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Estimation of energy balance components for surfaces with low vegetation

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Estimering af energibalancekomponenter for flader med lav vegetation

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After a merger between KVL, University of Copenhagen and The Danish Pharmaceutical University, the former KVL now continues it's activities as "Faculty of Life Sciences" under University of Copenhagen.

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Resumé

Opstilling af troværdige vandbalancer på markskala eller regionalt skala kræver nøjagtige estimater af størrelsen af balancens enkelte led. Det er konstateret, at beregnede afstrømningsopgørelser fra vandløbsoplande overstiger målte værdier med op til 30 %, og det er sandsynliggjort at størstedelen af denne fejlestimering stammer fra fejlberegnede nedbørs- og evapotranspirationsværdier. Fejl af denne størrelsesorden vil på mark- og oplandsskala medføre væsentlige fejlskøn i forbindelse med fx vandingsplanlægning, beregning af udvaskning af næringsstoffer og pesticider, bæredygtig forvaltning af grundvandsindvinding, vandressourceplanlægning mv. Selv såfremt blot halvdelen af vandbalancens afvigelse på 30 % stammer fra en fejlberegnet evapotranspiration, vil unøjagtigheder af denne størrelsesorden være afgørende i forbindelse med eksempelvis vandingsplanlægning af markafgrøder. Blot halvvejs gennem vandingssæsonen vil der være oparbejdet en afvigelse mellem faktisk og beregnet evapotranspiration på 30 mm. Ved et deficit af denne størrelsesorden vil afgrøder dyrket på sandede jorder være udsat for vandstress.

Ved beregning af evapotranspiration indgår et antal meteorologiske drivvariable. Usikkerheder i disse meteorologiske ligningsinput medfører betydelige usikkerheder og bidrager til fejlskøn af evapotranspirationens størrelse. Fejlestimering af det primære meteorologiske energiinput i form af nettostråling såvel som en undervurdering af merfordampningen fra afgrøder i forhold til en græsbevokset jordoverflade kan bidrage til en eventuel underestimering af evapotranspirationen. For at teste disse hypoteser blev der gennemført en analyse af eksisterende modellers tilstrækkelighed til at beregne langbølget indstråling. På baggrund heraf blev resultaterne fra såvel eksisterende som nyudviklede modeller til beregning af nettostråling analyseret og sammenlignet med op til fire uafhængige meteorologiske data sæt. Følsomheden af beregning af reference evapotranspiration, dvs. evapotranspirationen fra en veldefineret græsdækket jordoverflade, i forhold til den anvendte metode til at estimere nettostråling blev undersøgt. Endelig blev evapotranspirationen fra vinterhvede, målt vha. eddy covarians udstyr gennem afgrødens vækstsæson, sammenlignet med reference evapotranspiration med henblik på udledning af afgrødekoefficienter. De anvendte meteorologiske data sæt stammede fra Taastrup og Foulum i Danmark og Córdoba og Zaragoza i Spanien. For at sikre pålideligheden og gyldigheden af de meteorologiske data sæt, blev de forud for anvendelse underkastet et omfattende kvalitetskontrolprogram hvori eventuelle fejlmålinger blev identificeret og fejlene afhjulpet.

Resultaterne viste en markant forskel i resultaterne blandt de 20 afprøvede modeller til beregning af langbølget indstråling i dagtimerne under skyfri forhold. Sammenlignet med målte værdier af langbølget indstråling under skyfri betingelser i Taastrup og Foulum varierede systematiske fejl (MBE) fra -23 til 15 W m^{-2}. Resultaterne fra modellerne forbedredes ved 1) brug af meget klare dag (lav turbiditet af atmosfæren) og 2) ved at aggregere timeværdier til døgnværdier. Resultaterne viste derimod ingen eller lav følsomhed overfor 1) hvorvidt opvindsområderne var domineret af henholdsvis vegetation eller befæstede arealer, eller 2) årstid. I alt fire modeller blev fundet at give tilfredsstillende resultater.

I forbindelse med beregning af nettostråling blev modeller baseret på henholdsvis 1) regression, 2) fysisk-empirisk udledte sammenhænge baseret på Stefan-Boltzman teori og 3) reducerede sammenhænge baseret på Stefan-Boltzman teori, afprøvet. Modellerne blev anvendt under brug af såvel lokalt kalibrerede model koefficienter, koefficienter kalibreret til det pågældende klima som koefficienter hentet i litteraturen. Resultaterne viste, at nettostrålingsmodellerne som var baseret på regression udviste god overensstemmelse med målte værdier hvis de blev anvendt på samme lokalitet (med et uafhængigt data sæt) som hvor koefficienterne var beregnet, men stor variabilitet hvis de blev anvendt på andre lokaliteter. Modeller baseret på Stefan-Boltzman teori udviste mindre følsomhed overfor lokalitet, om end resultaterne blev forbedret ved at graduere modellernes koefficienter for skydække efter klimazone. Resultaterne var i bedst overensstemmelse med målte værdier i forårs- og sommermånederne, og når nettostrålingen blev beregnet på døgnbasis frem for på timebasis. Resultaterne fra modellerne baseret på reducerede Stefan-Boltzman sammenhænge var i overensstemmelse med resultaterne fra modellerne baseret på Stefan-Boltzman teori.

Følsomheden af beregningen af timeværdier for reference evapotranspiration i forhold til hvordan nettostråling var estimeret blev testet. Sammenlignet med reference fordampning beregnet under brug af målte værdier af nettostråling, blev der fundet en MBE ned til 0.01 mm pr. dag ved brug af beregnede værdier af nettostråling.

Der blev opnået en overensstemmelse mellem nettostrålingen beregnet fra energibalancen fra eddy covarians målingerne i en vinterhvede mark til en uafhængig måling af nettostråling foretaget i samme mark på mellem 55 og 75 %. Latent varmefluks for vinterhvede blev som gennemsnit over to års målinger i vækstsæsonen fundet til at være 1.14 højere end reference evapotranspiration. Denne værdi overstiger maximumtærsklen på 1.0 som tidligere har været anvendt i Danmark. Dette antyder, at evapotranspirationen har været underestimeret i tidligere vandbalancestudier.

Summary

Considerable uncertainties are involved when establishing a water balance at field or regional scale. Discrepancies of up to 30 % between calculated and measured stream flows have been observed. The major uncertainties arise from the estimations of precipitation and evapotranspiration components of the water balances. On field and regional scale this magnitude of error will cause considerable mistakes in estimations of e.g. irrigation and solute leaching, in sustainable ground water management, water resource planning etc. Even if only half the 30 % discrepancy is from an incorrect evapotranspiration estimation, uncertainties of this magnitude is intolerable for e.g. an irrigation scheduler. Only half way through the irrigation season a 30 mm discrepancy between the actual and calculated soil water content would occur. At this deficit crops would experience water stress on sandy soils.

Calculation of evapotranspiration includes a number of meteorological variables. Errors in net radiation, which is the main meteorological energy source in the calculations, and an underestimation of the actual evapotranspiration from crops compared to the evapotranspiration from a grass covered surface may contribute to the uncertainties in the water balances. In order to test these hypothesises an analysis of the performance of existing models to estimate daytime long wave incoming radiation was conducted. Based on this the predictions from existing and new models to calculate net radiation was analyzed and compared to measured values of net radiation from up to four independent meteorological data sets. The sensitivity of calculating reference evapotranspiration, i.e. the evapotranspiration from a well defined grass covered surface, to the method of establishing net radiation was analyzed. Finally was the evapotranspiration from winter wheat, measured using an eddy covariance system during the growing season, compared to reference evapotranspiration in order to establish crop coefficients. The meteorological data sets were from Taastrup and Foulum in Denmark and Córdoba and Zaragoza in Spain. The quality of the meteorological data set were scrutinized prior the analysis thereby identifying erroneous measurements. Measurements flagged for some kind of error by the quality control system were corrected or omitted from the analysis.

The results showed a large difference in the predictions of twenty models for calculating daytime long wave radiation. Compared to measured values of clear sky long wave irradiance in Taastrup and Foulum the mean bias error (MBE) of the model predictions varied from -23 to 15 W m^{-2}. The

predictions of the models were improved 1) when using very clear days with low atmospheric turbidity, and 2) when aggregating the daytime hourly input to daily daytime input. No improvements in the performance of the models were found when dividing the data set by 1) whether the upwind surface type was mainly urban or covered with vegetation, and 2) season. Four models were found to give consistent and satisfactory results.

When calculating net radiation models based on 1) regression, 2) physical-empirical equations based on full Stefan-Boltzman theory and 3) reduced Stefan-Boltzman type models were used. The models were used with locally calibrated model coefficients, model coefficients calibrated for the climate at the site as well as using coefficients from the literature. Models based on regression showed good agreement with measured net radiation values when the models were used at the same site as they were calibrated at (using a split dataset). However, the regression type models showed considerable variability when used with other calibration coefficients than the locally calibrated ones. Models based on Stefan-Boltzman theory were less sensitive to whether locally calibrated model coefficients were used. The performance of these models did, however, improve when the model coefficients accounting for cloud cover were adjusted with respect to climate regime. The models made better predictions during spring and summer, and when calculating net radiation for daily time steps rather than hourly time steps. The performance of the reduced Stefan-Boltzman models resembled the performance of the full Stefan-Boltzman theory models.

The sensitivity of the estimations of hourly values of reference evapotranspiration with respect to how net radiation was estimated was analyzed. Compared to reference evapotranspiration calculated using measured net radiation, MBE as low as 0.01 mm day^{-1} was found when using calculated values of net radiation.

