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Detailed documentation of the PLATIN (PLant-ATmosphere INteraction) model

Ludger Grünhage and Hans-Dieter Haenel
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Johann Heinrich von Thünen-Institut
Bundesforschungsinstitut für
Ländliche Räume, Wald und Fischerei (vTI)
Bundesallee 50, 38116 Braunschweig,
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Ludger Grünhage1 and Hans-Dieter Haenel2

1 Institute for Plant Ecology, Justus-Liebig-University, Heinrich-Buff-Ring 26-32, D-35392 Gießen, Germany

2 Institute of Agricultural Climate Research, Johann Heinrich von Thünen-Institut, Federal Research Institute for Rural Area, Forestry and Fisheries, Bundesallee 50, D-38116 Braunschweig, Germany
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Ludger Grünhage¹ and Hans-Dieter Haenel²

Abstract

The exchange of energy and matter between phytosphere and near-surface atmosphere is a complex process controlled by a number of influence factors. These not only comprise 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).

Modelling the underlying processes requires a more or less extensive reduction of their complexity. The degree of simplification depends on what is to be modelled and on the availability of data to operate the model. The SVAT model PLATIN (PLant-ATmosphere-INteraction) presented here belongs to the category of models to be used for practical purposes e.g. in agriculture (the need for irrigation, among other things) or to establish dose-response functions in ecotoxicology. Like numerous other SVAT models PLATIN is based on the big-leaf concept which 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 exchange of trace gases and fine-particle constituents is quantified. The vertical transport between an above-canopy reference height, for which air properties and concentrations of matter must be known, and the sinks and/or sources of the plant/soil-surface system is modelled using three resistances: the turbulent atmospheric transport resistance between reference height and the level of momentum sink; the quasi-laminar resistance between momentum-sink level and the surface of the big leaf to account for the differences between momentum transfer and transport of energy and matter; canopy resistance which in turn is modelled using a number of further resistances arranged in series and in parallel. In general, the resistance values depend on the type of entity transported (momentum, heat, gaseous species, particles).

PLATIN calculates numerous detailed results which agree quite well with measurements, demonstrating good model performance. 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 was 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.

¹ Institute for Plant Ecology, Justus-Liebig-University, Heinrich-Buff-Ring 26-32, D-35392 Gießen, Germany
² Institute of Agricultural Climate Research, Johann Heinrich von Thunen Institute, Federal Research Institute for Rural Areas, Forestry and Fisheries, Bundesallee 50, D-38116 Braunschweig, Germany
Zusammenfassung


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; Erisman 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). Figure 1 shows a separation of air constituents with respect to their deposition properties (particle size and mass, state). 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 Erisman 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. Erisman 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 in 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 $F_{c, \text{absorbed}}$ over time (cumulative flux density). In the context of acidification and...
eutrophication of terrestrial ecosystems, \( F_{c,\text{absorbed}} \) is the overall input of acidifying or eutrophying species into the system as a whole (\( F_{c,\text{absorbed}} = F_{c,\text{total}} \)). On the other hand, for \( \text{SO}_2 \) or for \( \text{O}_3 \) (in particular as phytotoxic agents), \( F_{c,\text{absorbed}} \) is only a part of the total flux: the total flux \( F_{c,\text{total}} \) must then be partitioned into fluxes (1) absorbed by the plant through the stomata and the cuticle (\( F_{c,\text{stom \& cut}} \)), and (2) deposited on external plant surfaces and the soil (\( F_{c,\text{non-stomatal}} \); 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 (\( \text{NH}_3 \)) 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 (\( \text{O}_3, \text{SO}_2 \)) 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 BIAFLUX joint programme (Biosphere Atmosphere Exchange of Pollutants; http://www.accent-network.org), presents the documentation of an extended version of the \textit{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. In the EXCEL version of the extended PLATIN model, the equations presented in this paper are cited in the Excel spreadsheets with their original equation numbers. 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 2. The core module solves the canopy energy balance while simultaneously providing all the resistances also relevant for trace gas exchange. The description of the underlying processes is given in Chapter 2 with special aspects more detailed in the Appendices C, E, and F. As the functioning of the core module is essential for the entire performance of PLATIN, a short subsection at the end of Chapter 2 provides an overview-like comparison of measured and modelled latent and sensible heat fluxes above a semi-natural grassland in 2004 at the Linden field site (for details see Grünhage et al., 1996; Jäger et al., 2003).

The solution of the energy balance requires auxiliary modules (radiation, soil moisture, ground heat flux, and CO2 exchange) being described in the respective Appendices (A, B, H, D, and G). Modelling canopy development is understood as another auxiliary module, which, however, is treated elsewhere (e.g. for wheat in Grünhage et al., 1999).

Chapters 3 and 4 will present the modules quantifying biosphere/atmosphere exchange of O3, SO2, NH3, nitric oxide (NO), nitrogen dioxide (NO2), nitric acid (HNO3) and nitric acid (HNO3) as well as of ammonium (NH4), nitrate (NO3) and sulphate (SO4) in fine particles. Chemical sinks and/or sources between reference height and canopy surface are not taken into account. For a discussion see Grünhage et al. (2000). A special submodule described in Chapter 3 estimates the stomatal O3 uptake of the sunlit leaf fraction yielding the leaf area-related stomatal conductance for sunlit parts of the plant stand and thus providing an interface to measurements on gas exchange on leaf level.

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 \( z = d + z_{0,\text{scalar}} \). It is convenient to assume that the roughness length for gaseous species e.g. \( z_{0,\text{H}_2\text{O}} \) equals the roughness length for sensible heat \( z_{0,\text{h}} \).

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. The same holds for the ground heat flux density \( G \) [W m\(^{-2}\)] the approximation of which is described in Appendix D.

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 3.

![Fig. 3](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 \( R_{\text{atmosphere}}(d+z_{0,\text{m}}, z_{\text{ref}}) \) [s m\(^{-1}\)], representing the atmospheric transport properties between the conceptual height of the momentum sink near the big leaf surface \( z = d + z_{0,\text{m}} \) and a reference height \( z_{\text{ref}} \) above the canopy, where \( d \) is the displacement height and \( z_{0,\text{m}} \) is the roughness length 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 the atmosphere can be approximated by $R_{\text{atmosphere}} \approx R_{\text{ah}}$, where $R_{\text{ah}}$ 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 $R_b$ [s m$^{-1}$] between momentum sink height $z = d + z_{0m}$ and the conceptual sink/source height for sensible heat and trace gases (including H$_2$O) at $z = d + z_{0h}$; and

(3) the bulk canopy or surface resistance $R_{\text{canopy}}$ or $R_c$ [s m$^{-1}$], describing the influence of the plant/soil system on the vertical exchange of trace gases (including H$_2$O).

2.1 Turbulent atmospheric resistance

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

$$R_{\text{ah}}(z_1, z_2) = \frac{\ln\left(\frac{z_2 - d}{z_1 - d}\right) - \Psi_h\left(\frac{z_2 - d}{L}\right) + \Psi_h\left(\frac{z_1 - d}{L}\right)}{\kappa \cdot u_*}$$

with

- $z_1$ e.g. momentum sink height $d + z_{0m}$ [m]
- $z_2$ e.g. reference height $z_{\text{ref,T}}$ for actual air temperature $t_a$ [°C] or reference height $z_{\text{ref,A}}$ for a trace gas or fine-particle constituent
- $L$ Monin-Obukhov length [m]
- $\kappa$ dimensionless von Kármán constant ($\kappa = 0.41$; cf. Dyer, 1974)
- $u_*$ friction velocity [m 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 $z_{0m} = 0.13 \cdot h$, 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 $z_{0h}$ is smaller than $z_{0m}$. According to Figure 4.24 in Brutsaert (1984) a typical value of $\ln(z_{0m}/z_{0h})$ is 2 for grass and corn so that we assume a value of 2 to be representative also for agricultural crops. For forests $\ln(z_{0f}/z_{0m}) = 1$ seems to be an acceptable value. Note that we use $\ln(z_{0m}/z_{0h}) = 2$ 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):
The friction velocity is given by:

\[ u_* = \frac{\kappa \cdot u(z_{\text{ref}})}{\ln \left( \frac{z_{\text{ref}, u} - d}{z_0} \right) - \Psi_m \left( \frac{z_{\text{ref}, u} - d}{L} \right) + \Psi_m \left( \frac{z_0}{L} \right)} \]

with \( u(z_{\text{ref}}) \) horizontal wind velocity at reference height \( z_{\text{ref}, u} \) [m/s], \( \Psi_m \) atmospheric stability function for momentum

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

\[ L = - \rho_{\text{moist air}} \cdot c_{p, \text{moist air}} \cdot \frac{\bar{\theta} \cdot u_*^3}{\kappa \cdot g \cdot H} \]

\[ \approx - \rho_{\text{moist air}} \cdot c_{p, \text{moist air}} \cdot \frac{\theta(z_{\text{ref}}) \cdot u_*^3}{\kappa \cdot g \cdot H} \]

with \( \bar{\theta} \) average potential temperature of the air layer under consideration [K], \( g \) gravitational acceleration (\( g = 9.81 \text{ m/s}^2 \)), \( \rho_{\text{moist air}} \) density of moist air [kg/m\(^3\)] at absolute temperature \( T \) [\( T = t_a + 273.15 \text{ K} \)] with \( t_a \) the actual air temperature (°C) measured at reference height \( z_{\text{ref}, T} \) (see Appendix F, eq. (F7)), \( c_{p, \text{moist air}} \) specific heat of moist air at a constant pressure [m\(^2\) s\(^{-2}\) K\(^{-1}\)] (see Appendix F, eq. (F9))

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 \( T(z_{\text{ref}}) \) according to (cf. Stull, 1988):

\[ \theta(z_{\text{ref}}) = T(z_{\text{ref}}) + (z_{\text{ref}} \cdot \Gamma_d) \]

with \( \Gamma_d \) dry adiabatic lapse rate [K/km]

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

As \( u_* \) and \( H \) implicitly depend on \( L \), \( L \) must be calculated iteratively. For this purpose a so-called neutral value is defined: \( L_{\text{neutral}} = 10^2 \text{ m} \). (In reality \( L \) should approach plus or minus infinity under neutral atmospheric stability conditions. However, from the mathematical point of view it is sufficient to use a very high positive value of \( L \) as is done with \( L_{\text{neutral}} \).) Iteration starts with \( L = L_{\text{neutral}} \). In the rather rare case that iteration does not converge \( L \) is finally reset to \( L_{\text{neutral}} \). This may be incorrect to some extent but is estimated to be more effective than to drop the complete evaluation of the current data set (which would interrupt the calculation of storage terms like the interception reservoir or the soil water content). Of course the neutral value \( L = L_{\text{neutral}} \) also applies in situations where \( H = 0 \).

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

The quasi-laminar layer resistance for water vapour \( R_{b, \text{H}_2O} \) 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):
\[ R_{b,H2O} = R_{b,heat} \cdot \left( \frac{Sc_{H2O}}{Pr} \right)^{2/3} = \frac{\ln \left( \frac{z_{0n}}{z_{0h}} \right)}{\kappa \cdot u_*} + \frac{\Psi_h \left( \frac{z_{0h}}{L} \right)}{Pr \cdot \left( \frac{Sc_{H2O}}{Pr} \right)^{2/3}} \]  \hspace{1cm} (7)

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

\( Sc_{H2O} \) Schmidt number for water vapour (the ratio of the kinematic viscosity of dry air and the molecular diffusivity of the respective trace gas)

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

For water vapour, \( (Sc/Pr)^{2/3} \) is 0.90 (cf. Table 2, Chapter 3).

2.3 Bulk canopy resistance for water vapour

The bulk canopy resistance \( R_{c,H2O} \) is a composite resistance describing stomatal and cuticular transpiration and evaporation. \( R_{c,H2O} \) can be approximated by a weighted combination of soil resistance \( R_{soil} \), bulk stomatal resistance \( R_{c, stom} \) and bulk cuticle resistance \( R_{c, cut} \), 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 \( LAI = LAI_{max} \) \( [m^2 \cdot m^{-2}] \)) to a sparse canopy:

\[
\frac{1}{R_{c,H2O}} = \left[ (1-\beta^*) \left( \frac{1}{R_{c, stom, H2O}} + \frac{1}{R_{c, cut, H2O}} \right) + \frac{\beta}{R_{soil, H2O}} \right]
\]  \hspace{1cm} (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 \( R_{soil} \) (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 \( R_{c, stom, H2O} \) and bulk cuticle resistance \( R_{c, cut, H2O} \) 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 \( c_{LAI} \) to describe the attenuation effect. They defined:

\[
1-\beta^* = 1 - e^{-c_{LAI} \cdot LAI_{non-senescent}}
\]  \hspace{1cm} (9)

and

\[
\beta = e^{-c_{LAI} \cdot LAI_{total}}
\]  \hspace{1cm} (10)

The expression \((1-\beta^*)\) may be interpreted as the fraction of radiation intercepted by non-senescent (green) leaves which is given by \( LAI_{non-senescent} \) (= one-sided leaf area index of non-senescent leaves; \( = \) 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 \( LAI_{total} \) (= non-senescent plus senescent leaves). For spring and winter wheat a parameterization to calculate \( LAI_{total} \) and \( LAI_{non-senescent} \) 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 $c_{LAI}$. However, Grünhage and Haenel (1997) made successfully use of a constant value for $c_{LAI}$, 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 $c_{LAI}$ is in the range of 0.3 to 0.6 (Ross, 1981). This includes $c_{LAI} = 0.4$ 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. cabbages, clover) $c_{LAI}$ approaches 1.8 as can be deduced from Monteith (1965).

PLATIN for Excel now incorporates a canopy radiation submodel (see Appendix B), 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 $c_{LAI}$ 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 $c_{LAI}$, but dependent on solar height:

$$k_b = \frac{k_{b,90^\circ}}{\sin \phi}$$  \hspace{1cm} (11)

with $k_{b,90^\circ}$ the 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 $c_{LAI}$ 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 $c_{LAI}$ in the calculation of $\beta$ by $k_{b,\text{max}}$ rather than $k_b$:

$$\beta = e^{-k_{b,\text{max}} \cdot SAI}$$  \hspace{1cm} (12)

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

and

$$k_{b,\text{max}} = \frac{k_{b,90^\circ}}{\sin \phi_{\text{max}}}$$  \hspace{1cm} (13)

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

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

$$1 - \beta^* = \frac{I_{c,\text{sunlit}} + I_{c,\text{shaded}}}{PAR}$$  \hspace{1cm} (14)
with $I_{c, \text{sunlit}}$ irradiance absorbed by the sunlit fraction of non-senescent leaves of the canopy [µmol m$^{-2}$ s$^{-1}$]

$I_{c, \text{shaded}}$ irradiance absorbed by the shaded fraction of non-senescent leaves of the canopy [µmol m$^{-2}$ s$^{-1}$]

PAR photosynthetically active radiation measured above the canopy [µmol m$^{-2}$ 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

$R_{\text{soil, H}_2\text{O}}$ 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, $R_{\text{soil, H}_2\text{O}}$ is parameterized in the following manner:

(a) For a fully wet soil, $R_{\text{soil, H}_2\text{O}}$ equals $R_{\text{soil, H}_2\text{O}, \text{min}} (= 100 \text{ s m}^{-1}).$

(b) For daylight hours (i.e. time intervals with global radiation $S_t \geq 50 \text{ W m}^{-2}$), $R_{\text{soil, H}_2\text{O}}$ is increased by a given fraction of $R_{\text{soil, H}_2\text{O}, \text{min}}$ if there is no precipitation:

$$ (R_{\text{soil, H}_2\text{O}})_n = (R_{\text{soil, H}_2\text{O}})_{n-1} + RX \cdot R_{\text{soil, H}_2\text{O}, \text{min}} $$

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. $R_{\text{soil, H}_2\text{O}}$ is bound by the upper limit of 4000 s m$^{-1}$, the choice of which is based on the results of Daamen and Simmonds (1996).

(c) At night $R_{\text{soil, H}_2\text{O}}$ stays constant at the value calculated for the last late-afternoon daylight hour, i.e. $(R_{\text{soil, H}_2\text{O}})_n = (R_{\text{soil, H}_2\text{O}})_{n-1}. $

(d) At any time interval with precipitation not reaching the ground, $R_{\text{soil, H}_2\text{O}}$ stays constant at the value calculated before, i.e. $(R_{\text{soil, H}_2\text{O}})_n = (R_{\text{soil, H}_2\text{O}})_{n-1}. $

(e) At any time interval with precipitation and/or dew reaching the ground, $R_{\text{soil, H}_2\text{O}}$ is decreased by a fraction $R_Y = a_{\text{soil}} \cdot W_{in}$ of $R_{\text{soil, H}_2\text{O}, \text{min}}$:

$$ (R_{\text{soil, H}_2\text{O}})_n = (R_{\text{soil, H}_2\text{O}})_{n-1} - a_{\text{soil}} \cdot W_{in} \cdot R_{\text{soil, H}_2\text{O}, \text{min}} $$

with $W_{in}$ amount of precipitation and/or dew reaching the ground (water input) [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 $INT_{\text{max}}$ [mm] is assumed to be proportional to total LAI:

$$ INT_{\text{max}} = b_{\text{INT}} \cdot LAI_{\text{total}} $$

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 $Precip$ and dew and depleted by evaporation. Dew formation and depletion of the reservoir is estimated due to potential evapotranspiration rate $E_{pot}$ [mm]
applying the Penman-Monteith approach (see Chapter 2.4) with $R_{c,H_2O} = 0$ s m$^{-1}$ assuming neutral atmospheric stratification. The interception $INT$ [mm] is parameterized according to

$$INT_n = Precip_n + INT_{n-1} - E_{pot,n}$$

with $0 \leq INT_n \leq INT_{\text{max}}$. The precipitation and dew reaching the ground $W_{in}$ is than given by:

$$W_{in,n} = Precip_n + (INT_{n-1} - E_{pot,n}) - INT_{\text{max}}$$

with $W_{in,n} \geq 0$ 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, $R_{\text{soil},H_2O}$ can be set to $R_{\text{soil},H_2O,\text{min}}$.

Note: $R_{c,H_2O}$ is set to zero if the interception reservoir is not empty. Comparisons of modelled evapotranspiration rates with measured fluxes show that setting $R_{c,H_2O}$ 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 exchange (Kerstiens and Lendzian, 1989a, b; Lendzian and Kerstiens, 1991; Kerstiens et al., 1992). According to the aforementioned authors $R_{\text{cut},H_2O}$ on leaf basis is $9 \times 10^4$ s m$^{-1}$ (cf. Table 3, Chapter 3.1). 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 $R_{c,\text{stom, min},H_2O}$, which is representative for a fully developed canopy (without senescent leaves) under optimum conditions for maximal transpiration, upscaling from leaf to canopy level is performed applying $k_{b,\text{max}}$ at maximum solar elevation of the year (summer solstice):

$$R_{\text{canopy}} = R_{\text{leaf, literature}} \cdot \left(1 - e^{-k_{b,\text{max, summer solstice}} \cdot LAI_{\text{leaf, literature}}}ight)$$

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 CO$_2$ is described according to the Jarvis-Stewart approach (Jarvis, 1976; Stewart, 1988):

$$R_{c,\text{stom},H_2O} = \frac{1}{R_{c,\text{stom, min},H_2O} \cdot f_1(S_t) \cdot f_2(t_a) \cdot f_3(d(VPD, SM)) \cdot f_5(\text{time}) \cdot f_6(\text{PHEN}) \cdot f_7(O_3) \cdot f_8(\text{CO}_2)}$$

or

$$R_{c,\text{stom},H_2O} = \frac{1}{R_{c,\text{stom, min},H_2O} \cdot f_1(S_t) \cdot f_2(t_a) \cdot f_3(d(VPD, SM)) \cdot f_5(\text{time}) \cdot f_6(\text{PHEN}) \cdot f_7(O_3) \cdot f_8(\text{CO}_2)}$$

(21)
\[ R_{c,\text{stom},\text{H}_2\text{O}} = \left( \frac{1}{R_{c,\text{stom},\text{min},\text{H}_2\text{O}}} \right)^{-1} \cdot f_1(S_t) \cdot f_2(t_a) \cdot f_3(VPD) \cdot f_4(SM) \cdot f_5(\text{time}) \cdot f_6(\text{PHEN}) \cdot f_7(O_3) \cdot f_8(\text{CO}_2) \]  

(22)

where \( R_{c,\text{stom},\text{min},\text{H}_2\text{O}} \) represents the minimum value of the stomatal resistance for water vapour of the respective ecosystem. Functions \( f_1(S_t) \), \( f_2(t_a) \), \( f_3(VPD) \) and \( f_4(SM) \) account for the effects of solar radiation \( S_t \) [W m\(^{-2}\)], air temperature \( t_a \) [°C], water vapour pressure deficit of the atmosphere \( VPD \) [hPa] and soil moisture \( SM \) [m\(^3\) m\(^{-3}\)] on stomatal aperture \((0 \leq f \leq 1)\). While eq. (22) is based on a multiplicative dependence of stomatal resistance on \( VPD \) and \( SM \) in PLATIN, a combined function \( f_{3d}(VPD, SM) \) is preferred for biological reasons. A combined function \( f_{3d}(VPD, SM) \) reflects the observation that increasing soil moisture deficits strongly influence stomatal closure due to \( VPD \). It is recommended to use measured soil moisture content \( SM \) for \( f_{3d}(VPD, SM) \) 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.

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):

\[
\lambda E = \frac{s_c \cdot (R_{\text{net}} - G) + \rho_{\text{moist air}} \cdot c_{p,\text{moist air}} \cdot \frac{VPD}{V_{\text{ref,2}}} \cdot R_{\text{sh}}(d + z_{0\text{m}}, T_{\text{ref,2}}) + R_{\text{b,heat}}}{s_c + \gamma \cdot \frac{R_{\text{sh}}(d + z_{0\text{m}}, T_{\text{ref,2}}) + R_{\text{b,\text{H}_2\text{O}}} + R_{\text{c,\text{H}_2\text{O}}} + R_{\text{b,heat}}}{R_{\text{sh}}(d + z_{0\text{m}}, T_{\text{ref,2}}) + R_{\text{b,heat}}}} \tag{23}
\]

with \( \gamma \) psychrometric constant (= 0.655 hPa K\(^{-1}\))

and

\[
s_c = \frac{e_{\text{sat}}(T_s) - e_{\text{sat}}(T_{\text{ref}})}{T_s - T_{\text{ref}}} \tag{24}
\]

with \( e_{\text{sat}} \) saturation water vapour pressure of the atmosphere [hPa]

(see Appendix F, eqs. (F1) and (F2))

and \( T_s \) absolute canopy surface temperature at conceptual height \( z = d + z_{0\text{h}} \) [K]

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 estimated by the simple Ohm’s-law formulation. In any case the solution of the energy balance can be achieved only iteratively, because the real unknown in eq. (1) is the surface temperature \( T_s \) which is involved in a non-linear way in the set of equations described above to solve eq. (1).
However, the method used in PLATIN offers the option to get rid of iterations by replacing eq. (24) by the slope of water vapour saturation pressure at reference-height air temperature (see eqs. (F5) or (F6) in Appendix F). This is the way the Penman-Monteith equation is often used as a kind of stand-alone model to estimate evapotranspiration, because it yields results only slightly different from eq. (23), cf. discussion in McArthur (1990). Another advantage of the method used in PLATIN 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 for Excel makes use only of eq. (23).

The surface temperature $T_s$ is related to potential canopy surface temperature $\theta_s$ [K]

$$\theta_s = \theta_{(z_{\text{ref}}, T)} + \frac{H \cdot \left( R_{\text{ah}}(d + z_{\text{ref}}, T) + R_{\text{b,heat}} \right)}{\rho_{\text{moist air}} \cdot c_{p, \text{moist air}}}$$ (25)

by eq. (6). As eq. (25) is part of the set of model equations needed to solve iteratively the energy balance (eq. (1)), a starting value of $\theta_s$ is needed which is assigned the value of the air temperature at reference height minus 0.1 K.

According to the conceptual structure of PLATIN, total latent heat flux density $\lambda E$ can be split up into the contribution by the vegetation $\lambda E_{\text{transpiration}}$ and the one coming from the soil $\lambda E_{\text{evaporation}}$ as given in eqs. (26) and (27):

$$\lambda E_{\text{transpiration}} = \frac{\lambda E}{1 + \beta \cdot \frac{R_{c, \text{cut}, \text{H}_2\text{O}} \cdot R_{c, \text{cut}, \text{H}_2\text{O}}}{R_{\text{soil}, \text{H}_2\text{O}} \cdot (R_{c, \text{cut}, \text{H}_2\text{O}} + R_{c, \text{cut}, \text{H}_2\text{O}})}}$$ (26)

$$\lambda E_{\text{evaporation}} = \lambda E - \lambda E_{\text{transpiration}}$$ (27)

With the latent heat of water vapourisation

$$\lambda \equiv (2.501 - 0.00237 \cdot t_s) \cdot 10^6 \text{ J kg}^{-1}$$ (28)

the actual evapotranspiration $E$ [mm m$^{-2}$] can be computed for the time interval $t_1$ to $t_2$:

$$E = \int_{t_1}^{t_2} \frac{\lambda E}{\lambda} \cdot dt$$ (29)

The water vapour flux density $F_c(\text{H}_2\text{O})$ [g m$^{-2}$ s$^{-1}$] is:

$$F_c(\text{H}_2\text{O}) = \frac{\lambda E}{\lambda \cdot 10^3}$$ (30)

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 CO$_2$/H$_2$O 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 \(z_0\), and bulk canopy resistance for water vapor \(R_{c, H_2O}\) can be found in Appendices I and J.

Figure 4 clearly illustrates that PLATIN is able to simulate measured fluxes adequately.

![Graph showing comparison of measured and modelled sensible and latent heat fluxes for June 2004 during daylight hours.]

**Fig. 4** Comparison of measured and modelled sensible heat (a) and latent heat (b) for June 2004 during daylight hours.

### 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:

\[
F_c(A) = -\frac{\rho_A(z_{\text{ref},A}) - \rho_A(d + z_{0m})}{R_{ah}(d + z_{0m}, z_{\text{ref},A})}
\]

\[
= -\frac{\rho_A(z_{\text{ref},A}) - \rho_A(d + z_{0c})}{R_{ah}(d + z_{0m}, z_{\text{ref},A}) + R_{b,A}}
\]

\[
= -\frac{\rho_A(z_{\text{ref},A}) - \rho_{A,\text{comp}}}{R_{ah}(d + z_{0m}, z_{\text{ref},A}) + R_{b,A} + R_{c,A}}
\]

with \(F_c(A)\) total vertical atmosphere-canopy flux of trace gas \(A\) \([\mu g \cdot m^{-2} \cdot s^{-1}]\), \(\rho_A(z_{\text{ref}})\) measured concentration (potential) of trace gas \(A\) at height \(z = z_{\text{ref},A}\) \([\mu g \cdot m^{-3}]\), \(\rho_A(d + z_{0m})\) concentration of trace gas \(A\) at the conceptual height \(z = d + z_{0m}\) \([\mu g \cdot m^{-3}]\), \(\rho_A(d + z_{0c})\) concentration of trace gas \(A\) at the conceptual height \(z = d + z_{0c} = d + z_{0h}\) \([\mu g \cdot m^{-3}]\), and \(\rho_{A,\text{comp}}\) canopy compensation concentration of trace gas \(A\) \([\mu g \cdot m^{-3}]\).

The respective resistance scheme is shown in Figure 5. It is based on the assumption that chemical sinks and/or sources between reference height and canopy surface can be neglected. Under this assumption it is allowed to adopt the turbulent atmospheric resistance \(R_{ah}\) calculated after eq. (2), and the quasi-laminar layer resistance \(R_{b,A}\) calculated after eq. (7) taking into account the respective \((Sc/A/Pr)^{2/3}\) (cf. Table 2; see Chapter 3.1).
The so-called canopy compensation concentration $\rho_{\text{A, comp}}$ is an effective concentration. It is defined by eq. (33) and represents, along with the bulk canopy resistance $R_{c,A}$ (see eq. (34)) a more complex network as depicted in Figure 6.

If a plant/soil system can be considered a perfect sink as is usually assumed for the exchange of O$_3$, SO$_2$, NO$_2$ and HNO$_3$, one can set $\rho_{\text{A, int1}} = \rho_{\text{A, int2}} = \rho_{\text{A, ext}} = \rho_{\text{A, soil}} = 0$ which leads to $\rho_{\text{A, comp}} = 0$. For NH$_3$ (and HONO) the scheme must allow for bi-directional fluxes which means that, in general, all the aforementioned concentrations are different from zero and have to be parameterized (for more details see Chapter 3.4).

The bulk canopy resistance $R_{c,A}$ represents a combination of resistances (Figure 6) characterizing the fluxes through the leaf stomata ($R_{c,\text{stom},A}$), into or out the mesophyll tissue ($R_{c,\text{mes},A}$), through the cuticle of the leaves ($R_{c,\text{cut},A}$), to or from external plant surfaces ($R_{c,\text{ext},A}$), and down to or up from the soil surface ($R_{\text{soil},A}$).
Parameters $\beta^*$ and $\beta$ account for the phenological stage of the canopy on vertical exchange of gaseous species between the plant/soil system and the atmosphere near the ground (see Chapter 2.3).

