Energy Fluxes above Three Disparate Surfaces in a Temperate Mesoscale Coastal Catchment

This study is part of the long-term catchment-scale hydrological observatory, HOBE, situated in the Skjern River catchment covering 2500 km² on the western coast of Denmark. To gain a more detailed knowledge of how evapotranspiration is controlled by the local surface and atmospheric processes, eddy-covariance systems have been installed over an agricultural field, over a spruce [Picea abies (L.) H. Karst.] plantation, and on wet grassland. Measurements started in fall 2008, and the first annual series showed large differences in evaporative response among the surfaces. The annual sum was about 500 mm for the wet grassland and spruce plantation, while it was about 300 mm for the irrigated agricultural site. In winter, the actual evapotranspiration rate of the grassland and the forest were much larger than the available energy evaluated from the radiation balance, while at the same time large-scale sensible heat flux directed toward the ground was measured. At the agricultural site, the evapotranspiration rate was controlled by crop development with a leaf area index ≥3 being the threshold where actual evapotranspiration reached the potential rate. At the forest site, transpiration rates were severely limited due to stomatal control, which could be related to soil moisture and vapor pressure deficits. The interception evaporation was very important for total evapotranspiration. At the meadow, summer low flow in the river coincided with drying out of the meadows, which limited evapotranspiration. In late summer, evapotranspiration rates at the meadow and forest sites again increased significantly compared with radiative available energy, and again sensible heat flux directed toward the ground was observed.

In many humid parts of the world, evapotranspiration accounts for the second largest contribution to the water balance, after precipitation. This is also the case in Denmark, where the present study was performed. Still, the parameterization of evapotranspiration from standard climate data is not trivial and models often give significantly different estimates of the potential evapotranspiration (Sommer and Jacobs, 2004; López-Urrea et al., 2006). This uncertainty makes it difficult to close the local hydrologic budget and to describe local and regional variations accurately. Furthermore, a precise accounting for the evapotranspiration term is crucial for our understanding of the hydrologic functioning of different ecosystem types under future climatic conditions.

Land use has long been known to have a profound effect on the catchment-scale water balance. In a comprehensive review of the effect of vegetation changes on water yield, Bosch and Hewlett (1982) showed that a reduction in vegetation cover increases catchment yield and decreases evapotranspiration. Likewise, the type of vegetation plays a role. A reduction in the surface cover of coniferous forests causes a large decrease in evapotranspiration, while a reduction in shrub or grassland cover promotes a smaller decrease. This calls for an explicit treatment of different land-use classes when working with physically based hydrologic models, and a lot of work has been done incorporating anything from simple empirical formulas to more advanced soil—vegetation—atmosphere transfer models (Dunn and Mackay, 1995; Twine et al., 2004; Mao and Cherkauer, 2008; Boegh et al., 2009).

Evapotranspiration can be partitioned into the transpiration of water through photosynthetically active leaves and direct evaporation from the surface in the form of soil evaporation and interception evaporation of rain caught by the canopy. The relative sizes of transpiration and evaporation depend on the type and coverage of the vegetation. Well-managed, short agricultural crops typically transpire at higher rates than woodlands due to the optimization for maximum yield in agricultural production, while woodlands...
typically have higher interception evaporation due to the much larger canopy structure (Oke, 1987).

On a weekly or monthly basis, the major controlling parameter for the transpiration rate of crops and deciduous vegetation is the development of the canopy, often expressed through the leaf area index (LAI) (Kristensen, 1974; Rosset et al., 1997; Burba and Verma, 2001; Testi et al., 2004). Assuming no significant water stress, transpiration will scale linearly with LAI from zero in the wintertime to a maximum rate close to the potential evapotranspiration when LAI increases to $\geq 3$ (Kristensen, 1974; Allen et al., 1998). Surface evaporation is very small in comparison, scaling from close to zero in winter when the soil is bare (spring crops) or the canopy structure is very small (winter crops) to about 10 to 15% of the total evapotranspiration during peak growth (Sadras et al., 1991).

For forests, the pattern is more complex. Although transpiration is the most important process in long-term balances, direct evaporation can account for about a third of the annual evapotranspiration (Herbst et al., 1999). In deciduous forest, leaf development remains an important controlling parameter for transpiration as well as canopy structure and its influence on interception, throughfall, and stem flow (Herbst et al., 2008). In contrast, evergreen coniferous forests have a stable, high LAI year-round and the major controlling parameters in the absence of water stress are radiation and air humidity for transpiration and the frequency of precipitation events for direct evaporation (Oke, 1987).

