PLATIN (PLant-ATmosphere INteraction) - a model of biosphere/atmosphere exchange of latent and sensible heat, trace gases and fine-particle constituents

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Abstract

The exchange of energy and matter between phytosphere and near-surface atmosphere is a complex process controlled by a number of influence factors. Modelling has not only to consider the state of the air above and within the plant canopy (temperature, humidity, flow velocity, gas or particle concentration in the air) and the air’s transport capability, but also several physical, physiological, and chemical properties of the vegetation (plant architecture, vertically varying capability to receive or emit energy and gases, water budget, chemical reactions).

The SVAT model PLATIN (PLant-ATmosphere-INteraction) presented here is, like numerous other SVAT models, based on the big-leaf concept in order to reduce modelling complexity. The big-leaf approach replaces the vertical resolution of sources and sinks within the plant stand (including the soil surface beneath) by the idea of a single big leaf with overall properties equivalent to those of the complete plant/soil-surface system. The core module of PLATIN is based on the canopy energy budget and calculates the exchange of sensible and latent heat between phytosphere and near-surface atmosphere. Coupled to this the vertical exchange of matter is quantified.

In order to improve the treatment of the influence of the vertical light distribution within the canopy as well as to provide an additional way to validate the model, PLATIN has been extended by a submodule to estimate the stomatal uptake of trace gases (e.g. ozone) by the two different categories of sunlit and shaded leaves. This is achieved by extending the big-leaf concept by subdividing the big-leaf into a sunlit and a shaded fraction. One of the results obtained by this submodule is the stomatal conductance for sunlit leaves normalized by the leaf area index. This stomatal conductance represents an interface to measurements of trace gas exchange on leaf level.

Keywords: biosphere/atmosphere exchange, big leaf approach, latent and sensible heat, trace gas, fine particle, modelling

Zusammenfassung

PLATIN (PLant-ATmosphere INteraction) - ein Modell zur Bestimmung des Austausches von latenter und fühlbarer Wärme, Spurengasen und Schwebstaubinhaltsstoffen zwischen bodennaher Atmosphäre und Phytosphäre


Schlüsselworte: Austausch zwischen Biosphäre und bodennaher Atmosphäre, big-leaf-Konzept, latente und fühlbare Wärme, Spurengase, Schwebstaubinhaltsstoffe, Modellierung

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1 Introduction

Classical air pollution problems caused by very high concentrations of sulphur dioxide ($SO_2$) and London-type smog have decreased to acceptable levels in most parts of Europe. Nevertheless, there are still a number of potential ecological threats such as acidification and eutrophication of terrestrial and aquatic ecosystems, increased tropospheric ozone ($O_3$) concentrations and stratospheric ozone depletion, as well as greenhouse effects and human health problems caused by suspended particulate matter. Reactive atmospheric nitrogen species contribute to all these phenomena (cf. Dämmgen and Sutton, 2001; Erismann et al., 1998; Graedel and Crutzen, 1995).

During the 1970s it was recognised that transboundary air pollution has ecological as well as economic consequences e.g. for the forest and fish industries (UNECE, 2004). As a consequence, the countries of the UNECE (UN Economic Commission for Europe) developed a legal, organisational and scientific framework to deal with these problems. In 1979 the UNECE Convention on Long-Range Transboundary Air Pollution (LRTAP) was signed; it entered into force in 1983 (UNECE, 1979). In this context, the so-called multi-pollutant multi-effect or Gothenburg protocol (UNECE, 1999) requires the quantification - or at least estimation - of fluxes of atmospheric reactive nitrogen and sulphur species as well as of ozone and particulate matter between the ecosystems under consideration and the atmosphere near the ground.

Ideally, fluxes should be measured continuously and in an area-covering manner. Of course, this is not feasible. Another problem is that for some air constituents the toxicologically relevant flux is only a part of the total flux. Therefore modelling of fluxes has become a useful tool. Measurement and modelling techniques separate into two main categories, according to the type of species under consideration and their deposition properties: gases and fine particles ($0.002 \, \mu m < d_p < 2.5 \, \mu m$, with $d_p$ the aerodynamic diameter of particles) on the one hand and coarse particles ($d_p \geq 2.5 \, \mu m$; Finlayson-Pitts and Pitts, 1986; Gallagher et al., 1997) on the other hand. 'Particles' in this context may be solid or liquid (including rain and cloud drops). In general, fluxes of inert gases or fine particles are governed by turbulent diffusion in the atmosphere, by molecular diffusion within the (quasi-laminar) boundary layer adjacent to plant and soil surfaces, and by chemical reactions at the surfaces. In case of reactive gases or fine particles, also chemical reactions in the air have to be taken into account. Fluxes of very large particles ($d_p > 100 \, \mu m$) are predominantly controlled by gravitational forces whereas fluxes of smaller particles ($d_p < 100 \, \mu m$) are a result of diffusive, gravitational and inertial effects (interception, including impaction and turbulent inertial effects), depending on particle size and density (cf. Slinn 1982, Grünhage et al. 1998). Overviews on monitoring and modelling of biosphere/atmosphere exchange of gases, fine and coarse particles as well as of wet deposition are given in Dämmgen et al. (1997), Grünhage et al. (2000), Krupa (2002), Dämmgen et al. (2005) and Erismann et al. (2005).

Modelling of biosphere/atmosphere exchange of gases and fine-particle constituents also depends on the resolution in space and time needed. Whereas local scale Soil-Vegetation-Atmosphere-Transfer (SVAT) models rely on the detailed description of the canopy energy balance of the ecosystem under consideration, regional or national scale models make use of simplifying and integrating assumptions and make use of typical deposition velocities rather than site-specific driving forces (cf. Erismann et al., 2005). At the European scale, flux estimates are based on large-scale modelled meteorology and concentration fields; ecosystem properties are replaced by those of a vegetation type (cf. Grünhage et al., 2004). Necessarily, the complexity of details and processes considered in flux modelling decreases with increasing scale in space and time. This means that those generalized approaches must be carefully calibrated by well validated local scale models.

SVAT models serve two purposes: (1) In agricultural and forest meteorology they are used to calculate water dynamics e.g. to predict irrigation; (2) in the context of the ecotoxicology of air constituents they are needed to derive dose-response relationships (cf. Dämmgen and Grünhage, 1998).

