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Key Points:

- Models and parameterizations for fogwater deposition are reviewed
- Literature data at forests highly scattered on heterogeneous landscapes
- Issues in model and experimental uncertainties have to be clarified

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Fogwater deposition modeling for terrestrial ecosystems: A review of developments and measurements

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Abstract Recent progress in modeling fogwater (and low cloud water) deposition over terrestrial ecosystems during fogwater droplet interception by vegetative surfaces is reviewed. Several types of models and parameterizations for fogwater deposition are discussed with comparing assumptions, input parameter requirements, and modeled processes. The relationships among deposition velocity of fogwater (V_d) in model results, wind speed, and plant species structures associated with literature values are gathered for model validation. Quantitative comparisons between model results and observations in forest environments revealed differences as large as 2 orders of magnitude, which are likely caused by uncertainties in measurement techniques over heterogeneous landscapes. Results from the literature review show that V_d values ranged from 2.1 to 8.0 cm s^{-1} for short vegetation, whereas V_d = 7.7–92 cm s^{-1} and 0–20 cm s^{-1} for forests measured by throughfall-based methods and the eddy covariance method, respectively. This review also discusses the current understanding of the impacts of fogwater deposition on atmosphere-land interactions and over complex terrain based on results from numerical studies. Lastly, future research priorities in innovative modeling and observational approaches for model validation are outlined.

1. Introduction

The importance of water and nutrient inputs via fog (defined as low clouds in contact with the Earth's surface) that is deposited to terrestrial ecosystems at high elevations and its role in the hydrological cycle has been recognized since the beginning of the last century [*Marioth*, 1906; *Linke*, 1916; *Grunow*, 1955]. In fog that occurs at the surface of the atmospheric boundary layer, liquid water droplets are transported downward by turbulence that is generated by wind shear, and eventually, these droplets are intercepted by the plant canopies [*Lovett*, 1984]. The intercepted water drips to the soil surface under the canopies in the form of throughfall or as stemflow in a phenomenon known as fogwater deposition [*Means*, 1927; *Byers*, 1953; *Azevedo and Morgan*, 1974]. In semiarid and arid regions, in particular, fogwater deposition has long been recognized as an important factor in determining the water balance of woody plants [*Armstrong*, 1990; *Schemenauer and Cereceda*, 1994; *Olivier*, 2002; *Del Val et al.*, 2006; *Hildebrandt and Eltahir*, 2007]. In terms of negative influences of fog, acidic fogwater deposition has been associated with forest decline in industrialized areas [*Anderson et al.*, 1999]. Fog chemistry projects conducted during the past few decades indicate that forest ecosystems are exposed to higher ion concentrations through fogwater than through precipitation [*Saxena and Lin*, 1990; *Kalina and Puxbaum*, 1994; *Schemenauer et al.*, 1994; *Fuzzi et al.*, 1996; *Aneja et al.*, 1998; *Kalina et al.*, 1998; *Wrzesinsky and Klemm*, 2000].

Several studies on the first estimates of fogwater deposition have emphasized that there is a significant contribution from fogwater deposition to total deposition fluxes of water and pollutants [*Lovett*, 1984; *Miller et al.*, 1993a, 1993b; *Pahl et al.*, 1994; *Burkard et al.*, 2003]. Fog is considered a substantial source of water input to cloud forests [*Vogelmann*, 1973] and coastal redwood ecosystems [*Dawson*, 1998] with positive effects on their water status. In cases with high pollution loadings, fogwater deposition is considered a negative effect directly related to forest decline [*Vong et al.*, 1991]. Now, several observational approaches are used for quantifying fogwater deposition; these include passive fog gauges [e.g., *Cavelier et al.*, 1996], the canopy water balance method [e.g., *Juvik and Nullet*, 1995], the eddy covariance approach [e.g., *Beswick et al.*, 1991], and the mass-balance approach [e.g., *Schmid et al.*, 2011]. However, such measurements have limitations because they cannot be carried out over long-term periods and still do not work well over complex terrain. For these reasons, many numerical simulation models for estimation of fogwater deposition to vegetation have been developed over the past several decades (Table 1); these models are reviewed in detail in this paper.

Table 1. References of 1-D Fogwater Deposition Models[a]

References	Model Type	Meteorological Input Data	Turbulent Calc.	DSD/Number of Size Bins	Impaction	Sedimentation	Canopy Evaporation	Edge Effect	Plant Obstacles/Leaf Width
Shuttleworth [1977]	Single layer	U, LWC, T, q, R_n	Log-normal	*Best* [1951]/?	*Ranz and Wong* [1952]	None	Included	Neglected	Needle-leaf
Lovett [1984]	Multilayer	U, LWC, T, q, R_n	Extinction coefficient	*Best* [1951]/3	*Thorne et al.* [1982]	Stokes's law	Included	Neglected	Needle-leaf, stems, branches
Mueller and Weatherford [1988]; *Mueller* [1991]	Multilayer	U, LWC, T, q, R_n	Multilayer	Observations	*Thorne et al.* [1982]	Stokes's law	Included	Included	Needle-leaf
Miller et al. [1993a]	Multilayer	U, LWC, T, q, R_n	Log-normal	*Best* [1951]/500	*Thorne et al.* [1982]	Stokes's law	Included	Neglected	Needle/broad-leaved
Pahl et al. [1994]	Multilayer	U, LWC	Log-normal	*Deirmendjian* [1969]/20	*Thorne et al.* [1982]	Stokes's law	None	Neglected	Needle-leaf
Slinn [1982]	Analytical	U, LWC	Analytical	Observations	*Slinn* [1982]	Stokes's law	Included	Neglected	Needle-leaf
von Glasow and Bott [1999]	Multilayer atmosphere-soil-vegetation	U, LWC, T, q, S_d, L_d, Pr, aerosol	Second-order closure	Explicitly calculated/50	*Thorne et al.* [1982]	Stokes's law	Included	Neglected	Needle-leaf
Katata et al. [2008]	Multilayer atmosphere-soil-vegetation	U, LWC, P, T, q, S_d, L_d, Pr, CO_2	Second-order closure	*Deirmendjian* [1969]/100	*Peters and Eiden* [1992]	Stokes's law	Included	Neglected	Needle-/broad-leaved, crop, grass

[a] U, horizontal wind speed; LWC, liquid water content of fog; P, atmospheric pressure; T, air temperature; q, specific humidity; R_n, net radiation; S_d and L_d, downward solar and long-wave radiation, respectively; Pr, precipitation; CO_2, CO_2 concentration; Aerosol, aerosol spectrum; and DSD, droplet size distribution.

Most models compute fogwater deposition as the sum of the following two independent processes: (1) turbulence-driven fog droplet interception and scavenging by forest elements (mainly by leaves) and (2) gravitational droplet sedimentation of larger droplets. Modeling the effects of these factors appropriately remains very challenging since fogwater deposition is influenced by numerous factors such as liquid water content (LWC), droplet size distribution, wind speed profiles above and within the canopies, the vertical profiles of leaf and stem/branch areas within the canopy, and the heterogeneity of the forest structure [*Lovett*, 1984; *Lovett and Reiners*, 1986]. As a result, various modeling approaches as summarized in Table 1 have been proposed to quantify fogwater deposition. Presently, there is no single model that performs best in all aspects and under all conditions where fogwater deposition is believed to be a relevant component in the ecosystem's hydrological and nutrient cycles.

Several papers have articulated the state of knowledge of fogwater deposition models [*Vong et al.*, 1991; *Gallagher et al.*, 1997; *Wesely and Hicks*, 2000; *Bruijnzeel et al.*, 2005; *Pryor et al.*, 2008]. *Bruijnzeel et al.* [2011] recently developed the first regional maps of modeled fogwater deposition and demonstrated difficulties of direct comparisons between modeled and point measurements due to local topography and variable vegetation cover. These uncertainties of the models are primarily associated with the prediction accuracy of one-dimensional (1-D) fogwater deposition models or the parameterization included as the surface boundary condition in three-dimensional (3-D) meteorological models. However, fogwater deposition is only crudely represented, or completely neglected, even in the state-of-art 3-D fog forecast models reviewed by *Gultepe et al.* [2007]. Hence, a module for fogwater deposition calculations must be considered to the surface boundary condition of a 3-D fog forecast model that is relevant to quantify the removal of fog droplets by vegetation from the atmosphere. While I will not address generalized fog forecast modeling in this review, the improvement of fogwater deposition modeling is both essential for stand-alone applications of such models and for their incorporation into a more complex 3-D fog forecast modeling system.

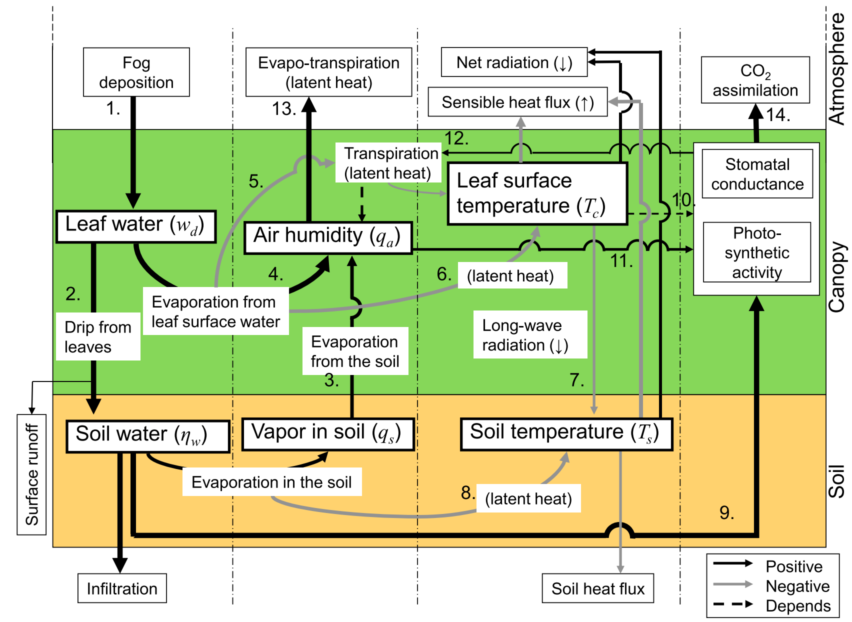

Figure 1. Schematic diagram of the interdependency between water, energy, and CO_2 exchange processes related to fogwater deposition in the multilayer atmosphere-SOiL-VEGetation model, SOLVEG (reprint from *Katata et al.* [2010]). Arrow thicknesses indicate their relative importance.

