detail in Pielke (1974), Mahrer and Pielke (1977) and McNider and Pielke(1981), only a brief description of this model is given in this paper.

Over flat horizontally homogeneous terrain, the predictive equations for wind vector, V, potential temperature, θ, specific humidity, q, and soil temperature, T_s, are as follows:

$$\frac{\partial V}{\partial t}=fK(V_g-V)+\frac{\partial}{\partial z}(Km\frac{\partial V}{\partial z})$$

$$\frac{\partial\theta}{\partial t}=\frac{\partial}{\partial z}(K_h\frac{\partial\theta}{\partial z})+R$$

$$\frac{\partial q}{\partial t}=\frac{\partial}{\partial z}(K_q\frac{\partial q}{\partial z})$$

$$\frac{\partial T_s}{\partial t} = \frac{K_s}{C_s} \cdot \frac{\partial^2 T_s}{\partial z^2} \tag{13}$$

where V_g is the synoptic wind, R is the radiation heating / cooling term due to shortwave and longwave radiation divergences, which are parameterized following the methods suggested by Atwater and Brown (1974) and Sasamori(1972). Here only radiation active constituents (water vapor and carbon dioxide) in the atmosphere are considered.

In the model, a surface heat energy equation is used to predict the surface temperature while the surface fluxes of momentum, sensible heat and latent heat are computed according to surface similarity law provided by Businger et al.(1971). The surface heat energy equation is:

$$R_n = H_s + LE + G \tag{14}$$

The model consists of 14 vertical levels in both air and soil, ranging from near surface to 7 km above ground and from surface to 0.7m into soil with grid interval 5 cm. The model levels for potential temperature and specific humidity and related initial values are given in Table 1. The soil input parameters are given in Table 2.

Table 1. The Numerical Model Levels for Potential Temperature and Specific Humidity and the Related Initial Values

levels (m)	10	32.5	75.0	200	400	600	800
	1050	1350	1750	2500	4000	6000	7000
q (g / kg)	5	5	5	5	5	5	3.4
	3.2	2.2	1.7	0.5	0.5	0.5	0.5
θ (℃)	10.0	10.1	10.3	10.7	11,4	12.1	12.8
	13.7	14.7	16.1	18.8	22.3	25.8	27.5

Table 2. The Soil Parameters Used in the Numerical Model Simulations

substance	volumetric fraction	C (cal cm^{-3}K^{-1})	K (m cal cm^{-1}K^{-1})
quartz	0.5	0.48	21
clay minerals	0.05	0.48	7
organic matter	0.05	0.60	0.6
water	0.4 m	1.00	1.37
air	0.4(1−m)	0.00030	0.06

III. RESULTS AND ANALYSIS

A set of numerical model simulations, designed to evaluate the impact of soil moisture availability on dispersion−related turbulent characteristics was carried out.

1. *The Impact of the Soil Moisture on Surface Friction Velocity*

The impact of m on u_* obtained from the numerical simulations is shown in Fig.1. Figure 1 describes that during the periods of about one hour of sunrise and sunset,the values of u_* increase rapidly during sunrise and decrease rapidly during sunset. During the most of daytime hours, the change of values of u_* with time, however, is relatively small (from 1000 to 1500 LST, the ratio of u_* / u_{*max} is within 90%) as compared to the changes during sunrise or sunset. Therefore the change of u_* with time during daytime can be neglected. The values of u_* decrease as increasing soil moisture availability for any daytime hours from our numerical simulated results as shown in Fig.1. The quantitative relationship between u_* and m is shown in inset of Fig.1, which can be approximately expressed as:

$$u_* = C_{u_*} - d_{u_*} m^{1/2} \qquad (15)$$

where the constants C_{u_*} and d_{u_*} are in unit m / s, with C_{u_*} =0.20and d_{u_*} =0.0167 at 1300 LST obtained from the present study. The constant C_{u_*} is corresponding to u_* at absolutely dry soil condition, and the value of $C_{u_*} - d_{u_*}$ represents the value of u_* at saturated soil moisture condition.The ratio of u_* between saturated and dry soils is about 0.9 ,which suggests that the impact of m on u^* is small and can be neglected as a first approximation used in following sections.