Any perturbation on plant or ecosystem level is a function of the absorbed dose, i.e. the integral of the absorbed flux density $F_{c,absorbed}$ over time (cumulative flux density). In the context of acidification and eutrophication of terrestrial ecosystems, $F_{c,absorbed}$ is the overall input of acidifying or eutrophying species into the system as a whole ($F_{c,absorbed} = F_{c,total}$). On the other hand, for $SO_2$ or for $O_3$ (in particular as phytotoxic agents), $F_{c,absorbed}$ is only a part of the total flux: the total flux $F_{c,total}$ must then be partitioned into fluxes (1) absorbed by the plant through the stomata and the cuticle ($F_{c, stom & cut}$), and (2) deposited on external plant surfaces and the soil ($F_{c,non-stomatal}$: combined non-stomatal deposition). Studies show that penetration through the cuticle can be neglected in comparison to stomatal uptake (cf. literature cited in Grünhage et al., 2000). For ammonia ($NH_3$) bi-directional fluxes have to be taken into account, because, dependent on the nitrogen status of the respective system, deposition or emission situations can occur.

Non-stomatal deposition of phytotoxic gases ($O_3, SO_2$) is toxicologically almost irrelevant under ambient conditions in Europe but nevertheless a considerable part of the total flux (Grünhage et al., 1998; Fowler et al., 2001; Gerosa et al., 2003, 2004). Modelling of stomatal behaviour is crucial for the establishment of dose-response relationships (cf. Dämmgen et al., 1997; Grünhage et al., 2004; Tuovinen...
et al., 2004). As illustrated by Grünhage et al. (2003), any parameterization of stomatal behaviour in SVAT models for this purpose has to be validated at least via measurements of canopy level water vapour exchange.

This paper, which is a contribution to the European BL-AFLUX joint programme (Biosphere Atmosphere Exchange of Pollutants; http://www.accent-network.org), presents important new aspects of an extended version of the big leaf SVAT model PLATIN (PLant-ATmosphere INteraction) published by Grünhage and Haenel (1997) for the estimation of the exchange of latent and sensible heat, trace gases and fine-particle constituents between the plant/soil system and the atmosphere near the ground. Already the former PLATIN model had been published (in a simplified version) as an EXCEL version (named WINDEP for Worksheet-INtegrated Deposition Estimation Programme, cf. Grünhage and Haenel, 2000) in order to allow users to easily reflect model structure and equations and to adapt the model to their own requirements. The new PLATIN model will be available as ‘PLATIN for Excel’ via download from: http://www.uni-giessen.de/cms/ukl-en/PLATIN

PLATIN consists of several modules, as is illustrated in Figure 1. Except for canopy development modelling (which is treated elsewhere, e.g. for wheat in Grünhage et al., 1999), all other modules and processes indicated in Figure 1 are presented in a full model description in the special issue 319 of this journal (Grünhage and Haenel, 2008). This special issue also comprises the modules quantifying biosphere/atmosphere exchange of O3, SO2, NH3, nitric oxide (NO), nitrogen dioxide (NO2), nitrous acid (HNO2) and nitric acid (HNO3) as well as of ammonium (NH4), nitrate (NO3) and sulphate (SO4) in fine particles. The full model description is completed by a list of data needed to calibrate and run PLATIN.

A comparison of measured and modelled flux densities of trace gases will be published elsewhere.

2 Biosphere/atmosphere exchange of latent and sensible heat

Vertical flux densities of energy are part of the typical entities governing structure and function of ecotopes (Dämmgen et al., 1997). Energy fluxes must be known to establish the biosphere’s energy budget, which, along with the budget of matter, is essential for the understanding of ecosystem behaviour. However, while energy fluxes between the near-surface atmosphere and the biosphere can be measured, it is far more difficult to derive the energy balance of the biosphere from measurements. Thus, a common approach has become to model the biosphere system. In general, this modelling is one-dimensional, i.e. based on the assumption that all properties be functions of height z only. A short description of model scheme principles is given in Grünhage et al. (2000).

The one-dimensional PLant-Atmosphere INteraction model (PLATIN) is based on the big leaf concept which assumes that the vertical distribution of sources and/or sinks of a scalar (sensible heat, latent heat, ozone or another trace gas) can be represented by a single source and/or sink at the big leaf surface located at the conceptual height z = d + z_{scalar}. It is convenient to assume that the roughness length for gaseous species e.g. \( z_{\text{scalar}}^{\text{O22}} \) equals the roughness length for sensible heat \( z_{\text{d}}^{\text{O22}} \).

The core module of PLATIN deals with the solution of the canopy energy balance defined for the big leaf surface by

\[ R_{\text{net}} = H + \lambda E + G \]  

with \( R_{\text{net}} \) net radiation balance [W·m\(^{-2}\)]
\( H \) turbulent vertical flux density of sensible heat [W·m\(^{-2}\)]
\( \lambda E \) turbulent vertical flux density of latent heat [W·m\(^{-2}\)]
\( G \) ground heat flux density [W·m\(^{-2}\)]

Net radiation balance \( R_{\text{net}} \) [W·m\(^{-2}\)] is preferably provided by measurements. Otherwise it can be estimated in parts
or completely as discussed in Appendix A (cf. Grünhage and Haenel, 2008). The same holds for the ground heat flux density \( G \) \([\text{W} \cdot \text{m}^{-2}]\) the approximation of which is described in Appendix D (cf. Grünhage and Haenel, 2008).

The calculation of the fluxes of sensible and latent heat, \( H \) and \( \lambda E \), (and of gas fluxes) is based on Ohm’s law making use of a resistance network as illustrated in Figure 2.

![Figure 2](image)

A resistance analogue for water vapour (modified from PORG, 1997)

There are three major resistance components (which will be discussed in more detail in subsequent chapters):

1. the atmospheric resistance \( R_{\text{atmosphere}}(d+z_{\text{om}}, z_{\text{atm}}) \) \([\text{s} \cdot \text{m}^{-1}]\), representing the atmospheric transport properties between the conceptual height of the momentum sink near the big leaf surface \( z = d + z_{\text{om}} \) and a reference height \( z_{\text{atm}} \) above the canopy, where \( d \) is the displacement height and \( z_{\text{om}} \) is the roughness height for momentum. (Atmospheric turbulence is driven both by mechanical and thermal forces. The latter intensifies the mechanically induced turbulence within periods of atmospheric heating during daylight hours (unstable atmospheric stratification), whereas it weakens mechanically induced turbulence during cooling periods especially in the night (stable atmospheric stratification). Atmospheric transport by molecular diffusion can be neglected under turbulent conditions. Therefore \( R_{\text{atmosphere}} \) can be approximated by \( R_{\text{atmosphere}} \cong R_{\text{atm}} \) where \( R_{\text{atm}} \) is the turbulent atmospheric resistance for sensible heat transfer including a correction for non-neutral atmospheric stability conditions.);

2. the quasi-laminar layer resistance \( R_{\text{quasi-laminar layer}} \) or \( R_{\text{h}} \) \([\text{s} \cdot \text{m}^{-1}]\) between momentum sink height \( z = d + z_{\text{om}} \) and the conceptual sink/source height for sensible heat and trace gases (including \( \text{H}_2\text{O} \)) at \( z = d + z_{\text{om}} \) and the quasi-laminar layer resistance for water vapour (modified from PORG, 1997) the atmospheric resistance (cf. Grünhage and Haenel, 2008). The same holds for the ground heat flux density \( G \) \([\text{W} \cdot \text{m}^{-2}]\) the approximation of which is described in Appendix D (cf. Grünhage and Haenel, 2008).

