Effective Dry Deposition Velocities for Gases and Particles over Heterogeneous Terrain

JIANMIN MA AND S. M. DAGGUPATY

Air Quality Modeling and Integration Research Division, Meteorological Service of Canada, Downsview, Ontario, Canada

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ABSTRACT

Dry deposition velocities of gases and particles are highly dependent on surface type. In a numerical model, each grid cell may contain multiple surface types, each with a different deposition velocity. Therefore, some kind of averaging technique generally is used to compute the average of the subgrid-scale deposition velocities within a grid cell. In this paper, effective surface parameters are suggested to relate the mean properties of concentration and wind speed to the mean surface fluxes. An effective deposition velocity is computed subject to these effective surface parameters and a weighted-average technique. This effective deposition velocity is compared with an alternate weighted-average deposition velocity that has been used widely in numerical air quality models. For particles, the effective deposition velocity can be significantly different from the weighted-average deposition velocity. For some gases, for which biological factors often control the deposition process, the difference between these two average deposition velocities can still be distinguished for typical gases and surface properties.

1. Introduction

For a realistic representation of the dry deposition process in meso-, regional-, and global-scale numerical models, with a typical horizontal grid resolution of about 10–100 km, it is necessary to consider the effects of underlying subgrid-scale surface properties on surface parameters and fluxes, because the earth’s surface rarely is homogeneous within a grid square. Rather, the land surfaces are characterized by different land uses, vegetation types, and soil types. To parameterize the effects of this land surface subgrid-scale heterogeneity, a variety of averaging techniques has been developed and applied in climate and regional numerical models to calculate surface momentum, heat, and moisture fluxes. These techniques combine the parameters for different land covers into single effective values for application in numerical models (e.g., Avissar and Pielke 1989; Kosta and Suarez 1992; Pitman 1994).

To model the dry deposition flux over heterogeneous terrain correctly in a numerical model, a specific parameterization for dry deposition velocity over various homogeneous subgrid areas of the model grid cells has to be developed to account for the effects of land surface heterogeneity. This development requires grid-representative (or effective) surface parameters, fluxes, and biological factors that dominate the dry deposition process over heterogeneous terrain. For particles, turbulent and molecular exchange processes control the dry deposition velocities. For gases such as sulfur dioxide (SO2) and ozone (O3), the vegetation characteristics, for example, leaf area index (LAI) and stomatal resistance, often play a more important role in the determination of the dry deposition velocities than do turbulent and molecular exchange (Pleim et al. 1984; Padro 1993). Here, the effective surface parameters (such as the friction velocity, aerodynamic roughness length, and Obukhov length) must represent a gridcell value and account for the dynamical influence of subgrid-scale heterogeneity. These effective surface parameters are evaluated using the blending-height concept (e.g., Wood and Mason 1991). Determination of the grid-representative dry deposition velocity can be accomplished by first evaluating effective aerodynamic parameters, such as the effective roughness lengths and the friction velocity introduced in the next section, and then using these effective values to calculate an effective dry deposition velocity.

The gridcell average of deposition velocity must be computed as a weighted average based on fractional land use categories within individual grid cells. This approach computes the deposition velocity for each subgrid “patch” in terms of land use categories and then averages them, namely,

\[ \overline{v_d} = \sum_i f_i v_d', \]

in which individual aerodynamic resistances are computed for each land type, where \( f_i \) is the gridcell frac-
tional area covered by the patch $i$ with the dry deposition velocity $v_{di}$. This approach has been used widely in numerical modeling of dry deposition velocities for different gaseous species and particles, for example, by Sheeh et al. (1979), Walcek et al. (1986), and Padro (1993).

Because each land type has a characteristic surface roughness and friction velocity, wind speed may deviate significantly from a grid-averaged value of each of these quantities. In reality, landscape variations often result in spatial gradients in surface momentum, heat, and concentration fluxes. An extensive body of European literature (e.g., Erisman et al. 1997) has shown that areas on the upwind side of forest plots receive considerably more particulate dry deposition than do other areas of the same plot. This fact was noticed by Walcek et al. (1986) in their study. They computed individual aerodynamic resistances for each land type encountered within an 80-km² area by assuming the product of wind velocity and the friction velocity to be a constant at the altitude where deposition velocity is calculated. To utilize this assumption, they also estimated a logarithmic-averaged aerodynamic roughness length.