This paper aims to review the recent progress in modeling fogwater deposition for terrestrial ecosystems. As an introduction, roles of fogwater deposition in atmosphere-land interactions are summarized based on several numerical modeling studies (section 2). Then, recent progress that has been made in 1-D fogwater deposition modeling is reviewed (section 3). Furthermore, comprehensive comparisons between these models and available data for fogwater deposition are made to show uncertainties of both models and observations (section 4). The studies for implementation of fogwater deposition scheme to meteorological (and fog forecast) models are also reviewed, and the difficulties of such modeling approach are discussed (section 5). Finally, future research needs for fogwater deposition modeling are identified through revealing the remaining uncertainties in our understanding (section 6). It should be noted that capture by the vegetative surface (mainly by plant leaves during the growing season but also by branches and stems when leaves are absent or few) of wind-driven fog or cloud droplets is widely referred to as fog drip [*Byers*, 1953], fog precipitation [*Nagel*, 1956], fogwater or occult precipitation [*Rutter*, 1975], fogwater deposition [*Dollard et al.*, 1983], or fog or cloud interception [e.g., *Bruijnzeel et al.*, 2005]. The term fogwater deposition is used throughout the present paper.

2. Roles of Fogwater Deposition for Ecosystems

Observational knowledge of influences of fogwater deposition (and fog itself) on atmosphere-land interactions such as water input, energy budget, evapotranspiration, plant growth, and ecosystem development within cloud forests has drastically increased in recent decades. A detailed review by *Bruijnzeel et al.* [2011] discussed the current knowledge of hydrometeorology in tropical montane cloud forests throughout the world. In contrast, there are still very few numerical studies related to this topic [*Hildebrandt and Eltahir*, 2007, 2008; *Hildebrandt et al.*, 2007; *Katata et al.*, 2010]. Although such studies are few in number, they are very helpful for systematically examining the relationship between fogwater deposition and atmosphere-land interactions over terrestrial ecosystems. In what follows, we will focus on the general aspects of atmosphere-land interactions related to fogwater deposition while introducing observational evidence. However, it should be noted that these generalizations are based on very few scientific studies and hence are associated with a high degree of uncertainty.

2.1. Influence on Water Exchanges

A summary of the atmosphere-vegetation-soil interactions affected by fogwater deposition is schematically shown in Figure 1 (from *Katata et al.* [2010]). When fogwater deposition occurs, the liquid water from the fog is retained by the leaves as leaf surface water (Figure 1, process no. 1). When the leaf surface water exceeds the water storage capacity of the leaves, liquid water retained on the leaf surfaces drips from the leaves to the soil (Figure 1, process no. 2). This water then infiltrates into the soil, thus increasing evaporation from the soil (Figure 1, process no. 3). Importantly, fogwater deposition not only keeps the soil layers near the ground surface wet but also provides water for the deeper soil layers, which is important for effective water uptake by plant roots [*Katata et al*., 2010].

Hildebrandt et al. [2007] demonstrated that soil infiltration is deeper when fog is present because of the low evaporative demand, which results in water being stored in the deeper layers where it is protected from soil evaporation and out of the reach of shallow-rooted plants such as grasses. This water is, however, still available for transpiration by deep-rooted plants such as trees, particularly during periods of high evaporative demand when surface soils may partially or completely dry out. Both studies agree that fogwater deposition provides an important water source, and trees with deep roots may suffer from competition with grasses in semiarid environments. *Hildebrandt and Eltahir* [2008] examined fogwater inputs to a seasonal semiarid cloud forest in Oman using a dynamic vegetation model coupled with an analytical fogwater deposition model [*Slinn*, 1982]. Their results demonstrate that the magnitude of fogwater deposition depends strongly on the vegetation and that sufficient amounts of fogwater to sustain forests can only be gathered under the prerequisite that forests are already present. Therefore, once a canopy has been established via reforestation and the roughness of the surface increases, forest trees may be able to capture sufficient amounts of fogwater to sustain themselves. This finding agrees with purely experimental evidence from the Mediterranean region [*Valiente et al*., 2011]. Such a feedback is generally only expected in water limited areas where cloud forests may have been established in historical times under a different climate regime or where horizontal precipitation (namely, the turbulent fraction thereof) is a relevant contribution to the water balance. In all cases, this feedback mechanism appears important for any such forests with temporarily elevated water demands (e.g., during the dry season) [*Hildebrandt et al*., 2007].

The increase in leaf surface water due to fogwater deposition causes enhanced direct evaporation from the leaf surface water (Figure 1, process no. 4), which results in reduced transpiration. This reduction in transpiration is not only directly related to the reduced energy still available after wet leaf evaporation [*Eugster et al*., 2006] but also to a minor extent indirectly related to decreases in the vapor pressure deficit (VPD) that drives the turbulent exchange of water vapor once liquid water (in the plant leaves) has vaporized (Figure 1, process no. 5). Related observational studies at a Puerto Rican elfin cloud forest [*Eugster et al*., 2006] show reductions in solar radiation inputs. The indirect effect of fogwater deposition on base flow through reduced evaporation losses may be a much more relevant effect that fog imposes on cloud forests in the tropics, and likely elsewhere, than the direct water gains from fogwater interception. Studies using sap flow techniques to quantify the suppression effect of fog on plant transpiration also show reductions of transpiration by 40–60% under foggy conditions as compared to fog-free conditions [*Hildebrandt and Eltahir*, 2007; *Reinhardt and Smith*, 2008; *Ritter et al*., 2009]. In a subtropical mountain cloud forest range in northeastern Taiwan, the water vapor flux is dominated by evaporation of intercepted fog and the contribution from transpiration is limited [*Mildenberger et al*., 2009].

2.2. Influence on Energy Budgets

Fogwater deposition decreases the leaf temperature in the canopy via latent heat due to evaporation from the leaf surface water (Figure 1, process no. 6), which results in decreased downward long-wave radiation entering the soil surface (Figure 1, process no. 7). In addition, fogwater deposition also decreases the soil temperature by increasing the latent heat flux of evaporation from the soil (Figure 1, process no. 8).

Some observational evidence is available for the proposed influence of fogwater deposition on the energy balance described above. *Klemm et al*. [2006] revealed that during foggy conditions, the short-wave radiation was strongly reduced and the long-wave radiation was balanced. In addition, sensible and latent heat fluxes are also strongly limited during foggy conditions. The field experiments in the Puerto Rican elfin cloud forest [*Eugster et al*., 2006; *Holwerda et al*., 2006] showed that downward long-wave radiation falling to the ground surface was small as a result of the deposited fog, which ultimately results in less net radiation. Latent heat fluxes due to evaporation from wet canopies and the soil decrease both the air and leaf temperatures (and hence, the energy balance), which is mentioned by *Hildebrandt et al*. [2007] along with their descriptions of experimental data.

2.3. Influence on CO_2 Exchange

In numerical simulations, the changes in soil water content (Figure 1, process no. 9), leaf temperature (Figure 1, process no. 10), and air humidity within the canopy (Figure 1, process no. 11) affect photosynthesis based on the relationship between the CO_2 assimilation rate and stomatal conductance. Changes in the stomatal conductance, however, cause no significant increase in transpiration (Figure 1, process no. 12) as the restricted transpiration caused by evaporation from leaf surface water compensates for the above effect (Figure 1, process no. 5). Evapotranspiration from the entire canopy then eventually increases because of the increased evaporation from the soil and leaf surface water (Figure 1, process no. 13). Fogwater deposition consequently enhances CO_2 assimilation by easing the stress level within the canopy caused by increased air humidity around the leaf surfaces and soil water (Figure 1, process no. 14).

There is some experimental evidence supporting the conjecture that the decrease in leaf-to-air VPD associated with cloudy conditions can enhance canopy photosynthesis [*Lamaud et al*., 1996; *Freedman et al*., 2001; *Rocha et al*., 2004; *Mildenberger et al*., 2009; *Zhang et al*., 2011] and thus enhance leaf photosynthesis [*Collatz et al*., 1991]. Field observations also suggest that fog can provide benefits other than increased diffuse light, including cooler leaf temperatures and reduced VPD, both of which decrease transpiration [*Gu et al*., 2002]. Although leaf wetness due to fogwater deposition could reduce the carbon gain because of the limited diffusion of CO_2 in water [*Novel*, 2005], water beading on hydrophobic leaf surfaces can result in reduced transpiration and increased photosynthetic carbon gain, i.e., greater stomata opening at low VPD [*Smith and McClean*, 1989]. This is because many plant species in areas with frequent leaf wetting have strategies for surface water repulsion [*Brewer et al*., 1991; *Brewer and Smith*, 1997]. Beyond the model predictions in Figure 1, observational studies confirm that fog typically increases the diffuse component of sunlight at the same level of Photosynthetic Active Radiation (PAR) and thus can also enhance ecosystem photosynthetic rates by increasing diffuse radiation within the canopy and in the understory [*Johnson and Smith*, 2006; *Brodersen et al*., 2008; *Still et al*., 2009]. Specifically, this is because plant leaves have higher radiation use efficiencies for diffuse PAR than for the direct PAR [e.g., *Price and Black*, 1990]. At the same time, decreases in PAR due to foggy conditions can reduce plant photosynthesis in tropical [*Letts and Mulligan*, 2005] and subtropical cloud forests [*Mildenberger et al*., 2009]. The variation in photosynthetic responses to foggy situations is likely due to differences in plant species [*Mildenberger et al*., 2009], forest structure (i.e., open or closed forests), and cloud optical thickness [*Min*, 2005]. The latest study revealed that fog can even affect microbial dynamics and belowground carbon cycling at a coastal area in the USA [*Carbone et al*., 2013]. Further numerical analysis of fogwater deposition may lead to a better understanding of the complicated linkage between fog or low cloud immersion and ecosystem carbon cycling.

3. Fogwater Deposition Modeling

In this section, the development of fogwater deposition models such as resistance, analytical, and more sophisticated models is reviewed. Although these mechanistic (process-based) models introduced in this section are strong tools for fogwater deposition simulation, the crucial meteorological and vegetation parameters necessary for such models are not always available. Moreover, difficulties of model validation due to uncertainties in modeling and observations represent issues that have not been completely resolved at this moment in time (section 4). Given such situations, a simple empirical parameterization in conjunction with less parameters and input data would be useful for hydrological and ecological assessments of fogwater deposition. Such a simple parameterization would also be useful for meteorological or fog forecast modeling studies to reduce computational costs. The existing parameterizations and implementation of those to meteorological models are described in section 5.

3.1. Resistance Model

Shuttleworth [1977], based on the work of *Hori* [1953] and *Merriam* [1973], proposed a steady state 1-D computational model that included the fundamental processes of fogwater deposition to, and evaporation from, a uniform vegetation canopy. The fogwater deposition model is based on that of *Monteith* [1965] using the analogy of electrical resistance in a direct-current circuit based on the "big-leaf" concept.