(2) the quasi-laminar layer resistance \( R_{\text{quasi-laminar layer}} \)

According to the Monin-Obukhov theory (Monin and Obukhov, 1954), the turbulent atmospheric resistance \( R_{\text{atm}} \) between two heights \( z_1 \) and \( z_2 \) \((z_1 < z_2)\) can be expressed by

\[
R_{\text{atm}}(z_1, z_2) = \kappa \cdot \frac{\ln(z_2 - d) - \ln(z_1 - d)}{z_2 - z_1} = \kappa \cdot \frac{\ln(z_2 - d)}{L} + \kappa \cdot \frac{\ln(z_1 - d)}{L} \tag{2}
\]

with \( z_1 \) e.g. momentum sink height \( d+z_{\text{om}} \),

\( z_2 \) e.g. reference height \( z_{\text{atm}} \) for actual air temperature \( T_a \) \([\text{°C}]\) or reference height \( z_{\text{atm},r} \) for a trace gas or fine-particle constituent

and \( L \) Monin-Obukhov length \([\text{m}]\)

\( \kappa \) dimensionless von Kármán constant \((\kappa = 0.41; \text{cf. Dyer, 1974})\)

\( u \) friction velocity \([\text{m} \cdot \text{s}^{-1}]\)

\( \Psi_h \) atmospheric stability function for sensible heat

For vegetation like wheat or forests the displacement height and the roughness length are usually approximated by \( d = 0.67 \cdot h \) and \( z_{\text{om}} = 0.13 \cdot h \), respectively, with \( h \) the canopy height (Brutsaert, 1984). A parameterization of canopy height \( h \) for spring and winter wheat as a function of phenological development is given in Grünhage et al. (1999).

Calculation of energy balance time series for growing agricultural crops requires also the definition of the roughness length for bare agricultural soil. There is no unique value for all types of soils and their possible surface states. Table 2.2 in Oke (1978) gives a range of 0.001 – 0.01 m (along with displacement height \( d = 0 \) m).

The roughness length for sensible heat \( z_{\text{om}} \) is smaller than \( z_{\text{om}} \). According to Figure 4.24 in Brutsaert (1984) a typical value of \( \ln(z_{\text{om}}/z_{\text{om}}) \) is 2 for grass and corn so that we assume a value of 2 to be representative also for agricultural crops. For forests \( \ln(z_{\text{om}}/z_{\text{om}}) = 1 \) seems to be an acceptable value. Note that we use \( \ln(z_{\text{om}}/z_{\text{om}}) = 2 \) also for bare soil.

Eq. (2) is based on Monin-Obukhov theory. Strictly, this theory is valid only above the roughness sublayer which may range up to 2 or 2.5 times the vegetation height over tall and very rough canopies. For discussion see e.g. Cellier and Brunet (1992). Except for maize it seems tolerable to use eq. (2) for agricultural crops without further
correction, because the height of the roughness sublayer is generally smaller than the typical agrometeorological reference height of 2 m (for e.g. air temperature measurements). Over forests, however, most often the reference height is located within or at least at the upper boundary of the roughness sublayer. For this case, PLATIN makes use of a modified resistance equation (see e.g. Sellers et al., 1986):

\[ R_{h,\text{forests}}(z_1, z_2) = \frac{R_a(z_1, z_2)}{2} \]

(3)

The friction velocity is given by:

\[ u_* = \left( \frac{\kappa}{\ell} \frac{\psi_m}{\ln \left( \frac{z_{\text{ref}, u} - d}{z_{\text{obs}}} \right)} \right) - \frac{\psi_m}{\ln \left( \frac{z_{\text{ref}, u} - d}{L} \right)} \]

(4)

with \( u(z_{\text{ref}}) \) horizontal wind velocity at reference height \( z_{\text{ref}, u} \) [m·s\(^{-1}\)]

\( \psi_m \) atmospheric stability function for momentum

The Monin-Obukhov length \( L \) (Monin and Obukhov, 1954) is defined as:

\[ L = \frac{\rho_{\text{m, moist air}}}{\kappa} \left( \frac{\bar{\theta} - u_*^3}{g} \right) \]

(5)

\[ \approx \frac{\rho_{\text{m, moist air}}}{\kappa} \left( \frac{\bar{\theta}(z_{\text{ref}}) - u_*^3}{g} \right) \]

with \( \bar{\theta} \) average potential temperature of the air layer under consideration [K]

\( g \) gravitational acceleration \( (g = 9.81 \text{ m/s}^2) \)

\( \rho_{\text{m, moist air}} \) density of moist air [kg·m\(^{-3}\)] at absolute temperature \( T \) \( (T = t + 273.15 \text{ K}) \) with \( t \) the actual air temperature \( (\text{°C}) \) measured at reference height \( z_{\text{ref}, T} \) (see Appendix F, eq. (F7) in Grünhage and Haenel, 2008))

\( c_{\text{p, moist air}} \) specific heat of moist air at a constant pressure \( (\text{m}^2·\text{s}^{-2}·\text{K}^{-1}) \)

(see Appendix F, eq. (F9) in Grünhage and Haenel, 2008))

It is sufficient to approximate the layer-average potential temperature by the potential temperature at reference height, \( \theta(z_{\text{ref}}) \), which is estimated from the actual air temperature \( \bar{T}(z_{\text{ref}}) \) according to (cf. Stull, 1988):

\[ \bar{T}(z_{\text{ref}}) = \bar{T}(z_{\text{ref}})_{\text{dry}} + \bar{T}(z_{\text{ref}})_{\text{dry}} \cdot \Gamma_d \]

(6)

with \( \Gamma_d \) dry adiabatic lapse rate \( \Gamma_d = -9.76 \text{ K·km}^{-1} \)

The atmospheric stability functions for momentum \( \Psi_m \) and sensible heat \( \Psi_h \) are given in Appendix C in Grünhage and Haenel (2008).