When comparing the eddy covariance measurements to an independent measurement of net radiation in the winter wheat field, there was an energy balance closure between 55 and 75 %. Evapotranspiration from the winter wheat was found to exceed reference evapotranspiration by 1.14. This value is higher than the maximum value of 1.0 commonly used in Denmark. This indicates that evapotranspiration has been underestimated in previous water balance studies.

List of original publications

I. Kjaersgaard, J. H., Plauborg, F. L., Hansen, S., 2007. Comparison of models for calculating long wave irradiance using long term dataset. Agric. For. Meteorol. 143, 49-63.

II. Kjaersgaard, J. H., Cuenca, R. H., Plauborg, F. L., Hansen, S., 2007. Long-term comparisons of net radiation calculation schemes. Bound. Layer Meteorol. DOI 10.1007/s10546-006-9151-8

III. Kjaersgaard, J. H., Cuenca, R. H., Martínez-Cob, A., Gavilán, P., Plauborg, F. L., Hansen, S., 2007. Comparison of the performance of net radiation calculation schemes. Submitted to Theoretical and Applied Climat.

IV. Kjaersgaard, J. H., Plauborg, F. L., Mollerup, M., Petersen, C. T., Hansen, S., 2007. Sensitivity of reference evapotranspiration calculation and crop coefficients for winter wheat in a temperate, sub-humid climate. Submitted to Agric. Water Manag.

In the following the papers will be referred to by their Roman numerals.

1 Introduction

The process of when a liquid substance is converted to vapour phase is termed evaporation. In cases where the phase change is from solid to vapour the phenomena is called sublimation. Vaporization of water through the stomata of living plants is termed transpiration. For a vegetated land surface the evaporation of water into the air from the soil and wet vegetation surfaces is difficult to separate from vegetation transpiration, and the two terms are often combined in the term evapotranspiration (ET) (Miller, 1977; Marschner, 1995).

1.1 Rationale

In a natural ecosystem water is continuously in circulation. Water evaporates into the atmosphere from land and sea surfaces, water vapour is transported with the wind before the vapour condenses and falls to the ground. Precipitation falling on a land surface is either intercepted by a vegetation canopy, infiltrates the surface, accumulates on the surface or run off laterally. Water accumulated at the soil surface and in the upper soil layers is either lost to the atmosphere as evapotranspiration from the soil or vegetation, or percolates to deeper soil layers and contributes to a change in storage of soil water or the formation of ground water. From here the water is stored or flows to streams, lakes and oceans (Brutsaert, 1982).

It is necessary to understand the mechanisms and driving forces behind each part of the hydrological cycle to fully encompass the mechanism and magnitude of the overall water circulation. One important part of the water cycle is the fate and function of water after arriving at the upper or lower boundary of the strata associated with growing land plants, normally a few metres above and below the ground surface. For a volume of soil at a suitable scale, i.e. a field or a region and using the principle of conservation of mass a water balance can be set up, keeping track of water added to and withdrawn from the system. Water is added to the system through precipitation, irrigation, upward capillary flow and lateral flow into the system. Water leaves the system through evapotranspiration, downward percolation out of the root zone and runoff. Water balances on regional scale is usually calculated by aggregation based on different vegetation cover, soil type, climate regime etc. within the region (Hillel, 1998).

At present, considerable uncertainties are involved when establishing a water balance on field or regional scale. The major uncertainties are in the terms for precipitation and ET (Plauborg et al., 2002; Refsgaard et al., 2003). Uncertainties arise from factors such as inappropriate measurement technique and equipment, calculation errors, and that estimations are normally scaled from one point to e.g. field or regional scale. In Denmark precipitation is measured at meteorological stations scattered throughout the country using rain gauges of the Hellman type. Its design causes a measurement loss from instrument wetting and aerodynamic effects. Corrections to account for these losses have been suggested (Allerup and Madsen, 1979; Allerup et al., 1998; Vejen, 2002), but may be insufficient (Refsgaard et al., 2003). Examples of errors in ET measurement are given in Paper IV.

The implications of incorrect estimation of the terms of the water balance can be large. When estimating nutrient leaching from agricultural fields during the National Land Monitoring Programme (LOOP), Grant et al. (2001) calculated nitrate leaching for a individual fields by multiplying an estimated percolation with a measured nitrate concentration. The magnitude of calculated nitrate loss was therefore highly dependent on percolation being estimated correctly. Percolation was estimated as the remaining unknown from a water balance set up for the test sites. Refsgaard et al. (2001, 2003) questioned the validity of the approaches for estimating evapotranspiration and precipitation. When using the percolation values reported by Grant et al. (2001) Refsgaard et al. (2001, 2003) found that modelled stream flows were 30 % higher than measured stream flows in the LOOP catchments. This suggested that Grant et al. (2001) overestimated the percolation. Based on these results Refsgaard et al. (2003) and Henriksen and Refsgaard (2003) recommended that the use and magnitude of correction factors for precipitation measurements and the estimation of evapotranspiration should be scrutinized. Even if incorrect evapotranspiration estimation only explains half of the 30 % discrepancy, ET under-predictions of a magnitude of 15 % is intolerable for e.g. an irrigation scheduler. A 30 mm discrepancy between actual and calculated soil water content would occur only half way through the irrigation season. At this deficit crops would start to experience water stress on sandy soils (Hillel, 1998; Twine et al. 2000).

As noted by e.g. Detlefsen and Plauborg (2001) and Plauborg et al. (2002) lack of reliable meteorological input has caused considerable uncertainty in ET estimations. Although they are considered to give better predictions of evapotranspiration, Penman-Monteith type equations (Penman, 1948; Monteith, 1965; Shuttleworth and Wallace, 1985; Mikkelsen and Olesen, 1991; Allen et al., 1998; ASCE-EWRI, 2005) has been opted out in favour of the more simple Makkink (1957) equation in a number of recent applications (Plauborg et al., 2002; Styczen et al., 2005). The main advantages of the Makkink equation are it requires fewer input data and is relatively less sensitive to the quality of the input meteorological variables compared to

Penman-Monteith type equations (Aslyng and Hansen, 1982; Hansen, 1984; Plauborg et al., 2002).

One of the meteorological variables contributing most to uncertainties when estimating ET is estimating the energy available to drive evapotranspiration, viz. net radiation and air temperature (FAO, 1990; Allen et al., 1998). Measurements of solar radiation and air temperature are part of the routine measurement programme at most weather stations, including the station network operated by the Danish Meteorological Institute (Scharling, 2000; Scharling and Kern-Hensen, 2002; Styczen et al., 2005). Net radiation is normally only measured at well equipped research weather stations (Hansen, 2000; Brotzge and Duchon, 2000; Alados et al., 2003). The commonly used Penman-Monteith type equations generally requires net radiation as one of its meteorological input. The Makkink (1957) model (and the Penman-Monteith based model proposed by Mikkelsen and Olesen (1991)) uses solar radiation as input. Hence, to enable the use of the Penman-Monteith type equations, a number of additional calculations for estimating net long wave radiation and net radiation must be carried out.

1.2 Objectives

Considering the large differences between measured and estimated water balances and that available water resources is becoming scarcer, a better understanding of the uncertainties in the estimation of evapotranspiration is of importance. The aim of this study was to evaluate the magnitude of errors associated with calculating incoming long wave radiation and devise models that can be used in subsequent calculations of net radiation. Another goal was to develop new models for calculation of net radiation and test their performance along with existing models from the literature against measured values in order to evaluate what magnitude of errors there are associated with net radiation estimations. As net radiation is commonly used in estimation of evapotranspiration, an other goal was to test the relative magnitude of error introduced when using calculated rather than measured net radiation estimations and to establish an estimate of the daily actual evapotranspiration for a winter wheat crop

The improved knowledge is required at appropriate scales to establish reliable water balances, water resources planning, in agricultural water management, studies of nutrient leaching from arable land etc.

2. General literature review

Ancient Greek and Chinese philosophers trying to resolve the mystery of why the oceans did not overflow when rivers ran into them established the first rational explanations of evapotranspiration on record. As early as in the eighth century B.C. the Greek Hesiode (1928, c.f. Brutsaert, 1982) made an attempt to describe the atmospheric part of the hydrological cycle. In his fourth century B.C. work “Meteorology”, the Greek Aristotle proposed that atmospheric moisture was a result of water evaporated by solar radiation or other heat source (Aristotle, 1952, c.f. Brutsaert, 1982). Some of the earliest experimental studies of evaporation were conducted by Perrault (1733, c.f. Brutsaert, 1982) and Halley (Philip, 1979). Based on their experiments Perrault and Halley suggested that not only solar energy but also wind and air temperature influenced evaporation of water in nature. In the beginning of the nineteenth century Dalton included air humidity as a factor controlling the rate of evaporation (Dalton, 1802, c.f. Brutsaert, 1982). Towards the end of the nineteenth century relationships between evaporation and meteorological variables were being established laying the basis for the development of the empirical and physical based evaporation and evapotranspiration calculation methods currently used (Brutsaert, 1982; Monteith and Unsworth, 1990).

2.1 Radiation at the surface of the Earth

At temperatures above absolute zero all objects emit radiation at intensities proportional to the fourth power of their temperature. The flux of radiation R over all wavelengths from an object is calculated from Stefan-Boltzmans law

$$R = \varepsilon \sigma T^{4} \qquad (1)$$

where σ is the Stefan-Boltzmann constant, ε is the emissivity of the object and T is its absolute surface temperature.

