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On Estimating Dry Deposition Rates in Complex Terrain

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ABSTRACT

In complex terrain, horizontal advection and filtration through a canopy can add substantially to the vertical diffusion component assumed to be the dominant transfer mechanism in conventional deposition velocity formulations. To illustrate this, three separate kinds of terrain complexity are addressed here: 1) a horizontal landscape with patches of forest, 2) a uniformly vegetated gentle hill, and 3) a mountainous area. In flat areas with plots of trees, the elevation of the standard area-weighted dry deposition velocity will likely depend on the product $hn^{1/2}$, where h is the tree height and n is the number of plots per unit area. For the second case, it is proposed that the standard "flat earth" deposition velocity might need to be increased by a factor like $[1 + R_a/(R_b + R_c)]^{1/2}$. For mountainous ecosystems, where no precise estimate of local dry deposition appears attainable, the actual dry deposition rate is probably bounded by the extremes associated with 1) the flat earth assumption involving aerodynamic, quasi-boundary layer, and canopy resistances as in conventional formulations, and 2) an alternative assumption that the aerodynamic resistance is zero. Such issues are of particular importance in the context of atmospheric loadings to sensitive ecosystems, where the concepts of critical loads and deposition forecasting are now of increasing relevance. They are probably of less importance if the emphasis is on air quality alone, because air quality responds slowly to changes in deposition rates. The issues addressed here are mainly appropriate in the context of air surface exchange that is not controlled by surface resistance (e.g., for deposition of easily captured chemicals such as nitric acid vapor, and perhaps for atmospheric momentum) and for chemicals that have no local sources. It is argued that dry deposition rates derived from classical applications of deposition velocities are often underestimates.

1. Introduction

Deposition from the atmosphere constitutes one of the sinks for air pollution (the sole sink for some pollutants such as particulates), and one of the sources of pollution to the terrestrial, aquatic, and biospheric environments. There is a dominant requirement to get the estimates of these exchange rates right. There is need to consider both wet and dry deposition, the former associated with the precipitation process and the latter with the turbulent exchange of trace gases and small particles and the gravitational settling of larger particles. Science has been concentrating on this task for many decades, with great success in some instances and with marginal success in others. Dry deposition falls into the latter category.

"Dry deposition" has different meanings in different communities. To some workers, it is the rate of transfer of pollutants to the surfaces of receptors on a per unit area basis regardless of whether it is the area of a leaf or of the surface of a building, or whether the surface is horizontal or tilted. To others, it is the rate of deposition of particles from the atmosphere, expressed as the flux across a unit area of the lower boundary of the atmosphere. In the present context, dry deposition is the flux of trace gases and particles from the air to the underlying surface that constitutes a lower boundary to the atmosphere—the atmosphere-to-landscape situation.

There are several reasons for the uncertainty that surrounds the estimation of dry deposition rates, including the undeniable fact that measurement techniques remain problematic. Recent advances in eddy correlation methods (e.g., relaxed eddy accumulation; see Meyers et al. 2006) have allowed a variety of chemical species to be addressed, but the new methodology still suffers the constraints of conventional micrometeorology—it does not measure the flux at the surface itself

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but at some convenient height above it. For the measurement at this convenient height to be the same as the exchange rate at the surface itself, there are stringent criteria that must be satisfied. As is well known in micrometeorology, conditions must be stationary in time. Second, the measurements must take place in horizontally homogeneous surroundings. Third, the measurements must be at a height that is above the region of direct influence of the separate roughness elements that populate the surface and below the height at which fluxes will change with height because of large-scale planetary boundary layer effects [i.e., above $10z_0$ and below about $0.10z_i$ (sometimes $0.07z_i$), where z_0 is the roughness length and z_i is the PBL depth]. All of these constraints are well known. If they are violated, then special efforts must be made to justify the tacit assumption that fluxes are constant with height. Hence, even though experimental techniques have improved considerably, the micrometeorological methodologies remain limited in the way in which they can be used.