Nitric acid is a special case, because bulk canopy resistance $R_{c, HNO3}$ is effectively zero in normal conditions. For numerical reasons a minimum value of $R_{c, HNO3}$ of 1 s m$^{-1}$ is enforced. Considering a low-temperature resistance $R_{\text{low}}$ at sub-zero air temperatures $t_a$ [°C] according to Wesely (1989)

$$R_{\text{low}} = 1000 \cdot \exp(-t_a - 4) \text{ s m}^{-1}$$

results in:

$$R_{c, HNO3} = \max(1 \text{ s m}^{-1}, R_{\text{low}})$$

Note: According to Wesely (1989) $R_{\text{low}}$ is added to $R_{c, \text{ext}}$ and $R_{\text{soil}}$ for all gases mentioned below with exception of sulphur dioxide. For the resistance of snow-covered surfaces we refer to the approximations given by Erisman et al. (1994).

### 3.1 Bulk stomatal, mesophyll and cuticular resistances for trace gases

The bulk stomatal resistance for a gaseous species A is related to that of water vapour by the ratio the respective molecular diffusivities $D$ (cf. Table 2):

$$R_{c, \text{stom}, A} = R_{c, \text{stom}, H2O} \cdot \frac{D_{H2O}}{D_A}$$

According to Wesely (1989), the canopy mesophyll resistance for any trace gas is given by:

$$R_{c, \text{mes}, A} = \left(\frac{H_A^*}{3000} + 100 \cdot f_{0,A}\right)^{-1}$$

with $H_A^*$ effective Henry’s Law value (cf. Table 3)

$f_{0,A}$ chemical reactivity factor (cf. Table 3)

This formulation allows for two parallel pathways to the sink: one dominant for a primarily water soluble species like SO$_2$, and the other one for a primarily chemically reactive species like O$_3$. The value 3000 was derived from considerations of carbon dioxide uptake by vegetation (for details see Wesely, 1989). There is experimental evidence that $R_{c, \text{mes}, \text{NO2}}$ calculated according to eq. (38) is much too small for conifer species as reviewed by Ganzeveld and Lelieveld (1995). They approximated the mesophyll resistance for NO$_2$ by:

$$R_{c, \text{mes}, \text{NO2}} = 0.5 \cdot R_{c, \text{stom}, \text{NO2}}$$

As mentioned above, the cuticular resistances for trace gases published in the literature are given on leaf basis (cf. Table 3) and must therefore be upscaled to canopy level by eq. (20).

### 3.2 Bulk external plant surface resistance for trace gases

Permeation through cuticles excluded, the amount of deposition on external plant surfaces depends on, for example, the wetness of the surface, the pH of the wetted surface or the surface temperature (e.g. Wesely, 1989; Erisman et al., 1994; Fowler et al., 2001; Zhang et al., 2002; Massman, 2004; Altimir et al., 2004, 2006).
Table 2
Diffusivities in air (Benner et al., 1988; Grünhage and Haenel, 1997; Massman, 1998) and related values

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<tr>
<td></td>
<td>$D_A [10^{-6} \text{ m}^2 \cdot \text{s}^{-1}]$</td>
<td>$T_0 = 273.15 \text{ K}$; reference air pressure $p_0 = 1013.25 \text{ hPa}$</td>
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<tr>
<td>H$_2$O (water vapour)</td>
<td>21.9 ± 0.4</td>
<td>21.78</td>
<td>21.9</td>
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<tr>
<td>CO$_2$ (carbon dioxide)</td>
<td>13.7 ± 0.4</td>
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<td>SO$_2$ (sulphur dioxide)</td>
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<td>O$_3$ (ozone)</td>
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<td>NH$_3$ (ammonia)</td>
<td>20.3 ± 0.7</td>
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<td>HNO$_2$ (nitrous acid)</td>
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</tbody>
</table>

$D_A(T, p) = D_A \cdot [(T \cdot T_0^{-1})^{1.81} \cdot (p_0 \cdot p^{-1})]$

<table>
<thead>
<tr>
<th></th>
<th>$S_{CA} (= \nu \cdot D_A^{-1})$</th>
<th>$(S_{CA} \cdot Pr^{-1})^{2/3}$</th>
<th>$D_{H2O} \cdot D_A^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>H$_2$O (water vapour)</td>
<td>0.61</td>
<td>0.90</td>
<td>1</td>
</tr>
<tr>
<td>CO$_2$ (carbon dioxide)</td>
<td>0.97</td>
<td>1.23</td>
<td>1.60</td>
</tr>
<tr>
<td>SO$_2$ (sulphur dioxide)</td>
<td>1.24</td>
<td>1.45</td>
<td>2.05</td>
</tr>
<tr>
<td>O$_3$ (ozone)</td>
<td>0.92</td>
<td>1.19</td>
<td>1.51</td>
</tr>
<tr>
<td>NH$_3$ (ammonia)</td>
<td>0.67</td>
<td>0.96</td>
<td>1.10</td>
</tr>
<tr>
<td>NO (nitric oxide)</td>
<td>0.74</td>
<td>1.03</td>
<td>1.22</td>
</tr>
<tr>
<td>NO$_2$ (nitrogen dioxide)</td>
<td>0.96</td>
<td>1.22</td>
<td>1.58</td>
</tr>
<tr>
<td>HNO$_3$ (nitric acid)</td>
<td>1.46</td>
<td>1.62</td>
<td>2.41</td>
</tr>
<tr>
<td>HNO$_2$ (nitrous acid)</td>
<td>1.53</td>
<td>1.67</td>
<td>2.52</td>
</tr>
</tbody>
</table>

kinematic viscosity of dry air $\nu = 13.3 \cdot 10^{-6} \text{ m}^2 \cdot \text{s}^{-1}$
Prandtl number $Pr = 0.71$

In general, three qualitative "surface wetness states" can be distinguished: wet, partially wet, dry. Based on plausible reasoning or on data from the literature, PLATIN assumes external plant surfaces to be wet if at least one of the following conditions applies:

- precipitation $Precip > 0$ mm
- relative interception reservoir charge $INT_n \cdot INT_{max}^{-1} \geq 0.2$
- relative humidity $rH > 90\%$ (according to Erisman et al., 1993)

The Unified EMEP model (Simpson et al., 2003) does not explicitly use a wet state, but extends the state of partial wetness gradually up to 100% $rH$. However, this does not meet e.g. observations of plant surfaces still wet after precipitation events while air humidity has already fallen below 100%. This discrepancy resolves when taking into account that the Unified EMEP model is to calculate surface resistances on a horizontal scale much greater than that of local SVAT models like PLATIN. On such
great horizontal scales completely wet surfaces are not likely to exist. Nevertheless, the concept of a
gradual approach from dry to wet and vice versa has often been reported in literature also for local con-
siderations, and it will explicitly be introduced in PLATIN where it has not yet been defined before (see
below in context with eqs. (46) to (49)). When not defined otherwise, PLATIN also adopts the upper
humidity threshold for the surface state "dry" as given by EMEP (Simpson et al., 2003): $rH = 85\%$ for
forests and $rH = 75\%$ for other canopies. These provisional thresholds are based upon results of wet-
ness measurements presented by Klemm et al. (2002).

Table 3

<table>
<thead>
<tr>
<th>substance</th>
<th>$H_A^*$ [M atm$^{-1}$]</th>
<th>$f_{0,A}$</th>
<th>$R_{c,\text{mes},A}$ [s$^{-1}$ m$^{-1}$]</th>
<th>$R_{\text{leaf,cut},A}$ [s$^{-1}$ m$^{-1}$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>H$_2$O (water vapour)</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>9.10$^4$</td>
</tr>
<tr>
<td>CO$_2$ (carbon dioxide)</td>
<td>4.4</td>
<td>0</td>
<td>682</td>
<td>1.10$^7$</td>
</tr>
<tr>
<td>SO$_2$ (sulphur dioxide)</td>
<td>1.10$^5$</td>
<td>0</td>
<td>0.03</td>
<td>2.10$^5$</td>
</tr>
<tr>
<td>O$_3$ (ozone)</td>
<td>0.01</td>
<td>1</td>
<td>0.01</td>
<td>3.10$^7$</td>
</tr>
<tr>
<td>NH$_3$ (ammonia)</td>
<td>2.10$^4$</td>
<td>0</td>
<td>0.15</td>
<td>3.10$^6$</td>
</tr>
<tr>
<td>NO (nitric oxide)</td>
<td>2.10$^3$</td>
<td>0</td>
<td>1.5.10$^5$</td>
<td>8.10$^7$</td>
</tr>
<tr>
<td>NO$_2$ (nitrogen dioxide)</td>
<td>0.01</td>
<td>0.1</td>
<td>0.10</td>
<td>2.10$^6$</td>
</tr>
<tr>
<td>HNO$_3$ (nitric acid)</td>
<td>1.10$^{14}$</td>
<td>0</td>
<td>3.10$^{-11}$</td>
<td>1.10$^5$</td>
</tr>
<tr>
<td>HNO$_2$ (nitrous acid)</td>
<td>1.10$^5$</td>
<td>0.1</td>
<td>0.02</td>
<td>2.10$^6$</td>
</tr>
</tbody>
</table>

$^1$ HNO$_2$ is treated as NO$_2$

The external plant surfaces resistance for ammonia $R_{c,\text{ext},\text{NH}_3}$ depends on relative air humidity $rH$
surface temperature, and the molar "acidity ratio" of SO$_2$ and NH$_3$ concentration. An approximation of
$R_{c,\text{ext},\text{NH}_3}$ taking into account all these three factors can be found in Simpson et al. (2003). While surface
temperature could be provided by PLATIN, SO$_2$ concentrations may not always be available. Thus
PLATIN keeps to the less data demanding approach given by Sutton and Fowler (1993) which is based
on $rH$ only:

$$
R_{c,\text{ext},\text{NH}_3} = 2 \cdot \exp\left(\frac{100 - rH}{12}\right)
$$

(40)

Evidently, there is no need to distinguish between different surface wetness states.

The external plant surfaces resistance for sulphur dioxide $R_{c,\text{ext},\text{SO}_2}$ depends on rain events, relative
humidity $rH$ and air temperature $t_a$ and is parameterized according to Erisman et al. (1994):

- if external plant surfaces are wet: $R_{c,\text{ext,SO}_2} = 1$ s$^{-1}$ m$^{-1}$
- if $t_a > -1$ °C and $rH > 81.3\%$: $R_{c,\text{ext,SO}_2} = 0.58 \cdot 10^{12} \cdot \exp(-0.278 \cdot rH)$ s$^{-1}$ m$^{-1}$
- if $t_a > -1$ °C and $rH \leq 81.3\%$: $R_{c,\text{ext,SO}_2} = 25000 \cdot \exp(-0.0693 \cdot rH)$ s$^{-1}$ m$^{-1}$
- if $-5$ °C < $t_a$ ≤ $-1$ °C: $R_{c,\text{ext,SO}_2} = 200$ s$^{-1}$ m$^{-1}$
- if $t_a$ ≤ $-5$ °C: $R_{c,\text{ext,SO}_2} = 500$ s$^{-1}$ m$^{-1}$
Similar to $R_{c,\text{ext, NH}_3}$, there is no need to distinguish between different surface wetness states.

Note: This parameterization does not take into account the control of $R_{c,\text{ext, SO}_2}$ by the NH$_3$ levels in the air (cf. Simpson et al., 2003). Here again, PLATIN makes use of the less data demanding approach.

Often, for ozone as well as for all other trace gas species, constant external plant surface resistances are defined separately for wet and dry surface conditions. For dry surface conditions $R_{c,\text{ext, dry, O}_3}$ is estimated by

$$R_{c,\text{ext, dry, O}_3} = 2000 \cdot \left( 1 - e^{-k_{b,\text{max, summer solstice}}} \right) \text{s} \cdot \text{m}^{-1} \quad (46)$$

For $k_{b,\text{max, summer solstice}}$ see Chapter 2.3, context of eq. (20). The leaf-level resistance $R_{\text{leaf, ext, dry, O}_3}$ of 2000 s m$^{-1}$ is adopted from Gao et al. (1993). For other gases than NH$_3$, SO$_2$ and O$_3$ $R_{c,\text{ext, dry, A}}$ is computed according to Wesely (1989):

$$R_{c,\text{ext, dry, A}} = \frac{R_{c,\text{ext, dry, O}_3}}{(10^{-5} \cdot H_A^*) + f_{0,A}} \quad (47)$$

As described in the PLATIN version published by Grünhage and Haenel (1997), the "wet" resistance for O$_3$ is calculated by

$$R_{c,\text{ext, wet, O}_3} = \left( \frac{1}{1000} + \frac{1}{3 \cdot R_{c,\text{ext, dry, O}_3}} \right)^{-1} \quad (48)$$

while for all gases other than NH$_3$, SO$_2$ and O$_3$ the resistance is estimated from:

$$R_{c,\text{ext, wet, A}} = \left( \frac{1}{3 \cdot R_{c,\text{ext, dry, A}}} + \frac{H_A^*}{10^7} + \frac{f_{0,A}}{3 \cdot R_{c,\text{ext, dry, O}_3}} \right)^{-1} \quad (49)$$

While not reported explicitly in the literature, a gradual shifting between "dry" and "wet" resistances as given by eqs. (46) to (49) seems plausible and is incorporated in PLATIN according to the scheme described in Simpson et al. (2003). It is based on the definition of a so-called humidity factor $F_{h\text{H}}$

$$F_{h\text{H}} = \begin{cases} 0 & \text{for } rH \leq rH_{\text{dry}} \\ \frac{rH - rH_{\text{dry}}}{rH_{\text{wet}} - rH_{\text{dry}}} & \text{for } rH_{\text{dry}} < rH \leq rH_{\text{wet}} \\ 1 & \text{for } rH_{\text{wet}} < rH \end{cases} \quad (50)$$

where, as mentioned above, $rH_{\text{wet}} = 90 \%$ while $rH_{\text{dry}} = 85 \%$ for forests and $rH_{\text{dry}} = 75 \%$ for other canopies. With $F_{h\text{H}}$ the resulting external-surface resistance for species A (including O$_3$, but not NH$_3$ and SO$_2$) reads:

$$\frac{1}{R_{c,\text{ext, A}}} = F_{h\text{H}} \cdot \frac{1}{R_{c,\text{ext, wet, A}}} + (1 - F_{h\text{H}}) \cdot \frac{1}{R_{c,\text{ext, dry, A}}} \quad (51)$$

Note: For O$_3$, relations (46) and (48) represent only a first estimate. As illustrated by Fowler et al. (2001), the external plant surface resistance for O$_3$ decreases as global radiation $S_t$ and therefore surface temperature increase:

$$R_{c,\text{non-stomatal, dry, O}_3} = -129.9 \cdot \ln(S_t) + 989.1 \quad (52)$$
This reduction of $R_{c,\text{non-stomatal, dry, } O_3}$ observed over moorland vegetation at solar radiation fluxes above 100 W m$^{-2}$ in southern Scotland, is regarded as evidence of thermal decomposition of O$_3$ at the leaf surfaces. In the context of O$_3$ exchange modelling, future research has to focus on the derivation of parameterizations of the non-stomatal O$_3$ resistance.

### 3.3 Soil resistance for trace gases

The deposition rates of trace gases on soil surfaces depend on their water solubility as well as on their reactivity. The soil surface resistance for any gaseous species $R_{\text{soil, dry, } A}$ is computed as

$$R_{\text{soil, dry, } A} = \left( \frac{H_A \ast}{10^5 \cdot R_{\text{soil, dry, SO}_2}} + \frac{f_{0, A}}{R_{\text{soil, dry, O}_3}} \right)^{-1}$$

according to Wesely (1989), where $R_{\text{soil, dry, SO}_2}$ is set to 500 s m$^{-1}$ and $R_{\text{soil, dry, O}_3}$ at 200 s m$^{-1}$ (Gao et al., 1993). Soil surface resistance under wet conditions (i.e. when $R_{\text{soil}} = R_{\text{soil, min}}$) depends on whether ozone or another species is concerned. It is calculated the same way as for external plant surfaces, i.e. according to eqs. (48) and (49). For the resistance of snow covered surfaces the approximations described in Erisman et al. (1994) can be used. In agreement with the observations for $R_{\text{soil, } O_3}$ summarized in Figure 3 in Massman (2004), the parameterization given above results in a soil resistance which is higher for wet conditions than for dry conditions.

### 3.4 Special treatment of NH$_3$

The estimation of NH$_3$ fluxes over vegetated land deserves special attention, because, in contrast to e.g. O$_3$ and SO$_2$, NH$_3$ fluxes over vegetated land are bi-directional, i.e. the net flux can combine simultaneously occurring deposition and emission. In addition, different parts of the plants and also the soil may serve as sources or sinks. Nemitz et al. (2000) found the siliques of oilseed rape to emit NH$_3$ while the leaves definitely act as a sink and the decomposing leaf litter at the ground again was identified to be a NH$_3$ source.

Another issue is the effect of fertilization. With mineral fertilization, arable fields are likely to be net sinks for reactive nitrogen, while application of slurry turns them into net sources (U. Dämmgen, von Thunen Institute, Braunschweig, personal communication). NH$_3$ may be released directly from the fertilizer due to thermal decomposition: $\text{NH}_4^+ + \text{HCO}_3^- \rightarrow \text{NH}_3 + \text{H}_2\text{O} + \text{CO}_2$. Urea decomposes at the surface partly due to enzymatic clearage. N from fertilizer, which passes through root and stem into the leaves, increases the concentration of ammonium in the apoplast (see below), leading to NH$_3$ emission from the leaves.

Another point is that emission and deposition rates depend to some degree on the results of preceding physical and chemical processes.

These issues have to be considered by modelling of NH$_3$ fluxes between the atmosphere and vegetated surfaces. In order to account for a vertical distribution of sources and sinks within the canopy and on the ground, multi-layer resistance models are likely to be the appropriate choice as pointed out by Nemitz et al. (2000). Dynamic effects can be included, in principle, by introducing capacitors into the model’s resistance network as was done by Sutton et al. (1998). More difficult is the treatment of slurry and mineral fertilizer application which, at the time being, no mechanistic modelling approach is available for. Therefore it is recommended (U. Dämmgen, von Thunen Institute, Braunschweig, personal communication) for the first seven days after slurry or mineral fertilizer application not to use a SVAT model like PLATIN but to estimate the cumulative NH$_3$ emissions from emission factors according to EMEP/CORINAIR (2007).
At the present state of development, PLATIN is a steady-state single-layer model which cannot cope with the multi-layer and dynamic requirements formulated above. However, PLATIN could be extended in future by introducing new branches into the resistance network depicted in Figure 6 to account for additional NH$_3$ sinks or sources like the oilseed rape siliques or a litter layer beneath the canopy. Also, the introduction of capacitors (the charge or discharge of which requires knowledge about the respective conditions during past times) would not impose any problems. This can be seen from the fact that PLATIN can already be driven with a rain interception reservoir, cf. Chapter 2.3, which in effect is nothing else than a capacitor (with respect to evaporation fluxes). As soon as a mathematical description is available for NH$_3$ flux resulting from fertilizing, it can be integrated in the PLATIN resistance network if the node can be identified where this flux is entering the network.

Nevertheless, already the present state of PLATIN represents an important tool to approach an overall nitrogen balance of vegetated surfaces. The base to estimate NH$_3$ atmosphere-canopy exchange is eq. (32) which requires knowledge of NH$_3$ concentration $\rho_{\text{NH}_3}(d+z_{0c})$ at the conceptual position $z = d + z_{0c}$ of the big leaf surface for gaseous fluxes. The concentration at this level has been called canopy "compensation point" by Sutton and Fowler (1993), but is rather to be considered the "net potential for NH$_3$ emission from the canopy" (cf. Sutton et al., 1998). According to Figure 5 the following equation for $\rho_{\text{NH}_3}(d+z_{0c})$ results:

$$
\rho_{\text{NH}_3}(d+z_{0c}) = \frac{\rho_{\text{NH}_3}^{3}(z_{\text{ref,NH}_3})}{R_{s}(d+z_{0c}, z_{\text{ref,NH}_3}) + R_{b,NH_3}} + (1-\beta) \cdot \frac{\rho_{\text{NH}_3}^{3}(z_{\text{ref,NH}_3})}{R_{s}(z_{\text{ref,NH}_3}) + R_{b,NH_3}} (1-\beta) \cdot \frac{\rho_{\text{NH}_3}^{3}(z_{\text{ref,NH}_3})}{R_{s}(z_{\text{ref,NH}_3}) + R_{b,NH_3}} + \beta 
$$

(54)

with $\rho_{\text{NH}_3}$, stom ammonia concentration in the sub-stomatal cavities [µg m$^{-3}$]

Because exchange of gaseous species through the plant's cuticle is of minor importance (cf. Chapter 2.3, Table 3) in the present PLATIN version, penetration of NH$_3$ through the plant's cuticle is neglected by completely dropping the respective flow branch from the resistance scheme. Additionally, emissions from external plant surfaces and from the soil beneath the canopy are disregarded by assigning zero value to $\rho_{\text{NH}_3}$, ext and $\rho_{\text{NH}_3}$, soil. With these simplifications eq. (54) modifies to:

$$
\rho_{\text{NH}_3}(d+z_{0c}) = \frac{\rho_{\text{NH}_3}^{3}(z_{\text{ref,NH}_3})}{R_{s}(d+z_{0c}, z_{\text{ref,NH}_3}) + R_{b,NH_3}} + (1-\beta) \cdot \frac{\rho_{\text{NH}_3}^{3}(z_{\text{ref,NH}_3})}{R_{s}(z_{\text{ref,NH}_3}) + R_{b,NH_3}} (1-\beta) \cdot \frac{\rho_{\text{NH}_3}^{3}(z_{\text{ref,NH}_3})}{R_{s}(z_{\text{ref,NH}_3}) + R_{b,NH_3}} + \beta 
$$

(55)

Note: $R_{s,ext}$, NH$_3$ and $R_{s,soil}$, NH$_3$ are parameterized as described in Chapters 3.2 and 3.3.

In order to apply eq. (55) it is necessary to calculate $\rho_{\text{NH}_3}$, stom. According to Sutton et al. (1994) the NH$_3$ gas phase concentration in the sub-stomatal cavities of the canopy $\rho_{\text{NH}_3}$, stom is related to the plant's nitrogen status and pH in the apoplast by the Henry and dissociation equilibria for NH$_3$ and NH$_4^+$ (cf. Flechard and Fowler, 1998):

---

1 Schaaf and Meesenburg (2005) who also used PLATIN present an equation which is quite similar to (55) but physically inadequate, because it neglects the important influence of canopy architecture represented by the factors $(1-\beta)$, $(1-\beta)$, and $\beta$. Schaaf and Meesenburg (2005) name $\rho_{\text{NH}_3}(d+z_{0c})$ the canopy compensation point, which may have been the reason for, or the outcome of, their Figures 5.1 and 5.2 which are contradictory regarding just the position of this compensation point within the resistance network. (Their Figure 5.2 identifies $\rho_{\text{NH}_3}$, ext with $\rho_{\text{ext}}$ in our Figure 5). In addition, the equation for $R_{s,ext}$ given in Schaaf and Meesenburg (2005, chapter 5.1.5) does not agree with their Figure 5.2 which does not display the unique potential at the model surface implicitly invoked by the respective equation. Schaaf and Meesenburg (2005) write they used an extended PLATIN version. It must be noted that all model features they describe had previously been introduced in PLATIN or related model versions by the authors of PLATIN, Grünhage and Haenel. Nevertheless, the latter are not responsible for possible errors in description and application of PLATIN by Schaaf and Meesenburg (2005).
\[ \rho_{\text{NH}_3,\text{stom}} = a_{\text{NH}_3} \cdot \frac{15.43 - \frac{4507.08}{T_s}}{T_s} \cdot \Gamma \]  

(56)

where \( \Gamma \) is the vegetation type-specific ratio of ammonium to protons in the apoplast [mol·mol\(^{-1}\)], \( T_s \) is the surface absolute temperature of the canopy [K], and \( a_{\text{NH}_3} \) is the dimension adaptation factor (= 1 µg·m\(^{-3}\)·K).

The vegetation type-specific ratio of ammonium to protons in the apoplast \( \Gamma \) represents a modelling concept rather than a measurable entity. However, Nemitz et al. (2001) determined \( \Gamma \) experimentally. As described by Flechard et al. (1999), \( \Gamma \) values published for growing arable crops are mostly in the range of 250 – 4000 mol·mol\(^{-1}\). Estimates for moorland vegetation are substantially lower: 80 – 3000 mol·mol\(^{-1}\). The authors describe a value of 180 mol·mol\(^{-1}\) for a moorland in southern Scotland. Sortebberg and Hov (1996) quote \( \Gamma \approx 950 \) mol·mol\(^{-1}\) for crops and grassland and \( \Gamma \approx 320 \) mol·mol\(^{-1}\) for other vegetated surfaces. For a semi-natural ungrazed short grassland \( \Gamma \) approximates 1000 mol·mol\(^{-1}\) (Spindler et al., 2001). Nemitz et al. (2001) indicate a value of 2000 mol·mol\(^{-1}\) for oilseed rape canopy. For the extremely eutrophied Speulder forest in the Netherlands a value of 8500 mol·mol\(^{-1}\) is adequate (according to private communication of J.W. Erisman, ECN Petten, and E. Nemitz, CEH Edinburgh, as quoted in Schaaf and Meesenburg, 2005). Accordingly the following default values of \( \Gamma \) are recommended for use in PLATIN:

- extremely eutrophicated ecosystems: \( \Gamma = 8500 \) mol·mol\(^{-1}\)
- intensively managed grassland and arable crops: \( \Gamma = 2000 \) mol·mol\(^{-1}\)
- extensively managed grassland and forest: \( \Gamma = 1000 \) mol·mol\(^{-1}\)
- N-limited ecosystems: \( \Gamma = 300 \) mol·mol\(^{-1}\)

### 3.5 Partitioning of total atmosphere-canopy flux

In case of deposition, the resistance network (Figures 5, 6) allows to partition the total atmosphere-canopy flux \( F_{c,\text{total}}(A) \) into (1) fluxes absorbed by the plant through the stomata and the cuticle \( F_{c,\text{stom}}(A) + F_{c,\text{cut}}(A) \), and (2) fluxes down to external plant surfaces \( F_{c,\text{ext}}(A) \) and the soil beneath the canopy \( F_{c,\text{soil}}(A) \). Studies show that penetration of gases through the cuticle \( F_{c,\text{cut}}(A) \) can be neglected in comparison to stomatal uptake \( F_{c,\text{stom}}(A) \) (cf. Chapter 2.3).

Combining \( F_{c,\text{ext}} \) and \( F_{c,\text{soil}} \) to \( F_{c,\text{non-stomatal}} \) and neglecting cuticular fluxes (i.e. approximating \( F_{c,\text{stom}} + F_{c,\text{cut}} \equiv F_{c,\text{stom}} \)) one obtains:

\[ F_{c,\text{total}}(A) = F_{c,\text{stom}}(A) + F_{c,\text{ext}}(A) + F_{c,\text{soil}}(A) \equiv F_{c,\text{stom}}(A) + F_{c,\text{non-stomatal}}(A) \]  

(57)

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

\[ PAD(A) = \int_{t_1}^{t_2} F_{c,\text{stom}}(A) \cdot dt = \int_{t_1}^{t_2} F_{c,\text{stom}}(A) \cdot dt \]  

(58)

For \( O_3 \) the integral of \( F_{c,\text{stom}} + F_{c,\text{cut}} \equiv F_{c,\text{stom}} \) over time \( t \) is called accumulated stomatal flux of ozone, \( AF_{\text{st}} \) (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 \( \rho_{O_3}(z_{ref}) \) 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
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L Grünhage and H-D Haenel

disposes only of one single canopy-representative laminar boundary layer the surface of which is located at \( d + z_{0m} \). The concentration at this height is calculated from:

\[
\rho_{\text{O}_3}(d + z_{0m}) = \rho_{\text{O}_3}(z_{\text{ref},0m}) + \left[ F_{c,\text{total}}(\text{O}_3) \cdot R_{\text{sh}}(d + z_{0m}, z_{\text{ref},0m}) \right] 
\]  

(59)

This is contrasted by the UNECE (2004, 2007) approach that the concentration at the upper surface of the laminar boundary layer of the sunlit upper canopy leaves be represented by the \( \text{O}_3 \) concentration at the top of the canopy \( z = h \). Within the M-O framework, this concentration is given by:

\[
\rho_{\text{O}_3}(h) = \rho_{\text{O}_3}(z_{\text{ref},0m}) + \left[ F_{c,\text{total}}(\text{O}_3) \cdot R_{\text{sh}}(h, z_{\text{ref},0m}) \right] 
\]  

(60)

Besides the fact that the effective upper surface of all the laminar boundary layers existing within the canopy is probably represented better by \( d + z_{0m} \) than by \( h \), application of eq. (60) 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. (61; see below) with \( \text{O}_3 \) concentrations from eq. (60) 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. (60) 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 \( z_{0m} \) by \( h = (d + z_{0m}) \cdot 0.8^{-1} \) according to Brutsaert (1984). In order to reveal the influence of the differing heights in eqs. (59) and (60), the stomatal uptake as calculated with \( \text{O}_3 \) concentrations from eq. (60) has then been normalised by the one obtained with \( \text{O}_3 \) concentrations from eq. (59). As stomatal uptake is proportional to \( \text{O}_3 \) concentration, this ratio of stomatal uptake turns out to be identical to the ratio of the respective \( \text{O}_3 \) concentrations. This concentration ratio is displayed in Figure 7.

