



Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

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Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman¹, R. P. Bartholomeus¹, S. E. A. T. M. van der Zee²,
M. F. P. Bierkens^{3,4}, and J. P. M. Witte^{1,5}

¹KWR Watercycle Research Institute, P.O. Box 1072, 3430 BB Nieuwegein, the Netherlands

²Soil Physics and Land Management, Environmental Sciences Group, Wageningen University, P.O. Box 47, 6700 AA Wageningen, the Netherlands

³Department of Physical Geography, Faculty of Geosciences, Utrecht University, P.O. Box 80115, 3508 TC Utrecht, the Netherlands

⁴Deltares, P.O. Box 85467, 3508 AL Utrecht, the Netherlands

⁵VU University, Institute of Ecological Science, Department of Systems Ecology, de Boelelaan 1085, 1081 HV Amsterdam, the Netherlands

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Correspondence to: B. R. Voortman (bernard.voortman@kwrwater.nl)

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Title Page

Abstract Introduction

Conclusions References

Tables Figures

◀ ▶

◀ ▶

Back Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Abstract

Coastal and inland dunes provide various ecosystem services that are related to groundwater, such as drinking water production and biodiversity. To manage groundwater in a sustainable manner, knowledge of actual evapotranspiration (ET_a) for the various land covers in dunes is essential. Aiming at improving the parameterization of dune vegetation in hydro-meteorological models, this study explores the magnitude of energy and water fluxes in an inland dune ecosystem in the Netherlands. Hydro-meteorological measurements were used to parameterize the Penman–Monteith evapotranspiration model for four different surfaces: bare sand, moss, grass and heather. We found that the net longwave radiation (R_{nl}) was the largest energy flux for most surfaces during daytime. However, modelling this flux by a calibrated FAO-56 R_{nl} model for each surface and for hourly time steps was unsuccessful. Our R_{nl} model, with a novel sub-model using solar elevation angle and air temperature to describe the diurnal pattern in radiative surface temperature, improved R_{nl} simulations considerably. Model simulations of evaporation from moss surfaces showed that the modulating effect of mosses on the water balance is species dependent. We demonstrate that dense moss carpets (*Campylopus introflexus*) evaporate more (5 %, +14 mm) than bare sand (total of 258 mm in 2013), while more open structured mosses (*Hypnum cupressiforme*) evaporate less (–30 %, –76 mm) than bare sand. Additionally, we found that a drought event in the summer of 2013 showed a pronounced delayed signal on lysimeter measurements of ET_a for the grass and heather surfaces respectively. Due to the desiccation of leaves after the drought event, and their feedback on the parameters of the Penman–Monteith equation, the potential evapotranspiration in the year 2013 dropped with 9 % (–37mm) and 10 % (–61 mm) for the grass and heather surfaces respectively, which subsequently led to lowered ET_a of 8 % (–29 mm) and 7 % (–29 mm). These feedbacks are of importance to water resources, especially during a changing climate with increasing number of drought days. Therefore, such feedbacks need to be integrated into a coupled plant physiological and hydro-meteorological model to accurately

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



simulate ET_a . In addition, our study showed that groundwater recharge in dunes can be increased considerably by promoting moss vegetation, especially of open structured moss species.

1 Introduction

Coastal and inland sand dunes are major drinking water production sites in the Netherlands. Approximately 23% of Dutch drinking water originates from aquifers in these dunes, which are replenished by both natural groundwater recharge and artificial infiltration of surface waters. Another ecosystem service of groundwater in dune systems is that shallow groundwater tables sustain nature targets with a very high conservation value. Such targets, like wet dune slacks and oligotrophic pools, are often legally enforced, e.g. by the European Habitat Directive and by the Water Framework Directive. Furthermore, a deep layer of fresh groundwater in coastal dunes protects the hinterland from the inflow of saline groundwater.

Under a warming climate, summers are expected to become dryer and the water quality of surface waters may degrade (Delpla et al., 2009), especially during dry periods with low river discharge rates (Zwolsman and van Bokhoven, 2007; van Vliet and Zwolsman, 2008). To maintain current drinking water quality and production costs, water production in the future may have to rely more on natural groundwater recharge. This implies that drinking water companies need to search for new water production sites or intensify current groundwater extractions, while protecting groundwater dependent nature targets.

For sustainable management of renewable groundwater resources, groundwater extractions should be balanced with the amount of precipitation that percolates to the saturated zone, the groundwater recharge. Knowledge of actual evapotranspiration (ET_a , here defined as the sum of plant transpiration, soil evaporation, and evaporation from canopy interception) for the various land covers is essential to quantify the amount of recharge. Inland dune systems are predominantly covered with deciduous and pine

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



where ET_p is the potential evapotranspiration [mm s^{-1}], Δ is the slope of the saturation vapor pressure vs. temperature curve [$\text{kPa } ^\circ\text{C}^{-1}$], R_n is the net radiation [J m^{-2}], G is the soil heat flux [J m^{-2}], ρ_a is the air density [kg m^{-3}], c_p is specific heat of moist air [$\text{J kg}^{-1} ^\circ\text{C}^{-1}$], e_s is the saturation vapor pressure of the air [kPa], e_a is the actual vapor pressure of the air [kPa], r_a is aerodynamic resistance to turbulent heat and vapor transfer [s m^{-1}], γ is the psychrometric constant [$\text{kPa } ^\circ\text{C}^{-1}$], λ is the latent heat of vaporization [J kg^{-1}], and ρ_w is the density of liquid water [kg m^{-3}]. Results of Irmak et al. (2005) suggest that estimates of ET_p on hourly time steps are more accurate than estimates on a daily timescale. Furthermore, Liu et al. (2005) showed that the use of daily input values leads to a systematic overestimation of ET_a , especially for sandy soils. Hence, energy fluxes in the Penman–Monteith equation are preferably simulated at sub-diurnal timescales. Furthermore, understanding and simulation of plant physiological processes requires knowledge of the diurnal variation of environmental variables (Nozue and Maloof, 2006). Therefore, field data was aggregated to hourly time steps to maintain the diurnal pattern and to analyze our field results at the same time interval as commonly available climate data.

In this paper evapotranspiration is defined as the sum of transpiration, soil evaporation, and evaporation from canopy interception, expressed in mm per time unit. Radiative and soil heat fluxes are expressed in W m^{-2} . Figure 1 shows the procedures followed to translate field data (Sect. 2.1) to sub-models of the Penman–Monteith equation (Sect. 2.2) and to subsequently calculate ET_p and simulate ET_a (Sect. 2.3).