In this study, the focus will be on the atmospheric exchange mechanisms responsible for the surface fluxes of momentum and species concentration. The objective of this paper is to develop an “averaging technique” for calculating an effective dry deposition velocity by applying a resistance approach to derive a total effective dry deposition velocity. In the next section, we shall demonstrate the process of deriving the spatially averaged effective parameters. We develop grid-representative parameters by using spatial averaging over the grid area. The aerodynamic parameters involved in the formulas are derived through the blending-height/height-scale concept (discussed in section 2d), and these are superscripted with “eff,” meaning effective. It is noted that the effective parameters are not derived by area averaging over the grid square. Rather, following Taylor (1987) and Mason (1988), we take spatial averaging of Eq. (3),

$$\langle F_d \rangle = \frac{k(u_{di})^{1/2}(c_i - \langle c \rangle)}{\ln \frac{z}{z_{0i}}} - \psi_c \left( \frac{z}{L_{eff}} \right),$$

so as to relate the mean concentration above an anticipated height scale (see section 2d for detail) with the mean surface concentration and mean surface concentration fluxes. In Eq. (4), $z_{0i}$ is the effective roughness length for concentration, $L_{eff}$ is the effective Monin–Obukhov length (m), and $\psi_c$ will be given later. From Eqs. (2) and (4), we have

$$u_{di}^{eff} = \frac{k(u_{di}^{eff})^{1/2}}{\ln \frac{z}{z_{0i}}} - \psi_u \left( \frac{z}{L_{eff}} \right).$$

Equation (5) shows that $u_{di}^{eff}$ from the surface exchange is determined entirely by the surface boundary layer parameters related to the nature of the transfer process in the atmosphere. Equation (5) is presented here merely to demonstrate the process of deriving the spatially averaged deposition velocity. In the next section, we shall apply a resistance approach to derive a total effective deposition velocity $u_{di}^{eff}$.

Note that, for the areally averaged friction velocity, $\langle u_{di}^{eff} \rangle \neq (\langle u_d \rangle)^2$ (Taylor 1987; Claussen 1990). Accordingly, the grid-representative wind velocity $u$ at a height $z$ ($u_z$) in the surface boundary layer follows from Wood and Mason (1991),

$$\langle u_z \rangle = \frac{\langle u_{di}^{eff} \rangle^{1/2}}{\kappa} \ln \frac{z}{z_{0i}} - \psi_u \left( \frac{z}{L_{eff}} \right),$$

where $z_{0i}^{eff}$ is the effective roughness length for momentum, and $\psi_u$ will be described later. Equation (6) can be used to determine the effective values of those surface parameters defined above. The estimate of the effective surface parameters will be described in section 2d.
b. Effective dry deposition velocity for gases

After the approach described by Wesely and Hicks (1977) and Hicks et al. (1987), the dry deposition velocity for gases can be written as

\[ u_d = \frac{1}{R_a + R_b + R_c}, \] (7)

where \( R_a \), \( R_b \), and \( R_c \) denote the bulk aerodynamic, quasi-laminar sublayer, and canopy (surface) resistances, respectively. Because, for most gases, the surface resistance is larger than the other two resistances, the averaging method expressed by Eq. (2) may not be appropriate and may lead to an incorrect computation of deposition flux. In this study, we will use the averaging technique given by Eq. (1), but we propose to apply a blending-height and effective roughness length approach to compute the aerodynamic resistance in Eq. (7). Over each subgrid surface type, the aerodynamic resistance can be estimated by

\[ R_a = \frac{\ln \left( \frac{z}{z_0} \right) - \psi \left( \frac{z}{L} \right)}{\kappa (u_b^*)^{1/2}}. \] (8)