![Figure 7](image)

**Fig. 7**
Box-and-whisker plot of \( \text{O}_3 \) concentrations at \( z = (d + z_{0m}) \cdot 1.1 \), \( z = h \) and \( z = h + 0.05 \) m in percent of \( \text{O}_3 \) concentrations at \( z = d + z_{0m} \) during daylight hours in June 2004

Estimating stomatal uptake from \( h \) rather than \( d + z_{0m} \) generally leads to overestimation of stomatal uptake which may not be negligible. This overestimation clearly depends on how much \( h \) differs from
$d + z_{0m}$, which is demonstrated by varying $h$ by plus 5 and $d + z_{0m}$ by a factor of 1.1. (Diminishing $h$, on the other hand, would mean to approach $d + z_{0m}$ and therefore reduce overestimation. However, $h$ should never reach $d + z_{0m}$, 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 $O_3$, $F_{c, stom}(O_3)$ is the toxicologically effective share of $F_{c, total}(O_3)$, flux-effect relationships should be based on that component which is given by:

$$F_{c, stom}(O_3) = \frac{\rho_{O3}(z_{ref,O3})}{R_{ah} + R_{b,O3} + \frac{R_{c, stom+mes,O3}}{1 - \beta^*} + \left[ R_{ah} + R_{b,O3} \right] \frac{R_{c, stom+mes,O3}}{1 - \beta^*} \left[ 1 - \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{\beta}{R_{soil,O3}} \right]}{\left( \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{\beta}{R_{soil,O3}} \right)}$$ (61)

with

$$R_{ah} = R_{ah}(d + z_{0m}, z_{ref,O3})$$

$$R_{c, stom+mes,O3} = R_{c, stom,O3} + R_{c, mes,O3}$$ (62)

The flux of $O_3$ through the cuticle $F_{c, cut}(O_3)$, which normally can be neglected, can be derived from eq. (63),

$$F_{c, cut}(O_3) = \frac{\rho_{O3}(z_{ref,O3})}{R_{ah} + R_{b,O3} + \frac{R_{c, cut,O3}}{1 - \beta^*} + \left[ R_{ah} + R_{b,O3} \right] \frac{R_{c, cut,O3}}{1 - \beta^*} \left[ 1 - \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{\beta}{R_{soil,O3}} \right]}{\left( \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{\beta}{R_{soil,O3}} \right)}$$ (63)

the deposition of $O_3$ on external plant surfaces $F_{c, ext}(O_3)$ and the soil beneath the canopy $F_{c, soil}(O_3)$, which are combined to non-stomatal deposition $F_{c, non-stomatal}(O_3)$, from eqs. (64) and (65):

$$F_{c, ext}(O_3) = \frac{\rho_{O3}(z_{ref,O3})}{R_{ah} + R_{b,O3} + \frac{R_{c, ext,O3}}{1 - \beta^*} + \left[ R_{ah} + R_{b,O3} \right] \frac{R_{c, ext,O3}}{1 - \beta^*} \left[ 1 - \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{\beta}{R_{soil,O3}} \right]}{\left( \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{\beta}{R_{soil,O3}} \right)}$$ (64)

$$F_{c, soil}(O_3) = \frac{\rho_{O3}(z_{ref,O3})}{R_{ah} + R_{b,O3} + \frac{R_{c, soil,O3}}{\beta} + \left[ R_{ah} + R_{b,O3} \right] \frac{R_{c, soil,O3}}{\beta} \left[ 1 - \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{1}{R_{soil,O3}} \right]}{\left( \frac{1}{R_{c, cut,O3}} + \frac{1}{R_{c, ext,O3}} + \frac{1}{R_{soil,O3}} \right)}$$ (65)

$F_{c, stom}(O_3)$ can further be subdivided into the flux entering the compartment of sunlit leaves, $F_{c, stom, sunlit}(O_3)$, and the flux taken up by the shaded-leaves compartment, $F_{c, stom, shaded}(O_3)$:

$$F_{c, stom}(O_3) = F_{c, stom, sunlit}(O_3) + F_{c, stom, shaded}(O_3)$$ (66)
The estimation of \( F_{c, \text{stom, sunlit}(O_3)} \) and \( F_{c, \text{stom, shaded}(O_3)} \) is controlled by the \( O_3 \) bulk resistances assigned to the two compartments and their relation to the bulk canopy resistance \( R^{*}_{c, \text{stom, } O_3} \). The latter is proportional to the \( H_2O \) bulk canopy resistance \( R^{*}_{c, \text{stom, } H_2O} \), cf. eq. (37):

\[
R^{*}_{c, \text{stom, } O_3} = R^{*}_{c, \text{stom, } H_2O} \cdot \frac{D_{H_2O}}{D_{O_3}} \quad (67)
\]

\( R^{*}_{c, \text{stom, } H_2O} \) is obtained from

\[
R^{*}_{c, \text{stom, } H_2O} = \frac{R_{c, \text{stom, H}_2O}}{1 - \beta^{*}} \quad (68)
\]

where \( R_{c, \text{stom, H}_2O} \) is given by eqs. (21) or (22) as the bulk resistance one obtains under neglection of the vertical distribution of within-canopy radiation extinction. In eq. (68), this extinction is accounted for by the correction term \( 1 - \beta^{*} \) (eq. (14)).

Similarly the resistances for the two compartments "sunlit leaves" and "shaded leaves", are defined by

\[
R^{*}_{c, \text{stom, sunlit, } O_3} = \frac{R_{c, \text{stom, } H_2O}}{1 - \beta^{*}_{\text{sunlit}}} \cdot \frac{D_{H_2O}}{D_{O_3}} \quad (69)
\]

and

\[
R^{*}_{c, \text{stom, shaded, } O_3} = \frac{R_{c, \text{stom, } H_2O}}{1 - \beta^{*}_{\text{shaded}}} \cdot \frac{D_{H_2O}}{D_{O_3}} \quad (70)
\]

where \( 1 - \beta^{*}_{x} \) (with \( x = \text{sunlit, shaded} \)) is given by (cf. Eq. (14)):

\[
1 - \beta^{*}_{x} = \frac{I_{c, x}}{PAR} \quad (71)
\]

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 \( \beta^{*}_{x} \) and \( \beta^{*}_{\text{sunlit}} \) or \( \beta^{*}_{\text{shaded}} \), respectively:

\[
\frac{F_{c, \text{stom, sunlit}(O_3)}}{F_{c, \text{stom}(O_3)}} = \frac{R^{*}_{c, \text{stom, O}_3}}{R^{*}_{c, \text{stom, sunlit, O}_3}} = \frac{1 - \beta^{*}_{\text{sunlit}}}{1 - \beta^{*}} \quad (72)
\]

\[
\frac{F_{c, \text{stom, shaded}(O_3)}}{F_{c, \text{stom}(O_3)}} = \frac{R^{*}_{c, \text{stom, O}_3}}{R^{*}_{c, \text{stom, shaded, O}_3}} = \frac{1 - \beta^{*}_{\text{shaded}}}{1 - \beta^{*}} \quad (73)
\]

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

\[
F_{\text{leaf, stom, sunlit}(O_3)} = \frac{F_{c, \text{stom, sunlit}(O_3)}}{LAI_{\text{sunlit}}} \quad (74)
\]

as required by the UNECE Mapping Manual 2004 (UNECE, 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. (75)

\[
g_{\text{leaf, stom, sunlit, } O_3} = \frac{1 - \beta^{*}_{\text{sunlit}}}{R_{c, \text{stom, } O_3}} \cdot \frac{1}{LAI_{\text{sunlit}}} \quad (75)
\]
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.

4 Biosphere/atmosphere exchange of fine-particle constituents

The transport of fine-particle constituents to the surface is usually described as proportional to a dry deposition velocity \( v_D(z_{ref}) \) [m s\(^{-1}\)]. According to Erisman et al. (1994) deposition velocities for fine-particle constituents can be obtained from parameterizations in terms of Monin-Obukhov length \( L \) and friction velocity \( u_\ast \) [m s\(^{-1}\)]. For low vegetation \( v_D \) is parameterized according to Wesely et al. (1985)

\[
v_D = \frac{u_\ast}{500}, \quad L \geq 0 \text{ m}\]  \hspace{1cm} (76)

\[
v_D = \frac{u_\ast}{500} \left[1 + \left(\frac{300}{-L}\right)^{2/3}\right], \quad L < 0 \text{ m}\]  \hspace{1cm} (77)

and for forests according to Erisman et al. (1997):

\[
v_D = \frac{1}{v_{D_h}} + \frac{1}{R_{ah}(50)}\]  \hspace{1cm} (78)

where \( v_{D_h} \) is estimated from:

\[
v_{D_h} = \frac{u_\ast}{u_h} \cdot E(u_\ast)\]  \hspace{1cm} (79)

with

\( R_{ah}(50) \) turbulent atmospheric resistance between canopy height \( h \) and \( z_{ref} = 50 \text{ m [s m}^{-1}] \)

\( u_h \) horizontal wind velocity at canopy height \( h \) [m s\(^{-1}\)]

\( E(u_\ast) \) \( u_\ast \)-dependent value for fine-particle constituents [m s\(^{-1}\)] as summarized in Table 4

<table>
<thead>
<tr>
<th>Species</th>
<th>Wet surface</th>
<th>Dry surface</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Relative humidity</td>
<td></td>
</tr>
<tr>
<td></td>
<td>( rH \leq 80 % )</td>
<td>( rH &gt; 80 % )</td>
</tr>
<tr>
<td>( NH_4^+ )</td>
<td>( a = 0.66 )</td>
<td>( a = 0.66 )</td>
</tr>
<tr>
<td></td>
<td>( b = 0.41 )</td>
<td>( b = 0.41 )</td>
</tr>
<tr>
<td></td>
<td>( c = 1 )</td>
<td>( 1 + 0.37 \cdot \exp\left(\frac{rH - 80}{20}\right))</td>
</tr>
</tbody>
</table>
Table 4 (continued)
Parameterizations of $E(u_*)$ values for different components and conditions (Erisman et al., 1997)

<table>
<thead>
<tr>
<th>Species</th>
<th>Wet surface Relative humidity</th>
<th>Dry surface Relative humidity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$rH \leq 80 %$</td>
<td>$rH &gt; 80 %$</td>
</tr>
<tr>
<td>SO$_4^{2-}$</td>
<td>$a = 0.08$</td>
<td>$a = 0.08$</td>
</tr>
<tr>
<td></td>
<td>$b = 0.45$</td>
<td>$b = 0.45$</td>
</tr>
<tr>
<td></td>
<td>$c = 1$</td>
<td>$c = \left[1 + 0.37 \cdot \exp\left(\frac{rH - 80}{20}\right)\right]$</td>
</tr>
<tr>
<td>NO$_3^-$</td>
<td>$a = 0.10$</td>
<td>$a = 0.10$</td>
</tr>
<tr>
<td></td>
<td>$b = 0.43$</td>
<td>$b = 0.43$</td>
</tr>
<tr>
<td></td>
<td>$c = 1$</td>
<td>$c = \left[1 + 0.37 \cdot \exp\left(\frac{rH - 80}{20}\right)\right]$</td>
</tr>
</tbody>
</table>

5 Input parameters needed

To run PLATIN, the following data are needed depending on what is to be calculated by the model. First of all, informations on site location and site characteristics must be defined:

- latitude $\varphi_{\text{geo}}$ [degree]
- longitude $\lambda_{\text{geo}}$ [degree]
- height above sea level [m]; required for site comparisons only
- vegetation type: short vegetation or forest (selector switch)
- soil moisture: site-specific field capacity $SM_C$ [m$^3$·m$^{-3}$]
- soil moisture: site-specific wilting point $SM_W$ [m$^3$·m$^{-3}$]

Additionally,

- vegetation type-specific minimum bulk stomatal resistance for water vapour $R_{c,\text{stom},\text{min},\text{H}_2\text{O}}$ [s·m$^{-1}$]; cf. Chapter 2.3
- vegetation type-specific attenuation coefficient at solar elevation of 90° $k_{b,90}$; cf. Chapter 2.3

must be given. The vegetation type-specific values of $R_{c,\text{stom},\text{min},\text{H}_2\text{O}}$ and $k_{b,90}$ must be adjusted via comparisons of measured and modelled latent heat flux densities. Furthermore, the following informations are needed:

- number of days in the respective year (365 or 366)
- difference between Local Standard Time and Greenwich Mean Time [h]
- duration of measurement interval [h]

This data set provides the framework to operate the various modules and submodules constituting PLATIN, cf. Figure 8.

As mentioned earlier (see Chapter 1), the PLATIN core module solves the canopy energy balance including all the resistances also relevant for trace gas exchange. The solution of the energy balance requires auxiliary modules (radiation, soil moisture, etc.). Results are passed over to modules which calculate the exchange of trace gases (especially O$_3$) and fine-particle constituents. Finally, a special submodule estimates the stomatal O$_3$ uptake of the sunlit leaf fraction yielding the leaf area-related
stomatal conductance for sunlit parts of the plant stand and thus providing an interface to measurements of gas exchange on leaf level.

To estimate the fluxes mentioned in Figure 8, the following data set is indispensably needed on a half-hourly or hourly base:
- canopy net radiation balance $R_{\text{net}}$ [W m$^{-2}$]; cf. Appendix A if $R_{\text{net}}$ cannot be measured
- global radiation [W m$^{-2}$]
- photosynthetically active radiation $\text{PAR}$ [µmol m$^{-2}$ s$^{-1}$]; if PAR is not available see Appendix L
- horizontal wind speed $u$ [m s$^{-1}$] at a reference height $z_{\text{ref},u}$ above the canopy
- air temperature $t_a$ [°C] at a reference height $z_{\text{ref},T}$ above the canopy
- relative air humidity $rH$ [%] at a reference height $z_{\text{ref},rH} = z_{\text{ref},T}$ above the canopy
- air pressure $p$ [hPa] at a reference height $z_{\text{ref},p}$ (default value: 1013.25 hPa) above the canopy
- precipitation $\text{Precip}$ [mm H$_2$O $\equiv 10^{-3}$ m$^3$ m$^{-2}$]

and on (at least) a daily base:
- soil moisture $\text{SM}$ in the uppermost soil layer (e.g. 0 - 16 cm) [m$^3$ m$^{-3}$]
- roughness length for momentum $z_{0m}$ [m] or at least canopy height $h$ [m]
- displacement height $d$ [m] or at least canopy height $h$ [m]
- leaf area index of non-senescent leaves $\text{LAI}_{\text{non-senescent}}$ [m$^2$ m$^{-2}$]
- total surface area of the vegetation $\text{SAI}$ [m$^2$ m$^{-2}$]

The estimation of vertical flux densities of trace gas or fine-particle constituents requires additional data to be provided on half-hourly or hourly base:
- concentrations $\rho_A$ of trace gases or fine-particle constituents [ppb or µg m$^{-3}$] at a reference height $z_{\text{ref},A}$ above the canopy
- vegetation type-specific ratio of ammonium to protons in the apoplast $\Gamma$ [mol mol$^{-1}$] if NH$_3$ exchange is to be modelled; for default values see Chapter 3.4
While all the data mentioned above serve to drive one or more modules of PLATIN, calibration of PLATIN requires the following flux densities to be known (i.e. to be measured) on half-hourly or hourly base:

- sensible heat $H$ [W m$^{-2}$]
- latent heat $\lambda E$ [W m$^{-2}$]
- total flux of trace gas species A, e.g. ozone, $F_c(A)$ [$\mu$g m$^{-2}$ s$^{-1}$]

In some cases, missing data can be substituted by calculated values.

To calculate canopy net radiation balance $R_{\text{net}}$, PLATIN provides parameterizations to estimate global radiation from astronomically maximum possible solar irradiation (depending on cloud-cover degree), the amount of short wave radiation reflected by the big-leaf surface (making use of short-wave albedo $\alpha$), the long wave downward radiation (depending on readily available data of air temperature and relative humidity at screen height 2 m), and finally the longwave radiation sent out from the big-leaf surface. Details are given in Appendix A. Note that in case of modelling $R_{\text{net}}$ short-wave albedo $\alpha$ is required as additional input parameter.

In order to calculate soil moisture $SM$, PLATIN makes use of a refined and extended force-restore model of soil water content, see Appendix H. This submodel needs additional input parameters:

- depth of soil layer under consideration $SLD$ [m]
- maximum possible vertical water flow from groundwater into rooted soil layer due to soil capillarity $C_{\text{cap}}$ [kg m$^{-2}$ s$^{-1}$]
- $RLD_{90}$ which is the depth of the soil layer containing 90 % of total root mass [m]

Note that, if concentrations of CO$_2$ are given, PLATIN is able to account for the influence of photosynthesis on the canopy energy balance. In general, photosynthesis is equivalent to an additional sink in the energy balance up to approx. 20 W m$^{-2}$. To estimate the CO$_2$ exchange between canopy and near-surface atmosphere additional data are needed:

- CO$_2$ concentration [ppm] at a reference height $z_{\text{ref, CO2}}$ on half-hourly or hourly base
- soil temperature $t_{\text{soil}}$ on half-hourly or hourly base [$^\circ$C]
- residual (irreducible) soil moisture [m$^3$ m$^{-3}$]
- soil moisture content at saturation [m$^3$ m$^{-3}$]
- nitrogen content of the aboveground sapwood biomass [g m$^{-2}$]

Calibration of the CO$_2$ exchange submodule requires measured data of

- carbon dioxide exchange $F_c(CO_2)$ [$\mu$mol m$^{-2}$ s$^{-1}$]

on a half-hourly or hourly base.

If PLATIN for Excel is to be applied to forest ecosystems, information on the state of the air above the forest canopy is required (wind speed, temperature, humidity, concentrations). If data are not available they have to be estimated from measurements above short vegetation near the forest (cf. Appendix K).

References


Erisman JW, Draaijers GPJ, Steingröver E, van Dijk H, Boxman A, de Vries W (1998) Assessment of the exposure and loads of acidifying and eutrophying pollutants and ozone, as well as their harmful influence on the vitality of the trees and the Speulder forest ecosystem as a whole. Water Air Soil Pollut 105: 539-571
Fowler D, Flechard C, Cape JN, Storeton-West RL, Coyle M (2001) Measurements of ozone deposition to vegetation quantifying the flux, the stomatal and non-stomatal components. Water Air Soil Pollut 130: 63-74


Massman WJ (1998) A review of the molecular diffusivities of H₂O, CO₂, CH₄, CO, O₃, SO₂, NH₃, N₂O, NO, and NO₂ in air, O₂ and N₂ near STP. Atmos Environ 32: 1111-1127


Sorteberg A, Hov Ø (1996) Two parametrizations of the dry deposition exchange for SO₂ and NH₃ in a numerical model. Atmos Environ 30: 1823-1840
Appendix A

Parameterization of net radiation

In cases where net radiation balance $R_{net}$ [W·m$^{-2}$] is not measured it can be estimated according to:

$$ R_{net} = S_t \cdot (1 - \alpha) + \varepsilon \cdot L_d - L_u $$

(A1)

with

- $S_t$ global radiation [W·m$^{-2}$]
- $\alpha$ short-wave albedo
- $\varepsilon$ effective long-wave emissivity of the canopy (default value: $\varepsilon = 0.97$)
- $L_d$ flux density of downward long-wave radiation of the atmosphere [W·m$^{-2}$]
- $L_u$ flux density of upward long-wave radiation of the atmosphere [W·m$^{-2}$]

Thermal radiation emitted by the canopy $L_u$ is determined by surface temperature $T_s$ [K] at $z = d + z_{oh}$ and its effective longwave emissivity $\varepsilon$:

$$ L_u = \varepsilon \cdot \sigma \cdot T_s^4 $$

(A2)

with

- $\sigma$ Stefan-Boltzmann constant ($= 5.669 \cdot 10^{-8}$ W·m$^{-2}$·K$^{-4}$)

If no measured data are available for $L_d$, it can be approximated, e.g. by parameterisations given in the literature (e.g. Brutsaert, 1984). In general, such parameterisations are based on a clear-sky approach combined with a function of cloud cover degree, cf. eqs. (6.18) and (6.24) in Brutsaert (1984) which adopted for the first version of PLATIN (Grünhage and Haenel, 1997). However, as it is not an easy task to provide cloud cover data routinely, we included an approximation of daytime cloud cover based on the relation of actual global radiation to the maximum possible global radiation, while the nighttime cloud cover was simply assumed to be equal to the latest daytime cloud cover.

During the vegetation periods 1998 and 1999 at the Linden field site at 50.53°N 8.69°E, this approach lead to daytime $L_d$ values considerably different from measured $L_d$ values as is demonstrated by some statistics shown in Table A1.

Table A1: Results of regression analysis of simulated (by former model) vs. observed $L_d$

<table>
<thead>
<tr>
<th></th>
<th>N</th>
<th>$R^2$</th>
<th>bias (W·m$^{-2}$)</th>
<th>standard deviation (W·m$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MAY - SEPTEMBER 1998</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>daylight hours ($S_t \geq 50$ W·m$^{-2}$)</td>
<td>3735</td>
<td>0.457</td>
<td>27.61</td>
<td>46.0</td>
</tr>
<tr>
<td>nighttime</td>
<td>2820</td>
<td>0.539</td>
<td>−20.75</td>
<td>39.2</td>
</tr>
<tr>
<td>MAY - SEPTEMBER 1999</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>daylight hours ($S_t \geq 50$ W·m$^{-2}$)</td>
<td>3830</td>
<td>0.474</td>
<td>38.05</td>
<td>53.1</td>
</tr>
<tr>
<td>nighttime</td>
<td>2804</td>
<td>0.532</td>
<td>−17.86</td>
<td>39.4</td>
</tr>
</tbody>
</table>

For regression analysis during daylight hours only data sets with $S_t \geq 50$ W·m$^{-2}$ were selected to avoid uncertainties during transition times.

Although the nighttime $L_d$ simulation is based on a comparably rough cloud cover estimate, the standard deviation and $R^2$ were still better than the respective values for the daytime parameterisation. Thus we decided to modify the cloud cover influence while we kept the clear-sky approach of Brutsaert (1984; eq. (6.18)) because of its high physical content (the validity of which was confirmed again only recently by Kjaersgaard et al., 2007). Our improved correction for cloud cover effect should not try to simulate cloud cover degree but should be directly based on screen-height data of global radiation and relative humidity for daytime estimates of $L_d$, and on relative humidity and air temperature for nighttime estimates of $L_d$. All input data were routinely recorded at the Linden field site at 50.53°N 8.69°E within BIATEX-2, a subproject of EUROTRAC-2 (cf. Midgley et al., 2003). The new approach and its prelimi-
nary calibration by $L_d$-measurements of one single month (May 1999) were published in Grünhage and Haenel (2000). Subsequently the new $L_d$ model has been calibrated and tested for the vegetation periods 1998 and 1999 (unpublished), which lead to some minor adjustments of the constants but left unchanged the structure of the $L_d$ model which is given below by eqs. (A3) through (A14). It should be kept in mind that our parameterisation is locally calibrated for Linden and its application to other sites should be accompanied by some sufficient measurements of $L_d$ (cf. discussion of a model similar to ours in Kjaersgaard et al., 2007).

Table A2 gives some statistics of the results obtained with our new parameterisation. Comparison with Table A1 reveals that nighttime simulation of $L_d$ could slightly be improved by the new approximation, but great progress was achieved in daytime $L_d$ estimates.

<table>
<thead>
<tr>
<th></th>
<th>N</th>
<th>$R^2$</th>
<th>bias $(W \cdot m^{-2})$</th>
<th>standard deviation $(W \cdot m^{-2})$</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>May - September 1998</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>daylight hours ($S_t \geq 50 \ W \cdot m^{-2}$)</td>
<td>3735</td>
<td>0.721</td>
<td>-0.91</td>
<td>22.3</td>
</tr>
<tr>
<td>nighttime</td>
<td>2820</td>
<td>0.580</td>
<td>-5.51</td>
<td>24.0</td>
</tr>
<tr>
<td><strong>May - September 1999</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>daylight hours ($S_t \geq 50 \ W \cdot m^{-2}$)</td>
<td>3830</td>
<td>0.734</td>
<td>0.89</td>
<td>21.6</td>
</tr>
<tr>
<td>nighttime</td>
<td>2804</td>
<td>0.617</td>
<td>5.60</td>
<td>25.1</td>
</tr>
</tbody>
</table>

The new $L_d$ parameterisation is given by the subsequent equations:

- **daytime $L_d$** (global radiation $S_t \geq 50 \ W \cdot m^{-2}$):

$$L_{d, S_t \geq 50 \ W \cdot m^{-2}} = 1.24 \left( \frac{e_{2m}}{T_{2m}} \right)^{0.7} \cdot \sigma \cdot T_{2m}^4 + A_{dd} + B_{dd} + C_{dd}$$  \hspace{1cm} (A3)

where $e_{2m}$ is the actual water vapour pressure [hPa] and $T_{2m}$ is the absolute air temperature [K] at $z = 2 \ m$ above ground.

and

$$A_{dd} = 16.9 - 70 \cdot \frac{S_t}{S_{t, ref}} \quad \text{for} \quad S_t \leq S_{t, ref}$$  \hspace{1cm} (A4)

$$A_{dd} = -53.1 \quad \text{for} \quad S_t > S_{t, ref}$$  \hspace{1cm} (A5)

$$B_{dd} = 2.3 + 1.33 \cdot (rH - 40) \quad \text{for} \quad rH \geq 40 \%$$  \hspace{1cm} (A6)

$$B_{dd} = 2.3 \quad \text{for} \quad rH < 40 \%$$  \hspace{1cm} (A7)

and

$$C_{dd} = -12.3 + 1.1 \cdot (rH - 70) \quad \text{for} \quad rH \geq 70 \%$$  \hspace{1cm} (A8)

$$C_{dd} = -12.3 + 0.6 \cdot (70 - rH) \quad \text{for} \quad rH < 70 \%$$  \hspace{1cm} (A9)

where $S_{t, ref}$ is the maximum possible global radiation according to eq. (A15) and $rH$ is the relative humidity [%] at $z = 2 \ m$ above ground.

The main contribution to daylight $L_d$ is given by the first term on the right hand side of eq. (A3). It is semi-empirical, but with a firm physical background (see derivation of eq. (6.18) in Brutsaert, 1984).
Term $A_{dn}$ represents a linear ad-hoc approach to estimate cloud cover influence by making use of the ratio of actual solar radiation to maximum possible solar irradiation. Term $B_{dn}$ was derived formally (with the help of some minor simplifications) from the definition of the lifting condensation level (cf. Stull 1988). Other than for $A_{dn}$ and $B_{dn}$, no direct physical explanation seems possible for the third term, $C_{dn}$, the contribution of which to the variance, however, is much smaller than that of the other terms. The constants in $A_{dn}$, $B_{dn}$, and $C_{dn}$ were separately adjusted to minimise the bias resulting from the variable part in each of these three terms.

- \textbf{nighttime $L_d$ ($S_t < 50 \ W \ m^{-2}$):}

$$L_{dn, S_t=0 \ W \ m^{-2}} = 1.24 \left( \frac{e_{2m}}{T_{2m}} \right)^{1/7} \cdot \sigma \cdot T_{2m}^4 + A_{dn} + B_{dn} \quad (A10)$$

with

- $$A_{dn} = 14 + 10.7 \cdot (rH - 92.5) \quad \text{for} \quad rH \geq 92.5 \% \quad (A11)$$
- $$A_{dn} = 14 \quad \text{for} \quad rH < 92.5 \% \quad (A12)$$

and

- $$B_{dn} = 11 - 20 \cdot \sqrt{T_{2m,n-1} - T_{2m,n}} \quad \text{for} \quad (T_{2m,n-1} - T_{2m,n}) > 0 \quad (n: \text{actual data set}) \quad (A13)$$
- $$B_{dn} = 11 \quad \text{for} \quad (T_{2m,n-1} - T_{2m,n}) \leq 0 \quad (A14)$$

where $e_{2m}$ is the actual water vapour pressure (hPa) and $T_{2m}$ is the absolute air temperature (K) at $z = 2$ m above ground.

The main contribution to nighttime $L_d$ is given by the first term on the right hand side of eq. (A10) which is known already from eq. (A3). Using a plot of the residual of the parameterised clear sky $L_d$ against measured $L_d$, term $A_{dn}$ was derived as a function of relative humidity. No other meteorological variable measured at Linden investigation site (like wind speed or air temperature) could significantly contribute to further improvement of the parameterisation. Term $B_{dn}$ takes into account that the decrease of nocturnal air temperature is well correlated with fractional cloud cover (depending of course on local surface conditions). The constants in $A_{dn}$ and $B_{dn}$ were separately adjusted to minimise the bias resulting from the introduction of the variable parts of $A_{dn}$ and $B_{dn}$. \[ \]

- \textbf{maximum possible global radiation $S_{t, \text{ref}}$}

$S_{t, \text{ref}}$ is the astronomic maximum possible global radiation at cloudless sky parameterised according to Kasten and Czeplak (1980)

$$S_{t, \text{ref}} = S_{t, \text{cloudless sky}} = a_{\text{Stref,1}} \cdot \sin \varphi + a_{\text{Stref,2}} \quad (A15)$$

where $a_{\text{Stref,1}}$ and $a_{\text{Stref,2}}$ are empirical coefficients describing the average atmospheric attenuation of short-wave radiation by water vapour and dust at a given site, and $\varphi$ is solar elevation. For the Linden field site at 50.53°N 8.69°E, $a_1$ and $a_2$ were adjusted to $a_1 = 1097 \ W \ m^{-2}$ and $a_2 = -54 \ W \ m^{-2}$.