Similar problems arise in the description of deposition processes in models. For example, the so-called inferential schemes for estimating dry deposition rates from air chemistry and selected supporting measurements (Hicks et al. 1987) are based on an assumption that conventional "flat earth" micrometeorology applies. In other words, these methods can be expected to yield accurate estimates of dry deposition rates only in the same conditions as constrain the measurement of fluxes by micrometeorology, as mentioned above. In the lack of data against which to evaluate the performance of such models in other circumstances, there must be questions about the accuracy of their predictions. It is the present goal to present arguments relating to the enhancement of dry deposition rates by surface inhomogeneities and to the specification of upper and lower bounds on deposition rates to land areas of varying topographic complexity. The discussion will relate to three specific cases: 1) a flat grassy plain with patches of forest, 2) a gentle hill in uniformly vegetated terrain, and 3) a forested steep mountaintop. In each case the discussion will revolve around considerations of the multiple resistance analogy (see Sheih et al. 1979), originally developed in agrometeorology and now familiar in discussions of air surface exchange in general.

The adaptation of the multiple resistance approach to the dry deposition question has simplified the communication of results and ideas about the interpretation and representation of experimental data. It provides a framework that simulates the relevant processes in an easily understood manner. Modern literature refers to an aerodynamic resistance R_a that describes transfer across the lowest layers of the atmosphere from some convenient height at which concentrations are specified, a quasi-laminar layer resistance R_b that takes molecular and Brownian diffusive factors into account, and a surface or canopy resistance R_c that combines the physical and biological processes associated with the actual capture of the depositing material at the surface itself. In this framework, the effective deposition velocity is then the reciprocal of the sum $R_a + R_b + R_c$. However, in common with all micrometeorological relationships, the multiple resistance framework applies best to a spatially extensive, flat, and homogenous surface. But the need is usually to represent the exchange between the air and a natural landscape, containing vegetation inhomogeneities and complex terrain. Classical micrometeorology and agrometeorology have been able to avoid many of the difficulties that then arise, because much of the focus is on the fluxes of water vapor (LE) and sensible heat H. The sum H + LE is limited by the surface radiation balance, after allowance for the relatively small storage components in the soil and biomass. There is no such constraint on the exchange rates of pollutants (or on the exchange rate of momentum).

It is acknowledged that the problem of complex terrain encompasses far more than the situations considered here. For example, in daytime some hillsides will receive more direct radiation than others, and therefore the sum of sensible and latent heat fluxes from sunlit slopes will be greater than from less sunlit. Other surface fluxes will also be affected. This is of great current importance in estimating spatial averages of carbon dioxide sequestration (e.g., Katul et al. 2006). Whether the spatial averages of strongly sunlit and shaded slopes can be approximated by expressions that relate to a horizontal surface is outside the present scope.

It is also clear that predictions of local air quality will not be greatly affected by changes in local dry deposition rates, for most air pollutants, even if there is a large error in the computation of deposition. Air concentrations of most common pollutants are only slowly affected by changes in the rate of dry deposition from the atmosphere. The importance of the present considerations is more associated with the need to assess the rate of influx of pollutants into ecosystems, and especially the sensitive components of them.

2. Patchy vegetation

Consider an area that contains plots of different vegetation. It is commonly assumed that the spatial average dry deposition rate will be equal to the area-weighted average of the rates to the individual plots (see Ma and Daggupaty 2000). This is clearly an appropriate approximation in the case of cropland growing two kinds of crops of similar physical characteristics, where we can write with considerable confidence that the spatial average flux F is

$$F = A_1 V_{d1} C_1 + A_2 V_{d2} C_2, \tag{1}$$

where A_1 and A_2 are the proportions of the area under vegetation 1 and 2, V_{d1} and V_{d2} are the corresponding deposition velocities, and C_1 and C_2 are the concentrations at the selected reference height above the two kinds of surface. In practice, it is a very close approximation to assume $C_1 = C_2$, since air concentrations respond very slowly to changes in surface fluxes. Hence, the spatial effective deposition velocity is then

$$V'_d = A_1 V_{d1} + A_2 V_{d2}. \tag{2}$$

Extrapolation to a surface with a large number of plots of different vegetation is straightforward; however, these simple concepts break down when the visualization of distinct patches of different vegetation is inappropriate. What should one do, for example, when the plots are of vegetation of different heights, or when there are no distinct plots of different kinds of vegetation but a heterogeneous mix of different species? As an obvious example, how might one address the case of a grassy plain with scattered plots of trees?