Instead of computing the friction velocity \( u_g \), at height \( z \) over each surface type, we compute the averaged surface stress at the blending height, which is defined as the height scale for momentum transfer over roughness changes for which the flow changes from equilibrium with a local surface to being approximately independent of horizontal position. In other words, below the blending height, wind speeds are in equilibrium with each surface type (implying that the wind profiles are logarithmic), but, above the blending height, wind speeds are approximately uniform over all surface types; the spatially averaged mean flow profile can be expected to be a logarithmic profile with a roughness length equal to \( z_{\text{eff}} \). At this height scale, the average turbulent resistance matches the average wind speed. This approach will handle effectively the influence of the spatial gradients in surface momentum flux resulted from subgrid-scale landscape variations on computations of the aerodynamic resistance. From Eq. (6), we have

\[ u_g = \frac{R_a}{(u_b^*)^{1/2}} \left[ \ln \left( \frac{L_{\text{eff}}}{z_{\text{eff}}} \right) - \psi \left( \frac{L_{\text{eff}}}{L} \right) \right], \] (9)

where \( L_{\text{eff}} \) is the blending height.

For a homogeneous surface, the quasi-laminar resistance can be parameterized as (Wesely and Hicks 1977)

\[ R_b = \frac{5}{u_g (D/\nu)^{2/3}}, \] (10)

where \( \nu \) is the kinematic viscosity of air, and \( D \) is the molecular diffusivity of the gas of interest. For each individual subgrid-scale surface type, in analogy to Eq. (8), the quasi-laminar resistance \( R_b \) can be computed by replacing \( u_g \) by \( (u_b^*)^{1/2} \).

The surface resistance, however, depends primarily on the nature of the surface and the characteristics of the depositing gas. The investigation of Bonan et al. (1993) shows that the influences of subgrid-scale heterogeneity in the LAI, stomatal resistance, and soil moisture on land–atmosphere interactions in a climate model are significant. The grid-averaged deposition velocity is estimated by substitution of Eqs. (8)–(10) [replacing \( u_g \) in Eq. (10) by \( (u_b^*)^{1/2} \)] into Eq. (7), and then Eq. (1) is used to find the average deposition velocity. To distinguish this averaged deposition velocity from the WBCW model (with no blending-height approach to be used), we refer to this grid-averaged deposition velocity as the effective deposition velocity, designated by \( \nu_{\text{eff}} \).

c. Effective dry deposition velocity for particles

A number of models for computing the dry deposition velocity of particles have been proposed (Ruijgrok et al. 1995). Though the formulation of the dry deposition velocity for particles is different from that for gases, one can derive an effective dry deposition velocity for particles in the same fashion as described in the last section.

The effective dry deposition velocities for particles have been generated using two different models (Slinn 1982; Hicks et al. 1987; Venkatram et al. 1988; Wesely et al. 1985). These two models use different formulations for turbulent and molecular exchange and often produce different dry deposition velocities (ENSR 1993). The first one was suggested by Slinn (1982), Pleim et al. (1984), and Hicks et al. (1987) based on a resistance scheme and considering the influence of particle settling velocity. This model is appropriate for particle diameters range from \( 10^{-3} \) to \( 10^{2} \) μm. Over each individual subgrid surface, the dry deposition velocity is defined as

\[ \nu_{di} = \frac{1}{R_{\text{ms}} + R_{si} + R_{bi} u_s + \nu_s}, \] (11)

where \( \nu_s \) is the gravitational settling velocity given by the Stokes law and is determined primarily by particle size and density (e.g., Pleim et al. 1984). There exists no spatial average of \( \nu_s \). Over each individual subgrid-scale surface, \( R_{\text{ms}} \) in Eq. (11) can be computed from Eqs. (8) and (9). For particles, special attention is paid to the quasi-laminar resistance \( R_b \), which is usually the controlling factor. Slinn (1982) derived a simplified expression for \( R_b \) that is inversely proportional to the surface stress. In keeping with Slinn (1982) but using the same averaging approach as is used in Eq. (10), we can define
Based on this model, the dry deposition velocity over each subgrid-scale surface can be written as

\[ R_{ai} = \frac{1}{(u_m^* L_{ai})^{1/2}} E_j, \]  

(12a)

where \( E_j \) is the collection efficiency, defined as

\[ E_j = Sc^{-2/3} + 10^{-3/St}, \]  

(12b)

where \( Sc \) is the Schmidt number and \( St \) is the Stokes number [\( St = (\nu/g)(\alpha^2_{1}/\mu) \)], \( \mu \) is the viscosity of air (\( \approx 0.15 \text{ cm}^2 \text{s}^{-1} \)), and \( g \) is the acceleration from gravity.

