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Factors affecting dewfall, its measurement with lysimeters, and its estimation with micrometeorological equations

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Summary

Introduction to the topic

Dew is a form of precipitation, which is usually not explicitly considered in hydrologic studies, because the amounts are small. However, in semiarid and arid regions dewfall can reach or even exceed all other forms of precipitation for extended periods or indeed a whole year. Even in humid regions it can be the biggest type of precipitation over shorter periods such as a week or a month, although dewfall normally contributes only a small percentage to the total annual precipitation. There is ample evidence that even small amounts of dewfall can be beneficial to plants, not only in arid and semiarid, but also in humid regions.

For these reasons it is worth to have a closer looks at some aspects related to dewfall. The specific objectives of this thesis are:

- to identify the meteorological factors affecting dew formation and to quantify their effect,
- to examine the weighing precision of the type of lysimeter used later in this study to asses its suitability for dew studies,
- to quantify the amount of dew precipitated on various types of vegetation in northern Germany during a night and during the course of a year using lysimeters,
- to assess different methods to compute dewfall, and
- to compare dewfall measured with lysimeters and dewfall computed with the Bowen ratio energy balance and the Penman-Monteith equation.

All field work for this thesis was carried out at the Falkenberg lysimeter station of the Department of Soil Physics of the Helmholtz Centre for Environmental Research - UFZ. The station is located in northern Germany, some 120 km north-west of Berlin.

Analysis of the effect of meteorological factors on dew formation

Dew forms when the temperature of a surface cools below the dew point temperature of the surrounding air. As a result vapour contained in this air condenses on the surface. The cooling of a surface is caused by a radiation loss. How much it cools down depends on the weather conditions.

The energy balance equation was analysed to identify the meteorological factors which determine the degree of cooling, and to assess their effect on dew formation. These factors were found to be air temperature (T_a), cloud cover (N), wind speed (u), soil heat flux (G), and relative humidity (h_r). The temperature of the surface (T_s) dew forms on is also important. However, it is not a meteorological factor, but determined by the aforementioned variables. Also, net radiation was not considered explicitly, but indirectly by looking at the effect of temperature on the radiation balance.

All other conditions equal, the analysis revealed that dewfall increases linearly as N or G decrease, or h_r increases. The effect of T_a and u on dewfall is non-linear. It first rises with T_a or u and then falls again. All five meteorological factors identified here have roughly the same impact. Each can lead to variations in dewfall between 0 and ~25 W m^{-2} (0 to ~0.04 mm h^{-1}), the precise magnitude of the impact depends on the value of the other factors. Dewfall is always highest if N = 0, G = 0 and h_r = 1. At which air temperature dewfall is highest depends on wind speed and vice versa.

Testing the precision of a weighable gravitation lysimeter

Tests were carried out to determine the weighing precision of a 2 m deep lysimeter with a 1 m^2 cross-sectional area and a total mass of 3,500 to 3,850 kg, depending on the soil water content. The weighing mechanism consists on three shear stress cells laid out for a load capacity of 1,320 kg each.

Mass changes as small as 20 g, which is equivalent here to a water gain or loss of 0.02 mm, can be measured with good accuracy and stability under favourable environmental conditions (low wind speed and relatively constant temperature). This precision does not depend on the position on the lysimeter where the mass change occurs, and is as good as the best values reported in the literature for other lysimeters.

To prevent water and debris from entering the cleavage between lysimeter vessel and pit casing, a rubber collar can be placed across the cleavage. It is attached to the casing and extends about 1 to 2 cm into the vessel. Although the collar is not supposed to touch the vessel, it does at a few points. This seriously lowers weighing precision, because this contact exerts forces on the vessel, which distort the true weight. Hence, one should refrain from using this type of collar and develop another one.

Weighing precision decreases with increasing wind speed, because wind exerts forces on the lysimeter vessel and can thus alter its apparent weight. It is temperature-dependent, too.

Effect of vegetation type and growth stage on dewfall, determined with high precision weighing lysimeters at a site in northern Germany

The amount and temporal distribution of dewfall on grass, maize and winter barley was measured with four high precision weighing lysimeters at Falkenberg in northern Germany during 2004 and 2005 to quantify the contribution of dewfall to the water balance of the region, and to assess how dewfall is affected by the vegetation cover.

Two lysimeters were under continuous grass, two were cropped (maize from April through September 2004, followed by winter barley until July 2005, fallow the rest of the time). Observed dewfall ranged from 27.1 to 31.8 mm per year, which was 5.5 to 6.9% of the annual rainfall. In several months of the study period dewfall was > 20% of the monthly precipitation.

On fallow lysimeters there were fewer nights with dewfall and less dewfall per event than on lysimeters with grass. After crops were planted the number of dewfall-nights and the amount of dewfall per event rose quickly and eventually surpassed that on the lysimeters with grass. After harvest both parameters dropped well below the values on the grass lysimeters again.

Assessment of four methods to compute dewfall

High precision weighing lysimeters are an effective tool to quantify dewfall, but they are not wide-spread due to their high cost. One alternative is to compute dewfall from meteorological data and under consideration of the properties of the surface in question. Four equations, which were shown in the literature to work for this purpose, were assessed.

Three of them, the energy balance (EB), turbulent vapour transport (TVT) and Penman-Monteith (PM) equation, contain a heat and/or vapour conductance term. To get a correct value for it requires a wind profile in equilibrium with the vegetation under investigation. This was apparently not the case under the conditions at the Falkenberg. Hence, the EB and TVT equation could not be used successfully without adjusting the conductance term. The PM equation was less beset by this problem, because the conductance term in it is small at high relative humidities (which correspond to low vapour pressure deficits) common during dew events. The Bowen ratio energy balance (BREB) equation was found to work best, because it lacks a conductance term.

The BREB, EB and TVT equation need the temperature of the surface, which is usually not available. This leaves the PM equation, from which it has been eliminated, as the only option then.

Comparison between measured and computed dewfall

Dewfall measured with a lysimeter was compared to dewfall computed with the Bowen ratio energy balance (BREB) and Penman-Monteith (PM) equation, which were previously identified as the most suitable equations for this purpose.

Measured and calculated data follow a similar tendency, but the calculated values are consistently and significantly higher. Also, the measured dewfall fluctuates considerably, while the calculations show continuous dewfall at a similar rate throughout the night. Finally, compared to the lysimeter record the two equations predict dewfall too early and for too long. Despite these differences the cumulative measured and calculated values correlate quite well for the hours when both the lysimeter and the equations show dewfall.

Estimates with the BREB and PM equation are in very good agreement, but the latter always yields slightly higher values. This largely disappears, if the slope of the saturation vapour pressure curve, which enters into the PM equation, is evaluated at the mean of air and surface temperature rather than just at air temperature, which is the normal procedure.

It was pointed out before that under the site conditions at Falkenberg it is difficult to get proper values for the heat and vapour conductance. The latter enters into the PM equation as a multiplier of the vapour pressure deficit. Because this was low during the observation period, the conductance had little effect on the results.

For the period considered here net radiation (R_n) and soil heat flux (G) were estimated with empirical equations, because no measured data were available. A comparison of measured values and data computed with these equations for a period, when measured data for R_n and G as well as all the parameters needed to compute them were available, revealed that errors in the estimation of R_n and G could be the reason for the observed deviations between measured and computed dewfall.

Further measurements are planned so that eventually a full set of measured data will be available to check the performance of the BREB and PM equation against lysimeter data.

Zusammenfassung

Einleitung zum Thema

Tau ist eine Form von Niederschlag, die in der Regel nicht ausdrücklich in hydrologischen Studien betrachtet wird, da die Mengen klein sind. In semi-ariden und ariden Regionen erreicht oder übersteigt Tau allerdings alle anderen Formen des Niederschlags über längere Zeit oder sogar ein ganzes Jahr. Selbst in humiden Regionen kann er über kürzere Zeiträume, wie eine Woche oder einen Monat, die größte Menge zum Niederschlag beitragen, obwohl Tau normalerweise nur einen geringen Prozentsatz der gesamten jährlichen Niederschlagsmenge ausmacht. Es gibt viele Hinweise, dass selbst kleine Mengen von Tau für Pflanzen vorteilhaft sein können, nicht nur in ariden und semi-ariden, sondern auch in humiden Regionen.

Aus diesen Gründen lohnt es sich, einen genaueren Blick auf einige Aspekte im Zusammenhang mit Tau zu werfen. Die spezifischen Ziele dieser Arbeit sind:

1) Die meteorologischen Faktoren, die Taubildung beeinflussen, zu identifizieren und ihre Wirkung zu quantifizieren.

2) Die Wägegenauigkeit eines Lysimetertyps zu überprüfen, der in dieser Studie später verwendet wird, um seine Eignung für Taustudien einzuschätzen.

3) Die Taumenge auf unterschiedlichen Vegetationsformen im nördlichen Deutschland in einer Nacht und im Laufe eines Jahres mit Lysimetern zu quantifizieren.

4) Verschiedene Methoden zur Berechnung des Taufalls zu bewerten.

5) Gemessene und berechnete Taumengen zu vergleichen.

Die Geländearbeiten für diese Dissertation wurden in der Lysimeterstation Falkenberg des Departments Bodenphysik des Helmholtz Zentrums für Umweltforschung – UFZ durchgeführt. Sie liegt in Norddeutschland, etwa 120 km nordwestlich von Berlin.

Analyse des Effekts meteorologischer Faktoren auf Taubildung

Tau bildet sich, wenn die Temperatur einer Oberfläche unter die Taupunkttemperatur der umgebenden Luft abkühlt. Infolgedessen kondensiert Wasserdampf, der in dieser Luft enthalten ist, auf der Oberfläche. Die Abkühlung einer Oberfläche wird durch einen Strahlungsverlust verursacht. Wie viel sie abkühlt, hängt von den Wetterbedingungen ab.

Die Energiebilanzgleichung wurde analysiert, um die meteorologischen Faktoren zu identifizieren, die den Grad der Abkühlung beeinflussen, und um ihre Wirkung auf die Taubildung zu bewerten. Als Faktoren wurden aufgezeigt: Lufttemperatur (T_a), Bewölkungsgrad (N), Windgeschwindigkeit (u), Bodenwärmestrom (G) und relative Luftfeuchtigkeit (h_r). Die Temperatur der Oberfläche (T_s), an der sich Tau bildet, ist auch wichtig. Jedoch ist sie kein meteorologischer Faktor, sondern wird durch die eben erwähnten Faktoren bestimmt. Die Nettostrahlung wurde nicht ausdrücklich betrachtet, sondern indirekt durch den Effekt der Temperatur auf die Strahlungsbilanz berücksichtigt.

Wenn alle anderen Bedingungen gleich waren, ergab die Analyse, dass die Taumenge linear ansteigt, wenn N oder G abnehmen oder h_r zunimmt. Der Effekt von T_a und von u auf Tau ist nicht linear. Zuerst steigt die Taumenge mit T_a oder u an und fällt dann wieder. Alle fünf meteorologischen Faktoren, die hier identifiziert wurden, haben ungefähr die gleiche potenzielle Auswirkung auf die Taubildung. Jeder kann zu Schwankungen zwischen 0 und ~25 W m^{-2} (0 bis ~0,04 mm h^{-1}) führen. Das genaue Ausmaß der Auswirkung hängt vom Wert der anderen Faktoren ab. Taubildung ist immer dann am höchsten, wenn N = 0, G = 0 und h_r = 1. Bei welcher Lufttemperatur die Taubildung am höchsten ist, hängt von der Windgeschwindigkeit ab und umgekehrt.

Prüfung der Genauigkeit eines wägbaren Gravitationslysimeters

Tests wurden durchgeführt, um die Wiegegenauigkeit eines 2 m tiefen Gravitationslysimeters mit einer Oberfläche von 1 m² und einer Gesamtmasse von 3.500 bis 3.850 kg, die abhängig vom Bodenwassergehalt ist, zu ermitteln. Der Wägemechanismus besteht aus drei Scherkraftzellen mit einer Nutzlast von je 1.320 kg.

Massenänderungen von nur 20 g, die hier einem Wassergewinn oder -verlust von 0,02 mm entsprechen, können unter günstigen Bedingungen (niedrige Windgeschwindigkeit und relativ konstante Temperatur) mit guter Genauigkeit und Stabilität gemessen werden. Diese Genauigkeit hängt nicht von der Position auf dem Lysimeter ab, an der die Massenänderung auftritt, und ist so gut wie die besten Werte, die in der Literatur für andere Lysimeter angegeben sind.

Um zu verhindern, dass Wasser und Schmutz in den Spalt zwischen Lysimetergefäß und Grubenummantellung eindringt, wird ein Gummikragen über den Spalt installiert. Er ist mit der Ummantellung verbunden und reicht 1 bis 2 cm über das Gefäß. Obwohl der Kragen nicht das Gefäß berühren soll, tut er es an einigen Punkten. Dieses senkt die Genauigkeit des Wiegens, weil dieser Kontakt Kräfte auf das Gefäß ausübt, die sein Gewicht verzerren. Folglich sollte man diesen Kragen nicht verwenden und einen anderen entwickeln.

Die Wägegenauigkeit verringert sich bei Zunahme der Windgeschwindigkeit, weil Wind Kräfte auf den Lysimeterbehälter ausüben und sein Gewicht verfälschen kann. Sie ist außerdem temperaturabhängig.

Effekt der Vegetationsart und ihres Wachstumstadiums auf Taumengen, bestimmt mit hochgenauen wägbaren Lysimetern an einem Standort in Norddeutschland

Die Menge und die zeitliche Verteilung von Taumengen auf Gras, Mais und Wintergerste wurden mit vier hochgenauen wägbaren Lysimetern an einem Standort in Norddeutschland während 2004 und 2005 gemessen. Ziel war es zu quantifizieren, wie hoch der Beitrag von Tau zur Wasserbilanz der Region ist, und zu beurteilen, wie Taubildung durch die Vegetation beeinflusst wird.

Zwei Lysimeter waren ständig unter Gras, zwei waren mit Feldfrüchten bepflanzt (Mais von April bis September 2004, gefolgt von Wintergerste bis Juli 2005, Brache den Rest der Zeit). Die ermittelte Taumenge reichte von 27,1 bis 31,8 Millimeter pro Jahr, was 5,5 bis 6,9% der jährlichen Niederschlagsmenge entsprach. In einigen Monaten des Studienzeitraums war der Taufall > 20% der monatlichen Niederschlagsmenge.

Es gab weniger Nächte mit Tau und weniger Tau pro Ereignis auf brachen Lysimetern als auf Lysimetern unter Gras. Nachdem Getreide gepflanzt wurde, stiegen die Zahl der Nächte mit Tau und die Taumenge pro Ereignis schnell an und übertrafen schließlich die auf den Lysimetern mit Gras. Nach der Ernte sanken beide Parameter wieder deutlich unter die Werte der Graslysimeter.

Beurteilung von vier Methoden zur Berechnung von Taumengen

Lysimeter sind ein geeignetes Instrument, um Taumengen zu quantifizieren, aber sie sind nicht weit verbreitet wegen ihrer hohen Kosten. Eine Alternative ist, den Taufall auf Basis meteorologischer Daten und unter Berücksichtigung der Eigenschaften der Oberfläche (Vegetation) zu berechnen. Vier Gleichungen, die in der Literatur als für diesen Zweck geeignet aufgezeigt wurden, wurden verwendet.

Drei davon, die Energiebilanz-Gleichung (EB), die Gleichung für turbulenten Wasserdampftransport (TVT) und die Penman-Monteith-Gleichung (PM), enthalten einen Wärme- und/oder Dampfleitfähigkeitsausdruck. Um einen korrekten Wert dafür zu erhalten, ist ein Windprofil im Gleichgewicht mit der Vegetation erforderlich. Dies war offenbar unter den Bedingungen in Falkenberg nicht der Fall. Folglich konnten die EB- und TVT-Gleichung nicht erfolgreich verwendet werden, ohne den Leitfähigkeitsausdruck anzupassen. Die PM-Gleichung war durch dieses Problem weniger betroffen, weil der Leitfähigkeitsausdruck darin bei den hohen relativen Luftfeuchtigkeiten (gleichbedeutend mit geringen Wasserdampfdruckdefiziten), die während eines Tauereignisses oft vorkommen, klein ist. Es zeigte sich, dass die Bowen-ratio Energiebilanz-Gleichung (BREB) am Besten funktioniert, weil sie keinen Leitfähigkeitsausdruck enthält.

Die BREB-, EB- und TVT-Gleichungen benötigen die Temperatur der Oberfläche, die in der Regel nicht verfügbar ist. Dann ist die PM-Gleichung, aus der sie entfernt wurde, die einzige Wahl.

Vergleich zwischen gemessenen und berechneten Taumengen

Mit einem Lysimeter gemessene Taumengen wurden mit Taumengen, die mit der Bowen-ratio Energiebilanz-Gleichung (BREB) und der Penman-Monteith-Gleichung (PM) berechnet wurden, verglichen. Diese beiden Gleichungen wurden vorher als die Besten für diesen Zweck identifiziert.

Gemessene und berechnete Daten folgen einer ähnlichen Tendenz, aber die berechneten Werte sind durchgehend und erheblich höher. Die gemessene Taumenge schwankt auch beträchtlich, während die Berechnungen ununterbrochene Taubildung mit einer ähnlichen Rate in der gesamten Nacht zeigen. Verglichen mit den Lysimeterdaten geben die zwei Gleichungen die Taubildung zu früh und für zu lange an. Trotz dieser Unterschiede korrelieren die kumulativen gemessenen und berechneten Werte ziemlich gut für die Stunden, in denen sowohl das Lysimeter als auch die Gleichungen Taubildung anzeigen.

Die Schätzungen mit der BREB- und der PM-Gleichung stimmen sehr gut überein, aber letztere bringt immer etwas höhere Werte. Dieser Unterschied verschwindet weitgehend, wenn die Steigung der Sättigungsdampfdruckkurve, die in der PM-Gleichung enthalten ist, für das Mittel der Luft- und Oberflächentemperatur berechnet wird, und nicht für die Lufttemperatur, was die normale Vorgehensweise ist.

Es wurde bereits vorher darauf hingewiesen, dass es unter den Bedingungen in Falkenberg schwierig ist, korrekte Werte für die Wärme- bzw. Wasserdampfleitfähigkeit zu erhalten. Letztere tritt in der PM-Gleichung als Multiplikator des Dampfdruckdefizits auf. Da dieses während des Beobachtungszeitraums niedrig war, hatte die Leitfähigkeit praktisch keine Wirkung auf die Ergebnisse.

Für den Betrachtungszeitraum hier wurden Nettostrahlung (R_n) und Bodenwärmestrom (G) mit empirischen Gleichungen geschätzt, weil keine gemessenen Daten vorhanden waren. Ein Vergleich der gemessenen Werte und der Daten, die mit diesen Gleichungen während eines Zeitraums berechnet wurden, als gemessene Daten für R_n und G sowie alle anderen notwendigen Parameter vorhanden waren, ergab, dass der Fehler bei der Schätzung von R_n

und G der Grund für die Abweichungen zwischen gemessenen und berechneten Taumengen in diesem Kapitel sein könnte.

Weitere Messungen sind geplant, um letztlich einen kompletten Satz gemessener Daten zu erhalten. Damit können dann Berechnungen mit der BREB- und PM-Gleichung mit Lysimeterdaten zuverlässig verglichen werden.

List of variables

Variable	Description	Units
a	constant (0.611)	kPa
b	constant (17.502)	dimensionless
c	constant (240.97)	K
c_p	heat capacity of air	J mol^{-1} K^{-1}
d	zero plane displacement	m
e_a	vapour pressure of the ambient air	kPa
e_s	vapour pressure of the air at the surface	kPa
$e_s(T_a)$	saturation vapour pressure at air temperature	kPa
$e_s(T_s)$	saturation vapour pressure at surface temperature	kPa
f_1	constant (9.35×10^{-6})	K^{-2}
f_2	constant (60)	W m^{-2}
G	soil heat flux	W m^{-2}
g_H	heat conductance of the air	mol m^{-2} s^{-1}
g_v	vapour conductance of the air	mol m^{-2} s^{-1}
H	sensible heat flux	W m^{-2}
H_u	upward sensible heat flux	W m^{-2}
h	crop height	m
h_r	relative humidity of the air	%
k	von Karman constant (0.41)	dimensionless
k_T	soil thermal conductivity	W m^{-1} k^{-1}
LAI	leaf area index	dimensionless
L_{in}	incoming long wave radiation	W m^{-2}
L_{out}	outgoing long wave radiation	W m^{-2}
M	energy released or required by plant metabolic processes	W m^{-2}
N	cloud cover	dimensionless
P	air pressure	kPa
R_n	net radiation	W m^{-2}
s	slope of the temperature - saturation vapour pressure curve	kPa K^{-1}
S_{in}	incoming short wave radiation	W m^{-2}

T	mean of air and surface temperature	K
T_a	air temperature	K
T_i	temperature inside the canopy	K
T_s	surface temperature	K
T_5	soil temperature at 5 cm depth	K
T_{10}	soil temperature at 10 cm depth	K
u	wind speed	$m\ s^{-1}$
z	height at which wind speed is measured	m
z_0	roughness coefficient	m
α	surface reflectance for short wave radiation (albedo)	dimensionless
β	Bowen ratio	dimensionless
γ	light extinction coefficient	dimensionless
Δe	vapour pressure difference between two points of measurement	kPa
Δe_s	saturated vapour pressure difference between two points of measurement	kPa
ΔT	temperature difference between two points of measurement	K
Δz	distance between two points of measurement	m
ε_s	emissivity of the surface for long wave radiation	dimensionless
ε_{sky}	emissivity of the sky for long wave radiation	dimensionless
λ	latent heat of vapourisation	$J\ mol^{-1}$
λE	latent heat flux	$W\ m^{-2}$
λE_{BREB}	latent heat flux (dewfall) estimated with the Bowen ratio energy balance equation	$W\ m^{-2}$
λE_{EB}	latent heat flux (dewfall) estimated with the energy balance equation	$W\ m^{-2}$
λE_{PM}	latent heat flux (dewfall) estimated with the Penman-Monteith equation	$W\ m^{-2}$
λE_{TVT}	latent heat flux (dewfall) estimated with the turbulent vapour transport equation	$W\ m^{-2}$
λE_u	upward latent heat flux (dewrise)	$W\ m^{-2}$
ρ_a	molar density of the air	$mol\ m^{-3}$
σ	Stefan-Boltzmann constant (5.67×10^{-8})	$W\ m^{-2}\ K^{-4}$

φ	adjustment factor	dimensionless
ψ	atmospheric stability factor	dimensionless
ω	proportionality factor	dimensionless

1. Introduction to the topic

1.1 The significance of dew

Atmospheric water vapour which has condensed on surfaces of objects is referred to as dew. Dew condenses when the temperature of a surface is at or below the dew-point temperature of the ambient air. Dew is helpful for plants and animals (Stone, 1957a, b; Wallin, 1967; Jacobs et al., 1999; Liu et al., 2001; Li, 2002). Many studies concerning natural condensation of atmospheric moisture are oriented towards agriculture and the study of the soil cooling (Awanou and Hazoume, 1997; Marek and Straub, 2001).

Dew is a form of precipitation, which is usually not explicitly considered in soil moisture studies, because the mass flux is small, rarely exceeding 0.3 - 0.5 mm per night on vegetation (Garratt and Segal, 1988). In most regions of the world dew contributes only a small percentage to the annual precipitation, but the presence of water from dew has important implications for several physical and biological processes, especially but not exclusively in semiarid and arid ecosystems, which are normally characterised by low or negligible rainfall and where water availability is the most important limiting growth factor. Dew has been investigated in a number of such countries, e.g. Tanzania (Nilsson, 1996), Israel (Zangvil, 1996; Kidron et al., 2000) and Bahrain (Alnaser and Barakat, 2000).

In an arid ecosystem the availability of water governs both the survival and survival strategies of plants and animals. Dewfall is a process whereby moisture from the atmospheric water reservoir condenses on the earth surface. Together with sporadic rainfall episodes the frequent occurrence of dew serves as an important source of moisture for animals (Acostav Baladón and Gioda, 1991; Degen et al., 1992; Moffett, 1985), plants (Evenari et al., 1982; Larmuth and Harvey, 1978; Simon et al., 1994), and biological crusts. The latter can contribute to the stabilisation of sand dunes (Danin et al., 1989; Lange et al., 1992). Subramaniam and Kesava Rao (1983) also commented on the possible role of dew in stabilising sand dunes in the Rajasthan desert of India. Furthermore, where the percentage of vegetation cover is low, the moisture can stabilise the upper layer of an arid soil (Jacobs et al., 1999).

In semiarid and arid areas dew can make up a great portion of the total annual precipitation. It can be a significant source of moisture for plants (Baier, 1966) and for the bacteria of biological crusts (Lange et al., 1992, 1998). In regions of Israel, for example, the deposition of dew on plants often approaches or exceeds the annual rainfall (Ashbel, 1949). In India Subramaniam and Kesava Rao (1983) found dew to be 37% of the actual seasonal rainfall in 1975 - 76, and 27% of the usual seasonal rainfall. This was about 14% of the actual potential evapotranspiration and 18% of the normal potential evapotranspiration. Knoche (1939) stated that sometimes the daily dewfall is comparable to a light rainfall, since dew can cause uniform wetting of a large area. Tuller and Chilton (1973) found that dewfall was normally 12 - 14% of the monthly rainfall, but 154% thereof in a dry period.

The ecological significance of dew has long been recognised, especially in the growing season of the vegetation. The prime benefits of dew are that it improves the internal water balance of plants, because transpiration is reduced during the morning hours when dew is still present on the leaves, and that it cools the leaves (Baier, 1966; Tuller and Chilton, 1973; Stewart, 1977; Barradas and Glez-Medellin, 1999). Dew also prolongs the survival of tree seedlings (Stone, 1957b; Fritschen and Doraiswamy, 1973), delays the wilting of leaves (Stone, 1957b; UNESCO, 1958) and increases the humidity in epiphyte microhabitats in tropical canopies (Stone, 1957b). Steubing and Casperson (1959) found that winter barley plants protected from dew were 6 cm shorter than plants exposed to dew. Differences in blooming time were very distinct, too. Mustard, lupine and potato plants protected from dew flowered earlier. It is known that drought hastens flower formation. Also, condensation releases a small amount of latent heat, which retards nocturnal cooling of exposed portions of a plant.

On the other hand, not all the effects of dew on vegetation are beneficial. Dew is related to some problems found in agriculture. Since condensed vapour results in free water on the leaf surface, the formation of a film of water on plant leaves contributes to the development of bacteria, fungal pathogens and plant epidemics (Wallin, 1963; Buttler, 1980; Pedro and Gillespie, 1982a, b; Royle and Butler, 1986; Garratt and Segal, 1988; Uehara et al., 1988; Jacobs et al., 1990; Horino et al., 1990; Lhomme and Jimenez, 1992; Wilson et al., 1999; Luo and Goudriaan, 1999, 2000). In Germany dew rather than rain is considered more important in the duration of leaf wetness and resulting apple scab development (van Eimern, 1964). Dew coincident with certain temperatures is an important factor in the development of Cercospora beticola on sugar beets (Mischke, 1960).

Another problem is the increased deposition of pollutants on a condensing surface. Studies on dew composition show that the pollutant concentration is usually very high, especially the concentration of weak acids and ammonia (Wisniewski, 1982; Janssen and Römer, 1991; Hughes and Brimblecombe, 1994; Okochi et al., 1996; Khare et al., 2000; Takenaka et al., 2003). At present there is also an interest in dew formation on canopies with regard to its potential significance as a surface sink in the dry deposition of water-soluble gases, e.g. sulphur dioxide. Subsequent evaporation of dew after sunrise may result in a residue of sulphur compounds attached to the foliage with potential harmful effects to the plant canopy (Garratt and Segal, 1988).

1.2 The physical process of dew formation

The formation of dew through the condensation of water vapour on many types of surfaces, e.g. grass and crops, roofs and vehicles, is an everyday experience. Dew normally occurs during night-time or early in the morning as a result of a radiative loss of heat from the soil or a leaf, followed by condensation of water vapour. Dew formation is favoured by light winds. Dew is visible only when atmospheric moisture condenses at a rate greater than that at which it evaporates again.

Radiative cooling may result in surface temperatures several degrees below the ambient air temperature and thus actually allow for dew to form when the relative humidity of the surrounding air is significantly below 100% (Monteith and Szeicz, 1961).

The amount of dew formed on a surface depends on how far the radiation loss is balanced by a gain through sensible heat flux, latent heat flux, heat conduction and heat storage (Madeira et al., 2002). The sensible heat flux is proportional to the temperature gradient between the surface of dew deposition and the atmosphere, while the latent heat flux is proportional to the vapour pressure gradient. These two fluxes also depend on a transfer coefficient, which in turn depends on wind speed (Burrage, 1972). Heat conduction and storage are determined by the thermal properties of the soil or plant material.

The magnitude of radiative cooling is mainly dependent on meteorological factors, but the thermal properties of the surface play an important role as well. In response to nocturnal radiative cooling the air temperature falls continuously, while the relative humidity increases. This is accompanied by a release of latent heat as dew condenses. The transfer of water vapour to the surface is essentially a function of the vapour pressure gradient and a transfer coefficient involving molecular diffusion near the surface of condensation, and of turbulent transfer from greater distances. A good analysis of the various factors affecting dew formation is given by Monteith (1957, 1963) and Garratt and Segal (1988).

There was a long controversy with respect to the origin of dew. Now it is well recognised that dew on soil or vegetation surfaces can originate from dewrise or distillation (flux of water vapour from the soil to a surface), dewfall or condensation (flux of vapour from the atmosphere to a surface), and guttation. Monteith (1957) was the first to make this important distinction. In conditions of very light wind (0.5 m/s or less at 2 m) and clear sky, which lead to rapid cooling, dew arises mostly from distillation (Monteith, 1957). Guttation is the exudation of water (or sap, to be more precise) by leaves (Newton and Riley, 1964). Since guttation accounts for the smallest amount of dew and has been found to be insignificant (Atzema et al., 1990), it is usually not considered in studies of dew formation.

Nocturnal radiative loss plays an important role in the dew formation process (Luo and Goudriaan, 2000). Hence, it is a direct or indirect input into many dew formation simulation models that use the energy balance approach (e.g. Goudriaan, 1977; Pedro and Gillespie, 1982a, b; Jacobs et al., 1990; Wittich, 1995; Wilson et al., 1999). Such models are used to simulate dew duration or dewfall. Dew determined by the direction of the latent heat flux. If it is directed towards the surface, dew occurs.

Besides the moisture and temperature condition of the air and the soil, as well as various meteorological factors, the characteristics of the surface dew may deposit on are an important consideration, too. In Germany, for example, Steubing (1952) expressed dewfall for different forest covers in per cent of dewfall on a meadow. Dewfall was highest on the meadow, while dew deposition in mid-forest was only 1% of that on the meadow. Wallin (1967) reports that dewfall was greater on loose than on compact soil.

The formation of dew is linked to aerodynamics and thermodynamics, especially near-surface meteorological parameters and surface properties. Micrometeorological conditions associated with the frequent occurrence of dew have been known for some time. Neumann (1956), Monteith (1957, 1963) and Long (1958) suggested that dewfall represents a flux of latent heat towards the surface, the opposite of evaporation, and discussed the weather conditions for dew formation. It was recognised that sufficient moisture in the air and intensive radiative cooling of the surface are two basic requirements for dew formation so that temperature and humidity can be inferred as two important meteorological factors. In addition, Baier (1966) indicated that clear skies and light winds are favourable for dew formation. Clear skies allow more outgoing radiation and thus accelerate radiative cooling (Newton and Riley, 1964). Light winds bring more moisture to the surface.

More recently other studies on the relationship between dew formation and various meteorological factors were carried out. Based on computations Garratt and Segal (1988) evaluated the dependence of dew amount on wind speed, air temperature, atmospheric stability, relative humidity, cloudiness and soil characteristics, all of which are significant factors. Their work provides some refinements to a similar study by Monteith and Szeicz (1961). Scherm and van Bruggen (1993a) carried out a sensitivity analysis to compare the effects of temperature, atmospheric humidity (vapour pressure), cloud cover, and wind speed on dew duration at various sites. They found that at coastal sites humidity and cloud cover were the most sensitive factors, while at inland sites and wind speed was.

1.3 Methods used in the dew studies

There is no universal procedure for the measurement of dew, although there is a long history of dew research (Richards, 2004). Because of the usually small amount, dew cannot be measured with standard rain gauges. Hence, various types of dew gauges have been developed (e.g Leick, 1932; Kessler, 1939; Duvdevani, 1947; Hirst, 1954, 1957; Lloyd, 1961; Nagel, 1962). Many materials have been used as a surface for dew to condense on (Nagel, 1962; Myers, 1974; Wales-Smith, 1983; Luo and Goudriaan, 2000a), but the results are not satisfactory (Richards, 2004). A widely used device is the Leick plate, a non-hygroscopic porcelain disk, which can be placed inside a crop canopy such as maize (e.g. Jacobs et al., 1994). Another frequently employed technique (e.g. Tuller and Chilton, 1973; Myers, 1974; Sudmeyer et al., 1994; Luo and Goudriaan, 2000a) is the visual assessment of dew formation on specially treated wooden blocks, which was developed by Duvdevani (1947).

Blotting is also a widely used method to observe dew formation. It has been used in studies of dew on grass and crop leaves (Collins, 1961; Burrage, 1972; Jacobs et al., 1994; Sudmeyer et al., 1994; Richards and Oke, 2002). In this method sheets of pre-weighed blotting paper are pressed onto wet leaves. Then the papers are weighed to determine the mass change caused by the absorption of water. The deficiency of this procedure is that it collects all moisture on a surface, but cannot distinguish dewfall and distillation (and guttation). For the hydrological balance dewfall is of more significance than distillation, since dewfall is a true addition to the hydrologic balance, while distillation is merely a redistribution of water already in the system.

In addition, there is also a visual and/or tactile assessment of dew. It is the least complex but most subjective method. It is included in many studies (Myers, 1974; Sudmeyer et al., 1994; Luo and Goudriaan, 2000a), but has been shown to lack precision and consistency (Nagel, 1962; Haines, 1980).

A very different and promising, but also rather expensive method for measuring dew is the use of lysimeters (Meissner et al., 2007). High precision weighing lysimeters can monitor mass changes continuously and with high resolution so that the course of dew formation during a day (or more to the point a night) can be followed. Lysimeter design and size vary greatly. Accuracy depends on how well conditions within the lysimeter mimic those of the surrounding undisturbed surface, normally soil and vegetation (Slatyer and McIllroy, 1961; Sharma, 1976; Severini et al., 1984; Grimmond et al., 1992; Jacobs et al., 2000; Meissner et al., 2007). One big advantage of lysimeters over other methods of measuring dew is that the type of vegetation to be investigated can be planted on them. In the absence of rain, fog and irrigation an increase in mass can be attributed to dewfall, because the sides and base of a lysimeter are sealed. In dew measurements with a lysimeter distillation and guttation are not recorded, since both are a redistribution of moisture from the soil to plant leaves within the lysimeter monolith (Richards, 2004). For dew studies a lysimeter requires a resolution of > 0.05 mm of water. This precision is reached by some lysimeters (Sharma, 1976; Meissner et al., 2007).

When measured dew data are not available, simulation with a model is a possible alternative. Some relatively simple models to estimate dew amount or duration are based on a correlation with climate data. For example, Davis (1958) used nomogramme involving net radiation, dew point depression and wind speed to indicate favourable and unfavourable dew conditions. Collins and Taylor (1961) developed a method in Australia to predict the onset of dew on the surface of a thin large leaf using relative humidity, temperature and radiation. Bootsma et al. (1973) and Getz (1981) attempted to estimate leaf wetness duration in plant canopies from relative humidity data from weather stations, but this was not found to be a good predictor. Crowe et al. (1978) used a multiple regression approach based on relative humidity, wind speed and minimum temperature to forecast dew duration with an accuracy of $\pm$ 3 h. Lhomme et al. (1992) estimated dew duration from standard weather station measurements of air temperature, humidity, wind velocity, solar radiation and cloud cover. On average the differences between observed and estimated values were < 1 h.

There are also some neural network models and complex numerical simulations of atmosphere - canopy systems, in which the amount of dew is a minor output parameter (Thompson, 1981; Francl and Panigrahi, 1997). Multi-layer models were established, too, which include the soil and contain various layers throughout the canopy (Goudriaan, 1977; Braden, 1982). However, the most widespread method used to model dew on crops is a single-layer energy balance approach (Pedro and Gillespie, 1982a, b; Jacobs et al., 1990; Wittich, 1995; Wilson et al., 1999; Madeira et al., 2002). This approach has been shown to be accurate. For example, Pedro and Gillespie (1982a) simulated leaf wetness duration for corn, soybean and apple leaves to within 0.5 - 1.0 h. Their model has been validated for other crops (onion, corn and lettuce) and locations, and as an operational tool for scheduling agricultural sprays (Gillespie and Barr, 1984; Bass et al., 1991; Scherm and van Bruggen, 1993b; Scherm et al., 1995). Severini et al. (1984) compared the dewfall on lawn measured by a weighable lysimeter and computed with the energy balance equation. There was a good agreement between observed and computed dew onset, dew end and dew duration.

Compared to studies on dew duration, there are few studies on dew amounts. Neumann (1956) established an equation for the turbulent diffusion of water vapour to estimate in an approximate manner the amount of dewfall on short grass or on a bare surface using routine meteorological data. Sudmeyer et al. (1994) used the Penman-Monteith equation to predict the potential amount of dew. Jacobs et al. (1994) estimated the amount of dewfall and dewrise in a maize canopy during one night, and the drying time during in the early morning. They were estimated using the Bowen ratio energy balance and the soil diffusivity technique. Inclán and Forkel (1995) evaluated the performance of the big-leaf model SiB and the multi-layer model Cupid on the basis of a comparison of measured and modelled energy fluxes. The results indicate that the energy flux estimated with Cupid and the Penman-Monteith equation are in good agreement with measurements.

1.4 Objectives of this thesis

The previous sections illuminated that dew can be a significant parameter in the water balance and in soil-plant-atmosphere interactions. They also pointed out that the factors which influence dew formation warrant further investigation. So do lysimeters as a measurement tool for the quantification of dew amounts, and various micrometeorological equations to estimate dew amounts in the absence of measurements.

With this background the objectives of this thesis are:

- to identify the meteorological factors affecting dew formation and to quantify their effect (chapter 2),
- to examine the weighing precision of the type of lysimeter used later in this study to asses its suitability for dew studies (chapter 3),
- to quantify the amount of dew precipitated on various types of vegetation in northern Germany during a night and during the course of a year using lysimeters (chapter 4),

- to assess different methods to compute dewfall (chapter 5), and
- to compare dewfall measured with lysimeters and dewfall computed with the Bowen ratio energy balance and the Penman-Monteith equation (chapter 6).

Each of these objectives is treated in a self-contained chapter with its own introduction, a description of the materials and methods used, a presentation of the results, their discussion, conclusions drawn from the work, and a list of references.

All field work was carried out at the Falkenberg lysimeter station of the Department of Soil Physics of the Helmholtz Centre for Environmental Research - UFZ. The station is located in northern Germany, some 120 km north-west of Berlin. The site is described in detail in chapter 4.

In this thesis the focus is on dewfall, because it is the only dew component which yields a true water input into the system (cf. section 1.2). Hence, in the following the term dew usually refers to dewfall, unless the context suggests otherwise.

1.5 References

Acostav Baladón A. N., Gioda, A. (1991): L'importance des precipitations occultes sous les tropiques secs. Sécheresse 2, 132-135.

Alnaser, W. E., Barakat, A. (2000): Use of condensed water vapour from the atmosphere for irrigation in Barhein. Appl. Energ. 65, 3-18.

Ashbel, D. (1949): Frequency and distribution of dew in Palestine. Geogr. Rev. 39, 291-297.

Atzema, A. J., Jacobs, A. F. G., Wartena, L. (1990): Moisture distribution within a maize crop due to dew. Netherlands J. Agric. Sci. 38, 117-129.

Awanou, C. N., Hazoume, R. P. (1997): Study of natural condensation of atmospheric humidity. Renew. Energ. 10, 19-34.

Baier, W. (1966): Studies on dew formation under semi-arid conditions. Agric. Meteorol. 3, 103-112.

Barradas V. L, Glez-Medellin M. G. (1999): Dew and its effect on two heliophile understory species of a tropical dry deciduous forest in Mexico. Int. J. Biometeorol 43, 1-7.

Bass, B., Savdie, I., Gillespie, T. J. (1991): Simulation of leaf wetness duration for field corn. Agric. For. Meteorol. 57, 69-84.

Bootsma, A., Gillespie,T. J., Sutton, J. C. (1973): Germination of Phyllosticta maydis Conidia in an incubation chamber with control of high relative humidities. Phytopathology 63, 1157-1161.

Braden, H. (1982): Simulationsmodell für den Wasser-, Energie- und Stoffhaushalt in Pflanzenbeständen. Berichte des Instituts für Meteorologie und Klimatologie der Universität Hannover, 23, pp. 149. Institut für Meteorologie und Klimatologie der Universität Hannover, Hannover.

Burrage, S. W. (1972): Dew on wheat. Agric. Meteorol. 10, 3-12.

Buttler, D. R. (1980): Dew and thermal lag: measurements and an estimate of wetness duration on cocoa pods. Q. J. R. Meteorol. Soc. 106, 539-550.

Collins, B. G. (1961): A standing dew meter. Meteorol. Mag. 90, 114-117.

Collins, B. G., Taylor, R. J. (1961): Conditions governing the onset of dew on large leaves. Australian J. Appl. Sci. 12, 23-29.

Crowe, M. J., Coakley, S. M., Emge, R. G. (1978): Forecasting dew duration at Pendleton, Oregon, using simple weather observations. J. Appl. Meteorol. 17, 1482-1487.

Danin, A., Bar-Or, Y., Dor, I., Yisraeli, T. (1989): The role of cyanobacteria in stabilization of sand dunes in southern Israel. Ecol. Mediterr. 15, 55-64.

Davis, P. A. (1958): An investigation of a method for predicting dew duration. Arch. Meteorol. Geophys. Bioklimatol., Ser. A., 10, 66-93.

Degen, A., Leeper, A., Shachak, M. (1992): The effect of slope direction and population density on water influx in a desert snail, Trochoidea seetzenii. Funct. Ecol. 6,160-166.

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

Evenari, M., Shanan, L., Tadmor, N. (1982): The Negev: the challenge of a desert. Harvard University Press, Cambridge, Mass.

Francl, L. J., Panigrahi, S. (1997): Artificial neural network models of wheat leaf wetness. Agric. For. Meteorol. 88, 57-65.

Fritschen, L. J., Doraiswamy, P. (1973): Dew: an addition to the hydrologic balance of Douglas fir. Water Resour. Res. 9, 891-894.

Garratt, J. R., Segal, M. (1988): On the contribution of atmospheric moisture to dew formation. Bound.-Layer Meteorol. 45, 209-236.

Getz, R. R. (1981): The determination of wetting duration in plant canopies. In: American Meteorological Society, 15th Conference on Agriculture and Forest Meteorology and 5th Conference on Biometeorology, April 1 - 3, 1981 (pp. 155-157). Soc., Boston, Mass.

Gillespie, T. J., Barr, A. (1984): Adaptation of a dew estimation scheme to a new crop and site. Agric. For. Meteorol. 31, 289-295.

Goudriaan, J. (1977): Crop micrometeorology: a simulation study. Centre for Agricultural Publishing and Documentation, Wageningen.

Grimmond, C. S. B., Isard, S. A., Belding, M. J. (1992): Development and evaluation of continuously weighing mini-lysimeters. Agric. For. Meteorol. 62, 205-218.

Haines, D. A. (1980): The adjective classification for dew: what does it mean? National Weather Digest 5, 47-50.

Hirst, J. M., 1954. A method for recording the formation and persistence of water deposits on plant shoots. Q. J. R. Meteorol. Soc. 80, 227-231.

Hirst, J. M. (1957): A simplified surface-wetness recorder. Plant Pathol. 6, 57-61.

Horino, O., El-Nemr, F., Yousef, M., Niki, H. (1990): Factors of the rice blast outbreak in Nile Delta, Egypt. JARQ-JPN Agr. Res. Q. 23, 176-181.

Hughes, R. N., Brimblecombe, P. (1994): Dew and guttation: formation and environmental significance. Agric. For. Meteorol. 67, 173-190.

Inclán, M. G., Forkel, R. (1995): Comparison of energy fluxes calculated with the Penman-Monteith equation and the vegetation models SiB and Cupid. J. Hydrol. 166, 193-211.

Jacobs, A. F. G., van Pul, W. A. J., van Dijken, A. (1990): Similarity moisture dew profiles within a corn canopy. J. Appl. Meteorol. 29, 1300-1306.

Jacobs, A. F. G., van Pul, A., El-Kilani, R. M. M. (1994): Dew formation and the drying process within a maize canopy. Bound.-Layer Meteorol. 69, 367-378.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (1999): Dew deposition and drying in a desert system: a simple simulation model. J. Arid Environ. 42, 211-222.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (2000): Dew measurements along a longitudinal sand dune transect, Negev Desert, Israel. Int. J. Biometeorol. 43, 184-190.

Janssen, L. H. J. M., Römer, F. G. (1991): The frequency and duration of dew occurrence over a year. Tellus B 43, 408-419.

Kessler, O. W. (1939): Der Tauschreiber Kessler-Fuess. Bioklim. Beibl. Meteorol. Z. 6, 23-26.

Khare, P., Singh, S. P., Kumari, K. M., Kumar, A., Srivastava, S. S. (2000): Characterization of organic acids in dew collected on surrogate surfaces. J. Atmos. Chem. 37, 231-244.

Kidron, G. J., Yair, A., Danin, A. (2000): Dew variability within a small arid drainage basin in the Negev Highlands, Israel. Q. J. R. Meteorol. Soc. 126, 63-80.

Knoche, W. (1939): Notes on the dew. An. Soc. Cient. Argent. 127, 150-151.

Lange, O. L., Kidron, G., Büdel, B., Meyer, A., Kilian, E., Abeliovich, A. (1992): Taxonomic composition and photosynthetic characteristics of the “biological soil crusts” covering sand dunes in the western Negev Desert. Funct. Ecol. 6, 519-527.

Lange, O. L., Belnap, J., Reichenberger, H. (1998): Photosynthesis of the cyanobacterial soil-crust lichen Collema tenax from arid lands in southern Utah, USA: role of water content on light and temperature responses of CO_2 exchange. Funct. Ecol. 12, 195-202.

Larmuth, J., Harvey, H. J. (1978): Aspects of the occurrence of desert plants. J. Arid Environ. 1, 129-133.

Leick, E. (1932): Zur Methodik der relativen Taumessung. Bot. Cent.bl. Beih. 49, 160-189.

Lhomme, J .P., Jimenez, O. F. (1992): Estimating dew duration on banana and plantain leaves from standard meteorological observations. Agric. For. Meteorol. 62, 263-274.

Liu, W. J., Zeng, J. M., Wang, C. M., Li, H. M., Duang, W. P. (2001): On the relationship between forests and occult precipitation (dew and fog precipitation). J. Nat. Resour. 16, 571-575

Li, X.Y. (2002): Effect of gravel and sand mulches on dew deposition in the semiarid region of China. J. Hydrol. 260, 151-160.

Lloyd, M. (1961): The contribution of dew to the summer water budget of northern Idaho. Bull. Am. Meteorol. Soc. 42, 572-580.

