Contribution of dew to the water budget of a grassland area in the Netherlands

Adrie F. G. Jacobs,[1] Bert G. Heusinkveld,[1] Roy J. Wichink Kruit,[1]
and Simon M. Berkowicz[2]

Received 22 February 2005; revised 6 December 2005; accepted 27 December 2005; published 15 March 2006.

[1] The annual amount of dew input to the water budget in the midlatitudes is mostly neglected, possibly because direct dew measurements are very difficult and time-consuming. As the Netherlands has a very high frequency of dew events, a grassland area was selected to determine whether dew input could be significant. The study site is situated within the Wageningen University meteorological station. Dew measurement experiments were carried out in 2004. Data were used to verify a surface energy dew model, which was then applied to an 11-year data set. A mean annual dew amount of 37 mm was obtained with a standard deviation of 8 mm, while the mean annual precipitation was 830 mm with a standard deviation of 200 mm. Dew contributed about 4.5% of the mean annual precipitation. The average number of dew nights per year was 250 (70%) with a standard deviation of 25 nights. This frequency significantly affects leaf wetness and possible vegetation diseases.

Citation: Jacobs, A. F. G., B. G. Heusinkveld, R. J. Wichink Kruit, and S. M. Berkowicz (2006), Contribution of dew to the water budget of a grassland area in the Netherlands, *Water Resour. Res.*, *42*, W03415, doi:10.1029/2005WR004055.

1. Introduction

[2] Fog, mist, and dew are meteorological phenomena that can provide moisture to a given surface. Although such phenomena contribute free water to the Earth's surface, in meteorology they are not considered as precipitation. The interception of fog in forested areas can reach many liters per square meter per episode [*Dawson*, 1998], while dew, formed through condensation, has a theoretical maximum of 0.8 mm per night [*Monteith*, 1957]. Such dew quantities are small and as a consequence are difficult to measure.

[3] Nevertheless, dew may provide free liquid water to living organisms [*Evenari et al.*, 1982; *Jacobs et al.*, 2000a, 2000b; *Steinberger et al.*, 1989] and can also contribute to the water budget [*Baier*, 1966; *Beysens*, 1995; *Malek et al.*, 1999]. Fog, mist, and dew can form drops or water films on plant leaves, causing so-called leaf wetness. Leaf wetness affects plant growth [*Wallin*, 1967] and favors the development of plant diseases [*Aylor*, 1986]. When water is deposited on leaves for critical periods and when temperatures are suitable, fungal spores and other pathogens may develop that can be extremely harmful to plant canopies.

[4] In arid and semiarid regions, however, dew can contribute considerably to the water balance [*Malek et al.*, 1999]. In the NW Negev desert of Israel, observations by *Evenari et al.* [1982] and *Zangvil* [1996] have shown that dew occurs about 200 evenings per year and can reach the equivalent of 30 mm of annual precipitation. In drought years, dew can exceed the annual precipitation [*Evenari et al.*, 1982] and monthly precipitation in summer-dry climates [*Tuller and Chilton*, 1973].

[5] In suitable arid and semiarid areas, specially designed passive fog collectors can typically collect 1–10 L m^{-2} d^{-1} and serve as a primary source of water for village communities [*Olivier and Rautenbach*, 2002; *Schemenauer and Cereceda*, 1994a, 1994b]. In Corsica, large dew collectors of 30 m^2 have been tested [*Muselli et al.*, 2006].

[6] The Netherlands, a midlatitude coastal country, has a very high frequency of dew events that are more or less evenly distributed during the year. Fog episodes also occur, but in contrast to dew, they have a very irregular pattern. Here we focus on dew in a Netherlands grassland region. Dew events and an estimate of annual dew input are compared with precipitation to assess the extent of dew contribution to the total annual water budget.

[7] Since dew condenses on a given surface, there is no standard way to measure it [*Berkowicz et al.*, 2001]. Thus proxy approaches have been attempted depending on the intended application. As an indicator for pathogens in agriculture, wooden blocks were designed by *Duvdevani* [1947] as an optical way of quantifying the dew drops that formed upon the wooden surface. A standard set of photos was used to calibrate drop size and equivalent dew depth. This technique was simple to execute in practice and therefore applied in various studies [e.g., *Tuller and Chilton*, 1973]. However, under arid and semiarid conditions this approach is highly unreliable [*Jacobs et al.*, 2000a; *Lomas*, 1965]. Other techniques to quantify nighttime dew amounts include the Hiltner dew balance [*Hiltner*, 1930; *Gelbe*, 1954] and the Kessler-Fuess recorder [*Kessler*, 1939; *Nagel*, 1962; *Baier*, 1966]. In the NW Negev desert, manual and recording microlysimeters have been used [*Jacobs et al.*, 2000b; *Heusinkveld et al.*, 2006].

[8] In the present research, dew was measured using small lysimeters, with the collected data used to calibrate

[1]Wageningen University, Department of Meteorology and Air Quality, Wageningen University, Wageningen, Netherlands.
[2]Minerva Arid Ecosystems Research Centre, Hebrew University of Jerusalem, Jerusalem, Israel.

0043-1397/06/2005WR004055$09.00

a dew model. The verified model was then applied to an 11-year meteorological data set to generate estimates of nighttime dew amounts during this period.

2. Materials and Methods

2.1. Theory

[9] Dew can occur when evening radiative cooling allows water vapor from the atmospheric water reservoir to condense on a given surface [*Garratt and Segal*, 1988; *Jacobs and Nieveen*, 1995]. In addition, dew can form when soil water evaporates during the night and is intercepted by a canopy [*Monteith*, 1957; *Garratt*, 1992] and through an internal plant water excretion process known as guttation. Guttation amounts, however, are small [*Long*, 1955] and will be neglected in the present study.