Solar elevation $\varphi$ is calculated depending on latitude, longitude and time according to Lenoble (1993):

$$\sin \varphi = \sin \varphi_{\text{geo}} \cdot \sin \Delta_x + \cos \varphi_{\text{geo}} \cdot \cos \Delta_x \cdot \cos \varphi_h \quad (A16)$$

with

- $\varphi_{\text{geo}}$ latitude [radians]
- $\Delta_x$ sun declination [radians]
- $\varphi_h$ hour angle [radians]
Sun declination $\Delta_{\text{sun}}$ is given by:

$$\Delta_{\text{sun}} = 0.006918 - 0.399912 \cdot \cos \phi_d + 0.070257 \cdot \sin \phi_d$$

$$- 0.006758 \cdot \cos 2\phi_d + 0.000907 \cdot \sin 2\phi_d$$  \hfill (A17)

where the day angle $\phi_d$ [radians] is:

$$\phi_d = 2 \cdot \pi \cdot \frac{\text{day of the year} - 1}{\text{number of days in the year}}$$  \hfill (A18)

The hour angle $\phi_h$ is given by:

$$\phi_h = \pi \cdot \left(1 - \frac{TST}{12}\right)$$  \hfill (A19)

where TST is the True Solar Time [h; decimal system] for the center of the time interval under consideration:

$$TST = \text{GMT} + \frac{\lambda_{\text{geo}}}{15} + \text{ET} - \frac{\text{DT}}{2}$$  \hfill (A20)

with
- GMT Greenwich Mean Time (for Germany: CET - 1) [h]
- CET Central European Time [h]
- $\lambda_{\text{geo}}$ longitude [degree]
- DT duration of time interval [h]

and the equation of time ET [h]:

$$\text{ET} = 3.819667 \cdot (0.000075 + 0.001868 \cdot \cos \phi_d - 0.032077 \cdot \sin \phi_d$$

$$- 0.014615 \cdot \cos 2\phi_d - 0.040849 \cdot \sin 2\phi_d)$$  \hfill (A21)
Appendix B

Radiation model

The sunlit and shaded leaf area indices of the canopy as well as the irradiance absorbed by the sunlit and shaded leaf fractions are parameterised according to de Pury and Farquhar (1997).

• Sunlit and shaded leaf area index

The sunlit leaf area index of the non-senescent leaves of the canopy \[ \text{m}^2 \cdot \text{m}^{-2} \] is

\[
LAI_{\text{sunlit}} = \frac{1 - e^{-k_b \cdot LAI_{\text{non-senescent}}}}{k_b}
\]

where \( k_b \) is a canopy-typical light attenuation coefficient:

\[
k_b = \frac{k_{b,90^\circ}}{\sin \phi}
\]

with \( k_{b,90^\circ} \) value for solar elevation of 90° (cf. Chapter 2.3)

\( \phi \) solar elevation (see eq. (A16) in Appendix A)

The shaded leaf area index of the canopy \[ \text{m}^2 \cdot \text{m}^{-2} \] is then calculated from:

\[
LAI_{\text{shaded}} = LAI_{\text{non-senescent}} - LAI_{\text{sunlit}}
\]

• Irradiance absorbed by the sunlit and shaded leaf fractions of the canopy

The irradiance absorbed by the canopy is calculated as the sum of direct-beam, diffusive and scattered-beam components of photosynthetically active radiation \( \text{PAR} \) measured above the canopy \[ \mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} \].

Direct beam irradiance \( I_b \) \[ \mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} \] is given by

\[
I_b = PAR \cdot (1 - f_d)
\]

and diffuse irradiance \( I_d \) \[ \mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} \] by:

\[
I_d = PAR \cdot f_d
\]

with \( f_d \) fraction of diffusive radiation (proportion of total attenuated radiation, beam plus diffuse) for cloudless skies

\( f_d \) is obtained by

\[
f_d = \frac{1 - a_{\text{PAR}}^m}{1 + a_{\text{PAR}}^m \cdot (1/f_a - 1)}
\]

with \( a_{\text{PAR}} \) atmospheric transmission coefficient of photosynthetically active radiation \( \text{PAR} \) for clear sky conditions

(default value: \( a_{\text{PAR}} = 0.72 \) according to de Pury and Farquhar, 1997)

where the optical air mass \( m \) is defined as the ratio of the mass of atmosphere traversed per unit cross-sectional area of the solar beam to that traversed for a site at sea level if the sun was directly overhead:
\[ m = \frac{p \cdot p_0^{-1}}{\sin \phi} \]  

with \( p \) actual atmospheric pressure [hPa]  
\( p_0 \) atmospheric pressure at sea level (\( p_0 = 1013.25 \) hPa)

and \( f_s \) is the proportion of attenuated radiation that reach the surface as diffuse radiation. Under cloudless skies it has been observed to range from 40 to 45% (Weiss and Norman, 1985). In accordance with de Pury and Farquhar (1997) \( f_s = 0.426 \) is chosen as a default value.

The total irradiance absorbed by the sunlit leaf fraction of the canopy \( l_{c, \text{sunlit}} \) [\( \mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} \)] is given as the sum of direct-beam, diffusive and scattered-beam components:

\[ I_{c, \text{sunlit}} = I_{c, \text{direct-beam, sunlit}} + I_{c, \text{diffuse, sunlit}} + I_{c, \text{scattered-beam, sunlit}} \]  

Direct-beam irradiance absorbed by sunlit leaves \( l_{c, \text{direct-beam, sunlit}} \) is given by

\[ I_{c, \text{direct-beam, sunlit}} = I_b \cdot (1 - \sigma_{\text{PAR}}) \cdot \{1 - \exp(-k_b \cdot LAI_{\text{non-senescent}})\} \]  

with \( \sigma_{\text{PAR}} \) leaf scattering coefficient of PAR  
(default value: \( \sigma_{\text{PAR}} = 0.15 \) according to de Pury and Farquhar, 1997)

Diffuse irradiance absorbed by sunlit leaves \( l_{c, \text{diffuse, sunlit}} \) can be calculated from:

\[ I_{c, \text{diffuse, sunlit}} = I_d \cdot (1 - \rho_{cd}) \cdot \{1 - \exp[-(k'_d + k_b) \cdot LAI_{\text{non-senescent}}]\} \cdot \frac{k'_d}{k'_d + k_b} \]  

with \( \rho_{cd} \) canopy reflection coefficient for diffuse PAR  
(default value: \( \rho_{cd} = 0.036 \) according to de Pury and Farquhar, 1997)

The extinction coefficient \( k'_d \) is given by

\[ k'_d = k_d \cdot \sqrt{1 - \sigma_{\text{PAR}}} = 0.78 \cdot \sqrt{1 - \sigma_{\text{PAR}}} \]  

with \( k_d \) diffuse PAR extinction coefficient  
(default value: \( k_d = 0.78 \) according to de Pury and Farquhar, 1997)

The scattered-beam irradiance absorbed by sunlit leaves \( l_{c, \text{scattered-beam, sunlit}} \) results from:

\[ I_{c, \text{scattered-beam, sunlit}} = I_b \cdot \left[ (1 - \rho_{cb}) \cdot \{1 - \exp[-(k'_b + k_b) \cdot LAI_{\text{non-senescent}}]\} \cdot \frac{k'_b}{k'_b + k_b} \right] \]

\[ - (1 - \sigma_{\text{PAR}}) \cdot \frac{1 - \exp(-2 \cdot k_b \cdot LAI_{\text{non-senescent}})}{2} \]  

with \( k'_b \) beam and scattered beam PAR extinction coefficient

\[ k'_b = k_b \cdot \sqrt{1 - \sigma_{\text{PAR}}} \]  

\( \rho_{cb} \) canopy reflection coefficient for beam PAR
The reflection coefficient $\rho_{cb}$ is defined as

$$\rho_{cb} = 1 - \exp\left[-2 \cdot \frac{\rho_h \cdot k_b}{1 + k_b}\right] \quad (B14)$$

with $\rho_h$ reflection coefficient for beam PAR of a canopy with horizontal leaves

$$\rho_h = \frac{1 - \sqrt{1 - \sigma_{\text{PAR}}}}{1 + \sqrt{1 - \sigma_{\text{PAR}}}} \quad (B15)$$

The total irradiance absorbed by the shaded leaf area $I_{c, \text{shaded}}$ [µmol m$^{-2}$ s$^{-1}$] is the sum of diffuse and scattered diffuse irradiance $I_{c, \text{diffuse, shaded}}$ and of scattered beam irradiance $I_{c, \text{scattered-beam, shaded}}$ absorbed by the shaded leaves:

$$I_{c, \text{shaded}} = I_{c, \text{diffuse and scattered diffuse, shaded}} + I_{c, \text{scattered-beam, shaded}} \quad (B16)$$

with

$$I_{c, \text{diffuse and scattered diffuse, shaded}} = I_d \cdot (1 - \rho_{cd}) \cdot \left\{ \frac{1 - \exp\left[-k_d' \cdot LAI_{\text{non-senescent}}\right]}{- \left\{1 - \exp\left[-\left(k'_d + k_b\right) \cdot LAI_{\text{non-senescent}}\right]\right\} \cdot \frac{k'_d}{k'_d + k_b}} \right\} \quad (B17)$$

and

$$I_{c, \text{scattered-beam, shaded}} = I_b \cdot \left\{ \frac{1 - \exp\left[-k'_b \cdot LAI_{\text{non-senescent}}\right]}{\left\{1 - \exp\left[-\left(k'_b + k_b\right) \cdot LAI_{\text{non-senescent}}\right]\right\} \cdot \frac{k'_b}{k'_b + k_b}} \right\} \quad \left\{1 - \exp\left[-k_b \cdot LAI_{\text{non-senescent}}\right] - \frac{1 - \exp\left(-2 \cdot k_b \cdot LAI_{\text{non-senescent}}\right)}{2}\right\} \quad (B18)$$
Appendix C

Parameterization of the atmospheric stability functions

The atmospheric stability functions for momentum \( \Psi_m \) and sensible heat \( \Psi_h \) are calculated as described in Haenel and Siebers (1995). Adopting the set of empirical coefficients given by Dyer (1974), i.e. \( a_u = a_s = 1, b_u = -16, \) and \( b_s = 5 \) in context with \( \kappa = 0.41 \), one obtains for \( \Psi_m \) and \( \Psi_h \):

- **unstable atmospheric stratification** (\( L < 0 \) m)
  \[
  \Psi_m(\zeta) = 2 \cdot \ln \left[ \frac{1}{\Phi_m(\zeta)} + 1 \right] + \ln \left[ \frac{1}{\Phi_m^2(\zeta)} + 1 \right] - 2 \cdot \arctan \left[ \frac{1}{\Phi_m(\zeta)} \right] \quad (C1)
  \]
  with
  \[
  \Phi_m(\zeta) = a_u \cdot (1 + b_u \cdot \zeta)^{-0.25} = (1 - 16 \cdot \zeta)^{-0.25} \quad (C2)
  \]
  \[
  \zeta = \frac{z - d}{L} \quad \text{with e.g. } z = z_1 = d + z_{0m} \quad \text{and } z = z_2' = z_{ref, u}
  \]
  and
  \[
  \Psi_h(\zeta) = 2 \cdot \ln \left[ \frac{1}{\Phi_h(\zeta)} + 1 \right] \quad (C4)
  \]
  with
  \[
  \Phi_h(\zeta) = a_s \cdot (1 + b_s \cdot \zeta)^{-0.5} = (1 - 16 \cdot \zeta)^{-0.5} \quad (C5)
  \]
  \[
  \zeta = \frac{z - d}{L} \quad \text{with e.g. } z = z_1 = d + z_{0m} \quad \text{and } z = z_2 = z_{ref, T}
  \]

- **stable atmospheric stratification** (\( L > 0 \) m)
  \[
  \Psi_m(\zeta) = \Psi_h(\zeta) = (1 - a_s) \cdot \ln \zeta - a_s \cdot b_s \cdot \zeta = -5 \cdot \zeta \quad (C7)
  \]
  with
  \[
  \zeta = \frac{z - d}{L} \quad \text{and e.g. } z = z_1 = d + z_{0m}, \quad z = z_2 = z_{ref, T}, \quad z = z_2' = z_{ref, u}
  \]

  Note that for stable conditions with \( \zeta \) approaching unity, stability functions of the type of eq. (C7) are not valid as briefly reviewed by Foken (2003). \( \Psi_m \) and \( \Psi_h \) should then attain a constant value in order to account for the fact that, for more and more increasing stability, the magnitude of turbulence elements is no longer determined by the height \( z \) but by the Monin-Obukhov length (so-called \( z \)-less scaling). Therefore, according to Handorf et al. (1999), for \( \zeta > 0.8 \) the atmospheric stability functions (C7) are restricted in order not to fall below the value of \(-4\).

- **neutral atmospheric stratification** (\( |L| \to \infty \))
  \[
  \Psi_m = \Psi_h = 0 \quad (C9)
  \]

  In the case of the estimation of flux densities of ozone, sulphur dioxide, nitrogen dioxide, carbon dioxide or fine-particle constituents, the height \( z = z_2 \) represents the concentration measurement height of these species.
Appendix D

Parameterization of ground heat flux density and energy exchange due to photosynthesis

The ground heat flux \( G \) at \( z = d + z_0 \) in the big-leaf approach is the total of three physical heat storage flux densities \( G_{\text{physical heat storage}} \) and the energy exchange due to photosynthesis:

\[
G = G_{\text{soil}} + S_{\text{in-canopy air}} + S_{\text{vegetation}} + Ph
\]

\[
= G_{\text{physical heat storage}} + Ph
\]  

\( (D1) \)

with

- \( G_{\text{soil}} \): soil heat flux density \([\text{W} \cdot \text{m}^{-2}]\)
- \( S_{\text{in-canopy air}} \): flux of energy due to changes in temperature and humidity of the air in the canopy \([\text{W} \cdot \text{m}^{-2}]\)
- \( S_{\text{vegetation}} \): flux of energy due to changes in temperature of the aboveground biomass \([\text{W} \cdot \text{m}^{-2}]\)
- \( Ph \): energy exchange due to photosynthesis and respiration of the aboveground biomass \([\text{W} \cdot \text{m}^{-2}]\)

For low vegetation (e.g. grass, agricultural crops) \( G_{\text{soil}} \) is the dominant part of \( G_{\text{physical heat storage}} \), allowing to neglect other storage fluxes in eq. (D1), while this is not possible for forests (cf. Frühauf, 1998).

- Physical heat storage flux density \([\text{W} \cdot \text{m}^{-2}]\)

In PLATIN, \( G_{\text{physical heat storage}} \) is parameterised as proportional to the net radiative flux density \( R_{\text{net}} \) (cf. Brutsaert, 1984).

\[
G_{\text{physical heat storage}} = a_{1,G\text{phs}} \cdot e^{-k_{b,max} \cdot SAI} \cdot R_{\text{net}} \quad \text{if} \quad R_{\text{net}} \geq 0 \quad \text{W} \cdot \text{m}^{-2} \]  

\( (D2) \)

\[
G_{\text{physical heat storage}} = a_{2,G\text{phs}} \cdot R_{\text{net}} \quad \text{if} \quad R_{\text{net}} < 0 \quad \text{W} \cdot \text{m}^{-2} \]  

\( (D3) \)

with

- \( k_{b,max} \): attenuation coefficient of the canopy at 12 h TST (true solar time)
- \( SAI \): total surface area of the vegetation \([\text{m}^2 \cdot \text{m}^{-2}]\)

This parameterisation is based on the fact that, beneath a plant stand, \( G_{\text{physical heat storage}} \) represents only a relatively small fraction of the energy balance and that it follows more or less the time course of net radiation \( R_{\text{net}} \) as the most important fraction of the energy balance. During daylight hours, \( R_{\text{net}} \) is driven by incoming solar radiation and therefore the parameterisation of \( G_{\text{soil}} \) must account for shortwave radiation extinction within the plant stand. This can simply be parameterised by \( \exp(-k_{b,max} \cdot SAI) \) as discussed in Chapter 2.3. During nighttime, on the other hand, \( R_{\text{net}} \) only consists of the longwave radiation balance so that there is no need to consider the plant stand architecture when parameterising nocturnal \( G \).

The empirical coefficients \( a_{1,G\text{phs}} \) and \( a_{2,G\text{phs}} \) should be determined from energy balance measurements for each specific vegetation type separately. However, considering the subordinate contribution of \( G \) to the energy balance as well as the simplicity of the approach given above for \( G \), we restrict ourselves to find reasonable values of \( a_{1,G\text{phs}} \) and \( a_{2,G\text{phs}} \) for low vegetation like or grassland and agricultural crops (including the asymptotic limit of bare soil) and forests.

Neglecting the fact that measured energy balances do never close completely due to measurement errors and different footprints of the contributing components (cf. Foken, 2003), we calculated the energy balance residual \( E_{\text{B residual}} \) which is assumed to be identical with \( G \):

\[
E_{\text{B residual}} = G = R_{\text{net}} - H - \lambda E
\]  

\( (D4) \)
From modelled $G$ according to eqs. (D2) and (D3), for the Linden semi-natural grassland $a_1, G_{phs}$ can be approximated by 0.55 and $a_2, G_{phs}$ by 0.9. These values are used in PLATIN as default values for low vegetation (e.g. grassland, crops).

From ratios of $G_{physical heat storage}/R_{net}$ published in the literature (cf. Fig. 4.21 in Oke, 1978) default values $a_1, G_{phs} = a_2, G_{phs} \approx 1$ are deduced for forest ecosystems.

**energy exchange due to photosynthesis [W m$^{-2}$]**

The energy needed to fix one mole of CO$_2$ is $477 \cdot 10^3$ J. Hence $Ph$ is approximated as:

$$Ph = 477 \cdot 10^3 \text{ J} \cdot \text{mol}^{-1} \cdot A_{net, canopy}$$  \hspace{1cm} (D5)

with $A_{net, canopy}$ net rate of canopy photosynthesis in $\mu$mol $m^{-2} \cdot s^{-1}$

which results in an energy quantity needed to fix 1 $\mu$mol $m^{-2} \cdot s^{-1}$ CO$_2$ in plant photosynthesis of approximately 0.5 W $m^{-2}$.

$A_{net, canopy}$ is calculated according to de Pury and Farquhar (1997) taking into account the sunlit and shaded leaf fractions of the canopy as described in Appendix G.
Appendix E

*Jarvis functions for radiation, temperature, water vapour pressure deficit of the atmosphere, soil moisture, phenology, ozone and carbon dioxide*

- **Jarvis-Stewart function for radiation**

\[
f_1(S_1) = \frac{S_1}{S_1 + S_2} \left( \frac{S_1 + S_2}{S_1} \right)
\]

(E1)

with 
- \( S_1 \) actual global radiation [W·m\(^{-2}\)]
- \( S_1 \) maximum global radiation (default value: \( S_1 = 1000 \) W·m\(^{-2}\))
- \( S_2 \) empirical coefficient governing the shape of the function (default value: \( S_2 = 100 \) W·m\(^{-2}\))

- **Jarvis-Stewart function for temperature**

\[
f_2(t_a) = \left( \frac{t_3 - t_1}{t_2 - t_1} \right) \left( \frac{t_3 - t_a}{t_3 - t_2} \right) \left( t_3 - t_1 \right)
\]

(E2)

with 
- \( t_a \) actual temperature [°C]
- \( t_1 \) vegetation-type dependent minimum temperature (default value: \( t_1 = 0 \) °C)
- \( t_2 \) vegetation-type dependent optimum temperature (default value: \( t_2 = 20 \) °C)
- \( t_3 \) vegetation-type maximum temperature at which stomata no longer remain open (default value: \( t_3 = 40 \) °C)

Fig. E1
Dependence of the relative stomatal resistance on solar radiation for various coefficients \( S_2 \) (left) and on air temperature for various coefficients \( t_2 \) (right)
• Jarvis-Stewart function for water vapour pressure deficit in the atmosphere

\[ f_3(VPD) = \min \left[ 1, \max \left( \frac{VPD - V_1}{V_2 - V_1}, V_3 \right) \right] \]  

(E3)

with

- \( VPD \) actual water vapour pressure deficit of the air [hPa]  
  (see Appendix F, eq. (F1 - F4))
- \( V_1 \) maximum value for \( VPD \) (default value: \( V_1 = 40 \) hPa)
- \( V_2 \) threshold value for \( VPD \) (default value: \( V_2 = 10 \) hPa)
- \( V_3 \) minimum threshold (default value: \( V_3 = 0.15 \))

• Jarvis-Stewart function for soil moisture

\[ f_4(SM) = \min \left[ 1, \max \left( \frac{SM - SM_w}{SM_c - W_1} - SM_w, W_2 \right) \right] \]  

(E4)

with

- \( SM \) actual soil water content [m\(^3\)m\(^{-3}\)] for a specific soil layer  
- \( SM_w \) site-specific wilting point [m\(^3\)m\(^{-3}\)]  
- \( SM_c \) site-specific field capacity [m\(^3\)m\(^{-3}\)]  
- \( W_1 \) site-specific threshold level [m\(^3\)m\(^{-3}\)] (default value: \( W_1 = 0.25 \cdot SM_c \))  
- \( W_2 \) minimum threshold (default value: \( W_2 = 0.15 \))

Fig. E2
Dependence of the relative stomatal resistance on water vapour pressure deficit of the atmosphere for different coefficients \( V_2 \) (left) and on soil moisture (right)
**Jarvis-Stewart function for VPD and soil moisture**

In general, the Jarvis-Stewart functions for water vapour pressure deficit in the atmosphere and soil moisture are combined multiplicatively. However, as plants under water stress react more sensitive to VPD, in PLATIN a combined function $f_{3,4}(VPD, SM)$ is used which accounts for the interaction effect:

$$f_{3,4}(VPD, SM) = \min \left[ 1, \max \left( \frac{VPD - (V_1 \cdot V_5)}{(V_2 \cdot V_5) - (V_1 \cdot V_3)}, V_3 \right) \right]$$

(E5)

with

$$V_5 = V_4 \cdot \frac{SM - SM_w}{(SM_c - W'_1) - SM_w}$$

(E6)

and $V_4$ empirical weighting coefficient (default value: $V_4 = 1$)

---

**Fig. E3**

Dependence of the relative stomatal resistance on water vapour pressure deficit of the atmosphere and soil moisture

---

**Jarvis-Stewart function for time of day**

A time dependent impact on stomatal resistance has been often reported in the literature (e.g. Körner, 1994; Uddling et al., 2004; Goumenaki et al., 2007). After Körner (1994), the same set of climatic conditions (e.g. light, temperature, $VPD$, soil moisture) could result in lower stomatal resistances during the morning than in the afternoon. This afternoon increase of $R_{stom, H2O}$ compared to the morning values is in the order of 20%.

For daylight hours after 2 pm the following relation, derived from the factors given in Körner et al. (1995) for *Picea abies* and *Pinus sylvestris* and in Goumenaki et al. (2007) for *Lactuca sativa*, can serve as a first approximation:

$$f_5(\text{time}) = -0.66 + 0.279 \cdot \text{time} - 0.01147 \cdot \text{time}^2$$

(E7)

with $\text{time}$ time of day [h; decimal fraction]

During daylight hours till 2 pm, $f_5(\text{time})$ is set unity. During night $f_5(\text{time})$ is of no practical use, but is set unity in order to avoid mathematical problems during a 24 hours model run.
• Jarvis-Stewart functions for phenology, ozone and carbon dioxide

$f_6$(PHEN) and $f_7$(O3) allow for the modifying influence of phenology and ozone on stomatal resistance: both senescence due to normal aging and premature senescence induced by ozone limit stomatal aperture.

With $f_8$(CO2) the influence of elevated CO2 on stomatal aperture is multiplicatively taken into account (cf. Oltchev et al., 1998):

\[
f_8(\text{CO}_2) = \left( \frac{c_1 \cdot \chi_{\text{CO}_2}}{c_2 + \chi_{\text{CO}_2}} \right) \cdot \left( \frac{\chi_{\text{CO}_2, \text{base}}}{\chi_{\text{CO}_2}} \right)
\]

(E8)

with \( \chi_{\text{CO}_2} \) actual mole fraction of CO2 at reference height [ppm]  
\( \chi_{\text{CO}_2, \text{base}} \) base CO2 mole fraction (default value: \( \chi_{\text{CO}_2, \text{base}} = 350 \) ppm)  
\( c_1, c_2 \) empirical constants describing dependence of photosynthesis rate on ambient CO2 concentration (\( c_1 \): dimensionless, \( c_2 \) in ppm)

$f_6$(PHEN) and $f_8$(CO2) must be derived for the respective vegetation by comparison of modelled and measured latent heat fluxes (default value: $f_6$(PHEN) = 1) and CO2 fluxes respectively (default value: $f_8$(CO2) = 1). $f_7$(O3) depends on stomatal uptake of O3 and must be derived experimentally (default value: $f_7$(O3) = 1).
Appendix F

Equations for water vapour pressure deficit of the atmosphere, slope of the water vapour pressure saturation curve and density and specific heat of moist air

- **water vapour pressure deficit of the atmosphere [hPa]**

\[ VPD = e_{\text{saturation water vapour pressure}} - e_{\text{water vapour pressure}} \]  \hspace{1cm} (F1)

with the saturation water vapour pressure of the atmosphere [hPa; after Magnus]:

\[ e_{\text{saturation water vapour pressure}} = 6.1078 \cdot e^{\frac{17.0805 \cdot t_a}{234.175 + t_a}} \quad \text{if } t_a \geq 0°C \]  \hspace{1cm} (F2)

\[ e_{\text{saturation water vapour pressure}} = 6.1078 \cdot e^{\frac{22.44294 \cdot t_a}{272.44 + t_a}} \quad \text{if } t_a < 0°C \]  \hspace{1cm} (F3)

and the water vapour pressure [hPa]:

\[ e_{\text{water vapour pressure}} = e_{\text{saturation water vapour pressure}} \cdot \frac{rH}{100} \]  \hspace{1cm} (F4)

where \( rH [\%] \) is the relative humidity measured at air temperature reference height \( z_{\text{ref}, T} \)

- **slope of the water vapour saturation pressure curve [hPa K\(^{-1}\)]**

\[ \frac{de_{\text{saturation water vapour pressure}}}{dt_a} = e_{\text{saturation water vapour pressure}} \cdot \frac{17.0805 \cdot 234.175}{(234.175 + t_a)^2} \quad \text{if } t_a \geq 0°C \]  \hspace{1cm} (F5)

\[ \frac{de_{\text{saturation water vapour pressure}}}{dt_a} = e_{\text{saturation water vapour pressure}} \cdot \frac{22.44294 \cdot 272.44}{(272.44 + t_a)^2} \quad \text{if } t_a < 0°C \]  \hspace{1cm} (F6)

Note that, under certain circumstances, (F5) or (F6) may serve as an approximation of \( s_c \) in the Penman-Monteith equation mentioned in Chapter 2.4.

- **density of moist air at absolute temperature \( T \) [kg m\(^{-3}\)]**

\[ \rho_{\text{moist air}} = \rho_{\text{dry air}} \cdot \left( 1 - 0.378 \cdot \frac{e_{\text{water vapor pressure}}}{p} \right) \]  \hspace{1cm} (F7)

with \[ \rho_{\text{dry air}} = \frac{p}{R_{\text{dry air}} \cdot (273.15 + t_a)} \cdot 100 \]  \hspace{1cm} (F8)

and \[ p \] air pressure [hPa]

\[ R_{\text{dry air}} \] gas constant for dry air \( (R_{\text{dry air}} = 287.04 \text{ J kg}^{-1} \cdot \text{K}^{-1}) \)

- **specific heat of moist air at constant pressure [m\(^2\) s\(^{-2}\) K\(^{-1}\)]**

\[ c_{p, \text{moist air}} = c_{p, \text{dry air}} \cdot (1 + 0.84 \cdot q) \]  \hspace{1cm} (F9)

with \[ c_{p, \text{dry air}} = 1004.67 \text{ m}^2 \cdot \text{s}^{-2} \cdot \text{K}^{-1} \]

and \[ q \] specific air humidity [g g\(^{-1}\)]

\[ q = \frac{0.622 \cdot e_{\text{water vapor pressure}}}{p - 0.378 \cdot e_{\text{water vapor pressure}}} \]  \hspace{1cm} (F10)
Appendix G

Biosphere/atmosphere exchange of carbon dioxide

The exchange of carbon dioxide between the reference height above the canopy and the phyto-sphere \( F_c(\mathrm{CO}_2) \, [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \) arises as the effect of canopy photosynthesis \( A_{\text{net}} \) (positive towards the canopy, \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \) and ecosystem respiration \( \text{Resp}^* \) (positive upwards, \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)).