The study was part of a series on surface energy fluxes and evapotranspiration from the HOBE hydrological observatory. We compared measurements from three sites in the Skjern River catchment with the simple Makkink potential evapotranspiration model (Makkink, 1957) that is extensively used in Denmark and has been adapted to the local climatic conditions. We also compared them with the widely used and more advanced Penman–Monteith model (Monteith, 1965), which has been verified in a number of studies in different climatic regions (Allen et al., 1998; Gavin and Agnew, 2004; López-Urrea et al., 2006). By pointing out weaknesses and strengths in the modeled outputs at different times of the year, we aimed to improve our understanding of the processes and main controls on evapotranspiration. Using eddy-covariance evapotranspiration measurements from three disparate surfaces within a small catchment having a uniform climate, we show that land-use classifications are indispensable for regional estimates of evapotranspiration. Future work within the HOBE observatory will use these measurements to compare up-scaling techniques with remote sensing data and to calibrate and develop a catchment-wide integrated hydrologic model based on the MIKE SHE code (Refsgaard and Storm, 1995).

Materials and Methods
Site Description and Instrumentation
The Catchment

Figure 1 shows the catchment location and the three field sites. The catchment covers about 2500 km$^2$ and extends 65 km inland. The landscape is very flat throughout the catchment, rising gradually only to about 100 m inland. The geology is dominated by loose glacial and meltwater deposits, with soils being comprised mostly of coarse sand. The climate can be characterized as maritime, with a mean annual temperature of 8°C and an annual precipitation of 850 mm. The weather is dominated by the prevailing westerlies from the Atlantic Ocean and the accompanying wandering cyclones. This creates a slight orographic effect, with about 10% more precipitation in the eastern part of the catchment than at the coastline (Frich et al., 1997).
Gludsted Plantation (Forest Site, 16% of Catchment)

Gludsted Plantation is an approximately 100-yr-old and 3500-ha spruce plantation consisting primarily of 15- to 20-m-tall Norway spruce of varying stand age. Single lines of approximately 25-m-high grand fir \([\text{Abies grandis} \ (\text{Douglas ex D. Don}) \text{ Lindl.}]\) are repeated at intervals of about 150 m. The parent material of the soil is sandy deposits left by the last glaciation. The texture is coarse sand with some larger rocks mixed in and a clay content of <1%. A 5-cm-thick organic horizon with a C/N ratio of 27 overlies the mineral soil (Gundersen et al., 2009). The forest floor vegetation is sparse and consists mainly of mosses. An instrument mast of 37.5-m height was erected to ensure proper separation between the eddy-covariance instruments and the canopy top. The mast was located near the center of the plantation to ensure that the entire fetch was covered by forest and to avoid potential edge effects resulting in untypically high evapotranspiration rates (Herbst et al., 2007).

The eddy-covariance (EC) instrumentation consisted of a sonic anemometer (R3–50, Gill Instruments Ltd., Lymington, UK) mounted at the top of the mast along an open path \(\text{CO}_2/\text{H}_2\text{O}\) analyzer (LI-7500, LI-COR, Lincoln, NE). Data were measured at a frequency of 10 Hz and stored on a datalogger (CR3000, Campbell Scientific, Logan, UT). Ancillary data include a four-component radiation sensor (NR01, Hukseflux Thermal Sensors B.V., Delft, the Netherlands) mounted 30 m above the forest floor, two temperature and relative humidity sensors (HMP 45C, Vaisala Oyj, Helsinki, Finland) at 30 and 15 m, respectively, two soil heat flux plates buried 5 cm below the surface (HFP01, Hukseflux Thermal Sensors), and a tipping bucket rain gauge mounted on top of the instrument container near the center of a small clearing (Rimco 7499, McVan Instruments, Mulgrave, VIC, Australia). All ancillary data except precipitation were measured once every minute, and 10-min averages were stored on the datalogger.

Voulund Farm (Agricultural Site, 68% of Catchment)

Voulund Farm is an 800-ha farm producing mainly pigs (\(\text{Sus scrofa}\)) and feed crops (barley [\(\text{Hordeum vulgare} \text{ L.}\)] and maize [\(\text{Zea mays} \text{ L.}\)]). The fields surrounding the instrument mast were planted with winter or summer barley in 2008 to 2009 (see Fig. 2 for more details). All fields were irrigated intensively from mid-April 2009 to the end of the growing season for winter or summer barley. The winter barley fields were harvested on 21 July and the spring barley fields during the first week of August. Soon after harvest, the fields were replanted with the catch crop \(\text{Raphanus sativus} \text{ L.}\), which remained until the end of the data period. The soil was a Spodosol consisting of coarse sand below a 0.25-m-deep organic topsoil. Soil porosities in the upper 1 m of the profile ranged between 0.35 and 0.40. The available soil water \([\text{pF} 2.0–4.2, \text{suction pF} = \log_{10}(\text{suction in centimeters of water})]\) was 19% (v/v) in the upper 20 cm of the plow layer and only 6% (v/v) in the remaining part of the root zone, necessitating frequent irrigation to maintain crop growth during most growing seasons (Schelde et al., 2011). The groundwater level was located well below the root zone, at a depth
of approximately 5 m. Irrigation took place on 7 d between 18 Apr. and 18 June 2009, and 22 to 25 mm was applied on each occasion.