[10] In this study, simple manual lysimeter containers were used to measure dew. The containers were weighed every 30 min during special periods. For each night, new samples were taken in order to avoid a deviation between the moisture balance of the container and its direct environment. Although this approach is very accurate, it is very laborious and time-consuming.

[11] Dew can also be estimated indirectly by using the eddy-covariance technique in which

$$\lambda_v E = \rho \lambda_v \overline{w'q'} \qquad \lambda_v E = \rho \lambda_v \overline{w'q'}, \tag{1}$$

where E (kg m^{-2} s^{-1}) is evapotranspiration or dewfall, ρ (kg m^{-3}) is air density, λ_v (J kg^{-1}) is the latent heat of vaporization, w (m s^{-1}) is the vertical velocity component, and q (kg kg^{-1}) is the specific humidity of the air. If $\lambda_v E$ is positive, the grass evapotranspirates; if it is negative, dew accumulates at the grass cover. Because of the nature of nighttime eddy-covariance measurements [*Nieveen et al.*, 2005], there is a considerable scatter and uncertainty in nighttime measurement data using this technique. This is why in practice, surface energy budget models are preferred instead [*Holtslag and de Bruin*, 1988].

[12] We start from the Earth's surface energy budget [*Garratt and Segal*, 1988]:

$$Q* - G = \lambda_v E + H, \tag{2}$$

where Q* (W m^{-2}) is net radiation, G (W m^{-2}) is soil heat flux, $\lambda_v E$ (W m^{-2}) is evapotranspiration, and H (W m^{-2}) is sensible heat, and combine this result with the free water evaporation/dew formation [*Garratt and Segal*, 1988]:

$$\lambda_v E = \rho \lambda_v \frac{q*(T_o) - q}{r_{av}}, \tag{3}$$

where $q*(T_o)$ (kg kg^{-1}) is saturated specific humidity at surface temperature T_o (°C), q (kg kg^{-1}) is specific humidity at a reference level, z_r (in our case z_r = 1.5 m above the grass cover), and r_{av} (s m^{-1}) is the aerodynamic resistance to vapor transport. Then the evaporation or dew formation of free liquid water is reached after using Penman's substitution [*Garratt*, 1992]:

$$\lambda_v E = \frac{s}{s+\gamma}(Q* - G) + \frac{\gamma}{s+\gamma}\frac{\rho\lambda_v \delta q}{r_{av}}, \tag{4}$$

where $s = dq*/dT$ (K^{-1}) is the slope of the saturation specific humidity curve, $\gamma = c_p/\gamma_v$ (K^{-1}) is the psychrometric constant, $\delta q = q*(T_a) - q$ (kg kg^{-1}) is the deficit specific humidity at reference level, and T_a is the air temperature. The aerodynamic resistance to vapor transport, r_{av}, is given by [*Garratt*, 1992]

$$r_{av} = \frac{\left(\ln\frac{z_r}{z_{ov}} - \Psi_v(\eta_r) + \Psi_v(\eta_{ov})\right)\left(\ln\frac{z_r}{z_o} - \Psi_m(\eta_r) + \Psi_m(\eta_o)\right)}{\kappa^2 u_r} \tag{5}$$

where $\eta_r = z_r/L$, $\eta_o = z_o/L$, $\eta_{ov} = z_{ov}/L$ with L Obukhov's stability length scale, z_o and z_{ov} are the roughness lengths for momentum and vapor, respectively. Here ln (z_o/z_{ov}) = 2 is assigned the classical value [*Garratt*, 1992]. Furthermore, u_r is the windspeed at the reference height, κ (=0.4) is von Karman's constant, and Ψ_m and Ψ_v are the integrated stability functions for momentum and vapor, respectively. The latter are defined by [*Garratt*, 1992]

$$\begin{aligned}
&\Psi_m(\eta) = -5\eta \text{ for } \eta \ge 0\\
&\Psi_m(\eta) = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2a\tan(x) + \frac{\pi}{2} \text{ for } \eta < 0\\
&\text{where } x = (1+16|\eta|)^{0.25}
\end{aligned} \tag{6a}$$

$$\begin{aligned}
&\Psi_v(\eta) = -5\eta \text{ for } \eta \ge 0 \quad .\\
&\Psi_v(\eta) = 2\ln\left(\frac{1+x^2}{2}\right) \text{for } \eta < 0
\end{aligned} \tag{6b}$$

Here we wish to estimate the evaporation or dew formation of free liquid water according to equation (4) using common meteorological variables, and $\eta_r = z_r/L$ from the same meteorological variables. In the present study the following relation was used [*de Bruin et al.*, 1999]:

$$\begin{aligned}
&\eta_r = \frac{z_r}{L} = Ri_B(z_r) \text{ for } Ri_B < 0\\
&\eta_r = \frac{z_r}{L} = \left|\frac{Ri_B(z_r)}{1 - c_1 Ri_B(z_r)}\right| \text{ for } Ri_B \ge 0 \quad ,\\
&\text{where } Ri_B = \frac{g}{T_{abs}(z_r)} \frac{(z_r - z_{oh})(T_a(z_r) - T_o)}{u^2(z_r)}
\end{aligned} \tag{7}$$

where Ri_B is the bulk Richardson number, g (m s^{-2}) is gravity, $T_{abs}(z_r)$ is the absolute temperature at the reference height, c_1(=5) is a constant, z_{oh} is the roughness length for heat, evaluated here as $z_{oh} = z_{ov}$, and T_o is the surface temperature. If the surface temperature T_o is not available, T_a at 10 cm can be substituted instead.