There is considerable theoretical as well as experimental evidence that edge effects can be important. Slinn (1982), Beier et al. (1992), and Weathers et al. (2000) focus on the removal of pollutants by forests as air moves through the canopy, a horizontal filtration process rather than vertical diffusion. At the leading edge of a forest, air penetrates the canopy for a distance of many times the tree height (depending on the density of subcanopy growth). There will be an enhancement of deposition near the edge, perhaps with a corresponding deposition "shadow" downwind of the trailing edge. Such effects are readily measured by throughfall methods (see Lovett and Lindberg 1984; Lindberg and Garten 1989), which measure the total deposition in rain collected at a forest floor and relate this to the deposition in rain incident at the top of the canopy, permitting the dry deposition component to be assessed from the difference (material that is previously deposited to the foliage and is washed off by the precipitation). Throughfall studies confirm that dry deposition rates are greatest at the upwind edge of a plot of trees, and decrease exponentially downwind, trending toward a constant rate corresponding to the micrometeorological expectation based on a deposition velocity specific for the canopy in question.

Hicks (1995) addressed the question of how to take such edge effects into account when assessing the rate of removal of air pollutants across a grid cell containing many plots of trees of different configurations. This analysis did not result in an easily applied general rule, but did indicate that for a landscape containing distinct plots of trees in grassland the amplification of spatially averaged dry deposition [on top of the area-weighted result expressed above by Eq. (1)] will depend on a quantity like $hn^{1/2}$, where h is the height of the trees and n is the number of plots per unit area (i.e., within a grid cell of a numerical simulation). The state of understanding is not such that this matter can be explored much further, as yet. It is partially for this reason that aircraft programs have been initiated to measure the exchange of trace gases between the air and natural heterogeneous surfaces (e.g., Crawford et al. 1996). The results of some of these aircraft programs will be considered later.

A key factor in all investigations of the importance of edge effects is the amount of crosswind edge per unit area of consideration. In general, it seems that the matter of edge effects calls for attention using fractals, and that estimating the actual dry deposition rate to a selected area will doubtlessly be site-specific. Equation (2) represents a likely lower bound on the spatial average deposition rate. An upper bound is not obvious, but for areas with many small plots of trees a factor of 2 seems possible.

3. Gentle hills

There have been many studies of flow fields around and across single hills, most addressing the way in which plumes emanating from smokestacks are affected. Wind tunnel studies have proved especially rewarding (e.g., Meroney 1980; Ohba et al. 1990; Snyder 1990). Field studies are far fewer. Bradley (1980) studied micrometeorological variables near the surface of a forested hill rising to some 170 m above the surrounding terrain. The results largely confirmed theoretical expectations (Jackson and Hunt 1975, among others listed by Bradley 1980), with a greatly increased momentum flux (by a factor of about 3–5) to the top of the hill than to the surrounding terrain. We might expect other fluxes to be elevated as well, but the magnitude of the increase will doubtlessly depend on the property being transferred. To first order, the amplification in stress by a factor of 3–5 implies a reduction in R_a to about 20%–30% of its expected value based on conventional micrometeorology. If the intent were to assess the dry deposition to a hilltop, then a corresponding reduction in R_a while holding the other resistances constant would seem to be a good first approximation.

Jackson and Hunt (1975) divide the flow above a gentle hill into two layers—an inviscid outer layer and an inner layer that accommodates the flux changes imposed by the presence of the hill. In this theory, the surface roughness is assumed to be the same across the entire area of consideration. The hill is a bump on the landscape, with the same surface characteristics as its surroundings. Finnigan and Belcher (2004) have extended the approach to a hill with a plant canopy, where local acceleration and deceleration of the wind impose the need to consider advection a major factor affecting air surface exchange. They conclude that "the estimates of air-surface exchange of carbon dioxide (CO_2) need to be corrected for the advective fluxes if they are not to be in serious error," a sentiment that clearly extends to the dry deposition case presently considered. They also question the applicability of the concept of a roughness length, on which much of the existing suite of resistance formations is based. Katul et al. (2006) extend the theoretical analysis for CO_2 to a uniformly forested hill and confirm that advection is a dominating factor in the overall air surface exchange. Ross and Vosper (2005) conclude that "dynamic interaction of forest canopies with the atmosphere over complex terrain . . . can lead to interesting behavior not captured by current roughness length schemes." The familiar multiple-resistance framework is one of those current schemes.