Long, I. F. (1958): Some observations on dew. Meteorol. Mag. 87, 161-168.

Luo, W., Goudriaan, J. (1999): Effects of altering water temperature on leaf wetness in paddy rice crops. Agric. For. Meteorol. 97, 33-42.

Luo, W., Goudriaan, J. (2000): Dew formation on rice under varying durations of nocturnal radiative loss. Agric. For. Meteorol. 104, 303-313.

Luo, W., Goudriaan, J. (2000a): Measuring dew formation and its threshold value for net radiation loss on top leaves in a paddy rice crop by using the dewball: a new and simple instrument. Int. J. Biometeorol. 44, 167-171.

Madeira, A. C., Kim, K. S., Taylor, S. E., Gleason, M. L. (2002): A simple cloud-based energy balance model to estimate dew. Agric. For. Meteorol. 111, 55-63.

Marek, R., Straub, J. (2001): Analysis of the evaporation coefficient and the condensation coefficient of water. Int. J. Heat Mass Tran. 44, 39-53.

Meissner, R., Seeger, J., Rupp, H., Seyfarth, M., Borg, H. (2007): Measurement of dew, fog, and rime with a high-precision gravitation lysimeter. J. Plant Nutr. Soil Sci. 170, 335-344.

Mischke, W. (1960): Untersuchungen über den Einfluß des Bestandsklimas auf die Entwicklung der Blattfleckenkrankheit *Cercospora beticola* im Hinblick auf die Einrichtung eines Warndienstes. Bayer. Landwirtsch. Jahrb. 37, 197-227.

Moffett, M. W. (1985): An Indian ant's novel method for obtaining water. Natl. Geogr. Res. 1, 146-149.

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

Monteith J. L., Szeicz G. (1961): The radiation balance of bare soil and vegetation. Q. J. R. Meteorol. Soc. 87,159-170.

Monteith, J. L. (1963): Dew: facts and fallacies. In: Rutter, A. J., Whitehead, F. H. (eds), The water relations of plants (pp. 37-56). Wiley, New York.

Myers, T. M. (1974): Dew as a visual indicator of the urban heat island of Washington, D.C. M.Sc. thesis, University of Maryland, Maryland.

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

Neumann, J. (1956): Estimating the amount of dewfall, Arch. Meteorol. Geophys. Bioklimatol., Ser. A, 9, 197-203.

Newton, O. H., Riley, J. A. (1964): Dew in the Mississippi delta in the fall. Mon. Weather Rev. 92, 369-373.

Nilsson, T. (1996): Initial experiments on dew collection in Sweden and Tanzania. Sol. Energ. Mat. Sol. Cells 40, 23-32.

Okochi, H., Kajimoto, T., Arai, Y., Igawa, M. (1996): Effect of acid deposition on urban dew chemistry in Yokohama, Japan. Bull. Chem. Soc. Jpn. 69, 3355-3365.

Pedro, M. J., Gillespie, T. J. (1982a): Estimating dew duration. I. Utilizing micrometeorological data. Agric. Meteorol. 25, 283-296.

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

Richards, K., Oke, T. R. (2002): Validation and results of a scale model of dew deposition in urban environments. Int. J. Climatol. 22, 1915-1933.

Richards, K. (2004): Observation and simulation of dew in rural and urban environments. Prog. Phys. Geogr. 28, 76-94.

Royle, D. J., Butler, D. R. (1986): Epidemiological significance of liquid water in crop canopies and its role in disease forecasting. In: Ayres, P. G., Boddy, L. (eds.), Water, fungi and plants (pp. 139-156). Cambridge University Press, Cambridge.

Scherm, H., van Bruggen, A. H. C. (1993a): Sensitivity of simulated dew duration to meteorological variations in different climatic regions of California. Agric. For. Meteorol. 66, 229-245.

Scherm, H., van Bruggen, A. H. C. (1993b): Response surface models for germination and infection of *Bremia lactucae*, the fungus causing downy mildew of lettuce. Ecol. Model. 65, 281-296.

Scherm, H., Koike, S. T., Laemmlen, F. F., van Bruggen, A. H. C. (1995): Field evaluation of fungicide spray advisories against lettuce downy mildew (*Bremia lactucae*) based on measured or forecast morning leaf wetness. Plant Dis. 79, 511-516.

Severini, M., Moriconi, M. L., Tonna, G., Olivieri, B. (1984): Dewfall and evapotranspiration determination during day- and nighttime on an irrigated lawn. J. Clim. Appl. Meteorol. 23, 1241-1246.

Sharma, M. L. (1976): Contribution of dew in the hydrologic balance of a semi-arid grassland.

Agric. Meteorol. 17, 321-331.

Simon, R. D., Abeliovich, A., Belkin, S. (1994): A novel terrestrial halophilic environment: the phylloplane of Atriplex halimus, a salt-excreting plant. FEMS Microbiol. Ecol. 14, 99-110.

Slatyer, R. O., McIllroy, I. C. (1961): Practical Microclimatology. Unesco, Paris.

Steubing, L. (1952): Der Tau und seine Beeinflussung durch Windschutzanlagen. Biolog. Zentralbl. 71, 282- 313.

Steubing, L., Casperson, G. (1959): Pflanzentemperatur und Taubeschlag. Angew. Meteorol. 3, 219-224.

Stewart, J. B. (1977): Evaporation from the wet canopy of a pine forest. Water Resour. Res. 13, 915-921.

Stone, E. C. (1957a): Dew as an ecological factor. I. A review of the literature. Ecology 38, 407-413.

Stone, E. C. (1957b): Dew as an ecological factor. II. The effect of artificial dew on the survival of Pinus ponderosa and associated species. Ecology 38, 414-422.

Subramaniam, A. R., Kesava Rao, A. V. R. (1983): Dew fall in sand dune areas of India. Int. J. Biometeorol. 27, 271-280.

Sudmeyer, R. A., Nulsen, R. A., Scott, W. D. (1994): Measured dewfall and potential condensation on grazed pasture in the Collie River basin, southwestern Australia. J. Hydrol. 154, 255-269.

Takenaka, N., Soda, H., Sato, K., Terada, H., Suzue, T., Bandow, H., Maeda, Y. (2003): Difference in amounts and composition of dew from different types of dew collectors. Water Air Soil Pollut. 147, 51-60.

Thompson, N. (1981): The duration of leaf wetness. Meteorol. Mag. 110, 1-12.

Tuller, S. E., Chilton, R. (1973): The role of dew in the seasonal moisture balance of a summer-dry climate. Agric. Meteorol. 11, 135-142.

Uehara, Y., Imoto, M., Sakai, Y. (1988): Study on the forecasting of the rice blast development using the weather data from AMeDAS. Bull. Hiroshima Prefecture Agric. Exp. Stn. 51, 1-15.

UNESCO (1958): Climatology and microclimatology: proceedings of the Canberra symposium. UNESCO, Paris.

van Eimern, J. (1964): Untersuchungen über das Klima in Pflanzenbeständen. Ber. Deut. Wetterd. 13, 76 S.

Wales-Smith, B. G. (1983): An inexpensive device for measuring net dewfall. Weather 38, 54-56.

Wallin, J. R. (1963): Dew, its significance and measurement in phytopathology. Phytopathology 53, 1210-1216.

Wallin, J. R. (1967): Agrometeorological aspects of dew. Agric. Meteorol. 4, 85-102.

Wilson, T. B., Bland, W. L., Norman, J. M. (1999): Measurement and simulation of dew accumulation and drying in a potato canopy. Agric. For. Meteorol. 93, 111-119.

Wisniewski, J. (1982): The potential acidity associated with dews, frosts, and fogs. Water Air Soil Pollut. 17, 361-377.

Wittich, K-P. (1995): Some remarks on dew duration on top of an orchard. Agric. For. Meteorol. 72, 167-180.

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

2. Analysis of the effect of meteorological factors on dew formation

2.1 Abstract

Dew forms when the temperature of a surface cools below the dew point temperature of the surrounding air. As a result vapour contained in this air condenses on the surface. The cooling of a surface is caused by a radiation loss. How much it cools down depends on the weather conditions.

The energy balance equation was analysed to identify the meteorological factors which determine the degree of cooling, and to assess their effect on dew formation. These factors were found to be air temperature (T_a), cloud cover (N), wind speed (u), soil heat flux (G), and relative humidity (h_r). The temperature of the surface (T_s) dew forms on is also important. However, it is not a meteorological factor, but determined by the aforementioned variables. Also, net radiation was not considered explicitly, but indirectly by looking at the effect of temperature on the radiation balance.

All other conditions equal, the analysis revealed that dewfall increases linearly as N or G decrease, or h_r increases. The effect of T_a and u on dewfall is non-linear. It first rises with T_a or u and then falls again. All five meteorological factors identified here have roughly the same impact. Each can lead to variations in dewfall between 0 and ~ 25 W m^{-2}, the precise magnitude of the impact depends on the value of the other factors. Dewfall is always highest if N = 0, G = 0 and h_r = 1. At which air temperature dewfall is highest depends on wind speed and vice versa.

2.2 Introduction

Dew formation is the result of nocturnal radiation loss and vapour condensation. During the night a natural surface (e.g. a soil or a plant) usually cools down, because more radiation is emitted from the surface than it receives from its surroundings. When the surface temperature has decreased to the dew point temperature of the air near the surface, water vapour contained in this air will condense on the surface.

Dew is a form of precipitation, but the amount is usually small, rarely exceeding 0.3 - 0.5 mm per night on vegetation (Garratt and Segal, 1988). Despite the small amounts, dew is an important moisture input in arid and semiarid ecosystems where rainfall is usually low, since it can supplement the rainfall and even surpass it at times (e.g. Malek et al., 1999; Kidron et al., 1999, 2000; Ninari and Berliner, 2002; Ye et al., 2007). Even in humid regions dew can contribute a notable percentage to the precipitation over a short period such as a week or a month, although it typically only accounts for a small fraction of the annual precipitation (Tuller and Chilton, 1973). In addition, dew which occurs on vegetation is a weather parameter of significance to the agricultural community, since for some crops dew can affect the choice of optimal harvest time and the efficacy of the application of pesticides and herbicides (Wallin, 1967). Moreover, dew is of major importance to the development of various foliar bacterial and fungal pathogens (Wallin, 1963).

Neumann (1956), Monteith (1957, 1963) and Long (1958) pointed out that dewfall represents a flux of latent heat towards the surface, which makes it the opposite of evaporation, and discussed the weather conditions favourable for dew formation. It was recognised that sufficient moisture in the air and intensive radiative cooling of the surface are two basic requirements for dew formation. Hence, temperature and humidity can be inferred as two important meteorological factors. In addition, Baier (1966) indicated that clear skies and light winds are favourable for dew formation. Clear skies allow more outgoing radiation and thus accelerate radiative cooling, while light winds can bring additional moisture to the surface.

Some studies have been conducted to analyse the relationship between dew formation and meteorological factors by using regression equations (e.g. Smith, 1958; Riley and Giles, 1965; Smith and Carpenter, 1966; Crowe et al., 1978; Ye et al., 2007). Air temperature, cloud cover, wind speed and relative humidity are usually taken as the key factors in dew formation. However, because of different conditions in various studies or at the sites where the studies were carried out, the effect of a given meteorological factor on dew formation differs significantly. For example, Shaw (1973) found a good correlation between dew formation and periods of relative humidity above 85%. In contrast, Smith (1958) gave a relative humidity of 90% as the threshold for dew occurrence. Also, there are controversies about the effect of wind on dew formation. Beysens et al. (2005) found that light winds or wind speeds near 0 m/s prompt dew formation. However, Monteith (1957) indicated that the turbulent transfer of water vapour from the air to the surface is negligible when the wind speed drops below 0.5 m/s, and that the turbulent flux of vapour to a surface increases with wind speed and reaches a maximum when the wind speed is 2 - 3 m/s. Muselli et al. (2002) found that a wind speed of 4.5 m/s is a threshold for dew formation; above it there was no dew. This shows that correlation analyses between dew formation and climate data are highly empirical and help little in understanding the underlying physical phenomena (Scherm and van Bruggen, 1993a).

The latent heat flux represents dew formation. It is a component of the surface energy balance, together with net radiation, sensible heat flux and soil heat flux. Hence, the best way to assess which meteorological factors influence dew formation and how is to apply the energy balance equation for a surface. This equation was used to study dew formation before. For example, Monteith (1957) analysed the change of components of the surface energy balance as dew condensed. Furthermore, some studies use the energy balance approach to predict dew occurrence. (e.g. Pedro and Gillespie, 1982a, b; Bass et al., 1991; Lhomme and Jimenez, 1992; Scherm and van Bruggen, 1993a)

The effect of some meteorological factors on dew formation was already studied by Monteith (1963) and Garratt and Segal (1988) using the Penman-Monteith equation. Surface temperature and surface vapour pressure are eliminated from this equation. In addition, for the most part in their studies the air was assumed to be saturated (i.e. the relative humidity to be 1), which means they looked at potential dew amounts. In our study we use the energy balance equation, which includes surface temperature and surface vapour pressure, and we allow the relative humidity to take different values. A computer program which solves the energy balance equation was written to analyse the effect of individual meteorological factors on dew formation.

2.3 Materials and methods

2.3.1 Energy balance equation

The energy balance equation for a soil or plant surface can be written as:

$$0 = R_n + H + G + \lambda E + M \tag{2.1}$$

where R_n = net radiation (W m^{-2}), H = sensible heat flux between atmosphere and surface (W m^{-2}), G = soil heat flux (W m^{-2}), λE = latent heat flux between atmosphere and surface (W m^{-2}), and M = energy released or required by metabolic processes (W m^{-2}). M is often small compared to the other components of the energy balance (Gates, 1965). Hence, this term is not taken into account here. Omitting it in Eq. 2.1 gives:

$$0 = R_n + H + G + \lambda E \tag{2.2}$$

The individual components of Eq. 2.2 are presented in more detail below. The terms are written such that fluxes directed towards the surface are positive and fluxes away from the surface negative.

Note that at the soil surface G can be split into an upward sensible (H_u) and latent heat flux (λE_u), as pointed out by Monteith (1963). λE_u represents dewrise. Using this split in Eq. 2.2 yields $0 = R_n + H + H_u + \lambda E_u + \lambda E$ (Garratt and Seagal, 1988). Rearranging this expression accordingly, dewrise (λE_u) and dewfall (λE) can be calculated separately, or as a sum ($\lambda E_u + \lambda E$). Since this thesis is only concerned with dewfall, G is not split.

Net radiation

Net radiation is the difference between incoming and outgoing radiation. Between sunset and sunrise, when dew mostly occurs, there is only long wave radiation. Hence, for R_n one can write:

$$R_n = L_{in} - L_{out} \tag{2.3}$$

where L_{in} = incoming long wave radiation (W m^{-2}), and L_{out} = outgoing long wave radiation (W m^{-2}). Long wave radiation can be computed with the Stefan-Boltzmann-equation. For L_{in} it reads:

$$L_{in} = \varepsilon_{sky} \cdot \sigma \cdot T_a^4 \tag{2.4}$$

where ε_{sky} = emissivity of the sky for long wave radiation of the sky (dimensionless fraction), σ = Stefan-Boltzmann constant ($5.67 \cdot 10^{-8}$ W m^{-2} K^{-4}), and T_a = air temperature at screen height (K). The equivalent expression for L_{out} is:

$$L_{out} = \varepsilon_s \cdot \sigma \cdot T_s^4 \tag{2.5}$$

where ε_s = emissivity of the surface for long wave radiation (dimensionless fraction), and T_s = temperature of the radiating surface (K). We set ε_s = 0.95, which is a typical value for soils and plants (Campbell and Norman, 1998).

For ε_{sky} we use an empirical expression proposed by Paltridge and Platt (1976):

$$\varepsilon_{sky} = f_1 \cdot T_a^2 + \frac{f_2 \cdot N}{\sigma \cdot T_a^4} \tag{2.6}$$

where $f_1 = 9.35 \cdot 10^{-6}$ K^{-2}, f_2 = 60 W m^{-2}, and N = cloud cover (dimensionless fraction). Note that ε_{sky} is not a direct function of T_a, as one may conclude from Eq. 2.6, but of the vapour pressure of the atmosphere (Campbell and Norman, 1998). However, there is a correlation between vapour pressure and T_a so that there is a correlation between T_a and ε_{sky}, too.

Sensible heat flux

Sensible heat flux is the transfer of heat away from or to a surface by turbulent transport and diffusion. Its intensity depends on a heat transfer coefficient (conductance) and the temperature gradient between the surface and the surrounding air and is given by:

$$H = c_p \cdot g_H \cdot (T_a - T_s) \tag{2.7}$$

where c_p = heat capacity of air (J mol^{-1} K^{-1}), and g_H = heat conductance of the air between the surface and screen height (mol m^{-2} s^{-1}). To compute heat conductance we employ the equation used by Davie (2003):

$$g_H = k^2 \cdot u \cdot \rho_a \cdot \left(\ln \frac{z-d}{z_0} \right)^{-2} \tag{2.8}$$

where k = von Karman constant (0.41, dimensionless), u = wind speed at height z (m s^{-1}), ρ_a = molar density of the air (mol m^{-3}), z = height at which wind speed is measured (m), d =

zero plane displacement (m), and z_0 = roughness coefficient (m). We evaluate d and z_0 with the equations from Campbell and Norman (1998) as d = 0.65·h, and z_0 = 0.1·h, where h = height of the crop (m). Eq. 2.8 shows that the heat conductance depends on wind speed and surface parameters.

Soil heat flux

Soil heat flux is very variable, since it depends on soil texture and structure as well as on soil moisture content (Nakshabandi and Kohnke, 1965; Rosenberg, 1974; Monteith and Unsworth, 1990; Evett, 2001). In different soils or in the same soil at different water contents the soil heat flux is different. Hence, in this study soil heat flux is used as an input variable to analyse how different soil heat flux values influence dew formation.

Latent heat flux

In the process of dew formation water changes form the gaseous to the liquid phase (condensation). During this phase change heat is released to the surface. This so-called latent heat flux depends on a vapour transfer coefficient (conductance) and on the vapour pressure gradient between the ambient air and the surface:

$$\lambda E = \lambda \cdot g_v \cdot \frac{(e_a - e_s)}{P} \tag{2.9}$$

where λ = latent heat of vapourisation (J mol^{-1}), g_v = vapour conductance of the air between the surface and screen height (mol m^{-2} s^{-1}), e_a = vapour pressure of the air (kPa), e_s = vapour pressure of the air at the surface (kPa), which equals the saturation vapour pressure at the temperature of the surface when dew condenses, and P = air pressure (kPa).

Following Brutsaert (1982) the conductances for heat and vapour are set equal, i.e. $g_H = g_v$. The saturation vapour pressure at surface temperature is calculated with the Tetens formula (Campbell and Norman, 1998):

$$e_s = a \cdot \exp\left(\frac{b \cdot T_s}{T_s + c}\right) \tag{2.10}$$

where a = 0.611 kPa, b = 17.502 (dimensionless), and c = 240.97 K. The ambient vapour pressure is also calculated with the help of the Tetens formula as:

$$e_a = h_r \cdot a \cdot \exp\left(\frac{b \cdot T_a}{T_a + c}\right) \tag{2.11}$$

where h_r = relative humidity of the air.

Synopsis

Substituting Eq. 2.3 to 2.11 into Eq. 2.2 the energy balance equation can be expanded and now written in the form:

$$\begin{aligned} 0 = &\left\langle \left(f_1 \cdot \sigma \cdot T_a^6 + f_2 \cdot N\right) - (\varepsilon_s \cdot \sigma \cdot T_s^4)\right\rangle + \left\langle c_p \cdot k^2 \cdot u \cdot \rho_a \cdot \left(\ln \frac{z-d}{z_0}\right)^{-2} \cdot (T_a - T_s)\right\rangle + \langle G \rangle \\ &+ \left\langle \lambda \cdot k^2 \cdot u \cdot \rho_a \cdot \left(\ln \frac{z-d}{z_0}\right)^{-2} \cdot \frac{h_r \cdot a \cdot \exp\left(\frac{b \cdot T_a}{T_a + c}\right) - a \cdot \exp\left(\frac{b \cdot T_s}{T_s + c}\right)}{P}\right\rangle \end{aligned} \tag{2.12}$$

Leaving aside the various constants (f_1, σ, f_2, c_p, k, ρ_a λ, a, b, c, P), plant or soil dependent parameters (ε_s, d, z_0), and the height of wind measurements (z), the expanded energy balance equation reveals that the meteorological factors affecting dew formation are (from left to right):

- air temperature T_a,
- cloud cover N,
- wind speed u,
- soil heat flux G,
- relative humidity hr.

The surface temperature T_s is not a meteorological factor, but determined by them. It therefore represents the dependent variable in Eq. 2.12, while T_a, N, u, G and h_r are the independent variables.

Note that the last term in Eq. 2.12 computes the latent heat flux, i.e. dewfall if it is positive, or evaporation if it is negative. So, if wind speed, relative humidity, air temperature and surface temperature are known, one can use this term to compute dew formation directly. In general, the surface temperature is not measured. However, if the required meteorological parameters are known, one can use the complete Eq. 2.12 to compute T_s. Once determined, it can be inserted into the last term to calculate dew formation. Alternatively, one can use T_s and the first three terms of Eq. 2.12 to do it.

Figure 2.1 illustrates the dependency of the latent heat flux (λE) on meteorological factors. The magnitude of λE is governed by the vapour conductance and the vapour pressure gradient. The direction of λE is determined by the vapour pressure gradient. If it is positive, i.e. directed towards the surface, latent heat flows towards surface, which results in condensation at the surface, and dew occurs. In contrast, if the vapour pressure gradient is negative, i.e. directed away from the surface, vapour evaporates from surface, which results in a latent heat flux into atmosphere.

Eq. 2.8 shows that the vapour conductance is associated with the physical nature of the surface, and wind speed. Aerodynamic roughness (z_0) and zero-plane displacement (d) are surface parameters. Only meteorological factors are considered in this chapter, the surface conditions are not varied. Under the same surface conditions vapour conductance is only influenced by wind speed.

The vapour pressure gradient is determined by the difference between the vapour pressure at the surface, which is assumed to be saturated, and the vapour pressure of the air at a certain height (here 2 m). The vapour pressure of the air is a function of air temperature and relative humidity, the saturated vapour pressure of the surface is only a function of surface temperature. Eq. 2.12 indicates that the surface temperature depends on air temperature, cloud cover, wind speed, soil heat flux and relative humidity.

In the next chapter computations will be made with Eq. 2.12 to assess in detail, how these meteorological factors influence dew formation.

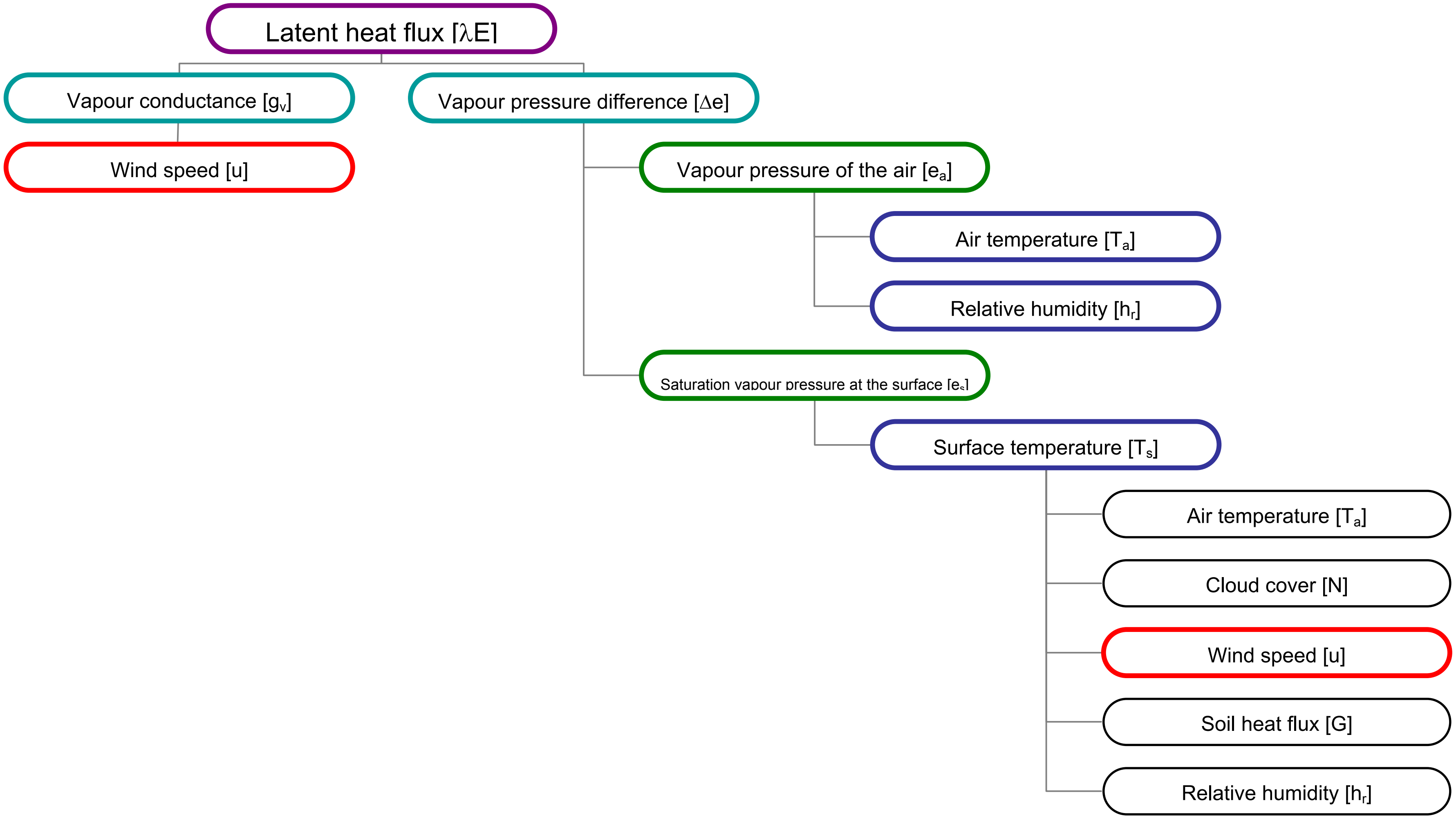


Figure 2.1: Dependence of latent heat flux on meteorological factors.

2.3.2 Computation procedure

For a given set of values for the meteorological parameters T_a, N, u, G and h_r a certain value of T_s will ensue to balance Eq. 2.12. If one of the meteorological parameters changes its value, T_s will change, too. The equation cannot be solved explicitly for T_s, so the binary search method was employed to find its value to an accuracy of 0.001°C.

In all computations the meteorological parameters are treated as known inputs. A base value is selected for each of them and the equation then solved for T_s. Next, the value of one parameter is varied over a specified range, while all others are kept constant. The equation is again solved for T_s. The T_s values and the meteorological data from which they result are now inserted into the last term of Eq. 2.12, which is an expansion of Eq. 2.9. This yields the latent heat flux, which represents dew formation if it is positive, and evaporation if it is negative. In this manner one gets the dew formation as a function of a chosen meteorological parameter for a constant set of values for the other parameters. The set of values chosen for the computations here are given in Table 2.1.

Table 2.1: Base values chosen for the computations and their subsequent range of variation.

Parameter	Base value	Range
T_a (°C)	10	2 - 30
N (fraction)	0.5	0 - 1
u (m s^{-1})	1	0.1 - 8
G (W m^{-2})	20	0 - 40
h_r (fraction)	0.85	0.6 - 1

Each of the five parameters in Table 2.1 is taken in turn as the “independent” variable, i.e. as the one which is varied over the whole range indicated in the table. Calculations with the independent variable are carried out for different values of the other four parameters. However, only one parameter is altered at a time, while the remaining three are assigned the base value shown in Table 2.1.

The results of the computations are presented in five sections, one for each of the parameters as the independent value. Unless stated otherwise the base value is used for each parameter. In a slight deviation to the order given in Table 2.1 wind speed u is looked at last.

2.4 Results

2.4.1 Air temperature

Figure 2.2 shows that the effect of air temperature on latent heat flux (λE) is non-linear for all combinations of values used in the calculations and that how air temperature affects latent heat flux depends on other meteorological factors, too. At certain values of N, G, hr and u there is little variation in latent heat flux with air temperature, while at others it is considerable. Also, at a given air temperature latent heat flux drops as cloud cover (Fig. 2.2a) or soil heat flux (Fig. 2.2b) increase, or relative humidity (Fig. 2.2c) decreases. (The effect of u will be discussed later.) This drop or increase is more pronounced the higher T_a is. For example, N, G and h_r have a bigger effect on latent heat flux at $T_a = 25°C$ than at $T_a = 5°C$.

For $N \leq 0.5$, all G values used here, and for $h_r \geq 0.7$ latent heat flux increases with air temperature up to a certain point, and then drops again (Fig. 2.2a - c). The temperature at which λE is highest falls with increasing cloud cover, increasing soil heat flux, and decreasing relative humidity. For $N > 0.5$ and for $h_r < 0.7$ latent heat flux decreases continuously with air temperature.

There is dewfall over the whole temperature range looked at here, if there are no clouds (N = 0). At 25% cloud cover (N = 0.25) dewfall gives way to evaporation at air temperatures > 29°C. This switch occurs at ever lower temperatures as the cloud cover increases. At full cloud cover (N = 1) there is no dewfall and only evaporation over the depicted temperature range. Similarly, there is dewfall over the whole temperature range considered, if $h_r \geq 0.9$. At $h_r = 0.8$ dewfall gives way to evaporation at air temperatures > 22°C. This threshold temperature drops to $T_a = 10°C$ for $h_r = 0.7$. For $h_r \leq 0.6$ there is no dewfall and only evaporation. For all values of G considered there is dewfall over the entire temperature range looked at here, except for G = 40 W m^{-2} and $T_a > 27°C$, when evaporation takes over.

The effect of wind speed latent on heat flux (Fig. 2.2d) is more complicated than that of the other three parameters just discussed. For a given air temperature latent heat flux increases with wind speed up to a certain value of u (2 m s^{-1} for $T_a \leq 20°C$ and 1 m s^{-1} for $T_a > 20°C$) and then decreases again.

For all wind speeds except 8 m s^{-1} dewfall increases with T_a up to a certain temperature and then drops again (Fig. 2.2d). The air temperature at which dewfall is highest decreases with wind speed. For u = 8 m s^{-1} latent heat flux decreases continuously with air temperature and changes from dewfall to evaporation at $T_a = 21°C$. For u = 4 m s^{-1} evaporation occurs at $T_a > 29°C$. At all other wind speeds looked at here there is dewfall over the entire temperature range considered.

2.4.2 Cloud cover

Figure 2.3 illustrates that latent heat flux declines linearly with cloud cover for all parameter combinations employed, and that the relationship between λE and N is influenced by other meteorological factors.

Figure 2.3a shows the effect of N on latent heat flux at various values of T_a, which was already displayed in a different way in Figure 2.2a. Both figures lead to the same conclusions, which makes a detailed discussion of Figure 2.3a and indeed the figure itself superfluous. However, it is included here, because the increasingly steeper slopes at rising values of T_a nicely demonstrate that cloud cover has a bigger influence on latent heat flux the higher T_a is.

In contrast, the rate of decline in λE with N is the same for all values of G or h_r, although the slope is different for the two parameters (Fig. 2.3b and c). The lower G or the higher h_r, the higher N at which dewfall is replaced by evaporation. At $h_r \leq 0.7$ there is only evaporation.

At a given cloud cover latent heat flux declines as soil heat flux increases (Fig. 2.3b) or relative humidity decreases (Fig. 2.3c). The magnitude of this decline is the same at each value of N, which implies that the change in latent heat flux with G or h_r is linear, but different for the two parameters.

Figure 2.3d illustrates that the effect of cloud cover on latent heat flux becomes more pronounced (i.e the slope steeper) as u goes up. Also, as u increases the cloud cover at which dewfall gives was to evaporation becomes less. Furthermore, at a given cloud cover latent heat flux increases up to a certain value of u and then declines again. This value depends on N. For example, it is 2 m s^{-1} at N = 0, but 1 m s^{-1} at N = 0.5, which means it declines as cloud cover increases.

2.4.3 Soil heat flux

A comparison of Figure 2.3 and 2.4 shows that the relationship between latent and soil heat flux and the interplay with the other parameters is similar to the relationship between latent heat flux and cloud cover. Hence, for the most part the same statements apply as for N in the previous chapter:

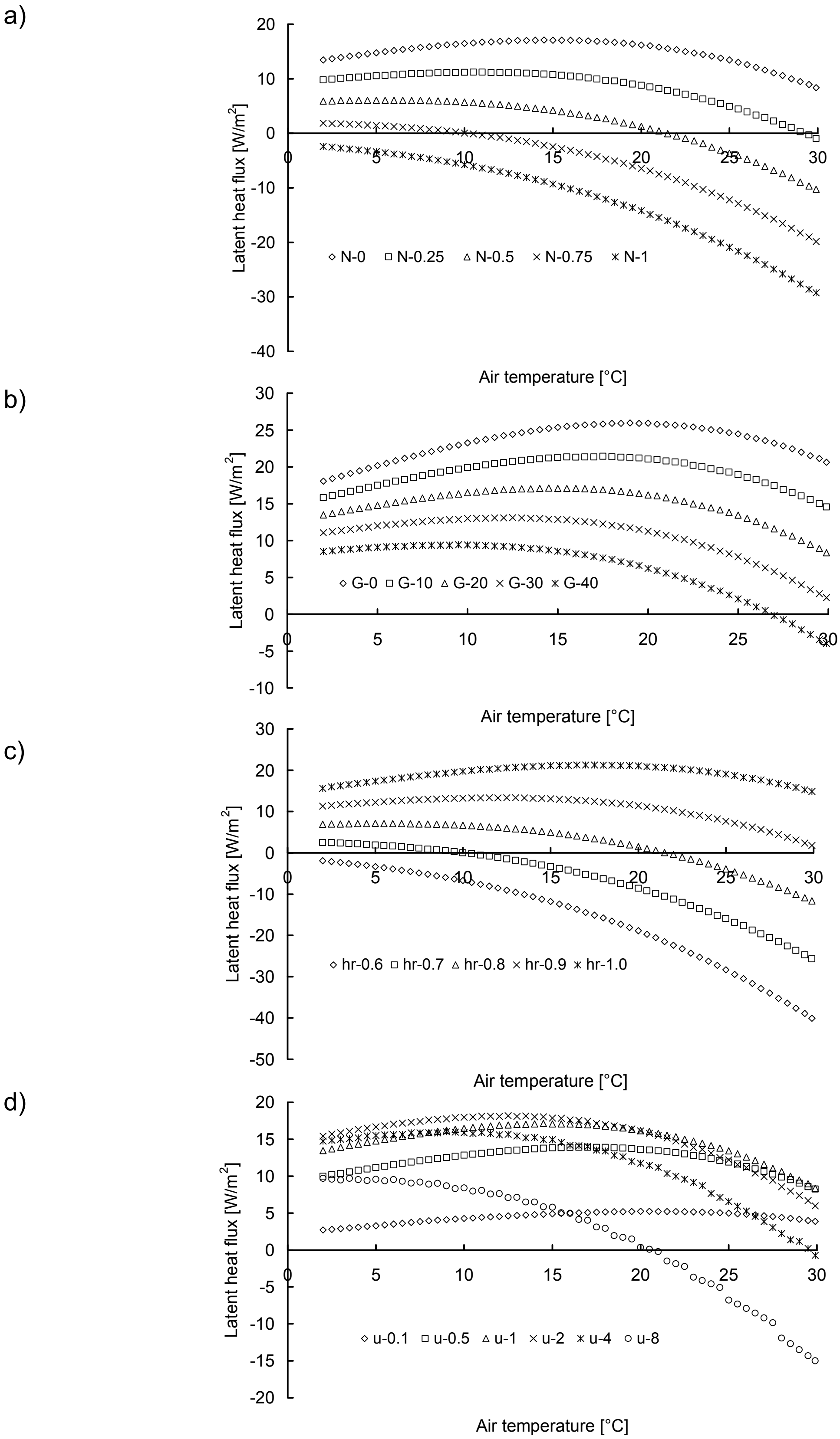


Figure 2.2: Effect of air temperature on latent heat flux at different values of a) cloud cover (N), b) soil heat flux (G), c) relative humidity (h_r), and d) wind speed (u).

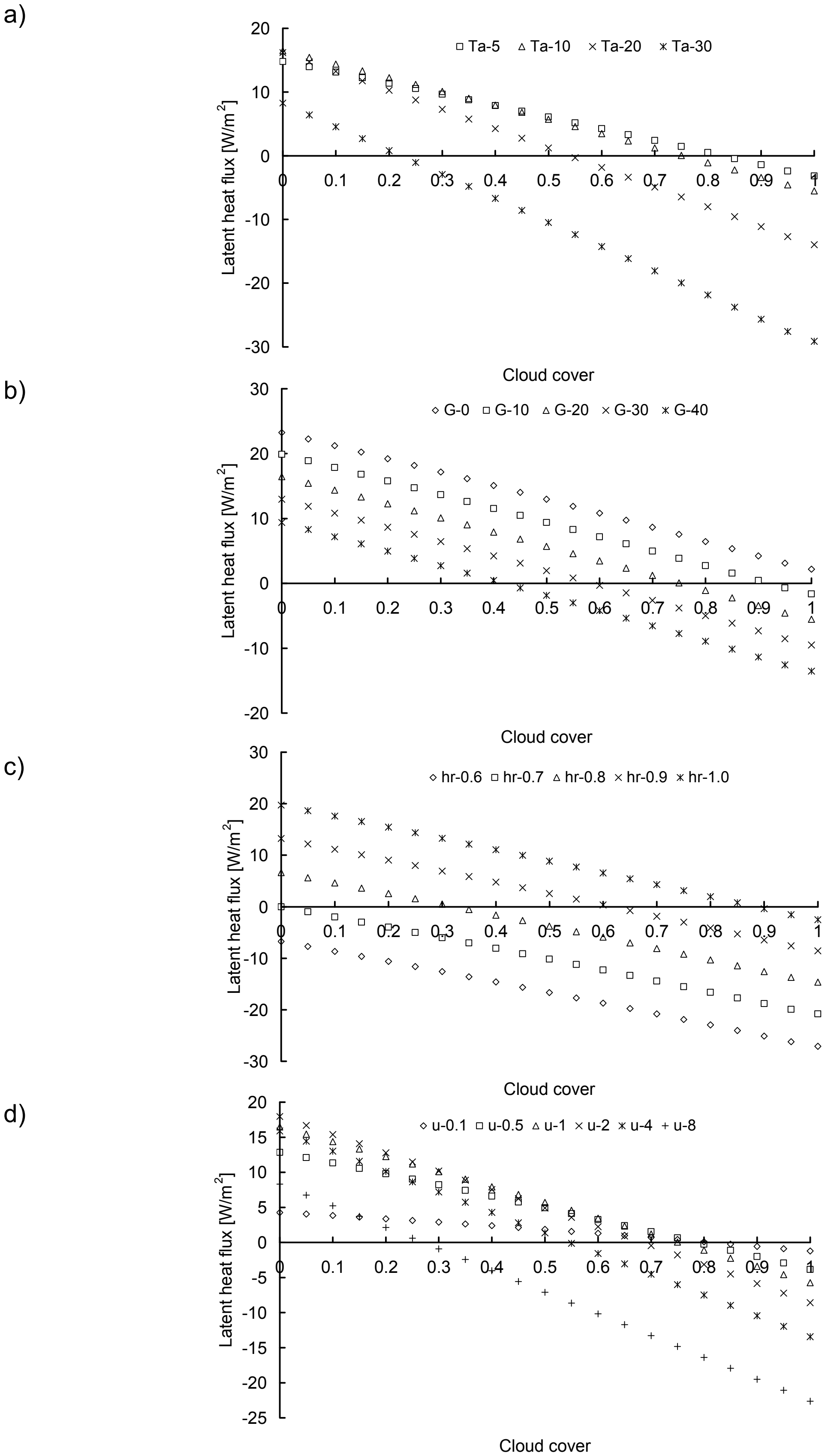


Figure 2.3: Effect of cloud cover on latent heat flux at different values of a) air temperature (T_a), b) soil heat flux (G), c) relative humidity (h_r), and d) wind speed (u).

Latent heat flux declines linearly with G and the relationship between λE and G is influenced by other meteorological factors. G has a bigger influence on λE as T_a increases, which was already discussed in connection with Figure 2.2b. The rate of decline in λE with G is the same for all values of N (which was already visible in Figure 2.3b) or h_r, although the rate is different for N and h_r (Fig. 2.4b and c). The lower N or the higher h_r, the higher G at which dewfall is replaced by evaporation. At $h_r \leq 0.6$ there is only evaporation. At a given G dewfall declines as N increases (Fig. 2.4b) or h_r decreases (Fig. 2.4c). The magnitude of this decline is the same at each value of G, but different for N and h_r.

The effect of G on λE is bigger (steeper slope) as wind speed goes up (Fig. 2.4d). Evaporation only occurs at G > 37 W m^{-2} and u = 8 m s^{-1}. For the range of soil heat flux values used here the latent heat flux at a given G increases with wind speed up to u = 2 m s^{-1} and then declines again.

2.4.4 Relative humidity

The previous illustrations have shown that λE goes down as N or G increase, but that it goes up as h_r increases. Hence, Figure 2.5 looks like a mirror image of Figure 2.3 or 2.4 reflected across the x-axis. It shows that latent heat flux increases linearly with relative humidity for all parameter combinations considered and that the relationship between λE and h_r is influenced by T_a, N, G and u. If hr $\leq$ 0.6 there is no dewfall for any parameter combination.

Like Figure 2.2a, Figure 2.5a illustrates that the higher T_a, the steeper the increase in λE with h_r. Also, the lower the humidity, the bigger the effect of air temperature on λE.

As already alluded to in previous figures (Fig. 2.3c and 2.4c), the rate of increase in λE with h_r is the same for all values of N (Fig. 2.5b) or G (Fig. 2.5c), although the rate is different for N and G. The lower N or G, the lower h_r at which dewfall is replaced by evaporation. Lastly, at a given h_r latent heat flux rises as N (Fig. 2.5b) or G decreases (Fig. 2.5c). The magnitude of this rise is the same for each incremental change in N or G, respectively, i.e. the rise is linear.

Similar to the influence of T_a, the rise in λE with h_r becomes steeper as wind speed goes up (Fig. 2.5d). In addition, the higher u the higher h_r at which dewfall gives way to evaporation. Lastly, latent heat flux increases with wind speed up to a certain value of u and then declines again. This value depends on h_r.

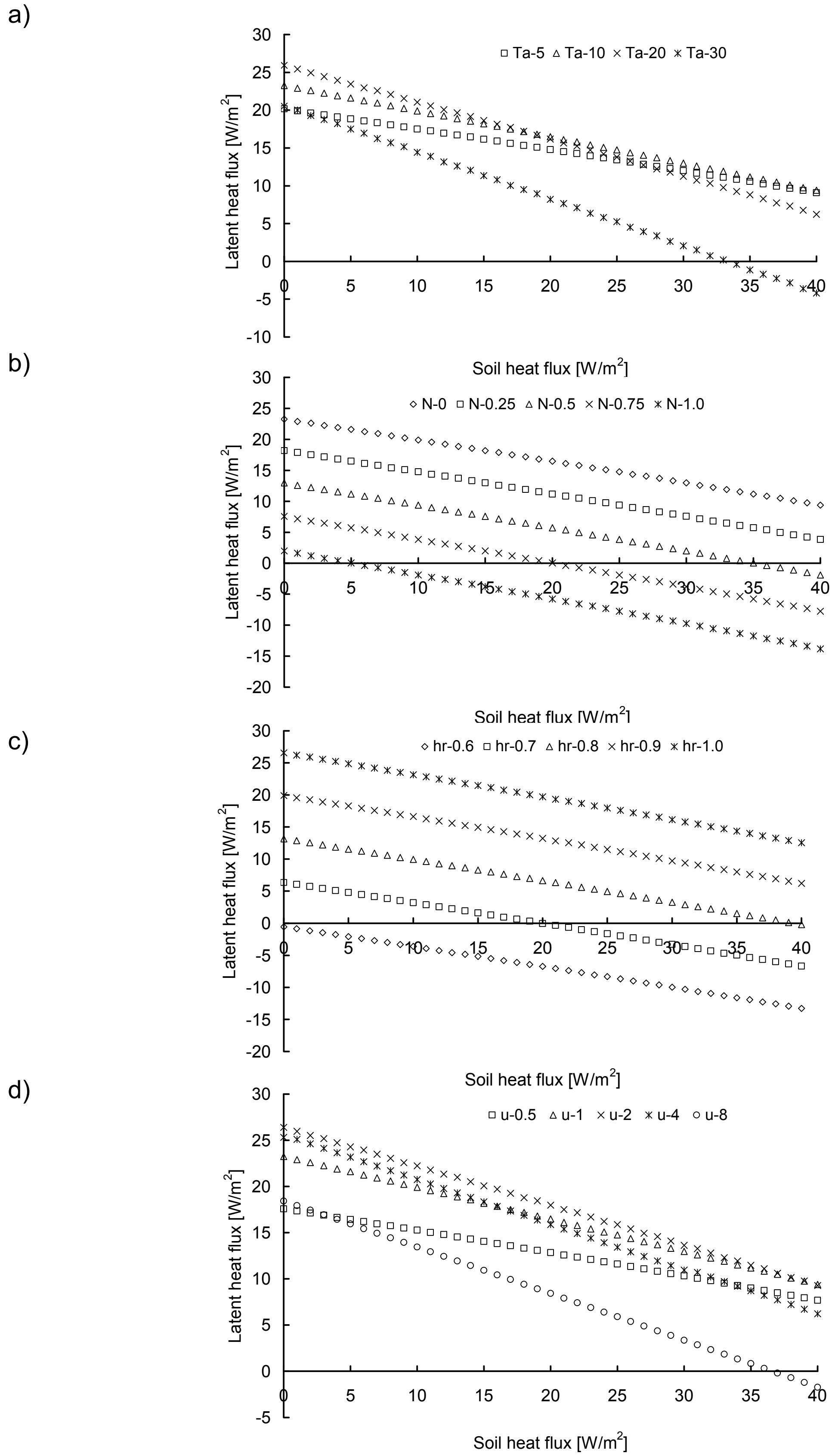


Figure 2.4: Effect of soil heat flux on latent heat flux at different values of a) air temperature (T_a), b) cloud cover (N), c) relative humidity (h_r), and d) wind speed (u).