[13] The accumulated amount of dew within the grass cover is calculated by summing the negative evaporation according to

$$\begin{aligned}
&D_{i+1} = D_i + E_i \Delta t \textit{ if } D_{i+1} \ge 0\\
&D_{i+1} = 0 \textit{ if } D_i + E_i \Delta t < 0
\end{aligned} \quad , \tag{8}$$

where D_{i+1} is the new accumulated dew amount, D_i is the former dew amount, E_i is the dew flux density calculated using equation (4), and Δt(= 600 s) is the time step. If

$D_i + E_i\Delta t < 0$, D_{i+1} is set to zero since this means that all free water on the leaves has evaporated. In addition, it must be noted that after D_{i+1} has been set to 0, the evaporation equation (4) cannot be applied anymore, since the crop resistance r_c must be taken into account.

[14] The above mentioned methods were used in this study. Because the Penman technique is relatively simple, we applied it to the grassland study area. Although there is a second energy budget technique available, it is best applicable for tall canopies only [*Pedro and Gillespie*, 1982a, 1982b; *Jacobs et al.*, 2005].

2.2. Experimental Setup

[15] Wageningen University operates a meteorological observatory, the Haarweg Station, in the center of the Netherlands (latitude 51°58′N, longitude 5°38′W, altitude 7 m above sea level (a.s.l.); www.met.wau.nl). The region has perennial grassland with the dominant plant species consisting of *Lolium perenne* and *Poa trivialis*. The grass cover is mowed weekly and has a mean height of 10 cm and a mean leaf area index (LAI) of 2.9 [*Snel*, 2004]. The LAI of the grass cover was estimated with a digital plant canopy imager (Licor, type CI-110). The soil at the site is predominantly heavy basin clay resulting from the back swamps of the Rhine River.

[16] An aspirated psychrometer measures the air temperature T_a and wet-bulb temperature T_w at reference height $z_r = 1.5$ m. At 10 cm height, the grass temperature T_a(10 cm) is measured with a shielded Pt-100 thermometer (homemade). The incoming (R_{gi}) and outgoing (R_{go}) shortwave radiation is measured with an aspirated pyranometer (Kipp & Zonen, model CM11). The incoming (R_{Li}) and outgoing (R_{Lo}) longwave radiation is measured with a pyrgeometer (Kipp & Zonen, model CG 1). The outgoing longwave radiometer is used to evaluate the surface temperature T_o, ($T_o = \sqrt[4]{R_{lo}/\varepsilon\sigma}$, where ε is emissivity and σ is Stefan Boltzmann's constant). A leaf wetness sensor (237 wetness sensing grid; Campbell Scientific, Inc.) rests at a height of 0.05 m.

[17] Soil temperatures T_s were measured by Pt-100 element at depths 0.05, 0.10, 0.20, 0.50, and 1.0 m. At 0.01, 0.10, and 0.50 m soil depths the soil moisture content is measured with a TDR (time domain reflectrometry) system. The soil heat flux is measured by a heat plate (TNO, WS 31-Cp) buried at a depth of 75 mm. The measured soil heat fluxes are corrected for instrumental shape correction, Φ, for soil heat flux sensors as proposed by *Mogensen* [1970]:

$$\Phi = \frac{1}{1 - 1.7tA^{-0.5}\left(1 - \varepsilon_p^{-1}\right)}. \tag{9}$$

Here t is the thickness of the plate, A is the area of the plate, and ε_p is the ratio of the thermal conductivity of the plate to that of the soil. Moreover, the measured soil heat fluxes are corrected for the soil heat storage above the heat plate [*Fritschen and Gay*, 1979] according to

$$G(0) = G(z_p) - C_s \int_{z_p}^{0} \frac{\delta T}{\delta t} dz, \tag{10}$$

where z_p is the depth of the soil plate and C_s is the volumetric heat capacity of the soil.

[18] Apart from standard agrometeorological measurements, turbulent fluxes of momentum, heat, and mass (H_2O and CO_2) are measured as well. A lattice tower is instrumented with an eddy-covariance system installed at a height of 3.5 m. This system includes a three-dimensional sonic anemometer (3-D Solent, Gill Instruments Ltd., model A1012R2), a fine wire thermocouple (homemade), and an open path infrared CO_2 and H_2O gas analyzer (IRGA) (LI-COR, Inc., model LI-7500). The 3-D sonic anemometer and the IRGA are set 0.05 m apart.

[19] Direct dew measurements were initiated in 2004 using small microlysimeters of the type described by *Boast and Robertson* [1982]. Seven microlysimeter containers (internal height 30 mm; internal diameter 70 mm) were buried randomly in the station at seven locations. They contained soil and grass samples and were manually weighed in the field every 30 min from 1800 UTC to 1200 UTC with a portable sensitive (±1 mg) balance (Mettler PM1200). New samples were taken daily in order not to disturb the water budget of the microlysimeters.

[20] The slow response meteorological instruments are sampled at 0.25 Hz. At 30-min intervals, data are averaged and stored in data loggers for subsequent processing. The fast response sonic anemometer, the IRGA system, and the fine wire thermocouple are sampled at 20.8 Hz. The raw data of the eddy-covariance system are stored on a PC and processed in 30-min intervals. More details about the experiments, the measurement site, and the data processing are given by *Jacobs et al.* [2003].

3. Results and Discussion

[21] Microlysimeter measurements were made over 20 nights in April and May 2004. Because of rain and fog events, however, only eight nights could be used in this study. Figure 1a displays the course of the dew amounts gathered with the lysimeters and measured with the eddy-covariance technique, along with the simulated amounts according to the surface energy budget model, for one selected night (29–30 May 2004). Results from the leaf wetness grid are also plotted in Figure 1a. The night selected was representative of the measurement nights and highlights the good agreement between the microlysimeter measurements and the simulated dew results. Moreover, the wetness period is well indicated by the wetness sensor. The sensor data are in arbitrary units with 0 = dry and > 0 = wet. Figure 1b displays the mean wind speed at 2 m height along with the friction velocity u^*, measured by the eddy-covariance system, and also shows that the night presented was relatively calm ($u^* \leq 0.1$ m s^{-1}).