G1 - Canopy photosynthesis

The net rate of canopy photosynthesis \( A_{\text{net}} \) is parameterized according to de Pury and Farquhar (1997). The so-called single layered sun/shade model described by the authors is a scaled version of a leaf photosynthesis model (cf. e.g. Farquhar et al., 1980; von Caemmerer 2000) integrating separately the sunlit and shaded leaf fractions of the canopy:

\[
A_{\text{net}} = A_{\text{net, sunlit}} + A_{\text{net, shaded}} = (A_{\text{sunlit}} + A_{\text{shaded}}) - (\text{Resp}_{\text{d, sunlit}} + \text{Resp}_{\text{d, shaded}}) \tag{G1}
\]

with \( A_{\text{net, x}} \) net rate of photosynthesis of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)

\( \text{Resp}_{\text{d, x}} \) mitochondrial respiration rate (leaf dark respiration) of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)

*Parameterisation of net rate of photosynthesis of sunlit and shaded leaf fraction of the canopy*

The net rate of photosynthesis of the sunlit and shaded leaf fraction of the canopy, \( A_{\text{net, sunlit}} \) and \( A_{\text{net, shaded}} \), is calculated based on a biochemical model of photosynthesis according to Farquhar et al. (1980):

\[
A_{\text{net, x}} = \min\{A_{c, x}, A_{j, x}, A_{p, x}\} - \text{Resp}_{\text{d, x}} \tag{G2}
\]

with \( A_{\text{net, x}} \) net rate of photosynthesis of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)

\( A_{c, x} \) Rubisco-limited rate of CO\(_2\) assimilation of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)

\( A_{j, x} \) electron-transport limited rate of CO\(_2\) assimilation of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \), i.e. rate of photosynthesis limited by Rubisco regeneration

\( A_{p, x} \) triose phosphate utilization-limited rate of CO\(_2\) assimilation of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)

\( \text{Resp}_{\text{d, x}} \) daytime mitochondrial respiration rate of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)

(default value see Table G1)

The Rubisco-limited CO\(_2\) assimilation rate of sunlit \( A_{c, \text{sunlit}} \) or shaded leaf fractions \( A_{c, \text{shaded}} \) of the canopy is:

\[
A_{c, x} = V_{c, x} \cdot \frac{p_{i, x} - \Gamma_*}{p_{i, x} + K_c \cdot (1 + \frac{O_{\text{i}}}{K_o})} \tag{G3}
\]

with \( V_{c, x} \) photosynthetic Rubisco capacity of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy \( [\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \)

\( p_{i, x} \) intercellular CO\(_2\) partial pressure of the sunlit (\( x = \text{sunlit} \)) or shaded (\( x = \text{shaded} \)) leaf fraction of the canopy [Pa]
\( \Gamma^* \) CO₂ compensation point of photosynthesis in the absence of mitochondrial respiration [Pa]
(default value see Table G1)

\( K_c \) Michaelis-Menten constant of Rubisco for CO₂ [Pa]
(default value see Table G1)

\( K_o \) Michaelis-Menten constant of Rubisco for O₂ [Pa]
(default value see Table G1)

\( O_i \) intercellular O₂ partial pressure (= 20.5 \( \times \) 10³ Pa)

### Table G1

Photosynthetic parameters for wheat at 25°C according to de Pury and Farquhar (1997)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Unit</th>
<th>activation energy ( E_a ) [J mol⁻¹]</th>
</tr>
</thead>
<tbody>
<tr>
<td>( K_c )</td>
<td>40.4</td>
<td>Pa</td>
<td>59400</td>
</tr>
<tr>
<td>( K_o )</td>
<td>( 24.8 \times 10^3 )</td>
<td>Pa</td>
<td>36000</td>
</tr>
<tr>
<td>( \Gamma^* )</td>
<td>3.69</td>
<td>Pa</td>
<td>29000</td>
</tr>
<tr>
<td>( k_n )</td>
<td>0.713</td>
<td>---</td>
<td></td>
</tr>
<tr>
<td>( V_{\text{leaf at top of canopy}} )</td>
<td>110</td>
<td>( \mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} )</td>
<td>64800</td>
</tr>
<tr>
<td>( \text{Resp}_{\text{d}} )</td>
<td>0.0089 ( V_c )</td>
<td>( \mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} )</td>
<td>66400</td>
</tr>
<tr>
<td>( J_{\text{max}} )</td>
<td>2.1 ( V_c )</td>
<td>( \mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} )</td>
<td>37000</td>
</tr>
</tbody>
</table>

1) photosynthetic Rubisco capacity per unit leaf area at top of canopy

2) \( \text{Resp}_{\text{d}} \): 0.01 \( V_c \) - 0.02 \( V_c \) (cf. von Caemmerer, 2000)

\( J_{\text{max}} \): 1.5 \( V_c \) - 2 \( V_c \) (cf. von Caemmerer, 2000)

The photosynthetic capacities of the sunlit and shaded leaf fractions of the canopy \( V_{\text{c, sunlit}} \) and \( V_{\text{c, shaded}} \) are calculated by integrating the leaf photosynthetic capacity \( V_{\text{leaf, x}} \) and the respective leaf area fraction taking into account the leaf nitrogen content (cf. de Pury and Farquhar, 1997).

\( V_{\text{c, sunlit}} \) is approximated as given in eq. (G4) and \( V_{\text{c, shaded}} \) as given in eq. (G5).

\[
V_{\text{c, sunlit}} = LAI_{\text{non-senescent}} \cdot V_{\text{leaf at top of canopy}} \cdot \frac{1 - \exp[-k_n - (k_b \cdot LAI_{\text{non-senescent}})]}{k_n + (k_b \cdot LAI_{\text{non-senescent}})} \quad (G4)
\]

with

- \( V_{\text{leaf at top of canopy}} \) photosynthetic Rubisco capacity per unit leaf area at top of canopy \([\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}]\); (default value: see Table G1)
- \( k_n \) coefficient of leaf-nitrogen allocation in a canopy (default value: see Table G1)

\[
V_{\text{c, shaded}} = V_c - V_{\text{c, sunlit}} \quad (G5)
\]

with

- \( V_c \) canopy photosynthetic capacity \([\mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}]\)

\[
V_c = LAI_{\text{non-senescent}} \cdot V_{\text{leaf at top of canopy}} \cdot \frac{1 - \exp[-k_n]}{k_n} \quad (G6)
\]
The Rubisco regeneration-limited or electron-transport-limited rate of the sunlit, $A_{j,\text{sunlit}}$, or shaded, $A_{j,\text{shaded}}$, leaf fraction of the canopy is:

$$A_{j,x} = J_x \cdot \frac{p_{i,x} - \Gamma_x}{4 \cdot p_{i,x} + 8 \cdot \Gamma_x}$$  \hspace{1cm} (G7)

with $J_x$ electron transport rate per unit leaf area of the sunlit ($x = \text{sunlit}$) or shaded ($x = \text{shaded}$) leaf fraction of the canopy [$\mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$]

$$J_x = \frac{I_{2,x} + J_{\text{max},x} - \sqrt{(I_{2,x} + J_{\text{max},x})^2 - 4 \cdot \theta \cdot I_{2,x} \cdot J_{\text{max},x}}}{2 \cdot \theta}$$  \hspace{1cm} (G8)

with $J_{\text{max},x}$ light-saturated (maximum) electron transport rate per unit leaf area of the sunlit ($x = \text{sunlit}$) or shaded ($x = \text{shaded}$) leaf fraction of the canopy [$\mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$]

and

$$I_{2,x} = I_{c,x} \cdot \frac{1 - f}{2}$$  \hspace{1cm} (G9)

with $I_{2,x}$ photosynthetically active irradiation absorbed by PS II of the sunlit ($x = \text{sunlit}$) or shaded ($x = \text{shaded}$) leaf fraction of the canopy [$\mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$]

$I_{c,x}$ photosynthetic active radiation absorbed by the sunlit ($x = \text{sunlit}$) or shaded ($x = \text{shaded}$) leaf fraction of the canopy [$\mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$]

$f$ corrects for spectral quality of light (default value: $f \sim 0.15$)

The export-limited or phosphate-limited canopy CO$_2$ assimilation rate $A_{p,x}$ of the sunlit ($x = \text{sunlit}$) or shaded ($x = \text{shaded}$) leaf fraction of the canopy is:

$$A_{p,x} = (3 \cdot T_p) \cdot \frac{p_{i,x} - \Gamma_x}{p_{i,x} - \left(1 + \frac{3 \alpha_c}{2}\right) \cdot \Gamma_x}$$  \hspace{1cm} (G10)

according to von Caemmerer (2000)

with $T_p$ rate of triose phosphate export from the chloroplast [$\mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$]

$\alpha_c$ fraction of glycolate carbon not returned to the chloroplast ($0 < \alpha_c < 1$)

or

$$A_{p,x} = \frac{V_{c,x}}{2}$$  \hspace{1cm} (G11)

according to Collatz et al. (1991)

IN PLATIN, eq. (G11) is implemented as a first guess. The phosphate limitation is relevant under conditions where plants are grown under elevated ambient CO$_2$ concentrations.

De Pury and Farquhar (1997) scaled daytime mitochondrial respiration rate (leaf dark respiration) $\text{Resp}_{d,x}$ of the sunlit ($x = \text{sunlit}$) or shaded ($x = \text{shaded}$) leaf fraction of the canopy to $V_{c,x}$ as:

$$\text{Resp}_{d,x} = 0.0089 \cdot V_{c,x}$$  \hspace{1cm} (G12)

Additionally, in PLATIN daytime canopy leaf respiration rate $\text{Resp}_{d,c}$ is parameterized by:
\[ \text{Resp}_{d,c} = \text{LAI}_{\text{non-senecent}} \cdot \text{Resp}_{d,\text{ref}} \cdot \sqrt{\frac{273 + t_s}{298}} \quad (G13) \]

with \( \text{Resp}_{d,\text{ref}} \) reference canopy leaf respiration rate (\( \mu \text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1} \))
\( t_s \) leaf temperature [°C]

\( \text{Resp}_{d,\text{ref}} \) is adjusted by multiple regression analysis to the sum of \( \text{Resp}_{d,\text{sunlit}} \) and \( \text{Resp}_{d,\text{shaded}} \). Taking into account the temperature dependency as described by eq. (G15), nighttime canopy mitochondrial respiration rate \( \text{Resp}_{c,\text{nighttime}} \) is estimated.

The temperature dependence of \( \Gamma^* \) is described by:

\[ \Gamma^* = \Gamma^*_0(25°C) + 0.188 \cdot (t_s - 25) + 0.0036 \cdot (t_s - 25)^2 \quad (G14) \]

The temperature dependence of \( \text{Resp}_{d}, \text{Resp}_{n} \) and the kinetic constants \( k_c, k_o \) and \( V_{c,x} \) is described by an Arrhenius function of the form:

\[ \text{Parameter}(t_s) = \text{Parameter}(25°C) \cdot \exp \left( \frac{(t_s - 25) \cdot E_a}{298 \cdot R \cdot (273 + t_s)} \right) \quad (G15) \]

with \( t_s \) leaf temperature [°C]
\( R \) universal gas constant (\( R = 8.314 \text{ J K}^{-1} \cdot \text{mol}^{-1} \))
\( E_a \) activation energy [J mol\(^{-1}\)], listed in Table G1

The temperature dependency of \( J_{\text{max}} \) is taken into account by the function:

\[ J_{\text{max}}(t_s) = J_{\text{max}}(25°C) \cdot \frac{((273 + t_s) - 298) \cdot E_a}{298 \cdot R \cdot (273 + t_s)} \cdot \frac{1 + \exp \left( \frac{298 \cdot S_J - H_J}{298 \cdot R} \right)}{1 + \exp \left( \frac{(273 + t_s) \cdot S_J - H_J}{R \cdot (273 + t_s)} \right)} \quad (G16) \]

with \( t_s \) leaf temperature [°C]; \( T = 273 + t_s \) [K]
\( R \) universal gas constant (= 8.314 J K\(^{-1}\)·mol\(^{-1}\))
\( S_J \) 710 J K\(^{-1}\)·mol\(^{-1}\)
\( H_J \) 220000 J mol\(^{-1}\)

**G2 - Approximation of belowground respiration and respiration rate of aboveground woody plant parts**

An estimation of belowground respiration \( \text{Resp}_{\text{belowground}} \) as well as of the respiration rate of the aboveground woody plant parts \( \text{Resp}_{w} \) are necessary for the computation of canopy net ecosystem exchange, with:

\[ \text{Resp}^* = \text{Resp}_{w} + \text{Resp}_{\text{belowground}} \quad (G17) \]

• **belowground respiration**

In PLATIN, the simple bulk parameterization approach to predict soil CO₂ efflux \( \text{Resp}_{\text{belowground}} \) described by Nikolov and Zeller (2003) is implemented. The approach considers the effect of soil moisture at 14 cm depth (\( SM_{14\text{cm}} \)), soil temperature at 10 cm depth \( t_{\text{soil,10cm}} \) and clay content (\( Cl \)) on the net CO₂ release only. The CO₂ evolution from the soil is estimated as:
\[
\text{Resp}_{\text{belowground}} = \frac{f_T + f_M - \sqrt{(f_T + f_M)^2 - 4 \cdot \omega \cdot f_T \cdot f_M}}{2 \omega} \cdot (1 - 0.0065 \cdot CI)
\]

with \( \text{Resp}_{\text{belowground, max}} \) rate of soil respiration under optimum temperature and moisture conditions [\( \mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1} \)]

\( f_T \) non-dimensional factor quantifying the limitations of soil temperature on CO\(_2\) efflux

\( f_M \) non-dimensional factor quantifying the limitations of soil moisture on CO\(_2\) efflux

\( \omega \) convexity coefficient defining the smoothness of the transition between \( f_T \) and \( f_M \) (\( \omega = 0.985 \))

\( CI \) clay content [%]

A \( \text{Resp}_{\text{belowground, max}} \) of 17 \( \mu\text{mol}\cdot\text{m}^{-2}\cdot\text{s}^{-1} \), derived from respiration rates from hardwood forests around the world, is suggested in Nikolov and Zeller (2003). The temperature factor is computed after Kirschbaum (1995):

\[
f_T = \exp\left[ -3.764 + 0.204 \cdot t_{\text{soil,10cm}} \cdot \left( 1 - 0.5 \cdot \frac{t_{\text{soil,10cm}}}{t_{\text{opt}}} \right) \right]
\]

with \( t_{\text{opt}} \) optimum temperature for CO\(_2\) evolution (default value: \( t_{\text{opt}} = 36.9 \, ^\circ\text{C} \))

Eq. (G19) is based on soil respiration data from 11 studies worldwide (cf. Kirschbaum, 1995) and implies a variable \( Q_{10} \). The effect of actual volumetric moisture on soil respiration is based on measurements and models by Schlentner and van Cleve (1984) and Grant and Rochette (1994):

\[
f_M = \frac{1}{1 + \exp[3 - 14 \cdot SM^*]} \quad \text{if} \quad 0 \leq SM^* \leq 0.74
\]

\[
f_M = \frac{1}{1 + \exp[3 - 76 \cdot (1 - SM^*)]} \quad \text{if} \quad 0.74 < SM^* \leq 1.0
\]

where \( SM^* \) is the relative moisture saturation at 14 cm depth:

\[
SM^* = \frac{SM - SM_{\text{res}}}{SM_{\text{sat}} - SM_{\text{res}}}
\]

with \( SM \) actual soil water content [\( \text{m}^3\cdot\text{m}^{-3} \)]

\( SM_{\text{res}} \) residual (irreducible) soil moisture [\( \text{m}^3\cdot\text{m}^{-3} \)]

\( SM_{\text{sat}} \) soil moisture content at saturation [\( \text{m}^3\cdot\text{m}^{-3} \)]

Nocturnal eddy covariance measurements provide information on the respiratory release of above-ground and belowground processes. Plant respiration \( \text{Resp}_{c, \text{nighttime}} \) is substracted from the nocturnal atmospheric CO\(_2\) flux to obtain the nocturnal belowground CO\(_2\) flux.

For crops and grassland systems \( \text{Resp}^* \) equals \( \text{Resp}_{\text{belowground}} \).

**aboveground woody respiration**

For ecosystems other than crops and grassland, aboveground woody respiration \( \text{Resp}_w \) is parameterized according to Nikolov and Zeller (2003). As described by the author’s, this total respiratory flux is partitioned into a maintenance \( \text{Resp}_{w, m} \) and growth \( \text{Resp}_{w, g} \) component.
Aboveground woody maintenance respiration $\text{Resp}_{w,m}$ is assumed to be linearly related to the nitrogen content of living tissue and varies exponentially with air temperature:

$$\text{Resp}_{w,m} = 0.2103 \cdot N_s \cdot \frac{t_{\text{lag3h}} - 20}{10}$$  \hspace{1cm} (G23)

with $\text{Resp}_{w,m}$ aboveground woody maintenance respiration [µmol·m$^{-2}$·s$^{-1}$]

$N_s$ nitrogen content of the aboveground sapwood biomass [g·m$^{-2}$]

$t_{\text{lag3h}}$ relative change of the respiration rate per 10 K temperature increase

As reported in the description of the FORFLUX model (Nikolov and Zeller, 2003) $N_s$ typically ranges between 1 and 13 g·m$^{-2}$. $Q_{10}$ is approximated as follows, taking into account an acclimation response:

$$Q_{10} = 3.25 - 0.077 \cdot t_{\text{days}} \quad \text{if} \quad t_{\text{days}} \geq 0 \degree C$$  \hspace{1cm} (G24)

$$Q_{10} = 3.25 \quad \text{if} \quad t_{\text{days}} < 0 \degree C$$  \hspace{1cm} (G25)

with $t_{\text{days}}$ mean daily air temperature of the past 7 days [°C]

Both, $t_{\text{days}}$ and $Q_{10}$ must be updated daily in the model.

The respiration rate from aboveground woody biomass due to construction of new tissue $\text{Resp}_{w,g}$ is assumed to be constant during a day and proportional to the net carbon gain of vegetation in the previous day (cf. Nikolov and Zeller, 2003):

$$\text{Resp}_{w,g} = f_g \cdot \delta_a \cdot \frac{A'_{\text{net}} - \text{Resp}'_{w,m} - \text{Resp}'_{r,m}}{86.4}$$  \hspace{1cm} (G26)

with $\text{Resp}_{w,g}$ aboveground woody maintenance respiration due to construction of new tissue [µmol·m$^{-2}$·s$^{-1}$]

$f_g$ construction cost ($f_g = 0.2$ mol·mol$^{-1}$)

$\delta_a$ carbon allocation coefficient

$A'_{\text{net}}$ 24-h integrated canopy net photosynthesis of the previous day [mmol·m$^{-2}$·d$^{-1}$]

$\text{Resp}'_{w,m}$ cumulative daily maintenance respiratory flux from aboveground woody biomass [mmol·m$^{-2}$·d$^{-1}$]

$\text{Resp}'_{r,m}$ cumulative daily maintenance respiratory flux from roots [mmol·m$^{-2}$·d$^{-1}$]

The factor 86.4 converts flux units mmol·m$^{-2}$·d$^{-1}$ to µmol·m$^{-2}$·s$^{-1}$. As summarized by Nikolov and Zeller (2003) the carbon allocation coefficient $\delta_a$ depends on species physiology, tree age and growth conditions. Typically it varies between 0.34 and 0.8 (cf. Cannell, 1985). Root maintenance respiration $\text{Resp}'_{r,m}$ is estimated from the daily integral of root total respiration $\text{Resp}'_{r,t}$ assuming roots have the same construction cost as the aboveground biomass:

$$\text{Resp}'_{r,m} = f_g \cdot \delta_a \cdot \frac{\text{Resp}'_{r,t} - f_g \cdot (1-\delta_a) \cdot (A'_{\text{net}} - \text{Resp}'_{w,m})}{1 - f_g \cdot (1-\delta_a)}$$  \hspace{1cm} (G27)

where $\text{Resp}'_{r,t}$ must be computed as a site-specific fraction of the total CO$_2$ efflux from soil $\text{Resp}_{\text{below-ground}}$ (cf. Bowden et al., 1992; Nadelhoffer and Raich, 1992).

The total aboveground woody respiration $\text{Resp}_w$ is:

$$\text{Resp}_w = \text{Resp}_{w,m} + \text{Resp}_{w,g}$$  \hspace{1cm} (G28)
**G3 - Coupling of biochemical and diffusion equations for biosphere/atmosphere CO2 exchange**

The net rate of canopy photosynthesis $A_{net}$ is related to the net ecosystem CO2 exchange, i.e. the vertical above-canopy CO2 flux $F_{c}(CO_{2})$ [µmol·m$^{-2}$·s$^{-1}$] by:

$$F_{c}(CO_{2}) = - \frac{c_{CO_{2}}(z_{ref}) - c_{CO_{2}}(d + z_{0CO_{2}})}{R_{ah} + R_{b,CO_{2}}}$$

$$= -(A_{net} - Resp_{w} - Resp_{belowground})$$

(G29)

$$= -(A_{net} + Resp^*)$$

with

- $c_{CO_{2}}(z_{ref})$ concentration (amount concentration) of CO2 at reference height [µmol·m$^{-3}$]
- $c_{CO_{2}}(d + z_{0CO_{2}})$ CO2 concentration at conceptual height $z = d + z_{0CO_{2}}$ [µmol·m$^{-3}$]
- $R_{ah}$ atmospheric resistance [s·m$^{-1}$]; c.f. chapter 2.1
- $R_{b,CO_{2}}$ quasi-laminar layer resistance for CO2 [s·m$^{-1}$]; c.f. chapter 2.2
- $Resp_{w}$ respiration rate of aboveground woody plant parts [µmol·m$^{-2}$·s$^{-1}$]
- $Resp_{belowground}$ belowground respiration rate (roots, microbes) [µmol·m$^{-2}$·s$^{-1}$]

In order to obtain $A_{net}$ as function of the reference-height concentration $c_{CO_{2}}(z_{ref})$ some additional considerations and approximations are needed. First of all, $A_{net}$ is rewritten as an analogue to Ohm’s law:

$$A_{net} = A_{net, sunlit} + A_{net, shaded}$$

$$= \frac{c_{CO_{2}}(d + z_{0CO_{2}}) - c_{i, sunlit}}{R_{c, stom, sunlit, CO_{2}}} + \frac{c_{CO_{2}}(d + z_{0CO_{2}}) - c_{i, shaded}}{R_{c, stom, shaded, CO_{2}}}$$

(G31)

with

- $c_{i, x}$ intercellular CO2 concentration of the sunlit or shaded leaf fraction of the canopy [µmol·m$^{-3}$]

The bulk stomatal resistance for CO2 exchange via stomata of the sunlit and shaded leaf fraction at a specific canopy development stage $R_{c, stom, CO_{2}}^{*}$ [s·m$^{-1}$] is given by

$$R_{c, stom, sunlit, CO_{2}}^{*} = \frac{R_{c, stom, H_{2}O}}{(1 - \beta \text{ sunlit})} \cdot \frac{D_{H_{2}O}}{D_{CO_{2}}} = \frac{R_{c, stom, H_{2}O}}{(1 - \beta \text{ sunlit})} \cdot 1.6$$

(G32)

$$R_{c, stom, shaded, CO_{2}}^{*} = \frac{R_{c, stom, H_{2}O}}{(1 - \beta \text{ shaded})} \cdot \frac{D_{H_{2}O}}{D_{CO_{2}}} = \frac{R_{c, stom, H_{2}O}}{(1 - \beta \text{ shaded})} \cdot 1.6$$

(G33)

taking into account the differences between the molecular diffusivity for water vapour $D_{H_{2}O}$ and molecular diffusivity of the CO2 in air (cf. Table 2) and the actual canopy development stage (cf. chapter 2.3).

Algebraic rearrangements of eqs. (G31) with (G29) yield an expression for $c_{i, sunlit}$ and $c_{i, sunlit}$:

$$c_{i, sunlit} = c_{CO_{2}}(d + z_{0CO_{2}}) - (A_{net, sunlit} \cdot R_{c, stom, sunlit, CO_{2}}^{*})$$

(G34)

$$= c_{CO_{2}}(z_{ref}) - [(A_{net} - Resp^*) \cdot (R_{ah} + R_{b,CO_{2}})] - (A_{net, sunlit} \cdot R_{c, stom, sunlit, CO_{2}}^{*})$$

$$c_{i, shaded} = c_{CO_{2}}(d + z_{0CO_{2}}) - (A_{net, shaded} \cdot R_{c, stom, shaded, CO_{2}}^{*})$$

(G35)

$$= c_{CO_{2}}(z_{ref}) - [(A_{net} - Resp^*) \cdot (R_{ah} + R_{b,CO_{2}})] - (A_{net, shaded} \cdot R_{c, stom, shaded, CO_{2}}^{*})$$
Conversion of CO₂ concentration [µmol m⁻³] in partial pressure [Pa] yields for \( \rho_i \) (cf. Appendix M):

\[
P_{i,\text{sunlit}} = (\phi_{\text{cf}} \cdot p) \cdot \left[ C_{\text{CO}_2}(d + z_{\text{CO}_2}) - (A_{\text{net,\ sunlit}} \cdot R^*_c,\ stom,\ sunlit,\ CO_2) \right]
\]
\[
P_{i,\text{shaded}} = (\phi_{\text{cf}} \cdot p) \cdot \left[ C_{\text{CO}_2}(z_{\text{ref}}) - [(A_{\text{net}} - R_{\text{Resp}} \cdot (R_{\text{ah}} + R_{b,\text{CO}_2})) - (A_{\text{net,\ sunlit}} \cdot R^*_c,\ stom,\ sunlit,\ CO_2) \right]
\]

\[\phi_{\text{cf}}\] atmospheric pressure [Pa]

\[\phi_{\text{cf}} = 22.4 \cdot 10^{-9} m^3 \cdot \mu mol^{-1}\]

The system of eqs. (G1), (G2), (G29), (G31) (G36) and (G37) are solved iteratively. The initial value is chosen as \( \rho_i = 0.7 \rho_{\text{CO}_2}(z_{\text{ref}}) \). This value is based on the fact that a constant ratio \( c_i/c_a \) in the order of magnitude of 0.7 for C₃ plants can be observed under non-limiting conditions even when ambient CO₂ concentration \( c_a \) is varied (cf. Ball and Berry, 1982; Drake et al., 1997; Kim and Verma, 1991; Wong et al., 1979). Only when stomata are nearly closed the ratio \( c_i/c_a \) deviates from this cardinal values tending towards unity. Note, eq. (G31) applies to steady-state conditions for \( R_{c,\ stom,\ x,\ CO_2} \) and \( A_{\text{net,\ x}} \), because responses of leaf biochemistry and stomatal resistance to changes in environmental conditions have quite different relaxation times (e.g. Leuning, 1990).

**G4 - Calculation of biosphere/atmosphere exchange of carbon dioxide**

The exchange of carbon dioxide between the reference height above the canopy and the plant/soil system \( F_{c}(\text{CO}_2) \) is calculated during times with solar elevation angle \( \phi > 0 \) as given in eq. (G38)

\[
F_c(\text{CO}_2) = -A_{\text{net}} + \text{Resp}^* \quad \text{(G38)}
\]

and during times with solar elevation angle \( \phi = 0 \) as given in eq. (G39):

\[
F_c(\text{CO}_2) = \text{Resp}_{c,\ \text{nighttime}} + \text{Resp}^* \quad \text{(G39)}
\]
Appendix H

Soil water model

PLATIN makes use of a so-called force-restore model (cf. Deardorff, 1978) to provide a measure of volumetric soil water content $SM$ [m$^3$ m$^{-3}$] which is needed to estimate stomatal resistance (cf. Chapter 2.3, Appendix E). Other than more complex modelling concepts, the force-restore approach is limited to predict the average soil water content of only one single layer the top side of which is the soil surface:

$$\frac{dSM}{dt} = \frac{1}{\rho_{H2O} \cdot SLD} \cdot \left[ - \frac{\lambda E_{SLD}}{\lambda} + \frac{W_{in}}{DTI} + C_{cap} \cdot \left( 1 - \frac{SM - SM_{W}}{SM_{C} - SM_{W}} \right) \right] \quad (H1)$$

with

- $\rho_{H2O}$ density of water [kg m$^{-3}$]
- $SLD$ depth of soil layer under consideration [m]
- $\lambda E_{SLD}$ turbulent vertical flux density of latent heat in case of non-zero evapotranspiration [W m$^{-2}$] originated from the SLD-layer
- $\lambda$ latent heat of water vaporisation [J kg$^{-1}$]
- $W_{in}$ precipitation and/or dew reaching the ground [kg m$^{-2}$ = mm H$_2$O]
- $DTI$ measurement time interval [s]
- $C_{cap}$ maximum possible vertical water flow from groundwater into rooted soil layer due to soil capillarity [kg m$^{-2}$ s$^{-1}$]
- $SM_{W}$ site-specific wilting point [m$^3$ m$^{-3}$]
- $SM_{C}$ site-specific field capacity [m$^3$ m$^{-3}$]

The value of $C_{cap}$ site-specifically depends on the vertical distance between ground water table and the soil layer under consideration and can reach up to 5 mm d$^{-1}$ (Scheffer and Schachtschabel, 1982) which is about 5.8 $\times$ 10$^{-5}$ mm s$^{-1}$.

Equations like (H1) are often used with $\lambda E_{SLD} = \lambda E$ and accordingly $SLD = RLD$ where $RLD$ is a measure of the depth of the rooted soil layer. The deeper the layer the lower is the amplitude of the diurnal course of soil water content calculated by eq. (H1). This dampening is an undesired result with respect to the influence of soil water content on the modelling of the diurnal variation of stomatal resistance. Therefore another way has been chosen in PLATIN: the soil water content is calculated for a given soil layer depth $SLD$ in the order of 0.1 m, because, at least for agricultural crops, the water content in such a thin layer (along with its pronounced diurnal variation) seems to be appropriate for stomatal resistance modelling. In addition, a special parameterization of $\lambda E_{SLD}$ has been developed$^2$ in order to account for the fact that, especially in case of high evaporative demand, transpiration should be fed not only by the roots in the thin subsurface soil layer but also by roots located below.

The basic idea of this parameterization is that plants tend to minimize the amount of energy needed to supply water for transpiration. The respective modelling concept in PLATIN is based on the assumption that the vertical distribution of soil water content does not influence root water uptake and that roots of deeper soil layers contribute to transpiration the lesser the lesser transpiration rate is itself, and vice versa.

To develop the approach, evapotranspiration $\lambda E$ is separated into bare soil evaporation $\lambda E_{evaporation}$ (the water for which is assumed to be always completely drawn from the SLD-layer) and canopy transpiration $\lambda E_{transpiration}$

$$\lambda E = \lambda E_{evaporation} + \lambda E_{transpiration} \quad (H2)$$

where $\lambda E_{evaporation}$ and $\lambda E_{transpiration}$ are given in Chapter 2.4, eqs. (26) and (27).