2.2 Quasi-laminar layer resistance for sensible heat and water vapour

The quasi-laminar layer resistance for water vapour \( R_{b, \text{H}_2\text{O}} \) is estimated according to the approach by Hicks et al. (1987) taking into account the empirical results for permeable rough canopies described by Brutsaert (1984); for details see Grünhage et al. (2000):

\[ R_{b, \text{H}_2\text{O}} = \left( \frac{S_{\text{C,H}_2\text{O}}}{\text{Pr}} \right)^{2/3} \]

(7)

with \( R_{b, \text{heat}} \) quasi-laminar layer resistance for sensible heat

\( S_{\text{C,H}_2\text{O}} \) Schmidt number for water vapour (the ratio of the kinematic viscosity of dry air and the molecular diffusivity of the respective trace gas)

\( \text{Pr} \) Prandtl number (the ratio of the kinematic viscosity of dry air and the molecular diffusivity of heat)

For water vapour, \( (S_{\text{C,H}_2\text{O}})^{2/3} \) is 0.90 (cf. Table 2 in Grünhage and Haenel, 2008).

2.3 Bulk canopy resistance for water vapour

The bulk canopy resistance \( R_{c, \text{H}_2\text{O}} \) is a composite resistance describing stomatal and cuticular transpiration and evaporation. \( R_{c, \text{H}_2\text{O}} \) can be approximated by a weighted combination of soil resistance \( R_{c, \text{soil}} \), bulk stomatal resistance \( R_{c, \text{stom}} \) and bulk cuticle resistance \( R_{c, \text{cut}} \) known for a fully developed canopy (without senescent leaves) under optimum conditions for maximal transpiration. The weights depend on the actual canopy development stage taking into account the transition from a dense canopy (one-sided leaf area index \( LAI = LAI_{\text{max}} \text{[m}^2·\text{m}^{-2} \)) to a sparse canopy:

\[ \frac{1}{R_{c, \text{H}_2\text{O}}} \left( [1 - \beta'] \left( \frac{1}{R_{c, \text{soil, H}_2\text{O}}} + \frac{1}{R_{c, \text{stom, H}_2\text{O}}} \right) + \beta \frac{1}{R_{c, \text{cut, H}_2\text{O}}} \right) \]

(8)

In order to keep as close as possible to the single-leaf representation of the biosphere in PLATIN, eq. (8) makes use of a weighted \( R_{c, \text{soil}} \) (cf. Grünhage et al., 2000) instead of an additional in-canopy scalar transport resistance,
where the coefficient $\beta$ must be unity for bare soil and approaches zero for a fully developed dense canopy. If all leaves could contribute to the energy and water exchange between canopy and atmosphere, the weight of the reciprocal sum of $R_{\text{stem, H2O}}$ and bulk cuticle resistance $R_{\text{c, out, H2O}}$ would be $(1 - \beta)$. However, as only non-senescent leaves are relevant, a modified weight $(1 - \beta^*)$ is introduced.

Grünhage and Haenel (1997) presented a plausible ad-hoc approach to estimate $(1 - \beta^*)$ and $\beta$. It was based on the fact that the vertical distribution of incoming radiation energy within the canopy is one of the main limiting factors for the total canopy energy and water budget. Grünhage and Haenel (1997) simply assumed the available radiation energy to decrease exponentially with increasing distance from the top of the canopy and introduced a vegetation-type specific coefficient $c_{\text{LAI}}$ to describe the attenuation effect. They defined:

$$1 - \beta^* = 1 - e^{-c_{\text{LAI}} \cdot LAI_{\text{non-senescent}}} \quad (9)$$

and

$$\beta = e^{-c_{\text{LAI}} \cdot LAI_{\text{total}}} \quad (10)$$

The expression $(1 - \beta^*)$ may be interpreted as the fraction of radiation intercepted by non-senescent (green) leaves which is given by $LAI_{\text{non-senescent}}$ (one-sided leaf area index of non-senescent leaves; $= \text{projected leaf area PLA according to UNECE (2004, 2007)}$). The weight $\beta$ estimates the fraction of radiation reaching the ground depending on one-sided total leaf area index $LAI_{\text{total}}$ (one-sided plus senescent leaves). For spring and winter wheat a parameterization to calculate $LAI_{\text{total}}$ and $LAI_{\text{non-senescent}}$ as a function of phenological stages is given in Grünhage et al. (1999).

As radiation distribution within the canopy is (at least) a function of the solar elevation angle $\phi$ and of leaf angle distribution, the same should hold for $c_{\text{LAI}}$. However, Grünhage and Haenel (1997) made successfully use of a constant value for $c_{\text{LAI}}$, only dependent on vegetation type. This constant value may be interpreted as an effective mean value. Coefficients averaged over all solar elevations are summarized for different vegetation types e.g. in Monteith and Unsworth (1990). For most vegetation types $c_{\text{LAI}}$ is in the range of 0.3 to 0.6 (Ross, 1981). This includes $c_{\text{LAI}} = 0.4$ for crops as described by Ritchie (1972) as well as 0.5 for spring wheat (Choudhury et al., 1987) and a maritime pine canopy (Granier and Louiseau, 1994). On the other hand, for canopies with predominantly horizontally arranged leaves (e.g. cabbage, clover) $c_{\text{LAI}}$ approaches 1.8 as can be deduced from Monteith (1965).

PLATIN now incorporates a canopy radiation submodel (see Appendix B in Grünhage and Haenel (2008)), which allows to calculate the vertical radiation energy distribution and related entities within the canopy. Therefore, the parameterizations of the weights $(1 - \beta)$ and $\beta$, i.e. eqs. (9) and (10) had to be reconsidered in so far, as it could be possible and reasonable to replace the externally given coefficient $c_{\text{LAI}}$ by an entity calculated by the canopy radiation model. However, as radiation distribution is only a predictor for the weights $(1 - \beta^*)$ and $\beta$, care had to be taken when adopting results from the new radiation model.