The various analyses of flow across hills substantiate the expectation that conventional eddy fluxes will be lowered in some areas and enhanced in others because of the presence of the local topographic relief. It remains to be seen whether these changes cancel each other out as a spatial average is constructed. In parallel with the considerations of forest edges discussed above, it is evident in the present case that advection needs to be considered as a prime contributor to deposition to natural landscapes.

The discussion above has special relevance to nocturnal situations, for which conventional dry deposition velocity formulations usually predict very small values because of the prevailing atmospheric stability regime (a prediction that is well supported by experiments over conventional micrometeorological field sites). The picture of nocturnal boundary layer behavior that is now evolving entails spatially organized circulations that will preferentially bring air from aloft into canopies at high elevations. The potential repercussions are obvious, but are not yet addressable in terms of either a modified spatially averaged deposition velocity approach or a bounding approach (as advocated here). The interested reader is referred to the work of Monti et al. (2002) and Katul et al. (2006), who address the upslope–downslope circulation issue in considerable detail.

Once again, Eq. (2) would appear to represent an appropriate lower bound. As above, the determination of an upper bound will probably be site-specific, but the discussion above indicates that reducing R_a to 20% of its normal value could well be an appropriate first step.

4. Mountains

There are often concerns about the pollution regimes of sensitive mountaintop ecosystems. Conventional expressions used to describe dry deposition in horizontal and homogeneous terrain are clearly inappropriate. As has been emphasized above, the situation is somewhat like that of a single tree standing on a grassy plain—the dominant mechanisms for transferring pollution from the air are advection and filtration, not vertical diffusion from aloft. Accordingly, the roughness length and displacement height concepts normally associated with micrometeorological depictions of air surface exchange are again of questionable applicability.

For the mountaintop case, C refers to the concentrations in the air impinging on the vegetation canopy. It was already shown, above, that R_a could be reduced to 20%–30% of its flat earth expectation even for a gentle hill. A standard micrometeorological rule of thumb is that separation can occur when the local slope changes by about one part in seven. In fact, this change in slope is now known to depend on the size and shape of the hill and its roughness (Mason and King 1984; Wood 1995). The classical value of about 1:7 corresponds to a small but forested peak. For the lack of something better, we now assume that this 1:7 rule also applies to the case in which the wind is likely to penetrate a mountaintop canopy, disregarding the obvious role of the transparency of the canopy.

There is clearly a range of behaviors that could be expected. One obvious extreme is the situation in which the terrain complexity has no influence on the deposition regime. The opposite extreme is the case in which the aerodynamic resistance is greatly reduced, say to zero. In the first case, transfer by turbulent diffusion may well be an appropriate concept, but the direction might not be defined by the gravitational vertical but rather by the local streamlines. In the second case, the concept of transfer by vertical diffusion is largely inappropriate, because the transfer is by advection and filtration. The foliage-dependent surface resistance R_c will be relevant in both instances. It is not immediately obvious that the quasi-laminar layer resistance R_b will be common, but for the moment equality will be assumed, based on analogy with the simple-surface case and the role of the quasi-laminar layer resistance for individual roughness elements.