The EC mast was placed in a 300- by 400-m barley field surrounded by agricultural fields with a similar crop height. During the middle of the growing season (May), 50% of the flux originated from a distance of <570 m, on average (Gash, 1986) and the distance of maximum contribution was 150 m during the daytime and 200 m at night (Schuepp et al., 1990). A 15- by 100-m plot near the center of the barley field was reserved for instrumentation and vegetated with short grass and weeds. The location of the EC mast in this instrumentation plot ensured that the fetch in all main wind directions (north, west, and southwest) was not influenced by the slightly different surface of the instrument plot.

The EC setup consisted of a sonic anemometer (R3–50, Gill Instruments Ltd.) mounted on top of the mast (12 m above ground level) with gas inlets leading to a closed path CO$_2$/H$_2$O analyzer (LI-7000, LI-COR). Data were measured at 10 Hz and stored on a datalogger (CR3000, Campbell Scientific). The instrumentation for the meteorologic data was similar to that at Gludsted Plantation, with the radiation, temperature, and humidity being measured at 4 m above the ground. Rainfall was measured at the southern end of the instrument plot at 1.5 m above the surface. Crop development was monitored by measuring the LAI every 3 wk on average (LAI-2000, LI-COR).

Skjern Meadows (Wet Grassland, 7% of Catchment)

During 1968, the course of the lower part of the Skjern River was straightened and pumps and embankments were built, draining 4000 ha of former peatland. In 2000, the river was restored to its former meandering course and pumping activity was greatly reduced, converting some 2200 ha to wetland status (Nielsen and Schierup, 2007). The upper part of the restored wetland, mainly covered by fens and lakes, is characterized by Histosols, whereas the floodplain around the study site, as well as the marshes close to the mouth of the Skjern River, are dominated by Fluvisols according to the FAO soil classification system. The meadows are grazed by cattle each summer to keep the vegetation short, and the grass was cut once during the experiment period, on 29 June 2009.

The instrument mast was placed in a 3-m-high hedgerow with a total sensor height of 7 m above the wetland surface. The fetch of the mast was dominated by “wet grassland,” but also the free water surface of the Skjern River was visible as well as the hedgerow and a gravel access road. Relative to the agricultural and forest sites, the meadow was considerably more heterogeneous. Considering the relatively short distance to the river from the EC mast location, the possible contribution of the hedgerow disturbed the wind field, giving systematically positive values for the vertical wind component ($\omega$) in the uncorrected data. The data were gap filled using a standardized method adopted by CarboEurope and FLUXNET (Moffat et al., 2007) (gaia.agaria.unitsit.it/database/eddyproc/, verified 18 Nov. 2010). The data records to be gap filled were selected using the quality classes proposed by Foken et al. (2004), which were calculated by the AltEddy software for each half-hour flux data record. Classes 1 to 3 were considered valid data and Classes 4 to 9 were gap filled. For the two open path sensors at Gludsted Plantation and Skjern Meadows, gap-filled data were used for 40 and 37% of the time, respectively, while for the closed path sensor at Voulund Farm, gap filling was used for 24% of the time. The amount of gaps is in line with other EC studies (Moffat

Data Processing

Raw EC data were processed using the AltEddy software package version 3.5 (Alterra, University of Wageningen, Wageningen, the Netherlands). Fluxes were calculated as 30-min averages, and the two-dimensional planar fit method was used to correct errors induced by the difference between anemometer tilt and the main streamline of the wind coordinate system (Wilczak et al., 2001). This was especially important at Skjern Meadows, where the presence of the hedgerow disturbed the wind field, giving systematically positive values for the vertical wind component ($\omega$) in the uncorrected data.
et al., 2007) and is the result of an inherent limitation of the EC method. The gaps originated primarily from periods with very low turbulence and, secondarily for the open path sensors, from periods with rain. Low turbulence mostly prevailed during the nighttime when the evaporative flux was expected to be small, and hence the impact of gap filling on the seasonal fluxes would be low. Similarly, we would not expect the ET to be high during rainfall, but interception evaporation immediately following rain can be very substantial. The open path sensors were mounted at an angle that ensured rain would drain off the sensor head. In our experience, the open path sensors are able to measure correctly within a half-hour time step after it has stopped raining, enabling the measurement of most of the evaporative flux directly related to a rain event.