[22] Figure 1a also reveals that the eddy-covariance results hardly agree with the microlysimeters. The nighttime eddy-covariance measurements almost always indicated a serious underestimation of the nighttime vapor fluxes. As mentioned earlier, because of the nature of nighttime eddy-covariance measurements [*Nieveen et al.*, 2005] a considerable amount of scatter and uncertainty can be expected for nighttime measurement data.

[23] This serious mismatch between the eddy-covariance and microlysimeter measurements, however, must have some physical basis. Hence the imbalance between the

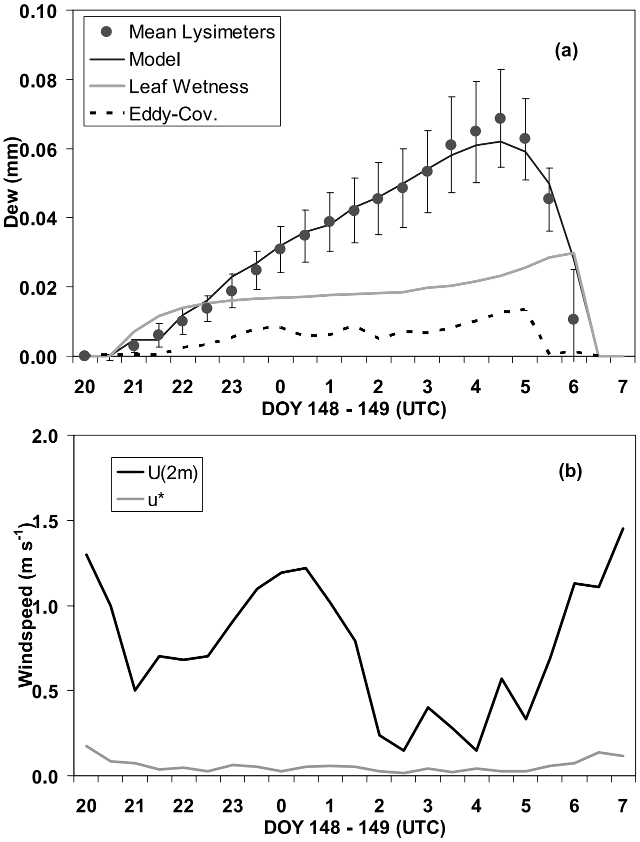

Figure 1. (a) Course of the cumulative dew amounts measured by the microlysimeters and the eddy-covariance techniques versus calculated dew amounts according to the surface energy budget model. Leaf wetness sensor data indicated in arbitrary units (0 = dry, > 0 = wet). Standard deviations are provided for the microlysimeter data. (b) Course of wind speed at 2 m height and friction velocity, u^*, 29–30 May 2004.

available energy ($Q^* - G$) and the turbulent fluxes ($H + \lambda_v E$) was analyzed for four consecutive days (27–30 May 2004) and the results were plotted in a scattergram (Figure 2). The linear regression forced through the origin was $y = 0.87x$ with $R^2 = 0.96$ ($N = 192$). This means that there is an imbalance of 13%, which is high but not uncommon by using the covariance technique [*Laubach et al.*, 1994; *Brotzge and Crawford*, 2003].

[24] Many groups are studying the surface energy imbalance. It is now suspected that the problem lies in correctly measuring the latent heat flux over vegetation, since under dry and very dry conditions this imbalance is much smaller and sometimes even zero [e.g., *Van de Wiel et al.*, 2003; *Heusinkveld et al.*, 2004; *Steeneveld et al.*, 2006]. According to the nighttime surface energy budget (see inset, Figure 2), however, the imbalance during the night is much larger. Only for relatively windy nights ($u^* > 0.1$ m s^{-1}) do the nighttime results follow more or less the daytime correlation. For relatively calm nights ($u^* \leq 0.1$ m s^{-1}), which are the nights that favor dew formation, it appears that the turbulent fluxes ($H + \lambda_v E$) are seriously underestimated. During relatively calm stable nights, turbulence is suppressed and the eddy-covariances become ill defined since these conditions are nonstationary and nonhomogeneous. It appears that the criterion $u^* \leq 0.1$ m s^{-1} is an appropriate threshold for not applying the eddy-covariance technique [*Van de Wiel et al.*, 2003]. The same nighttime underestimation as given above was found by *Laubach et al.* [1994] in central Germany. Note that the example in Figure 1 also was a calm night.

[25] Visual inspection of the eddy-covariance system during calm nights established that it becomes extremely wet, which also contributes to the uncertainty of the measured fluxes. For this reason, the gas analyzer system used in the meteorological station is heated artificially in order to prevent drop formation on the lenses. This heating improves the results somewhat but not sufficiently. Another possible reason for underestimation is that the eddy-covariance system measures fluxes at a height of 3.5 m whereas nighttime vertical divergence of fluxes can occur in the

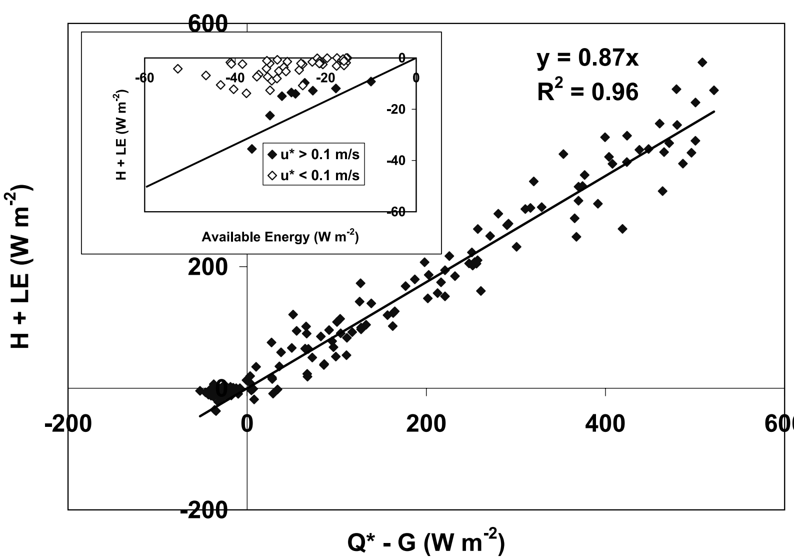

Figure 2. The surface energy budget closure for four consecutive days in May 2004. Top left frame shows an enlargement of nighttime conditions.