The second model is an empirical parameterization of dry deposition velocity from Wesely et al. (1985) and Ruijgrok et al. (1995). This model is based on flux measurements of sulfate particles over grass and can be used for estimation of the dry deposition velocity of small-size particles with diameters in the range 0.05–1.0 \( \mu \text{m} \).

Based on this model, the dry deposition velocity over each subgrid-scale surface can be written as

\[ v_{de} = \frac{1}{R_{de} + R_{ds}}, \]  

(13)

where \( R_{ds} = v_{ds}^{-1} \), and \( v_{ds} \) is the surface deposition velocity. Similar to Eq. (12), \( R_{ds} \) is given by

\[ R_{ds} = \frac{1}{\alpha_i (u_m^* L_{ai})^{1/2}}, \]  

(14)

where the nondimensional coefficient

\[ \alpha_i = \begin{cases} 0.002 & L_i \geq 0, \\ 0.002[1 + (-300/L_i)^{2/3}] & L_i < 0, \end{cases} \]

with \( L_i \) being the Obukhov length at a subgrid point. The aerodynamic resistance over each surface type \( R_{ai} \) is computed from Eqs. (8) and (9). After the dry deposition velocities expressed by Eqs. (11) and (13) associated with Eqs. (8)–(10), (12), and (14) are obtained, we again use Eq. (1) to generate the effective deposition velocities for particles.

d. Computational framework of effective values of surface parameters

Mason (1988) has provided a heuristic model to evaluate the effective roughness lengths for a neutral atmosphere. Ma and Daggupaty (1998) have extended this approach to a stratified boundary layer, which allows for different atmospheric stability (from stable to unstable) conditions to be involved in the estimation of blending height and other effective surface parameters. Using Eqs. (4) and (6), the effective roughness lengths for momentum and concentration are given by

\[ \ln \left( 1 + \frac{L_{bc}}{z_L} \right) = \psi_m \left( \frac{L_{bc}}{L} \right) \]  

(15a)

\[ \ln \left( 1 + \frac{L_{bc}}{z_L} \right) = \psi_c \left( \frac{L_{bc}}{L} \right) \]  

(15b)

where

\[ \psi_m = \int_{z_{L}}^{Z} [1 - \phi_m(x)] \, dx/L, \] \[ \psi_c = \int_{z_{L}}^{Z} [1 - \phi_c(x)] \, dx/L, \]

with \( Z = z/L \) and \( Z_{L} = z_{L}/L \); the nondimensional wind shear and concentration gradient are written as (Nieuwstadt and Duynkerke 1996)

\[ \phi_m(z/L) = \left( 1 - 16 \frac{z}{L} \right)^{-1/4} \text{ for } z/L < 0, \]

\[ \phi_c(z/L) = \left( 1 - 14 \frac{z}{L} \right)^{-1/2} \text{ for } z/L < 0, \]

and (Högström 1996)

\[ \phi_m = 1 + 5.3 \frac{z}{L}, \quad \phi_c = 1 + 8 \frac{z}{L} \text{ for } z/L > 0. \]

In Eq. (15b), \( L_{bc} \) is the height scale for pollutant transfer over heterogeneous surfaces. Further, \( L_h \) and \( L_{bc} \) are expressed as follows (Ma and Daggupaty 1998):

\[ L_{h} = \frac{\alpha_s \kappa \pi}{\phi_m(L_h/L)} \left( \frac{u_\theta}{u(L_h)} \right) \lambda \]  

(16a)

\[ L_{bc} = \frac{\alpha_s \kappa \pi}{P_r(L_{bc}/L)} \left( \frac{u_\theta}{u(L_{bc})} \right) \lambda, \]  

(16b)

where \( P_r = \phi_c/\phi_m \) is the turbulent Prandtl number, \( \lambda \) is the horizontal scale or wave length of the surface heterogeneities, and the coefficients \( (\alpha_s, \alpha_i) = (0.125, 0.031) \). Equation (16) indicates that both height scales are functions of the horizontal scale \( \lambda \) of the heterogeneous terrain.