2.2 Hydro-meteorological measurements

Four homogeneous sites of bare sand, moss (*Campylopus introflexus*), grass (*Agrostis vinealis*) and heather (*Calluna vulgaris*) were selected to measure actual evapotranspiration (ET_a), the net radiation (R_n), the soil heat flux (G) and the albedo. Other meteorological variables such as wind speed (u , at 2 m above the surface), relative humidity (RH, 1.5 m above the surface), air temperature (T_a , 1.5 m above the surface), and rain

2.3 Parameterization of the Penman–Monteith equation

2.3.1 Net radiation (R_n)

The net radiation (R_n) is defined as:

$$R_n = R_{ns} + R_{nl} = (1 - \text{albedo})R_{s\downarrow} + (\varepsilon_s R_{l\downarrow} - R_{l\uparrow}), \quad (2)$$

5 where R_{ns} is the net shortwave radiation, R_{nl} is the net longwave radiation, $R_{s\downarrow}$ is the incoming solar radiation, $R_{l\downarrow}$ is the downwelling longwave radiation from the atmosphere to the surface, $R_{l\uparrow}$ is the emitted longwave radiation by the surface into the atmosphere and ε_s is the surface emissivity representing the reflected downwelling longwave radiation. The albedo in Eq. (2) was determined by linear regression between measured
10 $R_{s\downarrow}$ and $R_{s\uparrow}$. Based on the albedo obtained this way, R_{nl} follows from measurements of R_n by subtracting calculated R_{ns} from measured R_n . Throughout this paper, this back-calculated R_{nl} is referred to as the measured R_{nl} .

In hydro-meteorological models, R_{nl} is commonly estimated under clear sky conditions and multiplied by a factor to correct for clouds (Irmak et al., 2010; Gubler et al.,
15 2012; Blonquist Jr et al., 2010; Temesgen et al., 2007). A similar approach was followed in this study in which the Stefan–Boltzmann law is substituted into Eq. (2) for $R_{l\downarrow}$ and $R_{l\uparrow}$ under clear sky conditions (Saito and Šimůnek, 2009; Van Bavel and Hillel, 1976) and multiplied by a cloudiness function to obtain R_{nl} :

$$R_{nl} = (\varepsilon_s \varepsilon_a \sigma T_a^4 - \varepsilon_s \sigma T_s^4) f_{cd}, \quad (3)$$

20 where ε_a is the clear sky emissivity of the atmosphere [–], ε_s is the surface emissivity [–], σ is the Stefan–Boltzmann constant ($5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-1}$), T_a is the air temperature [K], T_s is the surface temperature [K] and f_{cd} is a cloudiness function [–] (described later). For vegetated surfaces $\varepsilon_s = 0.95$ was used (based on Jones and Callow, 2004), and $\varepsilon_s = 0.925$ for bare sand (based on Fuchs and Tanner, 1968). Estimating ε_a has

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

[Title Page](#)

[Abstract](#)

[Introduction](#)

[Conclusions](#)

[References](#)

[Tables](#)

[Figures](#)

[⏪](#)

[⏩](#)

[◀](#)

[▶](#)

[Back](#)

[Close](#)

[Full Screen / Esc](#)

[Printer-friendly Version](#)

[Interactive Discussion](#)



by minimizing the root mean squared error (RMSE) by generalized reduced gradient nonlinear optimization. The $T_{s, \text{offset}}$ was determined as the average nighttime $T_s - T_a$ to limit the amount of parameters during the optimization. Eq. (6) was substituted for T_s in Eq. (3) to estimate R_{nl} . This novel approach to derive R_{nl} was compared to the R_{nl} model of the FAO-56 approach (Allen et al., 1998), originally derived to obtain daily estimates of R_{nl} (using minimum and maximum daily T_a divided by 2 instead of T_a in Eq. 7) but commonly applied at hourly timescales (ASCE-EWRI, 2005; Perera et al., 2015; Gavilán et al., 2008; López-Urrea et al., 2006):

$$R_{nl} = -\sigma T_a^4 (a - b \sqrt{e_a}) \left(1.35 \frac{R_s}{R_{s0}} - 0.35 \right), \quad (7)$$

where the first term between brackets represents the net emittance, which should compensate for the fact that T_s is not measured. The empirical parameters a and b can be calibrated for a specific climate and/or vegetation. The second term between brackets is a cloudiness function. The default parameter values for a and b are 0.34 and 0.14, respectively (Allen et al., 1998). We calibrated these parameters for every site by linear least squares regression for clear days ($R_s/R_{s0} > 0.9$) and compared the performance of both R_{nl} models (Eqs. 3 and 7).

2.3.2 Soil heat flux (G)

The soil heat flux is commonly expressed as a fraction of R_{nl} , particularly at large scales using remote sensing (Su, 2002; Bastiaanssen et al., 1998; Kustas et al., 1998; Kustas and Daughtry, 1990; Friedl, 1996). We adopted the same approach making a distinction between daytime (F_{day}) and nighttime (F_{night}) fractions, determined by linear least squares regression between R_{nl} and the average of the two sets of soil heat flux measurements.

2.3.3 Aerodynamic resistance (r_a)

The aerodynamic resistance under neutral stability conditions can be estimated by (Monteith and Unsworth, 1990):

$$r_a = \frac{\ln \left[\frac{z_m - d}{z_{om}} \right] \ln \left[\frac{z_h - d}{z_{oh}} \right]}{k^2 u_z}, \quad (8)$$

5 where z_m is the height of wind speed measurements [m], d is the zero plane displacement height [m], z_{om} is the roughness length governing momentum transfer [m], z_h is the height of the humidity measurements [m], z_{oh} is the roughness length governing transfer of heat and vapor [m], k is the von Karman's constant (0.41 [–]) and u_z is the wind speed at height z_m [m s^{-1}]. For grass, empirical equations are developed (FAO 56
10 approach) to estimate d , z_{om} and z_{oh} :

$$d = 0.66V, \quad (9)$$

$$z_{om} = 0.123V, \quad (10)$$

$$z_{oh} = 0.1z_{om}, \quad (11)$$

15 where V is the vegetation height. Wallace et al. (1984) found comparable coefficients for heather: $d = 0.63V$ and $z_{om} = 0.13V$ and therefore Eqs. (9)–(11) were applied for both surfaces using a constant vegetation height of 7 and 31 cm for the grass and heather surfaces respectively. For the moss surface, we used a vegetation height of 2 cm, which is equal to the thickness of the moss mat. For the bare sand surface we assumed $d = 0$ m, and used typical surface roughness values published by Oke (1978):
20 $z_{oh} = 0.001$ m and $z_{om} = z_{oh}$.

2.3.4 Surface resistance (r_s) and canopy interception

Canopy interception was simulated as a water storage which needs to be filled before rain water reaches the soil surface. A maximum storage capacity of 0.50 mm was de-

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



5 fined for heather following the study of Ladekarl et al. (2005). To our knowledge no literature value of the interception capacity of the specific grass species (*Agrostis vinealis*) is published. Considering the relatively low vegetation height we assumed a maximum interception capacity of 0.25 mm.

10 We distinguished wet (r_{swet}) and dry canopy surface resistance (r_s), since interception water evaporates without the interference of leaf stomata. During canopy interception (i.e. if the interception store is fully or partly filled) we used a surface resistance of 0 s m^{-1} , reducing Eq. (1) to the Penman equation (Penman, 1948; Monteith and Unsworth, 1990). After the canopy storage is emptied the surface resistance switches to r_s . The r_s was back-calculated for daytime periods for the heather and grass lysimeters by substituting measured R_n , G , ET_a , e_s and e_a and simulated r_a into Eq. (1) under non-stressed conditions (i.e. $ET_p = ET_a$). Nighttime evaporation was assumed to be equal to 0 mm. To make sure that the back-calculated r_s was based on days at which evapotranspiration occurred at a potential rate, it was back-calculated for every two consecutive days after precipitation events and after emptying of the (calculated) interception store. The surface resistance (r_s) of bare sand and moss was assumed to be equal to 10 s m^{-1} , i.e. similar to the surface resistance under well watered conditions of bare soil found by Van de Griend and Owe (1994).