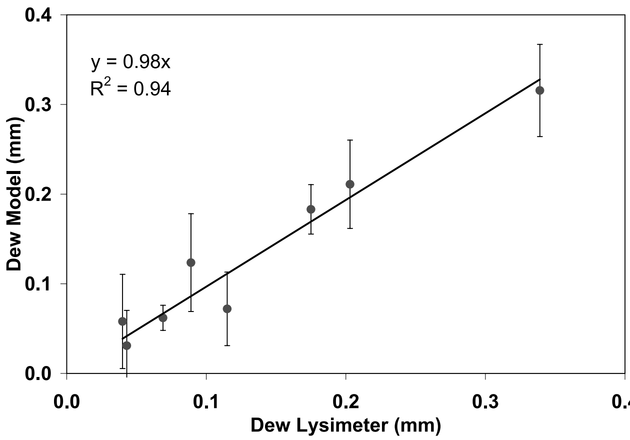

Figure 3. The modeled and measured dew amount, and standard deviations, during eight nights over a grassland area.

stable layer near the grass cover. In particular, vertical divergence of fluxes can occur during haze or fog conditions when small water drops float just above the surface. This issue is presently being investigated in our study area. For the present study most dew events occurred during relatively calm nights ($u^* \leq 1$ m s^{-1}) for which use of the eddy-covariance technique becomes questionable. Hence this technique was not applied in this study.

[26] In Figure 3 the scattergram compares the modeled dew amounts versus the measured dew amounts obtained with the microlysimeters for eight nights. The linear regression forced through the origin was $y = 0.98x$ with $R^2 = 0.94$ ($N = 8$), which means that the unexplained variance is 6%. As the simulated dew results agree well within 2% with the microlysimeter data, the model was considered acceptable for nighttime dew assessments in the study region. It must be noted, however, that although the model is verified mainly for the spring season only, it would apply to the fall season (mirror image of spring) and summer too. During summer and fall the meteorological conditions do not deviate extremely from the spring season. In winter, however, particularly under frost conditions or snow cover, the model would be less reliable. During frost and hoarfrost condition in winter we adapted s and γ in equation (4) for ice. However, it must be noted that this correction for ice did not affect our results much because the saturated specific moisture curve for vapor and ice deviate little from each other.

[27] To check that the dew model (equation (4)) performs well throughout all seasons, equation (4) was analyzed further. Equation (4) contains two different terms, an energy term and a deficit term. If the relative contribution of both terms is more or less equal, then this provides some confidence that the model performs well in all seasons. Accordingly, we analyzed the behavior of the ratio of both terms (deficit term divided by the energy term) throughout the year for the 11-year data set. This analysis showed no annual cycle in this ratio(= 0.58 ± 0.07). This result means that the energy term is the most important and that the standard deviation is relatively low. This supports the use of equation (4) for all seasons.

[28] Figure 4 displays the course of the annual dew and precipitation amounts based on the 11-year data record period. The averaged dew and precipitation during this period were 37 mm and 830 mm, respectively, with standard deviations of 8 mm and 200 mm, respectively. On average, dew contributes only about 4.5% of the mean annual precipitation. The mean annual dew amount is small in comparison to the mean precipitation and to the standard deviation of the precipitation.

[29] Figure 5 compares the mean monthly dew and precipitation amounts calculated for the selected 11-year data period, along with the trend functions (fourth-order polynomial). In order to correct for different days per month, the data in Figure 5 have been normalized for 30 days. The dew and precipitation amounts are more or less evenly distributed over the year. There appears to be a slight tendency for lower dew amounts during the longest daylight period (May to July), and vice versa during autumn. The standard deviation of the mean monthly dew is about 1 mm while the standard deviation of the mean precipitation is about 30 mm. Again, the contribution of dew to the mean monthly water balance is of minor importance.

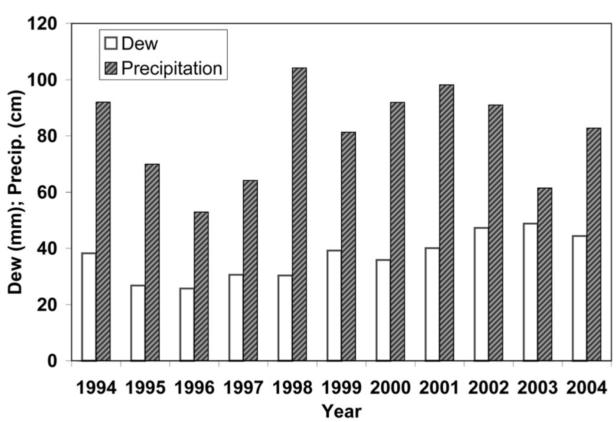

Figure 4. Course of the annual dew and precipitation amounts during the 11-year data period. Note to scale; dew in millimeters and precipitation in centimeters.