2.1.1 Short wave radiation

The spectrum of short wave radiation from the sun received at the top of the atmosphere resembles the spectrum from an object having a surface temperature of approximately 6000 K and ε=1. At the outer edge of the atmosphere the short wave radiation flux density is denoted the solar constant. Based on satellite measurements the solar constant has been estimated to be

in the 1367 – 1395 W m^{-2} range (Hickey et al., 1982; Miller, 1981; Brutsaert, 1982; Arya, 2001). The deviating estimates may be attributed to sun spot activity, improved measurement technique and variations in sun-Earth distance (Miller, 1981; Arya, 2001). Allen et al. (1998) recommended 1367 W m^{-2}.
As a result of the curvature of the Earth and its atmosphere, radiation from the sun is received at an angle ψ to the normal. Following Lambert's cosine law, the radiant energy absorbed by a surface is proportional to cos ψ. The short wave radiation flux density received at a horizontal plane at the top of the atmosphere is termed extraterrestrial radiation (Arya, 2001).
Short wave radiation flux density, spectral composition and angular distribution are modified as it passes through the atmosphere as a result of scattering, absorption and reflection by clouds, atmospheric gasses, particles and aerosols. Short wave radiation incident on the ground surface has two distinct directional properties, viz. direct radiation from the sun, and diffuse radiation scattered by the atmosphere (Jensen, 1996; Allen et al., 1998). Allen et al. (1998) noted that on a clear sky day with low atmospheric turbidity, potential clear sky radiation incident on the ground surface is approximately 75 % of extraterrestrial radiation. This percentage is smaller at low sun angles, as the beam travels through a larger quantity of air.
A fraction of the incoming short wave radiation S_i is reflected by the surface. The quantity reflected relative to S_i is termed albedo and varies from 0.05 for wet bare soil to 0.95 for fresh snow (Brutsaert, 1982). For short vegetation such as green grass and most field crops the albedo is in the 0.18-0.27 range (Fritschen, 1967; Kalma and Badham, 1972; Brutsaert, 1982; Meyer et al., 1999). Doorenbos and Pruitt (1977) recommended 0.25 for most field crops. FAO (1990), Allen et al. (1998) and ASCE-EWRI (2005) defined a reference surface and used an albedo of 0.23. The albedo value varies as consequence of sun angle, and shows diurnal and annual variations (Kalma and Badham, 1972; Paltridge and Platt, 1976; Dong et al., 1992). Also the leaf water status and the presence of water or rime on the vegetation alter its albedo (Carlson et al., 1971). Models to calculate albedo as a function of sun angle was developed by e.g. Dong et al. (1992), USDA-SCS (1993) and Iziomon and Mayer (2002).

2.1.2 Long wave radiation

Long wave radiation is the radiant flux resulting from emission of energy from gases or particles in the atmosphere or by vegetation or the ground surface. An accurate estimation of incoming long wave radiation L_i reaching the ground surface requires detailed information about the profiles and properties of emitting agents such as aerosols and gases, and temperature in the atmosphere (Kasten and Czeplak, 1980; Monteith and Unsworth, 1990; Niemelä et al., 2001). Since vertical profile measurements are rarely available, air temperature T_a at screen level (normally 2 m above the ground) is commonly used when downward L_i is calculated using Eq. 1. The error introduced when using screen level T_a measurements rather than atmospheric temperature profiles are small, as more than half the L_i received at the ground is emitted within the lowest 100 m of the atmosphere. The magnitude of flux is highly influenced by the temperature gradients that can be measured near the ground (Monteith and Unsworth, 1990; Crawford and Duchon, 1999). A number of empirical or physically based models to calculate the atmospheric emissivity under clear sky conditions based on T_a and/or air water vapour pressure, e_a has been proposed (e.g. Ångström, 1915; Brunt, 1932; Anderson, 1954; Swinbank, 1963; Idso and Jackson, 1969; Brutsaert, 1975; Satterlund, 1979; Idso, 1981; Prata, 1996; Crawford and Duchon, 1999; Iziomon et al., 2003). The model by Brunt (1932) has further been calibrated in a number of studies (e.g. Monteith, 1961; Swinbank, 1963; Sellers, 1965; Berger et al., 1984; Berdahl and Martin, 1984; FAO, 1990; Heitor et al., 1991; Korsgaard et al., 1991 and Iziomon et al., 2003). To account for the increased L_i induced by clouds, an additional term is normally added to the clear sky L_i calculation models (Crawford and Duchon, 1999; Korsgaard et al. 1991).

Most natural surfaces has emissivities in the range 0.9 - 1.0, and for grassy vegetation it is 0.97 - 0.98 (Brutsaert, 1982; FAO, 1990). Calculation of outgoing long wave radiation L_o using Eq. 1 requires detailed information about the emissive properties and temperatures of the surface (soil, vegetation etc). This information is rarely available (Dilley and O'Brien, 1998). A common assumption is therefore that the temperature of the ground is equal to air temperature at screen height (Jensen et al., 1990; Monteith and Unsworth, 1990; Crawford and Duchon, 1999). As a consequence temperature at the ground surface is considered equal to screen height temperature, which in turn is considered equal to the temperature of the atmosphere. Calculating net long wave outgoing radiation L_n using Eq. 1 thus depends on the net emissivity, i.e. the difference in emissivities between the ground and the atmosphere. Models for calculating net emissivity is shown as models 5 - 8 in Table 1. Jensen et al. (1990), FAO (1990) and Allen et al. (1998) based the calculation of net emissivity on the model by Brunt (1932) assuming a ground surface emissivity of 0.98. Hansen (2000) used the model for atmospheric emissivity developed by Brutsaert (1975) assuming a surface emissivity of 0.97. These procedures are shown as model 5 and 6 in Table 1. In both cases, the presence of cloud cover was adjusted using a regression function with the ratio of observed S_i to the potential solar radiation S_{i0}, as suggested by Wright and Jensen (1972). However, using the Wright and Jensen (1972) adjustment for cloud cover and the Brunt (1932) and Brutsaert (1975) atmospheric emissivity expressions involve applying a number of model coefficients. These model coefficients must be determined empirically, based on experiments. To reduce the number of calibration coefficients, model 7 in Table 1 was developed (Paper III) by forcing the Wright and Jensen (1972) cloud cover regression function through the origin and merge the Stefan-Boltzman constant, the T_a expression, the net emissivity function and the slope of the Wright and Jensen (1972) cloudiness function into a single model coefficient c_c. Dr. Slob from the Royal Netherlands Meteorological Institute (unpublished, c.f. De Bruin and Stricker, 1982) developed a similar L_n estimation method, shown as model 8 in Table 1. Slob used extraterrestrial radiation S_a in the numerator of the Wright and Jensen (1972) cloud cover function rather than S_{i0}.

Table 1. Net radiation R_n calculation models based on solar radiation, S_i, albedo α, minimum and maximum absolute air temperature T_{max} and T_{min}, respectively, actual water vapour pressure e_a, clear sky potential solar radiaion S_{i0} and extraterrestrial radiation S_a. Model calibration coefficients are shown in Paper II and III.

Model no	Source	Model
1	E.g. Jensen et al. (1990)	$R_n = a_1 S_i + b_1$
2	E.g. Jensen et al. (1990)	$R_n = a_2 (1-\alpha) S_i + b_2$
3	Irmak et al. (2003)	$R_n = \beta_0 + \beta_1 X_1 + \beta_2 X_2 + \beta_3 X_3 + \beta_4 X_4$
4	Paper II	$R_n = \beta_0 + \beta_1 X_1 + \beta_2 X_2 + \beta_3 X_3$
5	Allen et al. (1998)	$R_n = S_i (1-\alpha) - \sigma \left[\frac{T_{\max,K}{}^4 + T_{\min,K}{}^4}{2} \right] \left(a_l + b_l \sqrt{e_a} \right) \left(a_c \frac{S_i}{S_{i0}} + b_c \right)$
6	Hansen (2000)	$R_n = S_i (1-\alpha) - \sigma \left[\frac{T_{\max,K}{}^4 + T_{\min,K}{}^4}{2} \right] \left(a \left(\frac{e_a}{T_a} \right)^{1/7} - b \right) \left(c_1 \frac{S_i}{S_{i0}} + c_2 \right)$
7	Paper III	$R_n = S_i (1-\alpha) - \left(c_c \frac{S_i}{S_{i0}} \right)$
8	Slob (unpublished, c.f. De Bruin and Stricker, 1982)	$R_n = S_i (1-\alpha) - \left(c_s \frac{S_i}{S_a} \right)$

2.1.3 Net radiation

Net radiation R_n is the amount of energy available at the surface for energy consuming processes such as evapotranspiration ET and heating of soil and atmosphere. Models to calculate R_n are normally based on S_i. The simplest models calculate R_n using linear (Fritchen, 1967; Aslyng, 1974; Kaminsky and Dubayah, 1997) or multiple (Irmak et al., 2003) regression models. These models are simple to use and require only one or few explanatory variables. Other models estimates net radiation by estimating the terms in the radiation balance

$$R_n = S_i - S_o + L_i - L_o = S_i(1-\alpha) - L_n \quad (2)$$

individually. R_n is net radiation, S_i and S_o are incoming and outgoing shortwave radiation, respectively and L_i and L_o are incoming and outgoing long wave radiation, respectively, α is the albedo and L_n is net outgoing long wave radiation (here positive away from the surface). In ET contexts, it is common to express energy fluxes directed towards the surface positive (Allen et al., 1998). Some common net radiation estimation methods are shown in Table 1. Model calibration coefficients are shown in papers II and III. The diurnal cycle of short wave radiation is far more variable than long wave radiation, as shown in Fig. 1.