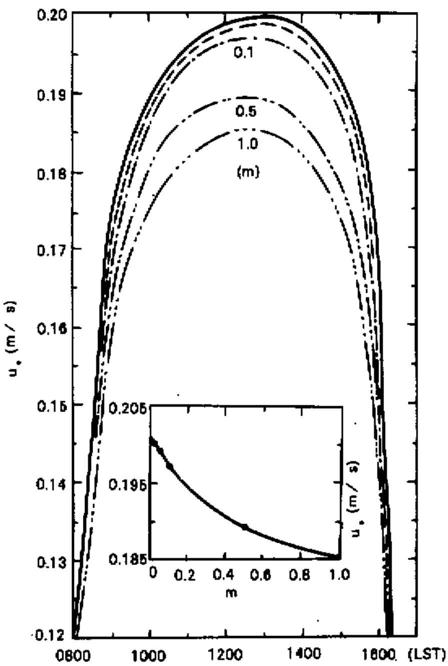

Fig.1. Numerical model simulated values of friction velocity (u_*) with differet values of soil moisture availability. In the inset, the change of u at 1300 LST, as a function of m is presented.

2. The Impact of m on T_*

The impact of m on T_* computed from present simulations is presented in Fig.2. Figure 2 shows that the values of T_* reach to their maximum values for all values of m at around noon hour. The values of T_* decrease as the soil moisture tends to drier. The relationship between T_* and m can be described as

$$T_* = d_h + C_h m^b \tag{16}$$

where constants C_h and d_h are in unit (K). For the current simulations at 1300 LST, we have: for $m \geqslant 0.1, C_h = -.177, d_h = -.25$ and $b = -0.4$; for $m \leqslant 0.1, d_h = -0.756, C_h = 0.6$ and $b = 1$ as shown in the inset of Fig.2. The value of d_h is the value of T_* at absolutely dry soil condition, and the value of $d_h + C_h$ is the value of T_* at saturated soil moisture condition. The ratio of T_* between saturated and dry soil is about 0.5. Therefore, the soil moistore availability exerts a strong impact on T_*.

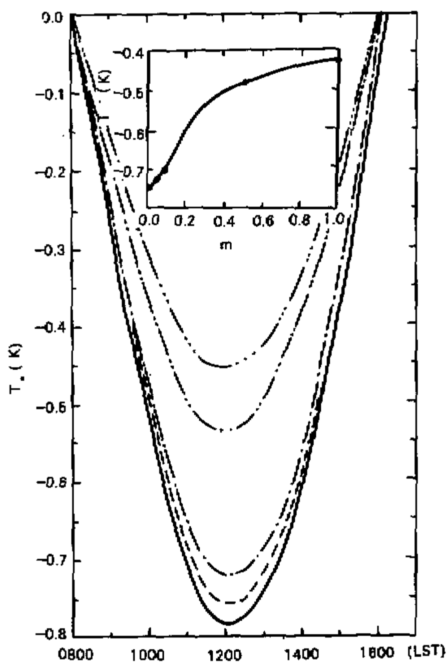

Fig.2. The same as in Fig.1, except for T_*.

3. The Impact of m on q_*

The impact of m on q_* shown by Fig.3 was simulated results, which illustrates that the influence of m on q_* is very sensitive, when the soil moisture availability changes its value from $m=0.01$ to $m=1.0$, the values of q_* are nearly changed by about two orders of magnitude. The relationship between q_* and m is described in the inset of Fig.3, and depicted as

$$q_* = C_q m^{0.6} \tag{17}$$

where $C_q = -4.69 \times 10^{-4}$.

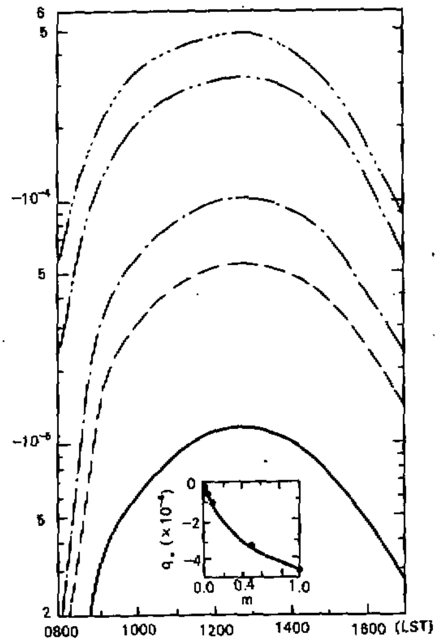

Fig.3. As in Fig.1, except for q.

4. *The Impact of* m on h

The time–dependence of h with different values of m from current simulations, based on the formula of Deardorff(1974),is presented in Fig.4a. Figure 4a illustrates that the PBL's height over very dry soil is about 40% higher than that over saturated soil.