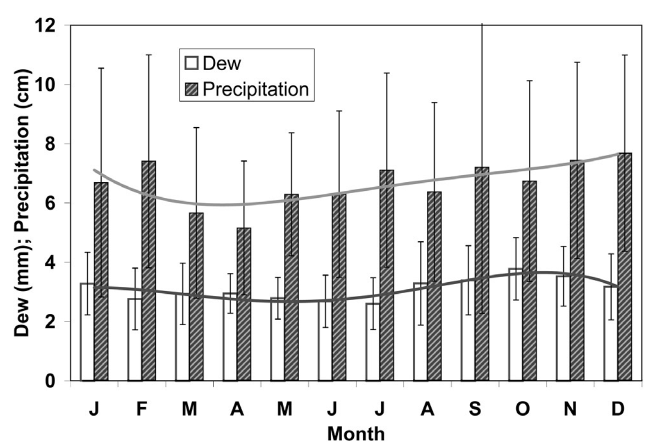

Figure 5. Course of the mean monthly dew and precipitation amounts (and standard deviations) during the 11-year data period. Note to scale; dew in millimeters and precipitation in centimeters.

[30] Figure 6 displays the number of annual dew nights and precipitation days for the 11-year data period. Here a dew night is defined as a night with more than 0.05 mm dew, and a precipitation day is defined as a day with more than 0.1 mm rain. On average the number of dew nights is 250 with a standard deviation of 25 nights, and the number of precipitation days is 190 with a standard deviation of 26 days. Dew occurs on nearly 70% of all nights, which is a very high frequency of dew events. The discrepancy between the total number of days with dew or rain, 440, versus 365 days per year, is that there were about 75 occurrences when it rained in the day but had conditions conducive to dew formation in the evening. However, 250 dew nights is not exceptional. In the northern Negev desert of Israel, *Evenari et al.* [1982] recorded about 200 dew nights per year. Although there was considerable dew research carried out by German scientists, in particular during the pre-1970 period, their publications focused primarily on measurement techniques and leaf wetness rather than on dew amounts. The latter issue is very important in the midlatitudes because of the development of fungal diseases.

[31] In the study region, precipitation occurs about 50% of the year. Figure 7 compares the distribution of the mean number of monthly dew nights and precipitation days. Moreover, the trend functions (fourth-order polynomial) have been plotted in Figure 7. The results are given in percentages in order to compensate for the different number of days per month. For the mean number of monthly averaged dew nights, a trend in the yearly cycle can be recognized. During the summer period (July to September) a maximum number of dew nights occur (about 25), while during the winter period (January until March) there is a minimum of about 17. The number of precipitation days tends to have an opposite pattern. In winter the polar jet stream lies over the region and brings many depressions accompanied by wet spells [*Peixoto and Oort*, 1992]. In summer, however, the path is more northern and causes fewer rainy spells. During the summer, however, there are more convective conditions with locally scattered convective showers during daytime. During nighttime, convective clouds disappear, leading to radiative cooling at the Earth's surface with accompanying dew and radiative fog formation.

[32] In spite of the meager contribution to the water budget, dew plays an important role in agriculture and ecology in the Netherlands. Leaf wetness and temperature combine to present conditions for pathogens and fungal and other foliar diseases that can endanger crop yield. Such diseases are often controlled by fungicide sprays. With increasing environmental awareness and the high cost of fungicides, there is a pressing need to curb excessive use of chemical control measures. Accurate determination of antecedent environmental conditions relevant to pathogen development can help to reduce fungicide use. Thus reliable estimates of leaf wetness duration will improve decision-making and assist in maximizing the efficiency of fungicide application.

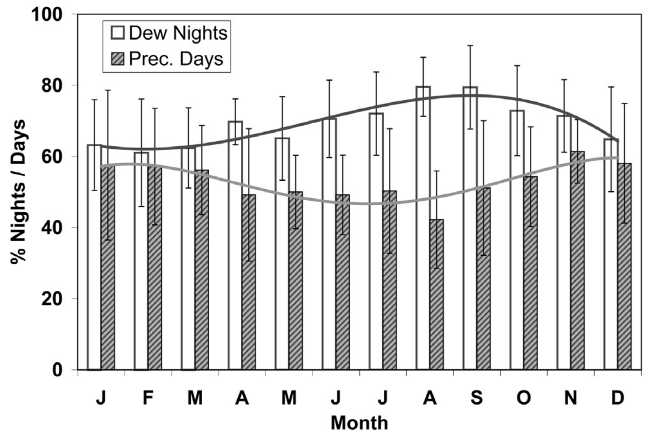

Figure 7. Percentage of monthly dew nights and precipitation days, and standard deviations and trend lines.

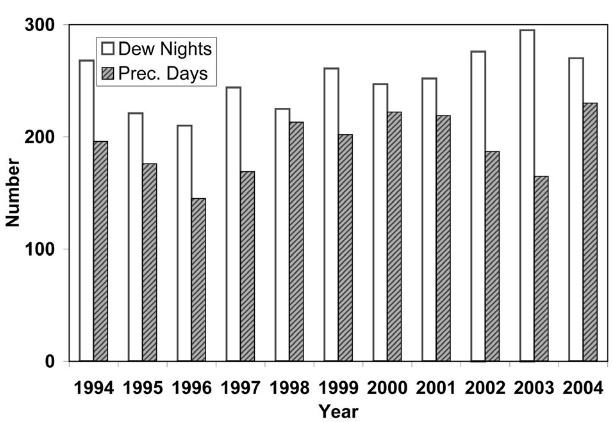

Figure 6. Number of annual dew nights and precipitation days during the 11-year data period.

4. Conclusions

[33] On the basis of field dew experiments and database analyses, nighttime dew amounts in a grassland region of the Netherlands were simulated for the spring, summer, and fall seasons using a surface energy budget dew model. During winter, the dew modeling results were less reliable due to frost conditions or snow cover.

[34] The eddy-covariance technique appeared to underestimate the nighttime fluxes under low wind conditions and very dewy nights and is thus unsuited for dew measurements.

[35] During the data period analyzed, the averaged annual dew amount was 37 ± 8 mm, which is about 4.5% of the mean annual precipitation of the study area (820 ± 200 mm). Hence dew is of minor importance in this region in terms of the total water budget.