The canopy radiation model allows to calculate an attenuation coefficient $k_b$ similar to $c_{\text{LAI}}$, but dependent on solar height:

$$k_b \frac{k_{b,90^\circ}}{\sin \phi} = \beta \quad (11)$$

with $k_{b,90^\circ} = k_b$ value for solar elevations of $90^\circ$

According to eq. (13) in Sellers (1985), $k_{b,90^\circ}$ is 0.5 for spherically arranged leaves, 0.27 for vertical and 1.23 for horizontal leaves. However, there is no use to replace $c_{\text{LAI}}$ in (10) by $k_b$ according to (11), because (11) is valid only for daylight hours while the weight $\beta$ is needed also during night. Therefore it was decided to replace $c_{\text{LAI}}$ in the calculation of $\beta$ by $k_{b,\text{max}}$ rather than $k_b$:

$$\beta = e^{-k_{b,\text{max}} \cdot SAI} \quad (12)$$

with $SAI$ total surface area of the vegetation $[m^2 \cdot m^{-2}]$ and

$$k_{b,\text{max}} \frac{k_{b,90^\circ}}{\sin \phi_{\text{max}}} = \beta \quad (13)$$

where $\phi_{\text{max}}$ solar elevation at 12 h TST (true solar time)

A minor adjustment is the formal replacement of $LAI_{\text{total}}$ in eq. (10) by $SAI$, the total surface area of the vegetation. $SAI$ is set equal to $LAI_{\text{total}} + 1$ for forests (Tuovinen et al., 2004) and to $LAI_{\text{total}}$ for short vegetation (crops, grassland).

The advantage of eq. (12) over (10) is that the annual course of solar height is now accounted for. The switch from $c_{\text{LAI}}$ to $k_{b,\text{max}}$ does not change the results significantly.

Stomatal behaviour of the plants strongly depends on irradiance absorption. This means that, when simulating stomata-related processes, the fraction of radiation intercepted by the non-senescent leaves should be taken into account explicitly within the parameterization of $(1 - \beta^*)$. With entities calculated by the canopy radiation submodel, $(1 - \beta^*)$ has been redefined for the current PLATIN version as follows:
\[ 1 - \beta^* \frac{I_{\text{sunlit}}}{\text{PAR}} + \gamma I_{\text{shaded}} \]

with
- \( I_{\text{sunlit}} \) irradiance absorbed by the sunlit fraction of non-senescent leaves of the canopy \([\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}]\)
- \( I_{\text{shaded}} \) irradiance absorbed by the shaded fraction of non-senescent leaves of the canopy \([\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}]\)
- \( \text{PAR} \) photosynthetically active radiation measured above the canopy \([\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}]\)

Eq. (14) is needed as base for other calculations like the fractioning of the total ozone stomatal uptake into stomatal uptake by the sunlit and shaded leaf fraction of the canopy during daylight hours. Clearly, eq. (14) is meaningless during night time. But as the nocturnal stomatal uptake is of inferior importance, \( \beta^* \) can then simply be replaced by \( \beta \) according to eq. (12), but calculated with the non-senescent LAI:

**Soil resistance for water vapour**

\( R_{\text{soil,H2O}} \) is a complex function of vertical soil water distribution. An important feature of evaporation from bare soil is a fast reduction due to the drying of the uppermost soil layer after rainfall. Therefore, \( R_{\text{soil,H2O}} \) is parameterized in the following manner:

(a) For a fully wet soil, \( R_{\text{soil,H2O}} \) equals \( R_{\text{soil,H2O, min}} \) \((=100 \text{ s.m}^{-1})\).
(b) For daylight hours (i.e. time intervals with global radiation \( S \geq 50 \text{ W.m}^{-2} \)), \( R_{\text{soil,H2O}} \) is increased by a given fraction of \( R_{\text{soil,H2O, min}} \) if there is no precipitation:

\[ (R_{\text{soil,H2O}})_n = (R_{\text{soil,H2O}})_{n-1} + \alpha \cdot (R_{\text{soil,H2O, min}}) \]

where \( n \) is the index of the data set under consideration and \( n-1 \) denotes the previous data set. RX is chosen to be 0.05 for half-hourly data sets and 0.1 for hourly data sets. \( R_{\text{soil,H2O}} \) is bound by the upper limit of 4000 \text{ s.m}^{-1}, the choice of which is based on the results of Daamen and Simmonds (1996).

(c) At night \( R_{\text{soil,H2O}} \) stays constant at the value calculated for the last late-afternoon daylight hour, i.e. \( (R_{\text{soil,H2O}})_n = (R_{\text{soil,H2O}})_{n-1} \).
(d) At any time interval with precipitation not reaching the ground, \( R_{\text{soil,H2O}} \) stays constant at the value calculated before, i.e. \( (R_{\text{soil,H2O}})_n = (R_{\text{soil,H2O}})_{n-1} \).
(e) At any time interval with precipitation and/or dew reaching the ground, \( R_{\text{soil,H2O}} \) is decreased by a fraction \( R_Y = a_{\text{soil}} \cdot W_{\text{in}} \) of \( R_{\text{soil,H2O, min}} \).

\[ (R_{\text{soil,H2O}})_n = (R_{\text{soil,H2O}})_{n-1} - a_{\text{soil}} \cdot W_{\text{in}} \cdot R_{\text{soil,H2O, min}} \]

with \( W_{\text{in}} \) amount of precipitation and/or dew reaching the ground (water input) \([\text{mm}]\)

For short vegetation (crops, grassland), the empirical constant \( a_{\text{soil}} \) is set to 10 mm\(^{-1}\) for half-hourly and 20 mm\(^{-1}\) for hourly data sets.

The amount of precipitation and/or dew reaching the ground depends on the interception reservoir capacity of the canopy. In PLATIN, this capacity \( INT_{\text{max}} \) \([\text{mm}]\) is assumed to be proportional to total LAI:

\[ INT_{\text{max}} = b_{\text{INT}} \cdot \text{LAI}_{\text{total}} \]

The constant \( b_{\text{INT}} \) is chosen as 0.2 mm according to Dickinson (1984), neglecting the fact that leaves become able to intercept more precipitation during senescence (cf. Braden, 1995).

The interception reservoir is filled by precipitation \( \text{Precip} \) and dew and depleted by evaporation. Dew formation and depletion of the reservoir is estimated due to potential evapotranspiration rate \( E_{\text{pot}} \) \([\text{mm}]\) applying the Penman-Monteith approach (see Chapter 2.4) with \( R_{\text{c,H2O}} = 0 \text{ s.m}^{-1} \) assuming neutral atmospheric stratification. The interception \( INT \) \([\text{mm}]\) is parameterized according to

\[ INT_{n} = \text{Precip}_{n} + a_{\text{INT}} \cdot E_{\text{pot},n} \]

with \( 0 \leq INT_{n} \leq INT_{\text{max}} \). The precipitation and dew reaching the ground \( W_{\text{in},n} \) is then given by:

\[ W_{\text{in},n} = \text{Precip}_{n} + (INT_{n} - b_{\text{INT}} \cdot \text{LAI}_{\text{total}}) \]

with \( W_{\text{in}} \geq 0 \text{ mm} \).