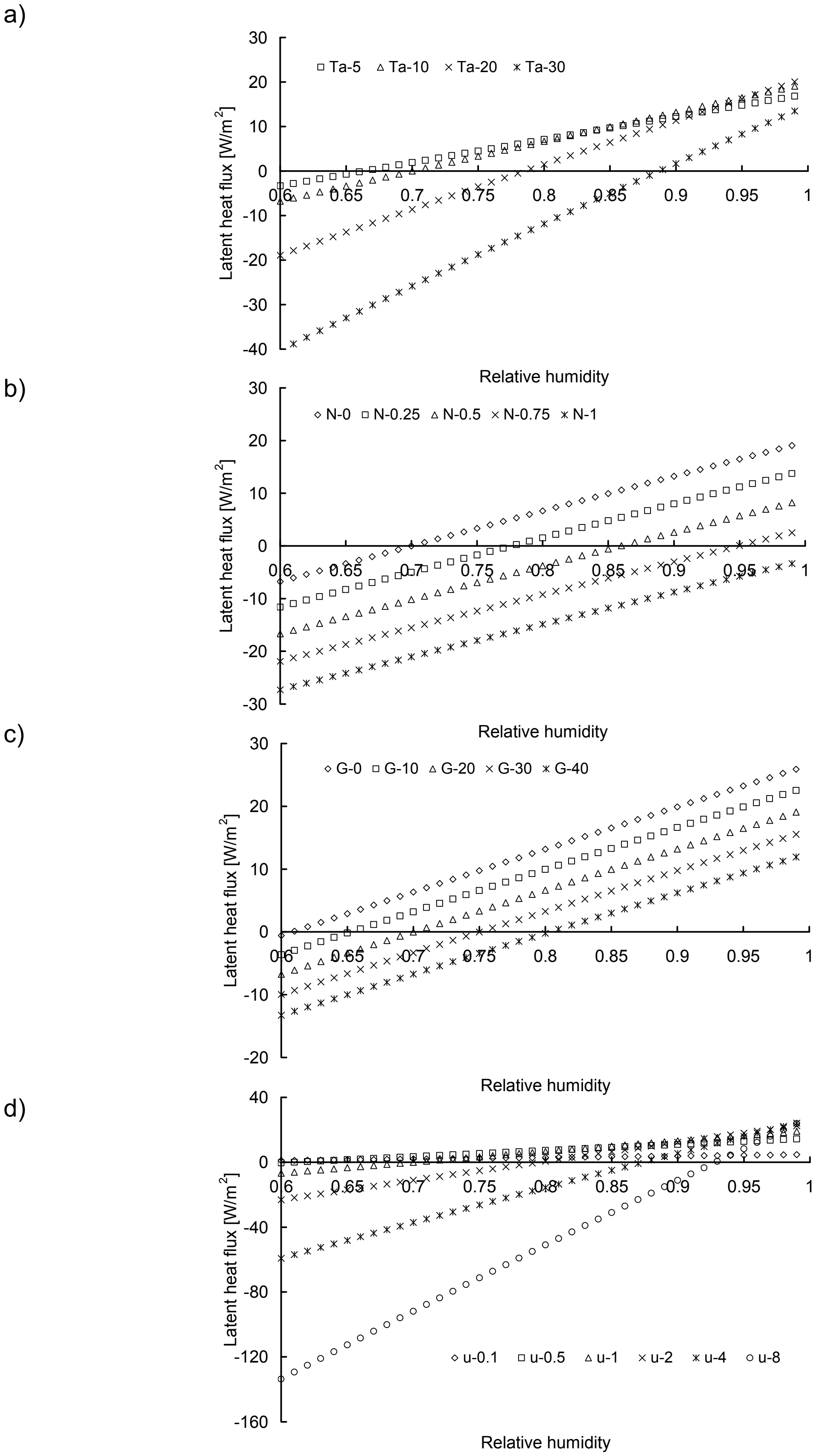


Figure 2.5: Effect of relative humidity flux on latent heat flux at different values of a) air temperature (T_a), b) cloud cover (N), c) soil heat flux (G), and d) wind speed (u).

2.4.5 Wind speed

The data in Figure 2.6a - d are the same as in Figure 2.2d to 2.5d, but plotted in a different fashion, which clarifies the points made in those figures. For all parameter combinations employed the relationship between latent heat flux and wind speed is non-linear (Fig. 2.6a - d). Also, with two exceptions (N = 1.0 in Fig. 2.6b, and h_r = 0.6 in Fig 2.6d) λE increases with u up to some value at which it begins to decline. Again, the precise nature of the relationship between λE and u is influenced by other meteorological factors.

The rate of increase in λE with u is roughly similar for all values of T_a (Fig. 2.6a), but the peak reached differs for the various temperatures. Also the subsequent rate of decline in λE with u is less the lower T_a. Furthermore, at a given wind speed the effect of T_a on λE is less the smaller T_a is, and the higher u the bigger the effect of T_a. Lastly, the wind speed at which dewfall gives way to evaporation declines as T_a goes up.

Except for N = 1, for which λE declines continuously as u goes up, initially in a concave, then in a linear fashion, the rate of increase in λE with u becomes steeper with decreasing N and the peak value higher (Fig. 2.6b). However, the subsequent rate of decline in λE with u is about the same, i.e. at any given wind speed after the peak for λE the change in λE with N is about the same. Finally, the lower N, the higher u at which dewfall changes into evaporation.

The influence of G on the relationship between λE and u is similar to that of N, except that the peaks are higher, and that even at the highest G looked at here there is an initial increase in λE followed by a decline (Fig. 2.6c).

With rising h_r the rate of increase in λE with u becomes steeper and the subsequent decline flatter (Fig. 2.6d). For h_r = 1 there is no decline; λE stays about constant once the peak has been reached. At a given u the change in λE with h_r is constant, but the higher u, the bigger this constant change. The more h_r moves away from 1, the lower the wind speed at which dewfall is supplanted by evaporation.

2.4.6 Importance of the various meteorological factors

All five meteorological factors identified here to influence dewfall can have roughly the same impact. Each of them can lead to variations in dewfall between 0 and ~ 25 W m^{-2}. The precise magnitude of the impact depends on the value of the other factors

The computations revealed that dewfall is always highest if N = 0, G = 0 and h_r = 1. At which air temperature dewfall is highest depends on wind speed, and vice versa.

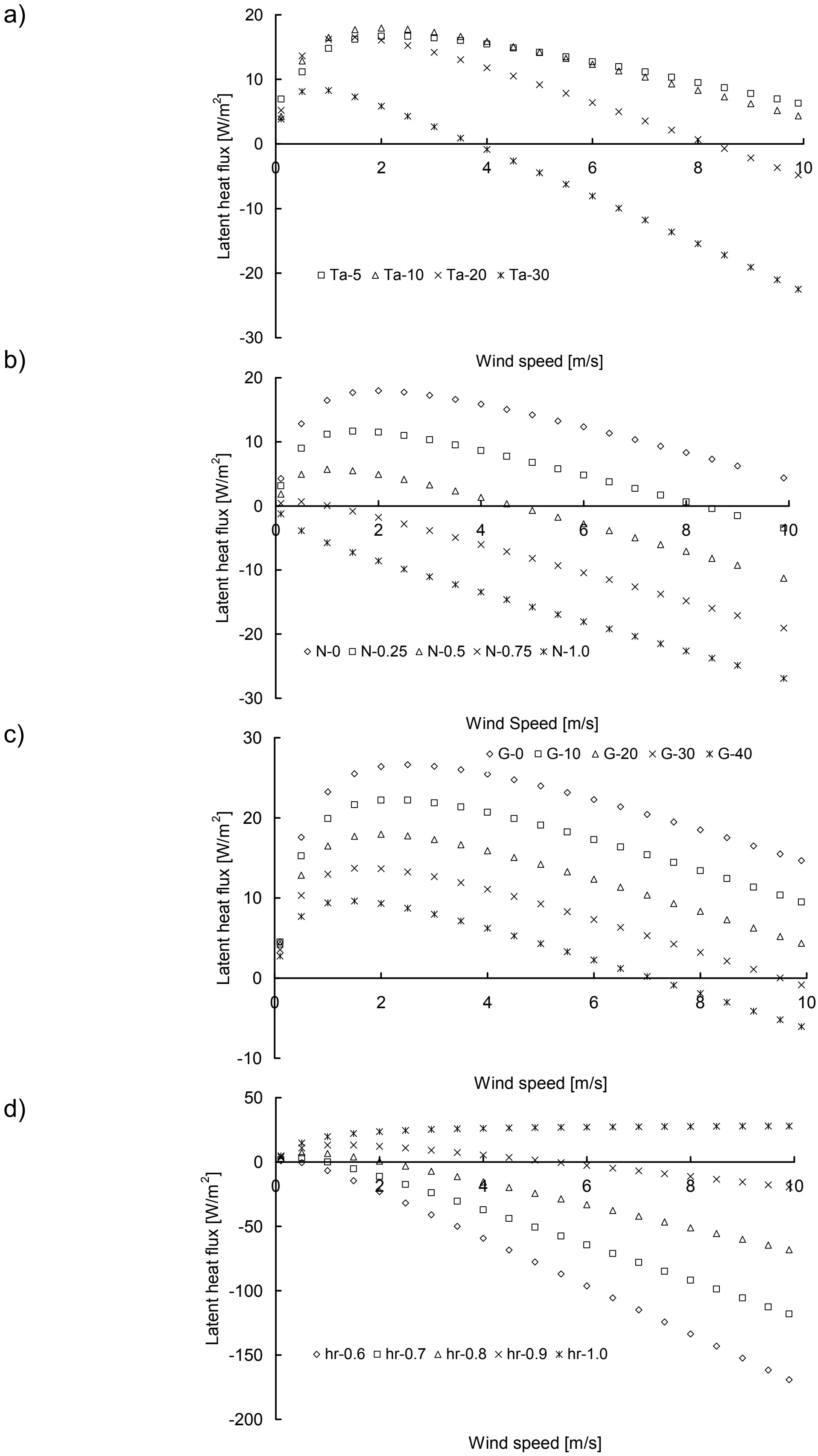


Figure 2.6: Effect of wind speed on latent heat flux at different values of a) air temperature (T_a), b) cloud cover (N), c) soil heat flux (G), and d) relative humidity (h_r).

2.5 Discussion

In section 2.4 the effect on latent heat flux of various meteorological factors was calculated. In this section, reasons for that are explained by analysing the variations of vapour conductance and vapour pressure gradient caused by a change in meteorological factors. The following statements are largely based on calculations with the equations in section 2.3

2.5.1 Air temperature

2.5.1.1 Vapour pressure gradient in relation to air temperature

According to Eq. 2.11 air vapour pressure (e_a) is determined by air temperature and relative humidity. Since in the calculations here relative humidity (h_r) was set at a certain value (0.95), e_a is only influenced by air temperature (Fig. 2.7a), which means that at the same h_r more moisture is contained in the same volume of air. In addition, Eq. 2.12 demonstrates that air temperature is also involved in determining surface temperature. Since a change in surface temperature results in a change of e_s, the latter is indirectly related to air temperature.

The change either in e_a or e_s gives rise to a variation of the vapour pressure gradient ($[e_a - e_s)] / \Delta z$), which leads to a change in latent heat flux. In this manner air temperature affects the latent heat flux.

Figure 2.7b shows that surface temperature (T_s) always increases with air temperature (T_a). Eq. 2.4 and 2.6 indicate that the long wave incoming radiation is a function of T_a and cloud cover. If cloud cover is given, as in the calculations here, then a higher T_a translates into more incoming radiation, which results in an increase of T_s. If the nature of the surface remains the same, the increased surface temperature leads to a greater e_s, as depicted in Figure 2.7b, too.

An increase in e_a makes the vapour pressure gradient larger. In contrast, an increase in e_s results in a smaller vapour pressure gradient. However, simultaneous increases in both make a change in the vapour pressure gradient with air temperature more complicated. Figure 2.7c displays the vapour pressure gradient change with air temperature. It initially increases up to a certain air temperature, but decreases thereafter. This tendency means that the increase in e_a is initially larger that that of e_s, but beyond a certain air temperature the situation reverses.

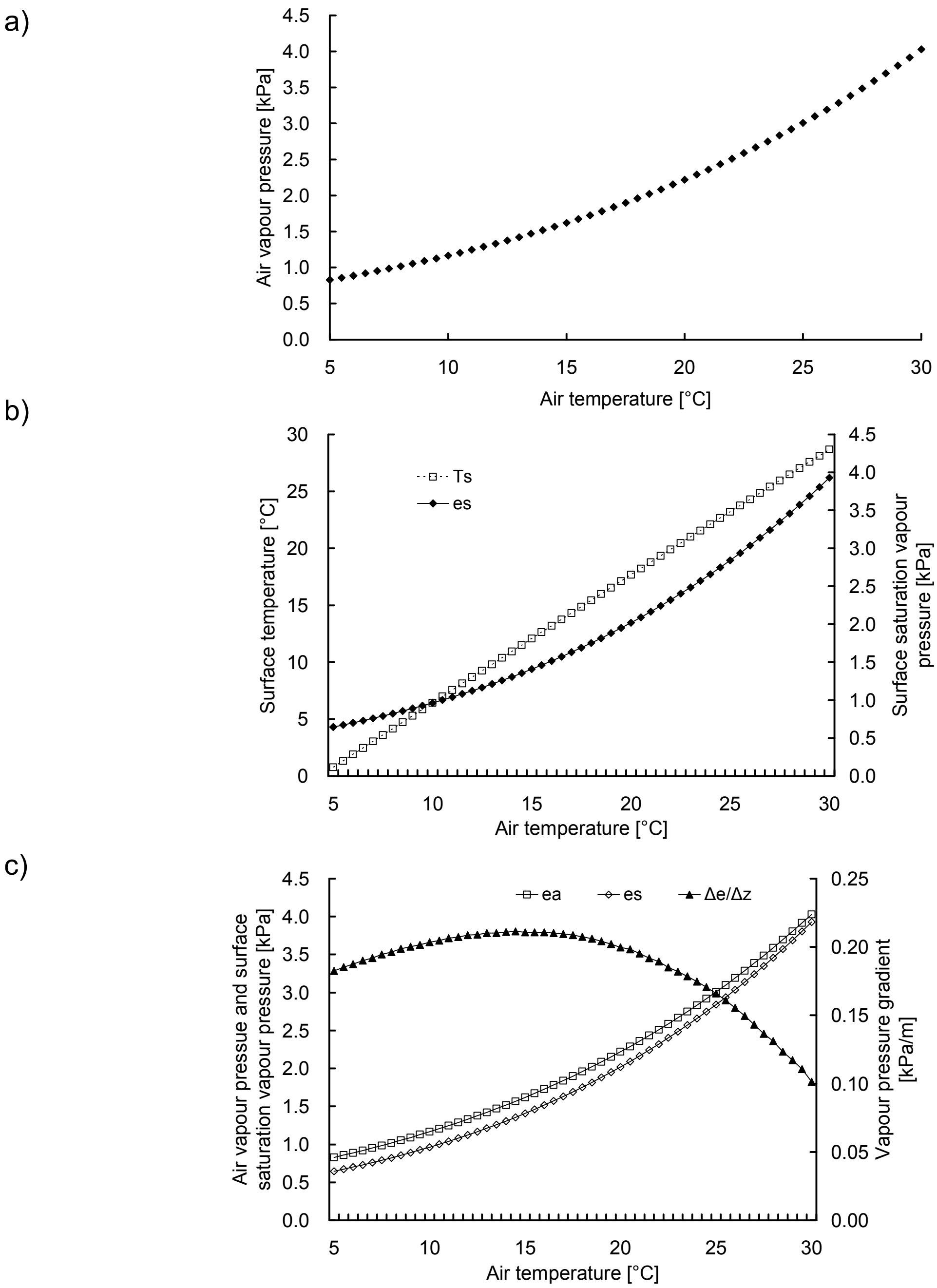


Figure 2.7: Saturation vapour pressure (a), surface temperature (b), surface saturation vapour pressure (b, c), air vapour pressure (c), and vapour pressure gradient (c) in relation to air temperature.

2.5.1.2 Latent heat flux in relation to air temperature

If wind speed is fixed, the vapour conductance is constant (cf. Eq. 2.8). Therefore, the change in latent heat flux follows the change in vapour pressure gradient (cf. Eq. 2.9). The change in vapour pressure gradient in turn varies in a curvilinear fashion with T_a (Fig. 2.7c). Consequently, latent heat flux initially increases with air temperature, but after a certain temperature drops again. This is demonstrated in Figure 2.2. A flow chart of how latent heat flux changes with air temperature is given in Figure 2.8.

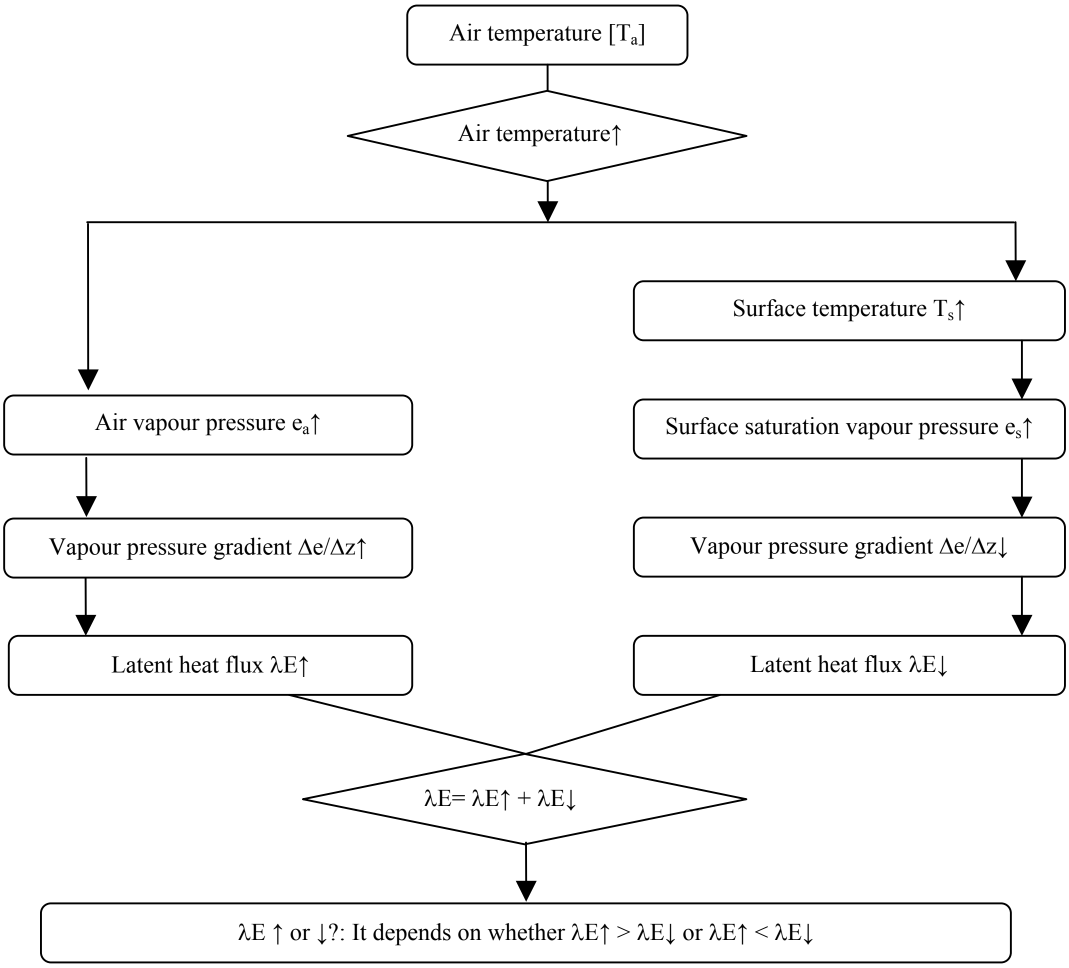
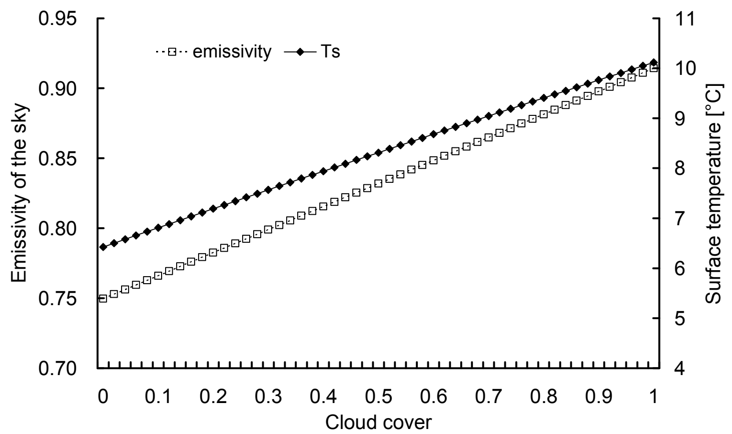


Figure 2.8: Flow chart of latent heat flux changing with air temperature.

2.5.2 Cloud cover

Surface temperature is also affected by cloud cover, since cloud cover is associated with the emissivity of the sky, which affects the incoming long wave radiation. This relationship is described by Eq. 2.4 and 2.6. Figure 2.9a shows that the emissivity always increases with cloud cover. Due to the increased sky emissivity, the nocturnal long wave radiation coming from the sky increases. Consequently, the radiational cooling weakens, which results in an increase in surface temperature.

a)



b)

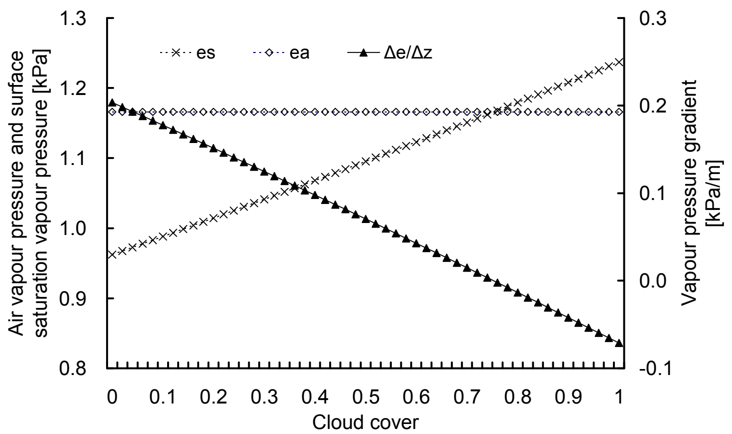


Figure 2.9: Emissivity of the sky (a), surface temperature (a), air vapour pressure and surface saturation vapour pressure (c), and vapour pressure gradient (c) in relation to cloud cover.

Without a change in wind speed a change in the vapour pressure gradient is the only cause leading to a variation of latent heat flux. Figure 2.9b shows that e_s and the vapour pressure gradient change with cloud cover. At a constant air temperature and relative humidity, e_a remains stable, but an increase in e_s, which is due to an increased surface temperature, leads to a decrease in the vapour pressure gradient. As explained in the foregoing, the decreased vapour pressure gradient causes λE to decrease. This is the reason for the result of Figure

2.3, which shows that λE falls with cloud cover. The flow chart of latent heat flux in relation to cloud cover is given in Figure 2.10.

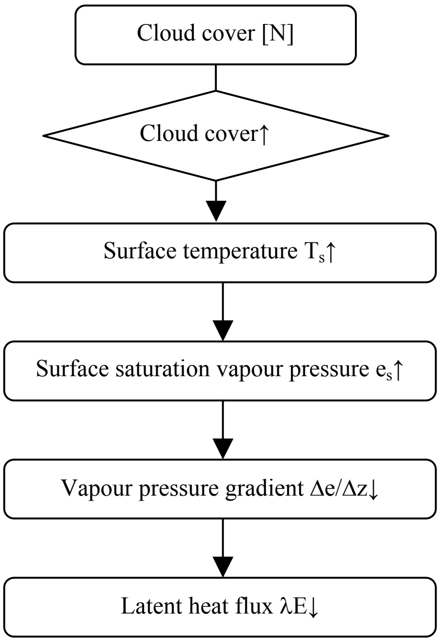


Figure 2.10: Flow chart of latent heat flux in relation to cloud cover.

2.5.3 Soil heat flux

At night the surface temperature generally decreases due to radiative cooling. However, a part of the nocturnal radiation loss can be offset by heat transferred from the soil. Therefore, as illustrated in Figure 2.11a, with an increased soil heat flux towards it the surface becomes warmer. This in turn leads to an increase in e_s. If e_a keeps constant, the vapour pressure gradient then declines with a rise in soil heat flux. Overall this leads to a decrease in latent heat flux (Fig. 2.11b). The flow chart of latent heat flux in relation to soil heat flux is depicted in Figure 2.12.

2.5.4 Relative humidity

An increase in relative humidity has a twofold effect on dew formation. On the one hand, a higher relative humidity leads to an increased surface temperature. This is a somewhat surprising result from Eq. 2.12, which is plotted in Figure 2.13a. As mentioned above, an increased surface temperature induces an increase in e_s, which is illustrated in Figure 2.13b. An increase in e_s is not favourable for dew formation, because it reduces the vapour pressure gradient, all other factors being equal. On the other hand, Figure 2.13b also indicates that with a rise in relative humidity, e_a increases. An increased e_a is favourable for dew formation, since this increases the vapour pressure gradient.

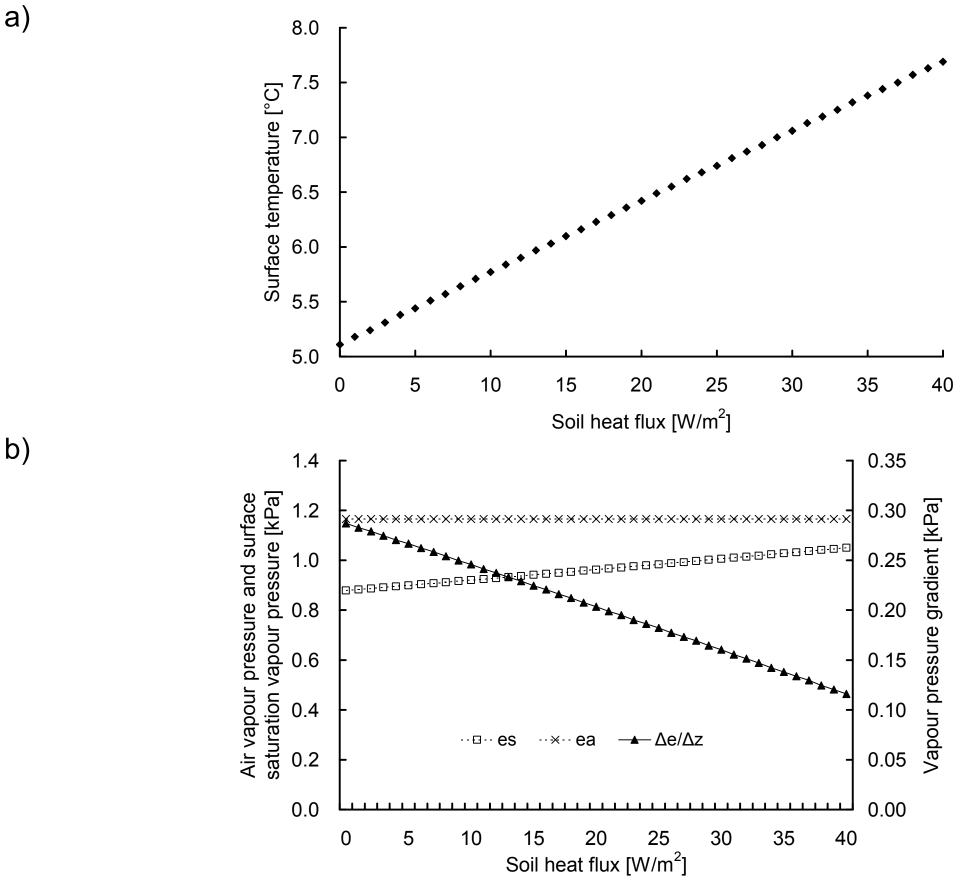


Figure 2.11: Surface temperature (a), air vapour pressure and surface saturation vapour pressure (b), and vapour pressure gradient (b) in relation to soil heat flux.

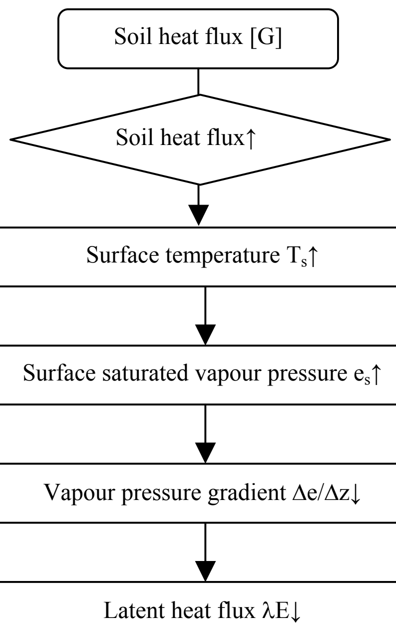


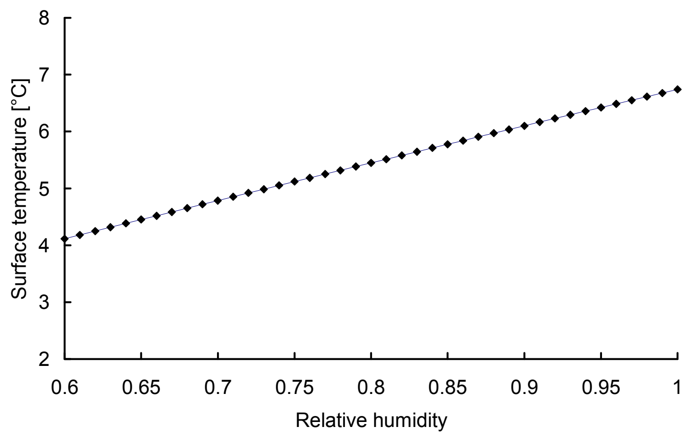
Figure 2.12: Flow chart of latent heat flux in relation to soil heat flux.

Because an increased relative humidity can lead to an increase in e_a as well as in e_s and both affect the vapour pressure gradient, the change in the vapour pressure gradient is decided by which increase is greater. Figure 2.13b indicates that, although e_s increases, the increase in e_a with relative humidity is faster. Hence, the vapour pressure gradient increases with relative humidity.

In addition, Figure 2.13b demonstrates that under the conditions in the calculations here, the vapour pressure gradient is only positive, if the relative humidity is > 70%. It this case e_a is larger than e_s and dew formation becomes possible.

Figure 2.14 gives the flow chart of latent heat flux in relation to relative humidity.

a)



b)

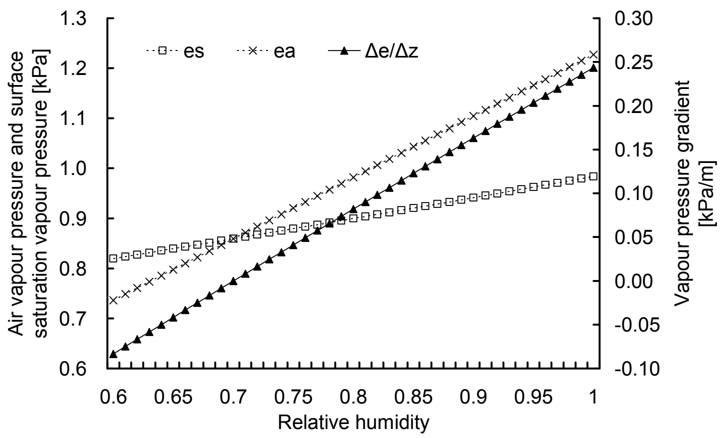


Figure 2.13: Surface temperature (a), air vapour pressure and surface saturation vapour pressure (b), and vapour pressure gradient (b)in relation to relative humidity.

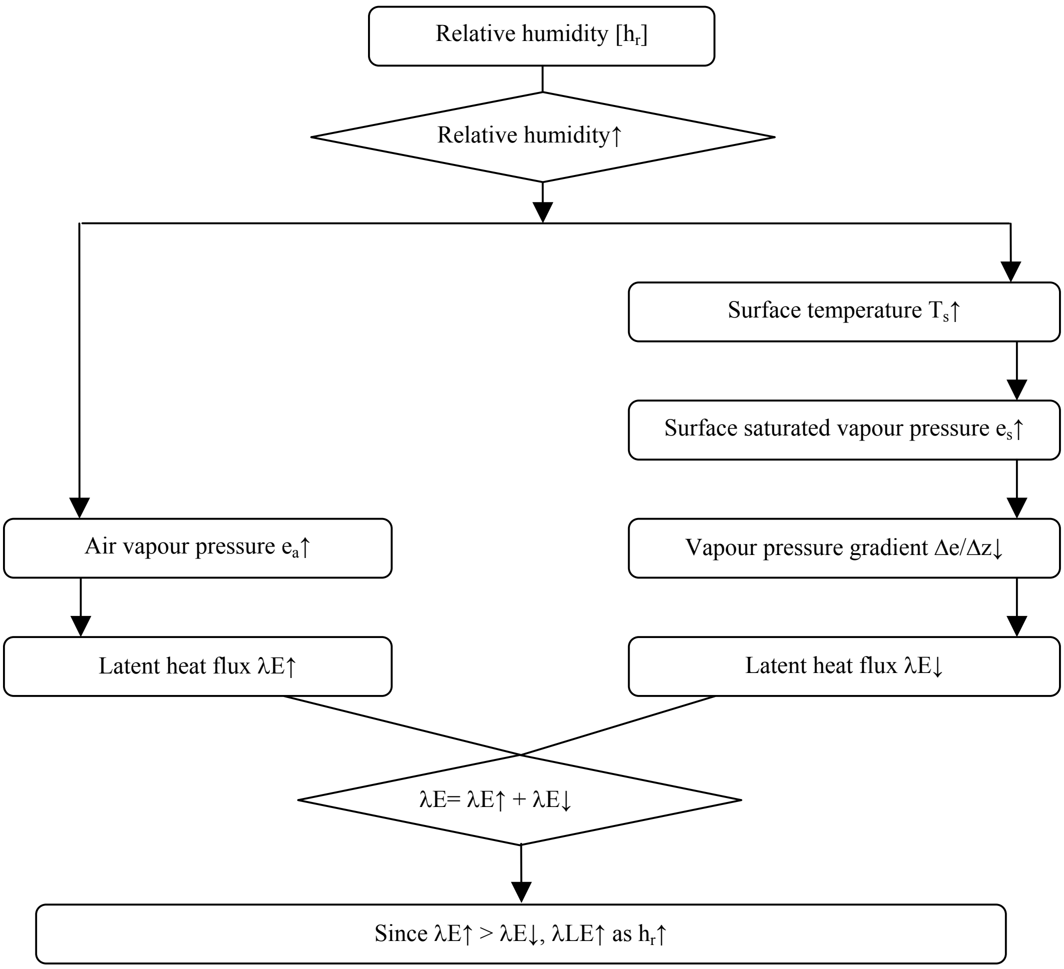


Figure 2.14: Flow chart of latent heat flux in relation to relative humidity.

2.5.5 Wind speed

Figure 2.1 indicates that wind speed affects both vapour conductance and the vapour pressure gradient. For these two aspects an analysis of the variation of latent heat flux arising from a change in wind speed is performed below.

2.5.5.1 Vapour conductance in relation to wind speed

When there is no change in the nature of a surface, only wind speed can lead to a variation in vapour conductance (cf. Eq. 2.8). Figure 2.15 demonstrates that the vapour conductance then always increases with wind speed. Furthermore, the figure shows a linear relationship between vapour conductance and wind speed; the rate of increase in vapour conductance with wind speed is constant. If this was the only effect, an increased wind speed would always facilitate dew formation.

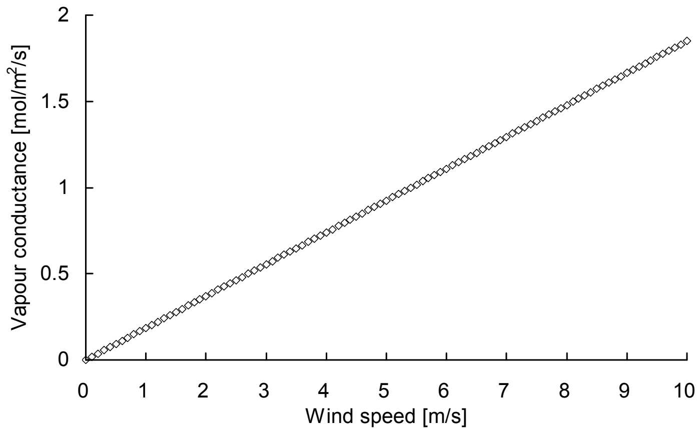


Figure 2.15: Vapour conductance in relation to wind speed.

2.5.5.2 Vapour pressure gradient in relation to wind speed

The vapour pressure gradient is determined by e_a and e_s. Figure 2.16a shows that a change in wind speed results in a change of surface temperature, which causes e_s to change since it is a function of surface temperature. The variation of e_s gives rise to a change in the vapour pressure gradient. The reason for the vapour pressure gradient changing with wind speed can be explained in detail as follows.

Figure 2.16a shows that T_s increases with wind speed, if the surface is cooler than the surrounding air, as in the calculations presented above. A higher wind velocity means increased convection, which in turn favours mixing of the air layers. Hence, it results in an increased surface temperature. However, when the wind speed is beyond a certain limit, there is no more change in surface temperature with wind speed, because a thorough mixing has been attained so that $T_s = T_a$.

Figure 2.16b shows how e_a and e_s respond to wind speed. If air temperature and humidity are kept constant, as in the calculations above, e_a does not change with wind speed. However, due to the increase of T_s with wind speed, e_s increases with wind speed. Moreover, Figure 2.16b indicates that e_s comes ever closer to e_a as wind speed increases. Consequently, the vapour pressure gradient decreases with wind speed (Fig 2.16c) and eventually reaches zero: At this point λE is also zero and dew formation ceases. Therefore, with respect to the vapour pressure gradient, increased wind speed hinders dew formation.

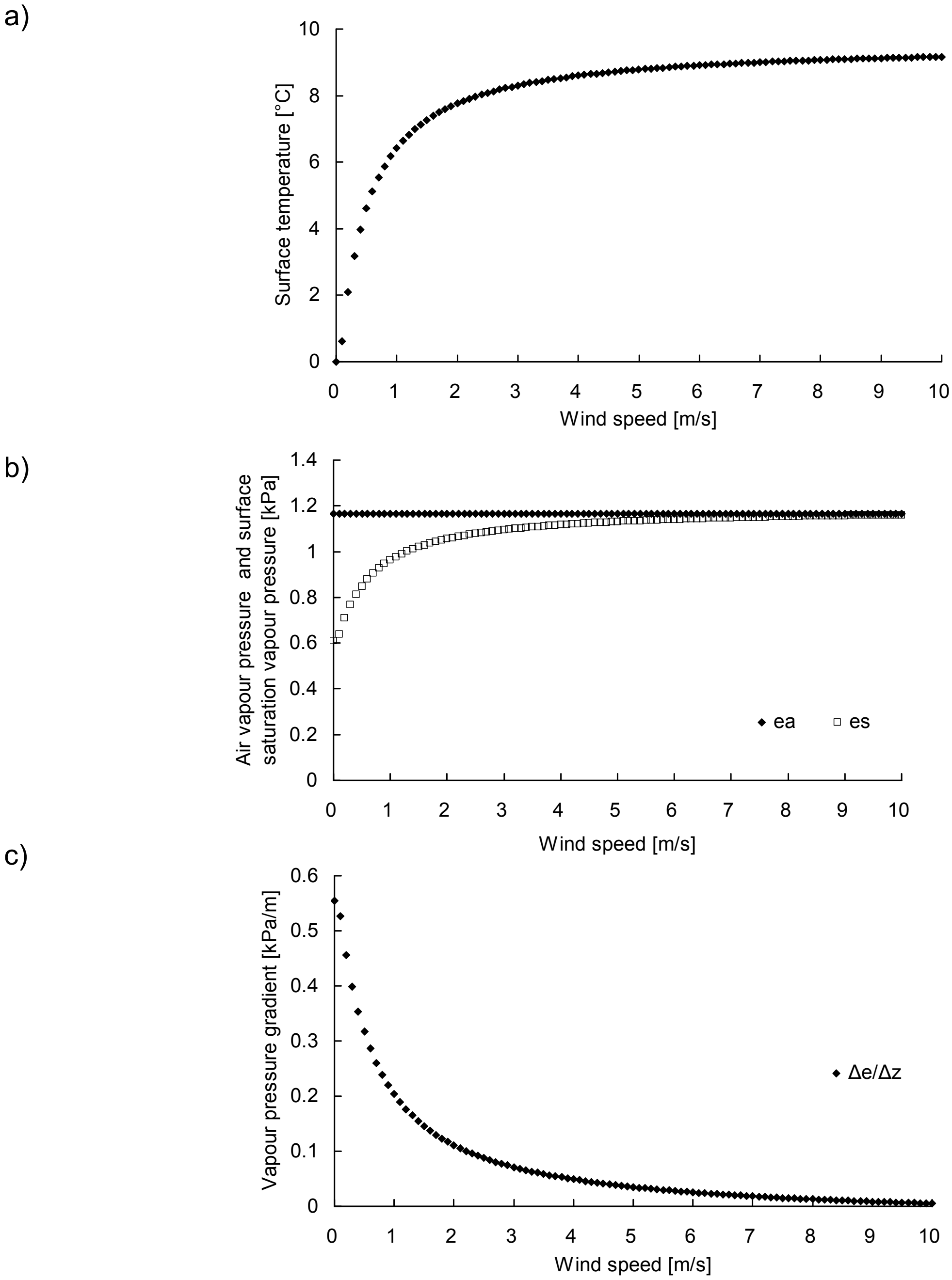


Figure 2.16: Surface temperature (a), air vapour pressure and surface saturation vapour pressure (b), and vapour pressure gradient (c) in relation to to wind speed.

2.5.5.3 Latent heat flux in relation to wind speed

Latent heat flux is the product of vapour conductance and vapour pressure gradient (Eq. 2.9). Figure 2.16c indicates that an increased wind speed leads to a decreased vapour pressure gradient, which induces a decrease in latent heat flux. However, Figure 2.15 shows that a rise in wind speed leads to an increased vapour conductance, which results in an increased in latent heat flux. How λE finally changes with wind speed therefore depends on the balance of these two phenomena.

Figure 2.6 indicates that dewfall initially increases and later decreases with wind speed. The reason is that at first the falling vapour pressure gradient is more than offset by the rising conductance, while later the rising conductance is more that offset by the falling vapour pressure gradient. The flow chart showing the dependence of latent heat flux on wind speed is plotted in Figure 2.17.

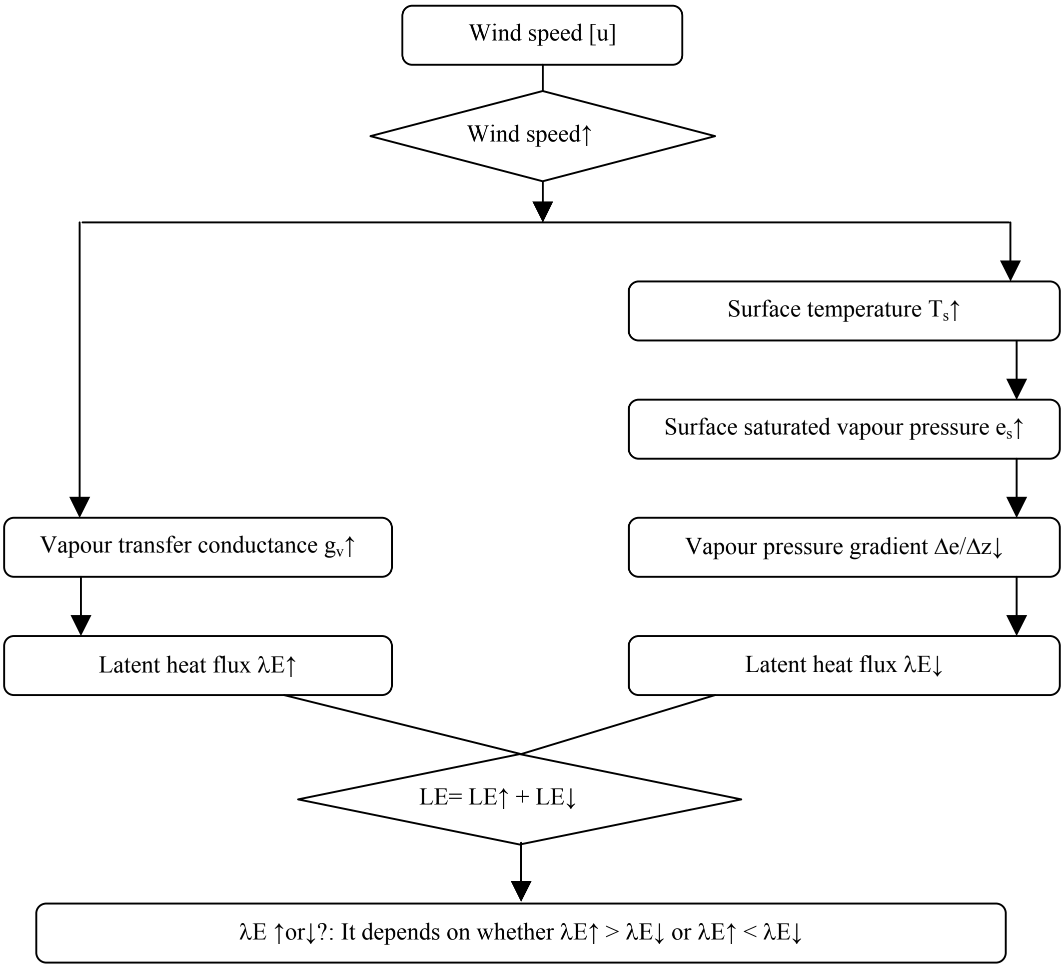


Figure 2.17: Flow chart of latent heat flux in relation to wind speed.

2.6 Conclusions

The five meteorological variables identified here to influence dewfall (T_a, N, G, h_r, u) have been mentioned in the literature before. However, their effect was generally not analysed as thoroughly. Net radiation (R_n), which is also frequently named as an important factor, was not considered explicitly here. Instead, the Stefan-Boltzmann equation was used to compute the incoming and outgoing long wave radiation with the help of air and surface temperature, respectively. (Since dewfall occurs mostly at night or in the early morning, short wave radiation was ignored.)

Three factors, namely N, G and h_r, appear only once and in a linear fashion in the version of the energy balance equation employed here (Eq. 2.12). Consequently, their effect on dewfall is linear, too. Dewfall increases as N or G decrease, or h_r increases.

Air temperature appears tree times in Eq. 2.12, twice in a non-linear form. Also, in one term an increasing T_a enhances dewfall, in two terms it depresses it, and vice versa. The relationship between T_a and dewfall is therefore non-linear. Dewfall initially rises, but later declines with T_a.

Wind speed appears two times in Eq. 2.12, once such that it benefits dewfall, and once such that it hinders it. As a result, the overall relationship between u and dewfall is non-linear, even though the u term itself is linear both times. Similar to T_a, dewfall initially rises, but later declines with u.

The potential magnitude of T_a, N, G, h_r and u on dewfall was found to be similar. The actual magnitude is determined by the values of the other factors. Dewfall is at its highest, if N = 0, G = 0 and $h_r = 1$. At which air temperature dewfall is highest depends on wind speed and vice versa.

2.7 References

Baier, W. (1966): Studies on dew formation under semi-arid conditions. Agric. Meteorol. 3, 103-112.

Bass, B., Savdie, I., Gillespie, T. J. (1991): Simulation of leaf wetness duration for field corn. Agric. For. Meteorol. 57, 69-84.

Beysens, D., Muselli, M., Nikolayev, V., Narhe, R., Milimouk, I. (2005): Measurement and modelling of dew in island, coastal and alpine areas. Atmos. Res. 73, 1-22.

Brutsaert, W. (1982): Evaporation into the atmosphere: theory, history, and applications. Reidel, Dordrecht.

Campbell, G. S., Norman, J. M. (1998): An introduction to environmental biophysics, 2nd ed. Springer, New York.

Crowe, M. J., Coakley, S. M., Emge, R. G. (1978): Forecasting dew duration at Pendleton, Oregon, using simple weather observations. J. Appl. Meteorol. 17, 1482-1487.