Open Path vs. Closed Path Gas Analyzers
While the open path (LI-7500) and closed path (LI-7000) sensors represent the state of the art in fast-response gas analysis, the different designs offer some unique advantages and disadvantages. When placed close to the sonic anemometer, the open path sensor gives a good direct measurement of the gas concentration in the parcel of air moving through the anemometer without any significant alterations to the air flow and with only very small lag times between the gas concentration and air movement samples. The disadvantage of the open path sensor type is poor performance during rain. The closed path sensor draws in air through an inlet placed near the anemometer and through a tube to the sensor placed inside an instrument hut. The obvious advantage of this design is the ability to measure during rain, as is evident from the better quality class distribution at Voulund compared with Skjern Meadows and Gludsted Plantation. The disadvantage, especially concerning water vapor fluxes, is the tube attenuation from adsorption and desorption of water vapor to the inner walls of the tube, acting as a low-pass filter effectively reducing the measured flux (e.g., Ibrom et al., 2007; Haslwanter et al., 2008). This high-frequency loss is not yet fully accounted for by the current version of AltEddy. The performance of the closed path sensor at Voulund was assessed by comparison to root zone integrating time domain reflectometry probes (Schelde et al., 2011), with the conclusion that the two methods agreed well with each other during the growing season. We are therefore confident to compare data from open and closed path systems, but it must be kept in mind that especially the apparent zero-ET during the winter at Voulund Farm must be treated with some caution.

Basic Meteorologic Parameters and Potential Evapotranspiration
Table 1 shows the monthly means of basic meteorologic parameters important for the ET rate. Overall the sites show very similar values, underlining the fact that the present data set presents a good opportunity to examine the control of the local surface conditions on the ET rate.

Incoming shortwave radiation was almost identical for the three sites, although there was a tendency for Skjern Meadow to receive slightly more radiation due to the smaller orographic effect and less cloud cover close to the sea. The available radiation was quite different, however. Using the incoming and outgoing shortwave radiation measurements, the calculated albedo for the forest was just 0.08, while the meadow and agricultural sites had albedos in the order of 0.18. Proximity to the sea also caused the meadow site to experience slightly higher
temperatures in winter, but the same seemed to hold true even during the summer. The agricultural site was the coldest, being slightly cooler than the forest for most months. The vapor pressure deficit seemed to be correlated to temperature to some extent. During winter, the deficit was larger at the warmer meadow site than at the agricultural site, but during summer, the relationship was reversed and the cooler fields experienced higher deficits than the meadow. The forest showed the largest variation, having the smallest deficits of all sites during winter and the largest deficits during summer. All sites experienced quite similar numbers of rainy days, although as expected the variation between the sites was largest in the high summer when rain mainly arrived in the form of convective showers.

**Makkink Potential Evapotranspiration**

Using the meteorologic parameters presented above, it is possible to estimate the potential evapotranspiration rate for each surface, afforded by the local available energy and atmospheric conditions without directly taking any surface controls other than albedo into account. This provides a reference to which the measured actual evapotranspiration (Ea) can be compared, allowing surface controls to be explored by looking at the deviations between the two. For the present study, the Makkink formula (Makkink, 1957) was applied for calculating the potential ET rate. The Makkink formula has been recommended for calculating potential evapotranspiration (Ep) in Denmark (Detlefsen and Plauborg, 2001) and is the formula currently applied by the Danish Meteorological Institute (Scharling, 2001). Its simplest formulation is

\[
\text{Ep} = \alpha \frac{\Delta}{\Delta + R_G}
\]  

where \(\Delta\) is the slope of the saturation vapor pressure vs. temperature curve, \(\gamma\) is the psychrometric constant, \(R_G\) is the global radiation, and \(\alpha\) is an empirical constant. It is a very simple formula, calculating the potential rate of evapotranspiration only from the available energy in terms of the global radiation and from the “efficiency” of water evaporation at a given temperature. The reliance on only a few meteorologic parameters makes it well suited for exploring surface controls. For the best precision, the empirical parameter \(\alpha\) can be varied by the time of year or daily mean temperatures (Mikkelsen and Olesen, 1991), but for this study, a constant value of \(\alpha = 0.65\) was used as a reference for the measured ET. The annual ET rates based on Makkink were 397, 503, and 457 mm yr\(^{-1}\) for Voulund Farm, Gludsted Plantation, and Skjern Meadow, respectively. From these data, we should expect the lowest actual evapotranspiration from the agricultural site (high albedo and low temperature), the highest value in the forest (low albedo and high temperature), and an intermediate value for the meadow site (high albedo and high temperature).

**Penman–Monteith Potential Evapotranspiration**

While this study focused on the relation between actual evapotranspiration and the Makkink estimated potential rate, the lack of process description in the Makkink formula makes it unsuitable for modeling and prediction of future ET rates. To address this, the widely used Penman–Monteith equation (Monteith, 1965) was included:

\[
\lambda \text{ET} = \frac{\Delta (R_n - G) + \rho_a c_p (\text{VPD}/\tau_a)}{\Delta + \gamma [1 + (\tau_a/\tau_s)]}
\]  

where \(R_n\) is the net radiation, \(G\) is the ground heat flux, \(\rho_a\) is the mean air density at constant pressure, \(c_p\) is the specific heat of air, VPD is the vapor pressure deficit, \(\tau_a\) is the atmospheric resistance, and \(\tau_s\) is the bulk surface resistance.