These two extremes define bounds (lower and upper: F_l and F_u) on the dry deposition rates:

$$F_l = C/(R_a + R_b + R_c) \quad \text{and} \tag{3}$$

$$F_u = C/(R_b + R_c). \tag{4}$$

Here F_l applies to when the slope is sufficiently gradual that transfer across the plane of the streamlines is by turbulence, and F_u corresponds to the case in which the terrain is so complex that the aerodynamic resistance is not a contributing factor. Neither extreme is likely to be appropriate as a spatial average; the appropriate spatial average might well be better defined as the geometric mean. In this event, it follows that the deposition velocity normally computed would need to be elevated as follows:

$$V'_d = V_d[1 + R_a/(R_b + R_c)]^{1/2}. \tag{5}$$

There are no known data with which to test this result. Moreover, the current considerations refer to complex terrain over which all conventional micrometeorological methods (gradient methods, eddy correlation, etc.) will not work because of the lack of spatial homogeneity. Watershed methods might be appropriate, but the standard methodologies result in the determination of very long averages, over months or more, from which confirming evidence would be hard to extract. Mass balance studies and throughfall approaches might be appropriate (see Weathers et al. 2000), but to resolve the questions that now arise these would be difficult to design and demanding to execute. With no clear experimental path to follow, it seems clear that understanding must build slowly, using whatever innovative methods might be promising.

In the meantime, the same lower bound identified above [based on Eqs. (2) and (3)] would seem appropriate. In this case, an upper bound is clear, resulting from taking R_a to be zero as in Eq. (4).

5. Recent aircraft studies

There have been many aircraft eddy-flux studies that provide relevant guidance. Lenschow et al. (1982) addressed the comparative deposition rates of ozone to forest and ocean, demonstrating the use of aircraft in dry deposition applications. Crawford et al. (1996) and Desjardins et al. (1997) reported on eddy fluxes measured over mixed boreal forest. Their studies concentrated on carbon dioxide exchange. Considerable variation was observed that might be considered an indicator of behavior for other chemical species. Several similar studies were summarized by Hicks (2001), where the focus was on ozone deposition as well as carbon dioxide exchange. A sequence of airborne eddy flux studies started with flights over flat land, with a checkerboard pattern of maize and soybeans, moving on to a study of mixed agricultural land, and ending with rolling terrain with patchy forests. The data presented are mostly based on line averages of eddy fluxes computed along low-altitude 30-km transects. The data confirm that the meteorological fluxes of sensible plus latent heat vary least. Fluxes of carbon dioxide vary much like fluxes of water vapor. Among the quantities reported, the dry deposition of ozone is the most affected by changes in the terrain. This sequence of observations is in accord with expectations: the sum of H and LE is tightly constrained, although the Bowen Ratio H/LE is free to vary widely. CO_2 exchange rates are governed by stomatal processes, in the same way as the exchange of water vapor. Hence, it would be expected that there might be some similarity between the spatial complexity of CO_2 and H_2O exchange rates. This is indeed the case. Ozone fluxes are comparatively free to vary, although also subject to the constraints corresponding to the influence of stomatal behavior. The aircraft data show that the variation in the ozone flux along a straight line path is about 6 times greater than that for H or LE, implying that the spatially averaged ozone flux to the surface is considerably greater than would be expected if the total resistance to the transfer of ozone were the same as that for sensible heat or the evaporation rate.

The aircraft results now available are tantalizing but not such that they strongly support the overall behavior predicted by the hypotheses presented here. Flying an aircraft low over complex terrain or mountains is not an attractive challenge. Efforts are now under way to utilize small unmanned research aircraft for such a mission.

6. Discussion

Figure 1 shows how an area of complex terrain might be addressed (based on an analysis by McMillen 1990). The area in question is in upstate New York, for the month of July. For the purposes of this illustration, elevation data on a 30-m grid have been used, analyzed in 1-km grid cells covering an 80 km × 80 km area. The various shadings illustrate the proportion of each grid cell that is classified as "complex," defined as when the local slope varies by more than one part in seven from its neighbors.