Davie, T. (2003): Fundamentals of hydrology. Routledge, London.

Evett, S. R. (2001): Water and energy balances at soil-plant-atmosphere interfaces. In: Warrick , A. W. (ed.), Soil physics companion (pp. 127-188). CRC Press, Boca Raton.

Garratt, J. R., Segal, M. (1988): On the contribution of atmospheric moisture to dew formation. Bound.-Layer Meteorol. 45, 209-236.

Gates, D. M. (1965):. Heat transfer in plants. Sci. Am. 213,76-84.

Kidron, G. J., Yaalon, D. H., Vonshak, A. (1999): Two causes for runoff initiation on microbiotic crusts: hydrophobicity and pore clogging. Soil Sci. 164, 18-27.

Kidron, G. J., Yair, A., Danin, A. (2000): Dew variability within a small arid drainage basin in the Negev Highlands, Israel. Q. J. R. Meteorol. Soc. 126, 63-80.

Lhomme, J .P., Jimenez, O. F. (1992): Estimating dew duration on banana and plantain leaves from standard meteorological observations. Agric. For. Meteorol. 62, 263-274.

Long, I. F. (1958): Some observations on dew. Meteorol. Mag. 87, 161-168.

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

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

Monteith, J. L. (1963): Dew: facts and fallacies. In: Rutter, A. J., Whitehead, F. H. (eds), The water relations of plants (pp. 37-56). Wiley, New York.

Monteith, J. L., Unsworth, M. H. (1990): Principles of environmental physics, 2nd ed. Arnold, London.

Muselli, M., Beysens, D., Marcillat, J., Milimouk, I., Nilsson, T., Louche, A. (2002): Dew water collector for potable water in Ajaccio (Corsica Island, France). Atmos. Res. 64, 297-312.

Nakshabandi, G. A., Kohnke, H. (1965): Thermal conductivity and diffusivity of soils as related to moisture tension and other physical properties. Agric. Meteorol.2, 271-279.

Neumann, J. (1956): Estimating the amount of dewfall. Arch. Meteorol. Geophys. Bioklimatol., Ser. A, 9, 197-203.

Ninari, N., Berliner, P. R. (2002): The role of dew in the water and heat balance of bare loess soil in the Negev Desert: quantifying the actual dew deposition on the soil surface. Atmos. Res. ,64, 323-334.

Paltridge G. W., Platt, C. M. R.(1976): Radiative processes in meteorology and climatology. Elsevier, Amsterdam.

Pedro, M. J., Gillespie, T. J. (1982a): Estimating dew duration. I. Utilizing micrometeorological data. Agric. Meteorol. 25, 283-296.

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

Riley, J. A., Giles W. L. (1965): Agricultural meteorology in relation to the use of pesticides in the U.S.A. Agric. Meteorol. 2, 225-245.

Rosenberg, N. J. (1974): Microclimate: the biological environment. Wiley, New York.

Scherm, H., van Bruggen, A. H. C. (1993a): Sensitivity of simulated dew duration to meteorological variations in different climatic regions of California. Agric. For. Meteorol. 66, 229-245.

Shaw, R. H. (1973): Dew duration in central Iowa. Iowa State J. Res. 47, 219-227.

Smith, L. P. (1958): The duration of surface wetness. Proc. Int. Hort. Congr. 15, 478-484.

Smith, H., Carpenter, R. C. (1966): Dew forecasting study for South Carolina. Agricultural weather studies for South Carolina. South Carolina Deptartment of Agriculture, Columbia.

Tuller, S. E., Chilton, R. (1973): The role of dew in the seasonal moisture balance of a summer-dry climate. Agric. Meteorol. 11, 135-142.

Wallin, J. R. (1963): Dew, its significance and measurement in phytopathology. Phytopathology 53, 1210-1216.

Wallin, J. R. (1967): Agrometeorological aspects of dew. Agric. Meteorol. 4, 85-102.

Ye, Y. H., Zhou, K., Song, L. Y., Jin, J. H., Peng, S. L. (2007): Dew amounts and its correlations with meteorological factors in urban landscapes of Guangzhou, China. Atmos. Res. 86, 21-29.

3. Testing the precision of a weighable gravitation lysimeter

3.1 Abstract

Tests were carried out to determine the weighing precision of a 2 m deep lysimeter with a 1 m² cross-sectional area and a total mass of 3,500 to 3,850 kg, depending on the soil water content. The weighing mechanism consists on three shear stress cells laid out for a load capacity of 1,320 kg each.

Mass changes as small as 20 g, which is equivalent here to a water gain or loss of 0.02 mm, can be measured with good accuracy and stability under favourable environmental conditions (low wind speed and relatively constant temperature). This precision does not depend on the position on the lysimeter where the mass change occurs, and is as good as the best values reported in the literature for other lysimeters.

To prevent water and debris from entering the cleavage between lysimeter vessel and pit casing, a rubber collar can be placed across the cleavage. It is attached to the casing and extends about 1 to 2 cm into the vessel. Although the collar is not supposed to touch the vessel, it does at a few points. This seriously lowers weighing precision, because this contact exerts forces on the vessel, which distort the true weight. Hence, one should refrain from using this type of collar and develop another one.

Weighing precision decreases with increasing wind speed, because wind exerts forces on the lysimeter vessel and can thus alter its apparent weight. It is temperature-dependent, too.

3.2 Introduction

Lysimeters are an important tool for water balance studies in agriculture, forestry and other environmental settings. In particular, they make it possible to quantify actual evaporation from a bare soil or actual evapotranspiration from a soil covered by vegetation. Moreover, seepage from lysimeters can be collected, which allows an assessment of the water loss from a soil profile and, thus, groundwater recharge. The seepage water can be analysed in the laboratory for its various constituents. Hence, lysimeters can be used to monitor the fate of solutes in a soil, too.

Lysimeters are classified as weighable or non-weighable. Non-weighable lysimeters are mainly useful to monitor seepage and solute leaching from a soil profile. Weighable lysimeters can monitor the mass continuously and thus provide detailed information about water storage changes in the soil for any time period. In conjunction with rainfall and seepage measurements water losses can then be specified as seepage or evapotranspiration.

A key parameter of a lysimeter is its weighing precision: the higher it is, the better the resolution of the weight measurements. This affects the questions which can be addressed with a lysimeter. For example, a high resolution makes it possible to chart seepage and evapotranspiration over short periods such as hours or less, while a low resolution may only allow daily values. Also, small mass inputs such as precipitation in the form of dew, fog or rime can be determined with a lysimeter of high weighing precision. Table 3.1 list the precision of various lysimeters as reported in the literature.

Table 3.1: Weighing precision of some lysimeters as reported in the literature.

Reference	Weighing system[a]	Cross-sectional area (m^2)	Depth (m)	Precision (g)	Precision (mm of water)
Pruitt and Angus (1960)	LC-W-CB	28.27	0.91	850	0.03
Fritschen et al. (1973)	HLC	10.75	1.2	650	0.06
McFarland et al. (1983)	LC-WO-CB	4.68	1.52	4680	1.0
Kirkham et al. (1984)	DS-WO-CB	2.31	1.30 and 1.60	50	0.02
Howell et al. (1985)	LC-W-FLAB	4.0	2.0	80	0.02
Sayler et al. (1985)	Load ring	0.89	1.37	270	0.3
Marek et al. (1988)	LC-W-CB	9.0	2.3	400	0.045
Allen and Fisher (1990)	LC-WO-CB	1.0	1.2	50	0.05
Howell et al. (1991)	LC-W-CB	9.0	2.3	450	0.05
Schneider et al. (1993)	LC-WO-CB	0.75	2.3	350	0.47
Plauborg (1995)	not specified	0.56	1.5	280	0.5
Qiu et al. (1996)	not specified	1.77	1.5	50	0.028
Young et al. (1997)	LC-W-CB	4.91	4.0	200	0.04
Schneider et al. (1998)	DS-WO-CB	2.25 and 3.0	2.44 and 1.60	220 and 60	0.1 and 0.02
Girona et al. (2004)	LC-W-CB	9.5	1.7	500	0.053
Meshkat et al. (1999)	LC-W-CB	0.44	0.8	10	0.025
Malone et al. (2000)	LC-W-CB	8.1	2.4	260	0.032
Gholam and Mohammad (2002)	LC-W-FLAB	3.0	1.75	840	0.28
Hunsaker et al. (2002)	LC-W-CB	1.0	1.6	30	0.03
Tyagi et al. (2003)	LC-WO-CB	3.94	1.98	200 - 3940	0.05 - 1.0
Unold (2003)	LC-W-CB	1.0	2.0	50	0.05
Yang et al. (2003)	not specified	1.77	1.6	50	0.028
Zhang et al. (2004)	LC-W-CB	3.0	2.5	60	0.02
Yoder et al. (2005)	LC-W-CB	4.0	1.8	200	0.05
Jia et al. (2006)	LC	2.32	1.37	280	0.12
Marek et al. (2006)	DS-WO-CB	3.0	2.5	10	0.0036
Gavilan and Berengena (2007)	CWPS	6.0	1.5	180	0.03
Meissner et al. (2007)	LC-WO-CB	1.0	2.0	30	0.03
Rupp et al. (2007)	LC-WO-CB	4.0	1.5	400	0.1

[a)] HLC: hydraulic load cell, LC-W-CB: load cell with counter-balance, LC-WO-CB: load cell without counter-balance, LC-W-FLAB: load cell with flexure-level action balance, DS-WO-CB: deck scale without counter-balance, CWPS: counter-weighted platform scale

In general, the absolute precision of a weighing mechanism, e. g. whether it can weigh to the nearest 1, 10 or 100 g, depends on the maximum load it is laid out for: the higher this load, the lower the absolute precision. For a lysimeter the cross-sectional area is of importance, too. For example, a lysimeter with an absolute weighing precision of 100 g and a cross-sectional area of 1 m² can register a water loss or gain of 0.1 mm. The same precision for an area of 2 m² translates into 0.05 mm, which is twice the precision in terms of a water loss or gain. Hence, given the same total mass and weighing precision, a shallower lysimeter with a bigger diameter has a better resolution for water balance studies than a deeper one with a smaller diameter.

The purpose of this paper is to test the precision of a type of weighable lysimeter installed at the Falkenberg research station of the UFZ - Helmholtz Center for Environmental Research to asses, whether it is good enough for use in the measurement of dew, fog or rime. Weighing precision in the context of this study has two aspects: 1) the accuracy a given mass change is measured with, and 2) the stability of the measurement over time.

3.3 Materials and methods

3.3.1 Description of the lysimeter

Figure 3.1 schematically illustrates the design of the type of weighable gravitation lysimeter investigated in this study. The circular lysimeter vessel is made of high density polyethylene, has a diameter of 1.13 m, leading to a surface area of 1.0 m^2, and a depth of 2.0 m. The bottom 25 cm are occupied by a filter layer (sand over coarse sand over gravel), the remainder is filled with a sandy soil material. TDR probes, tensiometers (combined with thermometers) and suction cups are installed at depths of 0.3, 0.9 and 1.5 m. The amount of seepage water leaving the bottom of the lysimeter is measured with a tipping bucket and then collected in a storage container from which water samples can be taken for chemical analysis.

The vessel is placed in a circular pit with a stainless steel casing. To prevent rainfall and other kinds of precipitation as well as dirt and dust from getting into the ~2 cm wide cleavage between vessel and pit casing, a rubber collar can be installed across the cleavage. It is attached to the casing and extends about 1 to 2 cm into the vessel (Fig. 3.2). It is designed not to touch the vessel at all, but in practice it does at a few points.

For weighing, the lysimeter vessel rests on three shear stress cells placed on top of aluminium pedestals. Each cell is laid out for a load capacity of 1,320 kg, which allows a total mass of 3,960 kg. Depending on the water content of the soil, the total mass of the vessel usually ranges from 3,500 to 3,850 kg.

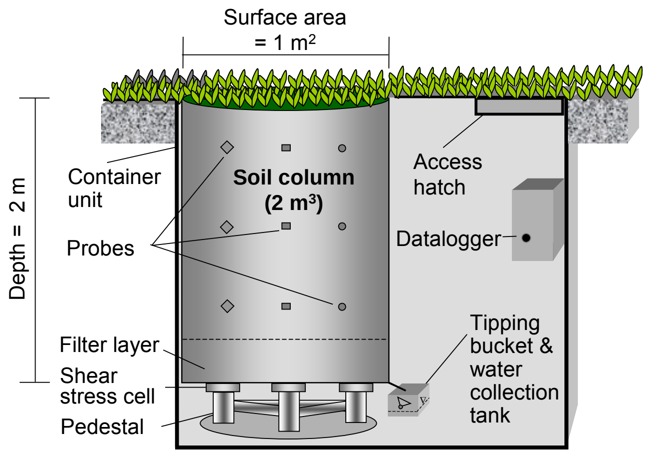


Figure 3.1: Schematic of the weighable gravitation lysimeter investigated in this study (● TDR, ■ Tensiometer/Thermometer, ◆ Suction cup).

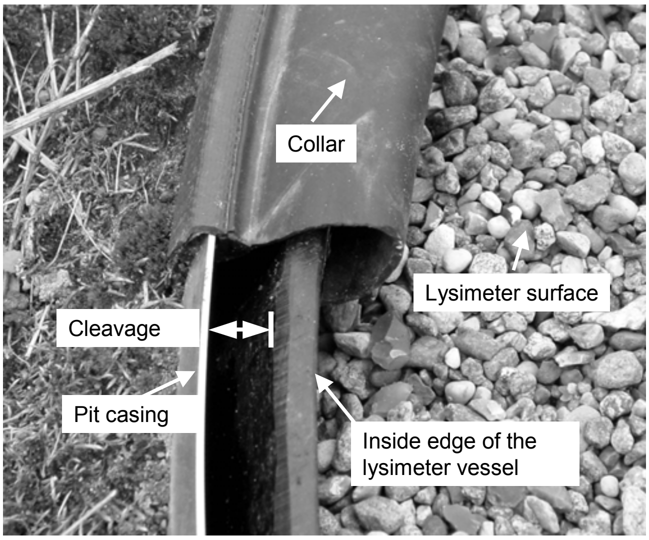


Figure 3.2: Detail of the collar across the cleavage between lysimeter vessel and pit casing.

The shear stress cells produce a current, whose magnitude depends on the load. The current is then transformed into a digital signal using an A/D-converter, which is adjusted to record the mass of the lysimeter vessel to the nearest 10 g. All measurements are stored in a data-logger, whose recording interval is chosen by the user. For our routine work the weight is registered every 10 seconds and the data then aggregated for recording as a 10-minute mean value.

For the study here a lysimeter without any vegetation was chosen in order to eliminate water loss by transpiration. To reduce evaporation from the soil, the surface was covered with a 4 cm thick gravel layer.

3.3.2 Testing the weighing precision of the lysimeter

All experiments were conducted during periods without precipitation and seepage from the soil profile, with little or no wind, small temperature variations over the measurement period, and low evaporative demand.

3.3.2.1 Smallest detectable mass change and measurement stability

As mentioned above, the weighing system here is arranged to record the lysimeter mass to the nearest 10 g. However, this does not imply a weighing precision of 10 g.

To find the smallest mass change the system can reliably detect, weights of 500, 200, 100, 50, 20 and 10 g were subsequently placed at the centre of the lysimeter for 22 minutes and then removed again. The data logger was set to read the lysimeter weight every 10 seconds and to store a mean weight for one minute intervals. The values for the first and 22nd minute were omitted in the later analysis to ensure that disturbances due to the placement and removal of the weights did not influence the results. The order in which the weights were applied was randomly selected. Before and after the trial with a given weight, the lysimeter mass without any weight applied was established for a five minute period, with the first and last minute again discarded, to obtain a reference mass.

This experiment was conducted twice: once without the aforementioned collar, which can easily be removed and attached again, and once with it.

3.3.2.2 Effect of load position

To investigate, if the smallest mass change the lysimeter can reliably detect depends on the position it occurs at, the aforementioned weights of 500, 200, 100, 50, 20 and 10 g were subsequently placed at 10, 23, 55, 77 and 100 cm along two perpendicular lines through the centre of the lysimeter (positions 1 to 9, Figure 3.3).

On all of these nine positions each weight was placed five times in succession for 3 minutes and then removed again. The data logger was again set to read the lysimeter weight every 10 seconds and to store a mean mass for one minute intervals. Only stored values for the 2nd minute of each 3 minute test period were used in the later analysis to ensure that disturbances due to the placement and removal of the load did not affect the results. Before and after the measurements with a given weight at a given position, a reference mass was obtained as described above. The weights and positions were selected in random order, but then retained for the five repeats at all nine points.

This experiment was also conducted without and with collar.

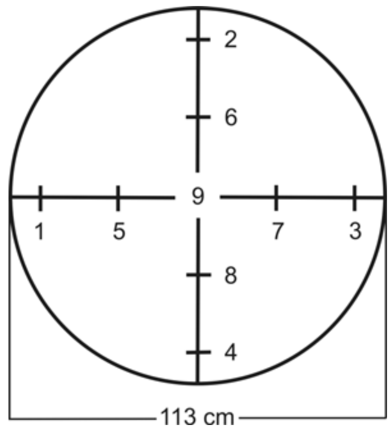


Figure 3.3: Location of the nine investigated load positions on the lysimeter vessel.

3.4 Results

3.4.1 Smallest detectable mass change and measurement stability

Figure 3.4 shows the recorded mass change over 20 minutes after the addition of different weights for the lysimeter without collar. For added weights from 500 to 20 g there are only two to four brief (1 minute) and small (10 g, except once 20 g) deviations from the true value over the 20 minute period. (Recall that the weighing system is adjusted to measure the mass to the nearest 10 g. Hence, deviations can only be registered in multiples of 10 g.) For the 10 g weight, there are considerably more deviations (8) of up to 30 g. This indicates that, without collar, mass changes as small as 20 g can be measured with good accuracy and stability. For the 1 m² cross-sectional area of our lysimeter this is equivalent to a water gain or loss of 0.02 mm.

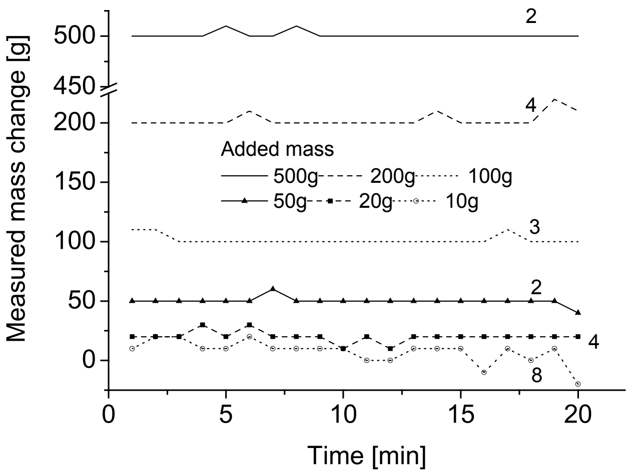


Figure 3.4: Recorded mass change over a 20 minute period after the addition of different weights for the lysimeter without collar. The values near the lines indicate the number of deviations from the true mass change.

This result is confirmed by Figure 3.5, where the mean measured mass for 20 measurements of 1 min each is related to the added mass for each of the 6 applied loads. There is a very good agreement between mean measured and added mass, as demonstrated by the close fit to the 1:1-line and the low standard deviations (≤ 5 g, except for the 10 g weight; see last column in Table 3.2).

The data from this experiment also indicate that the lysimeter can even detect a 10 g mass change under favourable conditions. However, the reading is not stable (Fig. 3.4), most likely for reasons discussed below.

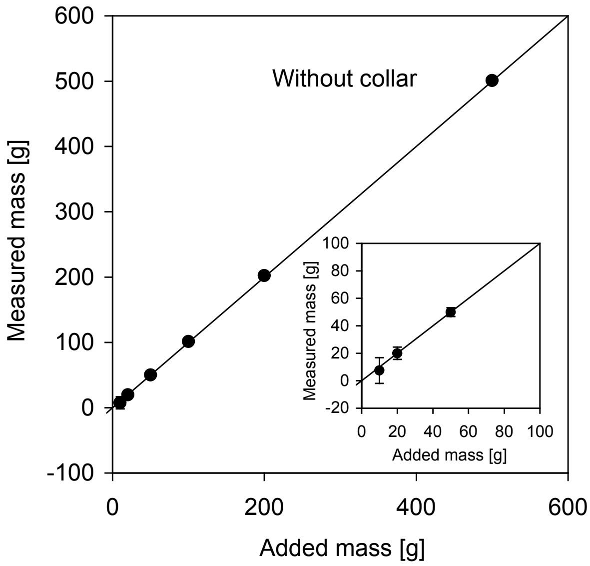


Figure 3.5: Measured versus added mass for the lysimeter without collar. Each point represents a mean of 20 measurements of 1 min each. The vertical lines are the standard deviations. If no standard deviation can be seen, it is smaller than the diameter of the symbol. The diagonal is the 1:1-line.

When they occurred, deviations from the added mass were the same (mostly 10 g) for all loads down to 20 g. This implies that measurement stability does not depend on the mass added. We do not know the exact causes for these deviations. However, wind exerts forces on a lysimeter and can thus alter its apparent weight. Also, the amount of current produced by a shear stress cell at a given load is slightly temperature-dependent. Hence, one or a combination of these factors is the most likely cause of the observed deviations. Since this experiment was carried out under conditions of low wind and low temperature variations, the deviations attributed to these factors were small in absolute terms. Yet, while a 10 g deviation from 500 g only amounts to a 2 % departure, the same deviation from 20 g represents a 50 % departure. This means that the relative weighing accuracy decreases as the added load gets smaller. However, as stated above, measurement stability does not depend on the added mass, at least for values ≥ 20 g.

For the lysimeter with collar (Fig. 3.6) the results are nowhere near as good. There are many more and greater deviations from the true value. In fact, it was hardly ever recorded. The measured values also fluctuate considerably. These statements hold for all tested weights from 500 to 10 g. The means for the 20 measurements for each weight also deviate markedly from the true value (Fig. 3.7), except for the 200 g weight. The standard deviations (between 16 and 41 g; see last column in Table 3.3) from all means are much greater than in the experiment without collar.

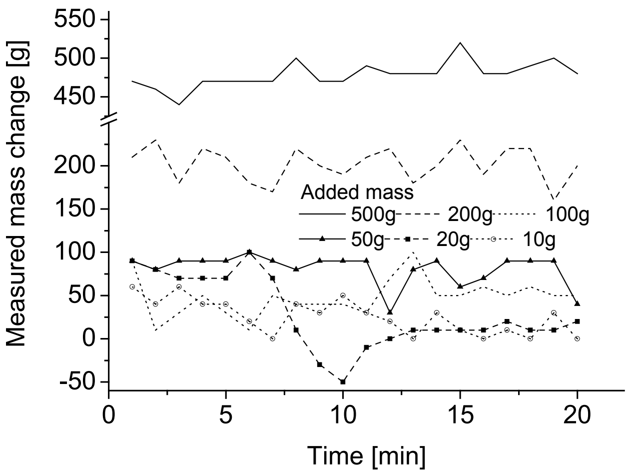


Figure 3.6: Recorded mass change over a 20 minute period after the addition of different weights for the lysimeter with collar.

It follows from these data that the use of the rubber collar seriously lowers weighing precision. The fact that it touches the vessel at a few points apparently exerts forces on the vessel, which distort the true weight. As a result only mass changes of 500 and 200 g were measured with an acceptable accuracy, and then only when averaged over the 20 min observation period, because the stability of the measurements was low.

3.4.2 Effect of load position

In the experiment without collar the mean measured mass varies somewhat between load positions for all added weights, but not in a systematic fashion (Table 3.2). Hence, there is no effect of load position on weighing precision. For most combinations of added mass and position the divergence of the measured from the added value (mostly < 20 g) and the standard deviation from the mean of five measurements at each position (0 - 43 g) are comparatively small.

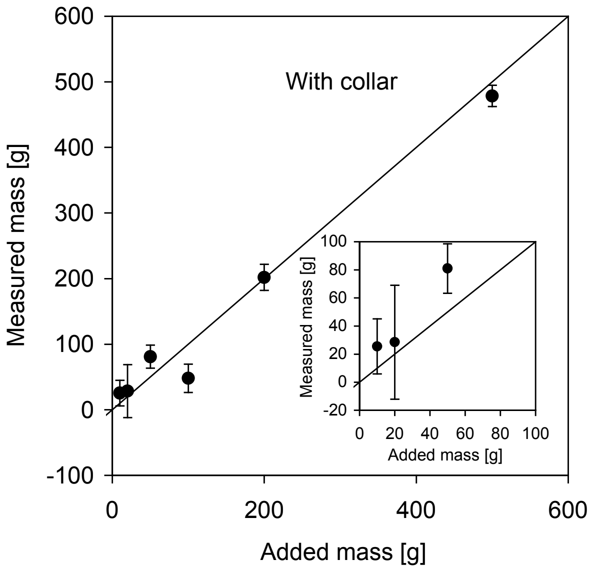


Figure 3.7: Measured versus added mass for the lysimeter with collar. Each point represents a mean of 20 measurements of 1 min each. The vertical lines are the standard deviations. The diagonal is the 1:1-line.

However, while the mean measured mass was about equally accurate for each added weight in both sets of experiments (cv. last two columns in Table 3.2), individual discrepancies and the standard deviations were much larger in the load position tests, most likely because winds were stronger and gustier than during the first set of experiments.

Table 3.2: Mean and standard deviation of the measured mass for various added weights as a function of load position (1 to 9) for the lysimeter without collar. For comparison the mean and standard deviation of the 20 measurements of 1 min each from the firsts experiment are given in the last column.

	Mean measured mass (g) ± standard deviation (g)										
Added mass (g)	1	2	3	4	5	6	7	8	9	All positions	20 min trial
500	512 ± 15	506 ± 12	520 ± 11	508 ± 19	470 ± 28	498 ± 7	502 ± 12	492 ± 4	503 ± 4	501 ± 20	501 ± 3
200	192 ± 4	192 ± 4	198 ± 4	208 ± 4	205 ± 5	206 ± 8	192 ± 4	190 ± 0	198 ± 4	198 ± 8	202 ± 5
100	184 ± 24	88 ± 6	77 ± 15	99 ± 5	112 ± 11	106 ± 9	90 ± 33	120 ± 11	95 ± 12	108 ± 34	102 ± 4
50	68 ± 19	74 ± 12	33 ± 9	34 ± 2	48 ± 16	85 ± 14	44 ± 17	64 ± 5	49 ± 14	55 ± 22	50 ± 3
20	3 ± 16	13 ± 12	68 ± 18	37 ± 12	6 ± 10	7 ± 5	14 ± 6	16 ± 10	27 ± 43	14 ± 31	20 ± 4
10	12 ± 10	24 ± 5	12 ± 12	4 ± 8	8 ± 4	14 ± 5	8 ± 4	8 ± 12	10 ± 11	11 ± 10	8 ± 9

Independent of the added mass the mean measured mass differs much more between load positions in the experiment with collar than in the one without. Again, these differences are not systematic (Table 3.3), which means that here, too, load position has no influence on weighing precision. The divergence from the added mass (frequently > 50 g) and the standard deviation from the mean of five measurements at each position (8 - 76) are rather large for quite a few combinations of added mass and position, when the collar is employed.

Table 3.3: Mean and standard deviation of the measured mass for various added weights as a function of load position (1 to 9) for the lysimeter with collar. For comparison the mean and standard deviation of the 20 measurements of 1 min each from the firsts experiment are given in the last column.

	Mean measured mass (g) ± standard deviation (g)										
Added mass (g)	1	2	3	4	5	6	7	8	9	All positions	20 min trial
500	504 ± 14	478 ± 12	478 ± 17	523 ± 8	625 ± 40	460 ± 16	498 ± 16	486 ± 22	478 ± 25	501 ± 48	478 ± 16
200	128 ± 17	162 ± 13	156 ± 35	358 ± 39	232 ± 46	228 ± 18	226 ± 14	224 ± 26	216 ± 53	214 ± 70	202 ± 20
100	112 ± 10	126 ± 8	120 ± 14	126 ± 8	120 ± 14	74 ± 19	86 ± 17	92 ± 26	110 ± 25	107 ± 25	48 ± 22
50	56 ± 14	40 ± 23	-22 ± 19	116 ± 39	92 ± 44	16 ± 36	50 ± 18	54 ± 58	32 ± 50	48 ± 53	81 ± 18
20	72 ± 29	36 ± 19	2 ± 20	30 ± 19	18 ± 21	102 ± 72	-34 ± 21	40 ± 54	28 ± 29	33 ± 52	28 ± 41
10	-12 ± 50	30 ± 23	66 ± 22	46 ± 62	-18 ± 76	-30 ± 28	-10 ± 13	36 ± 21	24 ± 29	15 ± 52	26 ± 20

3.5 Discussion

Under favourable environmental conditions and with no rubber collar attached, mass changes down to 20 g can be discerned with good accuracy and stability. This weighing precision, which is equivalent to 0.02 mm water gain or loss, is as good as the best values reported for other lysimeters (Table 3.1). However, since wind and temperature gradients affect the weight measurement, weighing precision decreases with increasing wind speed or temperature variation. The former is exemplified in Figure 3.8, which shows that deviations from the true mass are much greater during periods with higher wind speeds than during periods with lower ones. In addition, there seems to be a threshold wind speed of about 1 m/s for our lysimeter, below which wind has no effect on weight measurements. We have no suitable data to present a comparable figure for the effect of temperature.

The absolute deviations from the true weight caused by the above environmental parameters were the same for all added loads. Hence, in relative terms (% deviation from the true value) weighing accuracy becomes better the bigger the true weight change is.

The best results in testing weighing precision can be expected in the late evening hours, when wind speed is usually low and temperature changes are more gradual. In addition, temperature variations are generally smaller on cloudy days. Good results can also be achieved in the morning, but only after water, which condensed on the vessel during the night, has evaporated again.

There is always some evaporation from a lysimeter filled with moist soil. Testing should therefore be done in cooler periods such as late fall, winter or early spring, when potential evaporation is lowest. In theory it would be best to test the weighing precision before the lysimeter is filled with soil, to eliminate evaporation altogether. In practice this may not be feasible, however, since the expected gross weight of the soil-filled lysimeter would then have to be simulated temporarily with some non-evaporating material such as rocks.

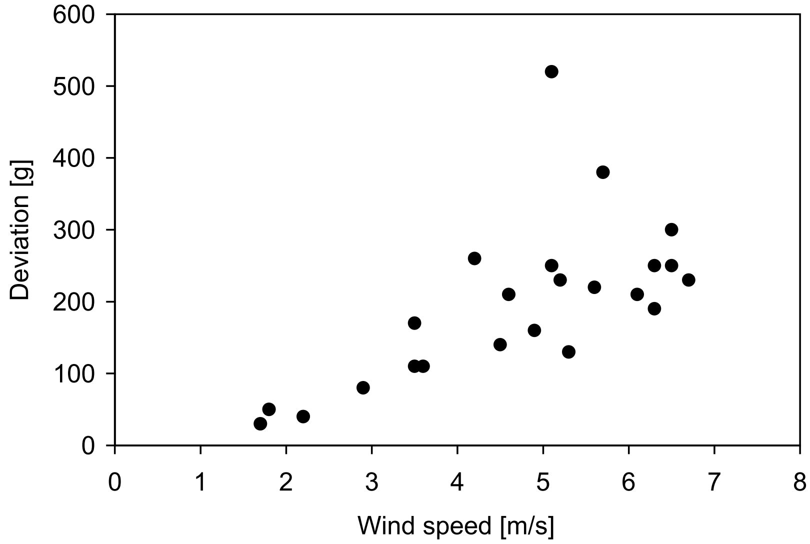


Figure 3.8: Maximum deviation from the true mass as a function of peak wind speed during one hour periods over the course of one specific day.

In nights when dewfall occurs in central Europe, it typically amounts to 0.02 to 1 mm (Hofmann, 1952; Burrage, 1972; Jacobs et al., 1990; Hughes and Brimblecombe, 1994; Meissner et al., 2007). For the 1 m² surface area of our lysimeter this translates into 20 to 1000 g. These amounts are large enough to be reliably detected with our lysimeter set-up. In addition, dewfall usually only occurs during periods of low to no winds, which eliminates one environmental influence on weighing precision.

3.6 Conclusions

With the investigated lysimeter, mass changes down to 20 g can be measured with good accuracy and stability under favourable environmental conditions.

Weighing precision does not depend on the position on the lysimeter, where a load change occurs, no matter if the rubber collar is attached or removed.

The type of rubber collar used here seriously lowers weighing precision in terms of accuracy and stability of the measurements, because it touches the lysimeter vessel at a few points. This exerts forces on the vessel, which distort the true weight. Hence, this type of collar should not be used. A type of collar needs to be developed, which does not influence the weighing process.

3.7 References

Allen, R. G., Fisher, D. K. (1990): Low-cost electronic weighing lysimeters. Trans. ASAE 33, 1823-1833.

Burrage, S. W. (1972): Dew on wheat. Agric. Meteorol. 10, 3-12.

Fritschen, L. J., Cox, L., Kinerson, R. (1973): A 28-meter douglas fir in a weighing lysimeter. Forest Sci. 19, 256-261.

Gavilan, P., Berengena, J. (2007): Accuracy of the Bowen ratio-energy balance method for measuring latent heat flux in a semiarid advective environment. Irr. Sci. 25, 127-140.

Gholam, A. B., Mohammad, J. K. (2002): A large electronic weighing lysimeter system: design and installation. J. Am. Water Resour. Ass. 38, 1053-1060.

Girona, J., Marsal, J., Mata, M., del Campo, J. (2004): Pear crop coefficients obtained in a large weighing lysimeter. In: Proceedings of the IVth International Symposium on Irrigation of Horticultural Crops (pp. 277-282). ISHS Acta Horticulturae 664.

Hofmann, G. (1952): Zur Methodik der Taumessung. Ber. deutsch. Wetterd. US-Zone 38, 360-364.

Howell, T. A., McCormick, R. L., Phene, C. J. (1985): Design and installation of large weighing lysimeters. Trans. ASAE 28, 106-117.

Howell, T. A., Schneider, A. D., Jensen, M. E. (1991) : History of lysimeter design and use for evapotranspiration measurements. In: Proceedings of the International Symposium on Lysimetry, July 23-25: Lysimeters for evapotranspiration and environmental measurements (pp. 1-9). Honolulu, Hawaii.

Hughes, R. N., Brimblecombe, P. (1994): Dew and guttation: formation and environmental significance. Agric. Forest Meteorol. 67, 173-190.

Hunsaker, D. J., Pinter, P. J. jr., Cai, H. (2002): Alfalfa basal crop coefficients for FAO-56 procedures in the desert regions of the southwestern U. S. Trans. ASAE 45, 1799-1815.

Jacobs, A. F. G., van Pul, W. A. J., van Duken, A. (1990): Similarity moisture dew profiles within a corn canopy. J. Appl. Meteorol. 29, 1300-1306.

Jia, X., Dukes, M., Jacobs, D. J. M., Irmak, S. (2006): Weighing lysimeter for evapotranspiration research in humid environment. Trans. ASABE 49, 401-412.

Kirkham, R. R., Gee, G. W., Jones, T. L. (1984): Weighing lysimeters for long-term water balance investigations at remote sites. Soil Sci. Soc. Am. J. 48, 1203-1205.

Malone, R. W., Bonta, J. V., Stewardson, D. J., Nelsen,T. (2000): Error analysis and quality improvement of the Coshocton weighing lysimeters. Trans. ASAE 43, 271-280.

Marek, T. H., Schneider, A. D., Howell, T. A., Ebeling, L. L. (1988): Design and construction of large weighing monolithic lysimeters. Trans. ASAE 31, 477-484.

Marek, T. H., Piccinni, G., Schneider, A. D., Howell, T. A., Jett, M., Dusek, D. A. (2006): Weighing lysimeters for the determination of crop water requirements and crop coefficients. Appl. Eng. Agric. 22, 851-856.

McFarland, M. J., Worthington, J. W., Newman, J. S. (1983): Design, installation and operation of a twin weighing lysimeter for fruit trees. Trans. ASAE 26, 1717-1721.

Meissner, R., Seeger, J., Rupp, H., Seyfarth, M., Borg, H. (2007): Measurement of dew, fog, and rime with a high-precision gravitation lysimeter. J. Plant Nutr. Soil Sci. 170, 335-344.

Meshkat, M., Warner, R. C., Walton, L. R. (1999): Lysimeter design, construction, and instrumentation for assessing evaporation from a large undisturbed soil monolith. Appl. Eng. Agric. 14, 303-308.

Plauborg, F. (1995): Evaporation from bare soil in a temperate humid climate - measurement using micro-lysimeters and time domain reflectometry. Agric. Forest Meteorol. 76, 1-17.

Pruitt, W. O., Angus, D. E. (1960): Large weighing lysimeter for measuring evapotranspiration. Trans. ASAE 3, 13-18.

Qiu, G. Y., Momii, K., Yano. T. (1996): Estimation of plant transpiration by imitation leaf temperature. I: Theoretical consideration and field verification. Trans. Japan. Soc. Irr. Drain. 64, 401-410.

Rupp, H., Meissner, R., Leinweber, P., Lennartz, B., Seyfarth, M. (2007): Design and operability of a large weighable fen lysimeter. Water Air Soil Poll. 186, 323-335.

Sayler, M. D., Allen, M. D., Burman, R. D., Smith, J. L. (1985): Design and installation of a weighable lysimeter. Final Report (Project No. 2-92384), submitted to the Wyoming Water Research Center, University of Wyoming, Laramie.

Schneider, A. D., Howell, T. A., Steiner, J. L. (1993): An evapotranspiration research facility using monolithic lysimeters from three soils. Appl. Eng. Agric. 9, 227-235.

Schneider, A. D., Howell, T. A., Moustafa, A. T. A., Evett, S. R., Abou-Zeid, W. (1998): A simplified weighing lysimeter for monolithic or reconstructed soils. Appl. Eng. Agric. 14, 267-273.

Tyagi, N. K., Sharma, D. K., Luthra, S. K. (2003): Determination of evapotranspiration for maize and berseem clover. Irr. Sci. 21, 173-181.

Unold, G. (2003): Anforderungen an die Sickerwasserbeprobung. BAL-Workshopbericht Sickerwassersammler vom 8. und 9. April 2002. UMS GmbH, Munich.

Yang, S. L., Aydin, M., Yano, T., Li, X. (2003): Evapotranspiration of orange trees in greenhouse lysimeters. Irr. Sci. 21, 145-149.

Yoder, R. E., Odhiambo, L. O., Wright, W. C. (2005): Evaluation of methods for estimating daily reference crop evapotranspiration at a site in the humid southeast United States. Appl. Eng. Agric. 21, 197-202.

Young, M. H., Wierenga, P. J., Mancino, C. F. (1997): Monitoring near-surface soil water storage in turfgrass using time domain reflectometry and weighing lysimetry. Soil Sci. Soc. Am. J. 61, 1138-1146.

Zhang, Y. Q., Yu, Q., Liu, C., Jiang, J., Zhang, X. (2004): Estimation of winter wheat evapotranspiration under water stress with two semi-empirical approaches. Agron. J. 96, 159-168.

4. Effect of vegetation type and growth stage on dewfall, determined with high precision weighing lysimeters at a site in northern Germany

4.1 Abstract

The amount and temporal distribution of dewfall on grass, maize and winter barley was measured with four high precision weighing lysimeters at a site in northern Germany during 2004 and 2005 to quantify the contribution of dewfall to the water balance of the region, and to assess how dewfall is affected by the vegetation cover.

Two lysimeters were under continuous grass, two were cropped (maize from April through September 2004, followed by winter barley until July 2005, fallow the rest of the time). Observed dewfall ranged from 27.1 to 31.8 mm per year, which was 5.5 to 6.9% of the annual rainfall. In several months of the study period dewfall was > 20% of the monthly precipitation.

On fallow lysimeters there were fewer nights with dewfall and less dewfall per event than on lysimeters with grass. After crops were planted the number of dewfall-nights and the amount of dewfall per event rose quickly and eventually surpassed that on the lysimeters with grass. After harvest both parameters dropped well below the values on the grass lysimeters again.

4.2 Introduction

Dew forms when the temperature of a surface (e.g. soil or a leaf) falls below the dew point temperature of the surrounding air so that vapour condenses on the surface. If the condensed water originated from the soil below, this is referred to as dewrise or distillation; if it originated from the ambient air, this is referred to as dewfall or condensation (Monteith, 1957). Only the latter represents a real input into the water balance, the former is merely a redistribution of water already in the system. The focus of this paper is on dewfall.

The amount of dewfall is determined by how much and for how long the temperature of a surface falls below the dew point temperature of the surrounding air, the moisture content of the air, and the ventilation (Monteith, 1957; Garratt and Segal, 1988; Camuffo and Giorio, 2003). It depends on the properties of a surface, too, and is therefore affected by the type of vegetation and its stage of development.

In semiarid and arid regions dewfall can reach or even exceed all other forms of precipitation for extended periods or even a whole year (e.g. Evenari et al., 1971; Kalthoff et al., 2006). In humid regions dewfall normally contributes only a small percentage to the total annual precipitation, but it may be the biggest component over shorter periods such as a week or a month (Tuller and Chilton, 1973). There is ample evidence that even small amounts of dewfall can be beneficial to plants, not only in arid and semiarid, but also in humid regions (e.g.

Hiltner, 1930; Went, 1955; Duvdevani, 1964; Baier, 1966; Wallin, 1967; Kerr and Beardsell, 1975; Kappen et al., 1980).

Although there is a long history of research on dew (in our paper this term always refers to the sum of dewrise and dewfall), there is no universal procedure for its measurement (Richards, 2004). Because of the usually small amount, it cannot be measured with standard rain gauges. Hence, various types of dew gauges have been developed (e.g Leick, 1932; Kessler, 1939; Duvdevani, 1947; Hirst, 1954, 1957; Lloyd, 1961; Nagel, 1962). They employ artificial surfaces like asbestos, gypsum, silicaceous earth, aluminium, artificial (plastic) grass mats or treated wood for dew to form on. Since these surfaces have different thermal properties, different gauges yield different dew amounts (e.g. Gelbke, 1955; Wallin, 1963; Kidron et al., 2000), which further depend on the height above the soil or plant surface a gauge is placed at (e.g. Leick, 1932; Fritzsche, 1934; Duvdevani, 1947, 1964; Evenari et al., 1971). The properties of the gauges differ from those of plants, too, so that the amount of dew deposited on them differs as well.

Furthermore, in a plant canopy leaves are spread across a range of heights, which leads to a complex distribution of surfaces for dew to accumulate on as well as a complex thermal regime within the canopy and, thus, complex conditions for dew formation. The leaf distribution changes as the plants grow, too. Therefore, the amount of dew recorded by a dew gauge cannot be expected to be the same as the amount deposited on a canopy. Also, the ratio of the amount deposited on a dew gauge and on a canopy is not constant, even if the plant surface does not change with time (Hofmann, 1955). Nevertheless, dew gauges are useful for comparative measurements between sites, provided the same type of gauge is employed and at the same height. The quantification of dew amounts is currently not part of the routine monitoring programme of the German National Meteorological Service (Deutscher Wetterdienst - DWD). The weather services in most countries do not routinely record dew amounts either.

For our purposes the main drawbacks of dew gauges are that they record dewrise plus dewfall, while only the latter is of interest in this study, and that they do not correctly reflect the amount of dew deposited on a canopy. In contrast, a weighing lysimeter only registers dewfall, since dewrise comes from within its confines and therefore does not cause a mass change. This makes it a suitable tool for measuring dewfall as demonstrated by Meissner et al. (2007), provided the weighing precision is high enough. Also, the type of vegetation of interest can be planted on a lysimeter, which allows for a direct measurement of dewfall on the chosen vegetation. (As pointed out by Hofmann (1955) some evaporation may occur even as dew forms. Strictly speaking a lysimeter therefore records net dewfall.)

Unfortunately, high precision weighing lysimeters are expensive, which precludes their widespread use. As a result there are only few studies which use them to investigate dewfall (see Table 4.3 in the results section). None of them look at a whole year or different types of vegetation. Hence, the objectives of this study are to use lysimeters 1) to quantify the amount and temporal distribution of dewfall over an extended period (2 years), and 2) to assess the influence of vegetation type and growth stage on dewfall.

4.3 Materials and methods

The study was carried out at the Falkenberg lysimeter station of the Department of Soil Physics of the Helmholtz Centre for Environmental Research - UFZ, which is located in northern Germany some 120 km northwest of Berlin. The site is 21 m above sea level, its mean annual precipitation is 588 mm with a maximum in July (69 mm) and a minimum in February (29 mm). Its potential annual evapotranspiration is 565 mm (Meissner et al., 1999), also with a maximum in July (106 mm) and a minimum in February (8 mm). The surrounding area is plain and mainly under grassland.

Four identical weighing lysimeters were employed in this study. Each has a surface area of 1 m^2, a depth of 2 m and is filled with a sandy soil. A detailed description of these lysimeters is given by Meissner et al. (2007). They can discern mass changes as small as 30 g, which for their 1 m^2 surface area corresponds to a depth of 0.03 mm of water. Their mass is recorded every 10 minutes.

The lysimeters are arranged in a rectangular pattern. To compare dewfall on different vegetation the two eastern lysimeters (No. 209 and 210) were planted with maize from April through September 2004 (they were bare prior to that), and with barley thereafter until July 2005. For the remainder of the year they were kept fallow. The western two (No. 211 and 212) were under continuous grass in both years. Table 4.1 contains further information on the vegetation cover. Note that the grass was mowed six times during the observation period to keep its height (5 - 25 cm) and density (85 - 100% ground cover) variation within a reasonably narrow range.

Dewfall results in a mass increase of a lysimeter. Hence, to identify dewfall the lysimeter records were surveyed for periods with mass increases. Since a mass increase may also result from rain or snow, periods with a mass increase were compared with the precipitation data collected at the site by a continuously recording tipping-bucket rain gauge. Mass increases not concurrent with rain or snow were finally classified as dewfall.

Table 4.1: The vegetation cover and its management on the four lysimeters during the study period.

Date	Lysimeter 209 and 210	Lysimeter 211 and 212
2004		
Jan 1	fallow	continuous grass
Apr 27	planting of maize	
May 24		1st mowing
Aug 2		2nd mowing
Sept 21	harvesting of maize	
Oct 6	planting of winter barley	3rd mowing
2005		
June 8		4th mowing
July 13	harvesting of winter barley, fallow thereafter	
Aug 18	tilling	
Aug 23		5th mowing
Oct 5		6th mowing

4.4 Results

The results of our study are summarised in Table 4.2. All values given in this table and in the following figures are means for two lysimeters (No. 209/210 and No. 211/212, respectively).

4.4.1 Number of nights with dewfall

As can be deduced from Table 4.2, dewfall occurred on 43% to 53% of the nights in a year and on 10% to 87% of the nights in a month.

Figure 4.1 depicts the time course of the number of nights with dewfall during the study period. In 2004 there were 36 more dewfall-nights on the grass lysimeters than on those with crops (Table 4.2). This arose from a much higher number of dewfall-nights on the grass lysimeters from March through June as well as in October and November. In 2005 there were also more dewfall-nights on grass, but the difference (13) was smaller. It was mostly due to more dewfall-nights on grass from September through November.

In both years the number of dewfall-nights on the lysimeters with grass showed two peaks, one in spring and one in autumn. This is a reflection of the meteorological conditions at the site, which are especially conducive to dewfall during these two seasons. In 2005 this was also observed on the cropped lysimeters, while in 2004 there was only one peak in spring.