Reliable field measurements of radiation are difficult to obtain. For ground based measurements of outgoing radiation or R_n it is difficult if not impossible to find a sample area where the surface represents the area of interest completely (Twine et al., 2000). For larger scale applications, satellite based measurements have shown promising results (Diak et al., 2000). Philipona et al. (1995) and Ohmaru et al. (1998) stated that well calibrated instruments should be able to measure short wave radiation with an error of less than ± 5 W m^{-2}, and long wave radiation with an error of less than ± 20 W m^{-2}. A common error in short wave radiation measurements is the cosine error, i.e. that the absorptive and reflective properties of the sensors itself depend on the angle of the incident beam (Coulson, 1975; Hansen et al., 1981). Coulson (1975) noted that cosine error at sun angles < 10° caused an error in the measurements of 25 % or more.

Net radiometers are delicate and require frequent maintenance and calibration (Allen et al., 1998; Alados et al., 2003). Most types of net radiometers are equipped with a dome, which normally must be cleaned for dust and possibly water droplets frequently. The dome is often

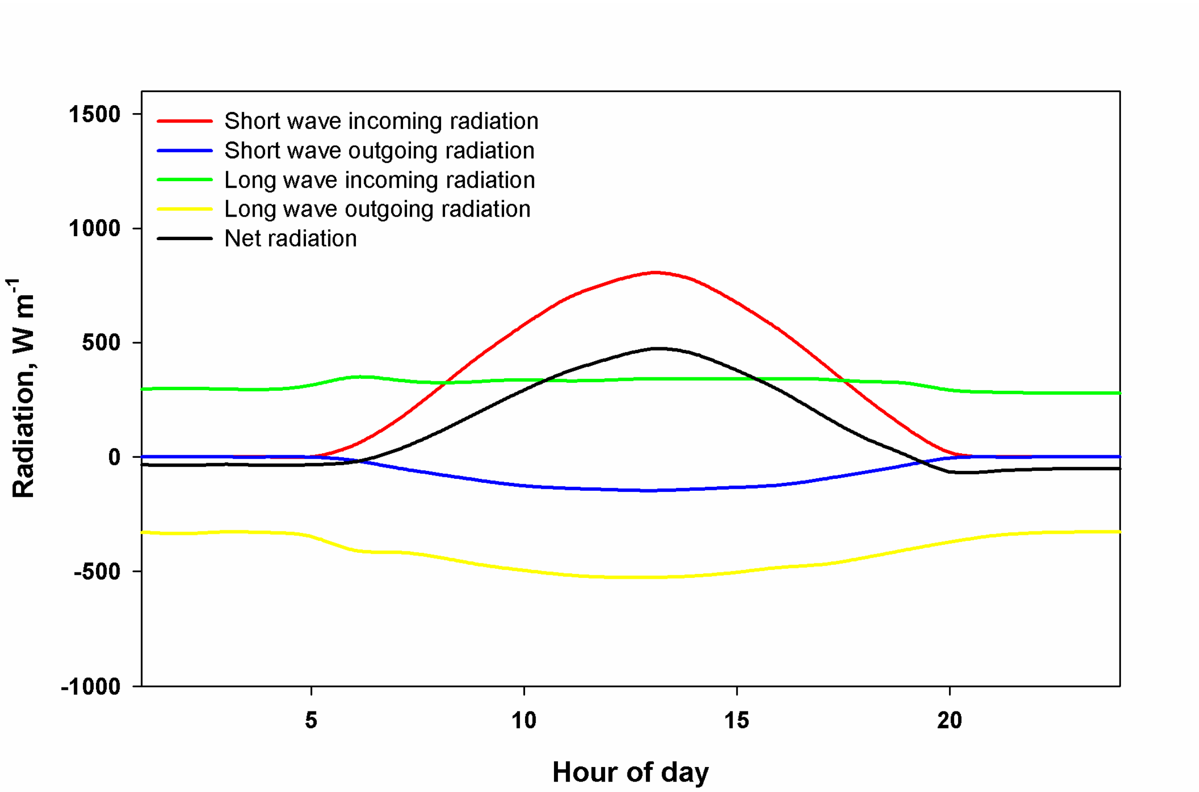

Figure 1. Daily radiation balance with long wave incoming radiation$_i$, long wave outgoing radiation, solar radiation, reflected solar radiation and net radiation under clear sky conditions on August 10 2004 in Taastrup.

made of polyethylene. The transmissivity of polyethylene gradually decreases after long term exposure, increasing the measurement error if not accounted for. Also, the domes are likely to be attacked and damaged by animals (Halldin and Lindroth, 1992). Condensation is likely to occur both inside and outside the domes. The design of the instrument may prevent condensation e.g. by being equipped with a heating ring to avoid the formation of dew and rime on the exterior (Jensen and Aslyng, 1967; Jensen, 1996). Condensation on the inside can be avoided by scrubbing moisture out of the air under the dome using a dessicant (Jensen and Aslyng, 1967; Halldin and Lindroth, 1992). In a field comparison, Halldin and Lindroth (1992) found differences between net radiometers of up to 20 %. Brotzge and Duchon (2000) found differences between two sensors varied ± 40 W m^{-2}. Domeless net radiometers were found to deviate even more from a reference sensor (Brotzge and Duchon, 2000). However, with proper calibration, installation and maintenance, the differences between well calibrated net radiometers are normally less than 6 % (Jensen, 1996; Twine et al., 2000).

2.2 Vaporisation of water from cropped surfaces

Water evaporates from a variety of surfaces such as e.g. open water bodies, soils and wet vegetation. Evaporation from a soil surface is effected by the degree of shading from vegetation and residues (Ortega et al., 2002; Thoma et al., 2005), and the amount of water available for evaporation (Hillel, 1998). Wetting of the surface occurs during rain, irrigation and from capillary rice from shallow ground water tables. When the soil surface is wet, the rate of evaporation is determined by meteorological conditions. As the surface dries out the rate of evaporation is determined by the hydraulic conductivity of the soil supplying water to the surface. During continuous drying water starts to evaporate from subsurface water menisci and is transported to the ground surface via diffusion or convection. With increasing drying of the soil evaporation may cease almost completely within a few days (Kristensen and Jensen, 1975; Allen et al., 1998). If the soil is covered by vegetation, the canopy intercepts part of the precipitation or overhead irrigation. The occurrence of wet plant surfaces reduces transpiration, as part of the available energy is used for evaporating the intercepted water and that a reduced water pressure deficit between the intercellular spaces of the leaf and the ambient air (Feddes et al., 1999).

If the water available to plants is limited, their growth and productivity can be impaired. This may be caused by closing of the stomata to reduce transpiration thus reducing photosynthesis activity (Marschner, 1995; Taiz and Zeiger, 1998), reproductive abortion during flowering (Setter et al., 2001; Liu et al., 2003) and premature senescence of the plants (Miller, 1977; Taiz and Zeiger, 1998). The majority of water taken up is transpirated into the atmosphere driving the transport of nutrients and photosynthesis assimilates through the plant (Taiz and Zeiger, 1998). Since large quantities of energy is required to change the phase of water from liquid to vapour, transpiration is also an effective means for cooling the leaves (Miller, 1977; Monteith and Unsworth, 1990).

Where rainfall is inadequate to cover the crop water requirements the plants are often irrigated to achieve greater production (Jensen et al., 1990). Turner (1997) estimated that as much as 70 % of all fresh water withdrawals world wide is used for irrigation purposes. In Denmark, about a quarter of the water consumption is used for irrigation in agriculture and horticulture (Refsgaard et al., 2003). As water availability is often limited, it is important to be able to predict how much water is required for irrigation (Allen et al., 1998).

The rate of evapotranspiration from a vegetation depends on a number of interlinked factors such as the physiological and morphological nature of the crop, management, climate, soil properties, local topography and the regional land use. It is difficult to measure the effects of each factor, and evapotranspiration estimation is largely based on empirical relations, with predictions using a small subset of these factors (Monteith and Unsworth, 1990; Feddes et al., 1999).

Water vaporizes in the intercellular spaces within the leaf. As the cuticle covering the leaf is almost impermeable to water, the degree of stomata aperture controls the majority of the vapour exchange. The relative humidity is 95 - 99 % in the intercellular spaces, while it typically is 50 - 90 % in the ambient air in sub-humid climate regimes such as Denmark (Jensen, 1996; Taiz and Zeiger, 1998). If the stomata are open, a net diffusion of water vapour out of the leaf occurs (Jensen et al., 1990; Taiz and Zeiger, 1998). Hence, regulating the stomata aperture changes the resistance against water vapour transport from the leaf. In addition, the laminar sublayer along the leaf also act as a resistance against transport into the ambient air. Both resistances are very variable and depend on the closure of the stomata and the thickness of boundary layer, respectively. If the plant is subject to water stress stomata will open only a little increasing the resistance. Similarly, under conditions with low mixing of air around the leaf the boundary layer gets thicker thereby increasing the resistance (Jensen et al., 1990; Taiz and Zeiger, 1998).