20 During the summer of 2013, a dry spell (from 4 until 25 July 2013) resulted in a vegetation dieback of grass and heather. Surface resistances were back-calculated for periods before and after the drought event. The drought event had 22 consecutive dry days with a cumulative reference evapotranspiration according to Makkink (1957) of 85 mm. Drought events of similar magnitude have been recorded 12 times during the past 57 years (from 1958 until 2014) at climate station “de Bilt” located in the center of the Netherlands (52.1° latitude, 5.18° longitude). The measurements in the heather vegetation started a week before the drought event. During this week, there were two days (30 June 2013 and 1 July 2013) for which r_s could be back-calculated. The estimated r_s for these days were 35 and 107 s m^{-1} respectively. We selected the r_s value of the second day to use in our model simulations (107 s m^{-1}) because it was

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Soil hydraulic properties in the hydrological model were described by the Van Genuchten relationships (Van Genuchten, 1980). Soil samples (100 cm^3) collected next to each lysimeter at 5 and 15 cm depth were used to derive the drying retention function. The average drying retention parameters (of the two samples collected next to each lysimeter) were used in the hydrological model taking hysteresis into account by assuming the wetting retention curve parameter (α_{wet}) to be twice as large as the drying retention curve parameter (α_{dry}) (Šimůnek et al., 1999). The unsaturated hydraulic properties (parameters l and K_0) were estimated using the Rosetta database and pedotransfer functions, providing the fitted drying retention curve parameters as input (Schaap et al., 2001). The hydraulic properties of the 15 cm long wick, guiding drainage water below the lysimeter into the tipping bucket, were taken from Knutson and Selker (1994).

Since mosses have neither leaf stomata nor roots, ET_a from the moss surface is limited by the capacity of the moss material to conduct water to the surface. This passive evaporation process is similar to the process of soil evaporation, i.e. evaporation becomes limited if the surface becomes too dry to deliver the potential rate. The unsaturated hydraulic properties of the dense *Campylopus introflexus* moss mat covering the lysimeter soil were based on the hydraulic properties derived by Voortman et al. (2014) and used in the first 2 cm of the model domain. Macro pores in the moss mat were neglected by Voortman et al. (2014), which implies that direct implementation of these hydraulic properties would result in large amounts of surface runoff generation or ponding, since the unsaturated hydraulic conductivity (K_0) of the moss mat is lower than 0.28 cm day^{-1} . Therefore, the dual porosity model of Durner (1994) was used to add 1000 cm day^{-1} to the hydraulic conductivity curve of Voortman et al. (2014) between -1 and 0 cm pressure head (Appendix A). This permits the infiltration of rain water at high intensity rain showers without affecting the unsaturated hydraulic behavior at negative pressure heads. Because of the complex shape of the retention function of the moss mat, hysteresis in the soil hydraulic functions in the underlying soil was neglected for the simulation of evaporation from moss surfaces. The sensitivity of this simplification

3 Results and discussion

3.1 Parameterization of the Penman–Monteith equation

3.1.1 Net shortwave radiation

The measured incoming and reflected solar radiation were used to compute the albedo of the four surfaces by linear regression (Fig. 5; Table 5). This single value for the albedo slightly overestimates the reflected solar radiation at large incoming solar radiation (Fig. 6) because of a dependency of the albedo on solar elevation angle β (Yang et al., 2008; Zhang et al., 2013). Nonetheless does the use of a single value for the albedo hardly affect the error in modeled R_{ns} : the mean difference (MD) between measured and modeled R_{ns} lies between -0.23 and 1.63 W m^{-2} (Table 1), which is equal to the energy required to evaporate 0.008 to $0.057 \text{ mm day}^{-1}$. The NSE for estimating R_{ns} is close to 1 (Table 1), showing almost a perfect match of modeled to observed data.

The dense moss mat *Campylopus introflexus* entirely covers the underlying mineral soil, which results in a low albedo (0.135) due to the dark green surface. The albedo of bare sand (0.261) is comparable to values found in literature for bare dry coarse soils (Qiu et al., 1998; Van Bavel and Hillel, 1976; Linacre, 1969; Liakatas et al., 1986) and the albedo for grass (0.179) is consistent with values reported in other studies during summer time (Hollinger et al., 2010) or for dried grass (Van Wijk and Scholte Ubing, 1963). Heather has a somewhat lower albedo (0.078) than was found in the literature: Miranda et al. (1984) report an albedo of 0.13 (*Calluna*, LAI ca. 4); Wouters et al. (1980) report an albedo of 0.102 (*Calluna*). The heather vegetation in our study was in a later successional stage with aging shrubs having a relatively large fraction of twigs and a smaller LAI (3.47) than found by Miranda et al. (1984). Furthermore the albedo data of heather vegetation was collected primarily past the growing season from September till November. The darker surface after the growing season and the lower LAI explains the small albedo compared to other studies.

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



3.1.2 Net longwave radiation

The fitted function of Eq. (6) describes the dynamics of the surface temperature relative to air temperature (Fig. 7; Table 5). All surfaces have a similar average nighttime surface temperature ($T_{s, \text{offset}}$) relative to T_a , ranging between -7.47 and -10.21 °C. The solar elevation angle at which the surfaces become warmer than the air temperature (μ_β), as well as the speed at which the surface warms up or cools down (σ_β), are comparable between the surfaces. The main difference between the surfaces is observed at high solar elevation angles. Sand and moss show a clear increasing slope during the day, while grass and heather are able to attenuate the increase in surface temperature, possibly due to a larger latent heat flux (Fig. 7). The moss surface shows the largest increase in surface temperature during the day. Although organic layers, e.g. dry peat, have a larger specific heat ($1600 \text{ J kg}^{-1} \text{ K}^{-1}$) than dry sand ($693 \text{ J kg}^{-1} \text{ K}^{-1}$) (Gavriliiev, 2004), the energy required to heat up the moss material is much smaller than for sand, because of the small dry bulk density of ca. 26.8 g L^{-1} (derived for *Campylopus introflexus* from Voortman et al., 2014). Therefore, the surface temperature and the emitted longwave radiation are largest for the moss surface.

Our R_{nl} model (Eqs. 3 and 6) simulates R_{nl} much better than the calibrated (Table 2) FAO-56 R_{nl} sub-model (Table 3). For the natural grass surface, the NSE even becomes negative using the calibrated FAO-56 approach. Several studies showed that the FAO-56 R_{nl} sub-model underestimates the magnitude of R_{nl} for reference grass vegetation and poorly describes the diurnal pattern (Matsui, 2010; Blonquist Jr et al., 2010; Yin et al., 2008; Temesgen et al., 2007). As mentioned, the FAO-56 R_{nl} sub-model was originally developed for reference grass vegetation under well-watered conditions for daily time steps, but is commonly applied at hourly timescales (ASCE-EWRI, 2005; Perera et al., 2015; Gavilán et al., 2008; López-Urrea et al., 2006; Irmak et al., 2005). At daily time steps, T_s is close to T_a , since the warmer daytime T_s is compensated by the cooler nighttime T_s . For hourly time steps, the assumption that T_s follows T_a is not valid, which explains the poor performance of the FAO-56 R_{nl} model for hourly time

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



steps. This poor performance cannot be compensated by calibrating the net emissivity parameters, since the diurnal pattern remains unaffected.