$^2$ This parameterization was presented already in Grünhage and Haenel (1997), where some equations had been slightly mistyped.
The fraction of $\lambda E_{\text{transpiration}}$ which is originating from the SLD-layer is called $\lambda E_{\text{transpiration, SLD}}$. If the density distribution of the transpiration-relevant roots is assumed to decay exponentially with soil depth and if the water uptake by the roots is directly proportional to the local root density, the ratio of $\lambda E_{\text{transpiration, SLD}}$ to $\lambda E_{\text{transpiration}}$ is given by

$$\frac{\lambda E_{\text{transpiration, SLD}}}{\lambda E_{\text{transpiration}}} = 1 - \exp\left(-2.303 \cdot \frac{SLD}{RLD_{90}}\right)$$

(H3)

where $RLD_{90}$ is the depth of the soil layer containing 90\% of total root mass. The vertical distribution of water uptake according to an exponential expression was used for example also by Lindström and Garde-lin (1992).

Eq. (H3) does not yet account for the desired influence of evaporative demand on root water uptake. To meet with this requirement, the constant $RLD_{90}$ is replaced by a variable layer depth $VLD$. As a first order guess, $VLD$ is assumed to be roughly proportional to the transpiration rate. However, $VLD$ should never tend to zero even when the transpiration rate does so, because the depth from which water is extracted by roots will never decline to zero even in case of very low or zero values of $\lambda E$. In order to keep the resulting ratio of $\lambda E_{\text{transpiration, SLD}}$ to $\lambda E_{\text{transpiration}}$ as simple as possible with respect to its mathematical formulation (see eq. (H5)), the following approach is chosen for $VLD$:

$$\frac{VLD}{VLD_{\text{max}}} = \left[1 - \ln\left(\frac{\lambda E}{\lambda E_{\text{max}}}\right)^{-1}\right]$$

(H4)

where $VLD_{\text{max}} = VLD(\lambda E = \lambda E_{\text{max}})$ corresponds to $RLD_{90}$ and $\lambda E_{\text{max}}$ will be specified below. Figure H1 shows $VLD \cdot VLD_{\text{max}}^{-1}$ as function of $\lambda E \cdot \lambda E_{\text{max}}^{-1}$.

Fig. H1
$VLD \cdot VLD_{\text{max}}^{-1}$ as function of $\lambda E \cdot \lambda E_{\text{max}}^{-1}$ (solid line) and the 1:1 reference line (dotted line)

The obvious deficiency of eq. (H5) in case of $\lambda E = 0$ is cured when combining eqs. (H3) and (H4) and replacing $RLD_{90}$ with $VLD$ in order to obtain a modified equation for the ratio of $\lambda E_{\text{transpiration, SLD}}$ to $\lambda E_{\text{transpiration}}$.
\[ \frac{\lambda E_{\text{transpiration,SLD}}}{\lambda E_{\text{transpiration}}} = 1 - \left( \frac{\lambda E}{\lambda E_{\text{ref}}} \right)^{2.303} \cdot \frac{\text{SLD}}{\text{VLD}_{\text{max}}} \]  

(H5)

where \( \lambda E_{\text{ref}} = e \cdot \lambda E_{\text{max}} \) may roughly be estimated as \( 10^3 \text{ W m}^{-2} \) implying \( \lambda E_{\text{max}} \) to be about \( 370 \text{ W m}^{-2} \). The latter seems to be a reasonable value under central European climate conditions. This can be concluded e.g. from the approach of Zhang and Lemeur (1995) who approximate the daily average of \( \lambda E \), \( \lambda E_{\text{day}} \), by

\[ \lambda E_{\text{day}} = \left( \frac{2}{\pi} \right) \cdot \lambda E_{\text{max}} \cdot \left( \frac{n_h}{24} \right) \]

where \( n_h \) is the number of daylight hours. Then for a clear day in June in the mid-latitudes with \( n_h = 16 \), \( \lambda E_{\text{max}} = 370 \text{ W m}^{-2} \) corresponds to \( \lambda E_{\text{day}} = 157 \text{ W m}^{-2} \) or 5.5 mm H2O, which is only slightly less than typical observed maximum values of about 6 mm (Schrödter, 1985). Figure H2 shows some results of eq. (H5) for \( \text{SLD} = 0.1 \text{ m} \) and different values of \( \text{VLD}_{\text{max}} \).

Finally, the entity \( \lambda E_{\text{SLD}} \) required to apply equation (H1) is obtained from:

\[ \lambda E_{\text{SLD}} = \lambda E_{\text{soil}} + \lambda E_{\text{transpiration,SLD}} \]  

(H6)

Application of the soil water model presented above requires proper selection of the site-specific parameters \( \text{SLD}, \text{VLD}_{\text{max}} = \text{RLD}_{90}, \) and \( \text{C}_{\text{cap}} \). In context with model calibration it may also be necessary to adjust the value of \( \lambda E_{\text{ref}} \) (or the fraction \( \lambda E_{\text{max}} = \lambda E_{\text{ref}} \cdot e^{-1} \)).

![Graph showing results of eq. (H6) for SLD = 0.1 m and different values of VLD_{max}.](image)
Appendix I

Estimation of displacement height and momentum roughness length from measured data

Friction velocity $u_*$ from eddy covariance measurements at height $z$, above ground, horizontal wind velocity $u$ measured at two heights $z_1$ and $z_2$ above ground (with $z_1 < z < z_2$), and Monin-Obukhov length $L$ based on eddy covariance data (for definition of $L$ see eq. (5)), can be used to estimate displacement height $d$ and momentum roughness length $z_0m$. All measurements are expected as half-hourly or hourly samplings. We recommend to apply the procedure described below on data sets measured during well established turbulence only: unstable atmospheric stratification, global radiation $S_\text{t} > 200 \ W\cdot\text{m}^{-2}$, horizontal wind velocity $u(z_1) > 1.5 \ \text{m/s}^{-1}$.

Starting point to derive a calculation procedure is the profile equation for horizontal wind velocity according to Monin-Obukhov theory:

$$ u(z) = u_* \cdot \left\{ \frac{u(z-d)}{z_0m} \right\} - \Psi_m \left( \frac{z-d}{L} \right) + \Psi_m \left( \frac{z_0m}{L} \right) $$  \tag{11}

Eq. (11) which contains both unknowns, $d$ and $z_0m$, applies to both wind velocity measurement heights. However, the roughness length can easily be eliminated by subtracting one equation from the other:

$$ u(z_2) - u(z_1) = u_* \cdot \left\{ \frac{u(z_2-d)}{z_1-d} \right\} - \Psi_m \left( \frac{z_2-d}{L} \right) + \Psi_m \left( \frac{z_1-d}{L} \right) $$  \tag{12}

As eq. (12) does not allow a closed mathematical solution for $d$, iteration is required. An iteration rule is obtained by simple re-arrangement of eq. (12):

$$ d_{i+1} = z_1 - \frac{z_2 - d_i}{\exp \left\{ \frac{\kappa \cdot [u(z_2) - u(z_1)]}{u_*} \right\} + \Psi_m \left( \frac{z_2-d_i}{L} \right) - \Psi_m \left( \frac{z_1-d_i}{L} \right) } $$  \tag{13}

Neglecting the atmospheric stability functions $\Psi_m$ in eq. (13) and setting $d_{i+1} = d_i = d_0$ delivers an equation for the initial value of $d$:

$$ d_0 = \frac{z_2 - z_1 \cdot \exp \left\{ \frac{\kappa \cdot [u(z_2) - u(z_1)]}{u_*} \right\}}{1 - \exp \left\{ \frac{\kappa \cdot [u(z_2) - u(z_1)]}{u_*} \right\} } $$  \tag{14}

In case of negative $d_0$ resulting from (14), set $d_0 = 0$.

Convergence of iteration by (13) depends on $u$, $u_*$, and $L$. We found by trial and error that for many cases convergence can be accelerated by using a modified iteration equation:

$$ d_{i+1} = d_i \cdot (1-r) + r \cdot \left[ z_1 - \frac{z_2 - d_i}{\exp \left\{ \frac{\kappa \cdot [u(z_2) - u(z_1)]}{u_*} \right\} + \Psi_m \left( \frac{z_2-d_i}{L} \right) - \Psi_m \left( \frac{z_1-d_i}{L} \right) } \right] $$  \tag{15}

with
\[ r = 1 + \frac{u_z}{u(z_2) - u(z_1)} \]  \quad (16)

Eq. (15) mostly converges over about ten or less iteration steps. For PLATIN, we use 15 steps to make sure the solution obtained to be stable. However, as Monin-Obukhov theory is not always met by real micrometeorological conditions, iteration sometimes fails yielding a non-convergent series of \( d \) or even negative \( d \). But experience shows that, in general, there are still enough successful iteration results to provide a daily average of \( d \) which is adequate to run PLATIN.

Once \( d \) is known, \( z_{0m} \) is the only unknown left in the profile equation (11). But again, there is no way to get a closed mathematical solution from (11). Thus, \( z_{0m} \) must be iterated, too. Appropriate rearrangement of eq. (11) provides the following iteration rule where \( z \) represents one of the two measurement heights for \( u, z_1 \) and \( z_2 \):

\[
z_{0m,i+1} = \frac{z - d}{\exp\left(\frac{\kappa \cdot u(z)}{u_*} + \Psi_m\left(\frac{z - d}{L}\right) - \Psi_m\left(\frac{z_{0m,i}}{L}\right)\right)} \quad (17)
\]

Iteration is initialized by

\[
z_{0m,0} = \frac{z - d}{\exp\left(\frac{\kappa \cdot u(z)}{u_*}\right)} \quad (18)
\]

which can be derived from eq. (17) by neglecting the stability functions. Usually the iteration takes less than ten steps (a number we generally use for PLATIN).

As there are two heights with wind velocity measurements, eq. (17) may be applied twice offering the chance for intercomparison of the results which ideally should coincide.
Appendix J

Estimation of bulk canopy resistance for water vapour from measured data

Eddy covariance measurements of sensible heat $H$, latent heat $\lambda E$, and friction velocity $u^*$, together with measurements of air temperature $t_a$ and relative humidity $rH$ can be used to estimate bulk canopy resistance for water vapour $R_{c, H2O}$ according to:

$$ R_{c, H2O} = \frac{0.622 \cdot \rho_{\text{moist air}}}{p} \cdot \frac{e_{\text{sat}}(d+z_{0h}) - e(d+z_{0h})}{(\lambda E \cdot \lambda)} \quad (J1) $$

with

- $\rho_{\text{moist air}}$ density of moist air at absolute temperature $T$ [kg m$^{-3}$]; cf. (F7) and (F8)
- $p$ air pressure [hPa]
- $e_{\text{sat}}(d+z_{0h})$ saturation water vapour pressure at $z = d + z_{0h}$ [hPa]; cf. eqs. (F2) and (F3)
- $e(d+z_{0h})$ actual water vapour pressure at $z = d + z_{0h}$ [hPa]
- $\lambda$ latent heat of water vaporisation at $z = d + z_{0h}$ [J kg$^{-1}$]; cf. eq. (28)

Canopy surface temperature $T_s$ needed for calculations is related to potential canopy surface temperature as described by eq. (25). Actual water vapour pressure $e(d+z_{0h})$ is given by:

$$ e(d+z_{0h}) = e(z_{\text{ref}}, T) + \frac{(\lambda E \cdot \lambda)}{0.622 \cdot \rho_{\text{moist air}}} \cdot \left( R_{\text{ah}}(d+z_{0m}, z_{\text{ref}}, T) + R_{\text{b, heat}} \right) \quad (J2) $$

with

- $e(z_{\text{ref}}, T)$ actual water vapour pressure at $z = z_{\text{ref}}, T$ [hPa]; cf. eq. (F4)
- $R_{\text{ah}}(d+z_{0m}, z_{\text{ref}}, T)$ atmospheric resistance [s m$^{-1}$]; cf. eqs. (2) and (3)
- $R_{\text{b, heat}}$ quasi-laminar resistance for sensible heat [s m$^{-1}$]; cf. eq. (7)

Atmospheric resistance $R_{\text{ah}}(d+z_{0m}, z_{\text{ref}}, T)$, quasi-laminar resistance for sensible heat $R_{\text{b, heat}}$, and Monin-Obukhov length $L$ (cf. eq. (5)), needed for the resistance estimations, are calculated using measured friction velocity $u^*$ and sensible heat flux $H$.

As described in Chapter 2.3, bulk canopy resistance $R_{c, H2O}$ is a composite resistance describing stomatal and cuticular transpiration and evaporation. In PLATIN, $R_{c, H2O}$ is approximated by a weighted combination of soil resistance $R_{\text{soil, H2O}}$, bulk stomatal resistance $R_{c, \text{stom, H2O}}$ and bulk cuticle resistance $R_{c, \text{cut, H2O}}$ known for a fully developed canopy (without senescent leaves) under optimum conditions for maximal transpiration. The weights $\beta^*$ and $\beta$ depend on the actual canopy development stage taking into account the transition from a dense canopy to a sparse canopy as given by eq. (8).

Consequently, for a given canopy development stage bulk stomatal resistance $R_{c, \text{stom, H2O}}$ or bulk stomatal conductance for water vapour $g_{c, \text{stom, H2O}}$ can be calculated by:

$$ g_{c, \text{stom, H2O}} = \frac{1-\beta^*}{R_{c, \text{stom, H2O}}} = \frac{1}{R_{c, H2O}} - \frac{1-\beta^*}{R_{c, \text{cut, H2O}}} - \frac{\beta}{R_{\text{soil, H2O}}} \quad (J3) $$

For a dense canopy, which may be assumed when evaporation from soil is below 5 % of total evapotranspiration, $R_{c, H2O}$ is often used as a first estimate of $R_{c, \text{stom, H2O}}$. We found that this assumption is associated with a mean error of approx. 10 %. Therefore, eq. (J3) represents a useful tool to derive stomatal resistance directly from measurement-based latent and sensible heat flux, friction velocity, Monin-Obukhov length and canopy resistance for water vapour. These measurement-based entities can serve for calibration of PLATIN during daylight hours with global radiation $S_t \geq 100$ W m$^{-2}$ if the subsequent quality criteria are met:

- consistency of measured data set indicated by: $e_{\text{sat}}(d+z_{0h}) - e(d+z_{0h}) > 0$ hPa
- consistency of measured $\lambda E$ as indicated by positive values during daylight hours
- no rainfall
- interception reservoir empty for current and previous data set (cf. eqs. (17) and (18))
• relative air humidity $rH < 75\%$
• integral turbulence characteristic (ITC) test according to Thomas and Foken (2002)
• stationarity tests for friction velocity, latent and sensible heat according to Foken and Wichura (1996)

Additionally, as PLATIN is based on the canopy energy balance, data sets to be used for model calibration are required to closely approach the energy balance closure. Therefore, the following criteria must be satisfied:

• closure of energy balance: $\text{ABS}(R_{\text{net}} - G - \lambda E - H) < 25 \text{ W m}^{-2}$

This holds also for nighttime, where only measured sensible heat fluxes can be used for model calibration.
Appendix K

Upscaling of meteorological data measured above short vegetation to a height of 50 m

If PLATIN for Excel is to be applied to forest ecosystems, information on the state of the air above the forest canopy is required (wind speed, temperature, humidity, concentrations). If data are not available they have to be estimated from measurements above short vegetation near the forest. While this does not impose serious problems for daylight hours with well established turbulence, transition times with growing internal boundary layers near the surface and especially the night with vertically high reaching, sometimes strongly stable stratification allow only for a rough estimation procedure.

- **daytime with unstable atmospheric stratification regime**

Once unstable stratification is established on daylight hours (MO length $L < 0$ m), vertical exchange leads to effective coupling between surface fluxes and fluxes in heights well above the surface. Neglecting occasional horizontal advection (which, as a matter of fact, can never be accounted for in one-dimensional models like PLATIN), it seems reasonable to extend the concept of height-constant fluxes up to heights of some tens of metres. Monin-Obukhov theory can then be used to predict wind speed, temperature, and humidity above the forest canopy from data measured at a nearby short-vegetation location. For the height up to which profiles are to be extended we arbitrarily select 50 m above ground in the field, making sure to be well above the canopy even in case of tall forests. The calculation procedure, which we shortly call upscaling procedure, will be described in the following.

The horizontal wind velocity $u$ at $z = 50$ m can be obtained from the vertical profile equation based on Monin-Obukhov theory:

$$u(z) = \frac{u_*}{\kappa} \cdot \left[ \ln \left( \frac{z - d}{z_{0m}} \right) - \Psi_m \left( \frac{z - d}{L} \right) \right] \left[ \ln \left( \frac{z_{0m}}{d} \right) - \Psi_m \left( \frac{z_{0m}}{L} \right) \right]$$

(K1)

The parameterisations for the atmospheric stability functions for momentum $\Psi_m$ and sensible heat $\Psi_h$ are described in Appendix C. The calculation friction velocity is computed from eq. (4).

Estimation of air temperature at $z = 50$ m above ground is based on the calculation of the potential temperature $\theta [K]$:

$$\theta(z) = \theta(d + z_{0m}) - \frac{H \cdot R_{ah}(z, d + z_{0m})}{\rho_{\text{moist air}} \cdot c_p, \text{moist air}}$$

(K2)

from which the absolute air temperature $T$ can easily be derived using the relation given in eq. (6):

$$\theta(z) = T(z) + (z \cdot \Gamma_d)$$

(K3)

with

$$\Gamma_d = -9.76 \, \text{K.km}^{-1}$$

assuming $\rho_{\text{moist air}}(z) = \rho_{\text{moist air}}(z_{\text{ref} T}) \equiv \rho_{\text{moist air}}(d+z_{0m})$

and $c_p, \text{moist air}(z) = c_p, \text{moist air}(z_{\text{ref} T}) \equiv c_p, \text{moist air}(d+z_{0m})$.

The turbulent atmospheric resistance $R_{ah}$ between the heights $z = 50$ m and $z = d+z_{0m}$ is given by:

$$R_{ah}(z, d + z_{0m}) = \frac{\ln \left( \frac{z - d}{z_{0m}} \right) - \Psi_h \left( \frac{z - d}{L} \right) + \Psi_h \left( \frac{z_{0m}}{L} \right)}{\kappa \cdot u_*}$$

(K4)
The potential air temperature $\theta$ at $z = d + z_{0m}$ follows from:

\[
H = \rho_{\text{moist air}} \cdot c_{p, \text{moist air}} \cdot \frac{\theta(d + z_{0m}) - \theta(z_{\text{ref}, T})}{R_{ah}(z_{\text{ref}, T}, d + z_{0m})}
\approx \rho_{\text{moist air}} \cdot c_{p, \text{moist air}} \cdot \frac{T(d + z_{0m}) - T(z_{\text{ref}, T})}{R_{ah}(z_{\text{ref}, T}, d + z_{0m})}
\]

with $z_{\text{ref}, T} \approx 2$ m above ground

The relative humidity $rH$ at $z = 50$ m above ground is defined by:

\[
rH(z) = \frac{e_{\text{water vapour pressure}(z)}}{e_{\text{saturation water vapour pressure}(z)}} \cdot 100
\]

The water vapour pressure $e$ at $z = 50$ m is calculated from modelled latent heat flux $\lambda E$ and the specific humidity of air $q$ at $z = 50$ m, which in turn is estimated from $q$ at $z = d + z_{0m}$ making use of the turbulent atmospheric resistance $R_{ah}$ between the two heights:

\[
e_{\text{water vapour pressure}(z)} = \frac{p \cdot q(z)}{0.622 + 0.378 \cdot q(z)}
\]

with $p$ the air pressure [hPa] neglecting the relatively weak height dependence of $p$ which is about 1/8 hPa·m$^{-1}$ in the lower 100 m of the atmosphere (e.g. Liljequist and Cehak, 1979).

The specific humidity of air at $z = 50$ m above ground is given by:

\[
q(z) = q(d + z_{0m}) - \frac{\lambda E \cdot R_{ah}(z, d + z_{0m})}{\lambda \cdot \rho_{\text{moist air}}}
\]

with $\lambda$ latent heat of water vaporisation (eq. (28))

and

\[
q(d + z_{0m}) = q(z_{\text{ref}, T}) + \frac{\lambda E \cdot R_{ah}(z_{\text{ref}, T}, d + z_{0m})}{\lambda \cdot \rho_{\text{moist air}}}
\]

assuming $\rho_{\text{moist air}}(z) \approx \rho_{\text{moist air}}(z_{\text{ref}, T}) \equiv \rho_{\text{moist air}}(d+z_{0m})$.

The saturation water vapour pressure of the atmosphere at $z = 50$ m is calculated according to eqs. (F2) and (F3) in Appendix F.

The ozone concentration $\rho_{O3}$ at $z = 50$ m above ground is approximated by:

\[
\rho_{O3}(z) = -\left[F_{\text{total}}(O_3) \cdot R_{ah}(z, d + z_{0m})\right] + \rho_{O3}(d + z_{0m})
\]

Like the other profile equations mentioned before, eq. (K10) results from Monin-Obukhov theory which assumes vertically constant fluxes. In case of ozone this means a drastic simplification as it completely neglects the influence of air chemistry. However, chemistry modelling would be far beyond the aim of simply estimating missing data above the forest canopy.

- nighttime with stable atmospheric stratification regime

During night, surface cooling leads to increasing atmospheric stability finally restricting the layer of presumably height-constant fluxes to a couple of metres above surface. The air above this shallow layer
is in a so-called z-less state (Mahrt and Vickers, 2003; Basu et al., 2006) which means that fluxes are
determined by local physics and, if ever, are only loosely correlated with surface fluxes. As Monin-
Obukhov theory is not applicable especially during situations with strong stable atmospheric stratification,
i.e. low wind velocities, no equations like that described above for daylight hours can be given for
the night. Nevertheless, there should be a way of roughly estimating the desired data at height 50 m.

In the first place, as a special case, we shall deal with very stable atmospheric stratification. We start
with the assumption that the air above ten metres height can generally be described by z-less scaling
(cf. e.g. Mahrt and Vickers, 2003). Combination of the definition of vertical momentum mixing length $l$,

$$ l(z) \equiv \frac{u_*}{\partial u/\partial z} \quad (K11) $$

where $z$ height $z$ above ground [m], $u_*$ friction velocity [m·s$^{-1}$], $u$ wind speed at height $z$ [m·s$^{-1}$]

and the z-less mixing length approximation according, e.g., to Mahrt and Vickers (2003),

$$ l = 0.5 \cdot \frac{u_*}{N} \quad (K12) $$

yields (with gravitational acceleration $g = 9.81$ m·s$^{-2}$ and potential temperature $\theta$)

$$ \frac{\partial u}{\partial z} = 2 \left( \frac{g}{\theta} \cdot \frac{\partial \theta}{\partial z} \right) \quad (K13) $$

when it is taken into account that the Brunt-Vaisala frequency $N$ is given by (cf. Stull, 1988):

$$ N = \sqrt{g \cdot \frac{\partial \theta}{\partial z}} \quad (K14) $$

As the potential temperature gradient can be assumed vertically constant for the height range con-
sidered (cf. e.g. large eddy simulation results in Basu et al., 2006), the partial derivatives in (K13) may
be replaced by differential quotients. After additional introduction of an effective mean potential tem-
perature $\bar{\theta}$, (K13) becomes:

$$ \frac{\Delta u}{\Delta z} = 2 \left( \frac{g}{\bar{\theta}} \cdot \frac{\Delta \theta}{\Delta z} \right) \quad (K15) $$

With the potential temperature gradient in the order of 1 K·100 m$^{-1}$ (cf. e.g. Figure 3 in Basu et al.,
2006) and a reasonable mean temperature of 280 K, (K15) leads to

$$ u(50 \text{ m}) = u(10 \text{ m}) + 1.5 \text{ m} \cdot \text{s}^{-1} \quad (K16) $$

Especially for calm weather conditions it may yield too high a value of $u(50 \text{ m})$. As the majority of the
nocturnal Linden grassland site wind data is characterized by low winds, we often found (K16) to yield
nocturnal $u(50 \text{ m})$ considerably exceeding $u(50 \text{ m})$ during subsequent daylight hours as calculated by
eq. (K1). Because the daily course of $u(50 \text{ m})$ should exhibit a certain continuity, we looked for another
way to estimate the nocturnal $u(50 \text{ m})$. Applying eq. (K1) for nighttime data sets, we found on the average
that for $u(10 \text{ m}) > 2 \text{ m} \cdot \text{s}^{-1}$ the ratio $u(50 \text{ m})/u(10 \text{ m})$ is about 1.5 ± 0.1 while it significantly decays
for lower values of $u(10 \text{ m})$. We assume that eq. (K1) is applicable in case of nocturnal $u(10 \text{ m}) > 2 \text{ m} \cdot \text{s}^{-1}$ and suggest to use the factor 1.5 for lower wind speed:
1.5 m) (10 m) (50 m) = \theta (10 m) + 0.4 K \tag{K18}

Unless in the very rare case of nocturnal air warming induced by horizontal advection, $\theta(50 \text{ m})$ should be limited by the last daylight-hour value.

As temperature is measured at the meteorological screen height, i.e. 2 m above ground, the potential temperature difference between 2 and 10 m has still to be estimated. Additionally, potential temperature has to be transformed into actual temperature, as the latter is needed as input to PLATIN. As the nocturnal forest energy balance will not be very sensitive to the exact temperature at 50 m height, we restrict ourselves to a rough order-of-magnitude estimate: The temperature gradient at higher levels above ground is of the order $10^{-2}$ K·m$^{-1}$, while it is of the order $1$ K·m$^{-1}$ near the ground. Therefore we assume an intermediate lapse rate of $10^{-1}$ K·m$^{-1}$ for the layer between 2 and 10 m. Combining this with (K18) and taking into account that actual temperature at 50 m height is about 0.4 K higher than potential temperature at this level (cf. eq. (K3)), we arrive at the final rough approximation equation for the actual temperature:

$$ T(50 \text{ m}) = T(2 \text{ m}) + 1 \text{ K} \tag{K19} $$

It has already been mentioned that, unless there is a temperature increase due to horizontal advection of warmer air, the temperature at 50 m height should not exceed the respective last daylight-hour value. This also holds for the actual temperature as estimated from (K19).

To approximate air humidity at 50 m height, we assume a homogeneous air mass with vertically constant mixing ratio between water vapour and dry air, which is practically the same as to assume vertically constant specific humidity. Therefore we suggest as a rough estimate:

$$ q(50 \text{ m}) = q(2 \text{ m}) \tag{K20} $$

Especially during night, the O$_3$ concentration at reference height can be reduced significantly by reaction with NO. The extent of reduction depends on the NO source strength of the ecosystem under consideration or e.g. of exhaust emissions. Taking into account a stable atmospheric stratification regime it can be assumed, that the mean O$_3$ gradient during night be greater than the mean gradient during unstable stratification. We calculate a mean ratio for $\rho_{O_3}(50 \text{ m}) : \rho_{O_3}(z_{ref})$ of approx. 1.1 under unstable atmospheric stratification regime and suggest this value as a rough estimate for nighttime where we accept a probable underestimation of the upscaled O$_3$ concentration at $z = 50$ m:

$$ \rho_{O_3}(50 \text{ m}) = \rho_{O_3}(z_{ref}) \cdot 1.1 \tag{K21} $$

- **transition times**

The micrometeorological processes during transition times would deserve a special treatment, because neither the upscaling procedure for daylight hours nor the one for nighttime can be applied to transition hours. It must be stated that there is no physically based model to describe the interrelation of the meteorological entities at 50 m height and those at the ground. For wind speed this problem is overcome by simply applying the scheme described in context of eq. (K17), i.e. by calculating wind speed at 50 m by eq. (K17) for $u(10 \text{ m}) \leq 2 \text{ m·s}^{-1}$ and by eq. (K1) for $u(10 \text{ m}) > 2 \text{ m·s}^{-1}$. Ozone concentration and air temperature at 50 m during transition times must be calculated by linearly interpolating 50 m data between day and night (in the late afternoon and early evening) and vice versa (in the early morning hours). Data gaps of relative humidity are closed by calculating relative humidity from temporally interpolated data of specific humidity and air temperature.
• **mean diurnal variation of micrometeorological parameters**

The mean diurnal variations of horizontal wind velocity, air temperature, relative humidity and ozone concentration measured at reference height and upscaled to a height of 50 m above ground according to the procedure described above are illustrated in Figure K1.

![Fig. K1](image)

Mean diurnal variations of horizontal wind velocity, air temperature, relative humidity and ozone concentration measured at reference height in June 2004 and upscaled to a height of 50 m above ground

• **reference surface for upscaling procedure**

In principle, the application of the "daytime" approach presupposes the knowledge of all the canopy characteristics (e.g. \(R_{\text{c, stom, min}}\), Jarvis factors, \(\text{LAI}\), development) and input parameters (e.g. soil moisture) as described in this manual. If these informations are not available, a reference surface should be applied. We recommend a grassland with an assumed height \(h\) of 0.12 m and a short-wave albedo \(\alpha\) of 0.23 as described in the FAO guideline for computing crop evapotranspiration (Allen et al., 2002). Soil moisture is assumed to be constant at 70 % of field capacity. Instead of a fixed surface resistance \(R_{\text{c, H2O}}\) we recommend to apply e.g. the canopy characteristics of the Linden grassland site. According to Allen et al. (2002) this "reference surface closely resembles an extensive surface of green, well-watered grass of uniform height, actively growing and completely shading the ground".