Figure 4b is computed based on Eqs.(10) and (11) where the sensible heat flux is obtained from the simulations. Comparing Fig.4a, where h is predicted utilizing a formulation introduced by Deardorff (1974), with Fig.4b, where the PBL's height is predicted by Eqs.(10) and (11), shows that although the PBL's heights predicted from rather simple formulae are a little lower than that computed from Deardorff(1974), the dependence of h on m for two predicted formula, however, is consistent and the development with time by the two prediction formulae are very similar.

From Eqs.(10) and (11), the following relation can be derived:

$$h(m2)/h(m1)=[(u_*T_*)_{m2}/(u_*T_*)_{m1}]^{1/2} \tag{18a}$$

Since $u_*(m2)/u_*(m1)$ is much smaller than $T_*(m2)/T_*(m1)$ according to the discussion in subsection 1, Equation (18a) then can be simplified as

$$h(m2)/h(m1)=[T_*(m2)/T_*(m1)]^{1/2} \tag{18b}$$

Equations (16), (18b) represent the influence of m on h. Computed result based on Eqs (16),(18b) indicates that the value of h over absolutely dry soil is about 40% higher than that over saturated soil.

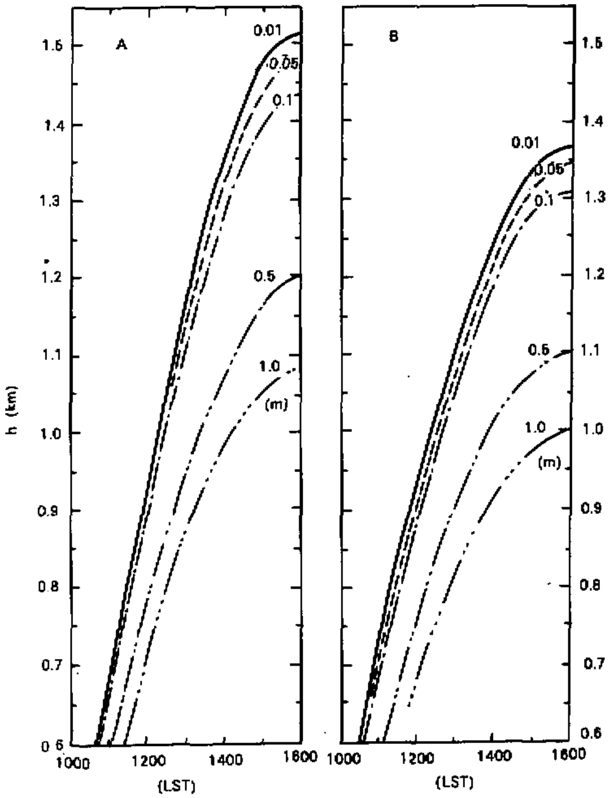

Fig.4. Numerical model simulated PBL's height (h):A, is based on Deardorff (1974), and B, based on Eqs, (10)and (11).

5. *Discussion*

In this subsection, only the dependences of L, μ, and w_* on m are discussed, since they are recently developed and have the great advantage in scaling the atmospheric ability of pollutant dispersion.

(1) *The impact of m on L*

Since the value L is negative during daytime unstable condition,therefore,the decrement of $1/L$ is corresponding to a more unstable atmosphere. Based on the discussion in subsection 1, the influence of m on u_* can be neglected. Then,Eqs.(3) and (6) can be expressed as

$$1/L = \alpha_{L1} T_* + \alpha_{L2} q_* \tag{19a}$$

where $\alpha_{L1} = gk / Tu_*^2$ and $\alpha_{L2} = 0.61gk / u_*^2$.

Equations (16),(17) and (19a) determine the relationship between $1/L$ and m. For most of atmospheric conditions the second term of the RHS (right hand side) in Eq.(19a) can be neglected as compared with the first term of RHS in (19a) as shown by Eqs.(16),(17). Therefore, Eq.(19a) is simplified as

$$1/L = \alpha_{L1} T_* \tag{19b}$$

The atmospheric stabilities over saturated soil classified by $1/L$ are about 56% and 60% of those over soil with $m=0.0$ and $m=0.01$, respectively, computed based on Eqs.(16) and (19b) with values of $u_* = 0.2\mathrm{m/s}$, $T_v = 290$ K (1300 LST).

The theoretical estimation is supported by Fig.5, which is computed from the numerical model simulations. Figure 5 presents that the variation of $1/L$ with time for different values of m is very similar to that of T_* as compared Fig.5 with Fig.2. The ratio of $1/L$ for $m=1.0$ with that for $m=0.01$ is 0.67, which is in agreement with the theoretical derivation where the ratio is 0.60.