Lovett et al. [1982] and *Lovett* [1984] adapted the Shuttleworth model by incorporating the vertical structure of forest canopies. This model consists of two modules: (1) a structure module that simulates the structure of the forest and the growth of trees and (2) a hydrology module that computes the evaporation rate and the rate of fogwater deposition resulting from impaction and sedimentation of fog droplets. The hydrology module is again comprised of two parts. The first part simulates the turbulent diffusion of fog droplets into the forest and the fogwater deposition to foliar and branch surfaces due to impaction and sedimentation effects. The second part simulates the evaporation and condensation processes during fog episodes over vegetation canopies. The sum of fluxes over all layers and all components yields the total flux of fogwater to the canopy. The Lovett model includes the effect of vertical variation in canopy structure through the construction of a multilayer model in which the vertical turbulent transport of droplets is controlled by the aerodynamic resistances between model layers and between the top layer and the air above the canopy. The droplet size distribution is that described by *Best* [1951]; droplets larger than 30 μm diameter are not included. The model was used to estimate the contribution of fogwater deposition to surface water budgets in subalpine balsam fir forests at several sites in the Appalachian Mountains. Essentially, the model assumes homogeneous terrain and steady state conditions in the atmospheric environment.

The Lovett model has been developed to investigate the contribution of fogwater deposition to the total acid deposition to ecosystems in the USA in the 1990s [*Vong et al*., 1991]. Several studies based on the work of *Lovett* [1984] have demonstrated the validity of the model through its application in practical cases with sensitivity tests. *Dasch* [1988] estimated fogwater deposition at Clingmans Peak in the USA using the Lovett model. *Sigmon et al*. [1989] applied the Lovett model to the Pinnacles of Shenandoah National Park in the USA. *Stogner and Saxena* [1988], *Saxena et al*. [1989], *Saxena and Lin* [1990], and *Lin and Saxena* [1991] applied the Lovett model to Mount Mitchell in the USA to estimate fogwater deposition. In later study, *Elias et al*. [1995] ran the Lovett model and estimated fogwater deposition at forests in Czech and Germany.

Mueller and Weatherford [1988] and *Mueller* [1991] modified the Lovett model to handle collection efficiency by treating it as a bulk collector (mainly leaf) rather than a multicomponent structure (stems, branches, etc.) and ran the model with observed droplet number and size distribution data in a spruce forest at Mount Whitetop in the USA. *Mueller* [1991] also modified the Lovett model to handle an explicit input of LWC to the size distribution. This model adjusts the input distribution to get a match with the measured LWC, but it preserves the distribution shape. Such an approach is supported by the fact that many observational studies indicate that the mode or mean diameter of fog droplets correlates with LWC [e.g., *Joslin et al*., 1990]. This approach was also used in later modeling studies [*Miller et al*., 1993a, 1993b; *Katata et al*., 2008]. *Vong et al*. [1991] summarized estimations of fogwater deposition on five mountains in the eastern USA by the Lovett model with modifications in leaf area and droplet collection efficiency.

Beswick et al. [1991] used the simplified version of the Lovett model with four canopy layers and a uniform leaf area profile to compare calculations from this model and the analytical model (section 3.2) with eddy covariance measurements. Their results suggest that the analytical model has an advantage over the Lovett model if relatively simple fogwater deposition models can be used with a relationship between momentum flux and droplet deposition velocity as a function of droplet size.

Miller et al. [1993a, 1993b] modified the Lovett model and applied it to a mixed forest at Mount Whiteface in the USA. Modifications were made for the extinction coefficient of wind speed, wind speed-dependent parameterization of displacement height and roughness length, capture efficiency for broad-leaved trees (birch) based on *Bache* [1979], and the fog droplet size distribution.

Pahl et al. [1994] applied the Lovett model to a coniferous forest in Germany. The following modifications of input data were made: a log-normal wind speed profile above the displacement height was assumed, and a modified Gamma distribution was used to approximate the droplet size distribution according to *Deirmendjian* [1969], with parameters fitted to observations, and leaf area index (LAI) derived from an empirical relation with tree diameters. *Kalina et al*. [1998], *Herckes et al*. [2002], *Zimmermann and Zimmermann* [2002], and *Vautz et al*. [2003] used this version of the Lovett model to investigate the sensitivities for meteorological and vegetation parameters and estimated long-term accumulation of fogwater deposition. *Baumgardner et al*. [2003] estimated fogwater deposition in the Appalachian Mountains using the same model with a larger number of droplet size bins (20 bins). *Klemm et al*. [2005] made the first direct comparisons in temporal changes of fogwater flux over the canopy between the model and eddy covariance measurements for a long-term period of 13 months.

3.2. Analytical Model

Slinn [1982] modeled the dry deposition of particles (including fog droplets) to vegetated canopy using approximate analytical equations of deposition velocity. In the analytical model, the fogwater deposition flux, F (kg m^{-2} s^{-1}), is represented as a product of the deposition velocity, V_d (m s^{-1}), and the LWC (kg m^{-3}) according to the inferential technique [*Hicks et al*., 1987; *Erisman*, 1994]:

$$F = \mathrm{LWC} \cdot V_d, \tag{1}$$

where F is positive for downward flux. *Slinn* [1982] derived the formation of V_d based on turbulent exchange theory and suggested a linear relation between V_d and the lateral wind speed. It is also known from later modeling studies [e.g., *Lovett*, 1984; *Katata et al*., 2008] that V_d often shows a nearly linear dependence on lateral wind speed above the canopies, U:

$$V_d = A_i U, \tag{2}$$

where A_i (nondimensional) is the removal efficiency of fog droplets by the canopy depending on vegetation characteristics, which are parameterized in section 5.2. Theoretically, this relationship is expected because both turbulent transport of fogwater droplets to the canopy and their droplet impaction to plant obstacles (represented using the Stokes number; section 4.2) increase with wind speed [*Slinn*, 1982]. Equation (2) is supported by many field observational studies [*Vermeulen et al*., 1997; *Burkard et al*., 2002; *Eugster et al*., 2006; *Holwerda et al*., 2006; *Klemm and Wrzesinsky,* 2007].

Although the analytical model was developed at almost the same time as the Lovett model and it has been widely used for dry deposition estimation, applications of the analytical model to fogwater deposition estimations have been limited [*Beswick et al*., 1991; *Hildebrandt and Eltahir*, 2008]. *Beswick et al*. [1991] showed that the analytical model, which is simpler than the Lovett model, produced better results than the Lovett model. This suggest that without a lot of tuning parameters (particularly for vegetation), the simple model may be more practical to use for estimating fogwater deposition compared with multilayer models. Compared with the other models, the analytical model has lower calculation costs, which is a useful feature for calculating fogwater deposition in 3-D meteorological models [*Hildebrandt and Eltahir*, 2008]. However, for the purpose of obtaining fogwater deposition estimates from routine observational data (e.g., wind speed and visibility), empirical parameterizations are more effective because the analytical model requires additional data representative of fogwater droplet size.

3.3. Sophisticated Atmosphere-Soil-Vegetation Model

There are more sophisticated models compared to the previously mentioned that can be used to investigate complicated atmosphere-land interactions under foggy conditions. The first such model called MIcrophysical FOG model for Vegetation (MIFOG-V) [*von Glasow and Bott*, 1999] is a 1-D boundary layer model for radiation fog simulation with tall vegetation. The model uses a multilayer structure for the atmosphere, soil, vegetation, microphysics, and radiation modules, and, in particular, it considers detailed interactions among heat, moisture, aerosol activation, and fog droplets. The microphysics module is based on MIFOG (MIcrophysical FOG model) developed by *Bott et al*. [1990]. The turbulence in the atmospheric module is calculated by the level 2.5 closure model of *Mellor and Yamada* [1982]. The deposition processes of fog by leaf surfaces are modeled with the experimental collection efficiency due to impaction provided by *Thorne et al*. [1982], and sedimentation is accounted for in a similar way as in the Lovett model. The particle growth including aerosol activation, which depends on the relative humidity and solubility of the aerosol particles, is calculated explicitly [*Bott et al*., 1990].

The second model is named SOLVEG, a one-dimensional multilayer atmosphere-SOiL-VEGetation model [*Katata et al*., 2008], which is also a sophisticated 1-D atmosphere-soil-vegetation model. This model is a multilayer model that consists of four modules for the atmosphere near the surface, soil, vegetation, and radiation within the vegetation canopy. The atmosphere module calculates the variables in each atmospheric layer by numerically solving 1-D diffusion equations for horizontal wind speed components, potential temperature, specific humidity, LWC of the fog, turbulent kinetic energy and length scale, and gas and aerosol concentrations by the turbulence closure model of *Yamada* [1982]. The soil module calculates the soil temperature, volumetric soil water content, and specific humidity of the air in the soil pores using equations for heat conduction, mass balance in liquid water, and water vapor diffusion, respectively [*Katata et al*., 2007].

Table 2. Model Response Ranked by Decreasing Sensitivity to Some Selected Parameterizations[a]

Parameterization	Response (%) SOLVEG	*Mueller* [1991]
Canopy homogeneity (edge effect)	150–300	300–400
Droplet size spectra	10–20	200
Droplet capture efficiency (needle-/broad-leaf or twig/whole tree)	20	100
Canopy structure (Leaf area index and canopy height)	<400	50

[a]The response was calculated as variation in gross and net fogwater deposition flux relative to the control case scenario. In *Mueller* [1991], the maximum uncertainty due to droplet capture efficiency was estimated using two parameterizations based on a spruce twig [*Thorne et al*., 1982] and a whole tree [*Joslin et al*., 1990]. The response for evaporation represents the difference between no evaporation and effect of maximum probable evaporation. Sensitivity tests using SOLVEG were made under the meteorological condition of a coniferous forest in Germany [*Katata et al*., 2008]. The edge effect was tested by setting the drag coefficient of leaves to zero. Three droplet size distributions [*Best,* 1951; *Deirmendjian,* 1969; *Klemm et al*., 2005] were tested to evaluate the response. The response for droplet capture efficiency was evaluated based on the tests of needle and broad-leaves with fixed values of LAI and canopy height. The response for canopy structure was derived from SOLVEG calculations in Figure 2.

The vegetation module calculates the leaf temperature, the water on the surface of the leaves (leaf surface water) for each canopy layer, and the vertical liquid water flux through the entire canopy. In this module, photosynthesis is also incorporated to calculate the CO_2 assimilation rate based on the relationship between stomatal resistance and the net CO_2 assimilation rate [*Nagai*, 2005]. The radiation module separately calculates direct and diffuse downward and upward fluxes of solar and long-wave radiation in the canopy and provides the radiation energy input for the heat budget calculations at the soil surface and in the center of the canopy layers [*Nagai*, 2003]. Both impaction and sedimentation processes are considered to calculate fogwater deposition. The empirical function of the Stokes number [*Peters and Eiden*, 1992] is applied to the collection efficiency due to impaction for the modified Gamma droplet size distribution [*Deirmendjian*, 1969].