At least the following input parameters must be measured at the monitoring station: global radiation, horizontal wind velocity at 10 m above ground, air temperature, relative humidity and ozone concentration.
Note:
Conversion of ozone concentration unit µg·m⁻³ into ppb at upscaling height z according to eq. (M1) presuppose an upscaling of atmospheric pressure \( p(z) \):

\[
p(z) = p(z_{\text{ref}}) \cdot \exp \left( -\frac{g \cdot (z - z_{\text{ref}})}{R \cdot T_m} \right)
\]

with
- \( g \) gravitational acceleration \( \approx 9.81 \text{ m·s}^{-2} \)
- \( R \) gas constant for dry air \( = 287.04 \text{ J·kg}^{-1}·\text{K}^{-1} \)
- \( T_m \) average temperature of the air layer under consideration \( \approx \frac{[T(z) + T(z_{\text{ref}})]}{2} \)
Appendix L

Calculation of photosynthetic photon flux density from global radiation

According to VDI 3786 sheet 13 (1993) PPFD can be calculated from the global radiation with an accuracy sufficient for practical purposes. For countries in the centre of Europe, the following conversion factors can be applied.

\[ PPFD = a \cdot S_i \]

with

- \( PPFD \) photosynthetic photon flux density \([\mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}]\)
- \( S_i \) global radiation \([\text{W} \cdot \text{m}^{-2}]\)
- \( a \) conversion factor \([\mu\text{mol} \cdot \text{W}^{-1} \cdot \text{s}^{-1}]\)

The variation of factor \( a \) with time of year is given in Table L1.

<table>
<thead>
<tr>
<th>Month</th>
<th>Means of all days ( a ) in ( \mu\text{mol} \cdot \text{W}^{-1} \cdot \text{s}^{-1} )</th>
</tr>
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<tr>
<td>January</td>
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<td>February</td>
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<td>March</td>
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<tr>
<td>April</td>
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<td>May</td>
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<td>June</td>
<td>2.07</td>
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<td>July</td>
<td>2.07</td>
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<td>August</td>
<td>2.10</td>
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<tr>
<td>September</td>
<td>2.07</td>
</tr>
<tr>
<td>October</td>
<td>2.07</td>
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<td>November</td>
<td>2.06</td>
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<tr>
<td>December</td>
<td>2.03</td>
</tr>
<tr>
<td>total of year</td>
<td>2.03</td>
</tr>
</tbody>
</table>
Appendix M

Conversion of units

The denotation follows IUPAC (1993).

gas concentration

\[ \rho_A = (\chi_A \cdot \alpha_{cf}) \cdot \left( \frac{M_A \cdot \beta_{cf}}{V_m \cdot \gamma_{cf}} \right) \cdot \left( \frac{T_0}{T} \right) \cdot \left( \frac{p}{p_0} \right) \]  \hfill (M1)

with

- \( \rho_A \) partial mass density (mass concentration) of a gaseous species A [\( \mu g \cdot m^{-3} \)]
- \( \chi_A \) mixing ratio or mole fraction of a gaseous species A [ppm]
- \( \alpha_{cf} \) mixing ratio conversion factor (\( \alpha_{cf} = 10^{-6} \, \text{mol} \cdot \text{mol}^{-1} \cdot \text{ppm}^{-1} \))
- \( M_A \) molar mass of a gaseous species A [g mol\(^{-1}\)]
- \( \beta_{cf} \) mass conversion factor (\( \beta_{cf} = 10^6 \, \mu g \cdot g^{-1} \))
- \( V_m \) molar volume (\( V_m = 22.4 \, l \cdot mol^{-1} \))
- \( \gamma_{cf} \) volume conversion factor (\( \gamma_{cf} = 10^{-3} \, m^3 \cdot l^{-1} \))
- \( T_0 \) standard temperature (\( T_0 = 273.15 \, K \))
- \( T \) actual absolute temperature [K]; \( \frac{T - T_0}{K} = \frac{t_a}{\degree C} \)
- \( t_a \) actual air temperature [\( ^\circ C \)]
- \( p_0 \) standard atmospheric pressure (\( p_0 = 1013.25 \, hPa \))
- \( p \) actual atmospheric pressure [hPa]

\[ p_A = (\chi_A \cdot \alpha_{cf}) \cdot p = c_A \cdot (V_m \cdot \gamma_{cf}) \cdot p \]  \hfill (M2)

with

- \( p_A \) partial pressure of a gaseous species A [Pa]
- \( \chi_A \) mixing ratio or mole fraction of a gaseous species A [ppm]
- \( \alpha_{cf} \) mixing ratio conversion factor (\( \alpha_{cf} = 10^{-6} \, \text{mol} \cdot \text{mol}^{-1} \cdot \text{ppm}^{-1} \))
- \( p \) atmospheric pressure [Pa]
- \( c_A \) concentration (amount concentration) of a gaseous species A [mol m\(^{-3}\)]
- \( V_m \) molar volume (\( V_m = 22.4 \, l \cdot mol^{-1} \))
- \( \gamma_{cf} \) volume conversion factor (\( \gamma_{cf} = 10^{-3} \, m^3 \cdot l^{-1} \))

\[ \rho_A = c_A \cdot (M_A \cdot \beta_{cf}) \]  \hfill (M3)

with

- \( \rho_A \) partial mass density (mass concentration) of a gaseous species A [\( \mu g \cdot m^{-3} \)]
- \( c_A \) concentration (amount concentration) of a gaseous species A [mol m\(^{-3}\)]
- \( M_A \) molar mass of a gaseous species A [g mol\(^{-1}\)]
- \( \beta_{cf} \) mass conversion factor (\( \beta_{cf} = 10^6 \, \mu g \cdot g^{-1} \))

\[ c_A = (\chi_A \cdot \alpha_{cf}) \cdot \left( \frac{1}{V_m \cdot \gamma_{cf}} \right) \cdot \left( \frac{T_0}{T} \right) \cdot \left( \frac{p}{p_0} \right) \]  \hfill (M4)

with

- \( c_A \) concentration (amount concentration) of a gaseous species A [mol m\(^{-3}\)]
- \( \chi_A \) mixing ratio or mole fraction of a gaseous species A [ppm]
- \( \alpha_{cf} \) mixing ratio conversion factor (\( \alpha_{cf} = 10^{-6} \, \text{mol} \cdot \text{mol}^{-1} \cdot \text{ppm}^{-1} \))
- \( V_m \) molar volume (\( V_m = 22.4 \, l \cdot mol^{-1} \))
γ_	ext{cf} \quad \text{volume conversion factor (γ_	ext{cf} = 10^{-3} \text{ m}^3\text{l}^{-1})}

T_0 \quad \text{standard temperature (T_0 = 273.15 K)}

T \quad \text{actual absolute temperature [K]; } \frac{T - T_0}{K} = \frac{t_a}{\circ \text{C}}

t_a \quad \text{actual air temperature [°C]}

\rho_0 \quad \text{standard atmospheric pressure (\rho_0 = 1013.25 hPa)}

\rho \quad \text{actual atmospheric pressure [hPa]}

**vertical flux density of gaseous species F_A**

\[ F(A) [\text{g} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] = F(A) [\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \cdot M_A \quad \text{(M5)} \]

with

- \( F(A) \) \quad \text{vertical flux density of a gaseous species A [g m}^{-2} \text{s}^{-1} \text{ or mol m}^{-2} \text{s}^{-1}] 
- \( M_A \) \quad \text{molar mass of a gaseous species A [g mol}^{-1}] 

**conductance for gaseous species g_A**

\[ g_A [\text{m} \cdot \text{s}^{-1}] = g_A [\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1}] \cdot \left\{ \frac{1}{V_m \cdot \gamma_	ext{cf}} \right\} \cdot \left\{ \frac{T_0}{T} \right\} \cdot \left\{ \frac{p}{\rho_0} \right\}^{-1} \quad \text{(M6)} \]