The principal meteorological driving forces for evaporation and transpiration are R_n, T_a, e_a and wind speed. The source of energy required to evaporate water is solar radiation and, to a lesser extent, sensible heat. The water vapour deficit drives the vapour flux between the evaporating surface and the surrounding air. By stirring up the air masses the wind maintains a vapour pressure gradient between the evaporating surface and the ambient air (Taiz and Zeiger, 1998; Feddes et al., 1999). For a bare soil and until after the emergence of vegetation nearly all of the ET comes from evaporation, while at full crop cover more than 90 % of ET comes from transpiration (Allen et al., 1998).

2.3 Estimation of ET

The partition of the available energy at the surface of the ground or vegetation, viz. R_n is described by the energy balance, which for a simple lumped system has the form

$$R_n = \lambda E + H + G \tag{3}$$

where λ is the latent heat of evaporation, E is the rate of evaporation, H is the flux of sensible heat into the atmosphere and G is soil heat flux. Other factors affecting the energy balance such as snow melt, freezing or thawing of water, photosynthesis, lateral advection and heat stored by the vegetation are usually neglected (Miller, 1977; Monteith and Unsworth, 1990), but may need to be considered depending on the scale of time and space for the application (Kessler and Jaeger, 1999). R_n is normally the dominating term in Eq. 3, as it usually has the greatest values and drives the other energy flux terms (Jensen et al., 1990).

The ratio of sensible to latent heat is termed the Bowen ratio, β, which for stable atmospheric conditions can be expressed as (Bowen, 1926; Lewis, 1995)

$$\beta = \frac{H}{\lambda E} = \gamma \frac{(T_s - T_a)}{(e_s - e_a)} \tag{4}$$

T_s and T_a is the temperature of the evaporating surface and air, respectively, e_s is the actual vapour pressure of air near the evaporating surface and e_a is actual vapour pressure of the ambient air. γ is the psycrometric constant which can be calculated as

$$\gamma = \frac{c_p P}{0.622\lambda} \tag{5}$$

where c_p is the specific heat of air at constant pressure and P is the atmospheric pressure.

If R_n and G have been established, E can be calculated if also the Bowen ratio (Eq. 4) is known.

The rate of water vapour flux away from the surface of a leaf is determined by diffusion over the laminar boundary layer of the leaf. This is usually expressed using Fick’s law, i.e.

$$\lambda E = -\frac{\rho c_p}{\gamma} K_v \frac{\partial e}{\partial Z} \tag{6}$$

where ρ and c_p are the density and specific heat of the air, respectively, K_v in the eddy transfer coefficient for vapour, $\partial e/\partial Z$ is the vertical gradient of atmospheric vapour pressure and γ is the psychrometric constant from Eq. 5. Analogously, the rate of sensible heat transfer are expressed as

$$H = \rho c_p K_H \frac{\partial T}{\partial Z} \tag{7}$$

where $\partial T/\partial Z$ is the temperature gradient and K_H is the transfer coefficient for sensible heat. When calculating evapotranspiration at field scale it is often useful to distinguish between different measures of evapotranspiration. Actual evapotranspiration ET_c is the actual amount of water that is lost to the atmosphere from a surface at the given water supply and content, field management, climatic conditions and vegetation development and density. Reference evapotranspiration ET_0 denotes the amount of water that is lost to the atmosphere from clipped, green, actively growing dense grass with an amble water supply as defined by Doorenbos and Pruit (1977) and Allen et al. (1998). Reference evapotranspiration is used to describe the atmospheric “demand” for water.

In the literature, the term potential evapotranspiration is sometimes used. Its definition is however ambiguous, as the expression is used interchangeably for evaporation from a surface resembling the definition of reference evapotranspiration (Penman, 1948; Hansen, 1984; Mikkelsen and Olesen, 1991; Plauborg, 2002) as well as evapotranspiration from any given crop with amble water supply (Aslyng and Hansen, 1982; Hillel, 1998). For this reason it has been strongly discouraged to use the term potential evapotranspiration (Allen et al., 1998). Calculation of ET_0 is tied to using the FAO Penman-Monteith equation (FAO, 1990; Allen et al., 1998). Other commonly used equations for calculating ET from short grass is the methods suggested by Penman (1948) and Makkink (1957).

2.3.1 The Penman combination method

Prior to 1948, equations based on either the Dalton mass transfer equation (Dalton, 1802, c.f. Brutsaert, 1982) or the energy balance equation (Eq. 3) were used to calculate evaporation (Penman, 1948). As input these equations require estimations of the surface temperature of the evaporating surface, which is difficult to measure. Penman (1948) combined the two equations to remove the surface temperature term.

Penman (1948) rearranged Eq. 3, divided it by λE and assumed that the air adjacent to the evaporating surface was saturated when using Eq. 4 for substituting β, and obtained

$$\lambda E = \frac{(R_n - G)}{(1+\beta)} = \frac{(R_n - G)}{1+\gamma\left(\frac{T_s - T_a}{e_s{}^0 - e_a}\right)} \tag{8}$$

where e_s^0 is the saturation vapour pressure of air near the evaporating surface. At small temperature differences between the evaporating surface and ambient air the gradient of the saturation water vapour pressure curve Δ can be approximated and rearranged as

$$(T_s - T_a) = \frac{\left(e_s{}^0 - e_a{}^0\right)}{\Delta} \tag{9}$$

where e_a^0 is saturation vapour pressure of air. Under humid climate conditions T_s is generally close to T_a (Penman, 1948). Hence, the gradient Δ can be calculated at air temperature. Substituting Eq 9 into Eq 8 produced

$$\lambda E = \frac{(R_n - G)}{1+\frac{\gamma}{\Delta}\frac{\left(e_s{}^0 - e_a{}^0\right)}{\left(e_s{}^0 - e_a\right)}} \tag{10}$$

For Eq. 10 T_s was required to calculate e_s^0. To remove the e_s^0 term Penman (1948) used Daltons mass transfer equation

$$\lambda E = f(u)\left(e_s{}^0 - e_a\right) \tag{11}$$

where *f(u)* is an empirical wind function. Penman (1948) also defined a new mass transfer function

$$\lambda E_a = f(u)\left(e_a{}^0 - e_a\right) \quad (12)$$

Dividing Eq. 11 into 12 and rearranging gave

$$\frac{\left(e_s{}^0 - e_a{}^0\right)}{\left(e_s{}^0 - e_a\right)} = 1 - \frac{\lambda E_a}{\lambda E} \quad (13)$$

Substituting Eq. 13 into Eq. 10 gave

$$\lambda E = \frac{\left(R_n - G\right)}{1 + \frac{\gamma}{\Delta}\left(1 - \frac{\lambda E_a}{\lambda E}\right)} \quad (14)$$

After being rearranged this produced the Penman (1948) equation

$$\lambda E = \frac{\Delta\left(R_n - G\right) + \gamma f(u)\left(e_a{}^0 - e_a\right)}{\Delta + \gamma} \quad (15)$$

The Penman (1948) equation was developed to calculate evaporation from wet surfaces such as open water surfaces, wet soils or wet crop surfaces. The empirical *f(u)* was based on linear regression using wind speed as the independent variable. Jensen et al. (1990) and Dodds et al. (2005) listed parameterizations of the Penman wind functions. The variation in the regression coefficients suggests that they have only limited spatial validity, and varies depending on the surface characteristics (Penman, 1956). Doorenbos and Pruitt (1977) noted, that using their parameterization of *f(u)* produced a systematic error of up to 30 % at some experimental sites.

2.3.2 The Penman-Monteith method

The Penman equation use meteorological data only and does not include any physiological behaviour of the plant. To include a physiological rather than a strictly environmental control of ET, Monteith (1965) and Rijtema (1965) applied resistances against water vapour transport to represent the water flux pathways from the intercellular spaces in the leaves and into the atmosphere. Using Eqs. 6 and 7 and using the approximation in Eq. 9 Monteith (1965) found

$$\lambda E = \frac{\Delta(R_n - G) + \rho_a c_p \dfrac{e_s^{\,0} - e_a}{r_a}}{\Delta + \gamma\left(1 + \dfrac{r_s}{r_a}\right)} \qquad (16)$$

which is referred to as the Penman-Monteith equation. r_s is surface resistance and r_a is atmospheric resistance. If appropriate values of r_s and r_a are specified the ET from any vegetation can be estimated (Wallace, 1995).

The use of resistances in transpiration calculations is analogous to resistances in electrical circuits. An important assumption when using these resistances is that the entire canopy is represented as one big leaf at height $d + z_{0h}$, where d is the zero-plane displacement height and z_{oh} is the roughness length for vapour transfer (Monteith, 1981).

The r_s describes the resistance against diffusional water vapour flow through stomata and cuticula and through the soil surface. r_s also contains some effects of vapour flow within the canopy. It may be subdivided into a stomatal resistance component and a component describing the combined resistance over the leaf boundary layer and turbulent transfer inside the canopy structure (Alves et al., 1998; Pereira et al., 1999). For a wet surface, r_s is essentially zero and increases with environmental stresses such as soil moisture deficit (Stewart, 1988; Hansen, 2002; Dodds et al., 2005).