Equations (15) and (16) show that the height scales and effective roughness lengths have to be resolved with an iterative procedure. The following iterative procedure is used to obtain these height scales and effective values for the roughness lengths, Obukhov length, and surface stress.
1) Calculate $L_a$ and $L_{bc}$ from Eqs. (16a) and (16b) by giving initial values of $z_0^\text{eff}$, $z_0^\text{eff}$ and $L^\text{eff}$.
2) Calculate $u(L_a)$ from the flux–gradient wind profile relationship.
3) Calculate $z_0^\text{eff}$ and $z_0^\text{eff}$ from Eq. (15).
4) Evaluate the stress $\langle u^2 \rangle$ from Eq. (9) and use the new value of $\langle u^2 \rangle$ to calculate $L^\text{eff}$.
5) Repeat steps 1–4 until the procedure converges.

After the effective surface parameters are obtained, we can estimate $u^\text{eff}_{ub}$ from Eq. (5) and $R_{ai}$ and $R_{bi}$ from Eqs. (8)–(10) [replacing $u_u$ by $\langle u^2 \rangle^{1/2}$ in (10)] for gases and Eqs. (8) and (12) for particles. Last, Eq. (1) is used to obtain the effective dry deposition velocities.

We wish to point out that, in the establishment of the model for dry deposition velocity, we do not consider so-called feedback effects, so the model does not account adequately for variations of concentration that occur over a surface with little uptake versus a surface with a large uptake rate.

### 3. Results and discussion

The two averaging methods are examined and compared in this section. In the application of the WBCW model, following Walcek et al. (1986), we assume the product of wind speed and the friction velocity $u_u u_a$ to be a constant for all surface types at 20-m height, and we compute values of the surface parameters (among which the friction velocity is the most important parameter) and individual aerodynamic resistances for each land type. After the deposition velocities for each land type are calculated, each subgrid location is obtained and Eq. (1) is used to find a weighted-averaged $u_c$. Three gases ($\text{SO}_2$, $\text{O}_3$, and nitric acid ($\text{HNO}_3$)) are selected. Although no specific particle type has been selected, the particle deposition calculations are performed over the particle size range 0.001–10 $\mu$m. Measurements and computations indicated that deposition velocities of trace gases and particulates to surfaces have considerable variability in terms of solubility, diffusivity, and reactivity. Much of this variability can be accounted for by meteorological variability and differences in surface characteristics. For the trace gases $\text{O}_3$ and $\text{SO}_2$, the dry deposition velocity over land are controlled largely by surface uptake; for $\text{SO}_2$ over water surfaces, $\text{HNO}_3$, and particles, the dry deposition velocities are controlled by turbulent transfer. The values of the surface parameters used in all computations are shown in Table 1. These surface parameters include surface roughness length, surface resistance, and land use category (LUC), which are obtained and computed from the Simple Biosphere Model (Sellers et al. 1996) and Wesely (1989). In Table 1, cases 1, 2, and 4 are extreme cases of rough–smooth surfaces. Case 3 describes smooth–smooth surfaces. Case 1, cases 1, 2, and 4 are extreme cases of rough–smooth surfaces. Case 3 describes smooth–smooth surfaces. Case 3 describes smooth–smooth surfaces. Case 3 describes smooth–smooth surfaces. Case 3 describes smooth–smooth surfaces. Case 3 describes smooth–smooth surfaces.