In this analysis a typical pattern in T_s relative to T_a is used to estimate T_s (Eq. 6), and subsequently R_{nl} (Eq. 3). This relationship (Fig. 7) is sensitive to local weather conditions, which implies that the parameters of Eq. (6) (Table 5) are not directly transferable to other locations or climates. The applicability of the presented approach to simulate R_{nl} should be tested before it is used for other surfaces or climates.

3.1.3 Soil heat flux

The soil heat flux G as fraction of R_n (F_{day} and F_{night}) decreases with vegetation cover (Table 5). The nighttime fractions are larger than the daytime fractions, as R_n becomes smaller in magnitude during the night, which simultaneously corresponds to a change in direction of R_n and G , from downward (positive) to upward (negative). Relatively small systematic errors are made using daytime and nighttime fractions of R_n to simulate G (MD between 1.92 and 0.69 $W m^{-2}$) (Table 4). In remote sensing algorithms G is often simulated as fraction of R_n , depending on the LAI or the fractional vegetation cover. In e.g. the SEBS algorithm, the soil heat flux fraction (F) is interpolated between 0.35 for bare soil and 0.05 for a full vegetation canopy (Su, 2002). These limits are close to the bare sand (0.270) and heather (0.066) F_{day} fractions (Table 5). The heather F_{day} (0.066) was close to the value found by Miranda et al. (1984) of 0.04.

The analysis of the relationship between R_n and G was based on the average of two sets of soil heat flux plates per surface. These sets of measurements showed on average a good agreement: a MD below 1.07 $W m^{-2}$ with a RMSE ranging between 5.02 and 9.40 $W m^{-2}$.

3.1.4 Energy balance

All the terms in the energy balance can be defined using daily lysimeter measurements of LE and an estimate of the sensible heat flux (H) as a residual term of the energy bal-

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



ance. For daytime measurements (between sunrise and sunset), the LE , H , G , R_{sf} and R_{nl} can be expressed as fraction of the R_{sl} . Table 6 summarizes the average fraction of R_{sl} attributed to these five different energy fluxes during the measurement campaign. The net longwave radiation is for most surfaces the largest energy flux during daytime (Table 6).

The LE of most surfaces is the second largest flux during daytime, which fraction increases with vegetation cover. Despite the large difference in albedo between bare sand and moss, the moss surface has only a slightly larger LE fraction than bare sand (Table 6). This is primarily caused by the larger R_{nl} flux of moss, which compensates the smaller amount of reflected solar radiation.

3.2 Potential and actual evapotranspiration

The modeled ET_a is in agreement with the measured ET_a , with some exceptions at the onset of dry out events (Fig. 8). In general, reduction of ET_p to ET_a is modeled a few days later than emerges from measurements. The cumulative $ET_{a, mod}$ over the measurement period (May–October 2013) deviates 21 mm (13%), -13 mm (-7%), 5 mm (2%) and -3 mm (-2%) from the measured ET_a of the sand, moss, grass and heather lysimeters respectively. The results of modeled vs. measured ET_a for non-water stressed ($ET_a = ET_p$) and water stresses conditions ($ET_{a, mod} < ET_p$) are summarized in Table 7.

We did not calibrate our model, e.g. by adjusting soil hydraulic properties, because several processes outlined by Allen et al. (1991) and wall flow (Cameron et al., 1992; Corwin, 2000; Till and McCabe, 1976; Saffigna et al., 1977) affect lysimeter measurements of ET_a and drainage. We suspect that wall flow caused the slightly earlier reduction of ET_p to ET_a at the onset of dry out events than was simulated by the model. Wall flow leads to a quicker exfiltration of rainwater and a subsequent lower moisture content in the lysimeter, and therefore a slightly earlier timing of drought compared to the model. Since wall flow does not occur in the undisturbed vegetation outside the lysimeters, calibrating e.g. soil hydraulic properties using measured surface and

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



drainage fluxes in the objective function could lead to biased characterizations of the soil hydraulic properties and erroneous simulations of soil water flow and ET_a .

One would expect oasis effects to occur in the vicinity of the lysimeters, because freely draining lysimeters must saturate at the bottom of the lysimeter tank before water drains out. This enlarges the water availability inside the lysimeters compared to its groundwater independent surroundings and occasionally leads to a situation in which the vegetation inside the lysimeters is still transpiring, while the vegetation outside the lysimeters becomes water stressed and heats up. In such situation advection of sensible heat generated in the vicinity of the lysimeters could contribute to the available energy for lysimeter evapotranspiration. However, calculated ET_p was seldom smaller than measured lysimeter ET_a , indicating that oasis effects were absent. Furthermore, if oasis effects were prominent, systematic underestimation of modeled lysimeter ET_a would occur, since we ignored the possible contribution of heat advection. Note that it is very unlikely that oasis effects affected the back-calculated surface resistances (Table 5), since these were based on days after rain events for which we may assume ET_a to be equal to ET_p for both the lysimeters and their surroundings.

Neglecting feedbacks of drought on the surface resistance (i.e. using a fixed r_s) of heather and dry grassland vegetation leads to an overestimation of cumulative ET_a of 7 to 9% for years with relatively severe drought (Table 7). The delayed drought response of these vegetation types is therefore of importance to water balance studies, especially when, according to the expectations, summers become dryer as a result of a changing climate. Longer recordings of ET_a in heathland and grassland are required to understand and parameterize the drought response of these vegetation types in coupled plant physiological and hydro-meteorological models.

To our knowledge, this paper describes for the first time the evaporation characteristics of a moss surface in a dune ecosystem in a temperate climate. The evaporation rate of the dense moss mat *Campylopus introflexus* is 5% larger than the evaporation rate of bare sand. *Campylopus introflexus* forms dense moss mats and of the moss species investigated by Voortman et al. (2014), it has the largest water holding capac-

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



measurements of understory evaporation in heathlands and grasslands are required to unravel the ecological interactions between mosses and vascular plants.

4 Conclusions

In this study the net longwave radiation (R_{nl}) appeared to be one of the largest energy fluxes in dune vegetation. The poor performance of the calibrated FAO-56 approach for simulating R_{nl} for hourly time steps illustrates that this energy flux has attracted insufficient attention in evapotranspiration research. The novel approach presented in this study to simulate R_{nl} outperformed the calibrated FAO-56 approach and forms an accurate alternative for estimating R_{nl} .