For forests, the coefficients RX and \( a_{\text{soil}} \) have not yet been properly adjusted. As a plausible working model, applicable to forests in Central Europe with generally non-drying soil, \( R_{\text{soil,H2O}} \) can be set to \( R_{\text{soil,H2O, min}} \).

Note: \( R_{\text{c,H2O}} \) is set to zero if the interception reservoir is not empty. Comparisons of modelled evapotranspiration rates with measured fluxes show that setting \( R_{\text{c,H2O}} \) to zero overestimates the real fluxes. Therefore, interception is not taken into account in latent heat flux modelling at present (see Chapter 2.4).

**Bulk cuticle resistance for water vapour**

Investigations of cuticular permeability of water vapour and other trace gases show that penetration through the cuticle can be neglected in comparison to stomatal ex-
change (Kerstiens and Lendzian, 1989a, b; Lendzian and Kerstiens, 1991; Kerstiens et al., 1992). According to the aforementioned authors \( R_{\text{cut, H2O}} \) on leaf basis is \( 9 \times 10^3 \text{ s}^{-1} \text{ m}^{-1} \) (cf. Table 3 in Grünhage and Haenel (2008)). According to Grünhage et al. (1999) resistances derived on leaf basis are upscaled to canopy level taking into account the PLATIN formulation of canopy architecture and radiation distribution within the canopy. Similar to the minimum value of the bulk stomatal resistance \( R_{c,\text{stom, min, H2O}} \), which is representative for a fully developed canopy (without senescent leaves) under optimum conditions for maximal transpiration, up-scaling from leaf to canopy level is performed applying \( k_{b,\text{max}} \) at maximum solar elevation of the year (summer solstice):

\[
R_{\text{canopy}} = \frac{R_{\text{leaf, literature}}}{\left( 1 - e^{-k_{b,\text{max, summer solstice}} \cdot LAI_{\text{leaf, literature}}} \right)^{20}}
\]

**Bulk stomatal resistance for water vapour**

The gas transfer through the stomata is by molecular diffusion. An inverse dependence of stomatal resistance on molecular diffusivity is generally accepted. In PLATIN, the dependence of stomatal resistance on radiation, temperature and the water budgets of atmosphere and soil as well as on modifying influence of time of day, phenology, ozone and \( \text{CO}_2 \) is described according to the Jarvis-Stewart approach (Jarvis, 1976; Stewart, 1988):

\[
R_{c,\text{stom, H2O}} = \left[ \frac{1}{R_{c,\text{stom, min, H2O}}} \right]^{-1} = f_1(S_s) \cdot f_2(t_s) \cdot f_3(VPD, SM) \cdot f_4(\text{time}) \cdot f_5(\text{PHEN}) \cdot f_6(\text{O}_3) \cdot f_8(\text{CO}_2)
\]

or

\[
R_{c,\text{stom, H2O}} = \left[ \frac{1}{R_{c,\text{stom, min, H2O}}} \right]^{-1} = f_1(S_s) \cdot f_2(t_s) \cdot f_3(VPD) \cdot f_4(SM) \cdot f_5(\text{time}) \cdot f_6(\text{PHEN}) \cdot f_7(\text{O}_3) \cdot f_8(\text{CO}_2)
\]

where \( R_{c,\text{stom, min, H2O}} \) represents the minimum value of the stomatal resistance for water vapour of the respective ecosystem. Functions \( f_1(S_s) \), \( f_2(t_s) \), \( f_3(VPD) \) and \( f_4(SM) \) account for the effects of solar radiation \( S_s \) [W m\(^{-2}\)], air temperature \( t_s \) [°C], water vapour pressure deficit of the atmosphere \( \text{VPD} \) [hPa] and soil moisture \( SM \) [m\(^3\) m\(^{-3}\)] on stomatal aperture (0 ≤ \( f_s \) ≤ 1). While eq. (22) is based on a multiplicative dependence of stomatal resistance on \( \text{VPD} \) and \( SM \) in PLATIN, a combined function \( f_{3/4}(\text{VPD}, SM) \) is preferred for biological reasons. A combined function \( f_{3/4}(\text{VPD}, SM) \) reflects the observation that increasing soil moisture deficits strongly influence stomatal closure due to \( \text{VPD} \). It is recommended to use measured soil moisture content \( SM \) for \( f_{3/4}(\text{VPD}, SM) \) or \( f_4(SM) \). If no \( SM \) data are available they must be simulated by a soil water model, a simple one is described in Appendix H in Grünhage and Haenel (2008). With \( f_{5}(\text{time}) \) a time-dependent impact on stomatal resistance can be taken into account (cf. Körner, 1994). \( f_6(\text{PHEN}) \) and \( f_7(\text{O}_3) \) represent the influence of phenology and ozone on stomatal resistance: both senescence due to natural ageing and premature senescence induced by ozone are limiting factors for stomatal aperture. The Jarvis-Stewart functions used are described in Appendix E in Grünhage and Haenel (2008). For \( S_s = 0 \), \( R_{c,\text{stom, H2O}} \) is set to 20000 s\(^{-1} \text{m}^{-1} \). Under ambient conditions with elevated \( \text{CO}_2 \) the influence of elevated \( \text{CO}_2 \) on stomatal aperture must be taken into account by an additional Jarvis-Stewart function \( f_8(\text{CO}_2) \).