#### 3a. Aerodynamic transfer

Because direct measurements of dry deposition velocities over heterogeneous terrain are not available, the
computed effective dry deposition velocities cannot be evaluated precisely. A qualitative evaluation, however, can be made by comparing the results obtained by the WBCW model and the method presented in the last section. Figure 1 shows the weighted average \( \overline{v}_{d,\text{turb}} \) computed by Eq. (1) and \( v_{d,\text{turb}}^{\text{eff}} \) by Eq. (5). Surface roughness lengths are chosen from case 3, and the effective roughness lengths are calculated by the iterative method presented in section 2d. The computation is performed from 0.22 to 100 m in the vertical with an averaged wind speed of 4.6 m s\(^{-1}\). It has been found (Wood and Mason 1991; Ma and Daggupaty 1998) that the effective roughness length for heat transfer is less than the logarithmic average of the local roughness lengths for heat transfer. This finding also applies to mass transport of pollutant concentration. Consequently, this would tend to decrease \( v_{d,\text{turb}}^{\text{eff}} \), in terms of Eq. (5), as we have observed from Fig. 1, which for all the boundary layer stratification shows \( v_{d,\text{turb}}^{\text{eff}} < \overline{v}_{d,\text{turb}} \). The largest difference between \( v_{d,\text{turb}}^{\text{eff}} \) and \( \overline{v}_{d,\text{turb}} \) appears at levels close to the ground (except for the stable case). The relative errors of \( v_{d,\text{turb}}^{\text{eff}} \) and \( \overline{v}_{d,\text{turb}} \) are about 10% at \( z = 0.22 \) m, and for \( z > 20 \) m the two deposition velocities are almost identical. This result is expected, because in Eq. (5) the effective roughness length is derived from the concepts of the height scales for momentum and concentration transfer over roughness changes. These height scales are, in general, about 1/100 of the horizontal scale of the surface heterogeneity, implying that the height scales are about 10–20 m for the selected horizontal scale of 1 km of the surface heterogeneity (Wood and Mason 1991; Ma and Daggupaty 1998). Above these vertical scales, the signatures of individual surface patches vanish according to the definitions of the height scales, so that the effect of aerodynamic transfer processes over inhomogeneous surfaces cannot be identified above these height scales.

b. Gases

For gases, the total dry deposition velocity is determined by atmospheric resistance, quasi-laminar resistance, and surface resistance. Because the surface resistance for most gases usually is greater than the quasi-laminar resistance, it is regarded as the controlling process. For smooth surfaces with small values of roughness length (except for the desert–water case) and small difference of the surface resistances over different surfaces, we can deduce that \( v_{d}^{\text{eff}} \) will not differ significantly from \( v_{d} \). For a surface covered by a pine forest, measurements indicate that about half the total resistance is due to atmospheric resistance (Pleim et al. 1984). It was found that the effective roughness length is weighted toward the larger roughness elements and is greater than the simple logarithmic average of the local roughness lengths for momentum (Wood and Ma-
As a result, the areally averaged surface stress is also weighted toward the same roughness elements according to the surface similarity theory. These arguments actually imply, from an aerodynamic point of view, that the resistance from aerodynamic transfer, defined by Eq. (8), will produce relatively larger values of the effective deposition velocity than that given by the WBCW model.

Figure 2 shows $v_{d}^{eff}$ and $\bar{v}_{d}$ values for $O_3$ and $SO_2$ for three different LUCs at 20-m height for the cases 1, 2, and 4 shown in Table 1. The differences between the two dry deposition velocities are due to the different techniques, as we discussed before, and change with underlying surfaces or LUCs. Because for these gases the surface resistance dominates the dry deposition velocities for selected LUCs, the distinction between the two deposition velocities would not be very evident. For example, in case 1 we can obtain the averaged sur-
face resistance for SO$_2$ as 4.7 s cm$^{-1}$, and from the model under neutral conditions with a wind velocity of 4 m s$^{-1}$ we can obtain the averaged aerodynamic and quasi-laminar resistances, given by Eqs. (8) and (10), as 0.206 and 0.103 s cm$^{-1}$, respectively. This result indicates clearly that the averaged dry deposition velocity would not be very sensitive to meteorological conditions. In case 4, $v_d^{\text{eff}}$ and $\overline{v}_d$ for SO$_2$ increase considerably with the increase of wind velocity (Fig. 2b). Because for SO$_2$ the surface resistance is zero over water, the aerodynamic resistance becomes dominant. The values of $v_d$ are consistently large because of this water solubility and obviously dominate the spatial average of the dry deposition velocity at this smooth-rougher surface. In this case, the large values of the dry deposition velocities for SO$_2$ over water (where the surface resistance vanishes) dominate the averaged deposition velocity over the mixed forest–water surfaces. In all cases, $v_d^{\text{eff}}$ tends to have similar profiles of $\overline{v}_d$ but has larger values than those for $\overline{v}_d$. This result becomes more evident for the cases in which the roughness lengths at two LUCs have significant gradients, because the averaged momentum flux is weighted toward rougher surfaces, which would lead to a smaller averaged aerodynamic resistance. Because the surface resistances control the total resistances, the deposition velocities (especially for O$_3$) derived from the two average methods are almost identical.