A classical approach would be to break up every conventional grid cell into smaller cells until every subcell

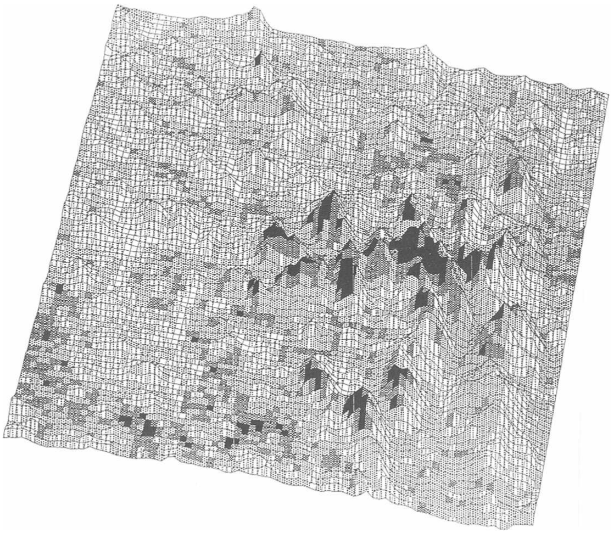

FIG. 1. The results of an analysis of a 100 km × 100 km area of northern New York, with the extent of likely amplification of standard dry deposition velocities indicated by the density of the shading (with black areas such that the aerodynamic resistance might best be considered to be zero). The initial horizontal resolution of elevation data was 30 m (see McMillen 1990), analyzed on a 1-km grid. Black areas are dominated by local slope changes exceeding 1 in 7. Unshaded areas have no local slope changes exceeding 1 in 7.

can be considered uniform. However, the question then would remain as to what description of air surface exchange should be used to address such small areas. Certainly, the standard micrometeorological constraints are violated in the shaded areas of Fig. 1. In this example, standard micrometeorology is not proposed for the grid cells identified as complex. Instead, and on the basis of the discussion above, the dark areas of the diagram are where dry deposition velocities would be expected to be highest, corresponding to deposition "hotspots." It is the current thesis that the areal average deposition across the 80 km × 80 km area represented by Fig. 1 could best be approximated using Eq. (5), perhaps with deposition to the identified subgrid-scale hotspots being best estimated using Eq. (4).

All of the above assumes that air concentrations remain largely unaffected by changes in local deposition rates. This is a defensible assumption for regions where deposition rates are low and sources are distant, but is certainly disputable when local deposition hotspots are suspected, as in Fig. 1. It is clear that there will be negative feedback involved; deposition to local hotspots will be somewhat reduced because the higher deposition velocities then appropriate will decrease the available air concentrations. Not only is this aspect of the problem beyond the present scope, but it is not clear that the issue can be addressed without a coupled observational and theoretical program that remains to be considered.

7. Conclusions

The focus of discussion about atmospheric deposition is slowly shifting, from its role as a sink for air pollution to its role as a contributor to the decline of already stressed parts of the terrestrial and aquatic environments. In the contemporary context of critical loads (e.g., see Langan et al. 2004), it is the actual deposition to the target area that is central to the decision making processes. It is unfortunate that many of these sensitive ecosystems are precisely the mountaintops and areas of complex terrain and surface heterogeneity that are poorly addressed using the models most commonly employed.

Conventional formulations of R_a are derived from basic micrometeorological principles that apply to a horizontal, homogeneous surface; the transfer from the atmosphere is correspondingly assumed to be via eddy diffusion through a plane defined such that there is no net transfer of dry air across it. If the surface is flat and horizontal, then R_a refers to turbulent exchange between some height at which concentrations C are measured and the zero plane level is defined by the wind profile and the flux of momentum. In the case of flow across and around a hill, there is reason to doubt the applicability of such micrometeorological convention.

For a landscape with patches of forest, it has been suggested that the height of the vegetation h and the number n of plots per unit area are key variables. The presence of the patches of forest elevates dry deposition rates beyond the level predicted by a simple area-weighted deposition velocity approach. Not surprisingly, the analysis indicates that the amplification factor to be applied to the conventional "flat earth" estimate of dry deposition might well scale as $hn^{1/2}$, where h is the height of the trees and n is the number of plots of trees per unit area of consideration.

For complex topography, there are two cases that are addressed above, both admittedly heuristically. The first is the case of hills of gentle enough relief that flow separation does not occur. The second is the case of mountains, where flow separation is likely. In both cases, there is accumulating evidence that the familiar roughness length framework is not properly applicable, and that the multiple-resistance approach of most dry deposition models is therefore questionable. Precise but general analyses do not appear feasible, because every situation is likely to be unique in some way. In the absence of a clear-cut way to address the issue of actual deposition rates, the considerations presented above can be used to identify upper and lower bounds on effective deposition velocities, both to the apex of hills and to areas of complex terrain.