The characteristic differences of these three types of models are summarized in Table 2. All models use input data of (typically hourly) horizontal wind speed and LWC above the canopy. The fewest input data requirements are found in the Lovett model modified by *Pahl et al*. [1994] because the model does not include the canopy evapotranspiration process. The other models need additional meteorological data such as air temperature, humidity, and radiation to calculate evapotranspiration. The approach of MIFOG-V and SOLVEG used to model the deposition mechanism is similar to that of the Lovett model; i.e., it is based on impaction and sedimentation processes. The important aspect of both the sophisticated models is to calculate the wind speed profile within the vegetation canopy by turbulence closure models [*Mellor and Yamada*, 1982; *Yamada*, 1982], although the Lovett model uses a simple exponential decay function of LAI. This method has a great advantage in that it represents the wind profile within the canopy better than the Lovett model. Moreover, SOLVEG explicitly calculates the relationship between stomatal resistance and CO_2 assimilation for each vegetation layer, which strongly influences evapotranspiration in wet canopies. In comparisons made between these models and eddy covariance measurements at a coniferous forest in Germany [*Katata et al*., 2008], SOLVEG predicted the observed deposition velocity better than the Lovett model. By considering complicated turbulent transfer within and evapotranspiration from wet canopies, such a detailed modeling approach seems to be appropriate for obtaining accurate predictions of fogwater deposition in the forest canopy.

4. Model-Observation Comparison

It has been emphasized before that the difficulty of model evaluation lies not only in determining the model limitations but also in identifying uncertainties associated with observational data [*Bruijnzeel et al*., 2005, 2011]. This is particularly true under conditions of strong wind in complex terrain. In this section, I first review past model-data comparison studies. Then, I evaluate model uncertainties based on sensitivity studies using the relatively recent fogwater deposition model (SOLVEG). Finally, I compare the calculations of fogwater deposition by SOLVEG with literature values at various forests from numerous data sources.

4.1. Past Model-Observation Comparison Studies

Observational data of throughfall and canopy water balance methods have been used for model validation of the multilayer resistance model (Lovett model) [*Lovett,* 1984] since the beginning. *Lovett* [1984] found a good agreement between model predictions and observations for several fog drip events in balsam fir forest canopies. However, the ensuing studies reported that the (even modified) Lovett model tended to overestimate the measurements. *Dasch* [1988] found that the Lovett model appears to considerably overestimate the contribution of fogwater deposition and suggested that this may be a result of uncertainties in input data such as the droplet size distribution of fog, LAI, and wind direction. *Lovett et al*. [1997] showed that a 25% overestimation of wind speed, LWC, and cloud immersion time would accumulate so that fogwater deposition would be overestimated by more than a factor of 2. *Mueller and Weatherford* [1988] confirmed that estimates of fogwater made with the Lovett model for forests appeared to be too high when compared with canopy throughfall data. *Mueller et al*. [1991] tested the Lovett model modified by *Mueller* [1991], and the model overestimated fogwater deposition under meteorological conditions with high values of LWC and horizontal wind speed. They attributed the reason for the overestimation to inadequacy of the modeled vertical transport process related to eddy diffusivity within forest canopy. *Miller et al*. [1993a, 1993b] found agreement between a tracer mass balance analysis of throughfall samples and model calculations, and the uncertainty of total (dry + fog) deposition estimates was on the order of ±25% in experiments at Mount Whiteface. However, they mentioned that the relatively good agreement does not guarantee that modeling of fogwater deposition will be accurate; the modeled fluxes, precipitation fluxes, throughfall, or soil water fluxes could all have compensating errors. *Pahl et al*. [1994] compared calculations by the modified Lovett model with observations of rain-free throughfall under spruce stands at Kleiner Feldberg in Germany. The calculated deposition fluxes in their model were clearly higher than the measured ones.

Since the 1990s, measurements using micrometeorological techniques have also been carried out and compared with model results. *Beswick et al*. [1991] compared size-resolved fluxes between the Lovett model and eddy covariance measurements at a spruce forest site in Scotland. They found that the simplified Lovett model generally overestimated the observed deposition velocity up to two times for droplets smaller than 20 μm in diameter. A study by *Klemm et al*. [2005] has also shown that calculations of fogwater flux over the forest canopy by the Lovett model overestimate measurements by the eddy covariance method by up to 32%.

In terms of sophisticated fogwater deposition models, *Katata et al*. [2008] compared the SOLVEG calculations with eddy covariance measurements at the coniferous forest site in Germany used by *Klemm et al*. [2005] over a long-term period of almost 1 year. Better predictions of measured turbulent and gravitational fogwater fluxes were obtained with the sophisticated model than with the Lovett model. The improvement in prediction accuracy could have been due to the detailed modeling of turbulent transport and canopy evaporation for fog droplets within the canopy. The model was also good at predicting the sensible and latent heat fluxes over the canopy [*Katata et al*., 2008], which is important for validation of the evapotranspiration process in a wet canopy. The MIFOG-V model is unfortunately not directly compared with relevant measurements because of the lack of data [*von Glasow and Bott*, 1999].

4.2. Model Uncertainties: A Sensitivity Analysis

A sensitivity analysis not only provides a framework for assessing the potential for bias and the extent of uncertainty in fogwater deposition estimates but also reveals significant factors that determine the extent of fogwater deposition. Many sensitivity tests have been carried out using original or modified Lovett models [*Lovett*, 1984; *Lovett and Reiners*, 1986; *Mueller*, 1991; *Herckes et al*., 2002; *Baumgardner et al*., 2003]. This paper overviews potential uncertainties in the models based on the sensitivity tests using SOLVEG and in the work of *Mueller* [1991], who examined the modified Lovett model to determine the relative importance of various parameterizations of model output values in fogwater deposition flux over coniferous forest (Table 2). Using the same model and throughfall data, *Mueller et al*. [1991] concludes that model performance is most sensitive to the wind speed profile and droplet collection efficiency schemes used, while the droplet size spectrum scheme and representativeness of the input meteorological data are also important.

According to Table 2, the horizontal heterogeneity in forest structure (hereinafter referred to edge effect) has the largest influence on fogwater deposition flux. In general, the 1-D models introduced in section 3 assume a horizontally homogeneous situation where downward transport of fog droplets is controlled by the process of vertical eddy diffusion between the vegetation canopy and the atmosphere. Such an assumption is probably reasonable for well-closed canopies when (in the absence of diffusion from above) the droplet concentration would fall off in an approximately exponential way against a horizontal distance on the order of 20 m away from the edge of a homogeneous tree stand [*Shuttleworth*, 1977]. However, these models are subject to potentially serious errors when applied to complex terrain because they are limited to representing the situation of turbulent transfer as only an entirely vertically diffusive process [*Mueller and Weatherford*, 1988]. It is known that fogwater deposition rates vary greatly as a function of elevation, slope, orientation, topographic exposure, canopy type, and the presence of edges along a large clearing or natural gap [*Lovett et al*., 1997]. In mountainous areas, the horizontal component of the wind vector may also dominate and steep slopes may expose some fraction of the canopy as a horizontal cross section to the wind [*Hicks and Meyers*, 1988]. Rapid changes in topography streamline compression as air flows over terrain features and the formation of quasi-steady wake eddies can force air streamlines to vary in height above the ground. When this happens, the potential exists for significant advection of fog-rich air downward into a canopy or fog depleted air from out of a canopy; thus, the premise that turbulent fogwater flux controls the fogwater deposition process is jeopardized [*Mueller et al*., 1991].

Lovett and Reiners [1986] showed with sensitivity tests of the Lovett model that the fogwater deposition flux near a forest edge or gap can increase by a factor of approximately five. *Dasch* [1988] reported on throughfall measurements at Clingmans Peak and found that a single exposed tree on the summit intercepted 10 times more fogwater than trees in the canopy. *Weathers et al*. [1995] showed, using a regression of deposition versus distance, that the deposition "half-distance," which is the point at which the rate of cloud water deposition is 50% of the rate at the windward edge of the forest, ranged from 2 to 36 m. They concluded that cloud forest landscapes are often highly heterogeneous, consisting of many edges or gaps, and thus, fogwater deposition models may seriously underestimate cloud deposition.

The edge effect, which is related to inflow and advection processes, is difficult to explicitly take into account with 1-D vertical models. In early studies [e.g., *Mueller and Weatherford*, 1988], a simple way is devised whereby the vertical aerodynamic resistance is set to zero to mimic the enhancement of fogwater deposition due to the edge effect (Table 1). In addition, the displacement height is assumed to be zero, which means the diabathic wind speed profile even within canopies. On the basis of this approach, *Mueller and Weatherford* [1988] ran the Lovett model with zero vertical aerodynamic resistance and showed that the advection of fog droplets caused an increase of up to 35–45% in deposition rates over those due to turbulent transport. The sensitivity test of SOLVEG with zero drag coefficient for all canopy layers shows fogwater deposition can increase up to 300% (Table 2).

The droplet size distribution can change fogwater deposition flux to some extent in the tests of SOLVEG with three parameterizations of *Best* [1951], *Deirmendjian* [1969], and *Klemm et al*. [2005], while it is the second most influential parameter for fogwater deposition estimation in *Mueller* [1991] (Table 2). Very large change of 200% in *Mueller* [1991] is attributed to his sensitivity tests that include the droplet size parameterization with fixed droplet diameter as unreasonable case in nature. In fact, *Herckes et al*. [2002] ran the Lovett model of *Pahl et al*. [1994] with two parameterizations [*Best,* 1951; *Deirmendjian,* 1969] and also found a small increase (30%) in fogwater deposition flux when using the parameterization of *Best* [1951], which is similar to the result of sensitivity test of SOLVEG (Table 2). In view of computational techniques, the number of fog droplet bins affects the calculation results of fogwater deposition (Table 1). *Miller et al*. [1993a, 1993b] show that when only three droplet size classes are used in the original Lovett model, deposition velocity calculated with the finest discretization can be underestimated by as much as 27% under meteorological conditions typical of the growing season. They recommend the use of 500 droplet size bins, which can provide a reasonable discretization of the droplet size distribution in terms of finding deposition velocity approaching that of the limiting deposition velocity to within 0.1%. *Baumgardner et al*. [2003] confirmed that the deposition velocity using 20 classes was within 1% of the deposition velocity using 500 classes. *Katata et al*. [2008] used 100 size bins with their detailed multilayer model, and they were successful in achieving a reasonable simulation of the observed turbulent fogwater deposition flux over coniferous forest.