**2.4 Latent and sensible heat flux densities**

It is straightforward to formulate \( H \) and \( \lambda E \) as analogs of Ohm’s law and to use them to operate a SVAT model like PLATIN. However, for PLATIN another way has been chosen. Inserting the resulting resistance-based formula for \( H \) in eq. (1) and solving for \( \lambda E \) yields the well-known Penman-Monteith equation (Monteith, 1965):

Once \( R_{\text{net}}, G, \) and \( \lambda E \) are known (\( \lambda E \) according to eq. (23)), their values are inserted into eq. (1) to obtain the sensible heat flux \( H \) as residual. This procedure exactly yields the same results as if both \( \lambda E \) and \( H \) had been esti-
is that any kind of $\lambda E$ estimate can be entered instead of eq. (23). This may be of interest e.g. in the case that measured values of $\lambda E$ are available and shall be tested within the modelling frame, or that not all relevant data are available to use eq. (23) so that a less data-demanding approach must be taken to obtain $\lambda E$. However, as far as not mentioned otherwise, PLATIN makes use only of eq. (23).

\[
\lambda E = \frac{s_c \cdot (R_{net} - G) + \rho_{\text{moist air}} \cdot \varphi_{p, \text{moist air}}}{s_c + \gamma} \frac{V P D}{R_{\text{ah}}(d + z_{0m}, \theta_{\text{ref}, T})} + R_{\text{b, heat}}
\]

with $\gamma$ psychrometric constant (= 0.655 hPa·K⁻¹)

\[
V P D = \text{water vapour pressure deficit of the atmosphere [hPa] (see Appendix F, eq. (F) in Grünhage and Haenel (2008))}
\]

and

\[
s_c \cdot e_{\text{sat}}(T_s) - e_{\text{sat}}(T(\theta_{\text{ref}, T})) \over T_s - \theta(\theta_{\text{ref}, T})
\]

with $e_{\text{sat}}$ saturation water vapour pressure of the atmosphere [hPa] (see Appendix F, eqs. (F2) and (F3) in Grünhage and Haenel (2008))

and $T_s$ absolute canopy surface temperature at conceptual height $z = d + z_{0m}$ [K]

\[
\theta_s = \theta(\theta_{\text{ref}, T}) + H \cdot \left( R_{\text{ah}}(d + z_{0m}, \theta_{\text{ref}, T}) + R_{\text{b, heat}} \right)
\]

2.5 Comparison of measured and modelled latent and sensible heat flux densities

At the Linden grassland site, friction velocity, latent heat, as well as sensible heat are measured using the eddy covariance method by means of a Solent R3 research ultrasonic anemometer (Gill Instruments Ltd, Hampshire, UK) in combination with a Li-7500 open path CO₂/H₂O gas analyzer (Li-COR Environmental, Lincoln, Nebraska, USA). To guarantee data sets of high accuracy several corrections and quality tests are applied (WPL correction, Schotanus/Liu correction, coordinate rotation, footprint analysis, test to check the fulfilment of stationarity and of well developed turbulence conditions; cf. Grünhage and Gerosa, 2008).

Model adjustment is based on data sets for which the energy balance residual is less than 30 W·m⁻². A description how to estimate displacement height $d$, roughness length for momentum $z_{0w}$ and bulk canopy resistance for water vapour $R_{c, \text{H}_2\text{O}}$ can be found in Appendices I and J in Grünhage and Haenel (2008).

Figure 3 clearly illustrates that PLATIN is able to simulate measured fluxes adequately.

![Figure 3](image-url)

Comparison of measured and modelled sensible heat (a) and latent heat (b) for daylight hours in June 2004 over a semi-natural grassland at the Linden field site (for details see Grünhage et al., 1996; Jäger et al., 2003)
3 Biosphere/atmosphere exchange of trace gases

The exchange of trace gas species $A$, $F_c(A)$, between the phytosphere and the atmosphere near the surface can be modelled by:

$$F_c(A) = -\frac{\rho_A(z_{ref,A}) - \rho_A(d + z_{on})}{R_{sh}(d + z_{on})}$$  \hspace{1cm} (26)

$$\approx \frac{\rho_A(z_{ref,A}) - \rho_A(d + z_{on})}{R_{sh}(d + z_{on})} + \rho_{h,comp}$$  \hspace{1cm} (27)

$$= \frac{\rho_A(z_{ref,A}) - \rho_{A,comp}}{R_{sh}(d + z_{on})}$$  \hspace{1cm} (28)

with $F_c(A)$ total vertical atmosphere-canopy flux of trace gas $A$ [µg·m$^{-2}$·s$^{-1}$]

$\rho_A(z_{ref})$ measured concentration (potential) of trace gas $A$ at height $z = z_{ref,A}$ [µg·m$^{-2}$]

$\rho_A(d + z_{on})$ concentration of trace gas $A$ at the conceptual height $z = d + z_{on}$ [µg·m$^{-2}$]

$\rho_{A,comp}$ canopy compensation concentration of trace gas $A$ [µg·m$^{-3}$]

Modelling details are given in Grünhage and Haenel (2008). In case of deposition, the resistance network allows to partition the total atmosphere-canopy flux $F_c(total)(A)$ into (1) fluxes absorbed by the plant through the stomata and the cuticle $F_c(stom\&cut) = F_c(stom) + F_c(cut)(A)$, and (2) fluxes down to external plant surfaces $F_c(ext)(A)$ and the soil beneath the canopy $F_c(soil)(A)$. Studies show that penetration of gases through the cuticle $F_{c,cut}(A)$ can be neglected in comparison to stomatal uptake $F_{c,stom}(A)$ (cf. Chapter 2.3).

Combining $F_c(ext)$ and $F_c(soil)$ to $F_c(non-stomatal)$ and neglecting cuticular fluxes (i.e. approximating $F_{c,cut}$ = $F_{c, stom}$) one obtains:

$$F_{c,total}(A) = F_{c, stom\&cut}(A) = F_{c, ext}(A) = F_{c, soil}(A)$$  \hspace{1cm} (29)

The integral of $F_{c, stom\&cut}$ over time $t$ is the pollutant absorbed dose, $PAD(A)$ [µg·m$^{-2}$] (Fowler and Cape, 1982):

$$PAD(A) = \int_{t_1}^{t_2} |F_{c, stom\&cut}(A)| dt$$  \hspace{1cm} (30)

For $O_3$ the integral of $F_{c,stom\&cut} \approx F_{c,stom}$ over time $t$ is called accumulated stomatal flux of ozone, $AF_{stom}$ (UNECE, 2004, 2007).