The equation defines a series of resistances to water vapor transport that are combined to form the bulk surface resistance and the aerodynamic resistance. The surface may control the surface resistance via stomatal control (transpiration), soil moisture, and the amount of bare soil visible (soil evaporation) or surface wetness status (interception evaporation). Surface roughness (canopy structure and size) controls the atmospheric resistance. With three-dimensional ultrasonic anemometer data, the atmospheric resistance may be calculated directly as the vertical wind speed (\(U\)) divided by the friction velocity (\(U^*\)) squared (for a discussion of this method vs. the usual wind profile derivation, see Herbst et al., 2008), leaving only the surface resistance as a variable. By assuming a fixed surface resistance of 100 s m\(^{-1}\), we calculated the Penman–Monteith potential evapotranspiration rate for each site. A surface resistance value of 100 s m\(^{-1}\) corresponds well to the observed daytime value for the three sites in the absence of significant soil moisture stress and with relative humidity not dropping below \(\sim 75\%\). This allows the calculated Penman–Monteith ET rate to be used as an “optimum conditions” benchmark, against which to analyze surface controls that make the actual evapotranspiration rate deviate from this optimum.

**Results**

**Annual Course of Water Fluxes**

Figure 3 presents the cumulative measured ET from the three sites. During the winter months from December to February, the forest and the meadow had a relatively steady ET rate of 0.42 and 0.26 mm yr\(^{-1}\), respectively, while the agricultural site showed almost no water vapor flux at all. In March and April, the ET rate for all sites increased markedly with the greening of the surface and the increase in available energy. The evergreen forest responded immediately to the increase in radiation and temperature, while the onset of significant ET at the meadow and agricultural sites was delayed by about 1.5 and 3 wk, respectively.

April 2009 was unusually dry with only about 15 mm of rain, and from late April the forest ET rate dropped to about 2.1 mm d\(^{-1}\)
on average—well below the average summer ET rates of the wet meadow and irrigated agricultural site of 2.5 and 2.4 mm d$^{-1}$, respectively. In July, the water vapor flux dropped off rapidly at the agricultural site in response to crop senescence and the halt of irrigation, and following the harvest of the spring crops in early August, the ET was much reduced. The forest exhibited a further...
decrease in the rate of ET in early July following a 2-wk dry spell but recovered later in July and August as the frequency and magnitude of rain events increased again. The meadow site also showed a slight decrease in the ET rate in July and early August as the wet grassland began to dry in response to summer low flow in the Skjern River but recovered strongly in mid-August following heavy rain events and rewetting of the area. Skjern Meadows and Gludsted Plantation showed a gradual decrease in the ET rate in September and October in response to the lower available energy.

The total sum of the measured Ea during the study period was 483 mm for Skjern Meadow, 460 mm for Gludsted Plantation, and 266 mm for Voulund Farm. Compared with the potential annual rates of 457 mm yr−1 for Skjern Meadow, 503 mm yr−1 for Gludsted Plantation, and 397 mm yr−1 for Voulund Farm, these results represent some rather large differences; observations were lower than expected at Voulund and Gludsted and higher than expected at Skjern Meadows.

Winter Evapotranspiration
Observed winter ET at the meadow and the forest exceeded the available energy from the radiation balance. The energy necessary to evaporate water was possibly provided by a downward flux of sensible heat. Figures 3b and 3c show a negative sensible heat flux (directed toward the ground) for the period from December through March when the Ea was larger than the Makkink Ep rate. Table 2 presents average daily values of Ea, Makkink Ep, and the measured sensible heat flux for December to February.

For the forest and meadow sites, the negative sensible heat flux was large enough to cover the difference between Ea and Ep. There were notable differences in sensible heat flux rates depending on whether the surface was dry or wet. During wet days, the sensible heat flux was several times larger than on dry days for both sites. In absolute values, the forest site received more than twice the amount of energy from the sensible heat flux than did the meadow site. This was also reflected in the ET rate during wet days when the forest evaporated at twice the rate of the meadow site. During dry days, the rates from the two sites were similar.

Spring and Summer Evapotranspiration
Gludsted Plantation
From early April, the ET rate changed from exceeding the Makkink estimate of Ep to falling behind the potential rate until the beginning of July (Fig. 4). Evident already from the end of March, Ea was inferior to the potential rate (Ea − Ep < 0) unless it had been raining within the previous 48 h. Throughout April,

<table>
<thead>
<tr>
<th>Precipitation condition</th>
<th>Gludsted Plantation</th>
<th>Makkink Ep</th>
<th>H</th>
<th>Skjern Meadows</th>
<th>Makkink Ep</th>
<th>H</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ea (mm)</td>
<td>Makkink Ep (mm)</td>
<td>H (MJ m−2 d−1)</td>
<td></td>
<td>Ea (mm)</td>
<td>Makkink Ep (mm)</td>
<td>H (MJ m−2 d−1)</td>
</tr>
<tr>
<td>Wet days</td>
<td>0.49</td>
<td>0.17</td>
<td>−2.2</td>
<td>0.24</td>
<td>0.16</td>
<td>−1.2</td>
</tr>
<tr>
<td>Dry days</td>
<td>0.24</td>
<td>0.16</td>
<td>−0.86</td>
<td>0.28</td>
<td>0.16</td>
<td>−0.34</td>
</tr>
</tbody>
</table>