A relatively simple hydrological model could be used to simulate evapotranspiration of dry dune vegetation with satisfactory results. Improvements in terms of climate robustness would be especially achieved if plant physiological processes were integrated in the hydro-meteorological model. Without considering the effects of dry spells on the surface resistance (r_s) of grassland and heathland vegetation, ET_a would be overestimated with 9 and 7% for years with relatively severe drought (drought events with a reoccurrence of once per five years).

Moss species are very economical with water. The evaporation of moss surfaces is comparable or even lower than bare sand. By promoting moss dominated ecosystems in coastal and inland dunes, the evapotranspiration could be reduced considerably, to the benefit of the groundwater system. Differences in evaporation between moss species are large and should be considered in water balance studies.

Long-term measurements of ET_a in heathland and grassland are required to study feedbacks between climate and plant physiological processes which allows the integration of the drought response of natural vegetation in coupled plant physiological and hydro-meteorological models. To understand the ecological interaction between mosses and vascular plants, detailed measurements of understory evaporation in heathlands and grasslands are required.

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

[Title Page](#)

[Abstract](#)

[Introduction](#)

[Conclusions](#)

[References](#)

[Tables](#)

[Figures](#)



[Back](#)

[Close](#)

[Full Screen / Esc](#)

[Printer-friendly Version](#)

[Interactive Discussion](#)



Appendix A: Soil hydraulic properties for the simulation of unsaturated flow with Hydrus-1D

Unsaturated flow in Hydrus 1D is described by a modified form of Richards' equation:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left(K \frac{\partial h}{\partial z} + K \right), \quad (\text{A1})$$

- 5 Where K is the unsaturated conductivity [L T^{-1}], z is the vertical coordinate [L] and t is the time [T]. The soil hydraulic properties were assumed to be described by the Mualum Van Genuchten functions:

$$\theta(h) = \theta_r + \frac{\theta_0 - \theta_r}{[1 + |\alpha h|^n]^m}, \quad (\text{A2})$$

$$K(\theta) = K_0 S_e^l \left[1 - \left(1 - S_e^{1/m} \right)^m \right]^2, \quad (\text{A3})$$

10 with

$$S_e(h) = \frac{\theta(h) - \theta_r}{\theta_0 - \theta_r}, \quad (\text{A4})$$

- where θ is the volumetric water content [$\text{L}^3 \text{L}^{-3}$], h is the soil water pressure head [L], θ_0 is an empirical parameter matching measured and modeled θ [$\text{L}^3 \text{L}^{-3}$], θ_r is the residual water content [$\text{L}^3 \text{L}^{-3}$] and α [L^{-1}] and n [-] are empirical shape parameters of the retention function, K is an empirical parameter, matching measured and modeled K [L T^{-1}], S_e is the effective saturation [-], l is the pore-connectivity parameter [-] and m ($= 1 - 1/n$) [-] is an empirical parameter. Drying retention data of two soil samples collected next to each lysimeter at 5 and 15 cm depth were used to fit a retention function with the RETC code (Van Genuchten et al., 1991). Hysteresis in the retention function was accounted for by assuming the retention curve parameter α for the wetting curve

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



(α_{wet}) to be twice as large as α of the drying retention curve (α_{dry}) (Šimůnek et al., 1999). The unsaturated hydraulic conductivity parameters l and K_0 were estimated using the Rosetta database and pedotransfer functions, providing the fitted drying retention curve parameters as input (Schaap et al., 2001). Average parameter values per lysimeter are summarized in Table A1.

The hydraulic properties of the 15 cm long wick, guiding drainage water below the lysimeter into the tipping bucket, were taken from Knutson and Selker (1994) who analyzed the same brand and type of wick, i.e. Peperell 1/2 inch. The K_0 of the wick was adjusted to correct for the smaller cross sectional area of the wick compared to the cross sectional area of the lysimeter in the 1-D model simulation (Table A1).

The heterogeneous pore structure of the moss material was described by the functions of Durner (1994):

$$S_e = w_1 (1 + [\alpha_1 h]^{n_1})^{-m_1} + w_2 (1 + [\alpha_2 h]^{n_2})^{-m_2}, \quad (\text{A5})$$

$$K(S_e) = K_s \frac{(w_1 S_{e_1} + w_2 S_{e_2})' \left(w_1 \alpha_1 \left[1 - \left(1 - S_{e_1}^{1/m_1} \right)^{m_1} \right] + w_2 \alpha_2 \left[1 - \left(1 - S_{e_2}^{1/m_2} \right)^{m_2} \right] \right)^2}{(w_1 \alpha_1 + w_2 \alpha_2)^2}, \quad (\text{A6})$$

Where w_1 and w_2 are weighting factors for two distinct pore systems of the moss layer; a capillary pore system (subscript 1) and a macro pore system active near saturation ($h > -1$ cm, subscript 2) and K_s is the hydraulic conductivity at saturation. Average hydraulic parameters of the capillary pore system and the volumetric portion of the macro pore system of the moss species *Campylopus introflexus* and *Hypnum cupressiforme* were taken from Voortman et al. (2014) (illustrated with dotted lines in Figs. A1 and A2). The α_2 parameter was fitted to the functions of Voortman et al. (2014) using $K_s = 1000 \text{ cm day}^{-1}$ and $n_2 = 2$ by minimizing the RMSE by generalized reduced gradient nonlinear optimization. Hydraulic parameter values are listed in Table A2.

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



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HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

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Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

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HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table 1. Model performance of R_{ns} simulations.

Surface	N	NSE	RMSE [W m^{-2}]	MD [W m^{-2}]	M%D [%]
Sand	218	0.998	5.99	-0.23	-0.10
Moss	1317	0.999	5.46	1.18	0.46
Grass	1203	0.998	7.78	1.63	0.55
Heather	407	0.999	3.00	0.24	0.09

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table 2. Calibrated net emissivity parameters of the FAO-56 R_{nl} sub-model (Eq. 7).

	<i>a</i>	<i>b</i>
Sand	0.31	−0.00
Moss	0.33	0.02
Grass	0.36	−0.06
Heather	0.24	0.02

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table 3. Model performance of R_{nl} simulations for hourly time steps.

Surface	N	NSE	RMSE [W m^{-2}]	MD [W m^{-2}]	M%D [%]
Using Eq. (3)					
Sand	5891	0.65	27.37	0.92	1.52
Moss	5997	0.74	28.57	3.73	5.19
Grass	6113	0.71	25.66	1.41	2.36
Heather	2424	0.63	27.63	-0.21	-0.40
Using FAO-56 Eq. (7)					
Sand	5891	0.41	35.39	4.34	7.14
Moss	5997	0.31	46.67	14.84	20.65
Grass	6113	-0.07	49.41	-18.23	-30.38
Heather	2424	0.29	38.24	10.50	19.54

[Title Page](#)
[Abstract](#)
[Introduction](#)
[Conclusions](#)
[References](#)
[Tables](#)
[Figures](#)
[Back](#)
[Close](#)
[Full Screen / Esc](#)
[Printer-friendly Version](#)
[Interactive Discussion](#)


HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table 4. Model performance of G simulations.