For HNO$_3$, the surface resistance is almost zero and is neglected in this study. Thus the aerodynamic resistance is the controlling resistance to the dry deposition process. The effective deposition velocities are larger than those given by the WBCW model (Fig. 3). In contrast to O$_3$ and SO$_2$, the deposition velocities produced from both techniques are very sensitive to meteorological conditions. Figure 3 shows that $v_d^{\text{eff}}$ and $\overline{v}_d$ increase linearly with wind velocity.

From the above analyses, the subgrid-scale land use variations will have the most pronounced effect on averaged deposition velocity under conditions when there is significant variation in land type in the averaging area. Meteorological factors will have their most pronounced effect on deposition velocities when the surface resistance to pollutant uptake is small. This result has been shown for nitric acid and sulfur dioxide dry deposition velocities over specific surface cover types, whereas the effect of subgrid-scale land use variations on the average dry deposition velocity has been shown for cases 1 and

![Graph showing deposition velocity vs. wind velocity for HNO$_3$.](image)
For O₃, because the surface resistance can lead to relatively weak surface stress and strong aerodynamic and quasi-laminar resistances as well as remarkable differences in the deposition velocities between the two surfaces. The total resistance over desert would be far greater than that over water. The deposition velocity over desert is consequently far less than that over water. In contrast to SO₂, the surface resistance for O₃ is much larger over water than over desert. The dry deposition velocities from the two models are shown in Fig. 4. It can be seen that both WBCW-model and effective deposition velocities for SO₂ are very sensitive to meteorological conditions and increase drastically with the increase of wind speed. This characteristic is, in fact, consistent with the variation of the deposition velocities over the water surface where, because the surface resistance is zero, the deposition velocity is almost directly proportional to the friction velocity. In other words, the deposition velocity over the water surface dominates the spatially averaged deposition velocity over the water–desert area. It is clear that the values of the effective deposition velocities are greater than the WBCW-model deposition velocities, indicating that the averaging technique to produce the effective deposition velocity gives much weight to the rougher surface for an averaged region from the point of view of the aerodynamic transfer. For O₃, because the surface resistance over the water surface dominates the total resistance, the two averaged deposition velocities have similar profiles.

c. Particles

First, we present the model result [Eq. (11)] based on Pleim et al. (1984). A comparison of \( v_{eff} \) and \( \bar{v}_\alpha \) at a height of 20 m for the selected friction velocities and particles with unit density and different sizes for case 1 is illustrated in Fig. 5. It can be seen that the greatest differences between \( v_{eff} \) and \( \bar{v}_\alpha \) occur for the small and large particles. For the medium-size particles, because the deposition velocity itself is very small, the deviation of the two models also is very small. Because the effects of biological factors on the dry deposition velocity and the dependence of it on foliar density are not taken into account for particles in this study [an improvement might be gained by accounting for the dependence of \( R_b \), defined by Eq. (12), on leaf dimension and LAI such as is described by Pleim et al. (1984)], each term of the denominator in Eq. (11) contains the information of the aerodynamic parameters. For very small and large particles, \( R_a \) and \( R_b \) in Eq. (11) make almost equal contributions to \( v_{eff} \); the effective values of aerodynamic parameters play a major role in measuring differences between \( v_{eff} \) and \( \bar{v}_\alpha \). Although, for the medium-size particles, \( R_b \) dominates the dry deposition process, Brownian (for particles) diffusivities control the magnitude of both \( v_{eff} \) and \( \bar{v}_\alpha \). It can be seen from Fig. 5 that, for small- and medium-size particles, \( v_{eff} \) is greater than \( \bar{v}_\alpha \), but, for large-size particles, \( v_{eff} \) is smaller than \( \bar{v}_\alpha \). Both \( v_{eff} \) and \( \bar{v}_\alpha \) are highly dependent on the variation of the friction velocity.