The appropriate choice of collection efficiency parameterizations can also influence fogwater deposition estimations (Table 2). Most models compute fogwater deposition by vegetation components as the sum of the following two independent processes: inertial impaction and sedimentation of fog droplets (Table 1). The

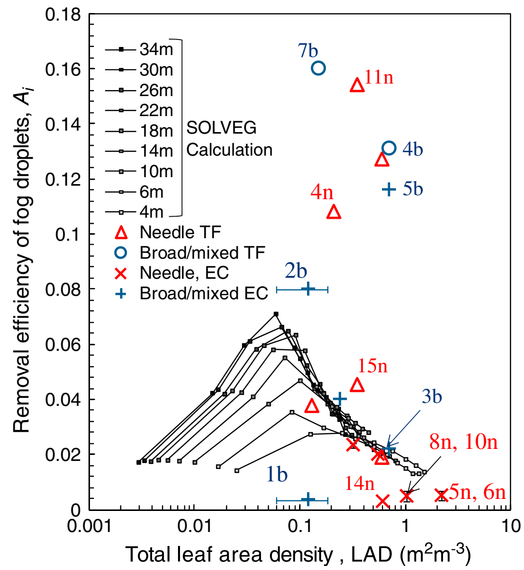

Figure 2. Changes of the removal efficiency of fog droplets by the canopy (A_i) with change in total leaf area density (LAD) for canopy heights of 4–34 m calculated by SOLVEG [*Katata et al*., 2008] (shaded squares with solid lines). Plots and numbers are given from field experiments done at various forest sites listed in Table 3.

processes depend on the factors such as the droplet size distribution and wind speed. *Mueller and Weatherford* [1988] carried out a sensitivity test of the fog droplet size distribution with a 10 μm mode diameter and found that fogwater deposition was roughly 90% due to inertial impaction and only about 10% due to sedimentation. However, a shift in mode diameter of fog droplets to 30 μm resulted in a dramatic shift in the relative importance of sedimentation [*Mueller and Weatherford*, 1988]. A similar situation where sedimentation can be important, even dominant, at low wind speeds < 2 m s^{-1} is observed in other sensitivity studies [*Lovett*, 1984; *Herckes et al*., 2002]. Typically, low wind speed conditions appear during radiation fog events. During such conditions, sedimentation is the dominant process for fogwater interception, whereas at higher wind speeds, interception prevails [*von Glasow and Bott*, 1999].

The collection efficiency of fog droplets by canopy surfaces is affected by tree morphology because it is often represented using the Stokes number, i.e., wind speed, droplet sizes, and obstacle (mainly leaf) sizes. In addition, leaf shape and size also influence the collection efficiency as an obstacle. In earlier studies, the morphological differences between spruce and fir needles (spruce needles are nearly cylindrical, whereas fir needles are more flattened) were not believed to be important given the experimental uncertainty in the wind tunnel data [*Thorne et al*., 1982]. In the same way, *Joslin et al*. [1990] suggested that the collection rate of whole tree collectors is best described as a simple linear function of horizontal fogwater flux (LWC multiplied with horizontal wind speed). However, *Mueller* [1991] examined the maximum uncertainty due to collection efficiency and found that it was very large (100%) when comparing parameterizations of both studies (Table 2). Several models include the difference in collection velocities due to vegetation species based on wind tunnel experiments [*Miller et al*., 1993a; *Katata et al*., 2008]. While *Baumgardner et al*. [2003] applied sensitivity tests and found that broad-leaf versus needle-leaf trees had only a minor effect on fogwater deposition flux, leaf characteristics still affect the deposition velocity to some extent (20 %) according to SOLVEG calculations (Table 2). Theoretically, broad-leaved trees generally exhibit smaller deposition velocities in comparisons to coniferous trees with the same canopy structure (i.e., LAI and canopy height) because the leaf size is larger and thus produces less capture efficiency.

Canopy structures (height and leaf area) strongly influence fogwater deposition flux, as well as canopy homogeneity such as edge effect (Table 2). As shown in next section (Figure 2), according to the result of sensitivity tests on SOLVEG data with varying canopy heights and LAI values, deposition velocity of tall and "moderate dense" trees is up to 4 times higher than that of short and extremely dense or sparse trees.

4.3. Comprehensive Model-Observation Comparison

Table 3 summarizes the variables related to fogwater deposition over short and tall vegetation surfaces measured by various methods. Measurements of fogwater deposition velocity, V_d, ranged from 2.1 to 8.0 cm s^{-1} (horizontal wind speed $U = 1.1$–9 m s^{-1}) for short vegetation. For forests, both values and variations of V_d were overall larger ranging from 7.7 to 92 cm s^{-1} ($U = 2$–11 m s^{-1}) and 0 to 20 cm s^{-1} ($U = 0$–15 m s^{-1}) as measured by throughfall or canopy water balance methods and the eddy covariance method (without any corrections described below), respectively.

High variability of deposition velocity for forests may be explained by uncertainties in measurement methods, which has been discussed at length. *Lovett* [1988] argues that the canopy water balance approach

Table 3. Existing Measurements Related to Fogwater Deposition by Various Methods Over Different Vegetation Types[a]

No.	References	Vegetation Species	Location/Method (Number of Collectors for TF and CWB)[b]	h (m)	Total LAI[c]	Total LAD	U (m s^{-1})	LWC (g m^{-3})	d_{pm} (μm)	V_d (cm s^{-1})	Removal Efficiency (A_i)
Grassland											
1g	*Dollard and Unsworth* [1983]	Grass	Mount St. Bernard Abbey, UK/G	0.08	–	–	1.1–4.2	0.05–0.22	20	2.8–6.4	0.0166
2g	*Gallagher et al.* [1988]	Moorland	Great Dun Fell, UK/G	0.2	2	10	7.0	0–0.24	6–10	4.2–6.7	0.0078
3g	*Fowler et al.* [1990]	Moorland	Great Dun Fell, UK/WL + G	0.2	5	25	9.0	0.24–0.40	10	2.1–3.9	0.0033
4g	*Cameron et al.* [1997]	Tussock grass	Dunedin, New Zealand/WL	0.8	3	3.75	4–9	0.14	14	4.0–8.0	0.0092
5g	*Thalmann et al.* [2002]	Grass & crop	Kerzersmoos, Switzerland/EC	0.3	≈3	≈10	1.1	0.061	10–20	2.7	0.0245
Needle-leaf forest											
1n	*Dasch* [1988]	Spruce-fir	Clingmans Peak, USA/TF (6)	10	6	0.6	3.7–4.5	0–0.28	> 6	7.7	0.0188
2n	*Dasch* [1988][d]	Spruce-fir	Clingmans Peak, USA/TF (7)	10	6	0.6	3.7–4.5	0–0.28	> 6	52	0.127
3n	*Mueller et al.* [1991][d]	Spruce	Mount Whitetop, USA/CWB (24)	15	2	0.13	7–11	0.17–0.26	–	34	0.0376
4n	*Mueller et al.* [1991]	Spruce	Mount Whitetop, USA/CWB (25)	17.5	3.6	0.21	4–7.5	0.05–0.25	–	62	0.108
5n	*Beswick et al.* [1991]	Spruce	Dunslair Heights, Scotland/EC	4.5	10	2.2	2.7–6.1	0.05–0.35	4.4–6.4	1.0–2.0	0.0034
6n	*Gallagher et al.* [1992]	Spruce	Dunslair Heights, Scotland/EC	4.5	10	2.2	15	0.2	5–9	10	0.0067
7n	*Pahl et al.* [1994]; *Wobrock et al.* [1994]	Spruce	Kleiner Feldberg, Germany/TF (2)	17	13	0.77	–	0.34	8–12	19–39	–
8n	*Vong and Kowalski* [1995]	Fir	Ceeka Peak, USA/EC	6.8	7.0	1.03	4–10	< 0.4	15	1–3	0.0029
9n	*Vermeulen et al.* [1997]	Fir	Speulderbos, Netherlands/EC	20	11	0.55	1–4	< 0.4	20	0–10	0.020
10n	*Kowalski and Vong* [1999]	Fir	Ceeka Peak, USA/EC	6.8	7.0	1.03	2–10	0.2–1.0	15–19	0–8	0.0067
11n	*Kobayashi et al.* [2002][d]	Cedar	Mount Rokko, Japan/TF (2)	13	4.5	0.35	2–8	0.1–0.3	–	92	0.154
12n	*Burkard et al.* [2002]; *Klemm and Wrzesinsky* [2007]	Spruce	Waldstein, Germany/EC	20	6.4	0.32	0–8	0.16	9–15	0–20	0.0242
13n	*Thalmann et al.* [2002]	Spruce	Waldstein, Germany/EC	20	6.4	0.32	3.3	0.066	9–15	7.5	0.0226
14n	*Beiderwieden et al.* [2008]	Spruce	Yuan Yang Lake, Taiwan/EC	10.3	6.3	0.61	1.5	0.18	5–25	4.4[e]	0.0029
15n	*Katata et al.* [2011][f]	Cedar	Mount Rokko, Japan/TF (2)	13	4.5	0.35	2–8	0.1–0.3	–	23–31	0.0452
Mixed and broad-leaved forests											
1b	*Burkard et al.* [2003][g]	Beech, spruce, ash, etc.	Lägeren, Switzerland/EC	30	1.7–5.5	0.06–0.18	2.5	0.06	10	0.9	0.0036
2b	*Burkard et al.* [2003][h]	Beech, spruce, ash, etc.	Lägeren, Switzerland/EC	30	1.7–5.5	0.06–0.18	0.6	0.14	14	4.8	0.080
3b	*Holwerda et al.* [2006]	Bignoniaceae etc.	Pico del Este, Puerto Rico/EC	3	2.1	0.7	6.3	0.08	13.8	14	0.022
4b	*Holwerda et al.* [2006]	Bignoniaceae etc.	Pico del Este, Puerto Rico/CWB (20)	3	2.1	0.7	6.3	0.08	13.8	83	0.131
5b	*Eugster et al.* [2006][i]	Bignoniaceae etc.	Pico del Este, Puerto Rico/EC	3	2.1	0.7	6.3	0.08	13.8	72	0.116
6b	*Schmid et al.* [2011]	Epiphytes, etc.	Monte Verde, Costa Rica/EC	21	> 5	> 0.24	2.6	0.13	7–12	10.4	0.040
7b	*Yamaguchi et al.* [2013][d]	*Betula ermanii*	Lake Mashu, Japan/TF (3)	10	1.5	0.15	3.7–5.7	0.04–0.07	12–18	75	0.160

[a] h, mean canopy height; U, wind speed above canopies; LWC, liquid water content of fog; d_{pm}, mode or mean droplet diameter; and V_d, deposition velocity. The removal efficiency of fog droplets by the canopy (A_i) was calculated simply by dividing the mean V_d by mean U assuming that both parameters were linearly related to each other.
[b] G = Gradient Method, WL = Weighting Lysimeter, EC = Eddy Covariance Method, TF = Throughfall and Rainfall, CWB= Canopy Water Budget Method.
[c] One-sided.
[d] Located near the forest edge.
[e] Calculated from turbulent flux only since no sedimentation flux is reported.
[f] Corrected No. 11n for the edge effect.
[g] Fog under advective influence, i.e., low clouds intercepted by the mountain range.
[h] Radiation fog.
[i] Corrected No. 3b for advection and condensation effects.