Surface	N	NSE	RMSE [W m^{-2}]	MD [W m^{-2}]	M%D [%]
Sand	6080	0.820	20.06	1.92	22.16
Moss	5335	0.901	12.02	1.65	24.29
Grass	6046	0.868	8.97	1.60	43.42
Heather	2028	0.641	11.39	0.69	40.27

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table 6. Average fractionation of the incoming shortwave radiation ($R_{s\downarrow}$) between different energy fluxes during daytime.

Surface	LE	H	G	$R_{s\downarrow}$	R_{nl}
Sand	0.22	0.13	0.10	0.26	0.28
Moss	0.24	0.17	0.09	0.14	0.36
Grass	0.27	0.21	0.06	0.18	0.29
Heather	0.35	0.20	0.05	0.08	0.32

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures



Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table 7. Modeled ET_p and ET_a for different surfaces in a lysimeter (lys.) and for a situation with deep groundwater levels (gw. ind.) for the year 2013.

	ET_p [mm]	ET_a lys. [mm]	ET_a gw. ind. [mm]
Bare sand	400	295	258
Moss (<i>Campylopus int.</i>)	468	312	272
Moss (<i>Hypnum cup.</i>)	468	–	182
Grass	392	350	333
Grass no dieback	429 (+9 %)	382 (+9 %)	362 (+9 %)
Heather	549	460	391
Heather no dieback	610 (+11 %)	499 (+8 %)	420 (+7 %)

[Title Page](#)
[Abstract](#)
[Introduction](#)
[Conclusions](#)
[References](#)
[Tables](#)
[Figures](#)
[Back](#)
[Close](#)
[Full Screen / Esc](#)
[Printer-friendly Version](#)
[Interactive Discussion](#)


HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table A1. Hydraulic parameter values of lysimeter soils.

	θ_r [-]	θ_0 [-]	α_{dry} [cm ⁻¹]	α_{wet} [cm ⁻¹]	n [-]	K_0 [cm h ⁻¹]	L [-]
Bare Sand	0.01	0.367	0.023	0.046	2.945	1.042	-0.401
Moss	0.01	0.397	0.019	–	2.335	0.734	-0.173
Grass	0.01	0.401	0.025	0.050	2.071	1.119	-0.278
Heather	0.01	0.392	0.018	0.036	2.581	0.679	-0.186
Wick	0.00	0.630	0.098	0.196	3.610	2.180	0.500

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Table A2. Hydraulic parameter values of the two moss species.

	θ_r [-]	θ_s [-]	α_1 [cm ⁻¹]	n [-]	K_s [cm h ⁻¹]	l [-]	w_2 [-]	α_2 [cm ⁻¹]	n_2 [-]
<i>Campylopus int.</i>	0.060	0.936	0.080	2.25	41.67	-2.69	0.371	45.89	2.00
<i>Hypnum cup.</i>	0.010	0.971	0.013	2.17	41.67	-2.37	0.800	16.61	2.00

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

⏪

⏩

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)**Table B1.** Parameters of the water stress reduction function used in the Hydrus 1D model.

P0 [cm]	P1 [cm]	P2H [cm]	P2L [cm]	P3 [cm]	r2H [mm h ⁻¹]	r2L [mm h ⁻¹]
-10	-25	-300	-1000	-8000	5	1

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

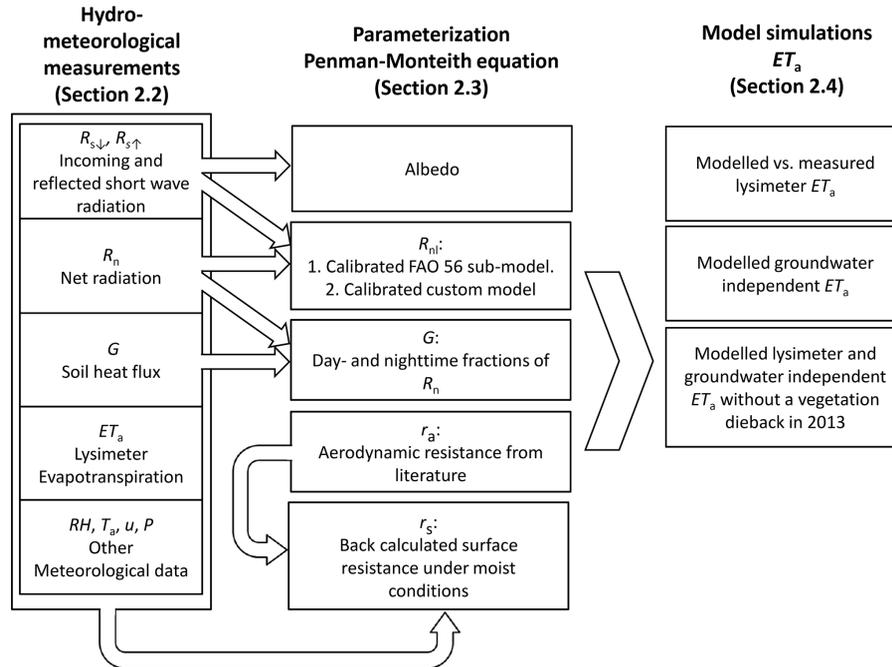


Figure 1. Organization of the research from measurements to model simulations.

Title Page	
Abstract	Introduction
Conclusions	References
Tables	Figures
◀	▶
◀	▶
Back	Close
Full Screen / Esc	
Printer-friendly Version	
Interactive Discussion	



Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

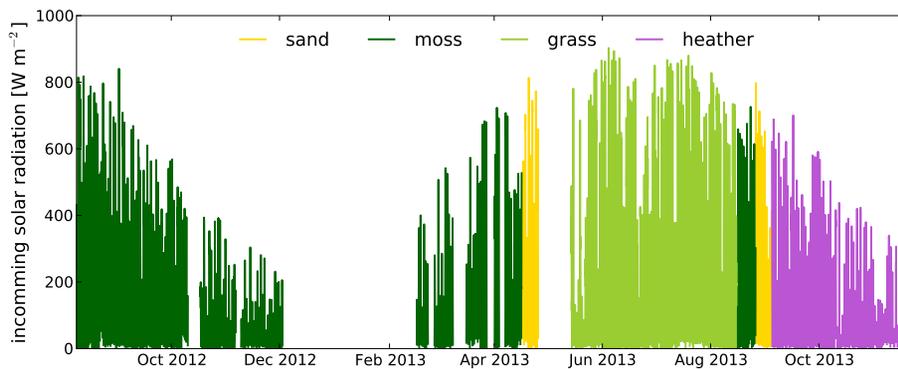


Figure 2. Measured incoming solar radiation R_{si} at the four different surfaces. Periods with snow cover or sensor maintenance were omitted.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

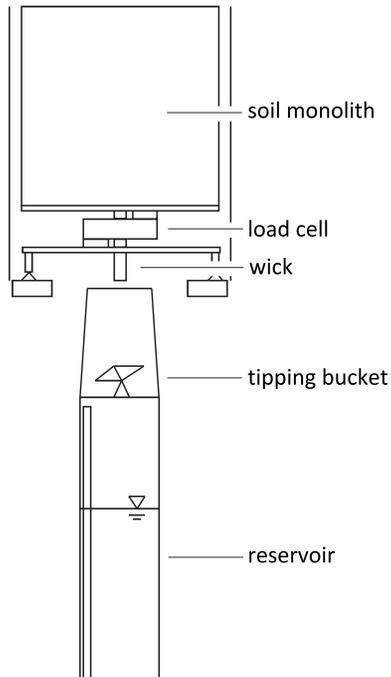


Figure 3. Lysimeter design.

HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

[Title Page](#)

[Abstract](#) | [Introduction](#)

[Conclusions](#) | [References](#)

[Tables](#) | [Figures](#)

[◀](#) | [▶](#)

[◀](#) | [▶](#)

[Back](#) | [Close](#)

[Full Screen / Esc](#)

[Printer-friendly Version](#)

[Interactive Discussion](#)



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

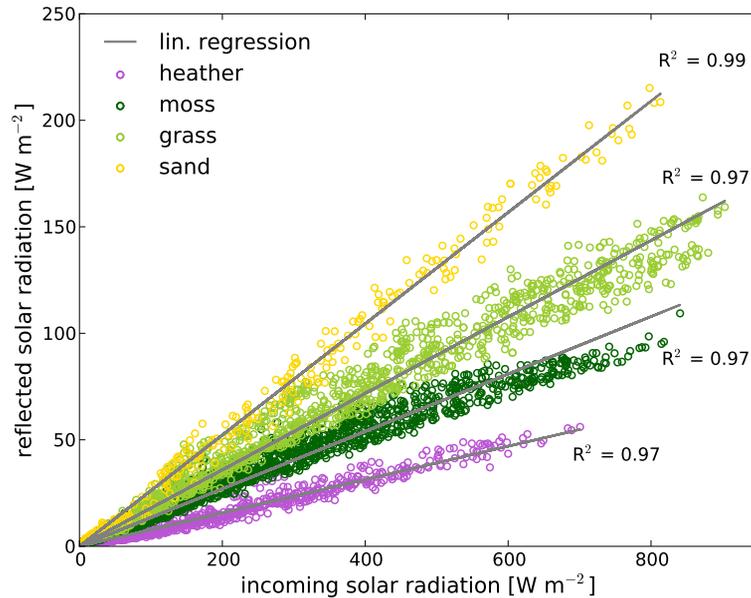


Figure 5. Linear regressions between incoming and reflected solar radiation.

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

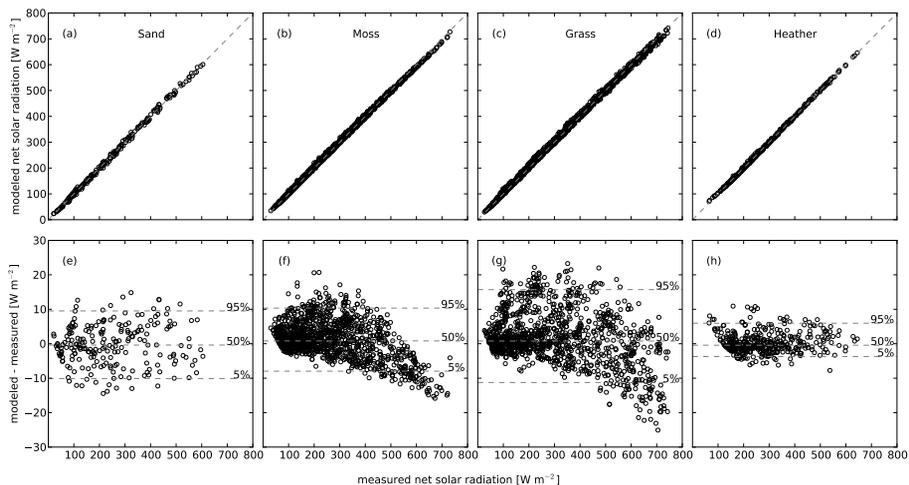


Figure 6. Modeled compared to measured net solar radiation (a–d, dashed lines are 1 : 1 lines) and deviations from the 1 : 1 line (e–h, dashed lines indicate 5, 50 and 95 percentiles).

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

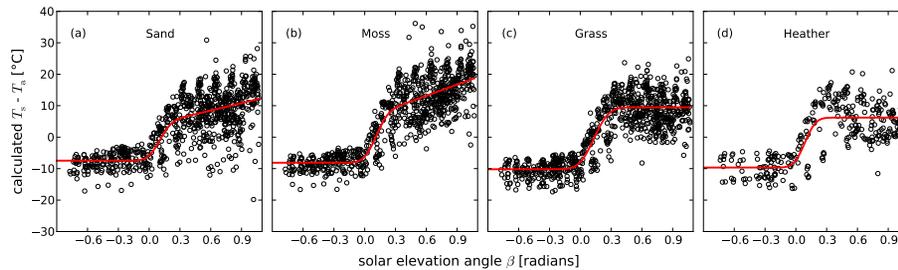


Figure 7. Measured surface temperature relative to air temperature ($T_s - T_a$) for clear hours ($f_{\text{od}} > 0.9$) as function of solar elevation angle β . Relationships (red lines) were fitted to the data using Eq. (6).

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[⏪](#)[⏩](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

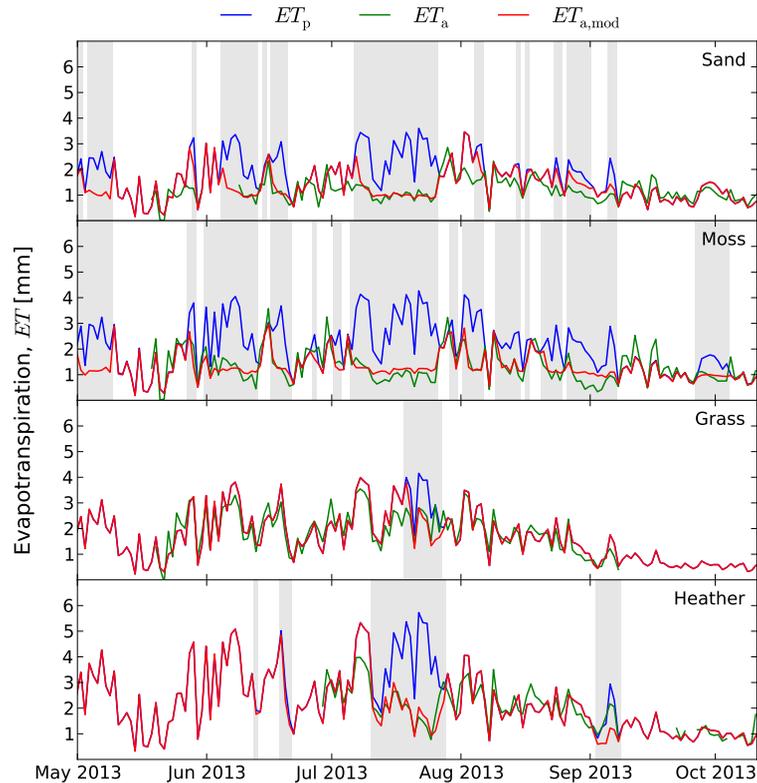


Figure 8. Measured and modeled daily ET for the four lysimeters. Grey bars indicate time periods where $ET_{a, \text{mod}}$ is smaller than ET_p , i.e. when evapotranspiration was water limited.