- For a forested mountaintop, dry deposition rates are likely to be at least as calculated using the familiar multiple resistance schemes with the aerodynamic resistance reduced to about a quarter of its flat earth value. [This is based on the field studies presented by Bradley (1980).] The corresponding upper limit would result from the use of a value of zero for the aerodynamic resistance.
- Based on the hypotheses now presented for an area of uniformly vegetated complex terrain, the appropriate deposition velocity is likely to be elevated above the value conventionally computed by a factor $[(1 + R_a/(R_b + R_c)]^{1/2}$.

In complex terrain at night, spatially organized circulations are likely to cause deposition to be focused on high-altitude slopes. As yet, there is not enough understanding of the phenomenology of the problem to permit estimation of the consequences. However, it is clear that the methods currently used to estimate dry deposition rates will likely underestimate loadings to high-altitude forests by an amount that could be substantial. There will be additional complications resulting from the effects of spatially organized exchange, resulting from katabatic flows. Such effects will further amplify areal dry deposition rates and will also be of most relevance for the uppermost slopes of mountains.

Regardless of the questions that inevitably arise, the results derived in simple situations are frequently applied elsewhere. The reason given is often that nothing better is available. Here, it is argued that errors arise in the case of some easily transferred pollutants and that these errors could be considerable. The effect of these conclusions on the debate about "critical loads" might also be considerable, since it seems probable that areal dry deposition rates of all chemicals (but especially those with small surface resistance) are underestimated by the assessment methods commonly employed. However, there are no simple micrometeorological methods with which to test the present conclusions. It seems likely that the throughfall methodologies used by Beier et al. (1992) and Weathers et al. (2000) might well prove the only way to verify the importance of terrain complexity in dry deposition applications.

In all cases considered here, it is proposed that the commonly accepted multiple resistance approach using conventional formulations for the various resistances will yield a lower bound on the actual deposition rate. This is a direct result of the consideration of deposition mechanisms additional to the vertical diffusion assumed by the usual methodologies. The question then arises as to whether conventional resistance formulations could be modified to include the effects of some of these additional mechanisms. At present, an answer is not obvious, but based on the considerations now put forward, it appears that the deposition velocity should generally be increased, perhaps most conveniently by the adoption of a parallel resistance term applied in the conventional multiple series resistance framework. Equation (5) shows how this might work in some particular cases.

Last, it is appropriate to point out that the present arguments do not apply to sensible or latent heat fluxes, nor to a number of chemical exchanges for which surface resistances dominate. However, it seems possible that the arguments presented here will prove relevant to the case of momentum exchange.

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REFERENCES

Beier, C., P. Gundersen, and L. Rasmussen, 1992: A new method for estimation of dry deposition of particles based on throughfall measurements in a forest edge. *Atmos. Environ.,* **26A,** 1553–1559.

Bradley, E. F., 1980: An experimental study of the profiles of wind speed, shearing stress and turbulence at the crest of a large hill. *Quart. J. Roy. Meteor. Soc.,* **106,** 101–124.

Crawford, T. L., R. J. Dobosy, R. T. McMillen, C. A. Vogel, and B. B. Hicks, 1996: Air-surface exchange measurements in heterogeneous regions: Extending tower observations with spatial structure observed from small aircraft. *Global Change Biol.,* **2,** 275–285.

Desjardins, R. L., and Coauthors, 1997: Scaling up flux measurements for the boreal forest using aircraft-tower combinations. *J. Geophys. Res.,* **102,** 29 125–29 134.

Finnigan, J. J., and S. E. Belcher, 2004: Flow over a hill covered with a plant canopy. *Quart. J. Roy. Meteor. Soc.,* **130,** 1–29.

Hicks, B. B., 1995: On the determination of total deposition to remote areas. *Acid Rain Research: Do We Have Enough Answers,* G. J. Heij and J. W. Erisman, Eds., Elsevier Press, 163–173.

——, 2001: Areal measurements of ozone, water, and heat fluxes over land with different surface complexity, using aircraft. *Water Air Soil Pollut. Focus,* **1,** 213–222.