is probably the best because it is the simplest and most direct method, it requires the fewest assumptions, and it is capable of integrating data over long time periods. At the same time, the confidence margins of the experimental field estimates are generally wide because of the accumulation and propagation of errors associated with the measurement of the individual components [*Holwerda et al*., 2006; *McJannet et al*., 2007]. Field measurements are always affected by local inhomogeneities of the leaf area and canopy structure above the collection points, in particular when only a small number of throughfall collectors (<10 numbers) are used [*Olson et al*., 1981; *Holwerda et al*., 2011]. In addition, there is the difficulty of adequately measuring precipitation inputs of wind-driven rain and drizzle [*Blocken et al*., 2006]. Moreover, there are also shortcomings in eddy covariance measurements especially over complex terrain. One important aspect to consider is the fact that the vast majority of eddy covariance studies focus on turbulent exchange over flat and horizontally homogeneous terrain, which is not the customary setting one finds in mountainous terrain where fogwater deposition is a relevant component of the water budget [*Eugster et al*., 2006]. An earlier study by *Vong and Kowalski* [1995] suggests that the effect of evaporation significantly influences the fluxes for small cloud droplets and can lead to upward fluxes for these particles. In their later studies, they found that eddy covariance measurements in sloping terrain may be underestimating fogwater deposition because of the condensation of water vapor as the air moves upslope [*Kowalski and Vong*, 1999]. In all cases, eddy covariance flux measurements quantify the net flux of fogwater over the canopy defined as the difference between upward flux and downward flux, whereas hydrologists are interested in the gross flux defined as water input due to fogwater deposition at the canopy and soil surface levels. Hence, measured net fluxes must be converted to gross flux for such a purpose, a procedure that typically involves model assumptions on how to quantify concurrent gross flux in the opposite direction. *Eugster et al*. [2006] concluded that the gross fogwater flux at the canopy level for an elfin cloud forest site in Puerto Rico was 1.7 times the net flux measured by the eddy covariance method. This value is similar to that obtained from comparisons between eddy covariance and the stable isotope tracer technique by *Schmid et al*. [2011]. The resulting difference in fogwater deposition at the height of measurement and that at the top of the canopy needs to be taken into account to obtain a realistic value of fogwater deposition [*Eugster et al*., 2006; *Holwerda et al*., 2006], which then can be directly compared with numerical model simulations. The conversion from net fluxes to gross fluxes should be also considered in the situation that net flux is actually not driven by the removal of fog droplets by the canopy but is driven by evaporation of small droplets before they even touch the vegetation. This would reduce the discrepancy between model results and measurements.

To investigate the high variability in observations, all data collected by various methods at forests (Table 3) are compared with calculation results [*Katata et al*., 2008] using the SOLVEG model. Figure 2 shows the dependence of the removal efficiency A_i (equation (2)) on total leaf area density (LAD). This figure shows changes of calculated and observed slopes of fogwater deposition velocity, A_i, in equation (2) according to changes in total leaf area density (LAD). It can be seen that the relationship between aerodynamic and canopy resistances for fogwater at a given LAD strongly influenced deposition velocity V_d (and thus A_i) in this numerical experiment. When LAD is small, A_i is also small because large canopy resistances due to less leaf surfaces inhibit increases of V_d. This limitation gradually disappears with an increase in LAD, based on the assumption that an increase of leaf surface area where fogwater can be intercepted increases the droplet removal efficiency of the canopy. The maximum deposition velocity is reached around a LAD of 0.1 m^2 m^{-3}. Several other modeling studies also have suggested that there is a maximum peak at a certain LAI value based on sensitivity tests whereby LAI is increased [*Lovett and Reiners*, 1986; *Herckes et al*., 2002]. Further increases in LAD decrease A_i because aerodynamic resistance begins to control V_d; in other words, fogwater in the air penetrates less through the canopy. As a result, high LAD values act to block the air and reduce total fogwater deposition. This decreasing tendency was also clearly observed in several measurements of A_i ranging from 0.01 to 0.06 ($R = -0.73$; Table 3), which indicates that the above discussion for dense canopies is most likely a valid representation of reality.

Meanwhile, the rest of the data in the range in A_i below 0.01 and above 0.06 are very scattered, and this makes it difficult to quantitatively validate the modeled deposition fluxes. As shown in Figure 2, the data collected by throughfall or the canopy water balance method were 5–10 times larger than the calculations (2n, 4n, 11n, 4b, and 7b in Table 3). The discrepancy for several sites (2n, 11n, and 7b in Table 3) can be partially explained by uncertainties due to the edge effect (section 4.3) and a small number of throughfall collectors. However, the fluxes observed within closed canopies at Mount Whitetop in the USA (4n in Table 3) are very large in comparison with those observed near the edge at the summit of the same mountain

(3n in Table), which is an area that one would expect to find higher interception rates than in the closed canopy. This may be an artifact of uncertainties in the input data for wind speed and LWC at both sites [*Mueller et al*., 1991]. The data collected by the canopy water balance method at Puerto Rico (4b in Table 3) also show large values despite the use of large number of collectors. In Figure 2, another two data points collected by the eddy covariance method without (3b in Table 3) [*Holwerda et al*., 2006] and with corrections for advection and condensation effects (5b in Table 3) [*Eugster et al*., 2006] at the same location are plotted. Although calculations of SOLVEG are close to the data of 3b, *Eugster et al*. [2006] and *Holwerda et al*. [2006] emphasize that there is a significant underestimation of these data because of the advection and condensation effects discussed above. Similar effects can be considered at sites where the eddy covariance net flux measurements were very small compared with calculated values (5n, 6n, 8n, 10n, and 14n in Figure 2). This indicates that a consistent method to estimate advection and condensation effects will be of key importance for achieving further improvements of model-data comparisons. At the coniferous site in Taiwan (14n in Table 3), the sedimentation flux is not included in the measurements, which should cause underestimation to some extent.

At a mixed forest site in Switzerland, deposition velocity of fog under advective influence (1b in Figure 2) is significantly smaller than that of radiation fog (2b in Figure 2). This difference is due to the fact that radiation fog has larger mean droplet diameters (14 μm, Table 3) and higher sedimentation rates [*Burkard et al*., 2003]. In the advective fog case, the net fluxes are not corrected for concurrent condensation and evaporation effects; namely, fog under advective influence is lifted orographically as clouds contacted the topography. However, the relevant correction is not needed in advective fog conditions because the air movements are along the mountain range at the site and there is much less orographic lifting than in Puerto Rico [*Eugster et al*., 2006].

5. Implementation in Meteorological Models

In this section, I review the prior studies of 2-D or 3-D meteorological (and fog forecast) models including the process of fogwater deposition. Then, I introduce the current fogwater deposition schemes (parameterizations) to show how to implement the fogwater deposition scheme to meteorological models. Finally, difficulties in appropriate treatment of fogwater deposition process in meteorological models, particularly when applying those to complex terrains such as in mountainous regions, are discussed.

It is noted that most of studies reviewed in this section use a simple parameterization of fogwater deposition with low computational costs. However, there are two studies that carried out offline coupling simulations a 3-D meteorological model with a sophisticated fogwater deposition model [*Katata et al*., 2010; *Shimadera et al*., 2011]. For example, *Katata et al*. [2010] estimated the amount of fogwater deposition to mountain forest in Saudi Arabia using SOLVEG. The spatial distribution of calculated cumulative fogwater deposition is depicted in Figure 3. Despite high computational load, this kind of simulation may provide a possibility for detailed analyses of fogwater deposition process with considering complicated forest structure.

5.1. Past Studies of Meteorological Models Including Fogwater Deposition Process

The first studies of two-dimensional (2-D) fogwater deposition calculations to complex terrain were performed by *Hill et al*. [1986, 1987] who focused on a cap cloud over an ideal hill. The studies investigated the spatial variation of fogwater deposition over a hill slope using cloud dynamical and microphysical models with a simple parameterization proposed by *Dollard and Unsworth* [1983] for deposition velocity. It was demonstrated that in such regions fogwater deposition is very sensitive to wind speed, the stability profile of the atmosphere, and the surface roughness. They also suggest that fogwater deposition might be at a maximum just to the lee of the summit of a hill, particularly during supercritical flow when the wind speeds tend to be at maximum in this region. *Hill et al*. [1987] examined fogwater deposition over the ideal hill using the same model and found that the deposition patterns were strongly affected by atmospheric stability and wind speed; the maximum fogwater deposition rate occurs somewhat upstream of the hill summit, depending on atmospheric stability. Further numerical studies at an ideal hill cap cloud have also been carried out by *Bower et al*. [1995].

Fogwater deposition modeling over real complex areas using 3-D models started in the 1990s. *Coe et al*. [1991] presented simulation results using a 3-D airflow model to investigate turbulent deposition of fogwater and dissolved species to regions of complex topography where fog or clouds are in contact with the ground. The model includes the size resolved deposition velocity over the forest and moorland proposed by

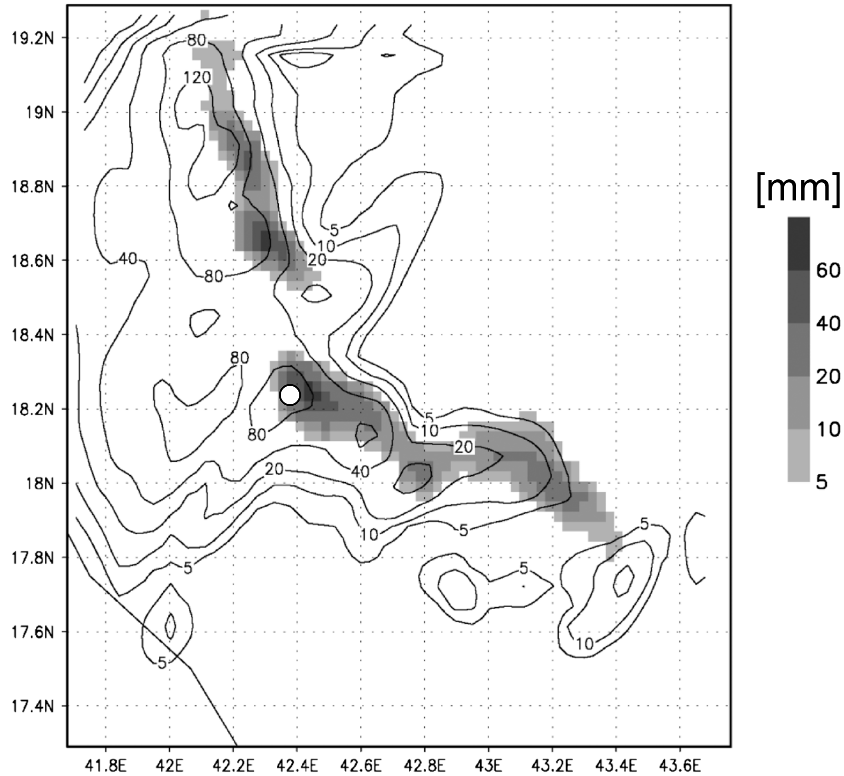

Figure 3. An example of 3-D fogwater deposition calculation: the spatial distribution of cumulative precipitation calculated by the Fifth-generation Penn State/NCAR Mesoscale Model, MM5 (contour lines) and fog deposition to vegetation by the multilayer atmosphere-SOiL-VEGetation model, SOLVEG (shaded areas) in Saudi Arabia (reprint from *Katata et al.* [2010]).