[36] The monthly dew amounts are evenly distributed throughout the year with an average of 3.1 ± 1.0 mm. Thus dew also has no impact on the monthly water budget in this region.

[37] The annual averaged number of dew nights during the analyzed period was 250 ± 25, which is a high frequency of occurrence. The number of dew events is somewhat higher (about 80%) in the summer half of the year and somewhat lower (about 60%) in the winter half.

Notation

Roman

A area soil heat flux plate, m^2.

c_1 constant.

Cs	volumetric heat capacity soil, J m^{-3} K^{-1}.
D	accumulated dew, kg m^{-2}.
E	evapotranspiration/dewfall, kg m^{-2} s^{-1}.
G	soil heat flux, W m^{-2}.
g	gravity, m s^{-2}.
H	sensible heat flux, W m^{-2}.
L	Obukhov's stability length, m.
Q^*	net radiation, W m^{-2}.
q	specific humidity of air, kg kg^{-1}.
q^*	saturated specific humidity of air, kg kg^{-1}.
R_{Li}	incoming longwave radiation, W m^{-2}.
R_{Lo}	outgoing longwave radiation, W m^{-2}.
R_{gi}	incoming shortwave radiation, W m^{-2}.
R_{go}	outgoing shortwave radiation, W m^{-2}.
Ri_B	bulk Richardson number.
r_{av}	aerodynamic resistance to vapor, s m^{-1}.
s	slope specific moisture saturation curve, 1 K^{-1}.
T_a	dry bulb temperature, °C.
T_a(10 cm)	temperature at 10 cm height, °C.
T_{abs}	absolute temperature, K.
T_o	surface temperature, °C.
T_w	wet bulb temperature, °C.
t	thickness soil heat flux plate, m.
u	wind speed, m s^{-1}.
u_r	wind speed at reference height, m s^{-1}.
w	vertical velocity component, m s^{-1}.
z	height, depth, m.
z_r	reference height, depth, m.
z_o	roughness length for momentum, m.
z_{oh}	roughness length for heat, m.
z_{ov}	roughness length for vapor, m.
z_p	depth soil heat plate, m.
Greek	
ε	emissivity.
ε_p	ratio thermal conductivity soil heat plate to soil.
Δt	time step, s.
δq	deficit specific humidity, kg kg^{-1}.
Φ	instrumental shape correction function.
η	Monin Obukhov parameter.
γ	psychrometic constant, Pa K^{-1}.
κ	von Karman's constant.
λ_v	latent heat for vaporization, J kg^{-1}.
ρ	density, kg m^{-3}.
σ	Stefan Boltzmann constant, W m^{-2} K^{-4}.
ψ_m	integrated stability function for momentum.
ψ_v	integrated stability function for vapor.

[38] **Acknowledgments.** R. J. Wichink Kruit was supported by the Royal Dutch Health Institute (R.I.V.M.), project M/725501. We thank the anonymous referees for their comments and suggestions.

References

Aylor, D. E. (1986), A framework for examining inter-regional aerial transport of fungal spores, *Agric. For. Meteorol.*, *38*, 263–288.

Baier, W. (1966), Studies on dew formation under semiarid conditions, *Agric. Meteorol.*, *3*, 103–112.

Berkowicz, S. M., B. G. Heusinkveld, and A. F. G. Jacobs (2001), Dew in an arid ecosystem: Ecological aspects and problems in dew measurement, in *Proceedings of the Second International Conference on Fog and Fog Collection*, edited by C. J. R. S. Schemenauer and H. Puxbaum, pp. 301–304, Int. Dev. Res. Cent., Ottawa, Ont., Canada.

Beysens, D. (1995), The formation of dew, *Atmos. Res.*, *39*, 215–237.

Boast, C. W., and T. M. Robertson (1982), A "micro-lysimeter" method for determining evaporation from bare soil: Description and laboratory evaluation, *Soil Sci. Soc. Am. J.*, *46*, 689–696.

Brotzge, J. A., and K. C. Crawford (2003), Examination of the surface energy budget: A comparison of eddy correlation and Bowen ratio measurement systems, *J. Hydrometeorol.*, *4*, 160–178.

Dawson, T. E. (1998), Fog in the California redwood forest: Ecosystem inputs and use by plants, *Oecologia*, *117*, 476–485.

de Bruin, H. A. R., R. J. Ronda, and B. J. H. van der Wiel (1999), Approximate solutions for the Obukhov length and the surface fluxes in terms of the bulk Richardson numbers, *Boundary Layer Meteorol.*, *95*, 145–157.

Duvdevani, S. (1947), An optical method of dew estimation, *Q. J. R. Meteorol., Soc.*, *73*, 282–296.

Evenari, M., I. Shanan, and N. Tadmor (1982), *The Negev: Challenge of a Desert*, Harvard Univ. Press, Cambridge, Mass.

Fritschen, L. J., and L. W. Gay (1979), *Environmental Instrumentation*, Springer, New York.

Garratt, J. R. (1992), *The Atmospheric Boundary Layer*, Cambridge University Press, New York.

Garratt, J. R., and M. Segal (1988), On the contribution to dew formation, *Boundary Layer Meteorol.*, *45*, 209–236.

Gelbe, W. (1954), Ergebnisse dreijaehrige Taumessungen, *Phys. Verh.*, *9*, 195–202.

Heusinkveld, B. G., A. F. G. Jacobs, A. A. M. Holtslag, and S. M. Berkowicz (2004), Surface energy balance closure in an arid region: Role of soil and heat flux, *Agric. For. Meteorol.*, *122*, 21–37.

Heusinkveld, B. G., S. M. Berkowicz, A. F. G. Jacobs, A. A. M. Holtslag, and C. A. M. W.Hillen (2006), An automated microlysimeter to study dew formation and evaporation in arid and semi-arid regions, *J. Hydrometeorol.*, in press.