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



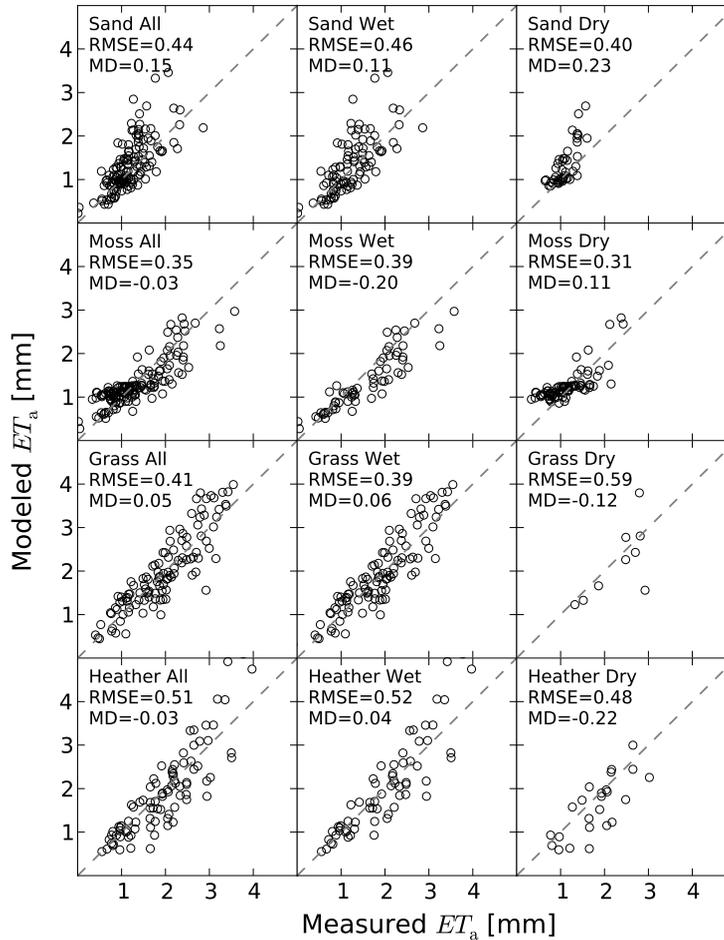


Figure 9. Measured vs. modeled ET_a of the lysimeters for all, wet ($ET_{a, \text{mod}} = ET_p$) and dry ($ET_{a, \text{mod}} \neq ET_p$) days. Dotted lines represent the 1 : 1 lines.

[Title Page](#)

[Abstract](#) | [Introduction](#)

[Conclusions](#) | [References](#)

[Tables](#) | [Figures](#)

[◀](#) | [▶](#)

[◀](#) | [▶](#)

[Back](#) | [Close](#)

[Full Screen / Esc](#)

[Printer-friendly Version](#)

[Interactive Discussion](#)



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

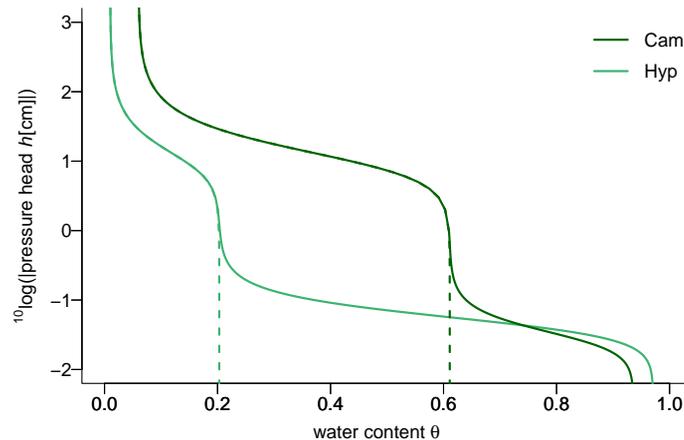


Figure A1. Water retention functions of two moss species: *Campylopus introflexus* and *Hypnum cupressiforme*. The dotted lines indicate the contribution of the capillary pore system characterized by Voortman et al. (2013).

[Title Page](#)[Abstract](#)[Introduction](#)[Conclusions](#)[References](#)[Tables](#)[Figures](#)[◀](#)[▶](#)[◀](#)[▶](#)[Back](#)[Close](#)[Full Screen / Esc](#)[Printer-friendly Version](#)[Interactive Discussion](#)

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

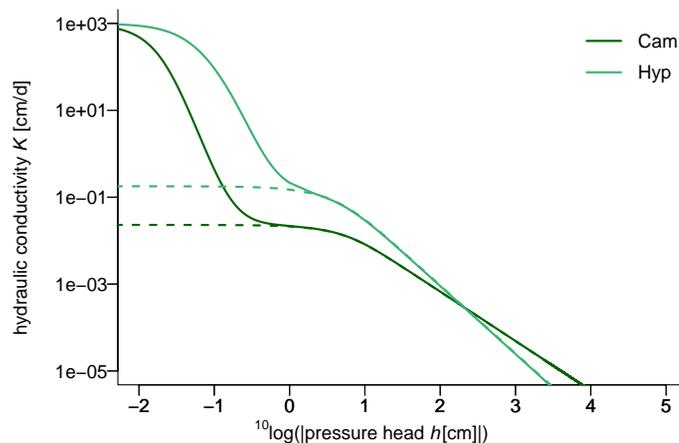


Figure A2. Hydraulic conductivity functions for two moss species: *Campylopus introflexus* and *Hypnum cupressiforme*. The dotted lines indicate the contribution of the capillary pore system characterized by Voortman et al. (2013).

Title Page

Abstract

Introduction

Conclusions

References

Tables

Figures

◀

▶

◀

▶

Back

Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion



HESSD

12, 4541–4594, 2015

Quantifying energy and water fluxes in dry dune ecosystems of the Netherlands

B. R. Voortman et al.

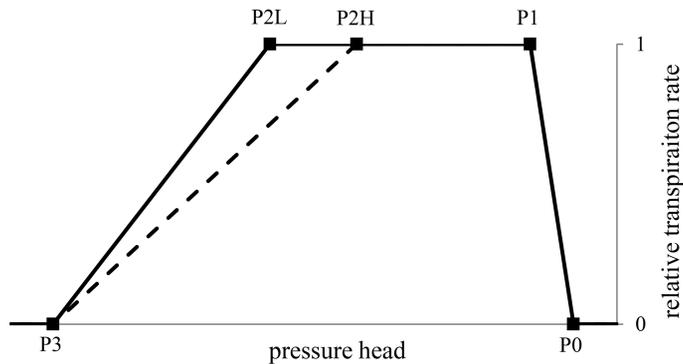


Figure B1. The relative transpiration rate as function of soil water pressure head according to Feddes et al. (1978).

[Title Page](#)

[Abstract](#)

[Introduction](#)

[Conclusions](#)

[References](#)

[Tables](#)

[Figures](#)

[⏪](#)

[⏩](#)

[◀](#)

[▶](#)

[Back](#)

[Close](#)

[Full Screen / Esc](#)

[Printer-friendly Version](#)

[Interactive Discussion](#)