Gallagher et al. [1992]. The results showed that the deposition was a very strong function of the position along the terrain in Scotland. Numerical simulations using the same model were also carried out over a moorland site in the UK [*Gallagher et al.*, 1992]. *Yin and Arp* [1994] investigated the surface water balance at forested watersheds in Canada using a forest hydrology model with simpler parameterization of deposition velocity [*Unsworth and Wilshaw*, 1989] adapted to open or partially forested areas. *Yanni et al.* [2000] also applied the same model to quantify the hydrological role of fogwater deposition in the Mersey River basin in Canada. *Walmsley et al.* [1996] proposed methodology using a numerical airflow model to estimate the hydrological input of fogwater deposition in mountainous terrain in the USA. They found that the fogwater deposition rate could be specified as a linear function both of terrain height above the cloud base and of wind speed, particularly near the summit. As a result, spatial patterns of fogwater deposition strongly reflected the pattern of topographic contours with some modifications being apparent because of spatial variations in wind speed. Similar results of topographic dependency in fogwater deposition can be found in studies at mountainous areas in Saudi Arabia [*Katata et al.*, 2010] and Japan [*Katata et al.*, 2011], where the regional meteorological models used included the FogDES scheme (sea next section).

In order to assess the influence of fogwater deposition to tropical montane cloud forests (TMCF) in terms of both hydrological and conservation reasons, a hydrological model including empirical parameterization of deposition velocity was applied widely throughout Latin America and parts of tropical Africa and Asia [*Bruijnzeel et al.*, 2011]. It was demonstrated that the areas where fog constitutes a significant hydrological input tended to be spatially restricted based on numerical simulation results. A detailed review of fogwater deposition for tropical montane cloud forests can be found in *Bruijnzeel et al.* [2011].

5.2. Fogwater Deposition Parameterizations for Meteorological Models

On the basis of the concept of deposition velocity (V_d) to calculate fogwater deposition flux represented as equation (1), several simple parameterizations based on aerodynamic approaches have been proposed in the past studies. *Unsworth and Wilshaw* [1989] first suggested that V_d could be represented as the reciprocal of the surface resistance for fog droplets, u_*^2/U, where U and u_* (m s^{-1}) are the horizontal wind and friction velocity over the tree canopy, respectively. This approach assumes that fog droplets are deposited by way of turbulent diffusion and sedimentation and that turbulent diffusion proceeds at a maximum rate determined by momentum transport. These assumptions are valid for large droplets 10–24 μm in diameter over short vegetation such as grassland [*Dollard and Unsworth*, 1983; *Gallagher et al.*, 1988]. Droplets smaller than 10–24 μm in diameter deposit less efficient than momentum transport because of the increasing importance of the boundary layer resistance around the individual elements of impaction, whereas droplets larger than 10–24 μm in diameter are transported less efficiently because their inertia does not allow them to follow the motion of the high frequency turbulent eddies close to the ground.

To avoid this shortcoming, *Gallagher et al.* [1992] proposed the following empirical equations for size-dependent V_d based on gradient and eddy covariance measurements over moorland and coniferous forest in Scotland and the UK, respectively:

$$V_d = \left[-0.0173\left(d_p/2\right)^2 + 0.315\left(d_p/2\right) - 1.30\right]u* \text{ for moorland and} \tag{3}$$

$$V_d = \left[-0.011\left(d_p/2\right)^2 + 0.311\left(d_p/2\right) - 1.41\right]u* \text{ for forest,} \tag{4}$$

where d_p is the droplet diameter in μm.

Vermeulen et al. [1997] proposed a simple parameterization without dependence on d_p, which was fitted with eddy covariance measurements of fogwater flux in a coniferous forest in the Netherlands:

$$V_d = 0.195u*^2. \tag{5}$$

Katata et al. [2008] proposed another formulation of V_d (namely, equation (2), which is shown again below) based on numerical simulations using SOLVEG at a coniferous forest in Germany. The equation is represented as a function of the horizontal wind speed over the canopy (U), LAI, and the canopy height (h):

$$V_d = A_i U, \tag{6}$$

$$A_i = 0.0164(\mathrm{LAI}/h)^{-0.5} \text{ for coniferous trees } (A_c). \tag{7}$$

Through equation (6), the calculations of A_c agreed with observations in various cloud forests with LAI/$h > 0.2$ [*Katata et al.*, 2008]. Recently, equation (6) was extended to sparse canopies and various vegetation species in order to develop a simple and accurate Fog Deposition EStimation (FogDES) scheme for meteorological models [*Katata and Nagai*, 2013]. The calculation result by *Katata et al.* [2008] for all combinations of h (4–34 m) and small LAI values (0.1–2) was fitted as a function of

$$A_c = 0.0095\mathrm{LAI}^3 - 0.05\mathrm{LAI}^2 + 0.0916\mathrm{LAI} + 0.0082. \tag{8}$$

Assuming that the relationship of equations (6) and (7) for dense and sparse canopies, respectively, holds for other vegetation types as well, the FogDES scheme can calculate the removal efficiency of fog droplets by the canopy, A_i, for each vegetation category, i, as

$$A_i = R_i A_c, \tag{9}$$

where R_i is the ratio of A_i to A_c (i.e., $R_i = 1$ for coniferous trees). According to comparisons in the calculated V_d between coniferous and broad-leaved forests with the same canopy height (h) and LAI [*Katata et al.*, 2008], the value of R_i for broad-leaved forests was 0.826. For short vegetation, additional SOLVEG calculations of the grassland with $h = 0.5$ m and LAI = 1 under the same simulation conditions of *Katata et al.* [2008] were carried out. In this calculation, the linear relationship between U and V_d used in equation (2) was also consistent ($R = 0.95$) and the R_i for grassland was determined to be 0.217 [*Katata and Nagai*, 2013]. For relatively smooth surfaces such as bare soil and water, although those are beyond the primary scope of this paper, the mechanism of gravitational settling is assumed to be dominant (i.e., R_i ~0). Thus, V_d is given as the gravitational settling velocity with a modified Gamma size distribution of fog droplets based on the data at a coniferous forest in Germany [*Katata et al.*, 2008]. The scheme is incorporated into the latest version of the meteorological model Advanced Research Weather Research and Forecasting (WRF) Version 3.5 [*Skamarock et al.*, 2008]. Changes in A_i for various vegetation types against the LAI values calculated using the FogDES scheme from equations (6)–(8) are illustrated in Figure 4a. Parameters used for the FogDES scheme are summarized in Table 4.

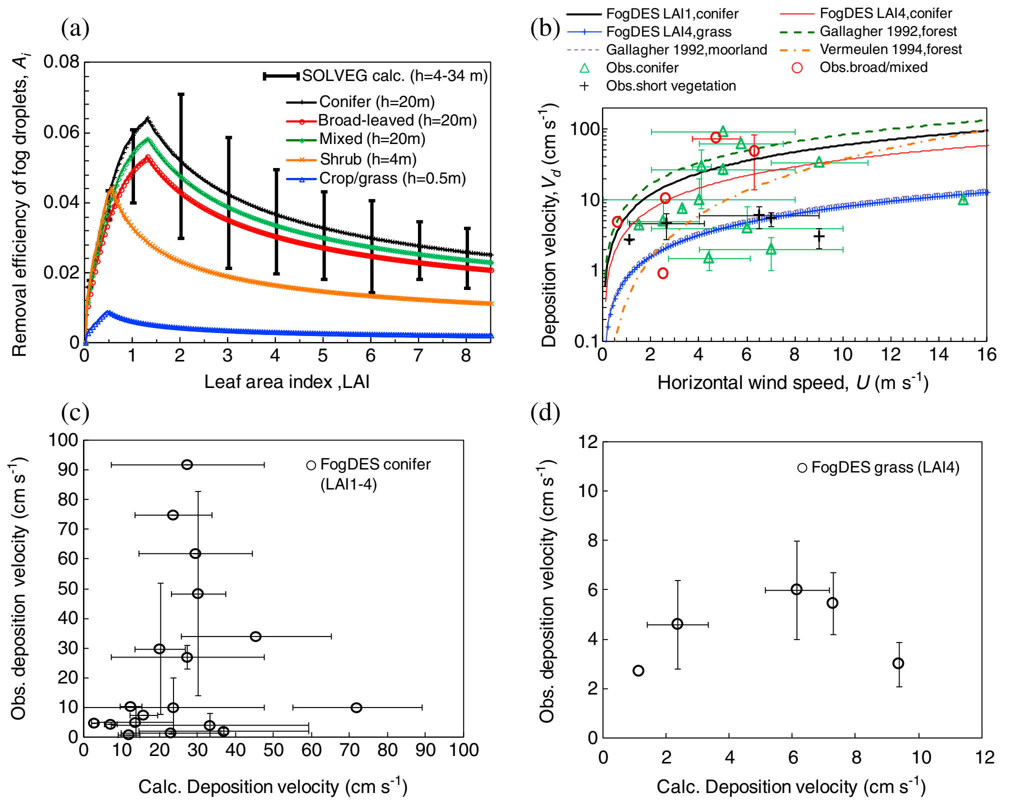

Figure 4. (a) Changes of the removal efficiency of fog droplets by the canopy (A_i) against leaf area index (LAI) for various vegetation categories in the simple and accurate Fog Deposition EStimation (FogDES) scheme [Eqs. (6)–(8)] and (b) fogwater deposition velocity by various parameterizations [FogDES, *Gallagher et al*., 1992; *Vermeulen et al*., 1997] for meteorological models introduced in section 5.2 against horizontal wind speed over the canopy. The SOLVEG calculations of A_i with canopy heights (h) of 4–34 m and leaf area index (LAI) values of 1–8 (black vertical bars), and measured deposition velocities at short and tall vegetation (open circles, triangles, and crosses) with the range from minimum to maximum values at the same site (bars) listed in Table 3 are also plotted in Figures 4a and 4b, respectively. (c and d) Scatter plots between observations and FogDES calculations for tall and short vegetation, respectively. Error bars in both graphs showed the range from minimum to maximum for observed wind speed and LAI = 1–4.