Hiltner, D. E. (1930), Der Tau und seine Bedeutung für den Pflanzenbau: Eine Theorie ueber die physiologische Bedeutung der Wasseraufnahme durch oberirdische Organe, *Wiss. Arch. Landwirt. A*, *3*, 1–70.

Holtslag, A. A. M., and H. A. R. de Bruin (1988), Applied modelling of the nighttime surface energy balance over land, *J. Appl. Meteorol.*, *27*, 689–704.

Jacobs, A. F. G., and J. P. Nieveen (1995), Formation of dew and the drying process within crop canopies, *Meteorol. Appl.*, *2*, 249–256.

Jacobs, A. F. G., B. G. Heusinkveld, and S. M. Berkowicz (2000a), Force-restore technique for ground surface temperature and moisture content in a dry desert system, *Water Resour. Res.*, *36*, 1261–1268.

Jacobs, A. F. G., B. G. Heusinkveld, and S. M. Berkowicz (2000b), Surface dew measurements across a linear dune transect, *Int. J. Biometeorol.*, *43*, 184–190.

Jacobs, A. F. G., B. G. Heusinkveld, and A. A. M. Holtslag (2003), Carbon dioxide and water vapour densities over a grassland area in the Netherlands, *Int. J. Climatol.*, *23*, 1663–1675.

Jacobs, A. F. G., B. G. Heusinkveld, and E. J. Klok (2005), Leaf wetness within a lily canopy, *Meteorol. Appl.*, *12*, 193–198.

Kessler, O. W. (1939), Der Tauschreiber Kessler-Fuess, *Bioklimatische Beil.*, *6*, 23–26.

Laubach, J., M. Raschendorfer, H. Kreilein, and G. Gravenhorst (1994), Determination of heat and water vapour fluxes above a spruce forest by eddy correlation, *Agric. For. Meteorol.*, *71*, 373–401.

Lomas, J. (1965), Note on dew duration recorders under semi-arid conditions, *Agric. Meteorol.*, *2*, 351–359.

Long, I. F. (1955), Dew and guttation, *Weather*, *10*, 128.

Malek, E., G. D. McCurdy, and B. Giles (1999), Dew contribution to the annual water balance in semi-arid desert valleys, *J. Arid Environ.*, *42*, 71–80.

Mogensen, V. O. (1970), The calibration factor of heat flux meters in relation to the thermal conductivity of the surrounding medium, *Agric. For. Meteorol.*, *7*, 401–410.

Monteith, J. L. (1957), Dew, *Q. J. R. Meteorol. Soc.*, *83*, 322–341.

Muselli, M., D. Beysens, and I. Mulimouk (2006), A comparative study of two large radiative dew water condensors, *J. Arid Environ.*, in press.

Nagel, J. F. (1962), On the measurement of dew, *Arch. Meteorol. Geophys. Bioklimatol., Ser. B*, *11*, 403–423.

Nieveen, J. P., D. I. Campbell, L. A. Schipper, and I. J. Blair (2005), Carbon exchange of grazed pasture on a drained peat soil, *Global Change Biol.*, *11*, 607–618.

Olivier, J., and C. J. Rautenbach (2002), The implementation of fog water collection systems in South Africa, *Atmos. Res.*, *64*, 227–238.

Pedro, M. J. Jr., and T. J. Gillespie (1982a), Estimating dew duration: II. Utilizing micrometeorological data, *Agric. Meteorol.*, *25*, 283–296.

Pedro, M. J. Jr., and T. J. Gillespie (1982b), Estimating dew duration: II. Utilizing standard weather data, *Agric. Meteorol.*, *25*, 297–310.

Peixoto, J. P., and J. P. Oort (1992), *Physics of Climate*, Am. Inst. of Physics, College Park, Md.

Schemenauer, R. S., and P. Cereceda (1994a), Fog collection's role in water planning for developing countries, *Nat. Resour. Forum*, *18*, 76–90.

Schemenauer, R. S., and P. Cereceda (1994b), A proposed standard fog collector for use in high elevation regions, *J. Appl. Meteorol.*, *33*, 1313–1320.

Snel, S. (2004), Soil-atmosphere exchange of CO_2, M.S. thesis, Dep. of Meteorol. and Air Qual., Wageningen Univ., Wageningen, Netherlands.

Steeneveld, G. J., B. J. H. van de Wiel, and A. A. M. Holtslag (2006), Modeling the evolution of the atmospheric boundary layer coupled to the land surface for a three day period in CASES-99, *J. Atmos. Sci.*, in press.

Steinberger, Y., I. Loboda, and W. Garner (1989), The influence of autumn dewfall on spatial and temporal distribution of nematodes in the desert ecosystem, *J. Arid Environ.*, *16*, 177–183.

Tuller, S. T., and R. Chilton (1973), The role of dew in a seasonal moisture balance of summer-dry climate, *Agric. Meteorol.*, *11*, 135–142.

Van de Wiel, B. J. H., A. F. Moene, O. K. Hartogensis, H. A. R. de Bruin, and A. A. M. Holtslag (2003), Intermittent turbulence in the stable boundary layer over land: III. A clasification for observations during CASES-99, *J. Atmos. Sci.*, *60*, 2509–2522.

Wallin, J. R. (1967), Agricultural aspects of dew, *Agric. Meteorol.*, *4*, 85–102.

Zangvil, A. (1996), Six years of dew observations in the Negev Desert, Israel, *J. Arid Environ.*, *32*, 361–371.

S. M. Berkowicz, Minerva Arid Ecosystems Research Centre, Hebrew University of Jerusalem, Earth Science Building, Safra Givat Ram Campus, Jerusalem, Israel 91904.

B. G. Heusinkveld, A. F. G. Jacobs, and R. J. Wichink Kruit, Department of Meteorology and Air Quality, Wageningen University, Duivendaal 2, NL-6701 APWageningen, Netherlands. (adrie.jacobs@wur.nl)