with

- \( g_A \) \quad \text{conductance for a gaseous species A [mol m}^{-2} \text{s}^{-1} \text{ or m} \cdot \text{s}^{-1}] 
- \( V_m \) \quad \text{molar volume (V_m = 22.4 m}^3 \text{ mol}^{-1}) 
- \( \gamma_	ext{cf} \) \quad \text{volume conversion factor (γ_	ext{cf} = 10^{-3} m}^3 \text{ l}^{-1}) 
- \( T_0 \) \quad \text{standard temperature (T_0 = 273.15 K)} 
- \( T \) \quad \text{actual absolute temperature [K]; } \frac{T - T_0}{K} = \frac{t_a}{\circ \text{C}} 
- \( t_a \) \quad \text{actual air temperature [°C]} 
- \( \rho_0 \) \quad \text{standard atmospheric pressure (\rho_0 = 1013.25 hPa)} 
- \( \rho \) \quad \text{actual atmospheric pressure [hPa]}
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<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Bezeichnung</th>
<th>unit/Einheit</th>
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<tr>
<td>α</td>
<td>short-wave albedo</td>
<td>kurzwellige Albedo der Erd- bzw. Bestandesoberfläche</td>
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<td>αcf</td>
<td>mixing ratio conversion factor</td>
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<td>fraction of glycolate carbon not returned to the chloroplast</td>
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<td>Wichtungsfaktor für den Entwicklungs-zustand des Bestandes basierend auf LAIf bzw. SAI</td>
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<td>vegetation type-specific ratio of ammonium to protons in the apoplast</td>
<td>vegetationsspezifisches Verhältnis von NH$_4^+$ und H$^+$ im Apoplasten</td>
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<td>Γ</td>
<td>CO$_2$ compensation point in the absence of mitochondrial respiration</td>
<td>CO$_2$-Kompensationspunkt der Brutto-photosynthese</td>
<td>Pa</td>
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<td>Zeitintervall</td>
<td>s</td>
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<td>z.B. Bestandestiefe</td>
<td>m</td>
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<td>evaporation</td>
<td>Evaporation</td>
<td>W m$^{-2}$</td>
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<tr>
<td>λE$_{transpiration}$</td>
<td>transpiration</td>
<td>Transpiration</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>λgeo</td>
<td>longitude</td>
<td>geographische Länge</td>
<td>degree, Grad</td>
</tr>
<tr>
<td>ν</td>
<td>kinematic viscosity of dry air</td>
<td>molekulare kinematische Viskosität der Luft</td>
<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>ρA(z0c)</td>
<td>partial mass density of a gaseous species A at height $z = z_0c$</td>
<td>Partialdichte des Spurengases A in der Höhe $z = z_0c$</td>
<td>µg m$^{-3}$</td>
</tr>
<tr>
<td>ρA(z0m)</td>
<td>partial mass density of a gaseous species A at height $z = z_0m$</td>
<td>Partialdichte des Spurengases A in der Höhe $z = z_0m$</td>
<td>µg m$^{-3}$</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
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<tr>
<td>$\rho_A(z_{ref},A)$, $\rho_A(z_{ref})$</td>
<td>partial mass density of a gaseous species A at height $z = z_{ref},A$</td>
<td>Partialdichte des Spurengases A in der Messhöhe $z = z_{ref},A$</td>
<td>$\mu g\cdot m^{-3}$</td>
</tr>
<tr>
<td>$\rho_A$, comp</td>
<td>canopy compensation concentration of gaseous species A</td>
<td>Bestandeskompensationskonzentration für die Spurengasspezies A</td>
<td>$\mu g\cdot m^{-3}$</td>
</tr>
<tr>
<td>$\rho_A$, soil</td>
<td>partial mass density of a trace gas A in the soil</td>
<td>Partialdichte des Spurengases A im Boden</td>
<td>$\mu g\cdot m^{-3}$</td>
</tr>
<tr>
<td>$\rho_A$, stom</td>
<td>partial mass density of a trace gas A in the substomatal cavities</td>
<td>Partialdichte des Spurengases A in den substomatären Räumen</td>
<td>$\mu g\cdot m^{-3}$</td>
</tr>
<tr>
<td>$\rho_{cd}$</td>
<td>canopy reflection coefficient for diffuse PAR</td>
<td>Bestandesreflektionskoeffizient für diffuse photosynthetisch-aktive Strahlung</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$\rho_{cb}$</td>
<td>canopy reflection coefficient for beam PAR</td>
<td>Bestandesreflektionskoeffizient für direkte photosynthetisch-aktive Strahlung</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$\rho_h$</td>
<td>reflection coefficient for beam PAR of a canopy with horizontal leaves</td>
<td>Bestandesreflektionskoeffizient für direkte photosynthetisch-aktive Strahlung für einen Pflanzenbestand mit horizontaler Blattausrichtung</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$\rho_{dry,av}$</td>
<td>density of dry air</td>
<td>Dichte der trockenen Luft</td>
<td>$kg\cdot m^{-3}$</td>
</tr>
<tr>
<td>$\rho_{moist,av}$</td>
<td>density of moist air</td>
<td>Dichte der feuchten Luft</td>
<td>$kg\cdot m^{-3}$</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>Stefan-Boltzmann constant</td>
<td>Stefan-Boltzmann-Konstante</td>
<td>$W\cdot m^{-2}\cdot K^{-4}$</td>
</tr>
<tr>
<td>$\sigma_{PAR}$</td>
<td>leaf scattering coefficient of PAR</td>
<td>PAR Blatt-Streuungskoeffizient</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$\phi$</td>
<td>solar elevation angle</td>
<td>Sonenhöhe</td>
<td>radians</td>
</tr>
<tr>
<td>$\phi_c$</td>
<td>combined conversion factor</td>
<td>Umrechnungsfaktor</td>
<td>$m^3\cdot \mu mol^{-3}$</td>
</tr>
<tr>
<td>$\phi_d$</td>
<td>day angle</td>
<td>Tageswinkel</td>
<td>radians, Bogenmaß</td>
</tr>
<tr>
<td>$\phi_{geo}$</td>
<td>latitude</td>
<td>geographische Breite</td>
<td>radians, Bogenmaß</td>
</tr>
<tr>
<td>$\phi_h$</td>
<td>hour angle</td>
<td>Stundenwinkel</td>
<td>radians, Bogenmaß</td>
</tr>
<tr>
<td>$\chi_A$</td>
<td>mole fraction of gaseous species A</td>
<td>Mischungsverhältnis bzw. Molenbruch der Spurengasspezies A</td>
<td>ppm, $\mu mol\cdot mol^{-1}$</td>
</tr>
<tr>
<td>$\gamma_h$</td>
<td>atmospheric stability function for sensible heat</td>
<td>integrierte Schichtungsfunktion für fühlbare Wärme</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$\gamma_m$</td>
<td>atmospheric stability function for momentum</td>
<td>integrierte Schichtungsfunktion für Impuls</td>
<td>dimensionless</td>
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<tr>
<td>$\omega$</td>
<td>convexity coefficient defining the smoothness of the transition between $f_T$ and $f_M$</td>
<td>Faktor, der den Übergang zwischen den $f_T$ und $f_M$ bestimmt</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$A_{1,Gphs}$</td>
<td>empirical constant</td>
<td>empirische Konstante</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$A_{2,Gphs}$</td>
<td>empirical constant</td>
<td>empirische Konstante</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$A$</td>
<td>canopy assimilation</td>
<td>Bestandesbruttophotosynthese</td>
<td>$\mu mol\cdot m^{-2}\cdot s^{-1}$</td>
</tr>
<tr>
<td>$A_{rub}$</td>
<td>Rubisco-limited rate of CO$_2$ assimilation</td>
<td>Rubisco-limitierte Bestandesbruttophotosynthese</td>
<td>$\mu mol\cdot m^{-2}\cdot s^{-1}$</td>
</tr>
<tr>
<td>$A_{et}$</td>
<td>electron-transport limited rate of CO$_2$ assimilation</td>
<td>Elektronentransport-limitierte Bestandesbruttophotosynthese</td>
<td>$\mu mol\cdot m^{-2}\cdot s^{-1}$</td>
</tr>
<tr>
<td>$A_{rel}$, $A_{rel,canopy}$</td>
<td>net rate of canopy photosynthesis</td>
<td>Bestandesnettophotosynthese</td>
<td>$\mu mol\cdot m^{-2}\cdot s^{-1}$</td>
</tr>
<tr>
<td>$A_{rel}$, $A_{rel,canopy}$</td>
<td>24-h integrated canopy net photosynthesis of the previous day</td>
<td>integrierte Bestandesnettophotosynthese des vergangenen Tages</td>
<td>$\mu mol\cdot m^{-2}\cdot s^{-1}$</td>
</tr>
<tr>
<td>$A_{ph}$</td>
<td>triose phosphate utilization-limited rate of CO$_2$ assimilation</td>
<td>durch Mangel an anorganischem P limitierte Bestandesbruttophotosynthese</td>
<td>$\mu mol\cdot m^{-2}\cdot s^{-1}$</td>
</tr>
<tr>
<td>$a_{1,Gphs}$</td>
<td>empirical constant</td>
<td>empirische Konstante</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$a_{rub}$</td>
<td>atmospheric transmission coefficient of PAR for clear sky conditions</td>
<td>atmosphärischer Transmissionskoeffizient für PAR bei wolkenfreiem Himmel</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$a_{sid}$</td>
<td>empirical constant</td>
<td>empirische Konstante</td>
<td>$mm^{-1}$</td>
</tr>
<tr>
<td>$a_{rel,1}$</td>
<td>empirical coefficient</td>
<td>empirischer Koeffizient</td>
<td>$W\cdot m^{-2}$</td>
</tr>
<tr>
<td>$a_{rel,2}$</td>
<td>empirical coefficient</td>
<td>empirischer Koeffizient</td>
<td>$W\cdot m^{-2}$</td>
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<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
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<td>--------</td>
<td>-------------</td>
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<td>--------------</td>
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<tr>
<td>$b_{IN}$</td>
<td>constant of proportionality</td>
<td>Proportionalitätskonstante</td>
<td>mm</td>
</tr>
<tr>
<td>$c_A$</td>
<td>concentration of gaseous species A</td>
<td>Konzentration des Spurengases A</td>
<td>mol m$^{-3}$</td>
</tr>
<tr>
<td>$c_i$</td>
<td>intercellular CO$_2$ concentration</td>
<td>CO$_2$-Konzentration in den Interzellen</td>
<td>$\mu$mol m$^{-3}$</td>
</tr>
<tr>
<td>$c_{I_AU}$</td>
<td>mean vegetation type-specific attenuation coefficient</td>
<td>mittlerer Strahlungsextinktionskoefizient des Pflanzenbestandes</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$c_{NH3}$</td>
<td>dimension adaptation factor</td>
<td>Faktor zur Dimensionsadaptation</td>
<td>K $\mu$g m$^{-3}$</td>
</tr>
<tr>
<td>$c_{p,d}$</td>
<td>specific heat of dry air at a constant pressure</td>
<td>spezifische Wärme von trockener Luft bei konstantem Druck</td>
<td>m$^2$ s$^{-2}$ K$^{-1}$</td>
</tr>
<tr>
<td>$c_{p,m}$</td>
<td>specific heat of moist air at a constant pressure</td>
<td>spezifische Wärme von feuchter Luft bei konstantem Druck</td>
<td>m$^2$ s$^{-2}$ K$^{-1}$</td>
</tr>
<tr>
<td>$c_{1,A2}$</td>
<td>empirical constant of the Jarvis CO$_2$ function describing dependence of photosynthesis rate on ambient CO$_2$ concentration</td>
<td>Konstanten, die die Abhängigkeit der Photosyntheserate von der CO$_2$-Konzentration in der Umgebungsluft in der Jarvis-CO$_2$-Funktion beschreiben</td>
<td>dimensionless, ppm</td>
</tr>
<tr>
<td>CET</td>
<td>Central European Time</td>
<td>mitteleuropäische Zeit</td>
<td>h</td>
</tr>
<tr>
<td>$C_l$</td>
<td>clay content</td>
<td>Tongehalt</td>
<td>%</td>
</tr>
<tr>
<td>$d$</td>
<td>displacement height</td>
<td>Verschiebungshöhe</td>
<td>m</td>
</tr>
<tr>
<td>$d+Z_{dc}$</td>
<td>sink height of a trace gas</td>
<td>Niveau der Spurengassenke</td>
<td>m</td>
</tr>
<tr>
<td>$d+Z_{dm}$</td>
<td>momentum sink height</td>
<td>Niveau der Impulssenke</td>
<td>m</td>
</tr>
<tr>
<td>$D_A$</td>
<td>diffusivity of a trace gas A in air</td>
<td>molekularer Diffusionskoeffizient der Spurengasspezies A in Luft</td>
<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>$d_p$</td>
<td>aerodynamic diameter of particles</td>
<td>aerodynamischer Partikeldurchmesser</td>
<td>$\mu$m</td>
</tr>
<tr>
<td>$DT$</td>
<td>duration of time interval</td>
<td>Dauer des Messintervalls</td>
<td>h</td>
</tr>
<tr>
<td>$E$</td>
<td>actual evapotranspiration</td>
<td>aktuelle Evapotranspiration</td>
<td>mm</td>
</tr>
<tr>
<td>$E_{pot}$</td>
<td>potential evapotranspiration</td>
<td>potentielle Evapotranspiration</td>
<td>mm</td>
</tr>
<tr>
<td>$E_a$</td>
<td>activation energy</td>
<td>Aktivierungsgenergie</td>
<td>J mol$^{-1}$</td>
</tr>
<tr>
<td>$E(u_*)$</td>
<td>$u_*$-dependent value for fine-particle constituents</td>
<td>schubspannungsgeschwindigkeitsabhängiger Wert für Schwebstaubstandteile</td>
<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>$EB_{residual}$</td>
<td>energy balance residual</td>
<td>Residuum der Energiebilanz</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$e_{sat,water vapour pressure}$</td>
<td>saturation water vapour pressure</td>
<td>Sättigungsdampfdruck</td>
<td>hPa</td>
</tr>
<tr>
<td>$e_{sat}$</td>
<td>equation of time</td>
<td>Zeitgleichung</td>
<td>h</td>
</tr>
<tr>
<td>$e_{water vapour pressure}$</td>
<td>water vapour pressure</td>
<td>Wasserdampfdruck</td>
<td>hPa</td>
</tr>
<tr>
<td>$f$</td>
<td>corrects for spectral quality of light</td>
<td>Lichtqualitätskorrekturfaktor</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$f_{R,A}$</td>
<td>chemical reactivity factor of trace gas A</td>
<td>chemischer Reaktivitätsfaktor der Spurengasspezies A</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$f_d$</td>
<td>fraction of diffusive radiation for cloudless skies</td>
<td>Anteil diffuser Sonnenstrahlung bei unbewölktem Himmel</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$f_b$</td>
<td>construction cost</td>
<td>&quot;Konstruktions&quot;kosten</td>
<td>mol mol$^{-1}$</td>
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<tr>
<td>$f_l$</td>
<td>factor quantifying the limitations of soil temperature on CO$_2$ evolution</td>
<td>Faktor, der die Limitierung der Bodentemperatur durch die Bodenfeuchte quantifiziert</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$f_{lamb}$</td>
<td>factor quantifying the limitations of soil moisture on CO$_2$ evolution</td>
<td>Faktor, der die Limitierung der Bodenfeuchte durch die Bodenprozesse quantifiziert</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$F$</td>
<td>turbulent vertical flux density</td>
<td>turbulente vertikale Flussdichte</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_s(A)$</td>
<td>flux density of a gaseous species including water vapour or an fine-particle constituent</td>
<td>Flussdichte einer Spurengasspezies einschließlich Wasserdampf oder eines Schwebstaubinhaltstoffs</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{c,absorbed}$</td>
<td>absorbed flux density</td>
<td>absorbierte Flussdichte</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{c,cut}$</td>
<td>flux density absorbed through the cuticle</td>
<td>durch die Cuticula absorbierte Flussdichte</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
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<td>-----------------</td>
<td>------------------------------------------------------------------------------</td>
<td>---------------------------------------------------------------</td>
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</tr>
<tr>
<td>$F_{\text{est}}$</td>
<td>flux density on external plant surfaces</td>
<td>Flussdichte auf externe Pflanzenoberflächen</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{c}}$</td>
<td>non-stomatal deposition</td>
<td>nicht-stomatäre Deposition</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{c, stom}}$</td>
<td>flux density absorbed through the stomata and the cuticle</td>
<td>durch Stomata und Cuticula absorbierte Flussdichte</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{c, stom, shaded}}$</td>
<td>stomatal uptake by the shaded leaf fraction of the canopy</td>
<td>stomatäre Aufnahme der Schattenblätter des Bestandes</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{c, stom, sunlit}}$</td>
<td>stomatal uptake by the sunlit leaf fraction of the canopy</td>
<td>stomatäre Aufnahme der Sonnenblätter des Bestandes</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{c, total}}$</td>
<td>stomatal uptake of the canopy per unit projected leaf area</td>
<td>Gesamt-Flussdichte</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{leaf, stom, sunlit}}$</td>
<td>stomatal uptake of the sunlit leaf fraction of the canopy</td>
<td>stomatäre Aufnahme der Sonnenblätter des Bestandes normiert auf eine Blattflächeneinheit</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{H}}$</td>
<td>humidity factor</td>
<td>Luftfeuchtigkeitsfaktor</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$F_{\text{H}}$</td>
<td>humidity factor</td>
<td>Luftfeuchtigkeitsfaktor</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$F_{\text{m}}$</td>
<td>flux density on the soil</td>
<td>Flussdichte auf Bodenoberflächen</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$F_{\text{soil}}$</td>
<td>flux density on the soil</td>
<td>Flussdichte auf Bodenoberflächen</td>
<td>g m$^{-2}$ s$^{-1}$, mol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$g$</td>
<td>gravitational acceleration</td>
<td>Gravitationsbeschleunigung</td>
<td>m s$^{-2}$</td>
</tr>
<tr>
<td>$G$</td>
<td>ground heat flux density</td>
<td>Energie, die in Bestand und Boden gespeichert bzw. von dort abgegeben wird</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$g_A$</td>
<td>conductance for a gaseous species A</td>
<td>Leitwert für Spurengasspezies A</td>
<td>mol m$^{-2}$ s$^{-1}$, m s$^{-1}$</td>
</tr>
<tr>
<td>$g_{\text{leaf, stom, sunlit, O$_3$}}$</td>
<td>stomatal conductance of sunlit leaves per unit projected leaf area</td>
<td>stomatärer Leitwert der Sonnenblätter des Bestandes normiert auf eine Blattflächeneinheit</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$\text{GMT}$</td>
<td>Greenwich Mean Time</td>
<td>Mittlere Greenwich-Zeit</td>
<td>h</td>
</tr>
<tr>
<td>$G_{\text{phys}}$</td>
<td>physical heat storage flux density</td>
<td>Energie, die physikalisch in Bestand und Boden gespeichert bzw. von dort abgegeben wird</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$G_{\text{soil}}$</td>
<td>soil heat flux density</td>
<td>Bodenwärmestrom</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$h$</td>
<td>canopy height</td>
<td>Bestandeshöhe</td>
<td>m</td>
</tr>
<tr>
<td>$H$</td>
<td>turbulent flux density of sensible heat</td>
<td>vertikale turbulente Flussdichte fühlaberer Wärme</td>
<td>W m$^{-2}$</td>
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<tr>
<td>$H_{\text{A}}$</td>
<td>effective Henry's law value of a trace gas A</td>
<td>effektiver Henry-Koeffizient der Spurengasspezies A</td>
<td>M atm$^{-1}$</td>
</tr>
<tr>
<td>$I_{2}$</td>
<td>photosynthetically active radiation absorbed by PSII of the leaves of the canopy</td>
<td>PAR absorbiert durch PSII derBlätter des Bestandes</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$I_{p}$</td>
<td>direct beam irradiance</td>
<td>direkte PAR</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$I_{c, shaded}$</td>
<td>irradiance absorbed by the shaded fraction of non-senescent leaves of the canopy</td>
<td>PAR absorbiert durch die nicht-seneszenten Schattenblätter des Bestandes</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$I_{c, sunlit}$</td>
<td>irradiance absorbed by the sunlit fraction of non-senescent leaves of the canopy</td>
<td>PAR absorbiert durch die nicht-seneszenten Sonnenblätter des Bestandes</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$I_s$</td>
<td>diffusive irradiance</td>
<td>diffuse PAR</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>INT</td>
<td>interception</td>
<td>Interception</td>
<td>mm</td>
</tr>
<tr>
<td>$\text{INT}_{\text{max}}$</td>
<td>interception reservoir capacity of the canopy</td>
<td>Bestandesinterzeptionskapazität</td>
<td>mm</td>
</tr>
<tr>
<td>$J$</td>
<td>electron transport rate per unit leaf area</td>
<td>Elektronentransportrate pro Blattflächeneinheit</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$J_{\text{max}}$</td>
<td>light-saturated (maximum) electron transport rate per unit leaf area</td>
<td>maximale Elektronentransportrate pro Blattflächeneinheit</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$k_o$</td>
<td>attenuation coefficient of the canopy beam radiation extinction coefficient of the canopy</td>
<td>Strahlungsextrinktionskoeffizient des Pflanzenbestandes</td>
<td>dimensionless</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
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<tr>
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<td>------------------------------------------------------------------------------</td>
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</tr>
<tr>
<td>$k'_b$</td>
<td>beam and scattered beam PAR extinction coefficient of the canopy</td>
<td>PAR-Strahlungsextinktionskoeffizient des Pflanzenbestandes</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$k_{b,90^\circ}$</td>
<td>attenuation coefficient of the canopy for solar elevation of 90°</td>
<td>Strahlungsextinktionskoeffizient des Pflanzenbestandes bei einer Sonnenhöhe von 90°</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$k_{b,max}$</td>
<td>attenuation coefficient of the canopy at 12 h TST</td>
<td>Strahlungsextinktionskoeffizient des Pflanzenbestandes um 12 Uhr wahrer Ortszeit</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$k_{b,max, summer solstice}$</td>
<td>attenuation coefficient of the canopy at 12 h TST on summer solstice</td>
<td>Strahlungsextinktionskoeffizient des Pflanzenbestandes um 12 Uhr wahrer Ortszeit am Tag der Sommersonnenwende</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$K_c$</td>
<td>Michaelis-Menten constant of Rubisco for CO₂</td>
<td>Michaelis-Menten-Konstante der Rubisco für CO₂</td>
<td>Pa</td>
</tr>
<tr>
<td>$k_d$</td>
<td>diffuse PAR extinction coefficient</td>
<td>Extinktionskoeffizient für diffuse photosynthetisch-aktive Strahlung</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$k'_d$</td>
<td>diffuse and scattered diffuse PAR extinction coefficient</td>
<td>Extinktionskoeffizient für diffuse und gestreute photosynthetisch-aktive Strahlung</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$k_n$</td>
<td>coefficient of leaf-nitrogen allocation in a canopy</td>
<td>Blatt-N-Allokationskoeffizient im Bestand</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$K_o$</td>
<td>Michaelis-Menten constant of Rubisco for O₂</td>
<td>Michaelis-Menten-Konstante der Rubisco für O₂</td>
<td>Pa</td>
</tr>
<tr>
<td>L</td>
<td>Monin-Obukhov length</td>
<td>Monin-Obukhov-Länge</td>
<td>m</td>
</tr>
<tr>
<td>LAI</td>
<td>one-sided leaf area index</td>
<td>einseitiger Blattflächenindex</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>LAI leaf, literature</td>
<td>leaf area index described in the literature</td>
<td>in der Literatur dokumentierter Blattflächenindex</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>LAI max</td>
<td>maximum leaf area index</td>
<td>maximaler Blattflächenindex</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>LAI non-senescent</td>
<td>one-sided leaf area index of non-senescent (green) leaves</td>
<td>einseitiger Blattflächenindex nicht-seneszent (grüner) Blätter</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>LAI shaded</td>
<td>one-sided leaf area index of the shaded leaf fraction of the canopy</td>
<td>einseitiger Blattflächenindex der Schattenblätter des Bestandes</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>LAI sunlit</td>
<td>one-sided leaf area index of the sunlit leaf fraction of the canopy</td>
<td>einseitiger Blattflächenindex der Sonnenblätter des Bestandes</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>LAI total</td>
<td>one-sided leaf area index of non-senescent and senescent leaves</td>
<td>einseitiger Blattflächenindex nicht-seneszent und seneszenter Blätter</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>$L_d$</td>
<td>flux density of downward long-wave radiation of the atmosphere</td>
<td>Flussdichte der langwelligen Gegenstrahlung der Atmosphäre</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>$L_{neutral}$</td>
<td>Monin-Obukhov length under neutral atmospheric stability conditions</td>
<td>Monin-Obukhov-Länge bei neutral geschichteter Atmosphäre</td>
<td>m</td>
</tr>
<tr>
<td>$L_u$</td>
<td>flux density of upward long-wave radiation of the atmosphere</td>
<td>Flussdichte der langwelligen, &quot;emittierten&quot; Ausstrahlung der Erd- bzw. Bestandesoberfläche</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>$M_A$</td>
<td>molar mass of gaseous species A</td>
<td>Molekulargewicht des Spurengases A</td>
<td>g mol⁻¹</td>
</tr>
<tr>
<td>$N_S$</td>
<td>nitrogen content of the aboveground sapwood biomass</td>
<td>N-Gehalt des oberirdischen Saftholzes</td>
<td>g m⁻²</td>
</tr>
<tr>
<td>O_i</td>
<td>intercellulat O₂ partial pressure</td>
<td>O₂-Partialdruck in den Interzellularen</td>
<td>PA</td>
</tr>
<tr>
<td>P</td>
<td>air pressure</td>
<td>Luftdruck</td>
<td>hPa</td>
</tr>
<tr>
<td>P₀</td>
<td>standard atmospheric pressure</td>
<td>Standardluftdruck</td>
<td>hPa</td>
</tr>
<tr>
<td>P₀ CO₂</td>
<td>CO₂ partial pressure</td>
<td>CO₂-Partialdruck</td>
<td>Pa</td>
</tr>
<tr>
<td>Pᵢ CO₂</td>
<td>intercelluläre CO₂ partial pressure</td>
<td>CO₂-Partialdruck in den Interzellularen</td>
<td>Pa</td>
</tr>
<tr>
<td>PAD(A)</td>
<td>pollutant absorbed dose of trace gas A</td>
<td>absorbierte Dosis des Spurengases A</td>
<td>µg m⁻³</td>
</tr>
<tr>
<td>PAR</td>
<td>photosynthetically active irradiation</td>
<td>photosynthetisch-aktive Strahlung</td>
<td>µmol m⁻² s⁻¹</td>
</tr>
<tr>
<td>$P_h$</td>
<td>energy exchange due to photosynthesis and respiration of the aboveground biomass</td>
<td>Energiefluss aufgrund von Photosynthese und Respiration der oberirdischen Bimasse</td>
<td>W m⁻²</td>
</tr>
<tr>
<td>PLA</td>
<td>projected leaf area (= LAI)</td>
<td>projizierte Blattfläche (= LAI)</td>
<td>m² m⁻²</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
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<tr>
<td>Pr</td>
<td>Prandtl number</td>
<td>Prandtl-Zahl</td>
<td>dimensionless</td>
</tr>
<tr>
<td>Precip</td>
<td>precipitation</td>
<td>Niederschlag</td>
<td>mm</td>
</tr>
<tr>
<td>q</td>
<td>specific air humidity</td>
<td>spezifische Luftfeuchte</td>
<td>g·g⁻¹</td>
</tr>
<tr>
<td>Q₁₀</td>
<td>relative change of a physiological process per 10 K temperature increase</td>
<td>relative Veränderung eines physiologischen Prozesses bei einer Temperaturerhöhung um 10 K</td>
<td>dimensionless</td>
</tr>
<tr>
<td>R</td>
<td>universal gas constant</td>
<td>universelle Gaskonstante</td>
<td>J·mol⁻¹·K⁻¹</td>
</tr>
<tr>
<td>Rₛ, atm, H₂O</td>
<td>bulk stomatal resistance for water vapour for a given canopy development stage</td>
<td>Bulk-Stomata-Widerstand für Wasserdampf bei einem gegebenen Entwicklungsstand des Bestandes</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ, atm, CO₂</td>
<td>CO₂ bulk stomatal resistance for a given canopy development stage</td>
<td>CO₂-Bulk-Stomata-Widerstand für ein gegebenes Entwicklungsstand des Bestandes</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ, atm, O₃</td>
<td>O₂ bulk stomatal resistance for a given canopy development stage</td>
<td>O₂-Bulk-Stomata-Widerstand für ein gegebenes Entwicklungsstand des Bestandes</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ, atm, shaded, CO₂</td>
<td>CO₂ bulk stomatal resistance of the shaded leaf area fraction for a given canopy development stage</td>
<td>CO₂-Bulk-Stomata-Widerstand der Schattenblätter für ein gegebenes Entwicklungsstand des Bestandes</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ, atm, shaded, O₃</td>
<td>O₂ bulk stomatal resistance of the shaded leaf area fraction for a given canopy development stage</td>
<td>O₂-Bulk-Stomata-Widerstand der Schattenblätter für ein gegebenes Entwicklungsstand des Bestandes</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ, atm, sunlit, CO₂</td>
<td>CO₂ bulk stomatal resistance of the sunlit leaf area fraction for a given canopy development stage</td>
<td>CO₂-Bulk-Stomata-Widerstand der Sonnenblätter für ein gegebenes Entwicklungsstand des Bestandes</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ, atm, sunlit, O₃</td>
<td>O₂ bulk stomatal resistance of the sunlit leaf area fraction for a given canopy development stage</td>
<td>O₂-Bulk-Stomata-Widerstand der Sonnenblätter für ein gegebenes Entwicklungsstand des Bestandes</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ(zₛ, A, dzm)</td>
<td>turbulent atmospheric resistance between the reference height zₛ and the momentum sink height z = dzm</td>
<td>turbulenter atmosphärischer Transport-säulenwiderstand zwischen Messhöhe 𝑧ₛ und dem Niveau der Impulssenke 𝑧 = 𝑑𝑧ₘ</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛₚ, forest</td>
<td>turbulent atmospheric resistance above a forest</td>
<td>turbulenter atmosphärischer Transport-säulenwiderstand über Wald</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛₚ</td>
<td>turbulent atmospheric resistance</td>
<td>turbulenter atmosphärischer Transport-säulenwiderstand</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, A</td>
<td>bulk canopy or surface resistance for water vapour or other gaseous species</td>
<td>Gesamtheit der Bulk-Transport- und Bulk-Reaktionswiderstände des Systems Vegetation Boden für Wasserdampf oder andere Spurengase</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, cut</td>
<td>bulk cuticle resistance</td>
<td>Bulk-Cuticular-Widerstand</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, cut, A</td>
<td>bulk cuticle resistance for water vapour or other gaseous species</td>
<td>Bulk-Cuticular-Widerstand für Wasserdampf oder andere Spurengase</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, ext</td>
<td>bulk external plant surface resistance</td>
<td>Bulk-Widerstand für die Reaktion an Pflanzenoberflächen</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, ext, A</td>
<td>bulk external plant surface resistance for a gaseous species A</td>
<td>Bulk-Widerstand für die Reaktion der Spurengasspezies A an Pflanzenoberflächen</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, ext, dry, A</td>
<td>bulk external plant surface resistance for a gaseous species A and dry external plant surfaces</td>
<td>Bulk-Widerstand für die Reaktion der Spurengasspezies A an trockenen Pflanzenoberflächen</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, ext, wet, A</td>
<td>bulk external plant surface resistance for a gaseous species A and wet external plant surfaces</td>
<td>Bulk-Widerstand für die Reaktion der Spurengasspezies A an feuchten Pflanzenoberflächen</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, mes</td>
<td>bulk mesophyll resistance</td>
<td>Bulk-Mesophyllwiderstand</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, mes, A</td>
<td>bulk mesophyll resistance for a gaseous species A</td>
<td>Bulk-Mesophyllwiderstand für Spurengasspezies A</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Rₛ₅, non-stomatal, dry, O₃</td>
<td>non-stomatal resistance for O₃ and dry external plant surfaces</td>
<td>nicht-stomatärer Widerstand für O₃ an trockenen Pflanzenoberflächen</td>
<td>s·m⁻¹</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
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</tr>
<tr>
<td>$R_{c, \text{ stom}}$</td>
<td>bulk stomatal resistance</td>
<td>Bulk-Stomata-Widerstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{c, \text{ stom}, A}$</td>
<td>bulk stomatal resistance for water vapour or other gaseous species</td>
<td>Bulk-Stomata-Widerstand für Wasserdampf oder andere Spurengase</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{c, \text{ stom}, \text{ min}, \text{ H}_2\text{O}}$</td>
<td>minimum value of bulk stomatal resistance for water vapour</td>
<td>minimaler Bulk-Stomata-Widerstand für Wasserdampf</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{c, \text{ stom}, \text{ max}, \text{ O}_3}$</td>
<td>$R_{c, \text{ stom}, \text{ O}<em>3} + R</em>{c, \text{ max}, \text{ O}_3}$</td>
<td>Widerstand für einen bestimmten Entwicklungsstand des Bestandes</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{c, \text{ stomatal}, \text{ H}_2\text{O}}$</td>
<td>bulk stomatal resistance for water vapour for a given stage of canopy development</td>
<td>Gesamtheit der Bulk-Transport- und Bulk-Reaktionswiderstände des Systems Vegetation Boden</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{canopy}, R_{c}}$</td>
<td>bulk canopy or surface resistance</td>
<td>Gaskonstante für trockene Luft</td>
<td>J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{dry, air}}$</td>
<td>gas constant for dry air</td>
<td>Gaskonstante für trockene Luft</td>
<td>J kg$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>$\text{rH}$</td>
<td>relative humidity</td>
<td>relative Luftfeuchte</td>
<td>%</td>
</tr>
<tr>
<td>$R_{\text{leaf}}$</td>
<td>leaf resistance</td>
<td>Blattwiderstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{leaf, cut}, A}$</td>
<td>leaf cuticle resistance for water vapour or other gaseous species</td>
<td>Blatt-Cuticular-Widerstand für Wasserdampf oder andere Spurengase</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{leaf, ext, dry}, A}$</td>
<td>leaf external plant surface resistance for a gaseous species A and dry external plant surfaces</td>
<td>Blattwiderstand für die Reaktion der Spurengasspezies A an trockenen Pflanzenoberflächen</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{leaf, literature}}$</td>
<td>resistance derived on leaf basis described in the literature</td>
<td>in der Literatur dokumentierter Blatt-Widerstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{low}}$</td>
<td>low temperature resistance</td>
<td>Widerstand bei niedriger Temperatur</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{molecular}}$</td>
<td>molecular atmospheric resistance</td>
<td>molekularer atmosphärischer Widerstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{net}}$</td>
<td>net radiation balance</td>
<td>Flussdichte der Gesamtstrahlungsbilanz</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$R_{\text{quasi-laminar layer, A}}$</td>
<td>quasi-laminar layer resistance for sensible heat, water vapour or other gaseous species</td>
<td>Transportsäulenwiderstand der quasi-laminaren Schicht für fühlbare Wärme, Wasserdampf oder andere Spurengase</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{soil}}$</td>
<td>soil resistance</td>
<td>Widerstand der Bodenoberfläche</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{soil, A}}$</td>
<td>soil resistance for water vapour or other gaseous species</td>
<td>Widerstand der Bodenoberfläche für Wasserdampf oder andere Spurengase</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{soil, dry, A}}$</td>
<td>soil resistance for gaseous species A under &quot;dry conditions&quot;</td>
<td>Widerstand der Bodenoberfläche für ein Spurengas A unter &quot;trockenen Bedingungen&quot;</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{soil, dry}, \text{ H}_2\text{O}, \text{ min}}$</td>
<td>minimum value of soil resistance for water vapour</td>
<td>minimaler Widerstand der Bodenoberfläche für Wasserdampf</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{soil, dry}, \text{ min}}$</td>
<td>minimum value of soil resistance</td>
<td>minimaler Widerstand der Bodenoberfläche</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{turbulent}}$</td>
<td>turbulent atmospheric resistance</td>
<td>turbulenter atmosphärischer Widerstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{rx, cut}}$</td>
<td>fraction of $R_{\text{soil, dry}, \text{ H}_2\text{O}, \text{ min}}$</td>
<td>Bruchteil von $R_{\text{soil, dry}, \text{ H}_2\text{O}, \text{ min}}$</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$R_{\text{rx, non-stomatal}}$</td>
<td>bulk or leaf cuticle resistance</td>
<td>Bulk- oder Blatt-Cuticulawiderstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{rx, stom} &amp; \text{ cut}}$</td>
<td>bulk or leaf non-stomatal resistance</td>
<td>nicht-stomatärer Bulk- oder Blatt-Widerstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{rx, stom}, R_{\text{rx, stomatal}}}$</td>
<td>combined bulk or leaf stomatal and cuticle resistance</td>
<td>kombinierter Bulk- oder Blatt-Stomata- und Cuticulawiderstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{rx, y, dry}}$</td>
<td>bulk stomatal resistance</td>
<td>Bulk- oder Blatt-Stomatawiderstand</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_{\text{rx, y, wet}}$</td>
<td>resistance of a dry surface</td>
<td>Widerstand trockener Oberflächen</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$\text{Resp}$</td>
<td>sum of belowground respiration and the respiration rate of the aboveground woody plant parts</td>
<td>Summe der Bodenatmung und Atmung der verholzten oberirdischen Pflanzenteile</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$\text{Resp}_{\text{belowground}}$</td>
<td>belowground respiration</td>
<td>Bodenatmung</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$\text{Resp}_{\text{d}}$</td>
<td>daytime mitochondrial respiration rate of the canopy</td>
<td>mitochondriale Bestandesatmung bei Tage</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$\text{Resp}_{\text{n, nighttime}}$</td>
<td>nighttime ecosystem respiration rate</td>
<td>nächtliche ökosystemare Atmung</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
</tr>
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<td>-----------------------------------------------------------------------------</td>
<td>---------------------------</td>
</tr>
<tr>
<td>Resp(r)</td>
<td>cumulative daily maintenance respiratory flux from roots</td>
<td>kumulierte tägliche Erhaltungsatmung der Wurzeln</td>
<td>(\mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1})</td>
</tr>
<tr>
<td>Resp(w)</td>
<td>respiration rate of the aboveground woody plant parts</td>
<td>Atmung der verholzten oberirdischen Pflanztfeile</td>
<td>(\mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1})</td>
</tr>
<tr>
<td>Resp(w)(_g)</td>
<td>aboveground woody maintenance respiration due to construction of new tissue</td>
<td>Erhaltungsatmung der verholzten oberirdischen Pflanztfeile aufgrund der Bildung von neuem Gewebe</td>
<td>(\mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1})</td>
</tr>
<tr>
<td>Resp(w)(_m)</td>
<td>aboveground woody maintenance respiration</td>
<td>Erhaltungsatmung der verholzten oberirdischen Pflanztfeile</td>
<td>(\mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1})</td>
</tr>
<tr>
<td>Resp(w)(_m)</td>
<td>cumulative daily maintenance respiratory flux from aboveground woody biomass</td>
<td>kumulierte tägliche Erhaltungsatmung der verholzten oberirdischen Pflanztfeile</td>
<td>(\mu\text{mol} \cdot \text{m}^{-2} \cdot \text{s}^{-1})</td>
</tr>
<tr>
<td>RLD(_{90})</td>
<td>depth of the soil layer containing 90 % of total root mass</td>
<td>Mächtigkeit der Bodenschicht, die 90 % der Wurzelbiomasse enthält</td>
<td>m</td>
</tr>
<tr>
<td>RY</td>
<td>fraction of (R_{\text{soil, H2O, min}})</td>
<td>Bruchteil von (R_{\text{soil, H2O, min}})</td>
<td>dimensionless</td>
</tr>
<tr>
<td>SAI</td>
<td>total surface area index of the vegetation</td>
<td>Oberflächenindex der Vegetation</td>
<td>m(^2) \cdot \text{m}^{-3})</td>
</tr>
<tr>
<td>(s_c)</td>
<td>difference quotient (\Delta e_{\text{sat}}/\Delta T)</td>
<td>Differenzenquotient (\Delta e_{\text{sat}}/\Delta T)</td>
<td>hPa K(^{-1})</td>
</tr>
<tr>
<td>(\text{ScA})</td>
<td>Schmidt number of a trace gas A</td>
<td>Schmidt-Zahl der Spurengasspezies A</td>
<td>dimensionless</td>
</tr>
<tr>
<td>(S_{\text{in-canopy, air}})</td>
<td>flux of energy due to changes in temperature and humidity of the air in the canopy</td>
<td>fühlbarer Wärmestrom im Bestandesraum</td>
<td>W \cdot \text{m}^{-2})</td>
</tr>
<tr>
<td>SM</td>
<td>soil moisture</td>
<td>volumetrischer Bodenwassergehalt</td>
<td>m(^3) \cdot \text{m}^{-3})</td>
</tr>
<tr>
<td>SM(_f)</td>
<td>field capacity</td>
<td>Feldkapazität</td>
<td>m(^3) \cdot \text{m}^{-3})</td>
</tr>
<tr>
<td>SM(_{res})</td>
<td>residual (irreducible) soil moisture</td>
<td>absolutes Minimum des Bodenwassergehaltes</td>
<td>m(^3) \cdot \text{m}^{-3})</td>
</tr>
<tr>
<td>SM(_{sat})</td>
<td>soil moisture at saturation</td>
<td>Bodenwassergehalt bei Sättigung</td>
<td>m(^3) \cdot \text{m}^{-3})</td>
</tr>
<tr>
<td>SM(_{w})</td>
<td>wilting point</td>
<td>perennialer Welkpunkt</td>
<td>m(^3) \cdot \text{m}^{-3})</td>
</tr>
<tr>
<td>S(_g)</td>
<td>global radiation</td>
<td>Globalstrahlung</td>
<td>W \cdot \text{m}^{-2})</td>
</tr>
<tr>
<td>S(_{\text{ref}})</td>
<td>astronomical maximum possible global radiation at cloudless sky</td>
<td>astronomisch maximal mögliche Globalstrahlung bei wolkenlosem Himmel</td>
<td>W \cdot \text{m}^{-2})</td>
</tr>
<tr>
<td>S(_1)</td>
<td>maximum global radiation for the Jarvis radiation function</td>
<td>maximal mögliche Globalstrahlung in der Jarvis-Funktion für Strahlung</td>
<td>W \cdot \text{m}^{-2})</td>
</tr>
<tr>
<td>S(_2)</td>
<td>empirical coefficient governing the shape of the Jarvis radiation function</td>
<td>Koeffizient, der die Kurvenform der Jarvis-Funktion für Strahlung bestimmt</td>
<td>W \cdot \text{m}^{-2})</td>
</tr>
<tr>
<td>subscript c</td>
<td>canopy</td>
<td>Bestand</td>
<td>---</td>
</tr>
<tr>
<td>subscript shaded</td>
<td>shaded leaf fraction of the canopy</td>
<td>Fraktion der Schattenblätter eines Bestandes</td>
<td>---</td>
</tr>
<tr>
<td>subscript sunlit</td>
<td>sunlit leaf fraction of the canopy</td>
<td>Fraktion der Sonnenblätter eines Bestandes</td>
<td>---</td>
</tr>
<tr>
<td>S(_{\text{vegetation}})</td>
<td>flux of energy due to changes in temperature of the aboveground biomass</td>
<td>Wärmespeicherung in der Biomasse</td>
<td>W \cdot \text{m}^{-2})</td>
</tr>
<tr>
<td>SLD</td>
<td>depth of soil layer</td>
<td>Mächtigkeit der Bodenschicht</td>
<td>m</td>
</tr>
<tr>
<td>(t)</td>
<td>time</td>
<td>Zeit</td>
<td>s, min, h, d, a</td>
</tr>
<tr>
<td>(T)</td>
<td>absolute air temperature</td>
<td>absolute Lufttemperatur</td>
<td>K</td>
</tr>
<tr>
<td>(T_0)</td>
<td>standard temperature</td>
<td>Standardtemperatur</td>
<td>K</td>
</tr>
<tr>
<td>(t(_a))</td>
<td>actual air temperature</td>
<td>aktuelle Lufttemperatur</td>
<td>°C</td>
</tr>
<tr>
<td>(t(_1))</td>
<td>vegetation-type dependent minimum temperature of the Jarvis temperature function</td>
<td>vegetationsspezifische Minimumtemperatur in der Jarvis-Funktion für Temperatur</td>
<td>°C</td>
</tr>
<tr>
<td>(t(_2))</td>
<td>vegetation-type dependent optimum temperature of the Jarvis temperature function</td>
<td>vegetationsspezifische Optimumtemperatur in der Jarvis-Funktion für Temperatur</td>
<td>°C</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
</tr>
<tr>
<td>--------</td>
<td>-------------</td>
<td>-------------</td>
<td>--------------</td>
</tr>
<tr>
<td>$t_s$</td>
<td>vegetation-type dependent maximum temperature of the Jarvis temperature function at which stomata no longer remain open</td>
<td>vegetationsspezifische Maximumtemperatur in der Jarvis-Funktion für Temperatur, bei der die Stomata nicht länger geöffnet sind</td>
<td>°C</td>
</tr>
<tr>
<td>$t_{3\text{days}}$</td>
<td>mean daily air temperature of the past 7 days</td>
<td>mittlere Lufttemperatur der letzten 7 Tage</td>
<td>°C</td>
</tr>
<tr>
<td>$t_{lag3h}$</td>
<td>ambient air temperature lagged by three hours</td>
<td>aktuelle Lufttemperatur vor drei Stunden</td>
<td>°C</td>
</tr>
<tr>
<td>$t_{opt}$</td>
<td>optimum soil temperature for CO$_2$ evolution</td>
<td>optimale Bodentemperatur für die Bodenatmung</td>
<td>°C</td>
</tr>
<tr>
<td>$T_p$</td>
<td>rate of triose phosphate export from the chloroplast</td>
<td>Rate des Triosephosphatexports aus den Chloroplasten</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$T_s$</td>
<td>absolute canopy surface temperature</td>
<td>absolute Bestandesoberflächentemperatur</td>
<td>K</td>
</tr>
<tr>
<td>$t_a$</td>
<td>surface temperature</td>
<td>aktuelle Oberflächentemperatur</td>
<td>°C</td>
</tr>
<tr>
<td>$t_{soil}$</td>
<td>soil temperature</td>
<td>Bodentemperatur</td>
<td>°C</td>
</tr>
<tr>
<td>TST</td>
<td>True Solar time</td>
<td>wahre Sonnenzeit</td>
<td>h</td>
</tr>
<tr>
<td>$u$</td>
<td>horizontal wind velocity</td>
<td>horizontale Windgeschwindigkeit</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$u_h$</td>
<td>horizontal wind velocity at canopy height</td>
<td>horizontale Windgeschwindigkeit an der Bestandesoberfläche</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$u_r$</td>
<td>friction velocity</td>
<td>Schubspannungsgeschwindigkeit</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$V_c$</td>
<td>photosynthetic Rubisco capacity of the canopy</td>
<td>Rubisco-Kapazität des Bestandes</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$v_0$</td>
<td>dry deposition velocity</td>
<td>Depositionsgeschwindigkeit</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$V_{l,\text{at top of canopy}}$</td>
<td>photosynthetic Rubisco capacity per unit leaf area at top of canopy</td>
<td>Rubisco-Kapazität pro Blattflächeneinheit an der Bestandesoberfläche</td>
<td>µmol m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$V_m$</td>
<td>molar volume</td>
<td>Molvolumen</td>
<td>l mol$^{-1}$</td>
</tr>
<tr>
<td>VPD</td>
<td>water vapour pressure deficit of the atmosphere</td>
<td>Wasserdampfsättigungsdefizit der Atmosphäre</td>
<td>hPa</td>
</tr>
<tr>
<td>$V_1$</td>
<td>maximum value of VPD Jarvis function</td>
<td>maximaler Wert für die VPD-Jarvis-Funktion</td>
<td>hPa</td>
</tr>
<tr>
<td>$V_2$</td>
<td>threshold value of VPD Jarvis function</td>
<td>Schwellenwert für die VPD-Jarvis-Funktion</td>
<td>hPa</td>
</tr>
<tr>
<td>$V_3$</td>
<td>minimum threshold of VPD Jarvis function</td>
<td>minimaler Schwellenwert für die VPD-Jarvis-Funktion</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$V_4$</td>
<td>empirical weighting coefficient of VPD Jarvis function</td>
<td>Wichtungsfaktor für die VPD-Bodenfeuchte-Jarvis-Funktion</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$W_1$</td>
<td>site-specific threshold level of the Jarvis soil moisture function</td>
<td>standortspezifischer Schwellenwert in der Jarvis-Funktion für Bodenfeuchte</td>
<td>m$^{-3}$ m$^{-3}$</td>
</tr>
<tr>
<td>$W_2$</td>
<td>minimum threshold of the Jarvis soil moisture function</td>
<td>minimaler Schwellenwert in der Jarvis-Funktion für Bodenfeuchte</td>
<td>dimensionless</td>
</tr>
<tr>
<td>$W_{ni}$</td>
<td>precipitation and/or dew reaching the ground</td>
<td>Niederschlag und Tau, der den Boden erreicht</td>
<td>mm</td>
</tr>
<tr>
<td>$z$</td>
<td>height above ground</td>
<td>Höhe über Grund</td>
<td>m</td>
</tr>
<tr>
<td>$z_{0,\text{scalar}}$</td>
<td>roughness length for a scalar (sensible heat, gases)</td>
<td>Rauhigkeitslänge für Skalar (fühlbare Wärme, Gase)</td>
<td>m</td>
</tr>
<tr>
<td>$z_{0c}$</td>
<td>roughness length for water vapour or other gaseous species</td>
<td>Rauhigkeitslänge für Wasserdampf oder andere Spurengase</td>
<td>m</td>
</tr>
<tr>
<td>$z_{0h}$</td>
<td>roughness length for sensible heat</td>
<td>Rauhigkeitslänge für fühlbare Wärme</td>
<td>m</td>
</tr>
<tr>
<td>$z_{0m}$</td>
<td>roughness length for momentum</td>
<td>Rauhigkeitslänge für Impuls</td>
<td>m</td>
</tr>
<tr>
<td>$z_{\text{ref}}$</td>
<td>reference height above the ground</td>
<td>Messhöhe über Grund</td>
<td>m</td>
</tr>
<tr>
<td>$z_{\text{ref, A}}$</td>
<td>reference height for a trace gas A</td>
<td>Messhöhe der Konzentration der Spurengasspezies A</td>
<td>m</td>
</tr>
<tr>
<td>$z_{\text{ref, p}}$</td>
<td>reference height for air pressure</td>
<td>Messhöhe des Luftdrucks</td>
<td>m</td>
</tr>
<tr>
<td>$z_{\text{ref, rh}}$</td>
<td>reference height for relative humidity</td>
<td>Messhöhe der relativen Luftfeuchte</td>
<td>m</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Bezeichnung</td>
<td>unit/Einheit</td>
</tr>
<tr>
<td>--------</td>
<td>-------------</td>
<td>-------------</td>
<td>--------------</td>
</tr>
<tr>
<td>$z_{ed, T}$</td>
<td>reference height for actual air temperature</td>
<td>Messhöhe der aktuellen Lufttemperatur</td>
<td>m</td>
</tr>
<tr>
<td>$z_{ed, u}$</td>
<td>reference height for horizontal wind velocity</td>
<td>Messhöhe der horizontalen Windgeschwindigkeit</td>
<td>m</td>
</tr>
</tbody>
</table>
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