Table 4.2: Number of nights with dewfall and amount of dewfall per dewfall-night and per month in 2004 and 2005. Each value represents the mean for two lysimeters (No. 209 and 210 for crops, No. 211 and 212 for grass).

Year	Month	Rainfall	No. of nights with dewfall		Mean dewfall per dewfall-night (mm)		Dewfall per month (mm)		Dewfall (% of rainfall)	
			crops	grass	crops	grass	crops	grass	crops	grass
2004	Jan	65.3	5	6	0.11	0.15	0.55	0.87	0.8	1.3
	Feb	17.5	5	5	0.15	0.19	0.76	0.93	4.3	5.3
	Mar	11.4	10	14	0.11	0.15	1.11	2.13	9.7	18.7
	Apr	37.5	13	22	0.09	0.19	1.23	4.20	3.3	11.2
	May	37.0	3	11	0.10	0.14	0.29	1.50	0.8	4.1
	Jun	57.1	9	13	0.08	0.11	0.68	1.49	1.2	2.6
	Jul	90.8	18	16	0.17	0.14	2.99	2.30	3.3	2.5
	Aug	62.2	19	17	0.20	0.12	3.80	1.97	6.1	3.2
	Sep	0.1	18	17	0.31	0.16	5.50	2.67	> 100.0	> 100.0
	Oct	11.0	17	26	0.13	0.19	2.24	5.05	20.4	45.9
	Nov	50.7	17	26	0.13	0.19	2.28	4.99	4.5	9.8
	Dec	21.4	23	20	0.25	0.18	5.66	3.65	26.5	17.1
	Sum	**462.0**	**157**	**193**			**27.09**	**31.75**	**5.9**	**6.9**
2005	Jan	61.5	14	13	0.17	0.19	2.36	2.46	3.8	4.0
	Feb	44.4	4	4	0.18	0.20	0.72	0.79	1.6	1.8
	Mar	32.7	6	4	0.16	0.23	0.97	0.94	3.0	2.9
	Apr	10.7	16	10	0.26	0.12	4.14	1.19	38.7	11.1
	May	68.3	21	24	0.26	0.15	5.54	3.63	8.1	5.3
	Jun	18.1	14	16	0.11	0.12	1.57	1.86	8.7	10.3
	Jul	98.1	13	12	0.11	0.15	1.49	1.75	1.5	1.8
	Aug	37.8	15	16	0.14	0.15	2.14	2.33	5.7	6.2
	Sep	49.7	18	23	0.12	0.13	2.13	3.04	4.3	6.1
	Oct	30.3	20	27	0.14	0.15	2.78	4.02	9.2	13.3
	Nov	30.1	19	22	0.15	0.21	2.94	4.66	9.8	15.5
	Dec	46.9	13	15	0.16	0.21	2.13	3.18	4.5	6.8
	Sum	**528.6**	**173**	**186**			**28.91**	**29.85**	**5.5**	**5.7**

Mowing did not obviously affect the time course of the number of dewfall-nights. However, this statement must be viewed with care, since there was no unmowed control to compare the data to.

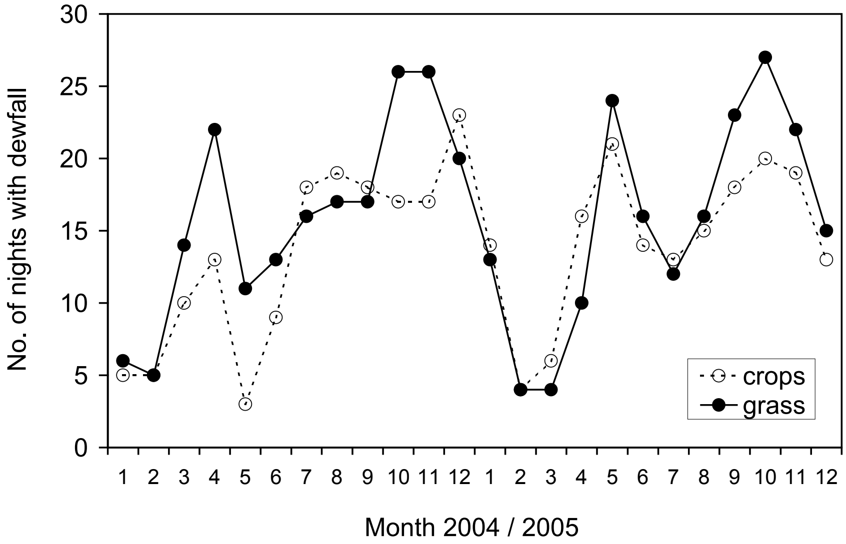


Figure 4.1: Number of nights with dewfall on the lysimeters with crops and with grass in 2004 and 2005.

4.4.2 Amount of dewfall

The maximum amount of dewfall we observed in one night was 0.6 mm, which is not far from the theoretical maximum of 0.7 to 0.8 mm/night (Monteith, 1957; Garratt and Segal, 1988). It occurred only once, but values around 0.5 mm/night five times. The mean for all nights with dewfall on grass or crops was 0.17 mm. On a monthly basis mean dewfall rates per dewfall-night varied between 0.08 and 0.31 mm (Table 4.2). Total dewfall per month ranged from 0.29 to 5.66 mm with a mean of 2.45 mm. Yearly dewfall was 27.1 to 31.8 mm.

Figure 4.2 displays the time course of the monthly means of the amount of dewfall per dewfall-night in 2004 and 2005. In both years dewfall per dewfall-night on grass did not vary as much as on crops. From July to September and in December 2004 as well as in April and May 2005 it was higher on crops than on grass. At all other times it was higher on grass.

Figure 4.3 shows the time course of the amount of dewfall per month in 2004 and 2005, which is the product of the number of dewfall-nights per month (Fig. 4.1) and the mean amount of dewfall per dewfall-night in a given month (Fig. 4.2). For the grass lysimeters the time course of the monthly dewfall is largely congruent with that of the number of dewfall-nights, because the amount of dewfall per dewfall-night did not change very much over the study period. Again, there was a peak in spring and autumn in both years, and mowing didn't influence the monthly dewfall in any apparent way.

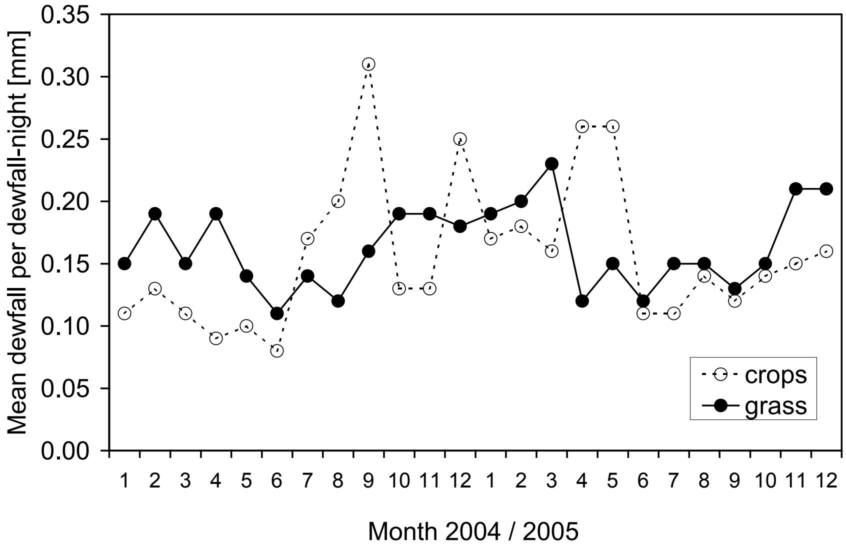


Figure 4.2: Monthly means of the amount of dewfall per dewfall-night on the lysimeters with crops and with grass in 2004 and 2005.

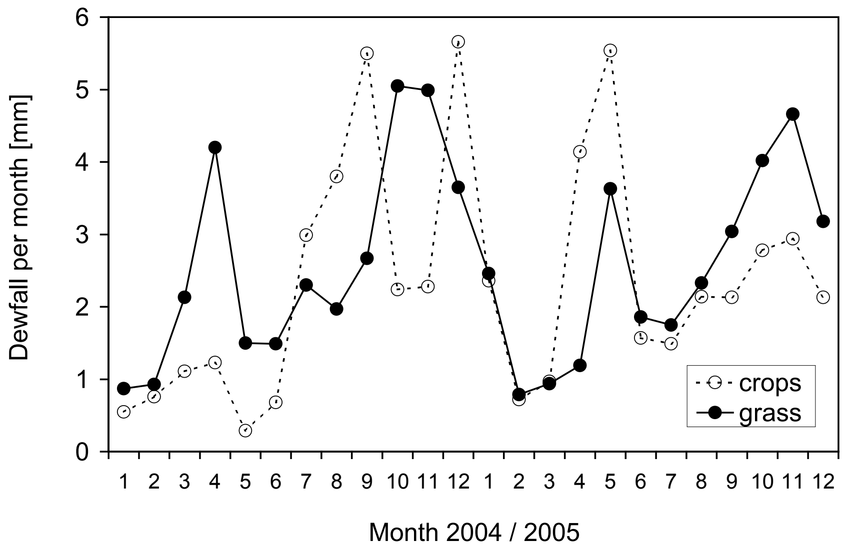


Figure 4.3: Amount of dewfall per month on the lysimeters with crops and with grass in 2004 and 2005.

For the crop lysimeters the monthly dewfall rose steeply from a minimum in May until September 2004 and then fell sharply. After a second rise from November to December it dropped again to an annual minimum in February 2005 from where it increased once more until May, followed by another decline. In July to September and December 2004 as well as in April and May 2005 there was more dewfall on crops than on grass. In all other months it was less than on grass.

As can be deduced from Table 4.2, in the whole of 2004 there was markedly more dewfall on the grass lysimeters than on the cropped ones (31.8 mm compared to 27.1 mm), while in 2005 the figures were similar (29.9 and 28.9 mm, respectively). However, in the period when maize was growing (April 27 to September 21, 2004) there was more dewfall on the cropped (11.8 mm) than on the grass lysimeters (9.7 mm). In the period when winter barley was growing (October 6, 2004 to July 13, 2005) there was also more dewfall on the cropped (25.7 mm) than on the grass lysimeters (24.3 mm).

4.5 Discussion

To put our dewfall values into perspective Table 4.3 shows the amount of dew recorded in different studies at various locations. Note that these figures comprise dewrise plus dewfall, unless they were obtained with lysimeters, in which case they state dewfall only. Also, in some of these studies rime or moisture deposited by fog were also counted as dew, even though they are not. This was the case in our study, too. In the autumn of both years there were some mornings with a few hours of fog. The occurrence and duration of fog, but not the amount of moisture deposited by it, are recorded at a DWD weather station 6.5 km northwest of the lysimeter station. Because it was rare and of short duration, we did not attempt to separate the amount of moisture deposited by fog from dewfall.

On nights with dew the mean amount in Table 4.3 varies from 0.03 to 0.24 mm, while the highest nightly values are 0.54 to 0.61 mm. The mean and peak nightly dewfall in our study (0.17 mm and 0.5 - 0.6 mm) are within these ranges. Studies, which follow dew formation over a whole year or longer, are rare. They reveal an annual dew amount of 0.5 to 48 mm. Our dewfall data (27 - 32 mm) lie within this range.

The differences in the values from the various studies are at least partly due to the different measurement techniques employed. Some methods (e.g. blotting paper, weighing of leaves) record dewrise, dewfall and guttation of sap on leaves, those using artificial surfaces (e.g. Leick dew plates, Duvdevani blocks) dewrise and -fall, lysimeters net dewfall. Also, the height of the instrument above the soil or crop surface affects the reading. Hence, different results even for the same environmental conditions can be expected, as already pointed out in the introduction.

Site conditions also affect dew formation. The exposition to the sun and wind leads to significant differences over short distances (e.g. Fritzsche, 1934; Steubing, 1951; Went, 1955; Kappen et al., 1980; Jacobs et al., 2000; Kidron et al., 2000). In the study of Raman et al. (1973) dew measurements were carried out with the same experimental set-up, but at different sites in India. From these data one may conclude that dew amounts are higher in more humid areas. However, this is not universally true. The proximity to water bodies, which can

Table 4.3: Amount of dew reported in various studies (n.a. = no information given).

Source	Observed period	Location	Method	Vegetation	Dew amount nightly range (mm)	nightly mean (mm)	yearly value (mm)
von Kerner (1892)	summer 1891	central Tyrol, Austria	Kappeller drosometer	n.a.	n.a. - 0.54	0.20	
Pickering (1913)	May 1912 - Apr 1913	Jamaica	iron plate	n.a.	n.a. - 0.40	0.12	34
Parchinger (1918)	36/41 days in 1917	2 sites in present day southwestern Ukraine	blotting paper	lawn	n.a. - 0.17	0.06 - 0.10	
Leick (1932)	a total of 15 days in Sept - Dec 1931	Greifswald, Germany	Leick dew plate	lawn	0.02 - 0.22	0.12	
Keller (1933)	Jun - Jul 1931	Transvaal, South Africa	blotting paper	pasture		0.08	
Fritzsche (1934)	Jul - Sept 1933	Hiddensee, Germany	Leick dew plate	lawn	0.06 - 0.39	0.13	
Duvdevani (1947)	1937 - 1944	northern Palestine	Duvdevani block	n.a.			26 - 33
Mäde (1954)	Sept 1949 - Aug 1953	Etzdorf, Germany	Kessler-Fuess dew recorder	lawn			8 - 11
Monteith (1957)	a total of 4 days in Aug, Sept, Oct 1953 and 1 day in Apr 1954	Harlington, England	blotting paper	lawn	0.03 - 0.13	0.09	
Duvdevani (1964)	1943 - 1947	13 stations in Palestine	Duvdevani block	n.a.			5 - 31
Aslyng (1965)	Apr - Nov in 1961 -1964	Seeland, Denmark	lysimeter	lawn	n.a. - 0.30		
Baier (1966)	1957 - 1958	Highveld region, South Africa	Kessler-Fuess dew recorder	lawn	0.05 - 0.40	0.08	13
Rosenberg (1969)	a total of 18 nights in Aug, Nov, Dec 1966	Mead, Nebraska	lysimeter	bare soil	0.03 - 0.42	0.18	
Burrage (1972)	Jul 1963	Ashford, England	blotting paper	wheat	0.02 - 0.33	0.16	
Evenari et al. (1971)	1963 - 1966	Negev, Israel	Duvdevani block	desert	n.a. - 0.35	0.17	26 - 37
Raman et al. (1973)	Oct - Mar in 1969 - 1970 (no dew Apr - Sept)	46 stations in India	Duvdevani block	n.a.	0.02 - 0.30	0.03 - 0.24	0.5 - 30
Tuller and Chilton (1973)	Jun, Jul, Aug, Oct 1970	Victoria, British Columbia, Canada	Duvdevani block	broom/short grass boundary	n.a. - 0.20	0.15	

Table 4.3: continued

Source	Observed period	Location	Method	Vegetation	Dew amount nightly range (mm)	nightly mean (mm)	yearly value (mm)
Sharma (1976)	Mar - Nov 1974 (no dew Dec - Feb)	New South Wales, Australia	lysimeter	grass	0.03 - 0.56	0.17	13
Hicks (1983)	44 days in late winter 1967	New South Wales, Australia	energy balance	n.a.	0.03 - 0.61	0.22	
Jacobs et al. (1990)	several days in Jun, Jul 1988	central Netherlands	Leick dew plate	maize	0.05 - 0.27	0.17	
Rönsch (1990)	Apr - Oct in 1977 - 1986	Harzgerode, Germany	Kessler-Fuess dew recorder	lawn	0.11 - 0.30	0.20	
Hughes and Brimblecombe (1994)	Apr - Nov 1985	Potter Heigham, England	filter paper	lawn	n.a. - 0.27	0.14	
Sudmeyer et al. (1994)	a total of 5 days in Jul - Oct 1989 and 2 days in Jul 1990	southwest Western Australia	weighing of small soil monoliths	pasture	0.002 - 0.10	0.03	
Wilson et al. (1999)	1 night each in Jul, Aug 1992 and Jul, Aug 1994	Wisconsin, USA	removal and weighing of leaves	potato	0.10 - 0.47		
Malek et al. (1999)	Oct 1993 - Sep 1994	north-eastern Nevada	Bowen ratio	desert shrubs			14
	Jan - Dec 1996	central Utah, USA	Bowen ratio	irrigated alfalfa			29
Jacobs et al. (2000)	5 weeks in Sep, Oct 1997	Negev, Israel	weighing of small soil monoliths	sand dune transect	0.08 - 0.31	0.15	
Kidron et al. (2000)	late summer and fall 1992	Negev, Israel	cloth covered glass plate	desert transect	0.07 - 0.31	0.23	
Luo and Goudriaan (2000)	a total of 14 nights in Feb, Mar, Apr 1994	Los Baños, Philippines	blotting paper	rice	0.11 - 0.21	0.15	
Jacobs et al. (2006)	1994 - 2004	central Netherlands	energy budget	grass			26 - 48
Kalthoff et al. (2006)	Dec 1999 - Dec 2003, Nov 2004	Elqui valley, Chile	Bowen ratio, Eddy correlation	desert vegetation, irrigated crops	0.01 - 0.25		5 - 10
Beysens et al. (2005)	Jan 2001 - Jan 2002	Ajaccio, Corsica, Fra.	plexiglass plate	n.a		0.07	8
	Jan 2002 - Jan 2003	Bordeaux, France	plexiglass plate	n.a.		0.05	10
	Jun 2000 - Jun 2001	Genoble, France	plexiglass plate	n.a.		0.04	4

serve as a source of moist air, is important, too. For instance, due to moist air moving in from the Mediterranean Sea, as much dew was observed in the Negev desert (Israel) as in humid Jamaica (Table 4.3). Hence, whether a site is classified as humid or arid allows no general conclusions about dew amounts.

The aim of all studies mentioned in Table 4.3 was to assess the amount of dew deposited in a given natural environment. There is also a large body of literature on the collection of dew to augment water supplies using artificial surfaces designed to maximise dew formation (e.g. Muselli et al., 2009). This is a very different objective. We therefore did not include any data from this literature in the table. However, the range of the nightly as well as the daily dew amounts observed in these studies is comparable to that in the studies in Table 4.3.

Our data indicate that the occurrence of dewfall is not only connected to meteorological conditions, but also to the type of vegetation and its growth stage. Before planting maize at the end of April 2004, the lysimeters 209 and 210 were fallow and there were fewer occurrences of dewfall (Fig. 4.1) and less dewfall (Figs. 4.2 and 4.3) than on the lysimeters 211 and 212, which were always under grass. This was still the case in May and June, while the maize plants were still small. As the maize grew the number of dewfall-nights and dewfall amounts rose quickly and by July had surpassed that on the lysimeters with grass. They remained above the values for the grass trials during August, when the maize reached its maximum height (~ 200 cm) and ground cover (> 95%), and September, but after harvest on September 21 fell far below them during October and November, when the barley planted on these lysimeters on October 6 was still in its early stages of development.

Figure 4.4 depicts the amount of nightly dewfall on maize in relation to plant height. Since the amount of dewfall depends on a number of factors (e.g. atmospheric conditions, day length), one cannot expect a perfect relationship. Nevertheless, the figure shows a clear tendency for nightly dewfall to increase with plant height. This arises for the following reasons. Soils have a greater heat content and a higher thermal conductivity than plants leaves. Hence, at night it takes longer for a soil surface than for a leaf to cool down (Lehmann and Schanderl, 1942) and a soil surface is usually warmer than a leaf. As a result, there is less dewfall on a bare soil than on plants. On soils covered with plants there is also a transfer of heat from the warmer soil to the cooler leaves, which reduces leaf cooling and, thus, dewfall on the leaves. As the plants grow in height the canopy becomes denser, too. Less sunlight now reaches and heats the soil during the day. Therefore, less heat is transferred from the soil to the leaves at night and the reduction of dewfall due to heat from the soil becomes smaller. A taller and denser canopy also means that this heat is distributed over a larger volume so that a given leaf receives less of it now.

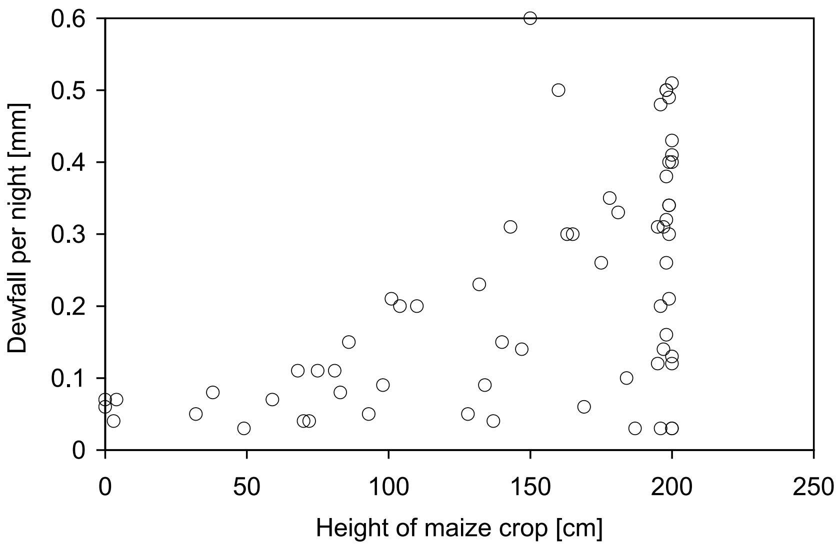


Figure 4.4: Amount of nightly dewfall on maize in relation to plant height.

For barley there was no tendency for the amount of nightly dewfall to increase with plant height. Furthermore, the pattern in the number of dewfall-nights was similar to that for grass. Yet, dewfall amounts followed the reasoning put forth for maize, though not as unequivocally. In December 2004 dewfall was much higher on barley than on grass, even though the barley had about the same height, but was less dense than the grass. The state of the vegetation did not change much in the next three month. However, in line with its lower density, there was now less dewfall on barley than on grass. In April and May the height of the barley crop surpassed that of the grass, as did the amount of dewfall. At the end of May the crop reached its maximum ground cover (> 95%). In June the barley reached maturity (and its maximum height of 80 - 90 cm) and turned yellow. While the height of the crop remained the same, its density decreased, which led to a drop in dewfall below the amount on grass. After harvesting the barley on July 13 the two lysimeters remained bare and dew amounts (and the number of dew events) stayed below the values on the neighbouring grass lysimeters.

Other studies also point to an effect of vegetation cover. Fritzsche (1934) and Duvdevani (1964) observed less dew over bare soil than over grass, which agrees with our observations. Tuller and Chilton (1973) recorded less dew at a forest edge and at the centre of a broom area than at a broom/grass boundary. As in our trials, Keller (1933) reported more dew, if the vegetation was denser, and Sudmeyer et al. (1994) found that the amount of dewfall increased with plant height and biomass. Leaf characteristics also play a role. For example, more dew forms on horizontal than on vertical leaves (Lehmann and Schanderl, 1942).

Currently dewfall is generally not considered in water balance studies in Germany, because the amounts are assumed to be very small. In our study dewfall on crops and grass reached

5.9% and 6.9% of the annual rainfall in 2004, while the figures for 2005 were 5.5% and 5.7%. In our opinion these values are too high to be ignored in the water balance. Furthermore, yearly figures mask the fact that dewfall can be a much bigger input during dry periods. For example, on the grass lysimeters in our experiments dewfall was equivalent to 46% of rainfall in October 2004. On the barley lysimeters it was 39% in April 2005. In September 2004 it was in essence the sole water input to all our lysimeters.

To put our data into perspective again, here are some values from other studies. At a different site in Germany dew amounted to just 1.9% of the annual rainfall (Mäde, 1954). Some other reported figures are 2% for dew in the Highveld region of South Africa (Baier, 1966), 2.5% for dewfall in southeastern Australia (Sharma, 1976), 5 to 10% for dew in the Elqui valley in Chile (Kalthoff et al., 2006), and 48% for dew at Advat in the Negev desert in Israel (Evenari et al., 1971). From these data one may conclude that dew(fall) is no significant input into the water balance in humid and semiarid regions, but can be an important one in arid regions. However, in these studies, too, dew was a much bigger input during dry periods. For example, in the data of Mäde (1954) dew surpassed 25% of the rainfall in some months, in the study of Sharma (1976) dewfall reached 30% of it in June. Also, in British Columbia Tuller and Chilton (1973) found that dew usually amounts to 12 - 14% of the monthly precipitation, but up to 154% of it during dry spells. This underlines that dew(fall) can be an important component of the water balance in humid and semiarid regions, too, at least during periods of low rainfall.

Still, in arid regions of the world dew is often a much more significant or sometimes even the dominant source of precipitation, at least for some periods. For instance, at Advat in the Negev desert (Israel) only 26 mm of rain were recorded in 1962/63, but 28 mm of dew (Evenari et al., 1971). Similarly, Kalthoff et al. (2006) noted that in the arid Elqui valley in Chile annual dew formation (5 - 10 mm) is in the same order of magnitude as the annual precipitation in dry years.

Nevertheless, in most environments the absolute amounts of water contributed by dewfall are small compared to the potential transpiration of plants and, consequently, not sufficient to supply the transpirational needs of a plant for very long. Dewfall alone is therefore not enough to prevent water stress. However, under otherwise equal conditions water-stressed plants, which receive dew, have higher leaf water potentials (Kerr and Beardsell, 1975), at least during the morning hours, grow better (Hiltner, 1930; Steubing, 1955; Duvdevani, 1964) and survive longer (Stone and Fowells, 1954) than plants deprived of dew. Even small amounts of dew can be sufficient to re-hydrate dry leaves to the point, where death by desiccation is averted. As a result, plants, which received dew, are more likely to be still alive and able to resume growth, once sufficient water becomes available again.

4.6 Conclusions

Our results indicate that dewfall makes a notable contribution to the water balance of crops and grass in northern Germany, since it amounted to 5.5 - 6.9% of the annual and, several times during the study period, to > 20% of the monthly precipitation.

The study also illustrates that the vegetation cover affects dew formation. There was consistently more dewfall on covered than on bare lysimeters. In addition, dewfall increased with crop growth, reflected in the rising frequency and amount of dewfall on growing crops compared to a continuous grass cover, and then fell again after harvest.

We gave a qualitative explanation, how vegetation and its state of development affects dew formation. In a future study we intend to collect data on net radiation, air, leaf temperature and soil temperatures, and ventilation, in addition to dewfall amounts, to provide a more quantitative description.

4.7 References

Aslyng, H.C., 1965. Rain, snow and dew measurements. Acta Agric. Scand. 15, 275-283.

Baier, W., 1966. Studies on dew formation under semi-arid conditions. Agric. Meteorol. 3, 103-112.

Beysens, D., Muselli, M., Nikolayev, V., Narhe, R., Milimouk, I., 2005. Measurement and modelling of dew in island, coastal and alpine areas. Atmos. Res. 73, 1-22.

Burrage, S.W., 1972. Dew on wheat. Agric. Meteorol. 10, 3-12.

Camuffo, D., Giorio, R., 2003. Quantitative evaluation of water deposited by dew on monuments. Bound.-Layer Meteorol. 107, 655-672.

Duvdevani, S., 1947. An optical method of dew estimation. Q. J. R. Meteorol. Soc. 73, 282-296.

Duvdevani, S., 1964. Dew in Israel and its effect on plants. Soil Sci. 98, 14-21.

Evenari, M., Shanan, L., Tadmor, N., 1971. The Negev. The challenge of a desert. Harvard University Press, Cambridge, Massachusetts.

Fritzsche, G., 1934. Vergleichende Messungen mit Leickschen Tauplatten. Bioklim. Beibl. Meteorol. Z. 1, 66-73.

Garratt, J.R., Segal M., 1988. On the contribution of atmospheric moisture to dew formation. Bound.-Layer Meteorol. 45, 209-236.

Gelbke, W., 1955. Untersuchungen zur Methodik der Taumessung anhand dreijähriger Taumessreihen in Greifswald. Abh. Meteorol. Hydrol. Dienst DDR 5, 40-47.

Hicks, B.B., 1983. A study of dewfall in an arid region: an analysis of Wangara data. Q. J. R. Meteorol. Soc. 109, 900-904.

Hiltner, E., 1930. Der Tau und seine Bedeutung für den Pflanzenbau. Eine Theorie über die physiologische Bedeutung der Wasseraufnahme durch oberirdische Pflanzenorgane. Wiss. Arch. Landw., Abt. A, 3, 1-70.

Hirst, J.M., 1954. A method for recording the formation and persistence of water deposits on plants shoots. Q. J. R. Meteorol. Soc. 80, 227-231.

Hirst, J.M., 1957. A simplified surface wetness recorder. Plant Path. 6, 57-61.

Hofmann, G., 1955. Die Thermodynamik der Taubildung. Ber. Deut. Wetterd. 3, 1-45.

Hughes, R.N., Brimblecombe, P., 1994. Dew and guttation: formation and environmental significance. Agric. For. Meteorol. 67, 173-190.

Jacobs, A.F.G., van Pul, W.A.J., van Dijken, A., 1990. Similarity moisture dew profiles within a corn canopy. J. Appl. Meteorol. 29, 1300-1306.

Jacobs, A.F.G., Heusinkveld, B.G., Berkowicz, S.M., 2000. Dew measurements along a longitudinal sand dune transect, Negev Desert, Israel. Int. J. Biometeorol. 43, 184-190.

Jacobs, A.F.G., Heusinkveld, B.G., Kruit, R.J.W., Berkowicz, S.M., 2006. Contribution of dew to the water budget of a grassland area in the Netherlands. Water Resour. Res. 42, W03415.

Kalthoff, N., Fiebig-Wittmaack, M., Meißner, C., Kohler, M., Uriarte, M., Bischoff-Gauß, I., Gonzales, E., 2006. The energy balance, evapo-transpiration and nocturnal dew deposition of an arid valley in the Andes. J. Arid Environ. 65, 420-443.

Kappen, L., Lange, O.L., Schulze, E.-D., Buschbom, U., Evenari, M., 1980. Ecophysiological investigations on lichens of the Negev Desert. VII. The influence of the habitat exposure on dew imbibition and photosynthetic productivity. Flora 169, 216-229.

Keller, H., 1933. Über Taumessungen im ariden Hochland Transvaals. Meteorol. Z. 50, 321-324.

Kerr, J.P., Beardsell, M.F., 1975. Effect of dew on leaf water potentials and crop resistances in a Paspalum pasture. Agron. J. 67, 596-599.

Kessler, O.W., 1939. Der Tauschreiber Kessler-Fuess. Bioklim. Beibl. Meteorol. Z. 6, 23-26.

Kidron, G.J., Yair, A., Danin, A., 2000. Dew variability within a small arid drainage basin in the Negev Highlands, Israel. Q. J. R. Meteorol. Soc. 126, 63-80.

Lehmann, P., Schanderl, H., 1942. Tau und Reif, Pflanzenwetterkundliche Untersuchungen. Deutsches Reich, Reichsamt für Wetterdienst, Wissenschaftliche Abhandlungen, Bd. IX, Nr. 4. Springer, Berlin.

Leick, E., 1932. Zur Methodik der relativen Taumessung. Bot. Cent.bl. Beih. 49, 160-189.

Lloyd, M., 1961. The contribution of dew to the summer water budget of northern Idaho. Bull. Am. Meteorol. Soc. 42, 572-580.

Luo, W.H., Goudriaan, J., 2000. Dew Formation on rice under varying durations of nocturnal radiative loss. Agric. For. Meteorol. 104, 303-313.

Mäde, A., 1954. Über die Höhe des nächtlichen Taufalles im mitteldeutschen Trockengebiet. Wiss. Z. Luther Univ. Halle, Math.-Naturw. 4, 185-190.

Malek, E., McCurdy, G., Giles, B., 1999. Dew contribution to the annual water balances in semi-arid desert valleys. J. Arid Environ. 42, 71-80.

Meissner, R., Seeger, J., Rupp, H., Schonert, P., 1999. Estimating the effects of set-aside on water quality: scaling-up of lysimeter studies. Land Degrad. Develop. 10, 13-20.

Meissner, R., Rupp, H., Seeger, J., Seyfarth, M., Borg, H., 2007. Measurement of dew, fog and rime with a high-precision gravitation lysimeter. J. Plant Nutr. Soil Sci. 170, 335-344.

Monteith, J.L., 1957. Dew. Q. J. R. Meteorol. Soc. 83, 322-341.

Muselli, M., Beysens, D., Mileta, M., Milimouk, I., 2009. Dew and rain water collection in the Dalmatian Coast, Croatia. Atmos. Res. 92, 455-463.

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

Parchinger, V., 1918. Eine neue Taumessungsmethode. Meteorol. Z. 35, 47-48.

Pickering, W.H., 1913. A year's record with a dew recorder at Mandeville, Jamaica. Q. J. R. Meteorol. Soc. 39, 326-327.

Raman, C.R.V., Venkataraman, S., Krishnamurthy, V., 1973. Dew over India and its contribution to winter-crop water balance. Agric. Meteorol. 11, 17-35.

Richards, K., 2004. Observation and simulation of dew in rural and urban environments. Prog. Phys. Geogr. 28, 76-94.

Rönsch, H., 1990. Tau und Reif in Harzgerode. Z. Meteorol. 40, 197-204.

Rosenberg, N.J., 1969. Evaporation and condensation on bare soil under irrigation in the East Central Great Plains. Agron. J. 61, 557-561.

Sharma, M.L., 1976. Contribution of dew to the hydrologic balance of a semi-arid grassland. Agric. Meteorol. 17, 321-331.

Steubing, L., 1951. Der Einfluß von Heckenanlagen auf den Taufall. Ber. Deut. Wetterd. U.S. Zone 32, 53-56.

Steubing, L., 1955. Studien über den Taufall als Vegetationsfaktor. Ber. Deut. Bot. Ges. 68, 55-70.

Stone, E.C., Fowells, H.A., 1954. Drought survival of ponderosa: pine seedlings treated with simulated dew survive by month non-treated controls in greenhouse tests. Calif. Agric. 8, 9.

Sudmeyer, R.A., Nulsen, R.A., Scott, W.D., 1994. Measured dewfall and potential condensation on grazed pasture in the Collie River Basin, southwestern Australia. J. Hydrol. 154, 255-269.

Tuller, S.E., Chilton, R., 1973. The role of dew in the seasonal moisture balance of a summer-dry climate. Agric. Meteorol. 11, 135-142.

von Kerner, F., 1892. Ein Instrument zur Messung des Tauniederschlages. Meteorol. Z. 9, 106-108.

Wallin, J.R., 1963. Dew, its significance and measurement in phytopathology. Phytopathology 53, 1210-1216.

Wallin, J.R., 1967. Agrometeorological aspects of dew. Agric. Meteorol. 4, 85-102.

Went, F.W., 1955. Fog, mist, dew and other sources of water. In: Stefferud, A. (ed.), Water. The yearbook of agriculture 1955 (pp. 103-109). U.S. Dept. of Agriculture, Washington, D.C.

Wilson, T.B., Bland, W.L., Norman, J.M., 1999. Measurement and simulation of dew accumulation and drying in a potato canopy. Agric. For. Meteorol. 93, 111-119.

5. Assessment of four methods to compute dewfall

5.1 Abstract

High precision weighing lysimeters are an effective tool to quantify dewfall, but they are not wide-spread due to their high cost. One alternative is to compute dewfall from meteorological data and under consideration of the properties of the surface in question. Four equations, which were shown in the literature to work for this purpose, are assessed in this chapter.

Three of them, the energy balance (EB), turbulent vapour transport (TVT) and Penman-Monteith (PM) equation, contain a heat and/or vapour conductance term. To get a correct value for it requires a wind profile in equilibrium with the vegetation under investigation. This was apparently not the case under the conditions at the site in this study. Hence, the EB and TVT equation could not be used successfully without adjusting the conductance term. The PM equation was less beset by this problem, because the conductance term in it is small at high relative humidities common during dew events. The Bowen ratio energy balance (BREB) equation was found to work best, because it lacks a conductance term.

The BREB, EB and TVT equation need the temperature of the surface, which is usually not available. This leaves the PM equation, from which it has been eliminated, as the only option then.

5.2 Introduction

Dew plays an important role in agriculture. This is discussed in more detail in chapter 2 and in the literature (e.g. Hiltner, 1930; Went, 1955; Duvdevani, 1964; Baier, 1966; Wallin, 1967; Evenari et al., 1971; Tuller and Chilton, 1973; Kerr and Beardsell, 1975; Kappen et al., 1980; Kalthoff et al., 2006). Dew amounts can be measured with various dew gauges, but there is still no standard way of measurement. Lysimeters are a very promising tool to quantify dewfall (Meissner et al., 2007). They have a high precision and can directly measure a mass change due to vapour condensation on a surface (dewfall). However, they are not readily available at every study site because they are expensive. So, one often needs to consider other ways to determine dewfall. One possibility is to compute it from meteorological data and under consideration of the properties of the surface in question.

Neumann (1956), Monteith (1957, 1963) and Long (1958) indicated that dewfall represents a flux of latent heat towards the surface, the opposite of evaporation. It is therefore possible to compute the amount of dew formation with methods developed for computing evaporation. Based on these methods there is a large body of literature on ways to compute dew formation from meteorological data. The most widely used methods to estimate dew amounts are the energy balance equation, the equation for turbulent vapour transport, the Penman-Monteith equation, and the Bowen ratio energy balance equation.

As dewfall represents the latent heat flux at a surface, dew amounts can be estimated with the energy budget for night-time heat flux within an atmosphere - canopy systems, if the other components of the energy balance are available either from measurement or calculation (e.g. Pedro and Gillespie, 1982a, b; Severini et al., 1984; Janssen and Römer, 1991; Madeira et al., 2001). Dewfall estimated in this way is generally in good agreement with measurements.

Neumann (1956) derived an equation for the turbulent transport of water vapour. Using routine meteorological data it can estimate dewfall on grass or on bare surfaces. Dewfall estimated with this equation was close to dewfall measured by a dew gauge at the same station.

The Penman-Monteith equation is also used to estimate the dew formation (Sentelhas and Gillespie, 2008). In this equation transport of energy and vapour within the atmosphere is taken into account. It was applied to studies on dew formation over various forms of vegetation and in different regions (e.g. Garratt and Segal, 1988; Sudmeyer, et al. 1994; Luo and Goudriaan, 2000; Jacobs et al. 1996, 1999, 2000, 2006, 2008). These studies found the Penman-Monteith equation to be a good method to assess dewfall.

Lastly, the Bowen ratio energy balance technique was also applied successfully to determine the above crop vapour flux towards the canopy, i.e. dewfall (e.g. Atzema et al., 1990; Jacobs et al., 1990, 1994).

The four models mentioned above are looked at in this chapter. The objective is to evaluate their performance in computing dewfall. (A comparison between measured and computed measured dewfall is the subject of chapter 6.) The meteorological data necessary for this were collected at the lysimeter station in Falkenberg. The details of the energy balance and the turbulent vapour transport equation were already described in chapter 2. The Penman-Monteith equation and the Bowen ratio energy balance will be derived here.

5.3 Materials and methods

5.3.1 Data

The data for the assessment here were collected on 19 to 21 November, 2009, with the meteorological instruments installed at the Falkenberg lysimeter station of the Department of Soil Physics of the Helmholtz Centre for Environmental Research - UFZ.

Air temperature and relative humidity were measured 2 m above a grass surface. Air temperature was measured with platinum resistance elements (809 LO-100, Wilh. Lambrecht

GmbH, Göttingen, Germany), ventilated in an Assmann psychrometer shield. Relative humidity was measured with a shielded sensor (Mela CPC1/5-ME, MELA Sensortechnik GmbH, Mohlsdorf, Germany) at a height of 2 m, and wind speed with a cup anemometer (Wind Sensor „Meteorology“ 14576-24V, Wilh. Lambrecht GmbH, Göttingen, Germany) at 10 m height at the same location. These measurements were logged automatically as 10 minute averages throughout the day.

To obtain the soil heat flux two temperature sensors (Pt100, Temperaturmeßtechnik Geraberg, Martinroda, Germany) were buried at 5 cm and 10 cm depth, and a thermal conductivity probe (Thermolink, Decagon Devices, Pullman, Washington, USA) at a depth of 7.5 cm. Soil temperature and thermal conductivity were recorded automatically every half-hourly and later employed to compute soil heat flux.

Net radiation was measured with a net-radiometer (NR Lite, Kipp and Zonen, Delft, The Netherlands) 1 m above the 20 cm tall grass. The surface temperature of the grass was measured hourly with a hand-held infrared (IR) thermometer (Rayngier MX4, Raytek Corporation, Santa Cruz, California, United States). The surface emissivity was set to 0.95. To obtain a more reliable value of surface temperature it was measured four times at four different positions. The four positions were situated in the direction of due south, due east, due north, and due west, and about 20 - 30 cm away from edge of the lysimeter (Fig. 5.1). In the following, surface temperature is given as the mean of the values measured at the four positions.

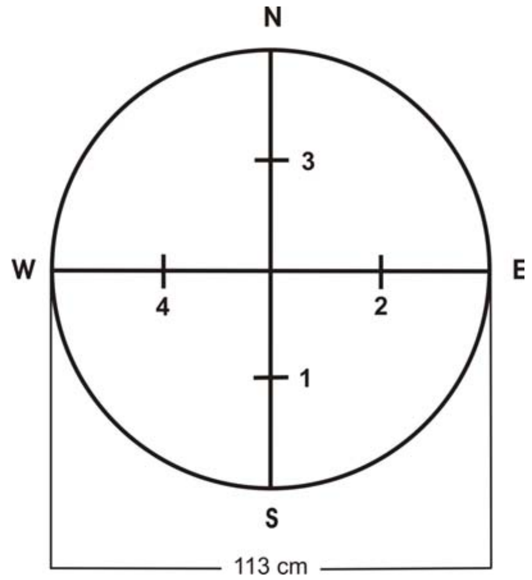


Figure 5.1: The four positions on the lysimeter where surface temperature was measured.

5.3.2 Equations

In the following, all equations are written such that fluxes towards the surface are positive.

5.3.2.1 Energy balance equation (EB)

A rearrangement of the energy balance equation (Eq. 2.2 in chapter 2) yields:

$$\lambda E = -(R_n + H + G) \tag{5.1}$$

This means that the latent heat flux (i.e. dewfall if this flux is directed towards the surface in question) can be computed as the sum of the other three terms in the equation. Directly measured values are employed here for R_n. H is computed on the basis of measured air and surface temperatures, wind speed and characteristics of the vegetation stand. The details were already presented in the section on sensible heat flux in chapter 2. G is calculated from measurements of soil temperature at 5 and 10 cm depth (T_5 and T_{10}) and thermal conductivity of the soil between the two temperature probes (k_T) as:

$$G = k_T \cdot \frac{T_{10} - T_5}{\Delta z} \tag{5.2}$$

where Δz = distance between the two soil temperature probes. Substituting this and Eq. 2.7 and 2.8 into Eq. 5.1 leads to the expression used for the computations here:

$$\lambda E = -(R_n + H + G) = -\left\langle R_n \right\rangle - \left\langle c_p \cdot k^2 \cdot u \cdot \rho_a \cdot \left(\ln \frac{z-d}{z_0} \right)^2 \cdot (T_a - T_s) \right\rangle - \left\langle k_T \cdot \frac{T_{10} - T_5}{\Delta z} \right\rangle \tag{5.3}$$

5.3.2.2 Equation for turbulent vapour transport (TVT)

As just described in the previous section, λE can be computed with the other terms of the energy balance equation. However, if air and surface temperatures, relative humidity, wind speed and characteristics of the vegetation stand are known, as in the case here, then λE can be computed directly with Eqs. 2.8 to 2.11 as:

$$\lambda E = \lambda \cdot g_v \cdot \frac{e_a - e_s}{P} = \lambda \cdot k^2 \cdot u \cdot \rho_a \cdot \left(\ln \frac{z-d}{z_0} \right)^{-2} \cdot \frac{h_r \cdot a \cdot \exp\left(\frac{b \cdot T_a}{T_a + c} \right) - a \cdot \exp\left(\frac{b \cdot T_s}{T_s + c} \right)}{P} \tag{5.4}$$

5.3.2.3 Penman-Monteith equation (PM)

The starting point for the Penman-Monteith equation is the equation for turbulent vapour transport. In a slightly altered form it reads:

$$\lambda E = \lambda \cdot g_v \cdot \frac{e_a - e_s(T_s)}{P} \tag{5.5}$$

where $e_s(T_s)$ is the saturation vapour pressure at the temperature of the surface (T_s). Since T_s is usually unknown, Penman (1948) introduced a way to eliminate it, which was refined later by Monteith (1973). The first step is to add and substract $e_s(T_a)$, i.e. the saturation vapour pressure at air temperature, from the vapour pressure term in Eq. 5.5:

$$\lambda E = \lambda \cdot g_v \cdot \frac{e_a - e_s(T_s) + e_s(T_a) - e_s(T_a)}{P} = \lambda \cdot g_v \cdot \frac{e_a - e_s(T_a)}{P} + \lambda \cdot g_v \cdot \frac{e_s(T_a) - e_s(T_s)}{P} \tag{5.6}$$

Now, the slope of the temperature - saturation vapour pressure curve (s) is given by:

$$s = \frac{\Delta e_s}{\Delta T} = \frac{e_s(T_a) - e_s(T_s)}{T_a - T_s} \tag{5.7}$$

from which follows that $e_s(T_a) - e_s(T_s) = s \cdot (T_a - T_s)$. Using this relationship in Eq. 5.6 leads to:

$$\lambda E = \lambda \cdot g_v \cdot \frac{e_a - e_s(T_a)}{P} + \lambda \cdot g_v \cdot \frac{s \cdot (T_a - T_s)}{P} \tag{5.8}$$

Eq. 2.7 states that $H = c_p \cdot g_H \cdot (T_a - T_s)$. Rearranging it yields:

$$(T_a - T_s) = \frac{H}{c_p \cdot g_H} \tag{5.9}$$

so that after inserting Eq. 5.9 into Eq. 5.8 one gets:

$$\lambda E = \lambda \cdot g_v \cdot \frac{e_a - e_s(T_a)}{P} + \frac{\lambda \cdot g_v \cdot s}{P} \cdot \frac{H}{c_p \cdot g_H} \tag{5.10}$$

As pointed out before, $g_v = g_H$ so that they can be cancelled out in the second term in the next step. The energy balance can also be written as H = - (Rn + G + λE). Application of this relationship in Eq. 5.10 produces:

$$\lambda E = \lambda \cdot g_v \cdot \frac{e_a - e_s(T_a)}{P} - \frac{\lambda \cdot s}{P} \cdot \frac{[R_n + G + \lambda E]}{c_p} \tag{5.11}$$

Finally, solving this expression for λE yields the Penman-Monteith equation:

$$\lambda E = \frac{-s \cdot [R_n + G] + c_p \cdot g_v \cdot [e_a - e_s(T_a)]}{s + P \cdot \frac{c_p}{\lambda}} \tag{5.12}$$

In the Penman-Monteith equation there is usually no minus sign before the (R_n + G) term and the order of e_a and $e_s(T_a)$ is reversed. These differences arise, because here a flux towards the canopy or soil surface was defined as positive. Hence, if there is dewfall λE is positive, and if there is evaporation it is negative.