The main variables controlling stomata and thus the magnitude of r_s are light intensity, leaf temperature, water vapour saturation deficit and plant water potential (Jarvis, 1976; Lhomme, 2001; Allen et al., 2006). Estimations of r_s based on meteorological variables (Jarvis, 1976; Stewart, 1988; Rochette et al., 1991; Todorovic, 1999; Alves and Peireira, 2000) or by inverting Eq. 16 (De Bruin and Holtslag, 1982; Lecina et al., 2003; Harris et al., 2004) have been suggested. However, because r_s have been found to fluctuate substantially over the course of a day, during a drying cycle and between plant species (DeHeer-Amissah et al., 1981; Lindroth, 1984; Lecina et al., 2003), it is difficult to establish r_s. In other studies (e.g. Szeicz and Long, 1969; Allen et al., 1989; FAO, 1990; Allen et al., 1998, 2006) a fixed value of r_s was found to be adequate.

The r_a describes the resistance against water vapour fluxes from the canopy to a reference level above the vegetation (Monteith, 1981; Allen et al., 1998).
Under neutral stability conditions r_a can be estimated from Garratt and Hicks (1973)

$$r_a = \frac{\ln\left(\frac{z_m - d}{z_{om}}\right)\ln\left(\frac{z_h - d}{z_{oh}}\right)}{k^2 u_z} \tag{17}$$

where z_m and z_h is the measurement height for the wind speed and air humidity, respectively, d is the zero plane displacement height, z_{om} is the roughness length governing momentum transfer, z_{oh} is the roughness length governing transfer of vapour and heat, k is the von Karman constant (0.41) and u_z is the wind speed at height z.

2.3.3 Need for standardized ET calculation method

A large number of methods for calculating ET based on different meteorological input have been suggested during the last several decades (Jensen et al., 1990; Allen, 2000). Many of these models are based on empirical relations with local calibrations of model coefficients and suffer from a limited spatial and temporal validity. When testing models using the same meteorological data sets Jensen et al. (1990) found very varying results. The deviating predictability is also clear from Fig. 2, where ET predictions for short grass using 6 commonly used models (Penman, 1948; Makkink, 1957; Priestley and Taylor, 1972; Hargreaves, 1975; Monteith, 1981; Allen et al., 1998) from June 18 to July 2 2005 at Taastrup compared to ET measured using a floating lysimeter (Aslyng and Kristensen, 1961) are shown.
The Penman and Penman-Monteith equations have been used extensively and gained videly acceptance (Allen et al., 2006), provided appropriate values for r_s and r_a is applied (Jensen et al., 1990; Wallace, 1995; Allen et al., 1998; van der Keur et al., 2001)
To obtain reliable estimates of ET in varying climates it is desirable to use a standardized method following some general guidelines. The Food and Agriculture Organization of the United Nations (FAO) developed such guidelines and recommended four different ET calculation methods (Doorenbos and Pruitt, 1977). Discrepancies in the results among these four models showed some weaknesses in the methodologies suggested by Doorenbos and Pruitt

(1977). In 1998, FAO published updated guidelines for calculating crop ET (FAO, 1990; Jensen et al., 1990; Allen et al., 1998).

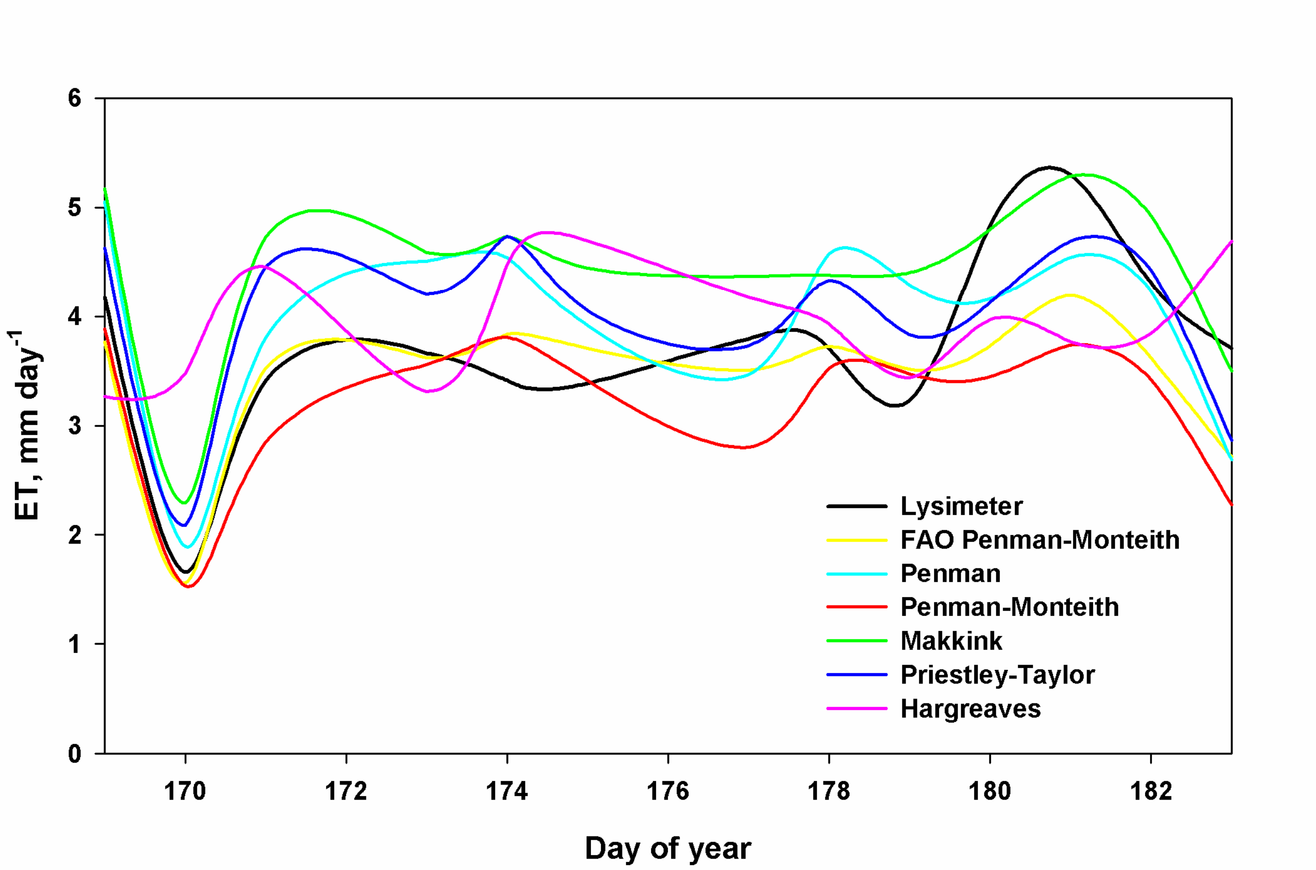

Figure 2. Daily evapotranspiration ET from short grass at Taastrup in 2005 measured using a weighable lysimeter and calculated using 6 commonly used models from the literature.

2.3.4 FAO ET_0 calculation method

To overcome the difficulties in establishing the magnitudes of r_a and r_s in Eq. 16 FAO (1990) and Allen et al. (1998) defined a reference surface as "*A hypothetical crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s m^{-1} and an albedo of 0.23*". This surface resembles the surface of green, actively growing grass with full ground cover and an abundant water supply. The r_s of 70 s m^{-1} was estimated based on Monteith (1981) and Allen et al. (1989). Itenfisu et al. (2003), Lecina et al. (2003) and Irmak et al. (2005) found a better agreement between ET_0 calculated on daily basis and 24-hour summed ET_0 when a lower value of r_s were used in the hourly calculations. This lead to a recommendation of using a r_s of 50 s m^{-1} during daytime and 200 s m^{-1} during night time for ET_0 calculations on an hourly time step basis (Itenfisu et al., 2003; ASCE-EWRI, 2005; Allen et al., 2006).

A main advantage of using a reference surface for calculating ET is that it is simpler to measure or estimate R_n for the reference surface and then relate ET_0 to other crops than to estimate R_n for each crop at all growth stages (Jensen et al., 1990).
Instability induced by large fluxes of sensible heat at the surface induces buoyancy which can influence the transport of heat and vapour from the canopy. In cases where the sensible heat flux is large, such as over dry or sparse vegetation, it may be necessary to correct the estimates of r_a from Eq. 17 for lack of stability in the boundary layer (Harris et al., 2004; Allen et al., 2006). No corrections for stability were carried out in the FAO methodology. Allen et al. (2006) argued, that under reference conditions sensible heat flux is small relative to ET. Hence, heat and vapour transfer is not strongly affected by buoyancy, and corrections for boundary layer instability could be ignored with little error (Allen et al., 2006). Another problem when applying one source ("big leaf") models is that using the $d + z_{0h}$ as the level of the evaporative surface may lead to an overestimation of r_a, as the top layer of the canopy is an important source of vapour fluxes (Alves et al. 1998). However, since the reference crop is low vegetation, this error is small (Pereira et al., 1999).
Allen et al. (1998) used d=2/3h where h is crop height (Monteith, 1981), z_{om}=0.123h (Brutsaert, 1982), z_{oh}=0.1z_{om} (Stricker and Brutsaert, 1978) and k=0.41 in Eq. 17 whereby r_a became 208/u_z. By substituting these expressions into Eq. 16 using the properties of the reference surface, the FAO Penman-Monteith equation was obtained (Allen et al., 1998)

$$ET_0 = \frac{0.408\Delta(R_n - G) + \gamma \frac{C_n}{T_a + 273} u_z \left(e_a{}^o - e_a\right)}{\Delta + \gamma(1 + C_d u_z)} \qquad (18)$$

ET_0 is reference ET (mm day^{-1}), R_n and G is measured in MJ m^{-2} day^{-1}, T_a in °C, e_a^0 and e_a is in kPa and u_2 is wind speed at 2 m height in m s^{-1}. C_n is a numerator constant that changes with calculation time step as C_n=900 for daily time steps and 37 for hourly time steps. C_d is a denominator constant that changes with the preferred value of r_s. Values for r_s of 50, 70 and 200 s m^{-1} corresponds to C_d values of 0.24, 0.35 and 0.96 s m^{-1}, respectively.