Fig. 4. Gludsted Plantation: difference between actual (Ea) and Makkink potential evapotranspiration (Ep) (evapotranspiration deficit) compared with daily precipitation and soil moisture content.
May, and June, this pattern continued and the deficit between the observed evapotranspiration rate and Makkink potential rate increased. This suggests an increasing environmental control of the dry-weather transpiration rate, coinciding with rising VPD and a dry spell in late June and early July. As the soil gradually dried out during the dry spell, there was also a gradual increase in the evapotranspiration deficit. From mid-June, the soil moisture recovered and the precipitation frequency remained high until the end of the measurement period, resulting in an Ea rate equal to or even slightly higher than the Makkink potential rate.

**Skjern Meadows**
The Ea cumulative sum at the meadow site remained much closer to the Makkink Ep cumulative sum during the entire spring and summer period than at any of the other sites. This can be explained by the permanent nature of the vegetation in combination with a relatively stable water supply from the Skjern River. From Fig. 5 it is evident, however, that when the water level in the Skjern River was above 600-mm local datum, corresponding to ~30 cm below the surface of the lower parts of the meadow, Ea was predominantly above the potential Makkink rate. When the water level dropped below 600-mm local datum, the relationship was reversed. The few days of Ea exceeding Ep during the period of low water level could all be attributed to local rain events temporarily rewetting the surface. As the water level rose above the 600-mm level again in August, we observed the highest evapotranspiration rates of the entire study period for all sites, peaking at 5.9 mm d\(^{-1}\) on 13 August. The elevated rate lasted for a couple of days as the meadow rewetted, after which a more typical rate of 2.5 to 3 mm d\(^{-1}\) was attained.

**Voulund Farm**
The course in the actual evapotranspiration at Voulund Farm was tightly coupled with crop development (Fig. 3d and 6). During March and April, Ea was consistently below the Makkink potential...
rate until the winter barley on the instrumented field reached an LAI of 3 in the beginning of May. During the first half of May, Ea agreed with Ep; from mid-May when the surrounding fields with spring barley reached LAI = 3, Ea exceeded the potential rate. Excepting two rainy periods, this held true until the last week of June when irrigation was halted and senescence of the winter barley accelerated.

**Penman–Monteith Potential Evapotranspiration**

The Penman–Monteith estimate of potential evapotranspiration largely agreed with the Makkink Ep estimate (Fig. 3b-d). During the low radiation conditions in the winter months, the Penman–Monteith equation underestimated vapor fluxes in the forest to the same extent as the Makkink equation. At the meadow site, the Penman–Monteith prediction was more in agreement with observations due to the inclusion of the VPD term in the equation, capturing a few situations when relatively dry air afforded high evapotranspiration rates. During spring and summer, the estimated Penman–Monteith evapotranspiration rates were considerably higher than the Makkink rates for the meadow and agricultural sites, but for the forest site they were almost the same. This suggests a different surface resistance at the forest than the two other sites. In general, the chosen surface resistance of 100 s m\(^{-1}\) caused the predicted evapotranspiration at the annual time scale to surpass the observed evapotranspiration at all sites.

**Discussion**

**Wintertime Fluxes at the Forest and Meadow Site**

There is an apparent discrepancy between the available energy estimated from the radiation balance and the actual measured evapotranspiration at Gludsted Plantation and Skjern Meadow. This may be attributed to a process already observed and discussed by Shuttleworth and Calder (1979). Using data from two coniferous forests in the UK, they demonstrated large-scale heat advection toward the ground and accompanying high evaporation rates. They focused mostly on rainy days and argued that the heat released by condensation in the lower atmosphere increased heat advection, while at the same time the amount of water readily available for evaporation was large due to interception by the canopy. Some of the same trends can be seen in the data in the present study. For the meadow and forest sites, the independently measured sensible heat flux toward the ground was about 2.5 times larger on wet days than on dry days. Only at the forest site was the evapotranspiration rate higher on wet days than on dry days. At the meadow, rates were almost the same. This indicates that interception evaporation was only important in the forest, while the meadow experienced a stable evapotranspiration rate regardless of whether there was intercepted water in the grass canopy or not. The most probable explanation for the lacking interception effect at the meadow site is that a rather large part of the meadow surface already experienced standing water during winter. The 0.25 mm d\(^{-1}\) evapotranspiration rate observed for the meadow site was probably close to the potential rate from a free water surface and thus primarily limited by the atmospheric resistance to water vapor transport. The forest had a much larger roughness and corresponding atmospheric exchange, which may explain why the average evapotranspiration rates from the wet forest canopy were twice the rate observed at the meadow site. Removing measurements with low turbulence (\(U^* < 0.1\)), the estimated average winter atmospheric resistance (\(r_a = U/U^*2\)) was 54 s m\(^{-1}\) for the meadow and 37 s m\(^{-1}\) for the forest, while for high-turbulence (\(U^* > 0.25\)) conditions \(r_a\) was 31 s m\(^{-1}\) for the meadow and only 17 s m\(^{-1}\) for the forest. Even during dry winter days when the most recent precipitation event was at least 24 h away, there was considerable evapotranspiration in the forest. We cannot decide to what extent this was attributable to lingering interception evaporation, transpiration, or soil evaporation, but at least some indication that transpiration is part of the answer even in the midst of winter can be seen from the co-measured CO\(_2\) fluxes (Herbst et al., 2011). On several occasions during the winter there was a net C uptake, indicating photosynthesis and by extension transpiration.