The value of s can be computed as:

$$s = \frac{b \cdot c \cdot e_s(T)}{(c+T)^2} \tag{5.13}$$

where b and c are the same empirical coefficients as in Eq. 2.10 and 2.11 and T = absolute temperature. A look at Eq. 5.7 reveals that s should be evaluated at the average of air and surface temperature, i.e. at $T = (T_a + T_s)/2$. However, in the application of the PM equation s is evaluated at air temperature, i.e. at $T = T_a$, because the whole point of the PM equation is to get rid of T_s. This introduces a small error into the equation. If the surface temperature is available, using s derived for $T = (T_a + T_s)/2$ is more precise.

5.3.2.4 Bowen ratio energy balance equation (BREB)

The Bowen ratio energy balance equation is a variation of the energy balance equation. Bowen (1926) defined:

$$\beta = \frac{H}{\lambda E} \tag{5.14}$$

The ratio β has since been called Bowen ratio. Reshuffling this equation to $H = \beta \cdot \lambda E$ and substituting it into the energy balance equation yields:

$$0 = Rn + H + G + \lambda E = Rn + \beta \cdot \lambda E + G + \lambda E \tag{5.15}$$

Solving Eq. 5.15 for λE gives:

$$\lambda E = -\frac{R_n + G}{(1+\beta)} \tag{5.16}$$

To evaluate β one can use Eq. 2.7 for H and Eq. 2.9 for λE, which leads to the following expression for the Bowen ratio:

$$\beta = \frac{H}{\lambda E} = \frac{P \cdot c_p \cdot g_H \cdot (T_a - T_s)}{\lambda \cdot g_v \cdot (e_a - e_s)} \tag{5.17}$$

Assuming that $g_H = g_v$, which is commonly accepted, Eq. 5.17 simplifies to:

$$\beta = \frac{P \cdot c_p}{\lambda} \cdot \frac{(T_a - T_s)}{(e_a - e_s)} = \frac{P \cdot c_p}{\lambda} \cdot \left(\frac{\Delta T}{\Delta e} \right) \tag{5.18}$$

5.4 Results

5.4.1 Plausibility of the meteorological data

Calculations can only yield good results, if the input data are good. Hence, the data entering into the equations are assessed first. The best way to do this would be to compare them to independent measurements with a second set of instruments. Such measurements do not exist. Hence, the only way to judge the data is to see if the different variables behave as they should.

Net radiation (R_n), air (T_a), canopy surface (T_s), in-canopy (T_i), and soil temperature at two depth (T_5 and T_{10}) as well as wind speed (u) and relative humidity (h_r) were measured with modern instruments. Soil heat flux (G) was computed from the soil temperatures and soil thermal conductivity. Cloud cover was estimated visually. Figure 5.2 displays the changes in these variables from 18^{30} on 19 November to 13^{00} on 21 November, 2009. Sunrise on these two days was at ~ 7^{45}, sunset at ~ 16^{15}. Following the convention stated above, any flux towards surface is positive, any flux away from it negative.

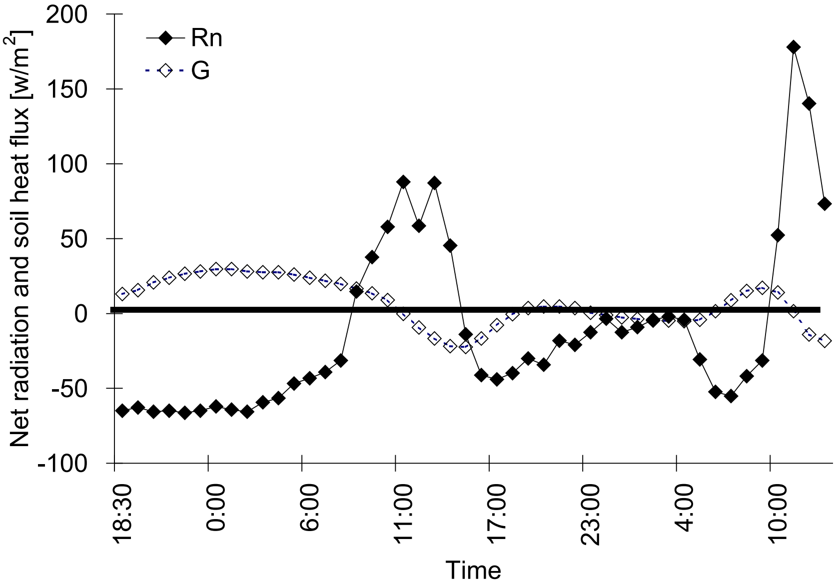
Net radiation was directed away from the surface (negative) during the first night (Fig. 5.2a). It remained more or less constant for most of the night, but then began to increase slowly as sunrise approached. From 8^{00} it increased fairly rapidly towards a peak around midday, from where it decreased again towards the evening. It became positive around 8^{30} and was negative again at 15^{00}. The dip in R_n at midday was caused by a brief rise in cloud cover (Fig. 5.2b).

At 18^{00} on the second night cloud cover increased markedly. In line with this R_n became less negative and approached zero, because the clouds diminished radiation loss. At 4^{00} cloud cover droped back; simultaneously R_n became more negative again. As the sun came up R_n increased steeply until 11^{00} and then fell due to another increase in cloud cover.

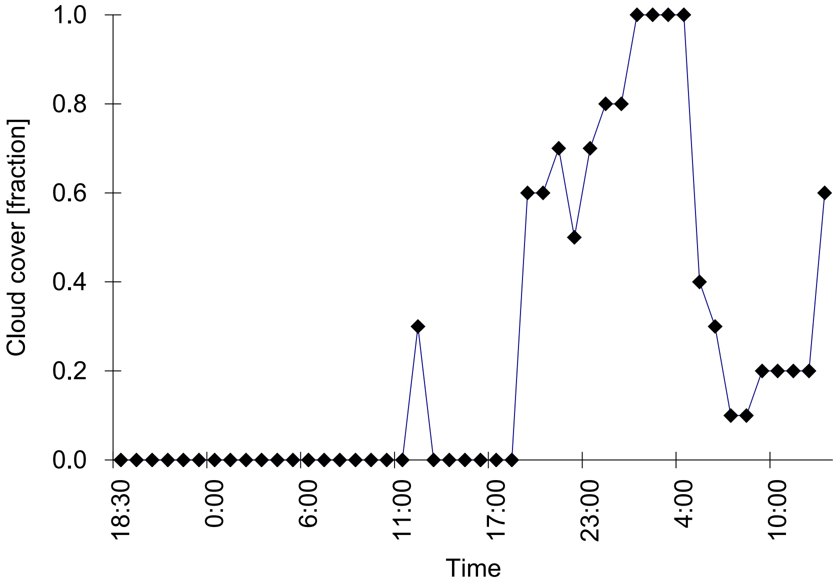
The observed night and day variation of net radiation and its alteration by cloud cover follows the pattern one would expect: radiation loss at night (less if there are clouds), gain by day (less if there are clouds).

Soil heat flux (G) was calculated from measured soil temperatures at 5 (T_5) and 10 cm (T_{10}) depth and soil thermal conductivity. The latter did not vary during the observation period so that G was determined by soil temperature changes (Fig. 5.2c). These in turn are a response to the energy flow across the soil surface. Positive R_n at the surface (or warm air moving across it) puts energy into the soil and leads to an increase in its temperature. Negative R_n or cold air moving across the surface means an energy loss and leads to a decrease in soil temperature. The temperature changes first occur near the surface and then, with a time lag, in greater depths. They are also greater near the surface.

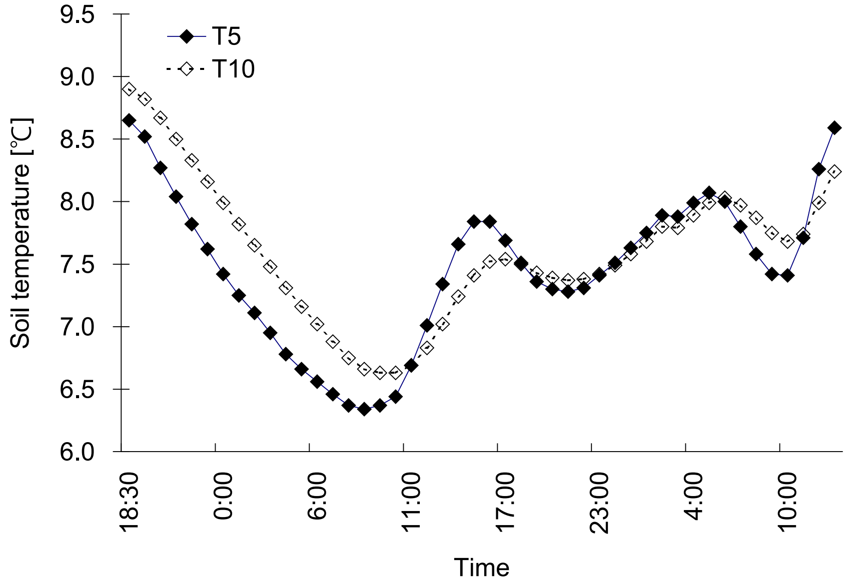
a)


Net radiation and soil heat flux [w/m2]
Rn
G
200
150
100
50
0
-50
-100
18:30
0:00
6:00
11:00
17:00
23:00
4:00
10:00
Time

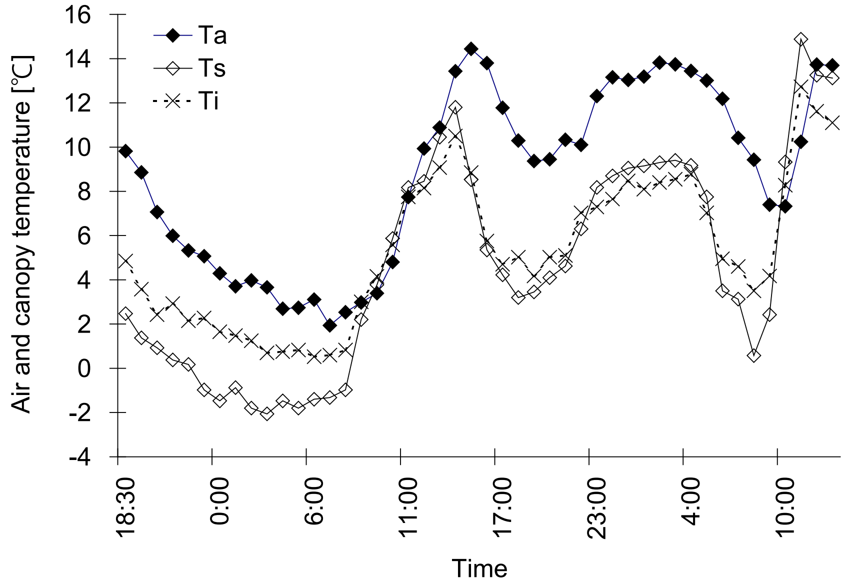
b)


Cloud cover [fraction]
1.0
0.8
0.6
0.4
0.2
0.0
18:30
0:00
6:00
11:00
17:00
23:00
4:00
10:00
Time

c)


T5
T10
9.5
9.0
8.5
8.0
7.5
7.0
6.5
6.0
Soil temperature [℃]
18:30
0:00
6:00
11:00
17:00
23:00
4:00
10:00
Time

d)


Ta
Ts
Ti
16
14
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2
0
-2
-4
Air and canopy temperature [℃]
18:30
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6:00
11:00
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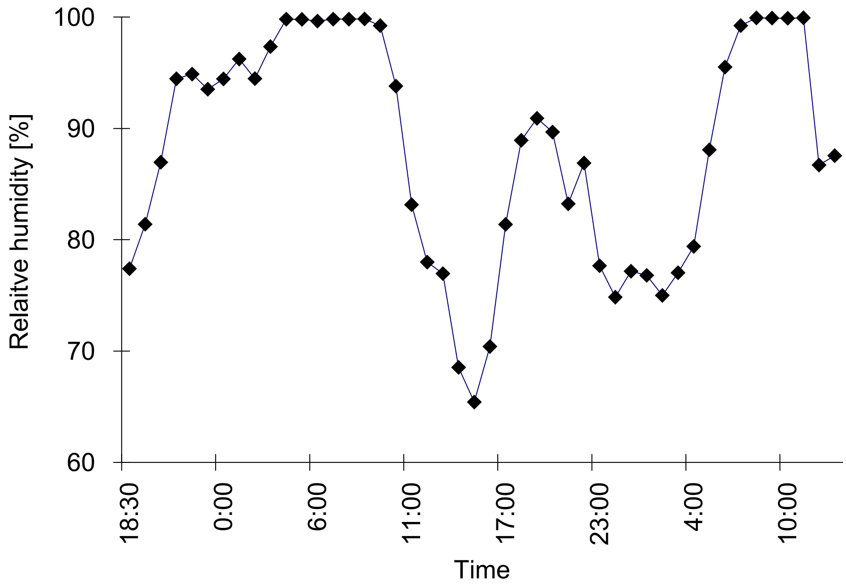
4:00
10:00
Time

e)



f)

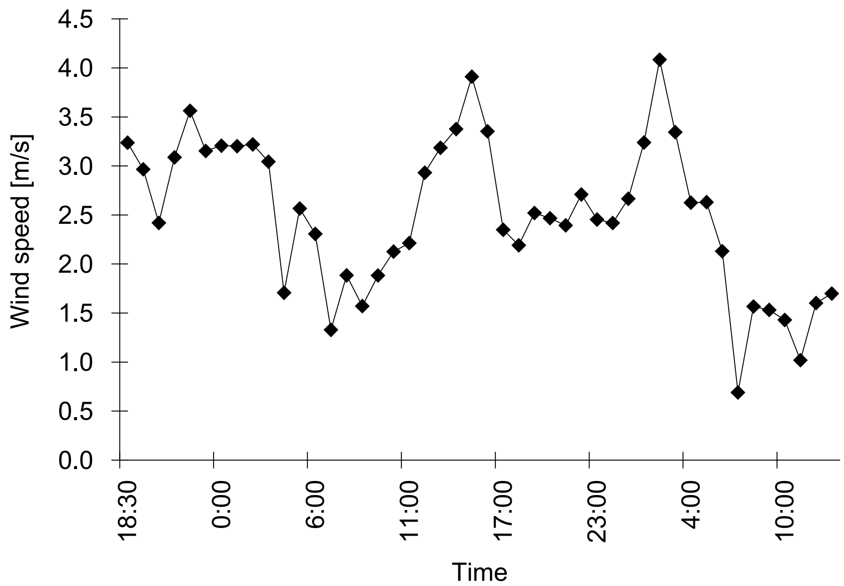


Figure 5.2: Time course of a) net radiation (R_n) and soil heat flux (G), b) cloud cover, c) soil temperature at 5 (T_5) and 10 cm (T_{10}) depth, d) air (T_a), canopy surface (T_s) and in-canopy temperature (T_i), e) relative humidity, and f) wind speed from 18^{30} on 19 November to 13^{00} on 21 November, 2009.

During the first night R_n was negative all the time. This caused T_5 to be significantly less than T_{10} (Fig. 5.2c), which gave rise to a distinct soil heat flux towards the soil surface (Fig. 5.2a). Initially it rose towards the middle of the night and later began to fall in response to an increasing and then decreasing soil temperature gradient. As a result of radiation input after sunrise, soil temperatures began to increase, more so and sooner 5 than at 10 cm depth, as one would expect. At 11^{00} T_5 surpassed T_{10} and G was now directed into the soil profile. Later in the day, as R_n declined again, T_5 followed suit with a few hours delay. T_{10} also declined, but later still and not as much as T_5. Consequently, heat flow into the soil diminished and became positive after 18^{00}. Beginning at this hour cloud cover increased (Fig. 5.2b), which reduced the radiation loss from the surface and prevented it from cooling much further. Hence, the decline in soil temperature was slowed. Since only a small temperature gradient towards the soil surface had developed so far (cf. Figure 5.2c) there was little heat flow towards the soil surface. Around 22^{00} warm air began to move in, which is visibly manifested in the temperature increases in Figure 5.2d (and was clearly felt by the author during the hourly canopy temperature measurement with the hand-held IR-thermometer). This energy input eventually raised T_5 slightly above T_{10} and induced a small heat flux into the soil. At 4^{00} the clouds (and the warm air) began to disappear so that radiation loss increased again (cf. Figure 5.2a). This in turn caused T_5 to fall below T_{10} and, thus, a distinct soil heat flux towards the soil surface. As R_n became positive again during the day, the soil temperatures were reversed once more, leading to a G now directed into the soil.

The variation of soil temperature and the associated soil heat flux also behaved as they should, namely broadly as a mirror image of the course of net radiation and in direct response to heat input by warm air, but with a smaller amplitude and a certain time lag arising from the heat storage capacity of the soil.

In response to the course of net radiation, but also with a certain time lag, above ground temperatures normally start to fall in the afternoon until some time after sunrise, when they start to increase again. This is precisely what happened from the beginning of the observations until 18^{00} the next evening. At that point the decline in temperatures was first halted due to increasing cloudiness, which reduced radiation loss, and later even reversed as a result of warm air moving across the area. The disappearance of the clouds and of the warm air after 4^{00} lead to a steep increase in radiation loss and, consequently, falling temperatures. They began to climb again as R_n increased after sunrise.

Apart from 9^{00} to 11^{00} on November 20, air temperature (T_a) was always higher than the surface temperature of the grass (T_s). During the night of 19/20 November the temperature inside the grass canopy (T_i) was always higher than that at the canopy surface (T_s), because of radiative heat loss from the surface and a heat gain from G inside the canopy. During most of the day of 20 November T_s was slightly greater than T_i because of a radiation gain at the canopy surface and some heat loss to the soil from inside the canopy. In the late afternoon T_s fell below T_i again as net radiation became negative (Fig. 5.2a). This should have stayed like this throughout the night, but the incoming clouds retarded the radiation loss from the canopy surface and the warm air moving across even heated it up. As a result T_s began to rise above T_i after 22^{00} and stayed higher until the clouds and the warm air had disappeared around 6^{00}. At that time T_s fell below T_i again due to radiative cooling. As R_n became positive again after sunrise, T_s climbed above T_i once more.

The response of T_a, T_s and T_i to net radiation, cloudiness and warm air was as expected, too.

If there is no vapour transferred into or out of a system, relative humidity is related to air temperature: it increases as the air temperature decreases and vice versa. In general relative humidity is therefore higher at night than by day, since the air temperature is lower at night. The change in relative humidity is displayed in Figure 5.2e. In the first night relative humidity increased as the air cooled down. It reached 100% at around 4^{00} and remained at that level until it became warmer after sunrise. With the rise in air temperature h_r dropped once more. This cycle began to repeat itself at 15^{00} on the second day. However, the cooling of the air and the associated increase in h_r was stopped and then even reversed by the passage of warm air. The normal cycle resumed once it and the clouds had gone.

Wind speed is often but not always higher during the day than at night. As a consequence there is no unique pattern the data should follow so that their validity cannot be ascertained from the time course observed here (Fig. 5.2f). However, there is nothing in the wind data to suggest that they are anything but alright.

In summary, all meteorological variables behaved as they should. Hence, there is no reason to doubt the quality of the data.

5.4.2 Performance of the equations

Due to different assumptions made in the various equations they cannot be expected to produce exactly the same values, but they should be fairly similar. Columns 2 to 5 in Table 5.1 show the results of computations with the four equations for the nights of November 19/20 and 20/21, 2009. They differ significantly among each other. The differences between the EB and TVT equation (Fig. 5.3) are particularly intriguing, because they are really just different sides of the same equation (cf. Eq. 5.1) and should therefore agree.

The divergence between the estimates with the EB and the TVT could be due to faulty data. However, this is unlikely, since the data are plausible, as pointed out in section 5.4.1. An-

Table 5.1a: Dewfall estimated with the energy balance (λE_{EB}), turbulent vapour transport (λE_{TVT}), Penman-Monteith (λE_{PM}), and Bowen ratio energy balance equation (λE_{BREB}). Columns 2 - 4 without, columns 10 - 13 with adjustment of the heat and/or vapour conductance. For λE_{PM}^* (column 13) the slope of the saturation vapour pressure curve was evaluated at the mean of air and surface temperature. Also presented are the conductances computed with Eq. 2.8 (g), the adjustment factor (φ) required for a given hour to achieve the best fit between λE_{EB} and λE_{TVT}, the product of $g \cdot \varphi$, and the conductance computed with a stability correction (g_{Ψ}). Data for 18^{30} on 19 November to 8^{00} on 20 November, 2009.

1	2	3	4	5	6	7	8	9	10	11	12	13
Time	λE_{EB} W/m^2	λE_{TVT} W/m^2	λE_{PM} W/m^2	λE_{BREB} W/m^2	g mol/m^2/s	φ dim.less	$g \cdot \varphi$ mol/m^2/s	g_{Ψ} mol/m^2/s	λE_{EB} W/m^2	λE_{TVT} W/m^2	λE_{PM} W/m^2	λE_{PM}^* W/m^2
18:30	-74.87	54.30	-2.88	15.60	0.59	0.29	0.17	0.263	15.28	15.28	19.31	17.65
19:00	-71.49	59.88	2.42	15.87	0.54	0.27	0.14	0.226	15.62	15.62	19.10	17.29
20:00	-35.22	43.57	10.64	15.83	0.44	0.36	0.16	0.171	15.82	15.82	18.48	16.88
21:00	-52.74	64.12	13.93	16.68	0.57	0.26	0.15	0.268	16.48	16.48	18.53	16.97
22:00	-59.66	66.32	12.65	15.95	0.66	0.24	0.16	0.349	15.96	15.96	17.79	16.38
23:00	-66.54	64.66	10.28	14.20	0.58	0.22	0.13	0.271	14.31	14.31	16.29	14.74
00:00	-68.06	61.90	8.93	12.40	0.60	0.20	0.12	0.285	12.25	12.25	13.92	12.63
01:00	-45.43	50.93	11.78	13.47	0.60	0.26	0.16	0.305	13.37	13.37	14.80	13.68
02:00	-63.67	61.12	11.22	14.16	0.60	0.23	0.14	0.286	13.98	13.98	15.90	14.41
03:00	-63.15	61.96	11.91	12.57	0.57	0.20	0.11	0.261	12.45	12.45	14.03	12.68
04:00	-9.91	26.94	12.90	11.88	0.32	0.44	0.14	0.109	11.90	11.90	12.96	12.02
05:00	-42.94	43.77	9.16	8.47	0.48	0.19	0.09	0.213	8.49	8.49	9.30	8.57
06:00	-37.46	39.79	8.54	7.99	0.43	0.20	0.09	0.178	8.03	8.03	8.78	8.10
07:00	-6.54	16.12	7.45	6.98	0.25	0.43	0.11	0.077	7.01	7.01	7.51	7.08
08:00	-24.55	25.32	5.08	4.80	0.35	0.18	0.06	0.138	4.66	4.66	5.02	4.71

Table 5.1b: Like Table 5.1a, but data for 18^{00} on 20 November to 8^{00} on 21 November, 2009.

1	2	3	4	5	6	7	8	9	10	11	12	13
Time	λE_{EB} W/m^2	λE_{TVT} W/m^2	λE_{PM} W/m^2	λE_{BREB} W/m^2	g mol/m^2/s	φ dim.less	g·φ mol/m^2/s	g_Ψ mol/m^2/s	λE_{EB} W/m^2	λE_{TVT} W/m^2	λE_{PM} W/m^2	λE_{PM}^* W/m^2
18:00	-42.49	60.67	12.13	17.08	0.40	0.28	0.11	0.136	17.06	17.06	19.63	17.87
19:00	-53.33	58.35	4.74	11.17	0.46	0.19	0.09	0.185	11.16	11.16	12.58	11.61
20:00	-40.94	48.01	5.34	11.99	0.45	0.25	0.11	0.185	12.08	12.08	13.57	12.64
21:00	-59.24	37.42	-10.16	4.59	0.43	0.12	0.05	0.171	4.54	4.54	5.35	4.99
22:00	-37.65	26.04	-6.00	5.58	0.49	0.21	0.10	0.243	5.61	5.61	6.35	6.08
23:00	-41.55	5.00	-18.58	1.03	0.44	0.20	0.09	0.199	1.15	1.15	1.88	1.87
00:00	-52.27	1.40	-26.47	0.11	0.43	0.08	0.03	0.189	0.15	0.15	0.46	0.48
01:00	-40.90	1.41	-20.02	0.37	0.48	0.27	0.13	0.232	0.02	0.02	0.98	1.06
02:00	-56.23	1.18	-28.44	0.21	0.58	0.18	0.11	0.317	0.18	0.18	0.99	1.05
03:00	-88.14	4.18	-44.47	0.37	0.73	0.09	0.06	0.443	0.38	0.38	1.00	1.03
03:30	-69.67	8.37	-33.34	0.67	0.60	0.08	0.05	0.326	0.67	0.67	1.10	1.10
04:00	-49.62	12.98	-20.50	1.70	0.47	0.13	0.06	0.221	1.51	1.51	2.06	2.02
05:00	-38.18	55.57	6.16	15.10	0.47	0.27	0.13	0.206	15.00	15.00	16.76	15.81
06:00	-47.10	96.99	25.84	25.32	0.38	0.26	0.10	0.119	25.43	25.43	28.90	25.83
07:00	19.52	27.02	26.15	23.30	0.13	0.43	0.05	0.018	23.18	23.18	25.86	23.38
08:00	-47.28	68.58	14.84	12.90	0.29	0.19	0.05	0.070	12.90	12.90	14.73	12.95

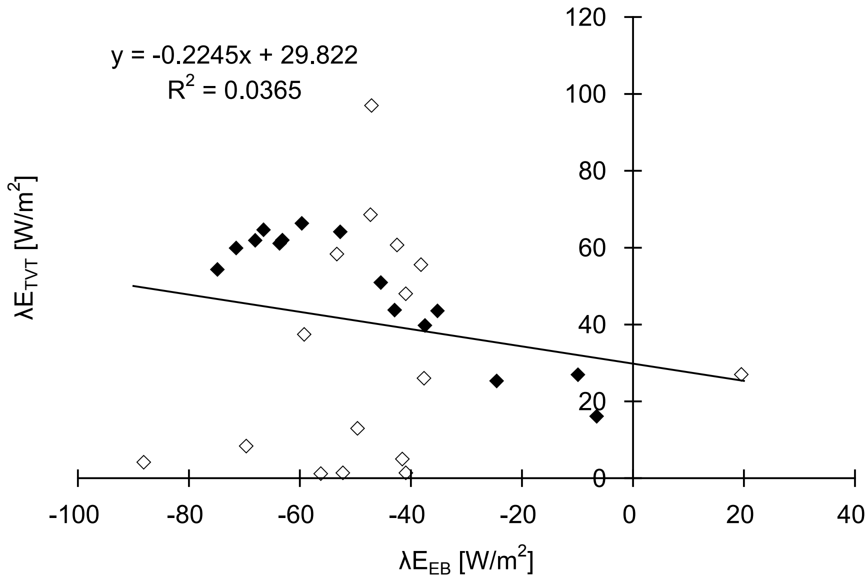


Figure 5.3: Correlation between the latent heat fluxes estimated with the energy balance (λE_{EB}) and the turbulent vapour transport equation (λE_{TVT}). Data for the nights of 19/20 (18^{30} - 8^{00}, ◆) and 20/21 November (18^{00} - 8^{00}, ◇).

other possible cause for errors are the empirical relationships used in the equations. The expressions for computing vapour pressure from temperature and relative humidity (Eq. 2.10 and 2.11) were proven to be good by various researchers (e.g. Buck, 1981; Campbell and Norman, 1998). In contrast, in the equation for heat (g_H) and vapour conductance (g_v) some coefficients are still under discussion. Furthermore, these conductances depend on atmospheric stability, which was not considered here. This makes g_H and g_v rather likely sources of error. This will be elaborated in the discussion section.

If the problem lies in the g terms, then multiplying them with a coefficient φ, whose appropriate value is to be found by iteration, should improve the agreement between the EB and TVT equation. Figure 5.4 indicates that after an adjustment the agreement is much better indeed. Using a single value for each night the best agreement was achieved with φ = 0.25 for the first and the second night. However, the fit around the 1:1 line is still not all that good. Using a different value for each hour results in a perfect agreement between the two equations (Fig. 5.5; Table 5.1, columns 10 and 11).

The perfect correlation between the EB and TVT equation after adjustment of the g terms suggests that they were indeed the cause of the initial disagreement. In the following the results obtained with the EB equation are compared with those obtained with the PM and BREB equation. In these comparisons the EB and the PM equation are used with unadjusted, nightly and hourly adjusted g terms. The adjustment factors applied are those derived by comparing the EB and TVT equation.

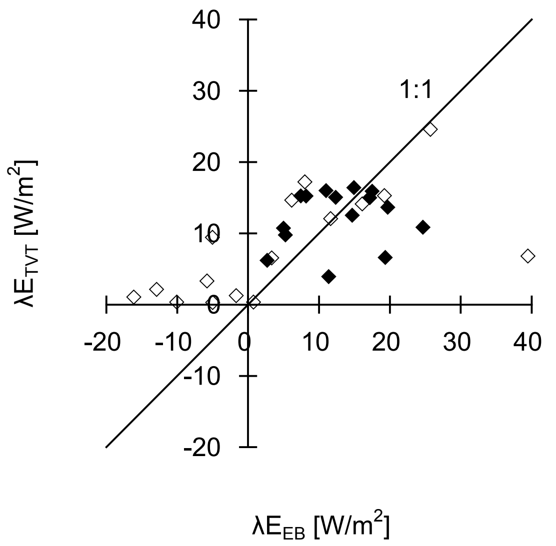


Figure 5.4: Correlation between the latent heat fluxes estimated with the energy balance (λE_{EB}) and the turbulent vapour transport equation (λE_{TVT}) after multiplying the g terms in both equations by 0.25. Data for the nights of 19/20 (18^{30} - 8^{00}, ◆) and 20/21 November (18^{00} - 8^{00}, ◇).

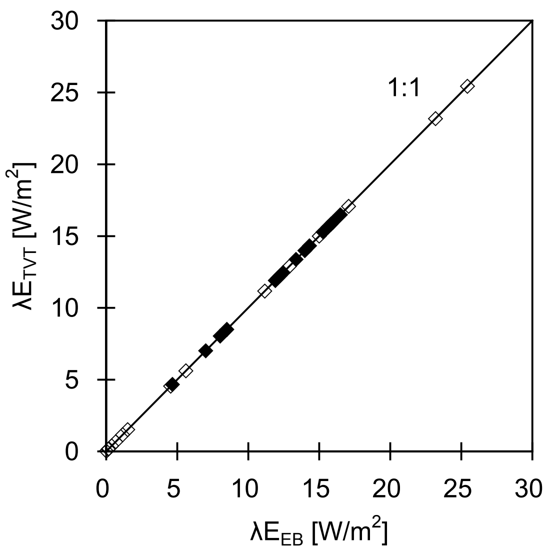
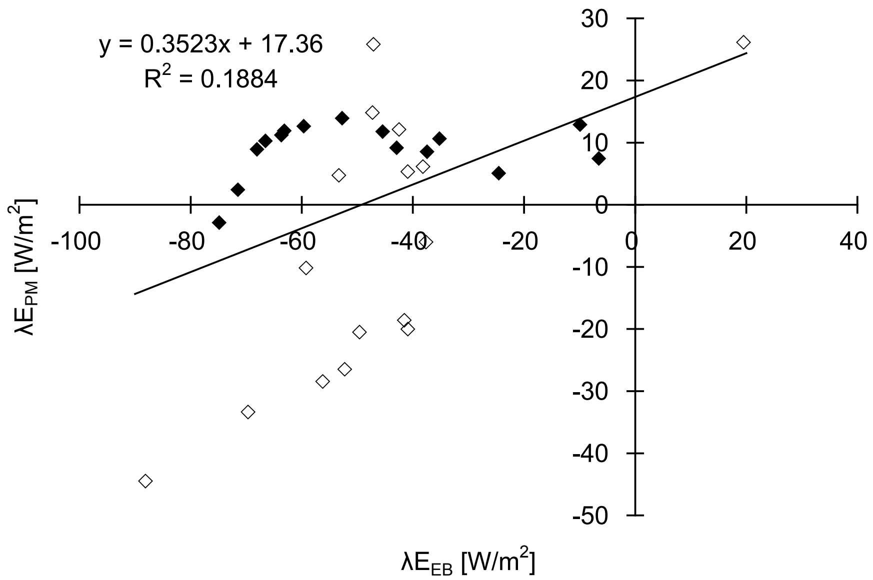


Figure 5.5: Correlation between the latent heat fluxes estimated with the energy balance (λE_{EB}) and the turbulent vapour transport equation (λE_{TVT}) after multiplying the g terms in both equations by a different value for each hour. Data for the nights of 19/20 (18^{30} - 8^{00}, ◆) and 20/21 November (18^{00} - 8^{00}, ◇).

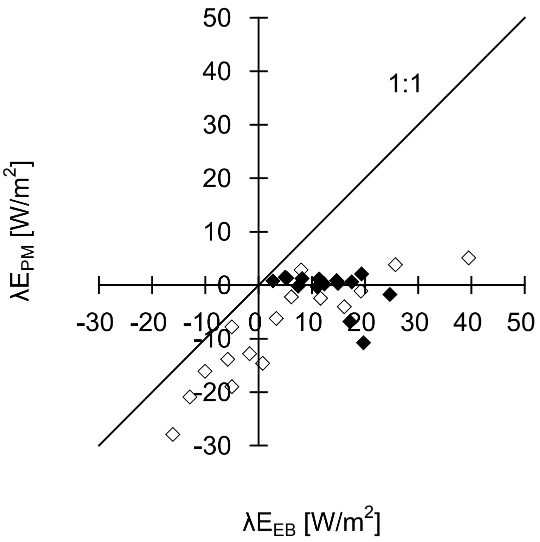
The PM equation (Eq. 5.12) contains the vapour conductance (g_v) in the second term of the numerator. Therefore, following the findings above, the coefficient φ must be introduced to adjust this term. However, if the relative humidity near the surface is 100%, which is often the case when dew forms, $e_a - e_s(T_a) = 0$ and the term containing g is inconsequential.

Figure 5.6a shows that there is no agreement between the EB and PM equations, if the g terms are not adjusted. There is some agreement after multiplying g with φ = 0.25 for each night (Fig. 5.6b). An hourly adjustment improves the correlation significantly (Fig. 5.6c), although it is not as good as the one between the EB and TVT equation.

a)



b)



c)

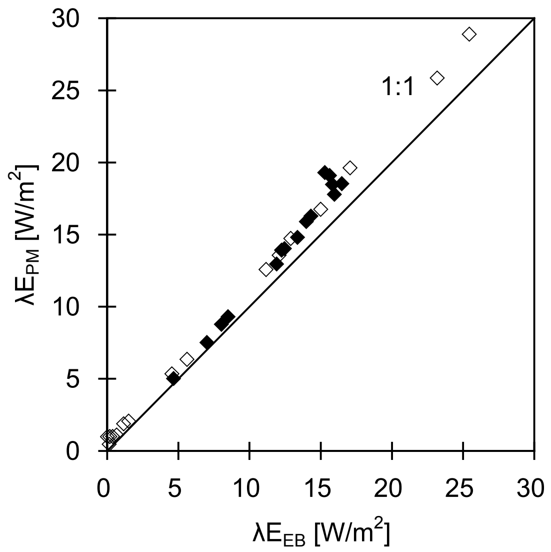


Figure 5.6: Correlation between the latent heat fluxes estimated with the energy balance (λE_{EB}) and the Penman-Monteith equation (λE_{PM}) a) without adjustment of the g terms, b) after multiplying the g terms in both equations by 0.25, and c) after multiplying the g terms in both equations with a different value for each hour. Data for the nights of 19/20 (18^{30} - 8^{00}, ◆) and 20/21 November (18^{00} - 8^{00}, ◇).

Figure 5.6c shows that the latent heat flux computed with the adjusted PM equation is consistently higher than the one computed with the adjusted EB equation. The possible reason for this bias lies in the calculation of s, i.e. the slope of the saturation vapour pressure curve (Eq. 5.7). So far it was evaluated at T_a, which is the intended procedure for the PM equation, but introduces some error (cf. section 5.3.2.3). Deriving it for $(T_a + T_s) / 2$ is more precise and here leads to a better agreement with the energy balance equation (Fig. 5.7).

In the BREB equation the vapour and heat conductance cancel out in the term β (cf. Eq. 5.17 and 5.18). Hence, calculations with this equation are not affected by any difficulties in getting a correct value for g. Figure 5.8 depicts the correlation between estimates with the EB and the BREB equation. Without adjusting the g term in the EB equation the correlation is very poor (Fig. 5.8a). Adjusting the g term with a single value for each night greatly improves the correlation (Fig. 5.8b). A near perfect one is achieved, if the g term is adjusted with a different value for each hour (Fig. 5.8c).

The close agreement between the EB equation, after adjustment of the g terms, and the BREB, which does not contain a g term and therefore needs no adjustment, is further evidence that there is a problem with the g terms. Hence, the absence of a g term makes the BREB equation a better alternative for computing dewfall than the EB, TVT or PM equation.

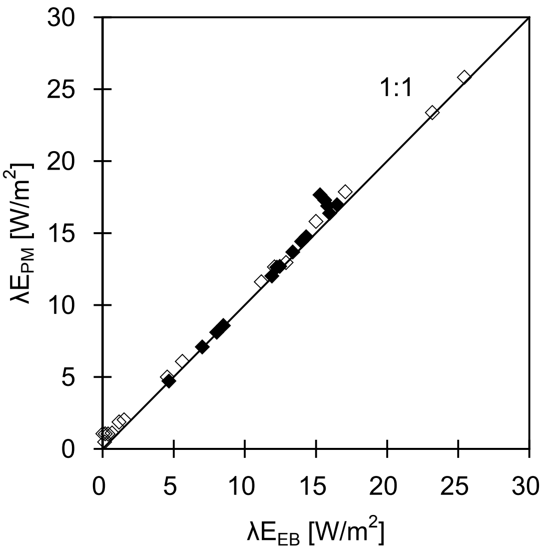
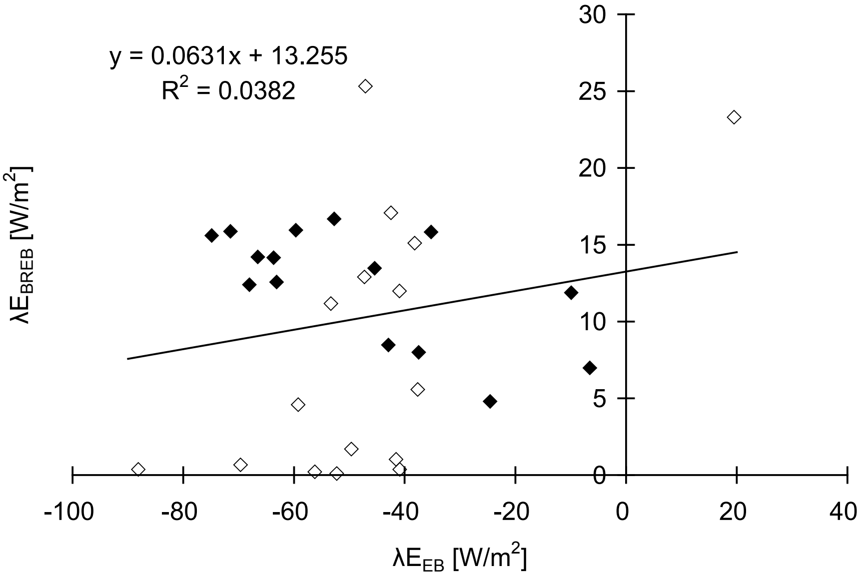
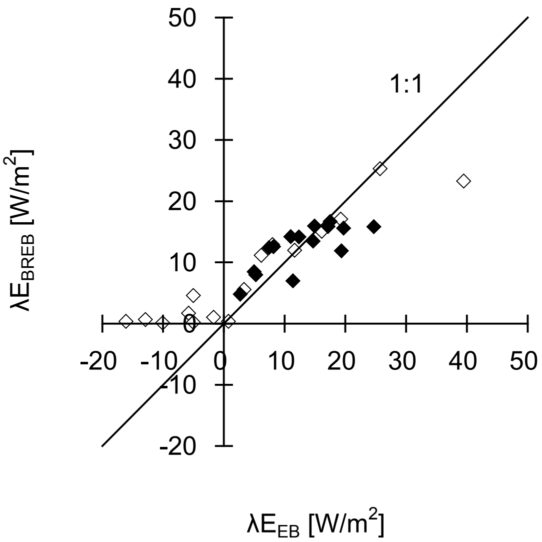


Figure 5.7: Like Figure 5.6c, but with the slope of the saturation vapour pressure curve in the PM equation evaluated at the mean of air and surface temperature.

a)



b)



c)

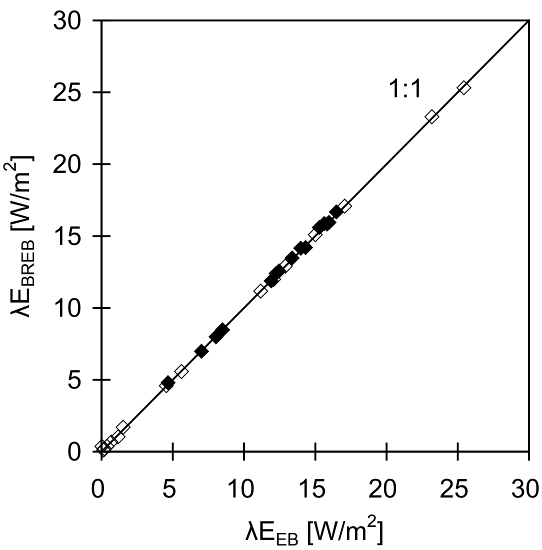


Figure 5.8: Correlation between the latent heat fluxes estimated with the energy balance (λE_{EB}) and the Bowen ratio energy balance equation (λE_{BREB}) a) without adjustment of the g terms, b) after multiplying the g terms in both equations by 0.25, and c) after multiplying the g terms in both equations with a different value for each hour. Data for the nights of 19/20 (18^{30} - 8^{00}, ◆) and 20/21 November (18^{00} - 8^{00}, ◇).

5.5 Discussion

While in principle all four equations can be used to compute dewfall, section 5.4 has shown that there is a problem in getting the correct value of g. This was clearly demonstrated by the fact that all four gave different values at first, but after adjustment of g agreed very well (cf. column 5 and 10 - 13 in Table 5.1). In light of this the BREB equation is the best choice, because it does not contain a g term. The second best choice is the PM equation. It does contain a g term, but this becomes ever smaller as e_a approaches e_s (i.e. as the relative humidity approaches 100%), at which point it reaches zero. When dewfall takes place the relative humidity is usually rather high. Hence, the g term poses a much smaller problem in the PM than in the EB or TVT equation.

An argument against the BREB equation is that it requires the temperature of the surface, which is often not available. (Recall that the EB and TVT equation do, too.) In many cases the PM equation will therefore be the only alternative, despite containing a g term and the inaccuracy in evaluating s at air temperature, since it does not require T_s.

In this chapter we presented in detail computations based on data from 19 to 21 November, 2009. For these days we had the most complete data set. However, we also carried out computations for four other dates. They showed the same result: the agreement between the four equations was poor without an adjustment of g, but very good with an adjustment. However, the most suitable value of the adjustment factor φ was different for each date and, averaged over the day in question, ranged from 0.25 to 3.6.

To identify the reasons for incorrect g values let us look again at Eq. 2.8, which was used here to calculate it: $g = g_v = g_H = k^2 \cdot u \cdot \left(\ln[(z-d)/z_0]\right)^{-2}$. This equation only holds for adiabatic (neutral) conditions. In this case there is only mechanical turbulence, which is always directed upwards and arises from wind and roughness of the surface. Now, if a surface becomes much warmer than the air above it, an upward thermal turbulence develops, which enforces the mechanical one. As a result, the conductance for heat and vapour increases and their turbulent transport into the atmosphere is enhanced. One speaks of unstable conditions. If a surface becomes much cooler than the air above it, a downward thermal turbulence ensues, which suppresses the mechanical turbulence. Consequently, the conductance, and with it the turbulent transport of heat and vapour into the atmosphere, decreases. One then speaks of stable conditions.

To account for the condition of the atmosphere the complete version of Eq. 2.8 contains a stability factor ψ and then reads $g = k^2 \cdot u \cdot \left(\ln[(z-d)/z_0] + \psi\right)^{-2}$. However, ψ is usually left out, because it can only be evaluated through an iterative approach. For further details on the stability correction and its computation the reader is referred to Monteith and Unsworth (1990) and Campbell and Norman (1998).

Dewfall typically occurs when the canopy is cooler than the surrounding air, i.e. under stable conditions. Hence, the conductance computed with the stability correction can be expected to be smaller than the conductance computed without it. This is illustrated in Table 5.2, which shows results from own calculations: the cooler the surface compared to the surrounding air, the larger the reduction in g due to downward thermal turbulence. This reduction becomes larger as wind speed decreases.

Table 5.2: Percent reduction in heat or vapour conductance due to the stability correction in relation to wind speed at 10 m height (u) and the difference between air temperature at 2 m height and the surface temperature of 21 cm tall grass (i.e. $\Delta T = T_a - T_s$). Air temperature was set to 8°C in all calculations.

	u (m/s)				
ΔT (°C)	1	2	3	4	5
1	60.7	38.7	25.6	17.7	12.7
2	69.6	50.2	36.7	27.4	20.9
3	74.0	56.6	43.5	33.9	26.8
4	76.7	60.7	48.3	38.7	31.3
5	78.6	63.8	51.9	42.5	35.0

To assess the influence of atmospheric stability on g on the nights of 19/20 and 20/21 November, 2009, the g values were computed with and without stability correction for the meteorological data during these two nights. The results are shown in Table 5.1. For these nights the g's with the correction (column 9) were on average 45% of the value without it (column 6). As stated above, the single value of φ which gave the best results for this period was 0.25. Hence, while not considering the stability correction undoubtably contributed to the incorrect values of g, this cannot be the sole cause.

In the conductance equation the value of the von Karman constant (k), the zero plane displacement (d) and the roughness length (z_0) are still subject to discussion. While k is normally taken as 0.40 or 0.41, Tennekes (1968) indicated its value to be 0.34. Businger et al. (1971) quoted is as ~ 0.35. Telford (1982) named the appropriate k to be 0.37, similar to results by Bergmann (1998), who gives a value of 0.3678.

There are also different proposals for evaluating the zero plane displacement (d) and roughness coefficient (z_0), too. Campbell and Norman (1998) propose that $d = 0.65 \cdot h$ and $z_0 = 0.1 \cdot h$, where h = crop height. These relationships were applied in the calculations presented here, unless stated otherwise. In other sources different relationships are mentioned for d and z_0. For example, Monteith and Unsworth (1990) give the representative value of d as 0.5·h to 0.7·h, and of z_0 as 0.08·h to 0.12·h. Maki (1975) indicated d = 0.75·h and z_0 = 0.1·h. Sudmeyer et al. (1994) expressed them as d = 0.63·h and z_0 = 0.11·h. Madeira et al. (2002) applied d = 0.64·h and z_0 = 0.13·h. In studies of Inclán and Forkel (1995), d and z_0 are calculated as 67% and 13% of the vegetation height. Furthermore, the value of d and z_0 are related to plant density (Campbell and Norman, 1998) and leaf area index (Shaw and Pereira, 1982; Biftu and Gan, 2000). Without considering these factors the determined value of d and z_0 may be off.