2.3.5 The Makkink method

Makkink (1957) proposed the equation

$$ET = A\frac{\Delta}{\Delta+\gamma}\frac{S_i}{\lambda} + B \tag{19}$$

to calculate ET from “standard” grass 8 to 13 cm high, extensively growing and well supplied with water. *A* and *B* are empirical model coefficients. Makkink (1957) suggested *A*=0.61 and *B*=-0.12 for The Netherlands. After further studies Makkink and van Heemst (1967) suggested using *A*=0.81 and *B*=-0.40. Based on 5 years of ET measurements using a floating lysimeter (Aslyng and Kristensen, 1961) in Taastrup Hansen et al. (1981) found *A*=0.59 and *B*=0.40. Aslyng and Hansen (1982) argued that the climatic conditions between The Netherlands and Denmark are similar and suggested using the average values between the latter coefficients, *A*=0.70 and *B*=0.0. These coefficients have been used in a number of applications (Hansen, 1984; Plauborg et al., 2002; Scharling and Kern-Hansen, 2002).

2.3.6 Decoupled calculation of crop ET

Calculating ET from Eqs. 18 or 19 acts as a measure of the atmospheric evaporative demand. Differences in plant height and leaf anatomy, stomatal characteristics, aerodynamic properties and albedo cause crop evapotranspiration to differ from that calculated from ET_0 under same meteorological conditions.

A common approach to estimate crop evapotranspiration under standard conditions ET_c is to relate ET_c to measured evaporation from an open water surface (Aslyng and Stendal, 1965; Kristensen, 1979) or calculated ET using a multiplication factor (Makkink, 1957; Doorenbos and Pruitt, 1977; Kassam and Kowal, 1975; Monteith, 1981; Allen et al. 1998).

FAO recommended for crops grown under standard conditions, i.e. the crop is growing in large fields, kept decease free, well watered and well fertilized, that ET_c is related to ET_0 (calculated using the FAO Penman-Monteith equation) using the empirical crop coefficient, k_c (Doorenbos and Pruitt, 1977; FAO, 1990; Allen et al. 1998). k_c is a bulk coefficient that integrates the differences between reference and actual vegetation characteristics, soil water evaporation and evaporation of intercepted water.

For crops growing under non-standard conditions caused by e.g. the presence of decease or pests, low soil fertility, water stress or water logging etc., the evapotranspiration may deviate from standard conditions. Crop coefficients for crops growing under non-standard conditions are adjusted by using a stress coefficient k_s to obtain adjusted actual evapotranspiration, $ET_{c\ adj}$ (Allen et al. 1998). The use of k_c to adjust the maximum obtainable evapotranspiration from a crop i.e. ET_c, is commonly used in crop growth and crop production modelling (Hansen, 2002; Styczen et al., 2005)

In a number of Danish studies k_c have been considered to have a value between 0 and 1 (Kristensen and Jensen, 1975; Aslyng and Hansen, 1982). A value larger than unity has been recommended abroad (Allen et al. 1998; Feddes, 1999; Howell et al., 2006) and in recent Danish publications (Hansen, 2000; Plauborg et al., 2002, Styczen et al., 2005).

When developing Eq. 19 Makkink (1957) used weighable lysimeters covered with short grass to calibrate the model. Makkink used a crop coefficient denoted *la faktoro f* to relate calculated ET to evaporation measured using a pan. Values of the factor *f* relating ET calculated using the Makkink equation to ET_c from crops, forests and wetlands have been published by e.g. Feddes (1987), Feddes et al. (1999) and Plauborg et al. (2002).

To account for differences in ET based on the phenological development of the crop, FAO differentiated values for k_c in four growth stages, viz. initial, development, midseason and late-season stage (Doorenbos and Pruitt, 1977; Allen et al. 1998). The initial stage was defined as the period from sowing to 10 % ground cover. Crop development stage was somewhat loosely defined as the period from 10 % ground cover to "*effective full cover*" i.e. when the ground is 70-80 % covered by vegetation. For cereals, the onset of flowering could also be used to indicate the end of the development stage (Allen et al., 1998). The midseason stage runs from effective full cover to start of maturity, while the late season stage runs from start of maturity to harvest or full senescence (Doorenbos and Pruitt, 1977; Allen et al., 1998).

In paper IV it was found that these definitions were inadequate to describe crop development satisfactorily. For winter wheat, flowering generally occurs around mid June in Denmark, which according to the FAO definitions marks the end of the crop development stage. Leaf area measurements, field records and visual observations of the winter wheat crop used in the present study indicated that effective full cover was reached four weeks prior flowering. This suggested that a more detailed crop development description should be used to determine the growth stages, while still allowing for local differences in development rate. For other agricultural purposes the phenological development of crops are described using the internationally recognized BBCH-scale (Zadoks et al., 1974; Lancashire et al., 1989). An overall outline of the scale for winter wheat and conversion between the four development stages set by the FAO and the BBCH scale used in Paper IV is shown in Table 2.

Table 2. The principal growth stages of the BBCH scale for the phenological development of winter wheat and the conversion to FAO crop development stage used in Paper IV.

BBCH Code	BBCH Principal growth stage	FAO Crop development stage
0	Germination	Initial
10	Leaf development	Initial/development
20	Tillering	Development
30	Stem elongation	Midseason
40	Booting	Midseason
50	Heading	Midseason
60	Flowering	Midseason
70	Development of fruit	Midseason
80	Ripening	Late season
90	Senescence	Late season

A main disadvantage of the single crop coefficient is it also incorporates soil evaporation. However, the mechanisms for the reduction of evaporation of soil water as the soil dries out is different than the plant characteristics ET_0 is based on. For use in detailed studies Allen et al. (1998) proposed using a dual crop coefficient as

$$k_c = k_{cb} + k_e \qquad (20)$$

where k_{cb} is a basal crop coefficient and k_e is a soil water evaporation coefficient. k_{cb} is the ratio of ET_c to ET_0 when the soil surface is dry, but the soil water content is adequate to sustain full crop transpiration. k_e is the is the evaporation from the soil surface. If the soil surface is wet, k_e may be large, but decreases sharply as the topsoil dries out. Estimation of k_e requires daily water balance estimations of topsoil water content which require detailed information about the soil water balance. Water intercepted by the canopy is included in k_{cb} (Allen et al. 1998). Partitioning evapotranspiration between transpiration and soil evaporation is important in applications such as crop production forecasting, as only transpiration is related to crop yield. Compared to the single coefficient approach, the dual crop coefficient is more suitable for daily irrigation scheduling or other research studies where daily variations in the soil surface wetness affect evapotranspiration and soil water fluxes (Wallace, 1995; Allen et al., 1998).

2.3.7 Coupled calculation of crop ET

Some controversy exists about estimating ET_c using crop coefficients. Wallace (1995) found that general crop coefficients were inadequate for sparsely vegetated areas or heterogeneous terrain, especially in arid and semi-arid environments. Shuttleworth and Wallace (1985) developed a dual source model where transpiration and evaporation from the surface was calculated individually using Penman-Monteith type equations. These equations were modified by breaking the aerodynamic resistance into a surface boundary resistance and a resistance against vapour exchange between the surface boundary layer and into the ambient air. The resistances were calculated based on crop height, wind speed and leaf area index (Shuttleworth and Wallace, 1985).

For sparse vegetation cover, going through the additional calculations involved in Shuttleworth and Wallace (1985) approach may be appropriate (van der Keur et al., 2001). The method is, however hampered by the difficulties in establishing the energy input (i.e. $R_n - G$) for these crops at all growth stages (Jensen et al., 1990) and the general difficulties in establishing appropriate atmospheric resistances (De Bruin and Holtslag, 1982; Jensen et al., 1990; Allen et al., 1998). Hence, for dense crop covers that prevails during the main growing seasons in sub-humid climates the single or dual crop coefficient approach is adequate (Jensen et al., 1990).

3 Overview of experimental studies

Meteorological data from up to four weather stations were used to test radiation and ET_0 estimation models, viz. Taastrup (55° 39´ N, 12° 20´ E, 36 m a.s.l.), Foulum (56° 29´ N, 9° 34´ E, 56 m a.s.l.), Córdoba (41° 43´ N, 0° 49´ W, 225 m a.s.l.) and Zaragoza (37° 51´ N, 4° 51´ W, 110 m a.s.l.). The data quality from the meteorological stations were scrutinized based on the recommendations by Allen (1996) and Allen et al. (1998) (Paper I, Fig. 1; Paper III, Figs. 1 and 2). Some outliers (< 25) in the radiation measurements were found and these outliers were omitted. No gap filling was performed on the datasets. In addition to the meteorological measurements, a growing season campaign measurement programme of ET_c was conducted during 2004 and 2005 using an eddy covariance (EC) system in a winter wheat field. Alongside the EC system measurements of R_n, G, precipitation and soil water content were conducted in the field. A set of statistical procedures were used to evaluate the predictions of the models compared to observed values (Paper I, Table 3) (Loague and Green, 1999; Vereckeen et al., 1991).