Elaborating further on the advection of sensible heat toward the surface, it is important to note that its size was considerably greater than the available energy deficit associated with observed ET rates at all times, excepting dry days at the meadow site. This implies that the ET rate was not primarily energy limited during wintertime but rather limited by surface and aerodynamic resistances. The negative sensible heat flux did not occur in response to evaporative cooling of the surface, but as a result of large-scale flow of relatively warm maritime air from the North Sea over the relatively cold land surface during winter. This effect has also been observed on a pasture in the UK by Harding et al. (2000) and Harding and Lloyd (2008), who concluded this to be a general feature of the UK climate. Indeed this might be a feature of most maritime climates, and as such it is something that is much in need of attention from the modeling community. It must be kept in mind that while the sensible heat flux may mitigate the energy restrictions to ET afforded by the low radiative energy, it is not directly coupled to the water vapor flux and care must therefore be taken if using the heat flux directly to model ET.

**Spring and Summer Fluxes**

**Sensible Heat Flux**

In general, the measured sensible heat fluxes agreed with the radiative available energy in spring and summer. The sensible heat flux changed sign in mid-March and became a net energy drain at all surfaces, transporting away excess energy from absorbed radiation not used for ET. In late summer, the sign of the sensible heat flux changed again. This time, however, the timing of the change was not the same for all surfaces. In the forest, the first change was in late August during a week with intense rain, and while the rate became slightly positive again for a short period in September, the general trend was toward negative values. This coincided with the period when the forest ET rate began to exceed the radiative available energy. The same was observed at the meadow where the sensible heat flux sign changed in mid-August following rewetting of the meadow and the return to above Makkink Ep rates. This
is comparable with Harding and Lloyd (2008), who saw a similar annual course of sensible heat flux with negative sign from early October to late March at a wet grassland site in Somerset Levels, UK. For the agricultural site, the sign did not change until the beginning of October, and even then, the flux was very low. This agrees with the finding that ET was very low at the agricultural site after harvest, and as such there was no significant evaporative cooling to act as a driver of heat flux toward the ground. Thus, in contrast to the winter situation when the negative sensible heat flux was mainly a result of large-scale movement of warm air over land, the late summer may show sensible heat flux driven by evaporative cooling of the surface. Evidence for this can be found by comparing the accumulated negative sensible heat flux with the ET surplus, i.e., the accumulated Ep subtracted from the equivalent accumulated Ea. For the meadow site from 12 August to 14 October, the Ea – Ep sum was 42 mm while the H sum was −32 mm, and for the forest site from 25 August to 14 October, the Ea – Ep sum was 20.6 mm while the H sum was −18 mm. These values are much more in line with the energy demand of the ET than during wintertime. It can also be argued that, in the absence of evaporative cooling, the sensible heat flux should not turn negative until the surface became cooler than the warm maritime air following the westerlies from the North Sea; this temperature transition took place in late autumn.

Controlling Parameters
This study examined three very different surfaces—two natural surfaces where, albeit in different ways, the supply of water was the primary limiting factor and a cultivated surface where crop development and senescence were the most important controls.

Even though the forest received the most energy due to the low albedo, this was not reflected in the observed ET rate, which strongly relied on interception evaporation in the forest. It only took a couple of days without precipitation for the ET rate to drop substantially below the Makkink potential rate. This can be interpreted as a low transpiration capacity of the coniferous canopy, which corresponds with the literature. Komatsu (2005) reviewed numerous published studies from forests and found average dry-canopy Priestley–Taylor coefficients of 0.54 ± 0.25 for temperate coniferous forest and 0.82 ± 0.16 for temperate broad-leaved forest. Gavin and Agnew (2004) found coefficients for temperate wet grassland ranging from 0.8 when the surface was dry to 1.25 when the area was inundated. Consequently, despite having 10% extra available energy in the forest, the typical summer ET rate was about 20% lower than at the meadow and agricultural sites. We observed that the dry-canopy difference between Ea and Makkink Ep became larger as summer went on. This means that the forest became less efficient in using the available energy as the radiation levels increased, implying a very strong stomatal control. This can partly be explained as a response to high VPD because days with large differences between Ea and Ep correlated well with large VPDs (data not shown), but soil moisture also seemed to have an effect. During the dry spell around the beginning of July, the Ea rates were unusually low—lower than during previous dry periods with similar high VPD values. This suggests that soil moisture stress increased the stomatal control on the water vapor flux beyond the usual VPD response.