The minimum and maximum values of the coefficients κ, d and z_0 suggested in the literature are listed in Table 5.3. It also gives the values used in the computations in this chapter, and the value of the term $k^2 \cdot (\ln[(z-d)/z_0])^{-2}$, which is the equation for g without wind speed (cf. Eq. 2.8), computed with the minimum, maximum and our values for the coefficients. The ratio of the minimum and maximum value of this term to the value used here is 0.64 and 1.09, respectively.

Table 5.3: Minimum and maximum value of the coefficients k (von Karman constant), d (zero plane displacement) and z_0 (roughness coefficient), the values used in the calculations in this chapter, and the variation of the conductance term without wind speed u when the different values are used (h = mean canopy height = 0.21 m, z = height of wind measurement = 10 m).

coefficient	minimum value	maximum value	value used here
k	0.34	0.41	0.41
d	0.5·h	0.75·h	0.65·h
z_0	0.08·h	0.13·h	0.1·h
$k^2 \cdot (\ln[(z-d)/z_0])^{-2}$	0.00284	0.00485	0.00444
ratio to value used here	0.64	1.09	1

The variation of k, d and z_0 is quite distinct, but one cannot judge which value is correct. Provided the minimum value is right, the term g in the calculations here is overestimated and the adjustment coefficient φ should be 0.64. If the maximum value is correct, g is underestimated and φ should be 1.09. If one applies the minimum values of the coefficients k, d and z_0 in addition to the stability correction, a combined reduction of g to $0.45 \cdot 0.64 = 0.29$ of its uncorrected value would result. This approaches the average adjustment required for a good fit (φ = 0.25), but is still to far off to yield a satisfactory agreement between the g's adjusted with a different φ for each hour and the g's corrected for stability and computed with the minimum values of k, d and z_0 (Fig. 5.9). (Note that the average φ is a latent heat flux weighted mean of the hourly φ values.)

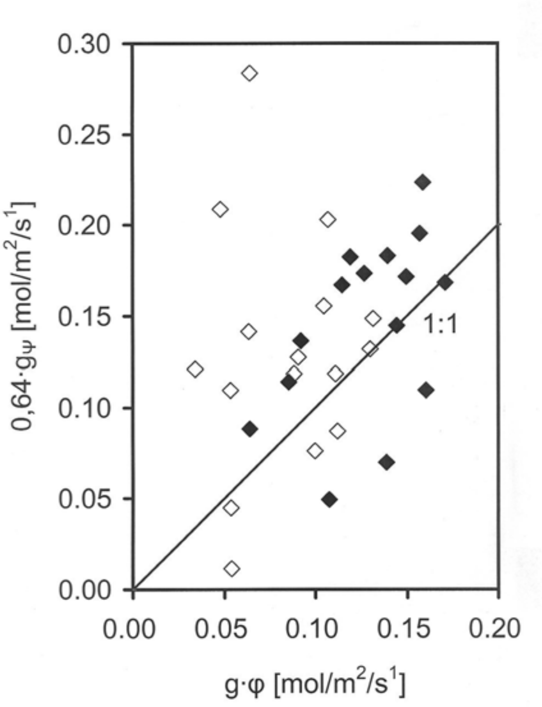


Figure 5.9: Comparison of the conductance computed with Eq. 2.8 without stability correction, but adjusted with a different factor φ for each hour, and the conductance computed with Eq. 2.8 with a stability correction and the minimum value of $k^2 \cdot (\ln[(z-d)/z_0])^{-2}$ given in Table 5.3. Data for the nights of 19/20 (18^{30} - 8^{00}, ◆) and 20/21 November (18^{00} - 8^{00}, ◇).

Applying the stability correction was obviously not sufficient to trace the hourly variation of the adjustment factor φ, even after figuring in the minimum values of k, d and z_0. Hence, there must be other reasons still, why the equation for conductance applied here does not produce good values.

The conductance equation assumes that the wind profile is in equilibrium with the surface under investigation. This requires the surrounding area to be under the same vegetation and large enough for an equilibrium to develop (sufficient fetch). The surface of interest here was 1 m² of ~ 21 cm tall grass. Being located at an experiment station, the vegetation around it is quite varied. Hence, an equilibrium is unlikely so that it should not come as a surprise that the equation does not give good results. This is further underlined by the fact that at another night φ was > 1, even though the atmosphere was stable so that a stability correction would have led to smaller g values than without the correction.

To clarify what exactly caused the difficulties in computing proper g values here requires further investigations, which are beyond the scope of this thesis.

5.6 Conclusions

There is ample evidence in the literature that the EB, TVT, PM and BREB equation can be used successfully to compute dewfall. However, under our conditions there was a problem in obtaining a correct value for the heat and vapour conductance. This severely hindered the use of the EB and TVT equation, in which they play a major role. The PM equation also contains a conductance term, but at high values of relative humidity typical during dewfall events it is much less important than in the EB or TVT equation. The inaccuracy arising from assessing s in the PM equation at air temperature is usually small. Hence, in most cases the PM equation can be expected to yield reasonable results. Due to the absence of a conductance term, the BREB equation turned out to be the best alternative.

However, the BREB (as well as the EB and TVT) equation requires the temperature of the surface, which is often not available. Then the PM equation, from which it has been eliminated, is the only option.

The precise reason for the difficulties in getting proper conductance values in this study is not certain. The most likely one it that the wind profile could not equilibrate with the vegetation on the lysimeter, because it was not surrounded by a large enough area with the same surface conditions (vegetation) to do so. This invalidates the equation for computing heat and vapour conductances. To confirm this it requires further investigations beyond the scope of this thesis.

5.7 References

Atzema, A. J., Jacobs, A. F. G., Wartena, L. (1990): Moisture distribution within a maize crop due to dew. Netherlands J. Agric. Sci. 38, 117-129.

Baier, W. (1966): Studies on dew formation under semi-arid conditions. Agric. Meteorol. 3, 103-112.

Bergmann J. C. (1998): A physical interpretation of von Karman's constant based on asymptotic considerations - a new value. J. Atmos. Sci. 55, 3403-3407.

Biftu, G. F., Gan, T. Y. (2000): Assessment of evapotranspiration models applied to a watershed of Canadian Prairies with mixed land-uses. Hydrol. Process. 14, 1305-1325.

Bowen, I. S. (1926): The ratio of heat losses by conduction and by evaporation from any water surface. Phys. Rev. 27, 779-787.

Buck, A. L. (1981): New equations for computing vapor pressure and enhancement factor. J. Appl. Meteorol. 20, 1527-1532.

Businger, J. A., Wyngaard, J. C., Izuml, Y., Bradley, E. F. (1971): Flux-profile relationships in the atmospheric surface layer. J. Atmos. Sci. 28, 181-189.

Campbell, G. S., Norman, J. M. (1998): An introduction to environmental biophysics, 2nd ed. Springer, New York.

Duvdevani, S. (1964): Dew in Israel and its effect on plants. Soil Sci. 98, 14-21.

Evenari, M., Shanan, L., Tadmor, N. (1971): The Negev. The challenge of a desert. Harvard University Press, Cambridge, Massachusetts.

Garratt, J. R., Segal, M. (1988): On the contribution of atmospheric moisture to dew formation. Bound.-Layer Meteorol. 45, 209-236.

Hiltner, E. (1930): Der Tau und seine Bedeutung für den Pflanzenbau. Eine Theorie über die physiologische Bedeutung der Wasseraufnahme durch oberirdische Pflanzenorgane. Wiss. Arch. Landw., Abt. A, 3, 1-70.

Inclán, M. G., Forkel, R. (1995): Comparison of energy fluxes calculated with the Penman-Monteith equation and the vegetation models SiB and Cupid. J. Hydrol. 166, 193-211.

Jacobs, A. F. G., van Pul, W. A. J., van Dijken, A. (1990): Similarity moisture dew profiles within a corn canopy. J. Appl. Meteorol. 29, 1300-1306.

Jacobs, A. F. G., van Pul, A., El-Kilani, R. M. M. (1994): Dew formation and the drying process within a maize canopy. Bound.-Layer Meteorol. 69, 367-378.

Jacobs, A. F G., van Boxel, J. H., Nieveen, J. (1996): Nighttime exchange processes near the soil surface of a maize canopy. Agric. Forest Meteorol. 82, 155-169.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (1999): Dew deposition and drying in a desert system: a simple simulation model. J. Arid Environ. 42, 211-222.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (2000): Dew measurements along a longitudinal sand dune transect, Negev Desert, Israel. Int. J. Biometeorol. 43, 184-190.

Jacobs, A. F. G., Heusinkveld, B. G., Kruit, R. J. W., Berkowicz, S. M. (2006): Contribution of dew to the water budget of a grassland area in the Netherlands. Water Resour. Res. 42, W03415.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (2008): Passive dew collection in a grassland area, The Netherlands. Atmos. Res. 87, 377-385.

Janssen, L. H. J. M., Römer, F. G. (1991): The frequency and duration of dew occurrence over a year. Tellus B 43, 408-419.

Kalthoff, N., Fiebig-Wittmaack, M., Meißner, C., Kohler, M., Uriarte, M., Bischoff-Gauß, I., Gonzales, E. (2006): The energy balance, evapo-transpiration and nocturnal dew deposition of an arid valley in the Andes. J. Arid Environ. 65, 420-443.

Kappen, L., Lange, O. L., Schulze, E.-D., Buschbom, U., Evenari, M. (1980): Ecophysiological investigations on lichens of the Negev Desert. VII. The influence of the habitat exposure on dew imbibition and photosynthetic productivity. Flora 169, 216-229.

Kerr, J. P., Beardsell, M. F. (1975): Effect of dew on leaf water potentials and crop resistances in a Paspalum pasture. Agron. J. 67, 596-599.

Long, I. F. (1958): Some observations on dew. Meteorol. Mag. 87, 161-168.

Luo, W., Goudriaan, J. (2000): Dew formation on rice under varying durations of nocturnal radiative loss. Agric. For. Meteorol. 104, 303-313.

Madeira, A. C., Gillespie, T. J., Duke, C. L. (2001): Effect of wetness on turfgrass canopy reflectance. Agric. Forest Meteorol. 107, 117-130.

Madeira, A. C., Kim, K. S., Taylor, S. E., Gleason, M. L. (2002): A simple cloud-based energy balance model to estimate dew. Agric. For. Meteorol. 111, 55-63.

Maki, T. (1975): Interrelationships between zero-plane displacement, aerodynamic roughness length and plant canopy height. J. Agric. Meteorol. 31, 7-15.

Meissner, R., Seeger, J., Rupp, H., Seyfarth, M., Borg, H. (2007): Measurement of dew, fog, and rime with a high-precision gravitation lysimeter. J. Plant Nutr. Soil Sci. 170, 335-344.

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

Monteith, J. L. (1963): Dew: facts and fallacies. In: Rutter, A. J., Whitehead, F. H. (eds), The water relations of plants (pp. 37-56). Wiley, New York.

Monteith, J. L. (1973): Principles of environmental physics. Arnold, London.

Monteith, J. L., Unsworth, M. H. (1990): Principles of environmental physics, 2nd ed. Arnold, London.

Neumann, J. (1956): Estimating the amount of dewfall. Arch. Meteorol. Geophys. Bioklimatol., Ser. A, 9, 197-203.

Pedro, M. J., Gillespie, T. J. (1982a): Estimating dew duration. I. Utilizing micrometeorological data. Agric. Meteorol. 25, 283-296.

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

Penman H. L. (1948): Natural evaporation from open water, bare soil and grass. Proc. R. Soc. Lond. A 193,120-145.

Sentelhas P. C., Gillespie, T. J. (2008): Estimating hourly net radiation for leaf wetness duration using the Penman-Monteith equation. Theor. Appl. Climatol. 91, 205-215.

Severini, M., Moriconi, M. L., Tonna, G., Olivieri, B. (1984): Dewfall and evapotranspiration determination during day- and nighttime on an irrigated lawn. J. Clim. Appl. Meteorol. 23, 1241-1246.

Shaw, R. H., Pereira, A. R. (1982): Aerodynamic roughness of a plant canopy: A numerical experiment. Agric. Meteorol. 26, 51-65.

Sudmeyer, R. A., Nulsen, R. A., Scott, W. D. (1994): Measured dewfall and potential condensation on grazed pasture in the Collie River basin, southwestern Australia. J. Hydrol. 154, 255-269.

Telford J. W. (1982): A theoretical value for von Karman's constant. Pure Appl. Geophys. 120, 648-661.

Tennekes, H. (1968): Outline of a second-order theory of turbulent pipe flow. Am. Inst. Aeronaut. Astronaut. J. 6, 1735-1740.

Tuller, S. E., Chilton, R. (1973): The role of dew in the seasonal moisture balance of a summer-dry climate. Agric. Meteorol. 11, 135-142.

Wallin, J. R. (1967): Agrometeorological aspects of dew. Agric. Meteorol. 4, 85-102.

Went, F. W. (1955): Fog, mist, dew and other sources of water. In: Stefferud, A. (ed.), Water. The yearbook of agriculture 1955 (pp. 103-109). U.S. Dept. of Agriculture, Washington, D.C.

6. Comparison between measured and computed dewfall

6.1 Abstract

Dewfall measured with a lysimeter is compared to dewfall computed with the Bowen ratio energy balance (BREB) and Penman-Monteith (PM) equation, which were identified as the most suitable equations for this purpose in the previous chapter.

Measured and calculated data follow a similar tendency, but the calculated values are consistently and significantly higher. Also, the measured dewfall fluctuates considerably, while the calculations show continuous dewfall at a similar rate throughout the night. Finally, compared to the lysimeter record the two equations predict dewfall too early and for too long. Despite these differences the cumulative measured and calculated values correlate quite well for the hours when both the lysimeter and the equations show dewfall.

Estimates with the BREB and PM equation are in very good agreement, but the latter always yields slightly higher values. This largely disappears, if the slope of the saturation vapour pressure curve, which enters into the PM equation, is evaluated at the mean of air and surface temperature rather than just at air temperature, which is the normal procedure.

It was shown in chapter 5 that under the site conditions at Falkenberg it is difficult to get proper values for the heat and vapour conductance. The latter enters into the PM equation as a multiplier of the vapour pressure deficit. Because this was low during the observation period, the conductance had little effect on the results.

For the period considered here net radiation (R_n) and soil heat flux (G) were estimated with empirical equations, because no measured data were available. A comparison of measured values and data computed with these equations for a period, when measured data for R_n and G as well as all the parameters needed to compute them were available, revealed that errors in the estimation of R_n and G could be the reason for the deviations between measured and computed dewfall observed here.

Further measurements are planned so that eventually a full set of measured data will be available to check the performance of the BREB and PM equation against lysimeter data.

6.2 Introduction

In chapter 5, four models to estimate dewfall were assessed, namely the energy balance (EB) equation, the equation for turbulent vapour transport (TVT), the Penman-Monteith (PM) equation, and the Bowen ratio energy balance (BREB) equation.

Except for the BREB equation, the equations contain a heat and/or vapour conductance term. This had to be adjusted by multiplying it with an extra coefficient to get good results. This problem was especially apparent in the EB and TVT equation. The PM equation also involves a conductance term. However, it is part of the term $c_p \cdot g_v \cdot [e_a - e_s(T_a)]$ so that its influence is dependent on the magnitude of the vapour pressure deficit $[e_a - e_s(T_a)]$. Therefore, the nearer the air is to being saturated, the smaller the effect of the conductance on the estimation. Since the air is normally near saturation as dew occurs, the vapour pressure deficit is small and the deviation due to an incorrect conductance in the PM equation is small.

There is no conductance in the BREB equation so that it appears to be the best method to estimate dewfall. However, as pointed out in chapter 5, it requires the temperature of the surface, which is usually not available. In that case the PM equation, from which it has been eliminated, is the only option.

Computations with models need to be validated by comparing them to direct measurements. Some methods to directly measure dew amounts were reviewed in chapter 1. Lysimeters are very promising instruments. Various types were already used to quantify dew (Slatyer and McIllroy, 1961; Sharma, 1976; Severini et al., 1984; Grimmond et al., 1992; Jacobs et al., 2000; Meissner et al., 2007).

In this chapter, dewfall on grass is measured with a high precision weighing lysimeter. These measurements are then compared to computations with the BREB and PM equation.

6.3 Materials and methods

To make a comparison between measurement and computation, several data collection campaigns were carried out at the Falkenberg lysimeter station. Unfortunately, the collection of data was far from ideal. A different problem appeared each time. Sometimes the weather was not beneficial for dew formation, at other times some data were unavailable due to the lack or the mechanical failure of an instrument. The best, albeit incomplete set of data so far was obtained for the period from 13^{00} on 9 October to 12^{00} on 10 October 2008. Sunset was at ~ 18^{30} on 9 October, and sunrise at ~ 7^{30} the next morning.

Dew was measured with the type of high precision weighing lysimeter already described in chapter 3. It has a surface area of 1 m^2, a depth of 2 m and is filled with a sandy soil. It can discern mass changes as small as 20 g, which for their 1 m^2 surface area corresponds to a depth of 0.02 mm of water. The mass is recorded every 10 min. The lysimeter was under grass, which was approximately 40 cm tall on 9/10 October, 2008.

Air temperature, relative humidity, wind speed and surface temperature were measured as described in section 5.3.1. However, on this date net radiation (R_n) and soil heat flux (G) could not be measured, because the instruments were not in place yet. They were estimated as follows.

During the night there is only long wave radiation, while in the early morning there is also some short wave radiation. Hence, the radiation balance is:

$$R_n = L_{in} - L_{out} + (1-\alpha)\cdot S_{in} \tag{6.1}$$

where L_{in} = incoming long wave radiation from the sky, L_{out} = outgoing long wave radiation from the surface, α = albedo or surface reflectance for short wave radiation, and S_{in} = incoming short wave radiation, which was measured with a pyranometer at the lysimeter station (Pyranometer 16130, Wilh. Lambrecht GmbH, Göttingen, Germany). Evett (2001) indicated that the albedo of well watered closed canopies is relatively constant. It was taken as α = 0.24 , which was suggested by Jones (1992) for grass.

Following Campbell and Norman (1998) the incoming and outgoing long wave radiation were calculated with the equations:

$$L_{in} = \varepsilon_{sky}\cdot\sigma\cdot T_a^4 \tag{6.2}$$

$$L_{out} = \varepsilon_s\cdot\sigma\cdot T_s^4 \tag{6.3}$$

where ε_{sky} = emissivity for long-wave radiation of the sky (dimensionless fraction), σ = Stefan-Boltzmann-constant ($5.67\cdot10^{-8}$ W m^{-2} K^{-4}), T_a = air temperature at 2 m height (K), ε_s = emissivity for long-wave radiation of the surface (dimensionless fraction), and T_s = surface temperature (K). We chose ε_s as 0.95, which is a typical value for soils or plants (Campbell and Norman, 1998).

For ε_{sky} we used an empirical expression proposed by Paltridge and Platt (1976):

$$\varepsilon_{sky} = f_1\cdot T_a^2 + \frac{f_2\cdot N}{\sigma\cdot T_a^4} \tag{6.4}$$

where $f_1 = 9.35\cdot10^{-6}$ K^{-2}, f_2 = 60 W m^{-2}, and N = cloud cover (dimensionless fraction).

During nighttime the soil heat flux G is generally directed upwards and primarily determined by heat conduction in the soil (Holtslag and de Bruin, 1988). It is usually measured with a soil heat flux plate or with a set-up described in section 5.3.1 and 5.3.2. In October 2008 a measurement of soil heat flux was not available. Hence, it was calculated based on a relationship between G and R_n. The main assumption in this procedure is that the magnitude of the ratio of G to R_n is essentially a function of the vegetation cover. Although there are some empirical equations, which treat the balance of long and short wave radiation inside canopies separately (e.g. van de Griend and van Boxel, 1989), the usually exponential decay of R_n has lead to a simple conceptual model of G/R_n versus leaf area index (LAI; e.g. Choudhury et al., 1987, Kustas et al., 1993):

$$\frac{G}{R_n} = \omega \cdot e^{-\gamma \cdot LAI} \tag{6.5}$$

where ω = a proportionality factor, and γ = light extinction coefficient, which varies somewhat with vegetation type and solar zenith angle (Monteith, 1973; Ross, 1981). Choudhury et al. (1987) chose the value of ω as 0.4, which is in broad agreement with observations of Fuchs and Hadas (1972, ω = 0.34) and Idso et al. (1975, ω = 0.22 to 0.51). Depending on the crop, observed extinction coefficients range from 0.45 to 0.65 (Monteith, 1973). Here β = 0.5 was used as in Choudhury et al. (1987). During the measurement period on 9/10 October, 2008, the leaf area index of the grass was 1.9. Substituting the values for ω, β and LAI just quoted into Eq. 6.5 yields $G = 0.155 \cdot R_n$.

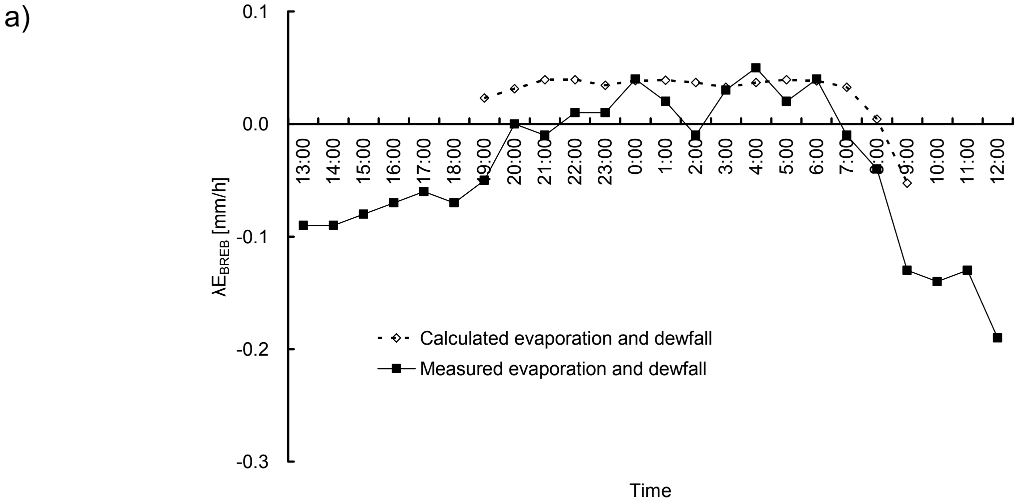
Note that the version of the PM and BREB equation used here do not include a canopy or stomatal resistance. Hence, they are only suitable for the estimation of dewfall at night and in the early morning before transpiration sets in, and for the estimation of evaporation of free water (dew) from the canopy surface in the morning. This is why computations were only carried out here for 19^{00} on 9 October to 9^{00} on 10 October, 2008, i.e. 1.5 hours after sunrise. By this time all dew accumulated over night had totally disappeared again.

6.4 Results

Figure 6.1 shows the measured and calculated hourly mass change of the lysimeter (i.e. latent heat flux, λE) during the observation period. For the reasons given above the estimates cover a shorter time span than the measurements.

Although both measured and calculated mass changes roughly follow a similar tendency, the individual hourly values differ significantly. The measured dewfall (positive λE) fluctuates considerably, while the calculations show continuous dewfall at a similar rate for each hour throughout the night. According to the measurements dewfall began at about 22^{00} and ended around 7^{00}. The calculated onset of dewfall was some three hours earlier, the end about two hours later than the measured one. Compared to the results from the lysimeter the two equations predict dewfall too early and for too long.

Figure 6.2 depicts the cumulative measured and estimated dew amount over the whole night from 19^{00} on October 9 to 7^{00} the next day. Both the BREB (Fig. 6.2a) and the PM equation (Fig. 6.2b) predict a certain amount of dewfall before any is recorded by the lysimeter.



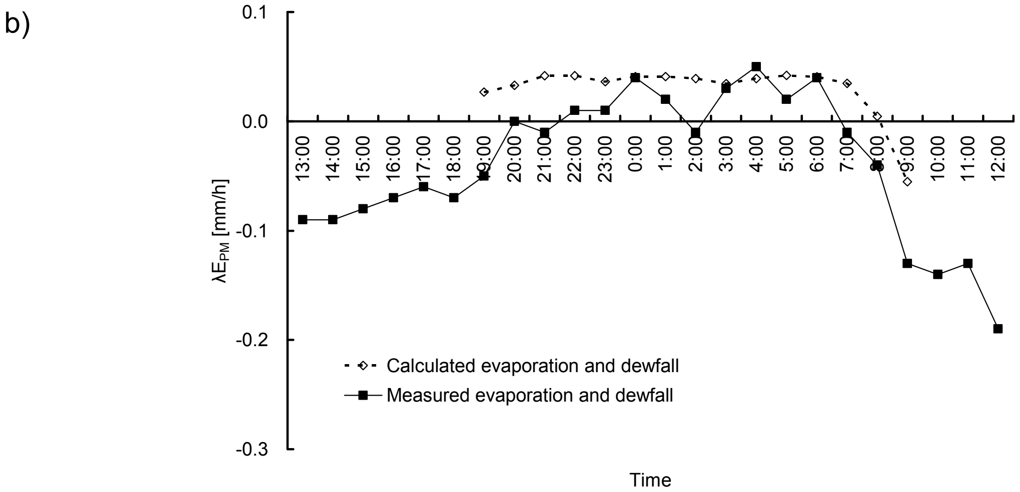


Figure 6.1: Latent heat flux (λE) measured with a lysimeter and calculated with a) the Bowen ratio energy balance equation (λE_{BREB}), and b) the Penman-Monteith equation (λE_{PM}). Hourly data for 13^{00} on 9 October to 12^{00} on 10 October, 2008. Positive λE values are dewfall, negative ones evaporation. Due to limitations of the equations (see text) the calculations cover a shorter time span than the measurements.

In addition, the estimated cumulative dewfall (0.46 mm for both equations) is markedly larger than the measured one (220 g, i.e. 0.22 mm). However, for the hours when both the lysimeter and the equations show dewfall, measured and calculated data correlate well. The correlation coefficient is very high and happens to be 0.9752 for both relationships.

The relationship between measurement and estimation has the form $y = a \cdot x + b$. Here y = cumulative estimated dewfall, x = cumulative measured dewfall, *a* = slope, and *b* = intercept. In Figure 6.2 the value of *a* for the two correlations is given as 1.42 and 1.51, respectively. This means the estimated dewfall is 1.42 and 1.51 times the measured dewfall. Both equation therefore overpredict considerably. The intercept *b* is 0.14 and 0.15, respectively, which means the equations predicted 0.14 and 0.15 mm accumulated dewfall before any was actually measured with the lysimeter.

a) b)

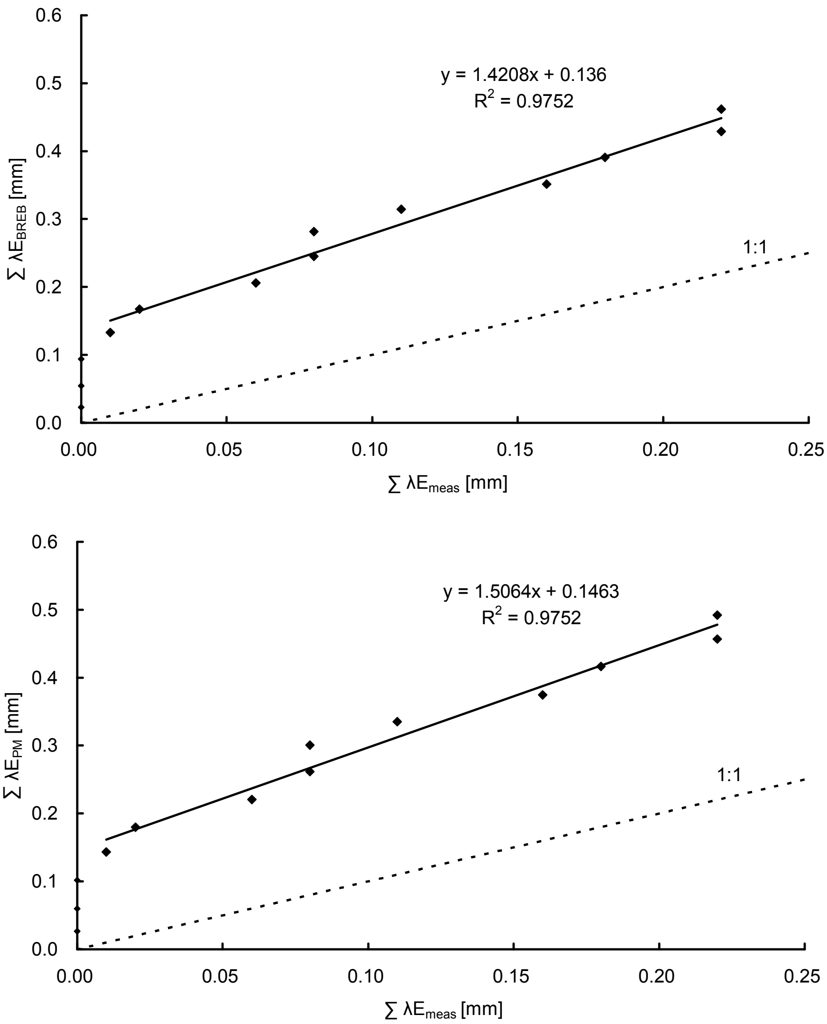
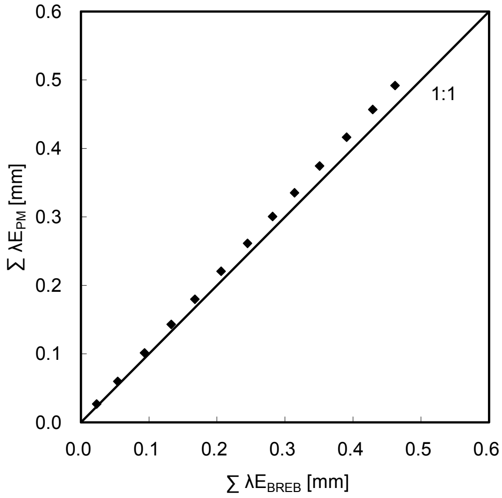


Figure 6.2: Correlation between cumulative measured dewfall ($\Sigma\lambda E_{meas}$) and cumulative dewfall calculated with a) the Bowen ratio energy balance equation ($\Sigma\lambda E_{BREB}$), and b) the Penman-Monteith equation ($\Sigma\lambda E_{PM}$). Hourly data for 19^{00} on 9 October to 7^{00} on 10 October, 2008.

In chapter 5 the BREB equation was identified to be a reliable method to estimate dew amounts. Using it as reference the correlation between dew amounts estimated with the BREB and the PM equation over the whole night is plotted in Figure 6.3a. It shows that the amounts estimated with the two models are quite close. The correlation between the estimations with the two models approximately fits the 1:1 line. However, the dewfall estimated with the PM equation is somewhat and persistently larger than the dewfall estimated with the BREB equation. This is consistent with the results in chapter 5, which also indicated that dewfall estimated with the PM equation is larger than estimates with the BREB equation.

As explained there, too, the PM equation requires the slope of the temperature - saturation vapour pressure curve (s). It is normally evaluated at air temperature, but it is more precise to evaluate it at the mean of air and surface temperature, i.e. at $T = (T_a + T_s) / 2$. This was done for Figure 6.3b, which demonstrates that by using the average of air and surface temperature to obtain s the correlation between the BREB and PM equation is closer to the 1:1 line.

a)



b)

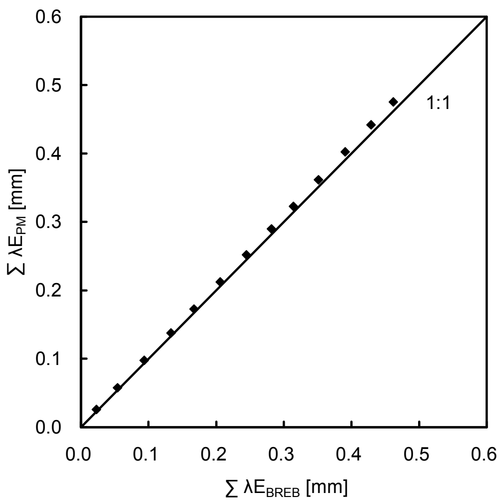


Figure 6.3: Correlation between cumulative dewfall estimated with the Bowen ration energy balance ($\Sigma\lambda E_{BREB}$) and the Penman-Monteith equation ($\Sigma\lambda E_{PM}$) with a) the slope of the saturation vapour pressure curve (s) evaluated at air temperature at 2 m height, and b) with s evaluated at the mean of air and surface temperature. Hourly data for 19^{00} on 9 October to 7^{00} on 10 October, 2008.

Recall from chapter 5 that the PM equation contains a conductance term and that there is a problem to get the proper value for it under the site conditions at Falkenberg. Hence, a bigger deviation between predictions with the BREB and PM equation would not have come as a surprise. However, also recall that the conductance term in the PM equation becomes insignificant, if the vapour pressure deficit of the air is low, i.e. the relative humidity is high. This was the case in the night of 9/10 October, 2008.

6.5 Discussion

The comparison between measured and computed dewfall presented here indicates that the estimated hourly and total dewfall is distinctly larger than the dewfall measured with the lysimeter. In addition, the estimated dewfall began earlier and the dew duration was longer than the measured one. Some possible reasons for this are given below.

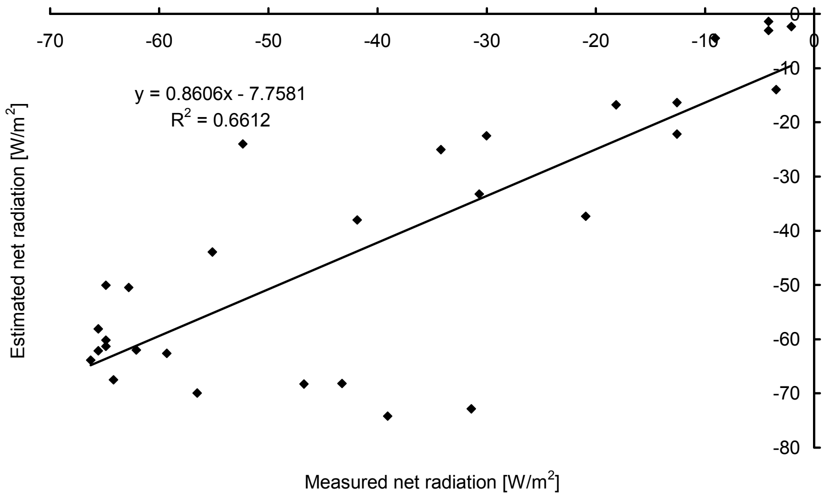
In chapter 2 the precision of the lysimeter used in this study was tested and found to be about 20 g, which is equivalent to 0.02 mm of water (dewfall). This means the lysimeter registers dewfall in steps of 0.02 mm, while the BREB and PM equation give continuous values. Hence, when the dewfall is < 0.02 mm, it may be identified by the models, but the lysimeter will not record it. This can lead to discrepancies between lysimeter data and calculations for one or several hours of light dewfall in succession, but the cumulative dewfall over a whole night should not differ by more than 0.02.

Net radiation and soil heat flux were estimated here with empirical equations, which is a big shortcoming of this study, even though these equations were proven to be reliable in some studies. Figure 6.4a shows the correlation between measured and estimated net radiation, and Figure 6.4b between measured and estimated soil heat flux for the nights of 19/20 and 20/21 November, 2009, when measured data for R_n and G as well as all the parameters needed to compute them were available. A substantial difference exists between measured and estimated R_n and G. Both variables are an input into the BREB and PM equation. Thus, if some error exists in the calculation of the R_n and G, a difference between measurement and estimate will result.

Figure 6.5 shows the time course of measured and estimated net radiation and soil heat flux from 18^{00} on 19 November to 7^{00} on 20 November, 2009. It can be seen that over the whole night soil heat flux is underestimated. Moreover, in the first part of the night the estimated net radiation is less negative than the measured one, in the middle of the night they agree quite well, while after 3^{00} the estimated net radiation is more negative than the measured values and the difference becomes larger and larger.

Net radiation and soil heat flux are important inputs into the BREB and PM equation (cf. Eq. 5.16 and 5.12). If the G entered is too small, both equations yield a latent heat flux (λE), i.e. dewfall in the case here, which is too large. Based on the evidence in Figure 6.5 it seems likely that G computed with Eq. 6.5 for the night of 9/10 October, 2008, was too small as well. This would explain why the computed dewfall was consistently and clearly higher than the measured one, as illustrated in Figure 6.1 and 6.2. If and how much R_n estimated with Eqs. 6.1 to 6.4 for the night of 9/10 October, 2008, may have been off cannot be inferred from the data in Figure 6.5. Consequently, this point cannot be elaborated on.

a)



b)

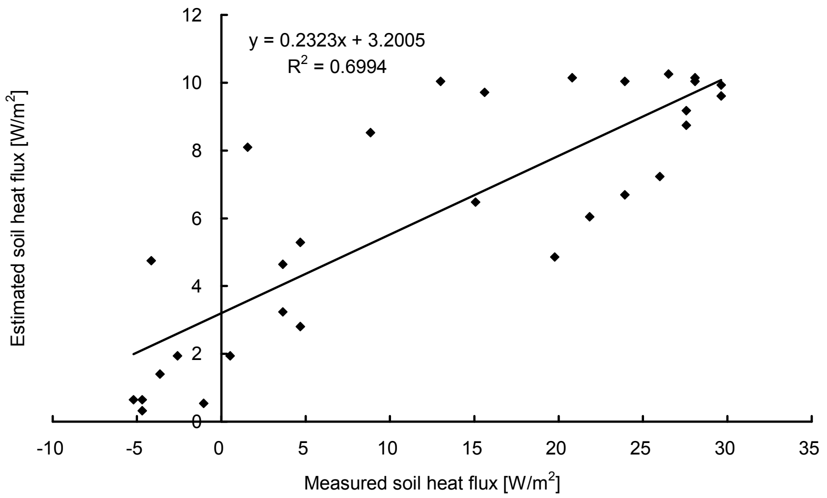


Figure 6.4: Correlation between a) measured and calculated (Eqs. 6.1 to 6.4) net radiation, and b) measured and calculated (Eq. 6.5) soil heat flux. Hourly data for 18^{00} on 19 November to 7^{00} on 20 November, 2009, and 19^{00} on 20 November to 7^{00} on 21 November, 2009. For measurement details see section 5.3.1.

Based on the above analysis, errors in the estimation of R_n and G could be the reason for the deviations between measured and computed dewfall observed in this chapter. Further measurement campaigns are planned so that eventually a full set of measured data will be available to check the performance of the BREB and PM equation against lysimeter measurements.

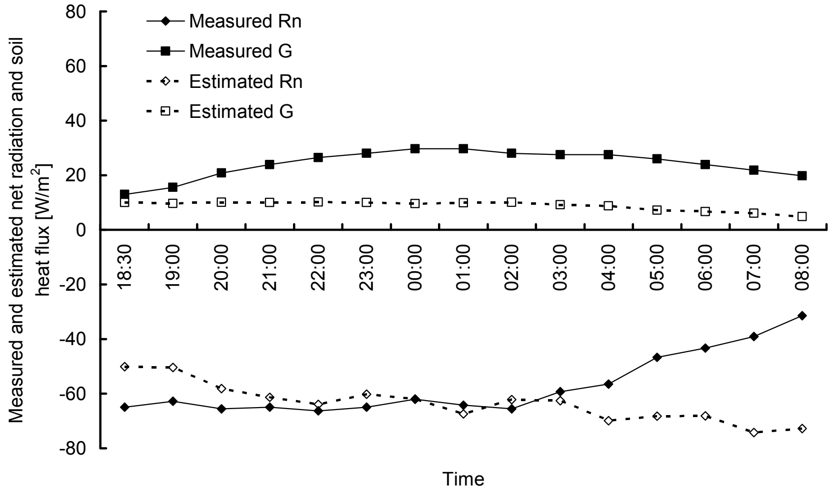


Figure 6.5: Time course of measured and calculated (Eqs. 6.1 to 6.4) net radiation (R_n), and measured and calculated (Eqs. 6.5) soil heat flux (G). Hourly data for 18^{00} on 19 November to 7^{00} on 20 November, 2009, and 19^{00} on 20 November to 7^{00} on 21 November, 2009. For measurement details see section 5.3.1.

A further potential contributor to the deviations between measured and calculated dewfall may be a horizontal transport (advection) of heat and vapour to the canopy from the surrounding area, or vice versa. This occurs when the micrometeorological conditions of the area of interest are different from those of the surrounding area and lead to horizontal gradients in air temperature and vapour pressure. Different meteorological conditions develop, if the surrounding area has a different vegetation, or if the area with the same vegetation is to small for a horizontal equilibrium to be reached (insufficient fetch). The vegetation within and around the Falkenberg research station is very heterogeneous, as already pointed out in chapter 5. So, advection may play a role. How much of a role can only be assessed, once a full set of good meteorological and lysimeter data is available so that poor data can be eliminated as a potential source of error.

Due to the discontinuity of the vegetation on and off the lysimeter, horizontal gradients are likely exist. Near the lysimeter boundary these gradients will be greatest and horizontal interactions (advection) will occur. Advection can affect the situation inside and outside the canopy, since it can bring in or take away moisture and heat and thereby influence dew formation.

6.6 Conclusions

For the assessment of the BREB and PM equation in estimating dewfall, i.e. a comparison of their predictions with dewfall measured with a lysimeter, it would have been preferable to have a complete set of measured meteorological data. However, to date no complete data set could be obtained on either of the dates it was attempted. Each time it failed for a different reason. So, for the period chosen for the comparison net radiation and soil heat flux had to estimated with empirical equations.

The estimated R_n and G were at least partly responsible for the less than perfect correlation between measured and calculated dewfall observed here. To what degree is unclear. Advection may have contributed to the disagreements as well. One needs to wait until measured data have been gathered for all variables in the two equations on the same night. A repeat of the comparison will then allow a better assessment of the equations and of the potential causes for erroneous predictions, should they still occur.

Nevertheless, the comparison here showed at least enough commonality between measured and computed dewfall to conclude that the BREB and PM equation can be applied in principle to estimate dewfall under the conditions at Falkenberg. In the literature they have already been shown to apply at other locations.

It was also shown here that the BREB and PM equation can produce very similar values under the right condition. During the observation period the relative humidity was near 100%, i.e. the vapour pressure deficit near zero. Hence, the vapour conductance, which is difficult to evaluate correctly under the site conditions at Falkenberg, had little bearing on the results, because it enters into the PM equation as a multiplier of the vapour pressure deficit. This finding also implies that with the right conductance value the two equations should give the same results for any relative humidity, especially if the slope of the saturation vapour pressure curve, which is used in the PM equation, is evaluated at the mean of air and surface temperature.

6.7 References

Campbell, G. S., Norman, J. M. (1998): An introduction to environmental biophysics, 2nd ed. Springer, New York.

Choudhury, B. J., Idso, S. B., Reginato, R. J. (1987): Analysis of an empirical model for soil heat flux under a growing wheat crop for estimating evaporation by an infrared-temperature based energy balance equation. Agric. For. Meteorol. 39, 283-297.

Evett, S. R. (2001): Water and energy balances at soil-plant-atmosphere interfaces. In: Warrick , A. W. (ed.), Soil physics companion (pp. 127-188). CRC Press, Boca Raton.

Fuchs, M., Hadas, A. (1972): The heat flux density in a non-homogeneous bare loessial soil. Bound.-Layer Meteorol. 3,191- 200.

Grimmond, C. S. B., Isard, S. A., Belding, M. J. (1992): Development and evaluation of continuously weighing mini-lysimeters. Agric. For. Meteorol. 62, 205-218.

Holtslag, A. A. M., de Bruin, H. A. R. (1988): Applied modeling of the nighttime surface energy balance over land. J. Appl. Meteorol. 27, 689-704.

Idso, S.B ., Aase, J. K., Jackson, R. D. (1975): Net radiation - soil heat flux relations as influenced by soil water content variations. Bound.-Layer Meteorol. 9,113-122.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (2000): Dew measurements along a longitudinal sand dune transect, Negev Desert, Israel. Int. J. Biometeorol. 43, 184-190.

Jones, H. G. (1992): Plants and microclimate: a quantitative approach to environmental plant physiology, 2nd ed. Cambridge University Press, Cambridge, UK.

Kustas, W. P., Daughtry, C. S. T., van Oevelen, P. J. (1993): Analytical treatment of the relationships between soil heat flux/net radiation ratio and vegetation indices. Remote Sens. Environ. 46, 319-330.

Meissner, R., Seeger, J., Rupp, H., Seyfarth, M., Borg, H. (2007): Measurement of dew, fog, and rime with a high-precision gravitation lysimeter. J. Plant Nutr. Soil Sci. 170, 335-344.

Monteith, J. L. (1973): Principles of environmental physics. Arnold, London.

Paltridge G. W., Platt, C. M. R.(1976): Radiative processes in meteorology and climatology. Elsevier, Amsterdam.

Ross, J. (1981): The radiation regime and architecture of plant stands. Junk, The Hague.

Severini, M., Moriconi, M. L., Tonna, G., Olivieri, B. (1984): Dewfall and evapotranspiration determination during day- and nighttime on an irrigated lawn. J. Clim. Appl. Meteorol. 23, 1241-1246.

Sharma, M. L. (1976): Contribution of dew in the hydrologic balance of a semi-arid grassland. Agric. Meteorol. 17, 321-331.

Slatyer, R. O., McIlroy, I. C. (1961): Practical Microclimatology. Unesco, Paris.

van de Griend, A. A., van Boxel, J. H. (1989): Water and surface energy balance model with a ultilayer canopy representation for remote sensing purposes. Water Resour. Res. 25, 949-971.

7. Concluding Remarks

This dissertation focused on (1) the effect of various meteorological factors on dewfall, (2) the precision of a certain model of a weighable gravitation lysimeter, (3) the effect of vegetation type and growth stage on dewfall, (4) the assessment of four methods to compute dewfall, and (5) a comparison between measured and computed dewfall. These points shall be briefly addressed again below.

7.1 Effect of various meteorological factors on dewfall

Dewfall represents a latent heat flux. Latent heat flux (λE) is a component of the surface energy balance, together with net radiation, sensible heat flux and soil heat flux. Hence, the best way to assess which meteorological factors influence dew formation and how is to analyse the energy balance equation for a surface.

An in depth analysis of the equation revealed that the meteorological factors affecting dew formation are: air temperature (T_a), cloud cover (N), wind speed (u), soil heat flux (G), and relative humidity (h_r). The surface temperature (T_s) is not an independent meteorological factor, but determined by aforementioned ones. It therefore represents the dependent variable in energy balance equation, while T_a, N, u, G and h_r are the independent variables.

A computer programme was written to calculate the energy balance. With this programme the surface temperature (T_s) was obtained using an iterative approach. T_a, N, u, G and h_r are inputs, and LE the ultimate output. This programme was used to analyse the effect of individual meteorological factors on dew formation.

The analysis indicated that the effect of an individual factor on dew formation also depends on the other meteorological factors. Furthermore, it showed that for every factor there is a threshold above or below which it there is no dewfall, if all other factors are kept constant. In addition, the relationship between air temperature and dewfall, and between wind speed and dewfall turned out to be curvilinear. The pattern is that dewfall initially increases with air temperature or wind speed, but then decreases again once air temperature or wind speed exceed a certain value. In contrast, there is a linear relationship between N, G, and h_r on the one hand, and λE on the other. Dewfall always decreases with rising cloud cover or soil heat flux, while it always increases as relative humidity goes up. Lastly, the potential influence of T_a, N, u, G and h_r was found to have roughly the same magnitude.

The effect of the various meteorological factors on dewfall was analysed quite thoroughly here. Hence, a need for further studies in that direction is not apparent.