——, D. D. Baldocchi, T. P. Meyers, D. R. Matt, and R. P. Hosker Jr., 1987: A preliminary multiple resistance routine for deriving dry deposition velocities from measured quantities. *Water Air Soil Pollut.,* **36,** 311–330.

Jackson, P. S., and J. C. R. Hunt, 1975: Turbulent wind flow over a low hill. *Quart. J. Roy. Meteor. Soc.,* **101,** 929–955.

Katul, G. G., J. J. Finnegan, D. Poggi, R. Leuning, and S. E. Belcher, 2006: The influence of hilly terrain on canopy–

atmosphere carbon dioxide exchange. *Bound.-Layer Meteor.,* **118,** 189–216.

Langan, S. J., J. Hall, B. Reynolds, M. Broadmeadow, M. Hornung, and M. S. Cresser, 2004: The development of an approach to assess critical loads of acidity for woodland habitats in Great Britain. *Hydrol. Earth System Sci.,* **8,** 355–365.

Lenschow, D. H., R. Pearson, and B. B. Stankov, 1982: Measurements of ozone vertical flux to ocean and forest. *J. Geophys. Res.,* **87,** 8833–8837.

Lindberg, S. E., and C. T. Garten, 1989: Sources of sulphur in forest canopy throughfall. *Nature,* **336,** 148–151.

Lovett, G. M., and S. E. Lindberg, 1984: Dry deposition and canopy exchange in a mixed oak forest as determined by analysis of throughfall. *J. Appl. Ecol.,* **21,** 1013–1027.

Ma, J., and S. M. Daggupaty, 2000: Effective dry deposition velocities for gases and particles over heterogeneous terrain. *J. Appl. Meteor.,* **39,** 1379–1390.

Mason, P. J., and J. C. King, 1984: Atmospheric flow over a succession of nearly two-dimensional ridges and valleys. *Quart. J. Roy. Meteor. Soc.,* **110,** 821–845.

McMillen, R. T., 1990: Estimating the spatial variability of trace gas deposition velocities. NOAA Tech. Memo. ERL-ARL-181, 37 pp.

Meroney, R. N., 1980: Wind-tunnel simulation of the flow over hills and complex terrain. *J. Indust. Aerodyn.,* **5,** 297–321.

Meyers, T. P., W. T. Luke, and J. L. Meisinger, 2006: Fluxes of ammonia and sulfate over maize using relaxed eddy accumulation. *Agric. For. Meteor.,* **136,** 203–213.

Monti, P., H. J. S. Fernando, M. Princevac, W. C. Chan, T. A. Kowalewski, and E. R. Pardyiak, 2002: Observations of flow and turbulence in the nocturnal boundary layer over a slope. *J. Atmos. Sci.,* **59,** 2513–2534.

Ohba, R., N. Ukeguchi, S. Kakishima, and B. Lamb, 1990: Wind tunnel experiment of gas diffusion in stably stratified flow over a complex terrain. *Atmos. Environ.,* **24,** 1987–2001.

Ross, A. N., and S. B. Vosper, 2005: Neutral turbulent flow over forested hills. *Quart. J. Roy. Meteor. Soc.,* **131,** 1841–1862.

Sheih, C. M., M. L. Wesely, and B. B. Hicks, 1979: Estimated dry deposition velocities of sulfur over the eastern United States and surrounding regions. *Atmos. Environ.,* **13,** 1361–1368.

Slinn, W. G. N., 1982: Predictions for particle deposition to vegetative surfaces. *Atmos. Environ.,* **16,** 1785–1794.

Snyder, W. H., 1990: Fluid modeling applied to atmospheric diffusion in complex terrain. *Atmos. Environ.,* **24A,** 2071–2088.

Weathers, K. C., G. M. Lovett, and P. Lathrop, 2000: The effect of landscape features on deposition to Hunter Mountain, Catskill Mountains, New York. *Ecol. Appl.,* **10,** 528–540.

Wood, N., 1995: The onset of separation in neutral, turbulent flow over hills. *Bound.-Layer Meteor.,* **76,** 137–164.