At the agricultural site, intense irrigation ensured that water stress was never an issue during the growing season. As expected, this caused crop development to be the primary control on ET. We saw an interesting apparent shift in ET rates from being below Makkink potential rates to being above potential rates at an LAI of 3. This is in good agreement with Kristensen (1974) who found LAI = 3 to be the threshold above which there was no additional increase in ET. It is also the commonly accepted threshold at which the crop coefficient should attain its maximum value if a crop coefficient approach is used to model Ep (Allen et al., 1998). The large control of crop development on the water vapor flux suggests that transpiration was by far the most important process at Voulund Farm. This is also supported by the fact that during the peak growth season, rainy days yielded ET rates similar to or even slightly lower than Ep. When considering the available radiative energy, the loss of transpiration by reduced light levels on rainy days was apparently greater than the interception evaporation gained. This response was functionally different from that of the forest, where the interception evaporation always resulted in ET rates larger than Ep. This is an important difference to keep in mind when working with model estimations, and it underlines the very different response of the surface types to the same weather events.

At Skjern Meadow, the interesting control was the apparent river water level influence on the Ea. Standing water on the meadow had disappeared already by mid-April, and as such, the water level in the river can be viewed as a proxy for the groundwater table within the meadow. To help interpret what happened at the 600-mm water level threshold, it is useful to look at the fluxes of CH₄ and CO₂ that were measured at the same site (see Herbst et al., 2011). The CH₄ flux was also strongly influenced by the river water level, with the 600-mm level functioning almost as an on–off switch. This introduction of aerobic conditions in the soil would indicate that significant drying out of the meadow takes place when the river stage drops below this level, and beginning water stress could explain the observed decrease in ET. For this to be true, we would expect transpiration to be a significant part of the total ET, but looking at the CO₂ flux reveals a mixed picture. The strongest CO₂ uptake of about 15 g C m⁻² d⁻¹ was seen in May when the water level was high, dropping slightly to about 12 g C m⁻² d⁻¹ in June when the water level dropped below 600 mm. This is in line with the ET observations, but on 29 June the grass was cut, turning the meadow into a net C source for about 3 wk until the vegetation recovered. Grass cutting was barely visible at all in the water vapor flux observations, suggesting that transpiration and
vegetation controls were not very important at the meadow, even during drying out in summer. But from where did the measured flux come, if not from the vegetation? A planned, more detailed study including water table measurements within the meadow, soil moisture observations, and a thorough footprint analysis may help shed light on this question.

Performance of the Penman–Monteith Formula, and Implications for Future Studies
The predictions of the Penman–Monteith formula with a fixed surface resistance performed similar to the predictions of the Makkink formula in terms of capturing the weekly and seasonal fluctuations in ET, although the magnitude of the Penman–Monteith prediction naturally changes depending on the applied surface resistance. This leads to the perhaps surprising conclusion that when tuned to Danish conditions (α = 0.65), the annual ET estimate from the very simple Makkink formula was as close to the measured values as the more complicated Penman–Monteith. This should not be seen as a general endorsement of the Makkink formula, however. Had detailed knowledge of the variation in surface resistance been available, the Penman–Monteith formula would certainly have given a better estimate, especially so with respect to process description and capturing the short-term and seasonal variations in ET. In future studies, we will aim at providing a parameterization of the surface resistance of the Penman–Monteith equation at short time scales—hourly to daily values. One way of approaching this is using the observed EC flux data to invert the Penman–Monteith formula and directly calculate the surface resistance. This will not only provide better surface-specific ET models for the HOBE catchment hydrologic model but also gain further insights into surface control of the ET process.

A special challenge will be to deal with the heat advection during winter. It is certainly possible to parameterize the Penman–Monteith model to reproduce the measured wintertime ET, even in the forest, by applying very low surface resistances; however, this may not be an accurate physical description of the ET process. We may need to add large-scale, laterally advected heat to the available energy term in the Penman–Monteith equation to get a true measure of the surface resistance, but how to accomplish this remains an important challenge.

Performance of the Radiation Balance “Available Energy” as a Modeling Concept
Most of the widely used simple ET models use radiative available energy in some form to drive the ET (e.g., Penman, 1948; Makkink, 1957; Priestley and Taylor, 1972). Based on the data presented here, estimating or even just constraining the ET to the available energy derived from radiation measurements seems to produce inaccurate results. Thirty years after the fundamental critique of Shuttleworth and Calder (1979) of using the radiation balance based Priestley–Taylor model in forests seeing significant rainfall, there is a pressing need to recognize the importance of the extra energy afforded by large-scale advection of sensible heat. This goes beyond forests. Looking at the present data, the predicted ET from both the forest and the meadow may be seriously underestimated during winter and late summer if heat advection is not included in the applied model.

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