3.1 Results and discussion

3.1.1 Daytime incoming long wave radiation estimation

The performance of twenty models (Paper I, Table 1) was tested against datasets containing 32 and 7 consecutive years of hourly data.

The results for the clear sky models showed no dependence on whether the models used T_a, e_a or both as input. This is in agreement with Swinbank (1963) who argued that both variables can be used as input based on the correlation between T_a and e_a, which is commonly expressed in the saturation pressure vs. T_a curve.

The performances of the individual models on an hourly basis were similar at both sites, with mean bias errors (MBE) ranging from -23 to 15 W m^{-2} (Paper I, Tables 4 and 5, Fig. 3). Model performance was improved when using 1) a more strict definition of clear sky, i.e. when using very clear sky conditions with low turbidity and 2) when aggregating the daytime hourly input into daily daytime input (Paper I, Fig. 5). No improvements in performance of the models were found when dividing the dataset by 1) whether the upwind surface type was mainly urban or covered with vegetation or 2) season. Considerable scatter and uncertainty was found in the predicted values of L_i for all models. The residuals between predicted and observed values showed some sign of periodicity and exceeded the residuals found by Crawford and Duchon (1999) and Niemelä et al. (1999).

The various parameterizations of the Brunt (1932) equation yielded very different results, ranging from the overall poorest performing model (using calibration coefficients suggested by Brunt (1932)) to one of the very best performing models, i.e. using the calibration coefficients suggested by FAO (1990). The good performance of the Swinbank (1963) model was in agreement with Skartveit et al. (1996). The two physical based models, viz. Brutsaert (1975) and Prata (1996) were also found to give good results.

Crawford and Duchon (1999) and Korsgaard et al. (1991) proposed adjustments for the presence of clouds. It was found, that the two all sky irradiance models performed with somewhat equal MBE and root mean square errors. Unlike the Korsgaard et al. (1991) model the Crawford and Duchon (1999) model did not require local calibration.

3.1.2 Net radiation calculation

Models to calculate R_n were tested, viz. the regression type models shown as models 1-4 in Table 1, Stefan-Boltzman type models including net emissivity and cloud cover functions shown as model 5 and 6 in Table 1, and reduced Stefan-Boltzman type models shown as models 7 and 8 in Table 1.

The regression type models made good predictions when used with model coefficients calibrated at the same site using a split dataset with daily meteorological data. Applying the model calibration coefficients at other sites increased the MBE, especially when using calibration coefficients from the literature (Paper III, Tables 3 - 6). The results confirmed that the model calibration coefficients from regression models were sensitive to even small regional meteorological differences, which confirmed the findings by Fritchen (1967), Alados et al. (2003) and Nandagiri and Kovoor (2005). This indicated that the meteorological information included in these equations was insufficient to cover the factors controlling R_n. For the multiple regression model Irmak et al. (2003) found that the inverse distance Earth-sun was a significant explanatory variable, which was confirmed in the present study. One may speculate however, if this distance is in fact a factor influencing R_n or if it reflects seasonal effects, such as higher degree of cloud cover at certain seasons or changes in albedo over the year. Also, using the formulation suggested by Irmak et al. (2003) included using minimum and maximum daily T_a. These two factors are likely to be correlated introducing a risk of multicollinearity (Der and

Everitt, 2002), which may increase the variance of the regression and make the prediction model less stable. In paper II it was found that the multicollinearity was relatively weak and it apparently did not affect the results. It was also found, that daily mean T_a could be used as a substitute for minimum and maximum daily T_a in the multiple regression model.

The Stefan-Boltzman type models showed considerable more consistency between regions compared to the regression models. The Allen et al. (1998) and Hansen (2000) models gave very similar results. Before applying these models, their calibration coefficients were tested against meteorological datasets from Taastrup and Zaragoza. The net emissivity the coefficients used by FAO (1990) and Allen et al. (1998), and Hansen (2000), respectively for the Brunt (1932) and Brutsaert (1975) net emissivity functions was found to give good results at the Taastrup and Foulum sites (Paper I). However, it was found that model performance was improved when using locally calibrated model coefficients for the cloud cover function (Paper II, Table 1; Paper III, Table 2) at both the Danish and the Spanish sites. The values of the locally calibrated coefficients were in agreement with Jensen et al. (1990). FAO (1990) also suggested using the values found in our studies for a sub-humid climate, while Allen et al. (1998) omitted this recommendation. This suggested that the calibration coefficients for the cloud cover function should be adjusted to match the climate under study. We also found, that it required at least five years of meteorological data to perform a local check of the cloud cover function coefficients. Model prediction were better when used on a daily rather than sub-daily basis. It is likely that incorrect night time cloud cover estimations for the sub-daily calculations caused this. Dividing the datasets by season indicated that during spring and summer the model predictions showed good agreement with observed values, while the performance were poorer during winter. This was probably attributed to a higher albedo values caused by the low sun angles during winter (Paper III, Fig. 4) (Dong et al., 1992; Kessler and Jaeger, 1999; Iziomon and Mayer, 2001).

No difference were observed whether using potential clear sky solar radiation or extraterrestrial radiation in the reduced Stefan-Boltzman models. The performance of the reduced Stefan-Boltzman models resembled the performance of the full Stefan-Boltzman models. As the former are simpler to use and easier to calibrate, they may be used as a substitute for the full Stefan-Boltzman models. This is in agreement with Allen and De Bruin (In prep).

3.1.3 Estimation of latent heat flux

The sensitivity of the calculation of hourly ET_0 to how R_n was established was tested. Calculated values using the R_n estimation method shown as model 5 in Table 1 and the reduced Stefan-Boltzman model shown as model 7 in Table 1 were compared to measured values. Using locally calibrated model coefficients for the Allen et al. (1998) R_n calculation method and the values for surface and atmospheric resistances suggested by ASCE-EWRI (2005) and Allen et al. (2006) gave an MBE 0.01 mm day^{-1}. This indicates that measured values of R_n can be used as a substitute of measured R_n values in ET_0 calculations. Latent heat flux from a winter wheat field was measured using an Eddy Covariance system. An energy balance closure in the range of 55 – 75 % of the eddy covariance system compared to an independent R_n measurements was found. Possible causes are discussed in Paper IV. The lack of closure was corrected following the recommendations of Twine et al. (2000). Daily ET_c for winter wheat was found to be 1.14 times ET_0 (Paper IV, Table 4 and Fig 5). This value is higher than the K_c value of 1.0 commonly used in Denmark. This indicates, that evapotranspiration has been underestimated in previous water balance studies.

3.2 Conclusions

Based on the experimental results, it was concluded that:

- There was a considerable deviation among models to calculate daytime incoming long wave radiation with MBE from -23 to 15 W m^{-2} compared to measured values. Four models were found have a consistent performance with low MBE. Generally model predictions improved when using very clear days with low atmospheric turbidity and when using the models on a daily basis rather than on an hourly basis. No improvement in model performance were found when dividing the data set by whether the upwind surface was mainly urban or covered with vegetation, or season.
- Models based on Stefan-Boltzman theory for calculating net radiation tracked measured values well and showed more consistent performance with lower MBE among sites compared to regression type models. The performance of using the Stefan-Boltzman equations were improved when adjusting the model coefficients by climate regime. Reducing the number of calibration coefficients from 4 to 1 in the Stefan-Boltzman type equations only rendered the model performance marginally.

- The sensitivity of the estimations of hourly values of reference evapotranspiration with respect to how net radiation was estimated was tested. Compared to reference evapotranspiration calculated using measured net radiation, MBE as low as 0.01 mm day^{-1} was found when using estimated values of net radiation.
- When comparing the eddy covariance measurements to an independent measurement of net radiation in the winter wheat field, there was an energy balance closure between 55 and 75 %. Evapotranspiration from the winter wheat was found to exceed reference evapotranspiration by 1.14.

3.3 Perspectives

Focus in ET research has for many years been on estimations of water consumption from individual crops at scales ranging from individual plants to field scale. Research has been abundant and provided large theoretical knowledge about the interrelationships between soil, vegetation and atmosphere. Estimations of ET have been used for irrigation scheduling, field water balance studies, estimations of nutrient leaching as function of crop type and similar.
For applications on larger scale, e.g. a region, having multiple types of surface cover and vegetation, the approaches used presently is often inadequate. Direct estimations of ET_c (e.g. using Penman-Monteith type equations) require estimations of crop specific parameters such as r_a and r_s. Estimations of these crop parameters are challenging, especially during periods of partial ground cover. Validation of these parameters require accumulation of extensive data sets and a standardized computational solution. Also, direct estimations of ET_c require measurements of meteorological variables for each surface cover type of the region in question. Based on these shortcomings for direct estimations of ET_c and $K_{c\ adj}$ the use of crop coefficients is the procedure to follow in the future. However, using crop coefficients will require further developments, such as

- Development of landscape coefficients rather than K_c, thereby incorporating the contributions from different sources in a landscape with varying vegetation cover.
- Solving the uncertainties in K_c values related to water content in the upper soil layers, e.g. the increased evaporation from soil (and vegetation) after an irrigation or precipitation incident.
- Further testing and development of descriptions of crop densities, such as based on leaf area index or a detailed description of the phenological development stage.

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Paper I

Comparison of models for calculating long wave irradiance using long term dataset

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Paper II

Long-term comparisons of net radiation calculation schemes

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Paper III

Comparison of the performance of net radiation calculation schemes

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Paper IV

Sensitivity of reference evapotranspiration calculation and crop coefficients for winter wheat in a temperate, sub-humid climate

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