Figure 4b shows the comparison of fogwater deposition velocities for coniferous forest against U estimated by the above three parameterizations. Note that u_* was converted to U based on a logarithmic wind profile with displacement and canopy heights of $d_0 = 13$ m and $h = 20$ m, respectively [*Brutsaert*, 1975]. Additionally, a roughness length of 1 m was used, as this is a typical value for forests [*Wieringa*, 1980]. For equations (3) and (4),

Table 4. Input Parameters for Calculations of the Removal Efficiency of Fog Droplets by the Canopy (A_i) for Various Vegetation Species used in the Simple and Accurate Fog Deposition EStimation (FogDES) Scheme Represented as Equations (6)–(8)

Land-Use Type, i	Canopy Height, h (m)	Ratio of the Removal Efficiency, R_i	Leaf Size, d_{leaf} (mm)	Notes and References
Coniferous forest	20.0	1.000	1	*Katata et al.* [2008]
Broad-leaved forest	20.0	0.826	30	*Katata et al.* [2008]
Mixed forest	20.0	0.913	1–30	Weighting average of coniferous and broad-leaved forest
Shrubland	4.0	1.000	1	R_i assumed to be same as coniferous forest
Cropland and grassland	0.5	0.217	2	*Katata and Nagai* [2013]
Mixed crop/grass/woodland	11.5	0.609	1–2	Weighting average of grassland and coniferous forest
Others (bare soil, water, city, etc.)	–	0.000	–	assumed R_i =0 and gravitational settling only

fog droplet diameter d_p was set to a typical value of 20 μm. Most of the parameterizations showed a linear increase of deposition velocity with U (Figure 4b). The parameterization of deposition velocity represented as equation (5) calculated smaller values in the range of the fitting data [*Vermeulen et al*., 1997], whereby $U < 4$ m s^{-1}. This may be explained by differences in fog types; *Vermeulen et al*. [1997] derives equation (5) by fitting data of low elevation advection or radiation fog, while the other parameterizations are based on measurements during hillside or mountain fog events.

Figures 4c and d depict the comparisons between observed and calculated deposition velocity for tall and short vegetation. Observational data for the comparisons are summarized in Table 2. The data collected at forests are highly scattered because of the differences in canopy structure and uncertainties in the measurement methods (section 4.3). In contrast, the data over short vegetation (black crosses in Figure 4b; Table 2, No. 1g–5g) show smaller variability but also indicate that the efficiency of this type of vegetation for removing fog droplets is smaller than that of voluminous and complexly structured forest canopies. The parameterizations for grassland predicted the observed deposition velocity reasonably well except for the one data of $V_d = 2.1–3.9$ at $U = 9.0$ from *Fowler et al*. [1990] (Figure 4d). This agreement between model results and observations for grassland may be explained by the fact that observations were made over relatively homogeneous terrain with very small effects from the canopy edge (see section 4.2).

5.3. Challenges in Implementation of Fogwater Deposition Scheme

Although the parameterizations introduced in the last section enable to estimate fogwater deposition in various spatial scales, there are several difficulties to accurately implement fogwater deposition. The first problem is how to treat the fogwater deposition near the forest edge (section 4.2). As shown in Table 2, the edge effect can produce 300–400% of fogwater deposition within the forest in the sensitivity tests of 1-D fogwater deposition models. This enhancement may be observed according to the recent studies of meteorological models that include a fogwater deposition parameterization. *Katata et al*. [2011] compared throughfall-based observations and SOLVEG calculations of fogwater deposition at Mount Rokko in Japan. They showed that although the trend of observed fogwater deposition events was reproduced by SOLVEG, the cumulative amount of modeled fogwater deposition was approximately one-quarter of that from throughfall observations at the end of the simulation. This large difference between calculations and observations could not be sufficiently explained by uncertainties in both the effects of LAI and the droplet size distribution. If the effect estimated by the empirical formula of *Draaijers et al*. [1994] over a detailed topographical map was excluded from throughfall observations, the values decreased and agreed with the calculations. Based on this, they suggested that fogwater deposition was strongly enhanced at the forest edge. More recently, an advanced approach with a 2-D fogwater deposition model has been proposed to directly simulate the edge effect of fogwater deposition [*Shimadera et al*., 2011]. Sensitivity tests in this paper showed that deposition velocity at the edge was 1.5–4 times larger than that in closed forest canopies. These results demonstrate that the edge effect can create individual "hot spots" in heterogeneous terrain where the fogwater deposition amount is very high [*Juvik et al*., 2011]. This causes a serious underestimation of local-scale fogwater deposition in complex terrains. The effect is considered to be significant for estimates of overall fog input over in high level of fragmentation of a cloud forest (i.e., the high ratio of forest edge length per unit surface area).

In addition to the edge effect, wind-driven rainfall also has a large impact on hydrological input to the canopy under strong wind and drizzle conditions. *Schmid et al*. [2011] revealed it based on stable isotopic measurements that throughfall is greatly affected by the wind-driven rainfall. As the edge effect, the water deposition of larger droplets (i.e., drizzle) may also cause the underestimation of the current fogwater deposition model [*Katata et al*., 2011]. This effect is completely missing in the current fogwater deposition model, and should be considered in future modeling studies.

Finally, the widely overlooked aspect of fogwater deposition to forest canopies is that it removes a large amount of cloud water from the atmosphere. This effect is rarely taken into account in the state-of-art 3-D fog models [e.g., *Tang et al*., 2009; *Müller et al*., 2010]. In general, they consider only the processes of gravitational settling of fog droplets and exchange of momentum, energy, and water vapor over the surface [*Gultepe et al*., 2007]. Since inertial impaction of fog droplets by the canopy is usually a dominant removal process from the atmosphere under high wind conditions (section 4.2), the parameterizations of fogwater deposition as introduced in the last section have an important role in LWC prediction at the lowest atmospheric layer [*Katata et al*., 2011]. Hence, the effect of cloud water removal should be included in meteorological models for accurate fog simulation.

6. Concluding Remarks

This paper reviewed recent progress in fogwater deposition modeling over terrestrial ecosystems. A large portion of the studies that investigated the role and impact of fogwater deposition on water, energy, and CO_2 exchanges in cloud forests with numerical simulations was based on observational evidence, while such studies appear to still be very limited. Further work is needed to understanding the complicated linkage between fog or low cloud immersion and ecosystem carbon and nitrogen cycles. Three different types of 1-D models (resistance, analytical, and sophisticated models) have been developed and improved to reduce uncertainties associated with the input data and model parameters. The sophisticated models are now capable of incorporating detailed atmosphere-land interaction processes. By comparing the sophisticated model calculations with literature values observed by eddy covariance and canopy water balance methods, it was confirmed that denser forest canopies tended to capture lower amounts of fog droplets, and the fogwater deposition velocity over forests was several times that over short vegetation such as grassland. A tendency for the deposition velocity to be proportional to the horizontal wind speed was found in both the model and observational results. However, qualitative comparisons were very difficult to make because of the very large variety in observational data ($V_d = 2.1$–8.0 cm s^{-1} for grasslands; $V_d = 7.7$–92 cm s^{-1} and 0–20 cm s^{-1} for forests by throughfall or canopy water balance methods, and the eddy covariance method, respectively). This variety was probably due to landscape heterogeneity, wind-driven rainfall, the number of collectors used (for throughfall measurements), the droplet size distribution, and advection and condensation effects (for eddy covariance measurements). Sensitivity analyses of data from multilayer models revealed key factors causing uncertainties in the model results such as the edge effect, canopy structure representation, droplet size distribution, collection rate by leaves, turbulent exchange processes within canopies, and evaporation from wet canopies. Meanwhile, simple parameterizations of deposition velocity have been applied to long-term estimates and spatial distribution analyses of fogwater deposition over complex topography. This approach may be most useful for mapping fogwater deposition at regional and global scales, as local-scale simulations with fine horizontal grid resolutions (<1 km) cannot capture "hot spot" regions of extremely high fogwater deposition without taking the edge effect into account.

Compared to mountain (or upslope) fog, data on the deposition flux of radiation fog [*Burkard et al.*, 2003] were quite limited. Theoretically, sedimentation is responsible for most of the fogwater deposition under the typical conditions associated with radiation fog (wind speeds less than 2 m s^{-1}) [*Dollard and Unsworth*, 1983; *Lovett*, 1984]. *Thalmann et al.* [2002] found that an average of 81% of the fogwater flux to the ground was associated with sedimentation for radiation fog at an agricultural site in Switzerland. More recently, *Shimadera et al.* [2011] made comparisons of deposition velocities obtained with a 2-D model and eddy covariance observations at the Lägeren site in Switzerland (Table 3) [*Burkard et al.*, 2003]. They found low prediction accuracy for the fogwater deposition model because radiation fog was dominant for most of the events, and such events are associated with low wind speeds at the site. More comparison studies between models and observations are required for situations involving radiation fog.

Over the past few decades, the relevant literature shows that significant progress has been made in regards to model applications involving several datasets in such as USA [e.g., *Mohnen*, 1988] and Europe [e.g., *Wobrock et al.*, 1994; *Klemm and Wrzesinsky*, 2007]. In addition, parameterizations for fogwater deposition estimation over large spatial scales have been developed and these have been extended to various vegetation types. Nevertheless, quantitative verification of these models and parameterizations is extremely difficult at the current stage of development, especially in mountainous areas because of uncertainties in both model calculations and observational data. To reduce the discrepancies between models and observations, it is necessary to model the edge effect appropriately. It has been suggested in recent decades that estimates of atmospheric deposition to mountainous terrain that do not include landscape heterogeneity may be inadequate [*Weathers et al.*, 1995, 2000]. To predict "hot spot" regions of extremely high fogwater deposition, landscape features such as forest edges, elevation, aspect, and vegetation type should be considered in future modeling efforts. While these features may be important for high-resolution simulation of fogwater deposition over complex terrain, it is still a very difficult task to describe complex structures of forest environments in 1-D models. The 2-D modeling approach [*Shimadera et al.*, 2011] is also challenging as it requires additional computational costs for large scale estimates of fogwater deposition at a high spatial resolution. As demonstrated in section 4.2, the classical approach for simulating the edge effect is that the aerodynamic resistance when the horizontal wind faces the edge is to set zero (or a very small number). If this is combined with remote sensing data resolving the area fraction covered with forest canopies, I believe this can become a valuable low-cost method for fogwater deposition estimation in complex terrain.

In addition to advanced modeling techniques, a new method to reduce uncertainties from experimental measurements for forests is also needed. There are two aspects that need to be addressed in particular. These include (1) reducing uncertainty of experimental estimates and (2) more precisely eliminating discrepancies between what the models actually model and what the measurements provide. In terms of (1), a consistent method to evaluate the uncertainties due to the artifact, edge effect, wind-driven rainfall under strong winds, advection, and condensation over upslopes should be developed from the datasets collected by two common approaches of canopy water balance and eddy covariance methods. A typical example of (2) is the problem with measured eddy covariance net fluxes versus gross fluxes at the leaf surface level that the model outputs, as discussed in section 4.3. Advanced models that are verified at the same (leaf surface) level using datasets with small uncertainties have the potential to contribute to increased knowledge regarding the impacts of future land-use and climatic change on cloud forests [*Still et al.*, 1999; *Foster*, 2001; *Williams et al.*, 2007; *Bruijnzeel et al.*, 2011].

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