(2) *The impact of m on μ*

The more unstable atmosphere during daytime PBL is corresponding to smaller values of μ according to Eq.(4). Since the influence of m on u_*, as discussed above, can be neglected as compared to the influence of m on T_* (i.e. u_* dependes on atmospheric condition and surface roughness), the following relation will be resulted in:

$$\mu = a_\mu T_* \tag{20a}$$

The parameterization of μ with m can be derived by substituting Eq.(16) into Eq.(20a); then we have

$$\mu = d_\mu + C_\mu m^b \tag{20b}$$

where $C_\mu = \alpha_\mu C_h$, $\alpha_\mu = gk / (T_v f u_*)$ and $d_\mu = \alpha_\mu d_h$, the value of $\alpha_\mu' = 676$ with $u_* = 0.2$ m/s, $T_v = 290$ K. The value of μ over saturated moisture soil is about 56% of that over absolute dry soil.

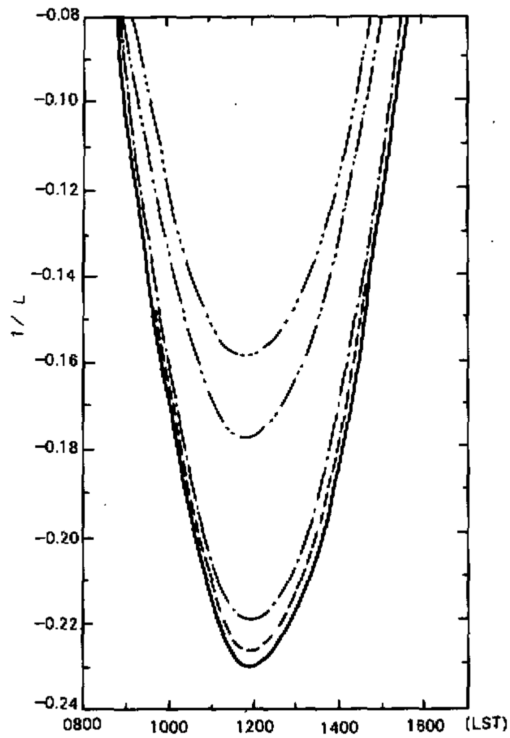

Fig.5. The same as in Fig.1, except for the reciprocal of the M–O length.

(3) *The impact of m on* w_*

The impact of m on w_* can be derived by substituting Eqs.(10,11) into Eq.(5), which results in:

$$w_*(m2)/w_*(m1) = [(u_* T_*)_{m2}/(u_* T_*)_{m1}]^{1/2} \tag{21a}$$

As a first approximation, u_* =constant is assumed as discussed in subsection 1. Then Eq.(21a) can be simplified as

$$w_*(m2)/w_*(m1) = (T_*(m2)/T_*(m1))^{1/2} \tag{21b}$$

Equations (16) and (21b) describe the relation of w_* with m. The value of w_* over absolutely dry soil is about 77% higher than that over saturated moisture soil.

IV. CONCLUSION

The present study has investigated the impact of soil moisture availability on dispersion–related turbulent characteristics and the atmospheric stability classified by Monin–Obukhov length, Kazanski–Monin parameter and convective velocity scale. The major conclusions of those evaluations are follows:

(1) The impact of soil moisture availability on surface friction velocity is insensitive. The influence of soil moisture availability on temperature characteristic scale (T_*) is evident. The value of T_* over saturated soil moisture is about half of that over absolutely dry soil. The impact of soil moisture availability on atmospheric humidity characteristic scale (q_*) is very strong: the value of q_* over saturated soil is nearly two orders of magnitude of that over absolutely dry soil.

(2) The planetary boundary layer height during daytime decreases when the soil moistore availability increases. The PBL's height over absolutely dry soil is about 40% higher than that over saturated soil.

(3) The soil moisture availability exerts a sensitive influence on atmospheric stabilities classified by Monin–Obukhov length (L), Kazanski–Monin parameter (μ) and convective velocity scale (w_*) by increasing the atmospheric stability as increasing the soil moisture availability: the atmospheric stability characterized by L^{-1}, μ and w_* increases about 30%–40% over saturated soil as compared to the one over absolutely dry soil.

(4) The impact of soil moisture availability on dispersion–related atmospheric stability is performed through the impact of soil moisture on temperature characteristic scale (T_*) as shown by Eqs.(19b), (20b) and (21b).

This study is available for LASG for its providing us computer time.

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