7.2 Precision of a weighable gravitation lysimeter

Small mass inputs such as precipitation in the form of dew, fog or rime can be determined with a lysimeter, as long as its weighing precision is good enough. There are different lysimeter models. To test whether the kind of lysimeter used in this study is suitable for the measurement of dew, fog or rime, two aspects of its weighing precision were investigated: 1) the accuracy and stability a given mass change is measured with, and 2) the effect of load position on the measurement.

On the lysimeter used here a rubber collar can be placed across the cleavage between lysimeter vessel and pit casing to prevent water and debris from entering the pit. It is attached to the casing and not supposed to touch the vessel, However, it does at a few points, which may affect the weighing precision. Hence, the experiments were conducted twice, once without the collar, which can easily be removed, and once with it.

To find the smallest mass change the system can reliably detect and the stability of its measurement, weights of 500, 200, 100, 50, 20 and 10 g were subsequently placed at the centre of the lysimeter for 22 minutes and then removed again. The data logger was set to read the lysimeter weight every 10 seconds and to store a mean weight for one minute intervals. To investigate if the smallest mass change the lysimeter can reliably detect depends on the position it occurs at, the aforementioned weights were subsequently placed at 10, 23, 55, 77 and 100 cm along two perpendicular lines through the centre of the lysimeter, whose diameter is 113 cm.

The trials demonstrated that under favourable environmental conditions mass changes down to 20 g can be discerned with good accuracy and stability. The mean measured mass varied somewhat between load positions for all added weights, but not in a systematic fashion. Hence, there is no effect of load position on weighing precision. The rubber collar was found to seriously lower weighing precision in terms of accuracy and stability of the measurements. If it touches the lysimeter it exerts forces on the vessel, which distort the true weight. If the highest possible weighing precision is required, the collar should be left off or a new design developed.

Strong wind can lead to pressure on a lysimeter, and large changes in temperature can affect the weighing mechanism. Wind and temperature gradients therefore affect the weight measurement. We found that the best weighing precision is achieved in the late evening hours (and presumably at night), when wind speed is usually low and temperature changes are more gradual.

Preliminary results show that wind speeds $< 1\ m\ s^{-1}$ do not disturb the precision of the lysimeters here, but that with increasing wind speed the effect on weighing precision increases. It is worthwhile to study the effect of wind on weighing precision in more detail. Ideally this may lead to a method to adjust lysimeter weights measured in windy weather or a way to protect the lysimeter from the influence of wind. A detailed study of the effect of temperature on the weighing mechanism is worthwhile for the same reasons.

7.3 Effect of vegetation type and growth stage on dewfall

Data in the literature suggest that on nights with dew the mean amount varies from 0.03 to 0.24 mm, while the highest nightly values are 0.54 to 0.61 mm. This means that the lysimeter used here is a suitable tool for measuring dewfall, since its weighing precision of 20 g (0.02 mm) is high enough to detect such amounts.

Using four high precision weighing lysimeters the amount and temporal distribution of dewfall on grass, maize and winter barley was determined for 2004 and 2005 to quantify the contribution of dewfall to the water balance of the region and to assess how dewfall is affected by the vegetation cover.

Our results indicate that dewfall makes a notable contribution to the water balance of crops and grass in northern Germany, since observed dewfall ranged from 27.1 to 31.8 mm per year, which amounted to 5.5 - 6.9% of the annual rainfall. In several months of the study period dewfall was > 20% of the monthly precipitation.

Additionally, the study also illustrated that the vegetation cover affects dew formation. There was consistently more dewfall on covered than on bare lysimeters. In addition, dewfall increased with crop growth, reflected in the rising frequency and amount of dewfall on growing crops compared to a continuous grass cover, and then fell again after harvest.

Regarding the reasons why vegetation and its state of development affects dew amounts we just gave a qualitative explanation. It was pointed out that there is more dewfall as the height and density of the vegetation increases, mainly because heat transfer from the soil to the canopy declines. Less of this heat transfer translates into more dewfall.

To quantify this, data on ventilation, radiation, vapour pressure, air, leaf temperature and soil temperatures, as well as plant characteristics (e.g. height and leaf area index) should be collected in and around the canopy, in addition to dew amounts. The latter should be broken down for the different layers in the canopy.

7.4 Four methods to compute dewfall

Dew amounts can be measured with various gauges, but there is still not a standard way of measurement. Lysimeters are a very promising tool for quantifying dewfall, but they are not widespread because of their high cost.

If measurements are not available, one alternative is to compute dewfall from meteorological data and under consideration of the properties of the surface in question. The most widely used tools to compute dew amounts are the equation for turbulent vapour transport (TVT), the energy balance equation (EB), the Bowen ratio energy balance equation (BREB), and the Penman-Monteith equation (PM).

Since the TVT and EB equation are really just different sides of the energy balance, results of computations with the two equations should agree. However, here dewfall estimated with the EB equation differed significantly from that estimated with the TVT equation, and from the results with the PM and BREB equation as well. The most likely reason for the deviations was erroneous values of the heat and vapour conductance. Hence, the EB, TVT and, to a lesser degree, the PM equation could not be used successfully without adjusting the conductance (g). After adjustment of g the results of all four equations agreed very well.

The BREB equation was identified as the best choice, because it does not contain a g term. The second best alternative is the PM equation, where g turns up in connection with the vapour pressure deficit. This is often near zero when dew falls, so that a wrong g has little effect. The PM equation is also the only one which does not require the temperature of the surface, which is usually not available. It is the only choice then.

The precise reason for the difficulties in getting proper conductance values in this study is not certain. The most likely one it that the wind profile could not equilibrate with the vegetation on the lysimeter, because it was not surrounded by a large enough area with the same surface conditions (vegetation) to do so. Yet, to be valid the equation the conduction is evaluated with demands this equilibrium.

To settle this issue requires the measurement of wind, temperature and vapour pressure profiles on the lysimeter at a various distances and directions around it.

This study named the BREB equation as the best method to estimate dewfall. However, the way the equation was used here assumed that the vapour pressure at the surface is a saturation. This is probably a reasonable assumption, if one is concerned with dewfall. If the vapour pressure is less than at saturation, the estimates with the BREB equation would be too low. There was no evidence of this here. Nevertheless, the assumption may not always hold, even during dew events. This should be investigated. To do this requires the measurement of the vapour pressure at the surface, besides its temperature.

An alternative to measuring vapour pressure and temperature at the surface and at 2 m height would be to measure it at two different heights above the canopy. This is the usual procedure to get the data for calculations with the BREB equation. The fact that the PM and BREB equation agreed well (see next section) suggests that our set-up works, too. A comparison between the usual and our experimental procedure would illuminate how good they are, respectively.

7.5 Comparison between measured and computed dewfall

In chapter 5 it was found that the deviations between the estimates with the various equations was mainly due to poor conductance values. When dew occurs on a surface, the vapour pressure deficit between the surface and the surrounding air is usually small so that the influence of an incorrect conductance in the PM equation is small. A conductance is not involved in the BREB equation. Therefore, these two equations appear to be the most appropriate methods to estimate dewfall. However, computations with these models need to be validated by comparing them to direct measurements.

So far only a preliminary comparison could be carried out, because to date no complete set of meteorological data was available. For the period considered for the comparison here net radiation (R_n) and soil heat flux (G) had to be estimated with empirical equations. Evidence was presented that these estimates of R_n and G may have been imperfect. It was argued that this was probably a major cause for the observed deviations between measured and computed dewfall. It was further argued that advection may have contributed to the disagreements as well.

The equations indicated dewfall earlier and for longer than the lysimeter record. Furthermore, the computed dewfall was always markedly higher than the one measured with the lysimeter, namely 1.4 to 1.5 times higher during the hours when both the equations and the lysimeter showed dewfall. Also, the calculations yielded continuous dewfall at a roughly similar rate throughout the night, while the measured values varied significantly. Nevertheless, measured and calculated data exhibited sufficient commonalities to suggest that the BREB and PM equation, which agreed very well with each other here, can be used to estimate dewfall at Falkenberg. In the literature they were already shown to work well at other places.

In the future data need to be gathered in the same night for all variables in the two equations, parallel to lysimeter data, of course. All necessary instruments are now up and running. This will allow to much more accurately compare measurements and estimates of dewfall. A repeat of the comparison will then yield a better assessment of the equations and of the potential causes for erroneous predictions, should they still occur.

One further point, which deserves more attention, is advection, It was not taken into account in our experimental set-up or in the equations used. In the studies presented here the surface fluxes were estimated for grass on a lysimeter surrounded by rather heterogeneous vegetation. Due to the discontinuity of the surface inside and outside the lysimeter, horizontal gradients in temperature and vapour pressure were likely to exist. Near the surface boundaries these gradients are greatest and horizontal interactions (advection) will occur. To take them into account it is necessary to consider a three-dimensional energy balance.

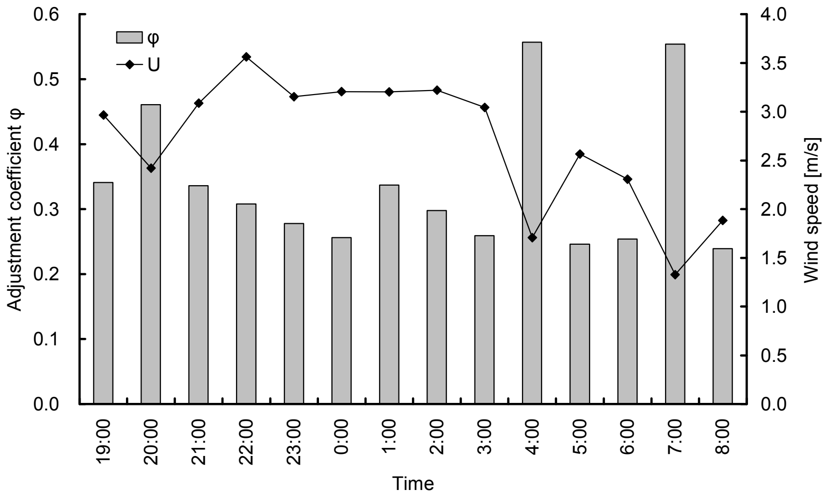
Advection is associated with wind speed. The higher the wind speed, the more heat and vapour can be transferred to or from the canopy. Figure 7.1 shows the value of the adjustment coefficients φ determined in chapter 5 for the nights of 19/20 and 20/21 November, 2009. It changes with time and there is, to some extent, a relationship between φ and wind speed. The coefficient is relatively large at low wind speed, for example at 20^{00}, 4^{00} and 7^{00} in the first night, and at 7^{00} in the second night. In contrast, the coefficient is relatively small at high wind speed, such as at 3^{00} of the second night. However, φ is not only affected by wind speed, since φ is not always the same at same wind speed.

The effect of advection should be investigated further by measuring the difference in wind, temperature and vapour pressure between the canopy surface on the lysimeter and the surrounding area. This should be done at a various distances and directions around the lysimeter and at two different heights, at least. This is the same procedure suggested in section 7.4 to probe whether the wind profile above the lysimeter has reached an equilibrium. With this kind of data at hand the equilibrium and the advection problem can be addressed.

Should it turn out that one or the other problem, or both, cannot be rectified, measurements and computations of dewfall should be repeated in a large field with homogenous vegetation, where insufficient fetch would then not be a problem. From the boundary of a field a certain distance is required before the wind profile reaches an equilibrium again. Ideally, this is where a lysimeter should be placed. There are methods to estimate this distance.

As in Falkenberg, lysimeters are usually situated in experiment stations, where the surrounding vegetation is rather heterogeneous. It may therefore be difficult to find a large lysimeter within a sufficiently large area of the same vegetation. Large lysimeter cannot be moved around either. A possible solution is to place a micro-lysimeter, which can be installed and removed easily, at the required location. Such lysimeters have been used successfully in past dew studies.

a)



b)

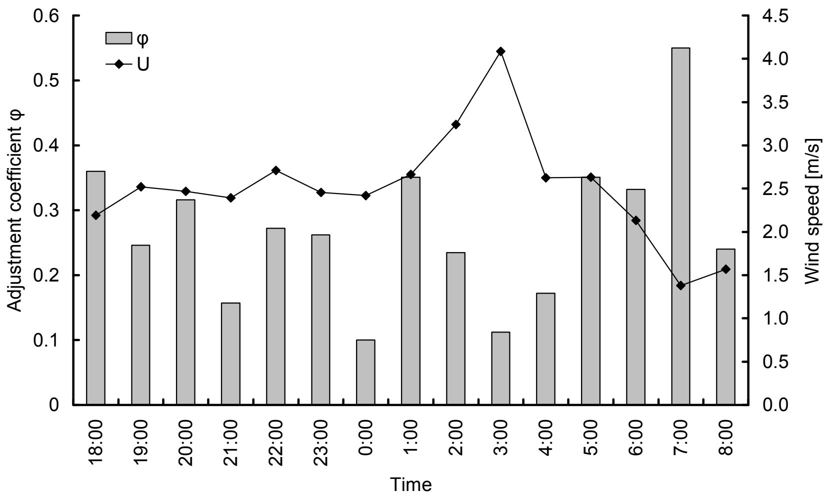


Figure 7.1: Hourly adjustment coefficient (φ) for the heat and vapour conductance in relation to wind speed for the period of a) 19^{00} on 19 November to 8^{00} on 20 November, 2009, and b) 18^{00} on 20 November to 8^{00} on 21 November, 2009.

If the measurement of dewfall and of the necessary meteorological and plant parameters are repeated at a site with sufficient fetch for the wind, temperature and vapour pressure profile to reach an equilibrium, the problems with advection and non-equilibrated profiles would disappear. This would allow a better assessment of the performance of the equations.

Note that advection and non-equilibrated profiles are not only a problem for studies on dew formation, but also on evaporation and transpiration. All three are a latent heat flux determined by the same meteorological and plant parameters, and all three are described by the same equations (though in slightly different versions).

8. References (all chapters)

Acostav Baladón A. N., Gioda, A. (1991): L'importance des precipitations occultes sous les tropiques secs. Sécheresse 2, 132-135.

Allen, R. G., Fisher, D. K. (1990): Low-cost electronic weighing lysimeters. Trans. ASAE 33, 1823-1833.

Alnaser, W. E., Barakat, A. (2000): Use of condensed water vapour from the atmosphere for irrigation in Barhein. Appl. Energ. 65, 3-18.

Ashbel, D. (1949): Frequency and distribution of dew in Palestine. Geogr. Rev. 39, 291-297.

Aslyng, H. C. (1965): Rain, snow and dew measurements. Acta Agric. Scand. 15, 275-283.

Atzema, A. J., Jacobs, A. F. G., Wartena, L. (1990): Moisture distribution within a maize crop due to dew. Netherlands J. Agric. Sci. 38, 117-129.

Awanou, C. N., Hazoume, R. P. (1997): Study of natural condensation of atmospheric humidity. Renew. Energ. 10, 19-34.

Baier, W. (1966): Studies on dew formation under semi-arid conditions. Agric. Meteorol. 3, 103-112.

Barradas V. L, Glez-Medellin M. G. (1999): Dew and its effect on two heliophile understory species of a tropical dry deciduous forest in Mexico. Int. J. Biometeorol 43, 1-7.

Bass, B., Savdie, I., Gillespie, T. J. (1991): Simulation of leaf wetness duration for field corn. Agric. For. Meteorol. 57, 69-84.

Bergmann J. C. (1998): A physical interpretation of von Karman's constant based on asymptotic considerations - a new value. J. Atmos. Sci. 55, 3403-3407.

Beysens, D., Muselli, M., Nikolayev, V., Narhe, R., Milimouk, I. (2005): Measurement and modelling of dew in island, coastal and alpine areas. Atmos. Res. 73, 1-22.

Biftu, G. F., Gan, T. Y. (2000): Assessment of evapotranspiration models applied to a watershed of Canadian Prairies with mixed land-uses. Hydrol. Process. 14, 1305-1325.

Bootsma, A., Gillespie,T. J., Sutton, J. C. (1973): Germination of Phyllosticta maydis Conidia in an incubation chamber with control of high relative humidities. Phytopathology 63, 1157-1161.

Bowen, I. S. (1926): The ratio of heat losses by conduction and by evaporation from any water surface. Phys. Rev. 27, 779-787.

Braden, H. (1982): Simulationsmodell für den Wasser-, Energie- und Stoffhaushalt in Pflanzenbeständen. Berichte des Instituts für Meteorologie und Klimatologie der Universität Hannover, 23, pp. 149. Institut für Meteorologie und Klimatologie der Universität Hannover, Hannover.

Brutsaert, W. (1982): Evaporation into the atmosphere: theory, history, and applications. Reidel, Dordrecht.

Buck, A. L. (1981): New equations for computing vapor pressure and enhancement factor. J. Appl. Meteorol. 20, 1527-1532.

Burrage, S. W. (1972): Dew on wheat. Agric. Meteorol. 10, 3-12.

Businger, J. A., Wyngaard, J. C., Izuml, Y., Bradley, E. F. (1971): Flux-profile relationships in the atmospheric surface layer. J. Atmos. Sci. 28, 181-189.

Buttler, D. R. (1980): Dew and thermal lag: measurements and an estimate of wetness duration on cocoa pods. Q. J. R. Meteorol. Soc. 106, 539-550.

Campbell, G. S., Norman, J. M. (1998): An introduction to environmental biophysics, 2nd ed. Springer, New York.

Camuffo, D., Giorio, R. (2003): Quantitative evaluation of water deposited by dew on monuments. Bound.-Layer Meteorol. 107, 655-672.

Choudhury, B. J., Idso, S. B., Reginato, R. J. (1987): Analysis of an empirical model for soil heat flux under a growing wheat crop for estimating evaporation by an infrared-temperature based energy balance equation. Agric. For. Meteorol. 39, 283-297.

Collins, B. G. (1961): A standing dew meter. Meteorol. Mag. 90, 114-117.

Collins, B. G., Taylor, R. J. (1961): Conditions governing the onset of dew on large leaves. Australian J. Appl. Sci. 12, 23-29.

Crowe, M. J., Coakley, S. M., Emge, R. G. (1978): Forecasting dew duration at Pendleton, Oregon, using simple weather observations. J. Appl. Meteorol. 17, 1482-1487.

Danin, A., Bar-Or, Y., Dor, I., Yisraeli, T. (1989): The role of cyanobacteria in stabilization of sand dunes in southern Israel. Ecol. Mediterr. 15, 55-64.

Davie, T. (2003): Fundamentals of hydrology. Routledge, London.

Davis, P. A. (1958): An investigation of a method for predicting dew duration. Arch. Meteorol. Geophys. Bioklimatol., Ser. A., 10, 66-93.

Degen, A., Leeper, A., Shachak, M. (1992): The effect of slope direction and population density on water influx in a desert snail, Trochoidea seetzenii. Funct. Ecol. 6,160-166.

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

Duvdevani, S. (1964): Dew in Israel and its effect on plants. Soil Sci. 98, 14-21.

Evenari, M., Shanan, L., Tadmor, N. (1971): The Negev. The challenge of a desert. Harvard University Press, Cambridge, Massachusetts.

Evenari, M., Shanan, L., Tadmor, N. (1982): The Negev: the challenge of a desert. Harvard University Press, Cambridge, Mass.

Evett, S. R. (2001): Water and energy balances at soil-plant-atmosphere interfaces. In: Warrick , A. W. (ed.), Soil physics companion (pp. 127-188). CRC Press, Boca Raton.

Francl, L. J., Panigrahi, S. (1997): Artificial neural network models of wheat leaf wetness. Agric. For. Meteorol. 88, 57-65.

Fritschen, L. J., Doraiswamy, P. (1973): Dew: an addition to the hydrologic balance of Douglas fir. Water Resour. Res. 9, 891-894.

Fritschen, L. J., Cox, L., Kinerson, R. (1973): A 28-meter douglas fir in a weighing lysimeter. Forest Sci. 19, 256-261.

Fritzsche, G. (1934): Vergleichende Messungen mit Leickschen Tauplatten. Bioklim. Beibl. Meteorol. Z. 1, 66-73.

Fuchs, M., Hadas, A. (1972): The heat flux density in a non-homogeneous bare loessial soil. Bound.-Layer Meteorol. 3,191- 200.

Garratt, J. R., Segal, M. (1988): On the contribution of atmospheric moisture to dew formation. Bound.-Layer Meteorol. 45, 209-236.

Gates, D. M. (1965):. Heat transfer in plants. Sci. Am. 213,76-84.

Gavilan, P., Berengena, J. (2007): Accuracy of the Bowen ratio-energy balance method for measuring latent heat flux in a semiarid advective environment. Irr. Sci. 25, 127-140.

Gelbke, W. (1955): Untersuchungen zur Methodik der Taumessung anhand dreijähriger Taumessreihen in Greifswald. Abh. Meteorol. Hydrol. Dienst DDR 5, 40-47.

Getz, R. R. (1981): The determination of wetting duration in plant canopies. In: American Meteorological Society, 15th Conference on Agriculture and Forest Meteorology and 5th Conference on Biometeorology, April 1 - 3, 1981 (pp. 155-157). Soc., Boston, Mass.

Gholam, A. B., Mohammad, J. K. (2002): A large electronic weighing lysimeter system: design and installation. J. Am. Water Resour. Ass. 38, 1053-1060.

Gillespie, T. J., Barr, A. (1984): Adaptation of a dew estimation scheme to a new crop and site. Agric. For. Meteorol. 31, 289-295.

Girona, J., Marsal, J., Mata, M., del Campo, J. (2004): Pear crop coefficients obtained in a large weighing lysimeter. In: Proceedings of the IVth International Symposium on Irrigation of Horticultural Crops (pp. 277-282). ISHS Acta Horticulturae 664.

Goudriaan, J. (1977): Crop micrometeorology: a simulation study. Centre for Agricultural Publishing and Documentation, Wageningen.

Grimmond, C. S. B., Isard, S. A., Belding, M. J. (1992): Development and evaluation of continuously weighing mini-lysimeters. Agric. For. Meteorol. 62, 205-218.

Haines, D. A. (1980): The adjective classification for dew: what does it mean? National Weather Digest 5, 47-50.

Hicks, B. B. (1983): A study of dewfall in an arid region: an analysis of Wangara data. Q. J. R. Meteorol. Soc. 109, 900-904.

Hiltner, E. (1930): Der Tau und seine Bedeutung für den Pflanzenbau. Eine Theorie über die physiologische Bedeutung der Wasseraufnahme durch oberirdische Pflanzenorgane. Wiss. Arch. Landw., Abt. A, 3, 1-70.

Hirst, J. M. (1954): A method for recording the formation and persistence of water deposits on plants shoots. Q. J. R. Meteorol. Soc. 80, 227-231.

Hirst, J. M. (1957): A simplified surface wetness recorder. Plant Path. 6, 57-61.

Hofmann, G. (1952): Zur Methodik der Taumessung. Ber. Deut. Wetterd. US-Zone 38, 360-364.

Hofmann, G. (1955): Die Thermodynamik der Taubildung. Ber. Deut. Wetterd. 3, 1-45.

Holtslag, A. A. M., de Bruin, H. A. R. (1988): Applied modeling of the nighttime surface energy balance over land. J. Appl. Meteorol. 27, 689-704.

Horino, O., El-Nemr, F., Yousef, M., Niki, H. (1990): Factors of the rice blast outbreak in Nile Delta, Egypt. JARQ-JPN Agr. Res. Q. 23, 176-181.

Howell, T. A., McCormick, R. L., Phene, C. J. (1985): Design and installation of large weighing lysimeters. Trans. ASAE 28, 106-117.

Howell, T. A., Schneider, A. D., Jensen, M. E. (1991): History of lysimeter design and use for evapotranspiration measurements. In: Proceedings of the International Symposium on Lysimetry, July 23-25: Lysimeters for evapotranspiration and environmental measurements (pp. 1-9). Honolulu, Hawaii.

Hughes, R. N., Brimblecombe, P. (1994): Dew and guttation: formation and environmental significance. Agric. Forest Meteorol. 67, 173-190.

Hunsaker, D. J., Pinter, P. J. jr., Cai, H. (2002): Alfalfa basal crop coefficients for FAO-56 procedures in the desert regions of the southwestern U. S. Trans. ASAE 45, 1799-1815.

Idso, S.B ., Aase, J. K., Jackson, R. D. (1975): Net radiation - soil heat flux relations as influenced by soil water content variations. Bound.-Layer Meteorol. 9,113-122.

Inclán, M. G., Forkel, R. (1995): Comparison of energy fluxes calculated with the Penman-Monteith equation and the vegetation models SiB and Cupid. J. Hydrol. 166, 193-211.

Jacobs, A. F. G., van Pul, W. A. J., van Dijken, A. (1990): Similarity moisture dew profiles within a corn canopy. J. Appl. Meteorol. 29, 1300-1306.

Jacobs, A. F. G., van Pul, A., El-Kilani, R. M. M. (1994): Dew formation and the drying process within a maize canopy. Bound.-Layer Meteorol. 69, 367-378.

Jacobs, A. F G., van Boxel, J. H., Nieveen, J. (1996): Nighttime exchange processes near the soil surface of a maize canopy. Agric. Forest Meteorol. 82, 155-169.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (1999): Dew deposition and drying in a desert system: a simple simulation model. J. Arid Environ. 42, 211-222.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (2000): Dew measurements along a longitudinal sand dune transect, Negev Desert, Israel. Int. J. Biometeorol. 43, 184-190.

Jacobs, A. F. G., Heusinkveld, B. G., Kruit, R. J. W., Berkowicz, S. M. (2006): Contribution of dew to the water budget of a grassland area in the Netherlands. Water Resour. Res. 42, W03415.

Jacobs, A. F. G., Heusinkveld, B. G., Berkowicz, S. M. (2008): Passive dew collection in a grassland area, The Netherlands. Atmos. Res. 87, 377-385.

Janssen, L. H. J. M., Römer, F. G. (1991): The frequency and duration of dew occurrence over a year. Tellus B 43, 408-419.

Jia, X., Dukes, M., Jacobs, D. J. M., Irmak, S. (2006): Weighing lysimeter for evapotranspiration research in humid environment. Trans. ASABE 49, 401-412.

Jones, H. G. (1992): Plants and microclimate: a quantitative approach to environmental plant physiology, 2nd ed. Cambridge University Press, Cambridge, UK.

Kalthoff, N., Fiebig-Wittmaack, M., Meißner, C., Kohler, M., Uriarte, M., Bischoff-Gauß, I., Gonzales, E. (2006): The energy balance, evapo-transpiration and nocturnal dew deposition of an arid valley in the Andes. J. Arid Environ. 65, 420-443.

Kappen, L., Lange, O. L., Schulze, E.-D., Buschbom, U., Evenari, M. (1980): Ecophysiological investigations on lichens of the Negev Desert. VII. The influence of the habitat exposure on dew imbibition and photosynthetic productivity. Flora 169, 216-229.

Keller, H. (1933): Über Taumessungen im ariden Hochland Transvaals. Meteorol. Z. 50, 321-324.

Kerr, J. P., Beardsell, M. F. (1975): Effect of dew on leaf water potentials and crop resistances in a Paspalum pasture. Agron. J. 67, 596-599.

Kessler, O. W. (1939): Der Tauschreiber Kessler-Fuess. Bioklim. Beibl. Meteorol. Z. 6, 23-26.

Khare, P., Singh, S. P., Kumari, K. M., Kumar, A., Srivastava, S. S. (2000): Characterization of organic acids in dew collected on surrogate surfaces. J. Atmos. Chem. 37, 231-244.

Kidron, G. J., Yaalon, D. H., Vonshak, A. (1999): Two causes for runoff initiation on microbiotic crusts: hydrophobicity and pore clogging. Soil Sci. 164, 18-27.

Kidron, G. J., Yair, A., Danin, A. (2000): Dew variability within a small arid drainage basin in the Negev Highlands, Israel. Q. J. R. Meteorol. Soc. 126, 63-80.

Kirkham, R. R., Gee, G. W., Jones, T. L. (1984): Weighing lysimeters for long-term water balance investigations at remote sites. Soil Sci. Soc. Am. J. 48, 1203-1205.

Knoche, W. (1939): Notes on the dew. An. Soc. Cient. Argent. 127, 150-151.

Kustas, W. P., Daughtry, C. S. T., van Oevelen, P. J. (1993): Analytical treatment of the relationships between soil heat flux/net radiation ratio and vegetation indices. Remote Sens. Environ. 46, 319-330.

Lange, O. L., Kidron, G., Büdel, B., Meyer, A., Kilian, E., Abeliovich, A. (1992): Taxonomic composition and photosynthetic characteristics of the “biological soil crusts” covering sand dunes in the western Negev Desert. Funct. Ecol. 6, 519-527.

Lange, O. L., Belnap, J., Reichenberger, H. (1998): Photosynthesis of the cyanobacterial soil-crust lichen Collema tenax from arid lands in southern Utah, USA: role of water content on light and temperature responses of CO_2 exchange. Funct. Ecol. 12, 195-202.

Larmuth, J., Harvey, H. J. (1978): Aspects of the occurrence of desert plants. J. Arid Environ. 1, 129-133.

Lehmann, P., Schanderl, H. (1942): Tau und Reif, Pflanzenwetterkundliche Untersuchungen. Deutsches Reich, Reichsamt für Wetterdienst, Wissenschaftliche Abhandlungen, Bd. IX, Nr. 4. Springer, Berlin.

Leick, E. (1932): Zur Methodik der relativen Taumessung. Bot. Cent.bl. Beih. 49, 160-189.

Lhomme, J .P., Jimenez, O. F. (1992): Estimating dew duration on banana and plantain leaves from standard meteorological observations. Agric. For. Meteorol. 62, 263-274.

Liu, W. J., Zeng, J. M., Wang, C. M., Li, H. M., Duang, W. P. (2001): On the relationship between forests and occult precipitation (dew and fog precipitation). J. Nat. Resour. 16, 571-575

Li, X.Y. (2002): Effect of gravel and sand mulches on dew deposition in the semiarid region of China. J. Hydrol. 260, 151-160.

Lloyd, M. (1961): The contribution of dew to the summer water budget of northern Idaho. Bull. Am. Meteorol. Soc. 42, 572-580.

Long, I. F. (1958): Some observations on dew. Meteorol. Mag. 87, 161-168.

Luo, W., Goudriaan, J. (1999): Effects of altering water temperature on leaf wetness in paddy rice crops. Agric. For. Meteorol. 97, 33-42.

Luo, W., Goudriaan, J. (2000): Dew formation on rice under varying durations of nocturnal radiative loss. Agric. For. Meteorol. 104, 303-313.

Luo, W., Goudriaan, J. (2000a): Measuring dew formation and its threshold value for net radiation loss on top leaves in a paddy rice crop by using the dewball: a new and simple instrument. Int. J. Biometeorol. 44, 167-171.

Madeira, A. C., Gillespie, T. J., Duke, C. L. (2001): Effect of wetness on turfgrass canopy reflectance. Agric. Forest Meteorol. 107, 117-130.

Madeira, A. C., Kim, K. S., Taylor, S. E., Gleason, M. L. (2002): A simple cloud-based energy balance model to estimate dew. Agric. For. Meteorol. 111, 55-63.

Mäde, A. (1954): Über die Höhe des nächtlichen Taufalles im mitteldeutschen Trockengebiet. Wiss. Z. Luther Univ. Halle, Math.-Naturw. 4, 185-190.

Maki, T. (1975): Interrelationships between zero- plane displacement, aerodynamic roughness length and plant canopy height. J. Agric. Meteorol. 31, 7-15.

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

Malone, R. W., Bonta, J. V., Stewardson, D. J., Nelsen,T. (2000): Error analysis and quality improvement of the Coshocton weighing lysimeters. Trans. ASAE 43, 271-280.

Marek, T. H., Schneider, A. D., Howell, T. A., Ebeling, L. L. (1988): Design and construction of large weighing monolithic lysimeters. Trans. ASAE 31, 477-484.

Marek, R., Straub, J. (2001): Analysis of the evaporation coefficient and the condensation coefficient of water. Int. J. Heat Mass Tran. 44, 39-53.

Marek, T. H., Piccinni, G., Schneider, A. D., Howell, T. A., Jett, M., Dusek, D. A. (2006): Weighing lysimeters for the determination of crop water requirements and crop coefficients. Appl. Eng. Agric. 22, 851-856.

McFarland, M. J., Worthington, J. W., Newman, J. S. (1983): Design, installation and operation of a twin weighing lysimeter for fruit trees. Trans. ASAE 26, 1717-1721.

Meissner, R., Seeger, J., Rupp, H., Schonert, P. (1999): Estimating the effects of set-aside on water quality: scaling-up of lysimeter studies. Land Degrad. Develop. 10, 13-20.

Meissner, R., Seeger, J., Rupp, H., Seyfarth, M., Borg, H. (2007): Measurement of dew, fog, and rime with a high-precision gravitation lysimeter. J. Plant Nutr. Soil Sci. 170, 335-344.

Meshkat, M., Warner, R. C., Walton, L. R. (1999): Lysimeter design, construction, and instrumentation for assessing evaporation from a large undisturbed soil monolith. Appl. Eng. Agric. 14, 303-308.

Mischke, W. (1960): Untersuchungen über den Einfluß des Bestandsklimas auf die Entwicklung der Blattfleckenkrankheit *Cercospora beticola* im Hinblick auf die Einrichtung eines Warndienstes. Bayer. Landwirtsch. Jahrb. 37, 197-227.

Moffett, M. W. (1985): An Indian ant's novel method for obtaining water. Natl. Geogr. Res. 1, 146-149.

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

Monteith J. L., Szeicz G. (1961): The radiation balance of bare soil and vegetation. Q. J. R. Meteorol. Soc. 87,159-170.

Monteith, J. L. (1963): Dew: facts and fallacies. In: Rutter, A. J., Whitehead, F. H. (eds), The water relations of plants (pp. 37-56). Wiley, New York.

Monteith, J. L. (1973): Principles of environmental physics. Arnold, London.

Monteith, J. L., Unsworth, M. H. (1990): Principles of environmental physics, 2nd ed. Arnold, London.

Muselli, M., Beysens, D., Marcillat, J., Milimouk, I., Nilsson, T., Louche, A. (2002): Dew water collector for potable water in Ajaccio (Corsica Island, France). Atmos. Res. 64, 297-312.

Muselli, M., Beysens, D., Mileta, M., Milimouk, I. (2009): Dew and rain water collection in the Dalmatian Coast, Croatia. Atmos. Res. 92, 455-463.

Myers, T. M. (1974): Dew as a visual indicator of the urban heat island of Washington, D.C. M.Sc. thesis, University of Maryland, Maryland.

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

Nakshabandi, G. A., Kohnke, H. (1965): Thermal conductivity and diffusivity of soils as related to moisture tension and other physical properties. Agric. Meteorol.2, 271-279.

Neumann, J. (1956): Estimating the amount of dewfall. Arch. Meteorol. Geophys. Bioklimatol., Ser. A, 9, 197-203.

Newton, O. H., Riley, J. A. (1964): Dew in the Mississippi delta in the fall. Mon. Weather Rev. 92, 369-373.

Nilsson, T. (1996): Initial experiments on dew collection in Sweden and Tanzania. Sol. Energ. Mat. Sol. Cells 40, 23-32.

Ninari, N., Berliner, P. R. (2002): The role of dew in the water and heat balance of bare loess soil in the Negev Desert: quantifying the actual dew deposition on the soil surface. Atmos. Res. ,64, 323-334.

Okochi, H., Kajimoto, T., Arai, Y., Igawa, M. (1996): Effect of acid deposition on urban dew chemistry in Yokohama, Japan. Bull. Chem. Soc. Jpn. 69, 3355-3365.

Paltridge G. W., Platt, C. M. R. (1976): Radiative processes in meteorology and climatology. Elsevier, Amsterdam.

Parchinger, V. (1918): Eine neue Taumessungsmethode. Meteorol. Z. 35, 47-48.

Pedro, M. J., Gillespie, T. J. (1982a): Estimating dew duration. I. Utilizing micrometeorological data. Agric. Meteorol. 25, 283-296.

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

Penman H. L. (1948): Natural evaporation from open water, bare soil and grass. Proc. R. Soc. Lond. A 193,120-145.

Pickering, W. H. (1913): A year’s record with a dew recorder at Mandeville, Jamaica. Q. J. R. Meteorol. Soc. 39, 326-327.

Plauborg, F. (1995): Evaporation from bare soil in a temperate humid climate - measurement using micro-lysimeters and time domain reflectometry. Agric. Forest Meteorol. 76, 1-17.

Pruitt, W. O., Angus, D. E. (1960): Large weighing lysimeter for measuring evapotranspiration. Trans. ASAE 3, 13-18.

Qiu, G. Y., Momii, K., Yano. T. (1996): Estimation of plant transpiration by imitation leaf temperature. I: Theoretical consideration and field verification. Trans. Japan. Soc. Irr. Drain. 64, 401-410.

Raman, C. R. V., Venkataraman, S., Krishnamurthy, V. (1973): Dew over India and its contribution to winter-crop water balance. Agric. Meteorol. 11, 17-35.

Richards, K., Oke, T. R. (2002): Validation and results of a scale model of dew deposition in urban environments. Int. J. Climatol. 22, 1915-1933.

Richards, K. (2004): Observation and simulation of dew in rural and urban environments. Prog. Phys. Geogr. 28, 76-94.

Riley, J. A., Giles W. L. (1965): Agricultural meteorology in relation to the use of pesticides in the U.S.A. Agric. Meteorol. 2, 225-245.

Rönsch, H. (1990): Tau und Reif in Harzgerode. Z. Meteorol. 40, 197-204.

Rosenberg, N. J. (1969): Evaporation and condensation on bare soil under irrigation in the East Central Great Plains. Agron. J. 61, 557-561.

Rosenberg, N. J. (1974): Microclimate: the biological environment. Wiley, New York.

Ross, J. (1981): The radiation regime and architecture of plant stands. Junk, The Hague.

Royle, D. J., Butler, D. R. (1986): Epidemiological significance of liquid water in crop canopies and its role in disease forecasting. In: Ayres, P. G., Boddy, L. (eds.), Water, fungi and plants (pp. 139-156). Cambridge University Press, Cambridge.

Rupp, H., Meissner, R., Leinweber, P., Lennartz, B., Seyfarth, M. (2007): Design and operability of a large weighable fen lysimeter. Water Air Soil Poll. 186, 323-335.

Sayler, M. D., Allen, M. D., Burman, R. D., Smith, J. L. (1985): Design and installation of a weighable lysimeter. Final Report (Project No. 2-92384), submitted to the Wyoming Water Research Center, University of Wyoming, Laramie.

Scherm, H., van Bruggen, A. H. C. (1993a): Sensitivity of simulated dew duration to meteorological variations in different climatic regions of California. Agric. For. Meteorol. 66, 229-245.

Scherm, H., van Bruggen, A. H. C. (1993b): Response surface models for germination and infection of *Bremia lactucae*, the fungus causing downy mildew of lettuce. Ecol. Model. 65, 281-296.

Scherm, H., Koike, S. T., Laemmlen, F. F., van Bruggen, A. H. C. (1995): Field evaluation of fungicide spray advisories against lettuce downy mildew (*Bremia lactucae*) based on measured or forecast morning leaf wetness. Plant Dis. 79, 511-516.

Schneider, A. D., Howell, T. A., Steiner, J. L. (1993): An evapotranspiration research facility using monolithic lysimeters from three soils. Appl. Eng. Agric. 9, 227-235.

Schneider, A. D., Howell, T. A., Moustafa, A. T. A., Evett, S. R., Abou-Zeid, W. (1998): A simplified weighing lysimeter for monolithic or reconstructed soils. Appl. Eng. Agric. 14, 267-273.

Sentelhas P. C., Gillespie, T. J. (2008): Estimating hourly net radiation for leaf wetness duration using the Penman-Monteith equation. Theor. Appl. Climatol. 91, 205-215.

Severini, M., Moriconi, M. L., Tonna, G., Olivieri, B. (1984): Dewfall and evapotranspiration determination during day- and nighttime on an irrigated lawn. J. Clim. Appl. Meteorol. 23, 1241-1246.

Sharma, M. L. (1976): Contribution of dew in the hydrologic balance of a semi-arid grassland. Agric. Meteorol. 17, 321-331.

Shaw, R. H. (1973): Dew duration in central Iowa. Iowa State J. Res. 47, 219-227.

Shaw, R. H., Pereira, A. R. (1982): Aerodynamic roughness of a plant canopy: A numerical experiment. ,Agric. Meteorol. 26, 51-65.

Simon, R. D., Abeliovich, A., Belkin, S. (1994): A novel terrestrial halophilic environment: the phylloplane of Atriplex halimus, a salt-excreting plant. FEMS Microbiol. Ecol. 14, 99-110.

Slatyer, R. O., McIllroy, I. C. (1961): Practical Microclimatology. Unesco, Paris.

Smith, L. P. (1958): The duration of surface wetness. Proc. Int. Hort. Congr. 15, 478-484.

Smith, H., Carpenter, R. C. (1966): Dew forecasting study for South Carolina. Agricultural weather studies for South Carolina. South Carolina Deptartment of Agriculture, Columbia.

Steubing, L. (1951): Der Einfluß von Heckenanlagen auf den Taufall. Ber. Deut. Wetterd. U.S. Zone 32, 53-56.

Steubing, L. (1952): Der Tau und seine Beeinflussung durch Windschutzanlagen. Biolog. Zentralbl. 71, 282- 313.

Steubing, L. (1955): Studien über den Taufall als Vegetationsfaktor. Ber. Deut. Bot. Ges. 68, 55-70.

Steubing, L., Casperson, G. (1959): Pflanzentemperatur und Taubeschlag. Angew. Meteorol. 3, 219-224.

Stewart, J. B. (1977): Evaporation from the wet canopy of a pine forest. Water Resour. Res. 13, 915-921.

Stone, E. C., Fowells, H. A. (1954): Drought survival of ponderosa: pine seedlings treated with simulated dew survive by month non-treated controls in greenhouse tests. Calif. Agric. 8, 9.

Stone, E. C. (1957a): Dew as an ecological factor. I. A review of the literature. Ecology 38, 407-413.

Stone, E. C. (1957b): Dew as an ecological factor. II. The effect of artificial dew on the survival of Pinus ponderosa and associated species. Ecology 38, 414-422.

Subramaniam, A. R., Kesava Rao, A. V. R. (1983): Dew fall in sand dune areas of India. Int. J. Biometeorol. 27, 271-280.

Sudmeyer, R. A., Nulsen, R. A., Scott, W. D. (1994): Measured dewfall and potential condensation on grazed pasture in the Collie River basin, southwestern Australia. J. Hydrol. 154, 255-269.

Takenaka, N., Soda, H., Sato, K., Terada, H., Suzue, T., Bandow, H., Maeda, Y. (2003): Difference in amounts and composition of dew from different types of dew collectors. Water Air Soil Pollut. 147, 51-60.

Telford J. W. (1982): A theoretical value for von Kármán's constant. Pure Appl. Geophys. 120, 648-661.

Tennekes, H. (1968): Outline of a second-order theory of turbulent pipe flow. Am. Inst. Aeronaut. Astronaut. J. 6, 1735-1740.

Thompson, N. (1981): The duration of leaf wetness. Meteorol. Mag. 110, 1-12.

Tuller, S. E., Chilton, R. (1973): The role of dew in the seasonal moisture balance of a summer-dry climate. Agric. Meteorol. 11, 135-142.

Tyagi, N. K., Sharma, D. K., Luthra, S. K. (2003): Determination of evapotranspiration for maize and berseem clover. Irr. Sci. 21, 173-181.

Uehara, Y., Imoto, M., Sakai, Y. (1988): Study on the forecasting of the rice blast development using the weather data from AMeDAS. Bull. Hiroshima Prefecture Agric. Exp. Stn. 51, 1-15.

UNESCO (1958): Climatology and microclimatology: proceedings of the Canberra symposium. UNESCO, Paris.

Unold, G. (2003): Anforderungen an die Sickerwasserbeprobung. BAL-Workshopbericht Sickerwassersammler vom 8. und 9. April 2002. UMS GmbH, Munich.

van de Griend, A. A., van Boxel, J. H. (1989): Water and surface energy balance model with a ultilayer canopy representation for remote sensing purposes. Water Resour. Res. 25, 949-971.

van Eimern, J. (1964): Untersuchungen über das Klima in Pflanzenbeständen. Ber. Deut. Wetterd. 13, 76 S.

von Kerner, F. (1892): Ein Instrument zur Messung des Tauniederschlages. Meteorol. Z. 9, 106-108.

Wales-Smith, B. G. (1983): An inexpensive device for measuring net dewfall. Weather 38, 54-56.

Wallin, J. R. (1963): Dew, its significance and measurement in phytopathology. Phytopathology 53, 1210-1216.

Wallin, J. R. (1967): Agrometeorological aspects of dew. Agric. Meteorol. 4, 85-102.

Went, F. W. (1955): Fog, mist, dew and other sources of water. In: Stefferud, A. (ed.), Water. The yearbook of agriculture 1955 (pp. 103-109). U.S. Dept. of Agriculture, Washington, D.C.

Wilson, T. B., Bland, W. L., Norman, J. M. (1999): Measurement and simulation of dew accumulation and drying in a potato canopy. Agric. For. Meteorol. 93, 111-119.

Wisniewski, J. (1982): The potential acidity associated with dews, frosts, and fogs. Water Air Soil Pollut. 17, 361-377.

Wittich, K-P. (1995): Some remarks on dew duration on top of an orchard. Agric. For. Meteorol. 72, 167-180.

Yang, S. L., Aydin, M., Yano, T., Li, X. (2003): Evapotranspiration of orange trees in greenhouse lysimeters. Irr. Sci. 21, 145-149.

Ye, Y. H., Zhou, K., Song, L. Y., Jin, J. H., Peng, S. L. (2007): Dew amounts and its correlations with meteorological factors in urban landscapes of Guangzhou, China. Atmos. Res. 86, 21-29.

Yoder, R. E., Odhiambo, L. O., Wright, W. C. (2005): Evaluation of methods for estimating daily reference crop evapotranspiration at a site in the humid southeast United States. Appl. Eng. Agric. 21, 197-202.

Young, M. H., Wierenga, P. J., Mancino, C. F. (1997): Monitoring near-surface soil water storage in turfgrass using time domain reflectometry and weighing lysimetry. Soil Sci. Soc. Am. J. 61, 1138-1146.

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

Zhang, Y. Q., Yu, Q., Liu, C., Jiang, J., Zhang, X. (2004): Estimation of winter wheat evapotranspiration under water stress with two semi-empirical approaches. Agron. J. 96, 159-168.

Darstellung der wissenschaftlichen Entwicklung

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- THESIS: Study on the Characteristics of a Super Absorbent Polymer and its Effect on Soil Structure and the Growth of Trees
- OVERALL GRADE: Excellent

07/99 **Bachelor of Science in Agriculture (B.Sc. Agr.)**

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Professional Experience

Since10/06 **Guest Scientist,** Helmholtz Center for Environmental Research - UFZ, Leipzig-Halle, Germany

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07/99-07/04 **Teaching Assistant,** Beijing Forestry University, Beijing, China

07/99-09/05 **Secretary of the Dean,** Beijing Forestry University, Beijing, China

Erklärung

Hiermit versichere ich, dass ich die vorliegende Arbeit selbständig und nur unter Verwendung der angegebenen Literatur und Hilfsmittel angefertigt habe.

Mit dieser wissenschaftlichen Arbeit wurden bisher keine vergeblichen Promotionsversuche unternommen.

Des Weiteren erkläre ich, dass gegen mich keine Strafverfahren anhängig sind.

Halle/Saale, den 08.04.2010

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