The effective dry deposition velocity, calculated according to Wesely et al. (1985) for neutral conditions, is shown in Fig. 6. It illustrates the variation of the two deposition velocities with the friction velocity at the 20-m height for case 1. Equation (13) indicates that the deposition velocities are determined entirely by meteorological conditions, so the analysis for the results of the deposition velocities for HNO₃ also is applicable for particles. As can be seen from Fig. 6, both deposition velocities increase linearly with the friction velocity. In analogy to HNO₃, \( v_{eff} \) produced by Eq. (13) is greater than \( \bar{v}_\alpha \), and the difference increases with increasing friction velocity. The variation of the effective and the WBCW-model deposition velocities with stability in case 4 is displayed in Fig. 7. For stable stratification, \( v_{eff} \) is slightly larger than \( \bar{v}_\alpha \), but both remain almost unchanged. For unstable stratification, however, \( v_{eff} \) is larger and presents a faster increase with instability than \( \bar{v}_\alpha \). As shown in Fig. 7, though the effective friction velocity increases with instability, one still can identify the sensitivity of the effective deposition velocity to the unstable or convective conditions.
4. Concluding remarks

A technique to determine an effective dry deposition velocity over heterogeneous terrain in a mesoscale or regional-scale numerical model has been presented. The technique was applied for several selected gases and particles in different ideal cases and compared with the WBCW model. The results for gases suggested that, when the surface resistance (which dominates the total resistance to the transport of gases over an averaged area) does not change considerably at subgrid points within the averaged area, the difference between the two techniques is not significant. Therefore, in such cases, the dry deposition velocities over heterogeneous terrain might be fairly insensitive to the averaging method.

The comparison of the two models indicated that, for the smooth–rougher surfaces, the WBCW model might, from an aerodynamic point of view, produce smaller values of the dry deposition velocity than the effective deposition velocity. Hence, for those gases affected only weakly by surface resistance, such as HNO$_3$, the values of the effective deposition velocity derived from the averaging technique developed in this study are greater than that from the WBCW model and are sensitive to meteorological conditions.

For particles, two models based on Pleim et al. (1984) and Wesely et al. (1985) were tested and compared with the WBCW model. With smaller and larger particle sizes, in the first model [Eq. (11)] the aerodynamic and quasi-laminar resistances make almost equal contributions to the total resistance to the transport of particles. There are considerable differences between the averaged dry deposition velocities produced by these two techniques. Because the aerodynamic processes dominate, the second model [Eq. (13)] predicts relatively large values of the deposition velocity in comparison with the WBCW model, agreeing with the fact that the surface stress is weighted toward rougher surfaces.

We have said very little about practical applications of the two models. One difficulty is that fewer field and laboratory measurements exist, so that it is not possible to compare modeled and measured spatially averaged dry deposition velocities. An investigation is being un-
dertaken to use the average method for dry deposition velocity developed in this study in a numerical transport model to check its effect on concentrations and then to compare the results with measurements. The results will be reported in a forthcoming paper. Physically, the differences between the two models are due essentially to the processes associated with aerodynamic parameters in the surface layer.

To seek a more rigorous solution, it may be desirable to use the new averaging technique to compute the effective dry deposition velocity over grid squares in a mesoscale or regional numerical model. This method requires estimates of effective values of aerodynamic parameters and the blending height for which the grid-square-averaged wind speed, temperature, and other relevant scalars must be computed. This necessity increases the computational cost but may lead to a more accurate estimate of the grid-scale concentration fluxes and the mean dry deposition velocity over a heterogeneous area.

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