Applying concentration- or flux-based critical levels of $O_3$ (UNECE, 2004, 2007) or the maximum-permissible $O_3$ concentration concept (Grünhage et al., 2001) requires $O_3$ concentration measured at reference height above the canopy $\rho_{O3}(z_{ref})$ to be transformed to concentration at the upper surface of the laminar boundary layer of the uppermost sunlit leaves. According to the single-leaf concept, PLATIN disposes only of one single canopy-representative laminar boundary layer the surface of which is located at $d + z_{on}$. The concentration at this height is calculated from:

$$\rho_{O3}(d + z_{on}) = \rho_{O3}(z_{ref}) \pm F_{c, total}(O_3) = R_{sh}(d + z_{on}, z_{ref, O3})$$  \hspace{1cm} (31)

This is contrasted by the UNECE (2004, 2007) approach that the concentration at the upper surface of the laminar boundary layer of the sunlit lower canopy leaves be represented by the $O_3$ concentration at the top of the canopy $z = h$. Within the M-O framework, this concentration is given by:

$$\rho_{O3}(h) = \rho_{O3}(z_{ref}) + F_c(total)(O_3) = R_{sh}(h, z_{ref, O3})$$  \hspace{1cm} (32)

Besides the fact that the effective upper surface of all the laminary boundary layers existing within the canopy is probably represented better by $d + z_{on}$ than by $h$, application of eq. (32) is prone to proper definition of $h$. Does $h$ e.g. represent the maximum or the average height of canopy elements above ground? Due to numerous irregularities in canopy architecture as well as wind-driven bending of the upper parts of a canopy it may be difficult to find a robust estimate of canopy height. Therefore it seems worthwhile to demonstrate the differences in stomatal uptake calculated according to eq. (33; see below) with $O_3$ concentrations from eq. (32) for varying $h$. This is done exemplarily using the daylight-hour data sets from June 2004 at Linden. Except for $h$, all data needed to evaluate eq. (32) were determined by precedent PLATIN runs using the aforementioned data sets.

In order to obtain a reasonable measure for $h$, we inverted the usual relations between displacement height and roughness length for momentum on one hand and the canopy height on the other hand, which allows to calculate $h$ from given displacement height $d$ and roughness length $z_{on}$ by $h = (d + z_{on}) - 0.8 - 1$ according to Brutsaert (1984). In order to reveal the influence of the differing heights in eqs. (31) and (32), the stomatal uptake as calculated with $O_3$ concentrations from eq. (32) has then been normalised by the one obtained with $O_3$ concentrations...
from eq. (31). As stomatal uptake is proportional to \( O_3 \) concentration, this ratio of stomatal uptake turns out to be identical to the ratio of the respective \( O_3 \) concentrations. This concentration ratio is displayed in Figure 4. Estimating stomatal uptake from \( h \) rather than \( d + z_{\text{on}} \) generally leads to overestimation of stomatal uptake which may not be negligible. This overestimation clearly depends on how much \( h \) differs from \( d + z_{\text{on}} \), which is demonstrated by varying \( h \) by plus 0.05 m and \( d + z_{\text{on}} \) by a factor of 1.1. (Diminishing \( h \), on the other hand, would mean to approach \( d + z_{\text{on}} \) and therefore reduce overestimation. However, \( h \) should never reach \( d + z_{\text{on}} \), because then \( h \) would not be representative for canopy height any longer.)

Even if a stomatal uptake approach as given by UNECE (2004, 2007) could be combined with guidance how to properly estimate canopy height \( h \), the problem still remains that \( h \) is not the best measure of the effective height of the canopy’s laminar boundary layer. Therefore we recommend to calculate stomatal uptake by a model like PLATIN which is calibrated by a number of water and energy balance quantities and which, therefore, is much less sensitive to the correct estimation of canopy height.

Because stomatal uptake of \( O_3 \), \( F_{c, \text{stom}}(O_3) \) is the toxically effective share of \( F_{c, \text{x} \sim}(O_3) \), flux-effect relationships should be based on that component which is given by:

\[
F_{c, \text{stom}}(O_3) = \frac{R_{\text{sh}}(d + z_{\text{on}}) + R_{c, \text{stom}, \text{mes}, O_3}}{R_{\text{sh}}(d + z_{\text{on}}) + R_{c, \text{stom}, \text{mes}, O_3} + R_{c, \text{stom}, \text{sunlit}, O_3} + R_{c, \text{stom}, \text{shaded}, O_3}} \quad (33)
\]

with

\[
R_{\text{sh}}(d + z_{\text{on}}) + R_{c, \text{stom}, \text{mes}, O_3} + \left( \frac{1}{1 - \beta^*} \right) \left( \frac{R_{\text{sh}}(d + z_{\text{on}}) + R_{c, \text{stom}, \text{mes}, O_3}}{R_{c, \text{stom}, \text{sunlit}, O_3} + R_{c, \text{stom}, \text{shaded}, O_3}} \right) = \left( \frac{1}{1 - \beta^*} \right) \left( \frac{1}{R_{\text{c, cut, O3}} + R_{c, \text{ext, O3}} + R_{\text{soil, O3}}} \right)
\]
As the $O_3$ concentration within the plant can be assumed to be zero, the relations of fluxes are simply given by the inverse ratio of the resistances involved which turns out to be a function of $\beta$, and $\beta\text{sunlit}$ or $\beta\text{shaded}$ respectively:

\[
\frac{F_{c, \text{stom, sunlit}}(O_3)}{F_{c, \text{stom, O}_3}} = \frac{R^*_{c, \text{stom, O}_3}}{R_{c, \text{stom, sunlit}}} \frac{1 - \beta_{\text{sunlit}}}{1 - \beta_{\text{O}_3}} \tag{41}
\]

\[
\frac{F_{c, \text{stom, shaded}}(O_3)}{F_{c, \text{stom, O}_3}} = \frac{R^*_{c, \text{stom, O}_3}}{R_{c, \text{stom, shaded}, O_3}} \frac{1 - \beta_{\text{shaded}}}{1 - \beta_{\text{O}_3}} \tag{42}
\]

Dividing $F_{c, \text{stom, sunlit}}(O_3)$ by LAI_{sunlit} (see Appendix B in Grünhage and Haenel (2008)) yields the flux of $O_3$ through the stomatal pores per unit projected leaf area (PLA)

\[
F_{\text{leaf, stom, sunlit}}(O_3) = \frac{F_{c, \text{stom, sunlit}}(O_3)}{\text{LAI}_{\text{sunlit}}} \tag{43}
\]

as required by the UNECE Mapping Manual 2004 (UN-ECE, 2004, 2007). The re-calculation of stomatal conductance of sunlit leaves $g_{\text{leaf, stom, sunlit}, O_3}$ from bulk stomatal resistance $R_{c, \text{stom, O}_3}$ according to eq. (44)

\[
g_{\text{leaf, stom, sunlit}, O_3} = \frac{1 - \beta_{\text{sunlit}}}{R_{c, \text{stom, O}_3}} \frac{1}{\text{LAI}_{\text{sunlit}}} \tag{44}
\]

provides a direct interface between canopy scale and leaf scale measurements as well as between micromet and impact research. Besides verification of the parameterization of stomatal conductance via measurements of canopy level water vapour exchange, big leaf stomatal conductance parameterization and water vapour fluxes can now be compared directly with porometer measurements on the leaf level. Upscaling algorithms from leaf to canopy level can be verified or adjusted.

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